

A case study of the Rusey Breccia:

Thrust zone brecciation, reorganisation and mineralisation

by

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A thesis for the fulfillment of the requirements of the University of London for the
Degree of Doctor of Philosophy.

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Abstract

The Rusey Breccia, 80m long and 2-3m thick, occurs in the footwall of the Rusey Thrust, beneath the launch point of a series of thrust ramps. The thrust is a major Variscan structure exposed in the Boscastle shales on the north Cornish coast.

The breccia forms a series of size graded layers, fining towards each layer boundary. The aligned clasts are surrounded in three dimensions by radial euhedral quartz. No fault gouge is present. The layering and re-brecciated clasts indicate repeated episodes of organisation and mineralisation. Clast organisation occurred in a dynamic environment resulting from episodic fluid injection associated with fault slip events, causing fluidisation of the clasts along the injection pathways. Mineralisation textures indicate relatively slow growth in a stable saturated medium capable of maintaining the clasts in suspension. Fluidisation is the dominant mechanism in the organisation of the breccia. Scale dependent differences in the behaviour of fluidised suspensions create the mineralising medium by forming a colloidal suspension of the fault gouge.

The Rusey Breccia is compared with a similar breccia outcropping in Kimmeridge Bay, Dorset, formed in a thrust system developed at shallow depths in a carbonate system.

The literature on Variscan and Cornubian geology is reviewed and the structural geology of the Rusey Headland is described and explained as part of the Variscan orogenic system. Photoelastic modelling of the Rusey Thrust geometry is used to support the model proposed for the formation of the Rusey Breccia. The processes involved in the reorganisation and mineralisation of the breccia are described and reviewed. The internal structure and matrix morphology of the Rusey and Kimmeridge breccias is described and a statistical analysis of the Rusey data presented. The field, lab and analytical evidence presented in the thesis is synthesized to produce an evolutionary model for the Rusey Breccia.

CONTENTS

Abstract	2
Contents	3
Figure list	8
Plate list	18
Thesis Introduction	21
Chapter One	
A literature review of the geology of the Variscan Orogen and the Cornubian Peninsula	24
1.1 Introduction	24
1.2 The European Regional Setting of the Cornubian Variscides	26
1.2:1 The plate tectonic framework	26
1.2:2 General work on the plate tectonic framework	30
1.3 The Variscides of the Cornubian Peninsula	32
1.3:1 The structural framework	32
1.3:2 The sedimentary framework	44
1.4 The geology of the Culm Basin and the Rusey area.	47
1.5 Summary	52
Chapter Two	
The Structure of Rusey Headland	55
2.1 Introduction	55
2.2 The Tectono-sedimentary Framework	57
2.3 The Structural Framework	60
2.3:1 Early Extension and Inversion	60
2.3:2 The Thrusts	64
2.3:3 The Rusey Thrust Zone and the Rusey Breccia	72
2.3:4 The Extensional Fault	78
2.3:5 The Strike-slip Faults	79
2.4 Discussion	81
2.5 Summary	83

Chapter Three

Photoelastic modeling of the Rusey Thrust system	86
3.1 Introduction	86
3.1:1 The Photoelastic Principle	87
3.1:2 Plane Polarised Light	87
3.1:3 Circularly Polarised Light	88
3.2 The Modeling Techniques	94
3.2:1 The Photoelastic Technique	94
3.2:2 Model Design and Preparation	96
3.2:2.1 Model Design	96
3.2:2.2 Model Preparation	96
3.3 The Analysis Techniques	97
3.3:1 The Principal Stress Orientations	97
3.3:2 The Sum of the Principal Stresses.	99
3.4 Modeling of the Rusey Thrust System	99
3.4:1 Isochromatic Data	99
3.4:2 Isoclinic Data	102
3.4:3 Discussion of the Isochromatic Data.	104
3.4:4 Discussion of the Isoclinic Data	109
3.5 Summary	110

Chapter Four

Field description and statistical analysis of the Rusey Breccia	115
4.1 Introduction	115
4.2 Organisational Features	115
4.2:1 Breccia layers	115
4.2:2 Clast grading	120
4.2:3 Oriented clasts	120
4.3 Morphological Features	125
4.3:1 Clast shape and structure	125
4.3:2 Crystal form and state	125
4.3:3 Voids	128
4.3:4 Interstitial Matrix	128
4.4 Spatial Arrangement of the Breccia	128

4.5 Late Deformation	134
4.6 Statistical Analysis of the Rusey Breccia	136
4.6:1 Introduction	136
4.6:2 Preferred Orientations	139
4.6:2a Methodology	139
4.6:2b Sample Analysis	148
4.6:2c Implications	149
4.6:3 Clast Density Variation	156
4.6:3a Methodology	156
4.6:3b Sample Analysis	157
4.6:3c Implications of variations in clast material distribution for breccia reorganisation mechanisms	159
4.6:4 Clast Size Grading	160
4.6:4a Methodology	160
4.6:4b Sample Analysis	161
4.6:4c Implications	163
4.7 Summary of the statistical analysis	163

Chapter Five

Fluidisation Mechanisms and Processes	165
5.1 Introduction	165
5.2 Fluidisation	165
5.2:1 Fluidisation (sensu stricto)	166
5.2:1a Fluidisation mechanism	166
5.2:1b Effects of variation in particle properties	173
5.2:2 Liquefaction	174
5.2:2a Liquefaction	174
5.2:2b Shear liquefaction	176
5.3 Rheology of fluidised beds	176
5.4 The comparative properties of fluidised beds and liquids	181
5.4:1 Temperature/ Velocity	182
5.4:2 Viscosity	183
5.4:3 Thermal Behaviour	183
5.4:5 Similarities in Behaviour between Fluidised Beds and Liquids	183

5.4:6 Differences in Behaviour between Fluidised Beds and Liquids	183
5.5 Experimental evidence for fluidisation	184
5.6 Field evidence of the occurrence of fluidisation in a geological environment	190
5.7 Summary	191

Chapter Six

Literature review of gel formation and behaviour	194
6.1 Introduction	194
6.2 Gel and Colloid Formation	194
6.2:1 The Formation of Silica Gel	194
6.2:2 The Formation of Colloidal Silica	196
6.3 Crystallisation in gels and colloids	198
6.4 Field evidence for the involvement of silica gels in geological processes	203
6.5 Summary	206

Chapter Seven

Constraints of Rusey Breccia mineralisation mechanisms using evidence from Kimmeridger Bay, Dorset.	208
7.1 Thrusting and Related Brecciation in Kimmeridge Bay, Dorset	208
7.1:1 Location and Aims	208
7.1:2 The Regional Structure	209
7.1:3 The Structure in Kimmeridge Bay	209
7.1:3a <i>The Thrusts</i>	210
7.14:3b <i>The Joints</i>	216
7.1:4 Mineralisation and Breccia Development.	218
7.2 Comparison of mineralisation mechanisms	221
7.3 Summary	226

Chapter Eight	
A model for the formation of the Rusey Breccia	 229
8.1 Introduction	 229
8.2 Comparison of Deformation at Rusey Headland and Kimmeridge Bay	 230
8.3 Properties of Fluidisation	 230
8.4 Properties of Colloids	 231
8.5 Synopsis of the field data	 231
8.6 The reorganisational model	 232
8.7 Evolutionary model for the formation of the Rusey Breccia	 233
8.8 Unresolved Issues	 237
8.9 Further work	 238
References	 240
Acknowledgements	 250

FIGURE LIST

Chapter One

Figure 1.1	
Composite location map showing the project field area relative to the European Variscides	 25
Figure 1.2	
Paleo-position of continents, showing paleolatitude measurements, (+ north, - south), Van der Voo, 1982	 26
Figure 1.3	
The Variscan fold belt of Europe, after Matte, 1986	 27
Figure 1.4	
Variscan cross sections, after Matte, 1986	
a : Interpretive cross sections through the Ardennes massif	
b : The recumbent fold nappes of the Montagne Noir, Massif Central	 29
Figure 1.5	
Correlation of Cornubia and the Harz Mountains across the Bray Fault, Holder and Leveridge, 1986	 30
Figure 1.6	
Variscan inversion of the Munster basin, Cooper et al. 1986	
a: Pre-Variscan basement with Paleozoic sediment fill on top	
b: Thrusting transports basement slice into sediment fill, causing folding and inversion of the Munster basin	
c: Deformation front moves north and second thrust forms	 31
Figure 1.7	
Geological map of the Rusey-Dartmoor area, Isaac et al. 1982	 33
Figure 1.8	
Thrust model for the Cornubian Variscides, Shackleton et al. 1982	 34
Figure 1.9	
Cornubian folds and cleavages, after Hobson and Sanderson, 1983	 35
Figure 1.10	
Cornubian refolded fold styles, Hobson and Sanderson, 1983	 36
Figure 1.11	
Backthrusting resulting from the development of a sticking point on the main detachment, Coward and Smallwood, 1984	 37
Figure 1.12	
The Variscan dextral megashear system, Barnes and Andrews, 1986	 38

Figure 1.13	
Pull apart basins development along dextral strike-slip faults, Aydin and Nur, 1982	 38
Figure 1.14	
Structure of the southern Culm overfold and a model for it's development, densities in g/cm ³ , Warr, 1989	 39
Figure 1.15	
Variscan evolution of the Cornubian peninsula, Pamplin, 1990	 40
Figure 1.16	
Cross sections of the Cornubian peninsula, from deep seismic sections, Le Gall, 1990	 42
Figure 1.17	
Tectono-sedimentary framework for the Cornubian basins inverted in the Variscan orogeny, Hartley and Warr, 1990	 44
Figure 1.18	
a: sedimentary framework for the Trevone basin	
b: sedimentary framework for the Culm basin	
c: schematic diagram of the deformation style for the Culm basin, after Selwood, 1990	 46
Figure 1.19	
Map and section of the Rusey headland, north Cornwall, after Freshney et al, BGS memoir, 1972	 48
Figure 1.20	
Restoration of slumped, folded and thrust beds, Bude, Whalley and Lloyd, 1986	 50
Figure 1.21	
Relationships between thrusting and extension, Mapeo and Andrews, 1991	 52

Chapter Two

Figure 2.1
Evolution of the Cornubian basins, Pamplin, 1990 55

Figure 2.2
Geological map of the Rusey area, North Cornwall. (Slip sense on interpreted faults is deduced from bedding orientations) 56

Figure 2.3
The Bouma sequence classification for turbidite flow deposits, after Collinson and Thompson, 1989, (originally Bouma, 1969) 59

Figure 2.4
a: Sedimentary deposition patterns in a continental half graben,
b: Schematic diagram of the deposition patterns in the Culm basin, prior to inversion 61

Figure 2.5
Schematic diagram of the path of time equivalent rocks during thrusting and later extensional faulting 62

Figure 2.6 Inverted extensional structures at Cutoff point 65

Figure 2.7 Incipient thrusts and thrust tip folds at Cutoff point. 66

Figure 2.8 Well developed thrust faults at Cutoff Point 67

Figure 2.9 Geological section of the Rusey headland, North Cornwall 69

Figure 2.10
Stereoplots of the poles to the thrust planes and the axial planes of minor folds recorded at Rusey Headland 70

Figure 2.11
Examples of multiple generations of veining in the shales surrounding the main breccia band 71

Figure 2.12
a Strike section of part of fault zone A showing a thrust duplex/stack.
b Dip section of part of fault zone A showing thrust planes and folds 73

Figure 2.13
Block diagram, showing the three dimensional form of the main Rusey breccia band 75

Figure 2.14
Composite synoptic diagram of the Rusey Breccia and its defining features. 77

Figure 2.15
Thrust bound and dissected folds, located on the Cliff Fault at the cliff top outcrop, Rusey 78

Chapter Three

Figure 3.1	
Plane polarisation of light	 88
Figure 3.2	
Circular polarisation of light	 90
Figure 3.3	
Development of coloured fringes in a photoelastic model	 92
Figure 3.4	
Annotated diagram of the photoelastic rig used in the modelling program	 95
Figure 3.5	
Conversion of raw isoclinic data to map of σ_1 orientations	 98
Figure 3.6	
Isochromatic data for R26 at 600psi	
a: Initial photographic data	
b: Contoured differential stress map	 100
Figure 3.7	
Contour maps of differential stress levels within the R series models:	
a: R5 at 300 psi load	
b: R10 at 300 psi load	
c: R26 at 300 psi load	
d: R26 at 600 psi load	 101
Figure 3.8	
Comparison of σ_1 orientations for the R series models	
a: R5	
b: R10	
c: R26	 103
Figure 3.9	
Mohr diagram to show the requisite stress states for shear and tensile failure	 105
Figure 3.10	
The thrust ramp model for breccia formation along the Rusey Thrust	 111
Figure 3.11	
Breccia formation due to extension within a dilational jog along the Rusey Thrust	 112

Chapter Four

Figure 4.1
Line drawing of sample R15, showing the possible presence of multiple layers 124

Figure 4.2
Sketch from thin section showing the change in grain size with distance from the clast 129

Figure 4.3
Schematic synoptic diagram of the Rusey Main Breccia Band 133

Figure 4.4
Internal structure of late stage slip planes within the Rusey Breccia 135

Figure 4.5
A comparison of rose diagrams produced by different methods of data input
a: hand drawing of long and short axes of each clast from a photograph
b: drafting package trace of a bitmap scan of a photograph
c: hand line drawing from a photograph 140

Figure 4.6a
A comparison of Orien plots with
a: with the largest clast element removed
b: complete data sets 143

Figure 4.6b
A comparison of Orien plots with
a: with the largest clast element removed
b: complete data sets 144

Figure 4.7
Clast preferred orientation plots of samples of the Rusey Breccia 145

Figure 4.8
Clast preferred orientation plots of samples of the Rusey Breccia 146

Figure 4.9
Clast preferred orientation plots of samples of the Rusey Breccia 147

Figure 4.10
Terminology used to describe the trend orientations from Rusey Breccia Orien plots 148

Figure 4.11
Grain rotation within a flowing medium 150

Figure 4.12
Diagram showing particle interaction, after Rees, 1968 152

Figure 4.13
Rose diagram showing the orientation of long axes of sand grains in a section cut parallel to the dune face, after Rees, 1968 153

Figure 4.14
a: Preferred grain orientations from aeolian sands
b: Preferred grain orientations from wind tunnel experiments 154

Figure 4.15
Flow orientation of non-equant clasts 156

Figure 4.16
Variation in clast volume with distance across a layer 158

Figure 4.17
Clast size distributions for the samples shown in the previous figure.
Variation trends marked in red 162

Chapter Five

Figure 5.1
The fluidisation process 167

Figure 5.2
Fluidisation styles after, Richardson, 1971 168

Figure 5.3
Flow patterns around an ascending bubble, showing the reversal
of flow in the area immediately below the bubble, after Collins, 1984 170

Figure 5.4
Fluidisation velocity curves, after Richardson, 1971 172

Figure 5.5
Liquefaction due to vibration 175

Figure 5.6
Liquefaction due to shear 177

Figure 5.7
Viscosity variation with increasing flow rate at constant applied shear 178

Figure 5.8
Variation in shear stress developed with increasing applied shear force
at constant flow rate (9cm/sec), after Schugerl, 1971 180

Figure 5.9
Variation in viscosity with percentage of fine grained material 181

Figure 5.10
Behaviour of particles in fluidised or sedimenting suspensions,
Richardson, 1971 182

Figure 5.11
Deformation sequence for fluidised sand beds, Nichols et al. 1994 184

Figure 5.12
Fluid escape structures in sand and cohesive layers after
Nichols et al. 1994 185

Figure 5.13
Fracturing prior to fluidisation in cohesive granular materials,187

Figure 5.14
Doming of layers and cavity development above a fluidised channel,
after McCallum, 1985188

Figure 5.15
Development of channelised flow after fluid dispersion through a coarse
grained layer 189

Chapter Six

Figure 6.1
Chemical bonding of monosilicic acid units to form a three dimensional network, thereby forming the solid phase of a gel, after Henisch, 1970 195

Figure 6.2
Aggregation of dispersed silica particles to form an interlinked three dimensional network, causing a very large increase in the viscosity of the liquid196

Figure 6.3
Potential energy curve for a colloidal suspension, showing the balance between attractive and repulsive forces, Elliston, 1984 197

Figure 6.4
Cusp formation around a crystal growing in gel, after Henisch, 1970 199

Figure 6.5
Development of sheaves of radially growing quartz fibres 201

Figure 6.6
Graph of variation through time of the conversion of colloidal silica to quartz, Oehler, 1976 202

Figure 6.7
Conversion rates of amorphous silica to quartz at different temperatures and pressures, Betterman and Liebau, 1975 203

Figure 6.8
Differences in colloid density depending on size range of colloid particles, Saunders, 1990 204

Chapter Seven

Figure 7.1
Sketch map of Kimmeridge Bay, Dorset, Leddra et al. 1987 208

Figure 7.2
Map of the thrusts outcropping in the Flats area of Kimmeridge Bay,
Dorset, Leddra et al. 1987 211

Figure 7.3
Thrust displacement in the Charnel area of Kimmeridge Bay 211

Figure 7.4
a: Stereoplots of thrust data from the Flats, Kimmeridge Bay, Leddra et. al. (1987),
showing only a very slight preferred orientation to the thrust planes
b: Stereoplot of all thrust data from the Flats, Kimmeridge Bay,
Leddra et. al. 1987 215

Figure 7.5
Pod development in cementstone bed as the thrust plane bifurcates across the central
region of the bed 220

Figure 7.6
Hydrofracture textures within the thrust plane breccia 221

Chapter Eight

Figure 8.1a, b and c
The early stages of the evolution of the Rusey Breccia 235

Figure 8.1 d and e
The final stages of the evolution of the Rusey Breccia. 236

PLATE LIST

Chapter Two

Plate 2.1	
The deformation surrounding the main thrust plane and fault zone A	 58
Plate 2.2	
Shale outcropping below the main breccia band	 58
Plate 2.3	
Photograph showing both brittle fracturing of layers and ductile folding within the same section, Cutoff Point, Exit Bay, Rusey	 63
Plate 2.4	
Inverted extensional structures at Cutoff point	 65
Plate 2.5	
Incipient thrusts and thrust tip folds at Cutoff point	 66
Plate 2.6	
Well developed thrust faults at Cutoff Point	 67
Plate 2.7a	
Extensional faulting in the rocks at Cutoff Point, Exit Bay	 68
Plate 2.7 b	
Extensional faulting in the rocks at Cutoff Point, Exit Bay	 68
Plate 2.8	
The Rusey Headland seen from the cliff top	 69
Plate 2.9	
Photograph of the Rusey Breccia showing veined clasts without fabric development	 74
Plate 2.10	
Well developed internal slip surfaces within the main breccia band	 76
Plate 2.11	
Well developed slickensides on the slip surfaces within the main breccia band	 76
Plate 2.12	
The lower cliffs of Middle Strand from the cliff top path	 80

Chapter Three

Plate 3.1
Isoclinic lines 89

Plate 3.2
Isochromatic patterns 93

Plate 3.3
Polygonal vein network, from the shales surrounding the Rusey Breccia 106

Plate 3.4
Ladder veining from the shales surrounding the Rusey Breccia 107

Chapter Four

Plate 4.1
The main Rusey Breccia band in outcrop 116

Plate 4.2a
The matrix textures in the breccia 117

Plate 4.2b
The matrix textures in the breccia 117

Plate 4.3
Black layer boundaries within the organised Rusey Breccia 118

Plate 4.4
White layer boundaries at the edge of the organised Rusey Breccia 118

Plate 4.5
White layer boundaries within the organised Rusey Breccia 119

Plate 4.6
Layer development without explicit layer boundaries 119

Plate 4.7
Channel development in the Rusey Breccia 121

Plate 4.8
'Negative' channel development in the Rusey Breccia 121

Plate 4.9a
Half of a breccia layer showing clast size fining towards the layer boundary 122

Plate 4.9b
The central layer exhibits clast size grading that fines outwards towards both layer boundaries 122

Plate 4.10a	
Transition from hydrofractured country rock into Rusey Breccia	 123
Plate 4.10b	
A sample where the preferred orientation of clasts is visually apparent	 123
Plate 4.11a	
A breccia clast bearing evidence of pre-brecciation veining events	 126
Plate 4.11b	
Rebrecciated breccia clast	 127
Plate 4.11c	
Protrusion of the larger clast from the upper layer into the lower layer	 127
Plate 4.12	
Large veins with euhedral quartz crystallisation and central space	 130
Plate 4.13a	
Breccia layers defined by layers of dark material along the layer boundaries	 131
Plate 4.13b	
Dark layers within corona mineralisation around clasts	 131
Plate 4.13c	
Dark material at edge of layer, forming the matrix which supports the clast/corona units	 132

Chapter Seven

Plate 7.1	
Extensional faulting cross cutting the mineralised thrust planes	 212
Plate 7.2	
Thrust fault passing downwards from the Flats Stone into the underlying shales	 213
Plate 7.3	
Polygonal development of thrusts	 214
Plate 7.4	
Development of cobblestone jointing	 217
Plate 7.5	
Development of jointing parallel to the thrust tip line	 219
Plate 7.6	
Matrix textures along a thrust plane in the Charnel area, Kimmeridge Bay	 222

Plate 7.7

Graded clast sequence along a thrust plane in the Charnel area of
Kimmeridge Bay

..... 223

Introduction and Aims

This thesis presents the results of a study of the tectonic setting and textures of the Rusey Breccia, which outcrops at one location on the North Cornish coast. The breccia body is a dominantly matrix supported band 80m long and 2-3m thick, located within the Rusey Thrust zone, a band of highly deformed shales which acquired an intense planar deformation fabric during the Variscan orogeny. The fabric detaches onto the surface of the mineralised breccia band and indicates that the timing of brecciation and mineralisation of the breccia was very early in the thrusting history. This is supported by published regional data which demonstrates that thrusting was the initial phase of deformation regionally and in some cases was syn-sedimentary.

The study aims to determine a mechanism and environment of formation which can account for certain features of the breccia, i.e. the presence of decimetre scale layering throughout the main band of breccia, the frequent presence of clast size grading within these layers and the alignment of the long axes of clasts visible in certain samples.

In addition an explanation of the matrix formation mechanisms and environment of formation was sought. The matrix is made up of pure, white, undeformed euhedral coronas of quartz developed radially in three dimensions around the clasts. There is no fault gouge associated with the breccia band and very little interstitial quartz matrix present to cement the clast/corona units together.

This assemblage of features, (layered and graded clast populations with three dimensionally euhedral matrix crystallization and an absence of fault gouge), is referred to in this work as Rusey texture.

The aim of the thesis is primarily to understand the environment and mechanisms that gave rise to the highly distinctive structural and mineralogical textures observed in the main body of the breccia. In the course of the study the local geological setting and history have been investigated in order to assess the broader setting within which the breccia formed and to determine the probable regional significance of the Rusey Thrust, the major thrust fault intimately associated with the Rusey Breccia. The limited outcrop available at Rusey Headland, North Cornwall, where the thrust and breccia crop out, combined with the difficulty of correlating the coastal structure with that further inland, make the regional significance of the structures at Rusey difficult to define. Therefore, whilst a certain amount of regional and local information on the structure and history of the area is required in order to understand the circumstances under which the thrust was active and the breccia formed, the majority of the thesis concentrates on understanding the how the more unusual features of the breccia were generated.

Chapters 1 to 3 deal with the proposed plate tectonic setting and the regional scale evolution of the system. Chapter 1, a survey of the literature on Cornubian geology, is intended to provide the reader with a picture of the large scale geological setting and the tectonic processes operating in the area studied during the formation of the breccia. Chapter 2 focuses more closely on the Rusey Headland, describing the structures, lithologies and probable history of the study area. The work is derived from the mapping carried out by the author and serves to determine the time period and geological environment in which the evolution of the breccia takes place. Chapter 3 describes an analogue model study, aimed at understanding the stress conditions under which the brecciated body formed. This study is intended to provide a description of the conditions under which the breccia developed and not the processes by which it was developed.

Chapters 4 to 8 are concerned with the main aim of the thesis which is to determine the conditions under which the breccia was modified from its original form to the form currently observed. Chapter 4 presents a detailed description of the features observed in the breccia, i.e. the layered and graded structure, the clast orientation and the mineralisation textures and a statistical analysis of the distribution of clast size and clast material and preferred orientation of clasts within single layers. The statistical analysis is included in Chapter 4 as it, together with the field observations, provide the data which must be accounted for by any model which is proposed. Chapters 5 and 6 are literature reviews of fluidisation and colloid and gel formation, two important processes that may be linked to the textural modification of the breccia. Chapter 7 assesses the relative merits of the two possible processes (gel and colloid formation) presented as candidates for generating the mineralising fluid. Field data from a second location in Kimmeridge Bay, Dorset is presented and used to justify the choice of one of these processes (colloid formation) in preference to the other. Chapter 8 summarizes the data and presents a model for the evolution of the Rusey Breccia. It concludes by assessing the significance of this model in understanding the formation of breccias in general and gives some suggestions for further work on the subject.

Chapter One

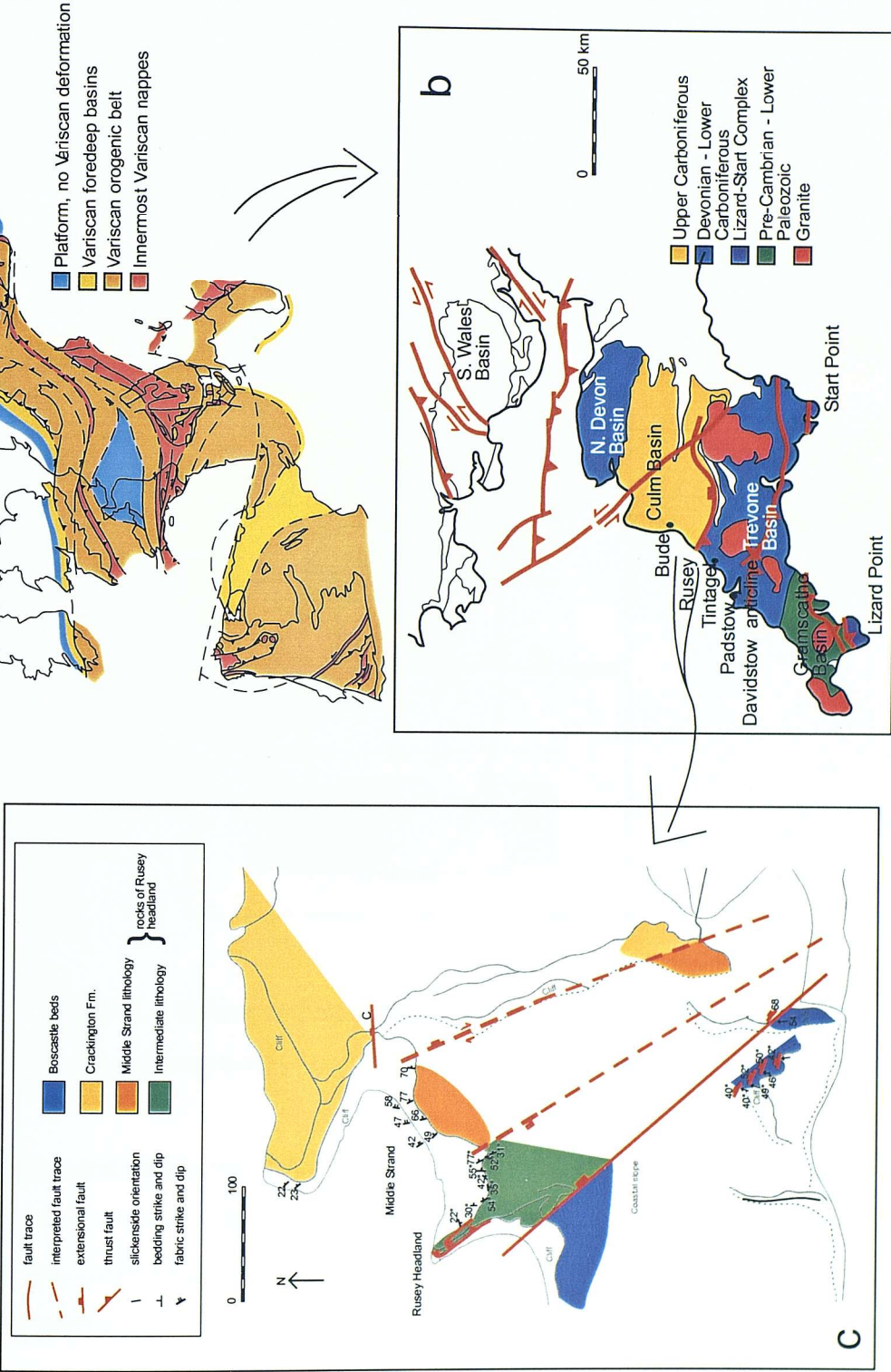
A literature review of the geology of the Variscan Orogen and the Cornubian Peninsula

1.1 Introduction

This chapter presents a summary of the literature that is relevant to this study. It is intended to create a general understanding of the structure and extent of the Variscan orogenic belt and its effect on the area studied, Fig. 1.1, thereby allowing an informed appreciation of the geological setting and history of the rocks that form the subject of this thesis. The first section describes the whole Variscan belt, which extends from the Appalachians east to the Urals and south to Spain, followed by a more detailed survey of the literature specific to the Cornubian peninsula. The next section deals with the sedimentary framework of the area and the final section is a description of those aspects of the local geology directly relevant to this study. The terms Variscan, Hercynian and Armorican are used almost interchangeably in the literature. They were coined for specific parts of the orogenic belt and when used precisely refer to deformation of different ages occurring in different parts of the belt. With increasing correlation between areas the terms have become confused to a certain extent. Within this thesis the orogen is referred to as Variscan. If the terms Hercynian, Rhenohercynian or Armorican are used they refer to geographical areas within the orogen, e.g. the Armorican Massif. The regional and local framework constructed in this chapter will be used as the basis of the model within which the structure of the Rusey headland will be examined (Chapter 2).

Figure 1.1: Composite location map showing the project field are a relative to the European Variscides.

- a : the Variscan belt of Europe
- b: the Variscides of the Devon and Cornwall
- c : the geology of Rusey headland



1.2 The European Regional Setting of the ^{Cornubian} Variscides

1.2.1 The plate tectonic framework

The oceanic and continental basin settings of the Cornubian Variscides have been placed in a plate tectonic setting by Van der Voo (1982), and Matte (1986). Van der Voo uses paleomagnetic data to constrain the widths of a major Devonian ocean between Gondwana and Laurussia, which by the Visean or Lower Carboniferous had shrunk to approximately 800km, Fig. 1.2.

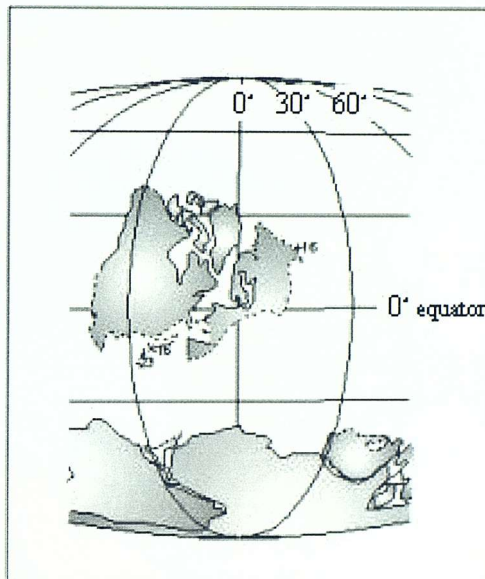


Figure 1.2 Paleo-position of continents, showing paleolatitude measurements, (+ north, - south), Van der Voo, 1982

Matte's 1986 paper details the plate-tectonic setting of the entire 3000km long, 7-800km wide Variscan/Rhenohercynian belt, Fig. 1.3. This paper covers a very large area, but is one of the few clear integrations of data from across the majority of the orogen and as such is very useful.

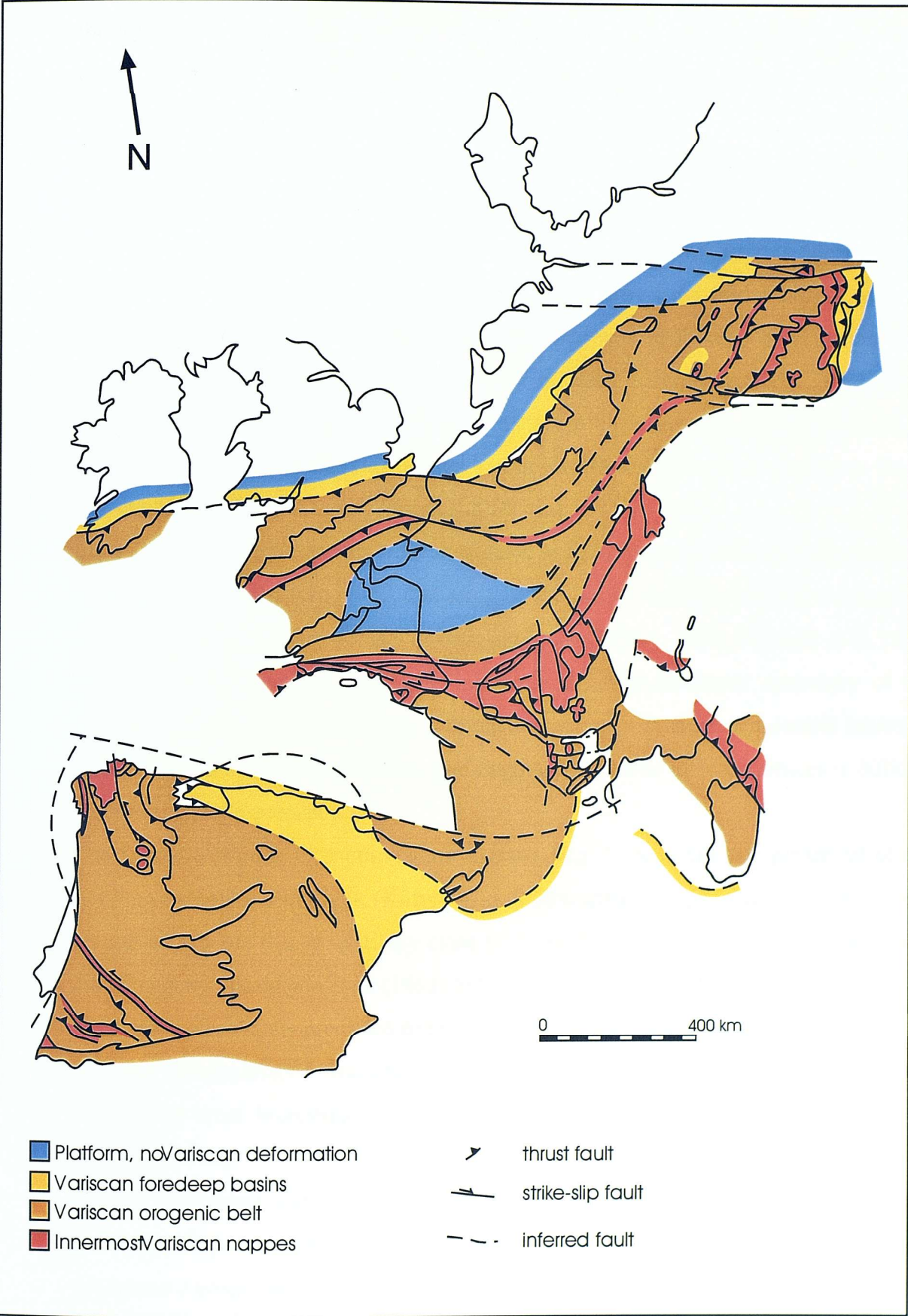


Figure 1.3 The Variscan fold belt of Europe, after Matte, 1986

The model described in the paper has symmetric development of thrusts and nappes throughout the entire belt, with the thrusts developed bi-directionally from a central zone. This broad central zone was flanked on both sides along the entire belt, by Devonian-Carboniferous sedimentary basins. These are now either low grade or unmetamorphosed, characterized by syn-tectonic flysch and molasse. Matte describes the external parts of the orogen, (such as the Cornubian peninsula), as characterized by one or more décollement which exploit stratigraphic horizons, leaving the basement undeformed, Fig. 1.4a, and the internal parts of the orogen by fold nappes, Fig. 1.4b, which are dated at 350-320 Ma. Fold nappe development temporally overlaps the onset of brittle/ductile strike-slip deformation that initiated at approximately 340 Ma and persisted until 300Ma. The central zone of metamorphic terranes and granites imply that up to 25km of overburden was created and eroded over an 80 Ma period from Upper Devonian, (360 Ma), to Lower Permian, (280 Ma).

For a long time the Variscides were considered to have no oceanic suture. More recently however it has been recognised that most of the mafic rocks present are remnants of oceanic crust, from either oceanic or back-arc basins, (Kirby, 1979, Bodinier et al. 1986, Barnes and Andrews, 1986). Matte interprets the well-developed symmetry of the orogen as the result of ocean closure by subduction on both sides of a central landmass to form paired ophiolite sequences. The original location of these sequences is difficult to determine with precision.

The Lizard Complex of southwestern England, Fig. 1.1b, is the best preserved of the northern ophiolitic sequences, (Kirby 1979). Radiometric dating by Davies, (1984) gives an age of 375 Ma that is relatively close to the probable time of obduction, which was considered by Shackleton et al. (1982) to be Lower Devonian.

Wide oceans are rarely predicted from ophiolites such as the Lizard Complex, which has isotopic characteristics transitional between pure mid-ocean ridge basalts and continental crust indicating that rifting was still in the initial stages, (Barnes and Andrews, 1986).

Crustal thickening during compression caused crustal flexure and foreland basin formation. Rapid erosion of the uplifted orogenic belt filled those basins with turbiditic sequences during the Carboniferous. The whole Variscan belt has been very strongly modified by synchronous and subsequent strike-slip and lateral extrusion.

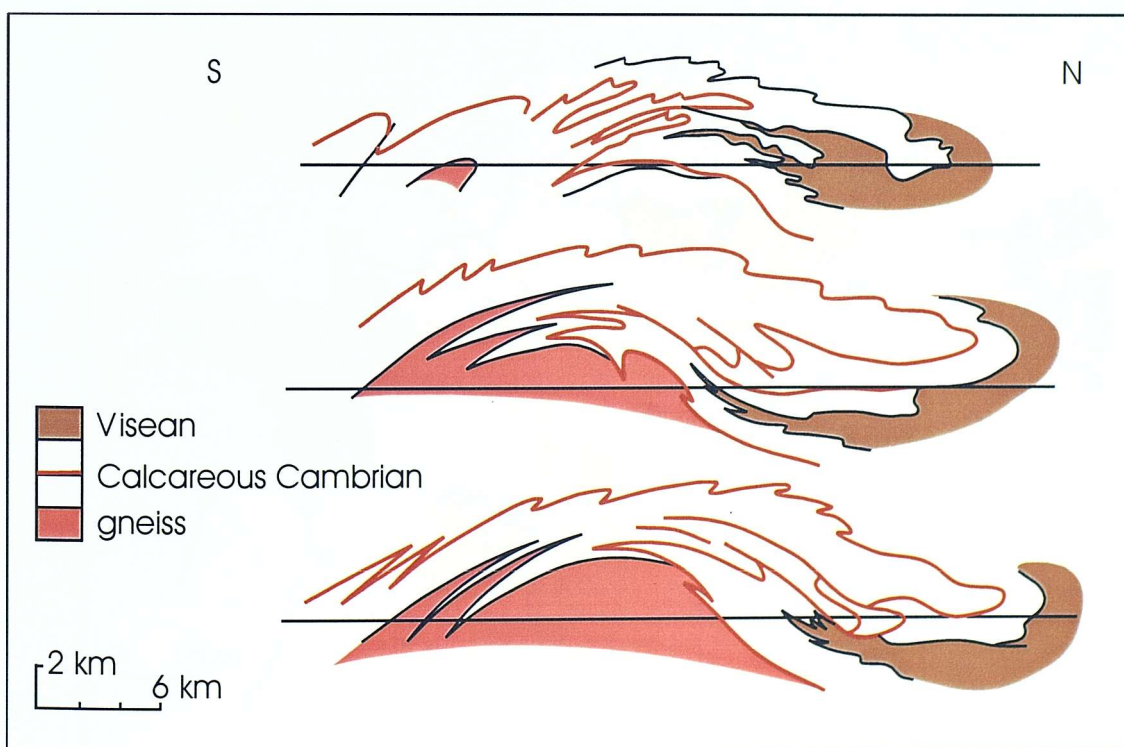
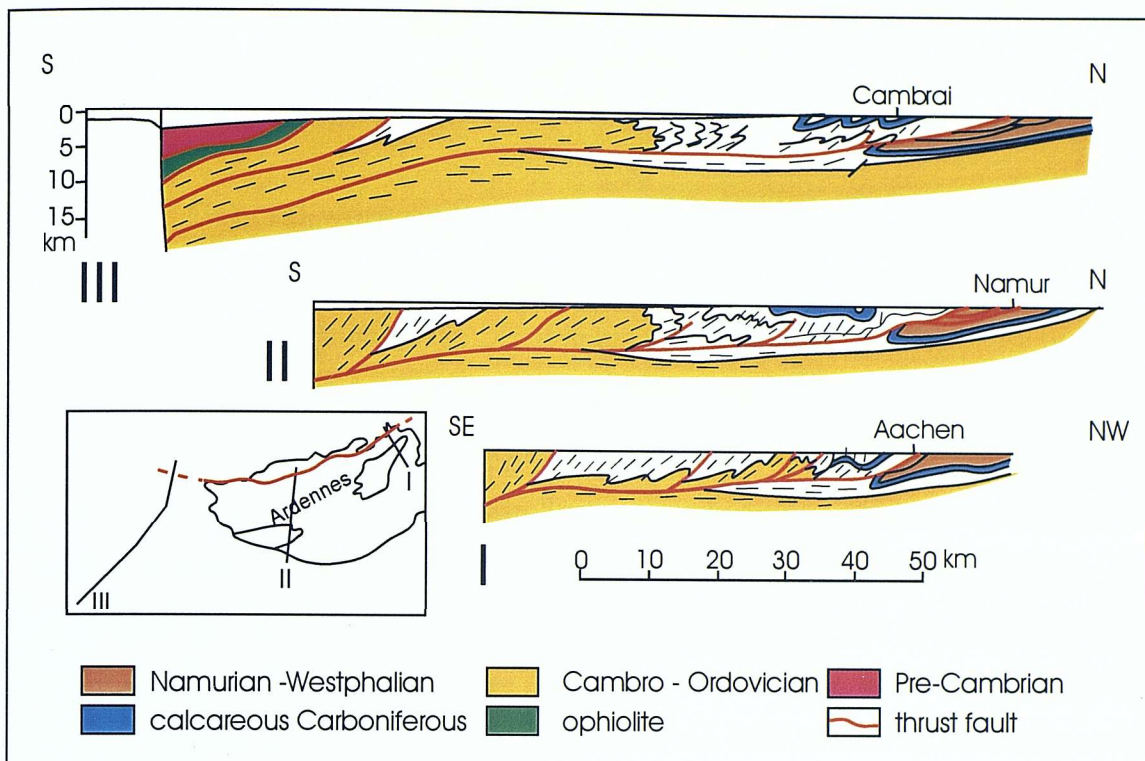


Figure 1.4 Variscan cross sections, after Matte, 1986
 a : Interpretive cross sections through the Ardennes massif
 b : The recumbent fold nappes of the Montaigne Noir, Massif Central

In summary, Matte (1986) clearly and concisely presents and justifies a classic plate tectonic model for the Variscan, consisting of obduction of oceanic crust followed by continental collision, with a complex intra-continental history after closure that he does not discuss in detail.

1.2.2 General work on the plate tectonic framework

Holder and Leveridge, (1986), carried out a correlation of Variscan structure across the Bray fault, Fig. 1.5, in which they emphasize the effects of dextral strike-slip faulting. They correlate the rocks of the Cornubian peninsula with those of the Harz Mountains in Germany, on the basis of the similarity in stratigraphy, provenance and geochemistry. Geophysical surveys have shown that basement lineations from both areas terminate against the Bray fault, Fig. 1.5. Holder and Leveridge use the occurrence of major dextral strike-slip faults and the presence of considerable transcurrent displacement in the Paris basin to suggest that Cornwall and the Harz mountains, now separated by the Ardennes Massif, were originally continuous.

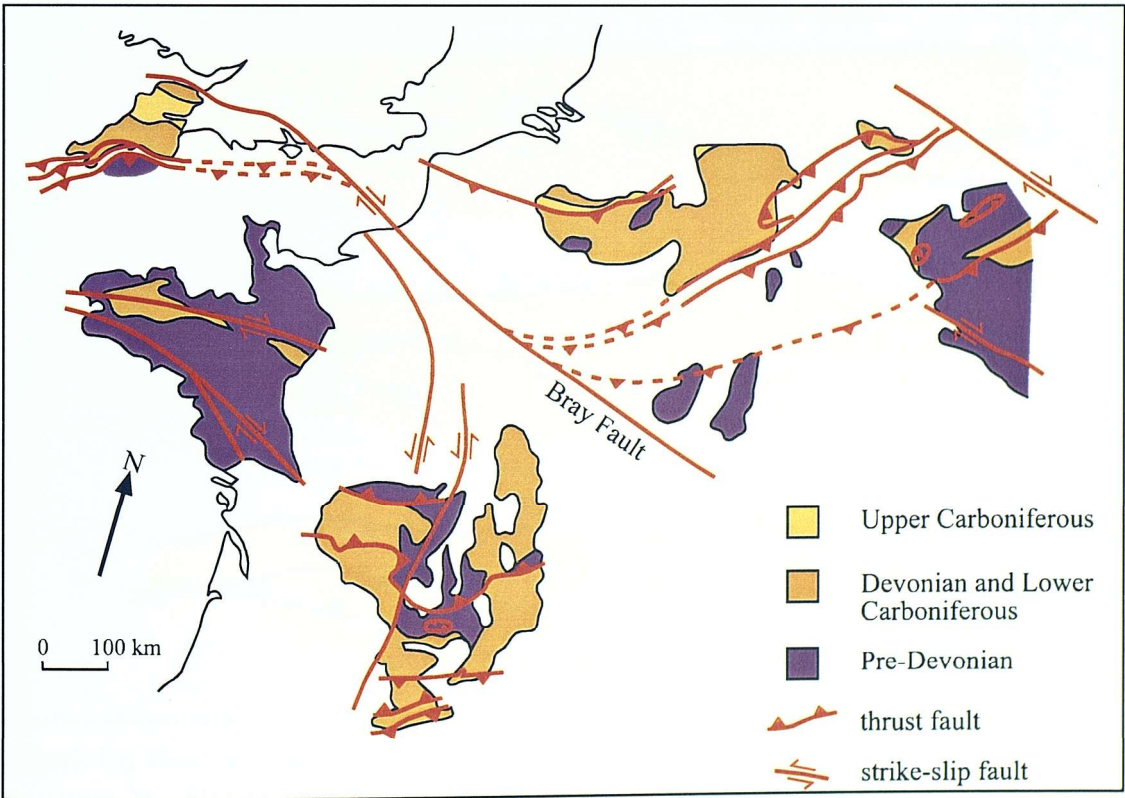


Figure 1.5 Correlation of Cornubia and the Harz Mountains across the Bray Fault, Holder and Leveridge, 1986

Paleomagnetic data (Scotese et al, 1979), is used to suggest that closure of the Cornish Rheohercynian basin in the Upper Devonian coincided with a shift in plate motion from normal convergence to oblique dextral convergence and that thrusting took place within a dextral transpressive regime

Cooper et al. (1986), document the outcrop patterns of the Irish Variscides, which are dominated by fold structures. Deformation is considered to have occurred above a low angle sole thrust, with initial layer parallel shortening subsequently overprinted by thrusting, Fig. 1.6. As in the Culm basin, there is a systematic change in facing direction from southward facing in the south through vertical, upward facing in the centre to northward facing in the north. North-south oriented dextral strike-slip faults have rotated the fold axes clockwise.

The model of Cooper et al. is summarized in Fig. 1.6, showing a deformation style and evolution very similar to that proposed for Devon and Cornwall, (Shackleton et al. 1982, Hobson and Sanderson, 1983, Pamplin, 1990).

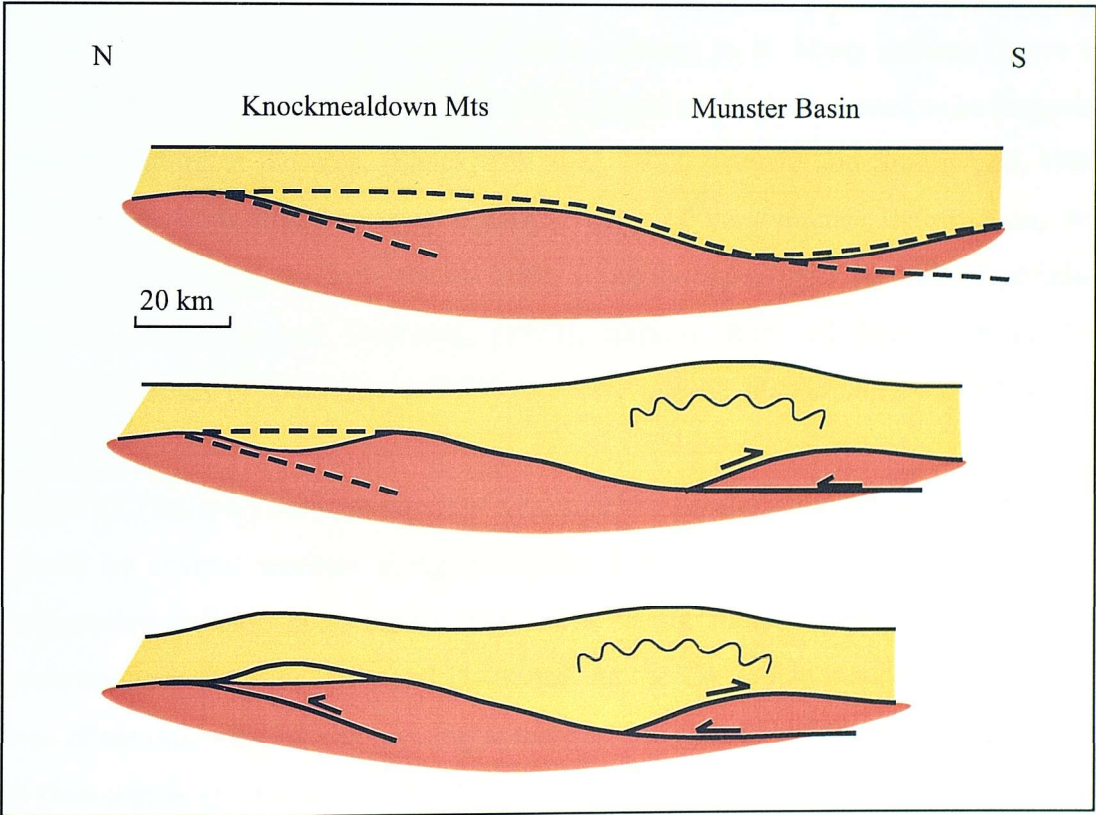


Figure 1.6 Variscan inversion of the Munster basin, Cooper et al. 1986
a: Pre-Variscan basement with Paleozoic sediment fill on top
b: Thrusting transports basement slice into sediment fill, causing folding and inversion of the Munster basin
c: Deformation front moves north and second thrust forms

1.3 The Variscides of the Cornubian Peninsula

1.3.1 The structural framework

Sanderson and Dearman's study (1973) is based on fold geometries and defines an area of south facing structures extending from south of Bude to Polzeath, (ten or twelve miles south of Boscawen). Their representation of the deformation in the Culm basin bears a striking resemblance to the model of Cooper et al. 1986, for folding in the Munster basin, Eire, (Fig. 1.6), indicating a broad along strike similarity in deformation styles for the northern extremity of the orogen. However, a structural interpretation which stretches across the Cornubian peninsula based only on the correlation of fold geometries and styles is unlikely to be entirely accurate, since the peninsula is segmented by synchronous and later thrust and extensional faulting.

Later work, e.g. Isaac et al. (1982), normally cites the Rusey Fault as the southern limit both of the Culm basin and the deformation relating to it. Many authors ignore the Rusey area altogether and suggest that the Tintagel high strain zone is the important structural feature of the area, (Shackleton et al, 1982, Coward and Smallwood, 1984). The Tintagel high strain zone is a narrow zone of very intense deformation, with rotation of fold axes to form sheath folds, (tongue-shaped/tight recumbent periclinal folds of Sanderson and Dearman, (1973), oblique folds of Sanderson (1973a), tergiversate folds of Dearman, (1969)) and very intense planar fabrics and lineations occur.

Isaac et al. (op. cit.) demonstrate that although the south facing structures have been mapped by several workers along the coast, it may not be possible to extend this interpretation inland. They report only north facing folds and northerly displacing thrusts between Bodmin Moor and Dartmoor. As the Cornubian peninsula consists of a series of tectonic nappes which young to the west the coastal interpretation of Sanderson and Dearman is as valid as the inland interpretation of Isaac et al. However, it highlights the need to accept not only that lateral variation may be a very important feature of the evolution of the area, but also that the coastal section may be composed of far more distantly related tectonic units than assumed by Sanderson and co-workers. Isaac et al. also define the Rusey fault as a late, low angle extensional fault, which dissects the

earlier formed nappes, Fig. 1.7, though they correlate it along strike with a southward displacing thrust fault. These field based papers contrast with the later work, which begins to show a more model driven emphasis, with the influence of plate tectonic concepts more in evidence in the interpretation of field data.

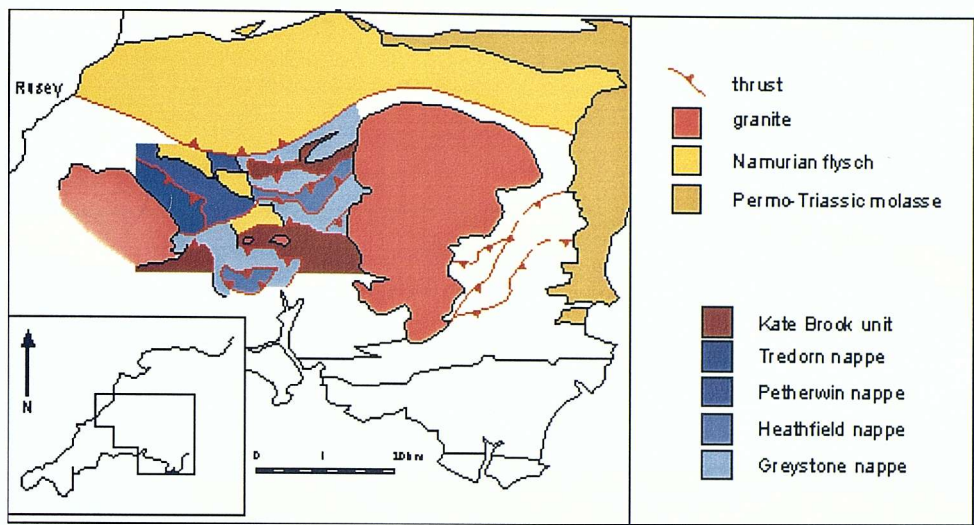


Figure 1.7 Geological map of the Rusey-Dartmoor area, Isaac et al. 1982

Shackleton et al. (1982), present a relatively simple model for Variscan deformation, Fig 1.8, whereby regional northerly directed thrusting occurs on a major low angle detachment, dipping southwards below the entire Cornubian peninsula. They describe two local zones of high intensity deformation, north facing at Ilfracombe and south facing at Tintagel.

Hobson and Sanderson, (1983), present a simplified but precise description of the geology of the Cornubian peninsula. However little importance is assigned to the thrusting which dominates the stratigraphy, Fig 1.9. Spatially, attention is concentrated on the coastal sections, particularly the northern coast between Bude and Padstow. Most of the major regional structures are discussed and the Tintagel High Strain Zone and the Rusey fault are identified as major tectonic junctions. However, Hobson and Sanderson ascribe very little importance to the role of faulting in general, and thrusting in particular in their regional model. This means that these structures don't receive nearly as much emphasis as the folds and cleavages that occur across the area, which are described and analysed thoroughly and clearly, Fig 1.10. The folding and inversion of the southern

NE

SW

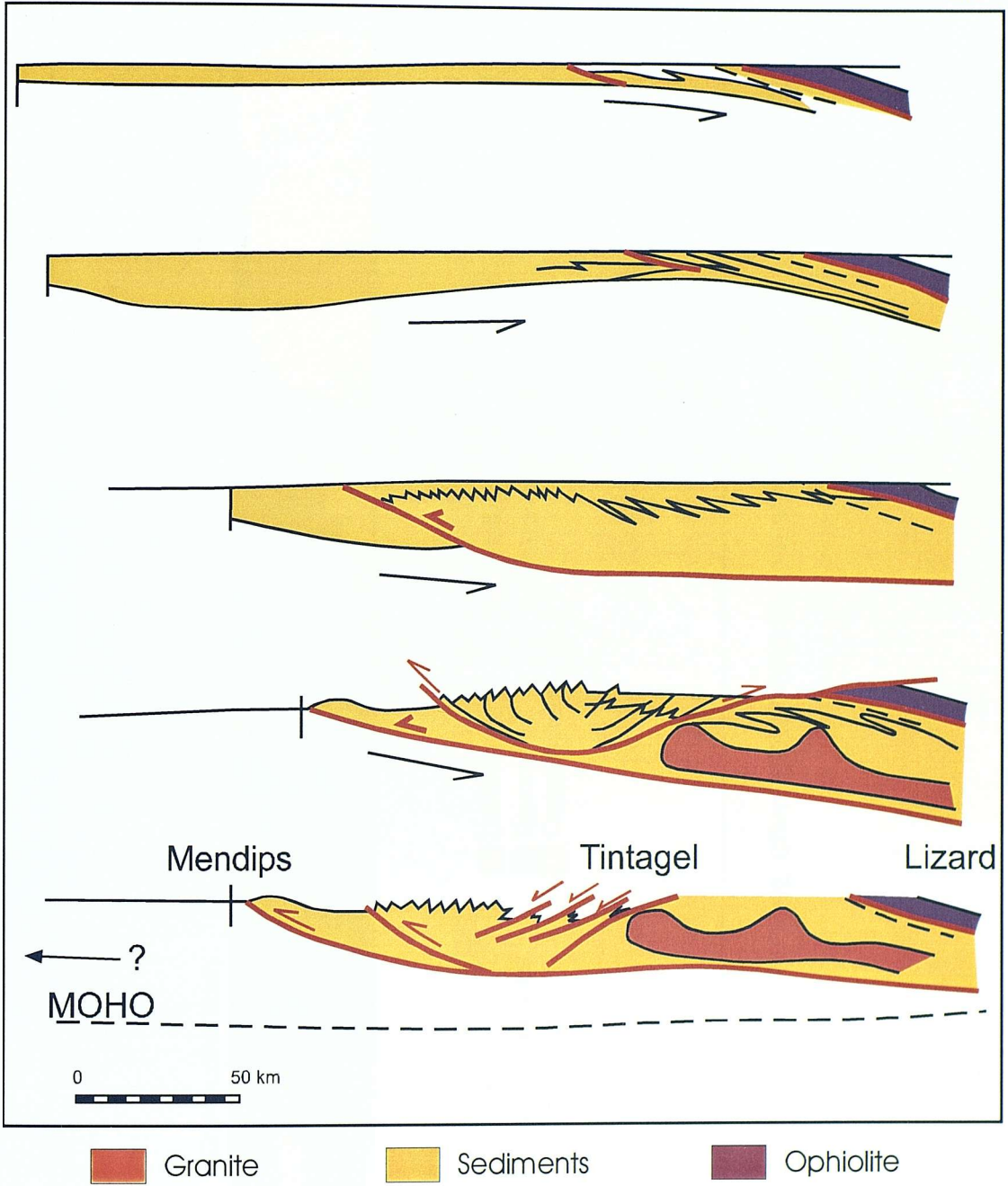


Figure 1.8 Thrust model for the Cornubian Variscides, Shackleton et al. 1982

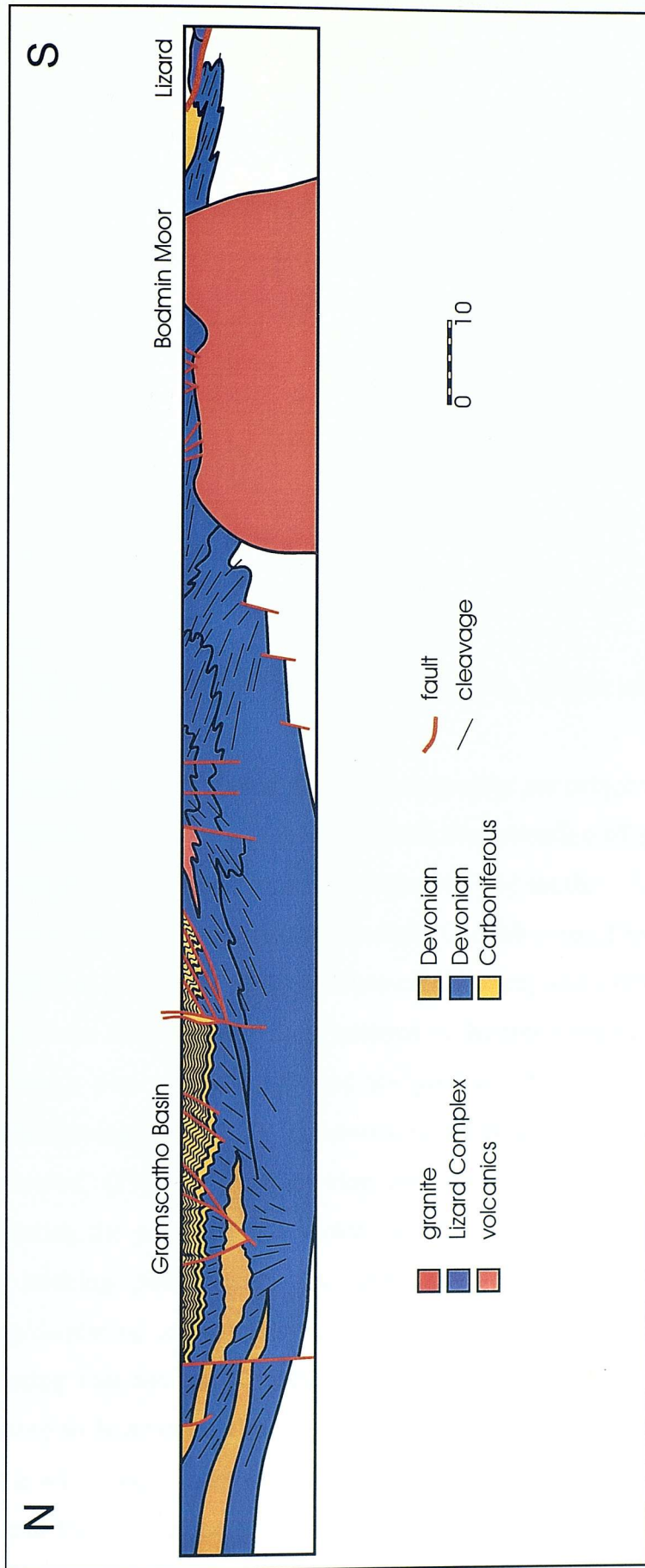


Figure 1.9 Cornubian folds and cleavages, after Hobson and Sanderson, 1983

basins, Gramscatho and Trevone, are assigned to a model of northwards directed inversion and transport on a south dipping low angle thrust zone.

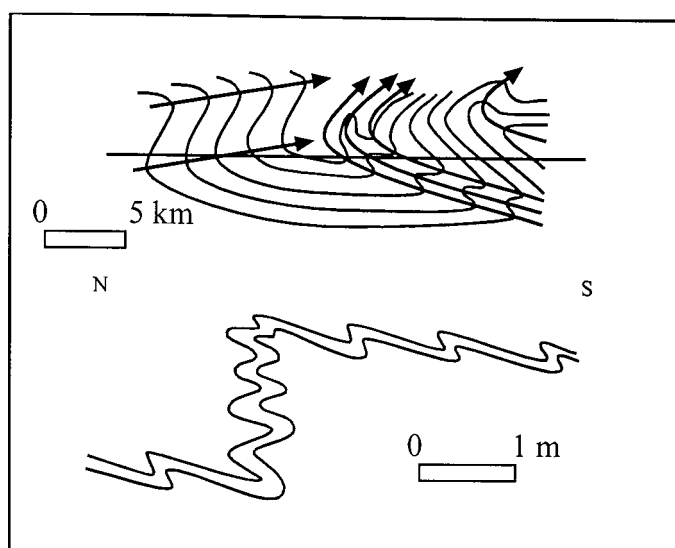


Figure 1.10 Cornubian refolded fold styles, Hobson and Sanderson, 1983

To the north of these basins, the Culm sequences are subjected to bi-directional, (east-west striking), deformation resulting from the formation of a major backthrust system allowing inversion and expulsion to both north and south.

Although not directly relevant to the north Cornish coast, Coward and McClay, (1983), support this model of northerly directed thrusting and elaborate on the diachronous nature of the deformation, which initiated in the south and migrated north. They help to establish a plate tectonic view of the geology of the peninsula by interpreting the regional geology within the framework of a foreland fold and thrust belt. Coward and Smallwood, (1984), further develop this theory with respect to the Culm basin by postulating the presence of a sticking point on their low angle detachment, Fig. 1.11. This sticking point causes the development of the backthrusts and the south facing/displacing structural style. They also introduce the concept of underthrusting, proposing that the rocks to the south have been wedged in under the Culm facies, (credited to Isaac et al. 1982). They again cite Isaac et al. in suggesting that the Rusey fault is a low angle extensional fault which defined the northern limit of underthrusting and that the development of this and other low angle normal faults in the area to be related to a Basin and Range-like period of extension at the end of the Variscan orogeny. Shackleton's 1984 paper on thin skinned tectonics and the Variscan front presents a

definition of the term 'orogenic front', as well as data specific to the Variscides. The flat lying cleavages, very high strains and shortening of the upper crust lead him to identify a

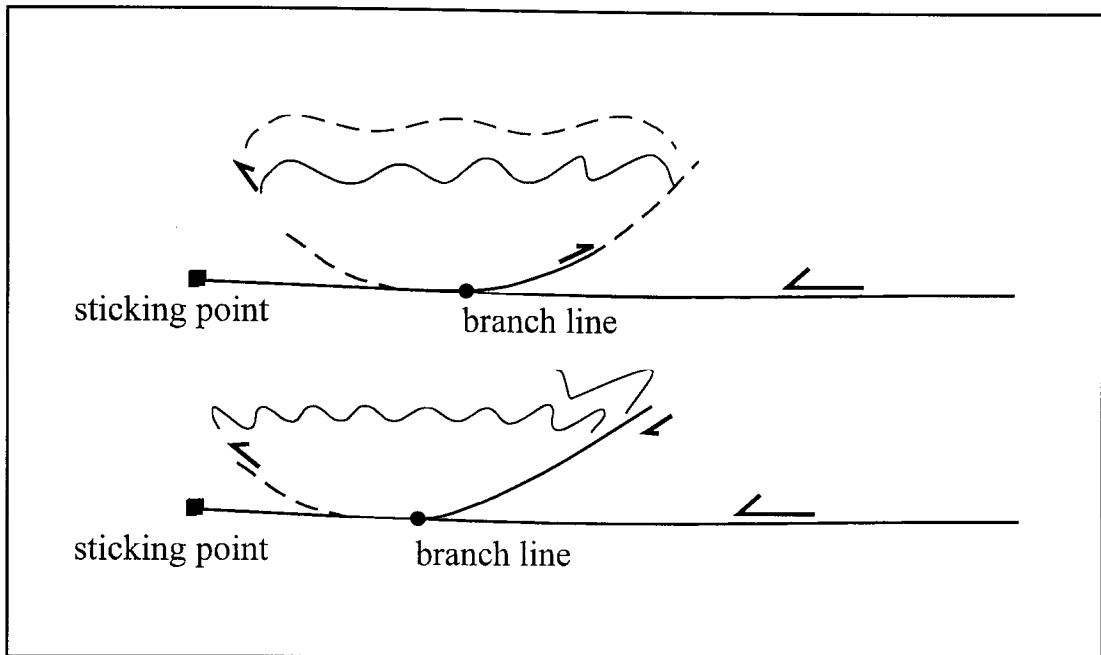


Figure 1.11 Backthrusting resulting from the development of a sticking point on the main detachment, Coward and Smallwood, 1984

geometric need for a basal décollement in the Cornubian Variscides. Such a décollement must be inclined and displacement on it must increase southwards.

The sedimentological data and interpretations proposed in Selwood and Thomas, (1986); describe a series of basins filling, being compressed and inverted and finally having their basin fills expelled by thrusting. This process initiates in the south, thereby creating the sediment supply for the sedimentary fill of a more northerly basin, which is in turn inverted and becomes a sediment source as the deformation front migrates northwards.

Complementing the work of Coward and McClay, (1983), Shackleton, (1984) and Coward and Smallwood, (1984), Barnes and Andrews, (1986), present data on the timing of ophiolite obduction in the Lizard complex, which they fit into a regional plate tectonic context. They propose that the Lizard Complex is a suture zone marking an episode of oceanic closure and continental collision. This collision event would have been responsible for the formation of the thrust belt proposed by Shackleton, Coward and McClay, Coward and Smallwood. The greenstones all have MORB affinities, but are transitional, implying that there was only limited rifting. The felsic rocks are derived from partial melting of continental crust. The isotopic data shows that the magmatic

evolution of the Lizard Complex is similar to that of the present day Red Sea. They place rift initiation in the upper Lower Devonian and obduction in the Lower Carboniferous as there is no high temperature metamorphic sole to the complex. Their structural setting for the development of the complex is similar to the present day San Andreas Fault zone, along which a major pull-apart basin forms the Salton Sea. The Rhenohercynian system is comprised of a dextral megashear, Fig. 1.12, along which small oceanic basins initiated at dextral offsets as pull-apart basins, Fig. 1.13. This would give the NNW sense of opening that is indicated by fieldwork.

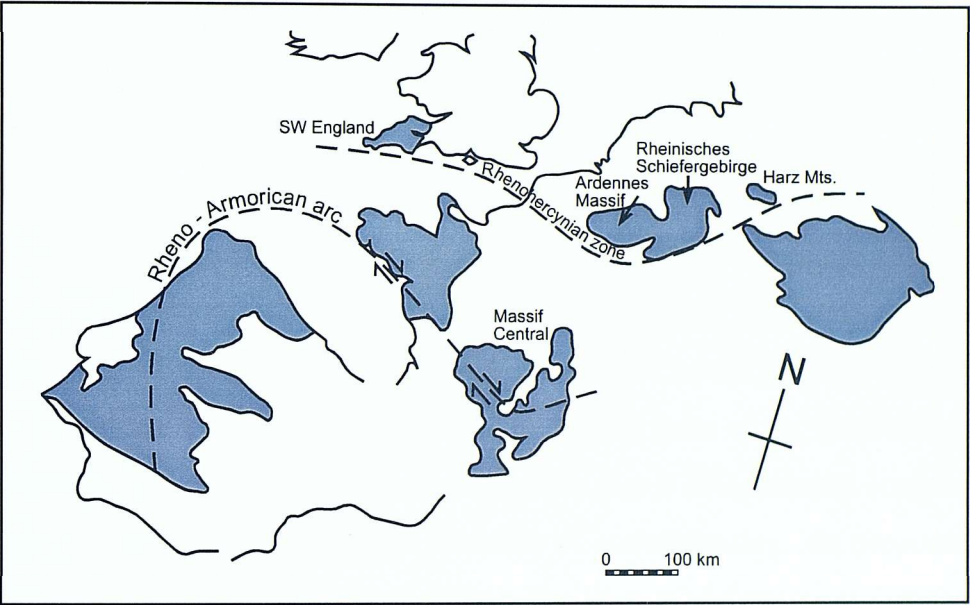


Figure 1.12 The Variscan dextral megashear system, Barnes and Andrews, 1986

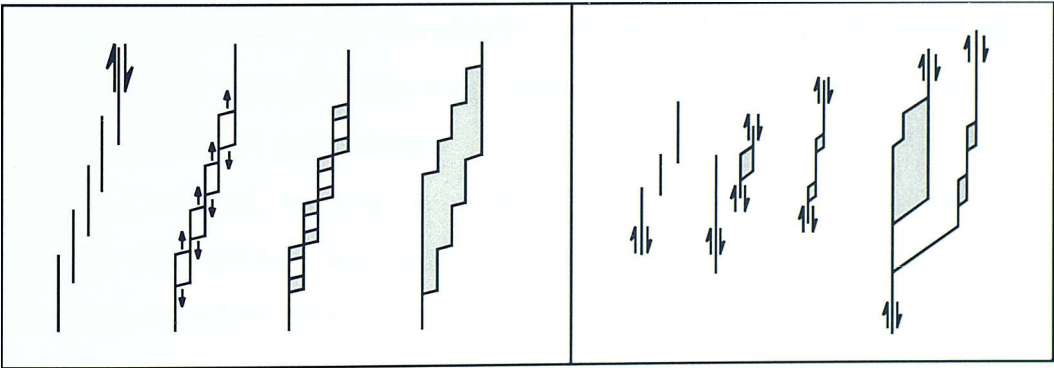


Figure 1.13 Pull apart basins development along dextral strike-slip faults, Aydin and Nur, 1982

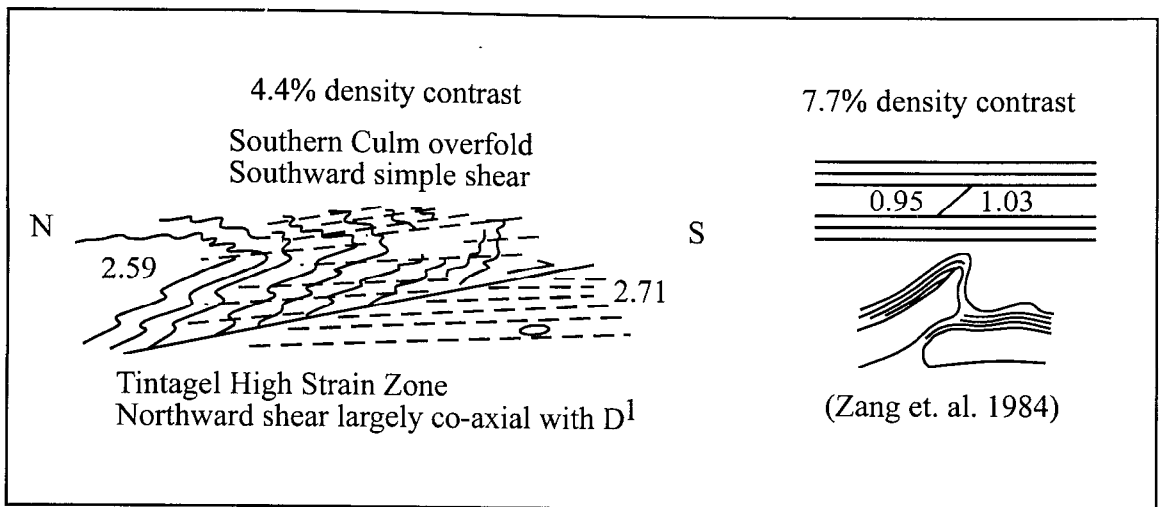


Figure 1.14 Structure of the southern Culm overfold and a model for its development, densities in g/cm³, Warr, 1989, (cited reference should be Zhang 1982)

Warr's 1989 paper on the development of the Davidstow anticline builds on previous work (Zhang, 1982) and gives a model for the progressive deformation of the Tintagel-Boscastle area. Warr describes a history of thrusting to the north that resulted in the underthrusting of Boscastle beds below the Devonian-Carboniferous Culm beds, as a result of significant density contrasts between the two units of the Boscastle and Culm successions, Fig. 1.14. Although the text specifies that it is the density contrast and not the absolute values that controls the initiation of underthrusting, the proposed density contrast of 4% is a very small, (Zang, 1984). The Warr model proposes that early, (315-300 Ma), syn-sedimentary thrusting created south facing folds within the Culm succession. Continued shortening, (300-295 Ma) initiated underthrusting towards the northwest and formed the Davidstow anticline. This was accompanied by an intensification of strain to create the Tintagel high strain zone, modifying the earliest folds into sheath folds and causing overfolding and backthrusting in the Culm sediments. Continued wedging caused late F₂ folding as the Tintagel high strain zone locked up. Displacement was then transferred down section to 'a lower décollement surface, passing below the Culm basin'. The only reason given by Warr for the siting of the main backthrust is the density contrast between Boscastle and Culm successions. Pamplin (1990) presents an extremely clear summary diagram for the evolutionary model of the Cornubian Peninsula, Fig. 1.15. Pamplin (1990) proposes that the Tintagel High Strain Zone is equivalent to a zone of distributed south directed backthrusting which is found further inland and that the concentration into a single zone is the result of a lateral thrust ramp at depth.

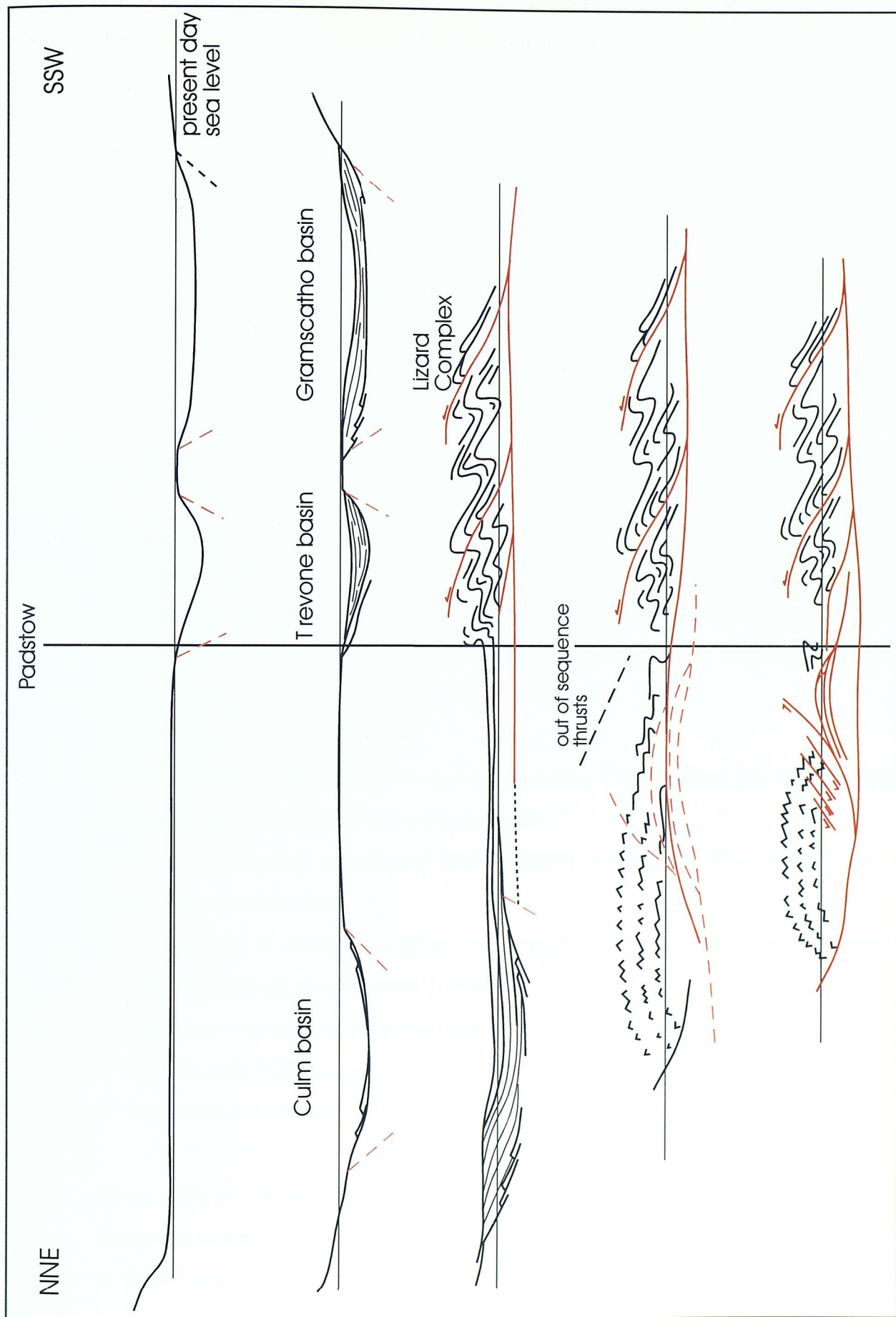


Figure 1.15 Variscan evolution of the Cornubian peninsula, Pamplin, 1990

The model presented by Pamplin, (Fig. 1.15), proposes, without going into great detail, a model similar to that generally suggested for the early development of the Cornubian peninsula, which involves the inversion of the Trevone and Gramscatho basins and development of the Culm Basin. Inversion of the Culm Basin occurs from both north and south, involving the Rusey backthrust, with folds developing well to the south, creating an extremely asymmetric basin profile. This was followed by telescoping of the southern part of the Culm Basin by out of sequence thrusting, which produced the Tintagel High Strain Zone. Although the summary diagram is comprehensive, the text goes little beyond the formation of the Tintagel High Strain Zone and the later extensional faulting is not discussed.

The last phase of Variscan related deformation is examined by Shail and Wilkinson, (1994). They conclude that the extension of Variscan basement must have been initiated no later than the Carboniferous i.e. by approximately 290/300 Ma. Low angle faulting associated with granite intrusion occurred at 280 Ma. There was repeated extensional reactivation of NNE and NNW Variscan basement wrench faults during the development of the post-Carboniferous Western Approaches basin. Shail and Wilkinson associate this basin development with lithospheric weakening related to orogenic collapse.

Finally Le Gall's 1990 paper is as good a summary of Cornubian geology as Matte's 1986 paper is of the geology of the entire orogen.

Le Gall drew several conclusions from structure mapped on deep seismic and its correlation with field data;

- that the area is sliced by a series of low-angle reflectors, interpreted as Variscan thrusts, which are south dipping, northwards directed structures.
- that these structures are overstepped by the north dipping Tintagel backthrusts.
- that this imbricated crustal stack is overridden by the Lizard complex, an ophiolitic suture that is rooted deep beneath the Moho.

Three units are defined for the crust of southwestern England. They are, in order of increasing depth;

- thick Upper Paleo/Meso/Cenozoic sedimentary basins:
- Paleozoic/pre-Cambrian basement, consisting of relatively transparent upper crust cut by low dipping reflectors:

- layered lower crust, along the Moho, above seismically featureless upper mantle.

The flatness of the lower crust lead Le Gall to suggest that any perturbations on the Moho have been re-equilibrated and the present Moho has no direct relationship to the structures above it. The lack of deep basement slices suggest that the thrust zones flattened out at depth, probably into a single décollement zone, forming a deep crustal imbricate stack and therefore may never have intersected the Moho.

Le Gall argues that the majority of the shortening must have been accommodated by underthrusting, causing a downward deflection of the overthickened crust which has

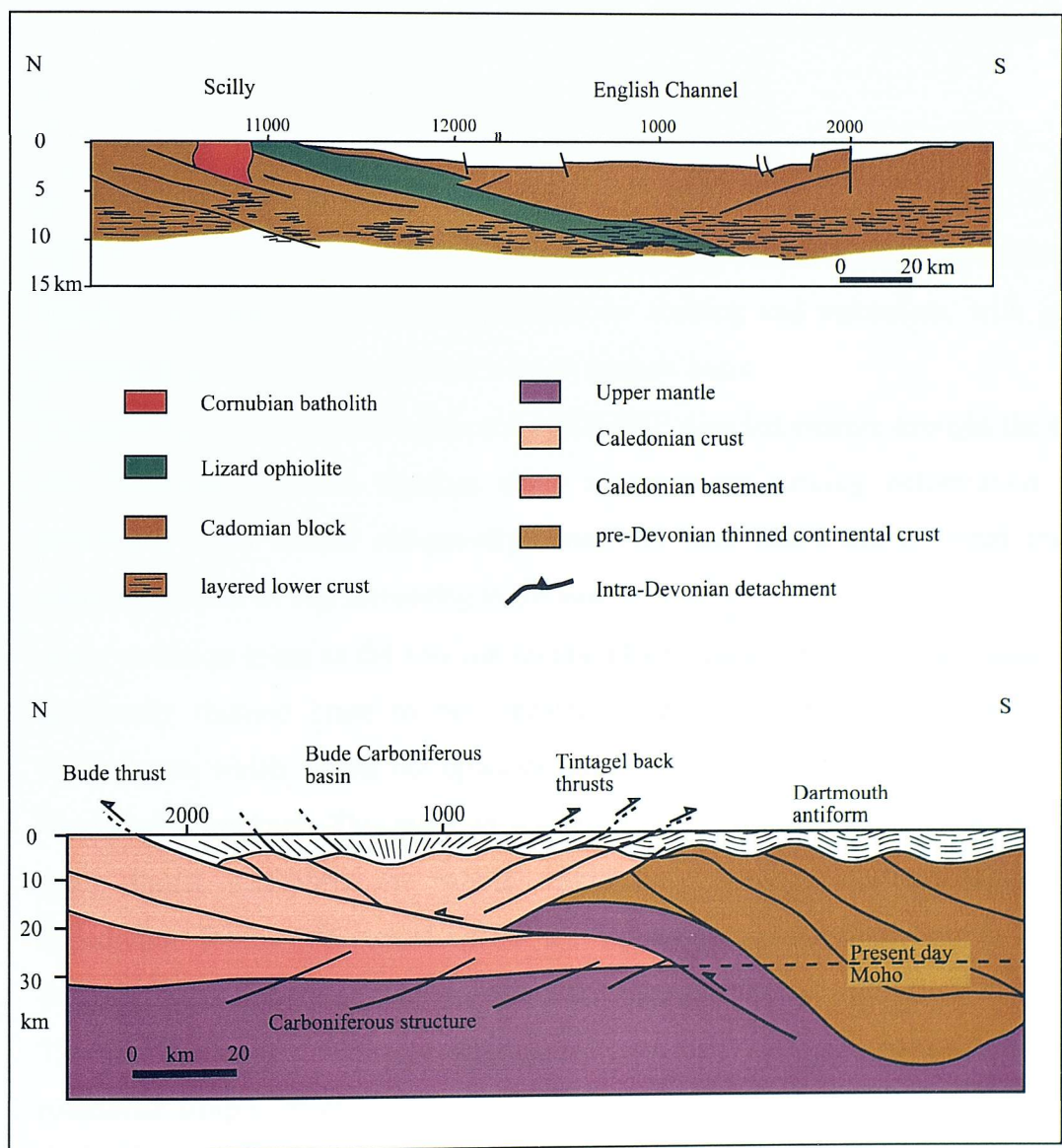


Figure 1.16 Cross sections of the Cornubian peninsula, from deep seismic sections, Le Gall, 1990

since been assimilated by the layered lower crust. This created both a thick skinned crustal duplex and a thin-skinned series of inverted basins, bounded to the north by the Bude thrust, Fig. 1.16, which subcrops along the edge of the Bristol Channel High.

Granite development is limited to those areas associated with deep-rooted thrusts, but the layered lower crust can clearly be seen underlying the plutons. Accordingly Le Gall interprets them as shallow sourced plutons. On evidence from gravimetric maps, wrench faults are interpreted which are oriented appropriately, (NW-SE striking), to have accommodated dextral strike slip motion. The Lizard suture is interpreted as a major south dipping thrust zone and is considered to be a Variscan suture of Upper Carboniferous age. This dating is based on the presence of Muscovian age breccias in the footwall to the Lizard Thrust, which indicates that the Lizard thrust is younger than most of the Variscan thrusting to the north.

In summary, the model proposed in this paper is

- extension in Devonian times in an open shelf setting, producing basinal sedimentary successions. There is sufficient extension for melting and volcanism, with greater stretching further south to produce a small oceanic basin.
- Obduction of the Lizard complex and NNW/SSE directed closure brought the Celtic and Continental blocks together along an east-west striking deformation zone, creating a major dextral oblique-slip zone. This zone had a displacement gradient with the amount of slip increasing to the east.
- major collision event as the two continental blocks continue to converge, causing the previously thinned crust to be repeatedly stacked at depth on south dipping underthrusts which flatten out upwards onto a flat lying roof thrust, lying under an 8km thick shear zone. This shear zone extends all the way under the Culm basin. The Bude thrust transports the whole sedimentary package northwards over the Caledonide foreland. External basins are buttressed against their bounding faults, forming tight, upright structures, whilst the more southerly, open shelf basins, e.g. Trevone/Gramscatho basins undergo translation along deep décollement and suffer rotational simple shear, creating flat lying cleavages, sheath folds and flat shear planes.

To the east a very dense network of deep northwards transporting imbricate thrusts is identified. Melting associated with this crustal thickening produced granite in the south

the Tintagel High Strain Zone. Further west the upper crust exhibits a more diffuse style of deformation, with more widely spaced thrusts, which all transport northwards. This agrees with the conclusions drawn by Pamplin, 1990. Le Gall interprets this to be associated with a large scale change in strike of the thrust planes, from 070° in the west to 100° in the east. The faults with 100° strike accommodate large scale dextral strike-slip during the later stages of the orogen.

1.3.2 The sedimentary framework

Hartley and Warr, (1990), present a regional description of basin evolution across the Cornubian peninsula. The stratigraphic record, as described by them, places the major extension in the mid-Lower Devonian with the majority of the sediments as syn-rift deposits, Fig. 1.17a. The syn-tectonic setting is supported by the presence of alkali basalt volcanics within the sediment pile. They place the D₁ deformation as Namurian.

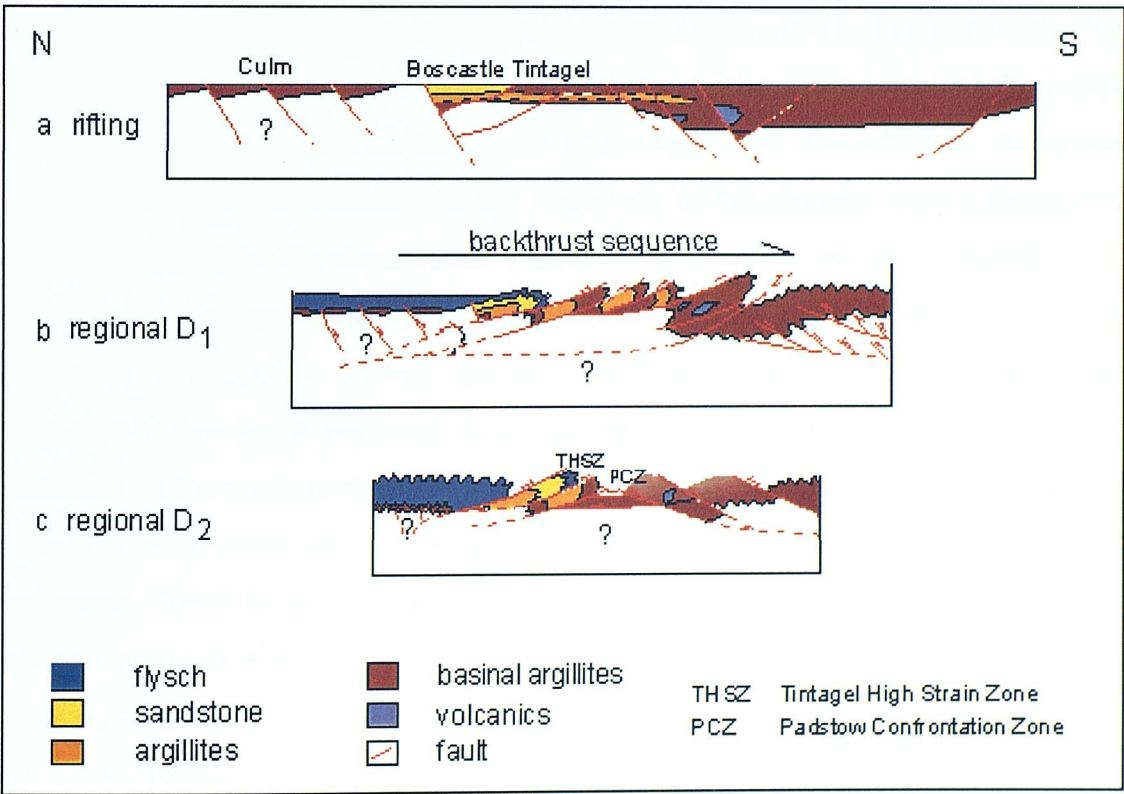


Figure 1.17 Tectono-sedimentary framework for the Cornubian basins inverted in the Variscan orogeny, Hartley and Warr, 1990

This deformation caused the inversion of the Trevone basin and the sedimentary successions. During and post D_1 the Culm Basin was filled with the proximal turbidites of the Crackington Formation, Fig. 1.17b.

Paleocurrent data indicate that drainage of the Culm Basin was axial, i.e. running east-west. The overlying deltaic sandstones of the Bideford Formation were sourced from the north and were succeeded by proximal turbidites and marine/lacustrine sandstones, also northerly sourced. D_2 caused dismemberment of F_1 by out of sequence thrusting, the development of lateral ramps and the occurrence of oblique slip thrust activity. The southern edge of the Culm Basin was underthrust and the distinctive bi-directional expulsion of the Culm Basin sediments achieved, Fig. 1.17c. Hartley and Warr conclude that both the sedimentation and the deformation were very strongly controlled by pre-Variscan structure.

This general theme is continued by Selwood, (1990), who reviewed basin development in central southwest England. He argued that Devonian sedimentation was initiated on continental crust forming the southernmost extension of the Caledonian continent. Most of the sediments are shelf basin and rise facies, within a half graben setting, with the inner (continental) shelf to the north and the outer (seaward) shelf to the south, Fig. 1.18a. Selwood describes facies changes in support of an east-west trending fault system, which controls the northern edge of the basin and olistostromes, slump and conglomeratic facies as evidence for the steepening of the southern margin throughout the Upper Devonian. The depositional system on the southern margin is limited to the north by a fault controlled carbonate rise, which overlooked the deepening Culm basin. This basin development starved the outer shelf and allowed extensive carbonate complexes to develop. A sea level rise during the early Carboniferous produced basinal conditions on the northern shelf and paralic facies in the south, Fig. 1.18b. These facies indicate a southwards source, being ponded by the carbonates before prograding northwards. Flysch facies covered the area by the Namurian.

Selwood proposes a model whereby the Culm basin was inverted with the original bounding fault acting as a buttress. This caused the basin fill to be expelled to the south, Fig. 1.18c, producing southerly directed thrusts and northerly directed shortcuts. D_2 overrode this with northerly transporting thrusts and finally D_3 produced gently north dipping extensional faults. In this model, the original basin architecture controls the final inverted structure.

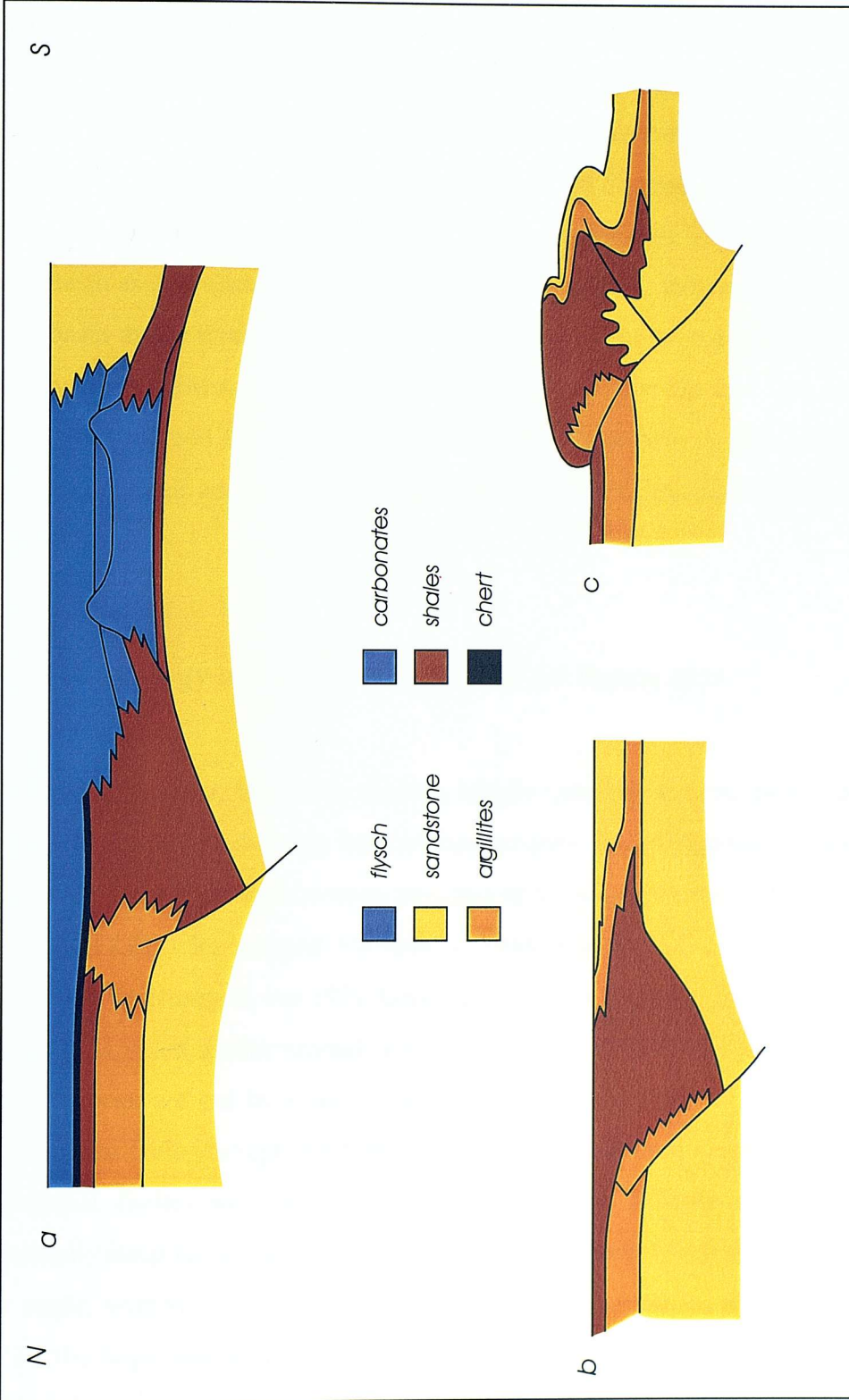


Figure 1.18

a: sedimentary framework for the Trevone basin
 b: sedimentary framework for the Culm basin
 c: schematic diagram of the deformation style for the Culm basin, after Selwood, 1990

Hecht, (1992), looks specifically at the development of the Culm basin, starting with a Dinantian deep water stage, during which black shales, carbonates and cherts were deposited and pyrite developed. These pass upwards into the Namurian Crackington Formation, comprising proximal turbidites shed into a deep east-west trending basin, then up into the laterally equivalent Bude and Bideford Formations, deposited during the Westphalian. They consist of shallow, storm influenced lake deposits and fluvio-deltaic sediments. The Bude Formation shows significant slumping and mass debris flows, increasing in frequency upwards, whilst the Bideford Formation indicates slow, steady subsidence. Hecht cites the subsidence path as good evidence for the development of the Culm Basin as a foreland basin, with an initially slow start, followed by rapid deepening of the basin as the thrust front approached. The sediment source is placed to the north, possibly from a Bristol Channel landmass, supported by the fact that similar Welsh deposits are sourced from the south. The soft sediment deformation indicates an east-west striking slope and basin axis, with the slumps and debris flows being derived from the north.

1.4 The geology of the Culm Basin and the Rusey area.

In a regional context, the Rusey fault is largely ignored, or used merely as a division between basins inverted during the Variscan orogeny. Interpretations of the nature of the fault vary from a high angle reverse structure to a low angle extensional structure. E. C. Freshney, BGS, who mapped the area in 1964, described a number of faults in the promontory at Rusey in the 1972 Geological Survey Memoir, Fig. 1.19. A low angle fault is interpreted as extensional and called the Rusey fault. This fault zone, and the Rusey breccia, are cut by a sub-vertical dextral wrench fault, as well as a number of steep normal faults, though the term steep appears to be used more to distinguish these extensional faults from the low angle one than to indicate that these faults are abnormally steep for an extensional fault. Freshney generally refers to the Rusey fault as low angle, without any definite sense of movement across it, stating that, (page 48, 1972), 'the large, low angle Rusey fault is associated with a sharp southward increase in deformation and a small southward increase in metamorphic grade. This suggests that the rocks to the south of the Rusey fault, although partly of the same age as those to the north, were brought up from a much deeper tectonic level, presumably during the low

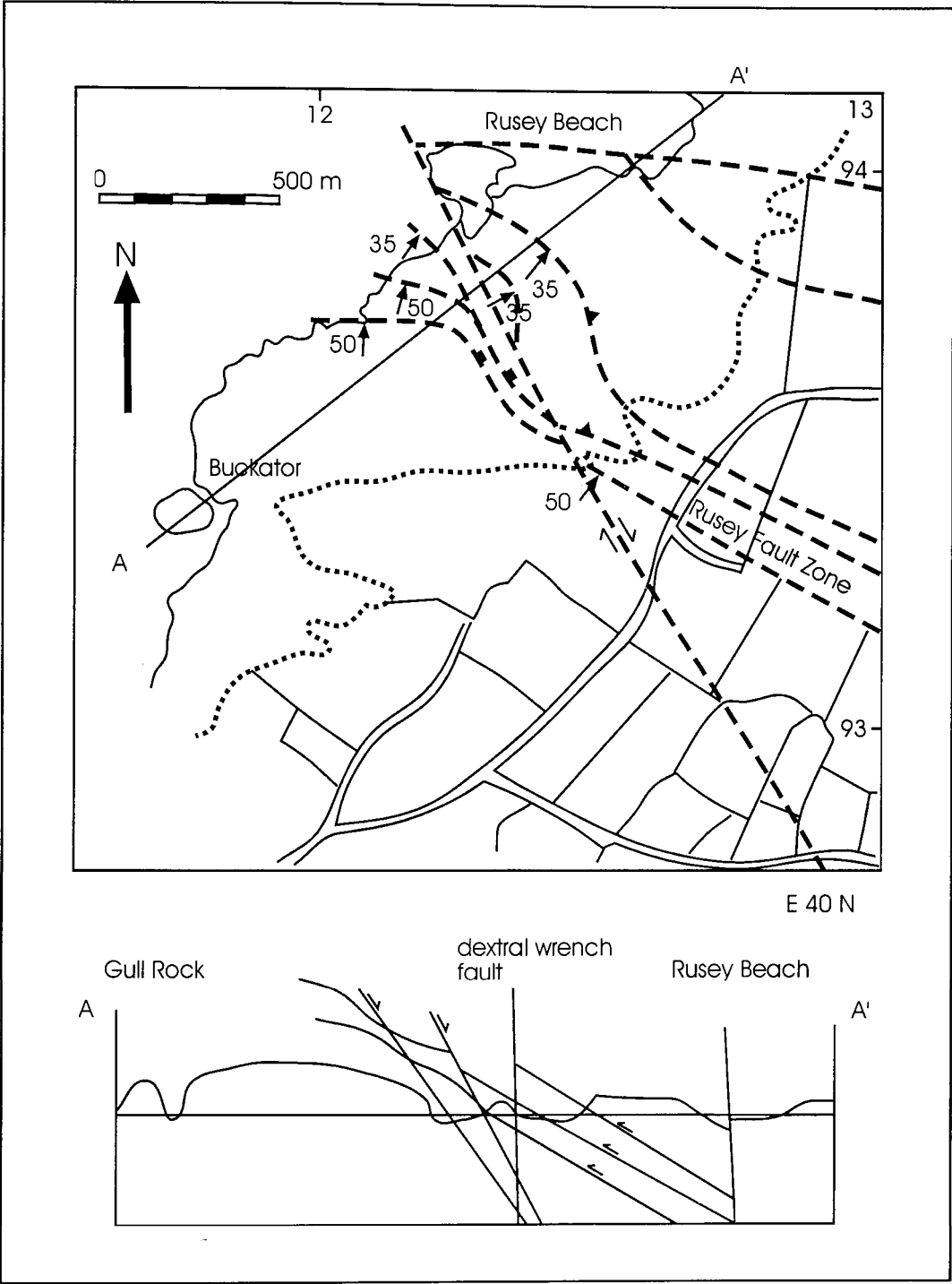


Figure 1.19 Map and section of the Rusey headland, north Cornwall after Freshney et al, BGS memoir 1972

angle normal faulting', and also that 'probably they, (i.e. the faults), acted as thrusts during the formation of the major southward facing recumbent folds and later as low angle normal faults.', (page 50). In the description of the Rusey outcrops Freshney never ascribes a sense of movement to the fault that he describes as representing 'a zone where repeated movement has taken place' and he uses descriptions such as 'on its northern side' (page 60/62, 1972) rather than in its hanging/footwall. He describes the breccia thus; 'bands of brecciated and rewelded powdered quartz are set in a sheared mass of black sooty shales with scattered chert bands and siltstones locally containing crinoid fossils', (page 60, 1972). He also describes steep, east-west trending large normal faults that cut the succession to the north of the breccia outcrop. Freshney et al. (1966), describe early south-facing folds, overturned to the south from Bude, beyond Tintagel and further south. These are assigned to a reverse slip phase, displacing top towards the south. Freshney et al. describe a phase of late, north-verging 'ruck' folds and a general sense of overthrusting to the north. From both the text and diagrams it is clear that the Rusey fault is being represented as a major boundary.

Since then field studies of Rusey are scarce, although Freshney's work is invariably cited by later workers.

The very earliest stages of Culm basin deformation are described in Enfield et al, (1985). They looked at high level sediments, the Bude and Crackington Formations, which underwent deformation from pre-lithification stage. They propose early northerly directed syn-sedimentary deformation, followed by late south verging back thrusting. They also describe sediment transport east-west, i.e. axial to the basin with sediments sourced from the north.

One of the most striking features of the upper part of the Culm succession is the frequent presence of slumping at several horizons. This deformation initially indicates collapse down a south dipping slope. This subsequently reverses to become northerly directed, indicating the arrival of the Variscan deformation front from the south and the onset of northerly directed thrusting. Enfield et al. propose that the main thrust underlying the Culm basin then sticks and initiates back thrusting. This caused the overprinting of north directed thrusts with south directed back thrusts. They assign the slump beds a degree of control over the location of thrust planes, being horizons that have been kept weak by elevated pore fluid pressures. However, their diagrams show the thrust running along a lithological junction some distance above the slump beds.

Whalley and Lloyd, (1986), present work from the same area, Fig. 1.20a and b, maintaining that thrusting occurred from the later stages of sediment deposition in the form of sediment slumps, providing early décollement. This is followed by thrusting, with subsequent kinking and folding of the thrust planes into chevron folds. They interpret these structures as initiating as classic sedimentary slumps, with sedimentary and tectonic deformation occurring as a continuum. As slumps slip down slope in response to slope instabilities the manner in which these horizons became, through continuous progressive deformation, reverse slip horizons is unclear. There is widespread evidence for high pore fluid pressures (in the form of fluidised beds and

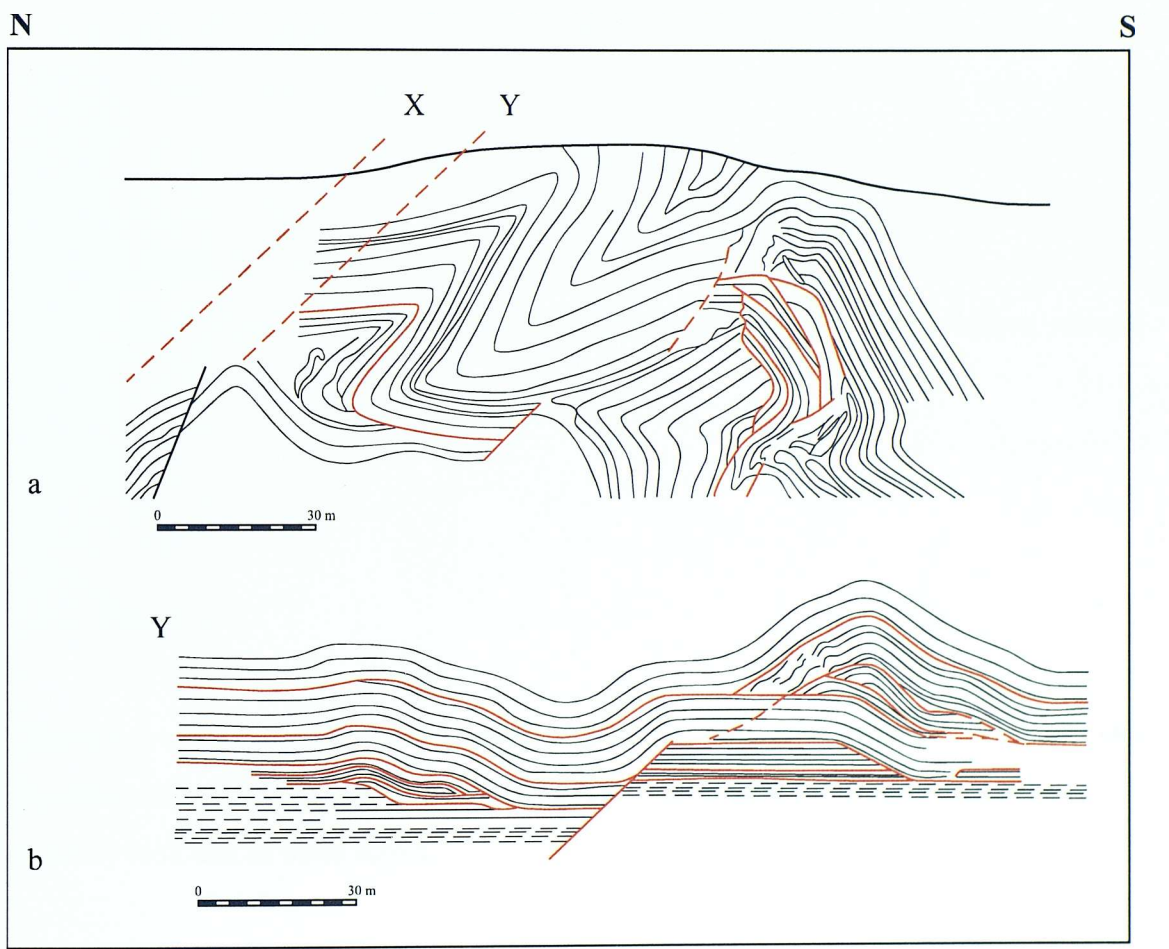


Figure 1.20 Restoration of slumped, folded and thrust beds, Bude,
Whalley and Lloyd, 1986

sand volcanoes) from a very early stage in the geological history. The northerly directed thrusting is later overprinted by southerly directed thrusting as backthrust geometries become active. Whalley and Lloyd conclude that there is a history of continuous

deformation from the final stages of deposition through to complete inversion of the Culm basin, with the southerly transporting thrusts occurring in the final stage.

Mapeo and Andrews, (1991), described complex deformation in the Bude Formation, comprising of synchronous extensional and thrust faulting, Fig. 1.21. They explain the presence of extensional faults as a geometric necessity from the occurrence of overlapping thrust strands or of the thrust cutting down section. This is not the earliest phase of deformation, but it is pre-folding and so forms an early part of D₁. The relationship is supported by the fact that the direction of transport on the thrust is parallel to the direction of transport on the extensional faults. A small number of the thrusts have been deformed by an extensional fault.

R. Jackson's 1991 paper relates small-scale structure to the larger tectonic regime, using vein arrays. He deduces that σ_1 was oriented northwest-southeast during early veining and fold amplification and therefore suggests that transpression was created by the imposition of oblique closure on an east - west basin margin. The transpression would be partitioned into horizontal shortening across the basin and dextral strike slip at the basin margins, reactivating the basin bounding faults as high angle oblique reverse faults.

1.5 Summary

This review of the literature has allowed the construction of a regional tectonic model consisting of a convergent event in the area that is now central Europe and Iberia, which probably took one of three forms.

1. simultaneous subduction of two oceans, the northern Rheic Ocean and the southern Massif Central Ocean. The former was no more than 80km wide, whilst the southern ocean was considerably wider, being 2-3000km wide prior to the onset of ocean closure. Closure of these oceans produced the double line of ophiolites seen in east and central Europe, with the Cornish Lizard ophiolite

2. closure of a single southern ocean, the Massif Central ocean, with initially normal closure being deflected into oblique closure as continental collision begins. Closure induced dextral strike slip along the northern margin of the

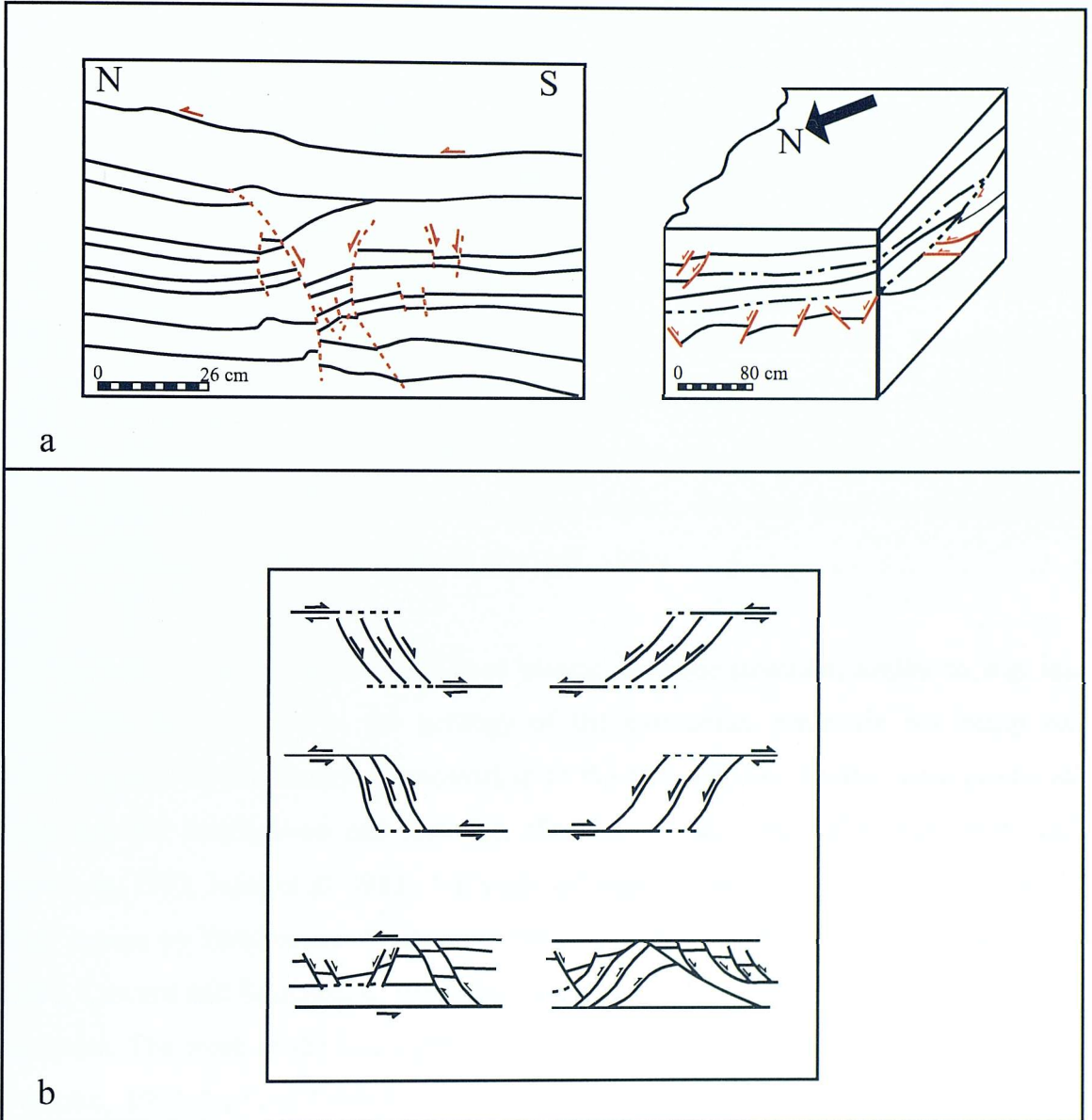


Figure 1.21 Relationships between thrusting and extension, Mapeo and Andrews, 1991

orogen and in Iberia, resulting in pull apart basins, several of which were sufficiently well developed to be floored by oceanic crust. Continued collision caused basin closure and obduction to form the northern belt of small, highly dismembered

ophiolites, (Holder and Leveridge, 1986, Scotese et al, 1979, Barnes and Andrews 1988).

3. Closure of a single large ocean with normal convergence from the south, but with the development of back arc basins along its northern margin, creating the northern ophiolites, (Van der Voo, 1982).

After ocean closure, the orogen progressed in the Cornubian region by the inversion of a series of foreland basins created and inverted on the edge of the Caledonian continent. These basins were opened with east west trending axes and were closed by east west striking thrusts, transporting to the north. Jackson, 1991 presents evidence for dextral shear from the syn-orogenic vein orientations, with σ_1 oriented NW-SE. Transpression was partitioned into dextral strike-slip along basin boundaries and a north-south compression within the basins and along major thrusts. Mid-crustal shortening was achieved by the development of an intracrustal duplex, detached from the thin skinned thrusting above by a major décollement, (Le Gall, 1990).

It is perhaps the lack of a clearly defined 'classic' orogenic structure, similar to, e.g. the Himalayas that resulted in the geology of the Cornubian peninsula not being set explicitly into a plate tectonic framework until the early eighties. Earlier work produced very detailed descriptions and histories, (Freshney et al. 1966, 1972, Sanderson and Dearman, 1973, Isaac et al. 1982), but a unified regional framework was not clear until, with papers by Hobson and Sanderson, 1983, Coward and McClay, 1983, Shackleton 1984, Coward and Smallwood, 1984, the field data became considered in terms of plate tectonics. The work of the late eighties and early nineties, (Barnes and Andrews, 1986, Pamplin, 1990, Le Gall 1990, Hartley and Warr 1990, Selwood, 1990 and Shail and Wilkinson, 1994), has established a model which is coherent though still incomplete, particularly as regards the form of the oceanic distribution at the beginning of the orogeny.

The framework within which the Rusey fault system is examined is therefore one of thrusting and inversion of basins in the external parts of the northwards propagating half of an orogen centred in mid-Europe. The thesis area is only tens of kilometres from the northern front of the orogen. During and after basin inversion, the area was subjected to

both dextral strike-slip and normal convergence. Renewed north-south extension occurred during orogenic collapse.

Chapter Two

The Structure of Rusey Headland

2.1 Introduction

The framework within which the Rusey fault system is examined has been described in detail in the previous chapter. It is one of thrusting and inversion of sedimentary basins in the external parts of the northwards propagating, east-west trending Variscan orogen centred in mid-Europe. The thesis area lies only tens of kilometres south of the northern front of the orogen. This study concentrates on the thrust which many workers have used to define the southern margin of the Culm basin, the Rusey Thrust. Of the four basins that form part of the northern margin of the Variscan belt in south-west England, the Culm basin is unique in having bi-directional inversion of its sediment fill and the Rusey thrust has been interpreted as a basin bounding regional backthrust, displacing top to the south (Andrews et al. 1988, Warr 1989, Pamplin, 1990) Fig. 2.1.

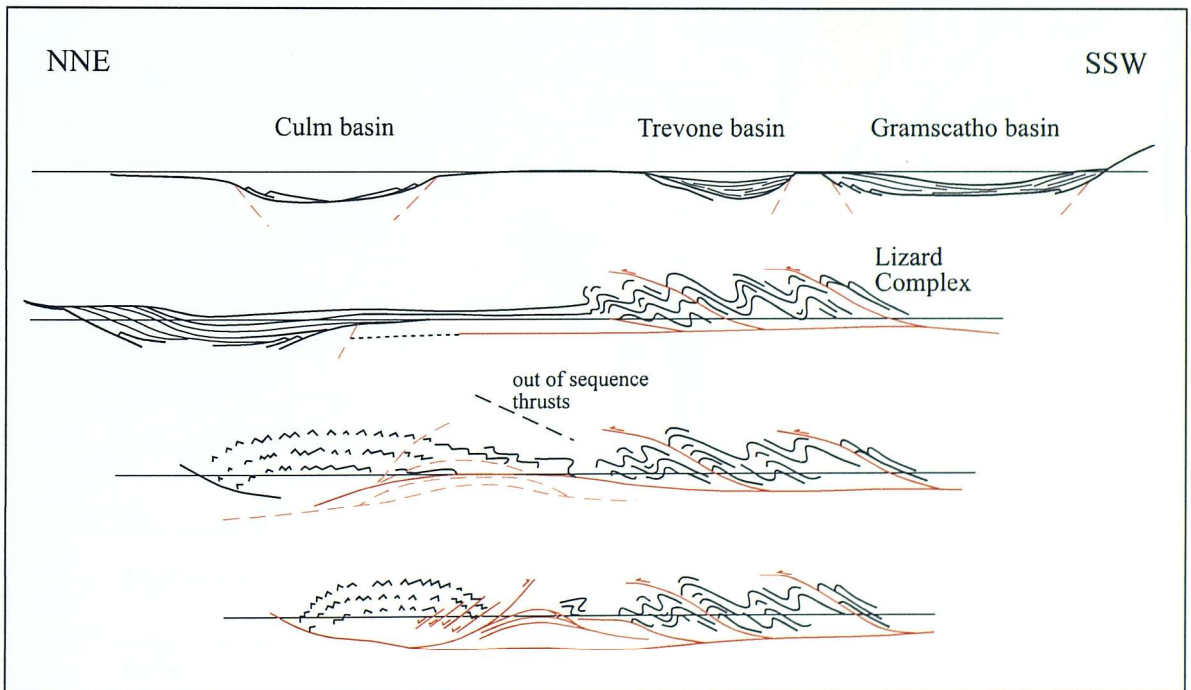


Figure 2.1 Evolution of the Cornubian basins, Pamplin, 1990

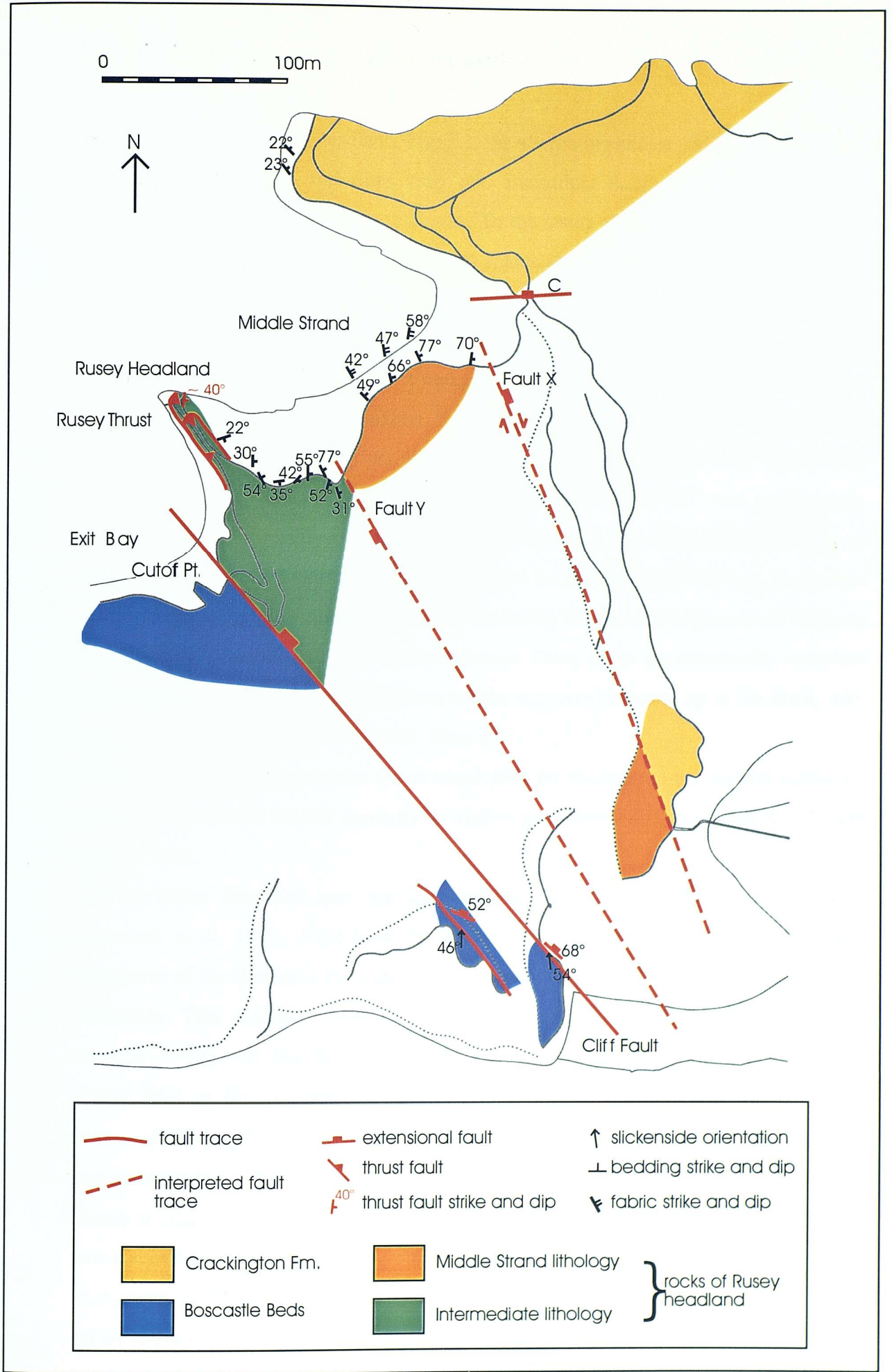


Figure 2.2. Geological map of the Rusey area, North Cornwall.
(Slip sense on interpreted faults is deduced from bedding orientations)

2.2 The Tectono-sedimentary Framework

To the north of the Rusey headland, Fig. 2.2, the alternating shales and sandstones of the Crackington Formation turbidites crop out. Individual sandstone beds within this formation may exceed a metre in thickness. To the south of the headland, the Boscastle Beds, which consist of blue-black silts and shales and volcanics, crop out. A zone of tectonic disturbance approximately 230m wide now separates these two formations, Fig. 2.2, occupying the exposures at Middle Strand and Rusey Headland. The 130m wide section exposed at Middle Strand, is composed of uniform mid-grey shales, with iron rich bands 2 to 3cm thick. In contrast, the 100m wide exposure which covers only the Rusey headland itself, comprises dark blue-grey to black shales with intercalated sandstones, each c. 10cm thick. Bedding strikes 019° and dips 17° - 27° east on the north-eastern side of the headland. On the south-western side the local deformation is so complex that a representative bedding dip cannot be determined, Plate 2.1. An intense planar fabric of highly variable orientation, formed by the parallel alignment of minerals such as mica, is associated with this deformation. These rocks are structurally underlain by shales which contain a band of quartz matrix supported breccia, up to 3m thick, with a mean strike and dip of $162^{\circ}/44^{\circ}$ east, Plate 2.2.

Within the limited exposure, the shales which host the breccias do not contain sandstone layers. They have an intense penetrative mineral alignment fabric, that strikes 016° and dips 59° east.

All the rocks described here are Carboniferous in age and palaeontological data, (Selwood et al. 1985) place both the Crackington Formation and the northernmost exposures of the Boscastle Formation, immediately to the south of the headland, in the Namurian. This coincides with the interpretation by E.C. Freshney (page 22, B.G.S. Memoir, 1972) who places the rocks both to the north and to the south west of the Rusey fault in the H_{1a} and E_{2c} goniatite zones respectively, dating these rocks as Namurian. From 91m south west of the Rusey Fault further south to Boscastle the rocks are assumed to be of Namurian age, though an occurrence of cherts two miles south of Rusey is suggested to be possibly Visean in age, again by Freshney, 1972. The sources cited above do not state explicitly that samples were taken from the zone of tectonic disruption. Therefore, whilst it is clear that the rocks immediately to the north and south of the tectonised zone are of the same age, this may not be true for the rocks within the



Plate 2.1 The deformation surrounding the main thrust plane and fault zone A. The complexity and difficulty of access can be seen. This plate can be compared with the field sketch of the lower half of fault zone A, just above its junction with the main thrust plane, figure 2.12.

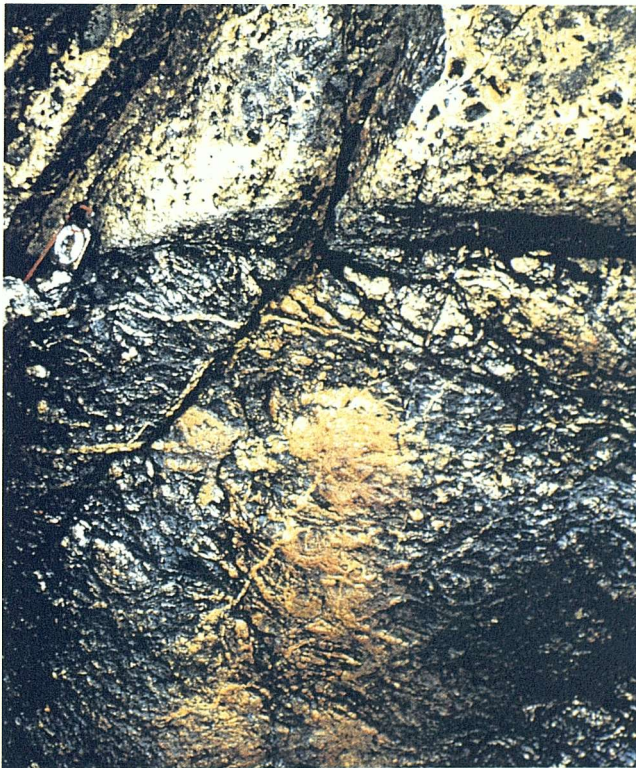


Plate 2.2 Shale outcropping below the main breccia band, showing an intensively developed fabric, the plane of which is inclined to the base of the breccia. The fabric curves asymptotically to the breccia base. The plane of the fabric dips with a very similar strike to the thrust but more steeply, implying that it is genetically related to thrust development.

zone. It is interesting to note, however, that although the Boscastle and Crackington formations are separated by a major dislocation zone, the beds within that zone are lithologically intermediate between the distal facies of the Boscastle beds and the more proximal facies of the Crackington Formation.

Facies analysis (Hecht, 1992) indicates a northern source, supplying material for the Crackington Formation turbidite flows into the basin from a Bristol Channel landmass (section 1.3:2). The Crackington Formation is made up of medial turbidites and contains well developed AB(C) sections of the Bouma sequence, Fig. 2.3. The Rusey beds are an intermediate sequence, classified as a CDE section with a high percentage of C units. The rocks are interbedded shales and sandstones, largely shales, with thin decimetre scale sandstones, now quartzites. The Boscastle Beds are generally considered to be sourced from the south and are thought to represent flysch deposits from the approaching thrust front (Freshney et. al., 1972, Selwood and Thomas, 1986, Selwood, 1990).

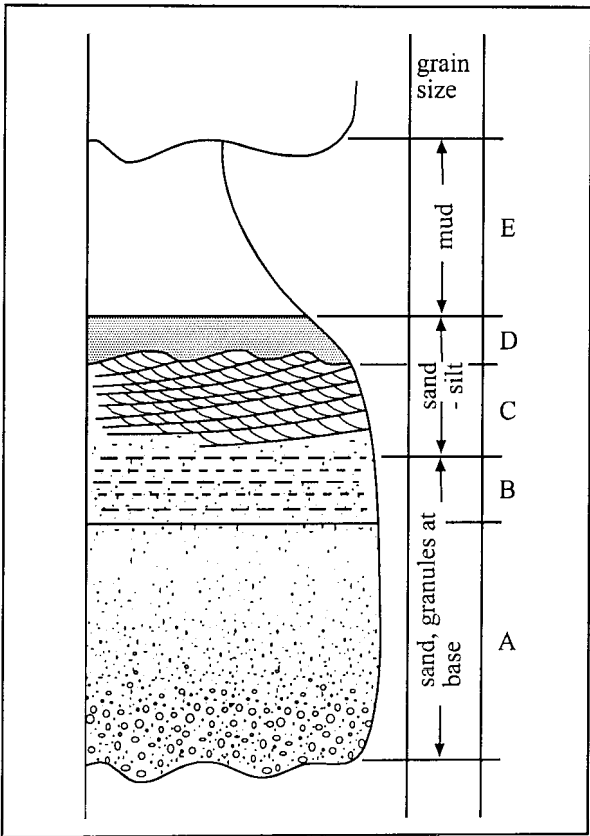


Figure 2.3 The Bouma sequence classification for turbidite flow deposits, after Collinson and Thompson, 1989, (originally Bouma, 1969)

From the palaeontological evidence referenced above it can be seen that the Crackington and Boscastle Formations are contemporary and therefore lateral equivalents. The tectono-sedimentary framework is seen as compatible with the sedimentary fill expected to develop within a half graben, Fig. 2.4. The Rusey lithologies, i.e. those lithologies which are exposed in the Middle Strand and Rusey Headland outcrops, are part of a fault-bounded sequence, the original thickness of which cannot be ascertained. However from their lithology they are transitional between and may originally have been comparable in thickness to the Crackington and Boscastle beds. The current areal extent of the Boscastle and Crackington Formations (they outcrop as uniform units over several miles of coastline), implies that the Culm Basin enjoyed a uniform depositional environment over a large area.

The fault bounded nature of the headland makes precise correlation between the units within the headland and those either side of it very difficult. However the close proximity of such distinct facies as the coarse grained Culm facies, medium-fine grained Rusey facies and fine grained Boscastle facies implies that considerable telescoping of the sequence has occurred, Fig. 2.5. Such tectonic telescoping would juxtapose sediments from different parts of the basin and it would be expected that the ages of these juxtaposed rocks would be different. The similarity in age of the three facies can be explained by the subsequent period of extensional faulting. The thrusts would place older rocks on younger, whilst the subsequent extension would drop the older rocks back down, Fig. 2.5. The current arrangement of facies places a medium-fine grained facies (Rusey beds) over the fine grained facies of the Boscastle beds. Both these facies would be expected to occur within the basin margins (Fig. 2.4). The Rusey thrust is therefore considered by the present author to be part of an intra-basin thrust system.

2.3 The Structural Framework

2.3:1 Early Extension and Inversion

Cutoff Point is an outcrop of the Boscastle Beds located in the centre of Exit Bay, Fig. 2.2. Whilst the topographically higher levels of the outcrop show only the isoclinal recumbent folds associated with the regional D_1 , the beach level exposure exhibits a

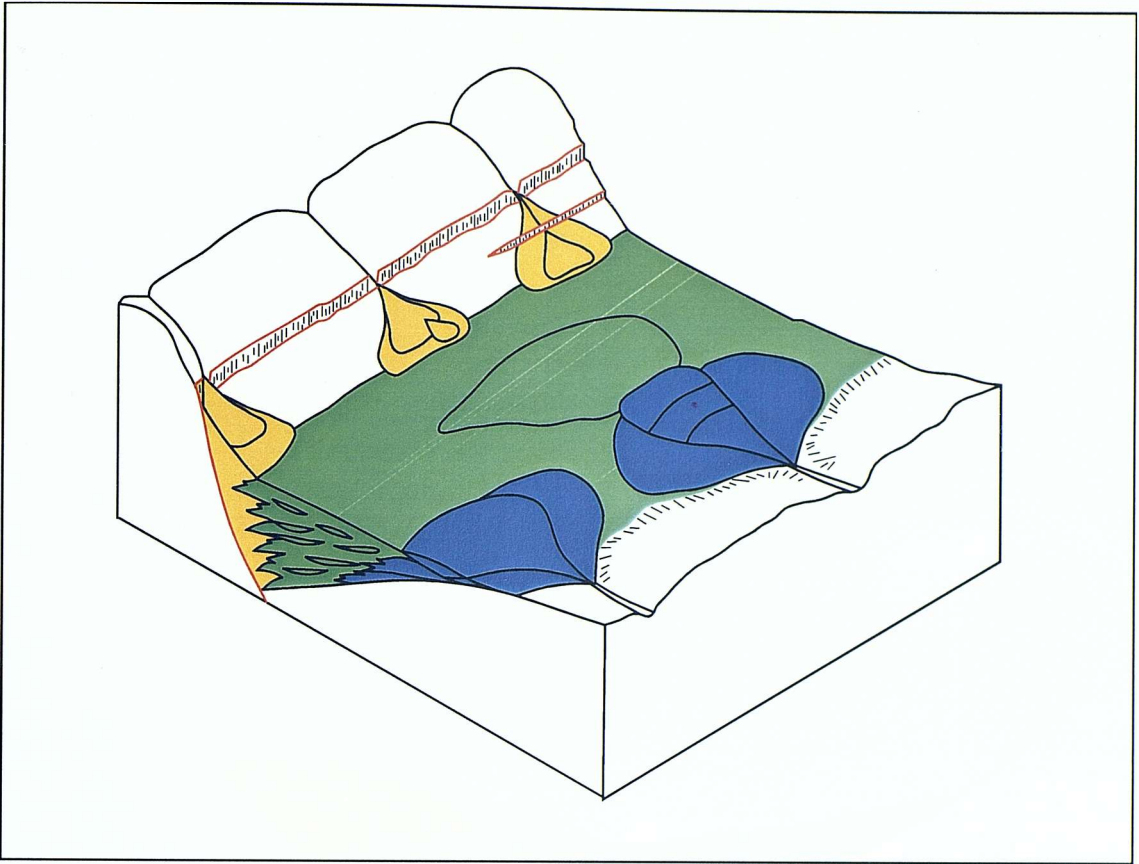


Figure 2.4a Sedimentary deposition patterns in a continental half graben, modified from Leeder and Gawthorpe, 1987 Key is as for 2.4b for ease of comparison.

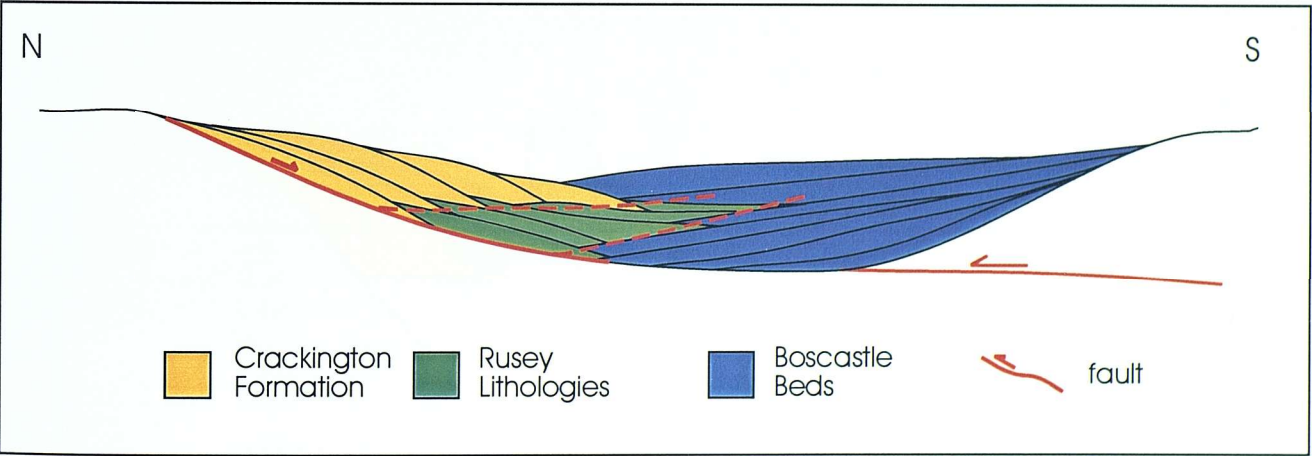


Figure 2.4b Schematic diagram of the deposition patterns in the Culm basin, prior to inversion. The blue lithologies are greatly overdeveloped in this model and the green ones underdeveloped, when compared to the schematic above, because of the much greater input of sediments from the approaching thrust front as opposed to those derived from a rolling over hanging wall.

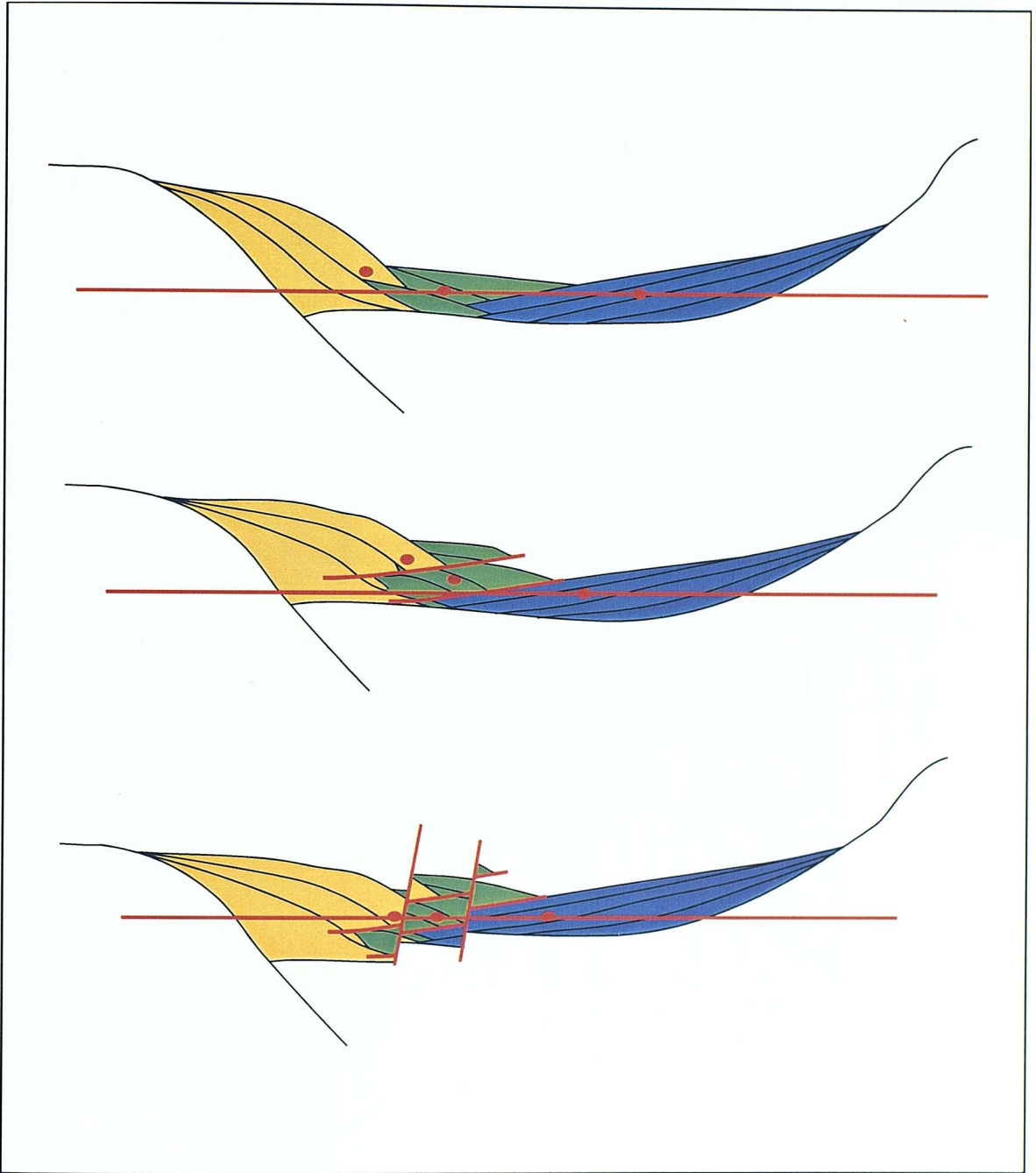


Figure 2.5 Schematic diagram of the path of time equivalent rocks, during thrusting and later extensional faulting.



Plate 2.3 Photograph showing both brittle fracturing of layers and ductile folding within the same section, Cutoff Point, Exit Bay, Rusey

more complex deformation history, Plate 2.3. The sequence shows the development of several gräben. Their bounding faults have been rotated and inverted, Fig. 2.6 and Plate 2.4. and a number of thrust faults have developed. These thrusts often developed above inverted extensional faults, Fig. 2.7 and Plate 2.5, and were possibly initiated at extensional fault tips lower in the sequence, Fig. 2.7 and Plate 2.5. This has been observed in the development of sand box models, (Keller and McClay, 1995). In the experiments, these faults generate thrust tip folds such as those seen in Fig. 2.7 and Plate 2.5, before becoming mature thrusts, Fig. 2.8 and Plate 2.6. There are also rare extensional faults which show no trace of inversion, Plate 2.7 a and b, which may belong to a later phase of extension. Whilst there is possible evidence of syn-sedimentary deformation in the rather chaotic structure of certain coarser grained layers, there is no indication that layers thicken with increasing proximity to the fault plane in the hanging wall of the extensional faults. These faults are therefore regarded as post-sedimentary.

2.3:2 The Thrusts

The rocks of the Rusey headland are cross-cut by a series of faults, Plate 2.8. The fault zone labelled A in Fig. 2.9 contains several displacement planes striking 020°-040° and dipping 20°-40° to the E and SE. The exposure is heavily weathered and difficult to access (Plate 2.1) hindering understanding of the extremely complex deformation adjacent to these planes. The lowermost fault below fault zone A is a well developed slip plane which is here termed the Rusey Thrust. Stereographic analysis of the thrust plane orientation data acquired in the field shows that all the thrust faults dip towards the east, though there is considerable variation in the strike, Fig. 2.10a. Fold axial planes show a fairly consistent dip to the south-east, Fig. 2.10b. Direct measurements of the minor fold axes were rarely possible and would not have been reliable. This was due to the state of the outcrop, which was in part dangerous and in part difficult to measure as at that level it was vertical and extremely unstable. Measurements of fold axial planes, though less accurate, could be obtained with less risk. Taken in conjunction with the sense of vergence, which indicates rotation of the fold limb was top to the north or north-west, this is interpreted to indicate a general transport of top to the north-west along the thrust fault. The data are scattered and this reflects the complex nature of the fault architecture at Rusey. The main Rusey Thrust fault is exposed 5-10m above sea

NW

SE

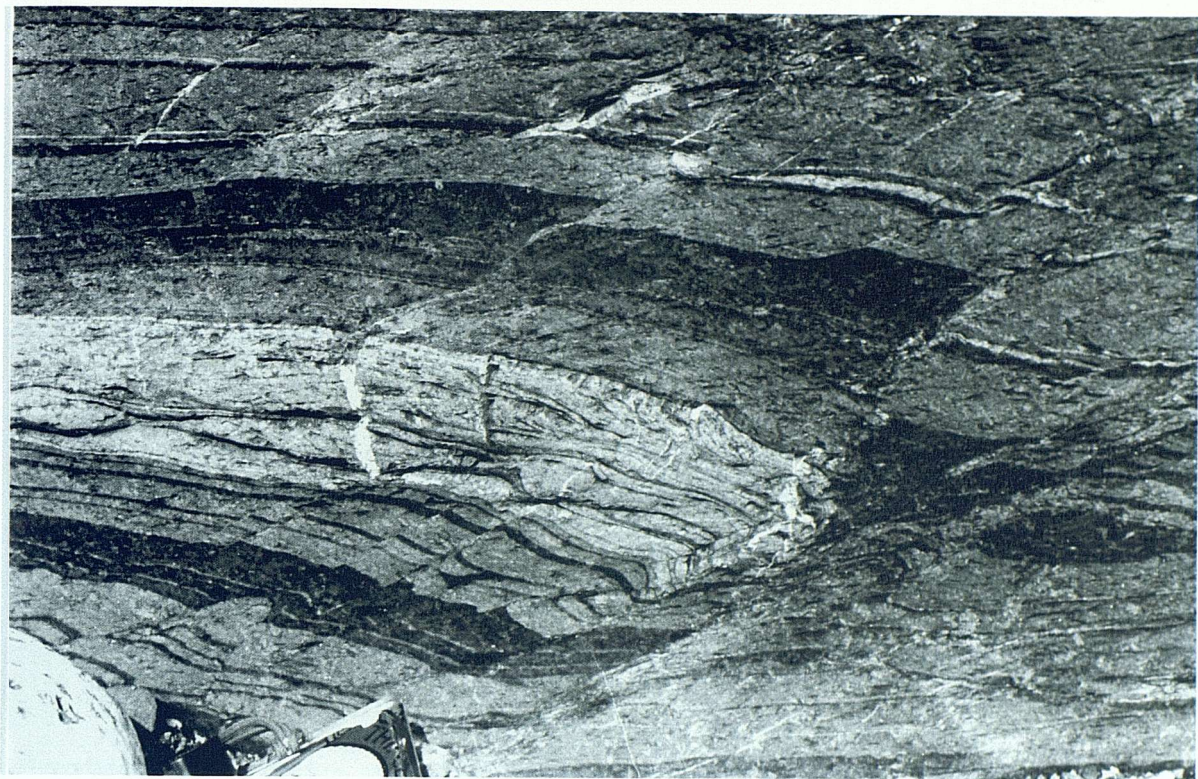


Plate 2.4 Inverted extensional structures at Cutoff Point.

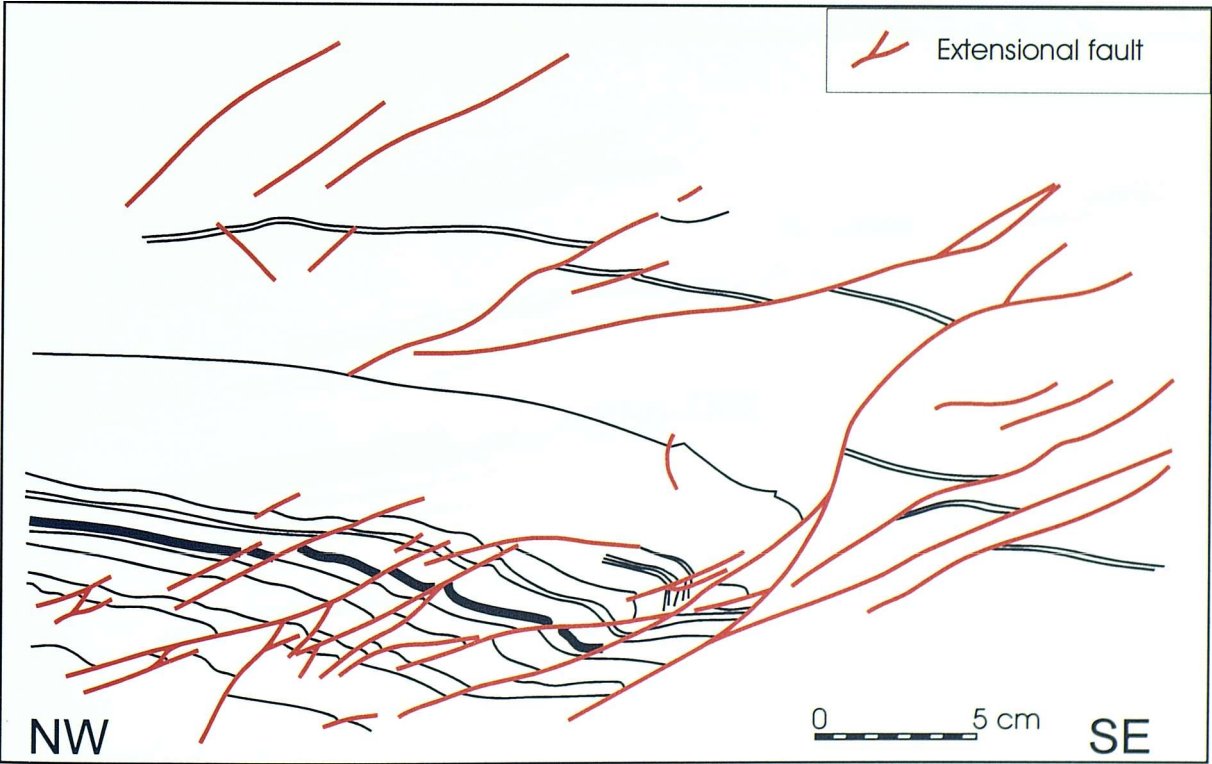


Figure 2.6 Inverted extensional structures at Cutoff Point.

NW

SE



Plate 2.5 Incipient thrusts and thrust tip folds at Cutoff Point.

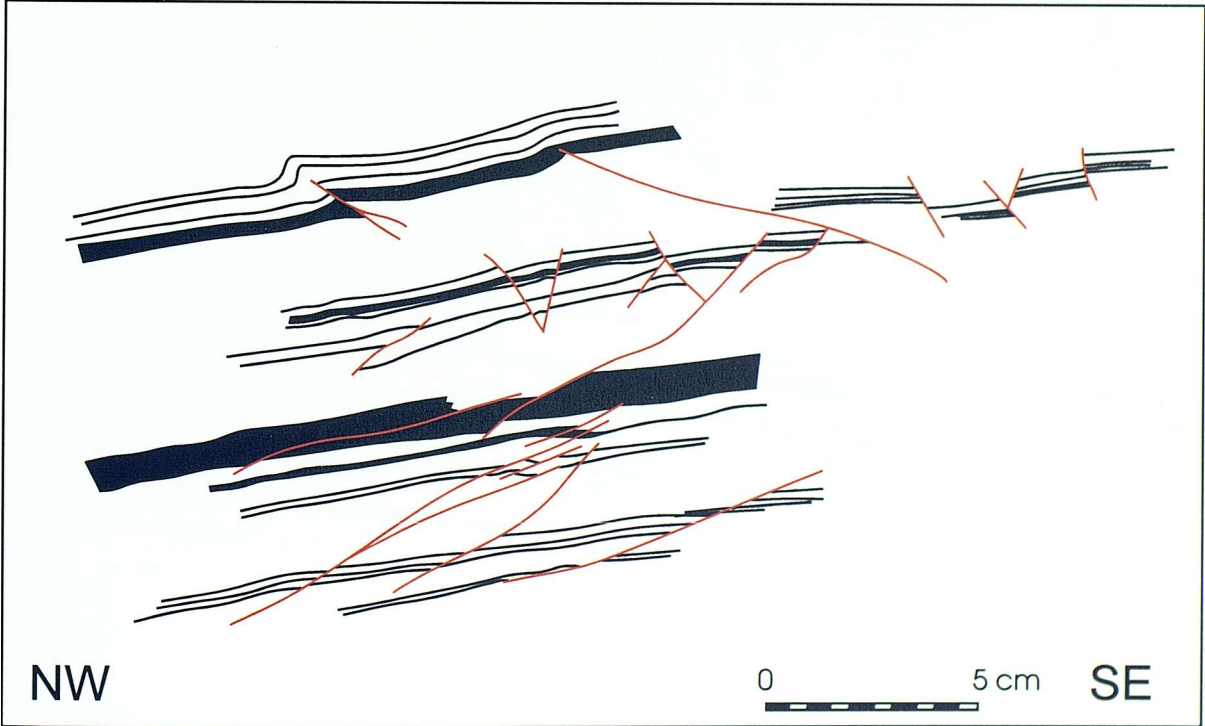


Figure 2.7 Incipient thrusts and thrust tip folds at Cutoff Point.



Plate 2.6 Well developed thrust faults at Cutoff Point.

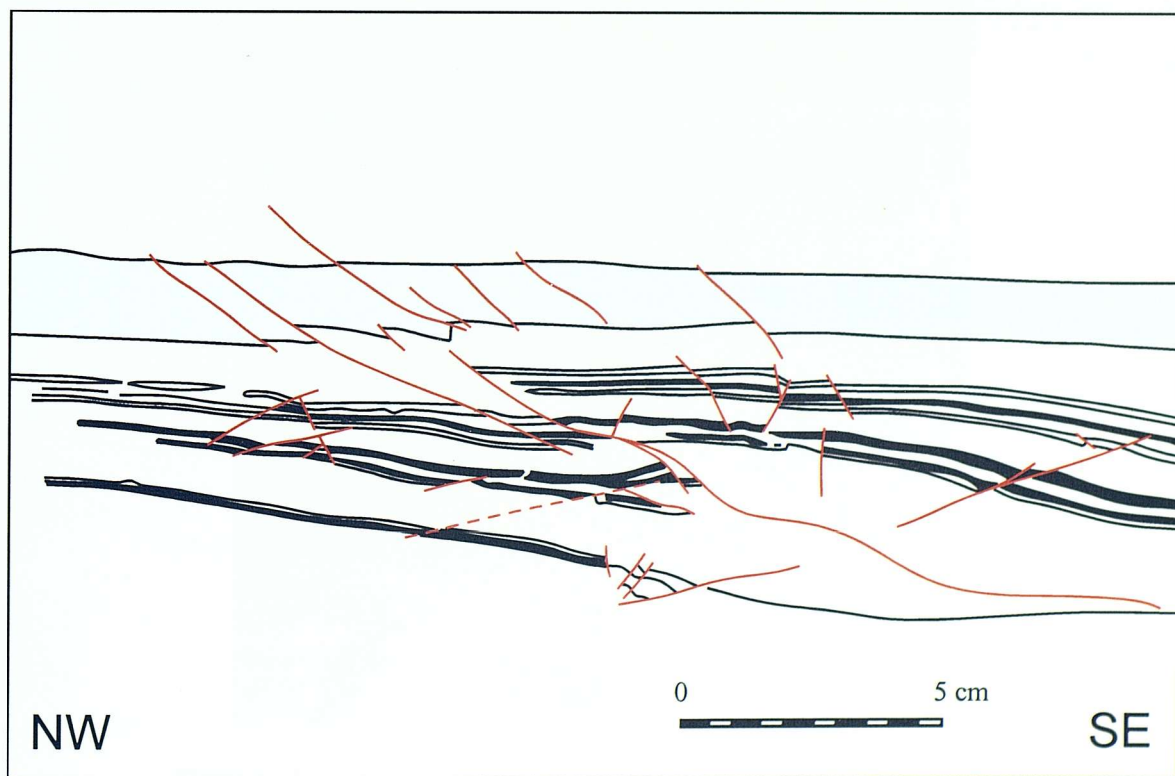


Figure 2.8 Well developed thrust faults at Cutoff Point.

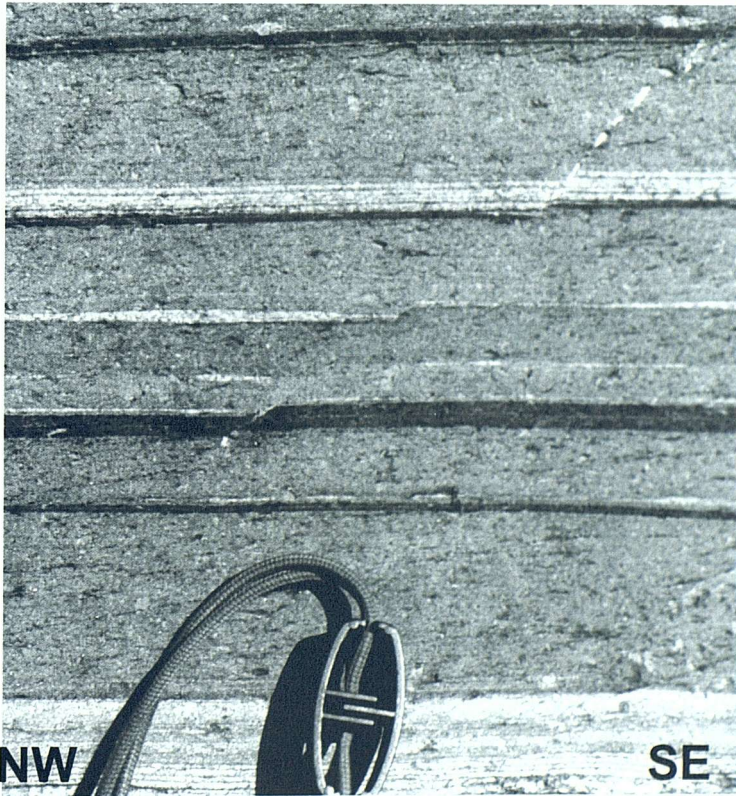
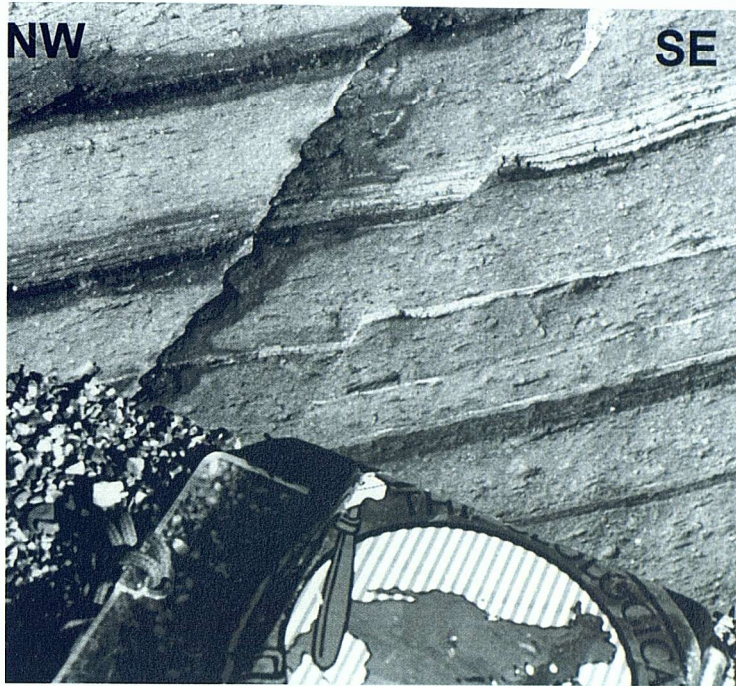


Plate 2.7a and b Extensional faulting in the rocks at Cutoff Point, Exit Bay. These faults show no sign of inversion.

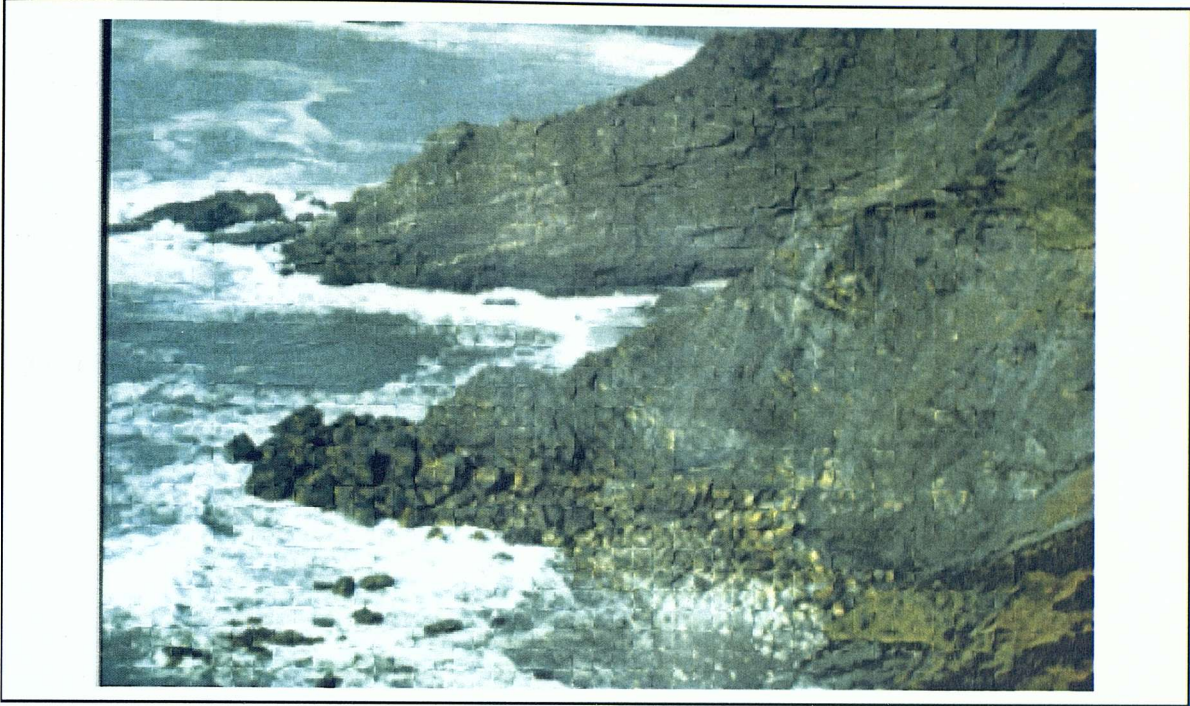


Plate 2.8 Section view of the Rusey headland, North Cornwall.

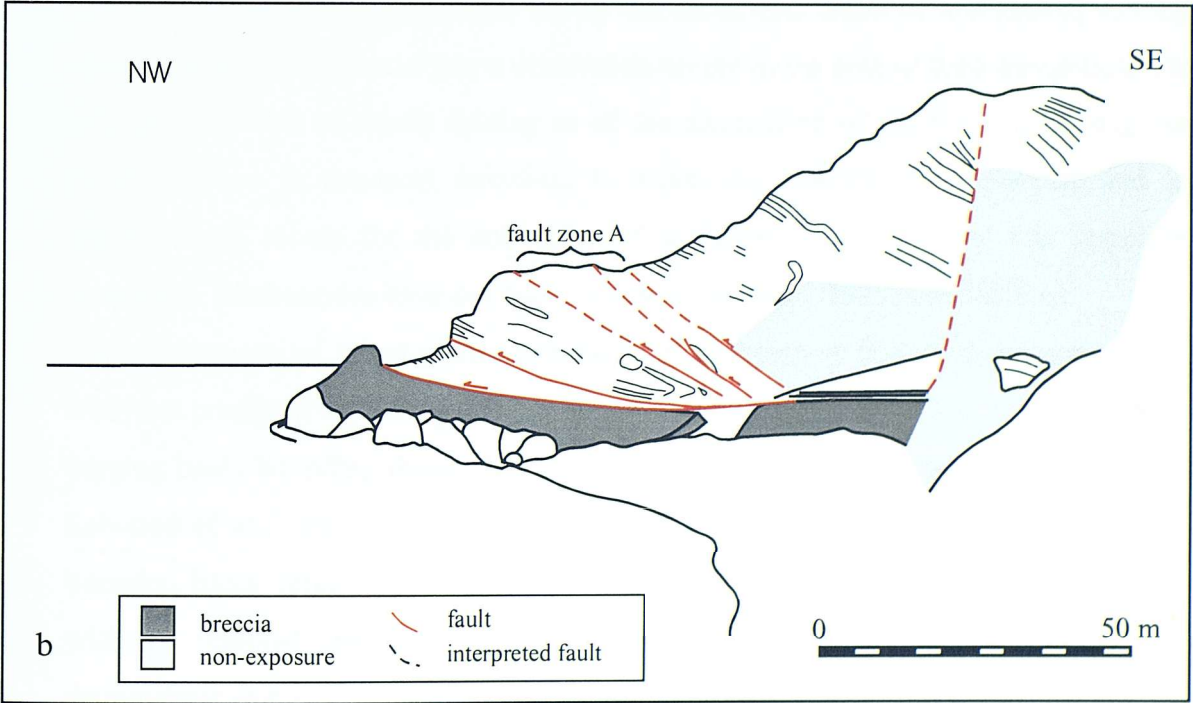


Figure 2.9 Geological section of the Rusey headland, North Cornwall.

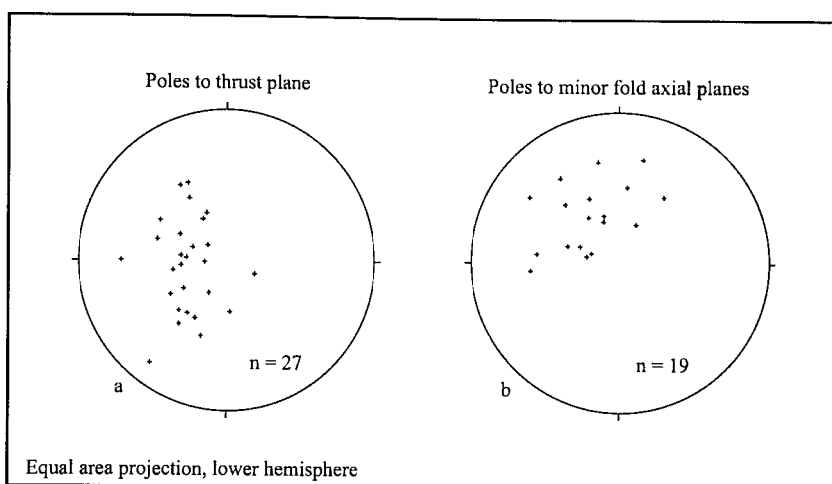


Figure 2.10 Stereoplots of the poles to the thrust planes and the axial planes of minor folds recorded at Rusey Headland

level and on the southwest side of the headland. Fault zone A merges down dip with the lower thrust plane. The interaction of fault zone A and the Rusey Thrust is responsible for the local complexity of the deformation, Plate 2.1. These faults are interpreted as thrusts, on the basis of fold axial plane orientation and sense of overturning given above, the dissection of folds into thrust bound fold hinges and fault/bedding relationships. These structures give a consistent top to the north-west sense of movement, although some variation in fold axial plane orientation occurs in the area of fault interaction. This may be the result of sheath folding or of the complexity of the faulting causing very local variation in transport direction. However the outcrop is sub-vertical and too dangerous to access for the collection of sufficient data to assess this possibility accurately. Slickensides have not been observed on any of these fault surfaces.

The orientations of thrust planes and thrust fault transport directions do not coincide with that predicted by the regional models, which interprets the Rusey Thrust as a south verging basin bounding thrust (Sanderson and Dearman, 1973, Shackleton et al. 1982, Selwood et al., 1985) though the position of the Rusey thrust system within a fault bounded block renders regional correlation uncertain. However a precise correlation with the regional geology is not necessary for this study. The present author will demonstrate that the present morphology of the Rusey Breccia relies more on the very local stress and fluid pressure conditions than on the regional tectonic environment.

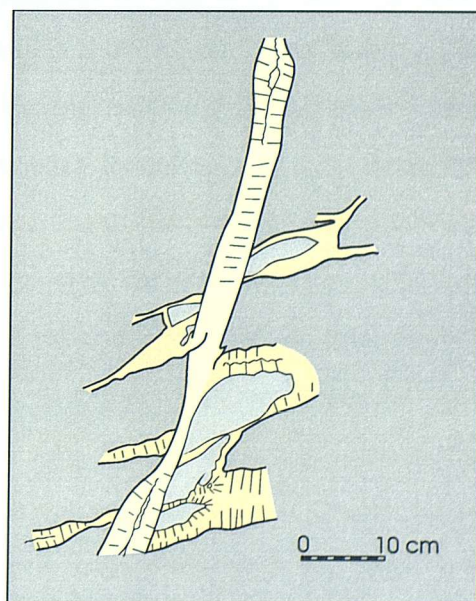
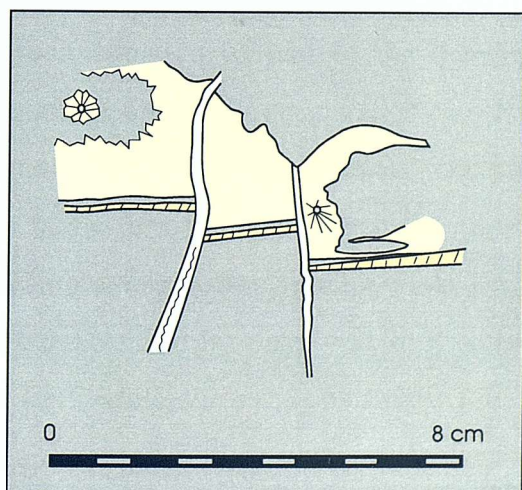
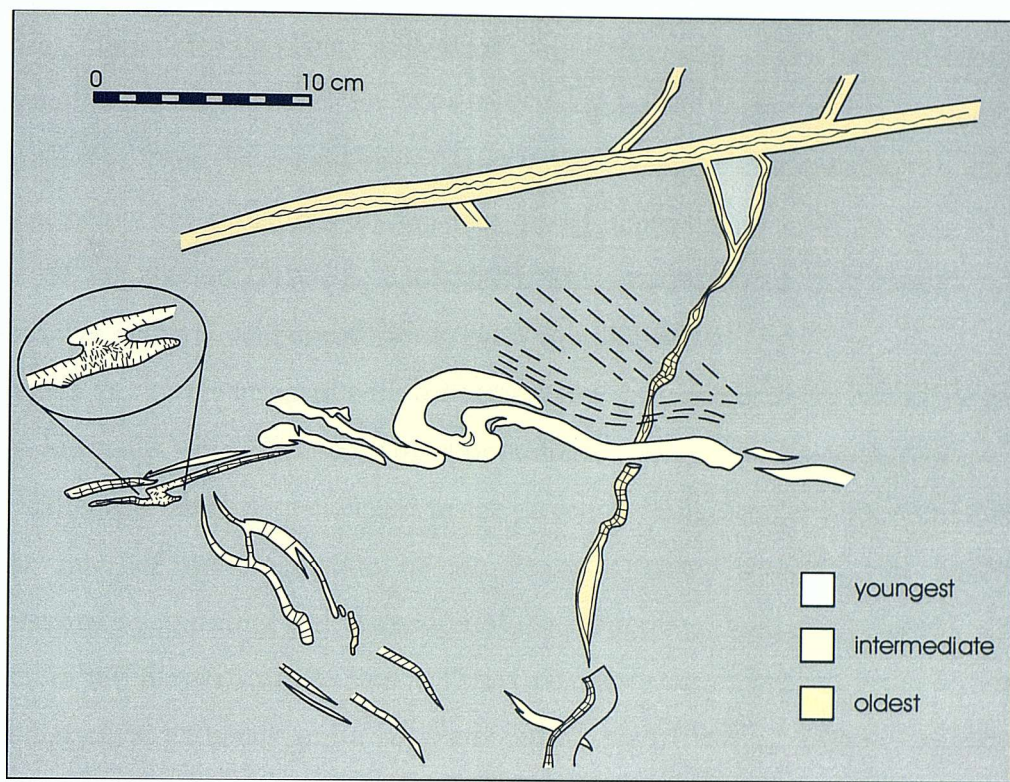


Figure 2.11 Examples of multiple generations of veining in the shales surrounding the main breccia band. The different generations of quartz are shown in different colours as in the key in the top diagram

2.3:3 The Rusey Thrust Zone and the Rusey Breccia

The rocks immediately below the Rusey Thrust are pelites that have been affected by multiphase deformation (Fig. 2.11). They have been extensively veined, followed by intense fabric development associated with thrusting. This has dissected and recrystallised the first generation of veins to the point where the fabric now defines their morphology. There is considerable thrust related veining and brecciation. The fabric present below the thrust has an apparent dip of approximately 52° to the ESE, in the plane of the section (NW-SE, 330° - 150°). The orientation of this fabric, an intense penetrative mineral alignment fabric, with a strike of 016° and true dip of 59° east (previous section) is oriented with the same strike as the beds in the hanging wall of the thrust but a dip which is steeper by approximately 40° . It is compatible with formation as a flattening fabric perpendicular to the direction of maximum compression within a thrust zone of the orientation of the Rusey Thrust. Thrusting was followed by extensional overprinting and disruption of the fabric, Fig. 2.12. The intensely deformed shales in the footwall of the Rusey Thrust contain a major breccia band, a number of subsidiary breccia bands and a large amount of quartz veining (Chapter 4). The fabric developed within the thrust zone is not seen either in the breccia clasts, Plate 2.9, or within elements of unbrecciated country rock shales which are encapsulated by the main breccia band. This indicates that the fabric had not formed at the time the brecciation took place and that the clasts were in some way protected from fabric forming processes by their situation as part of the breccia. Furthermore, the fabric planes change in orientation until they are parallel with the breccia boundaries as they approach those boundaries, indicating detachment onto the margins of the breccia body, again implying that fabric formation post-dated the development of the breccia. The clasts frequently contain pre-brecciation veining, Plate 2.9. The top and bottom of the main breccia band are characterised by sharply defined surfaces. The upper surface strikes 155° and dips 55° ENE whilst the basal one strikes 170° and dips 30° - 36° E. Whilst these orientations are an average of those obtained, the upper surface invariably has a steeper dip than the lower surface, indicating that the main breccia band may have a wedge or lensoid geometry, and is therefore of restricted spatial extent, Fig. 2.13. Similar sharply defined surfaces are developed within the band, Plate 2.10. These internal surfaces have orientations that are broadly coincident with those of the bounding surfaces of the main breccia band, striking 146° - 160° and dipping at between 40° - 65° ENE (typically 50°).

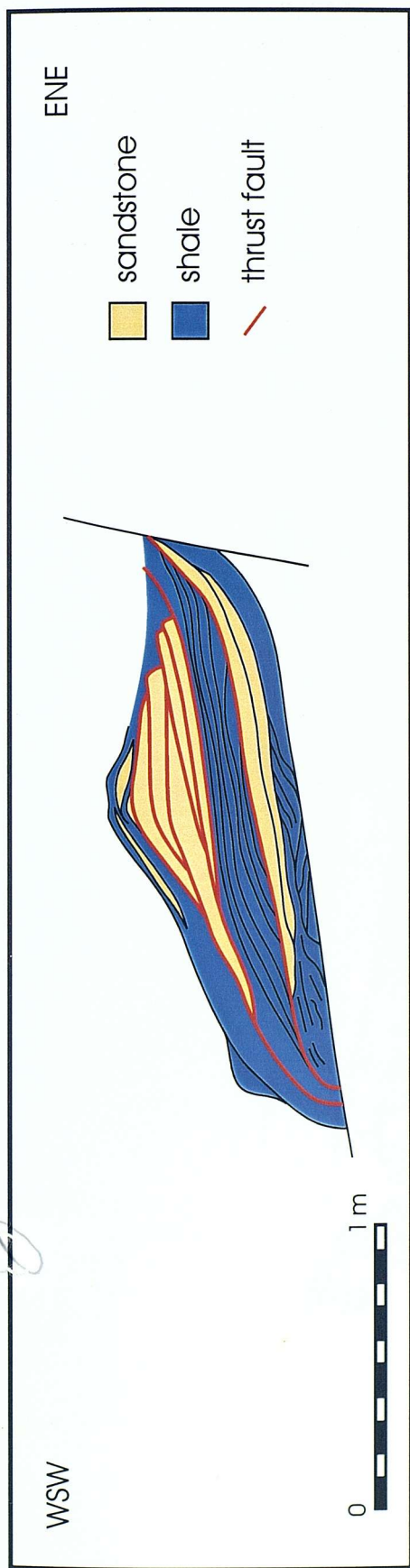


Figure 2.12a Strike section of part of fault zone A, see Fig.2.9, showing a thrust duplex/stack, with the top transporting toward the observer. The left side is undisturbed. The right side is extensionally displaced along the adjacent fault plane.

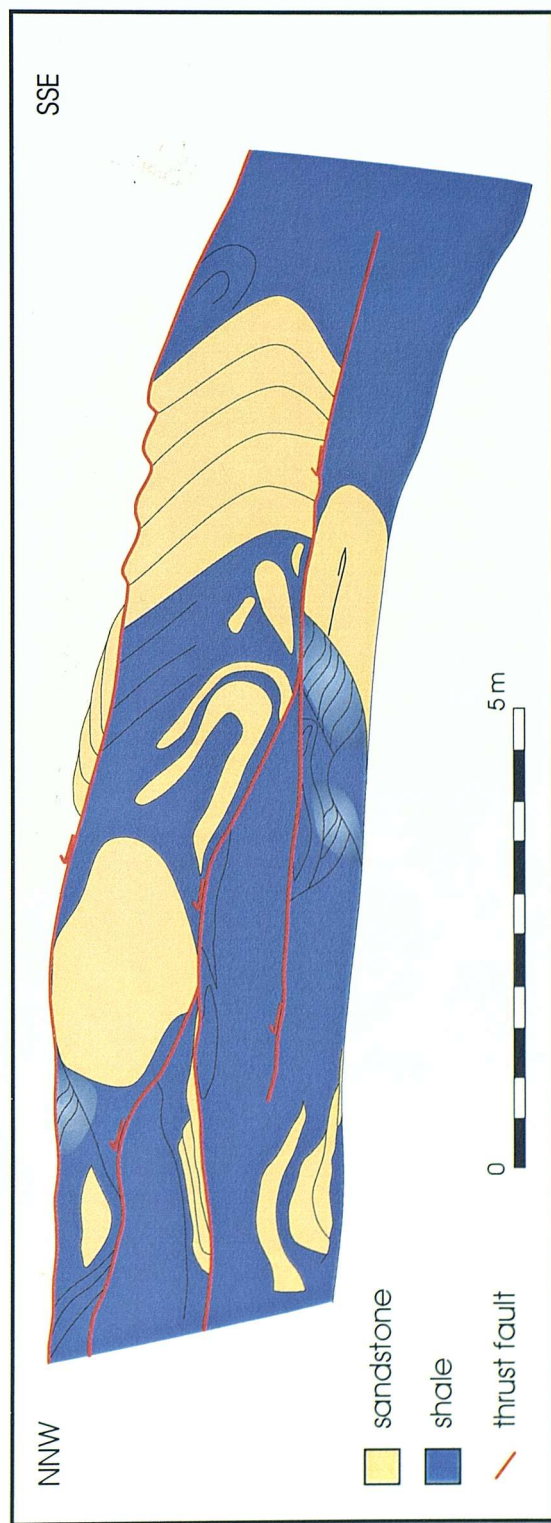


Figure 2.12b Dip section of part of fault zone A, see Fig.2.9, showing thrust planes and thrust folds, transport sense, top to the left. The thrust generated fabrics are extensionally offset in several places, (highlighted).

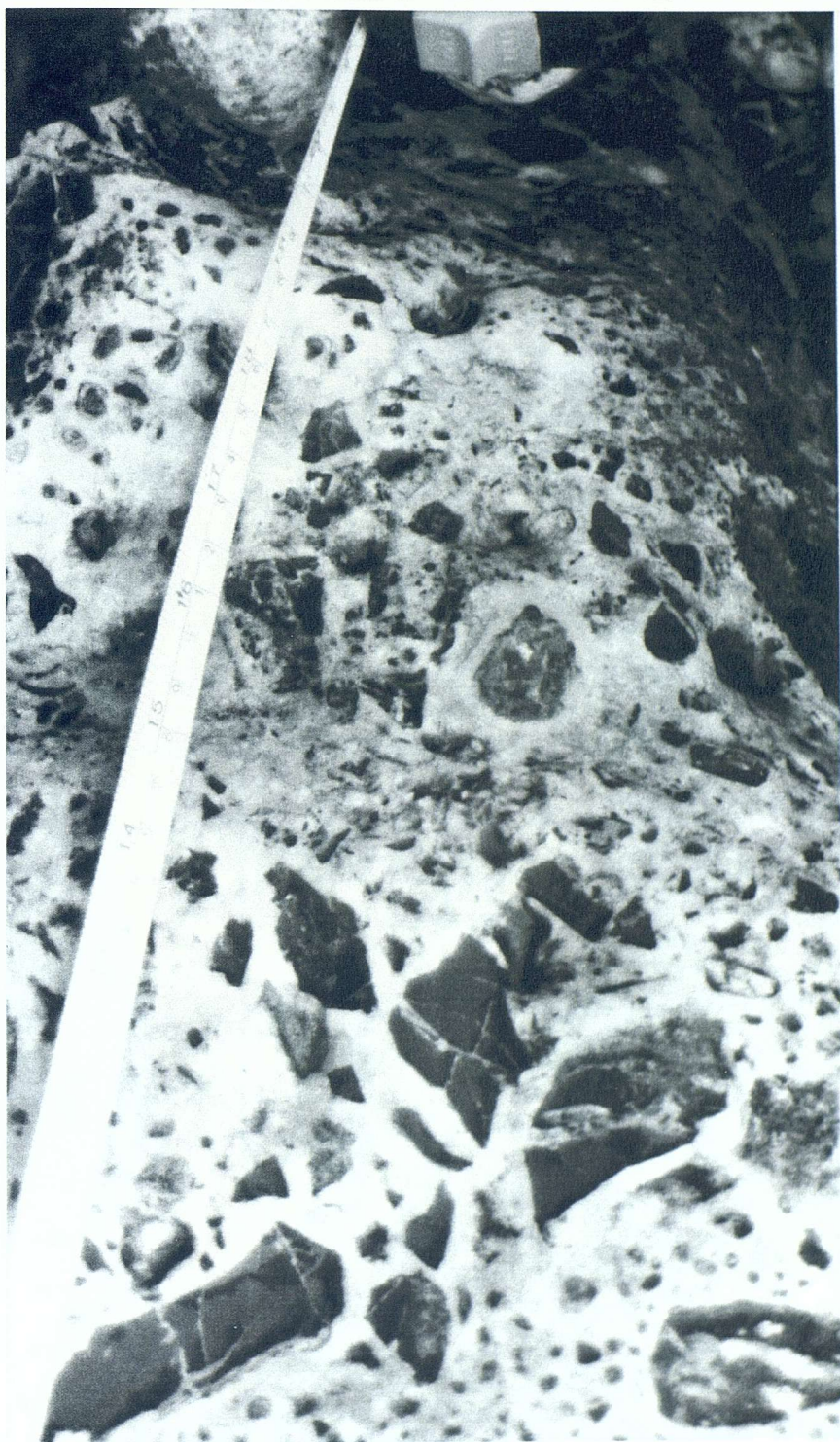


Plate 2.9 Photograph of the Rusey Breccia showing veined clasts without fabric development

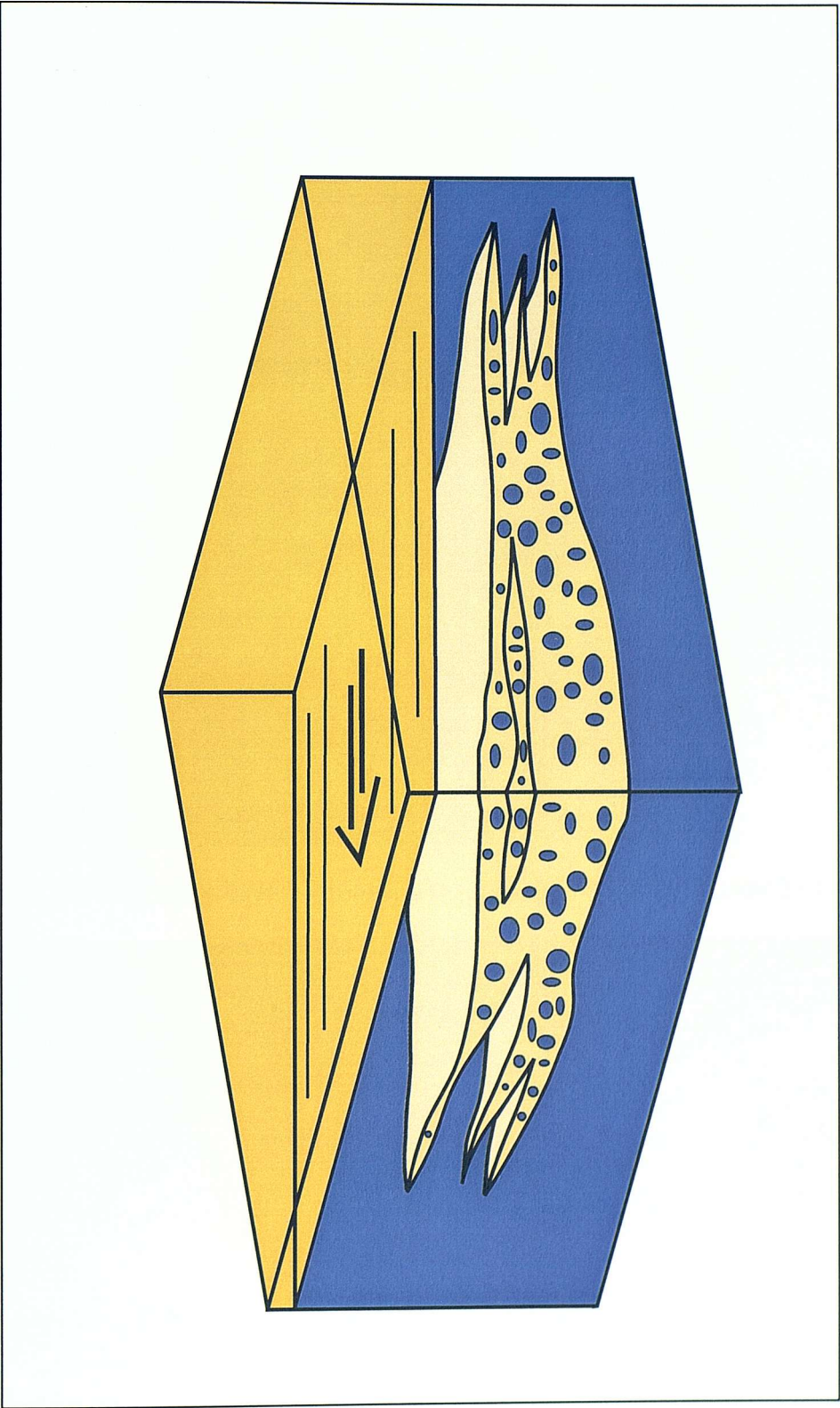


Figure 2.13 Block diagram, showing the three dimensional form of the main Rusey breccia band.



Plate 2.10 Well developed internal slip surfaces within the Rusey Breccia



Plate 2.11 Well developed slickensides on the slip surfaces within the Rusey Breccia

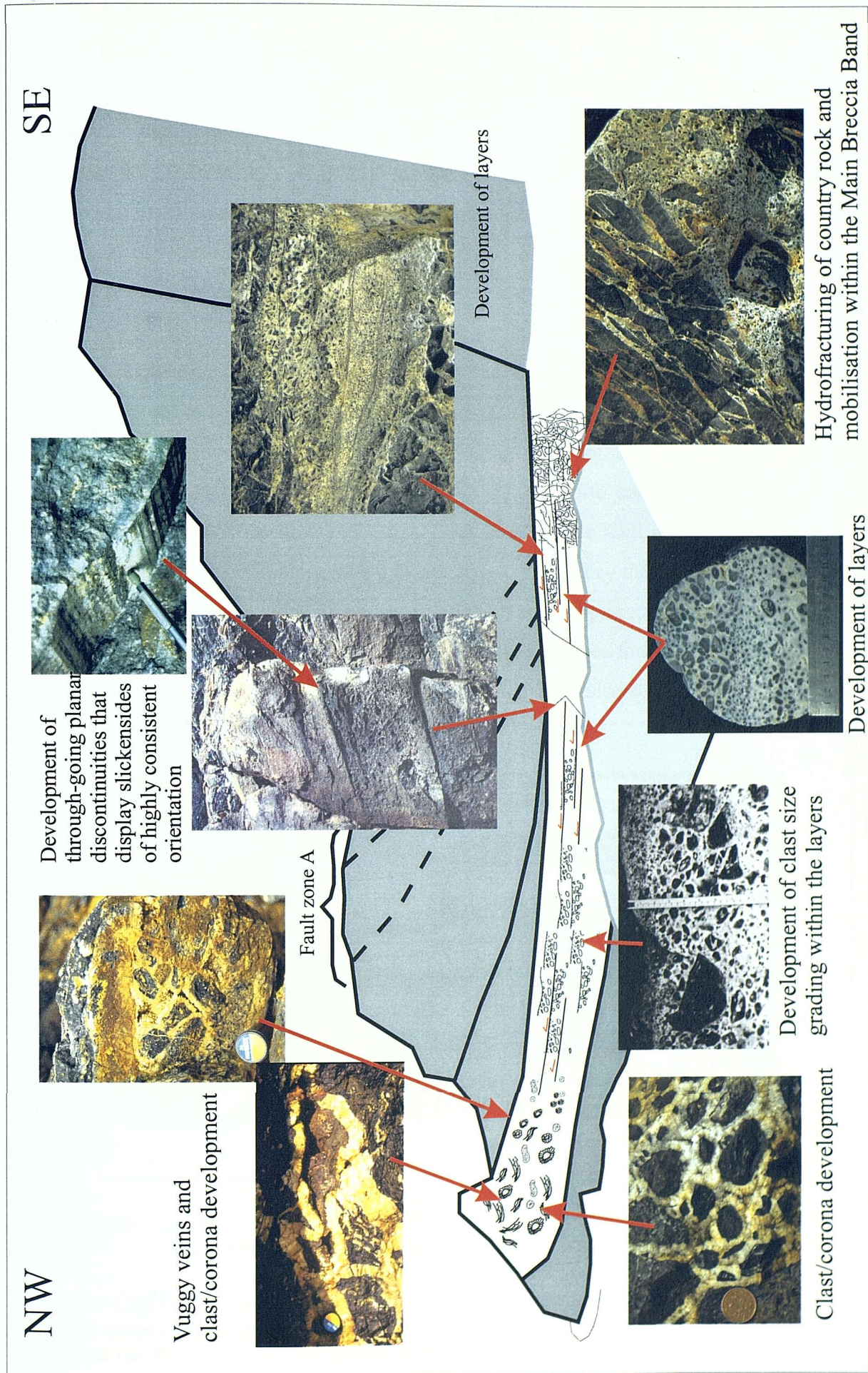


Figure 2.14 Composite synoptic diagram of the Rusey Breccia and its defining features

They have highly distinct subhorizontal slickensides, restricted in orientation to 150°-160°, Plate 2.11, indicating that movement along these surfaces was strike-slip. The absence of slickensides on the thrust surfaces and the form of these surfaces, cross-cutting the mineralised breccia, leads the author to place the generation of these features after the thrust activity. A general synoptic diagram is shown in Fig. 2.14, showing the spatial distribution of these features, with photographic illustration of the each feature.

2.3:4 The Extensional Fault

The slickenside orientation recorded for the intra-breccia slip surfaces (see above) matches those recorded for slickensides on an extensional fault plane, termed here the Cliff Fault, exposed at top and the base of Rusey cliff, Fig. 2.2. At the cliff top exposure the fault plane has a strike of 132°, a dip of 68° north-east and slickensides which plunge at 43°-48° towards 346°-352°. Beds abutting the fault on the north-east side of the fault are folded in a concave upwards sense where they make contact with the fault, suggesting that this side has undergone downwards displacement relative to the south west side of the fault. Layers of the Cliff Fault breccia (see footnote) exposed on the top of Rusey Cliff exhibit a rootless folding style, Fig. 2.15, also seen in the intermediate

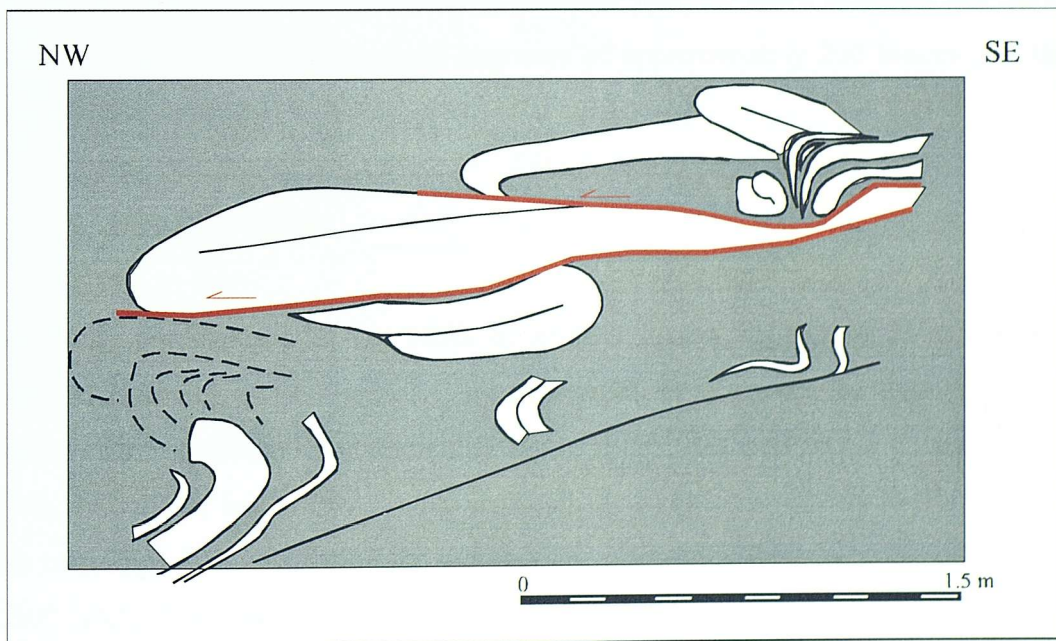


Figure 2.15 Thrust (red) bound and dissected folds, located on the Cliff Fault at the cliff top outcrop, Rusey, showing white breccia bands within blue shale.

The Cliff Fault breccia is a largely inaccessible discontinuous breccia body found along the trace of the Cliff Fault. Its presence is noted as having certain features in common with the Rusey Breccia, but greater similarity cannot be ascertained.

Rusey lithologies above the Rusey Thrust on the beach. The Cliff Fault can be located again at the base of Rusey cliff, where it occurs to the south-west of the headland, separating the Rusey lithologies and the main breccia band from the Boscastle Beds, Fig. 2.2. Although the fault plane is scree covered at the base of the cliff, the beds immediately

to its south-west have been intensely folded, such that the beds are convex upwards and dip approximately 45° north immediately adjacent to the fault plane, indicating that the Rusey headland is downthrown to the north-east. The fault can easily be traced across the foreshore as a distinct band of fault breccia cutting across the wave cut platform with a strike of 136° (cf. strike of the fault at the cliff top exposure). The folding of the beds and the slickenside orientations obtained from the fault plane at the cliff top exposure are used to interpret the motion on the fault as combining, in at least the final phase of slip, an important element of sinistral strike-slip motion with extensional motion. The large degree of down-warping of the beds at the base of the cliffs adjacent to the fault indicates substantial normal dip-slip displacement, on the order of decametres. The lack of displacement markers present on both sides of the fault and the poor exposure of the fault trace make a more accurate assessment of the amount of displacement across the fault impossible. However field evidence indicates that the Cliff fault is a major fault with a vertical exposure of approximately 200 metres and that it was oblique extensional in its last phase of activity.

2.3:5 The Strike-slip Faults

Two faults are exposed in the cliffs of Middle Strand, to the north-east of Rusey Headland (faults X & Y, Fig. 2.2). When viewed from above, the trace of the most easterly of these (fault X) is a distinct topographic feature, Plate 2.12 which strikes NNW, though the fault surface is not exposed. Bedding immediately to the west of the fault trace dips 70° - 80° to the east, whereas away from the fault trace the dip decreases to 50° NNE. This implies that there is an element of dip-slip on the fault, with the eastern side being downthrown. In addition, as the beds approach the fault trace, their strike progressively rotates through 50° , from 315° to 005° , indicating a significant element of dextral strike-slip, Fig. 2.2, fault X. Between fault X and the Rusey headland another fault occurs (Y, Fig. 2.2). There is no systematic rotation of bedding strike as



Trace of strike-slip fault
marked by linear dip in
topography

Plate 2.1.2 The lower cliffs of Middle Strand from the cliff top path

the trace of fault Y is approached from the west, though the beds dip towards the east and there is a progressive decrease in dip immediately adjacent to the fault (c.40°) indicating that the area to the west (the Rusey headland) has been downthrown by this fault. The trend of this fault is north-northwest but is much less defined than that of fault X. The dislocation appears subvertical, though the outcrop is covered in scree and this poor exposure makes it impossible to determine the exact slip vectors for fault X or Y. Since the headland is bounded both to the north-east and south-west by major steeply dipping faults and the lithologies are separated and entirely contained by these bounding faults, the direct field correlation of the rocks of the headland with the surrounding geology is not possible. The correlations which have been proposed in this study are those which require the minimum degree of complexity in the initial sedimentary architecture and the ensuing deformation.

2.4 Discussion

The three rock types exposed in the area are almost certainly genetically related. The exposure now available documents the sedimentary infilling and the compression and extensional dissection of a single sedimentary basin. The extensive dismemberment of the stratigraphy by faults prevents a precise correlation on the basis of field evidence, although a detailed palaeontological study may be able to define their relationship to each other more clearly.

However, because the rocks of the Rusey headland are isolated from the adjacent rocks by faults, it is difficult to determine their exact relationship to the regional metasediments. The occurrence of intensely developed deformation related to low angle faulting and brecciation at the junction of the Boscastle beds and the Rusey beds indicates that this boundary acted as an important displacement horizon. Kinematic indicators demonstrate that it has accommodated significant reverse dip-slip, which makes this fault compatible with the tectonic framework for the region (Chapter 1) in which major thrusting and inversion of the Culm basin occurred during the Variscan orogeny. The transport direction recorded on the thrust in its present orientation is to the northwest. This is not in accordance with the transport direction by the regional model (Sanderson and Dearman, 1973, Shackleton et al. 1982, Selwood et al., 1985). However the presence in the immediate locality of large faults which are suspected from field evidence to have accommodated strike-slip motion could have modified the orientation

of the thrust. It is also possible, given the extremely limited outcrop available for study, that the Rusey Thrust is part of a local perturbation along a regional thrust which is more compatible with the general regional model. The difference between the regionally predicted transport direction and that deduced for the Rusey Thrust may also be the result of movement on the series of thrusts labelled fault zone A on Fig. 2.9. The dangerous state of the outcrop prevented the acquisition of data from fault zone A from a level higher than approximately six feet above ground level. However visual assessment of the thrust transport direction indicated by minor thrust over-rides and fold vergence is apparently consistent with that of the Rusey Thrust. Fault zone A is interpreted as a large ramp thrust along the main Rusey Thrust. Fault X, to the north of the headland, is interpreted as large dextral strike-slip fault and the Cliff fault as a major oblique/extensional fault, downthrowing to the north-east. The final movement along the Cliff Fault, indicated by slickensides, is highly oblique sinistral, an uncommon sense of motion for deformation of Variscan age, with dextral strike and oblique slip motions being more commonly cited for late stage faulting (Shail and Wilkinson, 1994, Le Gall, 1990, Barnes and Andrews, 1986, Holder and Leveridge, 1986). Since the slickensides found on the Cliff fault and on surfaces cross-cutting the breccia are consistent in orientation with those found on extensional faults elsewhere along the coast, it is argued that, unlike the major thrust, the slickensided surfaces in the Rusey area have not been rotated by the local dextral strike-slip movements since they were last active and the slickensides were formed and therefore that the phase of activity to which they relate must post-date both thrusting and dextral strike slip faulting. The relationship between the rocks of the Rusey headland and those east of fault X and north of Middle Strand is unclear because the boundary between them is poorly exposed. However the abrupt change in facies across these boundaries and the field measurements of bedding strike and dip (Fig. 2.2, sections 2.3:3 and 2.3:5) indicate that these boundaries are faulted. Finally, an early phase of extensional faulting occurs prior to thrusting at Cutoff Point. These features may represent a purely local area of extension. The faulting style is inconsistent with it being syn-sedimentary or soft sediment deformation, as is common further north along the Bude beach section, as it lacks definite layer thickening into the fault planes. However the evidence of veining in these rocks prior to thrusting, as demonstrated by the veins in the breccia clasts, suggests that the rocks at the Rusey Headland were behaving in a more brittle manner than those further north prior to thrusting and it is quite possible that these early extensional features can be correlated with syn-sedimentary extension as documented by Whalley and Lloyd, 1986, or even

synchronous with thrusting as documented by Mapeo and Andrews, 1991, both in more northerly outcrops. The evidence of deformation in these rocks almost completely ends with the thrusting episode, despite being in the immediate vicinity of a large extensional fault that was active at a late stage in the deformation history of this particular area. This would certainly suggest that the area has been subject to a spatially limited stress regime that differed from the broader local one, making its variation from the expected pattern of deformation of less significance. It is unlikely that the initial extensional episode can be correlated to the orogenic collapse episode as there are no compressive episodes which postdate the orogenic collapse and which are intense enough to have caused the observed thrusting. This makes the rocks of Cutoff Point anomalous for the area, though whether this is a point of any regional or even local significance, except at the smallest field scale, is open to debate.

On the basis of the facies contained in the three units, it is probable that the Boscastle Beds, Rusey beds and Crackington Formation were deposited at different geographical locations within the same sedimentary basin, but that they could have been deposited penecontemporaneously. The palaeontological evidence indicates that the rocks are very close in age. As discussed earlier in this chapter, the present juxtaposition of these different facies can be explained by a combination of thrust stacking followed by extensional faulting, Fig. 2.5. The details of this tectonic scenario could be resolved by a careful integration of palaeontological and structural data that is beyond the remit of this study. As the sediments juxtaposed by the Rusey thrust are both mid-basinal, medium to fine grained facies and those basinal sediments extend both to the north and south, the exposure of the Rusey Thrust examined in this study is interpreted, by the present author, as located within the Culm Basin (cf. Fig. 1.24c, after Selwood, 1990) rather than being the southern basin bounding fault as proposed (Isaac et al, 1982) or accepted (Andrews et al, 1988, Pamplin, 1990) by previous workers.

2.5 Summary

The rocks at Cutoff Point record a deformation history that locally begins with extensional deformation of lithified sediments later disrupted by thrusting. The first major phase of deformation, both locally and regionally is thrusting as part of the Variscan orogeny. On both the local and regional scale D_1 thrusting would have

exploited weak zones defined by high fluid pressure and/or lithological variations (Enfield et al. 1985, Whalley and Lloyd, 1986). The thrust fault split the sequence into lithologically distinct units. Development of the Rusey Thrust involved catastrophic failure, which produced the protolith of the Rusey Breccia. The presence of quartz veins within the clasts (Plate 2.9) indicates that the rocks had fractured at P-T conditions under which quartz is mobile before thorough brecciation of the rocks took place. To occur under such conditions, the brecciation episode must have taken place post-lithification, giving it a tectonic rather than sedimentary origin. The possibility of resedimentation of lithified pebbles is not considered a viable alternative because of the consistency of composition between the clasts and present day country rocks (indicating close proximity of clast source area and deposition location) as well as the inconsistency of the lithofacies and the requisite coarse clastic deposition environments. Clasts of a sedimentary nature would have been deposited in a high energy sedimentary environment, e.g. beach facies. The black pyritic shales outcropping on the Rusey Headland do not match this lithofacies. Turbidite flows, which are developed in the geographic vicinity, do not commonly carry clasts of this size.

Variscan thrusting is the earliest post lithification phase of deformation in the area and the fabric development in the country rocks is interpreted as demonstrating that the breccia acted as a rigid body whilst the thrust fabrics were developed in the shales. Therefore the breccia is interpreted as having genetic links with the Rusey Thrust, with the breccia formation and its modification having taken place during a very early stage in the thrusting. The Rusey Thrust and its associated breccia were then isolated from their genetic environment by later faulting consisting of dextral strike slip faulting and a final major phase of extensional and oblique/extensional faulting. Although there is no direct evidence from the field area, it is clear from previous work that dextral strike-slip faulting was an important feature of the orogen during its development (Barnes and Andrews, 1986, Holder and Leveridge, 1986, Jackson, 1991) whilst extensional dissection took place after convergence had ceased (Freshney et al., 1972, Selwood et al., 1985). For this reason the strike-slip faults at Rusey are interpreted as being older than the extensional faulting, though the degree to which dextral strike-slip has impacted the present-day orientation of the structures at Rusey is not quantifiable from the present-day outcrop. Supporting evidence for the age relationships between strike-slip and extensional faulting is provided by the lack of a strike-slip fault in the headland north of Middle Strand, where the extrapolation of Fault X would be expected to crop

out. This has been truncated by an extensional fault which can be seen in the cliffs at the base of the northern headland.

Chapter Three

Photoelastic modeling of the Rusey Thrust system

3.1 Introduction

The presence of rectilinear and ladder vein geometries with pure tensile movement vectors on the margins of the Main Breccia Band implies that the area now occupied by the Rusey Breccia was initially deformed by tensile failure. The presence of large amounts of quartz vein fill would imply that this fracturing took place in the presence of significant amounts of mineralising solutions and that therefore the deformation mechanism is likely to be hydrofracture. Consideration of the stress regime likely to occur in a major orogenic thrust zone led the author to propose that the generation of a stress regime appropriate for tensile hydrofracture may occur as a result of the change in transport direction of the hanging wall at the thrust ramp. The author proposed that the change in orientation of the fault zone at the ramp lift-off point would create a stress focus and that the change in direction of transport of the hanging wall would result in a locally enhanced differential stress. It is proposed that the presence of fluids trapped beneath the fault plane would lower the effective stress in this area sufficiently that the area within the footwall stress focus would fail in tensile mode as the differential stress is increased as a result of fault slip events. If the presence of such a stress focus could be demonstrated then this would represent a valid explanation for the genesis and location of the well developed but highly localised Rusey Breccia. This model could produce the breccia body by deformation occurring on an observed structure. This would remove the need for the presence of unobserved structure (for instance a dilational jog model would require a fault which does not outcrop and whose existence cannot be proved one way or the other) in order to generate conditions appropriate for breccia formation.

Accordingly the author sought a method of modelling and visualising the stress state developed around a fault zone of the geometry of the Rusey Thrust and the thrust ramp. Photoelastic modelling fulfilled the requirements in being available, simple and generating modelling results which demonstrated directly and visually the required aspects of the stress regime, i.e the differential stress and principal stress axis orientation.

3.1:1 The Photoelastic Principle

The principle of photoelasticity as a modelling technique is based on the phenomenon of temporary double refraction. This occurs when an optically isotropic material is subjected to loading, bringing about a strain response. An unloaded optically isotropic material has refractive indices that are equal in all directions. When this material is strained the refractive indices rotate and become unequal, acquiring an orientation and magnitude coincident with the principal axes of stress. The difference in the magnitude of the refractive indices causes extinction of certain *wavelengths* of light, which enables the differential stress to be determined.

The orientation of the refractive indices coincides with the orientation of the principal stress axes. The reorientation of these axes under an imposed load causes reorientation of the refractive indices. This causes extinction of certain *orientations* of light wave, which enables the principal stress orientations to be determined. As long as the imposed load does not cause the stresses within the model to exceed the elastic limit of the material, (i.e. that when unloaded the model's strain state returns to zero), then the strain intensity and distribution in a linearly elastic material can be used to determine the stress magnitudes and orientations in the material when loaded. In the following section a brief description of the technique of photoelasticity is presented. For further detail, the reader is referred to texts by Hendry 1966, Dally and Riley 1991 and Connolly, 1996. The latter describes the technique which was used in this study and presents the technique and its justification as well as a modeling program proving its viability and applicability as a geological modeling tool. The technique uses light organized in two different ways, plane polarised light and circularly polarised light.

3.1:2 Plane Polarised Light

Plane polarised light is produced by shining a bright white light source through a standard polarising filter (the polariser) before it reaches the model. The filter will only transmit light waves vibrating parallel to the polarisation direction of the filter, Fig. 3.1. The redistribution of wave vibration directions induced by the model is determined by placing another plane polarising filter, (the analyser) on the observation side of the model. This filter is placed with its polarisation direction at right angles to the previous filter, Fig. 3.1. If the light passes through the model undisturbed, then none will pass through the analyser.

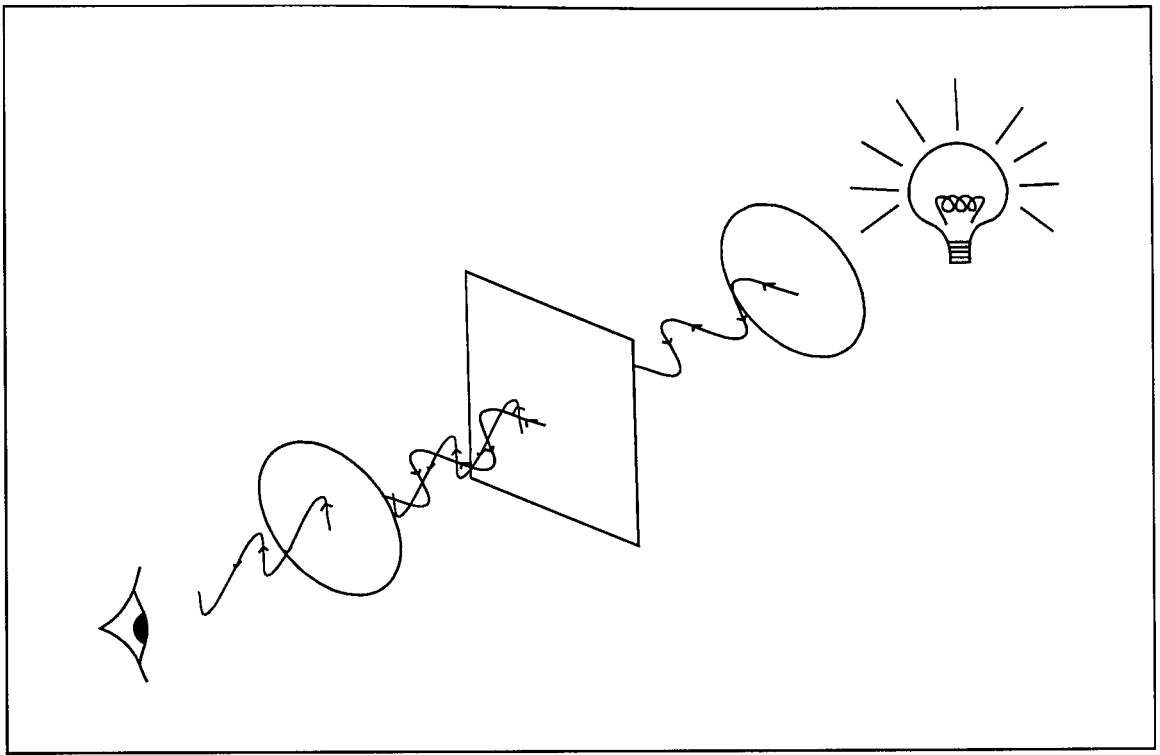


Figure 3.1 Plane polarisation of light

Light that is horizontally polarised by the first filter, (polariser), is split into two mutually perpendicular elements by the model. These two elements are recombined into one light ray on exiting the model and the second filter, (analyser), eliminates all but light rays vibrating vertically.

When plane polarised light passes through the model it splits into two elements, vibrating in mutually perpendicular directions. Elements vibrating in the planes of the principal stresses will not be reoriented and will therefore be eliminated by the analyser. This will produce black areas in the model, which usually take the form of curvilinear lines. These are the isoclinics, Plate 3.1, areas where the orientation of the principal stresses matches the orientation of the polarisation directions of the polarising filters. Rotation of the polarising filters by set increments allows the orientation of the principal stresses to be derived throughout the model.

3.1:3 Circularly Polarised Light

To create circularly polarised light two quarter wave plates are placed between the polarising elements and the model. The polarised light from the polariser passes through

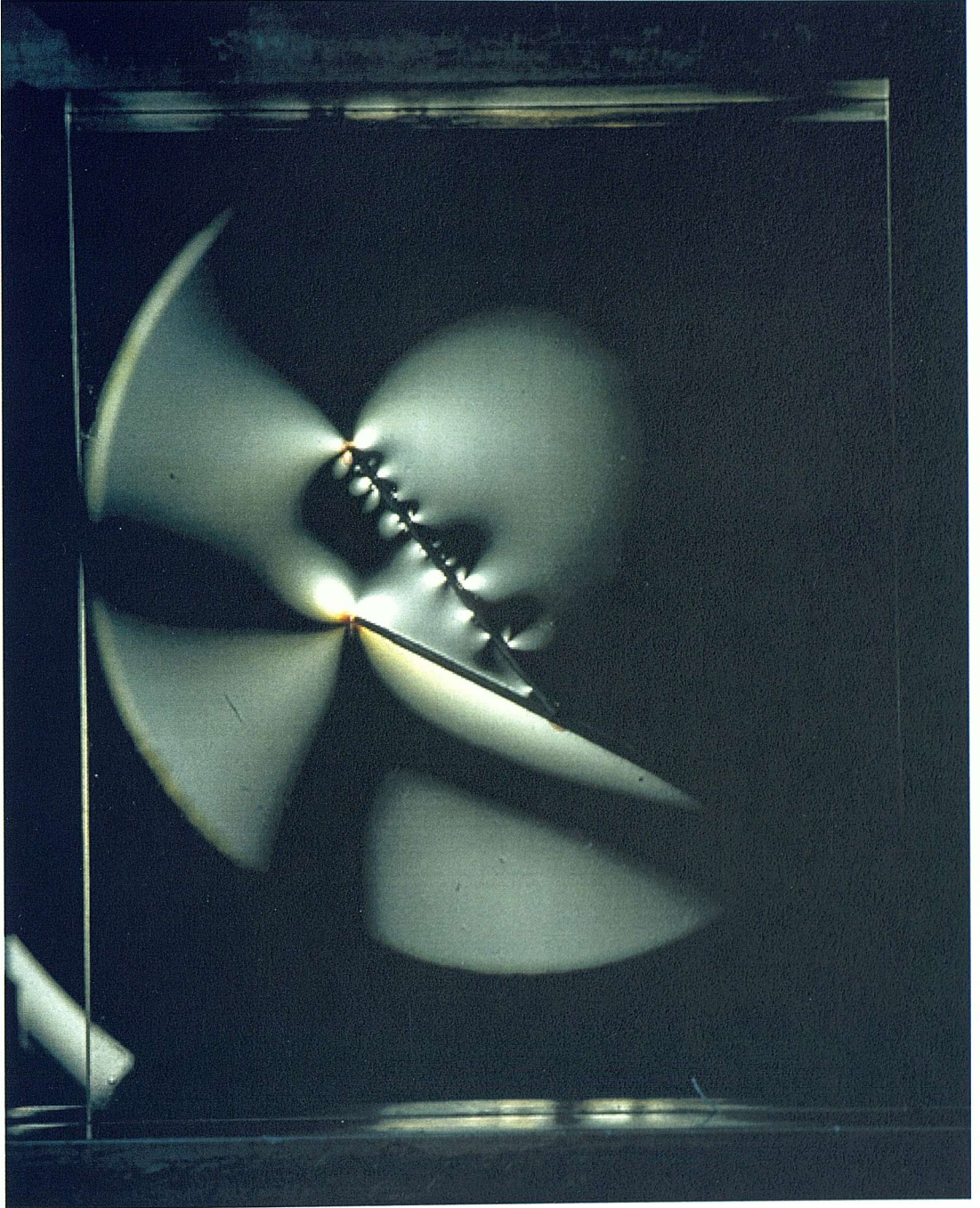


Plate 3.1 Isoclinic lines developed in a photo-elastic model under an imposed stress.

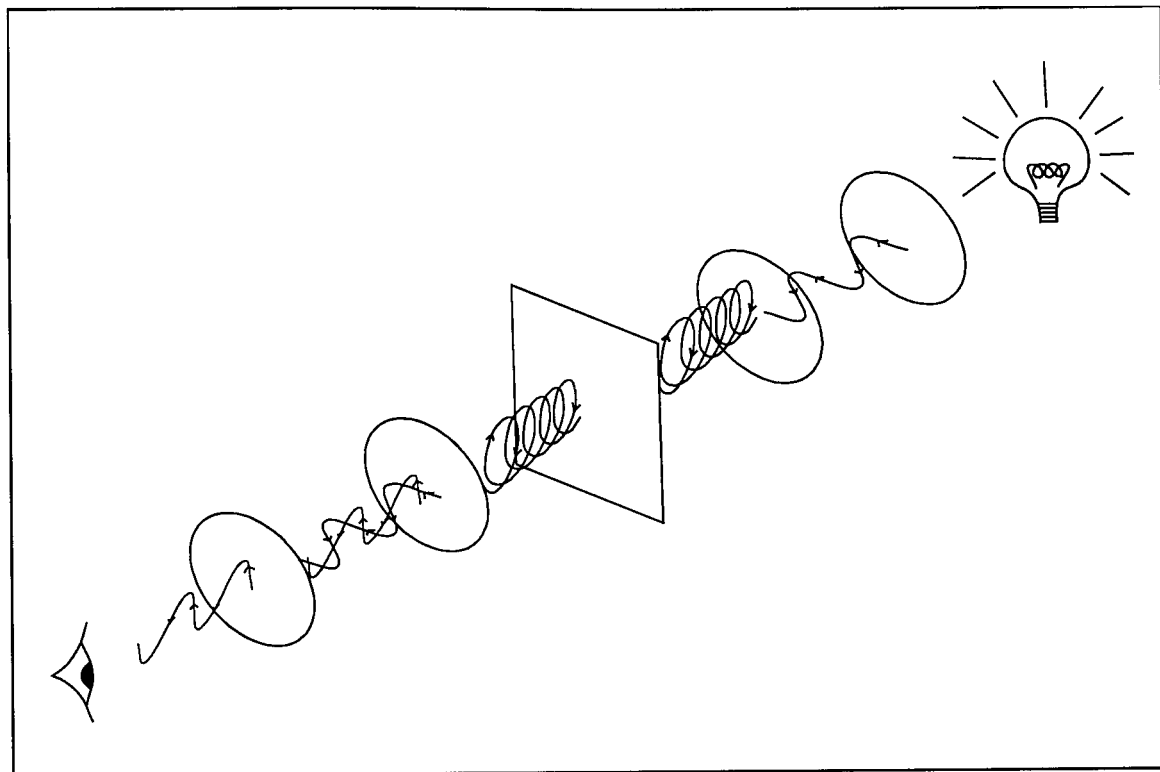


Figure 3.2 Circular polarisation of light.

The circular polarisation is imposed by two quarter wave plates are placed between the polarising elements and the model. The polarised light from the polariser passes through the first, set at a quarter offset, and is converted into light which vibrates perpendicular to the direction of propagation, but in such a way that a point on the light wave will describe a spiral as it moves forward. This effectively produces a wave that vibrates in all directions perpendicular to the direction of propagation. After passing through the model the light is passed back through another quarter wave plate set at three quarters offset, so that the circular motion is neutralised and the light waves consist of the disrupted remains of a plane polarised wave. This is passed through the analyser so that all undisrupted light, i.e. light retaining the orientation imposed by the polariser and which has passed through the model undisturbed, is removed. In fact, because the circular polarisation effect removes the angular dependence of light transmission, only light wavelengths which have undergone destructive interference have not been transmitted.

the first plate, set at a quarter offset. It is converted into light that vibrates perpendicular to the direction of propagation as it moves forward, i.e. a point on the light wave will describe a spiral as it moves forward, (it follows the same path as the tip of a corkscrew as it travels through a cork). This effectively produces a wave that vibrates in all directions perpendicular to the direction of propagation, Fig. 3.2. After passing through the model the light is passed back through another quarter wave plate set at three quarters offset, so that the circular motion is neutralised and the light waves consist of the disrupted remains of a plane polarised wave. This is passed through the analyser so that all undisrupted light, i.e. light retaining the orientation imposed by the polariser and which has passed through the model undisturbed, is removed. The light is split into two elements as it travels through the model. When they recombine on exiting the model, only light wavelengths that have undergone destructive interference have not been

transmitted. Because the light could have been vibrating at any angle as it hit the model, the recombined wave could also be vibrating at any angle and the initial polarising effect is removed. The analyser therefore removes very little light. The polarising filters are necessary for both plane and circularly polarised light observations because they organise the light that is passed through the model. This ensures that all the light that is observed has had its relevant characteristics, (i.e. the direction of vibration and relative retardation of the light waves), determined by the passage through the model.

White light contains light of all wavelengths. When circularly polarised white light passes through the model each wavelength of light is split into two elements that are aligned with the principal stress axes. In this situation the two elements of each wavelength have path lengths through the material which are proportional to the magnitude of the stress axes at that point and therefore when they emerge and recombine, one element of the wave has traveled further than the other, Fig. 3.3. Each wavelength of light therefore emerges with one element retarded relative to the other. When the elements recombine on exiting the material, they interfere. When this interference is destructive, the wavelength is extinguished, Fig. 3.3. This will occur to all the wavelengths of light passing through a given spot on the model, resulting in some wavelengths being extinguished at that point. Consequently the light observed no longer contains all wavelengths of light and is no longer white but coloured, Plate. 3.2. A given point in the model will have a given value for σ_1 and σ_2 . This will produce a given retardation effect for all wavelengths of light passing through that point. This will critically affect one wavelength of light. At another point in the model, where the model is under greater stress, the difference between the magnitude of σ_1 and σ_2 will be greater and therefore the retardation value will be greater, critically affecting light of a different wavelength. Therefore light of a different colour will be observed from that part of the model and a variety of coloured fringes will be visible across the model, Plate 3.2, determined by the level of stress in each part of the model.

In summary, the organisation of light into plane polarised light causes all light hitting the model to be vibrating in a single direction. Therefore when the principal stress axes in the model are oriented in that direction the light is unaltered and is totally removed by the second filter causing black areas to be observed on the model. This allows the identification of the stress axis orientations in all parts of the model. The organisation of light into circularly polarised light means that some part of all light waves penetrating the model will be transmitted by it. This is because the light is vibrating in different directions at different points along its path. Therefore only that part of the wave where the vibration direction coincides with the stress axes will be extinguished by the second polarising filter and the rest of the wave will be transmitted.

This creates a coloured image because each wavelength contained in the white light is split into two elements by its passage through the model and, when recombined, the wavelength will be either enhanced, attenuated or

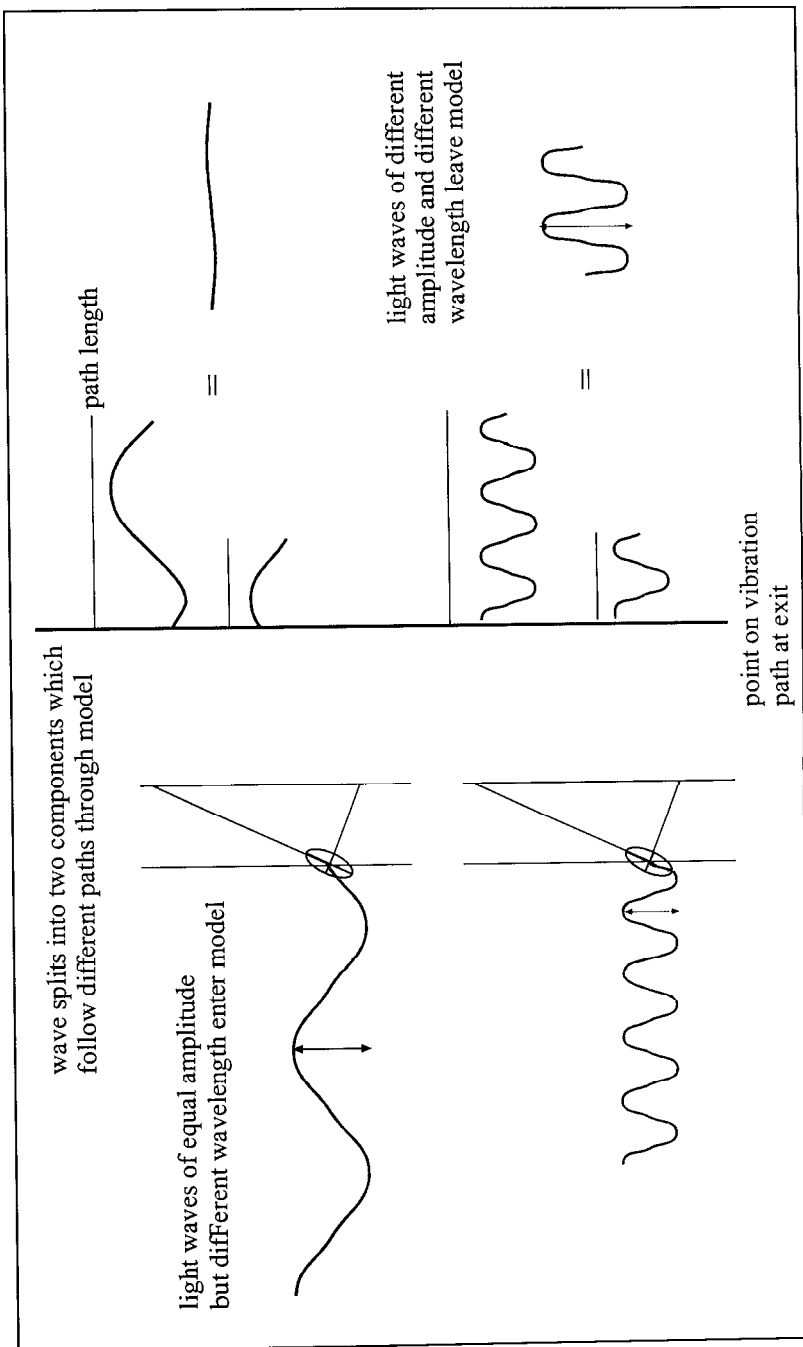


Figure 3.3 Development of coloured fringes in a photoelastic model. Two wavelengths of light, (red and green), are followed on their path through the model. Within that white light, each wavelength of light is split into two elements, aligned with the principal stress axes at the point at which the lightwave impinges upon the model, (stress ellipse shown at point of impact). Each wave element follows a path through the model dictated by the alignment of the wave element with the stress axis, resulting in different path lengths for each element as they traverse the model. When they emerge each element is at a different point on its vibration path and, in recombining, interferes either destructively, (red wavelength) or constructively, (green wavelength). This results in different wavelengths of light being received from different parts of the model. In the instance depicted above, the part of the model at which the light waves are penetrating the model would be observed as being green under this stress state.

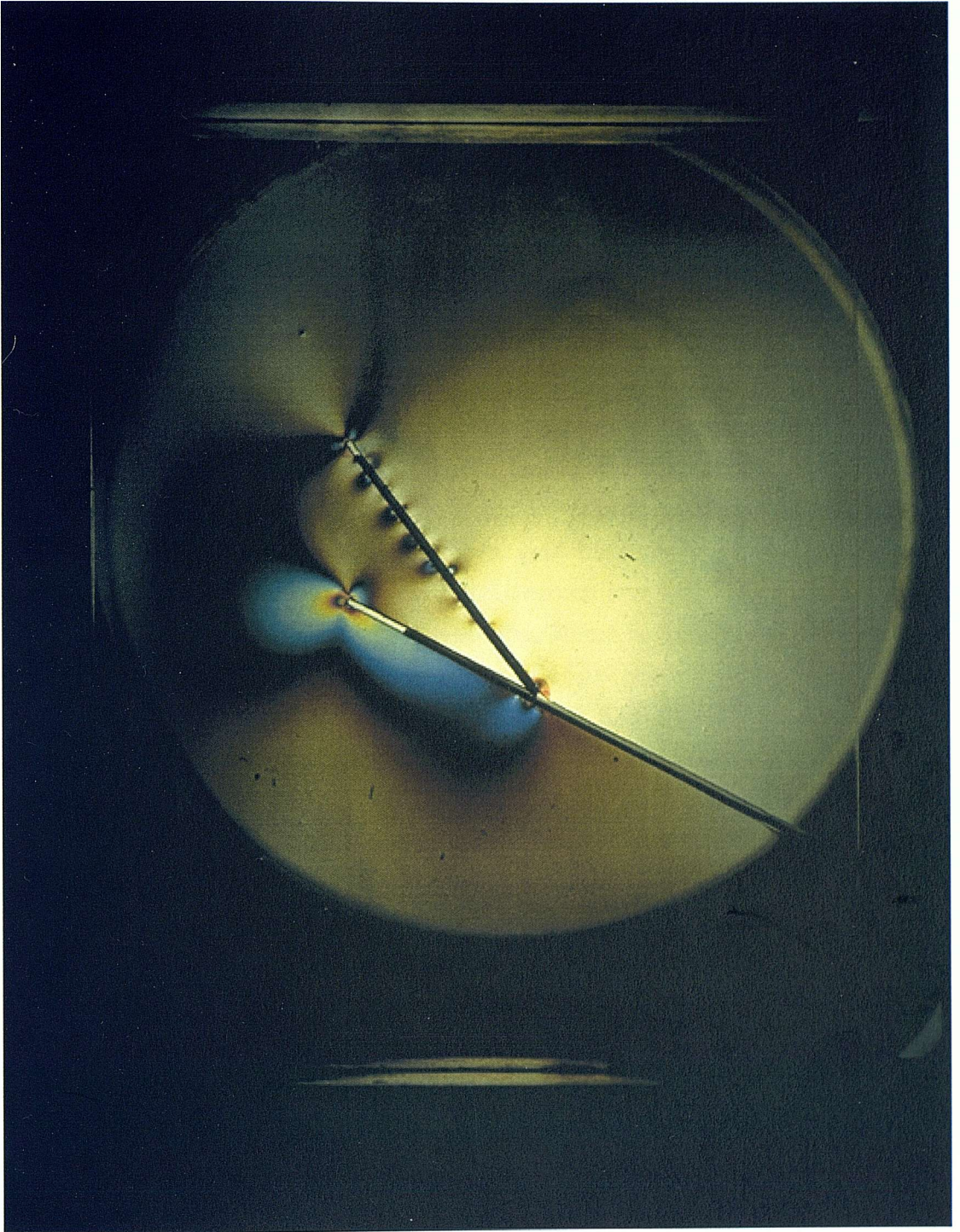


Plate 3.2 Isochromatic lines developed in a photo-elastic model under an imposed stress.

extinguished, Fig 3.3. Because this happens to a different degree to each wavelength of light in different parts of the model, each part of the model will transmit a different mixture of wavelengths of light and will therefore become coloured.

These two properties allow the orientation and magnitude of the principal stresses to be derived from the interference patterns recorded during loading. This enables the stress orientations and the variation in differential stress to be mapped out and a complete stress description for the model to be determined.

3.2 The Modeling Techniques

3.2:1 The Photoelastic Technique

The photoelastic rig used in the modeling program consists of a deformation rig with a polariser/quarter wave plate pair between the light source and the model and the analyser/quarter wave plate pair between the model and the camera, Fig. 3.2. The quarter wave plates can be removed from the light path, allowing either plane or circularly polarised light to be used. A screw gear links the polarising filters. This allows both filters to rotate together. The light source is a 500W tungsten bulb and Kodak Ektachrome tungsten film is used to produce 35mm colour slides.

The deformation rig, Fig. 3.4, consists of two pairs of platens at the top and base of the model. Each pair is clamped together with bolts, thus fixing the model in place. The basal pair is fixed to the rig by bolts. The upper pair is mounted on metal guides, which allow movement in the plane of the model only. This pair has a loading flange mounted on the top, which is pushed by a hydraulic piston, thus causing the upper platens to move sideways. This imposes a shear stress on the model.

During modeling, the load is recorded in pounds per square inch, (psi), by a gauge attached to the piston. This can be converted to the imposed shear stress by the formula;

$$\frac{\text{applied load}}{\text{piston area}} = \frac{\text{shear load on top surface}}{\text{area of top surface}} = \text{shear stress on top edge of model}$$

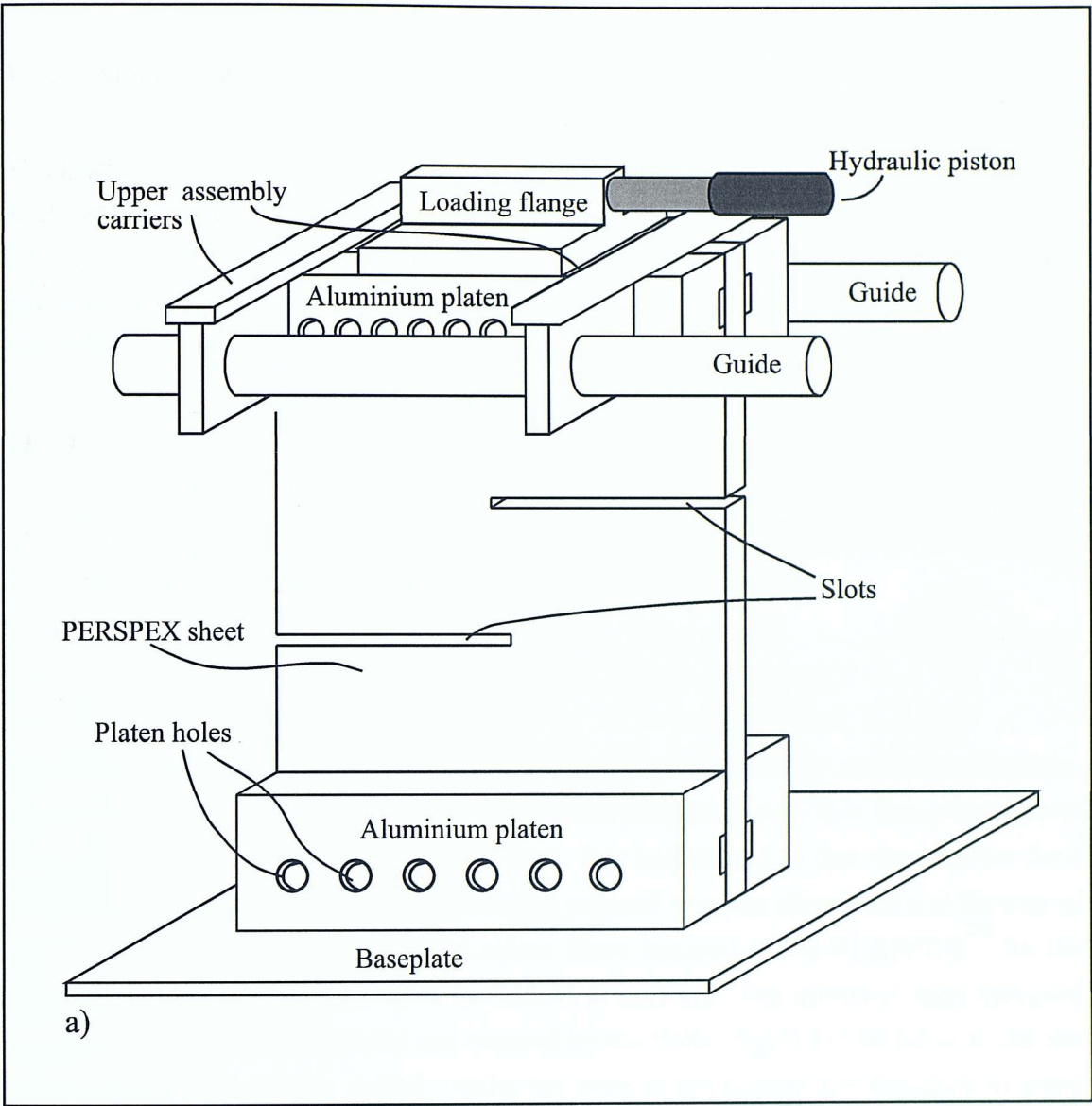


Figure 3.4 Annotated diagram of the photoelastic rig used in the modelling program

The normal load levels used and their conversions are given below:

applied load (psi)	shear stress (KPa)
200	385.07
250	481.34
300	577.61
350	673.88
400	770.15
450	866.42
500	962.68

3.2:2 Model Design and Preparation

3.2:2.1 Model Design

The models used in the modelling program were based on the fault geometries observed at Rusey. R26 has an inter-slot angle of 26° and an angle of inclination of 26° from horizontal for the main slot. These angles were derived directly from dip measurements made on the main thrust and thrust ramp faults in the field. R5 and R10 use the same interslot angle but have a main slot angle of 5° and 10° respectively. This was done in order to confirm that the field-derived geometry was not a special case and that similar behaviour could be observed for other geometries. The range of dip angles on the main slot also gave some measure of sensitivity testing of the modelling. This was regarded as advisable as the angle of the main slot was averaged from the dip measurements on the main thrust, which shows considerable variation in dip in the field.

3.2:2.2 Model Preparation

The models are prepared and mounted so as to produce an initially stress free situation. The blank has slots cut in it to mimic the desired fault geometry. It is then placed in an oven and raised to 120°C over several hours. It is maintained at that temperature for a minimum of twenty-four hours, before being reduced to room temperature at the rate of 10°C per hour. This will ensure that minor flaws induced in the PERSPEXTM by the cutting process are reannealed and the model is uniform. The model is then mounted between the two pairs of platens and secured by the bolts, Fig. 3.4. The holes at the top of the model are precisely drilled, whilst the ones at the bottom are designed to leave minor clearance to allow for adjustment, to ease the task of setting the model in a stress free position.

The top surface is then displaced by the piston. At the desired load photographs are taken, on the correct exposure and also stopped a half aperture stop under and over, to ensure usable photographic data. It is usual that the model is taken to a load of 200-250 psi before any data is recorded. At this load the isoclinics are generally well developed and quite sharp. They are photographed at 0° and then at 15° increments up to and including 75° . When this is done, the quarter wave plates are swung into place and the isochromatic data is recorded. The load is then increased and further isochromatic data recorded at higher load levels. Since the stress distribution is a function of the modeled geometry and that geometry does not change, the stress distribution does not change if the strain is kept below the elastic limit of the material. Therefore only one set of

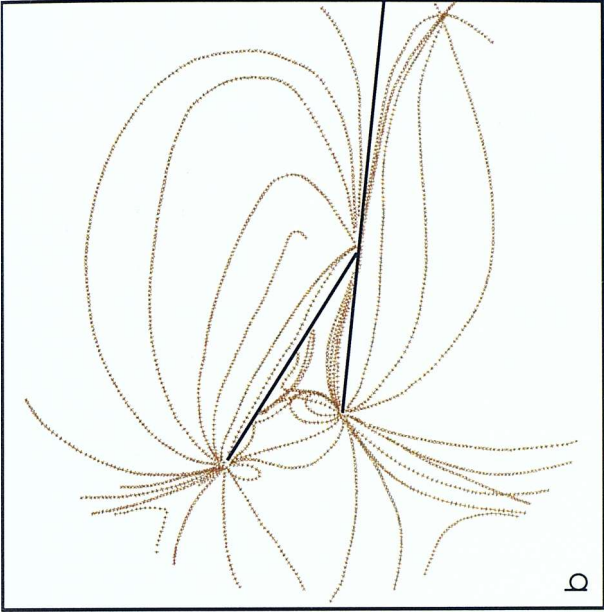
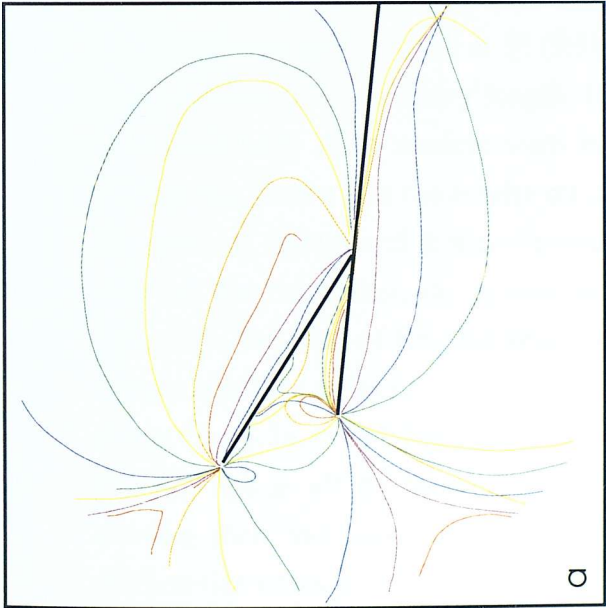
isoclinic data is taken, as soon as they are well defined. The stress magnitudes, however, do change with increasing load and they are recorded up to as high a load as possible.

3.3 The Analysis Techniques

3.3:1 The Principal Stress Orientations

The stress orientations are derived from the isoclinic data. This data consists of a set of slides of isoclinic lines taken at 0-75° polarising filter inclination. The lines show the position of the principal stress axes whose orientations match that of the analyser, Plate 3.1. The procedure is as follows;

- the slide of the model taken at 0° polarising filter inclination is projected onto a piece of paper and a line is drawn tracing each isoclinic line along the centre of the black extinction band. The isoclinic lines from the slide of each filter inclination are traced in the same way onto the same piece of paper, creating a composite image that defines the orientation of the principle stresses over the majority of the model.
- this image is scanned and imported into a drafting package. Each line is traced and color coded according to its polariser filter inclination, Fig. 3.5a.
- crosses are drawn which consist of a red line for σ_1 and a green line for σ_2 , Fig. 3.5b, (σ_2 is used here for the minimum principal stress because the modeling material is subjected to plane strain and is effectively two dimensional). These crosses are rotated until they match a given inclination angle of the polarising filters and are then placed, as closely spaced as possible, along the line which denotes points of extinction at that filter angle.
- when they have been placed along all the lines, that element of each cross which represents the σ_1 orientation is extended to meet its nearest neighbour to provide a map of the change in σ_1 stress orientations across the model and likewise for the σ_2 orientations. These are the stress trajectories for the model, Fig. 3.5c.



KEY

- 90° polarising filter inclination
- 75° polarising filter inclination
- 60° polarising filter inclination
- 45° polarising filter inclination
- 30° polarising filter inclination
- 15° polarising filter inclination
- σ_1 , (red) and σ_2 (green) orientation

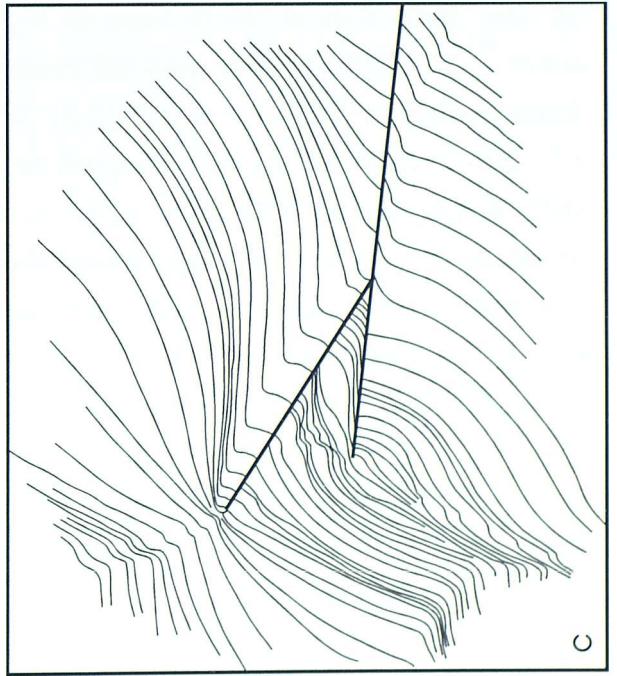


Figure 3.5 Conversion of raw isoclinic data to map of σ_1 Orientations.
a: composite map of isoclinic line stress axis orientations at different polariser inclinations
b: overlay of σ_1/σ_2 orientations along isoclinic lines
c: σ_1 orientations across model

3.3:2 The Sum of the Principal Stresses.

The technique for determining the sum of the principal stresses hinges on the fact that;

- the principal stresses at the boundary of the model are perpendicular and parallel to that boundary, because these boundaries are effectively free surfaces. Therefore, as the principal stress perpendicular to a free surface has a value of zero, the differential stress at that point must be equivalent to the principal stress oriented parallel to that surface. The differential stress is known from the points where the isochromatics intersect the model boundaries and the slots.
- photoelastic modeling materials are Laplacian in behaviour, i.e. an increase in stress produces an increase in strain in precisely the same relationship in either the x or y direction at any point in the model. Therefore a simple interpolation method, (finite difference) can be used to derive the values of the differential stresses for points falling between the points of known value lying along the isochromatic lines

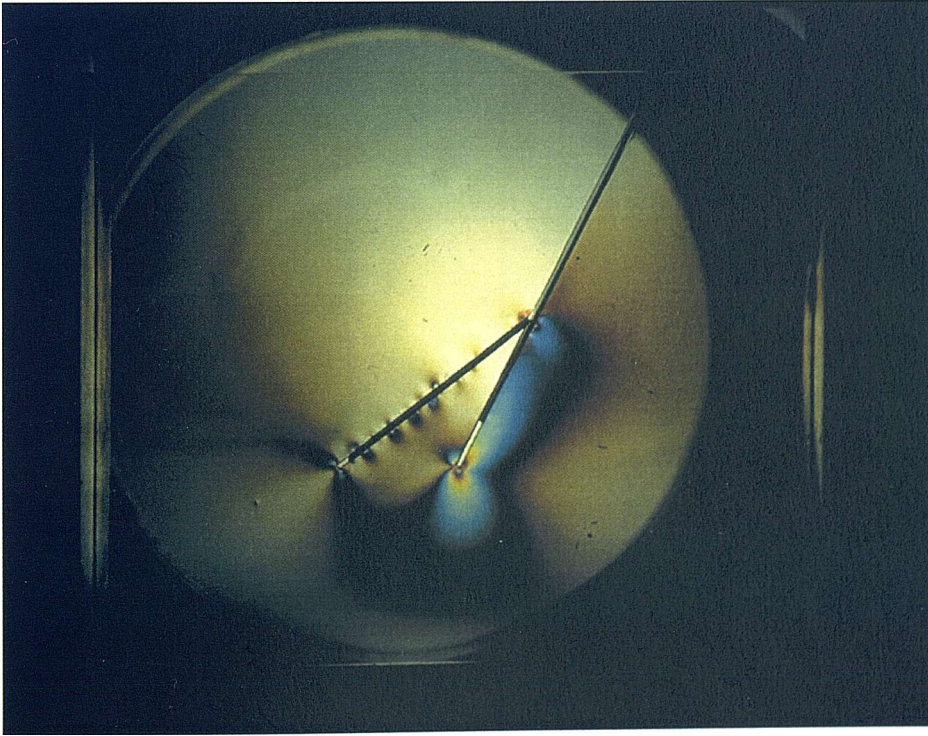
The result is a contoured plot of differential stress which looks virtually identical to the isochromatic image, Fig. 3.6a and b.

3.4 Modeling of the Rusey Thrust System

3.4:1 Isochromatic Data

Fig. 3.7 shows the differential stress maps drawn for models R5, R10 and R26 at 300psi. The models take the form of two slots cut at an angle of 26° to each other, with the upper most slot being the steeper. This geometry has been cut into PERSPEXTM sheets with the lower slot inclined at 5° (R5), at 10° (R10) and at 26° (R26) to the horizontal. The slots are of an arbitrary length. R26 was designed from the Rusey field data. For this study only three models were run, as previous work, (P.Connolly, unpub. PhD thesis), has shown that the results are precisely reproducible. The models were loaded to the point of failure and it was discovered that the greater the angle between the lower slot and the horizontal, the greater the applied load that the system could bear before failure. In the case of R5, this was a load of 300psi, for R10, 350 psi and for R26 the failure load was 600psi.

From Fig. 3.7a, b and c it can be seen that foci of high differential stress developed at the slot tips in all the models. The lower tip developed higher magnitude differential stresses than the upper one. In addition, at very high loads, a third focus of high differential stress developed in R26, located below the lower slot at the intersection of the two slots, Fig. 3.7d. Although initially a minor perturbation, this stress focus rapidly acquired a differential stress of a magnitude similar to those found at the slot tips. At 600psi, it has a stress concentration as high as that of the upper slot, extending over an area some four or five times greater. However the greatest stress concentration remained

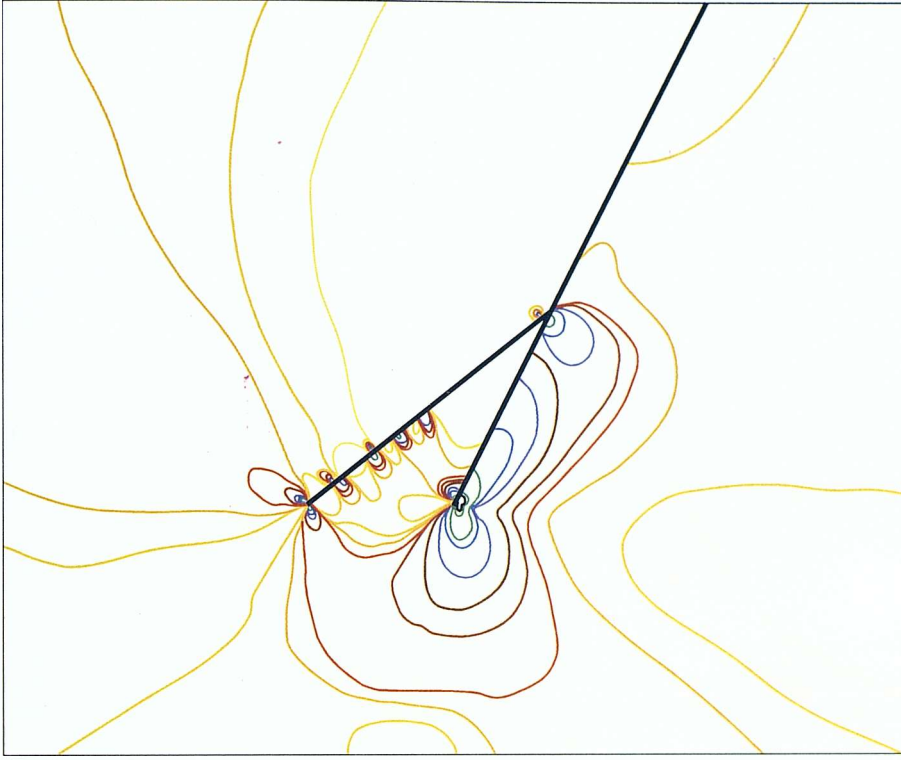


a

Figure 3.6 Isochromatic data for R26 at 600psi

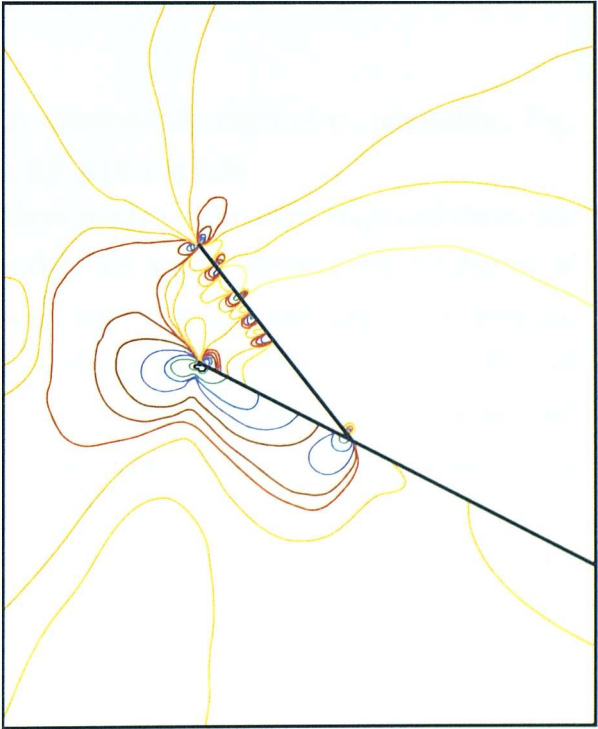
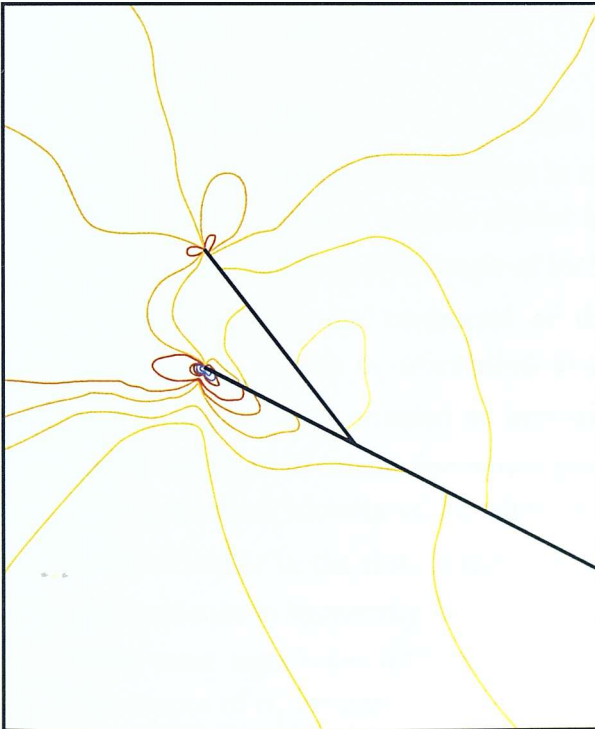
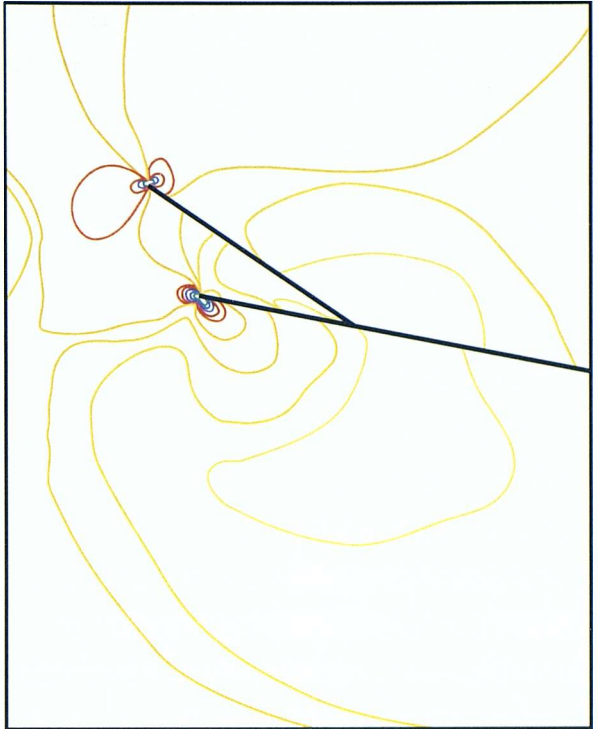
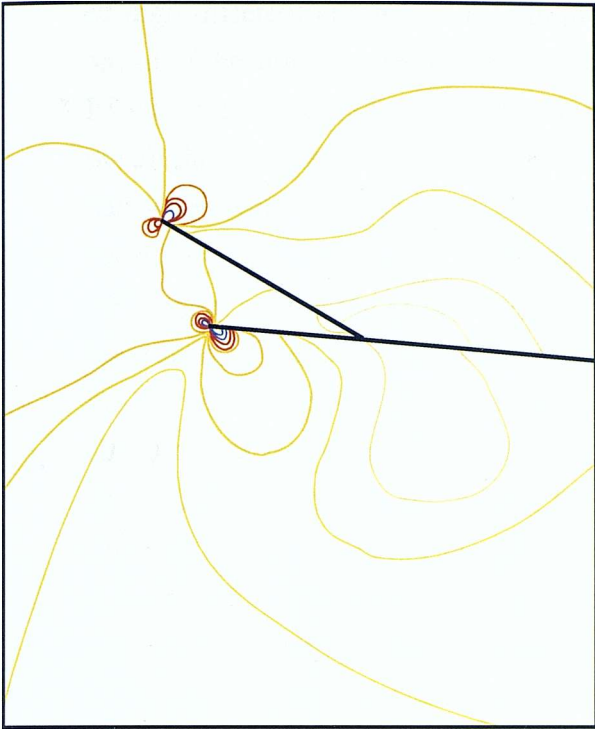
a: Initial photographic data

b: Contoured differential stress map.



b

Arrow indicates direction of increasing differential stress. Absolute magnitudes are scaled to the model and are not geologically relevant.



a

B

C

D

Figure 3.7 Contour maps of differential stress levels within the R series models:

- a: R5 at 300 psi load
- b: R10 at 300 psi load
- c: R26 at 300 psi load
- d: R26 at 600 psi load



Arrow indicates direction of increasing differential stress.

centred on the lower slot tip. In the R5 and R10 models, the differential stress foci remained limited in areal extent and the majority of the model is subjected to relatively low differential stress. In R26, the lower slot tip developed an extensive bi-lobate area of high differential stress that is limited to the area below the lower slot and the area ahead of the slot tips, (i.e. in Fig. 3.7, the area to the left of the slot tips).

Bi-lobate areas of high differential stress are also developed around the upper slot tip in the all three models. The high differential stress area developed around the upper slot tip had lobes which extended sub-perpendicular to the slot and this relationship remained approximately constant for all models. However the orientation of the lobes of high differential stress around the lower slot tip varied with the angle of inclination from the horizontal of the lower slot. In R5 the lobes extended at 25° degrees above and ahead of the slot tip and at 40° below and behind it. In R10 these orientations were parallel to and ahead of the slot tip and 27° below and behind it. In R26 they became 35° below and ahead and <10° below and behind the slot tip. Comparison of R10 and R26 at 300psi suggests that the extent of the area of high differential stress may have been somewhat greater in R26. The area of high differential stress that developed below the lower slot at the slot intersection point is perpendicular to the lower slot. There was another minor stress concentration in the hanging wall where the upper slot intersects the lower, Fig 3.7d, but it was of negligible magnitude.

3.4:2 Isoclinic Data

The isoclinic data, Plate 3.1, has been used to construct the maps of σ_1 orientation, Fig. 3.8a, b and c, surrounding the slots in models R5, R10 and R26.

The σ_1 distribution is broadly similar in all three models, with high angles of incidence between σ_1 and the slots. Angle of incidence does not appear to vary with the degree of inclination from the horizontal of the lower slot, although the degree of rotation experienced by the σ_1 orientation does vary with the slot inclination as the far field stress is invariably oriented at approximately 45° to the horizontal. In models R5 and R10, the far field orientation would produce a low angle of incidence with the slot but in the immediate vicinity of the slots, the σ_1 orientation rapidly rotates to an orientation perpendicular to the slot. In model R26 there is virtually no reorientation and the angle of incidence is apparently lower, at approximately 65-70°.

The most significant difference between the models is that R26 has a high angle of incidence of σ_1 orientation with the slots along their entire length, including the interslot area, whereas in R5 and R10 the σ_1 orientation undergoes rapid reorientation in the area below the lower slot and behind the slot intersection, and in the interslot area, immediately above the lower slot. Below the lower slot σ_1 reorients to a low angle relationship with the slot, Fig. 3.8a and b. In the interslot area σ_1 rotates until it is subparallel to the lower slot, Fig. 3.8a and b. It is noticeable that the stress trajectories

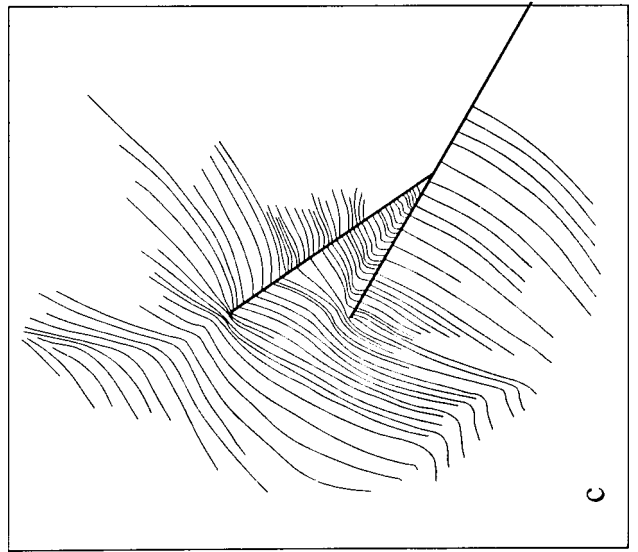
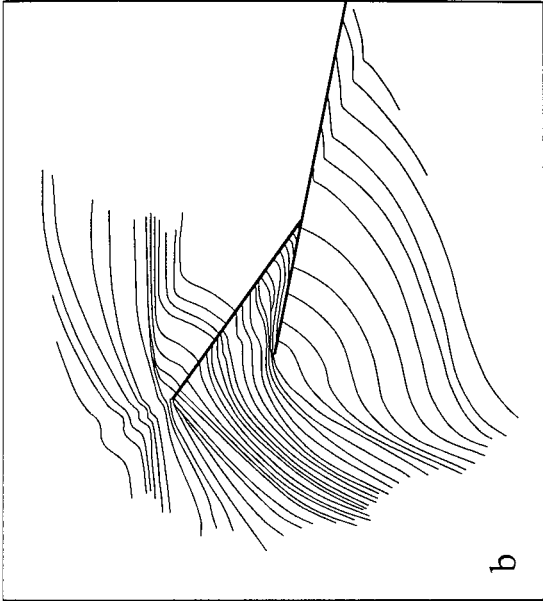
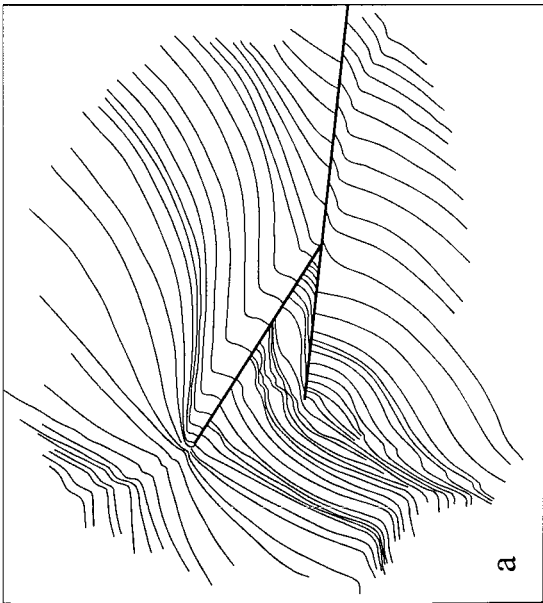


Figure 3.8 Comparison of σ_1 orientations for
the R series models
a: R5
b: R10
c: R26

are subject to a progressively lesser degree of minor orientation swings as the inclination of the system increases.

3.4:3 Discussion of the Isochromatic Data.

Areas of stress concentration are areas that would, if the system were taken beyond its elastic limit, be more likely to fail. Therefore the development of an area of high differential stress below the slot intersections in model R26 is of great interest as it coincides precisely with the area of the development of the Rusey Breccia, which is located in the footwall of the Rusey Thrust at the launch point of a ramp system.

The isochromatic data is useful in constructing a model for the structural evolution of the Rusey outcrop. The present author proposes that trapping of regionally evolved formation waters and metamorphic fluids by the thrust fault created a fluid reservoir in the thrust footwall. This reservoir, developed early in the evolution of the system, then became overpressured as the stress levels in the area built up. The trapping of fluid here, as a result of low cross-fault permeability, resulted in the development of thin veins filled with cryptocrystalline quartz and areas of quartz mineralisation showing diffuse margins, indicating mineralisation by replacement of the host rock rather than mineralisation of open fractures. If failure along the main thrust did not occur it can be seen that the lower, lattermost lobe of enhanced differential stress, (Fig. 3.7), would have extended back along the thrust footwall. It would be predicted that the fluids trapped under the thrust would raise the pore fluid pressure to levels where hydrofracture occurs within the area of footwall subjected to higher differential stresses. This would create an area of brecciation and veining, with high fluid pressures maintaining open fractures whilst quartz crystallised to form the Drusy textures seen preserved in the footwall to the north of the ramp launch point. Originally this fracturing and Drusy texture mineralisation extended back into the area below the ramp launch point as would brecciation associated with the very first stages of ramp formation. This hypothesis is supported by the evidence from Rusey which indicates the area now occupied by the Rusey Breccia was saturated with silica bearing fluids which have replaced parts of the country rock and were involved in multiple episodes of dilation and mineralisation. Polygonal fracture networks, Plate 3.3, and the presence of ladder veins, Plate 3.4, in otherwise apparently unfractured rock indicate that early failure episodes were purely dilational or took place under conditions of low differential stress, as may be expected if failure was facilitated by high fluid pressure as the local stress levels increased.

This is made more evident by the form of the polygonal fracture network. It can be seen from Plate 3.3 that the network shows a strong tendency to a rectilinear fracture pattern. These veins show crystal growth perpendicular to the wall and removal of the vein material would permit the rock to be reassembled, i.e. the rock fragments have not been

rotated, only separated. It would therefore seem that the rock has undergone a pure dilation. From the Mohr diagram, Fig. 3.9 it can be seen that for a rock to fracture in the tensile field both the effective stress, (stress minus pore fluid pressure) and the differential stress must be low, (blue circle). If the effective stress is high with a low differential stress, failure will not occur, (red circle), and if the differential stress is high, failure will occur in the regime of shear failure, (green circle).

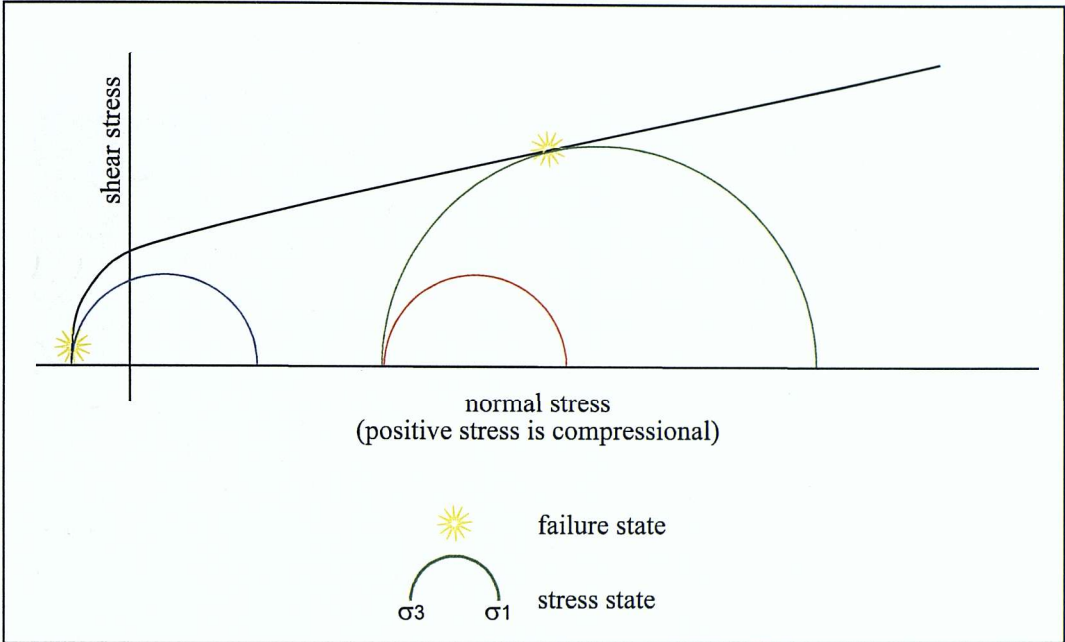


Figure 3.9 Mohr diagram to show the requisite stress states for shear and tensile failure.

For tensile failure to occur the stress state must satisfy the criteria,

$$\sigma_3 - p > T \text{ and } (\sigma_1 - \sigma_3) < 4T$$

i.e. that the least compressive effective principal stress must have become sufficiently negative to have overcome the tensile strength of the rock (T). and that the differential stress is low., (Price and Cosgrove, 1990). The $\sigma_3 - p$ term cannot exceed T if the differential stress is high, since an increase in p would decrease the effective magnitude of both σ_1 and σ_3 equally, causing the effective stress state to satisfy the shear failure criterion before $\sigma_3 - p$ could have acquired a sufficiently large tensile magnitude to exceed T and satisfy the tensile failure criterion. Since burial and lithification would only have imposed compressive stresses and the regional tectonic stress regime was compressional from the time these rocks were deposited it is unlikely that any of the imposed principal stresses were tensile. Therefore the pore fluid pressure in the rock must have been great enough to overcome the imposed compressive stress and exert an effective tensile stress, i.e. the effects of pore fluid pressure were sufficient to overcome the minimum principal stress and the cohesive strength of the rock, the defining

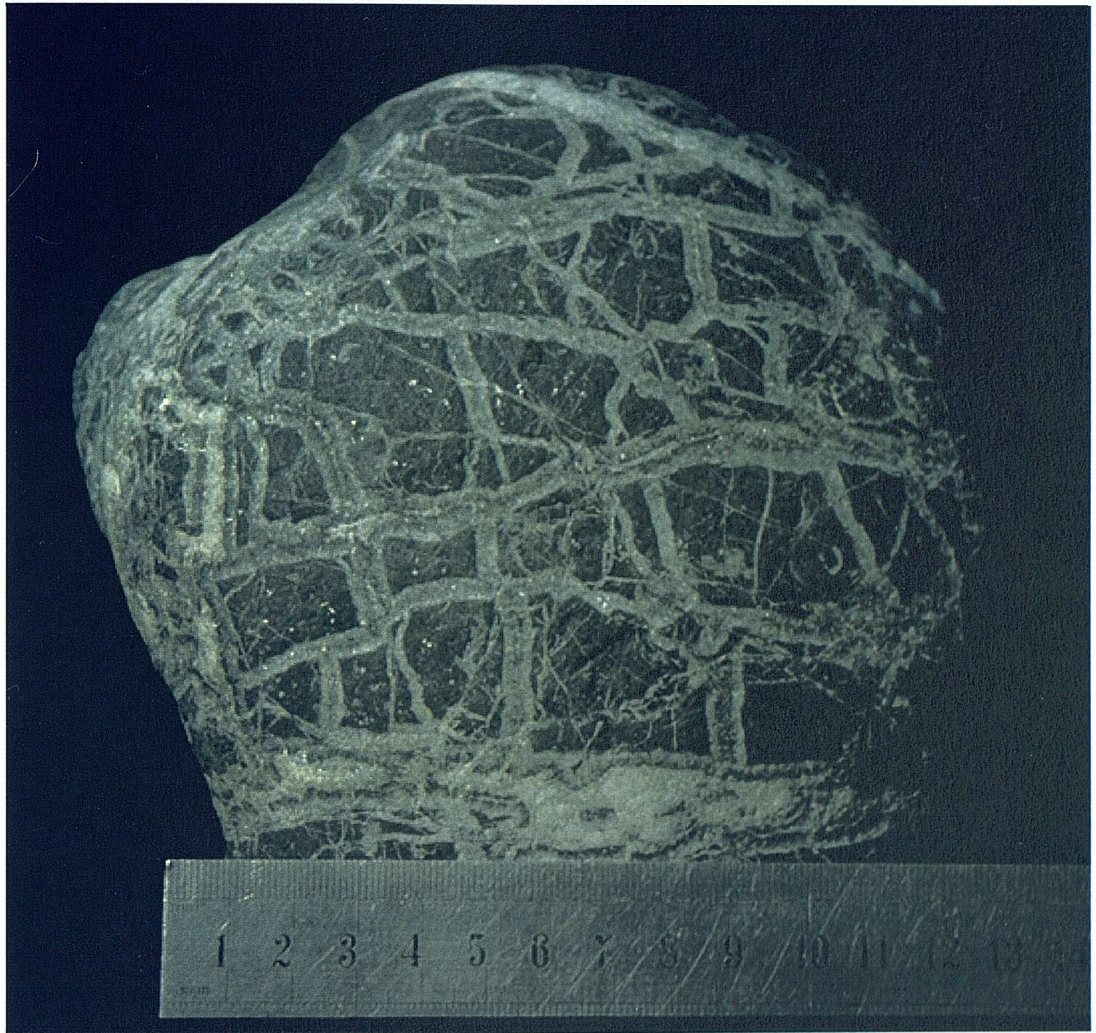


Plate 3.3 Polygonal vein network, from the shales surrounding the Rusey Breccia

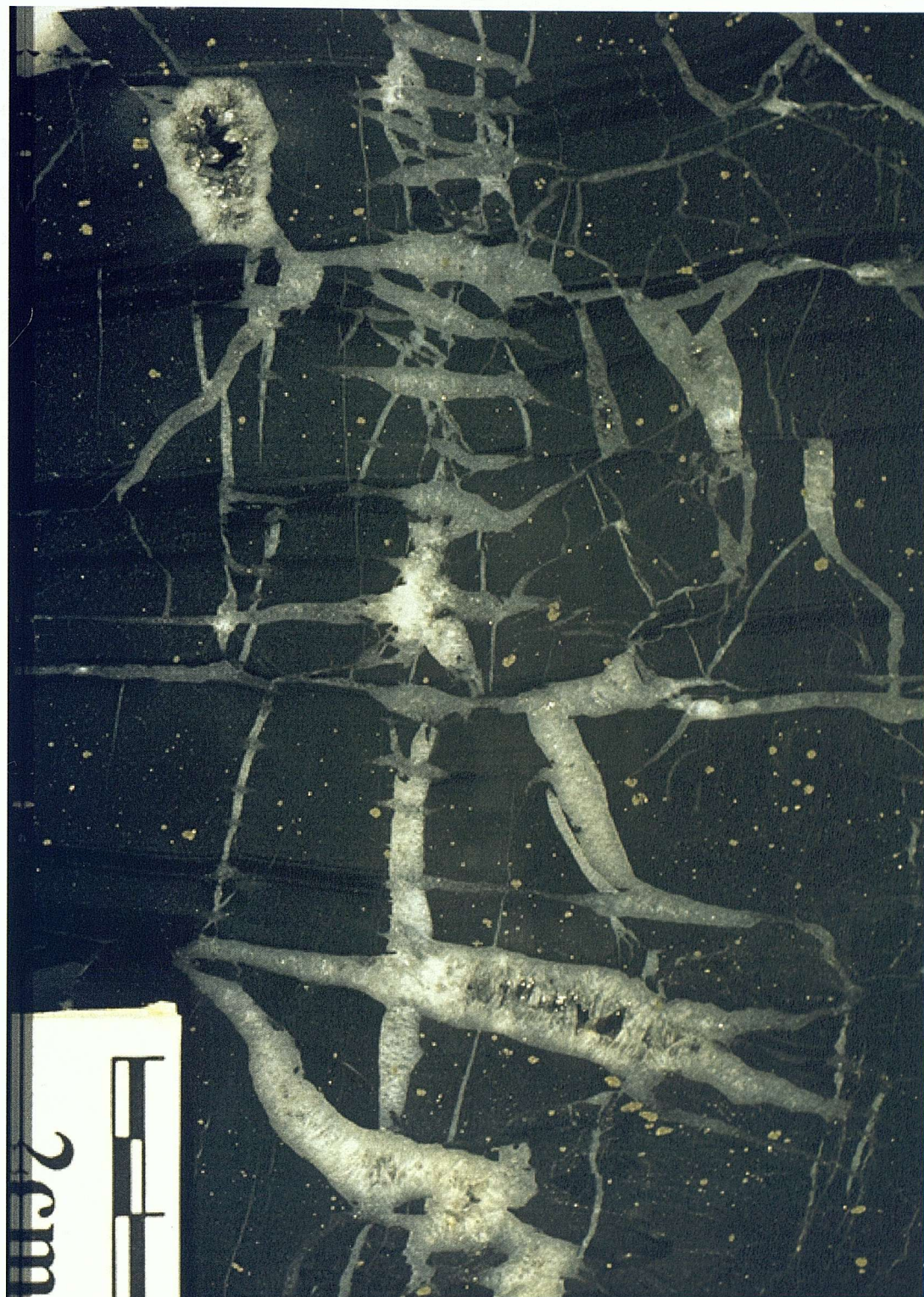


Plate 3.4 Ladder veining from the shales surrounding the Rusey Breccia

condition for hydrofracture, (Hubbert and Rubey, 1959, Gretner, 1965, Price and Cosgrove, 1990, Sibson, 1988). The mineralised state of the veins also implies that large amounts of fluid were available to maintain open fractures and import mineralising material. The polygonal and ladder geometries of the fracturing indicates that σ_2 and σ_3 were approximately equal. The presence of tensile fracturing indicates that the differential stress, $(\sigma_1 - \sigma_3)$, must have been low and the pore fluid pressure high to cause the effective stress state to satisfy the tensile failure criterion.

This would place the hydrofracture episode either early in the deformation history, i.e. before the differential stress reached values appropriate for shear failure or during the dynamic transient stress regime that operates during failure events, (Sibson 1988, 1992). As the area of the tensile fracturing is almost coincident with the area of the mature breccia development, the present author proposes that fault slip related hydrofracture of the area underneath the thrust ramp developed a brecciated area of rock which repeatedly failed under the stress concentrations associated with the thrust ramp development and that part of this area immediately beneath the ramp lift-off point developed into a modified organized brecciated unit in the footwall of the main thrust.

Numerical modeling, (Smith and Wiltschko, 1996), has shown that even without locally excessive pore fluid pressures, high fluid pressures may be expected to build up immediately underneath ramp thrusts as the thrust develops. They regard the buildup in pore fluid pressure as making the base of the thrust ramp a preferred site for failure. The presence of a high pressure fluid reservoir held in the brecciated volume of rock causes the system at Rusey to develop a significant variation in deformation style from the expected occurrence of failure along the thrust ramp. As the thrust ramp begins to develop, the presence of the weak brecciated rock, high fluid pressures and large amounts of fluid cause failure within the brecciated rock, rather than along the developing ramp zone. If fluid pressures were lower and less fracturing of the wall rocks had occurred, the ramp zone would be the weakest point and would fail first. However, at Rusey, the development of the ramp system is arrested whilst failure is relocated to within the breccia body. With each slip event, the dilatancy compaction mechanisms described by Segall, (1995), become effective. As slip initiates, the slip zone, (or critical slip patch), dilates, causing fluid to be drawn in. The dilation of the area causes the fluid pressure to drop. The viability of failure is critically determined by the fluid pressure. As fluid pressure drops, the breccia material 'stiffens' and failure is no longer possible. The slip patch becomes sub-critical and the velocity of displacement across it decreases, preventing the slip event from propagating. The fluid intake causes the development of Rusey type textures by mechanisms which will be described in Chapters 4, 5, 6 and 7. Once the area of the breccia located in the thrust footwall under the ramp launch point is mineralised, it is both stronger and less permeable. The drop in permeability restricts the fluid movement. This restricts the effectiveness of the dilatancy-compaction cycle and

the failure cycle is broken. Stress levels must increase before slip can be initiated and when failure occurs it is located within the developing ramp zone. The hydrofracture texture which developed in the footwall as slip ceased on the main thrust is reworked by the final stage of breccia development only in the area below the ramp and the situation observed at Rusey, i.e. a layered, graded breccia body surrounded by fractured and mineralised host rocks, is developed. The majority of the subsequent compressive deformation is accommodated by the thrust and the ramp and significant deformation of the brecciated area ceases.

The results of the photoelastic modeling are used to evaluate the evolution of the stress regime under which brecciation took place and the brecciated system evolved. They are of considerable help in defining the environment of formation and evolution of the Rusey breccia when combined with certain deductions regarding the geological environment of the Rusey Fault;

- that the combination of rapid burial and early metamorphism caused by the Variscan orogeny caused an unusual amount of formation water and evolved metamorphic water to be trapped under the Rusey thrust, generating fluid pressures greatly in excess of those occurring in thrust systems developed in more mature rocks
- that the high fluid pressure permitted extensive hydrofracture under the stress system observed in the models
- that the presence of high fluid pressure and a highly brecciated protolith allowed the dilatancy-compaction mechanisms proposed by Segall to become the controlling factor in the very early development of the thrust ramp.

The photo-elastic models therefore permit the proposition of a theoretical model for the complete generation and evolution of the Rusey Breccia.

3.4:4 Discussion of the Isoclinic Data

The high angle of intersection of the σ_1 orientation with the slots observed in the models is to be expected in that the slots are free surfaces. In such a situation the stress orientations must rotate until the axes of principal stress are parallel and perpendicular to the free surface. Whilst this is an artifact of the model, it is not necessarily an inaccurate representation of the modeled situation, (Connolly 1996 for a discussion of the model boundary conditions). Although natural faults would be expected to have a degree of cohesion and are therefore not free surfaces, there is evidence to suggest that there are certain conditions in which the faults are at best extremely weak. It should be noted that subsidiary faults of the San Andreas system in the vicinity of the major fault planes rotate in orientation from low to high angle with respect to the major fault plane (Byerlee, 1992). This has been explained in terms of the effect of high fluid pressures within the fault zone.

3.5 Summary

The model proposed for the Rusey Breccia states that appropriate conditions for tensile failure and brecciation were met in the early stages of ramp formation and are specifically related to the transient stress regimes existing *between* fault slip events. The presence of the developing ramp created and located the brecciation. The brecciation then briefly relocated the focus of deformation into the brecciated body until the breccia becomes mineralised. At this point the focus of deformation returned to the main fault planes and the developing ramp system. The early presence of the developing ramp and deformation by stick-slip activity on the fault planes

is a pre-requisite of the model. This is because the presence of the developing ramp creates the stress focus proposed to be responsible for breccia formation. Whilst the ramp development is abandoned during breccia re-organisation this is not a temporally significant part of the development of the thrust system. This is because the time scale of the processes for thrust system development is thousands to millions of years, whereas the time-scale of the processes for re-organising and mineralising the breccia body is proposed to cover minutes to days.

The photoelastic modeling of systems matching the geometry of the Rusey Thrust and its ramp system indicate that such a system could potentially produce intense hydrofracture and mineralisation in areas which match those in the natural example showing fracturing and veining. The geometry and properties of the natural fracture system indicate that tensile hydrofracture mechanisms were operating, requiring high pore fluid pressure and low differential stress. This model shown in Fig. 3.10 is proposed to account for the brecciation and mineralisation features seen at Rusey Headland. It utilises all the structures mapped in the field and does not require the existence of structures that have not been observed in outcrop. For these reasons it is preferred to other models, e.g. formation in a dilational jog, Fig. 3.11, which require additional faults for which there is no field evidence. The author accepts that these possibilities are valid, but wishes to propose a model of the evolution of the area based on mapped field evidence.

There is evidence that supports the idea that the stress regime around active faults with low cohesion is modelled accurately by the photoelastic models. The reorientation of the stress axes approaching the model slots has been supported by field evidence, Byerlee, 1992. This previous work associated the presence of fracture reorientation with the presence of high fluid pressures, which would cause fault weakening by reducing the frictional strength of the fault (Sibson, 1992). Since there is considerable evidence for high fluid pressures in the Rusey system, it is therefore proposed that the models

-  fluid reservoir
-  over pressured fluid concentration
-  highly over pressurised fluid concentration

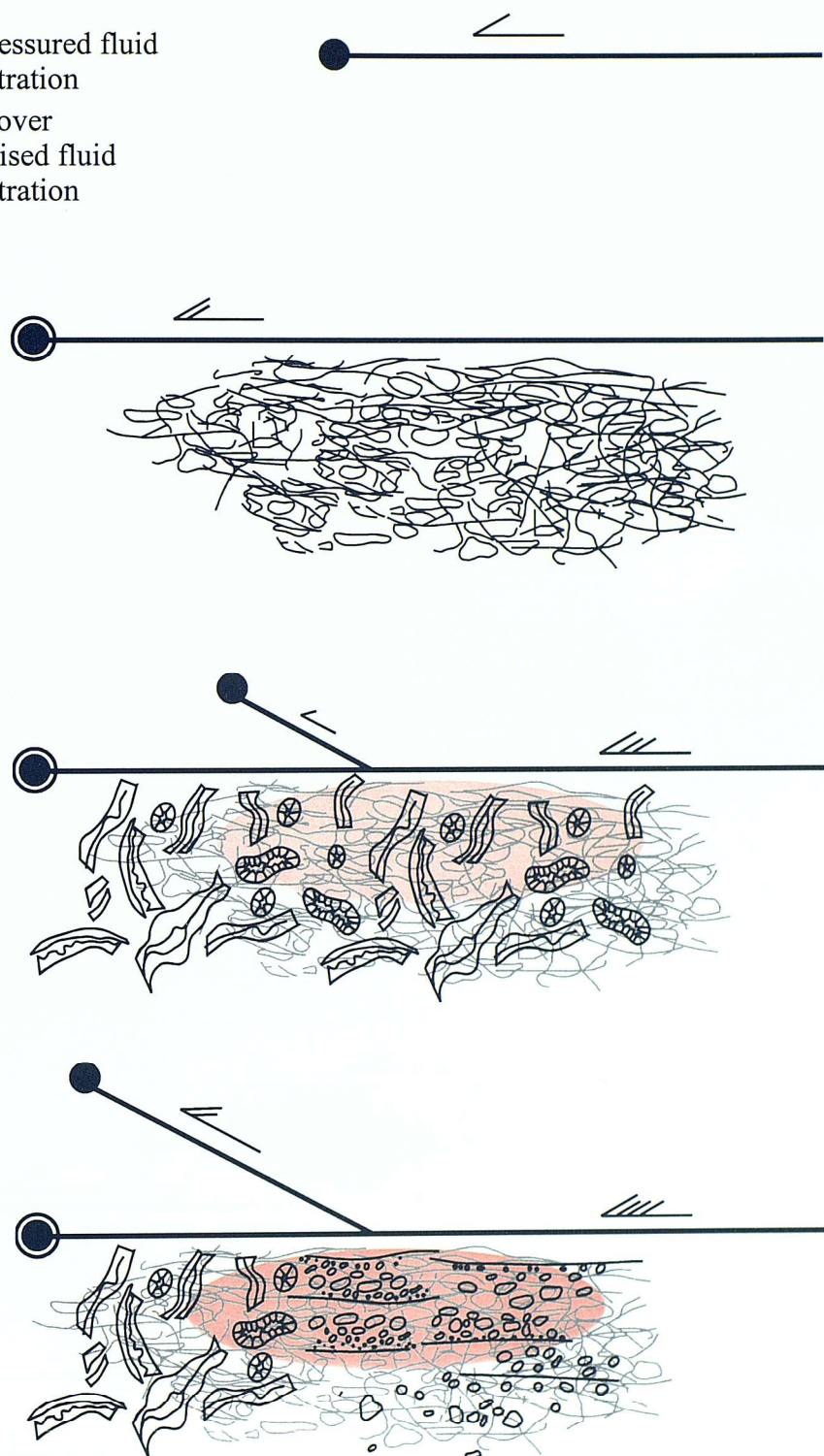


Figure 3.10 The thrust ramp model for breccia formation along the Rusey Thrust

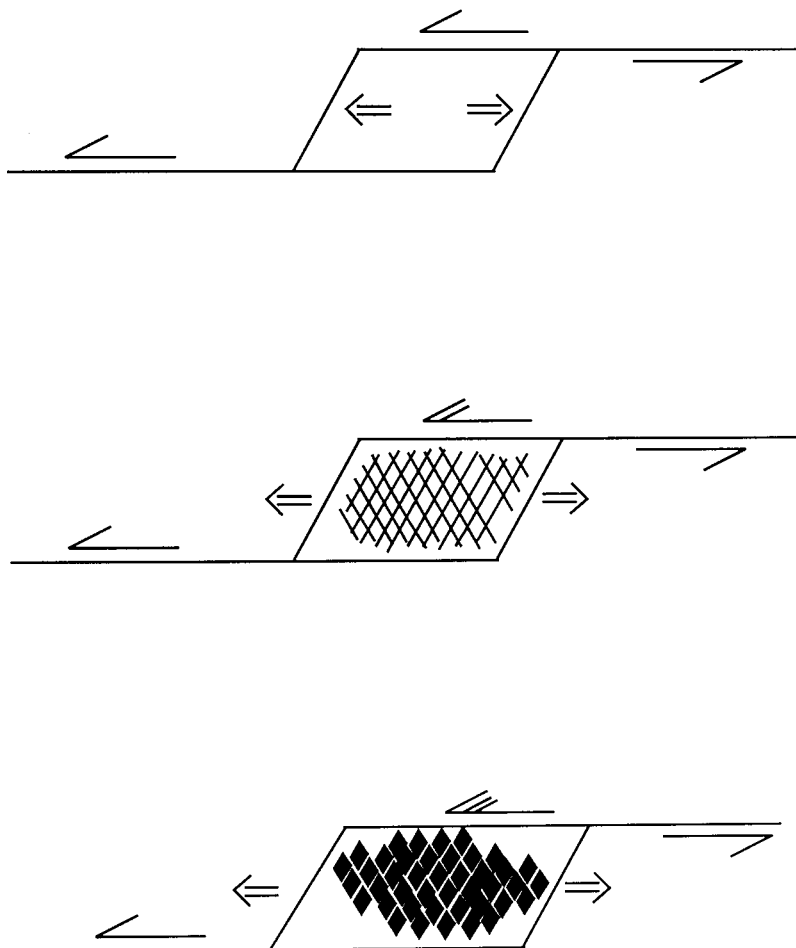


Figure 3.11 Breccia formation due to extension within a dilational jog along the Rusey Thrust

provide a valid approximation of the stress regime that operated at Rusey and can be used in constructing an evolutionary model. However this model provides little insight into the processes which reordered and mineralised the breccia after it was formed. The following chapters, (4, 5 and 6), present more detailed studies of the breccia and matrix textures and consider possible mechanisms that could generate them.

Description and Analysis of the Rusey Breccia

Chapter Four

Field description and statistical analysis of the Rusey Breccia

4.1 Introduction

The main breccia band, Plate 4.1, is between two and four metres thick and has an outcrop length of 80 metres. The clasts are shale or re-brecciated breccia and the matrix is monomineralic quartz. The clasts are matrix supported and don't make contact with each other in any section orientation, Plate 4.2a and b. In the samples assessed, the volume of clast material never exceeded 50% of the total volume and was generally approximately 30%. In the first part of this chapter the characteristics of the breccia will be described under two headings;

- structural and organisational features, i.e. the spatial arrangement of the breccia clasts
- mineralisation features, i.e. the texture and structure of the quartz mineralisation which forms the matrix of the breccia.

In the second part of this chapter, certain of the structural features will be analysed and described statistically, in order to provide a more quantitative assessment of their nature and significance.

4.2 Organisational Features

4.2:1 Breccia layers

The breccia exhibits a well-developed layered structure, defined by one of three types of boundary. The first is a very thin and black line, Plate 4.3. Under the microscope this line is shown to consist of quartz and globular clusters of opaque material, possibly the remains of organic matter. The opaque matter shows no evidence of deformation and the quartz is not ultramylonitic, although there is a grain size reduction when compared with the quartz in the adjacent layer. The second type of layer boundary is much thicker and white. It usually consists of a very sharp junction with the underlying layer from which surface a layer of euhedral, coarse quartz crystals grow, perpendicular to that boundary, Plate 4.4. There is a variation on this type, in which there is no well defined surface, Plate 4.5. The final layer boundary is one which is not a sharp surface but is a visual boundary defined by a sudden change in clast density or size, Plate 4.6. From the limited 3D exposure available it is seen that these layers defined by these boundaries have a strike apparently coincident with that of the main breccia band, (045°/055°), but a lower dip, only 20°/25° ENE, compared to 35°-50° for the main breccia band. Layer thicknesses are on the scale of tens of centimetres. There are also topographic



Plate 4.1 The main Rusey Breccia in outcrop



Plate 4.2a The matrix textures in the breccia



Plate 4.2b The matrix textures in the breccia

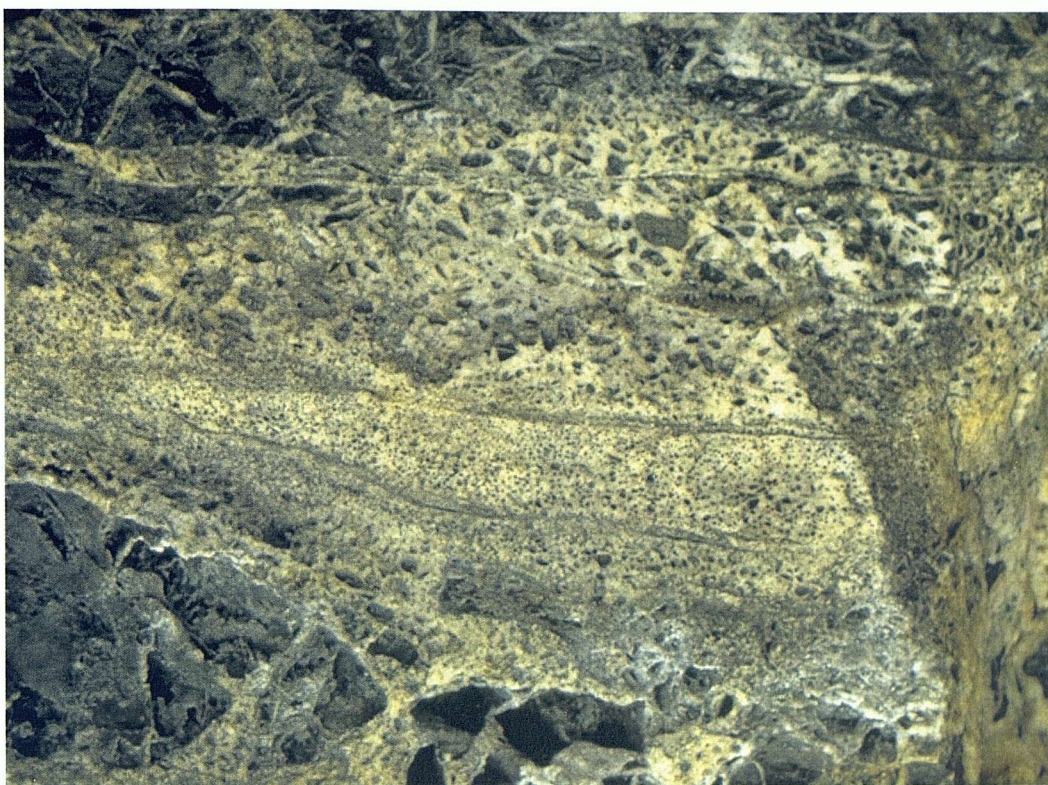


Plate 4.3 Black layer boundaries within the organised Rusey Breccia

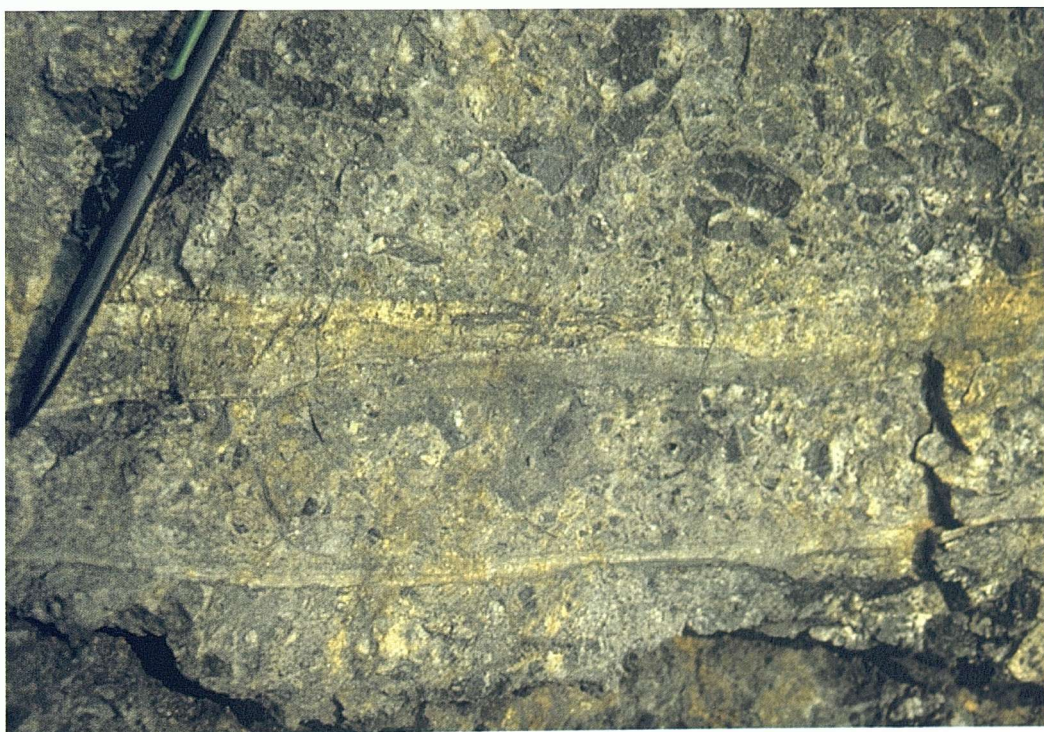


Plate 4.4 White layer boundaries at the edge of the organised Rusey Breccia



Plate 4.5 White layer boundaries within the organised Rusey Breccia

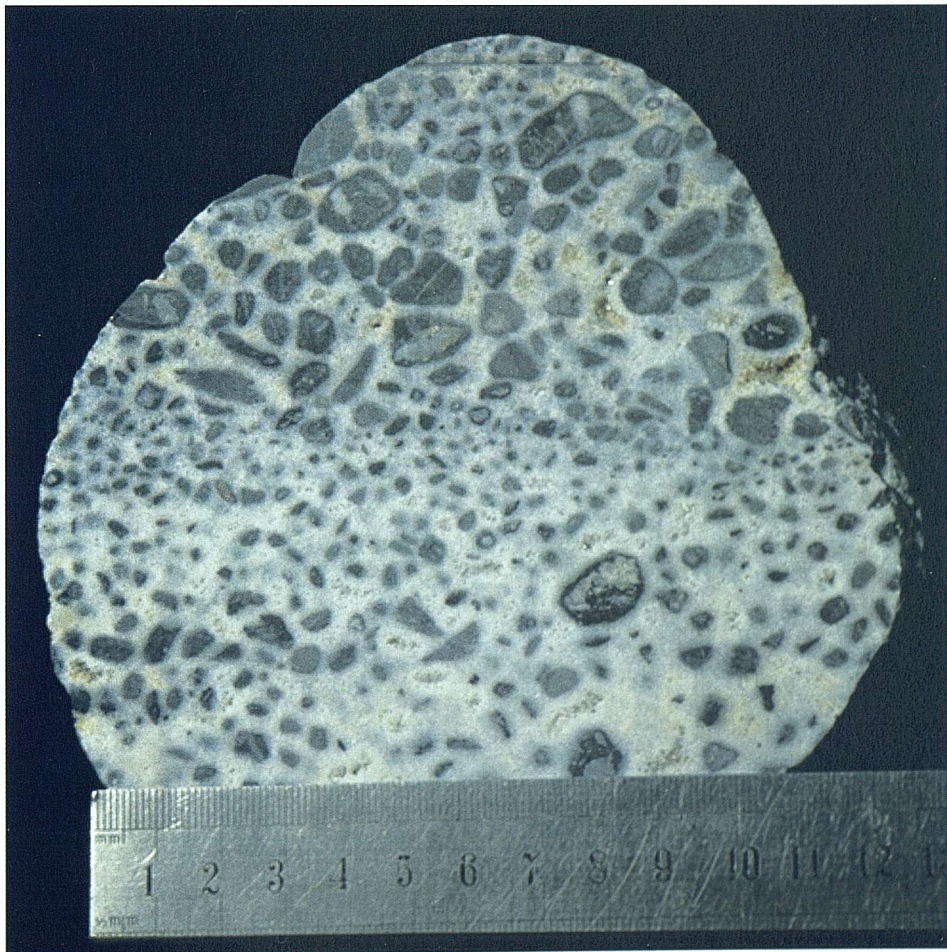


Plate 4.6 Layer development without explicit layer boundaries

irregularities in the base of these layers, often containing clasts, Plate 4.7, although they may have a layered fill consisting of mid to dark grey quartz. Under the microscope such dark quartz often derives its color from the presence of opaque material. Channels are relatively rare and usually contain the typical breccia materials, i.e. shale clasts with quartz matrix, taken from the edge of the breccia body. However there are extremely rare occurrences of 'negative' channels, Plate 4.8, where the channel consists of white quartz clasts, with a limited number of shale clasts, supported and separated by amorphous shale material. This is distinct from the material in the lower part of Plate 4.7, which is part of the country rock. Although this shows quartz fragments in a shale matrix, the shale material is not part of the layered unit and there is no ordering of those fragments in the manner seen in Plate 4.8, where the quartz clasts show a crude fining upwards in size. The shale matrix channel features pictured in Plate 4.8 are found within the layered main breccia band and show the same reordered, matrix supported textures as the quartz matrix layers.

4.2:2 Clast grading

Much of the central section of the breccia shows a systematic variation in clast size. This consists of a fining towards a layer boundary, Plate 4.9a, or a bi-directional grading, Plate 4.9b, fining out towards the edges of a single breccia layer. Whilst bi-directional grading is not common it occurs frequently enough to be regarded as a normal feature of this breccia. Its importance may be masked by weathered outcrop or poor exposure making it seem less common than it is. The grading is usually smooth but can occasionally show discontinuous variation in clast size. It can also define lens shaped bodies within the matrix, which have no definite boundaries within the matrix and are recognised only by a clast size grading sequence, Plate 4.6 and Fig. 4.1.

4.2:3 Oriented clasts

In addition to the ordering described above, there are examples of clasts being preferentially oriented parallel to the edges of larger clasts, Plate 4.10a. These smaller clasts show no sign of having been derived from the larger clast by brecciation. Elongate clasts show a statistically significant set of preferred orientations in the majority of samples analysed and this preferred orientation is often visually apparent, Plate 4.10b. This alignment contrasts with examples of clasts whose orientation appears to have been little modified and which can be easily restored to a pre-brecciation position as part of a larger clast or part of the wall rock, see Plate 4.10a.

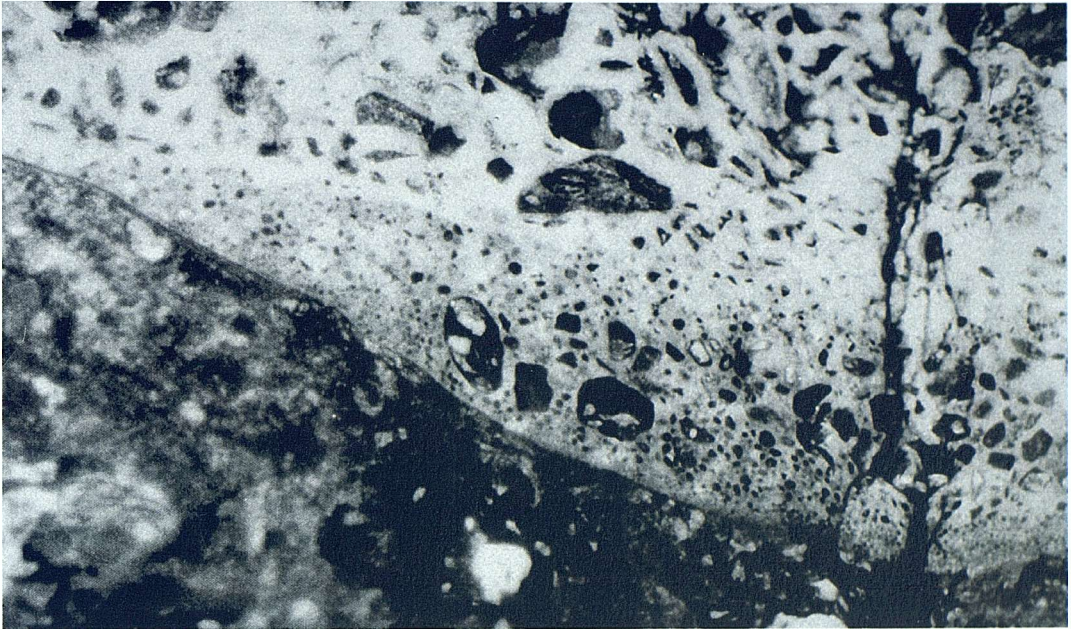


Plate 4.7 Channel development in the Rusey Breccia

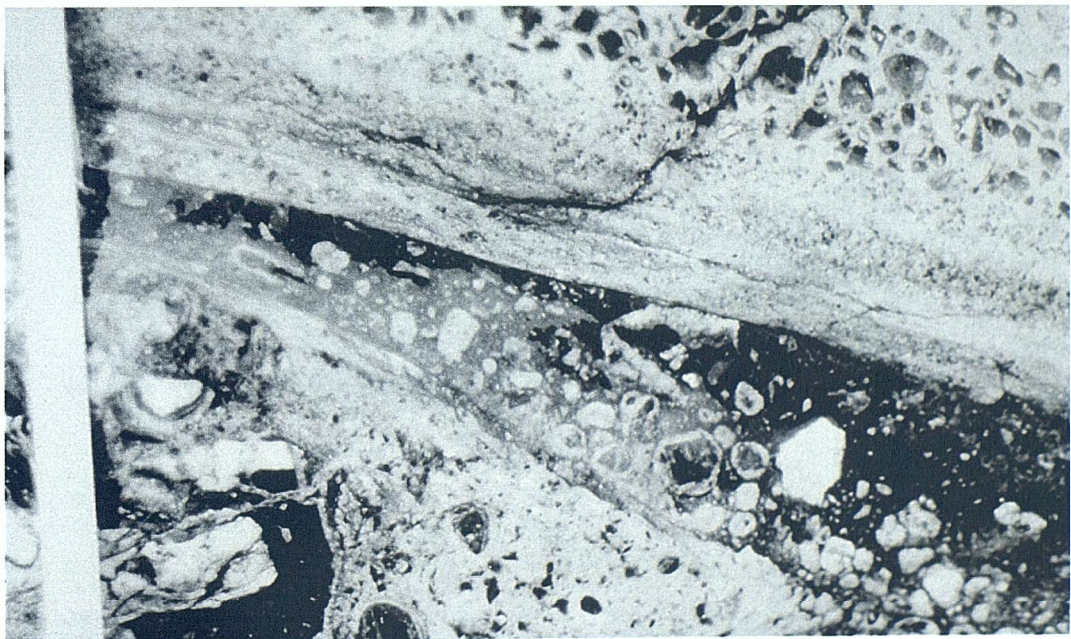


Plate 4.8 'Negative' channel development in the Rusey Breccia

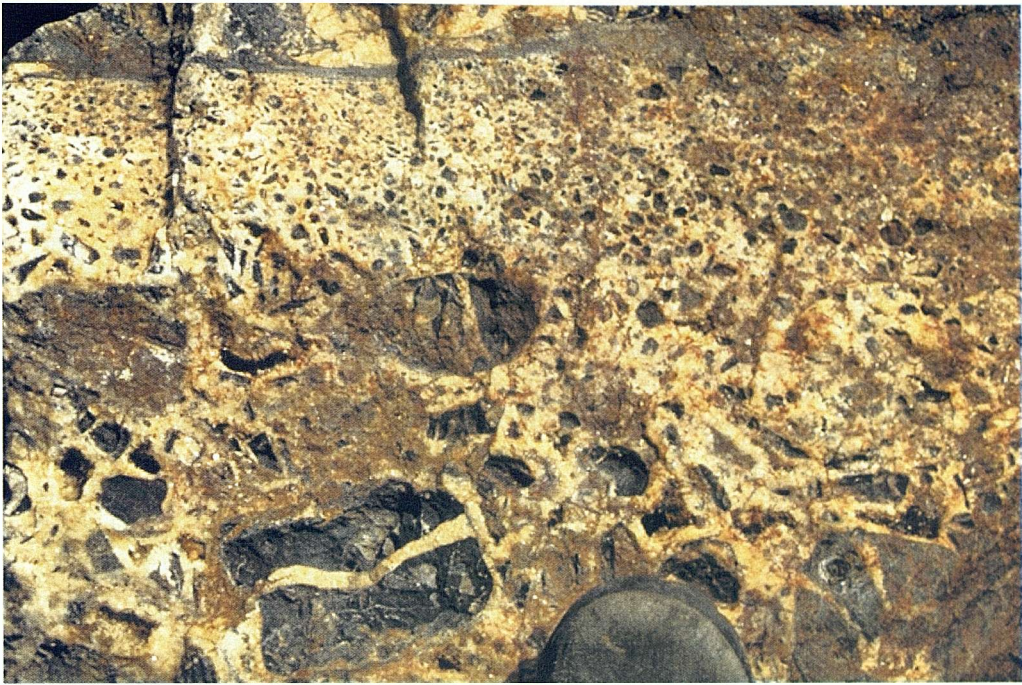


Plate 4.9a Half of a breccia layer showing clast size fining towards the layer boundary



Plate 4.9b The central layer exhibits clast size grading that fines outwards towards both layer boundaries

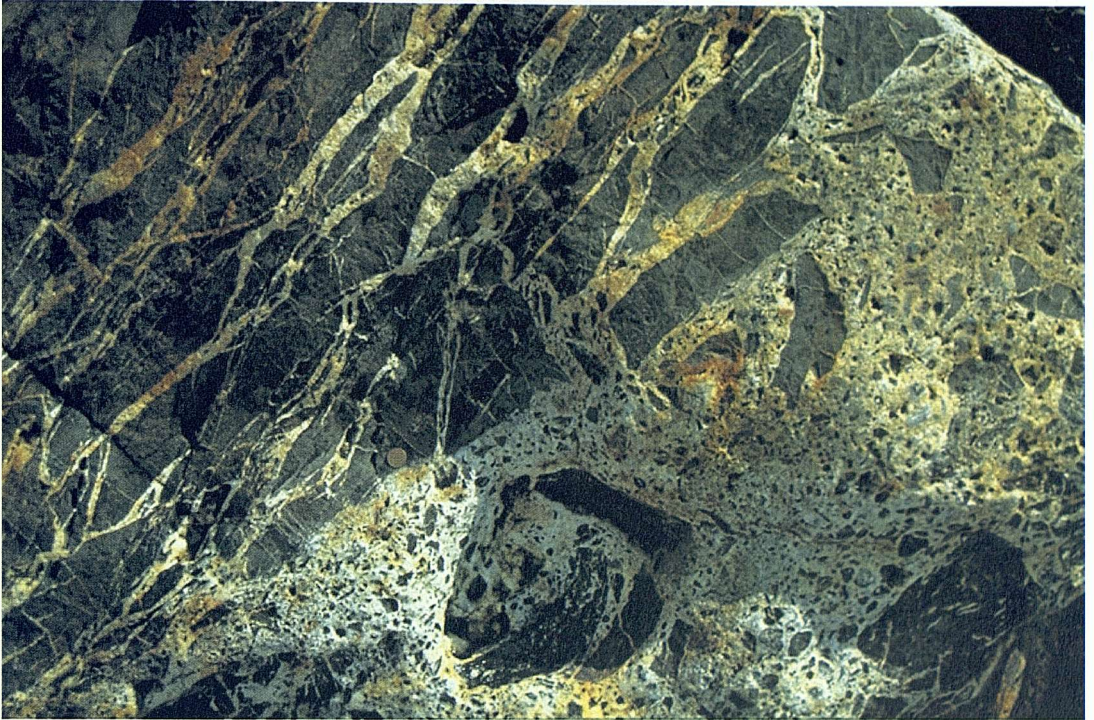


Plate 4.10a Transition from hydrofractured country rock into Rusey Breccia, (coin for scale, centre plate)

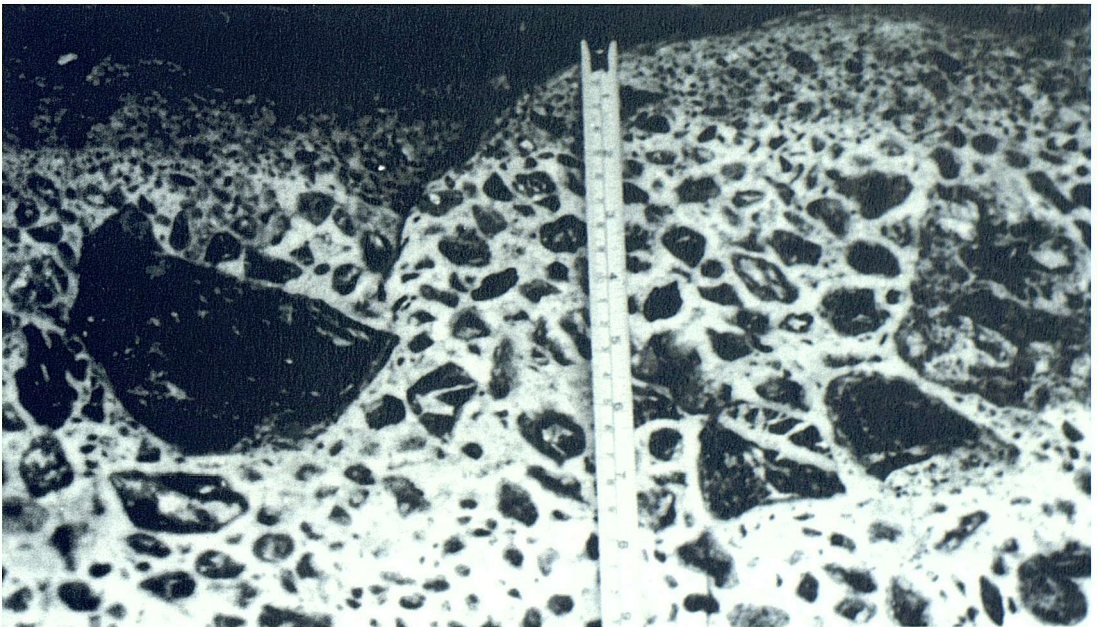


Plate 4.10b A sample where the preferred orientation of clasts is visually apparent

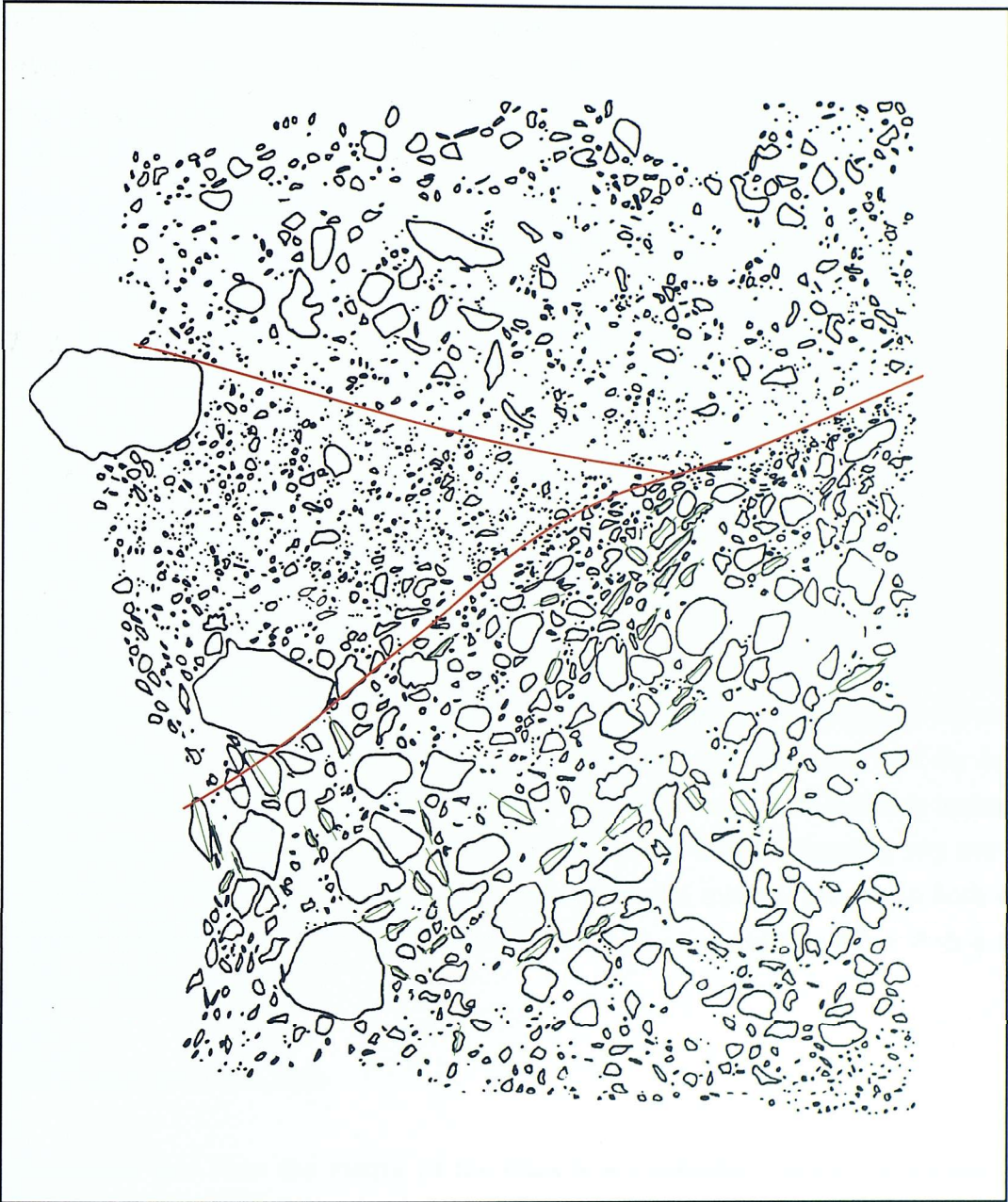


Figure 4.1 Line drawing of sample R15, showing the possible presence of multiple layers. The possible layer boundaries have been superimposed in red. The long axes of a number of clasts have been superimposed in green, showing a definite bimodal distribution of clast referred orientation. This is analysed more rigorously in section 4.4.

4.3 Morphological Features

4.3:1 Clast shape and structure

The majority of clasts show a rounded to sub-rounded profile and vary between equant and elongate. The re-brecciated breccia clasts fall into two categories; those which show earlier fracture events, preserving pre-existing veins, Plate 4.11a and those which preserve a previous generation of fully developed brecciation and mineralisation textures, Plate 4.11b. The former show the rounded profile typical of Rusey Breccia clasts, but the latter have a characteristic extremely rounded to globular profile. The clasts which show the same textures as the main band are rare, but those that show a lesser degree of deformation, consisting of fracturing and mineralisation, are common. The latter are composed largely of shale with quartz clasts or veins. They are often shot through with large veins, showing entrained clasts, which are sometimes wide enough to qualify as 'veins' of breccia. However, they are more often veined shale. These clasts are here regarded as country rock which had been veined during a previous deformation event, rather than true re-brecciated breccia clasts which have the same textural features as the main breccia band, as a whole, i.e. they contain matrix supported, rounded shale clasts, with matrix textures typical of this breccia, (see section 4.3:2). The re-brecciated breccia clasts often show the reordered textures described above and contain sections of pre-existing layers. A particularly unusual feature of the breccia, i.e. the rare occurrence of protruding clasts, is exhibited both by re-brecciated breccia clasts and by certain layers and is shown in Plate 4.11b and c. It consists of a large clast which is apparently an integral part of two distinct breccia units. The clast, without showing any previous mineralisation episodes, has acted as a nucleus for radial mineral growth in both units. The layer boundary between the units shows no sign of disruption other than a sharp truncation by the clast.

4.3:2 Crystal form and state

The crystals that form the matrix of the breccia are euhedral, though of varying size. They grow outwards from a clast or layer surface, perpendicular to that surface. This results in radial growth of quartz crystals around clasts to form clast/corona units, see Plate 4.2a and b, with the size of individual crystals varying with the size of the clast. Growth in vugs, veins and at layer boundaries is normal to the bounding surface regardless of whether the boundary is a clast boundary, (vugs), wall rock, (vugs/veins) or quartz breccia, (internal surfaces). Where the veins are only millimetres thick, crystal growth may not be euhedral and individual crystals are not visible with a hand lens. Under the microscope, undulose extinction is occasionally seen in large crystals and there is a distinct reduction in grain size towards well defined boundaries and in the

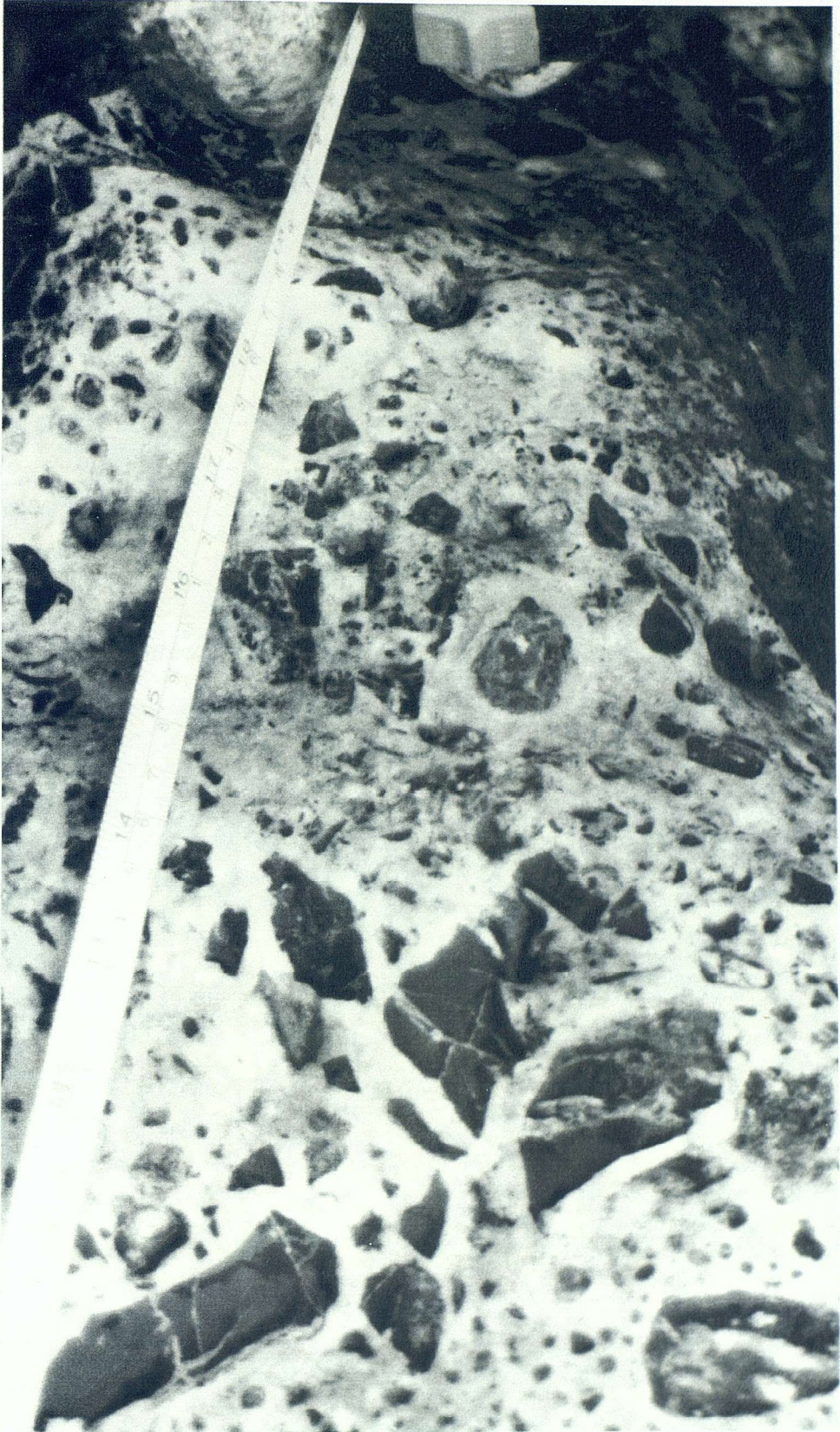


Plate 4.11a A breccia clast bearing evidence of pre-brecciation veining events



Plate 4.11b Rebrecciated breccia clast (compass-clino cord for scale)

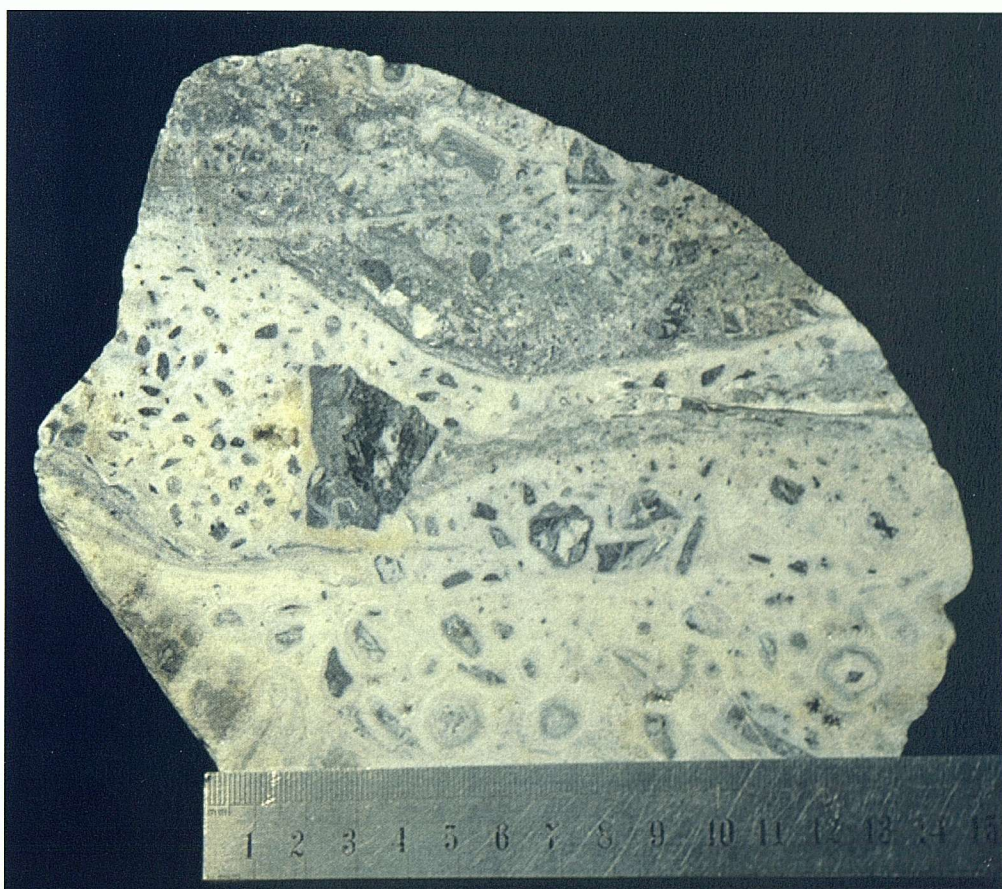


Plate 4.11c Protrusion of large clast from upper layer into lower layer

initial mineralisation around the clasts, Fig. 4.2. Where clast/corona units are in contact, there is some evidence of intergrowth, particularly when viewed under the microscope, but there is no evidence of pressure solution or other forms of recrystallisation. There is minimal interstitial quartz cementing the clast/corona units, see Plate 4.2a and b.

4.3:3 Voids

The amount of void space within the breccia is unusually high, see Plate 4.2a and b, due to the lack of interstitial matrix. The porosity of the rock is therefore high, though the permeability is untested. Even where the rock is 70-80% quartz, that quartz forms the coronas of quartz clasts, rather than infilling pore spaces between clasts. Vug development is more common at the north end of the outcrop. The vugs are rarely filled, Plate 4.12. Large vugs can be up to a metre across with well formed quartz crystals growing in from the walls.

4.3:4 Interstitial Matrix

The main breccia band shows little trace of rock flour or fine grained fault gouge, with the possible exception of the fine dark material which separates some of the layers, Plate 4.13a. This material sometimes appears between concentric zones of radial growth around the clasts, Plate 4.13b and, very rarely, forms a black residue in which the clast/corona units sit at the margin of some layers, Plate 4.13c. The majority of the breccia is coarse to extremely coarse clasts, (up to cobble sized, 64-256 mm), in a medium or coarsely crystalline quartz matrix. Type one layer boundaries, which contain dark material, appear, under the microscope, to be zones composed of fine grained quartz and a high concentration of opaque material. No presence of interstitial rock flour can be seen either in hand specimen, see Plate 4.2a and b or under the microscope. The entire matrix consists of crystalline quartz or opaque material. In areas of the breccia where the clast size is at the coarsest end of the range, there is very little interstitial quartz matrix.

4.4 Spatial Arrangement of the Breccia

The breccia demonstrates a spatial organisation in the development of the features described above, Fig. 4.3. To the south there is intense veining with randomly oriented quartz veins. The veins are thin and generally have an amorphous or cryptocrystalline mineral texture. There are amorphous patches of fine grained crystalline quartz. Further north, this passes from rock with a high concentration of veins into coarsely brecciated rock, possessing no remarkable degree of organisation. Towards the centre of the band, the degree of organisation increases markedly, with well defined layering and graded

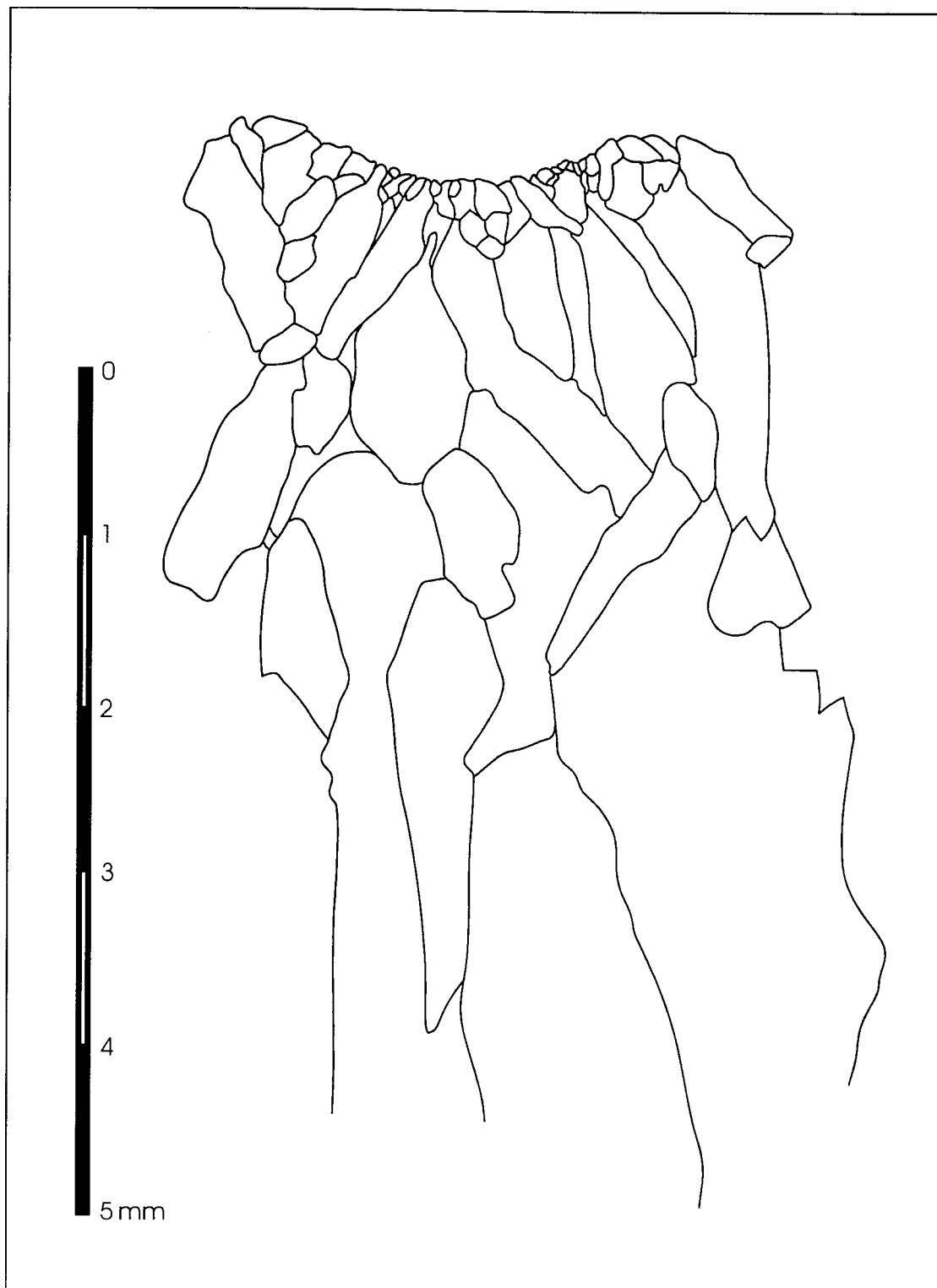


Figure 4.2 Sketch from thin section showing the change in grain size with distance from the clast. The mineralisation immediately adjacent to the clast is a layer of very fine grained quartz. The next layer of much larger and elongate grains grow from this layer, oriented perpendicular to the clast surface.



Plate 4.12 Large veins with euhedral quartz crystallisation and central space

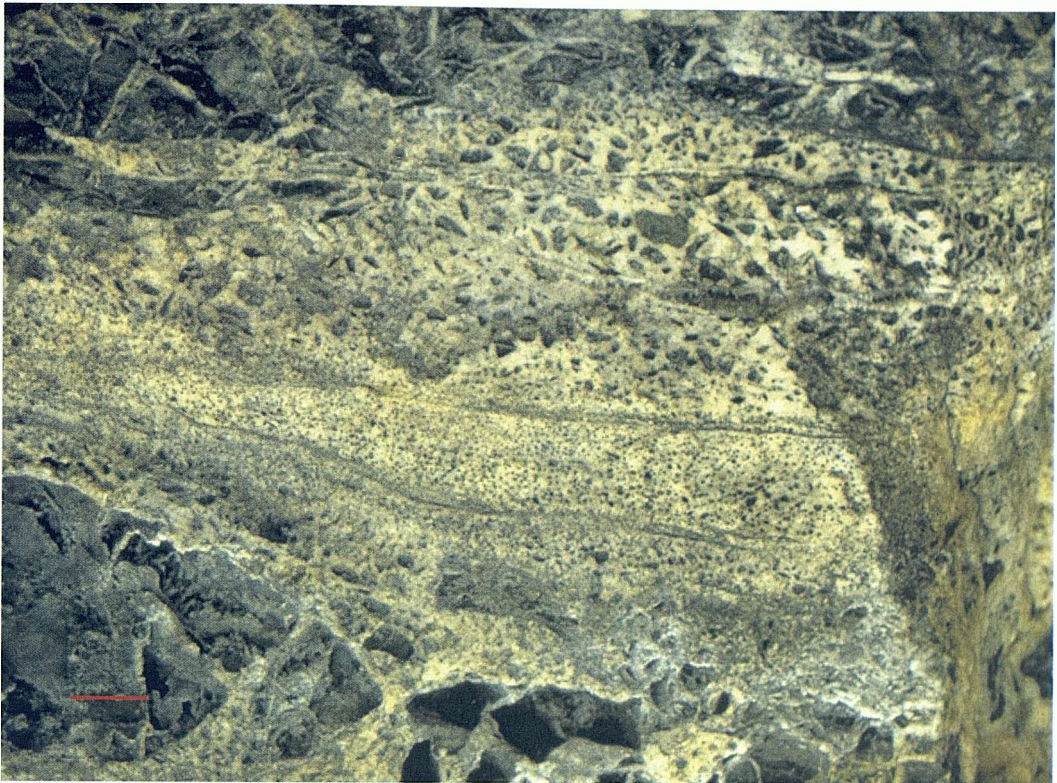


Plate 4.13a Breccia layers defined by layers of dark material at The boundaries (scale bar = 10cm)

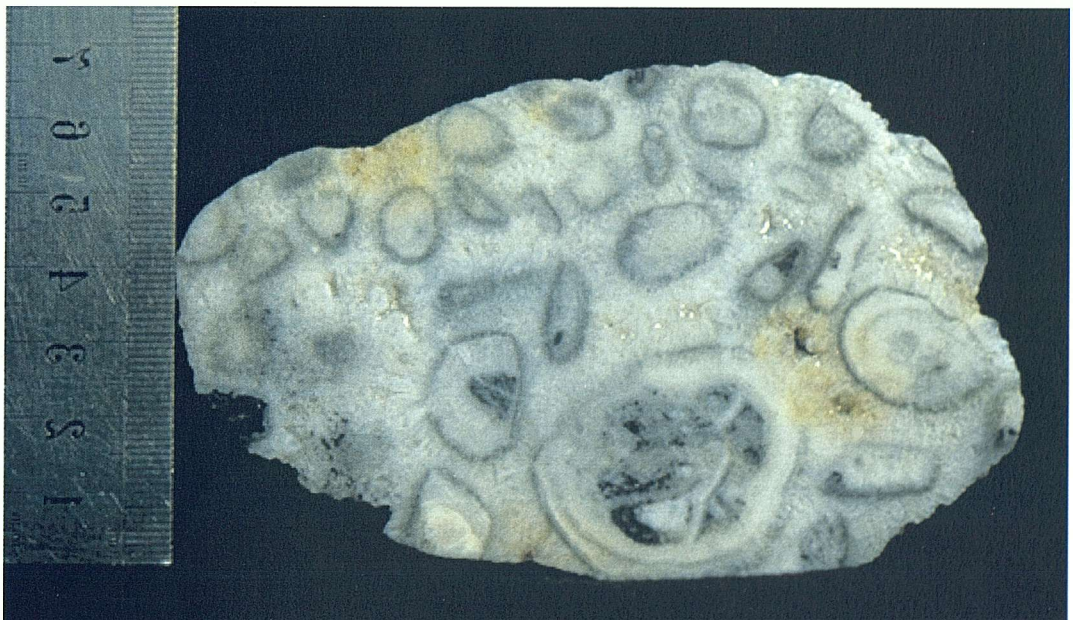


Plate 4.13b Dark layers within the corona mineralisation around clasts



Plate 4.13c Dark material at the edge of a layer, forming the matrix which supports the clast/corona units

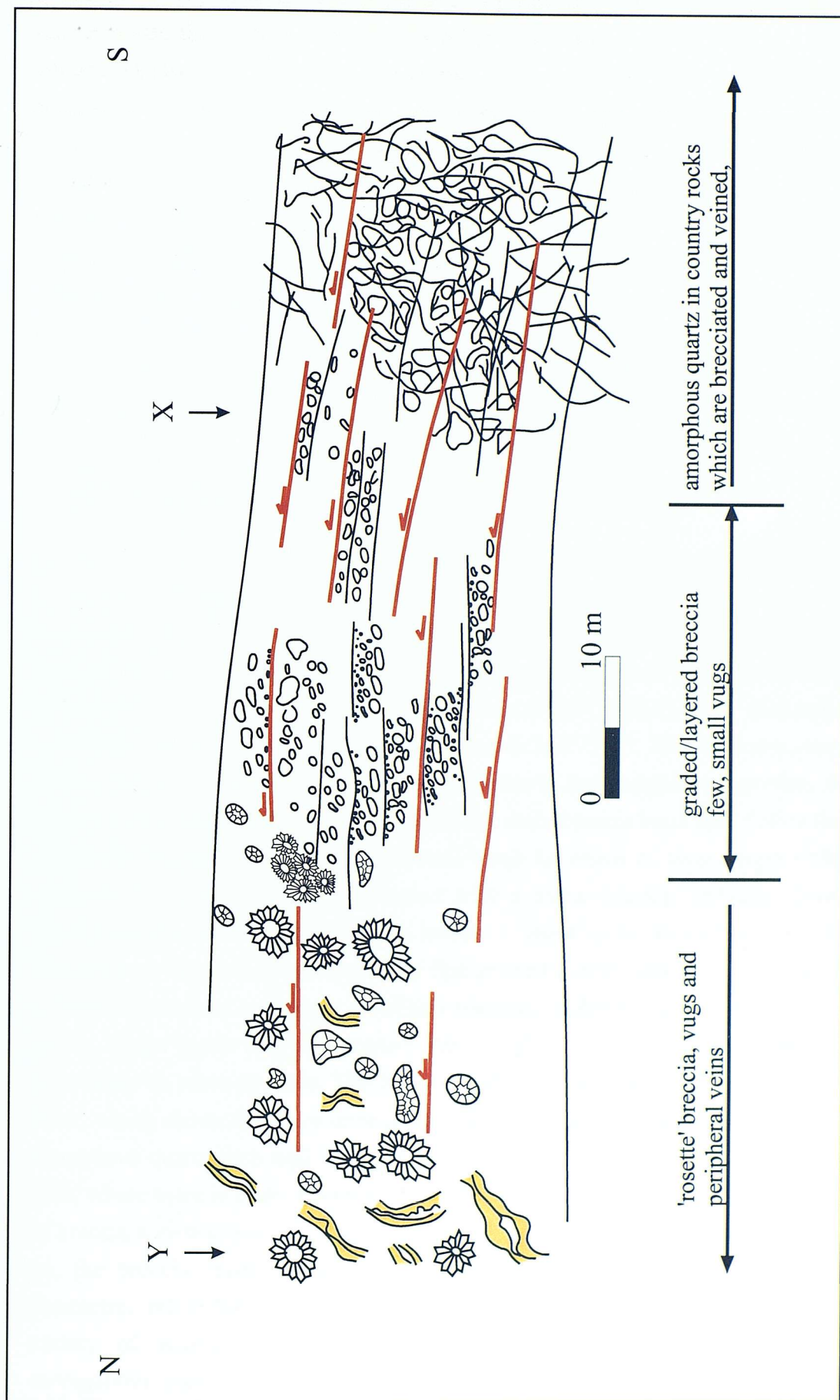


Figure 4.3 Schematic synoptic diagram of the Rusey Main Breccia Band. The diagram shows those brecciation features that are the concern of this thesis and their spatial variation through the breccia body

beds appearing and becoming persistent, Fig. 4.3. Clast size in this area is on the scale of millimetres to several centimetres and vug size is correspondingly small. Moving further north, the breccia returns to a disorganised state, but here there is a large void volume. Clast size increases, becoming several centimetres to several tens of centimetres and vugs can be nearly a metre across. The veining at this end of the breccia band is also randomly oriented, but much more dilatant, with extremely well developed euhedral, coarsely crystalline quartz crystals, up to two centimetres in length. The euhedral, radiating quartz growth textures are observed from the point at which the breccia begins to show ordered textures, (X, Fig. 4.3), as far north as breccia clasts are present as individual entities, (Y, Fig. 4.3). Where the breccia passes out into veined and brecciated country rock to the north, the quartz growth textures are still euhedral and perpendicular to the growth surface, though without separated clasts, radial textures do not form.

4.5 Late Deformation

There is evidence of later deformation of the main breccia band in the form of a number of surfaces cutting across the band. These surfaces are broadly coincident with the boundaries of the breccia body itself, (strike and dip approximately 163° , 45° ENE) generally striking between 146° - 160° and dipping at angles of between 40° - 65° to the ENE, with very well developed slickensides, (Plate 2.8/2.9). The slickensides are extremely consistent, plunging at 6° - 18° , towards 330° - 350° . This indicates almost pure strike-slip movement, provided these surfaces are in their original orientation. Surfaces similar to the slickensided surfaces within the main breccia band also define the upper and lower boundaries of the main breccia band for much of their length. Within the main breccia band these surfaces occur with a metre spacing vertically through the breccia band and are comprised of a series of zones up to several centimetres thick. These zones, Fig. 4.4, are comprised of fine grained quartz material, consisting of bands of both white quartz and a dark grey/black material, probably also largely, if not entirely, quartz. These bands may also contain clasts, both of shale and quartz, ranging in size from 5mm to sub-mm scale. The base of each zone is commonly a thin vein, 1-2mm thick, which shows the same form as the thicker veins in the shale units, i.e. a layer of amorphous quartz with well formed crystals growing out of it towards the centre of the vein, where there is often a central void, see Fig. 4.4. The zones may cut areas or layers of breccia with textures similar to the rest of the band but an order of magnitude smaller, i.e. the breccia clasts are millimetre to centimetre scale rather than centimetre to decimetre. When they occur within the breccia, they are often gently undulose, on a variety of scales, from centimetres to metres. The centimetre/decimetre scale corrugations parallel the slickenside directions precisely and are very well defined, (Plate 2.9).

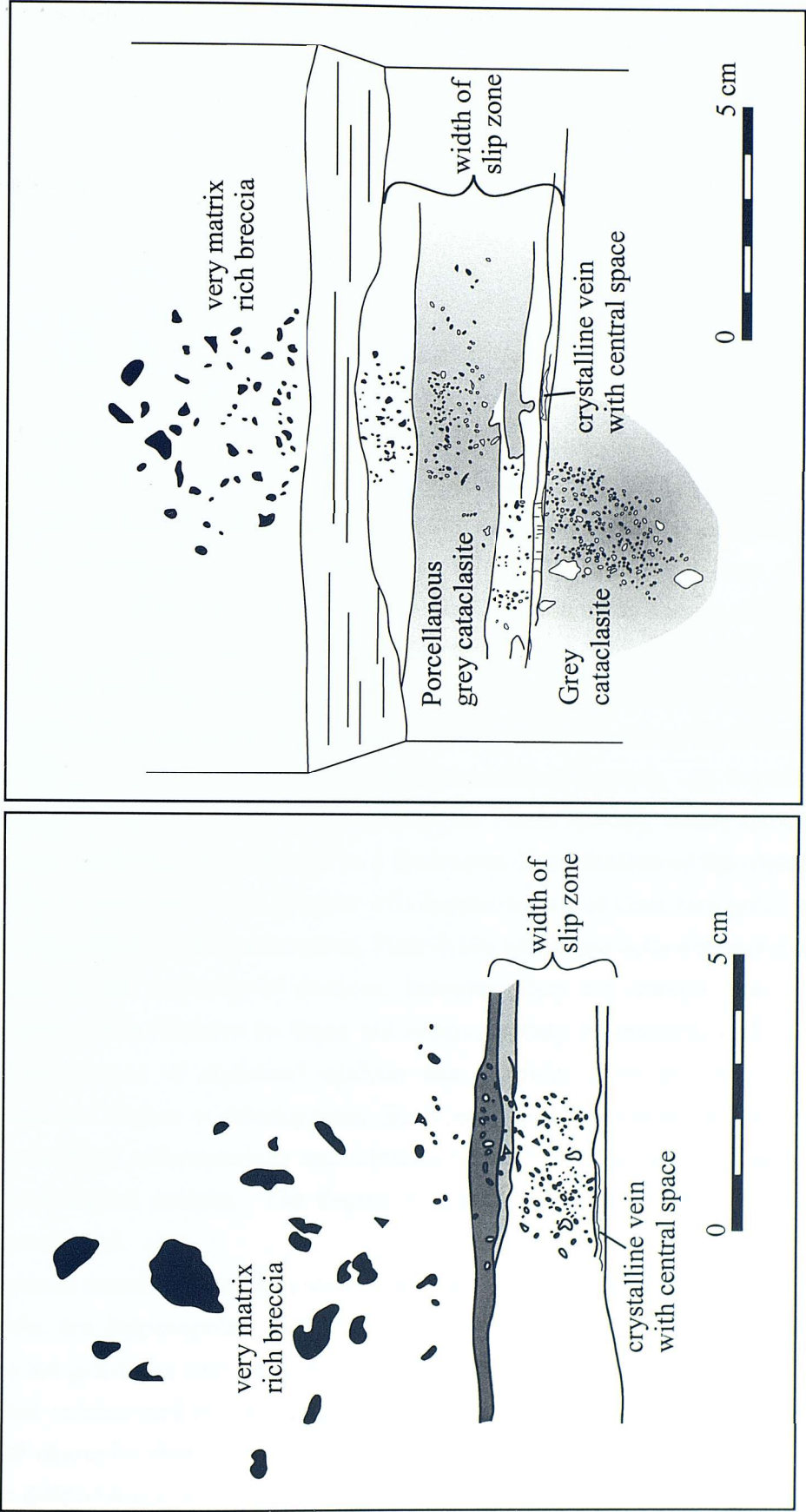


Figure 4.4 Internal structure of late stage slip planes within the Rusey Breccia

There are also late veins and small extensional faults which cut the main breccia band. The faults displace the breccia band, with displacement dying out rapidly as the fault penetrates into the surrounding shale. The veins often show some displacement, but crystal growth indicates dominantly tensile failure, rather than shear failure, i.e. the crystal growth is at right angles to the wall, not inclined across the vein.

4.6 Statistical Analysis of the Rusey Breccia

4.6:1 Introduction

Fault breccias are a complex class of rocks, usually consisting of transported, non-cohesive cataclastic rocks. They are formed by brittle fracturing of rock followed by granulation and comminution, accompanied by rotation. Most are poorly sorted with gouge fabric similar to that of glacial till or mudflow. Dilated faults often have a fill of angular to sub-rounded grains which may be fragment or matrix supported, but which rarely show sorting. If there is organisation it usually takes the form of larger more angular, more proximal clasts along the wall and rounded, matrix rich breccia in the centre of the zone, (Laznicka, 1988). However in the course of the field studies and examination of photographs, presented in the first half of this chapter, it was observed that the Rusey Breccia, whilst genetically linked to activity along the Rusey Thrust, (see Chapter 2), appeared to exhibit certain organisational features, e.g. layering, clast size grading and preferred orientation of clasts, (see Plates 4.2-10), which are not commonly associated with breccia formation in a fault zone. Examination of the plates and figures from the earlier part of this chapter will demonstrate that clast size grading, Plate 4.9a and b and preferred clast orientation, Plate 4.10a and b, are both well developed in many sections. In the majority of sections, however, they are present, but less markedly developed. Such features as these are relatively easy to quantify and can be studied using techniques of statistical analysis that provides a reliable assessment of their presence and degree of development. Such techniques have been in use for many years in the fields of sedimentology and fabric analysis. Their use here is drawn largely from sedimentological sources. The degree of post-ordering strain exhibited by the breccia seems minimal.

As a result standard structural strain analysis techniques, e.g. Fry or nearest neighbour analysis, are inappropriate as these are designed to be applied to granular materials where the grains are internally strained. Whilst there is evidence of clast abrasion in the samples represented in this chapter there is no evidence for internal clast strain and several examples that contradict the idea of the clasts being internally strained. Internal brittle deformation clearly pre-dates clast formation and although the clast material has undergone recrystallisation, Thin sections show no evidence of microfabric or ductile deformation within the clast. Standard structural particle analysis techniques are also

regarded by the present author as being inappropriate because they depend on the ability to unambiguously define the clast centre, Ramsey and Huber, 1967, "The technique can be used where we have an aggregate of objects which was initially fairly uniform in the sense that the distances between particle to particle centres were initially statistically constant and where the particle centres are relatively easy to define."

As can be seen from Plate 10a, there is no reliable evidence that either of these assumptions can be applied to the Rusey Breccia. The clasts are highly irregular and in some cases can be seen in close proximity to their original, pre-fracture location. In this situation the clasts are highly angular. It is therefore likely that the more rounded shape exhibited by the majority of the clasts was developed after the clast was formed. Since the angularities in the clast profile are removed rather than warped this transformation represents an erosion of the clast rather than a deformation. This would therefore invalidate techniques with the pre-requisites described above.

In addition the Fry and nearest neighbour techniques produce meaningful results because all the particles in a sample are strained in the same way and to the same degree. This is not the case for the Rusey Breccia, where the clasts have been mobilised, see Plate 4.1a/b. Each clast has an individual translation path, rather than having undergone the same strain path as its neighbour.

The minimum and maximum clast axes were easier to define with a higher degree of accuracy and accordingly they were initially used as a measure of clast shape. Although long/short axis analysis techniques are available, there remained a significant degree of uncertainty and human bias in selecting the axis orientation. There was also the problem that low axial ratio clasts would be reduced to only two possible orientation data and therefore represented an unacceptable (because avoidable) bias in the data. These reasons caused the author to prefer an entirely automated analysis procedure that utilized the clast boundary data. Whilst this creates some 'noise' in the analysis in the form of the presence of some data for all orientations it also ensures that all the clast orientation data available is used. The resulting plots demonstrate the minimum degree of alignment present in the sample and the present author regards this as preferable to portraying the maximum possible degree of alignment.

The application of many sedimentological techniques for particle analysis were rendered impossible due to the facts that;

- the samples are well mineralised and cannot be disaggregated for particle size analysis and;
- microscopic studies of the clasts could not be carried out with any reliability due to the severe difficulty and high unreliability involved in identifying the clast edge. This arose because the initial matrix mineralisation was of a very similar crystal size to the recrystallised meta-sediment of which the clast was comprised and the clast mineralogy was not substantially dissimilar to the matrix mineralogy and the clast boundary was not visible under the microscope.

For these reasons techniques were adopted which were a practical adaptation of the standard sedimentological techniques described by many workers, e.g. Tucker 1988. These techniques were adapted to be applied to photographic centimeter scale samples, as these gave greater data reliability than studies conducted at the microscopic scale. Where possible the analysis process was automated to increase speed and decrease human bias in selecting data.

Three dimensional studies of the breccia would have required oriented samples which were impossible to obtain from within the Main Breccia Band. The breccia is well mineralized and smoothly eroded. Oriented sample retrieval would have required the use of rock cutting equipment, which was not feasible in the field area.

As a result of the difficulties encountered in applying standard analysis techniques to the Rusey Braccia, techniques related to a comparison between a strained and unstrained state of solid rock, (e.g. Sanderson and Phillips, 1987) were combined with techniques developed to analyse the degree of organisation of loose granular materials, (Schwan, 1989, Rees, 1968). The philosophy of Sanderson and Phillips in applying length weighting to curved linear data was applied as a correction to the clast orientation data, although the method of presentation of the data is a standard rose diagram format, such as is generally used in sedimentology to present grain alignment data from coarse grained sediments. Clast area and clast size analysis techniques were modified from standard grain size analysis techniques, (Tucker, 1988).

Three of the ordered features of the breccia were subjected to analysis. These are;

- the presence of preferred clast orientation,
- the variation in clast density across a single layer and
- the presence of clast size grading.

These were chosen because statistical proof of their presence would be significant in determining if the system had been under the influence of a free flowing fluid, i.e. a fluid which flowed and transported the clasts as a sediment load rather than a fluid which permeated and flowed through a fixed bed of clastic material. The presence of structures similar in appearance to sedimentary channels, Plates 4.7 and 4.8, strongly suggested that free flowing fluids had been involved in the reorganisation of the breccia and evidence was sought to validate or eliminate this idea.

In the following sections all of the in-house computer software, (Orien, Wield, Propscan) was written by Dr. Richard Jolly, (Imperial College), whose help and advice was invaluable and is gratefully acknowledged.

4.6:2 Preferred Orientations

4.6:2a Methodology

The analysis program used, Orien, represents preferred orientation trends by splitting a clast outline into straight line segments and measuring the orientation of each line. These orientations are then plotted on a rose diagram. Since all clasts were represented by closed polygons, the program assigned the straight line segments a length dependent weighting. This ensured that longer lines were assigned a greater value and therefore that the orientation of the long sides of elongate clasts became relatively more important than the short sides. Without this weighting, no preferred orientations would have been apparent as closed polygons would have straight line segments in all orientations. The alternative would have been to split the polygon into equal length line segments, but this would have caused the orientation of a given segment to be averaged if it was not straight, introducing inaccuracies into the resulting plot.

Samples of the breccia were collected and photographs taken of sample areas too large to be removed. In order to analyse the preferred orientations it was necessary to determine the inherent directionality of each clast. Three techniques for isolating and mapping these data were tried. One method involved the identification of the long and short axes and two methods involved discretising the clast outlines for input into a computer analysis program.

- The photograph of the breccia was covered with an acetate sheet and the long and short axis of each clast was drawn on, Fig. 4.5a. This was accomplished by choosing the longest straight line possible across the clast and assigning that as the long axis. The short axis was assigned as the line at right angles to the long axis. This was an invalid method for determining the directionality of the clast because it only isolated two elements of directionality for each clast. For clasts of high aspect ratio this did not present an unacceptable restriction of the data. However for clasts of low aspect ratio, it imposed an element of directionality that was not an inherent feature of the clast. In addition, it was also an extremely time consuming method of data extraction.
- The photograph of the sample was scanned as a high resolution bitmap and then traced using the Corel TraceTM package, Fig. 4.5b. This technique was subject to the inherent limitations of the scanning resolution of the computer and its inflexibility in picking a value to define clast/non-clast boundaries. A single tonal value could not be used for images taken under different lighting conditions and assigning all black areas as clast resulted in shadows being picked as clast material, causing the traced image to be highly inaccurate. In shadowed areas a very high proportion of the photograph would be traced as falling within a clast boundary and closely spaced

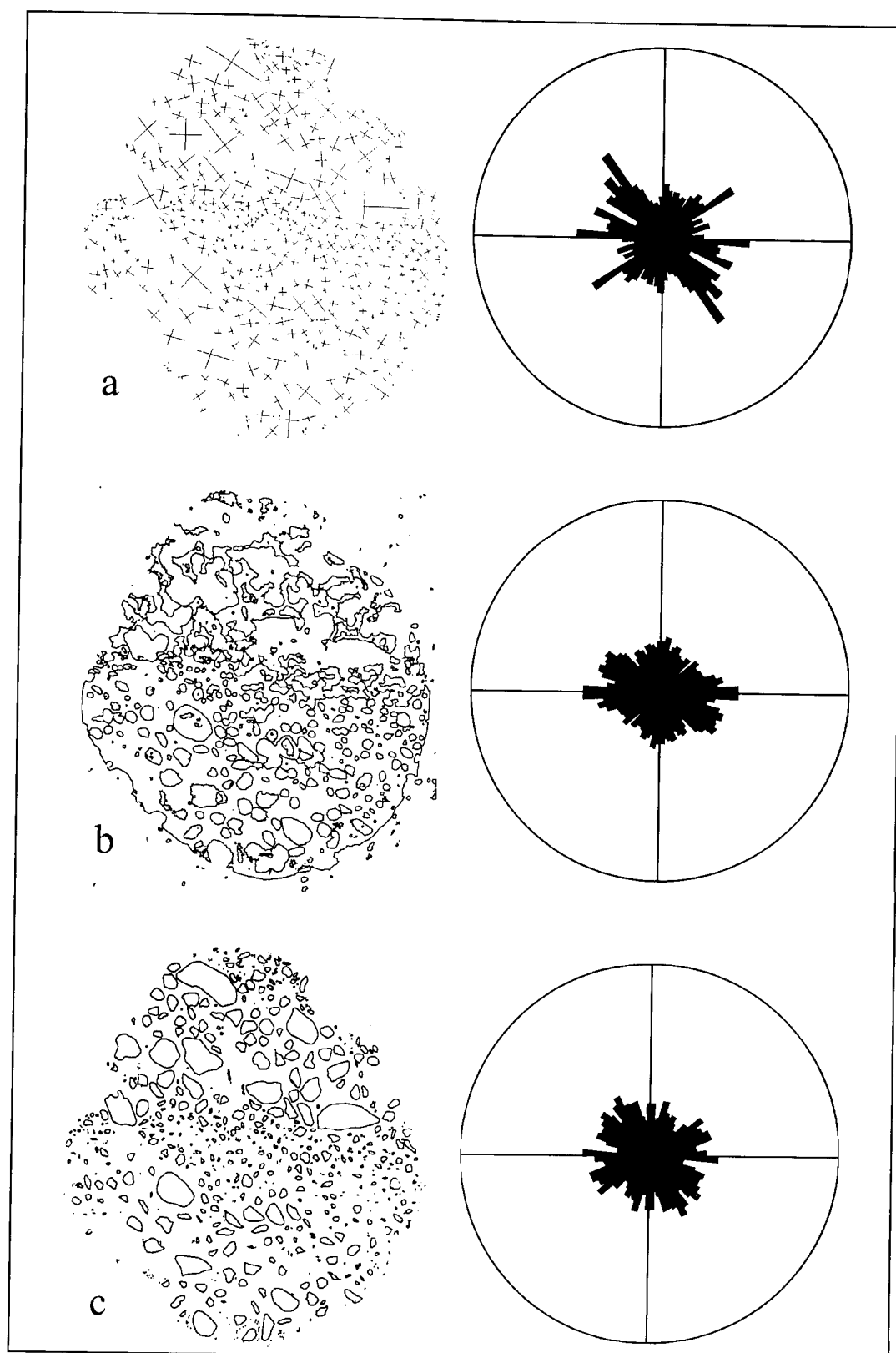


Figure 4.5 A comparison of rose diagrams produced by different methods of data input.
a: hand drawing of long and short axes of each clast from a photograph
b: drafting package trace of a bitmap scan of a photograph
c: hand line drawing from a photograph

clasts would merge into a single unit, see Fig. 4.5b. This created clast boundaries which did not exist and rendered the technique inaccurate. Low resolution scans caused pixilation of the image, introducing inaccuracies in the shapes of the traced outline.

- The third method involved placing an acetate sheet across the photograph and tracing, by hand, the outlines of each clast, Fig. 4.5c. Although somewhat time consuming it was significantly less so than drawing the clast axes, see Fig. 4.5a and imposed no unrealistic element of directionality on the sample. Such an element of directionality is imposed by drawing the long and short axes and isolating solely this information for the analysis as this restricts to just two the possible orientations which can be measured for that clast. By drawing the full clast outline all the parts of that outline contribute to the assessment of the clast orientation. Whilst this may not make a significant difference for the clasts of high aspect ratio, it does make a difference for the more equant clasts. In such a case the choice of two axis orientations out of all the possible range of orientations was entirely subjective and dependent on any subconscious bias on the part of the analyst. For example, if the long and short axes being drawn for a given sample appear to show a visually apparent preference for a particular orientation range, the analyst may be induced to subconsciously select a similar orientation for the axes of equant clasts. This would cause any alignment quantified by the Orien program to be unrealistically emphasised. If the full outline is used, the equant clasts would contribute straight lines segments oriented in all directions and would therefore show no preferred orientation. The equant clasts were not eliminated from the sample since the fewer the selection criteria imposed on the data, the less chance there was of causing the data to become skewed in favour of showing evidence of a possible preferred alignment. Therefore to include the full clast outline for every clast ensured that every possible piece of orientation data was supplied to the program.

The scans were converted from the encapsulated postscript (EPS) format output of Corel TraceTM to Hewlett Packard Graphical plotter file format (PLT) in Corel DrawTM. In the case of the long/short axes diagrams, these were placed directly into Corel DrawTM without using the tracing package and output as PLT files. These PLT files were then converted, using an in-house conversion program, Wield, to the PL3 format acceptable to the rose diagram plotting program, (Orien).

When inputting sample scans to Orien all scans were input with the layer boundary horizontal. This was done because many of the samples of well organised breccia were derived from beach samples and boulders and therefore their original orientation was unknown. Whilst this is a drawback, it does not render the analysis invalid as the layering within the in situ breccia is of a consistent orientation, i.e. subparallel to the margins of the breccia body, (Section 4:2.1). The grading packages are layer bounded

and it can be inferred that the processes causing reorganisation of that layer were restricted to that layer. Therefore the layer boundaries are the logical reference points from which to measure preferred orientations, since they were the boundaries of the active system. The clasts which are oblique to the layer boundary can be assigned a dip value with respect to that boundary but, as the in situ orientation of many of the samples is not known, they cannot be assigned a true dip or a dip direction. Therefore whilst the rose diagrams of the clast orientations can demonstrate that an alignment exists, this lack of precise orientation data for each sample renders it impossible to do any further analysis of flow directions within the breccia, such as is often carried out in sedimentary analyses.

Comparison of results from these three methods, see Fig. 4.5, shows that the input from the axis map gives a high definition plot, the input from the Corel traced photograph, see Fig. 4.5b, clearly shows little resolution of trends and the trace image shows that many of the clast outlines are obscured in Fig. 4.5b. The input from the line drawing of this sample demonstrates the presence of a weak alignment, though it is not the best developed example.

Given the significant element of subjective interpretation of the data when choosing the orientation of axes for clasts with low aspect ratios and the significant element of error clasts with an aspect ratio of 1 contributed to such plots, technique one was rejected, as was technique two, invalidated by the computers inability to accurately discretise the clast/matrix boundaries. The line drawing method was selected as being the only reliable method of data discretisation, selection and input. When transferring the data from acetate to computer it was again found that the resolution of the scanned image created errors, such that clasts smaller than a single pixel side length on the scanned image produced noticeable vertical and horizontal trends which were artifacts of the scan resolution rather than true data. This constituted the lower limit on the size of clast which could be analysed. The analysis program was modified to plot only elements of the data set over a certain length value, eliminating the smallest clasts from the data set. There was no upper limit on the size of clast which could be measured.

The effect of the largest clasts in producing a bias in the plots was investigated. The larger clasts were removed from the scans by painting them out in a bitmap editor, (Paint Shop ProTM), see Fig. 4.6. These bitmaps were then converted, by the method described above, for input into Orien. Comparison of the edited scans with the real data demonstrated that the larger clasts produced no bias in the preferred orientation. The examples presented in Fig. 4.6a-d include both those least and those most affected by the editing process and it can be seen that the difference is negligible.

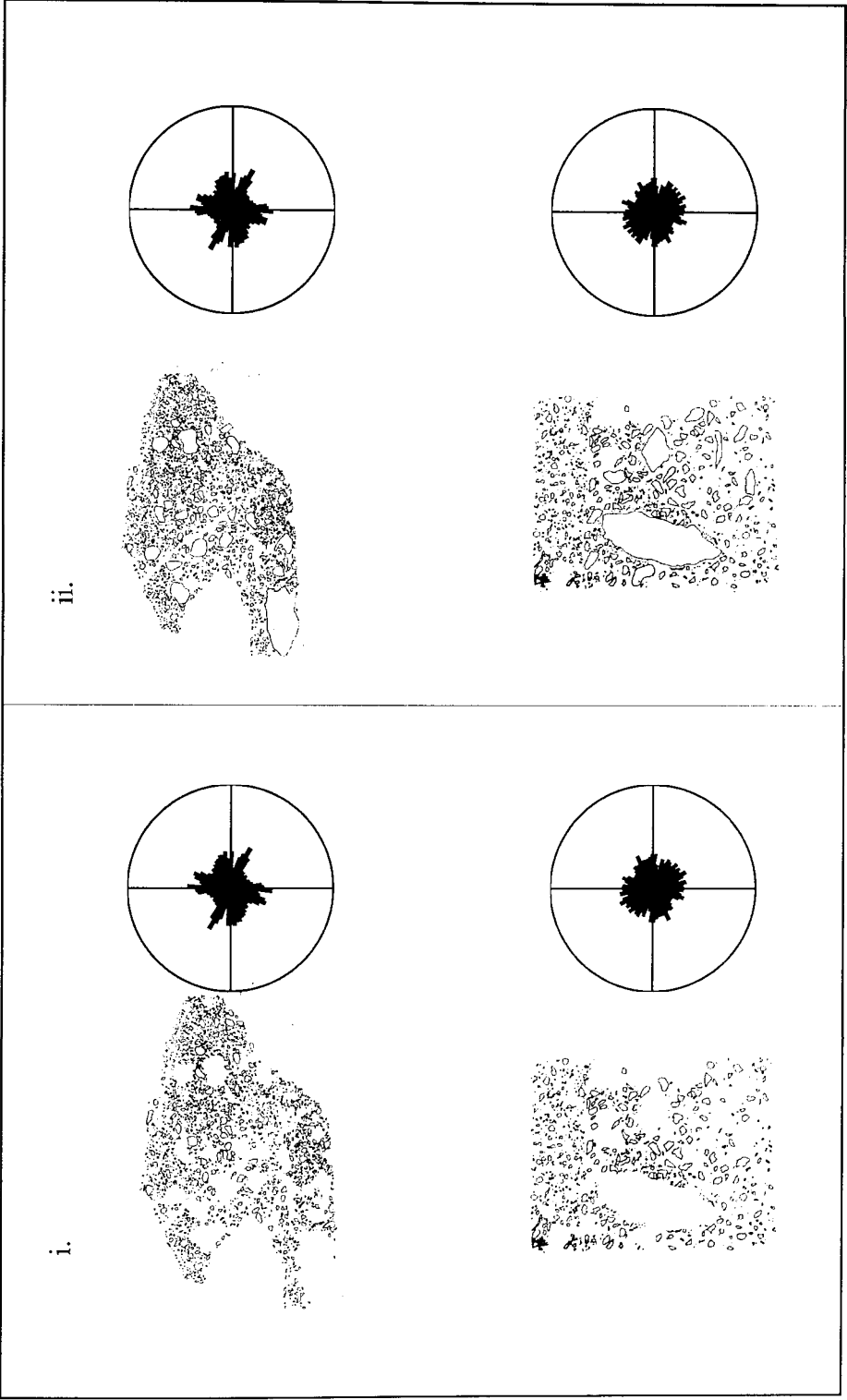


Figure 4.6a A comparison of Orien plots with
i: with the largest clast element removed
ii: complete data sets
This shows that the presence of the larger clasts within the data set in no way skews the distribution of the plots. The plots are true representations of the overall preferred orientations within the population.

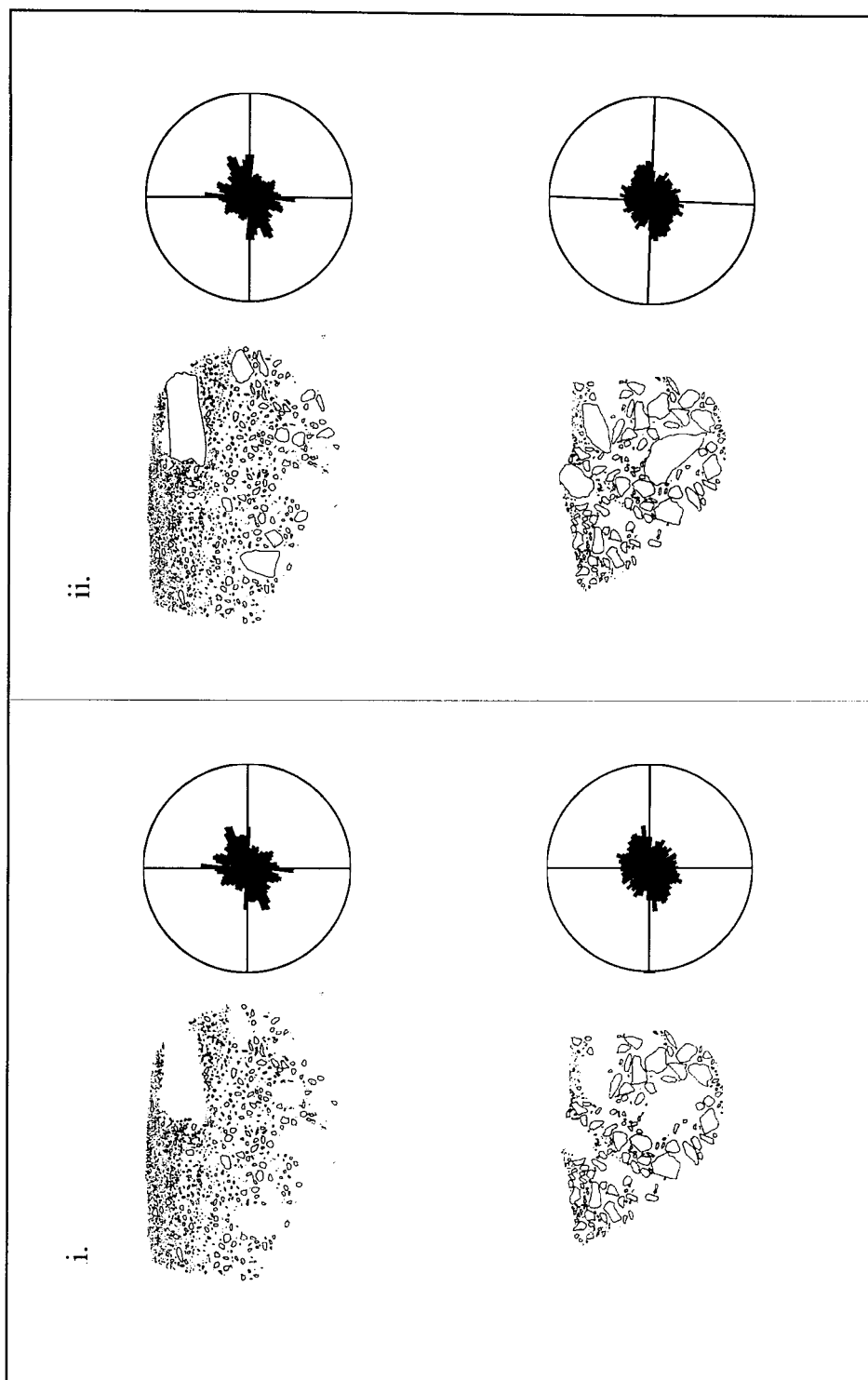


Figure 4.6b A comparison of Orien plots with

i: with the largest clast element removed

ii: complete data sets

This shows that the presence of the larger clasts within the data set in no way skews the distribution of the plots. The plots are true representations of the overall preferred orientations within the population.

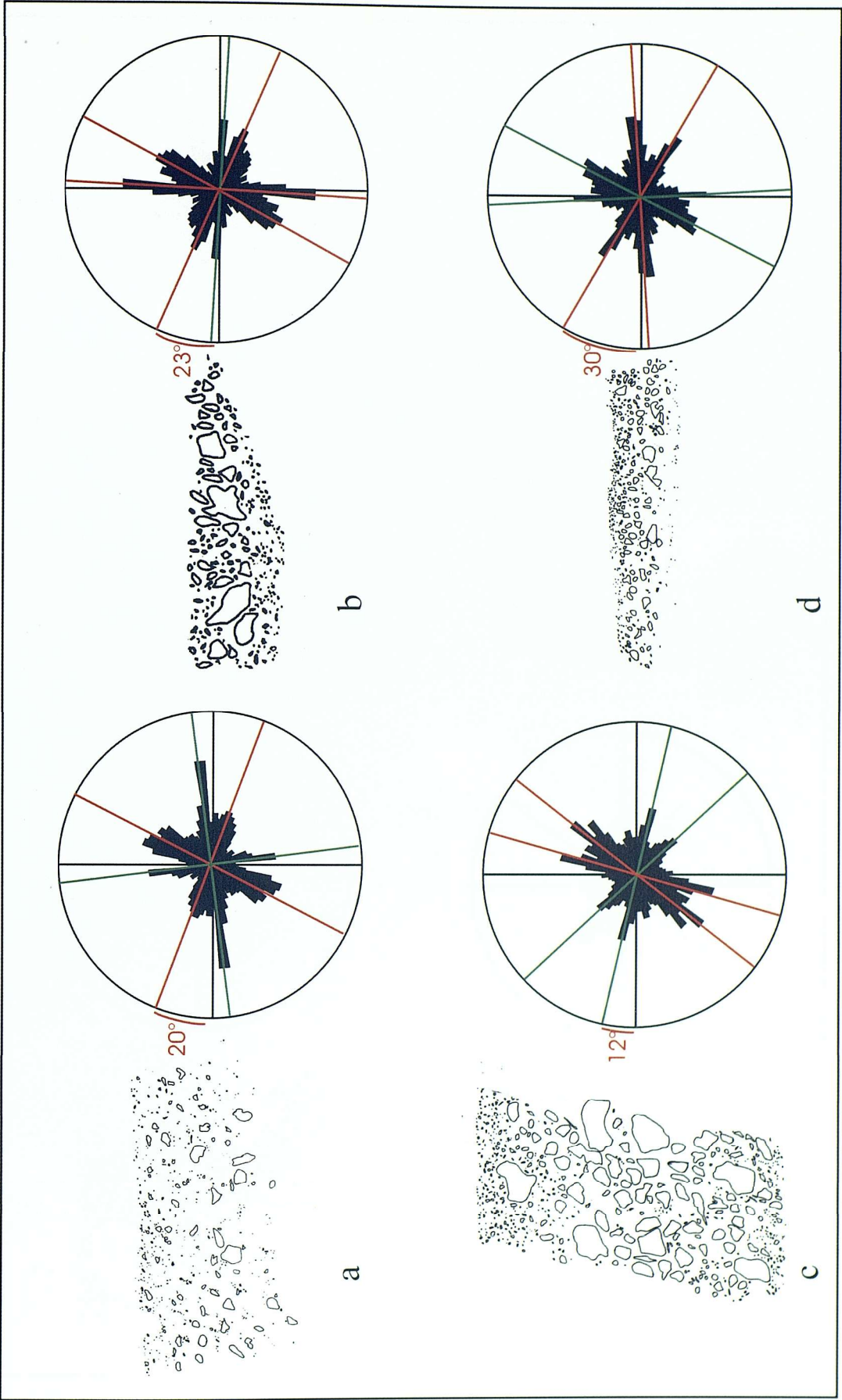


Figure 4.7 Clast preferred orientation plots of samples of the Rusey Breccia. usey Breccia.

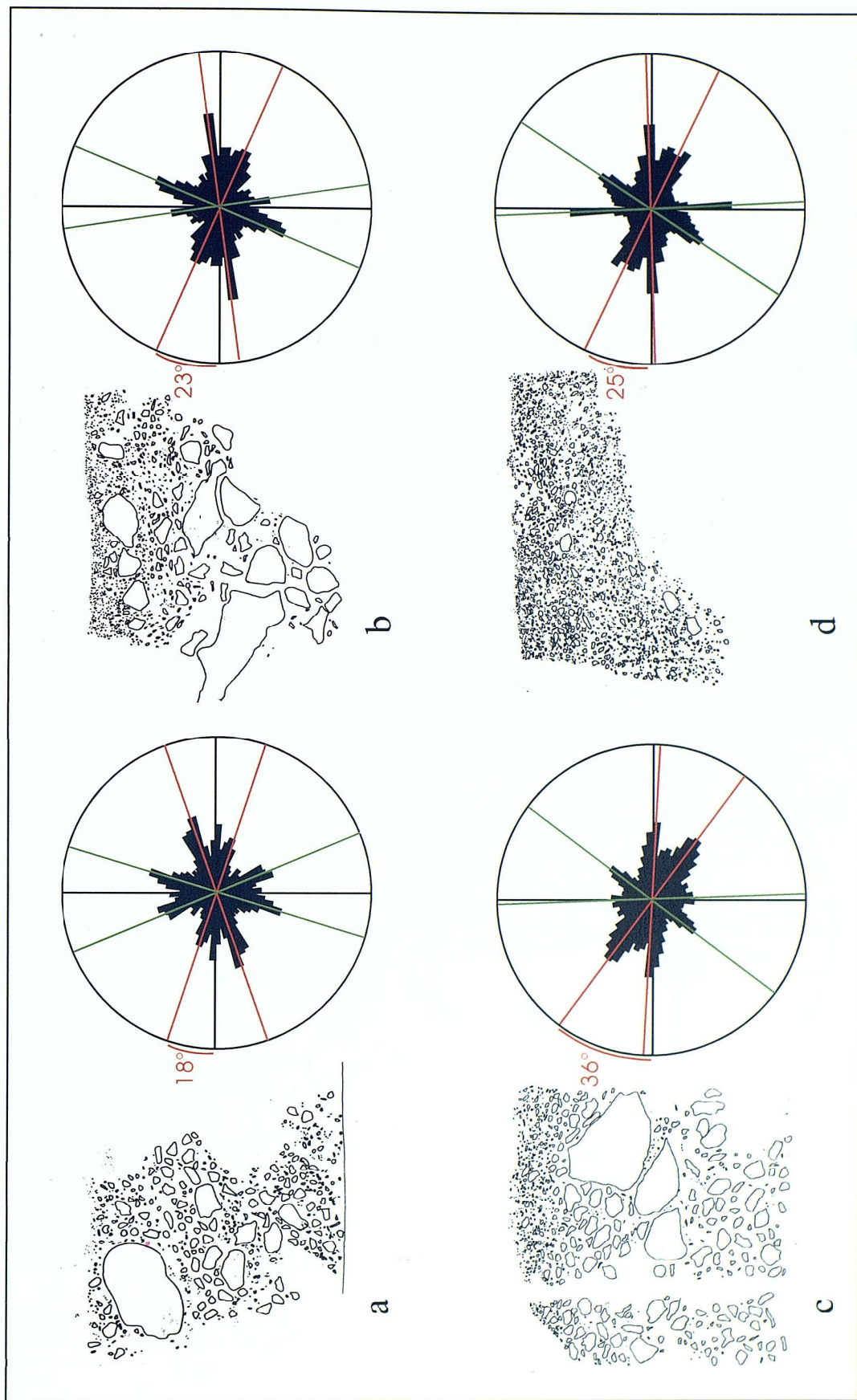


Figure 4.8 Clast preferred orientation plots of samples of the Rusey Breccia.

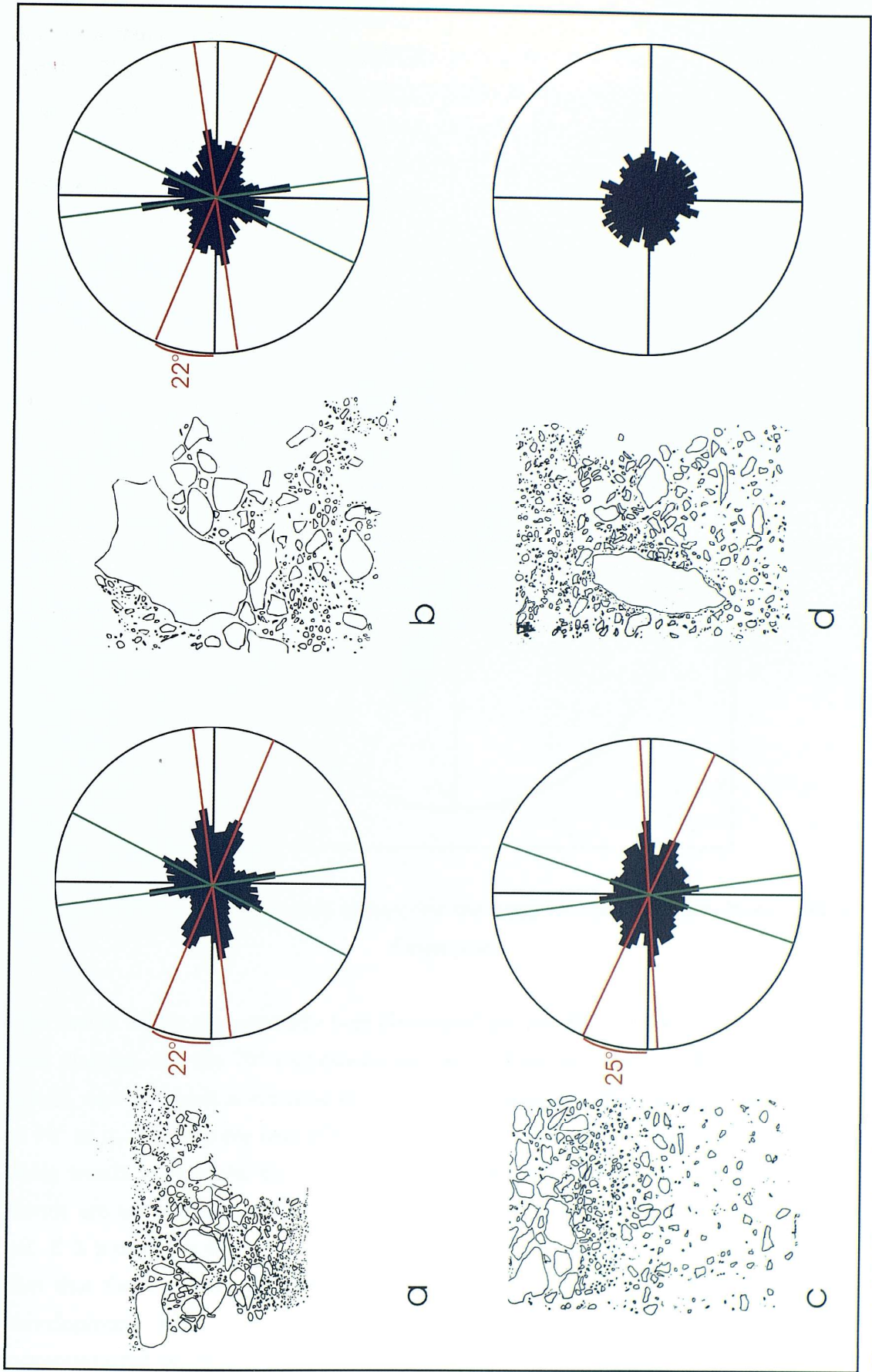


Figure 4.9 Clast preferred orientation plots of samples of the Rusey Breccia.

4.6:2b Sample Analysis

A representative selection of the preferred orientation plots are shown in Figs 4.7, 4.8 and 4.9. Thirty five samples were analysed of which 12 are shown here, illustrating the range of trend orientation and development exhibited by the full group. The terminology used in their description is illustrated in Fig. 4.10. All trends are referred to by their angular relationship to the upper half of the vertical axis of the plot, i.e. 30° clockwise, 60° anti-clockwise, as shown in Fig. 4.10. It can be seen that the majority of the samples show a distinct to very strong preferred orientation along 4 possible trends.

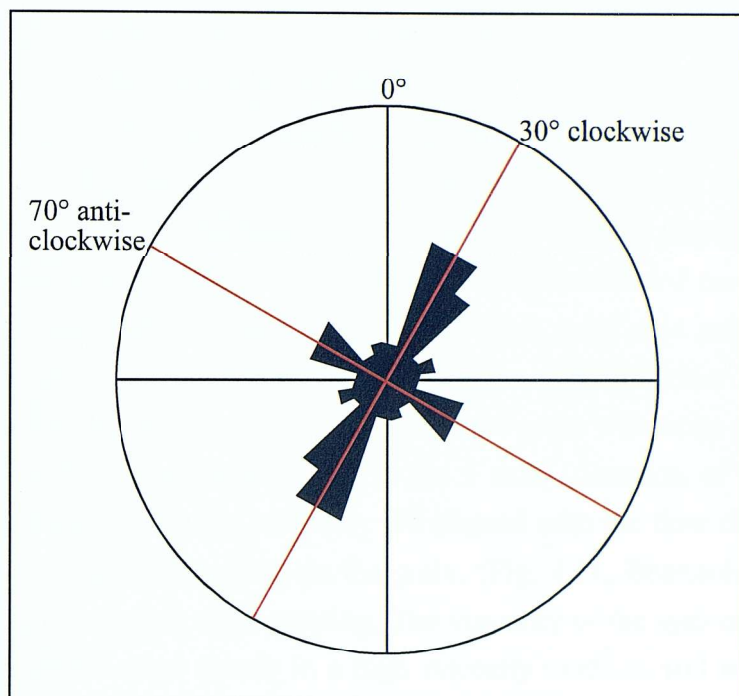


Figure 4.10 Terminology used to describe the trend orientations from Rusey Breccia Orient plots.

The trends which are generally best developed are the 30° clockwise orientation on the rose diagram and the 70° anti-clockwise trend. There are also two less well developed trends, one of which is oriented at 10° anti-clockwise to 10° clockwise and the other is at 90° to it, i.e. there are two sets of perpendicular trends. The degree of development of these trends is variable, but it is only in two samples, Fig. 4.9c and 4.9d, that all four trends are not present and only in one, Fig. 4.9d, that there are no discernible trends at all. It is a problem for the analysis that in situ samples are very difficult to obtain. The fact that the variation between trends in the samples is largely in the degree of their development rather than in their orientation is interpreted as indicating that the same organisational processes were operating in all the samples. Orienting all the samples with the layer boundary horizontal allows direct comparison of the results in terms of the preferred orientation within a layer. The results of this analysis give an assessment

of the preferred orientations of all the clasts in a single layer. It cannot assess any changes in the distribution of preferred orientations of clasts that may exist between different parts of a single layer.

4.6:2c Implications

Preferred orientations occur as a result of a particle being suspended in a moving fluid. Because the fluid is flowing there exists a velocity gradient between the fluid that is in contact with the channel sides and the fluid at the centre of the channel, such that flow is fastest in the centre of the channel. A velocity gradient is a pre-requisite for clast rotation. If all parts of the clast are being transported in the same direction at the same velocity the clast will translate but it will not rotate. The clast will only rotate if one part of the clast is moving at a greater velocity than another part. Work by Jeffrey, (1922), Bhattacharyya, (1966), and Gay, (1968a and b), has examined the behaviour of rigid particles, (Jeffrey, 1922 and Bhattacharyya, 1966) and viscous particles, (Gay, 1968a and b), both suspended in viscous media. They have demonstrated numerically and (in Gay's case) with some experimental evidence that high axial ratio grains rotate fastest when perpendicular to the flow direction and slowest when parallel to it, (Bhattacharyya, 1966), with the effect that an inclined grain will rotate about the z and y axis, Fig. 4.11, until it is aligned parallel to the x axis, (direction of flow). Therefore high axial ratio grains will rotate until they are aligned with the flow direction. Once in this position the shear couple acting on the grain, (Fig. 4.11, Bhattacharyya, 1966), no longer exists and the particle stops rotating. The viscosity of the system is significant in that a grain will rotate more slowly in a high viscosity medium and will remain in the aligned orientation longer. Bhattacharyya noted that this effect was best developed in flow through tubes. Gay (1968a) demonstrated that under pure shear a particle aligned with its long and short axes parallel to the stress axes will not rotate. All other particles will rotate and that the speed of rotation is dependent on the eccentricity or axial ratio of the grain. In his 1968b paper he also considered simple shear and demonstrated that particles rotate towards the shearing direction. This second paper considered systems where both particle and host medium were viscous and he demonstrated that the viscosity ratio between particle and host was critical in determining the deformation response. Experimental work done with rock clasts suspended in rock flour and compressed resulted in deformation in the clasts at very low stresses. The degree of deformation suffered by the clasts is a function of the viscosity contrast between rock clast and matrix, (Gay and Jaeger, 1975). In Gay, (1968b), it was demonstrated that with low viscosity ratios between the particle and the matrix, the particles rotated towards the shearing direction but the particle also deformed and would not be oriented parallel to the shear direction until shear strain was infinite. At high viscosity ratios however, the particle deforms far less and rotates more, such that it can rotate into parallelism with

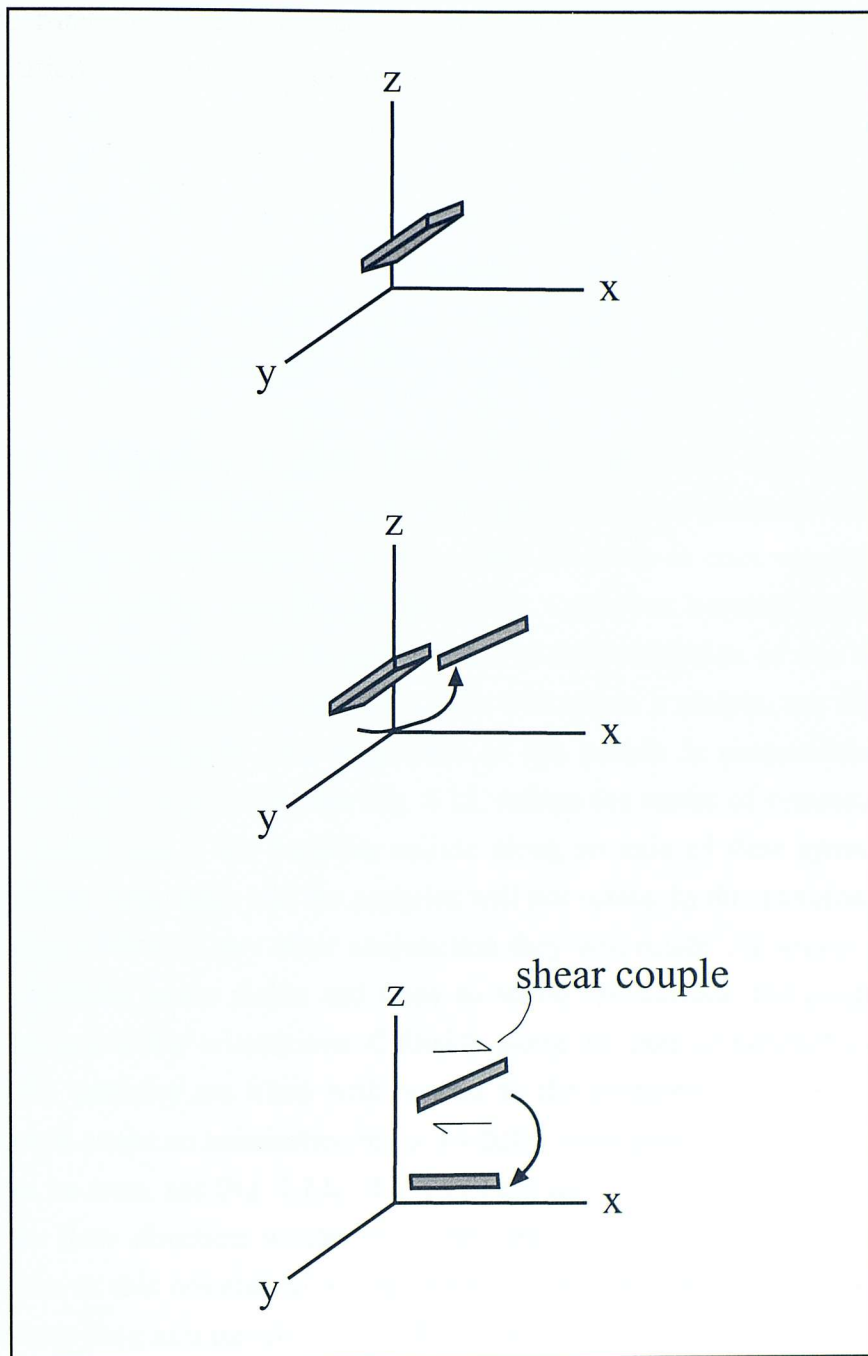


Figure 4.11 Grain rotation within a flowing medium
a: elongate grain oriented oblique to all three axes
b: Grain is rotated through the xy plane to lie parallel with the flow direction in the xy plane (flow is towards the right of the diagram along the x axis)
c: Grain is rotated through the xz plane to lie parallel to the flow direction in both xy and xz planes, after Bhattacharyya, 1966.

the shear direction. Gay (1968b) also presents some work on multiparticulate systems, where he noted that because they all rotated the same way a preferred orientation was developed and that the degree to which that preferred orientation was developed was dependent on the grain shape. The higher the eccentricity of the grain the longer it would take to rotate away from a shear direction-parallel orientation. This behaviour was also identified by Jeffreys (1922), for rigid particles in a flowing medium and the results of this arrested rotation has been observed in the field in igneous dykes, (Roberts and Sanderson, 1971, Ross, 1986, Rickwood, 1990) and clastic dykes, (Vitanage, 1954, Peterson, 1968). Gay, (1968b) also noted that the higher the particle density within the system the higher the bulk viscosity of the system, (cf. Elliston, 1984, Section 6:2.2, Change in viscosity of colloids with change in particle packing and Schugerl, 1971, Section 5:3, 5:4.1 and 2, Change in viscosity of fluidised beds with degree of expansion of bed)

Rees, 1968, notes that where the concentration of solid particles is high, (over 60%), collisional effects come into play. He explains the development of preferred orientations and the angle between that orientation and the flow direction in concentrated particle dispersions by the effects of intergranular collisions. Collisions between particles will create a force in each which will be proportional to the orientation of the tangential plane, Fig. 4.12a between the two particles. This will create a couple, see Fig. 4.12a, which will rotate the particle. The magnitude of the couple is proportional to the distance by which its line of action, see Fig. 4.12, misses the centre of symmetry of the particle. This means that if the particles collide along an axis of their symmetry the couple will have no magnitude and the particles will not rotate. In this position they are stable, but if they collide in any other conjunction they will rotate. He argues that this will preserve particles in the stable and close to stable orientations and progressively rotate all others into stable orientations. Collision along the axes of symmetry can only take place if the particles are tilted with respect to the direction of motion, see Fig. 4.12b. This would create an imbrication in the particles undergoing transport.

However it can be seen, see Fig. 4.12c, that there are two stable orientations. The long axis parallel to flow direction would be preferentially developed because collisions between particles in this orientation would cause less momentum transfer. This results in clasts with their long axis parallel to flow being less able to rotate than those oriented at 90° to the flow direction. Therefore it would be expected that there would be two sets of mutually perpendicular preferred orientations in a given flow system, (see data in Section 4.6.2b), with the one oriented closest to the flow direction being better developed.

Rees' work also showed that the collisions exert a dispersive force on the flowing system. The magnitude of this force is related to the angle of the impact such that the ratio of the tangential stress, (the stress imposed on each particle as a result of the impact, exerted in a direction perpendicular to the tangential line, see Fig. 4.12a), to the

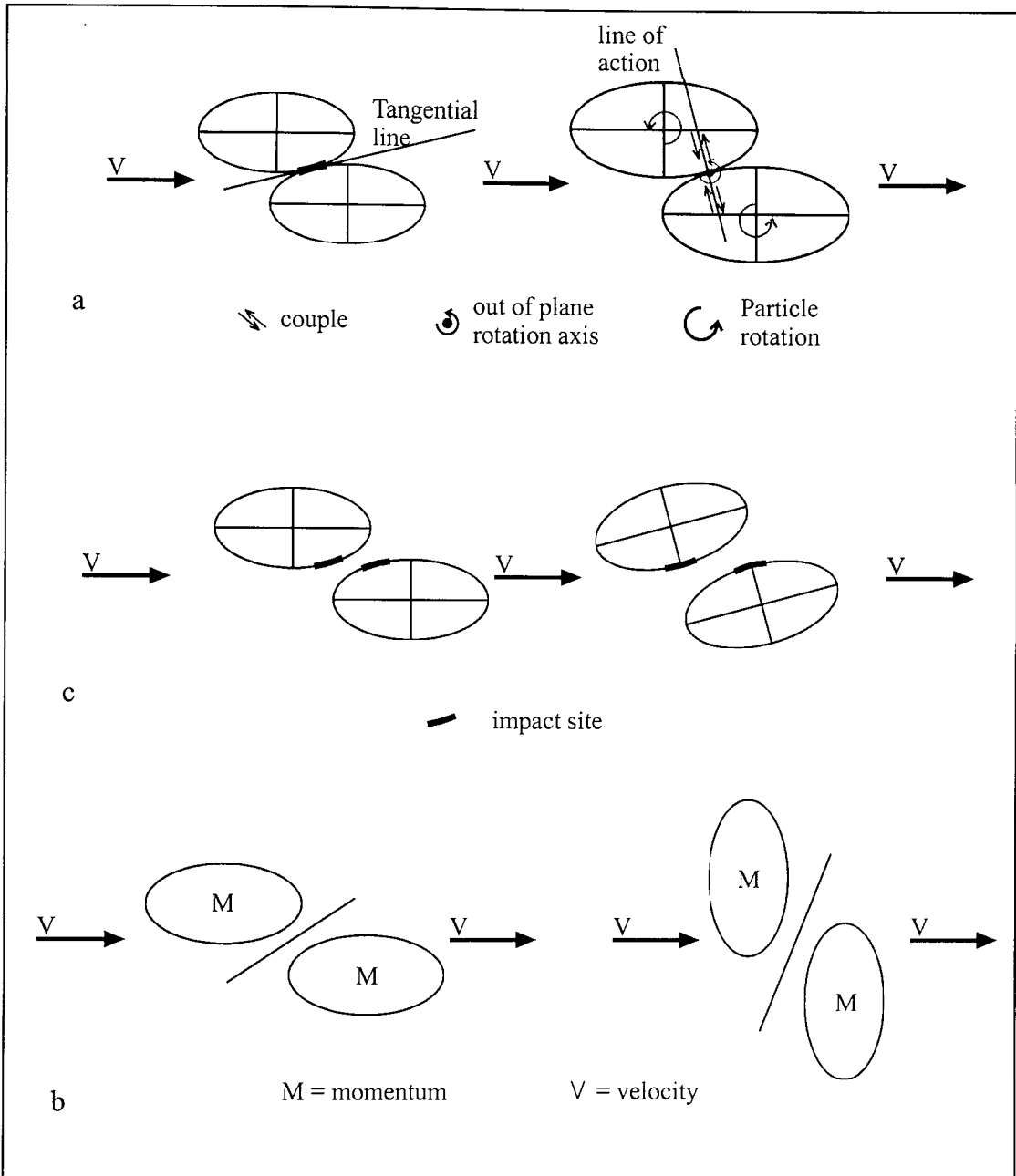


Figure 4.12 Diagram showing particle interaction

a: Creation of mechanical couple upon collision, causing rotation of the particles about an out of plane axis.

b: Requirement for the particles to be inclined before stable interaction, (i.e. impact where the forces created are coincident with the particle axes) can occur

c: Change in angle of impact for particles in stable orientations, showing the greater momentum imparted by an impact with the particle long axes oriented perpendicular to flow after Rees, 1968

dispersive force is a constant for a given fluid/solid system. For a real sedimentary system, this ratio is given as 0.32. $\tan^{-1}$ of this is 18° . This is the angle predicted as an average sediment imbrication angle, as this is the angle of rotation necessary to ensure that the resultant forces of a collision are acting parallel to the axes of symmetry of the particle. In his 1968 paper the angle of imbrication Rees derived from dune foreslope sampling is 11° , Fig. 4.13. Reineck and Singh, (1980), give the typical angle of imbrication for fluvial sediments as 15° to 30° and for beach deposits as 15° . They also note that preferred orientation of sediment grains is very sensitive to variations in flow direction and preferred orientations can vary significantly depending on local conditions. The Rusey data shows similar angles of preferred orientation, to both Rees' and Reineck and Singh's data, see Figs. 4.7, 8 and 9, although as the clasts are matrix supported this is not an imbrication texture, only an orientation texture.

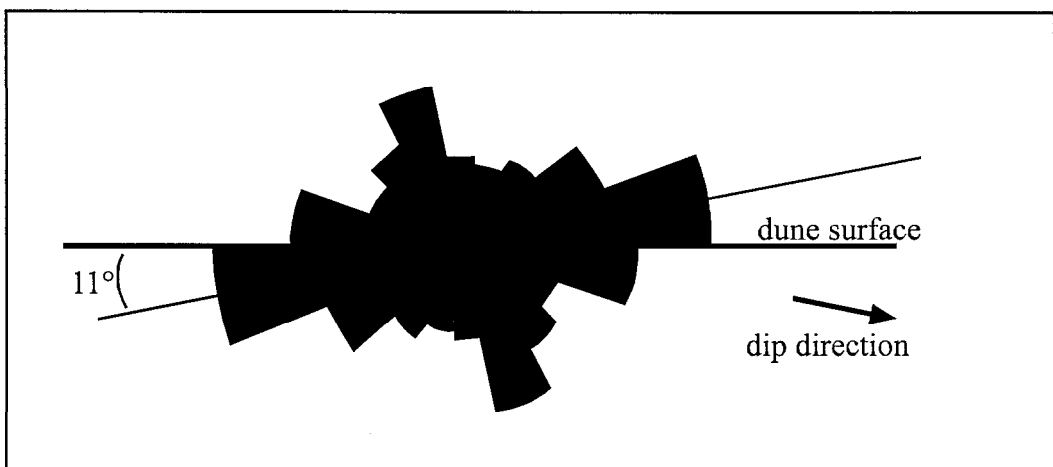


Figure 4.13 Rose diagram showing the orientation of long axes of sand grains in a section cut parallel to the dune face, after Rees, 1968

The majority of the Rusey preferred orientation data, (see Fig 4.7, 8 and 9) shows four trends. These usually consist of two well developed trends, marked in red, and two less well-developed trends oriented perpendicular to these, marked in green. Most of the samples with well developed trends show a major trend oriented at close to 20° from the horizontal, (the angular values are indicated on Figs 4.7, 8 and 9). The presence of minor trends perpendicular to the major trends may be indicative of an element of clast shape influence. This pattern could be caused if the clasts are approximately rectangular in profile. The long edges would provide the major trend and the shorter sides, the minor trend. However the clast profiles within the Rusey Breccia are generally sub rounded to rounded and angular clasts are not necessarily rectangular. This makes clast shape an unlikely factor in producing the subdivision of the preferred orientation directions into major and minor trends observed. Further clast shape analysis, (such as mapping the degree of shape variation from rectangular or square shapes in the Rusey

clast population) is required before the effect of clast shape on preferred orientation can be discounted completely.

The work of Schwan, (1989), compares natural and modelled grain fabrics. He shows that grain alignment can be adopted almost instantly when the grains are suspended within the moving fluid. In his experiments he uses an air current. The plots drawn from his work, Fig. 4.14a and b, show very similar distributions to the Orien plots drawn from Rusey data. The modelled data, see Fig. 4.14b, also show the tendency for the major preferred orientation to be slightly oblique to the flow direction, a feature which Schwan explains by permitting the air flow to meander, but which may be the result of the requirement for the grains to be oblique to the flow direction in order to acquire a stable position, as discussed above and illustrated in Fig. 4.12.

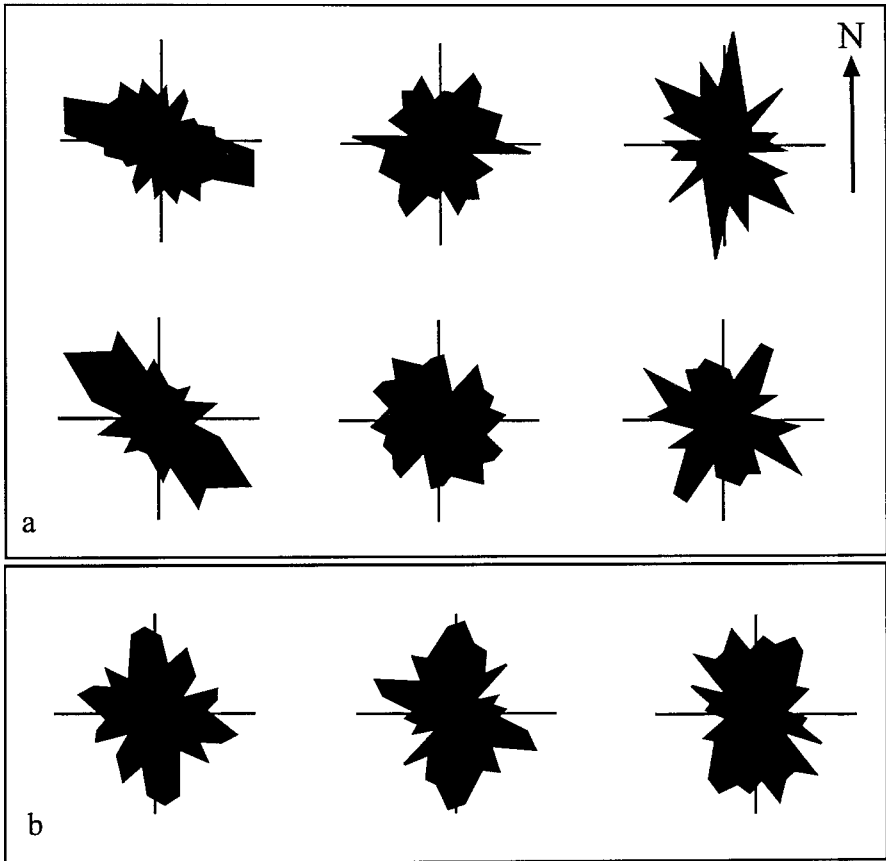


Figure 4.14
a: Preferred grain orientations from aeolian sands.
b: Preferred grain orientations from wind tunnel experiments.
Note that the plots are oriented with wind velocity directed towards N, after Schwan, 1989

The literature, (Jeffrey, (1922), Bhattacharyya, (1966), and Gay, (1968a and b), indicates that preferred orientations are caused by the clasts adopting an orientation within the fluid which requires the minimum amount of energy to maintain. When flow in the system ceases, there is a far greater probability of preserving clasts positioned in this orientation than in any other, (Bhattacharyya, (1966), and Gay, (1968b). For this self-orienting process to work the clasts must be suspended in a freely deformable

medium and that medium must have a velocity gradient across it making it possible for clasts to freely rotate and to impact each other. The experimental work of Gay and Jaeger, 1975 showed that deformation of rock clasts suspended in rock flour resulted in brittle deformation in the clasts rather than clast rotation. They numerically modelled the system by considering a 2 phase viscous system with the clast phase of a very high, (i.e. essentially solid), viscosity (Gay, 1968b). They found that where the 2 phases had a low viscosity contrast, (i.e. they approximated a rock clast/rock gouge system), the 'clast' phase deformed. It is only in a state of suspension within a fluid of much lower viscosity that the 'clasts' could be rotated into a common alignment without any internal damage. The main driving force for the development of such an alignment is the existence of a velocity gradient across the fluid body. Without this there is no reason for the clasts to change their orientation other than random grain-grain collisions which would not be likely to impose such a systematic and pervasive alignment. It would therefore seem that for the rotation and alignment of solid clasts the suspension of the clasts within a flowing fluid is the most realistic situation. The presence of preferred orientations in the Rusey Breccia is therefore taken as diagnostic of the involvement of a free-flowing fluid in the reorganisational process and specifically the involvement of a fluid that was capable of redistributing the clasts into a state of suspension.

Whilst the samples can be confidently aligned with respect to the layering, the lack of data about the sample orientation in the plane of the layering means that little can be said about the direction of flow of a fluid whose presence is deduced above. It is assumed, since the clast packages are layer bounded and clast orientations are so consistent when plotted with respect to the layer boundary that the fluid flow responsible for their reorganisation was contained within a given layer. It is also assumed from the range of angles exhibited by the preferred orientation with respect to the layer boundary that flow was directed parallel to the layer boundaries. This would mean that flow was also sub-parallel to the thrust and the boundaries of the main breccia body. However the lack of data on the orientation of the samples within the plane of the layering means that the direction of that flow cannot be assessed, i.e., it is not possible to tell whether the flow was up or down dip with respect to the layering or even if there was an out of plane element of movement. The direction of flow would be controlled by local stress gradients and may not be directly or simply related to the direction of movement on the thrust. The preferred orientations could not be used as a reliable indicator of the thrust slip vector. It is possible that the better developed trends in some sections, resulting from the fact that more of the clasts are in a particular orientation, shows that these sections are closer to the flow-parallel direction, whereas samples with the less developed trends are closer to the flow-perpendicular direction, Fig. 4.15 where the section in view can have a much wider range of orientations. This holds out the possibility that a more intrusive sampling method, such as cutting fresh samples directly

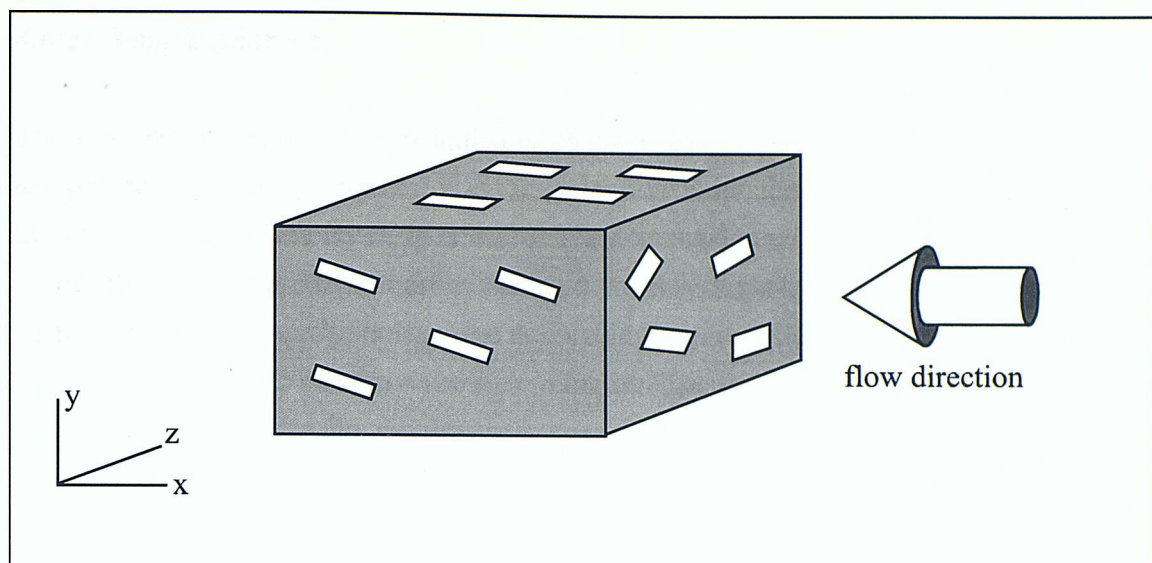


Figure 4.15 Flow orientation of non-equant clasts. When the flow direction is parallel to the xy/xz planes, the rectangular 'clasts' will be aligned in these planes. They will be randomly oriented in the yz plane. Therefore in the yz section, the clasts may show a variety of orientations, whereas in the xy and xz plane they are more aligned.

from the in situ outcrop, could define the direction of flow completely in three dimensions.

4.6:3 Clast Density Variation

4.6:3a Methodology

The technique used to estimate the amount of clast material in a given sample of breccia was very simple. The clast *density*, (i.e. the percentage of clast material in a given space) is estimated from the areal extent of the clasts in a sample section. The sample is represented (by hand) as a line drawing and the area occupied by clast material is coloured black. This drawing is then scanned at high resolution and the resulting image loaded into a program, Propscan, which scans a user-defined area and returns the percentage value of black pixels within that area. This technique was initially applied to the entire sample, but more informatively it has been applied to a sequence of strip profiles. Samples were oriented with respect to the layer boundary which was used as the reference datum. The strips provided a series of profiles which are presented graphically to show the variation in amount of clast material across each layer.

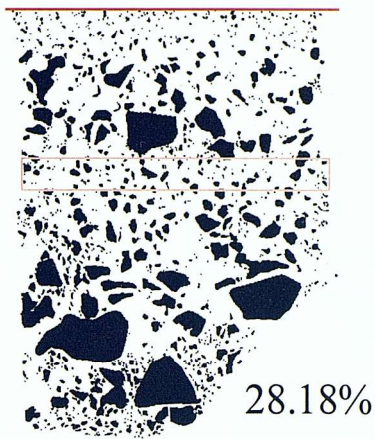
4.6:3b Sample Analysis

The analysis of whole samples indicated that the amount of clast material in any given sample rarely, if ever, exceed 50%, Fig. 4.16. In many cases it did not exceed 30% of the total volume of the layer. In addition to determining the percentage of clast material within the layer, the *distribution* of the clast material within the layer was studied using strip profiles. Eighteen samples were evaluated and the clearest examples are shown in Fig. 4.16.a, b and c. The scan profiles indicated that the total amount of clast material carried by the layer varies systematically across the layer. The variation in amount of clast material carried is smooth within the resolution of the sampling. Whilst the results are not as clear cut as for the clast *orientation*, it can be seen that the samples contain an ordered variation in amounts of clast material across the layer. Where layer boundaries are present it is observed that there is a systematic increase or decrease in amount of clast material with distance away from that boundary. This is particularly well illustrated by bw10 and bw1, see Fig. 4.16. If there is a causal link between layer boundaries and grading patterns, this would imply that;

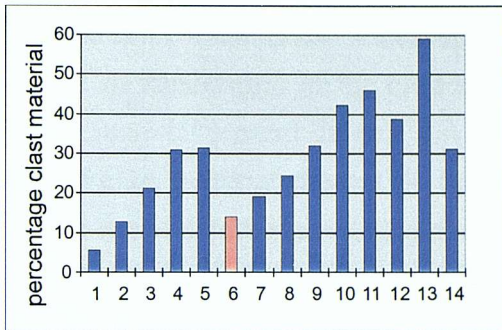
- layer bound processes control the variation in the amount of material present and
- the systematic variations seen in bw10 and bw1 are indicative of a layered structure in the sample which is not apparent to the naked eye.

The layering within the breccia is variably developed. It is often poorly defined and there are layers which are delineated only by visible differences in clast burden, see Plate 4.6. However, as mentioned above, even where there is no such obvious discontinuity, statistical analysis may reveal a ‘cryptic’ layering. It is not clear from the samples presented whether the variation can be regarded as being mono or bi-directional. Bw10 and bw12 seem to indicate mono-directional variation, (repetitive in the bw10 example), whilst bw1 tends more towards a bidirectionally ordered sequence. It should be made clear that whilst the layer boundaries are taken as the reference datum, the original way up of these samples is unknown and therefore the original direction of the mono-directional ordering in these samples is unknown.

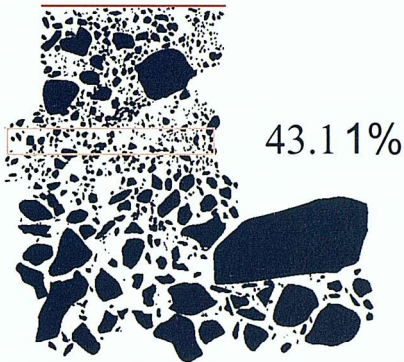
Where layers without distinct layer boundaries are visually apparent, it is generally the variation in clast *size* which makes them discernible. Fig. 4.16a and b show that the variation in the amount of clast material present is systematic, steadily increasing to a peak and then decreasing more or less abruptly to a minimum value and steadily rising again from there. Any variation in clast shape is likely to produce areas of each layer where the amount of clast material present is at a minimum. However when the layer boundaries can be detected systematic variations are observed across the layer and the author suggests that the presence of a systematic variation in the amount of clast material present, Fig 4.16c, can be used as an indication of the presence and orientation of a ‘cryptic’ layer boundary, (see Fig. 4.16a and c). It can be argued that clast size



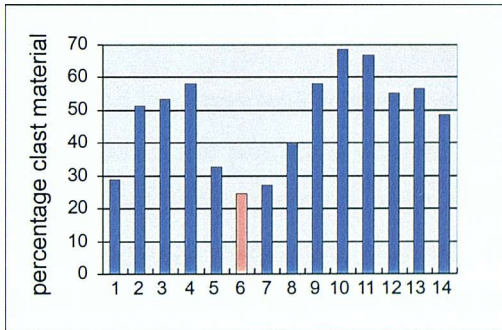
a



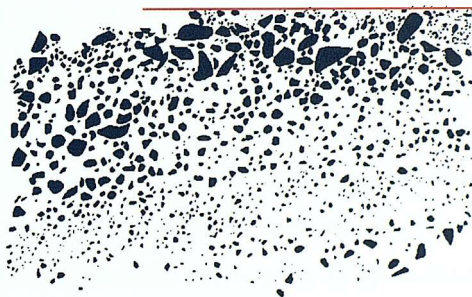
bw10



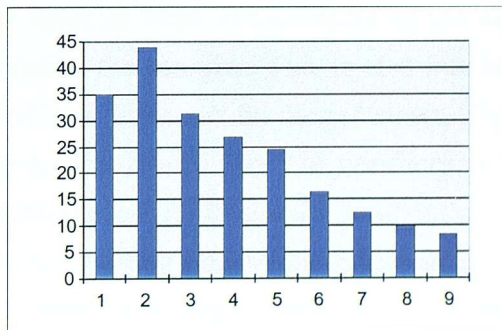
b



bw1



c



bw12

Figure 4.16 Variation in clast volume with distance across a layer, x axis numbers 1-14 corresponding to top to bottom strip profiles across each sample. Percentage value quoted is total percentage of clast material in the whole sample

variation would be a more logical choice for the detection of cryptic layer boundaries, given that one of the field criteria used for defining layers is the presence of clast size grading. However quantifying the clast size proved more complicated and very much more time consuming than a simple quantification of the amount of clast material present in a given sample. Since a comparison of the results suggests that the clast size and the amount of clast material present correlate well, the latter quantity was used by the present author to detect the orientation of the layer boundaries. Assessment of the clast size, as described in section 4.6:4, required an idealisation of the clast shape that the quantification method for the amount of clast material present did not. On this basis alone, the present author regards the amount of clast material present as being more accurately assessed and the distribution patterns derived to be more reliable. A more detailed discussion of the disadvantages of the assessment of clast size is presented in sections 4.6:4a and b.

4.6:3c Implications of variations in clast material distribution for breccia reorganisation mechanisms

The variation in the distribution of clast material and the proposed correlation with clast size, although not quite as clear cut as the presence of the preferred orientations, does provide supporting evidence for the hypothesis that these patterns were formed by organisation within a flowing fluid. The fluid flowing within the channel, (breccia layer), acquired a velocity gradient such that the fluid was moving most slowly where it was in contact with the channel sides and the highest velocity fluid was in the centre of the channel. The slower moving fluid would not be capable of transporting as high a bulk volume of clast material as the higher velocity fluid in the centre of the channel, thereby creating the volume gradient preserved in the samples. This is also true for clast size. Only the fastest flowing fluid will be able to transport the larger clasts, Fig. 4.17. The presence of mono-directional grading in the Rusey Breccia conflicts with this theory as flow in a channel would have a symmetrical flow velocity gradient across the channel. This should always give bi-directional grading and so the layers in the Rusey Breccia should always be bi-directionally graded. However clasts of the size found in the Rusey Breccia, (1-100mm) would require fluid velocities in the region of 1-10m/s to remain in suspension, see Fig. 4.17. If that velocity drops the clasts will be deposited, as they are during sedimentation. This gravitationally controlled process will produce a mono-directionally graded layer. Thus, if the original orientation of the samples is known it would be possible to determine whether the mono-directional grading could have been gravitationally produced. Unfortunately, sampling of in situ rock is extremely difficult and the original orientation of most samples is unknown. There are a number of possibilities for the creation of this grading. It could be that re-shear along the breccia layer margins produced reduction of clast size at the layer margins and that the change

in amount of clast material carried is due to the subsequent flushing out of the finer grained element (derived from the comminution of the clasts during shearing) from the clast assemblage. The surface forces acting between particles of very small diameter cause an increase in the cohesion of a granular bed at low particle sizes. Therefore the velocity required to mobilise the bed increases from a minimum value at intermediate particle sizes to maximal values as the particle size either increases or decreases. Any fluid moving fast enough to sever the greater interparticulate forces, e.g. capillary forces, exerted on the fine grained element would also be capable of mobilising the coarser grained element of a saturated granular assemblage and would impose the observed grading pattern. There seems no need to appeal to a previous shearing event to explain the textures displayed by the breccia if the flushing process would subsequently generate them anyway. It is also possible that such a grading could be produced by variable dissolution of clast material by invading fluids. This would imply that the smallest clast size would correlate to the greatest degree of dissolution. This would mean that the smallest clast size would coincide with the paths of maximum fluid flow, i.e. the centre of the flow channels, rather than the edges. This view of the layer structure is not supported by the field evidence. The field data show that the layers are often bounded by very well defined surfaces. A dissolution mechanism would be expected to produce a relatively poorly defined layer, since the amount of dissolution will be controlled by the penetration of the fluid. The penetration of fluid would be controlled by the variation in porosity and permeability, which will be controlled by the size, shape and packing of the grains. The penetration of fluid through a granular medium is unlikely to produce such sharply defined planar boundaries as seen in Plate 4.3/4.4. A dissolution mechanism offers no explanation for the development of a well defined surface in the centre of the fluid flow channel. There is also no mechanism to adequately explain the dissolution of clast material and the precipitation of quartz matrix synchronously at the same point. However if it is argued that the dissolution and precipitation occur separately, problems still arise in preserving the dissolution grading whilst mineralising the system.

4.6:4 Clast Size Grading

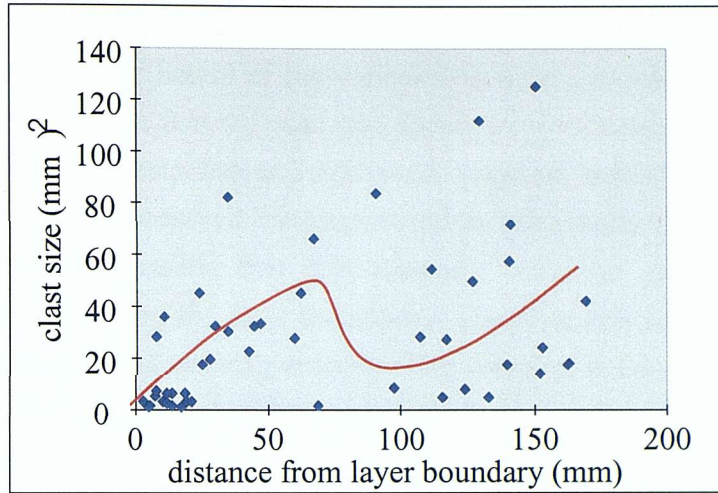
4.6:4a Methodology

The variation in clast size is readily apparent to the eye in many samples of the breccia, see Plate 4.9a and b. However the quantification of the variation proved more difficult. Grain size variation in sedimentology is usually examined by means of separating the samples into classes of grain size by passing the material through a number of sieves and weighing the fraction so derived, (see Tucker, 1988). This is a standard technique

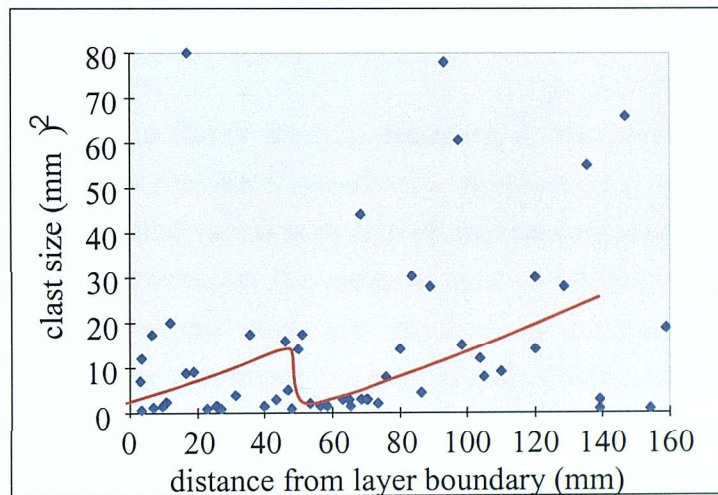
and is used to produce grain size histograms similar to those produced for amount of clast material in the previous section. It is inapplicable to this situation however as the Rusey Breccia clasts are not loose sediment and cannot be sieved. Therefore a means of assessing the area occupied by each clast section was required, this being the only clast size data available. It was not possible to use the proportional assessment technique from the previous section as it was the variation between individual clasts that was being examined, not the total amount of clast material within the scan area. Time constraints rendered it impossible to create a program which could calculate the area of the clasts as represented by the line drawings made for input into Orient, but it is suggested as part of the proposed further work from this project, that such a program should be written, as it would have applicability far beyond the scope of this project, particularly for strain and fabric analyses. A cruder technique was used to provide a initial quantitative assessment of the clast size grading seen at Rusey. The samples used were the same group used in the previous section and the layer boundary was once more taken as the reference datum. The method consisted of placing a sample line across the sample being assessed. Any clasts intersecting this sample line were assigned a long and short axis as described in section 4.6:2a, (technique one). An estimate of the clast size was derived simply by assuming the clast to be a rectangle and calculating the area of that rectangle. The area of the clast and the distance of the long and short axes intersection point were recorded and plotted against each other as a scatter graph, producing a graphical representation of the variation in clast size with distance from the layer boundary, Fig. 4.17. It is appreciated that equating the clast to an ellipse yields a more accurate measure of the grain size, (as proposed by Simigian and Starkey, 1986). However the analysis is intended to provide a description of the variation in clast size, rather than providing an accurate assessment of each individual clast area, i.e. its absolute size.

4.6:4b Sample Analysis

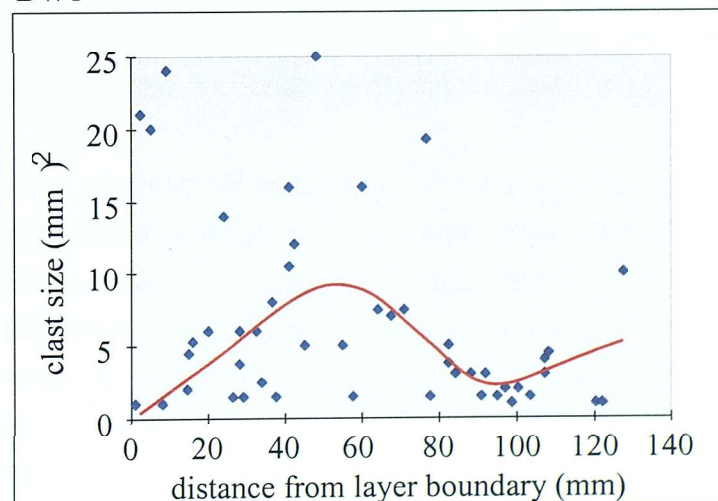
Fourteen samples were analysed, of which three are shown in Fig. 4.17. These are the same three samples as are shown in Fig. 4.16. It can be seen that the data shows much less well defined trends than are seen for the graphs of variation in total clast material. This is partly because line sampling permits a very much smaller percentage of the available data to be analysed than was possible with the previous analysis, emphasising the effect of anomalous data in the data set represented in Fig. 4.17. The larger clasts were found to be extremely few and evenly scattered throughout the data set and were eliminated from the plots in order to present the data at a scale which allowed the trends to be seen. Whilst the clast size variation can be seen to loosely follow the trends shown by the total clast material variation, a larger data set would be needed to confirm this conclusively.



BW10



BW1



BW12

Figure 4.17 Clast size distributions for the samples shown in the previous figure. Variation trends marked in red.

4.6:4c Implications

The data set is not well enough constrained for firm conclusions to be drawn regarding the significance or systematic nature of the variation in clast size. However the initial results of this analysis indicate that the clast size does vary systematically across layers. This is an important point to establish as a systematic variation requires the action of an organising mechanism. The observed variation could be adequately explained by fluid remobilisation. It is also possible that this variation could be explained by clast dissolution by fluid input along the layer boundaries. However this has been discussed (section 4.6:3c) and is regarded as being unlikely. The variation in clast size distribution parallels that for clast volume distribution (section 4.6:3) in that the finer element of the clast burden correlates with the least amount of material carried by the layer. However, further sampling would be necessary for this correlation to be confirmed.

4.7 Summary of the statistical analysis

The statistical analyses of the Rusey Breccia described in the previous sections are limited in nature and aim to provide a quantitative demonstration that the preferred orientation of clasts and a spatial variation in size of clast and amount of clast material exists within the layers of the breccia. The methods used to illustrate these features of the breccia are generally simple and are standard or derived from standard sedimentological techniques for examining variation in grain distribution. The exception is the use of closed polygons and line segment weighting in the assessment of grain orientation. This technique is used to avoid the imposition of an unrealistic element of directionality on individual clasts and the resulting inaccuracy in the rose diagram plots. The present author is indebted to Dr. Richard Jolly for drawing this technique to her attention and for providing the reference justifying its application, (Sanderson and Phillips, 1987).

Whilst the presence of a preferred orientation of clasts places certain significant constraints on the environment in which it was generated, the details of the orientation trends require further analysis. It would be necessary to examine both the trends disclosed and the possibility of artifacts being imposed on the distribution by the method of analysis before the source and significance of the trends could be definitively assessed. For example the presence of two sets of perpendicular clast edge orientation trends in many of the Orien plots, (see Figs. 4.7, 8 and 9), may be the result of clast shape or it may relate to the organisational processes which formed the orientation trends, (e.g. Rees, 1968). The analyses carried out by the present author does not enable one to determine which if either of these effects governs the textures observed. Nevertheless the textures observed and quantified in the breccia samples, namely;

- the presence of preferred orientations
- the presence of a systematic variation in the amount of clast material contained by different parts of individual layers
- a corresponding systematic variation in the size of clast present in different parts of individual layers

do provide evidence of organisation by transport in a free flowing fluid, i.e. a fluid which transports the clast burden rather than permeating and flowing through a clast-supported framework. The presence of clast preferred orientation is particularly important because the presence of a mobile fluid is an essential condition for preferred orientations to develop.

Chapter Five

Fluidisation Mechanisms and Processes

5.1 Introduction

In the previous chapter the Rusey Breccia has been described with particular reference to certain features, namely the degree of organisation exhibited by the breccia clasts and the style of the mineralisation which forms the supporting matrix. In chapter seven a model for the breccia evolution will be described, using the processes of fluidisation and gel formation. This chapter will discuss in detail the process of fluidisation, demonstrating that this process has both the capability to produce the organisational features observed in the breccia and has been observed as a geologically viable process, active in syndeformational environments characterised by high fluid pressure and fracture formation.

5.2 Fluidisation

Granular porous media can be fluidised by the input of moving fluid (Richardson, 1971) or the input of energy in the form of vibration or mechanical shock (Nichols, 1995). A state of fluidisation is achieved when the particles in the bed are supported entirely by the fluid and are no longer in physical contact. As shear resistance is the result of the frictional behaviour of grain-grain contacts the bed has no resistance to shear at this point and will behave in a very similar manner to a fluid. For this reason the process is known as fluidisation. Fluidisation is the process of achieving fluid-like behaviour in a bed of granular material by passing a moving fluid through the bed. However the term is often used to denote the 'family of processes' which can produce zero shear strength, and thereby pseudo-liquid behaviour, in a granular material (Nichols, 1995). The process of fluidisation *sensu stricto* will be covered in section 5.2:1a and b. Section 5.2:2a and b will cover the processes of liquefaction and shear liquefaction, which are included for completeness. They are not regarded as playing an important part in the modification of the Rusey Breccia because, as described in Chapter 4, the preferred orientation of clasts requires a moving liquid within which the clasts are oriented. In

neither liquefaction mechanism is the fluid itself moving. It is the grains which move through a stationary fluid, impelled by mutual collision. The orientation of the grain is therefore controlled by random grain-grain collision rather than by the systematic effect of the flow velocity gradient.

5.2:1 Fluidisation (*sensu stricto*)

5.2:1a Fluidisation mechanism

In the following description the sample being fluidised is subjected to fluid input at its base and fluid flow is upwards through the sample, Fig. 5.1a.

The initial state of the bed, resting grain against grain, with individual grain weight supported on the grain-grain contacts is termed the 'fixed bed' state, Fig. 5.1a.

When fluid is passed through the bed at low velocity the state remains fixed. As fluid velocity increases, the upward force of the fluid is transferred to the grain by fluid drag, with the result that the pressure exerted across the grain boundaries drops throughout the bed. Eventually, when the fluid velocity reaches a critical point the entire weight of the grains is supported by the fluid and the bed loses shear strength. The particles rearrange in such a way that the resistance to flow is minimised and the voidage (open space) in the bed increases. Further increase in fluid velocity results in further expansion and the particles become physically separated and are free to move with the flow, Fig. 5.1b. This is the point of incipient fluidisation and the velocity of the fluid at this point is the minimum fluidisation velocity (U_{mf}) for this bed (Richardson, 1971).

If the flow rate continues to increase the particles are transported with the flow, Fig. 5.1c. At this point one of two things may happen: either the bed will continue to expand uniformly or the excess fluid will pass through the system in the form of bubbles.

The former process is known as **particulate fluidisation** and is the norm for liquid/solid systems, Fig. 5.2a. The latter is known as **aggregative fluidisation** and is the norm for gas/solid systems, Fig. 5.2b (Zenz, 1971). The essential difference between the two is the maximum size of bubble which is stable within the system. The velocity with which bubbles rise increases as the bubble diameter increases. If the bubble rise velocity exceeds the terminal falling velocity of the particles, (the greatest velocity the particles will reach in sinking through the fluid under the influence of gravity), then the particles will be sucked into the base of the bubble as it passes (Richardson, 1971, Zenz, 1971).

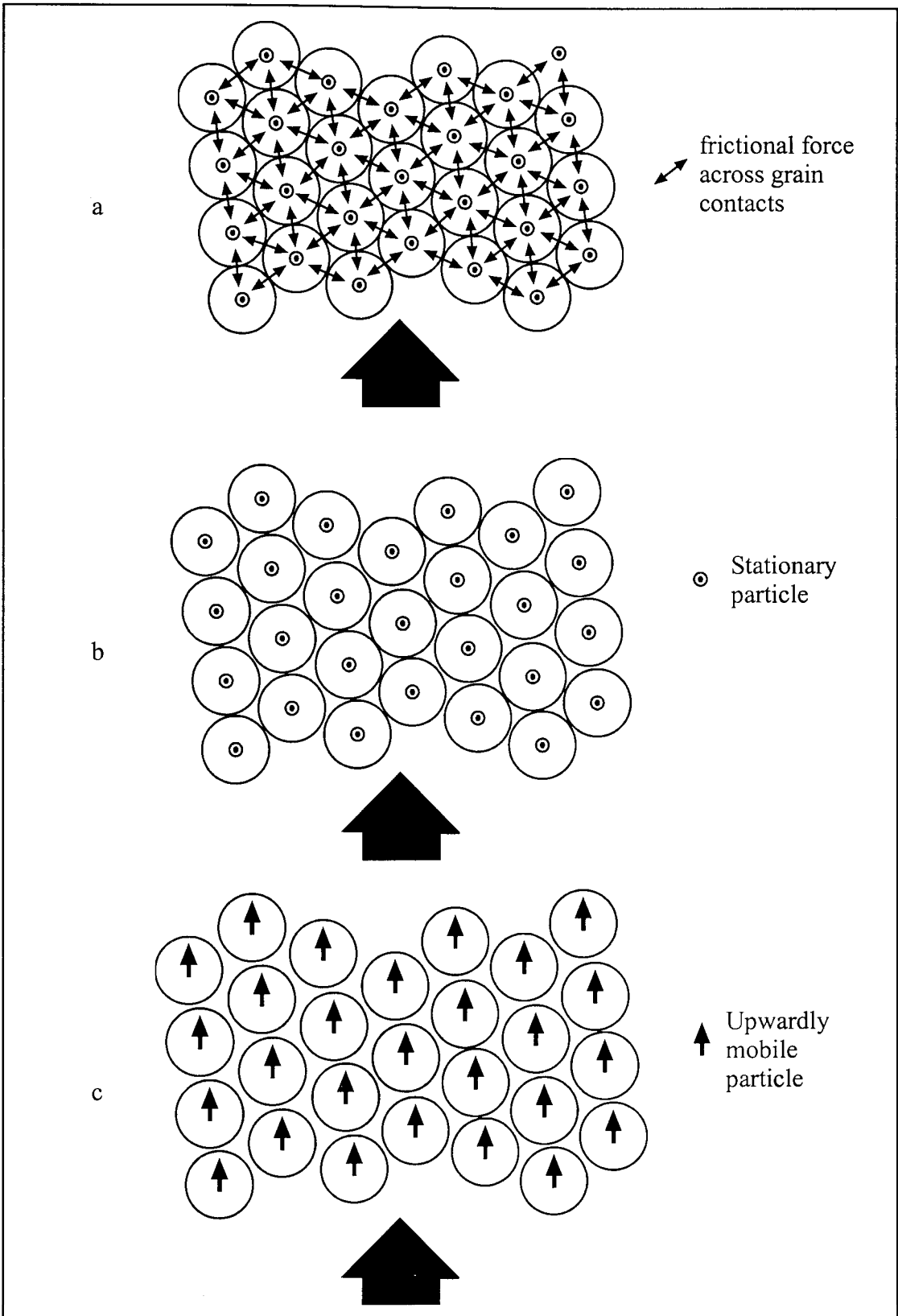


Figure 5.1 The fluidisation process;
a: Pre-fluidisation. Fluid flow is too slow to relieve the intergranular friction, produced by the pressure of grains stacked against one another
b: Incipient fluidisation. Fluid flow is strong enough to relieve the intergranular pressures and the bed has no shear strength, but as yet the grains will not move.
c: Fluidised state. The grains are supported and transported by the fluid flow'

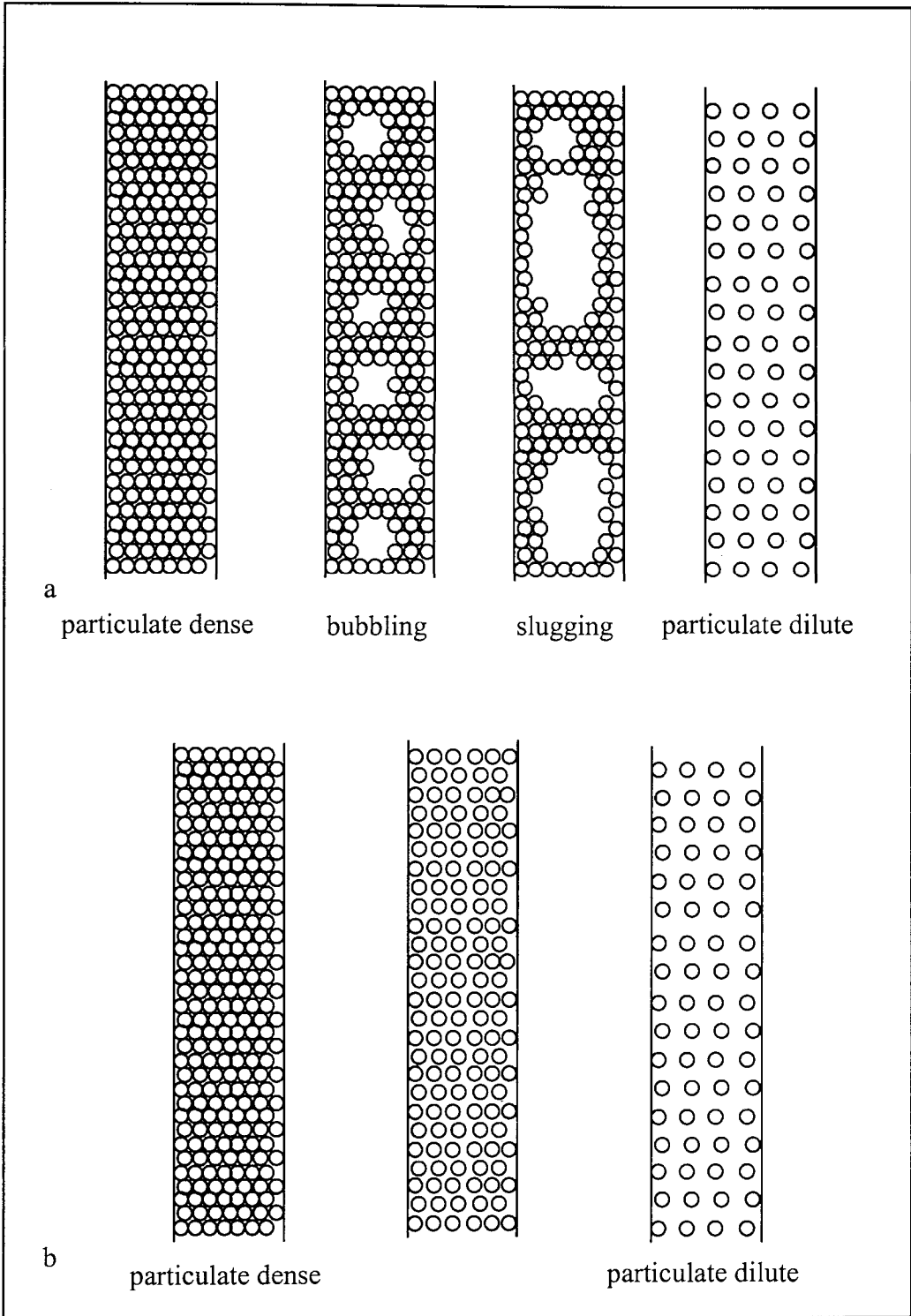


Figure 5.2 Fluidisation styles
a: aggregative fluidisation
b: particulate fluidisation
after Richardson, 1971

This will destroy the bubble. This process will occur in all fluidised systems and the difference between aggregative and particulate fluidisation is dependent on the size of the bubbles that can developed, though the precise definition is qualitative. If the bubbles are big enough to see then the fluidisation style is aggregative. If they are too small to see, then the particle dispersion is apparently uniform and the style is particulate. Thus the maximum bubble size is controlled by the terminal falling velocity of the particles, which is governed by the density contrast between the particle and the fluid. The higher the contrast, the higher the terminal velocity of the particles. This means that the bubble velocity must be much greater before the upward drag of the bubbles passage (Fig. 5.3, Collins, 1984) can overcome the downward velocity of the particle and draw particles into the bubble. As bubble velocity only increases with bubble diameter, a higher density contrast between fluid and particle will cause an increase in the maximum stable bubble size. Therefore the higher the density contrast between the fluid and solid phases, the more likely it is that aggregative fluidisation will occur. For this reason liquid-solid systems generally show particulate fluidisation. The cited studies of bubble growth and movement (Davidson and Harrison, 1971, Bugg et al, 1989) do not cite particle size as relevant and appear to have been carried out with grains of uniform size. However Stokes law,

$$V_0 = (1/18) * ((\rho_p - \rho_f)gd^2)/\eta$$

V_0 is terminal velocity, ρ_p is the density of the particle, ρ_f is the density of the fluid, g is acceleration due to gravity, d^2 is the square of the particle diameter, η is height of fluid column.

describing the terminal velocity of spherical particles falling through a viscous fluid, states that the terminal velocity is related to both the density contrast and the square of the particle diameter. This relationship is applicable to both sedimentary suspensions and to fluidised systems and predicts that the size of the particle determines the terminal velocity of the particle and therefore the maximum stable bubble size that can be developed. As studies of bubble development are uncommon outside the engineering literature and those studies do not concern themselves with particles of varying size, references specific to this situation are not known to the present author. The bubble's upward velocity (which is determined by the size of the bubble) must exceed the terminal velocity of the particle before it can draw the particle into the tail of the bubble, thereby destroying the bubble. It can be argued that the maximum bubble size in a given

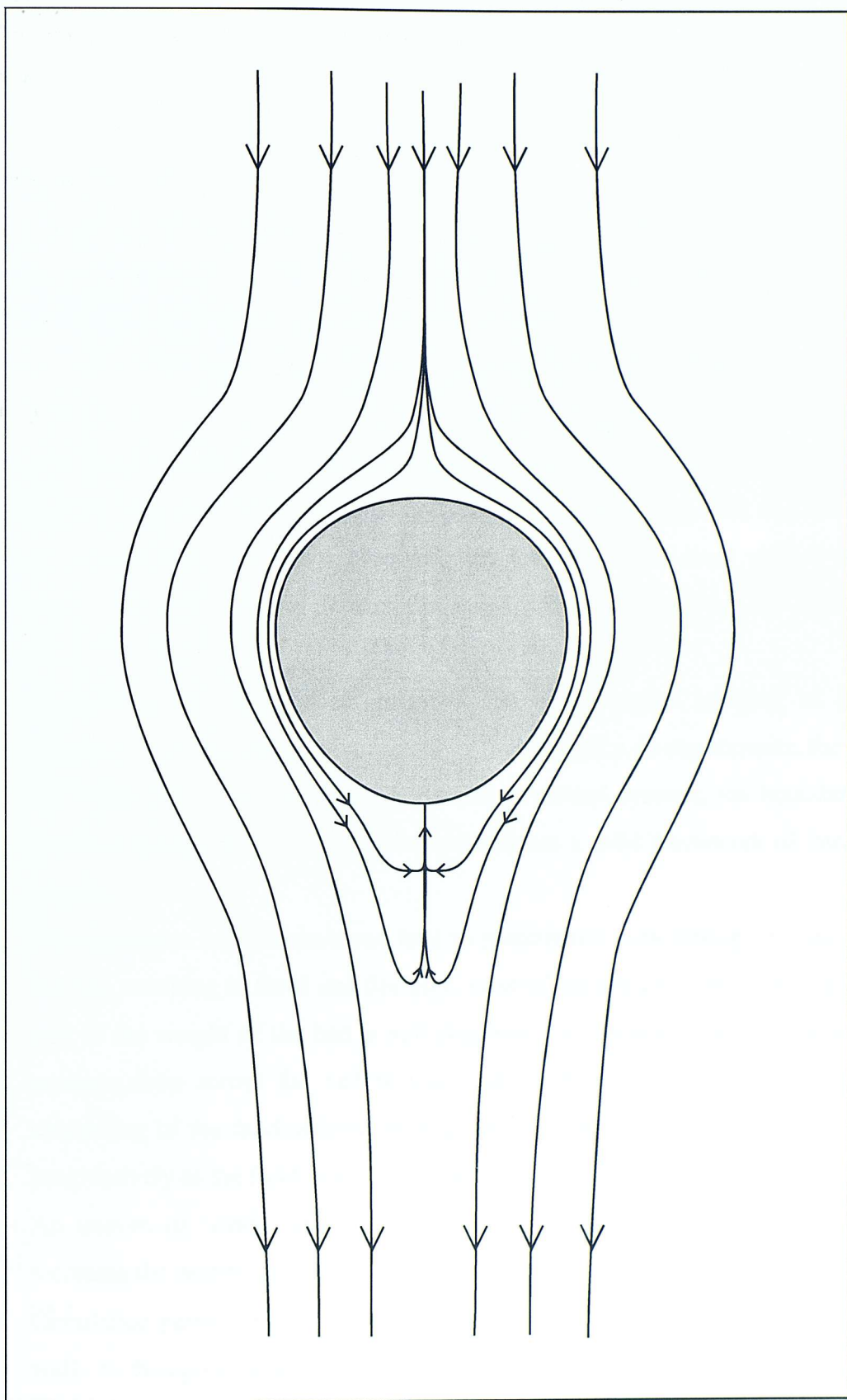


Figure 5.3 Flow patterns around an ascending bubble, showing the reversal of flow in the area immediately below the bubble, drawn from Bugg et al, 1998, see also Leeder, 1999, p. 87

system is therefore controlled by the smallest size of particle in the system, a proposal supported by the statement in an unrelated section of Richardson's 1971 work that very fine grained powders fluidise to form a very uniformly distributed suspension, implying that bubbles, if present, are stable only in the size range similar to that of the particles and are therefore far too small to be observed if they occur. However as soon as particles are drawn into the bubble the bubble will be destroyed and therefore particle size range will not be an important factor, only the value of the lower limit of that range. It has been observed in experiments (Nichols, 1995) that fluidisation of analogue models with gas tends to produce pipe structures, whilst fluidisation with a liquid produces an even reorganisation of all the particles in the bed and complete loss of original bed structure.

Ideally, the entire bed will become fluidised at a single fluid velocity, with a predictable expansion and a drop in the frictional pressure exerted between grains, Fig 5.4a. In real situations however this is not observed and the bed is fluidised over a range of velocities, Fig 5.4b. Several factors can cause this deviation from the ideal. These include:

- Interlocking of non-spherical particles can cause partial bridging of the bed, allowing frictional forces to be exerted by the walls, (i.e. in experiments, the walls of the model container, in larger experiments or natural systems, the boundary of the fluidised system, where the granular material has a solid framework of interlocking grains).
- Heterogeneous bed structure can lead to preferential flow through certain areas of the bed, resulting in fixed and fluidised areas of the bed coexisting. In this situation part of the weight of the bed is still supported on the grain-grain contacts and the pressure drop across the bed is less than predicted. This may result in severe channeling of the fluid pathway through the bed, because the channel will open up progressively as the fluid velocity increases.
- An uneven or widely spaced distribution of points of fluid input to the system increases the probability of channelling.
- Circulation patterns may be set up, causing a force to be exerted on the flow by the walls, in the opposing direction to flow.

Material assemblages capable of fluidising uniformly are also very difficult to fluidise initially. The properties conducive to forming a well-fluidised system are:

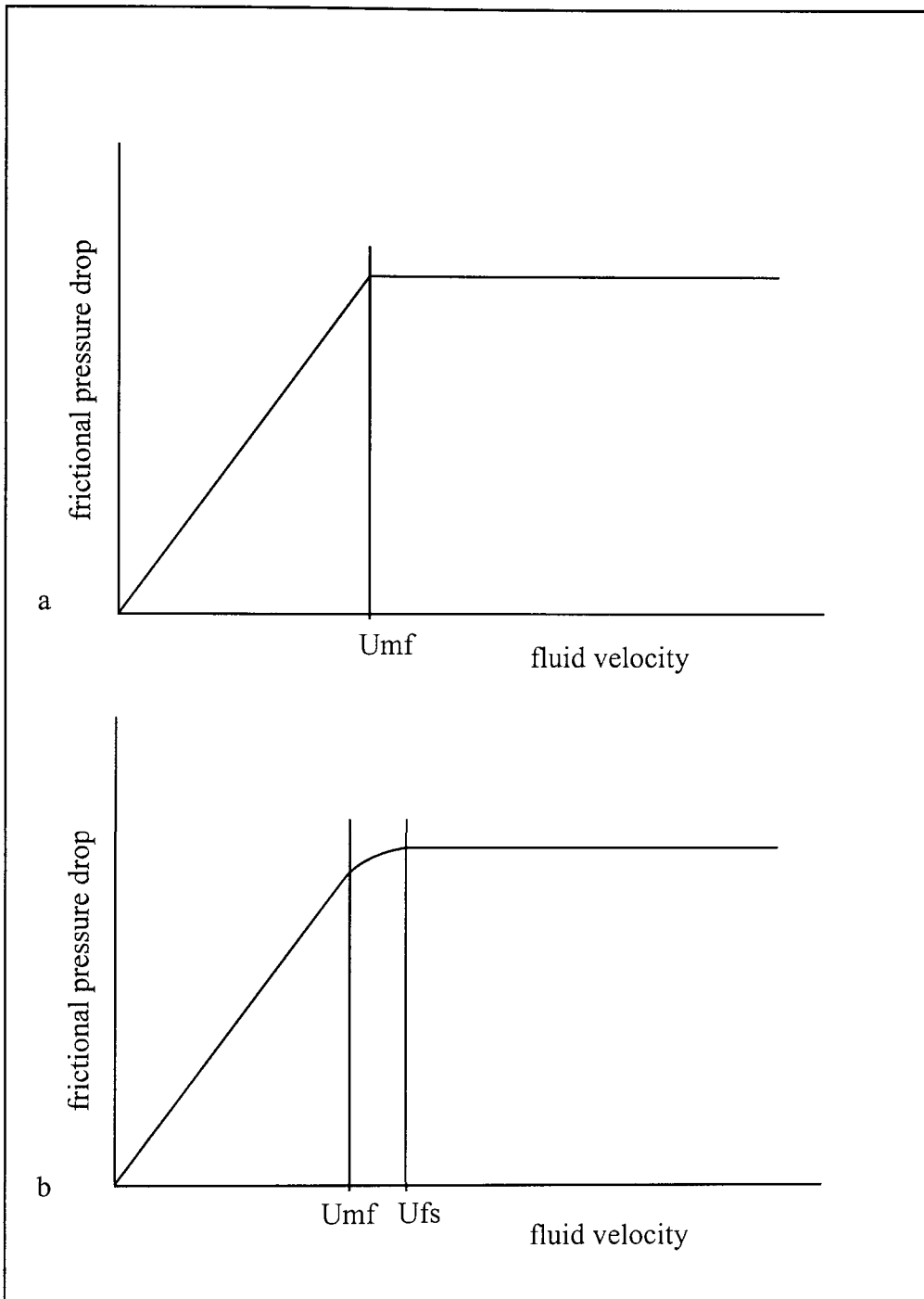


Figure 5.4 Fluidisation velocity curves, after Richardson, 1971

a: Ideal curve with a single fluidisation velocity

b: Real curve with graduated fluidisation velocity

Frictional pressure drop is the change in frictional resistance to shear across the grain boundaries.

U_{mf} is the fluidisation velocity

U_{fs} is the velocity of the fluid when total fluidisation of the layer is achieved.

- low particle/fluid density contrast,
- small particle size and size range,
- particle shape approaching spherical.

In particular, the low density contrast and small particle size makes the initial fluidisation difficult. This is because the smaller the particle size, the larger the surface area per unit volume of particle and the greater the influence of surface forces. Also the lower the density, the less the influence of gravity will aid in separating the particles. Thus the presence of small, low density particles in the bed undergoing flow leads to the occurrence of channelling, though if this can be prevented uniform fluidisation can be induced very effectively.

In real systems, as the fluid velocity exceeds U_{mf} (the minimum fluidisation velocity), pockets of fluidised material develop. Over a fairly small velocity range the majority of the bed is fluidised (U_{mf} - U_{fs} , Fig. 5.4b). However pockets of fixed material may persist until very high velocities are reached.

5.2:1b Effects of variation in particle properties

Liquid-solid systems generally exhibit fairly uniform expansion with increasing velocity, but where there is a spread of particle properties (i.e. size, shape, density etc.) there is a marked tendency for flow segregation to develop, as mentioned in the previous section. In fine grained systems there is a tendency for the flow to segregate into channels. The increase in voidage during channelisation is much less than would be associated with complete fluidisation. The channels represent low resistance paths for the movement of fluid through the system. The ratio of resistance to flow in the channel to that in the bed is very high and fluid flow is therefore very localised. Richardson (1971) argues that because the fluid velocity is high the fluid spends a shorter period of time in the system and it cannot expand all of the bed as a unit, only the channel. This effect is far less significant for coarse grained systems as fine grained systems are more cohesive and more readily form fractures which channelise the flow.

Where a variation in grain size exists within the system there is a very marked tendency for the particles to segregate during fluidisation according to size. In experiments performed with a range of grain sizes 0.05 - 0.1 mm (Richardson, 1971) the segregation was total, resulting in the concentration of larger grains and lower voidage at the base of the beds and smaller grains and higher voidage at the top. The tendency is reduced when

- the range of particle size is reduced,
- the particle/fluid density difference is reduced and/or
- the voidage decreased.

In the fluidisation of fine powders with a gas, it was found that the initiation of fluidisation was difficult (Richardson, 1971) but once mobilised, they were capable of expanding over a very large range before bubbling occurred. These systems are capable of forming stable channels such that the majority of the bed remains undisturbed. However it was found that agitation of the powder using a mixer pushed particles into the channels mechanically and destroyed them (Richardson, 1971). With the channels destroyed the bed expands to give a well developed particulate fluidised system. Fine silica powders in particular are capable of very large amounts of expansion.

The mechanisms of particle mixing in fluidised beds are complex. In experiments (Richardson, 1971) diffusion rates of fluid particles through the fluid phase were measured up to 1cm/sec. However it was found the radial mixing occurred to a much lesser extent than axial mixing (with respect to the flow direction). Similarly the experiments showed that the mean kinetic energy of the solid particles expended in the axial direction was twice that in the radial direction. A bulk circulation was found to be imposed on the random motion of individual particles, particularly if the fluid input was irregularly distributed. This resulted in mass movement of material upwards in the centre (with the flow) of the bed and downwards (against the flow) at the edges/walls of the system.

Liquid-solid systems are noted for a tendency to fluidise readily (Richardson, 1971) with a simple direct relationship between the resultant voidage of the bed and the velocity of the fluid causing fluidisation.

5.2:2 Liquefaction

5.2:2a Liquefaction

The imposition of a cyclic shear stress such as seismic shocks or continuous vibration causes the grains within a bed to become agitated, Fig. 5.5a. They collide with their neighbours and support their weight by the energy transfer of the grain-grain collisions, Fig. 5.5b. This results in the grains becoming temporarily suspended and the sediment loses shear strength, Fig. 5.5c.

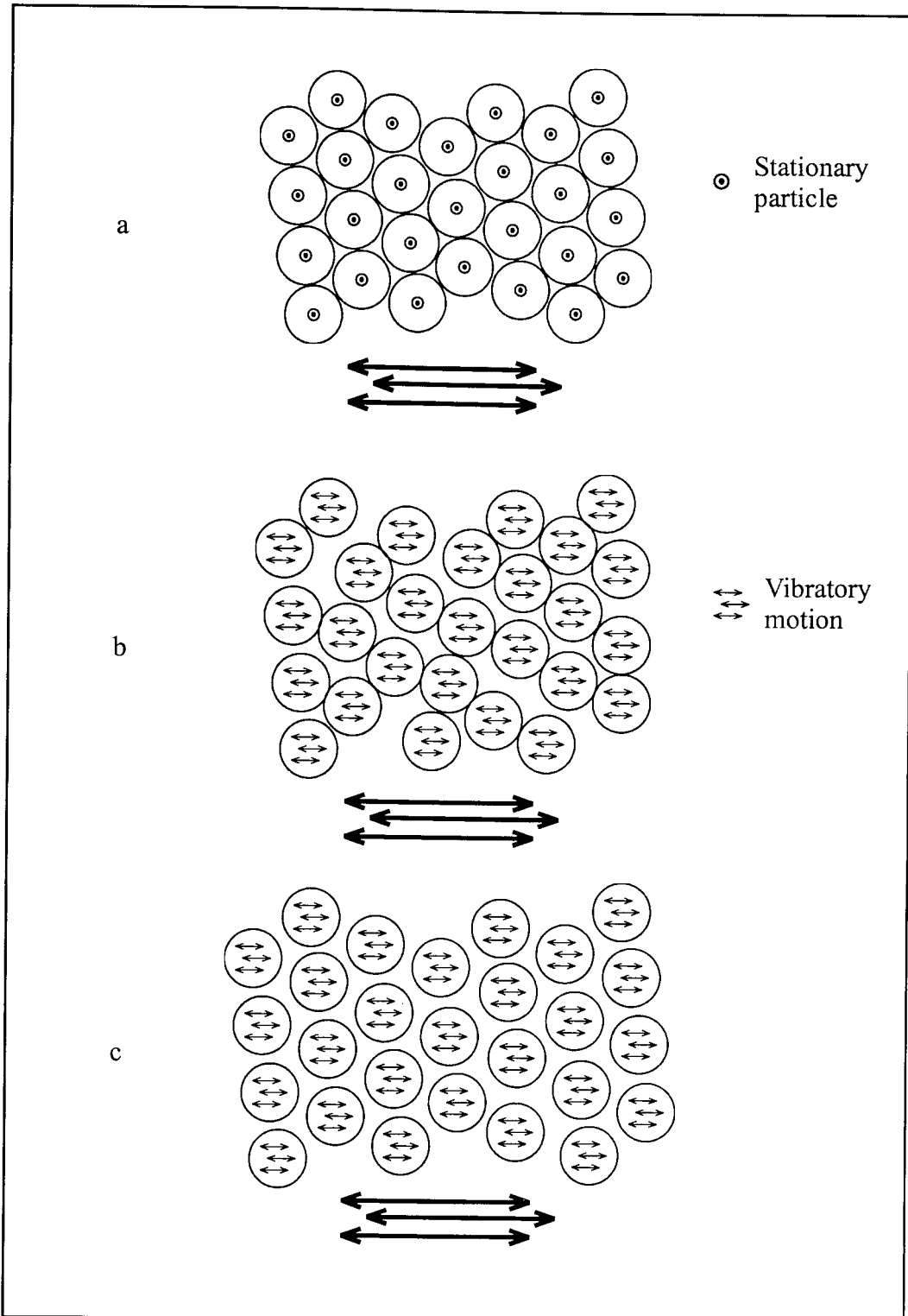


Figure 5.5 Liquefaction due to vibration

a: fixed bed

b: partial relief of intergranular forces by transfer of energy through collision.

c: total relief of intergranular forces by transfer of energy through collision.

5.2:2b Shear liquefaction

Shear liquefaction differs from simple liquefaction in that unidirectional shear stress is imposed, Fig 5.6a, rather than a bi-directional vibratory one. The applied shear stress causes the grains to collide, Fig. 5.6b, and kinetic energy transfer supports the weight of the grains as described in the previous section. If sufficient shear stress is imposed for grain-grain collisions to cause particle suspension the system will become liquefied and will fail, Fig 5.6c.

5.3 Rheology of fluidised beds

The rheological behaviour of a system is defined by the inter-relationship of inelastic deformation (flow) and the force causing it.

When subjected to a shear stress a body will resist therefore creating a shear stress in the body. In a fluid this resistance is called the viscosity of the fluid.

Fluidised systems also demonstrate a resistance to shear stress, as demonstrated by Schugerl, 1971, which is analogous to the viscosity of a liquid and can be regarded as a type of internal friction which operates between adjacent particles. It is the net result of the electrostatic forces, Van der Waals forces and capillary forces caused by residual water between particles. There is also an element of dynamic interparticulate forces which come into play only if the particles are moving relative to one another.

This apparently viscous behaviour was quantified by Schugerl (1971). Under the action of a shear couple a shear stress develops in a fluidised bed. Schugerl's experiments measured the shear stress generated within the bed and equated it to a measure of the apparent viscosity. True viscosity is measured by plotting the rate at which the body accumulates strain against the shear stress imposed upon the body.

However this approach and the experimental technique described by Schugerl (1971) leads to terminology which is confusing. The experiments were conducted in a Couette apparatus, whereby a shear stress is applied to the fluidised bed by means of a rotating cylinder in the centre of a circular apparatus. The torque applied to the rotating cylinder is directly proportional to the applied shear stress. For a particular velocity (rotation rate) the shear stress generated in the fluidised bed will be directly related to its 'viscosity', i.e. the apparent viscosity is measured as a shear stress within the bed.

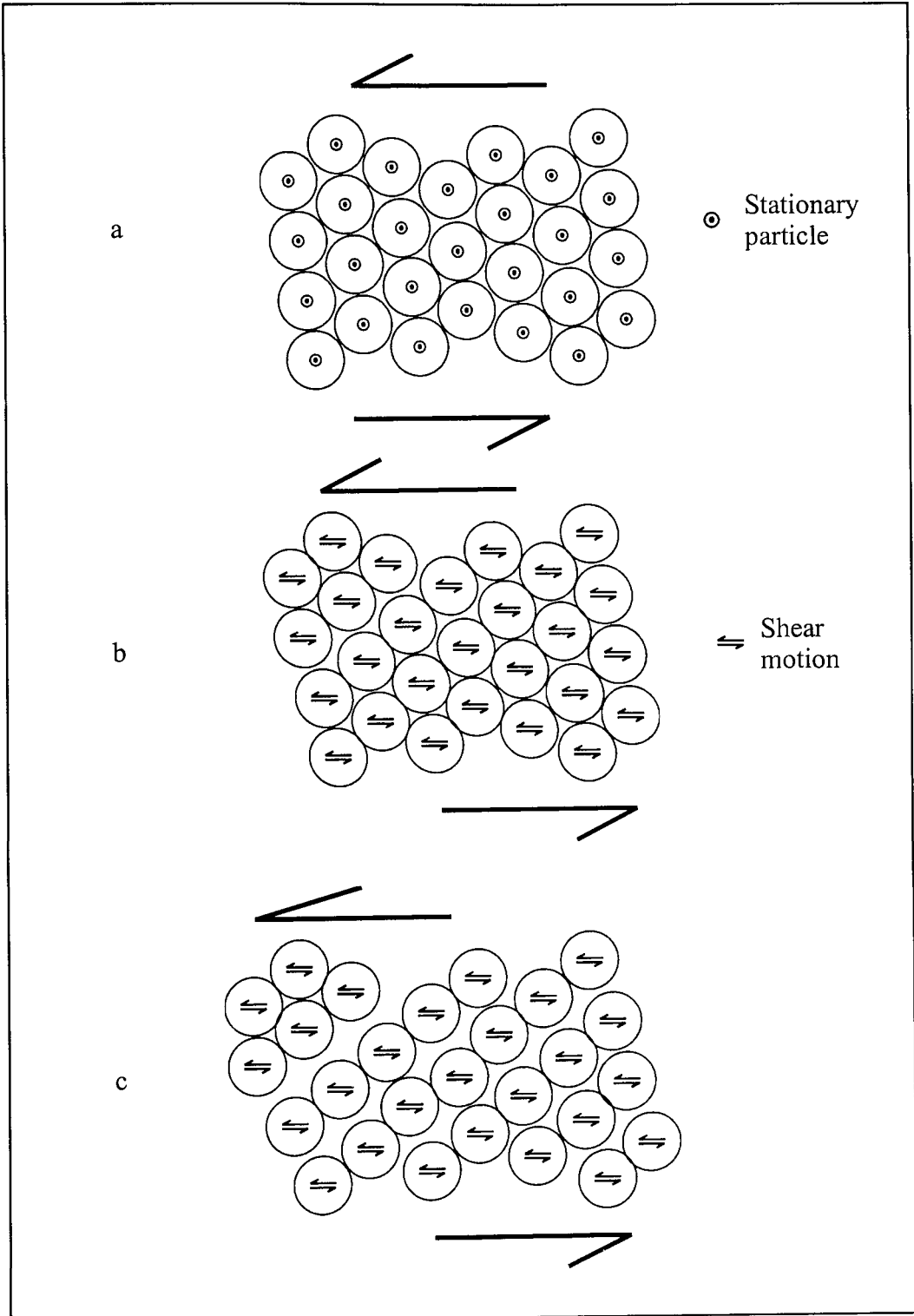


Figure 5.6 Liquefaction due to shear

a: fixed bed

b: partial relief of intergranular forces by transfer of energy through collision.

c: total relief of intergranular forces by transfer of energy through collision.

Accordingly the terms normally used when describing viscosity, shear stress and strain rate, are replaced, in Schugerls experiments, by torque in place of shear stress and shear stress (i.e. the shear stress generated within the bed) in place of the strain rate. This is an unusual and confusing description for the geologist and therefore, in the present study, the terms applied shear stress and generated shear stress ('viscosity') are used in place of torque and shear stress respectively.

In Schugerls experiments, at a given magnitude of applied shear stress (the flow limit) the particles will start to move in response to the applied shear stress and the bed will start to accommodate shear strain. At this point the bed has an apparent viscosity. This 'viscosity' is extremely sensitive to the velocity of the fluid causing fluidisation of the

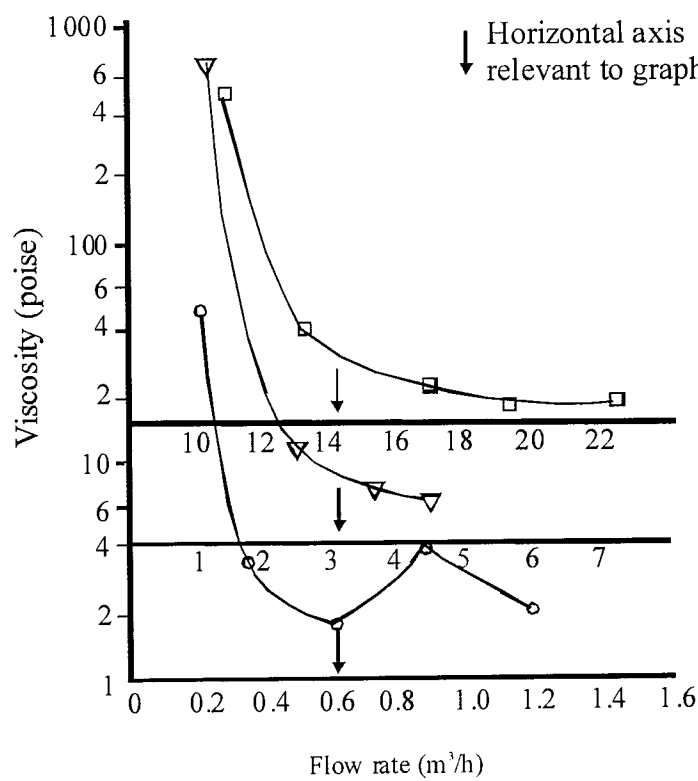


Figure 5.7 Viscosity variation with increasing flow rate at constant applied shear for

- sharp edged heavy quartz particles 450 µm
- ▽ sharp edged heavy quartz particles 105 µm
- smooth heavy glass beads 50 µm

after Schugerl, 1971

bed and Schugerl found that under a constant applied shear stress, the apparent viscosity decreases as the velocity of the fluid causing fluidisation increases, Fig. 5.7. The

apparent viscosity is therefore partially determined by the velocity of the fluidising liquid.

- The extent of the region undergoing shear strain;
when the bed first reaches the flow limit the region accommodating shear strain is restricted to the area immediately adjacent to the central cylinder. If the flow velocity of the fluid causing fluidisation is increased the extent of the region undergoing shear strain widens.
- The rate at which shear strain ceases;
in semi-fluidised beds the shear strain may take hours to cease when the applied shear stress is removed, whereas in well fluidised beds, shear strain will be accommodated very rapidly and the bed will cease to acquire strain when the applied shear stress is removed.

Other features determined by the flow velocity are:

A fluidised bed does not exhibit a linear relationship between the applied shear stress and the generated shear stress. This is analogous to the nonlinear relationship between the shear stress and strain rate demonstrated by Bingham fluids. Schugerl (1971) demonstrated that the rate at which the generated shear stress increases becomes slower as the applied shear stress increases, Fig. 5.8. This implies that the generated shear stress is accommodated by an increase in strain rate and therefore that the 'viscosity' decreases with increasing applied shear.

This change in behaviour is explained by Schugerl as being caused by a change in the orientation of the particles, creating a different bed structure, i.e. one more tightly packed. This creates only a very slight change in the density of the bed, but significantly diminishes the shear resistance.

During this process the particles within the fluid undergo rotation and translation (Jeffery, 1922). With increasing applied shear stress the grains begin to roll past each other rather than sliding over each other. For a given grain assemblage the value of rolling friction is less than that of sliding friction. As a result increasing the applied shear stress results in a decrease in the interparticular frictional forces, causing a decrease in the resistance to the shear stress, thereby causing a decrease in the apparent viscosity. Once the bed structure has changed and the grains are moving past each other entirely by rolling rather than sliding, the bed deforms in a steady state manner and the apparent viscosity value becomes constant.

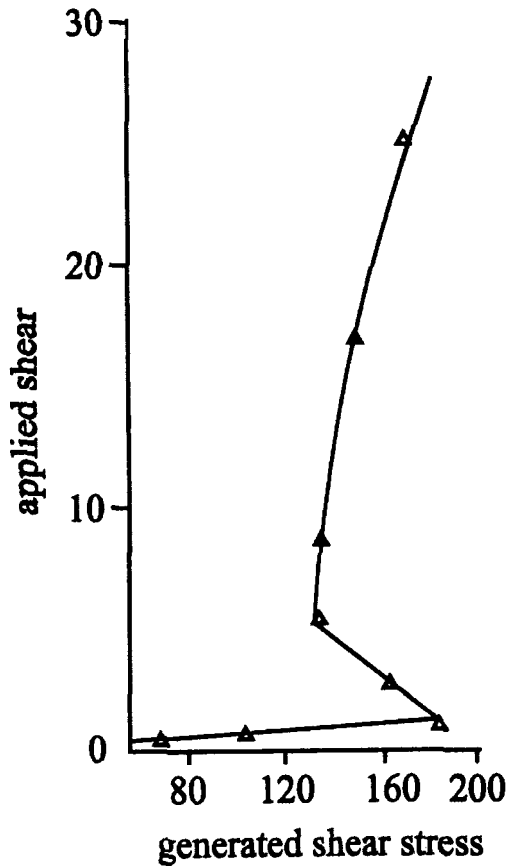


Figure 5.8 Variation in shear stress developed with increasing applied shear force at constant flow rate (9cm/sec), after Schugerl, 1971

Experiments on the dynamic viscosity of fluidised beds (Schugerl, 1971) have shown that smooth glass particles have viscosities in the range 1-5 poise and sharp quartz particles in the range 5-10 poise, Fig. 5.7. At low flow rates the viscosities are high, but as the flow rate increases the viscosity drops rapidly, until it approaches a constant value, Fig. 5.7. For coarse particles (450 microns, Schugerl, 1971) the flow velocities have to be very high before a constant value for the viscosity is achieved. With dense, coarse particles, the properties of the particle surface are not important, as demonstrated by the fact that in the size range 190-275 microns both sharp edged, irregularly shaped particles and smooth spherical particles produce a fluidised bed with a viscosity of 8-9 poise. Although the particle diameter does not seem to affect the viscosity, the density does, such that less dense particles create a fluidised bed of lower viscosity than denser particles (Schugerl, 1971). Particle diameter does affect the viscosity in systems where there is a range of particle sizes. There is a viscosity maximum at critical percentages of fine grained particles, Fig. 5.9. The exact percentage depends on the size of the finer grained element (Haygard and Sacerdote, 1966).

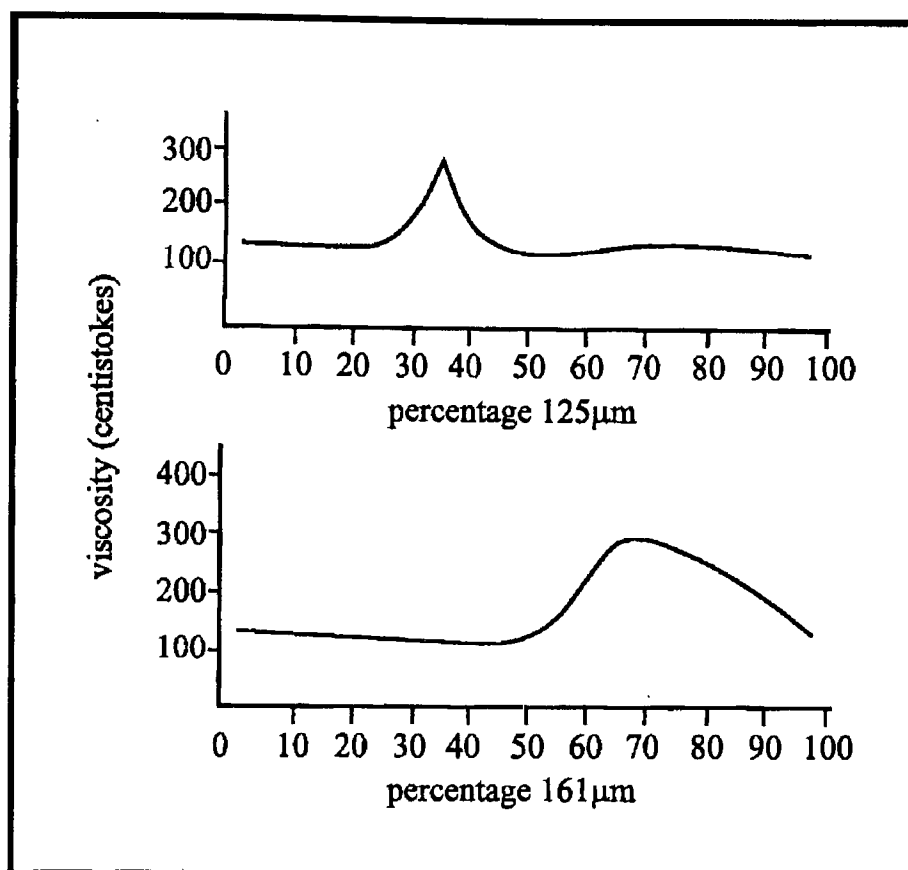


Figure 5.9 Variation in viscosity with percentage of fine grained material, after Haygard and Sacerdote, 1966

5.4 The comparative properties of fluidised beds and liquids

A bed at the point of incipient particulate fluidisation is hydrodynamically very similar to a sedimenting suspension, Fig. 5.10 (Gelperin and Einstein, 1971b). In the former system the particle is stationary (with respect to the system boundaries) within a flowing fluid and in the latter case a particle is falling through a stationary fluid. Both situations are governed by the physics of fluid flow around a solid object regardless of which phase is in motion. Because there is a large degree of interaction between the fluid and solid phases at all times during fluidisation, the flow paths of the fluid and the solid phase are very similar. As a fluidised bed behaves in a very similar way to a single phase fluid system, the behaviour of the particles can therefore be predicted using existing theories of flow behaviour (Gelperin and Einstein, 1971b).

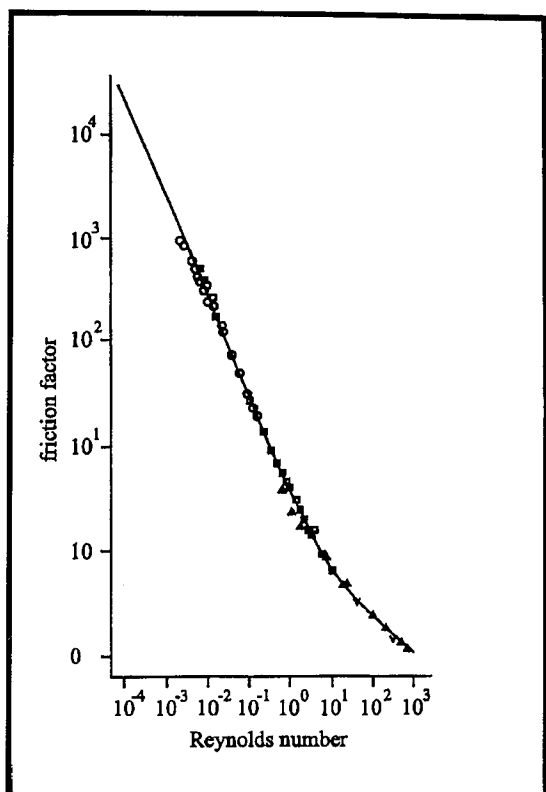


Figure 5.10 Behaviour of particles in

○ a sedimenting suspension

△ a fluidised suspension

after Richardson, 1971

The Reynolds number is a dimensionless number which is the ratio of the forces promoting turbulence, (velocity and distance), to those suppressing turbulence, (viscosity).

It has the formula $\frac{UL}{\nu}$

(where u = mean velocity, L = length, (a measure appropriate to the scale of the system, e.g. flow depth) ν = kinematic viscosity)

5.4:1 Temperature/ Velocity

The temperature of a fluid is closely analogous to the velocity of the fluidising liquid in a fluidised bed (Gelperin and Einstein, 1971) in that with increasing temperature the molecules of a liquid acquire increasing kinetic energy. As the fluid velocity within the bed increases, the kinetic energy of the particles supported by the flow also acquire greater kinetic energy. It follows that similar laws of behaviour will apply for both molecules and particles with increasing temperature and increasing fluid velocity. This means that the fluidisation of a bed is analogous to the melting of a solid body. The volume of the material expands in a similar manner during melting and during fluidisation and the relationship between volume expansion and temperature increase/velocity increase can be described by the same equations.

5.4:2 Viscosity

The viscosity of a liquid falls exponentially with increasing temperature. The viscosity of a fluidised bed behaves similarly with an increase in fluid velocity. Viscosity rises with increased particle size in fluidised systems and increased molecule size in liquids.

5.4:3 Thermal Behaviour

The thermal conductivity of a fluid is dependent on the molecular velocity just as the thermal conductivity of a fluidised bed is dependent on the particle velocity. The heat coefficient of each system increases with velocity/temperature because of the increase in activity of the particles/molecules. Heat transfer is very efficient in fluidised beds and heat is transferred from the fluid to the solid elements very quickly. Vigorous mixing of the particles which is a typical feature of fluidisation ensures that heat is distributed very rapidly. This creates an effectively isothermal bed (Gelperin and Einstein, 1971).

5.4:5 Similarities in Behaviour between Fluidised Beds and Liquids

Certain surface phenomena are demonstrated by both systems (Gelperin and Einstein, 1971). Surface tension is a feature of both as is bubble formation. Waves will appear on the free surface of a fluidised bed as they will on the free surface of a liquid if it is disturbed and if undisturbed the free surface of both fluidised beds and liquids will become horizontal.

5.4:6 Differences in Behaviour between Fluidised Beds and Liquids

In a fluidised bed the boundary effects are much more important than in a liquid and the behaviour of aggregates of particles can become very important at high velocities. Fluidised beds are a great deal more anisotropic than liquids with the directional difference in diffusion, mixing and expenditure of kinetic energy are extremely pronounced (section 5.2:1a).

5.5 Experimental evidence for fluidisation

Nichols et al. (1994) and Nichols (1995) describe a series of experiments designed to assess the conditions under which the three types of fluidisation (namely i) fluidisation

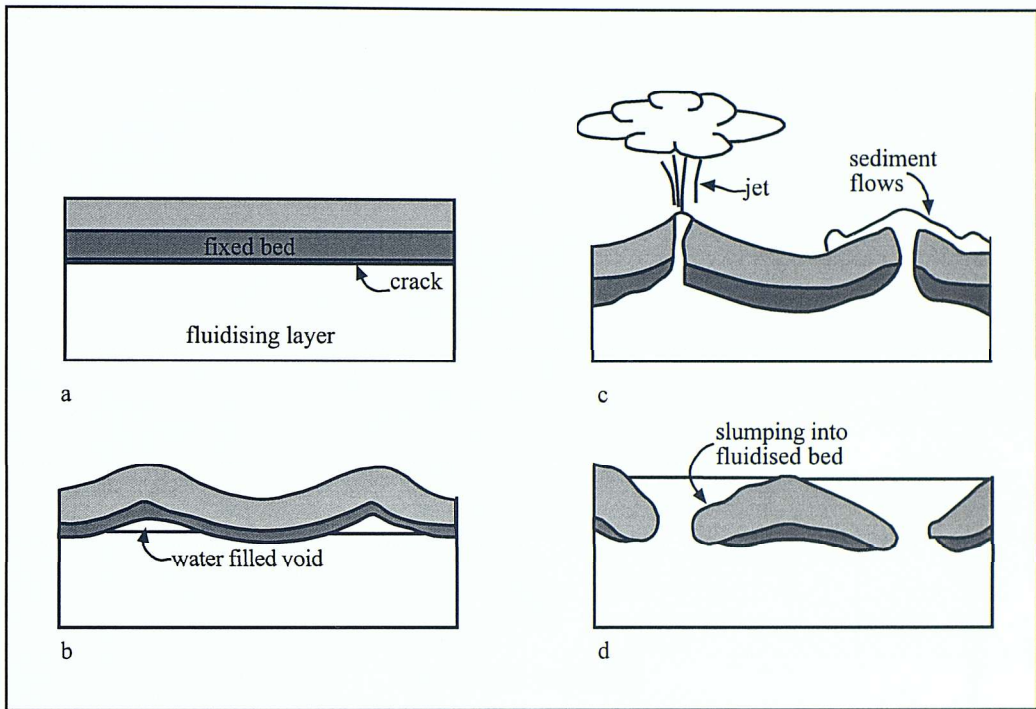


Figure 5.11 Deformation sequence for fluidised sand beds, Nichols et al. 1994

(sensu stricto) ii) liquefaction and iii) shear liquefaction), occur. He established that liquefaction by vibration resulted in cyclic increases in pore fluid pressure. He also argues for the expulsion of fluid after vibration events is the result of closer grain packing occurring when the grains fall out of suspension. In most natural systems where the pore fluid is liquid these two processes (i.e. cyclic increase in pore fluid pressure and closer grain packing) will occur in a complementary fashion.

The models, Fig. 5.11a, were repeatedly observed to split along layer boundaries, causing arching in the uppermost strata, Fig. 5.11b. Void development and axial breakthrough structures were observed beneath and within the antiforms respectively, Fig. 5.11c.

Nichols (1995) noted that the variation in the distribution of deformation during fluidisation depended on the source of the fluidising liquid. If the liquid is derived from within the fluidising bed, then the effects of remobilisation (i.e. void development, arching etc.) should increase in the direction of flow. If the fluid is derived from outside

the bed then the deformation can be uniform across the bed. Nichols et al. (1994) and Nichols (1995) also described certain differences in deformation style which are dependent on the material properties of the bed which is injected by the fluidised material. If it is comprised of loose sand then large cylindrical pipes with splayed ends are developed, Fig. 5.11c. If the material is more cohesive, e.g. clays, then the pipes are smaller and tend to develop very angular bends, Fig 5.12. Nichols et al. (1994) regard

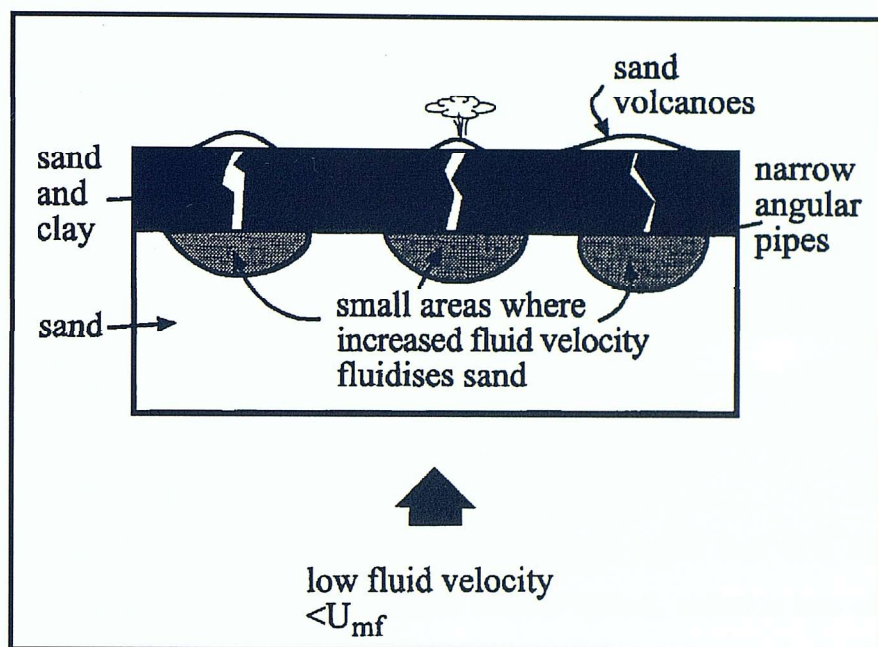


Figure 5.12 Fluid escape structures in sand and cohesive layers after Nichols et al. 1994

fluid escape structures as providing a record of pore fluid movement during depositional dewatering and early diagenesis and note that they also form permeable paths through otherwise impermeable sediments. The fluidised layer behaves as a fluid (section 5.4:5) with an initially planar surface that becomes increasingly undulose as the fluid velocity increases. They also comment on the tendency of a fluidised layer to settle into a tighter packing configuration. Areas that have been repacked will be more difficult to refluidise and so the fluids will remobilise a different section of the bed. When cohesive layers were introduced into the models, it was found that they broke up and sank through the fluidised layer. Small coherent fragments would circulate and became evenly distributed across the fluidised section. The fragments, both large and small remained intact over long periods of flow, although they did become more rounded as a result of erosion. The amount of fluid necessary to create fluidised structures was estimated. Given the modelling conditions of grain sizes in the range 74 - 5000 μm and fluid velocities of

0.3mm/s, it was found that it took a total fluid volume equivalent to a 20mm thick layer of water across the surface of the model 93mm thick to cause remobilisation and the formation of fluid escape structures. They claim that this would be equivalent to a 5% reduction of porosity during consolidation in a sandstone layer 1m thick generating enough excess fluid to cause fluid escape structures, though they don't say whether the sandstone layer is 1m thick at the beginning or end of consolidation.

McCallum (1985) carried out a series of experiments to evaluate the evidence for fluidisation processes in phreato-magmatic breccia pipe formation. Breccia pipes contain well rounded clasts and well developed segregational textures, with upwards and downwards vertical transport of material documented. These textures are difficult to explain using conventional magmatic or phreatic processes and a fluidisation mechanism with supporting experimental evidence is proposed by McCallum. He suggests that fluid build up could generate hydrofracture breccias, which with further increase in fluid pressure could fluidise and become reorganised. This model is supported by the occurrence of composite pipes where a layer of crackle (hydrofracture) breccia surrounds a pipe of rounded and rotated fragments.

In summarising the process of fluidisation McCallum notes that if a liquid is the fluidising agent then particle motion tends to be less violent, abrasion less effective and elutriation (the process of winnowing out the finer grained particulate element of the system) more effective. He defines a system which is not open at surface as being in a state of 'lean phase' fluidisation, forming a dispersed suspension. He describes this state as being extremely mobile. There are a number of processes which he identifies as potentially being very important in breccia modification. The first is channelling, the ability of a fluid to pass through a system without disturbing the surrounding particles. It is possible for the flow paths to either partially or completely traverse the fixed bed. As noted in section 5.2:1b, this is a phenomenon favoured by fine grained systems. Spouting is a circulation process involving fluid movement through a high speed zone. Clasts are dislodged, which fall and move into the fluid stream, where they are transported. This is a process favoured by coarse grained systems. Many geological systems will not have a free surface, which is an integral feature of the models. However McCallum holds that pressure release and up-arching of the overlying rocks are reasonable approximation of the modelled situation which has a free upper surface.

He described a number of breccia features in geological situations that any proposed model must explain. For example, he found that

- the rounded mineral fragments which made up the breccia clasts often demonstrated a high degree of polishing, pitting and striation, indicating that they had undergone physical erosion.
- there is commonly a core/margin segregation. This usually takes the form of smaller, rounder and better sorted fragments concentrating in the centre, whilst the margins are more crowded with clast material and the clasts are larger and more angular. However in some geological examples it was found that the clasts are larger towards the centre of the pipe.
- the matrix may comprise up to 70% of the rock.

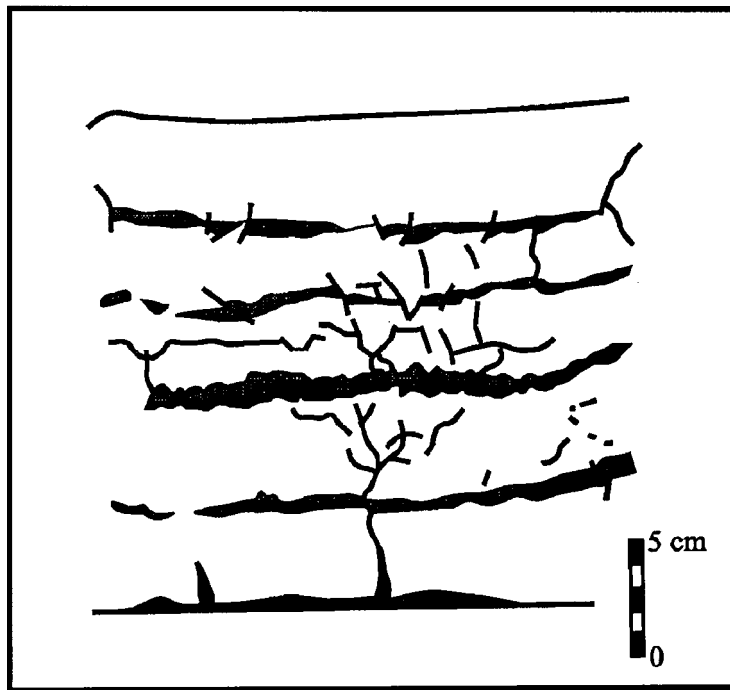


Figure 5.13 Fracturing prior to fluidisation in cohesive granular materials, after McCallum, 1985

His experimental work on fluidisation of loose granular materials showed that when cohesive effects are strong a complex branching fracture system develops, Fig. 5.13, which he tentatively equates to the crackle breccia texture. In low cohesion material the transition from fixed bed to the fluidised state occurs very rapidly. There is commonly a fluid filled bubble at the top of the system and concurrent doming of the material above, Fig. 5.14. Cavity or subsurface void development was particularly well developed under the more cohesive layer.

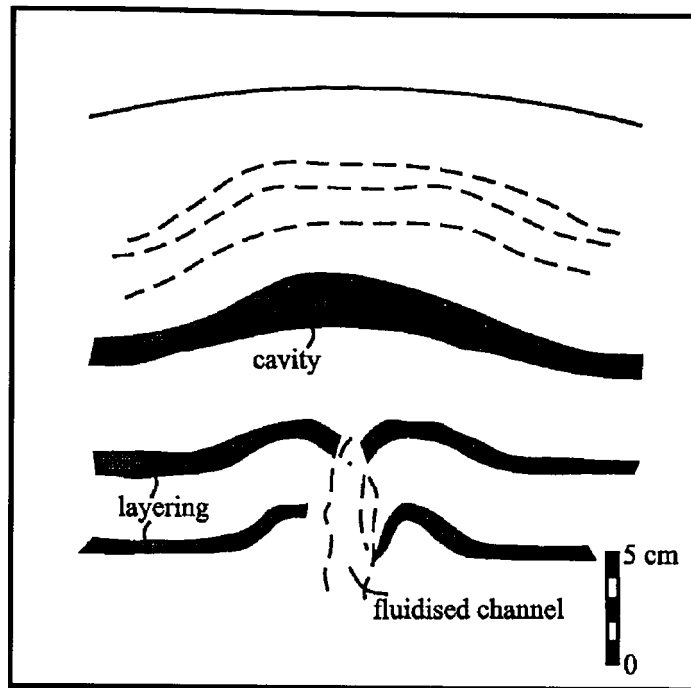


Figure 5.14 Doming of layers and cavity development above a fluidised channel, after McCallum, 1985

Differences in permeability produced different and reproducible effects. High permeability material has the effect of redistributing the fluid flow such that from a single fluid input there was lateral fluid flow and a far wider zone of the model was fluidised than in experiments on low permeability materials. Void or gas cavities frequently form at permeability contrast interfaces.

The models demonstrate that spouting can be a very important process, causing extremely thorough mixing of the clast population and significant rounding of the clasts, as also noted by Nichols (1995). This well mixed material was also forced out into finer fractures, a phenomenon also noted in field examples by Stel and Lankreyer, 1994 (next chapter). Continued flow was seen to cause enlargement of the fluidised conduit.

Annular segregation was a prominent feature of the McCallum models. He found that any system containing a variation in the clast size will exhibit some segregation characteristics. The usual form in nature is that the coarser clasts are on the outside and the finer ones in the middle and this was borne out by the majority of the experiments. However if the experiment was only fluidised for a very short period of time the reverse sense of grading was developed. McCallum explained this phenomenon by proposing that the finer particles were more effectively removed from the centre of the pipe, where the fastest velocities occurred, to the outside, where they were transported in slower moving fluid whilst the coarser clasts were transported in the fastest flowing fluid. This

would provide an initial fining out texture. With time or as the fluid velocity in the system drops, the larger clasts are 'sedimented out into the slower moving fluid at the edges of the pipe where they can no longer be transported and are trapped. This displaces the finer clasts into the centre of the pipe where they are still mobile and the direction of fining reverses.

Channelling is also often very prominent in the experimental models and McCallum (1985) found that it could be induced by the "distributor effect" of the fluid passing through a coarse, i.e. highly permeable layer, as predicted from early studies,

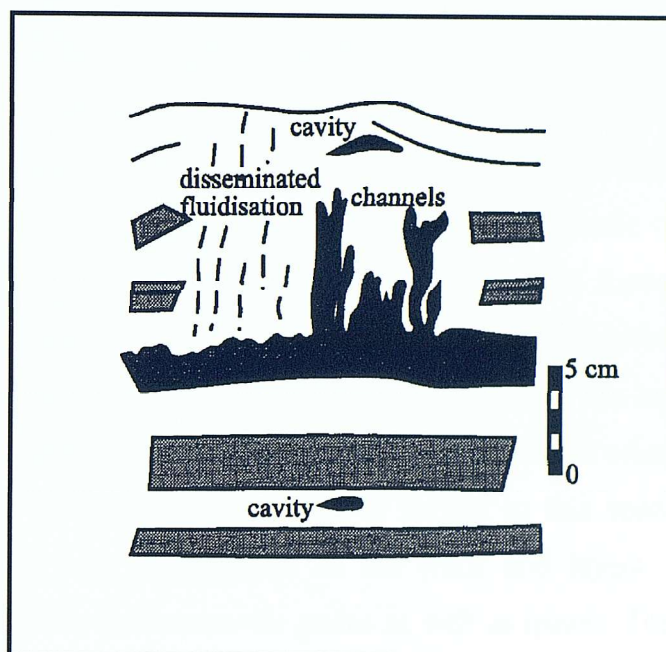


Figure 5.15 Development of channelised flow after fluid dispersion through a coarse grained layer. The coarse layer, having much greater permeability than the material below, allows lateral spread of the fluid. The fluid then forms a series of channels, penetrating upward from this broad base. This has the effect of distributing a single fluid input across the width of the model, hence the 'distributor' effect of coarse layers, after McCallum, 1985

(Richardson, 1971). The fluid is distributed laterally through the coarse layer and migrates upwards along numerous channels by irregular streaming, Fig. 5.15. The finer material is flushed out and redeposited higher up in the system, leaving the channels highly porous.

McCallum concluded that the reorganisational features of many breccia pipes were readily reproducible by fluidisation experiments. The initial form is determined by the cohesive strength of the material. In more cohesive material fracturing occurs prior to fluidisation, Fig. 5.13. The size grading is an intrinsic feature of fluidised systems with

fining out diagnostic of a short-lived system and the reverse grading becomes established as the system develops. The removal of the finer grained element is typical of systems which have undergone fluidisation.

5.6 Field evidence of the occurrence of fluidisation in a geological environment

The most convincing field evidence in support of the contention that fluidisation is a geologically viable process is that provided by clastic dykes, where the processes by which they formed have long been interpreted as flow remobilisation of sediment. Vitanage (1954) presents a description of dykes in the South Platte area of Colorado, USA, where the clastic dykes are spatially so closely associated with a major fault zone that in places the presence of the dykes is taken to be diagnostic of the presence of a fault. The dyke rock is fine to medium grained sandstone and shows colour banding. It occurs in narrow dykes in which wall parallel structures are observed. There are curved, layered structures and there is also a preferred orientation of the long axes of elongate quartz grains, oriented parallel to the layering. The rather weak orientation seen in hand specimen is more apparent when the rock is viewed in thin section. The grains are oriented with their long axis parallel to the walls and layers. This alignment is demonstrated by biotite and muscovite grains as well as quartz. The quartz matrix is a secondary overgrowth of quartz which forms an anhedral mosaic. The cement and pore space comprise no more than 15% of the rock. The implication Vitanage places on these features is that the sandstone was forced into open fractures as indicated by the sharply defined vertical layering, although the igneous/metamorphic nature of the host means that the sandstone must have been derived from above. He cites experimental evidence to support the theory (Jenkins, 1925) where mixtures of powders and clay were squeezed into cracks to form layered sequences, composed of thin sheets of segregated material.

Peterson's 1968 paper contains detailed descriptions of features in sandstone dykes in north western Sacramento, USA, which support a theory of emplacement by remobilisation and flow. The dykes are composed of massive fine to coarse grained sandstone, cemented with calcite or a mixture of calcite and fine grained clastic matrix. The sandstone possesses a preferred orientation of tabular grains, with their long axes parallel to the dyke walls. The percentage of aligned grains varies from a few percent to

near 100%. There is well defined layering oriented parallel to the dyke walls, consisting of bands of different grain sizes. The layer contacts vary from very sharp to gradational. Their degree of development varies not only between dykes but also from place to place within the same dyke. Individual layers commonly exhibit preferred orientation of the grains. Some of the layers are graded, showing a progressive change in grain size. The grading is perpendicular to the dyke wall and they may occur as single or multiple sequences within a layer and may be mono or bi-directionally graded. The dykes occasionally exhibit scours and channel fills. These have their long axes oriented vertical or subvertical, i.e. parallel to the proposed flow direction. This may indicate that sediment input was variable with regard to sediment grain size/viscosity/input velocity. The layers can pinch out laterally and some 'layers' may therefore be elongate lenses. It is also found that a series of long ridges and troughs, with an amplitude of approximately 2mm, occur on the layer surfaces. They too are oriented with their long axes vertical or subvertical. Rarely there are concretion-like structures, elongated into rods and oriented in the same manner as the ridges. Previous work, (Diller, 1889), has reported that in some dykes the margins are finer grained and better indurated than the centres. The generally vertical alignment of elongate structures implies that sediment injection was vertical and not lateral. Peterson (1968) concluded that the structures probably formed by the injection of 'sediment charged fluid' into pre-existing fractures. A combination of frictional drag against the walls, viscosity and velocity variations in the injected material and irregular rates of fracture opening resulted in the formation of the structures described above under conditions of laminar viscous flow.

5.7 Summary

To summarise, the remobilisation of unconsolidated granular materials is possible by the input of a moving fluid or by shaking a fluid saturated granular material. The former system is a fluidised system in the strict sense of the term and because the liquid phase of the bed has a velocity it is capable of relocating and reordering the granular material. In liquid-solid systems the type of fluidisation that usually occurs is particulate fluidisation, a uniform expansion and mobilisation of the bed. The ease of fluidisation and behaviour of the bed once it has been fluidised depend on particle properties such as size range and density and on the velocity of the fluid and level of externally applied

stress. Systems which contain finer grain sizes are more difficult to fluidise than coarser grain systems, but once fluidised, disperse more uniformly and remain more stable. They also tend to develop channelised flow more readily than coarser grained systems. Systems which contain a range of grain sizes will develop grain-size segregation, such that the gradation will initially fine outwards towards the boundaries of the system, but will, with time establish the reverse sense of grading which displays a grain size distribution which coarsens outwards.

On the basis of experimental work and field evidence, it is known that fluidised beds behave in a very similar way to both sedimentary suspensions and pure liquids. The interaction of fluid and particle is hydrodynamically the same for sedimentary suspensions and fluidised beds as both are concerned with the relative motion of a fluid past a solid object. Fluidised systems flow and come to rest in the same way as liquids with comparable viscosity. They also have similar properties to sedimentary suspensions, for example, the ability to use the suspended load to erode and scour, to form channels. In addition fluidised systems can develop organised bed structures such as grading and a preferred orientation of grains.

To summarise fluidisation can mobilise and transport material by forming and filling fractures and channels and is capable of deforming and rupturing constraining layers. The remobilised unit will show structures indicative of flow which are very similar to those seen in sediments, i.e. layering, channel and scour formation, size grading and attrition of suspended particles.

Not only does the fluidisation process have the potential to form the organised features observed in the layers of the Rusey Breccia, but it has been demonstrated by McCallums experimental models (1985) that it has the capability to do so. Field evidence from studies of volcanic brecciation and clastic dyke formation has shown that where fractures form and can be filled with a fluid carrying a suspended granular burden, the structures predicted by theory and produced experimentally can occur in nature. Thus although the process of fluidisation has not previously been intimately connected with large-scale fault activity, it has long been accepted as a viable process in smaller scale fractured systems. Theoretical work (Bhattacharya, 1966, Gay, 1968a, Gay and Jaeger, 1975) predicts that the development of preferred orientations in a geological situation can only occur by suspension of clasts in a liquid, implying that the fluidisation process must be involved during the development of preferred orientation of clasts. There is a great similarity between structures described for clastic dykes and vent breccias (both

formed by fluidisation, Jenkins, 1925, Vitanage, 1954, Peterson, 1968, McCallum, 1985) and the grading and orientation textures seen at Rusey. This similarity and the theoretical work cited above leads the present author to propose that fluidisation was the mechanism responsible for the reorganisation of the breccia. The geological environments showing evidence of fluidisation are characterised by high pore fluid pressure and fracture formation. Such an environment is proposed (Chapter 3) for the initial brecciation episode at Rusey. This proposal is based on photoelastic modelling and an interpretation of the stress regime responsible for the dilational fracturing and mineralisation seen in the field and it is argued that conditions capable of producing fluidisation events (i.e. abundant fluid, high fluid pressures, active deformation and an uncemented granular assemblage) were present at Rusey during the very early stages of compressive deformation and thrust development and that fluidisation is a viable mechanism for the reordering of the breccia.

Chapter Six

Literature review of gel formation and behaviour

6.1 Introduction

As stated in the introduction to Chapter 5 one of the main problems in explaining the Rusey Breccia is to provide a geologically feasible mechanism or process capable of generating the textures it displays. In addition to forming the various organisational textures seen in the rocks at Rusey the process must be capable of producing a matrix supported breccia. This process must involve a fluid capable of supporting in isolation a breccia clast whilst permitting euhedral quartz to be crystallised around it in all directions and it is suggested that the ability of gels and colloids to transform from very low viscosity solutions or separate solid and liquid phases to a highly viscous fluid or a solid material could make them potentially a very significant factor in the evolution of the breccia described. The fluidisation mechanism described in the previous chapter relies on a low viscosity fluid being present in order to be effective and the rotation of the clasts into a preferred orientation will be more efficient the lower the viscosity of the fluid by which they are mobilised. However such a low viscosity fluid would be incapable of providing support for clasts up to several centimetres long for the period of time necessary for the observed quartz coronas to grow. A much higher viscosity fluid would be required to prevent the gravitational settling of the clasts whilst the radial coronas of euhedral quartz developed in all directions around them. Because the formation of a gel or colloid provides a mechanism for holding the clasts in suspension whilst crystallisation of the matrix occurs the two processes are considered in this chapter with a view to determine if either is likely to occur in nature. This chapter ends with a description of some of the field evidence for the involvement of silica gels in fault and vein mineralisation.

6.2 Gel and Colloid Formation

6.2:1 The Formation of Silica Gel

The definitions of gels and colloids are often vague and imprecise and frequently leave materials which can be classed in either category. In this thesis a gel will be regarded as a two phase substance where the solid phase precipitates from solution to form a three dimensional chemically interlinked framework which is highly porous and permeated by the fluid phase. Such a material is similar to a polymer and some workers regard gels as a form of loosely interlinked polymer, (Henisch, 1970). The gelling process for silica gel involves the polymerisation of monosilicic acid solution, Fig. 6.1, and can be triggered in several ways;

- by cooling the solution
- by chemical action, e.g. changing the pH of the solution
- by addition of precipitating agents.

The length of time required for the gelling process to occur is very sensitive to the pH of the solution. Lowering the pH promotes linking of the silica groups with the expulsion of water, Fig 6.1. This elimination of water is called syneresis. There is an apparent time

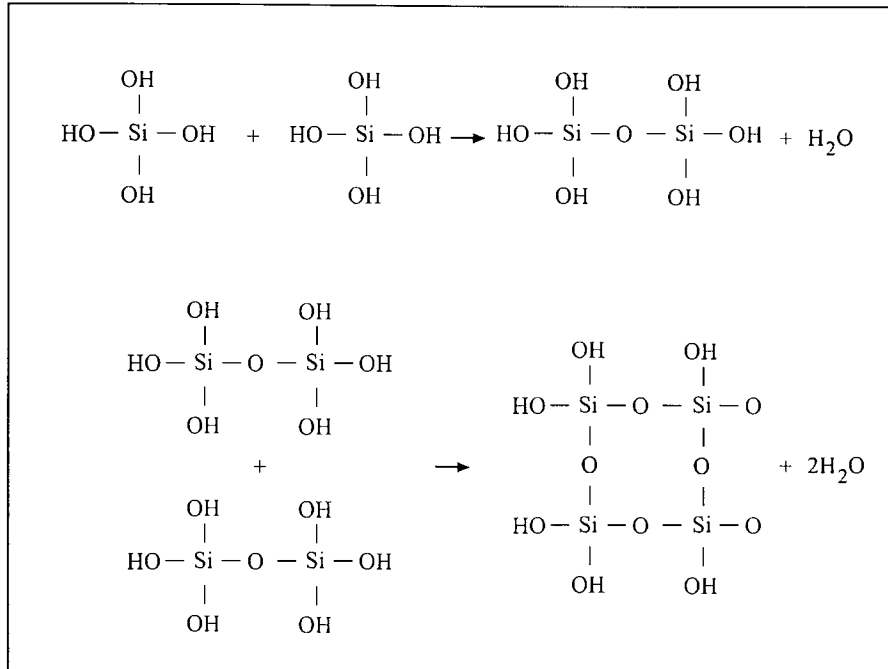


Figure 6.1 Chemical bonding of monosilicic acid units to form a three dimensional network, thereby forming the solid phase of a gel, after Henisch, 1970.

lag between applying the gelling catalyst and the increase in viscosity associated with gelation. Henisch, (1970), explains this by suggesting that initially the polymerisation process forms long chains and it is only with further polymerisation that cross linkages form between these chains, effecting the increase in viscosity. This cross linkage creates the three dimensional network of silica which forms the solid framework of the gel. The mechanical properties of the resulting gel are highly dependent on the concentration of silica in the gel. For example, gels of 1:40 silica:water ratio can be cut with a knife, whereas a 1:10 ratio gives a gel solid enough to be friable. Although Hench, (1970), regards chemical gelation as irreversible he does note that a partial reliquefaction, i.e. rupture of the cross-links, may occur upon mechanical agitation. However, once made, silicon-oxygen bonds are very stable and difficult to sever. Drying has only a small effect on the volume of the gel because a rigid network of silica has already been established. Iler, (1955), notes that repeated grinding of quartz will produce a powder which will react with boiling water to form monosilicic acid, the starting point for chemical gel formation. He also notes that crystalline quartz under pressure will react at high temperature (4-500°C) with water to form gelatinous silica.

6.2:2 The Formation of Colloidal Silica

The definition of a colloid used in this thesis is that of a dispersion of particles within a liquid medium in which the particle size is within the colloidal range, the colloidal range being defined as particles up to the maximum size that will not settle out of suspension, (Iler, 1955). The following literature review has been conducted with specific reference to colloidal suspensions of silica.

Liquids which are gel-like in behaviour can be made up of discrete particles rather than units bound by ionic bonds. X-ray scattering response indicates that many 'gels' are made up of spherical particles which can vary in size from 30-60 Ångstrom up to 30 millimicrons. These particles can effectively grow in size by aggregation, Fig. 6.2, and

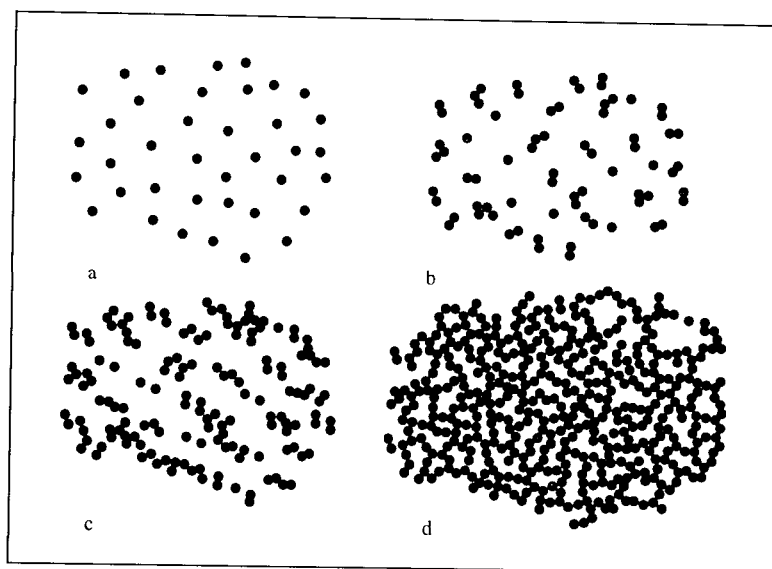


Figure 6.2 Aggregation of dispersed silica particles to form an interlinked three dimensional network, causing a very large increase in the viscosity of the liquid.

cause coagulation of the dispersion, in a similar way to the effects of polymerisation during gelation. Elliston, (1984), defines a colloid as a liquid containing 'matter in a finely divided state', defining matter as particles or units between 10 Ångstrom and 1 mm, and adds that 'the unique properties of the colloidal state lie in large surface area to volume ratio, such that surface energy becomes an important component of the total energy system'. He describes the various attractive and repulsive forces operating between particles.

In a colloid, surface forces are of great importance, unlike particle suspensions in a fluidised bed, (section 5.3). Elliston, (1984), describes two types of forces as being significant;

- Van der Waals forces
- electronic forces

When the colloid particles are sufficiently numerous, the particle density is such that the particles are close enough to each other for Van der Waal's surface forces to be

effective. These will pull the particles closer together. This may result in a self-densifying or synergetic colloid, (Elliston, 1984), self-densifying in that the solid phase of the colloid achieves a closer packing and synergetic in that this closer packing of the particles expels water. Densification or dispersion of the particles is critically dependent on their concentration. If there is enough water in the mixture it will stay as a liquid colloid. If the mixture has lost water, then the particles will adhere and form aggregates, Fig. 6.2, which will transform the liquid colloid into a much more viscous material and eventually to an effectively solid material as the closer packing between particles expels water.

As a result of the microparticles being similarly charged a repulsive force exists between them which opposes the Van der Waals forces. This repulsion keeps them in a stable dispersion as they approach each other as a result of Brownian motion.

Van der Waals forces between particles is given by the relationship K/L^3 , where K is a material constant for the specific colloid and L is the distance between particles. This relationship means that the Van der Waals forces are inversely proportional to the cube of the distance between particles. The electrostatic forces of repulsion are directly proportional to the square of the distance. The potential energy of the interaction, (V), is

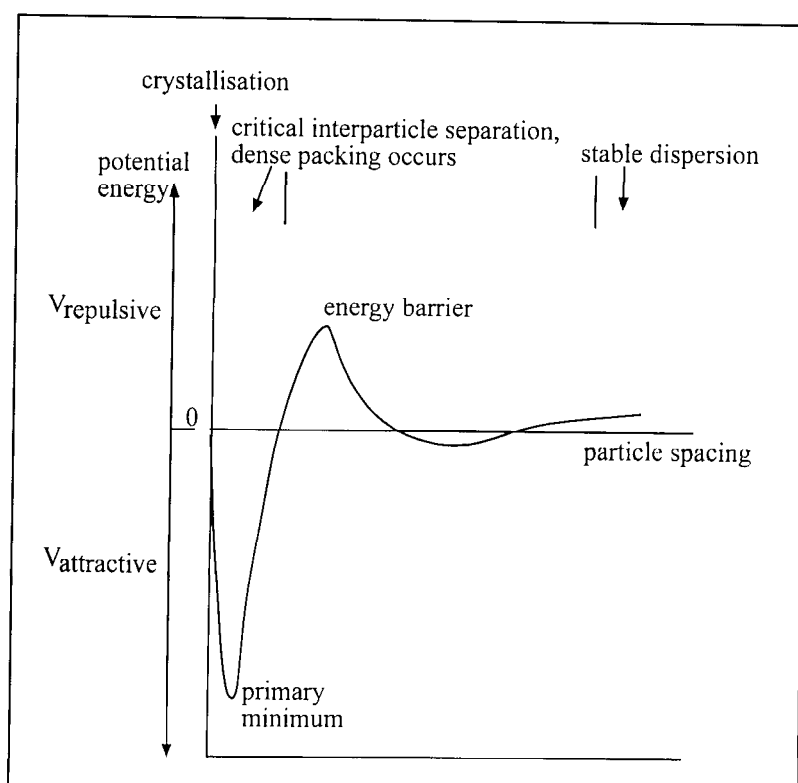


Figure 6.3 Potential energy curve for a colloidal suspension, showing the balance between attractive and repulsive forces. The energy barrier is the level of repulsion that must be overcome before the particles will begin to be attracted to each other. Once overcome the suspension rapidly achieves dense packing and crystallisation, Elliston, 1984

a function of the interparticle distance and is described by $V = V_{el} + V_{vw}$, Fig. 6.3. (V_{el} = potential energy of the electronic forces, V_{vw} = potential energy of Van der Waals forces), When the particles are in a close packed state, V is positive and the system would undergo a very rapid increase in viscosity, as the particles aggregate and become linked clusters forming a loose three dimensional network and coagulating the liquid, Fig. 6.2. Elliston, (1984), also discusses crystallisation from colloids. He points out that the solubility close to a surface may be lower than the average for the system, resulting in precipitation on the surface being preferred to bulk precipitation. This would explain why the presence of particulate contaminants in a gel will always act as nuclei for crystal growth. Also the process of tying the free particle into a crystalline structure is an exothermic reaction as a result of the release of the surface energy i.e. recrystallisation of the silica from its fine ground, suspended, high energy state to an ordered crystalline low energy state releases heat. The formation of strong silicon-oxygen bonds during crystallisation produces and expels water and, as a natural system may have other chemical constituents, any adsorbed ions which may be present. Elliston argues that this would create an electrolyte gradient, increasing towards the crystallising surface, which would impel charged particles/molecules towards the surface, as the expelled ions and the charged particles/molecules would have opposing charges and would therefore attract each other. In a colloidal silica system, this process would supply more silica to the crystallising surface and, in the high ionic concentrations immediately adjacent to the surface, the repulsive forces acting on the particles/macromolecules are further reduced, thus encouraging precipitation to occur. This causes further densification and crystal growth which in turn releases more electrolytes and the process self perpetuates. If the colloid undergoes thixotropic remobilisation this process can re-establish itself after the disruption and will occur whenever the colloid is static. Remobilisation is possible because colloids are thixotropic. i.e. they can undergo an isothermal liquefaction when subjected to mechanical shock, agitation or imposed shear which break the aggregate linkages. This process is repeatable and reversible. The expulsion of water as a result of the densification and precipitation processes keeps the colloid weak and wet and highly susceptible to reliquefaction.

It is apparent then that silica gels and colloids can be formed in a number of ways under conditions likely to be encountered in nature and specifically during the generation of a fault related breccia.

6.3 Crystallisation in gels and colloids

Gels and colloids share a number of properties when used as a medium for growing crystals. In the following section the term gel will be used, but the comments are equally applicable to colloids.

Gels are an extremely stable medium in which crystal growth can take place. They are extremely viscous and convection currents are suppressed, leaving diffusion as the only mechanism for supplying new solute to the crystallising face. The technique of growing crystals within a gel is ideal for producing perfectly formed crystals, a capacity provided by the 'cradling' effect of the gel, which supports the crystal and protects it, yielding as it grows without exerting any major forces upon it. A major role of the gel is also to suppress the nucleation of secondary nuclei, which would compete for the solvent and destroy the perfection of the crystal form. The problems involved in growing large crystals, i.e. greater than a few millimetres, are;

- limiting the nucleation of other crystals,
- keeping the crystal face supplied with solute,
- removing any waste products.

Gels will inhibit secondary nucleation and their highly porous nature will allow diffusion to supply new solute to the crystal face and the removal of waste materials. In the event of the solute supply becoming exhausted it is possible to transplant growing crystals into a new gel without causing any visible break in the crystal growth, although if this is not done carefully inclusions of the exhausted gel will be trapped along the new growth surface, Henisch, (1970).

Crystal growth within a gel is often associated with cusp formation, Fig. 6.4, (Henisch, 1970), where the gel has peeled away from the growing face of the crystal, creating a void. This is caused by pressure from the growing surface displacing the gel. The cusps are filled with solution and neighbouring cusps are connected, such that a growing crystal is almost totally surrounded by solution. As a result of this phenomenon the crystals which originally nucleated in gel grow increasingly from solution. This protects them from the interference of secondary nuclei.

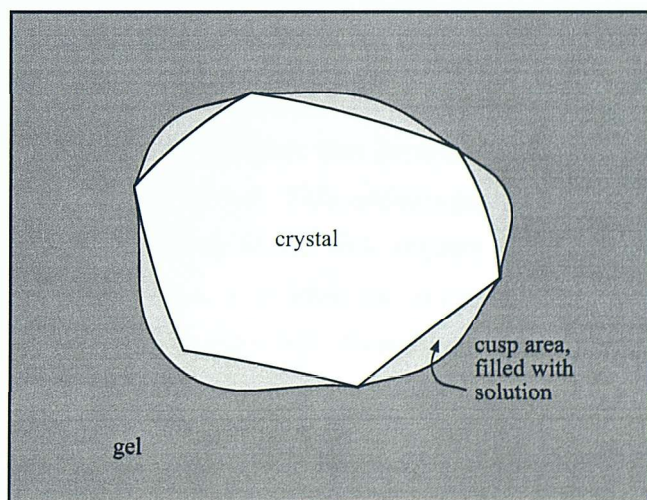


Figure 6.4 Cusp formation around a crystal growing in gel, after Henisch, 1970

Henisch studied crystallisation in gels of various compositions from the point of view of the use of the gel as a support and transport medium. Work done by Oehler, (1976), concentrated on the production of crystalline silica from colloidal silica, effectively consuming the rigid framework as the amorphous silica recrystallised to a more stable state. He used industrial silica colloid at 30.1 weight % in colloidal suspension at pH 10, stabilised by sodium ions, and caused it to coagulate by adding HCl, (reducing the pH to 7), and NaCl. Under this treatment the dispersed colloid became a solid, aggregated colloid in an hour.

Confined in silver capsules to prevent contamination or reaction, the gel was subjected to a pressure of 3kb at temperatures ranging between 100-300°C for 25 to 5,200 hours. At 200-300°C 83% of the sample had recrystallised to quartz after 50 hours and at 165°C crystallisation started after 100 hours and 52% of the sample had recrystallised to quartz after 170 hours, (7 days). Quartz was the only crystalline silica present in any sample which finished crystallising. X-ray diffraction showed some cristobalite present in intermediate stages, but it disappeared with time. The quartz crystallisation was found to be initially very slow followed by a phase of extremely rapid crystallisation when 60% of the quartz phase would be formed, followed by a very rapid decrease in crystallisation rate. When the sample was deliberately contaminated it was found that the contaminants, (algae), provided sites which were conducive to crystallite nucleation. The quartz crystallised as fibrous chalcedonic spherules on the order of 20 μm in diameter. Oehler describes the formation of the spherules as being the result of primary nuclei always giving rise to radially fibrous polycrystalline aggregates rather than single crystals. He claims that this is a result of the crystals branching at small angles from the fibre axis, rather than along preferred crystallographic axes. He identified this type of growth as occurring in multiphase systems in which the co-efficients of diffusion are small compared with crystal growth rates, conditions which are satisfied by impure solutions of high viscosity. Crystal growth in the impure solution takes only the pure material and leaves the impurities behind. The impurities cannot diffuse away and so form an uneven blanket across the surface of the crystal, exaggerating irregularities in the crystal shape as the crystal surfaces that protrude through the blanket are in purer solution and can grow faster, Fig. 6.5. This encourages frequent branching of the crystal and produces a sheaf of radiating fibres. This texture develops when the impermeable layer of impurities is thin. When it is thick the entire crystal is covered and branching occurs along preferred crystallographic directions. The thickness of the layer is determined by the diffusion rate and the crystal growth rate. Spherulite formation can take place even in very pure colloids and gels because polymers of low molecular weight can act as impurities relative to polymers of high molecular weight. It can also occur in colloids and gels of almost infinite viscosity, e.g. devitrified glass, and will occur in any colloid or gel above a threshold viscosity. Once initiated spherulite

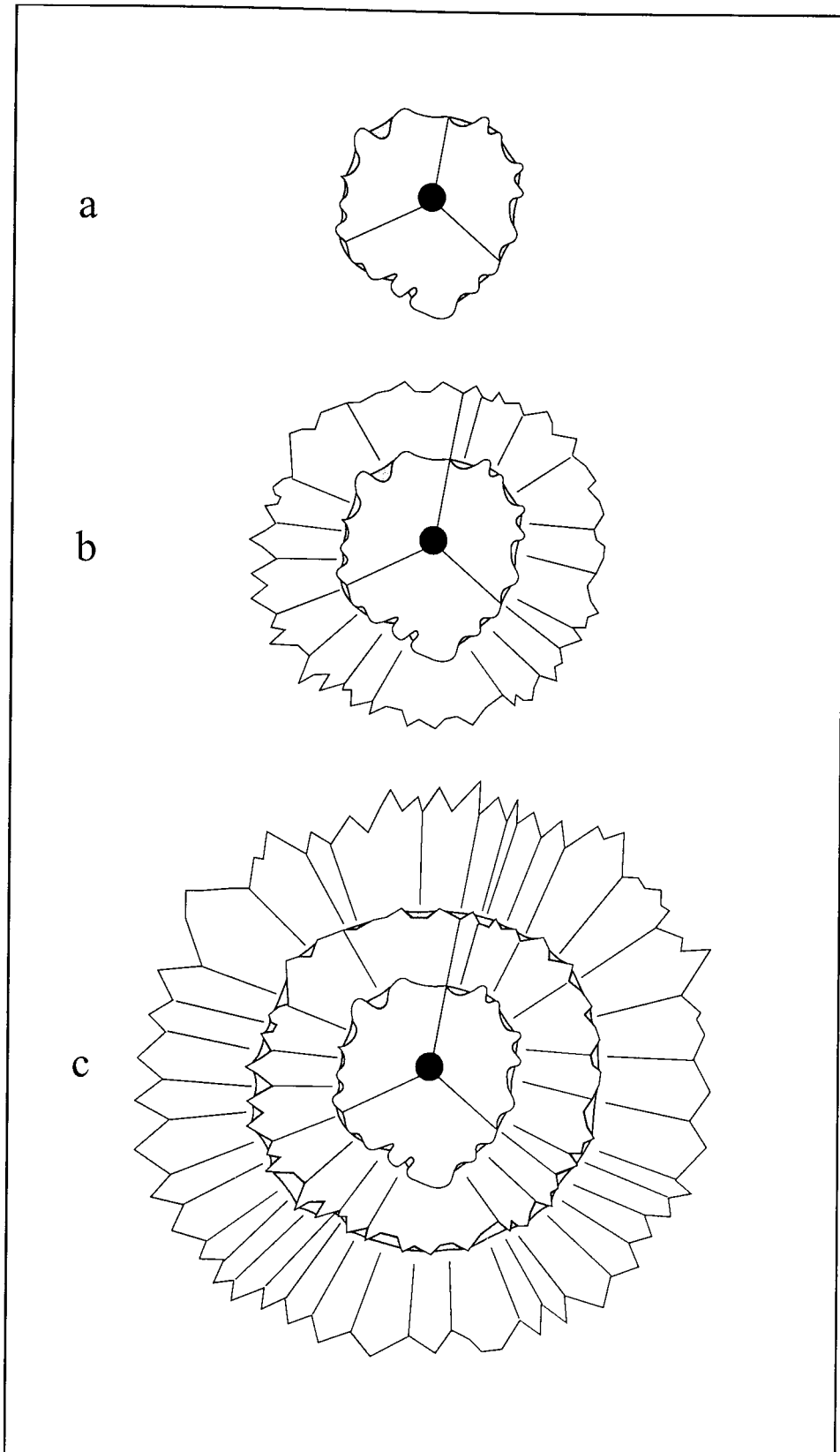


Figure 6.5 Development of sheaves of radially growing quartz fibres
a: initial crystallite forms and develops blanket of impurities
b: crystals divide and surfaces in pure gel grow faster than those covered by impurity blanket
c: further growth and subdivision

formation will continue until all the solid phase is consumed or until it is sufficiently diluted in the water expelled by syneresis during spherulite crystallisation.

In the products of experiments that produced greater than 60% quartz, double ended euhedral quartz crystals were found on the spherulite surfaces. Growth in a low viscosity dilute solution favours euhedral crystal growth. Oehler therefore proposes that the growth curve recorded in his experiments, Fig. 6.6, was governed by an initial sluggish growth of a quartz crystal nuclei, followed by very rapid growth during spherulite formation. This was in turn followed by slow growth when the colloid framework had been recrystallised to

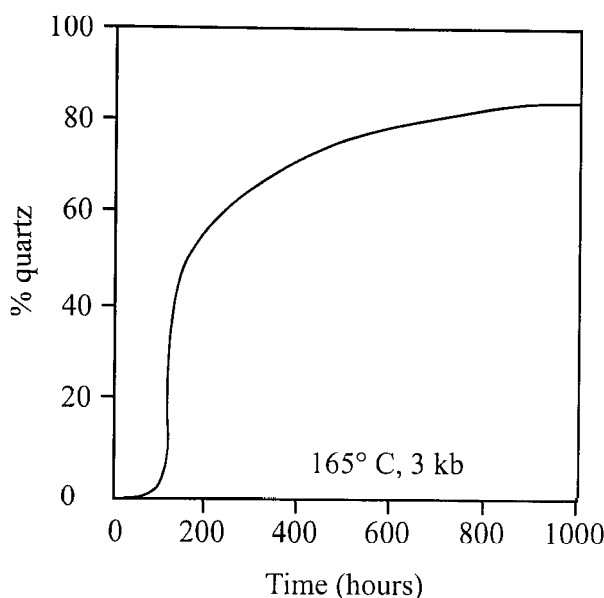


Figure 6.6 Graph of variation through time of the conversion of colloidal silica to quartz, Oehler, 1976

such an extent that the consumption of silica and expulsion of water had produced a suspension too dilute to support radial crystallisation. Euhedral crystals then grew from solution.

More detailed work on the crystallisation paths of the transition from suspended amorphous silica to crystalline quartz has been carried out by Betterman and Liebau, (1975). They used much lower confining pressures than Oehler, (1976), but much higher temperatures and mapped out in detail the transitions that Oehler, (1976), found only a trace of in his experiments. Betterman and Liebau found that at 3kb and above, the transition was from amorphous silica to quartz and that the quartz so formed was always low energy ordered quartz, with 1mm crystal size. However they found that varying the experimental parameters of pressure, temperature and pH had significant and variable effects. With a commercial colloid the effect of temperature variation was very small. With the authors own colloid, prepared from a finer ground silica dust, the effects of both pressure and temperature were much greater, with reaction times always decreasing with increasing temperature. As can be seen from Fig. 6.7 the reaction time for complete

transition from amorphous silica to quartz did not exceed 90 hours and at high temperatures was reduced to 10 hours. Betterman and Liebau defined a series of fields for the reaction paths. At low pressure, (1kb), and temperature, (130°C), the reaction path is amorphous silica to cristobalite to quartz. At 200-300°C and <2kb, the path is amorphous silica to cristobalite to keatite to quartz and a pressure >3kb was necessary before the direct transition of amorphous silica to quartz was recorded. Increasing the OH⁻ concentration of the solution,

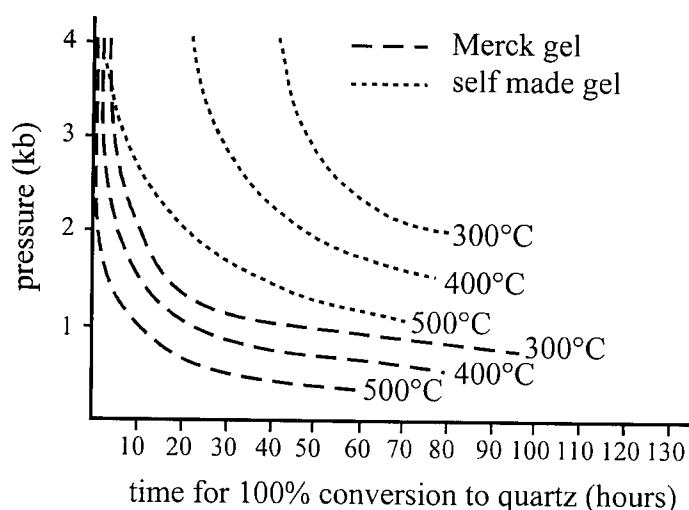


Figure 6.7 Conversion rates of amorphous silica to quartz at different temperatures and pressures, Betterman and Liebau, 1975.

i.e. increasing the pH, results in a transition path of direct transition from amorphous silica to quartz. In all the transition paths a low density silica phase is replaced with one of higher density and with increasing pressure the development of the intermediate density phases are suppressed and the highest density phase, low energy ordered quartz is formed at once. These fields were all identified using the commercial Merck silica colloid. With the other colloids used, differences in the initial amorphous silica produced large differences in the time required for complete transition and smaller differences in the pressures and temperature at which transitions occurred. This dependency of the rates of the various processes on the starting material is linked to the variations in the degree of contamination, differing particle and pore size and differences in structural state. It is clear from the experiments that the pre-crystallisation structure of the 'amorphous' silica is a significant determining factor in the ensuing transition to crystalline quartz.

6.4 Field evidence for the involvement of silica gels in geological processes

Most of the field studies involving silica gels have been carried out in the economic and mineralogical areas of geology. This is because colloidal silica is thought to be related to

the colloidal transport of gold. Hence an understanding of colloid behaviour in natural systems has a sound financial value. Saunder's 1990 paper was concerned with gold transport in epithermal systems. The system he studied was extremely shallow, 50-500m and very cool, approximately 160°C. The deposit is formed of a succession of bands of colloform silica (very finely banded silica) with variable gold content. The barren bands have equigranular to elongate crystals <0.1mm in diameter, grown with interpenetrating boundaries to form a jigsaw texture. There are localised developments of adularia, (KAlSi₃O₈), which resemble the chromatographic separation bands observed in gel separation techniques. The gold rich bands are layers of fine grained electrum, (an aluminium/gold mineral) and silica which conform to the substrate topography, filling in topographic lows in a pseudo-sedimentary style. There are local developments of coarse radial crystal growth, growing outwards into the vein. Also locally developed are textures similar in appearance to soft sediment deformation structures, with fragments of gold-silica bands broken up and redeposited elsewhere. They appear to be the result of gravitational instabilities in a very dense but still viscous precipitate.

The ore is simple in composition consisting largely of electrum, (the gold ore), silica and adularia with very minor amounts of iron sulphides. The textures suggest that the gold was transported attached to colloidal silica, which originally formed an amorphous precipitate. Post-depositional recrystallisation of the silica resulted in a fine grained ore which has preserved the banding features. Saunders suggests that these band features may have formed in two ways. Agglomerates of particles in colloidal suspension may have become too large to remain in suspension and as a result may have settled out and adhered to the vein walls. Alternatively the deposit may have been precipitated along the vein walls which attracted particles from the passing colloid, because of the influence surface forces have on the behaviour of colloidal particles.

Silica colloids with uniform particle size will be less well packed and will have a higher water content than colloids with variable particle size, Fig. 6.8. Saunders proposes that

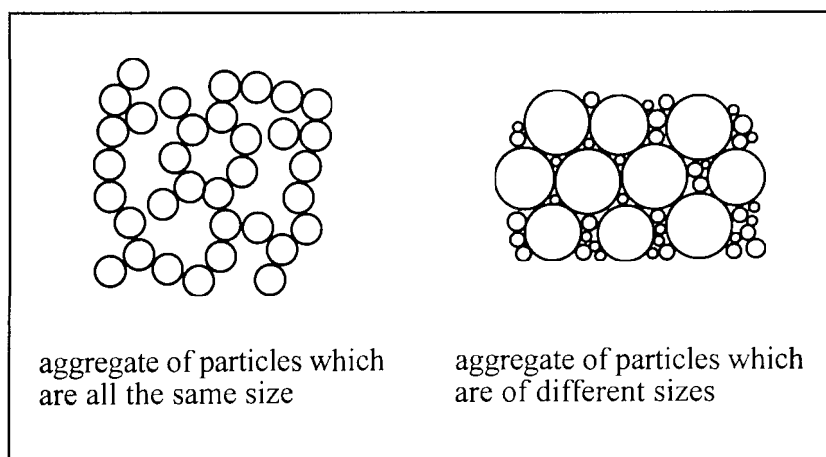


Figure 6.8 Differences in colloid density depending on size range of colloid particles, Saunders, 1990

the barren bands were originally a low water content, dense colloid and the gold rich bands were an amorphous precipitate. He suggests that the variations in water content, particle size and degree of size variation were responsible for the different recrystallisation paths, with the dense gel becoming quartz and the amorphous precipitate becoming opal. The presence of 1 weight % NaCl, typical of epithermal solutions, would greatly increase silica colloid formation and Saunders also suggests that the physical effects of boiling, (bubble formation, agitation), would enhance colloid formation due to increased collisional activity between particles. The gold ore was deposited in areas where the vein was wider. A fluid flowing through a channel, such as a vein, will decrease in velocity if the channel widens. This could allow the larger suspended aggregates to be sedimented, producing the topographically controlled form of the banding.

Herrington and Wilkinson, (1993), also studied gold-silica colloids, but in mesothermal systems. They proposed a pressure release mechanism for forming the silica colloid, whereby stick-slip failure patterns on faults produced significant pore fluid pressure drops. This created very rapid oversaturation of the pore fluids with respect to silica which resulted in the precipitation of amorphous silica, forming a colloidal suspension. The colloid would be unstable and undergo rapid recrystallisation to form quartz. Solid impurities would be expelled to the margins of the colloid with the syneretic water. The initial precipitate is amorphous silica because when supersaturation occurs suddenly, the least ordered form of silica will be produced. With time, this is recrystallised into more stable forms. Mesothermal veins often show low order silica polymorphs and overgrown chalcedonic spherules. Herrington and Wilkinson discount the notion of crack-seal vein formation, (Ramsay, 1980), or episodic vein opening/infill mechanism of Sibson, (1988), as being the possible formation mechanism in this situation because they produce oriented quartz textures which are rarely seen in mesothermal veins. Instead, they propose that stick-slip failure, in the manner proposed by Sibson, (1988), in the mesothermal environment, (i.e. depths of 9-12km and pressures of 2.5kb), would produce pressure drops in the region of 1.3kb. Silica in solution is very pressure sensitive and such pressure drops could cause silica oversaturation and colloid formation. They also cite the experimental work of Betterman and Liebau, (1975) and Oehler, (1976) in estimating the time taken for the transformation from colloid to quartz as being a matter of hours.

Stel and Lankreyer, (1994), used the properties of silica gels to explain certain features they described from a Scandinavian fault zone. They described three types of breccia-bearing silica veins occurring within the fault zone;

Type 1 exhibited a repetitive variation in the amount and size of clasts of the wall rock it contained. Locally this layering disappears and the clast load is unsorted. There are also offshoot injection veins similar in form to clastic dykes.

Type 2 veins show pronounced grading with the clast size decreasing across the vein. Stel and Lankreyer claim that the clast size decreases from one wall to the other, but it is not clear from their photomicrographs whether this occurs within one layer or is the result of different layers containing different clast sizes. The clasts are not aligned and the layers have variable clast content. There is evidence of solid state flow of individual layers, such as folds and boudinage. The quartz content varies from cryptocrystalline quartz with no clast burden to radiate, bladed quartz crystallised around the clasts with no intervening matrix.

Type 3 veins show a clast size grading, coarsening towards the centre of the vein.

Scanning electron microscope studies indicate that the quartz crystallised as radiating quartz sheaves in a matrix of chalcedony and fluid inclusion work showed that the last phase of fluid to be trapped was pure water.

Stel and Lankreyer, (1994), conclude that the deformation and flow textures were formed prior to quartz matrix crystallisation. The fluid inclusion data indicates trapping temperatures of 140-180°C. The fluid in which they formed could not have been a melt, yet from the radial growth of silicates on clast nuclei and the presence of folds and boudinage, they deduce that it had a high viscosity. They therefore deduced that it was a silica gel. Type one and two veins are largely clast supported, but the offshoots are matrix supported and the finer clast element in the main veins is also matrix supported, suggesting that settling was obstructed. They suggest that this feature was caused by a sudden rapid increase in viscosity. They describe a texture of interlocking bladed quartz generated by growth supported within a gel. The gel is proposed to have formed from the mixing of ground quartz with hot fluids drawn through the fault zone and coagulated by quenching the hot fluid on cold vein walls upon injection. Stel and Lankreyer proposed a time span of minutes to years for this process, and the transformation to crystalline quartz, to occur.

6.5 Summary

From the previous discussion it is clear that high viscosity liquids can be formed by either chemical gelation or mechanical particle aggregation. The terms gel and colloid are applied to such substances, but they are often used either interchangeably or with slightly different definitions. Gelation is defined in this thesis as a chemical reaction which results in the polymerisation of a dissolved species to form an interlinked network, which becomes the solid phase of the gel. Colloids are defined as a suspension of dispersed or aggregated particles with the degree of particle aggregation corresponding to the degree of molecule polymerisation in gels.

Silica gelation is held to be irreversible, (Henisch, 1970), as a result of the stability of the silica-oxygen bonds which create the solid phase framework. Colloidal silica, however, has the particles linked by surface forces, which can be broken by the

application of stress or shock, (thixotropy), and which can be reformed thereafter, (Elliston, 1984). Colloids can therefore be remobilised after coagulation has produced a highly viscous material.

One of the notable features of colloids is their ability to act as a medium for producing very well formed crystals with few defects, (Henisch, 1970). This is because of the colloids ability to 'cradle' the growing crystal, protecting it from secondary nucleation and supporting it while it grows without exerting any restriction on that growth. The high permeability of colloids and gels allow them to filter the liquid phase, maintaining a saturated solution at the crystal face, whilst expelling impurities to the margins of the system.

Of the two mechanisms described above, colloidal aggregation and gelation, the former shows a very strong resemblance to the process of fluidisation discussed in the previous chapter. The major difference between the two is not in their formation mechanisms, both of which involve the dispersal of a granular material in a fluid phase, but in the different behaviour of the resulting suspension, differences which are the result of the difference in size of the particles. In a fluidised bed the particles are large enough for surface forces acting between them to be unimportant. At the scale of colloidal suspensions however these forces are critical. Where a fluidised bed will always behave as a liquid, the fine grained suspension will undergo particle aggregation and will eventually coagulate, becoming highly viscous. Both gels and colloids are capable of being introduced to the system as a low viscosity fluid and of coagulating with extreme rapidity after a short period of time into a highly viscous media capable of supporting clasts in suspension.

Elliston's work, (1984), predicted that a gel or colloid of silica would always recrystallise into a lower energy form at a surface. Oehler's experiments, (1976), demonstrated that crystal growth always took place on the impurity surface if the gel was seeded with impurities. He identified the radial growth habit as the typical form of crystal growth from a seeded gel and explained it as being the result of the impurity blankets left by the consumption of gel during recrystallisation at the crystal face, causing growth along rapidly branching crystals. He showed that once nucleated, radial crystal growth was extremely rapid and that it took place under geologically valid temperatures and pressures. Betterman and Liebau (1975) showed in more detail that recrystallisation from a gel or colloid happened over a range of geologically valid temperatures and pressures and that at the PT conditions described for Rusey from fluid inclusion work, (Davies, 1993), of approximately 3kb and 350°C, the transition would be directly from the 'amorphous' (i.e. non-crystalline), silica to stable quartz. The conditions ascertained by Herrington and Wilkinson (1993) from fluid inclusion work on epithermal gold veins containing radial quartz growth textures match the Betterman and Liebau experimental results.

Chapter Seven

Constraint of Rusey Breccia mineralisation mechanisms using evidence from Kimmeridge Bay, Dorset

7.1 Thrusting and Related Brecciation in Kimmeridge Bay, Dorset

7.1:1 Location and Aims

The area studied is Kimmeridge Bay, located on the Dorset coast, 9 miles west of Poole (O.S. Outdoor Leisure sheet 15, G.R. 905790) Fig. 7.1. The bay is the type location for the Upper Jurassic Kimmeridge Clay, a series of gently south dipping interbedded shales and dark carbonates.

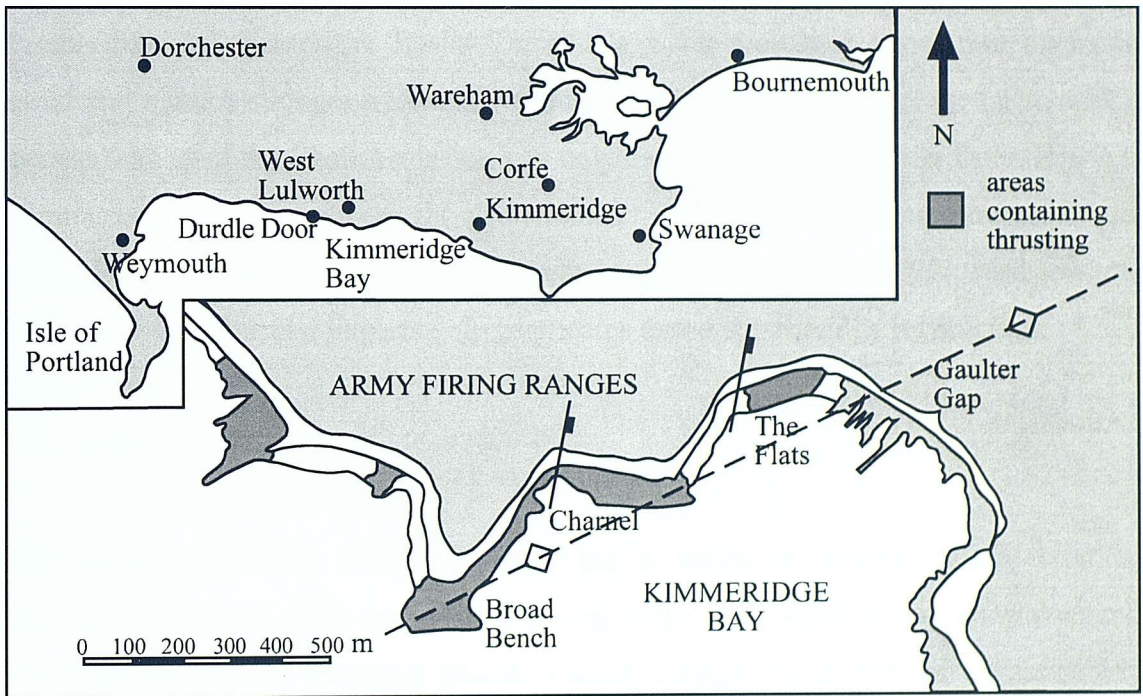


Figure 7.1 Sketch map of Kimmeridge Bay, Dorset, Leddra et al. 1987

The study was undertaken as a result of the discovery of breccia textures found along thrust planes, identical to the textures at Rusey, with matrix supported clasts, bearing radial crystallisation. These clasts very often showed no interstitial matrix between the clast/corona units, and in addition fault gouge was absent. The environment of

formation for these structures was clearly very different to that at Rusey, lacking both the intense plate tectonic activity during formation and great burial depths and temperatures which accompanies the latter. In addition the Kimmeridge Bay rocks are part of a carbonate chemical system rather than the silica system that operated at Rusey. Comparison of the mechanisms of carbonate gel and colloid formation and silica gel and colloid formation allows the evidence from Kimmeridge to be used to constrain the range of plausible mineralisation mechanisms operating in the field area at Rusey.

7.1:2 The Regional Structure

The regional structure is dominated by the Purbeck Anticline (B.G.S. Sheet 341) a gently folded, E-W trending structure and, to the north of the Anticline, the Purbeck monocline, a much more intensely deformed structure, also trending E-W. These two structures are the cover expression of reactivation on a major fault which marks the northern boundary of the Jurassic/Cretaceous basin and the southern margin of the Paleo-Oligocene Hampshire Basin. There is a weak pre-Chalk deposition inversion event and a much stronger post Chalk deposition event. The effects of the latter, which caused folding of the Chalk sequence, can only be clearly seen at Durdle Door, Fig. 7.1. Kimmeridge Bay lies between the Anticline and the Monocline, along the trace of the Broad Bench anticline. This fold runs from Broad Bench on a E30° N trend towards Kimmeridge village and imparts only gentle dips to the Kimmeridge Bay section.

7.1:3 The Structure in Kimmeridge Bay

The earliest structures within the bay are a series of growth faults, oriented approximately north-south and downthrowing to the east, which show later inversion. There have been two inversion phases, a local, limited one in the late Jurassic/early Cretaceous and a later, more intense one, related to the Alpine Orogeny, in the Mid Cretaceous (Scotchman, 1989, 1991). There is no direct evidence to link the small-scale thrusting and brecciation of the cementstone (which is discussed in the following paragraph and which is the main focus of this study) to either phase and its timing is therefore difficult to determine. Previous structural and geochemical studies (Bellamy, 1977, Irwin, 1980, Leddra et al, 1987) indicate that it occurred early in the burial history, probably at no more than 100-200m. This would place the thrust development

as late Jurassic to early Cretaceous (Scotchman, 1989, Scotchman, 1991, McQuaker and Gawthorpe, 1993). There is a phase of extension which produced normal faults which offset the thrust planes, Plate 7.1. This small-scale extension, shown in Plate 7.1, is almost certainly the same phase of extension which formed the mesoscale north-south faults which cut the Flats Stone Band between the Flats and Charnel areas and the Charnel and Broad Bench areas, shown on Fig. 7.1. These faults downthrow the Flats Stone Band to the east. As they cut the thrusts they were not formed during the same episode of extension as the syn-sedimentary faults which can be observed in the cliffs at the back of the bay.

7.1:3a The Thrusts

The thrust faulting occurs within one of the cementstones called the Flats Stone Band which, like the majority of the cementstones, consists largely of dolomite, Fig. 7.2. These thrusts occur in three outcrops, Broad Bench, Charnel and the Flats and have been studied by the present author in the areas known as the Flats and Charnel, Fig. 7.1. Late extensional faulting has downthrown the sequence to the east and it is probable that all three outcrops are formed from the same stratigraphic horizon. The Flats contain mineralised thrust planes which are largely but not entirely contained by the dolomite. The base of the beds rarely show offset exceeding a centimetre and the thrust planes can be traced for up to a metre as they pass into the shales and become subhorizontal, Plate 7.2. In the Charnel area, the top and bottom of the beds are both clearly displaced by several centimetres, Fig. 7.4, but no example was observed of the bedding parallel thrust planes cutting up or down into the surrounding shales.

In plan view the thrusts form a polygonal framework, Plate 7.3. Although this is partly apparent and the result of the overlapping of separate thrust planes of different orientations it can also be truly polygonal, where it is due to the change in orientation of a single thrust plane.

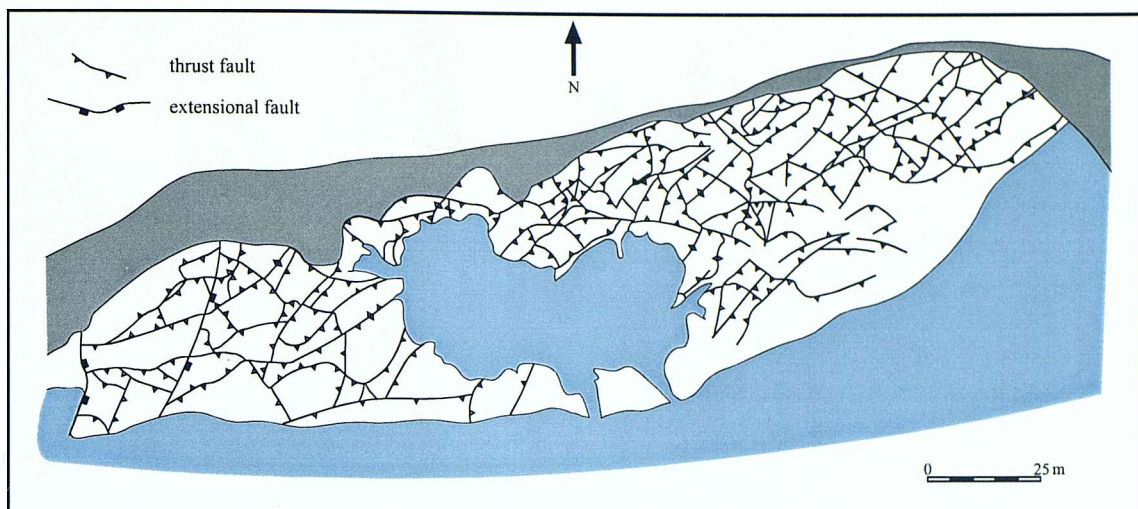


Figure 7.2 Map of the thrusts in the Flats area of Kimmeridge Bay, Leddra et al. 1987

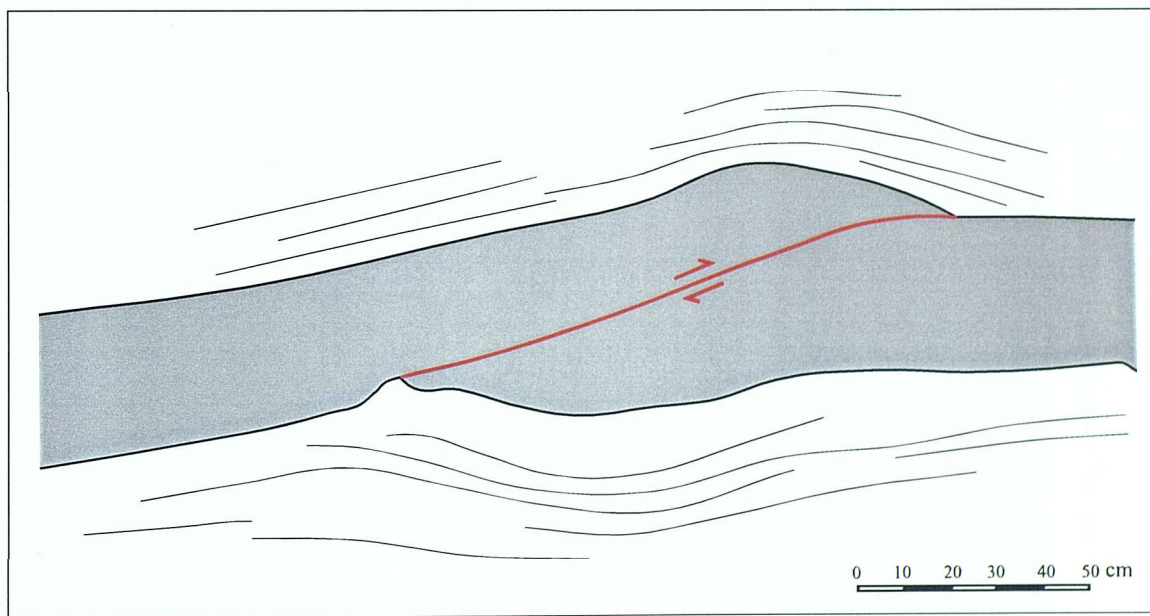


Figure 7.3 Thrust displacement in the Charnel area of Kimmeridge Bay. There is clear downwards displacement of the lower bed surface in the Flats Stone Band (shaded) indicated by the down warping of the fissility in the underlying shales and arching of the bed surface. This feature is not prominent in the Flats area to the east.



Plate 7.1 Extensional faulting cutting the mineralised thrust planes at Kimmeridge Bay, Dorset, scale bar = 5cm)

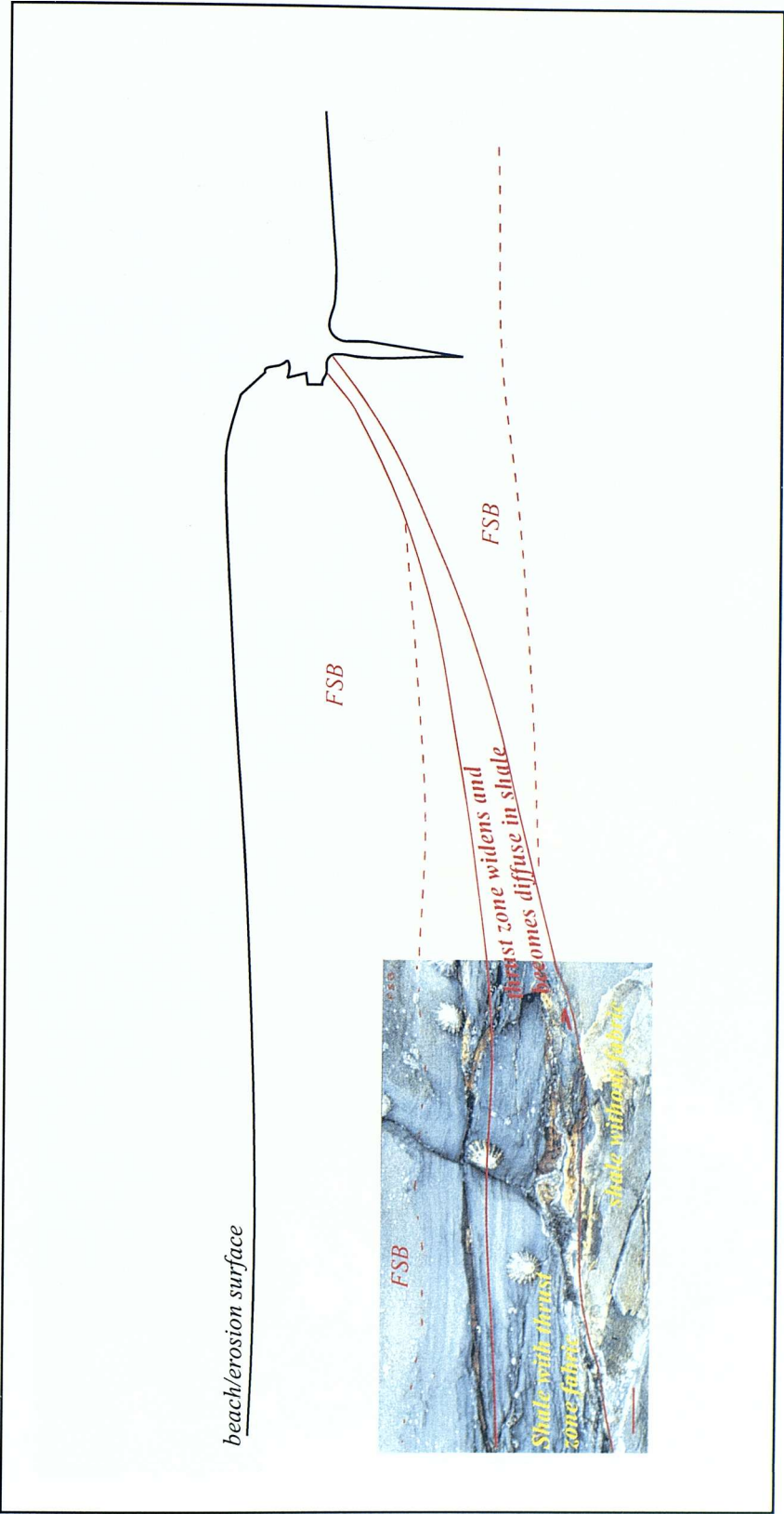


Plate 7.2 Thrust, (marked in red), passing from the Flats Stone Band, (FSB), into the underlying shale. Thrust plane widens into thrust zone which rapidly dies out to the left of the photograph. (schematic sketch from field data, scale bar = 10cm)

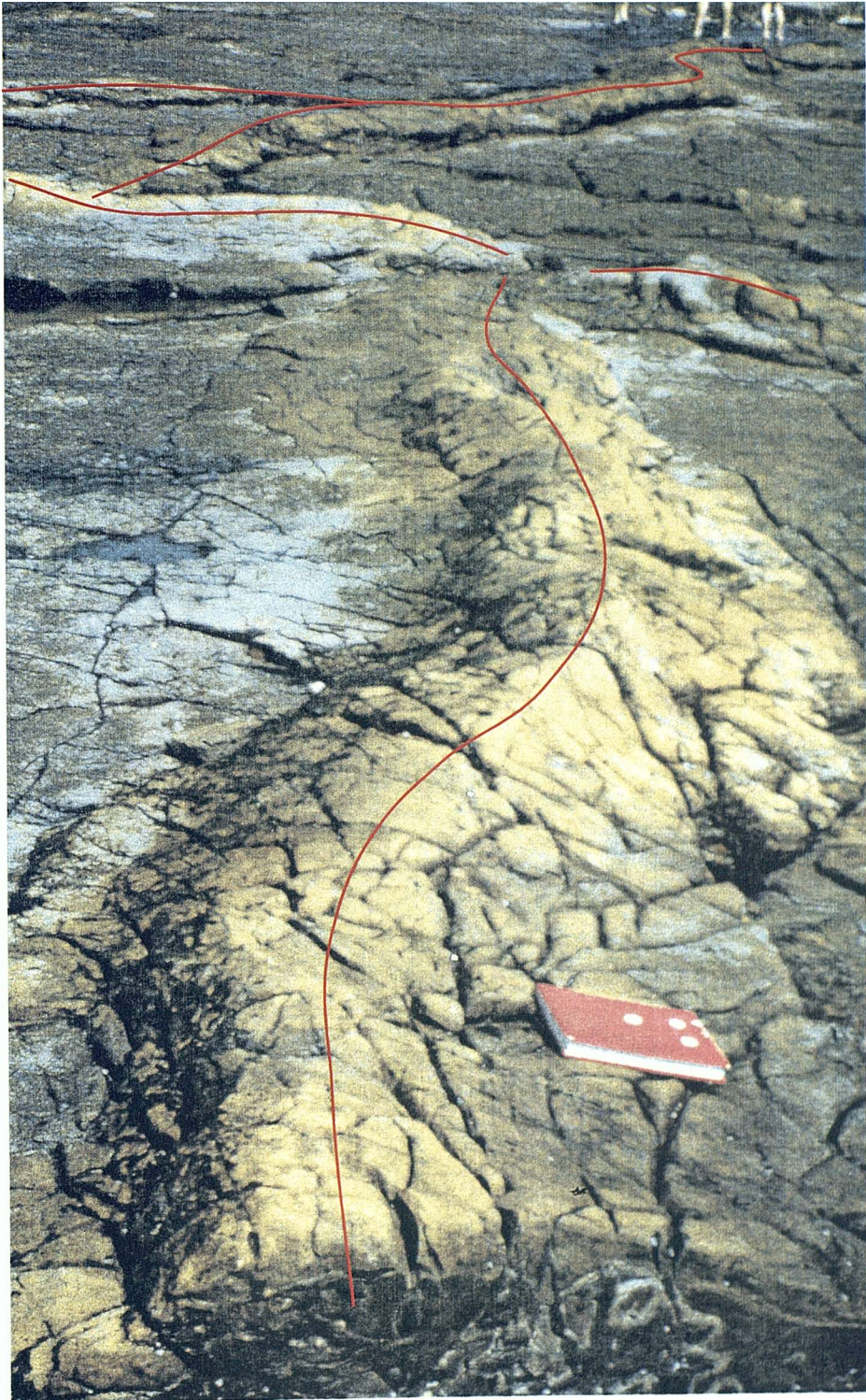


Plate 7.3 Polygonal development of thrusts, (marked by red lines)

Stereograms plotted by Leddra et al. (1987) show only a slight NW-SE trend, Fig. 7.4a, to the orientations of the slickenside plunge across the thrust planes and to the plane containing the poles to the thrust planes. They used this, together with the lack of similar layer parallel compression in the surrounding shales, as evidence for the non-tectonic origin for these structures. However, their data was split into East, Central and West areas of the Flats Stone Band and compilation of their data into a composite plot, Fig. 7.4b, reveals a more significant trend in the NW-SE orientation for the slickenside plunge orientations, although the range of orientations for the poles to the thrust planes is slightly increased.

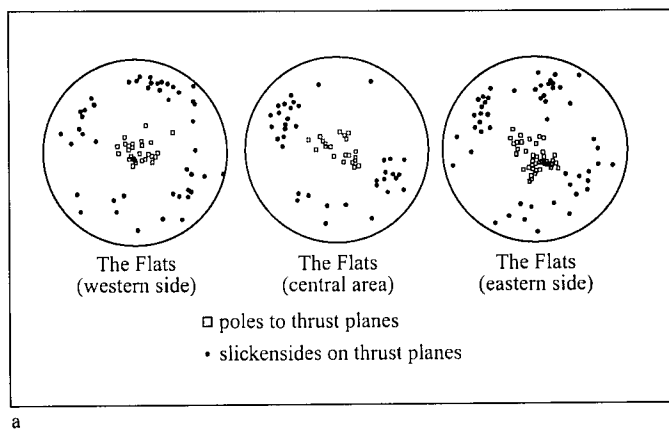


Figure 7.4a Stereoplots of thrust data from the Flats, Kimmeridge Bay, Leddra et. al. (1987), showing only a very slight preferred orientation to the thrust planes.

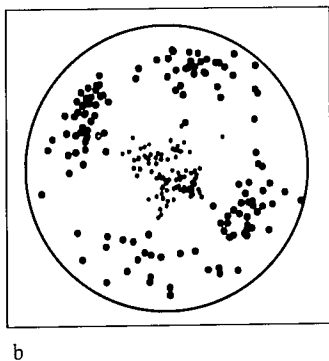


Figure 7.4b Stereoplot of all thrust data from the Flats, Kimmeridge Bay, Leddra et. al. 1987.

The rocks also show two episodes of jointing in addition to the thrusts and the later set of normal faults mentioned above. The oldest phase of jointing is termed here the cobblestone jointing, as the joints are randomly oriented or rectilinear in plan view, forming polygons in the order of 10-20cm diameter which weather in the more intensely jointed areas to produce a cobbled topography on the bed surface, Plate 7.4. This joint set is variably developed and exists largely within the dolomite beds. Within the shale beds above and below the dolomite the polygonal patterns persist for a distance of several millimetres into the bed. The planes are not quite well enough defined to be called joints, being little more than zones where the fissility of the shales has been disrupted. As the position and orientation of these features are identical to those of the jointing in the dolomite and as they only form in close proximity to these joints it seems likely that they were formed at the same time. If this is so then it is clear that during joint formation some cohesion existed between the dolomite and the shales. Their presence in beds both above and below the dolomite suggests that the jointing was initiated in the dolomite band and propagated in both directions vertically. This first joint set decreases in intensity westwards in both the Flats and Charnel areas and the thrusting varies likewise. In unthrust dolomite beds this jointing is not present.

The cobblestone jointing shows a number of significant variations in its development.

- the cobblestone jointing is only developed in areas of the Flats Stone Band which have also undergone thrusting.
- the intensity of the development of the jointing is less where the thrusts are fewer
- where the thrusts exist the jointing has a spatial relationship with the thrust fault such that it is poorly developed in the immediate proximity of the thrust fault. Where there are few thrust planes in a given area the cobblestone jointing is not well developed and decreases in intensity of development towards the thrust planes. Where there are a large number of thrust planes in a given area the cobblestone jointing is intensely developed but will still be relatively less well developed in proximity to the thrust plane, Plate 7.4. Where the thrust planes form complete polygons the cobblestone jointing appears to be most intensely developed in the central area of the polygons. This is because joints along the thrust front are closed, making them less susceptible to weathering processes and therefore less apparent.



Plate 7.4 Development of cobblescale jointing

A prominent fracture set has developed adjacent to the thrust planes. The fractures are long and curvilinear and are oriented parallel to the thrust front, Plate 7.5.

The second major set of joints ranging in orientation between 340°-020° cuts all beds and structures. It is equally well developed in all areas of Kimmeridge Bay. It is placed as the second episode of jointing on the basis of limited abutment data. These show that the uniformly developed joints terminate against the cobblestone joints and that there is a tendency for these planes to become far more irregular as they pass through patches of cobblestone jointing. It appears that propagation of the joints utilised appropriately oriented elements of the pre-existing fracture set. More detailed work would be required to prove these age relationships conclusively.

There is an episode of extensional deformation which postdates the two joint sets discussed above and which has given rise to normal faults striking north-south. On a much smaller scale, Plate 7.1, there is considerable evidence of this late phase of extension throughout the Bay, and the thrust planes are frequently cut by later extensional faults several centimetres in length and oriented north-south.

7.1:4 Mineralisation and Breccia Development.

Mineralisation and breccia development is restricted to the thrust planes within the Flats Stone Band. The thrust planes within the shale are bed parallel and rarely mineralised and at the outcrop scale they are most often defined by a zone of intense thrust/bed parallel fabric development.

Within the dolomite the thrust plane morphology ranges between two end members. The first takes the form of a thrust plane which contains a band of dolomite mineralisation several millimetres thick at the base of the cementstone band. As the thrust ramps up through the dolomite band, the thrust plane subdivides and widens into a diffuse network of millimetre scale veins which is developed across the centre of the band. At the top of the dolomite band this network narrows and the veins coalesce into a single plane, several millimetres thick. The bifurcation and divergence of these planes across the central part of the band forms a pod shape, Fig. 7.5. The planes generally contain crypto-crystalline dolomite mineralisation and do not contain breccia. However in some instances the area of carbonate bed between planes has undergone extensive hydrofracturing, Fig. 7.5 and 6. This can be seen to form part of an evolutionary sequence leading to the second end-member morphology. The hydrofractured lenses



Plate 7.5 Development of jointing parallel to the thrust tip, (fault tip marked in red)

contain clasts up to several centimetres in length, which are extremely angular and can be easily restored to their unfractured state by removal of the volume now occupied by mineralisation.

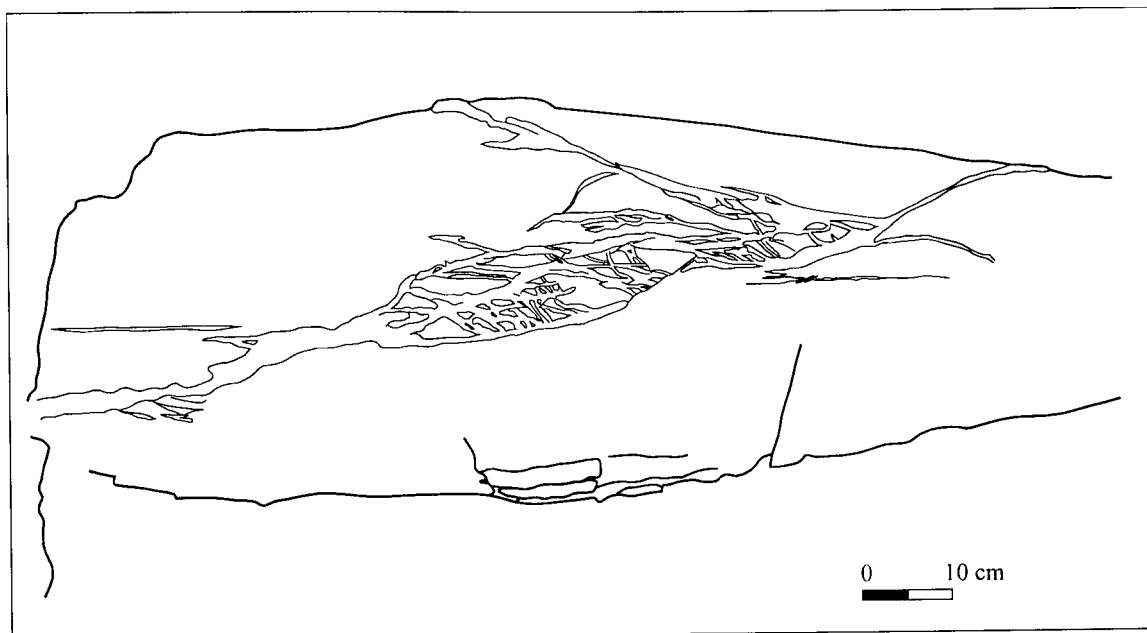


Figure 7.5 Pod development in cementstone bed as the thrust plane bifurcates across the central region of the bed.

The opposing end member morphology takes the form of a single thrust plane which can be several centimetres thick, Fig. 7.6 and Plate 7.6. The thrust plane contains large amounts of dolomite mineralisation and variable but significant amounts of dolomite breccia clasts. This breccia differs from the hydrofractured lenses, Fig. 7.5 and 6, in that the clasts are usually sub-rounded to rounded and appear to have no genetic relationship to their current location. In the Flats area these breccias occasionally show a rudimentary clast size grading which fines outwards towards the edges of the mineralised thrust plane. Mineralisation takes the form of concentric, crypto-crystalline layers or radial crystallisation of medium grained dolomite around the clasts. The clasts and their concentric mineralisation are set in an crypto-crystalline interstitial matrix of dolomite. Vug space is rare.

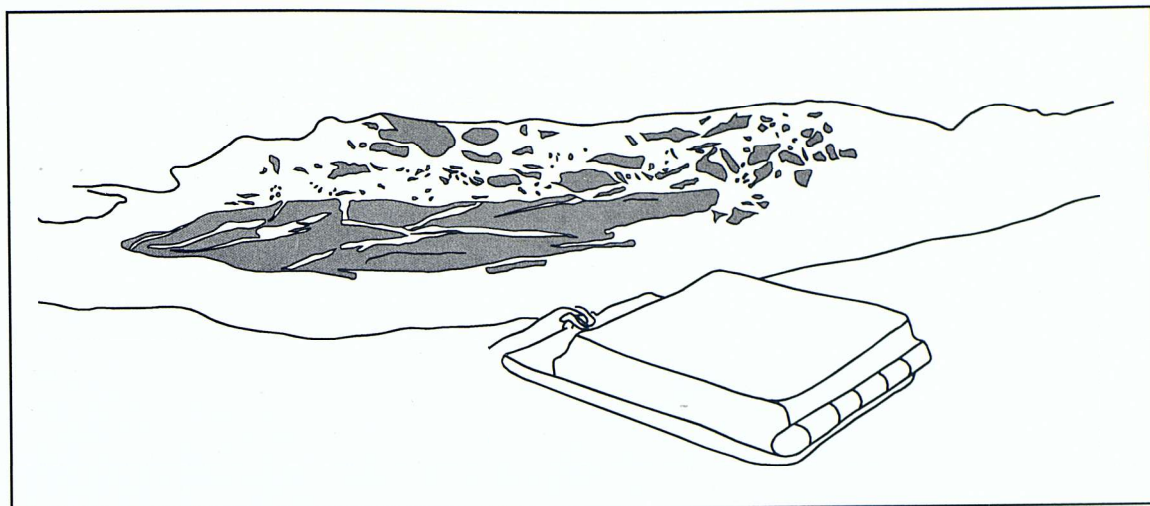


Figure 7.6 Hydrofracture textures within the thrust plane breccia. The large central clast would originally have been an unfractured area between two veins. It is gradually being fractured and broken into thrust breccia as the thrust zone develops and widens.

In the Charnel area the breccias show some different and highly distinctive features. Displacement on the thrusts appears greater than in the Flats area, with significant downward displacement by several centimetres of the lower bed surface, Fig. 7.4. This thrusting is accompanied by the formation of thicker thrust planes, in the order of fifteen or twenty centimetres thick, and the development of Rusey type textures in the eastern part of this area. These take the form of well developed grading, Plate 7.7, and medium to coarse grained, euhedral, radially crystallised clast coronas which have no interstitial matrix and as a result a significant amount of vug space, Plate 7.6. These textures are only present on the eastern edge of the Charnel and even there are only sporadically developed. There is no rock flour or fault gouge present in the breccias of either the Flats or Charnel areas.

7.2 Comparison of mineralisation mechanisms.

In this section the points relevant to distinguishing between the colloidal and gelling mechanisms are summarized and discussed. The conclusions of the discussion are used to choose the most likely mechanism to account for the breccia mineralisation features observed at both Rusey and at Kimmeridge Bay.



Plate 7.6 Matrix textures along a thrust plane in the Charnel area, Kimmeridge Bay. Although the photo is not well focused, it is possible to see the matrix supported nature of the breccia clasts and to pick out, particularly in the upper right hand corner, below the lens cap, the presence of an initial phase of mineralisation, crystallised around each individual clast, before the breccia assemblage was infilled by later mineralisation. This initial crystallisation around each clast consists of radial crystal growth, identical to that at Rusey



Plate 7.7 Graded clast sequence along a thrust plane in the Charnel area of Kimmeridge Bay

- Whilst silica gel formation can be triggered by a change in physical conditions such as change in pressure (Herrington and Wilkinson, 1993) readily expected in an active fault system, carbonate gel formation requires the addition of catalysts (Wilkinson, pers. comm.). The thrust faults at Kimmeridge form an apparently closed system, which would require the catalyst to have been present since deposition. This would imply that carbonate solutions should have reacted with the catalyst as they evolved and could not have remobilised the breccia clasts and then gelled.
- Although gel formation concentrates the dissolved mineral into a smaller volume, in order to provide the matrix at either Rusey or Kimmeridge, a gel would require either a higher concentration of dissolved mineral material than it is possible to take into solution or an extremely high degree of dilation of the brecciated area in order to accommodate the volume of gel required to crystallise the observed volume of matrix.
- Since brecciation processes in fault zones are commonly observed to produce a fine grained gouge surrounding the clasts, the lack of a fine grained fault product in both systems requires explanation. This is of particular importance at Kimmeridge, where the fault system is closed on the scale of observation and the fine grained material cannot have been transported out of the system. The fault gouge must therefore have been removed by a different and very efficient mechanism.
- Using the fault gouge as the solid phase of a colloid suspension is extremely efficient as it requires only sufficient fluid input to mobilise the gouge rather than either the input or throughput of a large volume of liquid containing dissolved mineral. This is of particular importance at Kimmeridge, where the thrust system appears to have been closed. It also creates, in situ, a mineralising fluid which contains an extremely high percentage of mineralising material, requiring very little new material to be added to the system in order to create the matrix.
- The process of colloid formation and the process of clast fluidisation are essentially the same process of fluidisation, demonstrating a scale dependent difference in behaviour. This would mean that the organisation and mineralisation features of both the breccias could be explained by the operation of a single process. At the scale of the fine grained suspension, interparticle forces are the dominant forces governing the rheology of the mixture, whereas at the scale of visible clasts, the

fluid dynamics of the clast-fluid interactions and clast-clast collisions are the dominant forces. At the larger scale, clasts over a millimetre in diameter, the surface area to volume ratio is so high that surface forces are too small to affect the behaviour of the particle, which is controlled by the mass of the particle and its interaction with the fluid and other clasts. The suspension therefore remains fluidised and capable of dynamic reorganisation until the fluid velocity drops. This occurs very rapidly as the fine grained particles begin to interact and coagulate. Fluidisation of fine grained suspensions is difficult without a mixing mechanism to break inter-grain bonds. The presence of the clasts would provide such a mechanism and once fluidised, the suspension is very stable, i.e. the particles have too little mass to settle out under gravity, enabling it to create an appropriate environment for the support of clast and for matrix crystallisation once it has coagulated.

With reference to the first two points made above, it is argued that an unacceptable degree of special pleading would be required to generate identical morphological features from two distinct chemical processes, one of which at least (the addition of the appropriate chemical catalyst uniformly throughout a liquid to produce gelation in what is apparently a chemically closed system) seems very unlikely. To accept the explanation of two chemical processes operating with identical results in systems of completely different chemistry when one physical process (namely the formation of a colloidal suspension from the fine grained fault gouge) operating in two systems of very similar physical conditions can produce the same results, seems unjustifiable.

In addition to being an efficient process, this mechanism avoids certain significant problems associated with attempting to derive the matrix from a gel, namely that the quantity of mineralising material in a gel is considerably less than that of the matrix material derived from it, i.e. although the formation of a gel concentrates the material available for mineralisation, it is still not concentrated enough to account for the very large volumes of mineralisation seen in the breccias at Rusey and Kimmeridge. The percentage of matrix of the breccia band is very high, over 50% at most points in both systems.

To generate this amount of mineralisation from aqueous solution would require the throughput of enormous volumes of water (the solubility of silica in water at 300°C $\approx$ 700mg/kg, (0.07%), Fournier, 1985). If the source of the mineral forming the matrix of the breccia were a gel then the mineral would be present in much higher concentrations

(up to $\approx 30\%$ for the silica/water system, Henisch, 1970, Iler 1955). However even so, for a gel to produce the amount of mineralisation seen at Rusey, where the matrix occupies up to 80% of the rock, would require the volume occupied by the gel during formation of the matrix to be in the region of twice the volume currently occupied by the breccia. There is no sign that either the Rusey or Kimmeridge systems have been subjected to such extreme dilation, though there is considerable pore space within the breccia.

However if the matrix were to crystallise from a colloidal suspension generated from a dispersed fault gouge, this volume problem does not arise, since a large volume of mineralising material is already present in suspension and the mechanisms for transforming it from ground material to recrystallised material have been documented by Elliston 1984. It was for this reason that the reader's attention is drawn to the fact that in both Kimmeridge and Rusey breccias the matrix is composed of a pure form of the major constituent of the host rock (which is 80%-90% quartz or dolomite) leaving little material to be lost during the recrystallisation process. Production of the matrix from the original fault gouge has the advantage that the original fault gouge would not have occupied a significantly different volume as a fault gouge than it does as a crystalline matrix and the only dilation needed is that necessary to accommodate the volume of water in which the gouge and clasts are suspended. The need for the suspension to have been dense enough for coagulation to occur means that the volume of water involved is likely to have been comparable to the volume of pore space, the abundance of the latter being remarkable in both systems.

7.3 Summary

On the basis of the evidence from Rusey, the existence of a remobilising fluid and its abrupt transition in state by either gelation or colloidal aggregation has been proposed. Unfortunately the field evidence does not enable one to determine which of the two processes, i.e. gel or colloid formation, gave rise to the dramatic increase in the viscosity of the fluid. The limited study of the thrust breccias at Kimmeridge Bay, Dorset was carried out in an attempt to resolve this problem.

As mentioned above only a limited amount of time was available to study the brecciation at Kimmeridge and as a result no attention has been given to the

mechanisms of formation for the thrust system at Kimmeridge, which is a matter requiring considerable further work. The work concentrated on ascertaining;

- that the deformation system being studied involved the production of fault breccias intimately related to thrust fault activity. (The breccias are entirely contained by the thrust planes within the dolomite band).
- that the deformation system involved the presence of significant amounts of fluid and the likelihood is that the fluid pressures involved in the system were high, as indicated by the presence of tensile fracturing and considerable mineralisation.
- that the clasts showed signs of having been remobilised after the initial brecciation event, as demonstrated by the rounded profile of the clasts and the lack of any obvious relationship to their immediate surroundings,
- that the system underwent considerable mineralisation, clearly indicated by the monomineralic dolomite matrix to the breccias.
- that the mineralisation consisted of extremely pure and well formed crystals, formed in three dimensional coronas about the clasts, Plate 7.6.
- that the mineralising material was of essentially the same composition as the host rock (Irwin, 1980, Leddra et al, 1987).
- that there was no fine grained fault product present in the system, Plate 7.6 and 7.7.

The above points relate to the comparison of the brecciation and mineralisation of the Rusey and Kimmeridge thrust systems and the study pursued by the present author at Kimmeridge Bay was limited to gathering sufficient data to ascertain the similarity of the two breccias with regard to these points.

It is concluded on the basis of the presence of these features in both locations that similar processes operated during the production of the both the Kimmeridge and Rusey fault breccias. Breccia formation occurred to a much more limited extent at Kimmeridge and later interstitial cements tend to obscure the crystal form at Kimmeridge. The deposition of this cement is regarded as being unrelated to the initial mineralisation event on the basis of its crypto-crystalline form and differing colour, the latter indicating a difference in chemical composition. The matrix is monomineralic dolomite in a dolomite lithology at Kimmeridge (Irwin, 1980, Leddra et al, 1987) and the matrix is monomineralic quartz in a siliciclastic lithology at Rusey.

Because the Rusey and Kimmeridge systems are of such different chemistry, any mechanism which produces a mineralising fluid with the ability to develop such similar

and unusual features in both systems should not depend on the occurrence of a particular process linked to the chemical composition of the rock (e.g. the formation of a silica gel). It is unlikely that two different chemical reactions in two significantly different geological environments could produce identical mineralising environments. Therefore, the presence of such similar features in the two systems has led the author to prefer a physical process (unrelated to the chemical composition of the rock) namely colloidal aggregation, for formation of the mineralising fluid. In addition the involvement of a colloid would avoid the requirement for a high fluid throughput to maintain the availability of solute for mineralisation by deriving all the matrix material from the breccia gouge. Since the mineral source is derived in-situ, the amount of dilation required is much lower.

The theory of colloid rather than gel formation is also preferred by the present author because when the breccia is remobilised it can be considered to be a system made up of a suspension of extremely fine grained material (gouge) which supports a suspension of much coarser material (clasts), thereby allowing the single process of fluidisation and suspension to produce both dynamic reorganisational environment and the stable mineralisation environment. This is because of a difference in the interaction behaviour of the suspended particles which is dependent on the size of the particles. In a system which is fluidised by water, the clasts would settle out as the fluid velocity drops, but in a system fluidised with a very fine grained suspended element, the coagulating suspension of fine grained material increases in viscosity as it decreases in velocity and the clasts remain in suspension. The difficulties of creating a well fluidised suspension of very fine grain size particles can be overcome by the mechanical effects of the clasts. The larger clasts disturb the finer grained, more cohesive material, initiating fluidisation (Richardson, 1971). Richardson (1971) showed that once fluidised, fine grained systems remain stable in the fluidised state. Therefore a single fluid input into a fault breccia, consisting of a clast/gouge mixture can both reorganise the breccia and create a mineralisation environment appropriate for the formation of the textures observed at Rusey.

Chapter Eight

A model for the formation of the Rusey Breccia

8.1 Introduction

It is clear that certain field observations, described in Chapter 4, can be correlated with certain aspects of the processes of fluidisation and crystal growth in gels and colloids described in Chapters 5 and 6. The summary sections of Chapters 5 and 6 present a justification for viewing these processes as geologically significant. This chapter will draw on the field evidence from Rusey (Chapter 4) together with the description of the processes (Chapter 5 & 6) to propose a model whereby the processes described in these chapters could have exerted an influence on the breccia zone to produce the features observed in the field.

The model set out is based on detailed field work, laboratory studies and a consideration of various processes associated with deformation and fluid flow.

Any model proposed for the formation of the breccia must explain how clasts of country rock come to be supported in a matrix of quartz. In addition it must account for the presence of layering within the breccia band, layer normal grading in clast size, the protrusion of clasts from one unit into another and the formation of the extremely pure matrix which exhibits well developed euhedral, three-dimensionally radial quartz growth textures around the clasts. The presence of identical breccia textures formed at Kimmeridge Bay (Chapter 7) in a carbonate dominated rather than a silica dominated system, under very different geological conditions, has been used to constrain the model derived for the mineralisation of the breccia at Rusey. The present author argues that, whilst the two locations had very different chemical conditions, the mechanical processes, which were primarily controlled by fluid pressure and stress, were very similar. A model is presented which is capable of generating the structures and textures observed within the Rusey breccia. The chapter closes with a suggestion of further work that might resolve some of the problems that still remain regarding the generation of the Rusey breccia and its textures.

8.2 Comparison of Deformation at Rusey Headland and Kimmeridge Bay

RUSEY	KIMMERIDGE
Silica chemistry	Carbonate chemistry
High pressure (3 kb)	Low pressure
Tectonic deformation (externally generated stress regime)	Diagenetic deformation (internally generated stress regime)
Thrust faults >1km long	Thrust faults <10m long
High fluid pressure generated by tectonic burial and prograde metamorphism	High fluid pressure generated by intake from adjacent dewatering shales, sealing and porosity destruction

Despite the differences in scale and driving mechanisms, once thrusting was initiated both the Rusey and the Kimmeridge systems began to respond in an identical manner. The high fluid pressure would have created low effective stresses in the area occupied by the Rusey breccia, preventing the degree of consolidation of the clast assemblage which would be predicted for a loose clast assemblage located on a major orogenic thrust. Low effective stresses would also allow easier remobilisation of the breccia clasts. This enabled the mineralizing suspension to form and coagulate in the same manner at both locations.

8.3 Properties of Fluidisation

- the fluid moves through the breccia by channelised free flow
- it is able to transport a suspended load of clasts, forming clast size grading and channeling features
- gravity is not significant in the initial mobilization period as the force it exerts is far exceeded by the force exerted by the high velocity fluid.
- coarse material causes the single fluid input to be 'streamed' into a number of different channels which provide low resistance paths through the breccia.
- channel acquires a velocity gradient imposed by drag along the channel walls. causes the characteristic bi-directional grading of the breccia clasts.
- the fluid is free flowing, carrying the clasts as a suspended load, fulfilling the conditions necessary for the development of the preferred orientations.

8.4 Properties of Colloids

- provides a mechanism whereby the clasts of the breccia can be held in suspension while the coronas of perfectly euhedral crystals grows around them,
- provides an explanation of the transition between an environment capable of generating flow textures within the breccia bands and an environment appropriate for matrix formation
- allows the formation of large volumes of matrix without the need for unreasonably large volumes of fluid to pass through the system and/or the need for unreasonably large amounts of dilation
- allows gravity to remain an insignificant force because the fluid velocity is only reduced by the fluid's rapid increase in viscosity. This process occurs on the time scale of seconds. The transition from fast moving fluid supporting the clasts to viscous fluid supporting the clasts occurs too fast for the clasts to settle out due to the effect of gravity.

8.5 Synopsis of field data

The quartz veins within the clasts at Rusey indicate that the brecciation did not take place at the surface and the breccia must therefore have been re-organised and mineralised at depth. Fluid inclusion work on the breccia matrix indicates that it formed at depths of approximately 8km. The presence of graded oriented clast material is indicative of its transport within a fluid of a relatively low viscosity. The only process capable of producing such fluid-borne reorganisation textures in a sub-surface environment is fluidisation, a process accepted as fundamental to the formation of clastic dykes and volcanic pipe breccias (McCallum, 1985, Nichols, 1994, 1995).

The experimental work of McCallum, 1985, shows that bi-directional grading is typical of fluidised systems. The fining outwards grading of the breccia layer observed at Rusey is typical of immature systems. This implies that the fluidised system did not flow for any great distance or period of time and that fluid velocity did not wane gradually, but that the flowing medium transporting the clasts passed very rapidly into a static medium which was able to hold the clasts in suspension during the crystallisation of the matrix. The present author holds that this must have happened by an in situ change in state of the mobilising fluid since;

Any fluid sufficiently viscous to be capable of supporting the clasts could not have flowed into the system, replacing a less viscous fluid, without displacing the suspended clast assemblage along with the less viscous fluid.

A less viscous fluid could not have been evacuated and replaced by a more viscous fluid without the clasts settling into a clast supported framework. A more viscous fluid could have permeated this framework but could not have resuspended the clasts.

The clasts could not have been mineralised by the viscous fluid and then ordered since the clasts show rounded to subrounded profiles consistent with attrition such as would have been produced by mobilisation and collision in a fluidised bed, whilst the clasts mineral coronas show pristine, often interlocking, crystals. This implies that the clasts underwent transport followed by mineralisation.

The evidence from Rusey, whilst providing good supporting evidence for the breccia having been reorganised by repeated fluid injection, also places some constraints on the probable physical properties of the mineralising fluid. The requirement for a high viscosity fluid is shown by the radial growth of the matrix crystals, which imply that the clasts must have been held suspended over greater lengths of time than would have elapsed during settling under gravity. In addition, because the matrix crystals show no sign of attrition the suspended clasts must have been held stationary. Two processes have been proposed to provide a suspension medium capable of holding the Rusey breccia clasts in suspension during the crystallisation of the matrix. These are gelation and colloidal aggregation. The gelation mechanism is a chemical process, whilst the aggregation mechanism is an essentially physical process. The work carried out at Kimmeridge Bay and presented in this chapter will be used to justify the present author's opinion that the physical process of colloidal aggregation is the more likely.

8.6 The reorganisational model

It is suggested that the driving mechanism for this cyclic fluid injection is the cyclic stick-slip associated with movement along the main thrust plane. The fault valving effects and stress drops involved in slip events (Sibson, 1988, 1992) could give rise to tensile fracturing, Fig. 7.1, which is spatially concentrated in the rocks underlying the Rusey thrust plane, where the stress field had been modified by the presence of the Rusey ramp system and where fluid pressure have been predicted to be at a maximum (section 3.4:3 and Smith and Wiltchko, 1996). As fault slip continued, the highly fractured rock in the thrust footwall acted as a fluid conduit into the body of rock immediately under the ramp lift-off point which had become a breccia by extensive fracturing and which could be remobilised by fluids drawn at high velocity along the fault plane.

The fluid injection took a channelised route into the brecciated body of rock that was determined by the packing and cohesion of the breccia. This created the breccia 'layers' and also graded and oriented the clasts. This reorganisation caused multiple collisions which generated the rounded to sub-rounded profile. The mobilising fluid underwent a

rapid transformation to a highly viscous state, either by gelation or by colloidal aggregation. The immature grading profile implies that this transition took place almost as soon as the layer was fluidised. This is consistent with the short time period recorded between the initiation of polymerisation/coagulation and the associated increase in viscosity identified for gels/colloids (Henisch 1970, Elliston, 1984) and with the process having been initiated by fluid injection and brought to a very rapid halt by coagulation. Mineralisation of that layer prevented it from being remobilised again and the next fluid injection event took a different path through the breccia, creating another layer.

The author wishes to emphasise that this would have been a very rapid process capable of reorganising and mineralising the breccia in maybe as short a period of time as the duration of a single fault slip event and its aftershocks. This would be effectively geologically instantaneous. To help distinguish the timescale of thrust development and breccia formation from the timescale of breccia reorganisation and mineralisation, the process is explicitly explained here and its occurrence is placed in the larger scale context in the next section.

8.7 Evolutionary model for the formation of the Rusey Breccia

The evolutionary model proposed for the formation of the Rusey Breccia, based on field work and discussions contained in this thesis is as follows:

During the late stages of the Variscan orogeny the Culm Basin, having been the foreland basin of the inverted Trevone basin to the south, began to undergo inversion itself. The thrusting exploited lithological boundaries in the sedimentary sequence and, on the southern margin of the Culm Basin, the Crackington formation, made up predominantly of turbidites, was thrust south over more distal turbidites, which, in turn, were thrust south over the Boscastle Beds, Fig. 2.5.

The Boscastle Beds underwent prograde metamorphism, producing large volumes of metamorphic brines. In addition the rapidity with which inversion succeeded deposition would have caused a large amount of fluid from diagenetic dewatering to be trapped within the thrust sheets. These fluids migrated up sequence until they encountered the structural and lithological barrier created by the Rusey Fault.

Propagation of the Rusey Fault created a permeability barrier which trapped metamorphic and formation waters within the Boscastle Shales. The Rusey Thrust tip then became stuck and displacement was transferred to a ramp which developed behind the thrust tip, creating the fault arrangement seen at the present day, Fig. 3.10. A photoelastic study of this fault system is described and discussed in Chapter 3. The study supported the theory that the area underneath the thrust became a fluid sink, which, as the main thrust locked and the ramp developed, became overpressured particularly around the ramp lift-off point, Fig 3.6d.

Locking of the thrust generated an enhanced differential stress, originating at the main thrust tip. A build up of enhanced differential stress then extended back along the thrust in the footwall. The present author proposes that this differential stress enhancement, in conjunction with the locally high fluid pressure, caused hydrofracturing and brecciation of the thrust footwall in a region extending backwards from the tip. Field observations show that the area in the footwall of the Rusey Thrust corresponding to the area of enhanced differential stress in the models became extensively hydrofractured creating the proto-Rusey Breccia, Fig. 8.1a.

During breccia formation it is argued that the brecciated body would be protected from the overburden load by the high fluid pressures, which maintained low effective stresses.

As the ramp developed, the presence of a high fluid pressure, mechanically weak area in the footwall of the main thrust caused slip to initiate within the breccia rather than within the proto-ramp zone. For the period of maybe no more than a single failure episode on the main thrust ramp system, deformation was relocated to within the breccia body. The sliding surface which is created in the breccia could dilate and draw in fluid easily. This created the fluid injection events which reorganised the breccia. This occurred by local fluidisation of the breccia and the formation of channels of remobilised clasts, Fig. 8.1b/c. Where the clasts were fluidised, their transport and reorganisation caused mixing of the fine grained fault gouge with the injected fluid to form a colloidal suspension.

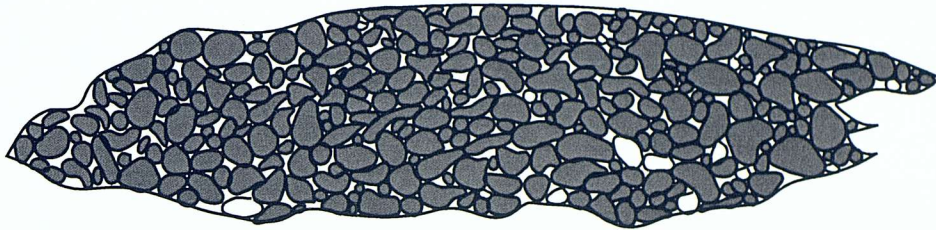
The silica held in suspension recrystallises from the unstable amorphous state to stable crystalline quartz, nucleating on and forming coronas around the suspended breccia clasts.

The formation of new remobilised and mineralised channels, (layers), within the breccia overprinted the hydrofractured breccia textures with the reorganised breccia textures beneath the ramp lift-off point.

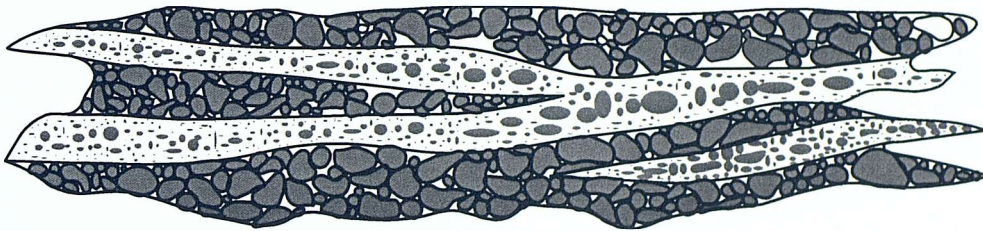
Recrystallisation of the suspended silica progresses to the point where the breccia body was sufficiently mineralised to heal the mechanical weakness that caused slip events to relocate within it. The mineral matrix restricted its ability to dilate and limited the permeability. These latter two effects acted to limit the intake of fluid and to reduce the velocity at which fluid entered the breccia because lesser dilation would have reduced the pressure gradient that controlled the fluid flux. Narrowing of the fluid pathways by mineralisation would have slowed the penetration of fluid. Remobilisation of the breccia clasts therefore ceased.

This may have required as little time as a week and would be geologically invisible in terms of its effects on the evolution of the larger ramp/thrust system.

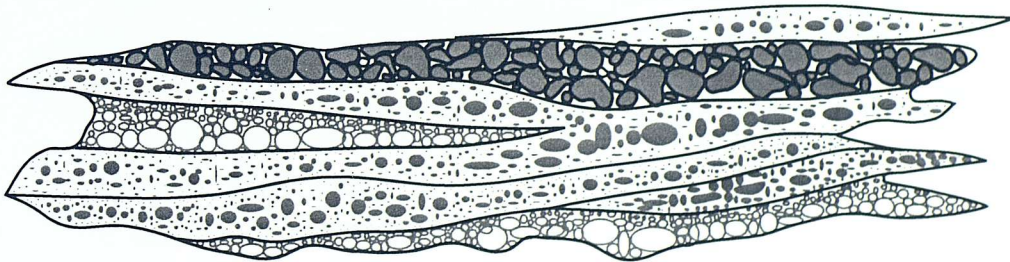
Failure relocated to the proto-ramp zone, which became the weakest element in the system under the prevailing stress regime. The development of the thrust system progressed in a manner similar to that which is more commonly recorded, whereby



a. Initial brecciation of the country rock

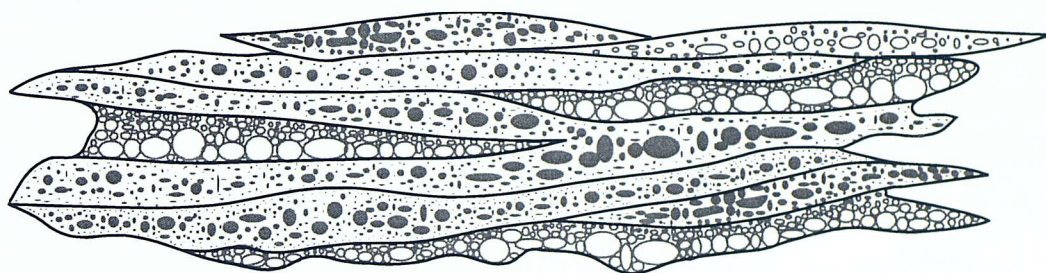


b. Reworking begins as further fault slip events cause fluid to be injected into the brecciated rock

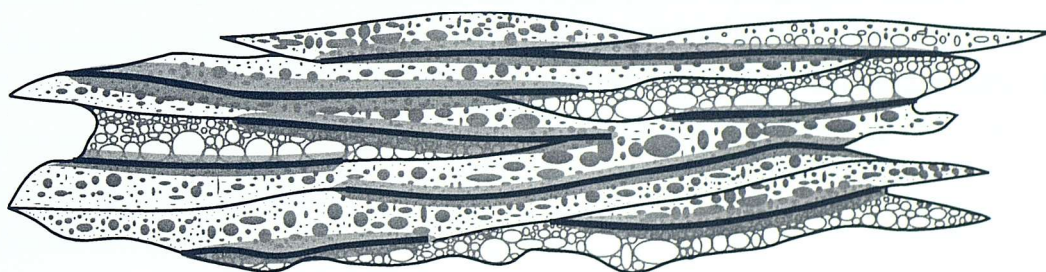


c. Reworking continues and the Rusey Breccia begins to acquire its layered textures. Layers of colloidally supported breccia begin to crystallise and the matrix textures begin to develop.

Figure 8.1 a, b and c The early stages of the evolution of the Rusey Breccia



d. Breccia has been completely reworked into an organised body.



e. Breccia is deformed by discrete slip planes.

Figure 8.1 d and e The final stages of the evolution of the Rusey Breccia.

brittle failure occurred along a clearly defined fault zone which became better defined with increasing slip. During this ongoing deformation the Rusey Breccia responded to the imposed stresses as a rigid body. Its boundaries acted as detachment surfaces along which subsequent thrusting took place. Fabrics compatible in orientation with the stress regime expected within the Rusey Thrust zone developed in the host shales (Chapter 2), and are clearly visible in outcrop. Thrusting ceased at the end of Variscan convergence. Dextral strike-slip movement along north-northwest striking faults rotated the thrusts and the breccia body into their present orientation, dipping south east, Fig. 2.1. The faults separate the intermediate lithologies exposed at Rusey Headland and Middle Strand from the Crackington Formation to the north, obscuring their original relationship.

Extensional faulting took place, extending the area in a north-south direction. A major extensional fault striking northwest and downthrowing to the northeast cuts through Exit Bay to the south of the Rusey Headland and separates the area from the Boscastle Beds to the south, obscuring the original relationship between the intermediate Rusey lithologies and the Boscastle Beds.

8.8 Unresolved Issues

On the basis of the field work and the discussions presented in this thesis the present author proposes that the presence of fault gouge and high fluid pressures, combined with periodic fluid injection, are sufficient to create all the unusual textures seen at Rusey and at Kimmeridge. It is argued that these textures will be associated primarily with thrust faulting and compressive regimes as these offer the greatest opportunities for overpressuring and the occurrence of pressure fluctuations large enough to create the necessary hydrofracture and dilation (Sibson, 1975). However very similar features have been recorded by Eichhubl, 1997 and observed by the present author to occur in a transpressive strike-slip setting at Jalama beach, California. It is accepted that, if the explanation presented for the formation of the breccia textures is correct, these textures should be more commonly recorded. However, it is relevant to note that despite having been studied by several previous workers, the textures in the breccias at Kimmeridge had not been noted before the present study. If the vugs separating the coronas are subsequently filled many of the matrix textures become obscure and easily overlooked. In addition the presence of noticeable grading requires the development of considerable size variation within the clast population and, although common at Rusey, this grading occurs only rarely at Kimmeridge. Furthermore, the amount of dilation required to permit fluidisation and colloid creation has not yet been quantitatively assessed and may prove to be of an unusually high order, although the numerical modeling of Rudnicki and Chen, 1988 demonstrates that the amount of dilation required in order for the stabilisation of a sub-critical slip-patch to occur is not necessarily very great. However,

crystallisation textures very similar to the mineralisation textures observed at Rusey and Kimmeridge have been documented in granites (Elliston, 1984) and the process of crystallisation from a gel/colloid may be applicable to many geological scenarios. In a fault zone setting, if the remobilised zone is connected to a larger fracture network or even to the surface, then the colloid may be diluted sufficiently to prevent the radial crystallisation textures developing (section 6.3, Oehler, 1976). If slip events release the overpressuring the breccia will compact and the resulting rock will be a clast supported breccia with interstitial matrix. There are therefore several reasons why these textures may not form or may not be preserved, although it is the opinion of the present author that these features may be more common than is currently acknowledged.

8.9 Further work

Whilst the processes described in this thesis are mechanically and geologically viable, there is considerable scope for experimental and field work to ascertain

- whether or not all these processes actually operate under geological conditions to create the observed structures.
- to quantify their controlling parameters, e.g. degree of dilation needed to suspend the fault breccia, the amount and velocity of fluid input required to remobilise the fault breccia and the relative ratio of fines to breccia clasts required to produce a colloid.
- to determine if these textures are actually particularly unusual or if they have merely been ignored or unnoticed in the past. That this is a possibility is demonstrated by the fact that the Kimmeridge textures are only here identified as being unusual despite significant work having been done on the thrusts where they occur.

Further study may quantitatively determine what the controlling factors are on the development of such textures. In this thesis the role of very high fluid pressures is emphasised and is regarded by the present author as the most significant factor. The modeling of Rudnicki and Chen has demonstrated that even during the pressure drops associated with failure the fluid pressure in the reservoir and in the dilatant zone must not drop below the level where the fluid pressure relieves a significant amount of the imposed stress. It may be that these textures are intrinsically related to compressive stress regimes because the magnitude of fluid pressures and stress fluctuations required to achieve appropriate conditions for fluidisation and colloid formation may only be obtainable in a compressive environment. The field observations upon which this thesis is based were all made in thrust systems, but the supporting observations made in California were made in a transpressive strike-slip regime and, although as noted earlier, these textures have been observed in strike slip regimes, to the authors knowledge, no such textures have been documented in extensional or transtensional settings.

It is also possible that a mechanical factor that enhances dilation, such as the mechanically weak area of brecciated rock at Rusey or the very high competency

contrast between the dolomite and shales at Kimmeridge, is required before a sufficient degree of dilation will occur. This would severely limit the number of structural situations in which these processes can operate to either a very small number of faults or to a very limited portion of a given fault plane or zone.

Quantitative assessment of any of these three factors would require extensive further field work and/or lab work.

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Acknowledgments

There have been many more people than are named here who helped, advised and encouraged me en route to and during my PhD, thank you for your help and your faith in me.

For help with my PhD, on the technical side, I thank John Hutchinson, Mark Thompson, Jim Dare and Peter Freemantle for help with equipment, rock cutting and making thin sections and Alan Jones, for being the man who arranges and organises, also Dick Campaign and Paul Grant for arranging my grant support and Paul Suddaby and Ella Hin for looking after the computers. On the academic side, I'd like to thank Joe Cartwright, Jamie Wilkinson, Chris Halls, Paul Garrard, Mike Coward and Doug Shearman for their help, advice and criticisms over the past few years. I also need to thank the structural group, past and present, Jacqui Booth, Matt Brown, Dave Thomas, Barney Philps, Salim al-Mahruqi, Simon Evans, James Cotton, Ed Blunt, Paul Wheeler for all the help and arguments and for being around to bounce ideas off and drink with. Jacqui, in particular, deserves gratitude and sympathy, together with Becky, Howard and Katie for being my housemates and for frequent cat-sitting. Once again Jacqui is thanked, along with Judi Lakin, Fran van Wyk de Fries and Alick Leslie, for risking life and limb as my field assistants. I thank Pete Connolly for his help and advice, for perfecting the photo-elastic modelling technique which provided the supporting evidence for my structural model and for editing the first part of this thesis, Dave Dewhurst for offering many pungent criticisms of my geology and thought processes and for more thesis editing and Richard Jolly, for writing the analysis programs and letting me abuse his computer and for yet more thesis editing. Also I'd like to thank Matt Brown for endless patience and the unstinting use of his computer and printer and Rachel Leslie for proof reading several chapters.

For the social life, alcohol and sympathy, I have to thank the above, plus Sarah, Louise, John, Dick, Kate, John, Kerry, James, Lidia, Gary, Craig, Barry, Scoff, Sally, Matt, Helen and the many other permanent and floating members of the Southside gang.

Most importantly I'd like to thank my supervisor, John Cosgrove, for taking me on in the first place, helping me find the grant, and for being a great supervisor. I want to

offer my heartfelt thanks for all I've learnt from him in the past few years, for his support and encouragement when things didn't work and for miraculously keeping his patience with me.

This PhD was funded by an Imperial College Research Bursary, which is gratefully acknowledged and the thesis was printed at Midland Valley Exploration Ltd, the use of whose facilities is very much appreciated.

There are a number of people to whom I owe personal thanks, Chris and Dom, Rachel and Alick, Fran and Ben, Sarah and Richard. Without these people I would have given up and I shall always be indebted to them.

Finally, this thesis is dedicated to my mother, in gratitude.