DEPOSITIONAL SYSTEMS OF THE PERMIAN VRYHEID FORMATION HIGHVELD COALFIELD, SOUTH AFRICA. THEIR RELATIONSHIP TO COAL SEAM OCCURRENCE AND DISTRIBUTION.

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DECLARATION

I declare that this dissertation is my own, unaided work. It is being submitted for the degree of Doctor of Philosophy in the Faculty of Science, University of the Witwatersrand, Johannesburg. It has not been submitted before for any degree or examination in any other University.

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This 21st day of February, 1995.

ABSTRACT

The Permo-Carboniferous Dwyka Group and Vryheid Formation of the Karoo Sequence in the Highveld Coalfield are analysed to determine the depositional systems operative during sedimentation. The investigation involves the processing of 629 borehole cores and logs in an area of 860km².

A genetic stratigraphy based on depositional sequences is constructed for the Highveld Coalfield. Each depositional sequence comprises a progradational phase of sedimentation followed by an aggradational phase of sedimentation and is terminated by a thin veneer of transgressive sediments. Five depositional sequences are identified and termed the No. 1, 2, 4, 5, and 6 seam depositional sequences. A wide spectrum of facies types are recognised and range from diamictite, through conglomerate, sandstone and siltstone, to coal.

Glacial, fluvial, deltaic, and shallow-marine depositional systems are interpreted. Glacial deposits represented by lodgement till, and glaciolacustrine deltaic and glaciofluvial sediments, constitute the No. 1 seam depositional sequence. The Nos. 1, 2 and 4 seam depositional sequences have facies associations interpreted as low-sinuosity, coarse-grained, bed-load dominated, fluvial systems. Two styles of deltaic sedimentation are documented namely, mouth-bar deltas and "Gilbertian" deltas. These deltaic deposits constitute the No. 2 to No. 3 seam interval, and the No. 5 and No. 6 seam depositional sequences respectively. Transgressive shallow-marine deposits overlie the No. 2, 4, 5, and 6 seam depositional sequences and represent the partial retention of microtidal barriers, shoreface deposition and shallow-marine sand ridges. The linking of sedimentary depositional systems to a well defined stratigraphic framework facilitated the compilation of the depositional architecture for each depositional sequence and portrays the changing styles of sedimentation with time.

Within the study area, palaeotopography, differential subsidence, and contemporary sedimentation and the depositional system present during peat accumulation, are important factors controlling the distribution and thickness of coal seams. Ash content, climate, vegetation, lithotype profiles and tectonic subsidence of Permo-Carboniferous coal deposits are reviewed. Tectonic setting which controls subsidence rates of peat swamps, and the position of coal seams within a foreland basin, are considered important factors influencing both the ash and maceral content of coal seams on a regional scale.

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TABLE OF CONTENTS

CHAPTER 1 1

1 INTRODUCTION 1

1.1 PREAMBLE 1

1.2 LOCATION 2

1.3 REGIONAL GEOLOGY 2

1.4 PREVIOUS WORK 11

1.4.1 Introduction 11

1.4.2 Stratigraphic Studies 11

1.4.3 Interpretive Studies 12

1.4.4 Applied Studies 15

1.5 OBJECTIVES AND METHODOLOGY 17

1.5.1 Objectives 17

1.5.2 Methodology 18

CHAPTER 2 20

2 STRATIGRAPHY 20

2.1 INTRODUCTION 20

2.2 GENETIC STRATIGRAPHY 23

2.2.1 Introduction 23

2.3 No. 1 SEAM DEPOSITIONAL SEQUENCE 31

2.4 No. 2 SEAM DEPOSITIONAL SEQUENCE 36

2.5 No. 4 SEAM DEPOSITIONAL SEQUENCE 42

2.6 No. 5 SEAM DEPOSITIONAL SEQUENCE 53

2.7 No. 6 SEAM DEPOSITIONAL SEQUENCE 66

2.8 INTERVAL ABOVE THE No. 6 SEAM 68

CHAPTER 3 71
3 FACIES DESCRIPTION AND INTERPRETATION . . . 71
3.1 FACIES CONCEPT 71
3.2 SEDIMENTARY FACIES 74
3.2.1 Diamictite Facies 74
3.2.2 Conglomerate Facies 76
3.2.3 Massive to Poorly Stratified Conglomerate . . . 77
3.2.4 Planar Cross-Bedded Granule Conglomerate . . . 80
3.2.5 The Sandstone Facies 82
3.2.6 Massive Sandstone 83
3.2.7 Planar Cross-Bedded Sandstone 85
3.2.8 Trough Cross-Bedded Sandstone 87
3.2.9 Planar Laminated Sandstone 90
3.2.10 Cross-Laminated Sandstone 93
3.2.11 Interlaminated Sandstone-Siltstone Facies . . . 96
3.2.12 Bioturbated Sandstone and Siltstone Facies . . . 102
3.2.13 Gravelly Siltstone Facies 106
3.2.14 Carbonaceous Siltstone Facies 109
3.2.15 Carbonate Facies 111
3.2.16 Coal Facies 114
3.3 SUMMARY 119

CHAPTER 4 121
4 DEPOSITIONAL SYSTEMS 121
4.1 INTRODUCTION 121
4.2 FACIES INTERPRETED AS GLACIAL DEPOSITS . . 123
4.2.1 Description of Glacial Deposits 123
4.2.2 Interpretation of Glacial Deposits 125

4.3 FACIES INTERPRETED AS FLUVIAL DEPOSITS . . 129
4.3.1 Description: No. 1 and No. 2 Seam Depositional Sequences 129
4.3.2 Description: No. 3 to No. 6 Seam Clastic Sequence. . 133
4.3.3 Interpretation of Fluvial Deposits 141
4.4 FACIES INTERPRETED AS DELTAIC DEPOSITS . . 152
4.4.1 Introduction 152
4.4.2 Description: No. 2 to No. 3 Seam Sedimentary Interval . 152
4.4.3 Description: No. 5-No. 6 Seam Depositional Sequences . 155
4.4.4 Interpretation of Deltaic Deposits 160
4.5 FACIES INTERPRETED AS TRANSGRESSIVE MARINE DEPOSITS 177
4.5.1 Introduction 177
4.5.2 Description of Transgressive Marine Deposits . . 177
4.5.3 Interpretation of Transgressive Marine Deposits . . 182

CHAPTER 5 186
5 DEPOSITIONAL ARCHITECTURE 186
5.1 INTRODUCTION 186
5.2 No. 1 SEAM DEPOSITIONAL SEQUENCE . . . 186
5.3 No. 2 SEAM DEPOSITIONAL SEQUENCE . . . 191
5.4 No. 4 SEAM DEPOSITIONAL SEQUENCE . . . 194
5.5 No. 5 SEAM DEPOSITIONAL SEQUENCE . . . 204
5.6 No. 6 SEAM DEPOSITIONAL SEQUENCE . . . 207

CHAPTER 6 212
6 COAL FORMATION AND DISTRIBUTION 212
6.1 INTRODUCTION 212
6.2 PEAT AND COAL FORMATION 213

4.3 FACIES INTERPRETED AS DELTAIC DEPOSITS . . . 215

4.3.1 Introduction: No. 1 and No. 2 Seam Depositional Sequences 215

4.3.2 Description: No. 3 to No. 4 Seam Clastic Sequence . . 216

4.3.3 Interpretation of Fluvial Deposits 218

4.4 FACIES INTERPRETED AS DELTAIC DEPOSITS . . . 220

4.4.1 Introduction: Seam 222

4.4.2 Description: No. 3 to No. 3 Seam Sedimentary Interval . 223

4.4.3 Description: No. 5-No. 6 Seam Depositional Sequences . 227

4.4.4 Interpretation of Deltaic Deposits 231

4.5 FACIES INTERPRETED AS TRANSGRESSIVE MARINE . 231

4.5 DISCUSSION: APPLICABILITY OF COAL DEPOSITIONAL

4.5.1 Introduction 239

4.5.2 Description of Transgressive Marine Deposits . . 239

4.5.3 Interpretation of Transgressive Marine Deposits . . 240

6.5.3 Climate and Flora 245

CHAPTER 5 6.5.4 Vertical Succession of Coal Facies 248

5 DEPOSITIONAL SYSTEMS 252

5.1 Introduction and Subsidence 255

5.2 No. 1 SEAM DEPOSITIONAL SETTING

CHAPTER 7 No. 2 SEAM DEPOSITIONAL ENVIRONMENT . . . 261

7 SUMMARY AND CONCLUSIONS 261

7.1 INTRODUCTION 261

7.2 STRATIGRAPHY 262

7.3 SEDIMENTARY FACIES 264

7.4 DEPOSITIONAL SYSTEMS 265

7.5 DEPOSITIONAL ARCHITECTURE 268

7.6 COAL FORMATION AND DISTRIBUTION . . . 270

REFERENCES 275

6.3 COAL THICKNESS AND DISTRIBUTION 216
6.3.1 Introduction 216
6.3.2 No. 1 Seam 216
6.3.3 No. 2 Seam 218
6.3.4 No. 3 Seam 220
6.3.5 No. 4 Lower Seam 222
6.3.6 No. 4 Upper Seam 225
6.3.7 No. 5 Seam 227
6.3.8 No. 6 Seam 231
6.4 COAL SEAM FACIES PROFILES 231
6.5 DISCUSSION: APPLICABILITY OF COAL DEPOSITIONAL MODELS 239
6.5.1 Introduction 239
6.5.2 Ash Content 240
6.5.3 Climate and Flora 245
6.5.4 Vertical Succession of Coal Facies 248
6.5.5 Depositional Systems 252
6.5.6 Tectonics and Subsidence 255

CHAPTER 7 261
7 SUMMARY AND CONCLUSIONS 261
7.1 INTRODUCTION 261
7.2 STRATIGRAPHY 262
7.3 SEDIMENTARY FACIES 264
7.4 DEPOSITIONAL SYSTEMS 265
7.5 DEPOSITIONAL ARCHITECTURE 268
7.6 COAL FORMATION AND DISTRIBUTION 270

REFERENCES 275

LIST OF ILLUSTRATIONS

Figure 1.1. The distribution of the Karoo Sequence within the Karoo Basin and elsewhere in Southern Africa. 3

Figure 1.2. Locality map of the Transvaal Coalfields and position of the study area. 4

Figure 1.3. The distribution of boreholes used in this study. 5

Figure 1.4. Cross-section through the Karoo Basin illustrating the stratigraphic relationships and position of the coal-bearing strata in the basin (modified after Cadle *et al.*, 1993). 7

Figure 2.1. The concept of depositional events and a depositional episode (modified after Frazier, 1974). 25

Figure 2.2. Type 1 sequences and systems tracts deposited in a basin with a shelf break (after, Van Wagoner *et al.*, 1988). 27

Figure 2.3. Type 2 sequences and systems tracts deposited in a basin with a shelf break (after, Van Wagoner *et al.*, 1988). 28

Figure 2.4. The stratigraphy and depositional sequences of the Vryheid Formation in the northwestern Highveld Coalfield. Columns (A) and (B) illustrate the variation in stratigraphy. 30

Figure 2.5. Locality map showing the positions of stratigraphic cross-sections. . 32

Figure 2.6. Stratigraphic columns of the No. 1 and No. 2 seam depositional sequences. Columns (A) and (B) illustrate the variation in stratigraphy for this sedimentary interval. 33

Figure 2.7. Westeast cross-sections of the stratigraphic interval between basement and the No. 2 seam. See Fig. 2.5 for positions of cross-sections. 34

Figure 2.8. Northsouth cross-section of the stratigraphic interval between basement and the No. 2 seam. See Fig. 2.5 for the position of the cross-section. 35

Figure 2.9. Isopach of the Dwyka Group within the study area. 37

Figure 2.10. Isopach of the No. 1 seam illustrating the lateral impersistence of the seam. The terms "thick and thin" are relative terms describing thickness variations of the seam. 38

Figure 2.11. Isopach of diamictite and sandstone between basement and the No. 2 seam. The sediments thin against basement lithologies in the west (see Figure 2.7). 40

Figure 2.12. Isopach of the No. 2 seam. The seam thins and pinches out in the west and reaches a maximum thickness in the north. The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam. 41

Figure 2.13. Isopach of the clastic parting within the No. 2 seam. The parting displays spay-like geometries and is largely confined to the western portion of the study area. 43

Figure 2.14. Stratigraphic columns of the No. 4 seam depositional sequence. Profiles (A) and (B) illustrate the variation in stratigraphy for this depositional sequence. 44

Figure 2.15. Isopach of the No. 2 to No. 4 Lower seam sedimentary interval. Note the north-south linear thickening of this interval. 45

Figure 2.16. Westeast cross-sections of the No. 4 seam depositional sequence. Note the complex splitting of the No. 4 seam and the lithological variation constituting the seam splits. See Fig. 2.5 for the location of cross-sections. 46

Figure 2.17. Northsouth cross-section of the No. 4 seam depositional sequence. The depositional sequence thickens to the south. See Fig. 2.5 for the location of the cross-section. 47

Figure 2.18. Isopach of the No. 2 to No. 3 seam sedimentary interval. The interval thickens to the east and thins towards, and pinches out against, basement in the west. 49

Figure 2.19. Isopach of the No. 3 seam. The seam is thin and laterally impersistent. The terms "thick and thin" are relative terms describing thickness variations of the seam. 50

Figure 2.20. Isopach of the No. 3 to No. 4 Lower seam interval. The interval thickens to the south. 51

Figure 2.21. Isopach of the No. 4 Lower-No. 4 Upper seam interval. Note the northsouth linear thickening of the interval. 52

Figure 2.22. Isopach of the No. 4 Lower seam. The seam thins and is absent in the southwest. There is also a linear thinning of the seam in the southern part of the study area. The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam. 54

Figure 2.23. Isopach of the No. 4 Upper seam. The seam thickens either side of the underlying No. 4 Lower-No. 4 Upper seam clastic parting (See Fig. 2.21). The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam. 55

Figure 2.24. Isopach of the clastic parting within the No. 4 Lower seam. Note the splay-like geometries of the parting. 56

Figure 2.25. Isopach of the clastic parting within the No. 4 Upper seam. Note the splay-like geometries of the parting. 57

Figure 2.26. Composite stratigraphic profiles of the No. 5 and No. 6 depositional sequences and the interval above the No. 6 seam. Profiles (A) and (B) illustrate the stratigraphic variations. 59

Figure 2.27. Westeast cross-sections of the No. 5 and No. 6 depositional sequences and the interval above the No. 6 seam. See Fig. 2.5 for the location of cross-sections. 60

Figure 2.28. Northsouth cross-section of the No. 5 and No. 6 seam depositional sequences and the interval above the No. 6 seam. See Fig. 2.5 for the location of the cross-section. 61

Figure 2.29. Isopach of the No. 5 seam depositional sequence. Note the thinning of the sequence to the west and thickening of the sequence to the south. 62

Figure 2.30. Isopach of the siltstone below the No. 5 seam. The siltstone is between 5m and 10m thick over much of the study area. 63

Figure 2.31. Isopach of the sandstone between the No. 5 seam and the underlying siltstone. The sandstone thins towards the west and thickens eastwards. 64

Figure 2.32. Isopach of the No. 5 seam. The seam is thin and laterally persistent. The terms "thick and thin" are relative terms describing thickness variations of the seam. 65

Figure 2.33. Isopach of the No. 6 seam depositional sequence. The sequence thickens eastwards. 67

Figure 2.34. Isopach of the No. 6 seam. The seam is laterally impersistent and thin. The terms "thick and thin" are relative terms describing thickness variations of the seam. 69

Figure 3.1. The diamictite facies comprising a siltstone matrix and extrabasinal pebbles. Note the faceting of pebbles. Figures (A) and (B) illustrate facies variations. 75

Figure 3.2. The massive to poorly stratified conglomerate facies comprising granule- grade quartz and feldspar grains (A). The massive to poorly stratified pebble conglomerate with a high matrix content (B). . . . 78

Figure 3.3. The cross-bedded granule conglomerate facies comprising well sorted quartz and feldspar grains. The facies is planar cross-bedded. . 81

Figure 3.4. The massive sandstone facies. (A) illustrates a well sorted medium-grained sandstone. (B) illustrates a massive sandstone with a high siltstone matrix. 84

Figure 3.5. The planar cross-bedded sandstone facies. The fining-upward foresets define the cross-bedding. 85

Figure 3.6. The trough cross-bedded sandstone facies defined by scoured surfaces overlain by concave-upward laminae. 88

Figure 3.7. The planar-laminated sandstone facies. Note the well defined horizontal to subhorizontal laminae. 91

Figure 3.8. The cross-laminated sandstone facies. (A) Wave-ripple laminated sandstone displaying both convex and concave foresets. (B) Cross-lamination produced by current ripples. 94

Figure 3.9. The flaser-laminated sandstone facies. Laminations are defined by siltstone drapes. 97

Figure 3.10. Interlaminated sandstone and siltstone facies displaying wavy lamination. 99

Figure 3.11. The lenticular-laminated siltstone facies. Lenses of sandstone within siltstone define the facies. 100

Figure 3.12. The bioturbated sandstone facies. Both *Siphonichnus* and *Skolithos* traces are present. 103

Figure 3.13. The bioturbated siltstone facies. The intense bioturbation is probably due to the high organic content within the siltstone. Note the well preserved *Siphonichnus* traces. 104

Figure 3.14. The gravelly siltstone facies. Note the dispersion of gravel grains in a siltstone matrix. 107

Figure 3.15. The carbonaceous siltstone facies. 110

Figure 3.16. The limestone facies. X-ray diffraction analysis confirmed calcite as the carbonate mineral constituting this facies. Note the development of large calcite crystals in the middle of the core. 112

Figure 3.17. The coal facies. Bright bands represent vitrinite-rich macerals and dull bands inertinite-rich macerals (No. 5 seam). 115

Figure 3.18. Rootlet remains beneath the No. 4 Lower seam. 118

Figure 4.1. Facies associations interpreted as glacial deposits. The facies associations comprise diamictite overlain by bioturbated siltstone, gravelly siltstone, bioturbated sandstone and granule conglomerate (Figs. 4.1A and B). These sequences are often terminated by carbonaceous siltstone (Fig. 4.1B) and/or coal. 126

Figure 4.2. Idealised profiles through the No. 1 and No. 2 seam clastic sequences. Note the fining upward facies arrangements which constitute channel-fill deposits. 130

Figure 4.3. Core of the No. 2 seam depositional sequence. Note the multistorey nature of the sequence and the general decrease in grain size upwards. 132

Figure 4.4. Core of the sandstone-dominated interval between the No. 3 and No. 4 coal seams. Note the erosive base to the sandstone, the fining-upward facies assemblages, and the dominance of the planar cross-bedded sandstone facies. 135

Figure 4.5. Core sequence of the interval between the No. 4 Lower and No. 4 Upper seams. Note the vertically stacked fining-upward facies assemblages capped by the interlaminated sandstone-siltstone facies. Overall, the interval fines upward in grain size. 136

Figure 4.6. Core sequences of the No. 4 Lower-No. 4 Upper parting where the interval thickness is less than 3m. Note the small-scale fining-upward facies assemblages, illustrated by core sequences (A) and (B), reflecting conditions of fluctuating discharge. 137

Figure 4.7. Isopach of the sandstone between the No. 4 Lower and 4 Upper coal seams. The sandstone displays a linear geometry, varies between 2-10km in width and is up to 15m in thickness. 139

Figure 4.8. Isopach of the sandstone between the uppermost No. 4 seam and the siltstone at the base of the No. 5 seam depositional sequence. Note the northsouth linear thickening along the western margin of the study area and the eastwest linear thickening in the south of the study area. . 140

Figure 4.9. Westeast and northsouth cross-sections through the No. 4 Lower/No. 4 Upper seam fluvial channel. The cross-sections illustrate profiles of the down-channel and cross-channel fills. Note the well defined fining-upward sequences which may represent third-order channel-fill sequences. 148

Figure 4.10. Cross-section of the No. 4 Lower-No. 4 Upper seam sedimentary interval. The cross-section illustrates the east-west variation in sedimentary styles from, in-channel, through channel-margin, to flood-basin subenvironments. 151

Figure 4.11. Borehole core of the No. 3 seam depositional sequence. The sequence comprises two coarsening-upward facies associations commencing with bioturbated siltstone and terminating with wave-ripple laminated sandstone. 154

Figure 4.12. Borehole core of the No. 5 seam depositional sequence. The sequence displays a classic coarsening-upward facies association. The cross-bedded coarse-grained sandstone distinguishes this sequence from the No. 3 seam sequence. 158

Figure 4.13. A highwall exposure of the No. 5 seam depositional sequence at Rietspruit Coal Mine. Note the tops of the compound foresets are steeply dipping. The foresets flatten out asymptotically in a basinward direction. Also illustrated is the coarsening-upward facies association of this depositional sequence. 159

Figure 4.14. Borehole core illustrating the sandstone-dominated coarsening-upward facies association of the No. 6 seam depositional sequence. Note, cross-bedding is the dominant sedimentary structure. . 161

Figure 4.15. Conceptual framework for a comparative study of deltas (after Elliot, 1986). 163

Figure 4.16. Classification of deltas prograding into low-energy basins (after Postma, 1990). 167

Figure 4.17. Core sequence illustrating the coarsening-upward facies association representing transgressive shallow-marine deposits. Note the *Skolithos* and *Siphonichnus* bioturbation of the interlaminated siltstone/sandstone and sandstone facies. 179

Figure 4.18. Core sequence of the transgressive shallow-marine deposit. The fluvial sequence overlying the coal seam is truncated by a 10cm thick granule conglomerate which in turn grades upwards into wave-ripple and trough cross-bedded glauconitic sandstone. 180

Figure 4.19. Core sequence of the highly glauconitic shallow-marine sandstone. The sandstone is structured by both wave-ripple lamination and trough cross-bedding (top). 181

Figure 5.1. The reconstruction of basement topography in the study area. . 188

Figure 5.2. Palaeogeography of the No. 1 seam depositional sequence after the retreat of the continental Dwyka glaciers. Note the development of a glaciolacustrine environment in the southeast and the deposition of glacial till in topographically low-lying areas. 189

Figure 5.3. Palaeogeography of the No. 1 seam depositional sequence, illustrating the construction of a delta plain upon which the No. 1 seam peat accumulated. The No. 1 seam is mainly present in the southeastern part of the study area. 190

Figure 5.4. Transgression of basinal waters terminated No. 1 seam peat accumulation. Note the position of basinal waters at maximum transgression. Reworked glacial sediment was subsequently transported and deposited by bed-load dominated fluvial systems and Gilbert-type deltas into an embayment situated along the eastern portion of the study area. 192

Figure 5.5. Palaeogeography of the No. 2 seam depositional sequence. Continued progradation and aggradation of fluvial and deltaic systems led to the development of an extensive outwash plain. The high-lying terrain situated in the north and east was still subaerially exposed. . 193

Figure 5.6. Cessation of coarse-clastic sedimentation allowed vegetation to proliferate. The No. 2 seam peat then accumulated. Note the pinch out of the No. 2 seam peat against palaeohigh areas situated in the west. . 195

Figure 5.7. Transgression of basinal waters over the No. 2 seam peat, terminating sedimentation of the No. 2 seam depositional sequence. . 196

Figure 5.8. Palaeogeography of the No. 3 seam interval illustrating the progradation of shallow-water, mouth-bar deltas into the basin. . . 198

Figure 5.9. The accumulation of the No. 3 seam peat on the abandoned delta platform. Note the sporadic development of No. 3 seam peat in the area. . 199

Figure 5.10. Palaeogeography of crevasse-splay sedimentation following No. 3 seam peat accumulation. 200

Figure 5.11. Palaeogeography during No. 4 Lower seam peat accumulation. Note a large bed-load dominated fluvial channel traversed across the swamp during the final stages of peat accumulation. . . . 201

Figure 5.12. Palaeogeography during, and subsequent to, No. 4 Upper seam peat accumulation. Renewed fluvial deposition followed No. 4 Upper seam peat accumulation. Limited compaction over the underlying fluvial channel resulted in subsequent fluvial sedimentation being offset to the west. 203

Figure 5.13. Palaeogeography at the termination of the No. 4 seam depositional sequence. Microtidal barrier-island systems transgressed northwards as a consequence of increased accommodation in the basin. . . . 205

Figure 5.14. Deposition of the No. 5 seam deltaic systems. Sediment was deposited by shallow-water Gilbert-type deltas characterised by steep delta-front profiles. The deltas were fed by bed-load dominated braided distributaries. The No. 5 seam peat accumulated upon the delta plain. . . 206

Figure 5.15. Transgression terminated No. 5 seam peat accumulation. Retreating microtidal barrier-island systems coupled with wave and tidal [illegible] deposited a series of shallow-marine sand ridges in the southern [illegible] of the study area. 209

Figure [illegible] Palaeogeography of the No. 6 seam deltaic systems. Gilbert-type deltas fed by bed-load dominated distributaries prograded basinwards and built a platform upon which the No. 6 seam peat accumulated. . . 210

Figure 5.17. Palaeogeography during No. 6 seam peat accumulation. Differential compaction coupled with renewed transgression of basinal waters resulted in the sporadic development of No. 6 seam peat. . . 211

Figure 6.1. Isopach of the No. 1 seam illustrating the distribution and thickness of the seam. The seam is generally thin and laterally impersistent. Low-lying terrain in the southeast, and palaeovalleys in the north of the study area were favourable sites for peat accumulation. . . . 217

Figure 6.2. Comparison of the distribution and thickness of the No. 2 seam (A) with the underlying sedimentary interval (B). Note the thinning of the seam in the west and the influence of the 20m-40m isopach of the underlying sediments on the thickness of the seam. 219

Figure 6.3. Comparison of the distribution and thickness of the No. 3 seam (A) with the isopach of No. 2 to No. 3 seam interval (B). Note the coincidence of the 20-30m isopach of the No. 2 to No. 3 seam interval with the presence of No. 3 seam coal. 221

Figure 6.4. Comparison of the distribution and thickness of the No. 4 Lower seam (A) with the isopach of the No. 2 seam to No. 4 Lower seam interval (B). Note the correlation between the north-south thinning of the seam and the 30m isopach of the No. 2 to No. 4 seam interval. . . . 223

Figure 6.5. Comparison of the isopach of the No. 4 Upper seam (A) with the No. 4 Lower/No. 4 Upper seam clastic parting (B). The seam thins over, and thickens either side of, the underlying No. 4 Lower/No. 4 Upper seam clastic parting. 226

Figure 6.6. Comparison of the No. 5 seam isopach (A), with the sandstone overlying the No. 4 Upper seam (B). The seam thins in those areas where the sandstone overlying the No. 4 Upper seam is >8m in thickness. 228

Figure 6.7. Comparison of the No. 5 seam isopach (A), with the deltaic sandstones underlying the No. 5 seam (B). Thick No. 5 seam coal is present either side of the fluvial channels (Fig. 6.6B) where the deltaic sandstone is between 5m and 10m thick. 230

Figure 6.8. Comparison of the No. 6 seam isopach (A), with the isopach of the No. 6 seam depositional sequence (B). Note the occurrence of the seam in many areas where the thickness of the sequence is between 30m and 40m. 232

Figure 6.9. Generalised coal lithotype profiles of the No. 1 to No. 6 coal seams. The Nos. 2, 4 and 5 seams display brighter lithotypes at the base of the seam which pass upward into dull lithotypes. The Nos. 1, 3 and 6 seams are thin and comprise alternating bright and dull lithotypes and shale. 234

Figure 6.10. Average ash analyses of the No. 2 coal seam from the Witbank Coalfield (after, Cadle et al., 1990).. 243

Figure 6.11. Cross-section through a foreland basin. Permo-Carboniferous coals form on both the orogenic and cratonic margins of foreland basins. Coals which form on the orogenic margin of these basins are generally vitrinite rich and coals which form on the cratonic margin of these basins are generally inertinite rich. 257

LIST OF TABLES

Table 1.1 The lithostratigraphic subdivision of the Karoo Sequence after S.A.C.S. (1980). 8

Table 4.1 Classification of the major-clastic depositional environments. After Galloway and Hobday (1983). 124

CHAPTER 1

1 INTRODUCTION

1.1 PREAMBLE

The importance of coal to South Africa's economy and to the energy requirements of South Africa cannot be over emphasized. It constitutes an estimated 88% of South Africa's energy requirements (van Zyl *et al.*, 1989). Total sales increased from 111,2 to 180,3 Mt during the period 1980-1992, and exports increased from 29,1 to 50,1 Mt over the same period (Mehliss, 1989; Tinney, 1993). During 1992 coal accounted for sales of R8,952,440,799 comprising 41% of non-gold mineral revenue (Schreuder, 1993). Coal is also the top revenue earner in terms of local mineral sales. It accounts for 55% of local mineral sales and generated revenue of R4,969,397,944 (Schreuder, 1993). In 1992, South Africa ranked third behind Australia and the U.S.A. in tonnes of coal exported and ranked fifth behind China, U.S.A., C.I.S. and India in terms of recoverable coal reserves (Tinney, 1993).

The present investigation focuses on determining the depositional systems of the Permian coal-bearing sediments within the Karoo Basin, South Africa. These depositional systems can be classified as process response models controlled by factors such as; source area, sediment type and rate of deposition, basin type and configuration, and tectonic and climatic factors. The depositional systems which result from the interaction of many variables frequently contain within them exploitable resources. In the case of this investigation the exploitable resource is coal, the most important revenue earner after gold. It is well known that coal type (maceral content), grade (degree of mineral matter), thickness and areal distribution are determined at the site of peat accumulation. Thus to understand these coal parameters it is imperative to

understand the depositional setting within which peat accumulates. The thrust of this investigation is the determination of the sedimentary framework within which the coal is situated and the resolution of the relationships between coal parameters and the depositional setting. The relationships between depositional systems and associated coal seams have application in either, future exploration or, mine planning.

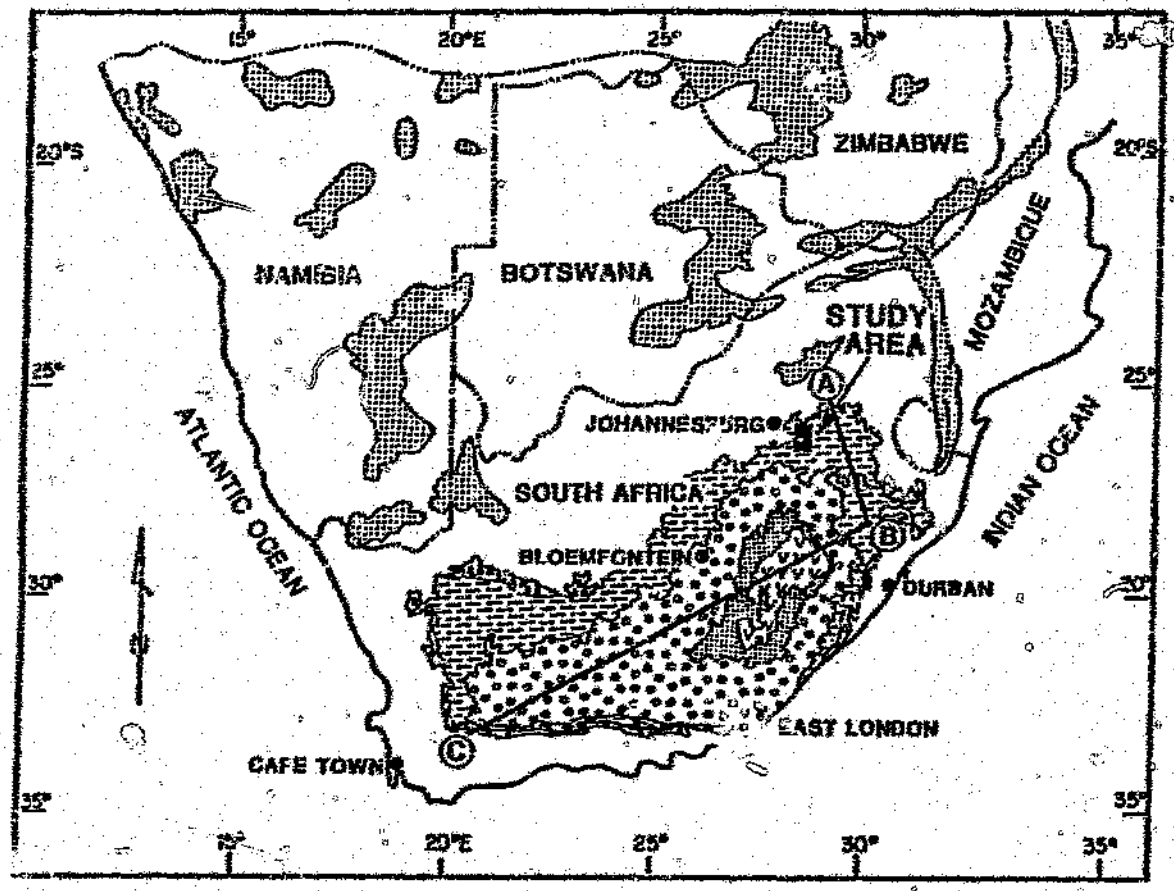
1.2 LOCATION

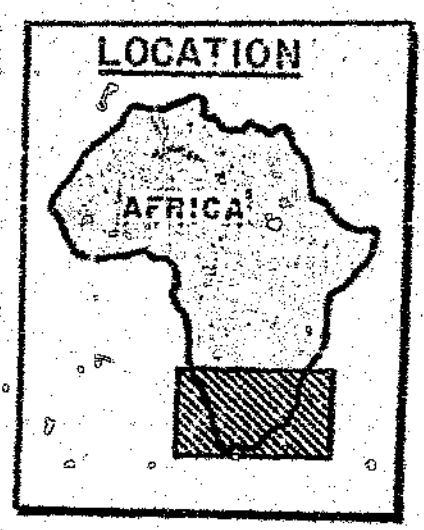
The study area is situated approximately 80km east-southeast of Johannesburg in the Eastern Transvaal Province of South Africa. The area is situated geologically within the northern portion of the Karoo Basin (Fig. 1.1) and more specifically within the northwestern Highveld Coalfield bordering on the Delmas and Witbank Coalfields. The geographic location of the study area and its relationship to the pre-Karoo geology is illustrated in Fig. 1.2.

The area of investigation covers about 860km^2 and was extensively core drilled for coal during a coal exploration programme. A total of 629 coreholes (Fig. 1.3) provide the data for the study, limiting the study to a subsurface analysis of the coal-bearing sediments. Restricted surface exposures in adjoining areas were used to supplement the data base.

1.3 REGIONAL GEOLOGY

The most complete assemblage of Carboniferous to Jurassic rocks in Southern Africa is preserved within the Karoo Basin which records a diverse assemblage of palaeoenvironmental settings. These range from glacial through submarine fan, slope, shelf, shoreline, deltaic, fluvial, lacustrine, to wet- and dry-desert environments. The





EXPLANATION

Karoo Basin

PERIOD	EPOCH	KAROO SEQUENCE		
JURASSIC				Lebombo
TRIASSIC				Stormberg Group
PERMIAN	LATE			Beaufort Group
	EARLY		Volksrust Formation / Vryheid Formation / Pietermaritzburg Formation	Ecca Group
CARBONIFEROUS				Dwyka Group

KAROO SEQUENCE ELSEWHERE IN SOUTHERN AFRICA

Figure 1.1. The distribution of the Karoo Sequence within the Karoo Basin and elsewhere in Southern Africa.

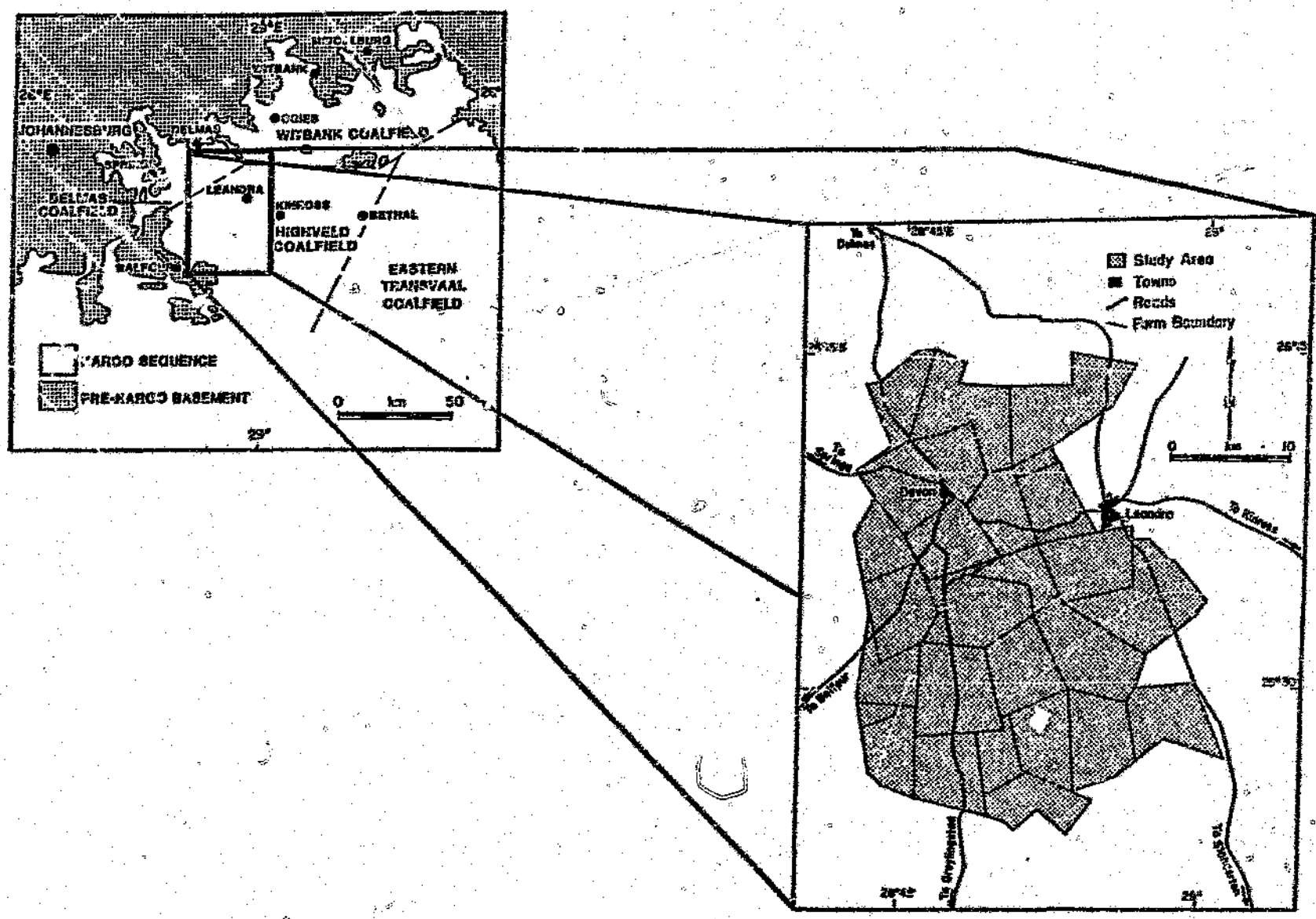
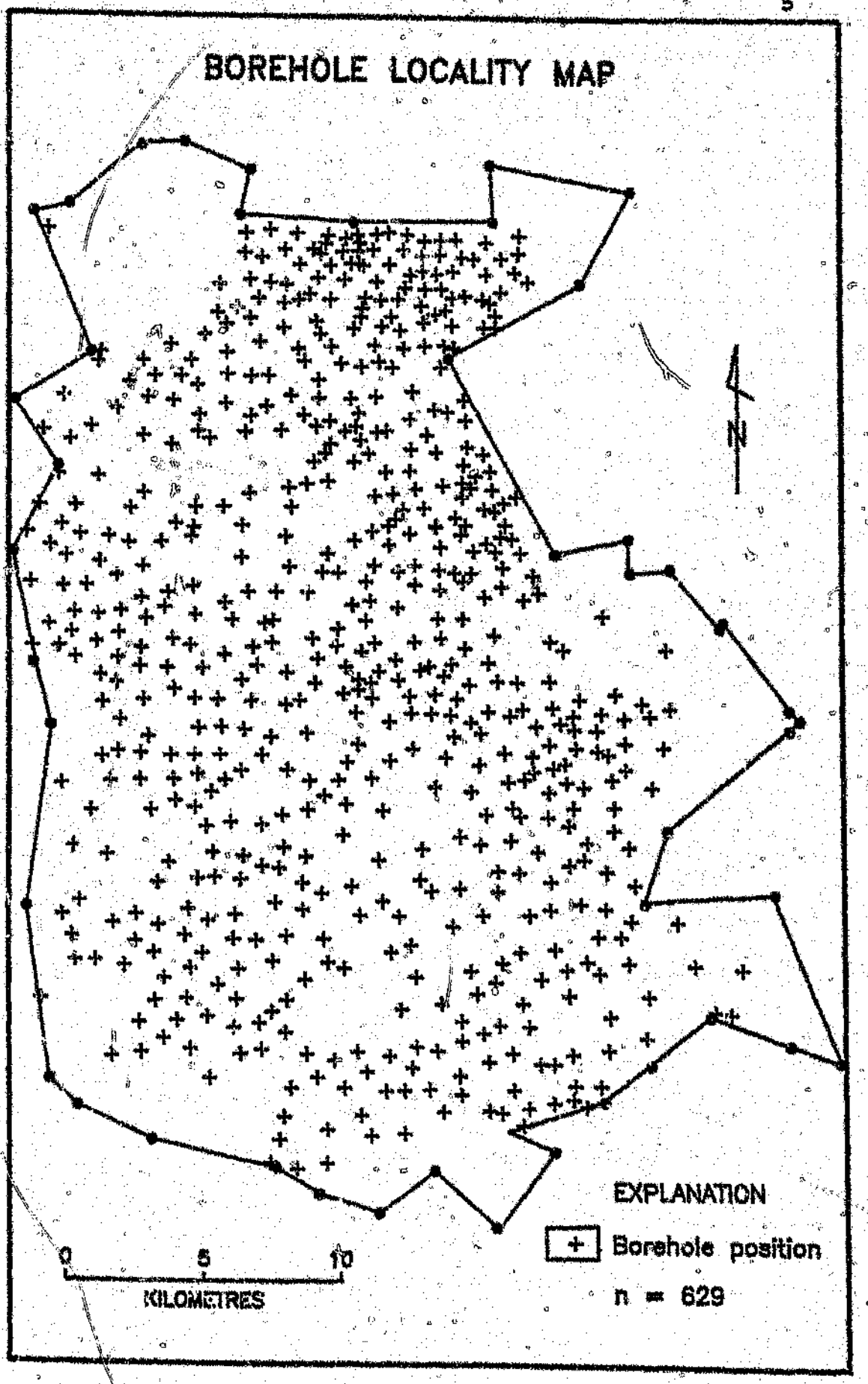


Figure 1.2. Locality map of the Transvaal Coalfields and position of the study area.



ure 1.3. The distribution of boreholes used in this study.

basin is asymmetrically filled (Fig. 1.4) and classified as a foreland basin (Allen *et al.*, 1986) as the depository formed between a mountain belt south of the basin, preserved today as the Cape Fold Belt, and the 3.0 Ga. Kaapvaal granitic craton to the north.

The rocks comprising the Karoo Basin are lithostratigraphically known as the Karoo Sequence (South African Committee for Stratigraphy, 1980). Johnson (1994) has recommended that the term Karoo Sequence be replaced by Karoo Supergroup however, this change in terminology has not been formalised. The term Sequence for the rocks within the Karoo Basin is used in this study. Within the Karoo Basin the rocks comprise from the base upwards the Dwyka Group, overlain by the Ecca and Beaufort Groups, which are in turn overlain by the Molteno, Elliot, and Clarens Formations (formally the Stormberg Group). For the purposes of this study these three formations are termed the Stormberg Group (see, Figs. 1.1, 1.2, and Table 1.1). This sedimentary succession is capped by the Lebombo Group volcanics (Table 1.1).

The Dwyka Group comprises diamictites, mudrocks and subordinate sandstones which thin toward the northern edge of the Basin (Fig. 1.4). The Ecca Group is dominated by a succession of grey mudrocks with subordinate sandstones. Due to differences in abundances of mudrock and sandstone, and in directions of sediment transport, the Ecca Group can be divided into three fill types. A southerly and southwesterly fill comprising largely mudrocks and fine-grained greywacke sandstones. This sequence of rocks was termed the southern and western Ecca facies by Ryan (1968). These sandstones shale-out northwards to form a thick fill of grey mudrocks which Ryan (1968) termed the central Ecca facies. In the northern portion of the basin the sedimentary fill was termed the northern Ecca facies by Ryan (1968). Here the Ecca Group which is the focus of this investigation, is subdivided into a lower Pietermaritzburg Shale Formation, a middle Vryheid Formation, comprising conglomerates sandstones grey mudrocks and economically exploitable coal seams, and an upper Volksrust Shale Formation (Table 1.1).

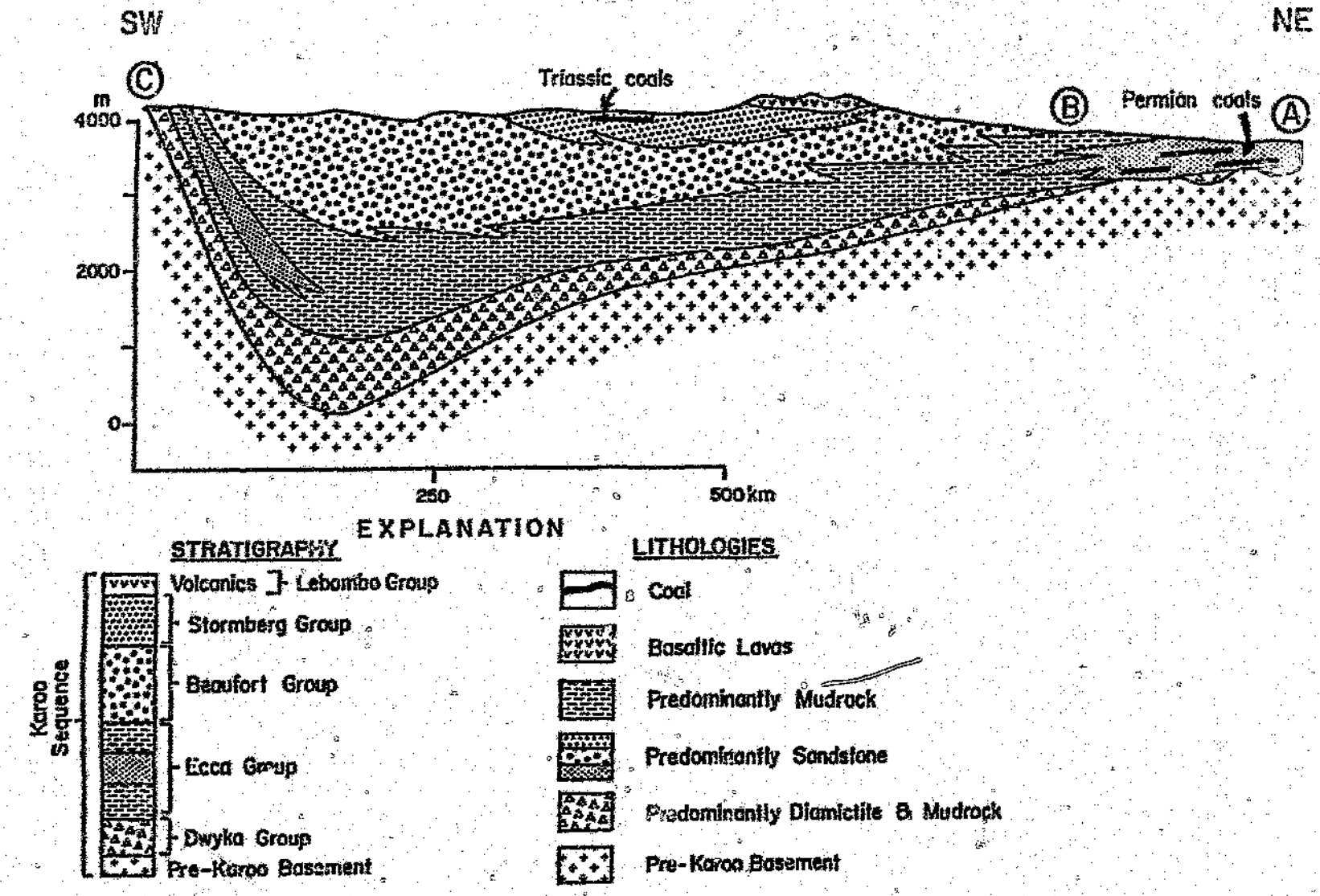


Figure 1.4. Cross-section through the Karoo Basin illustrating the stratigraphic relationships and position of the coal-bearing strata in the basin (modified after Cadle *et al.*, 1993).

Table 1.1 The lithostratigraphic subdivision of the Karoo Sequence after S.A.C.S. (1980).

SEQUENCE	GROUP	SUBGROUP	FORMATION			
			South Western Cape Province	Eastern Cape Province	Eastern Natal	Tshipise–Pafuri Area
KAROO	Lebombo			Drakensburg	Movene Jozini Letaba	Letaba
	Stormberg			Clarens Elliot Molteno	Clarens Nyoka Ntabene	Clarens Poabokpoort Klopperfontein
	Beaufort	Tarkastad		Burgersdorp Katberg		Solitude
		Adelaide	Teekloof Abrahamskraal	Balfour Middelton Koonap	Emakwezini	Fripp
	Ecca		Waterford Fort Brown Laingsburg Vischkuil Collingham Whitehill Prince Albert	(Waterford) Fort Brown Ripon Collingham Whitehill Prince Albert	Volksrust Vryheid Pietermaritzburg	Mikambeni Madzaringwe
	Dwyka					Tshidzi

The Beaufort Group is represented by a sequence of grey, green and maroon mudrocks alternating with fine-grained lenticular sandstones. The Molteno, Elliot and Clarens Formations (Stormberg Group) comprise coarse- to fine-grained sandstones and subordinate green and maroon mudrocks. The lithologies constituting the Beaufort Group and the latter three formations contain abundant fossil fauna and flora (Kitching, 1977; Kitching and Raath, 1984). The Lebombo Group sheet-flood basaltic rocks terminated Karoo sedimentation and record the final phase of basin infilling.

The rocks of the Karoo Sequence are only represented by the Dwyka Group and Vryheid Formation in the study area. This is due to the northward pinchout of the Pietermaritzburg Shale Formation, and the removal through erosion of the stratigraphic succession overlying the Vryheid Formation (Fig. 1.4). Sedimentation within the Karoo Basin began during the Carboniferous Period when continental ice sheets advanced and reated across the Karoo Basin. These sediments comprise subglacial tills, and proglacial conglomerates, sandstones, and laminated shales with dropstones (Visser, 1986; Visser and Loock, 1987). This assemblage of lithologies is termed the Dwyka Group. As Gondwanaland moved northwestward away from the south pole the climate began to ameliorate. As a consequence, by the Permian Period the ice-sheets had retreated to highlying terrains leaving behind a large water body termed the "Ecca sea".

The depository demarcated by the "Ecca sea" was filled by sediments sourced from both the southern and northern margins of the basin. At the time of Ecca deposition the southern basin margin formed a foredeep and the sedimentary fill comprised mudrocks and greywacke sandstones deposited in submarine fan, slope, shelf and deltaic environments (Johnson, 1976; Kingsley, 1981; Wickens 1989). Thus, the sedimentary rocks record the infilling and shallowing of the foredeep such that, by Beaufort times, fl· ·al systems were able to advance northwards across the basin. In the northern margin of the basin the sediments, which were sourced from the north, were deposited by proglacial, fluvial, deltaic, shoreline and shallow-marine depositional systems

(Cadle and Hobday, 1977; Vos and Hobday, 1977; Hobday 1978a; Le Blanc Smith and Eriksson, 1979). These sediments are termed the Dwyka Group, and the Vryheid Formation of the Ecca Group. During this phase of basin infilling, sedimentation took place in pulses. Initially, basin fill took place through progradation of sediments until equilibrium conditions were reached between subsidence and sediment supply. Thereafter, accommodation increased such that transgression by basinal waters occurred. A renewed period of progradation of sediments then followed transgression. Capping mc progradational sequences are laterally extensive coal seams reflecting that vegetation proliferated and optimum conditions for peat accumulation prevailed. Towards the end of the Permian Period the climate was warm temperate (Falcon, 1986) and the basin became tectonically more unstable. The change from relatively stable to unstable tectonic conditions accounts for the absence of coal seams towards the top of the Vryheid Formation and the presence of basinal mudrocks of the Volksrust Formation forming the uppermost unit of the Ecca Group.

The Beaufort Group, and the Molteno, Elliot and Clarens Formations, record a change to terrestrial depositional palaeoenvironments which was accompanied by a change in palaeoclimate fro warm temperate to arid (Falcon, 1986; Eriksson, 1981, 1984). This interval of sedimentation spans the Late Permian, Triassic and Jurassic Periods. In the northern Karoo Basin, the Beaufort sediments accumulated through the action of shallow-water deltas, and braided- and meandering-fluvial depositional systems, depositing their sediment load into lakes (Hobday 1978b; Turner, 1977; Turner, 1978; Van Dijk *et al.*, 1978; Stear, 1983). The Molteno Formation represents a wedge of sediment deposited by braided streams sourced from an area southeast of the present limits of the Karoo Basin (Turner, 1975, 1983). The overlying Elliot Formation comprises sediments deposited in a fluvial-floodplain environment subjected to periodic flooding separated by long periods of subaerial exposure (Turner, 1977; Turner, 1983; Eriksson, 1984). The Clarens Formation documents that the final phase of sedimentation took place in an extremely arid palaeoclimate. Sediments interpreted as

ephemeral stream, lacustrine, and aeolian deposits are documented by Eriksson (1981) and Beukes (1970) inferring that both wet-desert and dry-desert environments prevailed.

The outpouring of Jurassic basaltic lavas terminated sedimentation in the basin and heralded the initial phase of rifting which led to the break-up of Gondwanaland.

1.4 PREVIOUS WORK

1.4.1 Introduction

Geological investigations of the Karoo Sequence, and the Ecca Group rocks in particular, follow an evolutionary trend. The first investigations were essentially descriptive and stratigraphic in origin. From about 1970 to present the investigations have become interpretive and environments of deposition postulated. Over the past 15 years and particularly over the last 10 years applied studies relating to the coal-bearing Vryheid Formation have evolved.

1.4.2 Stratigraphic Studies

The initial geological investigations of the Karoo Sequence were stratigraphic in nature. The Dwyka Group is named after the diamictite (till) deposits present along the Dwyka River situated in the southern margin of the basin. Rubidge (1858) introduced the term Ecca to describe the dominantly mudrock succession exposed in the Ecca Pass in the eastern Cape. The term Beaufort is given to the alternating fine-grained sandstones and grey-green mudrocks present in the area around the town of Beaufort West. The Stormberg Group is named after rocks comprising coarse-glittering sandstones

(Molteno Formation), red shales (Elliot Formation) and thick, large-scale cross-stratified sandstones (Clarens Formation) exposed in the eastern Cape in the area of the Stormberg Mountains (Truswell, 1977; Tankard *et al.*, 1982).

In the northern Karoo Basin, the Ecca Series (Group) rocks were recognised to differ from those present in the southern margin of the basin. These rocks were termed the Lower, Middle and Upper Ecca Series (Du Toit, 19 , 1926) for what is now termed the Pietermaritzburg Formation, Vryheid Formation and the Volksrust Formation (S.A.C.S. 1980). The first detailed descriptions of the Vryheid Formation coal measures in the Transvaal Coalfield is recorded by Mellor (1906) who documented that the stratigraphy comprised conglomerates, coarse sandstones and minor shales which separated the Nos. 1, 2, 3, 4, and 5 coal seams. Some 34 years later, Blignaut and Furter (194 scribed in detail, the rocks exposed in northern KwaZulu/Natal. They divided the atigraphy into five zones; a lower transitional zone, a lower sandstone zone, a coal zone, and an upper sandstone and transitional zone. Later, Ryan (1968) informally named these zones formations. This subdivision by Blignaut and Furter (1940) of the Vryheid Formation is still used in coal studies (Roberts, 1986). In 1980 the South African Committee for Stratigraphy formally proposed that the Ecca rocks be given formation status and the Lower, Middle, and Upper Ecca Series be renamed the Pietermaritzburg Shale Formation, the Vryheid Sandstone Formation and the Volksrust Shale Formation in line with lithostratigraphic principles.

1.4.3 Interpretive Studies

Ryan (1968) in his basinwide study of the Ecca Group, subdivided the rocks into four facies namely, the northern, central, southern and western facies. This subdivision was based upon palaeocurrent data and recognisable differences in lithologies. Ryan (1968) was the first to document that the coal-bearing Ecca sediments were deposited by

fluvial and deltaic depositional systems. A very similar basinwide study was conducted by Stratten (1970) focusing on the Dwyka Group. This study analysed ice-flow directions and documented that ice advance and retreat took place from highlands situated north of the present coalfields.

The first detailed sedimentary model for the Vryheid Formation was undertaken by Hobday (1973) in the Tugela Ferry area of KwaZulu/Natal. This study was based upon the sedimentary facies approach to sedimentary modelling and, for the first time, interpreted that the rocks accumulated through progradation and switching of shallow-water, high-constructive deltas. Cadle (1974) working on the Ecca rocks from two separate areas namely, Groenvlei and Petdekop in the Eastern Transvaal, documented that the rocks comprised sediments deposited by both deltaic and fluvial depositional systems. The Vryheid Formation rocks overlying the Pietermaritzburg Shale Formation were deposited in a lower delta plain environment. These rocks were overlain by rocks deposited in an upper delta-plain/fluvial environment which were in turn overlain by rocks deposited in a lower delta-plain environment. Furthermore, Cadle (1974) identified the fact that the coal seams were associated with the sediments deposited by fluvial systems in upper delta-plain environments. Also documented was the north-south distribution of these environments and the southward thinning and pinchout of the Vryheid Formation.

Outcrop based studies undertaken by Mathew (1974), Hobday and Mathew (1975) and Hobday and Tavener-Smith (1975) in northern KwaZulu/Natal confirmed that the Vryheid Formation comprised a succession of sediments deposited by deltas overlain by sediments deposited by meandering fluvial systems. Hobday and Tavener-Smith (1975) were the first to identify trace fossil assemblages in the deltaic deposits which added a further refinement to the delta model. The earlier meandering fluvial model was expanded upon by later workers. In the Witbank Coalfield, Cairncross and Hobday (1979), Le Blanc Smith and Eriksson (1979) and Winter (1985) interpreted

anastomosed-fluvial deposits within the No. 2 and No. 4 coal seams. Le Blanc Smith (1980) also documented braided-fluvial systems and showed they gradually change in style downstream to become anastomosed. Later studies have shown that in the Witbank and Highveld Coalfields the dominant style of fluvial sedimentation is braided (Le Blanc-Smith, 1980; Cadle, 1982a; Winter, 1985; Cairncross, 1986; Cairncross and Cadle, 1988). A further variant to the braided-fluvial depositional system is documented from the base of the Vryheid Formation in the Witbank Coalfield. Here, overlying the till deposits, glaciolacustrine, glaciodeltaic and glaciofluvial sediments are described by Le Blanc Smith (1980), Winter (1985) and Cairncross (1986).

The recognition of shoreline and transgressive shallow-marine deposits has added to the fluvial and deltaic depositional systems first proposed. Vos and Hobday (1977) identified shoreline systems in Bothaville in the Orange Free State. In particular, lower and upper foreshore deposits were documented. In the Durban area of KwaZulu/Natal, Tavener-Smith (1982) described both shoreface and foreshore deposits and postulated a prograding shoreline complex overlain by back-barrier lagoonal sediments and thin coal seams. The identification of a variety of trace fossils from the Vryheid Formation has led to the refinement of depositional systems associated with the deltaic and transgressive shallow-marine environments. Studies of note are documented by Hobday and Tavener-Smith (1975), Stanistreet *et al.*, (1980), Turner *et al.*, (1981), Roberts (1986) and Christie (1988).

In summary, the depositional systems operative during Vryheid Formation deposition encompass a wide variety of terrestrial and transitional environments. These include; glaciolacustrine, fluvial, deltaic, shoreline and transgressive shallow-marine depositional systems.

1.4.4 Applied Studies

The interpretive studies undertaken during the 1970's provided a foundation for coal-based studies. During the 1980's a number of studies were published relating coal thickness, distribution, type and grade to depositional systems. Despite these recent studies, applied studies focusing upon the coal were undertaken during the early part of the century. These studies, which influenced thinking on coal formation for many years, did not have the hindsight of interpretive sedimentary techniques.

Mellor (1906) in the first really incisive study of the Transvaal Coalfield described the stratigraphy of the coal strata. He commented on the coarseness of sediment interbedded with the coal seams and suggested that rivers with "considerable force" transported the sediment. He also documented the presence of stone veins (the anastomosed-fluvial sandstones reported by Cairncross (1979, 1980) and Le Blanc Smith (1980)) and contemporaneous water channels (abandoned channel fills in the No. 2 seam). The drift origin of the coal was questioned due to; the lateral continuity and thinness of the No. 3 seam, the thin low-ash bands within the No. 2 seam and the presence of root markings at the base of the seam. Wybergh (1922) and Du Toit (1926) recognised that the coals formed from the *in situ* growth of vegetation and that adjacent to basement highs, type and grade of coal varied considerably. An important finding still applicable today is that they noted the base of coal seams were "purer" than the top of seams. The autochthonous origin for peat accumulation was further supported by Plumstead (1957, 1966). She identified *Glossopteris* and *Gangamopteris* floras which represented in excess of 90% of swamp vegetation.

The first coal depositional models began to evolve in the 1970's when Cadle (1974) identified thick, laterally persistent coal seams associated with sediments deposited by fluvial systems and thin laterally impersistent coals with sediments deposited in lower delta-plain environments. The importance of fluvial systems associated with thick

economically exploitable coal has been substantiated by Cadle and Hobday (1977), Steyn and Beukes (1979), Cairncross (1979, 1986), Le Blanc Smith (1980), Cadle (1982a), Winter (1985) and Cairncross and Cadle (1988).

Le Blanc Smith (1980) undertook a detailed study of the coal-bearing sedimentary succession in the Witbank Coalfield and evolved a genetic stratigraphy within which detailed studies of the coal-bearing sequences could be facilitated. He also noted the importance of palaeotopography as a controlling factor influencing coal type, grade, thickness and distribution for the No. 2 coal seam. The importance of clastic partings splitting coal into subseams through deposition of sediments by either, braided or, anastomosed fluvial systems is documented by Cairncross (1980), Le Blanc Smith (1980), Cadle (1982b), Winter (1985) and Cairncross (1986). Ash and calorific value data from coal seams adjacent to these fluvial channels reveal the deleterious influence of these channels on coal quality. In general, the coal seams adjacent to the channel deposits have relatively high-ash and low-volatile values. Conversely, the coal seams situated away from channel deposits have low-ash and high-calorific values.

Tectonic factors have also been documented as influencing coal thickness and maceral content. Cadle (1982a, 1982b, 1986) recognised excess differential compaction of underlying sediments as a factor causing the shale out of the No. 5 coal seam. Cadle (1982a) also postulated that stable tectonic conditions prevailed during the accumulation of the thick No. 2 and No. 4 seams and that unstable tectonic conditions prevailed in post No. 4 seam times resulting in the development of thin seams. More recently, a collaborative investigation of the No. 2 seam and underlying and overlying strata in the Witbank Coalfield was undertaken (Cadle *et al.*, 1989a). The study documented the depositional framework of No. 2 seam peat accumulation, the petrographic, palynological, geochemical and proximate and ultimate data of the seam. Palynological data confirm an Early Permian age for peat accumulation and petrography combined with proximate and ultimate data classify the seam, in terms of utilization, as a

high-grade bituminous steam coal. Mineralogically, kaolinite is the major ash mineral within the seam, specific trace elements are concentrated in the organic and inorganic fractions of the coal and high phosphorous concentrations are situated at the top of the seam.

1.5 OBJECTIVES AND METHODOLOGY

1.5.1 Objectives

Flat to undulatory topography, physiographically termed the "Highveld" typifies the landscape within which the Witbank and Highveld Coalfields are situated. The topography therefore precludes generous outcrop and consequently, most of the exploration for coal deposits was and is carried out by core drilling. These data are therefore basic to the evaluation of the coalfields. In this study, situated in the northwestern Highveld Coalfield, the data base comprises primarily drillhole cores complemented by the examination of outcrops within, and adjacent to, the study area (Figs. 1.2 and 1.3).

These data were synthesized and formed the basis for the subsurface sedimentological analysis. The analysis of the coalfields using this approach has been successfully applied to the central Witbank Coalfield (Le Blanc Smith, 1980), the eastern Witbank Coalfield (Cairncross, 1986, Holland, *et al.*, 1989) the northeastern Highveld Coalfield (Winter, 1985) and the southeastern Highveld Coalfield (Hagelskamp, 1987). These studies have successfully documented the stratigraphy of the various areas of the coalfields and proposed models of peat formation.

The primary objectives of the study are:

a. To provide a detailed stratigraphic framework for the coal-bearing Vryheid Formation in the northwestern Highveld Coalfield.

b. To identify depositional systems from the integration of vertical and lateral sedimentary facies associations.

c. To reconstruct the spatial and temporal depositional architecture for the study area from the distribution of depositional systems within depositional sequences.

d. To establish an in-seam stratigraphy for the Nos. 1-6 coal seams based upon coal lithotype profiles and to deduce what factors influence coal seam stratigraphy.

e. To formulate models of peat accumulation and document factors that influence coal seam thickness, distribution and maceral content.

1.5.2 Methodology

Data for the study were obtained from the 629 coreholes drilled over an area of 360km² (Fig. 1.3). Thus a spacing of about 1 corehole/km² is used to obtain the three dimensional sediment distribution. The cores drilled over the last twenty years and described by a number of different geologists were supplemented by cores personally examined. Utilizing the sedimentary facies approach of (Miall, 1977, 1978) coreholes were coded and graphically displayed utilizing the fortran IV computer programmes written by McCarthy (1981) and later Pope *et al.*, (1988). The best fit method of borehole correlation was applied to determine the stratigraphic framework for the

Coalfields. This approach was successfully applied by both Galloway (1972) and Ferm *et al.*, (1975). The construction of a stratigraphic framework allowed for the identification of discreet sedimentary units termed depositional sequences (Cadle, 1982b). Similar sedimentary units are also termed punctuated aggraded cycles (Goodwin and Anderson, 1985) or parasequences Van Wagoner *et al.*, (1988). The depositional sequences and specific lithologies within the sequences were mapped out using the General Purpose Contouring Package. These data together with sedimentological and coal cross-sections, graphically drawn by computer from in-house developed computer programmes, provided a three-dimensional appreciation of sedimentary facies associations. This information was then used to identify and construct depositional systems.

This method of detailed three-dimensional mapping of the Vryheid Formation allowed for the change in depositional styles with time to be documented and the influence of basin tectonics on sedimentation and peat formation to be postulated. These interpretations led to the documentation of a number of factors which influence peat accumulation. It is hoped that the understanding of depositional and peat forming processes, and the identification of those factors which influence coal thickness, distribution and coal type, will have application in coal exploitation.

CHAPTER 2

2 STRATIGRAPHY

2.1 INTRODUCTION

The construction of a stratigraphy for sedimentary rocks deposited within a basin provides a framework for a wide range of geological studies. Without a stratigraphic framework within which to work there is no way of determining younger from older rocks or rocks of similar age deposited within a basin. Therefore, the construction of a stratigraphy for geological investigations, especially those with an economic bias, is of fundamental importance. The more detailed the geological investigation, especially those based upon the concept of depositional systems (Fisher and McGowan, 1967), the more refined the stratigraphic approach required. Consequently, there has been a swing away from lithostratigraphic studies towards genetic stratigraphic studies. For the Vryheid Formation, this evolution from lithostratigraphy to genetic stratigraphy was necessitated due to the detailed investigations carried out in the coalfields in the recent past.

The rocks of the Karoo Basin were first stratigraphically subdivided by Rogers (1903) who used a time-stratigraphic subdivision. The rocks within the basin were termed the Karoo "System" and they were in turn subdivided into the Dwyka, Ecca, Beaufort and Stormberg "Series". Du Toit (1918) further refined the Ecca Series into the Lower Ecca Beds, the Middle Ecca Beds and the Upper Ecca Beds. These units were given formal lithostratigraphic status by the South African Committee for Stratigraphy (S.A.C.S. 1980) and, in the northern Karoo Basin, the terms Lower, Middle and Upper Ecca were renamed the Pietermaritzburg, Vryheid and Volksrust Formations respectively.

The Vryheid Formation which represents the main coal-bearing unit is described by S.A.C.S. (1980) as: "consisting essentially of sandstone, shale and subordinate coal beds, and has a maximum thickness of 500m". The use of the term Vryheid Formation is lithostratigraphically important as it defines a formation that extends over a large area, both on the surface and in the subsurface. However, once detailed studies of an interpretive nature are undertaken a more sophisticated stratigraphic framework is required. For example, using lithostratigraphy,

a. Where does one define the boundary between the Dwyka Group and Vryheid Formation where a glaciolacustrine delta deposits its sediment load within the basin?

b. How does one distinguish between separate sandstones, shales and coal seams deposited at the same stratigraphic position?

c. How does one identify eustatic rises and falls of sea level and sedimentary responses to basin tectonics?

This shortcoming in stratigraphic nomenclature was realised by Cadle (1974), who used a genetic subdivision of the Vryheid Formation, in order to define potentially favourable reservoir rocks for petroleum exploration. The Vryheid Formation in the Eastern Transvaal and northern KwaZulu/Natal was informally subdivided into a lower deltaic sedimentary sequence, a middle fluvial sedimentary sequence and an upper deltaic sedimentary sequence. A similar shortcoming in constructing a stratigraphic framework for detailed coal related studies was identified, by Le Blanc Smith (1980, 1981a), in the Witbank Coalfield. He proposed a formal genetic stratigraphy for this Coalfield based upon Genetic Sequences and Increments of Strata after Busch (1971). This method allowed for the detailed mapping of specific lithologies within the coalfield and provided a framework for inter-coalfield correlations of sedimentary packages.

Subsequent workers such as Cadle (1982b), Winter (1985) and Cairncross (1986) have not used the terminology proposed by Le Blanc Smith (1980, 1981a) as genetic increments of strata in the central Witbank Coalfield are not present in adjacent coalfields. Moreover, genetic sequences of strata present in the KwaZulu/Natal Coalfields are not present in the Witbank and Highveld Coalfields.

Cadle (1982b) introduced the term depositional sequence to identify progradational and aggradational sedimentary rocks separated by laterally persistent transgressive deposits. The informal terms No. 2 seam, No. 4 seam, No. 5 seam and No. 6 seam depositional sequences were introduced. This approach was followed by Winter (1985) in the Highveld Coalfield who subdivided the stratigraphy into the No. 2 Seam, No. 4 Seam and No. 5 Seam Genetic Sequences. Cairncross (1986) used the terms No. 2, No. 4 and No. 5 seam sequences to stratigraphically describe the sediments in the eastern Witbank Coalfield. Furthermore, Cairncross (1986) documented significant differences in lithologies below the No. 1 seam and between the No. 4 and No. 5 seams when compared to the genetic increments of strata described by Le Blanc Smith (1980) for the central Witbank Coalfield.

More recent sedimentologically based studies by Roberts (1986), Christie (1988) and Smith (1988) in northern KwaZulu/Natal have not rigorously pursued a genetic approach to the Vryheid Formation stratigraphy. These studies subdivided the Vryheid Formation into a Lower Zone, a Coal Zone, and an Upper Zone. This subdivision follows the lithostratigraphic approach of Blignaut and Furter (1940). In terms of depositional systems, the informal genetic subdivisions introduced by Cadle (1974) apply in the KwaZulu/Natal Coalfields i.e. lower deltaic, middle fluvial and upper deltaic sedimentary sequences.

In summary, it is apparent that a formal genetic stratigraphy for the Vryheid Formation can only be introduced once the regional variations and complexities from a genetic standpoint have been documented through process response studies of the Vryheid Formation.

2.2 GENETIC STRATIGRAPHY

2.2.1 Introduction

The delineation of depositional units of sufficient extent and appropriate scale for analysis is one of the more difficult tasks in genetic facies interpretation. In order to delineate these depositional units, processes of sediment fill must be recognised. Sedimentation within sedimentary basins or "depositional architecture" provides basic information on depositional processes, depositional systems and environments. Sedimentation takes place by:

a. **Aggradation;** the process of vertical filling of the basin.

b. **Progradation;** the process of infilling of a basin from the margin.

c. **Lateral accretion;** the process whereby sediment moving within the basin preferentially accumulates sediment against the margin.

These three mechanisms of sediment fill can be readily distinguished in the rock record. Aggradational sedimentation produces no inherent systematic textual trends and each bed may display varying texture and composition. Progradational sequences coarsen upward in grain size, which contrast with laterally accreting sequences which fine upward in grain size. Within the stratigraphic record these three types of sediment fill may coexist within a single depositional system or be stacked vertically as successive depositional increments within the stratigraphic record.

Frazier (1974) working on Quaternary sediments of the Gulf Coast Basin developed a conceptual model for three-dimensional stratigraphic studies. The model is based upon the principle that sedimentary basins are filled by alternating depositional and nondepositional intervals. Therefore a nondepositional interlude (hiatus) separates depositional units. Frazier (1974) also defined a **depositional event** as comprising a progradational phase, an aggradational phase and a transgressive phase. The progradational phase thickens basinward, the aggradational phase overlies the progradational phase and thickens landward, and the transgressive phase comprising a thin veneer of sediment terminates the depositional event. Stacked depositional events combine to form a major genetic stratigraphic unit termed a **depositional episode** (Fig. 2.1). The hiatal surfaces and transgressive deposits provide physical correlation markers that are used to define the boundaries of genetic units. Furthermore, these surfaces represent correlatable time lines. In summary, in order to achieve three-dimensional mapping of depositional systems, genetic stratigraphic units are determined through the recognition of the bounding surfaces defining depositional events and episodes.

Genetic stratigraphic concepts have also been applied in seismic stratigraphic analysis whereby reflection patterns outline sequences of conformable strata bounded by regional unconformities (Mitchum *et al.*, 1977, Vail *et al.*, 1977). On a regional scale, primary seismic reflectors tend to follow time-stratigraphic horizons, such as regional bedding surfaces or disconformities, and define both depositional events and depositional episodes. Van Wagoner *et al.*, (1988) outline the newly evolved concepts of sequence stratigraphy. The fundamental unit of sequence stratigraphy is the sequence which comprises a relatively conformable succession of genetically related strata bounded by unconformities and their correlative conformities (Mitchum, 1977). Parasequences and parasequence sets comprise the fundamental building blocks of sequences. A **parasequence** is defined as a relatively conformable succession of genetically related beds or bedsets bounded by **marine-flooding surfaces** and their

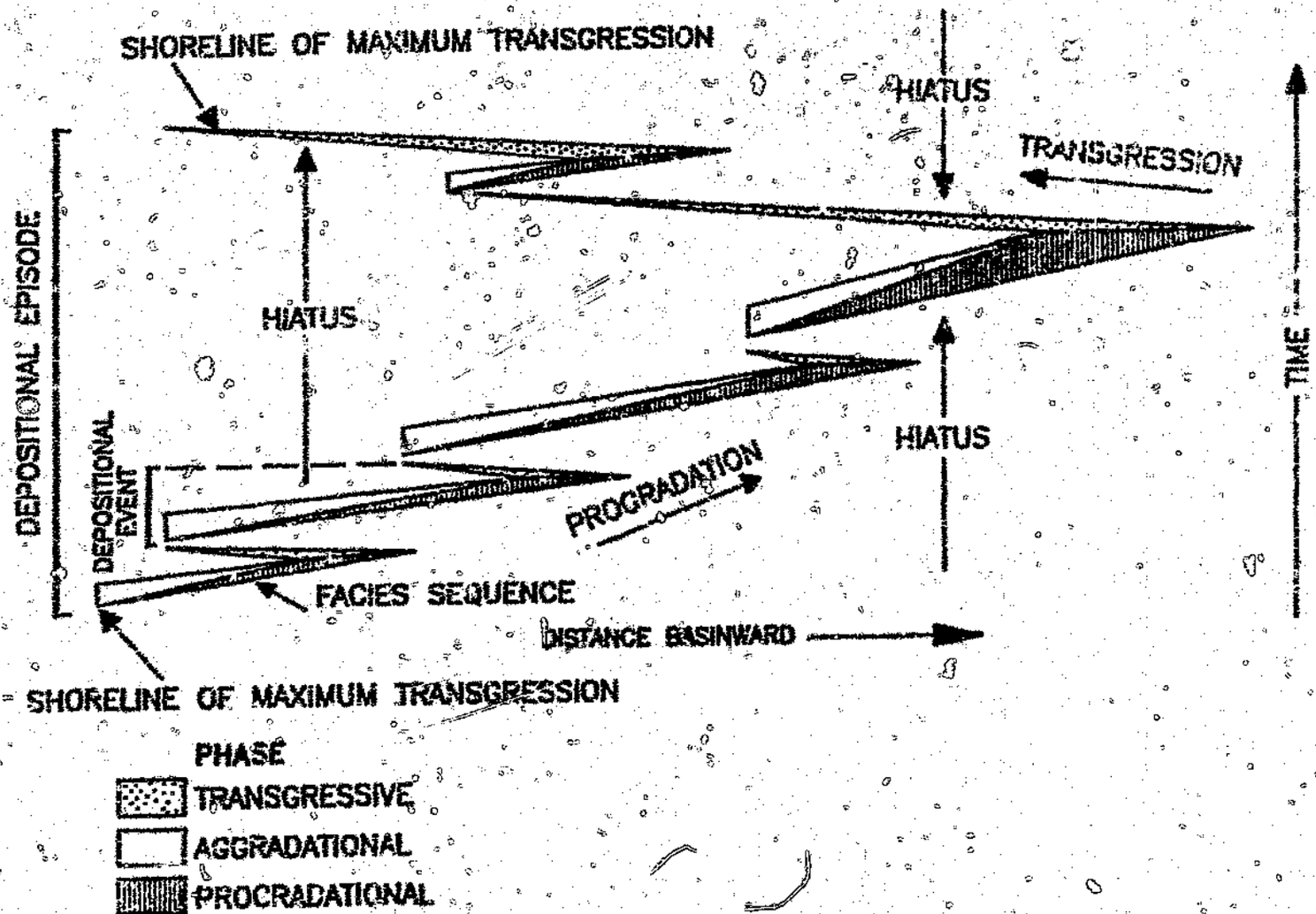


Figure 2.1. The concept of depositional events and a depositional episode (modified after Frazier, 1974).

correlative surfaces. The parasequence in siliciastic rocks are progradational and therefore shoal upward. A marine-flooding surface is a surface that separates younger from older strata across which there is evidence of an abrup increase in water depth. Parasequence sets, a succession of genetically related parasequences are bounded a major marine-flooding surfaces and their correlative surfaces.

Also defined are systems tracts which are linked contemporaneous depositional systems (Brown and Fisher, 1977). Van Wagoner *et al.*, (1988) recognise four systems tracts within seismic sequences. They are; the lowstand system tract, the shelf margin system tract, the transgressive systems tract and the highstand system tract. These system tracts are defined by the type of sequence boundary and parasequence sets. The lowstand system tract overlies a type 1 sequence boundary characterised by; subaerial erosion, stream rejuvenation, a basinward shift of facies, and a downward shift in onlap. Within the lowstand system tract, the following units are recognised; a basin-floor fan, a slope fan and a low-stand wedge (Fig. 2.2). The low-stand wedge comprises incised valley-fill deposits, and prograding parasequence sets representing shelf and slope deposits (Fig. 2.2). The shelf-margin system tract overlies a type 2 sequence boundary defined by a downward shift in onlap landward of the depositional shoreline break (Fig. 2.3). This tract type comprises one or more weakly progradational and aggradational parasequences. The transgressive systems tract is situated between the lowstand or shelf-margin system tracts, and the highstand system tract (Figs. 2.2 and 2.3). The transgressive system tract is characterised by one or more retrogradational parasequence sets. The highstand system tract is present landward of the other system tracts and characterised by one or more aggradational parasequence sets. The aggradational parasequence sets are succeeded by one or more progradational parasequence sets. The parasequences of the highstand system tract onlap landward on to the sequence boundary. This concept of genetic sequence stratigraphy is illustrated in Figures 2.2 and 2.3.

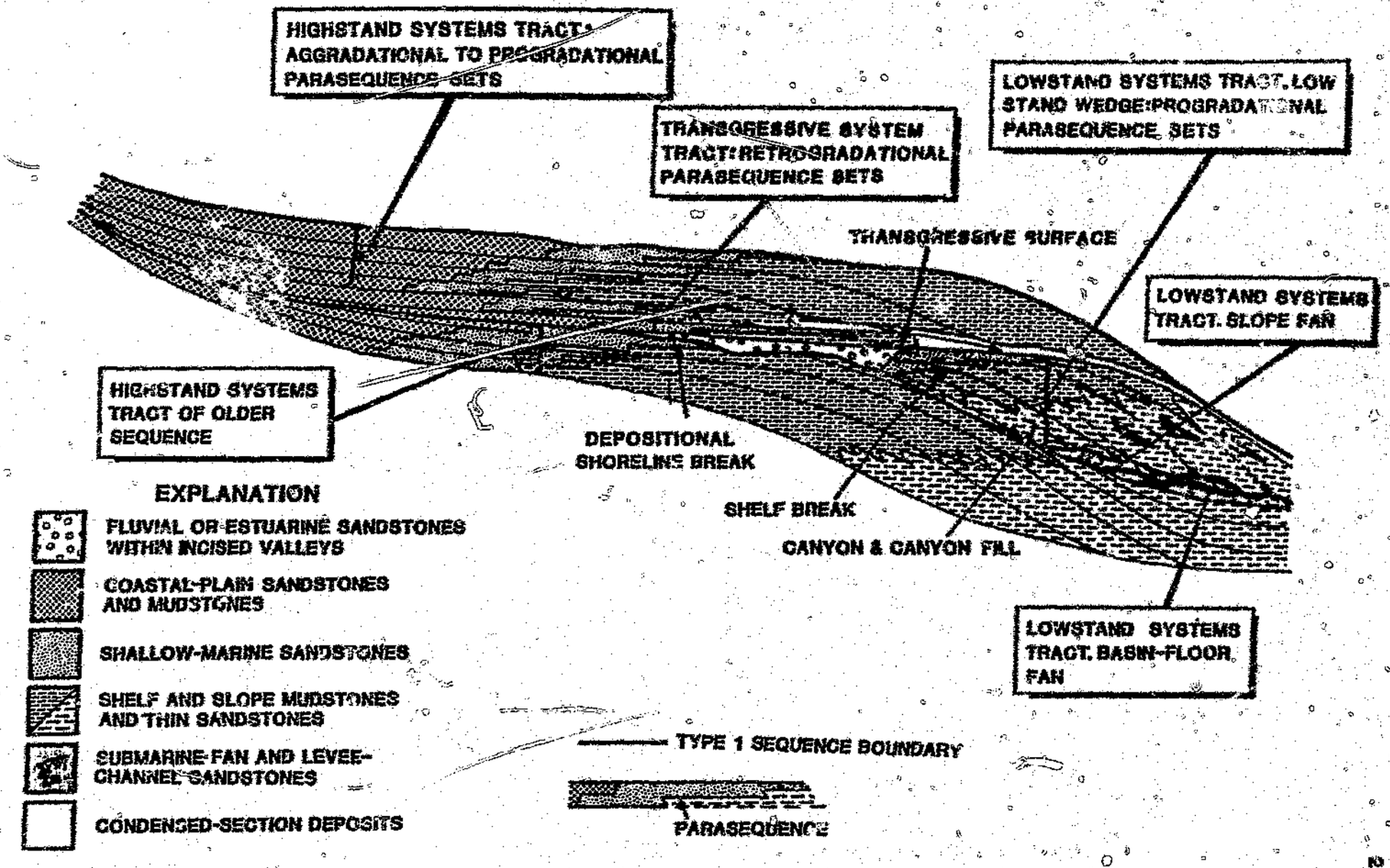


Figure 2.2. Type 1 sequences and systems tracts deposited in a basin with a shelf break (after, Van Wagoner *et al.*, 1988).

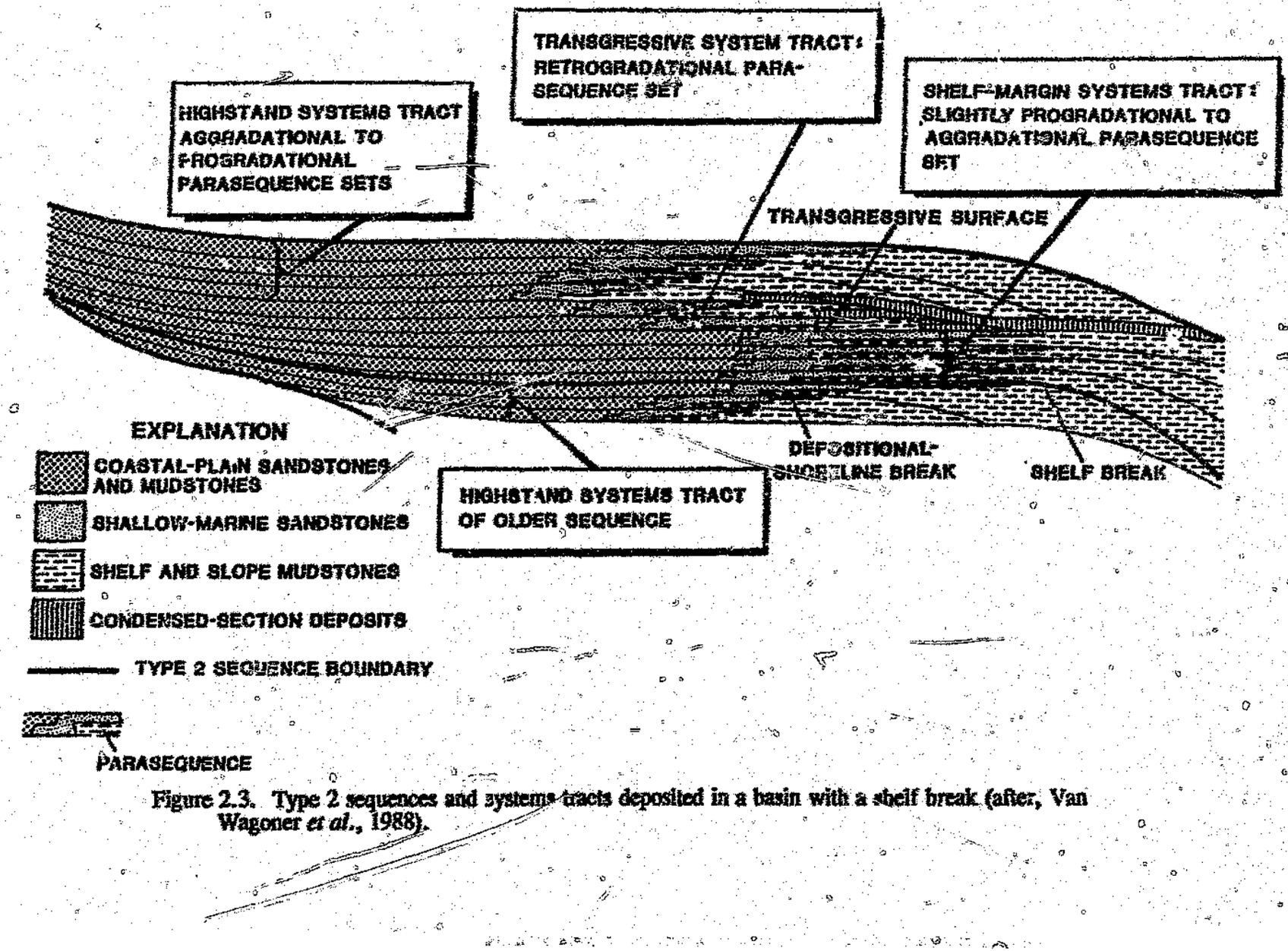


Figure 2.3. Type 2 sequences and systems tracts deposited in a basin with a shelf break (after, Van Wagoner *et al.*, 1988).

The applicability of sequence stratigraphy to sedimentary basins without the benefit of seismic sections can only be applied in part. Firstly, basin fill displaying unconformity surfaces, such as the Transvaal Sequence, would be ideal for this type of analysis (Clendenin, 1989). Secondly, sequence stratigraphy can also be applied to basin fill displaying shallowing-upward sequences and marine-flooding surfaces. However, certain refinements in sequence stratigraphy cannot be applied unless the rocks of the entire downdip section of a basin, from continental deposits through transitional and continental shelf to slope deposits, are well documented. For example, if rocks comprising coastal plain sandstones and mudrocks from the basin margin are studied it is difficult, if not impossible, to determine whether the rocks comprise part of a highstand systems tract, a shelf margin system tract or a transgressive systems tract.

With reference to the sedimentary fill within the Karoo Basin, which is situated within the Karoo foreland basin (Cadle *et al.*, 1990), no type 1 unconformities have been recognised to date. Consequently, type 1 sequence boundaries can only be placed at the base and top of the Karoo Sequence. Type 2 sequence bounderies may well be present within the basin but have not been recognised. At a parasequence level, however, both shallowing-upward sequences and marine-flooding surfaces are recognised within the Transvaal Coalfields. These parasequences are synonymous with the depositional events defined by Frazier (1974). Consequently, a genetic stratigraphy based upon parasequences or depositional events is constructed for the Vryheid Formation in the northwestern Highveld Coalfield.

Based upon the sedimentological interpretation of borehole logs and cross-sections depositional events or parasequences termed **depositional sequences** in this study are delineated (Cadle, 1982b, Cairncross and Cadle, 1987). A composite stratigraphic profile displaying the lithologies and genetic depositional sequences is presented in Figure 2.4. Five depositional sequences are delineated and are informally termed the **No. 1 seam depositional sequence, No. 2 seam depositional sequence the No. 4**

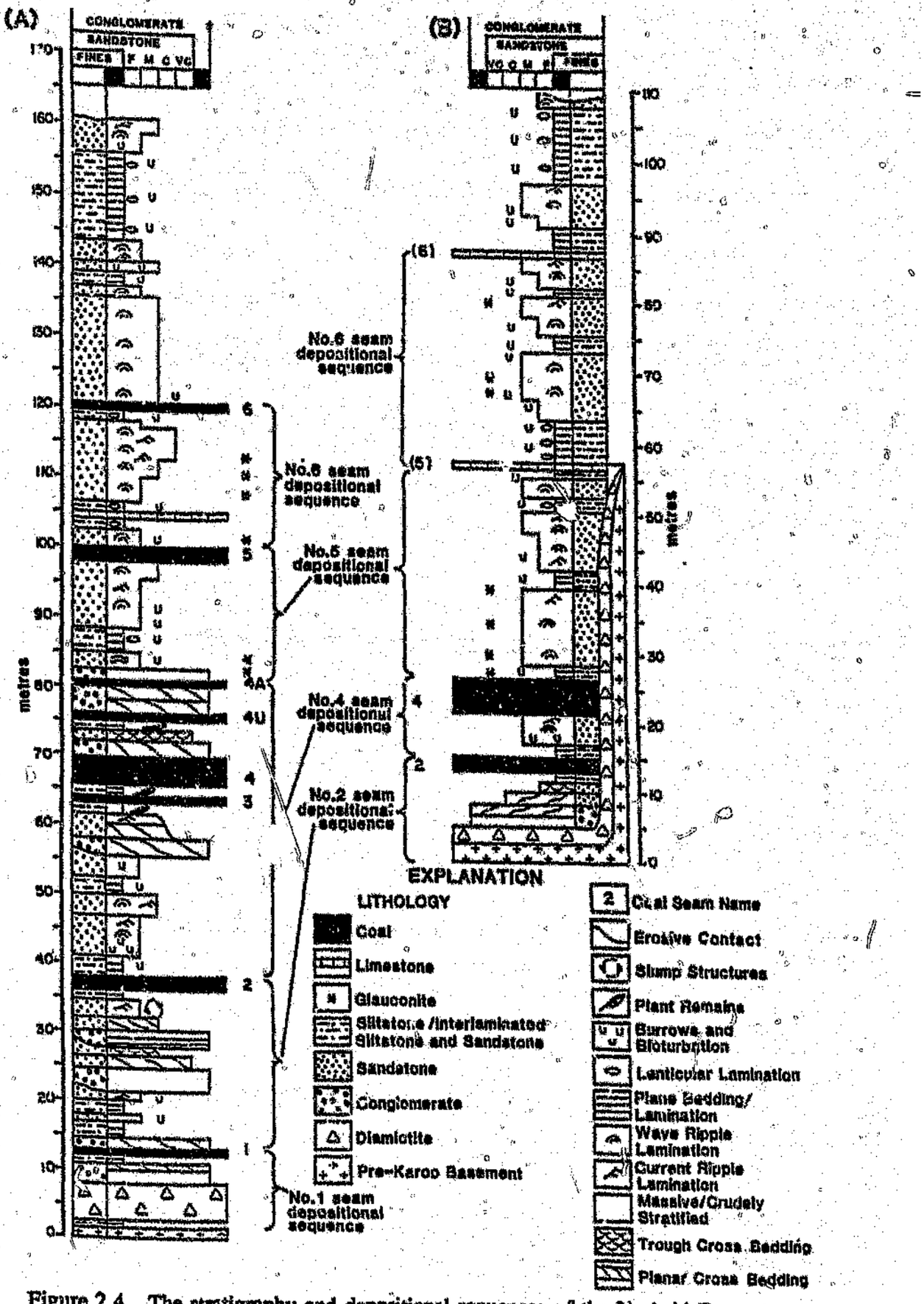


Figure 2.4. The stratigraphy and depositional sequences of the Vryheid Formation in the northwestern Highveld Coalfield. Columns (A) and (B) illustrate the variation in stratigraphy.

seam depositional sequence the No. 5 seam depositional sequence and the No. 6 seam depositional sequence. Each sequence comprises a progradational shallowing-upward sequence overlain by aggradationally deposited sediments and/or coal seams. The depositional sequences are separated by marine-flooding surfaces and a thin veneer of transgressive sediments.

The stratigraphy of the northwestern Highveld Coalfield is presented through the description of each depositional sequence. Borehole cross-sections, profiles and isopach maps are used to illustrate the lithological details of the depositional sequences. Figure 2.5 illustrates the position of four out of thirty-one cross-sections used to portray the variation in stratigraphy of the depositional sequences.

2.3 No. 1 SEAM DEPOSITIONAL SEQUENCE

The No. 1 seam depositional sequence is the first or basal sedimentary sequence deposited in the study area. The sequence comprises the stratigraphic interval between pre-Karoo basement and the top of the No. 1 coal seam or a correlative carbonaceous siltstone. The three-dimensional control of lithologies is not as well defined as the overlying depositional sequences due to the fact that, for many boreholes, drilling terminated after penetration of the No. 2 coal seam. However, 137 boreholes penetrated down to basement which have allowed the stratigraphy of the sequence to be determined. The sequence is characterised by the deposition of diamictite, conglomerate and coarse-grained sandstone (Fig. 2.6). Figures 2.7 and 2.8 illustrate the variation in thickness and sediment type. Lithologically, this sequence comprises:

a. clast-supported pebble, cobble, and granule conglomerate,

b. cross-stratified and massive, very coarse- to coarse-grained sandstone,

c. carbonaceous siltstone, mudstone and coal seams, and

d. diamictite.

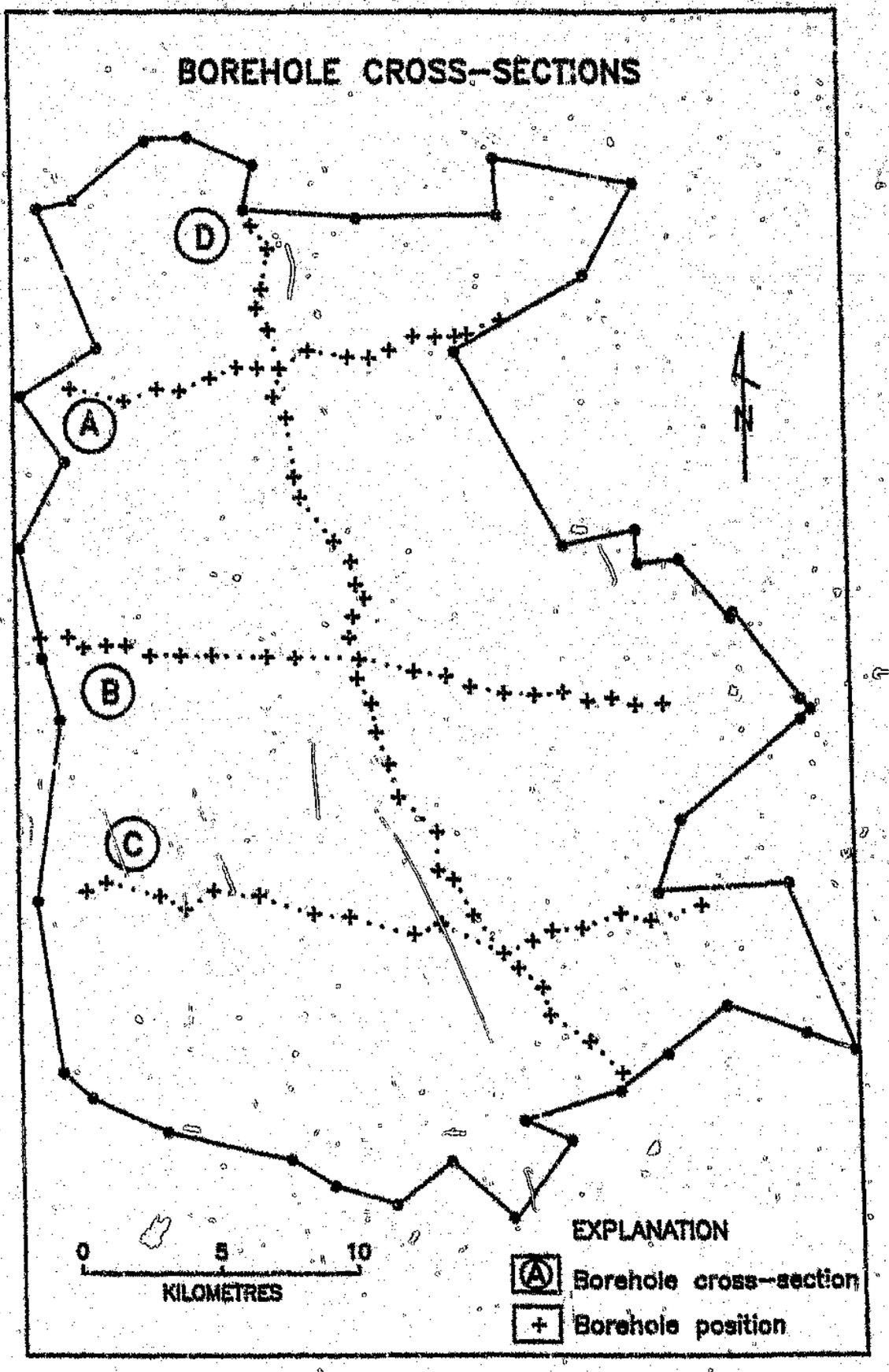


Figure 2.5. Locality map showing the positions of stratigraphic cross-sections.

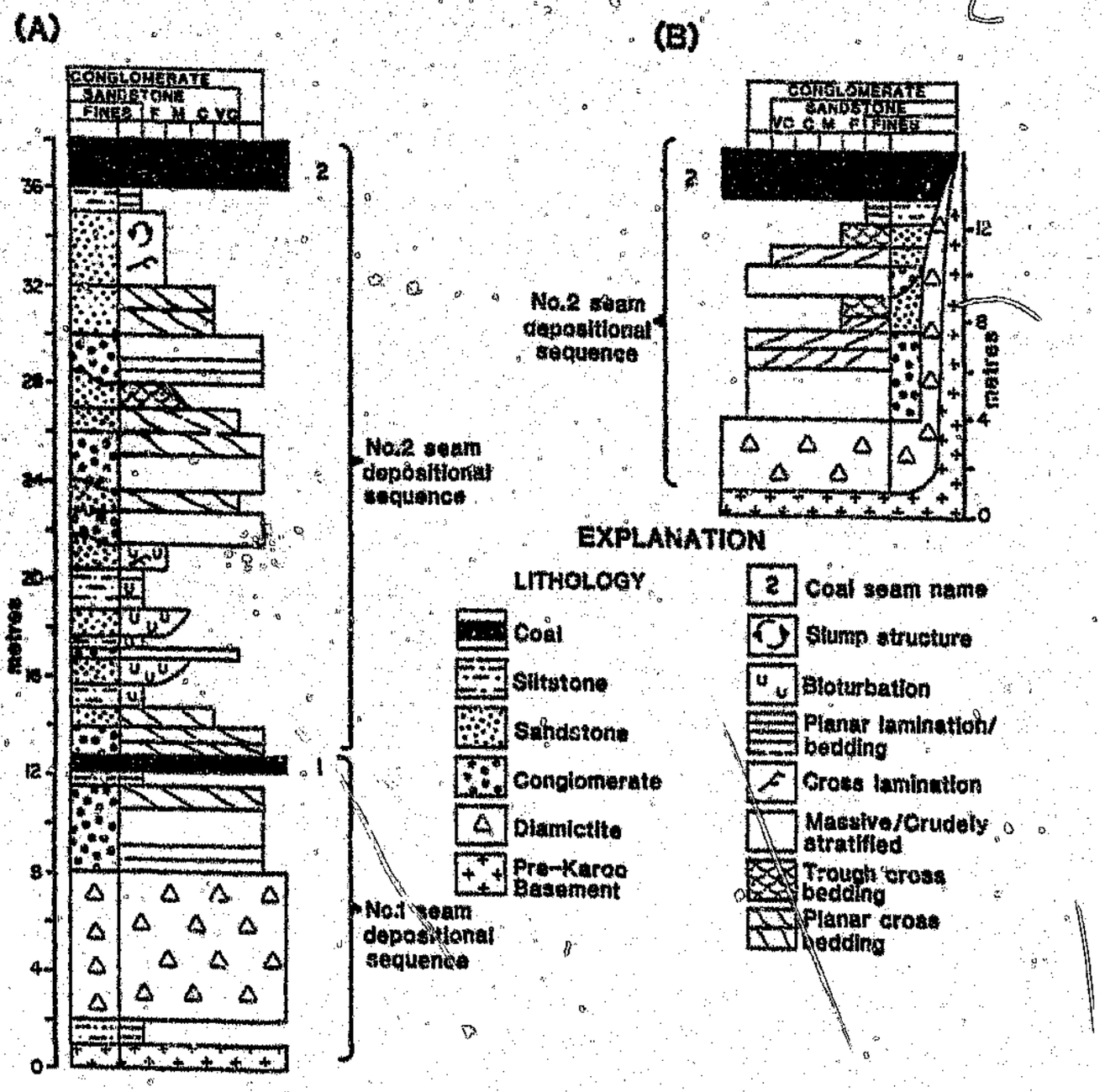


Figure 2.6. Stratigraphic columns of the No. 1 and No. 2 seam depositional sequences. Columns (A) and (B) illustrate the variation in stratigraphy for this sedimentary interval.

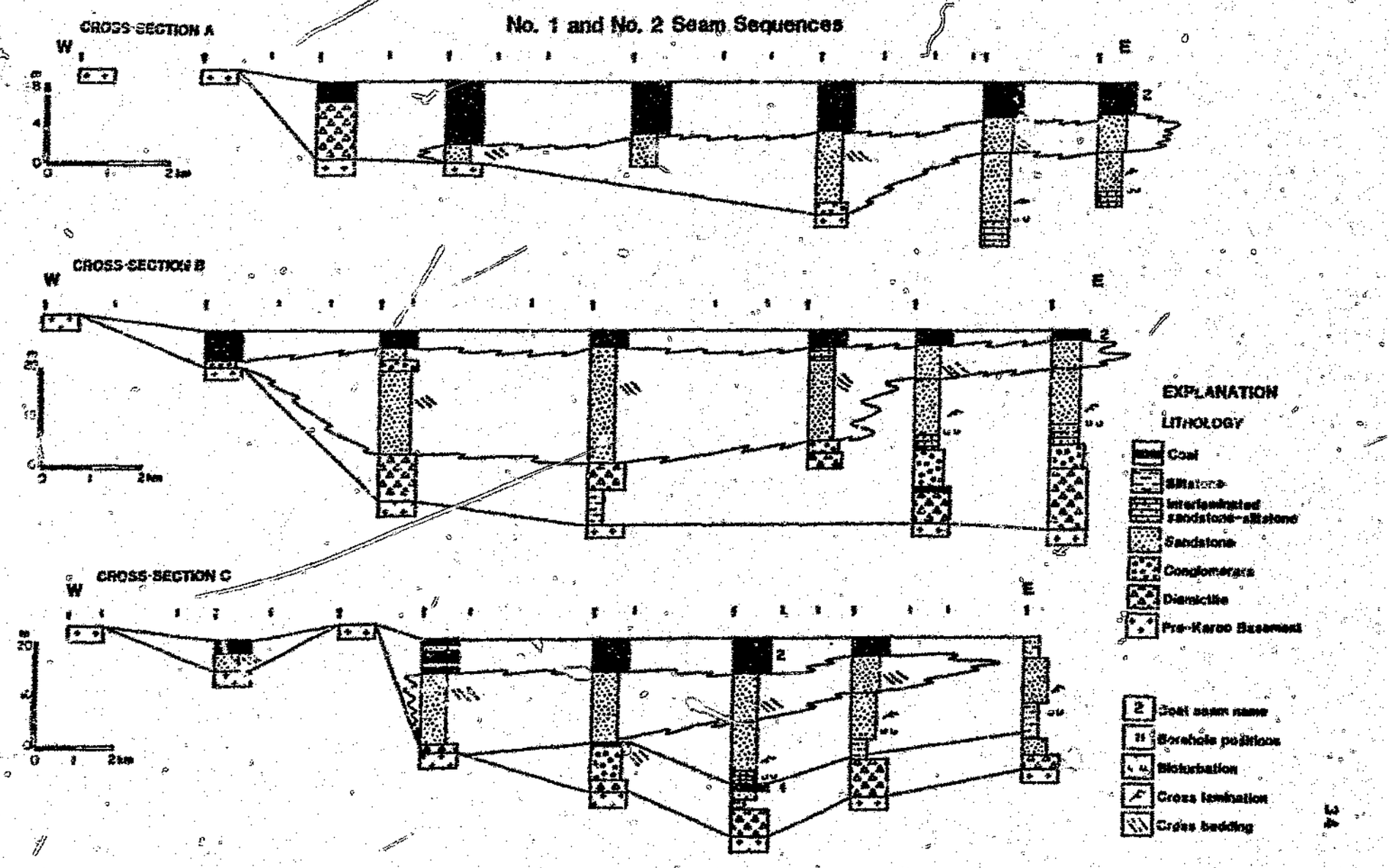


Figure 2.7. Westeast cross-sections of the stratigraphic interval between basement and the No. 2 seam. See Fig. 2.5 for positions of cross-sections.

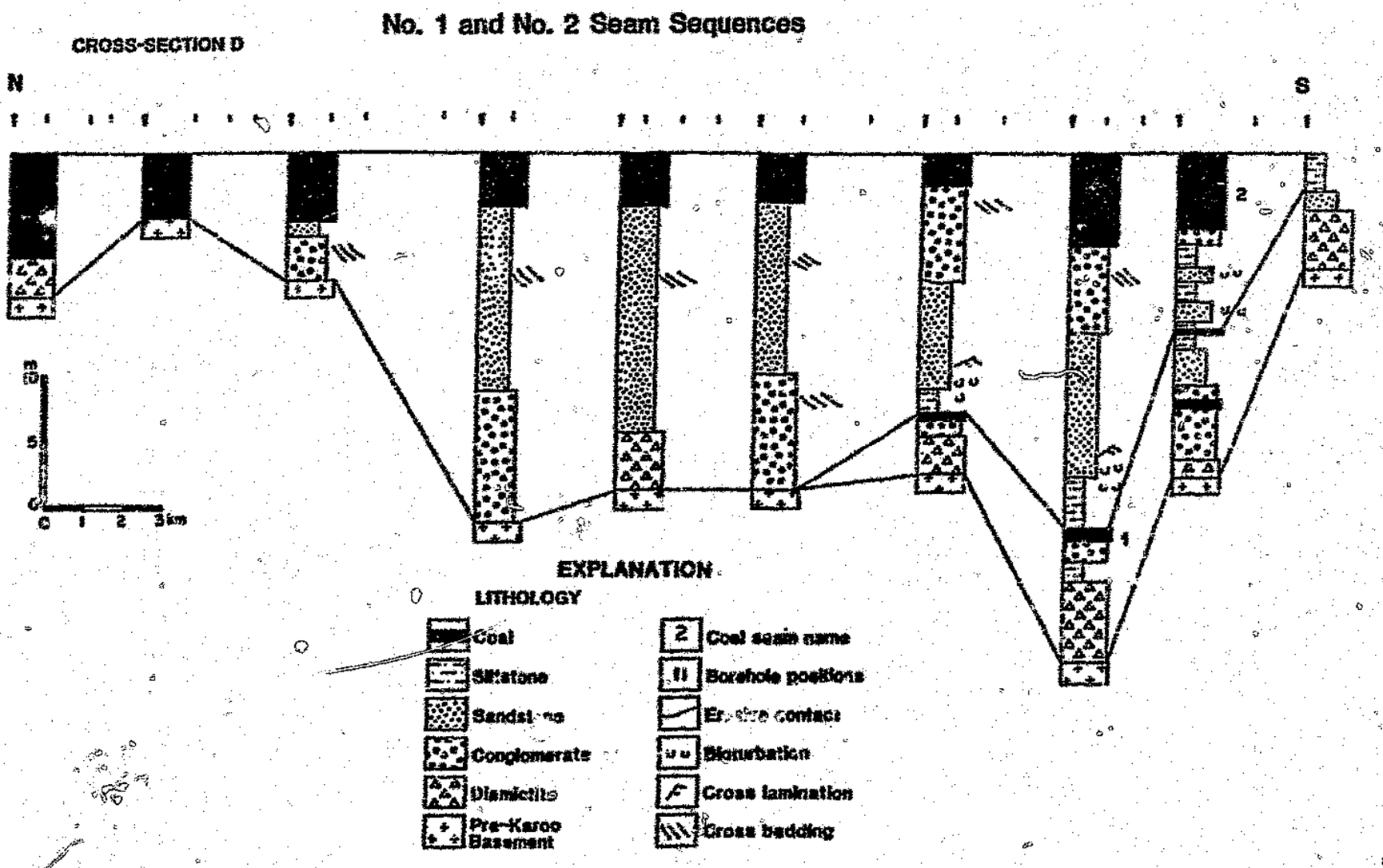


Figure 2.8. Northsouth cross-section of the stratigraphic interval between basement and the No. 2 seam. *See Fig. 2.5 for the position of the cross-section.*

In general, the sedimentary rocks of the No. 1 seam depositional sequence are only present in the southeastern portion of the study area as the rocks of this sequence pinch out against basement in the northwest (Figs. 2.7 and 2.8). The sequence comprises from the base upwards, interbedded diamictite (of the Dwyka Group) and carbonaceous mudstone, overlain by pebble and granule-grade conglomerate, carbonaceous siltstone, and the No. 1 coal seam. The Dwyka diamictite is confined to palaeotopographically low lying areas and attains its greatest thickness along the eastern margin of the study area where it comprises a 32m thick, palaeovalley fill (Fig. 2.9). Overlying the diamictite is a unit of conglomerate and very coarse-grained sandstone containing faceted pebbles. Vertically, these sediments display a fining upward in grain size (Fig. 2.6). Dark-grey to black carbonaceous siltstones or the laterally impersistent No. 1 coal seam cap this sedimentary interval. The isopach map of the No. 1 seam illustrates the distribution and thickness of the seam (Fig. 2.10). The seam reaches a maximum thickness of 2m in the southeast where it displays a laterally persistent geometry. Infrequently, pebble and granule-grade conglomerate and very coarse-grained sandstone split the No. 1 seam into two subseams (Fig. 2.8).

2.4 No. 2 SEAM DEPOSITIONAL SEQUENCE

The No. 2 seam depositional sequence comprises the stratigraphic interval between the top of the No. 1 seam and the top of the No. 2 seam (Fig. 2.6A). Where the No. 1 seam depositional sequence is not present, the sequence comprises the interval between basement and the No. 2 seam (Fig. 2.6B). The No. 2 seam depositional sequence differs lithologically from the equivalent stratigraphic interval in the Witbank and northeastern Highveld Coalfields, described by Le Blanc Smith (1980), Winter (1985) and Cairncross (1986). The lithological differences are that bioturbated siltstones and sandstones are present in the study area and not present in the Witbank and northeastern Highveld Coalfields.

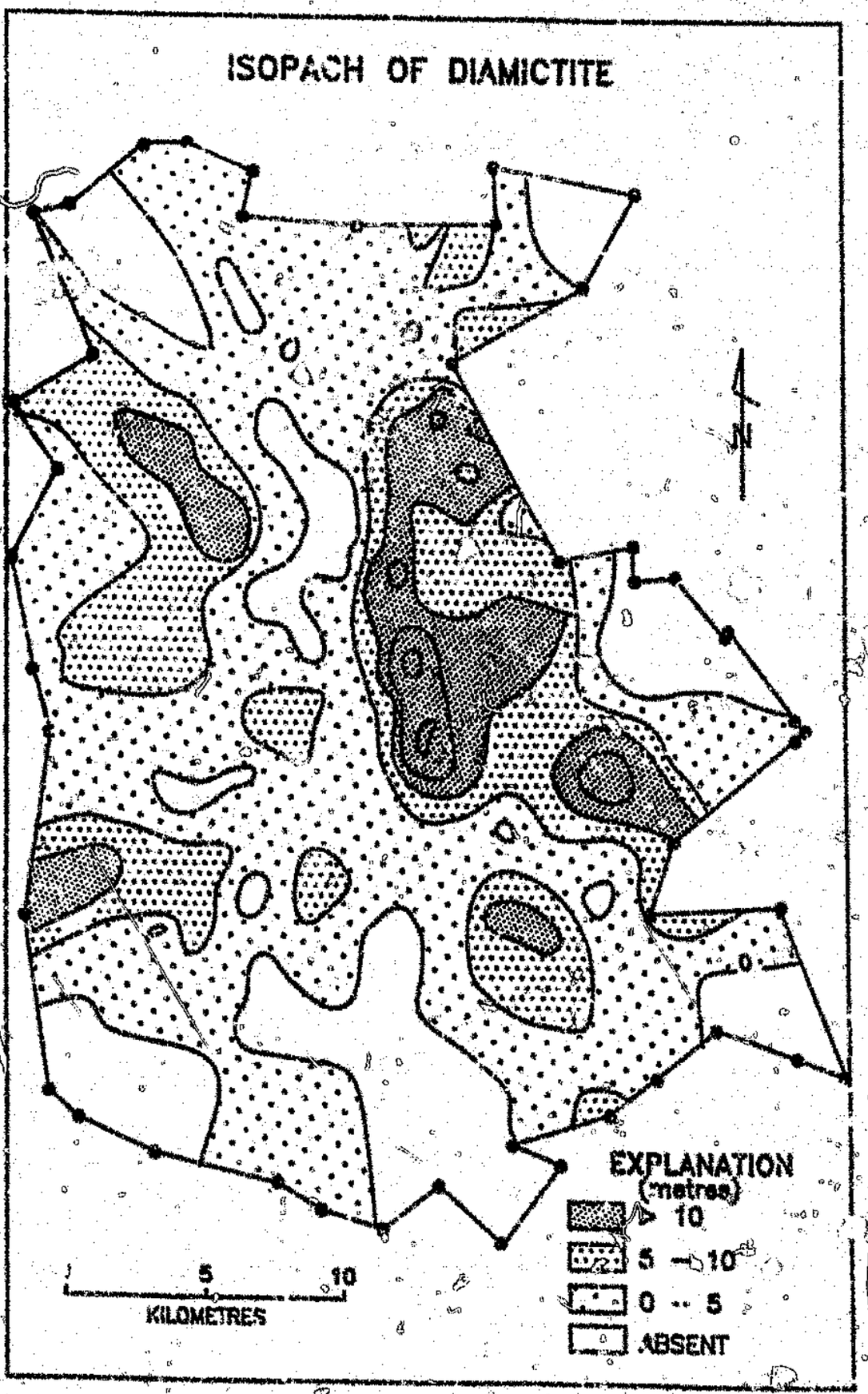


Figure 2.9. Isopach of the Dwyka Group within the study area.

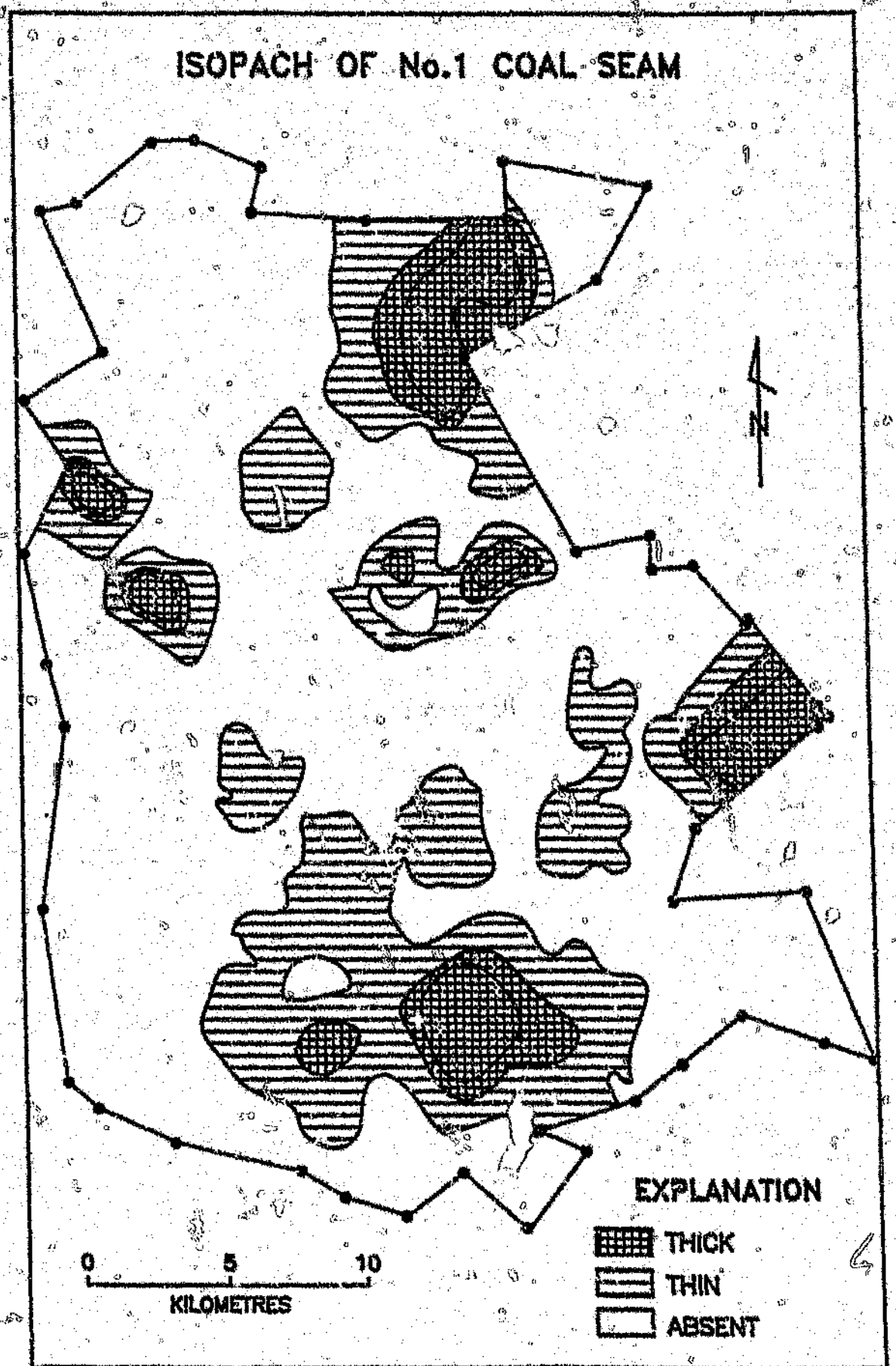


Figure 2.10. Isopach of the No. 1 seam illustrating the lateral impersistence of the seam. The terms "thick and thin" are relative terms describing thickness variations of the seam.

Two different lithological associations are present for the sequence in the study area (Fig. 2.6). In the northwestern portion of the study area the sequence comprises Dwyka Group diamictite overlain by between one and four units of massive and cross-bedded granule-grade conglomerates and very coarse-grained sandstones which fine upward into fine-grained sandstones (Fig. 2.6B). The conglomerates and very coarse-grained sandstones are erosively based and commonly contain coal spar and carbonaceous siltstone intraclasts. These sediments are overlain by carbonaceous siltstone and the No. 2 seam. In the eastern and southeastern portion of the study area, a different lithological association is present which comprises bioturbated carbonaceous siltstone and sandstone, and conglomerate. Vertically, the lithologies grade upward from siltstone, through fine-grained sandstone into granule-grade conglomerate (Fig. 2.6A). The northwest-southeast lateral change from coarse-grained lithologies to bioturbated lithologies is well illustrated in Figures 2.7 and 2.8.

The isopach map of the sandstone and diamictite between basement and the No. 2 seam illustrates how basement topography has controlled sedimentary thickness (Fig. 2.11). These lithologies thin and pinch out against basement in the west and northwest and thicken in an easterly direction. Also evident is a north-south thickening of these lithologies along the eastern margin of the study area which corresponds to the area of thickening of the underlying Dwyka Group diamictite. The No. 2 coal seam, which represents an aggradational phase of sedimentation, defines the top of the No. 2 seam depositional sequence. The seam attains its greatest thickness in the north and thins in a southerly and southeasterly direction (Fig. 2.12). The influence of basement topography is still evident as the seam pinches out against a north-south basement high situated along the western margin of the study area (Figs. 2.7 and 2.12).

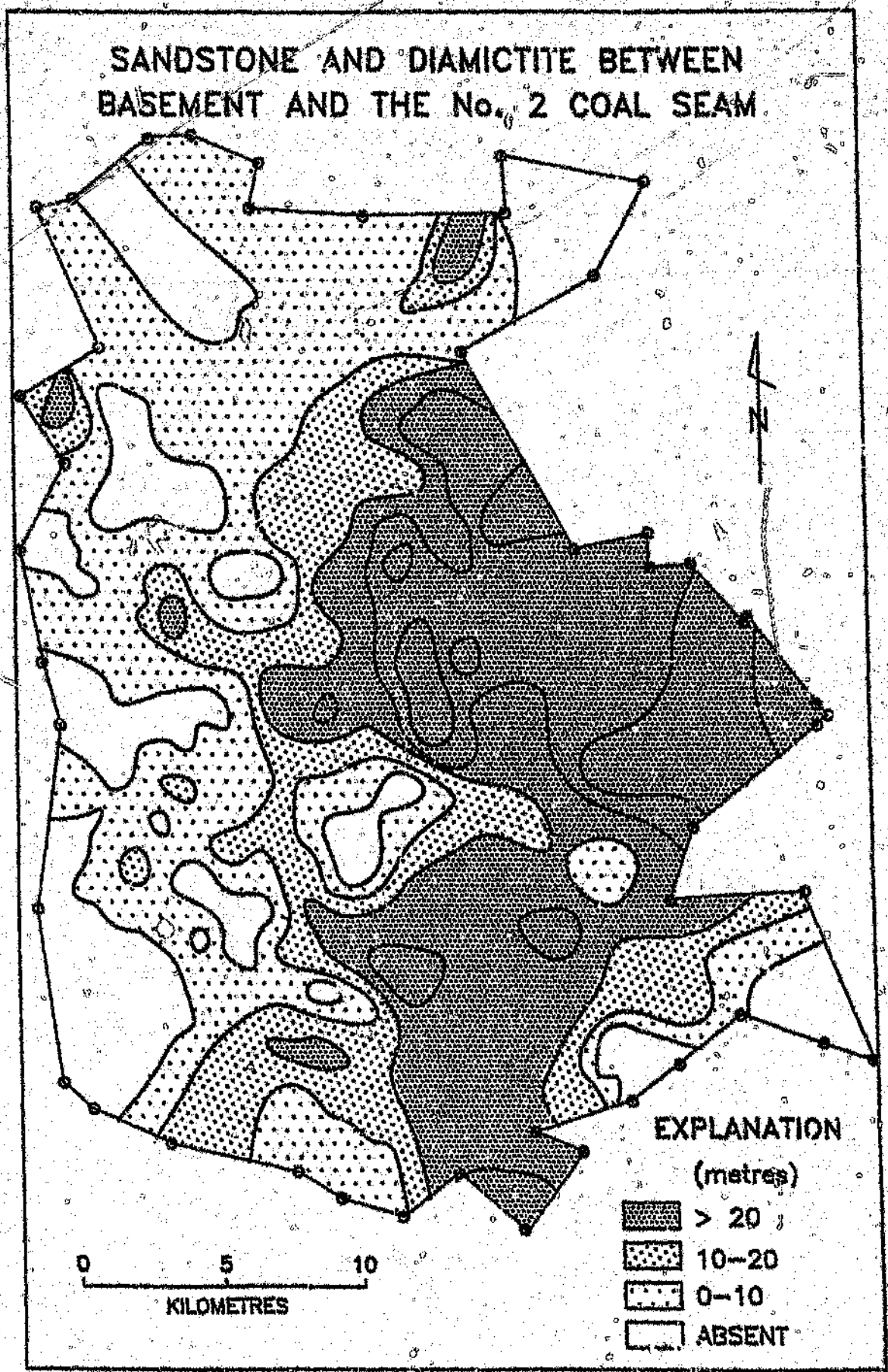


Figure 2.11. Isopach of diamictite and sandstone between basement and the No. 2 seam. The sediments thin against basement lithologies in the west (see Figure 2.7).

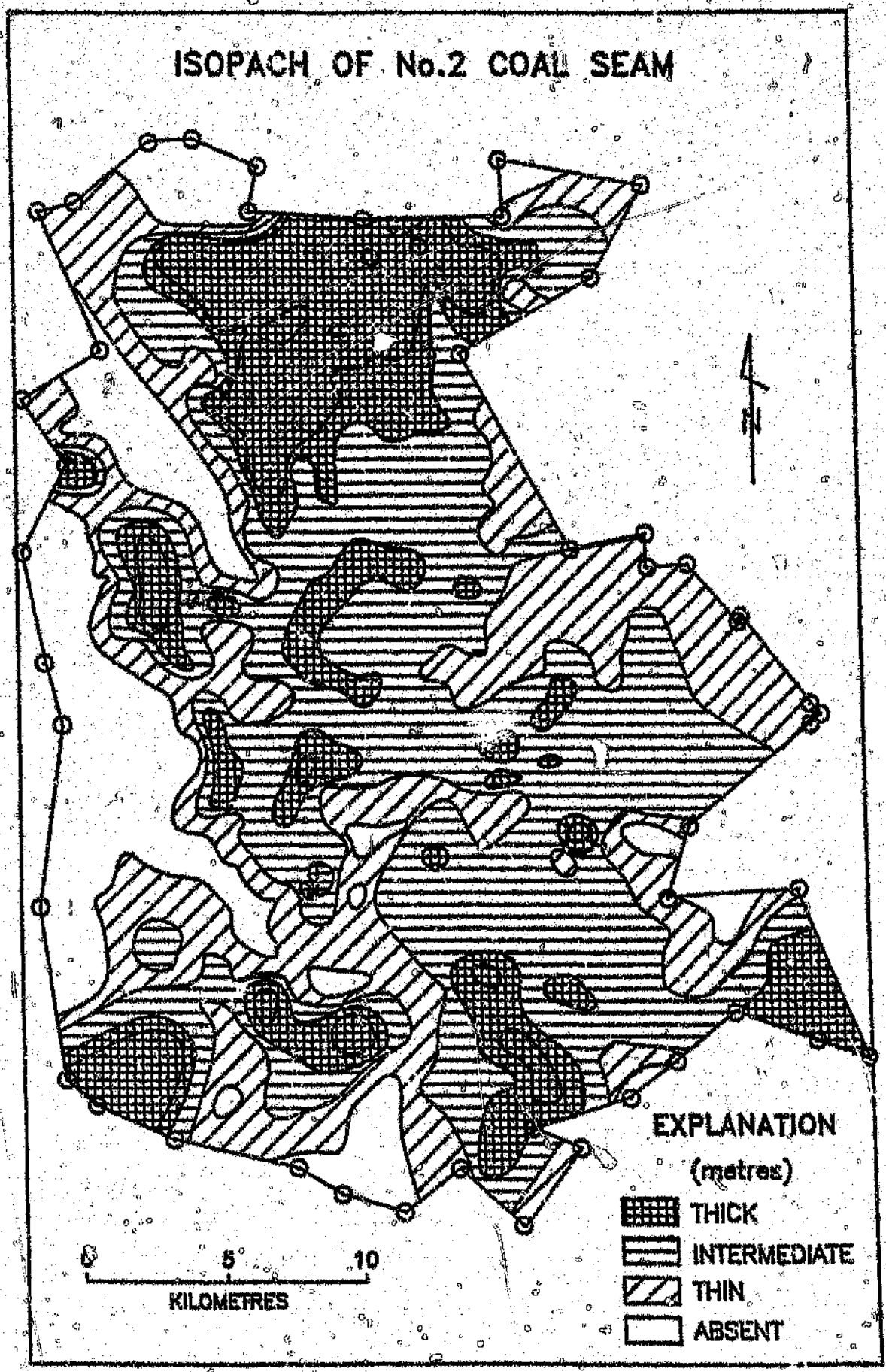


Figure 2.12. Isopach of the No. 2 seam. The seam thins and pinches out in the west and reaches a maximum thickness in the north. The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam.

The No. 2 seam is split by thin clastic partings into the No. 2 Lower and No. 2 Upper seams. The partings range in thickness from a few centimetres up to 5m and comprise siltstone and fine- to coarse-grained sandstone. The partings have splay-like geometries and are largely confined to the northwest and southwest of the study area (Fig. 2.13).

2.5 No. 4 SEAM DEPOSITIONAL SEQUENCE

The No. 4 seam depositional sequence is stratigraphically the most complex of the depositional sequences. It comprises the stratigraphic position between the top of the No. 2 seam and the top of the No. 4 seam or equivalent seam split (Fig. 2.14). Due to the complexity of the depositional sequence a number of stratigraphic intervals were recognised and isopached. These stratigraphic intervals are: the interval between the top of the No. 2 seam and the top of the No. 4 Lower seam, the interval between the top of the No. 2 seam and the top of the No. 3 seam, the interval between the top of the No. 3 seam and the top of the No. 4 Lower seam, and the interval between the top of the No. 4 Lower seam and the top of the No. 4 Upper seam (Fig. 2.14A).

The stratigraphic thickness between the No. 2 and No. 4 seams illustrates a 0-20m thickening of this interval from west to east across the study area (Fig. 2.15). The interval also thickens to the south were thicknesses of greater than 40m are attained. Also notable is the linear north-south thickening of the interval to greater than 30m in the southern half of the study area.

The interval between the No. 2 and No. 3 seams comprises either, one or, two upward-coarsening sequences (Figs. 2.16 and 2.17). The sequences commence with a carbonaceous siltstone and grade upwards through a fine-grained, bioturbated sandstone into a medium-grained, cross-laminated sandstone. Infrequently, these sandstones are

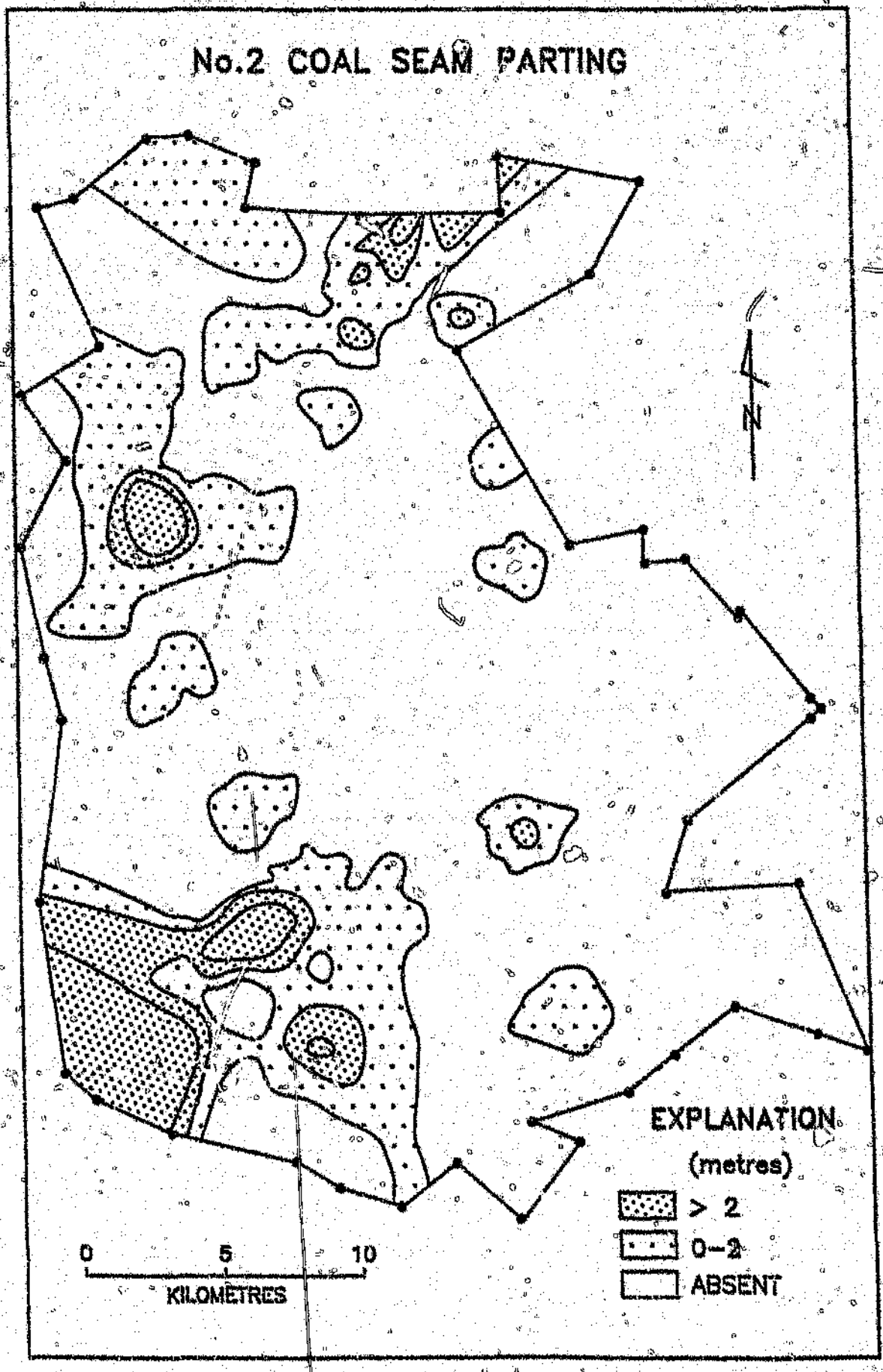


Figure 2.13. Isopach of the clastic parting within the No. 2 seam. The parting displays spay-like geometries and is largely confined to the western portion of the study area.

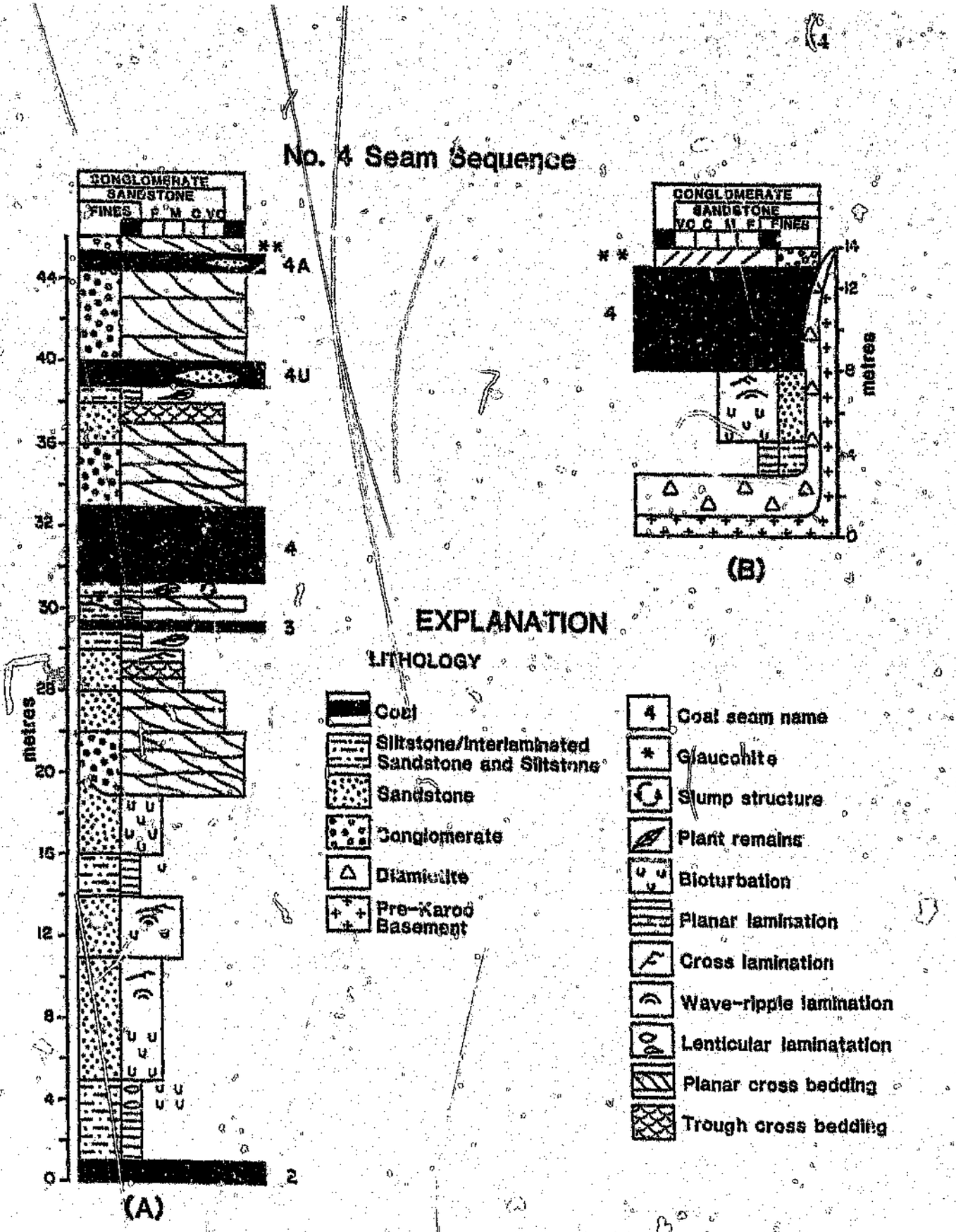


Figure 2.14. Stratigraphic columns of the No. 4 seam depositional sequence. Profiles (A) and (B) illustrate the variation in stratigraphy for this depositional sequence.

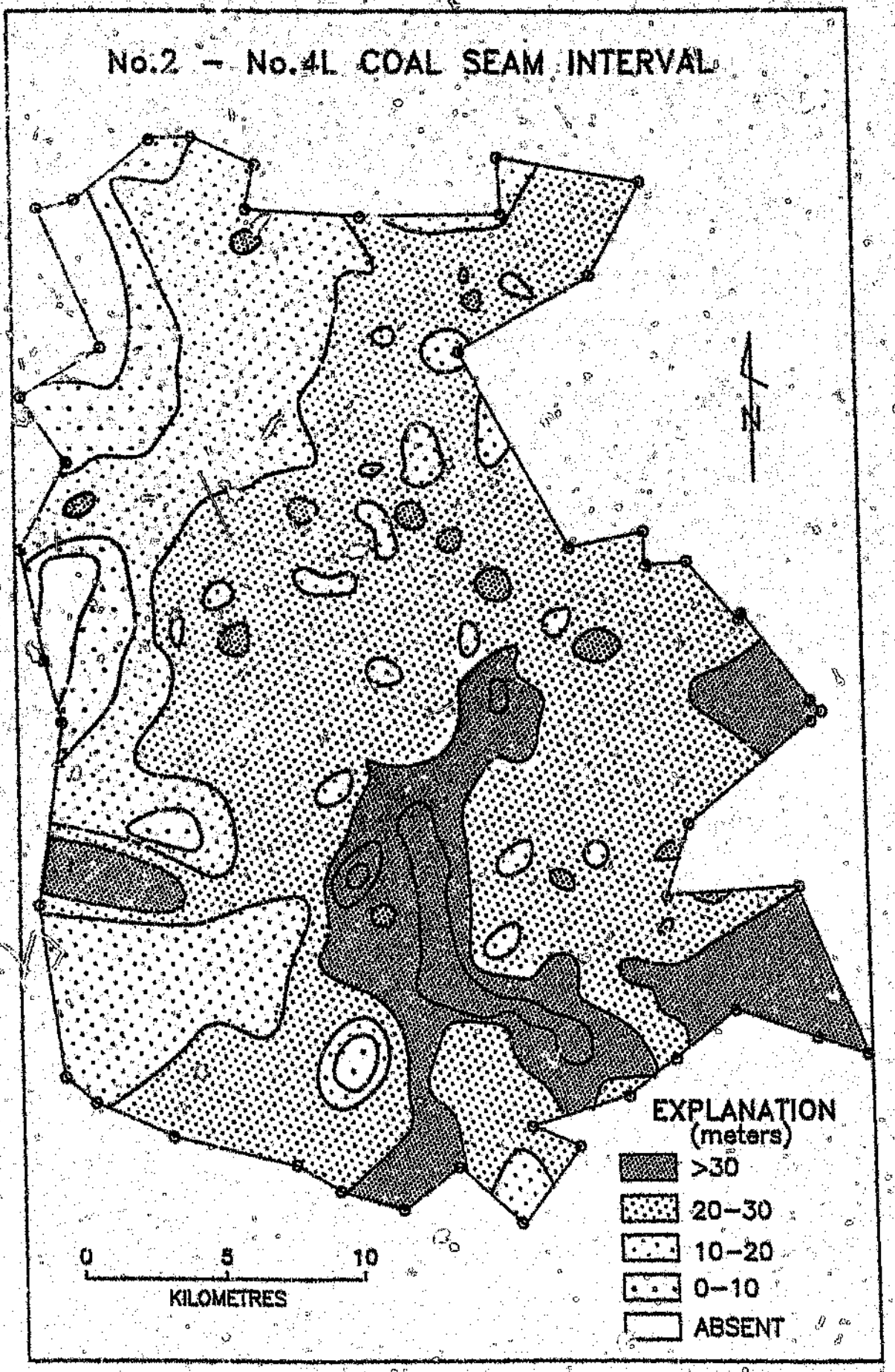


Figure 2.15. Isopach of the No. 2 to No. 4 Lower seam sedimentary interval. Note the north-south linear thickening of this interval.

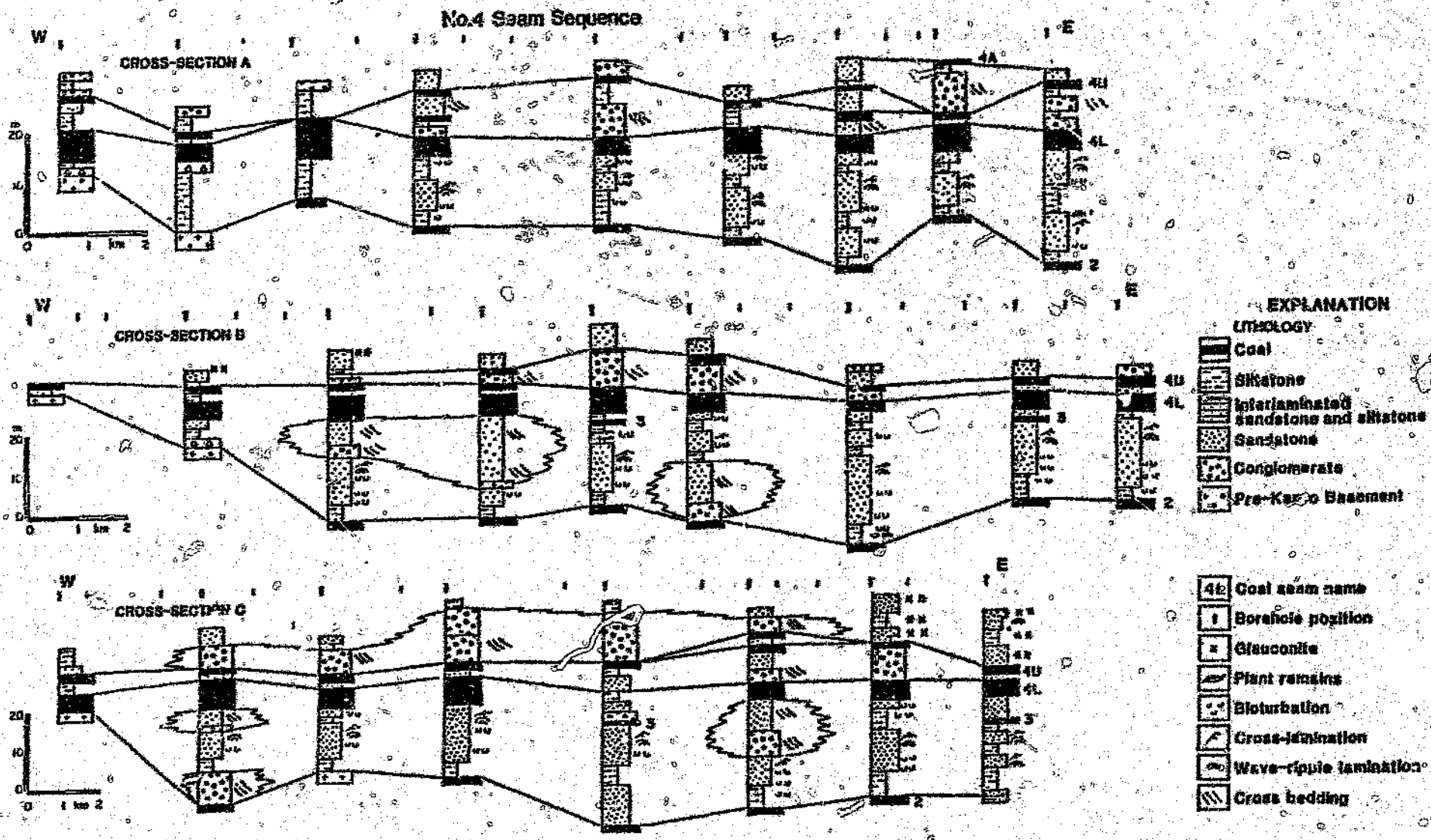


Figure 2.16. Westeast cross-sections of the No. 4 seam depositional sequence. Note the complex splitting of the No. 4 seam and the lithological variation constituting the seam. See Fig. 2.5 for the location of cross-sections.

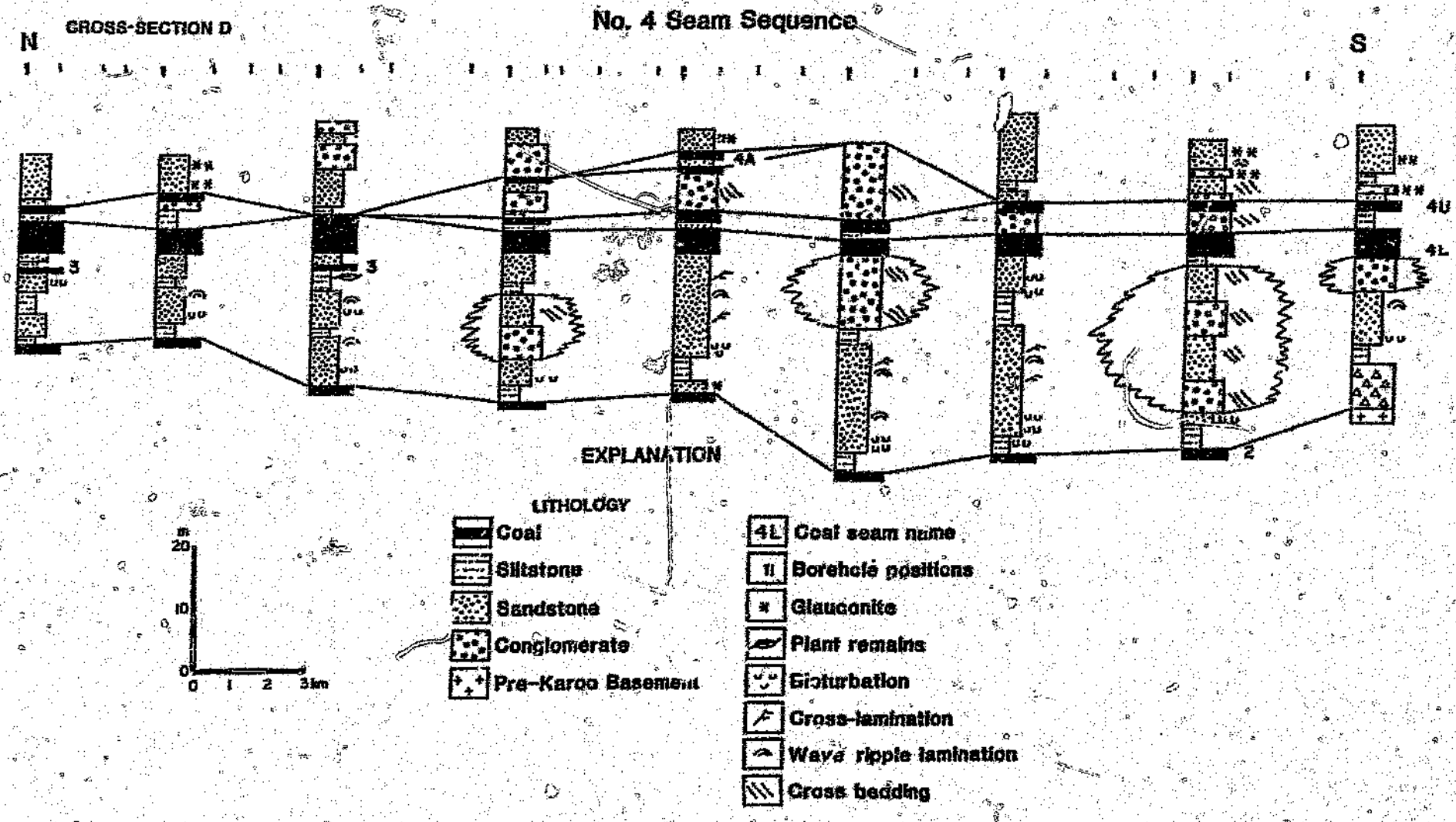


Figure 2.17. Northsouth cross-section of the No. 4 seam depositional sequence. The depositional sequence thickens to the south. See Fig. 2.5 for the location of the cross-section.

erosively overlain by granule-grade conglomerates and coarse-grained sandstones (Figs. 2.14 and 2.16). The distribution of sediment within this interval is illustrated in Figure 2.18. The sediments thicken to the east and pinch out against basement in the west. The No. 3 seam is laterally impersistent and < 1m thick (Fig. 2.19) consequently, in many parts of the study area, the No. 4 seam caps the No. 2-No. 3 seam interval (Figs. 2.14B, 2.16 and 2,17).

The interval between the No. 3 and No. 4 seams is lithologically variable. It commonly comprises a carbonaceous siltstone which grades upwards into an interlaminated sandstone-siltstone (Figs. 2.16 and 2.17). Figure 2.16C illustrates the variation in stratigraphy as here the interval comprises fining-upward granule conglomerate, fine-grained sandstone and carbonaceous mudstone. The distribution of the No. 3 to No. 4 seam interval is shown in Fig. 2.20. The interval is <4m thick and thickens to over 12m in the southeast. A variant to the intervals described occurs in the west of the study area, where the No. 2 to No. 4 seam depositional sequence thins to <10m and comprises diamictite overlain by only one coarsening- upward sequence capped by the No. 4 seam (Figs. 2.14B and 2.16B).

The No. 4 coal seam is split, by sedimentary partings, into three subseams termed; the No. 4 Lower, the No. 4 Upper, and the No. 4A seams (Fig. 2.14A). The No. 4 Upper seam is laterally extensive which allows the interval between the No. 4 Lower and No. 4 Upper seams to be isopached (Fig. 2.21). In general, the interval varies between 0 and 3m in thickness except for the area in the middle of the study area. Here, a north-south thickening between 3 and 15m is present which defines a linear body of granule conglomerate and sandstone (Figs. 2.16 and 2.21). The conglomerate and sandstone is planar cross-bedded and is overlain by laminated carbonaceous siltstone containing plant remains (Fig. 2.14A). Elsewhere, this interval comprises either, carbonaceous siltstone or an upward-coarsening sequence of siltstone grading upwards into medium- or coarse-grained sandstone (Figs. 2.16 and 2.17).

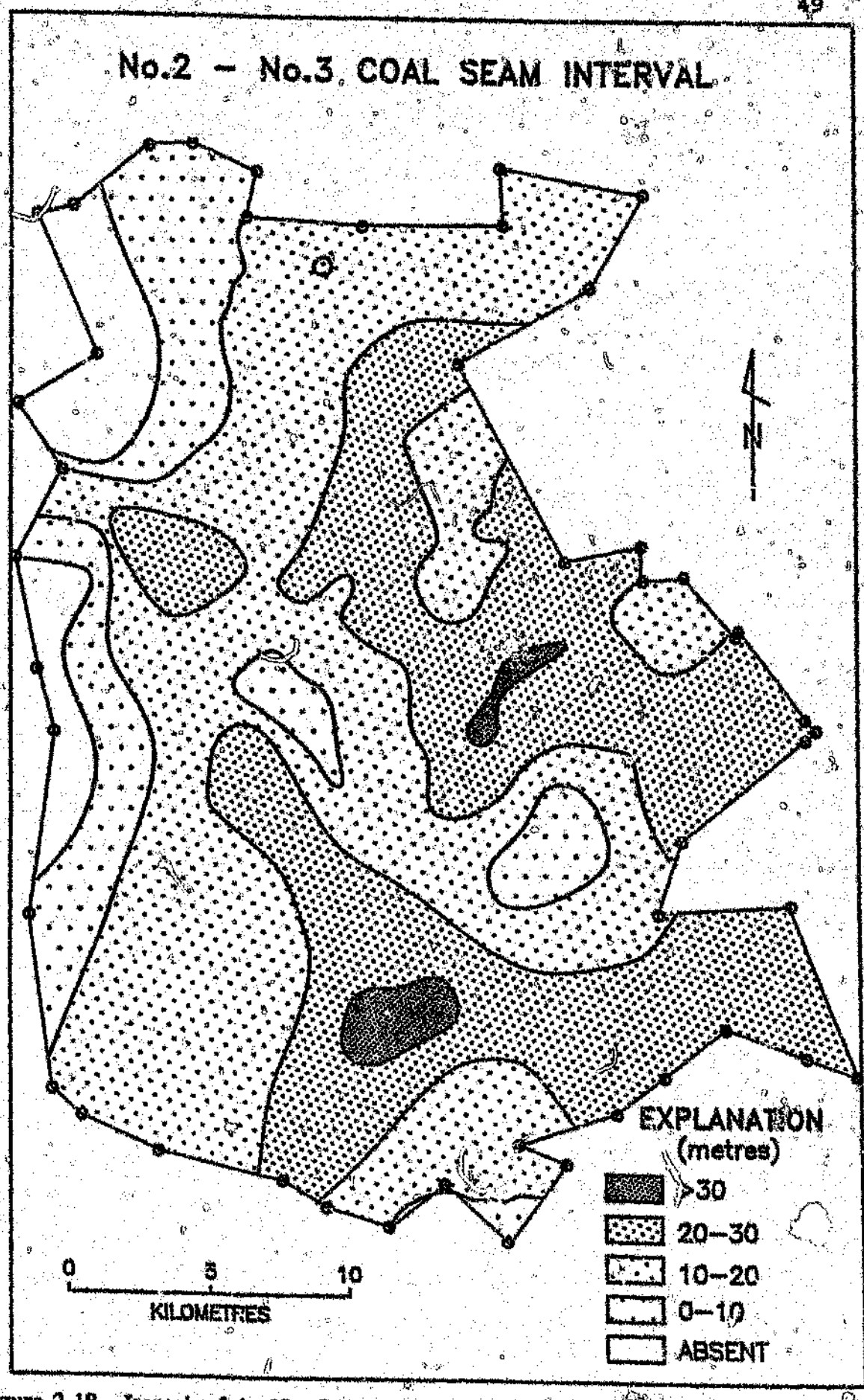


Figure 2.18. Isopach of the No. 2 to No. 3 seam sedimentary interval. The interval thickens to the east and thins towards, and pinches out against, basement in the west.

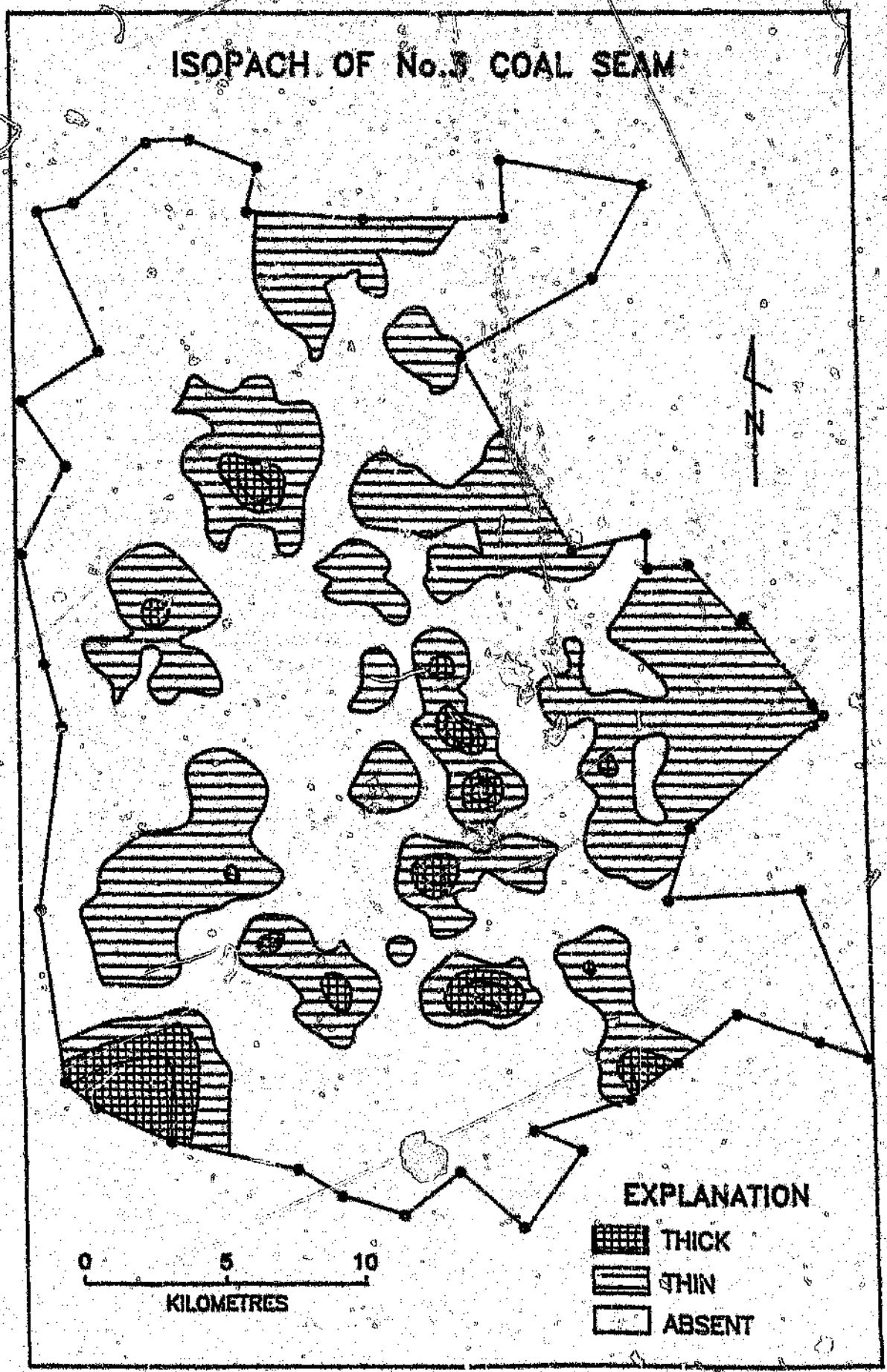


Figure 2.19. Isopach of the No. 3 seam. The seam is thin and laterally impersistent. The terms "thick and thin" are relative terms describing thickness variations of the seam.

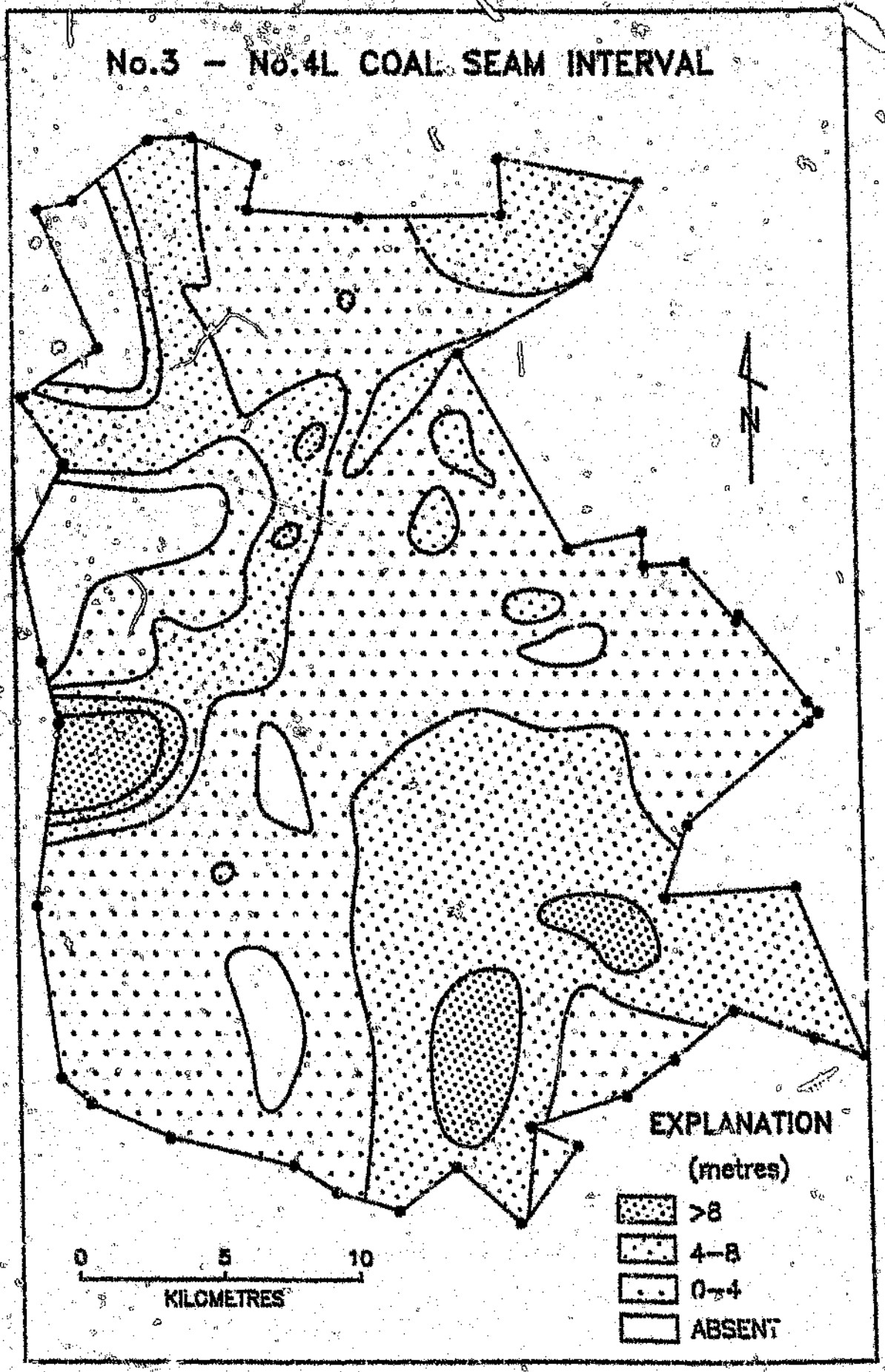


Figure 2.20. Isopach of the No. 3 to No. 4 Lower seam interval. The interval thickens to the south.

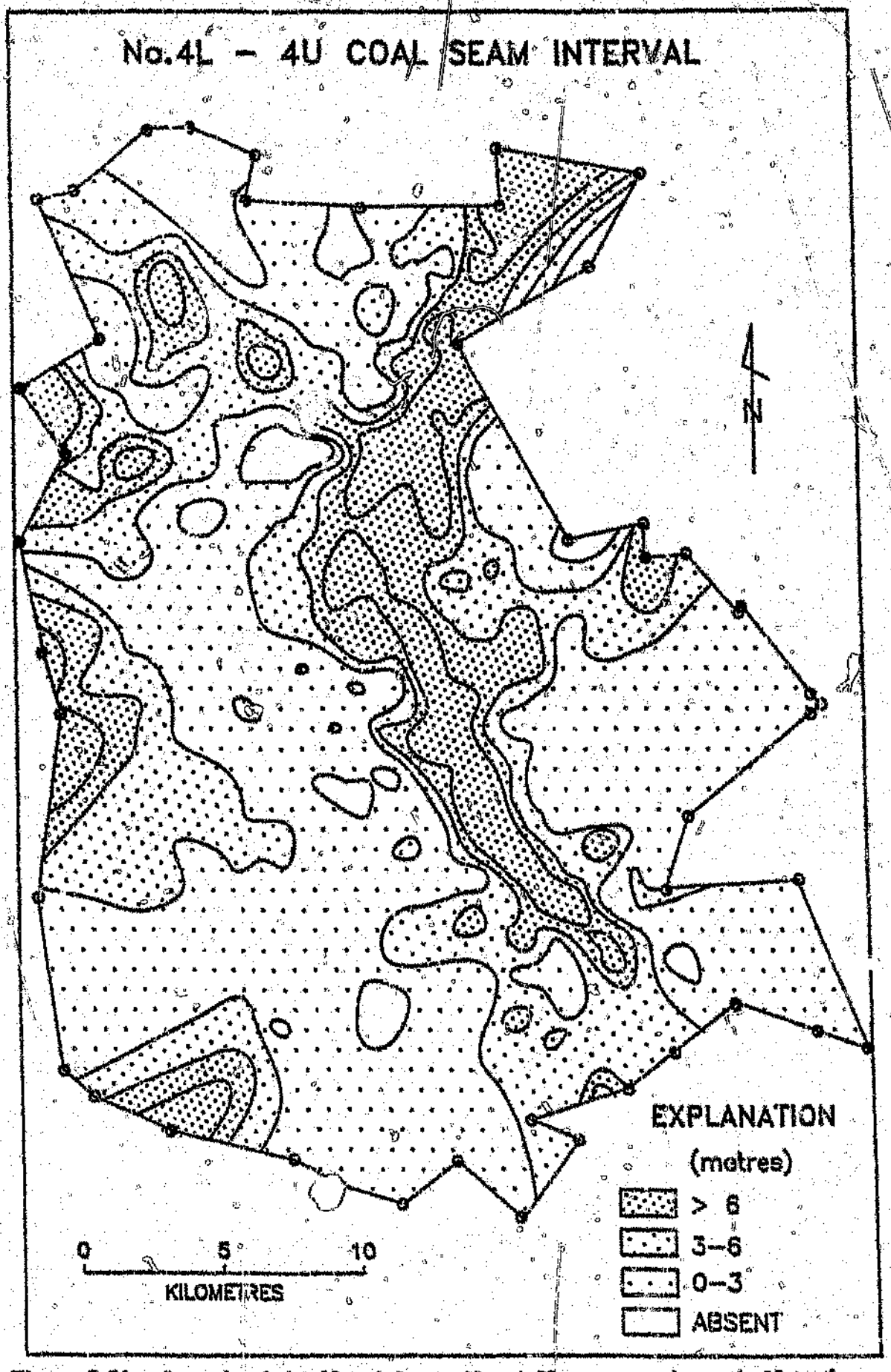


Figure 2.21. Isopach of the No. 4 Lower No. 4 Upper seam interval. Note the northsouth linear thickening of the interval.

sequence of conglomerates, coarse-grained sandstones and carbonaceous siltstones constitute the clastic parting between the No. 4 Upper and No. 4A seams (Figs. 2.14 and 2.16). However, in many instances, these lithologies are overlain by the glauconitic sandstones and siltstones of the succeeding depositional sequence. This is due to the lateral impersistence of the No. 4A seam (Fig. 2.17).

The No. 4 Lower seam is the thickest and most laterally extensive subseam of the No. 4 seam (Fig. 2.22). The seam is in general >4m thick and thins and pinches out against basement in the west. It also thins, and is absent, in a linear belt which corresponds to the thickening of the underlying No. 2 to No. 4 seam interval (compare, Figs. 2.15 and 2.22). The No. 4 Upper seam is thinner than the No. 4 Lower seam attaining a general thickness of >1m (Fig. 2.23). Two areas where the No. 4 Upper seam is >1m thick are present either side of the axis of maximum sediment fill for the No. 4 Lower-No. 4 Upper seam interval (compare Figs. 2.21 and 2.23).

Although the No. 4 seam is complexly split, small-scale clastic partings are present within the No. 4 Lower and No. 4 Upper seams. These clastic partings, display splay-like geometries, are usually <1m thick and, in general, comprise carbonaceous siltstone (Figs. 2.24 and 2.25). In areas where the partings are >1m thick the partings comprise upward-coarsening siltstone-sandstone couplets.

2.6 No. 5 SEAM DEPOSITIONAL SEQUENCE

The No. 5 seam depositional sequence represents one of the most laterally persistent depositional sequences in the Transvaal Coalfields (Cadle *et al.*, 1990). It is relatively easy to distinguish stratigraphically, as it represents a coarsening-upward sedimentary interval between the No. 4 and the No. 5 seams. The depositional sequence commences with, and is terminated by, glauconitic conglomerates, sandstones and siltstones

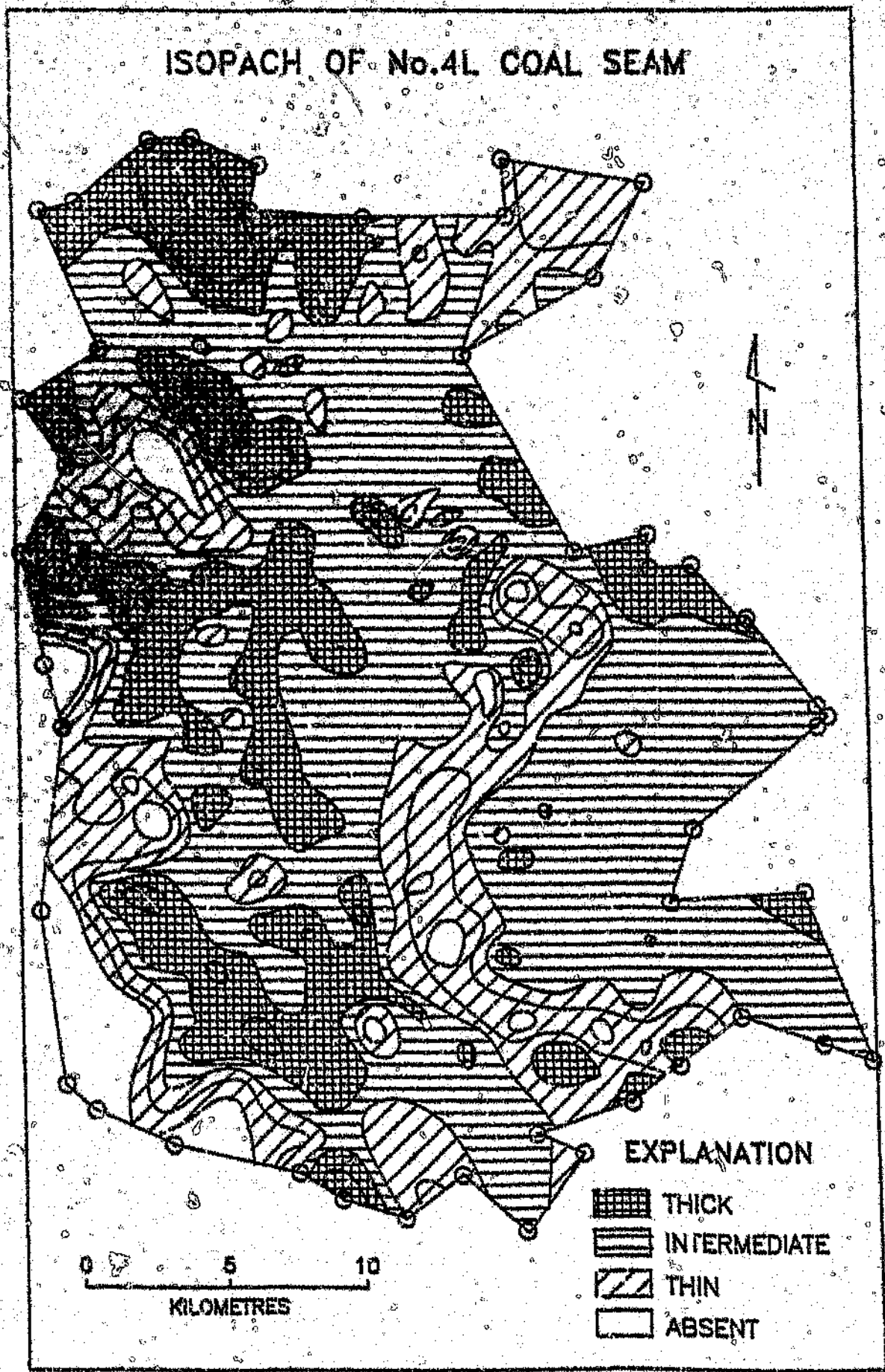


Figure 2.22.. Isopach of the No. 4 Lower seam. The seam thins and is absent in the southwest. There is also a linear thinning of the seam in the southern part of the study area. The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam.

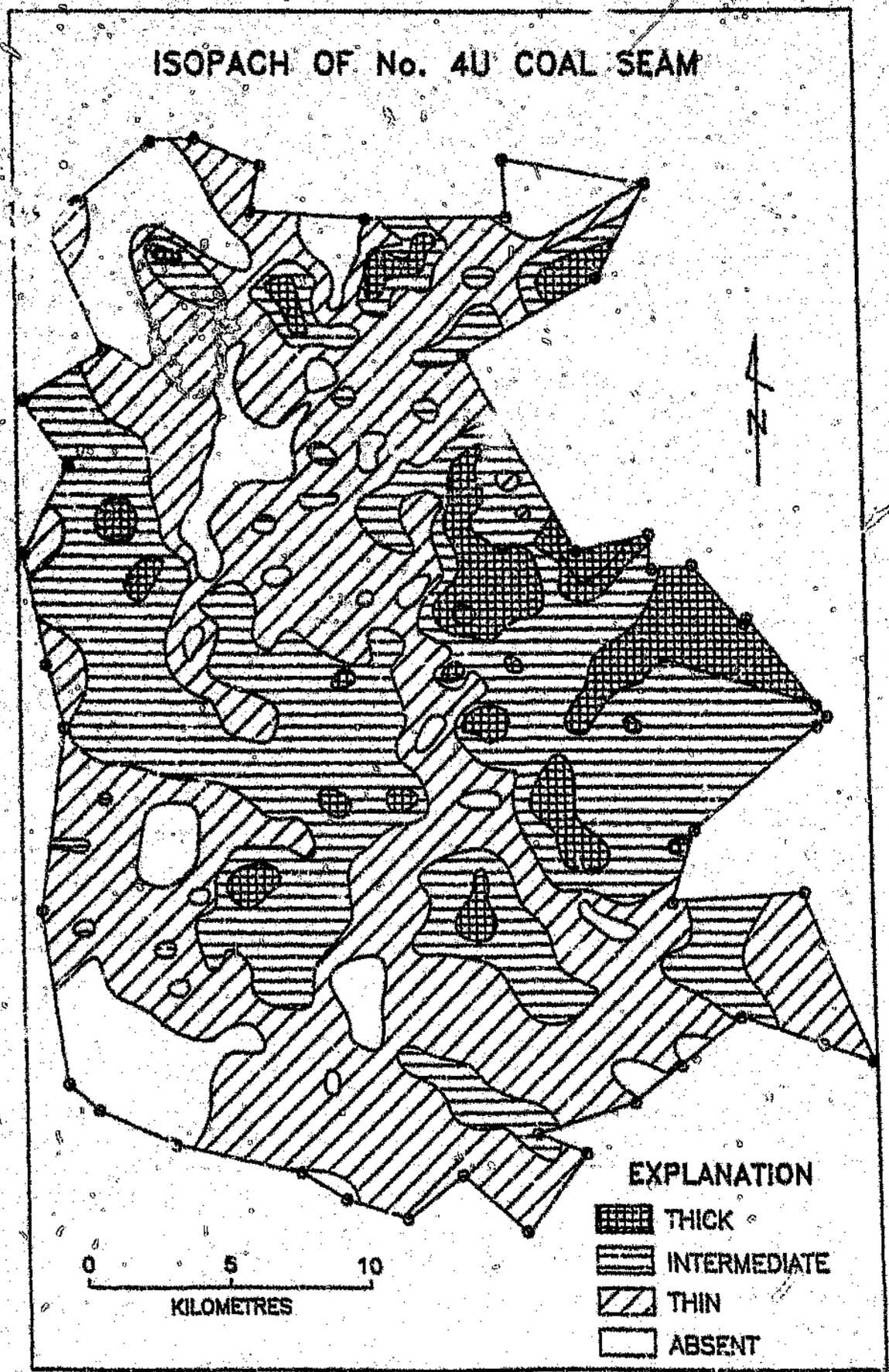


Figure 2.23. Isopach of the No. 4 Upper seam. The seam thickens either side of the underlying No. 4 Lower-No. 4 Upper seam clastic parting (See Fig. 2.21). The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam.

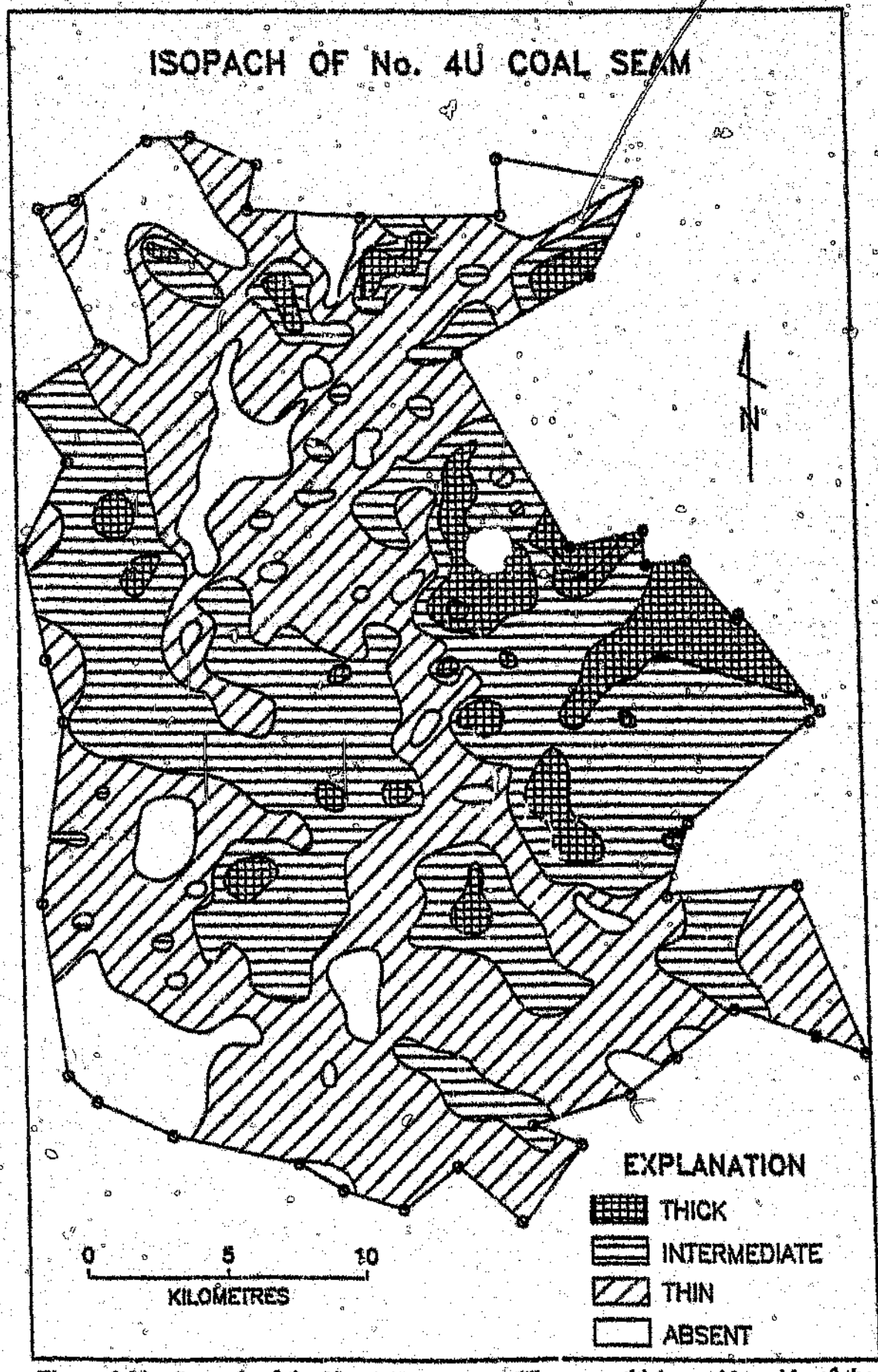


Figure 2.23. Isopach of the No. 4 Upper seam. The seam thickens either side of the underlying No. 4 Lower-No. 4 Upper seam clastic parting (See Fig. 2.21). The terms "thick, intermediate and thin" are relative terms describing thickness variations of the seam.

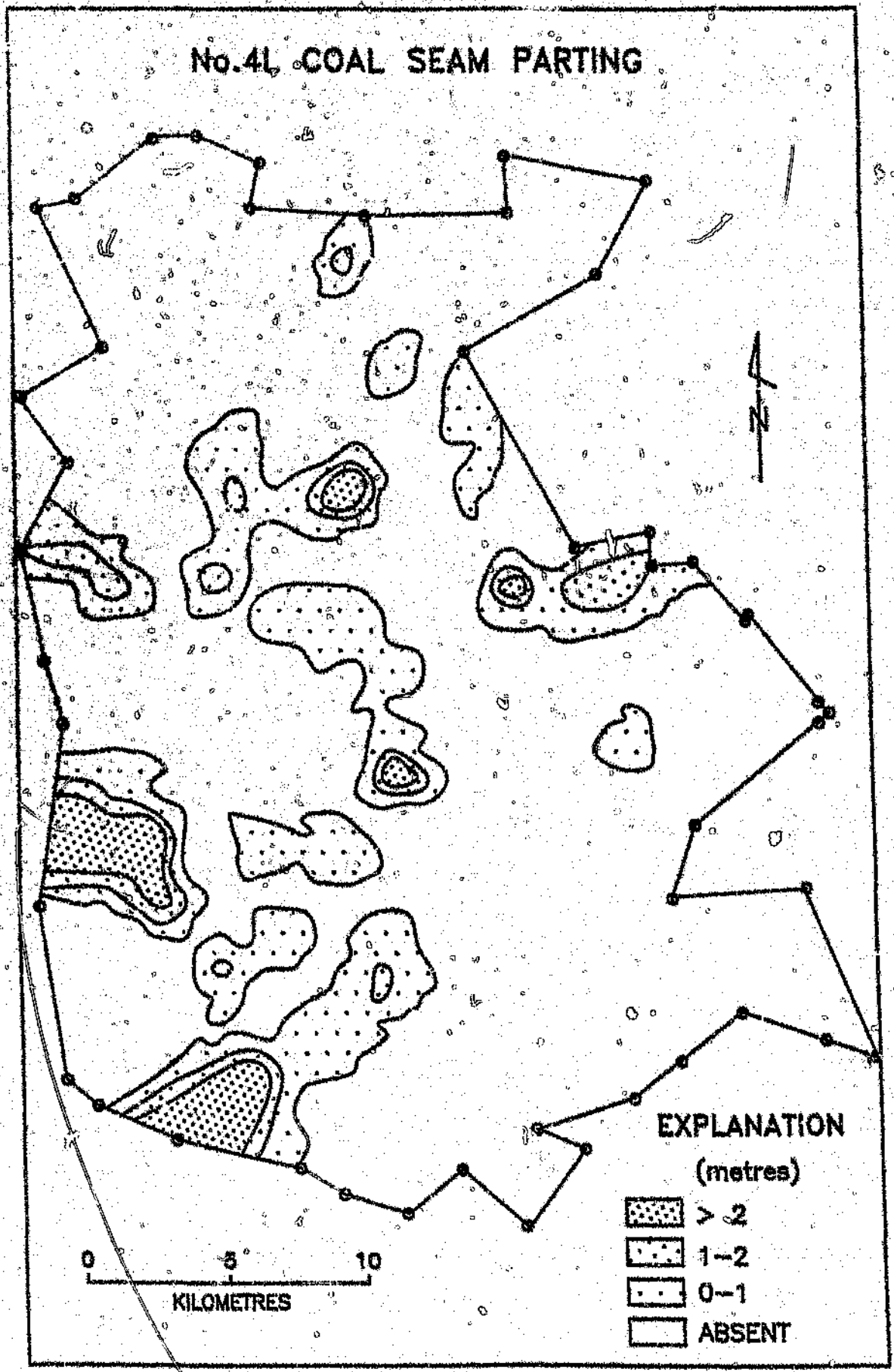


Figure 2.24. Isopach of the clastic parting within the No. 4 Lower seam. Note the splay-like geometries of the parting.

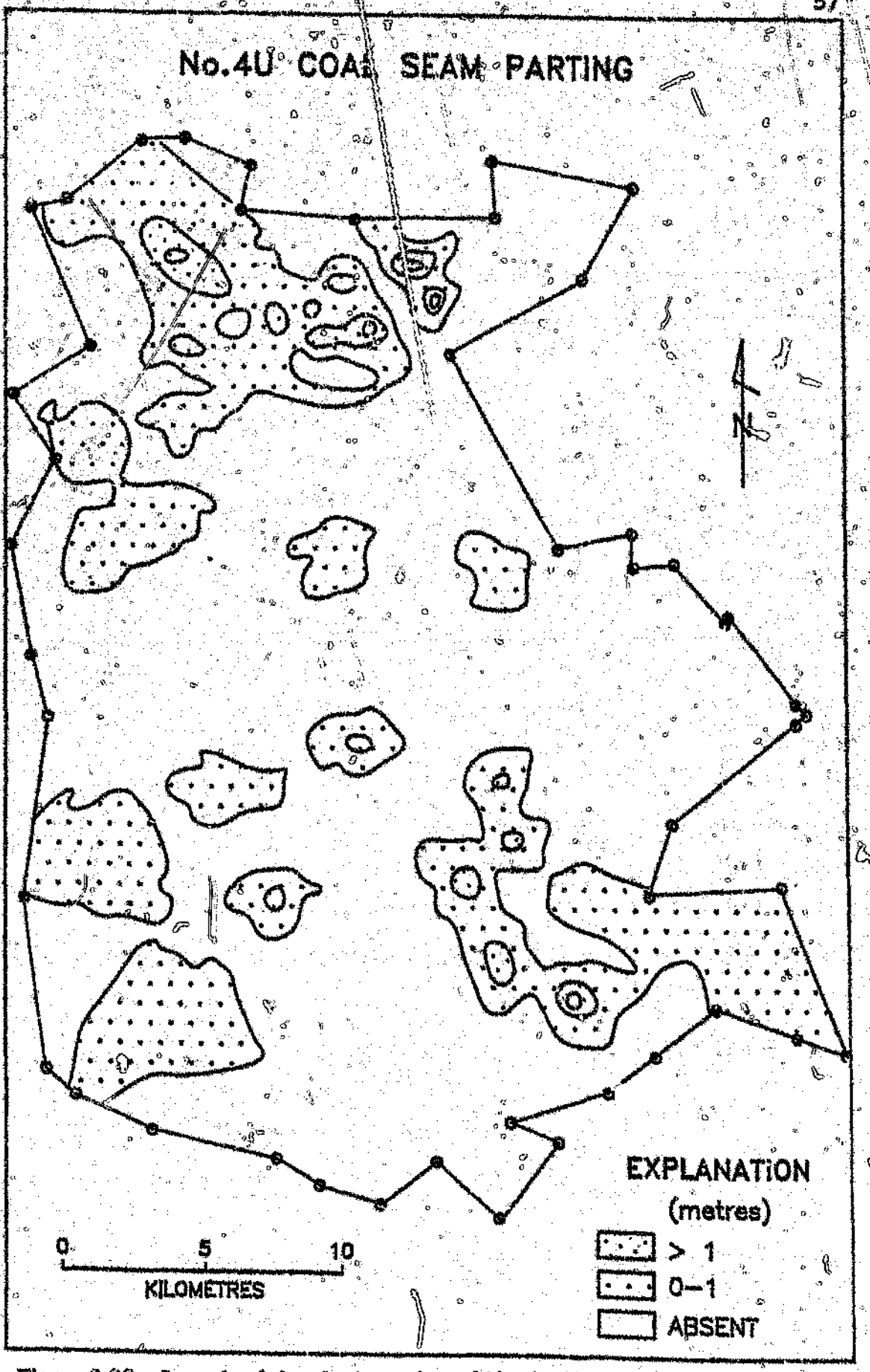


Figure 2.25. Isopach of the clastic parting within the No. 4 Upper seam. Note the splay-like geometries of the parting.

(Fig. 2.26). In the study area, the sequence commonly commences with either, a thin granule conglomerate or, a massive to cross-laminated sandstone or siltstone. These lithologies are often pale to dark green in colour due to the presence of glauconite. Overlying this basal sedimentary unit between one and three upward-coarsening sequences are present (Figs. 2.27 and 2.28). These sequences commence with lenticular and wave-ripple laminated carbonaceous siltstones and grade upwards into wave-rippled and trough cross-bedded sandstones which are bioturbated towards the base. The sequence is capped by the No. 5 seam. The depositional sequence thickens in a southerly direction where it reaches in excess of 40m locally (Fig. 2.29). The sequence thins and pinches out towards the west against basement lithologies. Also notable is a north-south thinning of this interval which corresponds with the area over which the underlying No. 4 Lower-No. 4 Upper seam interval thickens (compare Figs. 2.21 and 2.29).

Isopach maps of the siltstones and sandstones situated between, the basal glauconitic sediments overlying the No. 4 seam, and the No. 5 seam, illustrate the detailed fill of this depositional sequence (Figs. 2.30 and 2.31). The siltstones are uniformly distributed between 0-10m across the study area (Fig. 2.30). The sandstones however, thicken towards the east and south and thin towards the southwest. This easterly thickening of these sandstones may imply sedimentation from east to the west (Fig 2.31). Despite a thinning of the sandstones towards the southwest the influence of basement on the thickness of this depositional sequence is less pronounced than for underlying depositional sequences.

The No. 5 seam, the uppermost of the laterally persistent coal seams, ranges in thickness between 0 and 3m (Fig. 2.32). The seam is less than 1m thick over much of the area, and is absent in the southeast and in isolated patches along the western portion

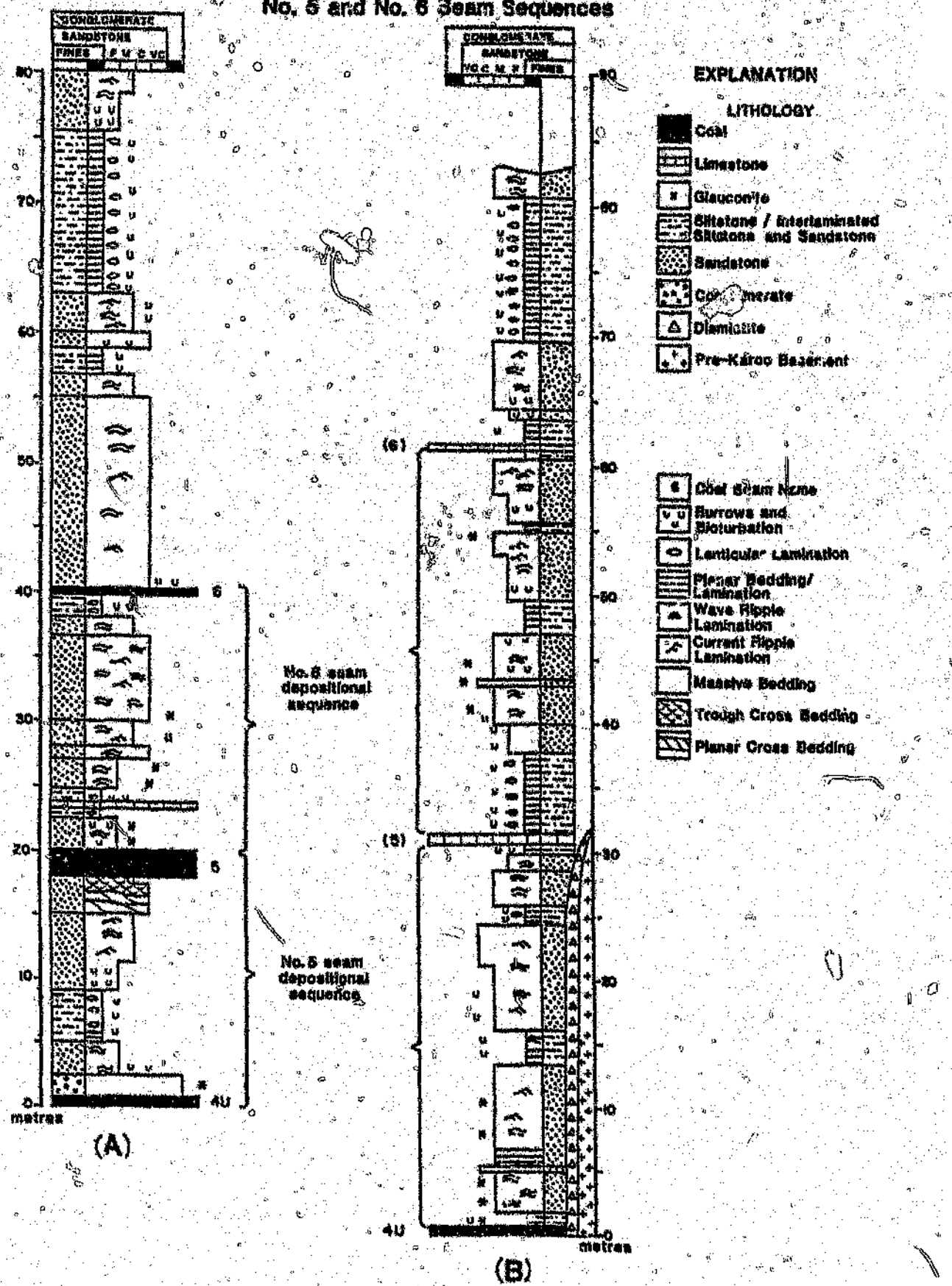


Figure 2.26. Composite stratigraphic profiles of the No. 5 and No. 6 depositional sequences and the interval above the No. 6 seam. Profiles (A) and (B) illustrate the stratigraphic variations.

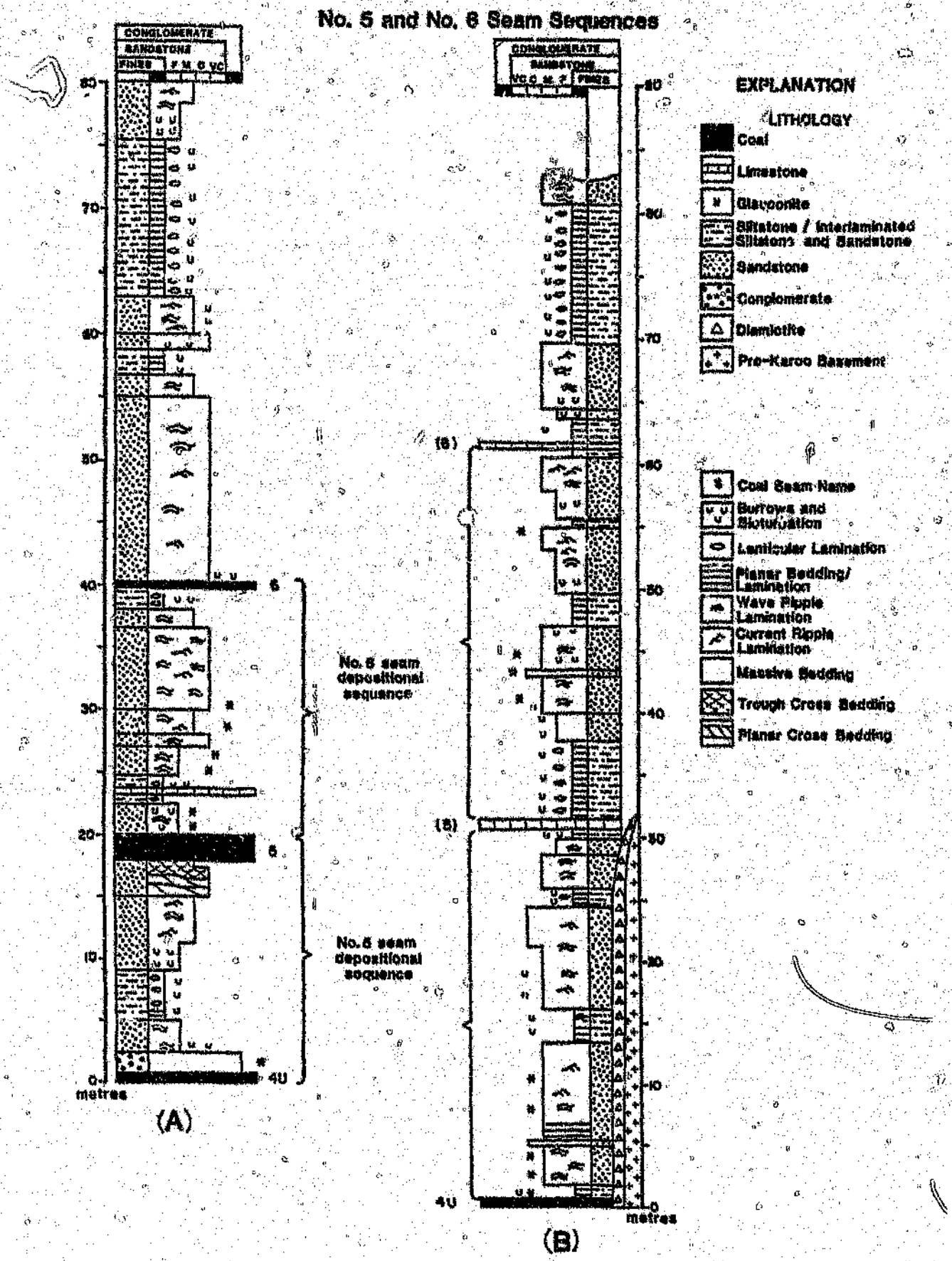


Figure 2.26. Composite stratigraphic profiles of the No. 5 and No. 6 depositional sequences and the interval above the No. 6 seam. Profiles (A) and (B) illustrate the stratigraphic variations.

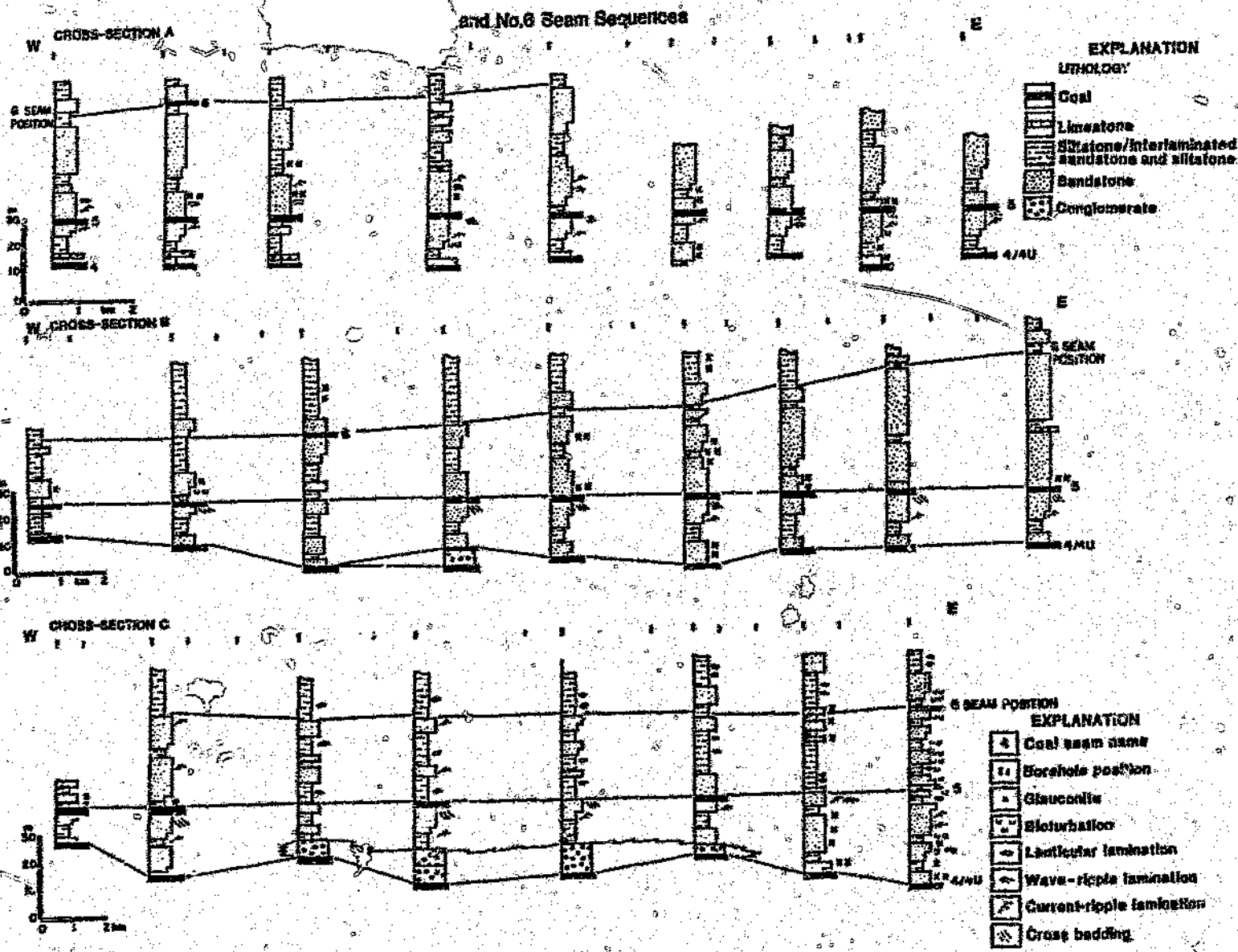


Figure 2.27. Westeast cross-sections of the No. 5 and No. 6 depositional sequences and the interval above the No. 6 seam. See Fig. 2.5 for the location of cross-sections.

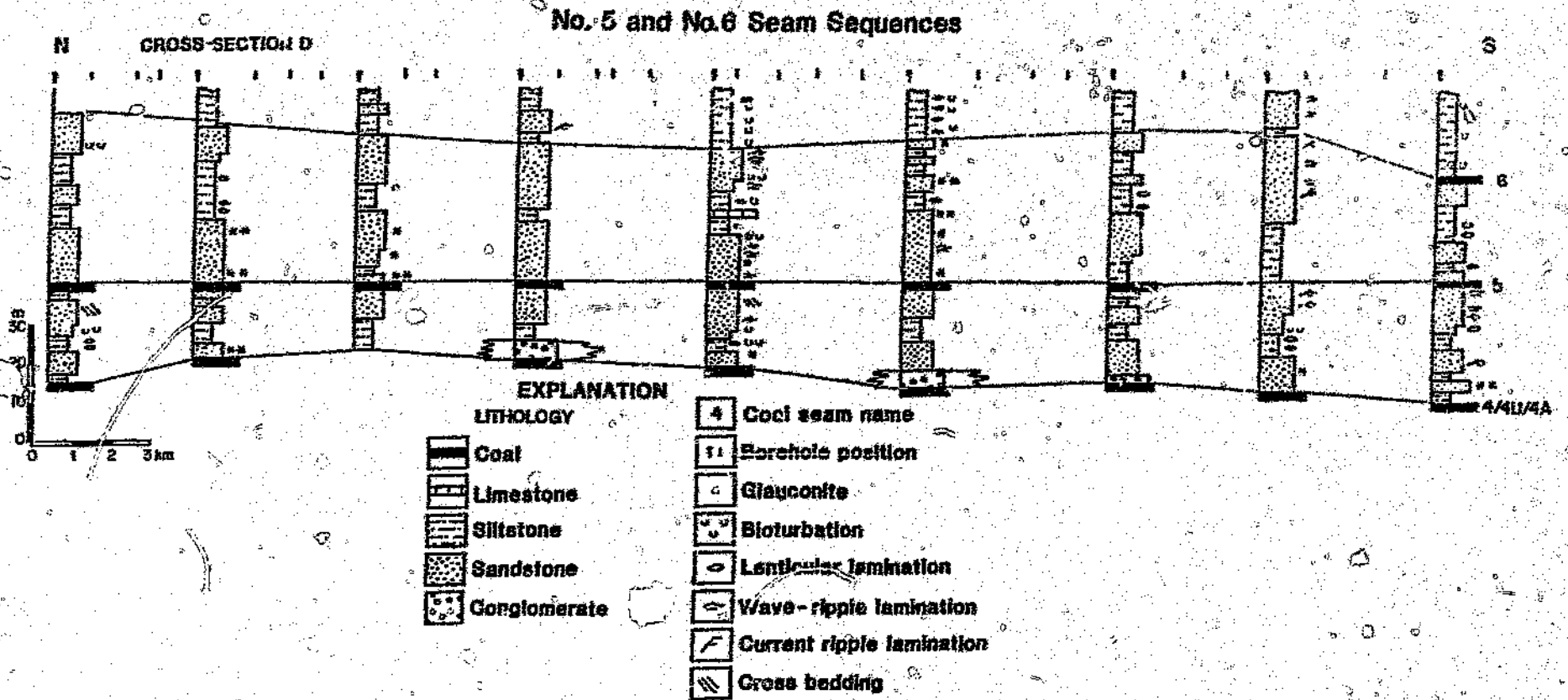


Figure 2.28. Northsouth cross-section of the No. 5 and No. 6 seam depositional sequences and the interval above the No. 6 seam. See Fig. 2.5 for the location of the cross-section.

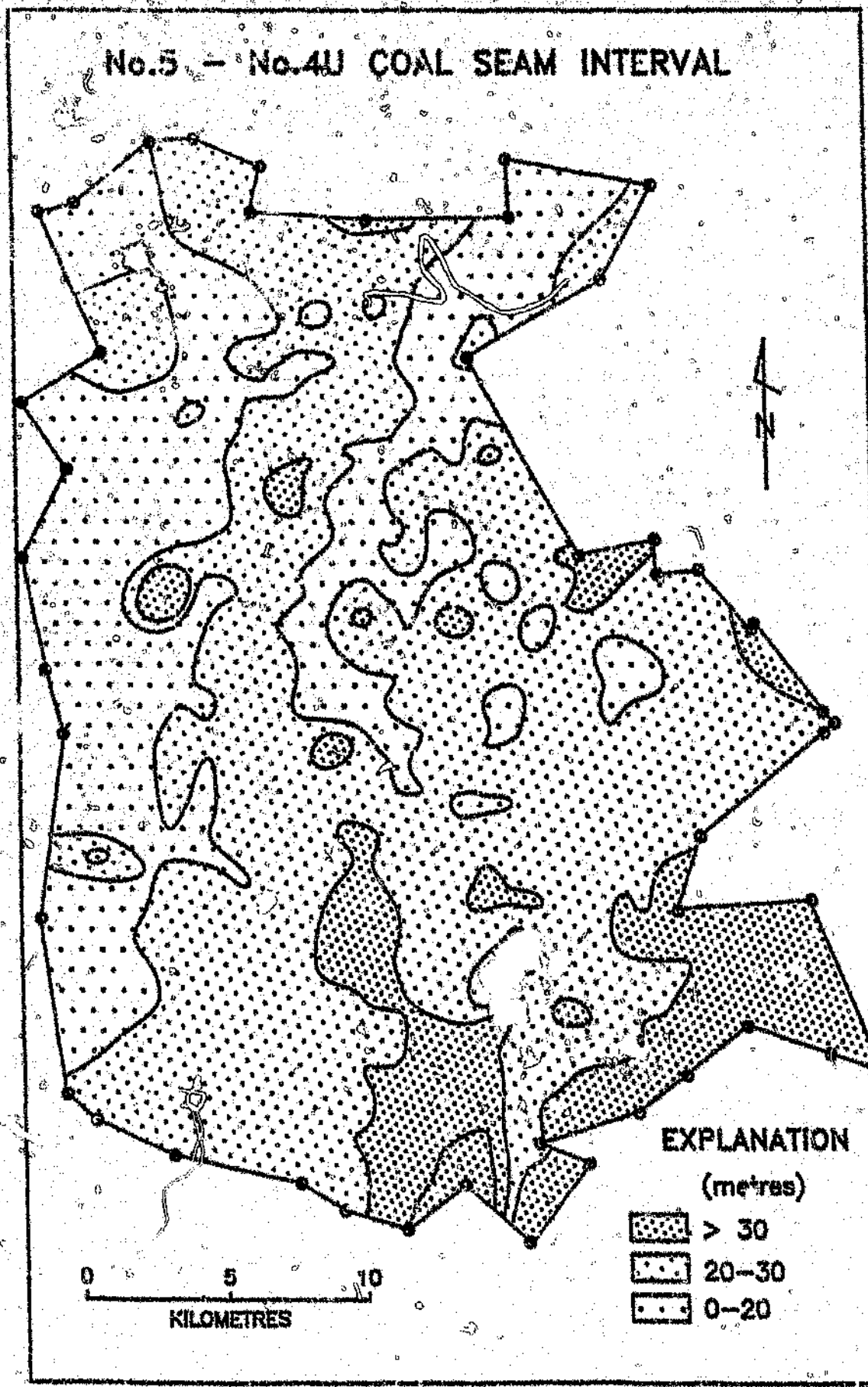


Figure 2.29. Isopach of the No. 5 seam depositional sequence. Note the thinning of the sequence to the west and thickening of the sequence to the south.

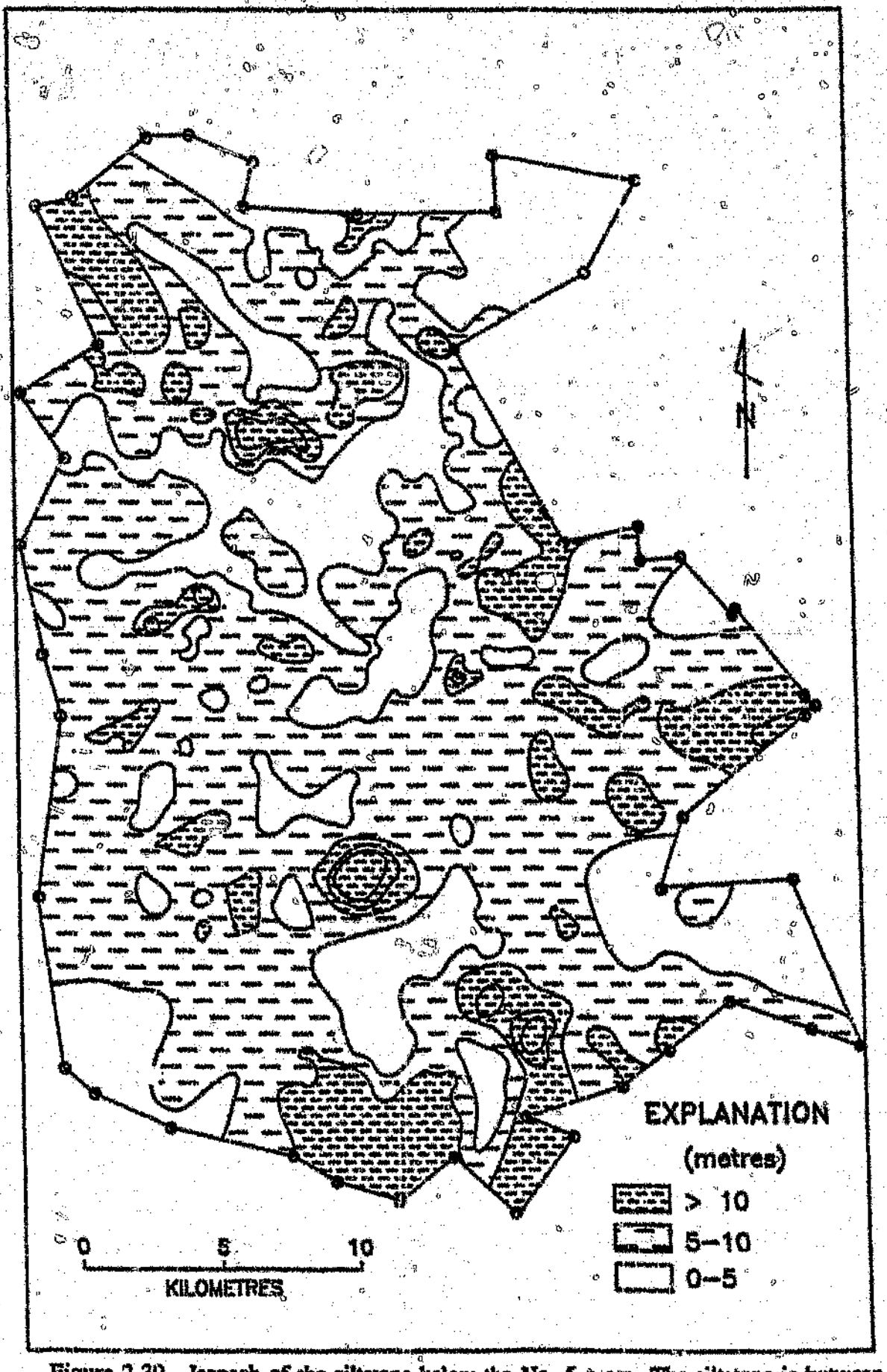


Figure 2.30. Isopach of the siltstone below the No. 5 seam. The siltstone is between 5m and 10m thick over much of the study area.

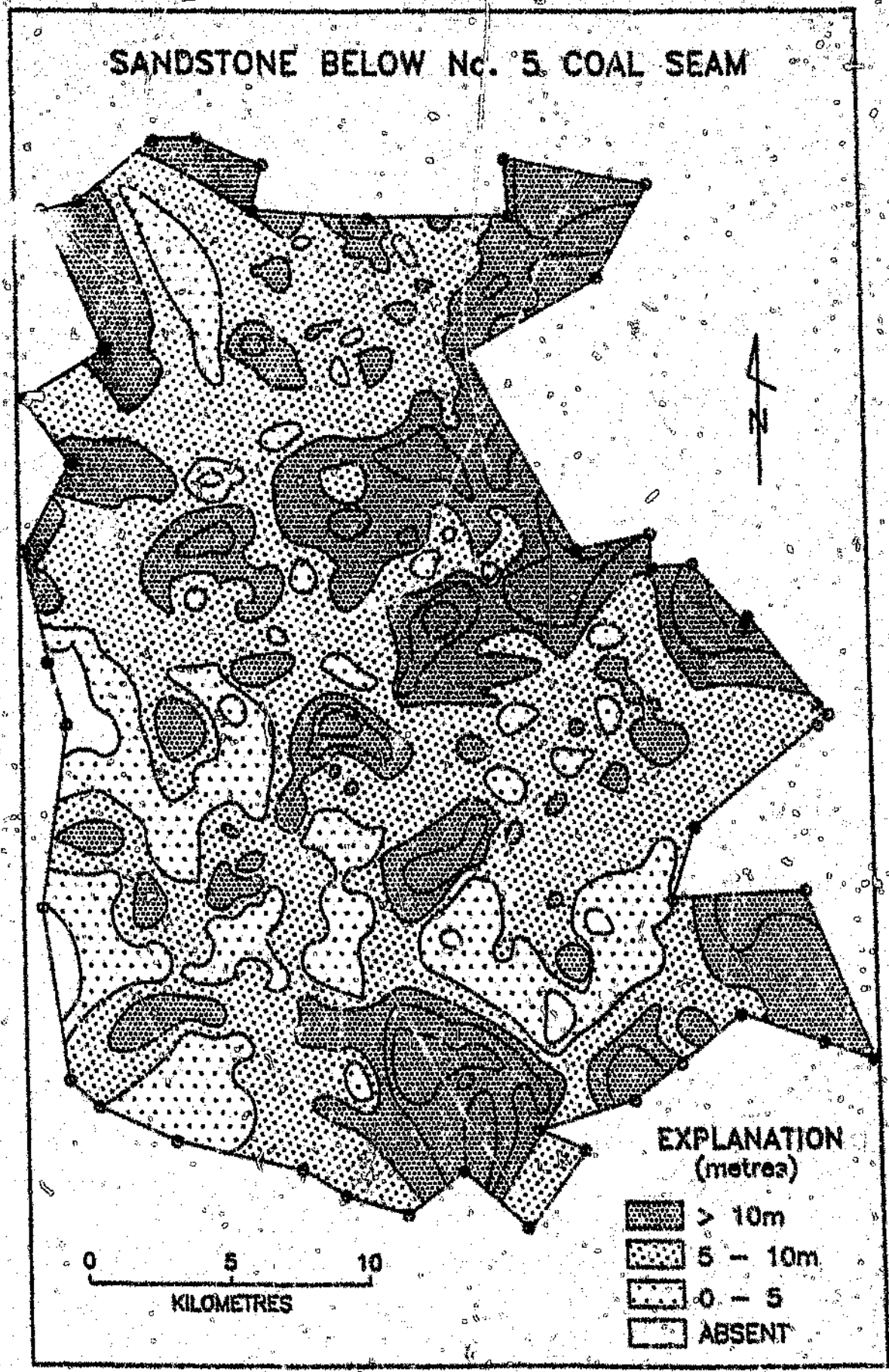


Figure 2.31 Isopach of the sandstone between the No. 5 seam and the underlying siltstone. The sandstone thins towards the west and thickens eastwards.

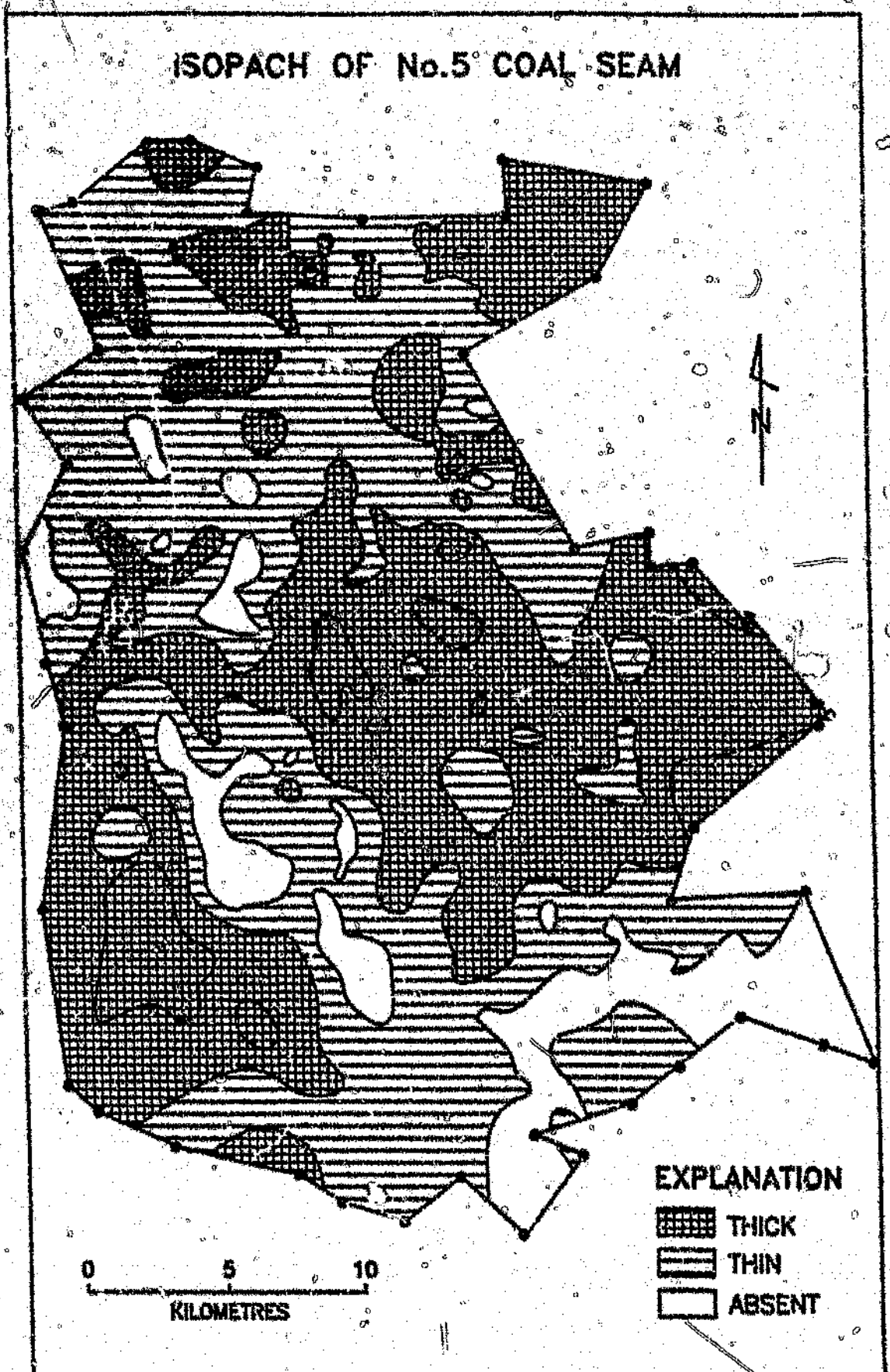


Figure 2.32. Isopach of the No. 5 seam. The seam is thin and laterally persistent. The terms "thick and thin" are relative terms describing thickness variations of the seam.

of the study area. In those areas where the No. 5 seam is absent, the equivalent lithology is a carbonaceous siltstone or limestone (Figs. 2.26 and 2.27).

2.7 No. 6 SEAM DEPOSITIONAL SEQUENCE

The No. 6 seam depositional sequence represents the interval between the No. 5 and No. 6 seams (Fig 2.26). The sequence is difficult to determine due to the fact that, the No. 6 seam is laterally impersistent and does not constitute a laterally persistent stratigraphic marker. Furthermore, these sedimentary rocks are lithologically similar to the rocks present both above and below the sequence (Fig. 2.26B).

The sequence commences with bioturbated glauconitic sandstones or siltstones which overlie the No. 5 seam. The sandstones range in thickness between a few centimetres and 15m. The siltstones range up to 5m in thickness and are interbedded with thin limestones. Overlying these lithologies are between one and three upward-coarsening sequences comprising siltstones and fine- to medium-grained sandstones (Figs. 2.27 and 2.28). The sandstones are commonly glauconitic and structured by wave-ripple lamination and trough cross-bedding. These sandstones are also massively bedded and display abrupt bases and tops. In these instances, the sandstones are well sorted and medium grained. The top of the depositional sequence is defined by either, the presence of the No. 6 seam or, where absent, by a thin bed of limestone or carbonaceous siltstone (Figs. 2.26 and 2.27C).

An isopach map of this depositional sequence is shown in Fig. 2.33. The areas of maximum thickness are situated in the northern and eastern portions of the study area which correspond with the areas of maximum thickness mapped out for the sandstones underlying the No. 5 seam (Fig. 2.31). The areal distribution of the No. 6 seam is

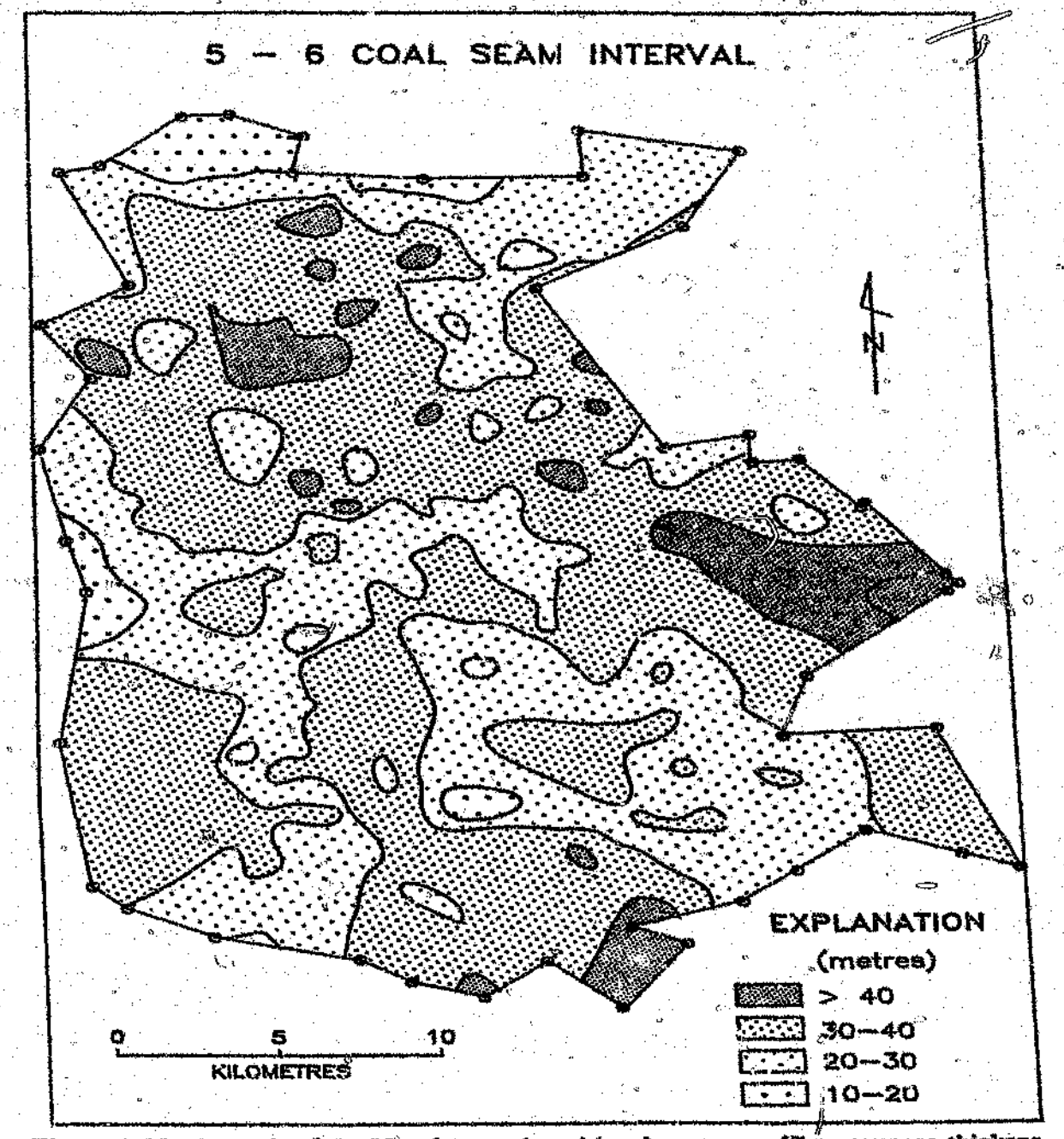


Figure 2.33. Isopach of the No. 6 seam depositional sequence. The sequence thickens eastwards.

shown in Figure 2.34. The seam is laterally impersistent and generally less than 1m thick.

2.8 INTERVAL ABOVE THE No. 6 SEAM

The interval above the No. 6 seam is characterised by the abundance of carbonaceous siltstone over sandstone (Fig. 2.26). Depositional sequences were not isopached due to the fact that:

a. in the northern part of the study area, much of this sedimentary interval is removed through erosion,

b. sandstone bodies are lenticular in geometry and have a random vertical distribution, and

c. there is no distinctive upper marker horizon to delineate depositional sequence boundaries.

The sedimentary rocks present above the No. 6 seam comprise either, an upward-coarsening sequence of carbonaceous siltstone grading up into a fine- to medium-grained sandstone or, a thick, well sorted medium-grained sandstone (Fig. 2.26). The sandstones are bioturbated at the base and structured by cross-lamination. Wave-ripple lamination dominates over trough cross-lamination. Unlike the sandstones above the No. 5 and No. 6 seams, these sandstones infrequently contain glauconite.

Overlying this basal sedimentary unit are several upward-coarsening siltstone-sandstone sequences (Figs. 2.26 and 2.27C). The siltstones are thick (up to 12m), bioturbated and display lenticular and planar lamination. The sandstones are bioturbated, and wave-ripple and trough cross-laminated. This interval of the Vryheid Formation stratigraphy is dominated by an increase in the abundance of bioturbation when compared with the underlying depositional sequences.

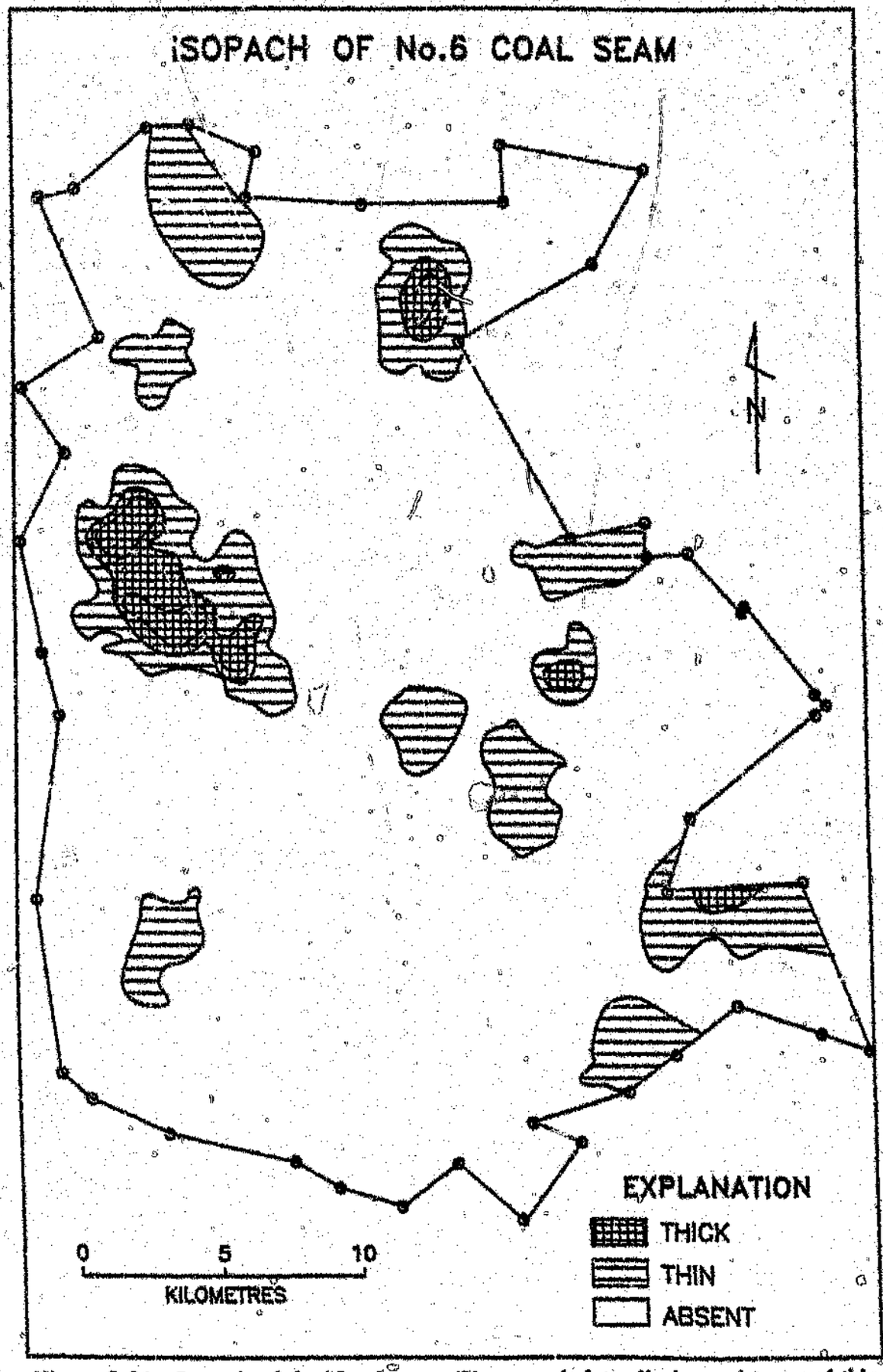


Figure 2.34. Isopach of the No. 6 seam. The seam is laterally impersistent and thin. The terms "thick and thin" are relative terms describing thickness variations of the seam.

In summary, the Dwyka Group and Vryheid Formation of the Ecca Group are subdivided into the No. 1, No. 2, No. 4, No. 5 and No. 6 seam depositional sequences. The identification and three-dimensional mapping of these depositional sequences provides a stratigraphic framework for the study area. The stratigraphic framework used in conjunction with detailed facies descriptions and interpretations provide the foundation upon which depositional environments are deduced.

CHAPTER 3

3 FACIES DESCRIPTION AND INTERPRETATION

3.1 FACIES CONCEPT

The facies concept used in sedimentary rocks has profoundly influenced the understanding of depositional sedimentary environments and depositional systems. Sedimentary facies are the primary building blocks upon which a variety of interpretations are based. Consequently, the importance of sedimentary facies cannot be over emphasised in any interpretation of sedimentary rocks.

The term facies is a Latin term meaning "face, figure, appearance, aspect, look or condition" (Hallam, 1981). The term, which signifies an abstract idea, was introduced into geology by Gressly (1838). The term was used to describe rock strata characterized by similar lithological and palaeontological criteria. Selley (1976), Middleton (1978), Walker (1984), Anderton (1985) and Reading (1986) discuss the ways in which the term facies is used. From its initial application as an objectively defined unit based upon observable differences in lithology and palaeontology, the term has been used in many different ways. Moore (1949) used the term descriptively to focus on "aerially restricted parts of a designated stratigraphic unit which exhibits characteristics significantly different from those of other parts of the unit". The objective approach to facies definition was continued by Selley (1976) who specified five basic parameters on which sedimentary facies could be defined; namely, lithology, sedimentary structures, palaeontology, geometry and palaeocurrent patterns.

The term facies is also used in an interpretative sense i.e. as turbidite facies and geosynclinal facies. Selley (1976) points out the danger of mixing objective and interpretive criteria, as interpretations can change due to the increase in knowledge with time.

In this study, the term facies is used in the same sense as that summarised by Middleton (1978). The facies are described objectively, based upon observable physical features in the rock. The application of the facies concept is also used in the same sense as outlined by Middleton (1978):

"... it is understood that they (facies) are units that will ultimately be given an environmental interpretation; but the facies definition is itself quite objective and based on the total field aspects of the rocks themselves .... The key to the interpretation of facies is to combine observations made on their spatial relations and internal characteristics (lithology and sedimentary structures) with comparative information from other well-studied stratigraphic units, and particularly from studies of modern sedimentary environments."

With the objective of using sedimentary facies to identify sedimentary environments and depositional systems, new facies definitions have evolved to address subsurface investigations in the petroleum industry. Galloway and Hobday (1983) discuss the application of wireline logs, isopach and isolith maps as facies tools in unravelling depositional systems. Mitchum *et al.*, (1977), and Sangree and Widmier (1977), defined and used seismic facies analysis to unravel and interpret the seismic response of sedimentary rocks in the geologic record. This technique, which geophysically records the physical signatures of rocks, is still used in an objective sense and thus still falls within the criteria and spirit of earlier facies definitions (Gressly, 1838; Selley 1976; Middleton, 1978; Walker, 1984; Reading, 1986).

Other workers have used lithofacies associations to interpret specific sedimentary environments. Cant and Walker (1976), Miall (1977, 1978 and 1980), Cant (1978) and Rust (1978) were the first to apply the concept of facies analysis to identify a variety of fluvial systems. This approach has also been successfully applied to submarine-fan systems by Bouma (1962) and Mutti and Ricci Lucchi (1972). During the 1980's, Allen (1983), Friend (1983), Ramos and Saponea (1983) and Miall (1985) introduced the concept of architectural-element analysis to identify fluvial systems. This approach combines lithofacies (Miall, 1978) with mesoform and macroform features such as the three-dimensional geometry of bar forms, compound bars, and channel and interchannel sediments. In this way, fluvial depositional systems are identified and classified.

In summary, two facies approaches are currently used in analysing sedimentary strata and identifying sedimentary environments or depositional systems. The first approach is that used by Miall (1978) whereby individual lithofacies are identified and then combined with mesoscale and macroscale features to arrive at the identification and description of a depositional system. The second approach is that described by Galloway and Hobday (1983) who identify macroscale features such as the three-dimensional characteristics of specific sedimentary rocks. Thereafter, smaller scale features such as lithofacies are identified and then depositional systems interpreted.

Within this study, elements of both approaches are used to define depositional systems. Although this is a subsurface study, the data were fortunately obtained from core holes. Consequently, the luxury of good lithofacies data were provided. Thus, the description and classification of lithofacies is used to identify depositional sequences or episodes (Galloway and Hobday, 1983). The depositional sequences and lithologies within the sequences were mapped out to provide a three-dimensional appreciation of lithofacies groupings. The integration of these data were used to identify sedimentary environments and depositional systems.

From the above discussion, on the application of different facies approaches to the unravelling of depositional systems, refinements to detail within depositional systems are provided by lithofacies. For this reason the lithofacies of the Dwyka Group and Vryheid Formations are described and interpreted. These data then form a basis upon which the interpretations of depositional systems are made.

3.2 SEDIMENTARY FACIES

3.2.1 Diamictite Facies

Description

Diamictite, or matrix-supported conglomerate, forms a thin veneer of sedimentary rock overlying pre-Karoo basement lithologies. The diamictite comprises a dark-grey to green mudrock within which pebbles and cobbles of amygdaloidal lava, chert, granite, and quartzite are scattered (Fig. 3.1A,B). The clast shapes are distinctive and frequently, one or more sides are faceted and striated. The diamictite is structureless and no fabric within the clasts is evident. A variety of lithologies under- and overlie the diamictite. The diamictite is commonly overlain by massive and cross-bedded granule conglomerate or coarse- to very coarse-grained cross-bedded sandstone. In the southeast of the study area, interbedded carbonaceous siltstone and bioturbated sandstone and siltstone under- and overlie the diamictite.

The facies is more thickly developed in palaeodepressions within the basement irrespective of basement elevation. The thickest development of diamictite is confined to a north-south trending palaeovalley situated along the eastern margin of the study

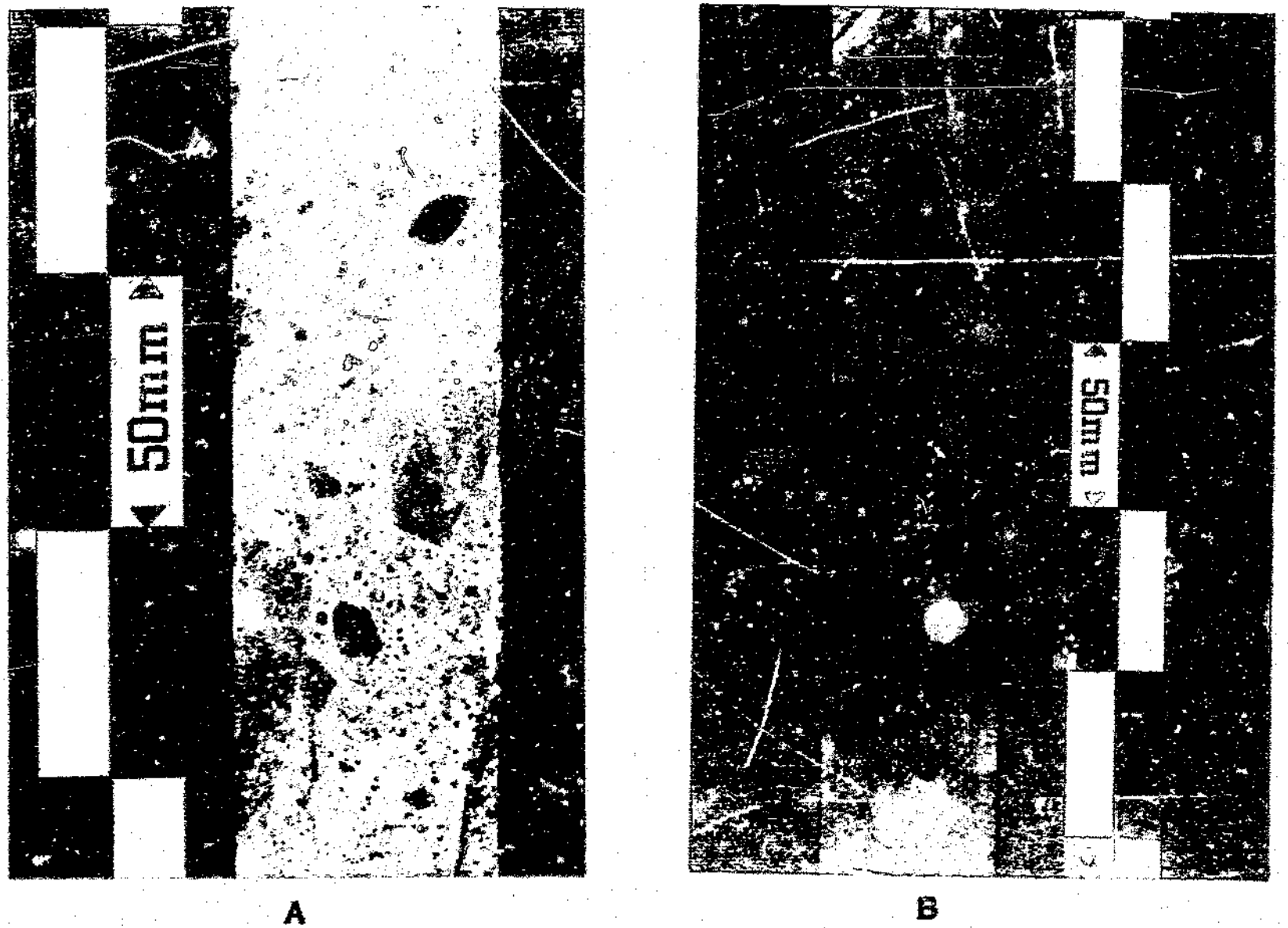


Figure 3.1. The diamictite facies comprising a siltstone matrix and extrabasinal pebbles. Note the faceting of pebbles. Figures (A) and (B) illustrate facies variations.

area, where a maximum thickness of 30m is recorded. In the west of the study area, where the diamictite is absent, the Vryheid Formation rests directly upon the pre-Karoo basement. Lithostratigraphically, the diamictite facies forms part of the Dwyka Group and represents the basal facies of the Karoo Sequence.

Interpretation

Diamictites are not indicative of any particular depositional environment. The sediments are primarily transported and deposited by processes of gravity and ice i.e. slumps, debris flows and melt-out tills. In the case of the Dwyka diamictites, the faceted pebbles, dropstones and striated pavements provide evidence for a glacial origin (Stratten. 1968 and 1970; Von Brunn, 1977; Le Blanc Smith, 1980, 1981b; Visser, 1986; Visser and Loock, 1987).

The lack of graded-stratified and glaciotectonised diamictites suggest the diamictites were deposited by wet-based glaciers rather than by thin inactive glaciers frozen to the substrate. Furthermore, the faceted pebbles suggest that debris was transported at the base of the glacier where intense abrasion between sediment and the substrate takes place (Boulton, 1978; Eyles *et al.*, 1983). Thus, the Dwyka diamictite from the study area is suggestive of a grounded-ice sediment, deposited by wet-based glaciers as a lodgement till.

3.2.2 Conglomerate Facies

This facies comprises sedimentary rocks that contain greater than 30% clasts larger than 2mm (Collinson and Thompson, 1982). An unusual grain-size feature in the rock sequence, is the abundance of clasts that fall within the granule-grade size range of

2-4mm. Two conglomerate facies can be distinguished by the presence or absence of sedimentary structures, namely: the massive to poorly stratified conglomerates, and the cross-bedded conglomerates.

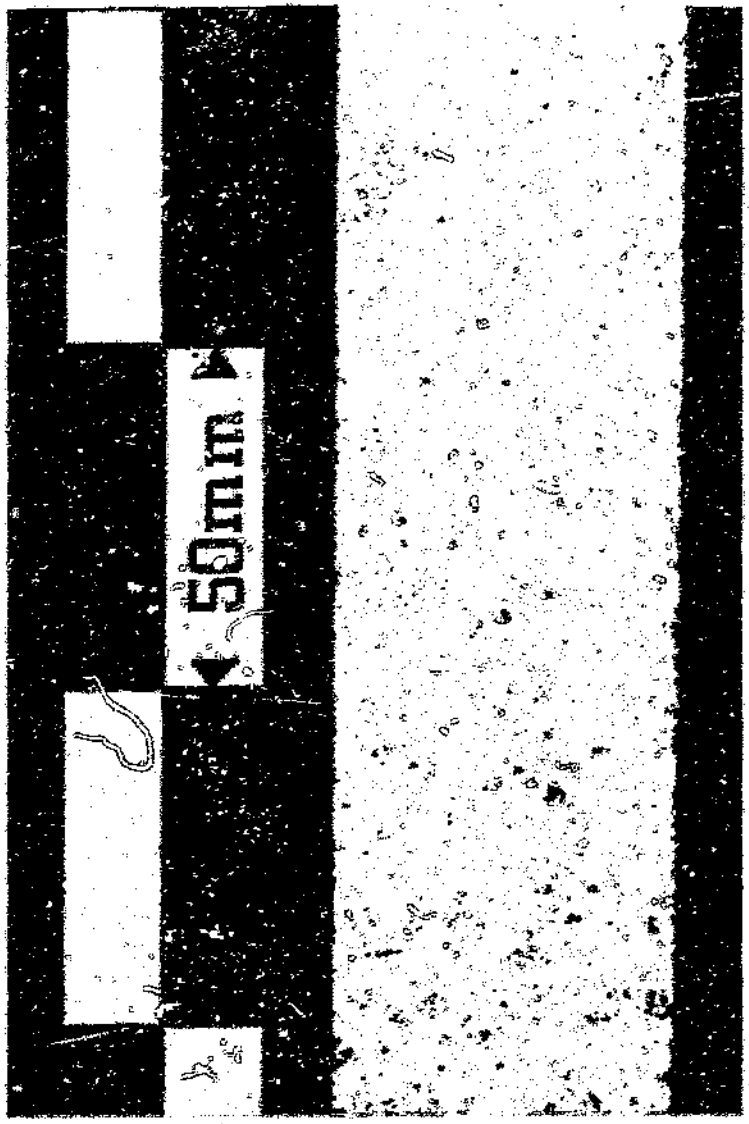
3.2.3 Massive to Poorly Stratified Conglomerate

Description

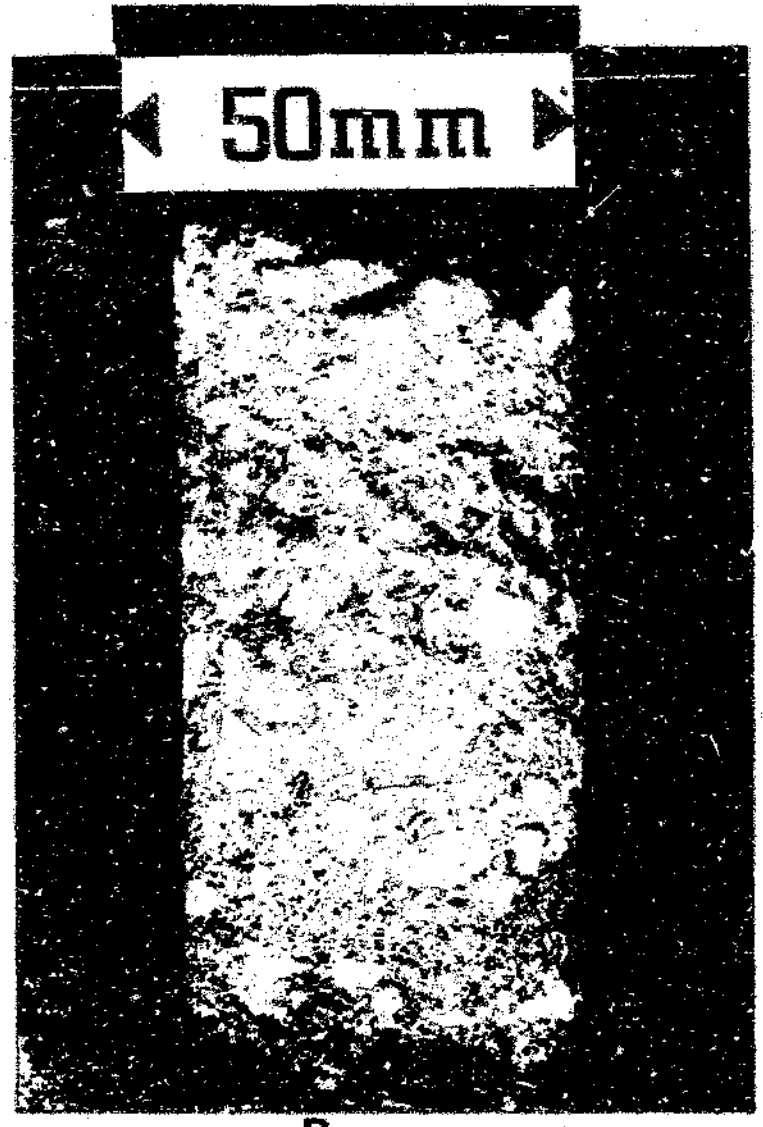
This facies is clast supported, grey to white in colour, and massive to poorly stratified. The matrix comprises either sandstone or siltstone. Based upon sorting, the abundance of pebble size clasts and matrix material, two types of conglomerates are distinguishable:

The first conglomerate type is a well-sorted, arkosic, granule conglomerate comprising subrounded quartz and feldspar clasts (Fig. 3.2A). The conglomerate is massive to crudely stratified. The interstices between clasts is often devoid of matrix. However, some conglomerates contain minor amounts of sandstone and carbonaceous siltstone matrix. The facies is 10cm-2m in thickness, and frequently interbedded with the cross-bedded conglomerate facies. Stratigraphically, this facies is present below the No. 2 seam and between the No. 4 Lower and No. 4 Upper seams.

The second conglomerate type has a high percentage of siltstone matrix and a clast size which varies between 2mm and 5cm in diameter (Fig. 3.2 B). Pebbles comprise, extrabasinal amygdaloidal lava, felsite, granite, vein quartz, and chert. Also present are clasts of intrabasinal siltstone and coal spar. Although primarily massive these



A



B

Figure 3.2. The massive to poorly stratified conglomerate facies comprising granule-grade quartz and feldspar grains (A). The massive to poorly stratified pebble conglomerate with a high matrix content (B).

conglomerates, which are between 1cm and 50cm thick, also display normal grading and planar bedding. The facies often displays a sharp to erosive base and comprises the base of fining-upward sequences. Stratigraphically, these sequences are situated below the No. 2 seam and between the No. 4 Lower and 4 Upper coal seams.

Interpretation

Structure within conglomerates is recognised from clast size variation, sorting, fabric changes and grading (Harms *et al.*, 1982). The conglomerates of this facies are well sorted, granule conglomerates and lack those parameters indicative of identifying structure or stratification in conglomerates. The texture of the sediment is source related and represents the weathered products of crystalline granitic rocks exposed to the north of the present basin margin. The infrequent crude stratification and association with cross-bedded conglomerate and sandstone, is suggestive of traction deposition. Harms *et al.*, (1982) suggest that within gravels, clasts experience restrictive forces relative to each other, and are unable to respond individually to fluid stresses, consequently, many gravels and conglomerates are massive in appearance. The crude stratification in the conglomerate is explained as a result of relatively high fluid and sediment discharge such that sediment is deposited downcurrent faster than it aggrades vertically (Hein and Walker, 1977).

The thin conglomerate beds, comprising extrabasinal pebbles, and a high matrix content were probably deposited by high-discharge events followed by rapid deposition. The association of this facies with coal seams and coal clasts suggests deposition in a continental setting where gravels are deposited by fluvial processes. Massive and crudely stratified gravels from fluvial environments have been described by Gustavson (1974), Miall (1977, 1978) and Rust (1978). Glaciofluvial processes are inferred from the association of massive conglomerate facies with Dwyka Group diamictites. Enyon

and Walker (1974), Gustavson (1974) and Ashley (1975), describe similar facies in modern glacial outwash sediments.

3.2.4 Planar Cross-Bedded Granule Conglomerate

Description

This facies comprises a light-grey granule conglomerate. In contrast to the massive conglomerate facies, well defined stratification is evident. Foresets are defined by fining-upward textural variations from conglomerate to coarse-grained sandstone or carbonaceous siltstone (Fig. 3.3). The conglomerate is planar cross-bedded with foresets displaying angles of 20° with set boundaries. Foreset thicknesses vary between 1-4cm and set thicknesses between 30cm-1,5m. The conglomerate contains scattered extrabasinal pebbles, and intraclasts of coal spar and siltstone. A three-dimensional exposure of this facies is present in outcrop at the farm Goedehoop in the Witbank Coalfield, 50km east of the present study area. Cosets of planar cross-bedding up to 10m in thickness are exposed. The sets are traceable for 30-100m in a downcurrent direction. In plan view transverse to flow, the foresets are sinuous, have wave lengths of about 10m, and are traceable for distances up to 60m (the extent of the exposure).

The basal 10-40cm of the facies is grey and has a carbonaceous siltstone matrix. This facies is interbedded with the massive conglomerate and cross-bedded sandstone facies and commonly overlies carbonaceous siltstone. Stratigraphically, this facies is present below the No. 2 seam and between the No. 4 Lower and 4 Upper coal seams.

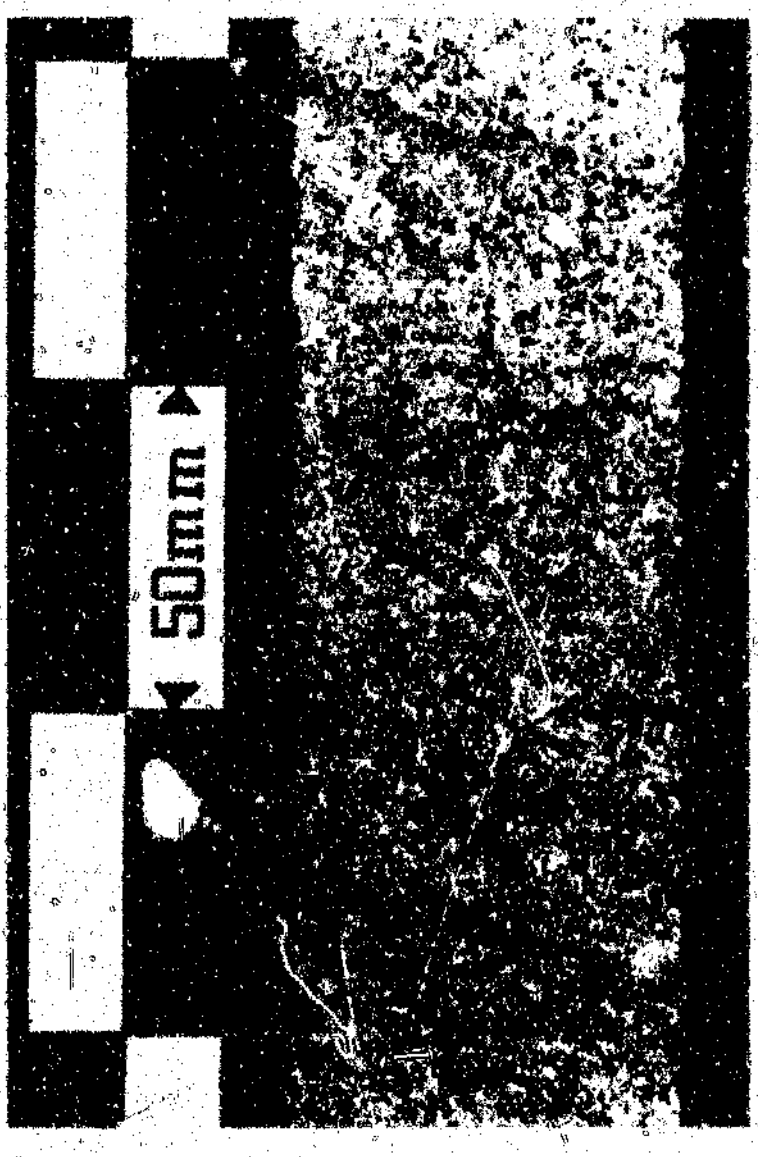


Figure 3.3. The cross-bedded granule conglomerate facies comprising well sorted quartz and feldspar grains. The facies is planar cross-bedded.

Interpretation

Unlike sand, the origin of cross-stratification in gravel has not been experimentally documented through flume experiments. However, by analogy, the lee face of bedforms are the most likely sites for foreset avalanching and grainfall. Planar cross-bedding is attributed to unit bar migration (Smith, 1972) and linguoid bar migration (Miall, 1978). Hein and Walker (1977) account for the stratification in gravels by low sediment and fluid discharge causing longitudinal bars to aggrade and build up steep downstream faces. Eventually, angle-of-repose cross-bedding results. Harms *et al.*, (1982) state that planar cross-bedding in sand originates from the downstream migration of two dimensional large ripples also termed sandwaves (Harms *et al.*, 1975). From flume experiments and observations undertaken in tidal current flow sandwaves in sand are the stable bedform at moderate velocities of 0,4-1m/sec (Harms *et al.*, 1982). The presence of coal intraclasts in this facies suggests sediment was deposited within a continental fluvial environment.

3.2.5 The Sandstone Facies

This facies comprises rocks composed of grains in the size range 0,65-2mm. The sandstone facies is important in that data concerning the types and strengths of currents, transporting and depositing sand grains, are known from flume experiments. In addition, these experiments have documented bedform migration and the generation of cross-stratification.

3.2.6 Massive Sandstone

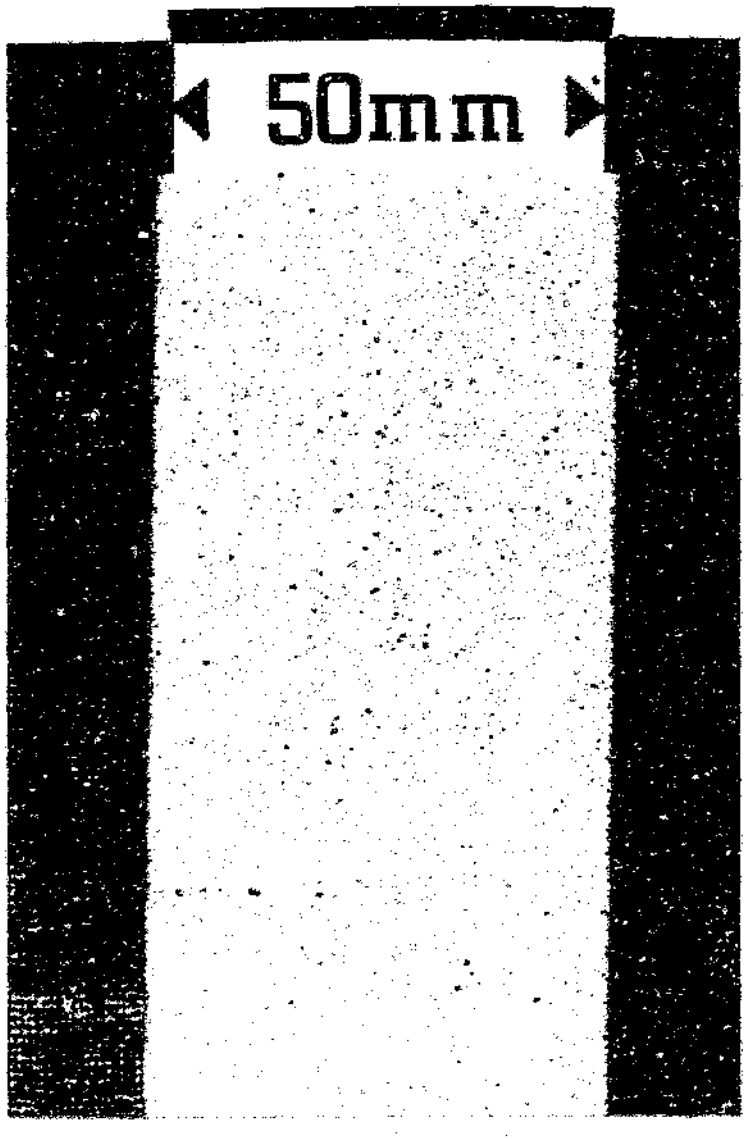
Description

This facies comprises a structureless sandstone with grains that vary from fine grained (0,25mm) to very coarse grained (2mm). The sandstone is well sorted, varying in colour between white, light grey, green and brown (Fig. 3.4A). The facies contains very little matrix and the brown variety is carbonate cemented. This facies also occurs as a poorly sorted sandstone, grey to brown in colour with a high percentage of micaceous siltstone matrix (Fig. 3.4B). The well sorted massive sandstone is frequently interbedded with cross-bedded and wave-ripple cross-laminated sandstone. The massive sandstone containing the high siltstone matrix is commonly interbedded with siltstone and alternating siltstone and sandstone beds.

Stratigraphically, the well sorted massive sandstone occurs immediately above the No. 2, No. 4 and No. 5 coal seams. It is also present as units up to 50cm thick towards the top of coarsening-upward units situated between the No. 2 and No. 3, and the No. 4 and No. 5 seams. The massive sandstone with the high percentage siltstone matrix, is frequently interbedded within thick siltstone-rich sequences. This facies association occurs above the No. 2 and No. 4 coal seams.

Interpretation

This facies is enigmatic to interpret. Hunter (1977), Blatt *et al.*, (1980), Harms *et al.*, (1982) and Collinson and Thompson (1989) invoke grainflow or fluidised flow processes to account for massive bedding. They document that secondary destructive processes such as liquifaction or bioturbation can destroy primary structure and render the sediment structureless.



A

B

Figure 3.4. The massive sandstone facies. (A) illustrates a well sorted medium-grained sandstone. (B) illustrates a massive sandstone with a high siltstone matrix.

The well sorted, carbonate-cemented sandstone is considered a traction produced sediment. This is inferred because the sandstone has very little matrix, displays a lack of textural variation and is interbedded with cross-bedded and cross-laminated sandstones. Hamblin (1962) demonstrated that in many instances well sorted sandstones, structureless to the naked eye, display stratification when X-rayed.

The massive sandstone with the high siltstone matrix, formed from the destruction of bedding through bioturbation by organisms. This view is supported by the association of this facies with bioturbated siltstone and sandstone. The facies probably represents a more intensely bioturbated sediment than the interbedded sandstone-siltstone facies (Cadle 1974; Blatt *et al.*, 1980).

3.2.7 Planar Cross-Bedded Sandstone

Description

This facies comprises fine- to very coarse-grained, arkosic sandstone. The sandstone commonly contains scattered subrounded pebbles and granules. The facies is structured by planar cross-bedding (Fig. 3.5). Set boundaries are flat, and foresets frequently grade from very coarse- to medium-grained sandstone and dip between 20° and 30°. Set heights vary between 20cm and 1,5m and cosets are up to 10m in thickness. This facies is commonly underlain by planar cross-bedded conglomerate and overlain by trough cross-bedded sandstone. Sandbodies that display this association of sedimentary facies have dimensions of up to 25m in thickness, 2-4km in width, and tens of km in length. Stratigraphically, these sandbodies are situated below the No. 2 seam and between the No. 4 Lower and No. 4 Upper seams.

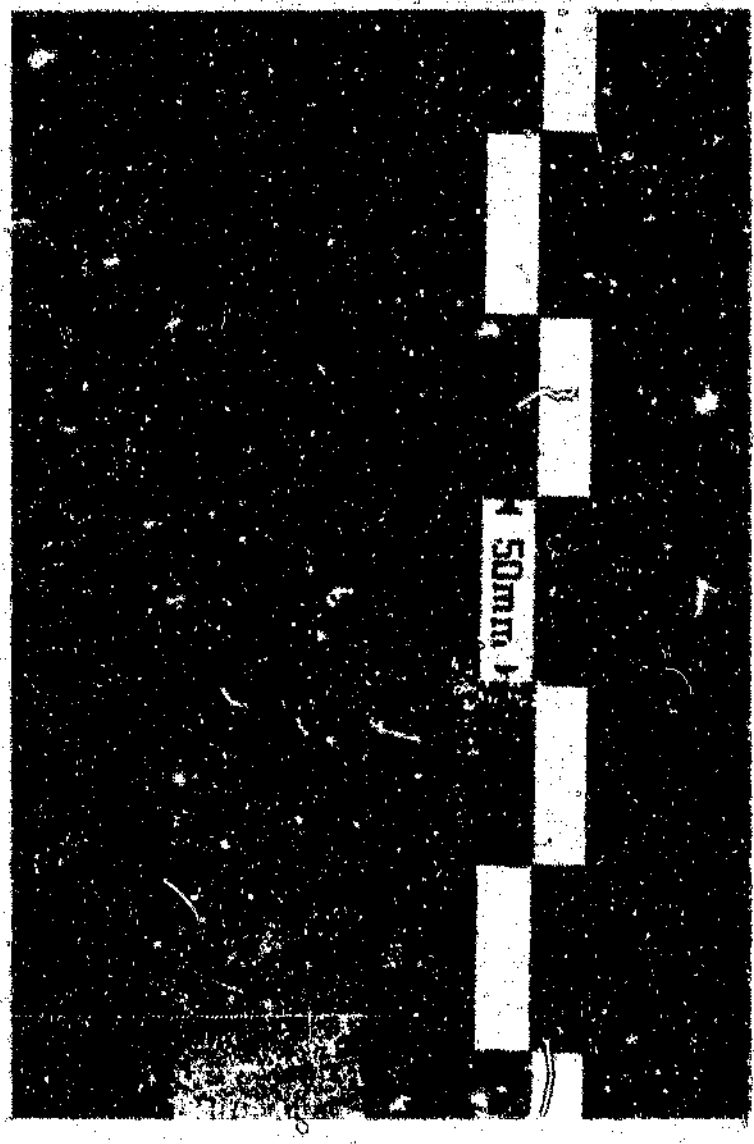


Figure 3.5. The planar cross-bedded sandstone facies. The fining-upward foresets define the cross-bedding.

Interpretation

Harms *et al.*, (1975, 1982) document that planar cross-bedding in sand-sized material forms from the downstream migration of sand waves (two-dimensional large ripples) under conditions of unidirectional flow. Based on flume experiments and observations in tidal-current flows, these bedforms form at velocities which range between 30-40cm/sec and 60cm-1,0m/sec for water depths of 10cm to 10m respectively (Harms *et al.*, 1982, Fig. 2-7). These velocities are higher than those required to transport sediment as small-scale ripples and lower than velocities required to transport sediment as dunes (three-dimensional large-scale ripples). The interpretation of a relatively low flow strength is supported by the steep and planar configuration

The association of this facies with cross-bedded conglomerate and its presence within linear sandbodies indicates that this facies formed by the migration of sand waves within braided-fluvial channels. The predominance of planar cross-bedding within braided-fluvial systems is well documented by Smith (1970), Cant and Walker (1976), Miall (1977, 1978) and Harms *et al.*, (1982).

3.2.8 Trough Cross-Bedded Sandstone

Description

This facies comprises a grey-white, fine- to medium-grained sandstone which varies in composition between an arkose and a quartz arenite. The distinguishing features of this facies are the troughed (festoon) and/or asymptotically based geometry of foresets. The erosive to undulating set boundaries and concave upward laminae which parallel the lower set boundaries is indicative of trough cross-bedding transverse to flow (Fig. 3.6).

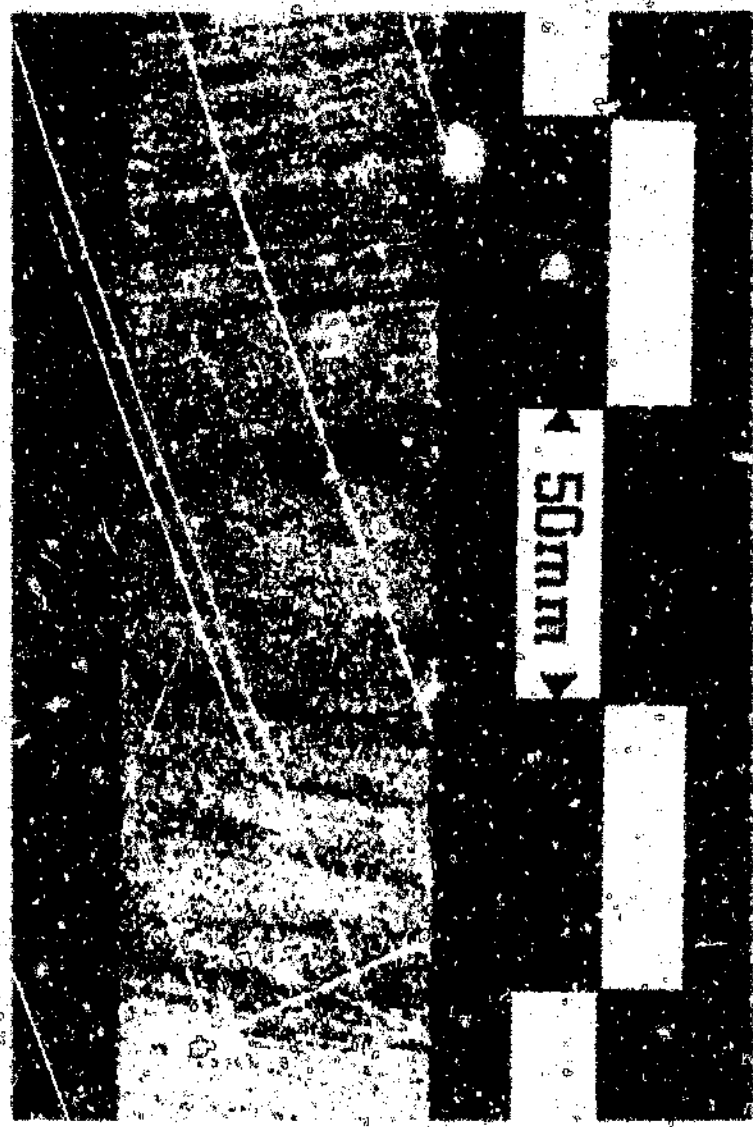


Figure 3.6. The trough cross-bedded sandstone facies defined by scoured surfaces overlain by concave-upward laminae.

Also identifiable in core are low-angled, asymptotically based foresets overlying scoured surfaces. This structure indicates trough cross-bedding parallel to flow. Set heights vary between 10-30cm and this facies either, overlies the planar cross-bedded sandstone facies or, is interbedded with the wave-ripple laminated sandstone facies. Stratigraphically, the arkosic sandstone occurs below the No. 2 and between the No. 4 Lower and No. 4 Upper coal seams. The arenite however, is stratigraphically situated within coarsening-upward sequences below the No. 3 coal seam, and below and above the No. 5 coal seam.

Interpretation

Viewed in isolation, the trough cross-bedded sandstone facies is not indicative of any particular environment. However, hydrodynamically, trough cross-bedded sandstone forms from the migration of dunes (large three-dimensional ripples) at relatively high velocities within the lower flow regime (Harms *et al.*, 1975 and 1982; Collinson and Thompson, 1989). At water depths of 10cm dunes form when velocities exceed 40cm/sec. Whereas, at water depths of 10m, dunes form when velocities exceed 80cm/sec (Harms *et al.*, 1982). The arkosic variant of this facies frequently overlies the planar cross-bedded sandstone and probably represents the migration of dunes over sandwaves within braided-fluvial channels.

The arenite variant is interbedded with wave-ripple laminated sandstone and is situated towards the top of upward-coarsening sequences below the No. 3 and No. 5 coal seams. In this setting, the facies represents dune migration within bay-fill and crevasse-splay environments.

3.2.9 Planar Laminated Sandstone

Description

The planar-laminated sandstone facies comprises light-grey, medium- to very fine-grained sandstone which displays parallel or sub-parallel laminae. The laminae vary in thickness between several millimetres and 2cm and are well defined by siltstone drapes (Fig. 3.7). The laminae often display grading and grade upwards from fine-grained sandstone to siltstone. These graded laminae vary in thickness from a few millimetres to 1cm (Fig. 3.7). Set thicknesses can vary between 5cm and 50cm and are defined by undulatory and discordant contacts. A distinguishing feature of this facies is that laminae are parallel to subparallel with underlying set boundaries. This facies is either, interbedded with wave-rippled sandstones or, overlies the planar or trough cross-bedded sandstone facies.

Stratigraphically, this facies occurs towards the top of coarsening-upward sedimentary units, below the No. 5 and No. 5 seams . It is also present overlying cross-bedded sandstones below the No. 2 and between the No. 4 Lower and No. 4 Upper coal seams.

Interpretation

The interpretation as to the mechanism of formation of this facies is variable. Planar lamination can be produced on a wide range of sediment size by a wide range of current velocities (Harms *et al.*, 1982). It can be:

a. Unidirectional flows in strong currents.

b. Unidirectional flows with low velocities.

c. Symmetrical oscillatory flows when orbital velocities are large.

d. Settling of fine-grained sediment out of suspension.

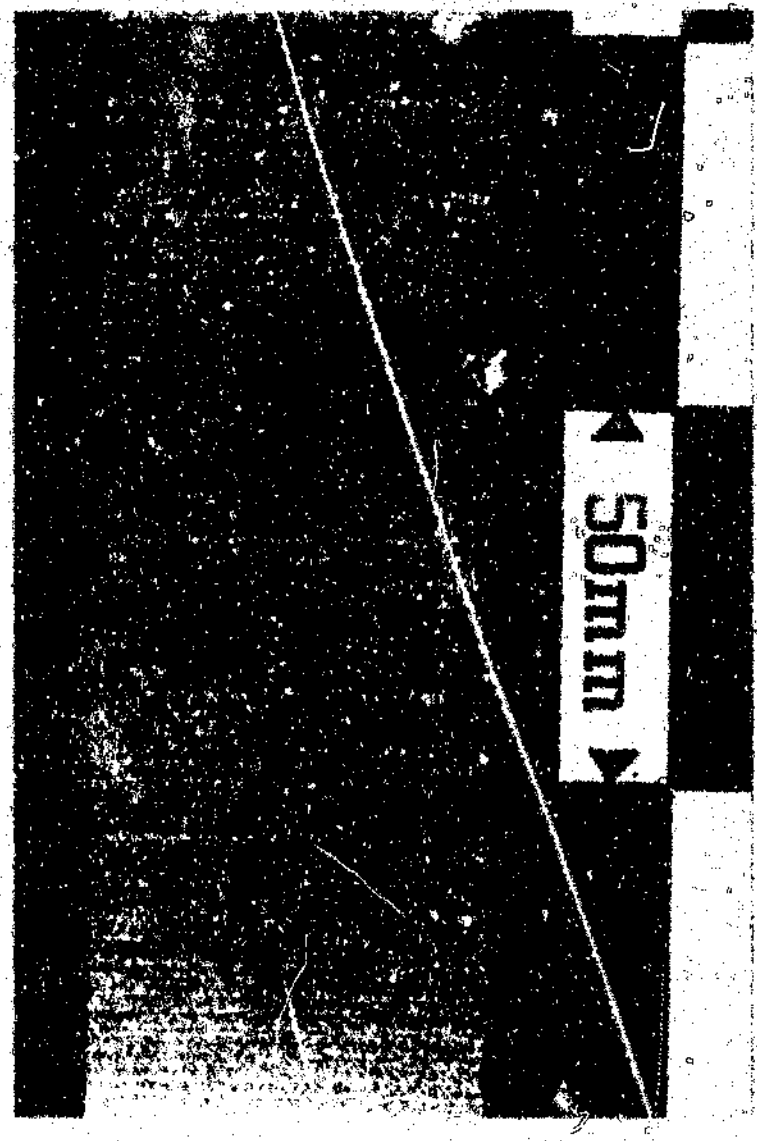


Figure 3.7. The planar-laminated sandstone facies. Note the well defined horizontal to subhorizontal laminae.

The association of this facies with wave-ripple lamination, trough and planar cross-bedding, erosive set boundaries and graded lamination gives insight into the processes which formed the facies. Of the four mechanisms of formation of planar lamination, three of these mechanisms probably account for the formation of this facies. These are; unidirectional flows in strong currents, symmetrical oscillatory flows with large orbital velocities, and the settling of fine-grained sediment out of suspension. Where the facies displays erosive set boundaries, graded laminations, and is interbedded with wave-rippled sandstones, the sediment was probably deposited during high velocities in symmetrical oscillatory flows. During reduction in velocity sediment settled out of suspension and formed graded laminations. Erosive set boundaries suggest fluctuating periods of high discharge during which erosion of the underlying sediment took place. The association of this facies towards the top of coarsening-upward sequences below the No. 3 and No. 5 seams implies this facies was deposited within a delta-front environment subject to wave action.

Where the facies overlies trough and planar cross-bedded sandstone no graded lamination is evident. This suggests deposition of sediment at high velocities in steady unidirectional flows. Deposition of this facies within braided-fluvial channels is implied from the presence of this facies towards the top of fining-upward sandstone sequences, situated within linear sandbodies over- and underlain by coal seams.

3.2.10 Cross-Laminated Sandstone

Description

The cross-laminated sandstone facies comprises very fine-grained and fine-grained micaceous sandstone. The facies is readily distinguishable in core as the sandstone is structured by small-scale cross-stratification and well defined silt-draped laminae. The set size of laminae ranges between 0,5cm and 5cm, and sets which are less than 1cm thick are estimated to have a cross-sectional dimensions of 5-10cm.

Two types of cross-lamination are distinguishable in core. The first type of cross-lamination, displays laminae which are both concave and convex upward, and set boundaries which are frequently erosive or discordant (Fig. 3.8A). Infrequently, sets display chevron-like interleaving laminae. The second type of cross-lamination, displays cross-laminae which are tabular parallel to flow (Fig. 3.8B). Laminae are also trough like transverse to flow, and overlie a concave-upward basal surface.

Stratigraphically, the first type of cross-lamination is present in the sandstone intervals between the No. 2 and No. 3 seams, the No. 3 and No. 5 seams and the No. 5 and No. 6 seams. The second type of cross-lamination is present in cosets 1-20cm thick, overlying planar cross-bedded sandstones, towards the top of fining-upward sandstones. These sandstones are situated below the No. 2 seam and between the No. 4 Lower and No. 4 Upper seams.

Interpretation

The stratification described forms from the migration of ripples under conditions of oscillatory flow, combined flow and unidirectional currents (Harms *et al.*, 1975; Harms *et al.*, 1982). The ripple type displaying convex and concave foresets, erosional

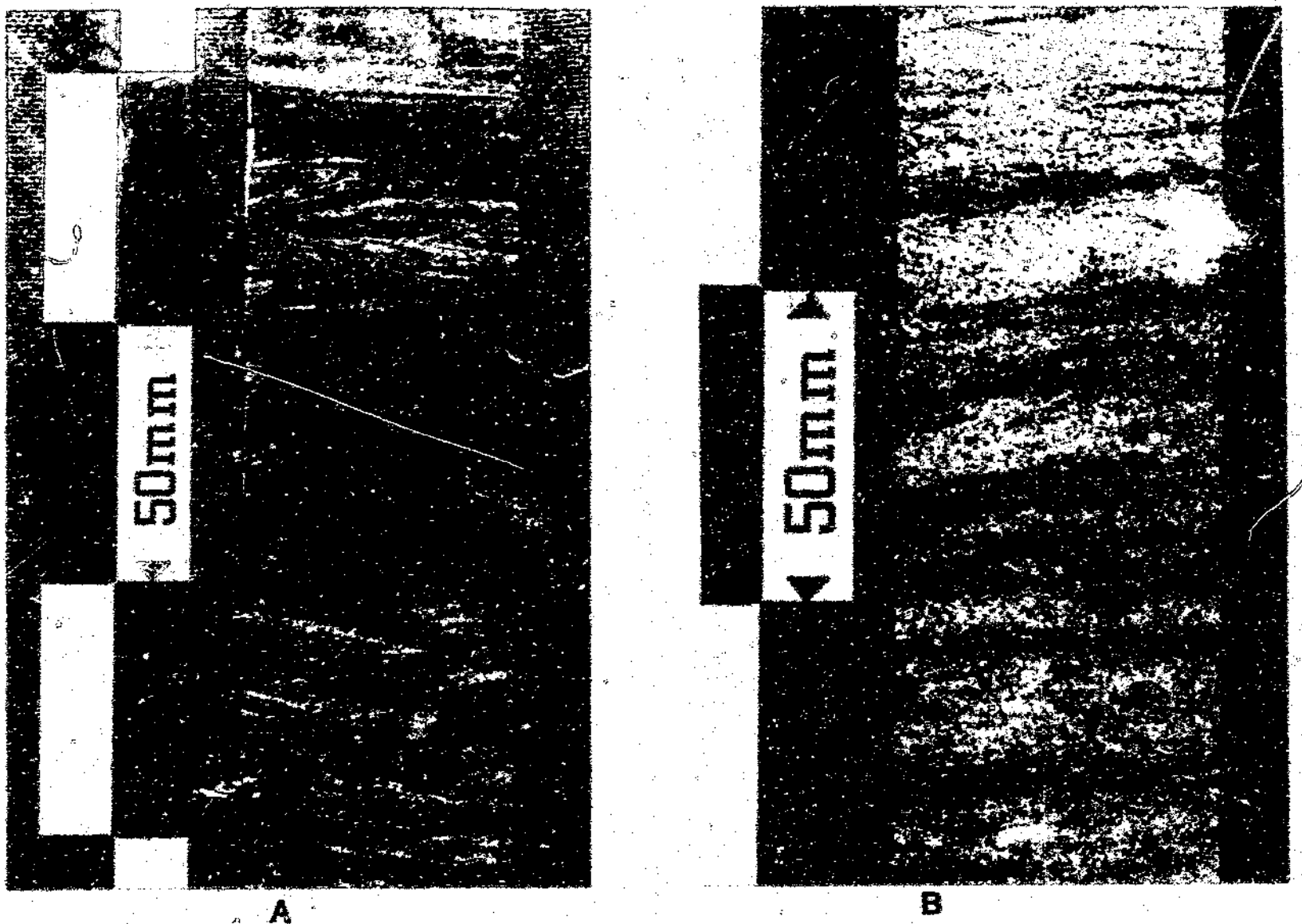


Figure 3.8. The cross-laminated sandstone facies. (A) Wave-ripple laminated sandstone displaying both convex and concave foresets. (B) Cross-lamination produced by current ripples.

and discordant set boundaries, and chevron like interleaved laminae, are formed from the migration of wave ripples (Harms *et al.*, 1975) or vortex ripples (Harms *et al.*, 1982; Collinson and Thompson, 1989). Where laminae are sigmoidal and convex upward interpretation becomes ambiguous. This type of lamination can form from either, migration and aggradation of symmetrical oscillation ripples or, the migration and aggradation of unidirectional current ripples with substantial fallout of sediment from suspension. The laminations which are concave parallel to erosive set contacts, form from the migration of current ripples with small angles of climb. This stratification type is termed small-scale trough cross-lamination.

The orbital speed and oscillation period for oscillation ripples are difficult to quantify due to limited experimental data. However, for grain sizes 0,15-0,21mm, Harms *et al.*, (1982, Fig. 2-14) illustrate that these ripples form at orbital speeds of 0,275-0,475m/sec and oscillation periods of 2-10sec. The interpretation of current ripples suggest current velocities are low (20-60cm/sec) and at least sufficient for traction transport. Large angles of climb indicate suspension sedimentation and provide little evidence about flow velocities.

Wave-ripple cross-lamination is present in coarsening-upward sequences between the No. 2 and No. 3 coal seams and within the No. 5 and No. 6 seam depositional sequences. At these stratigraphic positions, wave-ripple cross-lamination overlies the carbonaceous siltstone and alternating sandstone and siltstone facies. This facies association probably represents bay-fill and crevasse-splay sedimentation (Elliot, 1974). The wave-ripple cross-laminated sandstone represents delta-front sandstones reworked by low-energy wave processes. Similar sedimentary sequences have been documented for the Mississippi Delta by Coleman and Gagliano (1965).

The presence of small-scale oscillation-ripple lamination as the dominant cross-lamination type in the Vryheid Formation, suggests that wave processes were an important factor in reworking fine-grained sand. Small-scale trough cross-laminated sandstones, represent the migration of current ripples over dunes. These ripples form within unidirectional currents due to the reduction in any of the following variables; water depth, velocity and grain size (Harms *et al.*, 1982). Sandstones structured by small-scale trough cross-lamination are situated towards the top of upward-fining sandstones within the No. 2 and No. 4 depositional sequences. These cross-laminated sandstones document sedimentation within fluvial channels.

3.2.11 Interlaminated Sandstone-Siltstone Facies

Description

The interlaminated sandstone-siltstone facies is easily defined lithologically as it comprises alternating beds or laminae of sandstone and siltstone. The facies is intermediate between the lithological continuum of sand-dominated and silt-dominated facies. Within the interlaminated sandstone-siltstone facies three subfacies are distinguished:

a. Flaser-laminated sandstone,

b. Wavy-laminated sandstone-siltstone and,

c. Lenticular-laminated siltstone.

Stratigraphically, the boundaries between these three subfacies are difficult to define as these subfacies grade into one another. The flaser-laminated sandstone facies comprises fine- to medium-grained sandstone displaying thin siltstone laminae which drape troughs in foresets (Fig. 3.9). The siltstone laminae accentuate bedding. Reineck and

Figure 3.9. The flaser-laminated sandstone facies. Laminations are defined by siltstone drapes.

Wunderlich (1968) identified simple, bifurcated and wavy-flaser lamination. In this study, the term flaser includes all subdivisions of flaser lamination. A distinguishing feature of flaser lamination is that siltstone drapes are discontinuous over several centimetres. This facies is often interbedded with the wave-rippled sandstone. Wavy lamination comprises interlaminated sandstone and siltstone where laminae are laterally continuous for distances greater than several centimetres (Fig. 3.10). Dark- grey siltstone laminae 2mm-3cm in thickness drape over rippled sandstone surfaces. The sandstone laminae vary in thickness between 0,5mm-5cm. Lithologically, this facies is intermediate in terms of the siltstone and sandstone component when compared to the flaser-laminated sandstone and lenticular-laminated siltstone facies. Thus, the relative proportions of sandstone and siltstone laminae grade between sandstone-dominated and siltstone-dominated end members.

Lenticular-laminated siltstone facies comprises a dark-grey carbonaceous siltstone enclosing connected and isolated cross-laminated sandstone lenses (Fig. 3.11). Contacts between the siltstone and sandstone are sharp indicating alternating deposition of silt and sand. This facies frequently overlies the carbonaceous siltstone facies and underlies the wavy-laminated sandstone-siltstone facies. Intervals within the interbedded sandstone-siltstone facies display post-depositional structures such as; convolute bedding, load casts, and flame structures.

Stratigraphically, the three subfacies comprise units 2-3m thick above the No. 2, No. 3, No. 4, No. 5, and No. 6 coal seams. These subfacies are also present towards the base of coarsening-upward units and reflect an increasing sandstone component upwards. Furthermore, the lenticular-laminated siltstone and wavy-laminated sandstone-siltstone facies commonly occur as beds less than 5cm thick directly below coal seams.

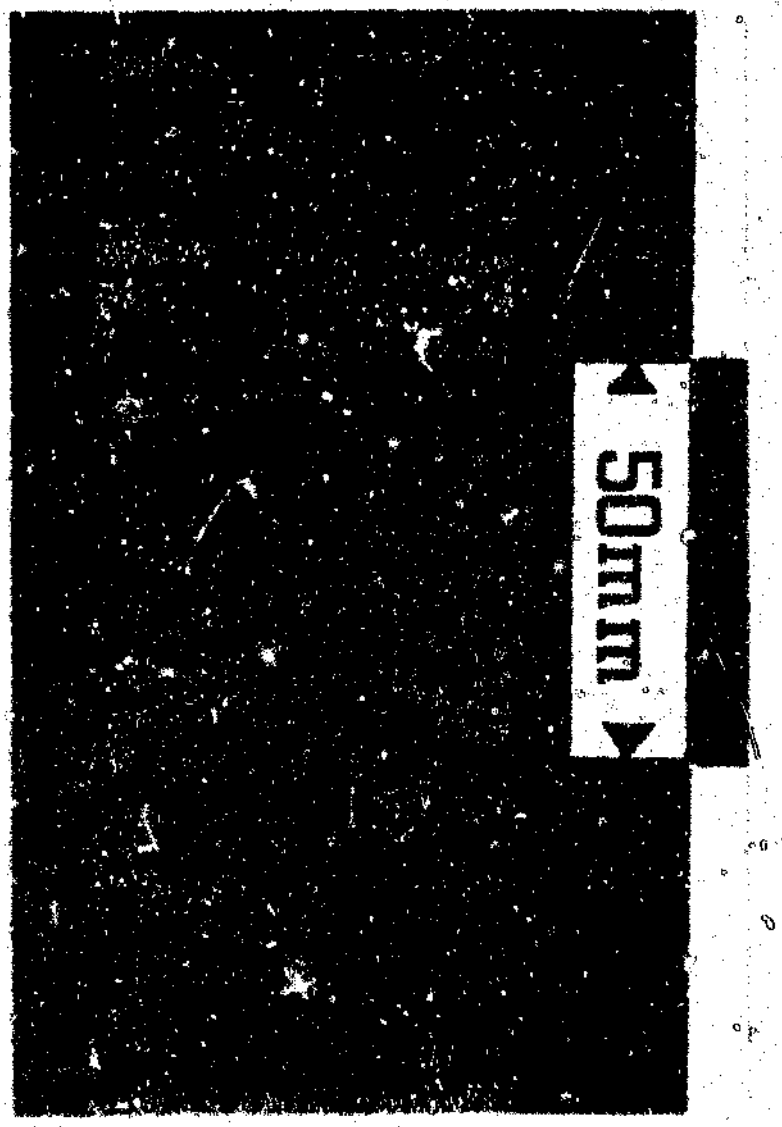


Figure 3.10. Interlaminated sandstone and siltstone facies displaying wavy lamination.

Figure 3.11. The lenticular-laminated siltstone facies. Lenses of sandstone within siltstone define the facies.

Interpretation

The repetition of sandstone and siltstone laminae suggest alternating periods of bedload transport followed by periods of suspension settling. The variation in thickness of laminae is controlled by the abundance of sediment in the system and the period over which current or suspension processes operated. Thus, currents are dominant during the deposition of flaser-laminated sandstone, both current and suspension processes are important during the deposition of wavy-laminated sediment and suspension processes are dominant during the deposition of lenticular-laminated siltstone.

These facies are recorded from a variety of environments such as subtidal and intertidal environments (Van Straaten, 1954; Reineck, 1963; Reineck and Wunderlich, 1968), delta-front and bay-fill environments (Coleman, 1976), and fluvial flood plains (Leeder, 1982). The association of this facies at the base of coarsening-upward sequences between the No. 2 and No. 3 seams and between the No. 5 and No. 6 seam depositional sequences suggests these facies were deposited within prodelta and delta-front environments. The association of this facies below and above coal seams implies deposition in overbank settings.

Deformation of the sediment to produce convolute bedding, load casts and flame structures suggest post-depositional processes. Processes such as liquifaction and compaction of lithologies with different densities and porosities account for these structures (Collinson and Thompson, 1989).

3.2.12 Bioturbated Sandstone and Siltstone Facies

Description

Bioturbation occurs preferentially within two lithologies. These are; fine- to medium-grained sandstone, and siltstone. The fine- to medium-grained sandstone is either, micaceous or, quartz arenitic in composition (Fig. 3.12). Where bioturbation has not obliterated sedimentary structures the sandstone is commonly wave-ripple laminated. The sandstone is however, frequently extensively bioturbated and massive in appearance. In these instances, the sandstone is micaceous and takes on the appearance of a silty sandstone.

The bioturbated siltstone facies comprises dominantly siltstone with minor amounts of interlaminated fine- to medium-grained sandstone. The intensity of bioturbation in this facies varies from minor, to complete, reworking of primary sedimentary structures (Fig. 3.13). Where present, the preserved structures display lenticular and planar lamination.

Bioturbation is preserved within both the sandstone and siltstone lithologies primarily as vertical burrows (Figs. 3.12, 3.13). Less commonly horizontal burrows and trails are preserved. Within borehole core, *Skolithos*, *Siphonichnus* and *Planolites* burrows are identified where bioturbation is not intense. Where the sediment is intensely bioturbated and takes on a massive appearance trace fossils cannot be identified.

Stratigraphically, the bioturbated sandstone facies occurs towards the top of coarsening-upward sequences between the No. 2 and No. 3 coal seams and within the No. 1, No. 5, and No. 6 seam depositional sequences. It is also present within thin, laterally extensive, sheet sandstones above the No. 2, No. 4 , No. 5 and No. 6 coal

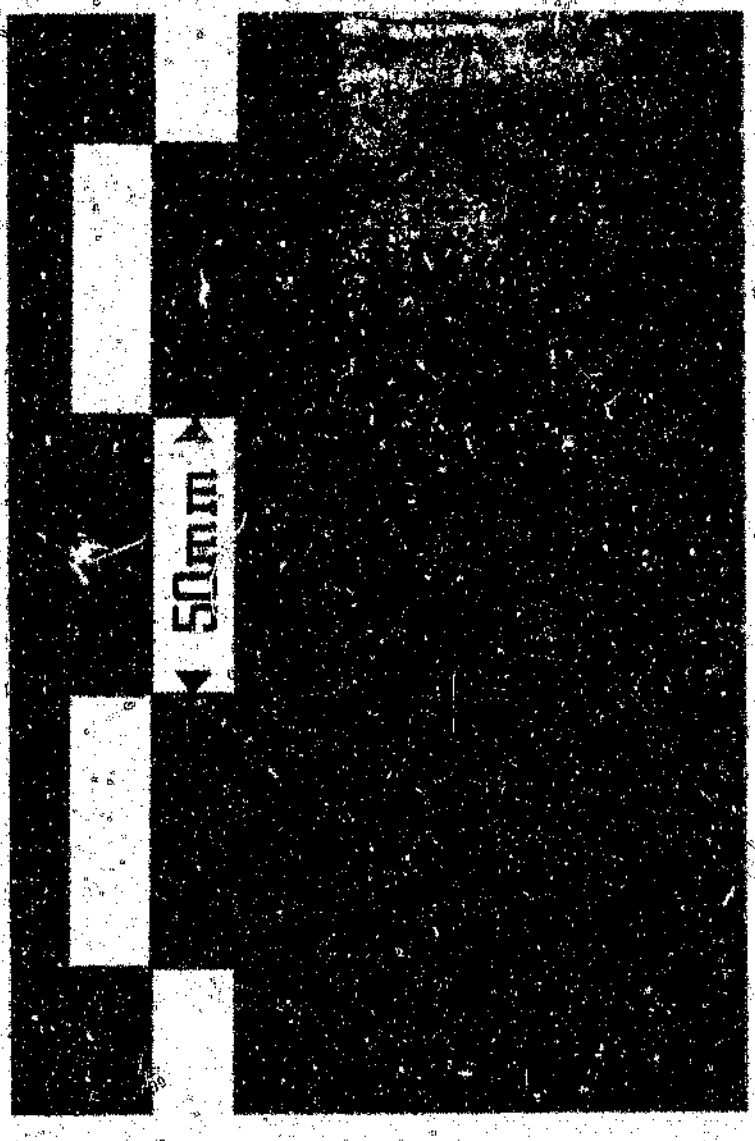


Figure 3.12. The bioturbated sandstone facies. Both *Siphonichnus* and *Skolithos* traces are present.

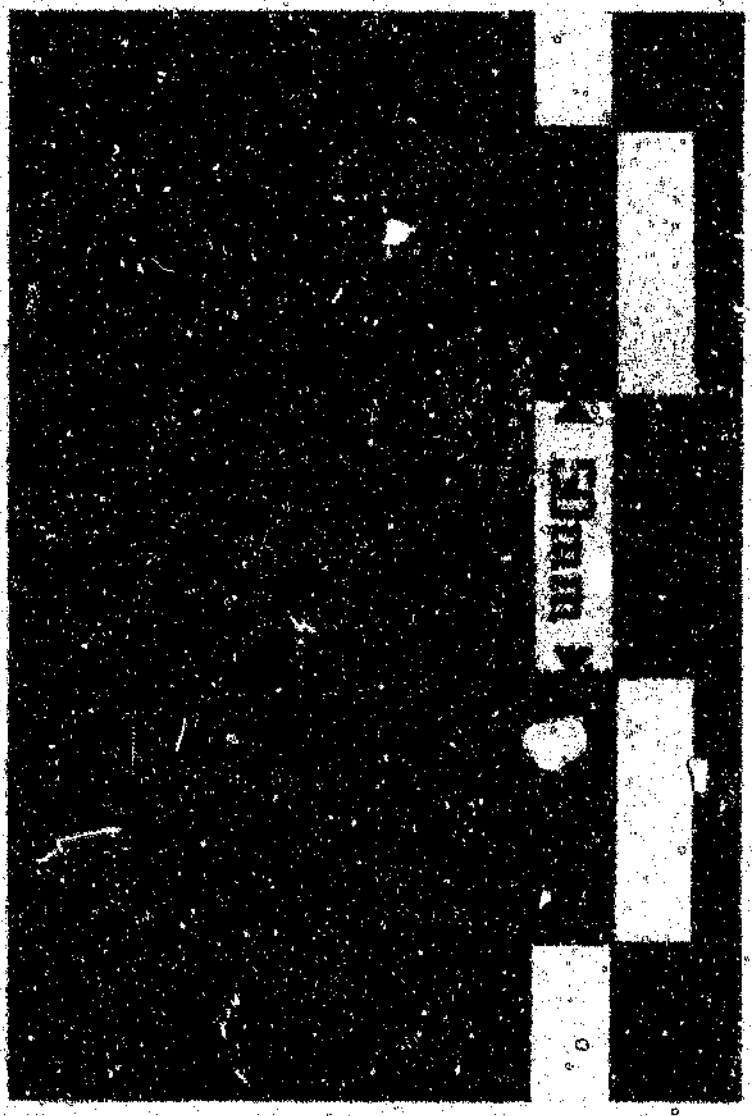


Figure 3.13. The bioturbated siltstone facies. The intense bioturbation is probably due to the high organic content within the siltstone. Note the well preserved *Siphonichnus* traces.

seams. The bioturbated siltstone facies is present towards the base of coarsening-upward sequences, between the No. 2 and No. 3 seams and within the No. 1, No. 5, and No. 6 seam depositional sequences. Siltstone beds which over- and underlie the No. 2, No. 3, No. 4, No. 5, and No. 6 seams, are also bioturbated.

Interpretation

The recognition of trace fossils within the Vryheid Formation is important as the traces represent a record of life and ev[illegible] that took place during, or soon after, deposition. The [illegible] that no body fossils are associated with these traces is probably due to the high organic content of shales and coal seams within the stratigraphic sequence. During diagenesis, humic acids are liberated from organic material and promote the dissolution of body fossils (Fisher *et al.*, 1969). Trace fossils and trace fossil assemblages are used to interpret bathymetry and nutrient supply (Seilacher, 1964; 1967; 1978). However, it is now accepted that variables such as substrate consistency, oxygen levels, salinity, temperature, sedimentation rates and ocean currents influence faunal distribution and behaviour (Howard, 1978).

Siphonichnus burrows which are considered by Stanistreet *et al.*, (1980) to possibly be a member of the *Skolithos* ichnofacies, together with *Skolithos* burrows are attributed to suspension feeders. These traces inhabit high-energy, shallow-water marine environments a setting which favours unbranched vertical burrows (Crimes, 1975; Chamberlain, 1978; Seilacher, 1978). The association of *Skolithos* and *Siphonichnus* burrows with wave-rippled sandstone suggests a mildly turbulent environment which allows food to be provided in suspension. *Planolites* trails are preserved in siltstone laminae and beds. This implies that the organic material is concentrated in greater abundances in the siltstone than in the sandstone. The presence of this ichnofacies in the silty sediments implies the trails were made by deposit feeders who lived in low-energy settings.

The bioturbated siltstone and sandstone facies which form coarsening-upward sequences, with cross-laminated and cross-bedded sandstones, formed in prodelta and delta-front environments. These facies associations are present between the No. 2 and No. 3 seam interval and the No. 5 and No. 6 seam depositional sequences. Similar bioturbated deposits are documented by Coleman and Gagliano (1965), and Basan *et al.*, (1979) from interdistributary-bay environments in deltas. Stanistreet *et al.*, (1980), Turner *et al.*, (1981), Mason *et al.*, (1983), Roberts (1986) and Christie (1988) describe bioturbation and similar trace fossils from delta-front and crevasse-splay environments in the Vryheid Formation. The association of trace fossils with carbonaceous siltstones adjacent to coal seams implies that these sediments accumulated in low-energy, nutrient-rich, backswamp environments.

3.2.13 Gravelly Siltstone Facies

Description

This facies is distinctive in appearance and comprises either, massive siltstone or, interlaminated carbonaceous siltstone and claystone which displays infrequent bioturbation. Conglomerate and sandstone interbeds range in thickness from 2 to 15mm and are randomly distributed within the facies. The facies is also distinctive in that granules and sandstone grains are scattered throughout the siltstone and claystone (Fig. 3.14).

The facies displays a variety of structures. The siltstone is often massive and bioturbated, and is commonly interbedded with interlaminated siltstone-claystone displaying planar lamination and graded bedding. The interbeds of granule conglomerate, sandstone, and siltstone display lenticular lamination.

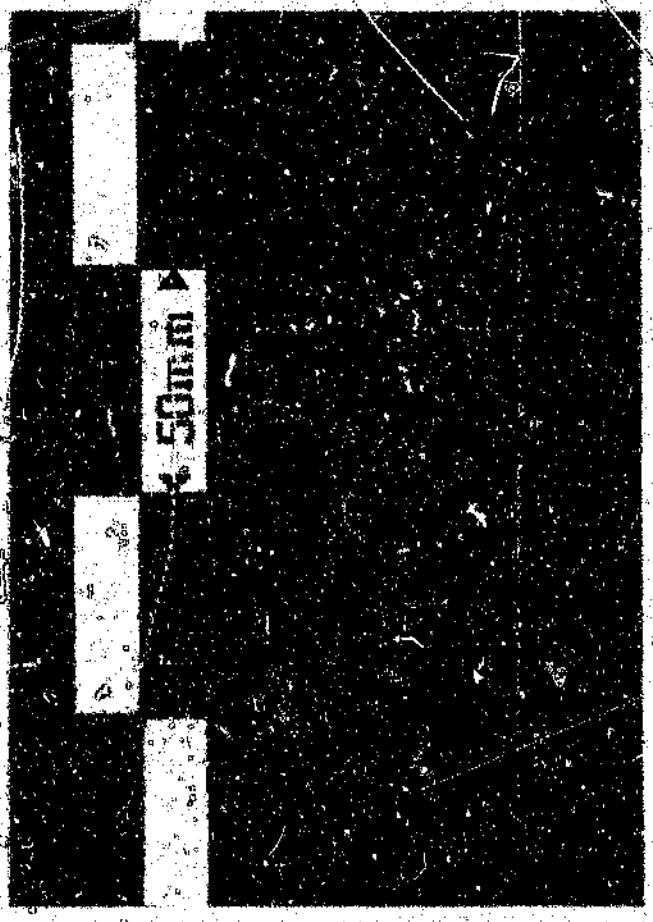


Figure 3.14. The gravelly siltstone facies. Note the dispersion of gravel grains in a siltstone matrix.

Stratigraphically this facies is present under- or overlying the diamictite facies, interbedded with the carbonaceous siltstone and bioturbated siltstone facies, and overlain by the cross-laminated sandstone facies. The facies is present in topographically low-lying areas at the base of the No. 1 and No. 2 seam depositional sequences.

Interpretation

The association of this facies with diamictite at the base of the Vryheid Formation suggests that deposition was associated with glacial processes. The laminated siltstone-claystone displaying graded bedding suggests deposition by subaqueous density underflow currents or turbidity currents. The association of this lithology with diamictite suggests that the interbedded siltstone-claystone and conglomerate-sandstone couplets, represent turbidity current deposition in proglacial lakes during periods of high meltwater discharge. High meltwater discharge takes place in summer. The claystone would deposit from suspension in winter. The conglomerate and sandstone would be derived from delta topsets and carried into the prodelta and lacustrine environments by density underflow currents (Cohen, 1983).

The scattered granules and sand grains present within the fine-grained sediments could have more than one possible origin. The first explanation is that the granules and grains could display their random distribution through the bioturbation of gravel and sand beds which is supported by the recognition of burrows associated with these deposits. The second explanation is that the granules and grains could represent coarse gravelly and sandy material which is ice rafted into a proglacial lacustrine or shallow-marine environment and subsequently deposited by suspension sedimentation (Eyles, *et al.*, 1985). Scattered granules, associated with the siltstone portion of a graded-bedded siltstone-claystone couplet, implies deposition of the siltstone and granules out of suspension by turbidity currents. A similar facies termed the pebbly mudstone facies is

documented by Le Blanc Smith and Eriksson (1979) at the same stratigraphic position in the Witbank Coalfield. These deposits are interpreted as turbidity current deposits in a proglacial setting.

3.2.14 Carbonaceous Siltstone Facies

Description

This facies is distinctive in core as it comprises a dark grey, carbonaceous and micaceous siltstone. It is either, massive in appearance or, displays planar lamination. (Fig. 3.15). The presence of pyrite crystals and *Glossopteris* plant leaves within the siltstone is distinctive of this facies.

The carbonaceous siltstone facies occurs as laterally continuous beds over areas in excess of hundreds of km^2. Stratigraphically, this facies is present above the No. 2, No. 4, No. 5 and No. 6 coal seams. At these stratigraphic positions the facies varies in thickness between 5cm and 4m. The facies also commonly occurs as units 10cm-1,5m thick below and above all coal seams. It is at these stratigraphic positions that *Glossopteris* plant leaves are encountered.

Interpretation

The carbonaceous siltstone facies structured by planar lamination and containing pyrite crystals suggests suspension deposition in a low-energy reducing environment. Carbonaceous muds have been documented from deep sea environments (Selley, 1976). However, the association of coal seams and *Glossopteris* plant leaves with this facies

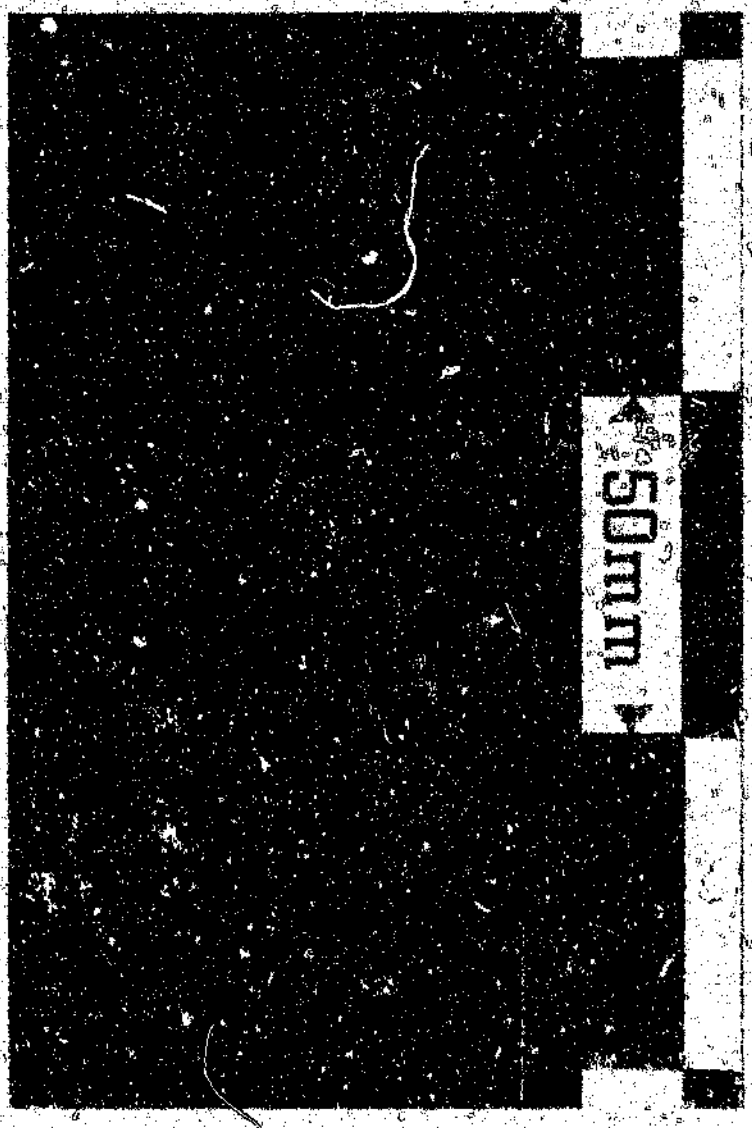


Figure 3.15. The carbonaceous siltstone facies.

precludes a deep-marine environment and suggests a shallow-water backswamp depositional environment. Coleman *et al.*, (1964) and Coleman and Gagliano (1965) have documented organic-rich black muds with silt laminations and plant remains from backswamp and prodelta environments of the Mississippi Delta.

The laterally extensive development of this facies at the base of coarsening-upward sequences above the No. 2, No. 4, No. 5 and No. 6 seams implies low-energy deposition following transgression. At these stratigraphic positions, this facies records marine-flooding surfaces and documents the commencement of progradational depositional sequences or parasequences (Van Wagoner *et al.*, 1988).

3.2.15 Carbonate Facies

Description

The carbonate facies is volumetrically an insignificant facies type. The carbonate occurs as laterally impersistent beds between several centimetres and 40cm thick at the stratigraphic position of the No. 5 and No. 6 seams. In addition, carbonate is present above the No. 5 seam at the position of transgression between the No. 5 and No. 6 depositional sequences. It is in the southeastern portion of the study area where the carbonate is present at these stratigraphic positions.

The carbonate which is identified as calcite through X-ray diffraction analysis is white in colour and is friable to coarsely crystalline in appearance (Fig. 3.16). The most common facies association is the interbedding of the carbonate with the carbonaceous siltstone facies. Carbonate also occurs as concretions, lenses and cleat fills within the coal seams.

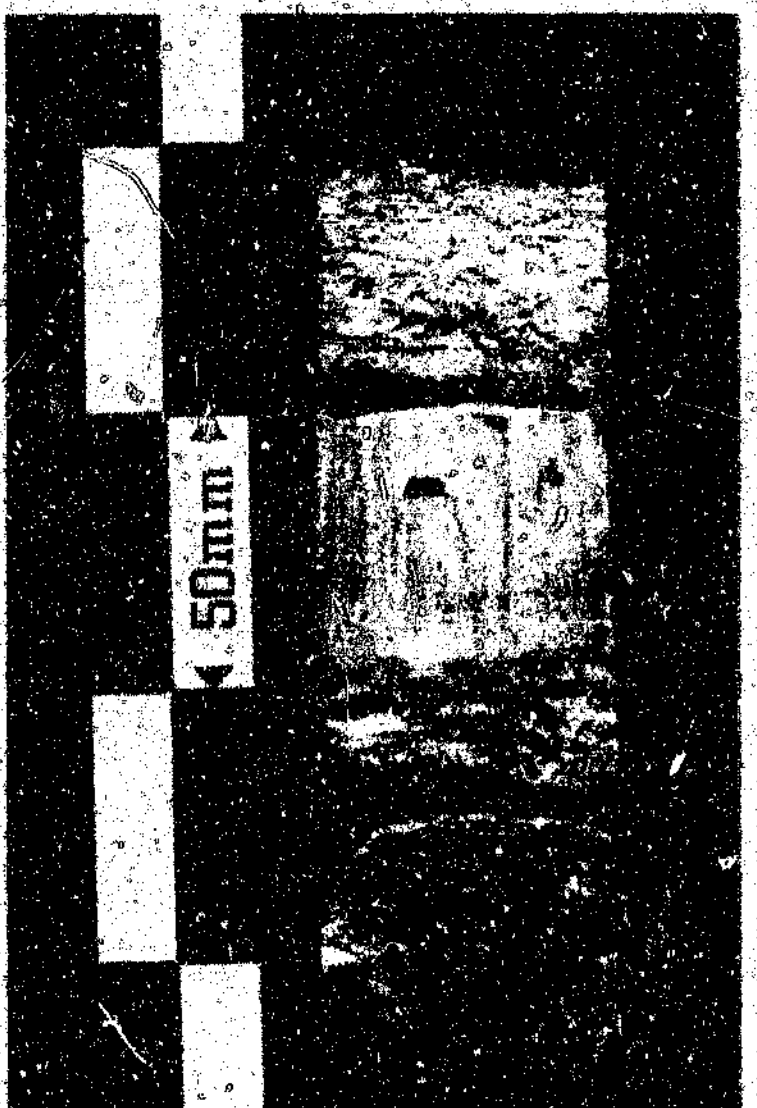


Figure 3.16. The limestone facies. X-ray diffraction analysis confirmed calcite as the carbonate mineral constituting this facies. Note the development of large calcite crystals in the middle of the core.

Interpretation

The occurrence of the carbonate at the stratigraphic position of coal seams and immediately above coal seams, suggests the carbonate precipitated out in either, shallow freshwater or, saline lacustrine environments. In lacustrine systems the sediments comprise, carbonate, organic matter, biogenic silica and detrital siliclastics. An increase in the relative abundance in any one of the components will result in a decrease in the remaining components. The major sources of carbonate in lakes are: (a) inorganically precipitated carbonate (b) photosynthesis induced, inorganically precipitated carbonate (bio-induced carbonate) (c) biogenic carbonate and (d) detrital material derived from carbonate in the drainage basin. Of these sources, the most likely mechanism for the carbonate precipitation is inorganic and bio-induced carbonate, as the likelihood of detrital and biogenic carbonate contributions would be insignificant.

Inorganic and bio-induced precipitation of $CaCO_3$ is enhanced by the removal of CO_2 through phytoplankton photosynthesis and fixation of CO_2 by bacteria and algae. The removal of CO_2 causes the increase in the pH of lacustrine waters and results in the supersaturation and precipitation of $CaCO_3$ (Megard, 1968; Wetzel, 1975).

The presence of carbonate at the stratigraphic positions of the No. 5 and No. 6 coal seams, suggests that these areas existed as shallow lakes prior to marine transgression which terminated peat accumulation. Here, through photosynthesis and algal and bacterial activity, $CaCO_3$ precipitated. The occurrence of torbanites at these stratigraphic positions documents algal activity during the accumulation of these peats (Cadle *et al.*, 1993). These data lend support to the hypothesis that the carbonates formed as inorganic and bio-induced carbonates. The fact that the Karoo peats are now medium-volatile coals implies the sediments have undergone diagenesis and the calcite has recrystallized to form large calcite crystals up to 3cm in length (Fig. 3.16).

3.2.16 Coal Facies

Description

The term "coal" is used in this study in the same way as defined by Schopf (1956). He produced a well thought out definition of coal:

"Coal is a readily combustible rock containing more than 50% by weight and 70% by volume carbonaceous material, formed from the compaction and induration of variously altered plant remains similar to those of peaty deposits. Differences in the kinds of plant material (type), in the degree of metamorphism (rank) and range of impurity (grade), are the characteristics of the varieties of coal."

In the study area, a number of coal seams are present within the Vryheid Formation and are termed from the base upwards, the Nos. 1, 2, 3, 4, 5 and 6 seams. Of these seams the No. 2 and No. 4 seams are usually exploitable due to their thickness and lateral continuity. The No. 2 and No. 4 seams commonly have thicknesses between 2-4m and are laterally extensive over 100's km². The No. 5 seam attains thicknesses of 1-2m for areas of 10-100's km², and the Nos. 1, 3, and 6 seams thicknesses of < 1m over areas that vary between several 10's m² and 10's km².

The coals of the Transvaal Coalfield (Fig. 1.3) are classified as inertinite rich, low rank, and high ash (Snyman, 1961 and 1976; Plumstead, 1966; and Falcon, 1978, 1986, and 1989). Macroscopically, the coal comprises alternating dull, lustrous and bright bands (Fig. 3.17). The bright bands are generally less than 10cm in thickness and are separated by dull and lustrous bands > 10cm in thickness. The No. 2 and No. 4 coal seams have a higher concentration of bright (vitrinite-rich) bands at the base of

Figure 3.17. The coal facies. Bright bands represent vitrinite-rich macerals and dull bands inertinite-rich macerals (No. 5 seam).

the seams and a higher concentration of dull (inertinite-rich) bands towards the top of the seams. The Nos. 1, 3, 5, and 6 seams are generally brighter coals than the No. 2 and No. 4 seams. These coals usually display a predominance of bright and lustrous bands over dull bands.

The most detailed petrographic analysis of a coal seam from the Vryheid Formation was published by Falcon (1989), who presented the petrography of the No. 2 seam in the Witbank Coalfield. The seam is classified as a low-rank bituminous coal (0.6-0.8% R_{ran}). Furthermore, Falcon (1989) subdivided the seam vertically into three zones. Zone A, at the base of the seam, is characterised by microlithotypes vitrite and intermediate alternating with inertodetrinite. Zone B, the middle zone, has the highest vitrite and intermediate content of all zones and is low in mineral matter. The uppermost zone, Zone C, has a high inertodetrinite content and a low vitrite and intermediate content. Each zone is separated by a carbominerite band.

Interpretation

Coal is formed from the compaction and induration of many different types of plants and parts of plants. When the accumulation of vegetable matter is balanced by subsidence, conditions are favourable for the accumulation of peat deposits. The water-table level is the factor that immediately controls mire formation and peat accumulation (Gore, 1983) and is assumed to control the degree of preservation of organic matter in a mire (Diessel, 1970; Stach *et al.*, 1982). With rapid descent of organic matter into reducing mire water, processes of microbiological oxidative decomposition are greatly reduced. With increasing depth of burial, these processes are totally arrested (Moore and Bellamy, 1973). With increasing temperature caused through burial, organic material is transformed into humic material termed "humines"

at depths of 20-400m. With further burial and increase in temperature, the organic material undergoes geochemical gelification. The coal, previously brown, becomes black and shows banding. At this stage coal is termed bituminous.

The Vryheid Formation coals are considered autochthonous in origin (Melkr, 1906; Wybergh 1922; Plumstead, 1957 and 1966; Falcon, 1986 and 1989). This conclusion is based upon the thickness and lateral continuity of seams, and maceral bands within seams. The presence of rootlet remains at the base of the seams provide further evidence for the autochthonous nature of the coals (Fig. 3.18). Based upon the identification of pollen and spore genera and species, Falcon (1986; Fig. 5) proposed that the Nos. 1 and 2 seams accumulated in a cold temperate climate. Thereafter, the climate ameliorated such that the No. 5 and No. 6 seams accumulated in a cool- to warm-temperate climate.

Based upon the petrographic study undertaken by Falcon (1989) on the No. 2 seam, the three in-seam zones are attributed to have formed in the following way:
Zone A (situated at the base of the seam), underwent two periods of relatively quiet, waterlogged reducing swamp conditions, interspersed with periods of organic detrital influx. Organic material that formed Zone B, (situated at the middle of the seam), accumulated in stable, quiet, water-logged conditions. During this period of peat accumulation no significant mineral matter was deposited and is reflected today by low-ash coal, comprising high-reactive macerals. Zone C (the upper zone), is characterised by abundant detrital organic matter and a brief return to quiet, water-logged conditions.

The origin of high inertinite content of the Vryheid Formation coal seams remains debateable. Falcon (1986; 1989) suggests that the flora and climate are the dominant factors in the formation of these inertinite-rich coals. Snyman (1961) proposed a

Figure 3.18. Rootlet remains beneath the No. 4 Lower seam.

"hypautochthonous" origin for the coal in which the organic matter formed as the result of limited transportation and redeposition. Cadle *et al.*, (1993) has postulated that tectonic setting is an important factor controlling coal type. The Ecca coals accumulated on the cratonic side of a foreland basin where minor subsidence of the peat swamp took place. In this tectonic setting, water-table fluctuations exceeded the rate of subsidence such that oxidation and reworking of peat took place giving rise to inertinite-rich coals.

3.3 SUMMARY

Certain sedimentary facies and facies associations present in the study area are distinctive to the Dwyka Group and Vryheid Formation. The Dwyka Group and base of the Vryheid Formation contain diamictites containing a variety of extrabasinal pebble- and cobble-sized clasts representing debris-flow sediments deposited during the Late Carboniferous-Early Permian glaciation of southern Africa.

A distinctive feature of the Vryheid Formation is the coarseness of sedimentary facies. Conglomerates and very coarse- to coarse-grained sandstones comprise a significant proportion of the stratigraphy. This suggests a predominance of coarse material in the source terrain and a limited distance of transport to the depository. High velocities for sediment transport are also implied. The finer grained sandstone facies are characterized by a dominance of wave-ripple cross-lamination over unidirectional current produced cross-lamination. The wave-ripple cross-lamination suggests that these sandstones were subjected to low-energy reworking by oscillatory currents.

The shale facies is distinguished by a predominance of siltstone over claystone, again reflecting the coarseness of sediment transported into the basin. The organic content of these sediments is high, consequently, where favourable conditions prevailed, these

sediments were bioturbated by a variety of organisms. Finally, the Vryheid Formation is unique within the Karoo Basin due to the presence of a number of relatively thick, laterally persistent coal seams. The seams are generally high in ash and inertinite contents, and low in vitrinite content.

CHAPTER 4

4 DEPOSITIONAL SYSTEMS

4.1 INTRODUCTION

Two approaches are used in the identification of sedimentary environments in the ancient rock record. The first approach is that of facies analysis whereby vertical facies arrangements and, where possible, lateral facies arrangements in the rock record are compared to facies sequences documented from modern environments, and ancient environments interpreted (Selley, 1976; Reading, 1986; Walker 1984). The second approach is that outlined by Galloway and Hobday (1983) who base environmental reconstruction of the rock record on the understanding of processes responsible for the erosion, transportation and deposition of sediment. These processes are reviewed in terms of the three-dimensional geometry and stratigraphic framework of genetic facies.

Walker (1984) who advocates the facies approach to environmental reconstruction has reviewed the facies model concept. A facies model is referred to by Walker (1984) as a general summary of a specific sedimentary environment, written in terms that make it usable as a norm for that environment, as a framework for future observations, as a basis for hydrodynamic interpretations, and as a predictive tool. This approach to facies modelling suggests that there is a finite number of environments, descriptive sedimentary facies and facies models. Anderton (1985) is sceptical of this approach as he views sedimentary facies as unique and consequently, each facies has a unique interpretation in terms of a facies model. Furthermore, Anderton (1985) views facies and environments as infinite. This standpoint is borne out by the development of fluvial facies models. Allen (1965) proposed four facies models to describe processes of fluvial sedimentation. These were: alluvial fan, braided, low sinuosity and

meandering. However, Miall (1977) proposed four facies models to describe low-sinuosity fluvial systems and in 1978 added two more models to the existing four in order to adequately classify the range of deposits attributed to the deposition of low-sinuosity (braided) fluvial systems. This suggests that facies models are variable and that facies models are inherently a simplistic summary of a variable range of deposits found in nature.

The concept of depositional systems has been informally proposed by Fisher and McGowan (1967). Depositional systems are considered as three-dimensional associations of lithofacies comprising one or more sedimentary environments related genetically within a stratigraphic framework. Consequently, one or more interpreted facies models could make up the depositional system. This approach views environments as process-response models in which environmental reconstruction is based upon process interpretation in the context of the three-dimensional geometry and stratigraphic position of genetic facies. Environmental interpretation of sedimentary rocks follows a hierarchical approach that first defines attributes of genetic sequences and then attempts site-specific interpretations. Galloway and Hobday (1983) suggest the following approach be adopted for the successful interpretation of depositional sequences:

a. Determine the framework of the depositional basin, then

b. identify principal genetic stratigraphic units, then

c. outline the three-dimensional depositional geometry and major depositional elements and component facies; then

d. identify recurrent vertical sequences within genetic sequences, then

e. recognise lateral and vertical facies associations cross-sectional geometry, then

f. interpret internal bedding, sedimentary structures, and textures of facies.

The more completely each level of description and interpretation can be concluded the more likely the a unique, predictive interpretation can be made at the next lower level (Galloway and Hobday, 1983).

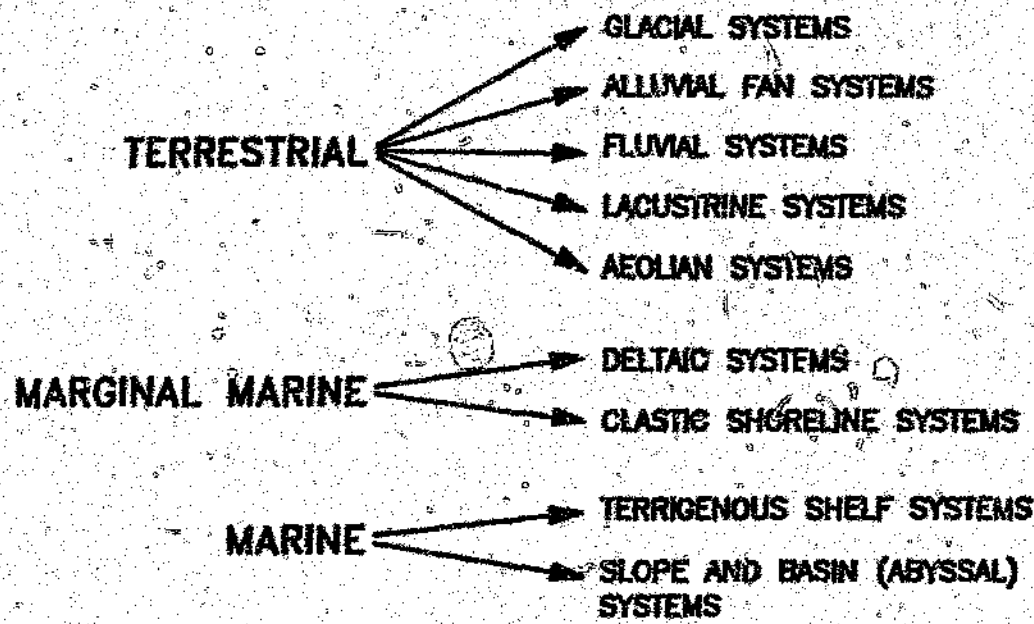
The major clastic depositional systems recognised are listed in Table 4.1. In this study, the approach of Galloway and Hobday (1983) to environmental reconstruction is followed. The genetic sequences are first identified, then depositional units are mapped out in three dimensions. Thereafter, individual facies, and vertical and lateral facies associations, are identified and interpreted from cores. This methodology is followed due to the data available for study. The data were obtained from borehole core, borehole logs and related outcrop adjacent to the study area. Unfortunately, due to the lack of outcrop within the study area, two-dimensional facies analysis determined from rock exposures was not possible. This limitation was offset by numerous borehole logs which permitted the three-dimensional analysis of facies sequences. These data enabled refined interpretations of sedimentary environments possible. Within the spectrum of depositional systems outlined in Table 4.1 the following depositional environments are interpreted; glacial depositional environments, fluvial depositional environments, deltaic depositional environments, and transgressive shallow-marine depositional environments.

4.2 FACIES INTERPRETED AS GLACIAL DEPOSITS

4.2.1 Description of Glacial Deposits

The most distinctive facies type attributed to glacial deposition is the diamictite facies. The facies frequently overlies basement and has a distribution controlled by palaeotopography. The diamictite forms a linear body of sediment over 20m thick situated, within a palaeovalley, along the eastern margin of the study area (Fig. 2.9).

Table 4.1 Classification of the major clastic depositional environments. After Galloway and Hobday (1983).



Away from the palaeovalley the facies thins gradually westwards and rapidly eastwards until it pinches out against basement elevated areas. In addition to the diamictite, a number of facies associations are present which suggest deposition by proglacial processes. These associations are :

a. Diamictite abruptly overlain by up to 10m of massive and planar cross-bedded granule conglomerate which grades upwards into sandstone and mudstone.

b. Interlaminated claystone and siltstone, up to 7m in thickness, overlain by diamictite.

c. Diamictite overlain by bioturbated siltstone, gravelly siltstone, and cross-stratified sandstone and granule conglomerate, capped by a coal seam (Fig. 4.1).

These facies associations which form part of both the No. 1 and No. 2 seam depositional sequences are illustrated in Figs. 2.6, 2.7, and 2.8. These figures show that the diamictite and granule-conglomerate facies association has a relatively widespread distribution. Conversely, the claystone-siltstone-diamictite facies association occurs as isolated deposits filling palaeodepressions. The third facies association, comprising diamictite overlain by interbedded bioturbated siltstone, gravelly siltstone and sandstone is best developed in the southeast of the study area where the basin deepens (Fig. 2.7).

4.2.2 Interpretation of Glacial Deposits

The pinch and swell geometry of diamictite filling palaeodepressions, as recorded in the study area, has also been documented by Le Blanc Smith and Eriksson (1979), Winter (1985), and Cairncross (1986) from the Witbank and Highveld Coalfields. The pinch and swell geometry of diamictite is a common feature of continental glacial deposits. The non-stratified nature of the diamictite and the presence of faceted clast shapes suggests deposition as a lodgement till by wet-based glaciers. A similar interpretation is invoked by Von Brunn (1977) for the tillite occurrences studied in

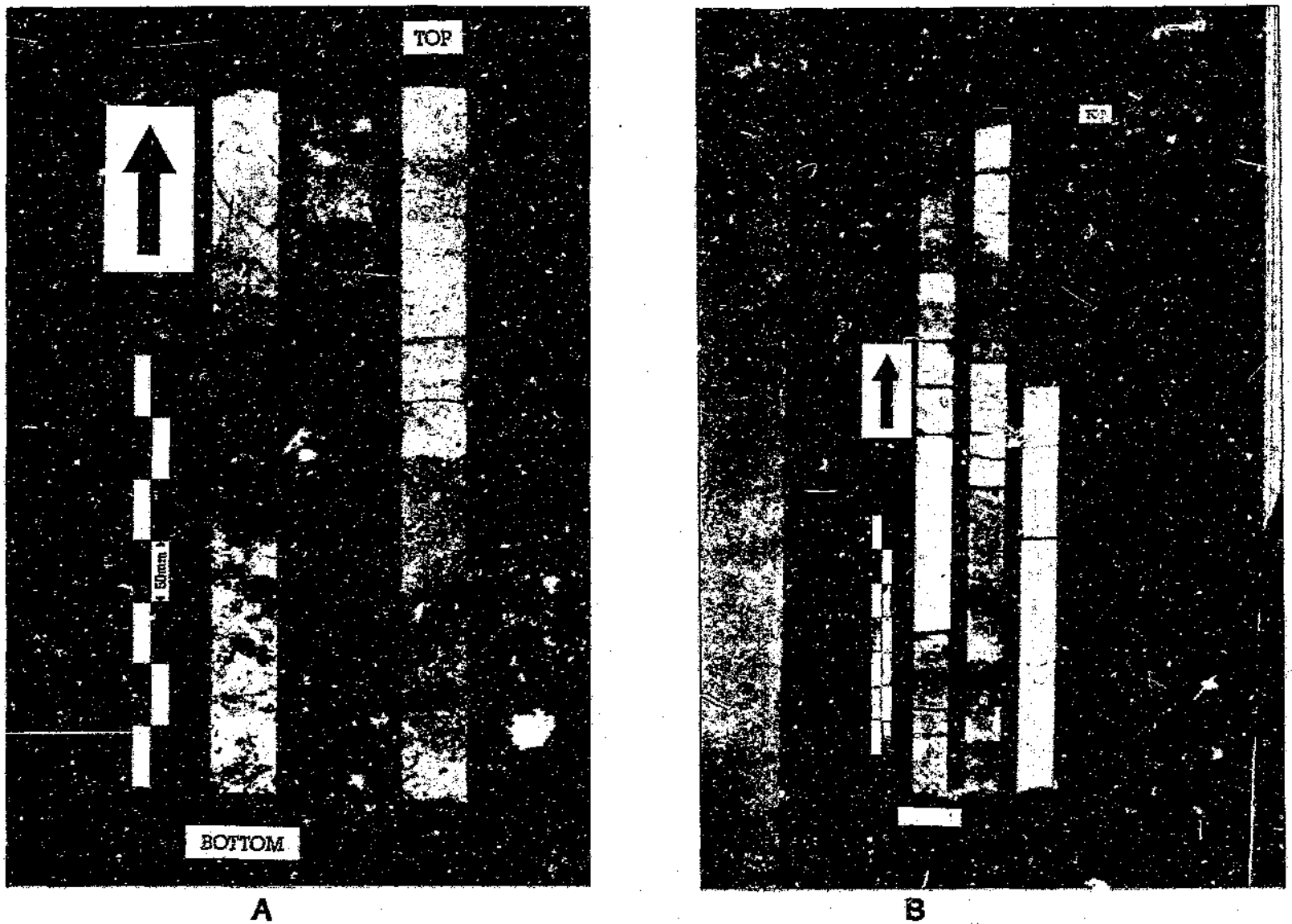


Figure 4.1. Facies associations interpreted as glacial deposits. The facies associations comprise diamictite overlain by bioturbated siltstone, gravelly siltstone, bioturbated sandstone and granule conglomerate (Figs. 4.1A and B). These sequences are often terminated by carbonaceous siltstone (Fig. 4.1B) and/or coal.

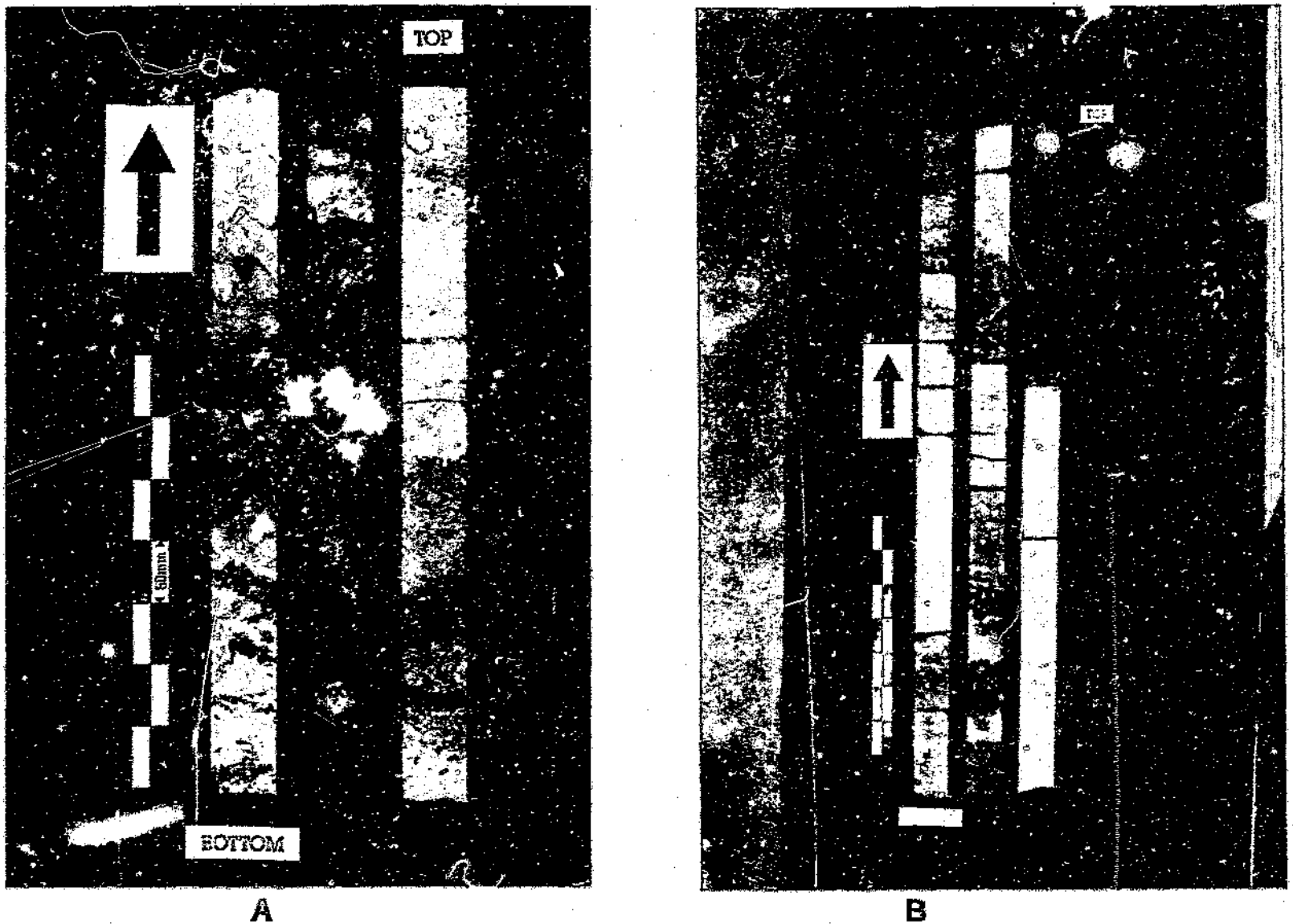


Figure 4.1. Facies associations interpreted as glacial deposits. The facies associations comprise diamictite overlain by bioturbated siltstone, gravelly siltstone, bioturbated sandstone and granule conglomerate (Figs. 4.1A and B). These sequences are often terminated by carbonaceous siltstone (Fig. 4.1B) and/or coal.

Natal. Thus, the diamictite is envisaged as a grounded ice deposit formed through glacial advance in a continental glacial environment. However, Cairncross (1986), argues that the transition from diamictite to overlying massive granule conglomerate suggests evidence for late-stage deposition of the diamictite.

Proglacial deposition is suggested by the diamictite-granule conglomerate, and the diamictite and fine-grained facies associations. The diamictite overlain by laterally extensive massive and planar cross-bedded granule conglomerate up to 10m thick, suggests deposition by braided-fluvial systems in a glaciofluvial environment. These deposits result from the reworking of glacial debris producing the aggradation of gravels and sandstones over the ice margin to form a sandur or outwash plain. In the proximal environments, situated in the adjoining Witbank Coalfield, the sediment is represented by pebble-sized conglomerates (Le Blanc Smith and Eriksson, 1979). However, these deposits have become somewhat less coarse grained further basinward and, in the study area, are represented by granule conglomerate. Modern outwash plains are well developed in Alaska, Arctic Canada, and Iceland (Rust 1972; Boothroyd and Ashley, 1975; Boothroyd and Nummedal 1978; Church 1983;). In these settings, the rivers are typically braided and of low sinuosity. The braided morphology of the river systems reflects the abundance of coarse-bedload material, variable discharge, and non-cohesive channel banks. In modern outwash plains, the river systems are only able to transport a small proportion of available bedload material except during floods (Ostrem, 1975). Similar processes are envisaged for the deposition of the granule conglomerate. The abundance of planar cross-bedding suggests that sediment was transported by the migration of sandwaves or transverse bars.

The facies associations comprising a) claystone-siltstone and interbedded diamictite and, b) diamictite overlain by bioturbated siltstone, containing the trace fossil *Siphonichnus*, grading upwards through gravelly siltstone, into cross-stratified sandstone and granule conglomerate (Figs. 2.6 and 2.7), suggest proglacial

sedimentation. Sedimentation into either, localised proglacial lakes or, a large brackish to shallow-marine water body is proposed. The laminated claystone-siltstone and gravelly siltstone facies accumulate from a combination of lacustrine suspension sedimentation, and turoidity or density underflow currents which occur during high meltwater discharge in summer. The siltstone and gravelly laminae represent turbidity deposition, the claystone proglacial suspension sedimentation, and the scattered granules in the claystone and siltstone, probable ice-rafted debris deposited from suspension into the lake-bottom. Coarsening upward ripple-laminated, poorly stratified and planar cross-bedded sandstones complete the lithofacies associations recording glaciolacustrine delta progradation over sites of diamictite deposition. Similar deposits are recorded by Biju-Duval *et al.*, (1981), for the late Ordovician glaciation of northern Africa, and Sudgen and John (1976), for the Pleistocene glacial deposits of North America and Europe. In addition, similar deposits have been described by Ashley (1975) from glacial lake Hitchcock, Massachusetts, and Gustavson (1975) from Malaspina Lake, southeastern Alaska.

In the southeast of the study area, the diamictite is overlain by several metres of interbedded bioturbated siltstone and gravelly siltstone. The siltstone is in turn overlain by cross-laminated and cross-bedded sandstones of the No. 2 seam depositional sequence (Fig 2.7C). This sequence records proglacial glaciodeltaic sedimentation and implies that the basin deepened to the southeast into the brackish to marginal marine Eocen "sea" (Cadle *et al.*, 1990).

4.3 FACIES INTERPRETED AS FLUVIAL DEPOSITS

4.3.1 Description: No. 1 and No. 2 Seam Depositional Sequences

The geometry and vertical facies stacking patterns of the conglomerates, and coarse-grained sandstones, present within the No. 1 and No. 2 seam depositional sequences are very similar, consequently, they have a similar description and interpretation. Lithologically, these coarse-grained sediments are dominated by the massive to poorly stratified conglomerate facies, the planar cross-bedded granule conglomerate and sandstone facies, and the trough cross-bedded sandstone facies. The interlaminated sandstone-siltstone facies, carbonaceous siltstone facies and coal (No. 1 or No. 2 seam) terminate the depositional interval. Vertically, the facies display individual to multiple fining-upward sequences. Detailed core logs reveal that individual fining-upward sequences vary in thickness between 1 and 5m and multiple sequences are up to 20m thick (Fig. 4.2).

The facies sequences commence with erosive surfaces floored with up to 50cm thick, laterally impersistent, massive to crudely stratified, medium- to small-pebble conglomerate, containing extrabasinal clasts and a carbonaceous matrix. Overlying this facies is a 3m to 5m thick sequence of massive and planar cross-bedded, granule conglomerate. This facies is the lithologically distinctive facies of these deposits. The basal 20cm to 1,5m of granule conglomerate displays a dark appearance due the incorporation of carbonaceous siltstone as the matrix and, in addition, carbonaceous siltstone intrabasinal clasts are scattered within the conglomerate. The dark colour of the granule conglomerate gives way to a white massive to planar cross-bedded granule conglomerate overlain by tabular cross-bedded, medium- to coarse-grained sandstone. Conglomerate and sandstone set thicknesses range between 20cm and 2m, and coset thicknesses up to 5m. Infrequently, trough cross-bedded sandstones with set thicknesses

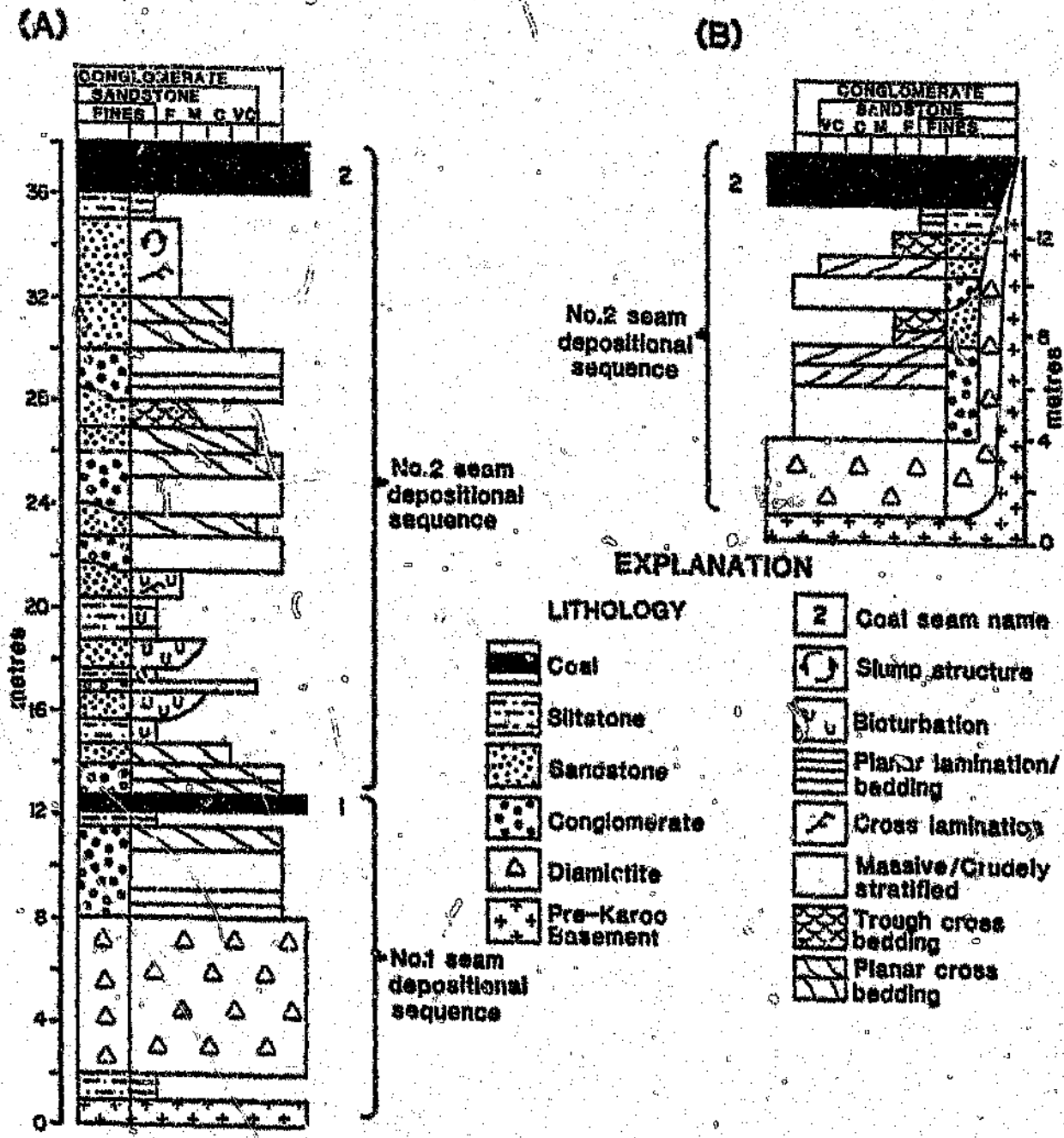


Figure 4.2. Idealised profiles through the No. 1 and No. 2 seam clastic sequences. Note the fining-upward facies arrangements which constitute channel-fill deposits.

between 10 and 50cm cap individual fining-upward sequences. Figure 4.3 is a core sequence through the No. 2 seam depositional sequence and illustrates many of the features described.

The finer-grained facies that overlie the conglomerate and cross-bedded sandstone facies associations are also well illustrated in Fig. 4.3. There is a gradual fining of grain size from cross-bedded sandstone into current-ripple cross-laminated sandstone and interlaminated sandstone and siltstone. The proportion of siltstone increases at the expense of sandstone and gives way to carbonaceous siltstone and coal. This finer grained facies association is restricted in areal distribution and the more frequent facies association is conglomerate and coarse-grained sandstone capped by the No. 2 coal seam.

Detailed isopachs through these deposits was not possible due to the lack of borehole control, however, selected borehole cross-sections through the stratigraphy reveal the general geometry of these deposits (Figs. 2.7 and 2.8). The facies underlying the No. 1 seam are thickest in the central portion of the study area. Here, the width to thickness ratio of these deposits, based on cross-section correlations, is >500. Elsewhere, the deposits are podiform in geometry and confined to localised palaeodepressions. The fining-upward coarse-grained sedimentary fill, between the No. 1 and No. 2 seams, also defines a sheet like geometry. Width to thickness ratios of these sediments vary between >800 in the north and >650 in the middle of the study area. The width to thickness ratio calculated for this interval in a north-south direction from Fig. 2.8, is >1000. These ratios, confirm the sheet-like geometry of this sedimentary interval.

Figure 4.3. Core of the No. 2 seam depositional sequence. Note the multistorey nature of the sequence and the general decrease in grain size upwards.

4.3.2 Description: No. 3 to No. 4 Seam Clastic Sequence

Overlying the No. 3 seam coal seam and separating the No. 4 Lower, No. 4 Upper and No. 4A coal seams, are sandstones interpreted as fluvial deposits. These sandstone bodies differ from the conglomerate and sandstone sequence below the No. 2 seam in two respects. Firstly, these deposits contain less conglomerate and more sandstone than the deposits below the No. 2 seam. Secondly, the deposits are lenticular in cross-section (2-6km), rather than sheet-like in geometry as documented below the No. 2 seam.

The similarity of the sandstones below and above the No. 4 Lower seam, in terms of texture, geometry and sedimentary structures, allows for a single description and interpretation for these deposits. These sandstone bodies are erosively based as evidenced by the carbonaceous siltstone matrix contained within the granule conglomerate and sandstones which floor the deposit. In addition, intraformational rip up clasts, of carbonaceous siltstone, are present in the basal 50cm of these sandstone dominated sequences. The most noticeable feature of these deposits are the stacked fining-upward facies sequences. Individual fining-upward facies sequences range in thickness between 50cm and 10m and multistoried fining-upward facies sequences range in thickness between 1m and 30m in thickness.

The fining-upward facies sequences commence with either, massive or planar cross-bedded granular conglomerate or, very coarse- to medium-grained planar cross-bedded sandstone. These facies fine upward into the coarse- to medium-grained trough cross-bedded sandstone, and fine-grained cross-laminated sandstone. The sequences are frequently terminated by either, the interlaminated sandstone-siltstone or, the carbonaceous siltstone facies. The erosive nature of the succeeding fining-upward facies sequence causes, the incomplete preservation, or removal of, the interlaminated sandstone-siltstone or carbonaceous siltstone facies. Thus, frequently, the fine-grained

facies only cap the uppermost fining-upward facies sequence. Consequently, the succession comprises incomplete facies sequences dominated by granule conglomerate and sandstone lithologies.

The most abundant sedimentary structure is planar cross-bedding. Set thicknesses range from 20cm to 1,5m and cosets up to 10m. Individual foresets display fining-upward graded laminae with particle size ranging from granule-grade gravel to medium-grained sand. The foresets dip between 20°-30°. Figure 4.4 illustrates the sandstone-dominated sequence between the No. 3 and No. 4 coal seams. The erosive basal contact of the sequence is evident as is the planar cross-bedded sets emphasised by micaceous laminae. Twelve erosive bases to the fining-upward facies sequences are documented from the core, illustrating the pulsating nature of sedimentation. Figures 2.16 and 2.17 show that these sandstone bodies thicken along axes and are erosive in nature. The erosive nature of the sandbodies is illustrated through the removal of the No. 3 coal seam and the deposits situated between the No. 2 and No. 3 seams.

Figure 4.5 is a representative core sequence between the No. 4 Lower and No. 4 Upper coal seams. This sequence is coarser grained than the underlying sandstone-dominated sequence as a greater proportion of granule conglomerate constitutes these lithologies. Planar cross-bedding is the dominant sedimentary structure and interlaminated sandstone-siltstone and carbonaceous siltstone facies are preserved at the tops of fining-upward facies sequences. The contrast in facies types, and the erosive floor at the base of fining-upward facies sequences, suggests periodic rapid fluctuations in discharge.

Figure 4.6 illustrates core sequences between the No. 4 Lower and No. 4 Upper coal seams where the parting between the two seams is less than 3m thick. These cores also illustrate small-scale, fining-upward facies sequences. Here, the facies commence with planar cross-bedded granule conglomerate or sandstone overlain by cross-bedded and

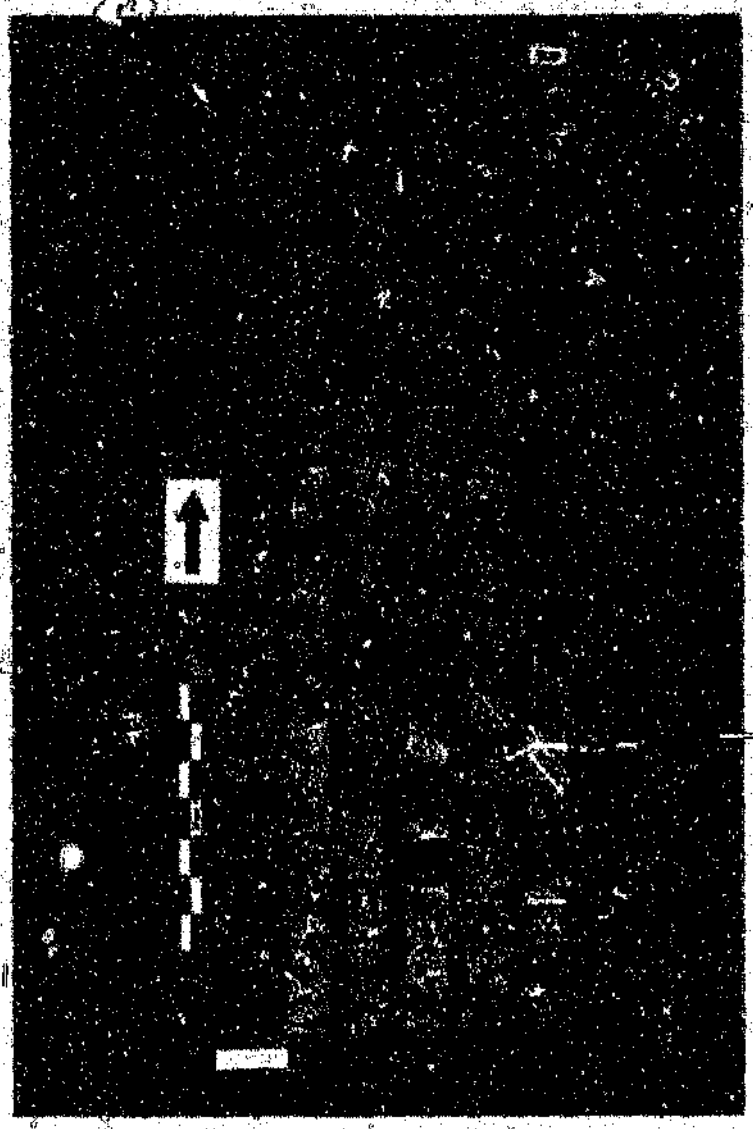


Figure 4.4. Core of the sandstone-dominated interval between the No. 3 and No. 4 coal seams. Note the erosive base to the sandstone, the fining-upward facies assemblages, and the dominance of the planar cross-bedded sandstone facies.

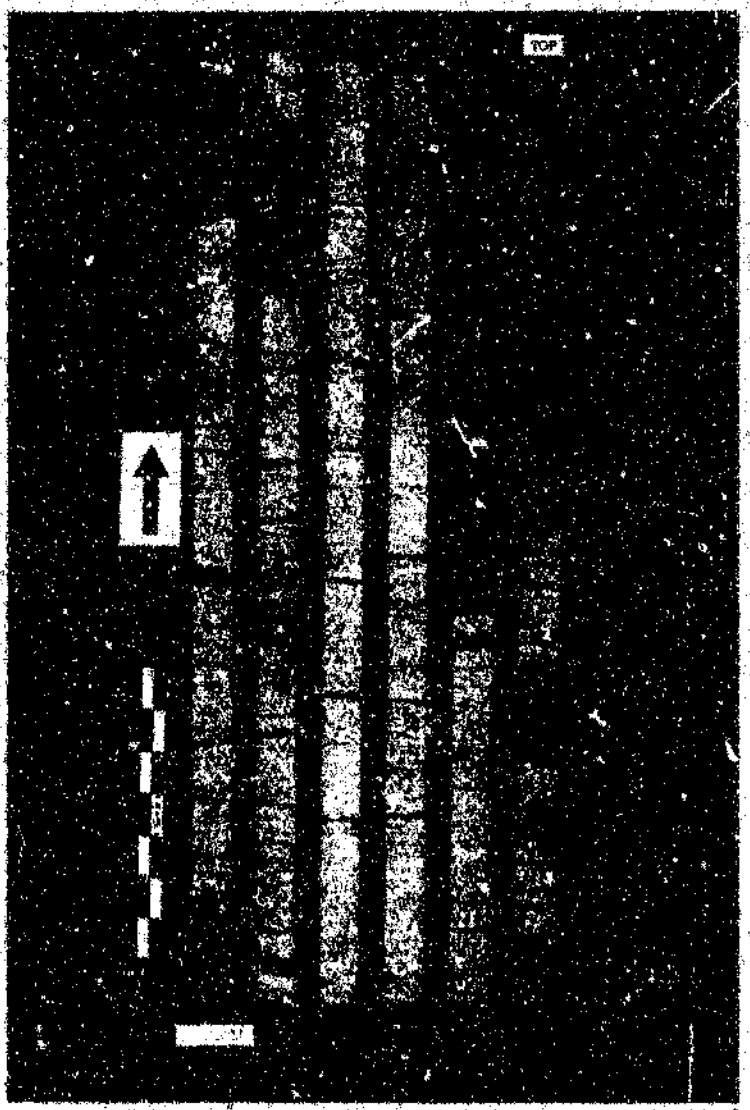


Figure 4.5. Core sequence of the interval between the No. 4 Lower and No. 4 Upper seams. Note the vertically stacked fining-upward facies assemblages capped by the interlaminated sandstone-siltstone facies. Overall, the interval fines upward in grain size.

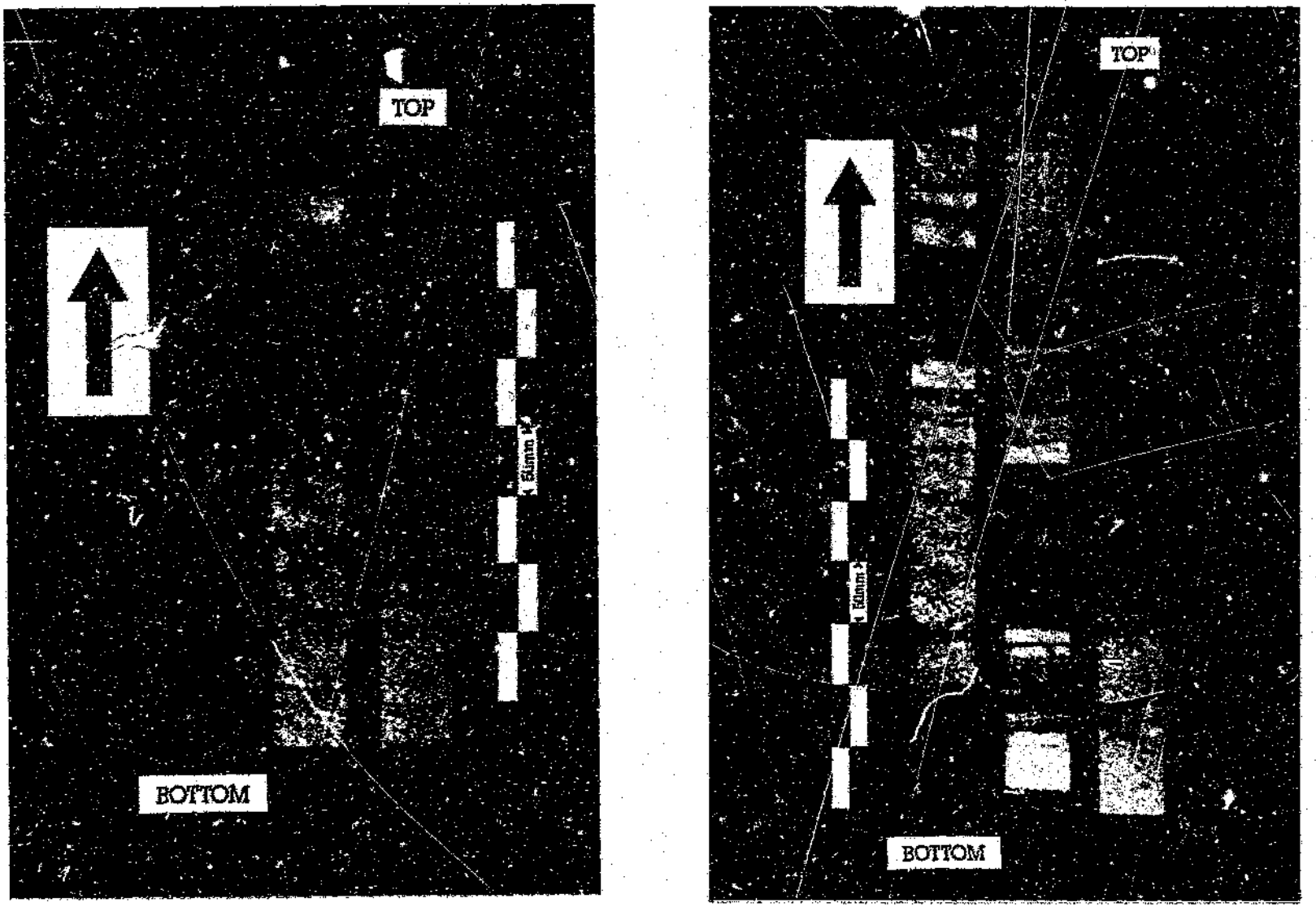


Figure 4.6. Core sequences of the No. 4 Lower-No. 4 Upper parting where the interval thickness is less than 3m. Note the small-scale fining-upward facies assemblages, illustrated by core sequences (A) and (B), reflecting conditions of fluctuating discharge.

planar laminated sandstone and are terminated by the interlaminated sandstone-siltstone facies. Figure 4.6A also illustrates plant rootlets at the top of the sequence suggesting the autochthonous origin for the No. 4 Upper coal seam while the small scale, fining-upward, facies sequences shown in Fig. 4.6B, indicate that fluctuating discharges operated in this sedimentary setting.

The geometry of these coarse-grained facies associations is distinctive. The sediments thicken along axes which are >35km in length. Figure 4.7 is an isopach of the sandstone between the No. 4 Lower and No. 4 Upper coal seams. The 3m contour defines the edge of the channel which attains a maximum thickness of 15m. The sandbody has a width that varies between 2km and 10km. Thus, the width/thickness ratios of this sandbody vary between 1:133 and 1:666.

Figure 4.8 illustrates the isopach of the sandstone interval between the No. 4 Upper coal seam and the shales at the base of the No. 5 seam depositional sequence. The sandstone isopach includes both the fluvial sandstones and transgressive shallow-marine sandstones which terminate the No. 4 seam depositional sequence. It was not possible from the majority of borehole logs to distinguish between these two sandstones therefore the entire sandstone interval was isopached. The transgressive sandstones, identified from borehole cores, are less than 4m thick therefore the interval isopached largely reflects the geometry of the channel sandstones above the No. 4 Upper seam. Two linear sandbodies are distinguished by the 8m contour. The first sandbody has a northsouth orientation and is situated immediately to the west of the No. 4 Lower and No. 4 Upper sandbody. The northsouth orientated sandbody intersects a sandbody with an east-west orientation present in the south of the study area (Figs. 4.8 and 2.16). The minimum lengths of the sandstone bodies are 31km and 24km respectively, and have width to thickness ratios (based upon the 8m contour) that vary between 1:138 and 1:500.

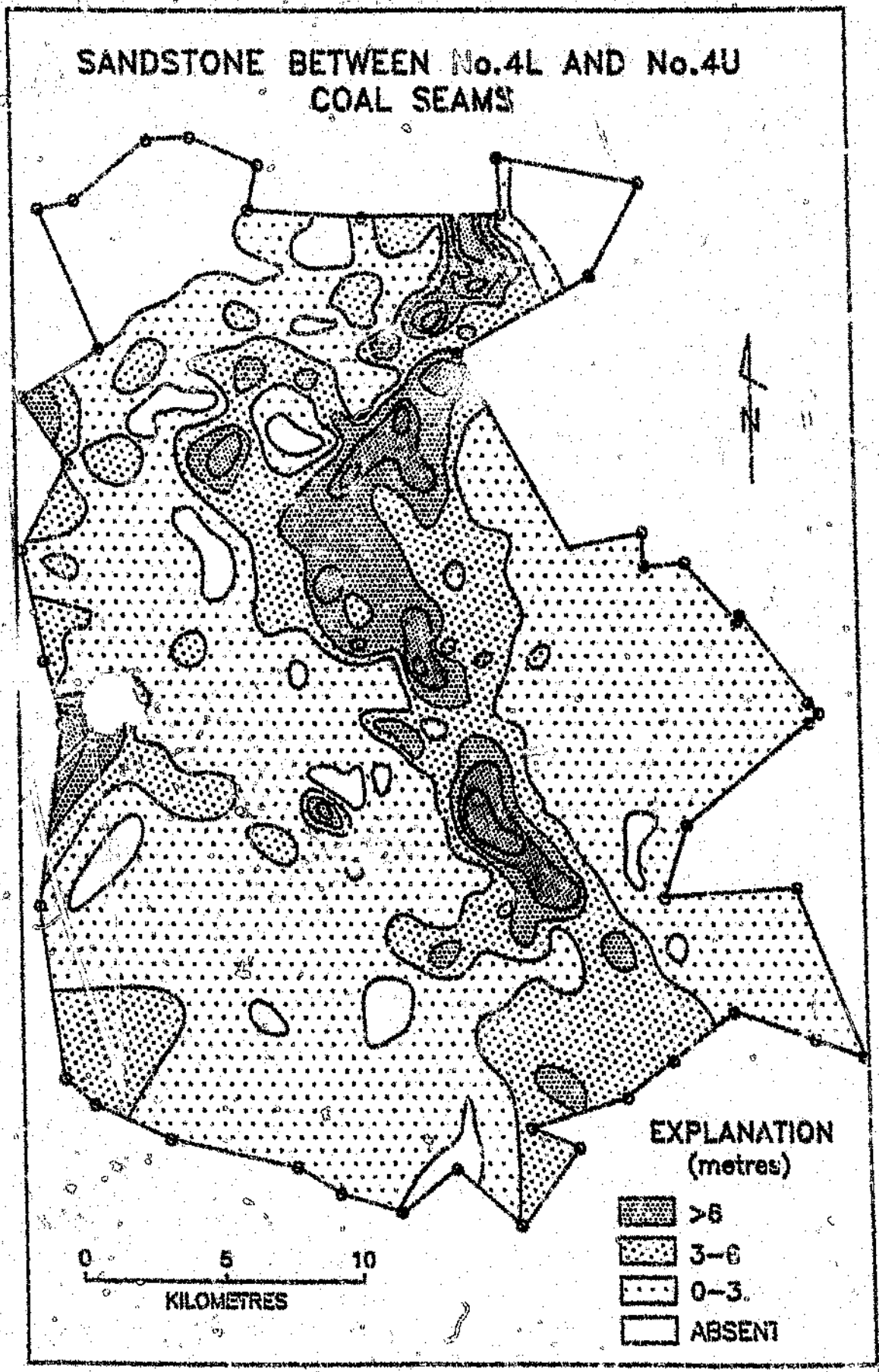


Figure 4.7. Isopach of the sandstone between the No. 4 Lower and 4 Upper coal seams. The sandstone displays a linear geometry, varies between 2-10km in width and is up to 15m in thickness.

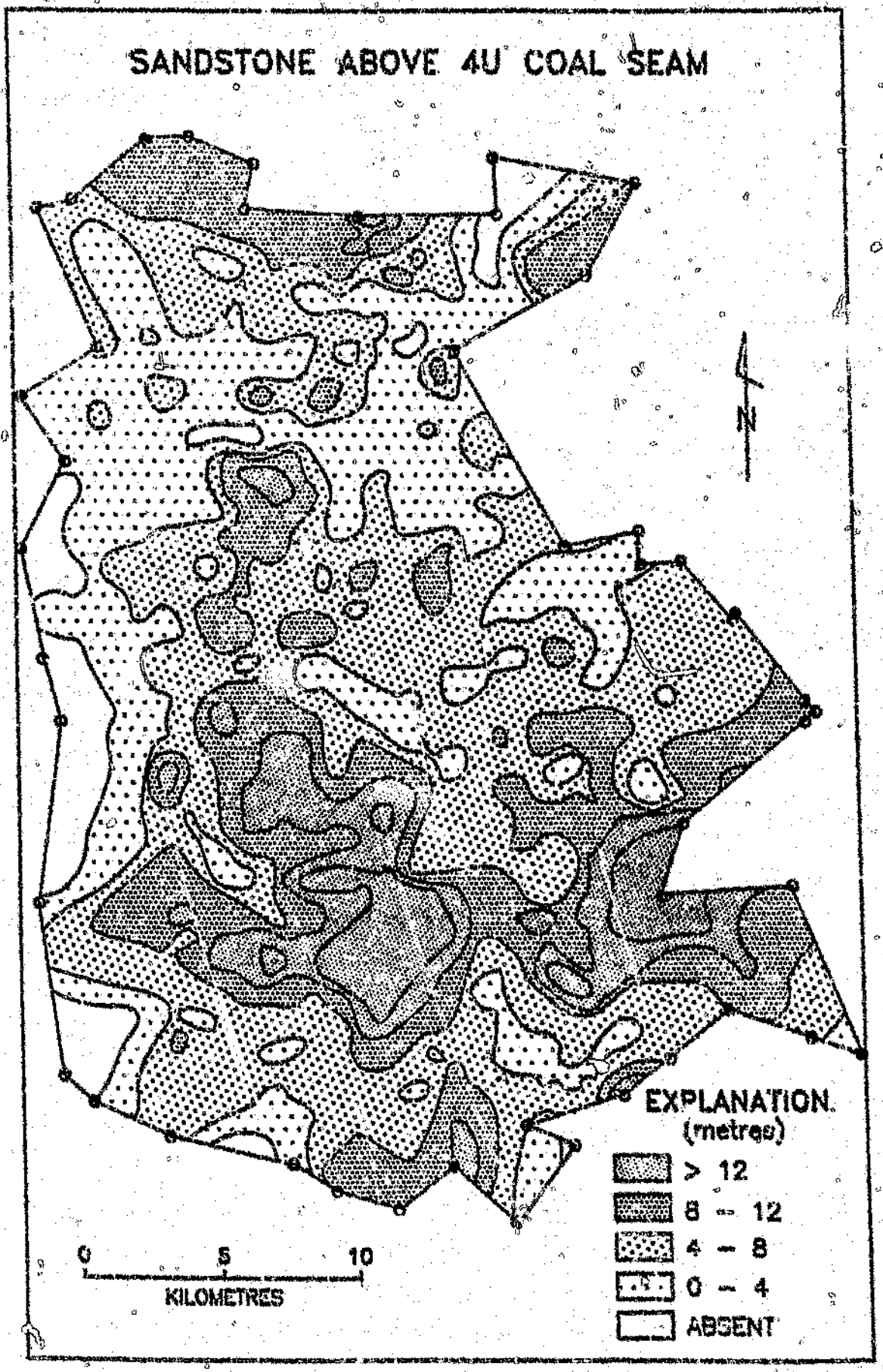


Figure 4.8. Isopach of the sandstone between the uppermost No. 4 seam and the siltstone at the base of the No. 5 seam depositional sequence. Note the north-south linear thickening along the western margin of the study area and the east-west linear thickening in the south of the study area.

4.3.3 Interpretation of Fluvial Deposits

Fluvial systems serve as conduits through which sediment is transported from a source area into lacustrine or marine basins. Fluvial style is a complex response to both autocyclic and allocyclic controls. Climate and tectonics are primary allocyclic controls and determine sediment load, discharge, slope, vegetation, and channel pattern. Geomorphologists and sedimentologists have attempted to classify fluvial systems using a variety of criteria. Channel pattern was proposed by Leopold and Wolman (1957) to determine river morphology. The following river patterns were identified; braided, meandering, straight, and anastomosed. This classification has been quantified by Rust (1978) who has proposed the use of numerical measures of channel sinuosities and of braiding.

Sediment load was used by Schumm (1968) for stable alluvial channels as a means of classifying fluvial systems. This scheme classifies river systems as; suspended-load channels (bed load <3% of total load), mixed-load channels (bed load 3-11% of total load and bed-load channels (bed load >11% of total load). Textural profiles and vertical facies sequences have also been used to classify fluvial deposits. Studies by Visher (1965), Allen (1965), McGowen and Garner (1970) and Jackson (1976) defined the fining-upward meandering-fluvial model, and the reviews by Miall (1977, 1978) and Rust (1978) postulated a variety of models for the braided-fluvial environment which highlighted the complexity of processes and deposits within braided-fluvial systems.

Jackson (1978) and Bridge (1985) have questioned the criteria upon which the fluvial models are based and this has led to workers pursuing different approaches to fluvial classifications. The recognition of the lack of two- and three-dimensional analyses of fluvial deposits has prompted a number of workers to pursue research in this area. This approach is referred to as "alluvial architecture" Miall (1985, 1988). Friend (1983)

recognised the differences in channel shape and internal channel fill and suggested a classification of five models of alluvial architecture. These models are based upon sheet-flood, fixed-channel, and mobile-channel belt geometries, combined with, coarse bed-load deposits, and coarse and fine bed-load and suspension-load deposits. Allen (1983) mapped out in two dimensions discordances within fluvial sandstones and identified the deposits of dunes, and simple, compound and composite-compound bars, and related these architectural elements to fluvial processes. This approach has been built upon by Miall (1985, 1988) and led to the identification of six orders of bounding surfaces which allow for the analysis of fluvial deposits. Thus, there has been, in the recent literature, a swing away from the facies model approach to fluvial systems, and more emphasis placed upon the description of deposits which illustrate the wide variety of fluvial depositional styles.

A further approach to the classification of fluvial deposits has been documented by Galloway and Hobday (1983). They recognise that fluvial systems consist of a mosaic of genetic facies comprising channel-fill, channel-margin and flood-basin deposits. The most diagnostic deposit is the channel facies and the recognition of these deposits is a key to unravelling the identity of a fluvial system. The emphasis of this classification is the identification of the three-dimensional geometry of channel-fill sequences. Their scheme defines channel-fill sequences as either, low-sinuosity or, high-sinuosity channels. Low-sinuosity channels encompass both sand-rich and mud-rich fluvial systems. The channels of bed-load, sand-rich fluvial systems are dominated by a variety of braid bars including lateral, transverse and longitudinal bars. Modern river systems of this type have been described for the Bijou Creek River by McKee *et al.*, (1967), the Brahmaputra River by Coleman (1969) and Bristow (1987), the Donjek River by Williams and Rust (1969) and Rust (1972), the Durance and Ardeche Rivers by Doeglas (1962), the Platte River by Smith (1970, 1971), the Tana River by Collinson (1970), the Scott River by Boothroyd and Ashley (1975), and the South Saskatchewan River by Cant and Walker (1978).

Mud-rich, low-sinuosity channels display highly convex channel cross-sections and both bed and bank accretion produce a symmetrical channel-fill deposit. Galloway and Hobday (1983) categorise the distributary-channel model discussed by Brown *et al.*, (1973), and the low-sinuosity channels described by Schumm (1968), Fisher *et al.*, (1969) and Smith and Smith (1980), as examples of this type of fluvial system.

High-sinuosity channels include meandering-fluvial systems. Included within this category of fluvial systems is the best known fluvial model termed the fine-grained, point-bar sequence as described from the Brazos River (Bernard *et al.*, 1970), the Mississippi River (Fisk, 1944; Frazier and Osanik, 1961; Davies, 1966), the Red River (Harms *et al.*, 1963) and the Wabash River (Jackson, 1975, 1976). Also included within the category of high-sinuosity channels is the chute-modified point-bar model or, coarse-grained point-bar model (Fisher and Brown, 1972) which accounts for the increasing variability of discharge and coarseness of sediment load in meandering fluvial rivers.

Based upon the preferred classification of Galloway and Hobday (1983) for fluvial systems, the coarse-grained deposits present below the No. 2 seam and below and above the No. 4 coal seam, including seam splits, are interpreted as low-sinuosity, bed-load dominated, sandy fluvial systems. This conclusion is drawn from detailed descriptions and analyses of the geometry of the channels, the vertical and lateral facies relationships within channel sequences, the texture of the channel-fill lithologies, the presence of fossil plant material, and the proximity of coal seams to these deposits.

Many of the studies on modern-day braided rivers describe processes and sedimentation on widely different scales. For example, the Platte River is 1-2m deep and 60-600m wide, the south Saskatchewan River is 3m deep and 600m wide, and the Brahmaputra River is 5m-35m deep and 3-18km wide. The sandbodies mapped out in this study are

2-10km wide and up to 15m thick. Consequently, they appear to have a geometry that would be approximately half as wide and deep as the Brahmaputra River. Two excellent studies have been undertaken on the Brahmaputra River by Coleman (1969) and, more recently, by Bristow (1987). Channel geometry and hierarchy, barforms, channel movement and style of deposition are documented which provide great insight into the processes that take place in the low-sinuosity reach of this river system and, by analogy, the deposits of the study area.

Based upon the study by Bristow (1987) the Brahmaputra River has a recorded maximum and minimum discharge of 76,500 m^3/sec and 3,950 m^3/sec respectively. The mean flow depth recorded was 5m, the maximum flow depth 35m, and the maximum surface velocity 3m/sec. The average width of the river is about 10 km, individual channels have width/depth ratios that vary between 50:1 and 500:1. Suspended load of the Brahmaputra River is significant with up to 4,544ppm recorded. Bed-load percentages, based upon estimates of suspended load, vary between 15% (Laursen, 1958) and 50% (Curray and Moore, 1971) of total load discharge. Measurements of bed-load grain size suggest a temporal change in grain size which varies from very fine-grained sand (low discharge) to fine- and medium-grained sand (high discharge).

The hierarchy of channels in the Brahmaputra is very different in both scale and processes from that described for the Donjeck River (Williams and Rust, 1969). Bristow (1987) documents three orders of channels. First-order channels include the entire river with the channel margin defined by the outermost river banks. Second-order channels are separated by medial bars and have both braided and meandering courses. These channels have constant patterns over at least ten years and are in apparent equilibrium with local discharge, slope, and sediment supply.

Second-order channels migrate independently of each other and represent an anastomosing reach within the braided system. Third-order channels are recognised within both braided and meandering courses and are present on bar tops.

Both bars and bar assemblages are documented by Bristow (1987). The following bar types are distinguished; lateral (point) bars, medial (braid) bars, diagonal bars, scroll bars, and tributary (chute) bars. Lateral bars occur on the inside of channel bends and in straight channel reaches. They are associated with subsidiary scroll bars and chute channels. The lateral bars form in first- or second-order channels and are modified by third-order channels. Medial bars are deposited at areas of flow divergence and tributary bars at channel junctions. The bar types identified, coalesce to form bar assemblages (megaforms) which are up to 30km in length and are associated with channel anastomosis.

The movement of channels with time are related to channel hierarchy. The first-order channel has avulsed once in 200 years and has migrated within a channel belt 20km wide, with an average migration rate of 70m per annum. Second-order channels have depths of up to 40m, maximum widths of 5km and migrate up to 1km per year. Importantly, most channel movement occur by lateral migration and to a lesser extent by channel switching. Third-order channels present within second-order channels, are hundreds of meters wide, migrate in response to local hydraulic conditions, and modify high-order bars during falling water stage. Styles of deposition recorded over a three year period within the Brahmaputra River are presented by Bristow (1987). These data provide new information on the amount and mechanism of deposition in low-sinuosity river systems. Sediment addition to pre-existing bars amounts to 53% of the area of deposition, whereas lateral accretion to the channel bank accounts for 19%, channel abandonment 15%, and new mid-channel bars 13%. Addition to pre-existing bars takes place by flank accretion (57%), downstream accretion (29%) and upstream accretion (14%). Lateral and upstream accretion to pre-existing bars takes place by the wrapping

around and up the flanks of the bar by scroll bars and megaripples. This mechanism of sedimentation has been postulated by both Allen (1983) and Bridge (1985). Downstream accretion to bars takes place during periods of reduced discharge where large flow separation zones are present. Here, deposits of climbing ripple lamination and mud drapes up to 2m thick are deposited. Lateral accretion to channel banks in the form of exceedingly large point bars also occur. The point bars are up to 10km long and 3km wide. The scale of these point bars, poses a major problem in identifying this type of feature in ancient deposits. Mid-channel bars have an elongate diamond shape, and could enlarge to form complex medial bars. Channel abandonment and subsequent channel-fill deposits only comprise 15% of deposition and preserve the fill of mainly second-order and third-order channels.

Based upon the study by Bristow (1987), the internal stratigraphy of the Brahmaputra River is controlled by second-order channel movements. The width to depth ratios of the channels are high (up to 500:1) consequently, lateral-accretion surfaces would prove difficult to recognise in the geologic record. A similar conclusion has been reached by Allen (1983) from his study of the Devonian Brownstones from Wales. Continual reworking of sediment within the channel belt will produce complex erosional and depositional surfaces. During falling water stage, suspended sediment is deposited on the downstream margin of medial and lateral bars. Smaller scale complications to the internal stratigraphy of a low-sinuosity sandbody will result from the low-stage reworking, entrenchment, and abandonment, of third-order channels.

A significant finding of the Brahmaputra River st[illegible], is the fact that the elements of the braided stream model, meandering stream model, and anasto[illegible] can all coexist in sandy braided rivers of large dimensions. It would not be surpris[illegible]g, if these three elements of fluvial sedimentation were not present in the fluvial deposits of the Highveld Coalfield. However, the lack of small-scale, two-dimensional data precludes detailed two-dimensional features from being identified. Cross-sections normal to the

channel axis of the No. 4 Lower-No. 4 Upper sandstone parting, illustrate features similar to those described from the Brahmaputra River (Fig. 4.9). Firstly, the channel has a sheet-like, lenticular cross-sectional geometry (Fig 4.7). Secondly, although the geometry of the channel sandstone parallel to the channel axis undulates in thickness, in general, the sandbody thins from north to south down the palaeoslope. Within the channel sandstone the carbonaceous siltstone facies, and cross-laminated interbedded sandstone-siltstone facies, could represent surfaces demarcating bar complexes or the edges of medial and lateral bars. Well defined fining-upward sequences, 1 to 3m thick, may represent channel fill deposits of third-order channels. These deposits comprise scoured bases overlain by small-pebble conglomerate and trough cross-bedded and cross-laminated sandstone. Sequences >5m thick, dominated by planar cross-bedded and planar bedded granule conglomerate and coarse-grained sandstone, represent the accumulation of transverse, linguoid and cross-channel bars which form either, complex sand flats (Cant and Walker, 1978) or, bar complexes (Bristow, 1987). Lateral-accretion deposits are not and cannot be distinguished from a study of this nature due to the lack of small-scale, two-dimensional data.

Differences in the geometry of the coarse-grained sediments are present below the No. 1 and between the No. 1 and No. 2 seams, and between the No. 4 Lower and 4 Upper seams. The deposits associated with the No. 1 and No. 2 coal seams are sheet-like in geometry and are tens of kilometres wide, while the sandstone bodies associated with the No. 4 seam vary in width between 2 and 8km. The differences in the geometry between these deposits is inferred to be controlled, in part, by the activity of vegetation in confining fluvial activity. In the case of the fluvial sediments present below the No. 1 coal seam, and between the No. 1 and No. 2 coal seams, fluvial activity was largely unconfined by peat growth. Hence, channels were flanked by uncohesive channel banks allowing for channel switching and the development of a well defined multilateral

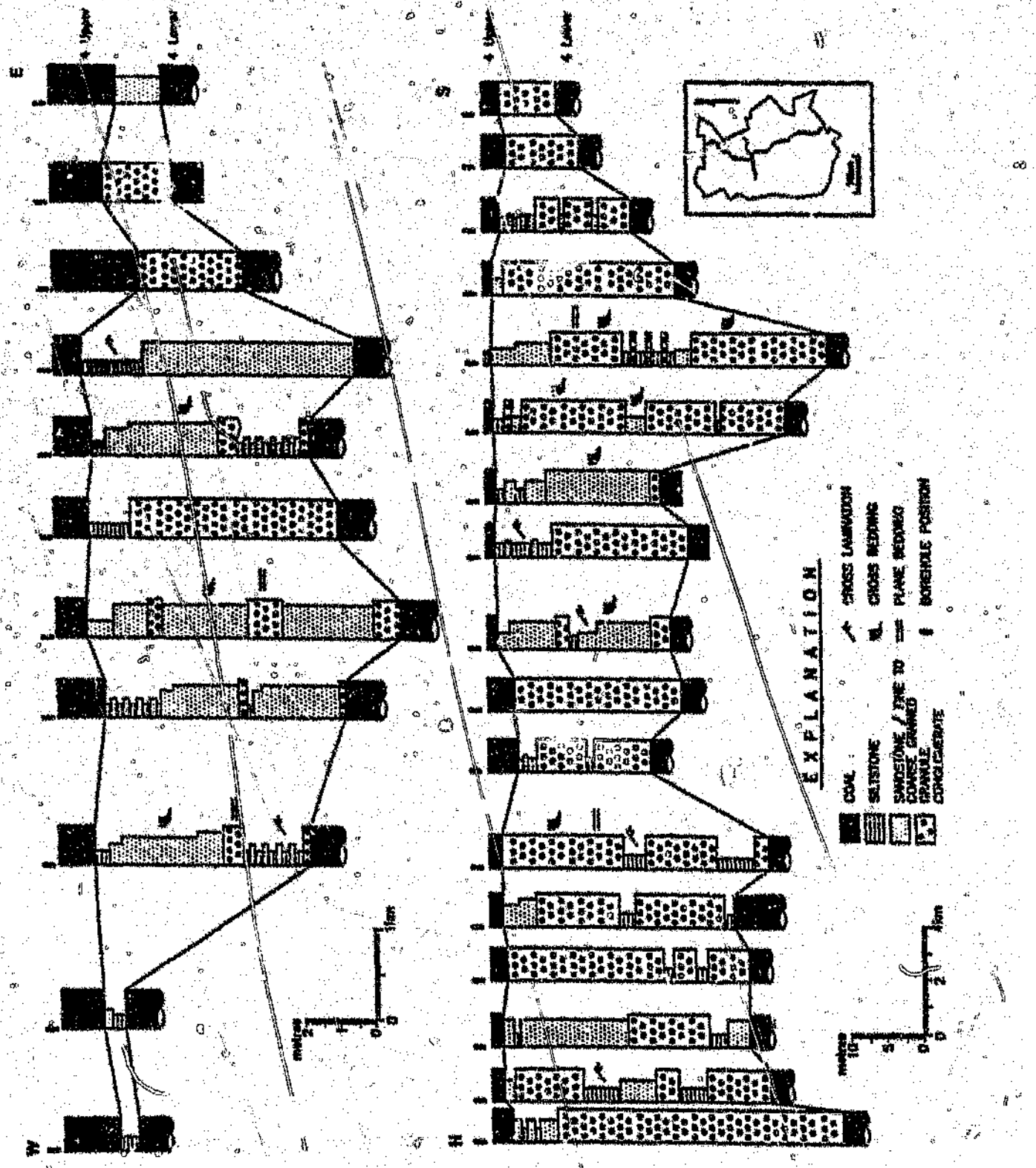


Figure 4.9. Westeast and northsouth cross-sections through the No. 4 Lower/No. 4 Upper seam fluvial channel. The cross-sections illustrate profiles of the down-channel and cross-channel fills. Note the well defined fining-upward sequences which may represent third-order channel-fill sequences.

sandstone body similar to a braid plain. Peat accumulation essentially followed abandonment of the fluvial tract and consequently was not a significant feature of channel confinement.

The No. 4 Lower-No. 4 Upper seam clastic sediments were deposited during the No. 4 seam peat forming event. The coarse-grained channel-fill deposits are defined by the 3m contour (Fig. 4.7) and the surrounding flood plain by <3m of fine-grained sediments including; the cross-laminated siltstone and sandstone facies, the carbonaceous siltstone facies and the No. 4 Upper seam peat. Hence, it is reasonable to assume that the lenticular cross-sectional geometry of the sandstone is in part due to the cohesive nature of the floodplain deposits. Smith (1976) has commented on the ability of plant communities to significantly confine flow in channels. The presence of vertical-accretion deposits including the accumulation of peat in floodplains is often used as a criterion for the existence of high-sinuosity rivers (Schumm, 1963). Thus, the No. 4 seam fluvial deposits exhibit criteria associated with high-sinuosity rivers (stable river banks and floodplain deposits) and low-sinuosity rivers (linear channel outline, coarse-bedload material, high width to thickness ratio). There is a similarity in grain size, and vertical and lateral facies arrangements, between the No. 1 and No. 2 seam fluvial deposits and the No. 4 seam fluvial deposits. This would imply that the No. 4 seam floodplain sediments were an important factor in confining the braided reach of these fluvial deposits.

The floodplain deposits of the No. 4 Lower-No. 4 Upper sedimentary interval form a sedimentary continuum from fining-upward crevasse-splay sandstones adjacent to the channel margin which grade laterally into interlaminated sandstones and siltstones and ultimately carbonaceous siltstones. The interlaminated sandstones and siltstones and carbonaceous siltstones are situated in areas remotely removed from the active channel margin. The presence of the No. 4 Upper coal seam and rootlets in the carbonaceous siltstone facies implies the floodplain comprised poorly drained swamps and lakes into

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The No. 4 Lower-No. 4 Upper seam clastic sediments were deposited during the No. 4 seam peat forming event. The coarse-grained channel-fill deposits are defined by the 3m contour (Fig. 4.7) and the surrounding flood plain by <3m of fine-grained sediments including; the cross-laminated siltstone and sandstone facies, the carbonaceous siltstone facies and the No. 4 Upper seam peat. Hence, it is reasonable to assume that the lenticular cross-sectional geometry of the sandstone is in part due to the cohesive nature of the floodplain deposits. Smith (1976) has commented on the ability of plant communities to significantly confine flow in channels. The presence of vertical-accretion deposits including the accumulation of peat in floodplains is often used as a criterion for the existence of high-sinuosity rivers (Schumm, 1963). Thus, the No. 4 seam fluvial deposits exhibit criteria associated with high-sinuosity rivers (stable river banks and floodplain deposits) and low-sinuosity rivers (linear channel outline, coarse-bedload material, high width to thickness ratio). There is a similarity in grain size, and vertical and lateral facies arrangements, between the No. 1 and No. 2 seam fluvial deposits and the No. 4 seam fluvial deposits. This would imply that the No. 4 seam floodplain sediments were an important factor in confining the braided reach of these fluvial deposits.

The floodplain deposits of the No. 4 Lower-No. 4 Upper sedimentary interval form a sedimentary continuum from fining-upward crevasse-splay sandstones adjacent to the channel margin which grade laterally into interlaminated sandstones and siltstones and ultimately carbonaceous siltstones. The interlaminated sandstones and siltstones and carbonaceous siltstones are situated in areas remotely removed from the active channel margin. The presence of the No. 4 Upper coal seam and rootlets in the carbonaceous siltstone facies implies the floodplain comprised poor drained swamps and lakes into

which coarse-clastic deposits were fed from the active channel margins. A mechanism of crevasse-splay sedimentation similar to that described by Coleman (1969) for the Brahmaputra River is envisaged for these deposits. A detailed east-west cross-section from the floodplain into the channel margin of the floodbasin deposits is illustrated in Fig. 4.10. In the east, the cross-section illustrates the proximal sandstone-dominated, fining-upward, crevasse-splay sequence structured by cross-bedding and cross-lamination. This sequence grades westwards into a more distal crevasse-splay deposit, comprising a coarsening-upward sequence. The coarsening-upward sequence comprises siltstone at the base overlain by fine-grained sandstone which, in turn, grades laterally into the interlaminated sandstone and siltstone and ultimately into carbonaceous siltstone. A similar crevasse-splay sequence has been documented [illegible] proximal fluvial deposits of the Saskatchewan River by Smith *et al.*, (1989).

In summary, these deposits of the study area are classified as coarse-grained, low-sinuosity, bed-load dominated fluvial systems. The fluvial systems in the lower part of the stratigraphy (No. 1 and No. 2 seam depositional sequences) were unconfined, tens of km in width, and resembled braid plains. The fluvial systems present below, within, and above, the No. 4 seam have a more linear geometry. These fluvial systems are between 2 and 10km in width and are both multistorey and multilateral in character. The restricted cross-sectional geometry of these sandstones, compared to the underlying fluvial deposits, suggests that overbank sediments contained the width of channel migration.

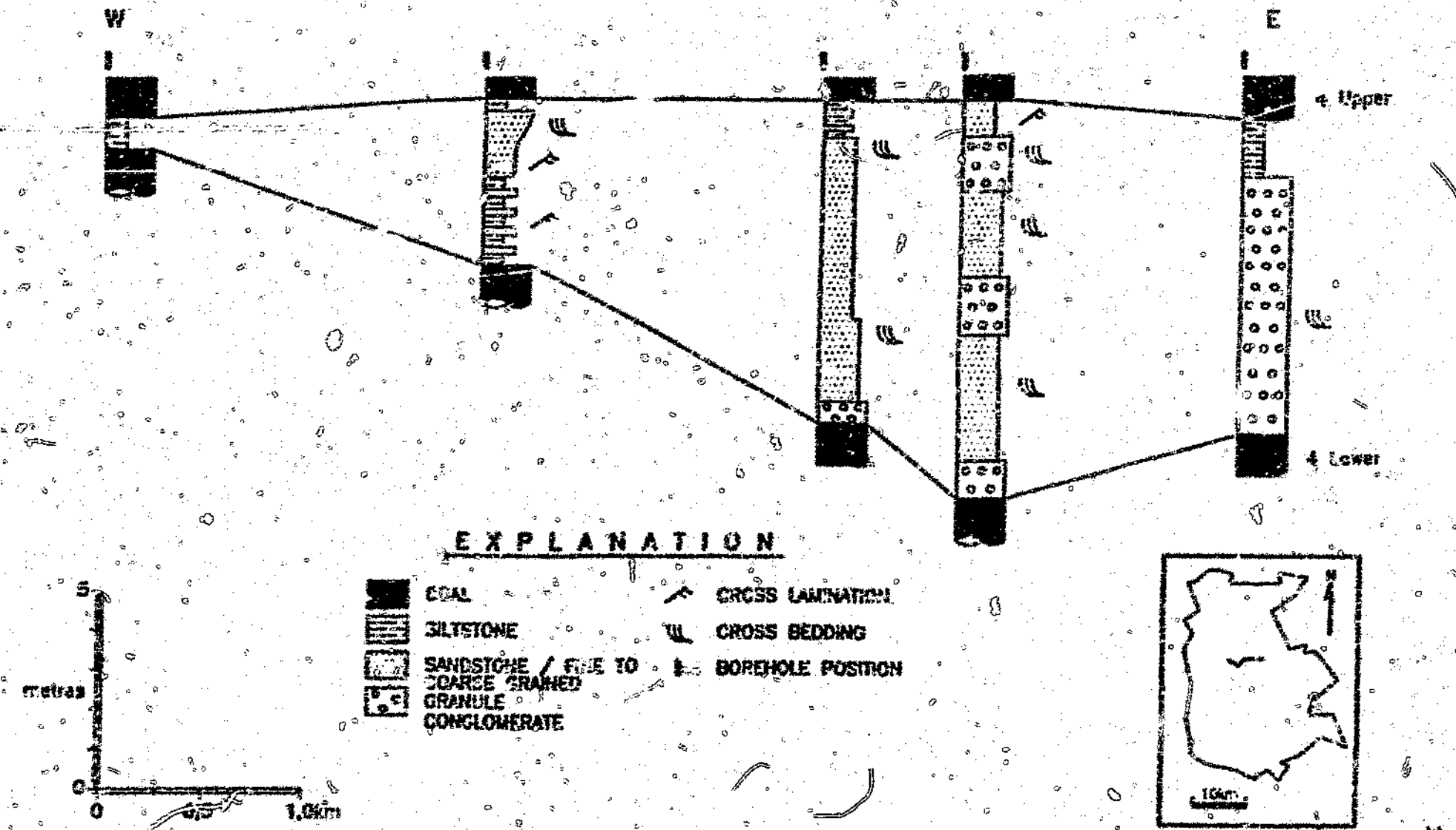


Figure 4.10. Cross-section of the No. 4 Lower-No. 4 Upper seam sedimentary interval. The cross-section illustrates the east-west variation in sedimentary styles from, in-channel, through channel-margin, to flood-basin subenvironments.

4.4 FACIES INTERPRETED AS DELTAIC DEPOSITS

4.4.1 Introduction

Three sedimentary intervals within the Vryheid Formation contain sedimentary sequences interpreted as deltaic deposits. These are the interval between the No. 2 and No. 3 coal seams, and the No. 5 seam and No. 6 seam depositional sequences. The sedimentary facies and facies sequences between the No. 2-No. 3 seam interval, and the No. 5 and No. 6 seam depositional sequences are sufficiently different that they are described separately.

4.4.2 Description: No. 2 to No. 3 Seam Sedimentary Interval

The No. 2 to No. 3 seam sedimentary interval is distinguished from the fluvial depositional sequences by the finer grain size, presence of bioturbation and the dominance of an upward-coarsening facies association. Due to the lateral impersistence of the No. 3 seam, this sedimentary interval is terminated either, by a carbonaceous siltstone or, the No. 4 coal seam (Fig. 2.14B). The interval comprises siltstone and sandstone lithologies which thin against basement highs to the west and thicken to the east and south in the form of two lobes (Fig. 2.18). These two lobes have a maximum thickness of up to 35m.

Cross-sections through the sequence (Figs. 2.16 and 2.17) illustrate three broad facies associations. These facies associations are also portrayed in Fig. 2.14 which illustrates the variation in the stratigraphic sequence for this sedimentary interval. The three facies associations are:

a. a single coarsening-upward facies association,

b. two, and occasionally three, coarsening-upward facies associations and,

c. a coarsening-upward facies association overlain by a fining-upward facies association.

The single coarsening-upward facies association, comprises a 3m to 5m thick sequence of carbonaceous siltstone facies or bioturbated siltstone facies which grades upwards through either, the flaser-laminated siltstone facies or, the interlaminated sandstone-siltstone facies. The remainder of the interval which varies from 5 to 15m in thickness, is a sandstone-dominated facies sequence. The sandstone-dominated facies sequence commences with, and is characteristically structured by, the wave-rippled sandstone facies which is frequently bioturbated. Towards the top of the sequence the wave-rippled sandstone facies is interbedded with current-ripple cross-laminated sandstone, planar bedded sandstone and occasional trough cross-bedded sandstone. In general, the sandstone grain size coarsens upward from very fine to fine grained. However, towards the top of the sequence the sandstone is medium grained when associated with the cross-laminated, planar bedded and trough cross-bedded sandstone facies.

The interval comprising between two and three coarsening-upward sequences has very similar facies assemblages to the single coarsening-upward sequence. Individual coarsening-upward sequences vary between 3 and 10m in thickness. These sequences are dominated by the carbonaceous siltstone, bioturbated siltstone, wave-rippled sandstone and bioturbated sandstone facies. Figure 4.11 illustrates a core sequence through two coarsening-upward facies associations. The base of the facies associations commence with carbonaceous siltstone and carbonaceous siltstone respectively. The sandstone, which abruptly overlies the siltstone, displays well defined wave-ripple lamination. The sandstones which overlie the carbonaceous siltstone of the second coarsening-upward sequence are extensively bioturbated, and both *Skolithos* and *Siphonichnus* traces are present. The No. 3 coal seam caps this sequence.

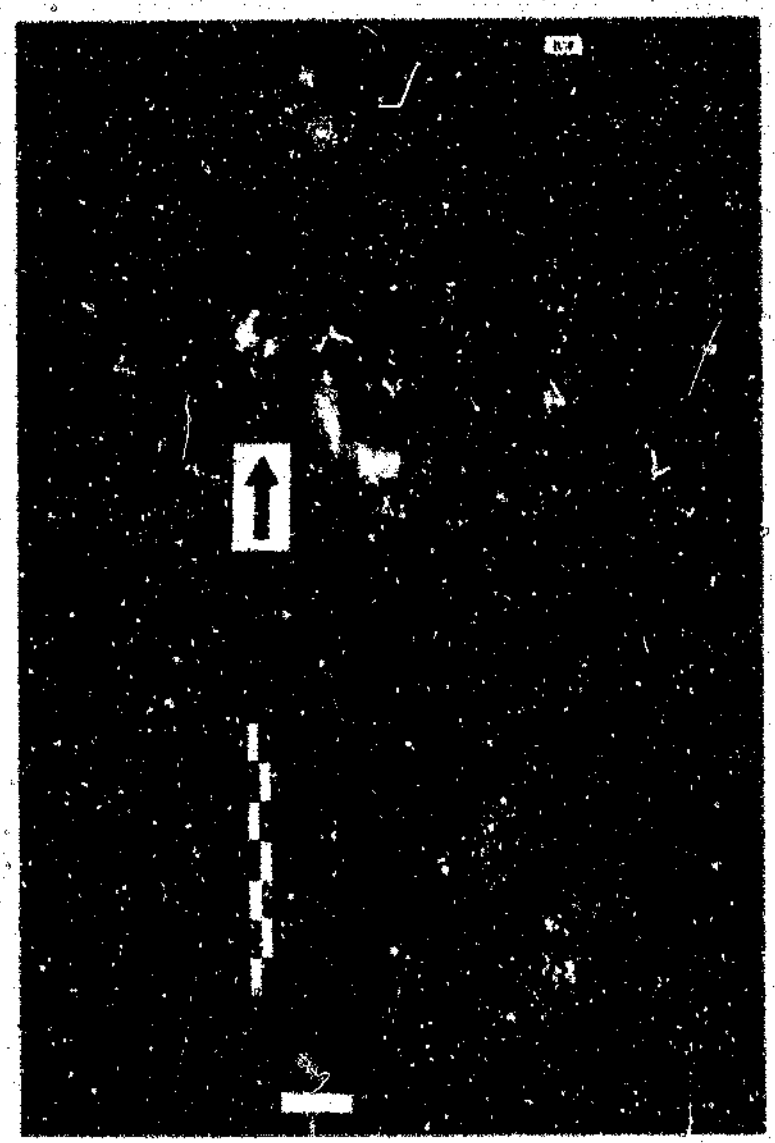


Figure 4.11. Borehole core of the No. 3 seam depositional sequence. The sequence comprises two coarsening-upward facies associations commencing with bioturbated siltstone and terminating with wave-ripple laminated sandstone.

The third facies assemblage comprises an incomplete coarsening-upward facies association terminated by a fining-upward facies association. The coarsening-upward facies sequence commences with the bioturbated and/or the carbonaceous siltstone facies which is overlain by wave-rippie laminated or current-ripple laminated very fine- to fine-grained sandstone. This coarsening-upward facies sequence is overlain by an erosively based small-pebble or granule conglomerate which defines the base of the fining-upward facies sequence (Figs. 2.16 and 2.17). The conglomerate is either, massive to poorly stratified or, planar cross-bedded. The conglomerate often contains intraclasts of carbonaceous siltstone and is overlain by medium- to fine-grained trough cross-bedded sandstone. These basal facies grade up into fine-grained, cross-laminated, sandstone which is infrequently bioturbated. Either, the carbonaceous siltstone facies or, the coal of the No. 3 or No. 4 seam cap the interval.

In summary, the No. 2 to No. 3 seam sequence is characterised by a coarsening-upward facies association dominated by bioturbated siltstone and cross-laminated sandstone which comprise, between one and three, coarsening-upward facies sequences. Infrequently, the coarsening-upward facies sequences are overlain by an erosively based, fining-upward facies association 5m to 10m thick. The upward-fining facies association is structured by cross-bedded conglomerate and sandstone and terminated by cross-laminated siltstone and coal.

4.4.3 Description: No. 5 and No. 6 Seam Depositional Sequences

The No. 5 and No. 6 seam depositional sequences, like the No. 2 to No. 3 seam depositional sequence, are distinguished by a well developed coarsening-upward facies association. These facies associations overlie either, a coal seam or, a thin laterally

persistent sandstone at the base of the depositional sequences. The No. 5 seam depositional sequence is terminated by the No. 5 coal seam, and the No. 6 seam depositional sequence by the No. 6 seam or a carbonaceous siltstone.

The geometry of the No. 5 seam depositional sequence illustrates that the sequence thins to <5m in the west over basement highs and thickens to >35m in the east and southeast (Fig. 2.29). The facies associations interpreted as deltaic deposits comprise, one and sometimes two, coarsening-upward sequences. Mapping of the siltstone and sandstone facies associations reveals the anatomy of this deposit in greater detail. The isopachs of the siltstone lithologies illustrate that the greater part of the study area is overlain by 5m-10m thick siltstone (Fig. 2.30). Only in the south does the siltstone attain a thickness of over 10m. Furthermore, no marked thinning of this sedimentary unit takes place towards the east. The isopach map of the sandstone facies illustrates the thinning of the unit to <5m in the west and thickening to >20m in the east and south of the study area (Fig. 2.31). Over much of the study area the sandstone facies are between 5 and 10m thick.

Cross-sections through this interval reveal that generally, the sequence comprises one coarsening-upward facies association (Figs. 2.27 and 2.28). Only in the south, where the interval thickness exceeds 25m, and in areas in the north and west, where the interval thickness is less than 15m, does this interval comprise two coarsening-upward facies associations.

The facies that comprise the coarsening-upward facies associations of the No. 5 and No. 6 seam depositional sequences are described in terms of two end members; a siltstone end member and a sandstone end member. The siltstone end member comprises at the base the carbonaceous siltstone facies which is frequently overlain by the lenticular-laminated siltstone facies and the interlaminated sandstone-siltstone facies. Any of these three facies types can display bioturbation. However, the intensity

of bioturbation is far less in these facies than in the basal facies of the No. 2 to No. 3 seam interval. The sandstone facies types gradationally overlie the siltstone facies types and comprise fine-grained, cross-laminated, sandstone facies displaying both wave- and current-ripple lamination. Towards the top of the sequence the sandstone becomes medium to very coarse grained and is structured by trough and planar cross-bedding with sets displaying a maximum thickness of 50cm. The trough cross-bedded sandstones are erosively based and stacked sets up to 2m thick are present. Infrequently, these trough cross-bedded sandstones are interbedded with coarse-grained planar cross-bedded sandstones. The uppermost 1,5m of the sandstone end member comprises fine-grained, cross-laminated, sandstone and siltstone. The sequences are terminated by the No. 5 and No. 6 coal seams.

Figure 4.12 is a core sequence illustrating the No. 5 seam coarsening-upward facies association. The sequence displays a facies association comprising carbonaceous siltstone, lenticular-laminated siltstone, interlaminated siltstone and sandstone, cross-laminated sandstone, and cross-bedded sandstone. The facies show slight to moderate bioturbation and the alternating siltstone and sandstone lithologies display soft-sediment deformation structures. Also of note are the thin siltstone beds 1-10cm thick interbedded with the sandstone facies types. These siltstone beds define large-scale foresets which are absent in the deposits of the No. 2 to No. 3 seam interval. The sandstone-dominated foresets are compound cross-stratified and internal foreset stratification comprises cross-lamination and cross-bedding. Individual foresets are under- and overlain by thin, bioturbated siltstone beds. Figure 4.13 illustrates a high-wall section through this sedimentary sequence at the opencast Rietspruit Coal Mine. The foresets, which are well defined by the siltstone beds, dip at 22° at the top of the foresets and flatten out asymptotically at the base of the foresets. Moreover, the foresets defined by the siltstone beds, display a wedge-shaped geometry and variable

Figure 4.12. Borehole core of the No. 5 seam depositional sequence. The sequence displays a classic coarsening-upward facies association. The cross-bedded coarse-grained sandstone distinguishes this sequence from the No. 3 seam sequence.

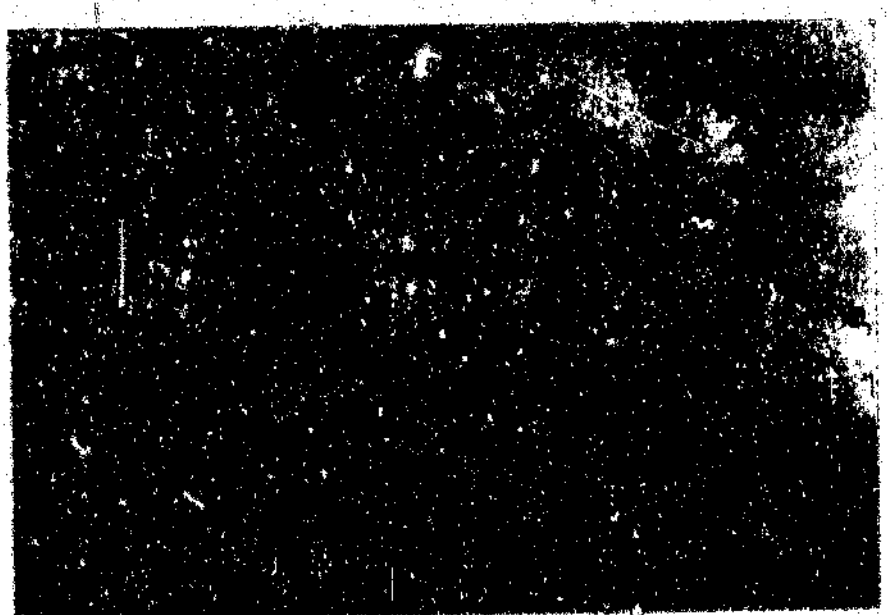


Figure 4.13. A highwall exposure of the No. 5 seam depositional sequence at Rietspruit Coal Mine. Note the tops of the compound foresets are steeply dipping. The foresets flatten out asymptotically in a basinward direction. Also illustrated is the coarsening-upward facies association of this depositional sequence.

facies types. At the top of the sequence the foresets are 40cm-1,5m thick and are dominated by sandstone facies types. The foresets thin and wedge out into siltstone-dominated facies types at the base of the sequence.

Figure 4.14 is a core sequence through the sandstone-dominated facies of the No. 6 seam depositional sequence. This core sequence compliments Fig. 4.13 as it illustrates the sedimentary structures of this sandstone cannot be seen in Fig 4.13. The sandstone is medium to coarse grained and displays both current- and wave-ripple lamination, and trough and planar cross-bedding. The cross-laminated facies are the dominant facies types within the lower half of the core sequence, and the cross-bedded facies the dominant facies types within the upper half of the core sequence. The sandstone is cleaner within the uppermost 2m, owing to the decrease in the number of siltstone laminae.

4.4.4 Interpretation of Deltaic Deposits

Introduction:

The coarsening-upward facies associations deposited within the No. 2-No. 3 seam interval and the No. 5 and No. 6 seam depositional sequences, are interpreted as sediments deposited by deltaic depositional systems. The concept of a delta goes back to 400 B.C. when Herodotus used the term to describe the triangular piece of ground between the branching channels of the Nile River in Egypt. Ever since this time, the term has been used to describe similar geographic features. Definitions of deltas do not differ substantially. Barrell (1912) defined a delta as "a deposit, partly subaerial, built by a river into or against a permanent body of water". As suggested by Moore and Asquith (1971), the entire subaerial and subaqueous contiguous sediment mass is included in the delta system. A more recent definition of deltas, which follows on the

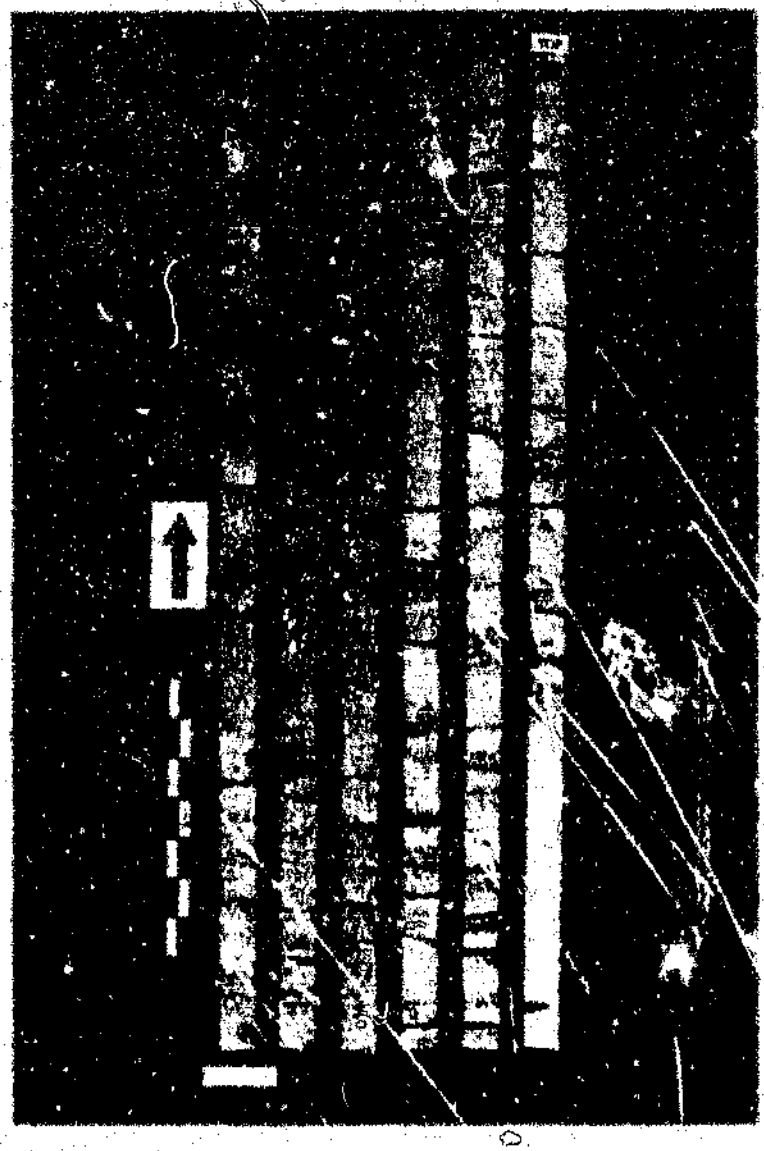


Figure 4.14. Borehole core illustrating the sandstone-dominated coarsening-upward facies association of the No. 6 seam depositional sequence. Note, cross-bedding is the dominant sedimentary structure.

numerous studies of the deltas in the 1960's and 1970's, is presented by Elliott (1986). Deltas are defined by Elliot (1986) as; " discrete shoreline protuberances formed where rivers enter oceans, lakes or lagoons and supply sediment more rapidly than can be distributed by basinal processes".

The appreciation during the 1980's of feeders to the deltas being fairly variable has led to the delta definition becoming more general. For example, the simple river delta has expanded to include the braidplain delta (McPherson *et al.*, 1987), the fan delta (Holmes, 1965) and the slope-apron delta (Busby-Spera, 1988). Moreover, non-alluvial deltas such as the lava delta and the pyroclastic delta (Holmes, 1965) are incorporated into delta terminology. Nemec (1990) appreciating the variability of delta types has defined the delta as, "a deposit built by a terrestrial feeder system, typically alluvial, into or against a body of standing water, either a lake or sea. The result is a localised, often irregular progradation of the shoreline controlled directly by the terrestrial feeder system, with possible modification by basinal processes, such as the action of waves or tides".

A conceptual framework for alluvial deltas was presented by Elliott (1986) in order to analyse the variability of modern and ancient deltas. He used the term "delta regime" to define the overall deltaic setting (Fig. 4.15). The delta regime is influenced by two major variables, the hinterland and receiving basin characteristics. These two variables are in turn controlled by climate, tectonics, subsidence and topography. The hinterland characteristics are reflected in the fluvial regime and the transported sediment load which are contested at the mouths of alluvial systems by the basinal regime. The basinal regime is defined by the shape, size, bathymetry, and dynamics of the basin. The interaction between the fluvial and basinal regimes controls the dispersal and deposition of the sediment defining the delta regime, and the vertical and lateral

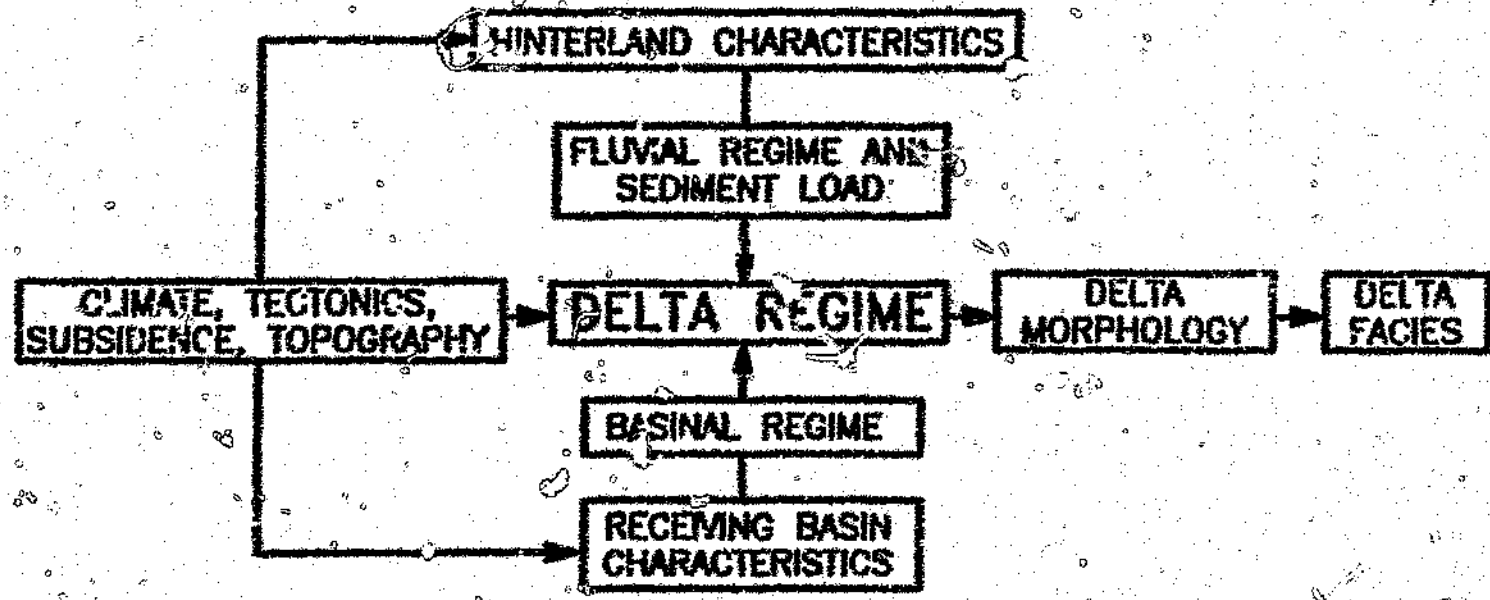


Figure 4.15. Conceptual framework for a comparative study of deltas (after Elliot, 1986).

distribution of deltaic facies. This study intends to demonstrate that the change in fluvial and basinal regime through time accounts for the variability of the deltaic facies and delta types present in the study area.

Delta classification:

The classification of deltas has been largely influenced by the study of large modern river systems. The link between processes and delta morphology provided a major break through in the understanding of deltaic systems. Fisher *et al.*, (1969) distinguished high-constructive deltas dominated by fluvial processes, from high-destructive deltas dominated by basinal processes. Lobate and birdfoot delta morphologies comprised the high-constructive deltas and the high-destructive deltas were subdivided into wave- and tide-dominated deltas. Coleman and Wright (1975) based upon the study of 34 present-day deltas classified deltas into six delta types based upon net sand distribution patterns in modern deltas. These delta types are described in terms of the dominant fluvial and basinal processes and morphologies. This classification scheme suffers certain weaknesses when applied to ancient deltas as firstly, ancient deltas are longer lived than their modern counterparts, consequently geometries may not be comparable and secondly, spatial thickness patterns also relate to differential subsidence and basin-floor topography as is evident in this study. Galloway (1975) produced a ternary diagram in which he built on the Fisher *et al.*, (1969) classification. The Galloway classification comprises three end members, fluvially dominated deltas, wave-dominated deltas and tide-dominated deltas and also caters for the full range of "mixed-type" deltas. This classification scheme plots deltas qualitatively and not quantitatively. This scheme is genetic rather than descriptive and relies on the skills of the researcher to interpret a deltaic deposit as fluvial rather than tide or wave dominated.

Orton (1988) recognised that the Galloway ternary diagram did not take grain size into account and his classification scheme adds a fourth dimension which includes and accommodates coarse-grained delta systems. This classification scheme however suffers the same qualitative and genetic weaknesses as the Galloway model. Two of the more recent classifications are presented by Nemec (1990) and Postma (1990). Nemec (1990) refers to Corner's classification (pers. comm.) in which all deltas are defined in terms of two variables, grain size and delta-face slope. Grain size is subdivided into mud-, sand- and gravel-dominated deltas, and delta-face slope into gentle or steep. The advantage of this classification scheme is the descriptive nature of the variables used and ease of identification in the field. Steep-sloped delta systems are identified by "megaforeset" cross-bedding, and mouth-bar dominated delta systems by micro-delta type cross-stratification. A weakness of the Nemec classification scheme is the inability to distinguish "megaforeset" cross-bedding using subsurface data such as borehole cores. The appreciation of the "megaforeset" cross-bedding in the No. 5 seam delta sequence was only recognised from highwall exposures of the sequence and not through the study of subsurface cores. The final classification scheme reviewed is that proposed by Postma (1990) for deltas prograding into low-energy, wave- and tide-dominated basins. This classification scheme also defines deltas in terms of two variables namely, feeder system type, and water depth. Deltas are defined in terms of shallow- or deep-water deltas and in terms of four feeder-system types. The feeder-systems types are defined as;

a. gravel-dominated alluvial systems with a delta-plain slope of greater than a few degrees,

b. gravel-dominated alluvial systems with numerous closely spaced bed-load channels with a delta plain having a steep gradient (+/- 0.4°),

c. gravel- and sand-dominated alluvial systems having closely spaced, relatively stable, channels and a moderate gradient, and

d. suspended-load dominated alluvial systems having widely spaced feeder channels of low gradient.

Twelve conceptual end members are constructed for the categorisation of deltas prograding into low-energy basins (Fig. 4.16). This classification scheme identifies the complexity of alluvial delta systems however, the draw-back of the system for the non-sedimentologist is that it is genetic in its approach as workers will have to determine both the feeder system type and the subtleties of delta-slope sediments in order to utilize this classification system.

Deltaic processes:

The processes operating on delta systems have not received the same amount of attention in the 1980's as the quest for expanding the variety of delta types. The Bates theory of jet flow has dominated the thinking of channel-mouth deposition since 1953. Bates (1953) contrasts situations in which the river water is more dense, equally dense and less dense than the basin water. These conditions of flow are termed hyperpycnal, homopycnal and hypopycnal respectively. Hyperpycnal flow causes the incoming dense water to flow beneath the basinal water as density currents, transporting sediment into the delta front, prodelta, and shelf or slope environments. Homopycnal flow results in the rapid three-dimensional mixing of the river and basinal waters, causing appreciable deposition at the river mouth. Hypopycnal flow, where the outflow is less dense than the basinal water, causes the incoming water to flow as a buoyancy-supported surface jet or plume over the basinal water. Wright (1977) in a comprehensive review of delta-mouth processes has built upon these early theories. Inertia-, friction-, and buoyancy-dominated river-mouth processes are considered in terms of outflow dispersion models. Inertial forces dominate where high-velocity bed-load rivers enter a fresh-water basin. Under these conditions, effluent spreading is minimal and rapid dumping of sediment takes place. Elliot (1986) considers that under conditions of sediment dispersal by turbulent jet under homopycnal flow elongate,

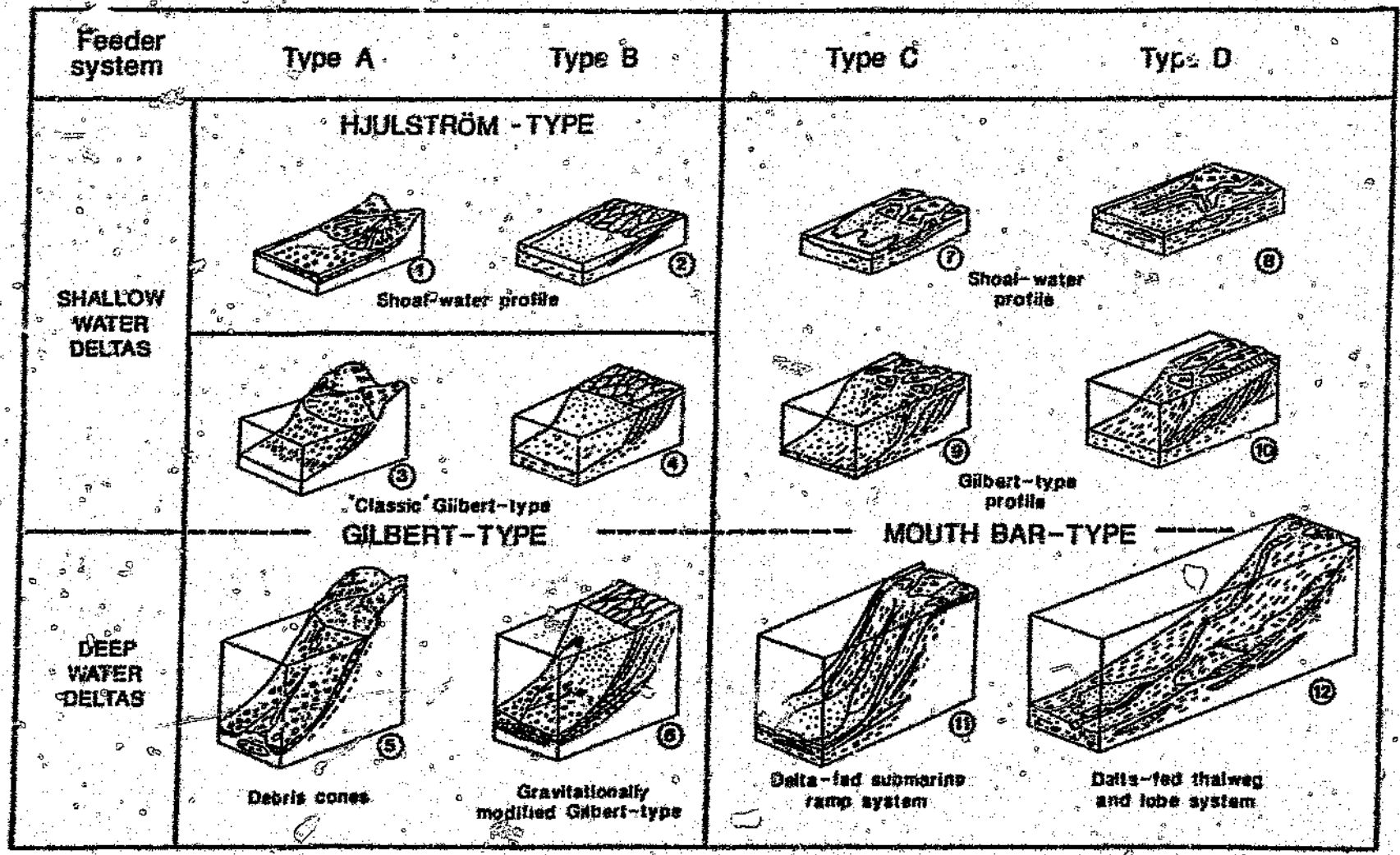


Figure 4.16. Classification of deltas prograding into low-energy basins (after Postma, 1990).

steep-fronted, Gilbert-type mouth bars will form. In contrast to this view, Postma (1990) suggests homopycnal conditions are less important than inertial forces at river mouths in forming Gilbert-type deltas, as these steep-fronted deltas occur in both fresh- and salt-water basins.

Under conditions of hypopycnal flow, where there are low offshore slopes, friction between the effluent water mass and the bottom sediments causes rapid deceleration and spreading of the outflow. Under these conditions, a triangular middle ground is formed which causes channel bifurcation and shifting. As progradation continues, new bars are formed at the mouths of the bifurcated channels and, in this manner, the delta advances into the basin. This is a common form of advancement of the Mississippi Delta crevasse splays. Buoyancy-dominated river mouths are favoured by fresh-water river systems entering marine basins. A salt-water wedge intrudes into the river mouth reducing the effect of frictional forces. This condition is optimised where there are deep distributary channels and deep water fronting the delta which reduce frictional forces. In this situation, coarse bed-load sediment is deposited at the river mouth and the finer material is transported in suspension further into the basin. Buoyancy-dominated river mouths produce a narrow elongate mouth bar as exemplified by the Mississippi Delta bar-finger sands. River-mouth processes of buoyancy, inertia, and frictional forces, seldom operate individually and all three forces may vary in importance due to fluctuating discharge and basinal conditions. During high-discharge periods, the river mouth is influenced more through inertia and frictional forces. During low-discharge periods buoyancy processes dominate. Both wave and tidal processes operating in the delta-front environment enhance the mixing of river and basinal waters promoting sediment deposition of the coarse sediment load at river mouths and the dispersion of finer grained material offshore. These processes therefore enhance the coarsening-upward grain size profile characteristic of delta depositional systems.

Discussion:

Differences in the two- and three-dimensional geometries, facies types and facies associations, between the No. 2 to No. 3 seam interval and No. 5 and No. 6 seam depositional sequences, are interpreted in terms of temporal variations in deltaic sedimentation.

The sediments of the No. 2 to No. 3 seam interval are interpreted as river-dominated, shoal-water deltas which built out into a basinal regime characterised by low wave energy. The carbonaceous siltstone, and bioturbated siltstone facies at the base of the sequence are interpreted as shelf and prodelta deposits. The interlaminated sandstone-siltstone facies represent basal delta-front deposits while the wave-ripple laminated and planar bedded sandstone facies are sediments deposited in the mouth-bar environment. The sedimentary interval mapped out is between 10m and 20m thick and the prodelta deposits are between 3 and 5m thick. These sedimentary thicknesses imply that the deltas are of the shoal water variety. Furthermore, the fact that the sequence comprises two coarsening-upward facies associations each <10m thick gives further support to the contention that the deltas built out into a shallow-water basin with a water depth of about 10m. Isopachs of the No. 2 to No. 3 seam interval illustrate a general thickening of the sequence to the east over a area of about 1,000km² (Fig. 2.18). The isopachs of this delta sequence, taking into account the thinning of the sediments against the basement highs in the west, suggest that the deltas have a lobate geometry.

The dominance of wave-ripple lamination over current-ripple lamination and planar bedding within the mouth-bar sandstones, indicates reworking of the mouth-bar environment by low-energy oscillatory waves. The waves would only have had the strength to rework the sands into less than 5cm high bedforms to produce this stratification. The current-rippled sandstones represent either, preservation of

unidirectional current flow by fluvial currents during flooding of the river mouths or, the development of shallow-water basinal currents. The presence of planar bedding towards the top of the mouth-bar sandstones suggests upper flow regime conditions were experienced periodically. These conditions could eventuate through the;

a. shoaling of water depth during tidal fluctuations, producing swash stratification at the tops of the mouth bar or,

b. impingement of high velocity storm-dominated currents on the mouth bar which reworked wave-rippled sands into planar bedded sands.

The abundance of bioturbation within both the pro-delta and delta-front siltstones and sandstones, suggests the deltas were fed nutrient-rich waters by updip fluvial systems which deposited their sediment load into a brackish or marine body of water. Similar bioturbated sediments have been recorded from the delta-front environments of the Mississippi Delta by Coleman and Gagliano (1965) and Coleman (1976).

Highwall exposures of this sequence reveal no megaforeset development of the delta front, consequently, this delta sequence is classified as a shallow-water, mouth-bar, delta as opposed to a "Gilbertian" shallow-water delta (Postma, 1990). Jopling (1965) modelled the development of foresets in a number of flume experiments. He concluded that, if the ratio of stream depth to basin depth is high and a significant concentration of suspended sediment is present in the river system, a low angle to the foreset slope is developed. These conditions would promote the development of a mouth-bar delta. When stream depth to basin depth ratios are 0,4 or greater, the foreset slope becomes tangential which is promoted by the deposition of fine-grained sediment at the bottomset. The channel-fill sequences, which are illustrated in Figs. 2.16B, C and 2.17, show the channels to comprise between half and three quarters the thickness of coarsening-upward sequences. This would suggest that the channel depth to basin depth ratios were of the order of 0,5 to 0,75.

The shallow depth of the basin, (minimum depth of 10m), the fine-grained sediment fill, the geometry of the delta sequence, and the abundance of wave-ripple lamination implies that frictional forces, buoyancy forces, and low energy wave action, promoted the formation of shallow-water, mouth-bar deltas. The deltas have a lobate geometry, with mouth-bar sands incised through by a number of distributary channels with depth ratios of >0,5 (Jopling, 1965). The deltas were subjected to low-energy wave reworking which spread the sediment load laterally. The lateral spreading of the sediment load was also enhanced by the dominance of frictional forces operating at the river mouths. This caused the deposition of sand to form "middle-ground" bars which enhanced channel bifurcation and spreading. The abundant bioturbation of both the siltstone and sandstone dominated facies suggest the sediments were deposited into a low-energy, brackish to marine basin.

The fact that the No. 2 to No. 3 seam interval often comprises two and sometimes three upward-coarsening sequences suggests that differential compaction of underlying sediments resulted in delta abandonment, subsidence, and renewed delta outbuilding. The differential compaction of the sediments was possibly promoted through the compaction of sediments during the early stages of sediment fill of the basin, coupled with the irregularities of the basement floor.

Smith *et al.*, (1989) comment upon the fact that within fluvial systems, large areas of sediment-fill result from avulsion of the fluvial system and the development of extensive crevasse splays which can cover over 500km². In addition, they make the inference that these crevasse-splay deposits could extend over several thousand km² and that their preservation potential is high. Within the Witbank and Highveld Coalfields (Fig. 1.2), the No. 2 to No. 3 seam deltaic sequence has been mapped out in detail over an area of approximately 4,000km² and the sequence is known from borehole drilling to extend over 10,000km². Moreover, Cadle (1982b) has correlated this deltaic interval into the Eastern Transvaal and Natal Coalfields which, if the correlations are

correct, would expand the area of deltaic deposition to about 30,000km². Although acknowledging that crevasse splays have similar thicknesses and vertical profiles as the No. 2 to No. 3 seam deltaic deposits (Coleman and Gagliano, 1964; Elliot, 1974), the great lateral extent of the unit throughout the Witbank and Highveld Coalfield and its possible extension into the Eastern Transvaal and Natal Coalfields, suggests deposition by delta progradation. The deltas are classified as shallow-water, mouth-bar deltas which were dominated by frictional and buoyant forces at the river mouth. The sediments were modified and reworked by low-energy wave action in a brackish to marine basin.

The No. 5 and No. 6 seam depositional sequences are interpreted as a shallow-water fluvially dominated "Gilbertian" deltas. These deltas were subjected to periodic high-discharge events, which fed sediment to the delta front through closely spaced braided channels. The lateral and vertical facies associations of these depositional sequences differ from the No. 3 seam deltas in terms of, the abundance and coarseness of facies types. The carbonaceous-siltstone and bioturbated-siltstone facies are interpreted as prodelta or shelf deposits (Gilbertian bottomsets). The interlaminated siltstone-sandstone facies, including the lenticular-laminated and wavy-laminated facies, displaying soft sediment deformation structures, represent sediment deposited at the base of the delta front. The current-ripple laminated, wave-ripple laminated and trough cross-bedded sandstones overlain by 2-5cm thick, bioturbated siltstone beds represent proximal delta front sedimentation (Gilbertian foresets). The Gilbertian topset deposits comprise the erosively based, interbedded trough and planar cross-bedded sandstones which are generally coarse to very coarse grained. The topsets fine upward into fine-grained, cross-laminated sandstones and siltstones. These deltaic deposits are terminated by the carbonaceous-siltstone facies or the No. 5 or No. 6 coal seam.

Detailed isopach mapping of the siltstone- and sandstone-dominated lithologies for the No. 5 seam depositional sequence, illustrates a uniform distribution in thickness, of between 5 and 10m, for both the siltstone and sandstone lithologies (Figs. 2.30 and 2.31). The sandstone thickens up to 20m towards the east and southeast illustrating a lobate geometry for the deltaic sandstones. The thickness of the No. 5 seam depositional sequence is between 20m and 30m over much of the study area, with the exception of the south where the interval thickens to over 40m. Cross-sections through the deposit illustrate that one coarsening-upward sequence is present over most of the study area except where the interval thickens in the southeast (Figs. 2.27 and 2.28). Here, three coarsening-upward sequences, about 10m thick, occur. Cross-sections of the No. 6 seam depositional sequence also illustrate coarsening-upward sequences up to 30m in thickness (Figs. 2.27 and 2.28). The inference drawn from these data is that the minimum water depth was probably between 20 and 30m. This water depth is significantly deeper than the minimum water depth of 10m, postulated for the basin during deposition of the No. 3 seam delta sequence.

The dominance of wave-ripple laminated fine- to medium-grained sandstones and siltstones that constitute the delta front suggest that a low-energy basinal regime was in existence during the deposition of the No. 5 and No. 6 seam delta sequences. However, the presence of trough and planar cross-bedded coarse- to very coarse-grained sandstones at the top of the delta sequence, suggest that sandy alluvial systems with closely spaced braided channels fed the delta front. The absence of mature arenites with well rounded grains, lenticular-laminated siltstones and sandstones, and siltstones displaying wavy bedding, implies that reworking of the delta by tidal process was not excessive. Also the infrequent occurrence of planar laminated sandstones would suggest that little reworking took place of the upper part of the delta front by wave action. The trough and planar cross-bedded sandstones are thus interpreted as the movement of

dunes and sandwaves in braided channels on the delta plain. The sediments of the delta front and delta plain imply a fluvially dominated delta system existed during the deposition of the No. 5 and No. 6 seam depositional sequences.

The presence of Gilbertian foresets in these delta sequences and their mechanism of formation are discussed. Axelsson (1967) in his paper on the Laitaure Delta puts forward the following factors which promote Gilbert-type delta profiles in shallow water:

a. the transport of sufficient quantities of bed-load material as far as the river mouth,
b. sufficiently large water depths immediately basinward of the river mouth and,
c. spreading of the effluent as an axial turbulent jet.

The presence of coarse-grained, trough and planar cross-bedded sandstones, interpreted as sediments deposited by braided distributaries, indicate that large quantities of bed-load material was transported to distributary mouths. The minimum basinal water depths are postulated to range between 20m and 30m. These water depths are two to three times the postulated water depths for No. 3 seam delta deposition. Conditions for axial turbulent jet spreading are enhanced where there are high outflow velocities through distributary channels, a sufficient depth of water, and small density contrasts between incoming and basinal waters. The fact that the delta foresets are draped by thin centimetre-thick siltstone laminae would indicate that the sand was deposited on the foresets during periods of high discharge favouring inertia-dominated deposition. The silt was deposited on the delta front during periods of low discharge when buoyancy-dominated conditions prevailed. The dominance of silt and mud deposition on the delta front of the Atchafalaya Delta during low discharge periods is well documented by van Heerden (1982). The Bates theory of homopycnal flow as a dominant causal mechanism for Gilbert-type delta profiles is questioned by Postma (1990), as Gilbertian deltas are present in both fresh- and salt-water basins.

The ratio of the thickness siltstone to sandstone in the No. 5 seam deltaic deposits can be estimated from the isopach maps of the No. 5 seam deltaic sediments. A thickness of 25m for the sandstones and 7,5m for siltstone implies a sandstone to siltstone ratio of 3,3 or a sandstone to sandstone-siltstone ratio of 0,77. It is tempting to use the sandstone to siltstone ratio of 3,3 as a crude maximum of the bed-load to suspended-load ratio and the ratio of 0,77 as a bed-load to total-load ratio. However, these ratios would be smaller assuming that the very fine-grained sandstone represents suspended-load material and that certain of the clay- and silt-sized material may well have been flushed a considerable distance away from distributary channel mouths by basinal processes. Bogen (1983) calculated the bed-load to total-load ratio of fjord deltas and compared these data to the delta foreset slopes. Deltas with a bed-load to total-load ratios of >0,4 and shallow initial basin depths of <30m have a classical Gilbert shape to the delta. These ratios and depths are supported by this study. The estimated bed-load to suspended-load ratio is 0,77 and basin depths are estimated between 20m and 30m.

Jopling (1965) shows from laboratory studies that where the channel depth to basin depth ratio is low <0,4 the foreset slope is steep and there is an angular contact between the foreset and bottomset. Based upon a channel depth of 5m for the braided distributaries and a water depth of 20m a depth ratio of 0,25 is calculated which supports the experimental data of Jopling (1965). The foreset geometry of deltas is also enhanced where the bed-load to suspended-load ratio is high and the suspended load constitutes less than 15% of the total load (Jopling 1965). Based upon thickness data of siltstone and sandstone within the No. 5 seam sequence and interpreting these sediments as representing the suspended load and bed load respectively, a suspended-load to total-load ratio of 23% is calculated. The true ratio would be less than 23% as it does not take into account the amount of sandstone in suspension and

the amount of fines flushed a great distance away from the river mouths. These data suggest that the bed-load to suspended-load ratio may approach 15% which Jopling implies promotes the development of foresets.

The upward-coarsening facies associations of the No. 5 and No. 6 seam depositional sequences are interpreted as sediments deposited by fluvially dominated Gilbertian deltas. The deltaic distributaries were closely spaced braided distributaries which flushed large quantities of bed-load material during periods of high discharge. During periods of low discharge silty material was flushed out through buoyancy forces onto the delta front and pro-delta environments. The thin siltstone drapes deposited on the delta front defined Gilbertian foresets. These foresets were developed through a combination of factors such as :

a. a high bed-load to total-load ratio (estimated as 0,77),

b. a low channel-depth to basin-depth ratio,

c. a relatively high basinal water depth of between 20-30m,

d. fluctuating discharge which would have promoted inertia-dominated turbulent jets causing rapid deposition of bed-load material.

The lack of bioturbation within the No. 5 seam deltaic deposits, when compared to the No. 3 seam deltaic deposits, suggests rapid rates of deposition which inhibited abundant bioturbation of the sediment. Moreover, the dilution of organic material by inorganic sediment could also cause a lack in bioturbation of this sedimentary interval. Similar reasoning is put forward by Van Straaten (1960) and Oomkens (1970) for the Rhone Delta, and Cotter (1975) for the deltaic deposits of the Upper Cretaceous Ferron Sandstone.

In summary, two different delta types are interpreted. The sediments of the No. 2 to No. 3 seam interval are interpreted as shallow-water, mouth-bar deltas dominated by frictional and buoyant forces at distributary mouths. Significant reworking of the

sediments by low-energy wave action took place. This is evidenced by abundant wave-rippie lamination and bioturbation of the sedimentary interval. The sediments interpreted as deltaic deposits within the No. 5 and No. 6 seam depositional sequences are classified as shallow-water, fluvially dominated "Gilbertian" deltas. These deltas display "megaforesets", are coarse grained, and experienced inertia-dominated turbulent jet discharge at distributary mouths. Differences in water depth, sediment type and river mouth discharge account for the differences in styles of deltaic deposition.

4.5 FACIES INTERPRETED AS TRANSGRESSIVE MARINE DEPOSITS

4.5.1 Introduction

At specific stratigraphic positions within the Vryheid Formation there are facies sequences interpreted as transgressive marine deposits. In each instance, these deposits overlie coal seams, mark the termination of coal accumulation, and separate depositional sequences. These transgressive marine deposits occur above the Nos. 2, 4, 5 and 6 seams. The similarity of facies types occurring at these different stratigraphic positions allows for these facies to be collectively described and interpreted.

4.5.2 Description of Transgressive Marine Deposits

Sediments interpreted as transgressive shallow-marine deposits, although present above the Nos. 2, 4, 5 and No. 6 seams, are well developed above the uppermost No. 4 seam and the No. 5 seam. Three lithofacies associations are documented; a coarsening-upward lithofacies association, a fining-upward lithofacies association and a well sorted sandstone lithofacies association.

The coarsening-upward lithofacies association commences with a grey or brown bioturbated siltstone facies containing disseminated pelletiferous glauconite which grades upward into an interlaminated siltstone-sandstone facies. This facies is overlain by a moderately to well sorted fine- to medium-grained sandstone which displays both current- and wave-ripple lamination. The sandstone is either, white or, light green in colour due to the presence of glauconite. Figure 4.17 illustrates the bioturbated interlaminated sandstone-siltstone facies grading up into the well sorted white sandstone. Carbonaceous siltstone or mudstone terminate the coarsening-upward lithofacies association. The coarsening-upward lithofacies association varies between 2m and 10m in thickness.

The fining-upward lithofacies association comprises at the base of the sequence either, a flat-based, arkosic, granule conglomerate or, a coarse-grained sandstone containing pelletiferous glauconite. These lithofacies vary in thickness between 2cm and 70cm, are weakly bioturbated and display planar bedding. Overlying the granule conglomerate and coarse-grained sandstone is a well sorted, white to light-green, medium-grained sandstone. The sandstone is structured by wave-ripple lamination and trough cross-bedding (Fig. 4.18). The sandstones display both *Skolithos* and *Siphonichnus* burrows. The lithofacies association can vary in thickness between 50cm and 3m.

The third lithofacies association is distinctive in that it comprises a white to light-green (glauconitic), well sorted, medium- to coarse-grained sandstone. This sandstone can vary in thickness between 5m and 15m and is predominantly structured by both wave-ripple lamination and trough cross-bedding. Figure 4.19 is an example of the glauconitic sandstone displaying both wave-ripple lamination and trough cross bedding. Occasionally, the sandstone is structured by planar cross-bedding. Granule conglomerate lags are interbedded within the sandstone. Due to the well sorted nature of the sandstone, it often appears massive. This facies association is particularly evident

Figure 4.17. Core sequence illustrating the coarsening-upward facies association representing transgressive shallow-marine deposits. Note the *Skolithos* and *Siphonichnus* bioturbation of the interlaminated siltstone/sandstone and sandstone facies.

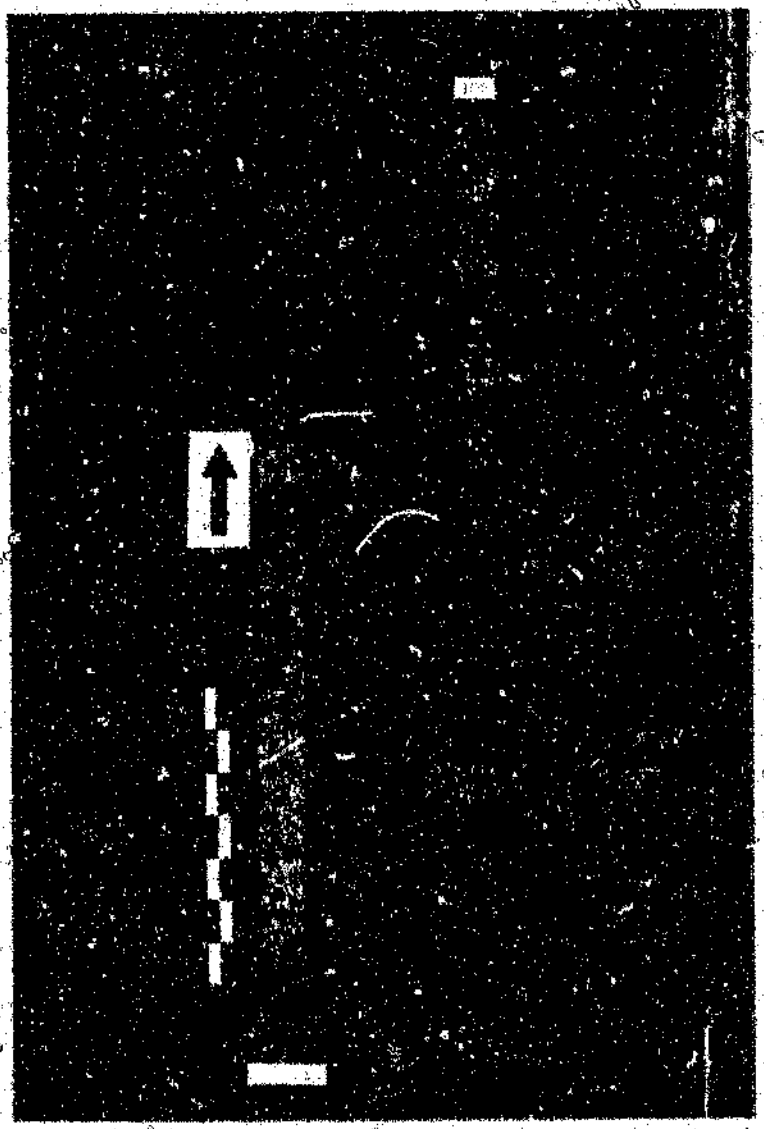


Figure 4.13. Core sequence of the transgressive shallow-marine deposit. The fluvial sequence overlying the coal seam is truncated by a 10cm thick granule conglomerate which in turn grades upwards into wave-ripple and trough cross-bedded glauconitic sandstone.

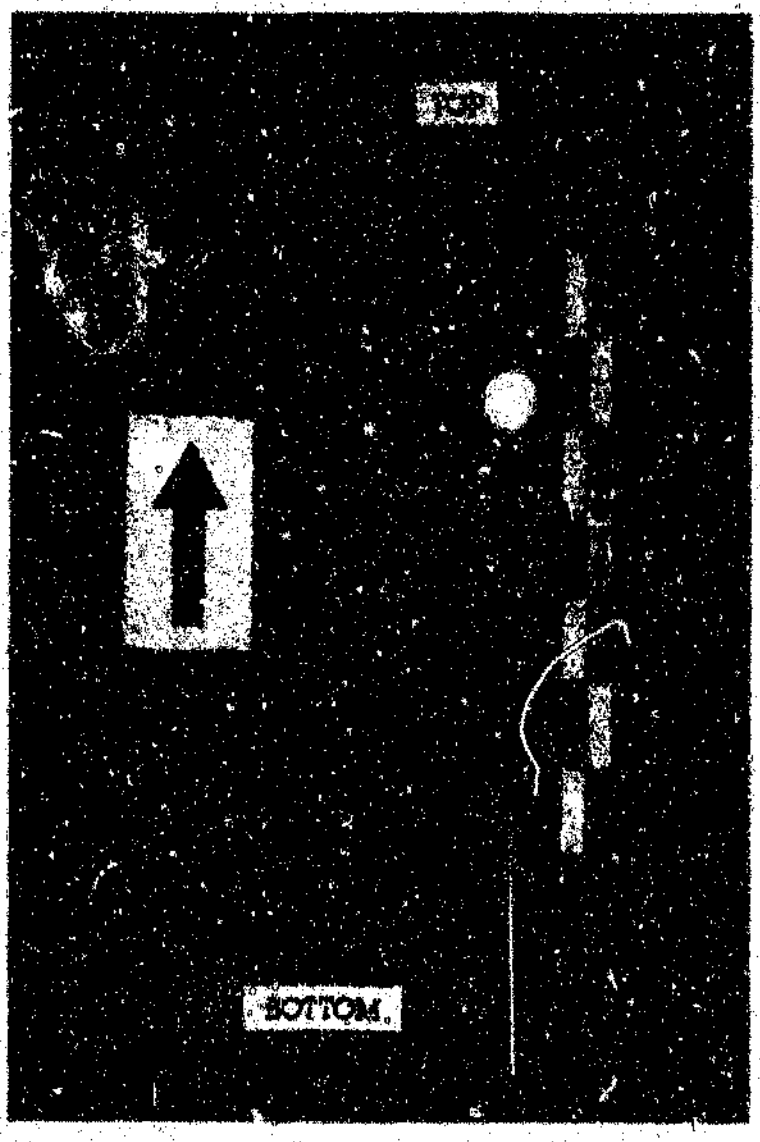


Figure 4.19. Core sequence of the highly glauconitic shallow-marine sandstone. The sandstone is structured by both wave-ripple lamination and trough cross-bedding (top).

above the No. 5 seam in the north and east of the study area (Figs. 2.27 and 2.28). The carbonaceous siltstone facies and thin laterally impersistent limestone facies overlie the sandstone (Figs. 2.26 and 2.28).

4.5.3 Interpretation of Transgressive Marine Deposits

The relatively thin sheet sandstones and siltstones present above the Nos. 2, 4, 5, and 6 coal seams are interpreted as transgressive shallow-marine deposits. These deposits indicate an abrupt change in environments from continental peat swamps (*in situ* coal seams) to shallow shelves (glauconitic siltstones, sandstones and granule conglomerates). In every instance, it is a transgressive event which terminates peat accumulation. These transgressive deposits can be correlated across the Witbank Coalfield (Le Blanc Smith 1980; Winter, 1985). In particular, the No. 4 and No. 5 seam transgressive deposits are correlated from the Natal Coalfields across the Witbank Coalfield to the Orange Free State Coalfields (Cadle *et al.*, 1993) and imply that the transgressive events are regional rather than local in extent.

Transgressive shallow-marine sediments are deposited during transgressive accommodation-dominated regimes (Swift and Thorne, 1991). Accommodation-dominated regimes occur whenever sediment is entering a shelf in lesser quantities than can be accommodated by compaction and sea-level rise or, can be swept away by tidal or storm currents. Vos and Hobday (1977) and Tavener-Smith (1982) have documented microtidal conditions for the deposition of regressive beach and barrier-island deposits situated in the Orange Free State and Natal Coalfields respectively. The predominance or wave-ripple lamination structuring the transgressive deposits, indirectly supports a microtidal regime within which transgression took place. Thus, the transgressive shallow-marine deposits are discussed within the framework of a retreating microtidal coastline.

Microtidal coastlines are distinguished by barrier systems which are long and continuous. Tidal inlets are characterised primarily by flood-tidal deltas located within the lagoon in a back barrier setting. These deltas obtain their sediment supply from sand brought in through the tidal inlets by the littoral drift system. Ebb-tidal deltas are small, or nonexistent, as wave attack redistributes the sediment. Barriers do not remain stationary but migrate landward with relative sea-level rise. Several modes of barrier retreat are documented in the literature. Leatherman (1983) has proposed a barrier roll-over model for barrier retreat whereby sand from the shoreface is welded onto the beach. Sand is then blown from the beach into dunes and then into the back-barrier environment. Storms also transport the sediment landward as washover fans. Furthermore, sediment is transported through tidal inlets and is deposited and stored in flood-tidal deltas. Thus, the barrier rolls over into the back-barrier environment with time. Swift *et al.*, (1991) document that discontinuous barrier retreat took place in the present Middle Atlantic shelf at time scales of several thousand years. Discontinuous barrier retreat is also documented from the Mississippi Delta (Penland *et al.*, 1985). These authors discuss how retreating barriers lose sand to the shelf to form a sand sheet seaward of the barrier. Barriers eventually starve and undergo barrier overstep. Thereafter, the barriers exist as submarine sand shoals before becoming buried by shelf mud.

The coarsening-upward lithofacies association of bioturbated carbonaceous siltstones, interbedded sandstones and siltstones, and cross-laminated sandstones represent the partial retention of transgressive barriers. The bioturbated siltstones represent fine-grained lagoonal fill. The sandstones represent advancing sediment brought into the lagoons either, through tidal inlets or, washover fans and subsequently reworked by both current and wave processes. These sandstones may well represent sedimentation

within flood-tidal deltas or the preservation of the landward facies of washover fans. Barrier overstep is suggested as a mechanism whereby the coarsening-upward lithofacies sequence is preserved.

The fining-upward lithofacies association of granule conglomerate overlain by well sorted, white to light-green sandstone represents transgressive shelf sedimentation. The granule conglomerate may well represent the debris of the erosional retreat process which accumulates within the shoreface. Much of the conglomerate associated with the transgressive event above the No. 4 seam was probably derived from the underlying braided fluvial channel system situated between the uppermost No. 4 seam and this transgressive event (Fig. 4.8). The conglomerate then represents a discontinuous basal transgressive lag. The overlying coarse-grained, wave-ripple laminated, and trough cross-bedded sandstones are possibly eroded from the underlying substrate as the shoreline underwent retreat in repose to relative sea-level rise. The discontinuous geometry of this facies association is similar to that recorded on modern continental shelves. In these modern environments, the sands vary in thickness between 0 and 10m (Belderson and Stride, 1966; Swift, 1976; Swift *et al.*, 1986), and are comparable in thickness to the sandstones of this study.

The third lithofacies facies association which comprises, medium- to coarse-grained, well sorted sandstone up to 15m in thickness represents shallow-marine sand ridges. These ridges form within the shoreface environment in response to storm flows. Swift and Field (1981) account for the origin of these sand ridges through the initial detachment of ridges from the shoreface zone. This is then followed by the headward erosion of the upcurrent swale terminations. The ridge then divides up to form smaller detached ridges. The isolated geometry of these sandstones within the study area supports this interpretation (Figs. 2.27 and 2.28). The predominance of trough cross-bedding and wave-ripple lamination indicate the movement of sediment by inshore currents and subsequent reworking of the sediment by wave action. The interbedding of

planar cross-bedded sandstone within the cross-laminated sandstone, implies the welding of smaller sandwaves onto the sand ridges. The interbedded granule-conglomerate lags within these sandstones suggest deposition during periods of peak current flow possibly generated during storm conditions. As the shoreline retreats landward with transgression, the ridges move from an active setting to a deeper offshore inactive setting. This is evidenced by the presence of carbonaceous siltstone and limestone facies overlying these sandstones (Figs. 2.27 and 2.28).

In summary, transgressive shallow-marine deposits occur overlying the No. 2, No 4, No. 5, and No. 6 coal seams. Three lithofacies associations are documented. The lithofacies associations are interpreted to represent sedimentation in shallow-marine settings. The coarsening-upward lithofacies association documents the preservation of lagoonal siltstones and the partial retention of tidal inlet and washover-fan sandstones. The fining-upward lithofacies association of conglomerate and well-sorted sandstone records deposition during shoreface retreat. The third lithofacies association of thick, medium- to coarse-grained sandstones represents the deposition of sand during the migration of shallow-marine sand ridges seaward of the shoreface zone. All three lithofacies associations record the termination of depositional sequences, and represent transgressive shallow-marine sedimentation which resulted from an increase in basin accommodation.

CHAPTER 5

5 DEPOSITIONAL ARCHITECTURE

5.1 INTRODUCTION

To portray the depositional history of the study area, the spatial and temporal depositional architectures of the Dwyka Group and Vryheid Formation are reconstructed. Cross-sections, isopach maps and borehole core are utilised to reconstruct the depositional architecture.

The depositional architecture is illustrated through the construction of a sequence of block diagrams which illustrate the changing palaeoenvironments with time. The boundaries of the study area are skewed to emphasise perspective. The horizontal scale is true, however, the vertical scale is variable to emphasise different increments of strata. The block diagrams illustrate the sequential addition of sediment to each of the depositional sequences or parasequences (Van Wagoner *et al.*, 1988). The depositional architecture of the No. 1, No. 2, No. 4, No. 5, and No. 6 seam depositional sequences is presented.

5.2 No. 1 SEAM DEPOSITIONAL SEQUENCE

With the retreat of the continental Dwyka glaciers to high-lying areas during the waning stages of the Permo-Carboniferous glaciation, much of the Karoo Basin was transgressed by the Ecca sea (Cadle *et al.*, 1993). The area of the Witbank Coalfield was largely subaerially exposed and coarse-grained, bed-load dominated fluvial systems rapidly deposited sediment over low-lying terrain (Le Blanc Smith, 1980).

However, in the study area, low-lying terrain situated in the southeast was inundated with water following glacial retreat. The sediment fill of this area resulted in the deposition of the No. 1 seam depositional sequence. Cross-sections, elevation data, and the isopach map of the Dwyka tillite, are utilised to reconstruct the palaeotopography of the study area (Fig. 5.1). The palaeotopography influenced sedimentation patterns from the No. 1 seam depositional sequence through to the No. 4 seam depositional sequence. In general, high-lying basement topography is situated along the northern and western margins of the study area and slopes away to lower elevations towards the east and southeast. The area of lowest topography is situated in the southeast. Several, fairly pronounced, north-south trending ridges and valleys are present along the western margin of the study area which give way to a more undulating topography in an easterly direction.

Figure 5.2 represents a reconstruction of the depositional systems present at the commencement of deposition of the No. 1 seam sequence. Following, the retreat of the Dwyka ice-sheets from the area, diamictite in the form of lodgement till was deposited in low-lying terrain at various topographic elevations. The thickest deposits of till are located along the eastern margin of the study area and in two palaeovalleys situated in the northeast and northwest. An embayment developed in the southeast into which sediment was fed. Glacial deposits situated in the subaerially exposed surrounding area were reworked by glacial outwash fluvial systems and fed into the embayment as coarse-grained, Gilbert-type, glaciolacustrine deltas (Bornhold and Prior, 1990; Corner *et al.*, 1990; Postma, 1990). Filling of the embayment led to the stabilisation of a glaciolacustrine delta plain. Reduction in sedimentation coupled with ground-water discharge and seasonal runoff, sustained plant growth which led to the development of the first peat swamp of the Vryheid Formation, termed the No. 1 Seam (Fig. 5.3). The

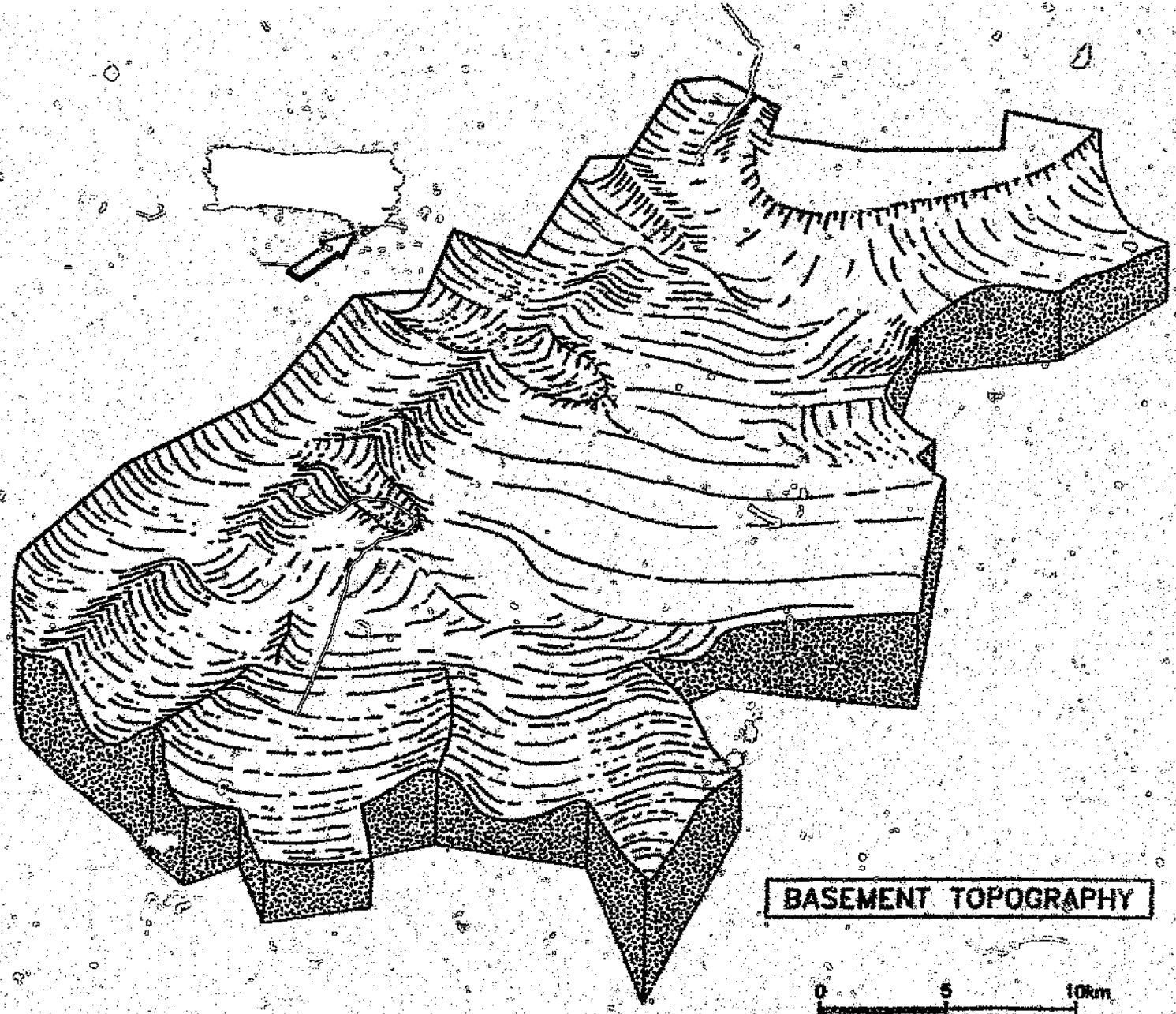


Figure 5.1. The reconstruction of basement topography in the study area.

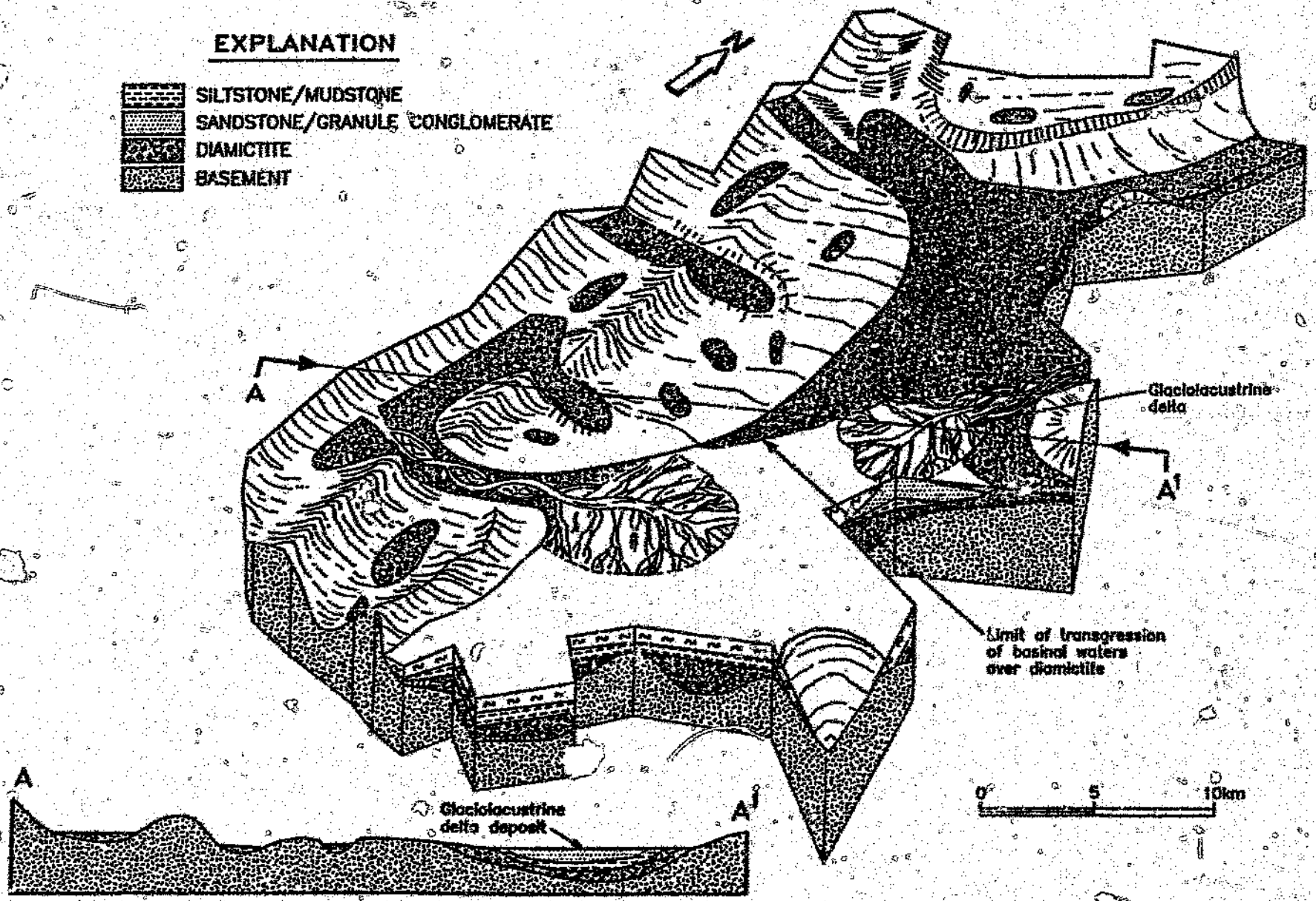


Figure 5.2. Palaeogeography of the No. 1 seam depositional sequence after the retreat of the continental Dwyka glaciers. Note the development of a glaciolacustrine environment in the southeast and the deposition of glacial till in topographically low-lying areas.

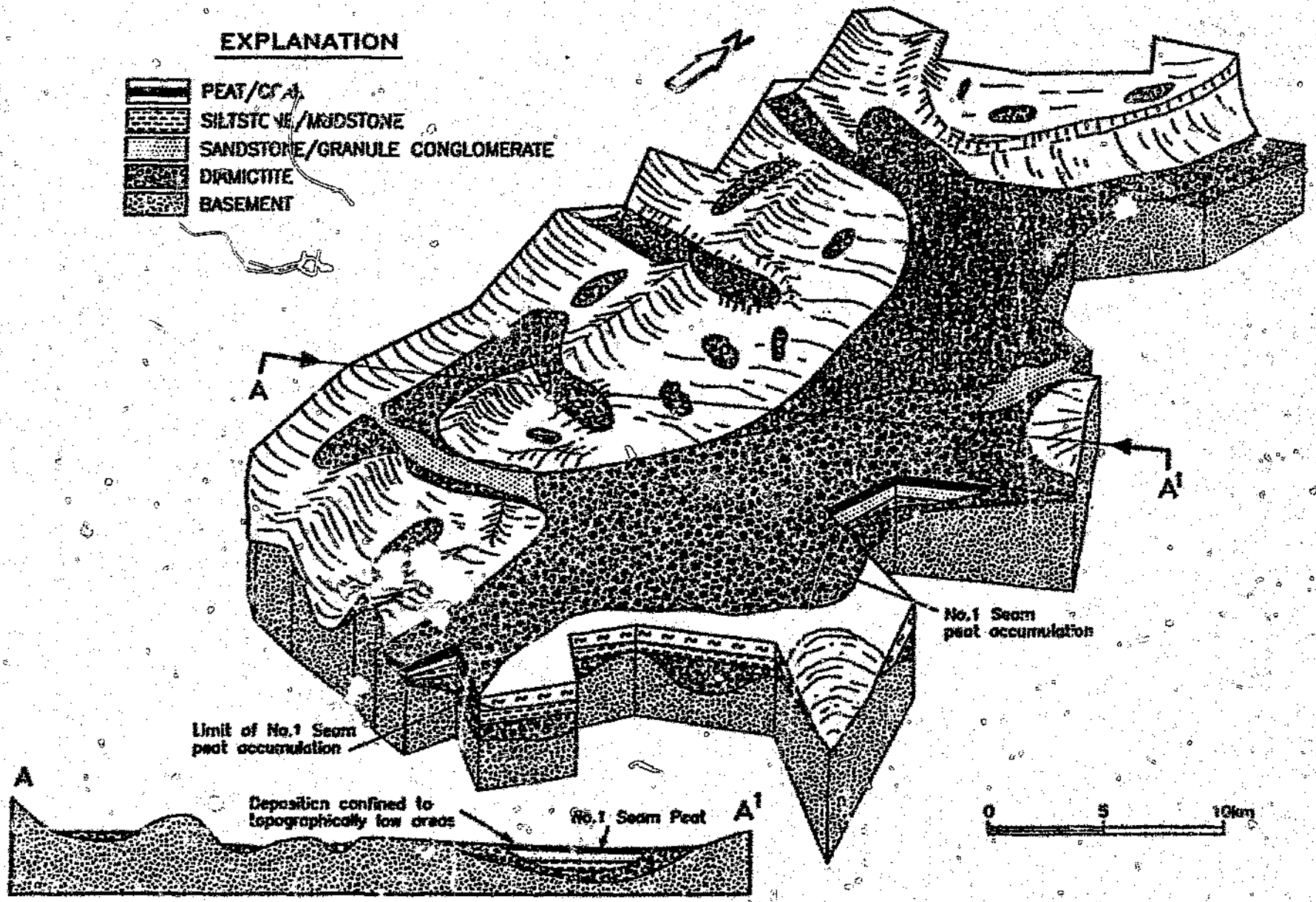


Figure 5.3. Palaeogeography of the No. 1 seam depositional sequence, illustrating the construction of a delta plain upon which the No. 1 seam peat accumulated. The No. 1 seam is mainly present in the southeastern part of the study area.

seam is thin, and restricted to the southeastern portion of the study area. This depositional sequence and the equivalent of this seam is not documented by Winter (1985) in the adjacent northern Highveld Coalfield.

5.3 No. 2 SEAM DEPOSITIONAL SEQUENCE

Transgression of basinal waters over the No. 1 seam delta plain terminated No. 1 seam peat accumulation. Figure 5.4 is a reconstruction of the palaeoenvironment at the position of maximum transgression of basinal waters over the No. 1 seam delta plain. The adjustment of the basin which gave rise to the transgression was followed by a renewed pulse of coarse-grained clastic sediments and resulted in the deposition of the No. 2 seam depositional sequence. Coarse-grained, bed-load dominated fluvial systems reworked glacially deposited sediment in subaerially exposed areas and transported the sediment to low-lying terrains. As much of the eastern portion of the study area was inundated with basinal waters, bed-load dominated fluvial systems initially flowed from the high-lying terrain in the west in an easterly direction. These fluvial systems transported and deposited sediment into the embayment to form glaciolacustrine Gilbert-type deltas (Postma, 1990).

Continued sedimentation led to the progradation of Gilbert-type deltas into the embayment and aggradation of fluvial sediments. Infilling of low-lying topography took place which led to the development of an extensive braided outwash plain, comprising coarse-grained sandstones and granule conglomerate, which advanced over the Gilbert-type deltas (Fig. 5.5). Similar sediment styles and depositional processes are recorded in the Witbank and Highveld Coalfields (Le Blanc Smith 1980; Winter, 1985) and in modern day post-glacial fjord-head deltaic deposits in northern Norway (Corner, 1975, 1990; Calles, 1977; Fjalstad, 1986). The thickness and distribution of the

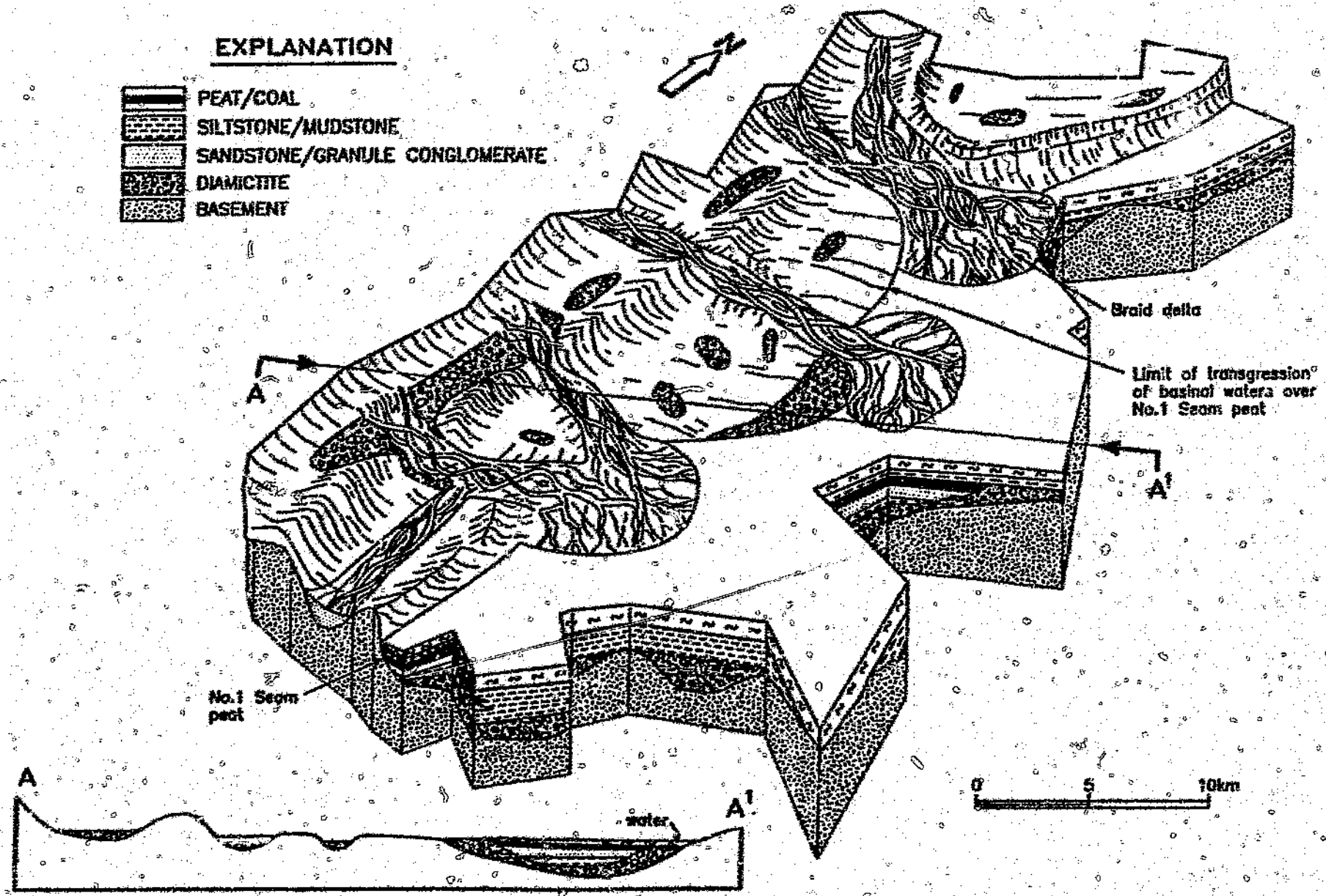


Figure 5.4. Transgression of basinal waters terminated No. 1 seam peat accumulation. Note the position of basinal waters at maximum transgression. Reworked glacial sediment was subsequently transported and deposited by bed-load dominated fluvial systems and Gilbert-type deltas into an embayment situated along the eastern portion of the study area.

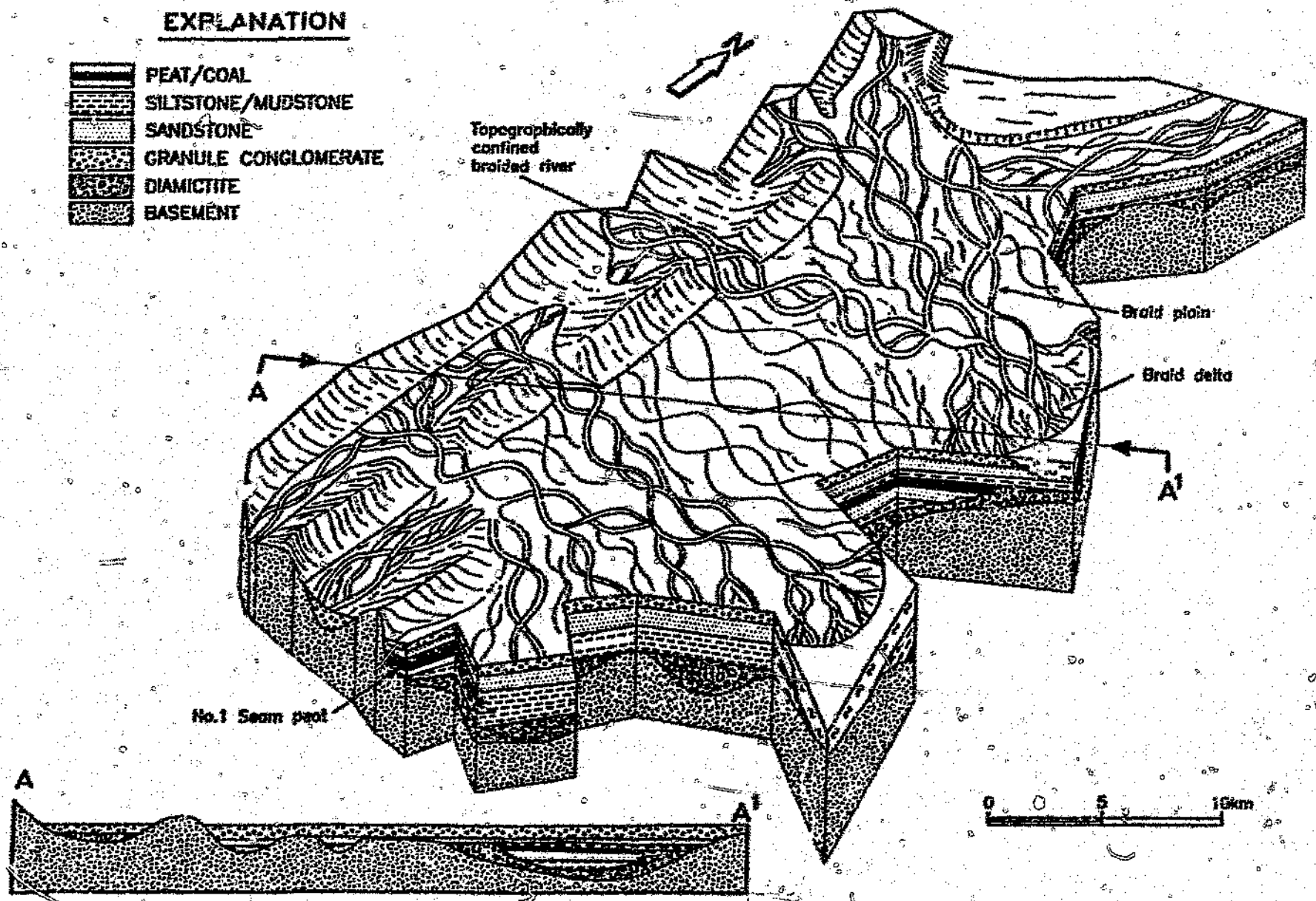


Figure 5.5. Palaeogeography of the No. 2 seam depositional sequence. Continued progradation and aggradation of fluvial and deltaic systems led to the development of an extensive outwash plain. The high-lying terrain situated in the north and east was still subaerially exposed.

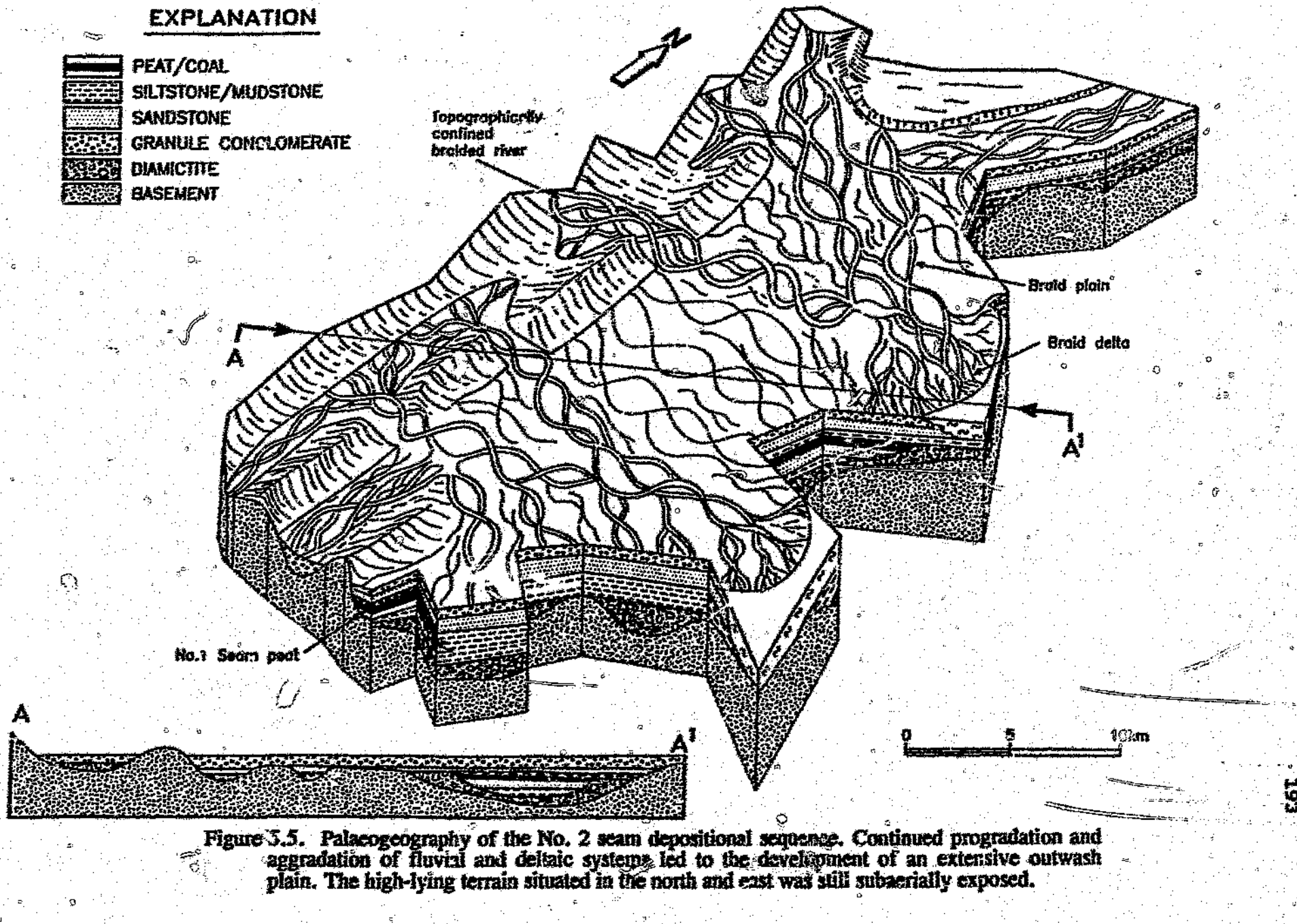


Figure 3.5. Palaeogeography of the No. 2 seam depositional sequence. Continued progradation and aggradation of fluvial and deltaic systems led to the development of an extensive outwash plain. The high-lying terrain situated in the north and east was still subaerially exposed.

sedimentary interval was still influenced by elevated basement along the northern, western and southwestern parts of the study area. Here, the sedimentary interval thins and pinches out against basement. The vertical change in lithologies from cross-bedded granule conglomerate to cross-laminated sandstones and siltstones documents a pronounced waning in sedimentation rates which enabled vegetation to become established and advance over the outwash plain. The termination of coarse-clastic sedimentation combined with the supply of ground and surface water allowed for the accumulation of thick peat which forms the No. 2 seam (Fig. 5.6). Elsewhere in the Witbank and Highveld Coalfields, both braided and anastomosed fluvial sedimentation took place during No. 2 seam peat accumulation (Le Blanc Smith, 1980; Winter 1985; Cairncross, 1986). Only minor clastic influx in the form of crevasse splay sands and silts entered the peat swamp from the north and southwest. The length and rate of peat accumulation reached optimal conditions in the northern Karoo Basin as the No. 2 seam is the thickest of all the seams and forms the bulk of South Africa's bituminous export coal (Tinney, *et al.*, 1992).

5.4 No. 4 SEAM DEPOSITIONAL SEQUENCE

No. 2 seam peat accumulation was terminated by a basin-wide transgression which drowned the peat swamp (Fig. 5.7). This basin-wide transgression is correlatable across the Witbank, Highveld and Eastern Transvaal Coalfields (Cadle, 1982b). The cause of the transgression is debateable, however, tectonic uplift of the active southern basin margin, as postulated by Quinlan and Beaumont (1984) for foreland basins, causing deepening of the basin and transgression, is favoured. In response to the deepening, shallow-water mouth-bar deltas prograded into the basin. Analysis of the fine-grained nature of the sediments, the depths of incision of the distributary channels, and the association of sedimentary structures favour the progradation of shallow-water

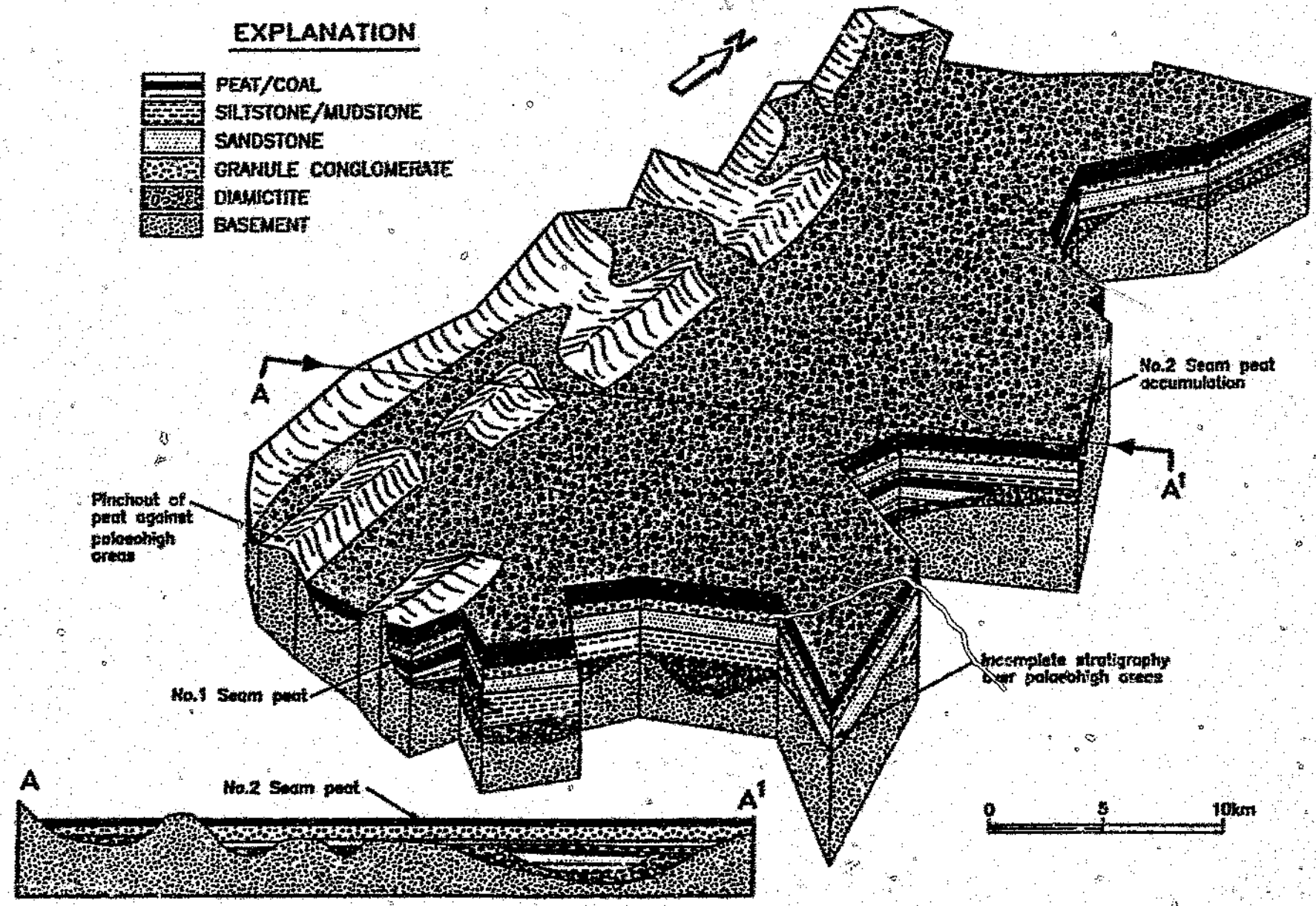


Figure 5.6. Cessation of coarse-clastic sedimentation allowed vegetation to proliferate. The No. 2 seam peat then accumulated. Note the pinch out of the No. 2 seam peat against palaeohigh areas situated in the west.

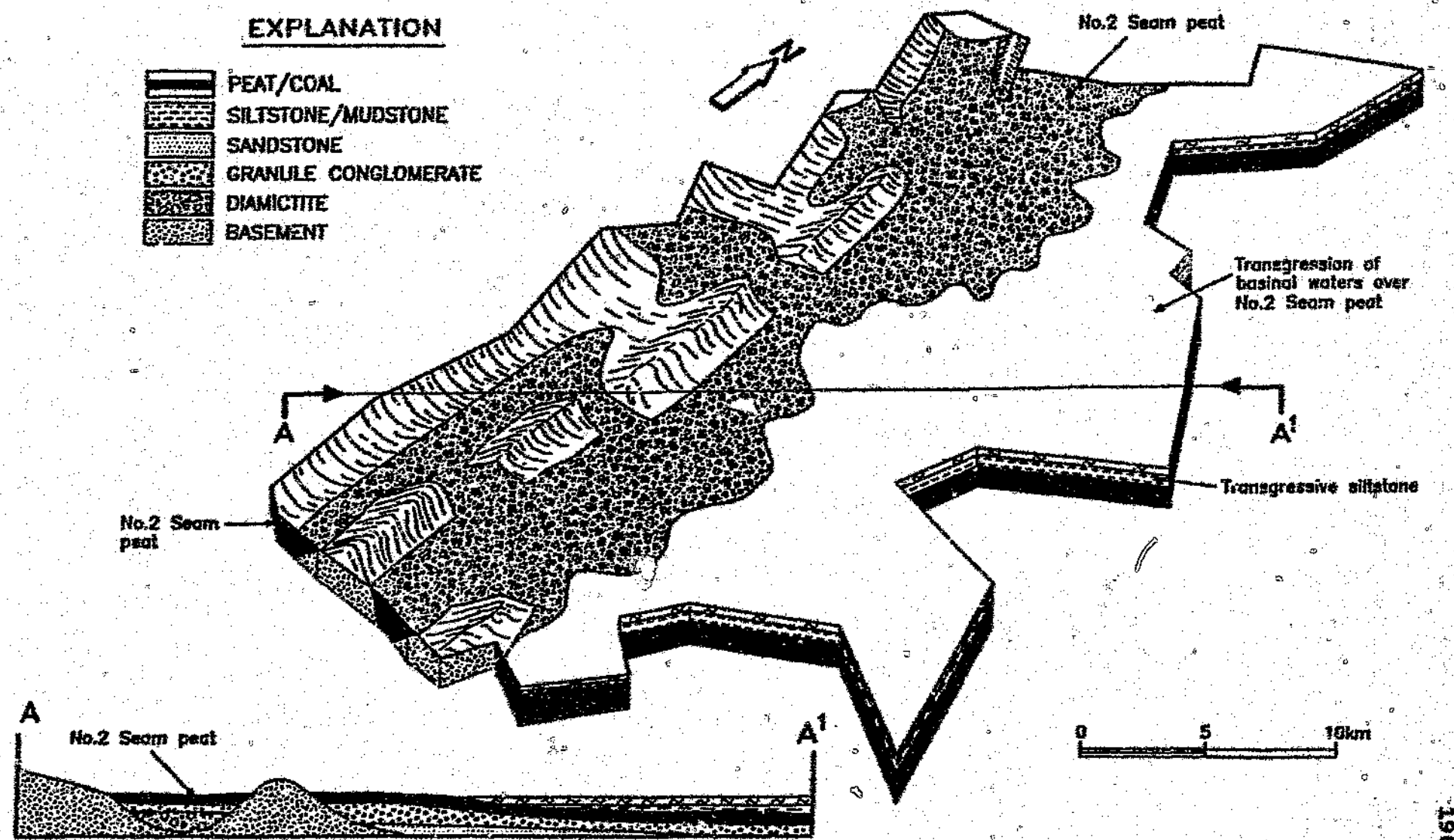


Figure 5.7. Transgression of basinal waters over the No. 2 seam peat, terminated sedimentation of the No. 2 seam depositional sequence.

mouth-bar deltas dominated by frictional and buoyant forces at the river mouth (Fig. 5.8). The deltas probably built out into a brackish to marine water body with water depths of about 10m. The deltas were fed nutrient-rich waters by updip fluvial systems which is supported by the extensive bioturbation of the delta-front siltstones. The extensive preservation of wave-ripple lamination within mouth-bar sands indicate that the mouth bars were reworked by low-energy wave action. Differential compaction of the shallow-water deltas is attested to by the delta plain comprising between two and three coarsening-upward sequences. The switching and subsidence of deltas created an unstable delta plain which was only vegetated in places. Figure 5.9 illustrates the distribution of the No. 3 seam peat at the conclusion of delta outbuilding. Thus, the No. 3 seam is represented as a thin, laterally discontinuous coal seam.

Continued subsidence led to drowning of the No. 3 seam peat swamp. Crevasse-splays, fed by advancing bed-load dominated fluvial channels advanced over the No. 3 seam delta plain (Fig. 5.10). These deposits comprise thin (3m thick) coarsening-upward sequences of carbonaceous siltstone and fine-grained, ripple-laminated sandstone. This interval of sedimentation resulted in the eventual stabilisation of the delta plain and allowed vegetation to proliferate to form the No. 4 seam peat swamp.

Cessation of coarse-clastic sedimentation at this time allowed for the No. 4 seam peat to accumulate over much of the study area (Fig. 5.11). Only in the topographically elevated areas in the southwest and west did the peat swamp pinch out against basement. Peat accumulation was interrupted by a pulse of coarse-grained clastic sediment which forms the clastic split between the No. 4 Lower and No. 4 Upper seams. This deposit of clastic sediment forms a linear axis 2km to 8km wide and up to 15m thick and has a north-south orientation (Fig. 5.11). These sediments were sourced from the northwest, north and northeast and deposited by a coarse-grained, bed-load dominated fluvial system which flowed in a southerly direction. The channel was

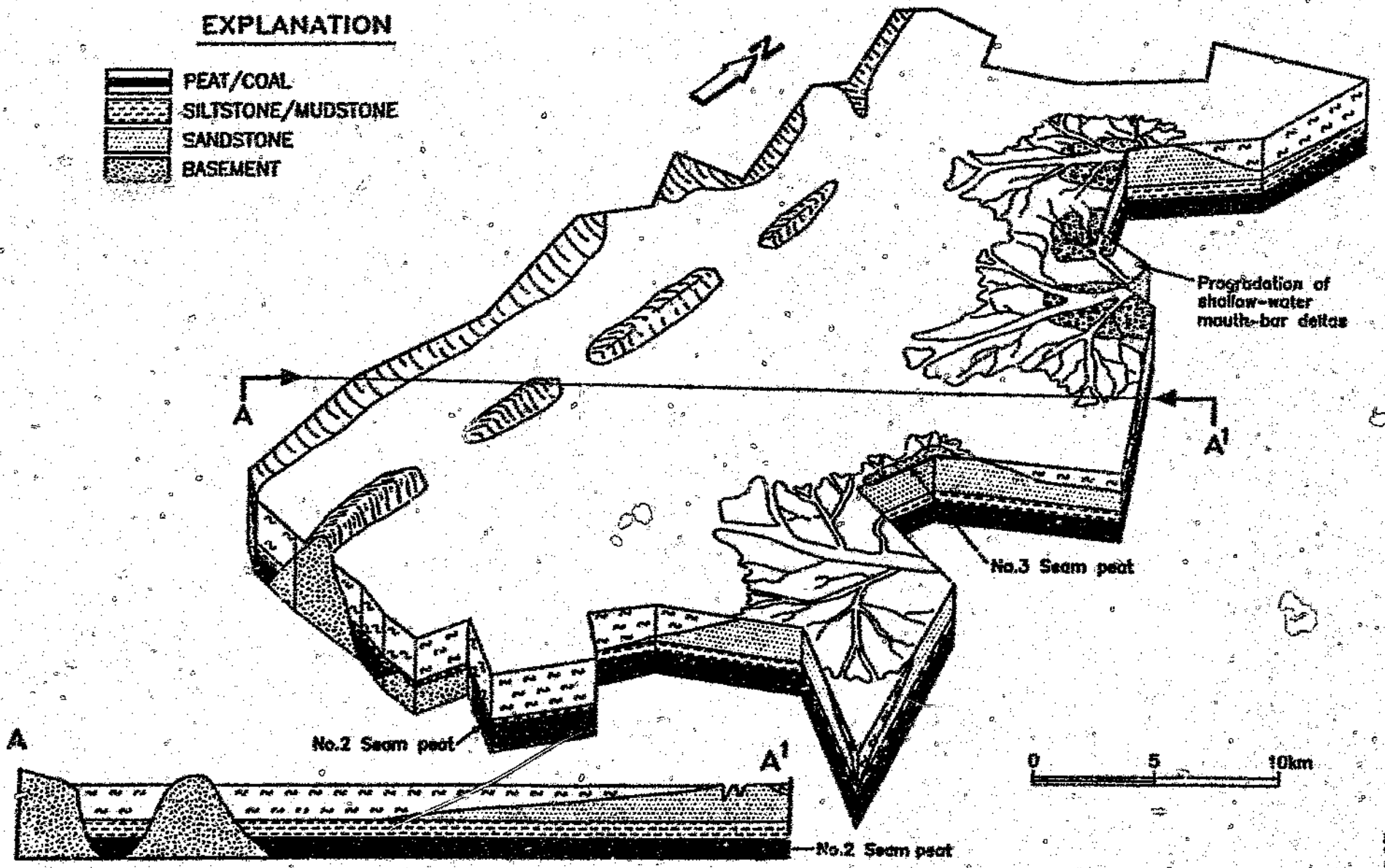


Figure 5.8. Palaeogeography of the No. 3 seam interval illustrating the progradation of shallow-water, mouth-bar deltas into the basin.

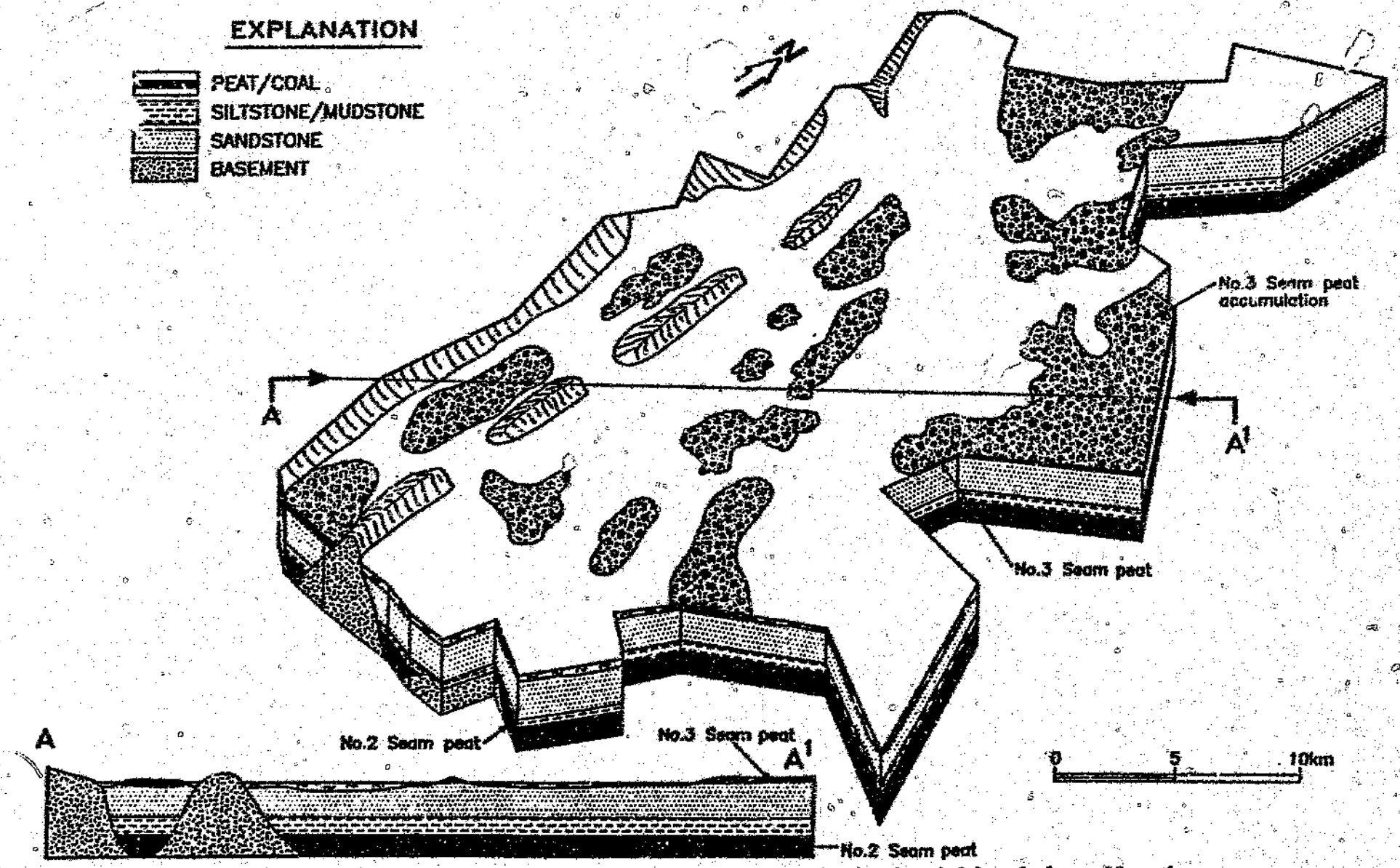


Figure 5.9. The accumulation of the No. 3 seam peat on the abandoned delta platform. Note the sporadic development of No. 3 seam peat in the area.

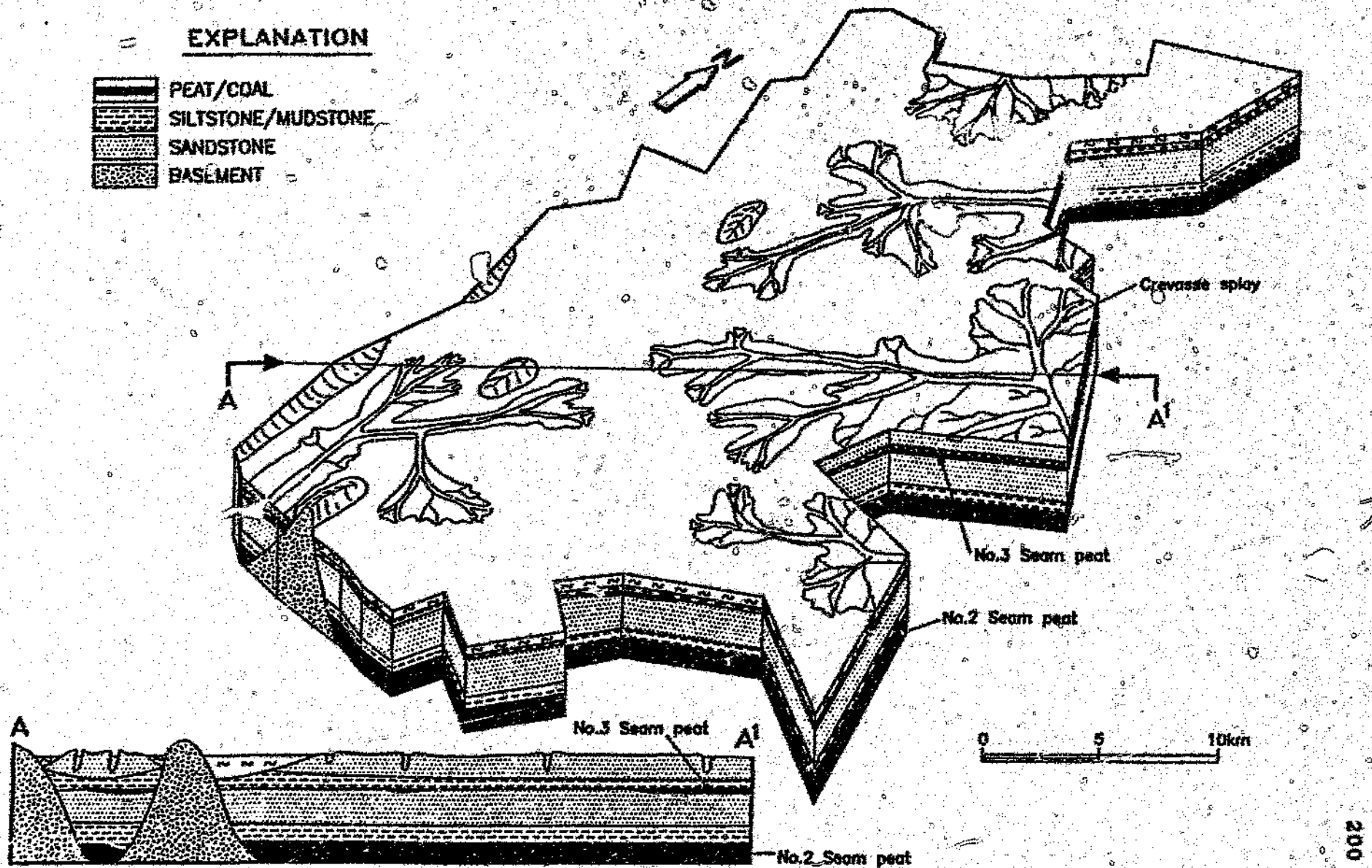


Figure 5.10. Palaeogeography of crevasse-splay sedimentation following No. 3 seam peat accumulation.

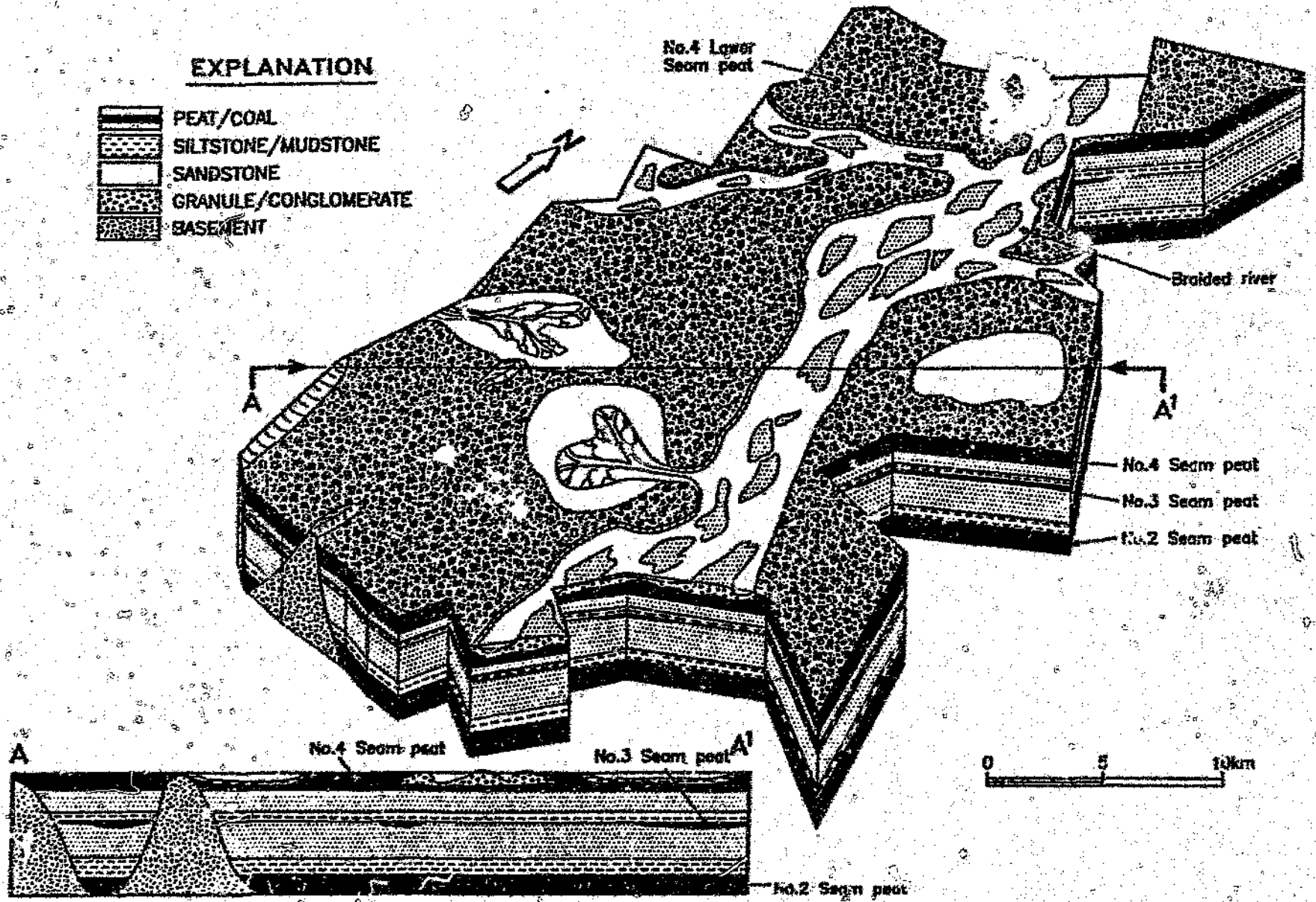


Figure 5.11 Palaeogeography during No. 4 Lower seam peat accumulation. Note a large bed-load dominated fluvial channel traversed across the swamp during the final stages of peat accumulation.

confined by vegetation and displays a well defined linear geometry. Periodic flooding of the channel, caused sediment to be funnelled through crevasse splays into the flood plain. Consequently, the clastic split between the No. 4 Lower and No. 4 Upper seams comprises proximal fining-upward crevasse-splay sediments adjacent to the channel margins. These sediments gradually grade into more distal coarsening upward crevasse-splay sediments which in turn grade into distal flood-basin silts. Figure 5.11 illustrates the palaeoenvironment during this period of sedimentation. The deposition of coarse-clastic sedimentation during No. 4 seam peat accumulation resulted in the thinning of the No. 4 Lower seam beneath the channel.

Following the cessation of this period of sedimentation, vegetation once again proliferated and accumulated to form the No. 4 Upper seam peat (Fig. 5.12). The final interval of sedimentation occurred when the No. 4 upper seam peat swamp was once again traversed across by a second bed-load dominated fluvial system with similar dimensions and geometry to the underlying fluvial system. Differential compaction of underlying sediments, resulted in a subtle topographic high existing over the location of the No. 4 Lower/No. 4 Upper seam fluvial channel. Thus the second fluvial system deposited sediment in a channel situated to the west of the underlying fluvial channel (compare Figs. 5.11 and 5.12). Isopachs of the second bed-load dominated fluvial system (Fig. 4.8) indicate that the channel was sourced from the north and northwest. In the southern part of the study area, two more tributaries flowing from the west and east join the main channel which exits the area in a southerly direction.

Basinal waters once again invaded this portion of the basin. This led to the drowning of the fluvial system and the No. 4 Upper seam peat, and termination of sedimentation constituting the No. 4 seam depositional sequence.

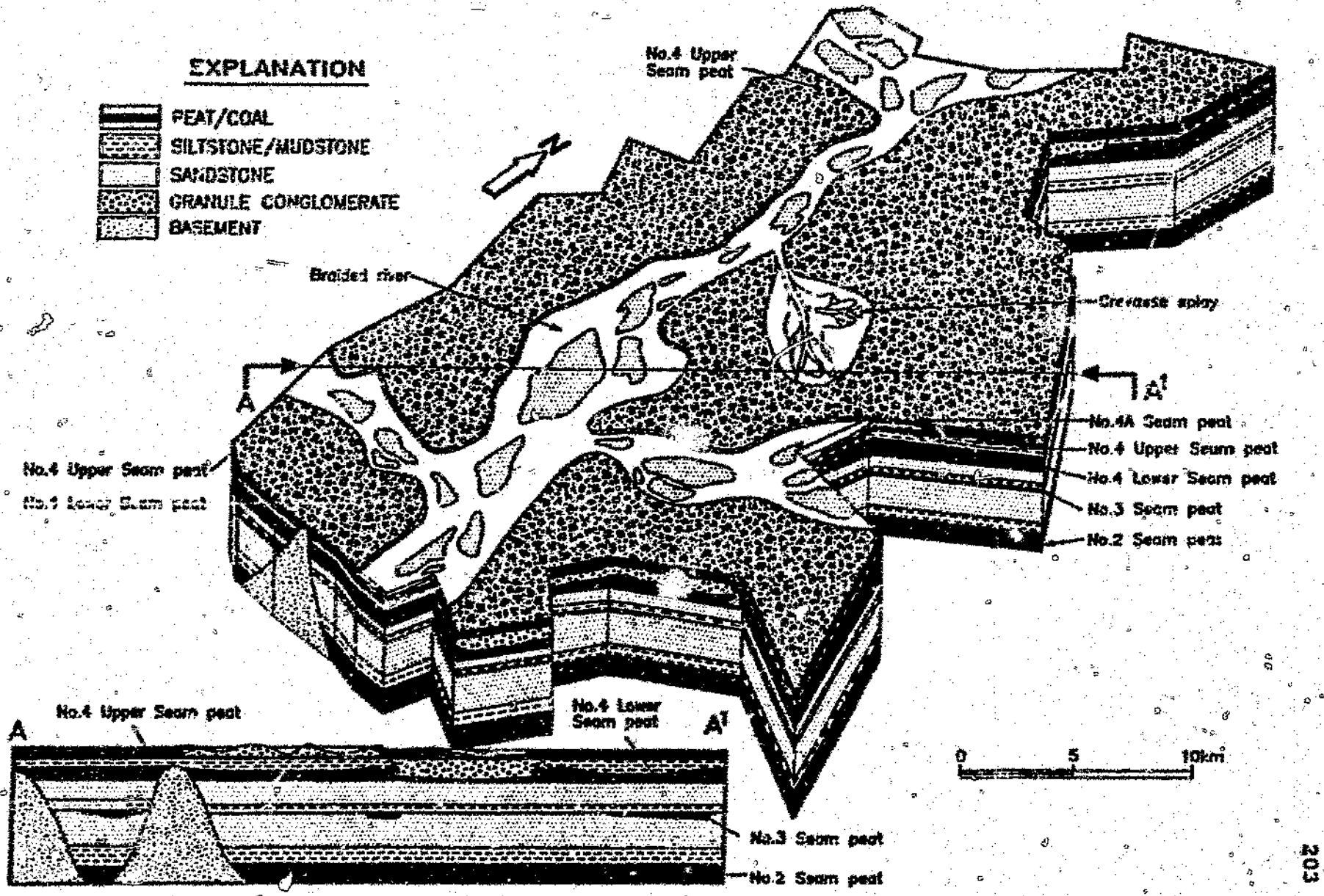


Figure 5.12. Palaeogeography during, and subsequent to, No. 4 Upper seam peat accumulation. Renewed fluvial deposition followed No. 4 Upper seam peat accumulation. Limited compaction over the underlying fluvial channel resulted in subsequent fluvial sedimentation being offset to the west.

5.5 No. 5 SEAM DEPOSITIONAL SEQUENCE

An extensive marine transgression terminated No. 4 seam peat accumulation. This transgression can be correlated across the northern portion of the Karoo Basin extending from the Natal Coalfields in the east, through the Witbank and Highveld Coalfields, to the Orange Free State Coalfield in the west (Cadle *et al.*, 1993).

Transgression occurs when sediment is entering a shelf in lesser quantities than can be accommodated by compaction or relative sea level rise. Both Vos and Hobday (1977) and Tavener-Smith (1982) have documented that microtidal conditions prevailed during the deposition of the Vryheid Formation. The glauconitic siltstones and sandstones overlying the No. 4 seam are consequently interpreted in terms of transgressive sediments that are deposited following the retreat of microtidal barrier-island systems. The retreat of the barrier-island system in the study area involved the drowning of the underlying braided-fluvial system and the reworking within the shoreface of granule conglomerate and coarse-grained sandstones into a thin, laterally discontinuous, glauconitic, sandstone-dominated deposit. The coarsening-upward glauconitic siltstones and sandstones represent the partial retention of back-barrier lagoonal muds and back-barrier sandstones as the barriers retreated in a northerly direction (Fig. 5.13).

Continued transgression led to the deposition of shelf mudstones which draped these deposits and formed the base of the No. 5 seam depositional sequence. Renewed delta progradation into the basin took place following the transgressive event. Shallow-water, Gilbert-type deltas prograded into the area from the east. These deltas differ in morphology and processes to the mouth-bar deltas associated with No. 3 seam deposition. These were coarse-grained deltas fed by closely spaced bed-load dominated distributaries (Fig 5.14). The deltas built out into water depths of between 20-30m. The deltas were subjected to fluctuating discharge which promotes the formation of

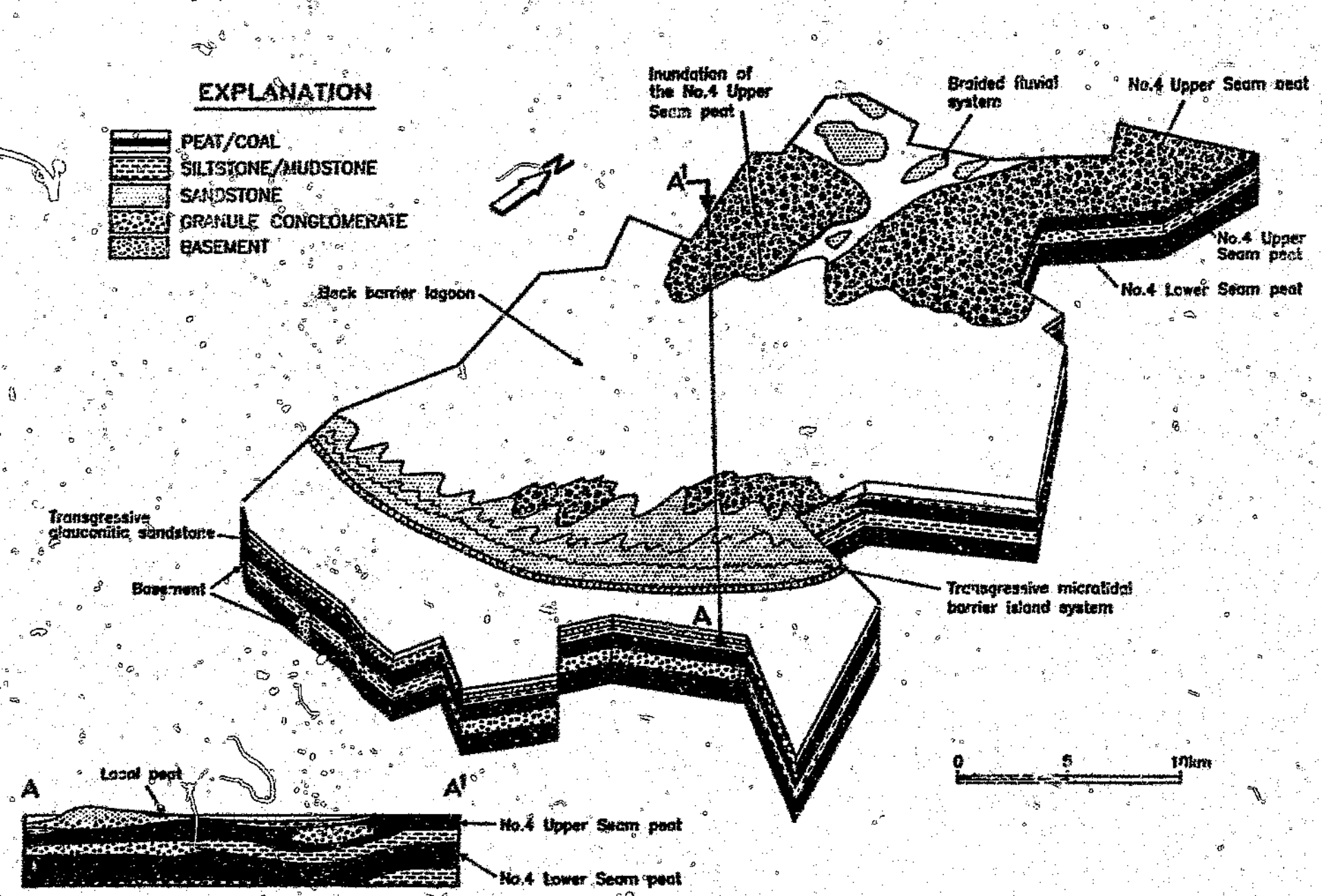


Figure 5.13, Palaeogeography at the termination of the No. 4 seam depositional sequence. Microtidal barrier-island systems transgressed northwards as a consequence of increased accommodation in the basin.

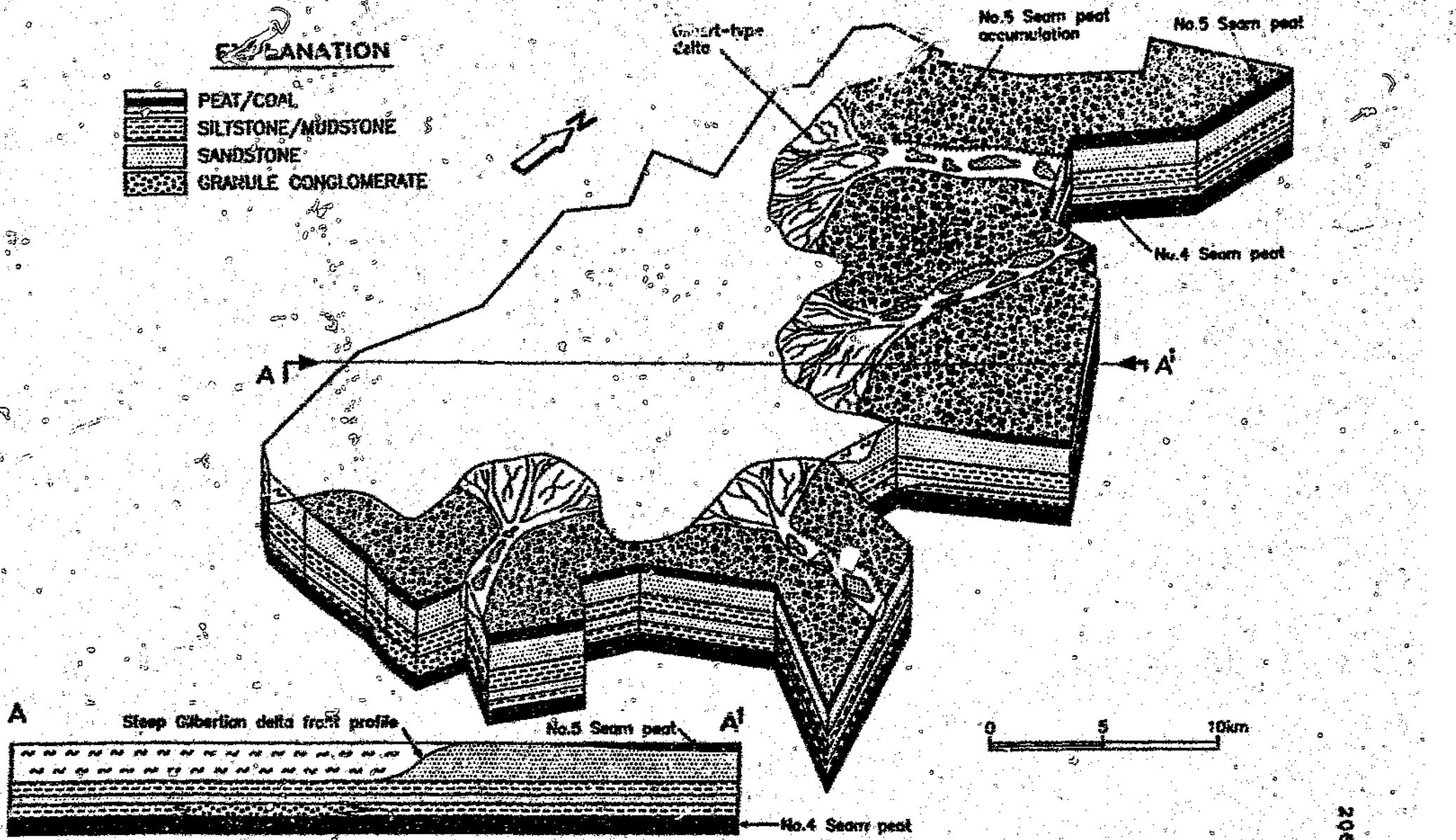


Figure 5.14. Deposition of the No. 5 seam deltaic systems. Sediment was deposited by shallow-water Gilbert-type deltas characterised by steep delta-front profiles. The deltas were fed by bed-load dominated braided distributaries. The No. 5 seam peat accumulated upon the delta plain.

inertia-dominated turbulent jets at distributary mouths. These turbulent jets discharging from distributary mouths caused the rapid deposition of bed-load material and the development of steep Gilbert-type delta foresets (Figs. 4.12 and 5.14). The locus of deposition is situated in the northeast and southeast of the study area. With continued progradation across the study area the deltas built up a stable platform of sediment forming the delta plain. This stable delta plain formed an ideal platform upon which vegetation proliferated to give rise to the No. 5 seam peat swamps. The thickest peat accumulated within interlobe areas where the ideal amount of differential compaction took place. Near-perfect conditions for peat accumulation existed, which led to the formation of the vitrinite-rich No. 5 seam (Holland *et al.*, 1989). Stable tectonic conditions prevailed during peat accumulation as the No. 5 seam coal displays a relatively uniform thickness.

5.6 No. 6 SEAM DEPOSITIONAL SEQUENCE

Tectonic instability of the basin causing an increase in the rate of accommodation resulted in drowning of the No. 5 seam peat swamps by basinal waters. This transgression is basin wide in extent and correlated across the northern margin of the Karoo Basin (Cadle *et al.*, 1993).

Sediments interpreted as shallow-marine deposits overlie the No. 5 seam. They occur as either, a coarsening-upward sequence of carbonaceous siltstone overlain by wave-ripple laminated, glauconitic sandstone or, as medium- to coarse-grained glauconitic sandstones up to 15m thick (Figs. 2.27 and 2.28). These shallow-marine sediments represent the reworked remains of microtidal barrier-island systems. The coarsening-upward sequence of siltstone and sandstone represent lagoonal siltstones overlain by relict barriers subjected to wave reworking. The thick glauconitic sandstones overlying the No. 5 seam are preserved as shallow-marine sand ridges.

These ridges are preserved during barrier retreat when sand ridges detach themselves from the shoreface as barriers retreat (Swift and Field, 1986; Swift *et al.*, 1991). Figure 5.15 illustrates the palaeogeography during barrier retreat.

Overlying the shallow-marine sediments are coarsening-upward sediments similar to the No. 5 seam depositional sequence. These sediments represent sedimentation by Gilbert-type deltas which prograded from the east (Fig. 5.16). The deltas were fed by bed-load dominated distributaries which flushed coarse bed-load material into the basin during periods of high discharge. The high bed-load ratio and dominance of inertia-dominated deposition promoted the formation of sand-dominated deltas with steep delta foresets.

A brief period of stabilisation of the delta plain is evidenced by the presence of the thin and laterally impersistent No. 6 seam (Fig. 5.17). Differential compaction of the delta plain coupled with relative sea level rise and transgression soon after the construction of the delta plain, accounts for the impersistent development of No. 6 seam peat. Tectonic instability of the basin took place at this time and is evidenced by the lack of coal seams and the presence of laterally impersistent sandstones and shales. This period of tectonic instability eventually led to extensive transgression and the deposition of the Volksrust Formation shelf shales which terminated Vryheid Formation sedimentation.

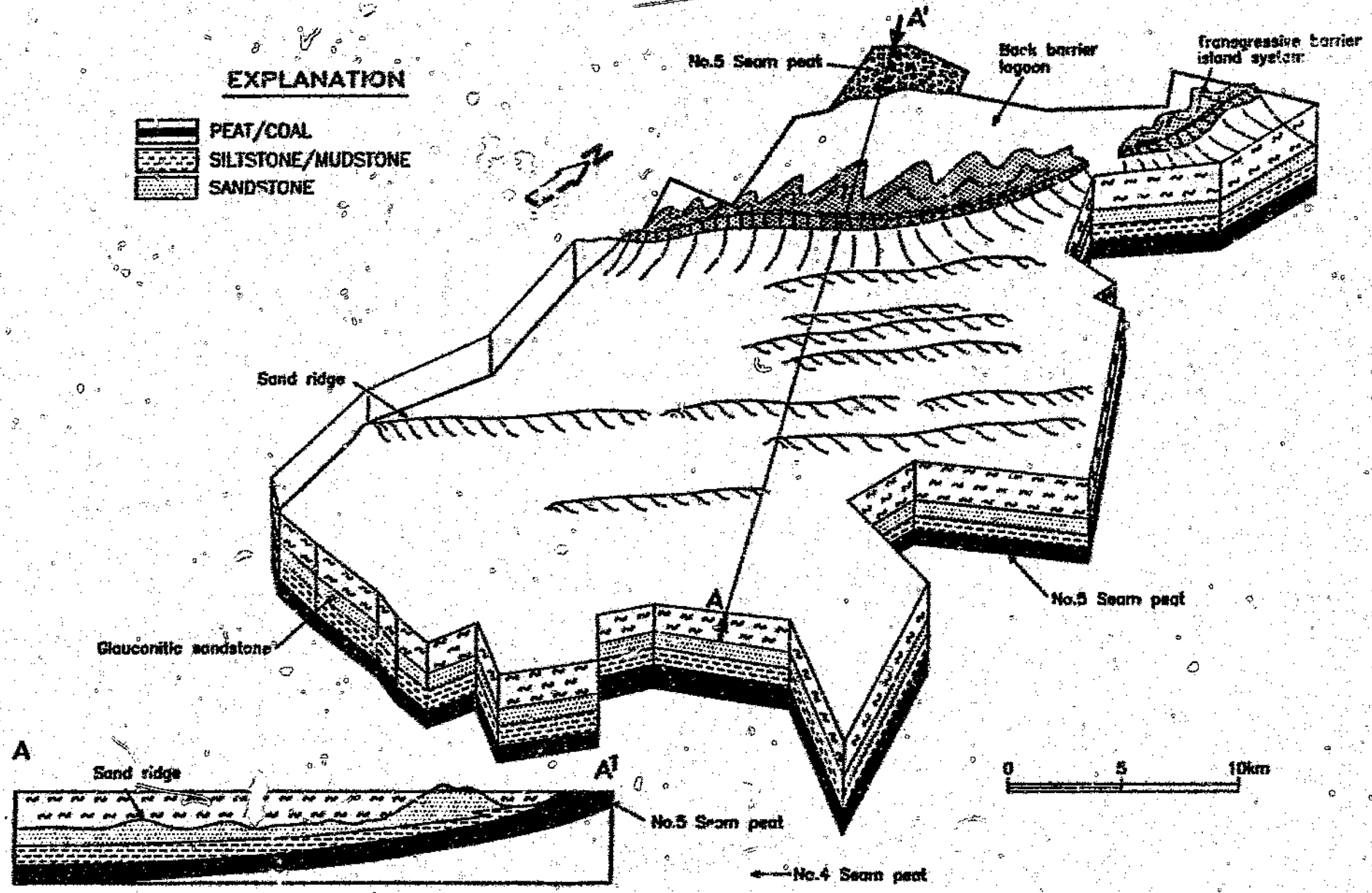


Figure 5.15. Transgression terminated No. 5 seam peat accumulation. Retreating microtidal barrier-island systems coupled with wave and tidal process deposited a series of shallow-marine sand ridges in the southern portion of the study area.

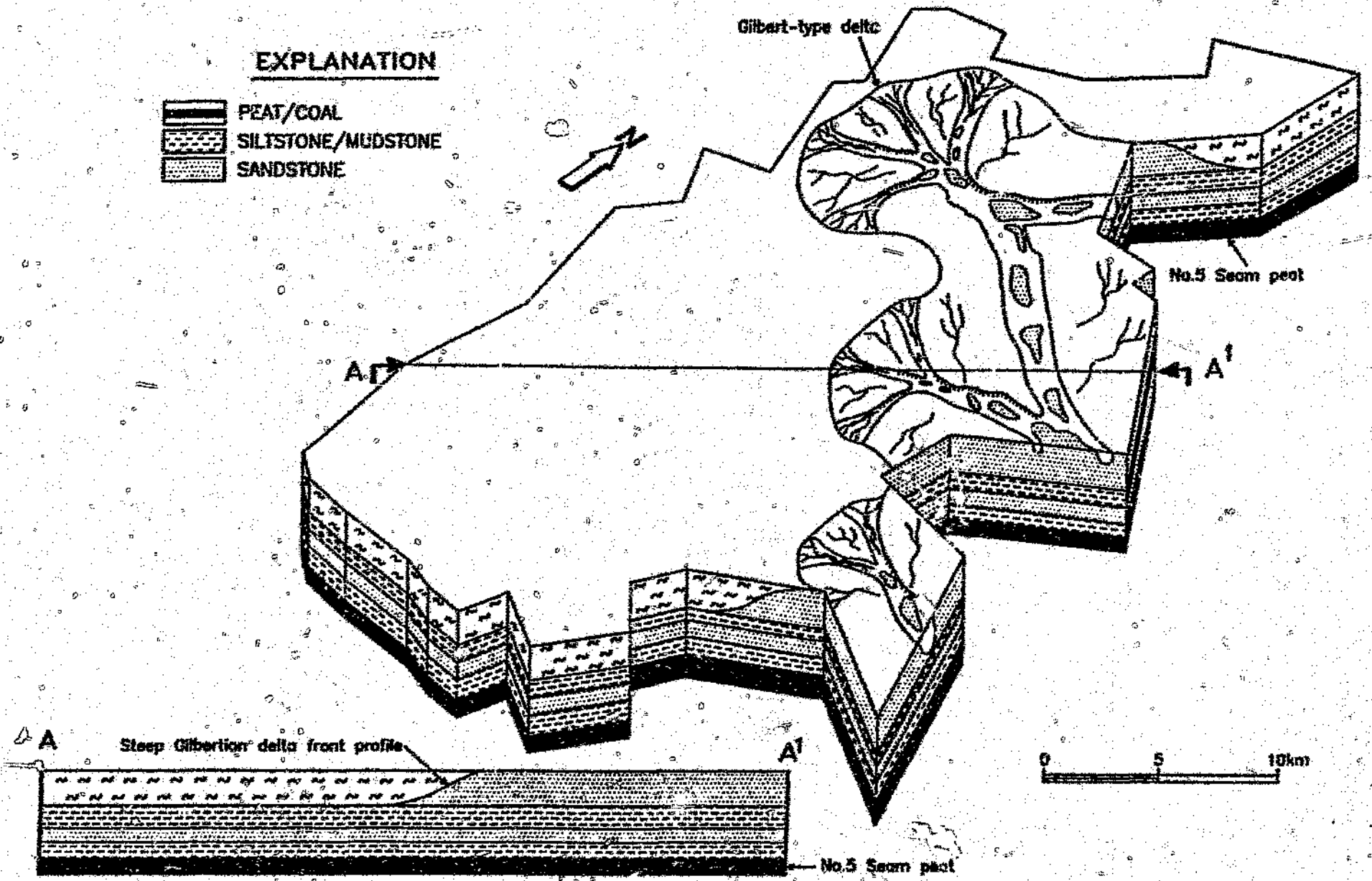


Figure 5.16. Palaeogeography of the No. 6 seam deltaic systems. Gilbert-type deltas fed by bed-load dominated distributaries prograded basinwards and built a platform upon which the No. 6 seam peat accumulated.

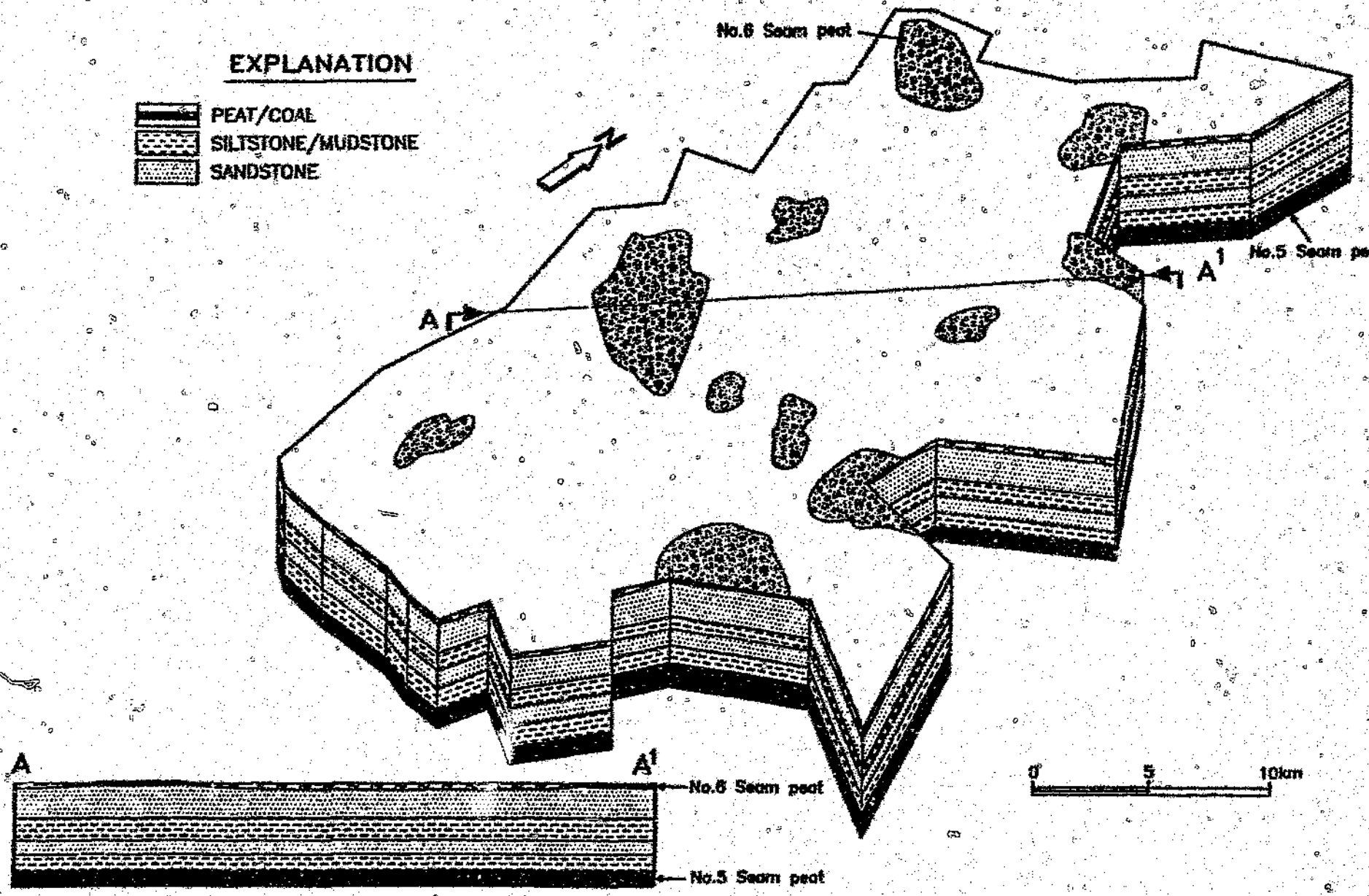


Figure 5.17. Palaeogeography during No. 6 seam peat accumulation. Differential compaction coupled with renewed transgression of basinal waters resulted in the sporadic development of No. 6 seam peat.

CHAPTER 6

6 COAL FORMATION AND DISTRIBUTION

6.1 INTRODUCTION

The importance of coal as a fossil fuel is underlined by the fact that 373,8 Mt of coal was traded on the world markets during 1992 (Tinney, 1993). Most of the world's coal resources are situated in the Upper Carboniferous and Permian rocks (40%) and in the Upper Cretaceous and Tertiary rocks (50%) (Fettweis, 1979). These data suggest that peat formed and was preserved as coal under specific environmental and tectonic regimes and were confined to specific periods of time during earth history.

For centuries coal has been exploited for its combustible qualities. Knowledge of this rock type is immense in terms of its chemical and physical properties, its change in behaviour on combustion and, more recently, the sedimentology of coal-forming environments. Yet very few authors specify what coal is, and how it is defined. McCabe (1984) describes coal as a sedimentary rock and focuses our attention on the relationship of coal to other sedimentary rocks while Diessel (1992), writes that coal is an organic sediment comprising coalified vegetal matter. A widely accepted definition of coal is that defined by Schopf (1956). This definition states that:

" Coal is a readily combustible rock containing more than 50 percent by weight and 70 percent by volume carbonaceous material, formed from the compaction or induration of variously altered plant remains similar to those of peaty deposits. Differences in the kinds of plant materials (type), in degree of metamorphism (rank), and range in impurities (grade), are characteristic of the varieties of coal".

Appreciation of the second part of this definition reveals that a great variety of coals can occur in nature. Coal therefore is classified by one or many parameters depending on the end use of coal. One aspect of Schopf's definition which is omitted is the aspect of the environment of coal formation. This is surprising considering that both the organic material and much of mineral matter constituting coal is a reflection of the environment during peat accumulation. Thus, studies of depositional models of coal-bearing strata have been formulated for use in both exploration and mining.

The lack of coal quality data has focused this study on the distribution and thickness of coal, and its relationship to depositional environments of peat accumulation. Furthermore, data pertaining to the lithological description of coal are used to infer the vertical stratigraphy of the coal seams in the study area. The applicability of these data to coal models are discussed.

6.2 PEAT AND COAL FORMATION

Coal forms from the accumulation of organic matter, primarily derived from plants, which has undergone both peatification and coalification. Peat formation is dependent on the excess of plant productivity over the respiratory processes of organisms. In present day peat-forming environments, peat accumulates where the rate at which detritivores and decomposers can consume available organic material is reduced. Factors which reduce the aerobic activity of microbes are; limited food resource and mineral elements, low pH and temperature, and low oxygen concentration. Of these, the lack of oxygen is considered a dominant cause of peat formation (Moore, 1989). During peat formation and accumulation, it is important that oxygen has only limited access to organic material to allow for maximum preservation. Thus, for peat to accumulate, a consistently high water table is required at, or above, the sediment surface (McCabe, 1984; Boron *et al.*, 1987).

Freshwater wetland systems in which peat accumulates are termed mires (Moore and Bellamy, 1973; Gore, 1983). Mires are subdivided into two types based upon their hydrology. Rheotrophic mires are fed by both ground-water flow and rain, whereas ombrotrophic mires receive water input only from rainfall. Various terms have been used in the literature to describe these two mire types.

Swamps are rheotrophic wetland ecosystems which have a water-table permanently above the peat surface and are subdivided into floating swamps, where vegetation is not rooted in sediment below, and swamp forests, where trees are an important constituent of the vegetation. In the United States of America, the term marsh is used to denote rheotrophic mires dominated by floating and herbaceous aquatics such as reeds and sedges (Martini and Glooschenko, 1985). The term fen is used to describe a rheotrophic ecosystem in which the dry season water table may be below the water table. It is regarded by Moore (1987) as a successional stage following swamp.

Bogs and bog forests are ombrotrophic peat forming ecosystems dependent on rainfall for their nutrient supply (Moore, 1989). Bogs are often dominated by plants of small stature such as mosses and sedges however, the boreal bogs of Canada and Scandanavia are vegetated by coniferous trees such as pine and spruce. Much has been made of raised bogs, where peat surfaces slope upwards towards the bog centre, as an environment within which low-ash peats and hence coals form (Fitch, 1954; Styan and Bustin, 1983; McCabe, 1984, 1987). Clymo (1987) draws an analogy between the forest bogs of Southeast Asia as present-day models of Carboniferous coal formation, and the boreal bogs of northern Europe, Canada and Siberia as models of Permian Gondwana coals of Australia.

Moore (1989) concludes from a study of modern day peats that no simple relationship exists between primary productivity of vegetation of the mire and the rate of peat formation. The most efficient sites of peat accumulation occur in high latitudes where high precipitation and low temperatures are experienced. High precipitation and low temperatures are factors which correspond to low primary productivity. Most high-productivity peatland vegetation areas have rapid decomposition rates such as tropical rain forests and are therefore not significant accumulators of peat. Cool climates promoting peat accumulation are supported by McCabe (1984) who concludes that the vast majority of peat is presently accumulating between the latitudes of 50°N and 70°N. Based upon the interpretations of palaeoclimate over the last 18 000 years in Canada and Britain, Haszeldine (1989) has speculated that world climatic cooling promotes regular rainfall in tropical and temperate zones, leading to extensive peat formation and hence coal. It is envisaged that favourable cool climatic conditions for peat formation existed during the Permian in South Africa. This is based upon the fact that South African Permian peats accumulated following a period of continental glaciation which existed during the Carboniferous.

The transformation of vegetable matter into peat and coal proceeds in two steps comprising the biochemical and geochemical stages. At the beginning of biochemical coalification, prior to vegetable material becoming waterlogged, oxidation and aerobic organisms degrade the source material. This process is termed humification during which organic material is converted to humic colloids by hydrolytic decomposition. These humic substances consist of a complex mixture of carboxylic, phenolic and other organic acids. With increasing depth of burial of peat, fungal and aerobic bacterial activity is replaced by anaerobic bacterial activity and both cellulose and lignin undergo further humification. Humification is followed by biochemical gelification during which humified substances undergo changes which include; partial or total loss of cell structure, peptidization, softening and plasticity, and homogenization. This process takes place partly in the peat stage but mainly in the stage of soft brown coal

(Teichmuller, 1989). Biochemical gelification is followed by geochemical gelification and comprises the coalification process during which the huminites are transformed into vitrinites. This transformation process takes place between the brown coal and hard coal stages and is attributed to increased temperature and pressure. Importantly, the colour of the coal changes from brown to black, its lustre from dull to bright and its hardness from soft to hard.

6.3 COAL THICKNESS AND DISTRIBUTION

6.3.1 Introduction

The thickness and distribution of the No. 1 to No. 6 coal seams is presented with reference to isopach data. The controls on the thickness and distribution of the seams are discussed in terms of palaeotopography, depositional environments and subsidence. All isopach data are presented in a relative sense. The terms thick, intermediate and thin are used to describe the relative thicknesses of the seams, irrespective of the fact that some seams are $\leq$ 1m thick and other seams $>$ 10m thick.

6.3.2 No. 1 Seam

The No. 1 coal seam is present in low-lying terrain located in the southeastern part of the study area (Fig. 6.1). The seam is also situated within palaeovalleys in the north of the study area. In general, the coal is thin and laterally impersistent (Fig. 6.1).

Stratigraphically, the No. 1 coal seam is present overlying the No. 1 seam depositional sequence where it has its greatest distribution (Fig. 2.8). It is also found overlying

Figure 6.1. Isopach of the No. 1 seam illustrating the distribution and thickness of the seam. The seam is generally thin and laterally impersistent. Low-lying terrain in the southeast, and palaeovalleys in the north of the study area were favourable sites for peat accumulation.

glacial tillite in low-lying depressions within palaeovalleys where the seam is sporadically distributed. The No. 1 seam peat preferentially accumulated in low-lying areas where both surface and ground water drainage took place such that raised water tables existed. In these localised areas, mires developed which resulted in the formation of the No. 1 seam peat.

Palaeotopography is considered the primary control on coal seam distribution. This is reasoned because the No. 1 coal seam is present only in low-lying areas where the water table and depositional surface were at a similar elevation. The thinness of the coal seam is attributed to the relatively short time of peat accumulation. Peat accumulation was terminated in the southeast by transgression of basinal waters and in the north by the deposition of outwash gravels of the succeeding No. 2 seam sequence.

6.3.3 No. 2 Seam

The No. 2 coal seam has a widespread distribution over the study area (Fig. 6.2A). The seam is present in low-lying terrain and is absent in the high-lying terrain situated along the western margin of the study area. In general, the No. 2 seam is thick and laterally persistent over much of the study area.

Stratigraphically, the No. 2 coal seam overlies coarse-grained sandstones and granule conglomerates of the No. 2 seam sequence (Fig. 2.6). Hence, the seam is present over the central and eastern portions of the study area. The No. 2 coal seam is also present within palaeovalleys situated along the western margin of the study area. The coal is thickly developed along a north-northwest, south-southeast trend in the middle of the study area. There is a general thinning and ultimate pinchout of the seam against the high-lying terrain in the west of the study area. The coal attains its thickest development along the northern margin of the study area (Fig. 6.2A).

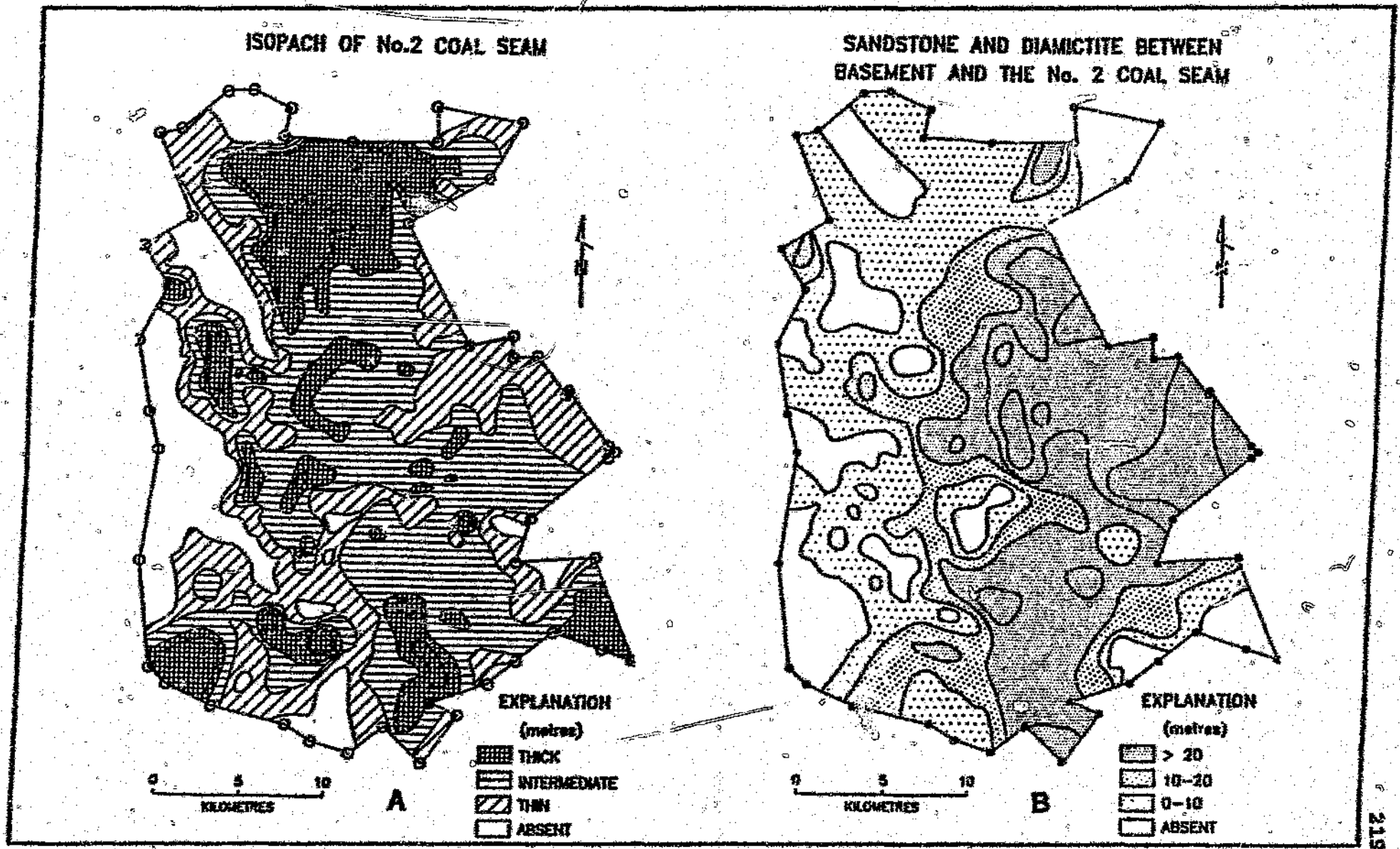


Figure 6.2. Comparison of the distribution and thickness of the No. 2 seam (A) with the underlying sedimentary interval (B). Note the thinning of the seam in the west and the influence of the 20m-40m isopach of the underlying sediments on the thickness of the seam.

Palaeotopography is considered the primary control on the distribution coal seam accumulation. Coal has accumulated in low-lying terrain and thins and pinches out against palaeohigh areas in the west (Fig. 6.2A). Differential subsidence of underlying sediments such that the water table remained at, or close to, the peat surface, may have resulted in the accumulation of intermediate and thick No. 2 seam peat. In the central and southern portions of the study area, where the underlying diamictite and sandstone is between 20m and 30m thick, the coal seam thickness is in general intermediate and thick (see Fig. 6.2A and B). In the east, where these lithologies exceed 30m in thickness, the coal is thin. Excess differential subsidence and drowning of the peat swamp accounts for the thinning of the seam in the east. The zone of thickest coal development is situated in the north of the study area and coincides with an area of raised basement topography and between 0-10m of underlying diamictite and sandstone (see Fig. 6.2A and B). It is difficult to reason why the coal attains optimum thicknesses in this area. The suggestion is made that a raised water table existed in this area during No. 2 seam peat accumulation which was coupled with optimum subsidence of the peat swamp.

6.3.4 No. 3 Seam

The No. 3 seam has an irregular distribution over the study area (Fig. 6.3A). The seam is absent along the western margin of the study area and displays a random distribution over the rest of the study area. In general, the seam is thin and laterally impersistent.

Stratigraphically, the No. 3 seam overlies the siltstones and fine-grained sandstones comprising the No. 3 seam deltaic interval of the No. 4 seam sequence (Fig. 2.14). The No. 3 seam interval is present over much of the study area with the exception of

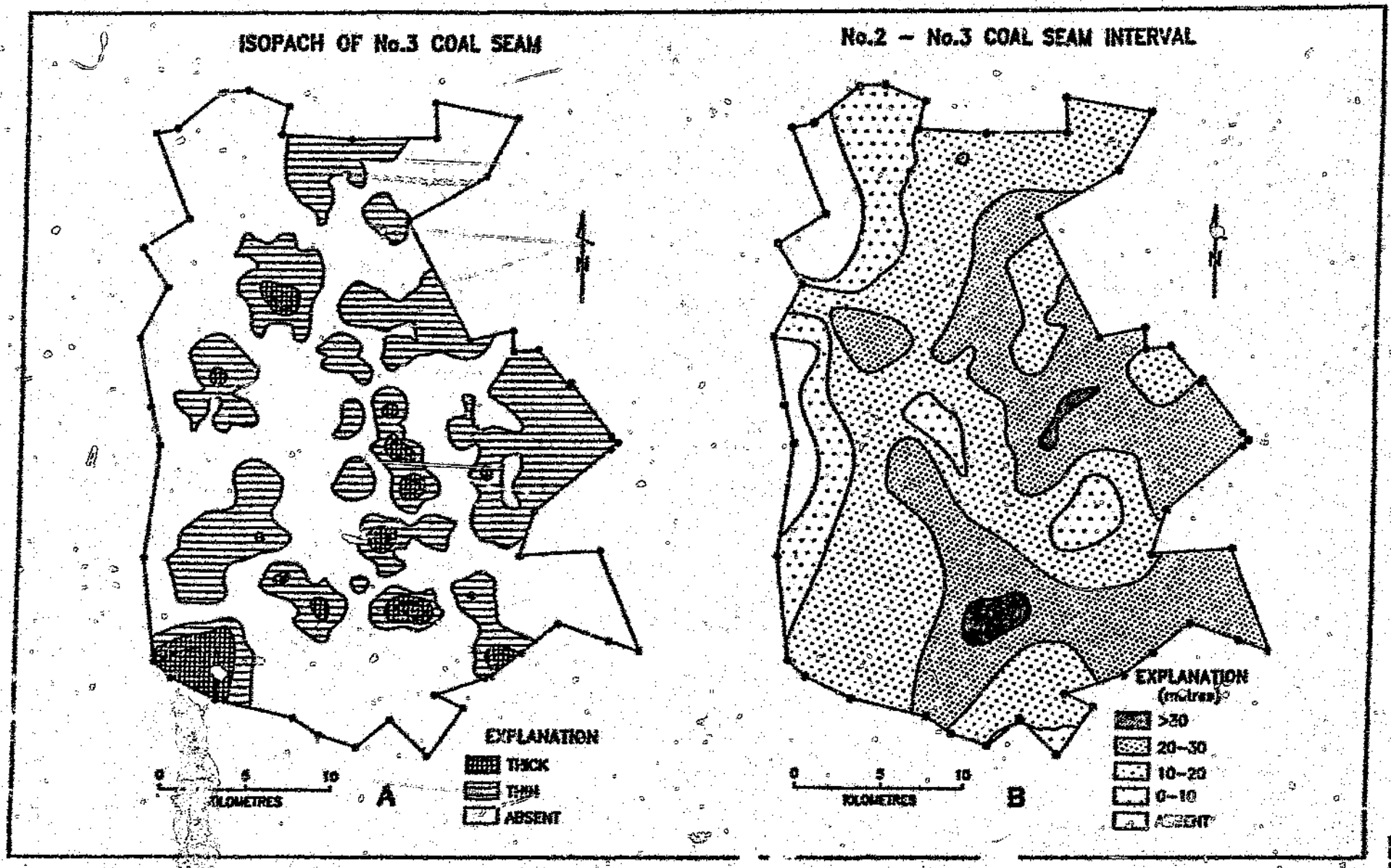


Figure 6.3. Comparison of the distribution and thickness of No. 3 seam (A) with the isopach of No. 2 to No. 3 seam interval (B). Note the coincidence of the 20-30m isopach of the No. 2 to No. 3 seam interval with the presence of No. 3 seam coal.

the high-lying ground in the west and this is reflected by the distribution of the seam (Fig. 6.3A). The No. 3 seam is widely distributed throughout the eastern portion of the study area where the No. 2 to No. 3 seam interval is between 20-30m (see Fig. 6.3A and B). However, in the western half of the study area the No. 3 seam is present where this interval varies between 10 and 20m.

Differential subsidence is an important factor accounting for the distribution and thickness of the No. 3 seam. The thinness of the seam is accounted for by fairly rapid differential subsidence of underlying sediments which allowed basinal waters to transgress the delta plain. This resulted in the drowning of incipient peat swamps and the termination of peat accumulation. This event was followed by a period of crevasse-splay sedimentation which ultimately led to the formation of a stable delta plain. The succeeding No. 4 seam accumulated on this plain. The presence of No. 3 seam peat in areas of thicker underlying sediments suggests that selected parts of these areas were, for a short period of time, sufficiently stable for peat accumulation. Palaeotopography is an obvious control on No. 3 seam distribution as the seam is absent in the elevated areas along the western margin of the study area.

6.3.5 No. 4 Lower Seam

The No. 4 Lower seam is a laterally extensive coal seam which is present over most of the study area however, the seam thins towards basement elevated areas along the western margin of the study area. In general, the seam is thick and laterally persistent over much of the study area (Fig. 6.4A).

Stratigraphically, the No. 4 Lower seam overlies the mudstones, siltstones and fine- to medium-grained sandstones which comprise both the No. 3 to No. 3 seam interval and

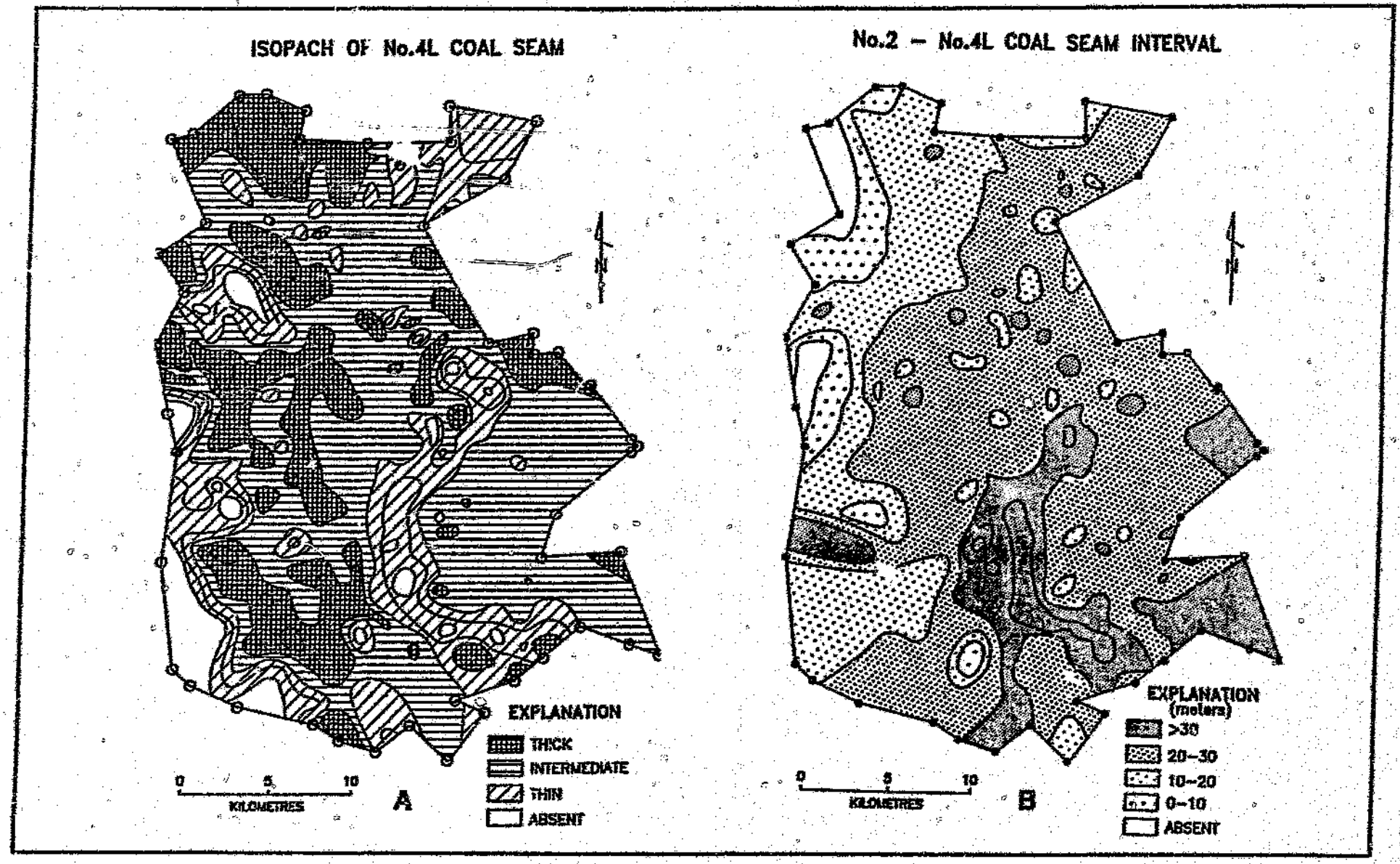


Figure 6.4. Comparison of the distribution and thickness of the No. 4 Lower seam (A) with the isopach of the No. 2 seam to No. 4 Lower seam interval (B). Note the correlation between the north-south thinning of the seam and the 30m isopach of the No. 2 to No. 4 seam interval.

the No. 3 to No. 4 seam interval. The crevasse-splay sediments which immediately underlie the No. 4 Lower seam provided a stable deltaic platform upon which the No. 4 Lower seam peat could accumulate. The No. 4 Lower seam thins in two main areas. Firstly, the seam thins and pinches out in places along the western margin of the study area. Secondly, a pronounced thinning of the seam is present as a northsouth linear area in the southern portion of the study area (Fig. 6.4A). The thickest occurrences of the No. 4 Lower seam are present as irregular northsouth trending belt adjacent to the western margin of the study area (Fig. 6.4A).

Basement topography and differential subsidence of the delta plain account for the distribution and variation in thickness of the No. 4 Lower seam. The thinning and pinchout of the No. 4 Lower seam along the western margin of the study area is attributed to the pinchout of the seam against elevated areas of basement. The thinning and absence of the seam in a linear zone in the southern part of the study area coincides with the thick accumulation of sediments within the No. 2 to No. 4 seam interval (see Fig. 6.4A and B). This zone underwent excessive subsidence and is envisaged as an embayment during the period of initial peat accumulation. With time the embayment filled and became stabilised hence, the thinness of the seam in this area. The thinning of the coal seam in the northeast corner of the study area is possibly due to erosion of the overlying fluvial channel sequence.

Areas which remained relatively stable during peat accumulation, and maintained the water table at or just above the surface of peat accumulation, accumulated thick peat. These conditions existed along the western margin of the study area in a belt flanking the most elevated basement high areas (Fig. 6.4A). The thickness of the No. 2 to No. 4 seam interval varies between 0m and 30m in these regions (Fig 6.4B).

6.3.6 No. 4 Upper Seam

The distribution of the No. 4 Upper coal seam is similar to that of the No. 4 Lower coal seam as the seam is laterally persistent and present over much of the study area. However, compared to the No. 4 Lower coal seam, the No. 4 Upper seam is absent in many more areas than the underlying seam. In general, the No. 4 Upper coal seam is a thin, laterally persistent coal seam (Fig. 6.5A).

Stratigraphically, the No. 4 Upper coal seam is separated from the underlying No. 4 Lower coal seam by a relatively thin sequence of braided-stream and crevasse-splay sedimentary rocks. These sedimentary rocks vary from mudstones to granule conglomerates. The influence of basement topography at this stratigraphic interval is not marked consequently, the seam is present over much of the western margin of the study area. The most striking feature of the No. 4 Upper coal-seam thickness is the northwest-southeast linear zone of thin coal, within which there are small areas where coal is absent (Fig. 6.5A). To the east and west of this zone the coal seam increases in thickness. The greatest thickness of coal is present as irregular areas situated mainly along the eastern margin of the study area.

Contemporaneous sedimentation during No. 4 Upper seam peat accumulation coupled with differential subsidence controlled seam thickness. Comparison of the sedimentary interval between the No. 4 Lower and No. 4 Upper seams (Fig. 6.5B), with the zone of thin No. 4 Upper seam coal (Fig. 6.5A), illustrates a remarkable correlation between the zone of thin coal and the 3m isopach of the sedimentary interval. These data coupled with cross-sections through this interval (Figs. 2.16 and 2.17), imply contemporaneous peat accumulation during deposition of braided-fluvial and associated crevasse-splay sediments (Fig. 5.11). With abandonment of the braided channel, peat encroached along the sides and over most of the channel. However, over the channel

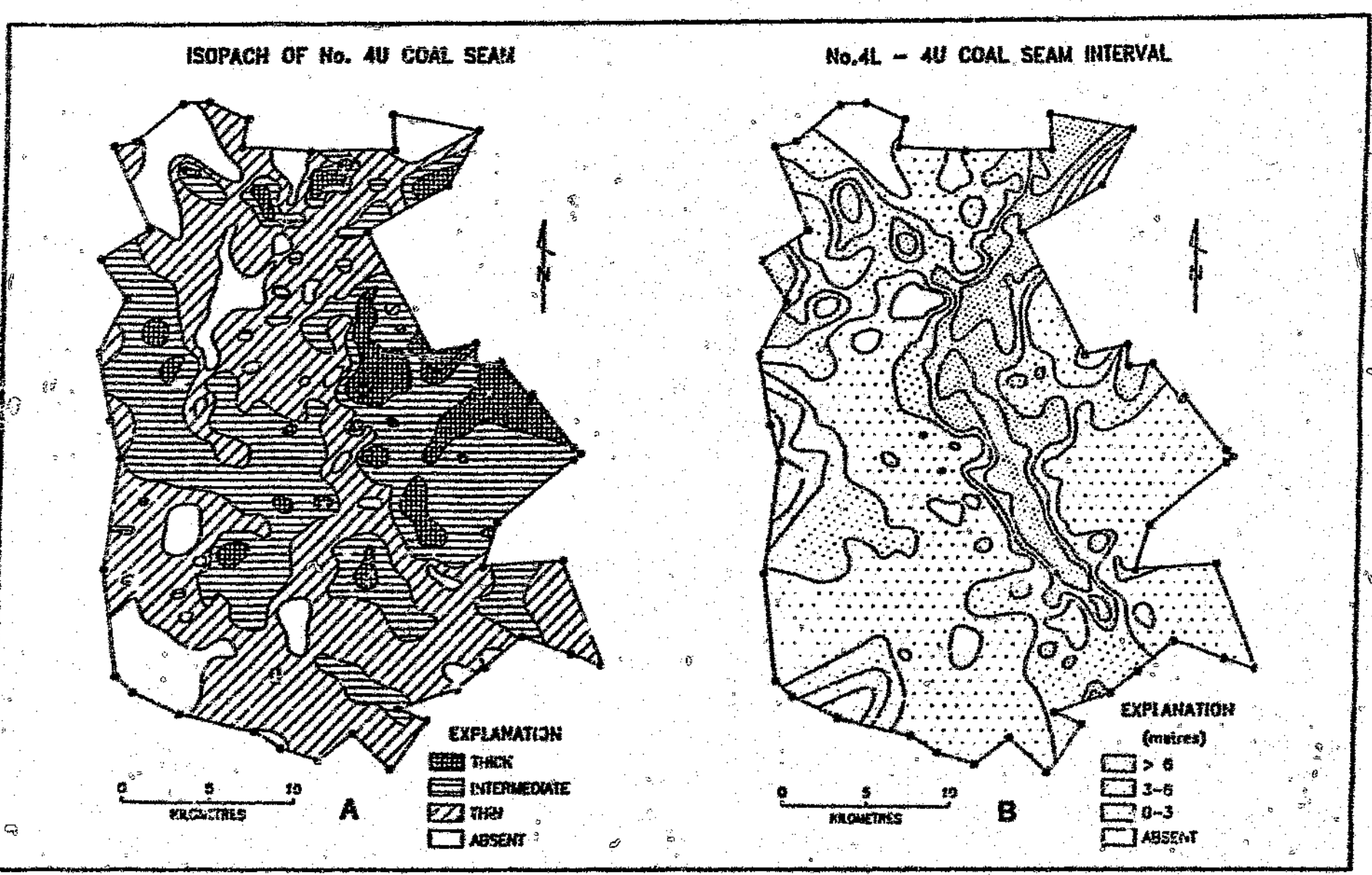


Figure 6.5. Comparison of the isopach of the No. 4 Upper seam (A) with the No. 4 Lower/No. 4 Upper seam clastic parting (B). The seam thins over, and thickens either side of, the underlying No. 4 Lower/No. 4 Upper seam clastic parting.

axis peat did not accumulate prior to the succeeding depositional sequence which accounts for the many small areas where the No. 4 Upper seam is absent (Fig. 6.5A). Thicker No. 4 Upper seam is present in areas where the No. 4 Lower-No. 4 Upper interval is between 0m and 3m in thickness. These areas are situated on either side of the braided channel (see Fig. 6.5A and B). At these localities, the fluvial flood plain was sufficiently stable that slow subsidence took place allowing for continuous peat accumulation.

6.3.7 No. 5 Seam

The No. 5 coal seam has a widespread distribution over the study area (Fig. 6.6A). At this level in the stratigraphy, elevated basement topography has little or no influence on the distribution of the No. 5 seam hence the seam is present over much of the western margin of the study area. Within the study area, the seam is a thin and laterally persistent.

Stratigraphically, the seam terminates the No. 5 seam depositional sequence which displays a characteristic coarsening-upward sedimentary sequence. The coarsening-upward sequence commences with mudstones and siltstones which grades upward into fine- and coarse-grained sandstones. However, crucial to the thickness variation of the No. 5 seam is the sandstone-dominated interval immediately overlying the No. 4 Upper coal seam and underlying the fine-grained clastics of the No. 5 seam depositional sequence. This sandstone interval represents a period of fluvial sedimentation following peat accumulation of the No. 4 Upper seam. This was followed by a period of shallow-marine sedimentation and reworking of the fluvial sediments following marine transgression of the fluvial flood plain (Figs. 5.12 and 5.13).

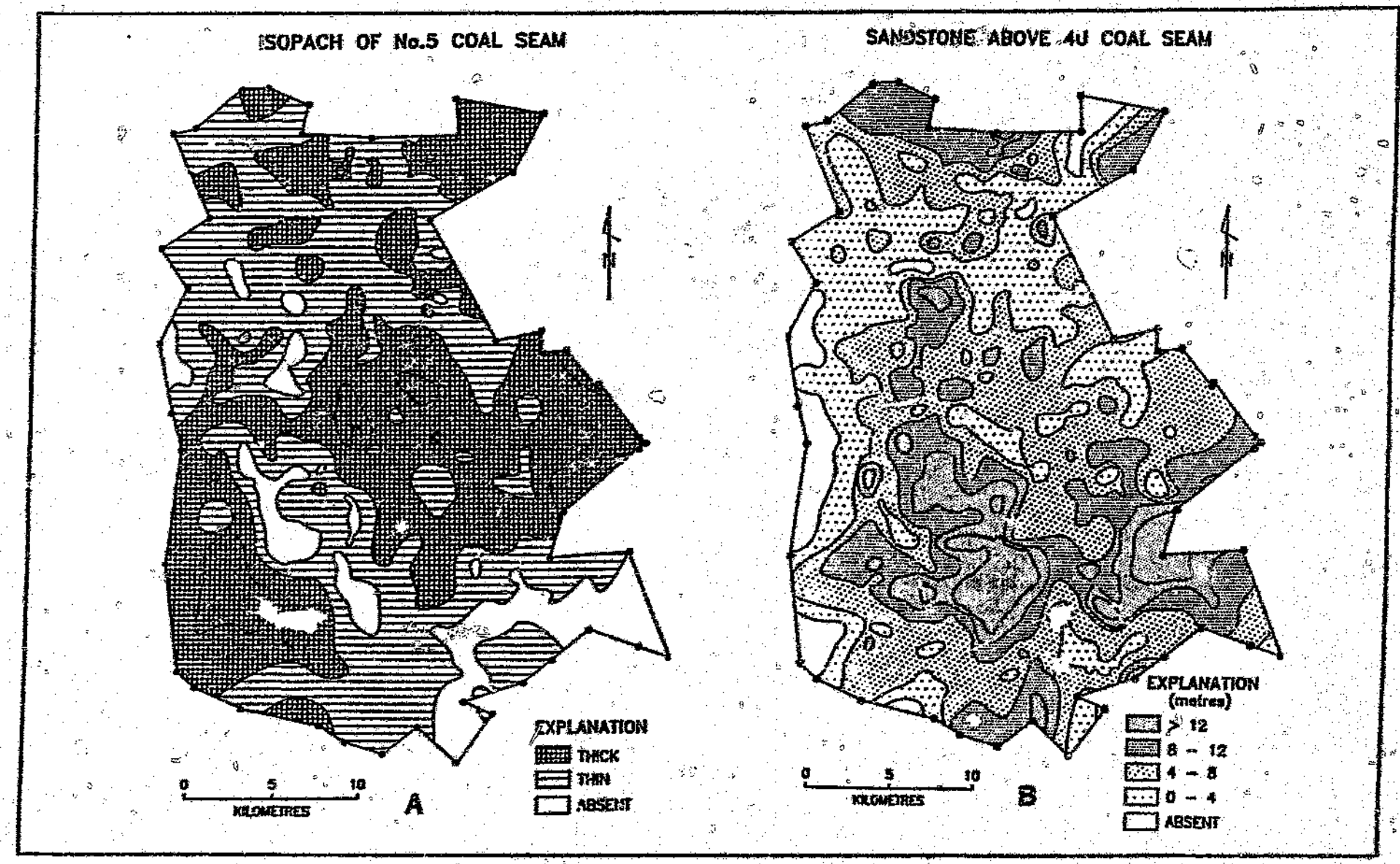


Figure 6.6. Comparison of the No. 5 seam isopach (A), with the sandstone overlying the No. 4 Upper seam (B). The seam thins in those areas where the sandstone overlying the No. 4 Upper seam is >8m in thickness.

The isopach map of the No. 5 seam illustrates an irregular thickness distribution (Fig. 6.6A). The seam is thin and absent in the southern and northwestern portions of the study area. In addition, a northsouth linear zone of thin coal and coal absence, is present adjacent to the western margin of the study area. Thick coal is present along the southwestern margin of the study area and within a large portion of the study area in the east (Fig. 6.6A).

Differential subsidence of underlying sediments has controlled both the distribution and thickness of the No. 5 coal seam. A most striking correlation can be made between the sandstone overlying the No. 4 Upper seam (Fig. 6.6B) and the distribution and thickness of the No. 5 seam (Fig. 6.6A). The 8m sandstone isopach defines the axes of two marine-reworked braided-fluvial channels which trend northsouth and eastwest respectively. The positions of these channels coincide with the zones of coal thinning and absence. A similar correlation does not exist between the No. 5 coal seam and the underlying siltstones and sandstones which comprise the coarsening-upward sequence which immediately underlies the seam. It is concluded that, during No. 5 seam peat accumulation, the combined weight of the underlying fluvial and deltaic sediments caused excess subsidence of the peat swamp to take place in those areas underlying the marine-reworked fluvial channels (Fig. 2.27, cross-section C). These areas remained embayments during the early phase of No. 5 seam peat accumulation. With time, these embayments filled with carbonaceous siltstones, swamps evolved and peat accumulated. Hence, the coal which presently overlies these channel sandstones is thin.

In the east of the study area, the No. 5 coal seam is thickly distributed in those areas adjacent to the marine-reworked fluvial channels and where the underlying deltaic sandstones are between 5m and 10m in thickness (Fig. 6.7 and Fig. 6.6B). Optimum conditions for peat accumulation took place in these areas where the rate of subsidence of the delta plain maintained the water table at or close to the peat surface. Tectonic

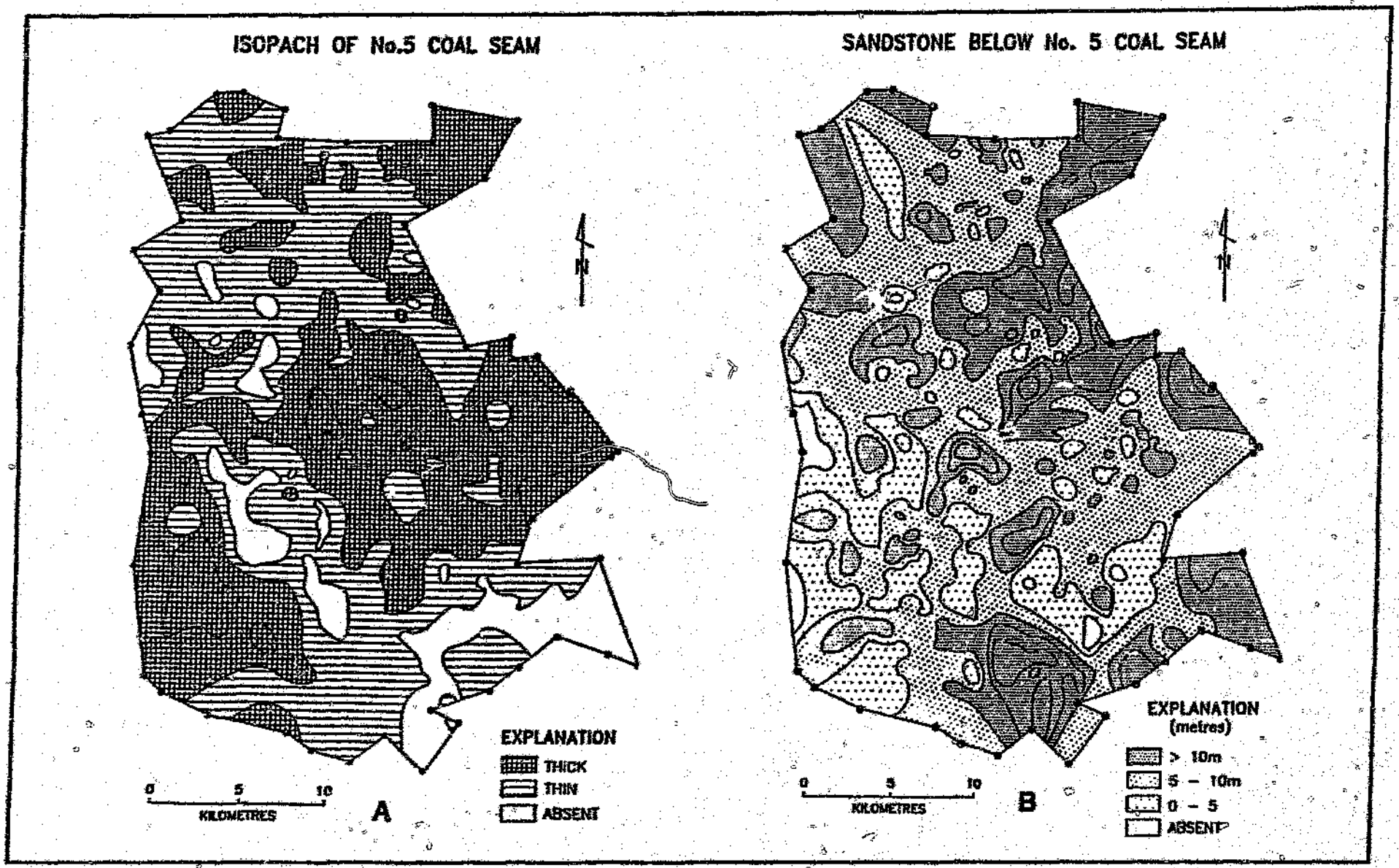


Figure 6.7. Comparison of the No. 5 seam isopach (A), with the deltaic sandstones underlying the No. 5 seam (B). Thick No. 5 seam coal is present either side of the fluvial channels (Fig. 6.6B) where the deltaic sandstone is between 5m and 10m thick.

instability of the basin caused transgression of the basinal waters over the No. 5 seam peat consequently, a short period of time existed for peat accumulation. Thus, compared to the No. 2 and No. 4 seams, the No. 5 seam is, in general, a thin seam.

6.3.8 No. 6 Seam

The No. 6 coal seam has an irregular distribution throughout the study area (Fig. 6.8A). The seam is laterally impersistent and thin. Stratigraphically, the No. 6 coal seam overlies a coarsening-upward sequence of siltstone, sandstone and granule conglomerate resulting from a period of deltaic sedimentation (Figs. 2.26 and 2.27). The coal seam is randomly distributed as isolated small occurrences throughout the study area and does not attain thicknesses that warrant economic exploitation.

Basin-wide tectonic instability probably accounts for the distribution and thickness of this seam. Many of the occurrences of the seam are situated at various positions along the 30m isopach for the No. 6 depositional sequence (see Fig. 6.8A and B). This probably represents the margin of a large delta lobe where peat formation began to take place. Tectonic instability of the basin shortly after deposition of the No. 6 seam delta complex and incipient peat formation, caused basinal waters to transgress over the delta complex and terminate any further peat accumulation. The No. 6 coal seam is therefore a thin, laterally impersistent coal seam.

6.4 COAL SEAM FACIES PROFILES

Facies profiles of the No. 1 to No. 6 coal seams from the study area are described. The coal facies are based upon macroscopic descriptions of the seams obtained from

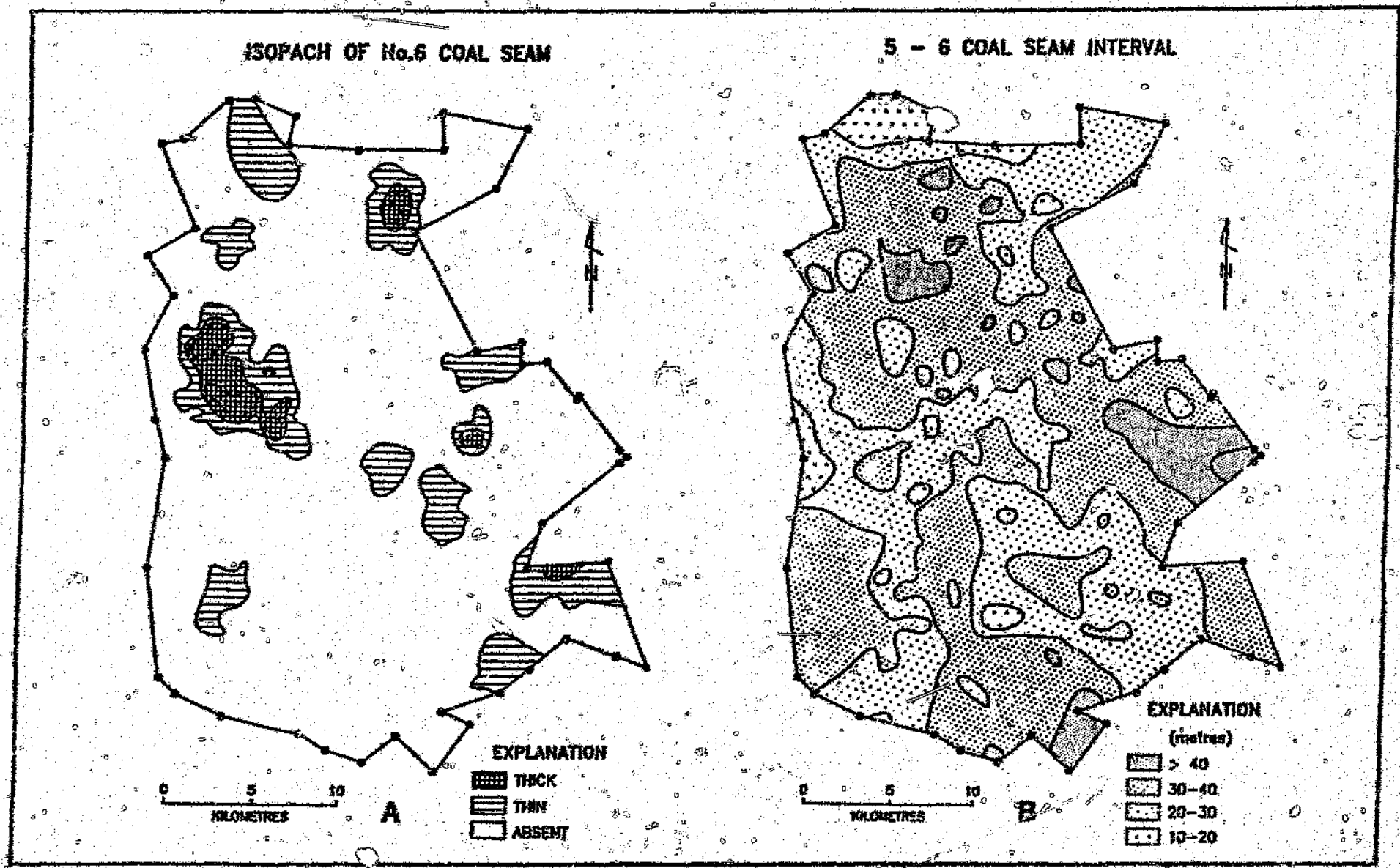


Figure 6.8. Comparison of the No. 6 seam isopach (A), with the isopach of the No. 6 seam depositional sequence (B). Note the occurrence of the seam in many areas where the thickness of the sequence is between 30m and 40m.

borehole logs. The seams are described as lithotypes in terms of the lustre of the coal which varies from dull to bright (Fig. 6.9). The relationship between macroscopic coal lithotypes and microscopic coal macerals is that dull-coal lithotypes represent predominantly the inertinite group of macerals and that bright coal lithotypes represent predominantly the vitrinite group of macerals (Stach *et al.*, 1982; Diessel, 1992). This interpretation is based upon the fact that there is a low percentage of liptinite macerals in the Highveld and Witbank Coalfields (Falcon, 1989). Thus a coal described as bright with dull bands is interpreted as a predominantly vitrinite-rich coal which is interbanded with subordinate inertinite-rich coal layers. Figure 6.9 illustrates generalised profiles of the coal seams. Two profiles of the No. 2 and No. 4 Lower seams are presented which illustrate the variation in coal lithotype in areas of thick and thin coal respectively. The Nos. 2, 4, and 5 seams are initially described as they are relatively thick seams and display similarities in their lithotype trends. Thereafter, the Nos. 1, 3 and 6 seams are described as they represent thin seams which display alternating bright and dull lithotypes.

The No. 2 coal seam is the thickest of all the coal seams and contains the greatest amount of bright coal (Fig. 6.9). In the north of the study area, where the seam is thickest, the overall vertical trend of the seam is from bright coal at the base to dull coal at the top of the seam. In detail, the seam displays a laterally impersistent dull-coal band at the base of the seam which is overlain by a bright-coal zone containing dull bands. This coal zone grades upward into a dull and bright interbanded zone which in turn passes upward into a carbonaceous siltstone towards the middle of the seam. Overlying the carbonaceous siltstone the coal is dull with bright bands which is overlain by a second bright coal zone with dull bands (Fig. 6.9). The top of the seam comprises a dull coal zone.

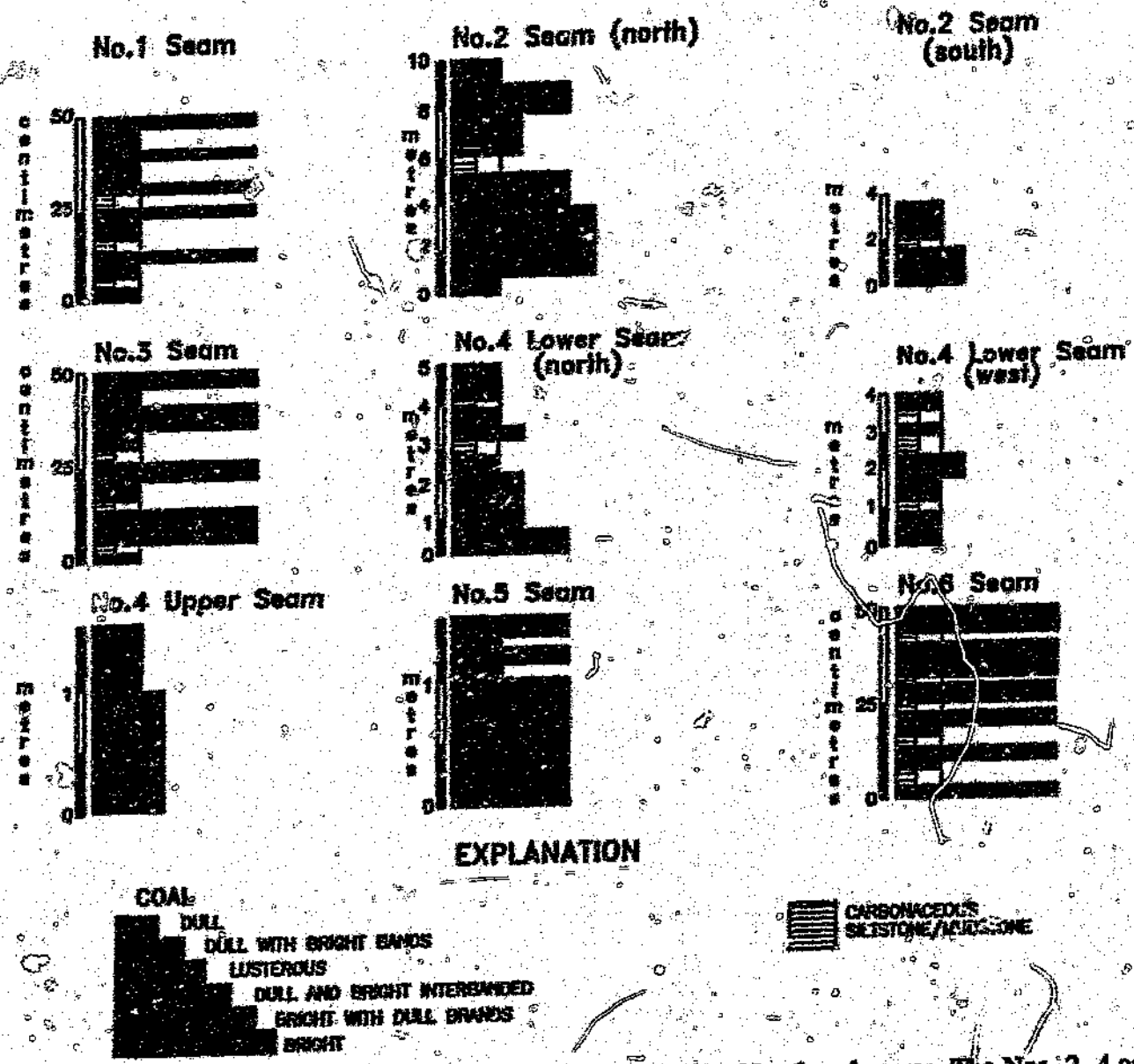


Figure 6.9. Generalised coal lithotype profiles of the No. 1 to No. 6 coal seams. The Nos. 2, 4 and 5 seams display brighter lithotypes at the base of the seam which pass upward into dull lithotypes. The Nos. 1, 3 and 6 seams are thin and comprise alternating bright and dull lithotypes and shale.

In the south of the study area, where the coal seam is thin and less than 4m in thickness, the seam is dull. Once again a vertical trend in which the seam becomes dull towards the top of the seam is evident. In detail, the seam commences with a dull coal containing bright bands overlain by a carbonaceous siltstone. The carbonaceous siltstone is overlain by dull coal which comprises the upper half of the seam. A feature of the No. 2 seam in the study area, is the presence of a laterally persistent carbonaceous siltstone and mudstone bed in the middle of the seam. This clastic bed separates a lower brighter coal zone from an upper duller coal zone. A second feature of the No. 2 seam is the presence of two bright coal zones. A thick bright coal zone is present at the base of the seam and thin bright coal zone is present towards the top of the seam.

Comparison of the No. 2 seam coal profile of the study area with that of the Witbank Coalfield reveals a remarkable similarity (Cairncross and Cadle, 1988; Falcon, 1989; Holland *et al.*, 1989). Detailed petrographic analysis of 16 coal profiles throughout the Witbank Coalfield is presented by Falcon (1989). In general, the seam is dominated by vitrite and intermediate microlithotypes (up to 70%) at the base of the seam. High percentages (up to 75%) of fusite/semifusite and inertodetrinite are present towards the middle of the seam, and high vitrite and intermediate percentages (up to 70%) towards the top of the seam.

The dull-coal bands at the base and top of the seam are also evident in the profiles of the No. 2 seam in the Witbank Coalfield. These dull-coal bands comprise the microlithotypes inertodetrinite and fusite/semifusite. Minor percentages of vitrites and intermediates are also present in these bands. All 16 profiles analysed in the Witbank Coalfield by Falcon (1989) contain the microlithotype liptinite. In most instances liptinite comprises less than 3% of the seam and in certain bands reaches a maximum of 5% of the seam. It can be concluded from the petrographic data presented by Falcon (1989) that liptinite is a minor constituent of the No. 2 seam.

It is evident from the lithotype and petrographic data presented, that optimum conditions for the preservation of organic matter occurred during the early and late stages of No. 2 seam peat formation. These two periods of optimum peat formation coincided with a relative water table rise. This relative increase in the water table probably occurred rapidly, and is indicated by an abrupt increase in the amount of vitrite and intermediate microlithotypes at these two positions within the seam (Falcon, 1989; Holland *et al.*, 1989). The water table then gradually dropped and this is reflected by the decreasing vitrite and intermediate contents within the two bright coal zones. Within the study area, the presence of the laterally persistent carbonaceous siltstone and mudstone bed, situated at the middle of the seam, implies subsidence and inundation of basinal waters over the peat swamp. The result of the incursion of basinal waters over the peat swamp resulted in the termination of peat accumulation and the deposition from suspension of silt and mud. Renewed shallowing of the embayment allowed peat accumulation to recommence.

The No. 4 Lower seam displays a similar lithotype trend to that of the No. 2 seam. The seam is brighter at the base and becomes dull towards the top of the seam (Fig. 6.9). In the north of the study area, the seam commences with a bright coal with dull bands and is overlain by dull coal with bright bands and dull coal. A laterally persistent carbonaceous siltstone is present at a position approximating the middle of the seam. Overlying the carbonaceous siltstone is a dull coal with bright bands which grades upwards into dull coal. The dull coal is split in two by a second laterally persistent thin carbonaceous siltstone. In the west of the study area the No. 4 Lower seam thins towards and over, and pinches out against, basement highs. At these localities the coal bands interfinger with carbonaceous siltstone and mudstone. The coal is described as either dull or dull with bright bands.

The No. 4 Upper seam represents renewed peat accumulation following a period of braided-stream and crevasse-splay sedimentation which terminated No. 4 Lower seam peat accumulation. In general, there is a decline in the brightness of the coal upwards (Fig. 6.9). This follows a similar trend described for the No. 2 and No. 4 Lower seams. However, the No. 4 Upper seam does not contain as much bright coal as either the No. 2 or the No. 4 Lower seams. The basal three quarters of the seam is a dull coal with bright bands. This lithotype is overlain by dull coal.

The only published petrographic profile of the No. 4 seam is presented by Holland *et al.*, (1989) in the eastern Witbank Coalfield. The seam analysed is the No. 4 Lower seam. The seam is about 2m thick and described as a dull lustrous to dull coal. Two bright bands are present in the seam, the first towards the middle of the seam and the second at the top of the seam. With refsrence to the No. 4 Lower seam in the study area the two brighter coal bands documented by Holland *et al.*, (1989) may be correlatives of the two bright bands at the base and middle of the seam in the study area. Petrographically, the No. 4 seam in the Witbank Coalfield comprises about 60% inertinite group macerals and less than 20% vitrinite group macerals. The two brighter coal zones contain between 30% and 45% vitrinite group macerals.

With reference to the No. 4 Lower seam, two relatively short-lived periods occurred when optimal conditions for peat preservation existed. This was at the commencement of No. 4 Lower seam peat accumulation and again towards the top of the seam following a period of carbonaceous siltstone deposition. The No. 4 Lower and No. 4 Upper seams accumulated in an upper delta plain environment during a period of very slow subsidence. It is postulated that a depressed or fluctuating water table existed during peat accumulation. These water table conditions enhanced oxidation of the peat and resulted in the formation of an inertinite-rich No. 4 coal seam.

The No. 5 coal seam is distinguished by a significant content of bright lithotypes (Fig. 6.9). The No. 5 seam also displays a vertical trend from brighter coal at the base to duller coal towards the top of the seam. The basal metre of the seam comprises a zone of bright and dull interbanded coal. This coal zone is overlain by a zone of dull coal interbanded with bright and dull coal. Unfortunately at present, there are no published petrographic data of the No. 5 seam. It is assumed however, based on the bright lithotypes, that the seam would be dominated by vitrite and intermediate microlithotypes. Stratigraphically, the No. 5 seam overlies the No. 5 seam deltaic sequence and it is likely that during stabilisation of the delta plain optimum conditions for peat accumulation prevailed. Subsidence of the delta plain kept pace with peat accumulation ensuring a raised water table was maintained in the mire. This would promote reducing conditions such that minimal oxidation of peat took place. Hence, the No. 5 seam displays a large proportion of bright lithotypes.

The Nos. 1, 3 and 6 coal seams are similar in thickness and lithotype profiles (Fig. 6.9). All three seams are <1m in thickness and are distinguished by alternating bright and dull coal bands, and carbonaceous shale. The No. 1 and No. 3 seams display thin brightening-upward trends of carbonaceous shale, dull coal, and bright coal. This contrasts with the No. 6 seam; which consists of alternating carbonaceous shale and bright coal. In the east Witbank Coalfield, a 75cm thick petrographic profile of the No. 3 seam is documented by Holland *et al.*, (1989). The seam is classified as a vitrinite-rich coal seam. The basal 59cm of the seam comprises 60% vitrinite group macerals and 20% inertinite group macerals. The top 14cm of the seam comprises 20% vitrinite group macerals and 42% inertinite group macerals. Liptinite group macerals average 8% throughout the seam. Therefore, it is assumed that for the Nos. 1, 3, and 6 seams, the bright-coal lithotypes are rich in vitrinite-group macerals and the dull-coal lithotypes are rich in inertinite-group macerals.

All three seams are thin and accumulated within mires undergoing excessive subsidence. Excessive subsidence resulted in inundation of the mire, termination of peat accumulation and the deposition of carbonaceous siltstone. Raised water tables during peat accumulation promoted reducing conditions and the preservation of vitrinite-rich coal bands. A similar lithotype stratigraphy is present on a grand scale within the Volksrust Formation coal seams in the Waterberg and Springbokflats Coalfields (Falcon, 1986). These seams accumulated in grabens which were subjected to periodic subsidence which may account for the alternating shale and bright coal stratigraphy.

6.5 DISCUSSION: APPLICABILITY OF COAL DEPOSITIONAL MODELS

6.5.1 Introduction

Over many years, debate and speculation has ensued over the differences in coal type (maceral content) between Carboniferous coals of the Northern Hemisphere and the Permo-Carboniferous coals of the Gondwana continents. One of the striking differences between the Northern and Southern Hemisphere coals is the variation in abundance of vitrinite and inertinite macerals in these coals. In general, Carboniferous coals of North America and Europe contain higher abundances of vitrinite and lower abundances of inertinite than their Gondwana counterparts (Falcon, 1977). This has led to speculation that differences between the Northern Hemisphere tropical climates and Gondwanan cool climates account for the variation in coal type (Falcon, 1986; Moore 1987; Clymo, 1987). Other authors believe the vegetation differences are the primary control accounting for coal type, with luxuriant forests responsible for peat accumulation in the Northern Hemisphere, contrasting with broad leaved trees and bushes and herbaceous plants responsible for peat formation in Gondwanaland (Phillips and Peppers, 1984;

Falcon 1986). More recently, Hunt, (1989) and Hunt and Smyth, (1989) have argued that the variations in coal type can be accounted for through tectonic control rather than through differences in climate and vegetation.

6.5.2 Ash Content

The ash or mineral matter content in coals is derived from three sources. It occurs, as a function of the original source plants, as sediment precipitated and deposited during peat formation, and through diagenetic processes which take place during coalification (Andrejko *et al.*, 1983a; McCabe, 1984).

Proponents of the theory of ash material derived from plants accounting for most of the mineral matter in peats and coal are; Cecil *et al.*, (1980) and Renton *et al.*, (1980). From studies of the ash content of peat from the Okefenokee and Snuggedy swamps these authors conclude that the source of the minerals in peat is the inorganic material contained in living plants. Moreover, most of the minerals are silicates which form from the amorphous alumino-silicate material that dominate the inorganic portion of the woody tissue of plants. Furthermore, Renton and Cecil (1979) discuss that the abundance of ash in a peat layer is a function of the amount of plant debris accumulated and the rate of degradation. The rate of degradation is controlled by the pH of the swamp water. At pH values below 4 microbial activity becomes increasingly suppressed. Over pH 4 the degradation processes increase. Thus, a low rate of degradation (acid conditions) would result in a low-ash peat. Conversely, a high rate of degradation (alkaline conditions) would result in a high-ash peat as plant debris is rapidly degraded and inorganic material released into the system. In this situation Renton and Cecil (1980) postulate the formation of a thick high-ash coals. This is a powerful argument in favour of the formation of high-ash coals so prevalent in

Gondwana coals. The equivalent of the No. 2 seam mined at the Sigma and New Vaal Collieries in the Orange Free State contains ash contents of over 30 % (Boshoff and Schoeman, 1993).

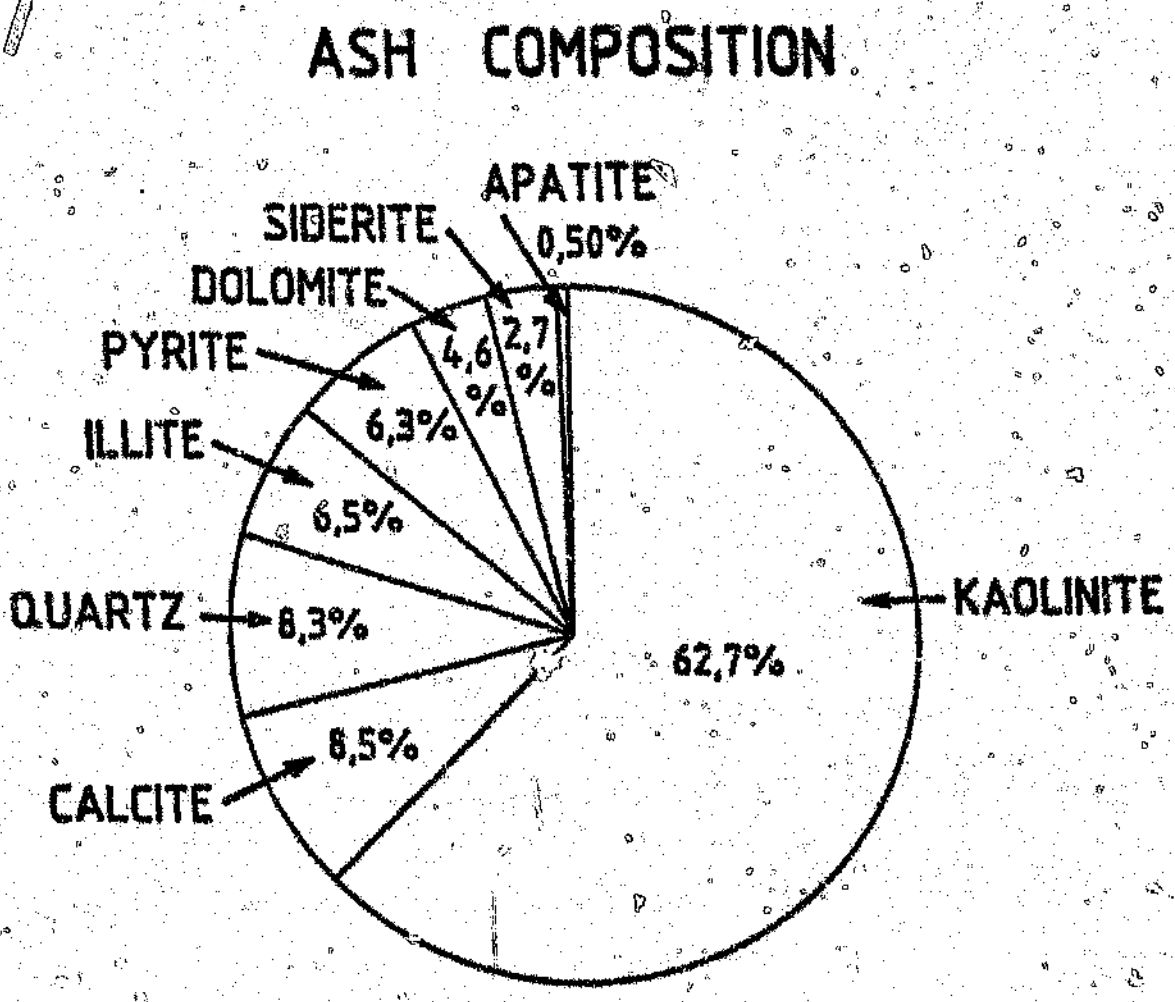
Andrejko *et al.*, (1983a) indirectly challenge the findings of Renton *et al.*, (1980) and Cecil *et al.*, (1980) based on their study of mineral matter examined from the peats of the Okefenokee, Snuggedy and Everglade swamps. All three swamps are dominated by the *Nymphaea* marshes. The Okefenokee is relatively free of non-siliceous minerals and dominated by authigenic, biologically derived, opaline silica. The Snuggedy swamp contains minerals derived from flood detritus and is dominated by clays and silt-sized particles. The Everglades are influenced by the inclusion of fresh-water authigenic carbonates. Andrejko *et al.*, (1983a) thus play down the importance of original vegetative depositional environments and believe an understanding of local and regional geology and the environment of the peat deposit can account for the mineral compositions within peat. They are firmly of the opinion that mineral matter in peat is derived from a combination of detrital, authigenic and diagenetic processes.

Reduction in the ash content during coalification from peat to coal is proposed by Kosters *et al.*, (1987). Peats of the Mississippi Delta contain average organic matter contents of 81.7% which imply high-ash values compared with Euramerican Carboniferous coals. Leaching experiments undertaken by Kosters and Bailey (1986) removes between a fifth and a third of the ash weight, suggesting a loss in inorganic material in early post-burial diagenesis. Andrejko *et al.*, (1983b) corroborate this view, and postulate that silica mobilisation is enhanced in the presence of low molecular weight organic acids generated during the humification process.

Cadle *et al.*, (1989a) quantify the amount and type of ash minerals in the No. 2 coal seam in the Witbank Coalfield. Two hundred and three samples were analysed and the

results are portrayed in Fig. 6.10. The ash content of the No. 2 seam, calculated from a composite of all samples, averages 23.23%. By far the most significant mineral constituting the ash content of the coal is kaolinite. Kaolinite averages 14.57% of the coal and comprises 62.7% of the ash content of the coal. Together with illite (1.51%) these two clay minerals comprise 69.2% of the ash minerals. The next most abundant mineral group are the carbonates, comprising calcite, dolomite and siderite which account for 15.8 % of the ash fraction. The remaining 15.1% of the ash consists of quartz (8.3%), pyrite (6.3%) and minor apatite (0.5%).

How is the kaolinite content of the No. 2 seam reconciled with the theory proposed by Renton and Cecil (1980)? In this theory, at low pH values amorphous silica undergoes dissolution and the dominant clay type to form would be kaolinite. At low pH values plant degradation would be inhibited and a low-ash kaolinite-rich peat would accumulate. This theory is not supported by the ash data from the Witbank Coalfield where high-ash, kaolinite-rich coals have formed. It is proposed that the source of the kaolinite is derived from detrital contribution, transformation of other minerals, or through precipitation from gel. It is well documented that both the No. 2 and No. 4 seams are split by braided-fluvial channels and crevasse-splay deposits, implying contemporaneous sedimentation during peat accumulation (Le Blanc Smith, 1980; Winter, 1985; Cairncross, 1986, Cadle *et al.*, 1993). Thus, there is certainly a detrital clay-mineral component added to the mire during peat accumulation. This is also supported by the association of clay minerals with inertodetrinite (Falcon and Snyman, 1986). The presence of clay minerals filling tellinite cell cavities, and occurring as small aggregates in vitrinite (Falcon and Snyman, 1986), supports the theory of neoformation (precipitation from gel) or kaolinite formation from vegetation at low pH values (Renton and Cecil, 1979). This process would take place during early diagenesis (Spears, 1987). Furthermore, the presence of small quantities of illite may suggest that kaolinite formed from transformation or alteration of detrital muscovite to



MINERALS	AVERAGE ASH%
KAOLINITE	14,57
CALCITE	1,97
QUARTZ	1,92
ILLITE	1,51
PYRITE	1,47
DOLOMITE	1,07
SIDERITE	0,63
APATITE	0,11
TOTAL	23,23

Figure 6.10. Average ash analyses of the No. 2 coal seam from the Witbank Coalfield (after, Cadle *et al.*, 1990).

illite (Rimmer and Erberl, 1982; Staub and Cohen, 1978). The abundance of inertinite macerals in these coals (up to 60 % (Falcon, 1977)) suggests high rates of degradation which would have provided void space in the peat within which authigenic kaolinite could precipitate.

The pyrite content of less than 1,5% is present in the No. 2 seam coal (Cadle *et al.*, 1989a). Compared to Euramerican coals the sulphur content in the coal is low. Casagrande (1987) maintains that all sulphur in coal can be accounted for in the peat-forming stage. Conditions which enhance the formation of pyrite and organic sulphur are; an anaerobic environment suitable for microbial sulphate reduction, ferrous iron, and peat organic matter. The fact that in general these coals are inertinite rich (Falcon, 1977) suggests oxidising conditions prevailed periodically, thus inhibiting the formation of high sulphur coals so prevalent in the vitrinite-rich Carboniferous coals of Euramerica (Teichmuller, 1989). From the analysis of the depositional systems, the No. 2 and No. 4 coal seams accumulated in fresh-water environments on fluvial flood plains. This environment is far removed from marine influenced peats where the sulphate content of the water provides the sulphate, a principal reactant for establishing pyritic and organic sulphur levels in peat (Casagrande, 1987).

The high carbonate content in the No. 2 seam (15.8% of the ash fraction) could be accounted for in a similar manner as proposed by Andrejko *et al.*, (1983a) for the carbonates present in the peats of the Everglades. The mechanism of carbonate precipitation is enhanced in fresh waters through the metabolic activities of fresh-water mollusks and colonies of blue-green algae. Carbonates are also formed during diagenesis whereby HCO_3^- is derived from organic fermentation and decarboxylation reactions in coal. The cations are derived from the diagenetic modification of detrital minerals such as the illidization of smectite and albitization of plagioclase (Spears, 1987).

6.5.3 Climate and Flora

Teichmuller (1989) is of the opinion that geologic age, palaeoclimate and associated ve ation have a significant role to play on the influence of coal facies. This state of affairs is largely due to the differences in coal facies associated with vitrinite-rich Carboniferous Euramerican coals, which accumulated in a tropical climate, compared to inertinite rich Gondwana coals, which formed in cool temperate climates.

Ziegler *et al.*, (1987) argue that consistent rainfall throughout the year and temperature as it effects evaporation are essential for peat formation. At the present time the Intertropical Convergence Zone has a width of about 10° latitude and is situated adjacent to the equator. This zone of the earth receives consistent rainfall throughout the year and it is here where tropical swamps occur. In terms of the reconstruction of Laurasia this zone coincides with abundant coal deposits in the Carboniferous (Ziegler *et al.*, 1981; Scotese and McKerrow, 1990). The Permian coal deposits of much of Gondwana are situated at palaeolatitudes greater than 50°S in a post-glacial to temperate climatic setting (Ziegler *et al.*, 1981; Falcon 1986; Scotese and Barrett, 1990). McCabe (1984) documents that many peats today accumulate in latitudes between 50°N and 70°N due to the low rates of peat decomposition which are favoured by cooler climates. Both Ziegler *et al.*, (1987) and Haszeldine (1989) suggest that coal deposits tend to be absent during periods of generally warm earth climate (Triassic to Palaeocene). These authors suggest that a warm hothouse earth produces climates with highly seasonal unreliable rainfall whereas a cold icehouse climate produces compressed climatic zones. This latter type of climatic setting experienced at present produces regular rainfall in the tropics and at the mixing zone of the tropical and polar air masses north and south of 50°N. This line of reasoning presented by Ziegler *et al.*, (1987) and Haszeldine, (1989), would account for the presence of abundant coal in Laurasia during the Carboniferous and in Gondwana during the Permian.

Variation in palaeoflora due to climatic change has been sited as a major factor in controlling coal type (Snyman, 1961; Falcon 1989 and Teichmuller, 1989). The important vegetation of Euramerican Carboniferous coals were the lycopod forests with subordinate abundances of pteridosperms, ferns and cordaitean trees. Phillips and Peppers (1984) reviewed coal ball studies and palynological investigations, and concluded that these coal swamps accumulated in a moist tropical climate during the Upper Carboniferous which experienced both wetter and drier periods. Lycopods dominated during the Westphalian which gave way to tree ferns during the Stephanian (Late Pennsylvanian). Although both the Westphalian and Stephanian experienced both wetter and drier climates the Stephanian experienced a generally drier climatic regime than during the Westphalian.

The vegetation of the Permian Gondwanan coals differed from the luxuriant flora of the Carboniferous forest swamps with tall tress. The Gondwana vegetation was stunted comprising herbaceous plants with shrub like broad-leafed trees and bushes of *Glossopteris* and *Gangamopteris* (Plumstead, 1966). Teichmuller (1989, p.44) is of the opinion that this is the primary reason that Gondwana coals are vitrinite poor. The Vryheid Formation coal swamps contain an extremely diversified and mixed assemblage of *Glossopteris-Gangamopteris* flora. These floras are characterised by abundant arborescent and herbaceous Glossopteridophyta, with minor components of lycopods, ferns, cordaites, and early gymnosperms. Based upon microfloral evidence the base of the coal measures are dominated by monosaccate-rich flora comprising *Virkkipollenites*, *Plicatipollenites* and *Potonieisporites* which infer a gymnospermous parent macroflora.

The monosaccate macroflora were rapidly replaced by nonstriate, disaccate-rich microflora (*Sulcatisporites*) comprising a highly varied suite of spores and pollens (Falcon, 1989). These floral changes reflect changing macroclimates and migration of vegetation belts with time. Falcon (1986) documents a cold-temperate macroclimate

which corresponds to the monosaccate microflora during which the Nos. 1 and 2 seam peats accumulated. The climate then ameliorated and changed from cold to cool/warm temperate. This climatic change corresponds to the dominance of nonstriate, disaccate-rich microflora during which the Nos. 4, 5, 6, seam peats accumulated. Falcon (1989) hypothesises that during cold-temperate climates, with short growing seasons plant growth and accumulation is slow and degradation of plant matter inhibited. The presence of high proportions of structured and partially gelified inertinites in the No. 1 and No. 2 coal seams are used by Falcon (1989) to support this hypothesis.

In contrast to the reasoning proposed by Falcon (1989), Hunt (1989) plays down the role of vegetation in controlling inertinite abundance in Permian coals. Palynological studies of Australian Permian coals reveal that glossopterids, other pteridosperms and the pteridophytes (true ferns) are co-dominant (Gould, 1975; Foster, 1979; McMinn, 1984). From these data, Hunt (1989) concludes that the flora provided a more or less heterogeneous input of woody material into Permian mires and, thus, flora is not regarded as a major control of inertinite type.

Collinson and Scott (1987) take a more philosophical approach when discussing the relationship between plant type and coal macerals. They question the conclusions drawn by Cohen and Spackman (1977, 1980) and Cohen *et al.*, (1987) concerning the deduction that the nature of the plant material forming peat will effect coal type as revealed by coal petrology. An example is presented whereby comparable lithotypes could be produced by very different plant communities e.g. exinite-rich coals could form from either, herbaceous fern or lycopodsid communities or, aborescent fern or lycopsid communities. Similarly, the "woody tissues" forming vitrinite may be derived from aborescent lycopsids, pteridosperms, cordaites or conifers and from bark or

secondary xylem of these plants. Collinson and Scott (1987) conclude that there need be no simple relationship between coal maceral or lithotype and an ancient plant community.

In conclusion, it is suggested that a more holistic view of the factors influencing coal type, including climate and flora, should be taken into consideration when determining those variables which control coal type.

6.5.4 Vertical Succession of Coal Facies

The vertical succession of coal facies, obtained from seams greater than a metre in thickness, displays a trend of bright-coal lithotypes at the base of seams which gradually pass upwards into dull-coal lithotypes with high-ash contents at the top of the seams. This trend is evident in the Nos. 2, 4 and 5 coal seams. In addition, these seams all contain a thin bright coal band towards the top of the seam. This trend is also well documented for the No. 2 and No. 4 seams in the Witbank Coalfield by Falcon (1989) and Holland *et al.*, (1989).

The trend of bright-coal lithotypes at the base of a seam passing upwards into dull-coal lithotypes is not unique to South African coals. Similar trends are evident from the major Carboniferous and Permian coalfields of the world. Hoffmann (1933) described the lithotypes and microlithotypes of the many seams in the Ruhr Coalfield of Germany. The base of the seams described, comprise thick bands of vitrite and liptinite-poor clarite, forming bright-coal bands. The vitrite content decreases upwards and the bright-coal bands are followed by an increase in semi-dull and dull bands consisting of liptinite-rich clarites and durite). Also documented is that, in many seams,

the amount of bright coal again increases towards the top of the seam. This sequence of coal is interpreted as a development from drier (forests) to wetter (marshes) peat-forming environments.

A similar coal seam profile to that described by Hoffmann (1933) was documented by Smith (1962, 1968) for the English Carboniferous coals. The seams commence with coal that is rich in vitrites and clarites, followed by trimacerites and finally durites rich in macrinite. This sequence is interpreted as evolving from, an initial wet rheotrophic forest swamp through, a less wet mire with open vegetation to, a relatively dry ombrotrophic high moor (raised bog). The Canadian Carboniferous coal seams described by Hacquebard and Donaldson (1969) have a similar vertical sequence when compared to those coal seams described by Hoffmann (1933) for the Ruhr Coalfield. The lower parts of the seams are rich in vitrites and clarites thought to be deposited in wet-forest swamps. The upper parts of the seams contain more dull layers comprising spore-rich clarites durites and clay partings deposited in reed marshes and open water. Like Hoffmann (1933), Hacquebard and Donaldson (1969) postulate a deepening of the peat-forming environment with time.

Smyth and Cook (1976) also document the trend from bright to dull coal within seams for Australian Gondwana coals. Based upon statistical methods, Smyth and Cook (1976) found that the vitrite plus clarite content decreases up the seam and that vitrite and clarite rich coal is the most likely lithotype to succeed a shale band. The top of the seam contains predominantly inertinite-rich bands and are rich in mineral matter. This type of coal profile is similar to the coal profiles described from the study area (Fig. 6.9), and for the No. 2 and No. 4 coal seams in the Witbank Coalfield described by Falcon (1989) and Holland *et al.*, (1989). The greater content of inertinite-group macerals towards the top of these coals implies the peat swamps underwent oxidation and became drier as the mire evolved.

There are diverse opinions on the interpretation of mire evolution with time. The interpretation by Smith (1962, 1968) on the evolution Carboniferous mires of England suggests that peat accumulation commences within wet-forest swamps which become relatively dry with time. This interpretation contrasts with the peat mire evolution presented by Hoffmann (1933) and Lutke (1985, 1987) for the evolution of the Ruhr Coalfield swamps and Hacquebard and Donaldson (1969) for the Canadian coals. These authors contend that peat accumulation commenced within wet-forest swamps and that the mire waters deepened such that peat accumulation terminated within reed marshes and open water.

Ecologists such as Walker (1970) and Tallis (1983) and Moore (1987) have demonstrated an evolution of mire type with time based upon studies of raised mires in northern Europe and eastern North America. The mires are initially rheotrophic and evolve into ombrotrophic bogs. Thus, peat initially accumulates within rheotrophic mires as swamps (aquatic emergent plants) which evolve into fens (dry-season water table below peat surface) and carrs (wooded swamp). These mires are then superseded by ombrotrophic raised mires which are bog peats. Ecologists have also speculated that the low-ash Carboniferous coals accumulated as peat under ombrotrophic conditions where organic material with a low-ash content prevails (Moore, 1987). This assumption has been supported by the work of Clymo (1987), McCabe (1987), Fulton (1987) and Bartram (1987) in their studies of peat and Carboniferous coal seams in Euramerica.

The rheotrophic to ombrotrophic evolution of mires is not supported by the data presented for South African and Australian Gondwana coals (Smyth and Cook, 1976; Falcon, 1989). These coals are rich in vitrite and intermediate microlithotypes at the base of the seams and change upward into coals rich in inertinite and carbominerite microlithotypes. The fact that these coals are rich in inertinite macerals implies that the mires were subjected to fluctuating water tables so enhancing the oxidation of peat. The

high carbominerite content towards the top of the seams suggests that the mires were not raised or domed but low lying. Thus, during floodings of the mires, mineral matter was deposited as peat accumulated. The petrographic data presented by Falcon (1989) and Holland *et al.*, (1989) from the coal seams in the Witbank Coalfield would suggest that the Vryheid Formation coal-measure peats accumulated in rheotrophic mires which became drier with time.

It is postulated that the No. 2, No. 4 and No. 5 coal seams commenced peat accumulation in rheotrophic mires with high water tables. With time, fluctuations of the water table allowed frequent periods of oxygenated conditions to exist in the peat swamp. The oxygenated conditions promoted microbial decomposition of the peat. This created void space in the peat and allowed authigenic and diagenetic minerals to precipitate. Furthermore, periodic flooding of the mires enabled the deposition of fine-clastic material during peat accumulation. Hence, these seams are vitrite and inertite rich at the base of the seams and become inertite and carbominerite rich towards the top of the seams.

These data presented for South African and Australian coal measures are in conflict with the interpretation of rheotrophic to ombrotrophic evolution of present-day mires and the interpretation of many low-ash Carboniferous coals as representing peat accumulation in ombrotrophic peat swamps. The problem of the recognition, properties and significance of coals deposited in rheotrophic or ombrotrophic mires still remains to be satisfactorily resolved.

6.5.5 Depositional Systems

Depositional systems and their relationship to coal seams commenced in the 1950's with the need by the major oil companies to better understand ancient sediments. The Yoredale cyclothems and associated coal seams were interpreted by Moore (1959) as deltaic deposits. In the United States, the work of Fisk (1960), Coleman and Smith (1964), and Frazier and Osanik (1969) on the Mississippi Delta misleadingly led workers to link coal-forming environments with deltaic depositional systems. This trend was followed in the 1970's by the work of Ferm (1970) and his co-workers on the coal deposits of the Appalachian Basin. These studies culminated in the benchmark paper by Horne *et al.*, (1978) which once again stressed the role of deltaic systems and coal accumulation.

However, since these studies coal deposits have been reported from a variety of depositional systems many of them terrestrial settings. Some examples of coal and peat deposits recorded from a variety of depositional systems are listed:

a. alluvial fan depositional systems documented from northern Spain (Heward, 1978) and the Canadian Cordillera (Long, 1981).

b. braided, meandering and anastomosed fluvial systems (Haszeldine and Anderton, 1980; Kaiser *et al.*, 1980; Smith *et al.*, 1989; Cadle *et al.*, 1993).

c. lacustrine depositional systems of the Tertiary basins of the Rocky Mountains (Ayers and Kaiser, 1984; Flores, 1989).

d. back-barrier depositional systems of comprising Carboniferous and Cretaceous coal-bearing deposits of the United States (Hobday, 1974; Ferm and Horne, 1979; Marley *et al.*, 1979).

Based upon the sedimentological investigations of coal-bearing depositional systems, coal depositional models have been postulated. One of the most rigorous models, based upon surface and subsurface sedimentological data, is presented by Horne *et al.*, (1978)

for the Carboniferous deposits of the Appalachian Basin. This research postulated that, back-barrier coals are thin, laterally discontinuous, high in sulphur and exhibit poor roof conditions. Lower delta plain coals are widespread, thin, and have an irregular sulphur and trace element distributions. Upper delta plain "fluvial" coals are thick, laterally discontinuous and are low in sulphur. Horne *et al.*, (1978) postulate that economic coals form in the transition zone between the lower and delta plains where the coals are thick, laterally continuous and low in sulphur. Fenn and Weisenfluh (1989) have modified the above model by suggesting that tectonics both on a basin-wide scale and locally plays a role in determining coal seam thickness.

Coal depositional models were challenged by McCabe (1984; 1987; 1992) who has argued, based upon data from Mississippi Delta peats, that peats accumulating adjacent to active sedimentation would, if preserved in the ancient record, ultimately form high-ash coals or carbonaceous shales. The alternative model proposed is that peat forms raised mires. These mires are situated above flood levels and do not receive water-transported mineral matter. This process of mire evolution would promotes the formation of low-ash coals. Moreover, McCabe (1987) reasons that coals accumulate as peats in mires well removed from active clastic sedimentation.

Based upon the documented studies of economic coals and associated depositional systems, it appears evident that peat can accumulate in most terrestrial and transitional environments where raised water tables are present. This is because external factors to the depositional system such as climate, subsidence, basement topography and sea level also influence the position of the water table in peat mires. It is speculative therefore, to postulate that a particular depositional environment as an optimum setting for peat accumulation.

With respect to this study, the No. 2 and No. 4 coal seams are thick, inertinite-rich, high-ash coals which accumulated in fluvial settings and are laterally continuous over 1,000's km² (Cadle *et al.*, 1993). In addition, the Nos. 3, 5 and 6 coal seams were deposited in deltaic settings and are thin, vitrinite rich, laterally impersistent, and lower in ash, than the No. 2 and No. 4 seam fluvial coals (Holland *et al.*, 1989; Cadle *et al.*, 1993). Thus, only part of the Appalachian coal depositional models apply to this study namely, the relationship of coal-seam thickness related to fluvial and deltaic environments. In contrast to the Horne *et al.*, (1978) model, the fluvial coals of this study are laterally extensive and deltaic coals laterally impersistent (Figs. 2.19, 2.22, 2.23, 2.32, 2.34). The concept that raised peat swamps are an important factor in promoting low-ash coals and preventing clastic incursions into the peat swamp (McCabe 1984, 1987 and 1992) does not appear to have played an important role in the accumulation of coal in the Witbank and Highveld Coalfields. The coals in these coalfields are generally high in ash (Boshoff and Schoeman, 1993) suggesting that peats can accumulate as low lying swamps over 1000's km². In addition, the small- and large-scale clastic partings within the No. 2 and No. 4 seams imply periodic clastic invasion of the peat swamp by both crevasse-splay and fluvial sediments (Figs. 2.13, 2.21)

The questioning by Ferm and Weir *et al.* (1989) that both depositional systems in combination with tectonic setting and subsidence could play a part in promoting the development of economic coals has relevance to this study. The role of tectonic setting and subsidence in promoting the formation of economic coal deposits is discussed in the following section.

6.5.6 Tectonics and Subsidence

Environmental wetness, determined by the water-table level, is the most important factor which controls mire formation and peat accumulation (Gore, 1983). The level of the water table in the mire also controls the degree of organic preservation (Stach *et al.*, 1982). Where subsidence is balanced by organic production optimal conditions for peat aggradation are achieved. Rapid subsidence implies a rapid rise in the water table and a wet environment. The balance between a terrestrial waterlogged environment and a flooded marine environment depends upon the difference between subsidence and the sediment accumulation rate (Jervey, 1988).

In the study area, the Nos. 1, 2, 4, 5, and 6 coal seams are located at the top of depositional sequences (Cadle *et al.*, 1993), or parasequences (Van Wagoner *et al.*, 1988), or depositional events (Galloway and Hobday, 1983). In each instance, the coarsening-upward clastic succession represents the progradational phase of a parasequence. The coal seams and associated clastic sediments represent the aggradational phase of a parasequence. In terms of the principles of sequence stratigraphy, and evidence from the No. 2 and No. 4 coal seams (Figs. 6.2A and 6.4A), the aggradational phase of a parasequence (coal seams) thickens towards the basin margin and is a function of the amount of time a parasequence experiences aggradation. The coals are considered to be dominantly regressive coals accumulating as the aggradational phase on upper and lower delta plains. The coals are not classified as transgressive coals, as the transgressive event is usually a geologically rapid event coinciding with a rapid rate of accommodation increase in the basin (Jervey, 1988; Posamantier *et al.*, 1988).

The coals of the Karoo Basin have accumulated in a foreland basin (Cadle *et al.*, 1993). The foreland basin setting is an obvious basin style in which economic coals accumulate and are preserved e.g. the Appalachian Basin of the United States

(Tankard, 1986; Ferm and Weisenfluh, 1989), the Sydney and Bowen Basins of Australia (Jones *et al.*, 1984; Hunt, 1989), the Saar-Nahe and Ruhr Basins of Germany (Diessel, 1992; Drozdzewski, 1993) and the Rocky mountain foredeep of Mesozoic and Tertiary time (Bustin and Smith, 1993; Flores and Hanley, 1984; Tankard, 1986). A major difference between the coal accumulations in the Vryheid Formation and the coals present in the other foreland basins, is the position of the coal seams with respect to the basin margin. The Karoo coal seams are located on the cratonic margin of the foreland basin whereas the coals of the foreland basins referred to, are located on the orogenic margin of the foreland basins (Fig. 6.11). It would be expected therefore, that over a given interval of time, the amount of subsidence on the cratonic margin of a foreland basin would be less than the amount of subsidence on the orogenic margin of the foreland basin.

It is proposed that the tectonic regime within which peats accumulate influence the general character of the coals in terms of coal type, and mineral content. Hunt (1989) and Hunt and Smyth (1989) present quantitative data to show that coal type varies with respect to the Permian coal basins of eastern Australia. Average vitrinite contents of the coal seams decrease from 80% to 60% across the outer and inner foreland basin (Sydney and Bowen Basins). The cratonic basins (Galilee and Cooper Basins) have vitrinite contents between 40% and 60%. A cross-section through the Cooper, Galilee and Bowen Basins (Hobday, 1987) suggests that the former two basins are tectonic remnants of a large Permian foreland basin. The Cooper and Galilee Basins are considered to represent the sedimentary remnant on the cratonic margin of a large foreland basin and hence are equated in terms of basin style with the cratonic margin of the Karoo Basin. Hunt (1989) also documented the proportion of structured to unstructured inertinite called the semifusinite ratio in the Permian coal basins. Again a spatial variation in the semifusinite ratio is present. The semifusinite ratio decreases from the depocenter (Sydney Basin) through the outer foreland basin to the Cooper

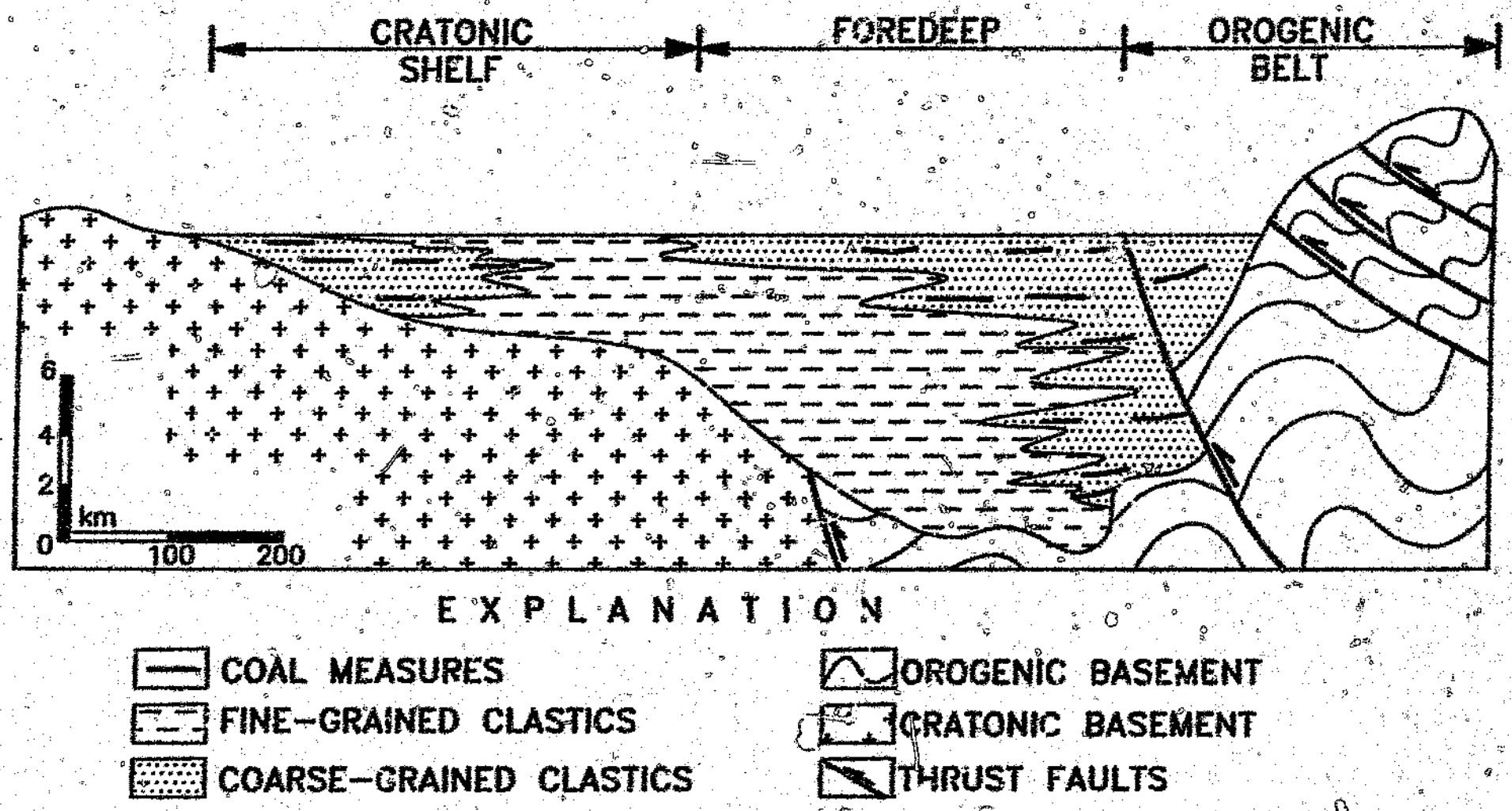


Figure 6.11. Cross-section through a foreland basin. Permo-Carboniferous coals form on both the orogenic and cratonic margins of foreland basins. Coals which form on the orogenic margin of these basins are generally vitrinite rich and coals which form on the cratonic margin of these basins are generally inertinite rich.

Basin. This decrease corresponds to a decrease in sediment accumulation rate (m/Ma) from 300 in the depocenter to 15 in the Cooper Basin. Moreover, band width, which measures the degree of interlayered mixing of vitrinite and inertinite macerals, varies spatially. The coarsely banded coals are present in the rapidly subsiding foreland basins and the more finely banded coals are present along the cratonic margin (Hunt, 1989, Fig. 16).

Thus, given a sufficiently humid climate, regional tectonic subsidence may be the most important factor controlling coal type in many coal basins. These data presented by Hunt (1989) imply that in sedimentary basins where peat accumulation takes place, the water table remains high in those parts of the basin undergoing rapid subsidence. This condition promotes the preservation of organic material and hence vitrinite-rich coals ultimately form. On those parts of the basin undergoing relatively slow subsidence water-table fluctuations are pronounced enhancing oxidation of organic material and hence inertinite-rich coals accumulate. Small-scale fluctuations in the water table are also accentuated under these tectonic conditions and this is reflected by the presence of finely banded coals. Based upon the observation that inertinite-rich coals are associated with high-ash contents it is postulated that during oxidation of peat, voids are created in the peat which are filled by both authigenic and diagenetic minerals. With respect to the No. 2 seam in the Witbank Coalfield, the "inertinite-rich" coal bands have a higher ash content than "vitrinite-rich" coal bands (Cadle *et al.*, 1989b).

Differential subsidence on a local scale may also influence coal type. Both the No. 3 and No. 5 coal seams are described as either bright or bright and dull interbanded coal seams. The seams display low-ash contents, and relatively high vitrinite contents and calorific values, when compared to the No. 2 and No. 4 coal seams (Holland *et al.*, 1989; Boshoff and Schoeman, 1993). From the analysis of the depositional environments both the No. 3 and No. 5 coal seam peats accumulated on delta plains and are underlain by prodelta and delta front shales. The No. 2 and No. 4 seams

accumulated upon alluvial plains comprising thick sequences of sandstones and granule conglomerates. Compaction of the underlying prodelta and delta front shales may have maintained high water-table levels within the No. 3 and No. 5 seam mires enabling the rapid preservation of organic matter which is reflected in the relatively high vitrinite contents of these seams. Holland *et al.*, (1989) have quantified the average vitrinite percentages of the Nos. 2, 3 and 4 coal seams from the same position in the Witbank Coalfield. Average vitrinite values for the Nos. 2, 3 and 4 coal seams are; 17%, 54% and 11% and for inertinite the values are; 73%, 30% and 67% respectively. These data would support the contention that differential compaction on a local scale may influence the abundance of vitrinite and inertinite in coals.

Differential subsidence on a regional scale has also taken place in the northern Karoo Basin influencing both coal type and thickness (Cadle *et al.*, 1993). The influence of differential compaction on coal type and thickness for the No. 2 and No. 4 coal seams is marked. When tracing the coal seams from the Orange Free State through the Transvaal Coalfield to the Natal Coalfield , the coal seams thin (Cadle *et al.*, 1993, Fig. 8), and become vitrinite rich (Cadle *et al.*, 1993, Fig. 5). The relatively high vitrinite content of the coals from the Natal Coalfield is attributed to the differential compaction of the peat swamp over a thick sequence of underlying sediments. Subsidence of these sediments maintained a raised water table during peat accumulation and enhanced the preservation potential of organic matter. Continued compaction of sediments coupled with greater basin subsidence caused local transgressions and the termination of peat accumulation. Hence, coal seams in the Natal Coalfield are frequently less than 2m in thickness.

Basin tectonics, a wet climatic regime and the degree of organic decomposition are considered the primary regional controls on coal type i.e. vitrinite- and inertinite-rich coals. The Vryheid Formation coals are classified as inertinite rich in comparison with the Permo-Carboniferous coals situated in Ruhr, Appalachian and Sydney Basins

(Falcon, 1977; 1986). This implies that organic matter was subjected to oxidising conditions and high rates of microbial organic decomposition during peat formation. These Permian peat swamps accumulated on the cratonic margin of the Karoo foreland basin. This tectonic setting promoted comparatively little subsidence of peat swamps, which allowed for the establishment of well drained swamps containing oxygenated waters, thus exposing the organic material to microbial activity. Hence, on a regional scale, the Vryheid Formation coals are inertinite rich. Variations in vitrinite and inertinite contents of the coals are attributed to subtle differential subsidence of the peat swamps. With respect to the Witbank and Highveld Coalfields, the No. 2 and No. 4 coal seams are inertinite rich and accumulated on fluvial flood plains which experienced relatively low rates of subsidence. The Nos. 1, 3, 5 and 6 coal seams however, represent vitrinite-rich coals which accumulated as peat on delta plains subjected to increased rates of subsidence. The vitrinite-rich coals of the Natal Coalfields accumulated in peat swamps that formed in a tectonic regime that experienced relatively high rates of subsidence (Cadle *et al.*, 1993).

CHAPTER 7

7 SUMMARY AND CONCLUSIONS

7.1 INTRODUCTION

The sedimentary rocks of the Dwyka Group and Vryheid Formation, situated in the Highveld Coalfield of the Karoo Basin, are analysed to determine the depositional systems that were active during sedimentation. The rocks of the Vryheid Formation contain coal seams thus, having determined the depositional systems, the relationships between depositional systems and coal thickness and distribution are deduced.

The study involved the processing a total of 629 borehole cores and logs over an area of 860km². These data provided a three-dimensional stratigraphic framework for analysis. The analysis of these data permitted the compilation of a detailed stratigraphy and the determination of the three-dimensional distribution of stratigraphic units. The identification of sedimentary facies, and their lateral and vertical association within specific stratigraphic units, enabled the interpretation of glacial, fluvial, deltaic and shallow-marine sedimentary depositional systems.

The linking of sedimentary depositional systems to a well defined stratigraphic framework facilitated the compilation of the depositional architecture through time. The relationships between depositional systems and coal seams have allowed for conclusions to be made on what factors influence the distribution, thickness and coal lithotypes of the coal seams within the study area. Comparison of the tectonic setting of the Karoo Basin coal seams with Permo-Carboniferous coal-bearing sequences from other basins

has led to the postulation that basin tectonics may influence coal type on a regional scale.

7.2 STRATIGRAPHY

Based upon the need to formulate a stratigraphy that enables detailed analysis of depositional systems and their associated coal seams, a genetic stratigraphy for the Highveld Coalfield is constructed. A composite stratigraphic column for the study area is compiled which illustrates the variation in stratigraphy encountered within the study area (Fig. 2.4). The subdivision of the stratigraphy into more detailed units is based upon depositional sequences (Cadle, 1982b) or parasequences (Van Wagoner *et al.*, 1988). The depositional sequences comprise a progradational, or shallowing-upward, phase of sedimentation followed by an aggradational phase of sedimentation. The progradational phase of sedimentation comprises siltstones which coarsen upward into sandstones and granule conglomerates. The aggradational phase of sedimentation is represented by siltstones, fine-grained sandstones and coal. Each depositional sequence is separated from the succeeding depositional sequence by a thin veneer of transgressive sediments or, a marine-flooding surface (Van Wagoner *et al.*, 1988). Within the study area five depositional sequences are delineated. They are informally termed from the base upwards; the No. 1 seam depositional sequence, the No. 2 seam depositional sequence, the No. 4 seam depositional sequence, the No. 5 seam depositional sequence and the No. 6 seam depositional sequence.

The No. 1 and No. 2 seam depositional sequences comprise diamictites, conglomerates, coarse-grained sandstones, and siltstones. The sequences thin and pinchout against elevated basement in the west and thicken towards the east and southeast. Each sequence is capped by a coal seam. The No. 1 seam depositional

sequence is present only in the southeast. However, the No. 2 seam sequence is present over much of the study area. The No. 1 coal seam is thin and laterally impersistent, and the No. 2 coal seam relatively thick and laterally persistent.

The No. 4 seam depositional sequence is subdivided into depositional intervals; the No. 2 to No. 3 seam depositional interval, and the interval between the No. 3 seam and the uppermost No. 4 seam, The No. 2 to No. 3 seam depositional interval comprises carbonaceous siltstones at the base overlain by fine- and medium-grained sandstones which are frequently bioturbated. The depositional interval thickens to the east and thins towards, and pinches out against, basement in the west. The No. 3 seam caps this interval and is thin and laterally impersistent. Overlying the No. 3 seam interval is a thin unit of siltstones and sandstones capped by the relatively thick, and laterally persistent, No. 4 coal seam. The No. 4 coal seam is complexly split by siltstones, sandstones and conglomerates such that the No. 4 seam is split into the No. 4 Lower, No. 4 Upper, and No. 4 A subseams. The No. 4 seam depositional sequence is terminated by a thin cover of glauconitic sandstone and siltstone.

The No. 5 and No. 6 seam depositional sequences display well defined coarsening-upward lithologies. The depositional sequences commence with carbonaceous siltstones and grade upward into fine- and coarse-grained sandstones. There is little influence of basement topography on the distribution and thickness of the sediments at this level in the stratigraphy. The Nos. 5 and 6 coal seams mark the termination of the respective depositional sequences. The No. 5 coal seam is relatively thin and laterally persistent whereas, the No. 6 coal seam is very thin (<50cm in thickness) and laterally impersistent. A distinctive unit of glauconitic siltstones and sandstones represents a marine transgressive event which separates the No. 5 from the No. 6 seam depositional sequence.

For the first time, an additional depositional sequence is recognised within the Witbank and Highveld Coalfields. This is the No. 1 seam depositional sequence. Previously, this interval of rocks was incorporated within the No. 2 seam depositional sequence (Le Blanc Smith, 1980; Winter, 1985; Cairncross, 1986). The recognition of this depositional sequence is made possible by the presence of carbonaceous siltstones which demarcate a marine-flooding surface separating the No. 1 and No. 2 seam depositional sequences.

7.3 SEDIMENTARY FACIES

Sedimentary facies identified from borehole cores are used, in combination with three-dimensional isopachs of facies assemblages, to identify depositional systems. A wide spectrum of facies types are present in the rocks studied. They range from diamictite, through conglomerate and sandstone, to siltstone facies types. Subdivision of these facies types is based upon sedimentary structures. Coal and limestone facies frequently, terminate depositional sequences. In total, 14 primary facies are identified.

The Dwyka Group, and Vryheid Formation have facies associations which are distinctive. The Dwyka Group is dominated by diamictites containing a variety of pebble and cobble-sized extrabasinal clasts. The Vryheid Formation is distinguished the coarseness of sedimentary facies, the carbon content of the finer grained facies types, and the presence of coal seams. The coarse-grained facies associated with the depositional sequences are typified by granule-grade conglomerates and very coarse- to coarse-grained sandstones. These facies imply a predominance of coarse-clastic material in the source terrain and a short distance of transport to the basin. The fine-grained facies are distinguished by the predominance of carbonaceous siltstones over claystones. The abundance of carbonaceous siltstones over claystones reflects the coarseness of sediment transported into the basin. The carbon content of the siltstones

implies they were deposited in reducing environments. The development of mires at the top of depositional sequences allowed for the accumulation of peat which resulted in the formation of economically exploitable coal seams.

7.4 DEPOSITIONAL SYSTEMS

The interpretation of depositional systems is based upon; the subdivision of the stratigraphy into depositional sequences, the three-dimensional mapping of stratigraphic intervals and gross lithological units, and the identification of lateral and vertical facies associations. This approach to the analysis of the stratigraphy has led to the interpretation of glacial, fluvial, deltaic, and transgressive shallow-marine deposits.

Facies interpreted as glacial deposits are largely associated with the No. 1 seam depositional sequence. These deposits are preferentially situated in palaeovalleys in elevated terrain, and in the topographically low-lying terrain in the southeast of the study area. The diamictite facies represents lodgement till deposited by wet-based glaciers. The facies association of diamictite overlain by granule conglomerates indicates deposition of the conglomerates by braided-fluvial systems in a glaciofluvial environment. In the extreme southeast of the study area, the facies association comprises diamictite overlain by a coarsening-upward association of siltstone and sandstone facies. This facies association documents glaciolacustrine deltaic sedimentation into a shallow water body.

The Nos. 1, 2 and 4 seam depositional sequences have facies associations interpreted as low-sinuosity, coarse-grained, bed-load dominated, fluvial systems. These facies associations are dominated by massively bedded and planar cross-bedded, granule conglomerates and sandstones. Cross-laminated sandstones and siltstones overlie these coarse-grained facies, and are present in minor abundances at the top of fining-upward

facies sequences. The fluvial systems within the No. 1 and 2 depositional sequences have a sheet-like geometry, were unconfined, tens of kilometres in width, and resembled braid plains. The fluvial systems present below, within, and above, the No. 4 coal seam display a linear geometry. These channel systems vary between 2km and 10km in width, are >35km in length and are both multistorey and multilateral in character. Away from the channel margins, the facies change to siltstones overlain by cross laminated sandstones representing crevasse-splay sedimentation. These crevasse-splay sediments were deposited in backswamp environments during periodic flooding. The restricted width of the fluvial systems associated with the No. 4 seam suggests that overbank sediments and vegetation contained the width of channel migration.

Facies associations interpreted as deltaic deposits are situated within the No. 2 to No. 3 seam interval of the No. 4 seam depositional sequence, and within the Nos. 5 and 6 seam depositional sequences. The facies associations from the No. 2 to No. 3 seam interval are interpreted as sediments deposited by river-dominated, shallow-water, mouth-bar deltas. The vertical facies association comprises from the base; carbonaceous siltstone and interlaminated sandstone-siltstone, which grades upwards into wave-rippled, cross-laminated and planar bedded sandstone. These facies were deposited by the progradation of deltas into water depths of about 10m. Outflow dispersion at distributary channel mouths was dominated by frictional and buoyant forces causing the deposition of sand to form mouth bars. The deltas have lobate geometries and mouth-bar sands are incised through by distributary channels with a channel depth to basin depth ratio of >0,5. Reworking of the sediments by low-energy wave action is evidenced by wave-ripple lamination and abundant bioturbation of the siltstones and sandstones.

A different style of deltaic sedimentation is interpreted for the facies associated with the No. 5 and No. 6 seam depositional sequences. The facies of these two depositional sequences are interpreted as shallow-water, fluvially-dominated "Gilbertian" deltas

subjected to periodic high-discharge events. Facies at the base of the deltas are similar to those of the No. 2 to No. 3 seam interval. However, the top of the delta sequences comprise trough and planar cross-bedded, coarse-grained sandstones. These sandstones fine upward into fine-grained, cross-laminated sandstones and siltstones overlain by [illegible]. The deltas prograded into estimated water depths of between 20m and 30m. Closely spaced distributaries flushed large quantities of bed-load sediment during periods of high discharge which favoured inertia-dominated deposition. During periods of low discharge, silt was flushed out through buoyancy forces onto the delta front. These thin silt drapes define Gilbertian foresets. Relatively high water depths, a high bed-load to total-load ratio, and fluctuating discharge promoting inertia-dominated turbulent jets at distributary mouths, caused rapid deposition of bed-load material. The combination of these factors which caused the rapid deposition of bed-load material, account for the development of Gilbertian foresets. Compared to the deltaic deposits of the No. 3 seam depositional sequence, the deltaic deposits of the Nos. 5 and 6 seam depositional sequences are not abundantly bioturbated. This is attributed to rapid rates of deposition inhibiting bioturbation of the sediment.

Transgressive shallow-marine deposits, overlie the Nos. 2, 4, 5, and 6 seam depositional sequences. Transgressive shallow-marine sediments are deposited in accommodation-dominated regimes. This occurs when sediment is entering the shelf in lesser quantities than can be accommodated by compaction or sea-level rise. These deposits comprise, a coarsening-upward facies association, a fining-upward facies association, and a well sorted sandstone facies association. The coarsening-upward facies association of glauconitic siltstones grading up into well sorted fine- to medium-grained sandstones represents the partial retention of microtidal, transgressive barriers. The fining-upward facies association of granule-grade conglomerates and well sorted glauconitic sandstones represents the reworking within the shoreface of fluvial deposits situated above the No. 4 Upper seam. The thick accumulation (15m) of well sorted, glauconitic, wave-rippled sandstones present above the No. 5 seam are