BASIN ANALYSIS OF THE BEAUFORT GROUP IN THE WESTERN PART OF THE KAROO BASIN

by

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Thesis submitted for the degree Philosophiae Doctor in the Department of Geology of the Faculty of Science, at the University of the Orange Free State, Bloemfontein

1990

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ABSTRACT

A basin analysis of the lower Beaufort Group in the Karoo Basin (Carboniferous-Jurassic) west of 24°E, cover an area of nearly 70000 km^2. The Karoo Sequence reaches a composite thickness greater than 7500 m, comprising diamictites, sandstones, shales and mudstones. Along the southern margin of the Basin the entire succession is intensely folded, becoming flat or gently undulating in the north. Minor displacements along dolerite dykes and small-scale thrusting are the only evidence of faulting in the Basin.

The transition from the uppermost Ecca to the lower Beaufort Group reflects a change from deltaic to fluvial depositional environments. The contact is taken above the uppermost deltaic sandstones, where the succession becomes distinctly argillaceous, and where therapsid fossils appear in the succession.

The lower Beaufort succession consists of alternating sandstones, and red and green mudstones forming cyclical, upwards-fining, fluvial sequences. Channel sandstones (greywackes and arkosic wackes) and intraformational conglomerates are overlain by red and green overbank mudstones with intercalated crevasse splay sandstone sheets and lenses. The mudstone facies (Fg, Fl, Fr and Fv) are commonly bioturbated and contain a variety of therapsid fossils, calcareous nodules, and palaeosols. Tuff beds are sparsely preserved. Freshwater molluscans, fish imprints and plant fossils occur at certain horizons.

The lower Beaufort comprises four upwards-fining megasequences, the uppermost incompletely preserved. The basal megasequence (A) overlies and grades diachronously into the Ecca Group from south to north across the Basin, thinning from more than 2500 m in the south to where it pinches out on surface, just south of Carnarvon in the north. In the unit which consists of a basal Combrinkskraal sandstone member overlain by the Leeu Gamka member, green and subordinate red mudstones are interbedded with thick, tabular sandstones. Megasequence B contains thick, tabular sandstones and green mudstones of the Koup member at the base, overlain by the predominantly red mudstones and thin sandstones of the Teekloof member. The unit is similarly wedge-shaped, thinning from nearly 900 m in the south to where it pinches out in the northeastern corner of the Basin. Megasequence C consists of a basal Nuweveld member, overlain by red mudstones of the Gifkop member. Megasequence D is sparsely preserved in the central part of the Basin, comprising the Leeukop member overlain by the Elandsberg Member. These megasequences correlate with similar stratigraphic units in the eastern sector of the Basin. Sandstone-hosted, syngenetic uranium orebodies occur in the arenaceous intervals of each megasequence.

The lowermost Beaufort shows northerly transport directions in the south, and subordinate, east-northeasterly directions in the central and northern parts of the Basin. These directions reflect transverse alluvial fan and longitudinal alluvial plain drainage systems, respectively. This pattern is repeated throughout the lower Beaufort, with the longitudinal drainage system becoming dominant in the uppermost units. Sandstone:mudstone ratio maps confirm the major transverse drainage system which entered the Basin from the south. Westerly to northwesterly transport directions in the easternmost part of the study area represent a separate, coeval drainage system in the eastern sector of the Basin.

The Beaufort Group was deposited as a molasse-type basin-fill in an orogenic foreland (retroarc) basin, yoked to a fold-thrust belt in the south. Isopach maps show that the Basin was strongly asymmetric initially, becoming more symmetric during deposition of the uppermost members. Sediment was derived from a mixed source, and uranium was introduced from weathered tuffs and granites. Uranium ore formation was controlled by the petrographic composition, presence of carbon, and permeability and diagenetic history of the host sands.

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1.0 INTRODUCTION

The Karoo Basin (Late Carboniferous - Early Jurassic) covers an area of approximately 600 000 km^2 and straddles all four provinces of the Republic of South Africa, the Kingdom of Lesotho, and the Republics of Transkei, Bophuthatswana and Venda. The central portion of the Basin in which the thickest succession of Karoo strata has been preserved, is referred to as the main Karoo Basin similar to the usage of the South African Committee for Stratigraphy (SACS, 1980). Karoo strata also occur in the neighbouring states Namibia-Southwest Africa, Botswana, Zimbabwe, Swaziland and Mozambique.

Geographically the Basin is bounded to the south and southwest by the Cape Fold Belt, and by the Indian Ocean to the east. The western and northern margins are erosional with small outliers of Karoo strata overlying older formations. Maximum thicknesses of Karoo rocks in the Basin occur immediately north of the Cape Fold Belt (Truswell, 1970). The present study area straddles this portion of the main Karoo Basin, situated approximately between 20°E and 24°E longitude and 30°45'S and 33°15'S latitude (FIG. 1.1). This geographic region is known as the Great Karoo.

The Dwyka Tillite Formation (Late Carboniferous - Early Permian) occurs at the base of the Karoo Sequence and succeeds the predominantly arenaceous Cape Supergroup conformably over much of the Basin, and unconformably only locally in the southern parts of the main Karoo Basin. The Cape Supergroup pinches out in the northern parts of the Karoo Basin, beyond where the Dwyka Formation directly overlies Precambrian basement. The Ecca Group overlies the Dwyka Formation conformably over most parts of the Basin, although Du Toit (1918) interpreted "stratigraphic breaks" and unconformities between these two units in the eastern half of the Basin. Regional stratigraphic relationships suggest a diachronous contact between the Dwyka Formation and Ecca Group (Visser, 1983).

The Ecca Group in the western Karoo Basin consists of alternating arenaceous and argillaceous successions, and is succeeded by the predominantly argillaceous Beaufort Group, with which it shows a marked diachronous relationship (Visser *et al.*, 1980; Jordaan, 1981). Only the lower Beaufort is present in the study area, and for the purposes of this dissertation the term refers to the "Lower Beaufort Stage" as used previously by Broom (1906), and the Adelaide Subgroup as

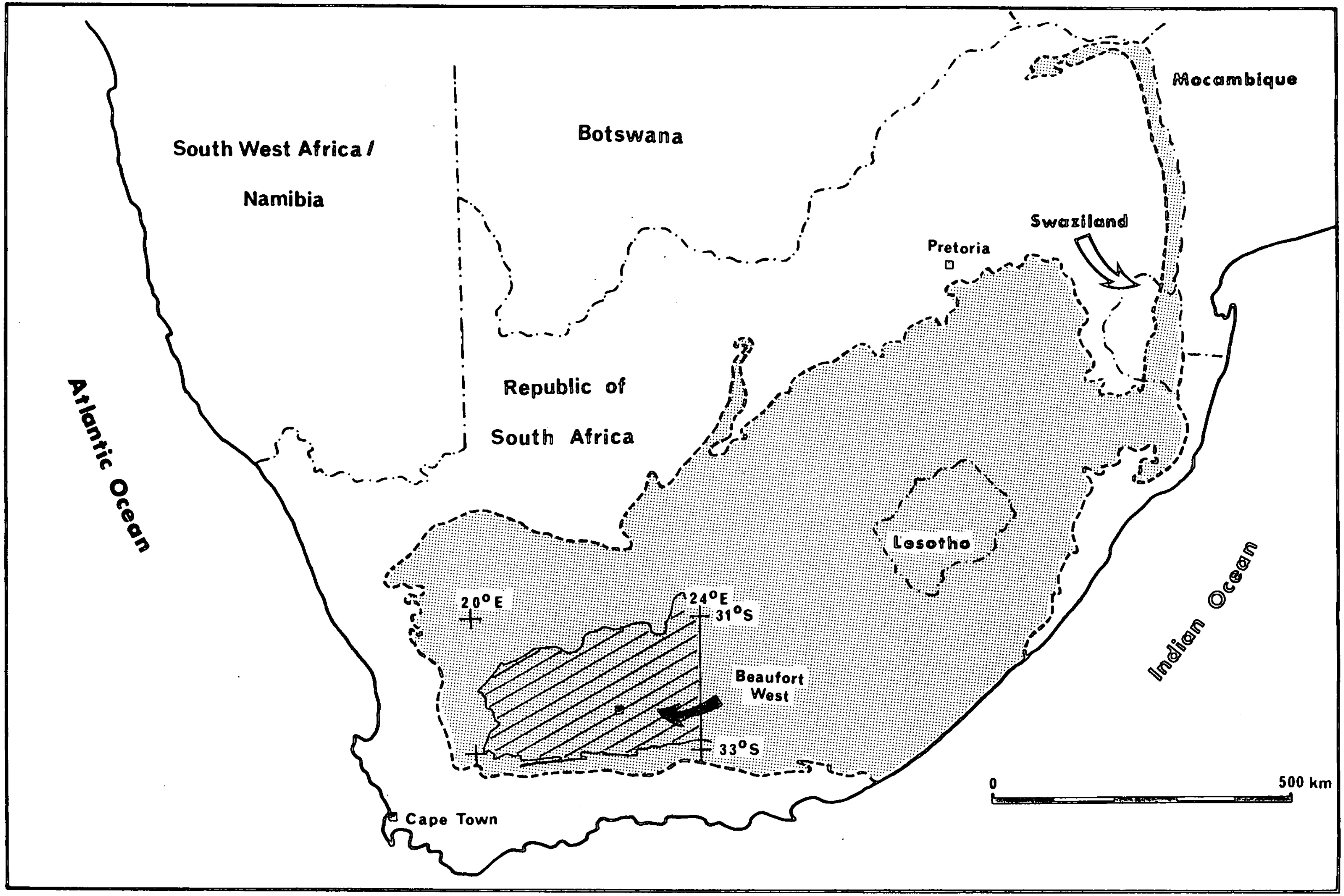

FIG. 1.1 *Locality plan showing the distribution of the Karoo Sequence in the main Karoo Basin (shaded), and the location of the study area (hatched).*

defined by SACS (1980). The Molteno, Elliot and Clarens Formations forming the uppermost part of the succession are confined to the central part of the Karoo Basin, and were probably never deposited in the southwestern parts of the Basin. Karoo sedimentation was terminated by widespread, Early Jurassic, Drakensberg volcanism (Truswell, 1970).

Although the Karoo Sequence attains a composite thickness of approximately 7500 m, maximum thicknesses probably never exceed 5000 m in any part of the main Karoo Basin. The lower Beaufort reaches a composite thickness of just less than 4000 m in the western Karoo Basin.

1.1 Physiography of the Study Area

The most striking physiographical features of the Great Karoo are the mountain ranges of the Cape Fold Belt and the Great Escarpment. The mountains forming the Great Escarpment are mainly the Roggeveld, Nuweveld and Kamdebo Ranges. These mountains are capped with dolerite and flanked by vast pediplains which are in turn dissected by recent, ephemeral streams. The extreme climate and tendency towards brief but heavy rainstorms, rarely as severe as the catastrophic Laingsburg floods in 1981, resulted in a high rate of denudation and good surface exposures of the Karoo strata. (Locations shown in FIG. 1.2, and listed in TABLE 1.1.)

The orogenic Cape Fold Belt comprises intensely folded Cape and Karoo strata, forming prominent mountain ranges such as the Witteberg and Swartberg along the southern margin of the Basin. Strata as old the Precambrian Matjies River Formation (Cango Group) were involved in the deformation, which ended in the Triassic (Rust, 1979). The east-west structural trend along the southern margin is replaced by a north-northwesterly trend, along the western margin of the Basin. Folding of the Beaufort strata was less severe along the western margin but the areas where the southern and western fold systems meet, such as in the Moordenaars Karoo and parts of the Koup (FIG. 1.2), are characterised by complex interference fold patterns.

The Great Escarpment follows an irregular course through the Great Karoo, from the Kamdebo just north of Aberdeen, to the Klein Roggeveld Mountains in the west. Beyond there it swings north along the Roggeveld escarpment. The Great Escarpment is an erosional landform which is

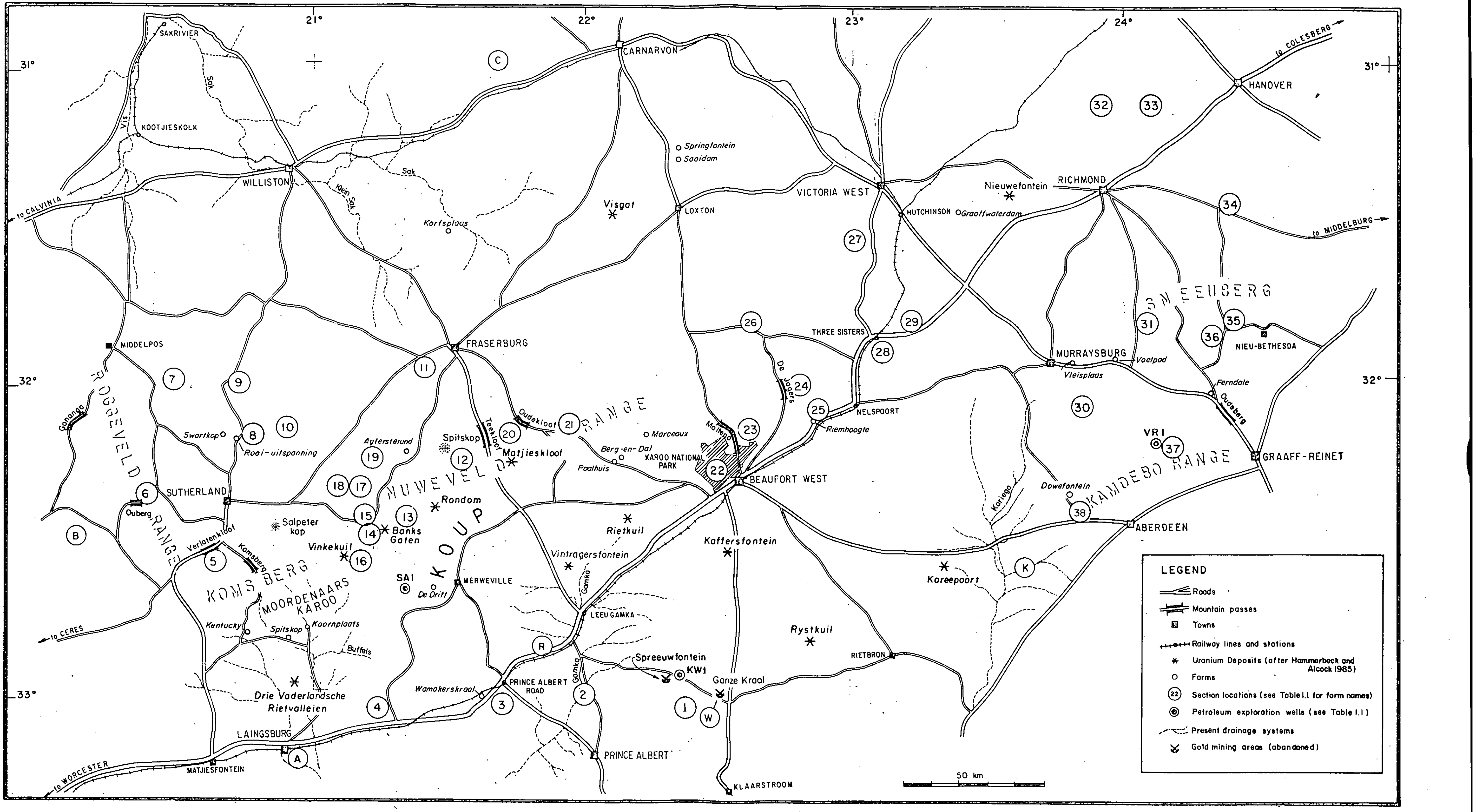

FIG. 1.2 Locality map of the study area showing farms, section locations, and other prominent geographic and cadastral features, including the locations of the major uranium ore deposits. (Locations of major ore deposits from Hammerbeck and Allcock, 1985.)

TABLE 1.1 List of stratigraphic sections and some locations shown in FIG. 1.2. Numbers and symbols as used in text, in Appendix 1, in TABLE 7.1, and in FIG. 7.1.

No.	Magisterial District	Location
1	Prince Albert	Caroluspoort 105, Spreeuw Fontein 26
2	Prince Albert	Kweekkraal 92, Combrinks Kraal 93/1, Blauwkranse 30
3	Prince Albert	Karee Kraal 88, Riet Kraal 89, Tuin Kraal 85, Goeiemoed 90, Elandsberg 87
4	Laingsburg	Becks Vlakte 112, Groot Fontein 113, Wilgebosch Fontein 62
5	Sutherland	Klip Drift 156, Klipbanks Rivier 155, Velaten Kloof 130, Gunstfontein 131
6	Sutherland	Oude Berg 111, Keerom 110
7	Sutherland	Wolve Dance 24, Branddekraal 22, Kareekas Berg 26
8	Sutherland	Wilgerbosch Kraal 32, Palmietfontein 42
9	Sutherland	Snyders Post 20, Hangindihak 18
10	Sutherland	Wilgerbosch Kraal 32, Palmietfontein 42
11	Fraserburg	Droogfoots Fontein 356
12	Beaufort West	Groot Tafel Bergsfontein 237, Goliads Kraal 240
13	Beaufort West	Overse Fontein 249, Bullekraal 251, Banks Gaten 250
14	Sutherland	De Kuilen 142
15	Sutherland, Fraserburg	De Kuilen 142, Rhenoster Valley 485
16	Beaufort West, Sutherland	Wilgebosch Kloof 2, Anneks Wilgebosch Kloof 3, Vinkekuil 143
17	Sutherland	Kruis Rivier 89, Bontberg 88
18	Sutherland, Fraserburg	Uitvlugt 90, Kruis Rivier 483
19	Fraserburg	Kruis Rivier 483, De Kom 474
20	Fraserburg	Wilgerbosch Kloof 449
21	Fraserburg	Rietvalley 452, Rondeheuvel 453, Adjoining Rondeheuvel 456

TABLE 1.1 (Continued)

No.	Magisterial District	Location
22	Beaufort West	Stols Hoek 182, Puttersvlei 190
23	Beaufort West	Williams Kraal 104, Kaffirs Kraal 105
24	Beaufort West	De Hoop 117, Welgevonden 70
25	Beaufort West	Rietfontein 122
26	Beaufort West	Drie Kop 53, Paardeberg 49, Adjoining Paardeberg 50
27	Victoria West	Jasfontein 214
28	Victoria West	Three Sisters 244
29	Murraysburg	Mordant Klaasenskraal 14
30	Murraysburg	Rietvalley 53, Annex Rietvalie 90, Toverwater 91, Groot Rivier 93
31	Murraysburg	Matjesfontein 31, Witteklip 32
32	Richmond	Schootelfountain 28
33	Richmond	Wortel Fountain 30
34	Richmond	Kriegar Fontein 73
35	Graaff-Reinet	Zuurplaats 35
36	Graaff-Reinet	Krugers Kraal 36, Klipfontein 47
37	Graaff-Reinet	De Vrede 286
38	Aberdeen	Uitkyk 100, Fontein Plaats 101
A	Laingsburg	Laingsburg Commonage
B	Sutherland	Bovenste Wagendrift 118
C	Carnarvon	T'kokoboos 500
K	Aberdeen	Ganna Leegte 159
W	Prince Albert	Spreeuw Fontein 26
R	Prince Albert	Riet Fontein 68

now controlled mainly by the distribution of resistant Karoo dolerites, and to a lesser extent the arenaceous zones in the Ecca and Beaufort successions. Much of the scarp face consists of soft-weathering, argillaceous Ecca and Beaufort strata. The Escarpment separates the geomorphologic provinces of the Highveld and Karoo (King, 1963).

The break-up of the Gondwana land mass during the Late Jurassic to Early Cretaceous period, led to the formation of the African continent. The Great Escarpment was also formed during this period, retreating inland from contemporary coastal areas to form the African erosion surface (King, 1963). Pediments of the younger (Late Cainozoic) erosional cycles, such as the Miocene cycle in particular, are locally preserved in lower lying areas of the Great Karoo. Semi-consolidated, calcareous, high-level gravel terraces along the course of the Leeu River near Letjiesbos are remnants of the Miocene erosional surface. Calcrete-covered palaeoweathering surfaces of the same age can also be seen on the flat plains east of the town of Beaufort West. Abandoned courses of the Salt River in this area are filled with up to 35 m of calcrete (Pretorius, 1977). Calcrete-covered palaeoslopes on The Three Sisters and surrounding hills are remnants of a more recent but wetter period of erosion, during which an abundance of scree was produced.

Headward erosion by the Kariega, Salt and Tankwa Rivers account for the present dissection of the African and Miocene surfaces, and also for the irregular shape of the Great Escarpment in the western Karoo. Ephemeral streams are depositing sand, calcareous gravels and rubble in recent alluvial valleys. The drainage directions in most of these valleys is controlled primarily by the structural grain of the Cape Fold Belt. The present arid climate and high rate of erosion are preventing the formation of extensive, residual soil horizons, and accompanying vegetation in the Karoo. Appreciable soil cover is present only locally in areas such as, for instance, parts of the Roggeveld.

1.2 Previous Investigations

Travellers such as Burchell and Lichtenstein, who visited the interior of southern Africa at the turn of the previous century, were amongst the first to record aspects of the geology of the country in their notes (Rogers, 1937). The monotonous Beaufort succession did not attract much attention until the discovery of vertebrate fossils in these strata. In the course of time these were followed by discoveries of gold and pseudocoal, and in recent years also the discovery of uranium. Petroleum exploration was carried out from time to time throughout the twentieth century. Although no economic discoveries have yet been made of these commodities, these investigations have contributed much to our present knowledge of the Karoo Basin.

1.2.1 Historical Review

The first recorded discovery of vertebrate fossils in Beaufort strata was made in 1827 near Beaufort West (Grisbrook, 1831). This was followed by the much publicised discoveries of "bidental" reptilian fossils near Fort Beaufort in the Eastern Cape, in 1838 (Bain, 1845). H.G. Seely, the first trained palaeontologist to collect fossils in the Karoo, proposed the first biozonation of what now constitutes the entire Karoo Sequence (Boonstra, 1969; Seely, 1892). The most notable contributions to the biozonation of the Beaufort Group in earlier years are summarised in TABLE 1.2.

The need to record fossil locations accurately for stratigraphic purposes was soon realised by Broom (1905). As more information became available Broom (1906) proposed a sixfold subdivision of the Beaufort Group. Watson (1914a) produced a sketch map of these biozones, of which the *Procolophon* zone was too thin to be shown on the map. Watson (1914b) later pointed out that the *Pareiasaurus* zone was unsuitable and his suggestion that it be replaced by the *Tapinocephalus* zone, was subsequently accepted. The latter scheme was shown later by Von Huene (1925) on his map of the lower Beaufort biozones.

More than half a century lapsed before Kitching (1970) proposed a number of modifications to this scheme. He pointed out that *Endothiodon* fossils were rare and that they are often found together with *Cistecephalus*. Kitching (1970) suggested that the *Endothiodon* zone be merged with the *Cistecephalus* zone, and introduced the *Daptocephalus* zone for the stratigraphic interval between the *Cistecephalus* and *Lystrosaurus* zones. He also proposed that the *Procolophon* zone be abolished. Keyser and Smith (1977-78) proposed a completely revised scheme comprising seven assemblage zones which could be linked to the present lithostratigraphic subdivisions of

TABLE 1.2 Historical review of the biozonations of the Beaufort Group in the main Karoo Basin.

H.G. Seely (1892)

Zone of the *Zonclodonts*
(Elliot and Clarence Formations)

Zone of the specialised *Theriodonts*
Zone of the *Dicynodonts*
Zone of the *Pareiasaurs*
Beaufort Group

Zone of the *Mesosaurs*
(Whitehill Formation)

R. Broom (1906)

Cynognatus Beds
Procolophon Beds
Lystrosaurus Beds
Cistecephalus Beds
Endothiodon Beds
Pareiasaurus Beds

D.M.S. Watson (1914a)

Cynognatus Zone
Procolophon Zone
Lystrosaurus Zone
Cistecephalus Zone
Endothiodon Zone
Tapinocephalus Zone

J.W. Kitching (1970)

Cynognatus Zone
Lystrosaurus Zone
Daptocephalus Zone
Cistecephalus Zone
Tapinocephalus Zone

A.W. Keyser and R.M.S. Smith (1977-78)

Kannemeyeria-Diademodon Assemblage-zone
Lystrosaurus-Thinaxodon Assemblage-zone
Dicynodon lacerticeps-Whaitsia Assemblage-zone
Aulocephalodon-Cistecephalus Assemblage-zone
Tropidostoma-Endothiodon Assemblage-zone
Pristerognatus Diictodon Assemblage-zone
Dinocephalian Assemblage-zone

SACS (1980). It was based on the distribution of the larger, more abundant, and readily identifiable reptilian fossils.

Cooper (1982) subdivided the entire Karoo Sequence into 15 biozones, all essentially assemblage zones. Cooper suggested that these applied to all Middle Permian to Lower Jurassic successions deposited on the ancient continent Pangaea, and related them to the lithostratigraphic units of that age in the main Karoo Basin.

The first geological subdivision of the Karoo Sequence was proposed in a report by Bain (1856), in which he illustrated the subdivisions on a geological sketch map and sections. Wyley (1859) compiled a stratigraphic column of the geological formations traversed on a journey across the Karoo. He also compiled an unpublished map and sections in which he assigned the Beaufort Group to the "Middle Coal Measures" (Rogers, 1937). T.R. Jones (quoted in Tate, 1867) suggested a fourfold subdivision of the Karoo Sequence comprising the Ecca, Koonap, Beaufort and Stormberg beds. He later merged his Koonap beds with the Ecca beds, thus suggesting a threefold scheme (Jones, 1884). Green (1883) in his discussion of the coals and stratigraphy of the Karoo Sequence introduced the term "Kimberley Shales" for what are now known as the Prince Albert and Whitehill Formations, but failed to identify their counterparts in the southern Karoo. Green interpreted his "Kimberley Shales" to occur somewhere above his Ecca beds in the Nuweveld and Kamdebo Ranges in the south (Green, 1883), although this interval is now known to consist of much younger Beaufort strata.

Molyneux (1881) investigated the potential for coal deposits in the Karoo and Stormberg (North-eastern Cape Province) regions and compiled a detailed section of the stratigraphic succession in the area between the Swartberg Mountains and Prince Albert Road. The pseudocoal occurrences of the Koup and Kamdebo regions were also investigated by Dunn (1879; 1886) who compiled three editions of a geological sketch map of the Cape Province (Rogers, 1937). Dunn (1886) showed the "Dwyka Conglomerate" to be continuous with the Griqualand West glacial deposits on his map and called it the "Dwyka Conglomerate (Glacial)". This knowledge enabled him to correlate the "Lower Karoo Beds" including the "Kimberley Shales" (Tierberg Formation) in the Orange Free State and Northern Cape, with their dark (Ecca) shale counterparts in the central part of the Karoo Basin. Sawyer (1893) drew several sections through the Karoo strata in the area, showing the correct stratigraphic and structural relationships.

The geology of the Karoo Basin received greater attention after the establishment of the Geological Commission of the Cape of Good Hope in 1896. After several field investigations Rogers (1903) proposed a fourfold subdivision for the Karoo Sequence, based on field investigations carried out until 1902. This scheme was essentially chronostratigraphic and was retained until recently, when it was replaced with the lithostratigraphic subdivisions accepted by SACS (1980).

Corstorphine (1897) proposed early in 1896 that the country surrounding Beaufort West be examined for the presence of coal, and the area was subsequently surveyed by Schwarz (1897). Rogers and Schwarz (1901) investigated the country between Beaufort West and Calvinia, and later the Beaufort West, Prince Albert, Sutherland, Fraserburg, Victoria West and Laingsburg Districts (Rogers, 1911; Rogers and Schwarz, 1903). A geological sheet map of the Beaufort West - Fraserburg area (Cape Sheet 13) was compiled in 1911.

The pseudocoal occurrences in the Laingsburg District were investigated on several occasions between 1897 and 1916 (Schwarz, 1897; Rogers and Schwarz, 1903; Rogers, 1917). A geological sheet map was compiled for the country surrounding Laingsburg (Cape Sheet 5). The southern Karoo was also investigated for the possible presence of petroleum during the First World War. The Karoo succession was briefly discussed by Rogers (1917) in a report based on those results.

A summary of previous subdivisions is given in TABLE 1.3.

1.2.2 Recent Investigations

The Geological Survey of the Union of South Africa was instructed at the outbreak of the Second World War to investigate the Karoo succession for the possible presence of petroleum. Parts of the Karoo Basin in the area immediately north of the Cape Fold Belt were amongst the areas isolated for further investigation (Haughton *et al.*, 1953). A reconnaissance survey was followed by detailed mapping of selected areas, including parts of the southern portion of the present study area, east of 21°30'E longitude and south of 32°30'S latitude. Three useful geological maps were subsequently

TABLE 1.3 Summary of previous stratigraphic subdivisions of the Karoo Sequence in the main Karoo Basin. (Partly after Corstorphine, 1904.)

A.G. Bain (1856)

Reptiliferous / Karoo (Lacustrine) Series	Sandstone and blue slaty beds	Reptilian remains, coal and wood
	Sandstones and blue slaty beds (and Beaufort Grit)	Reptilian remains, bivalves and plants
	Light-coloured slates	Plants
	Blue slates	Plants
	Claystone Porphyry	

A. Wyley (1859)

Coal Measures	Upper Coal Measures	Stormberg Beds
	Middle Coal Measures	Reptilian/ *Dicynodon* Beds
	Lower Coal Measures	Brown sandstone and shales
	Upper Karoo Shales	Ecca
	Trap Conglomerate	

T.R. Jones (in Tate, 1867)

Karoo Series	Stormberg Beds	
	Beaufort Beds	
	Koonap Beds	
	Ecca Beds	Upper Ecca Trap Breccia Lower Ecca

A.H. Green (1883)

Stormberg Beds	Volcanic Beds Cave Sandstone Red Beds Molteno Beds
Karoo Beds	
Kimberley Shales	
Ecca Beds	
Dwyka Conglomerate	

T.R. Jones (1884)

Triassic Karoo Beds	Upper	Cave Sandstone Red Beds Stormberg Beds
	Lower	Sandstones, shales Kimberley Shales and conglomerates (in the north)
Carboniferous Ecca Beds		Upper Ecca Beds Dwyka Conglomerate Lower Ecca Beds

E.J. Dunn (1887)

Triassic-Carboniferous	Stormberg Beds	Cave Sandstone Red Beds Coal Measures
	Upper Karroo Beds	
	Lower Karroo/ Ecca Beds	Kimberley Shales Koonap Beds Dwyka Conglomerate (glacial)

A.W. Rogers (1903)

Karoo System	Stormberg Series	Volcanic Beds Cave Sandstone Red Beds Molteno Beds
	Beaufort Series	Zone of specialised *Theriodonts* *Dicynodon* Beds *Pareiasaurus* Beds
	Ecca Series	Shales and sandstones Laingsburg Beds Shales
	Dwyka Series	Upper Shales Conglomerate Lower Shales

P.J. Rossouw *et al.* (1964)

Karoo System	Beaufort Series	Lower Beaufort Stage	Sandstones, mudstones, "chert", vertebrate fossils
	Ecca Series	Upper Ecca Stage	Sandstones, shales, mudstones
		Middle Ecca Stage	Shales
		Lower Ecca Stage	Sandstones, shales
	Dwyka Series	Upper Dwyka Stage	Shales
		Lower Dwyka Stage	Tillite

published: Sheet 166 Schoorsteenberg (Blignault *et al.*, 1948), Sheet 198 Merweville (Rossouw and De Villiers, 1952) and Sheets 3321B Gamkapoort and 3322A Prince Albert (Rossouw *et al.*, 1964).

The use of locally defined stratigraphic markers was attempted in those surveys to clarify the stratigraphy of the Beaufort Group. The "chert" beds were mapped by Blignault *et al.* (1948) in an attempt to solve the Beaufort Group stratigraphy in the Schoorsteenberg area. Rossouw and De Villiers (1952) used both the "Poortjie sandstone" and "chert" beds as markers in regional mapping programs in the Merweville area.

The petroleum exploration programs conducted in the Karoo Basin by the Southern Oil Exploration Corporation (SOEKOR) and a number of private groups since 1965, involved regional mapping and deep exploration drilling. Results were rarely published. During routine radiometric logging in 1967 radio-activity was detected in lower Beaufort strata in certain exploration wells (Turner, 1979). Uranium exploration commenced and surface mineralization was discovered just west of the town of Beaufort West in 1969. Numerous exploration companies were active in the area towards the mid-1970's. A considerable amount of information was gathered in the course of both routine exploration and specialised research programs. The information was gathered in an unco-ordinated manner by the different companies, and was largely lost after uranium exploration ceased in the early 1980's.

Mapping and revision of previous mapping of the Karoo strata, with the emphasis on the Beaufort Group, was undertaken by the Geological Survey of South Africa in the mid-1970's. Lithostratigraphic mapping of the Beaufort Group was carried out in conjunction with regional fossil collecting programs. On Sheets 3222 Beaufort West and 3322 Oudtshoorn the Beaufort Group was subdivided in the Abrahamskraal and Teekloof Formations of the Adelaide Subgroup (SACS, 1980). The Beaufort Group in the Sutherland area was further subdivided into units of between 60 and 120 m thick, starting with the K_31 above the Ecca-Beaufort contact and ending with the K_317 at the top (Sheet 3220 Sutherland).

Intensive investigations of the lower Beaufort were carried out in the course of uranium exploration programs over the last few years. Sedimentological and stratigraphic analyses of the uranium occurrences in the area southeast of Fraserburg were first reported by Kübler (1977). The uranium mineralization in the vicinity of Beaufort West was investigated geochemically by Pretorius (1977; 1982). The sedimentology and stratigraphy of an area southwest of Beaufort West were studied by Stear (1980a and b). Smith (1981) investigated the sedimentology and taphonomy of an area on the Nuweveld Range, west of Beaufort West. The Ecca Group along the western margin of the Basin was studied by Wickens (1984), and the palaeontology of the Ecca-Beaufort in the area between Prince Albert and Aberdeen was investigated by Rubidge (1983 and 1988). The sedimentology and stratigraphy of several uranium occurrences in the main Karoo Basin were studied in detail by Le Roux (1985).

1.3 Present Investigation

A sedimentological and stratigraphic investigation was carried out on the uraniferous lower Beaufort strata in the area west of 24°E longitude, and at selected localities to the east of this line. A suitable contact between the Ecca and Beaufort Groups had to be defined first, and so the study was extended to include the upper Ecca Group.

The primary objectives were to establish in the western Karoo Basin:

(a) the lithostratigraphic subdivisions of the lower Beaufort succession;

(b) the depositional framework, and palaeogeographic setting of the lower Beaufort drainage systems;

(c) the geometry, evolution and tectonic history of the Karoo Basin during lower Beaufort times; and

(d) the sedimentary and stratigraphic controls of uranium mineralization in the Beaufort sandstones.

The investigation involved the measuring of stratigraphic sections at suitable localities spaced across the Basin. Sections were initially spaced along the margins of the Basin, starting some distance below the Ecca-Beaufort contact and extending as far upwards in the succession as possible. Due to the lack of regionally defined marker horizons, and in order to achieve maximum traverse distances, localities had to be selected where the structure permits uninterrupted sections. Surface information was supplemented with drilling information where feasible. Fill-in sections of limited stratigraphic width were measured where permitted by topographic relief and suitable outcrops.

True stratigraphic thicknesses were measured directly in undisturbed strata, using a Jacob staff and Abney hand level (Kottlowski, 1965). Where dips of bedding exceeded approximately 35°, thicknesses were measured with a tape measure and restored trigonometrically. Palaeocurrent directions were measured with a Brunton pocket transit and corrected for magnetic declination. Rock samples (mainly sandstones) were collected routinely for thin sectioning and petrographic investigations, as well as for geochemical analyses. Rock colours were determined on fresh, wet, rock chips with the Rock-Color Chart of the Geological Society of America (Goddard *et al.*, 1975).

2.0 GENERAL GEOLOGY OF THE UPPER ECCA AND BEAUFORT GROUPS IN THE WESTERN KAROO BASIN

2.1 Stratigraphic Subdivisions of the Western Karoo Basin

The need for a lithostratigraphic subdivisioning of the Karoo Sequence has become apparent in recent years. Johnson (1966; 1976) proposed lithostratigraphic schemes for the Cape Supergroup and Karoo Sequence in the Eastern Cape Province, which were accepted with minor modifications by SACS (1980). Using this as a reference the stratigraphy in the western Karoo Basin was studied and classifications were proposed by several workers including Stear (1980b), Johnson (1979), Keyser and Smith (1977-78), Keyser *et al.* (1979), Johnson and Keyser (1979), Wickens (1984), Le Roux (1985), and others. The stratigraphic succession in the western Karoo Basin is depicted in FIG. 2.1.

Although certain intervals of the Karoo Sequence were studied in detail in parts of the Basin, a comprehensive basin analysis of the Karoo Basin has not been attempted to date. One of the reasons is that the Ecca and Beaufort Groups in the western and central parts of the Basin occupy a large interval of the Karoo Sequence, which was previously poorly understood. The present investigation was intended to bridge the gap in our knowledge of the sedimentology and stratigraphy of the lower Beaufort, as a background to a basin analysis of the western Karoo Basin. The regional distribution and correlation of lithostratigraphic subdivisions in the southwestern parts of the Karoo Basin are outlined in FIG. 2.2.

The Dwyka Tillite Formation (Late Carboniferous-Early Permian) at the base of the Karoo Sequence consists of a variety of glaciogene deposits, including diamictites, conglomerates, sandstones, and shales containing occasional dropstones (Visser, 1983). Shales become more common towards the top of the tillites and the succession grades vertically into the carbonaceous shales of the Prince Albert and Whitehill Formations, at the base of the Ecca Group.

The overlying, 30 m thick Collingham Formation consists of alternating silty sandstone beds, and tuffaceous and carbonaceous mudstones (cf. Martini, 1974). The "Matjiesfontein chert" is a regionally defined marker bed occurring near the base of the Collingham Formation, and the upper

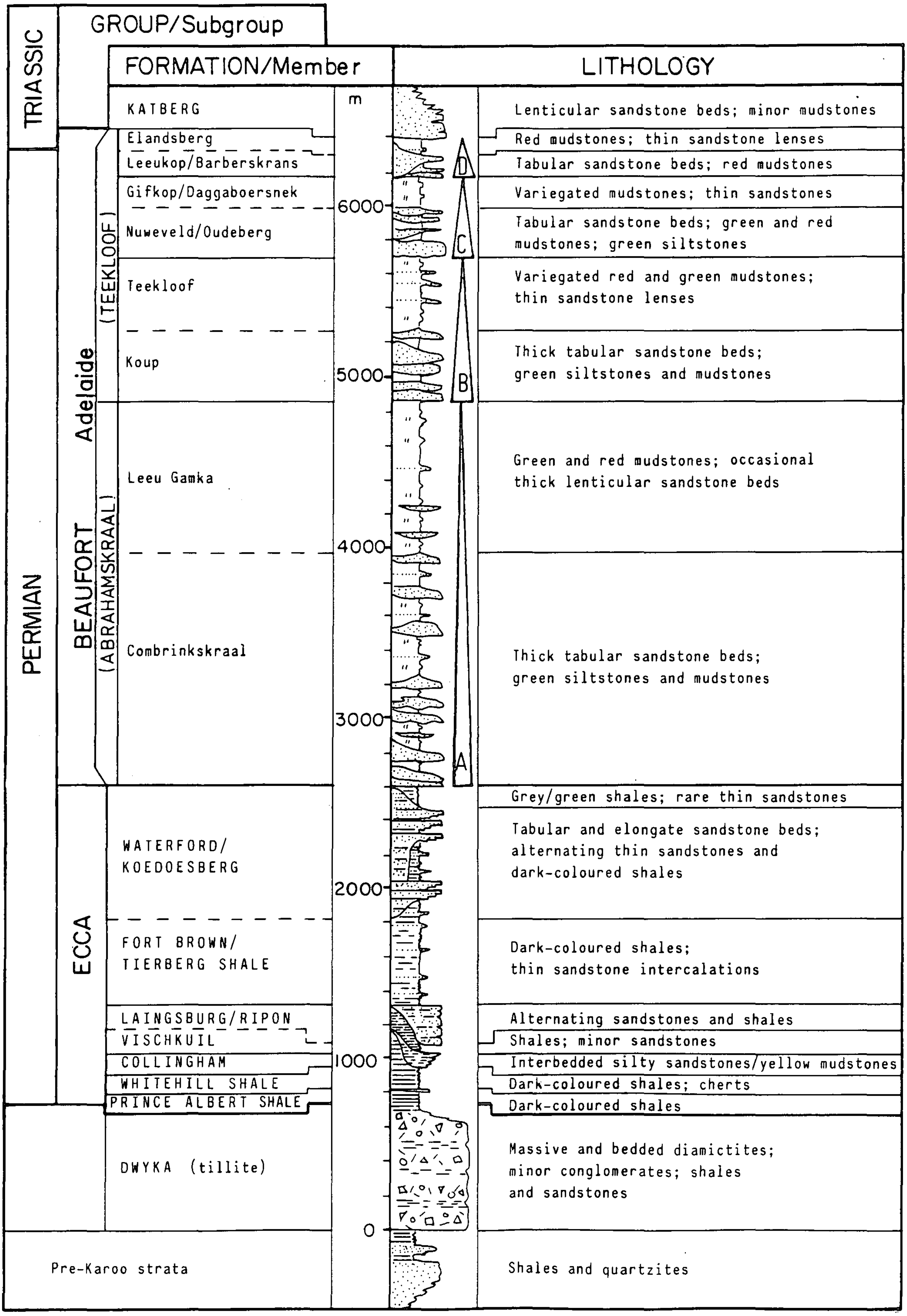

FIG. 2.1 Lithostratigraphic subdivisions of the Karoo Sequence in the main Karoo Basin, west of 24°E. (Partly accepted by SACS, 1980.)

contact is defined by the Laingsburg (Ripon) Formation. The Laingsburg Formation wedges out immediately north of Laingsburg but Collingham-type strata persist as far north as the Bloukrans Pass, south of Calvinia (pers. obs., 1979), beyond where these strata grade laterally into the Tierberg Formation. The shaly Vischkuil Formation underlying the Laingsburg Formation in the southwestern corner of the Basin, was proposed as a separate formation by Theron (1967). Theron considered the Vischkuil Formation a distal facies of the overlying, conspicuously arenaceous Laingsburg Formation, and interpreted both Formations as turbidite deposits. The entire Ecca Group up to here was deposited under relatively shallow marine conditions, in water depths not exceeding a few hundred metres (Visser and Loock, 1978).

2.1.1 Fort Brown Formation

The name "Fort Brown Shale Formation" was proposed by Johnson (1966) for the former "Middle Ecca Stage", and was subsequently accepted by SACS (1980). The succession consists of a monotonous alternation of thinly bedded mudstones and subordinate fine-grained sandstones, up to several hundreds of metres thick. True shales form a small proportion of the succession, which was deposited in a shelf to prodelta setting (Jordaan, 1981; Kingsley, 1977).

SACS (1980) suggested that the Fort Brown Formation did not extend beyond the Laingsburg Formation and that a cut-off point had to be determined in the area somewhere southwest of Sutherland. To the north of that point the Fort Brown Formation is replaced by the Tierberg Shale Formation, the latter name proposed by Nel (1977) and accepted by SACS (1980). Wickens (1979) proposed the name "Wadrif formation" for the strata equivalent to the uppermost Fort Brown Formation along the western margin of the Karoo Basin. He also proposed the name "Schoorsteenberg formation (member)" for the arenaceous succession at the base of the "Wadrif formation" (Wickens, 1979), previously known as the "Tankwa sandstone facies" (Winter and Venter, 1970). The Fort Brown and Tierberg Formations are thus stratigraphically partly equivalent and lithologically identical throughout the western Karoo Basin.

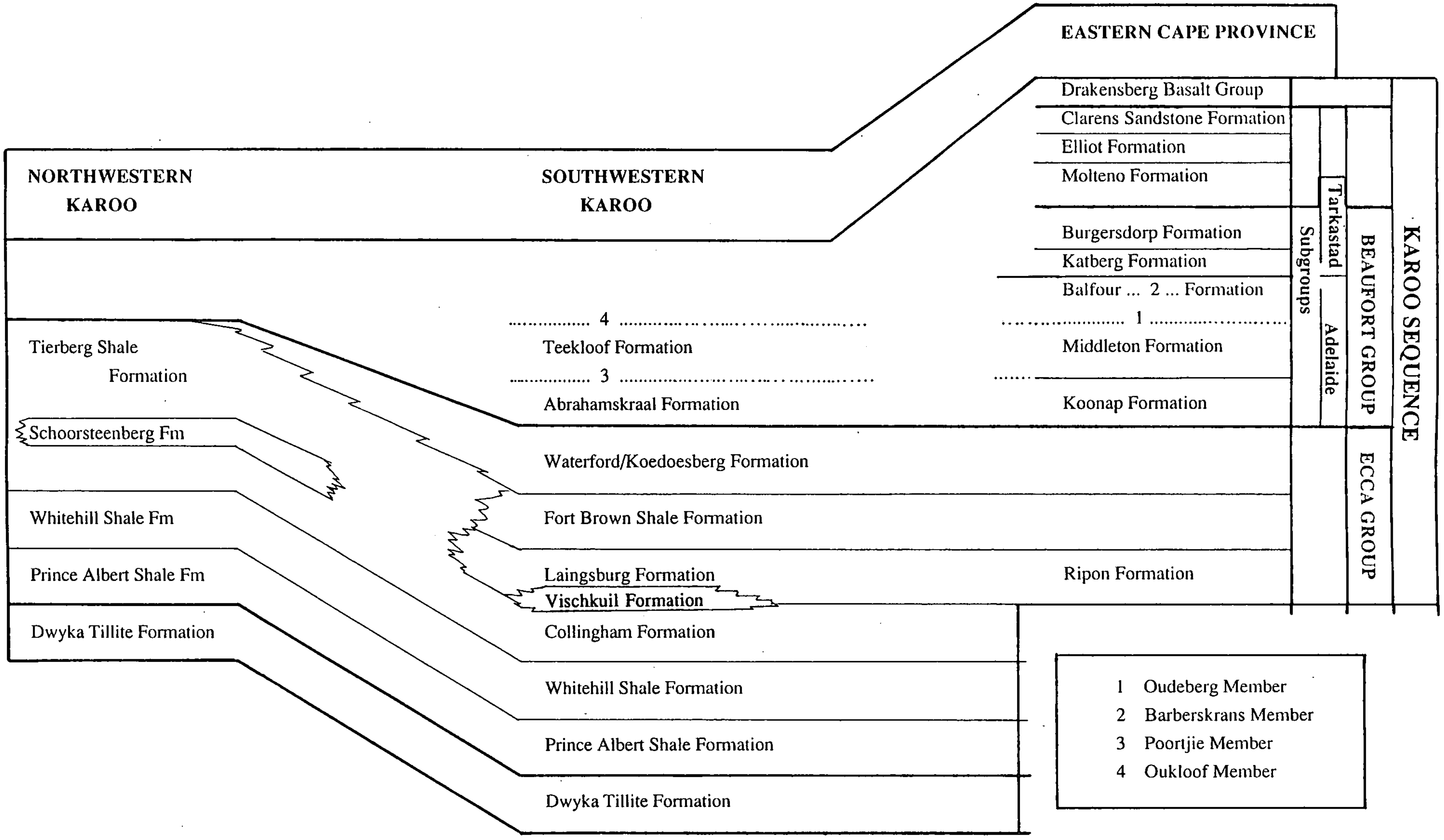

FIG. 2.2 Regional distribution and correlation of some of the major stratigraphic units and markers of the Karoo Sequence in the southeastern, southwestern and northwestern parts of the main Karoo Basin. (Not to scale.)

2.1.2 Waterford Formation

The name "Waterford Formation" was proposed by Johnson (1976) for the arenaceous succession overlying the Fort Brown Formation in the Eastern Cape. It therefore approximately constitutes the former "Upper Ecca Stage" (Rossouw *et al.*, 1964) in the Karoo Basin west of 24°E (SACS, 1980). The unit was named the "Koedoesberg formation" along the western margin of the Basin by Wickens (1979), and the "Carnarvon sandstone formation" in the northern part of the study area by Terblanche (1979).

The succession is conspicuously arenaceous and well developed throughout the western Karoo Basin, particularly to the west of a line running approximately through Carnarvon in the north and Klaarstroom in the south. The basal contact with the Fort Brown and Tierberg Formations is gradational, and these two units interfinger with the lowermost strata of the Waterford Formation. The Waterford Formation in the western Karoo Basin was deposited in the delta front and delta plain environments (Johnson, 1976; Jordaan, 1981).

2.1.3 Beaufort Group

The Beaufort Group consists of alternating feldspathic or lithic wackes, and brightly coloured red and green mudstones and siltstones. The lack of stratigraphic markers and unconformities, as well as the vast thickness of the succession, have hampered stratigraphic investigations over the years. Despite excellent outcrops the stratigraphy of the Beaufort Group was unknown for most of the western Karoo Basin, at the time when the present investigation commenced in 1978.

The Beaufort Group was until recently subdivided into the "Lower", "Middle" and "Upper" Beaufort, the latter referred to as the "Burghersdorp Beds" by Du Toit (1954). The Karoo Working Group of SACS (1980) has recently accepted a twofold subdivision comprising the Adelaide and Tarkastad Subgroups. The base of the Tarkastad Subgroup is situated below the first prominent sandstones of the Katberg Formation (Johnson, 1976). Du Toit (1918 and 1954), on the other hand, considered one of these (Katberg) sandstones as the top of the "Middle Beaufort Beds".

The Adelaide Subgroup was further subdivided into a lower Abrahamskraal and an upper Teekloof Formation by Keyser and Smith (1977-78) who tabled a number of criteria distinguishing these two formations. These subdivisions were accepted by SACS (1980). The "Poortjie sandstone" (Rossouw *et al.*, 1964) occurs at the base of the Teekloof Formation (Johnson and Keyser, 1979).

Johnson (1979) noted two groups of sandstones apart from the "Poortjie sandstone", the first below the "Poortjie sandstone" in the Sutherland area, and the second some 200 to 250 m above the "Poortjie sandstone" in the area south of Fraserburg. Turner (1979) mentioned the presence of sandstone concentrations in the Abrahamskraal and Teekloof Formations which he named the "Paalhuis" and "Oukloof" members, respectively. Van Tonder (1977) identified the "Wortelfontein", "Kommetjiesfontein" and "Blesberg" members in the Merriman - Hanover area, and correlated the "Wortelfontein sandstone member" with the Oudeberg Member identified by Keyser (1973) in the Graaff-Reinet area (SACS, 1980). Several more subdivisions have been proposed in the last few years, as discussed in following chapters.

2.2 Structural Geology

Although locally intensely folded, such as along the southern margins and to a lesser extent the western margins of the Basin, the Karoo Sequence is relatively undeformed. The intensity of folding decreases rapidly beyond the Basin margins with much of the interior of the Basin completely unaffected, notably the northeastern corner of the study area and beyond.

2.2.1 Jointing

Jointing is present in sandstones throughout the study area. They range from closed to partly open fractures which may be filled with quartz, calcite, calcrete or gravel. Joints occur in conjugate sets usually with a prominent set trending nearly north-south, and a less pronounced easterly or northeasterly trending set. These directions were also recorded in the area surrounding Fraserburg (Wilke, 1962), in the Koup (Haughton *et al.*, 1953), and in the southern Karoo (Rossouw *et al.*, 1964). Detailed measurements of a conjugate set of joints in the area northeast of Prince Albert

(Prince Albert goldfields), showed persistently north-northeasterly and north-northwesterly trends. The jointing appears to have formed as a result of deformation related to the Cape orogeny (cf. Section 9.3).

2.2.2 Folding

The Karoo strata along the southern margin of the Basin are intensely folded and locally overfolded, with fold axes trending east-west. These folds are distinctly asymmetric with the northern limbs dipping much steeper than the southern limbs. This style of folding allowed considerable exposure as well as uninterrupted stratigraphic sections of the lowermost members of the lower Beaufort, locally along the southern margin of the Basin. Only the lowermost strata of the Beaufort Group were affected by the folding, which was in places accompanied to a limited extent by thrust faulting.

A less pronounced, north-northwest to south-southeast folding trend is present along the western margin of the Basin. Interference of the western and southern folding trends gave rise to complex folding patterns particularly in the southwestern parts of the Basin. Gentle monoclinal folds resulted, as can be seen along the Klein Roggeveld escarpment, west of Merweville (FIG. 2.3). These folds "step down" towards the south, in contrast to the asymmetric folds which "step down" towards the north.

Lower Beaufort strata in the northern half of the study area dip gently to the south and southeast, towards the centre of the Basin. The centripetal dips in the western Karoo Basin were probably accentuated by differential compaction of interbedded argillaceous and arenaceous Ecca and Beaufort Group strata.

2.2.3 Faulting

No major faulting occurs in the western Karoo Basin. Minor thrusting associated with the folding, and has resulted in severing of fold limbs (usually at the hinges) and limited movement between limbs (Rossouw and De Villiers, 1952). The sense of movement is usually shown by slickensides

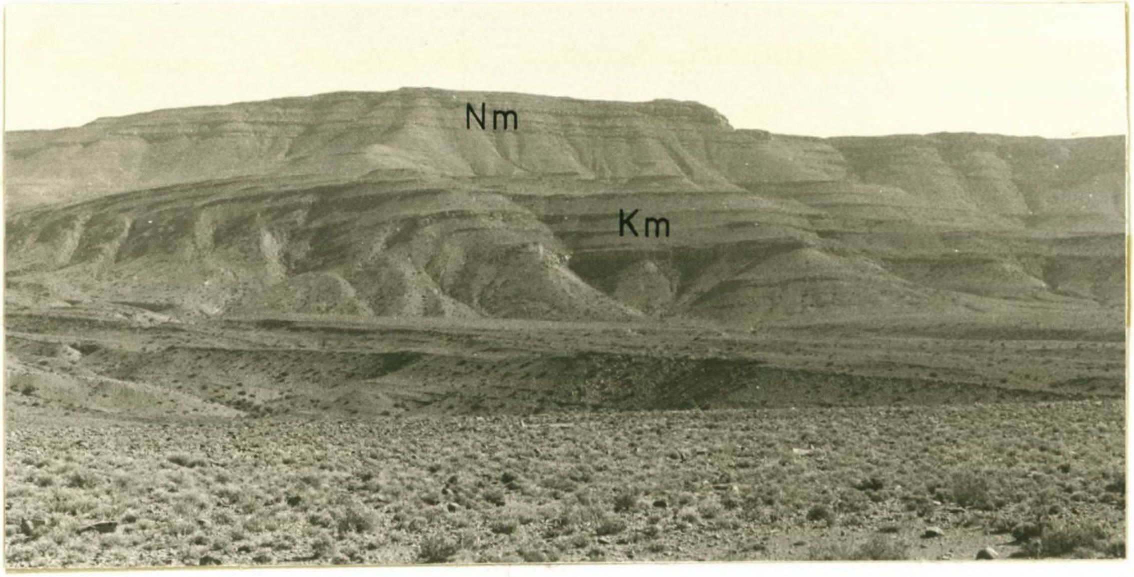

FIG. 2.3 Monoclinal folding of the Koup member (Km) in the Klein Roggeveld escarpment west of Merweville, against a backdrop of relatively undisturbed sandstones of the Nuweveld member (Nm).

FIG. 2.4 Normal faulting along thin a dolerite dyke (arrowed), showing displacement of the basal contact of the Koup member (dashed line) on the farm Rooi-uitspanning, north of Sutherland.

and vein quartz fillings. Faulting of this nature is limited to the more severely deformed, southern region of the study area. Strike-slip faults are present in the western parts of the Basin. Lateral displacements are commonly in the order of several metres to tens of metres, with significant but much smaller vertical displacements. Fault planes are nearly vertical. Downthrow is invariably towards the west or southwest. Examples of these faults were recorded on the farm Wilgerboschfontein (Laingsburg District), in the Verlatenkloof, and in the Ouberg Pass west of Sutherland (FIG. 1.2). These faults presented a serious problem in field measurements of detailed stratigraphic sections across the Ecca-Beaufort contact, particularly in the Ouberg Pass (location 6, FIG. 1.2).

Limited faulting is present along some of the dolerite dykes. The displacement appears to be independent of the thickness of the dyke. Vertical displacement of several metres occurs along a thin dolerite dyke north of Sutherland (Fig. 2.4). Uncertain displacement along an inclined dolerite sill in the Karoo National Park has hampered attempts to measure an uninterrupted section to the top of the Nuweveld escarpment at that locality.

2.3 Dolerite Intrusions

Dolerite intrusions, both dykes and sills, are restricted to the area situated generally to the north of 32°30'S latitude. Dolerite dykes were recorded in widths of between less than 1 m, up to 6 m. Laterally persistent sills occur in varying thicknesses, from less than 1 m on the farms Toorwater and Vleiplaats (location 30, FIG. 1.2), up to 96 m on the farm Vrede (location 37, FIG. 1.2). Contact metamorphic aureoles of limited widths (rarely exceeding 1 m) are restricted to the bases of sills, with virtually no discernable effects on the immediate hangingwalls, as for example on Vrede. In exceptional cases, such as the sills capping The Three Sisters (FIG. 1.2), thermal metamorphism resulted in the severe baking of several metres of the footwall.

As a rule dolerite sills did not present serious problems in field measurements except where traverse lines had to be transposed across transgressive sills, in poorly exposed areas such as on the Nuweveld escarpment in the Karoo National Park (location 22, FIG1.2). Rubble shed from these sills conceals large areas of the Karoo topography, occasionally leaving only limited exposures of

3.0 STRATIGRAPHIC SUBDIVISIONS OF THE LOWER BEAUFORT GROUP

3.1 Introduction

The Beaufort Group covers an area larger than any of the other geological units outcropping in South Africa. The succession is characterised by diagnostic red and green mudstones, on account of which its stratigraphic position was established very soon during earlier geological investigations of the Karoo Basin (cf. Bain, 1856; Wyley, 1859). Lithostratigraphic investigations were neglected in the past mainly due the vast thickness and lithological homogeneity of the lower Beaufort, and preference was given to biostratigraphic investigations. Where lithostratigraphic subdivisions were made these were often based on (or at least related to) known biostratigraphic subdivisions (Johnson and Keyser, 1979; Keyser and Smith, 1977-78; Smith, 1980).

Recent studies in the western Karoo Basin were mostly restricted to areas surrounding the town of Beaufort West. Investigations of the lower Beaufort in the remaining parts of the western Karoo Basin were discouraged by intense folding of strata in the south, the lack of continuous exposures (such as on the Aberdeen flats), the presence of disruptive dolerite intrusions (such as those capping the Nuweveld escarpment), uncertainty about the contact between the Ecca and Beaufort Groups, and the lack of regionally defined marker horizons within the lower Beaufort succession.

Stratigraphic profiles were used as a basis for the present investigation. The procedure involved the selection of section localities from aerial photographs, published geological maps, and topocadastral sheets. In the initial stages section localities were selected where maximum stratigraphic thicknesses of lower Beaufort strata were exposed, and where local marker horizons were expected in the succession. Marker horizons included the basal contact with the Ecca Group, the Poortjie sandstone as mapped by Rossouw and De Villiers (1952) and Rossouw *et al.* (1964), and the Oudeberg Member as mapped by Keyser (1973). A structural investigation of the southern margin of the Basin revealed a limited number of localities where uninterrupted traverses could be obtained from the Ecca contact through to the Poortjie sandstone. As the investigation proceeded and a fundamental lithostratigraphic subdivision became clear, these units were traced laterally into areas where correlation was complicated due to folding, lack of topographic relief, paucity of outcrops, or major facies changes. The field investigation was concluded by correlating lithostratigraphic units identified in the western Karoo Basin, with equivalent successions described in the eastern parts of the Basin by Dukas (1978), Johnson (1976), Tordiffe (1978), and Visser and Dukas (1979). Most of the subdivisions proposed by these authors have since been accepted by SACS (1980).

3.2 Previous Methods of Subdividing the Lower Beaufort

3.2.1 Stratigraphic Marker Beds

The first attempt to subdivide or to classify the Beaufort beds lithologically was by Schwarz (1897) who referred to certain sandstone beds along the Nuweveld escarpment, as either "defining" or "intermediate" sandstones. More recently "chert" beds were used to demarcate stratigraphic units (Haughton *et al.,* 1953; Rossouw and De Villiers, 1952). Rossouw and De Villiers (1952) distinguished between a lower group of reddish or grey "cherts", namely the "Droëfontein" or "Abrahamskraal cherts", and an upper group of two green "cherts" located just below the Poortjie sandstone. A third group of "cherts" occur just below the basal purple mudstones on the farm Botterkraal, north of Prince Albert, and were called the "Ecca cherts" by Rossouw *et al.* (1964).

The green "chert" beds and Poortjie sandstone were mapped as stratigraphic markers in the area situated between the Gamka and Kariega Rivers (Haughton *et al.*, 1953), a distance of some 200 km. As a result of their lenticular geometry and patchy distribution, these "chert" beds are absent in the three sections (locations 1, 2 and 3 on FIG. 1.2) traversing the southern margin of the Basin. The "chert" beds are restricted to the southernmost part of the Basin, as they are entirely absent in the northern half of the study area.

The Poortjie sandstone on its own has been used as a marker bed since it was first mapped in the Merweville area by Rossouw and De Villiers (1952). The unit has variously been described as a ". . . . prominent sandstone or group of sandstone bands . . . about 100 feet thick . . ." (Haughton *et al.,* 1953), and an unspecified sandstone 60 m thick and located approximately 1900 m above the first purple mudstone in the Beaufort Group (Rossouw *et al.,* 1964). Johnson and Keyser (1979) considered the Poortjie sandstone a mappable stratigraphic unit at the base of the Teekloof

Formation, while Turner (1979) classified the sandstone as the base of the overlying Paalhuis member (FIG. 2.2).

The Poortjie sandstone was indicated on the following geological sheet maps produced by the Geological Survey (Sheet 198 Merweville, Sheet 3321B Gamkapoort-3322A Prince Albert, Sheet 3220 Sutherland, and Sheet 3222 Beaufort West). The unit was accepted as a formal lithostratigraphic unit by SACS in 1980, and is presently known as the Poortjie Member. The geographic distribution of this unit within the study area, as well as its correlation with the lower Beaufort biozones, were discussed by Keyser and Smith (1977-78). The position of the base of the Poortjie Member is shown on Sheet 3222 Beaufort West and Sheet 3220 Sutherland.

3.2.2 Lithological Associations

The lack of regionally defined marker beds in the Beaufort Group has led to attempts to identify groups or zones of distinctive rock types in the succession. The Poortjie Member serves as an example. Initially described as a single sandstone or group of sandstones (Haughton *et al.*, 1953), of which individual sandstones were used as markers locally in the southern part of the Basin, the unit was subsequently found to be part of a much thicker zone of sandstones. It is shown here (FIG. 2.2) to form part of an arenaceous interval (Koup member) which is up to several hundred metres thick, but in which individual sandstones can be traced for no more than a few kilometres along strike. (Terminology in brackets refers to units named in this thesis.)

The Oudeberg Member (SACS, 1980) is a useful stratigraphic unit in the eastern sector of the Basin, but pinches out in the Kamdebo just beyond the eastern margin of the study area. The unit consists of two or three sandstone beds with subordinate mudstone in the area to the west and northwest of Graaff-Reinet (Visser and Dukas, 1979). The Oudeberg Member is just over ten metres thick on the farm Vrede (APP. 1.17), but thickens considerably towards the east and reaches a maximum thickness of approximately 180 m in the Great Fish River basin (Tordiffe, 1978).

Exploration for uranium ore deposits in the western Karoo Basin revealed that the larger orebodies were confined to stratigraphic intervals characterised by thick, laterally persistent sandstone bodies.

The two arenaceous successions which emerged as prime uranium exploration targets, are here referred to as the Koup and Nuweveld members. Field investigation and exploration drilling of these two mineralized units greatly contributed to our present understanding of the regional stratigraphy of the lower Beaufort in the study area.

3.2.3 Rock Colours

The abundance of red mudstones in the succession was used by Johnson (1966) to distinguish the Middleton Formation from the Koonap and Balfour Formations in the Eastern Cape Province. The Tarkastad Subgroup in the southern part of the Basin is similarly characterised by a higher proportion of red mudstone than the Adelaide Subgroup (Keyser and Smith, 1977-78). In the western Karoo Basin the abundance of red mudstone was thought to be diagnostic of the Teekloof Formation, distinguishing it from the Abrahamskraal Formation (SACS, 1980). This criterion is not valid in the northwestern part of the study area, where the green mudstones of the Abrahamskraal Formation are replaced entirely by red mudstones. In the western Karoo Basin mudstone colours are of little practical use in stratigraphic mapping of the lower Beaufort, including the Ecca-Beaufort contact (Jordaan, 1981).

3.2.4 Cyclic Stratigraphy

Alternating arenaceous and argillaceous successions were proposed by L. Rabie (quoted in Winter and Venter, 1970) as a basis for stratigraphic subdivisions in the Beaufort Group. Similar tendencies were noted by Van Niekerk (1977) and Van Tonder (1977) in the Beaufort Group to the north and northwest of Graaff-Reinet, respectively. Altogether five interbedded arenaceous and argillaceous members, each several tens or hundreds of metres thick, were mapped in the area west of Graaff-Reinet by Dukas (1978). These members were proposed to consist of one incomplete and two complete, upwards-fining "megacycles" by Visser and Dukas (1979). Large-scale cyclicity was also reported in the western sector of the Basin by Stear (1980b) who subdivided the lower Beaufort into seven megacycles, each consisting of an arenaceous unit at the base and an argillaceous unit on top.

3.2.5 Geographic Facies

A distinction has in recent years been drawn between stratigraphic subdivisions in the eastern and western parts of the Karoo Basin, subdivided roughly along the line 24°E (Keyser and Smith, 1977-78; SACS, 1980). Stear (1980b) referred to these areas as the eastern and western sectors, respectively. Ryan (1967) distinguished between three different geographic "facies" in the Ecca and lowermost Beaufort beds, on the basis of lithology and palaeocurrent directions. Ryan named these the "Western", "Southern" and "Eastern Facies". Visser and Loock (1979a) similarly suggested the stratigraphic equivalence of the Oudeberg Member near Graaff-Reinet and the Wortelfontein (Nuweveld) member near Richmond, both belonging to different "facies" characterised by differences in lithology and transport directions.

3.3 Biostratigraphic Subdivisions of the Beaufort Group

3.3.1 Historical Review

After numerous therapsid fossils had been collected in the Beaufort Group since their discovery more than a century ago (Grisbrook, 1831), a broad stratification of the beds which contained certain of these fossils, become apparent to the pioneer Karoo palaeontologists. The first biozonal scheme proposed by Seely (1892) was incomplete and was soon replaced by a more detailed, sixfold subdivision by Broom (1906). A minor modification was suggested by Watson (1914a) who proposed that the *Pareiasaurus* zone be replaced by the *Tapinocephalus* zone. These biozones were discussed in depth by Watson (1914a) and Von Huene (1925), both of whom presented sketch maps illustrating the distribution of these zones in South Africa.

Although minor inconsistencies became evident, Broom's system remained unchanged for many years until Kitching (1970) proposed some major changes. The paucity of *Endothiodon* fossils led Kitching (1970) to discard this zone, and the *Cistecephalus* zone was redefined to include only what he termed the "*Cistecephalus* band". The succession between the *Cistecephalus* band and the *Lystrosaurus* zone were assigned to his *Daptocephalus* zone, and the *Procolophon* zone was abolished due to the presence of these fossils throughout the *Lystrosaurus* zone. Kitching's work on the fossil reptiles in the Beaufort succession was followed by the publication of a fivefold subdivision in 1977. (These subdivisions are summarised in TABLE 1.1.)

The regional subdivisioning of the Beaufort Group was until that time the main priority of biostratigraphers. The basin-wide scale of their approach in many instances resulted in fossil localities having been recorded vaguely as farms, koppies or even magisterial districts. A new approach was introduced by Keyser (1973) who recorded fossil localities accurately on a geological map in the type area of the *Cistecephalus* zone, west of Graaff-Reinet. This was the first attempt to link biozones to the lithostratigraphic units of the Beaufort Group.

Keyser (1973) subdivided the *Cistecephalus* zone into a lower subzone and an overlying *Cistecephalus microhinus* subzone. This procedure largely eliminated uncertainty as to whether fossil localities indicated in the *Cistecephalus* zone referred to *Cistecephalus* zone fossils or *Cistecephalus microrhinus* itself. The lower subzone comprised a thickness of 180 m of grey, purple and red mudstone, with minor siltstone and sandstone beds, followed by the prominent Oudeberg Member on top. The *Cistecephalus microhinus* subzone was described as comprising 90 m of grey mudstone, with subordinate siltstone and red mudstone.

The fossil content of the Beaufort sequence in the Merriman - Hanover area was investigated by Van Tonder (1977). The distribution of *Cistecephalus microhinus* could not be established, but from other fossil finds the Wortelfontein (Nuweveld) member could be correlated with the Oudeberg Member near Graaff-Reinet (Visser and Loock, 1979a).

3.3.2 Present Biostratigraphic Subdivisions

In the course of regional fossil collecting programmes carried out in conjunction with a lithostratigraphic mapping program by the Geological Survey in the 1970's, a completely new system of biozonation was devised which showed a good correlation with the lithostratigraphic unit system applied at the time in the western Karoo (Keyser and Smith, 1977-78). Correlation with the lithostratigraphic units indicated some short range zone fossils and their associated faunal assemblages. The seven assemblage-zones (TABLE 1.1) recognised within the Beaufort Group by

Keyser and Smith (1977-78), were subsequently accepted by SACS (1980). Five zones represent the lower Beaufort in the western Karoo Basin, and the following summary of these zones is based on descriptions by Keyser and Smith (1977-78) and SACS (1980).

3.3.2.1 *Dinocephalian* Assemblage-zone

This zone occupies the stratigraphic interval above the Waterford Formation and below a green "chert" marker (occurring 120 m below the Poortjie Member), thus spanning most of the Abrahamskraal Formation as defined by SACS (1980). Fossils are generally less abundant in this zone than in the other zones. They include mainly the dinocephalians, large pareiasaurians, pristerognathid therocephalians and small dicynodonts. The zone replaces the old "lower *Tapinocephalus* zone" referred to by Boonstra (1969). *Tapinocephalus* is a rare fossil of which the range could not be determined with certainty, and was therefore found unsuitable as a zone fossil (Keyser and Smith, 1977-78).

3.3.2.2 *Pristerognathus - Diictodon* Assemblage-zone

This zone is characterised by the abundance of small dicynodonts and the absence of dinocephalians, medium or large dicynodonts, and pristerodonts (SACS, 1980). Despite their paucity the pristerognathids are diagnostic, and their disappearance marks the upper boundary of this zone. The zone occupies the upper 300 m of the old *Tapinocephalus* zone, the so-called "upper *Tapinocephalus* zone" of Boonstra (1969). The *Pristerognathus - Diictodon* Assemblage-zone therefore coincides with the Poortjie Member, as defined by SACS (1980).

3.3.2.3 *Tropidostoma - Endothiodon* Assemblage-zone

A remarkable change is noted in the faunal content of this zone, when compared to the zones below (SACS, 1980). The lower 60 m of the 240 m thick zone forms the acme zone for *Tropidostoma microtrema,* a zone fossil of which the skull is easily identifiable and which has a relatively short range (Keyser and Smith, 1977-78).

The lower boundary of this zone coincides with the base of the Teekloof Formation (SACS, 1980) and the upper boundary is marked by a sandy unit, above which *Aulocephalodon* is found. The sandy unit is therefore correlated with the Nuweveld member in the western Karoo, as well as the Oudeberg Member in the eastern Cape Province.

3.3.2.4 *Aulocephalodon - Cistecephalus* Assemblage-zone

This unit replaces or includes what was previously known as the upper part of the *Endothiodon* zone (Kitching, 1970), the lower part of Broom's (1906) *Cistecephalus* zone, the "*Cistecephalus* band" of Kitching (1977), and the *Cistecephalus microhinus* subzone of Keyser (1973). *Cistecephalus* occurs abundantly near the top of the zone, which can be termed an acme zone (R.M.H. Smith, pers. comm., 1981). These fossils become less abundant lower down in the succession and the range is therefore difficult to establish (SACS, 1980). *Aulocephalodon* has a well-defined range, and skull fragments are abundant and easily identifiable.

The *Aulocephalodon - Cistecephalus* Assemblage-zone zone appears to coincide with an arenaceous succession (Nuweveld member) referred to as the Oukloof member by Smith (1980), and which forms the Oudeberg Member in the Eastern Cape Province (SACS, 1980).

3.3.2.5 *Dicynodon lacerticeps - Whaitsia* Assemblage-zone

The lower boundary of this zone is marked by the appearance of *Dicynodon lacerticeps*, and the zone extends upwards to the position where the first *Lystrosaurus* fossils are encountered (SACS, 1980). The identity of *Dicynodon lacerticeps* and *Daptocephalus leoniceps* was determined by M.A. Cluver (quoted in Keyser and Smith, 1977-78) and this zone is therefore identical to the *Daptocephalus zone* proposed by Kitching (1970).

The distribution of this zone was described by Kitching (1977) and it appears to coincide with the

Steenkampsberg (Gifkop) member referred to by Smith (1980) in the western Karoo. The upper boundary occurs at at the base of the Palingkloof (Leeukop) Member in the Eastern Cape Province (Keyser and Smith, 1977-78).

3.3.4 Discussion

Biostratigraphic zones are continuous over considerable distances in the Karoo Basin. Studies of these zones in the western Karoo Basin have in the last few years been concentrated mainly on the well-exposed areas to the south and west of Beaufort West. The western and northwestern edge of the Basin, as well as the area known as the Aberdeen Flats (FIG. 1.2), are stratigraphically important areas that have not been investigated thoroughly. Work in these areas is hampered by poor outcrops and flat topography. Preliminary investigations by Keyser (1977) in one of these problematic areas, namely the area surrounding Sutherland, showed the stratigraphic relationship between the known uranium occurrences, the "Poortjie" sandstone, and *Tropidostoma microtrema.*

The influence of the present system of biozonation on the lithostratigraphic subdivisioning of the Beaufort Group in the western Karoo Basin is clear. The *Dinocephalian* Assemblage-zone coincides with the Abrahamskraal Formation (Johnson and Keyser, 1979). The faunal change just above the *Dinocephalian* Assemblage-zone (SACS, 1980) has largely determined the choice of a contact between the Abrahamskraal and Teekloof Formations close to that position (R.M.H. Smith, pers. comm., 1981). The correlation between the various biozones and lithostratigraphic units is yet to be determined throughout the western Karoo Basin, to gain some indication on whether they can be used confidently in stratigraphic correlations on a Basin-wide scale irrespective of the palaeoenvironmental setting of the host strata.

3.4 Present Lithostratigraphic Subdivisions of the Lower Beaufort

A number of stratigraphic units are being used or have been proposed in recent years in the lower Beaufort Group in the western Karoo Basin (Keyser and Smith, 1977-78; Smith, 1980; Stear, 1980a; Turner, 1979). Provisional subdivisions were published by SACS (1980). These units are briefly reviewed in terms of their overall geographic distribution, and some limitations on the general use of some of these units are pointed out.

3.4.1 Adelaide Subgroup

The Adelaide Subgroup (SACS, 1980) was introduced by Johnson (1976) in the Eastern Cape Province, for the interval referred to as the "lower Beaufort" by Broom (1906). The unit includes the entire Beaufort succession below the Katberg Formation (Johnson, 1976). The upper boundary (Katberg Formation) of the Adelaide Subgroup is not preserved in the western Karoo Basin, and the succession is therefore incomplete.

3.4.1.1 Abrahamskraal Formation

The Adelaide Subgroup is further subdivided into a lower Abrahamskraal Formation and an upper Teekloof Formation (SACS, 1980) in the western Karoo Basin. The lower boundary of the Abrahamskraal Formation occurs at the top of the Waterford Formation (Johnson and Keyser, 1979). The upper boundary was defined as the uppermost "chert" bed in the lower Beaufort succession by Keyser and Smith (1977-78), but subsequently taken at the base of the Poortjie Member by SACS (1980). A thickness of approximately 1800 m has been proposed for this unit along the southern outcrop belt by Keyser and Smith (1977-78). Stear (1980a and b) subdivided the formation into five megacycles, and calculated a thickness of some 1400 m for this interval. Considering that Strear (1980a) included his "Moordenaars sandstone" in this unit, a discrepancy in the order of 1000 m becomes evident between Stear's thickness and that reported by Keyser and Smith (1977-78).

3.4.1.2 Teekloof Formation

The Teekloof Formation is recognised as the uppermost unit of the lower Beaufort in the western Karoo Basin (SACS, 1980). The lower boundary coincides with the base of the Poortjie Member, and the upper boundary with the base of the Katberg Formation or equivalent strata (Keyser and Smith, 1977-78). That makes the Teekloof Formation equivalent to an interval including both the Balfour and Middleton Formations in the Eastern Cape Province (Johnson, 1979). The Teekloof Formation was subdivided into two megacycles by Stear (1980a), for which he reported a combined thickness of approximately 800 m. A minimum thickness of 1400 m was reported by Johnson and Keyser (1979), for the same succession along the Nuweveld escarpment. The lithological characteristics by which the Teekloof Formation can be distinguished from the Abrahamskraal Formation, were listed by Keyser and Smith (1977-78).

3.4.1.2.1 Poortjie Member

On account of it defining the base of the Teekloof Formation (SACS, 1980), the Poortjie Member could be viewed as the most important of the lithostratigraphic subdivisions proposed to date in the western Karoo Basin. First described as a sandstone band up to 30 m thick on the farm Rietfontein (Rossouw and De Villiers, 1952), the unit was later referred to as an arenaceous interval of unspecified thickness and lateral extent (Johnson and Keyser, 1979; Turner, 1979). The regional distribution of the Poortjie Member in the southwestern part of the study area, was shown on a provisional version of the Sheet 3220 Sutherland. Stear (1980b, FIG. 1.2) suggested a thickness of less than 200 m for this unit (base of his megacycle 6) and showed it to be underlain by the "Moordenaars sandstone".

The Poortjie Member is poorly exposed due to the flat topography, in the area southeast of Beaufort West. Throughout the western Karoo Basin it forms part of a prominent arenaceous interval several hundreds of metres thick, here referred to as the Koup member. The base of the unit coincides approximately with the Poortjie sandstone in the southern part of the Basin (Rossow and De Villiers,1952; Rossouw *et al.*, 1964), where it is underlain by a well-defined argillaceous interval more than 500 m thick (here referred to as the Leeu Gamka member).

3.4.1.2.2 Hoedemaker Member

The argillaceous Hoedemaker member was named by Smith (1980), and overlies the Paalhuis or Poortjie Member. The interval has not been described in much detail to date and is here referred to as the Teekloof member, as discussed below and in the description of the Teekloof member. (The Teekloof Formation - SACS, 1980 - is not considered a valid subdivision on the basis of the results of the present investigation.) A thickness of approximately 300 m was suggested by Stear (1980b).

3.4.1.2.3 Oukloof Member

The name "Oukloof member" was proposed by R.M.H. Smith (pers. comm., 1978), and described by Turner (1979) as an arenaceous unit approximately 120 m thick in the Teekloof Formation. The unit has been described as consisting mostly of thick, laterally persistent sandstones and subordinate "bluish-grey" or "maroon" mudstones (Turner, 1979).

3.4.1.2.4 Steenkampsberg Member

This member was proposed by Stear (1980a) to cap the lower Beaufort in the western Karoo Basin. The unit is exposed on the western slopes of Steenkampsberg where it consists mostly of red mudstones with few interbedded sandstone lenses.

3.4.2 Discussion

The contact between the Adelaide and Tarkastad Subgroups at the base of the Katberg Formation in the southern part of the Karoo Basin, becomes indistinct to the north of the study area and for that reason the rank of the Adelaide Subgroup is being questioned. The Katberg Formation loses its definitive arenaceous character and the contact becomes indistinct, in the area north of 31°S (pers. obs., 1978 and 1979). Strata equivalent to the distinctive Katberg sandstones in the southern Karoo Basin change to predominantly green and red mudstones and subordinate sandstones, in the southern Orange Free State. The stratigraphic position of the "massive yellow sandstone" suggested by Du Toit (1918) as being the base of the "Middle Beaufort" in the Orange Free State is uncertain, and the particular sandstone referred to appears to be absent over most of the southern Orange Free State (pers. obs., 1978 and 1979). It is thus suggested that all the arenaceous and argillaceous intervals of the Beaufort Group including the Katberg Formation as presently defined (SACS, 1980), conform rather to member status. Although the Adelaide and Tarkastad Subgroups are provisionally retained in this investigation, they may be required to be relegated to formational status, on condition that (1) the contact between these units can be mapped regionally with confidence, and (2) the overall degree of lithological change justifies the distinction. Despite the reservations expressed here, based primarily on observations of field relationships outside the study area, the issue is of a Basin-wide scale and it is not possible to make a recommendation on the basis of the results of the present investigation. The entire Beaufort Group in the western Karoo Basin falls below the Katberg Formation and is referred to informally and provisionally as the "lower Beaufort", pending clarification of the status of the constituent stratigraphic units.

The Abrahamskraal Formation shows very little lithological variation compared to the overlying Teekloof Formation. The "chert" beds which are characteristic of the Abrahamskraal Formation in the southern part of the Basin (Keyser and Smith,1977-78), are entirely absent in the western and northern parts of the study area. The Poortjie Member which defines the contact between these two units, has never been properly defined apart from a vague reference as ". . . a disconnected series of relatively arenaceous zones occurring at more or less the same stratigraphic level . . ." (SACS, 1980). Although the degree of lithological change required to distinguish units of formational status is somewhat subjective (cf. Hedberg, 1976), the changes between the Abrahamskraal and Teekloof Formations are negligible over the extent of the western Karoo Basin. The distinctive lithological characteristics of the succession in the southern part of the Basin change considerably in the northern half of the study area. The mudstone colours, for instance, change from predominantly green in the south to predominantly red in the north. On a regional scale no distinguishing characteristics could be found between the Abrahamskraal and Teekloof Formations, and their distinction cannot be justified in the present scheme.

3.5 Ecca-Beaufort Contact

A regionally consistent contact to subdivide the Ecca and Beaufort Groups had to be determined before a stratigraphic analysis of the Beaufort Group could be attempted. This aspect received little attention in the past, and a suitable contact had not been resolved by the time the investigation started in 1977. Several problems hampered such an investigation, namely (1) the lack of marker beds , (2) the diachronous Ecca-Beaufort transition, (3) the rapid changes in thicknesses and field appearances of these two successions, and (4) a disregard for the controls that the depositional environments had on the lithological character of the succession. The latter aspect is discussed in greater detail elsewhere, and only those aspects related to the lithostratigraphic subdivisioning of the succession are considered here.

3.5.1 Historical Review

The Beaufort Group was initially distinguished from the Ecca Group on the grounds of the presence of reptilian fossils (Bain, 1856) and ". . . purple, greenish and grey shales . . ." (Wyley, 1859). Few attempts have since been made to define the contact more accurately and the criteria used in different parts of the main Karoo Basin until now, were reviewed by Jordaan (1981).

The presence of reptilian fossils was used by Rogers (1925) north of Laingsburg to indicate the presence of Beaufort strata. The results were unsatisfactory as illustrated by the inclusion of some unmistakable Beaufort strata (Leeu Gamka member) northeast of Laingsburg as an inlier of Ecca (Sheet 5 Laingsburg). Detailed investigations of the area just north of Klaarstroom (Barry, 1974; Jordaan, 1981; Visser and Loock, 1979b) and elsewhere in the western Karoo Basin (Rubidge, 1983) have nevertheless shown that therapsid fossils make their appearance directly above the Waterford Formation, above the horizon favoured at present as the Ecca-Beaufort contact.

The exclusive use of lithological criteria to determine a contact in the southern Karoo was advocated by Haughton *et al.*, (1953), and the first occurrance of red mudstone above the Ecca Group was selected as an easily recognisable marker. The basal red mudstone was taken as the lower contact of the Beaufort Group in the Schoorsteenberg (Blignaut *et al.*, 1948) and Prince Albert (Rossouw *et* *al.*, 1964) areas. This contact has been questioned in recent years (Johnson, 1966; Jordaan, 1981) and found to be entirely inadequate for the purposes of a detailed stratigraphic analysis of the Beaufort Group.

The problems surrounding the interpretation of a stratigraphic contact in drillcore, amongst others, have led to suggestions that cyclical sequences characterised by the abuncance and paucity of sandstone (Winter and Venter, 1970) or thicknesses of individual sandstones (Van Biljon, 1967) may hold the key to a suitable contact. A number of lithological parameters were suggested by SACS (1980) as being diagnostic of the Ecca and Beaufort Groups, including the proportion, weathering and colour of argillaceous rocks, the presence of upwards-fining sedimentary cycles, reptilian remains, and the presence of cross-bedded sandstone. Inconsistencies in these criteria soon became evident: (1) The Waterford Formation, which forms part of the Ecca Group (SACS, 1980), consists predominantly of sandstone and not shale. The paucity or abundance of sandstones therefore does not serve as an indication of the stratigraphic position. (2) Cross-bedding is locally as common in the Ecca sandstones as in the Beaufort Group (Jordaan, 1981). (3) Trough cross-bedding is rare in the delta front deposits, but as abundant in the delta plain as in other fluvial deposits (Jordaan, 1981). (4) Red mudstones start appearing between several metres and several hundreds of metres above the Waterford Formation in the western Karoo Basin, and therefore do not appear on the same stratigraphic level everywhere.

These and other purely lithological criteria do not provide exclusive "fingerprints" of stratigraphic units or depositional environments, and neither can a contact be pegged to a single, laterally persistent bed. All parameters therefore had to be viewed in a broader context, and with reference to the palaeoenvironmental framework of the transitional sequence. The depositional environments of these vastly different stratigraphic units could have been expected to, and clearly had some control on the lithology of the transitional succession, and this aspect was used throughout as a *guideline* to determine a consistent and readily mappable contact.

3.5.2 Proposed Contact

The lower parts of the transitional sequence are characterised by dark Ecca shales with subordinate sandstones, grading upwards into the distinctly arenaceous Waterford Formation, followed on top by green and red Beaufort mudstones and interbedded sandstone lenses. Both the base and top of the Waterford Formation are sharp and the contacts clearly visible. Even where sandstone beds become rare and thin, such as in the area north of Klaarstroom, the Waterford Formation can easily be identified. When the genetic implications are considered there seems to be no justification in placing a major lithostratigraphic boundary at the base of the Waterford Formation (Hedberg, 1976, pp. 7, 17, 31 and 95; Jordaan, 1981). The contact between the Ecca and Beaufort Groups was therefore positioned at the top of the Waterford Formation, a horizon which broadly coincides with the first occurrences of red mudstone and therapsid fossils (Jordaan, 1981; Rubidge, 1983).

A thin, argillaceous unit at the top of the Waterford Formation, rarely more than 200 m thick, separate the Waterford Formation (and Britskraal Shale Member where present) from typical Beaufort Group sandstones and mudstones. The argillaceous unit consists of massive green mudstones, rarely massive red mudstones, and interbedded, thin siltstone and sandstone lenses. The first prominent Beaufort sandstones marking the top of the argillaceous unit occur up to 400 m above the Waterford Formation, as for instance on the Laingsburg Commonage (Jordaan, 1981). These strata are classified as part of the Beaufort Group as the lithological composition is identical, and interfinger with the lowermost arenaceous beds of the Beaufort Group.

3.6 Lithostratigraphic Subdivisions Used for the Lower Beaufort Group in the Western Karoo Basin

The need for regionally defined lithostratigraphic subdivisions which could be followed throughout the western Karoo Basin became clear from the previous discussion. The informal scheme used here is valid for the lower Beaufort (Adelaide Subgroup) throughout the study area and agrees in part with that accepted by SACS (1980). The scheme presented here was not intended to replace that accepted by SACS (1980), but is considered suitable to serve as a basis for future subdivions of the lower Beaufort in the western Karoo Basin. The subdivisions are based on lithology alone and the distribution of therapsid fossils in the succession was not considered. The correlation between lithostratigraphic and biostratigraphic units is an interesting and essential field for future stratigraphic analyses of this region. Certain diagnostic units have been described previously but have been redefined on the basis of the results of the present investigation. The proposed members are correlated with confidence with their counterparts described in the Adelaide Subgroup in the eastern part of the Basin (Dukas, 1978; SACS, 1980; Tordiffe, 1978; Van Tonder, 1977).

3.6.1 Combrinkskraal Member

The Combrinkskraal member forms the basal unit of the lower Beaufort in the western Karoo Basin, overlying the Waterford Formation of the Ecca Group conformably. The stratotype is located on the banks of the Gamka River, on the farms Combrinkskraal and Blauwkranse, situated some 25 km due south of Leeu Gamka (location 2, FIG. 1.2). A thickness of 1916 m was recorded at this locality. The basal contact coincides with the top of the shaly interval, just west of the homestead on Combrinkskraal (APP. 1.2), which resembles the Britskraal Shale Member described in the Eastern Cape Province by Johnson (1976). Elsewhere in the western Karoo Basin, where these shales are absent, the basal contact is placed at the topmost, thick, deltaic sandstone of the Waterford Formation. The upper boundary is transitional and occurs at the level where thick sandstones become rare, and the succession grades vertically into the Leeu Gamka member. The entire succession is intensely folded and even overfolded in the type area, dipping at gradients of more than 70° along much of the southern margin of the Basin (FIG. 3.1).

The succession consists typically of thick sandstone beds (up to 24 m), interbedded with green and red mudstones. Sandstone:mudstone ratios are of the order of 1:3 to 1:4. The thickest sandstones are found along the southern margin of the Basin between Laingsburg and Klaarstroom, and thick sandstones are very rare in the northern part of the Basin. Exceptions are two superimposed sandstones with a total thickness of 65 m, intersected in core drilling on Droogvoetsfontein, in the northern part of the Basin (location 11, FIG. 1.2). Sandstones comprise arkosic wackes and subwackes and in places they contain large, brown calcareous nodules ("koffieklip"). Occasionally entire sandstone beds are calcitised in the lowermost part of Combrinkskraal member north of Prince Albert, where they were described as "marls" by Rossouw *et al.* (1964). These sandstones show sporadic, high-grade but patchy uranium mineralization. In the southern part of the Basin the mudstones are predominantly green, and red mudstones are very rare. The mudstones gradually change to red where the succession is exposed along the western margin of the Basin (e.g.

FIG. 3.1 Steeply dipping channel sandstones, up to 17 m thick, near the base of the Combrinkskraal member on the farm Combrinkskraal.

FIG. 3.2 Gently folded, alternating mudstones and thin sandstones of the Leeu Gamka member on the farm Spreeuw Fontein, south of Beaufort West.

Verlatenkloof), becoming entirely red in the vicinity of Kareekasberg (location 7, FIG. 1.2), northwest of Sutherland.

The Combrinkskraal member reaches maximum thicknesses along the southern margin of the Basin, becoming much thinner in the northwestern part and pinching out north of Fraserburg and Loxton. The succession thins rapidly from nearly 2000 m on Combrinkskraal, to just more than 700 m on Wilgerbosfontein, over a distance of just over 70 km along strike. Thicknesses are uncertain just north of Laingsburg, due to complex folding, but a slight increase is evident in the Verlatenkloof. From there the succession thins gradually in a northerly direction. The entire Combrinkskraal member is represented by a 65 m thick sandstone unit on the farm Droogvoetsfontein (location 11, FIG. 1.2), from where it grades diachronously into the Ecca Group in a northerly direction. The diachronous transition is evident on surface in the vicinity of the farms Korfsplaas and Springfontein (situated approximately 70 km due north of Fraserburg; FIG. 1.2), and farther north on the farm T'kokoboos (location C, FIG. 1.2).

The Combrinkskraal member is poorly exposed in the southeastern part of the study area. The succession was studied on surface as far east as the Kariega River area (location K, FIG. 1.2), west of Aberdeen, where it hosts sporadic uranium mineralization. Estimates from a deep petroleum exploration well on the farm Vrede (location 37, FIG. 1.2) southwest of Graaff-Reinet, suggest a thickness of less than 500 m in this area. From there the Combrinkskraal member grades eastwards along strike into the Middleton Formation.

3.6.2 Leeu Gamka Member

The Leeu Gamka member is well exposed and reach maximum thicknesses (more than 500 m) on the banks of the Gamka River, on the farms Blauwkranse and Kweekkraal south of Leeu Gamka. The unit thins from more than 500 m along the southern margin of the Basin, to 60 m on the farm Droogvoetsfontein in the north. The Leeu Gamka and Combrinkskraal members become indistinct northeast of Fraserburg where they eventually grade into the Ecca Group, and subcrop beneath the Koup member in the area north and northeast of Loxton.

The succession is folded and consists of green mudstone, thick siltstone beds and subordinate sandstone at the type locality (FIG 3.2). Green mudstones predominate in the succession along the southern outcrop belt, but are gradually replaced by red mudstones to the north. Red mudstones become predominant in the northern and northwestern part of the Basin. Although thick sandstones (5 to 6 m) are rare they do occur at places in the southern part of the Basin (FIG. 3.3), notably on Karoluspoort and along the Roggeveld escarpment south of Sutherland. Some calcareous sandstones on the southern slopes of Elandsberg, in the Leeu Gamka member just east of Prince Albert Road, contain low-grade uranium mineralization.

The Leeu Gamka member constitutes the upper part of the Abrahamskraal Formation and *Dinocephalian* Assemblage-zone (SACS, 1980), and forms the most distinctive argillaceous interval in the Beaufort Group in the western Karoo Basin. The lower boundary is transitional and is positioned at the level above which thick sandstone beds typical of the Combrinkskraal member, become rare. The upper boundary is marked by the appearance of a distinctive arenaceous interval comprising thick, laterally persistent sandstone beds including the Poortjie sandstone as mappedby Rossouw and De Villiers (1952) and Rossouw *et al.* (1964). Except on the farm Wilgerbosfontein (location 4, FIG. 1.2), where the contact is transitional, the upper contact is sharp and can be mapped with ease throughout the study area. The contact is clearly seen near the top of Elandsberg (situated just east of Prince Albert Road - location 3, FIG. 1.2), near the top of Verlatenkloof Pass (location 5, FIG. 1.2) and Komsberg Pass (FIG. 3.3) south of Sutherland, near the top of Vaalberg situated west of Sutherland, on the northern slopes of Basterberg on the farm Snyderspos (FIG. 3.4), and along the western boundary of the farm Droogvoetsfontein (location 11, FIG. 1.2). The upper contact (with the Koup member) is slightly displaced by faulting along a thin dolerite dyke on the farm Rooi-uitspanning, north of Sutherland (FIG. 2.2). The inlier shown as Ecca Group northeast of Laingsburg, on Sheet 5 (Laingsburg) and the 1970 edition of the Geological Map of South Africa, was in the present investigation found to consist of a mudstone interval belonging to the Leeu Gamka member, surrounded by younger strata of the Koup member.

3.6.3 Koup Member

The Koup member in the type area is characterised by thick, laterally persistent sandstones,

FIG. 3.3 Sharp contact between the Leeu Gamka member (LGm) and the Koup member (Km) in the Komsberg Pass. Thick sandstone lenses can be seen near the top of the Leeu Gamka member.

FIG. 3.4 Sandstones of the arenaceous Koup member capping the argillaceous Leeu Gamka member in the Bastersberge, north of Sutherland.

alternating with green or (rarely) red mudstone containing numerous, thin siltstone and sandstone lenses. The sandstones are exceptionally thick in parts of the Koup, reaching thicknesses of up to 60 m on the farm Vintragersfontein (FIG. 1.2). The Rystkuil uranium deposit is hosted in a sandstone deposit of similar thickness (Eddington and Harrison, 1979), and a 37 m thick sandstone forms the base of the Koup member north of Aberdeen (location 38, FIG. 1.2). The upper boundary is usually indistinct and marked by a gradual decrease in the proportion of sandstones, except in the area situated between the farms Berg-en-Dal, Paalhuis and Marceaux, northwest of Beaufort West. Here the top of the Koup member is sharply demarcated by an interval containing an abundance of laterally persistent sandstone sheets (FIG. 3.5).

Over much of the southern part of the study area the Koup member consists of a distinctly arenaceous succession, up to 250 m thick. The 5th and 6th megacycles of Stear (1980a and b) form part of this succession. The Koup member has a similar arenaceous composition in the area between Sutherland and Fraserburg. Individual arenaceous zones within the Koup member could, as in the case of the Combrinkskraal member, not be correlated between profiles, particularly in folded areas such as the in the Moordenaars Karoo. Attempts to correlate clusters of sandstones were unsuccessful in the eastern part of the Koup, as a result of the flat topography of the area.

The Koup member occupies much of the central part of the Basin, south of the Nuweveld escarpment. The succession is complexly folded in the Moordenaars Karoo (FIG. 1.2), where the strike of the strata changes from east-west to north-south. The base of the Koup member ascends the escarpment south of Sutherland and follows the western margin of the Basin to the area north of Sutherland, where the northernmost Koup strata occur in Kareekasberg and Bastersberg (FIG. 1.2). The uppermost portion of the lower Beaufort succession becomes very arenaceous in the northwestern part of the Basin, and individual members are difficult to distinguish. The succession can be followed from the Sutherland District eastwards through the farms Palmietfontein and Droogvoetsfontein (localities 10 and 11, respectively, FIG. 1.2), to a poorly exposed area between Loxton and Carnarvon. The basal strata of the Koup member grade into the Ecca Group north of Loxton, and the transition beds can be seen on surface in the vicinity of the farms Saaidam and Springfontein (FIG. 1.2).

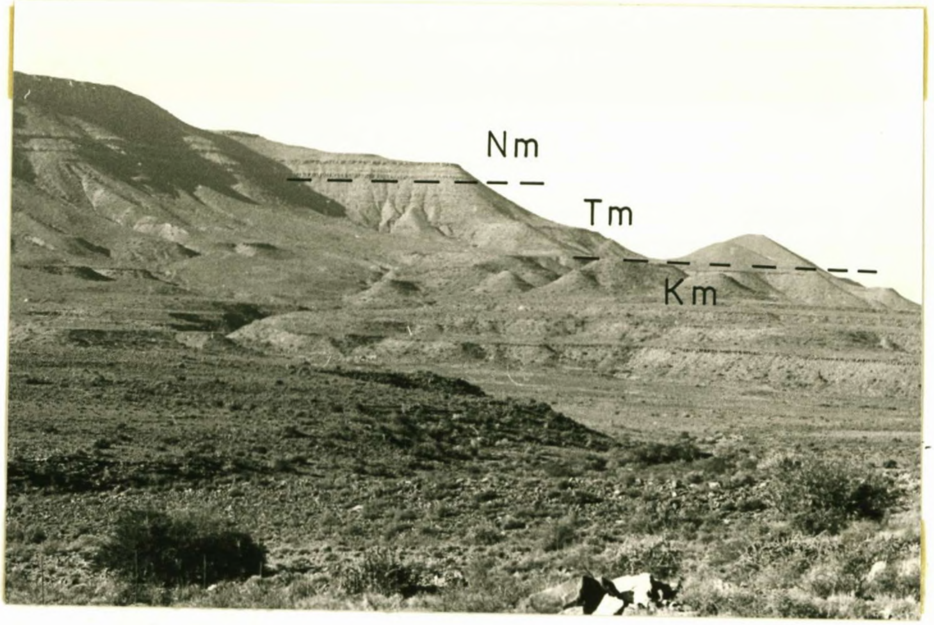

FIG. 3.5 Laterally persistent sandstone sheets in the Koup member (Km) followed on top by the Teekloof (Tm) and Nuweveld (Nm) members, on the Nuweveld escarpment west of Beaufort West.

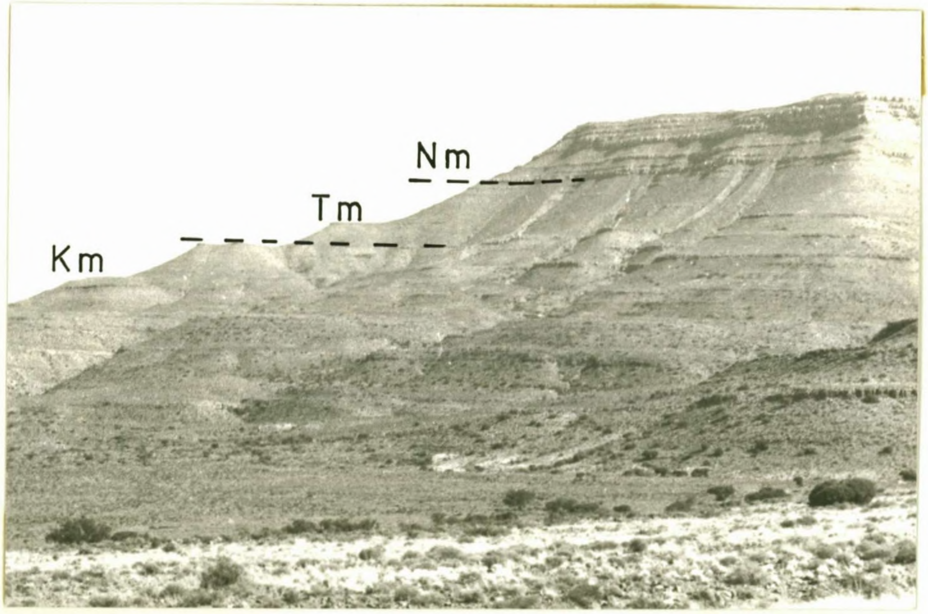

FIG. 3.6 Lower Beaufort succession in the Klein Roggeveld Mountains, southeast of Sutherland, consisting of the Koup member (Km) below, followed on top by the Teekloof (Tm) and Nuweveld (Nm) members.

The economic importance of the Koup member has resulted in several investigations. Kübler (1977) mapped five channel sandstones in detail, of which the "B", "C", "D" and "E" sandstones form part of the Koup member. Kübler (1977) concluded that his "B" sandstone was equivalent to the Poortjie sandstone, as defined by Rossouw and De Villiers (1952). Johnson (1979) referred to sandstones at the top of the Verlatenkloof (base of Koup member) and classified them as strata from the succession below the Poortjie sandstone. This succession (of which the Poortjie sandstone define the lower boundary) was named the Paalhuis member by Smith (1980). Both the Moordenaars (base of his 5th megacycle) and Poortjie sandstone (base of his 6th megacycle) described by Stear (1980b), belong to the Koup member.

The Koup member shows its maximum development in the regions known as the Moordenaars Karoo and the Koup, the latter being the type area (FIG. 3.6). The stratotype is located on the farm Wilgerbosfontein (location 4, FIG. 1.2), 35 km eastnortheast of Laingsburg, where the succession is fully exposed on the southern slopes of a series of prominent hills on the farm. A thickness of 768 m was measured at this locality. The unit can be traced eastwards from the Koup, along the foothills of the Nuweveld escarpment, to the foothills of the Kamdebo Range (north of Aberdeen) from where it grades laterally into the Middleton Formation in the Eastern Cape Province (Visser and Dukas, 1979). The unit could not be distinguished in drillcore from a petroleum exploration well VR1 located on the farm Vrede (location 37, FIG. 1.2) in the Graaff-Reinet District.

3.6.4 Teekloof Member

The Teekloof member consists of thick, red mudstone intervals with subordinate siltstone and sandstone intercalations. Mudstone intervals are up to 120 m thick on the farm Oversfontein (FIG. 3.7). The stratotype is located along the Nuweveld escarpment just east of the Teekloof Pass, 35 km south of Fraserburg (FIG. 1.2). The succession includes the interval stretching from the uppermost sandstone bed of the Koup member to the horizon where sandstones become abundant once more in the succession (Nuweveld member). The lower boundary is transitional over a few tens of metres, depending on the presence or absence of thick sandstones in the Koup member in a particular region (e.g. the foothills of the Nuweveld escarpment, northwest of Beaufort West). The upper boundary is defined by the basal sandstone of the arenaceous Nuweveld member.

FIG. 3.7 Thick succession of red mudstones in the type area of the Teekloof member in Teekloof Pass, south of Fraserburg.

FIG. 3.8 Lenticular habit of channel sandstones in the Nuweveld member on Spitskop, viewed from Tafelberg, south of Fraserburg.

Thicknesses of up to 190 m were measured on the Nuweveld escarpment (Wilgerboskloof, location 16, FIG. 1.2), the interval thinning to less than 30 m on Palmietfontein (location 10, FIG. 1.2), and becoming indistinct northeast of Fraserburg. It is questionable whether subdivision of this unit serves any purpose in that part of the Basin.

The Teekloof member is synonymous with the Hoedemaker member, and falls within the *Tropidostoma-Endothiodon* Assemblage-zone of Keyser and Smith, (1977-78). The name "Teekloof" is favoured in preference to "Hoedemaker" on the grounds that the Teekloof Pass represents an easily accessible type section and is a permanent geographical feature, as prescribed by the International Subcommittee for Statigraphic Classification (ISSC) (Hedberg, 1976).

3.6.5 Nuweveld Member

The Nuweveld member is the most prominent arenaceous succession outcropping along the Nuweveld escarpment, which also represents the type area (FIG. 3.8). The type section is located on the hillslopes of the Nuweveld Range, south of the farm house on the farm Oukloof, Fraserburg District (APP. 1.11). The Nuweveld member is identical to the Oukloof member and falls within the *Aulocephalodon-Cistecephalus* Asseblage-zone (Smith, 1980). The "*Cistecephalus* band" described by Kitching (1977) occurs near the top of the Nuweveld member on the farm Toorwater (FIG. 3.9). The Nuweveld member is equivalent to the Oudeberg Member in the Eastern Cape Province as noted by Visser and Loock (1979a). The present name (Nuweveld member) is favoured because of the economic connotation and the fact that the name "Oukloof member" is frequently confused with the Oudeberg Member. The location (Oukloof) could also be mistaken for the Ouberg Pass near Sutherland.

The Nuweveld member consists predominantly of moderately thick sandstones, alternating with green or red mudstones and siltstones. The sandstones comprise arkosic wackes but contain less feldspar than those of the Koup and Combrinkskraal members, and they are mottled white in places. Sandstones are commonly up to 20 m thick with a maximum of 28 m measured on Wilgerbosfontein (location 4, FIG. 1.2). These sandstones are concentrated in one or two arenaceous zones near the base of the unit. The upper boundary is transitional and located at a horizon where thick sandstones become rare. The interbedded mudstones and siltstones are predominantly green, occasionally red in the northwestern part of the study area.

The Nuweveld member is prominently developed as far east as the farm Toorwater, east of Murraysburg (location 30, FIG. 1.2). From there the unit can be traced in a northerly direction to the Richmond Allotment Area, and farther north to Groot Tafelberg and Wortelfontein (locations 32 and 33, respectively, FIG. 1.2) where it was locally referred to as the "Wortelfontein sandstone" (Van Tonder, 1977; Visser and Loock, 1979a).

To the west of Murraysburg the Nuweveld member is well developed on the farms Three Sisters and Mordant Klaasenskraal (locations 28 and 29, FIG. 1.2), and on the southern slopes of the Perdeberg Mountains (location 26, FIG. 1.2). From here the unit can be traced westwards along the Nuweveld escarpment, up to Karelskraal Pass and the top of the Klein Roggeveld Mountains (FIG. 3.5). The southernmost outcrop is found on Wilgerbosfontein, Laingsburg District (location 4, FIG. 1.2). The succession has been removed by erosion in the westernmost part of the Basin, up to a line just east of the town of Sutherland. From there it trends in a northerly direction along the foothills of Bontberg, Langeberg and Tafelberg (locations 17, 18 and 10, respectively, FIG. 1.2). The northern limits of the Nuweveld member in this region occur on the farm Droogvoetsfontein near Fraserburg, and farther to the east on the farms Nieuwefontein and Graaffwaterdam near Victoria West (FIG. 1.2). The sandstones in this area and towards Richmond show sporadic and patchy uranium mineralization. Uranium mineralization is present as small to medium-scale orebodies in the Nuweveld sandstones, in the area above the Nuweveld escarpment southwest of Fraserburg.

The unit is complete only in the central part of the Basin where the thickness remains fairly constant. Typical values are 170 m at Tafelberg (location 12), 180 m at Bontberg (location 17), 155 m at Karelskraal Pass (location 14), and 175 m at Oversfontein (location 13, FIG. 1.2). The thickness increases to 260 m on the farm Toorwater in the Murraysburg District, then decreases considerably towards the northern margins of the Basin to approximately 40 m in the vicinity of the farm Wortelfontein (location 33, FIG. 1.2).

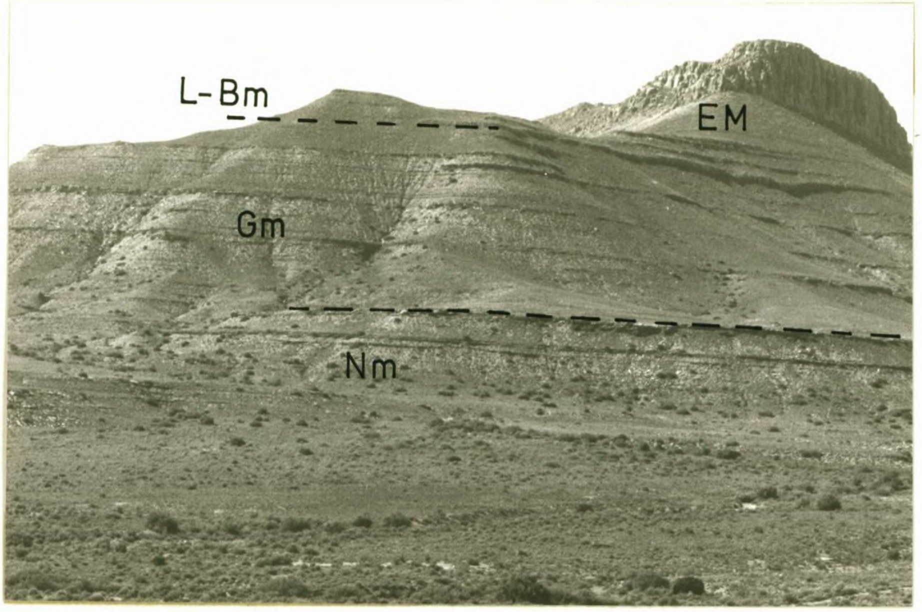

FIG. 3.9 The Nuweveld member (Nm) followed upwards by the Gifkop (Gm) and Barberskrans/Leeukop (L-Bm) members, and the Elandsberg Member (EM) near the top of the succession on the farm Toorwater (Murraysburg District).

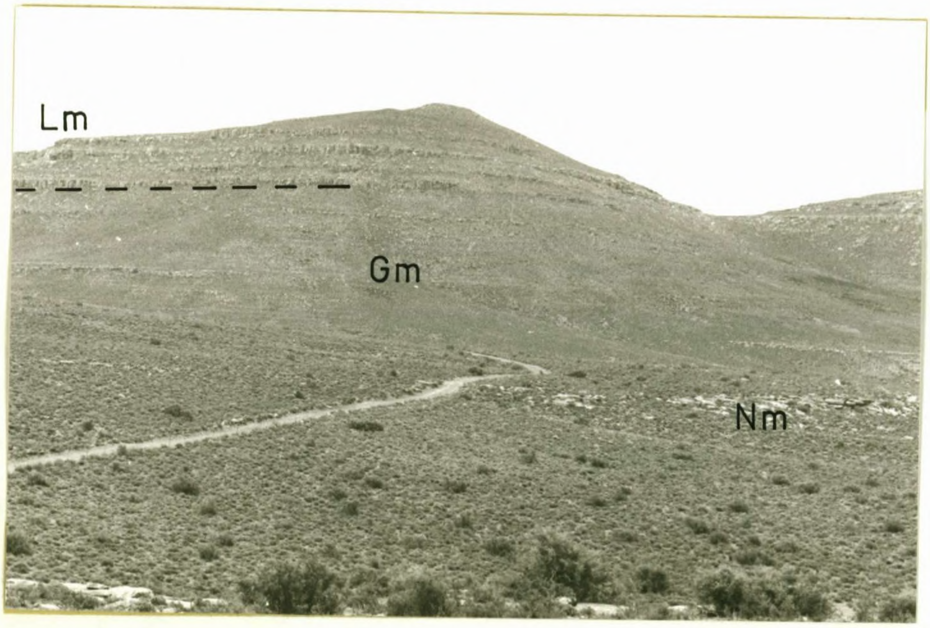

FIG. 3.10 The Leeukop member (Lm) on the farm Kruis Rivier, Fraserburg District. The Gifkop (Gm) and Nuweveld members (Nm) are poorly exposed in the foreground.

3.6.6 Gifkop Member

The Gifkop member consists typically of thick intervals of red mudstone, alternating with subordinate siltstone and sandstone lenses. Sandstones occur sparsely throughout the succession as multistoreyed lenses up to 12 m thick. The stratotype is located on the southern slopes of Gifkop, situated some 50 kilometres east of Sutherland (location 15, FIG. 1.2), where the unit attains a thickness of 185 m. The lower boundary of the Gifkop member is in places indistinct in the field, and occurs at the position where typical Nuweveld sandstones disappear and the succession changes upwards into predominantly red mudstones. The succession is rarely fully preserved, and then usually in the southern part of the Basin where it is preserved below the more resistant sandstone beds of the Leeukop member. In the northern part of the study area the unit is extensively eroded and preserved only as isolated hills consisting mostly of red mudstones, such as on Bloemfonteinkop near Victoria West (location 27, FIG. 1.2). The Gifkop member is correlated with the Daggaboersnek Member in the Eastern Cape Province (SACS, 1980), also referred to as the Boesmanskop mudstone by Visser and Dukas (1979). The Steenkampsberg member of Stear (1980a) was shown by him to overlie the Oukloof (Nuweveld) member, and it therefore forms part of the Gifkop member. The lower part of the Gifkop member coincides with the top of the *Aulocephalodon-Cistecephalus* Assemblage-zone, while the upper part falls into the *Dicynodon lacerticeps-Whaitsia* Assemblage-zone (Smith, 1980). Excellent exposures in the escarpment area surrounding Gifkop allowed good stratigraphic control but it appears as if correlation of the fossil assemblage-zones of Keyser and Smith (1977-78, Fig. 2.1) and the proposed lithostratigraphic units is somewhat problematic in that part of the Basin.

3.6.7 Leeukop Member

The lower Beaufort in the western Karoo Basin is capped by the Leeukop member (FIG. 3.9, 3.10 and 3.11). The stratotype is located on the southern slopes of Leeukop on the farm Die Kom, situated approximately 55 km southwest of Fraserburg (location 19, FIG. 1.2). The succession is structurally undeformed but extensively eroded, and preserved to a maximum thickness of 99 m on Langberg, east of Sutherland (location 18, FIG. 1.2).

The Leeukop member is a prominent arenaceous interval comprising thick sandstones alternating with red and green mudstones and siltstones. Individual sandstone beds are up to 20 m thick on Oversfontein (location 13, FIG. 1.2), and sandstones constitute up to 50 per cent of the Leeukop member. The lower boundary occurs where abundant thick sandstone lenses appear in the succession above the Gifkop member. Elsewhere the Leeukop member is preserved mainly along the Nuweveld escarpment where it can be seen on the farms Oversfontein and Tafelberg, in the Karoo National Park, on the farm Highlands, and as far north as the southern slopes of the Perdeberg Mountains (locations 13, 12, 22, 23 and 26, respectively, FIG 1.2; see also FIG. 3.11).

The Leeukop member is replaced by the Barberskrans Member in the northeastern part of the study area, and in the Eastern Cape Province (Tordiffe, 1978). The Barberskrans Member consists of up to three laterally persistent sheet sandstones in a zone up to 30 m thick in the Graaff-Reinet District, thinning to a single sandstone bed 16 m thick on Groot Tafelberg (location 32, FIG. 1.2). The Leeukop member is complexly interfingered with the Barberskrans Member in the area situated between the towns Graaff-Reinet, Murraysburg and Richmond. The Leeukop member is poorly developed on the farm Toorwater (Murraysburg District) where it is represented by a number of lenticular sandstones forming the base of the arenaceous succession in the hilltops known as Die Duiwel (location 30, FIG. 1.2). The top portion of this arenaceous unit consists of up to three laterally persistent sandstone sheets. Sandstone geometry, palaeocurrent directions and stratigraphic position (below the Elandsberg mudstones) suggest that these sandstones belong to the Barberskrans Member. Interfingering also occurs on Krugerskraal (Graaff-Reinet), where palaeocurrent measurements in a group of sandstones just below the Barberskrans Member showed that they form part of the Leeukop member (APP. 1.17). Although the Leeukop and Barberskrans Members occupy the same stratigraphic interval regionally, it appears that the Barberskrans Member is slightly younger and laps onto the Leeukop member in the area between Graaff-Reinet and Murraysburg.

3.6.8 Elandsberg Member

The Elandsberg Member was defined as an argillaceous sequence forming the top of the Balfour Formation in the Great Fish River basin (SACS, 1980; Tordiffe, 1978). The unit was described in

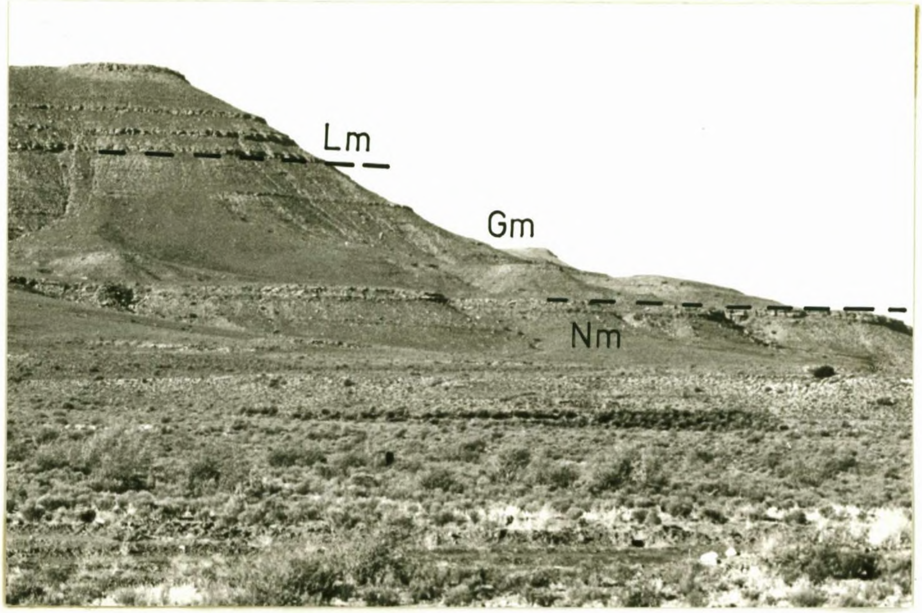

FIG. 3.11 The Leeukop member (Lm) overlying the Gifkop (Gm) and Nuweveld members (Nm) on the southern slopes of Perdeberg Mountain, southeast of Loxton.

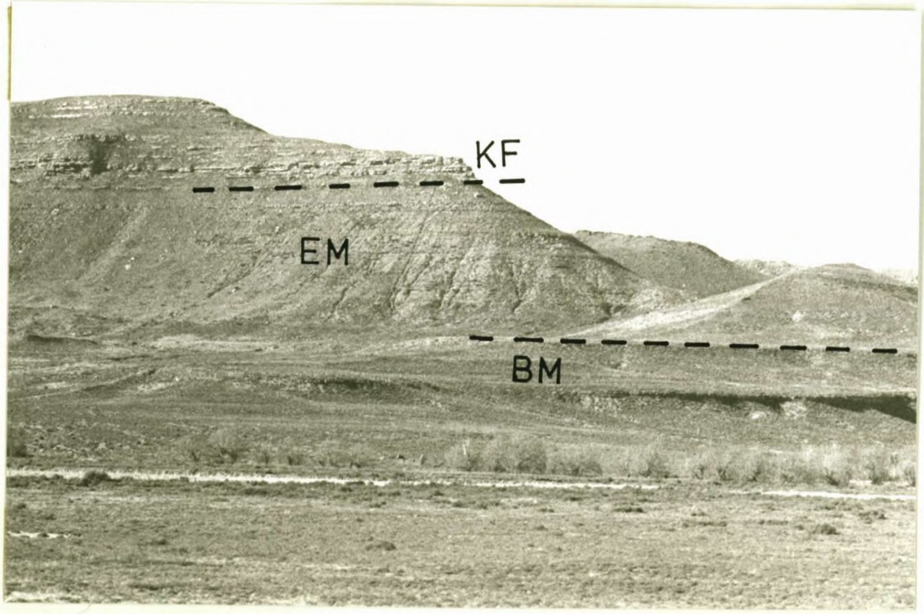

FIG. 3.12 The Barberskrans (BM) and Elandsberg Members (EM), capped by the Katberg Formation (KF), on the farm Ripplemead (Graaff-Reinet District).

the area adjoining the present study area to the east as an approximately 100 m thick argillaceous succession, located between the Barberskrans Member and Katberg Formation (Dukas, 1978). Strata equivalent to the Elandsberg Member are absent to the west of Murraysburg, and were probably eroded away.

The Elandsberg Member follows abruptly on the Barberskrans Member (FIG. 3.12), its thickness ranging between 120 m on Matjiesfontein and 135 m on Ripplemead (locations 31 and 35, respectively, FIG. 1.2). The succession consists of red mudstones with rare, thin (up to 4 m thick), sandstone and siltstone lenses. An abundance of thin sandstone lenses (showing a bleached appearance) are present near the top of the succession on Groot Tafelberg (location 32, FIG. 1.2), where the incompletely preserved Elandsberg Member is 115 m thick.

3.7 Cyclic Stratigraphy of the Lower Beaufort

3.7.1 Introduction

In the course of the present investigation few stratigraphic horizons could be traced in the field over distances of more than a few kilometres. The lower boundaries of the arenaceous members are the only regionally persistent marker horizons in the lower Beaufort in the western Karoo Basin. These boundaries are sharp and immediately obvious in the field, and where the succession is poorly exposed these contacts can be determined accurately by field correlation with known successions.

The upper boundaries of the arenaceous members are indistinct in places. The abundance of thick sandstones and lithological composition of the succession depend on the location of the section in the Basin, and vary considerably across the study area. The upper boundary of the Koup member serves as an example. This horizon is well-defined in the area between the farms Paalhuis and Marceaux, northwest of Beaufort West (FIG. 3.6), but becomes indistinct along strike. The overlying (argillaceous) Teekloof member is well-defined farther to the west, at Karelskraal Pass only 80 km away, but becomes indistinct in the area lying immediately to the north of Karelskraal Pass. In this area, surrounding Sutherland, the entire succession is distinctly arenaceous and the Teekloof member can hardly be distinguished. A similar problem is encountered in the northeastern part of the Basin where the contacts between the arenaceous and argillaceous members are indistinct in poorly exposed successions, such as on Bloemfonteinkop (location 27, FIG. 1.2). In these and many other such cases there were few benefits to be gained by rigidly subdividing the succession into discrete arenaceous and argillaceous units, and the need for a more practical scheme of subdivisions became evident. Large-scale upwards-fining sequences marked by arenaceous basal intervals and grading upwards into predominantly argillaceous intervals, thus became evident throughout the western Karoo Basin. The bases of these units are not necessarily marked by the thickest or most persistent sandstone beds but by the abundance of sandstones relative to adjacent argillaceous members.

3.7.2 Terminology and Nomenclature

Several orders of upwards-fining cycles are evident in the lower Beaufort. The most fundamental order of cyclicity is represented in the succession by cyclothems consisting of intraformational conglomerates (where present) overlying an erosive base, grading upwards into sandstones and ultimately mudstones or siltstones on top. A cyclothem represents a single sedimentary cycle (Duff *et al.*, 1967), and the Beaufort examples conform closely to those described by Allen (1964b) in the lower Old Red Sandstone. This was regarded as first-order cyclicity by Visser and Dukas (1979) in terms of the classification by Duff *et al.* (1967). The large-scale cycles in the lower Beaufort were described as ". . . regionally defineable first order . . ." cycles by Stear (1980b). Smith (1981) similarly referred to large-scale, first-order cycles based on the terminology of McLean and Jerzykiewicz (1978). In the terminology used here (after Duff *et al.*,1967) the large-scale cycles consist of a number of first-order cycles, and therefore constitute second-order cycles.

The four arenaceous and four argillaceous members constitute four large-scale, upwards-fining sequences in the western sector of the Karoo Basin, referred to here as *megasequences* after the terminology of Heward (1978). Sandstone-mudstone sequences locally developed within megasequences conform to what have been referred to as *sequences* by Heward (1978), and form part of the lower Beaufort *basin-fill sequence*. The genetic significance of these units is reviewed elsewhere.

3.7.3 Subdivisions Proposed for the Lower Beaufort in the Western Karoo Basin

Formal subdivisioning of the lower Beaufort sequence into regionally defined lithostratigraphic units would imply a fair degree of lithological homogeneity within each unit (cf. Hedberg, 1976, p. 31). That is mostly not possible and difficulties in attempts to subdivide sandstone and mudstone members in the field, even in certain well-exposed successions, have necessitated an alternative and more practical scheme of stratigraphic subdivisions. The upwards-fining megasequences are therefore proposed as informal stratigraphic subdivisions which can be identified throughout the western Karoo Basin, and correlated with similar stratigraphic units in the eastern sector of the Basin.

3.7.3.1 Megasequence A

This megasequence consists of the arenaceous Combrinkskraal member at the base, and grades upwards into the argillaceous Leeu Gamka member. The lower boundary is placed at the top of the Waterford Formation, and is therefore diachronous over most of the study area. The upper boundary is located at the sandstones marking the start of the Koup member.

Megasequence A attains a maximum thickness of 2454 m on the farm Combrinkskraal. The unit thins rapidly along strike to just under 1000 m on Wilgerbosfontein, northeast of Laingsburg. A slight thickening occurs in the Verlatenkloof area, then thinning to 920 m in the Ouberg Pass west of Sutherland. A maximum thickness of less than 600 m is calculated for the northwestern corner of the Basin, north of Sutherland. The unit is poorly exposed in the northern part of the Basin and stratigraphic control is restricted to percussion and core drilling results on Droogvoetsfontein, near Fraserburg. A thickness of just less than 200 m is calculated for this area but the succession as a whole grades diachronously into the Ecca Group to the north and northeast of Fraserburg, making accurate thickness estimates impossible. The extent and stratigraphic relationships in the eastern Cape Province are not clear but a maximum thickness of 450 to 500 m is estimated on the farm Vrede, southwest of Graaff-Reinet (location 37, FIG. 1.2).

3.7.3.2 Megasequence B

Megasequence B follows on megacycle A, and consists of the Koup member at the base and followed by the Teekloof member on top. The lower and upper boundaries are the same as for the constituent members. The succession occupies most of the southern slopes of the Kamdebo Range near Aberdeen, and grades laterally into the (undifferentiated) Middleton Formation in the eastern sector of the Basin. Except for an estimated thickness of 550 to 600 m in the Kamdebo Range northwest of Aberdeen, no reliable thickness estimates are available between this point and Wilgerbosfontein near Laingsburg in the west, where a maximum thickness of 840 m was recorded. A minimum thickness of just more than 400 m was measured in the vicinity of Beaufort West (location 22, FIG. 1.2), where the lower contact could not be determined accurately due to flat topography.

The succession thins rapidly to the north of Wilgerbosfontein and it is not much thicker than the 642 m measured on the farm Wilgerboskloof, on the Nuweveld escarpment. Megasequence B is not fully exposed in that area but a thickness of the order of 650 to 700 m is calculated by subtracting the estimated thickness of megasequence A (just more than 1000 m) in the petroleum exploration well SA1. The entire unit is preserved, and a thickness of 285 m recorded on the farm Palmietfontein, just north of Sutherland. The unit then thins towards the east to just over 200 m on the farm Droogvoetsfontein near Fraserburg, where it is also fully preserved. The unit becomes indistinct north of Loxton, and parts of it grade diachronously into the Ecca Group in the area northwest of the town of Richmond.

3.7.3.3 Megasequence C

The Nuweveld member forms the base of this unit, followed by the Gifkop member on top. The unit as a whole is correlated with megacycle 1 described by Visser and Dukas (1979) in the area surrounding Graaff-Reinet.

Megasequence C maintains a constant thickness throughout the western Karoo Basin. The thickness of the entire unit varies between 345 m just east of Murraysburg (APP. 1.15), 360 m in the Karoo National Park (APP. 1.12), 376 m on Tafelberg near Fraserburg (APP. 1.8), 338 m on Oversfontein (APP. 1.8), 340 m on Gifkop (APP. 1.9) and 356 m on Bontberg and Langberg (APP. 1.9 and 1.10, resp.). The succession is incomplete in the northeastern part of the study area but a minimum thickness of 198 m was determined at Klein Tafelberg (APP. 1.16) near Richmond. From here the succession persists in an eastern and northeastern direction beyond the study area.

3.7.3.4 Megasequence D

The succession is incompletely preserved in the western part of the Karoo Basin. The Leeukop member forms the base of the succession but due to erosion the topmost (Elandsberg) argillaceous member is absent in the western part of the study area. In the easternmost part the succession is capped by the Elandsberg Member, as for example in the Toorwater profile. Megasequence D is correlated with megacycle 2, described by Visser and Dukas (1979) in the Graaff-Reinet area. The entire unit is nearly 180 m thick in the area northwest of Graaff-Reinet (APP. 1.16), where it is fully preserved.

3.7.4 Discussion

The lower Beaufort megasequences are well-defined units which are easily recognised in the field. Despite considerable differences in thicknesses of the individual megasequences, they facilitate the subdivisioning of a considerable thickness (>4000 m) of lower Beaufort strata into useful stratigraphic intervals throughout the western Karoo Basin. Although the lower contact of the Beaufort Group is diachronous throughout the western Karoo Basin the contacts between megasequences are useful time-lines. Regional unconformities are not known to exist within the succession and the megasequences are therefore useful stratigraphic intervals for isopach studies, as well as for comparative palaeocurrent and lithological studies in that part of the Basin.

4.0 LITHOLOGY AND SEDIMENTARY STRUCTURES

The sedimentary structures, lithological composition and fossil content of sedimentary rocks form a basis for sedimentary facies analyses. These are the large-scale organisational elements of sedimentary rock strata that enable palaeoenvironmental reconstructions of sedimentary successions. Statistically significant, repetitive processes in sedimentary successions can be detected by using mathematical techniques, of which the Markov chain analysis is the most widely used (cf. Miall, 1973). The upwards-fining point bar model, for example, was resolved by observing vertical repetitions of related rock (facies) types in the rock profile (Allen, 1963a; Bernard *et al.*, 1962), in a manner similar to modern facies analyses.

Despite the apparent lithological monotony a variety of rock types and sedimentary structures exists within the mudstone and sandstone successions of the upper Ecca and lower Beaufort. Certain distinctive rock types such as the tuffs and "cherts" outcrop very rarely and they occur in restricted areas and in a specific stratigraphic interval. Similar examples are found also amongst sedimentary structures. Soft-sediment deformation structures, for instance, are abundant in the upper Ecca but rare in the lower Beaufort. Individual structures are rarely diagnostic on their own but in conjunction with other parameters they are useful in lithostratigraphic mapping, and to subdivide depositional subenvironments.

4.1 Petrology and Geochemistry

The uncomplicated lithological composition of the Beaufort succession has discouraged investigations of a specialised nature, such as geochemical and thin section petrographic studies. Although carried out routinely in the course of mineral exploration, mineralogical and geochemical studies were rarely reported (e.g. Kübler, 1977; Moon, 1977.) The benefits of investigations of this nature in provenance studies, amongst others, have become evident in recent years (Blatt, 1985).

In the present investigation textural and obvious compositional variations were logged in the course of field investigations. Rock samples were collected regionally across the study area, for petrographic and geochemical studies on a limited scale. Microscopic grain-size measurements were carried out in the course of sandstone petrographic examinations. A number of rock samples were also analysed by multi-element X-ray spectrography, X-ray fluorescence spectroscopy (TABLE 4.1), and semi-quantitative X-ray diffractography. Some delicate mineral constituents and rock fragments, as well as bulk geochemical compositions of rocks were altered by low grade metamorphism, particularly in the southern (deepest) parts of the Basin.

4.1.1 Rudites

4.1.1.1 Extrabasinal Clasts

4.1.1.1.1 Description

Conglomerate beds of extrabasinal origin are unknown in the lower Beaufort in the western Karoo Basin but isolated occurrences of exotic clasts have been reported from time to time. Broom (1912) first made reference to exotic clasts which he found associated with fossilised plant material on the farms Koornplaats and Spitskop, in the Moordenaars Karoo. These were up to the size of an "orange" (small cobble) and consisted of quartz porphyry, quartzite and schist. Local farmers pointed out several small (unidentified) pebbles at the base of a channel sandstone bed on the farm Agtersteland (Fraserburg District), to A.N. McLaurin (pers. comm., 1977). Access to this property was at the time restricted by exploration activity and the site could not be visited. The clasts have since been removed, except for one small quartzite pebble. In addition to these specimens large feldspar-porphyry and gneiss pebbles were found in a road cutting approximately 3 km south-west of Prince Albert Road by J.N.J. Visser (pers. comm., 1977). This section of rock has since been excavated in the process of road construction. A fine-grained quartzite pebble on the farm Voetpad (Murraysburg District), as well as a quartz-porphyry clast on the farm Dowefontein (Aberdeen), were found by W.P. Karpeta (pers. comm., 1985). Scrutiny of intraformational conglomerates for the presence of exotic clasts in these particular areas and elsewhere in the western Karoo Basin, has failed to yield more specimens.

The clast finds are not restricted to one particular area within the Basin, nor to a specific

TABLE 4.1 Geochemical compositions of selected whole-rock samples from the lower Beaufort.

No.	Rock Type	Element (ppm)*												Element (ppm)**			
		Zr	Mn	Ba	Pb	Zn	Sn	Be	Mo	Cu	Ni	Co	As	Rb	W	U_3O_8	ThO_2
701	Tuf: Beaufort Gp	120	200	950	10	12	5	0,8	<5	10	<10	<10	<100	187	<8	11	62
702	Sandstone: Koup mr	280	350	1500	<10	25	8	1,0	<5	8	30	<10	<100	70	<8	6	20
703	Sandstone: Ripon Fm	350	350	950	<10	20	5	0,8	<5	8	10	<10	<100	65	<8	8	27
704	Tuf: Ripon Fm	220	650	1200	<10	40	8	1,5	<5	5	<10	<10	<100	322	<8	15	55
705	Tuf: Ripon Fm	400	180	1500	40	40	5	1,0	<5	8	12	<10	<100	256	<8	20	69
706	Tuf: Ripon Fm	220	80	1500	10	40	5	0,5	<5	8	<10	<10	<100	256	<8	18	74
707	Sandstone: Waterford Fm	280	400	1500	<10	40	8	0,8	<5	5	10	<10	<100	53	12	11	40
708	"Chert": Collingham Fm	100	500	350	<10	<10	<5	<0,5	<5	8	<10	<10	<100	<5	<8	9	37
709	Green mudstone: Beaufort Gp	550	450	2000	10	45	8	1,5	<5	20	25	<10	<100	143	<8	<1	29
710	Sandstone: Combrinckskraal mr	150	300	800	<10	25	5	0,8	<5	5	<10	<10	<100	51	<8	4	20
711	Calcareous sandstone: Combrinckskraal mr	120	>5000	750	<10	30	5	<0,5	<5	25	<10	<10	<100	45	<8	41	15
712	Red mudstone: Koup mr	150	200	950	12	20	5	2,0	<5	8	25	<10	<100	229	<8	4	38
713	Uraniferous sandstone: Koup mr	220	>5000	1000	25	25	8	0,8	25	1000	25	10	900	38	13	224	17
715	Uraniferous sandstone: Nuweveld mr	120	180	1800	50	10	<5	0,5	100	5	15	<10	1500	107	20	4435	42
716	Uraniferous sandstone: Nuweveld mr	60	50	1500	15	<10	<5	<0,5	150	<5	<10	40	<100	69	<8	57	20
717	"Chert": Beaufort Gp	150	300	1500	<10	<10	5	1,5	<5	<5	<10	40	<100	108	<8	15	45
718	Uraniferous sandstone: Beaufort Gp (Edenburg District)	100	>5000	1000	10	15	8	0,5	280	<5	<10	40	<100	38	9	145	26

TABLE 4.1 (Continued)

No.	**Rock Type**	**Element (ppm)***												**Element (ppm)****			
		Zr	**Mn**	**Ba**	**Pb**	**Zn**	**Sn**	**Be**	**Mo**	**Cu**	**Ni**	**Co**	**As**	**Rb**	**W**	**U3O8**	**ThO2**
719	Uraniferous sandstone: Koup mr	120	>5000	900	30	30	8	0,5	30	<5	25	50	<100	76	<8	743	25
722	Uraniferous sandstone: Koup mr	280	<50	950	20	90	5	<0,5	<5	<5	20	50	<100	77	9	1262	34
723	"Chert": Beaufort Gp	180	200	1800	25	<10	5	0,5	<5	8	15	<10	<100	63	<8	8	23

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* Analyses by multi-element X-ray spectrography, by Anglo American Research Laboratories, Crown Mines.

** Analyses by X-ray fluorescence, by Anglo American Research Laboratories, Crown Mines.

Sb<50 ppm, Li<100 ppm, La<100 ppm, Ag<0,1 ppm for all samples

Abbreviations: Gp = Group, Fm = Formation, mr = member stratigraphic interval. All the clasts were found in the arenaceous members of the lower Beaufort. Those reported by Broom (1912) were located in the southwestern corner of the Basin in the Koup member where it reaches its maximum thickness. The clasts on Agtersteland and Voetpad were both located near the base of the Nuweveld member but were deposited by two different palaeodrainage systems, the first entering the Basin from the southwest and the second from the southeast. Those from Prince Albert Road came from near the top of the Combrinkskraal member. The clast from Dowefontein was collected in the Koup member. Significantly, the majority of the clasts came from near the bases of the upwards-fining Koup and Nuweveld megasequences where they are not only restricted to the intraformational conglomerate beds, but are also found as isolated clasts in the sandstones.

4.1.1.1.2 Discussion

The clasts were derived from a mixed source of sedimentary, igneous and metamorphic rocks. Broom (1912) argued that the abundance of fossil wood associated with the clasts he found, were proof that they had been rafted in by tree trunks carried in by rivers over considerable distances. Allen and Williams (1979) in their study of the Lower Old Red Sandstone in Wales, concluded that those conglomerates with exotic clasts were emplaced by major drainage systems originating outside the basin, while those lacking such clasts were deposited by local drainage systems eroding interfluvial ridges between major rivers within the basin. Considering the paucity and almost random distribution of the clasts in the lower Beaufort, their presence can best be explained by emplacement following catastrophic tectonic or flooding events in the source. It is possible that some of the quartzite clasts could have been derived from quartzitic strata of the Cape Supergroup located to the south and southwest of the study area.

4.1.1.2 Intraformational Conglomerates

4.1.1.2.1 Description

The lower Beaufort conglomerates consist of red and green mudstone and calcareous clasts set in fine-grained sandstone matrices. They occur as discrete conglomerate lenses or rarely as scattered clasts at or near the bases of major channel sandstone bodies. Occasionally small-pebble and granule layers line foreset beds in trough cross-bedded sandstones. Multistoreyed channel deposits commonly contain a number of conglomerate bands. At least eight superimposed storeys are marked by conglomerates in a 12 m thick channel sandstone in the Ouberg Pass (west of Sutherland). Individual beds may vary between a few centimetres and over 1 m in thickness. Kübler (1977) reported thicknesses of up to 2 m. Mean conglomerate thicknesses of between 0,25 and 0,35 m in major channel sandstone deposits appear to be largely independent of the thicknesses of channel sandstones or individual sandstone storeys.

The intraformational conglomerates in the Karoo were named "clay-pellet conglomerates" by Schwarz (1897), a term which has remained in popular use over the years. Clasts consist mostly of red and green mudstone but calcareous clasts in many instances predominate in the conglomerates of the Combrinkskraal and Koup megasequences. The latter conglomerates tend to be thicker than the mudstone-clast conglomerates, as a result of the durability of the calcareous clasts. Rolled fossil bones, plant fragments, and rare reptillian teeth are sparsely present amongst other intraclasts. An abundance of vegetal material, usually ferruginised, in places accompanies U_3O_8 mineralization. In those cases exceptionally high grades are evident, as for example on Vintragersfontein where much of the mineralization is hosted in thick intraformational conglomerates, associated with an abundance of fossil plant material. The matrix usually consists of fine-grained sandstone which in surface exposures is kaolinised, ferruginised, or calcitised, particularly where associated with U_3O_8 mineralization.

The intraformational conglomerates are clast supported and lack a distinct internal stratification. A crude horizontal or low-angle cross-stratification is rarely seen. Clasts are moderately sorted but as a rule show no discernable internal organisation. The lack of a fabric may be the a result of the almost equant clast shapes. A distinct clast imbrication was observed only once, near the top of a 19 m thick channel sandstone in Verlatenkloof.

The clasts range in size from small pebbles (<16 mm) to cobbles (> 64 mm) with a maximum size of just over 250 mm. (These do not include the isolated, large, angular blocks of green siltstone "floating" in sandstone.) Schwarz (1897) reported mudstone clasts several feet in diameter, but it is not clear whether they occurred as single blocks, or as clasts in intraformational conglomerates. A distinctive variety of intraformational conglomerate contains large, dark grey or black, fissile mudstone (shale) clasts. They are here informally referred to as the "black shale conglomerates" (FIG. 4.1). Clasts are up to 250 mm in length, subrounded, and in places show signs of soft-sediment contortion. Larger clasts are often seen to have been fragmented into clusters of smaller, angular clasts. Smith (1972) noted that delicate bedding planes (fissility) as seen in FIG. 4.1, would enhance disintegration of such clasts. Clasts had clearly undergone very little transportation and rounding prior to deposition. These conglomerates are restricted to a few thick channel sandstones in the Combrinkskraal and Koup members in the southern part of the study area.

Another unusual variety of intraformational conglomerate comprising angular, ferruginised (hematitic) clasts set amongst unaffected red and green mudstone intraclasts, was found in a sandstone bed in the Combrinkskraal member on the farm Karoluspoort. The hematite clasts are generally less than 30 mm in diameter. They occur in a well-packed conglomerate bed (40 to 50 cm thick) which is exposed along several hundreds of metres of strike length. No likely source rock or comparable clasts have yet been encountered anywhere in the study area, but their angularity, large size, and abundance seem to preclude a distal, extrabasinal source. They are likely to have been derived as partly consolidated, but already ferruginised sediments from the upstream sections of the alluvial system. The alluvium on these alluvial fans is in places exceptionally rich in ferromagnesian minerals which are unstable in oxidising conditions (Blatt, 1982). The ferrous iron is readily released to form Fe_2O_3 early during diagenesis.

FIG. 4.1 Cobble-sized, dark-grey shale intraclasts and fragments at the base of a channel sandstone in the Combrinkskraal member on the farm Kentuckey, Sutherland District.

FIG. 4.2 Large calcareous ("cannon ball") concretion in plane bedded and ripple cross-laminated sandstone. Note the sharp contacts of, and primary sedimentary structures preserved in the concretion.

4.1.1.2.2 Discussion

Intraclasts by virtue of their origin as cohesive but unlithified sedimentary fragments could not be expected to withstand transportation over great distances. Fagerstrom (1967) suggested that mud crusts formed by desiccation were the main source of mudstone clasts in intraformational conglomerates. Smith (1972), on the other hand, demonstrated experimentally that dried mud fragments underwent rapid attrition during transportation and that they were negligible sources of intraclasts. Smith also found that small mud clasts survived transportation over distances of no more than a few hundred metres from their source. Karcz (1969) described the formation of mud clasts in modern, ephemeral streams by mud cracking and collapsing of channel banks during dry periods. It is unlikely, though, that clasts formed in this way would remain preserved until the next flooding event. Rust (1984) concluded that desiccation cracking and bank collapse could be a significant source of intraclasts only if the mud clasts were large enough to remain moist inside, thus retaining a cohesive core for the duration of the transportation process.

Allen (1964a) considered cohesive levee and backswamp silts (muds) as the main source of intraclasts, formed by corrasion and collapse of channel banks (Fisk, 1947, quoted by Allen, 1964a). The few examples of possible collapsed overbank material (massive green mudstone blocks up to 30 cm in diameter) in the Beaufort strata also matched the overbank sediments in their immediate vicinity. Intraclasts could thus have been formed by further disintegration of the collapsed blocks of mud in streams. The paucity of these blocks (only two examples in 25 000 m of stratigraphic sections) in lower Beaufort channel sandstones furthermore suggests that the streamflow in these streams was powerful enough to dislodge and also disintegrate fairly cohesive mud fragments. Also, perennial rather than intermittent discharge took place in most lower Beaufort streams. A higher incidence of desiccation, erosion and burial of bank material during periods of falling water level, and more frequent preservation of collapsed mudstone blocks would have been expected in the event of intermittent discharge (cf. Gibling and Rust, 1984; Rust, 1984).

The formation of mudstone clasts by desiccation of muddy layers, as proposed by Fagerstrom (1967), is more likely to have taken place on floodplains away from major channelways. The abundance of small, angular mudstone clasts seen in a few crevasse splay sheet sandstones, for instance in the Nuweveld member on the farm Toorwater, could well have been derived in this manner from mud desiccation. The crevasse splays were formed in an environment (floodplain) where mudstone desiccation would have provided an abundance of such clasts, and which would have been incorporated with minimal reworking.

4.1.2 Arenites

4.1.2.1 Field Description

Sandstones occur mainly as discrete sheets and ribbons intercalated with mudstone, in ratios of sandstone to mudstone varying regionally between 1:2 and 1:6. The channel sandstones reach a maximum thickness of 60 m in the Combrinkskraal and Koup members. They occur mostly as composite, multistoreyed beds with individual storeys up to 6 m thick. Up to ten cyclical units occurring in such sandstones are bounded by subtle to well-defined, erosional contacts. Sandstones in the uppermost members in the succession, particularly in the northern half of the Basin, show at least two or three superimposed, upwards-fining sequences separated by high-relief erosional contacts.

Beaufort sandstones in outcrop typically weather in colours ranging between deep yellow and dull green, depending mainly on the presence of feldspar, ferromagnesian minerals, and argillaceous matrix. Red streaks and mottling are present in places and can be ascribed to weathered ore grains as well as bioturbation. In hand specimens the visible constituents are usually (weathered) feldspar, quartz, white mica (sericite), coalified vegetal fragments, and oxidised ore grains, set in a kaolinitic matrix. Feldspathic sandstones (e.g. the Koup sandstones in the type area) weather into sugary lumps whereas those with noticeably less feldspar (e.g. the Leeukop sandstones) typically weather as large, angular blocks. The deep yellow ("buff") colour is quite distinctive in certain areas, such as just west of Beaufort West where the Koup member was initially named the "zone of golden sands" by R.M.H. Smith, and referred to as such in Turner (1979).

Sandstones with an abundance of pyrite show extensive alteration following oxidation and leaching by acidic solutions generated in the process. The result is a bleached, kaolinitic sandstone showing Liesegang banding and, in many instances, brightly coloured secondary uranium and copper minerals. Fine carbonaceous fragments and lithic grains have become accentuated as a result of the bleaching of these sandstones.

Pods of calcareous sandstone, locally known as "koffieklip", are a conspicuous feature of major sandstones and are sporadically associated with uranium mineralization. They vary in nature from poorly defined replacement bodies, tens of metres long, to sharply demarcated, spheroidal nodules generally less than 2 m in diameter (FIG. 4.2). In exceptional cases the calcareous replacement may be more extensive. Two feldspathic sandstones in the Ecca-Beaufort transitional zone on the farm Groot Fontein (location 4, FIG. 1.2), 9 and 7 m thick, respectively, are extensively calcitised over considerable distances along strike. Dark, calcareous replacement bodies are abundantly present also in the lowermost sandstones of the Combrinkskraal member in the area just north and north-west of Prince Albert where this rock-type was described as "marl" by Rossouw *et al.* (1964).

Spherical calcareous nodules in sandstones are sometimes referred to as "kugelsandstein" (Pettijohn, 1975). They commonly show evidence of nucleation around intraclasts and fossil bone fragments. The continuity of sedimentary structures in the sandstones suggest that their formation took place before final lithification of the sandstone, probably during late-diagenesis as a result of carbonate segregation and pore filling (Pettijohn *et al.*, 1972). Pettijohn (1975) suggested that the sizes of these concretions are determined partly by the permeability of the host rock (at the time of formation of the concretion), a control which appears to have applied to the Beaufort sandstones as well. Here the feldspathic and immature sandstones of the Combrinkskraal and Koup members show extensive, irregular calcite replacements, whereas the texturally more mature sandstones contain well-defined nodules. Large, spherical nodules (up to 2 m in diameter) are nevertheless also found in the immature Waterford sandstones, together with irregular calcareous replacements. In these sandstones two spherical concretions may be partially connected to form peanut-shaped concretions.

The sandstones in areas such as in the Waterford Formation in the Verlatenkloof, the Nuweveld member at places along the Nuweveld escarpment, and notably the Koup member in the Kamdebo Range, show a very distinctive white mottling. Occasionally mottled sandstones show negative weathering of these textures (FIG. 4.3). The mottled sandstones commonly show flaser bedding in

FIG. 4.3 Negative weathering associated with mottling in a tuffaceous sandstone in the Nuweveld member, on the farm Oversfontein.

ripple cross-laminated intervals. Small-scale soft-sediment deformation (in places as perfect ball-and-pillow structures) is also present in intensely mottled sandstones on the farm Uitkyk (localion 38, FIG. 1.2) in the Aberdeen District. The presence of rusty coatings on some mottled sandstones suggests that the mottling is associated with an anomalous Fe and/or Mn content. Analyses done by Fuller (1970) suggested that the mottling was related to the presence of the zeolite laumontite, a low-grade metamorphic product of tuffaceous material, presumably deposited originally with the channel sands.

4.1.2.2 Petrography

The sandstones in thin section show fine to very fine-grained quartz, plagioclase, potassium feldspar, lithic fragments, mica, opaque ore grains, and occasional heavy minerals, set in a fine-grained matrix. In the case of "koffieklip" the intergranular voids are filled with calcite replacing the detrital matrix and in places the grains as well. The quartz grains usually show straining although clear, unstrained grains of possible volcaniclastic origin are also present (Ho-Tun, 1979). The feldspars had been altered to some extent and the fine-grained, chloritised and sericitised product is difficult to distinguish from the matrix. The plagioclase has been reported as andesine-labradorite (Kübler, 1977) and albite-oligoclase (Moon, 1974, quoted in Kübler, 1977), and some grains have been altered to sericite (Wallace and Van der Merwe, 1978). Lithic grains include chert, polycrystalline quartz, igneous fragments and small siltstone grains. The matrix consists of fine sericite, chlorite, carbonate, feldspar and quartz (Kübler, 1977).

Modal analyses by Hotton (1967), Kübler (1977), and Moon (1977) have shown that both the feldspar and matrix contents of Beaufort sandstones approach 25 per cent of the rock. Martini (1974) reported up to 30 per cent volcaniclastic fragments in sandstones from the lower Beaufort in the southern parts of the study area. Depending on the feldspar and lithic contents the sandstones can be classified arkosic or lithic wackes, according to the scheme of Dott (1964, modified by Pettijohn, 1975). The high matrix content has furthermore reduced the permeability and primary porosity of these sandstones, and has so impeded or even prevented circulation of epigenetic, uranium-bearing fluids.

4.1.3 "Cherts"

A fine-grained sedimentary rock type resembling chert in outcrop (Haughton *et al.*, 1953) occurs as thin (<1 m) beds intercalated with green mudstone. Individual beds are persistent over several kilometres (Rossouw *et al.*, 1964). They occur exclusively in the Leeu Gamka member and the basal parts of the Koup member, in the Koup region as well as along the southern margin of the Basin.

Fresh "cherts" are greyish-olive to light grey in colour, and in certain instances show a pinkish or reddish tinge. They are composed mainly of angular quartz and plagioclase grains in a sericite matrix, with subordinate apatite and mica (Rossouw and De Villiers, 1952). Martini (1974) reported the presence of vitric tuff in certain "chert" samples and proposed a volcaniclastic origin for all the Beaufort "cherts".

The Beaufort "chert" beds are associated with calcretes in places (Rossouw and De Villiers, 1952), and Smith (1981) concluded that they were silicified, calcareous evaporites. The "chert" beds nevertheless show current-generated sedimentary structures such as plane bedding and ripple cross-lamination, suggesting clastic deposition. On account of sedimentary structures and the dominant role of detrital sedimentation in the postulated depositional environment, it is unlikely that true cherts could have formed (cf. Blatt, 1982). The term "chert" is therefore unsuitable for the Beaufort examples but as it has by now become ingrained in the terminology it will be retained provisionally.

Assuming a detrital origin for the "cherts", the composition and mode of deposition of the sediments are seen as the most important factors which determine their physical characteristics. These fine-grained sediments were rapidly deposited without much sorting, and some clay was trapped in intergranular spaces. The dense packing prevented groundwater circulation and delicate tuffaceous particles were preserved as a result (Martini, 1974). The dense, siliceous texture of these rocks also suggests post-depositional silicification which could have resulted from *in situ* devitrification of felsic tuffaceous particles. The paucity of "cherts" and the anomalous Zn-Cu-Pb content of some of these beds suggest a slightly different origin to the enclosing rocks, the latter usually mudstones. As those parts of the Basin to which these "cherts" are confined, were most severely deformed, the effects of deep burial and tectonic stresses may have contributed to the presently indurated state of these rocks.

4.1.4 Mudrocks

4.1.4.1 Classification

The terminology of the rocks here referred to collectively as mudstones (Blatt *et al.*, 1972) has been reviewed often in recent years, despite their relatively simple lithological characteristics. Mudstones consist of grains with sizes generally finer than 0,0625 mm (Blatt *et al.*, 1972). Picard (1971) proposed a cumbersome classification based on the texture and composition of the clay minerals in mudrocks. Pettijohn (1975) emphasised the state of lithification, and proportions of clay (<0,0039 mm) and silt-sized (<0,0625 mm) grains. Potter *et al.* (1980) named all these rock types "shales" and classified them on the basis of texture, proportion of clay-sized grains, sedimentary structures, and degree of induration.

The fine-grained rocks are subdivided here according to a simple and practical scheme based on texture and field characteristics. Mudstones are massive, soft-weathering with an even texture. Shales, on the other hand, are distinctly laminated but occur very rarely in the lower Beaufort. In hand samples siltstones appear and feel coarse and granular, but clastic grains are subordinate to the argillaceous matrix. Although argillaceous rocks are rigidly classified according to grain sizes (reviewed in Pettijohn, 1975) it was pointed out by Krumbein and Sloss (1951) that this scheme is unpractical as "typical siltstones" change to "typical mudstones" at grain sizes smaller than approximately 0,01 mm, instead of 0,0039 mm. In order to avoid a confusing proliferation of terms all argillaceous rocks, except where specified otherwise, will here be referred to as "mudstones".

4.1.4.2 Composition

The Beaufort mudstones are composed of fine-grained quartz, feldspar, and clay minerals (Moon, 1977). The distinct lamination in certain Beaufort mudrocks is a result of both their mode of deposition (from suspension in standing bodies of water), and their content of clay plus mica (Pettijohn, 1975). The distinctive fissility of the "black shale" intraclasts in certain conglomerates is typical of shales with a high carbon content (Pettijohn, 1975).

4.1.4.3 Geochemistry

The Beaufort mudstones usually show consistent values for trace element content. Primary uranium concentrations are rare in mudstones (Cole, 1980) but elevated thorium concentrations (over 200 ppm) were recorded in red mudstones of the Gifkop member at a few places. Beaufort mudstones have a slightly higher thorium and rubidium content than sandstones, the latter element probably associated with the kaolinite content of the mudstones.

The argillaceous fraction of modern stream sediments derived from the Koup and Nuweveld members has an anomalous base metal content (Zn-Pb-Cu-Mo), and anomalous areas overlap with known uranium districts. Apart from Pb (some of which may be radiogenic), these value trends are apparently unrelated to stratiform base metal or sandstone-hosted uranium mineralization, but reflect a higher background content of these elements in the argillaceous country rocks. Visible, secondary copper mineralization is present in some mudstones, as for example some green mudstones near the base of the Combrinkskraal member in Ouberg Pass. The poorly defined base metal content (Zn-Pb-Cu-Mo) could have been derived from primary sources in the source area, or from interbedded tuffaceous sediments. "Chert" beds associated with occurrences of tuff in the Leeu Gamka member contain anomalous Zn values of up to 600 ppm, i.e. between 8 and 10 times the background content.

4.1.4.4 Sedimentary Structures

Beaufort mudstones typically lack small-scale internal structures but usually show a pronounced bedding on a scale of a few tens of centimetres. There is evidence that massive bedding in mudstones resulted mainly from homogenisation within individual beds, mainly as a result of bioturbation, pedogenesis, compaction, or dewatering. Soft-sediment deformation, notably convoluted lamination or load casting, contributed locally to the process. The effects of (differential) compaction, for instance, are illustrated by contorted sandy intercalations in otherwise massive mudstones, reaching a amplitude of more than 1 m in severe cases.

The shaly Britskraal Member is locally present at the base of the Combrinkskraal member in the southern parts of the study area, notably on the farm Combrinkskraal. Fine laminations are also present locally in mudstones of the Leeu Gamka member in this area, giving way in places to wavy laminations. Horizontal and wavy laminations are also seen in the basal mudstones of thin upwards-coarsening sequences in overbank sediments. The wavy laminations show a climbing tendency, a wavelength of 40 cm and amplitude of 8 cm occurs in one such example near the base of the Combrinkskraal member in Ouberg Pass, west of Sutherland.

4.1.4.5 Colour

Routine comparison of fresh, wet mudstone samples with the Rock-Color Chart of the Geological Society of America (Goddard *et al.*,1975) showed a narrow range of colours. Mudstones in outcrop are usually greyish-olive to olive-grey or rarely dark-grey, and greyish-red to dusky red, or variegated combinations of these colours. Fresh mudstones recovered from depth in exploration drilling are generally darker (usually medium-grey, or olive-grey) than those on surface. They are here referred to simply as "green", "red" or (rarely) "grey" in colour, and they are an important parameter used in the classification of the fine-grained facies. Mudstones of a particular colour are not stratabound, and predominantly green mudstones in the southern part of the Basin change to predominantly red mudstones in the northern part. The presence of a thin red mudstone interval (a few metres thick) near the top of megasequence B in the Palmietfontein section (location 10, FIG. 1.2), proved useful initially in locating the contact with the overlying megasequence (C), in an otherwise problematic area.

The colour of mudstones is controlled mainly by the clay, carbon and Fe contents (Potter *et al.*,1980). The carbon content largely controls the oxidation state of Fe and thus also the Fe^{2+}/Fe^{3+} ratio. Tomlinson (1916, quoted in Pettijohn, 1975) found that the total Fe content of mudstones remains approximately constant (*ca.* 5%) but that red mudstones generally have a lower Fe^{2+}/Fe^{3+} ratio than green mudstones. Friend (1966) determined by using X-ray diffraction techniques and chemical analyses that the red pigmentation in mudstones was caused by fine hematite, and concluded that the hematite formed *in situ*. Van Houten (1968) argued that the Fe in red mudstones was derived initially as iron-hydroxide from lateritic soils in the source area and was converted to hematite after burial, thus invoking an Fe-rich source. More recent information, however, shows that in adjacent red and green mudstones the total Fe content is in many cases lower in the green mudstones, suggesting loss of the more mobile Fe^{2+} during diagenesis (Blatt *et al.*, 1972).

The green mudstones owe their colours primarily to the presence of green phyllosilicates such as illite and chlorite. They usually contain small amounts of pyrite as well but according to Potter *et al.* (1980) this does not have an effect on the colour of these rocks. The darker colours of the buried mudstones are probably the result of the presence of finely divided carbon, effectively preserved below the level of oxidation.

4.1.4.6 Palaeoenvironmental Aspects of Mudstone Colouration

Red mudstones are popularly linked to an arid, terrestrial depositional environment (Pettijohn, 1975). The environment nevertheless had only an indirect control as the colours are more directly related to the carbon content of the sediments. Certain red mudstones of the Beaufort Group are pigmented as a result of pedogenic alteration. This process was in turn expedited by the effects of bioturbation and abundant vegetation (cf. Buurman, 1975), two factors which discount a very dry climate. The answer probably lies therein that the colour of mudstone is the result of several different processes and environmental controls, including pedogenic processes. The controls envisaged to have played a role include climate, bioturbation, the burial history, and diagenesis of the alluvium. It appears that other than mild oxidation and bleaching, the mudstones had undergone very few chemical changes since diagenesis (cf. Blatt, 1985).

4.1.4.7 Calcareous Concretions

Calcareous concretions and segregations are common in mudstones throughout the lower Beaufort. Two distincive types are found. The first (type 1) occurs as spherical or oblate shapes in red and green mudstone, usually in discontinuous layers and with the bedding of the host rock flowing around the nodules (FIG. 4.4). Most of these are of early diagenetic age and predate compaction of the sediment. Together with thin, calcareous laminae they occur in pedogenic horizons in the overbank successions (McPherson and Germs, 1979; Smith, 1981).

Septarian nodules are the second type and they are characterised by radial, calcite-filled cracks as seen in cross-section. They are commonly enclosed in a brown, powdery crust. The radial shrinkage cracks in septarian nodules were caused by case hardening, followed by chemical desiccation (Pettijohn, 1975). The upper parts of overbank mudstone beds in the Combrinkskraal and Leeu Gamka members contain continuous layers of these nodules (type 2). Some of these layers occur over areas of hundreds of square metres in parts of the southern Karoo underlain by the Leeu Gamka member, and nodules occur abundantly on weathered surfaces. They are considered to be of pedogenic origin, like all other calcareous nodules in mudstones, but it is uncertain what conditions were required to form septarian nodules.

4.1.8 Volcaniclastic Rocks

Following Fuller's (1970) report on the presence of laumontite of possible volcaniclastic origin in Karoo rocks, there had been much speculation but very little concrete evidence of volcaniclastic material in the Lower Beaufort succession.

Based on petrographic studies which showed the presence of glass shards Martini (1974) concluded that the "cherts" were reworked tuffs. Crystal tuffs were reported in the Ecca Group in the Eastern Cape Province by Lock and Johnson (1974). The uranium mineralization in the Beaufort Group was linked to the presence of tuffaceous material in these beds (Moon, 1977). Ho-Tun (1979) reported the presence of volcaniclastic material in calcareous, mineralized sandstones.

Despite the widespread presence of tuffaceous particles and their diagenetic products in the Karoo rocks, albeit in small quantities, there has in the past been little evidence of how this material was originally deposited. In this investigation a soapy, dusky yellow to light olive-grey, radioactive, benthonitic bed was found interbedded with green mudstones in the upper part of megasequence A, and is interpreted as a primary ash-fall deposit (FIG. 4.5). A similar origin is accepted for a similar type of rock intersected at depth in percussion drilling on Gannaleegte near Aberdeen (location K,

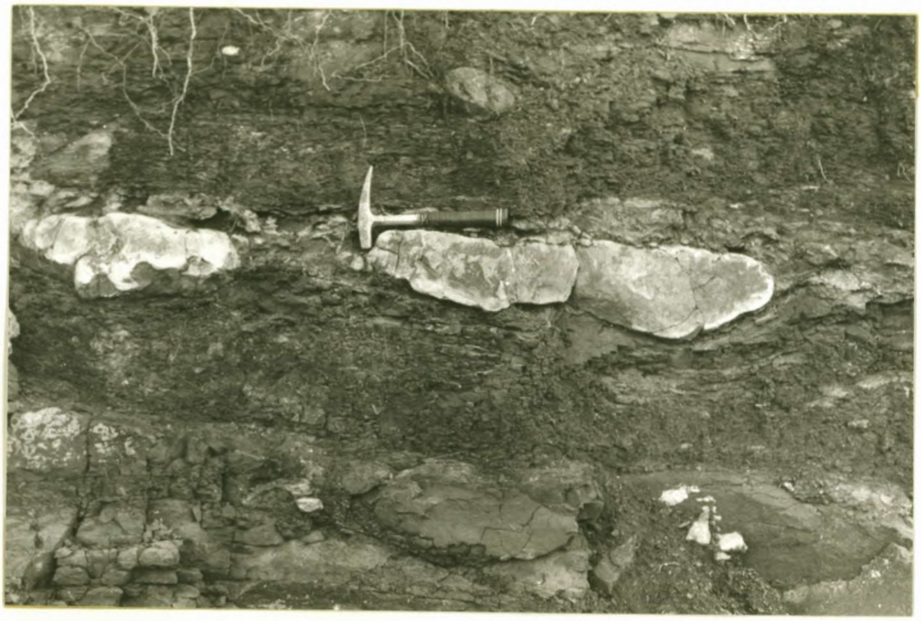


FIG. 4.4 Early diagenetic calcareous nodules in green overbank mudstones. Note traces of bedding planes curving around the nodules as a result of mudstone compaction.

FIG. 4.5 Primary volcanic ash-fall deposit (arrowed) interbedded with green, overbank mudstones in the upper part of megasequence A, south of Beaufort West.

FIG. 1.2). The fresh material intersected on Gannaleegte initially did not have a distinctive appearance but decayed rapidly into a friable, bentonitic substance resembling the surface exposures near Beaufort West. The Gannaleegte occurrence was not radio-active.

The trace-element content of the tuff near Beaufort West (TABLE 4.1) shows some resemblance to similar mudstones in the Collingham Formation, interpreted as tuffs by Martini (1974). Semi-quantitative X-ray diffraction analyses of the exposed tuff bed have indicated the presence of only quartz and illite (or sericite), both in equal abundances. Martini (1974) suggested that the illite associated with volcaniclastic material in Karoo rocks could represent original bentonitic deposits, in which the montmorillonite had been altered to sericite as a result of low-grade metamorphism.

The radio-activity of both the Collingham and Beaufort tuffs is attributed mainly to the presence of thorium, and to a lesser extent to uranium. These deposits show an elevated content of the "incompatible" elements rubidium, thorium and uranium. The parent magma had to have been sufficiently enriched in the volatile constituents to have caused explosive volcanism and the ejection of volcaniclastics including tuffs.

4.2 Sedimentary Structures and Textures

Sedimentary structures were used in the present investigation primarily for palaeoenvironmental interpretations and palaeocurrent analyses, and in special circumstances for stratigraphic mapping as well as for correlation purposes. They provide a method to study the relationship between the hydrodynamic conditions of streamflow and sediment transport, and resulting bedforms. Assemblages of sedimentary structures and in particular their vertical repetition in a succession, are diagnostic of depositonal environments (Collinson and Thompson, 1982). Although the presence of fluvial deposits in the Beaufort Group was established many years ago (Watson, 1913), modern sedimentological techniques have enabled reliable geomorphic and hydrodynamic reconstruction to be made of the alluvial systems (cf. Smith, 1987).

The classification of sedimentary structures is often based partly on morphological and partly on genetic aspects (Conybeare and Crook, 1968; Pettijohn, 1975; Selley, 1976). Simons and

Richardson (1961) classified current-generated bedforms according to the hydrodynamic conditions or "flow regimes" in which they formed in channels. Potter and Pettijohn (1977) assessed the directional significance of bedforms and sedimentary structures, and reviewed them along with the primary directional properties of sedimentary rocks in general. The scale and directional variability of the current-generated structures were used by Miall (1974) to rank them in as an indication of their reliability as palaeocurrent indicators.

Studies of bedforms and sedimentary structures in modern environments initially provided a basis for the interpretation and modelling of ancient alluvial deposits (e.g. Collinson, 1970; Williams, 1968). These processes were studied in considerably more detail in modern flume studies (Harms, 1975), the technical aspects of which are reviewed by Middleton and Southard (1984). On the basis of those results the sedimentary structures in the lower Beaufort strata were classified according to their scale, morphology, depositional environments, and inferred conditions of emplacement.

4.2.1 Sole Marks

Sole marks occur at the bases of sandstone beds, on their contacts with the underlying mudstones. Due to the fact that the basal contacts of sandstones are poorly exposed in the study area, sole marks were very rarely recorded and then mainly in the folded strata along the southern edge of the Basin. Scour-and-fills (gutter casts) are similar to sole marks but larger in scale. Certain soft-sediment deformational structures (e.g. loading structures) could also be classified as sole structures but are discussed elsewhere on the basis of their mode of origin.

The sole marks were scoured into soft, muddy beds under conditions of rapid streamflow, where the critical erosive stress of the cohesive sediment was exceeded (Leeder, 1982). Leeder noted mud composition, depositional history, and compaction as important controls in their formation. Observations in the lower Beaufort suggest that mud composition and pedogenesis of the muddy alluvium were important controls as well.

4.2.1.1 Gutter Casts

The term "gutter cast" was proposed by Whitaker (1973) for all scour-and-fill structures. These structures were previously referred to as scours, troughs, channels, etc., and the term "gutter cast" has now become widely accepted (e.g. Collinson and Thompson, 1982; Miall, 1984a). Gutter casts in the lower Beaufort occur either as sets of large-scale scours at the bases of major channel sandstones, or as isolated scours (pan scours) up to several metres wide in mudstone.

A fine example of gutter casts was recorded in the Nuweveld member along the Nuweveld escarpment (FIG. 4.6). They consist of steep-sided scours, sharply truncating the underlying green mudstones. In transverse view the scours are slightly asymmetric, irregularly spaced on a metre scale, usually paired, and locally up to 30 cm deep. Scours are aligned in the direction of streamflow, the latter deduced from trough cross-bedding. They occur too rarely in lower Beaufort sandstones to be measured routinely for palaeocurrent studies.

Gutter casts such as those depicted in FIG. 4.6 were produced during erosion of and sedimentation in the overlying channel. The exact hydrodynamic conditions which gave rise to their formation are uncertain, and Leeder (1982) noted that grooves have not yet been reproduced in the laboratory. It appears as if horizontal vortices in streams, and corrasion of the channel floor by suspended sand, were primarily responsible for their formation (Collinson and Thompson, 1982; Whitaker, 1973). Collinson and Thompson (1982) also suggested that pairs of vortices may form in the early stages and that one of them may become dominant at a later stage, resulting in oversteepening of one of the outer walls, and their typical asymmetrical cross-sectional shape. Whitaker (1973) found that where suitably exposed they proved to be of limited downstream extent, and to have smaller structures superimposed on their walls.

Common but less spectacular examples of gutter casts, also known as *pan scours* (Goldring and Aigner, 1982), occur as linear scours several metres wide and less than a metre deep, in green mudstones. They were cut in overbank alluvium during aborted channel avulsions or crevasse splaying. These scours were frequently seen in the floodplain deposits of the Combrinkskraal member in the southern Karoo, but never in red mudstones. A gutter cast showing several generations of scours was recorded in Ouberg Pass, and correspond to what Goldring and Aigner

FIG. 4.6 Gutter casts overlain by trough cross-bedding at the base of a channel sandstone in the Nuweveld member in the Nuweveld escarpment, north of Beaufort West. (Scale = 1,7 m.)

FIG. 4.7 Low-relief, triangular and deltoid flute casts at the base of a crevasse splay sandstone near the base of the Combrinkskraal member, on the farm Combrinkskraal.

(1982) termed *scour-in-scour* structures.

4.2.1.2 Flute Casts

These structures were appropriately described as ". . . heel-shaped hollows, scoured into mud bottoms . . ." by Selley (1976). As in the case with other sole marks they were not as abundant as expected, and were recorded only in the tilted strata in the southern part of the study area.

Two distinctive shapes of flute casts were recognised. The first occurs as bulbuous casts (Potter and Pettijohn, 1977) at the bases of fluvial channel sandstones. An unusual example of a series of bulbous flutes occurs on the scoured edges of a thin (<0,5 m) crevasse splay sandstone, on the farm Tuinkraal, Prince Albert District. These resemble flutes described on the outer walls of a braided channel sandstone by Gibling and Rust (1984). Triangular and deltoid flute casts were also recorded in a thin crevasse channel sandstone on the farm Combrinkskraal (FIG. 4.7). The flute casts all measured less than 8 to 10 cm long and 2 to 3 cm deep, the triangular casts being shallower than the bulbous ones.

Flutes are erosional structures created by vortices in rapidly flowing currents (Miall, 1984a) and they are not restricted to a particular environment. Dzulynski and Sanders (1959) suggested that their formation was controlled mainly by sediment discharge. They proposed that the absence of a thick saltation load (traction carpet) would encourage the formation of flutes, which explains why they are more common in crevasse splay sheet sandstones than in channel deposits. Experimental work suggests that flutes may evolve from longitudinal and meandering grooves at high streamflow velocities (Allen, 1969). Although not essential they are commonly initiated by bed defects (Allen, 1971). In the lower Beaufort it seems as if the triangular flutes are restricted to the crevasse splay deposits whereas the bulbous shapes occur both in channel and crevasse splay deposits. The triangular flutes thus appear to have formed in rapid, shallow flows whereas the bulbous flutes were formed in greater water depths.

Flute casts are useful directional structures in palaeocurrent studies (Potter and Pettijohn, 1977) but unfortunately they are either absent or too poorly exposed to be measured routinely in Beaufort channel deposits. A single flute cast was nevertheless the only directional structure that could be measured in the basal Beaufort channel sandstones on the Laingsburg Commonage, where other directional structures were obscured as a result of folding.

4.2.1.3 Groove Casts

Groove or substratal striation casts are rarely seen at the bases of crevasse splay sandstones in the lower Beaufort. They consist of sets of parallel grooves and ridges on a centimetre scale and were produced by scouring of cohesive substrata by rapidly flowing currents (Potter and Pettijohn, 1977).

Much is still to be explained about their origin. Hsu (1959) suggested on the basis of an investigation of turbidites that laminar rather than turbulent flow was required to form grooves. The groove casts seen in the Beaufort crevasse splay deposits (FIG. 4.8) do not show the striated (ornamented) edges common in marine turbidites but a similar mechanism of formation is accepted (cf. Potter and Pettijohn, 1977). The formation of grooves by engraving with a variety of tools has been discounted lately on grounds of the lack of physical evidence of such tools in the rocks (Potter and Pettijohn, 1977).

4.2.1.4 Shrinkage Crack Casts

Large-scale desiccation crack casts were recorded only once, in the Karoo National Park. These consist of 2 to 3 cm wide sandstone walls enclosing polygonal mudstone blocks. Small-scale desiccation cracks were recorded in a thin, dark-coloured mudstone layer in the Ecca-Beaufort transitional beds in the Ouberg Pass, west of Sutherland (FIG. 4.9). These structures are all interpreted as subaerial desiccation cracks, formed as a result of shrinkage of exposed cohesive sediment, followed by sand filling from above (Pettijohn, 1975). The environmental setting (subaqueous) and small scale of the Ouberg Pass occurrence (FIG. 4.9) has led to speculation as to whether they could be synaeresis cracks, an origin which seems unlikely in the light of published descriptions (Leeder, 1982, FIG. 11.6g; Selley, 1976, p. 230).

FIG. 4.8 Linear groove casts marking the base of a crevasse splay sheet sandstone. Delicate invertebrate trail marks are preserved, and faint obstacle scours above lens cap show current flow from lower left to upper right in picture.

FIG. 4.9 Small-scale desiccation cracks in the Ecca-Beaufort transitional sequence, Ouberg Pass, Sutherland District.

A variety of structures which could be mistaken for shrinkage cracks, are present in the lower Beaufort. Sandstone injections in the form of dykes can be distinguished on account of their scale and the observation that sandstone injections are rooted in sandstones (Oomkens, 1966). Nodular calcrete veins of pedogenic origin in red mudstones may resemble desiccation cracks. It is not always clear whether some of these structures are calcareous desiccation crack fillings, or whether they are entirely pedogenic. The paucity of desiccation cracks in mudstones is not easily explained but in regional comparisons the impression is gained that the Beaufort mudstones lost much of their original character (and sedimentary structures) as a result of post-depositional homogenisation, due to bioturbation, dewatering, and pedogenic processes.

4.2.2 Internal Bedding Structures

4.2.2.1 Massive Bedding

Massive or "homogeneous" bedding is defined as a primary bedding structure, lacking all evidence of post-depositional homogenisation. Many lower Beaufort sandstones show massive bedding locally in their basal parts, but its origin is doubtful. When Hamblin (1965) investigated the problem by means of X-ray techniques, well defined laminations were revealed in apparently massive sedimentary rocks.

Dewatering and compaction may have played a major part in the homogenisation of Beaufort sandstones (cf. Collinson and Thompson, 1982; Reineck and Singh, 1973). Intense bioturbation, on the other hand, is readily identified, and has caused sandstone homogenisation only locally. Massive bedding can also be formed by rapid deposition and burial of sand in streams before bedforms and bedding were shaped by tractive currents (Blatt *et al.*, 1972; Collinson and Thompson, 1982).

4.2.2.2 Plane Bedding

Plane bedding is more abundant than any of the other structures in lower Beaufort sandstones, both in terms of cumulative thickness and of frequency of occurrence. Individual strata are less than a centimetre thick and are therefore "laminae" rather than "beds", according to the classification of McKee and Weir (1953) and Ingram (1954). The term is synonymous with "flat bedding" as used by Selley (1976) and Allen (1965a), "even" lamination of Reineck and Singh (1973), and "horizontal lamination" or "laminated bedding" as used by Pettijohn *et al.* (1972). Use of the term "plane bedding" rather than "plane lamination" (Blatt *et al.*, 1972) is preferred as it identifies with the bedform in which the structures were formed, and distinguishes them from the fine laminations typical of suspension deposits.

Plane bedding in the Beaufort sandstones is usually accompanied by current lineations on bedding planes. Current lineations are diagnostic of plane bedding formed in the upper flow regime in alluvial channels (Allen, 1964b). Upper stage plane bedding forms at streamflow velocities too high for rough bedforms such as ripples, dunes and waves to remain stable (Harms, 1975). Flow depths are shallow, but not so shallow as to allow in-phase waves to form (Harms, 1975). Jackson (1976) found that in natural streams upper flow regime conditions were restricted to flow depths of less than 10 cm.

4.2.2.3 Large-scale Cross-bedding

Several types of cross-bedding are found in coarse-grained alluvial deposits. The diversity in scale and style of cross-bedding, of which some are diagnostic of certain depositional environments, has over the years led to several classification schemes.

The first systematic classification scheme was proposed by McKee and Weir (1953). They introduced certain qualitative terms such as "set" and "coset" for single and multiple sedimentation units. The character of the lower bounding surface and the attitude, shape and scale of cross-strata were the most important criteria. The main shortcomings were that (1) the classes were all large in scale, and that (2) the inherent nature of certain types of cross-strata to occur either as single or as grouped sets (cosets) were not taken into account. The latter aspect was included by Allen (1963b) in his classification which was based on six criteria, mainly the degree of lithological homogeneity of cross-strata, the scale, and the grouping of sets of cross-strata. Allen (1963b) distinguished altogether 15 different types of cross-stratification, denoted by letters of the Greek alphabet, in a scheme which has since become widely used.

Jacob (1973) pointed out that the nomenclature used by Allen (1963b) was unpopular, and that no distinction was made between cross-strata inclined at high and low angles. An alternative classification proposed by Jacob (1973) was found too cumbersome, and that of Allen (1963b) is still preferred. The lack of discrimination between shallow and steeply dipping foresets in Allen's (1963b) scheme nevertheless remains a serious constraint, and for that reason it was not used in the present investigation. A descriptive terminology of the types of cross-bedding found in the lower Beaufort is presented below, with reference where possible to Allen's scheme.

4.2.2.3.1 Low-angle Planar Cross-bedding

Low-angle cross-bedding showing slightly concave foreset beds, is rarely present in intraformational conglomerates (FIG. 4.10). They occur as solitary sets grading laterally into massive conglomerates. The examples studied were too poorly exposed for accurate palaeocurrent measurements. Cross-bedding of this nature forms in both longitudinal and diagonal gravel bars (Walker, 1975), usually in braided streams. Walker (1975) suggested that these bars form under conditions of high sediment and high fluid discharge. The mechanism of their formation in Beaufort meandering channels is not clear but taking into account the postulated (high-sinuosity) channel geometry, it is likely that this type of cross-bedding was formed in diagonal bars as a result of cross-flow.

4.2.2.3.2 Large-scale Trough Cross-bedding

Large-scale trough cross-bedding is present in channel sandstones in the upper Ecca and lower Beaufort beds throughout the study area. Trough cross-bedding usually occurs near the bases of

FIG. 4.10 Low-angle cross-bedding in conglomerates at the base of a channel sandstone, in the Koup member on the farm Bulkraal, adjoining the townlands of Beaufort West.

FIG. 4.11 Three superimposed sets of trough cross-bedding in a channel sandstone in the Nuweveld member on Tafelberg, Fraserburg. Total thickness is more than 1 m; pocket transit (70 mm high) for scale.

channel sandstones, and rarely as solitary sets in the top parts. Cosets comprising up to three or four superimposed, large-scale sets (FIG. 4.11) are rare, and occur mainly in the Koup member. Trough cross-bedding is present wherever moderately thick channel sandstones are present. Where sandstones are suitably exposed such as along the escarpment (on the farms Layton and Tafelkop), azimuths measured within individual channel sandstone storeys show unusual directional variability. Another feature of troughs frequently noted wherever they were exposed three-dimensionally, is their asymmetric shape of troughs as seen in plan view. The cross-bedding corresponds to Theta cross-stratification of Allen (1963b). Individual sets have mean thicknesses of 25 to 35 cm with a minimum of just more than 10 cm. Maximum set thicknesses rarely exceed 1,2 m, with one notable exception: a set 2,2 m thick in a sandstone of the Leeu Gamka member on the farm Spreeuw Fontein (Prince Albert District). Maximum set thicknesses are found in moderately thick (<7 m), meandering channel deposits. The thicker sandstones of the Koup and Combrinckskraal members often show no trough cross-bedding at all, or (rarely) solitary sets of average thicknesses.

Trough cross-bedding in fluvial channel sandstones was formed as a result of migrating, large-scale bedforms (Allen, 1967a). Troughs are up to 27 m wide and 1,2 m thick in Koup sandstones on the farm Tafelkop (location 12, FIG. 1.2). Trough cross-bedding described in recent alluvial sediments were related to large-scale, asymmetric, sinuous and lunate bedforms (Williams, 1968). Allen (1963c) suggested that they were rather formed by large-scale linguoid ripples (dunes). These bedforms, also referred to as "large ripples", were modelled in flume studies by Southard (1975), Middleton and Southard (1984), and others. Allen (1966) referred to all large ripples as "dunes", a term which has now become accepted by fluvial sedimentologists.

Dunes and other large-scale bedforms are never preserved in their original form anywhere in the lower Beaufort sandstones, and their dimensions and hydrodynamic parameters therefore had to be reconstructed from cross-bedding measurements. Allen (1966) showed that dune heights are roughly proportional to depth of streamflow. Yalin (1964), on account of both flume and river data, suggested that dune height does not exceed $^1/_6$ th of the flow depth, whereas Allen (1963c) suggested a ratio between $^1/_{10}$ and $^1/_5$. Allen (1967a) noted that large-scale cross-bedded units usually have erosional tops, which makes them unreliable for accurate depth estimates. Allen

(1967a) also found that height/depth correlations became unreliable in the case of three-dimensional dunes, and ratios as high as 0,8 were observed in natural rivers.

The problems in estimating flow depth are illustrated by an example in the Combrinkskraal member on the farm Spreeuw Fontein near Prince Albert. A 2,2 m thick set measured in a channel sandstone deposit less than 6 m thick translates to a flow depth of between 11 and 22 m, which greatly exceeds the total channel-fill thickness, as well as the depth of individual flows as estimated from the thickness of individual storeys (<3 m). The upper contact of the trough showed truncation by a later fluvial cycle, suggesting an even greater thicker set originally. These ratios are therefore unrealistic for estimating flow depths in ancient deposits as well.

Scheidegger and Potter (1967) suggested that dune height is related to fluid turbulence, and thus to sediment calibre and discharge. That relationship was expressed as follows:

$$\log D = -1{,}99 + 0{,}393 \log H$$

where:

D = median sediment grain size (mm), and

H = cross-bedding set thickness (cm).

The relationship does not apply satisfactorily to grain size (0,12 mm) and cross-bedding thickness (25 to 35 cm) data of lower Beaufort sandstones.

Foreset beds are inclined at maximum angles of up to 34°. These correspond to dips recorded in flume studies (Harms, 1975) and in dunes in modern fluvial deposits (Williams, 1968). In rare cases foreset beds show soft-sediment contortion of a nature which has been ascribed to gravity-induced sliding (Blatt *et al.*, 1972; Friend, 1965). The drag effect of currents flowing over dunes has been suggested as being partly responsible for the soft-sediment deformation (Conybeare and Crook, 1968), and pressure relief due to flow separation on the leesides of these bedforms makes this quite likely. The spatial association of these structures with convolute lamination in some lower Beaufort sandstones nevertheless suggests that they were formed mainly by sediment dewatering due to compaction (cf. Cant, 1982). Some may have formed as a result of soft-sediment, gravitational sliding.

4.2.2.3.3 Large-scale Planar Cross-bedding

Cross-bedding of this nature occurs as solitary sets less than 0,5 m thick. Foreset beds are homogeneous and dip fairly steeply (≥20°). Sets are bound by non-erosional surfaces, and they conform to Alpha cross- stratification of Allen (1963b). They are very rare but nevertheless present in both channel and crevasse splay sandstone deposits.

Planar cross-bedding is common in braided streams where it forms in transverse and longitudinal bars (Allen, 1966; Harms, 1975; Melvin, 1985). Transverse bars have been described in ancient meandering channel deposits by Miall (1984a) and Jackson (1976). In braided streams these bedforms reflect flow strengths lower than those of other large-scale bedforms, and they usually form as a result of the modification of other bedforms in the waning stages of high-level flooding events (Cant, 1982; Williams and Rust, 1969). A similar origin is accepted for the lower Beaufort examples, found in places in the thick channel sandstones of the Koup member in the type area. An example in a channel sandstone deposit at Riemhoogte is accentuated by red pigmentation (ferruginisation) of the foresets. The extreme paucity of these structures suggests that perennial rather than intermittent discharge prevailed during deposition of most of the lower Beaufort channel sandstones.

Planar cross-bedding of the same scale is seen in thin, crevasse splay sandstones deposited as sand sheets and lobes on floodplains adjacent to major streams. A similar origin was proposed by Saunderson and Jopling (1980) for tabular (planar) cross-bedding in recent sand deposits of the Brampton Esker in Ontario, Canada.

4.2.2.3.4 Epsilon Cross-bedding

This type of cross-bedding consists of sets of gently dipping, heterogeneous foreset beds, and was termed Epsilon cross-stratification by Allen (1963b). They are usually present as solitary sets in the top parts of channel sandstones. Sets are on average 2 to 3 m thick but range between less than 1 m to a maximum of over 6 m (FIG. 4.12). Cosets comprising several stacked sets, and reaching thicknesses of up to 8 m, represent several phases of channel migration and point bar accretion.

Foreset beds consisting of alternating sandstone and mudstone are several tens of centimetres thick, and are sigmoidal in cross section.

The formation of Epsilon cross-bedding by lateral accretion of point bars was proposed by Wright (1959) who referred to it as "meander bank" or "point bar" cross-bedding. The basis on which Wright (1959) made his interpretations does not appear to have been entirely correct, as he appears to have included other types of cross-bedding which are common in meandering channel deposits in his observations. The formation of Epsilon cross-bedding in point bars in meandering streams has nevertheless been shown clearly by Allen (1964a and 1965a), Leeder (1973), and Puigdefabregas and Van Vliet (1978).

Epsilon cross-bedding is neither abundant nor easily identifiable in lower Beaufort sandstones, a problem which was noted also by Allen (1966) and Graham (1975) in other ancient meandering channel deposits. An example of scroll bar relief in the lower Beaufort was described by Smith (1987) but spectacular examples such as those recorded by Puigdefabregas and Van Vliet (1978) in meandering channel deposits of the Tertiary of the southern Pyrenees are either absent or concealed in the lower Beaufort sandstones. Allen (1966) suggested that lateral deposition may be difficult to detect in cases where point bars were gently sloping, i.e. in relatively wide and shallow streams. It is also possible that the different accretionary units are not accentuated due to the fine and even grain sizes of the lower Beaufort channel sandstones, the result possibly of steady rather than intermittent discharge.

Palaeocurrent analyses show that much of the area in which field observations were made, namely the Nuweveld escarpment, is exposed in what represents a longitudinal section of the depository. Transverse sedimentary features such as Epsilon cross-bedding are therefore poorly exposed in such areas. However, where the succession is exposed more extensively and in transverse section, such as in the area northeast of Sutherland, Epsilon cross-bedding was frequently noted in the Koup member. Examples can be seen in these sandstones in road cuts of the main road linking Sutherland and Fraserburg, just outside Sutherland.

FIG. 4.12 Epsilon cross-bedding in the top part of a channel sandstone in the Koup member, on the Nuweveld escarpment. (Scale = 1,7 m)

FIG. 4.13 Climbing ripple cross-lamination in levee sandstones capping a channel deposit in the Koup member, Riemhoogte.

4.2.2.4 Small-scale Cross-lamination

Small-scale cross-lamination in a variety of styles and modes of occurrence is present in both the channel and floodplain deposits. The sandstones are fine to very fine-grained or silty, and show both fining and coarsening-upwards tendencies. Sets range in thickness between 1 cm and 8 cm in overbank siltstones and sandstones, but are fairly constant at between 2 and 4 cm thick in channel sandstones.

Ripple cross-lamination occurs in the top parts of channel sandstone storeys. Rib-and-furrow structures are ubiquitous on eroded surfaces. The overbank examples are convoluted in places, especially those in upwards-coarsening sequences. Cross-laminated sandstone intervals reach thicknesses of nearly 2 m in certain channel sandstones, particularly in the uppermost members of the lower Beaufort.

Small-scale cross-lamination was suggested to form as a result of "ripple drift" by Sorby (1859, quoted in Walker,1963), and the term *ripple-drift cross-lamination* was introduced by Walker (1963). Walker (1963) distinguished three morphological types (A, B and C) of cross-lamination based mainly on the degree of ripple stoss-side erosion, and the pattern of sediment aggradation as judged from the climbing tendencies of foresets. Jopling and Walker (1968) proposed a more practical classification of ripple-drift cross-lamination, and related them to their environments of deposition. The most complete analysis and classification of these structures to date, was that by Allen (1963b).

The different types of ripple cross-lamination in the lower Beaufort sandstones and siltstones are restricted to particular subenvironments, as reviewed in greater detail elsewhere. Nu cross-lamination (Allen, 1963b) is ubiquitous in the upper parts of channel sandstones, forming rib-and-furrow structures where suitably exposed. This type was referred to as *micro-cross-lamination* by Hamblin (1961). Kappa cross-lamination (Allen, 1963b) occurs in channel, levee and crevasse splay sandstone deposits. The example depicted in FIG. 4.13 was recorded in levee deposits capping a channel sandstone near the top of the Koup member at Riemhoogte (location 25, FIG. 1.2). Examples of a type resembling Mu cross-lamination (Allen, 1963b) were deposited in crevasse channels.

With the exception of Mu cross-stratification these structures are ubiquitous in fluvial deposits, where they are formed by trains of linguoid ripples (Allen,1965a and 1970; Blatt *et al.*, 1972; Friend, 1965). Examples of linguoid ripple marks were rarely preserved on the upper contacts of channel sandstones, notably in the Nuweveld member along the escarpment.

Climbing-ripple cross-lamination forms in sluggish streams as a result of the incomplete erosion of ripple stoss sides, coupled to a high rate of sediment aggradation (Blatt *et al.*, 1972; Jopling and Walker, 1968). Climbing ripple laminae have been recorded in recent channel deposits and natural levees (Wopfner, 1970). Mu cross-lamination was clearly formed by straight or slightly sinuous ripples (Allen, 1963b).

The formation of small ripples in streams is restricted to sand with grain sizes finer than 0,6 mm (Harms, 1975). Ripple dimension depends on grain size but is independent of flow depths (Allen, 1966). Ripples are formed in turbulent flows but differ from other bedforms in that they are restricted to hydraulically smooth flows (Allen, 1966). They form in the lower flow regime in streams where they coexist with other bedforms, occasionally superimposed on dunes (Friend, 1965; Harms, 1975). Rare examples of the latter are preserved in the lower Beaufort sandstones where ripple cross-lamination directly overlies and truncates large-scale sets of trough cross-bedding.

4.2.3 Bedding-plane Markings

These structures can be defined as current-generated structures occurring on surfaces between beds, and were classified as "surface marks" by Pettijohn (1975). Most notable are the ripple marks. Evans (1949) described the differences between wave and current ripples and attempted to relate them to depositional environments. Tanner (1967) introduced several dimensionless parameters to describe ripple geometries, by which they could be related to their depositional environments. Of these the ripple index (RI), the ripple symmetry index (RSI), and the bifurcation index (BI) are the most frequently used.

4.2.3.1 Current Lineations

Current lineations form on bedding planes in the upper flow regime. These lineations occur as two distinctive types. *Parting lineations* (Allen, 1964b) occur as series of irregularly spaced, parallel ridges and hollows on bedding planes. *Streaming lineations* (Conybeare and Crook, 1968) are shallow, small-scale grooves a few millimetres wide. Both types appear to form in the upper flow regime but whilst parting lineations are ubiquitous in channel sandstone deposits, streaming lineations occur very rarely. The latter also show a much greater variability in palaeocurrent directions. The genetic relationship between the two varieties is not entirely clear but Conybeare and Crook (1968) contended that the type Allen (1964b) produced experimentally was a streaming lineation, and not a parting lineation.

4.2.3.2 Ripple Marks

Current ripple marks are unusually rare considering the abundance of ripple cross-lamination in sandstones throughout the lower Beaufort succession. They are characterised by asymmetric profiles with sinuous, straight, lunate or linguoid ripple crests (cf. Selley, 1976). Straight-crested ripple marks were never recorded but sinuous (catenary or undulatory) ripples occur abundantly in overbank deposits, particularly in crevasse splay sheet sandstones. These have rounded crests and wavelengths of less than 5 cm, and although never verified in field observations they are considered responsible for the formation of Mu cross-lamination.

An increase in flow velocity in streams resulted in the destruction of ripples, and out-of-phase linguoid or lunate ripples formed in their place (Blatt *et al.*, 1972; Tucker, 1984). Allen (1969) attempted to model the shape of ripples under different flow conditions. Flume studies on small ripples by Banks and Collinson (1975) showed a complex relationship, suggesting that the shapes and dimensions were largely independent of the conditions of flow (e.g. flow depth), but that they depended rather on the boundary shear stress (i.e. the conditions prevailing at the sediment-fluid interface).

Linguoid ripples were occasionally observed in the lower Beaufort, invariably in channel sandstones of the Nuweveld member. In the exposures studied, these ripples in places show highly irregular shapes but the linguoid form and direction of migration are nevertheless clear. In exceptional cases these ripples are quite large and specimens more than 20 cm wide were measured in Nuweveld channel sandstones, in Oukloof.

Rib-and-furrows (FIG. 4.14) are the most abundant of the bedding plane markings in channel sandstones and certain crevasse splay deposits, representing a plan view of ripple cross-lamination. They proved to be very useful in palaeocurrent studies, and their presence can be used to distinguish distributary channel sandstones from other deltaic sandstone bodies in the Ecca-Beaufort transitional sequence. Individual troughs are 5 to 10 cm wide, with a maximum of 16 cm measured in the Nuweveld member on the farm Toorwater.

Incipient ripple marks are very rare and only a few examples were identified in lower Beaufort sandstones. As is often the case with bedding plane markings they are best exposed and preserved in the Nuweveld sandstones along the Nuweveld escarpment. They may be present, but not properly exposed, in other stratigraphic units as well. In each case they resemble hoof-shaped imprints embedded on bedding planes, truncating current lineations. The mode of preservation of these delicate imprints is not easy to explain. It is possible that stream discharge ceased soon after the transition from upper to lower flow-regime, but before fully developed current ripples were formed. This could have been the result of channel avulsion, during which the incipient ripples were buried by abandoned channel-fill muds.

Oscillation ripple marks occur mainly in the sandy and silty overbank deposits, and less commonly in active channel deposits. They are more abundant than current ripple marks and are distinguished on account of their symmetrical shapes, relatively straight and sharp crests, and ripple bifurcation (FIG. 4.15). Ripple wavelengths are typically 4 to 8 cm but exceptionally as small as 2 to 3 cm. In the case of ripples with shorter wavelengths the crests tend to be somewhat rounded. The larger ripples have ripple indices between 10 and 15.

Although conspicuous in the argillaceous Leeu Gamka member, ripple marks are not as abundant in the floodplain deposits elsewhere in the succession as expected. The absence of ponded bodies of water in which wave ripples could have formed, is the most likely explanation. Bioturbation and

FIG. 4.14 Large-scale rib-and-furrows capping a channel sandstone. Trough axes indicate the direction of stream flow, from left to right in picture.

FIG. 4.15 Interference ripple marks on a crevasse splay sheet sandstone, showing ripple bifurcation and ladderbacks. Note edge of a pan scour (P) in upper left corner of picture, and presence of ripple marks in the scour.

pedogenic alteration may also have obliterated ripple marks in alluvium, prior to or after burial.

Spectacular examples of *interference ripple marks* are found in parts of the lower Beaufort, and examples from the Beaufort West area are depicted in Miall (1984a). Although abundant in the Leeu Gamka member (FIG. 4.15) they occur in floodplain deposits throughout the lower Beaufort succession. These marks are also referred to as "ladderbacks" (Miall, 1984a) or "tadpole nests" (Selley, 1976). They may form as a result of the interference of either divergent wave and current ripples (Collinson and Thompson, 1982), or two directions of wave motion (Conybeare and Crook, 1968). They are of no particular diagnostic value, other than showing the presence of shallow bodies of ponded water.

4.2.4 Penecontemporaneous Deformation Structures

Soft-sediment deformation structures are not abundant, but a considerable variety of these structures is found in the lower Beaufort. Individual structures are not restricted to particular depositional environments (Pettijohn, 1975; Reineck and Singh, 1973), and they are of limited diagnostic value in the lower Beaufort sequence. Soft sediment deformation structures also lack directional significance and they cannot be used to determine the palaeoslope (Potter and Pettijohn, 1977). These structures are nevertheless abundant in the upper Ecca to the extent that soft-sediment deformation could be used confidently in stratigraphic mapping of the Ecca-Beaufort contact over large parts of the Basin (Jordaan, 1981; Visser and Loock, 1974). The palaeoenvironmental aspects of those occurrences are reviewed elsewhere, and the present discussion restricted to the soft-sediment deformation structures in the lower Beaufort.

4.2.4.1 Loading Structures

Although loading structures such as load casts, ball-and-pillows and sand balls are ubiquitous in the upper Ecca they also occur in the argillaceous intervals of lower Beaufort. Simple loading appears to have occurred at the bases of certain crevasse splay sheet sandstones, in some cases after limited compaction of the underlying mudstones (FIG. 4.16). The sand subsided into the mudstones

FIG. 4.16 Load casting at the base of a typical crevasse splay sheet sandstone. En echelon faulting in the kink zones suggests that mudstones were partly consolidated when loading commenced.

FIG. 4.17 Part of a large sandstone dyke, cutting across red and green mudstones at the base of the Combrinkskraal member on the farm Wamakerskraal, Prince Albert District.

below without notable liquefaction of the footwall muds. Proper ball-and-pillow structures were logged only once in overbank sandstones and in that instance the balls were formed entirely within a thin (15 cm thick) sandstone sheet, showing no deformation of the sandstone contacts. Deformation in all these cases was primarily as a result of gravitational loading of the sand, whilst liquefaction of the underlying mudstones played a lesser role.

An unusual example of ball-and-pillow structures accompanied by convolute laminations and slump structures occurs in a succession of thin (each <0,75 m thick) upwards-coarsening sequences in the Leeu Gamka member on Elandsberg, Prince Albert District (location 3, FIG. 1.2). The structures occur in argillaceous sandstones and siltstones overlying green mudstones, and probably resulted from dewatering and liquefaction of the mudstones.

Small-scale loading structures in the form of sand balls, a few millimetres in diameter, are present in some of the intensely mottled Koup sandstones in the Kamdebo. Their formation appears to be related to the desiccation of clay, formed by weathered tuffaceous material in the sand. The clayey matrix may have acted as a lubricant, and may have contributed to the liquefaction of wet sands.

4.2.4.2 Convolute Lamination

Convolute lamination accompanied by flame structures occurs in the thick channel sandstones of the Nuweveld member on Mordant Klaasenskraal, and the Koup member on Karoluspoort (FIG. 1.2). Convolute lamination is present also in ripple cross-laminated crevasse-splay sandstones, including the upwards-coarsening sequences on Elandsberg.

Selley (1976) drew a distinction between convolute lamination and convolute bedding. *Convolute lamination* is usually found in ripple cross-laminated sandstones, and most of the deformation took place during deposition of the sediment. This type of deformation was probably initiated by shear stresses exerted on the sediment-water interface (Potter and Pettijohn, 1977), which resulted in sediment dewatering and liquefaction, and deformation of the sand ripples (Reineck and Singh, 1973; Selley, 1976). *Convolute bedding* formed as a result of the deformation of entire beds on a larger scale (Selley, 1976). Deformation took place entirely within the bed, and the lower and upper contacts were left intact (Pettijohn, 1975). In beds that have undergone extensive soft-sediment deformation the latter aspect can be used to distinguish convolute bedding from other types of soft-sediment deformation. The occurrences on Mordant Klaasenskraal and Karoluspoort, cited previously, are typical examples. They appear to have been shock induced, most likely by seismic tremors.

4.2.4.3 Slump Structures

Slump structures, like most soft-sediment deformation structures, were initiated by gravitational loading (Pettijohn, 1975), but differ in respect of the lateral component of movement during their formation (Selley, 1976). They reflect deposition on an unstable slope, which in this case is the direction in which sediment accretion took place locally. Slump structures occur very rarely in the lower Beaufort, as for instance in the upwards-coarsening sequences in the Gamka member on Elandsberg. Those slump structures are shaped in the form of asymmetric slump rolls, all inclined in approximately the same direction.

The deformed trough cross-bedding foresets recorded in a prominent sandstone on Toorwater, are an unusual type of slump structure. Slumping was probably initiated by sediment dewatering, followed by gravitational sliding of the foreset beds.

4.2.4.4 Injection Structures

Injection structures occur on a variety of scales, mostly in the argillaceous successions. A large-scale sandstone dyke (FIG. 4.17) several metres long and up to 25 cm wide, cuts across the basal red mudstones of the lower Beaufort on the farm Wamakerskraal (Prince Albert District). The dyke shows very little evidence of ptygmatic folding. The host mudstones clearly underwent very little compaction subsequent to emplacement of the dyke, but they were still unconsolidated so as to allow liquefaction and intrusion of the sand (Selley, 1976). Liquefaction of the source bed could have taken place as a result of seismic shocks (Collinson and Thompson, 1982).

Small-scale injection structures occur in mudstones as well. The example depicted in FIG. 4.18 is one of a swarm of identical structures recorded in the Combrinkskraal member in the Verlatenkloof. Dragging of the mudstone walls and rafted mudstone flakes in the injected sandstone suggest that the mudstone was fairly cohesive but still unconsolidated at the time of injection. These structures are exposed in cross-section only, and they appear to be sandstone volcanoes rather than dykes. A plan view of similar structures, located by R.M.H. Smith (pers. comm., 1981) in lower Beaufort sandstones, is illustrated in Turner (1979).

4.2.4.5 Soft-sediment Folds

The thick, massive overbank mudstone successions are in most cases homogenised to the extent that no primary sedimentary structures are visible. Some evidence of compaction in the form of folds or contortions is seen in the rocks, which were involved the homogenisation. The sandstone layer interbedded with massive siltstones, shown in FIG. 4.19, is contorted with a fold amplitude of over a metre in places. Other signs of soft-sediment deformation, such as loading and detachment, as descibed by Leeder (1982), are absent in the sandstone layer despite the considerable degree of compaction and folding.

4.2.5 Biogenic Structures

These structures reflect the nature of the organic activity in the sediments at the time of deposition, and can be related indirectly to the depositional environments (Selley, 1976). Trace fossils are the most abundant and have the advantage over body fossils that they occur *in situ* in their host strata, the latter usually sandstones (Seilacher, 1967). They are present in environments where body fossils were either not deposited or not preserved (Frey, 1975).

The most common trace fossils are the *feeding trails* or *Fodichnia* (Seilacher, 1967), which were preserved as epichnial traces on bedding planes. Entire sandstone or siltstone beds were foraged occasionally, and in some cases completely bioturbated. The sedimentary features were completely destroyed in the process, and small protuberances on bedding surfaces accompanied by red mottling

FIG. 4.18 Small-scale sandstone injections in laminated red mudstones. Note the effects of drag on the mudstone walls, and rafted red mudstone fragments (arrowed).

FIG. 4.19 Soft-sediment contortion of a thin sandstone layer, due to compaction of massive overbank mudstones.

are the only evidence of organic activity. Fecal excretions are characteristic of sediment feeders but these were very rarely preserved in lower Beaufort strata.

Traces left by sediment-ingesting organisms are classified as the ichnogenera *Planolites* (Frey, 1975). The trace fossils of the lower Beaufort belong to the *Scoyenia* ichnofacies of Seilacher (1967), which is characteristic of a terrestrial environment (Frey, 1975). During the present investigation some of these fossils were recorded over a wide range of environments. Identical arthropod repichnial trackways resembling *Umfolozia* were, for instance, found both in abandoned channel-fill siltstones in the Leeukop member on the farm Toorwater, and in a distributary mouth bar sandstone in the Waterford Formation on Koedoesberg (locations 30 and B, resp., FIG. 1.2).

Plant fossil impressions are most abundant in sandstones, particularly in association with uranium mineralization, and are very rare in mudstones. They are restricted to leaf and stem impressions of *Glossopteris*, *Schizoneura* and *Phyllotheca*. Fine examples of *Glossopteris* leaf impressions are preserved in siltstones, approximately 600 m above the basal contact of the lower Beaufort, on the farm Tuinplaas (location 4, FIG. 1.2). Well-preserved equisetalians in various stages of disarticulation, are commonly present in sandstones. Almost complete specimens up to 2 m in length are associated with high-grade uranium mineralization in sandstones on the farm Swartkop near Sutherland (FIG. 4.20).

Vertical *tubes* are present near the top contacts of certain channel and crevasse splay sandstones. The tubes are less than 3 cm in diameter and a few tens of centimetres long (FIG. 4.21). They are cylindrical and straight, somewhat flattened, and they are usually filled with green mudstone. Rare examples of these structures branching upwards and filled with red mudstone have also been recorded in the top parts of channel sandstones. The sediment fill in the latter is massive and does not resemble the backfill left by sediment-ingesting organisms (cf. Simpson, 1975). These structures could not have been dwelling burrows as the postulated environment (alluvial plain) would not have supported suspension feeders (Seilacher, 1967). They do not show the characteristics of water escape structures described by Lowe (1975), and are interpreted as root and stem casts. They were formed as a result of the rapid burial of plants, followed by decay of the stems or roots and sediment infill from above.

FIG. 4.20 Equisetalian plant stem impression, 76 cm long, in an extensively altered and highly mineralized (U3O8) sandstone in the Koup member, on the farm Swartkop (Sutherland District).

FIG. 4.21 Vertical impressions left by roots or stems of unidentified plants in levee deposits capping a channel sandstone, in the Nuweveld member on the farm Toorwater (Murraysburg District).

Typical *rootlet beds* were recorded only once in red overbank mudstones, and never in green mudstones. Thin (<1 mm), irregularly shaped, greenish coloured tubes were preserved locally in red mudstones which were otherwise extensively altered by pedogenic processes. This example was recorded in the Koup member at Riemhoogte (location 25, FIG. 1.2). The green colouration is possibly the result of reduction of ferric pigment in the red mudstone in immediate contact with the roots.

Large-scale burrows up to 20 cm in diameter are present in massive siltstones in the Leeu Gamka member. Most of these burrows are irregular in shape and show possible scratch marks on their walls. They are now considered to have been dug by reptiles, possibly for breeding purposes (Smith, 1986). Vertebrate tracks are rare in the lower Beaufort, but examples of *Diictodon* from an undisclosed locality were reported by R.M.H. Smith (pers. comm., 1986).

5.0 ECCA-BEAUFORT TRANSITION

5.1 Introduction

The depositional environments and lithological composition of the Ecca and Beaufort Groups have been understood for many years. The Ecca Group consists mainly of dark, pelagic shales and interbedded turbidites at the base, and deltaic sandstones and shales at the top (Ryan, 1967; Wickens, 1984). The Beaufort Group consists of diagnostic red and green overbank mudstones, and interbedded fluvial sandstones. These two successions are separated by a transitional sequence of up to several hundred metres thick, which straddles the uppermost part of the Waterford Formation as defined at present (SACS, 1980). The transitional sequence has never previously been studied adequately in the study area. Chronostratigraphic markers are not present and a contact had to be defined which was regionally valid, before routine stratigraphic and sedimentological investigations could be undertaken.

Several authors suggested that the change in depositional environments should serve as a basis for the Ecca-Beaufort contact (Haughton *et al.*, 1953; Van Biljon, 1967). Indirect methods which could possibly be used to determine the position where this change took place include the first appearance of therapsid fossils in the succession (Rogers, 1925; Rubidge, 1983), and geochemistry (Kingsley, 1977). Lithological changes which mark that change in depositional environments include the proportion of sandstone, sedimentary structures of the arenaceous and argillaceous rocks, colour of the argillaceous rocks, and the presence of upwards-fining sequences and therapsid fossils (SACS, 1980). For the purposes of the present investigation a contact was chosen which separates the genetically unrelated Ecca and Beaufort successions but which at the same time obeys the preconditions set for the stratigraphic classification of sedimentary strata (Hedberg, 1976).

5.2 Transition from Marine to Terrestrial Deposits

The arenaceous Waterford Formation is prominently developed on the transition between typical Ecca and typical Beaufort strata. Apart from the fact that the Waterford Formation shows considerable changes in facies and thickness in the western Karoo Basin it was not clear initially whether it belonged with the Ecca or the Beaufort Groups. It varies in thickness from more than 800 m in the southwestern corner of the Basin, to less than 200 m in the area north of Klaarstroom.

5.2.1 Depositional Environments of the Transitional Sequence

The depositional environments of the Ecca Group were reviewed by Visser *et al.* (1980) and Wickens (1984). Visser *et al.* (1980) distinguished nine sedimentary facies as well as first, second and third-order upwards-coarsening cycles. The uppermost Ecca and lowermost Beaufort beds were investigated more closely by Jordaan (1981) in order to determine the palaeoenvironmental framework of the Ecca-Beaufort transition. The top of the deltaic interval was recommended as a feasible contact on account of the distinct lithological changes occurring at that position (Jordaan, 1981). This view was supported by Wickens (1984) who recommended after an extensive mapping program in the western Karoo Basin, that the Ecca-Beaufort contact be taken at the top of the Waterford Formation.

In the present investigation the Ecca-Beaufort transition was studied by means of detailed stratigraphic sections, facilitated by excellent outcrops along the western margin of the Basin. Detailed profiles were compiled from sections measured at six localities, namely the Laingsburg Commonage, Verlatenkloof Pass, Ouberg Pass, Koedoesberg, Keiskieberg, and the farm T'kokoboos west of Carnarvon (FIG. 1.2). The transitional beds were also studied cursorily at five more localities in sections through the upper Ecca and lower Beaufort Groups along the southern margin of the Basin (FIG. 1.2). A comprehensive description and interpretations are to be found in Jordaan (1981), and only a brief account will be given here.

Three of the facies (facies 1, 2 and 3) are predominantly argillaceous in composition. Facies 1 and 3 were deposited from suspension in an overbank setting on alluvial plains, and are distinguished primarily on the basis of the presence or absence of red mudstone. Facies 1 also occurs in the interdistributary bay and delta plain deposits. Facies 2 comprises thick intervals of alternating, thin sandstone and dark-coloured mudstone deposited in the prodelta, shelf and interdistributary bay environments.

Facies 4a (ripple cross-laminated sandstone), 4b (massive sandstone), 4c (plane bedded sandstone), 4d (low-angle cross-bedded sandstone) and 4e (through cross-bedded sandstone) were deposited as distributary mouth bar, distributary channel, crevasse splay and fluvial channel deposits in the delta front, delta plain and fluvial environments. The distributary mouth bar (facies 4b, 4c, 4d and rarely 4e), distributary channel (facies 4a, 4c and 4e), and fluvial channel deposits (facies 4a, 4b, 4c and 4e) consist of the same group of facies but arranged in different facies sequences. The distributary mouth bar deposits commonly show an upwards-coarsening tendency whereas fluvial channel sandstones show typical upwards-fining grain-size distributions.

Thus, while these facies proved to be of limited use individually in palaeoenvironmental interpretations, diagnostic facies associations became evident when studied in vertical succession. These associations show pertinent trends over limited intervals, but become indistinct on a broader scale due to the interfingering of depositional subenvironments.

5.2.1.1 Shelf/Prodelta Deposits

The shelf and prodelta deposits consist of thick intervals of alternating thin sandstone-mudstone couplets (centimetre scale), interbedded in places with massive or poorly laminated, dark-coloured shales. Fossil shells are absent, and horizontal trails and bioturbated beds are the only evidence of biological activity. The absence of fossil shells appears to be the result of slightly acidic conditions prevailing during and shortly after deposition, thought to have caused dissolution of shells (Scott and Fisher, 1969).

The prodelta deposits occur in the lowermost parts of the sections, where they are interbedded with distal bar deposits and, rarely, with thin distributary mouth bar deposits. The prodelta deposits are distinctly argillaceous and constitute most of the Fort Brown and Tierberg Formations, as described by SACS (1980).

FIG. 5.1 Ball-and-pillow structures in distributary mouth bar sandstones on the farm Rietkraal, Prince Albert District.

5.2.1.2 Delta Front Deposits

The delta front deposits occur as a distinctly arenaceous unit overlying the argillaceous prodelta deposits. The base of the unit is transitional over a few tens of metres and coincides with the basal contact of the Waterford Formation in the western Karoo Basin. These deposits reach a maximum thickness along the southern outcrop belt, north and northeast of Prince Albert, and thin considerably towards the north and east where they grade into argillaceous prodelta deposits.

The succession consists of upwards-coarsening cycles each comprising (facies 2) alternating mudstones and sandstones at the base, grading upwards into laterally persistent sandstone sheets showing low-angle planar cross-bedding and plane bedding. These sandstones are up to 16 m thick, have sharp to gradational lower contacts, and wave-ripple marks on their upper contacts. In places they exhibit spectacular examples of ball-and-pillow structures (FIG. 5.1) and other soft-sediment deformation structures (FIG 5.2). Body fossils in these sandstones consist of rare bivalve fossil shells, found to date only on the farm Zwartskraal in the Prince Albert District where

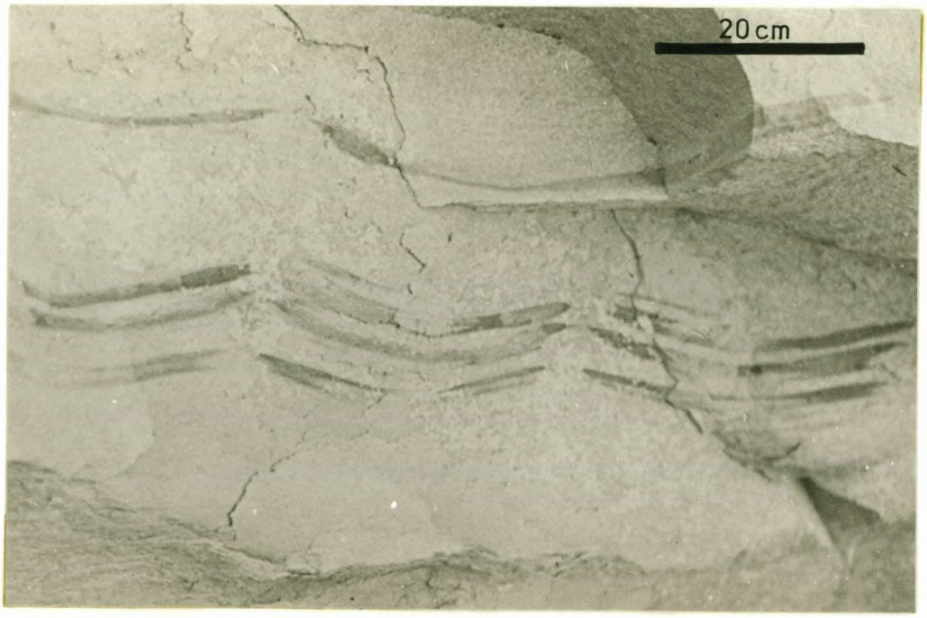


FIG. 5.2 Dish structures near the base of thick distributary mouth bar deposits in Verlatenkloof. (Bar scale = 20 cm.)

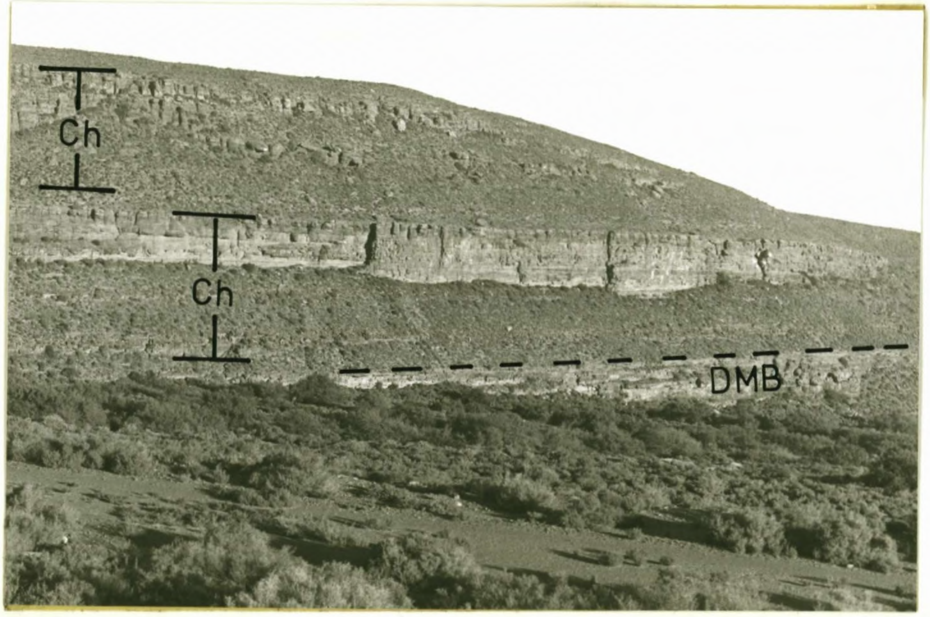


FIG. 5.3 Distributary channel deposits (Ch), each more than 20 m thick, capping distributary mouth bar deposits (DMB) in Verlatenkloof. Dashed line indicates delta front - delta plain contact.

they were discovered by R. Oosthuizen (Visser and Loock, 1979b). Trace fossils consist of invertebrate trails and worm burrows of the *Cruziana-Skolithos* ichnofacies of Seilacher (1967).

The upwards-coarsening sequences consist of prodelta and interdistributary bay deposits at the base, grading upwards into distributary mouth bar sheet sandstones (Jordaan, 1981). The basal contacts of distributary mouth bar sandstones in the western Karoo Basin are not gradational but sharp, similar to the distributary mouth bar deposits descibed in the eastern Karoo Basin by Hobday (1978). Sharp basal contacts are uncommon in distributary mouth bar sandstones (Coleman, 1976) and are usually formed by friction-dominated river discharge, which is normally the result of a small density contrast between fresh water river discharge and fresh basinal water (Hobday, 1978; Coleman, 1976). The lateral persistence of the distibutary mouth bar sandstones and the lack of erosion by distributary channels suggest on the contrary that buoyancy did play a significant role during deposition of the Ecca deposits. The second important factor controlling sedimentation at the river mouth in Ecca times, was inertia due to high outflow velocities (Coleman, 1976). The result was greater water depths seaward of the river mouth, and spreading of the effluent as a turbulent jet rather than a buoyant sheet. Friction between the effluent and sand beds occurring immediately seaward of the river mouth is a third factor affecting river mouth processes (Coleman, 1976). This process could not have played a significant role during Ecca deltaic sedimentation, as there is very little evidence of channel erosion of the distributary mouth bars. As is common in river-dominated deltas, delta front sedimentation was controlled by the interaction of more than one of the factors described. The evidence presented here for the interaction of buoyancy and inertia suggests that the Basin was saline or very brackish by the time the deltas were formed, but that outflow velocities of the river discharge was high presumably because of topographic relief between the hinterland and low-lying delta plains.

5.2.1.3 Delta Plain Deposits

The delta plain deposits overlie and, where prominently developed, laterally displace the delta front deposits. Total thicknesses range from a few metres to a maximum of nearly 150 m in the area southwest of Sutherland, namely in the Verlatenkloof. Elsewhere in the western Karoo Basin the delta plain deposits form a relatively thin capping (few metres or tens of metres) on top of the delta front succession.

The distributary channel deposits consist of thick, lenticular, "shoestring" sandstone bodies. Individual channel deposits reach a maximum thickness of more than 20 m but pinch out within a few hundred metres across strike. The channel sandstones have erosional bases and whereas the very thick deposits are multistoreyed, individual channel deposits are not easily distinguished in poor outcrops (FIG. 5.3). The Verlatenkloof examples were classified as low-sinuosity fluvial channel deposits by Turner (1978).

Distributary channel deposits consist of thick intervals of plane bedded sandstones overlying erosional bases, marked in places by thin (<0,5 m) lenses of intraformational conglomerates. Solitary sets of large-scale trough cross-bedding (Theta cross-stratification of Allen, 1963b) are present near the bases of individual storeys. Distributary channel sandstones are capped by varying thicknesses of ripple cross-laminated sandstones (facies 4a), which form distinctive rib-and-furrows on weathered surfaces. These features as well as the lack of noticeable soft-sediment deformation, distinguish them from the distributary mouth bar sandstones below. By virtue of their arenaceous composition the delta plain and delta front deposits (FIG. 5.3) can only be distinguished on the basis of their sedimentological characteristics. By contrast, the overlying alluvial plain succession is distinctly argillaceous throughout the western Karoo Basin.

The distributary channel sandstones are usually interbedded with facies 1 mudstones (Jordaan, 1981). Crevasse splay sandstone sheets and lenses up to 3 m thick are present locally in facies 1 overbank mudstones. Some of these crevasse splay deposits form thin (<3 m), perfect upwards-coarsening sequences. The upper contacts are in many instances bioturbated, otherwise ripple-marked.

A prominent shaly interval correlated with the Britskraal Member in the Eastern Cape Province (SACS, 1980) overlies the arenaceous delta plain deposits locally in the study area, notably on the farm Combrinckskraal. These deposits grade vertically into the lower Beaufort alluvial plain succession. Detailed stratigraphic correlation and palaeontological evidence by Rubidge (1983) have shown the shaly interval to belong to the deltaic Waterford Formation rather than the Beaufort Group. A lagoonal or paralic lacustrine environment of deposition is inferred. It is significant that these deposits are restricted to the central parts of the western sector of the Karoo Basin, near the postulated southern entry point into the lower Beaufort basin (cf. FIG. 8.1).

5.2.1.4 Fluvial Deposits

The transition from the delta plain to the alluvial plain deposits is abrupt, characterised by a sudden decrease in the proportion of sandstones relative to the argillaceous rocks in the succession. Only where the delta plain deposits reach a maximum thickness, namely along the western margin of the Basin in the Verlatenkloof area, was the contact found to be transitional over a few tens of metres.

The succession immediately overlying the delta plain deposits (up to 300 m thick) is distinctly argillaceous and consists mainly of green mudstones and minor sandstone intercalations. Prominent fluvial channel sandstones (>3 m thick) are rare. Mudstones are predominantly green, and red mudstones occur very rarely. The first therapsid fossils in the lower Beaufort succession occur in these mudstones (Barry, 1974; Rubidge, 1983). Rolled fossil bones and teeth are conspicuous in the intraformational conglomerates in the sandstones, and in many instances these bones are uraniferous. These beds form a transitional zone between the deltaic (lower delta plain) and fluvial (upper delta plain) deposits, and correspond in some respects to the transitional lower delta plain deposits described by Horne *et al.* (1978). (Terminology in brackets by Horne *et al.*, 1978.) These beds were classified as part of the Beaufort Group for the purposes of this investigation.

The fluvial deposits overlying the transitional beds consist of thick (>3 m), lenticular channel sandstones, interbedded with massive overbank mudstones and thin sandstone intercalations. The upper delta plain deposits described by Horne *et al.* (1978) form part of what are here considered fluvial deposits. Evidence of pedogenesis, usually calcrete nodules, is found in mudstones overlying the deltaic deposits. The first red mudstones which appear a few hundred metres higher up in the succession in the south, occur immediately above the deltaic deposits in the northern part of the study area. The fluvial deposits are reviewed in greater detail elsewhere.

5.2.2 Palaeoenvironmental Changes in the Transitional Interval

The interval between the top of the Fort Brown or Tierberg Formations and the base of the lower Beaufort reflects a gradual change from a shallow marine environment, with the maximum water depth up to a few hundreds of metres (Visser *et al.*, 1980), to a fluvial depositional environment. Comparison with modern river-dominated deltas (Horne *et al.*, 1978) has shown that this transition, from prodelta to alluvial plain, could have taken place gradually over a distance downstream of more than 50 km, a palaeoslope of nearly 1°. The abrupt change in lithological character between the prodelta and the delta front deposits, as well as contact relationships of the distributary mouth bar and distributary channel deposits, suggest that inertia and buoyancy dominated the behaviour of the river effluent. In order to account for the density contrasts required, the basinal water must have been saline or at least brackish throughout the period in which deltaic sedimentation took place in the western Karoo Basin.

Although no reliable palaeocurrent measurements could be made in the folded deltaic beds along the southern edge of the Basin, cursory observations suggest a northerly palaeocurrent direction. A pronounced easterly to east-northeasterly direction (vector mean = 80°) was measured in the thick distributary channel sandstones in Verlatenkloof, on the western edge of the Basin.

5.3 Options for an Ecca-Beaufort Contact

Detailed sedimentological analysis of the Ecca-Beaufort transition was carried out purely to determine a palaeoenvironmental framework to use as a reference throughout the study area. Structural deformation, weathering, and regional facies changes within the transitional interval have changed the field appearance of the succession to such an extent that visual comparison was of limited use in stratigraphic correlation. The lithological and palaeontological criteria used in the past to distinguish the Ecca and Beaufort Groups, are reviewed and some proposals based on field relationships are presented.

The diachronous nature of the Ecca-Beaufort sedimentation (Jordaan, 1981) is an important aspect of deltaic sedimentation in the Karoo Basin. The scale of diachronism in the Basin has an important bearing on isopach mapping and geometric reconstruction of the Basin. The options which emerged as possible stratigraphic contacts are reviewed below.

5.3.1 Base of Delta Front Deposits

This horizon coincides with the basal contact of the arenaceous Waterford Formation. The Waterford Formation varies in thickness from less than 200 m up to nearly 800 m along the southern outcrop belt, west of 24°E longitude. These thickness variations are due mainly to the basal portions of the delta front deposits grading laterally into argillaceous prodelta and shelf deposits (Fort Brown and Tierberg Formations). Elevation differences of up to 600 m could thus be expected in a longitudinal section of the Basin, over and above differences in stratigraphic thickness resulting from diachronism.

5.3.2 Base of Delta Plain Deposits

The delta plain distributary channel deposits are recognisable over most of the study area, forming part of the prominent arenaceous Waterford Formation, together with the underlying delta front deposits. The elongated channel and lenticular crevasse splay sandstone bodies in the delta plain succession are readily distinguishable from the tabular distributary mouth bar sandstones of the delta front environment. Stratigraphic profiles measured along the western margin of the basin show that in areas where the delta plain deposits reach their maximum thickness they are intertongued with coeval delta front deposits. Some specialised sedimentological knowledge will be required to map the contact between the delta front and delta plain deposits.

5.3.3 Top of Delta Plain Deposits

The delta plain deposits form the uppermost arenaceous interval of the Waterford Formation, as defined by SACS (1980), and in places amount to a thin veneer only a few metres or tens of metres thick. Prominent fluvial channel sandstones rarely overlie the delta plain deposits directly, as for instance at the top of Ganaga Pass on the western margin of the Basin. However, these sandstones are distinctly lenticular and can be distinguished readily from the delta plain deposits. Reptilian fossils (Rubidge, 1983) and pedogenic features, such as calcareous concretions and red mudstones, are indicative of a terrestrial setting.

5.3.4 First Prominent Fluvial Channel Sandstone

The first prominent fluvial channel sandstones above the deltaic deposits occur between a few tens and a few hundreds of metres above the upper contact of the Waterford Formation. These sandstones are distinctly lenticular in transverse section, and consequently of limited value in regional stratigraphic mapping.

5.3.5 First Red Mudstones

The first red mudstones in the succession were mapped as the Ecca-Beaufort contact by Haughton *et al.* (1953), Rossouw *et al.* (1964), and many others. Venter (1969) pointed out the inconsistency of this horizon as determined in petroleum exploration wells drilled by SOEKOR. In the present investigation the first red mudstones were found to differ in stratigraphic elevation by more than 1000 m over a few tens of kilometres southwest of Beaufort West (Jordaan, 1981). Although the first red mudstones appear immediately above the deltaic deposits in the northern parts of the study area, they are not consistent enough to be used in the southern part of the western Karoo Basin.

5.4 **Proposed Contact**

The question of a stratigraphic contact between the Ecca and Beaufort Groups involves firstly a choice between a lithostratigraphic and a chonostratigraphic contact. The latter proposition was favoured by Winter (1984) who mentioned the Poortjie Sandstone Member and Britskraal Shale Member (SACS, 1980), as possible candidates. The Britskraal Shale Member is very rarely developed along the southern margin of the Basin, and was identified only on the farm

Combrinckskraal. These strata were eliminated together with other potential chronostratigraphic markers in the north, as a result of the delta front and delta plain deposits grading diachronously into the argillaceous prodelta deposits (Tierberg Formation) in the northern part of the Basin. The basal contact of the Poortjie Member, on the other hand, occurs more than halfway up in the succession of typical lower Beaufort strata and the feasibility of placing a contact at that position is questioned.

With no regionally defined chronostratigraphic markers evident in the vicinity of the transition from the deltaic to alluvial plain deposits, one is left with a choice of one of the lithostratigraphic options outlined previously. The sharp lithological changes which mark the transition between the deltaic and the fluvial deposits are directly related to the change in the depositional environment. This horizon was used as a contact between the Ecca and Beaufort Groups throughout the present investigation. In the western Karoo Basin (where the Britskraal Shale Member or equivalent interval is largely absent), this contact is essentially the same as that recommended by SACS (1980), namely at the top of the Waterford Formation.

6.0 SEDIMENTARY FACIES AND DEPOSITIONAL ENVIRONMENTS OF THE LOWER BEAUFORT

6.1 Definition and Terminology

The facies concept remains controversial in sedimentology one and a half centuries after it was introduced by Gressly (1838). Studies of sedimentary facies relationships in fluvial deposits in particular have been used successfully to establish a framework to compare modern and ancient deposits (reviewed in Walker, 1979).

There has been considerable disagreement on the definition and exact meaning of the term "facies" (reviewed in Teichert, 1958, and in Weller, 1958). It is widely accepted that Gressly (1838) introduced "facies" as a purely descriptive term (Teichert, 1958). Romanovskij (1975), however, pointed out that Gressly (1838) took the environmental aspects of rocks into account as well by distinguishing, for instance, "marine littoral", "marine pelagic", and other environment-related facies. From numerous references it appears that in studies of ancient deposits the term is used mainly descriptively. The environmental connotation appears to have been added recently in studies of modern sedimentary environments, where depositional processes and products could be linked directly to the observed facies. As a result the term "facies" is now often used incorrectly as synonymous with "environment" or "deposit".

Reading (1978) suggested that facies should be used in a genetic as well as an environmental sense, on condition that the way in which it is used was clearly defined. Walker (1979) pointed out that facies defined in the field may have ambiguous interpretations, and that environmental interpretations should be reserved until the final stages of an investigtion. An example of this problem from the early years of uranium exploration in the lower Beaufort involved certain ripple-marked sheet sandstone deposits which were initially interpreted as tidalites, and the enclosing succession as marginal marine (tidal flat) deposits (Pretorius, 1977). The sandstones were subsequently shown to be crevasse splay deposits in an alluvial succession. An example of the extremes to which the interpretative aspect of facies is taken occasionally, is found in Eyles *et al.* (1983). They distinguished certain sedimentary facies on the strength of evidence for "current reworking", a purely subjective aspect of sedimentary rocks.

Teichert (1958) considered the terms "lithofacies" and "facies" to be synonymous but pointed out that use of the term "lithofacies" has merit when dealing with the purely inorganic aspects of a rock unit. Reading (1978) used the term "lithofacies" in a strictly observational sense. Preference is given here to the use of the term "facies" in a broader sense, including the mineralogical and petrographic as well as biological aspects, as intended by Gressly (1838). Following suggestions by Teichert (1958) the term is used here purely descriptively and never in an interpretative sense, but with the understanding that facies will ultimately be used in palaeoenvironmental interpretations.

6.2 Sedimentary Facies of the Lower Beaufort

6.2.1 Introduction

Altogether 14 sedimentary facies were described by Stear (1980b) in terms of a classification proposed initially by Miall (1977) for braided stream deposits. Cole (1980) recognised all 19 facies described by Miall (1978a), and identified the presence of Markov-dependent facies sequences as well as upwards-fining sequences. In order to keep the scheme of facies subdivisions simple as recommended by Miall (1973), only five coarse-grained and four fine-grained facies are distinguished here, based partly on the scheme of Miall (1978a).

Following a trend set in studies of braided stream deposits, on which Miall's facies scheme is based, the emphasis in most fluvial facies analyses has until now been placed mainly on the coarse-grained facies. The fine-grained facies have thus also been neglected in detailed studies of the Beaufort Group. Kübler (1977) distinguished between only one mudstone and one siltstone facies. Stear (1980b) recognised four fine-grained facies, two of which were identical but were distinguished on the grounds of the positions they occupied relative to channel sandstone bodies. Cole (1980) described six siltstone and one mudstone facies.

Individually the fine-grained facies, particularly when subdivided in too much detail, show a lack of organisation. When subdividing the fine-grained successions into facies assemblages certain regional trends become apparent, reflecting subtle variations in depositional subenvironments. Only four fine-grained facies are here distinguished in the lower Beaufort, but a more detailed investigation of the argillaceous intervals taking into account their fossil content, should reveal a more practical scheme.

6.2.2 Coarse-grained Facies

The coarse-grained (rudaceous and arenaceous) facies were subdivided on the basis of lithology and sedimentary structures. The parameters have been discussed in previous sections and only the general characteristics are listed here. The facies codes (in brackets) are as shown on stratigraphic profiles. Certain facies were not encountered in any of the sections (e.g. the planar cross-bedded sandstone facies) but were observed elsewhere in the lower Beaufort strata. Except where sandstones form part of the fine-grained facies, a minimum thickness of approximately 10 cm was set for the coarse-grained facies. Facies boundaries are erosional, sharp or gradational.

6.2.2.1 Erosional Scours (Se)

Erosional scours occur at the bases and within sandstone beds, throughout the lower Beaufort. The relief of scours changes from one type of channel sandstone to another. Some scouring is also associated with thin, lenticular, crevasse splay sandstone beds, but is not included in this classification. This facies was proposed by Miall (1978a).

Interpretation

The scouring associated with lower Beaufort sandstone beds was caused by channel erosion of more competent, argillaceous alluvial plain deposits, which formed the banks of fluvial channels.

6.2.2.2 Massive Conglomerate Facies (Gm)

The massive conglomerate facies is the only rudaceous facies distinguished, and comprises massive or vaguely stratified intraformational conglomerates (FIG. 4.10). This facies usually occurs as discrete beds from a few centimetres up to nearly 2 m thick. The lower contacts are invariably erosional whereas the upper contacts are sharp or gradational over a few centimetres, also rarely erosional where overlain by younger channel deposits.

The conglomerates are mostly clast-supported. No distinction was made between different clast assemblages but for detailed analyses it may be useful to distinguish assemblages dominated by (black) shale clasts, red and/or green mudstone clasts, and calcrete clasts. The "black shale" conglomerates are, for instance, restricted to thick channel sandstones of the Combrinkskraal and Koup members in the southern part of the Basin. Calcrete-clast conglomerates are associated with massive green mudstone successions (facies Fg), and detailed studies may yet reveal relationships between the conglomeratic, sandstone and fine-grained facies.

Interpretation

The massive conglomerates are "lag" deposits formed in fluvial channels. They were locally derived from channel banks, and deposited in the lower point bars as well as in longitudinal bars of meandering streams (Allen, 1965a; Miall, 1978a).

6.2.2.3 Plane Bedded Sandstone Facies (Sh)

The plane bedded sandstone facies is the most abundant of the coarse-grained facies, occurring in cumulative thicknesses of up to more than 10 m. Current lineations are conspicuous on bedding planes, but are often absent or invisible in thick sandstone deposits of the Koup member in the northwestern parts of the basin. The lower contacts are erosional or sharp, and the upper contacts sharp. In many instances thick plane bedded intervals are subdivideed by subtle, internal erosional surfaces. Individual facies Sh intervals rarely exceed a thickness of 1 m (cf. Turner, 1981), and is in places replaced along strike by facies St.

Interpretation

Plane bedded sandstones were deposited in the upper flow-regime of lower Beaufort fluvial channels.

6.2.2.4 Trough Cross-bedded Sandstone Facies (St)

The trough cross-bedded sandstone facies consists of solitary sets or cosets of large-scale trough cross-bedded sandstone, with sets ranging in thickness from just over 10 cm up to a maximum of approximately 1,2 m. Facies St occurs in the basal parts of moderately thick sandstone deposits, also very rarely right at the base of sandstone beds. In places facies St is found at the top of channel sandstone deposits in the Nuweveld member, substituting the cross-laminated sandstone facies (Sr). The overall abundance of facies St is low throughout the lower Beaufort, compared with fluvial deposits generally. The total thickness rarely comprises as much as 20 per cent of individual channel sandstones, and facies St is in many instances entirely absent from thick channel sandstone deposits.

Interpretation

Trough cross-bedded sandstones were deposited by migrating trains of linguoid dunes, which is formed in the upper part of the lower flow regime.

6.2.2.5 Massive Sandstone Facies (Sm)

As previously discussed, massive bedding of a primary nature is probably rare in sandstones of the lower Beaufort, and in many instances it appears to be the result of post-depositional homogenisation of bedded sands. There are numerous examples in the lower Beaufort sandstones where facies Sm grades laterally into facies Sh. Cole (1980) on the other hand found an association between massive bedding and trough cross-bedding in the lower Beaufort (megasequence B) channel sandstones he investigated. Jones and Rust (1983) suggested that in some instances massive bedding may be caused by the collapse of large bedforms in large channels. Massive bedding also forms as a result of very rapid deposition of sand from suspension, or as a result of deposition from concentrated sediment (sand) loads (Blatt *et al.*, 1972).

Interpretation

Massive bedding in lower Beaufort fluvial channel sandstones was formed by one or a combination of: rapid sedimentation, soft-sediment homogenisation of bedded sands (mainly as a result of dewatering), and bioturbation.

6.2.2.6 Small-scale Cross-laminated Sandstone Facies (Sr)

The small-scale cross-laminated sandstone facies comprises all the different types of ripple cross-lamination in sandstones, listed by Allen (1963b) and Jopling and Walker (1968). Facies Sr occurs almost invariably at or near the tops of intact channel sandstone storeys, and occurs abundantly in overbank sandstone sheets and lenses. In the latter case thin intervals of Sr were not specified separately in measured sections but were included in the fine-grained facies. In channel and crevasse sheet sandstones the lower contacts are sharp, and the upper contacts sharp to gradational. Facies St is often seen to replace facies St in the top parts of channel sandstones, in the Nuweveld member.

Interpretation

Facies Sr was deposited by (linguoid) current ripples in the lower flow-regime of currents during the waning stages of flooding in streams and crevasse splays.

6.2.3 Fine-grained Facies

The argillaceous successions interbedded with fluvial channel sandstones do not show systematic facies relationships. The fine-grained facies nevertheless show significant stratigraphic and palaeogeographic distributions when viewed on a broader scale. For this reason these deposits were subdivided into facies on the basis of the predominant lithology, internal stratification, and fossil content. The distribution and relative abundance of reptilian fossils in each of the facies were estimated partly from the locality maps of Kitching (1977) and Keyser and Smith (1977-78).

A minimum thickness in the order of 1 m was set for fine-grained facies, where this was feasible in field measurements. Facies boundaries are usually transitional, particularly of facies Fg (massive green mudstones) and Fr (massive red and green mudstones). Several fine-grained facies are often interbedded in the same overbank sequences.

6.2.3.1 Laminated Mudstone Facies (Fl)

Laminated and wavy-laminated mudstones very rarely occur in appreciable thicknesses (>1 m) in the lower Beaufort, and the paucity of these shaly rocks is one of the characteristics used to distinguish the Beaufort Group from the Ecca Group (SACS, 1980). Laminated mudstones occur sporadically in the lowermost stratigraphic unit (megasequence A) of the lower Beaufort in the southern parts of the Basin. Thicknesses of up to a few tens of metres were recorded on Combrinkskraal. Strata correlated with the Britskraal Member at the top of the Ecca Group, as for instance on the farm Combrinkskraal, consist mainly of facies Fl.

The facies consists of greyish-olive to olive-grey, finely laminated mudstones and silty mudstones. Wavy laminations are present in places. Minor soft-sediment deformation is commonly associated with sandy intercalations, as seen for instance on the farm Bloukrans, just north of Combrinkskraal (Prince Albert). The rocks are here referred to as "mudstones" rather than "shales" in order to avoid confusion with typical marine shales. Calcareous nodules, bioturbation, and animal and plant remains are absent.

Interpretation

Deposition took place mainly from suspension in open standing bodies of water. Wavy laminations are the result of weak tractive currents. The absence of bioturbation and large-scale homogenisation in facies Fl suggests that the environment was hostile (brackish or too deep?) to freshwater organisms which were very active in other pasrts of the lower Beaufort basin. The depositional environment was probably a floodplain which was completely inundated throughout the deposition of this facies.

The dark, fissile mudstones from which the "black shale" clasts in intraformational conglomerates were derived, were similarly deposited from suspension but in temporarily ponded segments of ephemeral streams. Their depositional setting (within channels), and consequent low preservation potential of these shales, account for their apparent absence from the succession.

6.2.3.2 Interbedded Green Mudstone and Sandstone Facies (Fg)

Facies Fg consists of interbedded greyish olive or olive grey to dark grey, massive mudstones and siltstones, as well as interbedded thin sandstones in a maximum cumulative thickness of up to several tens of metres. Thin (<2 m), plane-bedded, cross-bedded and ripple cross-laminated sandstones and silty sandstones represet up to 50 per cent of these successions, with an average of 20 to 30 per cent. The sandstone beds are both upwards-fining and upwards-coarsening. Rhythmically interbedded mudstone and sandstone deposits are present in places (FIG.6.1). Calcareous nodules occur abundantly as aggregates in sheets and lenses, with a large proportion of septarian nodules. Incipient palaeosols are present, as evidenced by pedogenic calcretes in green mudstone beds. Red mudstones are conspicuously absent. Tuffaceous beds have so far been identified only in facies Fg (FIG. 4.5).

The mudstones appear massive but bedding on a scale of a few centimetres and a few tens of centimetres can be seen in some fresh outcrops. Pervasive bioturbation in these beds is accompanied by numerous, small calcareous nodules up to 5 cm in diameter. Compaction and dewatering account for much of the homogenisation of individual mudstone beds, as shown by the presence of soft-sediment deformation (FIG. 4.19).

Reptilian and plant fossils are abundant, the latter associated with sandstone and siltstone intercalations. The only examples of *Glossopteris* recorded in the lower Beaufort succession were found as well-preserved, delicate leaf imprints in silty sandstones near the top of thin upwards-coarsening sequences on the farm Tuinplaas (Prince Albert District). Other fossils include ferruginised imprints of plant fragments, possibly of *Phyllotheca* and *Schizoneura*. Silicified wood fragments are conspicuous in interbedded sandstones of the Koup member in the area north of Aberdeen, and a 7 m long silicified tree trunk (*Dadoxalon?*) was observed in the Nuweveld member in the Murraysburg District. Freshwater mollusca are present in facies Fg on the farm Spreeuw Fontein. Several other finds of freshwater lamellibranchs in the lower Beaufort were described by Rossouw (1970) but the host rocks are uncertain.

Boonstra (1969) noted that a large proportion of the pareiasaurs in the old *Tapinocephalus* zone were found as complete skeletons in "blue" mudstone, buried in a standing position. Most of the lower portion of the *Tapinocephalus* zone consists of massive green mudstones of facies Fg, and *Bradysaurus* skeletons occurring in an upright position on the farm Rietfontein near Prince Albert (J.C. Loock, pers. comm., 1979) originated from facies Fg as well.

Interpretation

The massive green mudstones of facies Fg were deposited from suspension. Most of the mudstones largely lack evidence of deposition in standing water (e.g. wave ripple-marks) and were deposited subaerially in a well-drained environment, in the proximity of large rivers. The paucity of pedogenic features (e.g. calcareous nodules) suggests a rate of sedimentation too high to allow soil formation (Leeder, 1975a and b). A proximal alluvial floodplain setting is interpreted on account of the scale of the bedding in mudstones, the high proportion of interbedded sandstones, and the presence of terrestrial (reptilian) fossils. The interbedded sandstones were formed during flooding events when bedload spilled from channels onto the floodplains through crevasse splays in natural levees (described elsewhere). Unusual examples of facies Fg consisting of rhythmic alternations of sandstone and dark green mudstone (FIG.6.1), were deposited by periodic influx of sand into standing bodies of water. Topographic relief on the floodplain in the example depicted in FIG. 6.1, is estimated to have been in the order of 3,5 m, the total thickness of the flood basin deposit.

The erect position in which numerous pareiasaur skeletons were preserved in facies Fg mudstones, suggests that these reptiles were trapped in muddy ponds or lakes which existed locally on floodplains. The paucity of plant fossils in this facies seems to be in conflict with the large proportion of herbivorous reptilians in the lower Beaufort (Boonstra, 1969; Kitching, 1977). From the presence of large fossil tree trunks in interbedded channel deposits (latter not included in facies Fg), it appears as if vegetation was thickest in those parts of floodplains situated closest to the fluvial channels.

6.2.3.3 Interbedded Red Mudstone, Green Mudstone, and Sandstone Facies (Fr)

Facies Fr consists of interbedded massive, red and green mudstones with subordinate, thin sandstones and silty sandstone beds. The mudstones show irregular variations in colour and texture, and bedding on a scale between a few centimetres and more than 3 m thick. Facies Fr displays complex interfingering with either facies Fg or facies Fv, in successions which show rapid vertical and lateral changes in lithological character.

Sandstone intercalations occur as thin (<2 m) lenses in facies Fr, rarely comprising as much as 50 per cent but usually much less (15 to 25 per cent). Palaeosols, desiccation cracks, plant stem casts and calcareous nodules are abundantly present. Thin rootlets have been identified in this facies at Riemhoogte near Beaufort West. Irregularly shaped quartz and calcite inclusions resembling those described by Keyser (1966) in Beaufort Group mudstones in the Graaff-Reinet District, were also found in facies Fr in the Nuweveld member on the farm Toorwater (location 30, FIG. 1.2). Reptilian fossils are abundant, and trace fossils are more abundant or more conspicuous in this than in the other fine-grained facies. Silicified tree trunks are present and imprints of smaller plants as well as bioturbation are more abundant than in facies Fg.

Interpretation

Facies Fr was deposited mainly from suspension, by processes similar to those in which facies Fg was formed. However, the red mudstones and siliceous inclusions are indicative of a slightly more arid environment. Small plants were present and possibly even abundant in certain parts, as suggested by the presence of plant fossils and rootlet beds. The environment was probably too arid and the water table too deep to support the growth of large trees. The low rate of sediment aggradation, estimated at less than 3 mm per year (cf. Leeder, 1975a), allowed the formation of palaeosols. A flat, moderately vegetated, irregularly flooded, terrestrial environment such as a medial to distal floodplain, is interpreted for facies Fr. Situated some distance away from meander belts, those parts were only periodically inundated during high-level floods.

6.2.3.4 Variegated Mudstone Facies (Fv)

Thick successions of variegated green and red mudstones of facies Fv are characteristic of the uppermost members of the lower Beaufort, particularly the Teekloof and Gifkop members. Uninterrupted thicknesses of more than 100 m were logged along the Nuweveld escarpment. The mudstones consist of variegated red and green mudstone couplets 10 to 30 cm thick. Each couplet is defined by a sharp contact and consists of red mudstone at the base, grading upwards into green silty mudstone. The mudstones are massive and mottled on the transition from red to green mudstones. Thin sandstone intercalations are occasionally present in thick successions of facies Fv. Palaeosols as known elsewhere in the lower Beaufort have not been identified but small incipient calcareous nodules (few centimetres in diameter), are present in places. Reptilian fossils are very rare in these beds (R.M.H. Smith, pers. comm., 1981) and no plant fossils were noted in either the mudstones, or in the associated sandstones.

Facies Fv of the lower Beaufort shows a strong resemblance to the variegated redbeds in the Cathedral Bluffs Tongue of the Wasatch Formation in the United States (Braunagel and Stanley, 1977). The Cathedral Bluffs redbed couplets differ from the lower Beaufort examples insofar that the former appear to be coarser grained as well as upwards-fining. They consist of green, cross-laminated, clayey siltstones at the base overlain by red, horizontally laminated, silty mudstones.

Interpretation

Red and green mudstone couplets were deposited mainly from suspension during flooding events, occurring seasonally or even thousands of years apart (Braunagel and Stanley, 1977). The basal red mudstones of each couplet were deposited from suspension in floodwaters, ahead of weakly tractive currents which brought in slightly coarser-grained, green, silty mudstones on top. Ferrous iron as well as argillaceous material were leached from the topmost layers of each successive flood deposit by fluctuating groundwater, runoff drainage, and connate water (expelled during compaction). The iron and clay accumulated in the basal layer of each flood deposit, leaving a depleted upper portion. Redistribution and oxidation of the iron had probably been completed completed before the next flooding event, which effectively sealed the layer below from further oxidation. Vertical and lateral groundwater movements thus ceased shortly after deposition, which probably accounts for the preservation of the variegated textures.

Although the massive appearance of individual mudstone beds obscures the post-depositional history of these beds it is unlikely that pedogenic processes, as seen for instance in facies Fr, had much of an effect on the sediments after deposition. Conditions were too dry for widespread organic activity or near-surface groundwater fluctuation, normally required for soil formation (Buurman, 1975). An arid climate and resulting sparse vegetation would explain the absence of reptilian and plant fossils from these deposits.

6.2.4 Palaeoenvironmental Setting

The fluvial origin of the lower Beaufort facies association was discussed by Kübler (1977) and Cole (1980). A composite lower Beaufort facies sequence consists of an erosional scour (facies Se) with or without an overlying intraformational conglomerate (facies Gm), grading upwards into the sandstone facies Sh, St, Sm and Sr, and ultimately into facies Fl, Fg, Fr and Fv on top. These form distinctive upwards-fining facies sequences which compare well with the facies model proposed for the Devonian Old Red Sandstone in Britain, and the Devonian beds of the Appalachian region of the U.S.A. (Allen, 1964a). The coarse-grained members represent channel deposits, and the fine-grained facies represent overbank (floodplain and flood basin) deposits. Variations of the ideal upwards-fining facies model in the lower Beaufort, are discussed below.

6.3 Fluvial Architecture

The facies relationships of sedimentary successions as a rule deal mainly with their vertical profiles, which is essentially a one-dimensional facies arrangement. In order to reconstruct ancient alluvial systems such as those that gave rise to the lower Beaufort, the facies concept had to be modified. Friend (1983) proposed that the deposits of river systems (alluvial architecture) be classified on the basis of the identification of two- and three-dimensional, macro-scale features including bars, channels, sandstone geometries, and the behaviour (or absence) of channels. Miall (1985) proposed "architectural-element analysis" as an approach to the analysis of fluvial deposits, in which the nature of eight different elements such as channels, bedforms, etc., are determined. These features include the nature of the bounding surfaces, external shape, scale, and internal geometry of each of those elements, which also served as a basis for the field classification of the lower Beaufort fluvial deposits as outlined below. Smith *et al.* (1989) described the evolution of (meandering) river channelways, from crevasse splaying to the eastablishing of new meander belts, based on the processes active in the modern Saskatchewan River (Canada). They suggested four successive stages of floodplain evolution: (1) the initial avulsion stage, (2) the anastomosed stage, (3) the reversion stage, and (4) the single channel stage.

6.3.1 Fluvial Palaeochannel Deposits

The fluvial palaeochannel deposits form prominent sandstone "ribbons" and "sheets" up to a few tens of metres thick, interbedded with variable thicknesses of mudstone. Individual channel sandstones are up to 60 m thick and usually consist of several superimposed, upwards-fining sandstone sequences, referred to as "multistoreyed" channel deposits. The margins of individual channel sequences (FIG. 6.2 and 6.3) are rarely seen because channel sandstones are mostly elongated and exposed parallel to the Nuweveld escarpment. Channel margins may also be obscured as a result of the composite nature of the channel deposits. The original channel morphologies and bedforms can no longer be identified in fluvial deposits, and these are reconstructed on the basis of facies and geomorphic relationships.

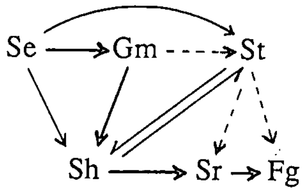
6.3.1.1 Channel Facies Relationships

Fluvial channel sequences all have sharply scoured, erosional contacts truncating overbank mudstones and older channel sandstones deposits. The sheet-like channel sandstones such as in the Koup member in the type area generally have low-relief contacts, whereas the lenticular channel deposits in the Nuweveld and Leeukop members all show high-relief contacts.

Scoured contacts are typically overlain by facies Gm. The contacts between the conglomerates and the overlying sandstones are sharp or transitional over a few centimetres, and in places these contacts are erosional too. The basal sandstone intervals are mostly plane bedded (facies Sh) or massive (facies Sm). In certain channel sandstones of the Nuweveld and Leeukop members the basal contact is overlain directly by facies St. Facies St is erosively emplaced in facies Sh or Gm, usually in the basal parts of channel sandstones. It is overlain by facies Sh or Sr, and has been observed to substitute facies Sr in the top parts of channel sequences at a number of localities in the Nuweveld member.

Channel sequences are capped by facies Sr intervals up to 2 or 3 m thick. They form rib-and-furrows on bedding planes. The upper contacts with the levee and floodplain deposits are sharp, although facies Sr rarely shows an upward transition into massive or ripple cross-laminated, calcareous, abandoned channel-fill siltstones.

The facies sequences in lower Beaufort meandering channel sandstones, based on observations in channel sandstones throughout the succession, are as follows (thicker arrows denote transitional probabilities higher than average, and dashed lines transitional probabilities lower than average):



6.3.1.2 Palaeochannel Sinuosities

The characteristics of alluvial channels are largely determined by their geometries. The planform shape of a channel is referred to as the "sinuosity" and defined as the ratio of the arc distance to half the meander length, measured between two inflection points (Leopold and Wolman, 1960). Sinuosity was expressed mathematically by Langbein and Leopold (1966, p.5), and calculated in terms of the maximum angular deviation in streamflow directions (measured in radians). The equation of Langbein and Leopold (1966) was converted by Miall (1976) for angular deviations (ø) measured in degrees instead of radians, as follows:

$$P = 1/[1 - (ø/252)^2]$$

The method used by Leopold and Wolman (1960) yields *sinuosity ratios* (R) slightly different to those using the method of Miall (1976), for similar angular deviations. The latter method was used throughout the present investigation. Values quoted here do not relate directly to those quoted by Leopold and Wolman (1960) and Schumm (1968) for meandering streams in North America, nor with those quoted by Le Roux (1979) for lower Beaufort sandstones, all of whom used the method of Leopold and Wolman (1960).

Angular deviations were calculated from the difference in mean current directions, in each individual sequence which could be identified within palaeochannel sandstones. Only maximum angular deviations were used to calculate channel sinuosities, and only in those channel deposits in which a representative number of measurements could be obtained for each sequence. Meandering channel deposits are not always exposed on surface at those positions where stream directions crossed each other at maximum angles. That means that the angular deviations in palaeocurrent directions, as shown for instance on the stratigraphic profiles (APP. 1), do not necessarily reflect the true channel sinuosity.

Low-sinuosity meandering channels in the lower Beaufort have sinuosity ratios (P) of less than 1,5. Channels with higher sinuosity ratios yielded deposits typical of high-sinuosity meandering streams, characterised by a lenticular geometry, evidence of lateral accretion, and multistoreyed upwards-fining sequences.

6.3.1.3 Facies Models

The ideal facies sequence cited previously is rarely fully developed, and the various channel facies sequences recognised in different parts of the succession are in many places incompletely preserved. The lower Beaufort channel deposits reveal an uncomplicated history of cyclical deposition. Altogether four different types of three major channel facies models are distinguished.

6.3.1.3.1 Type 1 Channel Deposits

Type 1 channel sandstones occur in the lowermost few hundred metres of the Combrinkskraal member along the southern margin of the Basin, in the area between Prince Albert and Klaarstroom. They occur as laterally persistent, multistoreyed deposits, typically up to 10 or 12 m thick. Up to ten superimposed sequences each less than 1,2 m thick have been observed in channel sandstones up to 8 m thick on the farm Combrinkskraal.

In these sandstones each channel sequence (storey) is sharply truncated by a succeeding sequence. Each sequence consists of facies Sh (or St) overlying a low-relief erosional contact, followed by facies St, or rarely facies Sr, on top. Typical upwards-fining sequences are rare. Accurate palaeocurrent measurements could not be made but parting lineations and poorly exposed trough axes suggest northerly transport directions.

Interpretation

The internal cyclicity, immature sandstone composition, and palaeocurrent directions suggest that type 1 channel sandstones were deposited during ephemeral flooding events in low-sinuosity (braided) streams. Their palaeogeographic and stratigraphic location at the base of the lowermost (Combrinckskraal) member southwest of Beaufort West suggests that these sandstones were deposited just beyond the distal reaches of alluvial fans which entered the Basin from the south. The considerable degree of scouring evident at the top of channel sequences, limited thicknesses of individual sequences, and the presence locally of trough cross-bedding at the top of sequences suggest that flow depth was somewhat greater than that suggested by the preserved channel sequence thickness. It is likely that complete channel sequences were deposited and then eroded during succeeding flooding events. Facies relationships and the lack of evidence for lateral accretion suggest that these sandstones are braided stream deposits comparable to the Bijou Creek-type of Miall (1977). However, the abundance of internal scours and facies St is more characteristic of the Malbaie-type (facies S_1) of Rust (1978).

6.3.1.3.2 Type 2 Channel Deposits

Type 2 channel deposits consist of thick sandstones dominated by facies Sh, in places with interbedded facies St (solitary sets), and facies Sr on top. The sandstones occur as laterally persistent, tabular sheets typically 7 to 13 m thick, with younger channels in places entrenched in older sheet deposits (FIG.6.2). Facies Gm occurs abundantly in thicknesses of up to 1,2 m, usually concentrated at the bases of sandstone sheets.

The type 2 deposits occur sporadically in the lower portion of the Combrinkskraal member in the southern parts of the Basin, and are typical of the Koup member in the Moordenaars Karoo (FIG.6.2). Individual channel sequences within the thicker channel sandstone deposits are

FIG. 6.3 Channel edge of multistoreyed, meandering channel sandstone deposit. Lower upwards-fining sequence wedging out towards right in picture. Note compaction and sagging of mudstones below channel base.

FIG. 6.4 Three lateral accretion deposits in a meandering (type 3b) channel sandstone. Direction of point bar accretion shown by arrows, and scoured contacts as dashed lines.

delineated by faint, internal, scoured surfaces. Stear (1985) showed that different pulses within a single flooding event can produce internal scouring of this nature. Channel sequences have sharp contacts with the overlying mudstones. Typical upwards-fining sequences such as those found in the type 3 channel sandstones, are rare. The lateral persistence of the sandstone sheets and lack of deeply scoured (channelised) features make it difficult to reconstruct the original channel geometry. Width:depth ratios of just more than 10 were calculated for some of the younger channels truncating the Koup sandstone sheets (cf. FIG. 6.2). Palaeocurrent measurements in channel sandstones in several areas showed consistent, unimodal current directions.

Interpretation

Fluvial sheet sandstones have become widely accepted as low-sinuosity channel deposits (e.g. Nami and Leeder, 1978). They were classified as mixed load channel deposits by Schumm (1968). Collinson (1978) questioned the assumption by Allen (1965a) and others, that fluvial sheet sandstones were produced by the lateral migration of low-sinuosity channels, and concluded that they were formed by the lateral coalescence of lenticular (point bar) deposits. The lower Beaufort examples lack the diagnostic cross-cutting erosional contacts which would have been expected if the sandstones had been deposited by coalescent point bars.

The younger, scoop-shaped channels entrenched in the older sheet deposits suggest a probable mode of deposition of the type 2 deposits. Initially small-scale channels were scoured into floodplain and pre-existing channel deposits by crevasse splays, or possibly even by sheetfloods (Collinson, 1978). This was followed by bank erosion and channel widening, as described in mixed-load channels by Schumm (1968). Bedload (sand and gravel) deposition followed during the waning stages of major flooding events. The sandstones are thus interpreted as braided stream deposits, and resemble the Bijou Creek-type of Miall (1977) and the S_1 facies of Rust (1978). An example of these channel deposits occurring on the farm Palmietfontein, south-southeast of Fraserburg, was shown in plan by Stuart-Williams (1981, PLATE 3).

6.3.1.3.3 Type 3 Channel Deposits

Type 3 channel deposits occur in the uppermost section of the Combrinkskraal member in the southern parts of the study area, in the whole of the Combrinkskraal member in the northeastern half of the study area, in the Koup member in the central and northern parts of the study area, and in the entire succession above the Koup member throughout the study area. They consist of multistoreyed sandstone deposits, each comprising up to three or four superimposed, upwards-fining channel sequences (FIG. 6.3 and 6.4). Each sequence (sandstone storey) is usually 2 to 3 m thick, rarely up to 6 m thick. Individual upwards-fining sequences are lenticular (FIG. 6.3) but usually stacked (FIG. 6.4) to form sandstone sheets varying in shape from lenticular to almost tabular. Individual channel sandstones are commonly connected. They reach a maximum thickness of just over 20 m in the Nuweveld member on the farm Wilgerbosfontein, northeast of Laingsburg.

Channel sequences consist of facies Gm overlying high-relief erosional contacts, followed upwards by facies Sh (with or without interbedded facies St), followed by facies Sr (replaced in places by facies St), and grade into massive overbank mudstones and siltstones at the top. Channel sandstones are in many instances capped by dark, abandoned channel-fill mudstones (FIG. 6.5). Fish fossils associated with type 3 channel deposits include *Atherstonia minor* (Woodward) found on the farm Klipfontein 30 km south-southwest of Fraserburg (Jubb and Gardiner, 1975), and *Elonichtys whaitsi* found on the farm Droogvoetsfontein near Fraserburg (Broom, 1918).

By using sandstone geometry and palaeocurrent measurements it was possible to distinguish two slightly different varieties of type 3 channel sandstones. These deposits are, however, closely related genetically and are not easily distinguished unless fully exposed. They had slightly different origins and mode of emplacement which account for the differences in their physical characteristics.

Type 3a channel sandstones are restricted to the upper part of the Combrinkskraal member, in the Leeu Gamka member, as well as in the Koup member in the central part of the study area. They occur as multistoreyed, lenticular sandstones in which upwards-fining sequences are present. Lateral accretion surfaces are only locally developed. Detailed palaeocurrent analyses at two

FIG. 6.5 *Abandoned channel-fill mudstones and sandstones 2,3 m thick (outlined in dashed lines), in a meandering channel deposit. Note asymmetric shape of channel fill.*

FIG. 6.6 *Cutbank levee deposit, grading from massive mudstones at the base into ripple cross-laminated sandstone at the top. Note load-casting near the top.*

different localities in the Combrinkskraal member revealed channel sinuosity ratios ranging between 1,39 and 1,41.

Type 3b channel sandstones occur in the uppermost part of the lower Beaufort member sequence, with the transition occurring at the base of the Nuweveld in the southwestern parts of the basin. The Koup channel sandstones in the Moordenaars Karoo (type 2 channel sandstones) change laterally into type 3b channel sandstones to the east and north of the type area. Evidence for lateral accretion is more conspicuous than in type 3a sandstones (cf. FIG. 6.6). In cross-sections the channel sandstones are seen to be replaced by new sandstone beds on slightly different elevations every few hundred metres along strike. Sandstone sequences are distinctly lenticular, laterally impersistent and upwards-fining.

By using the equation of Leeder (1973) to calculate bankfull channel widths, width:depth ratios of approximately 12 were calculated for these channels. The abandoned channel shown in FIG. 6.5 has a width:depth ratio of between 7 and 8. Calculations based on palaeocurrent measurements indicate a maximum sinuosity ratio of more than 2,0.

Interpretation

Type 3 sandstones are all typical meandering channel deposits as described by Allen (1965a), Leeder (1973), and many others. The type 3a sandstones are here classified as low-sinuosity meandering channel deposits. The type 3b sandstones are classified as high-sinuosity meandering channel deposits, also referred to as suspended load channel deposits by Schumm (1968). Although calculated channel sinuosities provide a reliable and practical method to distinguish between low and high-sinuosity channel deposits, other sedimentological parameters such as sandstone geometry and facies sequence are required in order to subdivide these deposits reliably.

6.3.2 Abandoned Channel-fill Deposits

Mudstone and silty sandstone lenses conforming to descriptions of abandoned channel fills (Allen, 1965a; Collinson, 1978) are sporadically present throughout the lower Beaufort sequence. Based on lithology and geometry several types of channel-fill deposits are related to two depositional processes, namely chute and neck cut-off.

Lenticular channel sandstones several metres thick and pinching out rapidly along strike, occur abundantly in the upper parts of the lower Beaufort. A fine example exposed in the Gifkop member on the farm Toorwater (location 30, FIG. 1.2) consists of a sandstone lens 7 m thick and less than 200 m wide. The sandstone consists predominantly of facies Sr and contains an abundance of disarticulated plant fossil impressions *(Phyllotheca?)* throughout the deposit. These sandstones all show evidence of gradual abandonment and are referred to as *partly abandoned channel-fill deposits.*

The second type of abandoned channel-fill sandstone resembles normal fluvial channel deposits but contains unusually high proportions of clay matrix, which impart a massive appearance and dark-grey to reddish colour to the rock. These sandstones are massive or ripple cross-laminated, 1 to 3 m thick, and commonly bioturbated on top. Intraformational conglomerates and soft-sediment deformation are largely absent.

A third type of abandoned channel fill is present in many of the type 3 channel sandstone deposits. The fill occurs as lenses of dark, carbonaceous siltstone commonly with interbedded sandstones (FIG. 6.5), and located in the top parts of channel sandstones. In the occurrence depicted in FIG. 6.5 the parameters of the channel fill closely resemble those calculated for the high-sinuosity meandering channels of the lower Beaufort. Calculations, using equation 1 of Leeder (1973) and substituting for the depth (2,3 m), yielded a channel width (24,5 m) close to the measured width of just over 20 m. The original meandering channel geometry was thus preserved during channel-fill deposition.

The sporadic presence of current-generated structures as well as lithological variations in these abandoned channel fills, suggest that channel abandonment was gradual, sand being introduced during sporadic floods, and mud deposition took place during periods of complete abandonment. According to Collinson (1978) and Allen (1965a) these channel fills are all characteristic of channel abandonment due to *chute cut-offs*.

Neck cut-offs are the result of rapid avulsion of meandering channels, followed by plugging of the abandoned channels with muds during normal overbank flooding (Allen, 1965a). In the lower Beaufort these deposits occur as incomplete channel sequences with sandstone at the base, grading upwards into massive, dark grey to olive grey, carbonaceous mudstones (facies Fg) up to several metres thick. Pedogenic calcareous nodules are in places found near the top of these channel fills. An example of this type of abandoned channel fill in the Koup member in the Verlatenkloof, consists of basal conglomerate 0,75 m thick, followed by 0,20 m of plane bedded sandstone grading abruptly into abandoned channel-fill mudstone.

Channel abandonment by chute or neck cut-off takes place when river gradients shoal to the extent that streams abandon their established channelways in order to shorten their courses and increase their gradients. Friedkin (1945) could demonstrate on the basis of experimental results and natural observations in the Mississippi River plains that chute cut-offs occur much more frequently than neck cut-offs. The latter occur very rarely, only when local obstructions by the way of migrating meanders result in different rates of cutbank erosion. In the process the upstream arm of a meander loop catches up with the downstream arm (Friedkin, 1945), and the channel diverts to the shortest and steepest route. Obstructions which caused neck cut-offs in the lower Beaufort were most likely lenses of more cohesive mudstones (e.g. existing abandoned channel-fills), interbedded with normal floodplain deposits. Mackin (1948) furthermore observed that shortening of meandering river courses, almost invariably by chute cut-off, is only a temporary arrangement and that soon afterwards the river extends its course to its previous length.

6.3.3 Natural Levee Deposits

Levee deposits are preserved very rarely as ripple cross-laminated sandstones and silty sandstones flanking meandering channel deposits. They occur as thin sandstone lenses, less than 1 m thick, dipping away from the major channel sandstone deposits. Climbing ripple cross-laminations are common, and in places the sandstones are bioturbated and mottled red (FIG. 4.13). A good example showing the geomorphic relationships of levee deposits relative to floodplain and channel deposits is exposed in the Teekloof Pass near Breakfast Neck. In this instance the levee sandstone (0,5 m thick) extends more than 100 m into the floodplain deposits before it pinches out. This type of levee deposit occurs on the outer banks of channels and is classified as a *cutbank levee deposit.*

Other examples of levee deposits consist of thin, ripple cross-laminated to massive, sandstone or siltstone sheets capping meandering channel sandstones. They rarely exceed 0,5 m in thickness, and are typically ripple marked on top which tends to obscure directional structures on the upper surfaces of channel sequences. These sandstone sheets are interpreted as *point bar (or convex bank) levee deposits*, as described in the Devonian sandstones of Spitsbergen by Moody-Stuart (1966), as well as in modern Mississippi River point bars by Fisk (1947).

Levee deposits represent the coarser material dumped on the banks of channels during flooding events. Levee construction usually results in the formation of alluvial ridges, and the local elevation of channelways relative to floodplains (Allen, 1965a). Fisk (1944; quoted in Allen, 1965a) reported levees more than 5 m high in the Mississippi River. Flood basin fills of up to 3,5 m thick on the farm Wortelfontein (FIG. 1.2) indicate the scale of lower Beaufort floodplain topography, resulting mostly from levee construction.

Cutbank levees have a very low preservation potential due to lateral migration of meandering channels (Collinson, 1978) and it can be assumed that most of the lower Beaufort cutbank levees were destroyed in this manner. Levees were in places stabilised by plant growth (FIG. 4.21) as was also noted in recent rivers in the United States (Leopold and Wolman, 1957). Many of the lower Beaufort examples of levee deposits show signs of severe bioturbation.

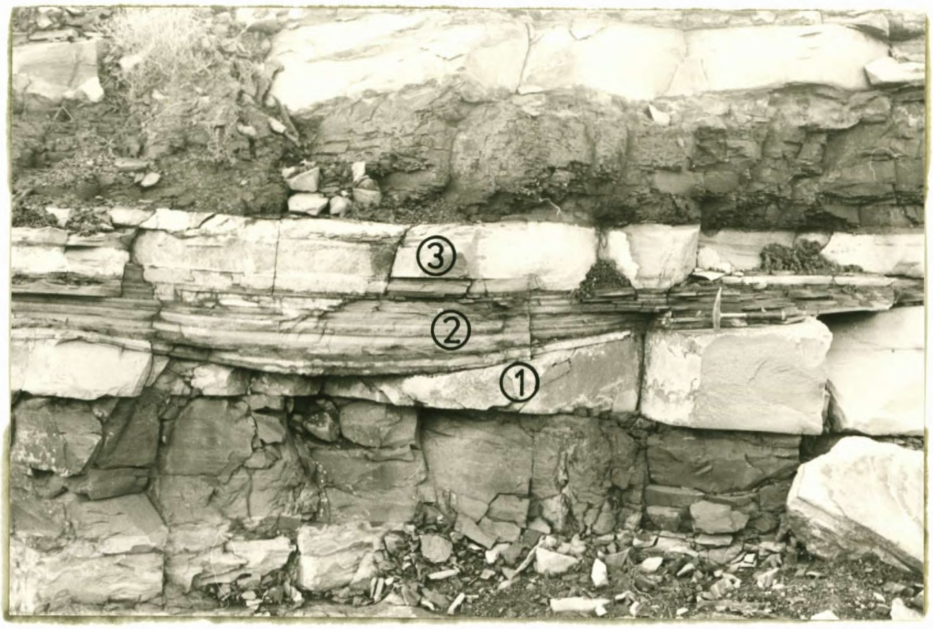


FIG. 6.7 Crevasse splay sheet sandstone showing multiple scouring. Three successive scouring and sedimentation events (numbered 1 to 3) are evident.

6.3.4 Crevasse Splay Deposits

The crevasse splay deposits can be assigned to three broadly defined classes, all of which show characteristics which are somewhat overlapping. They consist typically of sheets of fine-grained sandstone, generally less than 2 m thick, but which can extend up to several kilometres along strike. The different types of crevasse splay deposits are distinguished on the basis of contact and facies relationships, and their geometry.

The *crevasse splay sheet sandstones* (FIG. 4.16) are usually thin (0,5 m thick) although examples over 1 m thick are present in the Koup member south of Beaufort West. The sandstone sheets have low-relief erosional bases containing few scattered mudstone clasts but no conglomerates. In exceptional cases angular mudstone clasts are scattered throughout these sandstone sheets. Otherwise the sandstones are plane bedded at their bases grading upwards into ripple cross-laminated sandstones and siltstones, or they may be ripple cross-laminated throughout.

Planar cross-bedding is present, albeit very rare, in lower Beaufort examples resembling modern crevasse splay deposits described on the Clarence River floodplain in Australia by O'Brien and Wells (1986). The lower Beaufort examples are of similar scale (<0,5 m thick) and shape (lenticular) as those described by O'Brien and Wells. These deposits are well-exposed in the folded strata of the Combrinkskraal member along the southern margin of the Basin.

The sheet deposits formed where floods breached natural levees (Collinson, 1978), and bedload was washed onto dry floodplains. The sheet deposits mostly represent single flooding episodes but it is possible to identify several events within a single crevasse splay sheet sandstone (FIG. 6.7). Crevasse splay sandstones have never been found attached to the channel deposits from where they originated. Palaeocurrent measurements in one of these sandstones on the farm Toorwater showed a mean transport direction (126°) which deviates sharply from the mean channel transport direction (345°).

Crevasse channel sandstone deposits occur as erosively based and irregularly shaped sheets and lenses up to 1,5 m thick. Bases are frequently marked by a variety of flutes (FIG 4.7 and 4.8). Examples up to 0,5 m thick and less than 10 m wide occur in facies Fg siltstones. These deposits usually consist of single-storeyed, upwards-fining sandstones showing plane bedding, or rarely trough cross-bedding, at their bases and ripple cross-lamination on top. In places they show massive bedding due to bioturbation, dewatering and compaction. Crevasse channel deposits are distinguished from crevasse splay sheet deposits on account of the degree of scouring and the irregular shapes of the channel deposits.

The crevasse channels were formed during sporadic flooding (Collinson, 1978) and some may be the result of aborted channel avulsion (Bridge, 1984). The proximal examples occur in green mudstones as narrow deposits a few metres wide, and are never deeply scoured (<0,5 m). Examples in red mudstones reach thicknesses of more than 1,5 m. The thicker deposits are not easily distinguished from fluvial channel sandstones, as was noted also by Collinson (1978). The crevasse channel sandstones are considered the proximal equivalents of all the other crevasse splay deposits, although actual transitions were never observed in the field.

Microdelta deposits occur as lenticular upwards-coarsening sequences, up to 3 m thick as for instance in Ouberg Pass (location 6, FIG. 1.2). They consist of laminated green or red mudstones overlying sharp basal contacts, grading upwards into cross-laminated and wavy laminated siltstone and silty sandstone, followed by massive or ripple cross-laminated, fine-grained sandstone on top. Three thin, superimposed, upwards-coarsening sequences were recorded near the top of the Tuinkraal section east of Prince Albert Road. Load casting and ball-and-pillow structures occur in the sandy portions. In places the top contacts are wave ripple-marked, otherwise extensively bioturbated. Microdelta deposits occur throughout the lower Beaufort but are more abundant in the Leeu Gamka member, particularly along the southern margin of the Basin.

The upwards-coarsening sequences are interpreted as having been formed by crevasse splay and levee deposits prograding into submerged flood basins and other depressions (Miall, 1984b). Small-scale deltas similar to those described in the Brampton Esker by Saunderson and Jopling (1980), and in tidal and fluvial deposits by Dalrymple (1984), were formed in the process. The large-scale cross-bedding which is characteristic of the examples quoted, is absent in the lower Beaufort examples but the depositional processes were nevertheless similar. In the lower Beaufort examples the suspension deposits at the bases of successions were gradually replaced by poorly sorted sands on top, the latter deposited by weakly tractive currents. Examples of upwards-coarsening microdelta deposits in Verlatenkloof have been observed to grade laterally over a few tens of metres into upwards-fining crevasse splay sheet sandstones.

6.3.5 Overbank Deposits

The argillaceous overbank facies have been classified and described comprehensively in previous sections, and the depositional environments will be discussed briefly. The term *floodplain* refers in broad terms to the environment in which the mudstone successions, including the intercalated crevasse splay and levee sandstones, were deposited. Sedimentation on floodplains took place mainly as a result of vertical accretion from suspension during flooding events (Allen, 1964a). The presence and frequency of overbank sandstones as well as channel sandstones in the succession indicate the proximity of the floodplain environment to major channelways, and serve to distinguish *floodplain* and *flood basin* deposits. This terminology is used on stratigraphic profiles and throughout the text.

The overbank successions show vague upwards-fining trends. The meandering channel deposits grade upwards into green siltstone and mudstone with frequent sandstone intercalations, and eventually into red mudstones (where present). This tendency in floodplain deposits is ascribed mainly to gradual channel avulsion (Bridge, 1984).

6.3.5.1 Depositional Subenvironments

A number of depositional environments can be distinguished. The predominantly green mudstone successions (facies Fg) contain an abundance of reptilian and plant fossils and show much evidence of bioturbation. Organic activity indicates a hospitable environment in which plant and animal life abounded. A high rate of floodplain aggradation prevented the formation of mature palaeosols in this environment (Leeder, 1975b). The prevalence of overbank sandstone sheets and lenses indicates a setting in which an unusually large proportion (up to 40 per cent) of the succession represents levee and crevasse splay deposits. These regions were situated in close proximity to major channel belts (alluvial ridges), and are referred to as *proximal floodplains.*

Mudstone successions characterised by the abundance of red mudstone (facies Fr) and palaeosol, and the paucity of overbank sandstone and plant fossil remains, were deposited some distance away from major channel belts. Reptilian fossils are abundant as a result of the higher preservation potential of the remains of animals which perished in that environment. Suspension deposits are predominant in the succession and the rate of floodplain aggradation was low enough to permit soil formation in the alluvium. The environment is referred to as the *medial or distal floodplain.*

The *flood basin* deposits are characterised by thick successions of rhythmic argillaceous beds (facies Fv). Overbank sandstone intercalations are very rare and suspension deposits form thick (up to more than 100 m) uninterrupted successions. The considerable thickness and uniformity of these deposits show that conditions remained quiet and undisturbed over long periods of time.

Flood basins were situated in the most distal interfluvial regions of alluvial plains, and lack evidence of plant and animal life. Evidence of soil formation is absent despite inferred low rates of aggradation.

6.3.5.2 Pedogenic Processes

The formation of palaeosol was the most important alteration process affecting the lower Beaufort sediments immediately after deposition. Calcareous nodules in overbank mudstone were initially thought to form as a result of surface exposure only (Allen, 1965b). The nodules are now considered to be specifically of pedogenic origin, and that they were formed as a result of the downward migration of carbonates in the soil profile (Leeder, 1975a). Leeder (1975b) suggested that low floodplain accretion rates (in the order of less than 0,5 mm per year), as well as long periods of exposure, were required for soil formation. Retallack (1986) reviewed the interrelated factors controlling both fluvial sedimentation and soil formation. Bown and Kraus (1987) identified five stages of soil formation in alluvium, and proposed the term "pedofacies" to distinguish each different stage.

Pedogenic calcretes were first described in the lower Beaufort by McPherson and Germs (1979), who outlined several types of palaeosols in the succession. In the present investigation pedogenic calcretes were recorded throughout the lower Beaufort, but were rare in the uppermost mudstone members. They occur mainly as large, rounded nodules (including septarian nodules) in green mudstone (facies Fg), and calcareous laminae or smaller round nodules in red mudstone (facies Fr; cf. FIG. 4.4).

Mature palaeosols were rarely identified in weathered outcrops of the lower Beaufort, but fresh exposures (FIG. 6.8A), yielded spectacular examples. Several horizons characterised by pervasive bioturbation (FIG. 6.8A) are rooted in undisturbed, medium grey and dusky red mudstones (FIG. 6.8B). The upper surfaces are in places marked by thin calcareous layers (FIG. 6.8C), whereas the overlying mudstone contain large calcareous nodules. Reddish to brownish mudstone colours are characteristic. Faint, green rootlet traces were observed in associated mudstone at Riemhoogte

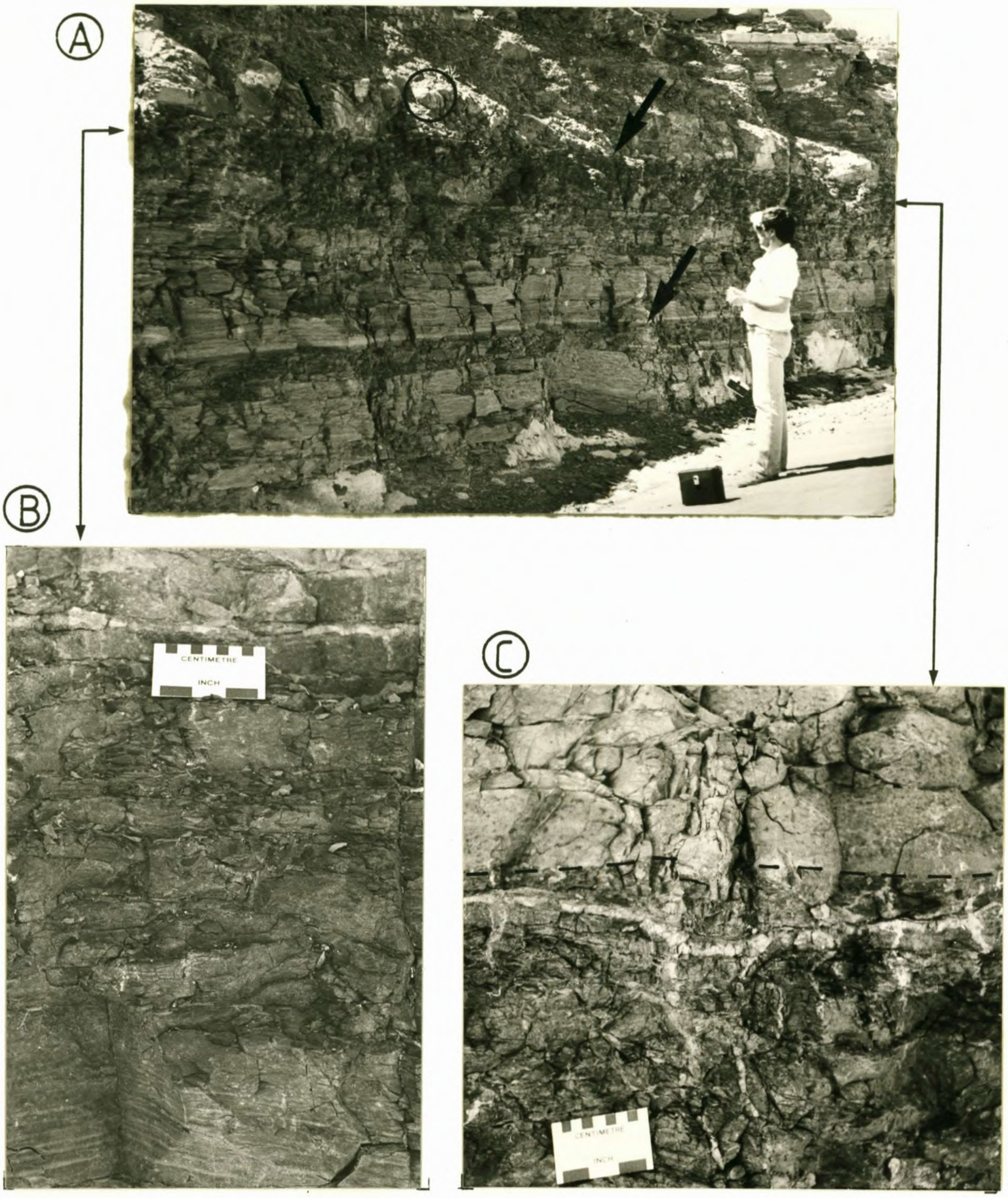


FIG. 6.8A Two palaeosols (large arrows) in overbank mudstone deposits. The upper unit grades upwards from laminated and cross-laminated siltstone into bioturbated siltstone, and is capped by a palaeocalcrete layer (small arrow) and calcareous nodules (circled). The unit is overlain by undisturbed mudstone.

FIG. 6.8B Close-up of upper palaeosol shown in FIG 6.8A. Note bioturbation in the top part of the palaeosol, grading downwards into laminated siltstone (near the bottom of the picture). Palaeocalcrete layer can be seen just above the scale.

FIG. 6.8C Another close-up of palaeosol and calcrete layer shown in FIG. 6.8A and 6.8B, showing sharp upper contact (dashed line) of palaeosol with overlying siltstone.

(location 25, FIG. 1.2), occurring approximately 50 m below the palaeosol layers depicted in FIG. 6.8.

Palaeosols are spaced at approximately 1 m intervals at Riemhoogte (FIG. 6.8A). Assuming a mean floodplain accretion rate of approximately 0,5 mm per year, it is estimated that in this case successive palaeosols were formed at intervals of some 200 years. Time spans of this order are much too short for extrabasinal controls such as climatic changes or tectonic activity (Retallack, 1986). Fluvial processes, such as periodic channel avulsions in particular, are considered to have been the most important controls in the formation of palaeosols in the lower Beaufort.

The presence of *incipient palaeosols* only in the proximal floodplain deposits, suggests that those areas were frequently flooded and never exposed for periods long enough to form mature palaeosols. Groundwater fluctuation played a large part in pedogenic processes in those areas, and formed hydromorphic soils as described by Buurman (1975).

The absence of mature palaeosols in the flood basin deposits can be addressed similarly. Buurman (1975) suggested that "biological ripening", which is the result mainly of disturbance by organisms and vegetation, was an important preparatory process in soil formation of this nature. There is no evidence of this process having affected the flood basin deposits but the variegated textures in these deposits are nevertheless the result of "chemical ripening", another important soil forming process (Buurman, 1975).

7.0 PALAEOCURRENT ANALYSIS

7.1 Introduction

The principal aim in the investigation of sediment dispersal patterns is to determine the direction of the palaeoslope in palaeogeographical studies (Potter and Pettijohn, 1977). Palaeocurrent measurements provided a direct method to determine palaeoflow directions locally, but aspects such as lithological composition and the geometry of the deposit provided additional insight into regional basin-fill patterns. In order to interpret palaeodrainage patterns the palaeocurrent models and depositional processes, as well as related statistical problems, had to be analysed first (Mader and Teyssen, 1985; Selley, 1968; Jizba, 1953).

Altogether 3325 palaeocurrent measurements were collected at 40 locations in the lower Beaufort, ranging in numbers of between a maximum of 266 readings at location 4 (Wilgerbosfontein) and a minimum of 25 readings at location 32 (Groot Tafelberg) (FIG. 1.2). An additional 26 measurements taken in the deltaic sandstones (distributary channel deposits) in the Verlatenkloof (location 5), as well as 18 measurements in the Katberg Formation sandstones on Ripplemead (location 35), were not included in the statistical processing of palaeocurrent data. The statistical parameters of all palaeocurrent readings are listed in TABLE 7.1.

7.1.1 Previous Work

The first regional palaeocurrent investigation of the Beaufort Group was carried out by Theron (1975). In this investigation the highest sampling density was achieved in the western Karoo Basin and although no stratigraphic control was applied in the sampling or processing of the data, most of the measurements in the area west of 24°E can be assumed to have been recorded in lower Beaufort strata. A general northerly drainage pattern was shown for the entire Beaufort Group (Theron, 1973, FIG. 6.1), with a westerly direction developed locally in the northeastern parts of the Basin. Theron (1970) showed the westerly (southwesterly) drainage direction to persist to the southwest as far as the Colesberg District.

Detailed palaeocurrent analyses by Kübler (1977) of selected sandstone units in the Koup member in an area west of Beaufort West, showed predominantly north-northeasterly drainage directions. By measuring palaeochannel trends Stear (1980b) found persistent easterly drainage directions in the area between Beaufort West and Merweville, including the area surveyed by Kübler (1977). Palaeocurrent analyses on a more limited scale (e.g. Le Roux, 1985; Smith, 1987; Stuart-Williams, 1981) yielded northeasterly directions in isolated areas in the central parts of the western Karoo Basin. Dukas (1978) mapped northwesterly trends in the Balfour Formation in the area adjoining the study area immediately to the east.

7.2 Methods

7.2.1 Sampling Procedures

Palaeocurrent measurements were collected primarily along stratigraphic sections. Measurements were made with a Brunton pocket transit. Detailed measurements were restricted, firstly, to palaeochannel sandstones and, secondly, to areas where the lower Beaufort succession was not excessively folded. In exceptional cases where dips exceeded 25°, the azimuths were restored by manual rotation on the outcrops, before they were measured. Stereographic rotation of azimuths using the method described by Potter and Pettijohn (1977) revealed a maximum correction of azimuths in the order of 3° for dips up to 25°. Drainage directions are expressed throughout in an opposite sense to meteorological convention (e.g. southerly meaning *towards* the south, and *not from* the south).

The number of measurements per site depended on the extent of the exposure and the type of sandstone (Potter and Pettijohn, 1977). A minimum of five to ten measurements from each type of sedimentary structure in each sandstone storey was usually found to be adequate, but in places high directional variability over a given area of surface outcrop required more measurements. Suitably exposed channel sandstones (e.g. the Kariega River area - location K, FIG. 1.2, 63 readings) in certain cases showed very low directional variabilities over areas of several square kilometers (TABLE 7.1). However, in multistoreyed meandering channel deposits where directional variabilities were high, the consistency of the measurements was always found to depend on whether the full succession, including all the composite sandstone storeys, was exposed and sampled.

In view of numerous inhomogeneities in fluvial drainage patterns, field measurements were carried out in sufficient detail to account for individual sedimentation (channeling) events. Directions measured on different types of sedimentary structures were recorded separately but these records were merged during final processing. Unweighted readings were then classified according to their different stratigraphic hierarchies (e.g. sandstone storey, megasequence, etc.) of origin, before statistical processing was carried out.

7.2.2 Ranking of Directional Structures

The suitability of various scales of directional features in fluvial deposits has been debated for many years (Allen, 1966), also in the case of the lower Beaufort (Stear, 1980b). Although channel sandstone orientations are the most reliable indication of drainage direction, they are suitably preserved in surface exposures in the western Karoo Basin only in exceptional cases (cf. Smith, 1987).

Allen (1966) pointed out that small-scale structures are usually superimposed on large-scale bedforms in fluvial channels, and that the directional variability increases as the variability of the lower order (large-scale) structures is inherited by the higher order (small-scale) structures. Fluvial bedforms were classified into four ranks or *orders of flow-vector fields* on the basis of their inherent variability, and their reliability as palaeocurrent indicators reviewed (Allen, 1966). The different preservation potentials of individual structures may also result in the elimination of some of the directional variability of the smaller scale structures as a result of the preferential preservation of only those structures aligned in a certain direction (Allen, 1967b), as for instance during major flooding events. These are some of the conditions which may have resulted in a degree of directional bias in the structures preserved in the lower Beaufort.

Miall (1974) proposed six ranks within the hierarchy of directional sedimentary structures of fluvial deposits, and indicated the expected variance (in degrees) of each rank. Miall at the same time suggested that large-scale directional structures were more reliable than small-scale structures, and that they should be weighted accordingly in calculations of mean directions.

Many other causes have been suggested for the directional variability of fluvial deposits but one which appears to be applicable to some lower Beaufort channel sandstone deposits, is the difference in variability between structures generated during high and low flow stages (Bluck, 1975/76). This aspect was not addressed directly by either Allen (1966) or Miall (1974) but is nevertheless related indirectly to the preservation potential of these structures.

The regional drainage patterns of fluvial deposits were summarised by Miall (1981). Altogether nine fundamental basin-fill patterns were identified based on the relationship of fluvial drainage directions relative to the axes of the basins, as well as the terminating environments of the rivers such as deltas, tidal flats or lakes.

7.2.3 Directional Structures in the Lower Beaufort

The sedimentary structures of lower Beaufort channel sandstones which are suitable for palaeocurrent measurements are fully reviewed elsewhere, and include in descending order of abundance: rib-and-furrow, current lineation, trough cross-bedding, flute cast, scour-and-fill including gutter casts, and clast imbrication. Of these, current lineation and scour-and-fill show orientations only, but not the direction of flow. Flute and gutter casts are very rare in the area due to poor exposure of channel sandstone bases, and they were of very little practical use in the present investigation.

Rib-and-furrows are present everywhere on the upper contacts of channel sandstone storeys except where these had been truncated by succeeding channels, or where these surfaces were covered by hard-weathering levee deposits (e.g. Nuweveld member in the Oukloof section and in Teekloof, FIG. 1.2). Due to their abundance and good exposures rib-and-furrows emerged the most important directional structures in the present investigation. No more than ten readings were recorded on rib-and-furrows on each outcrop studied, except where a significant deviation was evident along a limited strike length. Directional variance within channel sandstone storeys is low, with standard deviations as low as 11,3° (ten readings, location 30, FIG.7.1) being typical of high-sinuosity meandering channel deposits in the Nuweveld member.

Trough cross-bedding was measured only where the trough axis was fully exposed. Attempts to determine palaeocurrent directions in partially exposed troughs by means of stereographic plotting of individual foreset beds, as described by DeCelles *et al.* (1983), gave erratic results. This is partly due to the inherent directional variability of these structures in fluvial deposits (cf. Friend, 1965).

Potter and Pettijohn (1977, pp. 139-140) quoted several references in which large-scale cross-bedding corresponds very well to the general drainage directions in a number of environments. Allen (1966) and Miall (1974) suggested that lower directional variabilities could be expected in large-scale (rank 5) structures such as trough cross-bedding, than in small-scale (rank 6) structures. Although few measurements of trough axes were possible on each outcrop the directional variabilities in trough cross-bedding in the lower Beaufort sandstones are almost invariably high compared to the other structures. Ten measurements of well exposed trough axes in one specific sandstone storey in the Koup member on the farm Tafelberg (location 12, FIG. 1.2) yielded a spread of directions of 180° (18°-198°; standard deviation S = 50,5°) compared to a spread of 24° (284°-308°; S = 11,1°) measured on five rib-and-furrow measurements over the same area. The mean palaeocurrent directions of these two sets of readings are nevertheless comparable, being 320° and 299°, respectively. Friend (1965) noted a similar broad scatter of directions in trough cross-bedding in the fluvial deposits of the Devonian of Spitsbergen, measured over a short distance, and compared with directions measured on sole structures. It appears that in meandering channel deposits the directions measured on large-scale trough cross-bedding are not as consistent as the small-scale structures, and in this case there was no need to weight these readings according to their scale or rank.

Parting lineations are abundant in lower Beaufort sandstones and were measured routinely together with the other directional structures. They show a spread in orientation similar to that of rib-and-furrows but they do not indicate the direction of flow, and they were therefore of limited use in statistical analyses. By applying Student's t-test to compare the mean transport directions (Dixon and Massey, 1957; Till, 1974), it was determined that in the same sandstone storeys the mean directions for rib-and-furrows and parting lineations do not differ significantly at the 99 per cent confidence level. The orientations of parting lineations were thus of considerable value in confirming the directions as determined in other structures, particularly in multistoreyed meandering channel deposits.

Scour-and-fills are subject to the same limitations as parting lineations in palaeocurrent determinations they also show orientations rather than flow directions. They are rarely well exposed in channel sandstones, but the flow directions could be interpreted by comparison with other structures (FIG. 4.6).

Although *flute casts* are common in crevasse splay sheet sandstones, they were very rarely recorded in channel sandstones. They could be used only once in palaeocurrent determinations, namely in the basal group of sandstones of the Combrinckskraal member on the Laingsburg Commonage. Palaeocurrent measurements of flute casts show an east-southeasterly transport direction (128°), thus confirming the direction (132°) determined from three rib-and-furrows in an adjacent channel sandstone.

Clast imbrications were measured at only one locality, namely at the top of a 19,4 m thick channel sandstone in the Combrinkskraal member in the Verlatenkloof (location 5, FIG. 7.1). The mean direction as measured on four cobble-sized mudstone clasts was 182° (spread 170°-200°), compared with a mean of 177° (153°-212°) as measured on rib-and-furrows. Five elongate clasts, their long axes oriented horizontally, have their long axes at right angles to the directions listed before, namely a mean orientation of 86°/266° (255°-281°).

7.3 Statistical Processing of Directional Data

Palaeocurrent directions obtained by means of unidirectional structures are represented mathematically as vectors in three-dimensional space. The horizontal components of these vectors are represented by the palaeocurrent azimuths which show a circular normal distribution. The statistical methods used to determine parameters of linear distributions are not applicable to circular data in their standard format. Statistical analyses have shown, however, that the Von Mises distribution applies particularly well to unimodal palaeocurrent data (Mader and Teyssen, 1985; Mardia, 1972). Jones (1968) questioned the theoretical background and uniformity of the wrapped normal (Curray, 1956) and truncated normal distribution (Pincus, 1956), and pointed out the advantages of using the parameters associated with the circular distribution rather than those of linear distributions when dealing with orientation data.

The parameters used by Mardia (1972) and earlier workers to describe the circular normal (Von Mises) distribution were based on angular deviations expressed in radians rather than degrees. The converted equations (to degrees) used here were taken from Potter and Pettijohn (1977), Till (1974), and Karlstrom *et al.* (1983). The parameters of palaeocurrent distributions (TABLE 7.1) were calculated with the aid of a Hewlett Packard 41CV programmable calculator and a peripheral printer.

The mean azimuths of ungrouped data were calculated using the vector summation method of Reiche (1938), and the resultant directions referred to as the *vector means* (cf. Potter and Pettijohn, 1977). The mean co-ordinates of *n* angular measurements are calculated as follows:

$$V = \Sigma \cos \varnothing / n$$

and $$W = \Sigma \sin \varnothing / n,$$

where $\varnothing$ is the azimuth of each of *n* observations, in degrees. The resultant vector mean (X_V) was calculated from the equation:

$$X_V = \arctan W/V.$$

By assigning the computed angle to a quadrant based on the polarity of V and W, this angle could be expressed as an azimuth.

The *vector magnitude* R of the sum of n vectors is expressed as:

$$R = [V^2 + W^2]^{0,5}$$

and normalised as the *consistency ratio* L, as follows:

$$L = (R/n)\ 100$$

The consistency ratio (expressed as a percentage) is a measure of the dispersion of azimuths about the mean of the sample, with L = 100 per cent indicating no dispersion. This parameter is directly related to the standard deviation of linear distributions, as well as to the concentration parameter *kappa* of the circular normal (Von Mises) distribution. The consistency ratio can easily be displayed as the thickness or the lenght of an arrow on maps to show the clustering of values in a certain direction (cf. Karlstrom *et al.*, 1983). The *confidence limits* (95 per cent confidence level in this instance) about the estimated mean appear to give a clearer indication of dispersion, and these values are shown in TABLE 7.1. The two-tailed t-distribution was used to calculate the confidence limits for this interval, as follows:

$$X_V \pm [t_{\alpha/2}\, S_X]/[n^{0,5}]$$

where t_α is the value for a certain level of significance α of the t-distribution (Dixon and Massey, 1957). The *standard deviation* (S_X) used in the equation was calculated in degrees as:

$$S_X = 180/\pi\ [(-2 \log_e L/100)^{0,5}],$$

as used in Mardia (1972) and Karlstrom *et al.* (1983).

In order to determine whether the distribution of values differs significantly from the uniform distribution of an equal number of values, the *Rayleigh test of significance* (P) as used by Curray (1956) was applied to ungrouped sets of palaeocurrent readings. The parameter P calculated as follows:

$$P = e^{-nL \times 0,0001}$$

indicates that the distribution, consisting of n readings, is non-random at the 99 per cent confidence level for values of $P<0,01$.

The palaeocurrent measurements from each recording station are represented graphically as current rose diagrams in FIG. 7.1, and summarised as synopsis palaeocurrent rose diagrams for each megasequence in FIG. 7.2.

TABLE 7.1 Summary of palaeocurrent statistical data. Abbreviations and symbols as used in text; also explained below.

Location (No.)	Unit	n	X_v(°)	L(%)	S(°)	P	95% Conf. Lim.(°)
Karoluspoort (1)	B	42	114	88,8	27,9	-6	105-123
Combrinkskraal (2)	A	10	26	78,1	40,3	-3	357-055
	B	21	85	93,3	21,3	-8	75-95
Prince Albert Road (3)	A	38	57	54,8	62,8	-5	36-78
	B	7	66	91,2	24,6	-3	43-89
Wilgerbosfontein (4)	A	54	31	73,6	44,9	-13	19-43
	B	155	79	56,0	61,7	-22	69-89
	C	57	67	40,0	77,6	-4	46-88
Verlatenkloof (5)	E	26	80	68,5	49,8	-6	58-102
	A	77	109	51,2	66,3	-9	94-124
	B	38	127	64,0	54,1	-7	109-145
Ouberg Pass (6)	A	66	106	70,1	48,3	-15	94-118
Kareekasberg (7)	A	56	39	69,7	48,7	-12	26-52
Helpmekaar (8)	B	105	27	63,9	54,4	-19	16-38
	C	18	50	86,1	31,3	-6	34-66
Bastersberg (9)	A	6	78	96,0	16,4	-3	61-95
	B	33	352	23,0	98,2	-0,7	317-027
Palmietfontein (10)	B	47	345	40,4	77,1	-4	322-008
	C	51	1	77,1	41,3	-14	349-013
Droogvoetsfontein (11)	B	117	8	34,5	83,6	-7	353-023
	C	38	348	71,8	46,6	-9	333-003
Tafelberg (12)	B	128	11	55,5	62,2	-18	1-22
	C	89	88	54,5	63,1	-12	75-101
Oversfontein (13)	B	28	79	83,5	34,4	-9	66-92
	C	63	111	19,5	103,6	-11	85-137

TABLE 7.1 (Continued)

Location (No.)	**Unit**	**n**	**X_V(°)**	**L(%)**	**S(°)**	**P**	**95% Conf. Lim.(°)**
Oversfontein (13)	D	6	134	25,5	94,7	-0,2	35-233
Karelskraal Pass -	B	26	354	89,8	26,6	-10	5-343
Gifkop (14 - 15)	C	86	25	60,7	57,3	-14	13-37
	D	20	171	12,7	116,4	-0,2	117-225
Wilgerboskloof (16)	B	128	65	62,0	56,0	-22	55-75
	C	47	94	49,3	68,1	-5	74-114
Bontberg - Langberg	B	61	90	56,5	61,2	-9	74-106
(17 - 18)	C	106	76	41,2	76,3	-8	61-91
	D	26	93	81,8	36,3	-8	78-108
Leeukop (19)	C	46	42	68,7	49,6	-10	27-57
	D	36	56	47,0	70,4	-4	32-80
Oukloof (20)	B	64	26	27,5	92,1	-3	3-49
	C	22	138	25,0	95,4	-0,5	96-180
Layton (21)	B	44	76	38,4	79,3	-3	52-100
	C	37	46	68,9	49,5	-8	29-63
Karoo National Park (22)	B	88	65	64,2	53,9	-16	54-76
	C	69	87	56,8	60,9	-10	72-102
Highlands (23)	C	30	77	45,2	72,2	-3	50-104
	D	12	59	85,4	32,2	-4	39-79
De Jager's Pass (24)	B	36	103	16,5	108,8	-0,4	67-140
Riemhoogte (25)	B	48	53	55,0	62,7	-7	35-71
	C	30	37	82,2	35,9	-9	24-50
Booiskraal (26)	C	63	55	63,2	54,9	-11	41-69
	D	24	71	53,9	63,7	-4	44-98
Bloemfonteinkop (27)	C	62	38	17,9	106,3	-1	11-65
Three Sisters (28)	C	42	32	87,3	29,9	-14	23-41

TABLE 7.1 (Continued)

Location (No.)	**Unit**	**n**	**X_V(°)**	**L(%)**	**S(°)**	**P**	**95% Conf. Lim.(°)**
Mordant Klaasens-kraal (29)	C	50	36	52,5	65,0	-6	18-54
Toorwater (30)	C	119	335	5,5	138,0	-0,2	310-360
	D	35	172	75,6	42,9	-9	157-187
Matjiesfontein (31)	D	20	82	80,3	38,0	-6	64-100
	K	24	32	74,2	44,3	-6	12-52
Groot Tafelberg (32)	C	14	15	85,5	31,7	-5	357-033
	D	11	13	84,1	33,7	-4	350-036
Wortelfontein (33)	C	51	236	17,8	106,5	-0,7	206-266
	D	4	175	99,6	5,1	-2	167-183
Klein Tafelberg (34)	C	39	264	24,9	95,5	-1	233-295
	D	15	355	96,5	15,3	-7	347-003
Ripplemead (35)	D	18	355	93,1	21,7	-7	344-006
	K	18	339	92,0	23,4	-7	327-351
Krugerskraal (36)	C	80	338	26,7	93,1	-3	317-359
	D	13	27	62,1	55,9	-3	353-061
Vrede (37)		(No measurements)					
Uitkyk (38)	B	63	37	72,4	46,0	-15	25-49
Kariega River area (K)	A	63	64	86,4	31,0	-21	56-72
Spreeuwfontein (W)	A	77	344	41,0	76,5	-6	347-017
Rietfontein (R)	A	47	23	64,8	53,4	-9	7-39

TABLE 7.1 (Continued)

Explanation:

Location (No.) - sampling locality number (FIG. 1.2 and 7.1)

Unit - stratigraphic unit; A, B, C and D refer to lower Beaufort megasequences, E refers to Waterford/Koedoesberg Formation (Ecca Group), and K refers to Katberg Formation

n - number of measurements

X_V - vector mean (degrees)

L - consistency ratio (percentage)

S - standard deviation (degrees)

P - Rayleigh test of significance (10^x)

95% Conf. Lim.- 95 per cent confidence limits (degrees)

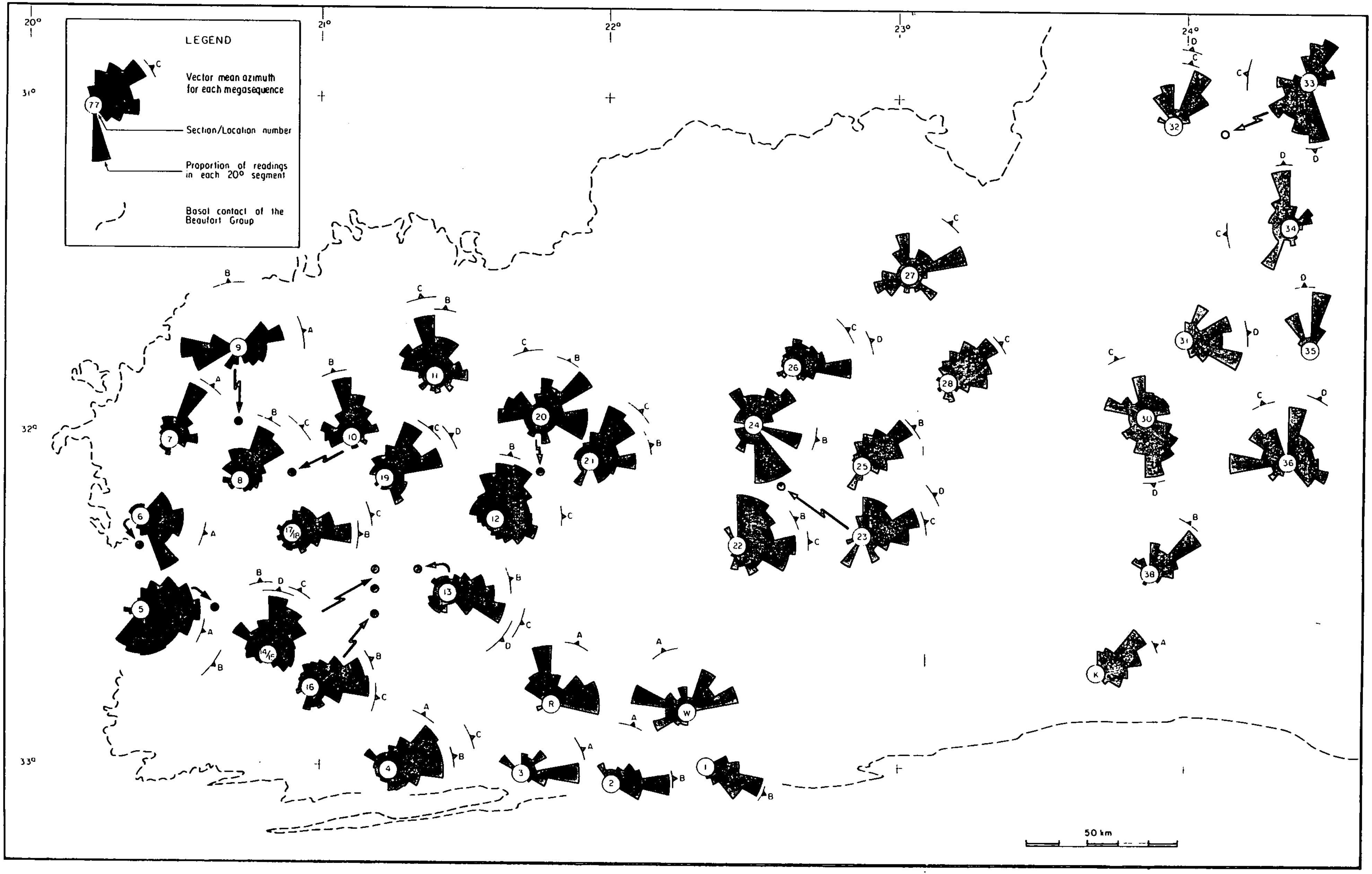


FIG. 7.1 Rose diagrams of palaeocurrent measurements in the lower Beaufort. Statistical parameters are summarised in TABLE 7.1.

7.4 Palaeocurrent Interpretation

The present interpretation is based on the rigid classification of palaeocurrent measurements according to the stratigraphic units from which they originated. Field measurements were collected throughout the various stratigraphic units, rather than from the arenaceous intervals in these successions only. After statistical processing of field measurements, palaeocurrent trends were interpreted but no further manipulation of results was undertaken. Where feasible the palaeocurrent trends were compared with lithological associations and other scalar features of the different stratigraphic intervals.

7.4.1 Megasequence A

The oldest strata in this unit, exposed at the base of the succession in the area south of 32° 30'S, show prominent northerly drainage directions. These directions could not be measured reliably or tested statistically, due to intense folding and overfolding of the strata over most of the area.

Comprehensive sampling of the the upper, gently folded sandstones in the succession was carried out at three localities shown on FIG. 7.1, namely on the farms Rietfontein (location R), Spreeuw Fontein (location W), and the Kariega River area (location K). A distinct polymodal distribution of palaeocurrent readings with up to three discrete subpopulations was evident in multistoreyed sandstones at the former two localities. Calculated vector means show northerly transport directions on Rietfontein and Spreeuw Fontein. A second, north-northeasterly to northeasterly trend evident in the upper sandstones of the succession at localities 2, 3 and 4 (FIG. 7.1) is less pronounced, but becomes more persistent towards the eastern boundary of the study area (location K).

The succession is incompletely exposed in the western and northwestern parts of the Basin, to the east of which it is overlain by strata of the younger megasequence B. Palaeocurrent directions change gradually from northeasterly to southeasterly over an area stretching from north (location 7) to south (location 5) across the Basin (FIG. 7.1). This southward change in palaeocurrent directions is furthermore confirmed by isolated readings of between 128° and 132° in the basal

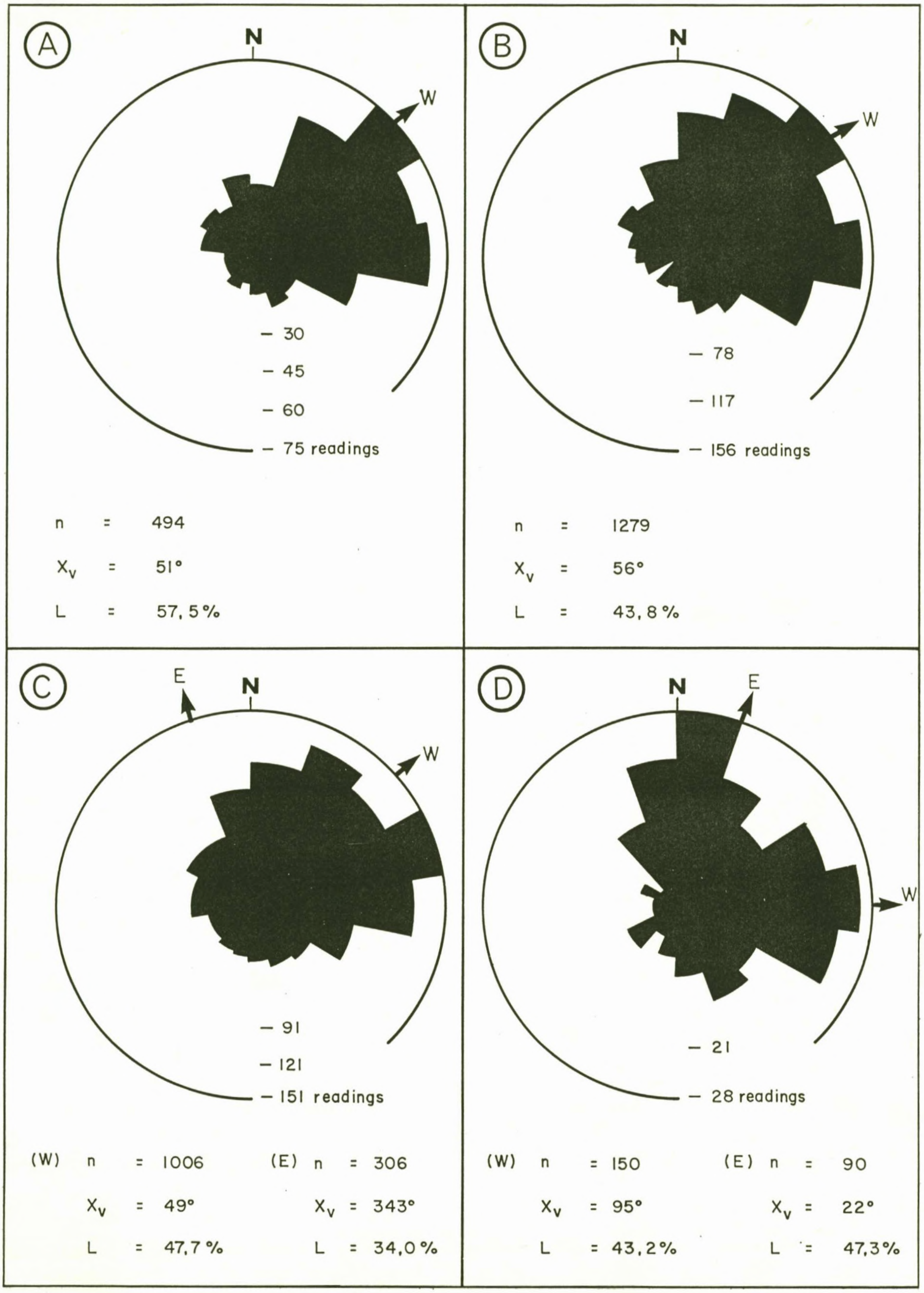


FIG. 7.2 Synopsis palaeocurrent rose diagrams (in 20° segments) of all measurements in megasequences A, B, C and D. Grand vector means for the western (W) and eastern sectors (E) shown as arrows on outer circles.

sandstones of the lower Beaufort (megasequence A) on the Laingsburg Commonage, as discussed previously. (The latter sandsones are equivalent to the uppermost interval of megasequence A farther to the east, namely at locations1, 2 and 3, on FIG. 1.2. The lower part of the Combrinkskraal member - at the base of megasequence A - is absent in the Laingsburg area.)

At least three major drainage systems can be identified (FIG. 7.3A). An older system flowed from the south to the north or northeast. A younger, radial (fan-shaped) drainage system debouched from the western margins and followed a trend roughly along the axis of the Basin. The third, northeasterly drainage pattern in the upper parts of the succession in the south appears to reflect the regional pattern of "run-off" drainage prevailing at the closing stages of the geomorphic cycle. All these patterns are repeated in the succeeding lower Beaufort megasequences.

7.4.2 Megasequence B

The northerly drainage system debouching from the south is very poorly developed and its presence could be established at only two localities, namely in the Karelskraal Pass - Gifkop area (locations 14/15, FIG. 7.1), and in the Tafelberg section (location 12, FIG. 7.1).

The radial drainage pattern in the western and northwestern parts of the Basin is more pronounced than in megasequence A (FIG. 7.3B). Drainage directions on the northern lobe of this system show a swing towards the north in the northwestern corner of the Basin. Individual sandstone storeys in certain channel sandstones in this area (location 9, FIG. 7.1) revealed mean transport directions (234°) comparable to the south-southeasterly directions shown in parts of this area by Theron (1973, Fig. 6.1). However, a representative palaeocurrent trend can only be determined after all the composite sandstone storeys in these channel sandstones have been sampled.

In the central parts of the Basin the easterly and northerly drainage systems appear to have become interdigitated, and as a result the radial (easterly) trends became somewhat obscured. The northeasterly trends in the southern, central and eastern parts of the study area are also more prominently developed than before. The directions measured along the southern margin of the Basin appear to curve progressively towards the south (locations 1, 2 and 3, FIG. 7.1), in a pattern

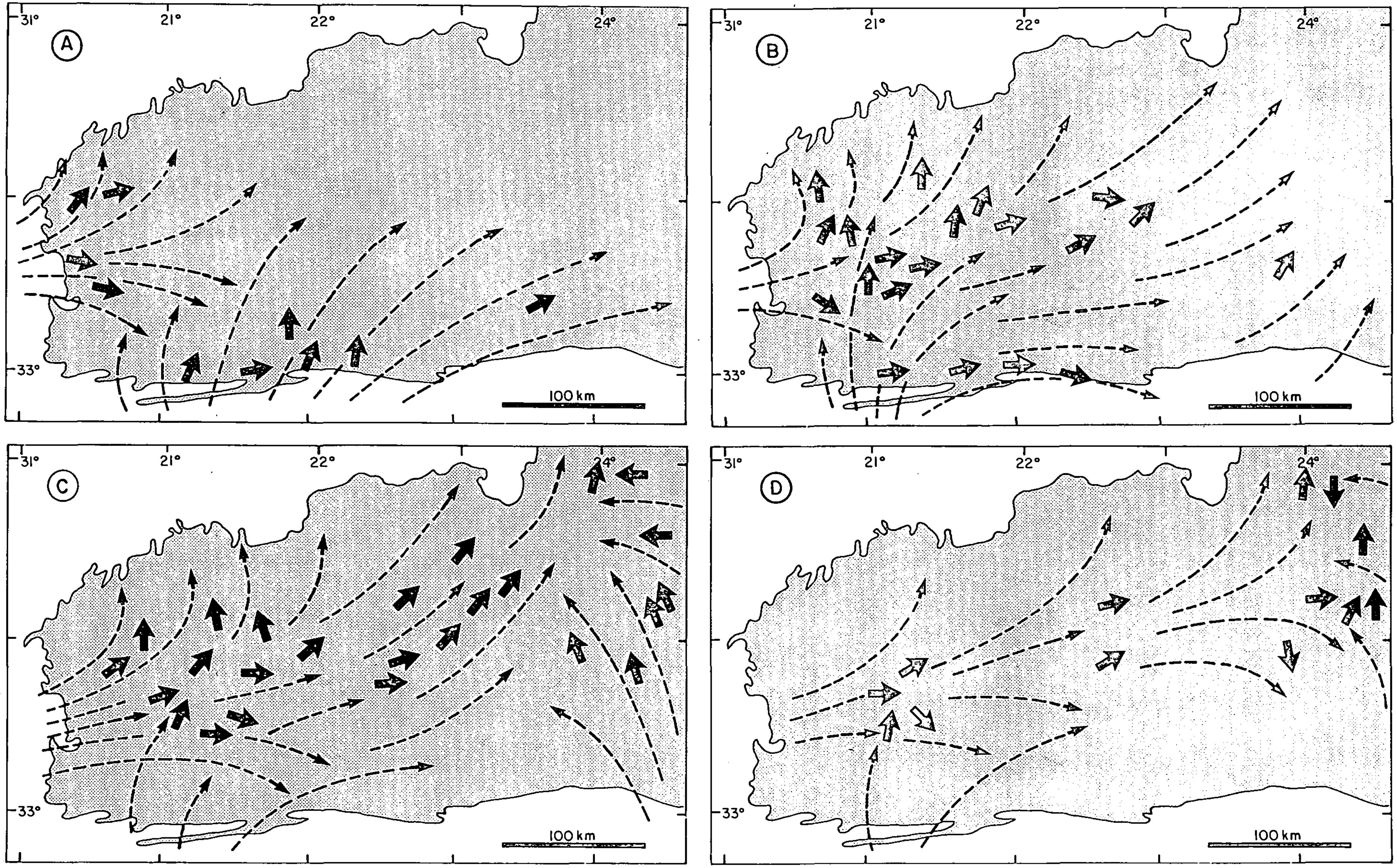


FIG. 7.3 Interpretation of regional drainage patterns in megasequences A, B, C and D. Vector means shown as solid arrows, and inferred drainage directions as broken arrows.

similar to those in the northern parts of the Basin curving towards the north. However, too few measurements are available to confirm this tendency beyond doubt.

7.4.3 Megasequence C

The radial drainage pattern seen in megasequences A and B is still evident in the westernmost parts (FIG. 7.3C), with the northerly deflection of directions in the northwestern parts of the drainage system still prominent. This direction was confirmed in the vicinity of the farm Eselsfontein, southwest of Fraserburg (FIG. 1.2), by drilling results which showed major channel sandstones near the base of the Nuweveld member to trend nearly due north. The northerly drainage system discharging from the southern margins was very weakly developed and present only in the Karelskraal Pass - Gifkop section (locations 14 /15, FIG. 7.1).

The east-northeasterly drainage direction seen throughout the western sector of the Basin became prominent in megasequence C. In the northeastern corner of the Basin this (east-northeasterly) drainage system was deflected slightly northward, towards the northeast (locations 26, 27 and 28, FIG. 7.1). Drainage in the northwestern corner of the Basin was deflected locally north -and westward, towards the north-northwest (locations 11 and 20, FIG. 7.1). From these directions collectively it is clear that drainage in the areas situated some distance to the north and south of the Basin axis, as defined by the longitudinal drainage direction, was deflected substantially from the direction (east-northeast) of the dip of the palaeoslope (FIG. 7.3C).

A new drainage system flowing towards the north-northwest is apparent in megasequence C in the southeastern parts of the study area (locations 30 and 36, FIG. 7.1). This system appears to be unrelated to the Oudeberg Member, which occurs just beyond the southeastern margins of the study area and which has a similar (north-northeasterly) drainage direction (Dukas, 1978). (The Oudeberg Member pinches out in the Kamdebo Range in the area situated approximately between locations 36 and 38, FIG. 7.1.) The westerly drainage direction measured at locations 33 and 34 (FIG. 7.1) is of uncertain origin, but runs parallel to that of the Oudeberg Member and associated sandstones which occur in the area to the east of the study area (Dukas, 1978).

7.4 Megasequence D

The northerly drainage system was evident only in the Karelskraal Pass - Gifkop section (locations 14 /15, FIG. 7.1). Elsewhere in the western sector of the Basin a prominent east-northeasterly direction is apparent. The tendency of palaeocurrent directions to deviate from the general palaeocurrent trend along the axial parts of the Basin is also evident in megasequence D (FIG. 7.3D). Conflicting palaeocurrent directions along the eastern margins of the study area may be as a result of (1) the presence of high-sinuousity meandering channel deposits, and (2) the interfingering of the fluvial deposits transported from the southwest and the southeast.

7.5 Palaeodrainage Model

The palaeocurrent directions in the western sector of the Basin reflect two principal fluvial drainage systems in a pattern which conforms to model 7 of Miall (1981, TABLE 1). The first system consisted of transverse alluvial plains which entered the western sector of the Basin from the south. These alluvial plains formed the distal environments of a transverse alluvial fan complex, flanking an elevated source area postulated to the south of the Basin. The second drainage system consisted of longitudinal alluvial plains (trunk rivers) flowing from the west to the northeast and east-northeast. The fluvial systems of the lower parts of the succession all terminated in deltaic depositional systems (Visser *et al.*, 1980). The palaeocurrent directions in the longitudinal channel deposits are almost parallel to those measured in the distributary channels (X_V=80°) in the Verlatenkloof (location 5, FIG. 7.1) at the base of megasequence A on the western margin of the Basin. The longitudinal drainage directions crudely define the axis of the Basin during deposition of the lower Beaufort (FIG. 7.3), and is commonly referred to as "axial" or "trunk" stream drainage (cf. Allen, 1965b; Link, 1984).

The radial palaeocurrent pattern seen in the western part of the Basin is the result of drainage away from a prominent alluvial ridge constructed in those proximal areas, or the run-off on the distal reaches of an alluvial fan system. This, together with the deviation of palaeocurrent directions away from the mean flow direction along the axis of the Basin, reflect a tendency by fluvial systems to break through the alluvial ridges and drain into lower lying areas adjacent to the alluvial ridges.

The variability of current directions in the different units is illustrated by the 95 per cent confidence limits about the vector means (TABLE. 7.1). Where sufficient readings are available the value distributions of stratigraphic units higher up in the succession (megasequences C and D) show confidence limits generally wider than those lower down in the succession (megasequences A and B). This tendency is the result of higher channel sinuosity and a greater scatter of palaeocurrent directions higher up in the succession.

The vector means of megasequences A, B and C are fairly consistent (FIG. 7.2), showing northeasterly trends in the western sector of the Basin. The differences in mean palaeocurrent directions are mainly the result of the geographical distribution of the areas in which the different units are preserved or exposed relative to the axis of the Basin. Most of the palaeocurrent readings in megasequence D, for instance, were collected in a linear zone close to the axis of the Basin, the area in which the uppermost parts of the succession are best preserved.

Although palaeocurrent readings from the western and eastern sectors were treated separately in calculations of grand vector means (FIG. 7.2), it is possible that some contamination of data could have taken place in the course of field measurements. The value distributions for each megasequence furthermore included measurements from both the transverse and the longitudinal drainage systems, which rendered them unsuitable for statistical purposes. Only in megasequence D were the palaeocurrent readings taken exclusively in channel deposits of the longitudinal drainage systems. Palaeocurrent statistical results are also biased to some extent towards trends prevalent in geographical areas which were more densely sampled. The grand vector means calculated for each of the intervals (FIG. 7.2) are therefore not a true indication of the trend of the depositional axis of the Basin. These are also collectively the reasons for the large deviation (49° to 95°) between grand vector means calculated for the various units in the western sector of the Basin.

The westerly and northwesterly drainage directions calculated for megasequences C in the far eastern sector of the Basin represent at least two separate drainage systems in that part of the succession (FIG. 7.3C). The first system deposited the Oudeberg Member in the southeasternmost parts of the study area, whereas a distinctly arenaceous lobe of the Nuweveld Member was emplaced at the same time by north-northwesterly draining streams in the vicinity of Murraysburg (location 30, FIG. 7.1). The co-existence of several drainage systems in such a limited area suggests that drainage patterns were affected by syndepositional tectonic activity in the Basin.

Only one drainage system is distinguished in megasequence D in the eastern sector of the Basin. The palaeocurrent trends are highly variable, due mainly to the variability in meandering channel directions and to fewer measurements taken at each sampling point. A westerly to northwesterly direction is predominant (FIG. 7.3D). A southward direction at location 33 (FIG. 7.1) is in conflict with the regional trend in the eastern sector, and is probably the result of sampling bias due to too few measurements at that locality. The high degree of variability in palaeocurrent directions in the area where the western and eastern sectors meet (FIG. 7.3C and 7.3D) is an indication of the complex interfingering over short distances of different depositional systems within megasequences C and D.

8.0 BASIN ANALYSIS

8.1 Introduction

The sediment dispersal patterns and palaeogeography of the Karoo Basin were to a large extent controlled by tectonism. The tectonic setting was also studied in order to assess the economic potential of the Basin. As the syndepositional structure of the Basin is usually obscured by post-depositional structural overprinting and erosion, sedimentological techniques were used where possible to reconstruct the geometry and palaeogeography of the Karoo Basin during lower Beaufort (Late Permian) times.

In preceding sections the depositional history of the Basin was reconstructed directly from field-based sedimentological observations. Sediment dispersal patterns, for instance, were deduced from measurements of directional features in channel sandstones. In the following sections certain scalar properties of the succession are included in a palaeogeographic reconstruction of the Karoo Basin during the lower Beaufort period.

8.2 Sandstone - to - Mudstone Ratio Map

8.2.1 Method

The ratio of channel sandstones to overbank deposits was calculated for each stratigraphic profile, and the values contoured for the entire western Karoo Basin. The primary objectives were to identify on contour facies maps regional sediment dispersal patterns in the Basin, and thus the direction of the palaeoslope. The technique is based on the assumption that the proportion of sandstone decreases in the distal parts of fluvial drainage systems.

Conybeare (1979) recommended that for stratigraphic intervals exceeding approximately 100 m in thickness, cumulative sandstone thickness rather than lithological ratio should be used for contouring. This approach is valid only in the case where the complete stratigraphic succession is present at each sampling point. In the present investigation the section localities are widely spaced and individual stratigraphic units (megasequences) are incomplete in many of the sections. As a result few lithological ratios are available for complete stratigraphic units throughout the Basin. The sandstone:mudstone ratio was thus calculated for the entire lower Beaufort succession traversed at each sampling locality, and the values contoured for the entire western Karoo Basin. Values were also calculated for individual stratigraphic units where a reasonable thickness is preserved and exposed. The following discussion is based on the regional contour map for the western Karoo Basin (FIG. 8.1). Sandstone:mudstone ratio values are listed in TABLE 8.1.

8.2.2 Discussion

The sandstone:mudstone ratio trends appear to conflict with drainage patterns indicated by palaeocurrent data (FIG. 7.3). Based on the assumption that the proportion of bedload sediments (sand and gravel) would have decreased in the distal reaches of the lower Beaufort drainage systems, the sandstone:mudstone ratio was expected to decrease in the northern part of the Basin. However, the considerable increase in this ratio in the northwestern corner of the study area seems to suggest the opposite, and a separate entry point could have been interpreted somewhere along the northern margin of the Basin (FIG. 8.1). Several arguments are presented here against such an interpretation. The use of lithological ratios in incomplete successions (such as in the present investigation), and without due consideration of the Basin architecture, is also shown to be incorrect.

An important aspect which is not apparent from this exercise is the decrease in the *total sandstone* content of the succession from south to north in the Basin. The lower Beaufort succession has a cumulative sandstone thickness of 560 m in the vicinity of Laingsburg (location 4), 217 m on Wilgerboskloof (location 16), and 100 m in the vicinity of Fraserburg (location 11). The sandstone content of individual stratigraphic units (megasequences) shows similar decreases from south to north. The Koup member has a cumulative sandstone thickness of 278 m and sandstone:mudstone ratio of 0,33 near Laingsburg in the south (location 4), compared with a cumulative sandstone thickness of 69 m and a sandstone:mudstone ratio of 0,34 near Fraserburg in the north (location 11). The sandstone:mudstone ratio thus increases, and the total sandstone content decreases considerably as the succession thins out in a northerly direction across the Basin.

TABLE 8.1 Sandstone:mudstone ratio values calculated for megasequences A, B, C and D in the western Karoo Basin. Ratio of incomplete units shown in brackets.

Profile (No.)	**Sandstone:mudstone Ratio**				
	Total Lower Beaufort	**Megasequence**			
		A	**B**	**C**	**D**
Karoluspoort (1)	0,25	0,22	-	-	-
Combrinkskraal (2)	0,22	0,21	-	-	-
Tuinkraal (3)	0,33	0,32	-	-	-
Wilgerbosfontein (4)	0,37	0,22	0,48	(0,73)	-
Verlatenkloof (5)	0,35	0,28	-	-	-
Ouberg Pass (6)	0,25	0,25	-	-	-
Kareekasberg (7)	0,35	0,33	-	-	-
Helpmekaar (8)	0,55	0,55	-	-	-
Bastersberg (9)	0,47	(0,12)	(1,54)	-	-
Palmietfontein (10)	0,55	-	(0,35)	(0,72)	-
Droogvoetsfontein (11)	0,59	0,51	1,66	-	-
Tafelberg (12)	0,37	-	(0,30)	0,42	-
Oversfontein (13)	0,39	-	(0,28)	0,41	(0,64)
Karelskraal Pass/ Gifkop (14/15)	0,41	-	(0,20)	0,48	(0,71)
Wilgerboskloof (16)	0,39	-	(0,29)	(1,20)	-
Puntkraal/Langberg (17/18)	0,48	-	(0,40)	0,42	(0,95)
Leeukop (19)	0,40	-	-	(0,24)	(0,80)
Oukloof (20)	0,35	-	-	(0,42)	-
Layton (21)	0,39	-	(0,27)	(1,06)	-
Karoo National Park (22)	0,24	-	(0,29)	0,19	-
Highlands (23)	0,27	-	-	(0,21)	(0,94)
De Jager's Pass (24)	0,34	-	(0,34)	-	-
Riemhoogte (25)	0,28	-	(0,22)	(0,36)	-
Booiskraal (26)	0,37	-	-	(0,27)	(0,91)

TABLE 8.1 (Continued)

Profile (No.)	Sandstone:mudstone Ratio				
	Total Lower Beaufort	Megasequence			
		A	B	C	D
Bloemfonteinkop (27)	0,21	-	-	(0,21)	-
Three Sisters/Mordant Klaasenskraal (28/29)	0,24	-	-	(0,22)	-
Toorwater (30)	0,23	-	-	(0,26)	(0,18)
Matjiesfontein (31)	0,09	-	-	-	0,09
Groot Tafelberg (32)	0,34	-	-	(0,21)	(0,31)
Wortelfontein (33)	0,31	-	-	(0,33)	-
Klein Tafelberg (34)	0,21	-	-	(0,14)	(0,26)
Ripplemead (35)	0,20	-	-	-	0,20
Krugerskraal (36)	0,29	-	-	(0,25)	-
Vrede (37)	-	-	-	-	-
Uitkyk (38)	0,43	-	(0,43)	-	-

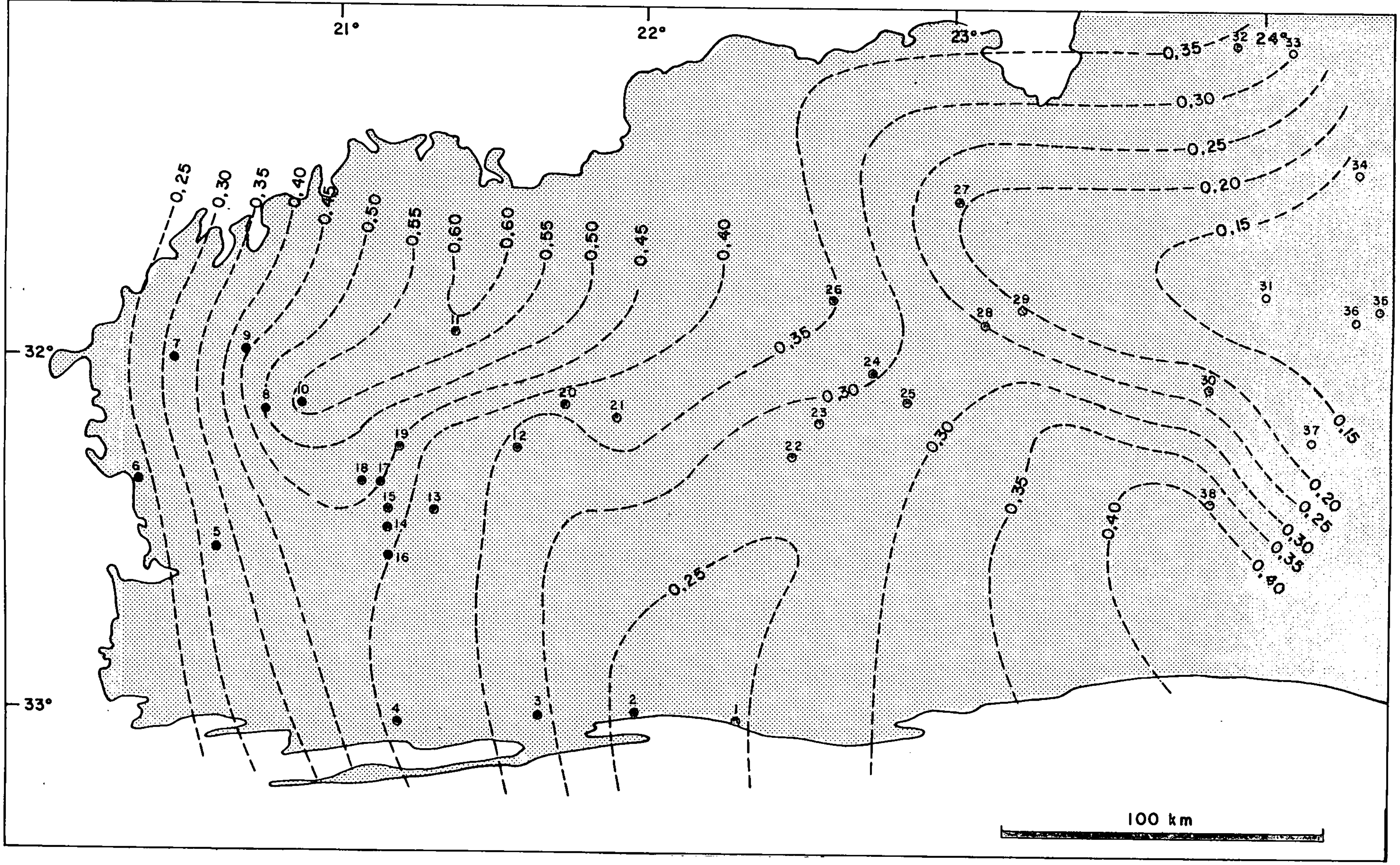


FIG. 8.1 Sandstone - to - mudstone ratio contour map of the total lower Beaufort Group (shaded) in the western Karoo Basin. Contour spacing is 0,05 units. (Section locations shown as closed circles; see TABLE 8.1 for individual values.)

A second aspect is the contribution and concentration of sandstone by longitudinal fluvial systems draining from west to east in the northern part of the Basin. The sections along the southern margin of the Basin (locations 1, 2 and 3), characterised by low a sandstone:mudstone ratio, consist mostly of strata (megasequence A) deposited by a transverse drainage system which entered the Basin from the south. A third drainage system in which rivers flowed towards the northeast (FIG. 7.3) resulted in an increase in the sandstone content in the southeastern corner of the study area (location 38). In this case it is clear that lithological ratios of strata (1) derived from several different geographic regions of the source area, and (2) deposited by more than one drainage system and (3) during different tectonic pulses, have some limitations in the reconstruction of regional drainage patterns. Isolith maps showing cumulative thickness of a single rock type would be of greater value in investigations of this nature, on the condition that a reasonable distribution of complete intervals of each stratigraphic unit in the succession were available.

Very few data points are available in the southern part of the study area, and as a result entry points could not be identified accurately. An increase in sandstone in the Wilgerbosfontein section (location 4) is considered to be the result of sand concentration both from an entry point along the southern margin of the Basin, and the contribution of sand by longitudinal drainage systems in the Koup and Nuweveld members. The increase in sandstone in the Kamdebo region north of Aberdeen (location 38) is similarly due to sand which was introduced by a new drainage system which entered the Basin from the south-southeast during deposition of megasequence C (cf. FIG. 7.3C).

The decrease in sandstone:mudstone ratio in the easternmost part of the area surrounding section locations 31, 34, 35, 36 and 37), is due mainly to the paucity of channel sandstones in the distal deposits of the Nuweveld, Oudeberg, Leeukop and Barberskrans members. The Oudeberg Member on the farm Vrede (location 37, FIG. 1.2) contains only 13,5 m of sandstone, whereas the stratigraphically equivalent Nuweveld member on the farm Toorwater (location 30, FIG. 1.2), situated only 40 km away, has a cumulative sandstone thickness of 80 m.

The interpretation of regional drainage patterns based on palaeocurrent measurements rather than lithological ratios, is preferred on the basis of the stratigraphic control that could be applied in the course of the palaeocurrent investigation. The increase in sandstone:mudstone ratio in a northerly direction is ascribed to the irregular sampling (stratigraphic) intervals, as well as contributions of sand by longitudinal drainage systems in the northern half of the study area.

8.3 Isopach Maps

Isopach contour maps were compiled for megasequence A (Combrinckskraal and Gamka members; FIG 8.2) and megasequence B (Koup and Teekloof members; FIG. 8.3). Stratigraphic control was by means of surface stratigraphic sections (Chapter 3), as well as to a limited extent, drilling results. Although stratigraphic contacts were not as clear in core as in surface sections, estimates based on core drilled on the farms Vrede (well number VR1, location 37, FIG. 1.2) and Sambokkraal (well number SA1, FIG1.2) were highly significant. As a result of present-day erosion of the upper two megasequences (C and D) too few complete sections were available, and spread over too small an area of the Basin for meaningful contouring. The contouring was done factually with very little interpretation, and on the scale of the present exercise no palinspastic adjustment seemed necessary for the area along the southern, folded margin of the Basin. Thickness values used in contour maps are summarised in TABLE 8.2.

Isopach maps indicate the geometry of a stratigraphic unit, reflecting variations in syndepositional thickness (Conybeare, 1979). Several factors may affect these thickness values after deposition and which may result in the distortion of isopach contouring trends. These may include unconformities within the succession, differential compaction, and diachronous contacts. The latter was the only serious problem in the present exercise as both megasequences A and B were found to grade diachronously into the Ecca Group within the limits of the study area. Winter (1984) questioned the use of the present Ecca-Beaufort contact and suggested that a chronostratigraphic datum be used instead. No such datum is known in that part of the succession (Jordaan, 1981 and 1985), and the present exercise was carried out within those constraints.

TABLE 8.2 Summary of stratigraphic thicknesses of megasequences A, B, C and D in the western Karoo Basin. Thicknesses of incomplete successions shown in round brackets, and calculated thicknesses in square brackets.

Profile (No.)	Stratigraphic Thickness				
	Total Lower Beaufort (m)	Megasequence			
		A (m)	B (m)	C (m)	D (m)
Karoluspoort (1)	2523	2334	(170)	-	-
Combrinkskraal (2)	2582	2554	(28)	-	-
Tuinkraal (3)	2055	(1865)	(82)	-	-
Wilgerbosfontein (4)	2472	996	842	(237)	-
Verlatenkloof (5)	1541	1012	(220)	-	-
Ouberg Pass (6)	950	922	(50)	-	-
Kareekasberg (7)	414	(407)	(7)	-	-
Helpmekaar (8)	344	(24)	(285)	-	-
Bastersberg (9)	321	(185)	(136)	-	-
Palmietfontein (10)	355	-	(137)	(218)	-
Droogvoetsfontein (11)	270	(17)	202	(51)	-
Tafelberg (12)	778	-	(234)	347	(197)
Oversfontein (13)	641	-	(221)	338	82
Karelskraal Pass/ Gifkop (14/15)	615	-	(220)	345	55
Wilgerboskloof (16)	774	-	(642)	(132)	-
Puntkraal (17)	374	-	(146)	(228)	-
Langberg (18)	295	-	-	(196)	(99)
Leeukop (19)	257	-	-	(163)	(94)
Oukloof (20)	320	-	(47)	(273)	-
Layton (21)	382	-	(297)	(85)	-
Karoo National Park (22)	893	-	(406)	360	(127)
Highlands (23)	368	-	-	(316)	(52)

TABLE 8.2 (Continued)

Profile (No.)	**Stratigraphic Thickness**				
	Total Lower Beaufort (m)	**Megasequence**			
		A (m)	**B (m)**	**C (m)**	**D (m)**
De Jager's Pass (24)	181	-	(181)	-	-
Riemhoogte (25)	395	-	(225)	(170)	-
Booiskraal (26)	331	-	-	(259)	(52)
Bloemfonteinkop (27)	261	-	-	(261)	-
Three Sisters (28)	264	-	(19)	(245)	-
Mordant Klaasenskraal (29)	297	-	-	(297)	-
Toorwater (30)	621	-	(10)	(391)	(220)
Matjiesfontein (31)	298	-	-	(35)	156
Groot Tafelberg (32)	344	-	-	(168)	(176)
Wortelfontein (33)	187	-	-	(153)	(34)
Klein Tafelberg (34)	306	-	-	(198)	(108)
Ripplemead (35)	208	-	-	(20)	166
Krugerskraal (36)	400	-	-	(383)	(17)
Vrede (37)	975	[500]	[460]	(14)	-
Uitkyk (38)	474	-	(474)	-	-

8.3.1 Megasequence A

Maximum sediment accumulation took place in a deep trough oriented parallel (east-west) to the Cape Fold Belt (FIG. 8.2). The depocentre was situated very close to the southern margin of the Basin (locations 1, 2, and 3, FIG. 8.2). Isopach contours and the presence of Beaufort strata in two tightly folded synclines south of the main body of Beaufort strata indicate that the contemporary southern margin of the Karoo Basin could have been situated to the south of the present margin.

The succession thins rapidly northwards from approximately 2500 m just north of Prince Albert, to less than 1000 m in the Moordenaars Karoo. No stratigraphic control exists between the Wilgerbosfontein section (location 4) and Verlatenkloof (location 5), where the succession thickens again to just more than 1000 m. In this area the stratigraphic thickening of the lower Beaufort is more clearly reflected in the underlying deltaic strata. The delta plain deposits, for instance, thicken from a few metres or tens of metres elsewhere in the western Karoo Basin to more than 100 m in the Verlatenkloof (Jordaan, 1981).

From Verlatenkloof the succession thins out gradually towards the north, eventually grading into the Ecca Group south of Carnavon. Megasequence A subcrops beneath the overlying succession (megasequence B) in a postulated east-southeasterly direction, south of Carnavon (FIG. 8.2).

Isopach contours along the western margin of the Basin are open, suggesting that the margin is erosional and that the Basin once existed farther to the west. Contour trends furthermore suggest a widening of the Basin to the west, and it appears to have been somewhat deeper in the southwestern corner. No information exists at present to determine the geometry of the postulated western extent of the Beaufort basin but from the present distribution of Dwyka and Ecca strata along the western margin, it seems that the Karoo Basin once persisted some distance to the west of the present limits of the Beaufort succession. An east-west trending basin, pear-shaped in plan with the widest end situated in the west, appears to have existed earlier during lower Beaufort sedimentation.

Isopach contouring in the southwestern part of the study area is poorly constrained, but stratigraphic control in the Vrede (location 37) and Uitkyk (location 38) sections shows a considerable decrease in thickness along strike from west to east across the Basin.

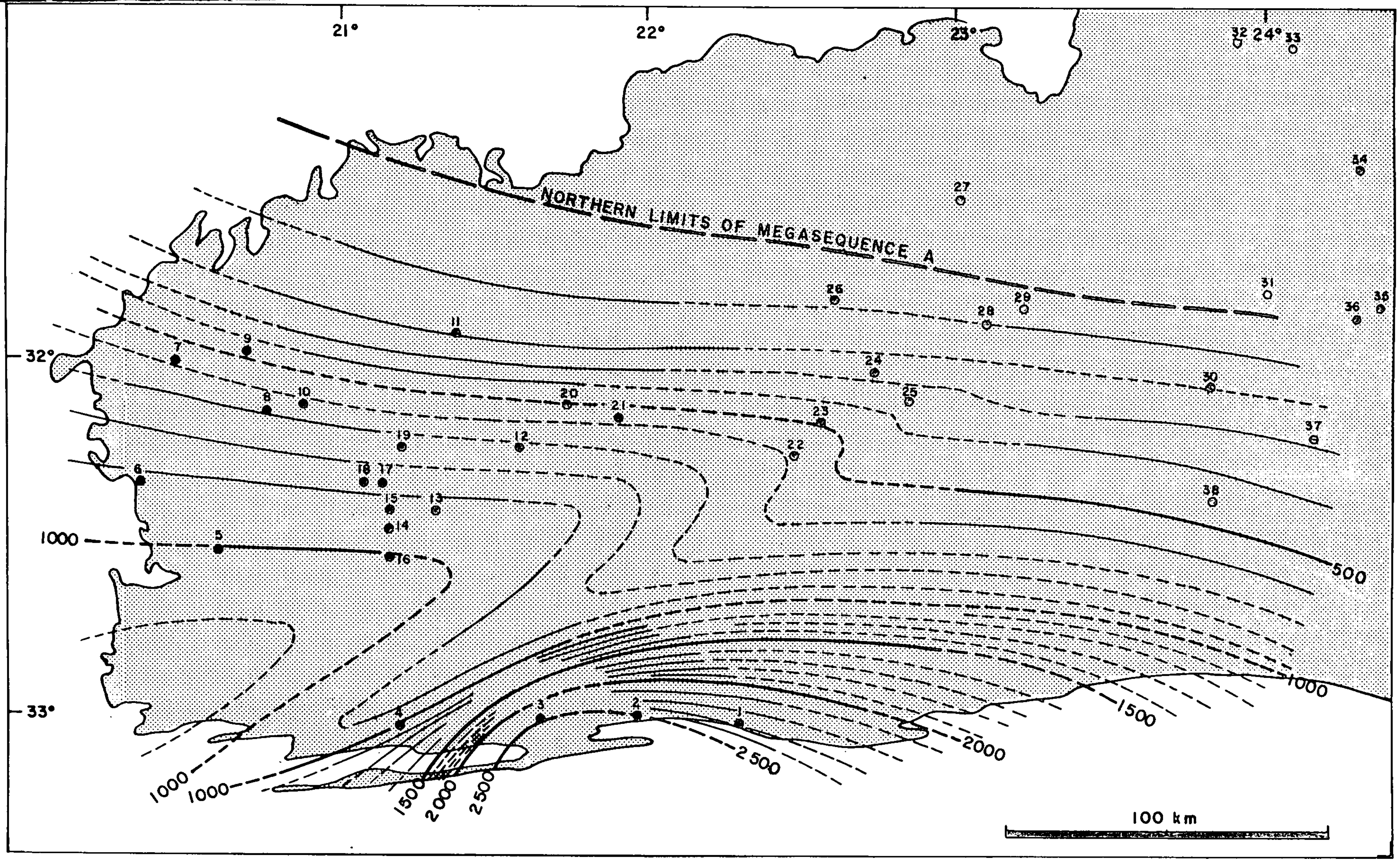


FIG. 8.2 Isopach contour map of megasequence A of the lower Beaufort Group (shaded) in the western Karoo Basin. Contour spacing is 100m. Locations (shown as closed circles) and stratigraphic thicknesses are listed in TABLE 8.2.

8.3.2 Megasequence B

The isopach map for megasequence B (FIG. 8.3) shows features similar to megasequence A, albeit apparently simplified due to fewer control points (TABLE 8.2). A maximum thickness of 861 m was recorded on the farm Wilgerbosfontein (location 4, FIG. 1.2). The complete succession is nowhere fully exposed in the central part of the Basin but stratigraphic thickness in well SA1 (Winter and Venter, 1970, TABLE 1) and the Wilgerboskloof section (location 16) suggest a thickness of some 650 m to 750 m for megasequence B in the area just west of Merweville. The unit gradually thins northwards to 285 m on the farms Helpmekaar (location 8) and Palmietfontein (location 10), reaching 202 m on Droogvoetsfontein (location 11).

Stratigraphic control in the Uitkyk section (location 38) and Vrede well (VR1, location 37) suggests a thickness in the order of 550 to 600 m for megasequence B in the southeastern part of the study area. The succession clearly does not show the marked decrease in thickness along strike from west to east seen in megasequence A. In the eastern part of the area megasequence B appears to thin gradually towards the north grading into the Ecca Group north of Richmond. The succession is overlain by sandstones of the Nuweveld member to the north and northeast of the farm Wortelfontein (location 33).

The contours along the western margin of the Basin appear to be open still, although lack of control in the southwestern corner of the Basin has hampered the investigation. Elsewhere in the Basin the contour lines appear to reflect the present geometry of the Basin more closely.

8.3.3 Discussion

The structure of the Basin during the early stages of lower Beaufort sedimentation is clearly reflected in isopach maps of megasequences A and B. The effects of folding are visible as gentle syndepositional warping inside the present margins of the Basin. Several elements typical of an orogenic foreland basin are identified on the basis of a reconstruction of the syndepositional structure of the Basin.

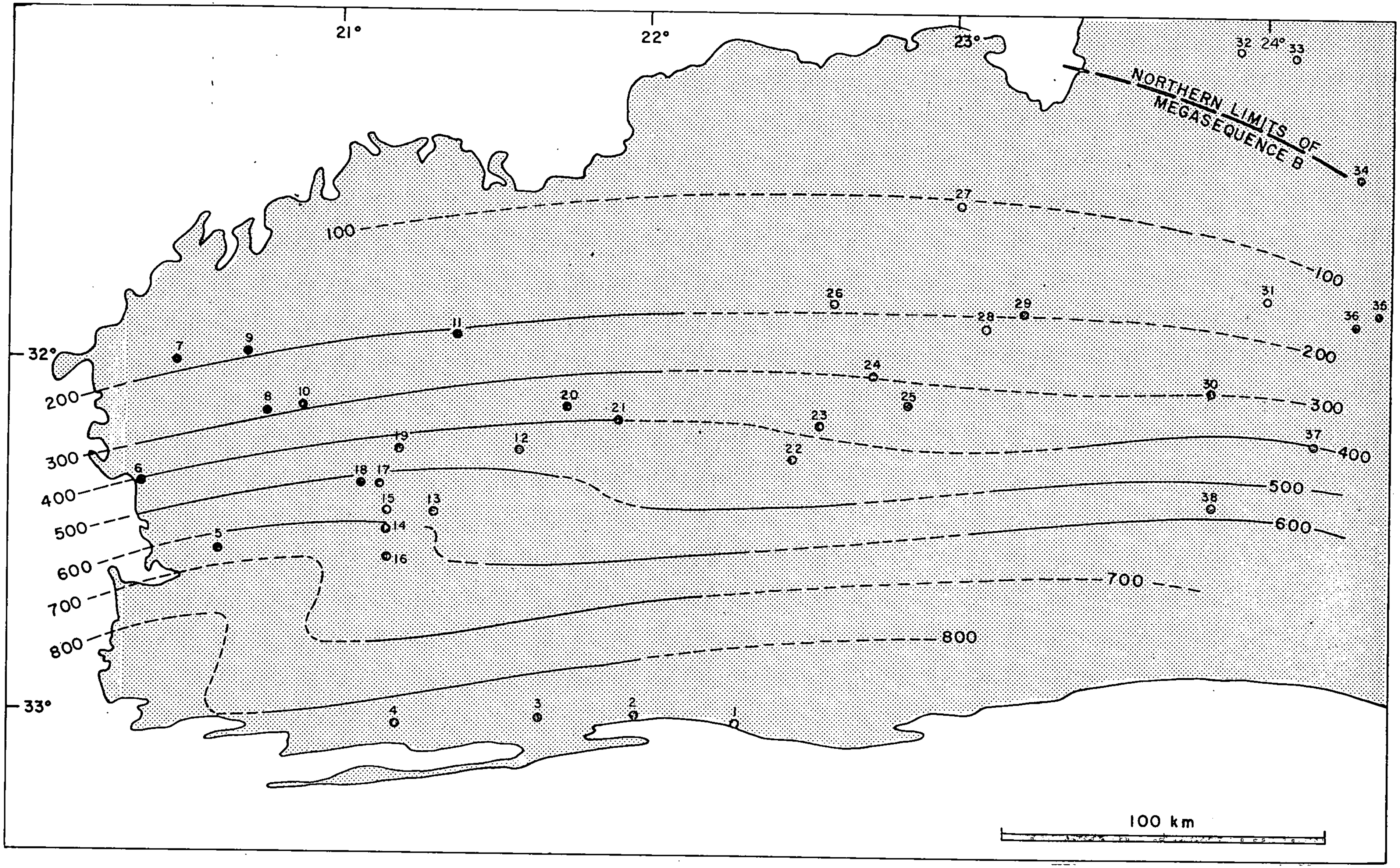


FIG. 8.3 Isopach contour map of megasequence B of the lower Beaufort Group (shaded) in the western Karoo Basin. Contour spacing is 100 m. Locations (closed circles) and stratigraphic thicknesses are listed in TABLE 8.2.

The Basin was asymmetric during the early stages of lower Beaufort sedimentation. Although not evident on isopach maps the depocentre is postulated to have been situated to the south of the present southern margin of the Basin. Isopach contours in the southern and central parts of the Basin were deflected by flexuring along an east-northeasterly axis which becomes less prominent in the eastern part of the study area. A shallow neogenic basin occupied the region between this anticlinal ridge and the northern margin of the Basin, during deposition of megasequence A. Although it was much less pronounced warping is inferred to have persisted at least until deposition of megasequence B.

The severe attenuation of megasequence A in the southeast may be due to a syndepositional, transverse fold structure colloquially known as the "Willowmore arch". The structure was referred to as the "Willowmore Transversal" by Winter and Venter (1970), and it was suggested to have affected Karoo sedimentation in the area situated roughly between Klaarstroom and Pearston since the Late Carboniferous (Dwyka Formation). In megasequence B there is no evidence of stratigraphic attenuation on the scale seen in megasequence A, and it can be assumed that the structure was not active, or fairly subdued, during that period. During deposition of megasequence C renewed upwarping of the Willowmore arch appears to have controlled sedimentation in the southeastern part of the study area, and sediment was shed into river systems draining north-northwestward.

The thickness determined for megasequence C is constant in the central part of the study area, varying between 338 m (Oversfontein section, location 13) and 347 m (Tafelberg section, location 12). A thickness of 391 m in the Toorwater section (location 31), and a thickness of approximately 465 m determined by Dukas (1978) in equivalent strata on the farm Ferndale in the Graaff-Reinet District, show a deepening of the Basin towards the east during deposition of megasequence C. From there megasequence C appears to thin to no more than 200 m at the northern margin of the Basin north of Richmond (Groot Tafelberg and Wortelfontein sections, locations 32 and 33, resp.).

An accurate thickness could not be determined for megasequence D anywhere in the western part of the Basin but a maximum thickness of 197 m was determined for an incomplete succession on Tafelberg (location 12) south of Fraserburg. In the eastern part of the study area a total thickness of 166 m was measured for the complete succession on the farm Ripplemead (location 35). A slightly greater thickness of 176 m was determined for an incomplete succession on Groot Tafelberg (location 32), the northernmost location where a significant thickness of megasequence D is preserved.

Based on these isolated measurements it appears as if the geometry of the western Karoo Basin became more symmetric in cross-section during deposition of the upper part of the lower Beaufort succession. This is supported by the observation by Visser (1978) that the shape of the Karoo Basin as a whole changed from asymmetric to a more symmetric shape during the Early Permian to Late Triassic period. The depocentre of the Basin shifted eastward during deposition of the lower Beaufort and attenuation of the succession over the Willowmore arch also became less noticeable, at least for the period during which megasequence C was deposited.

8.4 Tectonic Setting of the Late Permian Karoo Basin

8.4.1 Introduction

The mechanisms which applied to the formation of the main Karoo Basin within the context of Gondwana, were first pointed out by Du Toit (1937). Du Toit (1948) later expanded on his earlier views on the structural controls of the Basin, and proposed the formation of a "foreland" adjacent to the "foredeep" of the "Cape geosyncline". Newton (1973) could not identify an orogenic style of deformation in the Cape folding and suggested that the Karoo Basin formed as a result of gravitational subsidence. Newton at the same time proposed that the shape of the Basin was controlled by the structure of pre-Cape basement. Hobday (1978) suggested that the Beaufort Group was deposited as a molasse complex in a foreland basin.

8.4.2 Classification of Sedimentary Basins

Prior to the recognition of the roll of plate tectonics in the formation of depositional basins all sedimentary basins were classified according to the geosynclinal theory (Kay, 1948, quoted in Keiln, 1987). Dewey and Bird (1970) drew attention to the relationship between the different types of mountain belts and their plate tectonic settings, thus paving the way to a genetic classification of the (foreland) basins flanking these mountains. Dickinson (1974) made the first attempt to classify sedimentary basins in terms of their plate tectonic settings. Altogether 15 types of sedimentary basins were distinguished on the basis of the types of lithosphere, types of plate margins, and proximity of basins to plate margins. The classification of Bally and Snelson (1980) additionally took into account the rigidity of the lithosphere, and the style of subduction. Kingston *et al.* (1983) presented a geometric classification of basins with reference to their tectonic and depositional settings, as well as modifications which take place after the formation of such basins. Miall (1984a) placed much emphasis on the opening and closure of oceans as a basis for his classification of basins, which was presented as an expanded version of an earlier classification in which he linked basin tectonics to sediment-dispersal patterns (Miall, 1981). All the criteria used by earlier workers have been incorporated in a recent classification by Klein (1987), in which geodynamic processes are emphasised and various aspects such as thermal subsidence, rifting, stretching, and lithospheric flexure are brought in context with basin formation.

A *foreland basin* forms between an orogenic mountain belt and an adjacent craton (Allen *et al.*, 1986). Dickenson (1974) distinguished between two types of foreland basins:

(1) A *peripheral (foredeep) basin* is situated against the outer arc of a mountain belt during a continent-continent or continent-arc collision. In the event where continental crust or an arc is carried into a subduction zone, the latter eventually develops into a suture. Longitudinal drainage patterns are more likely to develop on account of the location of such basins between sutures and continents (Miall, 1981).

(2) A *retroarc (foreland) basin* is situated behind a magmatic arc generated by collision of oceanic and continental plates. The result of such a collision is subduction of the oceanic plate beneath the continental plate. Dickinson (1974) and Miall (1981) pointed out that retroarc basins were characterised by the prevalence of transverse basin-fill patterns.

Foreland basins are formed as a result of lithospheric loading in fold-thrust belts, followed by subsidence of the basin (Price and Mountjoy, 1970, quoted in Dickinson, 1974; Jordan, 1981). The behaviour of foreland basins is varied and is controlled mainly by the nature of the supporting lithosphere, including its thickness, tectonic grain, and thermal state (Allen *et al.*, 1986). Several models have been presented to date to explain the behaviour of a loaded lithosphere (i.e. due to thrust-loading) and its relevance to the formation of foreland basins. The purely *elastic model* explains lithospheric flexure as a function of the thermal state (Watts *et al.,* 1980), which in turn governs the thickness and rigidity of the plate (Royden and Karner, 1984). Quantitative modelling has shown that this model does not apply as well to the continental lithosphere as to the oceanic lithosphere (reasons are listed in Allen *et al.*,1986). The model was applied quantitatively and with satisfactory results to the foreland basin of the Rockies in the U.S.A. (Jordan, 1981). However, evidence of excessive variation in flexural rigidity along the strike of the Indian plate, based on gravimetric modelling by Lyon-Caen and Molnar (1985), contradicts the expected behaviour of an elastic lithosphere, at least in the case of the Himalayan orogeny (Allen *et al.*,1986). The *viscoelastic model* applied by Beaumont (1981) in his analysis of the Canadian Rockies and Alberta foreland basin (Canada) hinges on the questionable assumption that after loading the continental lithosphere would progressively soften and become more flexible with time, followed by subsidence of the basin. A third model *(thermo-rheological model)* constructed by Kusznir and Karner (1985) on the basis of the temperature and stress behaviour of lithospheric materials, has not yet been tested in actual case studies.

8.4.3 Diagnostic Features and Classification of the Late Permian Karoo Basin

The most outstanding feature of the Late Permian Karoo Basin is its distinctive asymmetric geometry, seen for instance on isopach maps of megasequences A and B of the lower Beaufort (FIG. 8.2 and 8.3). Even when the effect of the diachronous transition between the Ecca and Beaufort Groups is taken into account, the wedge shape of the Basin is clearly evident in transverse section (north-south). The interval between two chronostratigraphic horizons, taken in this instance at the top of the Whitehill Formation and the base of megasequence B, thins over a distance of just more than 200 km from 3700 m north of Prince Albert (Ecca thickness from Rossouw *et al.*, 1964) to an estimated thickness of less than 600 m in the vicinity of Carnarvon (Ecca thickness by J.C. Terblanche, pers. comm., 1987).

Isopach maps (of megasequence A in particular) and the distribution of the lower Beaufort strata in the Basin show some evidence for syndepositional warping within the Basin. The weakly defined flexure in the area just south of Sutherland (FIG. 8.2) resembles what Rocco and Jaboli (1958) referred to as "neogenic anticlines" in the Po basin. Along the northern margin of the Basin the lower Beaufort pinches out on the southern flanks of a subtle hinge-zone (Visser, 1978), interpreted as the peripheral forebulge of the Karoo foreland basin on the basis of descriptions of those features in the Appalachian and Cordilleran foreland basins of North America (Quinlan and Beaumont, 1984; Tankard, 1986). In foreland basins intrabasinal arching is considered the result of lithospheric flexure, initially, then basin subsidence in response to thrust loading (Quinlan and Beaumont, 1984). That is followed by flexural bulging during viscoelastic relaxation (softening) of the lithosphere, and the process is repeated during subsequent thrusting episodes (Tankard, 1986).

The Karoo Basin was yoked to a fold-thrust belt (Cape Fold Belt) in the south. Folding also occurred along the western margin, but this folding is of an uncertain age. Thrust faulting is not easily identified but is nevertheless present in the quartzitic Cape Supergroup rocks of the Cape Fold Belt in the south (Hiller and Snowden, 1983). Thrust faulting on a limited scale has also been identified in lower Beaufort strata in the southern part of the study area (Rossouw and De Villiers, 1952). The style of structural deformation along the southern margin of the Basin is typical of fold-thrust belts, and this deformation was clearly involved in the formation of the Karoo Basin. However, the structural relationship between the Basin and the Cape folding at the western margin is not entirely clear. The latter deformation may have controlled the transverse component of compression and folding which formed the Willowmore arch.

The large-scale cyclicity of the lower Beaufort succession is manifested over a large area of the Basin, and is clearly the result of extrabasinal, allocyclic sedimentary controls (terminology of Beerbower, 1964). The large-scale cycles, referred to here as "megasequences" after the terminology of Heward (1978), are interpreted to have formed in response to repeated tectonic pulses, as reviewed by Miall (1978b). The lower Beaufort consists of a cyclical succession of the distal deposits of thrust-derived, alluvial fan sequences, of which the proximal (coarse clastic) facies were destroyed by tectonic cannibalisation in a fashion typical of foreland basins (DeCelles *et al.*, 1987). The predominantly arenaceous members were thus deposited during periods of tectonic activity (thrusting) and source elevation, and the argillaceous members during periods of tectonic quiescence and peneplanation in the source, similar to the style of sedimentation in the molasse deposits of the Canadian Western Interior Plains (Stott, 1972, quoted in Miall, 1978b). The locations of alluvial fan apices through which the drainage systems entered the Basin were probably controlled by structural lows in the thrust front (Allen *et al.*, 1986).

Another diagnostic feature of the Karoo Basin, albeit not exclusive to the foreland setting, involves the regional drainage patterns. The lowermost Beaufort (megasequence A, FIG. 7.3) and underlying Ecca Group show distinct transverse (northerly) drainage patterns along the entire southern margin of the Karoo Basin (Ryan, 1967; Theron, 1970; also FIG. 7.3). A minor contribution of sediment from the Kaapvaal craton, transported southwards into the Basin during Ecca sedimentation, is also evident (Visser, in press). This pattern of predominantly cratonward (northwards) drainage, with a lesser component from the craton (southwards) towards the foreland, appears to be typical of retroarc (foreland) rather than of peripheral basins (Dickenson, 1974; Miall, 1981). Examples of foreland basins in which parts of the transverse basin fills have been reworked by longitudinal rivers, similar to that seen in the lower Beaufort, have been discussed by Miall (1981).

Tuff beds in the southern part of the Basin indicate the presence of syntectonic magmatism in the source, which was related to the Gondwanide orogeny by Elliot and Watts (1974). Barrett *et al.* (1972) considered these volcaniclastics to have emanated from the western part of Antarctica. The distribution of these volcanic rocks as indicated on a recent reconstruction of Gondwana by De Wit *et al.* (1988) confirms that position for the magmatic arc during the Cape orogeny. The paucity of volcanic rocks in the Basin furthermore points to a relatively cool (rigid) lithosphere at the time.

The entire Late Permian Karoo Basin is thus interpreted as a typical retroarc basin, as was suggested also by Visser (1978). The Basin was formed during collision of an oceanic plate (palaeo-Pacific plate) driven from the south into the craton (southwestern Gondwana plate). Folding and thrust-faulting of pre-Karoo rocks as well as the lowermost Karoo strata followed as subduction of the oceanic plate beneath the craton, proceeded. Lithospheric loading in the Cape Fold Belt led to subsidence of the Basin ahead of the thrust front.

The Basin is perched at high altitude at present, almost at the same elevation as the Cape Fold Belt.

Even though there is little evidence left of the once much more extensive fold-thrust belt which gave rise to a sedimentary basin of the dimensions of the Karooo Basin, certain conditions (e.g. previously extended foreland plate prior to overthrusting) exist under which thick overthrust wedges with very little topographic relief can develop next to foreland basins (reviewed in Allen *et al.*, 1986). Continued overthrusting and loading of the lithosphere furthermore resulted in an increasingly more rigid lithosphere in the foreland basin (Allen *et al.*, 1986), and these conditions were responsible for the preservation of the Karoo Basin-fill despite the reduction in the topographic load due to erosion of the Cape Fold Belt.

8.5 Palaeogeography of and Economic Implications for the Late Permian Karoo Basin

The western Karoo Basin consisted of a shallow, epicontinental sea in Middle to Late Permian times, and was bound to the south by an emergent Cape Fold Belt. The northern part of the Basin occupied the southern rim of the Kaapvaal Craton. The Basin was initially open (Middle Permian) towards the west, and extended eastwards into Antarctica. A palaeogeographic reconstruction of the Karoo Basin during lower Beaufort time, is shown in FIG. 8.4.

The Basin initially occupied a linear depression, striking almost east-west. During deposition of megasequence A the Basin was pear-shaped in plan, having been somewhat wider along the (presently eroded) western margin than is evident from the present distribution of lower Beaufort strata. Transverse drainage patterns (FIG. 7.3) indicate a central entry point in the south from which sediment was dispersed during the early history of the Karoo Basin. Sandstone:mudstone ratios show an unusual increase in the northern part of the Basin from which could be construed a northerly entry point. The drainage pattern shown by the palaeocurrent analysis is, however, accepted as being more reliable. The southernmost reaches of rivers entering the Basin from the south were ephemeral braided streams, and assumed a low-sinuosity meandering character downstream. The rivers debouched through river-dominated deltas into a shallow sea. This pattern is typical of a distal alluvial fan (fan-toe) setting.

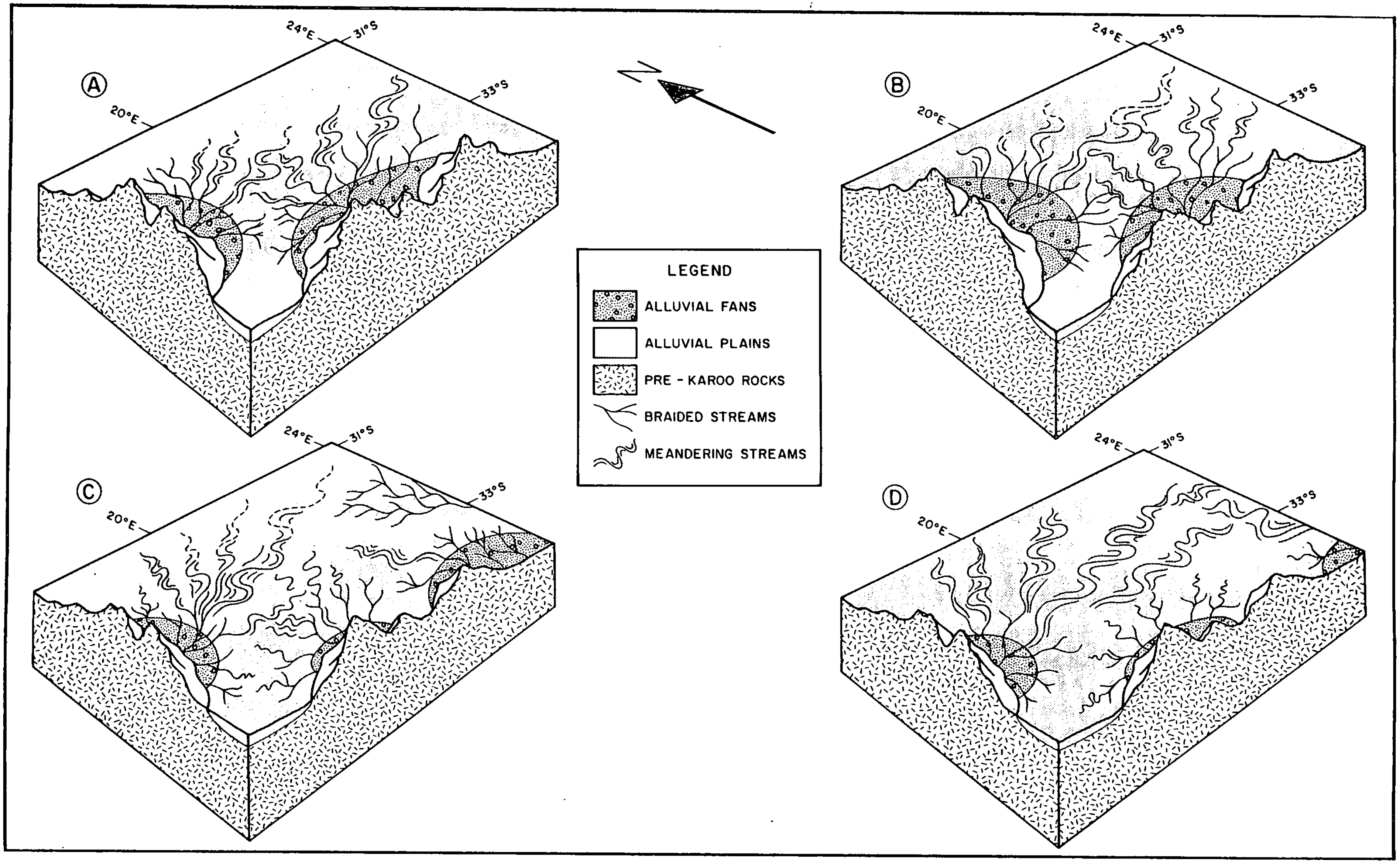


FIG. 8.4 Schematic reconstruction of the palaeogeography of the western Karoo Basin during the Late Permian. Alluvial depositional subenvironments and palaeodrainge patterns indicated for megasequences A, B, C and D.

Braided streams also entered the Basin from a west-southwesterly direction as part of a longitudinal drainage system which was active throughout the deposition of the lower Beaufort. These rivers changed abruptly downstream into high-sinuosity meandering streams. The presence of fish fossils in meandering channel deposits (Broom, 1918; Jubb and Gardiner, 1975) suggests that these streams may have been perennial, at least locally in the distal parts of the Basin. Longitudinal drainage became predominant later on in lower Beaufort times, when the southern hinterland became less prominent and eventually ceased to supply significant volumes of sediment to the Basin.

Sediments were also transported longitudinally along the axis of the Basin since early-Beaufort times. A proportion of the transverse basin fill was redistributed in the process, and extrinsic sediments were also introduced in the process. The indistinct radial drainage pattern seen in the western part of the Basin (FIG. 7.3) may therefore suggest longitudinal alluvial fan drainage. Extraformational clasts in sandstones in the Moordenaars Karoo, transported from the west, support the view that these sediments had been derived from a source beyond the western margin of the Basin. The western source probably consisted mainly of basement rocks, elevated as a result of thermal updoming of parts of Gondwana (Uruguayan Shield?) during the incipient stages of rifting of the continent. The southeastern source became more active during deposition of the Nuweveld member, shedding sediment into the Basin from two separate entry points. The Willowmore arch which caused the attenuation of megasequence A of the lower Beaufort in the easternmost part of the study area, became less prominent towards the end of lower Beaufort deposition. The structure nevertheless appears to have exerted some control on the supply and distribution of sediments during deposition of megasequence C, and possibly megasequence D as well.

A cool, humid climate prevailed initially in the southernmost part of the Basin, which was subjected to orographic rainfall. The Basin was inundated locally, freshwater lakes covering large areas in the south during deposition of the Leeu Gamka member. Vegetation was sparse and perhaps somewhat stunted. Larger plants such as trees, as well as therapsid reptiles, were concentrated in lacustrine regions. Calcareous palaeosols suggest that the northern part of the Basin was arid and sparsely vegetated, particularly during deposition of the topmost members of the lower Beaufort. Precipitation in that region was intermittent, interspersed with lengthy droughts. Hardened vegetation and animal life were restricted to areas that were irrigated by river systems flowing from the south and east. Some of the longitudinal rivers were probably perennial as they were fed by springs, emanating from a shallow groundwater table in the fan-toe region.

A combination of factors, mainly the tectonic setting in the orogenic foreland and the inhospitable climate, rendered the western part of the Karoo Basin an unfavourable host for coal and many other mineral deposits. Due to tectonic activity along the southern margin the Basin subsided at a rate too high, and the deltaic systems prograded too fast, to allow a lush vegetation to become established in one particular region for periods long enough to form peat swamps. The tectonic setting was favourable for uranium mineralization but the rates of Basin subsidence and sediment compaction were too high to permit the formation of large-scale sandstone-hosted uranium ore deposits. The arenaceous channel-fills were also too fine-grained and sparsely distributed in the succession, on account of the distal setting of the Basin, to allow large-scale epigenetic enrichment of the ore by circulating groundwater. The distribution of the Karoo rocks nevertheless permits a reconstruction of the orogenic system by which may be assessed the potential of the foreland and southern part of the Karoo foreland basin to host epithermal gold deposits.

9.0 ECONOMIC GEOLOGY

9.1 Introduction

The tectonic setting of the Karoo Basin and the sheer size of the area underlain by Karoo strata have made the Basin a frequent exploration target for minerals and energy materials during the last century. Although a variety of minerals occur in the area, no viable deposits have yet been proved in the western Karoo Basin. Two trial mining exercises to explore the Rietkuil and Rystkuil uranium deposits (FIG. 1.2), have been carried out in recent years.

Pseudocoal, a carbonaceous substance thought to be a petroleum derivative (Rossouw and De Villiers, 1952), was discovered in 1865 on the farm Leeuwrivierspoort situated 50 km northwest of Beaufort West (Haughton *et al.*, 1953). The pseudocoal occurrences are confined mainly to two areas, namely the Koup and the Kamdebo (FIG. 1.2). An exploration hole was drilled during the years 1879 to 1880 to explore for subsurface "coal" in the Kamdebo region (Rogers, 1917). Subsequent investigations of the economic potential of the pseudocoals in the Kamdebo region by Green (1883) and Dunn (1886) yielded negative results.

Gold discoveries followed soon afterwards in 1870 on the farm Spreeuw Fontein (FIG. 1.2), situated in what then became known as the Prince Albert goldfields (Dunn, 1871; Letcher, 1936). Mineralization was confined to the area comprising the farms Spreeuw Fontein, Varsch Fontein and Ganze Kraal (latter two adjoining Spreeuw Fontein). Parts of the area were proclaimed as a payable goldfield on the advice of Bain (1891). Although activities ceased early in this century, evidence of early prospecting is still present in the area, mainly as piles of vein quartz which have been excavated from the host sandstone. It appears as if much of the excavated vein quartz was stockpiled and never processed. Most of the almost 39 kg of gold recovered until 1893, when Sawyer (1893) visited the area, was recovered from alluvial excavations on the farm Spreeuw Fontein (Rossouw and De Villiers, 1952).

Field investigations into the possible presence of petroleum in the western Karoo Basin were first made during the First World War (Rogers, 1917), and again shortly after the start of the Second World War (Haughton *et al.*, 1953). SOEKOR as well as a number of private consortiums sank a number of petroleum exploration wells in the area in the 1960's and 1970's, all with negative results. Core drilled in the hole VR1, on the farm Vrede in the Graaff-Reinet District (location 37, FIG. 1.2), was used for stratigraphic purposes in the present investigation.

References were also made to copper mineralization on the farm Botterkraal (20 km northwest of Prince Albert; Sawyer, 1893), and in the area of the present town of Fraserburg (Molyneux, 1881).

Radioactivity was first detected in in the Karoo in 1967 in the course of downhole geophysical logging of petroleum exploration wells (Toens *et al.*, 1980). Sandstone-hosted uranium mineralization was subsequently discovered in 1969 in the vicinity of Beaufort West during a carborne radiometric survey of the main Karoo Basin (Moon, 1977). Vigorous exploration followed the initial uranium discoveries after the news became publicly known in *ca.* 1973, and persisted until the early 1980's. Significant concentrations of molybdenum were found associated with uranium mineralization (Cole, 1986). Several areas in the main Karoo Basin still remain prospective for uranium and molybdenum mineralization, the most important and best mineralized being the lower Beaufort succession in the western Karoo Basin. The nature and depositional controls of these deposits, as well as some recommendations for future work, are discussed here in as much detail as would be allowed by the confidential nature of the information.

9.2 Pseudocoal

9.2.1 Description

Pseudocoal occurs as cross-cutting veins and sills up to 2 m thick. It was studied at two localities in the Koup, namely on the farms Wilgerboskloof and De Drift near Merweville (FIG. 1.2). The pseudocoal on Wilgerboskloof occurs as irregular, lenticular, sill-like bodies on the slopes of the Klein Roggeveld mountain range. Those occurrences accessible from surface have all been depleted, and their nature and contact relationships therefore remain uncertain. The pseudocoal on the farm De Drift occurs as vertical or steeply dipping veins. These veins trend almost due north-south and cut sharply through the host rocks, showing evidence that the walls were dragged upwards during intrusion (FIG. 9.1). Pseudocoal is still being mined on a small scale, mainly for domestic heating purposes.

The pseudocoal is lustrous, brittle and has a lower specific gravity than domestic coal. Chemical analyses reported in Haughton *et al.* (1953) revealed low content of ash (0,9 to 5,2 per cent) and sulphur (0,5 per cent), but high fixed carbon (74,6 to 89,3 per cent) and volatile (7,4 to 17,4 per cent). Petrographic examination by A. W. Rogers of samples from the Koup (described in Haughton *et al.*, 1953) revealed that most of the ash, comprising lithic fragments, garnet, zircon, tourmaline, white mica, and clear quartz, was derived from wall rocks of the Karoo Sequence. The pseudocoal in this area is classified as "ipsonite", a petroleum derivative, and is therefore unrelated to coal and coal tar (Haughton *et al.*, 1953).

9.2.2 Formation of Pseudocoal

"Intrusive coals" similar to the pseudocoal are associated with dolerite intrusions in the Robb and Back coals (Carboniferous) in Scotland (Lumsden, 1967). As the Beaufort succession in the western Karoo Basin is devoid of coal beds or carbonaceous source rocks it appears likely that the pseudocoal was derived and emplaced from depth. Haughton *et al.* (1953) reported pseudocoal occurrences throughout the lower part of the Karoo Sequence in the Koup.

Although the Whitehill and Prince Albert Formations appear to have been the most likely source rocks for the pseudocoal, the environments of deposition of these strata remain uncertain. Based on comparisons of C^{13} and O^{18} isotopic data from the Karoo and the (marine) Passa Dois Group in southern Brazil, Keith (1967) suggested that the middle and upper Dwyka strata (now Prince Albert Shale and Whitehill Shale Formations) were deposited under predominantly fresh water conditions. The results of trace element (boron, manganese and lithium) studies of shales in the eastern part of the Basin were inconclusive (Kingsley, 1977) but suggested that the Ecca Group was deposited in an environment that was not typically marine but probably brackish. Subsequent sedimentological analyses (Visser *et al.*, 1980) nevertheless showed that shallow marine conditions prevailed in the Basin during the Early Permian, when the lower Ecca Group shales were deposited.

FIG. 9.1 Dragged walls of a sharply defined, north-south trending pseudocoal dyke 0,4 m wide, on the farm De Drift near Merweville.

FIG. 9.2 Vertical quartz veins hosting gold mineralization in sandstones on the farm Ganze Kraal, Prince Albert District. (Scale 1 m.)

Considering that the "intrusive coals" in Scotland (Lumsden, 1967) were derived by metamorphism of existing coal deposits, it is clear that the Karoo pseudocoals may not have been derived from petroliferous source rocks, as is generally believed. The pseudocoals were probably distilled as viscous liquids from coalified, vegetal material in the lower Karoo shales during a regional low-grade metamorphic event. These liquids were probably emplaced as diapirs, perhaps exploiting zones of structural weakness. The pseudocoals furthermore occur very patchily, and do not have the potential to become large-scale coal resources in the western Karoo Basin.

9.3 Gold Mineralization

9.3.1 Description of Gold Occurrences

The gold occurrences are associated with quartz-filled veins occurring in multistoreyed, type 3a channel sandstones deposited by low-sinuosity meandering streams. These sandstones occur as isolated lenses near the top of the Leeu Gamka mudstone member. Palaeocurrent measurements (n=77) in the mineralized sandstones throughout the Prince Albert goldfields (location W, FIG. 7.1) showed a mean northerly transport direction (344°). The sandstones are up to 6 or 7 m thick. Identical quartz veins in sandstones in the same stratigraphic position and similar structural settings at several other localities in the area were found to be unmineralized.

The strata in the area are gently folded in open, slightly asymmetric synclines and anticlines, with the northern limbs of anticlines dipping steeper than the southern limbs. The quartz veins run along the crests of anticlines, reaching widths of up to 30 cm, and terminate on the sandstone/mudstone contacts (FIG. 9.2). In cases where the veins cut through the sandstones they die out on reaching the contact with mudstone. The mudstones below a veined sandstone, exposed in an adit on the farm Ganze Kraal, showed no signs of disturbance other than crumpling due to tectonic folding.

The quartz veins run parallel to fold axial planes, trending east-west with a mean orientation of 97°/277° (n=23 measurements). A set of tension joints developed in fold hinges of mineralized sandstone runs almost parallel (96°/276°) to the veins. A second conjugate set of joints is oriented at acute angles (156°/336° and 45°/225°) to the vein quartz and tension joints. Slickensides on bedding planes are oriented nearly at right angles (003°/183°) to the quartz veins. Fold geometry as well as orientation of the small-scale structures suggest that the joints formed as a result of compression towards the north. The quartz fillings were formed as a result of opening of the fold hinge tension joints, the result mainly of competence contrast between the sandstone and interbedded mudstone.

The gold occurs as irregularly shaped but commonly crystalline grains (Rossouw and De Villiers, 1952). Although in the present investigation gold analyses by means of atomic absorption spectrometry (AAS) with a lower detection limit of 10 parts per billion failed to detect gold in vein quartz samples, sporadic gold concentrations of up to 31,2 g/t in several vein quartz samples were found by Sawyer (1893), who also reported small quantities of gold in sandstone in contact with quartz veins. AAS analyses of several sandstone samples taken from the sandstone host rocks also yielded negative results. Control sampling and analysis of quartz veins and sandstones in a similar structural setting and in the same stratigraphic position on the farm Rietfontein (location R, FIG. 1.2) also yielded negative results.

Most of the gold recovered to date occurred as alluvial gold in the form of irregularly shaped fragments, some slightly rounded grains, and nuggets weighing up to 67 g (Letcher, 1936; Rossouw and De Villiers, 1952).

9.3.2 Origin of Gold Mineralization

The Prince Albert gold occurrences have been explained as the product of hydrothermal mineralization, on the one hand, and detrital concentration, on the other hand. A volcanogenic, hydrothermal origin is highly unlikely as the host rocks show no signs of alteration, and the veins are developed only locally in the sandstones. The quartz veins consist exclusively of alpha-quartz, a low-temperature variety of the mineral (Rossouw and De Villiers, 1952). There is no evidence of igneous rocks in the vicinity, and a petroleum exploration well (KW1) on the farm Klein Waterval (portion of the farm Spreeuw Fontein; FIG. 1.2) penetrated more than 5 km of sedimentary strata without any sign of intrusive source rocks or hydrothermal alteration of the Karoo strata.

After investigation of the discovery sites, Schwarz (1904) suggested that the Prince Albert goldfields once formed a peneplain which was continuous with the Swartberg mountain range in the south. He concluded that the Prince Albert gold was derived from auriferous strata in the Table Mountain Group somewhere to the south, and similar to the gold occurrences in Table Mountain quartzites at Millwood (near Knysna). The gold was assumed to have been transported as detrital particles in a northerly direction, the auriferous sediment having entered cavities during subsequent dissection of the peneplain, and the cavities later became silicified (Schwarz, 1904).

Rossouw and De Villiers (1952) suggested that the gold was derived from the same source as the Beaufort sediments, and transported as particulate matter to the sites of burial. Their mechanism does not explain the association of the gold with the quartz veins, the limited geographical distribution of the gold mineralization, and why the gold occurs only in the top parts of the channel sandstones (where the auriferous veins are found).

It appears quite likely that the gold was derived from the same source rocks (including plutonic and volcanic) as some of the Beaufort sediments, which may have contained hydrothermal and/or epithermal gold mineralization. The gold was transported mainly as detrital particles, and less likely as metallic complexes in solution. In the latter case the gold could have been precipitated in low concentrations by carbonaceous material in the alluvial sediments, in much the same way as the uranium deposits formed elsewhere in the Beaufort succession. Subsequently the precipitated gold could have been leached by fluids generated during tectonic deformation and low-grade metamorphism of the lower Beaufort succession, and deposited in tension gashes together with quartz. The alluvial gold occurring in the same area was derived by mechanical weathering of gold from the mineralized rocks, and deposited with the sediments (mainly sand) in recent streams. Limited chemical redistribution of some of the gold could have taken place during exposure and pedogenesis of the sandstone host rocks and alluvium, as suggested by the presence of gold nuggets and crystalline gold grains in the alluvial deposits (Rossouw and De Villiers, 1952). The geographic distribution of these gold occurrences in an area near the southern margin of the Basin where the main transverse river system is postulated to have entered the Basin, was controlled by the mechanical processes of concentration of trace amounts of detrital gold in river sediments. A less likely explanation is that low concentrations of gold in solution in river effluent or seepage water were scavenged by carbon in the sediments. The gold, irrespective of its origin, was probably redistributed and concentrated during regional metamorphism of the Karoo Basin.

9.4 Uranium Mineralization

9.4.1 Introduction

The uranium deposits in the lower Beaufort succession share many characteristics with sandstone-hosted deposits in other parts of the world. They commonly occur (1) in fluviatile sandstones which are arkosic and associated with tuffaceous beds, (2) in foreland basins adjacent to orogenic belts, and (3) in intimate association with carbonaceous material preserved in the host sandstones (Adler, 1977; Gabelman, 1970; Kostov, 1977). Uranium deposits of this nature were formed at low temperatures and pressures, as suggested by the lack of evidence of alteration usually associated with hydrothermal activity (Hostetler and Garrels, 1962; Langmuir, 1978).

Detailed information on the Karoo uranium deposits is extremely scarce, contrary to the wealth of published information on sandstone-hosted uranium deposits elsewhere in the western world. The first descriptions of the Karoo deposits by Kübler (1977) and Moon (1977) were of a general nature. Descriptions by Anderson and Fraenkel (1979) and Eddington and Harrison (1979) gave some insight into the subeconomic Rietkuil and Ryst Kuil deposits (FIG. 1.2). During the present investigation large areas in the western Karoo Basin were inaccessible due to option agreements and exploration activities, thus severely hampering surface investigations. Valuable subsurface information gained in intensive drilling programs has also been lost in the confidential and competitive exploration environment. As a result very little is known about the geometric relationships of the major uranium deposits and their host rocks. These problems were overcome to some extent by Le Roux (1985) who gained access to and described seven ore deposits located throughout the main Karoo Basin.

The origin and mineralization of sandstone-hosted uranium deposits world-wide are reviewed in the following discussion as a background to a genetic interpretation of the deposits in the western Karoo Basin. The interpretation involves mainly the macroscopic features of the uranium deposits related to the stratigraphic positions, sedimentological characteristics, and palaeogeographic distribution of the ore in the Basin.

9.4.2 Nature of Sandstone-hosted Uranium Deposits

9.4.2.1 Geochemistry of Uranium

Uranium occurs in nature invariably as quadrivalent and hexavalent ions (Kostov, 1977). Due to its large size uranium is one of the crystallochemically incompatible elements and tends to become enriched in residual magmatic solutions together with Zr, Ti, Th, Nb and Ta (Adler, 1977). The ionic radius of U^{4+} (1,01 Å) is similar to that of Th^{4+}, Ce^{3+}, Y^{3+}, Ca^{2+} and Na^{+}, and substitutes those elements in rock-forming minerals. Hexavalent uranium, on the other hand, does not easily substitute, and by its oxyphile nature tends to convert to the uranyl group UO_2^{2+} (Kostov, 1977).

Sandstone-hosted uranium deposits conspicuously lack evidence of igneous activity and of host rock alteration, which as a rule is associated with hydrothermal ore deposits. The concept of derivation of uranium in low-temperature and low-concentration solutions derived from tuffs and granites, and concentration in sandstone by surficial processes, was made popular by Gruner (1956). Hostetler and Garrels (1962) investigated the transportation and precipitation of uranium in low-temperature solutions, and showed that uranyl-dicarbonate (UDC) and uranyl-tricarbonate (UTC) complexes are stable in groundwater saturated with CO_2 (high P_{CO_2}). Langmuir (1978) evaluated the chemical properties of 42 dissolved uranium species and 30 uranium minerals, and considerably refined the earlier results of Hostetler and Garrels (1962) regarding the behaviour of uranium in solution. Langmuir (1978) also showed that the primary minerals uraninite and coffinite are far less soluble in natural waters than previously accepted, which means that precipitation of these minerals could take place from highly diluted solutions.

Uranium, both in the quadrivalent and hexavalent states, tends to form complexes in aqueous solutions. Langmuir (1978) determined that uranyl (UO_2^{2+}) carbonate complexes are formed predominantly in slightly alkaline waters, and are highly stable in moderately oxidising conditions. The solubility of the primary ore minerals uraninite and coffinite is relatively high under these conditions but decreases rapidly in slightly acidic conditions.

9.4.2.2 Origin of Uranium

The origin of all the uranium in sedimentary ore deposits is tied to the primary distribution of the element in the crust. Sedimentary uranium deposits are not necessarily associated with source rocks that are anomalously enriched in uranium (Gruner, 1956), but the mode in which uranium occurs in these rocks determines their suitability as uranium sources (S.H.U. Bowie, quoted in Dahlkamp, 1977).

Granites and tuffs have been widely advocated as potential source rocks. Like the other oxyphile elements uranium tends to remain in a melt, and therefore becomes enriched both in granitic magmas and in acidic to intermediate tuffs (Adler, 1977; Dahlkamp, 1977; Kostov, 1977). Approximately 70 per cent of the uranium in granites worldwide occurs in refractory minerals such as zircon and monazite, with the balance occurring as interstitial material or adsorbed on accessory minerals such as apatite and hematite (S.H.U. Bowie, quoted in Dahlkamp, 1977). The adsorbed and interstitial uranium is most easily removed during weathering of granites, and examples are known in granites (Granite Mountains, Wyoming) where up to 75 per cent (and possibly much more) of all the uranium originally present has been leached to depths of up to 50 m (H.H. Adler, quoted in Dahlkamp, 1977). Experiments conducted by Szalazy and Samsoni (1969) also showed that natural waters leach uranium much more readily from granites than from any of the other common igneous rocks. Other potential source rocks associated with uranium ore deposits are Precambrian cratonic sequences (Kostov, 1977) and anatectic granites related to orogenic belts (Gabelman, 1977).

The actual contribution of uranium by granites to the ore deposits in adjacent basins is often questioned. In most instances the tuffs occurring in the proximity of sandstone-hosted uranium deposits appear to be a more likely source of the uranium. The best known example is the Catahoula Formation of the Texas Coastal Plain described by Galloway and Kaiser (1980). In this instance the original uranium content of the tuff was estimated at 10 parts per million (ppm), of which an estimated 50 per cent was mobilized through pedogenic and diagenetic processes. The alkalinity of groundwater is raised during the alteration of tuffs (pH 7 to 11), thus providing the chemical environment required for UDC and UTC complexes to form (Langmuir, 1978). Stable minerals such as zeolites and montmorillonite which form during subsequent reaction of fresh tuffs with groundwater (Huang, 1978) are present in most sandstone-hosted uranium deposits.

The significance of tuffs as a source of uranium, and behaviour of tuffaceous sandstone host rocks during pedogenesis and diagenesis, were discussed by Galloway and Kaiser (1980). They studied several uranium deposits in the fluviatile Catahoula Formation (Eocene to Pliocene) of the Texas Coastal Plain, and showed that the tuffs were the most important source of uranium, and also that the nature of post-depositional alteration of the volcanic glass may have a bearing on their releasing of uranium to ground water. Zielinski *et al.* (1980) showed that tuff which had undergone zeolitisation, usually in the phreatic zone, tend to retain rather than release its uranium in the process.

9.4.2.3 Precipitation of Uranium

Moore (1954) determined experimentally that uranium was precipitated most effectively by low-rank coals, and that under those conditions the uranium was apparently held by an irreversible process. Kochenov *et al.* (1965) suggested precipitation of uranium by coalified organic matter in a combination of sorption (adsorption) and reduction. Muto *et al.* (1965) found that clay minerals and carbonaceous matter were effective adsorbents but that the process was only a temporary fixing mechanism, prior to reduction of the uranium by carbon and sulphur. Muto *et al.* also showed that uranium, on the basis of its thermochemical properties, can be reduced directly from solution only at high temperatures such as in hydrothermal solutions. Titayeva (1967) found that uranium precipitated by organic material was bound to humic and fulvic acids, the former soluble in alkaline and the latter in acidic solutions (Squyres, 1980). Titayeva (1967) also suggested that once precipitated the uranium was retained by ion-exchange. Jennings and Leventhal (1977) proposed a structural model for humic material with ion-exchange and chelate sites, suitable to interact with uranium both under reducing and oxidising conditions.

Haji-Vassiliou and Kerr (1973) found that organic matter was intimately associated with all the major sandstone-hosted uranium deposits known in the United States at that time. Huang (1978) suggested on the basis of investigations of the Texas Coastal Plain uranium deposits, that the degree of maturity (carbonisation) of interbedded organic matter determined its efficiency to adsorb uranium in solutions. Spectroscopic studies by Koglin *et al.* (1978) confirmed earlier speculations that the humic acid content of the coalified matter is primarily responsible for the precipitation of uranium, but they suggested also that uranium in solution reacts chemically rather than becoming adsorbed to the carbonaceous matter. Leventhal (1980), on the other hand, found in primary uranium deposits of the Grants uranium belt (U.S.A.) that oxygen and other functional groups of uranyl complexes, and not the uranium itself, were bonded chemically to the organic material.

In order to understand the Colorado Plateau-type uranium deposits Etheridge *et al.* (1980) studied the interaction between humic acid and an aluminium-potassium-sulphate solution (and precipitate) experimentally in a porous medium. Certain of their observations, such as the precipition of the humic compound both parallel and perpendicular to the hydraulic slope, and the controls of decreasing flow rates on the formation of humic precipitates, apply to the tabular Karoo uranium deposits as well.

In summary, the two most important processes involved in the supergene concentration of uranium in fluvial sand deposits are (1) adsorption or dissolution of the uranyl complex (usually UDC or UTC depending on pH and P_{CO_2}), followed by (2) the reduction of U^{4+} by mobile reductants such as H_2S and/or CH_4 (Langmuir, 1978).

9.4.3 Description of Western Karoo Uranium Deposits

The major uranium deposits all occur in the arenaceous members of the lower Beaufort succession in the cental parts of the western sector of the Karoo Basin (FIG. 1.2). No significant uranium concentrations are known in the eastern sector of the Basin adjoining the study area. The deposits at Ryst Kuil, Rietkuil, Vindragersfontein and Drie Vaderlandsche Rietvalleyen all occur near the top of the Koup member, although a number of smaller deposits occur also in the Nuweveld member (e.g. Vinkekuil; location 16, FIG. 1.2). No large-scale deposits are known in the Combrinkskraal and Leeukop members. The latter unit may have been explored inadequately due to the inaccessibility of these sandstones on the highest hilltops in the area, where they are mostly covered by dolerite. Promising surface mineralization is present in Leeukop sandstones in the area southwest of Fraserburg, and rare but nonetheless ore-grade uranium concentrations occur in the Combrinkskraal member (e.g. the farm Wolve Kraal, adjoining the Prince Albert townlands to the north). Although the major ore deposits are restricted to the most arenaceous parts of the succession, surface mineralization is patchily developed in sandstones throughout the lower Beaufort.

The mineralization is of a primary (syngenetic) nature and restricted to fluvial sandstone beds. The ore occurs as tabular units or as isolated pods concentrated on vaguely defined horizons. Layers of ore are stacked vertically in zones up to 7 m thick, and oriented mostly parallel but also oblique or perpendicular to the longitudinal axis of host sandstones (Toens *et al.*,1980). Mineralization is best developed in the thickest parts of the host sandstones (Kübler, 1977). The thickest sandstone unit in a particular interval is preferentially mineralized. An example is the 60 m thick Ryst Kuil sandstone (Eddington and Harrison, 1979) which is anomalously thick for that part of the succession where the maximum sandstone thickness rarely exceeds 20 m. A number of thick, potentially favourable sandstones in the western Karoo Basin are nevertheless only weakly and sporadically mineralized. Examples are those at the bases of the Droogvoetsfontein (65 m sandstone) and Uitkyk (37 m sandstone) sections (locations 11 and 38 in FIG. 1.2, respectively). Also, no significant mineralization is known in that part of the succession with the highest proportion of sandstone in the Basin, namely the Koup and Nuweveld members in the area northeast of the town of Sutherland (FIG. 8.1).

Uranium ore bodies are invariably associated with abundant carbonaceous material in the host sandstones, but certain ore deposits in the Koup member in particular contain an abundance of calcium carbonate (Kübler, 1977; Le Roux, 1985; Turner, 1985). The uranium ore was classified by Jakob (1979) according to its content of organic carbon, carbon dioxide, carbonate and sulphide. The uranium content in ore zones is not always associated directly with carbon, and Jakob (1979) found proportionately less carbon at the expense of carbon dioxide in higher grade uranium ores. Despite the general association of uranium ore with the carbonaceous matter in the sandstone, the stratigraphic intervals where the major uranium deposits are found, are not always marked by an unusual abundance of fossilised plant material. An example is the Nuweveld member in the region to the north and northwest of Richmond where the Nuwejaarsfontein deposit is located (FIG. 1.2), and which contains an abundance of ferruginised plant remains but relatively poor uranium mineralization. The Koup member in the Kamdebo and in the region surrounding the town of Aberdeen also contains an unusual abundance of silicified plant fossils but no significant uranium mineralization on surface or underground.

Mineralized sandstone is as a rule slightly less mature in composition than barren sandstone, the former containing more feldspar and being less well-sorted than the latter, and resulting in a distinctive "sugary" weathering where exposed in outcrop. This relationship cannot be quantified as the composition and weathering habit change significantly within a mineralized sandstone. Mottling (cf. FIG. 4.3) of the nature ascribed to the presence of tuffaceous material (Fuller, 1970) is conspicuous in the *arenaceous intervals* containing mineralized sandstone (e.g. the Koup member in the Moordenaars Karoo, and the Nuweveld member southwest of Fraserburg), but is not always present in a particular *mineralized sandstone*. Uranium mineralization is also very poorly developed in the southeastern parts of the study area where tuffaceous sandstone beds of the Combrinkskraal and Koup members are intensely zeolitised, but which would qualify in many other respects (e.g. sheet-like geometry, high carbon content, rare internal mudstone, etc.) as favourable host rock.

The large-scale uranium deposits in the western Karoo Basin occur geographically in the proximity of and stratigraphically just above the tuff beds which have been identified on the basal contact of, or immediately below the Koup member, south of Beaufort West. The appearance of tuff beds and volcanic clasts in the succession (at the top of Combrinkskraal and in the Koup members) is seen as evidence of the beginning of a phase of increased volcanic activity which was accompanied by the extrusion of large volumes of tuff and lava in the postulated southern source area. Some of these tuffs are likely to have been incorporated into the normal sediment discharge into the Basin, and a genetic relationship with the uranium deposits in the Koup member appears likely.

The most conspicuous feature of the large-scale uranium ore deposits in the western Karoo Basin is their appearance in that part of the succession where longitudinal palaeodrainage became dominant over transverse drainage. Sandstone of the Koup member (Poortjie formation of Turner, 1985), which contains more than 50 per cent of the known uranium occurrences in the western Karoo Basin (Turner, 1985), was transported longitudinally from west to east whereas only a minor proportion of apparently unmineralized sandstone came from the south (cf. FIG. 7.2). The major uranium deposits occur mainly in type 3b (high-sinuosity) channel sandstones of the Koup and Nuweveld members, the only notable exception being the Drie Vaderlandsche Rietvalleyen ore deposit in the Koup member (braided stream sandstones). The uranium ore deposits were thus formed preferentially in the high-sinuosity meandering stream deposits, and to a much lesser extent in the low-sinuosity stream deposits formed in the upstream regions of the alluvial plains.

The Ryst Kuil uranium deposit (FIG. 1.2) is exceptional in two respects: (1) it is a single deposit much larger than any of the other deposits, with an ore reserve much larger than the combined reserves of all the other significant uranium deposits in the western Karoo Basin, and (2) it occurs on the downstream (eastern) side of the axis of the transverse drainage system through which material was discharged from the southern source area into the Basin (cf. FIG. 7.2 and FIG. 8.4). The Ryst Kuil host sandstone was deposited by the longitudinal drainage system which was active during much of the Koup period. The host sediments were therefore derived at least partly by reworking the transverse sediment input, the latter presumably enriched in tuffaceous material and thus also in uranium, supplied by active volcanism in the southern source area.

Epigenetic uranium mineralization took place only in exceptional cases, also on a very localised scale (few metres), and always in the absence of other metals such as molybdenum. High but patchy uranium concentrations are typical of ore exposed on surface. No significant concentrations of primary uranium mineralization have yet been located in the vicinity of epigenetic shoots and the two types of uranium concentrations appear to be unrelated. The epigenetic process was accompanied by severe alteration of the host rocks, with tongues of mineralization (exceptionally high-grade in places) shooting irregularly into unaltered (and unmineralized) sandstone. The ore, which contains a variety of secondary ore minerals, is associated with faint limonite staining, but uraninite and coffinite as well as carbonaceous material are conspicuously absent. Remobilisation took place during the present erosional cycle but the fact that the uranium is in radiogenic equilibrium with its decay products suggests that the mineralization process was concluded at least a few hundreds of thousands of years ago. At present this type of ore is of no economic significance.

9.4.4 Depositional Model for Karoo Uranium Deposits

The western Karoo uranium (plus molybdenum) deposits were formed very soon after deposition of the host sands, as the lack of alteration suggests that the sands were unlithified and offered free passage to mineralizing fluids. Some of the earlier mineralization was trapped in irregularly dispersed calcareous pods of high-grade ore. The bulk of the mineralization was concentrated as tabular bodies in the basal parts of channel sandstones, where the coarser grain sizes (including conglomerates) resulted in greater permeability, and where waterlogged carbonaceous matter was already concentrated. The complexed uranium in solution was precipitated by carbonaceous material, and then reduced by H2S generated by decaying organic material. As uranium and molybdenum behave similarly in the erosional cycle, and are closely associated in the Karoo uranium deposits, reference will be made here only to uranium.

The uranium in solution was derived mainly from altered tuffaceous material once present in the channel sands, as suggested by the spatial association of the major ore deposits with mottled sandstones and tuff beds. The extreme paucity of tuff beds and their weathering products in the overbank mudstones, such as montmorillonite and related zeolites, suggests that most of the tuffaceous material in the channel sandstones was derived extraneously and transported in by fluvial drainage, rather than from reworked primary ash falls in the Basin. Weakly alkaline solutions with slightly elevated uranyl carbonate contents were formed during pedogenesis and surficial weathering of channel sands. In the event where sand deposits contained moderate amounts of tuff and clay, freely circulating solutions formed uranium mineralization where they came into contact with interbedded carbonaceous material. Anomalously high proportions of tuff as well as severe pedogenic alteration (zeolitisation) of channel sands, on the other hand, severely restricted groundwater circulation. The release of uranium from the tuffs and syngenetic uranium mineralization were prevented under those circumstances. The sandstones of the Koup member in the vicinity of Aberdeen and in the Kamdebo serve as good examples of otherwise favourable sandstones in which zeolitisation of the volcanic glass (estimated from the degree of mottling) has prevented mineralization.

The tabular geometry, location, and orientation of ore bodies within mineralized sandstones were controlled to a large extent by groundwater movement, and the position of the water table in confined sand aquifers. The orientation in certain cases of ore bodies perpendicular to the channel transport directions, is likely to have resulted from the precipitation of uranium at the interface between ore-forming solutions and the basinal groundwater (Etheridge *et al.*, 1980), roughly perpendicular to the hydraulic gradient of the aquifer (Galloway, 1980). Limited post-depositional redistribution of uranium ceased as soon as lithification of the sand set in. The absence of mineralization in an otherwise potentially favourable host rock could signify that the water table was too deep in certain areas to allow the formation and free circulation of mineralizing solutions. The role of the palaeowater table in the mineralization processes is, however, impossible to interpret from the lithological record.

The abundance of carbonaceous material associated with the uranium ore was derived mainly from levee and proximal floodplain vegetation. Plant material was eroded and incorporated into the stream load during lateral migration of meandering channels, disintegrated by frequent reworking, and then buried as finely divided fragments in point bar deposits. This material was concentrated mainly in the basal sediment deposits (including intraformational conglomerate) of channels. The efficacy and promptness of the burial process in meandering streams, as opposed to braided streams, ensured the preservation of active carbon suitable for the precipitation of complexed uranium. The major ore bodies are concentrated in a limited area in the southwestern parts of the Karoo Basin where vegetation was stimulated either by orographic rainfall (Turner, 1985), or by frequent flood irrigation by perennial meandering streams.

9.4.5 Future Exploration

The guidelines for future uranium exploration in the western Karoo Basin are determined to a large extent by the economic feasibility to mine ore deposits of certain grades and tonnages at given depths. Although maximum ore grades are comparable to those of sandstone-hosted deposits world-wide, the mean grade and size of Karoo ore deposits are considerably lower than for instance those of Ambrosia Lake (New Mexico), Colorado Plateau, or Texas Coastal Plain in the United States (Grutt, 1971). Mining and extraction costs are the most important factors which would determine the future economic viability of uranium deposits of the grade and size known in the Karoo sandstone. Potentially economic ore deposits therefore need to be located at shallow depth and close to one another so that several deposits can share the same ore processing facility.

The Koup member in the entire area surrounding Beaufort West, bounded roughly by a line through the town of Merweville in the west, the Kariega River area in the east, and the Nuweveld escarpment in the north, has considerable potential to host more ore deposits. The eastern part of this area has greater potential on account of its position relative to the transverse entry point to the Basin, and the direction of longitudinal drainage and redistribution of tuffaceous sediments. The entire Koup member (a few hundred metres thick) can be penetrated over most of the area by conventional rotary percussion drilling, an efficient and relatively inexpensive drilling method.

The ore is in radiogenic equilibrium everywhere in the western Karoo Basin, and downhole radiometric logging is therefore a reliable technique for the detection of uranium ore intersections in exploration and water boreholes. The high degree of mobility of uranium in the weathering cycle makes hydrogeochemical sampling in water and exploration boreholes an attractive technique in addition to routine drilling programs, and despite the low permeability of the lower Beaufort sandstones. Radon-in-groundwater has proved itself in exploration as an even more valuable semi-quantitative technique to detect blind uranium ore deposits in the vicinity of water-bearing boreholes. Experience has shown, however, that count rates have to be compensated for drastically when comparing radon-in-groundwater results from newly drilled boreholes (e.g. for exploration purposes) with those from established water boreholes, as well as in the latter case, results obtained alternately on quiet or windy days. These corrections are carried out in addition to temporal corrections for radiometric decay losses of radon. The abundance of disseminated pyrite in uranium ore zones also makes electrical sounding an attractive tool to explore for extensions to known ore bodies. The induced polarization technique has been used with some success in the Karoo, but is too time-consuming to be used routinely for reconnaissance purposes.

In summary, renewed uranium exploration in the western Karoo Basin should be concentrated exclusively in the area outlined. Considering that additional surface investigations (including airborne radiometric surveys) are unlikely to yield more accurate or meaningful information in addition to that collected during the previous phase of exploration, comprehensive subsurface exploration will be the most viable alternative exploration method. Deep-level drilling of the Koup member, taking into account the sedimentological characteristics and spacing of the known Karoo uranium deposits, is recommended. In the past similar programs have been undertaken in an area to the east of that recommended, but were unsuccessful. Drilling should therefore be supplemented with regional hydrogeochemical and radon-in-groundwater sampling programs, as well as induced polarization surveys where promising signs of mineralization are found.

Exploration results have to date shown only one other element, molybdenum, to be present in ore grades, albeit in uneconomic quantities. Its close geochemical association with uranium allows the use of similar geochemical exploration techniques.

ACKNOWLEDGEMENTS

The author wishes to thank the following persons and institutions:

My wife Hester and family, for moral support and understanding.

The New Mining Business and Gold Divisions of Anglo American Corporation of South Africa Ltd., for financial and material assistance.

Prof. J.N.J. Visser for guidance and constructive criticism.

Mr. J.C. Loock for assistance and guidance with fieldwork.

Glenda Laas, Monika Franke and Karen van Vuuren who assisted with drafting of diagrams.

Richard Downing of De Beer and Associates, for assistance with layout and printing.

All the Karoo geologists, too many to name, with whom I had inspiring discussions and field trips in the course of my fieldwork.

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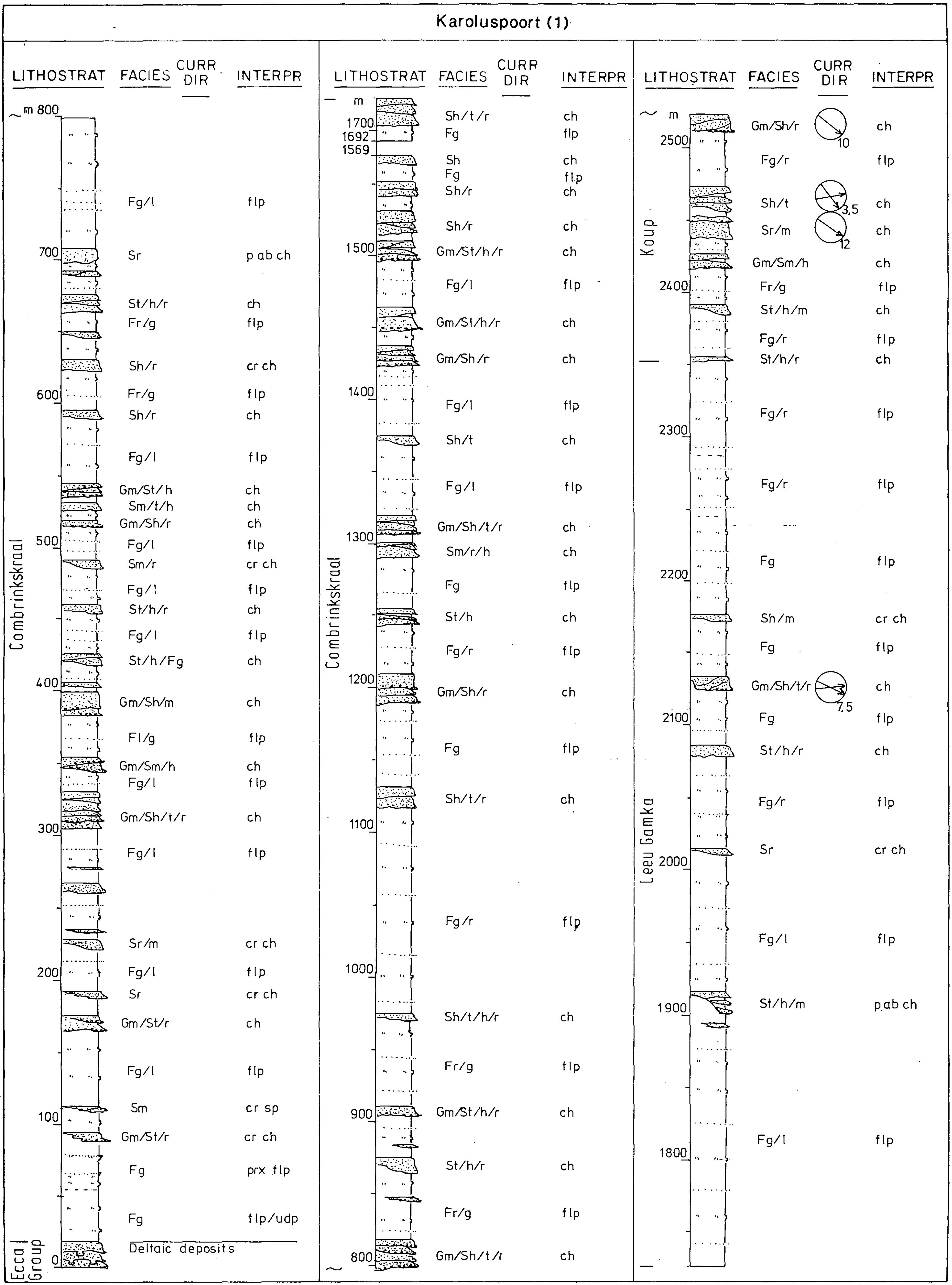
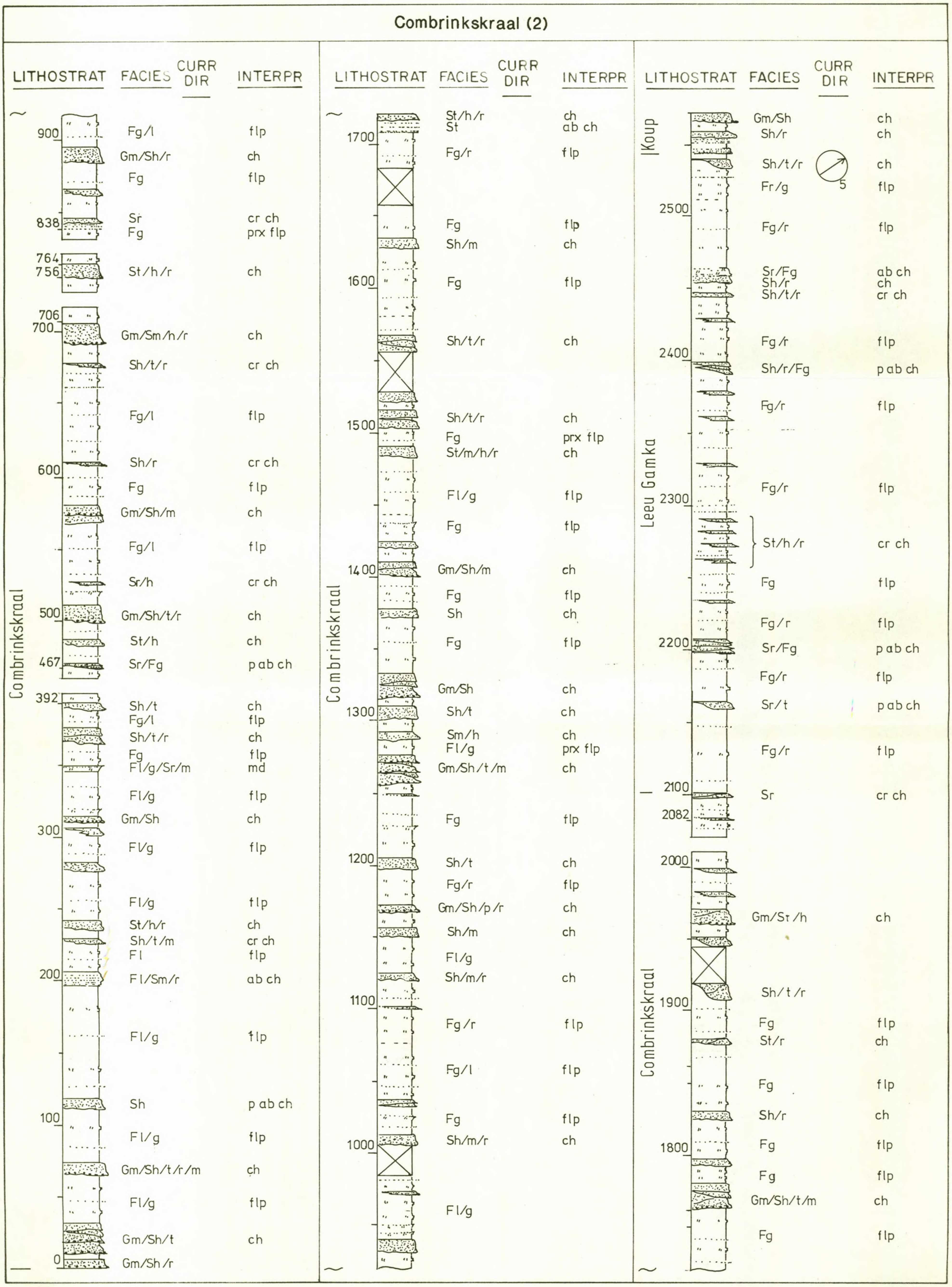
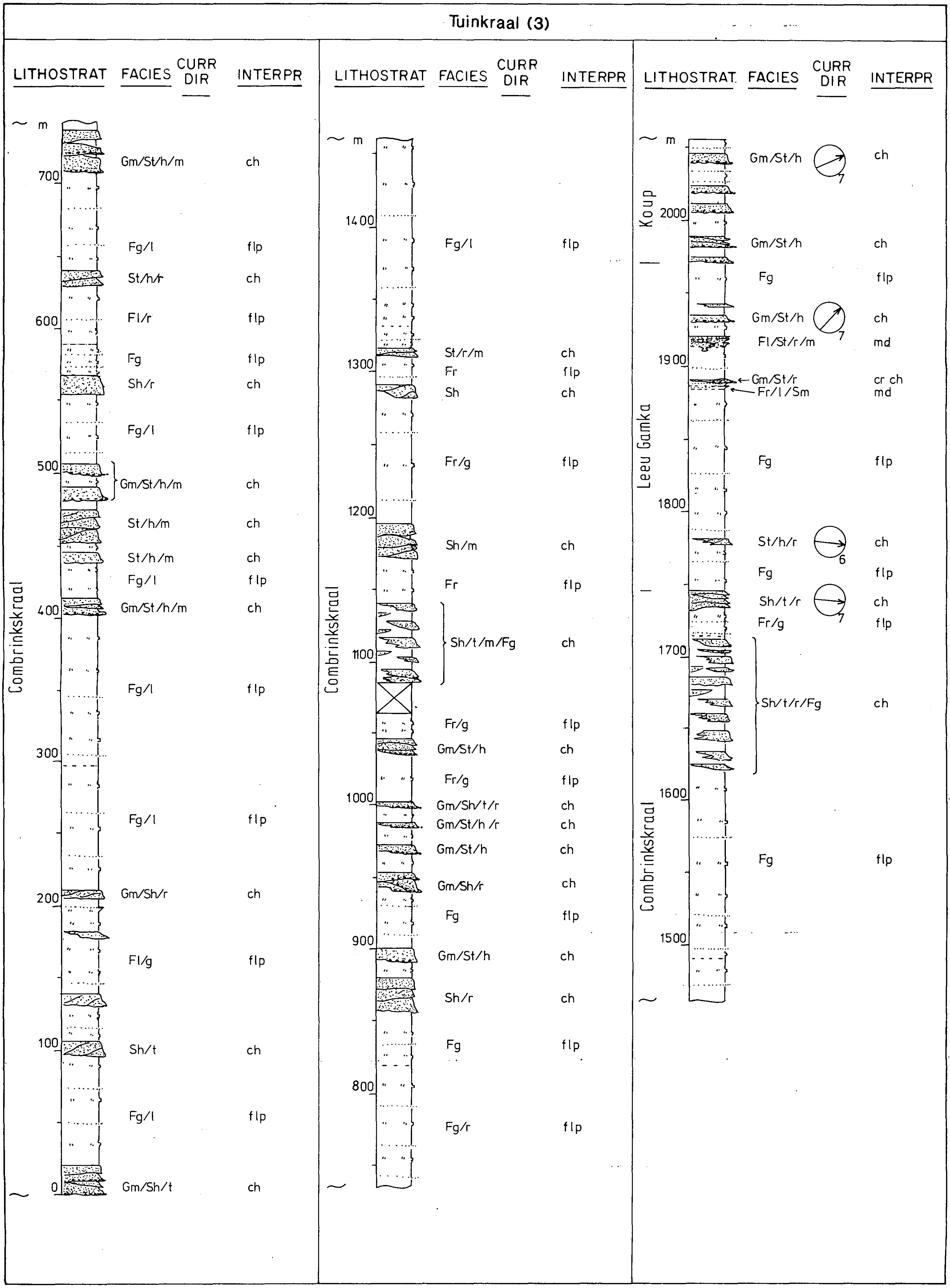
APPENDIX 1 Stratigraphic profiles measured at location numbers 1 to 38. (Scale = 1:2500; see FIG. 1.2 for locations, and APP. 2 for legend.)


Karoluspoort (1)

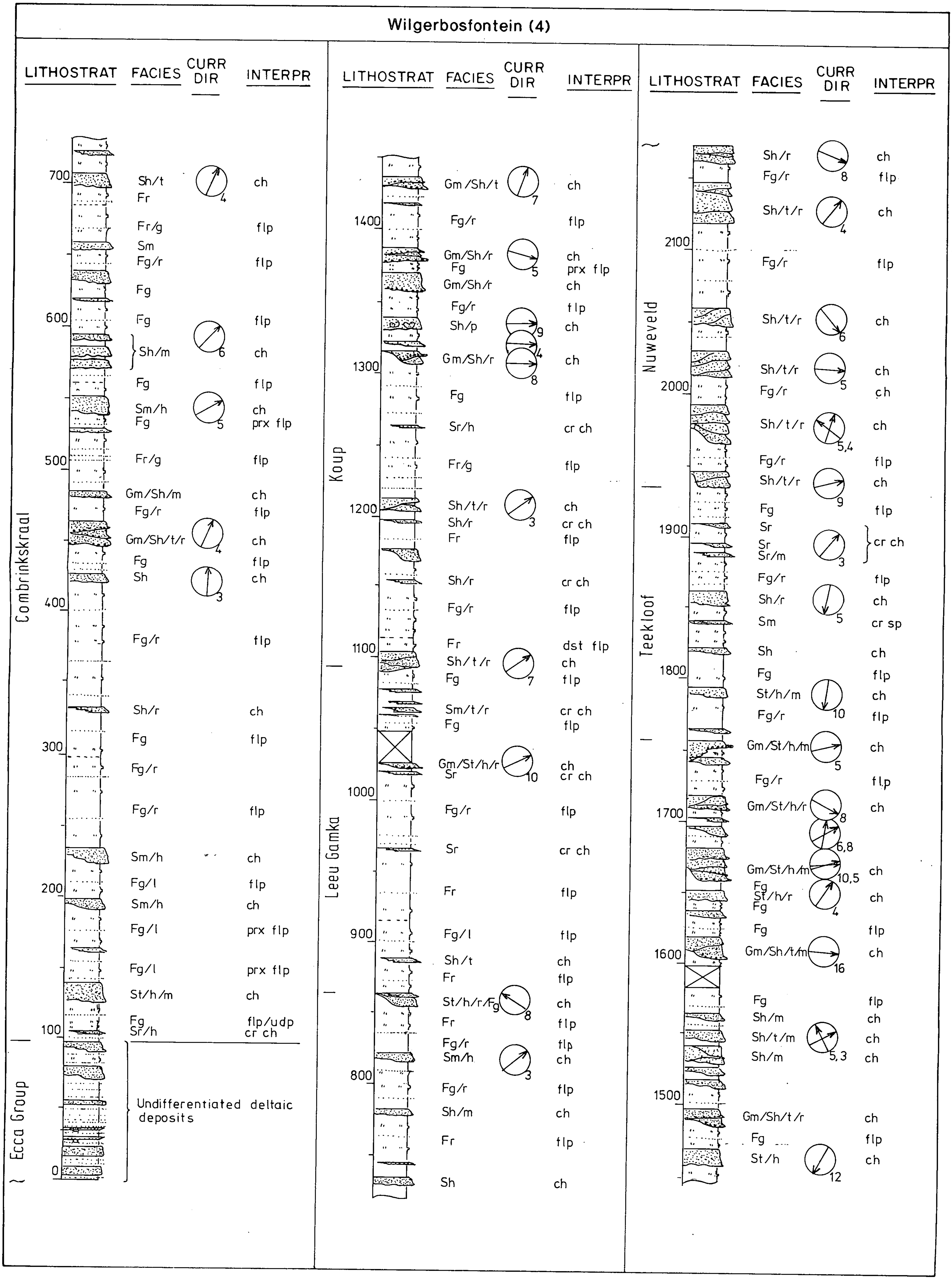
APP. 1.1 Stratigraphic profile measured on Karoluspoort (location 1).



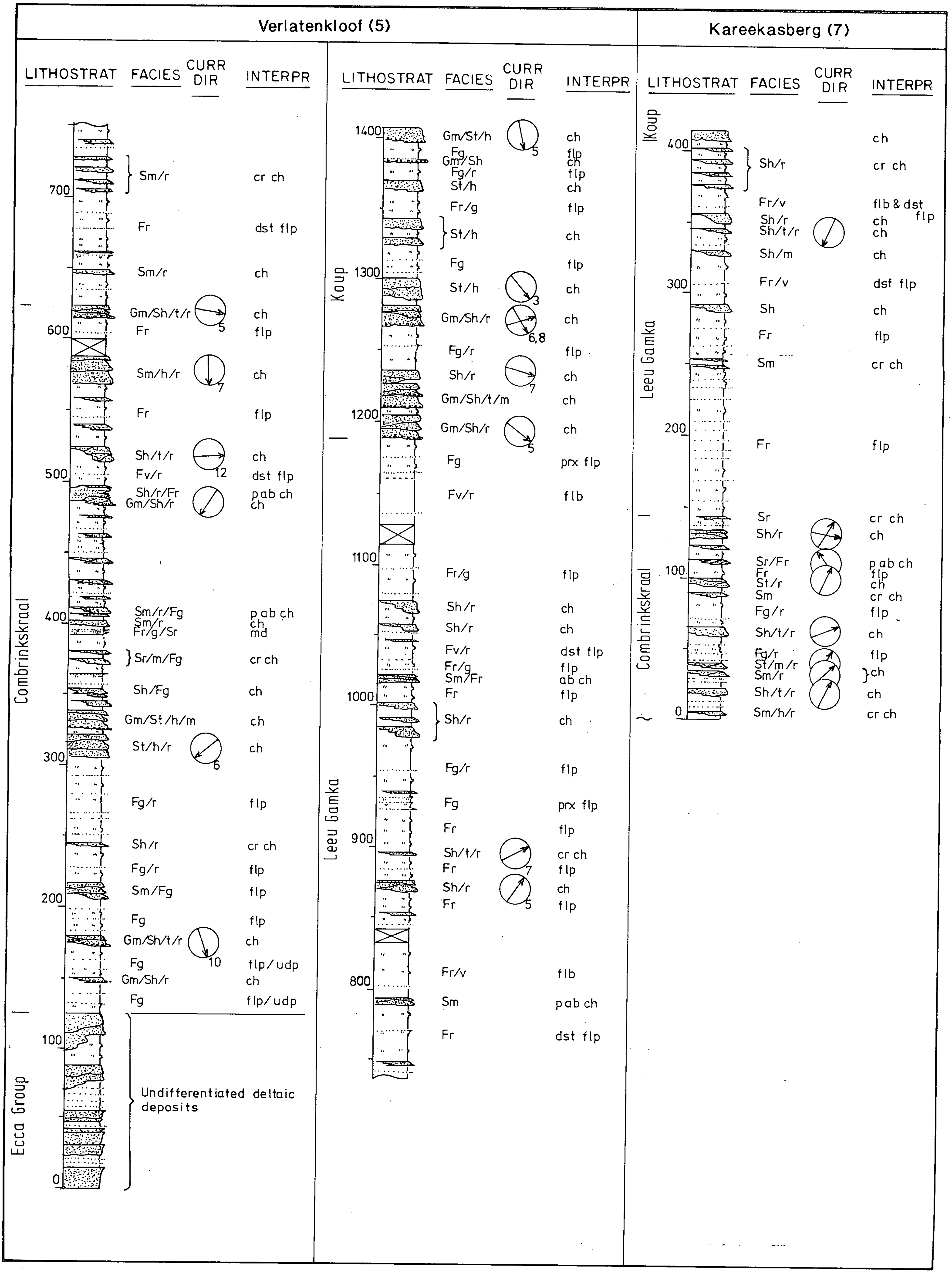
APP. 1.2 Stratigraphic profile measured on Combrinkskraal (location 2).



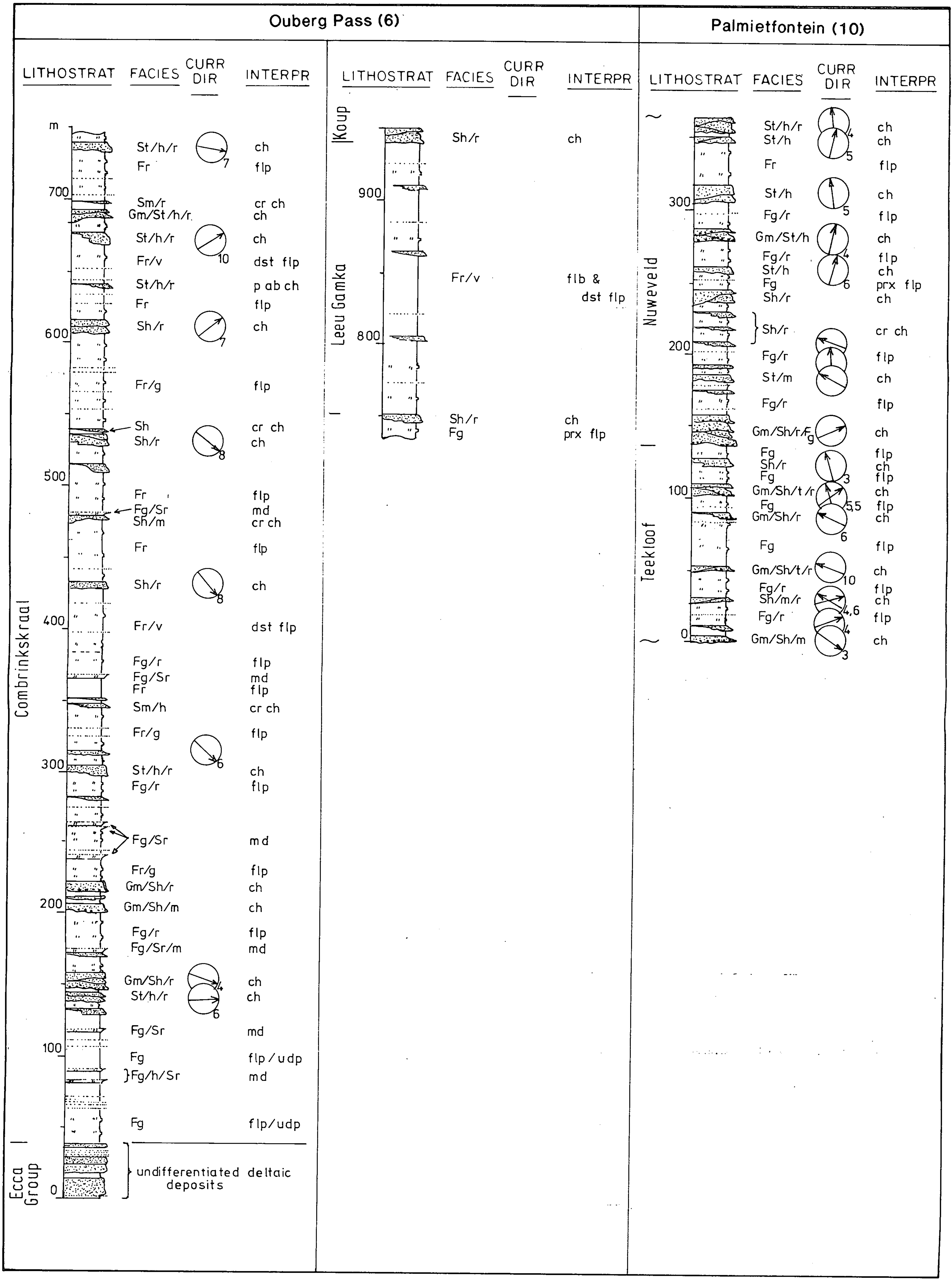
APP. 1.3 Stratigraphic profile of a traverse between Tuinkraal and Elandsberg (location 3).



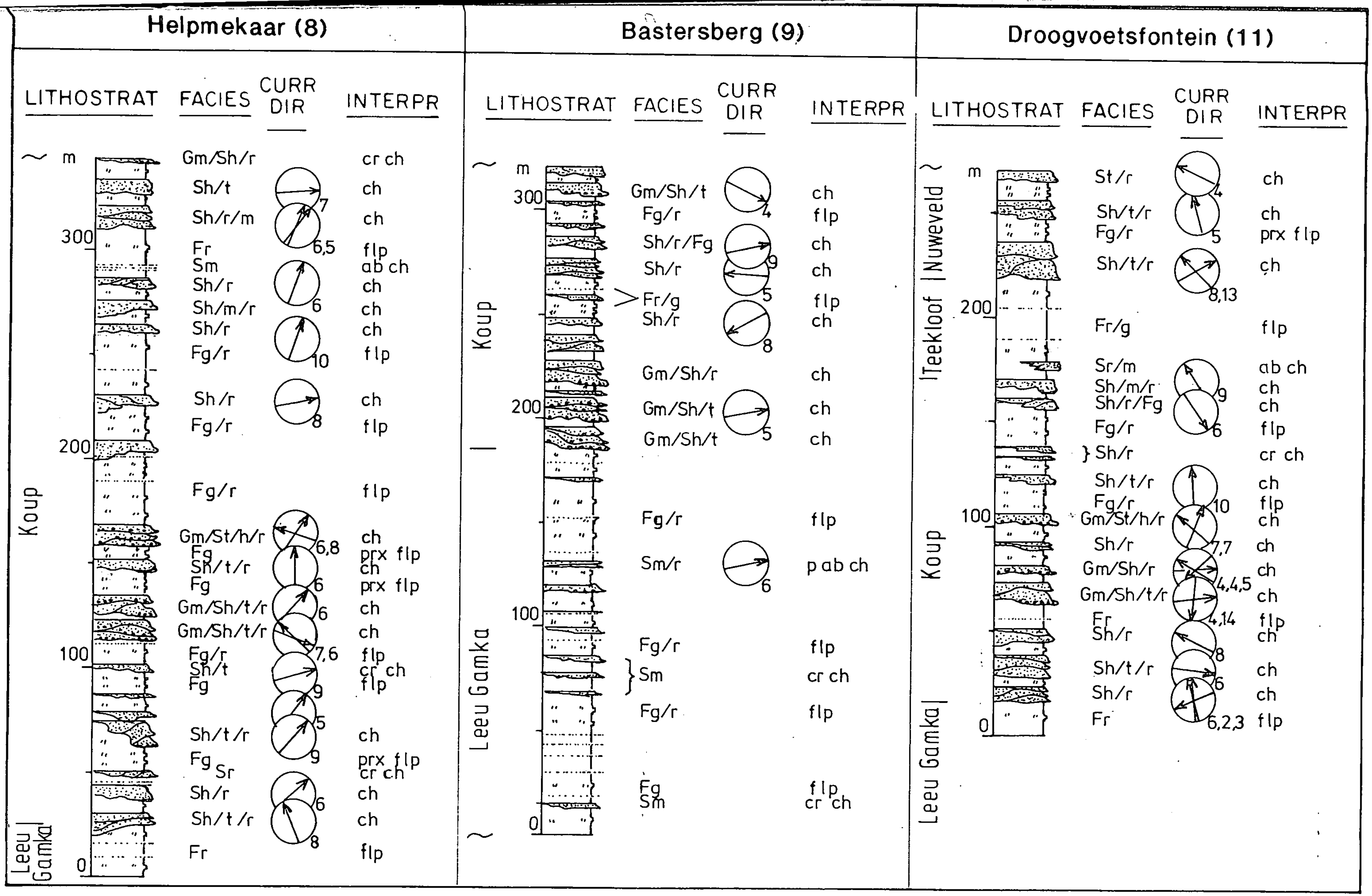
APP. 1.4 Stratigraphic profile measured on Wilgerbosfontein (location 4).



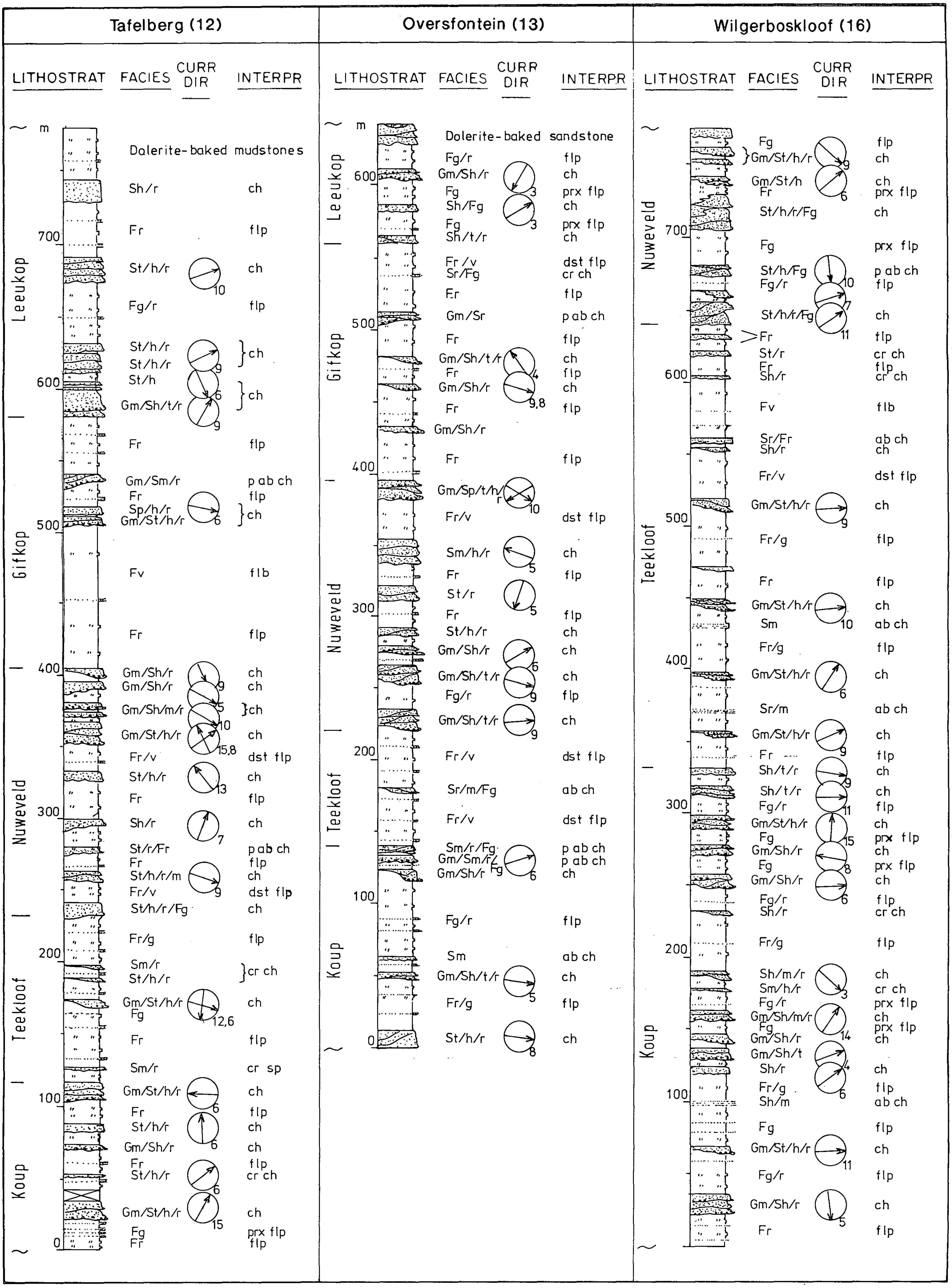
APP. 1.5 Stratigraphic profiles measured in Verlatenkloof (location 5) and on Kareekasberg (location 7).



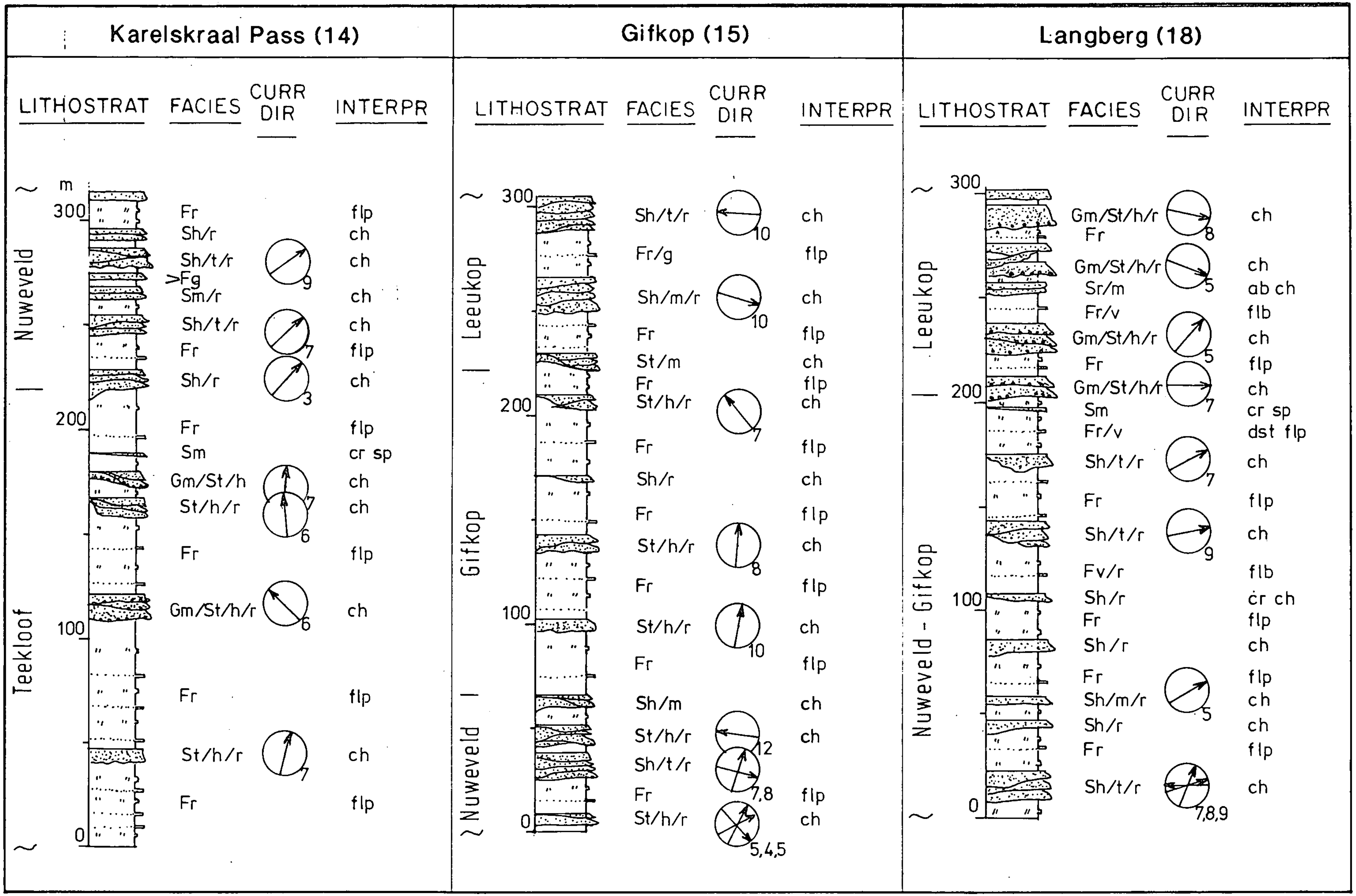
APP. 1.6 Stratigraphic profiles measured in Ouberg Pass (location 6) and on Palmietfontein (location 10).



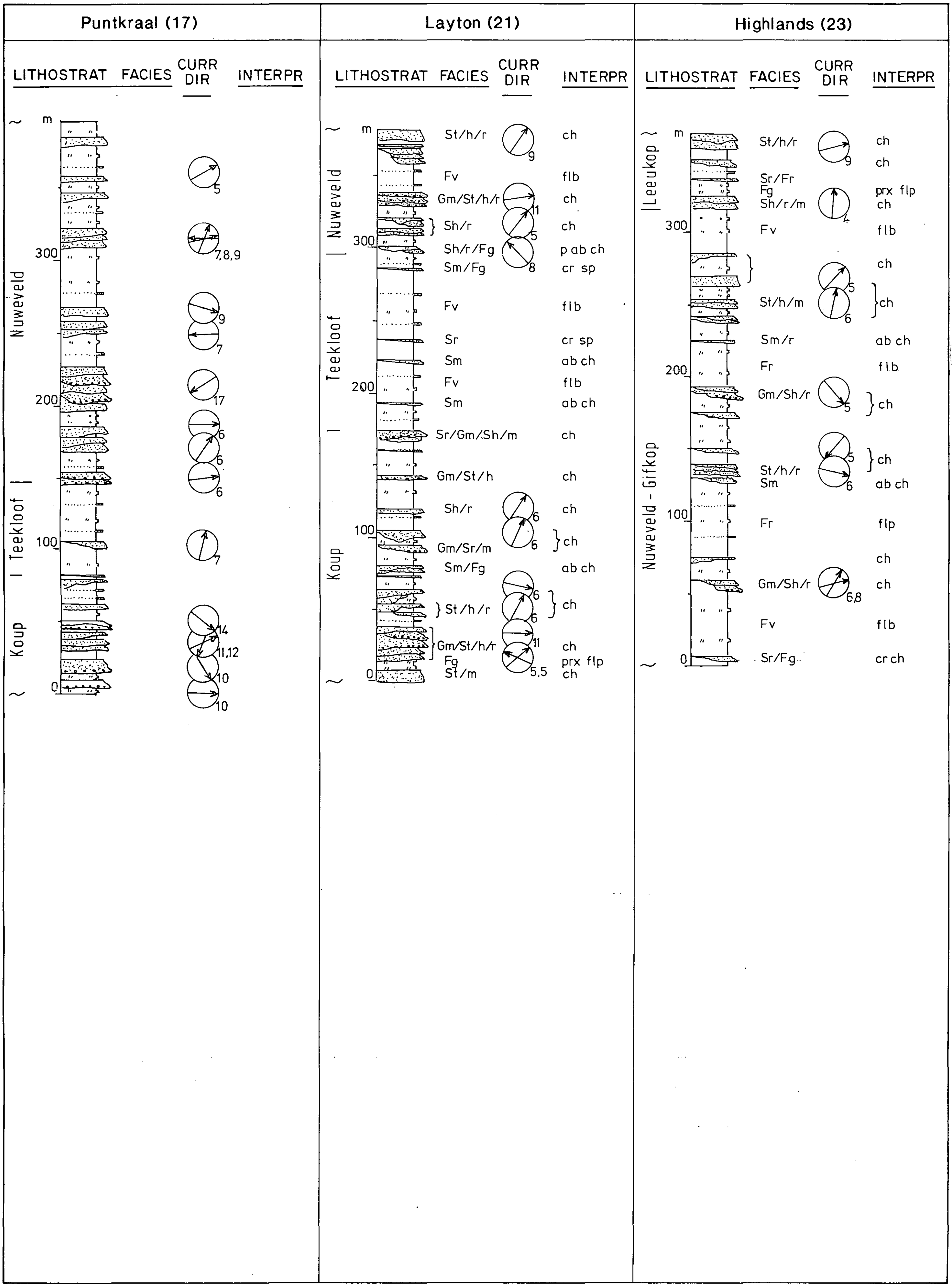
APP. 1.7 *Stratigraphic profiles measured on Helpmekaar (location 8), Bastersberg (location 9), and Droogvoetsfontein (location 11).*



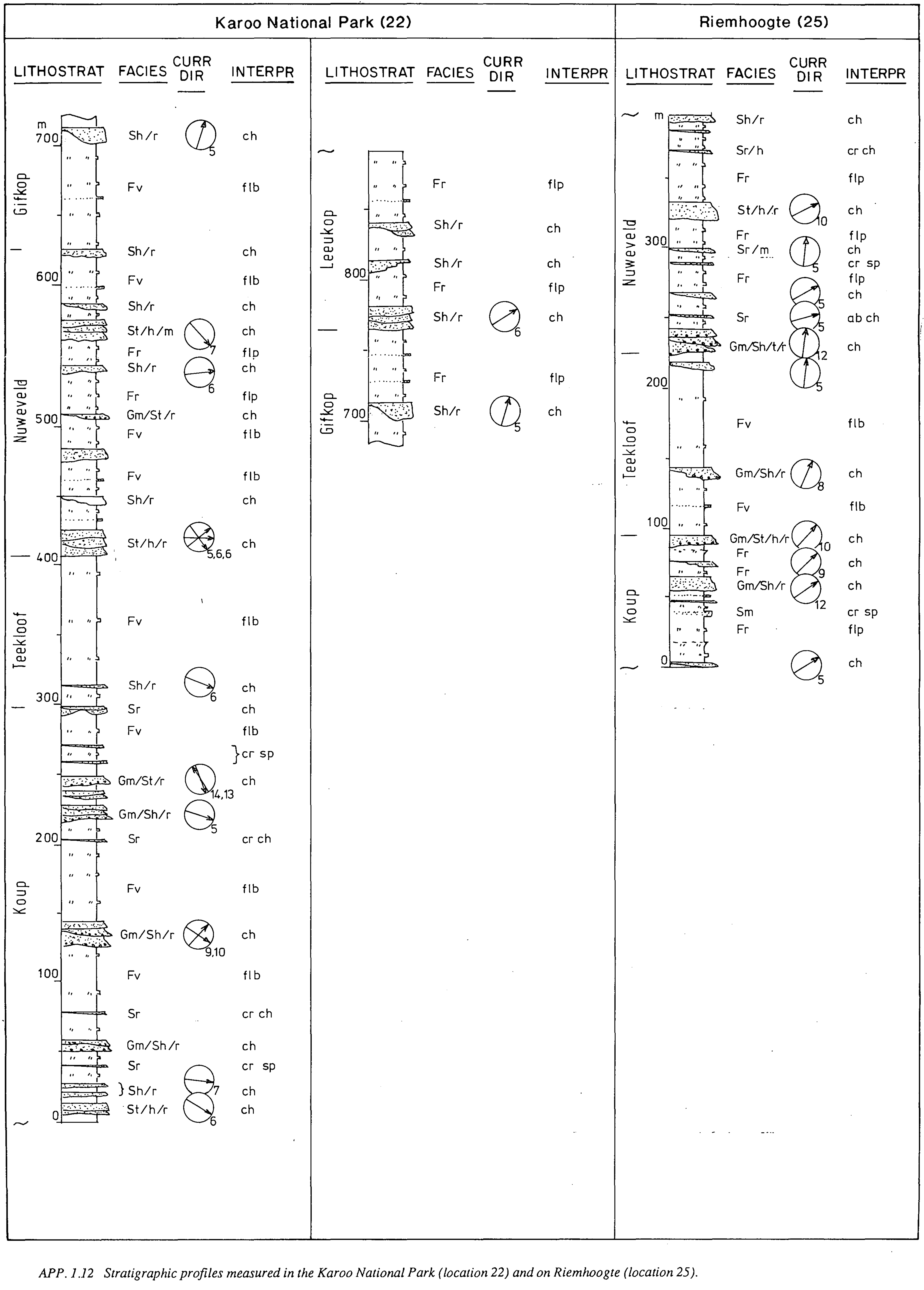
APP. 1.8 *Stratigraphic profiles measured on Tafelberg (location 12), Oversfontein (location 13), and Wilgerboskloof (location 16).*



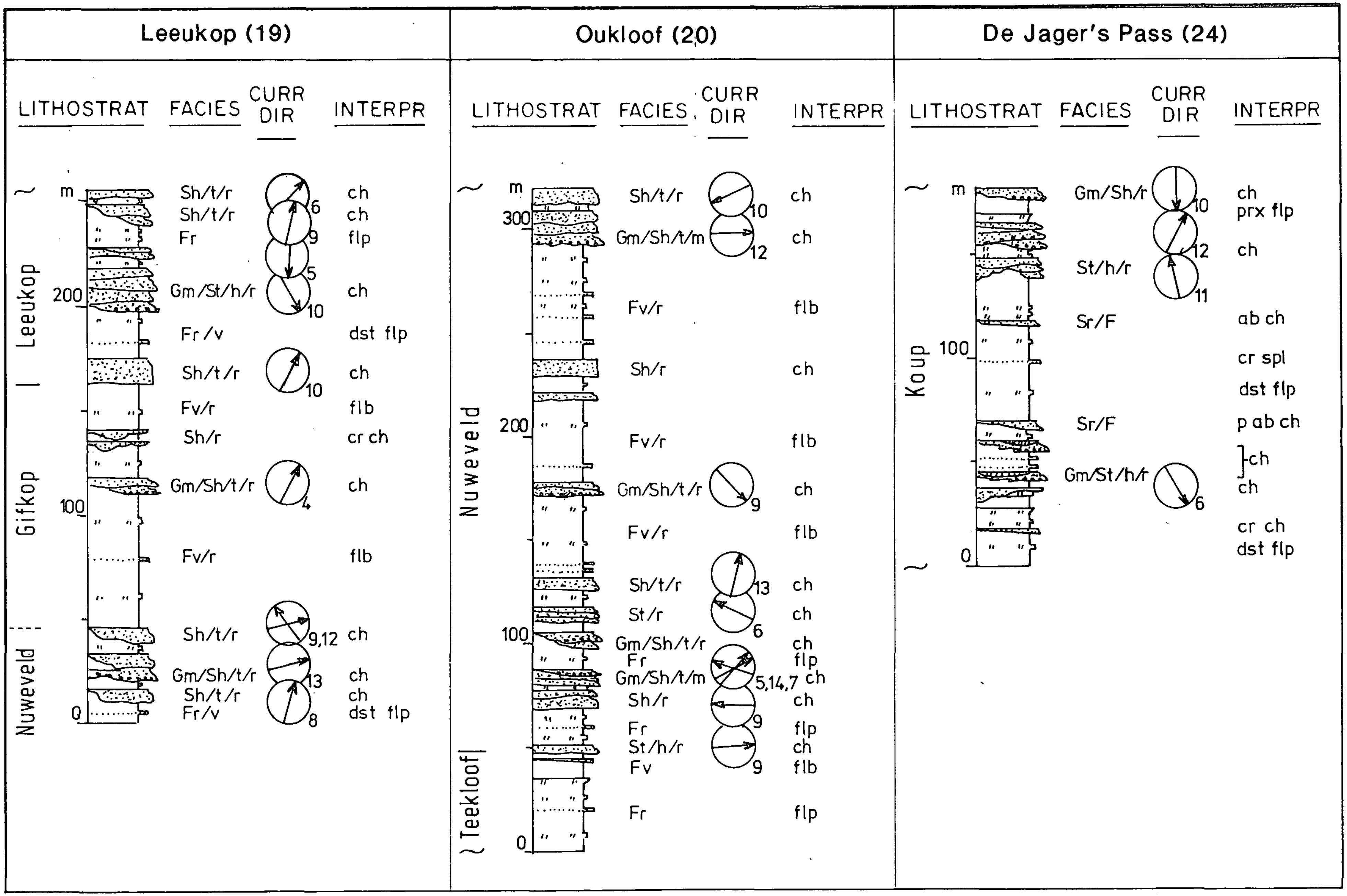
APP. 1.9 Stratigraphic profiles measured on Karelskraal Pass (location 14), Gifkop (location 15), and Langberg (location 18).



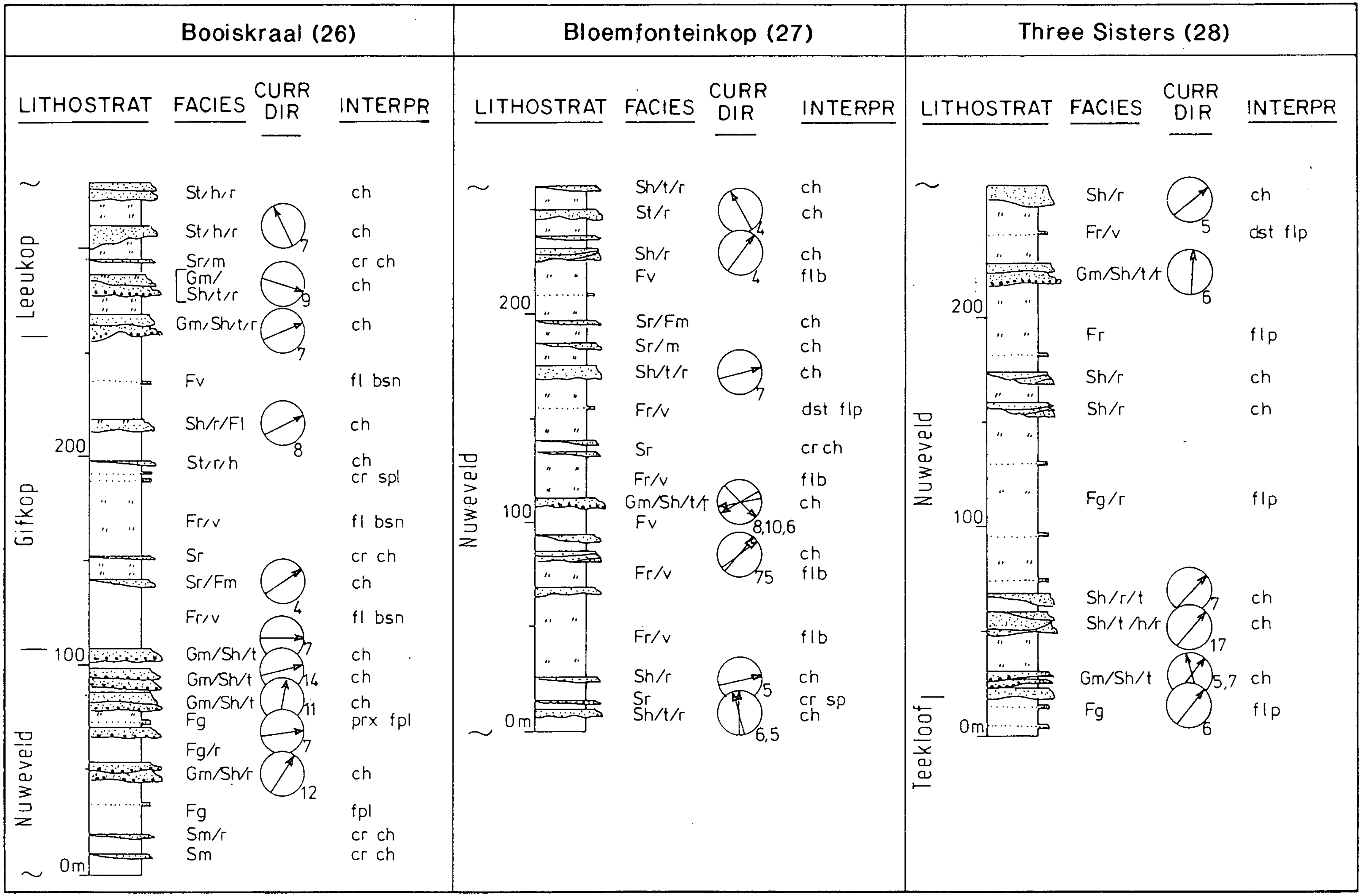
APP. 1.10 Stratigraphic profiles measured on Puntkraal (location 17), Layton (location 21) and Highlands (location 23).



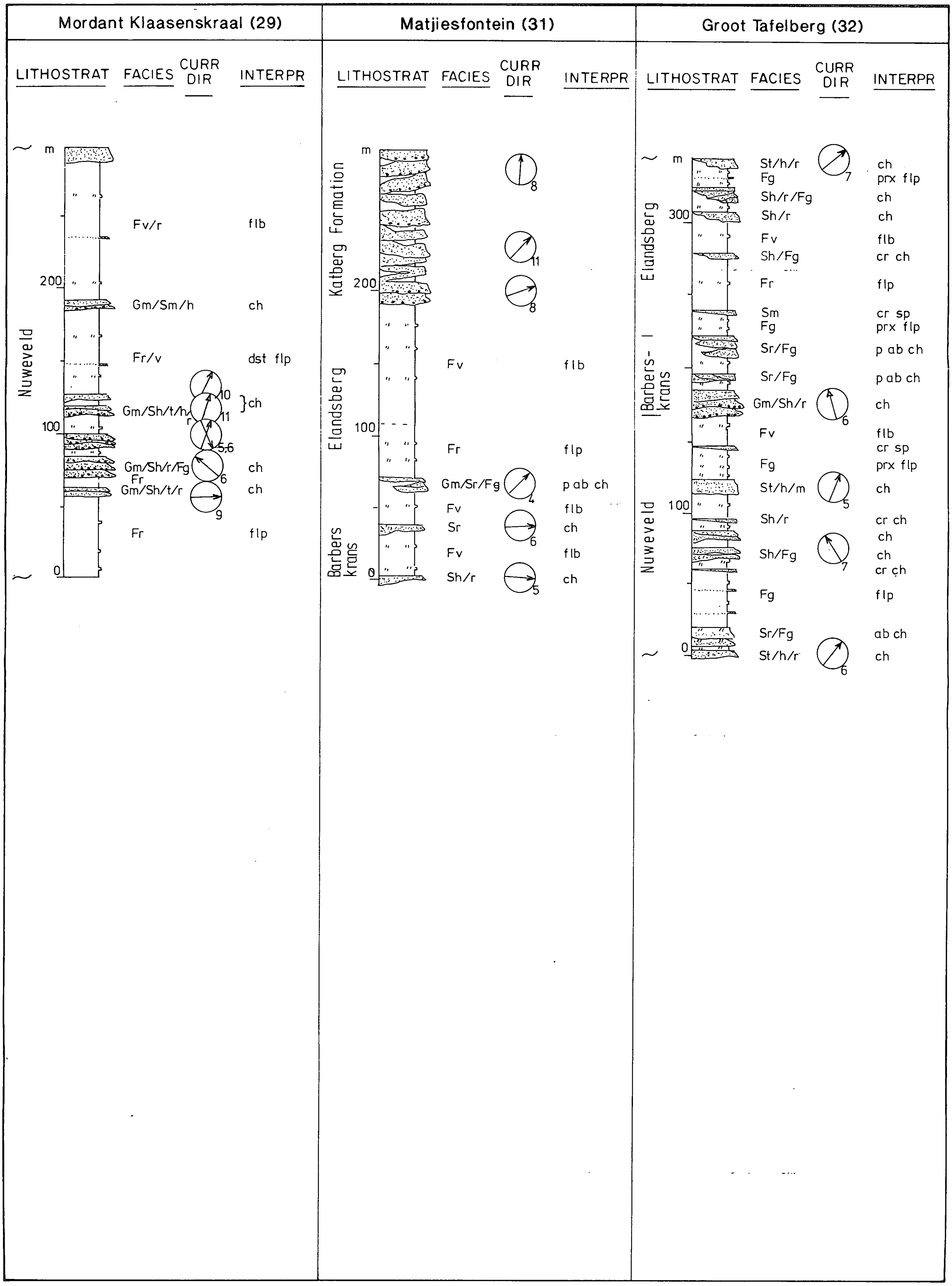
APP. 1.12 Stratigraphic profiles measured in the Karoo National Park (location 22) and on Riemhoogte (location 25).



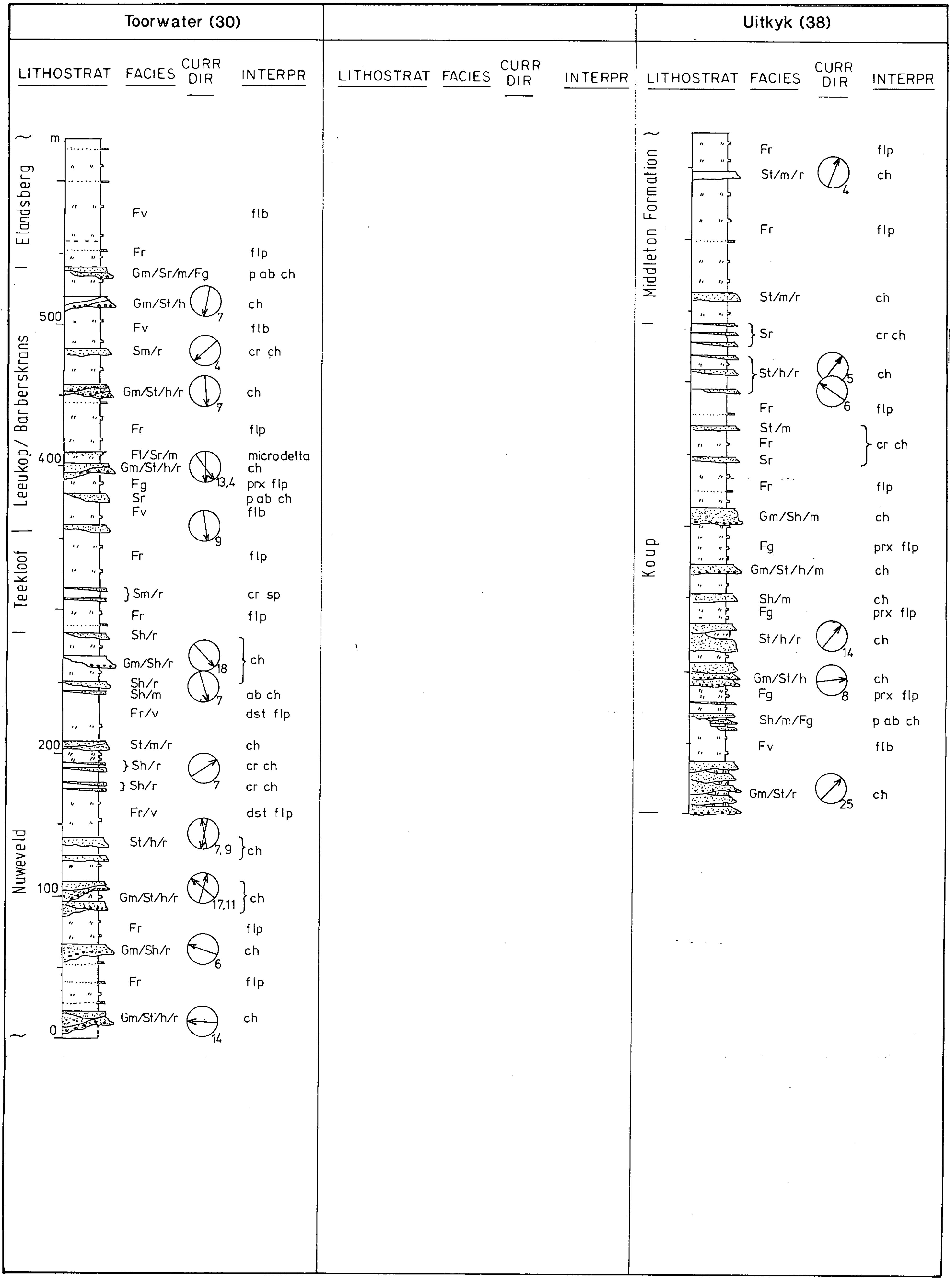
APP. 1.11 *Stratigraphic profiles measured on Leeukop (location 19), in Oukloof (location 20) and in De Jager's Pass (location 24).*



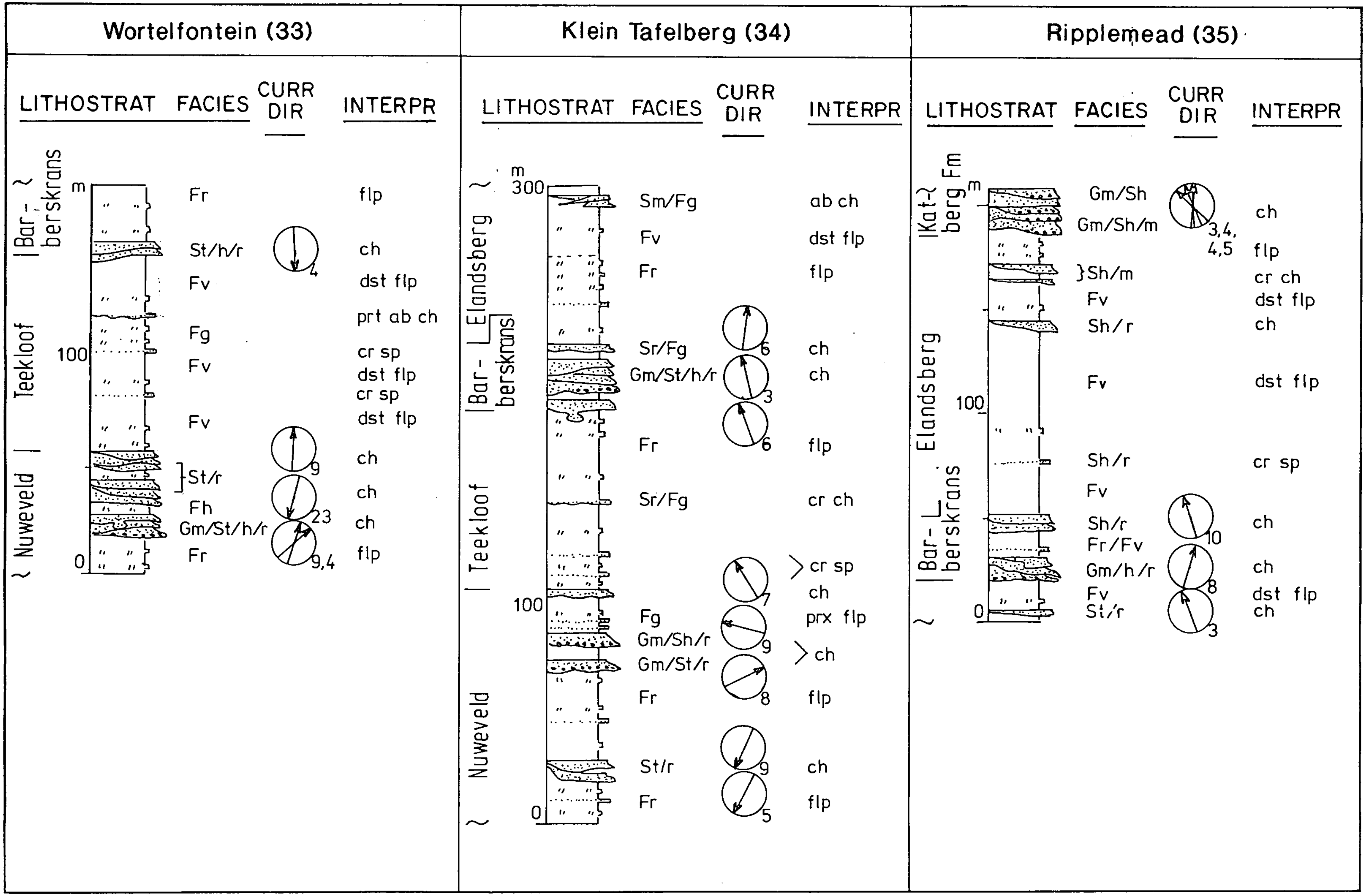
APP. 1.13 Stratigraphic profiles measured on Booiskraal (location 26), Bloemfonteinkop (location 27), and Three Sisters (location 28).



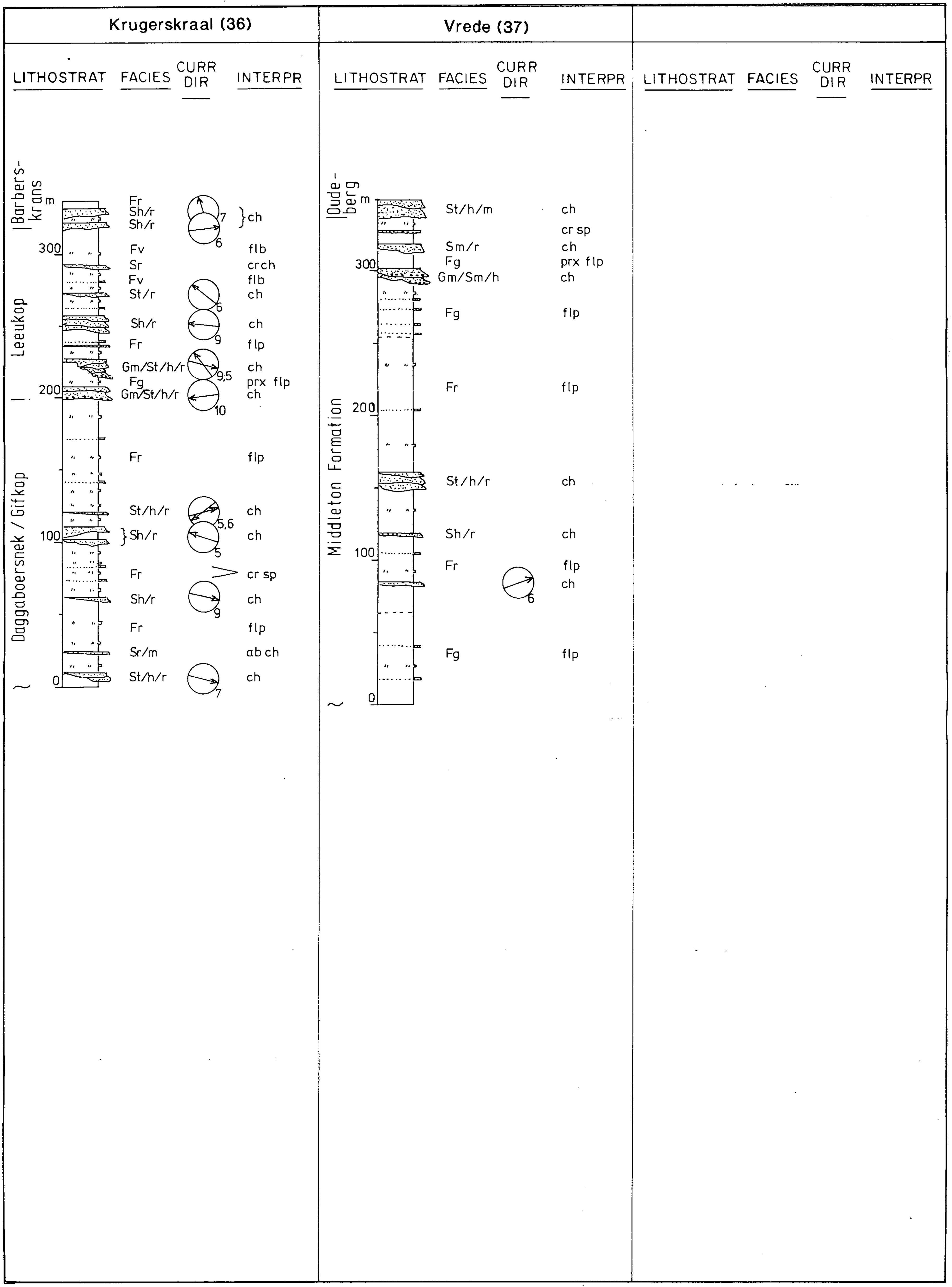
APP. 1.14 Stratigraphic profiles measured on Mordant Klaasenskraal (location 29), Matjiesfontein (location 31), and Groot Tafelberg (location 32).



APP. 1.15 Stratigraphic profiles measured on Toorwater (location 30) and Uitkyk (location 38).



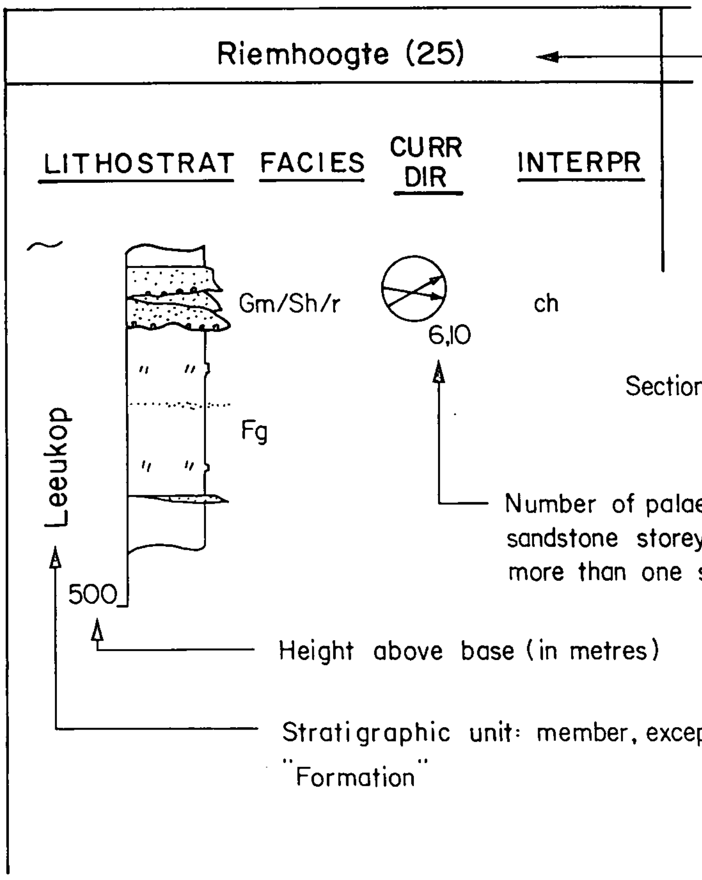
APP. 1.16 *Stratigraphic profiles measured on Wortelfontein (location 33), Klein Tafelberg (location 34), and Ripplemead (location 35).*



APP. 1.17 *Stratigraphic profiles measured on Krugerskraal (location 36) and Vrede (location 37).*

APPENDIX 2 Legend to stratigraphic profiles listed in APP. 1.

LEGEND



Section location and number (see TABLE 1.1)

Number of palaeocurrent readings per sandstone storey (clockwise where more than one set of readings)

Height above base (in metres)

Stratigraphic unit: member, except where specified "Formation"

Abbreviations

LITHOSTRAT	— Graphic log and lithostratigraphic subdivisions
FACIES	— Lithofacies (sequence); see text for explanation of symbols
CURR DIR	— Palaeocurrent vectors for each sandstone storey; True North facing top of page
INTERPR	— Depositional environment – related to facies:

ch	— fluvial channel
(p) ab ch	— (partly) abandoned channel
cr ch	— crevasse channel
cr sp	— crevasse splay
flp	— floodplain (dst= distal; prx= proximal)
flb	— flood basin
md	— microdelta

Lithology

▨	Sandstones (greywackes and arkosic wackes)
▭	Intraformational conglomerates
☐	Mudstones
// //	Siltstones
// // ⋯	Interbedded mudstones, siltstones and thin sandstones
▨	Soft-sediment deformation (mostly loading structures)
☒	Succession not exposed