THE UPPER BRACHINA SUBGROUP : A LATE PRECAMBRIAN INTERTIDAL DELTAIC AND SANDFLAT SEQUENCE IN THE FLINDERS RANGES, SOUTH AUSTRALIA .

(VOLUME I)

by

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LIST OF LOCALITY ABBREVIATIONS

1. SECTION LOCALITIES WITHIN THE STUDY AREA	(Subregion)
ABC – ABC Range (Reference Section)	Central I
AGG – Alligator Gorge (composite, using localities AGM, AGP, AGW)	Western II
ARK – "Arkaba" Station	Central I
ARR – Aroona "Ruins"	Central I
BGL – Black Gap Lookout (in Bunbinyunna Creek)	Central I
BJR – Black Jack Range	Central II
CCR – Chace Range	Central I
DDR – Druid Range	Central I
ELG – Eurilpa Gap	Eastern
EPB – East of Point Bonney (Wilpena Pound)	Central I
IKC – Ilka Creek (near Moralana)	Central I
MCH – Marchant Hill	Eastern
MDG – Middle Gorge (Subsidiary Section)	Western II
MKT – Moockra Tower	Western I
ODW – Oodlawirra	Eastern
PRT – "Partacoona" Station (Willochra Creek-Kanyaka Creek confluence)	Western II
PTG – Petanna Gorge	Western II
RMG – Richman Gap	Western II
SNH – South of "Narinna" Homestead	Central I
SPG – South of Parachilna Gorge	Central I
WKG – "Warrakimbo Gorge" (in Willochra Creek)	Western II
WMY – West of Mount Yappala	Central II
WNG – Warren Gorge	Western II
WWG – Wilkawillina Gorge	Central I

2. OTHER LOCALITIES WITHIN THE STUDY AREA	(Subregion)
AGH – Alligator Gorge: Hancock Lookout	Western II
AGM – Alligator Gorge: Mambray Creek	Western II
AGP – Alligator Gorge: Pine Track	Western II
AGW – Alligator Gorge: near Wilmington	Western II
AMC – Artimore Creek	Central I
AUB – Aubrey Creek (within Pichi Richi Pass)	Western II
BCC – Brachina Creek	Central I
BKG – Buckaringa Gorge	Western II
BNC – Bunyeroo Creek	Central I
BRG – Barunga Gap	Western III
BYR – Bunbinyunna Range (1 and 2)	Central I
CNC – Crow Nest Creek (near Black Jack Range)	Central II
DWS – Dawson	Eastern
EBH – East of "Buckaringa" Homestead	Western II
ECR – East End of Chace Range	Central I
GGC – Gorge Creek	Western II
HMG – Hanniman Gap	Western II
IGG – Ingram Gap	Western III
LCQ – Locheil Quarry	Western III
MLH – "Moralana" Homestead	Central I
MMC – Mernmerna Creek (on "Arkaba" Station)	Central I
MRC – Mary Creek (on "Arkaba" Station)	Central I
MTF – Mount Fergusson (near Port Pirie)	Western III
MTG – Mount Grainger (near Redcliff)	Western II
NAR – North of Aroona "Ruins"	Central I
NBC – North of Brachina Creek	Central I

NBG - North of Bunyeroo Gorge (1 and 2)	Central I
NBR - Nector Brook Range	Western II
NTO - "Narinna" Homestead Turn-Off	Central I
OBM - Oraparinna Barytes Mine (near Wilkawillina Gorge)	Central I
PCG - Parachilna Gorge	Central I
PRL - Prelinna (near Wilpena Pound)	Central I
PRP - Pichi Richi Pass (Saltia)	Western II
RDR - Red Range (between Wilpena Pound and Elder Range)	Central I
RNP - "Rawnsley Park" (near Wilpena Pound)	Central I
RNQ - Ridge North of Quorn	Western II
SAR - South of Aroona "Ruins"	Central I
SBC - South of Brachina Creek	Central I
SBG - South of Bunyeroo Gorge (1, 2 and 3)	Central I
SDC - Sacred Canyon (near Wilpena Pound)	Central I
SWG - South of Warren Gorge	Western II
SWH - South of Wonoka Hill (near Hawker)	Central I
TDM - The Dome (near Marchant Hill)	Eastern
TDP - Third Plain	Central I
UDR - Ulowdna Range	Central I
WCW - "Warcowie" Homestead	Central I
WKC - Waukarie Creek	Western II
WKH - Wonoka Hill (near Hawker)	Central I
WPC - Wilpena Creek	Central I
WSF - Woolshed Flat (within Pichi Richi Pass)	Western II

3. LOCALITIES OUTSIDE THE STUDY AREA

A. <u>Eyre Peninsula</u>

PTL - Point Lowly

SXB - Spinifex Bluff

B. Northern Flinders Ranges

CBC - Chambers Creek

MBR - Mount Bayley Range

PPG - Puttapa Gap

PTH - Patsy Hill

C. Mount Lofty Ranges and the Fleurieu Peninsula

HLC - Hallett Cove

MNR - Marino Rocks

OSB - O'Sullivan's Beach

SHC - South of Hallett Cove

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STATEMENT OF ORIGINALITY

This thesis contains no material which has been accepted for the award of any other degree or diploma in any University, nor, to the best of my knowledge, does it contain any material previously published or written by any other person, except where due reference is made in the text.

P.S.Plummer.

ABSTRACT

The stratigraphy of the late Precambrian upper Brachina Subgroup has been studied in detail throughout the southern and central Flinders Ranges of South Australia. Ten stratigraphically significant facies associations are readily recogniseable within which 18 separate and distinct lithotypes have been defined and described. The complex regional stratigraphic arrangement has been simplified by using a Markov Chain technique of analysis. The resultant lithotype stratigraphy is used as the base upon which the palaeogeographic history of the upper Brachina Subgroup is reconstructed.

A detailed sedimentologic analysis of each lithotype was undertaken in order to ascertain their individual palaeoenvironments of deposition. This involved a petrologic analysis of the arenaceous component of each lithotype, the analysis of the suite of sedimentary structures contained within each lithotype, and the analysis of all directional structures for palaeocurrent directions. For this latter analysis a new computer technique was developed whereby up to 3 individual populations can be separately analysed from any one distribution.

Deposition of the upper Brachina Subgroup succession was due to a phase of uplift tectonism and minor accompanying basic volcanism. Within this succession two distinct depositional episodes are readily discernable. During the first episode a massive sand influx flowed from a westerly source region (the Gawler Craton) into a shallow submerged, though possibly tidally influenced mudflat as a prograding deltaic succession (the "Alligator River Delta"). This initial delta developed in the western region of the Adelaide 'Geosyncline' as a fluvial and tide modified, wave-dominated system which was fed by stable outlet channels, protected by barrier-bars and surrounded by a low intertidal aerobic

mudflat. Preserved within this mudflat deposit are the probable body fossils of primitive cup-shaped coelenterates(?), which were possibly the ancestral organisms of the Ediacara assemblage. With continued sediment influx and basin shallowing, this initial delta system evolved to an unbarred fluvial modified, tide-dominated delta which was fed by migrating channels and surrounded by an intertidal mudflat. This mudflat was anaerobic, possibly due to the activity of abundant microscopic organisms.

The second depositional episode of the upper Brachina Subgroup developed when tectonic instability affected a portion of the basin's western margin (Uplift I). As a result, part of the previously deposited deltaic succession was eroded and reworked into a vast, thin intertidal sandflat which extended through the central region, and into the northern region of the Adelaide 'Geosyncline'. A second phase of tectonic instability (Uplift II) caused renewed activity along the basin's western margin, and also induced the emergence of at least two islands within the basin. Around these islands a thin, dominantly fluvial deposit was generated. The final phase of tectonic instability (Uplift III) affected only the western margin of the basin, and produced a narrow sand deposit of probable beach origin. Meanwhile, within the basin gradual subsidence induced the development of a shallow, possibly tidal aerobic mudflat and marked the end of the upper Brachina Subgroup phase of sedimentation.

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Mr Philip I. Leppard of the Statistics Department of this campus is thankfully acknowledged for programming the computer technique of palaeocurrent analysis, whilst Crosby Sensitizing are thanked for their reproductions of the stratigraphic columns found in the back pocket.

All samples collected from National Parks were done so under permits issued by the National Parks and Wildlife Service, whilst aid in establishing many of the section and sample localities was gained from aerial photographs supplied by the Department of Lands.

Lastly I wish to sincerely thank two people whose influence extended far beyond that required of them by the individual tasks they performed for this thesis. They are my sister, Mrs Sue Gibson, whose nimble fingers produced the typescript you are now reading, and Miss Priya Aiyar, who proof-read the entire text.

DEDICATION

This thesis is dedicated to my parents in the hope that it is at least partial repayment for their 25 years faithful support.

CHAPTER 1

INTRODUCTION

Within the Flinders Ranges of South Australia two extensive quartzitic units of late Precambrian age are well exposed and provide excellent opportunities for detailed study in an endeavour to resolve the many enigmas surrounding their origin. World-wide attention afforded to the Ediacara fossil assemblage (Glaessner and Wade 1966) induced Goldring and Curnow (1967) and Ford (1974) to undertake stratigraphic and sedimentologic studies of the youngest of these two quartzitic units - the Pound Subgroup (Jenkins 1975) - in the southern and central Flinders Ranges. The older of the two quartzitic units - the ABC Range Quartzite - however, remained unstudied and, along with its associated finer-grained facies, has become the centre of this thesis; the first major regional stratigraphic and sedimentologic study afforded to any rock unit within the Adelaidean succession..

THE QUARTZITE PROBLEM

Immense expanses of thick, mature to supermature deposits of sand typify regions of Precambrian to Early Palaeozoic sediment accumulation bearing close affinity to stable cratonic areas. Lithification of these sands by infiltration of siliceous fluids and subsequent cementation led Tieje (1921, p.655) to define such rocks as orthoquartzites - in opposition to 'paraquartzites' which originated through contact metamorphism. This term was popularised by Krynine (1941, 1945, 1948) and subsequently re-defined by Pettijohn (1957) to include all sandstones consisting of 95% or more detrital quartz. Gradually the contracted term quartzite gained common petrologic useage in sedimentary circles for "sandstones, cemented with silica which has grown in optical continuity around each detrital quartz grain" (Pettijohn *et al*. 1972, p.169). Confusion can arise, however, between such a sedimentary quartzite and "a granulose metamorphic rock, representing a recrystallized sandstone, consisting predominantly of quartz" (Holmes 1920, p.194). Thus, for reasons of clarity and to be in keeping with current trends in the literature, the term quartzite is used within this thesis as a 'field term' to distinguish extremely well indurated silica-rich rocks "which fracture through rather than around the constituent grains" (Pettijohn *et al.* 1972, p.169) from less well indurated sandstones, whilst the term quartzarenite is used in the petrologic sense to define sandstones "without appreciable matrix" that contain less than 5% feldspar and/or rock fragments (Pettijohn *et al.* 1972, p.169, after Williams *et al.* 1954, p.292). Other major rock terms used petrologically, based on the varying amounts of feldspar and/or rock fragments present, are given in figure 2-6.

Examples of extensive quartzitic deposits abound within the literature - *e.g.* the Baraboo Quartzite of Wisconsin (Brett 1955); the Tuscarora Quartzite of Pennsylvania and New Jersey (Folk 1960); the St. Peter Sandstone of the Upper Mississippi Valley (Dake 1921, Thiel 1935); the Table Mountain Group of South Africa (Visser 1974); the Lower Fine-Grained Quartzite of Scotland (Klein 1970a); the Eriboll Sandstone of Scotland (Swett *et al.* 1971); and the Pound Subgroup of South Australia (Ford 1974) to name but a few. Although such deposits do not appear to be forming at the present time, the theories pertaining to their palaeogeographic significance have been formulated on the basis of current environments producing mature sands.

Four factors hold the key to the understanding of these problematic quartzitic units, namely:-

(i) their almost monomineralic content;

(ii) their textural maturity - the component sand grains display well rounded surfaces and are well, to very well sorted;

(iii) their uniformity in composition and contained sedimentary structures throughout their (often elongate) sheet-like distributions; and

(iv) their cratonic derivation and depositional affinity to cratonic margins.

These four salient factors have induced many theories regarding the genesis of such quartz-rich deposits, including:-

(i) a series of marine advances and retreats involving periodic exposure of littoral sands to aeolian conditions (Thiel 1935);

(ii) intense chemical weathering of peneplained regions and the development of vast dune fields prior to deposition in shallow seas (Krynine 1941);

(iii) multicycle origins involving the reworking of pre-existing sand deposits (Krynine 1941);

(iv) desert deposition leading to the elimination of the clay fraction (Krynine 1942), producing the necessary grain roundness (Kuenen 1960) and developing the often present bimodal texture (Folk 1968);

(v) prolonged existence within the surf zone (Folk 1960);

(vi) lengthy transportation in river systems and/or along geosynclinal axes (Ketner 1968); and

(vii) the time of moon capture by the earth's gravitational field when the earth-moon distance was much less than at present, causing tidal velocities and ranges to be much greater, and consequently producing extensive sand waves in the littoral environment (Merrifield and Lamar 1968).

Such theories alone, however, have proved to be inadequate explanations of these quartzitic deposits, and their depositional environments and mechanisms of production remained enigmatic until the last decade.

The acceptance that a majority of the Precambrian and Early Palaeozoic quartzitic deposits are of marine origin - either proved by trace and/or body fossils (Seilacher 1968, Ford 1974) or facies associations - coupled with the abundant work being completed on recent tidal flat environments in Western Europe (e.g. Curray 1969, Klein 1967b, Klein and Sanders 1964, de Raaf and Boersma 1971, Reineck 1963, 1967, 1972, Reineck and Singh 1975, Reineck and Wunderlich 1967, van Straaten 1954a, 1954b, 1961, Wunderlich 1970), England (e.g. Evans 1965), the U.S.A. (e.g. Anderson 1973, Otvos 1965) and Nova Scotia (e.g. Balazs and Klein 1972, Klein 1963, 1966, 1967a, 1967b, 1970b) have led to the re-interpretation of ancient quartzitic deposits in Scotland (e.g. Anderton 1976, Klein 1970a, Sutton and Watson 1963, Swett *et al.* 1971), Ireland (e.g. de Raaf *et al.*1977), Norway (e.g. Banks 1973, Johnson 1975), the U.S.A. (e.g. Barnes and Klein 1975, Flores 1977, Ginsberg 1975, James and Oaks 1977, Klein 1975, Oaks *et al.* 1977) Canada (e.g. Wood 1973, Jansa 1975, Speer 1978), India (e.g. Chanda and Bhattacharyya 1974), South Africa (e.g. von Brunn 1974, von Brunn and Hobday 1976, Button and Vos 1977, Eriksson 1977, Tankard and Hobday 1977, Visser 1974, Vos and Eriksson 1977) and Australia (e.g. Draper 1977, Ford 1974) which has shed new light on this *terra incognita*.

This study is centred upon the palaeogeographic interpretation of the late Precambrian ABC Range Quartzite (Adelaide Supergroup) within the Flinders Ranges of South Australia.

GEOGRAPHIC SETTING

The highland belt of South Australia, extending from Kangaroo Island and the Fleurieu Peninsula, through the Mount Lofty, Flinders and Willouran Ranges to the Peake and Denison Ranges - a distance of 850km between latitudes 28°S and 36°S - is geologically known as the *Adelaide 'Geosyncline'* (Mawson and Sprigg 1950, after Browne; *see* David 1950, pp.693-694 and Fig. 201a-c; see Fig. 1-1 of this thesis).

The magnificance of this region is not governed, as in most highland regions of the world, by elevation, for St. Mary's Peak forms its cupola at a mere 1190m. It is the topographic expression, controlled by the underlying geology and varying with latitudinal changes in climate, which produces its scenic grandeur. In the Mount Lofty Ranges deep chemical weathering, induced by moderate rainfall and a temperate climate, leaves only scant outcrop upon the thickly vegetated, rounded hills, through which flow perennial streams. Northward, as the climate becomes more arid, mechanical weathering has attacked the broad, openly folded framework of the Flinders and Willouran Ranges. Here intermittent streams, generally displaying gallery forests of River Red Gums (Eucalyptus camaldulensis), dissect the resistant topography of massive quartzites, limestones and intervening shales, producing many hogback ranges (e.g. the ABC Range), pounds (e.g. Wilpena Pound) and awe inspiring gorges (e.g. Middle, Brachina and Parachilna Gorges) as they wind their way to the evaporitic drainage basins which encompass the northern extension of the ranges in a horseshow fashion. A poor soil cover and shallow weathering profile allows good outcrop and enables the bedrock lithology to strongly influence the vegetation distribution. Native pine (Callitris) and tussock grasses are found to be characteristic of shale units, whilst mallee eucalypts typify dolomitic horizons and porcupine grass (Triodia), commonly known as 'spinifex', blackboys (Xanthorrhoea) and curray bush (Cassinia) cover the quartzitic ridges.

HISTORIC REVIEW

Geologic comment during the last century regarding the numerous impressive quartzitic ranges of the highland belt of South Australia was restricted to isolated notes written mostly by Government surveyors (see Frome 1843, Burr 1846). One exception was Austin (1863) who, during a reconnaissance survey of the mines of South Australia, wrote of "a succession of curious hills ... called the ABC Range from an idea that the number of separate hills is the same as the letters of the alphabet." The quartzite forming this range was named the ABC Range Quartzite by Mawson (1939).

By the turn of the century interest in the geologic nature of the highland region as a whole was on the increase. Woolnough (1904) provided the first petrographic descriptions of thin-sections, including the (?)ABC Range Quartzite equivalent at Sellick's Hill south of Adelaide. Meanwhile Howchin (1907) assigned the sequence of purple shales and overlying quartzites above the Brighton Limestone to the Upper Cambrian (see Table 1-1) and, because of the red colouration and abundant crossbedding within the quartzites, ascribed them to an aeolian environment. David (1922) revised Howchin's allocation of these sediments to the Upper Cambrian by tentatively defining "all the strata from the base of the Archaeocyathinae limestones [Wilkawillina Limestone of the Hawker Group] to the basal conglomerates overlying the Archaean(?) schistose rocks of Aldgate in the Adelaide region" as the Adelaide Series and suggested "that they be classed, provisionally, as Proterozoic(?)." This revised definition was generally accepted, although contention over the age of the uppermost unit - the Pound Quartzite - was rife for the next 40 years.

Meanwhile, Mawson and Sprigg (1950) applied this time-based nomenclature to the sediments within the Adelaide 'Geosyncline', and incorporated the ABC Range Quartzite and underlying purple shales into the Marinoan Series of their Adelaide System (see Table 1-1). This time-based nomenclature was later abandoned, however, in favour of a lithostratigraphic nomenclature to overcome the lack of biostratigraphic control and likelihood of rock unit diachroneity (see Daily 1963), the ABC Range Quartzite being placed within the Wilpena Group (Dalgarno and Johnson 1964) of the Adelaide Supergroup (see Table 1-1).

The sequence measured and described by Mawson (1939) in Brachina

Creek in the central Flinders Ranges is now used as the type section for the Wilpena Group. Within this section Mawson designated the name ABC Range Quartzite to the "1,060 feet of sandy flags and quartzite of the ABC Range". In 1941 he formally recognised the existence within this unit of an uppermost arenaceous member (390 feet in Brachina Creek according to Dalgarno and Johnson 1964) and a lower member of "flaggy slate with some hard bands of sandstone". In their definition of the Wilpena Group Dalgarno and Johnson redefined the ABC Range Quartzite "to the 390 feet thick orthoquartzite at the top of Mawson's unit" and incorporated the lower "flaggy slate with some hard bands of sandstone" into the underlying Brachina Formation. Leeson (1970) recognised the existence of three members within the Brachina Formation in the northern Flinders Ranges. Plummer (1978: see Appendix VA of this thesis) reviewed the entire basal portion of the Wilpena Group (Nuccaleena Formation to ABC Range Quartzite inclusive) and classified these sediments on the basis of their lithostratigraphic relationships throughout the Adelaide 'Geosyncline'. The persistent sequence present enabled the definition of five formations within this horizon, whilst the facies interdigitation, both within and between these formations, necessitated their amalgamation into a new subgroup - the Brachina Subgroup. The nomenclatures of Mawson (1939, 1941), Dalgarno and Johnson (1964), Leeson (1970) and Plummer (1978) defining the basal portion of the Wilpena Group are presented in figure 1-2.

Notable contributions regarding the palaeoenvironmental significance of units within the Wilpena Group, other than Howchin (1907), were made by Mawson and Segnit (1948), who undertook a chemical and petrologic study of Howchin's "chocolate shale series" (the Bunyeroo Formation). Their interpretation of a terrestrial loessial accumulation of fine volcanic debris was based on the overall areal distribution, the angular shape of the component grains, and the presence of elements such as calcium, sodium and potassium in proportions higher than normally found in sediments that have been subjected to normal weathering conditions. Thomson (1969b) concluded that within the Brachina Formation the change from red to green colouration eastward (into the Ulupa Formation) was attributable "to an increase in the depth of water during sedimentation and the consequent absence of oxidizing conditions". The well sorted, current bedded quartzitic deposits were ascribed by Sprigg (1952) to continental terrace sedimentation during times of marginal stability, opposing the subaerial origin earlier proposed by Howchin. Thomson (1969b), concurring with Sprigg, described the eastwardly thinning ABC Range Quartzite as "a marginal shelf sediment...clearly derived from a western source area."

GEOLOGIC SETTING

Rifting and subsidence of the Adelaide 'Geosyncline' occurred between protracted lines of basement weakness - e.g. the Torrens Lineament, Paralana and Mount Nor'west Faults. Alderman (1973) writes of a connecting basin stretching northward from the Peake and Denison Ranges forming a great arc which curved westward around the stable cratonic Musgrave Block and extended through the Amadeus Basin into Western Australia. The basin's characteristics were, however, considerably modified in those areas and the thickness of sediments deposited was much less than in the region of the Mount Lofty and Flinders Ranges. Within the Adelaide 'Geosyncline' "subsidence and concominant sedimentation" (Sprigg 1952) during the Proterozoic and Early Palaeozoic allowed the accumulation of shallow to marginal marine sediments to an average thickness of 13 km (Schermerhorn 1974). Hypersubsidence (*sensu* Kamen-Kaye 1967) was, however, common throughout with the maximum thickness recorded being

33.5 km (Coats 1973). The 'geosyncline' was favourably compared by Sprigg (1952) with the 'miogeosyncline' of Kay (1947) because of the general lack of contemporaneous volcanism and the implied prolonged environmental stability during sedimentation. According to figures 1 and 2 of Sprigg (1952), however, the 'geosyncline' has much in common with a 'miogeocline' (*sensu* Stewart and Poole 1974) in which thick shallow water deposition occurred on a shelf outwardly bordered by an open ocean.

BASEMENT

The 'geosyncline' and its associated stable shelf regions are flanked to the west and northeast by a number of outcropping ancient cratonic nuclei (Fig. 1-1). Areally the most significant is the *Gawler Block* (Thomson 1969a). Here a phase of metamorphism and mountain building in early Carpentarian times (1700-1800 m.y. - the Kimban Phase) produced a basement of metasediments (quartzites, mica schists, amphibolitic rocks, dolomites and marble), migmatites, granite gneisses, and intrusive granites and basics - collectively forming the Cleve Metamorphics. A period of fluviatile to shallow marine sedimentation ensued depositing quartzites and conglomerates with heavy mineral and tuffaceous layers. Acid volcanic plug emplacement heralded the beginning of the Charlestonian Phase of tectonism, dominated by granitic intrusions dated at 1590±30 m.y. (Compston *et al.* 1966) and terminated by a series of gabbroic intrusions. A second thick conglomeratic sequence followed prior to the porphyritic, rhyolitic and tuffaceous extrusives of the Gawler Range Volcanics. Finally, at approximately 1475 m.y., the Wartakan Phase of tectonism and acid igneous intrusion cratonized the Gawler Block.

Glen *et al.* (1977) correlate the Gawler Block with the shield promontory spanning the South Australia-New South Wales border - the

Willyama Block (Campana and King 1958, Talbot 1967, Thomson 1969a). Within the high grade metamorphic succession of sillimanite garnet gneisses - associated with basic sills and plugs - and granitoid gneisses, and the lower grade sericitic schists, metasediments, amphibolites and iron formations, is recorded a complex series of tectonic events. As with the Gawler Block, early metamorphism and folding was followed by a series of igneous events - the entire cycle being known here as the Olarian Orogeny. The silver-lead-zinc deposits of Broken Hill - dated at 1675 m.y. - appear to be closely associated with this metamorphism. Granites and pegmatites, both preceded and succeeded by basic igneous intrusions, occurred at 1520 m.y. and are correlated with the Gawler Range Volcanics.

Other less major basement blocks include:-

(i) the Mount Painter Province (Thomson 1969a, Coats and Blissett 1971), now known to be part of the "inferred Curnamona Nucleus" (Thomson 1976). Here a series of metasediments are intruded by granites of the Terrapinnan Phase of tectonism, tentatively equated with the Charlestonian Phase on the Gawler Block;

(ii) the Peake and Denison Inliers (Reyner 1955, Thomson 1966, 1969a) - comprising metasediments, amphibolitic basics, migmatites and granitic gneisses preserved in graben structures;

(iii) the Houghton Inlier (Spry 1951, Cooper and Compston 1971);

(iv) the Anabama Inlier (Thomson 1969a); and

(v) numerous rafts within 'diapiric' structures scattered throughout the Flinders and Willouran Ranges (Mount (1975).

The localities of these basement blocks are shown on figure 1-1.

THE ADELAIDE 'GEOSYNCLINE'

The attainment of reliable commencement dates of Adelaidean sedimentation has been hampered by the general lack of biostratigraphic control.

The Roopena Volcanics - flanking and occurring within the rift zone - have provided the most widely quoted minimum basement age of 1340 m.y. (Thomson 1966). Coupled with this is a dating from the Cultana Granite of the Gawler Block at 1320 m.y. (Compston et al. 1966). However, recent work on the Houghton Inlier by Cooper and Compston (1971) and Cooper (1975) give an age of 867±32 m.y. for the final phase of basement metamorphism. Cooper (1975) also suggests that the Callanna Beds in the Mount Painter region are less than 900 m.y. old, which corresponds well with the 850±50 m.y. age obtained from the Wooltana Volcanics (Compston et al.1966) which lie within the Callanna Beds in the same region. Stratigraphically equivalent to the Wooltana Volcanics, however, are the newly defined Beda Volcanics (Mason et al. 1978), from which Webb and Hörr (1978) obtained a date of 697±70 m.y. It is thus suggested that Adelaidean sedimentation within the rift commenced approximately 700 to 800 m.y. ago, and continued with both major and minor breaks in the succession until the Middle Cambrian. Stabilization of the rift zone occurred during the phase of tectonism and metamorphism known as the Delamerian Orogeny (Mawson 1942), the final phase of which is placed in the Middle Ordovician at 465 m.y. by Compston et al. (1966).

STRUCTURE

The Adelaide 'Geosyncline' is readily divisible into three major structural zones. The largest of these zones, running from Kangaroo Island and the Fleurieu Peninsula (the Fleurieu Arc), through the Mount Lofty Ranges to the Southern Flinders Ranges (the Nackara Arc) and then eastward towards Broken Hill (the Olary Arc), is known as the South-Central Zone (Tipper and Finney 1965). This zone displays fold and fault trends which parallel its overall outline. Structural trends scribing a concave northward arc from the Mount Painter Block westward into the Willouran Ranges define the Copley Arc (Sprigg 1964), whilst wedged between these two is the third zone, the Northern Zone. Skirting the outer circumference of the Nackara Arc (i.e. Chace and Druid Ranges), is a zone of confluence between the geologic composition and topographic expression of the Northern Zone and the structural trends of the South-Central Zone known as the Outer Arc. Plummer (1978) noted a close correlation between various lithologic sequences through the Brachina Subgroup and the zone within which they were deposited. He suggested that the present day expression of the Adelaide Fold Belt is a direct reflection of the original structural morphology of the 'geosyncline'. Looking at the basin as a whole, he simplified the structural division to that shown on figure 1-3.

Within the 'geosyncline' three mechanisms of deformation have operated from the time of initial rifting and subsidence to the present day. Vertical movements along protracted lines of crustal weakness initiated rifting and sedimentation within the trough. Later, basement taphrogenesis produced contemporaneous faulting during lower Sturtian and Sturtian-Marinoan times (Binks 1971, Plummer and Gostin 1976) causing "sudden thickening, facies changes and overlap of sediments" (Coats 1962). As sediment accumulation continued gravitational foundering induced décollement deformation and the rising of 'diapiric' structures (Pierce 1966, Rutland and Murrell 1975). Often around these structures were small growth faults which display facies changes and thickness variations across them. These two styles of deformation - viz. vertical and décollement - reached maximum intensity in the Flinders and Willouran Ranges (ie. the central and northern regions on figure 1-3).

The third style of deformation - viz. lateral compression - accompanied the Delamerian Orogeny and affected the entire 'geosyncline'. This lateral compression by the enframing cratonic nuclei endowed upon the sedimentary pile its concentric fold pattern typified by narrow appressed anticlines (into which 'diapiric' structures often flowed - e.g. Arkaba 'diapir') and broad open synclines. An increase in intensity of folding occurs toward the Fleurieu Arc where syntectonic granites - the Palmer and Victor Harbour Granites - were intruded in the Late Cambrian to Early Ordovician (White et al. 1967, Wopfner et al. 1969, Milnes et al. 1977).

Despite the intensity and pervasiveness of these deformations the present day morphologic expression of the fold belt is largely due to Tertiary taphrogenesis. Accompanying this phase of uplift, reflecting deep-seated lateral basement compression, was broad and gentle arching which gained intensity toward the marginal areas of the ranges.

METAMORPHISM

Metamorphism of the cover rocks of the Adelaide Fold Belt occurred during the Delamerian Orogeny. Although described as generally unmetamorphosed (Thomson 1965, p.270) extensive chloritization, sericitization and re-crystallization of quartz are characteristic (Offler and Fleming 1968). A common feature possessed by the thick shale and siltstone units within the sedimentary pile is the change in colouration from red to green. Grim (1951) attributes such changes to the growth of metamorphic chlorite, although sedimentary facies changes from oxidized to unoxidized environments is commonly thought responsible (e.g. Thomson 1969b, p.74; Plummer 1978; see also Chapter 5 of this thesis). Zonation of metamorphic intensity decreases from the regions of exposed high grade metamorphic rocks at Palmer (east of Adelaide) and Mount Painter, where granites emplaced during the Delamerian Orogeny appear responsible. The final phase of metamorphism is placed at 465 m.y. by Compston et al. (1966).

SCOPE OF STUDY

This study is primarily concerned with a rigorous sedimentologic analysis of the late Precambrian ABC Range Quartzite throughout nearly 28,000 sq. km of the Flinders Ranges of South Australia. Because, however, this quartzite is spatially and temporally related to two finer-grained units, all three have been dealt with stratigraphically and sedimentologically in the regional synthesis, whilst only the quartzite has been studied in detail petrologically, for reasons outlined in Chapter 2. These data, coupled with those available from modern day environments, are used to establish a palaeogeographic model of origin for the sequence in the light of sedimentation and tectonism within, and adjacent to the Adelaide 'Geosyncline'.

CHAPTER 2

METHODOLOGY

Field data from a total of 77 localities were collected from approximately 28,000 sq. km. of the southern and central Flinders Ranges, the main area of study. These localities are shown on figure 1-1. In addition, a further 4 localities were visited in the northern Flinders Ranges, 4 in the Mount Lofty Ranges and on the Fleurieu Peninsula, and 2 on northern Eyre Peninsula to ensure the regional application of the upper Brachina Subgroup formation and facies stratigraphy. All locality abbreviations used are listed in the thesis preamble, and given on the fold-out at the back of each volume.

Within the southern and central Flinders Ranges 24 stratigraphic sections were measured in detail - including the reference and subsidiary sections of the Brachina Subgroup - and detailed mapping of lithotypes. was carried out in 3 stratigraphically important areas. Wherever possible, areas complicated by faulting and structural closures were avoided for the measurement of a stratigraphic section or readings of palaeocurrent directions. Such areas were, however, studied with the view of facies determination and spot samples were taken in an endeavour to obtain an integrated coverage of the entire area.

STRATIGRAPHIC SECTION MEASUREMENT

The measurement of stratigraphic sections was accomplished with a compass and 30m tape along traverses trending as near as practicable perpendicular to the strike of the beds, with a minimum number of offsets. Where possible, sections were measured along creek beds because of their greater percentage of outcrop than on the flanking hills, and also for their ease of access. Section localities were selected with the aid of aerial photographs which enabled an *a priori* determination of the most suitable creeks, regarding outcrop and accessibility, along with the added facility of establishing good coverage of the entire study area.

Observations were not restricted solely to the creek bed during the measurement of each section. All observations made were plotted directly onto a Bouma-like scroll at a scale of 1:500 allowing a resolution of 1m on the resultant stratigraphic columns. However, because of the great uniformity within facies - unless studied bed by bed on a centimetre scale - all stratigraphic sections (which range from 130 to 1750m thick) are presented on a scale of 1:7750 (see Appendix I). The reference and subsidiary sections are also presented at a scale of 1:1000 for greater clarity. Throughout each section field notes were made in parallel with the section scroll of the sedimentary structures present and their abundance, along with palaeocurrent directions and the structures from which they were taken.

ANALYTICAL TECHNIQUES

Palaeoenvironmental determinations of unfossiliferous sequences, such as the late Precambrian upper Brachina Subgroup, require a good knowledge of the environmental significance of the sedimentary structures contained within each facies, and a thorough analysis of palaeocurrent data. These results, coupled with the detailed analysis of the vertical and lateral facies associations and a petrologic analysis of each lithotype, lead to a palaeogeographic reconstruction of the sequence - the ultimate aim of many stratigraphic and sedimentologic studies.

FACIES ANALYSIS

Following the regional correlation of the upper Brachina Subgroup from the 77 localities visited throughout the study area, the sequence has been subdivided into 10 facies associations, comprising 18 separate and distinct lithotypes (see Chapter 3 for lithotype descriptions). Facies association 1 constitutes the Moorillah Formation and contains 2 lithotypes, whilst facies association 2 forms the Bayley Range Formation with only one lithotype. The ABC Range Quartzite is composed of facies associations 3 to 10 inclusive, within which are recognised 15 lithotypes (see Table 2-1).

In an endeavour to determine both the vertical and lateral stratigraphy of facies associations within the upper Brachina Subgroup 39 localities were selected where the complete sequence was known, and the lithotype transitions from each locality were plotted onto an 'observed lithotype transition flow diagram' (Fig. 2-1). This lithotype transition flow diagram, however, reveals a system too complex for all the major transitions to stand out, and too intricate to be of use in developing a meaningful facies association stratigraphy.

To overcome this problem such a system can be analysed by a Markov Chain technique which reduces the data to the important lithotype transitions throughout the sequence. Such methods are used increasingly in facies analysis studies, and examples are common in the literature (e.g. Gingerich 1969, Harms et al. 1975, Miall 1973, Selley 1970, Schwartzacher 1975, Turner 1974). The technique used here is based on that outlined by Harms et al. (1975, pp.69-73).

The data from the 'observed lithotype transition flow diagram' (Fig. 2-1) is transposed into an 'observed lithotype transition matrix' (t_{ij}), wherein the number of times facies i is overlain by facies j is given by t_{ij} (see Table 2-2). The row and column totals (t_i and t_j, where $\sum_{i=1}^{m} t_i = \sum_{j=1}^{n} t_j = p$: the total number of transitions) are then calculated and utilized to determine an 'expected lithotype transition matrix' (e_{ij}) (Table 2-3) using the equation $(t_i x t_j)/n$. This matrix represents a random lithotype sequence arrangement. By then subtracting the 'expected lithotype transition matrix' from the 'observed lithotype transition matrix' (i.e. $t_{ij}-e_{ij}$) a 'difference matrix' (d_{ij}) (Table 2-4) is determined. This matrix comprises positive terms where the observed lithotype transition occurs more frequently than in the expected random arrangement, and negative terms where the observed lithotype transition occurs less frequently than in the expected random arrangement.

It is the positive terms that are considered when developing models of facies relationships. These terms are viewed subjectively in an endeavour to overcome inconsistencies produced by such natural biasses as the occurrence of both cyclic and non-cyclic lithotype transitions within the same sequence, and the variation in areal extent of individual lithotypes and their subsequent probability of representation within any one section. The resultant 'major lithotype transition flow diagram' interpreted from the difference matrix (Fig.2-2) presents a model of both vertical and lateral development of lithotypes within the sequence which can be used, along with the sedimentologic, palaeocurrent and petrologic data of each lithotype, to develop a meaningful model of palaeogeographic history for the entire sequence.

PALAEOCURRENT ANALYSIS

As early as 1857 Sorby recognised that palaeocurrents, extricated from fossilized sedimentary structures, formed a major link between the aspects of stratigraphy, sedimentology and tectonics of an association of sedimentary rocks and the palaeogeographic history of that association. Sedimentary structures displaying palaeocurrent directions abound throughout the upper Brachina Subgroup. In all, measurements were taken from 2184 sedimentary structures at 46 outcrop localities in an endeavour to determine the palaeocurrent system of the sequence. As many readings as possible were taken off each type of structure at each locality, as, in general, at least 25 readings are believed to be necessary for meaningful analysis (Potter and Pettijohn 1963, p.263). Of these, 87.5% were from the ABC

Range Quartzite, 10% from the Bayley Range Formation and the remaining 2.5% from the Moorillah Formation. Readings from each type of sedimentary structure have been analysed separately to account for the structure hierarchy concept of Allen (1966) and Miall (1974) where, with an increase in the rank of the structure, there occurs a corresponding increase in the variance of the resultant palaeocurrent direction. The sedimentary structures analysed from each lithotype are listed in table 2-5 whilst the interpreted flow regimes are given in table 2-6.

Crossbedding formed the basis of the palaeocurrent analysis - 79% of all readings and 71% of the analyses. Several styles of cross-bedding are represented throughout the sequence, (Fig. 2-3; see Allen 1963 for crossbedding classification). Omikron crossbedding (Fig. 2-3A - formed by the migration of straight crested megaripples) is by far the most common, and dominates the ABC Range Quartzite. The average thickness of each set is approximately 13cm, with the average foreset slope being 23°. Minor set thinning and slope shallowing occurs toward the centre of the basin away from the region of sediment debouchment. Pi crossbedding (Fig. 2-3B - the result of repeated channel scouring and filling or the migration of lunate megaripples) is common within restricted zones of the ABC Range Quartzite, whilst alpha, beta and zeta crossbedding (Figs. 2-3C, 3D and 3E) are rare. Sands within the Bayley Range Formation are typified by gamma crossbedding (Fig. 2-3F), whilst xi crossbedding (Fig. 2-3G) is present toward the base of the Moorillah Formation. Beta crossbedding is also seen in both these latter formations, but only rarely.

Ripple cross-lamination (generally mu cross stratification - Fig. 2-3H) and straight asymmetric ripple marks are common throughout the entire sequence, comprising 14% of all directional readings and 17% of the analyses. Rare exposures displaying excellent lambda or kappa cross-stratification (Figs. 2-3I and 3J) are seen in the Bayley Range Formation. Exposures of ripple marks that reveal no asymmetry or cross-lamination, but show persistent crestal trends (ie. symmetric ripple marks) make up 2.3% of all readings and 4% of the analyses. Parting lineation (4% and 7% of the respective totals) and flute marks (0.3% and 1% respectively) complete the analyses. Readings from slump structures, eddy scours, groove marks and imbricated clasts comprised only 0.4% of all readings and were not statistically analysed.

Collection of Data. Palaeocurrent data were only collected from outcrops where unambiguous directions were determinable. Crossbedding azimuth directions were obtained by holding a stiff board parallel with the foreset face - this being either exposed or readily determined from its traces within the outcrop - upon which the direction of the current was marked. The board was then rotated parallel to bedding and correction for tectonic tilt was then accomplished, in the field, by restoring the board to the horizontal by rotation about the line parallel to the strike of the bedding. A corrected measurement of the current direction was then recorded between 000° and 360°. Readings from other uni-directional structures (e.g. ripple cross-lamination and flute marks) and bi-directional structures (e.g. symmetric ripple marks and parting lineation) were obtained by the same general procedure, although allowance for the lack of directional sense in the latter case being made in the analyses.

Analytical Technique. Many environments of sedimentation have a complexity of current systems acting upon their deposits. As a consequence, readings of palaeocurrent directions taken from preserved sedimentary structures often produce multimodal distributions. For palaeocurrent data, therefore, to assist in basin analysis, and the resolution of such local problems as the relation between individual directional sedimentary structures and the overall geometry of the containing lithologic unit, each distinct mode within the distribution must be analysed separately (as noted by Tanner 1959, p.221).

The palaeocurrent data from the lithotypes within the upper Brachina Subgroup display multimodal distributions. An analytical technique for such palaeocurrent data has been developed in collaboration with Mr. P.I. Leppard (Department of Statistics, The University of Adelaide) whereby mixtures of up to three populations within any one distribution can be analysed for mean direction, concentration about the mean, and proportion of each population to the entire distribution using ungrouped data statistics (see Plummer and Leppard in press). In the rare cases where quadrimodal distributions were present, the data was split into two bimodal distributions. Each of these distributions was analysed separately producing useful mean values, although the values obtained for the concentrations about each mean and each population proportion were not valid to the quadrimodal distribution.

Palaeocurrent data have been defined as circular normal distributions (see Pincus 1953, Mardia 1972) which can be expressed by the density function

$$f(x:\ p_i,\ u_i,\ k_i) = \frac{1}{2\pi} \sum_{i=i}^{m} \frac{p_i \exp(\cos(x-u_i))}{I_o(k_i)}$$

where m represents the number of populations;

p_i represents the proportion of population i in the mixed distribution, (where $\sum_{i=1}^{m} p_i=1$);

u_i represents the mean direction of population i, (where $0 < u_i \leqslant 2\pi$ radians);

k_i represents the concentration of population i about u_i, (where $k_i > 0$); and

I_o (.) is the modified Bessel function of the first kind and zero order.

The analytical technique is written in FORTRAN for the University of Adelaide's CYBER CDC 173 computer, and comprises two programmes. A full listing is given in Appendix II. Input data for the first programme consists of a header card for identification of each analysis, which is printed out at the top of the output; following this is a 'segment card' which selects the desired number of segments the distribution is divided into, starting from north and proceeding in a clockwise direction - limited to 1 (where 15^o segments are required), 2 (20^o segments), 3 (30^o segments) or 4 (45^o segments)[1] punched in the first column; this is then followed by the raw data of palaeocurrent directions. The raw data is punched, in degrees, in columns 9-80 on an I4 format, leaving columns 1-8 for data identification. This programme then groups the data into the chosen number of segments and produces a circular frequency display (Fig. 2-4A). From this display the user then estimates the number, and approximate position(s) of the population(s) within the distribution (Fig. 2-4B). This information is punched on a 'selector card' as 1's, 2's and 3's (in the first 8, 12, 18 or 24 columns of the card, depending upon the number of segments chosen) which defines - in a clockwise direction starting from north - which segments belong to which population. This selector card is placed between the segment card and the raw data cards, and is used by the second programme for initial starting values of the unknown parameters p_i, u_i, and k_i, for

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[1] All palaeocurrent analyses in this thesis are done using twelve 30^o segments because the author feels that any fewer divisions does not allow enough discrimination between populations, whilst with more divisions local irregularities within populations tend to be isolated as separate 'populations'. The option, however, of the number of segments the distribution is divided into was included in the programme to account for the personal preference of other workers who may use the technique at a later date (<u>see</u> Plummer and Leppard <u>in press</u>).

$i=1...m$. These values are then used by the programme to maximize the likelihood function

$$L = \prod_{j=1}^{n} f(x_j : p_i, u_i, k_i), \qquad \text{for } i=1...m,$$

(where n is the total number of readings) for p_i, u_i and k_i using routine ZXPOWL on the IMSL (1975) library of FORTRAN subroutines. Large sample estimates of variance of each parameter estimate are also calculated, along with a chi-square goodness-of-fit test of the model. Output of this programme includes circular frequency displays of both the observed data and the estimated circular normal distribution, along with a printout of the estimated values of p_i, u_i, k_i and the chi-square test (Fig. 2-5). Values of the chi-square test less than 0.05 are taken to indicate that the data differs significantly from the selected model of best fit.

The results of the palaeocurrent analysis are presented in Appendix III.

PETROLOGIC ANALYSIS

Throughout each stratigraphic section and at the majority of other localities, representative hand specimens were collected from each component lithotype of the upper Brachina Subgroup. All specimens were numbered and marked with dip and strike before being removed from their *in situ* position. All specimens cited in this thesis are held within the Geology Department of Adelaide University under accession number 469. The prefix TS refers to a thin section, whilst no prefix refers to a hand specimen. The method of specimen numbering is based on a 3 letter-3 number system. The 3 letters refer to the locality from which the specimen was collected. Of the 3 numbers, the first digit refers to the facies association to which the specimen belongs - except for facies association 10 which is identified by the first two digits of the 4 numbers present. (The identification of individual lithotypes within these facies associations by an extra letter or number was found to be unnecessary as each lithotype is very distinct). The last 2 digits are used to keep the specimens in stratigraphic sequence. For example, specimen 469/MDG511 was collected from Middle Gorge (MDG), facies association 5 (the first digit), and stratigraphically it lies above specimen 469/MDG510 and below specimen 469/MDG512.

Specimens from each lithotype were studied macroscopically, with the aid of a 10x lens and the rock colour chart of Goddard et al. (1963), and described following a modified scheme of Krynine (1948, p.157) using colour, dominant sedimentary structure, subtexture, varietal minerals and main textural name - these descriptions being presented in Chapter 3.

Detailed microscopic analysis of specimens was restricted to the sandy lithologies because, apart from the obvious grain size advantage displayed by arenites over argillites, the depositional texture is better preserved after compaction, diagenesis and metamorphism than in the muddy lithologies. Arenites, therefore, reveal a greater amount of information regarding source areas, depositional processes and environments than do argillites. Because, however, the degree of silica cementation within a majority of the sandy lithologies is such that, upon crushing, constituent grains shattered (thus rendering sieve analysis useless), thin section analysis formed the basis of the petrologic study. All thin sections were cut perpendicular to bedding, and of each thin section the composition was determined for the major components, using the visual percentage estimation charts of Terry and Chilingar (1955, after Shvetson 1954), and of the heavy mineral assemblage. Also visually estimated for each thin section were the general textural characteristics (sorting, grain roundness and sphericity, and grain size distribution) and its mean grain size. According to Friedman (1962), sedimentologic parameters, such as skewness and kurtosis, when estimated from thin sections, bear a poor correlation to the actual values obtained from sieve analysis, and as such were not determined in this study.

Of the major constituents, quartz was by far the most abundant. Of the quartz percentage, the crystallinity and extinction of the individual grains were classified, using a flat microscope stage, according to the scheme: monocrystalline straight extinction (single grains which go to complete extinction within a 5^{o} rotation of the microscope stage); monocrystalline undulose extinction (single grains which only go to part extinction within 5^{o}); and polycrystalline (composite grains). The nature of inclusions and overgrowths, along with grain size (range and mode), shape and roundness were also noted.

The percentage estimate of each constituent was allocated to classes based on the following 6-fold subdivision:-

1: less than 1%;

2: 1 to less than 3%;

3: 3 to less than 10%;

4: 10 to less than 25%;

5: 25 to less than 50%; and

6: 50% and greater.

The use of percentage classes for composition analysis has an added advantage over the use of actual percentage figures. The percentage classes chosen, when plotted onto a ternary QRF diagram, are good approximations to the major standard sandstone subdivisions (see Fig. 2-6). For a rock containing 95% quartz the term quartzarenite is used. The division between an arkose and a subarkose is drawn at the standard value of 25% feldspar, whilst a figure of 10% feldspar is used to distinguish between a subarkose and a feldspathic quartzose arenite. These same percentage values of rock fragments are used to similarly define litharenites, sublitharenites and lithic quartzose arenites, respectively. Using these percentage classes, therefore, instead of occupying a single point within a major region, any one analysis defines that major region. The added advantage of using percentage classes is that the class divisions can be analysed directly by computer techniques (see Fig. 2-6B). The data presented in this thesis are average figures of all specimens from each lithotype at each locality, and these have been diagrammatically summarized using computer assisted principal component analysis, using the programme PRINCA of Fitzgerald (1975), and cluster analysis, using the programme CLASS of Lance and Williams (1967) modified by Fitzgerald (1975). The results of these analyses are presented in Chapter 8.

In addition to this bulk composition analysis, point counting was undertaken upon the tourmaline fraction of the major sandy lithologies from nine stratigraphic sections selected to give a uniform coverage of the entire study area. Tourmaline possesses a great variety in grain size, colour and the nature of inclusions as determined by the mode of origin of each grain. This variety, along with the chemical and mechanical ultrastability of tourmaline, makes it an ideal aid in source rock determination. The analysis involved the counting of every tourmaline grain in each slide of the selected sections (abundance ranged from 0 to 150 grains, but averaged only 15 grains per slide) whereby colour, pleochroism and grain size, along with the nature of grain shape, inclusions and overgrowths were noted and applied to the criteria of source rock identification outlined by Krynine (1946). The results of this analysis are presented in Chapter 8 and Appendix IV.

CHAPTER 3

STRATIGRAPHY OF THE

UPPER BRACHINA SUBGROUP

Lying between the sediments of the Elatina Glaciation (Mawson 1949) and those of Cambrian age within the Adelaide 'Geosyncline' is the dominantly red bed sequence defined as the Wilpena Group (Dalgarno and Johnson 1964; see Table 1-1 of this thesis). This group ranges in thickness from 6500m in the northeastern Flinders Ranges to 3000m in the southern Flinders Ranges, whilst only 1000m is present in the Mount Lofty Ranges and 450m on the Stuart Shelf. Equivalents of this group are found in the northern Officer Basin near the Western Australia border, the Amadeus Basin in Central Australia, and the Barrier Ranges of western New South Wales.

The age of the Wilpena Group can be restricted to within the Vendian between 665±45 m.y. and 570±10 m.y. Stromatolites within the underlying Umberatana Group have been equated with the Late Minyar assemblage of the U.S.S.R. (950±50 m.y. to 680±20 m.y.) by Preiss (1971). Within the East kimberley region of Western Australia marine shales overlying the upper Precambrian Egan glacial deposits (which are equated to the Elatina glacials immediately beneath the Wilpena Group) have been dated at 665±45 m.y. (Compston and Arriens 1968).

Acritachs found within the lower Wilpena Group have been correlated with assemblages of Vendian age (680±20 m.y. to 570±10 m.y.) in the U.S.S.R. (Daily 1976, pers. comm.). Similarly the Ediacara fossil assemblage of the Pound Subgroup at the top of the Wilpena Group has been correlated by Glaessner (1971) with other remains of soft-bodied organisms in South-West Africa, Europe and the U.S.S.R. and classed as being Vendian in age.

Two coarsening-upward sequences comprise the Wilpena Group. The lower sequence - the Brachina Subgroup (Plummer 1978; see Fig. 1-2 of this thesis) - totals 2200m maximum thickness and is readily divisible into two on the basis of gross lithology. The lower Brachina Subgroup comprises a thick shale-siltstone sequence (the Moolooloo Formation) with a massive to laminated dolomite generally at its base (the Nuccaleena Formation). The upper Brachina Subgroup has three component formations (the Moorillah Formation, the Bayley Range Formation and the ABC Range Quartzite) which laterally intertongue with one another (Fig. 1-2).

The basal contact of the upper Brachina Subgroup is gradational to sharp, yet conformable with the Moolooloo Formation (see Chapter 4), whilst local disconformity, or intertonguing and/or gradational conformable contacts separate the upper Brachina Subgroup from the overlying Bunyeroo Formation (upper Wilpena Group; see Chapter 7).

The sequence present within the upper Brachina Subgroup is divisible into 10 facies associations comprising 18 separate and distinct lithotypes. Facies association 1 constitutes the Moorillah Formation and comprises 2 lithotypes, whilst the Bayley Range Formation is represented by facies association 2, comprising one lithotype. The remaining 8 facies associations and 15 lithotypes constitute the ABC Range Quartzite.

REGIONAL STRATIGRAPHIC RELATIONSHIPS

Regional stratigraphy is often complicated by sedimentary facies changes related to spatial variations in depositional environments over any specific time interval. Such changes are common throughout the upper Brachina Subgroup. This fact necessitated the designation of two reference sections (viz. the ABC Range and Middle Gorge sections) when defining the subgroup to include all component facies associations (Plummer 1978). The area covered during this study of the upper Brachina Subgroup is readily divisible into three regions on the basis of gross lithology (Fig. 3-1A). The central and eastern regions are dominated by fine-grained facies associations (i.e. finer than medium sand) - viz. the Moorillah and Bayley Range Formations - with only a thin development of coarse-grained facies associations (ie. medium sand and coarser) - viz. the ABC Range Quartzite - present in the central region (see sections Fig. 3-1B). Coarse-grained facies associations, however, dominate the western region.

Within these regions further division is necessary to account for lithotype variation and intertonguing within the facies associations, resulting in a six-fold subidivision of the study area (Fig. 3-1B). The eastern region remains undivided, whilst the central region is divided into two subregions, and the western region into three. Representative stratigraphic sections are given for each of these subregions, except WIII where no complete section was available for measurement.

Within the eastern region and subregion CI the Bayley Range Formation overlies the Moorillah Formation, whereas in subregion CII it intertongues with, and replaces the Moorillah Formation (see sections Fig. 3-1B; and Fig. 3-2). Subregion WI displays the same Moorillah-Bayley Range Formation succession as does the eastern region, except a thick sequence of ABC Range Quartzite is developed at the top. Westward this quartzite intertongues continually deeper into the succession until the entire sequence is dominated by coarse-grained facies associations in subregion WII (see sections Fig. 3-1B). These regional stratigraphic relationships are all shown on figure 3-2.

Subregion WIII displays only isolated outcrops of the coarse-grained facies associations - here termed the Barunga Sandstone. On the basis of field relationships no direct correlation is possible with any part of the upper Brachina Subgroup. Outcrop appearance, palaeocurrent data and petrologic characteristics, however, suggest correlation of this sandstone with lithotype 5A of the ABC Range Quartzite (see section headed "Lithotype Descriptions - Lithotype 5A"), and on this basis these sandstones from subregion WIII have been included in this study.

The Moorillah Formation, therefore, outcrops in the eastern region, and in subregions CI and WI. Its maximum recorded thickness of 531m occurs in subregion CI (Aroona Ruins section (ARR): see Appendix I, page A5) where it displays a red-purple colouration. A grey-green hue, however, typifies this formation in the eastern region and subregion WI. Although dominantly a shale-siltstone succession similar to that of the underlying Moolooloo Formation, the Moorillah Formation in these subregions is differentiated by the presence of occasional beds of dark greyish red soft-sediment deformed siltstone and sandstone (often tuffaceous) and/or banded dusky red, dusky blue grey and greyish purple, planar crossbedded fine to medium sandstone and quartzite.

Overlying the Moorillah Formation everywhere is the Bayley Range Formation. The maximum measured thickness of this formation is 833m in the eastern region (Marchant Hill section (MCH): see Appendix I, page A6) but minor fold complications may have influenced this figure. Generally the thickness of this formation lies between 300 and 450m. Toward subregion CII, and within subregion WI the Bayley Range Formation intertongues with, and eventually replaces the Moorillah Formation. Along the eastern edge of subregion WII, however, the Bayley Range Formation intertongues with, and is replaced by the ABC Range Quartzite.

As a consequence, the ABC Range Quartzite constitutes the entire upper Brachina Subgroup succession within subregion WII. It is within this subregion that the quartzite attains its greatest thickness, the maximum recorded being 1750m in a possibly incomplete section (Alligator Gorge section (AGG): see Appendix I, page A4). In subregion WI the ABC Range Quartzite forms only the top of the sequence, and here it has thinned to 500m and less, whilst in the central region up to only 100m of quartzite is present to complete the upper Brachina Subgroup. Where the ABC Range Quartzite is absent from the top of the sequence, and the Bayley Range Formation extends to the top of the succession (ie. in the eastern region and along the eastern edge of the subregion CII) quartzitic interbeds, thicker than the average seen in the Bayley Range Formation, are inevitably present between 50 and 100m below the top of the sequence. These interbeds, ranging in total thickness from 5 to 15m (e.g. Druid Range (DDR) and Oodlawirra (ODW) sections:see Appendix I, page A6) represent distal tongues of ABC Range Quartzite, as earlier suggested by Binks (1971).

FACIES ASSOCIATION STRATIGRAPHY

Quite a discriminating stratigraphic subdivision of the upper Brachina Subgroup has been accomplished which is useful both in the field on a macroscopic scale and in the laboratory on a microscopic scale. This stratigraphic subdivision is presented in table 2-1.

The upper Brachina Subgroup is readily divisible into three formations on the basis of gross lithology and colouration. Dominating subregion WII and capping the succession within subregion WI and the central region of the study area are sandy sediments which constitute the ABC Range Quartzite, whilst throughout the eastern region and the lower majority of the succession in subregion WI and the central region are the dominantly fine-grained deposits of the Moorillah and Bayley Range Formations. These latter two formations are readily differentiated on the basis of colouration and diagnostic lithologies, with the Bayley Range Formation exhibiting a grey-green hue and an abundance of pale greenish yellow lenticular sandstone interbeds, whilst the Moorillah Formation, although grey-green in the eastern region and subregion WI (and absent from subregion CII), exhibits a red-purple colouration in subregion CI, and throughout its distribution displays beds of soft-sediment deformed dark greyish red siltstones and lacks the lenticular sandstone interbeds so prevalent in the Bayley Range Formation.

Within these two finer grained formations, which are generally between 300 and 600m thick, the average bed thickness is in the order of 3 to 5cm (though it ranges from thin laminae to thick beds of 1.5m). Because of the uniformity with which all component lithologies are repeated throughout these formations, no subdivision is possible (or necessary), except that based on the regional colour change within the Moorillah Formation. The finer grained sediments of the upper Brachina Subgroup, therefore, which constitute the Moorillah and Bayley Range Formations, are designated facies associations 1 and 2, respectively, for the purposes of facies association analysis. Within facies association 1 lithotypes 1A and 1B are defined on the basis of the regional colour variation from red-purple to grey-green respectively, whilst facies association 2 remains undivided.

Within the coarse-grained succession of the upper Brachina Subgroup (ie. the ABC Range Quartzite) eight major subdivisions are made on the regional scale (defining facies associations 3 to 10 inclusive) based on the presence or absence of lithologic cyclicity, outcrop pattern, colour and general lithology. The latter two of these criteria are also used to distinguish lithotypes within each facies association. Cyclicity is restricted to facies associations 5 (between the quartzites of lithotype 5A and the shales of lithotype 5B) and 7 (between the quartzites of lithotype 7A and the silty sandstones of lithotype 7B), whilst very bold outcrops (reflecting mineralogic homogeneity) characterize facies associations 4 (pink quartzites of lithotype 4A and yellowish grey quartzites of lithotype 4B) and 10 (pale pink quartzites of lithotype 10A and pale red granular quartzites of lithotype 10B). Of the remaining facies associations (viz. 3, 6, 8 and 9) all display red-purple colouration, but they are differentiated by their stratigraphic position and general grain size and texture. Facies association 3 always lies below facies association(s) 4 and/or 5 and comprises medium quartzites (lithotype 3A) and massive to laminated siltstones (lithotype 3B). Both facies associations 6 and 8 lie above facies association 5, but the former comprises rubbly outcropping siltstones whilst the latter consists of coarse sandstones. Facies association 9 lies above facies association 7 and is composed of siltstones and microconglomeratic coarse sandstones (lithotype 9A), shales, and medium sandstones and quartzites (lithotype 9B) and coarse, to very coarse sandstones (lithotype 9C). Facies associations 3, 4, 5, 6, 8 and 10 are all restricted to the western region, whilst facies associations 7 and 9 outcrop within the central and eastern regions.

GENERAL FACIES ASSOCIATION SUCCESSION

The aim behind the establishment of an accurate stratigraphy (both vertically and laterally) for the upper Brachina Subgroup is to interpret from it the environmental conditions responsible for its deposition. The basic 'building block' of the stratigraphy, therefore, should not only be readily recognisable in the field and mappable, but also have discrete palaeoenvironmental significance. Each lithotype defined above by its outcrop pattern, colour and general lithology also displays a unique suite of sedimentary structures which is environmentally diagnostic, and as such renders it palaeoenvironmentally significant.

Because of the regional scale of this study, more than one set of environmental conditions of deposition were active within the study area at any one time. Consequently, when coupled with fluctuations both in and between the rates of sediment input, sediment accumulation and basin subsidence, overlapping and intertonguing of lithotypes often occurred, producing the complex stratigraphic arrangement shown in figures 2-1, 3-2 and 3-3. To simplify this arrangement to the major lithotype succession of the stratigraphy a Markov Chain technique of analysis is applied (see Chapter 2 for method). By then utilizing the lateral lithotype intertonguing to estimate a probable 'time succession' for the sequence, the resultant lithotype arrangement (Fig. 2-2) can be plotted to represent a diagrammatic 'time-related' cross-section of the sequence (Fig. 3-4), upon which the positions of selected time intervals are shown which are used as base lines for the palaeogeographic reconstructions presented in Chapters 4, 5, 6 and 7.

LITHOTYPE DESCRIPTIONS

The most fundamental aspects of sedimentology are mineralogy, texture and sedimentary structures. Because, however, the upper Brachina Subgroup is dominated by supermature sandy lithotypes, neither the mineralogy nor the texture of these lithotypes alone reveals much in regard to the environments of deposition. The varied suites of sedimentary structures of each lithotype, on the other hand, are the main criteria by which depositional environments are determined. As such, sedimentary structures are discussed in Chapters 4, 5, 6 and 7, where they are directly related to palaeocurrent data and the palaeoenvironmental interpretations.

These lithotype descriptions, therefore, include only the areal distribution of each facies association and/or lithotype, its thickness and outcrop pattern, mineralogy and texture, and dominant sedimentary structure where applicable. The bed thickness classification used is that of Ingram (1954, p.938, Table 2; i.e. less than 0.3cm is thinly laminated; 0.3 to 1.0cm is laminated; 1 to 3cm is very thinly bedded; 3 to 10cm is thinly bedded; 10 to 30cm is medium bedded; 30 to 100cm is thickly bedded and greater than 100cm is very thickly bedded). The facies association name is designed for ready identification in the field using colour (based on the colour chart of Goddard et al. 1963) and a grain size-based genetic name (using the sandstone (-1 to 4Ø), siltstone (4 to 8Ø) and shale (8 to 12Ø) divisions of the modified Wentworth grade scale (see Dunbar and Rogers 1957, p.161). Here the term sandstone refers to a rock which breaks around its constituent grains, whilst the term quartzite is used for one which breaks across its constituent grains). The more complete lithotype name follows a modified version of the scheme defined by Krynine (1948, p.157), using the descriptive sequence of colour (Goddard et al. 1963), dominant sedimentary structure, subtexture (using the subdivisions of the major grain size classes on the modified Wentworth grade scale), and varietal minerals and main name (determined from the percentages of the major constituent minerals - see Fig. 2-6B).

Facies association 1: purple or green shale and siltstone. Rubble-strewn, hilly outcrops typify this facies association. Two lithotypes are readily identified on the basis of colouration and the presence or absence of sand-sized detritus.

Lithotype 1A: greyish purple laminated fine to coarse lutite and pink to dusky red fine feldspathic quartzose arenite. Outcropping throughout subregion CI (Fig. 3-5A) and ranging in thickness from 200 to 550m, this lithotype is dominated by very thinly bedded, finely laminated shales, siltstones and fine sandstones, often related in fining upward sequences. Parting lineation, ripple marks and synaeresis cracks are common sedimentary structures found within these beds. Diagnostic of this lithotype are medium to thick beds of siltstone which display an abundance of soft-sediment deformation, and banded, crossbedded sandstones. These sandstones are composed predominantly of monocrystalline quartz, with orthoclase and plagioclase feldspar totalling 7 to 10% (Fig. 3-6A). Polycrystalline quartz, microcline feldspar, chlorite, muscovite, biotite, chert, opaques, tourmaline and zircon are all trace constituents. Locally tuffaceous layers are common as either massive, crossbedded or intraformational pebble conglomerate units cemented by a haematitic matrix.

Lithotype 1B: greyish olive fine to coarse lutite. Ranging in thickness from a feather edge in the west to 450m (section ODW: see Appendix 1, page A6), this lithotype is found throughout the eastern region and subregion WI (Fig. 3-5A). It is typified by very thin fining upward beds of siltstone to shale, within which structures induced by loading are often found. Occasional beds of purple soft-sediment deformed siltstone and sandstone occur which are composed predominantly of monocrystalline quartz. Orthoclase and plagioclase feldspar total about 5%, whilst chlorite and volcanigenic fragments both average 2% (Fig. 3-6A), and polycrystalline quartz, chert, muscovite, biotite, tourmaline, zircon and opaques occur in trace amounts. Locally up to 5% carbonate can be present. Cementation is by the combined action of secondary quartz and a clay matrix.

Facies association 2: green shale, siltstone and off-white sandstone. Outcropping as a rubble-strewn hilly topography, facies association 2 is found throughout the eastern and central regions (Fig. 3-5B), where it ranges in thickness from 200 to 833m, although structural repetition may have influenced this latter figure (see section MCH; page A6, Appendix I). Within the western region this facies association intertongues with the quartzitic deposits of facies association 5 (see Fig. 3-5B). The greyish green fine to coarse lutites are laminated to thinly bedded and composed primarily of chlorite, a quartzo-feldspathic fraction and muscovite, with biotite, sericite and opaques common. On the other hand, the thin to medium interbeds of lenticular pale yellowish green fine to medium feldspathic quartzose arenite are composed of between 90 and 95% well rounded monocrystalline quartz, whilst both orthoclase and plagioclase feldspar combine with minor chlorite, muscovite and trace amounts of polycrystalline quartz, microcline feldspar, biotite, chert, volcanigenic fragments, opaque minerals, tourmaline and zircon to comprise the remaining 5 to 10% (Fig. 3-6B). These sandstones are cemented by secondary quartz. Occasionally, very thin lenticular interbeds of coarse sand, rich in polycrystalline quartz are encountered (e.g. Artimore Creek (AMC) and Prelinna (PRL) localities). Common sedimentary structures within facies association 2 include ripple marks, soft-sediment deformation structures, synaeresis cracks, raindrop impressions and sand-filled scour-and-fill structures.

Facies association 3: purple siltstone and quartzite. Moderate outcrop of facies association 3 is restricted to subregion WII, where thicknesses of 45m are attained (Middle Gorge section (MDG); see pages A4 and A9, Appendix I), although rubbly thick interbeds occur within lithotype 1B in subregion WI (Fig. 3-7A).

Lithotype 3A: greyish red purple trough crossbedded medium feldspathic quartzose arenite. East-west elongate belts, widening and thinning in an eastward direction until they intertongue with lithotype 1B, are composed of about 95% monocrystalline quartz. The remaining 5% comprises dominantly orthoclase feldspar and opaque heavy minerals (Fig. 3-6C), althouth polycrystalline quartz, plagioclase feldspar, chert, volcanigenic fragments and carbonate, along with tourmaline and zircon, are present in trace amounts. A pervasive quartz cement binds these components into a strongly indurated quartzite, althouth generally the original well rounded grain outlines are preserved by rims of either dust-like inclusions or haematite. Within the ubiquitous trough crossbedding, recumbent foreset faces are quite common.

Lithotype 3B: dusky red purple or olive grey massive to laminated coarse lutite to fine subarkose. Lying adjacent to belts of lithotype 3A within subregion WII is lithotype 3B, dominated by thick beds of well laminated siltstone. This siltstone is composed of about 85% monocrystalline quartz, 12% feldspar (orthoclase, plagioclase and microcline), 3% opaque minerals and trace amounts of polycrystalline quartz, chert, muscovite, carbonate, zircon and tourmaline (Fig. 3-6C). Occasionally volcanigenic fragments are common in laminae. These components are cemented by a mixture of secondary haematite and quartz. Characteristic sedimentary structures of some of these siltstones are large scale crossbedding (about 1m thick but of an indeterminable nature), and scour structures filled with siltstone and rich in intraformational clasts. Where these siltstones display a green hue the iron cement has been replaced by chlorite, whilst the opaque minerals are seen in a festoon cross-laminated arrangement.

Facies association 4: pink quartzite. As shown in figure 3-7B, facies association 4 is restricted to narrow linear, north-south trending belts found within subregion WII. Often two generations of this boldly outcropping quartzite are present, each averaging about 15m in thickness, but reaching a maximum of 50m (see section MDG; pages A4 and A9, Appendix I).

Lithotype 4A; moderate pink planar crossbedded medium to coarse quartzarenite. The well rounded medium sand, composed predominantly of monocrystalline quartz, occurs in medium beds and has a significant tail of coarse sand scattered throughout. Polycrystalline quartz, orthoclase feldspar, chert, muscovite, biotite, tourmaline, zircon and opaque minerals are rare (Fig. 3-8A). Haematitic rims are sometimes present outlining the original grain boundaries from the pervasive quartz cement. Often cosets of the ubiquitous planar crossbedding are delineated by thin, very pale green shaly partings, within which are commonly developed straight, sand-filled synaeresis cracks. Flute rill marks, current crescents and raindrop impressions are environmentally evidential sedimentary structures.

Lithotype 4B: very pale yellowish grey massive medium to coarse quartzarenite. Occurring as wedges within lithotype 4A, this quartzarenite occasionally displays large scale (up to 3m) planar crossbeds. Compositionally it consists of about 95% unstrained monocrystalline quartz, whilst polycrystalline and strained monocrystalline quartz, orthoclase feldspar and trace amounts of chert constitute the remaining 5% (Fig. 3-8B).

Facies association 5: cyclic purple shale and quartzite. The ABC Range Quartzite is dominated by this facies association throughout the western region (Fig. 3-9A). Its thickness reaches a maximum of about 1000m in the southwest of subregion WII (section AGG: see Appendix I, page A4).

Lithotype 5A: greyish red purple planar crossbedded medium feldspathic quartzose arenite. Monocrystalline quartz accounts for about 90% of this lithotype, whilst orthoclase feldspar and polycrystalline quartz comprise most of the remaining 10% (Fig. 3-8C). Plagioclase and microcline feldspar, chert, volcanigenic fragments, muscovite, chlorite and the heavy minerals zircon, tourmaline, rutile and opaques are all present, but only in trace amounts. A pervasive quartz cement binds these constituents into an interlocking mosaic, although often dust-like inclusions outline the original well rounded grain boundaries. Apart from the ubiquitous planar crossbedding (upon the foreset faces of which are invariably found intraformational shale pebbles), ripple marks, raindrop impressions and mudcracks (both desiccation and synaeresis) are common sedimentary structures within these medium bedded arenites. Outcrop is moderate, but rubbly because of the common presence of purple shaly partings.

Lithotype 5B: dark greyish red fine to coarse lutite and pale purple fine to medium subarkose. Poorly outcropping finely laminated shales and siltstones with lenticular, often rippled, thin sandstone interbeds typify this lithotype.

Lithotype 5A is cyclically interposed with lithotype 5B in cycles which range in thickness from 3 to 12m. In an eastward direction lithotype 5B is replaced by facies association 2, whilst lithotype 5A becomes slightly finer and less feldspathic.

Facies association 6: purple siltstone and sandstone. Rubbly, subdued outcrops within a restricted zone of subregion WI and WII (see Fig. 3-9B) typify this facies association. Here, a dusky red purple flatly laminated, shale pebble rich very fine subarkose (Fig. 3-11A) ranges in thickness from a feather edge in the east to a maximum of 52m (section MDG; see Appendix I, pages A4 and A9). The subangular to subrounded quartz (monocrystalline)-feldspar (orthoclase and plagioclase) components are bound by a mixture of a clay matrix and quartz cement. Polycrystalline quartz, microcline feldspar, biotite, chert, volcanigenic fragments, opaque minerals, tourmaline and zircon are present in trace amounts. The rubbly nature of this facies association is a direct result of the alignment parallel to bedding of both the micas and the abundant intraformational shale pebbles, providing a natural and ready plan of parting.

Facies association 7: cyclic pale yellow quartzite and green silty sandstone. Within the central region (Fig. 3-10A) the ABC Range Quartzite is dominated by facies association 7. Ranging in thickness from a feather edge in the east to an estimated maximum of 80m (maximum measured thickness is 68m from the ABC Range reference section (ABC); see Appendix I, pages A5 and A8), this facies association comprises two lithotypes cyclically interposed in cycles which range from 8 to 25m thick. Outcrop is bold where the quartzites (lithotype 7A) occur, but intervened by subdued, rubbly outcrop over the silty sandstones (lithotype 7B).

Lithotype 7A: pale yellowish grey planar crossbedded fine to medium feldspathic quartzose arenite. The medium to thick beds of this lithotype, aside from the common planar crossbedding, frequently display ripple marks, parting lineation and also occasional trough crossbeds. Separating these beds is often a pale green shaly parting within which are commonly found shrinkage (synaeresis) cracks. As shown in figure 3-11B, these sands comprise predominantly monocrystalline quartz, with an average of about 7% feldspar (mostly orthoclase, but with trace amounts of plagioclase and microcline, all of which are often sericitized). Polycrystalline quartz, chert, volcanigenic fragments, muscovite, biotite, chlorite and the heavy minerals tourmaline, zircon, rutile and opaques occur in trace amounts. A pervasive quartz cement binds the framework grains, but invariably their original outlines are preserved by rims of dust-like inclusions. Locally lenticular medium beds of coarse to very coarse sand are encountered (e.g. Parachilna Gorge (PCG)) containing about 20% polycrystallic quartz and 3% rock fragments.

Lithotype 7B: pale greenish yellow rippled lutitic fine to medium quartzarenite. This lithotype is typified by thin to medium beds which grade from a fine to medium sandstone, through a silty fine sandstone, which is often rippled, and capped by a shaly siltstone. On the base of the overlying sandstone are commonly found the casts of synaeresis shrinkage cracks developed in the underlying shaly siltstone. Despite the presence of a chloritic matrix, averaging about 5% of the total rock, the composition of these silty sandstones is a little more quartzose than those of lithotype 7A (Fig. 3-11C). Monocrystalline quartz constitutes between 90 and 95%, whilst the remaining 5 to 10% comprises dominantly chlorite, muscovite and orthoclase feldspar. Plagioclase and microcline feldspar, polycrystalline quartz, chert, volcanigenic fragments, tourmaline, zircon, and opaques are present in trace amounts. Cementation is by combination of secondary quartz with the chloritic matrix.

Facies association 8: purple to brown sandstone. Outcropping poorly within subregion WII (Fig. 3-10B), this facies association ranges in thickness from a feather edge in the east to a maximum at section AGG (Appendix I; although no actual thickness was determinable within this section, close by, at Waukarie Creek (WKC), a thickness of 96m was measured). The component greyish red purple to light brown planar crossbedded coarse litho-feldspathic quartzose arenites (see Fig. 3-13A) are found in medium beds which sometimes display upward fining and ripple marks at the top. Texturally they are bimodal, consisting of a well rounded coarse to very coarse sand fraction, dominated by polycrystalline quartz and totalling between 20 and 80% of the lithotype, set in a fine to very fine sand matrix of unstrained monocrystalline quartz bound by a quartz cement. Sedimentary rock fragments (mainly chert and silicified (often (?)oolitic) carbonates) and feldspar (dominantly orthoclase) total about 10%, whilst muscovite, zircon, tourmaline and opaque minerals are present in trace amounts.

Facies association 9: purple shale, siltstone and locally conglomeratic sandstone. Lying as a thin cap to the ABC Range Quartzite throughout much of the central region, and extending into the eastern region (Fig. 3-12A), this facies association ranges in thickness from a feather edge to 15m (Ilka Creek section (IKC); see Appendix I, page A5). Outcrop is generally moderate, though often rubbly.

Lithotype 9A: dusky red laminated coarse feldspathic lutite to fine subarkose and blackish red purple to brown coarse to conglomeratic sublitharenite. Poorly outcropping medium beds of poorly sorted, well rounded coarse sand, containing a pebble-sized fraction which occasionally defines planar crossbed sets, are typical of this lithotype. Monocrystalline quartz is the major component of these sands, but both polycrystalline quartz and rock fragments (mostly chert, silicified (often (?)oolitic) carbonates, quartzose siltstones and shales, with some volcanigenic, doleritic and carbonate fragments) each total up to 20% (Fig. 3-14A). Feldspars average less than 2% whilst muscovite, chlorite, biotite, opaques, tourmaline and zircon are all present in trace amounts. Locally the micas can constitute up to 5%, whilst intraformational shale pebbles commonly comprise 5% of the lithotype. Cementation is by a combination of quartz (binding the finest material) and haematite (binding the coarser grains).

Locally adjacent to these conglomeratic sandstones are coarse siltstones to fine sandstones which display either a parallel lamination or an undulatory bedding surface. Commonly trains of coarse sand lie within this lamination. Although monocrystalline quartz greatly dominates the composition of these siltstones and fine sandstones, feldspar (mostly orthoclase and plagioclase) averages about 12% (Fig. 3-14A). Muscovite, opaques, tourmaline and zircon are common accessories within this quartz cemented framework.

Lithotype 9B: dark greyish red shale and light grey to greyish red purple fine to medium feldspathic quartzose arenite. A finely laminated shale dominates this lithotype, although locally this shale forms partings between thinly bedded, often rippled sandstones. These sandstones are composed predominantly of well rounded monocrystalline quartz, with up to 10% orthoclase and plagioclase feldspar (Fig. 3-14B). Trace amounts of polycrystalline quartz, microcline feldspar, chert, chlorite, muscovite, tourmaline and opaques are present, all bound by a dominantly quartz cement.

Lithotype 9C: blackish red purple trough crossbedded coarse feldspatho-lithic quartzose arenite. The medium to very thick beds of this lithotype, although typified by trough crossbedding, also commonly display grading, planar crossbedding, scour-and-fill structures or a near parallel lamination. As with the coarse sand fraction of lithotype 9A, well rounded monocrystalline quartz dominates these poorly sorted sands, although polycrystalline quartz comprises up to 20% and rock fragments (again mostly chert, silicified (often (?)oolitic) carbonates, quartzose siltstones and shales, with some carbonate and volcanigenic fragments) reach 10% (Fig. 3-14C). Chlorite and orthoclase feldspar both commonly average 2%, whilst muscovite, opaques, tourmaline and zircon are only rarely seen. Again as with lithotype 9A, cementation involves the binding of the finer fraction by quartz, whilst the coarser grains are held by haematite.

Facies association 10: off-white quartzite. This facies association occurs as bold, cliffy outcrops within subregions CII, WI, WII and the southeastern part of CI (Fig. 3-12B). Its thickness ranges from a feather edge in the east to a maximum of about 50m at Buckaringa Gorge (BKG). Lithotypes 10A and 10B are separated by between 2 and 30m of dark greyish red shale of the Bunyeroo Formation.

Lithotype 10A: pale pink massive coarse quartzarenite. The thick to very thick beds of well rounded medium sand (bearing a significant tail of coarse grains) generally appear structureless because of the almost monomineralic nature of this lithotype (averaging about 95% monocrystalline quartz) and the very pervasive quartz cementation. Feldspar (dominantly orthoclase), rock fragments (mainly chert) and polycrystalline quartz constitute the remaining 5% (Fig. 3-13B), whilst muscovite, chlorite, tourmaline, zircon and opaque minerals are only very rarely seen in thin section. Occasionally large scale (up to 1.5m) planar crossbeds and ripple laminations are seen.

Lithotype 10B: pale red massive coarse to granular lithic quartzose arenite. The very thick structureless beds of well rounded coarse sand contain a secondary mode (totalling 25 to 50% of the lithotype) of very coarse sand and granules, all cemented by secondary quartz. Occasionally, lenses of microconglomerate are present. Unstrained and strained monocrystalline quartz constitute between 90 and 95% of the lithotype, whilst rock fragments (mainly chert, carbonate, silicified (often (?)oolitic) carbonate, quartzose siltstone and rare porphyries) comprise up to 15% of the coarser fraction (Fig. 3-13C). Feldspar muscovite, opaques and zircon are very rare.

CHAPTER 4

PRE-UPPER BRACHINA SUBGROUP PALAEOGEOGRAPHY

PALAEOGEOGRAPHY OF THE LOWER BRACHINA SUBGROUP

Sedimentation was continuous from the shallow submerged deposits of the uppermost Umberatana Group (the glacigene Elatina Formation) to the lowermost deposits of the Brachina Subgroup (the Nuccaleena Formation), except locally where disconformable contacts occur. The Nuccaleena Formation comprises a regionally lenticular basal dolomite, which reaches 10m in thickness and is readily divisible into three distinct sedimentary facies, and an overlying shaly facies with dolomite lenses which reaches 60m in thickness (see Plummer in prep.: see Appendix VB of this thesis). Where the basal dolomite lenses out toward the margins of the Flinders Ranges it is replaced by the normally overlying shaly facies. Rowlands (1973, p.98), however, maintains that dolomite typical of the basal Nuccaleena facies fills "irregularities in the pre-Cambrian land surface" upon the stable Stuart Shelf to the west.

The basal dolomite facies contain a near parallel, or locally stromatolitic lamination (Plate 4-1a) and are characterized by the presence of large tepee structures (Plate 4-1b) typical of intertidal and supratidal conditions (see Assereto and Kendall 1977). Plummer (in prep.) concludes that these peritidal conditions were induced by the presence of an island (or series of small islands) within the centre of the basin (where the present-day Oraparinna, Enorama and possibly Blinman 'diapir' complexes occur) and a prominant tectonic high in the northeast (near the Mount Chambers 'diapir' complex), both of which possibly developed as a result of isostatic rebound following the Elatina glaciation. Following this rebound, subsidence took place within the Adelaide 'Geosyncline' causing the shaly Nuccaleena facies to replace the earlier dolomite Nuccaleena facies and act as a passage bed into the overlying shales and siltstones of the Moolooloo Formation.

Olive grey shales and fine siltstones dominate the Moolooloo Formation within the central region. Common interbeds within this sequence are dusky red to greyish red purple shales and cross-laminated siltstones, with minor pale red to reddish brown fine sandstones. A thickness of 630m is attained in the reference section near Bunyeroo Gorge (see Plummer 1978). In subregion WII thick greyish red shales, or beds (1 to 2cm thick) of greyish red siltstone grading to dark greyish red shale dominate the sequence (Plate 4-1c). Thin (less than 1cm thick) very pale red fine sandstones are found as isolated lenses between these graded beds, whilst occasionally interrupting the sequence are beds (up to 30cm thick) of ripple-drift cross-laminated greyish red siltstone and pale red fine sandstone (Plate 4-1d). Within the eastern region and subregion WI the Moolooloo Formation is composed of beds which grade (over less than 10cm) from greyish olive siltstone to yellowish olive grey shale.

The Moolooloo Formation is divisible into two major zones on the basis of colouration and the presence or absence of fine sand. These two zones are also palaeogeographically distinct. Within the zone typified by red-purple colouration and the presence of fine sand (i.e. within the central region and subregion WII) the Moolooloo Formation appears to have been dominated by low energy sedimentation from suspension interrupted by episodic flows of turbid bottom currents. These currents were erosive, leaving shallow scour structures, and flute and tool marks on the underlying clay surfaces (Plates 4-1e and 1f). The greyish red silts deposited from traction display ripple cross-lamination, sometimes depicted by non-continuous laminae of very pale red fine sand, grading into greyish red silts and clays which settled out from the turbid upper portion of the current. Load structures (Plates 4-2a and 2b) and dewatering outlets (Plate 4-2c), sometimes culminating in small, poorly developed monroe structures probably developed as the sediment compacted under gravity. Brittle deformation also occurred, as indicated by occasional siltstone dykes (Plate 4-2d). Rare small scale structures resembling longitudinal, to frondescent ridges produced by current scouring (Plate 4-2e; cf. fleur-de-lys patterns) have two possible origins. The most likely explanation is that they developed as one of the many turbid traction currents scoured into the underlying clay layer. An alternative origin, supported by the rarity of these structures, is by current activity produced during the rapid withdrawal of exceptionally high tides or storm water (Conybeare and Crook 1968, p. 284). An origin by loading, however, is not discounted. Thin isolated lenses of ripple cross-laminated very pale red fine sandstone found between the graded beds of siltstone and shale possibly represent a weak flaser bedding - again suggestive of possible tidal influence. Deposition of the Moolooloo Formation within this zone, therefore, took place under oxygenated conditions in a shallow submerged environment. Sediment accumulation occurred either above wave base and/or within the zone of tidal influence, or within a circulatory current system under very shallow subtidal conditions (Fig. 4-1).

Deposition within the zone typified by grey-green colouration of the Moolooloo Formation and the absence of fine sand (i.e. within the eastern region and subregion WI) appears also to have been dominated by the settling out of silt and clay from turbid bottom currents. The resultant beds show bases which display occasional scour structures, along with groove and flute marks, and frondescent- to toroid-like load structures (Plates 4-2f, 2g and 2h). These structures are filled with siltstone, showing no internal structure, which then passes upward into a finely laminated siltstone which, in turn, grades into shale. These grey-green beds appear to have been deposited under anaerobic subtidal conditions of slightly greater depth than those displaying red-purple colouration, and/or in a shallow environment lacking oxygenated circulatory currents (Fig. 4-1).

MOOLOOLOO FORMATION - UPPER BRACHINA SUBGROUP BOUNDARY

The Moolooloo Formation is overlain by one of five different upper Brachina Subgroup lithotypes - lithotype 1A in subregion CI, lithotype 1B in the eastern region and subregion WI, facies association 2 in subregion CII, and either lithotype 3A or 3B in subregion WII. A transitional boundary occurs between the Moolooloo Formation and lithotypes 1A, 1B and 3B. Lithotypes 1A and 1B (the Moorillah Formation), like the Moolooloo Formation, are dominantly shale-siltstone lithologies, but thick interbeds of banded very pale red and moderate red, or dusky red, dusky blue grey and greyish purple crossbedded fine to medium sandstone and quartzite, dark greyish red soft-sediment deformed siltstone and/or tuffaceous intraformational shale pebble conglomerate are diagnostic. The boundary between the Moolooloo Formation and these lithotypes, therefore, is taken to be where these diagnostic lithologies become prevalent. Lithotype 3B, on the other hand, is primarily composed of massive to laminated coarse siltstones to fine sandstones, and the boundary between it and the Moolooloo Formation is taken to be the point where these deposits dominate the sequence. Each of these transitional boundaries is marked by a change in outcrop expression from a subdued valley topography of the Moolooloo Formation, to an undulatory, hilly topography.

A gradational boundary is displayed between the Moolooloo Formation and facies association 2 (the Bayley Range Formation). A change in colour from the red-purple of the former to the grey-green of the latter takes place over a 10 to 15m interval. A minor amount of pale greenish yellow fine to medium sandstone in thin to lenticular interbeds is present within facies association 2. No obvious change in outcrop expression is discernable across this boundary.

Lithotype 3A is an extremely diagnostic facies - a greyish red purple trough crossbedded, medium to coarse quartzite. In some areas it appears as interbeds within lithotype 3B, and in these areas, therefore, displays the same boundary characteristics as does lithotype 3B. Elsewhere, however, a sharp boundary is observed. Although from the limited continuous outcrop along which the contact can be traced it appears conformable, a very shallow disconformable relationship is not discounted.

In general, then, the Moolooloo Formation-upper Brachina Subgroup boundary is conformable and transitional, although locally gradational and possible disconformable contacts occur. Temporally this boundary represents the ingression of coarse silt- to sand-sized detritus into a vast mudflat environment. This mudflat was aerobic, shallowly submerged, and possibly tidally influenced along the western flank and within the central region of the basin, whilst in the eastern region a gentle palaeoslope descended into an anaerobic subtidal mudflat environment (Fig. 4-1).

CHAPTER 5

PALAEOGEOGRAPHY OF THE UPPER BRACHINA SUBGROUP : DELTAIC PHASE

Deposition of the upper Brachina Subgroup commenced with the debouchment of an immense amount of sand into the basin from a westerly source region (the Gawler Craton). This influx produced a thick sand accumulation (the ABC Range Quartzite) extending from Alligator Gorge (AGG) northward to "Warrakimbo Gorge" (WKG), and spreading eastward, eventually to Moockra Tower (MKT) (see Fig. 1-1 for localities). Initially surrounding this vast sand pile, and covering the remainder of the study area, typical Moolooloo muds continued to steadily accumulate. These deposits, which constitute the Moorillah Formation are, however, differentiated from those of the Moolooloo Formation by the presence of occasional medium to thick beds of silt- to sand-sized sediment which display an abundance of well developed soft-sediment deformation structures.

REGIONAL LITHOTYPE ARRANGEMENT

The first sand to encroach upon the Moolooloo mudflats was restricted to three narrow, roughly west-east trending elongate zones of greyish red purple, heavy mineral rich, trough crossbedded medium to coarse quartzarenite (lithotype 3A). In subregion WII these narrow zones are flanked by broad expanses of dominantly dark greyish red massive to laminated siltstone which comprise lithotype 3B. Toward, and into subregion WI, however, as the gentle palaeoslope from the basin's western flank to the deeper eastern region was encountered, these narrow zones of lithotype 3A fanned out into thin, broad aprons of banded dusky red, dusky blue grey and greyish purple, planar crossbedded fine to medium quartzarenite, which mark the base of lithotype 1B and lie interbedded with the dominant grey-green shale-siltstone lithology of this latter lithotype (Fig. 5-1A). Further eastward, away from the influence of lithotype 3A,medium to thick beds of dark greyish red, soft-sediment deformed siltstone mark the lower boundary of lithotype 1B within subregion EI.

Northward, in subregion CII, the boundary between the Moolooloo Formation and the basal unit of the upper Brachina Subgroup (lithotype 1A of the Moorillah Formation) is defined by the presence of thickly bedded, banded very pale red and moderate red, crossbedded medium sandstones, dark greyish red intraformational shale pebble conglomeratic tuffaceous siltstones, and, again, dark greyish red soft-sediment deformed siltstones, all interbedded within a greyish red to olive grey laminated to thinly bedded shale-siltstone succession. Lying between the regions of lithotype 1A and lithotype 3A accumulation (i.e. within subregion CII), is found a narrow zone where greyish green shales and siltstones, with thin to lenticular, pale greenish yellow sandstone interbeds (facies association 2 - the Bayley Range Formation) replaces the greyish red Moolooloo deposits (Fig. 5-1A).

Immediately above lithotype 3B throughout the western portion of subregion WII, though still cut by the restricted zones of lithotype 3A (now down to two in number), sand deposition became predominant and a thick succession of greyish red purple, planar crossbedded, medium feldspathic quartzose arenites (lithotype 5A) were laid down cyclically interposed with dark greyish red shales, bearing thin to lenticular siltstone and sandstone interbeds (lithotype 5B). The eastern limit of this cyclic succession is bordered by a narrow, roughly north-south trending belt of moderate pink, planar crossbedded, medium quartzarenites (lithotype 4A), which intertongues with the remnant deposits of lithotype 3B (Fig. 5-1B). Basinward, away from the direct influence of the large sand influx, the narrow belt of facies association 2 extended right around the region of sand accumulation, thus separating it from the deposits of lithotype 1B. Still accumulating in subregion CI were the deposits of lithotype 1A.

With the continued and persistent influx of sand into subregion

WII the cyclic facies association 5 deposits steadily prograded into the basin. The north-south trending belt of lithotype 4A - now with wedge-shaped interbeds of very light grey massive coarse quartzarenite (lithotype 4B) - migrated with the prograding edge of facies association 5. Meanwhile the siltstones of lithotype 3B and quartzarenites of lithotype 3A were replaced by the deposits of facies association 5, and the narrow belt of facies association 2 steadily prograded basinward replacing the deposits of lithotypes 1A and 1B (Figs. 5-1B, 2A and 2B).

Eventually facies association 5 prograded across the entire area of subregion WII and intertongued with facies association 2 in subregion WI. Facies association 2 continued to replace lithotypes 1A and 1B (Figs. 5-3A and 3B) until only these two facies associations (viz. 2 and 5) were being deposited. The sandstones of the Barunga and Hummock Ranges in subregion WIII are correlated with those of facies association 5 (see Plummer 1978) and it is suggested that the full extent of facies association 5 deposition included this subregion, as shown on figure 5-3. The remainder of the study area (i.e. the eastern and central regions) was completely dominated by facies association 2 deposition, although rare thin lenses of coarse sandstone petrologically similar to those of facies association 9 (see Chapter 6) are found toward the top of facies association 2 in subregion CI at locality PRL and South of "Narinna" Homestead (SNH). The importance of these lenses will become obvious when the palaeogeographic significance of facies association 9 is discussed in Chapter 6.

SEDIMENTARY STRUCTURES, PALAEOCURRENTS AND PALAEOGEOGRAPHY

A large suite of sedimentary structures is found throughout the nine facies associations of this first phase of upper Brachina Subgroup deposition. Because of the environmental significance of a number of these structures, with many also providing palaeocurrent information, these sedimentary structures offer the necessary means from which palaeogeographic interpretations of the lithotype maps can be attempted (see Figs. 5-1 to 3).

Deposits of two contrasting types divide the study area into two during this first phase of sedimentation. In the western region sand-dominated lithotypes accumulated, whilst throughout the eastern and central regions shale and siltstone deposits were dominant.

SEDIMENTOLOGY - TIME INTERVAL I

Submerged Distal Delta (Lithotypes 3A and 3B)

The initial influx of sand-sized sediment from the Gawler Craton to the west entered the basin through three restricted zones which extended across subregion WII and into subregion WI, where they fanned out and possibly coalesced. Within subregion WII the deposits of these restricted zones (lithotype 3A)show an abundance of trough crossbedding (pi cross-stratification of Allen 1963), which forms by repeated current scouring and filling. Within this crossbedding recumbent foresets are seen (Plate 5-1a). Although such structures have been suggested to form by movement under gravity of oversteepened crossbedding, or from earthquake shocks, they are interpreted here as being indicative of drag by strong sediment-laden currents upon water-saturated, quasi-stable crossbeds (see Hendry and Stauffer 1977). Support for the presence of strong currents within these restricted zones is given by the abundance of heavy minerals contained within the quartzites compared to adjacent lithotypes. Palaeocurrent data from this lithotype are hard to obtain due to the scarcity with which bedding planes are exposed (these being necessary to determine accurate readings from trough crossbedding). Enough readings to analyse, however, were gained from Alligator Gorge: near Wilmington (AGW) and South of Warren Gorge (SWG). At both these localities a northeast to southeast current dominates (80% of the readings), whilst a minor secondary mode is suggested toward the southwest (see Fig. 5-4A).

As these lithotype 3A sands entered subregion WI the ubiquitous trough crossbedding was replaced by a planar cross-stratification, described by Allen (1963) as xi cross-stratification. In this subregion these quartzites form broad, thin aprons which intertongue with the dominantly greyish olive shale-siltstone sequence of lithotype 1B. No palaeocurrent information was obtained from these quartzites.

Adjacent to the restricted zones of lithotype 3A within subregion WII were deposited the siltstones of lithotype 3B. Although generally either structureless or flat- to shallowly cross-stratified (Plate 5-1b), occasional large-scale cross-stratification (set thickness averaging 1m) of an indistinguishible nature is seen (e.g. section MDG). Scour structures filled with locally derived shale and siltstone intraclasts set in a coarse siltstone matrix are locally present, disturbed bedding (load and flame structures) is not uncommon. Ripple cross-lamination (mu cross-stratification) is usually only seen when depicted by opaque minerals, but indicates current activity.

This sequence of deposits, and the sedimentary structures contained therein, are indicative of a subaqueous deltaic environment (see Reineck and Singh 1975, p.269), although rare desiccation cracks suggest occasional subaerial exposure. Restricted zones of high energy, coarse sediment-laden, eastwardly flowing currents, depositing lithotype 3A, represent distal distributary channel currents. These fanned out and deposited distributary mouth shoals as they dissipated within the region of lithotype 1B accumulation (see below). The deposits adjacent to the distributary channels (i.e. lithotype 3B) have characteristics typical of subaqueous levee deposits (Fig. 5-5).

Deltaic Bottomset (Lithotype 1B)

The region into which the distributary mouth shoals intertongue extends across subregion WI and throughout the eastern region. The greyish olive shales and siltstones which dominate lithotype 1B display the same general sedimentary structures as does the underlying Moolooloo Formation within these same areas. Graded beds, containing groove marks, flute marks and small-scale load structures, are indicative of deposition from turbid bottom currents. The grey-green hue of these sediments, due to the presence of ferrous iron, is suggestive of sedimentation under reducing conditions. The occasional presence of limonite cubes after pyrite within the eastern region (e.g. section MCH) supports this interpretation.

According to Berner (1970) the formation of sedimentary pyrite within marine sediments occurs when fine-grained detritus, accompanied by an abundance of dead organic matter, accumulates under very quiet conditions. Borchert (1960) explains that the waters within such quiet environments have three recognisable zones. The surface and nearshore waters are oxygenated, and within such waters almost complete oxidation of all organic substances occurs. Bottom waters in the deeper parts of the basin, on the other hand, contain an abundance of H_2S due to the anaerobic bacterial sulphate reduction of organic matter. Existing between these two kinds of water is a transition zone having CO_2 rich waters which only partially oxidize organic matter. It is within this latter zone that iron is dissolved and mobilized from the fine-grained sediments. When this dissolved iron moves into the zone of the bottom waters it reacts with the abundant H_2S present and a black, non-crystalline form of FeS is produced. Some of the H_2S, however, becomes oxidized, either inorganically or by bacterial activity, thereby releasing elemental sulphur which then combines with the black FeS to produce pyrite (FeS_2). Although this same general process of sedimentary pyrite formation often takes place below the sediment-water interface within well oxygenated environments where an abundance of microscopic organic activity occurs (e.g. intertidal mudflats), the sedimentary structures found within lithotype 1B, and the palaeoenvironmental significance of adjacent lithotypes, support deposition within a subtidal environment.

The pyrite-bearing shales and siltstones of lithotype 1B within the eastern region, therefore, suggest deposition within the zone of Borchert's bottom waters, whilst subregion WI, apparently lacking in pyrite but still reducing, would correlate with the transition zone. Also found within the eastern region, but apparently absent from subregion WI, are intermittent interruptions of the sequence by thick beds of dusky red purple siltstone and very pale red to moderate red, fine to medium sandstone. The colouration of these beds indicates that they were deposited under oxygenated conditions. Contained within them is an abundance of soft-sediment deformation structures (Plate 5-1c) suggestive of rapid sedimentation from turbid bottom currents. It is believed that these sediment-laden currents were generated from the oxygenated sediments bordering the prograding subaqueous delta within subregions WI and WII which then flowed down the palaeoslope and into the eastern region.

Lithotype 1B, therefore, lying seaward of a distal subaqueous prograding delta system, represents the deposits of an anaerobic deltaic bottomset environment. Mud deposition dominated, but periodic flows of oxygenated, sediment-laden bottom currents, generated from the distal reaches of the prograding distributary mouth shoals, interrupted the sequence (Fig. 5-5).

Shallow Subtidal Mudflat (Lithotype 1A - basal)

Within subregion CI of the study area the initial coarse sediment influx is marked by the presence of banded very pale red and moderate red, fine to medium sandstone interrupting the typical dark greyish red shale-siltstone Moolooloo sequence. These sandstones are commonly medium bedded and display either pi (Plate 5-1d) or xi crossbedding, from which no palaeocurrent data was obtainable. These sandstones can also be flatly laminated and associated with a pale green, fine to medium sandstone, from which parting lineation directions were obtained. A bimodal distribution is suggested from the two analyses completed, with one mode trending north-south and the other west-northwest - east-southeast (Fig. 5-4). Occasionally the red sandstones are seen as slurry-slump beds (Plate 5-1e), or contained within beds of massive dusky red purple siltstone as disrupted load casts and ball-and-pillow-like structures (Plate 5-1f). The siltstones themselves often display internal deformation (Plate 5-1g), or the presence of locally derived shale and siltstone clasts, generally highly ferruginous, in beds of intraformational conglomeratic tuffaceous siltstone (Plate 5-1h).

These beds demarcate the contact between the underlying dark greyish red shales and greyish red siltstones of the Moolooloo Formation and the overlying dominantly flat, thinly bedded sequence of greyish red purple shales, greyish red siltstones and very pale red to moderate and dusky red, fine to medium sandstones of lithotype 1A (the Moorillah Formation). These two sequences (viz. Moolooloo and basal lithotype 1A) were deposited within the same general environment - viz. a shallow submerged, possibly tidally influenced mudflat (Fig. 5-5). Support for the tidal influence is taken from the roughly east-west palaeocurrent trend gained from the parting lineation. The energy conditions of deposition, however, were sporadically greater for basal lithotype 1A than for the underlying Moolooloo Formation. Because of the presence of a prograding, though still submerged delta system within close proximity of the central region of the basin, the turbid bottom currents which flowed northeastward across this mudflat carried a greater amount of sediment into the region, including a coarse fraction previously not available. The drag of these sediment-laden currents, plus the increased density of their deposits upon the previously laid down, still water-saturated sediments, generated conditions suitable to the development of soft-sediment deformation. The erosive power of the currents carrying a high percentage of coarse detritus is seen in the sharp erosive contacts at the base of the crossbedded sandstones (see Plates 5-1d and 2a).

PALAEOGEOGRAPHY - TIME INTERVAL I

The coarse detritus debouched from the Gawler Craton into subregion WII of the basin entered through three broad, distal submerged distributary channels (lithotype 3A) which cut across the shallow submerged Moolooloo mudflat and deposited subaqueous levee deposits in the intervening areas (lithotype 3B) (see Fig. 5-5). Upon reaching the gentle palaeoslope leading down into the anaerobic subtidal mudflat of the eastern region and subregion WI, these distributary channels fanned out, and possibly coalesced, as they deposited their remaining coarse sediment as distributary mouth shoals. These shoals intertongue with the deeper, anaerobic subtidal deltaic bottomset deposits of the eastern region (lithotype 1B). Northward, in subregion CI of the basin, the shallow submerged, possibly tidally influenced Moolooloo mudflat persisted, with the addition of occasional pulses of coarser detritus (basal lithotype 1A).

This initial development of a distal submerged deltaic sequence within the western region of the basin heralded the development and progradation of a thick pile of deltaic deposits in the ensuing period of upper Brachina Subgroup sedimentation.

SEDIMENTOLOGY - TIME INTERVAL II

Intertidal Deltaic Plain (Lithotypes 5A and 5B)

With continued coarse sediment influx into the basin from the Gawler Craton, this submerged distal deltaic system was overlain by two cyclically interposed lithotypes, namely a moderate greyish red purple, planar crossbedded, medium feldspathic quartzose arenite (lithotype 5A) and a greyish red shale and siltstone, with very pale red lenticular, fine to medium sandstone interbeds (lithotype 5B).

A typical section of one lithotype 5A-5B cycle is presented in figure 5-6. As can be seen from this section the arenaceous crossbedded sets of lithotype 5A are often separated from one another by shaly partings. These shaly partings provide a natural plane of dissection of the sequence, and, with the aid of mechanical weathering processes, reveal a wealth of environmentally significant sedimentary structures.

Sedimentary Structures and their Palaeoenvironmental Significance

A number of sedimentary structures are found within lithotype 5A suggestive of periodic exposure to the atmosphere. Well developed large polygonal mudcracks are common. Some exposures reveal the complete crack infill (Plate 5-2b), whilst others appear to have lost the lower parts of the crack infill (Plate 5-2c). The polygonal nature of the (now eroded clay) plates and the constant width of each crack suggest a desiccation origin.

Such mudcracks cover only a portion of the relevant bedding plane upon which they are found. Sketched next to one such mudcrack occurrence in the column presented in figure 5-6 is the relationship between the shape of the underlying and overlying sandstone beds and the position of the mudcrack. As can be seen, the mudcrack occurs in a hollow in the underlying sandstone layer. Clay settled out of suspension from any water trapped within this hollow, which then cracked upon exposure and drying. The overlying sandstone layer thickens above the mudcrack where more sand was required to fill the hollow.

Commonly found upon the mould of the (now eroded desiccated clay) plates of these mudcracks is an abundance of irregularly shaped casts (Plates 5-2c and 2d). These casts possibly represent the sandy infills of impact structures formed by hailstones when the clay layer was still soft. The preservation potential of hailstone imprints in clay would be higher than for raindrop impressions because of their greater depth of penetration and the cohesive nature of the clay. Subsequent burial by a sand layer would readily produce casts of such melted hailstone imprints. Another possible origin for such casts is by dissolution and sand replacement of either salt or gypsum crystals which crystallized out of the brine trapped within a depression in the depositional surface, where salinity steadily increased as evaporation proceeded. The shape of these casts, however, resembles neither the typical cubic nor hopper crystal of salt (see Conybeare and Crook 1968, p.270, Plate 90) nor the somewhat elongate gypsum crystal (see Conybeare and Crook 1968, p.214, Plate 64c) and these origins therefore, seem unlikely.

Scattered talus pieces of lithotype 5A reveal a rough, bumpy upper surface (c.f. Plate 5-11c). The non-penetrative nature of this surface expression rules out the possibility of formation by differential compaction, and as Miller (1975) suggested, such surface features represent the impression of raindrops which fell upon a sand surface immediately prior to its burial. Where more than 1mm of rain has fallen the individual crater imprints are partially destroyed by other raindrops, leaving a surface displaying "irregular, partly connected depressions...separated by narrow ridges" (Reineck and Singh 1975, p.52), as seen on plate 5-2e which shows a beach sand (at Maslin's Beach south of Adelaide) after a short, heavy shower on 29/8/75. A similar structure, though usually with less relief, is produced by the migration of windblown foam across an exposed sand surface (see Reineck and Singh 1975, p.53, Fig. 71). Single bubble tracks blown across the sediment surface leave tracks similar to that shown on plate 5-2f (see Reineck and Singh 1975, p.53, Fig. 74). Both raindrop impressions, and foam and bubble tracks are indicative of subaerial exposure of the sediment surface.

Possible moulds of mudballs (Plate 5-2g) were found at locality BKG. The origin of mudballs is ascribed to the erosion of blocks of clayey material and their rounding during transportation in rapidly flowing water, such as swift streams and flood runoff (Bell 1940; Ojakangas and Thompson 1977). Often armouring of the mudballs occurs by the adhesion of sand, granules and pebbles upon their outer surfaces. The lack of any armour upon the mudballs at locality BKG (none evident from the moulds) suggests a lack of material of granule size and coarser within the sand-sized sediment over which the mudballs rolled. Mudballs are found in a variety of modern environments, ranging from fluvial, to lacustrine and marine beach, intertidal, submarine channels, and even urban environments (see Ojakangas and Thompson 1977).

On a smaller scale, shale pebbles and pebbles of sandy shale are very common throughout the arenites of lithotype 5A. Flat, and ranging in shape from small ovoid discs (Plate 5-3a), to quite large irregular to ovoid clasts (reaching 30cm diameter), these shale pebbles are often found toward the base of crossbedding foreset faces where the crossbed set is underlain by a shale layer (see Fig. 5-6). Throughout the lithotype 5A sequence these shale pebbles are seemingly more abundant in the crossbeds which have an easterly palaeocurrent component (see later). Shale pebbles originate either by desiccation, forming flakes which are later incorporated into the sandy sediments of similar hydraulic character (clay galls), or by the subaqueous ripping up of a clay layer (or part thereof) by a strong current. Although such structures, therefore, are not diagnostic of any specific environment of deposition they do tend to be more commonly found in environments subjected to desiccation.

Ripple marks, although present throughout lithotype 5A, are usually found at the top of each sand cycle where they are directly overlain by the shales and siltstones of lithotype 5B. This was noted by Klein (1970) who defined 'B-C-D sequences' (i.e. crossbedded sands (interval B) overlain by ripple cross-laminated sands (interval C) and followed by muds (interval D)) as being typical of intertidal conditions of deposition. Some ripple marks show distinctly flattened crestlines, eroded off by the action of very small water ripples ("capillary waves") produced by very strong winds on the water surface (see Reineck and Singh 1975, p.369, Fig. 529) during emergence of the sediment surface through a falling water level (Tanner 1958). These structures are further evidence for periodic exposure of the depositional surface, and support for an intertidal environment of deposition. Of the ripple marks which are not flat-topped, both symmetric and asymmetric wave ripples (wavelength averaging 8cm) and asymmetric current-dominated combined-flow ripples (wavelength averaging 11cm) are found (see Harms 1969, Fig. 14 for ripple interpretations using crest spacings). Occasionally ripple marks are found upon crossbedding foreset faces (Plate 5-3b) which indicate sudden changes in current strength (Klein 1971). Interference ripples are also found, as shown in plate 5-3c, where a set of asymmetric current ripples and a set of asymmetric wave-dominated combined-flow ripples are superimposed. Flute marks also testify to the presence of current activity, but these are only rarely seen.

By far the most common sedimentary structure found within lithotype 5A is planar (omikron) crossbedding. As seen in figure 5-6 most individual arenite beds are crossbedded. Within subregion WII the average set thickness is 14cm, with an average foreset slope of 23^{o}. Eastward within subregion WI, however, the crossbedding has an average set thickness of 12cm and an average foreset slope of 18^{o}. This decrease in foreset slope and set thickness is typical of crossbedding the further away it is from the source region (Hamblin 1961) and/or as energy conditions decrease (Brush 1958).

A typical crossbedded outcrop is shown in plate 5-3d in which herringbone, or chevron crossbedding is not uncommon. Herringbone cross-bedding comprises "crossbedded units with opposite direction of foreset laminae in adjacent layers" (Reineck and Singh 1975, p.86) and is indicative of tidal conditions (Reineck 1963). Although a majority of the crossbedded sets are uniform in thickness, and direction and angle of foreset slope for the length of the outcrop (up to 10m) some crossbedded sets show a variable relationship with its adjacent sets and/or within itself (Fig. 5-7A). Some sets degenerate into ripple-drift cross-lamination (Fig. 5-7A(i)), whilst others wedge out into an underlying set, leaving an apparent trough crossbed where outcrop is poor (Fig. 5-7A(ii)). Other sets truncate an underlying set with a convex upward surface (Fig. 5-7A(iii)) and Plate 5-3e). These "reactivation surfaces" (Collinson 1970, Klein 1970) dip at an angle shallower than the foreset laminae and are typical of recent fluvial and tidal deposits. Where crossbedded sets directly overlie lithotype 5B the leading edge of the set can show an intertonguing relationship, as depicted in figures 5-6 (between 2.95 and 3.10m) and 7A(iv).

A discordant relationship exists between the bedding plane direction and that of the lithologic boundary between the cyclically interposed lithotypes 5A and 5B. This suggests that the migration of discrete areas of sand deposition, separated by discrete areas of mud accumulation, was responsible for this cyclic sequence. The nature of these discrete areas, however, is not obvious from the typically occurring cross-sectional outcrop pattern. Rare plan view exposures of lithotype 5A display straight, to wavy-crested megaripples, with superimposed asymmetric current-dominated combined-flow ripples (Plate 5-3f). The migration of such megaripples in shoals, with muds accumulating between each shoal would produce such a cyclic arrangement, as depicted in figure 5-7B. As each megaripple migrates, minor scouring occurs at the base of each foreset face, eroding off the superimposed ripple marks on the preceding (and hence underlying) megaripple, (note the lack of ripple marks directly in front of each megaripple foreset on plate 5-3f). In general, therefore, the only set of ripple marks likely to be preserved are those on the back of the last megaripple in the shoal, which is directly overlain by the muds accumulating behind the shoal. Where minor mud accumulations occur in hollows between individual megaripples within a shoal, however, ripple marks can be preserved. This agrees well with the sequence presented in figure 5-6. In some places the megaripple form is that of a scour, or lunate megaripple (Plate 5-3g), suggesting slightly greater current velocities than required for the straight-crested megaripples (Reineck and Singh 1975, p.35).

Rarely seen in outcrop are recumbent and slumped (Plate 5-4a) crossbedding foreset faces. Such structures are produced by the drag developed as a current flows across saturated sediment. "Liquefaction and earthquake shocks have little, if any, control on the origin of deformed crossbedding" (Hendry and Stauffer 1977).

Other structures indicative of sediment saturation and deformation include load structures (Plate 5-4b) and shrinkage cracks (Plate 5-4c). A sand layer deposited upon a hydroplastic mud layer induces loading, dewatering and upward displacement of the mud (i.e. flame structures) as the sand founders to form load structures (see Reineck and Singh 1975, p.76 Fig. 125). The abundant shrinkage cracks found, normally having a straight to weakly sigmoidal shape and going from a point at one end, thickening toward the middle, then thinning and terminating in a point, are often isolated and bear no resemblance to cracks formed by desiccation. They have been described within the literature as 'fusiform structures' attributable to the activity of burrowing organisms (e.g. Bose 1977). Donovan and Foster (1972, p.309), on the other hand, attribute such structures to "subaqueous shrinkage by a synaeresis mechanism". Such an interpretation is applicable here.

The mechanism of formation involves the dewatering and subsequent shrinkage and 'tearing' of a saturated mud layer. Plate 5-4d shows a series of well formed small-scale load casts of sand which foundered into an underlying mud layer (now eroded away). The cracks lying between the load casts must have developed simultaneously, since for the loading to have occurred after desiccation is impossible as the mud layer would no longer be saturated and susceptible to hydroplastic deformation, and for desiccation to have occurred after hydroplastic deformation is also impossible as the mud layer would no longer have been subaerially exposed. Thus the loading and dewatering of a saturated mud layer by an overlying sand layer can generate the conditions conducive to the development of shrinkage cracks. The sand layer need not necessarily be a continuous bed, but may take the form of a connected lenticular layer, as shown on plate 5-4e.

Sinuous mudcracks (Plate 5-4f) are occasionally found within lithotype 5A. As with the 'fusiform' shrinkage cracks, these 'vermiform' structures have also been previously described as trace fossils. Where continuous, these sinuous structures have been classified under the genetic term Manchuriophycus Endo 1933. However, their close association with ripple marks (ie. they are normally found within ripple troughs) led Häntzschel (1949) and Wheeler and Quinlan (1951) to consider them as the casts of drying cracks. Where such structures display a spindle shape they have been termed Rhysonetron Hofmann 1967, but "despite the convincing organic aspect of some specimens, Rhysonetron is a sedimentary-diagenetic structure, resulting from shrinkage crack filling" (Hofmann 1971, p.39).

An incomplete annuloid-shaped structure, 63mm in diameter and 8mm broad, was found on the basal surface of a lithotype 5A cycle in section MDG (Plate 5-4g: see page A9, Appendix I). This structure resembles that of either a horizontal worm burrow, or the terminal portion of a feeding burrow similar to that described by Seilacher (1967, p.421) as a "horizontal protrusive type of rhizocorallid spreite burrow" and attributed to an unknown crustacean. The solitary nature of this structure, however, disfavours an organic origin. Running through the incomplete part of the annulus is a current lineation trending 087°-267°. An inorganic explanation of this structure is that it represents the sand infill of an eddy scour which developed behind an obstacle sitting on the underlying mud layer (perhaps a large mud flake). If this is the case, the current forming the lineation and the scour was directed toward 087°, a direction which lies within one of the two modal classes of palaeocurrent directions obtained from the crossbedding data at this locality (see later). Such an inorganic origin is favoured for this structure.

Lithotype 5B is composed dominantly of flat, to wavy laminated shale and siltstone. Sandstones are present as lenticular laminae to very thin beds. This type of bedding has been described by Wunderlich (1970) as "tidal bedding". The basal surface of the thin sandstone beds and lenses range from very flat, through slightly undulose with abundant casts of shrinkage cracks, to load casted where the sand foundered after being deposited upon a saturated muddy layer. In some instances the sand load casts have become isolated within the underlying muddy layer to form ball-and-pillow structures, or pseudonodules (Plate 5-4h). The upper surfaces of the thin sandstone beds and lenses usually display a ripple mark pattern; generally wave ripples, although current ripples and isolated ripples are occasionally seen.

The cyclically interposed lithotype 5A-5B succession, therefore, was deposited within an environment which suffered reversals in the bedload transport directions (producing herringbone crossbedding), alternating bedload transport with suspension settlement (forming lenticular and tidal bedding), emergence with sudden changes in flow direction at extremely shallow water depths (leaving flat-topped ripple marks, interference and symmetric ripple marks, ripple marks on crossbedding foreset faces, and 'B-C-D' sequences), differential loading and hydroplastic adjustment (causing load structures, pseudonodules and shrinkage cracks) and periodic subaerial exposure (as evidenced by large polygonal mudcracks, and possible raindrop and hailstone impressions). Klein (1971) lists these features as being indicative of intertidal sequences.

The cyclic lithotype 5A-5B arrangement, therefore, developed in response to the migration of megaripple shoals (lithotype 5A), separated by slightly elevated areas of mud accumulation (lithotype 5B), within an intertidal environment.

Estimate of Palaeotidal Range

Klein (1971) used such cyclic sequences to determine palaeotidal ranges of ancient successions. Within each cycle he defines four zones, each of which is dominated by different transport processes. The basal zone involves tidal bedload transport and produces a subtidal sand. This is followed by a combination of tidal bedload transport and emergence runoff prior to exposure, depositing a low tidal flat sand. Alternating bedload and suspension sedimentation follows leaving a mid-tidal flat sequence of interbedded muds and sands. The uppermost zone comprises a mud deposited from suspension in a high tidal flat. The palaeotidal range is determined from the thickness of the sequence between the contact of zones 1 and 2, and the top of zone 4.

Using section MDG, where the best exposure of the cyclic facies association 5 sequence occurs, the cycles were studied in an attempt to determine the feasibility of palaeotidal determinations. From the included sedimentary structures within each cycle, lithotype 5A represents only the low tidal flat sand deposit, the basal subtidal sand being absent. Also, lithotype 5B represents only the mid tidal flat deposit of interbedded sand and mud, the uppermost high tidal flat mud deposit generally being absent (see Fig. 5-6). The total palaeotidal range, therefore, cannot be accurately determined as both its lower and upper bounding surfaces are not represented within the cycles.

Also, these cycles range in thickness from 3 to 12m. No general trend of either upward thickening or thinning is evident, which suggests that the thickness variation is not due to a steady increase or decrease in the actual palaeotidal range acting upon a stable sediment surface.

The apparent randomness of the variation is more likely to be the result of fluctuations in the rates of sediment influx and basin subsidence. For example, assuming a fairly constant rate of sediment influx, a rapid rate of basin subsidence would produce a thick cycle (provided that the rate of sedimentation remained greater than the rate of subsidence), suggestive of a large palaeotidal range, whilst a decrease in the rate of basin subsidence would produce a thin cycle, suggestive of a small palaeotidal range.

It is therefore apparent that estimations of palaeotidal ranges from ancient sedimentary sequences should not be attempted unless, firstly, all four zones defined by Klein (1971) are present, and secondly, accurate estimates of the rates of sedimentation and basin subsidence are known.

Deposition of the cyclic facies association 5 persisted within subregion WII throughout the first phase of upper Brachina Subgroup sedimentation. The adjacent lithotypes, however, varied as the basin shallowed due to the high rate of sediment influx and accumulation. Consequently, two principal facies association arrangements resulted as the palaeogeographic configuration evolved with the change in energy conditions induced by this basin shallowing.

Barrier-Bar (Lithotypes 4A and 4B)

When the first medium sands which constitute lithotype 5A prograded across subregion WII, burying the subaqueous levee deposits of lithotype 3B, their progress was halted at the shallow palaeoslope which descended gradually eastward into the reducing, submerged deltaic bottomset environment (lithotype 1B). At this eastern limit a north-south trending, elongate sand body developed (facies association 4).

Sedimentary Structures and their Palaeoenvironmental Significance

Sedimentary structures testifying to periodic subaerial exposure, such as rare large polygonal desiccation cracks and surfaces displaying possible raindrop impressions (Plate 5-5a), are found within the quartzites of lithotype 4A, as they are within lithotype 5A. Layers rich in shale pebbles are generally found in close proximity to interbedded shaly layers. Shrinkage cracks are also found (Plate 5-5b). These cracks are generally linear in shape and have been previously described as the casts of ice crystals (Udden 1918). Symmetric wave ripples (average wavelength of 4 cm), slightly asymmetric wave-dominated combined-flow ripples (average wavelength of 6cm) and casts resembling those of flute rill marks (Plate 5-5c; see Reineck and Singh 1975, p.66) are sometimes found. Omikron crossbedding, however, dominates these quartzites. The average set thickness is 12cm, whilst the average foreset slope angle is 22° - slightly thinner and shallower than that of lithotype 5A.

Wedges of very light grey quartzite (lithotype 4B) are occasionally found within this dominantly moderate pink succession. These quartzites display either massive bedding or rare large-scale (up to 3m) planar (alpha) crossbedding, with either an undulatory upper surface, or one displaying shallow mounds.

Where lithotype 4A is cut by the distributary channels of the delta system (lithotype 3A) a transition occurs from the omikron cross-bedding of lithotype 4A to the pi crossbedding of lithotype 3A. The transition is typified by beds of well-formed omikron crossbedding inter-posed with beds showing a pi-type crossbedding, sometimes displaying recumbent foreset faces (Plate 5-5d). A rapid thinning of the sequence also occurs from lithotype 4A into lithotype 3A, suggestive of stronger more erosive current activity within the distributary channel zone.

The superposition of intertidal muds and sands (facies association 5) upon the submerged distal deltaic system (facies association 3) within subregion WII indicates the progradation of deltaic topset beds into the basin. The absence of deltaic foreset deposits is suggested as being due to the shallow nature of the basin within this area. Where the distal reaches of the intertidal deltaic plain encountered the easterly facing palaeoslope leading to the deltaic bottomset environment, wave activity reworked the sand fraction into a north-south trending, elongate body representing a barrier-bar system (facies association 4).

Palaeocurrents of the Intertidal Deltaic Plain and Barrier-Bar

Palaeocurrent data from the crossbedding of lithotypes 3A, 4A and 5A at this early stage of delta progradation and development confirm the interpretations thus far made. The trough crossbedded distributary channel deposits of lithotype 3A have a dominant current direction toward the northeast, and a minor secondary current toward the southwest (Fig. 5-4B). Upon the cyclic sand-mud covered intertidal deltaic plain (facies association 5) the palaeocurrent pattern parallels that of the distributary channels. One mode, a dominantly eastwardly directed current at Alligator Gorge (localities AGH and AGM), changes in a northward direction to a north-easterly current, with a southerly flowing secondary current, and in a southward direction to a southeasterly flowing current, with a secondary northerly current (Fig. 5-4B). Ripple mark directions were obtainable from one locality only (Woolshed Flat (WSF)). Here a bimodal distribution is present, but with a variation of 30^{o} to the crossbedding modes for this locality, although the proportionality of each mode remained about the same as for the crossbedding. This supports the change in current flow direction said to occur with emergence of the depositional surface during ebb tide, and held as indicative of intertidal sequences (Klein 1971).

The palaeocurrent information taken from the narrow, north-south trending belt of sand reworked at the distal edge of the intertidal deltaic plain (i.e. lithotype 4A) shows a complex current arrangement. The initial configuration shows the same eastwardly flowing current system, fanning both to the north and south as it leaves the basin margin, and also evidence of the same secondary, either northerly or southerly current system. However, a third current system is also present, flowing parallel to the trend of the sand body (Fig. 5-4B). This dominantly north flowing 'longshore' current is suggested as being the result of persistent wave attack from the southeast across the shallow sea which persisted at this time. The occasional reversals in direction, which occur on the northerly side of the distributary channel outlets, are probably caused by the eddy effect induced when a powerful stream of water enters a relatively still body of water.

Later, as the delta steadily built upward and general basin shallowing took place, these barrier-bar sand bodies migrated with the prograding edge of the delta, but eventually disappeared when the shallow sea situated to the east and southeast emerged into the intertidal zone (see later - facies association 2). Palaeocurrent data taken from a later generation of these protective barrier-bars show the same three-fold current system, plus a fourth current directed shoreward, and possibly representing either tidal flood currents or washover fan currents (Fig. 5-8A). Also contained within this later generation of barrier-bar sands are the large, isolated wedge-shaped interbeds of lithotype 4B, within which are occasional large-scale planar (alpha) crossbeds. These wedges are interpreted as possible aeolian dunes which developed upon the barrier-bars in a manner comparable to those described by Bart (1977).

Low Intertidal Mudflat (Lithotype 1A - upper)

Whilst seaward of the barrier-bar system rapid accumulation of fine-grained sediment occurred under quiet reducing conditions, episodically interrupted by oxygenated coarse-grained sediment-laden turbid bottom currents (producing lithotype 1B), to the north, within subregion CI of the basin, the very shallow submerged mudflat gradually emerged, with basin shallowing, into the low intertidal zone. The sediments deposited within this environment (constituting lithotype 1A) are host to numerous environmentally diagnostic sedimentary structures.

Sedimentary Structures and their Palaeoenvironmental Significance

The typical bedding found within lithotype 1A consists of flatly laminated to thinly bedded intercalations of greyish red shale, dark greyish red siltstone and very pale red to moderate red, fine to medium sandstone (Plate 5-5ë). Interrupting this flatly bedded sequence are medium to thick beds of dark greyish red siltstone within which an abundance of soft-sediment deformation has occurred (see bottom of Plate 5-5e). The surface expressions of such layers reveal extensive foundering of silty layers within clayey layers (Plate 5-5f), highly distorted surfaces (Plate 5-6a) and/or occasional mounds (Plate 5-6b - cf. monroe structures). An examination of these deformed beds showed a large variety of soft-sediment deformation structures, encompassing the four main groups outlined by Reineck and Singh (1975, p.75), namely convolute bedding, load structures, ball-and-pillow structures and slump structures.

Convolute bedding is not common, but where present it involves layers of silty shale which are both underlain and overlain by flatly laminated shale and siltstone (Plate 5-1g), indicating the penecontemporaneous nature of the structure. Liquefaction of a hydroplastic layer induced by the shearing action of a current flowing over the surface

(Sanders 1965), differential loading from overlying sediments (McKee and Goldberg 1969), lateral intrastratal flow induced by differential liquefaction (Williams 1960), local liquefaction induced by water expulsion and sediment compaction during intertidal exposure (Wunderlich 1967) and seismic shock are all mechanisms which have been forwarded as explanations for convolute bedding. Because of the lithologic uniformity between the convoluted layers and the bounding layers a mechanism involving lateral intrastratal flow induced by the shearing action of a current and causing differential liquefaction seems the most likely mechanism. The intertidal deposition of this lithotype (see later), however, along with the rarity with which this structure is encountered, suggest that Wunderlich's thesis should not be altogether discounted. Where the sediment cohesion has been overcome by the shearing action of the current, the bedding becomes disrupted, as shown in plate 5-6c.

Ball-and-pillow structures (commonly called pseudonodules) are very common throughout lithotype 1A (Plates 5-6d and 6e). Contorted internal laminations are often well preserved within these structures. Ball-and-pillow structures which have been weathered out often display wrinkle marks (Plates 5-5f and 6f) or cracks (synaeresis; Plate 5-6g) upon their curved surfaces, and sometimes surface markings suggestive of flowage (Plate 5-6h). Possible lines of flowage are also occasionally seen on what appear to be 'unloaded' surfaces (Plate 5-7a). Ball-and-pillow structures involve the vertical displacement of isolated load structures down into a hydroplastic layer (Kuenen 1958), although a degree of lateral displacement seems to have occurred with some of the structures. Such structures are known from both shallow and deep water environments.

Irregular bedding surfaces (e.g. Plate 5-7b) are suggested as being the result of differential loading, as load structures are very common within this lithotype with all size scales being represented (see Plates 5-1f and 7c). Load structures and their accompanying flame structures originate when sediment is deposited upon a hydroplastic layer causing differential loading and adjustment by vertical displacement (see Reineck and Singh 1975, p.76, Fig.125). Small-scale structures resembling Bhargava's (1972) "tadpole nests" (Plate 5-7d) are found to be the basal moulds of such loaded sediment (cf. Plates 4-2a and 2b).

A special type of load cast, namely 'load-casted ripples' (Plate 5-7e) result from differential deposition caused by the formation of isolated ripple crests which then sink into the underlying layer (Dzulynski and Kotlarczyk 1962). Load structures displaying a preferred orientation (Plate 5-7f) are occasionally seen suggesting either minor slumping of a hydroplastic layer, or its deformation by current shear (cf. convolute bedding). A wide variety of depositional environments are conducive to the formation of load structures, including fluvial, channels within tidal mudflats, and those typified by turbidity currents.

The deformed sediments within lithotype 1A do not always display perfect formation of one of the above-mentioned structural types. Often mixtures are found, such as poorly formed ball-and-pillow structures and highly contorted laminations (Plate 5-7g). Plate 5-1f shows both ball-and-pillow structures and 'load-casted ripples'. Also seen in this same outcrop, but looking down upon the bedding plane, are long, thin and wavy, upward sand injection structures (Plate 5-7h) originating from the 'load-casted ripples'. Downward sand injection structures of synaeresis origin are very common (Plate 5-8a), being formed by the dewatering and 'tearing' of a mud layer when loaded by the deposition of an overlying coarser layer. Evidence of cracking on a larger scale is rare, but a possible siltstone dyke is shown in the top right-hand-corner of plate 5-6e, suggesting more brittle deformation of the surrounding flatly laminated siltstone.

Another structure not common within lithotype 1A, but one that also supports the movement of cohesive sediment, is "Runzelmarken" (Häntzschel and Reineck 1968), or 'wrinkle marks'. Reineck (1969) proposed that such structures formed by wind acting upon the surface of wet, fine-grained sediment either exposed to the air, or covered by no more than 1cm of water. Within lithotype 1A these structures are found both upon flat bedding surfaces and within the troughs of ripple marks (Plate 5-8b).

Surfaces displaying ripple marks are abundant throughout lithotype 1A. Common types are symmetric wave ripples, with an average wavelength of 6cm (Plate 5-8c), polygonal interference ripples (Plate 5-8d) and microripples (sometimes flat-topped), having a wavelength of less than 5mm (Plate 5-8e), indicative of very shallow water (Singh 1969). Flat-topped ripples, indicative of exposure (Plates 5-8f and 8g), are also common. The ripples shown in plate 5-8f were asymmetric, but erosion of the crestlines has left a current lineation upon the exposed silty sediment. Plate 5-8g shows symmetric wave ripples (flood tide oriented) which have been planed-off by the ebb tide. Late stage runoff has produced a secondary ripple set within the troughs leading to the adjacent shallow channel within which are found ebb tide oriented asymmetric current ripples.

Rare isolated lenses of sand occurring at regular intervals along the bedding plane are occasionally found which give the impression of a set of isolated (or starved) ripple marks (Plate 5-9a). A close examination of the internal structure of both these lenses and the adjacent layers, however, reveals a very different origin to that of starved ripples. A truncated parallel lamination is present within these isolated sand lenses, not the expected ripple cross-lamination. Adjacent silty layers generally show parallel lamination, or planar cross-lamination. Occasional silty layers, however, display ripple cross-lamination. Currents forming ripple cross-lamination normally erode sediment from the stoss side of the ripple marks and deposit sediment upon the lee side. In this case stoss side erosion scoured down through the underlying layer, leaving only a series of regularly spaced, isolated lenses ('pseudoripples') preserved by lee side deposition of the overlying ripple cross-laminated silty layer.

Lenticular bedding with flat, connected lenses of sandstone is not uncommon (Plate 5-9b). According to Reineck and Singh (1975, p.101) lenticular bedding is dominantly of current origin, but alternating periods of slack water are essential. As such, tidal environments are strongly favoured for the formation of such bedding, and, in fact, Wunderlich (1970) termed such bedding "tidal bedding". However, this type of bedding has also been recorded from delta front and subtidal environments.

Solitary sets of beta crossbedded sandstones are occasionally found (Plate 5-9c and Fig. 2-3D). Although generally composed of well sorted sand, occasional large shale clasts are incorporated, often where local current activity has scoured down into the underlying layer (Plate 5-9d). Rare occurrences are found where very well rounded, small siltstone cobbles are incorporated within the well-bedded shale and siltstone sequence (Plate 5-9e). Such occurrences are common in present day beach and tidal flat environments. Also, rare structures are found for which no ready explanation of origin is available. Such structures are classed as problematica (e.g. Plate 5-9f).

Palaeocurrents of the Low Intertidal Mudflat

Palaeocurrent data for lithotype 1A is limited because attention was concentrated mainly upon the sandy facies of the ABC Range Quartzite (facies associations 3 to 10 inclusive). Nevertheless, analyses of asymmetric ripple marks indicate a bimodal palaeocurrent distribution (Fig. 5-4B) which, when coupled with such ripple mark associations as shown in plates 5-8f and 8g, suggest a strong tidal influence upon this mudflat. Further, the presence of wrinkle marks, microripples and flat-topped ripples indicate extremely shallow water and periodic exposure, signifying an intertidal depositional environment. The palaeocurrent bimodality results from wave activity directed toward the west (i.e. shoreward) during flood tide and current activity in broad, shallow channel zones directed eastward (i.e. basinward) during ebb tide (see Plate 5-8g).

Two analyses of crossbedding data both produced unimodal palaeocurrent distributions. At section ARR a westerly directed current is indicated, possibly related to the flood tide process, whilst at section BGL a northerly directed current is present. This may represent a distal deltaic current transporting sand northward from the area of delta development in subregion WII. Alternatively, the palaeocurrent directions from these two crossbedding analyses may be related to local basin topography. An isopach map of the upper Brachina Subgroup (Fig. 5-10) shows within subregion CI two local depressions separated by a narrow ridge, and all trending north-northwest - south-southeast. As can be seen from the locations of sections ARR and BGL, these two crossbedding palaeocurrent directions trend toward the deepest portion of one of the local depressions. Against this interpretation, however, is the presence of a slurry-slump structure (Plate 5-1c), found at Chace Range (CCR), which is directed westward, away from the centre of the depression. Two analyses of parting lineation produce a bimodal distribution with modes parallelling the northerly crossbedding direction and the easterly ripple mark direction.

PALAEOGEOGRAPHY - TIME INTERVAL II

Three major depositional environments constitute the palaeogeography of the basin at this time (Fig. 5-11). Within each of these environments sediment accumulation occurred under different energy conditions. Within subregion WII high energy conditions prevailed producing an extensive intertidal deltaic plain (facies association 5), cut by relatively stable distributary channel outlets (lithotype 3A). Wave activity, generated above the deltaic bottomset environment situated to the east (lithotype 1B), reworked the distal sands of the intertidal deltaic plain to produce a protective barrier-bar system (facies association 4).

Such delta systems, when characterized by high input of fine-grained suspended sediment, produce extensive subaqueous bottomset deposits composed of high water content, unstable muds, and hence typified by a wide variety of soft-sediment deformation structures. In this case, however, because of the shallow nature of the receiving basin and the presence of a strong northerly littoral drift current, much of the fine-grained material brought into the basin was transported into subregion CI where it accumulated within a dominantly low intertidal mudflat environment (lithotype 1A).

EVIDENCE OF PRECAMBRIAN LIFE(?)

Newly Discovered Evidence(?)

Within lithotype 1A in the central Flinders Ranges four *in situ* localities have been found where there are occurrences of an unusual sedimentary structure (Fig. 5-9). The first of these localities (BYR1) was found by Mr. K. Moriarty in 1975 whilst on a field excursion with Flinders University School of Earth Sciences. The remaining localities (*viz.* BGL, CBC and MRC) were all found by the author during the course of this study. In addition, from a fifth locality (BNC) one sample of the same structure was recovered from creek debris.

In section parallel to bedding these structures display a circular to ovoid outline (Plate 5-10a). Their shape in vertical section is commonly cylindrical, with either a flat, U-shaped base (Plate 5-10b), sometimes lopsided and/or with a shallow central indentation (Plate 5-10c), or a rounded to hemispherical base. A conical shape is also sometimes displayed, these often opening out in an upward direction like the mouth of a trumpet. Where complete, the width of these structures (disregarding the trumpet-like upper rim) is usually less than their depth, averaging 8 to 9cm, but ranging from 1.5 to 15cm. Often their surface is nearly smooth, but on some a series of fine longitudinal ridges extends along the cylinder or cone and fans out on the trumpet-like upper rim. Occasionally these ridges continue across the base of the structure as a series of radial markings (Plate 5-10d). The significance of these markings is questionable, however, as they are often indistinguishable from, and grade into slicken-sides, a feature which is common within lithotype 1A.

The structures are found both within medium bedded massive siltstones (see Plate 5-10a), and shales interbedded with thin siltstones and fine sandstones (see Plates 5-10c and 10e). In the latter instance the structures generally terminate their base at an underlying siltstone or fine sandstone layer (Plate 5-10e), but structures resting solely within shale are not uncommon (see Plate 5-10c). Occasionally the lamination within the surrounding sediment has been noted to be downwarped immediately adjacent to the edge of these structures. Invariably the structures terminate upward into a current laminated siltstone or fine sandstone layer, and this same sediment is found to comprise part of the structures' infill. At the bottom of the infill are frequently found sand and intraclasts of shale and siltstone (Plate 5-10b).

The nature of the infill indicates that these structures were present as cavities within the sedimentation surface prior to the deposition of their infill and the subsequent overlying layers. These cavities thus acted as traps for any coarse sediment being carried in traction before being buried by the finer sediment deposited from suspension. This syngenesis precludes an origin of these structures by post-depositional slumping, or by diagenetic water escape (e.g. cylindrical pipes and sand volcanoes, see Hawley and Hart 1934, Gabelman 1955, Conybeare and Crook 1968 (p.242, Plate 77); or spring pits, see Quirke 1930). Also, these structures clearly did not have an epigenetic concretionary origin. Such syngenetic structures could, therefore, only be formed either inorganically as toroids, or organically either as burrows (dwelling or hiding) of actinian-like creatures, and preserved as trace fossils such as Bergaueria Prantl 1945, Conostichnus Lesquereus 1876 and Solicyclus Quenstedt 1879, or by the burial of such cup-shaped coelenterate(?) body fossils as Arumberia Glaessner and Walter 1975, Baikalina Sokolov 1972, the "Erniettomorpha" of Pflug (1972), or Namalia Germs 1968.

Toroids according to Conybeare and Crook (1968, p.34 and Plate 54, p.190), are the casts of circular scour pits formed by whirlpool action generated along the boundary between two currents flowing in opposite directions. As such they are found in high energy shallow water environments of both marine and non-marine affinities. In cross-section parallel to bedding toroids are circular in shape and their diameter is usually greater than their depth. The external morphology of their casts is characteristically swirled, like that of a "folded bun", generally with a smooth surface, although the formation of roughly concentric patterns by scouring is common. Internally, such toroid casts display structural and textural homogeneity.

The cylindrical structures described by Dionne and Laverdiere (1972) are believed to be a special type of toroid, being filled immediately after the scouring phase by the sediment initially eroded from the hole. This immediate resedimentation occurs whilst whirlpool motion is still present in the hole, resulting in an internal 'cone-in-cone' arrangement. Dorr and Kauffman (1963) describe "rippled toroids" from the Mississippian Napoleon Sandstone of southern Michigan, in which the base of the structures display a series of ridges radially divergent from a central indentation. These "rippled toroids", however, are considered to be excellent examples of the trace fossil Astropolithon Dawson 1878, believed to be the feeding burrow of a coelenterate (Alpert 1977, p.6, Table 1; Crimes et al. 1977, p.112).

As can be appreciated from the description of the structures found within lithotype 1A given above, their morphology does not resemble that of a "folded bun", nor does the infill display an internal 'cone-in-cone' arrangement. Also, the radial patterning occasionally seen upon the base of these structures seems an unlikely feature to be formed by the swirling action of water, and is otherwise known only to occur on structures produced organically and preserved as either trace fossils or body fossil impressions.

Analagous Recent organically formed structures are known to be produced by the living sea anemone Phyllactis. In a study on the burrowing activity of this anemone, Mangum (1970, p.324) states that by a series of rhythmic contractions of the column musculature around a hydrostatic skeleton, fluid is alternately forced into the sandwithdrawn from an inflatable pedal disc, forcing it into the sand. "After the pedal disc reaches a hard substratum and sand surrounds the column [burrowing] ceases".

Also, such actinians produce burrows which are deeper than they are wide (Frey 1970, p.308), and they cause a downwarping of the sediment lamination (*see* Shinn 1968, p.891, Plate 112).

From the above considerations it is believed that the structures found within lithotype 1A were organically produced, and represent either the burrow of a primitive actinian-like creature, or the body fossil of a cup-shaped coelenterate(?). The full significance of these structures is to be presented in a later publication (Plummer *et al.* *in prep.*)

Further Evidence(?)

The structure known as *Bunyerichnus dalgarnoi* Glaessner 1969 (Plate 5-10f), found by Glaessner toward the base of lithotype 1A within Bunyeroo Creek (BNC) in the central Flinders Ranges in 1966, was originally interpreted as the track of a naked mollusc (Glaessner 1969, pp.376-379). Germs (1973, p.70) similarly interprets the trail *Buchholzbrunnichnus kröneri* Germs 1973, found within the late Precambrian Nama Group in South-West Africa. Both specimens, however, are solitary, and their metazoan origins have been questioned, with the form of *Bunyerichnus* being described by Jenkins (1975, p.19) as "a unique accidental set of markings made by a tethered implement moved by a current". Also, recent work on the role of vorticity in the development of lineations by wind erosion indicates that structures very closely resembling both *Bunyerichnus* (*see* Whitney 1978, p.11, Fig.4B) and *Buchholzbrunnichnus* (*see* Whitney 1978, p.11, Fig. 4E) can be readily produced by vorticity along lines of wind flow. Nevertheless, the general overseas opinion holds these two structures as being trace fossils, and hence the records of pre-Ediacaran macroscopic organisms.

Also found during the course of this study, in the section East of Point Bonney (EPB), was a sample bearing what appear to be the impressions of small medusoids (Plate 5-10g). Although the sample was from float, it was from approximately the same stratigraphic level as Bunyerichnus. The presence of both positive and negative features upon the one face of this sample eliminates such an inorganic explanation of their origin as by water escape, and suggests a possible derivation by the burial of small soft-bodied organisms. These impressions (Plates 5-10g and 10h) lie on the basal surface of a ripple cross-laminated fine sandstone layer which originally lay above a shale layer. This form of preservation is typical of many medusoid and annelid fossils of the famed Ediacara assemblage (Wade 1968) found within the uppermost Precambrian Pound Subgroup of the Adelaide Supergroup.

It seems probable, therefore, that life forms ancestral to the Ediacara fossil assemblage existed during the early stages of the upper Brachina Subgroup, some 1,200m below the Ediacara Member of the Rawnsley Quartzite (Pound Subgroup), at an age approximated by Jenkins (1978, *pers. comm.*) to be 640 m.y.

SEDIMENTOLOGY - TIME INTERVAL III

Intertidal Mudflat (Facies Association 2)

As high sedimentation rates continued to exceed the rate of basin subsidence, basinwide shallowing occurred. Eventually, the vast mudflat regions surrounding the intertidal deltaic plain emerged to well within the intertidal zone. As a consequence, a new set of energy conditions now controlled the sediment accumulation process, and both lithotypes 1A and 1B were gradually replaced by the shale, siltstone and sandstone deposits of facies association 2.

Sedimentary Structures and their Palaeoenvironmental Significance

The suite of sedimentary structures displayed by the deposits of facies association 2 has much in common with that of lithotype 1A.

Although dominantly a shale and siltstone sequence, lenticular developments of fine sandstone are ubiquitous. These lenticular sandstones range in appearance from thin beds, to thick isolated lenses (Plate 5-11a) and thin isolated lenses with parallel lamination or ripple cross-lamination (Plate 5-11b). It is these lenticular sandstones that display the majority of the diagnostic sedimentary structures found within facies association 2.

Structures indicative of periodic subaerial exposure are quite commonly found. The upper surface of some of the sandstone lenses have an irregular, bumpy appearance interpreted as the preserved impressions of raindrops (Plate 5-11c). Sets of ripple marks often display flattened crestlines (Plate 5-11d) eroded off during the final stages of emergence through a slowly falling water level (e.g. an ebbing tide). Occasionally a ripple set is found to have planed-off crestlines, but present within the troughs, immediately adjacent to each flattened crest, is the eroded sand which has not been flushed away. The sample shown in plate 5-11e has the eroded sand lying within the ripple trough on the shoreward side of each ripple crest (in this case toward the southwest). This suggests that, as the tide ebbed toward the east, small water ripples, generated by northeasterly winds, were responsible for the planing-off process. This agrees well with Tanner's (1958) thesis on the origin of flat-topped ripple marks. Also occasionally seen are flat-topped ripples with a secondary, unaltered parallel set developed within the troughs (Plate 5-11f), and ripples with rounded crests and pointed troughs (Plate 5-11a), both suggestive of falling water level and ultimate emergence (Reineck and Singh 1975, pp.368-369).

Sets of ripple marks showing no alteration to their form, however, are more commonly found. In general, such ripple mark sets were wave generated, displaying symmetric (average wavelength of 7cm) to slightly asymmetric (average wavelength of 10cm) form (Plate 5-11h). Interference wave ripples are commonly produced when two wave directions have acted upon the same sediments (Plate 5-12a; cf. Reineck and Singh 1975, p.368, Fig. 526i). Some interference ripple sets are encountered which involve current ripples as the major set, but with wave ripples generated simultaneously within the troughs (Plate 5-12b; cf. Reineck and Singh 1975, p.367, Fig. 526g). Combinations of current and wave activity along with "capillary wave" action produce a variety of ripple structures. An example of the combination of wave, late-stage runoff and "capillary wave" activity is seen in plate 5-12c. Here, late-stage runoff down the troughs of a set of wave ripples has produced a secondary set of symmetric ripples. "Capillary waves" have then planed-off the major set during emergence.

Both the exposures shown in plates 5-12c and 12d show the casts of shrinkage cracks generated within a thin shaly siltstone by dewatering and 'tearing' induced by loading from an overlying sandstone layer. Such features only develop when a 'coarse' layer is deposited upon a cohesive, relatively finer grained layer, for when such 'coarse' sediments are deposited upon a dried or compacted finer grained layer no such synaeresis cracking is induced and an essentially flat boundary is preserved (see Plate 5-11b). In some cases, however, displacement of a saturated underlying finer grained layer by downward 'coarse' sediment protruberances (load structures) occurs without causing synaeresis cracking (Plate 5-12e). The type of boundary, therefore, preserved between two layers of differing grain size (and where the overlying layer is coarser) shows a sequence of structures which is dependent upon the amount of water contained within the finer, underlying layer. Naturally, combinations of these three states occur, as seen in plate 5-4d. Here, initially, the underlying shale layer was saturated, allowing the development of load structures from the overlying sandstone layer. With this sand intrusion water was expelled from the underlying clay, decreasing its cohesion and causing synaeresis cracking during the final stages of the loading process.

Within facies association 2 such vertical penecontemporaneous deformation of saturated sediment is common. Load and flame structures developed within siltstones (Plate 5-12f), ball-shaped load structures still connected to the generating sandstone layer (Plate 5-12g) and isolated ball-and-pillow structures (Plates 5-12h and 13a) are all readily found. Occasionally deformation by lateral movement of a cohesive layer along its bedding plane occurs, producing a wrinkle structure (Plate 5-13b).

Many sandstone lenses within facies association 2 display irregular bases suggestive of erosion (Plates 5-13c and 13d). The basal sandstone layer often contains an abundance of incorporated, locally derived shale clasts (Plates 5-13e), which represent a type of lag deposit. Shale clasts are also common within scour-and-fill structures present within the same sandstone lenses (Plate 5-13f). These lenticular sandstones are interpreted as channel-fill sands. The internal structure of the infilling sand varies from ripple drift cross-lamination (kappa or lambda cross-stratification of Allen 1963; see Fig. 2-3, and Plate 5-13d), planar (gamma) crossbedding (Plate 5-13c), or a herringbone arrangement of planar (beta) crossbedding, often capped by ripple (mu) cross-lamination (Plate 5-13g). In plate 5-13g the bottom crossbed set is oriented shoreward (i.e. toward the west), and probably represents deposition during a flood tide, whilst the overlying crossbed set is aligned toward the southeast. Overlying the foreset lamination of this latter crossbed set is a parallel lamination which is interpreted as representing the topset lamination of the megaripple which produced the crossbedding. This is then overlain by a ripple cross-lamination, the surface expression of which shows a series of planed-off crestlines. This sequence of crossbed foreset, crossbed topset, ripple cross-lamination and planed-off ripples suggests formation by a current steadily decreasing in strength as shallowing, and finally emergence occurred. The southeasterly orientation of the sequence suggests an ebb tide current as being responsible. Other styles of crossbedding are only rarely seen within facies association 2, including the xi cross-stratification shown in plate 5-13h.

In a few localities within subregion CI lenticular developments of coarse, to very coarse sandstone are found within the top 70m of facies association 2 (Fig. 6-2A). Petrologically these sandstones are similar to those comprising facies association 9 (see Chapter 6) and their significance will become apparent when the palaeogeography of the second phase of the upper Brachina Subgroup is discussed.

Palaeocurrents of the Intertidal Deltaic Plain and the Intertidal Mudflat

Ripple marks and crossbedding provided the palaeocurrent data analysed from facies association 2, whilst from the adjacent and intertonguing lithotype 5A crossbedding, ripple marks and rare flute cast directions were analysed. The intertonguing relationship between facies association 2 and lithotype 5A is shown in figure 3-2. The results of the palaeocurrent analyses are presented on figure 5-8B.

As with the basal portion of lithotype 5A a bimodal palaeocurrent distribution is present in the quartzites which dominate the upper portion of lithotype 5A in the western region. The dominant mode trends from the west and fans out both to the north and south as it leaves the basin margin in the region of Alligator Gorge. A second such current is present at Barunga Gap (BRG) to the south (see Fig. 5-8B) where crossbedded sandstones outcrop which are presumed to be equivalent to lithotype 5A of the ABC

Range Quartzite. Between these two eastwardly flowing current systems, at Mount Fergusson (MTF), a bimodal palaeocurrent distribution is present having northerly and southerly modes. These two current directions are preserved, though not together except at MTF, throughout the distribution of lithotype 5A within the western region, and are thought to represent the bottom currents produced by "the entrainment of underlying salt water by the rapidly flowing fresh water effluent [that] converge at the river mouth" (Coleman and Wright 1975, p.112). As the rapidly flowing fluvial current leaves the basin margin and fans out, it loses its intensity and drops its coarse sediment load forming a delta. The presence of occasional scour (or lunate) megaripples (Plate 5-3h) and flute casts within the intertongues of lithotype 5A within facies association 2 suggest slightly stronger current activity than present across the remainder of the lithotype 5A deltaic plain, which is dominated by straight, to weakly sinusoidal-crested megaripples (Plate 5-3g). These intertongues represent shallow, migrating channel zones within which the fluvial current was concentrated.

Upon the surrounding intertidal mudflat (facies association 2) a bimodal palaeocurrent pattern is seen in both the ripple marks and the crossbedding (Fig. 5-8B). In both cases the dominant mode trends toward the southeast whilst the minor mode is oriented toward the northwest. These opposing directions are interpreted as indicative of the tidal influence upon sedimentation. Although many readings were taken from wave ripples, and hence are revealing the palaeowind direction, it is within such tide-dominated environments that such palaeocurrent distributions occur in association with such sequences of sedimentary structures as shown, for example, in plate 5-13g.

In the region of intertonguing between facies association 2 and lithotype 5A, confluence between the deltaic and tidal current systems is reflected within the palaeocurrent data, producing trimodal, and rarely quadrimodal distributions (see Appendix III, page A19).

The Problem of Colour

Iron oxides are known to be the source of most pigmentation within fine-grained rocks. The oxidation state of the iron, however, is important in determining just what colour the rocks take. Tomlinson (1916) showed that the ferric/ferrous ratio of red and purple slates was greater than 1.0, whilst for green slates it was less than 0.5, although the total iron content for both slates was about the same. In a predominantly green sequence, where diagenetic iron reduction, or secondary chloritization has not taken place, the origin of the green colouration, and hence the low ferric/ferrous ratio, must be considered in the light of environmental conditions of deposition. Haematite is generally the ferric oxide responsible for the red colouration found in continental and other readily oxidized sediments, whilst within their marine equivalents a green colouration is often present because of the "reduction of colloidal ferric iron by the abundant carbonaceous organic matter on the sea floor and redistribution of the ferrous ions during authigenesis of pyrite, glauconite, [chlorite and illite]. The amount of organic matter in marine sediments decreases sharply a few inches below the water-sediment interface so that most such reduction must be a surface phenomenon rather than a diagenetic one" (Blatt et al. 1972, p.368).

From the sedimentary structures and palaeocurrent data contained within facies association 2 an intertidal, and hence readily oxidizing environment of deposition is indicated. The dominant shale-siltstone lithology of this facies association, however, displays a green colouration which suggests that the depositional environment was reducing, or either diagenetic reduction of secondary chloritization of an originally red and purple lithology has occurred.

Diagenetic reduction of originally red and purple fine-grained rocks, according to van Houten (1961, 1968), generally involves a loss of total iron, whilst the ferrous iron content remains fairly constant. The process involves the reduction of ferric iron, and the flushing away of soluble ferrous iron by formation waters affected by small-scale increases in pH either locally, forming green ovoid reduction spots, or along porous bedding and/or joint planes, forming green layers and borders. Such a process, however, is not considered pervasive enough to cause such a uniformly even colour change necessary to have produced the hue of facies association 2.

No conclusive evidence is available to support secondary chloritization as the cause of the green hue of facies association 2. Although chlorite is the dominant green mineral within these shales and siltstones, much of it is present as the alteration product of devitrified volcanic glass (see Chapter 8). A greater percentage of devitrified volcanic glass altered to chlorite, however, is present within lithotype 1A, which still exhibits a deep red-purple hue. Also disfavouring secondary chloritization is the persistence of regional sedimentary facies arrangement throughout the Flinders Ranges. Although the entire shale-siltstone sequence of the Brachina Subgroup (i.e. Moolooloo, Moorillah (lithotype 1B) and Bayley Range (facies association 2) Formations) are green within the southeastern Flinders Ranges (the eastern region and subregion WI), only facies association 2 is green within the central and western Flinders Ranges (subregion WII and the central region). Lithologically there is no explanation as to why any secondary chloritization process should select only this very broad horizon, yet ignore the two lower, lithologically very similar horizons (which are both red-purple). Also, the overlying shales of the Bunyeroo Formation are everywhere red-purple - even in the southeastern Flinders Ranges where they should also have been invaded by secondary chlorite, if secondary chloritization was responsible for the green colouration of the sediments within this region. The green colouration of facies association 2, therefore, is believed to have been due to the presence of ferrous iron within the sediments at the time of deposition within a reducing environment.

By studying the included sedimentary structures of each component lithotype of the upper Brachina Subgroup specific palaeoenvironmental conditions of deposition are determinable. As shown above, using this method facies association 2 was undoubtedly deposited within an intertidal environment -- an environment within which oxidized sediments are usually found (displaying red-purple colourations such as those of lithotypes 1A, 5A and 5B). The presence of ferrous iron within facies association 2, therefore, suggests that either free oxygen was not present - incorrect because of the oxidized sediments both immediately below and above - or free oxygen was present, but being utilized by some other process and hence not available to oxidize the sediments.

Here, supporting Blatt <u>et al</u>.'s (1972, p.368) thesis that organic reduction of collodal ferric iron enables the authigenesis of such minerals as illite and chlorite, studies of present day intertidal mudflat environments analagous to that postulated for facies association 2 (<u>i.e.</u> areas where sedimentary structures are well preserved and the sediment is predominantly mud) show that very little macroscopic organic activity occurs to bioturbate, and hence oxidize the sediment. In these areas, however, abundant microscopic organic activity occurs utilizing whatever free oxygen is available with the result that the sediments are reduced and black. It is herein suggested that the reduction of the facies association 2 deposits was due to an abundance of microscopic organic activity at the time of deposition within an intertidal mudflat environment. In contrast, the red-purple intertidal mudflat sediments of lithotype 1A were deposited within a subenvironment not conducive to the proliferation of abundant microscopic organic matter - perhaps the low intertidal zone where energy conditions were too great and/or sedimentation too rapid.

PALAEOGEOGRAPHY - TIME INTERVAL III

With continued rapid ingression of material into the basin from the west and sedimentation rates exceeding those of basin subsidence, an overall shallowing of the basin occurred. With this shallowing the mudflat expanses of the central and eastern regions of the basin emerged well into the intertidal zone (facies association 2). As a consequence of this, the energy conditions which controlled sedimentation changed from being wave-influenced to being tide-influenced. The lack of wave energy acting upon the distal sands of the intertidal deltaic plain of subregion WII (facies association 5) resulted in the abandonment of the protective barrier-bar system (facies association 4) and the replacement of the relatively stable outlet channels (lithotype 3A) by a migrating outlet channel arrangement, thus allowing rapid progradation of the intertidal deltaic plain across the entire western region (Fig. 5-12).

EVOLUTION OF THE DELTA SYSTEM

Using the criteria of Coleman and Wright (1975) and Galloway (1975) the variations in the delta forming processes and the resultant evolution in delta morphology can be traced for the delta system which developed during this first phase of upper Brachina Subgroup sedimentation.

The sediment pile which comprises the delta accumulated within an intracratonic basin, and as such, rests solely upon continental crustal material. The lack of extensive marine evaporites beneath such deltaic piles, coupled with the lack of associated deep-water fan and pelagic facies, led Audley-Charles et al. (1977) to define such systems as "enclosed intracratonic deltas". The intracratonic basin into which this delta system of the upper Brachina Subgroup developed was esentially open-ended (i.e. a large basin extended seaward of the region of delta development) and had an active, though steadily abating zone of subsidence lying seaward of the shoreline (see Coleman and Wright 1975, p.115, Fig.9 configuration IV).

As seen above, by using such criteria as sedimentary facies association relationships, environmentally significant sedimentary structures and palaeocurrent distributions, this enclosed intracratonic deltaic system of the upper Brachina Subgroup can be divided into two major types.

The initial delta type developed upon a very shallow submerged platform. Progradation of the deltaic topset deposits (the intertidal deltaic plain - facies association 5) halted at the break of slope leading into a region of very quiet subtidal deposition (the deltaic bottomset environment - lithotype 1B). As such, therefore, an open sea effectively lay adjacent to the delta. Across this open sea a significant amount of wave power was generated to act upon the distal reaches of the delta and build a protective barrier-bar system (facies association 4). Moderate fluvial and tidal activity, along with a strong northerly directed littoral drift, caused the delta to grow parallel to the depositional strike forming a sand distribution pattern similar to Coleman and Wright's (1975, p.123, Fig.10) type VI (see Fig. 5-13). Much of the fine-grained detritus debouched into the basin through the delta system was carried by the northerly littoral drift into the central region where a vast, low intertidal mudflat developed (lithotype 1A).

Such wave-dominated deltaic systems typically display a 'single' type of distributary channel pattern (see Coleman and Wright 1975, p.109, Fig. 7c) consisting of relatively few channels (lithotype 3A) all originating from a nearly common point at the head of the delta. Progradation of such delta systems is by channel extension (see Coleman and Wright 1975, p.114, Fig.8) whereby one of the distributary channels carries the majority of the effluent at any one time, and hence progrades, whilst the other distributaries are attacked by wave action, allowing little progradation, but building protective barrier-bars. Such systems are defined as 'fluvial and tide modified, wave-dominated deltas' (see Galloway 1975). The Burdekin and Nile Deltas are modern day equivalents (see Fig. 5-13).

With sedimentation rates exceeding those of basin subsidence a general shallowing of the basin occurred causing variations to occur in the delta forming processes (see Table 5-1). With these changes a new deltaic system gradually evolved from the initial fluvial and tide modified, wave-dominated system. This basin shallowing caused the offshore slope to steadily decrease, which, in turn, led to the submerged deltaic bottomset environment to yield to a vast intertidal mudflat (facies association 2). This change produced an increase in the influence of tidal activity upon the system at the (near total) expense of wave activity and littoral drift. Consequently, sand was now distributed parallel to the depositional slope in a pattern resembling Coleman and Wright's (1975, p.123, Fig.10) type II (see Fig. 5-13). With the lack of wave activity available to rework the distal reaches of the delta into barrier-bars, tidal sand ridges accumulated between the distributary channels, which now flowed across the intertidal deltaic plain (facies association 5) in a complex 'rejoining' pattern (see Coleman and Wright 1975, p.109, Fig.7b). Delta progradation occurred by a channel switching procedure, involving shifts in the major distributary channel far upstream in the intertidal deltaic plain and a corresponding new course for the river (see Coleman and Wright 1975, p.114, Fig.8). This system represents a 'fluvial modified, tide-dominated delta' (see Galloway 1975), not unlike the present day Mahakam Delta (see Fig. 5-13).

Due to the lack of coarse detritus present within the lower Brachina Subgroup throughout the basin of deposition, uplift of a formerly peneplained Gawler Craton is suggested as the cause for the huge sediment influx responsible for the development of the upper Brachina Subgroup. The very well rounded nature of the sands in a majority of the sandy lithotypes (i.e. 4A, 4B and 5A) can be explained in terms of incessant and prolonged reworking by wave and tidal activity. However, the sands of lithotype 3A (distributary channels) also display very well rounded grain outlines, yet no (or at most, very little) wave or tidal activity upon these sands is suggested by the abundant trough crossbedding present with its (almost) unimodal fluvial palaeocurrent distribution. The presence of haematitic rims on many of the very well rounded grains within this lithotype (see Chapter 8) indicate that they passed through a cycle of aeolian abrasion prior to possible lengthy transportation within a river system (the "Alligator River") draining a newly uplifted region far inland. Lengthy drainage systems are common for intracratonic deltas, and are termed "Amerotype drainage" by Audley-Charles et al. (1977).

PALAEOGEOGRAPHIC EVOLUTION THROUGH THE DELTAIC PHASE OF THE UPPER BRACHINA SUBGROUP

The palaeoenvironmental significance has been outlined for each suite of sedimentary structures contained within each of the nine component lithotypes of the first phase of the upper Brachina Subgroup. In collaboration with the determined palaeocurrent directions and the spatial and temporal relationships between these lithotypes, this sedimentology enables the palaeogeographic evolution of the sequence to be developed.

Prior to the ingression into the basin of the sand-sized sediment which produced the upper Brachina Subgroup (i.e.during lower Brachina Subgroup time), the palaeogeographic configuration of the basin was one of total, though shallow submergence. Tectonic activity during this period was minimal, with a stable, almost peneplained landmass to the west surrendering only fine-grained sediment to the gently subsiding basin. Turbid bottom currents deposited this fine sediment under aerobic conditions within shallow submerged mudflats along the western flank of the basin and within the central region. As these currents flowed toward the east and southeast they encountered a gentle palaeoslope which led them into a very quiet submerged mudflat where they deposited the remaining very fine-grained material of their load under anaerobic conditions (Fig. 4-1).

With the generation of tectonic instability upon the Gawler Craton to the west, and the subsequent production of abundant coarse-grained sediment, the scene was set for the change in style of sedimentation and a consequent palaeogeographic rearrangement within the basin, which is stratigraphically recorded by the upper Brachina Subgroup succession.

The initial influx of this coarse sediment into the basin occurred in the southwestern Flinders Ranges as the submerged distal portion of a delta system. The coarsest sediment was restricted to three broad zones (lithotype 3A), representing distal distributary channels. These were flanked by broad subaqueous levee deposits (lithotype 3B), which developed across the remaining portion of the shallow submerged mudflat. Upon reaching the gentle palaeoslope leading into the slightly deeper eastern region of the basin, these channels fanned out into broad, possibly coalescing distributary mouth shoals which intertongue with the fine-grained deposits of this submerged mudflat (deltaic bottomset - lithotype

1B) (Fig. 5-5).

With continued coarse sediment influx a deltaic plain developed, comprising delta topset sands and muds (facies association 5), cut by relatively stable distributary channels (lithotype 3A), which rapidly prograded out into the basin until the gentle palaeoslope was encountered. The halt in progradation at this break of slope, coupled with the high rate of sedimentation and the very shallow nature of the basin along this western flank, soon resulted in the distal deltaic subenvironments emerging into the intertidal zone. Wave activity, generated upon the shallow sea to the east and southeast, reworked the distal deltaic topset sands into a protective barrier-bar system (facies association 4), through which passed the major distributary channels. To the northeast, within the central region of the basin, fine-grained deposits accumulated in a vast mudflat within the lower reaches of the intertidal zone (lithotype 1A) (Fig. 5-11).

Gradually, as the high sedimentation rate caused basinwide shallowing, a vast mudflat emerged well into the intertidal zone (facies association 2) to completely surround the delta. With this emergence, the accompanying change from wave to tidal activity operating upon the distal reaches of the delta caused major morphologic changes in the delta configuration. The protective barrier-bar system disappeared and, as a consequence, the relatively stable distributary channels were replaced by a migrating outlet arrangement (Fig. 5-12).

This first phase of upper Brachina Subgroup deposition, therefore, was induced by tectonic instability upon the Gawler Craton to the west and involved the development of a barred, fluvial and tide modified, wave-dominated delta system fed by relatively stable outlet channels and surrounded by subtidal, and tidal to low intertidal mudflats. This system then evolved, with the emergence of the basin well into the intertidal zone, to a non-barred, fluvial modified, tide-dominated delta system having a migrating outlet channel system and surrounded by a vast intertidal mudflat (Fig. 5-13). This latter palaeogeographic configuration persisted until mild tectonic activity (Uplift I) uplifted a portion of the basin margin causing the erosion and redistribution of part of the deltaic sequence, a possible minor relocation of the main drainage system outlet and a complete rearrangement of the palaeogeographic configuration of the basin.

CHAPTER 6

PALAEOGEOGRAPHY OF THE UPPER BRACHINA SUBGROUP: SANDFLAT PHASE

Tectonic instability upon the Gawler Craton was responsible for the generation of the enormous amount of coarse detritus debouched into the basin through the "Alligator River Delta" system forming the deltaic phase of the upper Brachina Subgroup. The extension of this instability across the craton and into the basin ended the deltaic phase of sedimentation by causing a major palaeogeographic rearrangement. This new phase of upper Brachina Subgroup deposition - the sandflat phase - was subjected to two relatively major episodes of uplift, each followed by peneplanation and/or subsidence producing a unique set of sedimentary facies related to the pervasiveness of each episode across the basin.

The sequence of tectonic unrest during this sandflat phase of the upper Brachina Subgroup is shown diagramatically in figure 6-1. Uplift I involved uplift of the basin's western margin and the brief emergence of small islands within the basin. Peneplanation and/or subsidence then followed producing facies associations 6, 7 and 8. Again uplift of the basin's western margin occurred during Uplift II, but this time it was apparently accompanied by two possible pulses of island emergence within the basin. Again peneplanation and/or subsidence occurred, this time producing facies association 9, lithotype 10A and the basal deposits of the Bunyeroo Formation.

REGIONAL LITHOTYPE ARRANGEMENT

Due to the effects of Uplift I, lying within the vast area of facies association 2 accumulation within the central region is a broad, tongue-shaped sheet of cyclically interposed pale yellowish grey planar crossbedded, fine to medium feldspathic quartzose arenites (lithotype 7A) and pale greenish yellow rippled, silty fine to medium litho-feldspathic quartzose arenites (lithotype 7B). Within facies association 2 away from this sand sheet, but at approximately the same stratigraphic level, are occasionally found small lenses of coarse sandstone petrologically similar to those of facies association 9 (see later). Separating this facies association 7 sand sheet from the newly uplifted basin margin within subregion WII is a narrow belt of dusky red purple, shale pebble-rich siltstones and fine sandstones (facies association 6), whilst to the south of the uplifted margin a series of greyish red purple to light brown planar crossbedded, coarse litho-feldspathic quartzose arenites accumulated (facies association 8), (see Fig. 6-2A). These coarse sandstones migrated northward, burying facies association 6, as Uplift I subsided and the shoreline retreated (Fig. 6-2B).

Toward the top of facies association 7 are found a second pulse of lenticular coarse sandstone developments petrologically similar to those of facies association 9 (Fig. 6-2B). These mark the onset of Uplift II. This tectonic episode produced the conglomeratic blackish red purple trough crossbedded coarse feldspatho-lithic quartzose arenites of facies association 9 which are widely distributed upon facies association 7 in the central region, and either capping facies association 2, or found within the overlying Bunyeroo Formation in the eastern region (Figs. 6-3A, 3B and 4A). Meanwhile, uplift of the basin's western margin generated the light pink coarse quartzarenites of lithotype 10A. This lithotype migrated basinward infront of an advancing shoreline across parts of facies associations 7, 8 and 9 (Figs. 6-2B, 3 and 4).

SEDIMENTARY STRUCTURES, PALAEOCURRENTS AND PALAEOGEOGRAPHY

Although the facies associations of this second phase of upper Brachina Subgroup sedimentation do not contain the wealth of sedimentary structures, nor the abundance of palaeocurrent data as do those of the earlier deltaic phase, quite detailed palaeogeographic interpretations can still be postulated.

SEDIMENTOLOGY - TIME INTERVAL IV

Skirting the newly uplifted portion of the basin margin were deposited the dusky red purple, flatly bedded siltstones and fine sandstones of facies association 6 (Fig. 6-2A). Because of the very poor outcrop of this unit very few sedimentary structures are discernable. Except for some rare very low angle crossbedding, an abundance of flat shale pebbles (Plate 6-1a) is the only sedimentary structure seen in the parallel laminated siltstones. Although shale pebbles are not diagnostic of any specific environment of deposition, their formation from cohesive muddy layers being due either to desiccation, or to scouring by swiftly flowing currents, the rapid decrease in abundance of these shale pebbles within this facies association away from the region of uplift, along with the palaeoenvironmental significance of adjacent facies, indicates the depositional environment as being within the intertidal zone.

South of the newly uplifted portion of the basin margin and its siltstone skirt, in the vicinity of Alligator Gorge (Fig. 6-2A), a sequence of greyish red purple to light brown, planar crossbedded coarse to very coarse sandstones (facies association 8) were laid down conformably upon facies association 5. Planar (omikron) crossbedding, with set thickness and foreset slope averaging 8cm and 28^{o} respectively, is ubiquitous (Plate 6-1b). The presence of occasional herringbone arrangements to this crossbedding suggest deposition from traction currents within a tidally influenced environment. Occasional beds showing grading from granules and small pebbles at the base to medium sand at the top (see Plate 6-1b) indicates a turbid nature to some of these currents. The only other sedimentary structures seen within these sandstones are rare symmetric wave ripples (Plate 6-1c), occasionally flat-topped (Plate 6-1d), with wavelengths ranging from 2 to 5cm, and shale pebbles. Evidence of subaerial exposure was minimal thus indicating a dominantly shallow submerged, though possibly tidally influenced depositional environment; possibly deltaic.

Gradually these sandstones of facies association 8 spread northward, burying the siltstones of facies association 6, to form a belt parallelling the shoreline (Fig. 6-2B). Palaeocurrent data from the abundant crossbedding within facies association 8 produces a bipolar distribution with modes parallelling the shoreline (Fig. 6-5) suggestive of longshore current activity.

<u>Intertidal Sandflat</u> (Lithotypes 7A and 7B)

Lying immediately basinward of the facies association 6 skirt to the newly uplifted portion of the basin margin is a cyclic arrangement of pale yellowish grey, planar crossbedded quartzite (lithotype 7A) and pale greenish yellow, rippled silty sandstone (lithotype 7B) (Plate 6-1e). This cyclic facies association forms a north-south trending elongate tongue, ranging in thickness from a feather edge to about 80m (maximum measured thickness being 68m; see section ABC, pages A5 and A8, Appendix I), lying within the grey-green intertidal mudflat deposits of facies association 2 (Fig. 6-2A and Plate 6-1f).

<u>Sedimentary Structures and their Palaeoenvironmental Significance</u>

As with the underlying deposits of facies association 2, the cyclic facies association 7 deposits display a suite of sedimentary structures suggestive of periodic exposure to the atmosphere, the presence of both wave and current activity, and also load-induced hydroplastic deformation.

Flat-topped ripple marks (Plate 6-2a) are the most common structure found within lithotype 7A that is diagnostic of periodic subaerial exposure. One float sample was seen where the flattened crestlines displayed a bumpy, irregular surface not present within the troughs, probably indicative of formation by rain which fell whilst water still lay within the troughs. Elsewhere entire surfaces display the impressions of raindrops. Structures indicative of exposure and high evaporation are salt crystal casts. Salt crystals are cubic in shape and typified by step-wise hollowed-out faces (the hopper effect; see Conybeare and Crook 1968, p.270, Plate 90). Plate 6-2b shows possible sand casts of such crystals.

Rare ripple mark sets are found displaying rounded crestlines and pointed troughs, suggested by Reineck and Singh (1975, p.368, Fig.528), to be due to modification by swash water during the time of subaerial emergence. Falling water levels are further suggested by the presence of ripple mark sets which are composed of a major set, within which are formed a parallel secondary set (cf. Reineck and Singh 1975, p.368, Fig. 527 caption). Symmetric (wavelength averaging 5cm), asymmetric (wavelength averaging 8cm) and interference wave ripples (Plate 6-2c) are common. Asymmetric (catenary) current ripples (Plate 6-2d) are also seen, but rarely.

Occasionally structures are seen resembling symmetric ripple marks, but which grade laterally into small-scale load structures and rarely pseudonodules (Plate 6-2e). Load structures are quite common, both on the small and large scales (Plates 6-2f and 6-2g). Such features always occur where a sand layer directly overlies a silty layer, and thus appear to be the result of hydroplastic displacement induced by overloading (as explained in Chapter 5). Rare thin, elongate structures are seen which resemble organically produced trails (Plate 6-2h). However, despite the similarity between these structures and those typical of late Precambrian surface worm trails (see Glaessner 1969, p.380, Figs. 5b and 5f), the presence of abundant shallow load structures immediately adjacent to them suggests that an inorganic mechanism similar to that which produces the abundant shrinkage cracks (Plate 6-3a) was responsible. This mechanism of overloading, differential compaction by hydroplastic displacement and dewatering of a silty layer by an overlying sandy layer is well shown in plate 6-3b. As this loading and dewatering process continues, the silty layer loses its plasticity within the immediate vicinity of the intruding sand load and begins to tear, producing (in the case of plate 6-3b, radiating) shrinkage cracks.

Tool marks, according to Reineck and Singh (1975, p.69), fall into one of three catagories; namely stationary tool marks, obstacle marks, or moving tool marks. All three types appear to be represented within lithotype 7A. Plate 6-3c shows the basal surface of a slab of lithotype 7A which displays a variety of irregular features possibly representing the marks of stationary tools such as irregularly shaped mud pebbles. Where such an object obstructed current flow, current crescents (Peabody 1947) have resulted (Plate 6-3d). Sengupta (1966, p.367 and Fig. 5a) discusses the current flow pattern and sedimentation mechanism responsible for the development of current crescents. Where current activity was strong enough to move the tool, groove marks (Plate 6-3e) and bounce marks (Plate 6-3f) have been produced. Tool marks, however, are not diagnostic of any specific depositional environment, being as common in intertidal mudflat deposits as in deep water turbidite deposits.

Parting lineation (Plate 6-3g), graded bedding and abundant planar (omikron) crossbedding (with average set thickness and foreset slope being 13cm and 23^{o}, respectively) are other sedimentary structures indicative of current activity. Occasionally herringbone crossbedding (Plate 6-4a) and reactivation surfaces are seen which, as stated earlier, suggest tidal conditions of deposition. Often the crossbedding is a mixture of omikron and pi cross-stratification resulting in complex outcrop patterns. Although generally unifrom in grain size throughout a crossbed set, toward the base of the foreset slope can be found accumulations of either clay and silt, or coarse to very coarse sand. Trains of coarse to very coarse sand are also commonly seen in plane-bedded layers, and may indicate a windblown origin. Rare recumbent (Plate 6-4b) and deformed crossbed foresets are recorded, indicating current drag over a saturated megaripple. Large-scale scour-and-fill structures, often multi-generation and possibly representing tidal channels (Plate 6-4c), are not uncommon.

The suite of sedimentary structures contained within lithotype 7A suggest that the environment of deposition was dominated by current activity under shallow conditions, but periodically subjected to subaerial exposure; i.e. an intertidal environment.

The pale greenish yellow silty sandstones of lithotype 7B invariably display a sequence of sedimentary structures which vary with grain size and current strength. At the base of the sequence is a plane-bedded to massive medium sandstone which invariably lies upon a shaly siltstone of the underlying sequence with a flat, to shallowly undulating contact. This contact commonly reveals shrinkage cracks and occasional small load structures. This basal layer is overlain by a small-scale planar crossbedded medium sandstone, which is often capped by a thin, plane-bedded sandstone, possibly representing the crossbedding topset laminae. Above this, ripple cross-lamination of silty sandstone occurs, over which lies a shale pebble-bearing shaly siltstone. Invariably the sequence is incomplete, but where complete it represents a decrease in current strength through the entire lower flow regime, possibly in response to waning tidal currents. These lithotype 7B sandstones and siltstones are the coarse-grained equivalents of facies association 2 which lies adjacent to the facies association 7 sand sheet within the centre of the basin. As with facies association 2, the greenish hue of the sediments is believed to be in response to iron reduction by microscopic organic activity.

Palaeocurrents of the Intertidal Sandflat

Directional measurements from planar crossbedding formed the majority of the palaeocurrent date of lithotype 7A, totalling 87% of all readings, with parting lineation completing the analysis. A bimodal distribution dominates most of the individual analyses with the dominant mode having a southeasterly trend, averaging 160°, and the secondary mode trending westward toward 265° (see Fig. 6-5). Occasionally a trimodal distribution is present, the third mode trending toward the north-north-east.

The two dominant modes roughly parallel those present in the underlying and surrounding mudflat deposits of facies association 2, which were laid down by tidal currents in an intertidal environment. This consideration, in conjunction with the environmental significance of the sedimentary structures contained within facies association 7, suggest a tidal ebb- and flood-current origin for the palaeocurrent distribution. Along the eastern edge of the sand sheet palaeocurrent readings were obtained from two localities (viz. DDR and SNH) which reveal a bimodal current system parallelling the edge of the sand sheet (Fig. 6-4) indicative of a 'longshore' current system.

PALAEOGEOGRAPHY - TIME INTERVAL IV

The sandflat phase of upper Brachina Subgroup deposition was generated when the tectonic instability, earlier active only on the Gawler Craton, extended across the craton and uplifted a portion of the basin margin causing a new shoreline to develop (Uplift I). Although this uplift occurred within the northern part of the region previously subjected to fluvio-deltaic deposition, the position of the main river outlet appears to have remained relatively stable, debouching the coarse detritus of facies association 8 directly above the thickest accumulation of deltaic facies association 5 detritus in the vicinity of Alligator Gorge. Because of the new shoreline position, the sands forming facies association 8 were forced out into the basin (see Fig. 6-6) in contrast to those of facies association 5 which were reworked alongshore.

This tectonic instability appears to have mildly affected the central portion of the central region within the basin - along the region outlined as a tectonic high on the isopach map of the upper Brachina Subgroup (Fig. 5-10). Thin, isolated lenses of coarse sandstone of a different provenance are found within facies association 2 at localities close to the present day Blinman complex (e.g. locality AMC) which, it is suggested, rose to form a shallow island from which these coarse sands were shed at approximately the same stratigraphic level as the initial generation of facies association 7 in the southern portion of the central region (i.e. about 70m below the boundary between the upper Brachina Subgroup and the Bunyeroo Formation; see Figs. 6-2A and 6).

The sands eroded from the previously deposited deltaic plain (facies association 5) were dispersed within the basin by tidal and longshore current activity into a vast intertidal sand tongue (facies association 7; Fig. 6-6). The cyclic arrangement of lithotypes 7A and 7B within this tongue ranges from 5 cycles in the centre (Plate 6-1e) to one prominant lithotype 7A tongue within facies association 2 along its edges. As with the cyclic arrangement of facies association 5 discussed earlier, the lithologic boundary between lithotypes 7A and 7B is discordant to the bedding, suggesting an origin by the migration of megarippled sand shoals under relatively high energy conditions, where well sorted sediment accumulated (lithotype 7A), separated from one another by regions of slightly higher elevation and lower energy conditions where less well sorted sediment accumulated (lithotype 7B). In these latter regions microscopic organic activity was prolific causing the reduction of the sediment and the green colouration. Again as with facies association 5, estimates of the palaeo-tidal range could be made from the cycle thicknesses, but the same arguments apply as outlined earlier which render such results extremely speculative.

Initially separating this intertidal sandflat from the newly uplifted shoreline was a zone of (?)high intertidal mud deposition (facies association 6; see Fig. 6-6). The coarse sand eroded off the craton and debouched into the basin through the "Alligator River" to the south of this new shoreline zone (facies association 8) formed a lobate distribution pattern lying perpendicular to the depositional strike similar to Coleman and Wright's (1975, p.123, Fig.10) type IV sand distribution.

With the cessation of Uplift I, the new shoreline gradually subsided and retreated. Although the intertidal sandflat further within the basin remained within the intertidal zone, the narrow (?)high inter-tidal mudflat belt separating it from the now retreating shoreline sank to possibly just below low tide level, allowing the coarse sands forming facies association 8 to migrate northward by longshore current activity to form a shallow submerged to possibly intertidal coastal sandflat (possibly a beach; Fig. 6-9).

Toward the top of facies association 7 in the northern part of subregion CI lenticular interbeds of coarse sandstone (Plate 6-4d) are found in the vicinity of the Blinman 'diapir' complex (e.g. Parachilna

Gorge (PCG) and Third Plain (TDP); see Fig. 6-2B). These coarse sandstones herald the second episode of tectonic instability to effect the basin (Uplift II) which caused renewed coastal erosion and deposition along the western flank of the basin, and island emergence within the basin itself.

SEDIMENTOLOGY - TIME INTERVALS V AND VI

This second episode of tectonic instability (Uplift II) displays an asymmetry across the basin. Activity upon the western margin involved a continuous episode of uplift and erosion, whilst within the basin it is seemingly recorded as two separate pulses (see Fig. 6-1).

The first of these pulses is represented by the microconglomeratic, coarse to very coarse sandstones and dusky red purple siltstones of lithotype 9A which are found in the centre of the central region to the west and northeast of the area today occupied by the Blinman-Enorama-Oraparinna 'diapir' complexes (Fig. 6-3A). Crossbedding, dominantly trough but with some planar, and abundant shale pebbles, displaying some imbrication, are the only sedimentary structures seen within the microconglomeratic sandstones. Palaeocurrent data from the crossbedding was only obtainable from one locality (South of Aroona "Ruins" (SAR)) where a bimodal distribution was found, having modes trending toward the south (186°) and west (262°) (see Fig. 6-5B). The dusky red purple siltstones associated with these microconglomeratic sandstones display a planar lamination or a highly irregular, undulatory surface (Plate 6-4e). Isolated trains of coarse sand are common throughout, possibly of windblown origin.

The restricted distribution of this lithotype, along with its coarseness and the abundant pebbles of reworked shale lithologies (Plate 6-4f), indicate that the source area lay close by. The diversity of rock types forming the clasts of the microconglomerates, ranging from green shales and siltstones, brown sandstones, pink quartzites, grey cherts and silicified (?)oolitic limestones, to volcanigenic clasts and dolerites (Plate 6-4f and 4g), coupled with the palaeocurrent data, suggest that the region now occupied by the Blinman-Enorama-Oraparinna 'diapir' complexes (which contain such a diversity of exotic rock types - see Mount 1975) was exposed above sea level as a shallow island (or series of islands).

Basinward of this region, especially toward the north and south-east, overlying the deposits of facies association 2, is a series of dark greyish red shales - the basal Bunyeroo Formation. Very little internal structure is discernable from these shales because of their very uniform grain size and a pervasive cleavage which splits them into 'pencil' shapes. The red-purple colouration of these shales indicate an oxidizing environment of deposition in contrast to the reducing conditions responsible for facies association 2, which was induced by a high activity of microscopic organisms. The onset of Uplift II, therefore, appears to have produced an oxidizing environment within which microscopic organisms could not thrive.

During the period of tectonic 'quiescence' separating the two pulses of Uplift II within the basin, the red-purple Bunyeroo shales steadily advanced toward the western margin (see Figs. 6-3B and 4A). A minor amount of sand was eroded from the area today occupied by the Oraparinna 'diapir' complex, producing the sandy subfacies of lithotype 9B. These sands range from blackish red purple irregularly laminated, medium to coarse sandstones, through thick laminae to thin beds of medium to coarse sandstone interbedded with dark greyish red siltstones (Plate 6-5a), to thin beds of medium sandstone, greyish red purple at the base, but grading upward to very light grey, where a ripple cross-lamination is often preserved, and overlain by a thin dark greyish red shale. A crude sand-shale cyclicity is suggested which, by analogy with facies associations 5 and 7, is indicative of an origin by the migration of sandwave shoals, separated by shale accumulation on regions of slightly higher elevation, under tidal to intertidal conditions. Palaeocurrent data obtained from these cyclic sandstones reveal a distribution which parallels the earlier tidal current systems of facies associations 2 and 7, with modes trending toward the east-southeast and west-northwest (Fig. 6-7A). Almost parallel modes were obtained from an occurrence of asymmetric (catenary) ripple marks developed in shale (Plate 6-5b). Elsewhere within the central region dark greyish red shales typical of the Bunyeroo Formation were deposited.

The second pulse of Uplift II within the basin caused a large amount of coarse to very coarse and pebbly sand to flood out across the vast lithotype 9B-Bunyeroo mudflat. Within these sandstones (which constitute lithotype 9C) crossbedding, graded bedding (Plate 6-5c) and shale pebbles are ubiquitous, whilst scour structures, parallel lamination (Plate 6-5d) and thicknesses of structureless sandstone are common. Well defined alpha crossbeds are rare, the dominant style being a scour-and-fill pi cross-stratification, from which very few accurate palaeocurrent readings were obtained because of the nature of the outcrop. The readings that were obtained produce a complex pattern, with modes toward the northeast, southeast, southwest and northwest all being present (Fig. 6-7B).

Where these sandstones have flowed out across the remnant intertidal sandflat of facies association 7, the contact is locally erosive (Plate 6-5e). The presence of cobble-sized fragments of lithotype 7A within the basal few centimetres of the overlying lithotype 9C (see Plate 6-5e) suggest that a certain degree of lithification of facies association 7 had occurred prior to the deposition of lithotype 9C. As with the microconglomeratic sandstones of lithotype 9A, these lithotype 9C deposits appear to have been derived from areas within the basin, their distribution and thickness indicating the Blinman-Enorama-Oraparinna 'diapir' complex area as one likely source area. A second source, which shed locally microconglomeratic material at this time, is postulated to have emerged in the region of the Worumba 'diapir' (see Fig. 6-4A).

The palaeoenvironmental significance of facies association 9 is difficult to ascertain solely from the contained sedimentary structures. However, when considered in the light of the overall conditions of sedimentation throughout the basin (i.e. regressive) and the associated lithotypes (i.e. dominantly deposited within intertidal environments), along with the distribution and thickness data for each of the component lithotypes (see Figs. 6-3A, 3B, 4A and 8), a palaeogeographic picture can be constructed.

A comparison between the microconglomerates of lithotype 9A and present day creek-rock deposited within the Arkaba 'diapir' complex of the central Flinders Ranges (Plate 6-5f) reveals such a remarkable similarity between grain size distribution and imbricated clast arrangement that similar fluvial conditions of deposition are envisaged for the two rock types. The presence of both pi and alpha crossbedding, and graded bedding within such poorly sorted sands supports such an interpretation of a fluvial environment of deposition. Similar deposits have been described by McGowen and Groat (1971), Campbell (1976) and Miall (1977).

The distribution and palaeocurrent data for facies association 9 indicate the source region lay close by and within the basin, whilst thickness data and the presence of such included exotic rock fragments as dolerites indicate that the areas today occupied by the Blinman-Enorama-Oraparinna and Worumba 'diapir' complexes acted as source areas.

Miall (1977, pp.47-48) describes a Platte-type braided river deposit as being one that is typified by trough and planar crossbedded sand facies, with interbeds of horizontally bedded, rippled and scoured sands (representing the deposits of shallow channels), gravels (channel lag deposits) and laminated muds and silts (overbank deposits). Such deposits are "thought to represent very shallow rivers or those without marked topographic differentiation". Lithotype 9A, therefore, would appear to have been deposited within such a system, with the microconglomeratic sandstones representing the shallow channel deposits, and the laminated to undulatory dusky red purple siltstones the overbank deposits.

McGowen and Groat (1971) divide the braided and alluvial Van Horn Sandstone (late Precambrian to Ordovician) of West Texas into three palaeoenvironmental zones, namely proximal, mid and distal alluvial fans. As seen above, the deposits of lithotype 9A equate to the proximal to mid fan braided channel and overbank environments, whilst the coarse sandstones of lithotype 9C would equate to the mid fan environment where dominantly trough crossbedded (dominated by fluvial braided stream conditions; e.g. at locality ABC), and to the distal fan environment where dominantly planar crossbedded (dominated by deposition from the fluvial system as it enters a shallow sea; e.g. at locality IKC - see McGowan and Groat 1971, p.38, Fig.30).

Meanwhile, along the western margin of the basin, Uplift II caused the shoreline to advance into the basin. With this advance the westernmost deposits of facies associations 5, 6 and 8 were subjected to erosion and were reworked along with newly debouched material into a skirt which preceded the shoreline advance. These reworked sands constitute lithotype 10A. Their almost monomineralogic nature invariably imparts a very massive outcrop appearance within which very little internal structure can be seen. Large-scale planar (alpha) crossbedding was encountered at one locality (East of "Buckaringa" Homestead (EBH); Plate 6-5g) from which a unimodal palaeocurrent distribution was obtained. This palaeocurrent direction trends toward the northeast (48°; Fig. 6-7B), parallelling the shoreline and hence suggestive of a longshore current system. Some ripple cross-lamination and small-scale scour-and-fill structures were the only other sedimentary structures encountered.

To the south of the advancing shoreline it appears as if the coarse sands forming facies association 8 were still being debouched into the basin at this time.

PALAEOGEOGRAPHY - TIME INTERVALS V AND VI

Within the advent of renewed tectonic activity within the basin (Uplift II), the emergence of islands and the advance of the western shoreline into the basin produced local variations in the environmental conditions of deposition, although the major palaeogeographic configuration remained similar to that of time interval IV - i.e. a vast intertidal sandflat (facies association 7) surrounded by an intertidal mudflat (facies association 2) with a (?)delta (facies association 8) formed in the south-west (Fig. 6-6). With the emergence of islands within the intertidal sandflat, microconglomeratic material and coarse sands were shed and deposited within braided fans - lithotype 9A representing the proximal to mid fluvial-dominated fans (Fig. 6-9), whilst lithotype 9C represents the deposits of mid fluvial-dominated, to distal marine-dominated fans (Fig. 6-10).

Basinward of these islands and their braided alluvial deposits the tectonic 'relaxation' separating Uplift II Pulse A from Uplift II Pulse B caused a general subsidence which persisted from that time onwards (except within the zone of the islands during Pulse B). The subsequent change in environmental conditons of deposition resulted in the apparent extermination of all microscopic organisms (prolific during the deposition of facies association 2 and responsible for the green colouration of those deposits), hence allowing free oxidation of the muds deposited from suspension within a vast mudflat. Locally within this vast mudflat (the basal Bunyeroo Formation) small tidal to intertidal sandflats trapped any sand eroded from the remnant islands within the basin, producing lithotype 9B.

The other local variation in depositional environment occurred as the basin margin was uplifted during Uplift II. In front of the advancing shoreline a northerly directed longshore current system produced a coastal skirt (possibly a beach deposit) of supermature coarse sandstone (lithotype 10A) which encroached upon the previously deposited sediments with continued shoreline advance into the basin (Figs. 6-9 and 10).

PALAEOGEOGRAPHIC EVOLUTION THROUGH THE SANDFLAT PHASE OF THE UPPER BRACHINA SUBGROUP

Tectonic instability upon the Gawler Craton produced a vast amount of detritus which made its way into the flanking, steadily subsiding basin through, what must have been, an extensive river system (the "Alligator River") to form the upper Brachina Subgroup. Where this river debouched its load into the basin a delta eventually formed (facies association 5) which was surrounded by a vast, organic rich intertidal mudflat (facies association 2). This same tectonic instability appears to have gradually extended itself across the craton, and episodically entered the basin to partially destroy this deltaic system and create the sandflat phase of upper Brachina Subgroup sedimentation.

As this tectonic instability initially neared the basin, the detritus being debouched through the river system steadily became coarser grained, and gradually facies association 5 was replaced by facies association 8. When the first episode of tectonism reached the basin (Uplift I) it caused partial erosion of the northern reaches of the deltaic system. These eroded deposits were reworked into an intertidal sandflat tongue (facies association 7) which spread northward across the organic rich intertidal mudflat (facies association 2). Initially separating this intertidal sandflat from the newly uplifted shoreline was a (?)high intertidal mudflat belt (facies association 6; see Fig. 6-6). The possibility that this instability caused small islands to appear within the intertidal mudflat is suggested by the presence of isolated lenses of coarse sandstone bearing exotic rock fragments typically found within the many 'diapiric' structures of the Flinders Ranges, which are known to have been unstable during Adelaidean times. As this uplift subsided and the basin margin retreated, the coarse sands forming facies association 8 migrated northward as a coastal sandflat generated from the (?)deltaic lobe to the south.

The second episode of tectonic activity (Uplift II) renewed uplift and erosion along the basin's western margin, and also reworked the deposits of facies associations 6, 8 and again the earlier deltaic deposits of facies association 5 into a coastal sandflat (lithotype 10A) which skirted the basinward migrating shoreline. This uplift also caused the emergence of (at least) two islands within the intertidal sandflat in the vicinity of the present day Blinman-Enorama-Oraparinna and Worumba 'diapir' complexes, apparently in two separate pulses. Pulse A is recorded by the fluvial proximal fan, to mid fan braided channel and overbank deposits of lithotype 9A around the Blinman 'diapir' complex (Fig. 6-9), whilst pulse B is recorded by the fluvial mid fan braided channel and marine-dominated distal fan deposits of lithotype 9C around both the Blinman and Worumba 'diapir' complexes (Fig. 6-10). Separating these two pulses was a period of minor subsidence which had a pronounced effect upon the entire depositional system and marked 'the beginning of the end' of the upper Brachina Subgroup. During this period of subsidence conditions favourable to abundant microscopic organic activity upon the surrounding intertidal mudflats ceased to prevail, and facies association 2 was replaced by the basal dark greyish red shales of the Bunyeroo Formation, deposited within a tidally influenced (and in places intertidally influenced) mudflat environment. These basal Bunyeroo shales intertongue with the intertidally deposited sandstones eroded from the remnant islands (lithotype 9B). With the cessation of Uplift II deposition of the Bunyeroo shales occurred throughout.

CHAPTER 7

POST-UPPER BRACHINA SUBGROUP PALAEOGEOGRAPHY

PALAEOGEOGRAPHY - TIME INTERVAL VII

As the effects of Uplift II subsided and all source areas of coarse detritus were submerged, the dark greyish red shales of the Bunyeroo Formation covered both lithotypes 9C and 10A. Within the centre of the basin thin isolated lenses of coarse sandstone petrologically similar to that of facies association 9 are found within the basal 5m of these shales (Plate 7-4c) indicating conformity between the two units. The apparent sharpness of the boundary, however, suggests that the remnant islands within the basin were of low topography and readily flooded during the basinwide subsidence and transgression which followed Uplift II.

Also found, but only rarely, within the basal Bunyeroo shales in subregion CI (e.g. at Mernmerna Creek (MMC)) are siderite concretions. According to Blatt et al. (1972, p.400) siderite is "almost exclusively authigenic" in origin. Krauskopf (1967, p.83) explains that siderite precipitates when either ferrous iron rich, and carbonate or bicarbonate rich solutions mix, or a ferrous iron and bicarbonate rich solution becomes alkaline. However, because much of the free iron released by weathering processes is in the ferric form and the "amount of ferrous iron in most surface waters is vanishingly small ... unless unusual conditions are encountered where either the supply of ferrous iron is large or a reducing environment is maintained by abundant organic matter", siderite is rarely found as a sediment.

Nevertheless, siderite concretions are common in a variety of sedimentary rocks, and, as Shannon (1977) reports for similar concretions in shallow marine rocks of Ireland, the siderite concretions found within the Bunyeroo shales are believed to have originated in a manner similar to that of green reduction spots (see van Houton 1961, 1968), which are also found in these Bunyeroo shales (e.g. North of Bunyeroo Gorge (NBG)).

The mechanism involves an early diagenetic reduction of ferric iron to mobile ferrous forms which then react with carbonate rich waters under conditions of negative Eh and low sulphide activity beneath the sediment-water interface (Curtis and Spears 1968, p.261).

In subregion WII the dark greyish red Bunyeroo shales sharply overlie lithotype 10A. However, along the western edge of this subregion the final episode of tectonic unrest occurred (Uplift III; see Fig. 6-1). The erosion and reworking of facies associations 5, 6, 8 and lithotype 10A produced a lag deposit of granular, coarse to very coarse quartzite (lithotype 10B; Plate 7-1a) which was deposited in front of an advancing shoreline as a coastal sandflat in a manner analogous to lithotype 10A (Figs. 7-1, 3B and 3D). To a greater degree than with lithotype 10A, the very boldly outcropping nature of this lithotype hides from view all trace of any internal structure that may be present. Commonly seen, however, are lenticular pebble conglomerates comparable with beach rock deposits (*see* Conybeare and Crook 1968, p.90, Plate 5). These indicate that lithotype 10B possibly represents a beach deposit analogous to lithotype 10A. Elsewhere at this time the dark greyish red shales of the Bunyeroo Formation were being laid down in a shallow marine mudflat environment (Fig. 7-2).

With the cessation of Uplift III and the final retreat of the shoreline by subsidence of the western edge of the basin, the dark greyish red Bunyeroo shales gradually covered the entire basin. Initially silty in the western region, these deposits show some evidence of traction current activity with the development of flute marks (Plate 7-1b) and small groove marks (Plate 7-1c). Thomson (1969, p.76) reports of "sun cracks" being recorded from this formation and concludes that conditions of "shallow water to subaerial exposure and mild tectonism" were responsible for its deposition. These shales are often typified by a pencil-shaped appearance in outcrop, caused by the intersection of bedding and cleavage planes, with the length often determined by joint plane spacing. At Hanniman Gap (HMG) 'pencils' reaching 75cm in length were found (Plate 7-1d), although the average length is in the order of 10cm.

Found at locality BKG immediately above lithotype 10B, but only in float, is an abundance of dark greyish purple granular coarse sandstone. Whether this actually belongs to lithotype 10B or represents a channel lag deposit within the Bunyeroo Formation is impossible to ascertain because of the complete lack of outcrop. Colouration suggests the latter of the two possibilities, whilst texture supports the former.

THE UPPER BRACHINA SUBGROUP-BUNYEROO FORMATION BOUNDARY

As seen above, throughout most of the study area the upper Brachina Subgroup is overlain by a thick, dominantly dark greyish red shale and siltstone sequence defined as the Bunyeroo Formation. Locally, however, deep scour structures (possible submarine canyons) cut through the Bunyeroo Formation and into the upper Brachina Subgroup. These are filled with limestones of the Wonoka Formation which lies stratigraphically above the Bunyeroo Formation. The boundary between the upper Brachina Subgroup and the Bunyeroo Formation is very complex, displaying both conformable and locally disconformable contacts. All contacts, however, are accompanied by a change in outcrop expression from the bold ridge and hilly topography of the upper Brachina Subgroup to the subdued valley topography of the overlying units.

A marked disconformable contact is present in the western portion of subregion WII (Fig. 7-3A). Here the dark greyish red shales and siltstones of the Bunyeroo Formation rest upon facies association 5 of the ABC Range Quartzite. The actual disconformity surface is only rarely seen, but exposures at two localities (viz. Petanna Gorge (PTG) and BKG) suggest a complex sequence of erosion and deposition.

One locality (see section PTG - Appendix I) displays a lenticular development of deeply weathered conglomeratic material outcropping above a shale band 1.5m above the uppermost bed of massive quartzite (lithotype 5A) and immediately below a thick development of greyish red shale and siltstone of the Bunyeroo Formation (Fig. 7-3B). The conglomerate clasts consist predominantly of shale and quartzite and reach up to 10cm in size (Plate 7-1e and 1f). These clasts appear to be of similar lithology to the sediments immediately below. Pebbles of green shale and red porphyrytic material were also noted, but in small amounts (Plate 7-1f). The unit has a thickness of 1.5m within the creek and extends up the creek bank parallel to the dip of the ABC Range Quartzite for about 3m. It is considered therefore, that this conglomerate belongs to the upper Brachina Subgroup-Bunyeroo Formation succession, and is not merely present day creek-rock.

According to definition, and on lithologic criteria, the boundary between the ABC Range Quartzite and the Bunyeroo Formation at this locality would be placed at the top of the uppermost development of quartzite. The conglomerate, therefore, is wholly contained within the Bunyeroo Formation 1.5m above the base. Because of the poor outcrop and deep weathering of this unit, however, its exact significance is difficult to ascertain from this one locality alone.

Outcrop of the disconformity between the ABC Range Quartzite and Bunyeroo Formation also occurs immediately to the south at locality BKG (Fig. 7-3C). The complexity of this boundary is only realised when viewed in relation to the complete sequence of events over the entire region of study (see Chapter 6 and earlier this chapter). The sands which constitute lithotype 10A were generated by the gradual uplift and erosion of the basin margin immediately west of this locality (Fig. 7-4B). This uplift and erosion (Uplift II) continued and steadily encroached upon the basin, pushing lithotype 10A before it in a continually reworked sand skirt. The stabilization, and then gradual subsidence of the basin margin then enabled the first Bunyeroo shales to be deposited above the lithotype 10A sands and on top of the eroded surface exposure of facies association 5 (Fig. 7-4C). A following phase of uplift and erosion (Uplift III) to the immediate west produced a second sand skirt - lithotype 10B - which prograded out onto the thin Bunyeroo shale deposit (Fig. 7-4D). Thin red purple shaly partings in the basal 2m of lithotype 10B indicate conformity between these sands and the underlying shales. As with lithotype 10A, these sands were continually being reworked by coastal erosion and nearshore processes along the leading edge of the encroaching uplifted margin. The overlap of lithotype 10B past the point of cutback from Uplift II suggests the possible presence of two disconformable surfaces, as shown in figure 7-3C.

The Bunyeroo Formation lying immediately above the lower disconformity and between lithotypes 10A and 10B is a shale lithology only, whereas above both the upper disconformity and lithotype 10B the Bunyeroo Formation comprises both shales and siltstones (Fig. 7-3C). The thin conglomeratic lens at section PTG, separating a thin shale band from a thick shale and siltstone sequence, therefore, possibly represents the upper disconformity. The lower disconformity would therefore be placed at the boundary between the uppermost quartzite of facies association 5 and the thin shale band lying below the conglomerate.

In all other localities where both lithotypes 10A and 10B are present and exposed (i.e. MKT, Richman Gap (RMG) and Waukarie Creek (WKC);

see Fig. 7-3A) the thickness of the Bunyeroo shale between the two lithotypes decreases toward the region of uplift, as would be expected in the above interpretation (Fig. 7-4).

It must be stressed that the exposure of this boundary is not good enough to fully ascertain its exact significance. The above interpretation was deduced not only from the limited amount of first hand information available, but also from the regional palaeogeographic interpretations (see Chapter 6 and earlier this chapter). The possibility that the shale band below the conglomerate at section PTG is correlative to lithotype 5B, with the conglomerate representing either a local channel feature or the lower disconformity, is discounted because lithotype 5B contains a significant amount of siltstone and fine to medium sandstone which is totally absent from the shale band. A second possible alternative interpretation sees the conglomerate as a local channel structure within the Bunyeroo Formation totally unrelated to any disconformity(ies) - the only disconformity present being between the uppermost quartzite and the shale band. Field evidence is not sufficient to prove or disprove this possibility, but the regional palaeogeographic interpretations favour the conglomerate as lying upon a second disconformity.

Immediately surrounding this region of disconformity throughout the remainder of subregion WII, and also throughout subregion WI and the southern portion of the central region, is a conformable intertonguing contact between facies association 10 and the Bunyeroo Formation. Where the contact is exposed lithotype 10A is overlain sharply, though with apparent conformity, by the Bunyeroo Formation. However, in subregion WI and the southeastern portion of subregion WII, resting conformably above between 2 and 30m of Bunyeroo shales are the granular quartzites of lithotype 10B. The upper contact between this lithotype and the overlying

Bunyeroo Formation is invariably covered, but by thickness considerations appears to be conformable.

Throughout most of subregion CI, and extending into the eastern region, the upper Brachina Subgroup is capped by the coarse sandstones and shales of facies association 9. The relationship between this facies association and the deposits which immediately underlie and overlie it is one of extreme complexity.

In general within the central region facies association 9 lies directly upon the quartzites and sandstones of facies association 7 with a locally erosive contact (Plate 6-5e). In some localities lenticular developments of coarse sandstone petrologically similar to those of facies association 9 are found interbedded within lithotype 7A (e.g. PCG and TDP; see Plate 6-4d). On this basis facies association 9 belongs genetically to the ABC Range Quartzite. However, lithotype 9B - an interbedded very light grey to greyish red purple sandstone and dark greyish red shale sequence - grades laterally northward into a dark greyish red shale which is lithologically identical to the basal shales of the overlying Bunyeroo Formation. Where lithotype 9C thins and eventually wedges out these lithotype 9B shales cannot be distinguished from the basal portion of that formation. Lithotype 9B, therefore, appears, at least in part, to be an intertongue of the Bunyeroo Formation within facies association 9. Everywhere contacts are seen within the central region between coarse sandstones and greyish red shales they are sharp, yet conformable. At locality PCG the lithotype 9B (shale)-9C (coarse sandstone) contact displays a load casted surface (Plate 7-1g) indicating sand deposition upon a saturated clay layer, whilst at localities CCR, MMC and South of Bunyeroo Gorge (SBY2) thin lenticular developments of coarse sandstone are found within the basal 2m of the overlying Bunyeroo Formation (Plate 7-1h). The boundary between the upper Brachina Subgroup and the Bunyeroo Formation in the central region, therefore, is intertonguing and conformable.

Locally within the eastern region the coarse sandstones of lithotype 9C overlie a thin development of lithotype 7A (e.g. Eurilpa Gap section (ELG); see Appendix I, Page A6), whilst at other localities a thin development of lithotypes 9B and 9C directly overlies facies association 2 (e.g. The Dome (TDM)). In each case lithotype 9C is sharply, though apparently conformably overlain by the Bunyeroo Formation. Elsewhere, lenticular developments of lithotype 9C occur well within the Bunyeroo Formation (e.g. section ODW, see Fig. 3-2; and Dawson (DWS)). At one locality (section MCH) the coarse sandstones of lithotype 9C lie within dark greyish red shales and siltstones in highly disrupted lenticular developments above a local disconformity between facies association 2 and the Bunyeroo Formation (see Fig. 3-3).

Where facies association 9 is absent in the eastern region the upper boundary of the upper Brachina Subgroup is marked by a rapid decrease in the presence of grey-green siltstone and pale greenish yellow lenticular sandstone coinciding with a rapid colour change of the shales, over a 5m interval, from the greyish green of facies association 2 to the dark greyish red of the Bunyeroo Formation.

It appears, therefore, that within the southeastern portion of the eastern region and the northern portion of subregion CI deposition of the Bunyeroo Formation had started prior to the generation of facies association 9. In the southern and central portions of the central region, and extending down into the northern portion of the eastern region, however, facies association 9 was deposited prior to the shales of the Bunyeroo Formation. The temporal significance of the upper Brachina Subgroup-Bunyeroo Formation boundary in these areas, therefore, is that the Bunyeroo Formation represents a transgression which began to the southeast and north of the study area and gradually encroached upon the central and western regions. The boundary is, therefore, diachronous. Meanwhile, within subregions WI and WII, and the southern portion of the central region a disconformable to intertonguing conformable boundary is present between facies associations 5 and 10 of the upper Brachina Subgroup and the Bunyeroo Formation. Temporally this boundary represents local reworking of the ABC Range Quartzite by basin margin uplift and their redeposition upon the adjacent Bunyeroo mudflat.

CHAPTER 8

PETROLOGY OF THE UPPER BRACHINA SUBGROUP

The upper Brachina Subgroup comprises two formations dominated by shale and siltstone (the Moorillah and Bayley Range Formations) and a laterally equivalent, to overlying formation dominated by sandstone and quartzite (the ABC Range Quartzite). The petrologic study of these formations was restricted to the sandy lithologies because of the obvious grain size advantage and greater provenance information revealed by arenites over argillites, as outlined in Chapter 2.

The shale-siltstone formations are composed predominantly of clay minerals, mica and a quartz-rich siliceous fraction. Lithotype 1A of the Moorillah Formation contains muscovite mica and a haematitic matrix, whilst chlorite and a clay-rich matrix are present within lithotype 1B of the Moorillah Formation and also the Bayley Range Formation. Sandstones form only a minor component of these formations and they are discussed below with the sandstone-quartzite lithologies of the ABC Range Quartzite.

VOLCANIGENIC LITHOLOGIES WITHIN THE MOORILLAH FORMATION

The Moorillah Formation equates to Mawson's (1938, p.261) unit 2 of his "Tuffaceous Series". A great variety of tuffaceous sediments are present, ranging from massive (generally dusky red), crossbedded (usually banded moderate red and pale yellowish grey) or current-bedded (dusky red and greyish purple) tuffaceous sandstones, (Plates 5-2a and 8-1a), to dusky purple massive, crossbedded or intraformational conglomeratic tuffaceous siltstones and sandstones (Plates 5-1h and 8-1b), and greyish purple soft-sediment deformed tuffaceous siltstones (Plates 8-1c). Pure tuff bands, up to 5cm thick, are occasionally seen, and these are usually associated with the intraformational conglomeratic tuffaceous siltstones, the intraclasts being chips and flakes of tuff (see Plate 8-1b).

Beds of these tuffaceous sediments occur scattered throughout the Moorillah Formation (lithotype 1A) in the southwestern portion of subregion CI (e.g. Red Range (RDR)). It is here, also, that the sandy tuffaceous sediments occur. Commonly found within these coarser tuffaceous deposits are diagenetic chert nodules (Plate 8-1d). In the centre of subregion CI only tuffaceous siltstones occur (predominantly in the basal half of lithotype 1A - e.g. sections ABC and ARR), whilst in the northeast (e.g. locality CBC) tuffaceous beds appear to be absent. Within lithotype 1B in the eastern region, and lithotypes 3A and 3B of subregion WII, tuffaceous material is present in very small amounts within the siltstone and sandstone lithologies. The origin of such tuffaceous sediment is by ash fall into an active sedimentary environment, and the grain size and areal distributions described above suggest a source region to the west.

The tuffs and tuffaceous sediments are all devitrified and highly ferruginized, and hence would be of little value for radiometric dating (except perhaps using zircons - but this aspect has not been pursued in this study). The grains of volcanic origin take one of two general forms, either highly angular, whispy flakes (shards), or either irregular, or ovoid to 'dumb-bell'-shaped spherulite-like grains. The shards have been altered either to muscovite or chlorite and are preserved within mud-sized sediment. In silty beds they lack prominant orientation or distortion (Plate 8-1e), but they display a considerable degree of flattening and alignment in shales which are interlaminated with fine sand (Plate 8-1f). The spherulite-like grains, on the other hand, (the shape of which is believed to have originated in a similar manner to that of tektites) are found in sediment of all size grades up to medium sand, but are often larger (by up to 3 times) than the containing sediment (see Plate 8-1f). A variety of alteration products have preserved these spherulite-like fragments within a variety of lithologies (Plates 8-1g and

1h). The apparent sequence of alteration is presented in figure 8-1. The original alteration product appears to be chlorite, which can then be replaced, to varying degrees, by either carbonate or silica (as either chert or quartz), or mixtures of both. Rarely prehnite(?) and the zeolite laumonite are seen. In most cases a haematite rim, either complete or 'blotchy', develops around each 'spherulite' during the chlorite stage of alteration, and tends to be preserved throughout any subsequent alteration phases (see Plates 8-1f, 1g and 1h). The alteration products, viz. chlorite, carbonate, muscovite and the iron-rich matrix, of these volcanigenic sediments are typical of basic volcanics. A few measurements of the extinction angle on included subangular plagioclase feldspar grains indicates an andesine-labradorite composition, also typical of basic volcanics.

As stated above, the grain size and areal distribution of these tuffaceous sediments suggest a source region to the west of the basin of deposition. Recent radiometric dating of core material upon the Stuart Shelf has revealed a 697±70 m.y. old basic volcanic deposit, newly defined as the Beda Volcanics (see Mason et al. 1978, Webb and Hörr 1978). These spilitic flows are separated from the Roopena Volcanics (1340 m.y., Thomson 1966) by the Pandurra Formation and Backy Point Beds, and are correlated with the Wooltana Volcanics and Callanna Beds of the northern Flinders Ranges. Despite the lack of direct correlation of the Brachina Subgroup with known volcanic deposits, the persistence of basic volcanism centred around the Cultana Inlier (see Mason et al. 1978, p.8, Fig.3) for such a lengthy period of time, tends to suggest the same source as being responsible for the volcanigenic deposits present within the Moorillah Formation.

'Spherulitic' structures are also found within the sandy deposits of lithotypes 3A (carbonate is also found in these sands; although no obvious shape is present, the very high energy of origin of this lithotype suggests that this carbonate was derived as an alteration produce of basic volcanic detritus), basal and uppermost 5A, 6, and basal and uppermost 7A of the ABC Range Quartzite. The structures are defined by an ovoid, cylindrical or 'dumb-bell' shaped haematite ring within which is a quartz mosaic, each unit of which displays the optical characteristics of one of the quartz grains which surround the haematite ring (Plate 8-1h). As with the 'spherulites' within lithotype 1A, these structures are larger than the containing sediment, again by up to 3 times.

Occasional purple beds within the Bayley Range Formation were found to be tuffaceous. Within the typically green Bayley Range lithologies, however, only rare spherulite-like structures were seen as isolated fragments and of similar size as the containing sediment. These fragments are believed to have been reworked, and not derived directly from an ash fall.

The stratigraphic position of these 'spherulites' within the upper Brachina Subgroup sequence suggests that basic volcanism accompanied not only the initial phase of tectonic uplift upon the Gawler Craton (lithotypes 1A, 3A, 3B and basal 5A), but also both Uplifts I (uppermost lithotype 5A, lithotype 6 and basal lithotype 7A) and II (uppermost lithotype 7A). No such 'spherulites' were observed within the reworked lithotypes of facies associations 4 and 10, nor within the sediments of facies association 9, locally derived from islands within the basin .

SANDSTONE PETROLOGY

The coarse-grained lithotypes of the upper Brachina Subgroup show varying degrees of compositional and textural maturity ranging from (almost) monomineralic, very well sorted quartzarenites to very poorly sorted sublitharenites (see Figs. 3-6, 8, 11, 13 and 14). Between these end members a continuous series exists which, as seen below and in figure 8-2, relates to nearness to source, the number and type of sedimentary environments through which the sands have passed (i.e. the degree of reworking and abrasion suffered by the sands) and the environment into which the sands were finally deposited.

Firstly, however, the mineral components of these sandy lithotypes are discussed.

COMPOSITION

Quartz:- In all sandy lithotypes of grain size smaller than coarse sand (i.e. those of facies associations 1, 2, 3, 4, 5, 6 and 7) unstrained monocrystalline quartz dominates to the near exclusion of all other possible constituents. Of the remaining constituents strained monocrystalline quartz averages about 5%, polycrystalline quartz is present in trace amounts, and about 5% consists of feldspar, rock fragments, secondary minerals and heavy minerals. Inclusions are occasionally seen within the quartz grains, including rutile needles, euhedral and rounded tourmaline and zircon, muscovite, biotite, magnetite, ilmenite and very rarely kyanite and microcline feldspar. When the original grain boundaries are preserved through the pervasive quartz cement by either dust-like inclusions or haematitic coatings (Plates 8-2a and 2b), these quartz grains display very well rounded outlines.

The dominance of unstrained monocrystalline quartz within these fine to medium lithotypes suggests prolonged abrasion (destroying the polycrystalline and strained monocrystalline quartz), whilst the inclusions indicate a dominantly granitic primary source, with some addition from metamorphic rocks.

The sandy lithotypes of facies associations 8, 9 and 10 contain a component fraction of coarse sand to granule size which is dominated by polycrystalline and strained monocrystalline quartz. Where this very coarse fraction forms a separate grain size mode, totalling between 35 and 90% of the total rock (i.e. facies association 8 and lithotypes 9A, 9C and 10B) polycrystalline quartz constitutes up to 25% of the entire rock, whilst strained monocrystalline quartz can reach 40%. Lithotype 10A, on the other hand, contains only a 'tail' of very coarse sand, and is dominated by coarse-grained unstrained monocrystalline quartz, with less than 10% polycrystalline quartz, suggesting a higher degree of abrasion than in the truly bimodal lithotypes. In these bimodal lithotypes the quartz cement has grown in optical continuity with the finer grained fraction, thus leaving the boundaries of the coarser grains free, revealing their very well rounded nature.

Feldspar:- Although ubiquitous, feldspar generally totals less than 5% of the composition within the majority of the sandy lithotypes. With decrease in grain size to very fine sand and silt, the feldspar content increases, occasionally to subarkosic proportions (e.g. in facies associations 1, 2, 6 and the very fine sandstone subfacies of lithotype 9A). Throughout the entire sequence orthoclase feldspar dominates, whilst plagioclase and microcline feldspar are present in trace amounts. As with the quartz grains, the feldspars are very well rounded, and are often readily identified by the lack of optical continuity with the surrounding quartz cement. Invariably both pure and sericitized fresh feldspars, along with altered and cloudy feldspars are present, suggesting the presence of both close and distant primary (granitic) source rocks.

Rock Fragments:- Rock fragments are only present in trace amounts throughout the fine to medium sandy lithotypes of the upper Brachina Sub-group. Detrital chert is ubiquitous, but porphyritic and silicified siltstone fragments are also seen, though very rarely. In the facies associations typified by coarse sand (i.e. facies associations 8, 9 and 10) rock fragments total up to 20% of the entire rock. In these rocks silicified (?)oolitic carbonates (Plate 8-2c), fine-grained siliceous clastics and detrital chert are common, whilst porphyritic fragments are only rarely seen. Also present within facies association 9 are cherts (silicified carbonates) and also rare carbonate, dolerite and volcanic fragments.

Other Light Minerals:- Micas are found in trace amounts within all the sandy lithotypes. Muscovite is ubiquitous, whilst chlorite and biotite are occasionally seen. Rarely up to 3% muscovite and 15% chlorite (TS. 469/BGL951) are present. These micas are of secondary origin, growing between the detrital grains prior to final cementation. Carbonate is very rarely present (e.g. TS. 469/NBR301) but, as stated earlier, it is believed to be an alteration product of basic volcanic 'spherulites'.

Heavy Minerals:- A supermature heavy mineral suite is present throughout the coarse-grained lithotypes of the upper Brachina Subgroup, a suite typical of quartzite deposits throughout the world. The detrital heavy mineral fraction locally reaches 2%, but on average is present only in trace amounts. It is dominated by opaques, although both tourmalines and zircon are ubiquitous. Rutile is only rarely seen, whilst xenotime and beryl are very rare. The grain size of these minerals is usually an order of magnitude smaller than the containing sediments, with a concentration apparent in the fine sandstone and siltstone lithotypes. The grain outlines of these minerals are very well rounded (Plate 8-2d), although broken edges are common.

Secondary heavy minerals are present in the form of specular haematite (Plate 8-2e), found within facies association 9 and where faulting has shattered the rock allowing iron rich solutions to filter through, and authigenic tourmaline, present in nearly all sandy lithotypes as overgrowths on detrital tourmaline grains (Plate 8-2f).

According to Krynine (1946) the colour of detrital tourmaline can be used as an indicator of its mode of origin. In broad terms he recognises four distinct modes of origin. Tourmaline displaying pleochroic blue colouration is diagnostic of a pegmatitic origin, whilst those that have a colourless-blue pleochroism are of authigenic origin within a sedimentary environment. Granitic terrains are indicated by green, brown, and pink to black tourmaline, whilst pleochroic colourless to yellow and light brown tourmaline is typical of metamorphic terrains. According to my colleagues Messrs. N. Manktelow, A.C. Purvis and G. Teale, however, pale brown and yellow tourmalines are also found in the low grade metamorphic rocks of the Adelaidean and Cambrian of Fleurieu Peninsula and the Mount Painter Province, and the Archaean near Kalgoorlie. Within metamorphic rocks of amphibolite grade and higher, on the other hand, tourmalines are typically dark brown and green.

In an endeavour to ascertain the provenance of the detrital heavy mineral fraction within the sandy lithotypes of each facies association within the upper Brachina Subgroup a count was made of every tourmaline grain within each slide from nine measured sections (see page A33, Appendix IV for localities). A total of 1758 tourmaline grains were counted from 122 thin sections, at an average of only 14.5 grains per slide. The total number of grains counted from each facies association of each section along with the percentage breakup of this total into granitic and high grade metamorphic, pegmatitic, low grade metamorphic, and authigenic tourmaline is given on page A35 of Appendix IV. Also, percentage columns have been constructed for the variation within each of the major facies associations at each section (see Appendix IV, pages A38 and A39) except sections AGG and MDG where too few grains were present. Chi-square tests were conducted on this data at four different levels, namely:-

1. each facies association in each section against the average for that facies association;
2. the average of each section against the average for the entire sequence;
3. each facies association in each section against the average for that section; and
4. the average of each facies association against the average for the entire sequence.

The results of these tests are summarized on pages A36 and A37 of Appendix IV. Although significant differences occur at each test level, no trends were forthcoming linking, for example, the depletion of one type of tourmaline from one facies association with its enrichment in another facies association, thus possibly indicating separate sources; or statistically identical tourmaline counts, suggesting possible reworking of one facies association to produce another.

Some inferences, however, can be drawn from the detrital heavy mineral fraction present within the upper Brachina Subgroup. These are:

1. that the major primary source was a granitic and high grade metamorphic terrain although pegmatitic, low grade metamorphic and pre-existing sedimentary source rocks all contributed in significant, though minor, amounts; and
2. that the well rounded nature of the detrital heavy minerals, especially authigenic tourmaline, indicates a multicyclic origin for the sediments prior to their deposition within the upper Brachina Subgroup.

CEMENT

A distinctive and problematic feature of many thick and extensive quartzite deposits throughout the world is the presence of a pervasive silica cement, usually quartz. Such cements often hide all textural characteristics of the deposits, leaving an interlocking quartz mosaic. Invariably the boundaries between the quartz within this mosaic are angular to irregular, and this earlier led to the belief that pressure solution phenomena were responsible for the generation of the secondary quartz needed for the silica cementation to produce such mosaics (Skolnick 1965). It was later noted, however, from thin section work on quartzites in which the original detrital grain outlines were preserved by thin rims of haematite, gas bubbles or dust-like inclusions, or more recently by cathodoluminescence where such rims are absent, that the original grain boundaries were invariably not in contact, and not only were pressure solution phenomena rarely seen, such as sutured grain boundaries, micro-stylolites and/or stylolites (see Heald 1955), but the amount of secondary quartz needed for cementation was far greater than previously thought. Pressure solution is now believed to be only an end-product of quartz enlargement (Sippel 1968), and the problem of "from where was all the secondary cementing quartz derived?" still remains.

Cementation within most of the sandy lithotypes of the upper Brachina Subgroup (the sands of lithotypes 1A, 2, 3A, 4A, 4B, 5A, 6, 7A, 7B, 8, 9B, 10A and 10B) is typically siliceous and pervasive. Secondary quartz has precipitated in optical continuity with each detrital quartz grain, but invariably rims of dust-like inclusions, and occasionally haematite, outline many of the original grain boundaries (Plates 8-2a and 2b). Only rarely are sutured grain boundaries, or the optical effects of pressure solution seen (Plates 8-2g and 2h).

Within lithotypes 9A and 9C iron (haematite) cementation is typical. In some cases haematite forms the only cement, but in many cases both haematite and quartz cements are present. In such cases quartz invariably binds the finer fraction (silt and fine sand), whilst haematite binds the coarser grains. Often in these rocks a three stage cementation process is evident, with an initial quartz phase, followed by one of haematite, and finally completed by a second quartz phase. Other sandy lithotypes within which an original haematite cement is present, though very minor, are those of facies associations 1, 3, 6 and 8. The haematite cement within facies association 1 is invariably derived from basic volcanigenic debris (see earlier), whilst the remainder were all subjected to either strong fluvial influence, or periods of prolonged subaerial exposure.

TEXTURE

The sandy lithotypes of the upper Brachina Subgroup fall into one of four broad textural classes - viz. multimodal, bimodal, unimodal (moderately sorted, but with a prominant coarse tail) or unimodal (very well sorted). As shown in figure 8-2, this grain size modality can be broadly related to the position of the depositional environment within the basin and the sedimentary processes each lithotype has been subjected to prior to its final deposition. It appears that with increased distance from the source and, more importantly, prolonged abrasion (dominantly within tidal environments) the coarser grains of the multimodal and bimodal lithotypes are rapidly abraded (in the case of fairly resistant unstrained monocrystalline quartz) and/or broken up (in the case of labile components such as rock fragments and polycrystalline quartz) until a very well sorted unimodal grain size distribution is attained. This trend toward unimodality, therefore, is parallelled by a trend toward complete dominance of composition by unstrained monocrystalline quartz. The ultimate end member of the quartzite series, therefore, would be a very well sorted sand composed of almost 100% unstrained monocrystalline quartz, with trace amounts of tourmaline, zircon, rutile, opaques and possibly detrital chert.

An attempt was made to numerically summarize the petrologic data of the upper Brachina Subgroup sandy lithotypes to aid the comparison and contrast study by using computer assisted cluster and principal component analysis techniques. The programmes CLUSTA and PRINCA (see Fitzgerald 1975) were used on coded data involving the textural characteristics and percentage groupings of the major components. Such analyses are often useful in diagrammatically delineating any trends which occur in the data. Examples of the results of these analyses are presented in figure 8-3 and, as can be readily seen except for the isolation of the coarser grained bimodal lithotypes of facies associations 8, 9 and 10, no obvious trends are present within the petrologic data.

All distinctions relating to source areas of these sandstones have been gleaned from both 'positive' and 'negative' attributes scattered throughout the entire sequence, and such numerical techniques cannot be critical enough to make such distinctions.

PROVENANCE

The sediments of the upper Brachina Subgroup were initially deposited within a deltaic system (the "Alligator River Delta") which gradually migrated out into the basin from a western shoreline. The sands of this system display a bimodal grain size distribution in the most proximal reaches of the delta (facies association 8) and a unimodal grain size distribution in the more distal reaches (facies association 5). Throughout the delta, the presence of both fresh and altered feldspars along with the very well rounded nature of the heavy mineral fraction (especially that of the authigenic tourmalines) and the rare presence of worn quartz overgrowths, suggest that, apart from the nearby and distant original primary granitic and, to a lesser extent, pegmatitic and metamorphic source rocks (as indicated by the tourmalines and inclusions within quartz grains), pre-existing sedimentary rocks also supplied detritus into the deltaic system. These data, coupled with the areal distribution and palaeocurrent patterns of the major facies associations, suggest that the initial upper Brachina Subgroup sediments were derived from the Gawler Craton situated to the west of the Adelaide 'Geosyncline'.

Within these deltaic sands, haematitic coatings are occasionally seen separating the original detrital grain outline from the quartz cement (see Plate 8-2b). Such coatings are believed to originate within desert environments by pervasive intrastratal solution of iron silicates (Walker 1967, Hubert and Reed 1978). Also, according to Folk (1968b), desert environments are responsible for the bimodal texture of many supermature sand deposits. Thus the presence, though rare, of both haematitic rims and a bimodal texture within the sands of the "Alligator River Delta" system suggests a probable period of aeolian deflation upon the Gawler Craton prior to their inclusion into the deltaic system.

The sediments of facies association 9 derived from small islands within the basin, on the other hand, show virtually no haematitic rims on any of the component grains, and also comprise a greater amount of labile constituents, such as rock fragments and polycrystalline quartz. As such, no phase of aeolian deflation prior to deposition is suggested. Their very well rounded grain boundaries and restricted source areas, therefore, indicate that their derivation was primarily from pre-existing sedimentary rocks. The apparently anomalous doleritic and volcanigenic rock fragments are diagnostic of the many 'diapiric' complexes found scattered throughout the central Flinders Ranges, and, as indicated in Chapter 6, two such regions were exposed as island source areas for the deposits of this facies association.

CONCLUSIONS

"...whereas many different criteria may throw light on the source of sands and especially on the environment of their deposition, few of them are sufficiently diagnostic to be used alone, but when they are considered together and balanced one against another they may permit safe inferences to be drawn..."

Dunbar & Rogers (1957)

The Adelaide Supergroup was deposited in a huge basin of shallow water sediment accumulation (the Adelaide 'Geosyncline') which separated one cratonic nucleus to the west (the Gawler Craton) from another to the northeast (the inferred 'Curnamona Nucleus' of Thomson (1976)). Continued, though non-uniform subsidence of this basin throughout the late Precambrian and into the Lower and Middle Cambrian, interrupted by phases of uplift tectonism, produced a thick succession of dominantly shallow marine deposits within which are displayed both major and minor erosive breaks.

One tectonically defined phase of sedimentation was responsible for the deposition of the Brachina Subgroup. Immediately following the uppermost Precambrian Elatina glaciation a thin dolomite (the basal Nuccaleena Formation) was deposited under predominantly intertidal conditions, although supratidal conditions prevailed locally around islands within the centre of the basin. As the post-glacial isostatic rebound abated, general subsidence occurred throughout the basin. This subsidence, accompanied by an influx of fine-grained detritus surrendered from the tectonically stable, almost peneplained enframing cratonic regions, enabled the thick Moolooloo Formation to develop within shallow submerged mudflats (see Fig. 4-1).

Deposition of the Moolooloo Formation within these shallow submerged mudflats persisted until a phase of uplift tectonism and minor accompanying basis volcanism occurred upon the Gawler Craton. This phase of tectonic activity generated a vast amount of sand-sized detritus which, when debouched into the basin, produced the deltaic phase of upper Brachina Subgroup sedimentation. Prior to the debouchment of this sand into the basin, however, it was seemingly subjected to a phase of aeolian deflation upon the craton.

Initially this sand-sized detritus entered the basin through three distal distributary channels (lithotype 3A), flanked by subaqueous levee deposits (lithotype 3B). These channels extended eastward across the aerobic Moolooloo mudflat and deposited their load in distributary mouth shoals upon, and at the base of the gentle palaeoslope leading into the anaerobic Moolooloo mudflat of the basin's eastern region (see Fig. 5-5). With continued sand influx a deltaic plain quickly developed. This deltaic plain, dominated by a fluvial current system and comprising topset beds of sand and mud (facies association 5) and cut by relatively stable distributary channels (lithotype 3A), rapidly prograded into the basin until the gentle eastward palaeoslope was encountered. At the base of this palaeoslope deltaic bottomset deposits accumulated (lithotype 1B) within an anaerobic subtidal mudflat. Wave energy generated above this deltaic bottomset reworked the distal deltaic sands into a protective barrier-bar system (facies association 4; see Fig. 5-11). Meanwhile, to the northeast of the delta, within the central region of the basin, fine-grained deposits accumulated within a low intertidal mudflat (lithotype 1A). Upon this mudflat cup-shaped coelenterates(?) ancestral to the Ediacara fossil assemblage flourished. The initial deltaic configuration therefore, was that of a fluvial and tide modified, wave-dominated system fed by stable distributary channels and protected by a barrier-bar complex.

Gradually, the massive detrital influx induced the sedimentation rate to exceed the rate of basin subsidence, causing the emergence of the depositional environments into the intertidal zone throughout the basin. The change in energy conditions from wave-dominated to tide-dominated across the now ebb-tide dominated, intertidal mudflat surrounding the delta (facies association 2) resulted in the abandonment of the protective barrier-bar system, and the subsequent change from a stable, to a migrating distributary channel outlet system. The deltaic configuration thus evolved, with basin shallowing, to a fluvial modified, tide-dominated system (see Fig. 5-12).

This latter deltaic configuration persisted until tectonic instability episodically affected the western margin and, briefly, the centre of the basin to produce the sandflat phase of upper Brachina Subgroup sedimentation. The initial episode of tectonic uplift to affect the basin margin (Uplift I) caused partial erosion of the northern reaches of the previously deposited deltaic succession. The eroded deposits were reworked by a roughly east-west tidal current system into an intertidal sandflat (facies association 7), and flanked basinward by an intertidal mudflat (facies association 2) and landward by a high intertidal mudflat (facies association 6). In the region of the "Alligator River" outlet coarser sand-sized detritus was being debouched (facies association 8; see Fig. 6-6).

A second episode of uplift tectonism (Uplift II) affected not only the basin's western margin, but also caused the emergence of at least two islands within the centre of the basin. Continued deposition of the coarse sands within a deltaic environment occurred to the south of the newly uplifted margin, whilst flanking this margin a narrow sand deposit of probable beach origin developed (lithotype 10A). Within the centre of the basin, in the vicinity of the present day Blinman-Enorama-Oraparinna and Worumba 'diapir' complexes, islands emerged shedding thin deposits of coarse sand through fluvial braided channel to marine-dominated distal fan systems (facies association 9; see Figs. 6-9 and 10).

Eventually tectonic uplift ceased within the basin and gradual subsidence resumed, enabling deposition within a shallow aerobic, possibly tidal mudflat to prevail (producing the Bunyeroo Formation). Along the western margin of the basin, however, a third and final episode of tectonic uplift occurred (Uplift III) producing, as before, a narrow sand deposit of probable beach origin (lithotype 10B; see Fig.7-2). By this time, however, the influx of coarse detritus through the "Alligator River" had seemingly ceased and, upon cessation of Uplift III, deposition of very fine-grained detritus, again surrendered from the once more tectonically stable, almost peneplained enframing cratonic regions, prevailed throughout the Adelaide 'Geosyncline'.

REFERENCES

ALDERMAN, A.R. (1973) Southern Aspect: An Introductory View of South Australian Geology. 158pp. S.Aust. Museum: Adelaide.

ALLEN, J.R.L. (1963) The classification of cross-stratified units with notes of their origin. Sedimentology 2: 93-114.

ALLEN, J.R.L. (1966) On bedforms and palaeocurrents. Sedimentology 6: 153-190.

ALPERT, S.P. (1977) Trace fossils and the basal Cambrian boundary. In: Crimes, T.P. and Harper, J.C. (Eds.): Trace Fossils 2. Spec. Issue geol. J. 9: 1-8.

ANDERSON, F.E. (1973) Observations of some sedimentary processes acting on a tidal flat. Mar. Geol. 14: 101-116.

ANDERTON, R. (1974) Tidal-shelf sedimentation: An example from the Scottish Dalradian. Sedimentology 23: 429-458.

AUDLEY-CHARLES, M.G., CURRAY, J.R. and EVANS, G. (1977). Location of major deltas. Geology 5: 341-344.

AUSTIN, J.B. (1863) The Mines of South Australia. 109pp. Platts et al.: Adelaide.

AUSTRALIAN CODE OF STRATIGRAPHIC NOMENCLATURE (1973) J. geol. Soc. Aust. 20: 105-112.

BALAZS, R.J. and KLEIN, G. deV. (1972). Roundness-mineralogical relations of some intertidal sands. J. sedim. Pet. 42: 425-433.

BANKS, N.L. (1973) Tide-dominated offshore sedimentation, Lower Cambrian, north Norway. Sedimentology 20: 213-228.

BARNES, J.J. and KLEIN, G. deV. (1975) Tidal deposits in the Zabriskie Quartzite (Cambrian) eastern California and western Nevada. In: Ginsburg, R.N. (Ed.): Tidal Deposits. Springer-Verlag: New York. pp.163-169.

BART, H.A. (1977) Sedimentology of cross-stratified sandstones in Arikaree Group, Miocene, southeastern Wyoming. Sediment. Geol. 19: 165-184.

BELL, H.S. (1940) Armored mud balls - their origin, properties and role in sedimentation. J. Geol. 48: 1-31.

BERNER, R.A. (1970) Sedimentary pyrite formation. Am. J. Sci. 268: 1-23.

BHARGAVA, O.N. (1972) A note on structures resembling molds of tadpole nests in the Upper Jurassic Tal Formation, Simla Himalaya, India. J. Sedim. Pet. 42: 236-238.

BINKS, P.J. (1971) Geology of the ORROROO 1:250,000 map area. Rep. Invest. geol. Surv. S.Aust. 36: 114pp.

BLATT, H., MIDDLETON, G. and MURRAY, R. (1972) Origin of Sedimentary Rocks. 634pp. Prentice-Hall Inc.: New Jersey.

BORCHERT, H. (1960) Genesis of marine sedimentary iron ores. Trans. Inst. Mining Metal. Lond. 69: 261-279.

BOSE, U. (1977) Fusiform structures in Sallavai Sandstone - biogenic or abiogenic? J. geol. Soc. India 18: 347-348.

BRETT, G.W. (1955) Cross-bedding in the Baraboo Quartzite of Wisconsin. J. geol. 63: 143-148.

von BRUNN, V. (1974) Tidalites of the Pongola Supergroup (early Precambrian) in the Swart-Mfolozi area, northern Natal. In: Kröner, A. (Ed.): Bull. contrib. Precamb. Geol., S. Af. 15: 107-122.

von BRUNN, V. and HOBDAY, D.K. (1976) Early Precambrian tidal sedimentation in the Pongola Supergroup of South Africa. J. sedim. Pet. 46: 670-679.

BRUSH, L.M., Jr. (1958) Study of cross-stratification in a large laboratory flume. Bull. geol. Soc. Am. 69: 1542.

BURR, T. (1846) Remarks on the geology and mineralogy of South Australia. In: Cawthorne, W.A. (Ed.): Legend of Kuperree and other pamphlets Pamphlets Aborigines 1: 32pp.

BUTTON, A. and VOS, R. (1977) Subtidal and intertidal clastic and carbonate sedimentation in a macrotidal environment: An example from the lower Proterozoic of South Africa. Sediment. Geol. 18: 175-200.

CAMPANA, B. (1958) The Flinders Ranges. In: Glaessner, M.F. and Parkin, L.W. (Eds.): The geology of South Australia. J. geol. Soc. Aust. 5: 28-45.

CAMPANA, B. and KING, D. (1958) Regional geology and mineral resources of the Olary Province. Bull. geol. Surv. S.Aust. 34: 133pp.

CAMPBELL, C.V. (1976) Reservoir geometry of a fluvial sheet sandstone. Bull. Am. Assoc. Petrol. Geol. 60: 1009-1020.

CHAMBERLAIN, C.K. (1971) Morphology and ethology of trace fossils from the Ouachita Mountains, southeast Oklahoma. J. Palaeont. 45: 212-246.

CHANDA, S.K. and BHATTACHARYYA, A. (1974) Ripple-drift cross-lamination in tidal deposits: Examples from the Precambrian Bhander Formation of Maihar, Satna District, Madhya Pradesh, India. Bull. geol. Soc. Am. 85: 1117-1122.

CLARKE, E. deC. (1938) Middle and West Australia. Reg. Geol. der Erde, Bd. 1: Abstr VII. Akad. Verlags: Leipzig.

COATS, R.P. (1962) The control of sedimentation by faulting in the Adelaide System. Quart. geol. Notes, geol. Surv. S.Aust. 4: 6-7.

COATS, R.P. (1965) Diapirism in the Adelaide Geosyncline. J. Aust. Petrol. Explor. Assoc. 5: 98-102.

COATS, R.P. (1973) Explanatory notes on the COPLEY 1:250,000 geological map. Geol. Surv. S.Aust.: 70pp.

COATS, R.P. and BLISSETT, A.H. (1971) Regional and economic geology of the Mount Painter province. Bull. geol. Surv. S.Aust. 43: 426pp.

COLEMAN, J.M. and WRIGHT, L.D. (1975) Modern river deltas: Variability of processes and sand bodies. In: Broussard, M.L. (Ed.): Deltas: Models for exploration. Houston geol. Soc.: 99-149.

COLLINSON, J.D. (1970) Bedforms of the Tana River, Norway. Geografiska Annaler 52A: 31-56.

COMPSTON, W. and ARRIENS, P.A. (1968) The Precambrian geochronology of Australia. Can. J. Earth Sci. 5: 561-583.

COMPSTON, W., CRAWFORD, A.R. and BOFINGER, V.M. (1966) A radiometric estimate of the duration of sedimentation in the Adelaide Geosyncline, South Australia. J. geol. Soc. Aust.13: 229-276.

CONYBEARE, C.E.B. and CROOK, K.A.W. (1968) Manual of sedimentary structures. Bull. Bur. Mineral Res., Geol. Geophys., Aust. 102: 327pp.

COOPER, J.A. (1975) Isotopic datings of the basement-cover boundaries within the Adelaide "Geosyncline". Abstr. 1st Aust. Geol. Conv., geol. Soc. Aust.: 12.

COOPER, J.A. and COMPSTON, W. (1971) Rb/Sr dating within the Houghton Inlier, South Australia. J. geol. Soc. Aust. 17: 213-219.

CRAWFORD, A.R. (1962) Absolute age determinations in South Australia. Quart. geol. Notes, geol. Surv. S.Aust. 2: 6-7.

CRIMES, T.P., LEGG, J., MARCOS, A. and ARBOLEYA, M. (1977) ?Late Precambrian-low Lower Cambrian trace fossils from Spain. In: Crimes, T.P. and Harper, J.C. (Eds.): Trace Fossils 2. Spec. Issue geol. J. 9: 91-138.

CURRAY, J.R. (1969) Estuaries, lagoons, tidal flats and deltas. In: Stanley, D.J. (Ed.): The new concepts of continental margin sedimentation. Am. geol. Inst.: JC-3-1 - JC-3-30.

CURTIS, C.D. and SPEARS, D.A. (1968) The formation of sedimentary iron minerals. Econ. Geol. 63: 257-270.

DAILY, B. (1956) The Cambrian in South Australia. In: El Sistema Cambrico, su palaeogeographica Y el problema de su base. Rep. 20th int. Geol. Congr., Mexico, 21: 91-147.

DAILY, B. (1963) The fossiliferous Cambrian succession on Fleurieu Peninsula, South Australia. Rec. S.Aust. Museum 14: 579-601.

DAKE, C.L. (1921) The problem of the St. Peter Sandstone. Bull. Missouri School Mines Metal., tech. Ser. 6: 228pp.

DALGARNO, C.R. and JOHNSON, J.E. (1964) The Wilpena Group. In: Thomson, B.P. et al.: Precambrian rock groups in the Adelaide Geosyncline: A new subdivision. Quart. geol. Notes, geol. Surv. S. Aust. 9: 9-16.

DAVID, T.W.E. (1922) Occurrence of remains of small Crustacea in the Proterozoic(?) or Lower Cambrian(?) rocks of Reynella, near Adelaide. Trans. R. Soc. S.Aust. 46: 6-8.

DAVID, T.W.E. (and BROWNE, W.R.) (1950) The Geology of the Commonwealth of Australia. Vol. I. 745pp. Edward Arnold & Co.: London.

DAWSON, J.W. (1878) Supplement to the second edition of Acadian Geology. In: Acadian Geology. 3rd Edn. Macmillan: London.

DAWSON, J.W. (1890) On burrows and tracks of invertebrate animals in Palaeozoic rocks, and other markings. Quart. J. geol. Soc. Lond. 46: 595-618.

DIONNE, J.C. and LAVERDIÈRE, C. (1972) Structure cylindrique verticale dans un dépôt meuble Quarternaire, au nord de Montréal, Quebec. Can. J. Earth Sci. 9: 528-543.

DONOVAN, R.N. and FOSTER, R.J. (1972) Subaqueous shrinkage cracks from the Caithness Flagstone Series (Middle Devonian) of northeast Scotland. J. sedim. Pet. 42: 309-317.

DORR, J.A., Jr. and KAUFFMAN, E.G. (1963) Rippled toroids from the Napolean Sandstone Member (Mississippian) of southern Michigan. J. sedim. Pet. 33: 751-758.

DRAPER, J.J. (1977) Environment of deposition of the Carlo Sandstone, Georgina Basin, Queensland and Northern Territory. J. Bur. Mineral Res, Geol. Geophys., Aust. 2: 97-110.

DUNBAR, C.O. and ROGERS, J. (1958) Principles of Stratigraphy. 356pp. Wiley & Sons: New York.

DZULYNSKI, S. and KOTLARCZYK, J. (1962) On load-casted ripples. Ann. Soc. géol. Pologne 32: 148-159.

ENDO, R. (1933) Manchuriophycus, nov.gen., from a Sirian formation of South Manchuria. Jap. J. geol. Geogr. 11: 43-48.

ERIKSSON, K.A. (1977) Tidal deposits from the Archaean Moodies Group, Barberton Mountain Land, South Africa. Sediment.Geol. 18: 257-281.

EVANS, G. (1965) Intertidal flat sediments and their environments of deposition in The Wash. Quart. J. geol. Soc. Lond. 121: 209-245.

FITZGERALD, M.J. (1975) AURPAK - A computer package to process the results of whole rock analyses. Analyt. Geochem.Rep. 4/75 (unpublished). Dept. Geol. Mineral., Univ. Adelaide.

FLORES, R.M. (1977) Marginal marine deposits in the upper Tesnus Formation (Carboniferous), Marathon Basin, Texas. J. sedim. Pet. 47: 582-592.

FOLK, R.L. (1960) Petrography and origin of the Tuscarora, Rose Hill and Keefer Formations, Lower and Middle Silurian of eastern West Virginia. J. sedim. Pet. 30: 1-58.

FOLK, R.L. (1968a) Petrology of Sedimentary Rocks. 170pp. Hamphill's Book Store: Texas.

FOLK, R.L. (1968b) Bimodal supermature sandstones. Product of the desert floor. Proc. 23rd int. Geol. Congr., Czechoslovakia, 8: 9-22.

FOLK, R.L., ANDREWS, P.B. and LEWIS, D.W. (1970) Detrital sedimentary rock classification and nomenelature for use in New Zealand. New Zd. J. Geol. Geophys. 13: 937-968.

FORD, C. (1974) Study of the sedimentology, palaeontology and stratigraphy of the Pound Quartzite, southern Flinders Ranges. Unpublished Rep., Dept. Geol. Mineral., Univ. Adelaide.

FORBES, B.G. (1971) A table of Adelaidean and Cambrian stratigraphic names. Quart. geol. Notes, geol. Surv. S.Aust. 38: 5-6.

FREY, R.W. (1970) The lebensspuren of some common marine invertebrates near Beaufort, North Carolina. II Anemone burrows. J. Palaeont. 44: 308-311.

FRIEDMAN, G.M. (1962) Comparison of moment measures for sieving and thin-section data in sedimentary petrological studies. J. sedim. Pet. 32: 15-25.

FROME, E.C. (1843) Report on the country to the north of the sources of the Light and Gilbert. Rep. Surveyor General 3.

GABELMAN, J.W. (1955) Cylindrical structures in Permian(?) siltstone, Eagle County, Colorado. J. Geol. 63: 214-227.

GALLOWAY, W.E. (1975) Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. In: Broussard, M.L. (Ed.): Deltas: Models for exploration. Houston geol. Soc.: 87-98.

GERMS, G.J.B. (1968) Discovery of a new fossil in the Nama System, South West Africa. Nature 219: 53-54.

GERMS, G.J.B. (1973) Possible Sprigginid worm and a new trace fossil from the Nama Group, South West Africa. Geology 1: 69-70.

GINGERICH, P.D. (1969) Markov analysis of cyclic alluvial sediments. J. sedim. Pet. 39: 330-332.

GINSBURG, R.N. (Ed.) (1975) Tidal Deposits: A Casebook in Recent and Examples and Fossil Counterparts. 428pp. Springer-Verlag: New York.

GLAESSNER, M.F. (1969) Trace fossils from the Precambrian and basal Cambrian. Lethaia 2: 369-393.

GLAESSNER, M.F. (1971) Geographic distribution and time range of the Ediacara Precambrian fauna. Bull. geol. Soc. Am. 82: 509-514.

GLAESSNER, M.F. and WADE, M. (1966) The late Precambrian fossils from Ediacara, South Australia. Palaeont. 9: 599-622.

GLAESSNER, M.F. and WALTER, M.R. (1975) New Precambrian fossils from the Arumbera Sandstone, Northern Territory, Australia. Alcheringa 1: 59-69.

GLEN, R.A., LAING, W.P., PARKER, A.J. and RUTLAND, R.W.R. (1977) Tectonic relationships between the Proterozoic Gawler and Willyama orogenic domains, Australia. J. geol. Soc. Aust. 24: 125-150.

GODDARD, E.N., TRASK, P.D., DeFORD, R.K., ROVE, O.N., SINGEWALD, J.T., Jr. and OVERBANK, R.M. (1963) Rock-colour chart. Geol. Soc. Am: 16pp.

GOLDRING, R. and CURNOW, C.N. (1967) The stratigraphy and facies of the late Precambrian at Ediacara, South Australia. J. geol. Soc. Aust. 14: 195-214.

GRIM, R.E. (1951) The depositional environment of red and green shales. J. sedim. Pet. 21: 226-232.

HAMBLIN, W.K. (1961) Palaeogeographic evolution of the Lake Superior region from Late Keweenawan to Late Cambrian time. Bull. geol. Soc. Am. 72: 1-18.

HÄNTZSCHEL, W. (1949) Zur Deutung von Manchuriophycus endo und ähnlichen Problematika. Mitt. geol. Staatsinst. Hamburg 19: 77-84.

HÄNTZSCHEL, W. and REINECK, H.E. (1968) Faziesuntersuchungen im Hettangium von Helmstedt (Niedersachsen). Mitt. geol. Sattsinst. Hamburg 37: 5-39.

HARMS, J.C. (1969) Hydraulic significance of some sand ripples. Bull. geol. Soc. Am. 80: 363-396.

HARMS, J.C., SOUTHARD, J.B., SPEARING, D.R. and WALKER, R.G. (1975) Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Short Course Soc. econ. Palaeont. Mineral., Tulsa 2: 161pp.

HAWLEY, J.E. and HART, R.C. (1934) Cylindrical structures in sandstone. Bull. geol. Soc. Am. 45: 1017-1034.

HEALD, M.T. (1955) Stylolites in sandstone. J. Geol. 63: 101-114.

HEINRICH, E. Wm. (1965) Microscopic Identification of Minerals. 414pp. McGraw-Hill: New York.

HENDRY, H.E. and STAUFFER, M.R. (1977) Folded crossbedding 1972 - In: Bolton, T.E. (Ed.): Current research in the geological sciences in Canada, May 1976-April 1977. Can. Geosci. Council, Paper 77-5: 96.

HOFMANN, H.J. (1967) Precambrian fossils(?) near Elliot Lake, Ontario. Science 156: 500-504.

HOFMANN, H.J. (1971) Precambrian fossils, pseudofossils and problematica in Canada. Bull. geol. Surv. Can. 189: 146pp.

HOLMES, A. (1920) The Nomenclature of Petrology. 284pp. Murby & Co.: London.

van HOUTON, F.B. (1961) Climatic significance of red beds. In: Descriptive palaeoclimatology. Interscience Publ. Inc: New York: 89-139.

van HOUTON, F.B. (1968) Iron oxides in red beds. Bull. geol. Soc. Am. 79: 399-416.

HOWCHIN, W. (1907) A general description of the Cambrian Series of South Australia. Austn. Assoc. Advmt. Sci. 11: 414-422.

HOWCHIN, W. (1918) The Geology of South Australia. 543pp. Govt. Printer: Adelaide.

HUBERT, J.F. and REED, A.A. (1978) Red-bed diagenesis in the East Berlin Formation, Newark Group, Connecticut Valley. J. sedim. Pet. 48: 175-184.

INGRAM, R.L. (1954) Terminology for the thickness of stratification and parting units in sedimentary rocks. Bull. geol. Soc. Am. 65: 937-938.

INTERNATIONAL MATHEMATICAL AND STATISTICAL LIBRARIES, INC. (IMSL). (1975) 4th Edn., Houston: Texas.

JAMES, W.C. and OAKES, R.Q., Jr. (1977) Petrology of the Kinnikinic Quartzite (Middle Ordovician), east-central Idaho. J. sedim. Pet. 47: 1491-1511.

JANSA, L.F. (1975) Tidal deposits in the Monkman Quartzite (Lower Ordovician), northeastern British Columbia, Canada. In: Ginsburg, R.N. (Ed.): Tidal Deposits. Springer-Verlag: New York. pp.153-161.

JENKINS, R.J.F. (1975) The diversification of early metazoans. Abstr. 1st Aust. geol. Conv., geol. Soc. Aust.: 19-20.

JOHNSON, H.D. (1975) Tide and wave-dominated inshore and shoreline sequences from the late Precambrian, Finnmark, North Norway. Sedimentology 22: 45-74.

JOPLING, A.V. and WALKER, R.G. (1968) Morphology and origin of ripple-drift lamination, with examples from the Pleistocene of Massachusetts. J. sedim. Pet. 38: 971-984.

KAMEN-KAYE, M. (1967) Basin subsidence and hypersubsidence. Bull. Am. Assoc. Petrol. Geol. 51: 1833-1842.

KAY, M. (1947) Geosynclinal nomenclature and the craton. Bull. Am. Assoc. Petrol. Geol. 31: 1289-1293.

KELLING, G. (1969) The environmental significance of cross-stratification parameters in an Upper Carboniferous fluvial basin. J. sedim. Pet. 39: 857-875.

KETNER, K.B. (1968) Comparison of Ordovisian eugeosynclinal and miogeosynclinal quartzites in the Cordilleran geogeosyncline. In: Geological Survey Research 1966. Prof. Pap. U.S. geol. Surv. 550-C: C54-C60.

KLEIN, G.deV. (1963) Bay of Fundy intertidal zone sediments. J. sedim. Pet. 33: 844-858.

KLEIN, G. deV. (1966) Directional relationships between primary structures and current systems in a tide-dominated environment. Bull. Am. Assoc. Petrol. Geol. 50: 621.

KLEIN, G.deV. (1967a) Palaeocurrent analysis in relation to modern marine sediment dispersal patterns. Bull. Am. Assoc. Petrol. Geol. 51: 366-382.

KLEIN, G.deV. (1967b) Comparison of recent and ancient tidal flat and estuarine sediments. In: Lauff, G.H. (Ed.): Estuaries. Am. Assoc. Advmt. Sci. Publ. 83: 207-212.

KLEIN, G.deV. (1970a) Tidal origin of a Precambrian Quartzite - the Lower Fine-grained Quartzite (Middle Dalradian) of Islay, Scotland. J. sedim. Pet. 40: 973-985.

KLEIN, G.deV. (1970b) Depositional and dispersal dynamics of intertidal sand bars. J. sedim. Pet. 40: 1095-1127.

KLEIN, G.deV. (1971) A sedimentary model for determining paleotidal range. Bull. geol. Soc. Am. 82: 2585-2592.

KLEIN, G.deV. (1975) Tidalites in the Eureka Quartzite (Ordovician), eastern California and western Nevada. In: Ginsburg, R.N. (Ed.): Tidal Deposits. Springer-Verlag: New York. pp. 145-151.

KLEIN, G.deV. and SANDERS, J.E. (1964) Comparison of sediments from Bay of Fundy and Dutch Wadden Sea tidal flats. J. sedim. Pet. 34: 19-24.

KRAUSKOPF, K.B. (1967) Introduction to Geochemistry. 721pp. McGraw-Hill: New York.

KRYNINE, P.D. (1941) Palaeogeographic and tectonic significance of sedimentary quartzites. Bull. geol. Soc. Am. 52: 1915-1916.

KRYNINE, P.D. (1942) Differential sedimentation and its products during one complete geosynclinal cycle. Anales 1st Congr. Panam. Ingen. de Mines y Geol., Geol. pt. 1, 2: 537-561.

KRYNINE, P.D. (1945) Sediments and the search for oil. Producers Monthly 9: 12-22.

KRYNINE, P.D. (1946) The tourmaline group in sediments. J. geol. 54: 65-87.

KRYNINE, P.D. (1948) The megascopic study and field classification of sedimentary rocks. J. Geol. 56: 130-165.

KUENEN, Ph. H. (1958) Experiments in geology. Trans. geol. Soc. Glasgow 23: 1-28.

KUENEN, Ph.H. (1960) Sand. Sci. Am. 202: 94-110.

LANCE, G.N. and WILLIAMS, W.T. (1967) A general theory of classificatory sorting strategies. 1. Hierarchical systems. The Computer J. 9: 373-380.

LEESON, B. (1970) Geology of the Beltana 1:63,360 map area. Rep. Invest. geol. Surv. S. Aust. 35: 92pp.

LESQUEREUX, L. (1876). Species of fossil marine plants from the Carboniferous measures. Ann. Rep. geol. Surv. Indiana 7: 134-145.

MANGUM, D.C. (1970) Burrowing behaviour of the sea anemone Phyllactis. Biol. Bull. 138: 316-325.

MARDIA, K.V. (1972) Statistics of Directional Data. 357pp. Academic Press: New York.

MASON, M.G., THOMSON, B.P. and TONKIN, D.G. (1978) Regional stratigraphy of the Beda Volcanics, Backy Point Beds and Pandurra Formation on the southern Stuart Shelf, South Australia. Quart. geol. Notes, geol. Surv. S.Aust. 66: 2-9.

MAWSON, D. (1927) Geological notes on an area along the northeastern margin of the northeastern portion of the Willouran Range. Trans. R. Soc. S.Aust. 51: 386-390.

MAWSON, D. (1938) Cambrian and sub-Cambrian formations at Parachilna Gorge. Trans. R. Soc. S.Aust. 62: 255-262.

MAWSON, D. (1939) The late Proterozoic sediments of South Australia. Rep. Aust. N.Z. Assoc. Advmt. Sci. 24: 79-88.

MAWSON, D. (1940) The Adelaide Series. Aust. J. Sci. 3: 25-27.

MAWSON, D. (1941) The Wilpena Pound Formation and underlying Proterozoic sediments. Trans. R. Soc. S.Aust. 65: 295-303.

MAWSON, D. (1942) The structural character of the Flinders Ranges. Trans. R. Soc.S.Aust. 6: 262-272.

MAWSON, D. (1947) The Adelaide Series as developed along the western margin of the Flinders Ranges. Trans. R. Soc. S.Aust. 71: 259-280.

MAWSON, D. (1949) Elatina glaciation: A third recurrence of glaciation evidenced in the Adelaide System. Trans. R. Soc. S.Aust. 73: 117-121.

MAWSON, D. and SEGNIT, E.R. (1948) Purple slates of the Adelaide System. Trans. R. Soc. S.Aust. 72: 276-280.

MAWSON, D. and SPRIGG, R.C. (1950) Subdivision of the Adelaide System. Aust. J. Sci. 13: 69-72.

McBRIDE, E.F. (1963) A classification of common sandstones. J. sedim. Pet. 33: 664-669.

McGOWEN, J.H. and GROAT, C.G. (1971) Van Horn Sandstone, west Texas: An alluvial fan model for mineral exploration. Rep. Invest. Bur. econ. Geol., Univ. Texas 72: 57pp.

McKEE, E.D. and GOLDBERG, M. (1969) Experiments on the formation of contorted structures in mud. Bull. geol. Soc. Am. 80: 231-244.

MERRIFIELD, P.M. and LAMAR, D.L. (1968) Sand waves and early Earth-Moon history. J. geophys. Res. 73: 4767-4774.

MIALL, A.D. (1973) Markov chain analysis applied to an ancient alluvial plain succession. Sedimentology 20: 347-364.

MIALL, A.D. (1974) Palaeocurrent analysis of alluvial sediments: A discussion of directional variance and vector magnitude. J. sedim. Pet. 44: 1174-1185.

MIALL, A.D. (1977) A review of the braided-river depositional environment. Earth-Sci. Reviews 13: 1-62.

MILNES, A.R., COMPSTON, W. and DAILY, B. (1977) Pre-to syn-tectonic emplacement of Early Palaeozoic granites in southeastern South Australia. J. geol. Soc. Aust. 24: 87-106.

MINCHAM, H. (1964) The Story of the Flinders Ranges. 306pp. Rigby Ltd.: Adelaide.

MOUNT, T.J. (1975) Diapirs and diapirism in the Adelaide 'Geosyncline', South Australia. Unpublished Ph.D. thesis. Dept. Geol. Mineral., Univ. Adelaide.

OAKS, R.Q., Jr., JAMES, W.C., FRANCIS, G.G. and SCHULINGKAMP, W.J.,II (1977) Summary of Middle Ordovician stratigraphy and tectonics, northern Utah, southern and central Idaho. Wyoming geol. Assoc. Guidebook, 29th Ann. Field Conf.: 101-118.

OFFLER, R. and FLEMING, P.D. (1968) A synthesis of folding and metamorphism in the Mount Lofty Ranges, South Australia. J. geol. Soc. Aust. 15: 245-266.

OJAKANGAS, R.W. and THOMPSON, J.A. (1977) Modern armored mud balls in an urban environment. J. sedim. Pet. 47: 1630-1633.

OTVOS, E.G., Jr. (1965) Sedimentation-erosion cycles of single tidal periods on Long Island Sound beaches. J. sedim. Pet. 35: 604-609.

PEABODY, F.E. (1947) Current crescents in the Triassic Moenkopi Formation. J. sedim. Pet. 17: 73-76.

PETTIJOHN, F.J. (1957) Sedimentary Rocks. 2nd Edn. 718pp. Harper & Row: New York.

PETTIJOHN, F.J., POTTER, P.E. and SIEVER, R. (1972) Sand and Sandstone 618pp. Springer-Verlag: New York.

PFLUG, H.D. (1972) Zur Fauna de Nama-Schichten im Südwest-Afrika. III. Erniettomorpha, Bau und Systematik. Palaeontographica A139: 134-170.

PIERCE, W.G. (1966) Jura tectonics as a décollement. Bull. geol. Soc. Am. 77: 1265-1276.

PINCUS, H.J. (1953) The analysis of aggregates of orientation data in the earth sciences. J. Geol. 61: 482-509.

PLUMMER, P.S. (1978) Stratigraphy of the lower Wilpena Group (late Precambrian), Flinders Ranges, South Australia. Trans. R. Soc. S.Aust. 102: 25-38.

PLUMMER, P.S. (in prep.) Palaeoenvironmental signficance of the Nuccaleena Formation (late Precambrian), central Flinders Ranges, South Australia.

PLUMMER, P.S. and GOSTIN, V.A. (1976) Faulting contemporaneous with Umberatana Group sedimentation (late Precambrian), southern Flinders Ranges, South Australia. Trans. R. Soc. S.Aust. 100: 29-37.

PLUMMER, P.S., JENKINS, R.J.F. and MORIARTY, K. (in prep.) Remains of the earliest animals?

PLUMMER, P.S. and LEPPARD, P.I. (in press) An analytical technique for uni-, bi-, and tri-modal palaeocurrent data. Computers and Geosciences.

POTTER, P.E. and PETTIJOHN, F.J. (1963) Palaeocurrents and Basin Analysis. 296pp. Springer-Verlag: Berlin.

PRANTL, F. (1945) Dvě záhadné Zklameněliny (stopy). z vrstev chrustenickch-d δ2, Rozpravy II. Tridy České Akad. 55: 3.

PREISS, W. (1971) Stromatolite biostratigraphy in the late Precambrian. Quart. geol. Notes, geol. Surv. S.Aust. 38: 3-11.

QUENSTADT, F.A. (1879) Petrifactenkunde Deutschlands. 1. Abt., V.6: Korallen Die Röhren - und Stern-korallen. 1093pp. Leipzig.

QUIRK, T.T. (1930) Spring pits, sedimentation phenomena. J. Geol. 38: 88-91.

de RAAF, J.F.M. and BOERSMA, J.R. (1971) Tidal deposits and their sedimentary structures (seven examples from Western Europe). Geol. Minjb. 50: 479-504.

de RAAF, J.F.M., BOERSMA, J.R. and van GELDER, A. (1977) Wave-generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland. Sedimentology 24: 451-483.

REINECK, H-E. (1963) Sedimentgefüge im Bereich der südlichen Nordsee. Senckenberg. naturf. Gesell. Abh. 505: 138pp.

REINECK, H-E. (1967) Layered sediments of tidal flats, beaches and shelf bottoms of the North Sea. In: LAUFF, G.H. (Ed.): Estuaries. Am. Assoc. Advmt. Sci. Publ. 83: 191-206.

REINECK, H-E. (1969) Die Entstehung von Runzelmarken. Senckenberg. marit. 1: 165-168.

REINECK, H-E. (1972) Tidal flats. In: Rigby, J.K. and Hamblin, W.K. (Eds.): Recognition of ancient sedimentary environments. Spec. Publ. Soc. econ. Palaeont. Mineral., Tulsa 16: 146-159.

REINECK, H-E. and SINGH, I.B. (1975) Depositional Sedimentary Environments: With Reference to Terrigenous Clastics. 439pp. Springer-Verlag: New York.

REINECK, H-E. and WUNDERLICH, F. (1967) A new method to measure rate of deposition of single lamina on tidal flats and shelf bottoms. Proc. 7th int. Sediment. Congr.

REYNER, M.L. (1955) The geology of the Peake and Denison region. Rep. Invest. geol. Surv. S.Aust. 6: 23pp.

ROWLANDS, N.J. (1973) The Adelaidean System of South Australia: A review of its sedimentation, tectonics and copper occurrences. Proc. Belt Symp. 1973, 1: 80-113.

RUTLAND, R.W.R. and MURRELL, B. (1975) Tectonics of the Adelaide fold belt. Abstr. 1st Aust.geol. Conv., geol. Soc. Aust.: 3-5.

SANDERS, J.E. (1965) Primary sedimentary structures formed by turbidity currents and related resedimentation mechanics. In: Middleton, G.V. (Ed.): Primary sedimentary structures and their hydrodynamic interpretation. Spec. Publ. Soc. econ. Palaeont. Mineral., Tulsa 12: 192-219.

SCHERMERHORN, L.J.G. (1974) Tectonic framework of late Precambrian supposed glacials. In: Wright, A.E. and Moseley, F. (Eds.): Ice Ages: Modern and Ancient. Spec. Issue geol. J. 6: 241-274.

SCHWARTZACHER, W. (1975) Sedimentation models and quantitative stratigraphy Develmt. Sediment. 19: 382pp.

SEGNIT, R.W. (1939) The Pre-Cambrian - Cambrian succession: The general and economic geology of these systems in portions of South Australia. Bull. geol. Surv. S.Aust. 18: 191pp.

SEILACHER, A. (1967) Bathymetry of trace fossils. Mar. Geol. 5: 413-428.

SEILACHER, A. (1968) Origin and diagenesis of the Oriskany Sandstone (Lower Devonian, Appalachians) as reflected in its shell fossils. In: Muller,G. and Friedman, G.M. (Eds.): Recent Developments in Carbonate Sedimentology in Central Europe. Springer-Verlag: New York. pp.175-185.

SELLEY, R.C. (1970) Studies of sequence in sediments using a simple mathematical device. Quart. J. geol. Soc. Lond. 125: 557-581.

SENGUPTA, S. (1966) Studies on orientation and imbrication of pebbles with respect to cross-stratification. J. sedim. Pet. 36: 362-369.

SHANNON, P.M. (1977) Diagenetic concretions from the Ribband Group sediments of County Wexford, Ireland. Geol. Mag. 114: 127-132.

SHINN, E.A. (1968) Burrowing in recent lime sediments of Florida and the Bahamas. J. Palaeont. 42: 879-894.

SHVETSON, M.S. (1954) Concerning some additional aids in studying sedimentary formations. Bull. Moscow Soc. Naturalists, Geol. Sect. 29: 61-66 (in Russian).

SINGH, I.B. (1969) Primary sedimentary structures in Precambrian quartzites of Telemark, southern Norway, and their environmental significance. Norsk. geol. Tidsskr. 49: 1-31.

SIPPEL, R.F. (1968) Sandstone petrology, evidence from luminescence petrography. J. sedim. Pet. 38: 530-554.

SKOLNICK, H. (1965) The quartzite problem. J. sedim. Pet. 35: 12-21.

SOKOLOV, B.S. (1972) Vendskiy etap v istorii zemli (Vendian Stage in the history of the Earth). Mezhd. geol. kongr., Dokl. sov. geol. Prohl. 7: 114-123 (in Russian).

SORBY, H.C. (1857) On the physical geography of the Tertiary Estuary of the Isle of Wight. Edinb. New Phil. J., n.s. 5: 275-298.

SPECHT, R.L. (1972) The Vegetation of South Australia. 2nd Edn. 328pp. Govt. Printer: Adelaide.

SPEER, J.A. (1978) The stratigraphy and depositional environment of the Aphebian Snyder Group, Labrador. Can. J. Earth Sci. 15: 52-68.

SPRIGG, R.C. (1952) Sedimentation in the Adelaide Geosyncline and the formation of the continental terrace. In: Glaessner, M.F. and Rudd, E.A. (Eds.): Sir Douglas Mawson Anniversary Volume: Univ. Adelaide. pp.153-159.

SPRIGG, R.C. (1964) The South Australian continental shelf as a habitat for petroleum. J. Aust. Petrol. Expl. Assoc. 4: 53-63.

SPRY, A.H. (1951) The Archaean complex at Houghton, South Australia. Trans. R. Soc. S.Aust. 74: 115-134.

STEWART, J.H. and POOLE, F.G. (1974) Lower Palaeozoic and uppermost Precambrian Cordilleran Miogeosyncline, Great Basin, western United States. Spec. Publ. Soc. econ. Palaeont. Mineral. Tulsa 22: 28-57.

van STRAATEN, L.M.J.U. (1954b) Composition and structure of recent marine sediments in the Netherlands. Leid. geol. Meded. 19: 1-110.

van STRAATEN, L.M.J.U. (1961) Sedimentation in tidal flat areas. J. Alberta Soc. Petrol. Geol. 9: 203-226.

SUTTON, J. and WATSON, J. (1963) Structures in shallow-water Precambrian sediments from northwest Scotland. Sedimentology 2: 207-214.

SWETT, K., KLEIN, G.deV. and SMIT, D.E. (1971) A Cambrian tidal sand body of northwest Scotland: An ancient-recent analog. J. Geol. 79: 400-415.

TALBOT, J.L. (1967) Subdivision and structure of the Precambrian (Willyama Complex and Adelaide System), Weekeroo, South Australia. Trans. R. Soc. S.Aust. 91: 45-58.

TANKARD, A.J. and HOBDAY, D.K. (1977) Tide-dominated back-barrier sedimentation, Early Ordovician Cape Basin, Cape Peninsula, South Africa. Sediment. Geol. 18: 135-159.

TANNER, W.F. (1958) An occurrence of flat-topped ripple marks. J. sedim. Pet. 28: 95-96.

TANNER, W.F. (1959) The importance of modes in cross-bedding data. J. sedim. Pet. 29: 221-226.

TERRY, R.D. and CHILINGAR, G.V. (1955) Summary of "Concerning some additional aids in studying sedimentary formations" by M.S. Shvetsov. J. sedim. Pet. 25: 229-234.

THIEL, G.A. (1935) Sedimentary and petrographic analysis of the St. Peter Sandstone. Bull. geol. Soc. Am. 46: 559-614.

THOMSON, B.P. (1965) Geology and mineralization of South Australia. In: Geology of Australian Ore Deposits. 8th Commonwlth. Min. Metall. Congr. 1: 270-284.

THOMSON, B.P. (1966) The lower boundary of the Adelaide System and older basement relationships in South Australia. J. geol. Soc. Aust. 13: 203-228.

THOMSON, B.P. (1969a) Precambrian crystalline basement. In: Parkin, L.W. (Ed.): Handbook of South Australian Geology. Geol. Surv. S.Aust.: 21-48.

THOMSON, B.P. (1969b) Precambrian basement cover: The Adelaide System. In: Parkin, L.W. (Ed.): Handbook of South Australian Geology. Geol. Surv. S.Aust.: 49-83.

THOMSON, B.P. (1976) Tectonics and regional geology of the Willyama, Mount Painter and Denison Inlier areas. In: Knight, C.L. (Ed.): Economic geology of Australia and Papua-New Guinea. 1. Metals. Monogr. Austn. Inst. Min. Metall. 5: 469-476.

THOMSON, B.P., COATS, R.P., MIRRAMS, R.C., FORBES, B.G., DALGARNO, C.R. and JOHNSON, J.E. (1964) Precambrian rock groups in the Adelaide Geosyncline: A new subdivision. Quart. geol. Notes. geol. Surv. S. Aust. 9: 19pp.

TIEJE, A.J. (1921) Suggestions as to the description and naming of sedimentary rocks. J. Geol. 29: 650-666.

TIPPER, D.B. and FINNEY, W.A. (1965) ORROROO-PARACHILNA airborne magnetic and radiometric survey, South Australia. Rec. Aust. Bur. Mineral. Res., Geol. Geophys. 1965/229.(unpublished).

TOMLINSON, C.W. (1916) The origin of red beds. J. Geol. 24: 153-179.

TURNER, P. (1974) Lithostratigraphy and facies analysis of the Ringerike Group of the Oslo Region. Norges geol. Unders. 314: 101-131.

UDDEN, J.A. (1918) Fossil ice crystals. Bull. Univ. Texas 1821: 8pp.

VISSER, J.N.J. (1974) The Table Mountain Group: A study in the deposition of quartz arenites on a stable shelf. Trans. geol. Soc. S. Af. 77: 229-237.

VOS, R. and ERIKSSON, K.A. (1977) An embayment model for tidal and wave swash deposits occurring within a fluvially dominated middle Proterozoic sequence in South Africa. Sediment. Geol. 18: 161-173.

WADE, M. (1968) Preservation of soft-bodied animals in Precambrian sandstones at Ediacara, South Australia. Lethaia 1: 238-267.

WALKER, T.R. (1967) Formation of red beds in modern and ancient deposits. Bull. geol. Soc. Am. 78: 353-368.

WEBB, A.W. and HÖRR, G. (1978) The Rb-Sr age and petrology of a flow from the Beda Volcanics. Quart. geol. Notes, geol. Surv. S.Aust. 66: 10-13.

WEBB, B.P. and HORWITZ, R. (1959) Notes on the boundaries of the Marinoan Series of the Adelaide System. Aust. J.Sci.21: 188-189.

WHEELER, H.E. and QUINLAN, J.L. (1951) Precambrian sinuous mudcracks from Idaho and Montana. J. sedim. Pet. 21: 141.

WHITE, A.J.R., COMPSTON, W. and KLEEMAN, A.W. (1967) The Palmer Granite: A study of a granite within a regional metamorphic environment. J. Pet. 8: 29-50.

WHITNEY, M.I. (1978) The role of vorticity in developing lineation by wind erosion. Bull. geol. Soc. Am. 89: 1-18.

WILLIAMS, E. (1960) Intra-stratal flow and convolute folding. Geol. Mag. 97: 208-214.

WILLIAMS, H., TURNER, F.J. and GILBERT, C.M. (1954) Petrology. 406pp. Freeman: San Fransisco.

WOOD, J. (1973) Stratigraphy and depositional environments of the Upper Huronian rocks of the Rawhide Lake - Flack Lake area, Ontario. In: Young, G.M. (Ed.): Huronian stratigraphy and sedimentation. Spec. Paper geol. Assoc. Can 12: 73-95.

WOOLNOUGH, W.G. (1904) Petrographical notes on some South Australian quartzites, sandstones and related rocks. Trans. R. Soc. S.Aust. 28: 193-211.

WOPFNER, H., THOMSON, B.P. and LUDBROOK, N.H. (1969) Palaeozoic era. In: Parkin, L.W. (Ed.): Handbook of South Australian Geology. Geol. Surv. S.Aust.: 84-132.

WUNDERLICH, F.(1967) Die Entstehung von "convolute bedding" an Platenrändern. Senckenberg. leth. 48: 345-349.

WUNDERLICH, F. (1970) Genesis and environment of the Nellenköpfchenschichten (Lower Emsian; Rheinian Devonian) at locus typicus in comparison with modern coastal environments of the German Bay. J. sedim. Pet. 40: 102-130.

Key to Localities

A. Within Study Area.

ABC - ABC Range
AGH - Alligator Gorge: Hancock Lookout
AGM - Alligator Gorge: Mambray Creek
AGP - Alligator Gorge: Pine Track
AGW - Alligator Gorge: near Wilmington
AMC - Artimore Creek
ARK - "Arkaba" Station
ARR - Aroona "Ruins"
AUB - Aubrey Creek
BCC - Brachina Creek
BGL - Black Gap Lookout
BJR - Black Jack Range
BKG - Buckaringa Gorge
BNC - Bunyeroo Creek
BRG - Barunga Gap
BYR - Bunbinyunna Range (1 & 2)
CCR - Chace Range
CNC - Crow Nest Creek
DDR - Druid Range
DWS - Dawson
EBH - East of "Buckaringa" Homestead
ECR - East End of Chace Range
ELG - Eurilpa Gap
EPB - East of Point Bonney
GGC - Gorge Creek
HMG - Hanniman Gap
IGG - Ingram Gap
ILK - Ilka Creek
LCQ - Locheil Quarry
MCH - Marchant Hill
MDG - Middle Gorge
MKT - Moockra Tower
MLH - "Moralana" Homestead
MMC - Mernmerna Creek
MRC - Mary Creek
MTF - Mount Fergusson
MTG - Mount Grainger
NAR - North of Aroona "Ruins"
NBC - North of Brachina Creek
NBG - North of Bunyeroo Gorge (1 & 2)
NBR - Nector Brook Range
NTO - "Narinna" Turn-Off
OBM - Oraparinna Barytes Mine
ODW - Oodlawirra
PCG - Parachilna Gorge
PRL - Prelinna
PRP - Pichi Richi Pass
PRT - "Partacoona" Station
PTG - Petanna Gorge
RDR - Red Range
RMG - Richman Gap
RNP - "Rawnsley Park"
RNQ - Ridge North of Quorn
SAR - South of Aroona "Ruins"
SBC - South of Brachina Creek
SBG - South of Bunyeroo Gorge (1, 2 & 3)
SDC - Sacred Canyon
SNH - South of "Narinna" Homestead
SPG - South of Parachilna Gorge
SWG - South of Warren Gorge
SWH - South of Wonoka Hill
TDM - The Dome
TDP - Third Plain
UDR - Ulowdna Range
WCW - "Warcowie" Homestead
WKC - Waukarie Creek
WKG - "Warrakimbo Gorge"
WKH - Wonoka Hill
WMY - West of Mount Yappala
WNG - Warren Gorge
WPC - Wilpena Creek
WSF - Woolshed Flat
WWG - Wilkawillina Gorge

B. Outside Study Area.

CBC - Chambers Creek
HLC - Hallett Cove
MBR - Mount Bayley Range
MNR - Marino Rocks
OSB - O'Sullivan's Beach
PPG - Puttapa Gap
PTH - Patsy Hill
PTL - Point Lowly
SHC - South of Hallett Cove
SXB - Spinifex Bluff

Stratigraphic Section Localities are underlined.