GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

BULLETIN 128

GEOLOGY OF THE BANGEMALL GROUP

THE EVOLUTION OF AN INTRACRATONIC PROTEROZOIC BASIN

DEPARTMENT OF MINES
WESTERN AUSTRALIA

GEOLOGY OF THE BANGEMALL GROUP

FRONTISPIECE

Angular unconformity between the Top Camp Dolomite of the Bangemall Group and the underlying metamorphosed sedimentary rocks of the Ashburton Formation (Wyloo Group).

The view is to the southwest, at latitude 23° 45' 30'' longitude 117° 17' 15'', close to the Top Camp gold diggings. Here, H. P. Woodward in 1890, first recognized the unconformity and the Bangemall sequence as a comparatively young and separate entity.

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BY

P. C. MUHLING AND A. T. BRAKEL

AND

STROMATOLITES AND OTHER ORGANIC REMAINS IN THE BANGEMALL BASIN

BY

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FOREWORD

Systematic mapping, with a special emphasis on the Bangemall Basin as an entity, commenced in 1973 and concluded in 1977. A considerable amount of geological mapping peripheral to the basin had already been undertaken in the previous decade.

From the collective data of some twelve 1:250 000 geological sheets, there emerged a coherent picture of basin evolution that involved regional and temporal variations in sedimentary facies and tectonic style. This warranted collation into a bulletin which will complement other bulletins on important tectonic units of Western Australia.

The study makes an important contribution to the knowledge of the Precambrian evolution of Western Australia and, most importantly, will be of value in drawing comparisons with other intracratonic basins of similar facies and age that host important base-metal deposits. It is becoming increasingly apparent that the origin of such sedimentary mineral deposits is tied closely to the sedimentational and tectonic aspects of basin evolution, particularly in regard to basement interaction. It is therefore expected that this bulletin will be of value for the next phase of mineral exploration, which will rely less on surface mineral occurrences, and more on geological models.

A. F. Trendall
DIRECTOR

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SUMMARY

The Bangemall Basin is an intracratonic sedimentary basin (ca. 1 100 m.y. old) deposited on a rifted Proterozoic orogen and an unidentified platform, both of which lie between the cratonized Archaean Pilbara and Yilgarn Blocks.

Sedimentation began in the western part of the basin, which overlies the core of the orogen—the Gascoyne Province. Subsidence created horsts and grabens by reactivation of steeply dipping faults which had been the loci of earlier Proterozoic tectonic activity. Alluvial fan deposits flanking horst blocks are overlain by a transgressive marine sequence of pyritic black shale, sandstone, and stromatolitic dolomite. These were deposited in shelf, barrier-bar and lagoon environments. The older metamorphic rocks were the source of the terrigenous clastic sediment.

The transgressive sequence ended with deposition of the Discovery Chert, which covers an area of 38 000 km^2 and is the main stratigraphic marker in the western basin. Molds of gypsum crystals and traces of barite indicate that the chert was deposited in hypersaline water. Microtextures show the chert was deposited as a silica gel, which has subsequently undergone at least eighty per cent compaction. Macrolamination of carbonaceous organic detritus is preserved, but microscopic laminae have been modified by plastic-style deformation parallel to bedding.

A shallow-marine platform sequence overlies the Discovery Chert and represents shoals of carbonate interfingering with terrigenous shale and sandstone. An ensuing phase of subsidence is marked by turbidite sheets overlain by marine-shelf shale.

Sedimentation was controlled by the relative tectonic activity of basement blocks and has produced two major facies changes that define three facies: the western facies adjoins the northern facies, which deposited as a stromatolitic carbonate sequence in lagoon and tidal-flat environments on a stable basement formed by the Ashburton Fold Belt. The western facies also passes eastward on to a stable cratonic platform to form the eastern facies, which is a transgressive sequence of sandstone overlain by lutite deposited on a shallow-marine shelf. This latter facies change broadly corresponds to a major northwesterly trending zone of basement arches and faults—the Tangadee Lineament. Tectonic reactivation on an old zone of deformation in the basement along the southeast edge of the platform produced fining-upward clastic sequences interpreted as deltaic deposits.

The end of sedimentation in the basin is marked by regressive sequences: in the west, a lutite-sandstone-lutite sequence corresponding to lagoon, barrier-bar and shelf environments covers the successions. In the eastern facies, on the eastern edge of the platform, a sandy sequence deposited as subtidal tongues and shoals on a shallow tide-swept sandy shelf prograded westwards, covering the previous sequences. Much of this regressive sequence on the platform was deposited as a series of tidal current ridges, of which spectacular examples occur on the southern edge of the platform.

Dolerite sills of contintental tholeiitic composition crop out over about 143 000 km^2. The sills, injected prior to folding, are concentrated in zones of greatest basement instability, which lie in the western part of the basin and along the southeast edge of the platform in the east.

Felsic volcanic rocks appear to be related to zones of faulting and basement arches. Highly potassic rhyolite flows occur close to the Tangadee Lineament, while a felsic lava and dyke occur close to a northeasterly trending lineament near Mount Palgrave. A small basaltic flow in the eastern part of the basin is the only sign of mafic volcanicity. Scattered deposits of highly potassic sediments are believed to be related to distant felsic pyroclastic volcanism of unknown location.

Regional evidence indicates that the basement and the Bangemall Group deformed together. Structural provinces in the basin reflect the tectonic activity of basement blocks: a fold belt lies over much of the core of the older orogen and an undeformed platform cover overlies cratonic basement of unknown nature. The first deformation was by vertical movement of basement horsts and grabens to produce drape folds in the cover. Stretching of the sediments was achieved by bedding-plane slip and intrafolial folds in incompetent shale and chert, and by steeply dipping normal faults and spaced cleavage in brittle sandstone. A second deformation resulted in lateral shortening which was achieved by concentration of strain in ductile rocks in the basement which became tightly folded with the overlying cover. Normal faults forming boundaries of some basement blocks were reactivated, and became the site for reverse faulting, which with continued shortening, penetrated the cover. These faults indicate the mechanism of lateral shortening was by the Gascoyne Province moving towards the Pilbara Craton. Total shortening estimated from a cross section constructed in the western part of the basin is 13.4 km or 10 per cent of the original distance.

The Bangemall Basin contains small deposits of copper, lead, zinc, manganese, and phosphate. Many of these occurrences are epigenetic with a structural control. Zones of highly deformed Bangemall group adjacent to highly deformed basement have potential for this style of mineralization. The tectonic and sedimentation history provides some potential for sedimentary base-metal occurrences. Some stratabound mineralization occurs close to basin hinge lines, and basement highs have contributed to some base-metal occurrences.

CHAPTER 1

Introduction

SCOPE

The Bangemall Basin is a Middle Proterozoic (ca 1 100 m.y.), intracratonic sedimentary basin which occupies about 145 000 km^2. It unconformably overlies older Precambrian sedimentary basins and metamorphic complexes of the Western Australian Shield in the northwest of Western Australia (Fig. 1). This bulletin is a synthesis of the regional stratigraphy and structure of the Bangemall Group and summarizes the work of many geologists who have contributed to the 1:250 000 regional mapping programme within the basin (Fig. 2).

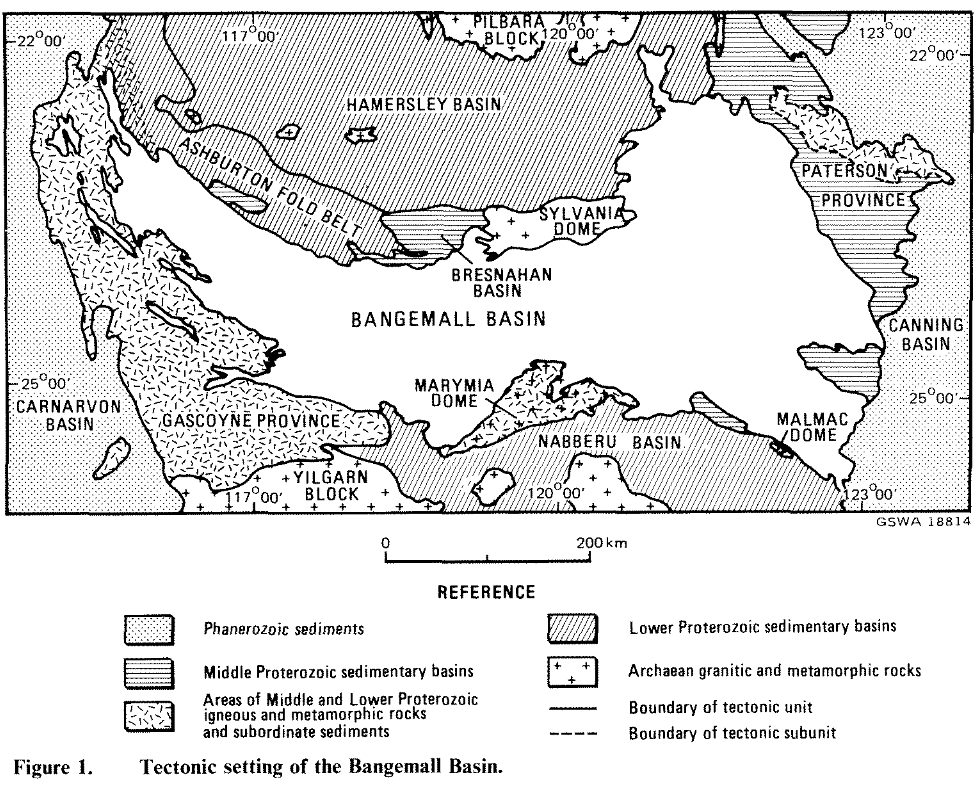

Figure 1. Tectonic setting of the Bangemall Basin.

As well as the general regional geology, three aspects have been emphasized.

1. The environment of deposition of the sediments.
2. The Discovery Chert—a remarkably persistent chert unit in the western part of the basin.

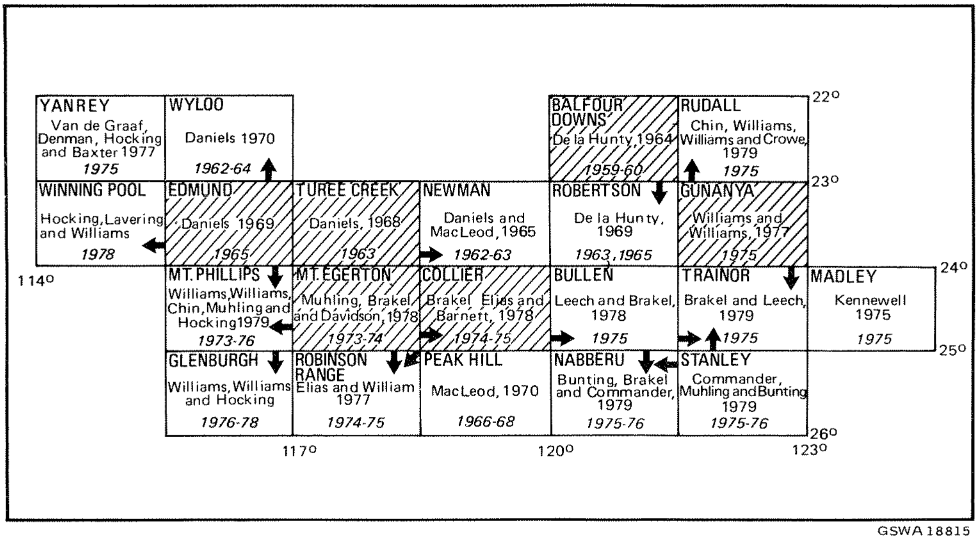

Figure 2. **Diagram of 1:250 000 map sheets covering the Bangemall Basin showing date of mapping (italic), and author(s) and date of publication (upright style) of the explanatory notes. The shaded sheets are those which resulted in innovations in stratigraphic nomenclature; the arrows show adoption of stratigraphic nomenclature from adjacent sheets.**

3. The relation between the sedimentation and structure of the Bangemall Basin and the basement.

LOCATION AND ACCESS

The Bangemall Basin occupies an arcuate area between latitudes 22° to 25°40′S and longitudes 115°30′ and 123°10′E. The Great Northern Highway passes through the central part of the basin, but the North West Coastal Highway lies to the west. Both of these highways are sealed all-weather roads, which are linked by subsidiary graded roads between pastoral stations. Such roads are impracticable during or immediately after heavy rain. Rocky ranges extending from Mount Minnie to Mulgul Station (Plate 1) form a natural barrier, over which access is difficult even by four-wheel-drive vehicle. Most of the region west of the Great Northern Highway is accessible by tracks that connect bores and wells or follow fence lines, but a few areas, such as parts of EDMUND* and TUREE CREEK, are still inaccessible.

Access to the eastern part of the basin is poor. A few station tracks and graded roads exist west of 120°E longitude and in the far southeast, but the remainder is uninhabited desert with widespread sandplain and dunes. A graded road, the Gun Barrel Highway, was constructed in the 1950s to link Carnegie Station on the eastern edge of the Bangemall Basin to Alice Springs in the Northern Territory. Lack of maintenance has rendered the road impassable to all but four-wheel-drive vehicles.

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*To avoid confusion with place names, 1:250 000 sheet names are written in full capitals throughout the bulletin.

There are no towns within the Bangemall Basin. The western part is served by the coastal towns of Carnarvon and Onslow and the small settlement of Gascoyne Junction on the main (unsealed) road linking Carnarvon to Meekatharra. The central part of the basin is served by the town of Meekatharra, which is on the Great Northern Highway. The iron-ore towns of Newman and Paraburdoo lie north of the basin margin. Wiluna is the nearest town to the eastern part of the basin.

Carnarvon, Paraburdoo, and Newman, have a daily jet service from Perth; Meekatharra has a light-aircraft service from Perth; and most pastoral stations have airstrips capable of receiving light planes. All towns and most stations have telephone or radio communication.

PREVIOUS INVESTIGATIONS

The first visitors to the area were exploring for new pastoral country, and most of them recorded natural features, including brief mention of rocks. The western part of the Bangemall Basin was investigated by Frank Gregory in 1858, who named Mount Augustus. In 1874, John Forrest collected rocks, which were later described by Huddleston (1883), from the headwaters of the Lyons River. Forrest also travelled through NABBERU and STANLEY and into South Australia. Ernest Giles, in 1876, in his expedition to the Overland Telegraph Line journeyed through MOUNT EGERTON to the Ashburton River, before turning east to pass through COLLIER, BULLEN and TRAINOR over a route about one degree of latitude further north than that of Forrest. The first expedition whose main purpose was the recording of scientific data was the Calvert Scientific Exploring Expedition of 1896 led by Lawrence Wells, who commented on sandstone and shale in parts of STANLEY and TRAINOR.

RECOGNITION OF THE BANGEMALL GROUP

The first recognition of the Bangemall sequence as a comparatively young and separate Precambrian entity was in 1890 by H. P. Woodward, the Government Geologist; his discoveries were summarized on the State geological map, which was published in 1894. The discovery of gold on the south side of the Ashburton valley in 1890 prompted Woodward to include the diggings in his trip that year. Here he noted for the first time the unconformity between the "clay slates" of the Ashburton valley and the gently dipping dolomite of the overlying Bangemall sequence. He also correctly identified the outcrop width of the basin by correlating the dolomite with the rocks extending south-west to the Barlee Range, where they meet "... crystalline schists and granite ...". The southern boundary of the sedimentary rocks was placed at the Teano Range (on MOUNT EGERTON) by Woodward, and all the tightly folded rocks south of the Lyons River, which are now know to be part of the Bangemall Group, were regarded as basement. Except for minor refinements, this mapping of the southern boundary persisted until the 1966 edition of the State Geological Map.

The boundaries of the sequence described by Woodward were refined and extended by Maitland (1909) during an inspection of the Bangemall gold-mining centre. Maitland correctly believed that the unconformity at the foot of Mount Phillips was the base of the sequence at Bangemall, which he called "Bangemall

Beds" or "Bangemall Series" and which he correlated with the succession to the northeast as far as the Ashburton gold diggings. By using the composite term "Bangemall-Nullagine Series" he also correlated the sequence with rocks now known to belong to the Hamersley Basin. Talbot (1910) recognized the southeastern boundary of the Bangemall Basin, at the Carnarvon Range, as well as that between the Parker Range (Bangemall Group) and the "Lee Steere Range slates" (Earaheedy Group) as an unconformity.

By 1919, the broad features of the southern boundary (except the Teano Range—Mount Labouchere zone) were outlined, but the Bangemall rocks were not recognized at this stage as a distinct unit and merely included in the vast area of "Nullagine Series" extending north to the Pilbara.

The Sylvania Dome, as well as the regional structure of the "Nullagine Series" on COLLIER and NEWMAN were mapped by Talbot (1920), who later extended the southeastern boundary further east (Talbot, 1926). Johnson (1950) recognized that the rocks of the Sawback Ranges were "Nullagine Series" and that they were younger than the group of metasediments further east, which are now assigned to the Padbury and Glengarry Groups. However, his mapping did not extend far on to MOUNT EGERTON, and the Sawback Range sequence was regarded as a southern outlier of the "Nullagine Series".

Regional mapping of the Mount Bruce Supergroup by the Geological Survey in 1960-63 led to subdivision and abandonment of the "Nullagine Series" and established the basis for the northern limits of the Bangemall Basin; this is summarized by Halligan and Daniels (1964). Subsequent work has defined the southern boundary of the basin (Teano Range—Mount Labouchere area) and the eastern boundary (Paterson Province and Officer Basin).

STRATIGRAPHIC SUBDIVISION

The name, Bangemall Group, was first applied by Halligan and Daniels (1964), who also set up three constituent formations on TUREE CREEK. De la Hunty (1964) defined the "Manganese Group" which he correlated with the Bangemall Group. Further stratigraphic subdivision was undertaken by Daniels (1969) in the western half of the basin, and by Brakel and Muhling (1976) and Williams and others (1976) in the eastern half. To this basic framework, other formations and members have been added by various workers from time to time. MacLeod (1970) described lithological units of the Bangemall Basin on PEAK HILL and established that they were unconformable on the Mount Labouchere sequence, but he also correlated with the Bangemall Group south of the Marymia Dome rocks that are now assigned to the Glengarry Group. In this bulletin the "Manganese Group" of de la Hunty (1964) is reclassified as a subgroup of the Bangemall Group, and five new subgroups (Edmund, Mucalana, Collier, Diebil and Kahrban Subgroups) are defined.

Regional appraisals of structure and stratigraphy of the Bangemall Basin have been produced by Daniels (1966b, 1975) and Brakel and Muhling (1976), all of whom recognized the basement control of folding in the western part of the

Bangemall Basin, and by Gee (1979b). A few aspects of the basin have been reported in comparative detail. Palaeocurrents on EDMUND and breccias were described by Daniels (1965; 1966). Manganese nodules in the "Manganese Group" were report by de la Hunty (1966). Stromatolites from the Bangemall Basin were described by Glaessner and others (1969). Marshall (1968), in an unpublished thesis, described the regional setting of the western part of the basin as well as macrofossils, microfossils, and mineralization.

There are few mineral deposits known in the Bangemall Basin. Gee (1975) provides the most recent summary on manganese, lode copper, and syngenetic sulphide mineralization. The Bangemall gold-mining centre was documented by Maitland (1909). Copper mineralization has been described by Talbot (1914), Low (1963), and more recently by Marston (1980). Small lead, copper and zinc occurrences are noted by Blockley (1971) who also gives a summary of the syngenetic sulphide mineralization described by Marshall (1968). De la Hunty (1963) has written on manganese deposits in both the "Manganese Group" and the Bangemall Basin.

GEOPHYSICS

Bouguer anomaly maps at scale 1:250 000 and 1:2 500 000 are available for the entire Bangemall Basin. Contoured aeromagnetic maps (scale 1:250 000) are available only for the southern edge of the basin covering the ROBINSON RANGE, PEAK HILL, NABBERU and STANLEY 1:250 000 sheets. Both gravity and aeromagnetic maps have been produced by the Bureau of Mineral Resources.

PHYSIOGRAPHY

The Bangemall Basin straddles a major continental divide, defined by Mulcahy and Bettenay (1972) as the boundary between external and internal drainage. Each of these drainage regions also has distinct physiography. Most of the Bangemall Basin lies more than 500 m above sea level (Fig. 3). However, dissection in the area of external drainage has produced relief more rugged than that in the area of internal drainage to the east.

EXTERNAL DRAINAGE REGION

The external drainage region covers most of the basin west of longitude 120° plus the western fringe of the basin in ROBERTSON and BALFOUR DOWNS. Here, the physical features can be described as uplands separated by wide colluvial slopes and plains and major drainages.

Uplands comprise dissected strike ridges with narrow intervening valleys and a dendritic drainage pattern. Rocks resistant to weathering, such as sandstone and chert, form the tops to prominent ridges and cuestas, whereas shale, siltstone, and, to a lesser degree carbonate, underlie valleys and scree slopes. Major hills in these zones are Mount Egerton (994 m above sea level) and Mount Palgrave (704 m) which rise about 400-500 m above the adjacent broad valley floors. Large

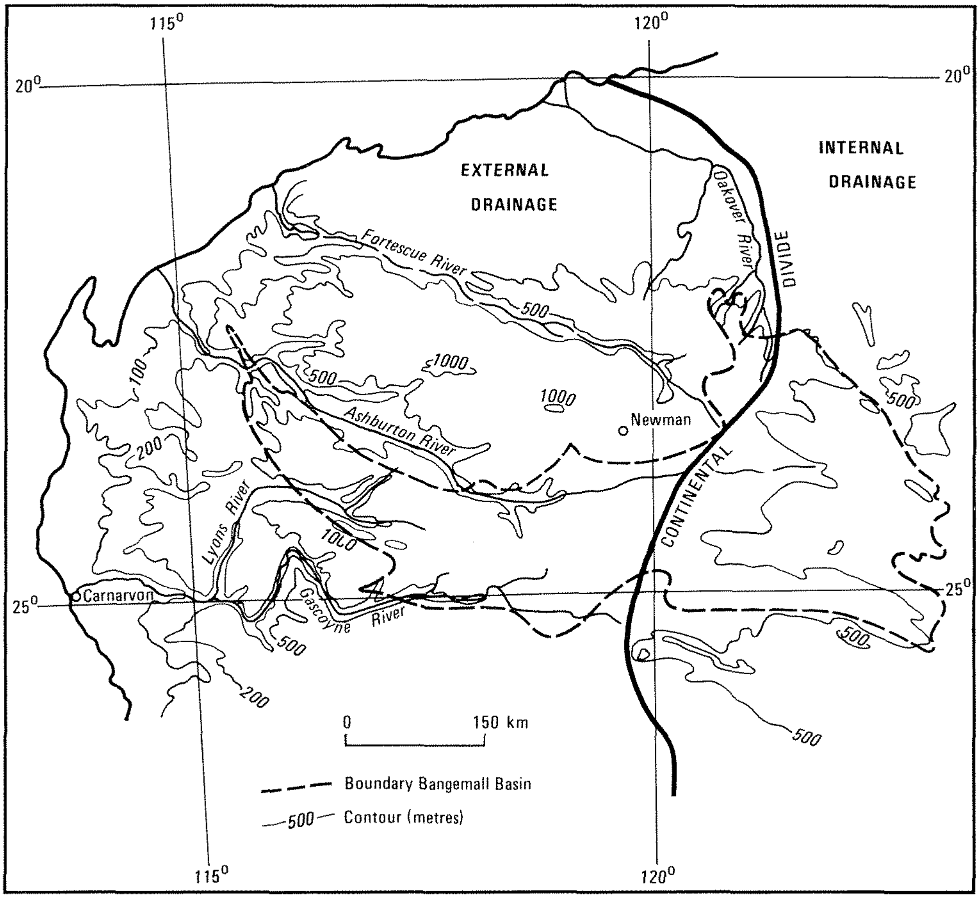

Figure 3. Physiographic setting of Bangemall Basin.

monadnocks such as Mount Augustus (1 106 m) and Mount Phillips (780 m) rise to 600 m above the major drainage channels. Although many of the rocks are weathered, and the shales have been bleached and silicified, there are only local areas of laterite and silcrete on drainage divides.

Colluvial slopes and plains fringe the major drainages which lie between upland zones. They range from scree accumulations at their highest points downslope to broad, gently sloping, sheetwash plains engraved with small creeks. The common deposit of the colluvial slopes is hardpan—an indurated colluvium derived by erosion from laterite profiles.

The major drainages, which flow into the Indian Ocean after heavy rain, are the Oakover, Fortescue, Ashburton and Gascoyne. The Oakover is a mature river, although there is evidence of rejuvenation since the Tertiary (de la Hunty, 1964). The Fortescue River is also mature and displays only local headward erosion. The Ashburton River, although mature, shows such signs of rejuvenation as a sandy course incised into hardpan, and benches and buttes of calcrete and hardpan in its upper reaches. The Gascoyne River and its tributary, the Lyons, appear to be more mature than the Ashburton, and although their courses are incised in hardpan, they also dissipate, in places, into numerous distributaries and swamps. Calcrete in the Gascoyne drainage forms low mounds and does not show the same degree of dissection that is evident in the Ashburton.

INTERNAL DRAINAGE REGION

The internal drainage region comprises parts of the Canning and Eucla Drainage Divisions of Mulcahy and Bettenay (1972). The relief is low and the topography is more gentle than that in the external drainage area, due to lack of dissection on active drainage lines. The internal drainage area is characterized by extensive, gently undulating sandplain, overlying a relict duricrusted landscape (van de Graaff and others, 1977) on which the weathered rock is exposed only where dissection has formed breakaways at the head of local drainages.

Low hills, cuestas and mesas project about 100 m above the sandplain. The largest are the Carnarvon Range (Mount Methwin, 908 m above sea level) and the Durba Hills. Most of these hills are monadnocks of comparatively unweathered rock, and stand up to 500 m above the relict weathering surface of Tertiary to Early Cretaceous age (van de Graaff and others, 1977). Much of the eastern area is covered with sand dunes. These are mainly simple, longitudinal seif dunes, although chain and net styles similar to those described by Crowe (1975) are also present. The dunes, which are fixed by vegetation, are up to 20 km long and 3-35 m high. Their trends range from northwesterly in the north of the area to southwesterly in the south and are oriented parallel to the prevailing wind directions. However, more complex wind patterns near hills and depressions result in net-style dunes and short longitudinal dunes of variable trends.

Most drainages are short and ephemeral, and empty into either salt lakes or sandplain. Many are in palaeovalleys of the old land surface. The largest are Savory Creek, which drains easterly into Lake Disappointment, and Kahrban Creek which empties into Lake Burnside (Oneahibunga). Savory Creek is a well-defined, incised, and braided channel with freshwater pools west of McFadden Ranges, but to the east it becomes a series of saline flats which drain into Lake Disappointment. Kahrban Creek occupies a wide, very gently sloping palaeovalley with extensive calcrete deposits and discontinuous, incised channels which become a single watercourse near Lake Burnside. Many lakes are slowly migrating east, as indicated by their profiles which show small escarpments and exposed bedrock on the eastern edges, and wind-blown gypsiferous sand on the western edges.

The sandplain, being a modified relict landscape, contains palaeodrainages defined by wide valleys, calcrete deposits, and salt lakes (van de Graaff and others, 1977). These are part of a very large system which emptied into the Indian Ocean to the northwest in Early Tertiary times, and some palaeodrainages may date back to the Permian. Flow ceased due to climatic changes in Early Tertiary times (van de Graaff and others, 1977).

CLIMATE

The climate is semi-arid to arid, and is characterized by a low and unreliable rainfall, high evaporation rates, mild temperatures during winter, and hot summers.

The rainfall ranges from 200 to 250 mm per year, and the eastern desert areas receive slightly less. The wettest six-monthly period is usually January to June. Most rain falls in summer from monsoonal depressions, though the southwest part may receive further rain from southerly low-pressure systems during winter. There are also occasional tropical cyclones in summer which may bring heavy rain. Evaporation from a free-water surface ranges from 2 400 to 2 800 mm per year.

The hottest months are December and January, and the coldest are July and August, when occasional frost may be experienced.

VEGETATION

The region is entirely within the Eremaean Botanical Province, which lies between the summer-rainfall area of the north and winter-rainfall area of the southwest. The vegetation is classed as woodland, high and low shrubland, and hummock grassland (Aplin, 1977). In general, tall trees are absent. The western part of the basin is characterized by an open woodland of mulga trees *(Acacia)*, high shrubs of *Cassia* and *Acacia* species, and low shrubs such as poverty bush *(Eremophila)* as well as grassland. Widespread but uncommon shrubs are sandalwood *(Santalum spicatum)* and quandong *(Santalum acuminatum)*. Tall trees (30 m) are found along the major watercourses: the common species are river red gum *(Eucalyptus camaldulensis)*, coolibah *(Eucalyptus microtheca)* and varieties of paperbark trees e.g. cadjeput (*Melaleuca leucadendron*). In the eastern parts of the region, there is extensive sandplain; spinifex (*Triodia* and *Plechtrachne*) and related grasses are the dominant vegetation, but scattered mulga trees also occur. Bloodwood (*Eucalyptus dichromophloia*) occurs in low-lying areas where there is some moisture and is also a common species on the sand dunes. Mallee eucalypts (e.g. *E. Kingsmillii*) up to about 5 m high are widespread, but not common. Native cypress (*Callitris columellaris*) is common in some rocky hills such as the Carnarvon Ranges. Salt lakes and drainages support plants adapted to saline conditions; saltbush (*Atriplex*) and samphire (*Arthrocnemum*) are the most common types. The vegetation of the eastern part of the basin has been mapped and described by Beard (1974).

TERMINOLOGY

A summary of the classifications used in this bulletin appear below.

BEDDING THICKNESS

The thickness of bedding units is classified as follows:

Bedding term	*Splitting term*	*Average thickness (mm)*
Laminated	Fissile	< 10
Very thinly bedded	Flaggy	10-30
Thinly bedded	Flaggy	30-100
Medium bedded	Slabby	100-300
Thickly bedded	Blocky	300-1000
Very thickly bedded	Massive	1000 <

CROSS-BEDDING

The size of cross-bedding units has been classified according to Conybeare and Crook (1968) and are:

Descriptive term	*Approximate thickness*
Small scale	< 5 cm thick
Medium scale	5 cm-2 m
Large scale	2 m-8 m
Very large scale..	8 m <

Types of cross-beds have been classified according to the scheme of Allen (1963, Fig. 4).

CLASSIFICATION OF SEDIMENTS

The size of grains (Table 1) in clastic sediments is classified, with only slight modification, according to the scheme of Wentworth (1922) quoted in Pettijohn (1975). The classification of sedimentary clastic rocks is as set out by Williams and others (1954). The proportion of clasts necessary for a rock to be a conglomerate follows the classification of Pettijohn (1975). Table 1 shows the classification used in the Bangemall Basin. The terms *sandstone, arenite,* and *wacke* are all intended to have the distinctions shown in Table 1, and the reader should note that arenite and wacke are different types of sandstone which is itself used as a general term.

TABLE 1. GRAIN SIZE AND CLASSIFICATION OF SEDIMENTARY CLASTIC ROCKS

<table>
<tr><th>Size (mm)</th><th>Clasts</th><th colspan="5">Rock name</th></tr>
<tr><td>256</td><td>Boulder</td><td rowspan="4">CONGLOMERATE (a)</td><td colspan="2">Boulder conglomerate</td><td colspan="2">boulders 10% (b)</td></tr>
<tr><td>64</td><td>Cobble</td><td colspan="2">Cobble conglomerate</td><td colspan="2">cobbles 10% (b)
boulders 10% (b)</td></tr>
<tr><td>4</td><td>Pebble</td><td colspan="2">Pebble conglomerate</td><td colspan="2">pebbles 10% (b)
cobbles 10% (b)
boulders 10% (b)</td></tr>
<tr><td>2</td><td>Granule</td><td colspan="2">Granule conglomerate</td><td colspan="2">granules 90% (b)</td></tr>
<tr><td rowspan="2">½
¼
1/16</td><td rowspan="2">Coarse sand
Medium sand
Fine sand</td><td rowspan="2">SANDSTONE</td><td colspan="2">< 10% matrix</td><td colspan="2">> 10% matrix</td></tr>
<tr><td>Arenite</td><td>coarse
medium
fine</td><td>Wacke</td><td>coarse
medium
fine</td></tr>
<tr><td rowspan="2">1/256</td><td rowspan="2">Silt</td><td rowspan="3">LUTITE</td><td colspan="2">Fissile</td><td colspan="2">Non-fissile</td></tr>
<tr><td colspan="2" rowspan="2">Shale</td><td rowspan="2">MUD-STONE</td><td>Siltstone</td></tr>
<tr><td></td><td>Clay</td><td>Claystone</td></tr>
</table>

(a) Total rudite > 25% of rock (b) Named size as percentage of rudite

CHAPTER 2

Regional Setting

TECTONIC SETTING

The region appears to have become an unstable zone within Archaean terrain by the beginning of the Proterozoic Era about 2 500 m.y. ago. The Hamersley Basin (2 500-2 200 m.y.) formed south of the present Pilbara Block on stable crust and was subsequently subjected to mild deformation. The locus of later sedimentation (2 000-1 700 m.y. ago) migrated southwards to the Ashburton Trough and into the region between the Pilbara and Yilgarn cratonic blocks, and also migrated eastwards to the Paterson Province. All these regions experienced episodes of igneous, metamorphic, and tectonic activity about (1 800-1 600 m.y. ago) which increased in intensity towards the southwest, thus forming the metamorphic complexes in the Gascoyne Province. The Ashburton Trough was deformed to form the Ashburton Fold Belt. This tectonism declined in severity eastwards. Another Proterozoic basin, the Nabberu Basin (2 000-1 700 m.y.) developed along the northern margin of the Yilgarn Block. Thick, shelf and trough sequences (Earaheedy and Glengarry Groups) were deposited in this basin, the northern parts of which were later deformed by folding and thrusting in the Stanley Fold Belt (Bunting and others, 1977).

Following a period of erosion, the Bresnahan and Mount Minnie Groups were deposited as alluvial-fan and braided-stream deposits in isolated basins on the fold belt that evolved from the Ashburton Trough. These terrestrial deposits were probably a response to block faulting in the underlying metasediments. The precise age of these sediments is unknown, but they are unconformably overlain by the Bangemall Group, and therefore are taken to be part of the basement of the Bangemall Basin. The Bangemall Group thus came to be deposited as a younger cover on a complex basement; in the west it straddles the transition from Gascoyne Province mobile belt to the less metamorphosed Ashburton Fold Belt; and in the east, it laps across various tectonic units, where much of it appears to be underlain by stable cratonic basement. The nature of this very stable basement in the far east is uncertain.

BASIN LIMITS

The Bangemall Basin is delineated by the present-day extent of the Bangemall Group, whose limits are mainly governed by erosion. The extent of the original sedimentary basin is conjectural, but it probably did not reach far beyond the present boundaries. It is not known how far the unit extends beneath the Permian and Cretaceous sediments in the Officer Basin. The occurrence of the Cornelia Inlier and outcrops of Yeneena Group in this eastern area suggest that the

Bangemall Basin is shallowing to the east and that the eastern limit may lie not far to the east, below the cover. The Bangemall Group therefore may not physically connect to the Proterozoic sequences of similar age in the Amadeus Basin and Musgrave Block.

Isolated synclinal keels of Bangemall Group such as the Mangaroon and Ti Tree Synclines, and the smaller occurrences near Trilbar and south of the Murchison River on ROBINSON RANGE (Elias and Williams, 1977), occur within the Gascoyne Province and are stratigraphically similar to the Bangemall Group, and are regarded as part of the Bangemall Basin. The most southwesterly of such outcrops occurs 80 km away from the present margin of the basin, at Hectors Bore (on GLENBURGH) (Williams and others, 1980). Further south, and along the western margin of the Yilgarn Block are several disconnected infaulted outcrop areas which may be correlatives of the Bangemall Group. These are outliers at Yalbra and Coordewandy Hills (GLENBURGH), the Badgeradda Group and Nilling Beds, the Dudawa Beds, and possibly the Wenmillia Formation and Yandanooka Group. Although these may once have been physically continuous with the Bangemall Group, it is more likely they were deposited within separate basins.

The Uaroo Group on YANREY (van de Graaff and others, 1977) has a succession of basal sandstone, followed by dolomite, sandstone, shale with chert, and dolomite with shale. The sequence is broadly similar to the lower Bangemall Group, and probably marks an extension of the basin at least 60 km west of its present limits.

Specification of the northeastern extent of the Bangemall Basin requires discussion of some stratigraphic problems, some of which have been resolved, but others of which are still outstanding. The "Manganese Group" named on BALFOUR DOWNS by de la Hunty (1964) consisted of four formations (Table 2), whose type localities have not been examined since the original mapping by de la Hunty (1964), and hence, equivalence to the Bangemall Group has not been proved. The names are retained pending remapping of BALFOUR DOWNS. Future work will probably show that these formations belong to the Bangemall Group, and they are treated as such in Chapter 3.

The "Bocrabee Sandstone" on NULLAGINE (Hickman, 1978) was mapped as Bangemall Group, but is now believed to be part of the Yeneena Group (Williams and others, 1976). The "Bocrabee Sandstone" on northeast BALFOUR DOWNS although reported as unconformably overlying "Manganese Group" actually overlies the Fortescue Group of the Hamersley Basin. However, further south, the "Bocrabee Sandstone" does correspond to Bangemall sediments. Thus, as mapped, the "Bocrabee Sandstone" corresponds to large areas of sandstone in both the Bangemall and Yeneena Groups. In view of the uncertainty as to which rock unit the name is meant to apply, and in the absence of a formal definition, it is recommended that the name "Bocrabee Sandstone" be discarded.

The "Waltha Woora Beds," mapped on NULLAGINE as Bangemall Group (Hickman, 1978), are also now believed to be part of the Yeneena Group, whereas the "Waltha Woora Beds" on ROBERTSON are now known to be part of the Bangemall Group.

It is now apparent that the northeastern extent of the Bangemall Group is more restricted than has been depicted on earlier geological maps. Evidence of syndepositional gravity faulting along parts of the present boundary on ROBERTSON and BALFOUR DOWNS, together with the absence of outliers to the north, suggest that the original depositional extent of the basin corresponds closely to its present outcrop extent.

In summary, the present northern and eastern boundaries may correspond, in a regional sense, with the present outcrop boundaries, but the original basin may have extended well to the south. The data from lithofacies, presented later in this bulletin, indicate an open ocean to the southwest and therefore support the above conclusion.

EVIDENCE OF AGE

A poor Rb-Sr isochron of about 1 080 m.y. was obtained by Compston and Arriens (1968) from felsic rocks, believed to be a dyke and related flow, at Mount Palgrave, and the same workers obtained the same age from shale in the Curran Formation, which lies about 600 m above the sub-Bangemall unconformity. Gee and others (1976) report an isochron of 1 098 ± 42 m.y. from rhyolite in the Coobarra Formation, 60 m above the basal unconformity, but the rock has an unusually high K_2O content, indicating alteration after eruption, so that the isotopic ratios date either early devitrification of the volcanic glass or early burial metamorphism. Although the reliability of each of these dates can be questioned, each is independent of the others, and when taken together they are remarkably consistent. The age of the Bangemall Group can thus be stated to be about 1 100 m.y.

Palaeontological evidence is confined to stromatolites. Walter (1972) notes that a stromatolite which occurs in the Irregully Formation at the base of the Group has a Middle Riphean age (approximately 1 350 to 950 m.y.). However, another stromatolite from the Skates Hills Formation was previously thought to range from about 900 m.y. to Early Cambrian (Preiss, 1976).

CHAPTER 3

Stratigraphy

REGIONAL FACIES DISTRIBUTION

Three major facies, the western, northern, and eastern, each with distinctive lithostratigraphic assemblages, are recognized in the Bangemall Basin; (Brakel and Muhling, 1976). The southeastern region could constitute a fourth facies. Their distribution is shown in Figure 4. The western facies has the thickest and most varied succession. The transition between the western and northern facies is the result of both the rapid lensing out of some units from the west, and changes in lithology along strike of others. The change from the western to the eastern facies is due to the easterly lensing out of dolomite, chert and sandstone units and the persistence of shale and siltstone beds. The eastern and northern facies are juxtaposed by faults, except east of Bulloo Downs, where two shale formations are laterally equivalent, and distinguishing between them becomes arbitrary. Although it is now desirable that most of the formations should be organized into formally defined stratigraphic subgroups, the regional facies nomenclature is used for more general reference, especially as their boundaries do not coincide exactly with the new subgroups, and they reflect sedimentological changes. The distribution of

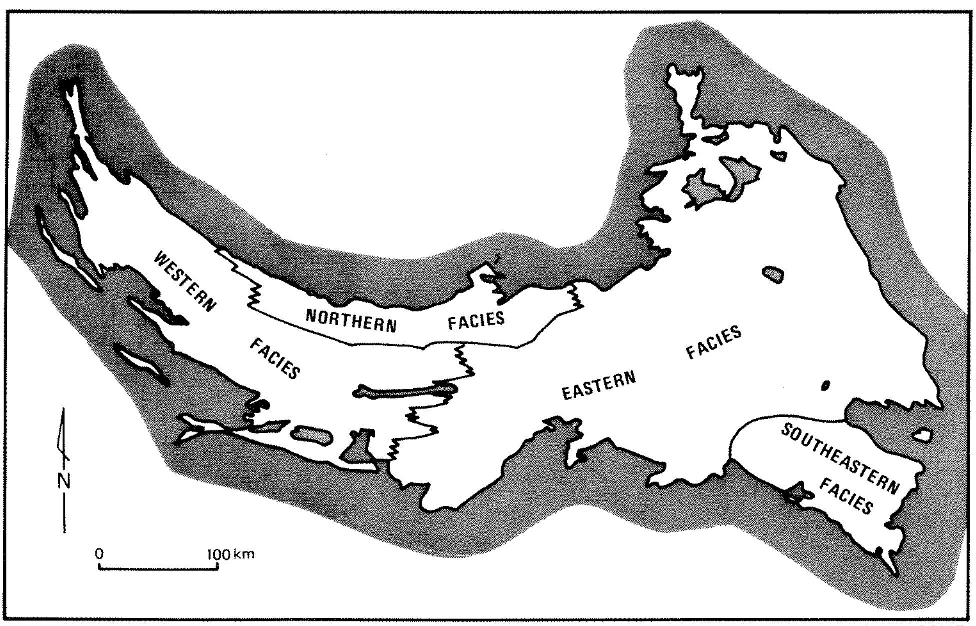

Figure 4. **Distribution of facies in the Bangemall Basin.**

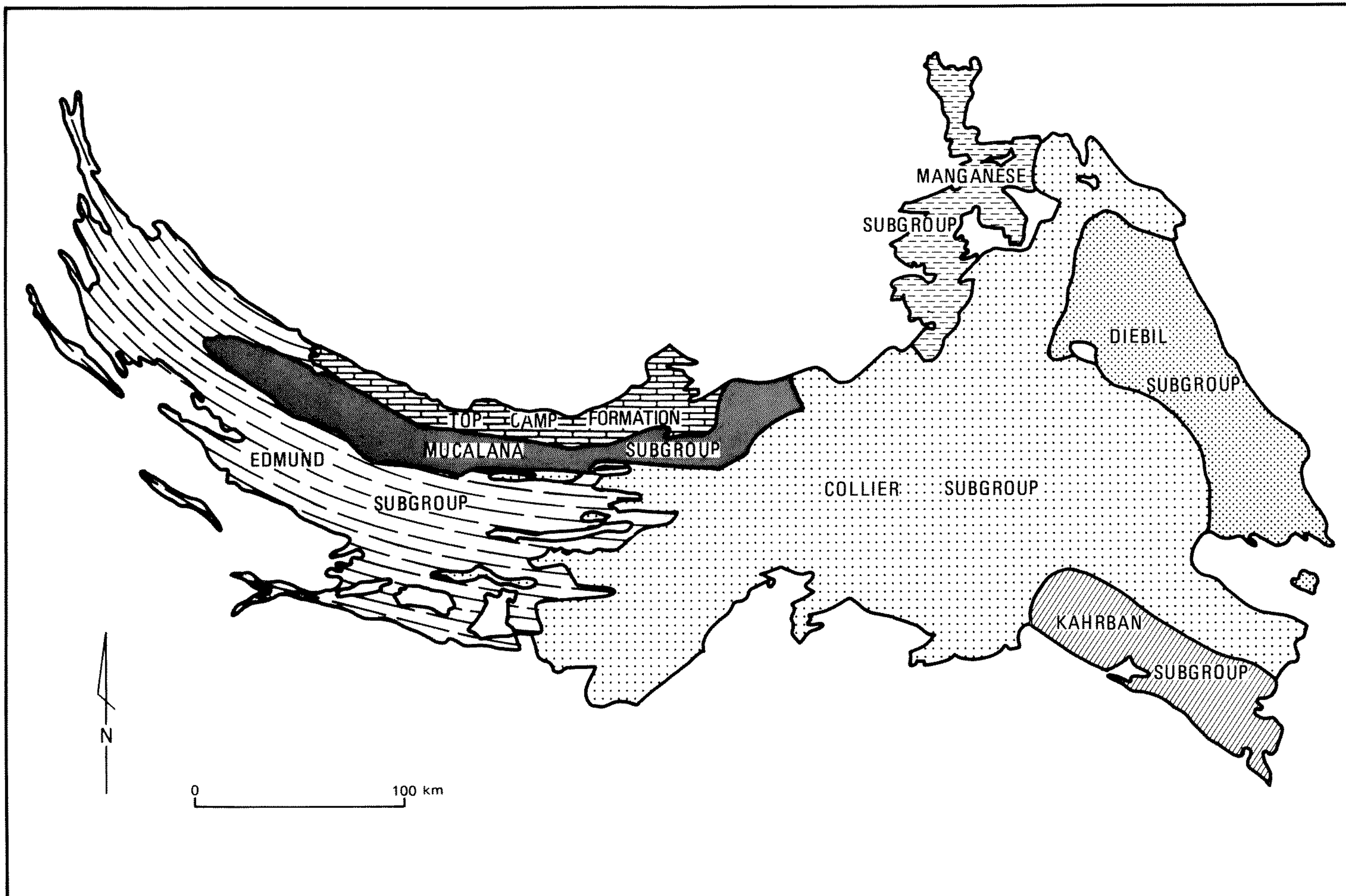

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Figure 5. Distribution of subgroups in the Bangemall Basin.

subgroups is shown in Figure 5, and the correlation of formations and subgroups is given in Table 2.

BANGEMALL GROUP

Derivation of Name: Bangemall Mining Centre (lat. 24°13′S, long. 116°28′E).

Constituent subgroups and ungrouped formations: Edmund, Mucalana, Collier, Diebil, Kahrban, and Manganese Subgroups, and Top Camp Formation; the Kahrban Subgroup is included subject to future verification that no major unconformity exists between it and the overlying Calyie Sandstone.

Thickness: Variable. Exceeds 8 km on EDMUND-MOUNT EGERTON.

Stratigraphic relationships: Occupies the entire Bangemall Basin and unconformably overlies older tectonic units; it is overlain by the Durba Sandstone and the Permian Paterson Formation near the eastern margin of the basin, and by Cainozoic superficial cover.

EDMUND SUBGROUP

Derivation of name: Edmund River.

Constituent formations: Mount Augustus Sandstone, Tringadee Formation, Coobarra Formation, Irregully Formation, Kiangi Creek Formation, Jillawarra Formation, Discovery Chert, Devil Creek Formation, Nanular Sandstone, Ullawarra Formation, Curran Formation, Coodardoo Formation.

Thickness: Variable; exceeds 4 000 m on EDMUND.

Distribution: Constitutes most of the western facies (Brakel and Muhling, 1976); it thus occupies the southwestern portion of the Bangemall Basin, from the Tangadee-Mingah Springs area to the Parry Range district at the northwestern extremity of the basin, and includes the northwestern basin margin as far as Turee Creek.

Stratigraphic Relationships: The Edmund Subgroup is the older subgroup in the western half of the basin and transgresses unconformably the various older tectonic units of the basement. It is conformably overlain by the Mucalana Subgroup. There is a physical eastwards continuity with the Collier Subgroup, the transition between the two being marked by the lensing out of the dolomitic, cherty and arenaceous formations of the Edmund Subgroup, so that the lutaceous formations of the latter come together as the Backdoor Formation of the eastern facies.

The Edmund Subgroup is noted for the interfingering and gradational relationships between most of its formations, its lithological diversity, and the presence of the unusual Discovery Chert.

MOUNT AUGUSTUS SANDSTONE

Derivation of name: Mount Augustus (lat. 24°19′30″S, long. 116°15′30″E).

Type area: Mount Augustus.

Lithology: Coarse sandstone, commonly with megaclasts and minor conglomerate lenses.

TABLE 2. CORRELATION OF FORMATIONS AND SUBGROUPS WITHIN THE BANGEMALL BASIN

NORTHERN FACIES		WESTERN FACIES		COLLIER, BULLEN AND PEAK HILL SHEETS (EASTERN FACIES)		ROBERTSON AND BALFOUR DOWNS SHEETS (EASTERN FACIES)		SOUTHEASTERN FACIES		EASTERN DESERT REGION (EASTERN FACIES)	
											DURBA SANDSTONE
MUCALANA SUBGROUP	KURABUKA FORMATION	MUCALANA SUBGROUP	KURABUKA FORMATION	COLLIER SUBGROUP	ILGARARI FORMATION						
	MT. VERNON SANDSTONE		MT. VERNON SANDSTONE		CALYIE SANDSTONE	COLLIER SUBGP	CALYIE SANDSTONE	COLLIER SUBGP	CALYIE SANDSTONE	DIEBIL SUBGROUP	McFADDEN SANDSTONE
	FORDS CREEK SHALE		FORDS CREEK SHALE		BACKDOOR FORMATION	MANGANESE SUBGROUP	NOREENA SHALE	KAHRBAN SUBGROUP	MARLOOYANOO FORMATION		SKATES HILLS FORMATION
	TOP CAMP FORMATION	EDMUND SUBGROUP	COODARDOO FORMATION								
			CURRAN FORMATION								
			ULLAWARRA FORMATION								
			NANULAR SANDSTONE								
			DEVIL CREEK FORMATION								
			DISCOVERY CHERT								
			JILLAWARRA FORMATION				BALFOUR SHALE		COONABILDIE FORMATION		
			KIANGI CREEK FORMATION		WONYULGUNNA SANDSTONE		BEE HILL SANDSTONE				
							COONDOON CONGLOMERATE				
			IRREGULLY FORMATION								
			MT. AUGUSTUS SANDSTONE, TRINGADEE AND COOBARRA FORMATIONS								

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Thickness: Estimated to be at least 600 m in the type area. Usually ranges from 200 to 400 m along the Cobra Synclinorium.

Stratigraphic Relations: Unconformably overlies schist and granitic rocks at the northwestern end of Mount Augustus. The contact has a relief of 60 m and is in places vertical, possibly because of later faulting. The top of the formation is not present here. It is overlain by poorly outcropping, well-cleaved siltstone of the Irregully Formation along the southwestern side of the Cobra Synclinorium, and the two formations appear to be conformable.

Description: The typical rock in the type area is coarse-grained quartz arenite, commonly containing scattered granules, pebbles, cobbles, and boulders exceeding 0.6 m across. Locally, the megaclasts are sufficiently numerous to form pockets and thin beds of granule conglomerate and coarser sandy conglomerate, which, though forming a minor portion of the succession, occur at any level. Vein quartz is by far the dominant clast type, but schist, phyllite and a quartz-tourmaline rock are also widespread. The clasts are mostly subangular to subrounded. There is no basal conglomerate, but pebbles of the basement rocks occur in the basal sandstone. The rocks are white when fresh, usually thickly bedded, and in places, dark heavy-mineral grains outline the cross-bedding foresets. Tourmaline and lithic grains are minor components of the rocks, and feldspar is rare. Lutite interbeds are absent from the sequence.

Although medium-grained sandstone is sparse in the type area, it comprises the largest part of the succession along the Cobra Synclinorium. Coarse- and fine-grained arenite, as beds and laminations, are also present. Isolated hills south west of Mount Augustus contain interbeds of white, fine-grained wacke and siltstone. Locally, the sandstone is arkose derived from underlying granitic basement.

TRINGADEE FORMATION

Derivation: Tringadee Bore (lat. 24°56′45″S, long. 117°36′30″E).

Type section: Northern limb of Sullivan Anticline from lat. 24°52′30″S, long. 117°23′30″E to lat. 24°51′45″S, long. 117°23′00″E. The base of this section is faulted against pre-Bangemall schistose sandstone, but only a small basal interval is believed to be missing. The unconformity over gneiss, and the basal beds, are exposed near Mount Remarkable (lat. 24°56′S, long. 117°23′E).

Lithology: Dominantly sandstone, mostly pebbly, with interbeds of conglomerate, siltstone, and near the top, dolomitic rocks.

Thickness: Difficult to obtain in the Sullivan Anticline, because of faulting at the base of the type section, and elsewhere there is mesoscale folding or lack of outcrop; a thickness of 1 100 m is estimated from air photographs in the type section. The formation is lenticular in gross geometry.

Stratigraphic relationships: Rests unconformably on rocks of the Gascoyne Province, and passes conformably upwards into the Irregully Formation. The top is defined by the base of the first persistent dolomite member. It is abrupt in the type section, but transitional in some other places. The Tringadee Formation is discontinuous on the basal unconformity. It is probably equivalent to the Mount Augustus Sandstone, although it is unlikely there was ever any physical continuity between them.

Description: The Tringadee Formation consists dominantly of interbedded and interlaminated, medium- and coarse-grained arenites, with subordinate granule and pebble conglomerate. Most of the arenites are pebbly; they vary from well to poorly sorted and contain small amounts of feldspar, which in a few places comprise 25 per cent of the rock. Clasts in the conglomerate include vein quartz, quartzite, and quartz-tourmaline rock. The formation shows an upward decrease in average grain size, although coarser grained lithologies are interbedded even at the higher levels. Fine-grained arenite and siltstone beds and laminations increase in frequency towards the top. Medium-scale planar cross-bedding is moderately common.

Beds and lentils of dolomitic rocks appear near the top, and consist of sandstone with dolomite cement, conglomerate with quartz, quartzite, and phyllite clasts in a sand and dolomite matrix, and dolomitic siltstone. The upper contact in the type section is abrupt from quartz arenite and siltstone to fine-grained dolomite of the Irregully Formation, but in other places is transitional through a sequence of interbedded dolomitic and non-dolomitic clastic sediments.

COOBARRA FORMATION

Derivation: Coobarra Creek, a major tributary of the Ethel River between Mulgul and Tangadee homesteads.

Type area: The area between lat. 24°35′S, long. 118°46′E, 25 km southwest of Tangadee homestead, and lat. 24°39′S, long. 118°50′E. No satisfactory type section can be specified because of discontinuous outcrop.

Lithology: Sandstone, mostly coarse-grained; conglomerate; lesser siltstone and shale, and subordinate rhyolite.

Thickness: Unknown; at least 500 m is estimated in the type area.

Age and stratigraphic relationships: Nonconformable on an inlier of sheared granitic rock. Overlain conformably, at least in part along its southern margin, by the Kiangi Creek Formation, where the boundary between the two is taken as the base of a 200 m-thick siltstone-shale unit which appears to persist for at least 10 km, or, where this is absent, the base of the lowest dolomite. Farther east, the top contact is obscured, and it is not known if the Kiangi Creek Formation persists, or if it lenses out to leave the Backdoor Formation resting on the Coobarra Formation. The unit is thought to be laterally equivalent westwards to the lower part of the Kiangi Creek Formation.

The unit is included with the Edmund rather than the Collier Subgroup because it is overlain by the Kiangi Creek Formation, and because it is a coarse, clastic, lenticular, terrestrial deposit on the basal unconformity analogous to, though probably somewhat younger than, the Mount Augustus Sandstone and Tringadee Formation. The basal Wonyulgunna Sandstone of the Collier Subgroup is by contrast a shoreline deposit, and has only a very small proportion of conglomerate.

Rhyolite 60 m above the base of the formation gave a Rb-Sr isochron of 1098 ± 42 m.y. (Gee and others, 1976).

Description: Coarse-grained arenite is by far the most prevalent lithology, but interbedded with it are medium- and fine-grained arenites, which tend to be more common in the top half of the unit. Also present towards the top are thin bands of wacke, siltstone and shale. The lutites, which also form thick local lenses, are white, earthy rocks containing little or no mica. The arenite near the base is commonly pebbly and contains up to 5 per cent feldspar and 1 per cent dark grains, such as magnetite. The upper arenite has more feldspar and lithic grains, which together constitute up to 25 per cent of the rock. These lithic grains consist of chert and red jaspilite. In general, the arenites are moderately to well sorted, massively bedded or laminated, and contain occasional layers of intraclast. Cross-bedded medium-scale units, usually of planar tangential type, form a minor proportion of the sequence.

The most striking rock type, however, is conglomerate. In the type area it makes up about half of the lower sequence, and contains well-rounded pebbles, cobbles, and boulders of vein quartz, together with accessory clasts of coarse-grained siliceous siltstone, quartz arenite, jaspilite-bearing conglomerate, and, rarely, quartz-tourmaline rock. Some of these rocks resemble those in the Padbury and Glengarry Groups (Elias and Williams, 1977; Gee, 1979a) on the southern margin of the Bangemall Basin. Some clasts are angular to subrounded, suggesting a local derivation such as from a subsurface continuation of the basement inlier north of Mulgul. The average proportion of conglomerate in the formation is estimated at less than 20 per cent, and in most places it is much less. The upper half of the sequence contains only thin units of pebble and granule conglomerate within the arenite.

Six rhyolite bodies occur about 60 m above the base of the formation at lat. 24°37′S, long. 118°47′E, and have been described by Gee and others (1976). They form a line of plugs that were extruded as small viscous domes and were later subjected to erosion and covered by conglomerate.

IRREGULLY FORMATION

Derivation of name: Irregully Creek.

Type section: Irregully Creek from lat. 23°19′S, long. 116°31′E to lat. 23°31′30″S, long. 116°35′00″E.

Lithology: Dolomite; shale and siltstone; lesser amounts of sandstone, conglomerate, breccia, and chert.

Thickness: 1 300 m estimated from air photographs in the type section, but it varies considerably over its extent.

Stratigraphic relationships: Lies on the basal unconformity of the Bangemall Group, except where the Mount Augustus Sandstone and Tringadee Formation are present conformably below it. It is usually conformable beneath the Kiangi Creek Formation, although in the type section, the latter is represented by only a few metres of quartz arenite and shale. In areas where the Kiangi Creek Formation is missing from the sequence, for example near Mount Candolle, the Irregully Formation is overlain by the Jillawarra Formation. The Irregully and Kiangi Creek Formations are lateral facies equivalents, in whole or in part depending on the locale, though the Irregully Formation is the older in a vertical profile. The Irregully - Kiangi Creek boundary is regarded as the base of the lowest laterally persistent sandstone unit. Sandstone lenses lower in the sequence are thus members of the Irregully Formation, whereas higher dolomites are considered members of the Kiangi Creek Formation.

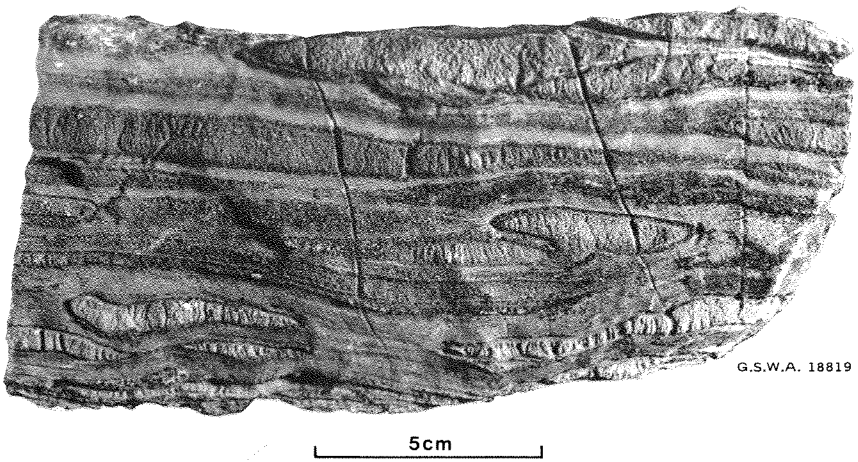

Figure 6. **Pods and layers of limestone (bold relief) within dolomite (subdued relief). Specimen 50485, from Irregully Formation 9.5 km west-northwest of Staten Hill.**

Named members: The Yilgatherra, Wongida Dolomite, Gooragoora Sandstone, Wannery, Chubilyer, Weewoddie Dolomite, Yeelingee, Warrada Dolomite and Revels Corner Sandstone Members are named in the sequence in Wyloo, but cannot be traced far due to their lenticular nature.

Description: The Irregully Formation consists mainly of dolomite, shale, siltstone, and variable amounts of sandstone, conglomerate, breccia and chert. The dolomite is grey, dark brown, yellow brown, buff or white on exterior surfaces, but is pink, or dark or light grey internally. It is well bedded, finely laminated, microcrystalline and usually unfossiliferous; laminations are conspicuous on weathered surfaces. Abundant cubes of goethite and hematite, ranging from less than 1 mm to 35 mm, on edge are presumably pseudomorphs after pyrite. Clusters of cubes forming nodules 50 mm in diameter have also been found. Limestone is rare, but has been recorded as thin bands and pods, in which internal laminations are discontinuous with those in the enclosing dolomite, probably because of compaction effects (Fig. 6.).

Small columnar stromatolites are known from two localities at the unconformity on MOUNT EGERTON, and two on ROBINSON RANGE, but otherwise stromatolites have only been observed north of the latitude of Mount Augustus. They are common in this last region, where they are prolific in some beds (*see* Appendix A). Algal lamination is also common in the type section. The forms *Baicalia capricornia* and *Conophyton garganicum australe* have been identified (Walter, 1972, and Grey, 1977), as well as an unidentified form (Figs 116A, B) previously unrecorded from the Bangemall Basin (Grey, 1977).

Grey shale and siltstone, which form a substantial part of the sequency in many areas, include micaceous varieties and those with a carbonate cement. Locally, such as near Mount Candolle, lutite predominates. Sandstone occurs in Irregully Creek, usually as thin beds of quartz arenite, but some thicker beds exceeding 1 m in thickness contain phases of granule and pebble conglomerate with clasts of quartz and dolomite. Arenite lenses of variable thickness are present elsewhere, and may have a dolomitic cement. Twenty kilometres west of Mulgul, an impersistent arenite occurs near a basement high; it is the only equivalent of the Kiangi Creek Formation.

Some interbeds of breccia, up to 0.3 m thick, consist of tabular, fine-grained dolomite clasts in a dolomite or dolarenite matrix. A much thicker occurrence has also been noted (Daniels, 1965), which contains large, angular clasts of algal dolomite closely resembling the dolomite in a bed some distance below it, except that the clasts are silicified. Outcrop conditions made it impossible to decide whether this breccia is an interbed or a product of much later, perhaps Tertiary, silicification. Large areas of similar breccia have been described by Daniels (1965), who considered them to be syndepositional, and therefore Proterozoic in age. Breccia 60 m thick is developed at Mount Price, where it overlies Bangemall rocks, and where it contains dolomite blocks in its lower part, and current-bedded sandstone blocks higher up. However, field relations with the Bangemall Group are difficult to interpret, and, on many hills, there is considerable doubt as to whether one is dealing with a primary breccia or a Cainozoic deposit.

Daniels (1970) subdivided the formation on WYLOO into the nine members described below. Similar subdivision has not been attempted elsewhere.

YILGATHERRA MEMBER

The lowest member of the Irregully Formation is the Yilgatherra Member, which consists of sandstone and quartz-pebble conglomerate, of which, the latter is frequently cemented by iron oxides.

The Yilgatherra Member varies considerably in thickness. It is approximately 300 m thick on the western side of the Parry Range, but is absent on the eastern side. Over much of the northwest part of the Bangemall Group, this basal unit is very thin, generally less than 30 m. It is frequently present only as a 0.3 m-thick grit or pebbly dolomite.

WONGIDA DOLOMITE MEMBER

The Wongida Dolomite Member either conformably overlies the Yilgatherra Member, or otherwise forms the base of the Irregully Formation in the northwestern part of the Bangemall Basin. It has a thickness of approximately 640 m.

The basal 150 m is a massive white-weathering dolomite, which underlies well-bedded, white, cream, mauve, and grey dolomite with some edgewise conglomerate, sandy dolomite, dolomitic sandstone and chert. The upper part carries abundant stromatolites, which are variably silicified.

GOORAGOORA SANDSTONE MEMBER

In the Mount Florry region, the Gooragoora Sandstone Member, which overlies the Wongida Dolomite Member, is a thin-bedded, cross-bedded sandstone that frequently carries small iron-oxide pseudomorphs after pyrite. Farther north, in the southern part of the Parry Range, the member consists of mudstone, shale, thin sandstone bands, and occasional thin dolomite horizons. Farther north still, the member becomes arenaceous and consists of quartz arenite and silty sandstone, interbedded with shale and variegated shaly sandstone.

WANNERY, CHUBILYER, WEEWODDIE AND YEELINGEE MEMBERS

Immediately north of Gooragoora Pool on the Wannery Creek in the southeast part of WYLOO, the Gooragoora Sandstone Member is conformably overlain by the Wannery Member, which in this locality is entirely shale.

To the southeast near Mount Florry, the Wannery Member thickens considerably and three equivalent members can be identified. These are, from the base upward, the Chubilyer Member, the Weewoddie Dolomite and the Yeelingee Member.

The Chubilyer Member consists of approximately 180 m of shale and mudstone with thin sandstone bands. This body of dolomite is lens shaped and is not of large extent.

The Weewoddie Dolomite Member is approximately 150 m thick and consists dominantly of black-weathering, dark blue-grey dolomite and a few thin sandstone layers. This body of dolomite is lens shaped and of no great extent.

The Yeelingee Member is about 210 m thick and consists of finely banded mudstone, sandstone, and thin dolomite bands. It overlies the Weewoddie Dolomite Member near Mount Florry, but rests on the Chubilyer Member a little to the northwest.

The Wannery Member and its lateral equivalents appear to be missing from the section in the southeastern part of the Parry Range.

WARRADA DOLOMITE MEMBER

North of Ashburton River, in the southern part of the Parry Range, the Gooragoora Sandstone Member is overlain by the 640 m-thick Warrada Dolomite Member. Its lower half comprises massive, cream-weathering, light-grey dolomite carrying sand grains and bands of calcarenite. The upper portion is a well-bedded dolomite containing bands of edgewise conglomerate and thin silicified horizons which become more common up the sequence. The unit is not developed south of the Ashburton River.

REVELS CORNER SANDSTONE MEMBER

This member overlies the Warrada Dolomite Member in the southern part of the Parry Range and is the youngest unit exposed in that region. It has not been identified south of the Ashburton River. The member is at least 330 m thick and consists of brown-weathering, thin-bedded sandstone, some bands of which carry clay pellets. On weathering the sandstone becomes porous. It may be equivalent to the Kiangi Creek Formation.

KIANGI CREEK FORMATION

Derivation of name: Kiangi Creek, 25 km west-northwest of Ullawarra homestead.

Type area: Headwaters of Kiangi Creek. No suitable type section has been identified but good exposures in a west-flowing tributary 8 km west of Mount Palgrave can be used as a reference section. This area can be reached by a mineral-exploration track. The base of the unit is not exposed and may be faulted out.

Lithology: Quartz arenite, shale, siltstone, and subordinate dolomite, conglomerate, and in the type area, two cherty members.

Thickness: At least 700 m in the type area. Varies greatly over its extent—up to 1 800 m has been estimated from air photos in the Mount Egerton district and Daniels (1969) gives 2 000 m as a possible thickness on EDMUND.

Stratigraphic relationships: Conformably overlies the Irregully Formation, except on the southern part of EDMUND where it lies directly on the basal unconformity. Near the Egerton mining centre, it coalesces with the Tringadee Formation because the Irregully Formation lenses out. In the Coobarra Creek district it overlies the Coobarra Formation. It underlies the Jillawarra Formation conformably. The unit is diachronous and interfingers with both vertically adjacent formations, or grades into them along strike from sandstone, through sandy dolomite or wacke, to dolomite or siltstone. In some areas the formation is entirely replaced laterally by the Irregully or Jillawarra Formations due to sandstone lensing out of the succession. Alternatively, the Kiangi Creek Formation may substitute for the others. The upper boundary is regarded as the top of the highest major sandstone member of the interval. In the type area, this boundary is a gradation from the normal medium-grained arenite through 0.2 m of fine-grained arenite, and then 0.1 m of coarse-grained quartz siltstone, into softer, fine siltstone and shale. Some 1 m-thick beds of arenite occur several metres higher up, but these are included with the Jillawarra Formation, because they occur in a dominantly lutite interval. The boundary with underlying formations is similarly defined and is discussed under the respective formations.

The formation thins to a few metres in Irregully Creek and lenses out to the southeast.

Isotopic Rb-Sr age determinations on a felsic rock near Mount Palgrave, probably an intrusive, gave a date of 1 008 ± 80 m.y. (Compston and Arriens, 1968).

Named members: Marshall (1968) subdivided the sequence in the type area into five members, the highest of which is now equated with the Jillawarra Formation. As the others have not been published and have not been examined individually by us, it is inappropriate to formalize them here.

The Glen Ross Shale Member has been named in the Mount Vernon Syncline (Muhling and others, 1976).

Description: Quartz arenite is the characteristic lithology of the Kiangi Creek Formation, but siltstone and shale are also important and in places may be more abundant than sandstone. Differential erosion of these rock types means that a sequence may at first sight appear to be entirely arenite but, on more detailed inspection, may be found to comprise over 50 per cent lutite. There are many areas, however, such as the Teano Range, where lutite is less than 5 per cent of the formation. Typical arenites are fine or medium grained, well sorted and contain little feldspar, generally less than 10 per cent but in rare cases more than 20 per cent. Coarse-grained varieties are common, and the sorting also varies. Wackes are very subordinate. Some rocks have a lithic and dark grain component (up to 5 per cent in exceptional cases) including tourmaline, an ubiquitous accessory. Beds containing platy molds of intraclasts of former lutite fragments are few. The rock is usually hard, white, and siliceous, resembling metaquartzite, but this is because of secondary surficial silicification. In dissected areas, such as the headwaters of Glen Ross Creek, the unit is soft and clayey. Bedding planes are irregular and frequently a metre or more apart, and the beds appear massive rather than laminated. Though most beds do not show cross-bedding, it is locally abundant. Scattered quartz pebbles and granules, and lenses of conglomerate, are widespread in some sandstone beds. The poorly outcropping shale and siltstone of the formation are light grey to brown and micaceous. Five kilometres northwest of Nicholl (Bamboo) Spring (COLLIER) is an unusual sequence of fine-grained, laminated to massive, fissile to medium-bedded, greenish to pinkish-grey arenite and wacke, containing variable amounts of mica, feldspar, lithic grains, and dolomitic cement.

Dolomite members are lithologically similar to those of other formations, and may also include dolomitic sandstone, sandy dolomite, and dolomitic shale. The only member known to contain stromatolites (*Baicalia capricornia*), is about 15 km north of Woodlands homestead. The largest areas with interbedded dolomite are the Mangaroon Syncline, the area west and east of Mount Egerton, and the area between Glen Ross Creek and the Ethel River.

In the type area there are two cherty members which consist of fine-grained quartz-feldspar-sericite rock.

GLEN ROSS SHALE MEMBER

This member is named after Glen Ross Creek, and crops out along the margins of the Mount Vernon Syncline. The upper portion is best seen in the northern limb, in the gorge of Glen Ross Creek and adjacent gullies, where the top contact with the overlying quartz arenite is sharp. The lower portion crops out only along the southern limb, where east of the creek it can be seen to grade downwards into a fine-grained wacke-arenite sequence similar to that 5 km northwest of Nicholl Spring. The estimated thickness is 125 m.

Most outctop consists of light-brown, weathered siltstone, shale and cherty beds, but in the gorge of Glen Ross Creek the true nature of the rocks is revealed as unsilicified, dark-grey, laminated siltstone and shale. Diamond-drill cores obtained in the area by Westfield Minerals NL show that the fresh rock at depth is black and carbonaceous. Overhangs in cliffs are commonly stained by malachite and white efflorescence of pickeringite (Simpson, 1923), alum, and halite, or yellow copiapite. Some lentils of fine-grained, laminated dolomite occur in the northwest of the syncline.

JILLAWARRA FORMATION

Derivation of name: Jillawarra Bore (lat. 24°40′45″S long. 118°24′15″E).

Type section: From lat. 24°40′15″S, long. 118°24′30″E southwards along the line of streams to the top of the formation at Jillawarra Bore. The base of the formation is faulted out in this area, but it may be inspected 4 km southeast of Mount Egerton, in small exposures amid scree. It is also well exposed at lat. 23°22′45″S long. 115°54′30″E, in the other type area Kiangi Creek Formation.

Lithology: Dominantly shale and siltstone, with subordinate claystone, chert, dolomite and sandstone.

Thickness: 1 020 m measured in the type section, which is incomplete; 600 m is estimated from air photographs in the Hells Doorway Syncline, 4 km southeast of Mount Egerton; 60 m was measured in the Discovery Creek district (lat. 23°12′15″S, long. 115°49′00″E), about 100 m estimated in the Barlee Range; and about 150 m estimated 8 km west of Mount Palgrave.

Stratigraphic relationships: The Kiangi Creek Formation and Discovery Chert respectively underlie and overlie the unit conformably. It is laterally equivalent to the Kiangi Creek in part, and even replaces it completely in some areas such as Mount Candolle. Eastwards, it passes into the Backdoor Formation of the Collier Subgroup, where the Discovery Chert cuts out. On EDMUND the unit has been informally referred to as the 'Prospect Shale', but it is much thinner there than in the type area, and was not designated a separate formation by Daniels (1966b). Subsequent checking found it present even as far as the northwest corner of EDMUND, pointing to its existence everywhere beneath the Discovery Chert, although in Irregully Creek it is represented by only about 2 m of shale. The basal contact is defined as the top of the highest major underlying sandstone bed, while the top boundary is that level in a transition zone where chert becomes the dominant lithology. Lenses of sandstone just above the fault in the type section suggest that the contact with the Kiangi Creek Formation was not far below that level.

Description: The formation consists dominantly of grey, white, brown, and black, silty shale, silstone and claystone, together with minor chert, dolomite and sandstone. The lutites usually contain fine detrital muscovite and, rarely, biotite. In the less dissected areas, lutite is commonly chert-like in outcrop, probably because of surface silicification of a rock already high in primary silica. Interbedded kaolinitic claystone in the type section is especially common in the lower portion.

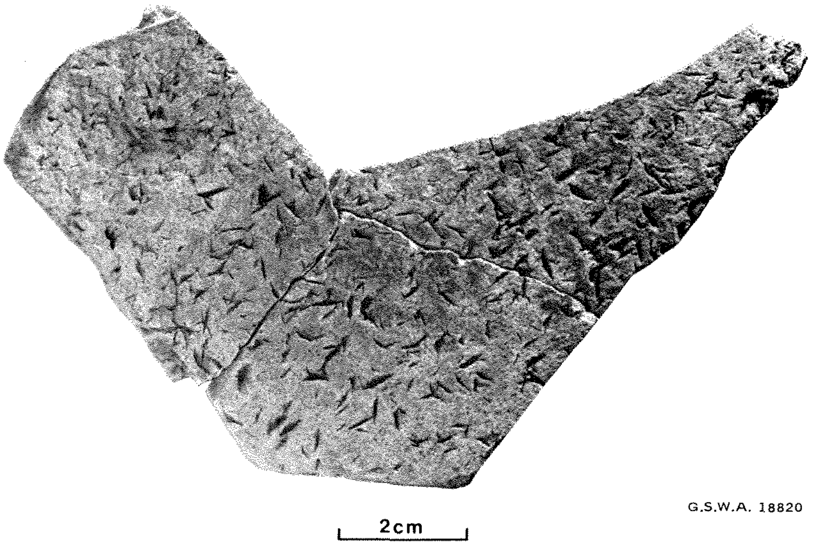

Figure 7. Molds of ?gypsum crystals on bedding surface of siliceous shale, Jillawarra Formation.

Fresh rock from drill cores is usually black, carbonaceous, and pyritic. The carbon content, while generally less than 5 per cent, can be as high as 10 per cent. Cubic crystal molds, up to 25 mm on edge but usually less than 10 mm, are plentiful in some beds. Many molds contain powdery iron oxides, showing that they are after pyrite. Smaller, elongate crystal molds filled with white clay are locally abundant, for example towards the top of the formation 41 km north of Mulgul homestead near Artesian Bore. Many such molds are rhomboid or biconvex in shape, and when found in surface outcrop contain kaolin and alunite, but in fresh rock consist of pyrite or pyrrhotite. The sulphide may have replaced original gypsum, as some shapes (Fig. 7) are strongly suggestive of gypsum (Carozzi, 1960, Fig. 65, p.338).

The most extensive chert member has a strike length of at least 25 km, and occurs near the top of the formation in the Lyons River Anticline. It is light coloured and is probably a silicified shale. Two others, lying at different levels in the sequence on the northern side of the Hells Doorway Syncline, are indistinguishable from the younger Discovery Chert. Only three chert bands, 0.25 to 0.75 m thick, occur in the type section, in the upper portion of the sequence. A shale bed containing numerous lentils of chert up to 30 mm long is a feature of the formation on ROBINSON RANGE and near Mount Deverell.

Sandstone members range from fine-grained wacke to medium and coarse-grained arenite, and are usually beds less than 1 m thick. The most prominent is a thick quartz arenite wedge close to the top of the formation in the Lyons River Anticline. Dolomite occurs as rare 10 mm-thick bands, and in the Woodlands-Mulgul region as thick lenses.

DISCOVERY CHERT

Derivation of name: First published by Daniels (1966b) and presumably named after Discovery Creek, 35 km northwest of Ullawarra homestead.

Type section: The Discovery Creek region is unsuitable for a type section because of strong cleavage which obscures the typical features of the unit. The type section chosen instead is the excellent exposure in Devil Creek (lat. 23°25′S, long. 116°23′E), 29 km east-northeast of Ullawarra. This is reached by taking the four-wheel-drive track via Doolgarie Creek and Blue Billy Bore, driving overland southeasterly to Devil Creek, and then upstream.

Lithology: Massive, laminated or fissile chert, with localized subordinate shale.

Thickness: 51 m in type section. Usually in the range 65 to 75 m and rarely outside the range 50 to 80 m.

Stratigraphic relationships: Conformable between the underlying Jillawarra Formation and the overlying Devil Creek Formation, or where these units are absent, between the Kiangi Creek and Ullawarra Formations respectively. The Discovery Chert does not interfinger with adjacent units. In Devil Creek, the top has a 0.9 m-thick transition zone of interbanded chert, dolomitic chert and dolomite. Elsewhere, the top zone is a mixture of chert and siltstone or shale, and the basal transition zone is similar. The boundaries of the formation are set at the levels in these zones were chert becomes the dominant lithology, and can be picked to within 10 mm where the outcrop permits the hardness of each band to be tested.

On weathered surfaces, however, this contact is not conspicuous. The chert merges eastwards with the Backdoor Formation of the Collier Subgroup by grading from chert to siliceous shale. Its transition into the Top Camp Formation has not been examined.

Description: The Discovery Chert is a distinctive, remarkably persistent unit which forms the best marker horizon in the Bangemall Group. Its most striking lithology is black, massive chert, which can appear homogeneous or have diffuse, light-coloured laminae that are planar, wavy or contorted. Colour variations include grey, white and maroon. A streaky texture visible with a hand lens is characteristic. Cut-offs of one group of laminae by others are common in some beds. Porous bands, probably representing weathered pyritic layers may be present. Possible gypsum molds and pseudomorphs also occur in several localities, but appear to be most abundant on EDMUND, where Marshall (1968) notes they are common towards the upper and lower contacts. Bedding is generally 10 mm to 0.5 m thick, and irregular swells and depressions form hummocky surfaces. Shale may be interbedded with chert. In many localities in the central part of the basin, the more massive black chert is underlain and overlain by well-bedded, lighter coloured chert, which is fissile and splintery in places. A more detailed description of the Discovery Chert is given in Chapter 5.

DEVIL CREEK FORMATION

Derivation of name: Devil Creek, 29 km east-northeast of Ullawarra homestead.

Type section: Devil Creek (lat. 23°25′S long. 116°23′E) starting where the Discovery Chert type section ends.

Lithology: Dolomite and shale, with occasional siltstone fine-grained sandstone, breccia and chert.

Thickness: 31 m measured in type section. Thicknesses of 365 m (Daniels, 1969) and > 800 m (Muhling and others, 1976) occur elsewhere.

Stratigraphic relationships: Overlies the Discovery Chert and is in turn overlain by the Ullawarra Formation. The base is transitional with the Discovery Chert, and is taken at the level where chert ceases to be the dominant rock type. The upper contact is taken as the top of the youngest major dolomite member, or the level where the proportion of dolomite falls below 10 per cent in a thinly bedded sequence. However, this is not always easy to define in an area such as the Range Creek Syncline, where, not only is there a substantial thickness of shale before more dolomite appears in quantity, but the upper dolomite level passes into shale along strike and the change is not photo-interpretable. The boundary then becomes arbitrary, and it is more convenient in practice to include the whole dolomite-bearing sequence in the Devil Creek Formation. This is one instance of the interfingering and lateral gradation with the Ullawarra Formation. Extreme instances of this variation occur 14 km northwest of Talga Pool (EDMUND) and west of Mount Sanford (MOUNT EGERTON), where no dolomite is developed in the sequence, and the Devil Creek Formation is therefore absent, having been substituted laterally by the Ullawarra. Southeast of Irregully Creek, as the three immediately underlying formations seem to cut out, the Devil Creek Formation amalgamates with the Irregully Formation to form the Top Camp Formation, and there is only equivalent to part of the Pingandy Dolomite Member.

Description: The formation has a distinctive striped photo-pattern which makes it a good marker in regional mapping, but it is not a very resistant unit, and, in many areas such as the Cobra and Mount Vernon Synclines, it outcrops poorly on the floors of valleys.

It is predominantly a dolomite and silty shale sequence, with minor siltstone, fine-grained sandstone, breccia and chert. The dolomite is thinly to thickly bedded and similar to the finely laminated, microcrystalline dolostone of the Irregully Formation, although recrystallization has coarsened some grains to 1 mm in size. Cross-bedding, while uncommon, is abundant locally, and consists of low-angle small- to medium-scale troughs and planar forsets, as well as climbing ripples. Erosion scours up to 0.1 m deep are present in places. Rarely, platy mudclasts and oolite bands up to 10 mm thick exist in some beds. Dolarenites, dolorudites and dololutites are very minor constituents. The rudaceous variety typically consists of subangular to subrounded slabs of fine-grained, laminated dolostone, up to 0.3 m long, in a coarse-grained dolomite-sand matrix cemented by carbonate. Dolarenite is similar, but variable in grain size, and without the megaclasts. It consists of both packstone and grainstone in Dunham's (1962) classification. Dolomite breccias occur in the area of the type section, but have rarely been found elsewhere. Four separate interbeds, 0.2 to 1.6 m thick, lie in the top half of the Devil Creek section, and are composed of slabs and fragments (up to 0.5 m long) of fine, laminated dolomite in a similar dolomite matrix. The matrix may also contain warped laminated bands. One of these breccia beds has a sharp, erosionally scoured base, while others have undefined, gradational boundaries. The coarsest and thickest breccia we are aware of contains boulders, at the base of the formation in Irregully Creek. Daniels (1966b) has recorded the only known stromatolitic beds, from the same area.

The shale, siltstone, and fine-grained sandstone beds are white, yellow, brown, pale green, purplish brown, and locally have alternating maroon and cream laminations. Small load casts on bedding surfaces are widespread. Some of these rocks have a dolomitic cement, and contain layers of dolomite a few millimetres thick. The proportion of shale to dolomite is laterally and vertically variable, and the dolomite itself may form discrete lenses.

Hematite cubes (up to 5 mm on edge) after pyrite are wide-spread, especially in dolomite, but fresh pyrite cubes are rarely found in outcrop. Nodules (up to 20 mm in diameter) consisting of small cubes and octahedra after pyrite also exist.

Some regional variations in sedimentation have taken place in the Devil Creek Formation. Daniels (1966b) has noted that west of Devil Creek the lower half of the sequence is a distinctive, very finely banded, blue-grey, brown-weathering dolomitic mudstone, whereas the upper half consists of thin-bedded, yellow-weathering dolomite with frequent maroon and brown stains. Near Wanna Station, the formation comprises thin-bedded dolomite with some thin sandstone and chert bands. In the Cobra Synclinorium, a yellow-weathering siltstone is developed at the base.

NANULAR SANDSTONE

Derivation of name: Nanular Creek, just west of Mount Clere homestead.

Type section: From lat. 25°08'30"S, long. 117°48'00"E to lat. 25°09'30"S, long. 117°47'30"E across the ridge south of the Sawback Range, and continued across the western end of a smaller ridge 2 km to the southeast.

Lithology: Medium- to fine-grained quartz arenite, grading to coarse-grained quartz siltstone in places. Interbeds of shale and wacke.

Thickness: 2 200 m estimated from air photographs, but the true thickness could be twice this value, because contacts are concealed. There may also be repetition by faulting in the section between the two ridges.

Stratigraphic relationships: The unit occurs between the Devil Creek and Backdoor Formations. North of Mount Deverell it appears to be laterally equivalent to the basal beds of the Backdoor Formation.

Description: The formation occupies a series of ridges on northern ROBINSON RANGE and southeastern MOUNT EGERTON.

It is a thick sequence of quartz arenite, usually medium to fine grained and moderately sorted, and containing accessory tourmaline, muscovite and zircon. Quartz overgrowths cement the grains. Coarse-grained feldspathic varieties and wacke beds are present, and in places the rock grades to coarse-grained quartz siltstone. Interbedded shale and siltstone is common. Ripple marks and cubic crystal molds after pyrite are abundant locally.

ULLAWARRA FORMATION

Derivation of name: Ullawarra Homestead (lat. 23°28'45"S, long. 116°06'45"E).

Type area: No adequate section through the formation is known. The type area is around Ullawarra homestead.

Lithology: Shale, siltstone and fine-grained arenite, with minor interbeds of wacke, claystone, dolomite and chert.

Thickness: Difficult to measure because of the large number of dolerite sills and the strong folding. Daniels (1966b) estimated a maximum thickness of between 1 500 and 1 800 m (5 000 to 6000 feet) near Coodardo Gap, but at least two-thirds of this is dolerite. An incomplete sequence of 1 100 m of sediment exists 20 km northeast of Staten Hill. In the centre of the Hells Doorway Syncline, 2 600 m is estimated, a thickness so great that it probably includes rocks equivalent to, but lithologically different from the Fords Creek Shale, and hence may more properly be called the Backdoor Formation, the lateral equivalent of the Ullawarra Formation.

Stratigraphic relationships: The Ullawarra Formation conformably overlies the Devil Creek Formation or Discovery Chert, and is conformably overlain by the Curran Formation. The lower boundary is regarded as the top of the youngest major underlying dolomite member, or, in thinly bedded sequences, the level at which the dolomite becomes less than 10 per cent of the total. The Ullawarra and Devil Creek formations interfinger with and laterally replace one another. The Ullawarra upper boundary is the first persistent cream-weathering, siliceous black shale of the Curran Formation. The unit pinches out east of Irregully Creek, and is replaced by interbedded chert breccia, shale, and siliceous laminated mudstone up to about 3 m thick, which probably continue for a considerable distance to the southeast. The Ullawarra is equivalent to part of the Backdoor Formation of the Collier Subgroup, one formation passing into the other. Demarcation between them is made where the Ullawarra can no longer be distinguished from the Fords Creek Shale or Jillawarra Formation, after intervening units have cut out.

Description: The formation is a variable sequence, chiefly of shale, siltstone, and fine-grained arenite. The lutites can be grey, black, brown, white or green, and are silty and well-bedded, ranging from (usually) laminated up to 0.5 m thick. Very thin- to medium-bedded shales, with alternating maroon and cream, or chocolate and pale green banding are common, but are not found east of the Mount Vernon Syncline. Load casts are locally abundant. Small amounts of hard, grey or white, coarse-grained quartz silstone are common in the sequence, but in some areas, for example the Berala Syncline, it is the dominant rock type. It typically breaks into blocky rubble, may be laminated, massively bedded, or cross stratified, and grades into fine-grained sandstone in places. Minor interbeds consist of wacke, claystone, dolomite and chert. Cubic crystal casts after pyrite occur sporadically, but are not as abundant as in older formations.

The most striking feature of the Ullawarra Formation is the large number of sills which intrude it. Generally they tend to be concordant with the bedding, yet they are frequently transgressive. It is not unusual for dolerite to make up more than two-thirds of the formation interval, and to form a continuum of dolerite in which lie discontinuous sedimentary stringers. Some of the fine-grained rock could possibly be lava, but the only evidence for this is a possible volcanic bomb (Daniels, 1969). A small occurrence of basaltic lava is known from the Calyie Sandstone (Leech and Brakel, 1978), showing that the basic magma sometimes broke through to the surface. It would not be surprising if this also happened in the Ullawarra Formation, considering the large volume of dolerite. Of interest is a sill with a fine-grained, amygdaloidal and vesicular top, indicating that the intrusion was shallow.

CURRAN FORMATION

Derivation of name: Curran Well (lat. 23°24′00″S, long. 116°11′15″E).

Type section: Along east fork of Wannery Creek at Curran Well.

Lithology: Grey to black, cream-weathering shale and mudstone, and traces of chert, limestone dolomite and fine-grained sandstone.

Thickness: 100 m estimated in type section from air photographs. Thickens to 250 m to the southeast.

Age and stratigraphic relationships: Rests conformably on the Ullawarra Formation, or the bedded breccia which replaces it east of Irregully Creek, and lies conformably below the Coodardoo Formation. At its most easterly point the unit is faulted out against the Mount Vernon Fault system, but just west of the fault it starts to loose its distinctive appearance, presumably never extending much farther east as an identifiable formation. On the northern margin of the basin it can be traced on air photographs towards the Fords Creek area, where it appears to be equivalent to a shale bed within the Pingandy Dolomite Member of the Top Camp Formation.

The lower boundary of the formation lies at the first appearance of a characteristic cream-weathering, dark-grey to black mudstone. The upper boundary lies in the transition zone with the Coodardoo Formation at the level where the first sandstone bed occurs in the sequence. This transition zone is better exposed at Coodardoo Gap than in the type section.

Isotopic Rb-Sr dating of black shale from the unit obtained an isochron of 1 080 ± 80 m.y. (Compston and Arriens, 1968).

Description: Most of the type section is thinly bedded black aphanitic mudstone, which is usually fined laminated but, in places, alternates with massive bands. Small lentils of medium to coarse silstone are present, especially in the upper half. The mudstone weathers to a distinctive, cream-coloured, siliceous rock, but the silica is probably secondary. A characteristic white speckled appearance is common. Minor chert and fine-grained sandstone are present in places, as well as rare, thin dolomite and metre-long lenses of grey limestone.

The transition zone at Coodardoo Gap begins where the typical light-weathering mudstone gives way to 2.5 m of brownish, interbedded medium to coarse siltstone and aphanitic mudstone. The siltstone in this interval occurs in beds more than 0.36 m thick, in contrast to laminar bands lower in the sequence. These beds are then overlain by the dominantly sandstone-mudstone interval at the base of the Coodardoo Formation.

The Curran Formation is split by a sill for most of its length, which gives it a distinctive double-banded photopattern.

COODARDOO FORMATION

Derivation of name: Coodardoo Gap (lat. 23°48′15″S, long. 116°38′00″E).

Type section: Coodardoo Gap (Fig. 8).

Lithology: Dominantly arenite and wacke, with subordinate siltstone, shale and aphanitic mudstone.

Thickness: About 200 m estimated in type section from air photographs. Daniels (1966b) records a maximum thickness of 370 m.

Stratigraphic relationships: The youngest formation of the Edmund Subgroup, it is conformable between the Curran Formation below and the Fords Creek Shale above. It lenses out in MOUNT EGERTON and merges with the Pingandy Dolomite Member at the top of the Top Camp Formation on TUREE CREEK. Air-photograph interpretation indicates that dolomite of the Pingandy Member comes to underlie and overlie the Coodardoo Formation where the sandstone lenses out. The lower boundary lies in the transition zone with the Curran, while the upper boundary is between the highest sandstone bed and the interbedded shale and carbonate of the Fords Creek Shale. A transition zone of interbedded sandstone and shale is developed in some areas, in which the highest major sandstone bed is the top of the Coodardoo.

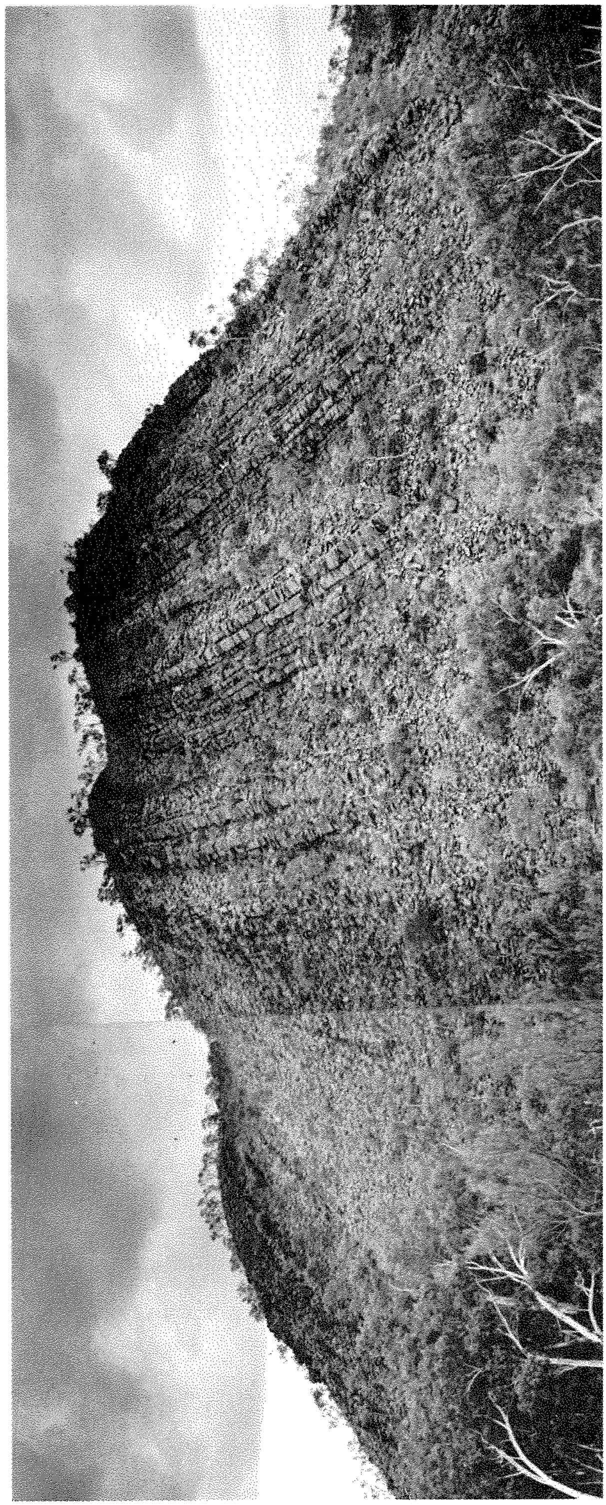

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Figure 8. Type section of the Coodardoo Formation (Coodardoo Gap—looking west). Differential erosion has left the hard sandstone standing out from the soft siltstone and mudstone.

Description: At the base of the type section, the formation consists of interbedded sandstone and mudstone, in beds up to 0.7 m thick, roughly in the proportion 3:2. The sandstone comprises fine- and medium-grained, slightly micaceous, quartz wacke and moderately sorted quartz arenite. Higher up, graded bedding and dewatering structures (Fig. 9) appear, and occur at intervals throughout the rest of the sequence. At the centre of the unit, shale, siltstone and aphanitic mudstone form only about 20 per cent of the sequence, and arenite and wacke occur in equal

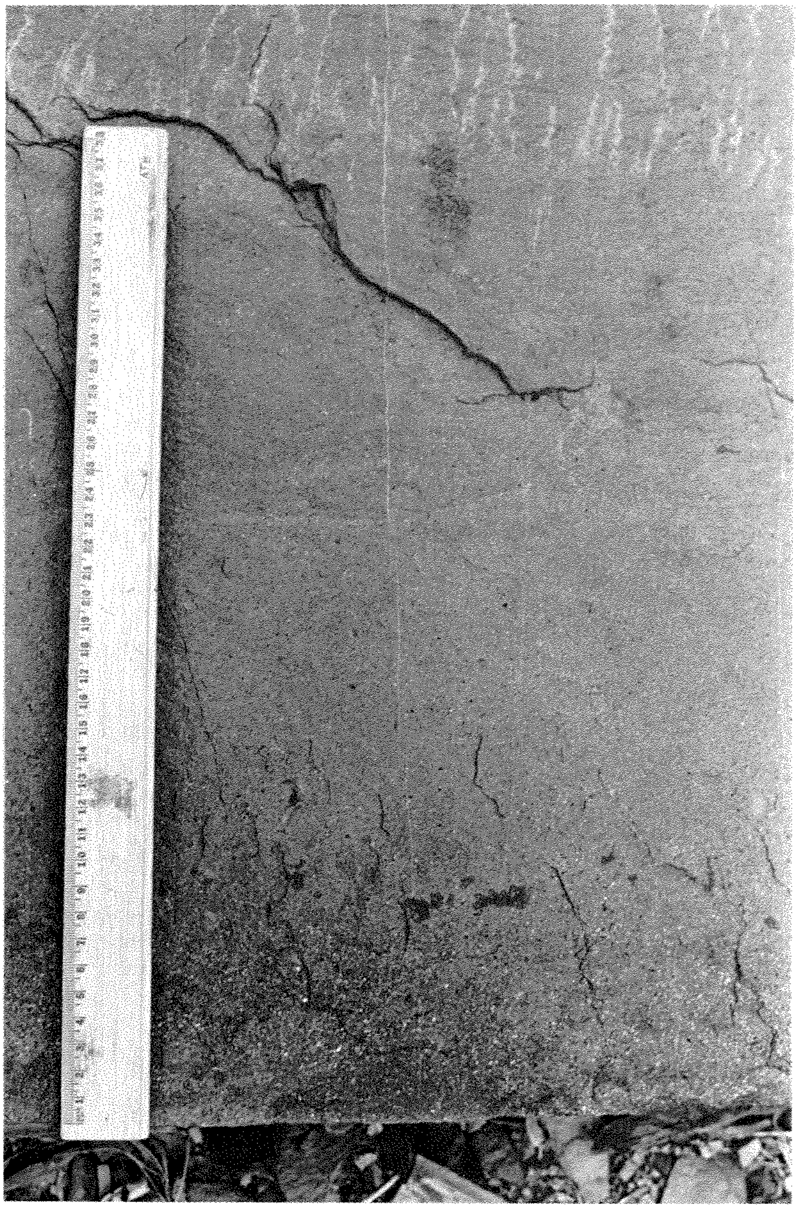

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Figure 9. **Graded bedding, Coodardoo Formation at Coodardoo Gap. The graded bedding interval (0-23 cm on ruler) is overlain by an interval of faint, horizontal lamination, from which dewatering columns arise.**

amounts, often grading into each other. There is more fine-grained than medium-grained sandstone. The interbedded shale is pale green, and the mudstone, brown. Uncommon medium-scale cross-bedding and layers of granule conglomerate up to 150 m thick are present. At the top of the section the arenite and wacke are thickly bedded and accompanied by laminated, micaceous siltstone.

Sorting of the formation improves, and average grain size increases towards the southeast; this results in diminished proportions of mudstone, shale, wacke and fine-grained arenite. In the Dooley Downs area, the unit consists mainly of thick-bedded, medium-grained, moderately sorted quartz arenite; beds are massive, and graded bedding is absent.

MUCALANA SUBGROUP

Derivation of name: Mucalana Creek, 11 km northwest of Mount Vernon.

Constituent formations: Fords Creek Shale, Mount Vernon Sandstone, and Kurabuka Formation.

Thickness: Composite thickness of the constituent formations in their respective type sections is at least 2 955 m.

Distribution: Occupies a belt of country along the main axial region of the Bangemall Basin from the head of the Edmund River to Bulloo Downs Station.

Stratigraphic relationships: The Mucalana Subgroup, the youngest in the western half of the basin, rests conformably on the Edmund Subgroup at its southwestern flank and on the Top Camp Formation at its northern flank. It thus lies in both the western and northern facies provinces. The group is laterally equivalent to the upper part of the Backdoor Formation and overlying formations, although it is usually faulted against the Collier Subgroup. Physical continuity between these two subgroups exists only in the northeast corner of COLLIER, where the distinction between them is arbitrary.

The Mucalana Subgroup is characterized by simple stratigraphy and lateral persistence of its formations, which do not display rapid changes in thickness. The lutites of the subgroup are usually greenish when fresh.

FORDS CREEK SHALE

Derivation of name: Fords Creek, on the southeast corner of TUREE CREEK.

Type section: Designated as the detailed traverse by R. Halligan along the southwest branch of Fords Creek between lat. 23°50′15″S, long 117°12′30″E, and the Kenneth Range at lat. 23°55′00″, long. 117°09′45″E (Halligan and Daniels, 1964). The base of the formation can be inspected at Coodardoo Gap.

Lithology: Greenish shale and mudstone, with lesser arenite, wacke, dolomite and chert.

Thickness: Approximately 1 740 m in type section (Daniels, 1968).

Stratigraphic relationships: The Fords Creek Shale is the lowest formation of the Mucalana Subgroup, and lies conformably on the Coodardoo, Curran and Top Camp Formations in various places. It is overlain conformably by the Mount Vernon Sandstone. At its eastern extent near Bulloo Downs, it grades downwards into the Top Camp Formation because of the downwards increasing proportion of dolomite, and becomes difficult to distinguish from it. In the same area the unit is laterally equivalent to the Backdoor Formation. The limits of these two formations are arbitrary, but have been set at a break in outcrop at the internal drainage divide, east of Bulloo Downs.

The bottom contact of the Fords Creek Shale at Coodardoo Gap is between thick-bedded sandstone of the Coodardoo Formation and an overlying interbedded shale-carbonate sequence. In the Dooley Downs area, there is a transition zone of interbedded sandstone and shale at the base, and the boundary is placed at the level where shale becomes dominant. The upper contact lies at the change from dominantly lutite to dominantly arenite. The Fords Creek Shale interfingers with the Mount Vernon Sandstone in some areas.

Named member: Jeeaila Sandstone Member.

Description: The Fords Creek Shale consists essentially of micaceous shale and mudstone. It is green, greenish grey or black when fresh but weathers to brown, greenish brown, maroon, buff or white. The green colour is due to chlorite, but the absence of organic carbon is the real reason for the colour difference of the Jillawarra Formation. The rocks are laminated and massively bedded, with widespread load casting. Pyrite is rare, but occurs in a few beds, where it may be up to 4 mm in size. Interbeds of dolomite and dolomitic shale are present locally at the base, and in the Bulloo Downs district, higher in the sequence. These are usually thinly bedded, laminated, fine grained, pink, grey brown, or dark grey weathering to white or dark brown, with rare thin intercalations of light grey limestone.

Sandstone members are plentiful; most are quartz arenite with a low feldspar content, and some contain lithic grains. Impressions of shale fragments are widespread. The arenites are most numerous northeast of Dooley Downs homestead. Wacke is less common, and is medium bedded, dark grey or dark green, and may have a few intraclast molds. Thin sandstone beds occur at the top of the formation wherever a transition zone is developed.

Chert is comparatively rare. Varieties observed include a mid-grey, streaky textured rock with small white specs, probably a silicified shale, and also a light grey, algal laminated type.

JEEAILA SANDSTONE MEMBER

The most important member of the Fords Creek Shale is a prominent cliff-forming unit up to 80 m thick outlining folds in the northwest corner of MOUNT EGERTON and extending 90 km to the east-southeast where it is cut off by the Mount Vernon Fault system. It is named after the Jeeaila River. The principal section is 23 km southeast of Dooley Downs homestead, where a track southeast from Eden Bore passes through a gap in the ridge (lat. 24°12′15″S, long. 117°19′00″E). The sequence below the member can be seen 9 km to the northwest near the nose of a syncline, and the sequence above can be inspected 1.5 km farther northeast where the member is crossed by a track.

The unit consists chiefly of medium-grained, well to moderately sorted arenite, but finer-grained and less sorted varieties also occur. The rocks are grey when fresh, laminated or massive, with occasional cross-bedding, very thickly bedded, and have zones with intraclasts of shale and siltstone. It overlies chert, interbedded with shale and mudstone, all of which are grey, green or black when fresh. The chert, has both undulose and flat bedding. It is persistent over at least 10 km in the reference area, but absent towards the Mount Vernon district, where black, laminated shale takes its place. the Jeeaila Sandstone Member is overlain by a green shale and siltstone sequence with sandstone interbeds.

MOUNT VERNON SANDSTONE

Derivation of name: Mount Vernon (lat. 24°8′45″S, long. 118°1′45″E).

Type section: Glen Ross Creek, 5 km north of Mount Vernon homestead (lat. 24°11′S, long. 118°14′E).

Lithology: Fine- and medium-grained arenite with rare conglomerate; interbedded siltstone and shale locally.

Thickness: 225 m in the type section.

Stratigraphic relationships: Interbedded between the Fords Creek Shale below and the Kurabuka Formation above. It is correlated with the Calyie Sandstone to the east. The lower contact is the change from dominantly lutite to dominantly arenite, and is usually well marked. In the type section, a dolerite sill has intruded along this horizon. However, in some areas, of which the northern side of the Lofty Range is the best example, the formation interfingers with the Fords Creek Shale, resulting in a thick transition zone of interbedded arenite and green shale. In such a case, the stratigraphic boundary is taken as the first appearance of a major sandstone member. The upper contact is an abrupt change from sandstone to greenish lutite of the Kurabuka Formation.

The Mount Vernon Sandstone was formerly regarded as the basal member of the Kurabuka Formation (Daniels 1968, 1969).

Description: The Mount Vernon Sandstone is a prominent cliff and ridge-forming formation and therefore a good marker unit for its 300 km extent.

In the type section it consists of white or grey, fine- and medium-grained quartz arenite, with rare feldspathic arenite and conglomerate. Beds of quartz siltstone and shale occur in the basal 30 m. Fine-grained arenite constitutes about 85 per cent of the total thickness of sandstone, the remainder being medium grained except for one coarse bed. The rocks are well sorted, usually with less than 2 per cent feldspar, but rarely with as much as 20 per cent. Laminated and fissile beds are more common than massive beds. Up to one per cent glauconite occurs in a micaceous interval (36 m thick) near the middle of the formation. Cross-bedding, ripple marks, and current lineations are common. Interclasts, up to 10 cm long, of shale, siltstone and fine sandstone are widespread. Thin conglomerate and pebbly sandstone layers contain pebbles of dark greenish-grey chert, vein quartz and arenite.

Outside the type section, medium and coarse-grained arenites are more common in some areas. Coarse-grained quartz siltstone is particularly important in the thick transition zone on the north side of the Lofty Range. Also in the Lofty Range is a small occurrence of green diamictite, comprising coarse, well-rounded quartz sand grains dispersed in a matrix of finer sand, silt and clay.

KURABUKA FORMATION

Derivation of name: Kurabuka (or Koorabooka) Creek, southeast corner of EDMUND.

Type section: There is no well-exposed section available anywhere, so a composite type section is specified in the Mount Vernon district where the highest level of the unit is preserved. Outcrop is generally poor, but is sufficient to denote the nature of the succession. The section starts at the end of the Mount Vernon Sandstone type section, (lat. 24°11′30″S, long. 118°25′45″E), and ends at the Mount Vernon Fault System (lat. 24°17′00″S, long. 118°16′00″E).

Lithology: Shale, mudstone and chert, usually greenish, with minor limestone and calcareous siltstone.

Thickness: > 990 m measured in the type section, but the top is faulted out and shallow dips make calculations imprecise.

Stratigraphic relationships: The unit is the highest formation of the Mucalana Subgroup, and conformably overlies the Mount Vernon Sandstone. It is equated with the Ilgarari Formation of the Collier Subgroup.

Description: The Kurabuka Formation is a sequence of shale, siltstone and aphanitic mudstone, usually greenish-grey or green in colour due to chlorite. It weathers brown or khaki. The upper 15 per cent of the unit contains numerous rhythmic alternations of green and white-weathering rock, which result in a characteristic white and grey striped photopattern. The two varieties have no visible difference except colour and a slight variation in fissility. Much-divided white mica is common. Near the top of the measured sequence are minor bands, up to 50 mm thick, of fine-grained grey calcareous siltstone, which commonly has lustre mottling due to the coarsely crystalline carbonate cement. Limestone occurs in a 5 m interval towards the middle of the section. This rock is mid-grey, finely to coarsely crystalline and contains blebs and crystals up to 2 mm across (20 mm in some bands) of dark-brown siderite. Dolomite may also be a constituent. The lower part of the formation is characterized by a light photopattern, because the rocks although dark grey to black or green where fresh, generally weather to white, cream, fawn, or brown colours. Interbedded chert varies from dark grey to greenish, and from laminated to massively bedded. It contains detrital quartz and sericite and appears to be silicified lutite. Rare cubic molds after pyrite, and small white specs probably after an iron sulphide, occur in some chert and mudstone bands. Lenses of quartz arenite and laminae of sandstone in lutite occur locally.

TOP CAMP FORMATION

Derivation of name: Top Camp gold workings (lat. 23°44′45″S, long. 117°16′30″E).

Synonomy: Formerly called the Top Camp Dolomite (Halligan and Daniels, 1964), but because the proportion of dolomite decreases easterly and becomes subordinate, the term formation is more appropriate. The unit was abandoned by Daniels (1966b) and replaced by several formations in EDMUND, but most of these formations do not occur in the northern facies, and the name Top Camp Formation is retained.

Type section: Designated as the detailed traverse by R. Halligan along Fords Creek between the unconformity at lat. 23°45′45″S, long. 117°17′15″E and lat. 23°50′15″S, long. 117°12′30″E (Daniels, 1968).

Lithology: Dolomite, limestone, shale, dolomitic shale, chert, quartz sandstone and breccia, with minor conglomerate lenses at the base.

Thickness: Approximately 954 m recorded by Daniels (1968) in the type section; 2 500 m is given by Muhling and others (1976) from MOUNT EGERTON.

Stratigraphic relationships: The Top Camp Formation is unconformable on the Wyloo and Bresnahan Groups (the Top Camp Unconformity of Maitland, 1909). The unit is overlain conformably by the Fords Creek Shale of the Mucalana Subgroup, and the contact occurs between the highest dolomite member and overlying green shale.

The formation is laterally equivalent to the Edmund Subgroup. Some formations, such as the Kiangi Creek and Jillawarra Formations, appear to lens out from the Edmund Subgroup, while others, such as the Irregully and Devil Creek Formations, continue into the Top Camp Formation. The Pingandy Dolomite Member at the top of the Top Camp is equivalent to the Devil Creek, Curran and Coodardoo Formations, while a chert directly beneath the Pingandy Dolomite Member may represent the Discovery Chert, and the remainder of the Top Camp Formation is probably continuous with the Irregully Formation. A possible correlation scheme is shown in Figure 10. The change from the nomenclature of the Edmund Subgroup to the Top Camp Formation is somewhat arbitrary. The non-dolomite formations of the Edmund Subgroup are traced southeasterly until they either lens out or can no longer be recognized. The remaining sequence is then referred to as Top Camp Formation, this change taking place in the area between Wandarry Creek and Fords Creek, near the EDMUND - TUREE CREEK boundary.

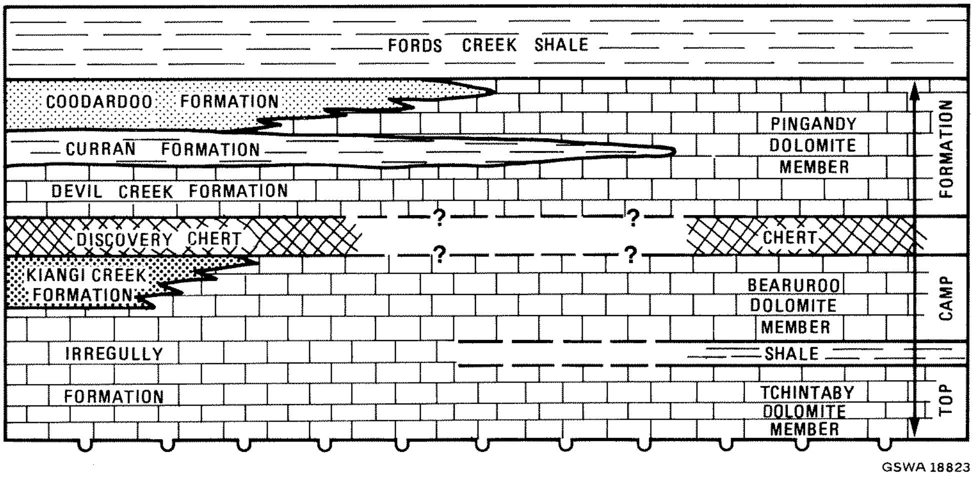

Figure 10. **Diagrammatic section showing transition from Edmund Subgroup (left) to Top Camp Formation (right). Not to scale.**

North of Bulloo Downs homestead, the formation is absent, and a succession of shale, sandstone, and minor dolomite of the Fords Creek Shale, overlies the basal unconformity. Extensive Cainozoic deposits largely obscure this change, but it appears to be a facies change in which dolomite members lens out eastwards.

Named members: Tchintaby Dolomite Member, Bearuroo Dolomite Member, Pingandy Dolomite Member, and Prairie Downs Member.

Description: The formation extends along the northern margin of the Bangemall Basin from TUREE CREEK to NEWMAN, and occupies over half of the northern facies.

At its base is the Tchintaby Dolomite Member, which has lenses of conglomerate and finer clastics near the unconformity. This is overlain in the type section by dark-grey shale, which gives off a white or yellow efflorescence in some exposures. The same interval in the Tunnel Creek area (northeast corner of MOUNT EGERTON) contains grey, pale-brown and olive shale and siltstone, with interbedded grey to black, laminated and streaky textured chert that is probably a silicified shale. Minor lenses of sandstone are also present. The next unit, the Bearuroo Dolomite Member, is thinner than the others and is succeeded by about 25 m of chert in the Fords Creek section. At Tunnel Creek however, a distinctive unit of fine-grained quartz arenite overlies the Bearuroo Member. This forms hills covered by blocky rubble but crops out poorly. Bedding surfaces are commonly marked by current lineations, groove casts, flute molds and current ripples, while intraformational conglomerate, slump structures and lenses of algal-laminated dolomite are also present. This rock unit is unlike any in the Edmund Subgroup. Last in the formational sequence is the Pingandy Dolomite Member, which is overlain by the Fords Creek Shale.

TCHINTABY DOLOMITE MEMBER

This basal member is named after Tchintaby Pool in the Ashburton River (lat. 23°59′45″E, long. 117°57′00″E). Although we have not examined Halligan's section in the Fords Creek, it appears to be the best location for the type sections of this and the higher two dolomite members. The Tchintaby Dolomite Member is defined as the dolomitic and associated rocks making up the first 216 m of this section as recorded by Daniels (1968). Its upper contact is with shale, except in NEWMAN, where it is overlain disconformably by the Prairie Downs Member.

The unit contains, thinly bedded, pale fawn, yellow-brown and white dolomite, thickly bedded stromatolitic dolomite, and fine- to medium-grained siliceous quartz sandstone. There are also minor shale beds, each with a basal granule conglomerate. Small-pebble and granule conglomerate comprise the basal 6 m of the member, but these vary, and Woodward (1890) has remarked on the absence of conglomerate, and the presence of dolomite resting directly on the unconformity in the Top Camp area.

BEARUROO DOLOMITE MEMBER

Named after Bearuroo Spring (lat. 23°54′S, long. 117°27′E), this unit occurs near the middle of the Top Camp Formation, between the dark-grey shale and the 25 m-thick chert in the Fords Creek section. No thickness is given by Daniels

(1968), but it is likely to be much more than the 85 m estimated for the unit from air photographs in the Tunnel Creek area. Most of it is thinly bedded dolomite, and dolomitic shale with some thin sandstone and chert bands, overlying interbedded, blocky to massive buff- and grey-weathering stromatolitic dolomite and fine-grained, buff, siliceous quartz sandstone. These rocks are succeeded by about 2 m of massive limestone, and then 4 m of white quartz arenite and medium to coarse, brown-weathering quartz sandstone with well-developed slump structures. Finally, 3 m of brown-weathering, massive grey dolomite forms the top of the member.

PINGANDY DOLOMITE MEMBER

This top member, whose name is derived from Pingandy homestead (lat. 23°59′45″S, long. 117°31′30″E), extends to Bulloo Downs station. It probably correlates with the Devil Creek Formation. The type section is specified as the 380 m interval overlying the 25 m-thick chert in Fords Creek, but the top of the member is better exposed near Pingandy Spring (lat. 24°04′30″S, long. 117°40′45″E).

In Fords Creek, it has shaly limestone and pebbly limestone at its base. This is overlain by an alternation of pale grey, cream and brown-weathering dolomite, within which are intraformational breccia beds, 0.3 to 1.2 m thick, which contain fragments oriented roughly parallel to bedding and up to 0.6 m long. The top half of the Pingandy Member comprises grey, buff, and fawn shaly dolomite, dolomitic shale, minor dark blue-grey dolomite, and banded chert. Shale is interbedded in places.

Numerous *Conophyton garganicum australe* stromatolites, up to 5.5 m in diameter, occur in a dolomite at lat. 24°01′45″S, long. 117°51′30″E (Grey, 1977), which is probably near the base of the Pingandy Member.

PRAIRIE DOWNS MEMBER

A thick wedge of conglomerate and arkose, referred to as "Prairie Downs Beds" by Daniels and MacLeod (1965), exists in the southern part of NEWMAN. It is named after Prairie Downs homestead (lat. 23°33′00″S, long. 119°8′45″E). For over half its extent it is unconformable on the Bresnahan Group, but elsewhere it rests disconformably on the Tchintaby Dolomite Member. Mapping on COLLIER, however, has established that the Tchintaby Member also overlies the Prairie Downs Member, indicating that the latter is a wedge which extended southwesterly into the area of dolomite deposition. On NEWMAN, the Prairie Downs Member is overlain by shale.

The type area is 10 km south of Prairie Downs homestead. Here the sequence begins with a conglomerate, at least 300 m thick, which consists mainly of boulders of jaspilite. Pseudomorphs after riebeckite have been recorded, suggesting derivation from the Hamersley Group to the north or northeast. The conglomerate grades laterally and vertically into arkose with thin interbanded quartz arenite, and these arenaceous rocks constitute the greater part of the unit. The rapid southwesterly thinning of the member suggests that deposition was controlled by contemporaneous activity along the Prairie Downs Fault (Daniels and MacLeod, 1965).

COLLIER SUBGROUP

Derivation of name: Collier Range.

Constituent formations: Wonyulgunna Sandstone, Backdoor Formation, Calyie Sandstone and Ilgarari Formation.

Thickness: The composite thickness is at least 6 400 m.

Distribution: Occupies most of the eastern half of the Bangemall Basin.

Stratigraphic relationships: Lies with major unconformity on units of the Hamersley and Nabberu Basins, the Sylvania and Marymia Domes, and the Cornelia Inlier. It is laterally equivalent to the Edmund and Mucalana Subgroups to the west, the Diebil Subgroup to the northeast, and the Kahrban Subgroup to the southeast. Its relationships to the Manganese Subgroup in the north is unclear.

The Coodardoo, Curran and Devil Creek Formations and Discovery Chert of the western facies lens out eastwards, so that the Fords Creek Shale, Ullawarra and Jillawarra Formations merge into one uniform lutite sequence called the Backdoor Formation, whose thickness is comparable to the combined thickness of the Jillawarra-Fords Creek interval (Fig. 11).

It was previously uncertain whether the Calyie Sandstone was equivalent to the Mount Vernon Sandstone or represented a lower stratigraphic level, analogous to the Coodardoo Formation (Brakel and Muhling, 1976). However, subsequent mapping in adjacent BULLEN and NEWMAN leaves little doubt that Calyie Sandstone and Ilgarari Formation are respectively equivalent to the Mount Vernon Sandstone and Kurabuka Formation. There are, however lithological differences between these units of the Collier and Mucalana Subgroups. Whereas the Kurabuka Formation is largely made up of greenish shale, mudstone, and chert, the Ilgarari Formation contains white and brownish shale, mudstone and fine-grained sandstone. Green rocks are rare in the Ilgarari Formation and have not been found at depths of up to 60 m in the Ilgarari mine. The Calyie Sandstone east of the Neds Gap Fault is thinner than the Mount Vernon Sandstone to the west, and has no corresponding transition zone of fine-grained sandstone at its base. This is explained

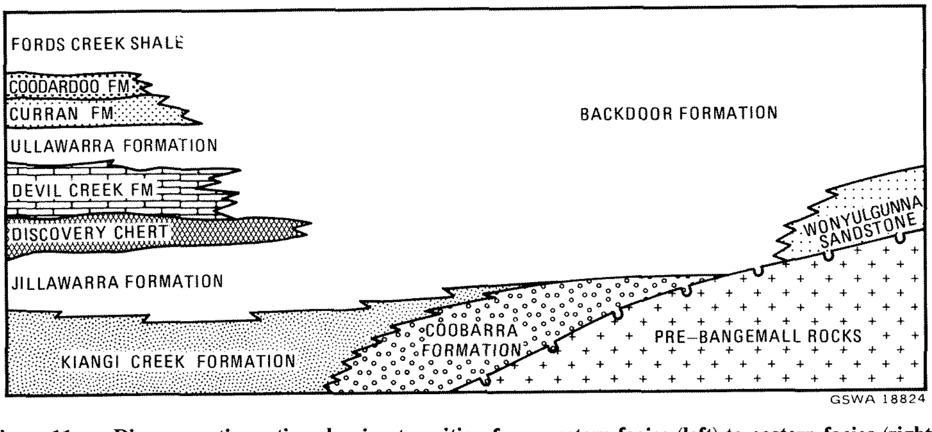

Figure 11. Diagrammatic section showing transition from western facies (left) to eastern facies (right). Not to scale.

by a facies change coincident with the Neds Gap Fault, because air photographs also show that beds at the base of the Mount Vernon Sandstone grade into Fords Creek Shale along strike on the north side of the Lofty Range. The Fords Creek Shale itself is represented by some green shale near the top of the Backdoor Formation in the Tangadee Syncline and west of Bunningunna Bluff. Green shale is absent farther south, and uncommon east of Neds Gap, where black and grey shale occur throughout the unit.

WONYULGUNNA SANDSTONE

Derivation of name: Wonyulgunna Hill (lat. 24°49′00″S, long. 119°45′45″E).

Type area: Wonyulgunna Hill ridge.

Lithology: Quartz arenite with minor cobble conglomerate forms the bulk of the formation; siltstone, shale and chert are more restricted.

Thickness: Estimated from air photographs to be at least 800 m.

Stratigraphic relationships: Rests unconformably on schist, banded iron-formation, and granite. It is conformably overlain by the Backdoor Formation. The unit merges with the Calyie Sandstone in the northern Carnarvon Range, where the Backdoor Formation lenses out eastwards. The unconformity is well exposed at lat. 24°48′15″S, long. 119°53′00″E, while the upper contact is present near the eastern end of the Yibbie Range where the sandstone passes upwards into coarse-grained quartz siltstone of the Backdoor Formation. The Wonyulgunna Sandstone lenses out westwards, and may be equivalent to sandstone members near the base of the Backdoor Formation in the Batthewmurnarna Hill district.

Description: The Wonyulgunna Sandstone forms an arcuate series of prominent ridges 100 km long. Sandstone, which forms the bulk of the formation, directly overlies the unconformity in the Wonyulgunna Hill and Mount Essendon areas. The most common type of sandstone is whitish, coarse-grained, moderately sorted, thickly bedded quartz arenite with a low feldspar content (generally 2 per cent or less). Cross-bedding and intraclast molds are widespread but not abundant. Conglomerates are uncommon; the best examples are at lat. 24°48′15″S, long. 119°53′00″E, where pebble and granule conglomerate rests on the unconformity, and 5 m higher, where a 10 m-thick cobble conglomerate is present. The clasts are vein quartz, quartzite, and occasional silicified banded iron-formation fragments. In most places, however, there is no basal conglomerate.

The area south of the Yibbie Range contains a lutite sequence which is not present elsewhere below the main sandstone interval. These rocks are dominated by coarse-grained, hard, quartzitic siltstone, but soft, earthy siltstone and shale are also common. Laminations, intraclast molds and rare pyrite cubes are present. Minor interbeds of sandstone are usually fine-grained. All these rocks vary in colour from grey, white and greenish to brown and pink. At the unconformity, immediately overlying foliated granitoid in the northeast corner of PEAK HILL and the adjacent part of COLLIER, is a black, massively bedded chert which weathers to off-white and light grey. The chert is laminated in places, but does not have the streaky texture of the Discovery Chert. The top of the lutite sequence is visible in two hills near the eastern end of the Yibbie Range, where coarse-grained siltstone grades up through fine-grained to medium-grained sandstone. These rocks are also found in outliers on the metamorphic-granitic basement in the southeast corner of COLLIER.

BACKDOOR FORMATION

Derivation of name: Backdoor Hills (lat. 24°36′S, long. 119°06′E), about 15 km northwest of the Collier Range.

Type area: From north of the Backdoor Hills to the western end of the Collier Range.

Lithology: Dominantly shale and siltstone, with lesser chert, claystone, dolomite and sandstone.

Thickness: No complete section measured, but estimated to be about 3 700 m from air photographs and a partial section at the end of the Collier Range.

Stratigraphic relationships: The lower contact in the type area is obscured. The unit is thought to be conformable on the Coobarra Formation, but the Kiangi Creek Formation may lie between the two. At Wonyulgunna Hill and eastwards, the unit overlies the Wonyulgunna Sandstone, but east of Batthewmurnarna Hill and near the Sylvania Dome it is unconformable on basement. It is conformably overlain by the Calyie Sandstone. The Backdoor Formation lenses out eastwards near Mount Essendon, while to the west it is equivalent to the Jillawarra-Fords Creek interval of the Edmund and Mucalana Subgroups. Here the unit usually overlies the Devil Creek Formation, but extends further down the sequence where the latter and the Discovery Chert have lensed out. It appears to be laterally equivalent to the Manganese Subgroup.

Rocks in the Mount Vernon Syncline, Hells Doorway Syncline and Brumby Creek Anticline previously mapped as Ullawarra Formation by Muhling and others (1976) and Brakel and others (1978) have now been assigned to the Backdoor Formation, because they probably include Fords Creek Shale equivalents, the Curran and Coodardoo Formations being absent.

Description: The Backdoor Formation occurs mainly on COLLIER and PEAK HILL. Being less resistant than the Calyie Sandstone, it forms subdued strike ridges and hills, and level country.

The shale and siltstone are usually bedded, laminated, and dark grey when fresh, weathering to brownish, yellow-brown, white and maroon colours. Green and olive-grey shales, similar to the Fords Creek Shale, occur in the centre of the Tangadee Syncline and approximately 15 km west of Bunningunna Bluff below the top of the formation. Siltstones, both laminated and massive, are of two types: a soft, earthy rock similar to the shale, and a hard grey-white, siliceous quartzitic type, which grades into fine-grained, feldspathic arenite.

Subordinate bands of chert, sandstone and claystone are interbedded in the succession. Chert bands are dark grey to black, and some are laminated and streaky-textured like the Discovery Chert. Other types appear to be silicified shale. Black, planar-bedded chert occurs near the top of the formation in the central part of COLLIER and has also been noted in parts of the Calyie Syncline. It is well exposed at the western end of the Collier Range, where chert is also rhythmically interbedded with shale. Another prominent chert member lies lower in the sequence at Beyondie Bluff. Sandstone members are present as thin beds in several places, but the best development of quartz arenite interbeds is in the Batthewurnarna Hill region. A few thin bands of quartz wacke appear at the top of the formation in the western Collier Range, where a grey-brown wacke also forms a persistent basal member of the Calyie Sandstone. Lenses of fine-grained, laminated dolomite containing rare, thin interbeds of limestone and calcareous sandstone occur at widely separated localities.

Pyrite is present as small, irregular patches, nodules, or weathered-out cubic molds up to 2 cm on edge, but is plentiful only in the southwestern and northeastern corners of COLLIER (Fig. 12).

Layers of alunite up to 8 cm thick occur in shales near Beyondie Bluff, and were probably kaolin bands altered by dilute sulphuric acid derived from the weathering of pyrite.

CALYIE SANDSTONE

Derivation of name: Calyie Hills, 19 km east of Mulgul homestead.

Type section: The Gap in the Calyie Hills and Flat Top Range at Mingah Spring, 7km north of Mingah Springs homestead.

Lithology: Quartz arenite with minor wacke and siltstone in the type section. Elsewhere minor conglomerate and rare shale and dolomite are present.

Thickness: 1 220 m in type section.

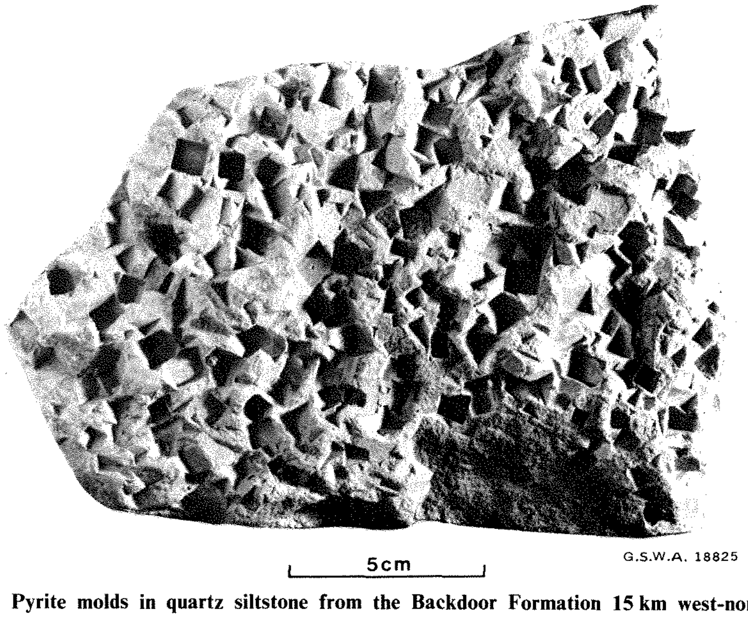

Figure 12. Pyrite molds in quartz siltstone from the Backdoor Formation 15 km west-northwest of Mingah Springs homestead, COLLIER.

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Figure 13. Angular unconformity between iron-formation of the Frere Formation and the overlying massive sandstone of the Calyie Sandstone in the Carnarvon Range (29°09'00"S 120°45'40"E) NABBERU.

Stratigraphic relationship: Overlies the Backdoor Formation and underlies the Ilgarari Formation in the type section. It is unconformable on pre-Bangemall rocks on NABBERU (Fig. 13), TRAINOR and RUDALL, and overlies the Coonabildie and Marlooyanoo Formations in apparent conformity. Laterally, it is equivalent to the McFadden Sandstone. The upper and lower boundaries of the formation are defined by the youngest and oldest laterally persistent sandstone members in the interval.

The sandstone on Mount Sanford and the southern side of the Lofty Range Fault can be correlated with either the Mount Vernon or Calyie Sandstones, but are assigned to the later because they overlie the Backdoor Formation.

Description: The Calyie Sandstone extends over most of the eastern half of the Bangemall Basin. On COLLIER it forms ridges, plateaux and mesas. Farther east, it is usually exposed as low, rocky hills jutting through windblown sand.

A generalized type section is given in Table 3. The bulk of the unit is whitish, fine- and medium-grained, siliceous quartz arenite, interbedded with minor wacke towards the base. Most of the rock is moderately sorted with up to 1 per cent feldspar. More than 2 per cent feldspar is present in the 270-500 m interval, and one bed contains over 5 per cent. The cement is silica, rarely clay. Apart from some de-watering structures, most beds are massive, but a subordinate number contain internal lamination and cross-bedding.

The formation varies from this section over its regional extent, so that medium-grained arenites are the most numerous. The sorting is generally moderate to very good, but in the TRAINOR-GUNANYA area the sorting is poorer. Clay cement is present in many unsilicified rocks. Beds less than 0.1 m thick tend to become more common to the east while those over 0.3 m decrease in abundance. The dominant rock on COLLIER is whitish, massive and structureless, but this decreases in abundance eastwards as yellow-brown laminated and cross-bedded outcrops become prevalent. Some of these variations are due in part to a westerly increase in surface silicification, which tends to obscure primary structures. The uncommon silicified beds of the desert region are always at the tops of outcrops left as remnants of a former silicified surface, now largely sand-blasted away.

TABLE 3. TYPE SECTION OF CALYIE SANDSTONE

Cumulative Thickness (m)	*Unit Thickness (m)*	*Description*
		Ilgarari Formation
1 218	70	Quartz arenite, white or grey; fine-grained, some medium-grained; very well to moderately sorted; bedding usually massive, but with alpha and pi cross-bedding and abundant intraclasts in some beds; sporadic exposure
1 148	106	No outcrop
1 042	40	Quartz arenite, fine- to medium-grained; moderately to very well-sorted; interbedded with coarse-grained quartz siltstone; both rocks commonly laminated with small and medium-scale cross-bedding and some intraclasts
1 002	510	No outcrop
492	95	Quartz arenite, coarse- to fine-grained but mostly medium-grained, moderately to well-sorted; massively bedded; some de-watering structures
397	71	No outcrop
326	85	Quartz arenite, fine- to medium-grained; moderately to well-sorted; massive; intraclasts and de-watering structures in some beds; pyrite molds in lowest bed
241	14	Siltstone, white or light grey, coarse-grained, laminated; small-scale trough cross-bedding in places; sporadic exposure
227	77	Quartz arenite, fine- to coarse-grained; moderately sorted, massive; rare intraclasts, de-watering structures and small-scale cross-bedding
150	28	No outcrop, except for 2 m of fine-grained massive quartz arenite
122	19	Quartz arenite, light grey, and olive, fine-grained, moderately and well-sorted, massive, some de-watering structures; interbedded with micaceous and quartz siltstone, with some small scale cross-bedding, and minor fine-grained wacke
103	23	No outcrop except for 2 m of medium to coarse, massive quartz arenite, and minor siltstone
80	80	Quartz arenite, fine- to coarse-grained, moderately to well-sorted, massive or roughly bedded; de-watering structures common in places; rare intraclasts and groove molds; interbedded with lesser fine-grained wacke and minor sandstone-siltstone laminite; exposure sporadic in parts
		Backdoor Formation

Pyrite, as cubes or formless masses 1-10 mm across, has been observed sporadically only on COLLIER. Intraclasts are widespread but rare, and are confined to specific beds. They are generally less than 3 cm in size, but range up to 25 cm where they are more abundant. Usually only the weathered-out molds remain, but those still preserved consist of siltstone, shale or silicified mudstone.

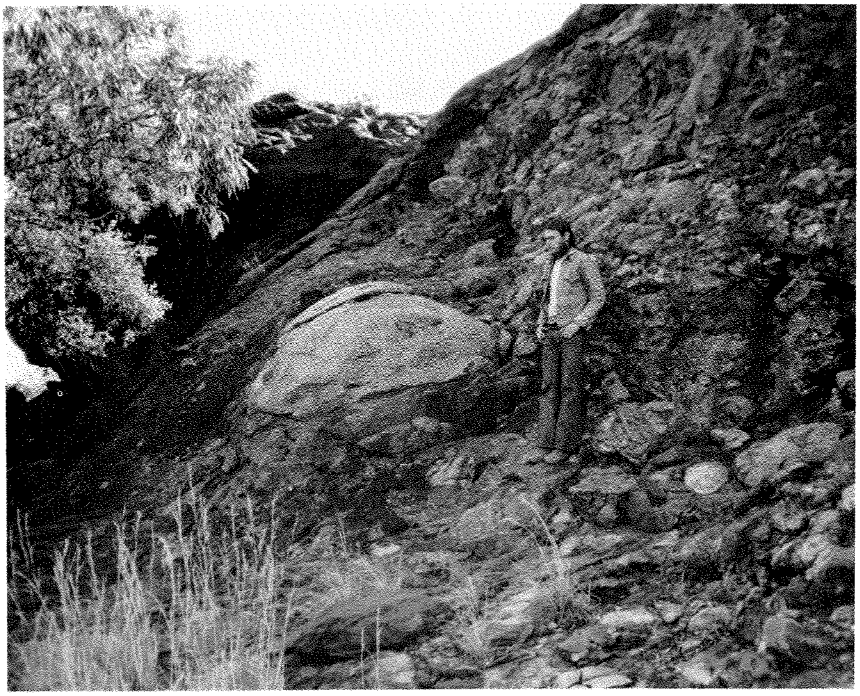
Pebbly and granule-bearing sandstone, conglomerate, siltstone, and shale, form a small proportion of the formation, generally as interbeds 0.1 m thick, though some are much thicker. Clasts are normally well rounded to subangular, and are composed of vein quartz, quartzite, light and dark chert, jasper, siltstone and agate. The most spectacular conglomerate occurs on the northeast corner of ROBERTSON, in at least two beds along the faulted margin of the formation north from Coondra Coondra Spring. These beds contain rounded boulders, up to 2 m across, of older conglomerate, sandstone, vein quartz and varicoloured cherts (Fig. 14). A glauconitic sandstone occurs near the base of the formation in the Kumarina district, and this was mistakenly described by Chapman (1933) as containing Cretaceous microfossils. Fine-grained dolomite is present 26 km north of Lake Wilderness, and de la Hunty (1969) mentions sandstone on ROBERTSON in which an original carbonate matrix is replaced by silica.

Cross-bedding is usually abundant, but, although sets more than 2 m thick are uncommon, they are a feature of the formation. Such large sets exist at localities 25 km northeast of Bullen Hill (up to 6 m thick), 15 km northwest of Yanneri Lake (2.5 m), 30 km north of Yanneri Lake (3.5 m), two localities near Kahrban Creek (2.5 m), 5 km north of Bocrabee Hill, and in the adjacent corner of RUDALL (2.5 m), and in the Carnarvon Range-Mount Davis area (8 m).

Carnarvon Range-Mount Davis sequence: This area features large and spectacular cross-bedding in a unit over 60 m thick, which starts about 40 m above the sub-Bangemall unconformity. Individual cross-bed sets range in thickness from 0.5 to 8 m (Fig. 15), but most are less than 2 m thick. In longitudinal sections the foresets are planar to tangential in geometry as in omicron cross-bedding, but in plan or transverse view they are seen to be very broad, shallow troughs up to about 250 m wide, in which the strike of the foresets can change through 70° of azimuth from side to side. The troughs scour into underlying sets forming sharp cut-off surfaces at low angles (5°-20°) to earlier laminae (Fig. 16). In plan, some sets have straight foreset traces for about 200 m, which curve sharply upward at one end within a distance of 1 m just before being cut-off by an overlying set; these are probably troughs, open at one end (scoops of Allen, 1963). The foreset laminae vary from 1 to 30 mm thick, and usually consist of coarse-grained, moderately sorted quartz arenite without feldspar. Medium-grained and granule-bearing layers are common, though not typical.

Between this cross-bedded interval and the basal unconformity is a variable sequence. A hill 1.5 km south of Mount Davis displays a 40 m cliff section of planar laminated, medium- and coarse-grained quartz arenite at the base of which is a 2 m thick conglomerate containing clasts, up to 0.5 m long, of vein quartz and schist derived from the underlying basement. The planar-bedded unit is scoured to depths of over 1 m at its top contact.

A



B

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Figure 14. Boulder conglomerate in the Calyie Sandstone near Coondra Coondra Spring, ROBERTSON. A—Exposed 2 m boulder. B—Typical crude bedding.

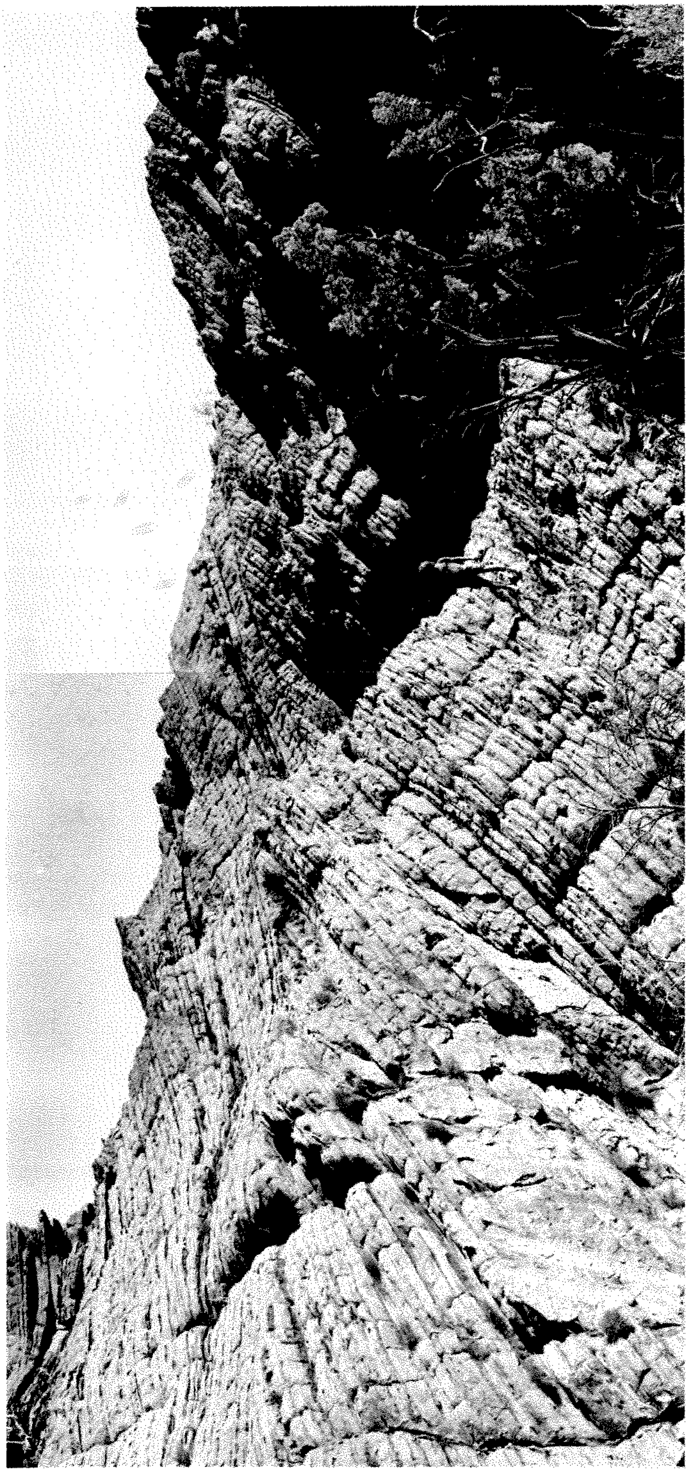

Figure 15. Large-scale cross-bedding in the Calyie Sandstone of the southern Carnarvon Range (25°15′40′′S, 120°38′20′′E). Sets are up to 8 m thick.

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Figure 16. Sharply defined scour contact between sets of large-scale cross-bedding in Calyie Sandstone near Mount Davis, NABBERU. Compass gives scale.

At Mount Davis itself, only the equivalent of the basal planar-bedded unit is present. Scattered, rounded pebbles of vein quartz and cherty siltstone occur singly throughout the sequence, or as ill-defined lenses and mono-pebble layers. Medium-scale pi cross-bedding is common. In a similar sequence near Trig Station M6, conglomeratic phases are more common and contain some clasts of an older conglomerate which itself contains pebbles of granular iron-formation from the Frere Formation of the Nabberu Basin.

A section at the southwestern end of the Carnarvon Range is exclusively sandstone, in which symmetrical ripple marks are locally abundant. This sequence is planar-bedded with minor alpha and omicron cross-bedding, and it grades upwards, by the increase in frequency and thickness of cross-bed sets, into the large-scale cross-bedded sequence. The lithology is laminated, coarse-grained, moderately sorted quartz arenite with lesser granule-bearing and medium-grained laminate.

ILGARARI FORMATION

Derivation of name: Ilgarari mine (lat. 24°22′S, long. 119°34′E).

Type area: The vicinity of Mardinidya Well, 12 km northwest of the Ilgarari mine. No type section measured.

Lithology: Siltstone, claystone, shale, and fine-grained sandstone.

Thickness: Unknown, because top of formation is not preserved. Measurements from air photographs suggest at least 650 m in the type area.

Stratigraphic relationships: Conformably overlies the Calyie Sandstone. Lateral equivalent of the Kurabuka Formation.

Description: The unit occurs only in the centre of synclinal areas west of the 120°E meridian, such as around Ilgarari, the Calyie Syncline, and the northern side of the Gascoyne River on PEAK HILL. It generally forms poor outcrop.

The dominant rocks are laminated and thinly bedded siltstone, claystone, and shale, which weather to a variety of whitish, brownish and yellowish colours. Beds of hard, siliceous, coarse-grained siltstone and fine-grained sandstone are subordinate. Buff-brown, bedded and faintly laminated chert occurs near the base of the unit at Mardinidya Well.

DIEBIL SUBGROUP

Derivation of name: Diebil Hill (lat. 23°37′S, long. 122°23′E).

Constituent formations: Skates Hills Formation and McFadden Sandstone.

Thickness: Unknown, but may be at least 1 000m.

Distribution: Northeastern Bangemall Basin, southwest of Lake Disappointment, from the Horsetrack Range to the Cornelia Range.

Stratigraphic relationships: Unconformable on the Cornelia Sandstone and the Yeneena Group, and in turn unconformably overlain by the Proterozoic Durba Sandstone and the Permian Paterson Formation. Westwards, it is equivalent to the Calyie Sandstone of the Collier Subgroup, into which it grades laterally.

SKATES HILLS FORMATION

Derivation of name: Skates Hills (lat. 24°34′S, long. 123°19′E).

Type Section: Lat. 24°33′15″S, long. 122°42′15″E, about 14 km north of the Cornelia Range (Brakel and Leech, 1979). The basal contact is not exposed.

Lithology: Dolomite, sandstone, shale, siltstone and conglomerate.

Thickness: About 50 m (estimated) in type section.

Age and stratigraphic relationships: Unconformably overlies the pre-Bangemall Cornelia Sandstone with an angular difference of up to 100°. The unit passes conformably upwards into the McFadden Sandstone, and the boundary between the two is taken as the top of the highest persistent dolomite member. It lenses out westwards.

The stromatolite *Acaciella* form indet. cf. *australica* is prolific in some beds and the same 'genus' has also been described from the Woolnough Hills diapir (Officer Basin). *A. australica* is considered to have an age range of Late Proterozoic (ca. 900 m.y.) to Early Cambrian (Preiss, 1976). This may mean that the easternmost Bangemall Basin is 200 m.y. younger than the central part, but we believe it is more likely that this occurrence extends the age range of this stromatolite back to the Middle Proterozoic. Grey (1978) likewise calls it the oldest known occurrence of this fossil group in Australia.

Description: The Skates Hills Formation is found in hills in the southeast quadrant of TRAINOR and in the Skates Hills (MADLEY).

Conglomerate lenses are developed sporadically at its base; the most spectacular one, at Phenoclast Hill (lat. 24°36'S, long. 122°40'E), contains rounded cobbles and boulders of pre-Bangemall quartz arenite and, occasionally, grey chert. This deposit is at least 30 m thick. These lenses are overlain by a sequence of fine-, medium- and coarse-grained, moderately sorted quartz arenite, with phases of wacke, shale and occasional glauconitic wacke-mudstone laminite.

The overlying interval comprises dolomite beds intercalated with soft, pink, dolomitic siltstone. The thickest dolomite is 10 m thick, but, along strike, units are thinner and tend to be more numerous; four beds, up to 3 m thick, occur at one locality. The dolomite is commonly laminated and bedded, and contains prolific columnar stromatolites in at least three zones. Below some of these zones, algal lamination, which locally contains irregular, thin nodules of chalcedony up to 15 cm long, resembling birdseye structure, is developed. There are also rare cubes of goethite after pyrite, up to 1 cm across. The youngest dolomite bed is overlain by siltstone, and then sandstone, of the McFadden Sandstone.

McFADDEN SANDSTONE

Derivation of name: McFadden Range, 25 km west of Lake Disappointment.

Type area: McFadden Range (lat. 23°21'30"S, long. 122°18'30"E) (Williams and Williams, 1977). The unconformity at the top of the formation is best exposed for about 20 km along the western side of the Durba Hills.

Lithology: Laminated, flaggy and massive, feldspathic quartz arenite and wacke, conglomerate, siltstone and claystone.

Thickness: Unknown, but may be at least 1 000 m.

Stratigraphic relationships: Rests with strong angular unconformity on the Yeneena Group and Cornelia Sandstone except on part of the southeast quadrant of TRAINOR, where it is conformable on the Skates Hills Formation. It is unconformably overlain by the Proterozoic Durba Sandstone and outliers of the Permian Paterson Formation. To the west it appears to be coeval with the Calyie Sandstone, the boundary between the two being arbitrarily taken where the McFadden Sandstone gradually loses its characteristic flaggy nature. On the west side of GUNANYA around Savory Creek it seems to overlie the Calyie Sandstone in part.

Description: The McFadden Sandstone occupies an area extending from RUDALL to TRAINOR, southwest of Lake Disappointment. It crops out in low hills, or in the sides of rugged mesas and buttes.

In the type area, the formation consists predominantly of laminated, flaggy sandstone, which is typically feldspathic and lithic and which has a little mica. Silt and clay are common matrix constituents, and many outcrops comprise interlaminated arenite and wacke. The proportion of feldspar (10-20 per cent) and lithic grains (chert and jasper) varies. Scattered pebbles and granules are prevalent in some beds. The sand ranges from fine to coarse, and the sorting from very good to poor. The sandstone is whitish or grey when fresh, but weathers to brown, or a white-speckled, purplish banded appearance when feldspathic. Thick, massively bedded arenite is intercalated in the sequence.

Southwards, all the distinctive features except flagginess are lost as the rocks grade to better sorted quartz arenites. Northwards there is also a trend to better sorted and less feldspathic sandstones, but lithic detritus remains common and some poorly sorted beds are still present, and increasing proportions of massive sandstone are interbedded with the flaggy rocks. The gradation westwards into the Calyie Sandstone is similar to that southwards, with flagginess in addition being lost. Large-scale pi cross-bedding, with broad scoop-shaped sets up to 8 m thick (Fig. 17), is abundant. Beds with siltstone and claystone intraclasts are widespread. Weathered pyrite balls 1 cm across are recorded from one outcrop.

A. D. T. Goode and W. D. M. Hall (B.H.P. Co., pers. comm.) have observed volcanic grains as major and minor components in thin sections of sandstone from the Skates Hills-McFadden Range area. These grains are embayed, corroded, and broken quartz euhedra with distinctive scalloped margins which could not be due to abrasion. In specimens taken from the same area during Geological Survey of Western Australia mapping, such grains are rare (Brakel and Leech, 1979). It would appear that these sandstones received a contribution of unevenly distributed ash-fall tuffs from distant volcanic activity.

There are subordinate beds and lenses of conglomerate, siltstone, and claystone. Clasts in the conglomerate and pebbly sandstone range up to cobble size and consist of quartz, chert, sandstone and siltstone/pelite, with lesser granite and jasper. The

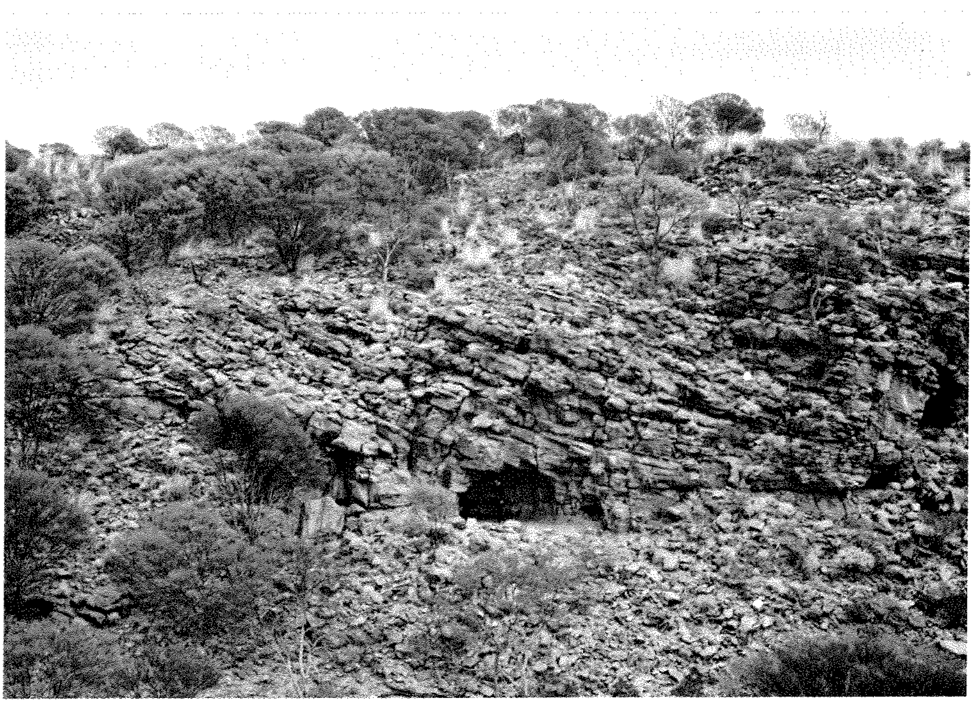

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Figure 17. McFadden Sandstone (24°21′00″S, 122°39′30″E) showing 8 m-thick cross-bedding.

coarsest conglomerates, which contain boulders up to 0.6 m across, are developed in northwestern GUNANYA (Williams and Williams, 1977) where they occur in channel-like bodies with erosional bases. The deposits can be traced in outcrop for lengths of over 100 m, and they are probably related to the even coarser conglomerates in the Calyie Sandstone at Coondra Coondra Spring 60 km to the west-northwest.

A small exposure of grey, laminated, fissile calcareous siltstone and shale exists alongside a salt pan marginal to Savory Creek (lat. 23°21′S, long. 122°15′E). The beds dip eastwards beneath the McFadden Sandstone, and are either an interbed of the formation or an equivalent of the Skates Hills Formation 150 km to the south-southeast.

KAHRBAN SUBGROUP

Derivation of name: Kahrban Creek.

Constituent formations: Coonabildie Formation and Marlooyanoo Formation.

Thickness: The composite thickness of the formations is a minimum of 1 700 m.

Distribution: Lies in the southeast corner of the Bangemall Basin between the Cornelia Inlier and the Nabberu Basin.

Stratigraphic relationships: Lies unconformably on rocks of the Nabberu Basin. Its relationship to other units, and even the relations between its constituent formations, are uncertain because contacts are obscured, so that it may possibly be pre-Bangemall. However, it is interpreted to be unconformable on the 10 km-thick, near-shore marine succession of the Scorpion Group, and conformable beneath the Calyie Sandstone. The contact between the Coonabildie and overlying Marlooyanoo Formation is similarly believed to be conformable (Commander, and others, 1979).

COONABILDIE FORMATION

Derivation of name: Coonabildie Range (lat. 25°41′S, long. 122°42′E), about 29 km west-northwest from Carnegie homestead.

Type area: From Coonabildie Range west to Mount Moore.

Lithology: Dominantly siltstone with lesser medium-grained sandstone; subordinate shale, chert and carbonate.

Thickness: No complete section measured, but estimated from air photographs to be a minimum of about 1 100 m.

Stratigraphic relationships: The lower contact is obscured but is thought to be an unconformity on the Earaheedy Group of the Nabberu Basin. It is a structurally simple sequence which cuts across the complex structure of the Earaheedy Group, and aeromagnetic mapping (Wyatt, 1977) shows the Earaheedy extending in the subsurface beneath the Coonabildie Formation at Mount Moore. The Coonabildie Formation is unconformably overlain by the Permian Paterson Formation, in the type area. The upper contact against the Marlooyanoo Formation is obscured, but the two units are probably conformable.

Description: The Coonabildie Formation occurs mostly on STANLEY with small areas on the adjacent sheets to the north and west. Quartz arenite forms cuestas and ridges, and siltstone underlies the escarpments and scree slopes. The Coonabildie Formation comprises interbedded siltstone and quartz arenite overlying a distinctive basal member of the cream-weathering, laminated and bedded chert. In the lower half of the formation sandstone bands are commonly less than 2 m thick and comprise less than 30 per cent of the sequence. This proportion increases towards the top, where sandstone bands are thicker and siltstone is subordinate. Siltstone is typically micaceous, purple in fresh exposures, and weathers to yellow and purple. Discontinuous laminae of fine-grained sandstone are common as well as subordinate interbeds of shale. In the area south from Glenayle homestead, the typical siltstone sequence is absent, and dark-green siltstone and shale contain beds, up to 3 cm thick, of fine- and medium-grained crystalline limestone as well as green, fine-grained sandstone laminae. Carbonate beds also occur at the top of the formation 27 km east-southeast of Glenayle.

The quartz arenite is medium grained with plentiful siltstone pellet molds, and with bedding 3 mm to 0.3 m thick. Most of the arenite is well sorted, pale grey, and has less than 5 per cent clay feldspar. A few bands are green, moderately sorted, and contain a few lithic fragments. Arenite bands in the lower half of the sequence contain discontinuous laminae of moderately sorted, coarse-grained arenite and granule conglomerate. Planar cross-beds are plentiful throughout the sequence. Ripples, current lineations, current crescents, flute casts and channels are abundant locally. Mud cracks and ice crystal casts occur in the type area (Figs 54, 55).

MARLOOYANOO FORMATION

Derivation of name: Marlooyanoo Hill (lat. 25°20′30″S, long. 122°30′30″E), about 58 km north-northwest from Carnegie homestead.

Type area: Between Marlooyanoo Hill and Lyn Bore on STANLEY.

Lithology: Siltstone and shale, usually laminated, and arenite, subordinate carbonate near the top.

Thickness: Estimated from air photographs to be at least 600 m.

Stratigraphic relationships: The upper and lower boundaries with other units of the Bangemall Basin are not exposed, but the Marlooyanoo Formation is inferred to lie conformably between the underlying Coonabildie Formation and the overlying Calyie Sandstone. It is overlain unconformably by the Permian Paterson Formation.

Description: The Marlooyanoo Formation is restricted to STANLEY. In the type area, where the sandstone is subordinate, it is composed of a series of upward-coarsening cycles. Each cycle starts with shale, but there is an upward increase in the proportion of interlaminated siltstone and fine-grained arenite, until the shale is capped with either fine- or medium-grained arenite. The siltstone is micaceous and varies from purple, where comparatively fresh, to yellow and pink, where more weathered. The weathered shale ranges from green-grey to white. A feature of most of the siltstone and shale is a regular lamination which only rarely has small scour surfaces (Fig. 57) and contorted zones.

The arenites mostly form a small proportion of the sequence with beds no more than about 3 m thick. Locally, such as just northwest from Carooil Bluff, they comprise the bulk of a zone at least 100 m thick. The arenite is typically yellow-brown where weathered, but green where fresh. It is well sorted and contains 5 to 15 per cent feldspar and up to 5 per cent interstitial clay. Grains are subrounded to subangular. The rock is usually well laminated due to regular variation in clay content. Elsehwere the bedding is 0.1 to 1 m thick. Both planar and trough cross-beds (Fig. 56) are well developed and have sets up to 1 m thick. Distorted trough cross-beds (Fig. 58) and cross-cutting zones of convolute bedding are common. There are some ripple marks.

MANGANESE SUBGROUP

The Manganese Subgroup occupies the region from Robertson Range homestead to the Davis River, and was named as a group by de la Hunty (1964) on BALFOUR DOWNS, presumably after Manganese Bore, 13 km west-northwest of Jiggalong. The Coondoon Conglomerate, Bee Hill Sandstone, Balfour Shale and Noreena Shale, with a composite thickness of about 525 m, were given as the constituent formations. The undifferentiated 'Manganese Group' was later mapped on adjacent ROBERTSON (de la Hunty, 1969), but recent work has shown that this unit is the lateral equivalent of the Backdoor Formation of the Bangemall Group, except for the Wadara Range area which is occupied by the older Yeneena Group. The type localities of the constituent formations of the 'Manganese Group' have not been examined since the original mapping by de la Hunty (1964), and the names are retained pending remapping of BALFOUR DOWNS. These formations probably form part of the Bangemall Basin, and we regard de la Hunty's 'Manganese Group' as a subgroup of the Bangemall Group.

The Manganese Subgroup was laid down unconformably on the Yeneena Group, rocks of the Hamersley Basin, and Archaean rocks. The contact with the overlying Calyie Sandstone is usually faulted, but a conformable transition between the two can be seen in a few places, such as Watch Point.

COONDOON CONGLOMERATE

The Coondoon Conglomerate is named after its type locality, immediately north of the junction of the Coondoon Creek with the Davis River. It lies on the basal unconformity and ranges from 0.15 to 15 m thick. It comprises conglomerate containing angular fragments and boulders, and interbedded, sandy, ferruginous, and manganiferous shales. The formation is known only from central northern BALFOUR DOWNS.

BEE HILL SANDSTONE

Overlying the Coondoon Conglomerate, and somewhat difficult to distinguish from it, is the Bee Hill Sandstone. On ROBERTSON it is intermittently developed at the base of the subgroup. The maximum thickness is "possibly 250 ft" (76 m), 28 km west of Jiggalong (de la Hunty, 1969). In the type locality, 3 km southeast of Bee Hill and extending south for 10 km, it contains pebble bands, granule conglomerate, sandstone and sandy shale, and a distinctive type of sandstone which has prominent red to purple and white bands caused by heavy-mineral layers repeated every 5 to 10 mm. This rock also has cross-bedding, in sets about 0.6 m thick, and white spots up to 25 mm in diameter. Where the unit lies on the basal unconformity, it contains a few boulder-bearing phases equivalent to the Coondoon Conglomerate.

BALFOUR SHALE

The Balfour Shale is widely distributed over the whole of the Manganese Subgroup. It overlies the lower formations, or, in places, the basal unconformity. It is well exposed at the type locality immediately northeast of Mount Trew and is named after Balfour Downs homestead. The shale is typically green and laminated, but varies to grey green and grey; it also varies in composition from micaceous to sandy and calcareous. It also contains some manganese. Some chert beds and thin-bedded glauconitic sandstone are present. On ROBERTSON, thin-bedded dolomite occurs near the base, but some dolomite outcrops in this region could be inliers of pre-Bangemall formations. The thickness exceeds 150 m; a figure of 1 500 feet (460 m) given by de la Hunty (1969) is from an area now suspected to be Fortescue Group.

NOREENA SHALE

The Noreena Shale is limited in distribution to central and northern BALFOUR DOWNS, where it conformably overlies the Balfour Shale. The formation is named after its type locality near Noreena Creek. It is about 30 m thick.

The prevailing colour of the Noreena Shale is chocolate, but parts are purplish to red-brown, because of iron and manganese oxides. The rock ranges from blocky shale to siltstone, and some is micaceous. The outstanding feature of much of the shale is the presence of pellets of primary braunite along the bedding planes. These pellets range in diameter from 2 to 25 mm and average about 5 mm.

DURBA SANDSTONE

Derivation of name: Durba Hills, 30 km southwest of Lake Disappointment.

Type area: Northern end of Durba Hills, lat. 23°43′00″S, long. 122°29′30″E (Williams and Williams, 1977).

Lithology: Quartz arenite with conglomerate lenses at the base.

Thickness: The top of the formation has been removed by erosion. About 100 m is present in the Durba and Diebil Hills.

Age and stratigraphic relationships: Unconformable on the McFadden Sandstone. The contact is usually an angular unconformity, but locally approaches paraconformity. This erosion surface has a present day extent of 140 km^2 and appears to be the base of a very shallow basin aligned parallel to structural trends in the older rocks of the Paterson Province to the northeast. The angular discordance between the two formations is at least in part due to a large-scale cross-bedding in the McFadden Sandstone, so that the duration and significance of the time break between the two is uncertain. The Durba Sandstone though post-Bangemall, is probably still of Middle Proterozoic age. It represents a last, brief return of deposition at the close of Precambrian sedimentation in the region.

Description: The formation is a cliff-forming sandstone which caps a series of scattered mesas, buttes and hills southwest of Lake Disappointment.

The rock is a massive, well-sorted, medium- to coarse-grained quartz arenite, with scattered pebbles and cobbles of quartz in places. Most beds are over 0.3 m thick and some are cross-bedded. Feldspar is accessory, rarely comprising as much as 5 per cent. The fresh arenite is white, but weathers externally to orange-brown. Intraclasts and shale interbeds are uncommon.

Lenses of conglomerate occur in erosional scours at the base of the unit. They are usually less than 0.5 m thick, but may range up to 2 m. The clasts include well-rounded pebbles and cobbles of quartz, quartz arenite, grey chert, and rare mudstone, jasper and oolite.

CHAPTER 4

Sedimentation

The sedimentary development of the Bangemall Basin has varied markedly from region to region. The major facies variations, which are identified as the western, northern and eastern facies, are reflected in the subdivision of the stratigraphic succession into formal subgroups. Except for some basal lenses, the sequence is marine. In future it may be possible to define distinct sub-basins, but, in the absence of geophysical and subsurface information, this is not attempted here.

The Discovery Chert, being an unusual formation of special interest, is treated separately in Chapter 5. The Manganese Subgroup in its type area on BALFOUR DOWNS has not been examined by the writers and is not discussed in this chapter.

ALLUVIAL-FAN DEPOSITION

TERRESTRIAL ALLUVIAL-FAN FACIES ON BASEMENT

The three, thick, wedge-shaped, coarse-grained units overlying the unconformity at the base of the Bangemall Group in the western half of the basin have the lithology and geometry of alluvial fan deposits (Table 4). Coarse-grained, moderately to poorly bedded sandstone dominates over lesser grain sizes, and bands of conglomerate and pebbly sandstone, which indicate closeness to the source area, are common throughout. This is corroborated by the angular to subrounded nature of much of the detritus, the moderate to poor sorting of some beds, and the presence of feldspar and lithic clasts which vary in abundance and composition according to the nature of the nearby basement. Water-rounded cobbles and boulders, over 0.6 m across, occur scattered through some sandstone beds, and, as they lack the extreme angularity of talus, point to alluvial fan deposition.

The conglomerates in the Mount Augustus Sandstone are usually ill-defined lenses of granule conglomerate that comprise a small proportion of the sequence; coarse, thick fanglomerate, typical of a fanhead (Bull, 1964), is lacking. Isolated megaclasts, particularly in the Mount Augustus Sandstone, probably represent boulders which were washed down the fan slope. The complete absence of lutite at Mount Augustus argues against a fan base; thus a mid-fan environment is interpreted. Siltstone and fine-grained wacke lenses, probably laid down in backwaters of braided streams farther down the fan, are interbedded to the southwest of Mount Augustus. In the same area, trains of symmetrical ripples with wavelengths of less than 50 mm, indicate water only centimetres deep (Tanner, 1971). Some of these are interference ripples, which consist of two ripple sets of different wavelengths, and were formed by successive wind-generated surface-wave trains from different directions. The sequence along the Cobra Synclinorium is dominated by medium-grained sandstone, suggesting either distal fan regions, or smaller fans flanking lower topography.

TABLE 4. DEPOSITIONAL FEATURES AND INTERPRETATION, ALLUVIAL-FAN FACIES

Feature	*Interpretation*
Coarse, pebbly sandstone sequences, poorly bedded moderately to poorly sorted with conglomerate lenses; regional wedge-shaped geometries	Mid-fan deposition on alluvial fans
Lithic clasts and feldspar; many angular and subangular clasts	Nearby source area
Rounded pebbles, cobbles and boulders	Water action
Siltstone lenses in sandstone sequence	Distal fan deposition, overbank deposits
Scour channels	Stream action on fans
Poorly sorted siltstone and wacke layers with erosional bases containing granules and pebbles	Subaerial sheet-wash mud-flow deposits laid down during waning stages of floods on fans
Alpha and omicron cross-bedding in sandstone	Straight-crested sand waves in stream channels
Pi cross-bedding in sandstone	Lunate ripples in stream channels
Small symmetrical ripples and interference ripples	Very shallow water in channel pools; wind-generated waves impinging on bottom sand
Dolomitic and shale lenses towards tops of formations	Marine interbeds in fluctutating marine-terrestrial conditions, where fans discharged into sea
Pyrite in some beds towards tops of formations	Areas of anaerobic conditions during sedimentation, probably marine or brackish
Load casts	Compaction structures

Sparse cross-bedding data from various levels of the Mount Augustus Sandstone indicate a southeasterly palaeocurrent direction, but one outcrop to the southwest shows a northwesterly trend. This may indicate fans on opposite sides of a basement high. The cross-beds themselves are medium-scale (Conybeare and Crook, 1968) alpha, omicron, and pi types, consistent with formation in streams. They are outline by laminations of dark heavy-mineral grains indicating grain transport as concentrations in ripple troughs.

The Tringadee Formation has a greater incidence of siltstone interbeds, some of which are 0.2 m thick, poorly sorted, have erosional bases, and contain scattered granules and pebbles. These are regarded as mud-flow deposits, similar to those described from alluvial fans by Blissenback (1954). Small scour channels (Fig. 18) are also present between beds of ubiquitous granular, pebbly, and moderately sorted arenites (Fig. 19). Other sedimentary structures are current crescents, ripples and medium-scale planar and trough cross-bedding, some of whose foresets are outlined by dark heavy-minerals. The evidence of strong current activity, the predominantly pebbly sandstone sequence, the mudflows, and the rapid lateral thinning of the formation in all directions, confirm an alluvial-fan environment. Foreset dip directions are only available from a few outcrops, but are mostly consistent with a provenance in older rocks to the south. In some narrow zones the formation may have been deposited in valleys in the basement, with alluvial contributions from either side. The core of basement in the Sullivan Anticline has not influenced current directions, and appears not to have been a topographic high at the time of deposition.

The top portion of the Tringadee Formation, with its lenses of dolomite and dolomite-cemented clastics, marks the onset of marine conditions. The apron of the alluvial fan is thought to have debouched into the sea, where wave action reworked some sand into fringing barrier bodies, behind which carbonate precipitated in sheltered lagoons. Pyrite in some carbonate-poor bands marks zones which were anaerobic. The coarser arenite and conglomerate bands in this top sequence suggest that the position of the shoreline fluctuated as the fans advanced and retreated because of varying sediment discharge or sea-level changes.

The Coobarra Formation differs from the other two units in having a larger proportion of conglomerate, locally up to 50 per cent of the sequence. Clasts of jaspilite-bearing conglomerate and quartz arenite show that early Proterozoic sediments form part of the provenance. These source rocks are likely to be part of the same sequence as the pre-Bangemall rocks exposed to the west in the faulted core of the Coolina Anticline. Festoon (pi) cross-bedded pebbly sandstones attest to quite violent transport conditions.

Cross-bedding directions (Fig. 20), measured from alpha and omicron foresets and pi trough-axes, display average southwesterly directions in the western half of the outcrop area, indicating a topographic high to the north of the Bujundunna Fault. A southeasterly trend, obtained from farther east and higher up the

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Figure 18. Cross-bedding and scour channels in coarse, pebbly sandstone of the Tringadee Formation 15 km south-southeast from Waldburg Station, MOUNT EGERTON.

sequence, may represent the eastern flank of the fan apron. Here conglomerate lenses and pebbles are rare, and the sequence consists of very thickly bedded, massive or faintly laminated, feldspathic lithic arenite, and thin siltstone and shale layers. The sandstone itself grades upwards from fine grained to coarse grained and granule bearing, before some fine- to medium-grained interbeds re-appear. The only sedimentary structure is rare alpha cross-bedding in sets up to 1 m thick. This sequence may represent part of an advancing alluvial fan, supplied mainly with sand-sized detritus.

Overlying the Coobarra Formation is a 200 m-thick siltstone-shale unit, which is laterally equivalent to dolomite lenses and is probably marine; smaller lutite lenses occur below this level. The situation was similar to that at the top of the Tringadee Formation, except that lutite instead of dolomite was deposited in fringing lagoons. Some lutite lenses, of course, may have been laid down in braided-

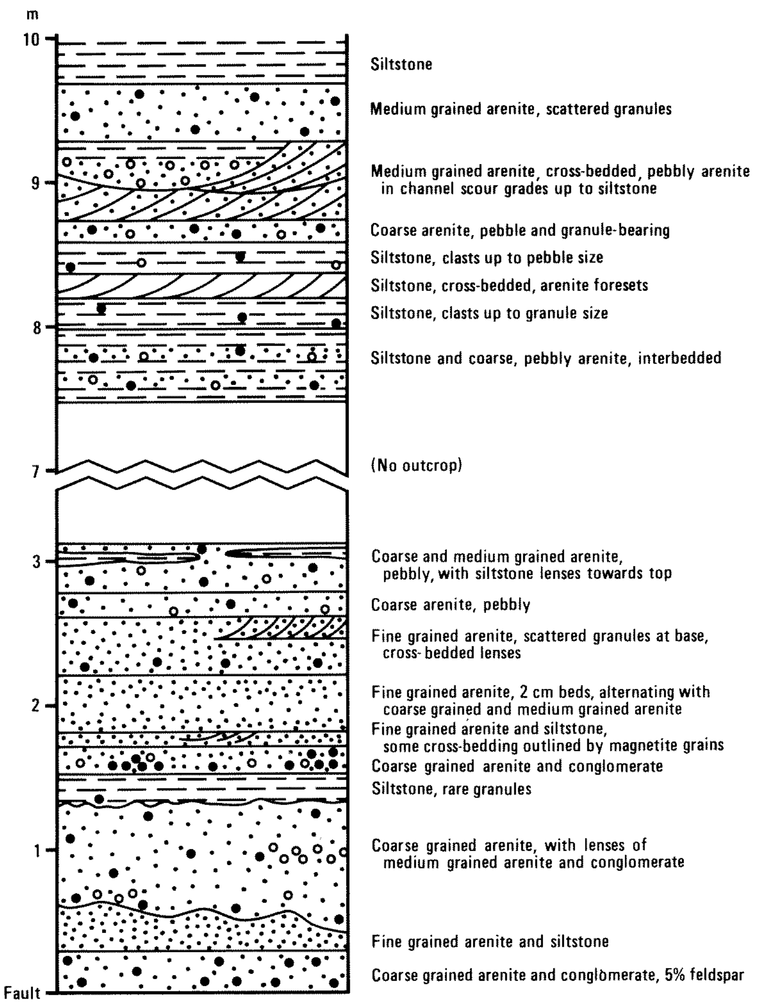


Figure 19. **Measured section of part of the Tringadee Formation near its base, showing scour channels filled with pebbly arenite and mud-flow deposits of pebbly siltstone.**

stream backwaters, or as small overbank deposits. A few thin bands of fine-grained wacke and sandy siltstone in this upper sequence may record subaerial sheet-wash mudflows.

The overlying, more mature, marine Kiangi Creek sands are inferred to be derived from the alluvial fan detritus by reworking in turbulent surf environments of beaches and barrier bars.

ALLUVIAL-FAN FACIES ADVANCING INTO LAGOONAL ENVIRONMENT

The thick, 70 km-long wedge of conglomerate, arkose and quartz arenite in the Top Camp Formation near Prairie Downs has the typical geometry and lithology of an alluvial fan. The basal jaspilite-boulder conglomerate thins southwards, as does the unit as a whole, suggesting deposition controlled by activity along the Prairie Downs Fault. The composition of the detritus, which includes pseudomorphs after riebeckite, indicates derivation from the Hamersley Basin to the north. Accumulation began directly on the unconformity with the Bresnahan Group, but encroached upon lagoonal carbonates as the fan extended to the southwest. As deposition on the fan waned, the lagoonal sediments transgressed over it.

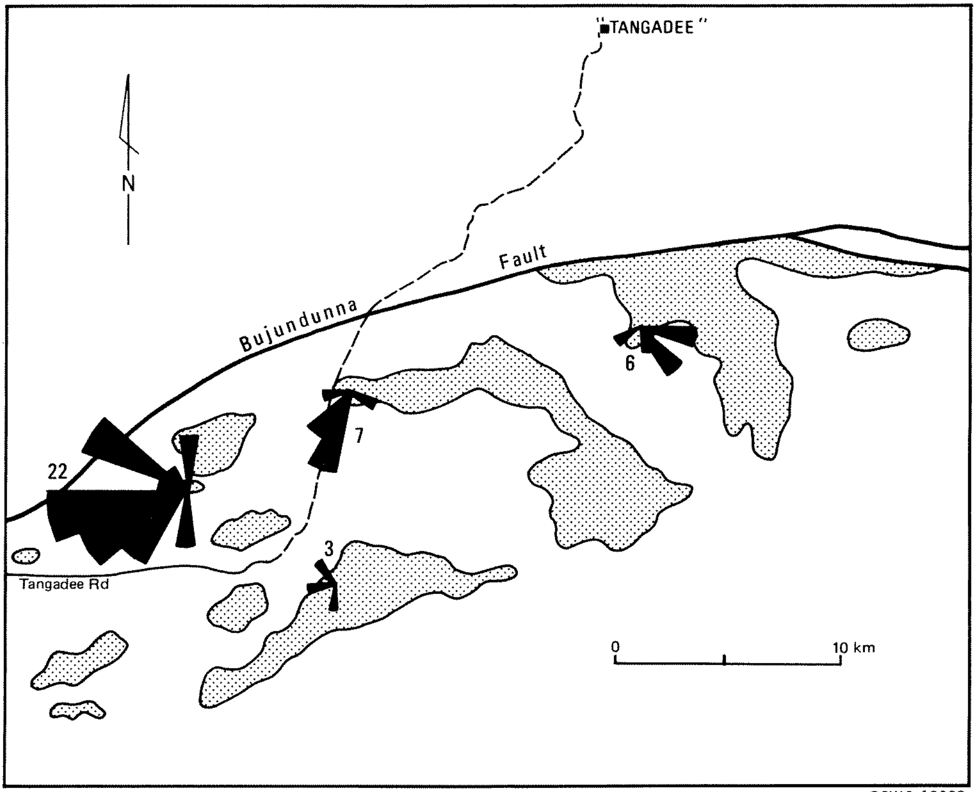


Figure 20. Palaeocurrent directions in the Coobarra Formation suggest there was a basement high north of the Bujundunna Fault. Rose diagrams are drawn according to the number of measurements in each 20-degree class interval. Total number of measurements for each locality is shown.

TRANSGRESSIVE LINEAR-SHORELINE DEPOSITION

The Top Camp - Irregully, Kiangi Creek - Wonyulgunna, and Jillawarra Formations correspond roughly to dolomite-lutite, arenite-lutite, and laminated lutite facies respectively. It is clear from extensive interfingering that these facies were deposited in adjoining environments, which migrated laterally to produce vertical sequences according to Walther's Law.*

The three, broad environments correspond well to the X, Y, and Z zones of the general theory of epeiric marine shelf sedimentation proposed by Shaw (1964) and Irwin (1965). The X zone is the open-marine shelf, where fine-grained sediments are deposited from suspension below effective wave base. The Y zone marks the high-energy environment of a barrier bar or beach system, and the Z zone corresponds to the sheltered lagoon and tidal flat area between the barrier bar and the mainland. The essential factors promoting these zones are high and low tide levels and the wave base, so that on shelves with gentle bottom gradients, the zones can be quite wide. The Y and Z zones tend to lie in linear belts parallel to the shoreline, and can be tens and hundreds of kilometres wide respectively (Selley, 1976, p.295). The X zone may cover thousands of square kilometres.

Each of these major facies is examined below.

LOWER LAGOONAL FACIES

This facies, developed in the western half of the basin (Fig. 27), includes most of the Top Camp and Irregully Formations, and the dolomitic portions of the Kiangi Creek Formation. It is characterized by fine-grained, laminated dolomite and lutite, but also contains subordinate coarser grained material. Its depositional structures are listed in Table 5.

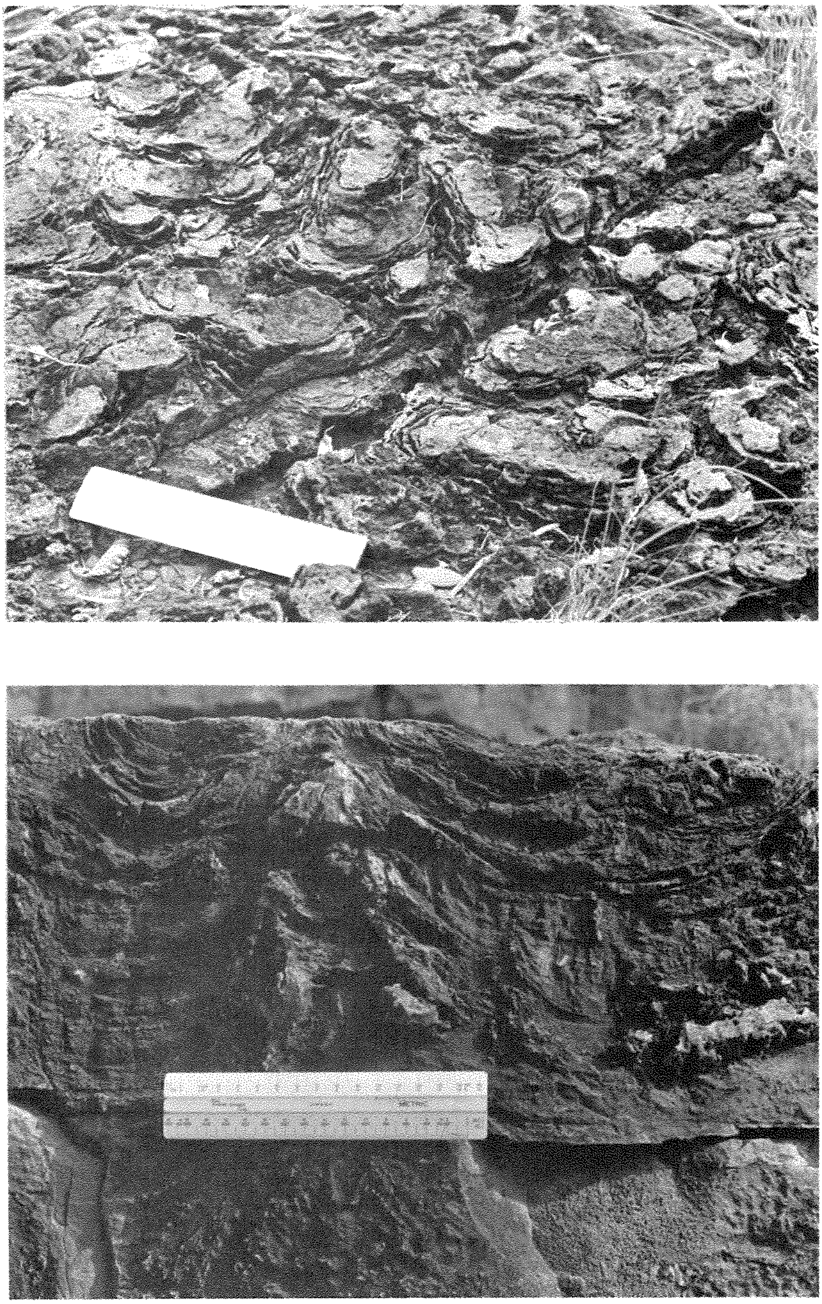
The abundance of carbonate over such a large area points unambiguously to marine deposition, whereas symmetrical ripples, interfering ripples, birds-eye structure (Grey, 1977), and desiccation cracks, attest to shallow water and sub-aerial exposure. Both high- and low-energy sub-environments are represented. These regimes are common to a fringing coastal lagoon system, in which most of the carbonate was laid down in quiet water, but in which some marginal tidal flats, beaches, shoals, and wash-over fans also occur.

Grey (1977) recognized a range of subtidal to intertidal environments which were conducive to the growth of particular stromatolite forms. A large colony of *Conophyton garganicum australe* in the Top Camp Formation 22 km northwest of Mount Vernon contains individual conical forms with diameters of up to 5.5 m. A cone of this size had a minimum growth relief of 9 m, which indicates the minimum water depth below wave base. Some huge bioherms are also developed in the Irregully Formation (Walter, 1972; Grey, 1977). Clear water was required by these algal and/or bacterial colonies, and the terrigenous content of the dolomite is, in fact, low.

*A conformable vertical sequence was generated by a lateral sequence of geographically related environments.

TABLE 5. DEPOSITIONAL FEATURES AND INTERPRETATION, LOWER LAGOONAL FACIES

Feature	*Interpretation*
Laminated, fine-grained dolomite with some interbeds of lutite	Quiet water in lagoons or in restricted bays and ponds; low clastic input at most times
Conophyton stromatolites	Quiet water, subtidal
Small cumulate stromatolites and algal lamination	Probably tidal flat environment
Baicalia stromatolites	Upper subtidal to lower intertidal regime; some formed large bioherms
Biostromes in outcrops at several levels	Repeated subsidence
Scoured stromatolite columns and edgewise conglomerate of algal dolomite	Erosion and redeposition by wave action in the intertidal zone
Breccia bands of algal dolomite fragments	High energy episodes of littoral erosion or undercutting of meaderbend bank; some bends may represent forereef talus from algal patch reefs
Rare oolites and oncolites	High energy conditions in intertidal zone, possibly in a channel
Polygonal saucer shaped structures (Fig. 21)	Cracked and curled-up algal mat of intertidal zone
Isolated quartz and lithic sand grains in dolomite	Clastics introduced by turbulent suspension or blown in by wind
Fine-grained muscovite in some dolomite	Terrigenous material which settled from suspension in calm conditions
Hematite clusters (up to 5 cm across) and pseudomorphs (up to 2 cm) after pyrite	Syngenetic or diagenetic pyrite formed in reducing conditions below the sediment
Occasional rounded hematite cubes after pyrite	Erosion of enclosing sediment, abrasion of cubes during transport, and re-deposition
Scour surfaces	Erosion in high energy events
Thin quartz arenite bands in dolomite	Storm sand layers
Symmetrical ripples, wavelength up to 8 cm (Fig. 22)	Wave-formed or modified ripples
Interfering sets of symmetrical ripples	Shallow water acted on by winds of variable direction
Desiccation cracks in mud infilled by sand, and in sand infilled by silt, up to 2 cm wide (Fig. 23)	Sub-aerial exposure
Rare, zoned cubic carbonate crystals in thin section	Replacement of evaporitic halite
Tepee structures (Fig. 21)	Salt crystallizing in intertidal sediments during evaporation, and deforming lamination by expansion pressure
Rare limestone lenses, bands and pods	Relicts of original carbonate sediment
Dolarenite lenses, some with slabs of fine-grained dolomite	Turbulent sub-environment e.g. tidal channels and shoals
Draping over scours in dolarenite	Suspension settling during slack-water period
Arenite, pebbly arenite, conglomerate and rare wacke interbeds	Intertonguing with shoal, beach, wash-out fan and barrier facies
Edgewise conglomerate of sandstone clasts	Beach regime
Small symmetrical ripples in sandstone (wavelength 25 mm)	Wind action on water only centimetres deep
Alpha and omicron cross-bedding in pebbly arenite	Sand wave deposition by strong tidal currents
Dolomitic sandstone interbeds	Diagenetic infiltration of carbonate into sand; some may be beach-rock



A

B

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Figure 21. Tepee structures in the Irregully Formation, Irregully Creek gorge. A—Polygonal saucer-shaped structure formed by desiccation cracking and curling up of algal mat. B—Cross section of tepee structures formed by expansion of sediment during periods of desiccation, wetting and crystallization of salts (Assereto and Kendall, 1977).

Polygonal, saucer-shaped structures (Fig. 21), in a bed overlying mudcracked ripples (Fig. 22), occur in the Irregully Formation. They closely resemble algal mat described by Kendall and Skipworth (1968), which becomes desiccated and cracked in the intertidal zone, curls up and traps sediment under the edges.

Examples of eroded stromatolites in the form of beds of stromatolitic breccia and scoured, *in situ* columns are known. These can be accounted for by eustatic sea level changes, storms, or the undercutting of tidal channel banks opposite a point bar. Interbedding of dolomite with sandstone, dolarenite, and granule and edgewise pebble conglomerate indicates an alternation with energetic marginal or channel conditions. The quartz arenite bands in dolomite are probably storm-sand layers (Reineck and Singh, 1973). Isolated, well-rounded pebbles of quartz, quartzite, and phyllite, up to 0.1 m long, were washed into deeper water during very turbulent episodes, but isolated quartz sand grains may have been blown in from adjacent beaches and barrier islands.

Modern examples of primary dolomite precipitation are all on a small scale (Pettijohn, 1975, p. 365), in contrast to the wide area of deposition in the Bangemall Basin, where alizarin red staining of dolomites reveals small grains of dispersed calcite; these may be original carbonate. There are a few examples of lenses and thin bands of limestone, where laminations in the enclosing dolomite bend around the lenses and are unrelated to those within the limestone, (Fig. 6). Therefore, the dolomite is considered to have been deposited as limestone.

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Figure 22. Symmetrical shallow-water ripple marks in sandstone showing desiccation cracks filled with fine-grained sandstone, Irregully Formation, Irregully Creek gorge.

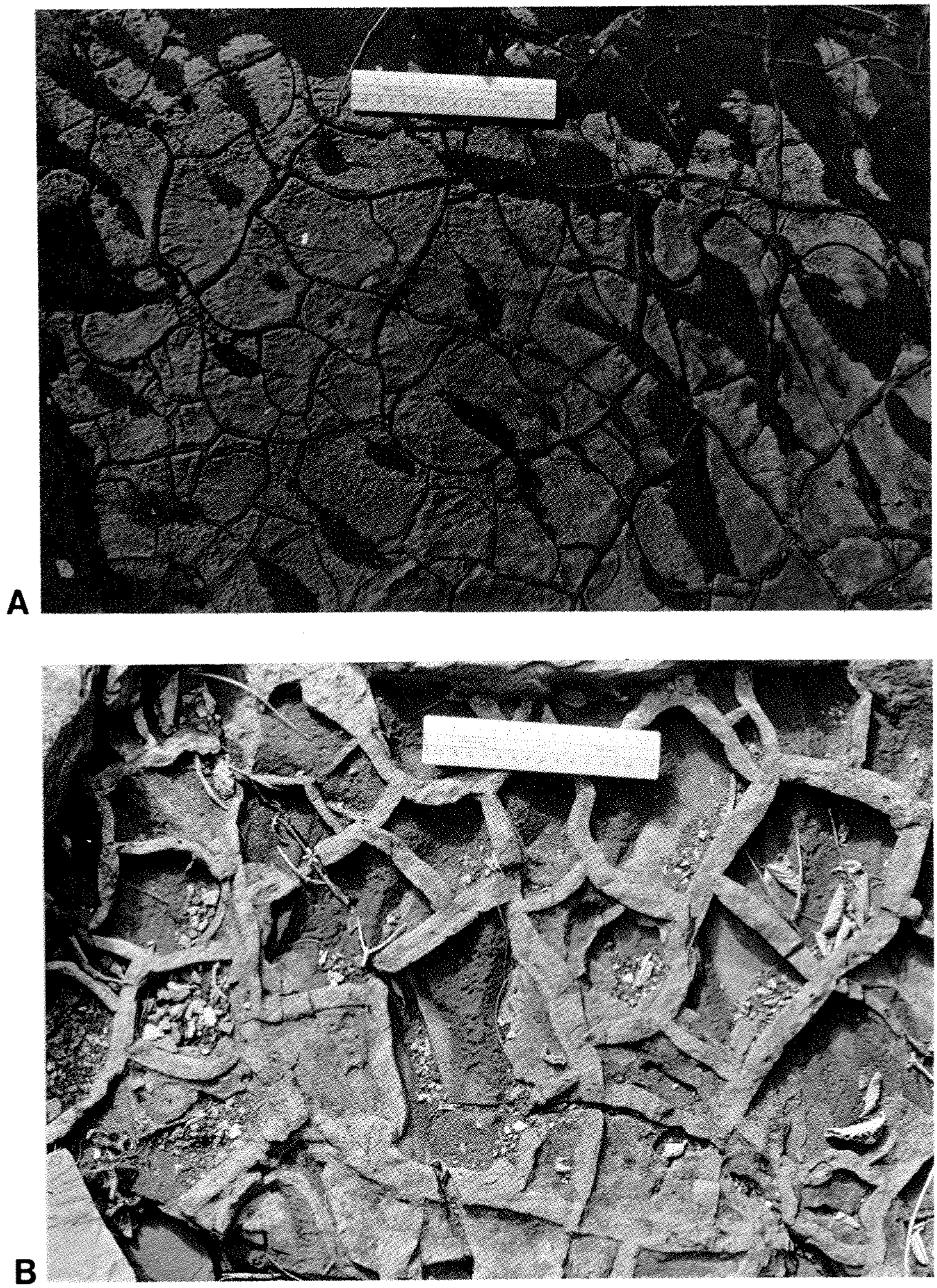

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Figure 23. **Desiccation cracks, Irregully Formation, Irregully Creek gorge. A—Cracks in dolomitic sandstone, infilled by mud. B—Cracks in sandstone, infilled by coarse silt. A sandy sediment such as this would only acquire desiccation cracks on subaerial exposure.**

Cross-bedding directions in pebbly arenites near the base of the sequence in Irregully Creek display bipolar northwest-southeast palaeocurrent trends, and symmetrical ripples higher in the sequence in this area are aligned to this direction. This is consistent with tidal surges moving along the basin axis, parallel to the shoreline, and implies that this part of the basin was not open to the southwest. To the west, Daniels (1966b) found that palaeocurrents in the basal Yilgatherra Member, which is probably a transgressive, sandstone-conglomerate shoreline deposit, were deflected around the Henry-Telfer Granite, which must have formed a peninsula, an island, or a mound along the margin of the basin. Elsewhere on EDMUND, current directions display the variability of tidally-influenced regions.

The Top Camp Formation in the Tunnel Creek area contains a thick, 100 km-long body of fine-grained laminated quartz arenite overlying the Bearuroo Dolomite Member. Numerous current lineations, small flute molds and groove casts, as well as sporadic, thin, mudflake conglomerate, lunate current ripples (wavelength 10 cm), and current shadows behind coarse sand grains are present. A lenticular, algal dolomite is interbedded about halfway up the sequence. A current-swept sub-environment is indicated, probably a tidally emergent shoal-bar system.

The occurrence of bioherms at several levels in the lagoonal facies is evidence that favourable conditions for their growth were repeated during slow subsidence of the basin.

LOWER BARRIER-BAR FACIES

The sandstone of the Kiangi Creek Formation, Wonyulgunna Sandstone, and arenite members near the base of the Backdoor Formation, constitute a facies in the western basin of more extensive outcrop area than the lagoonal facies. The lithology and the dominance of laminated and massive beds point to the high-energy conditions of a barrier-bar or beach environment.

The rocks are usually massive and, except for normally faint laminations, sedimentary structures are sparse (Table 6). Evenly laminated bedding can originate in several ways in various sub-environments, for example, suspension settling from wind, swash, or outer surf zone deposition (Clifton, Hunter, and Phillips, 1971), and backshore deposition including washover-fan formation (Reineck and Singh, 1973, p. 298). Erosion of tidal channel banks, desiccated mudflats, and mud layers deposited in ponds on washover fans and the backshore, provided material for thin, sporadic beds of intraclasts. Occasional conglomerate and pebbly sandstone beds, some containing well-rounded quartz clasts up to 8 cm across, are indicative of the very high energy conditions of tidal inlets. Sporadic herringbone cross-bedding (alternating sets of planar foresets dipping in diametrically opposed directions) suggests the reversing currents of a tidal regime. Some small symmetrical ripples would have formed in water only centimetres deep (Tanner, 1971).

Feldspar grains (up to 20 per cent of sandstone) are regarded as first-cycle detritus from granitic sources, and, at least in the most feldspathic rocks, their residence time in the highest energy zones must have been short. Subordinate wackes probably mark backshore zones, and result either from the mixing of waterlaid mud and airborne sand in lagoons or from poor sorting by sluggish

TABLE 6. DEPOSITIONAL FEATURES AND INTERPRETATION, LOWER BARRIER-BAR FACIES

Feature	*Interpretation*
Evenly laminated bedding	Lower part of upper flow regime (washover fan, swash, or surf zone); eolian or sub-aqueous suspension settling
Alpha and omicron cross-bedding	Sand wave deposition by currents, e.g. inner offshore zone (Clifton and others, 1971), tidal inlets, washover fans
Pi and nu cross-bedding	Migrating troughs of inner surf zone, outer nearshore zone and inner offshore zone (Clifton and others, 1971); lunate ripples in backshore zone
Occasional herringbone cross-bedding	Alternating tidal currents
Small symmetrical ripples	Waves in water only centimetres deep
Asymmetrical ripples	Current ripples of lower flow regime
Sporadic swash ripples	Swash and backwash zone of beach
Current lineations	Lower part of upper flow regime
Intraclastic layers	Short distance transport of mud flakes or chips
Conglomerate lenses and pebble sandstone	Very high energy regime, possibly channels of tidal inlets
Feldspathic sandstone beds	First cycle detritus from granitic sources, with short residence time in highest energy zones
Pseudomorphs after pyrite	Reducing conditions within the sediment
Lutite and dolomite interbeds	Intertonguing with adjoining facies; some backshore pools
Wacke beds	Poor sorting in backshore areas, or deltas
Intraformational folding	Compaction or slumping

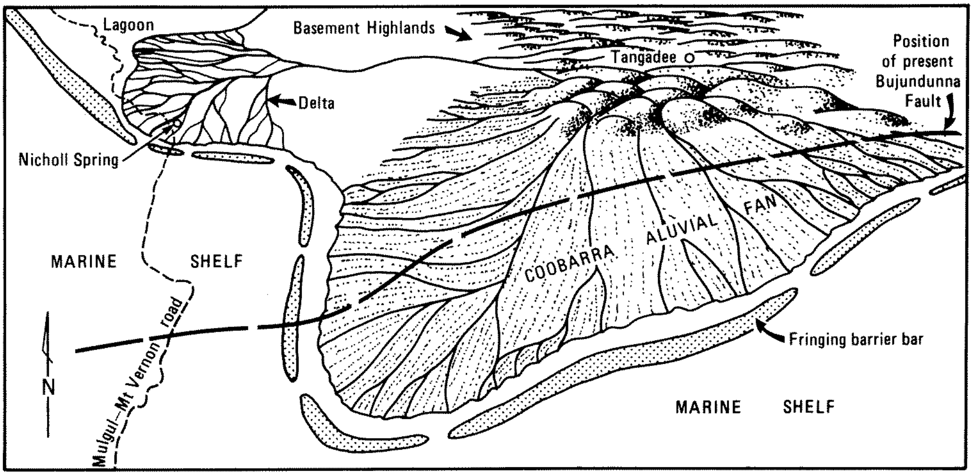


Figure 24. Palaeogeography of the Nicholl Spring—Coobarra Creek—Tangadee area in Kiangi Creek Formation time. The Coobarra alluvial fan and the deltaic subfacies may not have been strictly contemporaneous.

currents near the mouths of streams. A good example of the latter situation may be the sequence 5 km northwest of Nicholl Spring (COLLIER) where wacke dominates over arenite in places, and both have conspicuous mica, feldspar and lithic components. Dolomitic interbeds in this area can be attributed to fluctuating lagoonal influence. The wackes were probably laid down in a delta built out over interfingering shoal sand, lagoonal carbonates, and fine clastics. The sub-facies extends over a distance of 22 km, but its source area is unknown. We conjecture that it lay to the east, and was the same basement high, north of the Bujundunna Fault, which supplied debris to the Coobarra Formation (Fig. 24).

Most of the lutite interbeds in the barrier-bar facies can be looked upon as intertonguings with adjacent marine-shelf and lagoonal environments. Transitions between the Kiangi Creek and Jillawarra Formations either display such alternations, or a fining upwards representing decreasing energy conditions. Transition to sandstone facies from the underlying Irregully Formation is by a lithological gradation from dolomite through sandy dolomite to sandstone, or by interbedding of discrete dolomite and arenite units.

The barrier-bar facies could have formed a variety of coastal landforms, including barrier islands, submerged shoals, and mainland beaches and spits, punctuated at intervals by estuaries or deltas. The Barlee Range district may illustrate the mainland beach condition. The facies is locally absent in areas such as the northeastern Geegin Syncline and near Mount Candolle.

The barrier-bar facies is not found along the northern margin of the basin, where the Top Camp Dolomite records almost continuous lagoonal sedimentation. A 3 m-thick sandstone in Fords Creek, and the fine-grained arenite above the Bearuroo Dolomite Member in the Tunnel Creek area may represent tendencies to develop barrier bars here. Therefore, a well-developed barrier-bar system did not extend far beyond the present main synclinal axis of the Bangemall Basin.

The eastward extent of the facies is less clear, but the Wonyulgunna Sandstone, and sandstone members near the base of the Backdoor Formation are regarded as part of it. The coastal sand system was here attached to the mainland, because off-shore slopes were too steep for an intervening lagoonal system to form. Only in a few areas were shale, siltstone, and chert formed below the thick sandstone units; these areas appear to represent protected embayments in the basement surface. Occasional current lineations and intraclast layers are attributed to tidal activity. At the entrances to the bays, the muds came to be overlain by sand spits extending from the beaches. The basal lutite-carbonate sequence in the Bryah district (PEAK HILL) is probably lagoonal.

The Wonyulgunna Sandstone contains conglomerate lenses and is interpreted as being a beach facies. The conglomerate clasts of vein quartz and quartzite are typical of mature deposits; the common basement lithologies of schist and granite are not represented because these less stable components were eliminated by prolonged erosion in a high-energy surf zone. Feldspar is usually absent from the basal beds for the same reason. However, higher in the sequence, phases with up to 10 per cent feldspar do occur, suggesting less rigorous conditions. The moderate sorting of much of the unit, and the mixture of rounded and angular grains, also correspond with a less turbulent regime than that near the base. Planar tangential cross-bedding shows a spread of directions from easterly and westerly (longshore tidal currents) through to southerly (landward dip typical of the outer nearshore zone, Clifton and others, 1971). Variable directions of symmetrical and asymmetrical ripple marks reflect diverse origins and changeable currents.

LOWER OPEN-MARINE-SHELF FACIES

The Jillawarra Formation forms an extensive laminated mud sheet corresponding to the X zone of Irwin (1965). It is thickest in the centre of the basin between the Hells Doorway Syncline and Jillawarra Bore, where it exceeds 1 km, and thins gradually to the northwest (Fig. 25). If the overlying Discovery Chert is regarded as a time line, formation thickness can be equated with duration of shelf deposition, and the shoreline would have transgressed to the northwest during this time. The Jillawarra is equivalent to the basal Backdoor Formation in some areas, showing that there was also an easterly transgression. The area of greatest subsidence up to Discovery Chert time would correspond roughly with the position of the depocentre near Jillawarra Bore.

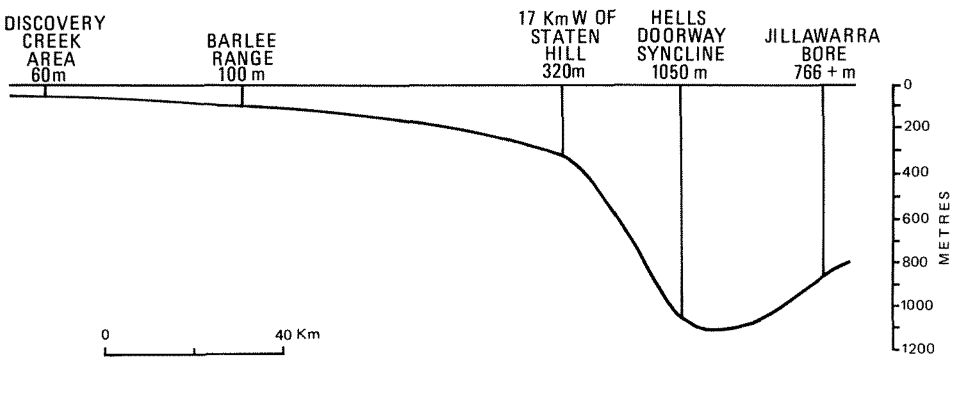


Figure 25. Thickness of the Jillawarra Formation between Discovery Creek and Jillawarra Bore.

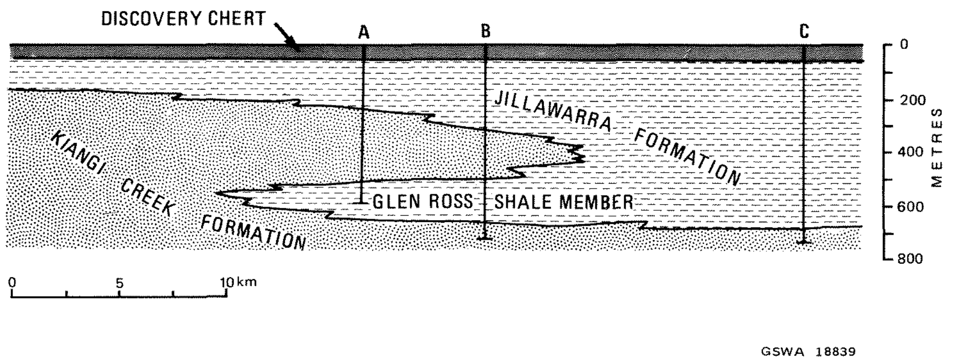


Figure 26. Stratigraphic relation between the Glen Ross Shale Member and adjacent units, in pre-deformation cross section. A—Vertical profile in north limb of Mount Vernon Syncline. B—Profile in south limb of Mount Vernon Syncline. C—Profile in north limb of Range Creek Syncline.

The Glen Ross Shale Member, a thick and extensive member of the Kiangi Creek Formation in the Mount Vernon Syncline, is lithologically so similar to the Jillawarra rocks that it is accounted for as a tongue of marine-shelf deposits. The overlying sandstone marks a temporary regression in the overall transgressive sequence (Fig. 26). This need not have been caused by eustatic change, but may have resulted from an increase in available sediment supply (Ryer, 1977), perhaps associated with the building of the delta west of Tangadee.

The laminated shale and siltstone facies, with sparse dolomite and chert (Table 7) is typical of a marine-shelf environment on the seaward side of a barrier system (Fig. 27) where mud settles from suspension below effective wave base (Selley, 1976, p.225). Marine conditions are corroborated by the Ga-B-Rb triangular plots of samples from diamond-drill cores in the Glen Ross Shale Member (Davy, and others 1978). Similar plots of the carbonate zone near the base of the member, however, imply a very shallow-marine or brackish influence, possibly hinting at proximity to a river mouth. In all these samples the B and Ga are associated with illite and chlorite, consistent with a marine regime.

Some organic material has been found in thin sections from all cores and is best explained as comminuted, filamentous debris washed out to sea from shore-line algal mats. Some may have originated from planktonic organisms. The reducing environment in the sediment during deposition is almost certainly due to the organic material itself.

There are a few indications that, in exceptional circumstances, wave action affected the shelf sediment. These include scour-and-fill structures and a calcarenite bed of rounded algal dolomite pellets. The latter signifies an agitated zone in

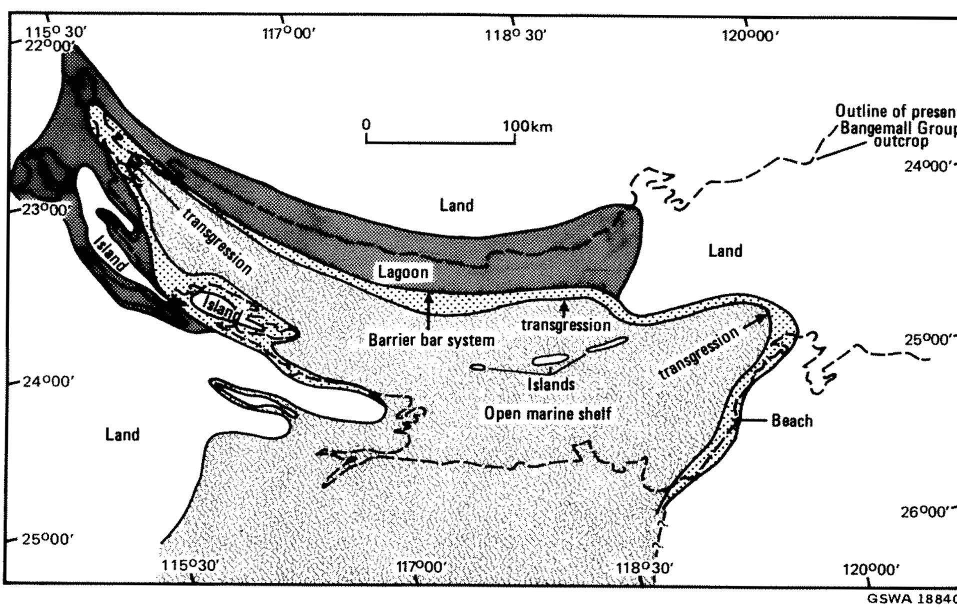


Figure 27. Generalized depositional environments in the Bangemall Basin during deposition of the Jillawarra Formation. The marine shelf may have extended further east than shown.

shallow water not far from the barrier-bar system. Sandstone interbeds also indicate occasional high-energy events. These beds are usually less than 0.1 m thick, and in places are merely trains of clastic quartz in shale. Contacts between sandstone and shale are normally knife sharp, demonstrating sudden, episodic suspension settling following heavy storms. Storm-sand layers extending 40 km offshore in the North Sea (Reineck and Singh, 1973) are regarded as the present day analogue of most of the beds. The thicker sandstone beds probably represent interfingering in the transition zone with the barrier-bar facies.

TABLE 7. DEPOSITIONAL FEATURES AND INTERPRETATION, OPEN-SHELF FACIES

Feature	*Interpretation*
Laminated shale and siltstone	Settling out of suspension below wave base
Dolomite members; illite and chlorite clay minerals	Marine regime
B-Rb-Ga ratios	Saline water, with local tendencies to brackish
Sharp contacts between siltstone and shale	Sudden changes in grain size of clastic supply
Carbon content up to 10 per cent	Supply of organic material in reducing conditions of deposition
Organic filaments	Algal mat debris washed out to sea from shoreline; algal mat growth in shallow sea, in absence of grazers, also possible
Pyrite blebs and cubes	Reducing conditions in sediment
Hematite cement in some unweathered rock	Oxidizing environment and atmosphere
Occasional scour and fill structures	Wave base impinging on sea floor during heavy storms
Traces of small-scale (1-3 mm) cross-bedding	Small current ripples
Trains of clastic quartz and sandstone beds up to 1 m thick with knife-sharp contacts	Suspension cloud settling following heavy storms; some interfingering with barrier-bar facies
Sporadic calcarenite layers	Local agitated zone in shallow water
Fragmental carbonate rocks; trough cross-bedding; scours; small wavelength symmetrical ripples in Peedawarra Synclinorium	Local shoaling and nearshore shallow water in absence of barrier-bar quartz-sand facies

Lenses of coarse-grained arenite and pebbly granule conglomerate at the base of the Jillawarra Formation 18 km west-northwest of Woodlands Station appear to be cross-sections of finger-like bodies and are inferred to be subtidal channel deposits that lie normal to the shoreline on the seaward side of tidal inlets.

A quartz arenite wedge at the top of the Jillawarra in the nose of the Lyons River Anticline contains some shale intraclasts and was probably deposited as a bar near the coast. This would imply that the water depth was within the range of normal wave action at the end of Jillawarra deposition in this area. Further evidence of similar water depths in localized areas is provided in the middle of the formation in the southwest limb of the Peedawarra Synclinorium by small wavelength symmetrical ripples (Fig. 28), dolarenites, doloconglomerates, and laminated dolomite with low-angle trough-cross-bedding and scours. This area is exceptional, however, because a barrier-bar sand unit is missing from the succession, so that the clastic carbonates and sedimentary structures mark the position of shoals.

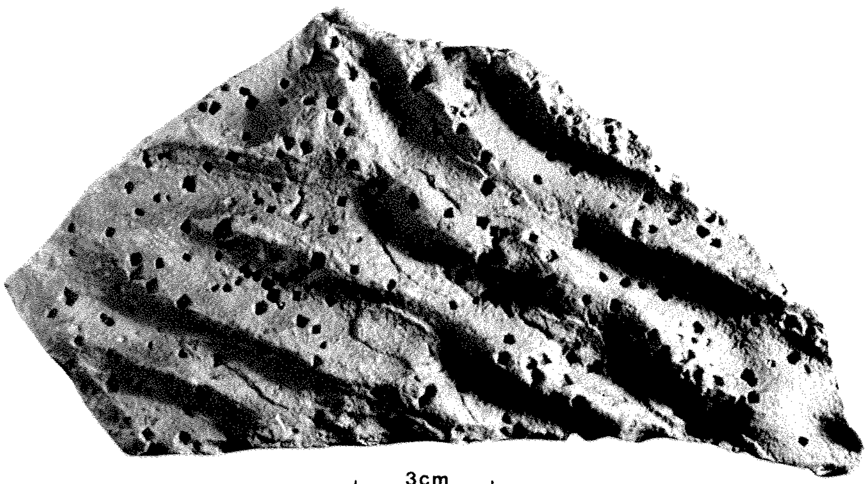


Figure 28. **Symmetrical ripple marks (average wavelength 20 mm) in siliceous siltstone of the Jillawarra Formation, southern limb Peedawarra Syncline. The bedding surface is studded with molds of cubic pyrite crystals. The ripples have been steepened by load deformation during compaction.**

The indications of gypsum crystals near the top of the Jillawarra Formation (and in the overlying Discovery Chert) points to hypersalinity in some areas, implying that circulation in the basin was becoming restricted towards the end of Jillawarra deposition.

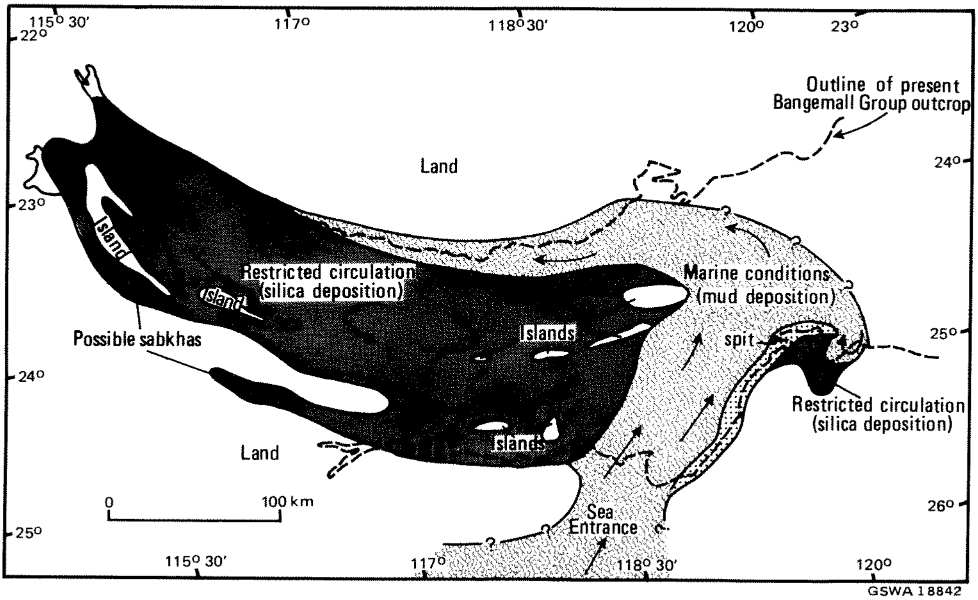


Figure 29. **The Bangemall Basin in Discovery Chert time. Inferred circulation pattern shown by arrows. The small eastern area of silica deposition may not have been contemporaneous with that of the Discovery Chert. There was some intermittent silica deposition in the normally marine area of the Backdoor Formation.**

SILICA DEPOSITION (DISCOVERY CHERT)

The deposition of the Discovery Chert is treated in detail in Chapter 5, but the salient features can be summarized here.

The onset of restricted circulation towards the end of Jillawarra time was probably caused by a barrier to the open sea forming along the southern margin of the basin west of Mount Deverell (Fig. 29). The barrier may have been a basement arch resulting from uplift along this zone. An opening between Mount Deverell and the Marymia Dome allowed access of marine water to the eastern part of the basin which appears to have been separated from the area of deposition of Discovery Chert by several islands and a submerged ridge.

The area in which the Discovery Chert was deposited lay below the limit of wave turbulence. The input of clastic material was very small, and evaporation concentrated Na^+ ions and SiO_2 in the water. Silica-secreting organisms were absent, but pelagic algae bloomed cyclically, and the organic remains decayed to produce a high CO_2 concentration, low pH, and reducing conditions. This chemical enviroment was favourable for the precipitation of primary silica gel, which became chert on diagenesis.

MARINE PLATFORM AND TURBIDITE DEPOSITION

CARBONATE PLATFORM FACIES

The Devil Creek Formation is a carbonate-lutite association similar to the Irregully Formation (Table 8). The main differences are far fewer stromatolites, fewer breccias, more dolarenite, and an apparent absence of desiccation features.

These fine-grained, laminated platform-facies carbonates settled out of suspension in quiet water. The few stromatolites grew in upper subtidal to lower intertidal conditions. Other indicators of shallow water are channel scours (Fig. 30), intraclasts, and dolobreccia (Fig. 31), cross-bedding (Fig. 32) and rare oolites.

Intraformational breccia beds in the Devil Creek type section have both erosional and non-erosional bases. Some rounded slabs signify transportation for some distance, while others show bedding broken to form free blocks, implying a slump origin. The breccia fragments are considered to be derived from a mudflat.

Dolarenites and dolorudites, which are widespread in the centre of the basin, indicate scattered, carbonate shoals. Climbing ripples (lambda and kappa cross-bedding, Fig. 32B) have a similar distribution to the dolarenites, and show unvaried current directions at each locality. They require, for their formation, rapid deposition of material from suspension under current or wave action. The current directions in the Devil Creek Formation are southerly to westerly (Fig. 33), coming from the axial region of the basin. The reason for this is not clear.

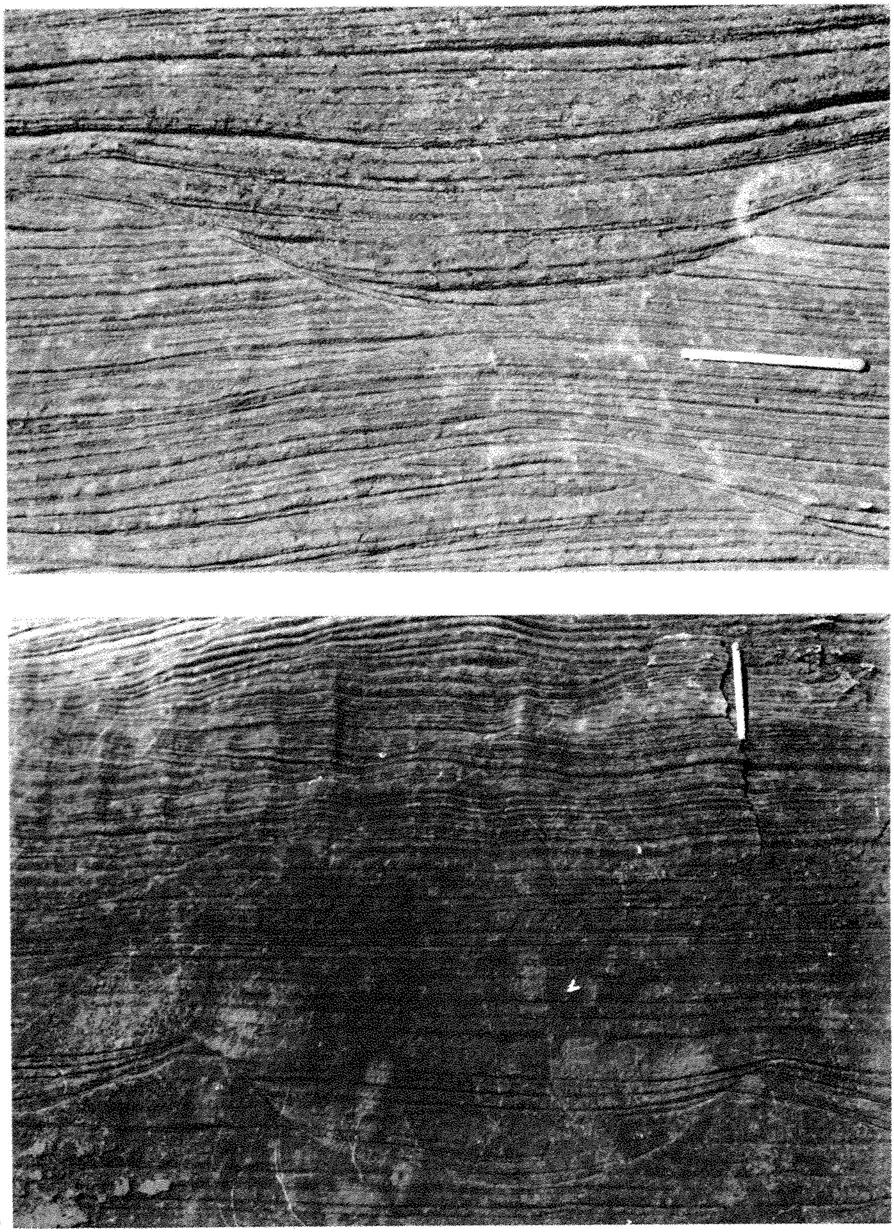
Intraformational folding is recorded in the Cobra Synclinorium and 9 km northwest of Milgun. At the latter locality there appears to have been slumping towards the north-northwest, probably triggered by tectonic movements on basement faults to the south.

TABLE 8. DEPOSITIONAL FEATURES AND INTERPRETATION, CARBONATE PLATFORM FACIES

Feature	*Interpretation*
Laminated, fine-grained dolomite with interbedded laminated lutite	Quiet water in lagoons or in restricted bays and ponds
Baicalia stromatolites	Upper subtidal to lower intertidal regime
Breccias consisting of boulders and slabs of dolomite in a micritic matrix	Littoral and tidal-flat erosion, or undercutting of meander-bend bank
Occasional oolites	High-energy conditions in intertidal zone or channel
Dolarenite beds, some with slabs of fine-grained dolomite up to 30 cm long; some calcirudite lenses	Turbulent regime of shoals or tidal channels
Channel scours in laminated sediment	Erosion events in a normally low-energy environment
Climbing ripples (lambda and kappa cross-bedding) in dolomite	Deposition from suspension under the influence of unidirectional currents
Occasional small wavelength (20 mm) ripples	Water only centimetres deep
Alpha, pi and nu cross-bedding	Migration of various ripple bedforms in conditions of little or no suspension deposition
Dolomite and dolomitic mudstone with clay platelets	Erosion from mudflat and redeposition in quieter water
Sporadic fine- to medium-grained quartz arenite lenses, some with climbing ripples and trough cross-bedding	Estuarine channel deposits
Fine-grained muscovite in some rocks	Terrigenous material which settled out of suspension in calm conditions
Pseudomorphs after pyrite cubes, pyrite balls, and, rarely, pyrrhotite; fresh pyrite is rare	Reducing conditions during diagenesis
Greenish dolomite and lutite near Tangadee and Mulgul	Chlorite component, possibly derived from local basement highs
Occasional flame structures and load casts	Loading features
Intraformational folding is rare	Slumping, probably triggered off by nearby active faulting
Calcite grains (up to 40 per cent) in dolomite, and occasional limestone pods	Relics of original carbonate sediment

Many dolomites contain up to 40 per cent calcite grains. One calcarenite showed less than 10 per cent dolomite. A rock, consisting of limestone clasts in a dolomite matrix and rare pods of limestone, is also known. These examples, like the carbonates of the Irregully Formation, are probably dolomitized primary limestones.

The general succession, which passes conformably upwards from the Discovery Chert into the Ullawarra Formation, does not indicate a marine transgression as for earlier formations. Most of this facies was deposited below wave base, on what was probably a sheltered, gently sloping marine platform subject to scouring by storm waves. Intertidal mud flats, estuaries and shoals also existed. The area of deposition coincided closely with that of the Discovery Chert and implies persistence of the restricted circulation which affected the latter. The change from silica to carbonate and fine-grained clastic sedimentation was sudden and widespread, and its cause is conjectural. The barrier, present to the south in Discovery Chert times, may have subsided to be replaced by a sandy bar, and the islands on western COLLIER may



A

B

G.S.W.A. 18843

Figure 30. Channel scours in dolomite of Devil Creek Formation 4 km southwest of Cobra homestead. A—Scour with smooth, gently sloped profile. The channel was filled by horizontally laminated sediment before interfluves were covered. Part of another shallow scour is visible at lower right. B—Scour with 40-degree slopes and peaked interfluves. Channel was only partly filled before continuous laminations were deposited.

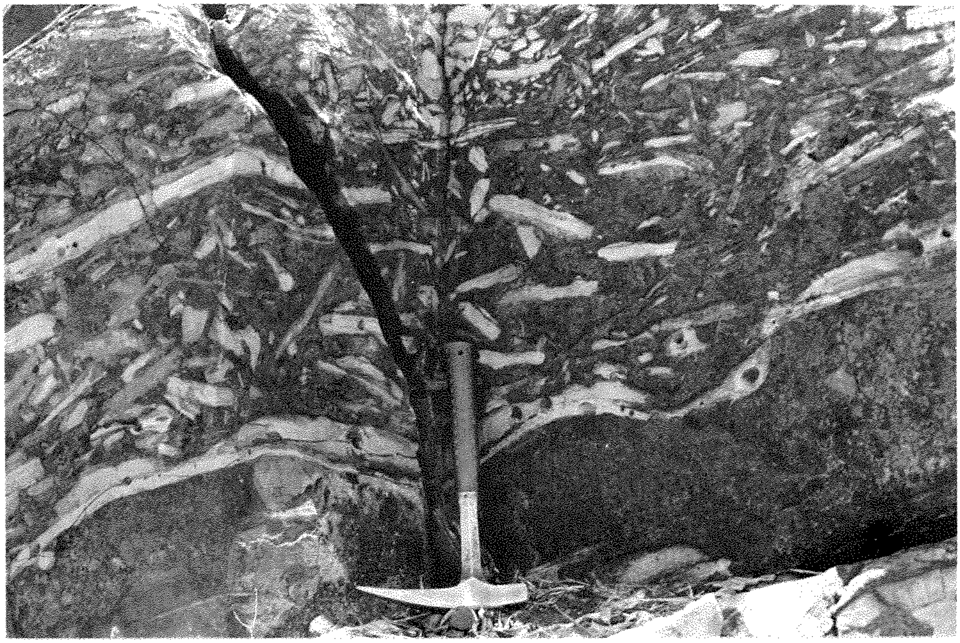


Figure 31. Basal part of the youngest breccia bed in the type section of the Devil Creek Formation. Slabs of fine-grained, laminated dolomite up to 0.5 m long are set in a matrix of fine-grained dolomite. The base of the bed is erosional.

also have been reduced in size or submerged. Greater marine influx through wider inlets would have affected the pH, salinity and circulation sufficiently to tip the balance from silica to carbonate precipitation. The normal shelf-mud sequence was not laid down because the supply of terrigenous clastic material was limited. There appears to be no exact modern analogy to the situation described.

SOUTHERN BARRIER FACIES

The Nanular Sandstone, which is confined to the Mount Clere - Mount Deverell region (Fig. 35), is a medium- to fine-grained, moderately well-sorted quartz arenite. Local phases of the sandstone are coarse grained or feldspathic. Mudflake layers, nu cross-bedding, current striations and ripples are present. All these features are similar to those of the Kiangi Creek Formation, and a corresponding barrier-bar facies is inferred.

The lithofacies appears to have formed adjacent to the Devil Creek Formation, and to have migrated northwards over the carbonate facies. It lenses out at the base of the Ullawarra Formation. Presumably the barrier sand formed on the subsiding land barrier which had been present in Discovery Chert time. It maintained a restricted circulation in the area of Devil Creek deposition, but allowed, by means of inlets, a greater marine influence than in Discovery Chert time. The Nanular Sandstone did not transgress over the rest of the basin to form a sheet analogous to the Kiangi Creek Formation, because it was located on a basement ridge and drowned by subsidence.

A

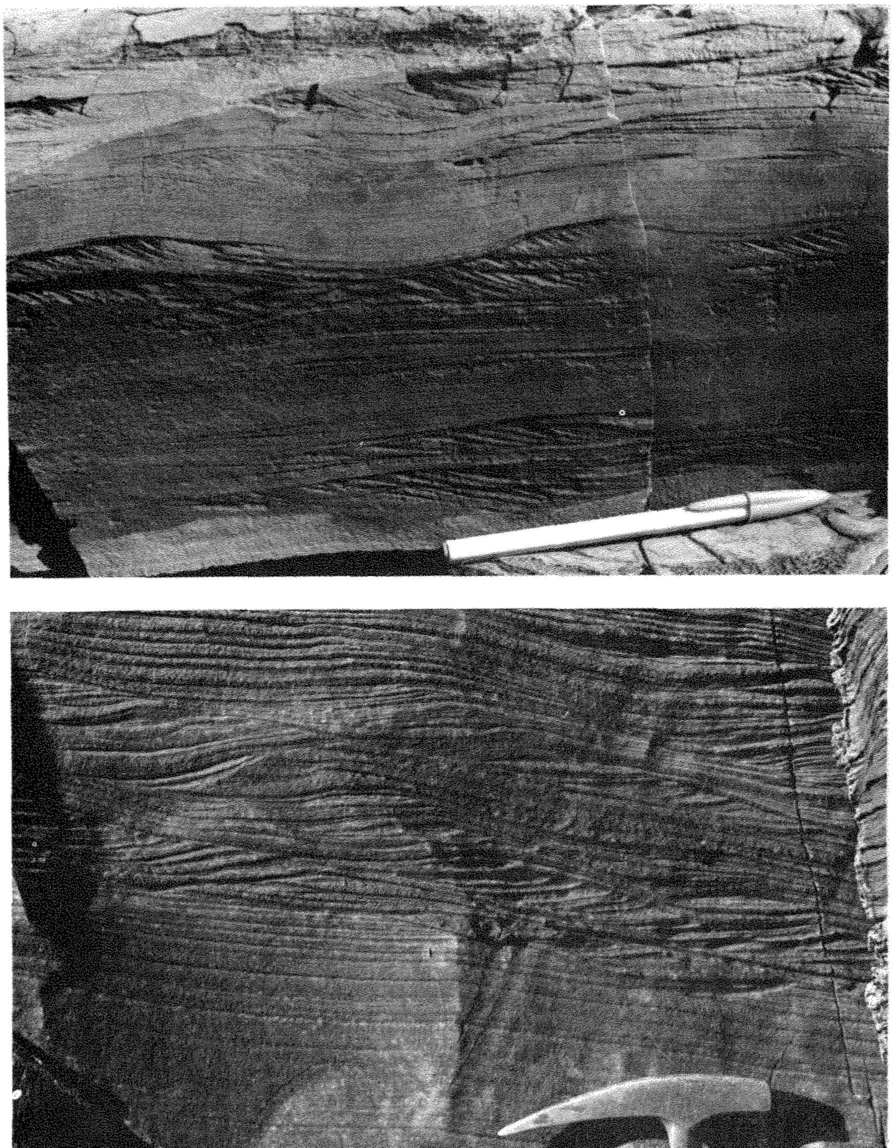
B


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Figure 32. Ripple cross-bedding in dolomite of the Devil Creek Formation. A—Internally cross-laminated symmetrical ripples 10 km west-southwest of Mulgul homestead. The cross-bedding dips persistently to the right (west), indicating that westerly currents were dominant. A weaker opposing current may have modified the ripple shapes to the symmetrical before quieter periods of suspension-settling occurred. B—Climbing ripple lamination 4 km southwest of Cobra homestead. The structures were initiated as internally cross-laminated ripples similar to those in A, but continued to grow vertically and laterally by suspension deposition while the current continued to act. Note the planar erosion surfaces between sets.

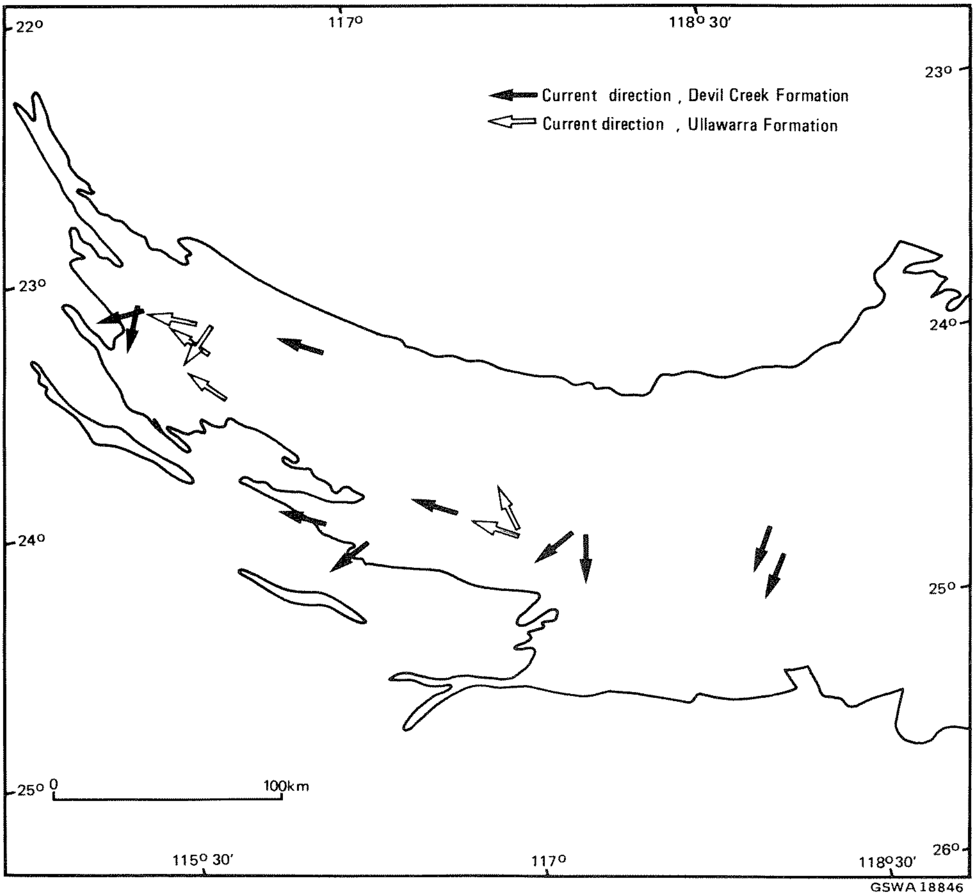


Figure 33. Palaeocurrents in the Devil Creek and Ullawarra Formations. Data from EDMUND after Daniels (1969).

LUTITE PLATFORM FACIES

The carbonate and lutite lithofacies of the Devil Creek Formation gave way to the lutite lithofacies of the Ullawarra Formation, due to an increased influx of fine-grained clastic material on the shallow-marine platform. Interfingering of the two facies shows that increased mud deposition took place at different times in different areas. In some areas, such as the Range Creek Syncline, clastic input stopped for a while, allowing a temporary resumption of carbonate deposition. In a few other places, for example, west of Mount Sanford, there was continuous deposition of mud after the Discovery Chert and no carbonate facies was laid down.

Many features of the Ullawarra Formation (Table 9), and equivalent parts of the Backdoor Formation, are similar to the Devil Creek Formation in style of clastic deposition. However, in place of carbonate shoals there were shoals of sand and silt. The feldspathic nature of some shoals suggests minimal reworking and rapid burial. Most of the current ripples, climbing ripples and cross-bedding occur in the coarse-grained siltstone of the Berala Syncline. The minimum flow depth calculated from one ripple set by Allen's (1970) method is 0.4 m. This silty and sandy sediment,

TABLE 9.
DEPOSITIONAL FEATURES AND INTERPRETATION, ULLAWARRA FORMATION

Feature	*Interpretation*
Well-bedded and laminated shale and siltstone	Deposition from suspension below wave base
Dolomite bands; illite and chlorite clay minerals	Marine regime
Pyrite molds	Reducing conditions in sediment
Coarse-grained siltstone and fine-grained sandstone subfacies	Shoal or intertidal sub-environment
Occasional medium- and coarse-grained arenite lenses	High-energy zones in shoal, tidal inlet or intertidal areas
Alpha cross-bedding in medium-grained arenite	Sand-wave deposition by currents of lower flow-regime
Climbing ripples (kappa cross-bedding) in fine-grained sandstone and siltstone	Deposition from suspension, dominating lateral transport by currents
Current ripples in coarse-grained siltstone	Dominantly lateral transport of material by lower flow-regime currents
Wacke beds up to 0.5 m thick	Poorly sorted deposit near estuary mouth, or turbidite
Occasional algal lamination with gypsum crystal pseudomorphs	Intertidal algal mat subjected to evaporitic desiccation
Groove, bounce and brush casts from thin wacke beds	Erosional currents of turbidity flows

where widespread, may correspond to the intertidal zone at Jade Bay on the North Sea (Reineck and Singh, 1973, p. 357) where the coarsest sediment is deposited around low-tide level. This coarse-grained siltstone is less common in the main basin synclinorium, suggesting a hinge line roughly along the North Lyons River with renewed basin subsidence to its north.

Algal-laminated mudstone, in which platy gypsum crystals grew across the lamination (Fig. 34) occurs 7 km northeast of Ullawarra homestead (EDMUND). This demonstrates that at least locally (south of the postulated hinge line) an algal mat grew in an embayment or pool which was subsequently evaporated.

Daniels (1966b) reported cross-bedding, groove casts, bounce casts and brush casts from thin greywacke beds, and a 32° difference between erosional and depositional currents. These are attributed to turbidity flows expanding over the basin bottom (Potter and Pettijohn, 1963, p. 131).

The Curran Formation, consisting of laminated siltstone and shale, is also the result of slow deposition from suspension in a shelf setting. Lithologically it is more uniform than the variable Ullawarra Formation, the rocks are black instead of green, and the regional extent is quite limited compared to that of the Ullawarra-Backdoor Formations (Fig. 35).

It is generally without current structure, but there are occasional bands, a few centimetres thick, of silt lenses within finer grained mud-drape layers. This structure is produced by alternating periods of current wave activity and slack water, and is common in tidal regimes (Reineck and Singh, 1973, p.100). However, Coleman and Gagliano (1965) also report it from shelf and prodelta environments.

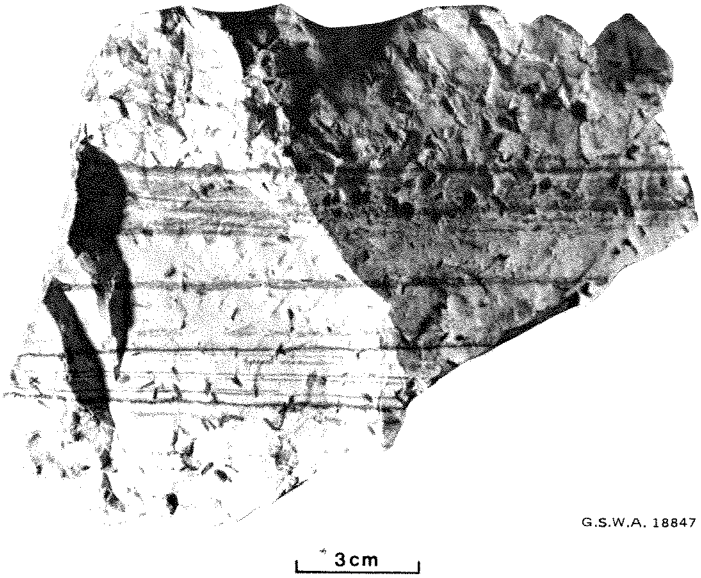


Figure 34. Pseudomorphs after gypsum cutting across bedding in laminated mudstone from the Ullawarra Formation 7 km northeast of Ullawarra homestead, EDMUND.

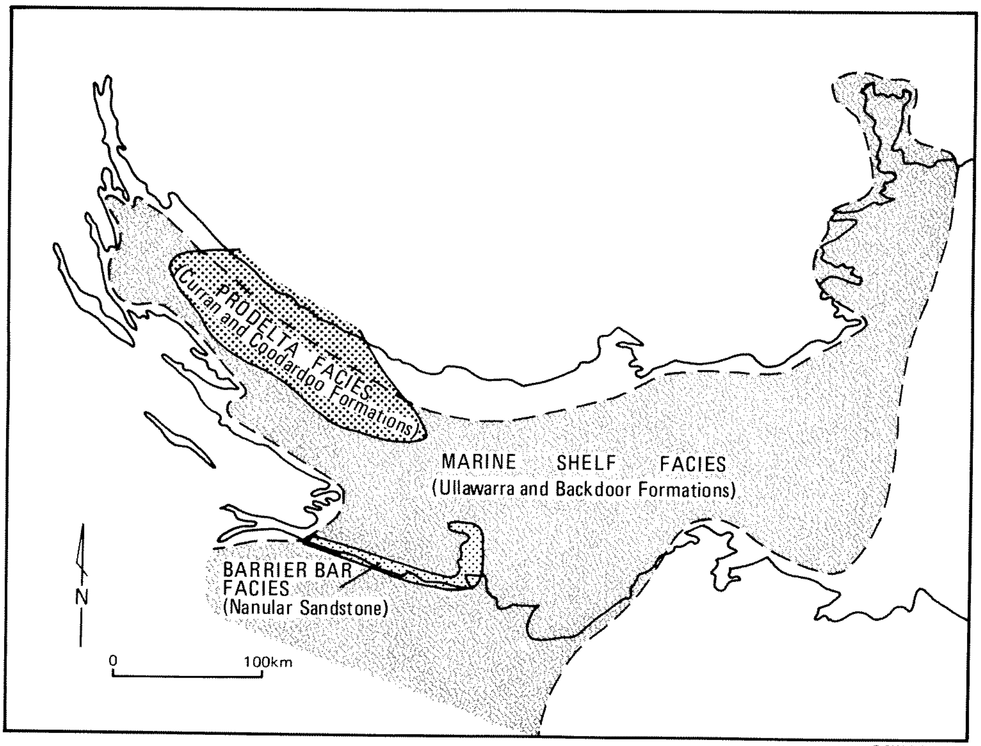


Figure 35. Generalized depositional environments of the Nanular Sandstone, Ullawarra, Backdoor, Curran and Coodardoo Formations. The Nanular Sandstone was deposited before the others as a barrier bar which had subsided before Ullawarra deposition started. The Curran and Coodardoo Formations, which were deposited in front of a delta located to their north, prograded over part of the Ullawarra marine shelf.

Also present are mu cross-bedding cosets up to 0.25 m thick, generated by the migration of trains of small current ripples, and some thin turbidite beds, with mudstone intraclasts. Although there are occasional thin dolomite lenses and limestone concretions, there is no evidence for carbonate shoaling as in the underlying formations.

The Curran Formation is interpreted as a distal prodelta shelf deposit. The unit is always present below the prodelta turbidite of the Coodardoo Formation (*see* below) and extends beyond it. This close association between the two units, both in a limited area in the northwest of the basin, implies a related origin. Distant deltaic influence explains the deposition of a mud blanket of more uniform lithology than the underlying muds. The black colour of the Curran Formation is due to carbon. This is attributed to comminuted organic debris washed out to sea from the deltas. A northerly to northwesterly source of sediment is inferred as for the Coodardoo Formation.

TURBIDITE FACIES

The Coodardoo Formation is a turbidite facies, which occupies approximately the same region as the Curran Formation. In the type section (Coodardoo Gap) it is characterized by cyclic deposition (Fig. 36). A typical cycle starts with quartz arenite which grades upwards into wacke and, eventually, siltstone, fine-grained mudstone, and shale. The base of each cycle is sharply defined and clearly erosional, with scours up to 0.2 m deep; several surfaces display current lineation, fluting or groove molds. The upwards increase in matrix content is accompanied by a decline in grain size. This is most striking where the base of a cycle is coarse-grained (Fig. 9). A few beds with faint, wispy lamination in the top half are typical Bouma sequences laid down by turbidity currents. De-watering structures (Fig. 9) formed as liquid escaped upwards during the settling and compaction of the turbidites. In some cycles the basal arenite is absent; in others, arenite follows the wacke unit. The boundary below the upper lutite of a cycle may be sharp, indicating that the upper lutite is not part of a turbidite unit, but settled from suspension between turbidite events. Irregular, vertical fluctuations in matrix content, and sharp boundaries between some sandstones are compatible with multistorey turbidites. Some cycles have inversely graded bases, which can also form in the same environment (Sanders, 1965). The sandstone intervals of cycles tend to thicken upwards in the sequence, and the sandstone/lutite ratio ranges between 1.5 and 4.0, suggesting proximal turbidite (Pettijohn, 1975). Mudclasts, including slabs up to 30 cm long, illustrate the erosive power of the turbidity flows. Between zones of graded bedded units are massive sandstones without grading, which may have been reworked by deep wave turbulence or currents. Evidence of traction currents is alpha cross-bedding in a 15 cm-thick granule conglomerate.

The Coodardoo Formation forms a sand sheet 175 km long, and palaeocurrents were to the southwest. Increasing distance from the source is accompanied by a stratigraphic thinning and a decreasing proportion of wacke, until southeast of Dooley Downs (MOUNT EGERTON) only arenite is present. The sandstones of this latter region have no graded bedding and are structureless. The water may not have been very deep (less than 30 m) and the turbidite sands may have been winnowed by deep wave turbulence or bottom traction currents.

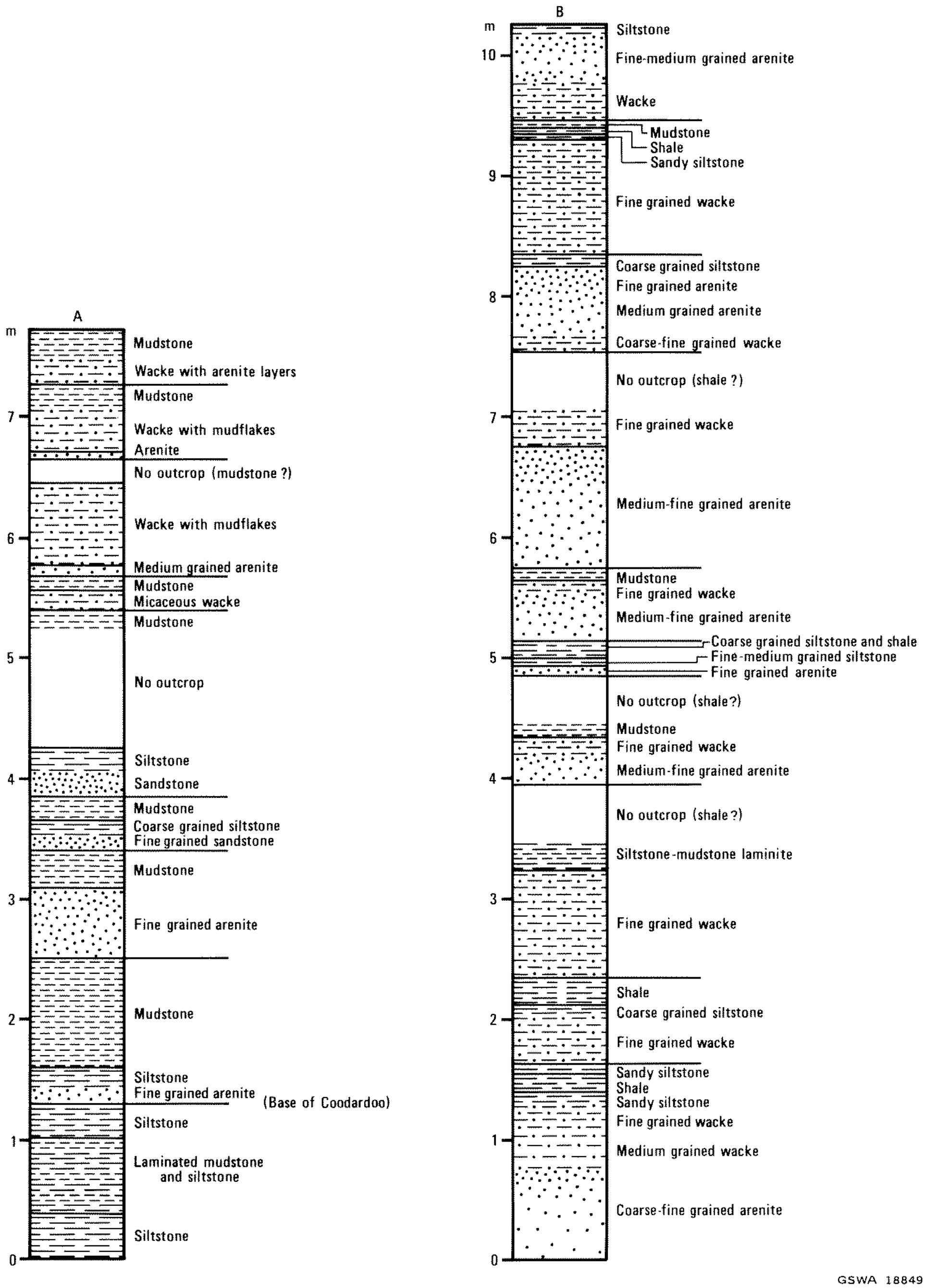


Figure 36. **Measured sections in the Coodardoo Formation type section, Coodardoo Gap. A—Base of formation. B—Middle of formation. The sequences consist of stacked, fining-upwards cycles, whose boundaries are indicated to the right of the columns. Graded bedding is dominant. The proportion of sandstone in the cycles increases upwards suggesting an advancing turbidite fan.**

The most likely source of the turbidity currents was the front of a sandy delta to the north or northwest, where depositional instabilites led to intermittent slumping.

REGRESSIVE LINEAR-SHORELINE DEPOSITION

The three major facies of the Mucalana Subgroup resemble the Irregully-Jillawarra interval in reverse, and are interpreted as a regressive linear clastic shoreline sequence. Differences in detail are the near absence of carbonates in the Kurabuka Formation as compared with the Irregully, and continuing turbidite deposition from Coodardoo times into Fords Creek Shale times. The lagoonal environment of the Kurabuka is the youngest preserved in the western facies of the basin, but continued marine regression would have produced deltaic and fluvial environments before the basin was filled.

UPPER OPEN-MARINE-SHELF FACIES

The Fords Creek Shale is predominantly a sheet of laminated lutite, deposited below wave base on a marine shelf (Table 10). Shoal areas are indicated by a few dolarenite members. Algal lamination (Fig. 69) in chert northeast of Coodardoo Gap indicates a water depth of probably less than 50 m (Brock, 1976). Possible gypsum-crystal-molds and small quartz-filled pores, which may represent fenestral

TABLE 10. DEPOSITIONAL FEATURES AND INTERPRETATION, UPPER MARINE-SHELF FACIES

Feature	*Interpretation*
Well-bedded, laminated, fine-grained clastics	Suspension settling in quiet water
Sparse pyrite	Reducing conditions in sediment
Glauconite in some sandstone and siltstone	Marine derivation
Limestone and dolomite members	Marine conditions
Sparse medium- to coarse-grained dolarenite	Derived from shallow turbulent water, but may have been transported to deeper water
Algal lamination in a chert member	Intertidal or subtidal algal growth within the photic zone
Thin, fine-grained arenite beds and mudstone-arenite laminite	Storm sand layers
Wacke beds, with intraclasts, groove and flute casts on soles, parting lineation and ripples, occasional slump folds, and cross-bedding at shaly top of units	Turbidity current deposits
Intraformational conglomerate, 2.5 m thick, of pebbles and rafts up to 2 m long of dolomite, siltstone, and subordinate chert in an arenite matrix.	Slumping of partially consolidated sediments
Arenite beds, with intraclasts, mu cross-bedding and occasonal ripples, usually less than 0.5 m thick	Probable turbidites, many reworked by traction currents
Cycles with arenite beds thickening upwards from cycle to cycle, shale dominant	Distal turbidite sequences

("birdseye") fabric, suggest subaerial exposure. An analyzed sample of Fords Creek Shale (Appendix D) plots in the marine field of the Ga-B-Rb triangular diagram.

Sandstone interbeds are common in the western half of the lithofacies, and are usually less than 0.5 m thick. The thickest is the Jeeaila Sandstone Member (80 m). The tendency of the Coodardoo Formation to have less sandstone and a lower proportion of wacke to arenite in the east is present also in the Fords Creek Shale. For example, wacke and arenite are most abundant near Dooley Downs, while in Mucalana Creek there is no wacke though arenite is common; near Mount Vernon arenite beds are few, until in COLLIER, no sandstone is present. The sandstone beds feature some characteristics of turbidites (Table 10), and any intraclasts usually occur towards the base of units. However, no graded bedding has been reported.

The wacke beds could be turbidites, but the more abundant arenites present a problem in that they have spread in thin sheets over a greater area. Quartz arenites are known from turbidite basins (Sturt, 1961), but their origin is controversial (Selley, 1970). Hubert (1964) has argued that modern, well-sorted, deep-sea sands are transported by ocean-bottom currents, some perhaps being reworked turbidites. Analogy with the Coodardoo Formation suggests that the sandstones of the Fords Creek Shale were emplaced as turbidity flows, but reworked by bottom traction currents or storm waves so that any muddy material was winnowed out. Two examples from MOUNT EGERTON of arenite bands occurring in thickening upwards cycles with a sandstone:shale ratio of much less than 1.0, indicate distal rather than proximal turbidites (Pettijohn, 1975). Very thin beds may be storm sand layers, as postulated for the Jillawarra Formation. Some of the massive mud-rock beds may also be turbidity current deposits. Turbidites can be deposited in

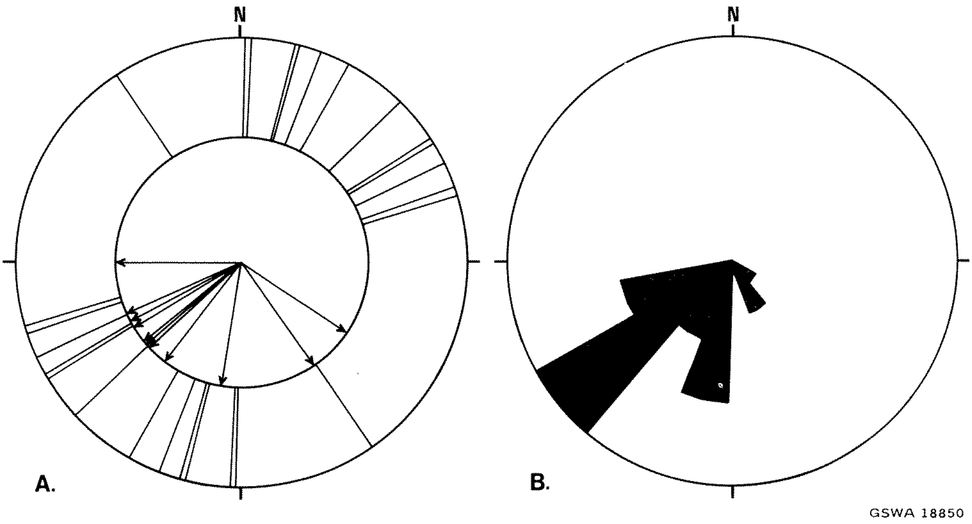


Figure 37. Palaeocurrent data from the Fords Creek Shale. A—Cross-bedding (inner circle) and bounce and groove casts (outer circle), including data from Daniels (1966b). B—Rose diagram of same data, assuming that cross-bedding gives correct direction of movement for other measurements. Class interval 20°.

shallow water, and the interpretation does not conflict with shoal areas being present elsewhere in the region of Fords Creek Shale deposition.

The Fords Creek Shale is thus seen as a marine-shelf facies, similar to the Jillawarra Formation, but with the intermittent addition of turbidites in the western half of its extent.

Palaeocurrent measurements are few, but define a southwesterly trend with a subsidiary southeasterly component (Fig. 37). The source to the northeast which contrasts with that of the Coodardoo, is interpreted as a series of deltas, whose fronts oversteepened and slumped from time to time to generate turbidity currents. These spread downslope to the basin axis, and then turned southeasterly down the gentle trough which had existed since Ullawarra times. Farther east, deposition from suspension of shelf mud was undisturbed by such events and, locally, limestone was laid down when the supply of clastic material was interrupted.

UPPER BARRIER-BAR FACIES

Depositional features of the Mount Vernon Sandstone (Table 11) resemble those of the Kiangi Creek Formation, for which a barrier-bar environment is interpreted, but there are differences in detail. Intraclast beds and herringbone cross-bedding are more abundant in the Mount Vernon Sandstone, probably because of better preservation in a regressive marine sequence than in a transgressive one where the deposits are reworked. Shrinkage cracks (Fig. 38) resemble those pictured by Conybeare and Crook (1968, Plate 51B) and suggest subaerial exposure. Occasional trains of straight, subdued ripples with rough intercrestal areas (Fig. 38B) may be swash ripples. They are up to 8 cm in wavelength, vary from symmetrical to highly asymmetrical (RSI = 12), and have ripple indices ranging down to 13. This RI figure is much less than the usual range for swash ripples (40 to 300), however van Straaten (1953) reported examples with RI less than 10. A small outcrop of diamictite occurs in the Lofty Range; it consists of coarse-grained, well-rounded quartz grains in a green matrix of fine sand, silt, and clay, and could be a turbidity-current deposit, probably originating from instability at the front of a local delta. Well-developed flute molds nearby may be due to turbidity current scouring.

The main lithological differences, as compared with the Kiangi Creek sandstones, are a higher proportion of fine-grained arenite (about 85 per cent of the type section), less feldspar (usually no more than 2 per cent), and fewer wacke beds. Some beds contain up to 1 per cent glauconite affirming a marine setting.

The diachronous nature of the formation and its interfingering with the Fords Creek Shale are clearly evident in the Lofty Range. Beds in the Mount Vernon Sandstone, which can be interpreted as time lines, transgress the formation from the top contact with the Ilgarari Formation, to the bottom contact with the Fords Creek Shale. The Lofty Range exposure is a section subparallel to depositional strike, and the direction of progradation of the barrier-bar facies varied at different times and places. Palaeocurrent directions, as indicated by cross-bedding and current ripples, tend to be bipolar at a given locality (Fig. 40). In the type section, easterly and westerly currents are evident, implying tidal surges parallel to the basin axis.

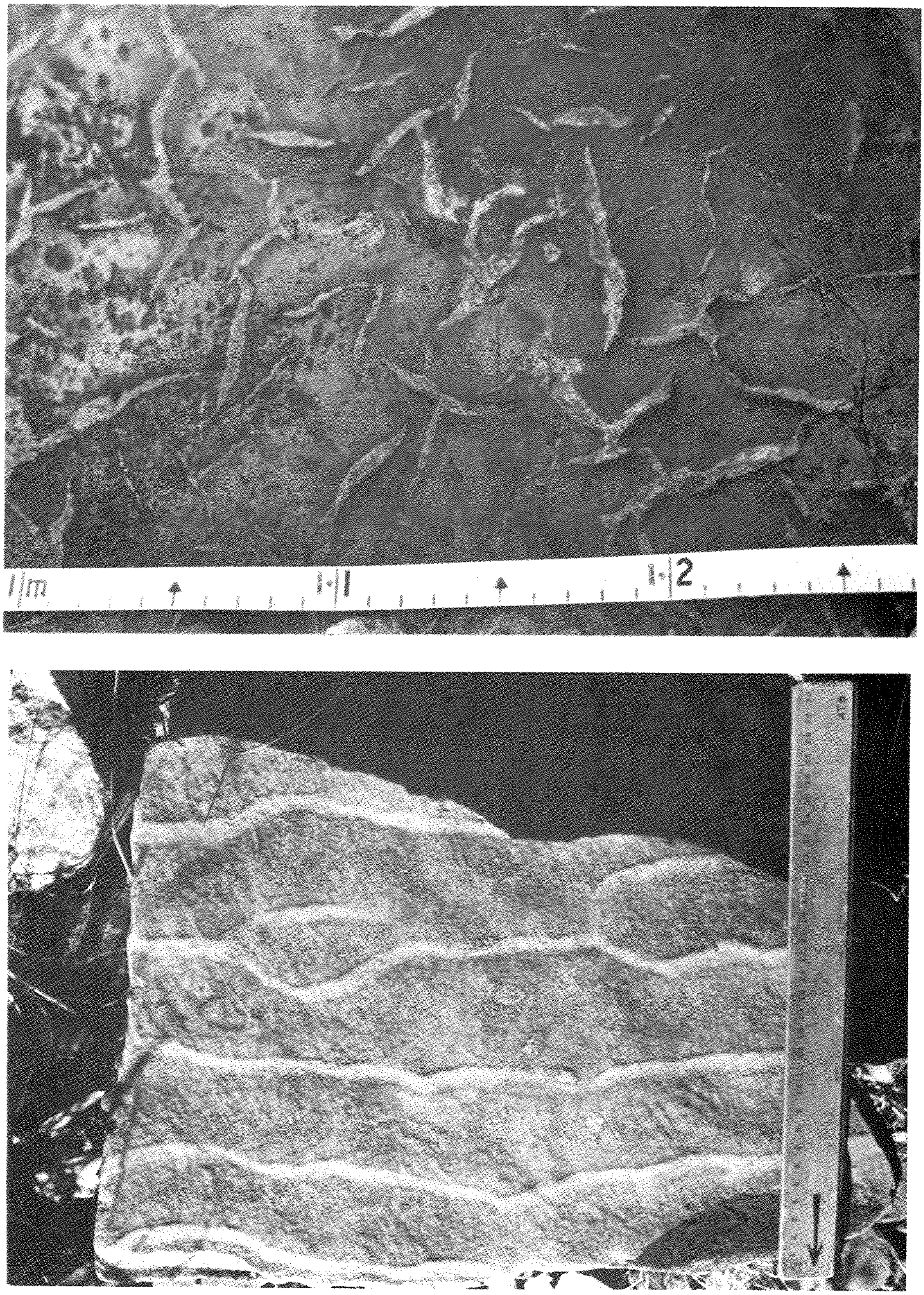
TABLE 11. DEPOSITIONAL FEATURES AND INTERPRETATION, UPPER BARRIER-BAR FACIES

Feature	*Interpretation*
Laminated bedding	Lower part of upper flow-regime, washover fan, swash, or surf zone; eolian or aqueous suspension settling
Alpha and omicron cross-bedding	Sand wave deposition by currents, e.g. inner offshore zone (Clifton and others, 1971) tidal inlets, washover fans
Pi cross-bedding	Migrating troughs of inner surf zone, outer nearshore zone, and inner offshore zone (Clifton and others, 1971)
Widespread, herringbone cross-bedding	Alternating tidal currents
Small symmetrical ripples (wave-length 5 cm)	Waves in water only centimetres deep
Strongly asymmetric ripples	Lower flow-regime currents
Occasional possible swash ripples (Fig. 38B)	Swash and backwash zone of beach
Interference ripples (Fig. 39A)	Wind-caused waves in very shallow water
Current lineations and rare bounce marks (Fig. 39B)	Lower part of upper flow-regime
Intraclastic sandstone and intraformational conglomerate layers	Short distance transport of mud flakes or chips
Occasional pebbly sandstone and granule or pebble conglomerate lenses	Very high energy conditions (e.g. tidal inlets and channels, nearshore zone)
Sparse, incomplete shrinkage cracks in fine-grained arenite (Fig. 38A)	Dehydration on subaerial exposure
Glauconitic beds	Marine derivation
Low-angle scours	High energy events (e.g. storm turbulence, rapid flow in tidal channels, nearshore wave disturbance)
Lutite members	Interfingering with adjacent environments; some backshore pools
Rare diamictite	Mudflow or turbidity current
Coarsely micaceous and feldspathic beds	First cycle detritus from granitic sources, with short residence time in highest energy zones
Deformed cross-beds and convoluted bedding	Current drag and compaction or slumping
Occasional flute molds	Eddies in traction or turbidity current
Wacke beds	Poor sorting in backshore areas or near small deltas

UPPER LAGOONAL FACIES

The lagoonal facies of the regressive marine linear clastic shoreline model applied to the Mucalana Sub group is represented by the Kurabuka Formation (Table 12).

Most of the sedimentary pile is a monotonous sequence of bright-green to olive shale and mudstone, some of it finely micaceous. Lamination indicates quiet water, which allowed continuous deposition of silt and clay from suspension. Occasional small ripples and interference ripples indicate areas of very shallow water. Pyritic chert, fine-grained glauconitic sandstone, and carbonate lenses, are consistent with marine conditions. These features point to a marginal marine lagoon.



A

B

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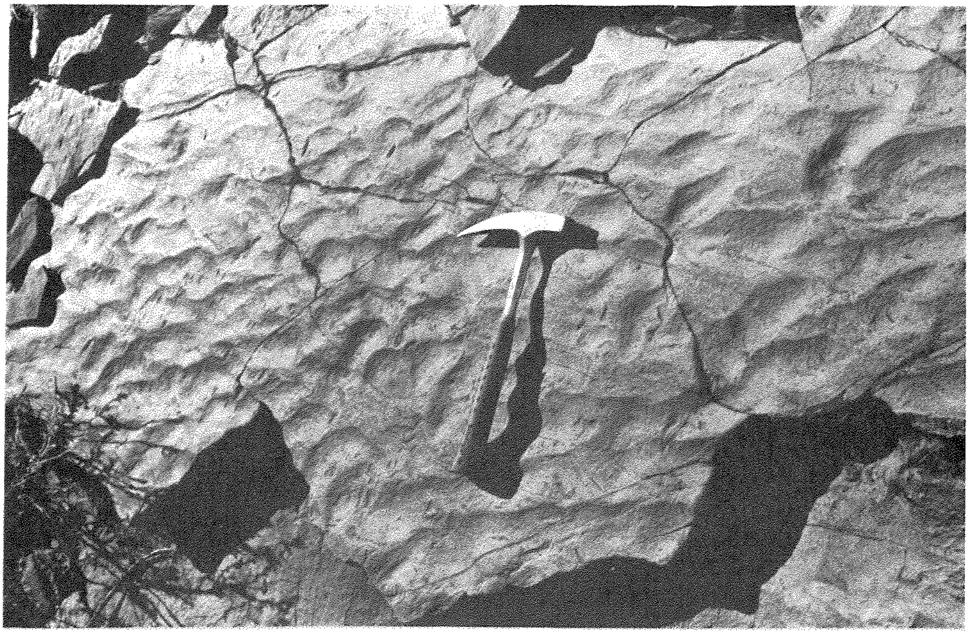
Figure 38. **Structures formed in shallow water, Mount Vernon Sandstone type section. A—Shrinkage cracks in fine-grained arenite indicate subaerial exposure. Compare with Conybeare and Crook (Fig. 516, 1968). B—Possible swash ripples in fine-grained quartz arenite.**



A

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Figure 39. A—Interference ripples (slightly deformed by loading) in fine-grained arenite in the type section of Mount Vernon Sandstone. B—Casts of current shadows and bounce marks on a bedding surface near that of A. Current direction was from left to right.

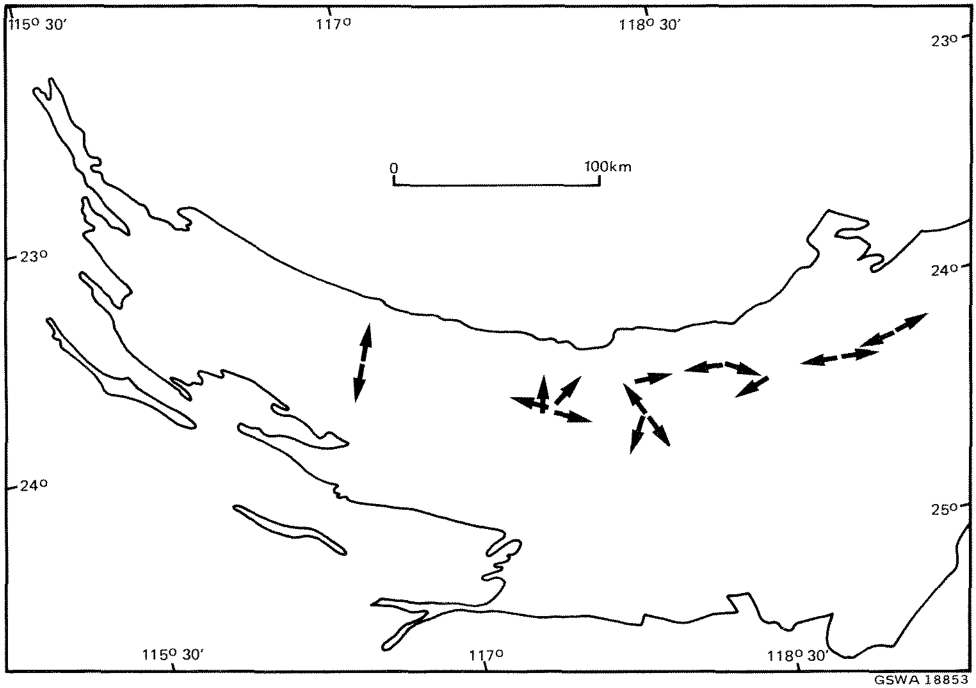


Figure 40. **Palaeocurrent directions in the Mount Vernon Sandstone. Diverse and opposing directions within the same outcrop are common. In many places, the bipolar directions are parallel to depositional strike and are consistent with the tidal circulation.**

TABLE 12. DEPOSITIONAL FEATURES AND INTERPRETATION, UPPER LAGOONAL FACIES

Feature	*Interpretation*
Laminated, fine-grained clastics	Suspension settlement in calm water
Occasional small symmetrical ripples (wavelengths 8-12 mm)	Water only centimetres deep
Scattered interference ripples	Wind action on very shallow water
Scattered faint current lineations	Lower part of lower flow-regime
Curled mud plates several centimetre in size	Subaerial desiccation
White specks in mudstone	Replacement of iron sulphide (reducing conditions) or of evaporite mineral
Occasional glauconitic fine-grained sandstone near base of formation	Marine derivation
Pyrite molds in chert	Reducing conditions during diagenesis
Limestone interbed	Marine environment and pause in clastic sedimentation
Calcareous siltstone interbeds with lustre mottling and some cone-in-cone concretions	Infiltration and growth of calcite in silt

DEPOSITION OF THE EASTERN FACIES

The Backdoor, Calyie and Ilgarari Formations had similar environments to those of their western facies counterparts, and thus correspond to the open-marine shelf, barrier-bar and lagoonal zones of a regressive marine epeiric shoreline.

Towards the eastern and northeastern extremities of the Bangemall Basin, deltaic, alluvial fan, beach and tidal deposition have been recognized in the Calyie, McFadden, and Skates Hills Formations.

EASTERN SHELF MUD FACIES

The Backdoor Formation (Table 13) resembles the Jillawarra (Table 7), and indicates a similar open-marine shelf. Its higher levels are laterally equivalent to the Ullawarra Formation and Fords Creek Shale, and it shares some of the features typical of these units, in particular the coarse-grained quartz siltstone (grading to very fine quartz arenite) and green shale. The top of the Backdoor, however, unlike the Fords Creek Shale, contains black chert, discussed in Chapter 5. The absence of green shale in the Backdoor Formation south of the latitude of the Glen Ross Anticline evidences a different composition of supplied sediment.

No drill-core samples of the formation are available. Black shale is occasionally present in outcrop, especially towards the top of the unit, and suggests that the usual shale is graphitic at depth like its Jillawarra counterparts. No algal filaments or other debris of definite organic derivation are known.

TABLE 13. DEPOSITIONAL FEATURES AND INTERPRETATION, BACKDOOR FORMATION

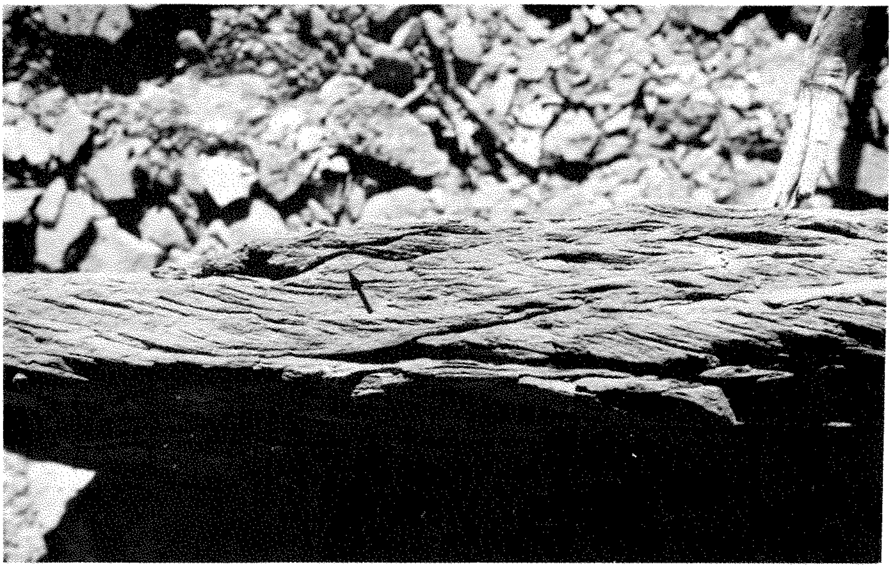
Feature	*Interpretation*
Laminated shale and siltstone	Slow settling out of suspension below wave base
Laminated dolomite members	Marine regime, below wave base
Sharp contacts between shale and siltstone layers	Sudden changes in grain size of clastic supply or rate of deposition
Pyrite cubes and blebs	Reducing conditions in sediment
Current lineations	Lower part of upper flow-regime currents
Occasional small scale (1 cm) cross-bedding	Small current ripples
Very occasional flute molds	Erosive traction or turbidity current
Occasional scour-and-fill structures	Wave base impinging on sea floor during heavy storms
Thin sandstone beds with sharp contacts	Storm sand layers
Occasional herringbone cross-bedding, in sandstone lenses	Reversing tidal currents
Mudstone intraclasts in sandstone lenses	Short transport distance of mud chips in shallow water
Large cone-in-cone concretions of limestone within shale	Diagenetic growth within sediment

A few small areas of laminated dolomite, and occasional dolomite beds up to 1 m thick, represent isolated localities on the shelf where quiet conditions were favourable to carbonate accumulation. Rarely, thin bands of the original limestone are present. Large cone-in-cone concretions, up to 1 m in diameter, of limestone within shale have formed in the centre of the Tangadee Syncline and 20 km east of Nicholl Spring. These are diagenetic structures that formed within a few metres of the sediment-water interface (Pettijohn, 1975).

Thin sandstone bands in the sequence are mostly regarded as storm-sand layers, and appear to be more common near the top and bottom of the formation, offshore from the barrier-bar and beach environments of the Calyie and Wonyulgunna-Lower Backdoor Formations. In the Mundiwindi district, quartz arenite lenses record tidal currents in herringbone cross-bedding, flute molds, current lineation and mudstone intraclasts derived by erosion of mudchips from nearby tidal flats.

Farther north at Watch Point, 10 km northeast of Jiggalong, the Balfour Shale of the Manganese Subgroup is correlated with the Backdoor Formation. No detailed sedimentary analysis has been done in this district, but general features of the sequence (Table 14) suggest a deltaic enviroment. Marine prodelta muds, deposited out of suspension below wave base, are succeeded upwards by coarser clastics, indicating a progressive shallowing of water as the delta built up. Climbing-ripple lamination, formed by deposition from suspension in a current, (Allen, 1963), contains lenses of reverse-dipping foresets (Fig. 41) ascribed to tidal action. The dominant flow direction is southerly to southeasterly, resulting from seaward outflow in the estuary. Also present are convolute beds, a common feature on the steeper slopes of sand bars in tidal enviroments (Wunderlich, 1967). Mudcracks and mudflake intraclasts imply the presence of channel-flanking tidal flats.

The sequence near Robertson Range homestead, 20 km south, may be interpreted similarly. It contains very broad, gentle trough cross-lamination and erosive scours (Fig. 42), which are due to the migration of lunate megaripples, and which are common in distributary mouth bar-deposits.



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5cm

Figure 41. **Climbing ripple lamination (Type A of Jopling and Walker, 1968) formed by deposition from suspension under the influence of a current. The two reverse-dipping forests (arrowed) were formed by flood-tide currents, as the dominant flow direction to the right (southeast) represents seaward outflow in an estuary. Siltstone of the upper Manganese Subgroup, Watch Point district.**

TABLE 14. SEQUENCES IN THE UPPER MANGANESE SUBGROUP AND BASAL CALYIE SANDSTONE, JIGGALONG AREA

Watch Point district	*Robertson Range homestead district*
Arenite, fine-grained; 10 per cent feldspar; 1 cm beds; cross-bedding	Arenite, fine- to medium-grained; massive to thickly bedded; thickness variable laterally because of erosion into underlying rock (Fig. 42)
Arenite, fine-grained cross-bedded, contorted	Arentite, fine- to medium-grained; laminated to thickly bedded, some contortion; broad, gentle, trough cross-bedding
Arenite, fine-grained; 5 per cent clay; 5-15 per cent feldspar	
Arenite, fine-grained; 1 m thick 10 per cent feldspar	Arenite, fine-grained; siltstone, coarse-grained; 1 to 10 cm beds; local intraclasts and ripples
Sandstone, 1 cm beds, contorted	Shale and siltstone, interlaminated green, micaceous; cut-offs and ripples in places; discontinuous bands of massive siltstone up to 10 cm thick
Arenite, fine-grained, green, micaceous, rippled; some mudcracks and intraclasts	
Siltstone, micaceous; mostly coarse-grained; mud-flake intraclasts; climbing ripples; reversals of current direction (Fig. 41)	
Mudstone and siltstone, fine- to coarse-grained, laminated, olive green, micaceous; many cut-offs and erosional bases in massive siltstone layers	

No implied correlation between columns.

G.S.W.A. 18855

Figure 42. Broad, gentle trough cross-bedding formed by lunate megaripple migration on the distributory mouth bar of a delta, and later eroded by channelling. Arenties of the basal Calyie Sandstone near Robertson Range homestead.

The general impression given by the Balfour Shale is of a southward-prograding, tidally influenced, deltaic shoreline. However, in the absence of more detailed information on sedimentary features and the spacial distribution of rock bodies this interpretation is tentative.

EASTERN SAND FACIES

The deposition of the Calyie Sandstone extended over an area of almost 100 000 km^2 and resulted in the laying down of a quartz arenite sheet 1 km thick. More than one environment was involved.

The sedimentary structures are mostly typical of subaqueous deposition (Table 15). There is not xi cross-stratification, and the style of cross-bedding is not that expected of an eolian regime. The vast majority of cross-bedding sets are usually 0.1 to 0.3 m thick. One instance of backflow current ripples on the toe of an asymptotic foreset was noted; these are formed by reverse eddies on the leeside of a megaripple and do not form in eolian conditions. A few re-activation surfaces (Fig. 44) occur: this structure is considered by De Raaf and Boersma (1971) to be characteristic of tidal deposits. Kumar and Sanders (1976) record them in connection with inlet sequences formed by the migration of a barrier island, and they can also form in

TABLE 15. DEPOSITIONAL FEATURES AND INTERPRETATION, EASTERN SAND FACIES

Feature	*Interpretation*
Evenly laminated bedding	Lower part of upper flow regime in swash or surf zone; suspension settling; slip face of tidal current ridge
Alpha, beta, and omicron cross-bedding	Sand-wave deposition by currents, (e.g. tidal inlets, flanks of tidal ridges)
Pi and nu cross-bedding	Migrating linguoid ripples of inner surf zone, outer nearshore zone, and inner offshore zone
Herringbone cross-bedding and outcrops with reversed current directions	Opposed tidal currents
Reactivation (discontinuity) surfaces within cross-bedded sequences	Tidal activity
Symmetrical ripples and interference ripples	Wave oscillation ripples formed in shallow water
Flat-topped ripples	Ripples planed off by waves in very shallow water
Current lineations (Fig. 43)	Lower part of upper flow regime
Asymmetrical ripples on toe of foresets with steep sides facing foresets	Current ripples formed under water by backflow in eddy in front of sand wave slip face
High variance (=7207) of palaeocurrents; occasional lenses of dolomite, dolomite-cemented sandstones, and glauconitic sandstone	Marine regime
Conglomerate lenses; pebbly and cobbly sandstones, broken waterworn cobble	Turbulent and violent conditions
Shale and siltsone lenses	Suspension settling in quiet water, (e.g. small lagoons behind sand bars, parts of tidal shelf protected from erosive currents)
Siltstone intraclasts	Short distance transport of mud flakes or chips
De-watering structures	Upwelling and expulsion of interstitial water during compaction of sand
Pyrite masses, small, irregular, rare	Reducing conditions within sediment

G.S.W.A. 18856

Figure 43. Current lineations on the bedding surface in sandstone, Calyie Sandstone, Coondia Coondia Spring (ROBERTSON). The pen indicates the direction of the palaeocurrent towards the southwest as deduced from some small V-shaped markings.

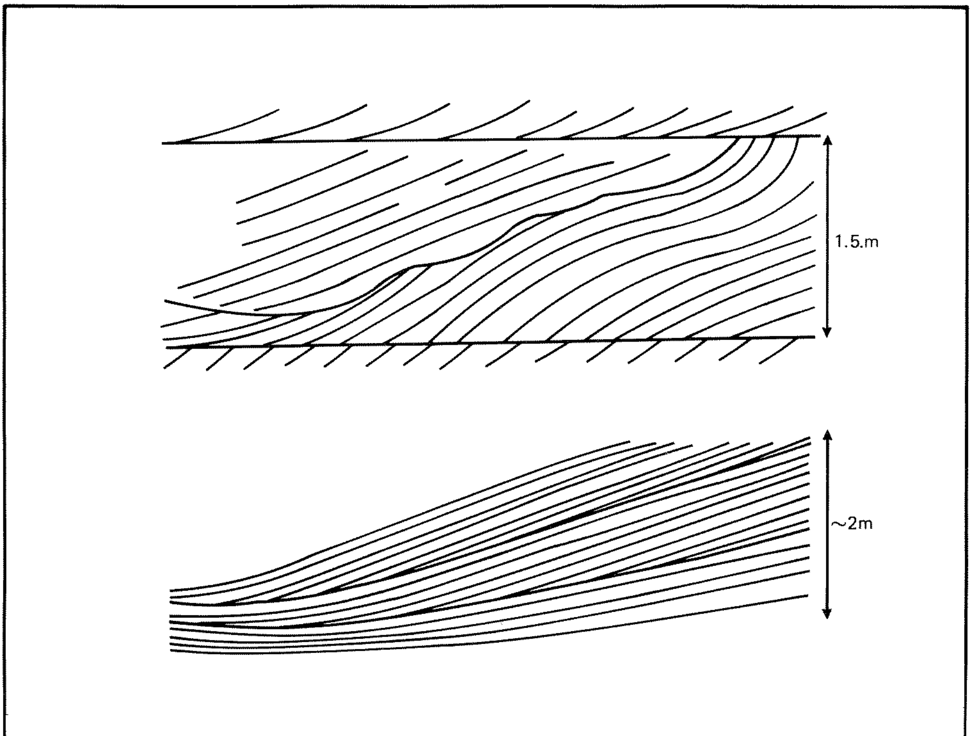

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Figure 44. Channel and other surfaces (probably resulting from tidal activity) within cross-bedded Calyie Sandstone from northwestern BULLEN. Sketches of vertical exposures.

rivers (Collinson, 1970). The presence of carbonate-cemented sandstone (de la Hunty, 1964), dolomite and glauconite in local but widely separated areas argues that most, if not all, of the formation is marine. The irregular pattern of foreset directions (Fig. 45) and their high variance is inconsistent with a fluvial or eolian origin (Potter and Pettijohn, 1963), and also points to marine conditions.

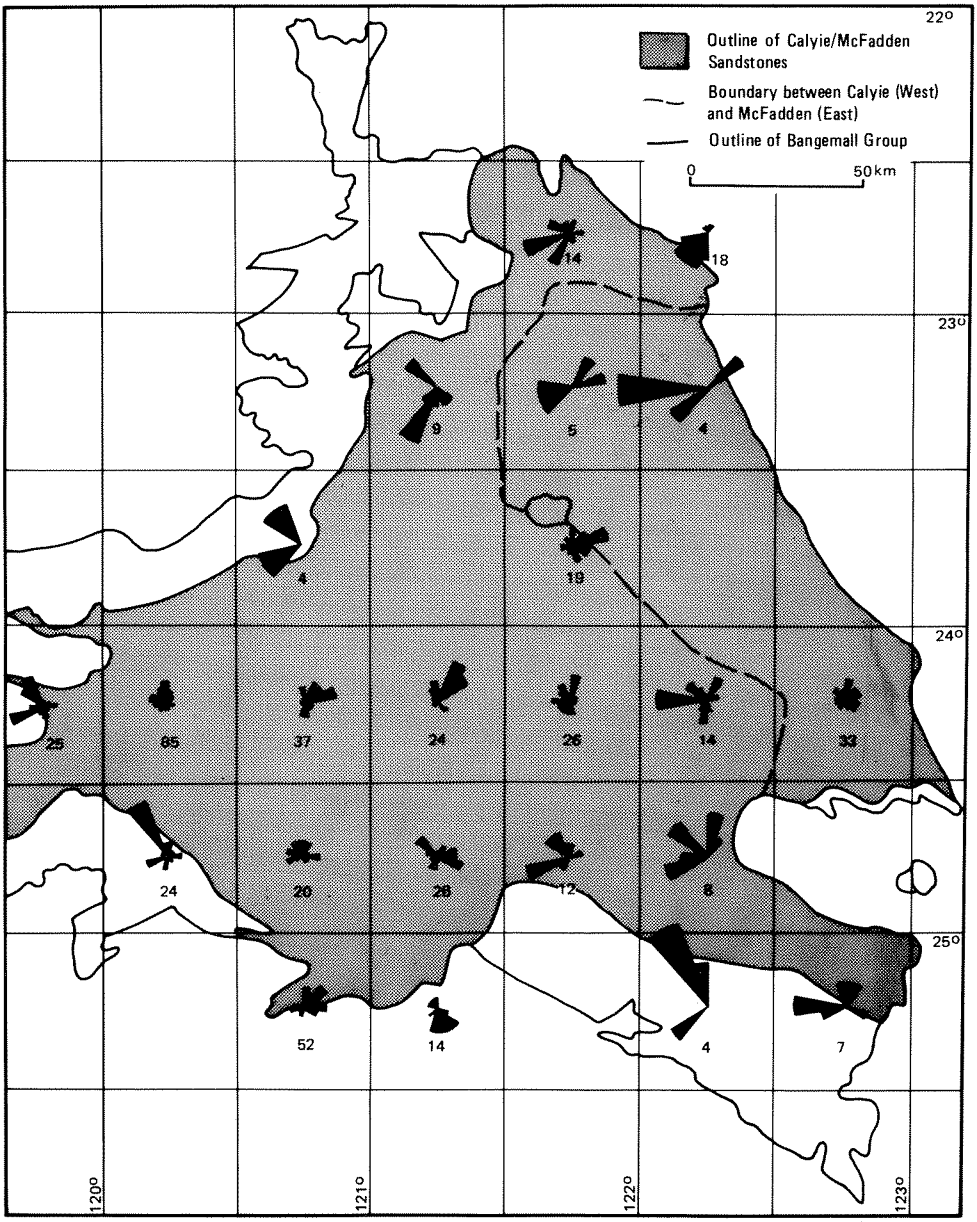

Figure 45. **Palaeocurrent roses for the Calyie and McFadden Sandstones, eastern Bangemall Basin, illustrating the diversity of current directions and lack of consistent pattern. Each rose sums the data from cross-bedding in its ½° quadrangle. Class intervals are 20°, and lengths from the centres are proportional to the percentage of data in each class. The number of measurements in each quadrangle is shown below the rose.**

Symmetrical ripples, including some with wavelengths less than 5 cm, and interference ripples, are widespread and indicate shallow water. Minimum water depths can be calculated from cross-bedding thicknesses (Allen, 1970, p. 78-79). These depths were 1-2.5 m for the usual 0.1-0.3 m thick sets, and 40 m for a 6 m thick set. Topset beds are rarely preserved, so that the present set thickness is less than the original height of the megaripple. In two sets where topset beds had not been scoured away, the minimum flow depths were 2 m and 2.5 m.

The pebbles and conglomerate lenses in the sandstone, required high-energy conditions for their transportation, and they could not have been carried far offshore. Cobbles and boulders can be moved by violent storms; violent events are corroborated by one *in situ* clast, which consisted of half of a well-rounded cobble showing an unworn fracture surface. Subrounded, subangular and angular clasts testify to short transport distances. Most of the clasts are vein quartz and quartz arenite, which could have been derived from units underlying the Bangemall Group, such as the Earaheedy Group, Scorpion Group, Cornelia Sandstone, and Yeneena Group. A possible setting for these conglomerates was on the current-swept channel floors of tidal inlets between barrier islands. Other possible inlet sequences, similar to those noted by Kumar and Sanders (1976), are present in various places in the Calyie Sandstone. The boulder conglomerates of northeastern ROBERTSON, which contain clasts of conglomerate and sandstone up to 2 m across, must have been derived from their immediate vicinity and may be fluvial.

Intraclasts of soft siltstone could not have survived more than a few hundreds of metres from their sites of origin. Those as large as 25 cm long would have been moved only a few metres. They were probably derived from dried, silty mud which was broken by desiccation cracks or wave action, and strewn along a shallow, sandy bottom. Lenses of siltstone and shale, usually less than 0.5 m thick, are widespread over the region, although they are not abundant, probably because many were eroded to become the source of intraclasts. Such bodies would have been deposited in sheltered areas, for example behind bars in small lagoons.

In eastern BULLEN, form lines (Fig. 46) indicate a sequence dipping consistently northeasterly for about 100 km into a shallow structural basin. If the form lines represent horizontally deposited bedding, a very great thickness of about 10 km would be present. More reasonable is the deposition of the sandstone in this region as a series of wedges, analogous to the formation of cross-bedding on a smaller scale. Such wedges may be generated by the migration of tidal current ridges of the type described from the North Sea by Houbolt (1968). These ridges occur in groups on a tide-swept shelf floor, and are up to 40 m high and 65 km long. They are not readily accessible to study, but sparker profiles of the North Sea ridges show tabular internal cross-bedding. The ridges lie at a small angle to tidal current directions and migrate slowly at right angles to their long axes. Sand is transported obliquely towards the crests as megaripples, and then down the slip faces. In BULLEN, current directions fit this pattern by usually trending either at a low angle to the form lines, that is, northeasterly (Fig. 46). The bedding would thus represent the slip faces of tidal current ridges which migrated northeasterly.

The Calyie Sandstone was deposited variously in shoreline to tidal-shelf regimes, but the unit is diachronous and all these conditions did not prevail over the whole of its extent. The basal portion, where it is laterally equivalent to the

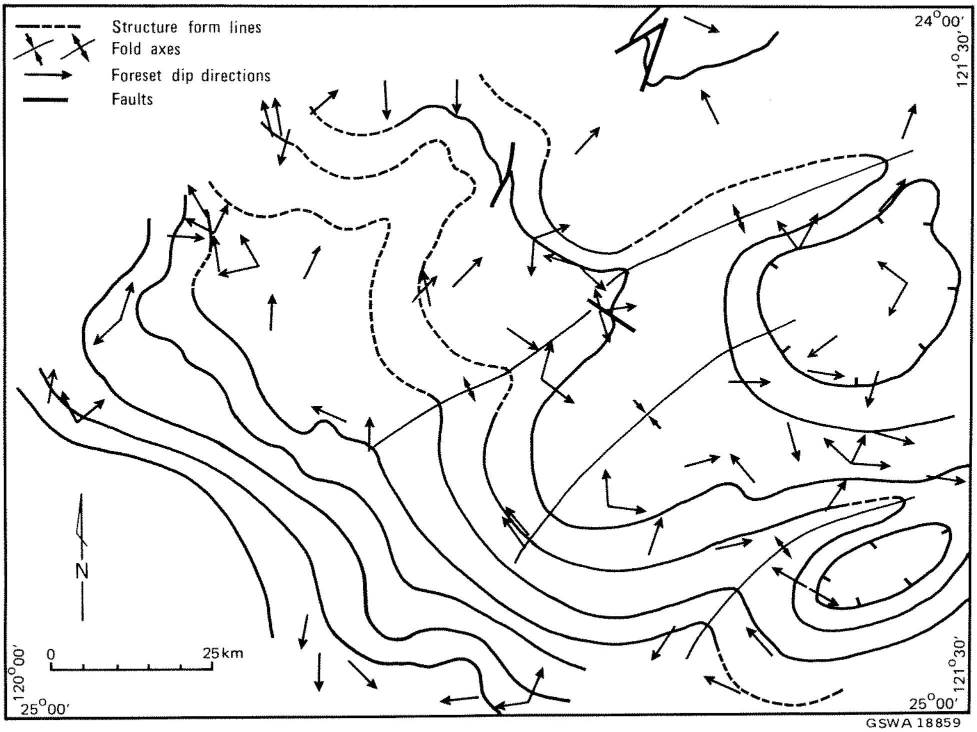

Figure 46. **The relation between current bedding and bedding in the Calyie Sandstone suggests that the sandstone was deposited as a series of tidal ridges. The dip of most foresets is either subparallel to the structural form lines (parallel to the tidal current) or in a direction normal to them, usually to the northeast (parallel to the slip faces of ridges) migrating northeast perpendicular to the tidal current. The structural form lines, extrapolated from the strike of bedding, are dashed where the data are insufficient.**

Wonyulgunna Sandstone, is interpreted as a transgressive beach phase. Following a stand-still of the shoreline the remainder of the sequence was laid down during a regressive marine period. The latter phase is all that is present in the type section and the outcrop area west of longitude 120°30'E. The lensing-out of the offshore facies of the Backdoor Formation about half way across BULLEN suggests that the area of sand deposition extended a considerable distance from the coast in the eastern region and implies that the bulk of the sedimentary input was sand. This would have come from older sandstones and granitic rocks in the basement. This sand was probably first deposited as a transgressive coastal sand system, and then reworked by tidal currents.

The great thickness, 1 250 m, in the type section is difficult to explain by reference to modern analogues, and may mean that conditions in the Proterozoic were markedly different from the present, in respect to tidal ranges and greater land erosion.

DUNE CROSS-BEDDED SUBFACIES

The large-scale cross-bedding of the southern Carnarvon Range-Mount Davies region was formed in sand dunes at least as high as the cross-bedding set thicknesses, and indicates either a tidal-current or eolian environment. If eolian, the

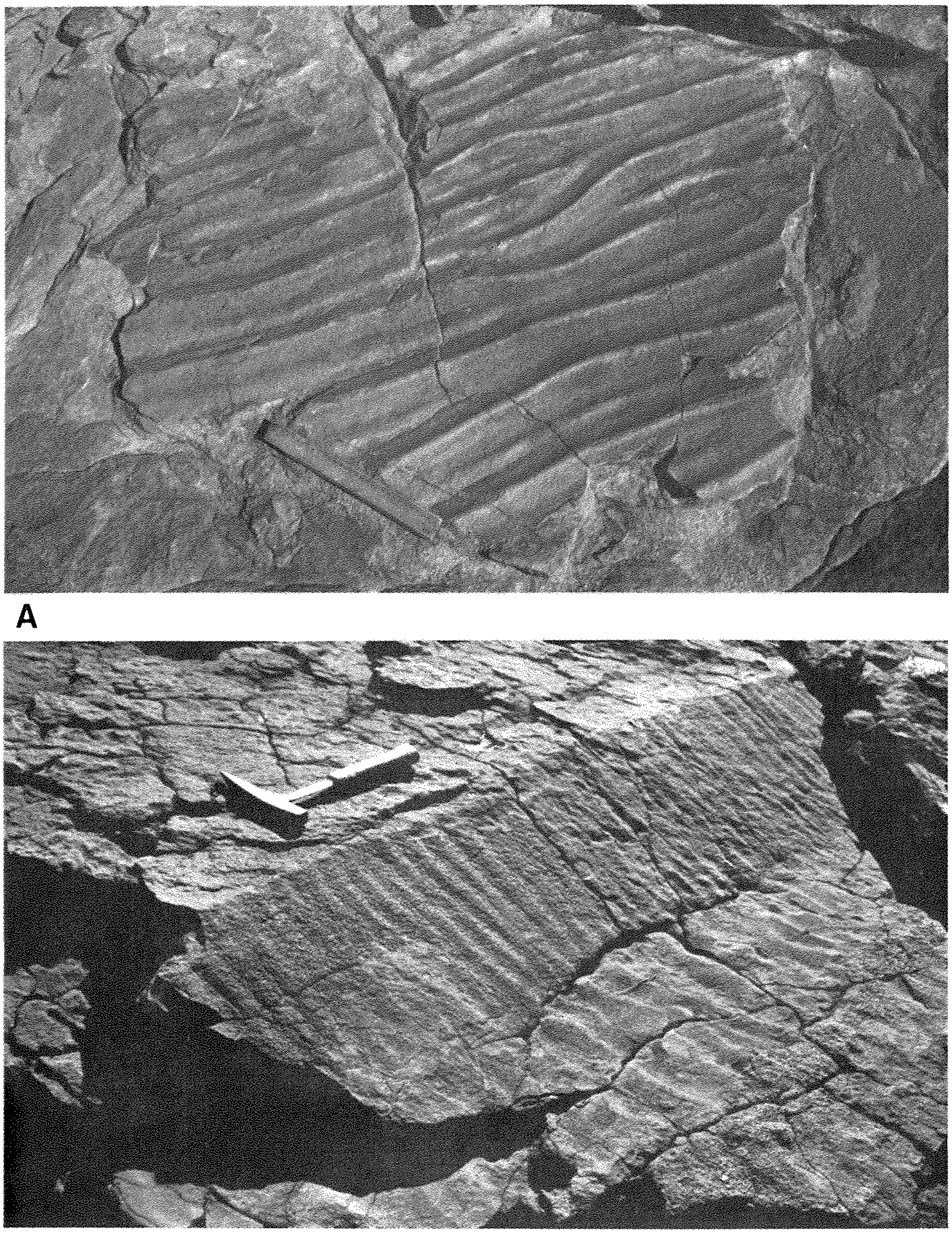

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Figure 47. Symmetrical ripple marks in the basal sandstone unit below the dune cross-bedded subfacies, southern Carnarvon Range. These were formed by the oscillation of waves in shallow water. Some other symmetrical ripples in the same unit were wind generated. A.—Wave oscillation ripples with pointed crests (ripple index RI=9.5; ripple symmetry index RSI=1.1; continuity index=12) Scale is 0.3 m. B.—Wave ripples (RI=12.5, RSI=1.0) formed on a forest slope.

dunes would have been coastal and not desert features, because the subfacies lies in a narrow zone along the margin of the basin. The adjancent region of the Calyie Sandstone was marine, and the laminated, mostly flat-bedded, basal sandstone sequence below the subfacies was deposited in water.

Evidence for deposition in water of the basal unit includes lenses of pebbly sandstone and conglomerate, symmetrical wave-formed ripples (Fig. 47), interference ripples, and current lineations. Some well-rounded, water-worn conglomerate clasts are up to 30 cm in diameter, and must have been deposited near the shore line. Most ripple marks in this lower sequence when plotted on the RI vs RSI diagram of Tanner (1967) lie in the wave-ripple field (Fig. 48). A significant number, however, plot as wind ripples and appear to be genuine, considering that their position on the diagram could not result from the compaction of coarse sand. Three examples of flat-topped ripples suggest planing-off by waves in very shallow water (Tanner, 1967), or by wind (Steidtmann, 1974). Continuity indices (crest length divided by average wavelength) obtained for 2 ripple trains indicate formation by waves in water. The wavelengths in some of the trains are less than 5 cm, demonstrating very shallow water (Tanner, 1971), as do the interference ripples. Hence the depositional conditions of the basal beds included shallow water and some wind action, such as on a beach or off-shore barrier bar.

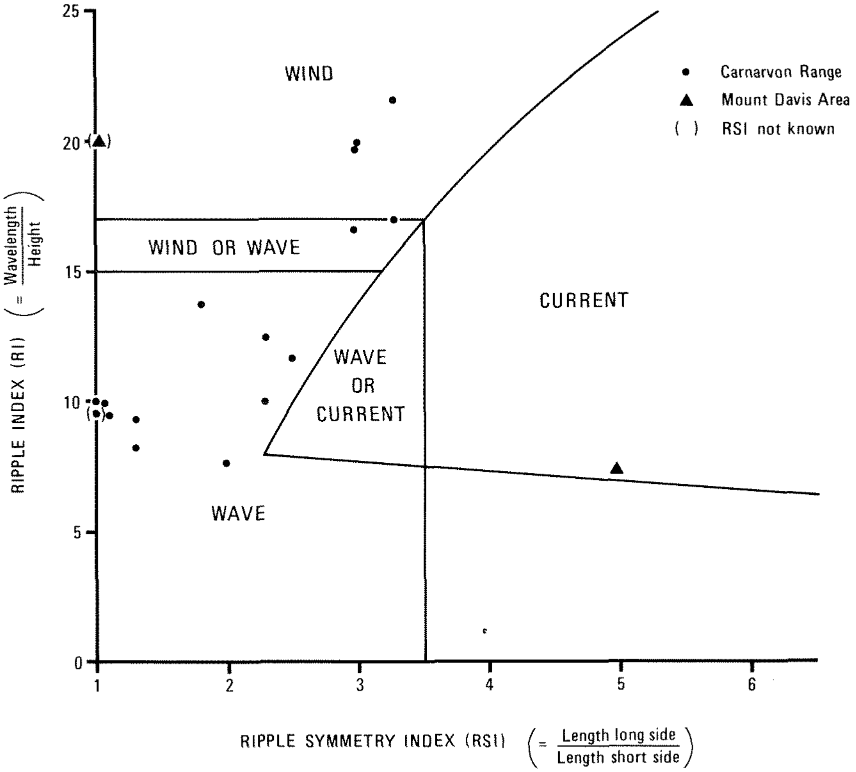

Figure 48. **Measured ripples from the Calyie Sandstone of the Carnarvon Range—Mount Davis area are plotted on an RI vs RSI diagram to determine types and orgins. The sole current ripple is from a lens within the dune cross-bedded subfacies near Mount Davis. The rest are from the basal unit underlying the dune cross-bedded subfacies.**

At the time of writing (1977), reliable criteria to discriminate between eolian and tidal-current cross-bedding do not exist. The genesis of even the well known Navajo Sandstone of the western U.S.A., which is similar to the dune cross-bedded subfacies, is controversial (Freeman and Visher, 1975). Absence of shale interbeds (Selley, 1970) and mica (Glennie, 1970) has been held to point to an eolian origin, but this can also be true for tidal environment. The presence of well-sorted, rounded and frosted grains is likewise not diagnostic (Freeman and Visher, 1975). Festoon cross-bedding occurs in coastal dunes (McKee, 1957), but similar cross-bed sets have been reported by van de Graaff (1972) from beds analogous to the subtidal current deposits of the North Sea. Efforts have been made to use settling velocities of grains (Steidtmann, 1974) and log-probability plots of grain size (Freeman and Visher, 1975) to distinguish eolian and tidal deposits, but these methods cannot be used with silica-cemented sandstone.

Although several features of the dune cross-bedded subfacies could be construed as favoring an eolian origin, we believe that a tidal current origin is more likely, for the following reasons.

(1) The sandstone is only moderately sorted, whereas eolian deposits are generally very well sorted at least within individual laminations, (Glennie, 1970; Reineck and Singh, 1973, p.290). The grain size ranges commonly from 0.5 to 1 mm, and some layers contain granules, in contrast to typical wind-blown sand which is fine to medium grained (Reineck and Singh, 1973), or according to Allen (1970), has a mean size of 0.2 to 0.4 mm. Some of the grains in foresets are smeared with white clay, contrary to what would be expected in wind-blown sand.

(2) The maximum foreset dip after correction for tectonic tilt is 30°, and less than a quarter of the readings exceed 25°. This is compatible with the maximum repose angle of subaqueous sand of about 30° (Hoyt, 1967) to 35° (Conybeare and Crook, 1968), and contrasts with dips ranging to 30-40° in coastal dunes (Reineck and Singh, 1973).

(3) Near Mount Davis, a bedding surface in a flat-bedded lens within the dune subfacies, contains asymmetrical current ripples showing that this at least was deposited under water. No wind ripples occur on any of the large-scale cross-bedding surfaces.

(4) A sandstone dyke (Fig. 49) intrudes the large cross-bedding near Mount Davis. This implies a mobile sand-water mixture intruding cohesive sand.

(5) The convex-upward foreset laminae and soft-sediment deformation which are common in coastal dunes (Reineck and Singh, 1973) were not observed.

(6) The low variability of foreset dip directions at a given locality (Fig. 50) is expected for eolian conditions, while in a tidal environment, bipolar directions may be present (De Raaf and Boersma, 1971); but if the main tidal ebb-and-flood currents follow mutually exclusive paths, only one direction may be preserved. This is the case elsewhere in the Calyie Sandstone. Also, in the Carnarvon Range the palaeocurrent direction lies along the presumed basin margin, but near Mount Davis the direction is normal to this. Onshore winds are unlikely to vary as much in this short distance.

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Figure 49. Sandstone dyke intrusive into a large cross-bed set in the Calyie Sandstone near Mount Davis.

The vertical profiles exposed in the Carnarvon Range-Mount Davis region are thus thought to record a beach deposit being transgressed by tide-swept shelf sands, rather than representing a beach environment passing into coastal dunes.

EASTERN LAGOONAL FACIES

The Ilgarari Formation, overlying the Calyie Sandstone, represents the lagoonal facies of the regressive marine shoreline, interpreted for the Collier Subgroup. The poor exposure and restricted area of occurrence has led to few sedimentary structures being found, but the well-bedded and laminated nature of the unit, and the prevailing shale-siltstone-claystone lithology are consistent with lagoonal deposition. Fine-grained sandstone and coarse-grained quartz siltstone contain isolated, small-scale tangential cross-bedding produced by low-energy currents. This rock may record an intertidal or slightly deeper environment analogous to that in Jade Bay on the North Sea (Reineck and Singh, 1973).

EASTERN BASAL FACIES

The Skates Hills Formation, on the eastern margin of the Bangemall Basin, has a basal, lenticular, boulder-bearing cobble conglomerate, up to 30 m thick, which consists of well-rounded clasts of the underlying Cornelia Sandstone. This deposit is interpreted as a series of small alluvial fans coming off the basement sandstone ridges (Table 16). The sandstone sequence overlying the conglomerate

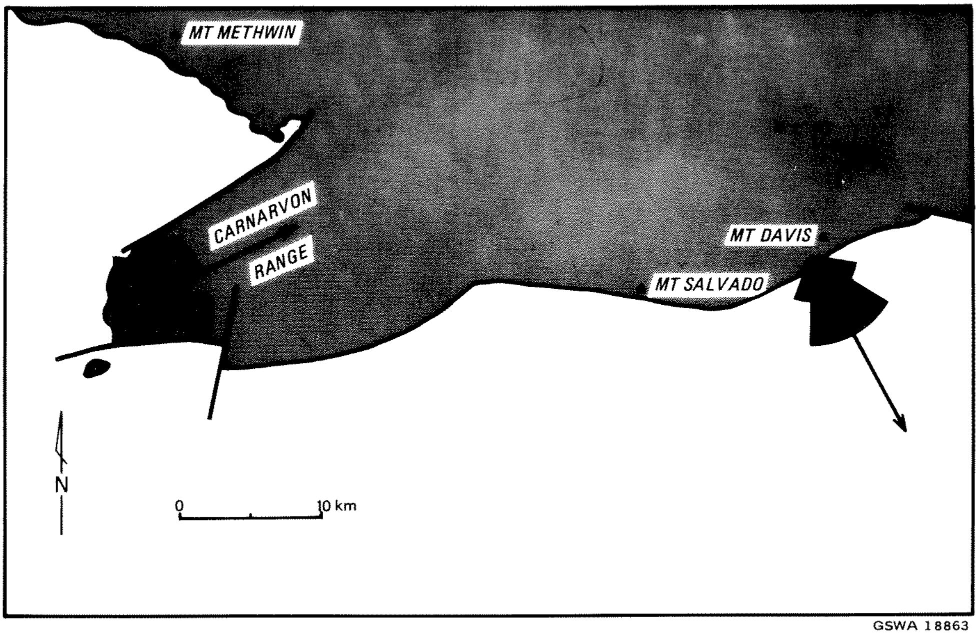

Figure 50. Cross-bedding azimuths in the dune cross-bedded subfacies of the Calyie Sandstone (NABBERU). Rose diagrams show mean direction; class intervals are 20°—lengths are proportional to the percentage of data in each class. Stippling indicates the Calyie Sandstone.

comprises fine- to coarse-grained, moderately to well-sorted, laminated quartz arenite. The presence of some current lineations and lack of cross-bedding suggests that the sand, probably supplied by the alluvial fans, was reworked in a beach zone.

TABLE 16. DEPOSITIONAL FEATURES AND INTERPRETATION, EASTERN BASAL FACIES

Feature	*Interpretation*
Basal conglomerate, up to 30 m thick	High-energy deposits of alluvial fans
Sandstone lenses, up to 10 per cent clay cement	Incomplete sorting in cutoff channel or channel pool on alluvial fan
Sandstone, current lineation	Transitional flow regime or lower part of lower flow regime
Laminated sand	Beach deposition
Glauconite in laminite lens	Marine origin
Dolomite, algal lamination with birdseye structures	Stratiform stromatolites (algal mats), subject to subaerial exposure.
Acaciella, columnar stromatolites	Shallow subtidal to lower intertidal growth conditions
Stromatolite columns, aligned; edgewise intraclasts between columns	Wave action
Goethite cubes after pyrite, rare, up to 1 cm on edge	Reducing conditions during diagenesis
Interference ripples and small symmetrical ripples (wavelength 2 cm) in overlying sandstone	Very shallow water

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Figure 51. Part of a stromatolitic bioherm in the Skates Hills Formation (24''48'S, 122°58'E) on southeastern TRAINOR. Both plan and vertical exposure are visible. The stromatolites have been identified as *Acaciella* cf. *australica* by Grey, 1978.

The top of the formation is a stromatolitic dolomite which usually rests on shale. The dolomite occurs variously as a single unit 10 m thick (B), or as four beds (A) separated by siltstone (Table 17). Algal-mat lamination and columnar stromatolites (Fig. 51), interpreted by Grey (1977) as shallow subtidal and lower intertidal, are common. At one outcrop an alignment of the columns was recorded, probably resulting from prevailing wave direction as in modern stromatolites at Shark Bay (Playford and Cockbain, 1976). Edgewise intraclasts between some columns also indicate wave action, but there are also intercolumnar layers of very fine-grained micrite, indicating sheltered spots. Irregular birdseye structures (fenestral fabric) up to 15 cm long and filled with chalcedony demonstrate subaerial exposure of the algal mats (Wilson, 1975).

The sandstone-shale sequence of the basal part of the overlying McFadden Sandstone contains interference ripples and small symmetrical ripples (wavelength 2 cm), again demonstrating very shallow water.

The Skates Hills Formation thus displays a succession of environments from alluvial fan, through beach deposition, to intertidal and subtidal algal buildups.

TIDAL SAND-TONGUE AND CHANNEL FACIES

The eastern and northeastern regions of the Bangemall Basin contain diverse sandstone types, mapped as McFadden Sandstone and adjacent parts of the Calyie Sandstone.

TABLE 17. EXAMPLES OF DOLOMITE PROFILES, SKATES HILLS FORMATION

A	*B*
Sandstone, medium- to coarse-grained, clay cement, interference ripples and small ripples	(silcrete)
Siltstone	
Dolomite, 0.3 m; stromatolites	Dolomite, 6 m, buff and pink, finely crystalline or aphanitic; columnar stromatolites at top; algal-mat layering with birdseye (fenestral) structures and pyrite pseudomorphs (rare) below
Siltstone, soft, pink	
Dolomite, 2 m, stromatolites	
Siltstone, calcareous	
Dolomite, 3 m, white and pink; stromatolites, large at base	
Siltstone, calcareous	
Dolomite, 3 m	
Shale	Sandstone, fine-grained; some fine- to coarse-grained wacke

No implied correlation between columns.

The central part of this region, around the McFadden Range and Durba Hills, has a distinctive lithology not found elsewhere in the basin. The main rock type is flaggy, thinly interbedded arenite and wacke, with varying proportions of martrix, pebbles, and lithic and feldspar grains. Southwards, westwards, and northwards many of these features are lost as the lithology changes. These trends are unlikely to result from environmental variation alone. The unusual composition of the rocks in the area is thought to be due to a great deal of mud, feldspar and lithic detritus from a nearby stream mouth. The only older rocks known to the east, however, are sandstones of the Yeneena Group, but non-outcropping interbedded shales, or schist inliers of the Rudall Metamorphic Complex may have provided the sediment. Diagenetic alteration or weathering of volcanic ash may have been responsible for some of the clay matrix, as volcanic quartz grains of ash-fall origin are known in some beds (A. Goode and M. Hall, BHP, pers. comm.). This contribution is small, however, because tuffaceous beds appear to be few, and most feldspar is fresh.

The poorly sorted sediments were deposited in an environment that did not promote the winnowing of mud (for example, tidal flats not subjected to strong wave action). Wackes are known from intertidal environments (Thompson, 1975). Indicators of shallow water (Table 18) are present, and the dispersion of cross-bedding directions (Figs. 45 and 52) indicates tidal rather than fluvial or non-tidal deltaic deposition. The deposits may have originated on a tidal coast characterized by sand tongues and channels. Sand tongues are seaward extensions of intertidal flats into the subtidal zone and are located between channels. As the channels migrate laterally, accretion on the advancing flank of the tongues may generate large-scale cross-bedding such as found in the McFadden Sandstone. Giant ripples in the channels can also deposit large-scale cross-bedding. Evenly laminated sand is present in sand tongues, especially above wave base. A modern analogue may be the region of the Nordergrunde on the German North Sea coast (Reineck and Singh, 1973).

On TRAINOR, where the sands were better sorted, the formation has many of the features of Facies "C" in the lower Bald Eagle Formation (Ordovician) of the Central Appalachians (Thompson, 1975). This was interpreted as a delta front distributary channel deposit, but differs from the McFadden in that cross-bedding is unimodal. Possibly the McFadden Sandstone is, in part, such a deposit, but it was influenced by tides, so that the outer delta margins were reworked into sand-tongue and channel bodies. The large southeasterly trending channel-like bodies of cobble conglomerate on northwestern GUNANYA (and perhaps the related boulder conglomerate in the Calyie Sandstone near Coondra Coondra Spring) represent fluvial channels, which were probably developed on prograding deltas, advancing to the southeast. The strong southwesterly cross-bedding trend in the adjacent Calyie Sandstone on RUDALL may be due to ebb-dominated tides rather than to derivation from the northeast, because the sandstone contains clasts of jasper. Jasper is not known from the Paterson Province, but is present in the Hamersley Basin to the northwest.

TABLE 18. DEPOSITIONAL FEATURES AND INTERPRETATION, SAND TONGUE AND CHANNEL FACIES

Feature	*Interpretation*
Granule to cobble conglomerate and pebbly sandstone in thin beds	Scour lenses and channel infills
Occasional symmetrical and interference ripples; shrinkage cracks	Lag deposits in channels; shallow-water areas; some subaerial exposures
Wacke layers and clay matrix in arenites; some graded bedding in wacke laminations	Proximity to river mouth(s) discharging mud; sedimentation below wave base; possible tuffaceous component
Abundant large to very large (8 m or more thick) pi and omicron cross-bedding	Lateral accretion on migrating sand tongues between channels; megaripple cross-bedding of giant ripples in channels; and subtidal current ridges
Medium-scale cross-bedding	Migrating ripples
Rare theta cross-bedding	Scouring during high-energy periods (e.g. strong tides and storms) and subsequent filling
High variance of cross-bedding (= 10 175)	Variable currents of tidal regime
Bi-polar palaeocurrents at some outcrops	Ebb and flood currents
Sandstone well-bedded, laminated, flaggy	Subtidal to intertidal sand tongue deposits
Well-sorted laminations	Supply of clean sand and/or deposition above wave base
Chaotic bedding affecting some outcrops of cross-bedded sandstone	Slumping of over-steep slopes of sand tongues due to removal of hydraulic support when exposed above water, undermining by migrating channels, etc.
Microfaulting with 1 cm displacements	Failure in unconsolidated but cohesive sand, probably related to slumping
Shale and siltstone intraclasts	Derivation from nearby tidal mudflat
Minor lutite interbeds	Slack-water sedimentation
Volcanic grains in some outcrops	Layers with ash-fall tuff component
Iron-oxide concretions (rare) up to 1 cm across	Pyrite balls formed during diagenesis
Large, channel-like bodies of cobble conglomerate trending southeasterly (GUNANYA)	Channels on a delta advancing from the northwest

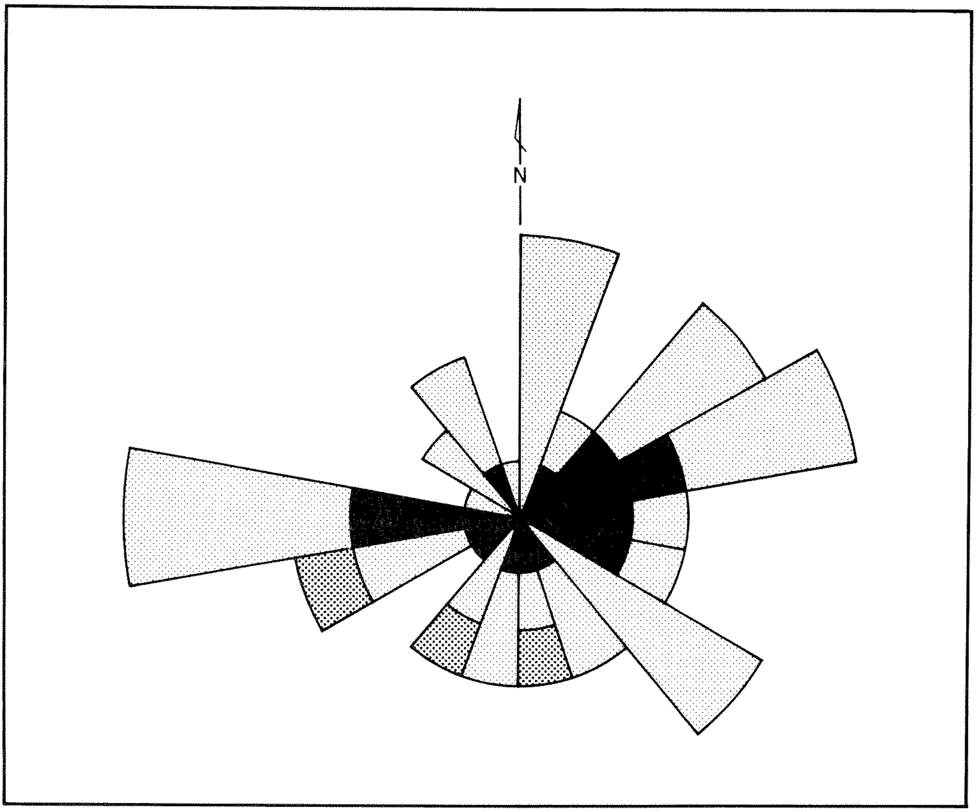

Figure 52. **Rose diagram of cross-bedding dip directions in the McFadden Sandstone; 54 measurements, 20° class intervals. GUNANYA readings are plotted in black, TRAINOR readings as light stipples, and RUDALL readings as dark stipples. GUNANYA has a strong bipolar distribution, reflecting tidal currents. The diagram as a whole is less strongly bipolar because of greater current variability in the larger region.**

The McFadden Sandstone is thus envisaged as recording a sandy intertidal enviroment which was succeeded by deltaic and eventually fluvial deposition. As the shoreline migrated westwards, the coastal conditions changed to the beach, barrier-bar and tidal-inlet environments of the Calyie Sandstone, with tidal current ridges being present on a sandy tide-swept shelf farther offshore.

TABLE 19. DEPOSITIONAL FEATURES AND INTERPRETATION DURBA SANDSTONE

Feature	*Interpretation*
Scouring at base of formation	Fluvial erosion
Conglomerate lenses, including boulders	Channel lag deposits
Claystone lenses near base	Suspension settlement in backwaters of braided stream system
Occasional Interference ripples	Shallow water in pools
Occasional intraclasts	Dried mud flakes derived from minor overbank or backwater deposits
Asymmetrical ripples	Current ripples
Omicron and pi cross-bedding	Migrating large ripples in channel beds

DURBA SANDSTONE

Too little is known about the sedimentology of the Durba Sandstone to allow any but the most tentative environmental interpretation. Fluvial deposition is consistent with the known facts (Table 19).

DELTAIC DEPOSITION OF THE KAHRBAN SUBGROUP

SOUTHEASTERN DELTAIC FACIES (LOWER)

Southeast of Glenayle, the basal member of the Coonabildie Formation is a chert which presumably formed by inorganic silica precipitation in a restricted, stagnant, saline basin, in a manner similar to that of the Discovery Chert (Chapter 6). Some thin limestone beds near the base of the Coonabildie Formation indicate a marine environment.

The formation shows a general coarsening upward, as sandstone beds become thicker and the proportion of sandstone to mudrock increases. Upward coarsening from siltstone to sandy siltstone or arenite is also exhibited by some individual beds.

TABLE 20. DEPOSITIONAL FEATURES AND INTERPRETATION, SOUTHEASTERN DELTAIC FACIES (LOWER)

Feature	*Interpretation*
Sandstone and siltstone, interlaminated, commonly lenticular or as stringers	Variable current strengths during delta-front sedimentation, mainly as a result of fluctuations in discharge
Siltstone-sandstone beds, coarsening upwards, about 1 m thick	Crevasse-splay deposits
Ripples, symmetrical and asymmetrical RI = 7 to 12	Wave oscillation and current ripples in shallow water
Ripples, symmetrical, wave length 3 cm or less	Very shallow water, only centimetres deep
Pi cross-bedding	Migrating lunate ripples in distributary channels and on delta front
Alpha, omicron, and mu cross-bedding	Deposition by sand waves and ripples in channels and on delta front
Groove marks; current lineations; current crescents, and current shadows	Strong current activity in channels
Channels, up to 30 m wide	Cross-sections of erosive scouring in distributaries or tidal creeks
Occasional conglomerate lenses, pebbles in sandstone	Lag gravel in distributaries
Abundant intraclasts in some sandstone beds	Short distance of transport of mud flakes or chips (e.g. tidal flat, channel)
Flute molds	Eddy scouring by traction or turbidity currents
Limestone, 2-3 cm	Marine environment; low clastic input
Wacke beds	Poor sorting at delta front
Sheet-like geometry of sandstone units	Delta-front sands of high-constructive lobate delta
Mudcracks, Figure 54	Sub-aerial desiccation on inter-distributary or marginal tidal flat
Ice crystals, Figure 55	Occasional temperature drop below freezing
Load casts	Compaction structure

These and other features (Table 20) suggest a prograding-delta facies. Interlaminated siltstone and sandstone and thin stringers and lenses of one lithology in the other, are common in parts of the depositional area, and typical of delta-front sedimentation as displayed in cores through modern deltas (Coleman and Gagliano,

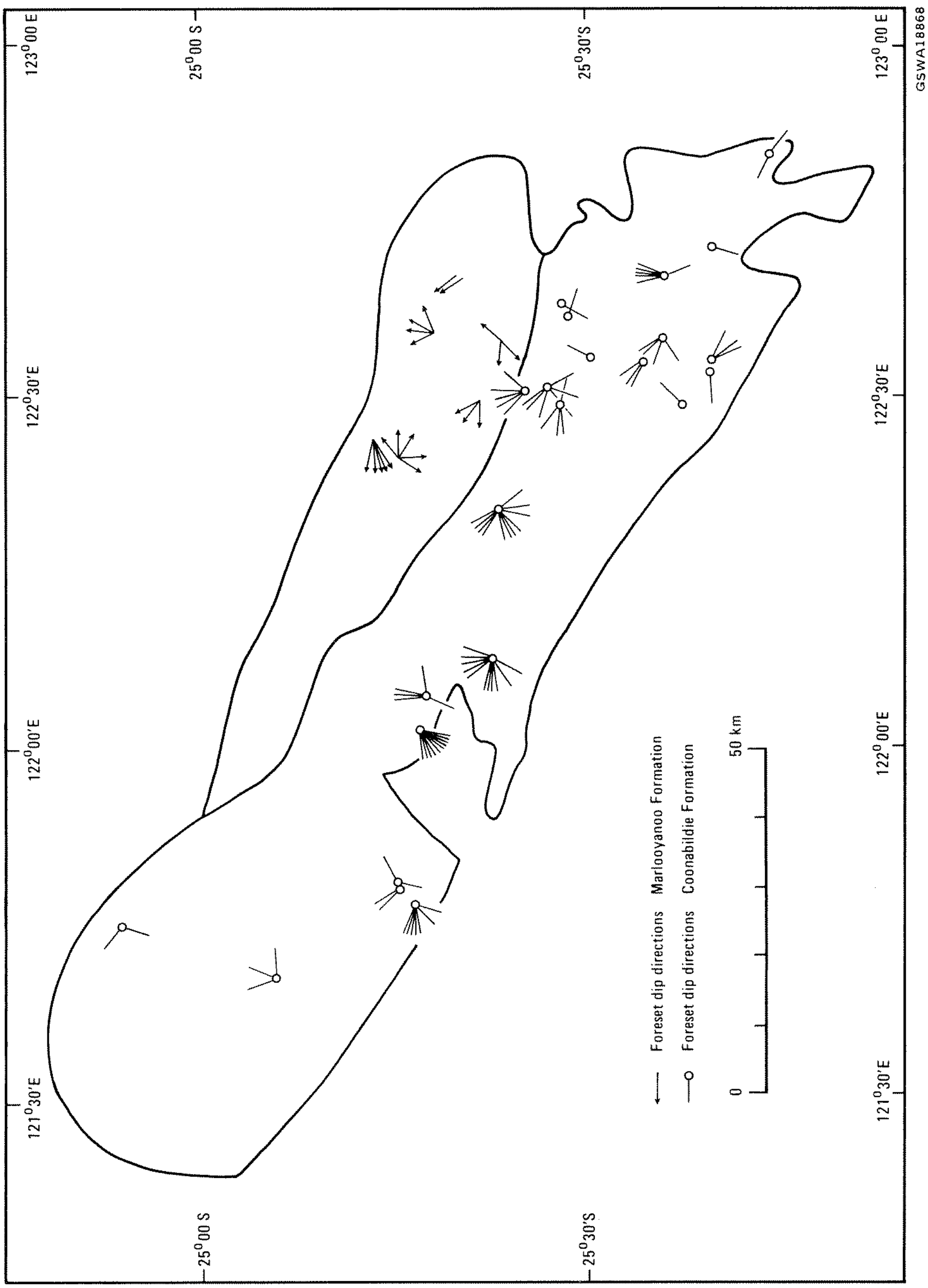

Figure 53. **Palaeocurrent directions in the Kahrban Subgroup according to cross-bedding. Individual foreset measurements are given, and show that at a given outcrop the directions are unimodal except for occasional opposing directions which are attributed to tidal influence.**

1965; Donaldson and others, 1970; Coleman and Wright, 1975). Cross-bedding throughout the region reveals a general trend from east to west (Fig. 53). The palaeocurrents are more consistent than for most marine situations, but more varied than for most fluvial regimes. This was due to diversely trending distributaries, and probable tidal influence. Turbidity currents may have formed off the delta front and flowed into the prodelta zone. Bedding-plane structures interpreted as ice crystal casts (Fig. 55) are the only indication of cold temperatures recorded from the Bangemall Basin. At the top of the formation, 27 km east-southeast of Glenayle are some dark carbonate beds, which point to a brief return to marine, probably lagoonal, conditions as delta-building stopped and compaction settlement occurred.

4cm

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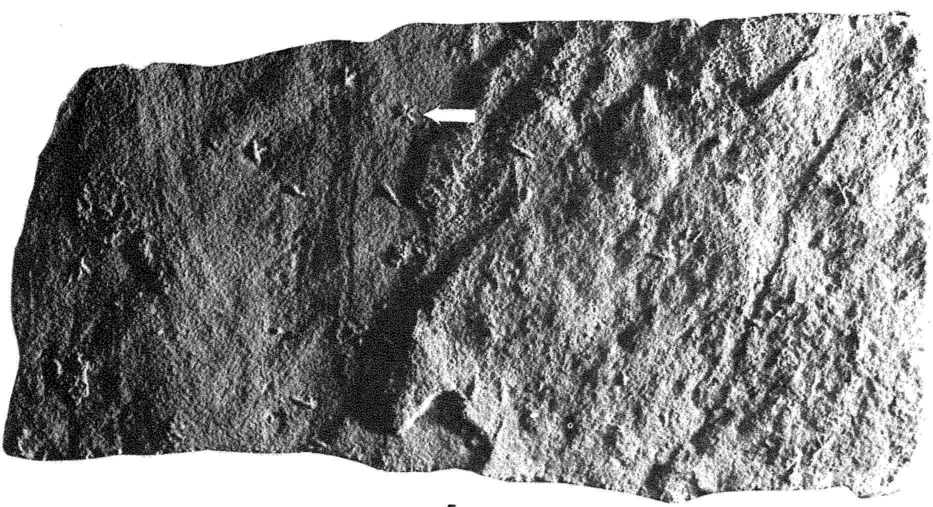
Figure 54. Mudcracks; probably formed by subaerial desiccation, Coonabildie Formation.



5cm

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Figure 55. Star-like structures on a bedding surface (Coonabildie Formation) interpreted as ice-crystal casts. A six pointed star.

SOUTHEASTERN DELTAIC FACIES (UPPER)

The Marlooyanoo Formation also appears to be a deltaic deposit, as evidenced by coarsening-upwards cycles and interlaminated shale, siltstone and sandstone resembling delta-front sedimentation (Table 21).

Flaser, wavy, and lenticular bedding, associated with symmetrical ripples containing unidirectional cross-bedding (Fig. 56A) suggest some tidal activity. The subordinate volume of tidally-influenced sediments, however, and the absence of indentifiable fringing barrier-bar or beach-sand bodies, argues against the sequence representing tide-dominated or wave-dominated destructional phases of the delta cycle. Like the Coonabildie Formation, a constructional delta is indicated, but the two formations differ in that the Marlooyanoo has more mud and silt. The local presence of thick sandstone units which may have been bar-fingers of distributaries, for example just northwest of Carooill Bluff, suggests that the Marlooyanoo may have been deposited as an elongate or "birdfoot" delta. The Coonabildie Formation on the other hand may represent a lobate delta.

TABLE 21. DEPOSITIONAL FEATURES AND INTERPRETATION, SOUTHEASTERN DELTAIC FACIES (UPPER)

Feature	*Interpretation*
Interlaminated shale-siltstone and siltstone-sandstone	Variable current strengths during delta-front sedimentation
Clay-rich sandstone beds	Variable energy conditions during deposition, followed by infiltration of clay into pores between sand grains
Coarsening-upwards cycles from shale into sandstone	Advancing delta lobes and/or crevasse splays
Symmetrical ripples with unidirectional internal cross-bedding, associated with mud flasers (Fig. 56A)	Tidal modification of current-formed ripples, draped by mud during high-water still-stand or periods of low current and wave activity
Flaser bedding, wavy bedding, and lenticular bedding	Alternating current activity and slack water, due to tide reversals, spring-neap tide cycles, and perhaps variable discharge at delta front
Alpha and omicron cross-bedding	Deposition by sand waves in channels and on delta front
Pi and nu cross-bedding (Fig. 56B)	Migrating lunate and linguoid current ripples of various sizes in channels and on delta front
Cut-offs in laminated siltstone and shales (Fig. 57)	Scouring in a normally low-energy environment (e.g. interdistributary bay affected by storms)
Siltstone intraclasts in some sandstone beds	Short distance of transport of semi-consolidated silt fragments from tidal flat
Convolute lamination and distorted cross-bedding (Fig. 58)	Liquifaction of sediment caused by de-watering or differential overloading, or current shear stress on sediment
Interlaminated red siltstones and grey-green shales	Alternating oxidizing and reducing conditions
Scattered granules in sandstone (rare)	Bed load of distributary channel
Feldspathic content (up to 20 per cent) of many sandstones	First cycle detritus, with only short or no residence time in rigorous regimes

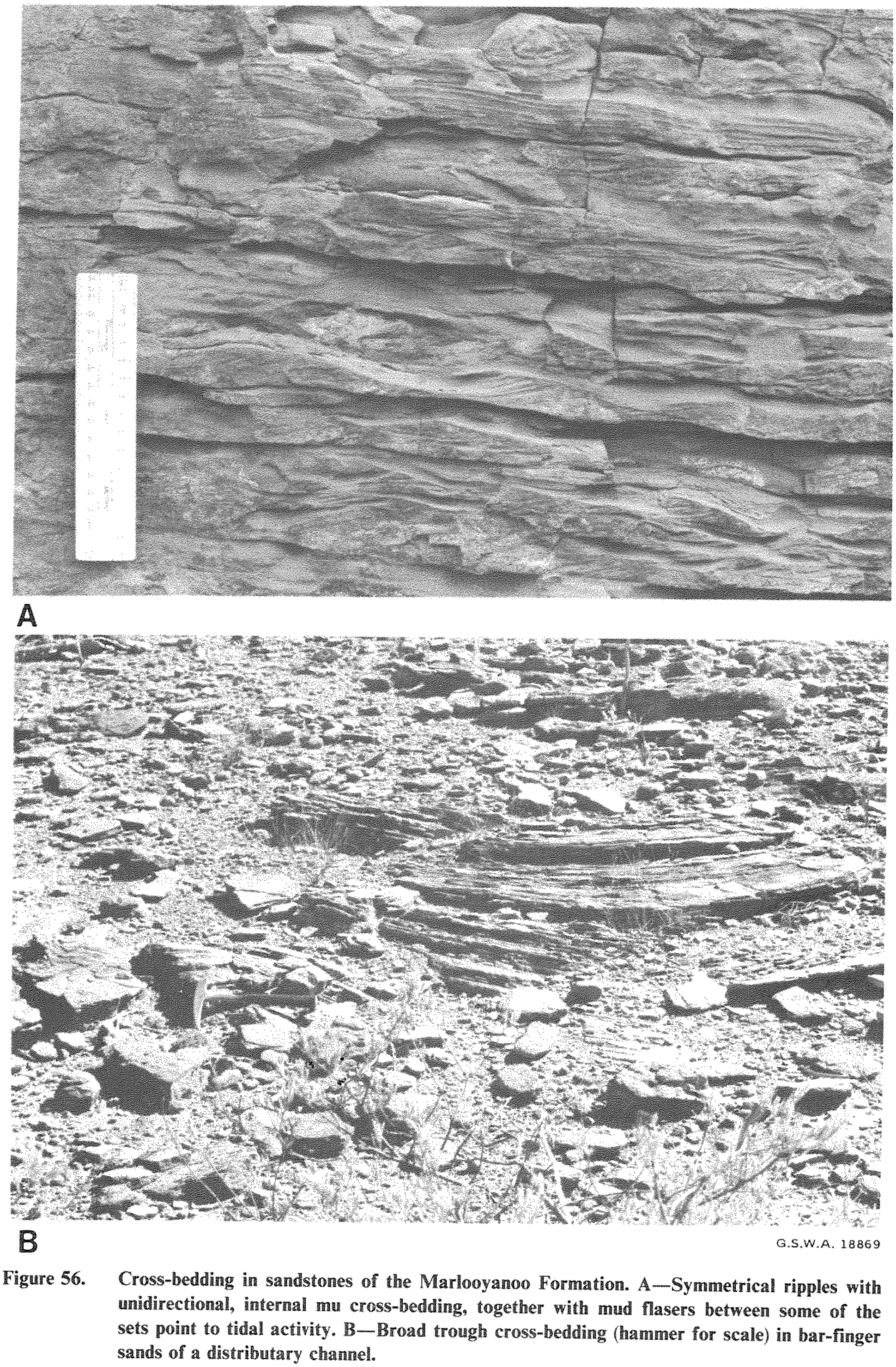

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Figure 56. Cross-bedding in sandstones of the Marlooyanoo Formation. A—Symmetrical ripples with unidirectional, internal mu cross-bedding, together with mud flasers between some of the sets point to tidal activity. B—Broad trough cross-bedding (hammer for scale) in bar-finger sands of a distributary channel.

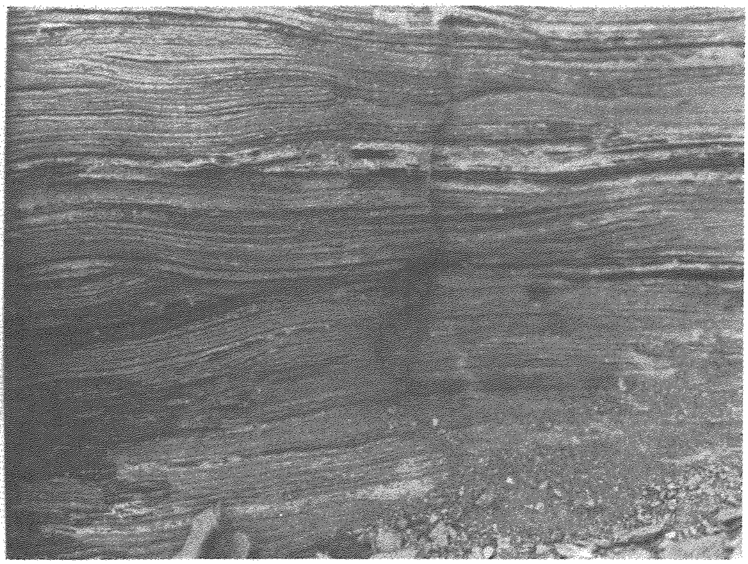

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Figure 57. Scour cut-off in laminated siltstone and shale of the Marlooyanoo Formation.

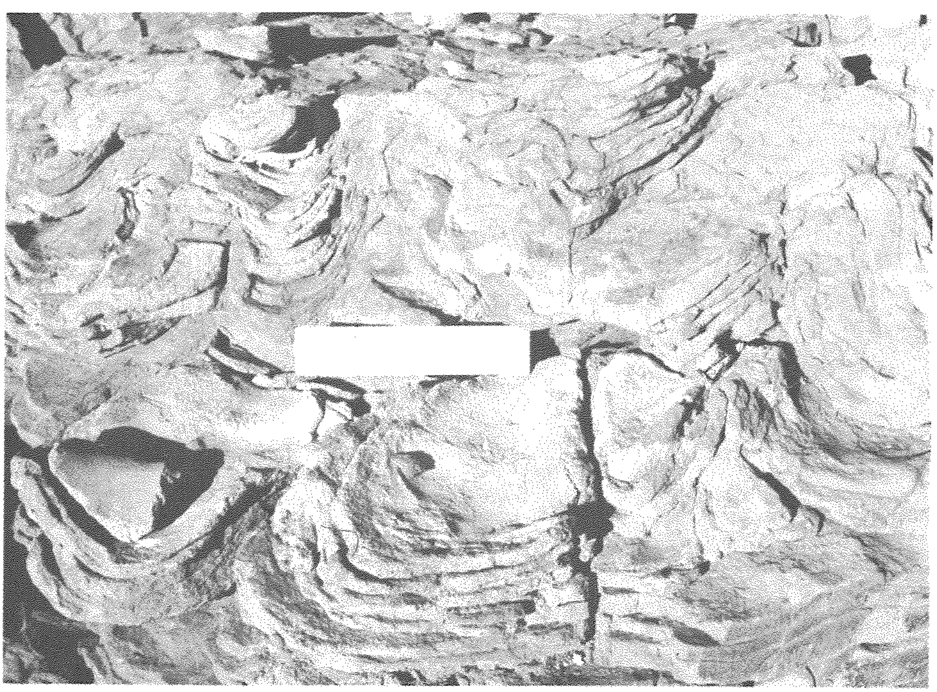

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Figure 58. Distorted bedding in the bottom of a trough cross-bed in feldspathic sandstone, Marlooyanoo Formation. The distortion is due to liquefaction of the sediment, a common process in the distributary channels of a delta.

SYNTHESIS OF KAHRBAN SUBGROUP DEPOSITION

Assuming that the Coonabildie Formation, Marlooyanoo Formation and Calyie Sandstone constitue a conformable sequence, a model for the deposition of the Kahrban Group can be put together as follows.

When the region was first inundated by the sea, transgression was too rapid to leave a record of shoreline deposits. The embayment formed was shallow and had poor circulation, resulting in the deposition of chert. A barrier bar of sand, represented now by the adjacent Calyie Sandstone, was responsible for the restricted circulation by cutting off direct access between the embayment and the open sea. The shallow embayment was later filled by a lobate delta advancing from the east. There was a slow relative rise in sea level during this time, to allow over 1 km thickness of Coonabildie Formation to accumulate, but the amount of terrigenous material brought in was more than able to keep pace with it. The barrier bar at the entrance to the open sea may have persisted, so that the Calyie Sandstone came to lie laterally against the Coonabildie Formation's western extent. When deltaic deposition stopped temporarily, the rising sea level, aided by compaction settlement of the delta pile, allowed the Calyie barrier bar to begin transgressing the Coonabildie, while some minor carbonate collected in the lagoon which formed to the east of the migrating barrier. Delta building resumed, possibly after uplift or climatic change in the eastern hinterland. If the delta complex was large relative to the outcrop area of the Kahrban Subgroup, the resumption of deposition could have been due to delta-switching. The sediments now being supplied contained less sand and more mud, promoting the construction of a "birdfoot" type of delta.

When delta building ceased, the Marlooyanoo delta foundered and its upper beds were reworked by wave action into sandy barrier bars and beaches. These sands became the base of the Calyie in this area. Near Mount Sir Gerard the basal Calyie contains subordinate limestone bands suggesting some transient lagoonal conditions.

CHAPTER 5

The Discovery Chert And Other Cherty Rocks

REGIONAL ASPECTS

The Discovery Chert occurs in the western facies of the Bangemall Basin as a sheet 350 km long by 110 km wide (Plate 1). It occupies about 38 000 km^2; but because the chert has been folded and eroded, its original extent was more than this. To the east and north, the formation loses its identity by grading into siliceous shale and merging with adjoining sequences. Its original limits to the south and west are unknown, but are believed to have been not far beyond the present boundary.

The chert overlies older formations as a blanket, regardless of the pre-existing paleogeography, and thus rests variously on marine-shelf muds, barrier-bar sand and lagoonal dolomite (Fig. 59). The thickness is remarkably uniform over most of

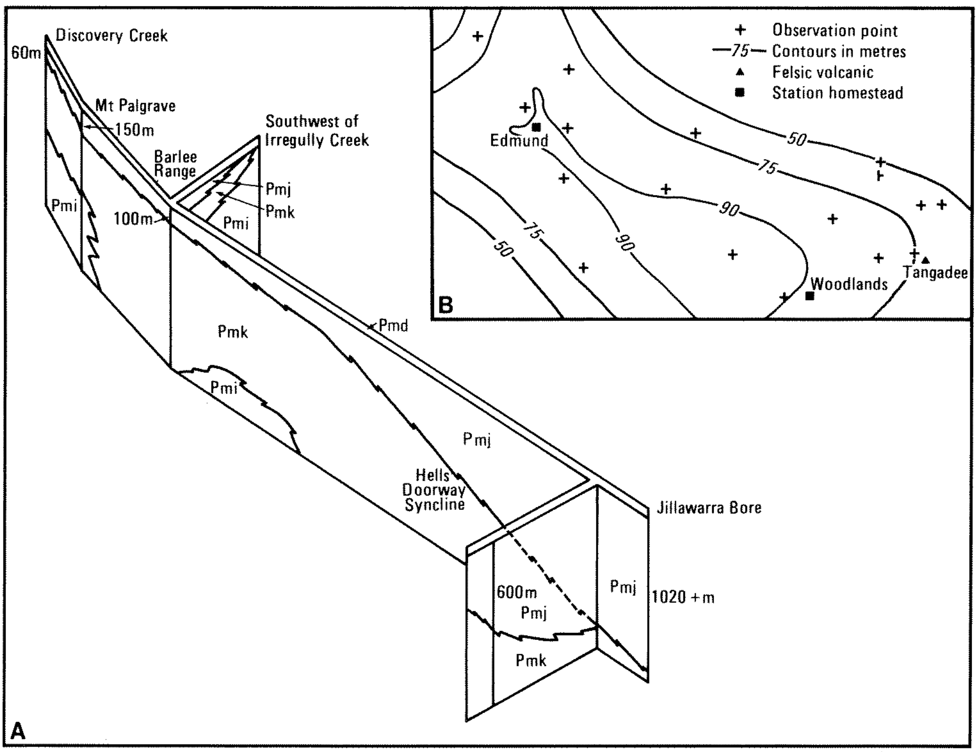

Figure 59. Relation of Discovery Chert to older units and variation in thickness. A—Relation of chert to underlying units, and thickness variation in the Jillawarra Formation. B—Third-order trend surface of thickness of Discovery Chert. The high-valued contours correspond to the greatest thickness of chert and suggest the depocentre is between Edmund and Woodlands.

its extent. Of twelve measured thicknesses, ten lie between 50 and 80 m. The two other measurements, 129 m and 16 m, were obtained from the margins of the outcrop area. Daniels (1966) quotes more varied thicknesses for EDMUND, some up to 580 m, but we believe that such 'anomalous' figures include chert members and intervening shales of the Jillawarra and Kiangi Creek Formations. A third-order trend-surface, which can be regarded as a smoothed isopach map, was calculated from the acceptable measurements and shows an axis of greatest thickness from Edmund to Woodlands (Fig. 59B). Mineral exploration company reports on the Ti Tree and Mangaroon Synclines indicate thicknesses of about 45 metres; these thicknesses also indicate the chert thins to the southwest. The linear depocentre would correspond to the axis of greatest water depth and/or subsidence, for deposition.

The upper and lower contacts of the Discovery Chert are transitional over intervals of about 1 m. Typical basal transitions occur at Jillawarra Bore, Devil

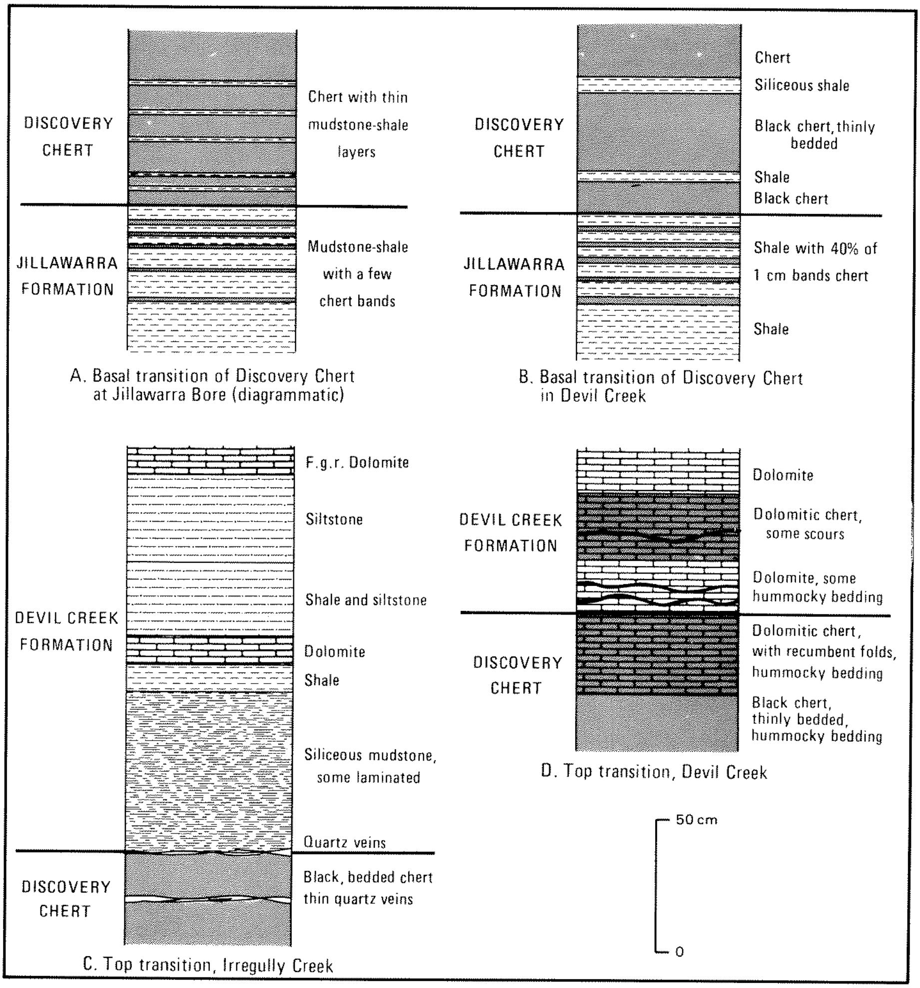

Figure 60. Transitional upper and lower boundaries of the Discovery Chert. A—is diagrammatic, the remainder are to scale.

Creek (Fig. 60A, B) and the Mount Vernon Syncline (Fig. 86), in which shale with a few chert bands gives way abruptly to chert with minor shale. Another type of transition in Brumby Creek, COLLIER, shows a gradation from shale through siliceous shale to fissile chert.

Top contacts of the formation tend to be sharp; bedded chert is commonly overlain by bedded siltstone or fine mudstone of the Devil Creek, Ullawarra, or Backdoor Formations (Fig. 60C), but similarities in surface weathering and bedding thickness of the chert and overlying rocks can make these contacts inconspicuous. For example in a section in the Brumby Creek Anticline, 37 m of whitish chert without fissile interbeds, overlie the more normal chert which contains fissile interbeds. The white chert is a surface silicification product of a soft, white claystone which occurs in its unoxidized state 12 km east along strike. A direct transition from chert to dolomite exists in the type section of Discovery Chert in Devil Creek, where normal black chert passes into carbonate-bearing chert and then into fine-grained cherty dolomite.

STRUCTURES AND TEXTURES

PETROGRAPHY

The Discovery Chert is composed of microcrystalline quartz, opaque carbonaceous material, and subordinate kaolin, illite/sericite, clastic quartz, and pyrite. The secondary minerals, hematite, jarosite and alunite are also present, as are trace amounts of carbonate, gypsum and barite. Lamination is preserved; but most microtextures in the quartz have been destroyed by recrystallization, and carbonaceous material has been redistributed along quartz-grain boundaries.

The weathering profile within rocks of the Bangemall Group has been extensively dissected, and therefore surface samples are from various points within the profile, as well as from relatively fresh rocks. Leaching and weathering effects in surface samples include hematite pseudomorphs after pyrite, and quartz- and clay-filled crystal molds after pyrite and gypsum. Kaolin, hematite, jarosite, and alunite, which are in the chert, are typical of weathered black shales of the Jillawarra Formation (Davy, 1980). The presence of similar minerals in surface samples of Discovery Chert indicates weathering similar to that in carbonaceous shales. All samples are siliceous and have a 'cherty' aspect at the surface, and it is likely that pyrite and carbon have been diminished or removed in many surface samples.

PETROGRAPHIC TYPES

Three petrographic types of chert have been distinguished and there are indications that these reflect different degrees of oxidation of a single parent rock type.

Black carbonaceous chert is probably the least altered type. It typically has laminae 2-4 mm wide, alternately rich and poor in black carbonaceous matter and all set in a matrix of microcrystalline quartz with minor clastic quartz and sericite.

Brown chert is uncommon. It is black in hand specimen, but consists of brown microcrystalline silica with an average grain diameter of 10 μm. The brown colour is probably organic material, as it is resistant to hydrofluoric acid and does not produce an effect on the electron microprobe, as would be expected if the colouration was iron oxide. Also, there is free carbon present (Appendix B). The brown cherts which come from the same localities as the black variety show the same microstructures and contain numerous secondary minerals. Therefore, it is likely that the brown carbonaceous chert is a weathered form of the black chert.

Pure siliceous chert is common in outcrop. It contains microcrystalline quartz and traces of clay and secondary hematite that outline faint laminations which display structures similar to those in carbonaceous chert (Fig. 73). These structures, and the secondary minerals indicate that the pure siliceous chert is a leached carbonaceous chert. Further evidence of leaching is visible in outcrops and in drill core; the evidence shows that shale and chert which are black at depth, crop out as a creamy, white chert, deficient in carbon (Davy, 1980).

MINERALOGY

QUARTZ

Most of the chert consists of microcrystalline quartz grains which have an average diameter of about 5 μm, but there are patches of coarser grains (to 30 μm). The grains are irregular in three dimensions and have no preferred orientation or elongation.

CARBONACEOUS MATERIAL

The carbonaceous material forms wispy aggregates ranging from 0.01 to 0.2 mm thick. The finest aggregates are composed of a three dimensional network of black rods, about 1-2 μm long, interstitial to microcrystalline quartz crystals.

CLASTIC PARTICLES

Clastic quartz and sericite grains are a subordinate (<1 per cent), but widespread, component of the chert. The quartz clasts (average diameter 0.04 mm) are subangular to subrounded and occur as scattered, isolated grains. Ragged sericite flakes lie parallel to the lamination. Clastic grains are concentrated towards the base and top of the Discovery Chert and decrease towards the centre. This variation reflects a gradual cessation of clastic sedimentation allowing chemical deposition followed by a resumption of clastic deposition.

Wholly clastic chert is rare (Fig. 68), but is present at the base of the Discovery Chert in the Brumby Creek Anticline.

CARBONATE

Carbonate ranges up to 10 per cent of the chert, but is uncommon. It occurs as rhombs or veins. Rhombs about 1.3 mm on edge may be scattered throughout the rock (Fig. 61A), may form crystal aggregates parallel to lamination (Fig. 61B), or may occur in stylolites. The rhombs are now filled with yellow-brown iron oxide, fibrous kaolin, or microcrystalline quartz. Rhombs have grown across pellets in

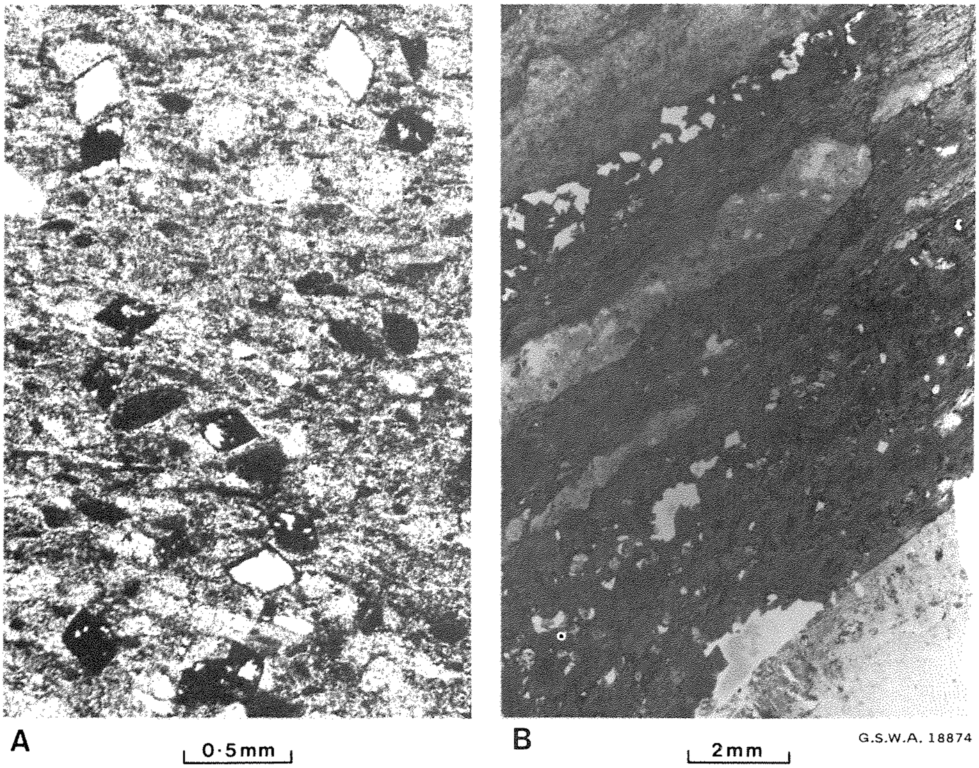

Figure 61. Photomicrographs of carbonate in Discovery Chert. A—Iron oxide and clay pseudomorphs after carbonate in clastic textured chert (specimen 34501 B), Discovery Chert, Brumby Creek Anticline, COLLIER. B—Crystal molds, after carbonate (specimen 50467), Coobarra Creek, COLLIER.

clastic laminated chert and across intrafolial folds, but appear to have formed prior to recrystallization of silica into the microcrystalline form. Therefore the rhomb carbonate formed late in the diagenesis of the chert. A few specimens contain wavy veins of calcite which cut across all features in the chert, and here, calcite was the last mineral to form.

GYPSUM

Patchy concentrations of crystal molds up to 5 mm long cut across laminae in the Discovery Chert. Some are now filled with quartz and kaolinite. Many molds have monoclinic shapes and bear a striking resemblance to the lensoid crystals of gypsum described by Kinsman and others (1976).

Fresh gypsum is rare and has been recognized only in the Coobarra section on COLLIER where it forms irregular concentrations of euhedral crystals. These crystals formed before crystallization of the microcrystalline quartz.

PYRITE

Pyrite is the common sulphide mineral in the Discovery Chert. It crystallized after formation of laminations and after compaction. Most of the pyrite was in cubic crystals ranging from 4 μm to about 5 mm on edge. Almost all pyrite has been replaced by hematite, goethite, or less commonly, microcrystalline quartz or kaolin.

Pyrrhotite is present in chert samples from the Barlee Range.

Hematite is a widespread though minor secondary mineral, and most is oxidized pyrite. It occurs either as scattered grains or aggregates. The scattered grains are either anhedral or, commonly, cubes averaging about 4 μm on edge. The aggregates are present as discontinuous laminae up to 0.1 mm wide, ovoid concentrations (average 0.9 x 0.4 mm) or framboidal clusters.

MINOR SECONDARY MINERALS

Jarosite is a minor mineral in carbonaceous chert. It occurs in lenticular layers composed of randomly oriented, fine-grained, tabular, orange crystals, as well as veins which cut laminations. Jarosite is a typical mineral in oxidized black shale (Davy, 1980) and seems likely to have resulted from the interaction of clay minerals with acid produced by oxidation of pyrite.

Alunite forms isolated tabular crystals (about 8 x 4 μm) scattered throughout a few samples of chert. It probably formed in a similar manner to the jarosite.

EVIDENCE FOR A GEL

SOFT-SEDIMENT STRUCTURES

The Discovery Chert contains numerous minor depositional and deformation structures, including slump folds, load casts, concentric wrinkles of bedding and syneresis cracks. In the slump folds, planar undisturbed bedding caps the folds (Fig.

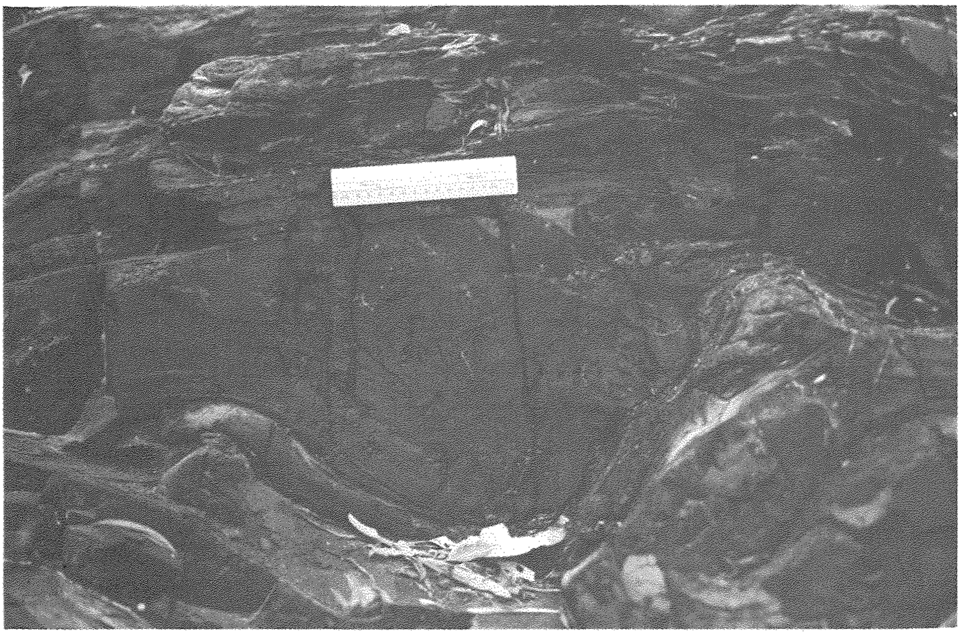

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Figure 62. Slump fold, capped by undisturbed bedding, Devil Creek, EDMUND. Scale is 15 cm long.

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Figure 63. Small-scale load casts, illustrating density differences in the Discovery Chert when still plastic, Brumby Creek Anticline. Scale in millimetres.

62) and indicates relevelling of a gel, sol, or hydroplastic material at the sediment-water interface. Load casts (Fig. 63) are evidence of the differential mobility of plastic material. Concentric wrinkles of laminae show no cracking in their crests and form an annular pattern around low mounds in the bedding (Fig. 64). These mounds are cut by irregular cracks, which, in places, are filled with microcrystalline silica and which could be syneresis (shrinkage) cracks (Pettijohn, 1975), such as occur in gel-like material. These structures demonstrate the Discovery Chert was plastic and gel-like in its early history.

CHERT SPHERES

Closely packed chert spheres which are rimmed with carbonaceous material, are common in carbonaceous chert (Fig. 65). There are two sizes of spheres. The smaller are 5 μm in diameter and are a pale, cloudy brown. Their shapes vary from

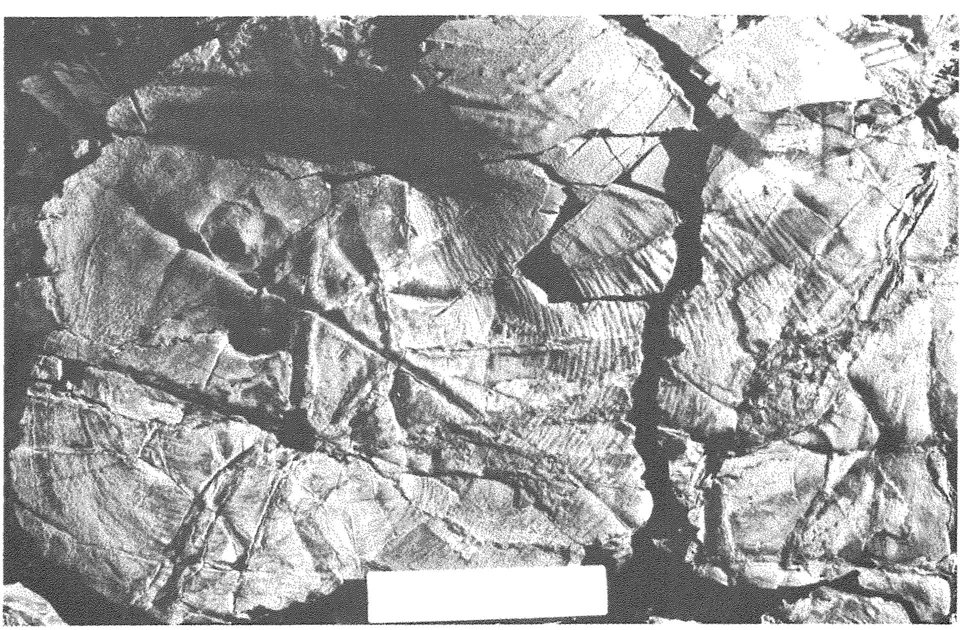

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Figure 64. Concentric wrinkles of laminations, which contain no cracking in their crests, indicate that the chert was in a plastic condition, Mount Vernon Syncline, MOUNT EGERTON.

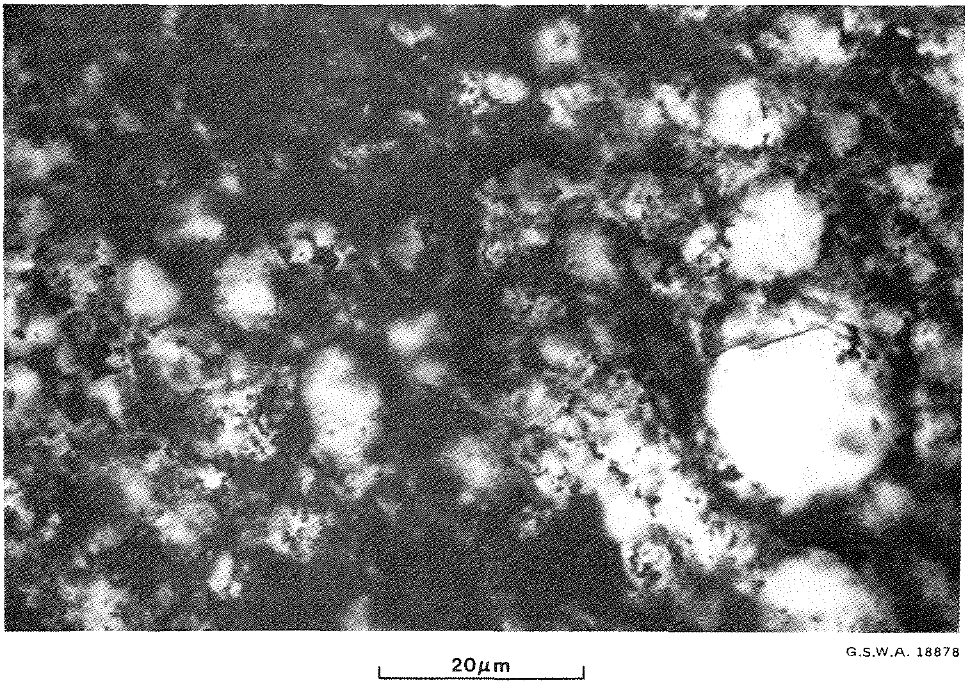

Figure 65. **Discovery Chert, showing large and small silica spheres. The lamination trends from upper left to lower right and the long axes of the oval bodies are at an angle to the lamination. Also note that the boundaries between some spheres are flat. Both points suggest that the final shape of the sphere is determined by crystallization forces rather than sedimentation or compaction (Specimen F11127).**

spherical to ovoid. The long axes of the ovoids may be at any angle to the carbonaceous laminations (Fig. 65) though many are parallel to them. The larger are consistently circular in cross section, average 20 μm in diameter, and consist of clear quartz.

The 20 μm spheres have features listed below which suggest they formed by *in situ* crystallization rather than by accumulation of sedimented particles:

(1) They are present in both carbon rich and carbon poor layers, and are not controlled by the edges of laminae.

(2) A planar 'flattened' boundary between adjacent spheres indicates mutual interference during growth, especially where the boundary is perpendicular to the laminations.

(3) The spheres have not been affected by compaction. The Discovery Chert has undergone considerable compaction, as shown by the carbonaceous laminae, which wrap around clastic quartz grains. The carbonaceous laminae do not wrap around the spheres, which have therefore formed *in situ* after compaction.

The spheres now consist of microcrystalline quartz. The quartz within some chert spheres has a fibrous radial pattern of extinction, which does not cross sphere boundaries, and which suggests that the spheres initially crystallized as chalcedony spherulites. The small silica spheres have optical properties which are similar to those of the large spheres and probably also crystallized as chalcedony.

The principles of spherulitic crystallization, summarized by Oehler (1976), show that in natural aqueous systems at low temperatures, chalcedony spherulites form only from highly viscous siliceous solutions. To attain sufficient viscosity, the solutions would have to be colloidal. Therefore it seems likely the spheres in Discovery Chert crystallized from a silica gel. This idea is supported by the similarity of the shapes and sizes of the spheres in the Discovery Chert to the spheres crystallized from silica gel in experiments by Oehler (1976).

DIAGENESIS

Modification of the silica gel has been by de-watering, compaction, and finally, recrystallization. Comparison of the thickness of silica laminae between carbonaceous shreds, with the thickness of laminae adjacent to clastic quartz grains indicates a compaction of at least 60 per cent. Another calculation of compaction can be made from the ration of arc length to chord length of convoluted quartz-clay veins that are perpendicular to bedding If it is assumed that the veins were originally straight, a minimum compaction of 40 per cent has occurred.

BEDDING AND LAMINATION

Layering on a scale of millimetres and decimetres is a feature of the Discovery Chert. The thicker layering (bedding) is mostly less than 50 cm thick and is expressed either by different colours or by alternate fissile and non-fissile layers. Colour banding is due to subtle differences in quantities of carbonaceous material, clays and iron oxides, which are accentuated by weathering. Fissile cherts contain planes of parting less than 1 mm apart and generally form beds of poor outcrop

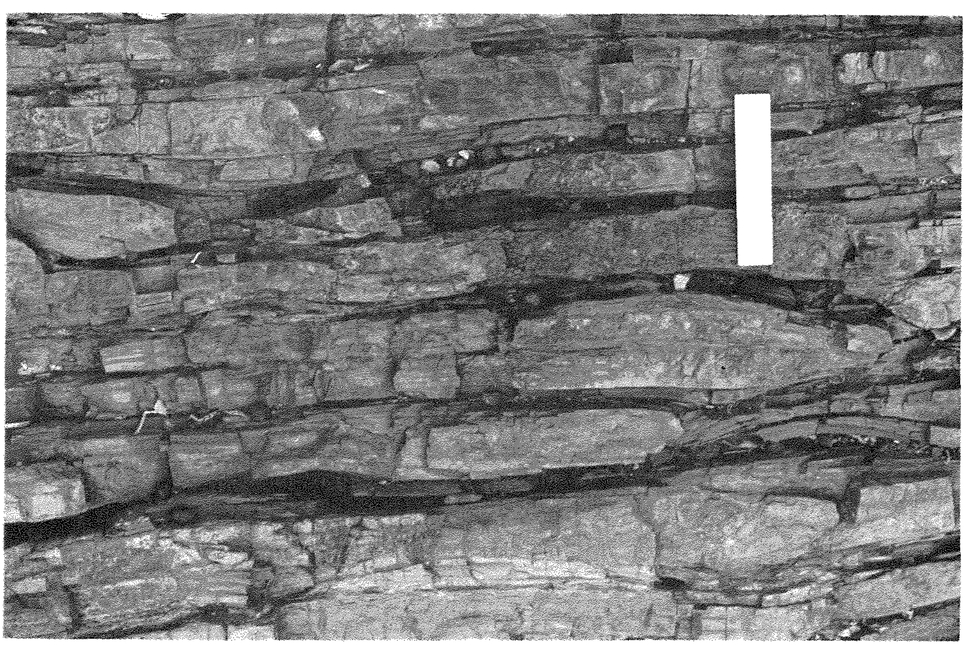

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Figure 66. Beds of fissile and non-fissile chert, Discovery Chert, Brumby Creek Anticline, COLLIER. Scale 15 cm long.

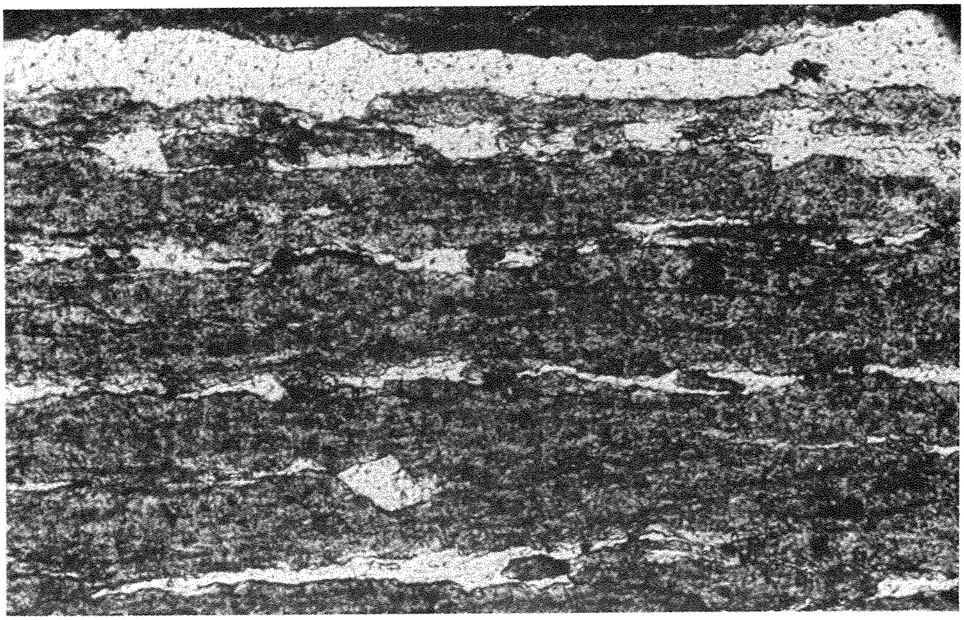

Figure 67. **Photomicrograph of fissile chert with streaks of opaline silica. Section perpendicular to bedding. Specimen 34501 F, Brumby Creek Anticline, COLLIER.**

between more resistant non-fissile chert (Fig. 66). The fissile chert shows a very fine, discontinuous, (diffuse) lamination in thin section (Fig. 67) which contrasts with the more open, widely spaced and well-defined laminations of the non-fissile chert (Fig. 71). The planes of fissility are marked in thin section by discontinuous streaks of opaline silica, 0.1 mm thick, parallel to bedding. The origin of the opaline silica is not known. It may have replaced a more soluble mineral such as gypsum.

LAMINATION

Almost all types of chert display lamination ranging from 0.01 mm to 1 cm in width. Two types of laminae are recognized, clastic and carbonaceous.

Clastic laminae are formed by layers of different sized clasts. They are uncommon, but are present at the base of the Discovery Chert in the Brumby Creek and Mount Vernon areas. Clasts of laminated and massive pure chert and carbonaceous chert are present in addition to clastic sericite and quartz and quartz-sericite-tourmaline rock. The chert clasts range from rounded equant grains to those which were plastic during compaction (Fig. 68). Many clasts contain laminae of organic material, and this, together with the similarity of their grain size of silica to most of the Discovery Chert, indicates they were derived by penecontemporaneous erosion of chert.

Carbonaceous lamination in the most general terms comprises dark bands rich in carbonaceous material alternating with lighter coloured siliceous bands. Both bands contain microcrystalline quartz and also may include kaolin, illite and hematite. Carbonaceous lamination is continuous and roughtly parallel at hand-

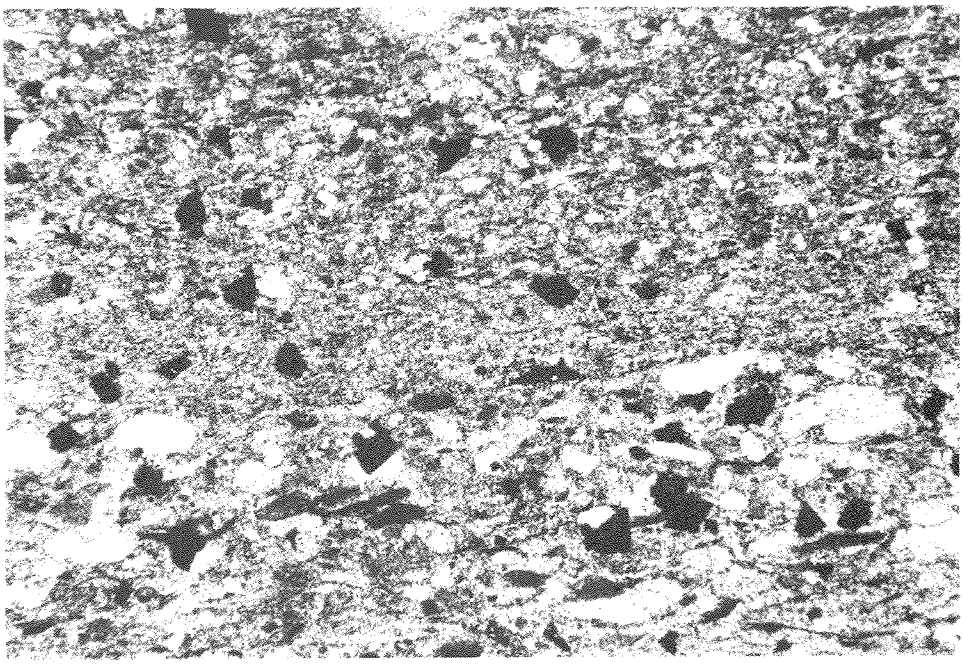

Figure 68. Photomicrograph of clastic laminated chert, Brumby Creek Anticline, COLLIER. The pale, elongate casts are of chert; the black crystalline shapes are goethite pseudomorphs after carbonate. Specimen 34501 B.

specimen scale, and ranges in width from 2 to 4 mm. Within the mesoscopic carbonaceous laminae are two different scales of finer layers, rich and poor in carbonaceous material. These have been termed microlamination and fine microlamination and their relation to each other and to the mesoscopic lamination is set out in Table 22.

TABLE 22. RELATION BETWEEN LAMINATION, MICROLAMINATION AND FINE MICROLAMINATION

Name	*Thickness (mm)*	*Features*
Mesoscopic lamination	< 2.0 - 4.0	Continuous and roughly parallel; visible in hand specimen
Microlamination	0.1 - 0.2	Discontinuous; visible with hand lens
Fine microlamination	0.01 - 0.02	Discontinuous; visible microscopically

These subdivisions form a hierarchy. Fine microlamination is formed by the finest microscopically visible layers, alternately rich and poor in carbon. Groups of fine microlaminae together form coarser layers—the microlaminations which in turn form groups that define the mesoscopically visible lamination. Carbonaceous lamination and fine microlamination show little variation, but the microlamination shows great variety in texture.

The discontinuous nature of both the dark and light microlaminae imparts a distinctive streaky, lenticular fabric to the rock (Fig. 72). The individual lenticles are about 1.0 mm long, parallel to the microlamination, 0.2 mm wide, and consist of lenses of quartz, sericite and clay which are enclosed by shreds of carbonaceous material. This fabric is not so well developed in the more siliceous cherts, where there are only aggregates of clay or chlorite and no carbonaceous material occurs with quartz. On lamination surfaces, the lenticles are amoeba-shaped with no lineation visible. Although most are lenticular in cross-section a few are approximately equant. Streaky texture is the earliest recognizable texture in the Discovery Chert and has been modified or affected by crystallization of quartz, soft sediment folding and stylolites. The microlaminae and streaky fabric can be subdivided into three morphological styles.

(1) Lenticular, carbonaceous microlamination is a common style. Lenses (0.1 mm wide) of quartz, sericite, clay, and subordinate carbonaceous matter are enclosed in carbonaceous wisps (Figs. 69, 70). Such lenses occur in lamination-rich or lamination-poor zones in carbonaceous material. This microlamination is present in chert from the Barlee Range (EDMUND), and Marshall (1968) has noted a similar style in the Mount Palgrave area. However, the best examples of this style of microlamination are in black chert from the Fords Creek Shale.

Significant features of the texture are the crinkling of the lenticular microlamination, the presence of gypsum crystal molds, and possible relic fenestral texture. The crinkles display the following features:

(a) A cyclic sequence of development. The degree of crinkling is more intense at the base of a lamination (2-3 mm wide) and decrease gradually in intensity upwards to flatter microlamination (Fig. 69).

(b) Crinkled microlaminae have been cut off by overlying laminae.

(c) Laminae contain fragments of finely wrinkled microlaminae derived by erosion of the immediately underlying lamination.

All these features suggest the crinkles are a primary structure formed during deposition of the chert.

The crystal molds perpendicular to the microlaminae are now filled with microcrystalline quartz. Their shape is suggestive of lenticular gypsum crystals.

Lenticular patches of clear, microcrystalline quartz (Fig. 70) are interpreted as infillings of original voids that were orientated parallel to the microlamination (fenestral texture). These patches contrast with most siliceous lenses in the chert as they contain very little sericite or clay.

It is not known whether this carbonaceous microlamination resulted from accumulation of pelagic organisms or from growth *in situ* as an algal mat. The gypsum and fenestral textures are common but not diagnostic features of algal mats. The primary crinkly texture is probably the result of desiccation during exposure, but there are no other features, such as mudcracks to indicate exposure. The massive to poorly layered bands (Fig. 69) resemble densely calcified zones in recent algae (e.g. Monty, 1976, Fig. 7), but these zones do not seem to be directly related to the form of the algal mat and so may be a response to particular chemical conditions.

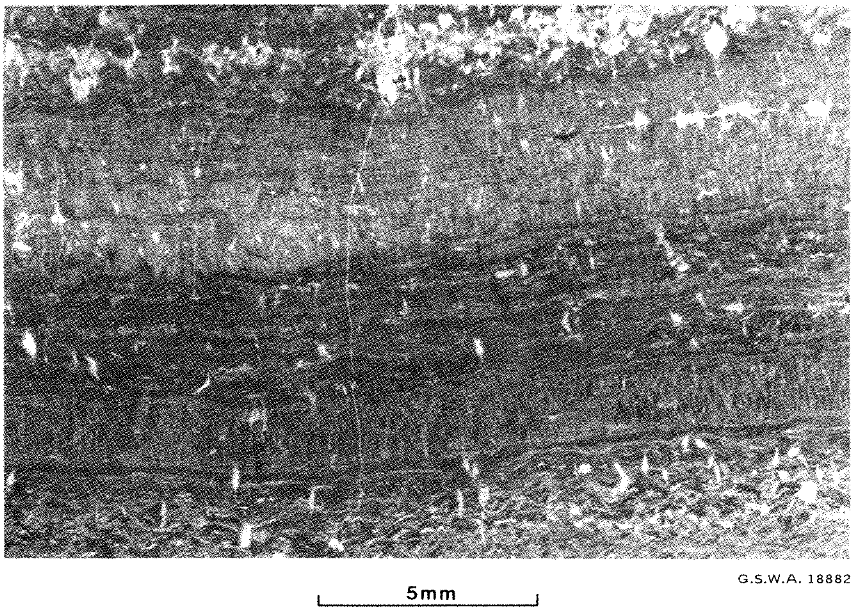

Figure 69. **Photomicrograph showing laminations 1-6 mm wide of light and dark layers. Microlaminae shown as fine, light and dark layers less than 1 mm wide, are intensely crinkled at the base of each lamination, and become flatter towards the top. Note also the poorly microlaminated bands which resemble calcified zones in modern stromatolites. Chert from Fords Creek Shale 1 km northeast of Coodardoo Gap, EDMUND.**

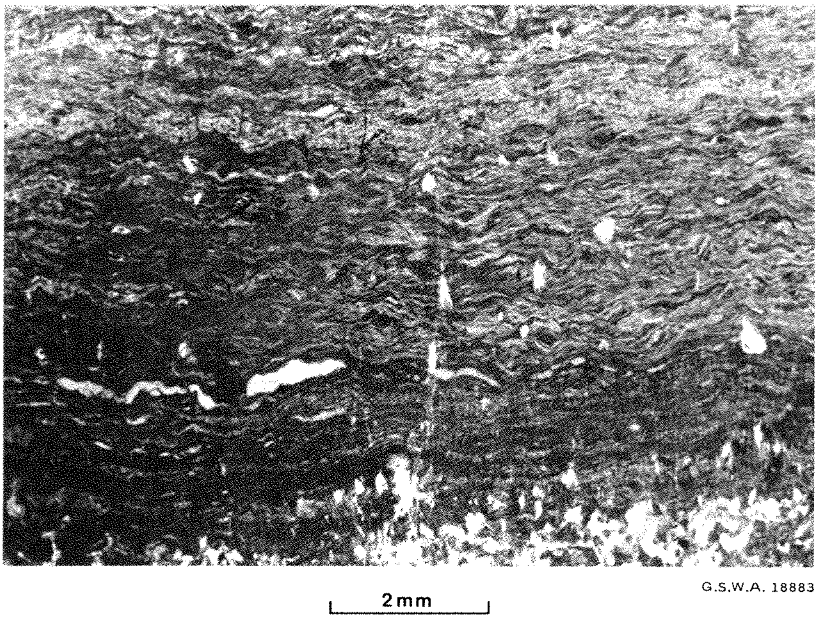

Figure 70. **Lenticular carbonaceous microlaminations in chert of Fords Creek Shale, 1 km northeast from Coodardoo Gap, EDMUND. The microlaminae are light and dark strands less than 0.5 mm wide. The clear patches parallel to the microlaminae may be fenestral texture. Specimen 50444.**

(2) Parallel carbonaceous microlamination is common. Laminations which are continuous over distances of centimetres contain parallel microlamination. Even finer internal microlamination is usually well developed within the broader microlaminae. This fine microlamination is outlined by concentrations of blebs of carbonaceous material about 5 μm in diameter. The fine microlaminae meet both the top and base of the larger scale bands in low-angle discontinuities (Fig. 71).

(3) Braided carbonaceous microlamination is the most common style. In hand specimen, broad, parallel laminae up to 5 mm wide contain a fine streakiness parallel to the laminae (Fig. 72). The fine streakiness comprises sets of parallel microlaminations about 0.3 mm long which truncate each other at low angles (Fig. 73B) resulting in a discontinuous wavy texture (Fig. 73A). Fine microlaminae are present within the microlaminae. Therefore, although the texture is more irregular than in the parallel-laminated chert, the same scales of lamination are present. This suggests that the various scales of lamination in the braided style had the same origin as their counterparts in the parallel-laminated chert. Therefore, because the scales of lamination were established prior to the truncations, the texture is not the result of large-scale de-watering or compaction.

Instead, tiny folds adjacent to some truncations (Fig. 73B). suggest the braided microlamination is a deformation texture produced by disruption, at a low angle, to the lamination of parallel laminated chert. The time of disruption is, therefore, after most of the compaction which produced lamination, but while the silica was still plastic and in a form which would not show evidence of

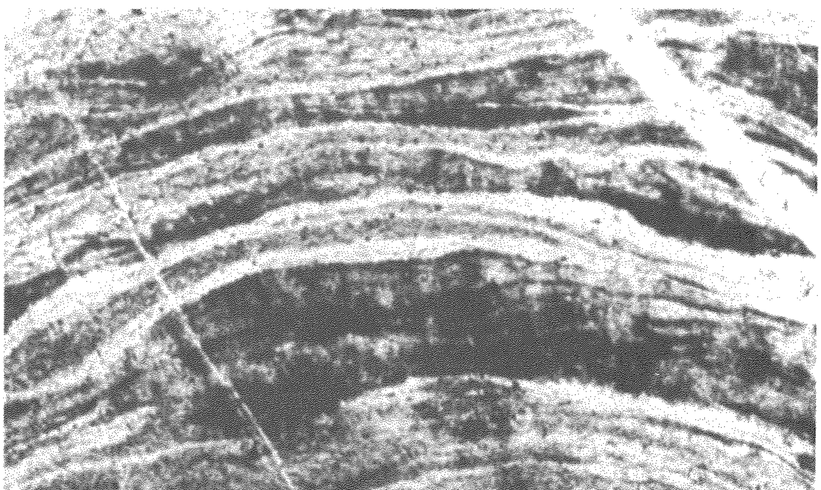

Figure 71. **Parallel microlamination about 0.5 mm wide containing fine microlamination about 0.1 mm wide, Discovery Chert, Coobarra Creek, COLLIER. Specimen 34504 C.**

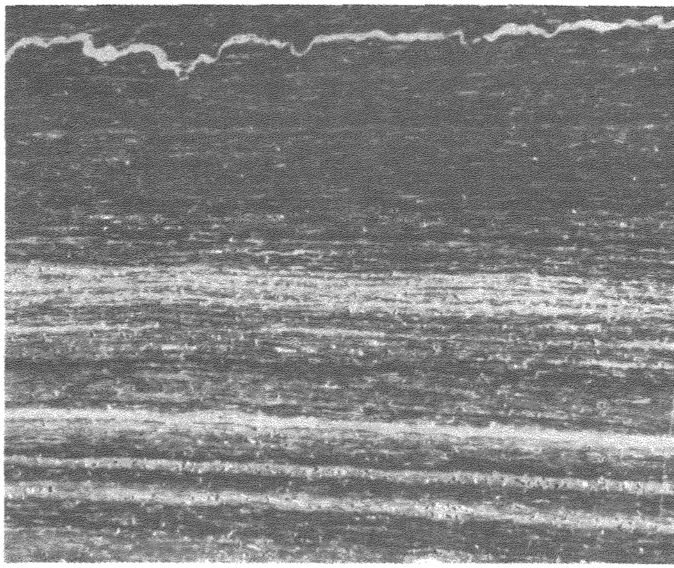

Figure 72. Braided microlamination shows in hand specimen as a fine streakiness about 0.1 mm wide, within the lamination, which is 1 to 5 mm wide. Brumby Creek Anticline, COLLIER. Specimen 34501 M.

strain. Deformation at a low angle to lamination then continued in the chert, folding the disrupted lamination into isoclinal, intrafolial folds associated with low-angle faults.

FOLDS

Zones of minor, intrafolial, recumbent folds and associated, small, low-angle faults are common in the Discovery Chert (Fig. 74) and also occur in black shale of the Curran Formation. The zones are 5-20 cm thick, parallel to bedding, and continuous over distances up to 15 metres. The folds range from open to isoclinal, and have deformed both lamination and streaky texture. Many zones are complexes of recumbent folds, breccia, and small nappes, separated from each other by thrusts (Fig. 75). Fold-axis surfaces are plane or curved and lie approximately parallel to regional bedding. The azimuth of plunge of the fold hinges, in a structurally homogeneous area (e.g. one limb of a major fold), is spread over a wide range. Some fold-hinge lines are gently curved, and the convex outlines of adjacent fold crests commonly point in opposite directions.

Low-angle faults, which truncate intrafolial folds, have been traced in outcrop for 5 m. In a few zones, slices of folds may be matched with their complementary portion on the opposite side of a fault.

FOLD MICROTEXTURE

In thin section, folds are outlined by laminations, streaky texture, and clastic sericite grains orientated parallel to the laminations. Microcrystalline quartz grains show no orientation, either parallel to fold-axis surfaces or to laminations. However,

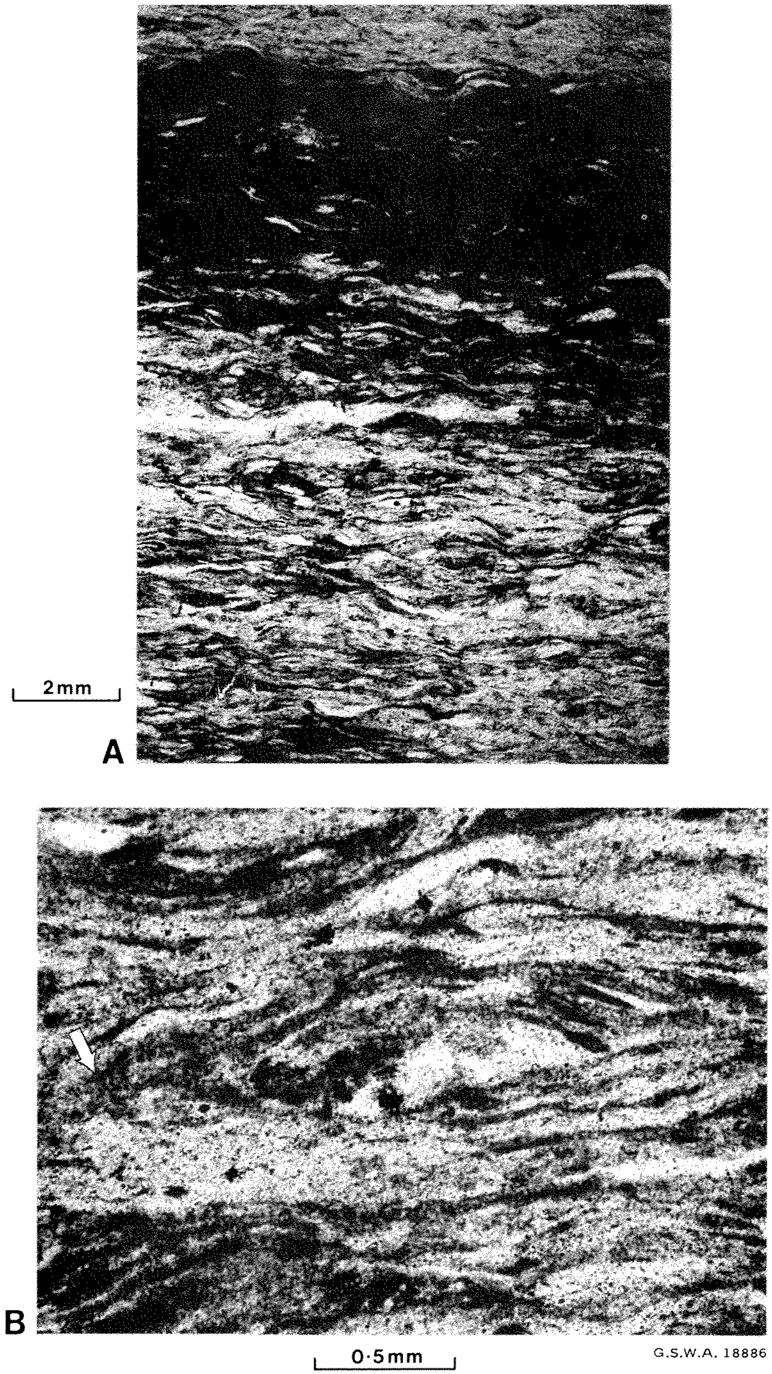

Figure 73. **Braided microlamination. A—Light laminations about 6 mm wide contain braided microlamination: narrow, light and dark, discontinuous, curved streaks about 0-5 mm wide. B—Braided microlamination; each microlamination about 0.3 mm wide is composed of parallel light and dark microlaminae 0-0.05 mm wide. Note sets of fine microlaminae are truncated, also fold-nose (arrowed) suggesting truncation is a zone of movement.**

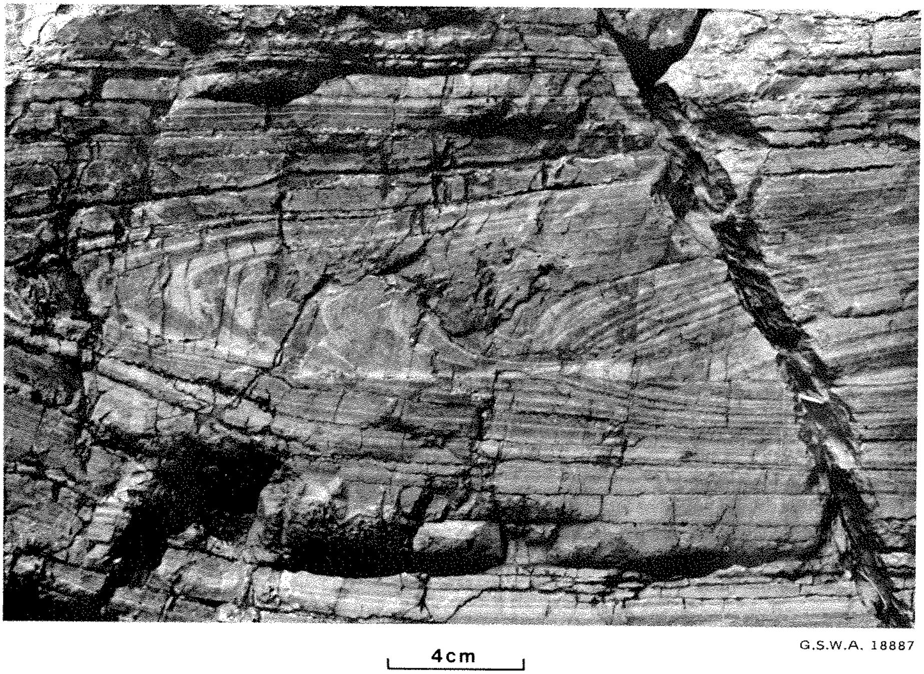

Figure 74. Intrafolial fold and thrust, Discovery Chert, Brumby Creek Anticline, COLLIER.

axial-plane cleavage is developed in the tight and isoclinal folds and is shown by very fine sericite and by edges of quartz-sericite lenses trending parallel to the fold axial plane. The laminations and quartz-sericite-clay lenses define asymmetric, parasitic folds in the crest (Fig. 76). The quartz-sericite-clay shapes behave as strain markers, and in a fold profile are thin, elongate lenses in fold limbs and are thicker in fold crests (Figs. 77, 78).

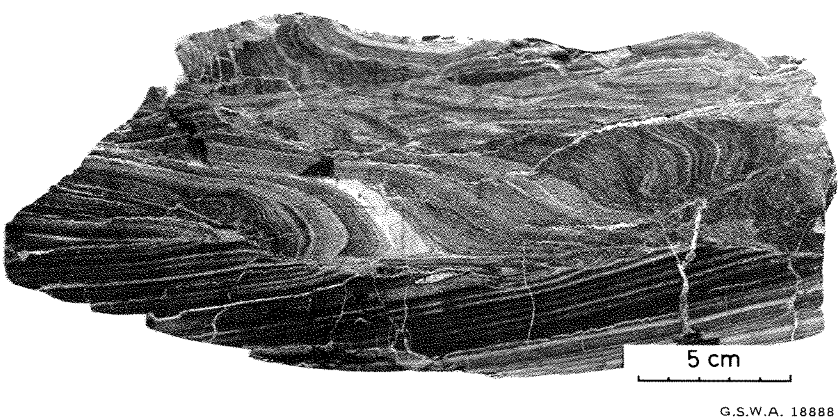

Figure 75. Zone of recumbent intrafolial folds, nappes, and thrust, Discovery Chert, Brumby Creek Anticline, COLLIER. Specimen 34501 M.

1cm

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Figure 76. **Basal contact of intrafolial fold zone, showing small nappes slicing into and slightly compressing the underlying chert. Specimen 34501 M.**

On a lamination surface the lenses have irregular, platy shapes in the fold limbs and show a slight elongation parallel to the fold axis. Although the original shape of the lenses is not known, the parasitic folds indicate the intrafolial folds were probably initiated as buckles which were then modified by flattening.

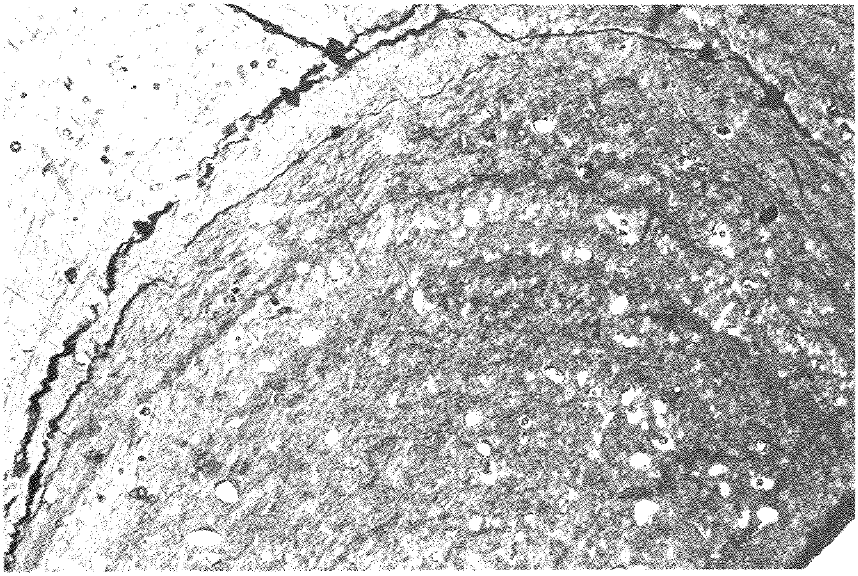

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Figure 77. **Photomicrograph of fold core with parasitic folds in limb and crest. Specimen 34501 K.**

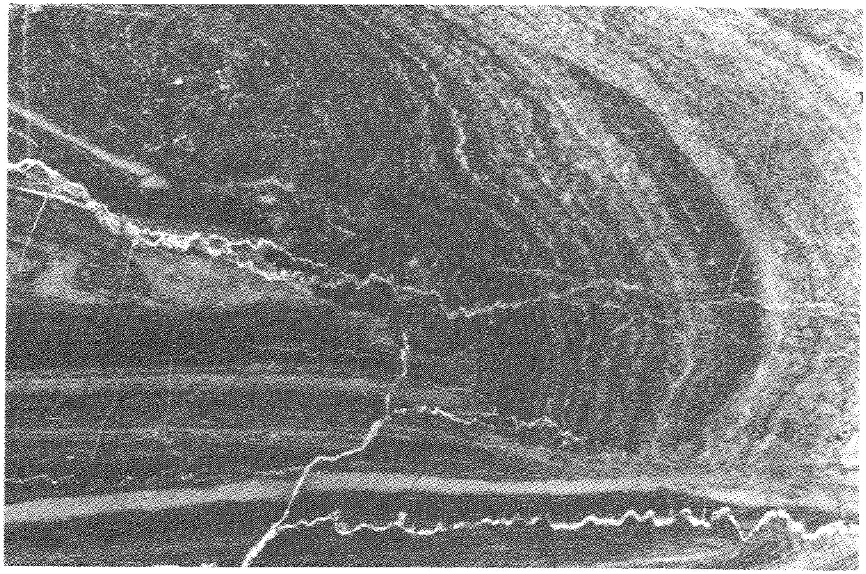

Figure 78. **Fold profile, showing flattened streaks in fold limb and thicker streaks in fold crest. Specimen 34502 M.**

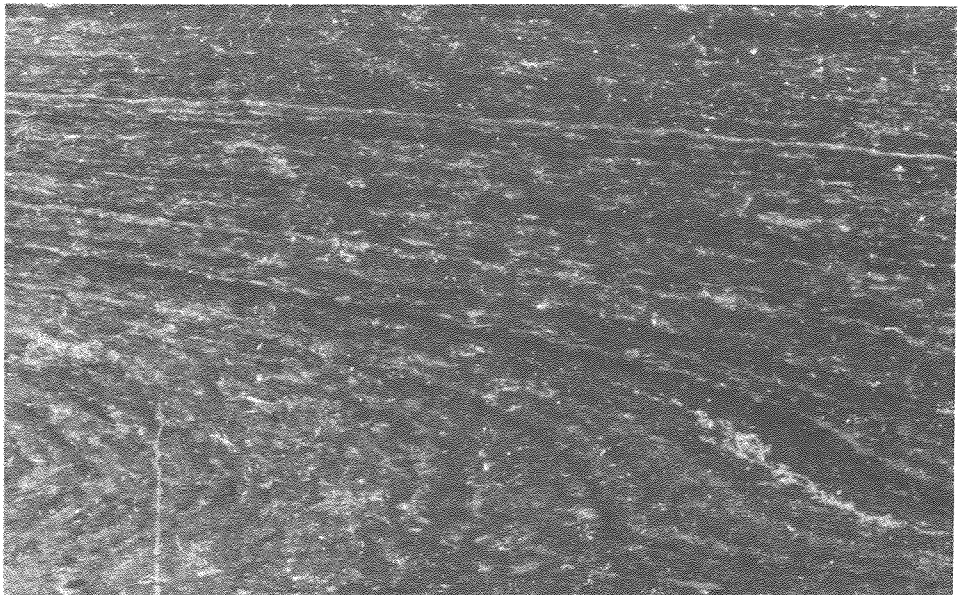

Figure 79. **Fine laminations and streaky texture which are widely spaced on the right side, become thinner as they drape over the intrafolial fold.**

ORIGIN OF FOLDS

The folds formed either by soft-sediment deformation (syn-sedimentary) or by tectonic deformation. These processes are, clearly, not mutually exclusive, and there are numerous criteria for distinguishing between the two (Potter and Pettijohn, 1963; Williams and others, 1969; and Hobbs and others, 1976). Many criteria have been found to be unsatisfactory. Thus, criteria such as style (Woodcock, 1976), confinement to a single layer (Hobbs and others, 1976), overturning in opposite directions (Williams and others, 1969), and the presence of axial-plane cleavages (Williams and others, 1969; Corbett, 1973) have all been demonstrated to be ambiguous for determination of the time of formation of folds.

In the Discovery Chert, the absence of joints and veins that can be directly related to the folding and the absence of bent clastic sericite grains in fold closures suggest the chert was soft during folding. Such plasticity may be interpreted as evidence for folding early in the history of the chert. However, there are three features which indicate the intrafolial folds are different from the early open cast slumps illustrated in Figure 63.

(1) Axial planes of folds are at a low-angle to bedding.

(2) The folds die out parallel to bedding by a transition from isoclinal recumbent fold, to asymmetric fold, to monocline (Fig. 79). Fine laminae overlying the monoclinal section are widely spaced and distinct but become more closely spaced on top of the isocline (Fig. 79), where the individual laminae lose their identity and the chert has a homogenized streaky texture formed by remnants of disrupted thin laminae.

(3) Braided microlamination is thought to be derived from the disruption of parallel microlamination and suggests that the braided texture formed late in the compaction of the chert. Intrafolial folds deform the braided microlamination and presumably formed late in the history of compaction.

All these points suggest the folds were controlled in their attitude and formation by an overlying constraining bed. The sharp basal contact of the fold complex in Figure 78 indicated that chert under the folds was relatively competent. Above the fold zones the laminae are undisturbed and were also competent. Intrafolial folds seem to have formed in zones where plasticity was maintained while the remainder of the sediment was undergoing more advanced compaction and lithification. Such zones may have been rich in connate water. The sense of movement of the folds as deduced from the direction in which the fold noses point, is variable within any one horizon. Consequently the folds are not the result of a uniformly directed force, such as gravity, but, rather, more localized forces, to which the main response was movement parallel to bedding and flattening of folds in a direction perpendicular to bedding. Such a strain pattern suggests the intrafolial folds were generated by the movement of the slabs of lithified chert along the zones of plastic chert.

TIME OF FOLD FORMATION

The fine lamination and the low, clastic content indicate that the chert was deposited in a quiet, stable environment which would not experience stresses to

Figure 80. Stereo-pair of undulose bedding in the floor of a gorge, Discovery Chert, Peedawarra Syncline.

trigger movement of large slabs of chert. Breccias as well as folded, braided microlamination in the intrafolial fold zones formed after most chert had been compacted. Therefore, the low-angle faults and intrafolial folds formed when the chert still retained plastic zones which had persisted until the basin became less stable. Intrafolial folds have been folded by minor folds which are part of the tight regional folds. Hence, the intrafolial folds formed before the tight regional folds and probably during the earliest phase of broad, regional folding which was a response to movement on basement horst and graben structures (*see* A MODEL FOR BASIN DEFORMATION).

UNDULOSE BEDDING

Undulose bedding is a bedding-surface feature expressed as irregular swells and depressions from 0.2 to 2 m across. Amplitude ranges to about 0.1 m (Figs 80, 81). It occurs in zones 1 m or more thick, and may pervade a section through a chert unit from top to bottom. Most sections through the Discovery Chert have at least one such zone. Undulose-bedded zones grade vertically into normal planar bedding. Most affected beds have a pinch-and-swell geometry in vertical section, but disconnected pods are also seen.

Undulose bedding is not restricted to the Discovery Chert. It occurs in a chert member of the Kiangi Creek Formation on Staten Hill. It is also present in cherts of the upper Backdoor Formation at the western end of the Collier Range and chert

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Figure 81. Vertical section of undulose bedding in Discovery Chert, Peedawarra Syncline, MOUNT PHILLIPS.

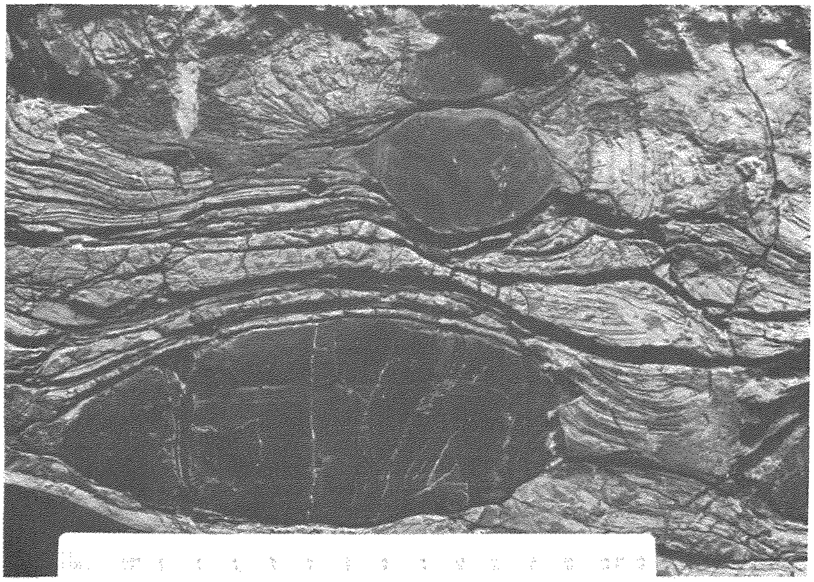

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Figure 82. Chert pods, which have replaced dolomite prior to compaction, Irregully Formation. Irregully Creek gorge. An example of type II undulose structure.

and lutite of the Jillawarra Formation and rarely, in dolomite of the Irregully Formation (Fig. 82). Undulations are irregular and have no preferred orientation except in the Collier Range chert, which has elongated undulations at two levels a couple of metres apart, in mean orientations of 096° and 020° (Fig. 83). The 096° direction is close to one of the main structural trends of the region, suggesting a partial tectonic control of the undulation orientation. Steinitz (1970) reports a

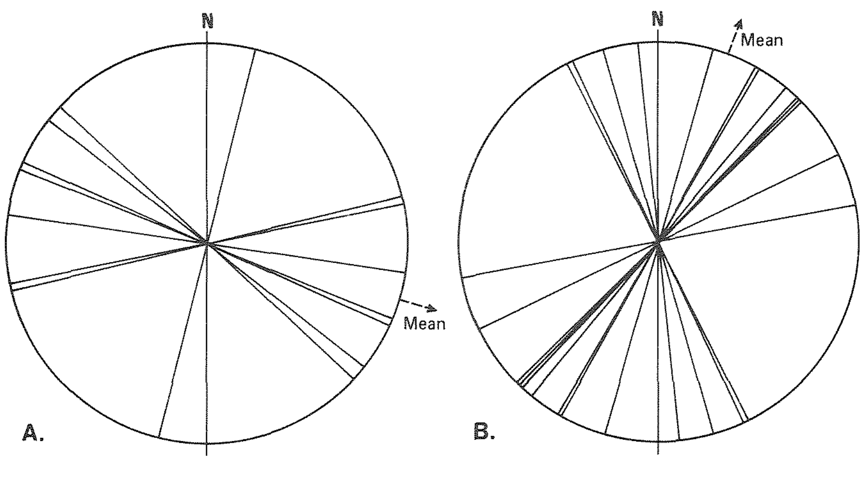

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Figure 83. Orientation of elongate undulose bedding in two chert beds in the upper Backdoor Formation, west end of Collier Range. A—Upper bed. B—Lower bed.

similar correlation between structural trends and elongate protrusions from chert beds. The 020° average does not correspond to any prominent structural direction, though it is almost normal to the 096° trend and may have been influenced by the same stress field.

CLASSIFICATION

Undulose bedding is classified according to three criteria:

(1) the presence or absence of lamination between bounding surfaces;

(2) the concordance or discordance of bounding surfaces to exterior lamination;

(3) the relation of interior lamination to bounding surfaces. Five types of undulose bedding have been observed (Table 23). Types II and IV, recorded from the upper Backdoor Formation in the Collier Range, are not known from the Discovery Chert.

TABLE 23. CLASSIFICATION OF UNDULOSE BEDDING

	Bounding surface concordant with exterior lamination	*Bounding surface discordant with exterior lamination*
	Lamination between bounding surfaces	
Interior lamination conforms to bounding surface shape	TYPE I	TYPE III
Interior lamination does not conform to bounding surface shape	TYPE II	Not observed
Exterior lamination continuous with internal lamination	Not possible	TYPE IV
	No lamination between bounding surfaces	
	TYPE V	Not observed

DESCRIPTION AND ORIGIN OF TYPES

Type I undulose bedding is by far the most common, and normally consists of a stacked array of irregularly pinching and swelling layers, where a swell in one is compensated by a necking in the vertically adjacent layer. Non-fissile layers may be in direct contact, but close inspection will usually reveal a thin interbed of fissile rock, perhaps only millimetres thick. The fissile interbed can range in thickness up to about that of the non-fissile ones. Occasionally adjacent non-fissile layers conform to each other and resemble folded strata; this is an accommodation of both layers to a neighbouring swell and happens at random for only limited distances. Lensing-out of layers and podding are rare and pass laterally into fissile beds. Internal laminations, being conformable to the bounding surfaces, are usually thicker in swelled areas where, under the microscope, streaky texture is seen to be thicker than the streaks in pinched areas.

The swells and depressions resemble large-scale load casts in plan, so that it is simplest to envisage them as a compaction structure. The varying thickness of internal streaky texture supports such a plastic flow mechanism, but it still needs to be explained why most chert does not have this feature. Affected zones would need to consist of hydroplastic layers of slightly different viscosities or densities. This variation may be due to higher than normal interstitial water content, or possibly because some zones were originally of the less dense mineral, opal, as suggested by Kirchmayer (1959) for Nevadan Carboniferous cherts. Where adjacent beds consisted of material with uniform physical properties, undulose bedding could not be developed.

Type II undulose bedding is uncommon. It differs from Type I in that the interior lamination terminates against the upper and lower bounding surfaces (Fig. 82), or is asymptotic to them. Stylotites within swells end the same way. The upper bounding surfaces can be either planar or undulose. Disconnected pods are more common in Type II than in Type I undulose bedding. Weathered-out lenses display upper surfaces where the internal lamination meets them to give a contour effect. Lenses of black chert in the Collier Range appear to have been less compactable than the enclosing grey fissile chert.

Isolated pods of black chert in dolomite in the Irregully Formation (Fig. 82) are a variant of Type II undulose bedding and provide the clue to the formation of this structure. Silica is precipitated in chemically favourable places as layers or lenses which cut across the original bedding and partly replace the original rock. Tectonic stress may have influenced the orientation of some lenses. The unsilicified enclosing rock was then compacted, resulting in a tight wrapping of lamination around the cherty core, within which the original attitude of lamination was preserved. At the bounding surfaces, plastic deformation of incompetent rock against resistant chert smeared out laminations to give an apparently disconformable contact. In many cases, stylolitic solution at the bounding surface accentuated the disconformity. Tectonic stress influenced the orientation of some lenses by differential stylolitic solution, as in the Collier Range (Fig. 83). Where the enclosing rock is a fissile chert, an enclosing shale was affected by a second, post-compaction silicification. The silica for this second silicification was probably derived from stylolitic solution in nearby layers. The light colour of fissile chert is due to leaching by weathering of the original dark colour.

Type III structures are uncommon. They consist of laminations conformable with a lower bounding surface which cuts across underlying lamination. A well-defined upper bounding surface is commonly absent, and the bedding tends to grade upwards into normal planar bedding. It strongly resembles low-angle scour structures occasionally found in shale and sandstone, including the shale underlying the chert in the Collier Range. The feature can thus be interpreted as a scour structure, infilled by suspension draping, and then compacted.

Type IV undulose bedding consists of lamination continuous across a bounding surface defined by a change in lithology. Only one example of this type in which black, non-fissile chert is overlain by light-grey fissile chert is known. The laminations thicken upwards on entering the black chert to form a hump. This structure is attributed to post-depositional silicification, forming a lens cutting across the original bedding, and the subsequent compaction of unaffected sediment around the lens to a lesser extent than during the formation of Type II structures.

Type V structures are similar to Types I and II, but internal lamination is missing. They are moderately common. Many undulose beds appear to be Type V at first sight, but on close inspection show faint lamination. Others which are massive in hand specimen may reveal lamination in thin section. Type V beds are regarded as Types I and II, in which the layer involved was unlaminated. Their origins can therefore be those of either Types I or II.

Some Type V structures are known from the Jillawarra Formation, involving massive mudstone and shale. In this case there is no silicification as postulated for Type II, and generation by compaction under loading of beds of differing densities and viscosities is favoured.

CONCLUSION

Undulose bedding of Types I, III and V, which occur in the Discovery Chert and other formations, can be attributed to loading and compaction of sediments in which adjacent beds differed in some way, e.g. density, water content, greater thickness in a scour fill.

Types II and IV, which are recorded only from the upper Backdoor Formation of the Collier Range, require the additional process of silicification to form a lens cutting across the original bedding, before the sediment was compacted.

MINOR STRUCTURES

RIPPLE MARKS

Ripples are almost non-existent in the cherts. Only two occurrences have been noted, both in the Discovery Chert.

One of these, 17 km north-northwest of Mulgul, has both deformed and undeformed symmetrical ripple sets, the latter having wavelengths of 120 mm and a ripple index of 13.3. Some bifurcations are present. They are either wave-oscillation ripples or tidally modified current ripples, each implying shallow water.

The second example is from the Cobra Synclinorium and consists of symmetrical ripples of 70 mm wavelength and ripple index of 14. No bifurcations were visible in this small exposure. Faint internal laminations are asymptotic to, or cut off by the ripple surface, suggesting erosion. They are interpreted as longitudinal ripples formed in shallow water, parallel to tidal currents.

MACROSPHERES

The Discovery Chert east of Quartzite Well (Range Creek Syncline) has a few layers containing silica spheres up to 30 mm in diameter. They consist of aphanitic, massive silica, lighter in colour than the enclosing chert. Some of these bodies are lens- or bean-shaped, and aligned along the bedding. The surrounding laminations are not transgressed, except perhaps some centrally disposed ones, but bend smoothly to pass around the spheres. The spheres were probably formed by the replacement of silica gel by sulphide before diagenesis. Later differential compaction wrapped the laminations around them, and finally silica replaced the sulphide in the spheres.

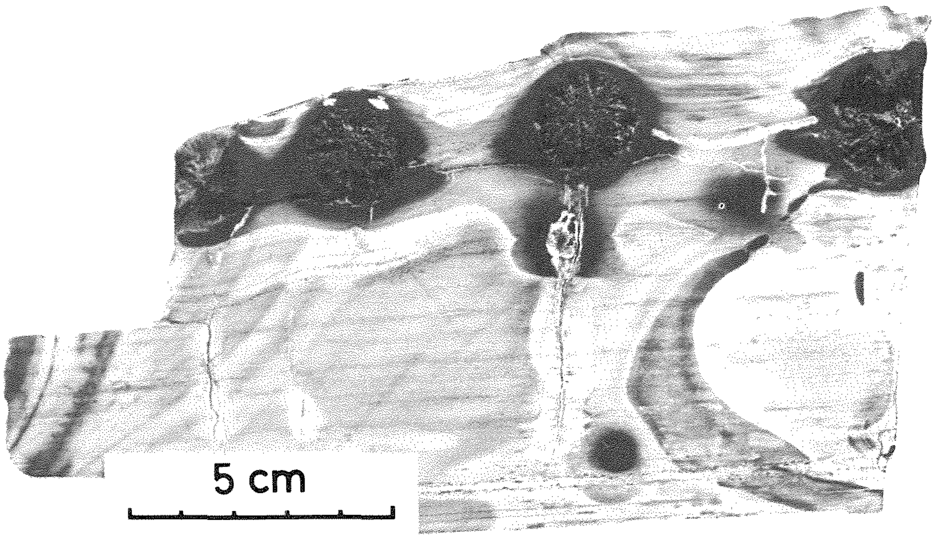

Figure 84. ***Goethite spherulites after ?marcasite in weathered, silicified claystone of the Backdoor Formation, Brumby Creek Anticline, COLLIER.***

Support for this origin is given by macrospheres of similar size in silicified claystone of the Backdoor Formation in the Brumby Creek Anticline. These have well developed internal radiating fabric of goethite and have grown across the lamination in the rock (Fig. 84). Some bending of lamination can be attributed to compaction. These spheres may have grown as balls of marcasite or evaporite minerals in the clay matrix but did not undergo later replacement by silica.

BRECCIA

Breccia textures are rare in the Discovery Chert. A few examples are in the upper part of the chert profile in the eastern area of deposition, namely the Mount Vernon syncline, Brumby Creek, and Coobarra sections. The breccia forms lenses broadly concordant to bedding, up to 30 cm wide and 2 to 3 m long. In detail, the breccia is intrusive into laminations at both upper and lower contacts. Angular fragments of black, laminated, streaky chert, some of which clearly once fitted together (Fig. 85) are scattered in a microcrystalline quartz and clay matrix. The breccia is comparatively late in the diagenesis of the chert; it was formed after compaction and development of the streaky texture, as well as after formation of some stylolites which are confined to fragments and are at a high angle to the chert fragment boundary.

The breccia at Brumby Creek is associated with intrafolial folds and low-angle faults. It may represent a brittle-style adjustment of consolidated chert during movement along the intrafolial fold zones. The breccia does not seem likely to have formed by collapse resulting from dissolution of evaporite minerals or carbonate because no relicts of any of these minerals have been recognized.

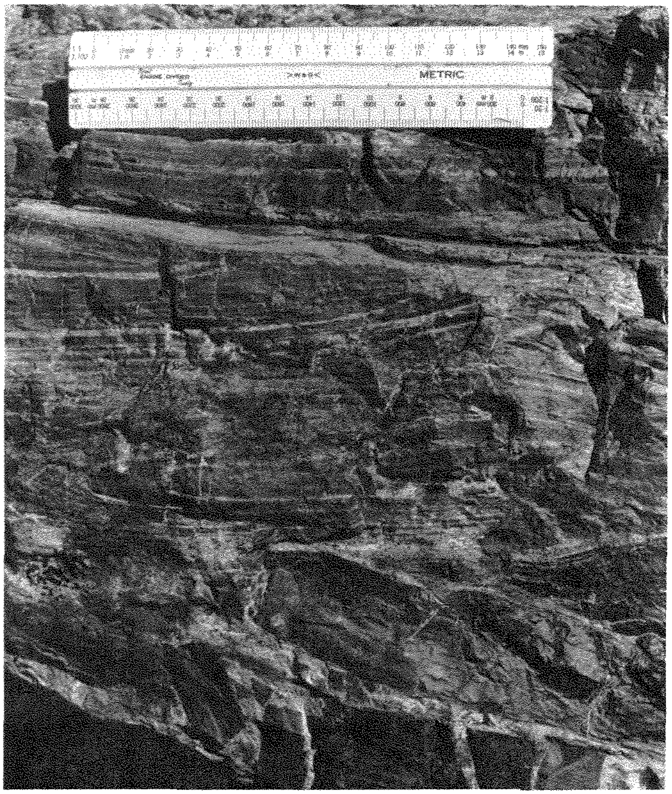

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Figure 85. Breccia underlying an intraformational fold zone, Brumby Creek, COLLIER.

STYLOLITES

Stylolites are irregularly distributed and widespread in the Discovery Chert. They have cut across intrafolial folds and low-angle faults. Stylolites have formed both before and after brecciation in the chert. Although stylolites have clearly modified many of the textures and structures in the chert, it is thought no structures are due solely to stylolitic solution and therefore this aspect has not been studied here any further.

ORIGIN

Any theory of the origin of the Discovery Chert must account for the following features.

(1) The Discovery Chert was deposited in shallow water as indicated by the geological setting of the adjacent formations.

(2) Deposition was a uniform event which took place simultaneously over a wide area of the Bangemall Basin. This is shown by the chert's blanket-like relation to underlying formations, its continuity, its small range in thickness, and its uniform composition.

(3) The chert was precipitated in quiet water, as witnessed by the fine lamination, absence of coarse material, the dearth of current structures, and the geometry of the isopachs which outline a linear depocentre along the contemporary basin axis.

(4) Water circulation was restricted, resulting in anaerobic and locally hypersaline conditions. Preservation of organic material, crystallization of some minor gypsum and carbonate, and the lack of tidal or other currents can be cited as evidence.

(5) There was negligible clastic input.

(6) Conditions favourable for the deposition of the chert were not basin-wide. To the east, normal marine shelf conditions prevailed in the Backdoor Formation.

Features which rule out a volcanic origin are:

(1) The absence of volcanic rocks anywhere near the Discovery Chert and presence of only very minor volcanic rocks in the Bangemall Basin.

(2) The isopach map (Fig. 59) shows a thickness variation incompatible with the derivation of the chert from the only known volcanic rocks which are in the Tangadee area.

(3) No volcanic textures, such as shards, have been recognized in the chert.

(4) The major oxide and trace element analyses in both the Discovery Chert and the top of the Jillawarra Formation have features typical of fine grained sediments, free from volcanic detritus (Pettijohn, 1975). For example K_2O is mostly greater than CaO, Na_2O:K_2O ratios are low, and there are no indications of mafic volcanic content such as high MgO, and CaO.

(5) The top of the Jillawarra Formation in the Mount Vernon Syncline contains an unusually high proportion of jarosite, which implies large amounts of pyrite were present. Elsewhere in the Jillawarra Formation, pyrite has sulphur isotopes compatible with a sedimentary origin (Marshall, 1968) so volcanic or exhalative pyrite is probably not present.

If the chert were produced from seawater, then silica either replaced sediment or was precipitated from solution by an organic or inorganic mechanism.

REPLACEMENT OF SEDIMENT

As the Discovery Chert was a gel in its early stages, it could only have replaced unlithified sediment.

Black shale: There are a number of features common to black shale and chert, which might suggest the chert may be due to replacement of a black shale. Both rocks show streaky texture, and structures common in chert, such as small scale recumbent folds and associated low angle faults. Also silicified shale is moderately common and can be traced along strike into unaffected shale. However, silicified shale has a clastic texture, with detrital quartz and sericite-muscovite grains, and is distinct from chert which contains a very small component of clastic quartz and phyllosilicate.

The basal contact of the Discovery Chert in the Mount Vernon syncline is a transition zone which consists of chert layers, up to about 5 mm thick, delicately interlaminated with shale (Fig. 86). The chert layers have sharp boundaries and the intervening shale layers show no siliceous alteration or cross-cutting silica veins. A zone (Fig. 88) displaying similar features is present in the Collier Range. Such relationships between chert and black shale indicate that replacement of the shale to form chert is unlikely.

Carbonate: There are numerous documented examples of the formation of chert by the replacement of carbonate sediment, viz. chertification of the Redwall Limestone, Arizona (McKee, 1960), the Coomberdale Chert, Western Australia (Logan and Chase, 1961), chert nodules in the Thornton reef complex, Illinois (Weiner and Koster van Groos, 1976) and cherts of the Tsumeb Subgroup, southwestern Africa (Germs, 1975). Replacement can occur before or after lithification (Steinitz, 1970).

Replacement of carbonate has taken place on a small scale in the Bangemall Group. Stromatolites, breccia beds and laminations in the Irregully Formation have been silicified prior to compaction (Fig. 87). There are also small occurrences of carbonate in the Discovery Chert. Rhombs, after carbonate, appear at a number of localities near the edge of the area of chert. However, many rhombs appear to be late rather than relict in the chert as they cut across chert pellets and intrafolial

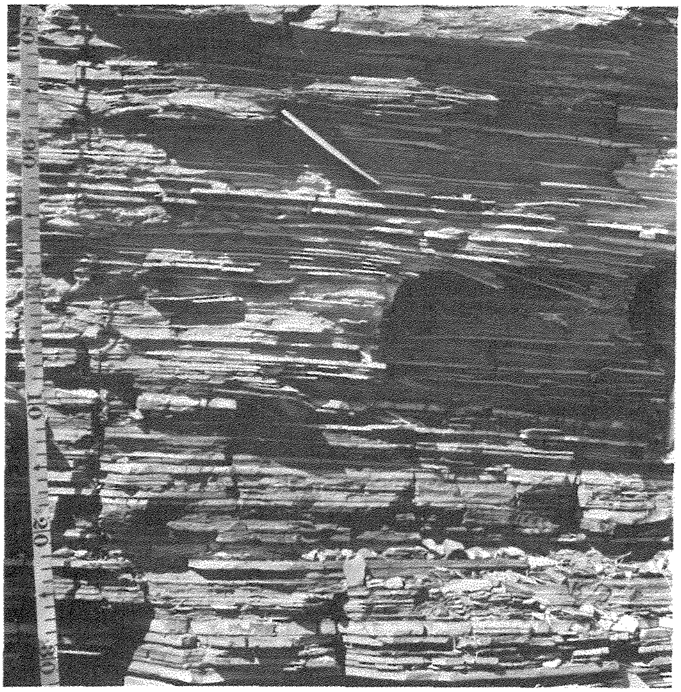

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Figure 86. Interlaminated chert and shale in the transition zone between Jillawarra Formation and Discovery Chert, Mount Vernon Syncline, MOUNT EGERTON.

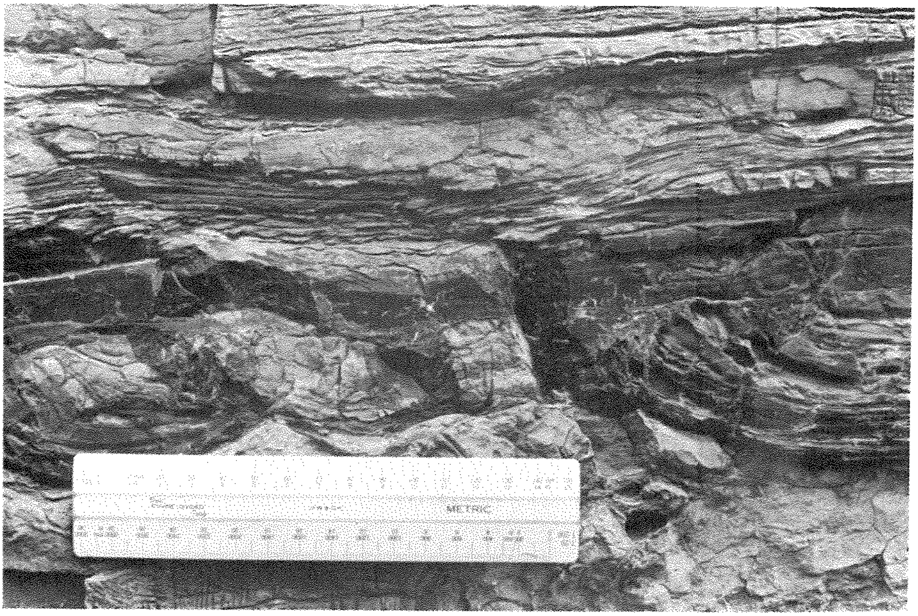

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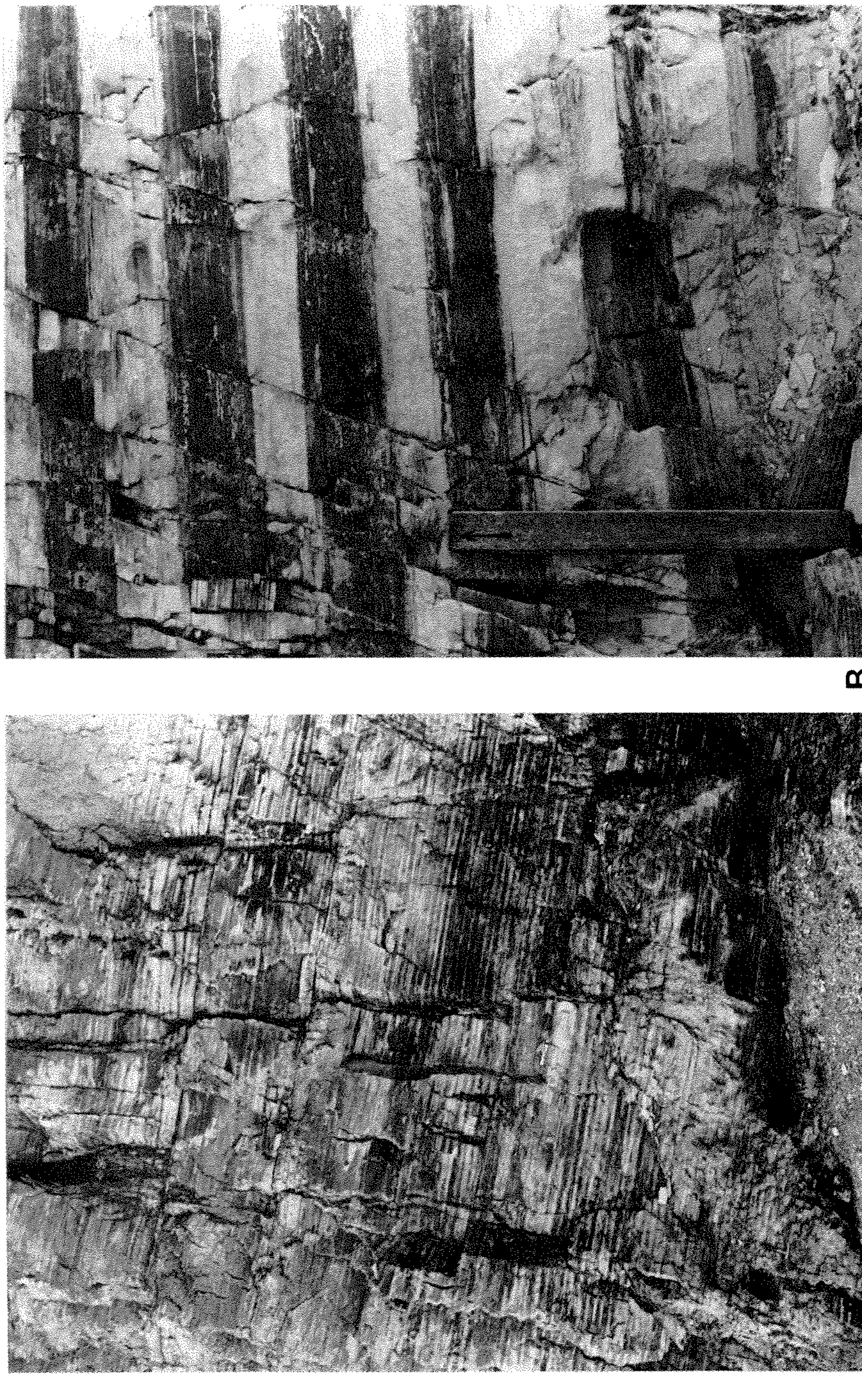
Figure 87. Black chert has partially replaced pale dolomite breccia, Irregully Creek gorge, EDMUND.

folds. Marshall (1968) reports 'incompletely replaced dolomite laminae and scattered dolomite crystals' in thin sections of the Discovery Chert. These appear to be a local occurrence.

Regional features of the Discovery Chert show that chert did not replace a carbonate. Where the Discovery Chert is close to a major carbonate unit there is no evidence of chert infiltration into carbonate. In addition, the chert is one of the few formations which has no carbonate members and which does not pass laterally into carbonate. Finally, the dolomites in the basin have variable, commonly lenticular, shapes up to hundreds of metres thick and quite clearly do not resemble the blanket geometry of the chert.

The Discovery Chert contains almost no CO_2, low CaO, and low MgO (Table 26). These values are similar to those in primary cherts (Table 25) and in nodules and layers in carbonate rocks, so absolute values of the oxides are of little assistance in distinguishing between cherts which are replaced carbonates and those which are primary (Table 24). A ratio between carbonate CaO, MgO, CO_2, and non-carbonate Al_2O_3, Fe_2O_3 components was examined for cherts of various origins (Cressman, 1962). The ratios for cherts with carbonate rocks are high and form a group distinct from the low ratios typical of cherts not associated with carbonates. The Discovery Chert ratios are low and group with those cherts not associated with carbonates (Table 24).

Replacement of Evaporite: Microscopic grains of fresh gypsum and barite are present near the eastern edge of the Discovery Chert at Coobarra Creek (COLLIER). These grains are enclosed in microcrystalline quartz and so predate the formation of the chert. It is possible these grains are relics of an evaporite sequence. However, we have seen no evidence that the entire Discovery Chert is a



B

A

Figure 88. **Rhythmic layering in the upper part of the Backdoor Formation, western end of Collier Range. A—View of cliff showing vertical extent of rhythmite. (Hammer in lower left gives scale). B—Close-up of layering. The light-tone layers are shale, whereas the dark layers consist of chert-shale laminae.**

TABLE 24. PARTIAL ANALYSES OF CHERT NOT FORMED BY REPLACEMENT OF CARBONATE

	1	*2*	*3*	*4*	*5*	*6*	*7*	*8*	*9*	*10*
Al_2O_3%	0.11	13.01	14.85	15.37	1.39	10.65	1.70	1.80	0.10	5.23
Fe_2O_3%	0.00	5.96	6.51	6.85	0.69	4.02	0.78	0.36	3.40	2.47
CaO%	0.00	1.87	2.26	1.86	0.00	0.48	0.41	0.49	0.60	11.67
MgO%	0.00	2.34	2.58	3.58	0.26	0.79	0.23	0.64	0.19	0.81
CO_2%	0.00	0.20	0.51	0.70	0.00	0.00	0.06	0.60	0.40	7.90
C/NC	0.00	0.23	0.25	0.28	0.13	0.09	0.28	0.80	0.34	2.65

Source: Cressman, 1962, 1. Novaculite, Table 1, 2. Diatom ooze, Table 3. 3-7. Radiolarian rocks, Table 4. 8-10. Spicular chert, Table 5.

$C/NC = CaO + MgO + CO_2 / Al_2O_3 + Fe_2O_3$

TABLE 25. PARTIAL ANALYSES OF CHERT IN CARBONATE ROCKS

	1	*2*	*3*	*4*	*5*	*6*	*7*	*8*	*9*
Al_2O_3%	0.14	0.32	0.45	1.67	0.07	1.92	0.12	0.35	0.9
Fe_2O_3%	0.40	0.09	0.02	1.03	0.07	0.88	0.00	0.60	0.3
CaO%	1.11	1.25	12.90	10.80	1.25	5.40	0.16	2.25	2.7
MgO%	0.71	0.21	1.88	5.45	0.49	0.20	0.01	0.10	0.8
CO_2%	1.38	1.17	12.04	13.90	1.20	4.10	0.07	1.45	3.6
C/NC	5.93	6.41	57.06	11.17	21.00	3.46	2.00	4.00	5.92

Source: Cressman, 1962, Table 7. $C/NC = CaO + MgO + CO_2 / Al_2O_3 + Fe_2O_3$

TABLE 26. PARTIAL ANALYSES OF DISCOVERY CHERT AND CHERT IN THE JILLAWARRA FORMATION

	Discovery Chert				*Jillawarra Cherts*			
GSWA NO	*50701*	*50707*	*50710*	*50407*	*50730*	*50731*	*50708*	*50718*
Al_2O_3%	0.3	0.2	1.3	2.0	4.1	6.8	3.7	0.3
Fe_2O_3%	0.4	0.0	0.4	2.4	0.4	1.3	0.2	0.3
CaO%	0.0	0.0	0.06	0.12	0.79	2.0	0.0	0.0
MgO%	0.10	0.10	0.17	0.08	0.08	0.53	0.41	0.05
CO_2%	0.31	0.12	1.45	0.06	0.33	1.63	0.81	0.20
C/NC	0.58	1.1	0.99	0.29	0.44	0.51	0.31	0.42

No 50710 contains carbonate veins derived from Devil Creek Formation

No 50731 contains carbonate and is from a 20 mm lens of chert in shale

$$C/NC = CaO + MgO + CO_2/Al_2O_3 + Fe_2O_3$$

replaced evaporite. There are local indications of shallow-water deposition such as clastic-textured chert produced by penecontemporaneous erosion, but this together with carbonate rhombs and abnormally high clastic content appear to be local features near the edge of the Discovery Chert.

Features commonly accompanying evaporite sequences such as desiccation cracks, halite casts, layers of evaporite minerals (or their replacement textures) have not been recognized in either the chert or adjacent formations. The gypsum and barite occur with structures indicating shallow-water deposition at the eastern edge of the chert, but neither these minerals nor the evidence for shallow water appear to extend into the main basin of chert deposition.

ORGANIC OR INORGANIC DEPOSITION

Sedimentary silica can be formed by the accumulation of the skeletons of silica secreting organisms. Although carbonaceous layers and acritarchs (Marshall, 1968) indicate organisms were present during deposition of the Discovery Chert, no relict siliceous organic material has been recognized. Also, the chert spheres (*see*, chert spheres) indicate that the early form of the Discovery Chert was a gel and not derived by recrystallization of a deposit of siliceous remains *in situ*, in the manner suggested by Ernst and Calvert (1969) for the Monterey Formation. Complete dissolution of a deposit of siliceous tests to form a gel is unlikely because the rate of dissolution of siliceous material in seawater is very slow (Blatt and others, 1972), and there is no evidence (such as corroded clastic quartz grains) to indicate silica had been taken into solution during formation of the Discovery Chert. All these points indicate the silica for the chert was not deposited by silica secreting organisms and that the alternative is inorganic precipitation.

Inorganic silica deposition: Ideas on precipitation of silica in Precambrian rocks are based on the chemistry of modern seawater. The concentration of silica in modern seawater is about 1-2 ppm (Krauskopf, 1967). This low level is due to continuous removal of silica either by the action of silica secreting organisms such as diatoms or sponges; the precipitation of silicates (e.g. glauconite, authigenic feldspar) as suggested by Krauskopf (1967) or the adsorption of silica by clay minerals (Garrels and MacKenzie, 1971).

There are good indications that these factors did not operate during the time of deposition of Discovery Chert and so allowed the concentration of silica in seawater to rise. Silica extracting organisms were present at least as far back as the Cambrian (Cressman, 1962) and possibly the Late Proterozoic. However, all the available evidence indicates silica extracting organisms did not contribute to the Discovery Chert.

The adsorbtion of silica by clay minerals and the precipitation of silicates was probably negligible during deposition of the chert, as almost all influx of clastic silicate minerals and clays had ceased.

An additional factor assisting the concentration of silica, is evaporation. Gypsum crystal molds in the Discovery Chert indicate evaporation was significant.

If the silica concentration of present day surface seawater is raised to about 30 ppm, a sepiolite-like phase precipitates (Mackenzie and others, 1967). However, Drever (1974) suggests that with the low activity of Mg^{+2}, amorphous silica is likely to be a more stable precipitate in water with pH lower than 7.7.

Inorganic deposition of silica in a closed lake basin has been recorded by Peterson and van der Borch (1965). Algal activity increases the pH of the water, thus dissolving clastic silicate grains, producing silica in solution. Silica is deposited from solution when the pH is lowered by flooding. This demonstrates inorganic deposition of silica, but is inadequate to explain the amounts of silica required for the Discovery Chert because the main rock deposited in the lake is carbonate with silica comprising only 1-6 per cent.

Early diagenetic silica is commonly associated with carbonates (e.g. replacement of stromatolitic layers in the Irregully Formation), which suggests that conditions for deposition of both carbonate and silica should be very generally similar. The problem is to concentrate and precipitate chert without deposition of carbonate.

A significant factor determining which mineral would precipitate is pH: a more acid environment encourages deposition of silica and dissolution of carbonate. The Jillawarra Formation is dominantly black shale, with some pyrite, pointing to deposition under anaerobic conditions similar to those where black shales are forming today—stagnant marine basin—the bottom of the Black Sea, and deep spots in fjords (Krauskopf, 1967). The pH of water in these localities is lower than normal seawater, between 5.5 and 7 (Krauskopf, 1967) where it is in contact with black muds.

Therefore, chert may be produced by the combined effect of a small amount of evaporation to precipitate silica and a low enough pH (5-7) to keep carbonate in solution.

Origin of lamination: The formation of laminations poses a major problem: whether the carbonaceous material grew at the site of chert deposition (i.e. algal mat), or accumulated as a deposit of pelagic organisms.

A few sections of Discovery Chert (Barlee Range, EDMUND), show features of lenticular, carbonaceous microlamination suggestive of an algal mat. Marshall (1968) also proposed an algal-mat origin for cherts with similar features in the Mount Palgrave area. However, most of the Discovery Chert does not show these features. Therefore, are the characteristics of the 'algal-mat facies' due to a distinctive species of organism different from those forming most of the carbonaceous layers in chert with which it is laterally continuous, or the result of a slightly different environment acting on the same organisms which form the bulk of the carbonaceous layers in chert? Little evidence has been recognized: no well-preserved algal filaments are persent (Marshall, 1968; Grey, 1978), because recrystallization of the quartz has redistributed carbonaceous material so that no diagnostic shapes, such as algal filaments remain. The different subdivisions of lamination described above (*see* Lamination) are present in all the carbonaceous laminated cherts. This implies that, despite the different styles of microlamination, there was a common factor controlling deposition of carbonaceous material througout the basin at about the same time. This suggests the carbonaceous matter is the remains of pelagic organisms which were distributed throughout the basin. The chert with features of algal mats may, therefore, have been derived from pelagic organisms deposited in an environment (very shallow water) suitable for development of gypsum crystals, crinkling of laminae, very small scale penecontemporaneous erosion and growth of diagenetic carbonate.

CONCLUSION

The only feasible theories of the origin of the cherts are direct organic or inorganic deposition of silica rather than replacement of pre-existing rocks. On the evidence available we prefer inorganic deposition of silica because no remains of siliceous organisms have been recognized in the chert. This is consistent with the findings of many other workers on Precambrian cherts (e.g. Marshall, 1968; Trendall and Blockley, 1970). Microspheres in the Discovery Chert are identical to those produced experimentally by crystallization of a silica gel (Oehler, 1976). Such a gel is almost certainly the result of primary inorganic precipitation. Although silica-secreting organisms were not present, the experiments of Oehler (1976) suggest that algae may have catalysed the precipitation of silica.

The inorganic mechanism is an attractive one, explaining all the observations in a simple internally consistent model, without recourse to conjectural methods of introducing silica. It explains the Discovery Chert as a widespread, simultaneous response to uniform chemical conditions in a shallow, stagnant body of water receiving little clastic material and protected from direct marine influence. Locally around the shallow fringes of the basin, hypersaline or even evaporitic conditions prevailed. Other black cherts in the Bangemall Basin are inferred to have formed in similar conditions, or more limited extent.

CHAPTER 6

Igneous Rocks

The Bangemall Basin contains basic intrusives and volcanogenic rocks. The intrusives represent enormous volumes of basaltic magma and were injected chiefly as sills, but also as dykes. Rhyolite and basalt occur rarely. Some bodies of high potash, fine-grained rock may be altered tuffs.

DOLERITE SILLS

MORPHOLOGY AND OCCURRENCE

Dolerite sills crop out over an area of about 143 000 km^2 in the Bangemall Basin, thus making it one of the larger intrusive tholeiite provinces of the world. Sills can exceed 100 m in thickness, and may be simple tabular bodies, or irregular bodies that vary in thickness quite markedly. The most extensive single sheet has a length of over 60 km. Sheets are generally concordant with the bedding of the country rock but may locally cross-cut it in zones that vary from less than a metre to hundreds of metres. The Fords Creek Shale in the Mount Vernon region, for example, contains two intrusive sheets which gradually and completely transgress the entire 1 700 m-thick formation.

The sills occur preferentially in shale, siltstone, and chert, and are uncommon in sandstone. The Discovery Chert and the Curran Formation, both relatively thin units, are commonly split in two by a sill. Dolerite is especially abundant in the Ullawarra Formation, usually exceeding the sediments in bulk to such an extent that sediments have been left as screens in a sea of igneous rock. Over two-thirds of the area of the Ullawara Formation on EDMUND is occupied by dolerite. Sills are absent from the Coodardoo Formation and from the coarse sandstone and conglomerate units overlying the basal unconformity. In the Irregully Formation they are present only on EDMUND. The relation between intrusives and host lithology can be attributed to the relative ease of injection of magma into the more fissile rocks at a specific stratigraphic level.

The intrusives are almost entirely confined to the Bangemall Basin, and are found in surrounding older provinces only rarely; this signifies that the sills are closely connected with the tectonic evolution of the basin. It is clear that the sills were intruded before folding of the Bangemall Group because they are folded themselves and, in some fold cores, are cemented to chlorite schist.

REGIONAL DISTRIBUTION

In addition to the stratigraphic control, a regional control is also evident (Fig. 89). The region west of the Tangadee Lineament is characterized by a high concentration of sills, which contrasts with the low concentration, or absence, to the

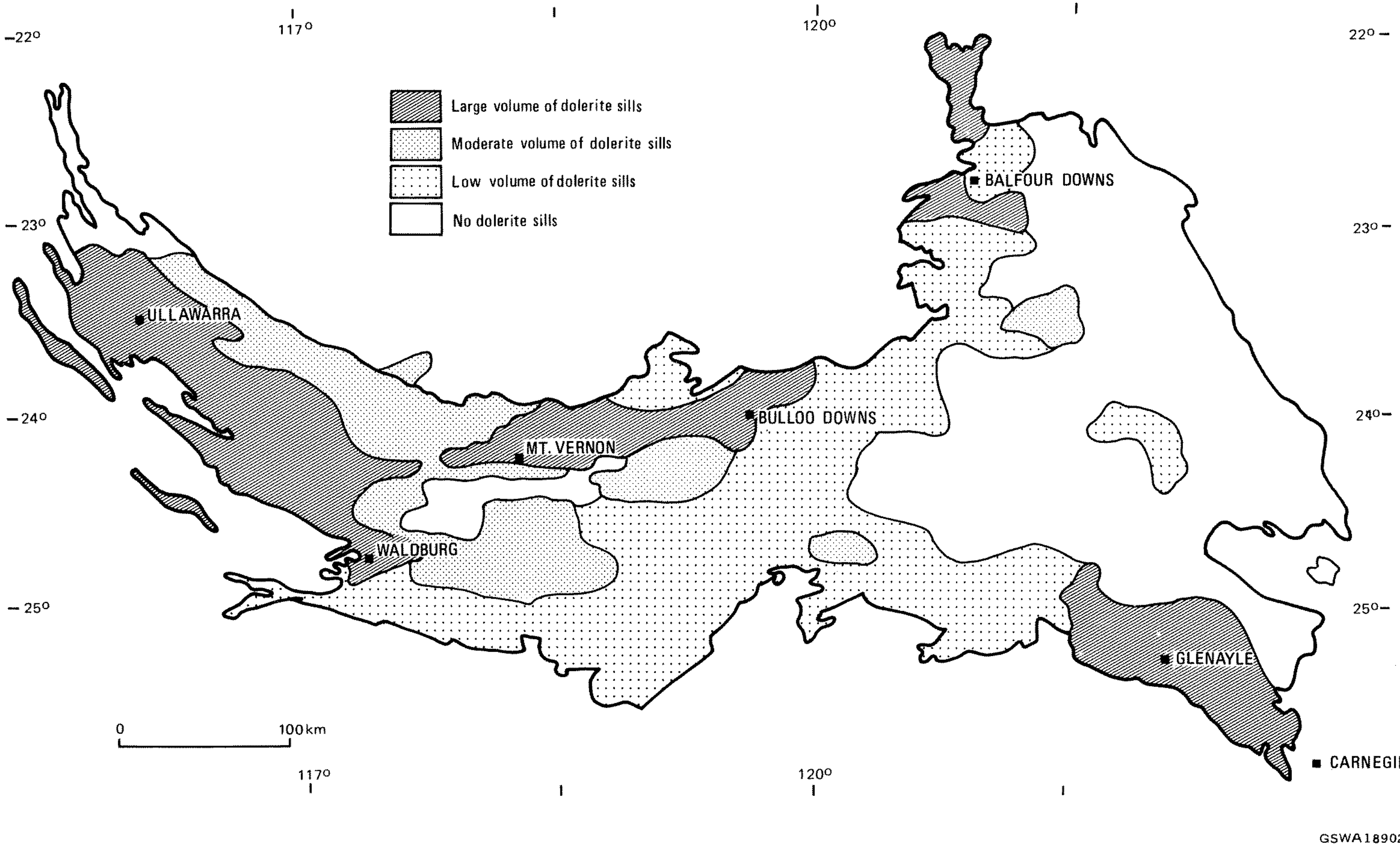


Figure 89. Abundance and distribution of dolerite sills.

east. This region correlates broadly with the greatest intensity of tectonism. Four areas in the basin have particularly large volumes of dolerite:

(1) The Ullawarra-Waldburg area coincides with the Edmund Fold Belt, the most deformed part of the Bangemall Basin. More magma was generated beneath this region than any other.

(2) The Mount Vernon-Bulloo Downs area has, by contrast, a simple structure. The area lies on the extension of the Flint Hill Lineament, but the lineament is more conspicuous to the southwest, where it is not associated with abnormal amounts of dolerite.

(3) The Balfour Downs area is not severely deformed, but it has a greater abundance of dolerite than rocks of the same lithology and stratigraphic level to the south. It thus seems to overlie a region of greater magma generation.

(4) The Glenayle-Carnegie area contains much lutite in contrast to the adjoining sandstone regions, but there is nevertheless more dolerite than can be explained by a lithology favourable for intrusion. This area, too, was supplied with copious magma.

PETROLOGY

The primary mineralogy of a typical specimen is plagioclase, augite, pigeonite or orthopyroxene, magnetite, subordinate quartz and orthoclase in interstitial granophyric intergrowths, pyrite, and, rarely, olivine. Occasionally, pyrite can comprise 5 per cent of the rock. This mineral assemblage is typical of tholeiitic rocks. Deuteric alteration to biotite, sericite, hornblende, clinozoisite, leucoxene, chlorite and bastite is ubiquitous. During the slow cooling of some of the magma, pigeonite has inverted to orthopyroxene containing exsolved blobs of calcic clinopyroxene, and the subsequent alteration of the orthopyroxene produced bastite pseudomorphs containing unaltered clinypyroxene blobs.

The texture is subophitic, and plagioclase laths are wholly or partially enclosed in larger pyroxenes. The grain size is usually less than 3 mm. Apart from fine-grained chilled margins and gabbroic phases in a few thick sills, there is little variation in the medium-grained texture. Occasional pegmatite zones are present as well as a few pink, quartz-feldspar, granophyric veins. The only known differentiated sill on ROBINSON RANGE has a higher than normal proportion of plagioclase near its top.

The petrography of the intrusive rock is uniform over the region, indicating an unvarying origin and rapid emplacement without assimilation of country rock.

Contact metamorphism of wall rocks is not usually visible in hand specimens. Xenoliths of wall rock are rare, but examples occur at a contact with Fords Creek Shale (lat. 24°19′S, long. 117°49′E), where a fine-grained chilled zone contains numerous slabs up to 0.3 m long. Flow banding is visible around the xenoliths, which themselves have been rounded and plastically deformed (Fig. 90). Daniels (1969) reported a sill whose fine-grained amygdaloidal top carries fragments of the overlying sediments. All these features suggest that at least some dolerite sills were emplaced near the surface, in incompletely lithified sediments.

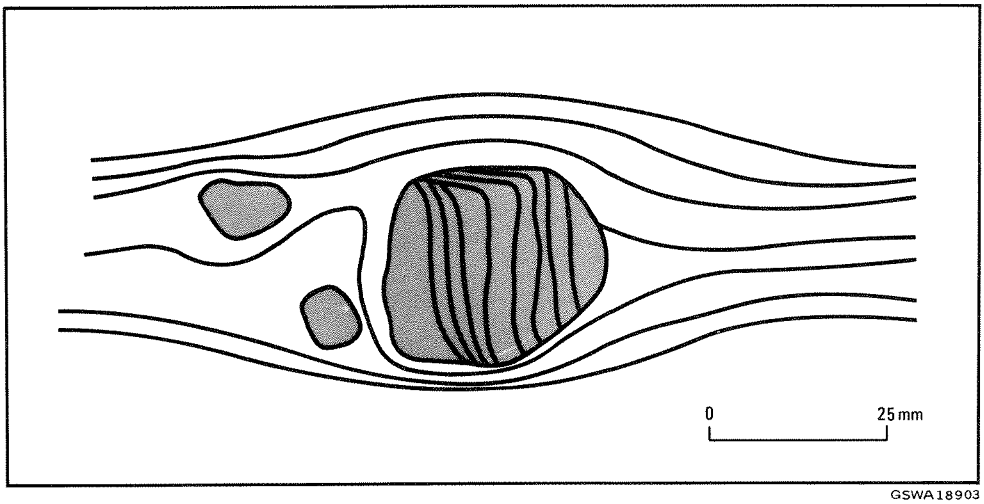


Figure 90. Wall-rock fragments of laminated shale (shaded) in the flow-banded margin of a dolerite sill (lat. 24°19′S, long. 117°49′E). Deformation of shale laminations indicate that the large inclusion was in an unconsolidated state.

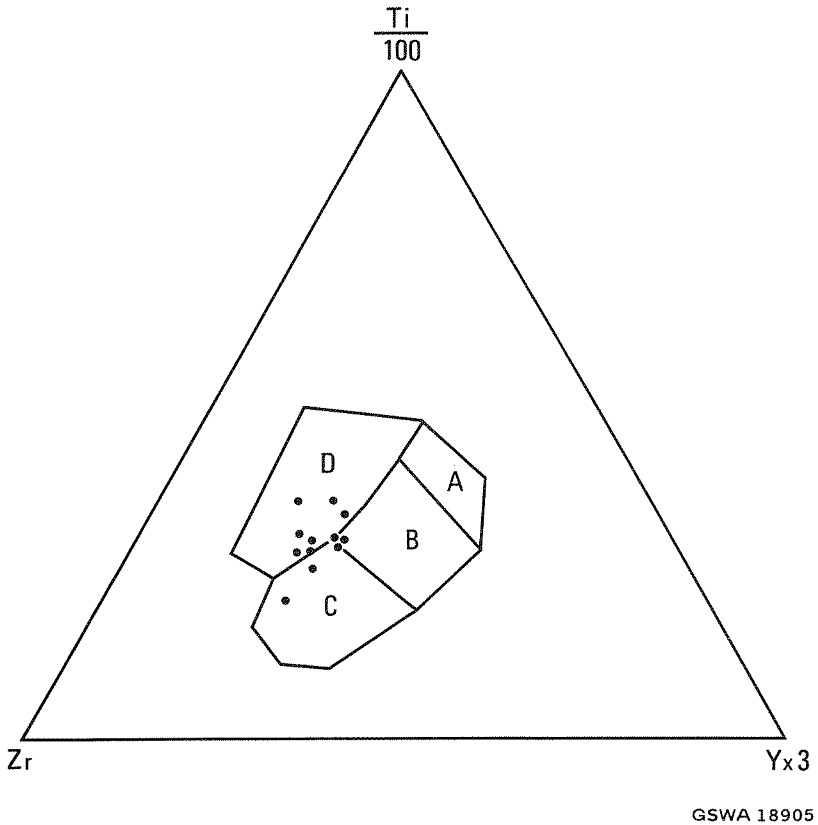


Figure 91. Diagram to discriminate between basic magma types using Ti, Zr, and Y (after Pearce and Cann 1973). Ratios for Bangemall dolerites plot in or near the field D which contain "within plate" basalts (ocean, island or continental basalts). Fields A, B, and C contain plate margin basalts (ocean-floor and volcanic arc basalts).

TABLE 27. CHEMICAL COMPOSITIONS AND C.I.P.W. NORMS OF BASALT TYPES AND OF BANGEMALL DOLERITES

Oxide (wt%)	1	2	3	4	5
SiO_2	51.57	46.53	50.8	48.0	56.6
TiO_2	0.80	2.28	1.59	1.94	1.56
Al_2O_3	15.91	14.31	14.0	13.3	14.2
Fe_2O_3	2.74	3.16	2.0	2.9	2.7
FeO	7.04	9.81	10.10	11.90	8.40
MnO	0.17	0.18	0.18	0.23	0.16
MgO	6.73	9.54	5.04	5.97	2.39
CaO	11.74	10.32	9.24	8.56	5.74
Na_2O	2.41	2.85	2.53	2.86	3.72
K_2O	0.44	0.84	1.26	0.57	1.66
P_2O_5	0.11	0.28	0.18	0.14	0.28
Rest (a)	0.45	0.14	1.74	3.57	2.44
TOTAL	100.11	100.24	98.7	99.9	99.9
Q	2.26	—	2.53	—	10.66
or	2.60	5.28	7.45	3.37	9.81
ab	20.39	20.04	21.41	24.20	31.47
an	31.29	23.63	23.13	21.77	17.15
ne	—	2.20	—	—	—
di	21.32	20.89	17.65	16.33	7.86
hy	16.05	—	18.37	19.76	12.89
ol	—	18.48	—	2.71	—
mt	3.97	4.53	2.9	4.20	3.91
il	1.52	4.41	3.02	3.68	2.96
ap	0.26	0.67	0.43	0.33	0.66
ca	—	—	0.11	0.05	0.07

1. Typical tholeiite (Carmichael and others, 1974, Table 2-1).
2. Typical alkaline olivine basalt (Carmichael and others, 1974, Table 2-1).
3. Typical dolerite, (No 37459), Bangemall Basin.
4. Dolerite poor in silica (No 50721) Bangemall Basin.
5. Dolerite rich in silica (No 37461) Bangemall Basin.
   (a) For Bangemall dolerites, the *rest* comprises H_2O^+, H_2O^- and CO_2. SO_3 was not determined.

Note: Analyses of 13 dolerites from the Bangemall Basin are given in Appendices F and G.

In the Cobra and Berala Synclines, all prominent veins occur close to, or within, dolerite sills and seem to be genetically related to them. Copper mineralization along northeasterly- trending shear zones in the Kumarina, Ilgarari, Bulloo Downs districts appears to be associated with dolerite in close proximity.

CHEMISTRY

The variation in chemistry of the Bangemall dolerites is small (Table 27) and is consistent with their uniform appearance in the field and the lack of evidence of differentiation. Comparison of the chemical composition and C.I.P.W. norms of typical tholeiitic and alkaline olivine basalts elsewhere in the world with those of the

Bangemall dolerites confirms that the latter are tholeiitic (Table 27). Normative quartz is present in all but one analysis, and normative hypersthene is plentiful, even in sample 50721 which contains olivine.

The tectonic setting of the Bangemall dolerites is clearly continental; they intruded a sedimentary sequence that lies on a basement forming part of an upper crustal layer, 10 to 15 km thick (Drummond, 1979), which appears to have behaved as a single plate since the Archaean (Embleton, 1978). A discrimination diagram (Fig. 91), using Ti, Zr and Y, shows that the Bangemall dolerites plot in the field of 'within plate' basalts (modern oceanic-island and continental basalts) rather than in the field of 'plate margin' basalts (volcanic-arc and ocean-floor basalts) (Pearce and Cann, 1971).

The Bangemall dolerites (Table 28) are similar to continental tholeiites in that they have high K_2O, a high Fe_2O_3:FeO ratio, and high Ba, Rb, Pb, Th and U compared with oceanic tholeittes. However, the K:Rb ratio and the amount of CaO, Sr and Zr in the Bangemall dolerites are not diagnostic of either oceanic or continental tholeiites and the level of P_2O_5 is more appropriate to an oceanic tholeiite suite. In summary, the chemistry of the Bangemall dolerites could be regarded as that of high-TiO_2 continental tholeiite.

TABLE 28. GEOCHEMICAL RANGES FOR OCEANIC AND CONTINENTAL THOLEIITES COMPARED WITH BANGEMALL DOLERITE

	Oceanic tholeiite	*Continental tholeiite*		
		Karoo	*Oenpelli*	*Bangemall*
K_2O (per cent)	0.08-0.57	0.6-0.95	0.38-1.06	0.42-1.93
TiO_2	1.26-2.03	0.45-1.29	0.95-2.66	1.17-2.97
P_2O_5	0.12-0.23	0.03-0.17	0.08-0.37	0.14-0.50
Fe_2O_3/FeO	0.25-0.36	0.02-0.88	0.09-0.33	0.18-0.81
K/Rb	230-1020	130-170	224-378	145-470
Ba ppm	5-16	16-360	149-305	180-1000
Rb	1.14-22	20-38	9-32	10-95
Sr	90-320	140-400	216-338	75-300
Pb	0.56-1.29	< 10	< 2-8	10-80
Th	0.13-0.15	80-160	< 2-40	<10-10
U	0.09-0.16	8	< 2-2	<1-2
Zr	45-160	85-90	46-170	105-420

Source of data:

Oceanic tholeiite—from Tables 8-1 and 8-2, columns 1-3 and 5-8, Carmichael and others (1974)
Karoo—from Table 9-2, columns 2-6, Carmichael and others (1974)
Oenpelli—from Table 2, Stuart-Smith and Ferguson (1978)
Note: Complete analyses of Bangemall dolerites are given in Appendices F and G.

DOLERITE DYKES

Dykes are of similar mineralogy to that of the dolerite sills. Most are 1 to 2 m wide, but the largest observed is 50 m wide and 15 km long. They are especially common in the Henry River region and in the Fords Creek Shale. Two generations of dykes exist. The earlier dykes are seen to be feeders to the sills, but later dykes transect the sills and folds in the sedimentary formations and post-date the main period of tectonism. The Henry River swarm, and at least some of the dykes in the Fords Creek Shale, belong to the later generation and trend between north and east-northeast.

There are dykes in some of the north-northeasterly lineaments of the eastern part of the basin, and most of these lineaments probably contain unexposed dykes. The longest line of dykes and lineaments stretches 160 km from Lake Disappointment southwards to STANLEY.

Contact effects are visible only within 0.3 m of where dykes intrude green shale. A darkening in colour, and hardening due to slight recrystallization are the only expressions of metamorphism.

BASIC VOLCANICS

Some altered basalt occurs within the Calyie Sandstone in the northwest of BULLEN (lat. 24°12′20″S, long. 120°04′30″E). The flow is approximately 4 m thick, has a sharp basal contact on pebbly quartz arenite, and is overlain by coarse-grained quartz arenite. The rock is light grey, has a trachytic texture formed by decussate laths about 0.5 mm long, and passes upwards into a fine-grained variety which contains amygdales of chalcedony near its top.

Although the original texture is well preserved, the primary minerals have been completely replaced. The laths, presumably after plagioclase, are filled by a colourless, almost opaque mineral, possibly chlorite or opal. The groundmass consists of brown devitrified glass. There are small scattered grains of leucoxene, and equant pseudomorphs (0.1 to 0.2 mm) of serpentine after pyroxene or olivine. The fine-grained chilled, vuggy top has a similar mineralogy, but rounded pseudomorphs after mafic minerals consist of chlorite as well as serpentine.

Unequivocal evidence of basic flows elsewhere in the basin is lacking. Some of the voluminous mafic rock in the Ullawarra Formation is fine grained but there is no evidence of extrusion except for a possible volcanic bomb (Daniels, 1969). If sizable quantities of lava had poured out into a sub-aqueous environment, pillow structure would have developed. Clearly, basalt eruptions were rare in Bangemall times.

SIGNIFICANCE OF MAFIC MAGMATISM

The isolated small basalt flow on BULLEN demonstrates that tholeiite intrusion had started while sedimentation was still going on. Vesicles in the upper part of some sills in the Ullawarra Formation imply a shallow depth of intrusion, probably less than the 3.5 km of overburden which eventually came to overlie the formation at the end of Bangemall Group deposition. Thus intrusion into the Ullawarra probably took place while sediments were being deposited higher in the sequence. Extensive intrusion into semi-consolidated sediments cannot be reconciled with the rarity of basalt flows if the sills were actively injected by a pressurized magma (because such pressure would need to have been greater than the weight of overlying sediments, and the magma would have had enough hydrostatic head to reach the surface). A passive, flotational intrusion must be inferred, in which dense, mafic magma flowed gravitationally under less dense sediments, so that the roof rocks effectively floated on the magma (Bradley, 1965).

Tholeiitic magmas are mantle-derived (Green and others, 1967), but form at shallower depths than alkali basalts. Basic magma compositions are determined by the depth of liquid segregation from a rising crystal-liquid mush in the case of tholeiites, less than 15 km. This depth in turn is related to the depth of original partial melting, which may be 50-200 km. The presence of a major tholeiite province where mantle-derived magma was passively emplaced at shallow depths in a subsiding, infilling basin must indicate a tensional regime in the crust during sedimentation. The distribution of sills (Fig. 89) shows that the greatest concentrations occur in the western part of the basin, and along the southern and northwestern margins of the wide eastern part of the basin. The pattern suggests an embryonic rift valley which bifurcates into two smaller arms east of a triple junction at the 120°E meridian, north of the Marymia Dome. The Bangemall Basin thus appears to have been formed by subsidence about a rift system, just as the Palaeozoic Michigan Basin and the modern Chad Basin have developed on the sites of rifts (Burke, 1976, Fig. 4).

FELSIC LAVAS

In recent years there have appeared, in both unpublished mineral exploration reports and published literature, references to acid volcanics in the Bangemall Basin. A review by Gee and others (1976) refers to these unsubstantiated occurrences. Daniels (1969) mentions sandstones, with a small tuffaceous content, thought to be present in the Kiangi Creek Formation on EDMUND, and acid volcanics near Mount Palgrave. Compston and Arriens (1968) dated an acid lava by the Rb-Sr method at approximately 1 080 m.y.; the location and petrology of the rock are not documented, but it is believed to be from a dyke and related flow in the Jillawarra Formation near Mount Palgrave. This occurrence was not located in the present study. Exploration company reports lodged with the Geological Survey, and Marshall (1968), state that there are lavas and tuffs in the Cobra, Geegin (or "Kurabuka"), Peedawarra, Berala ("Thumbla"), Isabella, Candolle and Mount Vernon Synclines on MOUNT PHILLIPS and MOUNT EGERTON at about the same stratigraphic level as the Mount Palgrave reported occurrence.

The presence or otherwise of felsic volcanics is important because of the implications they have in the search for base-metal mineralization. During regional mapping south of latitude 24°S therefore, special attention was given to identifying any rock which had even a remote chance of being volcanogenic. Thirty two speciments of apparent volcanogenic aspect were collected, all but two from west of the 120°E meridian. Only two definite occurrences of rhyolite and two probable tuffs were found, and these were all east of previously alleged volcanogenic areas. Twelve of the rocks could be either epiclastic or pyroclastic, although an epiclastic origin is favoured. The remainder were various cherty or argillaceous sedimentary rocks. The Discovery Chert (Chapter 5), also shows no evidence of a volcanic origin. It can therefore be concluded that felsic volcanics played a very minor role in the evolution of the Bangemall Basin.

The identified volcanogenic rocks, and some of the more unusual rocks of doubtful origin are discussed in the following section. The specimens were examined by petrologists J. R. Drake, W. G. Libby, and D. F. Blight.

RHYOLITE

The most significant rhyolite occurs 27 km southwest of Tangadee as an east-northeasterly trending line of six bun-shaped extrusions (Fig. 92, locality 1) 60 m above the base of the Coobarra Formation. The bodies have been described by Gee and others (1976), and occur within pebbly sandstone and conglomerate of the Coobarra Formation. They have concordant bases but disconformable walls and eroded tops.

The rock consists of euhedral quartz and feldspar phenocrysts in a cryptocrystalline groundmass of mainly quartz and microcline, with accessory biotite, chlorite, zircon, and opaques. The overwhelming predominance of quartz and microcline in the mode is confirmed by X-ray diffraction.

The quartz phenocrysts are hexagonal or rhomboidal in section, preserving the bipyramidal shape of β-quartz. Most phenocrysts are embayed by the groundmass, and many have become well rounded. Fragmentation of phenocrysts is rare, although many show healed cracks.

The feldspar phenocrysts now consist of granular intergrowths of quartz and microcline with small rosettes of chlorite and are euhedral with a monoclinic pseudohexagonal outline. These are considered to have been sanidine phenocrysts. Quartz and former sanidine phenocrysts amount to about 5 per cent of the rock, and commonly form glomeroporphyritic aggregates.

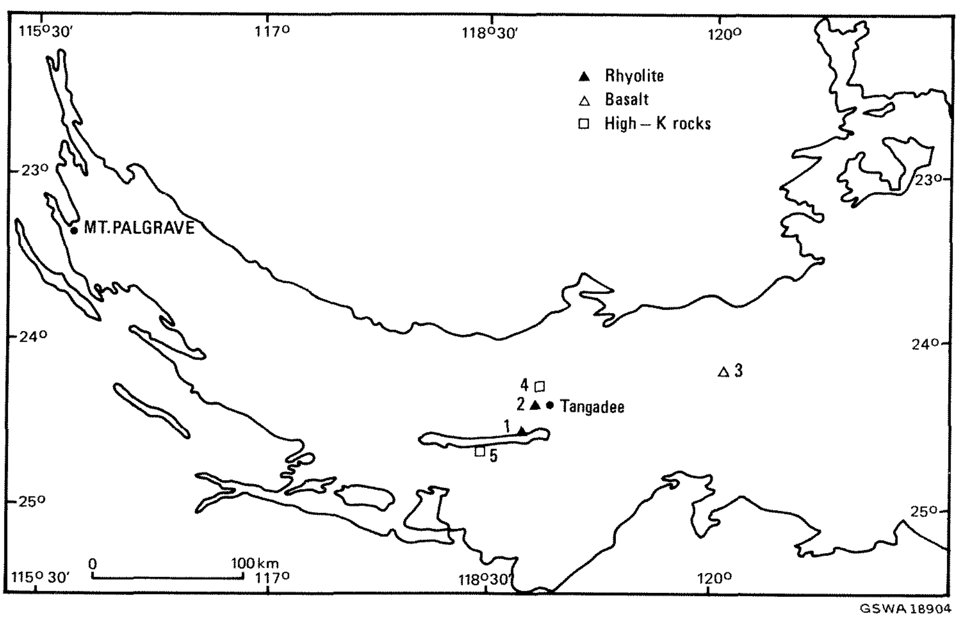


Figure 92. Localities of volcanogenic rocks.

Irregular or distincly sperical-shaped lithophysae, generally less than 2 mm in diameter, are sparsely distributed. These bodies generally contain quartz and microcline with a diffuse concentric structure, but also contain chlorite, carbonate, ilmenorutile, and sulphide. Dark square spots consist of poikilitic pyrite (now oxidized) intergrown with quartz, and this structure is encircled by concentric rings of a quartz and microcline mosaic. Fresh skeletal pyrite with intergrowths of quartz especially in the cores, is present. Lithophysae are wrapped by the flow banding, and therefore were in existence as solid objects while the lava was still flowing.

Despite the perfect preservation of the thin laminae that are conspicuous on polished faces; in thin section, the groundmass is a featureless cryptocrystalline

TABLE 29. ANALYSES OF RHYOLITES NEAR TANGADEE

Oxide (Wt %)	*1.*	*2.*	*3.*	*4.*	*5.*
SiO_3	78.0	76.1	74.9	73.66	71.34
Al_2O_3	11.1	12.3	12.3	13.45	12.64
Fe_2O_3	0.12	0.12	0.13	1.25	2.73
FeO	0.49	0.37	0.84	0.75	1.72
MgO	0.02	0.02	0.23	0.32	0.79
CaO	0.10	0.06	0.04	1.13	0.60
Na_2O	0.15	0.15	0.20	2.99	2.87
K_2O	9.06	9.21	10.45	5.35	6.13
H_2O^+	0.51	0.80	0.29	0.78	0.34
CO_2	0.02	0.02	0.02	....	< 0.01
TiO_2	< 0.01	< 0.01	< 0.01	0.22	0.31
P_2O_5	< 0.01	< 0.01	< 0.01	0.07	0.30
SO_3	0.04	0.03	0.03	....	....
Cr_2O_3	0.32	0.26	0.04	....	< 0.01
V_2O_5	< 0.01	< 0.01	< 0.01	....	0.01
MnO_2	< 0.01	< 0.01	< 0.01	0.03	0.10
TOTALS	99.97	99.57	99.87	....	....
Elements (ppm)					
Cu	45	100	110		
Li	< 5	< 5	< 5		
Mo	<10	<10	<10		
Pb	110	110	65		
Rb	150	160	180		
Sn	< 20	20	< 20		
Sr	20	20	20		
W	5	< 5	< 5		
Zn	17	16	15		

1, 2, 3: G.S.W.A. Samples 41839A-C, rhyolite near Tangadee, Analyst: Government Chemical Laboratories. Major oxides by classical methods, trace elements by atomic absorption and colorimetric methods.
4: Average calc-alkali rhyolite from Nockolds (1954).
5: Hilda Rhyolite of the Cassidy Group from the Warburton Range (Daniels, 1974).

mosaic probably resulting from the devitrification of glass. There is no evidence of shards or perlitic cracking.

These rocks were initially extruded as viscous, glassy lava that contained phenocrysts of β-quartz and sanidine. The origin of the flow banding is uncertain because the orginal geochemical and textural contrasts, if any, between laminae are not known.

The chemistry of the rock, as is emphasised by extremely high K_2O (up to 10.5 per cent) and extremely low Na_2O (about 0.15 per cent), is most unusual for a rhyolite (Table 29). Clearly the rock has been chemically reconstituted, but the timing and mechanism of this event are conjectural.

The rhyolite appears to have erupted along a fissure, forming a line of viscous domes that grew upward and outward as the earlier extruded rhyolite was shouldered aside by later extrusions and slumped laterally. The plugs were then subjected to active scouring that removed most of the associated debris. Based upon this model, the total volume of rhyolite extruded would not exceed 0.02 km^3. The amount removed by erosion is unknown, but could be of the same order of magnitude.

Another felsic volcanic rock, which was not found *in situ*, is worthy of description because of the rarity of such rocks in the Bangemall Group. It was a boulder found about 200 m north of Kellys Bore (Fig. 92, locality 2) in an area of no immediate outcrop where the surrounding rocks are shale, mudstone, and dolomite of the Devil Creek Formation, and dolerite sills.

The rock is an amygdaloidal, weakly porphyritic rhyolite. The amygdales are slightly elongate, up to 7 mm long, and have rims of zeolites and cores of crystalline carbonate. The zeolites have been tentatively identified optically as stilbite and heulandite; some chabazite could also occur. Altered potash feldspar phenocrysts have the form of sanidine, but they have been replaced by an intergrowth of quartz, microcline and chlorite. The groundmass shows no relict flow textures or structures, but consists of very fine-grained quartzofeldspathic material with minor carbonate, chlorite, and rutile. Spherulites, which indicate that the rock was originally glassy, are common in the groundmass.

ENIGMATIC ROCKS OF VOLCANOGENIC APPEARANCE

HIGH-POTASH ROCKS

Fine-grained highly potassic rocks were found at two localities. One of these is 12 km northwest of Tangadee (Fig. 92, locality 4) where greenish-white bands, 2 cm thick, occur in shale and siltstone of the Backdoor Formation. In thin section, fine-grained quartz and specks of clay are scattered through a cryptocrystalline matrix, which X-ray diffraction (XRD) analysis showed to be microcline with minor quartz and clay. Partial analysis of the rock by atomic absorption spectroscopy gave: K_2O, 9.0%; Na_2O, 1.9%, CaO, 5.8%; MgO, 0.6% (analyst, R. Davy). The CaO content is attributed mainly to calcite veins.

High-potash sedimentary rocks have been widely discussed in the literature, and a recent review of the subject was given by Davy (1975). Three origins have been proposed: dolomitization of adjacent illitic limestones, which released K to the shaly beds (Swett, 1968); diagenetic feldspathization of clayey sediments, involving ion exchange with seawater (Bowie and others, 1966); and alteration of glassy tuff beds (Weiss, 1954; Sheppard and Gude, 1969; Croxford, 1965).

The absence of dolomite associated with the beds north of Tangadee rules out the first theory. The second implies a higher K content in seawater in the late Precambrian than at present (Bowie and others, 1966) or other special conditions. The third option implies occasional brief ashfalls, whose pyroclastic textures were subsequently destroyed by feldspathization; with this option, the high K content is still a problem, but Battey (1955), Simons (1962), and Lipman (1965) have all noted that devitrified rhyolite glasses show K enrichment, although the mechanism is not known. The devitrified rhyolite from southwest of Tangadee also has a high content of potash (7-10.5 per cent K_2O). An origin by alteration of glassy tuff is therefore favoured.

Equally contentious is the nature of a rock, again within the Backdoor Formation, occupying a hill 14 km north of Mulgul (Fig. 92, locality 5). The rock is grey, bedded, laminated, and contains some 0.1 m-thick lenses of fine-grained

TABLE 30. ANALYSES OF POTASSIC ROCKS; 14 KM NORTH OF MULGUL

GSWA No.	(a) *39445*	*41691*	*41693*	(b) *39445*	(c) *Glauconite*
SiO_2	55.1	(c) 55.6	52.9	52.7	43-49
Al_2O_3	13.0	13.1	9.0	9.9	3-18
Fe_2O_3	1.1	0.7	1.1	6.7	6-25
FeO(d)	3.73	3.47	5.44	0.0	
MgO	7.3	(d) 5.97	(d) 8.95	11.1	3-4.5
CaO	8.37	9.38	13.85	13.4	0-1
Na_2O(d)	0.65	0.30	1.19	n.d.	0.1-0.2
K_2O	7.5	8.5	4.1	4.3	5.6-7.5
H_2O+(d)	1.78	1.78	2.01	n.d.	
H_2O- (d)	0.19	0.00	0.00	n.d.	
CO_2(d)	0.05	0.00	0.00	n.d.	
TiO_2	1.02	0.58	0.61	1.5	0.1-0.2
P_2O_5	0.22	0.09	0.12	0.2	
MnO	0.14	0.17	0.30	0.2	
TOTALS	100.2	99.7	99.5	100.0	

Analyses in per cent.
(a) Average of two analyses
(b) Heavy mineral separate
(c) Summary of analyses from Deer, Howie and Zussman (1962)
(d) Analysis by chemical methods
n.d. = not determined. All other analyses by X-ray fluorescent techniques. Analyses by W.A. Government Chemical Laboratories.

carbonate. Thin-section examination showed sparse poikiloblastic microcline in a cryptocrystalline mosaic of minerals too fine grained to identify optically. Analysis (XRD) disclosed that various samples contain from 5 to 50 per cent of poorly crystallized clinopyroxene, whereas most of the remaining portion is microcline; subordinate chlorite, epidote, plagioclase, and quartz are present. Chemical analysis of a heavy mineral separate (Table 30) indicates that the pyroxene is aluminous.

A clue to the origin of this unusual mineralogy is provided by the interlensed carbonate, which, under the microscope, is revealed as a fine-grained, equigranular, lepidoblastic rock composed of blades of colourless amphibole (probably tremolite) set in a matrix of calcite and subordinate quartz and epidote. This assemblage can be derived from low-grade metamorphism, or perhaps diagenesis, of a quartz-bearing dolomite by the reaction:

$$\text{Dolomite} + \text{quartz} + H_2O = \text{tremolite} + \text{calcite} + CO_2$$

(Winkler, 1976, p. 113).

A similar reaction at slightly higher temperatures:

$$\text{Dolomite} + \text{quartz} = \text{diopside} + CO_2$$

can generate clinopyroxene which could possibly be aluminous if clay were present. The low partial pressure of CO_2 that these reactions require to proceed at low temperatures (around 300°C) is realized in shallowly buried wet sediments. The high potash content, now expressed as microcline, could have been added as authigenic glauconite, or by the alteration of glassy tuffs. The high boron content in two samples (Table 31) may indicate metasomatism or an evaporite environment.

The simplest explanation for the formation of this rock is that tuff from a small, nearby volcanic vent was deposited in a carbonate environment and became enriched in potash during alteration; boron was added by hydrothermal action, and heat from the local igneous activity metamorphosed the deposit. The uniqueness of the rock in the Bangemall Basin would therefore be due to abnormal and local circumstances. The tuff would have been felsic, because the trace-element chemistry (Table 31) shows high Ce, La and Rb which are at normal levels for felsic rocks, while the Cr, Cu, Ni, Pb and Sc contents are too low for mafic rocks or even for admixtures of mafic tuff. The very high Ba content is not confined to this occurrence, but is present in the Jillawarra Bore region generally, and may be metasomatic like the boron.

The composition of the rock is also consistent with low-grade metamorphism of a glauconite-dolomite sediment. A postulated non-volcanic setting in which glauconite was forming could explain the high boron levels by evaporitic interludes, but it would not offer any intrinsic reason for the metamorphism; the heat source, which was not present elsewhere in the basin, would have to have fortuitously affected only this deposit of unusual chemical composition. However, glauconite can form from volcanic glass in seawater (Takahashi, 1939), and its formation could have been the step which added potash to the rock in the volcanogenic hypothesis.

In summary, high-K rocks in the Bangemall Basin are as rare as the lavas. The occurrence north of Tangadee is believed to have been laid down in a mud-depositing environment as felsic tuff from a non-local source, and subsequently

TABLE 31. TRACE ELEMENTS IN ROCKS 14 KM NORTH OF MULGUL

GSWA No.	*41691*	(a) *41692*	*41693*
Ag	<10	<10	<10
As	5	<5	<5
B	1 300	230	20
Ba	1 800	<20	1 600
Bi	<15	<15	<15
Cd	<10	<10	<10
Ce	230	60	40
Co	15	10	25
Cr	80	10	50
Cu	20	10	15
Ga	10	2	8
Ge	<10	<10	<10
La	130	30	20
Li	20	5	40
Mn	—	1 800	—
Mo	1	0.5	2
Ni	50	10	50
Pb	10	<10	<10
Rb	190	<10	100
Sb	<20	<20	<20
Sc	<5	<5	<5
Sn	10	35	<5
Sr	80	70	120
Te	<20	<20	<20
Th	25	<5	<5
Ti	3 100	1 000	3 400
U	<1	<1	<1
V	30	20	45
Y	60	30	20
Zn	100	30	65

Analysis by W.A. Government Chemical Laboratories. Results in ppm.

(a) Carbonate lens.

altered. The deposit north of Mulgul is best explained as the result of felsic tuff from a nearby vent falling into a carbonate setting and undergoing low-grade metamorphism after burial.

PROBABLE SEDIMENTARY ROCKS

A number of minor rock types encountered in the field are quite siliceous and have a streaky blotchy appearance, outwardly resembling felsic volcanic rocks. However petrographically there is no evidence for a volcanic affinity.

These rocks vary in colour, but are usually shades of grey or greenish-grey. All are fine grained, either completely aphanitic or consisting of small mineral grains or blebs (resembling phenocrysts) in aphanitic matrix. Lamination may be present and a streaky texture, visible under a hand lens, is common in both laminated and non-laminated varieties. A blotchy appearance, formed by small, irregular, light-coloured patches, is also common.

These rocks are composed of silt-size grains of quartz and accessories in a cryptocrystalline matrix. No evidence of shards or true phenocrysts is present, but the grains are too fine to determine whether the rocks are epiclastic or pyroclastic. Whole-rock X-ray diffraction traces of most samples show that the matrix is quartz with very small amounts of muscovite, kaolin and, rarely, goethite. Accessory amounts of tourmaline, opaque oxides, zircon and ?sphene are generally present. The well-rounded tourmaline, which is ubiquitous in definite epiclastic rocks from the Bangemall Basin, suggests an epiclastic origin for these rocks.

The structures and textures which, in hand specimen, suggest that the rocks are volcanic are not apparent in thin section. The white blebs resembling phenocrysts are shadowy white clay or cryptocrystalline quartz pseudomorphs, possibly after pyrite or gypsum crystals. The streaky texture, suggestive of lavas or ignimbrites, is caused by interlayering of diffuse lenticles, from 0.5 to 1.0 mm long, of clay and cryptocrystalline quartz; it is a diagenetic texture, also common in cherts.

Most of the investigated occurrences of possible acid lavas and tuffs reported by companies from MOUNT PHILLIPS and MOUNT EGERTON belong to this group of rocks.

CHAPTER 7

Structure

REGIONAL DEFORMATION PATTERNS

IDENTIFICATION OF TECTONIC UNITS

The Bangemall Basin can be subdivided into three tectonic units based on variation in the style and intensity of folding, as well as the trend of the axial traces of folds. These units are: the Edmund Fold Belt, a curvilinear area distinguished by zones of linear, tight folds; the Pingandy Shelf, an area of gently dipping sediments; and the Bullen Platform, a gently deformed area (Plate 2). These units are a revision of the structural provinces introduced by Brakel and Muhling (1976).

Folds in the Bangemall Group range from open drape-folds with a box profile to tight, upright folds with an approximately similar (Ramsay, 1968) profile and axial-plane cleavage; they have been classified according to size as regional (half wave length, 5-10 km), major (1-2 km), meso (hectametres) or minor, (metres to centimetres).

Each tectonic unit has been subdivided into structural domains, areas within which there is a consistent fold style or orientation of fold axes. Definition of the domains facilitates description of the regional fold patterns, and assists in the analysis of the structural behaviour of the basement.

BULLEN PLATFORM

Folds within the Bullen Platform are broad and open, have gentle dips, and axes that trend northeast or north-northeast. There are also areas of dome-and-basin folding which result from interference between northeasterly and northwesterly fold trends. The Bullen Platform has been subdivided into structural domains, each with its distinctive trend of fold axes and fold style (Plate 2). There is no evidence for superposition between folds of adjacent domains.

STRUCTURAL DOMAINS A-F

Domain A has broad folds which are related to dome-and-basin folds in the adjacent domain (M) and which are due to the combination of east to east-southeasterly trends typical of the Edmund Fold Belt and the northeasterly trend of much of the Bullen Platform (Plate 2).

Domain B is an area which dips gently off the northeast end of the Marymia Dome and grades north into an area of northeasterly trending linear folds. The consistent northerly dips in the southern area are not interpreted as a homoclinal structure in a thick sedimentary sequence, but as an original depositional dip of northwest-trending tidal current ridges on a flat platform.

The northeastern part of domain B extends as a narrow zone of linear folds to the northeastern boundary of the basin, and interrupts the dome-and-basin folds of domains E and F.

Domain C contains dome-and-basin folds whose long axes trend north-northeast.

Domain D is characterized by east-southeasterly trending folds.

Domain E is distinguished by dome-and-basin folds which result from the interaction of northeast- and northwest-trending folds. The northeasterly trend can be seen in isolated folds along the western edge of the domain. The northwest direction is linked with a basement ridge which is a continuation of the Cornelia Inlier. A significant feature is the rim syncline around the small inlier immediately northwest from the Cornelia Inlier. This suggests that basement along the ridge was pushed up during deformation.

Domain F contains north-northeasterly trending linear folds and subsidiary dome-and-basin folds close to the Boondawarri and Balfour Inliers. These basement inliers are in a zone which extends northwest into the basement, where Archaean granitic rocks also form arches under the Proterozoic Fortescue Group. The zone does not appear to form a major structural boundary between two structural units, but rather an interruption to the folds in domain F. A feature of the central part of this domain is a broad belt of lineaments, dolerite dykes, and faults that trend north-northeast.

ROBERTSON FAULT SYSTEM

The Robertson Fault System forms the boundary between the Calyie Sandstone to the east, and both the Manganese Subgroup and the Yeneena Group (contained in the Balfour Inlier) to the west (Plate 2). It separates the northeasterly trending linear folds of domain F from the more variable fold directions of domains I and J. Hence, the Robertson Fault System is interpreted to continue into the Tangadee Lineament. The fault is indicated by zones of brecciation, silicification and steep bedding. The amount of displacement is uncertain; however, it has cut folds in the Calyie Sandstone; and scattered remnants of basal Calyie Sandstone, which conformably overlie the Manganese Subgroup west of the fault, indicate the western block has moved up by an amount equivalent to the thickness of the Calyie Sandstone (at least 700 metres). This sense of movement is confirmed by the downward drag of bedding in sediments on the west side of the fault. The dip of the fault zone is probably near-vertical as shear zones and minor faults dip steeply to the east or west. The fault pattern (Plate 2) resembles that associated with normal block faulting.

PINGANDY SHELF

The Pingandy Shelf is a relatively undeformed apron of sediments flanking and overlying the Ashburton Fold Belt and the block-faulted and gently folded sediments around the Sylvania Dome and Hamersley Basin. The southern boundary of the Pingandy Shelf is taken as the appearance of the tight linear folds typical of the Edmund Fold Belt, or the appearance of the Mount Vernon—Lofty Range

Fault System. The eastern boundary is either the Robertson Fault System or the first appearance of consistent north-northeasterly trending folds east of the Tangadee Lineament.

STRUCTURAL DOMAINS G-J

Domain G is the arc of sediments that dips gently west and south off the Ashburton Fold Belt.

Domain H is characterized by southeast-trending open folds, block faults, and gentle bedding dips. It marks the swing of bedding trends from the easterly direction typical of domain G to a northeasterly trend. The northern boundary of the domain is defined by the Sylvania Dome.

Domain I has northeasterly trending linear folds, as well as some of dome-and-basin style. Bedding dips are gentle.

Domain J contains blocks of upfaulted basement and mostly east and east-northeasterly trending folds in the cover.

EDMUND FOLD BELT

The Edmund Fold Belt has open to tight folds with steeply dipping axial surfaces. Tight folds in elongate, discrete zones are the distinctive feature of the fold belt. Typically most folds are elongate, and doubly plunging with axes plunging less than 30°. Individual anticlines and synclines range from 20 to about 80 km in length.

STRUCTURAL DOMAINS K-M

Domain K extends from the edge of the Pingandy Shelf, which is the main synclinorial axis of the Bangemall Basin, to the southwestern margin of the Basin. The eastern boundary of the domain is the Tangadee Lineament. Fold axes trend in an arc from southeast to easterly, parallel to major lithological and structural trends in the Hamersley Basin and Ashburton Fold Belt.

Domain L has mostly tight folds with arcuate axial traces trending from northeast to east. The east and west boundaries of the domain are defined by the northeasterly trending Tangadee and Flint Hill Lineaments respectively.

Domain M occurs between the Tangadee Lineament and the Marymia Dome, and its northern boundary is the limit of zones of tight folds. The domain contains elongate dome-and-basin folds, which have resulted from the interaction of easterly fold trends (extending from the west) with the northeasterly trends of the Bullen Platform. Although many folds are open, their linear aspect and the presence of a few tight folds suggests this domain belongs to the Edmund Fold Belt.

ZONES OF OPEN AND TIGHT FOLDS

The Edmund Fold Belt contains alternating linear zones of open and tight folds, which are generally oriented parallel to the trend of the main fold axes (Plate 4). The zones of tight folds are concentrated in structural domains K and L, but some extend across the Tangadee Lineament into domain M where they form discrete areas of steep dips and cleavage within gently folded rocks. In the eastern part of the Edmund Fold Belt the zones of tight folds become narrow and pass into regional faults.

The zones of open and tight folds are evident in the overall cross-section (Plate 3) of the Bangemall Basin, which shows segments of broad, open folds separated by narrow zones of tighter folds, commonly accompanied by reverse faults. The characteristics of each type of fold are given in Table 32.

TABLE 32. CHARACTERISTICS OF TIGHT AND OPEN FOLDS

Tight	*Open*
Close to tight (dihedral angle 30°-70°)	Open to gentle (dihedral angle 70°-180°)
Mostly as complexes of major folds with no well-defined, single axial zone; rarely present as a single regional fold	Mostly a single regional fold outlined by long, gently arcuate trends with major folds subordinate
Single folds have a narrow axial zone with a rounded to cuspate profile	Single folds have a broad axial zone in which there are commonly two hinges forming a box fold
Dips generally steeper than 30°	Dips generally less than 30°
May have axial plane cleavage	No associated cleavage

On parts of EDMUND and MOUNT PHILLIPS, erosion has only partly removed the cover of Bangemall Group rocks, and, therefore, direct comparison of basement and cover structures can be made. Zones of tight folds are seen to be underlain, either by zones of medium to low-grade metasediments, or by shear zones in the basement. The Ti Tree Syncline mostly overlies the low-grade metasediments of the James Formation. A similar situation is apparent on WYLOO, where the Mount Minnie Group forms a zone along strike from the tight syncline of Bangemall Group in the Parry Range area. It seems likely that the incompetent metasediments would deform more readily than the surrounding granitoid, thus producing a discrete zone of tight folding in the cover. A different situation is apparent in the Mangaroon Syncline, where tightly folded Bangemall Group overlies granitoid, migmatite, and gneiss. The western margin of the syncline coincides with a zone of shear and retrogression in the basement. Here, a zone of weakness in the basement probably enables tight folding of the cover.

Zones of open folds in the cover, correlate with a basemant of granitoid rocks. An example is the open folds extending from Mount Padbury on EDMUND, southeastward to Mount Augustus.

Therefore, the mechanical properties of the rock in the basement have controlled the style of deformation in the cover.

REGIONAL LINEAMENTS

Regional lineaments are lines of photolineaments, joints, faults, dolerite dykes, and small folds, which cut across the basin. They coincide with changes in sedimentary facies, and so appear to be lines along which tectonism has influenced sedimentation as well as deformation. The Tangadee Lineament extends from the Egerton Inlier (Plate 2) northeast, via the Sylvania Dome, to the Balfour Inlier. Between the Egerton Inlier and the Sylvania Dome the lineament is defined by faults and major folds which trend northeasterly. It is also expressed by the change from open-style northeast to north-northeasterly folds typical of the Bullen Platform to linear, easterly trending folds of the Edmund Fold Belt. Interference between these trends has produced dome-and-basin folds and the western limit of influence of the northeasterly trend is at the lineament. Early control of the sedimentation along the lineament is shown by conglomeratic sediments flanking the eastern ends of the Egerton and Mulgul Inliers.

Northeast of the Sylvania Dome, the lineament is expressed by the Robertson Fault System and also the change from north-northeast-trending folds typical of domain F to the north-northeast to east-trending folds and dome-and-basin folds which occur with epeirogenic faults in domains I and J.

The Flint Hill Lineament marks the transition between northwest-trending folds of domain L to northeast-trending folds of domain M (Plate 2). More precisely, it is defined by a fault zone and quartz veins in the southwest and extends north-easterly into domain L, where it is marked by a prominent photo lineament and dolerite dyke. Tectonism has influenced sedimentation on the lineament to produce a sandstone facies directly beneath the Discovery Chert in the Lyons River Anticline.

There is no evidence of superposition of the fold trend of domain L upon that domain M, although the style of folding and the patchy development of axial-plane cleavage makes the age relationship between the fold directions inconclusive. The lack of interference structures and the retention of the linear fold styles and trends in each structural domain suggests they develop synchronously, but independently of each other. The Flint Hill Lineament is on line with a zone of divergence between two major structural directions of metamorphic foliations in the basement; the west-northwest trend typical on MOUNT PHILLIPS and the northeasterly trend on southeast GLENBURGH. These basement trends are parallel with, and along strike from, those in the Bangemall Basin. The lineament, therefore, is the expression in the cover of the zone of divergence in the basement.

During deformation of the basin, the lineament was the boundary between two basement zones upon which the cover was folded on trends following those in the basement.

Subordinate northeasterly trending lineaments are present on EDMUND and MOUNT PHILLIPS. They comprise zones in which the northwesterly trending folds terminate. The lineaments are also defined by northeasterly trending joints, faults and dolerite dykes. Facies change within the Bangemall sediments also occur along these lineaments in EDMUND (Daniels, 1969; Marshall, 1968).

REVERSE FAULT SYSTEM

The central part of the Bangemall Basin has been laterally shortened by a series of important reverse faults (Plates 2, 3) most of which trend east or east-southeast. A few subsidiary faults trend east-northeast. The largest faults are in the eastern section of the Edmund Fold Belt where the axes of regional folds trend easterly, and the faults die out to the west in the area where the fold axes swing to become northwest or southwest. All the important reverse faults are accompanied by a vertically dipping normal fault which occurs within a kilometre to the north of the main fault.

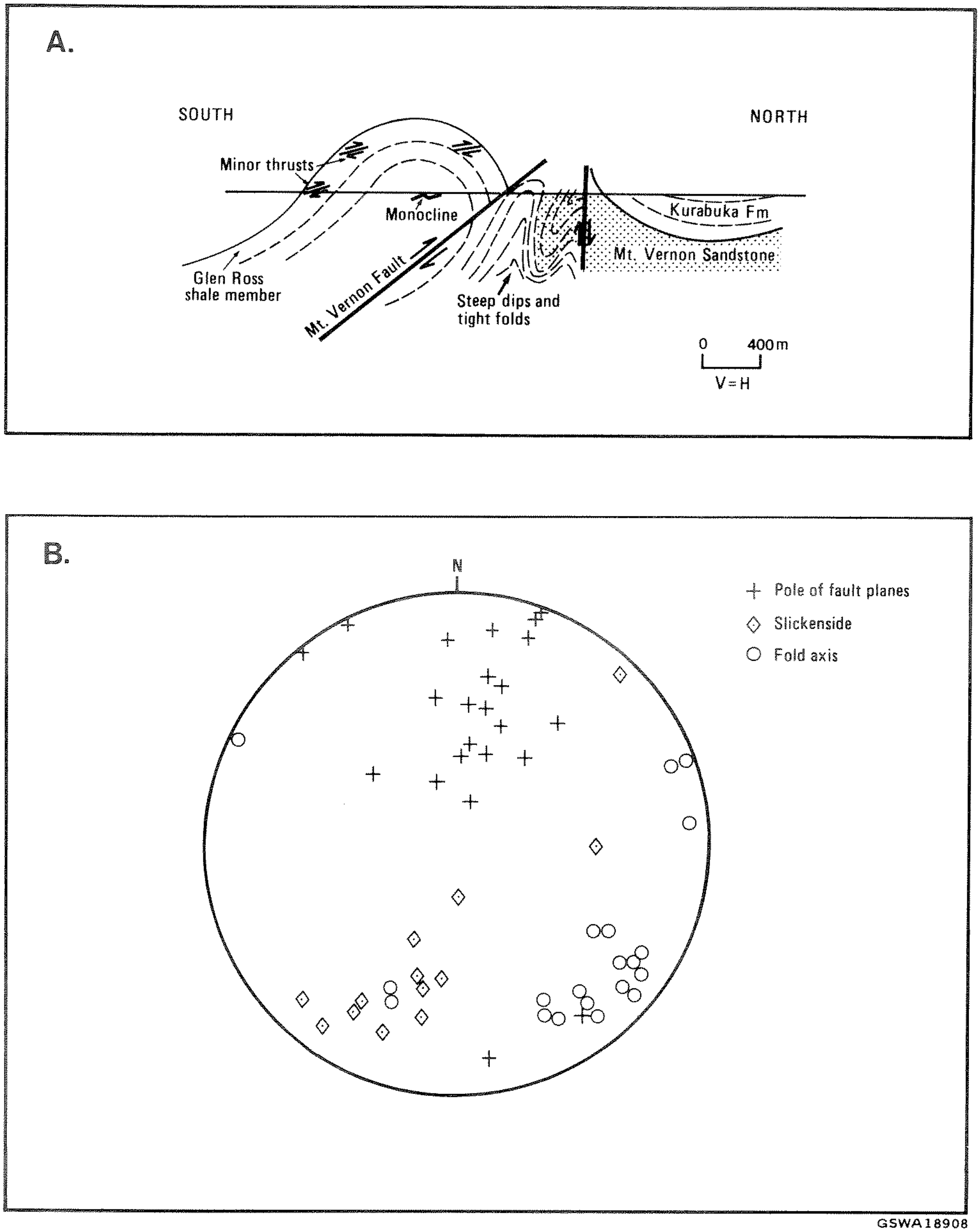


Figure 93. A—Cross section of the Mount Vernon Fault zone near Glen Ross Creek, MOUNT EGERTON (Busk construction south of fault). B—Orientation of minor thrusts, meso and minor folds, and slickensides on thrusts, Mount Vernon Fault zone, same area as cross section. All data from upthrust plate.

MOUNT VERNON—LOFTY RANGE

This fault system is a major east-trending tectonic zone; it forms part of the boundary between the Pingandy Shelf and the Edmund Fold Belt, and is at least 180 km long. The fault system consists of the Mount Vernon Fault, which extends from MOUNT EGERTON to COLLIER, and the Lofty Range Fault, which occurs 6 km to the north of the Mount Vernon Fault on COLLIER. The fault system dies out to the west in an anticline and is terminated in the east by the Neds Gap Fault, which trends northeast and forms part of the Tangadee Lineament.

The main reverse fault plane is not exposed, but is located by major breaks in the stratigraphy, by minor thrust faults, and by meso-scale asymmetric folds. The greatest displacement on the fault is in the west, where the Glen Ross Shale Member in the Kiangi Creek Formation is faulted against Fords Creek Shale, at least 3.4 km of vertical movement is represented. There is less displacement at the eastern end of the fault, where the Ilgarari Formation is against the Calyie Sandstone. Nevertheless, meso-scale folds in this area confirm the presence of the fault zone.

The main features of the Mount Vernon Fault are summarized in a cross section (Fig. 93A). The displacement of stratigraphy together with the orientation of the minor thrusts, meso-scale asymmetric folds and slickensides (Fig. 93B) show that the south block was thrust over the north block in a direction of 025°. However, the dip of the main fault is unlikely to be such a low angle (10°-20°) as the minor thrusts, because it would cut an extremely thick section of competent sandstone of the Kiangi Creek Formation (Plate 3). Therefore, a fault dipping 50° south and cutting directly across the sandstone is more probable. Such a fault, at depth, could either lie along the base of the sandstone or penetrate basement.

JEEAILA FAULT

The Jeeaila Fault forms part of the boundary between the Pingandy Shelf and the Edmund Fold Belt. It is not directly linked at the surface with the Mount Vernon-Lofty Range Fault, but is in an *en echelon* relationship, and they are probably connected at depth. The Jeeaila Fault strikes easterly and has brought gently folded Fords Creek Shale to overlie tightly folded Fords Creek Shale and Mount Vernon Sandstone. The Jeeaila Fault cuts out part of the northern limb of a regional anticline when the fold moved over the northern limb (Fig. 94). This sense of movement is confirmed in the highly folded zone in the footwall block at the eastern end of the main fault where minor thrust faults and axial planes of minor asymmetric folds dip south (Fig. 95).

The eastern end of the fault comprises a zone of brecciated and folded Fords Creek Shale which dips between 30° and 50° south. The brittle style of deformation in the zone is shown by the breccia (Fig. 96) and the angular disharmonic style of minor folds in the zone (Fig. 95). The orientation of minor thrusts, slickensides and axes of minor folds show that the direction of movement at the west end of the fault was northerly, whereas at the east end it was northeasterly (Fig. 94). Hence the overall direction of movement was about north-northeasterly.

Splay faults are common in the eastern end of the overthrust block (Fig. 94), where they are expressed as slickensided, siliceous breccia zones (Fig. 96) that trend northeast and dip southeast. The slickensides indicate strike-slip movement on some

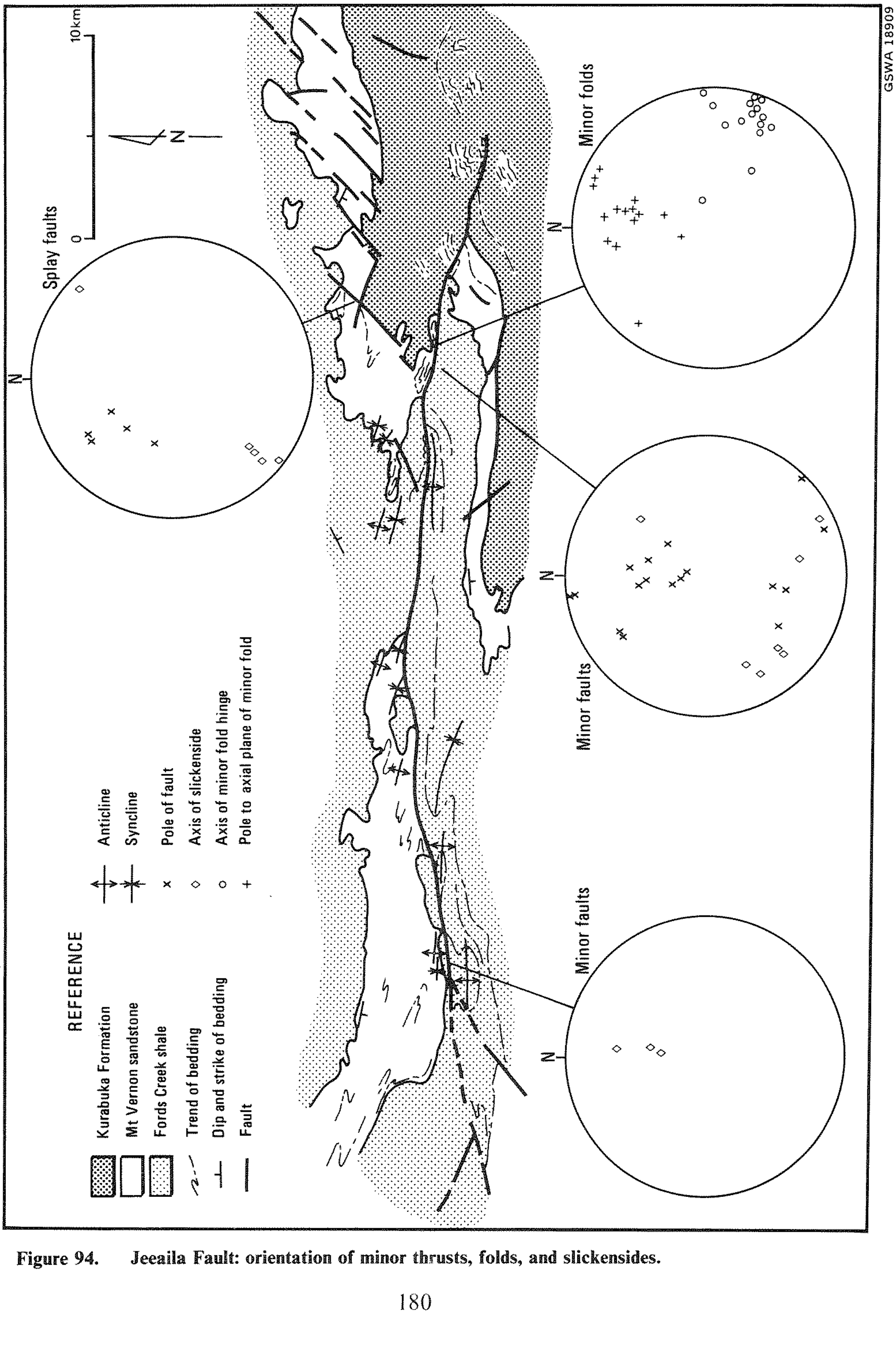
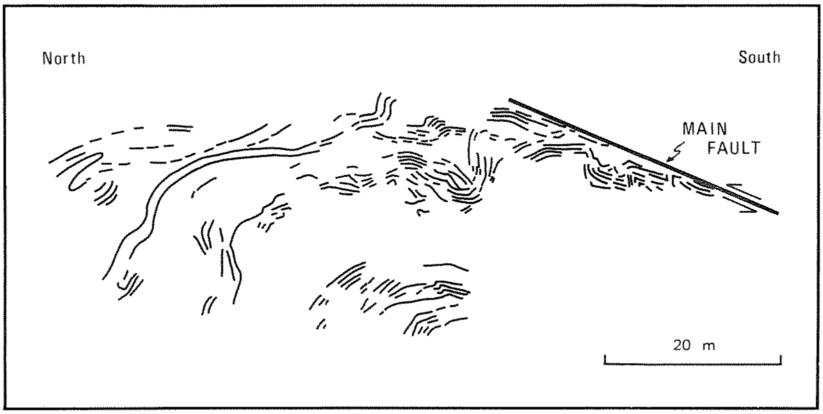


Figure 94. Jeeaila Fault: orientation of minor thrusts, folds, and slickensides.

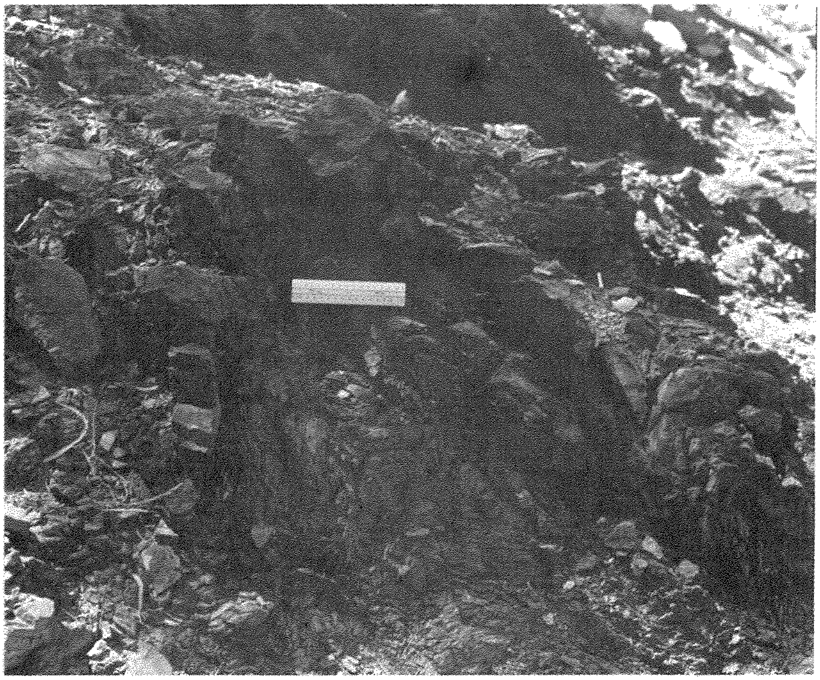
splay faults, and displacements were probably less than 100 m. Secondary splay faults trend parallel to the main reverse fault and dip south.

QUARTZITE WELL FAULT

The Quartzite Well Fault forms the northern boundary of the Mulgul Inlier against Bangemall Group, and provides evidence of participation of the basement in



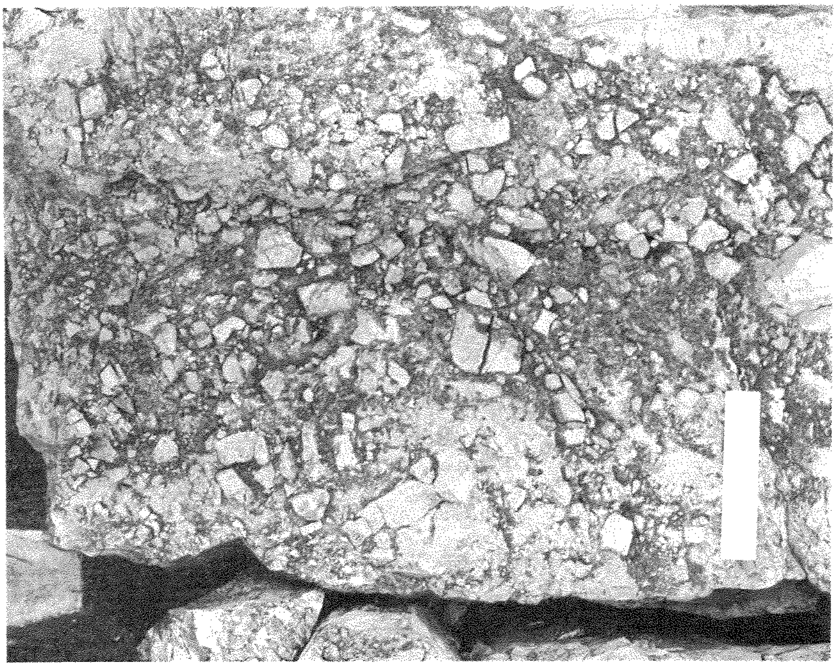
A



B

G.S.W.A. 18910

Figure 95. Structural style in the overthrust block, Jeeaila Fault. A—Cross section of footwall, Gorge Creek West, showing disharmonic box folds and asymmetrical folds with south-dipping axial planes. Trends of bedding planes traced from photographs. Section is 100 m long. B—Disharmonic folds in interbedded sandstone and shale. Looking east, axial plane dips south.



G.S.W.A. 18911

Figure 96. Siliceous breccia in splay-fault zone in overthrust block, Jeeaila Fault.

faulting. Shale, dolomite, sandstone, slate and phyllite possibly belonging to the Glengarry Group, have been brought against rocks of the Kiangi Creek-Jillawarra-Discovery Chert-Devil Creek Dolomite interval. The fault has an arcuate trace, convex to the south. At the crest of the arc, there is a single, narrow, fault zone; however, further west, the fault zone shows the same structural style as the other reverse faults and two faults are present. The main fault zone dips 40°-60° south near Quartzite Well. The minimum displacement along the fault plane is 3.4 km (Plate 3). Slickensides, minor asymmetrical folds, and thrusts, indicate a direction of movement from the south-southwest.

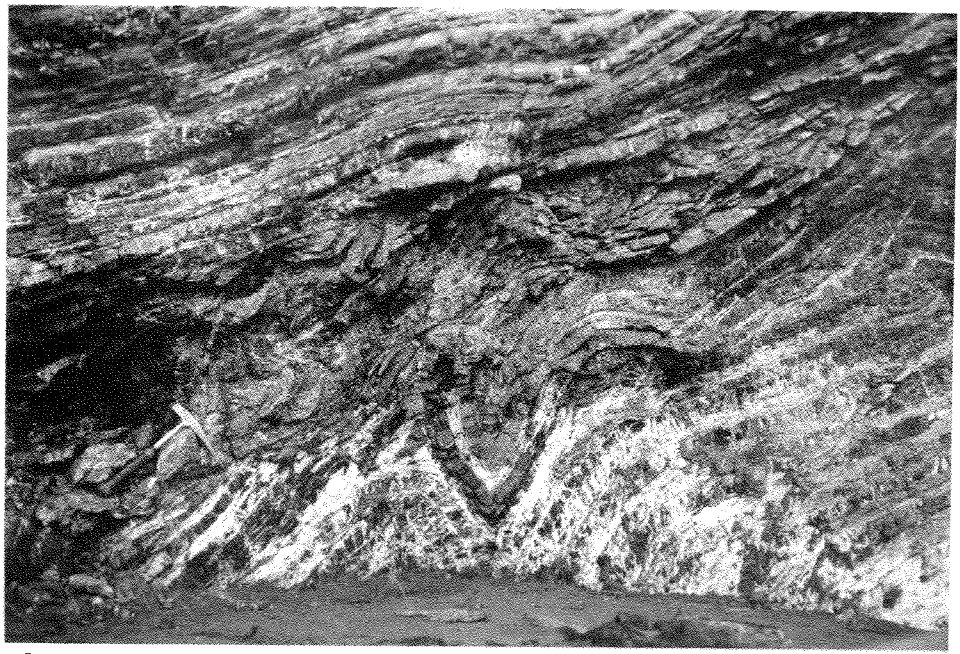
FOLDING OF FAULTS

Minor thrust faults accompanying the main Mount Vernon Fault have been folded. For example, in Figure 97, minor thrust faults separate three zones of differential deformation. Axial planes of box folds and isoclines in the lowest zone have been folded together with the overlying thrust to produce open folds which are evident in the middle structural zone, but not in the top zone where the top thrust is unfolded. This sequence of formation and deformation of folds and faults indicates there was continued activity on the Mount Vernon Fault.

FAULTING ON THE SOUTHERN MARGIN

Much of the southern boundary of the Bangemall Basin, especially on ROBINSON RANGE, STANLEY and NABBERU is marked by faults. The evidence for faulting is:

(1) Truncation of structure in the Bangemall Group along the boundary (Plate 2).



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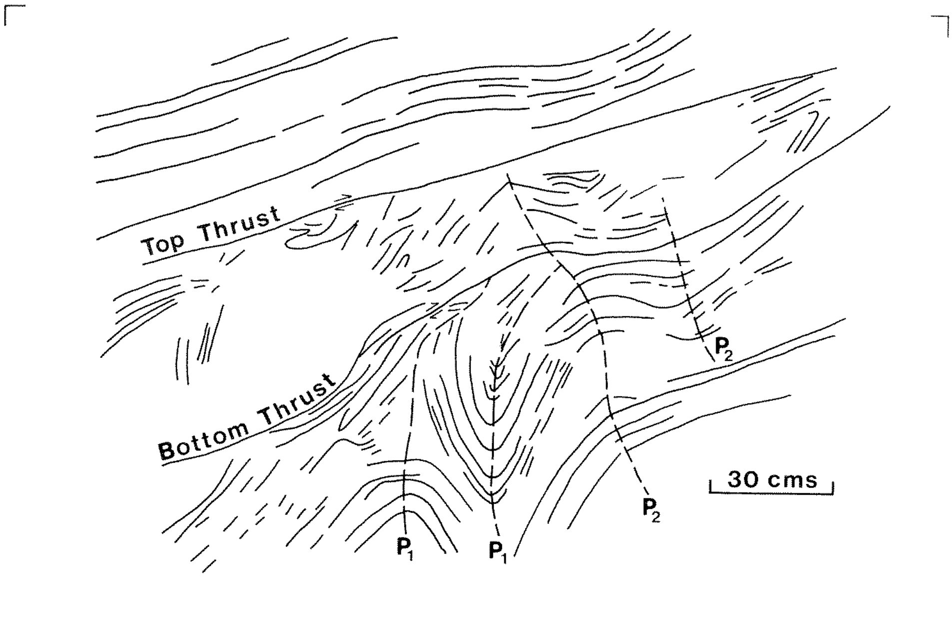


B

G.S.W.A. 18912

Figure 97. A—Folded minor thrusts in interbedded chert and shale from the hanging wall, Mount Vernon Fault, MOUNT EGERTON. B—Sketch from A showing traces of early (P_1) and later (P_2) fold axes.

(2) Truncation of Bangemall stratigraphic units high in the sequence against basement (Plate 1).

(3) Truncation of aeromagnetic patterns of the basement at the boundary of the Bangemall Group. Where there is no fault, basement aeromagnetic patterns extend beneath the cover sequence.

(4) Regional facing of the Bangemall sequence into the basement. This is characteristic on ROBINSON RANGE, where the sequence dips and faces south, off the Hibernian Anticline toward the basement.

Most of these faults dip steeply into the basement, and clearly, the basement has moved up and over the cover rocks. The faults bounding the southern edge of the Bangemall Basin lie in comparatively long segments trending east-southeast to easterly, or in short segments trending northeast. On ROBINSON RANGE, east-southeast-trending faults have sliced the southern fringe of the Bangemall Group into a series of narrow blocks, within which the bedding is crumpled into asymmetrical minor folds. Each block has a degree of deformation different from its neighbour. The actual boundary fault is exposed west from the Coolinbar Syncline, where stromatolitic dolomite dips steeply south and faces into granitoid from which it is separated by a quartz-veined shear zone dipping about 70° south.

On the southern margin on NABBERU and STANLEY, the Scorpion Group has been uplifted by an east-southeast trending fault against the Bangemall Group; however, the dip of the fault is not known.

One northeasterly trending fault zone is exposed at the northeast end of the Marymia Dome (Plate 2), where granitic and metamorphic rocks have been thrust to the southeast over crumpled Bangemall Group along a fault that dips 60° north. Although this is a different direction of thrusting from that in the other southern boundary faults, the style of movement is similar, in that basement has been thrust over the cover.

The western boundary also shows evidence for movement of the basement over the Bangemall rocks. Northeast from Maroonah (EDMUND), the base of the Bangemall Group is marked by tight, overturned, similar-style minor folds, whose slaty cleavage dips 60° to 80° southwest. Small-scale thrusting and local overturning are also mentioned by Daniels (1969).

RELATION OF REVERSE FAULTS TO REGIONAL FOLDING

The reverse faults are parallel to, and located adjacent to, tight regional folds; however, the evidence is that the faults generally formed in the cover after the folding. The Mount Vernon and Jeeaila Faults truncate regional folds trending 110° and 120°, and a splay fault from the Mount Vernon Fault cuts the Mount Vernon Syncline. The Quartzite Well Fault has also cut the south limb of the Range Creek Syncline. An exception to this general relationship is where the Lofty Range Fault curves around the nose of the Brumby Creek Anticline and has participated in the folding.

Folds associated with the thrust faults are asymmetric, disharmonic, and box-like in profile, with a general brittle style. All these features contrast with the more ductile style of regional symmetrical folding. Some of the minor symmetrical folds have been redeformed by retightening and the development of break-thrusts due to the thrust deformation, e.g. Fig. 98.

The contrast in styles of deformation could be caused by progressive dewatering of the sediments during the early stages of deformation; higher rates of strain and shallower depths of folding during the later deformation. Most tight

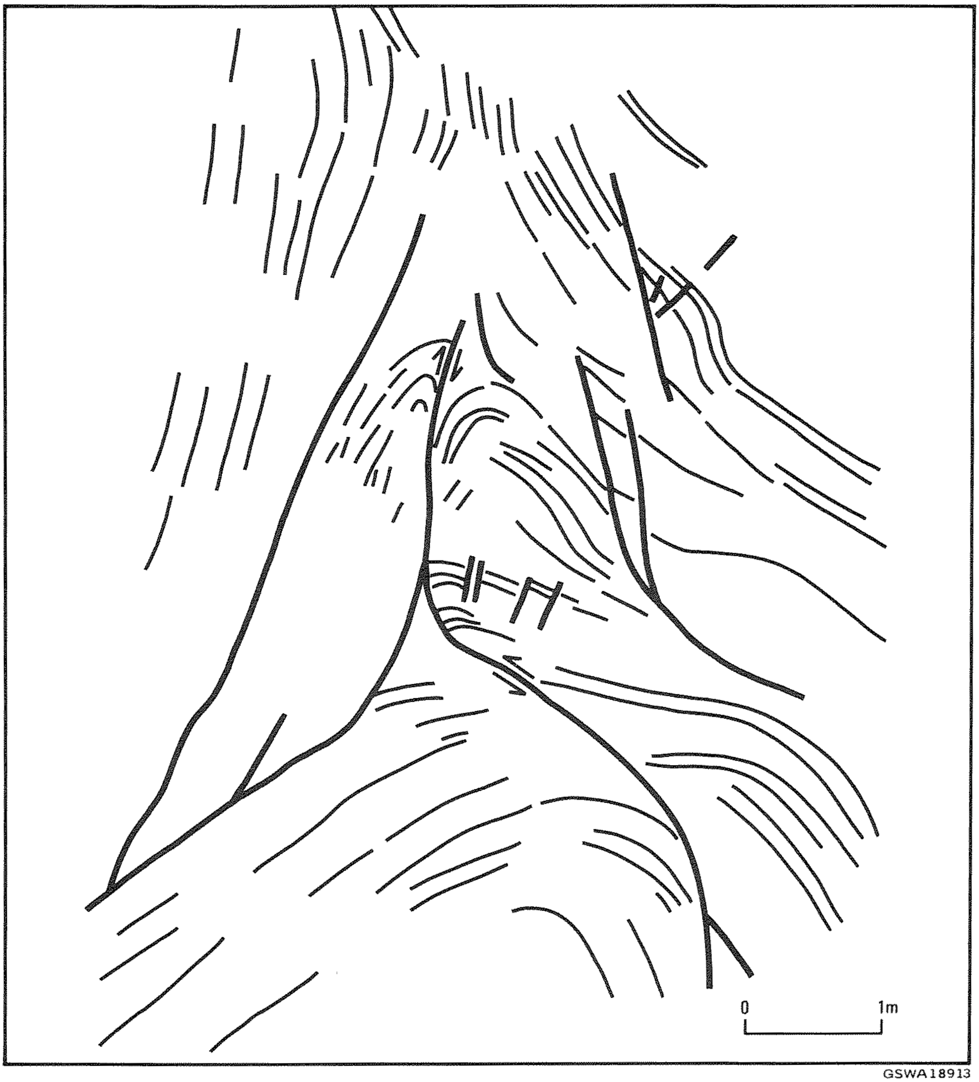


Figure 98. **The disharmony between the upper and lower part of the axial zone of the anticline indicates later movement on bedding planes and faults to produce a peak in the crest of the anticline. This has effectively shortened a pre-existing anticline enclosed between the Mount Lofty and the Mount Vernon Faults. The diagram is traced from a photograph approximately parallel to the axis of an anticline in shale of the Jillawarra Formation, Brumby Creek, COLLIER.**

regional folds are in grabens and exhibit a good axial-plane slaty cleavage, which is absent in the thrust fault structures. This suggests the regional folds formed at greater confining pressure (and therefore greater depth) than the faults.

The reverse faults have come up from the basement and probably reflect boundaries between segments of different lithology or old faults in the basement. Significant movement on the reverse faults only extended into the cover during the last stages of deformation and uplift after the regional folds had formed.

ORIGIN OF REVERSE FAULTS

Four features are important in the consideration of the origin of the reverse faults. These are:

(1) A close spatial relation between zones of tight regional folding and major reverse faults (Plate 4).

(2) With the exception of the southern boundary faults, all the major reverse faults are accompanied by a normal fault immediately to the north (e.g. Fig. 93). The normal fault does not fit the strain pattern related to the reverse fault, but does mark the limit of the deformation associated with the main fault.

(3) The Quartzite Well Fault has cut basement, and as the other major reverse faults show the same features, they too can be presumed to cut basement.

(4) The style of folds in the cover is controlled by basement lithology.

The reverse faults are interpreted to have formed as follows:

(1) The normal fault formed early in the deformation of the basin, at a boundary between a relatively competent basement block to the north and an incompetent block to the south.

(2) During lateral shortening of the basin, the incompetent basement block and its cover of Bangemall Group were shortened by folding.

(3) With further shortening, the boundary between basement blocks was reactivated and became the site for reverse faults. The incompetent part of the basement and the cover were pushed against and overrode the northern block, to the north, which acted as a buttress.

STRUCTURAL CROSS-SECTION

A detailed structural traverse was made along the line of section shown in Plate 2 with the object of constructing an accurate cross-section of the Edmund Fold Belt and Pingandy Shelf. Several methods have been proposed for constructing fold profiles at depth—most of these follow Busk (1929). The Buskian and related methods all presume parallel (or concentric) fold geometry and a constant arc-length of beds during folding. It is easily shown that constructions based on this geometry always led to room problems in cores of anticlines (Ragan, 1973). Carey (1962) has argued that parallel (or concentric) folding demands detachment of beds at lower structural levels, and further implies considerable lateral transport of sedimentary piles. The question of decollement therefore arises in any deformed sedimentary basin and this aspect is discussed in a later section of this chapter. Buskian sections are commonly used to estimate the amount of lateral shortening and this question is also discussed later in this chapter.

Within the limits of known stratigraphic thicknesses, it is not possible to construct a section based on the Buskian tangent-arc method, because of the rapid changes in style across strike and the alternation of zones of open and tight folding. It is clear both from field observations of fold styles and construction of this cross-section that fold profiles do not correspond to arcs of circles.

In the zones of open folds, the method that Hills (1963, p.232) describes as the method of tangents produces mega kink-folds with similar-style geometry across each inflection. Fold inflections are arbitrarily rounded to appear realistic, and the profiles are terminated at depth, at postulated normal faults which are known to occur in places at the unconformity.

In the zones of tight folding, profiles are constructed using axial projections along the plunge of outcrop data off the section. This enables some account to be taken of a flattening component of buckle-style folds. Where sufficient information does not exist the fold profiles are continued to basement using the Buskian tangent-arc method and the inferred thickness of the buried stratigraphic units.

These constructions were made at a scale of 1:50 000 and presented here (Plate 3) at a scale of 1:250 000. This reduction renders the cross-section diagrammatic, but is still a useful reference for discussion of the regional structure.

BASEMENT-COVER RELATIONSHIPS

The sedimentary pile in the Edmund Fold Belt forms a blanket 200 000 km^2 in area, and 10 km thick, resting on a basement of granitoid, gneiss and metasedimentary rocks. It has been deformed, quite strongly in places, and thus presents the question—what is the deformation mechanism of the cover, and how does it relate to movement in the basement? It was noted earlier that certain geometrical considerations require decollement at depth and imply lateral transport of upper strata over older rocks.

DECOLLEMENT OR CRATONIC DEFORMATION

Deformation in the cover may be the superstructural expression of bulk deformation of the crust in an orogenic zone. However, this model cannot apply to the Bangemall Group which shows negligible metamorphism. Structures in the basement, which can be related to deformation in the cover, are non-metamorphic and brittle in style, and the basement appears to have behaved as a cold, rigid block. The concept of decollement absolves the need for basement participation.

The classic area for decollement is the Jura Mountains where Mesozoic and Tertiary sediments, initially unconformably overlying a Palaeozoic crystalline basement were able to deform into box folds of parallel-style geometry by movement along a highly ductile evaporite layer. The concept of decollement has been widely applied to fold belts, especially those called "shallow fold and thrust belts" (Hobbs and others, 1976). A practical difficulty with this model is the difficulty of identification of high-strain zones, and the distinguishing of them from local features unrelated to bulk strain. A theoretical difficulty is the shear-strength of a thin sedimentary blanket, which would be too low to transmit stress across a basin the size of the Bangemall Basin.

To some extent this theoretical problem can be overcome by invoking gravity sliding or gravity creeping (i.e. down-slope movement under the influence of gravity). This can produce a variety of structures such as gravity-collapse parasitic folds, foreland thrust belts, gently folded sedimentary sheets, recumbent fold and nappes, all of which require a decollement surface.

In the Bangemall Basin, there is no evidence for deformation by decollement. There is no suitably incompetent gliding layer anywhere in the sequence, and there are no known examples of duplication of stratigraphy by thrusts. Although the thrust-fault systems are consistently orientated, there is no overall asymmetry to the folded basin, and no significant overturning of the strata. In fact, some of these faults cut the basement at a high angle.

BASEMENT DEFORMATION

Basement control of cover deformation is well documented in the Rocky Mountain foreland in Wyoming and in the Colorado Plateau, where movements of buried fault blocks have produced drape folds in the cover. A drape fold is one in which the geometry in relatively ductile sedimentary strata is determined by the differential uplift of underlying brittle rocks of structural basement (Matthews, 1978). The resultant folds are commonly asymmetrical, with the steeply dipping limb draped over the faulted edge of a basement block. Minor structures in the steeper limb of the folds indicate thinning and extension have taken place, either by faulting or by flow. Such structures have been described (Stearns, 1978; Reches, 1978) and reproduced in experimental work (Logan, Friedman and Stearns, 1978). The participation of basement blocks in these folds have been determined by drilling, geophysics and field mapping (e.g. Matthews and Work, 1978; Borg, 1962 *in* Badgley, 1965). The attitude of the faults is variable at the surface and ranges from low-angle thrusts to high-angle faults. Many cross-sections show that faults steepen at depth, and most authors agree that there were substantial vertical movements.

Unlike the inferential evidence for gravity sliding, there are clear criteria which can be applied to cratonic deformation models: for example, where cover and basement are seen to be folded together, or where structures in basement are continuous with those in the cover. Unfortunately, in the Bangemall Basin, such evidence is uncommon because good exposure of the unconformity between cover and basement is rare. Indirect evidence, such as folds in the cover with cores of basement, and parallelism of basement and cover structures, implies participation of basement in folding of the cover.

There are four points which considered together demonstrate cover-and-basement deformation.

(1) The axial trends of regional and major folds in the Bangemall Basin are parallel to structure in the basement. For example, fold axes and cleavages in the northwest part of the Edmund Fold Belt trend 130°, parallel to trends of

metamorphic and granitic rocks in the nearby basement. Further south, this trend swings west-northwest and meets the Flint Hill Lineament, which is not only a boundary between west-northwest and northeast-trending folds in the cover, but also lies close to the boundary between those same trends in the basement. The same relationship applies to subsidiary structural trends. In the basement beneath the Ti Tree Syncline on MOUNT PHILLIPS (Williams and others, 1979), subsidiary structural trends of 110° and 045° have overprinted the main trend of 130°. The Ti Tree Syncline shows a similar sequence of overprinting of structures on these same trends: an early direction of tight, similar style folding (Fig. 110) and axial-plane cleavage trending 130°, crenulated by cleavages trending from 50° to 110°.

(2) Boundaries of structural domains in the cover also mark boundaries of basement segments. For example, the Jeeaila River Fault, together with the western end of the Mount Vernon Fault (and its west-northwest splay fault) accurately define the boundary between zones of open and tight folds. The cross-section (Plate 3) shows these zones of tight folds bounded by faults, most of which probably penetrate basement. Hence tight folds are located either alongside faults which mark the edge of basement blocks, or over basement segments bounded by faults.

(3) Basement and cover are observed to be folded together in two regional anticlines, the Sullivan and Coolina Anticlines. Both anticlines have a core of basement metasediments with an envelope of Bangemall Group rocks. Bedding planes in basement and cover are parallel with each other (Fig. 99).

(4) There are discrete zones of high strain in the cover, expressed by zones of steep cleavage in open folds, zones of major folds which trend across regional folds, and the zones of tight folding. Examples of this are:

    (a) The regional folds, Range Creek Syncline and Glen Ross Anticline, trend easterly, but smaller major folds on their limbs trend 110° (Fig. 100). Some of the smaller folds, although very open, display a weak axial-plane

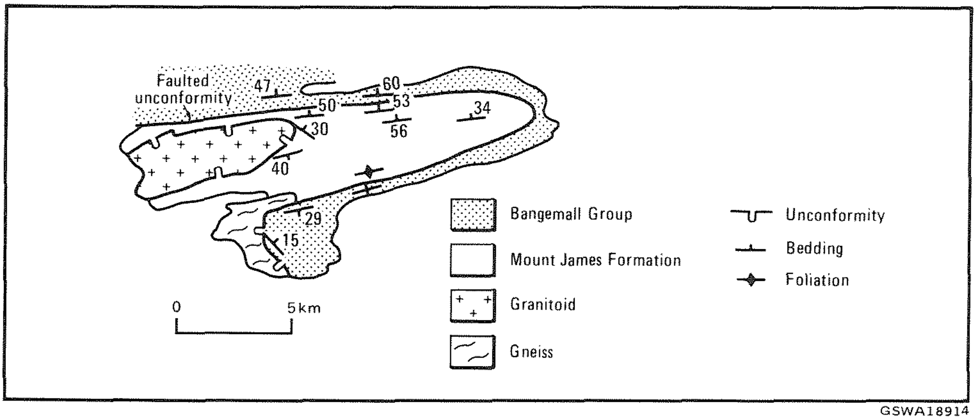


Figure 99. **Sullivan Anticline, MOUNT EGERTON. Bedding in basement anticline is parallel to that in the cover of Bangemall sediments indicating that the basement and the cover have been folded together.**

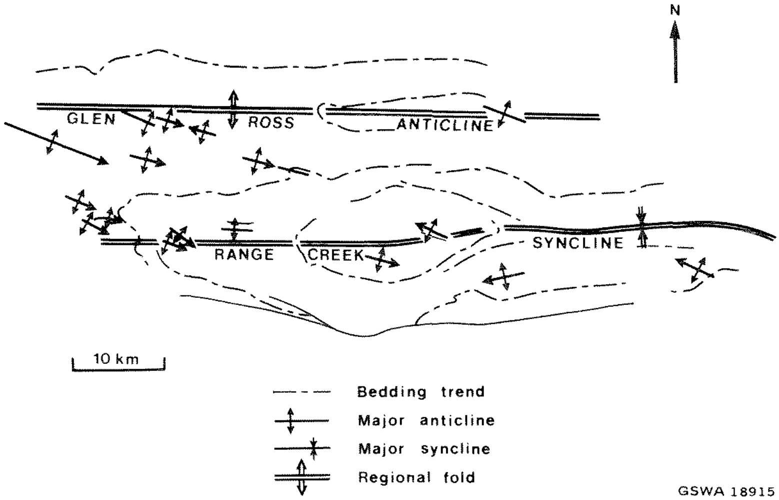


Figure 100. **The axial traces of the major folds trend mostly northwest across the regional folds and appear to be independent of them.**

cleavage, which even at the microscopic scale has a spaced nature (Fig. 101). Cleavage is not a feature of the broad regional folds. These major folds have a trend and symmetry which does not relate to the easterly trending regional folds, and it appears that the major folds have been imposed after formation of the regional folds.

(b) Discrete zones of tight regional folds are enclosed between open folds, particularly in domains L and M, in which dips are generally low. The style of folding is not governed by the mechanical properties of the rocks.

A

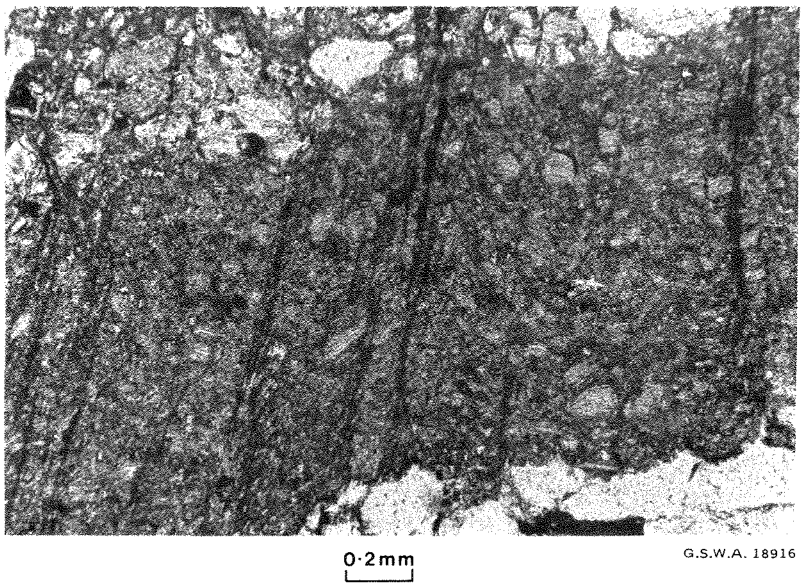


Figure 101. **Photomicrograph of spaced cleavage cutting interlaminated sandstone and siltstone in the core of a major fold in the Kiangi Creek Formation, Glen Ross Anticline, MOUNT EGERTON.**

For example, comparatively incompetent shale is present in both the open Range Creek Syncline and the tight Hells Doorway Syncline; competent sandstone can be traced from the Range Creek Syncline to the tight Coolina Anticline.

The work of Hubbert and Rubey (1959) has shown that a cover sequence in compression would have insufficient shear strength to transmit stress across the basin. According to the current elastic-viscous theories for buckle folds, zones of tight folds near the basin edge should first close before stress could be transmitted to the centre of the basin—yet zones of tight folds occur in the centre of the basin. It is also difficult to imagine the production of successive zones of open and tight folds by lateral compression acting independently from the basement. Therefore, zones of high strain in the cover resulted from discrete zones of stress, which would have to be transmitted upward from the basement. This conclusion is consistent with structures and fold styles of the Edmund Fold Belt which are described in the following section and show how the cover has adjusted to basement movements.

FOLD STYLES OF THE EDMUND FOLD BELT

ZONES OF OPEN FOLDS.

The MOUNT EGERTON cross section (Plate 3) shows that open folds form either as a broad box syncline or a narrow, rounded anticline. Most folds are upright and symmetrical. The only asymmetrical folds are the Mount Arapiles Anticline and the Watts Well Syncline, both of which are at the southern side of the section and have axial planes dipping steeply south.

MINOR STRUCTURES

Minor structures in the open folds, in particular the Glen Ross Anticline and Range Creek Syncline, show the behaviour of the Bangemall sediments during movements of the basement and indicate how the open folds developed. The structures are described in their approximate order of formation.

Slump folds are present in shale, siltstone and sandstone (Fig. 103). They are small-scale, asymmetrical to recumbent folds in which laminae vary considerably in thickness. Their sense of movement is down the limbs of the Glen Ross Anticline and indicates that the anticline crest was a structural high early in the deformation of the Bangemall Basin.

Disharmonic zones are stratabound zones of slides, boundinage, wedges, and undulatory folding in the Glen Ross Anticline (Fig. 102). These zones are up to 5 m thick and consist of laminated sandstone and subordinate shale with disturbed bedding, and lie between massive, uniformly dipping beds of medium- and coarse-grained arenite 2 to 3 m thick. The disharmonic zones have steeper, more irregular dips and tighter minor folds than the massive arenite which is very gently folded. The variable dips in the disharmonic zones occur where the upper part of a zone has moved down dip over the lower part. Minor structures indicate that the movement of the sediment was complex in detail. Some movement has been on slides—i.e. bedding-plane faults which separate discordant strata in one locality but which

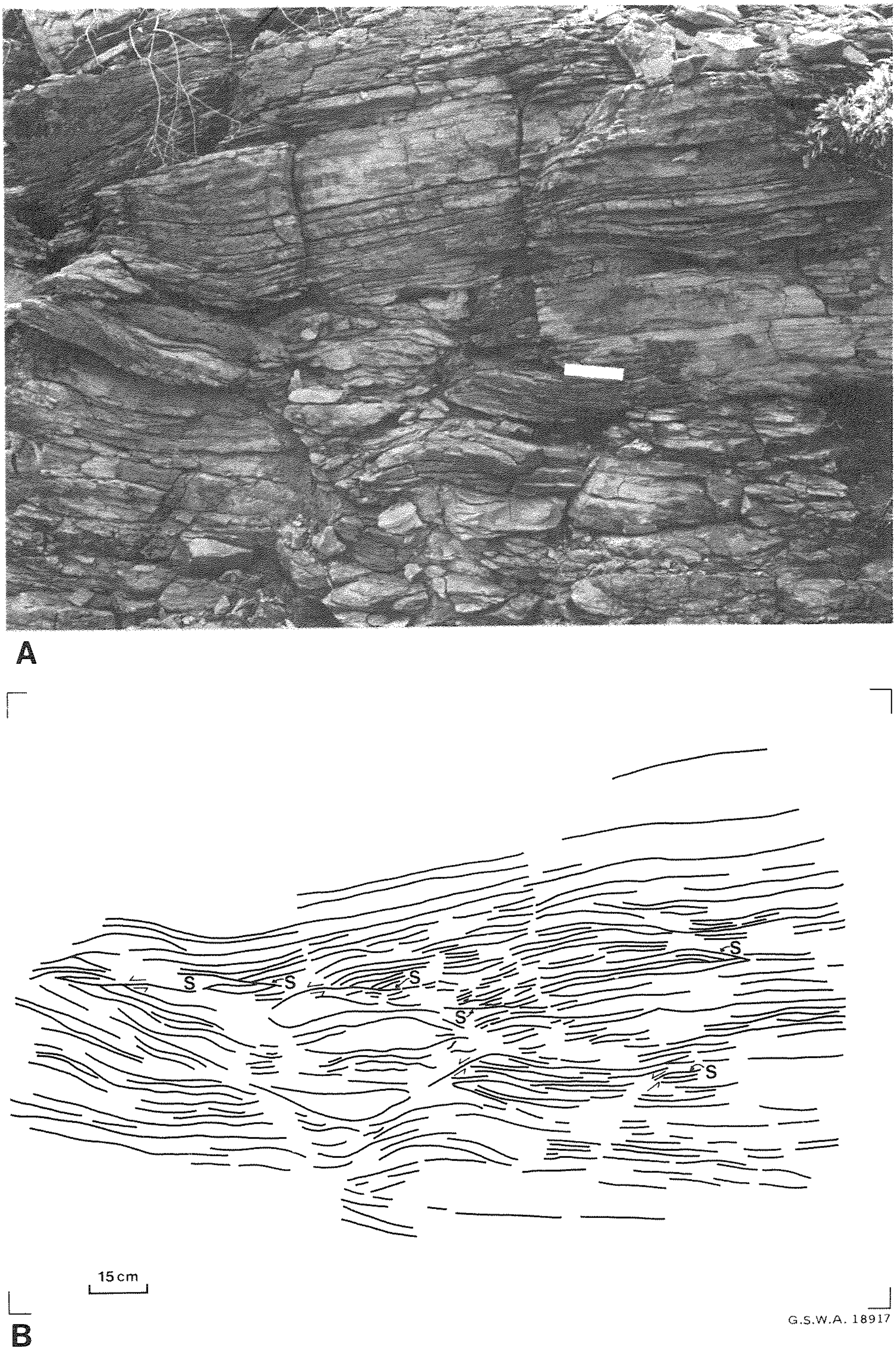


Figure 102. **A—Detail of part of disharmonic zone showing disharmony between upper and lower beds; slides and boudinage in laminated sandstone and minor shale, Kiangi Creek Formation, Glen Ross Anticline, MOUNT EGERTON. B—Tracing of photo in A. The upper beds have moved to the left, down the north limb of the anticline. Slides marked by "S".**

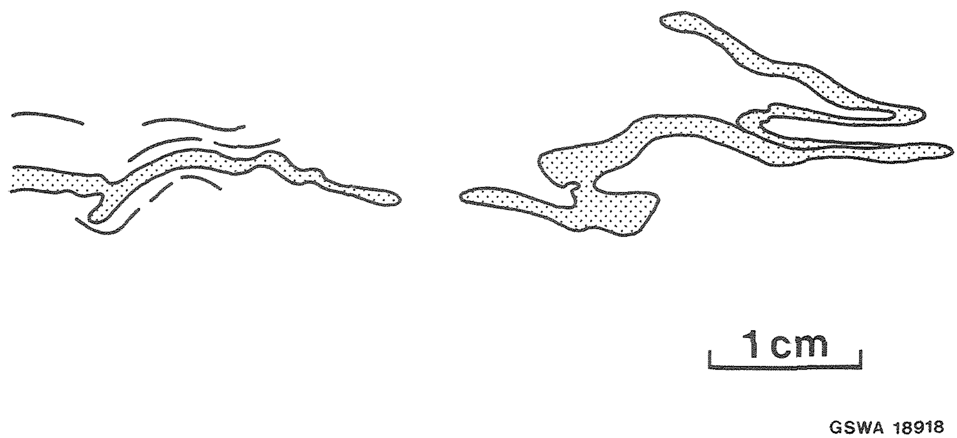


Figure 103. Slumped sandstone lamination in fine sandstone and siltstone of the Kiangi Creek Formation, Glen Ross Anticline. The slump direction is towards the Range Creek Syncline. The variation in the thickness of the lamination and tightness of some of the folds suggest the rock was plastic at the time of slumping. (Traced from photo).

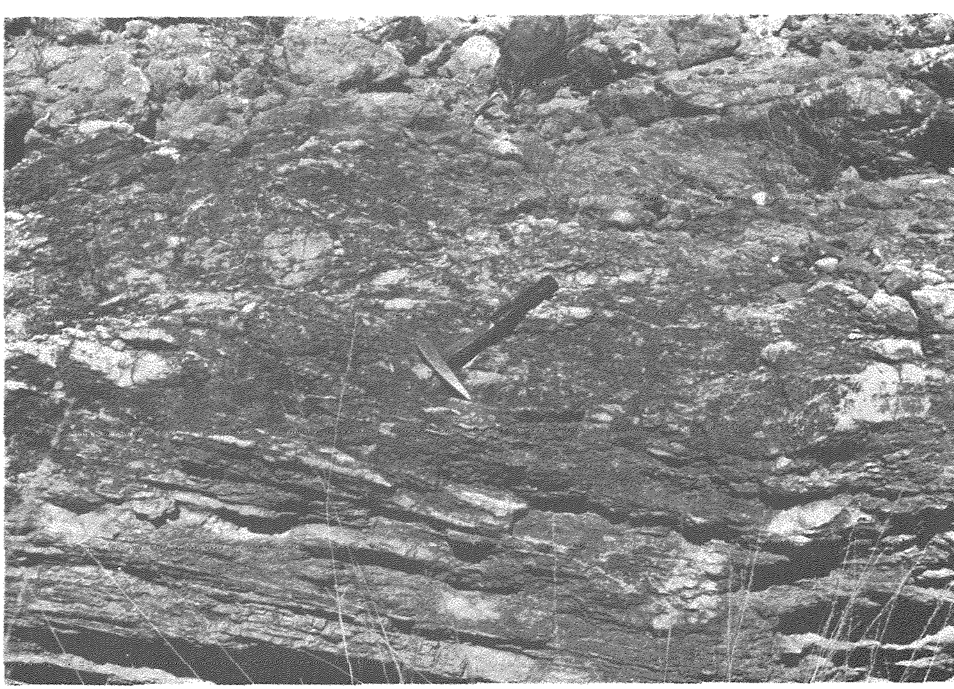


Figure 104. Low-angle fault-zone breccia, Kiangi Creek Formation, Glen Ross Anticline. The breccia blocks are of dolomite (white), sandstone and shale in a sandstone matrix. The beds under the fault zone dip down the limb of the anticline, whereas those beds above the fault dip back into the anticline. This configuration suggests that the top beds moved downslope compared to the lower beds.

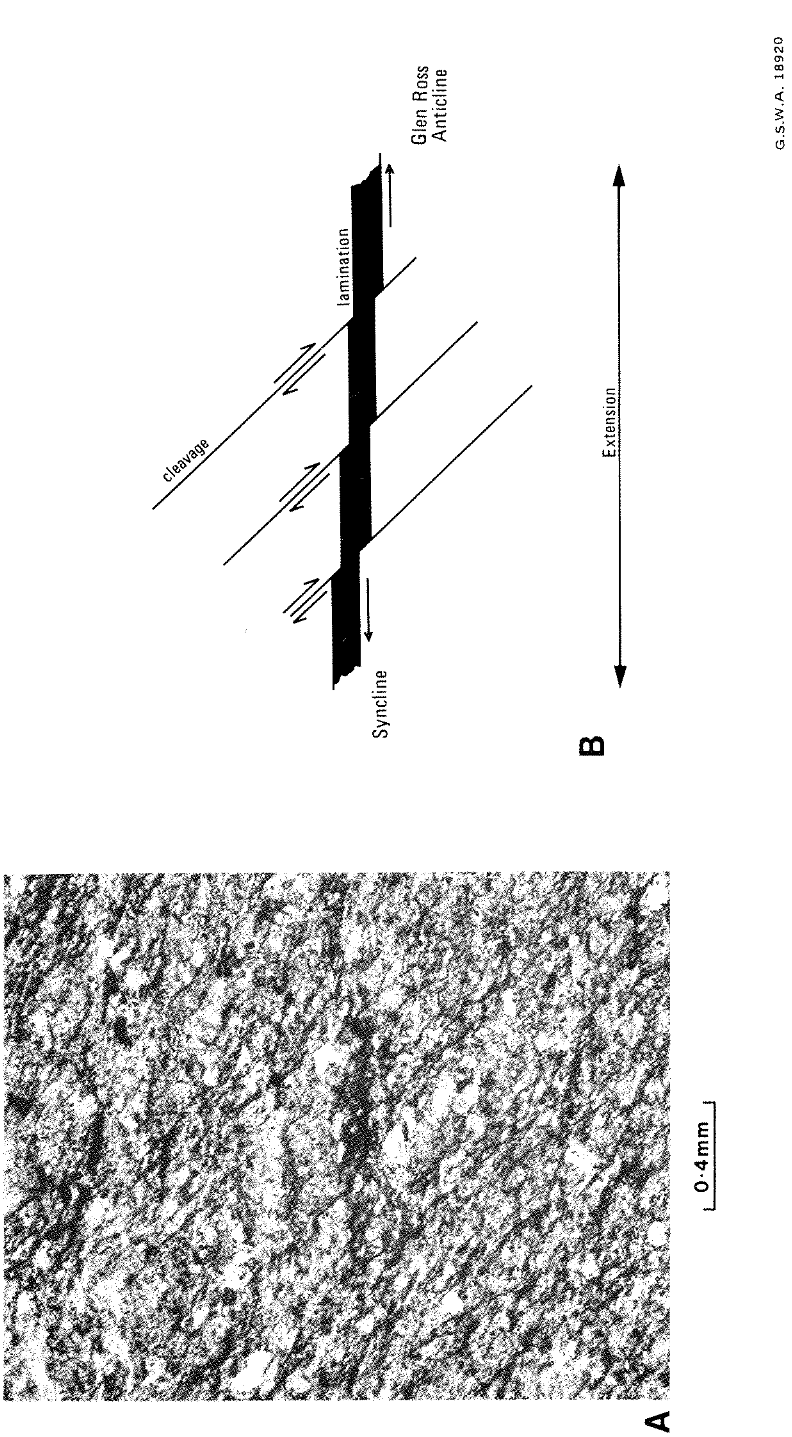
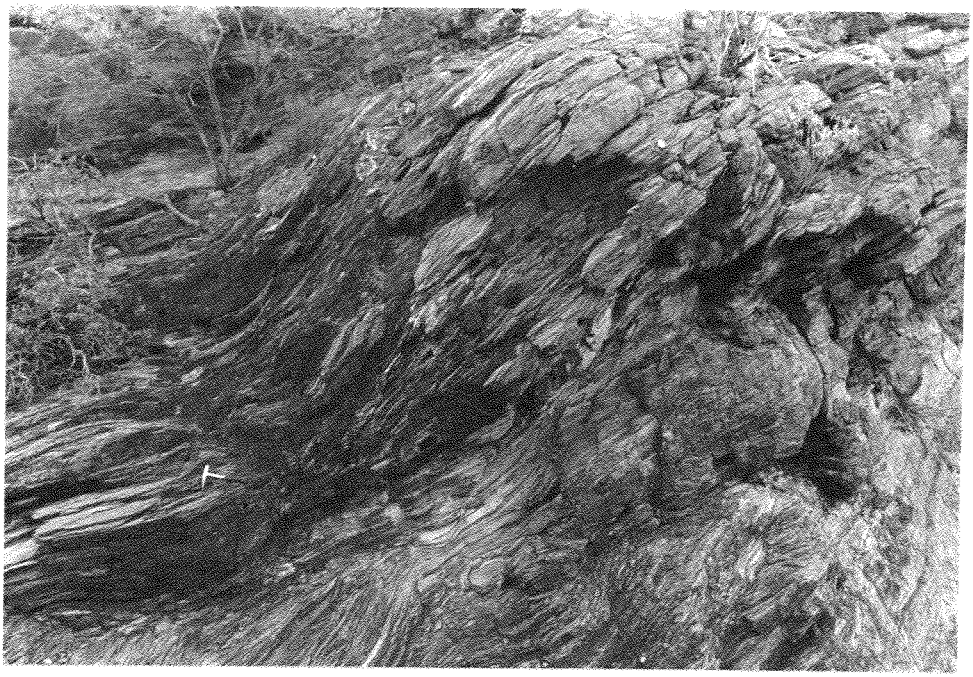


Figure 105. A—Photomicrograph of low-angle crenulation cleavage in shale lamination, Kiangi Creek Formation, Glen Ross Anticline. The left side of the photo is down dip. The lamination (horizontal) has been displaced by the cleavage such that each microlithon has moved up relative to its neighbour on the right (or anticline) side. Relatively large-scale displacement is visible in the centre. B—The sense of movement indicates that the shale has been laterally extended (Specimen 48569).



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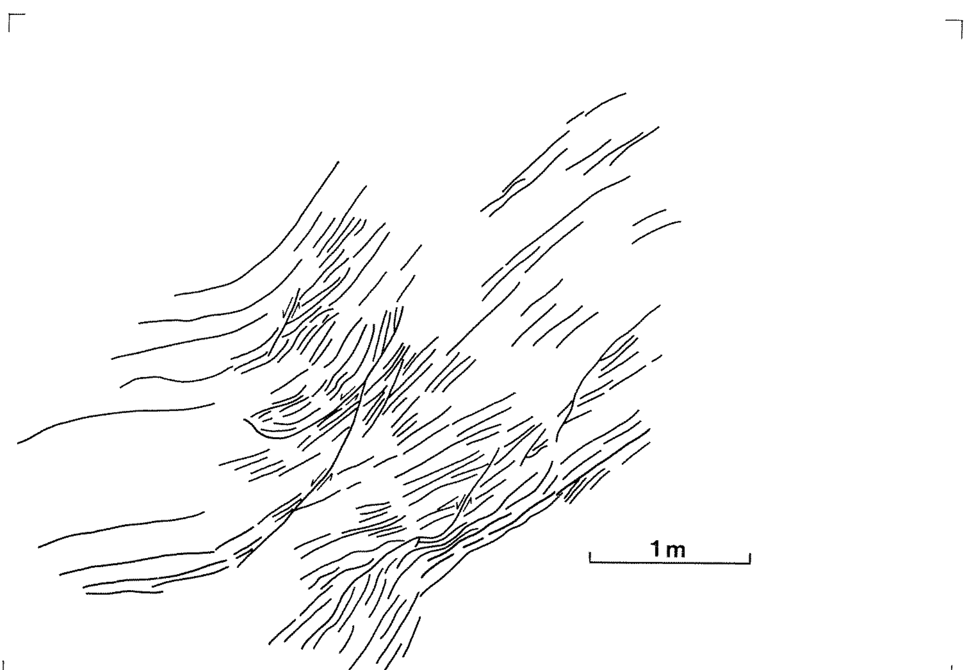


B

Figure 106. A—Monocline and associated high-angle normal faults, in laminated sandstone, Kiangi Creek Formation, Glen Ross Anticline. B—Tracing from photo.

elsewhere become parallel to bedding. The term slide is used in a similar sense to that by Davies and Cave (1976) who described bedding-plane faults formed in plastic sediments. The attenuation and bending of laminae near the slides suggest the laminae were still slightly plastic. Stretching parallel to the bedding was also achieved by boudinage which produced lenses of sandstone. Boudinage was caused by local extension, contemporaneous with the slides. Hence, the mass of sediment moved down the tectonic dip, and layers of differing plasticity moved at different rates. The gross structure in disharmonic zones where the upper beds appear to have moved over the lower is similar to a wedge (Cloos, 1961) in which low-angle shear planes cut beds into wedge-shaped slices. Bulk deformation of the rock could occur by transient extensional and compressional movements by these wedges during gravity creep. The slides and wedges in the Glen Ross Anticline have moved down the northern flank of the anticline, in the same direction as the slump folds.

Low-angle faults are characterized by brecciation and are interpreted to have formed later than the slides which involved plastic deformation. The faults provide evidence of gravity induced movement within the limbs of broad, open folds. The best example is the southern limb of the Glen Ross Anticline, where a fault dipping 25° south separates a sequence of medium-grained arenite dipping gently north from an underlying laminated fine-grained arenite dipping south (Fig. 104). The fault is marked by a breccia zone, 35 cm wide, which contains fragments of sandstone similar to those in the underlying block, as well as elongated lenses of dolomite in a sandstone matrix. The attitude of the beds on either side of the fault suggests gravity induced movement.

Low-angle crenulation cleavage is developed in shales interbedded with the laminated sandstone in the disharmonic zones in the northern limb of the Glen Ross

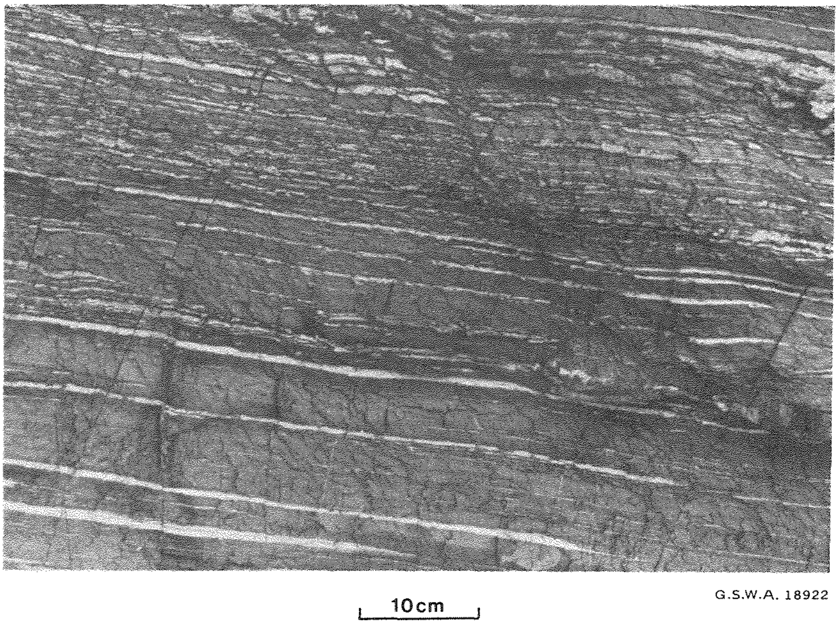


Figure 107. High-angle strain-slip cleavage showing normal sense of movement and resultant lateral extension in laminated sandstone, Kiangi Creek Formation, Glen Ross Anticline.

Anticline. It dips uniformly south and is independent of small-scale folds. No recrystallization of quartz has occurred although there has been bending and alignment of sericite and muscovite. The cleavage has produced a series of microlithons in which any northern microlithon has moved up in relation to its southern neighbour (Fig. 105). The form of the cleavage suggests it is the result of simple shear and it has the strain pattern of a multitude of micro normal-faults; the features indicate that the shale horizon has been laterally extended.

High-angle normal faults dip more steeply than 45° towards the synclines. Figure 106 shows an example from the south limb of the Glen Ross Anticline, where normal faults are associated with a monocline. These faults strike parallel to the regional fold axis and generally dip 70° south. Most faults show little shear or quartz veining, although one contains a breccia of vein quartz, sericite, iron oxides, and recrystallized quartz and carbonate.

A strain-slip cleavage, parallel to the steeply-dipping faults, is defined by aligned sericite and trails of iron oxides in the cleavage plane; it also has kinked detrital muscovite flakes. Many cleavage planes show small displacements (Fig. 107) in the same sense as the faults and, therefore appear to be due to simple shear associated with the normal faults. This cleavage contrasts, both in form and origin, to the slaty-style cleavage in the tight regional folds which shows good mineral orientation and is interpreted to have formed by flattening.

The significant feature in Figure 106 is that the normal faults and cleavage are concentrated in the steep limb of the monocline, which, therefore, appears to have formed by stretching. Similar zones of faults and cleavage are present across the top of the Glen Ross Anticline, and their orientation indicates the sediments have laterally extended by sagging away from the anticline.

FORMATION OF OPEN FOLDS

The minor structures in the sediments range from early plastic-style to late brittle-style features. Only a few small slump folds, which imply highly mobile, partly lithified rock, are present; most of the structures formed when the sediments were either plastic with a very high viscosity, or brittle. The structures fall into two groups: those which have formed by the downslope movement of layers (or groups of layers) over one another, and those formed by lateral extension of the sediments. Slumps, slides, boudinage, and low-angle faults form the first group. The slumps, slides and local boudinage all have features indicating that the sediment was plastic during their formation, whereas the low-angle faults are characterized by brecciation and probably formed when the sediments were brittle.

All these structures result in the transfer of material from the top of an anticline towards the synclines, and so produce thinning of the sediments in anticline crests. The process is an adjustment of the cover to a rising anticline. The second group includes low-angle crenulation cleavage, monoclines, normal faults, and high-angle strain-slip cleavage. The formation of the monocline also involves stretching of the sediments, roughly parallel to bedding, while the other structures cut across bedding. All these structures involve extension of the sediments by sagging off the anticline towards the synclines.

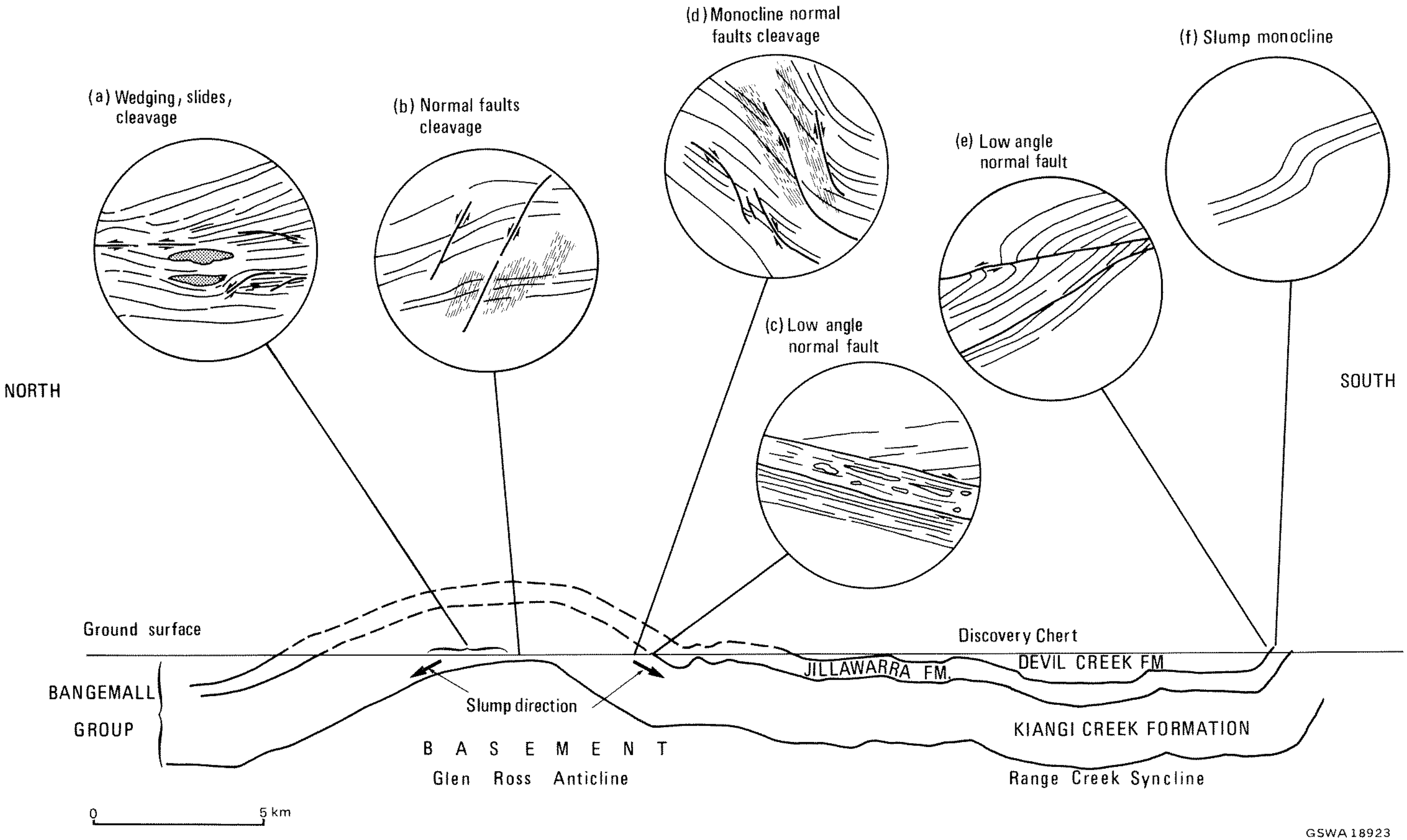


Figure 108. Relation of minor structures to open folds. Scale for inset diagrams all similar except for (D).

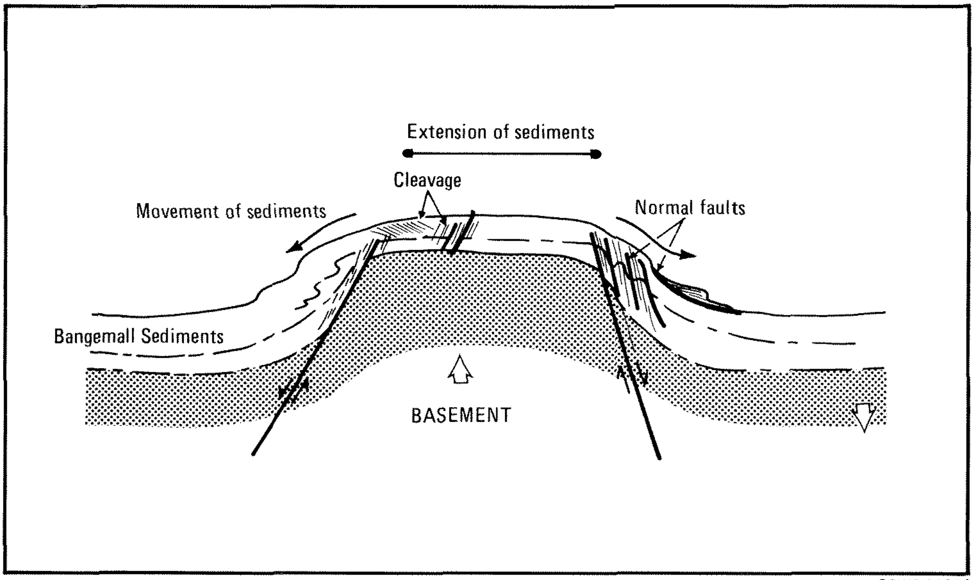


Figure 109. Diagram of formation of the Glen Ross Anticline by the stretching of the cover over the basement rocks.

Figure 108 summarizes the location and relationships between the minor structures and the broad folds. The sequence of events is interpreted to be:

(1) Minor gravity spreading of plastic sediments downdip towards the syncline.

(2) Slides and boudins formed in the more finely laminated arenaceous layers due to one bed moving downslope over another.

(3) Movement on low-angle crenulation cleavage, due also to down-slope movement of sediment, resulted in extension of the sequence.

(4) Further extension of the rocks took place by movement on steeply dipping normal faults and cleavage, as well as by stretching to form monoclines.

The formation of open folds can be deduced from the strain pattern in the Glen Ross Anticline. The location of zones of monoclines, normal faults and cleavage show that most strain is concentrated in the limbs of the anticline. The box shape of the anticline and the location of maximum strain in the limbs show that the anticline is a drape fold formed by the upward movement of basement blocks bounded by faults (Fig. 109). Thus, a typical syncline is a broad box fold resembling a long trough which has few folds, bounded by short limbs formed in response to normal faulting. The hinge lines are zones about which the short limbs have been rotated and stretched against the sides of rising basement horsts. The sediments responded initially by moving downslope towards falling centres (now synclines) and later by draping and stretching over basement blocks.

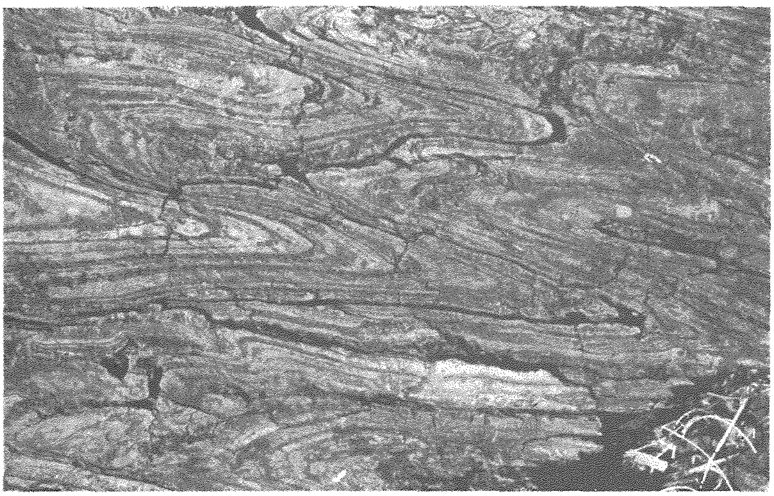
ZONES OF TIGHT FOLDS

Regional-scale tight folds are upright and symmetrical. Regional synclines are complex and contain numerous tight, major folds and a well-developed axial-plane cleavage. Profiles constructed from maps as axial-projection profiles of tight synclines are only approximate as the variation of plunge at depth is unknown. However, these profiles probably indicate the geometry of the synclines where the synclines are steeply plunging and maintain the same profile throughout much of the hinge zone. The profiles consistently show that layers are thicker in synclinal hinges than in limbs. Competent layers of sandstone or dolerite (where consistently parallel to bedding) are approximately class 1C (Ramsay, 1967), in which the curvature of the outer layer is less than that of the inner. Incompetent shales show approximately class 2 or 3 (Ramsay, 1967) profiles in which the curvature of the outer boundary of the layer is greater than that of the inner. The combination of those two shapes of layers results in a roughly similar (Hobbs and others, 1976) style fold and for convenience these are referred to here as similar-style folds.

SYNCLINAL OUTLIERS

The outliers such as the Mangaroon and Ti Tree Synclines are tighter than most of the synclines within the main outcrop of the Bangemall Group and show best the style of tight synclines.

The Ti Tree Syncline is 70 km long and narrow; its average width is 4 km. The fold is very tight, almost isoclinal and bedding dips are 70°-90°. Minor folds in chert bands in the axial zone of the Ti Tree Syncline have the same axial-plane cleavage as the main syncline and hence should be an approximate representation of the style of the larger fold in the competent rocks. These minor folds (Fig. 110) have a similar (Ramsay, 1967, class 2,) profile. In detail, (Fig. 111), the layers have either a class 2 profile, or alternate layers have class 1C and class 3 profiles, which together produce a class 2 shape. The axial-plane cleavage common to these minor folds in the Ti Tree Syncline is composed of aligned sericite and stretched quartz grains which form a true slaty cleavage at a low angle to the bedding.



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2cm

Figure 110. Tight minor folds in the Discovery Chert, axial zone of the Ti Tree Syncline.

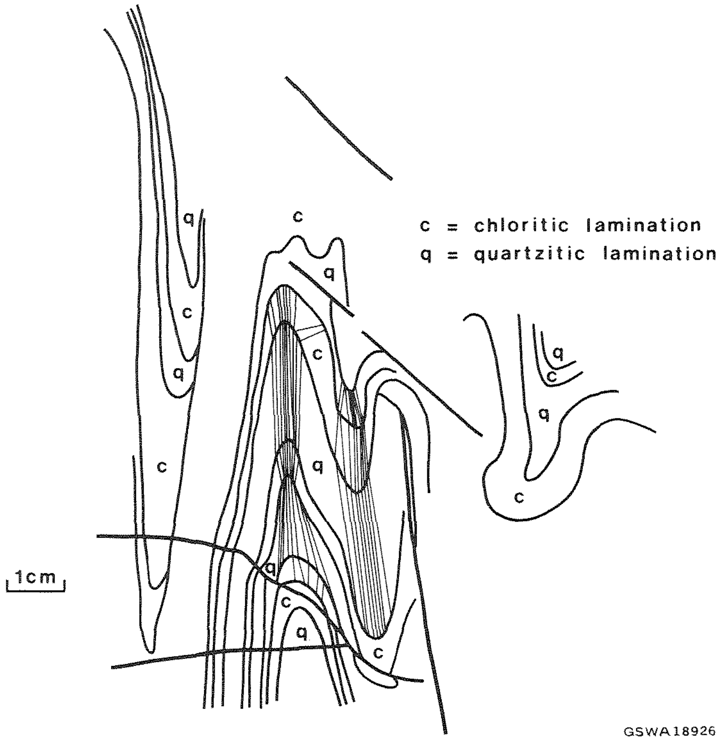


Figure 111. **Profile of similar-style minor folds. Isogons show that chlorite layers have a Class 1C shape and that quartzite layers display mostly a Class 2 or 3 shape. However the field overall has a Class 2 profile.**

The Coolinbar Hill Synclinorium (Plate 2) is a gently-plunging complex of anticlines and synclines whose limbs dip 60° or more. Axial-plane cleavage in the synclines is similar to that in the Ti Tree Syncline. It is expressed in chert as aligned sericite, and elongate, microcrystalline quartz grains. Cleavage in the anticlines of shale and chert is more weakly developed. The silty laminae in shale have cleavage which is defined by fractures spaced 1 to 2 mm apart as well as elongate blebs of microcrystalline quartz and clay, whereas in the fine mudstone layers, sericite parallel to bedding is crenulated, and some microcrystalline quartz is elongated parallel to the cleavage. Therefore, the cleavage in the anticlines is also slaty, and cuts minor, open buckle-style folds.

The style of minor structures within the tight synclines suggests that they formed either by flattening of initially parallel folds or by flow parallel to the axial surface. Minor folds in the Ti Tree Syncline do not have a perfect similar profile, but display profiles of flattened parallel folds (Ramsay, 1967). Bedding surfaces in tightly folded arenite show slickensides perpendicular to fold axes, indicating bedding-plane slip during folding. These points suggest that the Coolinbar Hill Syncline, which contains open buckle folds and a weak slaty cleavage, represents an early but arrested stage in the development of the tight regional folds such as the Ti Tree Syncline.

TIGHT ANTICLINES

Tight, regional anticlines are comparatively simple folds with few accompanying major folds and a poorly developed axial-plane cleavage. There are only two examples—the Coolina and the Sullivan Anticlines (Plate 3). Both folds have cores of basement metasediments which are folded along with the cover. Wherever visible, the contact between cover and basement is a fault zone. The fault at the northern contact in each anticline is steeply dipping and cuts across bedding whereas the fault at the southern contact is steeply dipping and subparallel to bedding in cover and basement or to foliation in basement.

The cross-sections suggest that the basement anticline was present prior to deposition of the Bangemall sediments and that it was tightened during deformation of the cover. However, this process alone seems incapable of producing bedding in the basement subparallel to that in the cover. The cross-sections show that, in both anticlines, the core of basement moved up along faults into the cover. The upwards movement of the core would have produced slip on the unconformity, by stretching the cover over the rising basement, and so bringing the bedding in both basement and cover to near parallelism.

Movement of basement up into cover is also indicated in the Mulgul Anticline, where a discordant slice of basement wacke is enclosed in Irregully Formation. It appears that the wacke was detached from the basement and incorporated into the cover during formation of the anticline. This suggests that the whole anticline was able to make room by upward movement when the flattening phase produced constriction in the core. This idea is consistent with the poor development of cleavage and lack of tight major folds in the regional anticlines and indicates that they formed under conditions of less strain and lower confining pressures than the tight synclines.

A MODEL FOR BASIN DEFORMATION

It therefore seems that the deformational phase of the basin was initiated by the sinking of basement blocks to form horsts and grabens (Fig. 112). These blocks were formed along structural boundaries of different lithologies or along old fault zones in the basement which were established during a previous orogeny. Sedimentation was also controlled by the initial phases of basement segmentation.

The first regional deformation of the sediments was by gravity spreading, in response to vertical movement of basement horsts and graben. Gravity spreading involved two processes which operated together: movement on sediments downslope and stretching of the sedimentary cover. The sediments moved downslope towards synclinal zones and slump folds characterize the early stages of movement. As the comparatively competent clastic sediments become lithified they continued to move off the anticlinal zones by a series of complex local adjustments involving slides, boudinage and wedging. Continued development of the horsts produced drape folds in the cover, and extension of the sediments which lay directly over basement fault zones by movement on steeply dipping normal faults and associated cleavage. The less competent sediments, such as black shale and chert, were deformed by bedding-plane slip while still plastic, disrupted fine depositional structures and formed streaky textures in chert and some shale. With further lithification, the black shale and chert developed lithified zones separated by layers of high plasticity, perhaps where connate water was trapped. During gravity spreading and later extension of the sediments these zones of high plasticity were the weak layers which became zones of high strain.

Following the vertical movements, the basement was laterally shortened. The cover rocks over rigid blocks of basement were only slightly affected; however, sediments over incompetent or ductile basement became highly deformed by two processes: the tightening of basement grabens and the rise of basement horsts to form antiforms. Thus synclines in the cover have been squeezed tightly between competent basement blocks to produce flattened parallel folds with an axial-plane slaty cleavage. Antiforms in the basement were also squeezed between competent basement blocks, but their position as horsts enabled easy upward release. This produced a diapiric effect with a rising core of basement mantled by a stretched cover.

The final process is an imbrication of the cover by reverse faults. The normal faults, and the zones of tight folding in the footwall blocks of the reverse faults, indicate they were initially zones of normal faulting. The continuation of basement shortening pushed the antiformal zones up over the adjacent buttresses of rigid basement, resulting in overriding, stretching and production of reverse faults, which are thus an expression of the last stages of folding. This process of overriding is also evident in most of the cross-section, (Plate 3), where the northern edge of each segment of cover has slightly overridden and is higher than the adjacent segment.

BULK STRAIN IN THE EDMUND FOLD BELT

Accurately constructed profiles of a fold belt, such as Plate 3, are often used to estimate bulk strain by comparing the arc-length of a marker horizon with its chord-length. From these measurements estimates of crustal shortening are often made. Such a simple analysis can be made in the Edmund Fold Belt, using the Discovery Chert as a marker horizon, providing the stretching effects of the early drape folds are considered.

Bulk deformation of the sedimentary cover is made up of three components, which from youngest to oldest are:

(1) Shortening recorded by the heave of reverse faults. In the cross-section, the sum of the heaves of the reverse faults is 4.4 km, a figure that can be applied to both the cover and basement.

(2) Buckling and flattening of cover by flattening in ductile zones in the basement. As there is no thrusting observed along the unconformity, this component also applies to both the cover and basement. The sum of the excess of arc-length over chord-length in all the zones of tight folding is 9 km. This figure does not take into account the flattening component which has modified parallel-style folds into similar-style folds. Stretching of the arc-length would exaggerate the shortening, but at low values of strain, this would be compensated by the shortening due to the flattening itself. It is reasonable to conclude that crustal shortening during this stage was about 9 km.

(3) Stretching of the cover during drape folding. The sum of the differences between arc-length and chord-length in the zone of open folds gives 6.4 km, a figure which refers only to the stretching (both vertical and horizontal) of the cover. This extension is postulated to occur on steeply inclined normal faults which in horst-and-graben terrains probably dip at no less than 60°. On this assumption, the total crustal extension is given by $6.4 \times \cos 60° = 3.2$ km. Three kilometres is probably a reasonable estimate of the crustal extension at this early stage in the structural development.

All this deformation occurred across the 100 km width of the Edmund Fold Belt, which before crustal shortening, was 123 km wide. The bulk strain can therefore be estimated as an initial extension of 3 km or 2.5 per cent and a later shortening of 13.4 km or 10 per cent. Although merely simple estimates, these figures provide constraints on the magnitude of rifting and collision movements and general plate-tectonic models which are discussed in Chapter 9.

CHAPTER 8

Economic Geology

The Bangemall Basin is a Proterozoic intracratonic basin containing mature sandstone; pyritic, carbonaceous, and evaporitic shale; stromatolitic carbonates; and subordinate volcanic rocks. By analogy with other Proterozoic sequences of platform sediments, it should be prospective for stratabound deposits of copper, lead, zinc, and, possibly, gold and uranium. Some encouragement is given by small copper, gold, and lead-zinc prospects; however, no major mineral deposits have yet been located and no uranium mineralization has been recorded. Most mineralization is associated with faults and quartz veins (Blockley, 1971, Marston, 1980), which are secondary features.

It should be useful to examine mineralization in the Bangemall Basin from two aspects: firstly to see if there is a pattern to the known mineralization; secondly to evaluate possible tectonic and sedimentary controls of potential stratabound or stratiform mineral deposits.

MINERALIZATION IN THE BANGEMALL BASIN

Mineralization known in the Bangemall Basin can be divided into the following groups:

(1) Stratabound zinc with minor copper

(2) Fault-zone deposits of lead and zinc

(3) Fault/fissure deposits of copper at Kumarina and Ilgarari

(4) Gold

(5) Veins of copper and lead in carbonate rocks

(6) Surface enrichment of managanese

(7) Stratabound phosphate.

STRATABOUND ZINC-COPPER

Stratabound zinc-copper occurrences are present near the top of the Jillawarra Formation. They are small, uneconomic concentrations of fine-grained disseminated pyrite, and subordinate sphalerite and covellite in laminated black shale at Mount Palgrave and Mount Vernon (Marshall, 1968), or in carbonaceous siltstone and siderite at Quartzite Well (Davy, 1980). Typical drilling results from around Mount Palgrave were less than 0.2% copper and 0.5% zinc. The best drillhole intersection at Mount Vernon assayed 1.2% zinc and 0.1% copper over 11 metres. The occurrences are at a consistent stratigraphic level, and this, together with the sulphur isotopes and lack of comparable metal concentrations in lower stratigraphic levels indicates they are syngenetic (Marshall, 1968).

The abundances of trace metals in the Jillawarra Formation at Mounts Vernon and Palgrave are similar to those in other shallow-marine, black-shale sequences (Davy, 1980). In addition, the metal abundances in the Jillawarra Formation correlate with carbon and sulphur at Mounts Vernon and Palgrave (Davy, 1980) and so, are related to the reducing potential of the environment rather than the clastic fraction. Therefore, the Jillawarra Formation represents an environment recognized elsewhere in the world as suitable for the fixing and concentration of metals (i.e. reducing conditions in the presence of carbonaceous material of organic origin, and little input of clastic material).

The presently known distribution of metals in the Jillawarra Formation was discovered by exploration companies (Union Miniere Development and Mining Corporation Ltd, and Westfield Minerals (W.A.) N.L.), who demonstrated that geochemical stream-sediment sampling is a valid exploration method for detection of mineralization within the Bangemall Basin (Smith and Davy, 1979). Most zinc-copper occurrences lie close to boundaries between structural provinces in the Bangemall Basin, which probably coincide with sedimentary hinge lines controlled by basement movements. The continuity of fine lamination and the fine grain size of the Jillawarra Formation suggests these movements must have been slow and comparatively small. However, there are some indications of emergent basement shown by thinning of the Jillawarra Formation over basement arches (Marshall, 1968) and deposition of sand units towards the top of the formation in the Mount Vernon Syncline and at the east end of the Lyons River Anticline. These movements on basement structures may have produced areas of enhanced reducing potential which fixed the zinc-copper minerals.

The source of the metals is unknown. There is no evidence for widespread volcanism during sedimentation. Detrital copper sulphide at Quartzite Well (Davy, 1980) indicates that movement on basement structures had exposed a mineralized basement which was eroded to contribute to the stratabound mineralization.

LEAD-ZINC

Pyrite-lead-zinc occurrences are in the breccia of the Quartzite Well Fault, where low-grade metamorphosed basement comprising sandstone, dolomite, shale, slate and chlorite-magnetite-quartz-phyllite, has been thrust over Jillawarra Formation, Discovery Chert, and Devil Creek Formation (Plate 3). The basement sediments may possibly correlate with the Glengarry or Padbury Groups (Bunting and others, 1977; Gee, 1979a) on PEAK HILL and GLENGARRY.

At the surface a ferruginous, siliceous gossan extends over 1 400 m. At depth, galena and sphalerite are associated with late-stage veins of microcrystalline silica and pyrite. Amoco Minerals of Australia Ltd reports the best mineralization intersected so far (June, 1979) was 34 m assaying 1.99% Pb, 1.07% Zn and 27.9 g/t of Ag. The source of the mineralization is not clear.

FAULT-FISSURE COPPER DEPOSITS

Copper deposits at Kumarina and Ilgarari (COLLIER) are supergene enrichments of sulphide-quartz fault and fissure fillings, which are commonly situated along northeasterly trending contracts between shale and dolerite (Marston, 1980). These copper deposits appear to be related to the basement, as indicated by the following points:

(1) The mineralization is independent of stratigraphic level; it occurs in the Fords Creek Shale, and the Ilgarari and the Backdoor Formations.

(2) All deposits lie within a broad zone trending north-northwest (Plate 2), which also extends to copper prospects in the pre-Bangemall metasedimentary sequences.

(3) Most of the mineralization is located in northeasterly trending faults which probably penetrate the basement of the Bangemall Group.

The source of the copper is unknown. The surrounding black shale is only weakly anomalous in copper (Marston, 1980), and dolerite sills elsewhere in the basin contain no copper mineralization. Point two suggests the source of the copper is the pre-Bangemall sequences such as the Wyloo, Padbury, and Glengarry Groups. The production from the principal mines is given below.

Copper Ores & Conc.	*Cupreous Ore & Conc.*	*Av. Grade*	*Contained Copper*
1907.70 t	—	30.76%	586.86 t
—	1252.81 t	16.19%	202.82 t

GOLD

Gold in significant quantities has been discovered only at the Bangemall Mining Centre in the Cobra Synclinorium, where quartz veins forming saddle reefs and veins parallel to axial-plane cleavage contain free gold, pyrite, and carbonate (Maitland, 1909). The reefs and veins are in cleaved shale and dolerite sills near the top of the Jillawarra Formation. Deformation and low-grade metamorphism appear to have controlled the emplacement of these deposits. Known gold mineralization is restricted to this occurrence and indicates a unique source. Basement metasediments which are probably correlates of the Glengarry Group, contain gold in the core of the Hibernian Anticline (Plate 2). Similar rocks beneath the Cobra Synclinorium could provide a source from which gold was moved during periods of deformation and dolerite intrusion within the Bangemall rocks. Another possible source of gold could be placer deposits from alluvial fans in the Mount Augustus Formation which should extend beneath the Cobra Synclinorium. Production from the main mine, the Carnarvon Gem, is given below.

Dollied	*Ore*	*Gold*	*Total Gold*	*Period*
0.194 kg	179.18 t	4.735 kg	4.928 kg	1897, 1899

COPPER-LEAD(-ZINC) IN CARBONATE ROCKS

Minor showings of copper and lead are scattered throughout carbonates of the western Bangemall Group (Plate 2). The mineralization is associated with quartz veins, some of which occupy shear zones in dolomite (e.g. McCarthy Find), or quartz veins in dolerite intrusive into dolomite, e.g. Latham prospect (Blockley, 1971). Primary minerals are commonly galena and chalcopyrite, and secondary minerals are cerussite, anglesite, malachite, and chrysocolla. The metals were presumably concentrated from the surrounding carbonate.

The largest of the minor deposits is the Joy Helen prospect (EDMUND), which also contains zinc (sphalerite). The mineralization forms irregular lodes within brecciated, silicifed algal dolomite (Blockley, 1971). The age and origin of the breccia is uncertain. Daniels (1965) has presented evidence that similar breccia on WYLOO is Proterozoic rather than Cainozoic. Most breccia occurs in the Irregully Formation, but some is interbedded with the Devil Creek Formation (Fig. 31). It is uncertain whether the mineral lodes owe their origin to the breccia or to the dolomite.

Blockley (1971) has classified the deposit as a Mississippi Valley type. Details of the form and grade of all these occurrences are given by Blockley (1971).

MANGANESE

Most manganese deposits have developed on the Manganese Subgroup and were formed by superficial enrichment related to Tertiary river systems. They were described by de la Hunty (1963). The protore is manganiferous sediments (mostly Noreena Shale) of the Manganese Subgroup, and common minerals are pyrolusite, cryptomelane and braunite. The deposits are mostly small and scattered. The largest deposit, that at Balfour Downs is one of the largest in Western Australia, with reserves of 1.72 Mt grading 35% to 40% Mn, but the grade is too low to be economic.

Minor surface enrichments of manganese minerals are common in the Mucalana and Collier Subgroups: the eastern area of the Fords Creek Shale, the Kurabuka Formation, the northeastern Backdoor Formation, and the Ilgarari Formation. The largest of these is developed on shale of the Ilgarari Formation and is uneconomic. The Woodlands and Mulgul occurrences (de la Hunty, 1963) are associated with dolomite and shale respectively. These rocks are now considered to be either the Padbury or Glengarry Groups, which lie unconformably beneath the Bangemall Group.

STRATABOUND PHOSPHATE OCCURRENCES

Variscite, a green, hydrated aluminium phosphate, occurs in the Jillawarra Formation and the lower part of the Backdoor Formation which is equivalent to the Jillawarra Formation. The main deposits are on MOUNT EGERTON (Muhling and others, 1978) and ROBINSON RANGE (Elias and Williams, 1977). In both areas the variscite is in narrow (<100 mm) veins in silicifed mudstone and shale or in irregular masses in brecciated chert. Both deposits are close to dolomite and chert. Small amounts have been mined for semi-precious stone from near Milgun on MOUNT EGERTON.

The variscite is clearly secondary; however, the factors responsible for concentration of the mineral are not known. Variscite appears to be restricted to the Jillawarra Formation—Backdoor Formation interval and may indicate an anomalous content of primary phosphate.

POSSIBLE ANALOGUES

McARTHUR RIVER, N.T.

The McArthur River H.Y.C. lead-zinc deposit lies in the McArthur Basin, N.T. a gently deformed, unmetamorphosed platform-cover sequence of Proterozoic age (1.5-1.6 b.y.), unconformably overlying a lower Proterozoic orogenic zone (1.8-2.0 b.y.) consisting of low- to high-grade metasediments, dolerite, and anatectic granitoids, as well as post tectonic granitic rocks (Plumb and Derrick, 1975). The following summary is based on work by Murray (1975), Croxford (1968), Walker and others (1978) and Muir (1979).

Within a carbonate sequence there is a mineralized black pyritic shale (H.Y.C. Shale) that appears to have formed in a local sub-basin controlled by faulting of the basement. This shale passes laterally into dolomite breccias which are believed to be talus that accumulated against an active fault. Fine-grained, stratiform lead-zinc mineralization occurs in the pyritic shale, and most workers seem to be agreed upon a syngenetic origin involving precipitation of metal sulphides in micro-laminae in a euxinic environment, possibly by sulphate-reducing bacteria. Coarse-grained stratabound lead-zinc mineralization occurs as fracture fillings in the dolomite breccia, and may represent either pools of syngenetic sulphidic muds that were remobilized during brecciation, or permeation of solution-collapse breccias by metal-rich brines which were channelled by the fault. Volcanism does not appear to have played a direct role in the mineralization, or to have been a major influence in the environment; however, there were intermittent volcanic ash falls up until the time of mineralization.

There are no localities in the Bangemall Basin with sedimentational and environmental features precisely similar to those at McArthur River. There is little evidence for evaporitic rocks contemporaneous with volcanicity and subsidence. However, there are two areas in the Bangemall Basin where volcanism was contemporaneous with subsidence.

In the first area, west from Tangadee, tuffaceous bands occur in the Backdoor and Devil Creek Formations; rhyolite also occurs in the basal Coobarra unit at Tangadee (Gee and others, 1976).

However, only those units which contain black shale in local sub-basins would be suitable environments to fix metals. Only the Backdoor Formation and the Jillawarra Formation—Discovery Chert interval contain black shales. Sub-basins related to contemporaneous tectonism and containing black shale are mostly restricted to pre-Discovery Chert units, namely the Jillawarra Formation and its lateral equivalent. The lower part of the Backdoor Formation, especially where adjacent to a tectonic lineament (Tangadee Lineament), could provide a setting similar to that at McArthur River.

The second area, near Mount Palgrave, contains a felsic dyke and flow (Marshall, 1968), and anomalous copper-zinc concentrations in the lateral equivalent of the Jillawarra Formation, the "Prospect Shale". However, this unit is thin; and, although large thickness variations are unlikely, the most likely areas for the development of small sub-basins are where the Jillawarra is between northeasterly trending basement arches.

KUPFERSCHIEFER

The Permian Kupferschiefer is a bituminous, calcareous, or dolomitic shale, of which, an area of 20 000 km^2 is greatly enriched in copper, lead and zinc (Dunham, 1964). It is part of a series of unmetamorphosed, and relatively undeformed platform sediments that were deposited adjacent to a block-faulted foreland of granite of the Variscan orogenic belt. Continental sediments containing sabkha, eolian, and fluviatile deposits with locally significant areas of volcanic rocks (Glennie, 1970) form the basal sequence. During transgression of a lagoonal facies over the continental sediments, the Kupferschiefer was deposited slowly in quiet, anaerobic conditions with enrichment in copper, lead and zinc sulphides in local areas of stagnation (Dunham, 1964). The sulphides form minute, disseminated spheres in lenses parallel to bedding. Framboidal pyrite is interpreted as biogenic by Love (1962) as cited by Dunham (1964). Overlying the Kupferschiefer is a blanket of magnesian limestone, which was laid down in hypersaline conditions and this in turn is overlain by an evaporite series.

The source of the metals is unknown, though Dunham (1964) estimates the necessary quantities could have been eroded from the Variscan highlands. An alternative interpretation (Jacobsen 1975) is that the stratigraphy represents a sabkha situation (Renfro, 1974) whereby constant evaporation in the hypersaline environment draws fluids up through the underlying rocks. Metals in the basal continental sediments and volcanics are leached by the rising fluids and fixed in the reducing environment of lagoonal black shales. Hence, slow sedimentation, a reducing environment, and hypersaline to evaporitic conditions, are essential features. The presence of underlying continental sediments (including red beds) and volcanics provides a suitable source of metals.

Few of the Bangemall black shales are underlain by continental sediments and overlain by evaporitic carbonates. Continental sediments are represented by the Mount Augustus, Tringadee, and Coobarra Formations as well as the Prairie Downs Member. Only the Tringadee Formation contains shales towards the top and is overlain by carbonates—the lagoonal sediments of the Irregully Formation. In the shale, interbeds of cross-bedded, fine-grained arenite and siltstone suggest sedimentation was not particularly slow nor was the environment stagnant.

Some dolomites in the Bangemall Group show characteristics of sabkha deposits, namely, algal mats and 'birdseye' texture, and most were deposited in either the intertidal or subtidal zone. However, they contain no anhydrite (or gypsum), one of the diagnostic features of sabkhas, pointed out by Wood and Wolfe, (1969), and cited by Selley, (1976). Evaporites imply a regression which produced a continental environment during a period of aridity. The lower part of the Bangemall Group has resulted from a slow steady transgression. The only indication of a regression is on NEWMAN, where dolomite remnants of Top Camp Formation overlie the Bresnahan Group. The dolomite was probably extensive prior to deposition of the Prairie Downs Member, but has been partly eroded, and overlain by the Prairie Downs Member (Daniels and MacLeod, 1965) which are probably continental sediments. In general, the potential for sabkha deposits seems low in the Bangemall Basin.

ZAMBIAN COPPER BELT

The Zambian Copper Belt lies in 1 100 m.y. old, weakly metamorphosed platform-cover sediments adjacent to the Congo Craton (Jacobsen, 1975). The basement, which is unconformably overlain by a sequence of metasedimentary rocks, comprises schists, paragneisses and granitoids (Garlick, 1961). Rifting of the basement (Brock, 1961) was followed by deposition of the cover sequence of fluvial and shallow-marine clastic sediments and carbonates, some of which are algal reefs. Inter-reef argillites contain pyrite, which is stratigraphically the lowest sulphide and which is followed upwards by chalcopyrite, then bornite. This zoning is also expressed horizontally: bornite occurs at the reef fringe; and pyrite, in deeper water shales. The basal unconformity has considerable relief, and basal members were deposited between hills of granite (Stanton, 1972). The stromatolite reefs occurred in the very shallow water, and the optimum mineralization occurs in the carbonaceous shale that was deposited in sheltered embayments fringing the reefs. Hence, there is a relation between mineralization and the palaeo-coastline, but biological activity, reduced circulation, and sedimentary transgressions were also important. Anhydrite layers, as well as magnesite and dolomite, suggest a hypersaline environment during either deposition or diagenesis (Jacobsen, 1975), and eolian sandstone occurs in the basal sequence. Such an association indicates that metals may have been deposited from brines where they met the reducing carbonaceous bands, either directly as in the sabkha model proposed by Renfro (1974) or indirectly during diagenesis (Jacobsen, 1975).

Although there are some basic similarities, precise analogies do not appear to be present in the Bangemall Basin. The basal unconformity is of more subdued relief especially on the northern edge of the basin, although a few places on the southern edge such as the basal contact of the Mount Augustus Formation show appreciable palaeotopography. The basement area north of the Sullivan Anticline probably was a topographic high during sedimentation and a source for the fluviatile sediments of the Tringadee Formation. However the unconformity at the base of the Tringadee Formation around the Hibernian Anticline has low relief, and, therefore, the palaeotopography was subdued. The southern and western end of the Resolution Synclinorium contains a few of the requisite features of Copper Belt geology (i.e. fluviatile sediments overlain by dolomitic sediments containing shales). However, it is not known if these shales are carbonaceous.

If source area is important, then the Backdoor Formation near the northeast end of the Marymia Dome offers possibilities for metal fixing. Here, black shale unconformably overlies basement that contains copper in quartz veins. However, the relief on the unconformity was low, and it is difficult to envisage significant concentration of metals in this environment.

MARINE PHOSPHATES

Marine phosphate deposits range in age from Precambrian to Tertiary. Their significant points as summarized in Christie (1978) are:

(1) The tectonic setting ranges from geosynclines to platforms. However, most phosphates form on marine shelves which are generally of small extent in geosynclines, but form the major part of platform areas.

(2) The distribution of sedimentary facies and the occurrence of modern phosphate deposits indicate that most phosphates formed in water of normal salinity at depths of less than 500 m.

(3) Organic material is abundant in most phosphorites and indicates that a reducing environment is usual, but not essential. Organic material also influences the alkalinity of the environment, which may be a primary control for phosphate deposition (Cook, 1976). The rich assemblage of flora and fauna indicates a source of phosphate which is rich in nutrients, such as estuarine waters or upwelling cold oceanic waters.

(4) Sedimentary textures, such as rounded fragments of phosphatic brachiopod shells, suggest that some phosphate is primary. However, the presence of phosphatized calcareous fossils as well as the discrepancy between the low primary phosphate content of fecal pellets in contrast with their high phosphate content in phosphate deposits, together indicate diagenetic upgrading of many deposits. Mechanical reworking by currents has also enriched deposits.

(5) Terrigenous sediments are of minor importance and are fine grained, and sediment from adjacent cratonic areas has not diluted the components of chemical deposition and concentration of phosphate. For example evaporites and red beds on the cratonic side of the Phosphoria Formation suggest an arid climate, which has ensured little transport of clastic sediment into the phosphate.

The miogeosyncline and platform tectonic models of phosphate deposition have aspects applicable to the Bangemall Basin. The Permian Phosphoria Formation of the western United States, which exemplifies the miogeosynclinal setting, was deposited in a flexure zone between a stable shelf and the Cordilleran Geosyncline (Christie, 1978). The deposits are associated with organic-rich, black chert and argillaceous sediments, as well as carbonates, especially dolomite. The phosphate was deposited in a reducing environment relatively far from shore.

Rocks of this lithology, with variscite, occur in the Jillawarra Formation, which was deposited in a marine-shelf environment. Chert and carbonate within the Jillawarra attest to pauses in clastic sedimentation. The P_2O_5 content of the black shale and chert from the Jillawarra (Appendix C) ranges from below, to significantly above, the average levels for black shales, but is not as high as in the Phosphoria Formation. An unfavourable aspect of the Jillawarra Formation is that the main period of low clastic input (just prior to deposition of the Discovery Chert) corresponded to hypersaline rather than normal marine conditions. This situation may have been due to the barring of the basin from the sea, which would prevent upwelling oceanic water contributing phosphate to the basin. Therefore, more prospective areas may be near carbonate and chert zones low in the Jillawarra Formation, especially on ROBINSON RANGE and southwest MOUNT EGERTON.

Features of platform deposits are shown in the Georgina Basin in Australia. Phosphorite was deposited in Cambrian rocks close to shore and adjacent to basement highs that were inactive during sedimentation. The phosphorite is associated with chert, siltstone, and limestone. Current structures are common. Carbonaceous sediments show that some deposits (Duchess) were formed under reducing conditions (Cook, 1975); however, others (Lady Annie) are characterized by coquinitic cherts and light-coloured phosporite (Rogers and Keevers, 1975) and were presumably deposited in oxidizing conditions. Application of the platform model to the Bangemall Basin suggests prospective areas may be in zones of chemical and fine clastic sedimentation close to former basement highs. Present basement highs are probably close to the position of original basement highs, many of which were only mildly active during sedimentation and thus provide suitable areas for phosphatic sediments. The absence so far, of phosphates in these areas suggests the Jillawarra Formation may be the better prospect.

CONCLUSION

The Bangemall Basin contains minor deposits of copper, lead, zinc, gold, and manganese, but no major deposit of these metals or uranium has been discovered. The tectonic and sedimentation history provides some potential for sedimentary base-metal occurrences in the Bangemall Basin. However, in analogy with the models presented above, some desirable metallogenic factors did not operate in the Bangemall Basin. Deposition in the Bangemall Basin was as much as 400 m.y. after the last major tectonic event in the Gascoyne Province, and the Bangemall Group was deposited on a mature land surface. The basin was initiated by slow formation of horsts and graben together with a slow marine transgression, which allowed only small areas of continental sediments to form that would be suitable as a source of metals. Such a transgression also seems to have prevented the formation of local pockets of intensely reducing conditions adjacent to possible metal sources in continental sediments or basement rocks. The slow tectonic tempo during sedimentation possibly inhibited volcanism depriving the sediments of a ready source of metal ions.

The absence of basal sabkhas in this setting is puzzling, though the climate may have been too wet. Also, many sabkhas result from periods of regression, which encourage continental evaporitic conditions. The only major halt to transgression in the Bangemall Basin was in the Jillawarra-Discovery Chert interval which resulted in a hypersaline marine environment, well removed from continental sediments.

Many of the known metal occurrences in the Bangemall Basin are epigenetic and show a structural control. Thus, many of the metals now in fault zones were probably mobilized during the deformation of the basement that occurred simultaneously with faulting and folding of the cover. Therefore, the more promising areas for epigenetic mineral deposits are zones of strongly deformed Bangemall Group adjacent to strongly deformed basement.

A few metal occurrences are controlled by stratigraphy and thus affected by basement highs and basin hinge lines, and therefore indirectly influenced by movements on basement structures. Debris from a few basement highs did contribute to mineralization during deposition of the Jillawarra Formation. However the majority of sulphide has been precipitated in place (Davy, 1980) and the source cannot be determined. The significance of the occurrence of stratabound mineralization close to basin hinge lines is uncertain. Such hinge lines were controlled by basement faults. Mineralization may have been introduced via the faults into a suitable sedimentary environment. If so, the Jillawarra Formation may be prospective adjacent to basement faults which penetrate the cover, especially where syndepositional movement on the fault may have created locally deeper sections within the formation.

CHAPTER 9

Synthesis

TECTONIC CONTROLS ON THE BANGEMALL BASIN

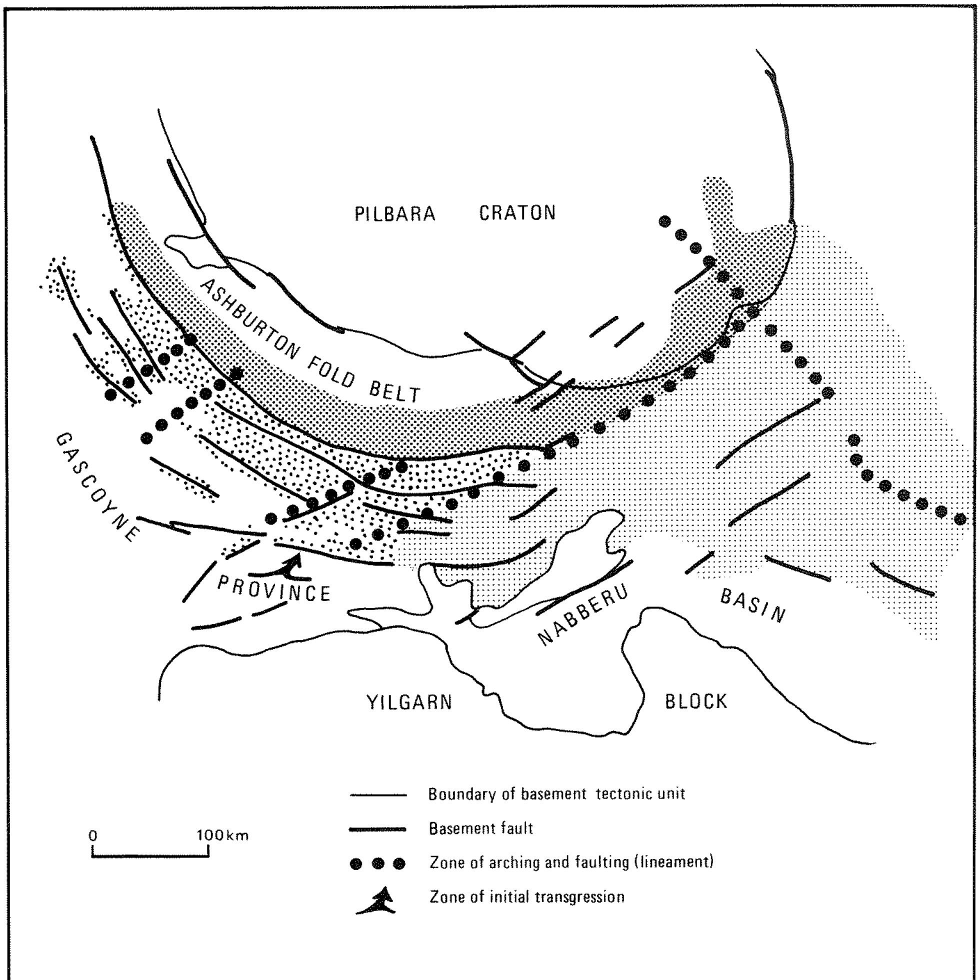
The Bangemall Basin (about 1 100 m.y.) was the last phase of activity in the orogenic belt which separates the Yilgarn and Pilbara cratons. It formed initially over two basement components, namely the Gascoyne Province and the Ashburton Fold Belt, and then spread to the east. The Gascoyne Province was formerly (2 000-1 600 m.y. ago), the locus of sedimentation, deformation, plutonic and metamorphic activity, during which the Wyloo Group was subjected to medium- to high-grade metamorphism. By contrast, in the Ashburton Fold Belt (formerly a major sedimentary trough), the Wyloo Group was folded but metamorphosed to only a low grade. The difference in tectonic style and activity between these two components of the orogenic belt influenced subsequent sedimentation and deformation events in this orogenic zone and was a major cause of contrasting sedimentary and structural styles of the three main structural zones of the Bangemall Basin—the Pingandy Shelf, the Edmund Fold Belt and the Bullen Platform.

About five hundred million years after the orogenic peak in the Gascoyne Province, the Bangemall Basin was initiated by a period of crustal extension and subsidence. Steeply dipping normal faults were generated by reactivation of old faults and shears. Crustal extension did not lead to separation of the blocks or to significant volcanism and the Bangemall Basin was established entirely on continental crust.

SEDIMENTATION

A stable marine shelf (Pingandy Shelf) developed on the Ashburton Fold Belt, which was a tectonically stable area at this time. By contrast, the Gascoyne Province was the site of an unstable marine basin that developed over a horst-and-graben terrain (Fig. 112). The boundary between the two zones was a faulted crustal warp, marked by faults which trend southeast to easterly.

The horst-and-graben terrain was composed of elongate blocks trending southeast to east-southeasterly and terminated by northeasterly trending faults, the most active of which lay along the Tangadee and Flint Hill Lineaments. The uplifted horst blocks were eroded concurrently with sedimentation and so a subdued topography was produced. Most grabens contained slowly accumulating shallow-water sediments and only adjacent to the more active horsts were alluvial fans developed. In the deeper levels of the fault zones associated with the Tangadee Lineament, small quantities of melt, which extruded as rhyolite flows, were produced.



BASEMENT TECTONIC STYLE	LITHOSTRATIGRAPHIC UNITS : (subgroup unless stated otherwise)	SEDIMENTARY UNIT STYLE	COVER PROVINCE
SHELF	Mucalana; Edmund; Top Camp Formation; Collier; Manganese	CONTINUOUS, SIMPLE UNITS : carbonate, shale, sandstone	PINGANDY SHELF
HORST AND GRABEN	Edmund	DISCONTINUOUS, COMPLEX, INTERFINGERING UNITS : shale, sandstone, carbonate, conglomerate	EDMUND FOLD BELT
PLATFORM	Collier; Diebil; Kahrban	CONTINUOUS, SIMPLE UNITS sandstone, minor shale LOCALLY COMPLEX UNITS ONLY IN KAHRBAN SUBGROUP : shale, sandstone	BULLEN PLATFORM

GSWA18927

Figure 112. Relation between tectonic features of the basement and the style of sedimentary units and structural provinces in the cover.

Marine transgression started in a complex graben which was bounded by the Tangadee and Flint Hill Lineaments. During subsidence there was slow transgression to the east and northwest, during which terrigenous sediments were derived from the Gascoyne Province to the west of the basin. The source area was of low relief and so the supply of sediment was insufficient for deltas to build, but sufficient detritus was transported by marine currents along the coast to form barrier-bar sands. Landward of the barrier-bars, little terrigenous sediment was deposited, thus enabling precipitation of carbonate in lagoons and tidal flats. Seawards of the barrier-bar, clastic muds were laid down on a marine shelf. With the rise in sea level the various lagoon, barrier-bar and shelf facies transgressed the horst-and-graben terrain. Where the land surface was gentle, the complete sequence of lagoon - barrier-bar - shelf facies was deposited, but where the slope of the basement surface was steeper, lagoons did not develop and the sands were deposited as beaches. Except for small alluvial fans fringing horsts, terrigenous sandstone and shale were largely confined to grabens in the basin, while on the gentle rises of submerged horsts the supply of terrigenous sediment was low and lagoonal carbonate deposited. Along the Tangadee Lineament, faulting and warping was sufficient to produce quantities of sediment too great to be removed by marine currents and a sandy delta grew westward into the adjacent graben.

With further transgression to the northwest, faulting and warping along northeasterly trending lineaments divided the horst-and-graben terrain by a series of basement arches which influenced later sedimentation. Thus uplift along the Flint Hill Lineament supplied detritus for sandstone and conglomerate which extended into shelf muds of the Jillawarra Formation. Also sand and silt sequences are thinner over rising basement arches than in the adjacent grabens. By the time the shelf muds had transgressed the barrier sands and lagoonal carbonates, most basement irregularities had been fully covered by sediment.

Extensive tidal flats and lagoons then developed on a gently south-sloping marine shelf at a level above the horst-and-graben terrain. However, along the Tangadee Lineament and the southern edge of the basin, basement horst ridges continued to be active and remained as islands and shoals that formed a barrier between the basin and the open seas, which presumably lay to the south and east. With the cessation of tectonic activity, only fine terrigenous material was carried into the basin. Tides and currents were weak and unable to redistribute the material through the basin and so it accumulated along the west and southwest margins of the basin. The western basin at this stage became stagnant and hypersaline; and the pelagic micro-organisms died, resulting in acid reducing conditions. In the absence of both terrigenous material and silica-extracting organisms, silica gel precipitated to form the Discovery Chert. The chert contains minor silt, clay, carbonate and gypsum where the water was shallower along the southwestern and eastern margins of the horst-and-graben terrain.

Silica precipitation ended when tectonic movements in the basement enabled water from the open sea to circulate through the basin, initiating a shallow marine platform with shoals of carbonates, clastics, and mudflats. The western basin was then subjected to a second phase of crustal extension which resulted in subsidence along the zone separating the Ashburton Fold Belt and the Gascoyne Province. For the first time areas to the northwest and north of the basin contributed large amounts of clastic sediment to both the unstable marine basin and the marine shelf, and a build-up of deltas resulted. The prodelta mud of the Curran Formation spread east and in turn was overlain by turbidites of the Coodardoo Formation, which were generated by slumping down the delta front. The extension of the crust also allowed injection of large volumes of tholeiitic magma into the sediments.

The rate of subsidence of the western basin then diminished and produced a marine-shelf environment on which was deposited mud and at times sand from turbidity currents.

Meanwhile, during deposition of the lagoon—barrier-bar— marine-shelf facies in the western basin, the sea transgressed north on to the Sylvania Dome and the eastern edge of the Pilbara Craton, which formed the eastward continuation of the marine shelf on the Ashburton Fold Belt. Both the shelf and the platform provided a basement of low relief on which a sequence of mud transgressed beach sands and spread eastward and northeast eventually lensing out just north of the Carnarvon Range and west of the basement highs on GUNANYA and TRAINOR.

Deposition of mud on the platform continued during the stillstand in the western basin until a marine-shelf environment was established over all the basin west of the Cornelia-Balfour zone of basement arches. East of the zone of basement arches, a sandy sequence was deposited on the unconformity. Only in the Skates Hills district were alluvial fan, beach, and intertidal (stromatolitic) carbonate sequences developed. The sandy facies on the eastern margin (McFadden Sandstone) was deposited as subtidal sand tongues and shoals on the seaward side of tidal flats. Volcanic activity of unknown source contributed a tuffaceous component to some beds in the east of the basin.

Along the southeastern edge of the platform there developed a west-northwesterly trending zone, tectonically more active than the rest of the platform. A rapid marine transgression, which left no sedimentary record, was followed by a stillstand during which chert and shale at the base of the Kahrban Subgroup were deposited in shallow water. Sediment eroded from land uplifted in the southeast was then deposited as a delta which advanced to the northwest. Ultimately, settling of the delta led to the area being covered by subtidal sands (Calyie Sandstone), which to the north lapped on to the subaerial Cornelia Inlier. To the northwest, in the Robertson Range—Jiggalong area, sandy deltas supplied by erosion off the Pilbara Craton advanced southeast on to the platform.

Regression of the sea from east to west across the platform brought the eastern sandy sequence to cover the marine-shelf silts and clays of the Backdoor Formation. The platform now formed a tide-swept sandy shelf similar in setting to the present day North Sea. With further retreat of the sea, sedimentation east of the 120° meridian ceased, leaving the fluvial Durba Sandstone on top of the progradational sequences.

The western half of the basin, where muds of the Fords Creek Shale were being laid down on the shelf concurrently with sand on the platform, was also affected by the regression. The sedimentary response was the reverse of the basal succession. A barrier-sand system migrated over the shelf deposits as the sea withdrew, succeeded by sediments of the adjacent lagoons. These coastal lagoons were well endowed with terrigenous mud, but some limestones were also formed. This environment is the youngest now preserved in the western facies province.

DEFORMATION

The tectonic activity displayed by the three main basement segments during deposition increased in tempo during deformation of the basin. The Pingandy Shelf overlies a tectonically inert basement, and did not participate in the deformation affecting the Edmund Fold Belt which overlies the Gascoyne Province. The northeast part of the shelf overlay a stable basement of the Pilbara Craton (and its stable cratonic cover—the Hamersley Group), the edge of which is marked by the Robertson Fault System. Here, Bangemall sediments were deformed by faults and open folds due to block faulting in the basement.

The stable marine shelf forming the eastern part of the basin became the Bullen Platform. Here, broad open folds (many of which were domes and basins) formed by the draping of sediments over basement blocks controlled by northeast and northwest-trending faults. The northeasterly trend is dominant, except for a major zone of faulting and uplift which produced a series of *en echelon* arches (Cornelia-Boondawarri-Balfour) trending northwesterly into the Hamersley Basin. The horst-and-graben terrain which developed on active basement became the Edmund Fold Belt. North of the Flint Hill Lineament, south easterly shear zones were dominant in the basement and interacted with a more widely spaced, subordinate northeasterly set. East of the lineament, northeast and easterly trending shear zones controlled the deformation. In the Edmund Fold Belt, the horst-and-graben blocks had continued to move on normal faults, stretching the overlying sediments to form drape folds. The crustal shortening in the basement, possibly related to movement of the Gascoyne Province towards the Pilbara Craton, caused lateral shortening of the cover by squeezing the fold belt against the Pingandy Shelf. Strain was concentrated in ductile zones of metasediments in the basement and along faults forming basement-block boundaries. The metasediments were incompetent in comparison with the intervening blocks of crystalline rocks and so became folded together with the cover. Crystalline rocks underwent no shortening, unlike their sedimentary cover. Normal faults forming the boundaries of some basement blocks were reactivated and became the sites of reverse faults. With continued lateral shortening, ductile zones were squeezed between rigid blocks and rode up to the north, producing reverse faults in the cover. Total shortening across the western part of the basin is about 19 km. Such a small movement is compatible with the palaeomagnetic data of McElhinny and Embleton (1976) which indicates that the Pilbara and Yilgarn Blocks have maintained their relative positions since stabilization of the Archaean cratons.

Finally, a period of northwest-southeast brittle extension, saw the emplacement of basic dykes on northerly to easterly trends throughout the region.

LATE GEOLOGICAL HISTORY

The subsequent history has been largely one of erosion. There was Permian glacial activity in the east (Brakel and Leech, 1979) and the west (Grey and others, 1977), which may once have extended over the whole region, but most of these glacial deposits, and any Mesozoic sediments of the Canning and Officer Basins which extended westwards, have now been stripped off. Superficial units accumulated as a veneer during the Tertiary and Quaternary periods, and these are now being dissected west of the inland drainage divide.

APPENDICES

APPENDIX A

Stromatolites And Other Organic Remains In The Bangemall Basin

by

KATHLEEN GREY

INTRODUCTION

A study of fossil material from the Bangemall Basin was undertaken in an attempt to assess the palaeoecological and biostratigraphical potential of stromatolites and other organic remains. Stromatolites are abundant in the dolomitic carbonates of the Bangemall Group, although their taxonomic diversity is low (Grey, 1982). They are present in the lower part of the succession in the western and northern facies, and are particularly common in the Irregully Formation. They also occur in the Top Camp Dolomite, Kiangi Creek Formation and Devil Creek Formation. A stromatolitic dolomite is present near the top of the Skates Hills Formation, a basal unit of the eastern facies. Isolated stromatolitic outcrops of possible Bangemall Group are known from the Gascoyne Province.

Poorly preserved organic material occurs in chert and silicified siltstone in several formations. Various structures of possible organic origin, now altered to such an extent that their original nature is difficult to determine, are present in many samples, especially in the Coobarra Formation, Irregully Formation, Kiangi Creek Formation, Jillawarra Formation, Discovery Chert, Fords Creek Shale and Backdoor Formation. Abundant amorphous carbon of probable organic origin was released by maceration of some samples in hydrofluoric acid, and spheroids of uncertain origin have been obtained from a few samples. A fairly diverse assemblage could possibly be obtained by more selective sampling. However, contamination cannot be ruled out for these samples (Cloud and Morrison, 1979), and the results reported here should be regarded as a preliminary indication of areas in which more rigorous studies could be carried out.

Possible calcareous chlorophytes from the Devil Creek Formation and discoid megafossils from the Irregully Formation were described by Marshall (1968). No similar material was found during the present study.

PREVIOUS STUDIES

A history of stromatolite studies is given by Walter (1972), and by several authors in Walter (1976). The presence of stromatolites in the Bangemall Group seems to have gone unnoticed until systematic regional mapping was undertaken by the Geological Survey. They are first recorded by Halligan and Daniels (1964) and figured by Daniels (1966).

Preliminary results of biostratigraphic studies of stromatolites in Australia (including some from the Bangemall Basin) were published by Glaessner and others (1969). In 1972, Walter published the first systematic studies of Bangemall stromatolites. Further studies (Grey, 1977a; 1978a) on the Bangemall Group have now been carried out. The biostratigraphic distribution of stromatolites in the Bangemall Basin and their relationships to other stromatolite occurrences in Western Australia are discussed by Grey (1979; 1982). Possible Bangemall Group outliers containing stromatolites have also been reported (Grey, 1977b; van de Graaff and others, 1977).

Microfossils were first discovered in the Bangemall Group by J. Schupp (Daniels, 1975) and were studied by Marshall (1968). Marshall noted the presence of abundant acritarchs in the Irregully Formation, the Jillawarra Formation, and the Discovery Chert. He concluded that the acritarchs were of probable planktonic origin because of their wide distribution.

Marshall (1968) also reported the presence of discoid megafossils of possible medusoid affinities in the Irregully Formation. In addition, he recorded several types of calcareous microfossils from the Devil Creek Formation, which he believed showed affinities with complex chlorophytes. Marshall's localities were not revisited during the present studies, but a brief examination was made of his original material, which is housed in the Princeton University collection.

Marshall's findings suggested the presence of organisms of much greater complexity than are generally believed to have existed at this time (Schopf and others, 1973). However, some of the specimens show a striking resemblance to *Microcodium* described by Esteban (1974) from an Eocene caliche from Spain, and are of doubtful Proterozoic age. The structures described as medusoids by Marshall are most probably of non-biogenic origin and may well be derived from gypsum rosettes. Other samples contain concentrically-banded, oval structures of considerable morphological complexity. These appear to be biogenic and more detailed examination is required to determine their nature. The significance of the Bangemall Group structures is therefore uncertain. No evidence of similar material was observed in the GSWA collection.

METHODS OF STUDY

Stromatolite samples were collected from approximately 60 localities, either by the author, or by members of the Bangemall Basin mapping parties. Methods of study, both field and laboratory, are basically those outlined by Walter (1972) and Preiss (1976a.)

The literature on stromatolitic carbonates is extensive, two recently published books (Ginsburg, 1975; Walter, 1976) present comprehensive reviews of various aspects of the subject, and a two-part bibliography has been published (Awramik and others, 1976; Awramik and others, 1979).

In addition to the study of stromatolites, 87 cherts, silicified shales and other silicified rocks were examined in thin section to determine their organic content and selected samples were macerated with hydrofluoric acid. Few samples were collected specifically for palaeontological examination, and better preserved material could probably be obtained by more selective sampling.

PRESERVATION AND COMPOSITION OF THE BIOTA

Stromatolites and other organic remains occur mainly in the western part of the Bangemall Basin and are mostly absent from the more arenaceous facies to the east, with the exception of a carbonate unit in the Skates Hills Formation. Stromatolites are most common in the Irregully Formation and its lateral equivalents. Although stromatolites are numerically abundant, their taxonomic diversity is low (Grey, 1982). Many are stratiform or domed and are of little value for stratigraphic purposes. Poor preservation and extensive silicification, particularly along the western margins of the basin, prevents the identification of many branching columnar forms.

Baicalia capricornia Walter (Fig. 113) is the most common of the branching columnar forms and is present throughout the Irregully Formation, and ranges upwards into the Devil Creek Formation. It is also present in the Kiangi Creek Formation, and has recently been collected from the Mulya Dolomite in the adjacent Gascoyne Province. *Conophyton garganicum australe* Walter (Fig. 114) occurs in the lower part of the Irregully Formation and in the Top Camp Dolomite.

Conical structures are present along the western margin of the basin but they are poorly preserved and highly silicified. Although they were originally considered to be forms of *Conophyton*, (Grey, 1979; 1982) more recent investigations, particularly of those occurring in the Milly Springs area, suggest that some of them are tepee structures, while others are draped stromatolitic structures similar to those described by Vlasov (1977).

A new stromatolite form, consisting of straight, unbranched, club-shaped columns (Figs 115 and 116), was collected from the middle part of the Irregully Formation in Irregully Gorge, but has not yet been described. Poorly preserved *Conophyton garganicum australe* occurs in a possible Bangemall outlier near Hectors Bore (Grey, 1977b).

Acaceilla cf. *australica* occurs in the Skates Hills Formation (Grey, 1978a), which crops out on the eastern margin of the basin.

Amorphous carbon of probable organic origin, and poorly preserved structures of doubtful organic origin, are present in the Coobarra Formation, Irregully Formation, Kiangi Creek Formation, Jillawarra Formation, Discovery Chert, Ullawarra Formation, Fords Creek Shale, and the Backdoor Formation. Marshall (1968) reported the occurrence of acritarchs in the Irregully, Jillawarra, Discovery Chert, and Devil Creek Formations. None of the material examined in this study seems to be as well-preserved as that described by Marshall. In many of the samples the carbonaceous material occurs as minute particles which are either finely disseminated throughout the sample, or are concentrated along crystal boundaries. In some samples opaque spheres are present, but they are usually poorly preserved and have diffuse boundaries indicating recrystallization. Most of these spheres range in size from 5-8 μm but larger, very diffuse spheres up to 30 μm are also present.

Although no identifiable microfossils are present in any of the thin sections examined, some problematic structures occur which are believed to have developed from original organic material. These include grey or brown spheres with radiating or concentric structures, (Fig. 117a), trails of diffuse carbonaceous material (Fig. 117B), and 'pods' of coarse silica crystals (Fig. 117C), which are bounded by concentrations of organic material, and in which each quartz grain has an opaque carbonaceous spheroid at its centre. These pods may have developed as a result of the differential crystallization of silica around microbial filaments, similar to those described from the Waterton Formation by Oehler (1976). In the specimens from the Bangemall Group, recrystallization has progressed further than that described from the Waterton Formation. Organic material has moved along the front of crystallization to form a structure which has 'walls' and contains spore-like bodies.

In some samples, spheres and cubes of iron oxide are present. These occasionally occur linked in chains and have a similar mode of preservation to material in the 1 500 m.y. old H.Y.C. Pyritic Shale Member of the Barney Creek Formation in the McArthur Group of northern Australia (Oehler, 1977).

Problematic tubular structures with terminal spheroids (Fig. 117D-F) are abundant in a sample from the Discovery Chert (Grey, 1978b). M. R. Walter (pers. comm.) pointed out their resemblance to the trails of ambient pyrite grains (Tyler and Barghoorn, 1963; Knoll and Barghoorn, 1974). Although the Discovery Chert structures have some features in common with ambient grain trails, they exhibit several puzzling features, and the origin of the structures remains doubtful (Grey, 1978b).

Selected samples from several formations were macerated, and in most cases the residues were found to contain abundant amorphous carbon. Some samples from the Discovery Chert, and in particular from the Backdoor Formation, contain organic material of spheroidal or oval shape (Fig. 117G-I), some of which have thick walls and granular to reticulate ornament. Some specimens show budding and others occur as diads or tetrads. The affinities of these micro-organisms have not been determined, and, as has already been discussed, it is possible that these structures are contaminants. They are however, recovered in great abundance in some samples, and this suggests that further investigation should be carried out.

STRATIGRAPHIC DISTRIBUTION

COOBARRA FORMATION

A single specimen of brown chert with abundant silica spheres contained finely disseminated carbon particles and small opaque spheres. 'Pod' structures are common.

IRREGULLY FORMATION

The Irregully Formation, a major carbonate unit in the western facies of the Bangemall Basin contains abundant stromatolitic beds, particularly in the Parry Range, Henry River and Irregully Creek areas.

Stromatolites are common in the carbonates of the Parry Range area, but preservation is usually poor, due to considerable alteration and surface silicification. At many localities, only the overall shape and broad banding remain to indicate an original stromatolitic origin. At some localities silicified bands occur in close proximity to unaltered carbonate and laminae can be traced across the boundary. Few laminae are preserved where silicification has occurred. Instead, the silicified portions of the stromatolites contain broad bands which apparently follow the original laminations, but which terminate in sharp crests. The reasons for the difference between the two lithologies is not clear, but some of the V-shaped terminations may be of secondary origin, and are probably tepee structures. Others may resemble the stromatolitic structures described by Vlasov (1977).

Columnar and cone-shaped stromatolites may be present at some localities, but because of poor preservation it is difficult to distinguish them from forms with a crested or V-shaped termination, or from deformation structures. In the northern Parry Range area identification of the columnar forms is not possible. Very little can be determined about the environment of deposition, other than to suggest that it was most probably intertidal.

In the southern part of the Parry Range there is greater development of columnar and conical forms although deformation and silicification are also noticeable in this area. Extensive outcrops occur in the vicinity of Wongida Well and Milly Springs and require more detailed study. Silicification in this area has destroyed many details of lamination, making identification difficult. In particular, it has not been possible to make a detailed study of crested and V-shaped forms which occur in this area. Club-shaped columnar stromatolites, resembling an undescribed form from Irregully Gorge, are found in unsilicified carbonate.

In the Milly Springs area there is an apparent development of cyclic sequences comprising a basal unit of intraclastic or, more rarely, a possibly oncolitic dolomite, overlain by a dark blue-grey limestone containing club-shaped, columnar stromatolites. This passes upwards into a silicified unit containing flat-laminated and crested stromatolitic banding. Similar association of stromatolite-rich sequences have been described from the Proterozoic of the Canadian Shield (Hoffman, 1976) but further studies are necessary before more detailed comparisons can be made between the Irregully Formation and the cycles described by Hoffman.

The columnar stromatolites most probably formed in quiet water, lagoonal conditions, while the silicified, crested stromatolites may indicate a shallow-water intertidal to possibly supratidal phase.

The type locality of *Baicalia capricornia* Walter, is an extensive outcrop near the Henry River East Branch. *Baicalia capricornia* also occurs in a second outcrop near the type locality, and at an outcrop near Pindanni Bore, although preservation at these localities is poorer than at the type locality. Stromatolite formation occurred in a series of overlapping bioherms, probably in shallow subtidal or lagoonal environments. Other outcrops in this general area contain only poorly preserved, non-columnar types.

Approximately 1 300 m of Irregully Formation is exposed at Irregully Gorge and upstream of the gorge along the banks of Irregully Creek. Much of the succession is stromatolitic. Preservation is remarkably good in some parts of the gorge, and this area would merit further study.

Walter (1972) recorded *Conophyton garganicum australe* and *Baicalia capricornia* from near the base of the Irregully Formation at the north end of the gorge. Flat-laminated, undulatory and pseudocolumnar forms occur commonly throughout the sequence; and, at the southern end of the gorge, an unnamed form with unbranched, club-shaped columns occurs. *Baicalia capricornia* is present near the top of the Irregully Formation in the vicinity of the White Cliffs.

A somewhat variable environment of deposition is indicated. Flat-laminated stromatolitic banding probably indicates upper intertidal to supratidal conditions while *Baicalia capricornia* probably grew in moderately high-energy conditions in the shallow subtidal to intertidal regime. Deeper water and quieter conditions are indicated by the presence of *Conophyton garganicum australe* and the club-shaped form.

Miscellaneous outcrops of stromatolites in the Irregully Formation occur at Coolinbar Hill, near Peedawarra Flats, and on Mulgul and Woodlands stations. Some of these localities have columnar forms, which can be tentatively identified as *Baicalia capricornia* Walter, and show evidence of fairly high-energy conditions at the time of deposition since interspace filling frequently consists of platey intraclasts, often stromatolitic in origin. Other stromatolites are either non-columnar or too poorly preserved to allow identification. A shallow, subtidal to intertidal environment, probably with a moderately high-energy regime is indicated.

The discoid megafossils of possible medusoid affinity reported by Marshall (1968) were collected from a locality in the Irregully Formation 27.2 km south-southwest of Ashburton Downs homestead. This locality was not examined in the present study and no comparative material was found elsewhere.

Two samples of chert from the Irregully Formation contained finely disseminated carbon particles, but no identifiable microfossils. Marshall (1968) reported the presence of acritarchs, but only rare, poorly preserved opaque spheres were observed in the present study. Spheres and cubes of iron oxide are also present.

KIANGI CREEK FORMATION

Only one stromatolite locality has been reported from this formation and this contains a columnar form tentatively identified as *Baicalia capricornia*. A single chert sample contains abundant, finely disseminated carbon particles, but no identifiable microfossils.

JILLAWARRA FORMATION

Samples from this formation contain abundant, finely disseminated carbon particles and, occasionally, poorly-preserved, opaque spheres with an average diameter of 6 μm. Cubes and spheres of iron oxide, which may have replaced organic material, are also present.

DISCOVERY CHERT

Many of the samples examined from this formation have undergone considerable recrystallization, and although abundant amorphous carbon is present in most samples, no identifiable microfossils were observed in thin section. The carbon particles may be either finely disseminated throughout the sample, or concentrated around crystal boundaries. In some samples the carbon is concentrated in lenses. Small, dark spheres, often showing concentric or radiating structures indicative of recrystallization, are sometimes present. Iron oxide particles, sometimes in the shape of spheres and cubes, which are possibly replacing organic material, are common and may be linked together in chains. Elongate 'pods' are common in some samples. Some poorly laminated brown cherts contain large patches of fine-grained, brown material concentrated in spheres which may be similar to the elongate pod-shaped structures in black cherts.

A sample from near Coobarra Creek contains tubular structures of uncertain origin resembling ambient pyrite grains. Laminated structures resembling stromatolites occur in one or two samples, but the laminae are more regular than is usual for stromatolites, and they are composed of clay particles, suggesting an inorganic origin. Abundant amorphous carbon, and spheroids of variable size, ranging up to 16 μm in diameter were recovered by maceration from brown chert. The affinities of these spheres have not been determined and the possibility of contamination cannot be ruled out.

DEVIL CREEK FORMATION

Only one stromatolite locality with *Baicalia capricornia* has been recorded from the Devil Creek Formation, a locality near White Cliffs in Irregully Creek. Marshall (1968) reported calcareous microfossils in the Devil Creek Formation, at a locality near Cobra homestead, and from two other localities where material is poorly preserved. Several different types of microstructure are present and Marshall suggested that they showed affinities with structurally complex chlorophytes. Marshall's localities were not visited during the present study, and no similar material was collected from other localities.

ULLAWARRA FORMATION

Samples from this formation were strongly recrystallized and contained no evidence of organic material, with the exception of a single chert sample containing abundant amorphous carbon.

FORDS CREEK SHALE

A fine-grained, laminated, brown chert was examined and found to contain diffuse carbon particles and small dark spheres of possible biogenic origin. Iron oxide spheres, and chains of spheres also occur. Laminations in the chert are wavy-banded and have the typical alternating light and dark banding associated with stromatolites.

TOP CAMP DOLOMITE

Stromatolites have been recorded from Fords Creek (Daniels, 1968; Walter, 1972) and from near Pingandy Creek. At Fords Creek *Conophyton garganicum australe* Walter occurs about 40 m above the base of the succession in the Tchintaby Dolomite Member. Very large forms of *Conophyton garganicum australe* occur in a dolomite unit of uncertain stratigraphic position between the Bearuroo Dolomite Member and the Pingandy Dolomite Member near Pingandy Creek.

An extensive outcrop of large conical forms is found at this locality, probably representing a series of bioherms, consisting of several successive layers of cones with an average maximum observed diameter of 4 m.

Following Donaldson's (1976) suggestion for calculating the relief of individual cones the minimum relief of the Pingandy specimens would have ranged from 3 m to 9 m. Donaldson suggests that the formation of such large cones could only have occurred in quiet, relatively deep-water conditions by bacterial precipitation. At Irregully Creek and Fords Creek, the specimens are smaller than at Pingandy Creek, and the water depth may have been correspondingly shallower.

Other stromatolite localities in the Top Camp Dolomite contain only flat-laminated or domed forms.

BACKDOOR FORMATION

Samples of the Backdoor Formation consist of a fine-grained, poorly laminated, brownish chert containing abundant spheres with a high birefringence, probably jarosite. Reddish-brown and dark-grey spheres are also present, and may be of organic origin, but this cannot be demonstrated in thin section.

Abundant acritarchs of uncertain affinities were obtained by maceration. They consist of spheroids and ovoids of variable size and shape. A large number of spheriods are from 6-10 μm in diameter, but others may be up to 30 μm in diameter. Most specimens have thick walls and may have granular to reticulate ornament. The acritarchs may occur singly or in groups. Some spheres have been broken, and appear hollow. These structures may well be contaminants (Cloud and Morrison, 1979).

SKATES HILLS FORMATION

A stromatolitic dolomite occurs near the top of the Skates Hills Formation, a basal unit of the Bangemall Group in the eastern facies, (Williams and others, 1976). Specimens from two localities consist of extremely well-preserved columns of *Acaciella* cf. *australica* (Grey, 1978a). Specimens from a third locality consist of cumulate forms, which may be basal portions of the columnar forms, but which are not sufficiently developed for accurate determination. A moderately high-energy environment of deposition is indicated by the lateral elongation of the columns in one of the samples to form structures resembling ripple marks.

PROBABLE BANGEMALL GROUP OUTLIERS

An outlier of probable Bangemall Group occurs near Hectors Bore on GLENBURGH (Williams and others, 1980), and contains *Conophyton* and stratiform stromatolites (Grey, 1977b) with some development of tepee structures. Preservation of the *Conophyton* is too poor for the form to be determined, but it has affinities with *Conophyton garganicum australe*.

In the Uaroo Group, to the west of the Bangemall Basin in the Yanrey area, the Mulya Dolomite and the Tinkers Dolomite have both been described as algal dolomites (van de Graaff and others, 1977). Deformed and silicified bioherms of

Baicalia capricornia are present in the Mulya Dolomite supporting a possible correlation between the Uaroo Group and the Bangemall Group (van de Graaff and others, 1977).

BIOSTRATIGRAPHY

Grey (1979, 1982) has summarized the data on biostratigraphic relationships of stromatolites in Western Australia. *Baicalia capricornia* and *Conophyton garganicum and australe* are known only from the Bangemall Group. The group *Baicalia* has a range from 1 350 to ?900 m.y. (Preiss, 1976) while *Conophyton* occurs in the late Archaean and throughout the Proterozoic and in early Cambrian (Krylov and Semikhatov, 1976). However, the form *Conophyton garganicum* is thought to be restricted to the period from 1 600 m.y. to about 1 000 m.y.

The previously known range of *Acaciella* is from less than 1 000 m.y. to early Cambrian (Preiss, 1976). *Acaciella* cf. *australica* from the Skates Hills Formation is probably older than this (Grey, 1978a) however, the age of the Skates Hills Formation is uncertain, although it is correlated with rocks younger than the 1 100 m.y. stromatolitic sequence in the western part of the Bangemall Basin (Williams and others, 1976).

The potential of microfossils for biostratigraphy is still a matter for debate although several important trends have been recognized in recent years (Schopf, 1977). The poor preservation and possible contaminant origin of the Bangemall Group material makes detailed comparisons with other assemblages difficult, and no conclusions are possible at present.

CONCLUSIONS

Stromatolites are abundant in the western part of the Bangemall Basin and probably played an important role in the deposition of the lower part of the Bangemall Group. Taxonomic diversity is low and the forms present are known only from the Bangemall Basin and outliers of probable Bangemall age. The environment of deposition was mainly shallow subtidal to intertidal. Poorly preserved material of possibly biogenic origin is present in cherty sediments in the Bangemall Group. Better preserved material could probably be obtained by more selective sampling, and the present results are inconclusive because of possible contamination.

REFERENCES

Awramik, S.M., Haupt, A., Hofman, H.J, and Walter, M.R., 1979, Stromatolite Bibliography 2: Precambrian Research, 9, p. 105-166.

Awramik, S.M., Hofman H.J. and Raaben, M.E., 1976, Bibliography *in* Walter M.R. (ed.), Developments in Sedimentology, 20: Stromatolites: Amsterdam, Elsevier, p. 705-771.

Cloud, P., and Morrison, K., 1979, On microbial contaminants, micropseudofossils, and the oldest records of life: Precambrian Research, 9, p. 81-91.

Daniels, J.L, 1966, Revised stratigraphy, palaeocurrent system and palaeogeography of the Proterozoic Bangemall Group: West. Australia Geol. Survey Ann. Rept 1965, p. 48-56.

————1968, Turee Creek, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

————1975, Bangemall Basin, *in* Geology of Western Australia: West. Australia Geol. Survey Mem. 2, p. 147-158.

Donaldson, J.A., 1976, Paleoecology of Conophyton and associated stromatolites in the Precambrian Dismal Lakes and Rae Groups, Canada, *in* Walter, M.R. (ed.), Developments in Sedimentology, 20: Stromatolites: Amsterdam, Elsevier, p. 523-534.

Esteban, M., 1974, Caliche textures and 'Microcodium': Soc. Geol. Italiana, Bollettino 91 (Suppl. 1973), p. 105-125.

Ginsburg, R.N., (ed.), 1975, Tidal Deposits: a casebook of recent examples and fossil counterparts: New York, Springer.

Glaessner, M.F., Preiss, W.V., and Walter, M.R., 1969, Precambrian columnar stromatolites in Australia, morphological and stratigraphic analysis: Science, 164, p. 1056-1058.

Grey, K., 1977a, Middle Proterozoic stromatolites from the Bangemall Basin, Western Australia: West. Australia Geol. Survey Palaeont. Rept 45/77 (unpublished).

————1977b, Stromatolites and cryptalgal laminate from Hectors Bore, Glenburgh 1:250 000 Sheet: West. Australia Geol. Survey Palaeont. Rept 59/77 (unpublished).

———— 1978a, *Acaciella* cf. *australica* from the Skates Hills Formation, eastern Bangemall Basin, Western Australia: West. Australia Geol. Survey Ann. Rept. 1977, p. 71-73.

——— 1978b, Problematic microstructures in the Discovery Chert, Bangemall Group, W.A.—ambient grains or microfossils?: West. Australia Geol. Survey Palaeont. Rept. 61/78 (unpublished).

———— 1979, Preliminary results of biostratigraphic studies of Proterozoic stromatolites in Western Australia: West. Australia Geol. Survey Record 1979/2.

————1982, Aspects of Proterozoic stromatolite biostratigraphy in Western Australia: Precambrian Research, 18, p. 347-365.

Grey, K., van de Graaff, W.J.E., and Hocking, R.M., 1977, Precambrian stromatolites as provenance indicators in the Permian Lyons Formation, Carnarvon Basin: West. Australian Geol. Survey Ann. Rept. 1976.

Halligan, R., and Daniels, J. L., 1964, The Precambrian geology of the Ashburton valley region, North-West Division: West. Australia Geol. Survey Ann. Rept 1963.

Hoffman, P., 1976, Environmental diversity of Middle Precambrian stromatolites, *in* Walter, M. R. (ed.), Developments in sedimentology, 20: Stromatolites: Amsterdam, Elsevier, p. 599-611.

Knoll, A. H., and Barghoorn, E. S., 1974, Ambient pyrite in Precambrian chert: New evidence and a theory: U.S.A. Nat. Acad. Sci. Proc., p. 2329-2331.

Krylov, I. N., and Semikhatov, M. A., 1976, Table of time-ranges of the principal groups of Precambrian stromatolites, *in* Walter, M. R. (ed.) Developments in sedimentology, 20: Stromatolities: Amsterdam, Elsevier, p. 693-694.

Marshall, A. E., 1968, Geological studies of the Proterozoic Bangemall Group, N. W. Australia: Princeton University Ph. D. thesis (unpublished).

Oehler, J. H., 1976, Possible silicified algae from the Late Precambrian Waterton Formation of Canada: Jour. Paleontology, v. 50, p.778-782.

————1977, Microflora of the H.Y.C. Pyritic Shale Member of the Barney Creek Formation (McArthur Group) middle Proterozoic of northern Australia: Alcheringa, v.1, p. 315-349.

Preiss, W. V., 1976a, Basic field and laboratory methods for the study of stromatolites, *in* Walter, M. R. (ed.), Developments in sedimentology, 20: Stromatolites: Amsterdam, Elsevier, p. 5-14.

————1976b, Intercontinental correlations, *in* Walter, M. R. (ed.), Developments in sedimentology, 20: Stromatolites: Amsterdam, Elsevier, p. 359-370.

Schopf, J. W., 1977, Biostratigraphic usefulness of stromatolitic Precambrian microbiotas a preliminary analysis: Precambrian Res., v. 5, p. 143-173.

Schopf, J. W. Haugh, B. N., Molnar, R. E., and Satterthwait, D. F., 1973, On the development of metaphytes and metazoans: Jour. Paleontology, v. 47, p. 1-9.

Tyler, S. A., and Barghoorn, E. S., 1963, Ambient pyrite grains Precambrian cherts: American Jour. Sci., v. 261, p. 424-434.

van de Graaff, W. J. E., Denman, P. D., Hocking, R. M., Baxter, J. L., 1977, Explanatory notes on the Ningaloo-Yanrey 1:250 000 Geological Sheet, W.A.: West. Australia Geol. Survey Rec. 1977/9.

Vlasov, F. Ya., 1977, Dokembriiskie stromatolity iz satinskoi svity Yuzhnogo Urala. (Precambrian stromatolites from the Satka Formation in the Southern Urals), *in* Materialy po paleontologii srednego paleozoya Urala i Sibiri: Sverdlovsk. Inst. Geol. Goekhim. Tr. 126: p. 101-128.

Walter, M. R., 1972, Stromatolites and the biostratigraphy of the Australian Precambrian and Cambrian: Palaeontological Association London, Spec. Pap. in Palaeontology no. 11.

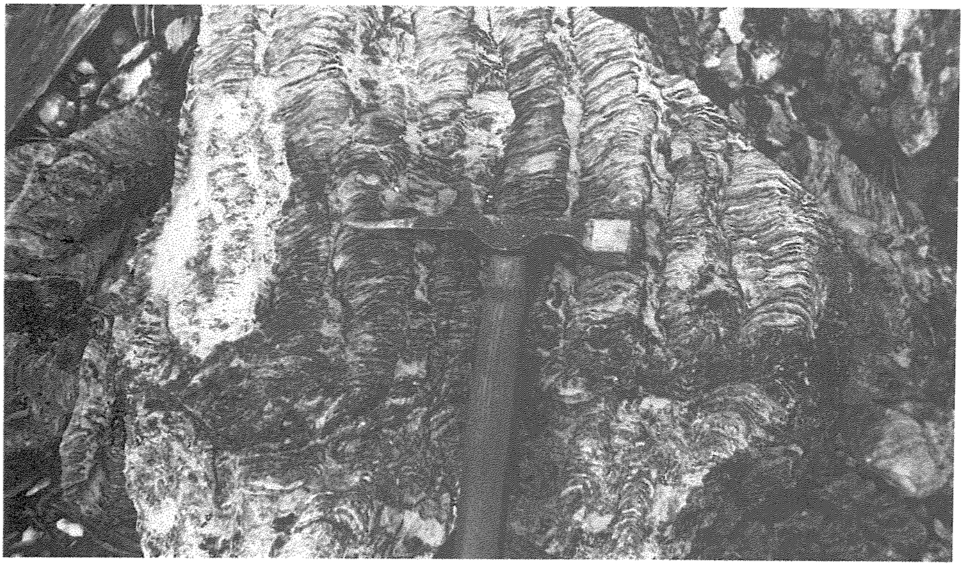
Walter, M. R. (ed.), 1976, Developments in sedimentology 20: Stromatolites: Amsterdam, Elsevier.

————1977, Interpreting stromatolites: American Scientist v. 65, no. 5, p. 563-571.

Williams, I. R., Brakel, A. T., Chin, R. J., and Williams, S. J., 1976, The Stratigraphy of the Eastern Bangemall Basin and Paterson Province: West. Australia Geol. Survey Ann. Rept 1975, p. 79-83.

Williams, S. J., Williams, I. R., Hocking, R. M., and Chin, R. J., 1980, Explantory Notes on the Glenburgh 1:250 000 Geological Sheet, W.A.: West. Australia Geol. Survey Record, 1980/4.

Figure 113. ***Baicalia capricornia*** **Walter. A—Specimen at type locality. B—GSWA F9908 from the type locality.**

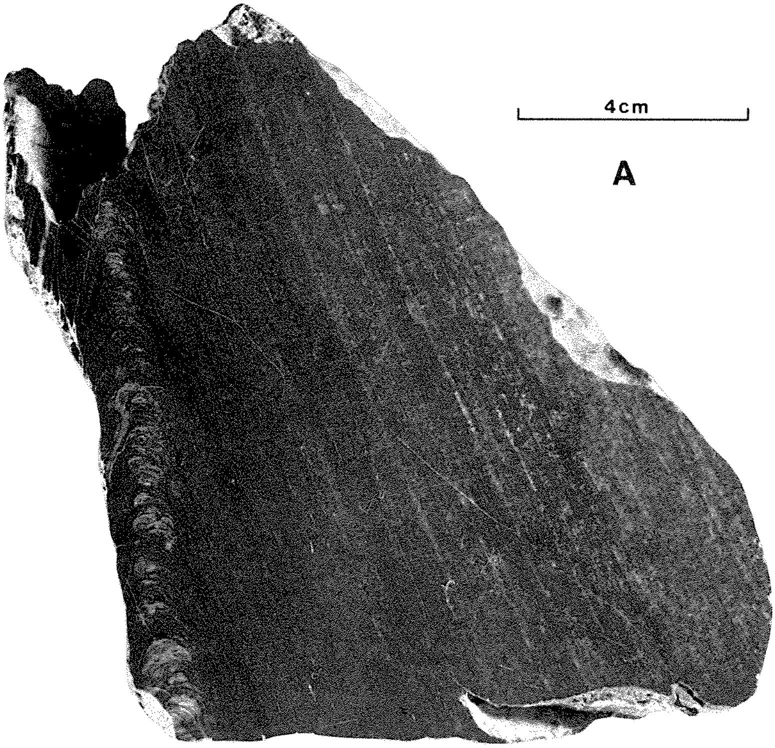


A

B

G.S.W.A. 18928

Figure 114. ***Conophyton garganicum australe*** **Walter from near Pingandy Creek. A—GSWA F9936. B—GSWA F9935.**


4cm
A

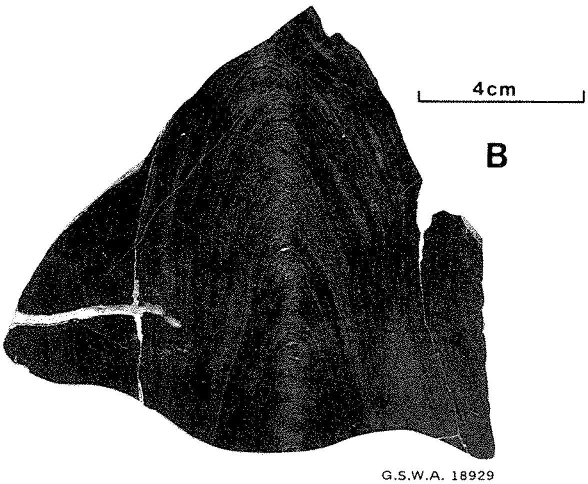

4cm
B
G.S.W.A. 18929

Figure 115. Unbranched club-shaped stromatolite (group and form indet.) from Irregully Creek gorge. A—GSWA F9914. B—GSWA F9915.

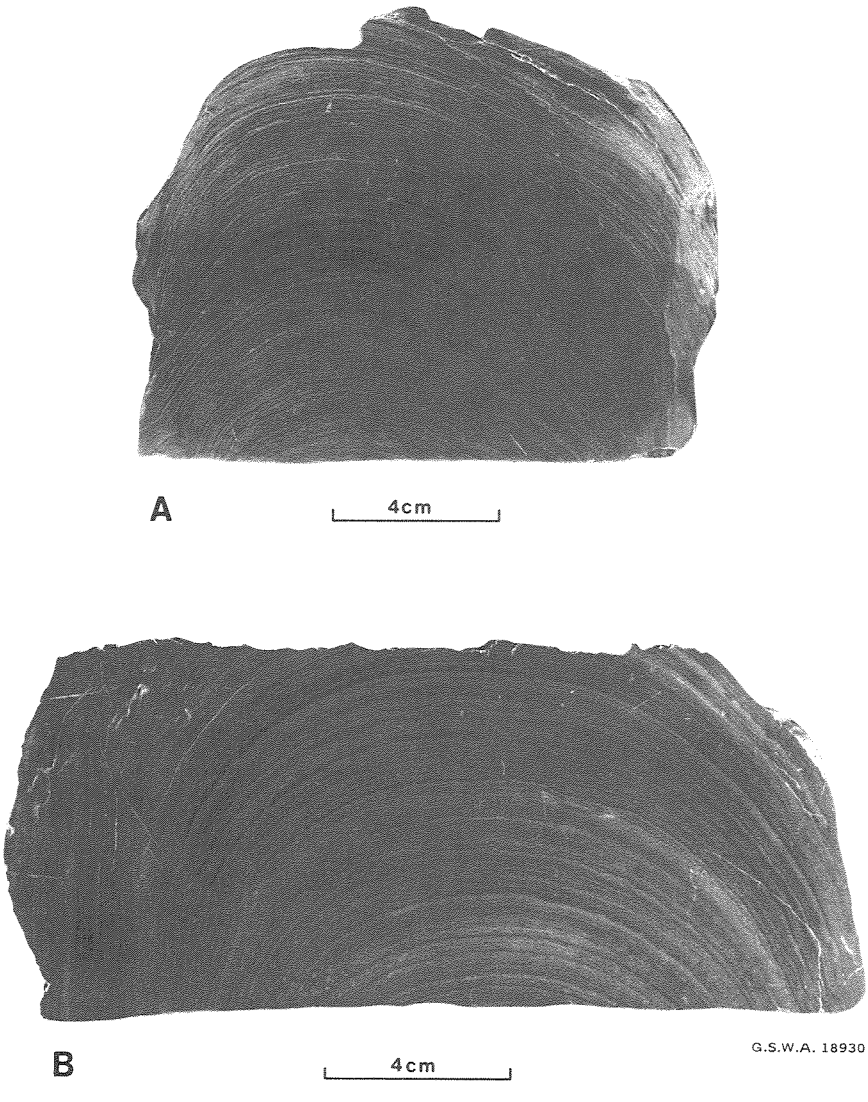
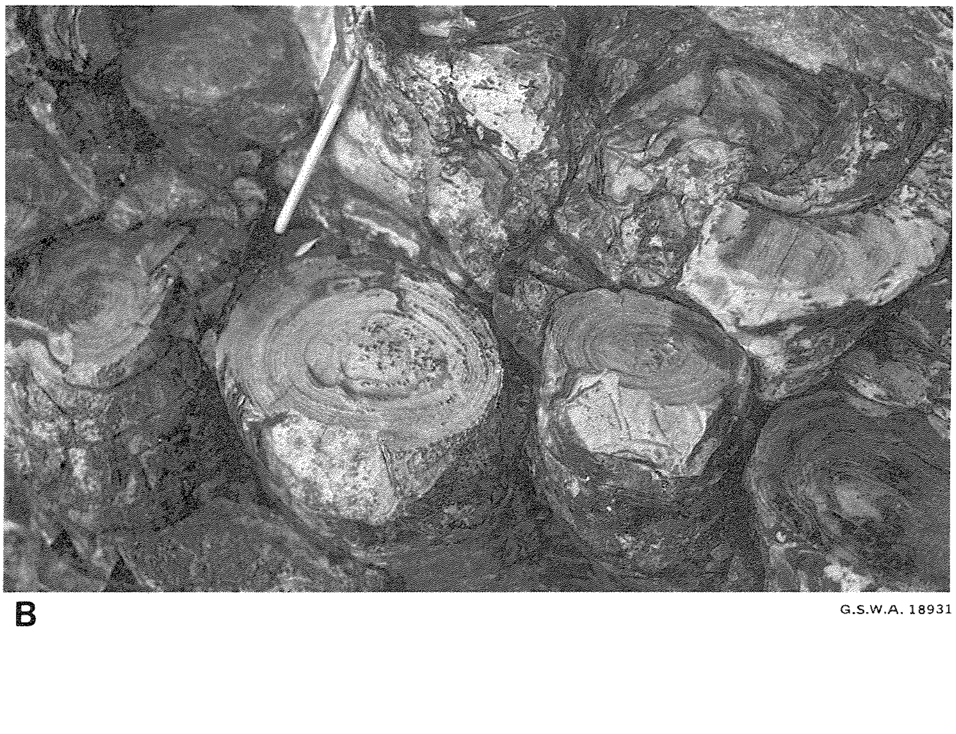

A
4cm
B
4cm
G.S.W.A. 18930

Figure 116. Unnamed club-shaped stromatolites in dolomite of the Irregully Formation, Irregully Creek gorge, 63 km north of Wanna homestead. A—Transverse section parallel to bedding. B—Outcrop showing three dimensional shape.

A



B

G.S.W.A. 18931

Figure 117. Microstructure of possible biogenic origin in the Bangemall Group. Bar scale in all figures represents 10 μm. GSWA fossil registration numbers follow description. Co-ordinates on Leitz Ortholux microscope number 587962 are given in brackets.

A—Sphere with radiating concentric structure, Discovery Chert. GSWA F11127 (27.9, -100.0).

B—Trail of diffuse carbonaceous material, Discovery Chert. GSWA F11176 (69.8, -096.8).

C—'Pod' of coarse silica crystals bound by possible organic material, Discovery Chert. GSWA F11140 (41.9, -101.6)

D—Problematic tubular structure showing terminal spheroid, Discovery Chert. GSWA F11143 (41.5, -111.4).

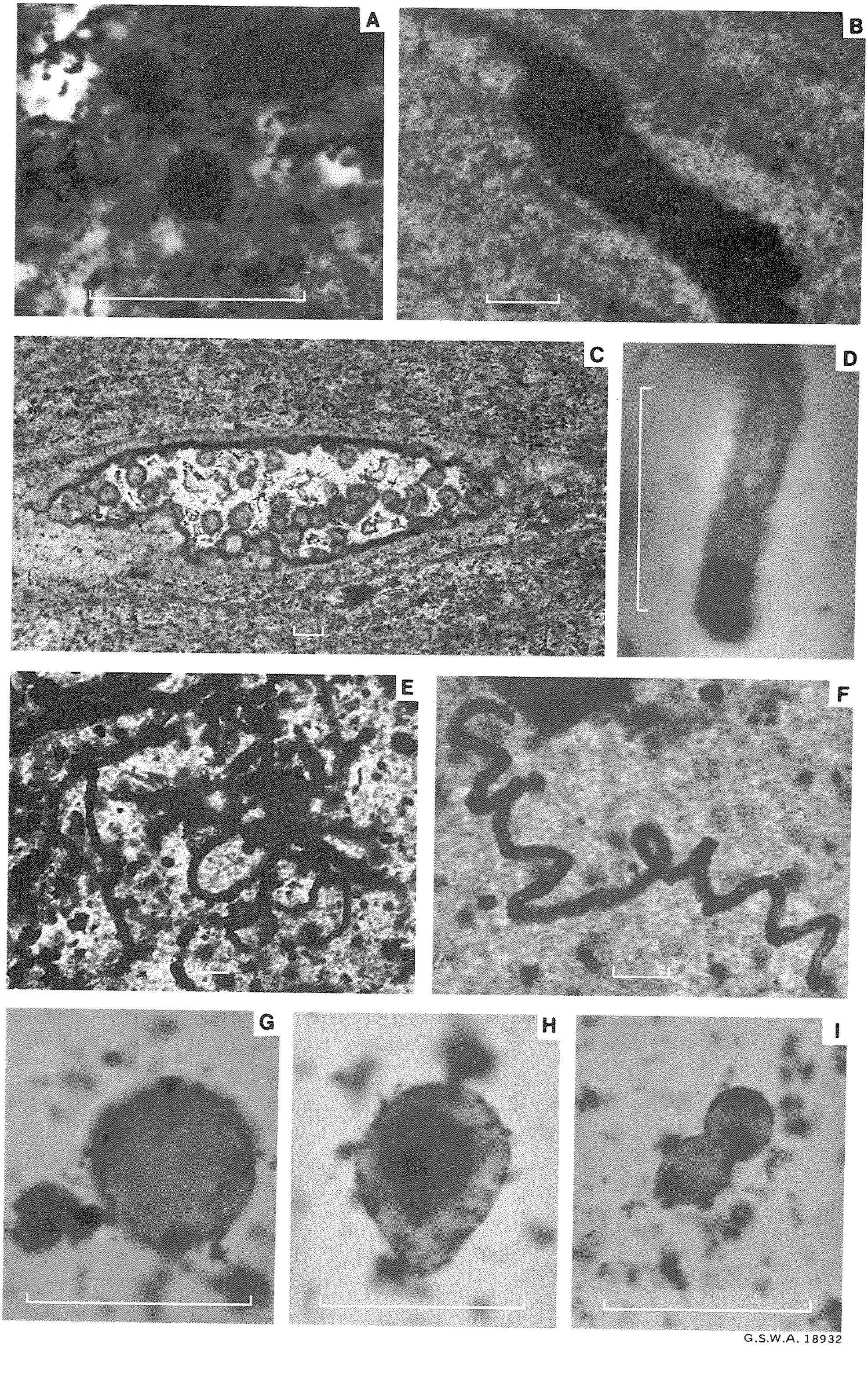
E—Cluster of problematic tubular structures, Discovery Chert. GSWA. F11143 (26.5, -112.3).

F—*Problematic tubular structure showing sinuous nature*, Discovery Chert. GSWA F11143 (26.3, -105.8).

G—Spheroidal body obtained from maceration, Backdoor Formation. GSWA F11150 (40.9, -114.0).

H—Elongate spheroidal body obtained from maceration, Backdoor Formation. GSWA F11150 (44.9, -114.0).

I—Linked spheres obtained from maceration, Backdoor Formation. GSWA F11150 (52.7, -113.4).



G.S.W.A. 18932

APPENDIX B. ANALYSES OF DISCOVERY CHERT AND OTHER CHERTS

	1	*2*	*3*	*4*	*5*	*6*	*7*	*8*	*9*	*10*	*11*	*12*	*13*	*14*	*15*
SiO_2	96.1	86.4	95.2	98.3	95.8	97.1	97.2	93.4	93.7	89.9	88.7	99.5	82.9	87.2	91.4
Al_2O_3	1.6	2.0	1.1	0.3	1.2	0.2	0.2	3.7	1.3	3.7	2.1	0.17	0.1	1.86	0.9
Fe_2O_3	0.3	2.4	0.7	0.4	0.4	0.4	0.0	0.2	0.4	2.3	1.5	0.12	3.4	1.39	0.5
FeO(a)	0.40	0.58	0.47	0.79	0.81	0.61	0.51	0.26	0.58	—	—	—	—	—	—
MgO(a)	0.08	0.08	0.24	0.03	0.18	0.10	0.10	0.41	0.17	0.5	0.4	0.05	0.19	0.14	0.8
CaO	0.02	0.12	0.30	0.01	0.46	0.00	0.00	0.00	0.06	0.3	0.2	0.09	1.60	0.05	2.7
Na_2O(a)	0.06	0.11	0.04	0.02	0.11	0.01	0.01	0.03	0.08	0.7	0.3	0.15	0.65	—	—
K_2O	0.3	1.0	0.5	0.1	0.3	0.0	0.1	1.4	0.6	0.7	0.5	0.07	1.40	—	—
H_2O^+(a)	0.86	0.16	1.19	0.64	0.70	0.44	0.05	0.64	0.12	—	—	0.12	} 0.33	—	—
H_2O^-(a)	0.20	0.33	0.17	0.01	0.08	0.04	0.03	0.02	0.09	—	—	—		—	—
CO_2(a)	0.11	0.06	0.17	0.10	0.22	0.31	0.12	0.18	1.45	—	—	—	0.40	—	—
TiO_2	0.20	0.16	0.11	0.02	0.10	0.01	0.04	0.17	0.08	0.2	0.2	—	0.27	—	0.02
P_2O_5	0.06	0.05	0.14	0.10	0.06	0.00	0.00	0.00	0.00	0.9	—	—	0.8	—	—
FeS_2	0.06	0.02	0.41	0.11	0.34	—	—	—	—	—	—	—	—	—	—
SO_3	—	1.98	—	—	—	—	—	—	—	—	—	—	—	—	—
MnO	<0.01	—	<0.01	<0.01	0.01	0.00	0.00	0.00	0.00	0.1	—	—	—	—	—
Total C(a)	0.03	4.03	0.20	0.26	0.24	0.07	0.11	0.08	1.00	—	—	—	—	—	—
Total S(a)	—	—	0.21	0.055	0.17	0.04	0.01	0.01	0.37	—	—	—	—	—	—
Ba	150	50	50	250	100	120	85	370	60						
Ga	< 2	3	< 2	< 2	< 2	< 2	< 2	4	< 2						
Rb	16	40	20	2	12	4	5	65	23						
Sr	24	20	38	112	26	85	20	20	15						
B	25	15	15	10	25	2	10	25	5						
Cu	145	110	60	40	120	40	100	20	40						
Pb	50	20	40	30	120	< 20	< 20	< 20	< 20						
Zn	60	10	60	35	70	40	20	20	20						
U	4	3	2	2	2	4	< 1	< 1	1						
V	50	290	310	110	< 20	120	130	250	90						

Oxides in per cent; trace elements in parts per million.
(a)—analysis by chemical methods, remainder by X-ray fluorescence samples 1-9.
The analyses in Appendix B are of the following material:

1. Discovery Chert, No. 34501F, Brumby Creek Anticline, COLLIER
2. Discovery Chert, No. 50407, Mount Egerton
3. Discovery Chert, No. 50418 Devil Creek, EDMUND
4. [illegible]
7. Discovery Chert, No. 50707, Lyons River Anticline, MOUNT EGERTON
8. Discovery Chert, No. 50708, Lyons River Anticline, MOUNT EGERTON
9. Discovery Chert, No. 50710, Devil Creek, EDMUND
10. Average radiolarian chert, Cressman (1962)
11. Average spicular chert, Cressman (1962)
12. Novaculite, Rockport, Arkansas, in Pettijohn, 1975

APPENDIX C. CHEMICAL COMPOSITION OF BLACK SHALE FROM THE TOP OF THE JILLAWARRA FORMATION MOUNT VERNON SYNCLINE

	50401	*50402*	*50403*	*50404*	*50405*	*50408*	*50409*	*50406*	$\overline{X}$	*S*
SiO_2	76.8	68.6	54.5	70.9	75.9	77.8	77.5	81.3	72.9	8.5
Al_2O_3	4.1	5.2	5.3	9.2	7.3	6.2	7.2	4.4	6.1	1.7
Fe_2O_3	3.7	9.1	19.2	4.3	8.4	7.9	6.5	4.3	7.9	5.0
FeO	0.36	0.13	0.45	0.32	0.32	0.45	0.64	0.58	0.41	0.16
MnO	0.03	0.03	0.03	0.02	0.02	0.03	0.03	0.02	0.03	0.01
CaO	0.16	0.08	0.04	0.04	0.08	0.11	0.06	0.05	0.08	0.04
MgO	0.17	0.29	0.27	0.27	0.21	0.23	0.19	0.18	0.23	0.05
Na_2O	0.19	0.24	0.19	0.24	0.13	0.11	0.15	0.11	0.17	0.05
K_2O	1.16	2.5	3.0	2.0	1.3	1.2	1.2	1.4	1.8	0.7
TiO_2	0.60	0.58	0.31	0.35	0.25	0.25	0.25	0.20	0.35	0.16
P_2O_5	0.08	0.10	0.10	0.06	0.07	0.07	0.06	0.05	0.07	0.02
Total S (as SO_3)	2.00	4.90	8.88	3.00	0.79	0.35	0.73	0.95	2.70	2.92
Total C	7.91	4.68	2.36	6.20	2.65	1.73	3.33	4.38	4.16	2.09
CO_2	0.12	0.12	0.05	0.04	0.09	0.07	0.04	—	0.08	0.04
H_2O^+	2.79	3.69	4.56	3.72	2.81	2.84	3.10	1.99	3.19	0.78
H_2O^-	0.65	0.60	0.36	0.56	0.39	0.41	0.45	0.32	0.47	0.12
B	35	40	30	40	30	35	40	25	34	6
Cu	400	200	110	580	520	760	500	170	405	228
Ga	10	24	16	10	10	10	10	<10	12.5	5
Pb	90	40	80	50	40	50	35	50	54	20
Rb	110	140	150	115	80	75	75	95	105	29
U	3	1	1	8	7	8	8	8	5.5	3.3
V	420	860	640	880	620	540	620	700	660	154
Zn	35	15	65	90	60	90	70	20	56	29

Oxides in per cent; trace elements in parts per million.

APPENDIX D. CHEMICAL COMPOSITION OF SEDIMENTS FROM THE ULLAWARRA FORMATION

	50411	*50412*	*50413*	*50414*	*41681*	*41682*
SiO_2	51.0	67.4	70.4	70.6	71.6	53.0
Al_2O_3	9.1	10.4	7.9	11.5	14.7	10.3
Fe_2O_3	4.9	3.8	2.9	5.4	1.32	27.0
FeO	4.0	7.1	9.4	1.9	0.33	<0.05
MnO	0.08	0.09	0.12	0.05	<0.05	<0.05
CaO	7.7	0.36	0.34	0.34	0.08	0.03
MgO	6.6	3.5	2.6	0.53	1.20	0.59
Na_2O	0.11	0.11	0.10	0.13	0.47	0.40
K_2O	2.6	1.6	0.5	2.0	4.32	2.04
TiO_2	0.89	0.43	0.30	0.83	0.81	0.78
P_2O_5	0.19	0.14	0.16	0.10	0.02	0.04
Total S	0.03	0.08	0.04	0.12		
Total C	2.65	0.03	0.04	0.03		
CO_2	9.46	0.06	0.06	0.06	0.07	<0.05
H_2O^+	3.33	4.42	3.90	4.32	3.40	3.66
H_2O^-	0.35	0.33	0.23	0.55	1.57	1.99
B	60	45	20	90	195	210
Cu	30	120	50	30		
Ga	14	14	12	14		
Pb	55	50	50	60		
Rb	100	70	25	100		
U	<1	<1	<1	2		
V	170	120	70	160		
Zn	80	165	95	10		

Oxides in per cent; trace elements in parts per million.

50411—Dolomitic shale
50412—Shale
50413—Green mudstone
50414—Laminated shale and siltstone
41681—Green mudstone
41682—Maroon mudstone

APPENDIX E. CHEMICAL COMPOSITION OF BLACK SHALE FROM THE CURRAN FORMATION AND FORDS CREEK SHALE

	Curran Formation			*Fords Creek Shale*
	50416	*50416A*	*50417*	*50149*
SiO_2	81.6	77.4	78.0	67.7
Al_2O_3	10.2	11.4	10.7	15.0
Fe_2O_3	0.8	1.5	1.0	1.6
FeO	0.19	0.9	1.0	5.3
MnO	0.02	0.02	0.02	0.35
CaO	0.04	0.28	0.21	0.21
MgO	0.66	0.50	0.50	1.3
Na_2O	0.12	0.13	0.12	0.65
K_2O	3.4	3.6	3.0	3.8
TiO_2	0.59	0.55	0.60	0.45
P_2O_5	0.12	0.12	0.16	0.11
Total S	0.04	0.02	0.04	0.01
Total C	0.80	0.76	0.77	0.03
CO_2	—	< 0.01	0.14	0.05
H_2O^+	1.88	2.5	2.5	3.0
H_2O^-	0.28	0.19	0.33	0.06
B	70	80	120	350
Cu	45	400	105	85
Ga	16	16	16	18
Pb	50	60	50	55
Rb	170	180	160	170
U	2	1	1	< 1
V	410	—	—	—
Zn	40	30	20	70

Oxides in per cent; trace elements in parts per million.

APPENDIX F. CHEMICAL COMPOSITIONS AND CIPW NORMS OF BASIC INTRUSIONS, BANGEMALL BASIN

	(a) 37456	37457	37458	37459	37460	37461	37462	37463	37464	39409	39435	41604	50721
SiO_2	49.6	50.2	50.8	50.8	53.0	56.6	49.8	51.1	51.4	49.8	51.2	48.9	48.0
TiO_2	1.17	1.61	1.58	1.59	2.66	1.56	2.47	1.77	2.26	2.02	1.42	2.97	1.94
Al_2O_3	14.9	14.4	13.8	14.0	11.3	14.2	12.7	13.6	14.5	13.0	13.6	12.4	13.3
Fe_2O_3	3.0	3.7	3.2	2.0	6.5	2.7	7.1	3.3	5.6	4.0	1.6	7.1	2.9
FeO	8.18	8.18	8.90	10.10	10.50	8.40	8.76	8.76	7.94	9.40	8.83	9.33	11.90
MnO	0.19	0.19	0.18	0.18	0.25	0.16	0.21	0.17	0.16	0.21	0.18	0.20	0.23
MgO	7.46	7.46	6.80	5.04	2.07	2.39	4.91	5.17	3.65	5.50	5.37	4.11	5.97
CaO	10.64	10.64	9.18	9.24	4.89	5.74	8.92	8.34	7.82	10.50	6.12	7.01	8.56
Na_2O	2.29	2.29	2.36	2.53	5.19	3.72	2.37	2.40	2.56	1.67	1.64	3.02	2.86
K_2O	0.59	0.59	1.27	1.26	0.79	1.66	0.93	1.78	1.90	0.42	1.31	1.43	0.57
P_2O_5	0.15	0.15	0.20	0.18	0.50	0.28	0.26	0.21	0.19	0.20	0.17	0.26	0.14
H_2O^+	1.75	1.57	1.30	1.33	2.08	2.06	1.29	2.67	2.03	3.07	4.77	2.40	3.30
H_2O^-	0.27	0.33	0.33	0.36	0.52	0.35	0.61	0.63	0.42	0.35	0.32	0.43	0.25
CO_2	0.14	0.03	0.05	0.05	0.05	0.03	0.11	0.05	0.06	0.03	3.74	0.04	0.02
TOTAL	100.3	100.5	100.0	98.7	100.3	98.8	100.4	99.9	100.5	100.2	100.3	99.6	99.9
K/Na	0.29	0.3	0.6	0.56	0.17	0.50	0.44	0.83	0.83	0.28	0.9	0.53	0.21
K/Rb	245	244	300	263	330	145	308	269	197	233	218	264	470
Rb/Sr	0.1	0.12	0.14	0.19	0.27	0.50	0.13	0.23	0.27	0.08	0.42	0.17	0.04
Q	0.73	1.82	2.35	2.53	6.57	10.66	8.53	4.60	7.95	8.57	20.35	6.11	—
C	—	—	—	—	—	—	—	—	—	—	7.42	—	—
or	3.49	4.31	7.51	7.45	4.67	9.81	5.50	10.52	11.23	2.48	7.74	8.45	3.37
ab	19.38	23.27	19.97	21.41	43.91	31.47	20.05	20.31	21.66	14.13	13.88	25.55	24.20
an	28.64	24.80	23.31	23.13	5.21	17.15	21.27	21.08	22.47	26.74	5.61	16.06	21.77
di	18.13	16.00	16.83	17.65	13.09	7.86	16.70	15.30	12.04	19.60	—	13.69	16.33
hy	20.72	19.47	20.13	18.37	8.51	12.89	10.65	16.09	9.71	15.07	26.6	10.28	19.76
ol	—	—	—	—	—	—	—	—	—	—	—	—	2.71
mt	4.35	5.36	4.64	2.90	9.42	3.91	10.29	4.78	8.12	5.80	2.32	10.29	4.20
il	2.22	3.06	3.00	3.02	5.05	2.96	4.69	3.36	4.29	3.84	2.70	5.64	3.68
ap	0.36	0.45	0.47	0.43	1.18	0.66	0.62	0.50	0.45	0.47	0.40	0.62	0.33
ca	0.32	0.07	0.11	0.11	0.11	0.07	0.25	0.11	0.14	0.07	8.51	0.09	0.05

APPENDIX G. ABUNDANCE OF TRACE ELEMENTS IN BASIC INTRUSIONS, BANGEMALL BASIN

	(a) 37456	*37457*	*37458*	*37459*	*37460*	*37461*	*37462*	*37463*	*37464*	*39409*	*39435*	*41604*	*50721*
Ba	180	250	450	500	220	600	300	1000	450	240	350	400	180
Cd	< 1	< 1	< 1	< 1	< 1	< 1	< 1	< 1	< 1	< 1	< 1	< 1	< 1
Ce	20	60	40	40	80	60	60	40	80	20	20	80	40
Co	40	50	50	45	35	30	50	40	50	40	30	60	50
Cu	180	120	160	850	70	50	200	170	95	70	90	210	150
Ga	15	18	18	18	15	21	21	18	18	15	12	18	15
La	< 20	< 20	< 20	40	40	40	< 20	< 20	60	< 20	< 20	40	< 20
Pb	40	20	10	30	20	10	80	20	10	10	20	10	10
Mo	1	1	1	1	1	1	1	1	1	1	< 1	< 1	< 1
Ni	85	70	140	80	< 5	15	60	80	30	50	60	55	60
Nb	5	5	5	5	20	10	5	< 5	< 5	< 5	< 5	5	5
Rb	20	25	35	40	20	95	25	55	80	15	50	45	10
Sc	40	35	35	35	45	35	40	35	30	50	50	40	45
Ag	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5	< 5
Sr	200	210	260	210	75	190	200	240	300	200	120	260	240
Th	< 10	< 10	< 10	< 10	< 10	10	< 10	< 10	10	< 10	< 10	< 10	< 10
Sn	22	8	6	70	2	2	80	6	4	6	4	4	16
Li	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 20	< 20
U	< 1	< 1	< 1	< 1	< 1	< 1	2	< 1	< 1	< 1	< 1	< 1	< 1
V	280	320	280	340	180	280	420	300	460	420	360	550	440
Y	20	25	25	30	50	35	35	30	30	30	25	30	25
Zn	110	110	100	115	98	110	140	115	82	110	94	125	130
Zr	105	150	165	180	240	240	420	180	210	150	120	225	135

(a) GSWA Sample numbers; results in parts per million.

SAMPLE	LATITUDE	LONGITUDE	LITHOLOGY
37456	24°57′40″S	118°40′15″E	Dolerite
37457	24°08′15″	118°45′30″	Dolerite
37458	24°14′40″	117°05′50″	Dolerite
37459	24°16′00″	119°28′45″	Dolerite
37460	24°36′15″	118°10′45″	Gabbro
37461	24°18′45″	117°19′20″	Leucodolerite
37462	24°14′00″	118°14′15″	Fine-grained dolerite
37463	24°55′15″S	120°05′30″E	Dolerite
37464	25°11′40″	121°53′40″	Dolerite
39409	24°09′30″	116°26′45″	Dolerite
39435	24°25′15″	117°18′30″	Dolerite
41604	24°16′45″	118°36′00″	Dolerite
50721	23°41′15″	115°36′30″	Dolerite

REFERENCES

Allen, J. R. L., 1963, The classification of cross-stratified units with notes on their origin: Sedimentology, v. 2, p. 93-114.

————1970, Physical processes of sedimentation. An introduction: London, G. Allen & Unwin.

Anderson, J. E., 1970, Recognition and time distribution of texturally altered tuff: Geol. Soc. America Bull., v. 81, p. 287-292.

Aplin, T. E. H., 1977, The vegetation of Western Australia, *in* Western Australia Yearbook 1977: Perth, Government Printer.

Assereto, R. L. and Kendall, C.G.S.C., 1977, Nature, origin and classification of peritidal tepee structures and related breccias: Sedimentology, v. 24, no. 2, p. 153-210.

Badgley, P. C., 1965, Structural and tectonic principles: Tokyo, Harper and Row & John Weatherhill Inc.

Battey, M. H., 1955, Alkali metasomatism and the petrology of some keratophyres: Geol. Mag., v. 92, p. 104-126.

Beard, J. S., 1974, The vegetation of the Great Victoria Desert Area: Explan. memoir map sheet 3, Vegetation Survey of West. Australia: Nedlands, Univ. West. Australia Press.

Blatt, H., Middleton, G., and Murray, R., 1972, Origin of sedimentary rocks: New Jersey, Prentice Hall.

Blissenback, E., 1954, Geology of alluvial fans in southern Nevada: Geol. Soc. America Bull., v. 65, p. 175-190.

Blockley, J. G., 1971, The lead, zinc, and silver deposits of Western Australia: West. Australia Geol. Survey Mineral Resources Bull. 9.

Bowie, S. H. U., Dawson, J., Gallagher, M. J., and Ostle, D., 1966, Potassium-rich shales in the Cambrian of north-west Scotland: Institution Mining Metall. Trans., v. 75, pt. B125, pt. 125-145.

Bradley, J., 1965, Intrusion of major dolerite sills: Royal Soc. of New Zealand, Transactions (Geology), v. 3, p. 27-55.

Brakel, A. T., and Muhling, P. C., 1976, Stratigraphy, sedimentation and structure in the Bangemall Basin, Western Australia: West. Australia Geol. Survey Ann. Rept 1975, p. 70-79.

Brakel A. T., Elias, M., and Barnett, J. C., 1978, Explanatory notes on the Collier 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1978/8.

Brakel, A. T., and Leech, R. E. J., 1979, explanatory notes on the Trainor 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1978/14.

Brock, B. B., 1961, Structural setting of the copperbelt, *in* The geology of the Northern Rhodesian copperbelt (F. Mendelssohn, ed.): London, Macdonald.

Bull, W. B., 1964, Alluvial fans and near surface subsidence in western Fresno County, California: U.S. Geol. Survey Prof. Pap. 437-A.

Bunting, J. A., Brakel, A. T., and Commander, D. P., 1978, Explanatory notes on the Nabberu 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1978/12.

Bunting, J. A., Commander, D. P., and Gee, R. D., 1977, Preliminary synthesis of Lower Proterozoic stratigraphy and structure adjacent to the northern margin of the Yilgarn Block: West. Australia Geol. Survey Ann. Rept 1976, p. 43-48.

Burke, K., 1976, The Chad Basin: an active intra-continental basin: Tectonophysics, v. 36, p. 197-206.

Busk, H., 1929, Earth flexures—their geometry and analysis in geological section with special reference to the problem of oil finding: London, Cambridge Univ. Press.

Button, A., 1975, The Gondwanaland Precambrian Project: Econ. Geol. Res. Unit, University of Witwatersrand Ann. Rept 1974, p. 37-54.

Carey, S. W., 1962, Folding: Alberta Soc. Petroleum Geol. Jour., v. 10, no. 3, p. 95-144.

Carozzi, A. V., 1960, Microscopic sedimentary petrography: New York, Wiley.

Carmichael, I. S., Turner, F. J., and Verhoogen, J., 1974, Igneous Petrology: McGraw Hill.

Chapman, F., 1933, On fossiliferous grits and cherts, presumably of Cretaceous age associated with the Nullagines of Western Australia: Royal Soc. Victoria Proc., v. 46, p. 60-65.

Chin, R. J., Williams, I. R., Williams, S. J., and Crowe, R. W. A., 1979, Explanatory notes on the Rudall 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1978/17.

Christie, R. L., 1978, Sedimentary phosphate deposits—an interim review: Canada Geol. Survey, Pap. 78-20.

Clifton, H. E., Hunter, R. E., and Phillips, R. L., 1971, Depositional structures and processes in the non-barred, high energy near shore: Jour. Sed. Pet., v. 41, p. 651-670.

Cloos, E., 1961, Bedding plane slips, wedges and folding in layered sequences: Bull. Comm. Finland No. 196, p. 105-122.

Coleman, J. M., and Gagliano, S. M., 1965, Sedimentary structures: Mississippi River deltaic plain, *in* Primary sedimentary structures and their hydrodynamic interpretation (G. V. Middleton, ed.): Soc. Econ. Palaeontologists Mineralogists Spec. Pub. 12, p. 133-148.

Coleman, J. M., and Wright, L. D., 1975, Sedimentation in an arid macrotidal alluvial river system: Ord River, Western Australia: Jour. Geol., v. 86, p. 621-642.

Collinson, J. D., 1970, Bedforms of the Tana River, Norway: Geograf. Ann., v. 52A, p. 31-56.

Commander, D. P., Muhling, P. C., and Bunting, J. A., 1979, Explanatory notes on the Stanley 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1978/15.

Compston, W., and Arriens, P. A., 1968, The Precambrian geochronology of Australia: Canadian Jour. of Earth Sciences, v. 5, p. 561-583.

Conybeare, C. G. B., and Crook, K. A. W., 1968, Manual of sedimentary structures: Australia BMR Bull. 102.

Cook, P. J., 1975, Georgina Basin phosphatic province, Queensland and Northern Territory—a regional geology, *in* Economic geology of Australia and Papua New Guinea, Metals (C. L. Knight, ed.): Australasian Inst. Mining Metall. Mon. 5, p. 245-250.

————1976, Sedimentary phosphate deposits, *in* Stratabound and stratiform ore deposits (K. H. Wolf, ed.): Amsterdam, Elsevier, pt. 5, p. 505-535.

Corbett, K. D., 1973, Open cast slump sheets and their relationship to sandstone beds in an Upper Cambrian flysch sequence, Tasmania: Jour. Sed. Pet., v. 43, p. 139-160.

Cressman, E. R., 1962, Data of Geochemistry: Non-detrital siliceous sediments: U.S. Geol. Survey Prof. Pap, 440-T, p. 1-22 (6th edition).

Crowe, R. W. A., 1975, The classification, genesis and evolution of sand dunes in the Great Sandy Desert: West. Australia Geol. Survey Ann. Rept 1974, p. 46-49.

Croxford, N. J. W., 1965, Origin and significance of volcanic potash-rich rocks from Mt Isa: Institution Mining Metall. Trans., v. 74, p. 33-43.

————1968, A mineralogical study of the McArthur River lead-zinc-silver deposit: Australasian Inst. Mining Metal Proc., 226, p. 97-108.

Daniels, J. L., 1965, Breccias associated with the Proterozoic Bangemall Group: West. Australia Geol. Survey Ann. Rept 1964, p. 34-36.

———1966a, The Proterozoic Geology of the North-West Division of Western Australia: Australasian Inst. Mining Metall. Proc. 219, p. 17-26.

———1966b, Revised stratigraphy, palaeocurrent system and palaeogeography of the Proterozoic Bangemall Group: West. Australia Geol. Survey Ann. Rept 1965, p. 48-56.

———1968, Turee Creek, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

———1969, Edmund, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

———1970, Wyloo, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

———1974, The geology of the Blackstone Region, Western Australia: West. Australia Geol. Survey Bull. 123.

———1975, Bangemall Basin, *in* Geology of Western Australia: West. Australia Geol. Survey Mem. 2 p. 147-159.

Daniels, J. L., and MacLeod, W. N., 1965, Newman, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

Davy, R., 1980, A chemical and mineralogical study of low-grade zinc mineralization at three localities in the Proterozoic Bangemall rocks of Western Australia: West. Australia Geol. Survey Rept 10.

———1975, A geochemical study of a dolomite-BIF transition in the lower part of the Hamersley Group: West. Australia Geol. Survey Ann. Rept 1974, p. 88-100.

Davy, R., Brakel, A. T., and Muhling, P. C., 1978, The value of B-Ga-Rb diagrams in determining depositional conditions in Proterozoic rocks in the northwest of Western Australia: West. Australia Geol. Survey Ann. Rept 1977, p. 80-85.

Davies, W., and Cave, R., 1976, Folding and cleavage determined during sedimentation: Sedimentary Geol., v. 15, p. 89-133.

Deer, W. A., Howie, R. A., and Zussman, J., 1962, An introduction to the rock-forming minerals, v. 3, Sheet Silicates: London, Longmans.

de la Hunty, L. E., 1963, The geology of the manganese deposits of Western Australia: West. Australia Geol. Survey Bull. 116.

———1964, Balfour Downs, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

———1966, Manganese nodules in Middle Proterozoic shale in the Pilbara Goldfield, Western Australia: West. Australia Geol. Survey Ann. Rept 1965, p. 65-68.

———1969, Robertson, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

Dennis, J. G., 1972, Structural geology: New York, Ronald Press Co.

De Raaf, J. F. M., and Boersma, J. R., 1971, Tidal deposits and their sedimentary structures (seven examples from Western Europe): Geol. Mijnbouw, v. 50, p. 479-504.

Donaldson, A. C., Martin, R. H., and Kanes, W. H., 1970, Holocene Guadalupe delta of Texas Gulf Coast, *in* Deltaic sedimentation; modern and ancient: Soc. Econ. Palaeontol. Mineral. Spec. Pub., no. 15, p. 105-137.

Drever, J. L., 1974, Geochemical model for the origin of Precambrian banded iron formations: Geol. Soc. America Bull., v. 85, p. 1099-1106.

Drummond, B. J., 1979, A crustal profile across the Pilbara and northern Yilgarn Blocks, Western Australia: BMR Jour. of Australian Geol. and Geophys. v. 4, no. 3, p. 171-180.

Dunham, R. J., 1962, Classification of carbonate rocks according to depositional texture, *in* Classification of carbonate rocks (William E. Ham, ed.): American Assoc. Pet. Geol. Mem. 1, p. 108-121.

Dunham, K. C., 1964, Neptunist concepts in ore genesis: Econ. Geol., v. 59, p. 1-2.

Elias, M., and Williams, S. J., 1977, Explanatory notes on the Robinson Range 1:250 000 geological sheet, W.A.,: West. Australia Geol. Survey Rec. 1977/6.

Embleton, B. J. J., 1978, The palaeomagnetism of 2 400 m.y. old rocks from the Australian craton and its relation to Archaean-Proterozoic tectonics: Precambrian Research 6, p. 275-291.

Emery, K. O., 1956, Sediments and water of the Persian Gulf: American Assoc. Pet. Geol. Bull., v. 40, p. 2354-2383.

Ernst, W. G., and Calvert, S. E., 1969, An experimental study of the recrystallization of porcellanite: American Jour. Sci., v. 267A, p. 114-133.

Freeman, W. E., and Visher, G. S., 1975, Stratigraphic analysis of the Navajo Sandstone: Jour. Sed. Pet., v. 45, p. 651-668.

Garlick W. G., 1961, Muva System, *in* The geology of the Northern Rhodesian copperbelt (F. Mendelssohn, ed.): London, Macdonald, p. 21-30.

Garrels, R. M., and MacKenzie, F. T., 1971, Evolution of sedimentary rocks: New York, W. W. Norton and Company.

Gee, R. D., 1975, Bangemall Basin—regional geology, *in* Economic Geology of Australia and Papua New Guinea, Metals (C. L. Knight, ed.): Melbourne, Australasian Inst. Mining Metall. Mon. 5, p. 528-529.

————1979a, The geology of the Peak Hill area: West. Australia Geol. Survey Ann. Rept 1979, p. 55-62.

————1979b, Structure and tectonic style of the Western Australian shield: Tectonophysics, v. 58, p. 327-369.

Gee, R. D., de Laeter, J. R., and Drake, J. R., 1976, The geology and geochronology of altered rhyolite from the lower part of the Bangemall Group near Tangadee, Western Australia: West. Australia Geol. Survey Ann. Rept 1975, p. 112-117.

Germs, G. J., 1975, Silicification structures from the Upper Otavi Group and the significance of chert in this unit: Geol. Soc. South Africa, Trans., v. 78, p. 67-70.

Glaessner, M. F., Preiss, W. V., and Walter, M. R., 1969, Precambrian columnar stromatolites in Australia: Morphological and stratigraphical analysis: Science, v. 164, no. 3883, p. 1056-1058.

Glennie, K. W., 1970, Desert sedimentary environments: Development in sedimentology, v. 14: Amsterdam, Elsevier.

Green, J. E., 1970, Snowflake texture not diagnostic of devitrified ash-flow tuffs—Discussion: Geol. Soc. America Bull., v. 81, p. 2527-2528.

Green, T. H., Green, D. H., and Ringwood, A. E., 1967, The origin of high-alumina basalts and their relationships to quartz tholeiites and alkali basalts: Earth Planet. Sci. Lett., v. 2, no. 1, p. 41-51.

Grey, K., 1977, Middle Proterozoic stromatolites from the Bangemall Basin, Western Australia: West. Australia Geol. Survey Palaeont. Rept 45/77 (unpublished).

————1978, *Acaciella* cf. *australica* from the Skates Hills Formation, eastern Bangemall Basin, Western Australia: West. Australia Geol. Survey Ann. Rept 1977, p. 71-73.

Halligan, R., and Daniels, J. L., 1964, Precambrian geology of the Ashburton Valley region—North-West Division: West. Australia Geol. Survey Ann. Rept 1963, p. 38-46.

Harrison, J. V., and Falcon, N. L., 1934, Collapse structures: Geol. Mag., v. 71, p. 529-539.

Hickman, A. H., 1978, Nullagine, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

Hills, E. S., 1963, Elements of structural geology: Methuen & Co. Ltd.

Hobbs, B. E., Means, W. D., and Williams, P. F., 1976, An outline of structural geology: New York, John Wiley & Sons Inc.

Hocking, R. M., Williams, S. J. and Lavaring, I. H., 1983, Explanatory notes on the Winning Pool-Minilya 1:250 000 Geological Sheet, W.A.: West. Australia Geol. Survey Rec. 1982/9.

Houbolt, J. J. H. C.,. 1968, Recent sediments in the southern bight of the North Sea: Geol. Mijnbouw, v. 47, p. 245-273.

Hoyt, J. H., 1967, Barrier island formation: Geol. Soc. America Bull., v. 78, p. 1125-1136.

Hubbert, M. K., and Rubey, W. W., 1959, Role of fluid pressure in mechanics of overthrust faulting—I. Mechanics of fluid-filled porous solids and its application to overthrust faulting: Geol. Soc. America Bull., v. 70, no. 2, p. 115-166.

Hubert, J. F., 1964, Textural evidence for deposition of many western North Atlantic deep-sea sands by ocean bottom currents rather than turbidity currents: Jour. Geol., v. 72, p. 757-785.

Huddleston, W. H., 1883, Notes on a collection of fossils and of rock-specimens from West Australia, north of the Gascoyne River: Geol. Soc. London Quart. Jour., v. 39, p. 582-595.

Irwin, M. L., 1965, General theory of epeiric clear water sedimentation: American Assoc. Pet. Geol. Bull. v. 49, p. 445-459.

Jacobsen, J. B. E., 1975, Copper deposits in time and space: Minerals Sci. Eng., v. 7, p. 337-371.

Johnson, W., 1950, A geological reconnaissance survey of part of the area included between the limits lat. 24°00'S and lat. 22°00'S and between long. 115°30'E and long. 118°30'E including parts of the Yalgoo, Murchison, Peak Hill and Gascoyne Goldfields: West. Australia Geol. Survey Bull. 106.

Jopling, A. V. and Walker, R. G., 1968, Morphology and origin of ripple-drift cross-lamination with examples from the Pleistocene of Massachussets: Jour. Sed. Petrology v. 38, p. 971-984.

Kendall, C. G. St C., and Skipworth, P. A. d'E, 1968, Recent algal mats of a Persian Gulf lagoon: Jour. Sed. Pet., v. 38, p. 1040-1058.

Kennewell, P. J., 1975, Madley, W.A.: Australia Bur. Mineral Resources 1:250 000 Geol. Series Explan. Notes.

Kinsman, D. J. J., and Park, R. K., 1976, Algal belt and coastal sabkha evolution, Trucial Coast, Persian Gulf, *in* Stromatolites (M. R. Walter, ed.): Amsterdam, Elsevier.

Krauskopf, K. B., 1967, Introduction to geochemistry: New York, McGraw Hill.

Kumar, N., and Sanders, J. E., 1976, Characteristics of shore-face storm deposits—Modern and ancient examples: Jour. Sed. Pet., v. 46, p. 145-162.

Leech, R. E. J., and Brakel, A. T., 1978, Explanatory notes on the Bullen 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1978/4.

Lipman, P. W., 1965, Chemical comparison of glassy and crystalline volcanic rocks: U.S. Geol. Survey Bull. 1201-D, 24p.

Logan, B. W., and Chase, R. L., 1961, The stratigraphy of the Moora Group, Western Australia: Royal Soc. West. Australia Jour., v. 44, p. 14-31.

Logan, J. M., Friedman, M., and Stearns, M. T., 1978, Experimental folding of rocks under confining pressure—Part VI Further studies on faulted drape folds, *in* Laramide folding associated with basement block faulting in the western United States: Geol. Soc. America Mem. 151, p. 79-99.

Love, L. G., 1962, Biogenic primary sulphide of the Permian Kupferschiefer and Marl Slate: Econ. Geol., v. 57, p. 350-366.

Low, G. H., 1963, Copper deposits of Western Australia: West. Australia Geol. Survey Mineral Resources Bull. 8.

Lowe, D. R., 1976, Non glacial varves in Lower Member of Arkansas Novaculite (Devonian), Arkansas and Oklahoma: American Assoc. Pet. Geol. Bull., v. 60, p. 2103-2116.

Mackenzie, F. T., Garrels, R. M., Brisker, O. P., and Bickley, F., 1967, Silica in seawater: control by silica minerals: Science, v. 155, p. 1404-1405.

MacLeod, W. N., 1970, Peak Hill, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

Maitland, A. G., 1909, Geological Investigations in the country lying between 21°30′ and 25°30′S latitude and 113°30′ and 118°30′E longitude, embracing parts of the Gascoyne, Ashburton and West Pilbara Goldfields: West. Australia Geol. Survey Bull. 33.

Marshall, A. E., 1968, Geological studies of the Proterozoic Bangemall Group, N. W. Australia: Princeton University Ph.D. Thesis (unpublished).

Marston, R. J., 1980, Copper mineralization in Western Australia: West. Australia Geol. Survey Mineral Resources Bull. 11.

Matthews, V., 1978, Preface, *in* Laramide folding associated with basement block faulting in the western United States: Geol. Soc. America Mem. 151.

Matthews, V., and Work, D. F., 1978, Laramide folding associated with basement block faulting along the northeastern flank of the Front Range Colorado, *in* Laramide folding associated with basement block faulting in the western United States: Geol. Soc. America Mem. 151, p. 101-124.

McElhinny, M. W., and Embleton, B. J. J., 1976, Precambrian and Early Palaeozoic palaeomagnetism in Australia: Royal Soc. London, Phil. Trans. Ser. A, v. 280, p. 417-431.

McKee, E. D., 1957, Primary structures in some recent sediments: American Assoc. Pet. Geol. Bull., v. 41, p. 1704-1747.

McKee, E. D., and Gutschick, R. E., 1969, History of the Redwall Limestone of northern Arizona: Geol. Soc. America Mem., no. 114.

Monty, C. V. L., 1976, The origin and development of cyrstalgal fabrics, *in* Stromatolites (M. R. Walter, ed.): Amsterdam, Elsevier, p. 193.

Muhling, P. C., Brakel, A. T., Davidson, W.A., 1978, Mount Egerton, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

Muir, M. D., 1979, Sabkha model for part of the McArthur Group and implications for mineralization: BMR Jour. of Australian Geol. and Geophys., v. 4, p. 149-165.

Mulcahy, M., and Bettenay, E., 1972, Soil and landscape studies in Western Australia—(1) The major drainage divisions: Geol. Soc. Australia Jour., v. 18, p. 349.

Murray, W. J., 1975, McArthur River H.Y.C. lead-zinc and related deposits, N.T., *in* Economic Geology of Australia and Papua New Guinea, Metals (C. L. Knight, ed.), Australasian Inst. Mining Metall. Mon. 5, p. 329-339.

Nockolds, S. R., 1954, Average chemical compositions of some igneous rocks: Geol. Soc. America Bull., v. 65, p. 1007-1032.

Oehler, J. H., 1976, Hydrothermal crystallization of silica gel: Geol. Soc. America Bull., v. 87, p. 1143-1152.

Pearce, J. A., and Cann, J. R., 1971, Ophiolite origin investigated by discriminant analysis using Ti, Zr and Y: Earth Planet. Sci. Lett., v. 12, no. 3, p. 339-349.

Peterson, M. N. A., and van der Borch, C. C., 1965, Chert: modern inorganic deposition in a carbonate precipitating locality: Science, v. 149, p. 1501-1503.

Pettijohn, F.J., 1975, Sedimentary rocks: New York, Harper and Row.

Playford, P. E., and Cockbain, A. E., 1976, Modern algal stromatolites at Hamelin Pool, a hypersaline barred basin in Shark Bay, Western Australia, *in* Stromatolites (M. R. Walter, ed.): Amsterdam, Elsevier.

Plumb, K. A., and Derrick, G. M., 1975, Geology of the Proterozoic rocks of the Kimberley to Mt Isa region, *in* Economic Geology of Australia and Papua New Guinea, Metals (C. L. Knight, ed.): Australasian Inst. Mining Metall. Mon. 5, p. 217-252.

Potter, P. E., and Pettijohn, F. J., 1963, Palaeocurrents and basin analysis: Berlin, Springer-Verlag, p. 152-155.

Preiss, W. V., 1976, Proterozoic stromatolites from the Nabberu and Officer Basins, Western Australia and their biostratigraphic significance: South Australia Geol. Survey Rept Invest., v. 47, p. 1-51.

Price, N. J., 1977, Aspects of gravity tectonics and the development of listric faults: Jour. Geol. Soc., v. 133, p. 311-328.

Price, R. A., and Mountjoy, E. W., 1970, Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca Rivers—a progress report: Geol. Assoc. Canada Spec. Pap. 6, p. 7-26.

Ragan, D. M., 1973, Structural geology: An introduction to geometrical techniques: Wiley, 2nd Edition.

Ramsay, J. G., 1967, The folding and fracturing of rocks: New York, McGraw-Hill.

Reches, Z., 1978, Development of monoclines: Part 1. Structure of the Palisades Creek branch of the East Kaibala monocline, Grand Canyon, Arizona, *in* Laramide folding associated with basement block faulting in the western United States (V. Matthews, ed.): Geol. Soc. America Mem. 151, p. 235-272.

Reineck, H. E., and Singh, I. B., 1973, Depositional sedimentary environments: Berlin, Springer-Verlag.

Renfro, A. R., 1974, Genesis of evaporite associated stratiform metalliferous deposits—a sabkha process: Econ. Geol., v. 69, p. 33-45.

Rogers, J. K., and Keevers, R. E., 1975, Lady Annie-Lady Jane phosphate deposits, Georgina Basin, Queensland, *in* Economic Geology of Australia and Papua New Guinea, Metals (C. L. Knight, ed.): Australasian Inst. Mining Metall, Mon. 4, p. 251-265.

Ryer, T. A., 1977, Patterns of Cretaceous shallow marine sedimentation, Coolville and Rockport areas, Utah: Geol. Soc. America Bull., v. 88, p. 177-188.

Sanders, J. E., 1965, Primary sedimentary structures formed by turbidity currents and related resedimentation mechanisms, *in* Primary sedimentary structures and their hydrodynamic interpretation (G. B. Middleton, ed.): Soc. Econ. Palaeo. Min. Spec. Pub. 12, p. 192-219.

Selley, R. C., 1970, Ancient sedimentary environments: London, Chapman and Hall.

————1976, An introduction to sedimentology: London, Academic Press.

Shaw, A. B., 1964, Time in stratigraphy: New York, McGraw Hill.

Sheppard, R. A., and Gude, A. J., 1969, Diagnosis of tuffs in the Barstow Formation, Mud Hills, San Bernadina County, California: U.S. Geol. Survey Prof. Pap. 634.

Simons, F. S., 1962, Devitrification dikes and giant spherulites from Klondyke, Arizona: American Min., v. 47, p. 871-885.

Simpson, E. S., 1923, Secondary sulphates and chert in the Nullagine Series; Royal Soc. West. Australia Jour., v. 9, p. 45-53.

Smith, R. E., and Davy, R., 1979, Exploration geochemistry at the Mount Palgrave and Mount Vernon Cu-Zn locations, Bangemall Basin: West. Australia Geol. Survey Ann. Rept 1978, p. 92-100.

Spencer, E. W., 1969, Introduction to structure of the earth: New York, McGraw Hill.

Stanton, R. L., 1972, Ore petrology: New York, McGraw Hill.

Stearns, D. W., 1978, Faulting and forced folding in the Rocky Mountains foreland: *in* Laramide folding associated with basement block faulting in the western United States (V. Matthews, ed.): Geol. Soc. America Mem. 151, p. 1-38.

Steidtmann, J. R., 1974, Evidence for eolian origin of cross-stratification and sandstone of the Casper Formation, southernmost Laramie Basin, Wyoming: Geol. Soc. America Bull., v. 85, p. 1835-1842.

Steinitz, G., 1970, Chert 'dike' structures in Senomian chert beds, southern Negev, Israel: Jour. Sed. Pet., v. 40, p. 1241-1254.

Stuart-Smith, P. G., and Ferguson, J., 1978, The Oenpelli Dolerite—a Precambrian continental suite from the Northern Territory, Australia: BMR Jour. of Australian Geol. and Geophys. v. 3, p. 125-133.

Sturt, B. A., 1961, Discussion *in* Some aspects of sedimentation in orogenic belts: Geol. Soc. London Proc. no. 1587, p. 78.

Sutton, J., 1969, Rates of change within orogenic belts, *in* Time and Place in Orogeny: Geological Soc. London, p. 239-250.

Swett, K., 1968, Authigenic feldspars and cherts resulting from dolomitization of illitic limestones—a hypothesis: Jour. Sed. Pet., v. 38, p. 128-135.

Takahashi, J., 1939, Synopsis of glauconitization, *in* Recent marine sediments (P. D. Trask, ed.), Tulsa Okla.: American Assoc. Pet. Geol., p. 503-512.

Talbot, H. W. B., 1910, Geological observations in the country between Wiluna, Halls Creek and Tanami: West. Australia Geol. Survey Bull. 39.

————1914, The Ilgarari mineral leases and Humphrey's Copper Find: West. Australia Geol. Survey Bull. 59.

————1920, The geology and mineral resources of the North West, Central and Eastern Divisions between long. 119° and 122°E, and lat. 22° and 28°S: West. Australia Geol. Survey Bull. 83.

————1926, A geological reconnaissance in the Central and Eastern Divisions between 122°30′ and 123°30′E longitude and 25°30′ and 28°15′S latitude: West. Australia Geol. Survey Bull. 87.

Talbot, J. L., 1969, The effects of the Palaeozoic orogeny on the Precambrian shield of South Australia, *in* Age relations in high-grade metamorphic terrains: Geol. Assoc. Canada Spec. Pap. 3, p. 59-66.

Tanner, W. F., 1967, Ripple mark indices and their uses: Sedimentology, v. 9, p. 89-104.

————1971, Numerical estimates of ancient waves, water depth and fetch: Sedimentology, v. 16, p. 71-88.

Thompson, A. M., 1975, Clastic coastal environments in Ordovician molasse, Central Appalachians, *in* Tidal Deposits (R. N. Ginsberg, ed.): Berlin, Springer Verlag, p. 135-143.

Trendall, A. F., and Blockley, J. G., 1970, The iron formations of the Precambrian Hamersley Group, Western Australia: West. Australia Geol. Survey Bull. 119.

van de Graaff, W. J. E., 1972, The Wanna Beds—an analogue of Recent North Sea sediments: West. Australia Geol. Survey Ann. Rept 1971, p. 100-102.

van de Graaff, W. J. E., Crowe, R. W. A., Bunting, J. A., and Jackson, M. J., 1977, Relict early Cainozoic drainages in arid Western Australia: Z. Geomorph N. F. 21 no. 4, p. 379-400.

van de Graaff, W. J. E., Denman, P. D., Hocking, R. M., and Baxter, J. L., 1977, Explanatory notes on the Ningaloo-Yanrey 1:250 000 geological sheet, W. A.: West. Australia Geol. Survey Rec. 1977/9.

van Straaten, L. M. J. U., 1953, Rhythmic pattern on Dutch North Sea beaches: Geol. Mijnbouw, v. 15e, p. 31-43.

Walker, R. N., Logan, R. G., and Binnekamp, J. G., 1978, Recent geological advances concerning the H. Y. C. and associated deposits, McArthur River, N. T.: Geol. Soc. Australia Jour., v. 24, p. 365-380.

Walter, M. R., 1972, Stromatolites and the biostratigraphy of the Australian Precambrian and Cambrian: Palaeontological Assoc. London, Spec. Pap in Palaeontology no. 11.

Weiner, W. F., and Koster van Groos A. F., 1976, Petrographic and geochemical study of the formation of chert around the Thornton reef complex, Illinois: Geol. Soc. America Bull., v. 87, p. 310-318.

Weiss, M. P., 1954, Feldspathized shale from Minnesota: Jour. Sed. Pet., v. 24, p. 270-274.

Williams, P. F., Collins, A. R., and Wiltshire, R. G., 1969, Cleavage and penecontemporaneous deformation structures in sedimentary rocks: Jour. Geol. v.77, p.415-425.

Williams, I. R., 1968, Yarraloola, W.A.: West. Australia Geol. Survey 1:250 000 Geol. Series Explan. Notes.

Williams, I. R., Brakel, A. T., Chin, R. J., and Williams, S. J., 1976, The stratigraphy of the eastern Bangemall Basin and the Paterson Province: West. Australia Geol. Survey Ann. Rept 1975, p.79-82.

Williams, H., Turner, F. J., and Gilbert, C. M., 1954, Petrography—an introduction to the study of rocks in thin sections: San Francisco, Freeman.

Williams, I. R., and Williams, S. J., 1977, Explanatory notes on the Gunanya 1:250 000 Geological Sheet, W.A.: West. Australia Geol. Survey Rec. 1977/7.

Williams, I. R., Williams, S. J., and Hocking, R. M., 1980, Explanatory notes on the Glenburgh 1:250 000 geological sheet, W.A.: West. Australia Geol. Survey Rec. 1980/4.

Williams, S. J., Williams, I. R., Chin, R. J., Muhling, P. C., and Hocking, R. M., 1979, Explanatory notes on the Mount Phillips 1:250 000 geological sheet, W.A.: West Australia Geol. Survey Rec. 1978/13.

Wilson, R. C. L., 1975, Some examples of shoaling deposits from the Upper Jurassic of Portugal, *in* Tidal deposits (R. N. Ginsberg, ed.): New York, Springer, p. 363-371.

Winkler, H. G. F., 1976, Petrogenesis of Metamorphic Rocks: New York, Springer.

Wood, G. V., and Wolfe, M. J., 1969, Sabkha cycles in the Arab/Darb formation off the Trucial Coast of Arabia, *in* Lithification of carbonate sediments, 2, Sedimentology, v. 12, no. 3-4, p. 165-191.

Woodcock, N. H., 1976, Structural style in slumps: Geol. Soc. London, Jour., v. 132, p.399-415.

Woodward, H. P., 1891, Geognosy. A general description of the Victoria, Murchison, Gascoyne, Ashburton, Fortescue, Roebourne and De Grey districts: West. Australia Ann. General Rept of the Government Geologist for the year 1890, p.6-12.

Wunderlich, F., 1967, Die Entstehung von 'convolute bedding' and Platenrandern: Senchenbergiana Lethaea, v. 48, p. 345-349.

Wyatt, B. W., 1977, Aeromagnetic survey of Glenburgh, Robinson Range, Peak Hill, Nabberu and Stanley 1:250 000 Sheet areas, Western Australia 1972: Australia BMR Rec. 1977/35.

SUBJECT INDEX

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AUTHOR INDEX

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REFERENCE
CAINOZOIC
QUATERNARY
PALAEOZOIC
PERMIAN
PROTEROZOIC
MIDDLE PROTEROZOIC
BANGEMALL GROUP
ARCHAEAN AND EARLY PROTEROZOIC
Q Alluvium, colluvium and saline lake deposits : Gravel, talus, sand, silt, clay, mud, salt and wind-blown gypsum
Qs Wind-blown sand
Cz Hardpan (consolidated colluvium and alluvium), laterite and lateritic detritus mixed with wind-blown sand; minor silcrete
Czk Valley calcrete
P PATERSON FORMATION : glacigene diamictite and conglomerate
b Basic sills : dolerite; minor gabbro and basalt phases; includes DAVIS DOLERITE
WANDARRY CREEK – BULLOO DOWNS
PARRY RANGE – TANGADEE
TANGADEE – KELLY RANGE AND ROBERTSON RANGE
ROBERTSON RANGE – DAVIS RIVER
KELLY RANGE AND ROBERTSON RANGE – CARNEGIE AND McFADDEN RANGE
MUCALANA SUB-GROUP
TOP CAMP FORMATION
EDMUND SUB-GROUP
COLLIER SUB-GROUP
MANGANESE SUB-GROUP
KAHRBAN SUB-GROUP
ᎮS DURBA SANDSTONE : Quartz arenite; minor basal conglomerate lenses
ᎮMq KURABUKA FORMATION : Shale, mudstone and chert, usually greenish; minor limestone and calcareous sandstone
ᎮMz ILGARARI FORMATION : Siltstone, claystone, shale and fine-grained sandstone
ᎮMy CALYIE SANDSTONE : Quartz arenite, cross-bedded over most of its extent; minor wacke; siltstone, shale, conglomerate and rare dolomite
ᎮMn MT VERNON SANDSTONE : Fine and medium-grained arenite; rare conglomerate; local siltstone and shale
ᎮMs FORDS CREEK SHALE : Green and dark grey shale and mudstone; subordinate arenite, wacke, dolomite and chert
ᎮMm McFADDEN SANDSTONE : Laminated feldspathic and quartz arenite and wacke; lesser conglomerate, siltstone and claystone; prominent cross-bedding
ᎮMb BACKDOOR FORMATION : Shale and siltstone; lesser chert, claystone, dolomite and sandstone
ᎮMg MANGANESE SUB-GROUP : Conglomerate and sandstone (COONDOON CONGLOMERATE and BEE HILL SANDSTONE); green to grey shale with some sandstone and chert (BALFOUR SHALE); chocolate shale and siltstone with braunite pellets (NOREENA SHALE)
ᎮMc COODARDOO FORMATION : Arenite and wacke with graded bedding and dewatering structures; lesser siltstone, shale and aphanitic mudstone
ᎮMu CURRAN FORMATION : Grey to black, cream-weathering shale and mudstone, very minor chert, limestone, dolomite, and fine-grained sandstone
ᎮMt TOP CAMP FORMATION : Dolomite, limestone, shale, dolomitic shale, chert, quartz sandstone, breccia; minor conglomerate lenses at base
ᎮMl ULLAWARRA FORMATION : Shale, siltstone and fine-grained arenite; minor wacke, claystone, dolomite and chert
ᎮMf NANULAR SANDSTONE : Quartz arenite; lesser quartz siltstone and wacke
ᎮMv DEVIL CREEK FORMATION : Dolomite and shale, minor siltstone, fine-grained sandstone, breccia and chert
ᎮMh SKATES HILLS FORMATION : Algal dolomite, sandstone, shale, siltstone and conglomerate
ᎮMd DISCOVERY CHERT : Massive, laminated and fissile chert, minor shale
ᎮMr MARLOOYANOO FORMATION : Laminated siltstone and shale, feldspathic and quartz arenite
ᎮMx COONABILDIE FORMATION : Siltstone and medium-grained sandstone; minor shale, chert and carbonate
ᎮMw WONYULGUNNA SANDSTONE : Quartz arenite, minor cobble conglomerate; local siltstone, shale and chert
ᎮMj JILLAWARRA FORMATION : Shale and siltstone; lesser claystone, chert, dolomite and sandstone
ᎮMk KIANGI CREEK FORMATION : Quartz arenite, shale and siltstone; minor dolomite, conglomerate and chert
ᎮMi IRREGULLY FORMATION : Dolomite and shale, lesser sandstone, conglomerate, breccia and chert
ᎮMo COOBARRA FORMATION : Arenite, commonly coarse-grained; coarse conglomerate; lesser siltstone, shale, wacke and minor rhyolite
ᎮMe TRINGADEE FORMATION : Arenite, commonly pebbly; interbeds of conglomerate, siltstone, and near the top, dolomitic rocks
ᎮMa MT AUGUSTUS SANDSTONE : Coarse-grained quartz arenite; local medium and fine-grained arenite, wacke, arkose and siltstone
ᎮMtp PRAIRIE DOWNS MEMBER : Quartz arenite, arkose and conglomerate
ᎮU UAROO GROUP : Sandstone, dolomite, shale and chert
ᎮB BRESNAHAN GROUP, MOUNT MINNIE GROUP AND CAPRICORN SANDSTONE : Sandstone, conglomerate, local dolomite and shale
ᎮO SCORPION GROUP : Dolomite, sandstone, siltstone, shale and conglomerate
ᎮC CORNELIA SANDSTONE : Siliceous quartz sandstone; minor siltstone, shale and granule conglomerate
ᎮY YENEENA GROUP : Quartz sandstone, conglomerate, siltstone, shale and dolomite
ᎮL OLDER SEDIMENTARY BASINS (NABBERU BASIN, ASHBURTON TROUGH AND HAMERSLEY BASIN): Shale, siltstone, sandstone, iron formation, dolomite, limestone, conglomerate, chert, basalt and dacite, in part metamorphosed
A-Ꮾ METAMORPHIC AND GRANITOID PROVINCES (GASCOYNE PROVINCE, MARYMIA DOME, SYLVANIA DOME and other inliers): Gneiss, metasediments, migmatite, granitoid and greenstone
SYMBOLS
Geological boundary
Fault
Dolerite dyke
Quartz dyke
Syncline
Anticline
Overturned anticline
Overturned syncline
Mining centre
Mine, not being worked
Mineralization, occurrence or prospect
Asbestos A
Copper Cu
Gemstone Gs
Gold Au
Lead Pb
Manganese Mn
Uranium U
Zinc Zn
Highway with route marker 95
Road or track
Homestead Billinnooka
Watercourse
Spring or soak
Bore or well
INDEX TO 1 : 250 000 SHEETS
YANREY
WYLOO
MT BRUCE
ROY HILL
BALFOUR DOWNS
RUDALL
TABLETOP
WINNING POOL
EDMUND
TUREE CREEK
NEWMAN
ROBERTSON
GUNANYA
RUNTON
KENNEDY RANGE
MT PHILLIPS
MT EGERTON
COLLIER
BULLEN
TRAINOR
MADLEY
WOORAMEL
GLENBURGH
ROBINSON RANGE
PEAK HILL
NABBERU
STANLEY
HERBERT
LOCALITY MAP
WESTERN AUSTRALIA
Roebourne 248 km
Carnarvon 265 km
Newman 27 km
Meekatharra 115 km
LAKE DISAPPOINTMENT
HON. P.V. JONES, M.L.A.
MINISTER FOR MINES
A.F. TRENDALL, DIRECTOR, GEOLOGICAL SURVEY
GEOLOGICAL MAP
OF THE
BANGEMALL BASIN
SCALE 1:1 000 000
KILOMETRES

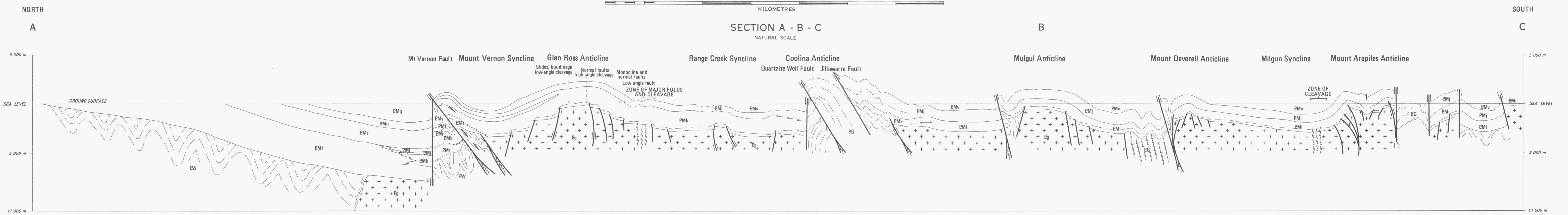

HON. P.V. JONES, M.L.A.
MINISTER FOR MINES
A.F. TRENDALL, DIRECTOR, GEOLOGICAL SURVEY
STRUCTURE OF THE BANGEMALL BASIN
SCALE 1 : 1 000 000
20 0 20 40 60 80 100 120
KILOMETRES
REFERENCE
Unconformity
Fault
Accurate
Inferred
Normal, apex of triangle shows dip
Low – angle reverse fault (dip < 45°), apex of triangle shows dip
High – angle reverse fault (dip > 45°), apex of triangle shows dip
Folds
Anticline with plunge
Overturned
Syncline with plunge
Overturned
Regional anticline
Bedding plane trend
Structural form line
Strike and dip of bedding
1° – 14°
15° – 29°
30° – 75°
Dolerite dyke
Quartz dyke
q
Structural domain
A
Major lineament
Flint Hill
Minor lineament
Basement inlier
MULGUL
Line of structural cross – section
A B
Prospect
Mineral occurrence
Copper Cu
Gold Au
Lead Pb
Manganese Mn
Uranium U
Variscite (phosphate) P
Zinc Zn
BALFOUR
BOONDAWARRI
CORNELIA
EGERTON
MULGUL
SYLVANIA
DOME
MARYMIA
SCORPION
Flint Hill
Tangadee
Hibernian Anticline
Coolinbar Syncline
Mt Vernon
Lofty Range
Bromby Creek
LOCALITY MAP
WESTERN AUSTRALIA
TECTONIC UNITS WITHIN THE BANGEMALL BASIN
SCALE 1: 7 500 000
100 0 100 200 300 400
KILOMETRES
REFERENCE
Fault
Anticlinorium
Synclinorium
Boundary of tectonic unit
Basement
PINGANDY SHELF
EDMUND FOLD BELT
BULLEN PLATFORM
INDEX TO 1 : 250 000 SHEETS
YANREY
WYLOO
MT BRUCE
ROY HILL
BALFOUR DOWNS
RUDALL
TABLETOP
WINNING POOL
EDMUND
TUREE CREEK
NEWMAN
ROBERTSON
GUNANYA
RUNTON
KENNEDY RANGE
MT PHILLIPS
MT EGERTON
COLLIER
BULLEN
TRAINOR
MADLEY
WOORAMEL
GLENBURGH
ROBINSON RANGE
PEAK HILL
NABBERU
STANLEY
HERBERT

STRUCTURAL CROSS-SECTION BANGEMALL BASIN

SCALE 1:250 000

5 0 5 10 15 20 25 30

KILOMETRES



NOTES

1. Plane of section vertical
2. Line of cross-section A-B-C on plate 2

REFERENCE

Bangemall Group

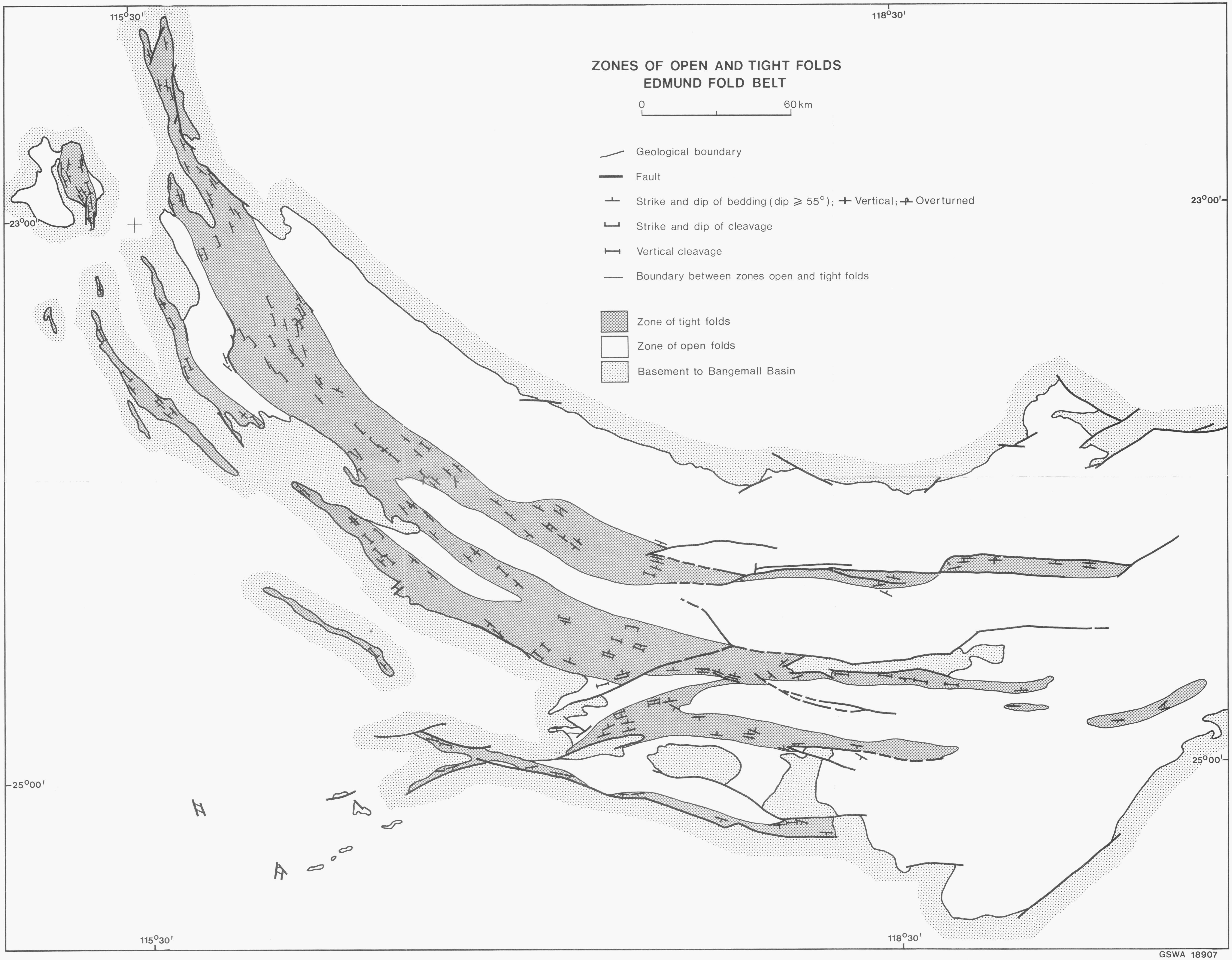
PMq	KURABUKA FORMATION
PMn	MOUNT VERNON SANDSTONE
PMs	FORDS CREEK SHALE
PMl	ULLAWARRA FORMATION
PMv	DEVIL CREEK FORMATION
PMj	JILLAWARRA FORMATION AND DISCOVERY CHERT
PMk	KIANGI CREEK FORMATION
PMi	IRREGULLY FORMATION
PMt	TOP CAMP FORMATION
Pg	Granitoid and gneiss
PG	GLENGARRY GROUP
PW	WYLOO GROUP

SYMBOLS

Shear belt

Facies change

Bedding plane


ZONES OF OPEN AND TIGHT FOLDS
EDMUND FOLD BELT
0
60 km
Geological boundary
Fault
Strike and dip of bedding (dip ≥ 55°); Vertical; Overturned
Strike and dip of cleavage
Vertical cleavage
Boundary between zones open and tight folds
Zone of tight folds
Zone of open folds
Basement to Bangemall Basin
115°30′
118°30′
23°00′
23°00′
25°00′
25°00′
115°30′
118°30′

GSWA 18907