GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

BULLETIN 131

GEOLOGY OF THE EASTERN PART OF THE NABBERU BASIN WESTERN AUSTRALIA

DEPARTMENT OF MINES
WESTERN AUSTRALIA

GEOLOGY OF THE EASTERN PART
OF THE NABBERU BASIN WESTERN AUSTRALIA

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FRONTISPIECE

False-colour, enhanced Landsat image of the western part of the Earaheedy Sub-basin. Compare with geology as shown on Plate 1. The dark iron-formations of the Frere Formation are clearly visible along the northern and southern margins and western closure of the sub-basin. The Teague Ring Structure lower right disrupts the southern margin. Rubbly, vegetation-covered hills of Princess Ranges Quartzite show as a brown tone at right side of image, centre of synclinorium. In the lower left corner lateritized shale and dolerite of the Glengarry Group have a patchy green pattern, while immediately to the right (separated by a north-south creek) lateritized Archaean greenstones have a similar pattern. Dune fields (yellow-brown) surround hills of Bangemall Group sandstone (dark-grey) at centre top.

The Landsat image is reproduced by the courtesy of the Broken Hill Proprietary Company Limited.

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BULLETIN 131

GEOLOGY OF THE EASTERN PART OF THE NABBERU BASIN WESTERN AUSTRALIA

by

J. A. BUNTING

AND

STROMATOLITES AND BIOGENIC ACTIVITY IN THE NABBERU BASIN

by

KATHLEEN GREY

GOVERNMENT PRINTING OFFICE

PERTH 1986

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FOREWORD

This bulletin represents another in the series of documentary works on the geology of major regions in Western Australia. The work was undertaken between the years 1975 and 1977, and was associated with the regional mapping program in the area between the Yilgarn and Pilbara Blocks.

The work synthesizes the geological information of the eastern part of the Nabberu Basin, giving particular emphasis to sedimentation and structural development. Of specific interest, more from the aspect of wider correlations, is the occurrence of granular iron-formations whose textures and structures are fully described in this bulletin.

It is expected that this regional synthesis will provide the basis for the geological models used in further mineral exploration of the area.

December, 1984.

A. F. TRENDALL

Director

Geological Survey of W.A.

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SUMMARY

The sedimentary and minor volcanic rocks of the Nabberu Basin unconformably overlie Archaean basement (Yilgarn Block to the south, and several smaller inliers), and are unconformably overlain by the Proterozoic Bangemall Basin (depositional age about 1.0 Ga) and the Phanerozoic Officer Basin. The Glengarry Group (depositional age about 1.8-2.0 Ga), occupying most of the western part of the basin, contains a shelf sequence (quartz arenite, shale, carbonate) in the south, and a thicker trough facies (greywacke, shale, arkose, basalt) in the north. The group was folded prior to deposition of the Padbury Group (iron-formation, shale, carbonate) in the west and the partly contemporaneous Earaheedy Group in the east.

The Earaheedy Group (depositional age about 1.8 Ga) is between 5000 m and 6000 m thick. Deposition began in the northwest with a marine transgression which progressed southeastwards, leaving a veneer of sandstone, shale, and carbonate(Yelma Formation) over the whole Earaheedy Sub-basin. An extensive marine shelf persisted, on which iron-formation, shale, chert and carbonate were deposited (Frere Formation). The iron-formation consists essentially of chert and hematite, recrystallized to quartz and magnetite (now martite) in the north. Textures are characteristically granular, with peloidal, intraclastic, oolitic and stromatolitic varieties—all indicative of shallow-water deposition above wave base. Localized banded iron-formation indicates deeper water. The percentage of iron, the iron-formation: shale ratio, and the amount of banded iron-formation all increase to the northwest, suggesting deeper water and a source of iron in that direction, whereas clastic input was from the southeast. A regressive phase followed (Windidda Formation), during which lagoonal carbonates, muds and local evaporites were deposited in the southeast part of the sub-basin, culminating in partial emergence. Shelf conditions, represented by shales, persisted elsewhere.

A second transgressive cycle followed, with ferruginous sandstones and shale (Wandiwarra Formation) being deposited on a deepening shelf. A diachronous regressive phase (Princess Ranges Quartzite) resulted in mature quartz arenite and siltstone forming in a shallow-tidal (locally deltaic?) environment, to be followed by very shallow-water deposition of lithic and feldspathic sandstone and shale (Wongawol Formation) in an enclosed basin. Carbonate lenses developed, and eventually became a major component (Kulele Limestone), forming cycles of stromatolitic limestone and fine-grained clastics. Finally a thin sequence of sandstone with minor shale was deposited (Mulgarra Sandstone). Palaeogeographic evidence from most of these formations indicates a shoreline to the east and south, with deeper water to the northwest.

The overall structure of the basin is an arcuate asymmetrical synclinorium, trending approximately east-west. To the south of the axis the rocks are gently folded (Kingston Platform) where they overlie stable Yilgarn Block. To the north they are more tightly folded and cleaved in the Stanley Fold Belt, which, to the west, merges with the complex mobile belt of the Gascoyne Province.

The Earaheedy Sub-basin is poorly endowed with mineral occurrences. Patchy iron enrichment is present in the Frere Formation, galena occurs in stromatolitic dolomite of the Yelma Formation, and there is minor manganese mineralization. Several environments have been explored for uranium, including unconformities (compare with Pine Creek Geosyncline deposits), conglomerates (Blind River model), calcretes, and the cryptoexplosive Teague Ring Structure.

The Nabberu Basin can be considered as an aulacogen—an incipient continental rift that failed to develop—related to possibly more active tectonic opening in the Gascoyne Province and beyond. The situation is comparable with aulacogens in Proterozoic basins in other parts of the world, particularly North America, and is characteristic of this transitional period between the Archaean and Phanerozoic styles of crustal evolution.

CHAPTER 1

Introduction

The Nabberu Basin is an arcuate belt of Proterozoic (about 1.6-2.0 Ga) sedimentary and minor volcanic rocks flanking the northern margin of the Archaean Yilgarn Block. This bulletin describes the author's work on the stratigraphy and sedimentation of the eastern part of the basin (Earaheedy Sub-basin) and, for the purpose of providing a regional framework, summarizes work by the author and others on the remainder of the basin. A chapter on basin development summarizes the depositional and tectonic history of the basin. Stromatolites, which are a significant component of some units, are described in Appendix 1 by K. Grey.

LOCATION, COMMUNICATIONS AND ACCESS

The area of the Nabberu Basin is sparsely populated, and the few permanent residents are engaged in the pastoral industry at a number of widely scattered sheep and cattle stations (Fig. 1). The nearest permanent towns are Wiluna (population 221; 1981 census) and Meekatharra (population 989), which lie to the south of the basin, on the Yilgarn Block. Both towns were once important gold-mining centres, but now are mainly regional service centres. Small mining centres (gold, manganese, and copper) proliferated in the western part of the area, and a substantial mining town existed at Peak Hill; however all of these are now abandoned, although some gold is still mined around Peak Hill itself. The eastern part of the Nabberu Basin, which is the main subject of this bulletin, contains no identified economic mineral deposits.

The Great Northern Highway, passing through the western part of the Nabberu Basin, is a good sealed road that connects Perth and Newman. Good graded-gravel roads link Wiluna with Meekatharra and other towns to the south. Most pastoral stations are serviced by graded roads, and a network of station and exploration-company tracks provides moderately good local access to most areas of interest. Off-road driving by suitably equipped four-wheel-drive vehicles is feasible, except in the most rugged hill areas; however many parts of the area are remote and lack surface water, and extreme care and preparation are needed. This is especially true of the northern and eastern parts of the area, which form the fringes of the Gibson and Great Victoria Deserts. Access in these areas is by rough tracks, no longer maintained, which lead to Halls Creek (Canning Stock Route), Warburton (Gunbarrel Highway) and Laverton.

CLIMATE AND VEGETATION

The climate is semi-arid to arid; the mean annual rainfall is between 200 and 240 mm, and is unreliable. The area is subject to periods of drought as well as

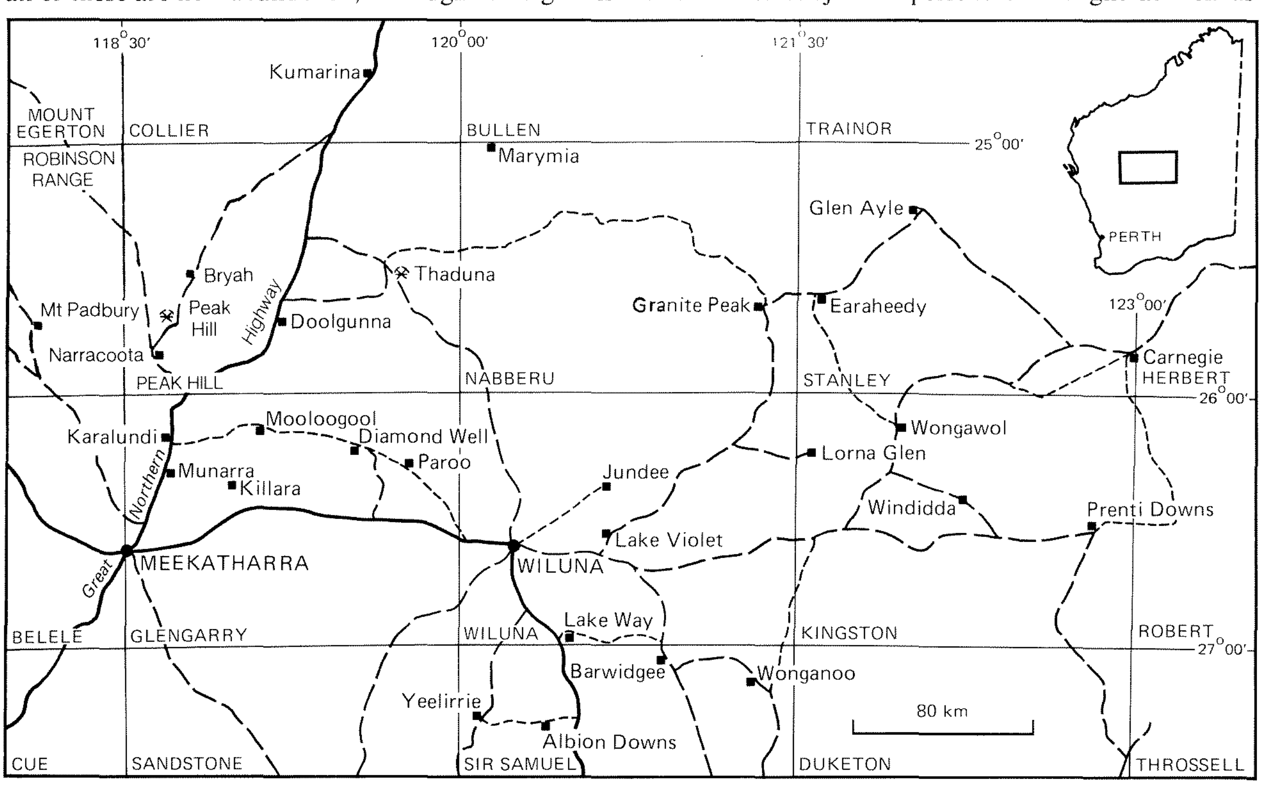

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Figure 1. Access and index to 1:250 000 sheets.

localized short-term floods, and rain may come from summer cyclones and winter depressions. Average potential evaporation is between 2 400 and 3 000 mm.

Summers are very hot, with an average January maximum of between 35 and 40°C, and a minimum of between 20 and 23°C. Winters are mild, with an average July maximum of 20°C and minimum of 6°C. Frosts are a common feature on cloudless nights in winter.

Vegetation communities (mapped by Beard, 1974) correspond closely to physiographic units. Stony hills and colluvium support low mulga (*Acacia aneura*) and other small shrubs, or soft spinifex (*Plectrachne melvillei*) and wattle (*Acacia* sp.). Sheetwash plains and flood plains are covered with open mulga woodland and a ground cover of grasses, whereas major watercourses are commonly lined with tall river gums (*Eucalyptus camaldulensis*) and have an understory of *Eremophila* sp. Desert sand plain is covered with spinifex (*Triodia* sp.) and scattered low mallee (*Eucalyptus* sp.). Areas marginal to salt lakes support halophytes such as samphire (*Arthrocnemum* sp.) and saltbush (*Atriplex* sp.), with *Casuarina* sp. in alluvial channels and calcrete areas.

PHYSIOGRAPHY AND CAINOZOIC GEOLOGY

The land surface is characterized by generally low relief, and elevation rises from about 440 m in the low-lying areas around Lake Carnegie to about 600 m in some of the ranges to the west. Local relief is seldom more than 100 m, although sandstone of the Bangemall Group in the Carnarvon Range rises some 300 m above the surrounding plain, to a maximum elevation of 903 m (Mount Methwin).

Resistant rocks, such as quartzite and iron-formation, form cuestas where the rocks are gently dipping (e.g. Princess Ranges, Frere Range, Timperley Range), or elongate hills and ridges where they dip more steeply (e.g. Mudan Hills, Lee Steere Range). Low ground between cuestas and ridges generally corresponds to shaly units. Shale also forms low rounded hills, whereas carbonate units weather to more rugged topography. Mesas and breakaways (cliffs up to 20 m high) form where shale is deeply weathered, and are most characteristic of the outliers of flat-lying Permian sediments. The mesas and breakaways are capped by duricrust which represents remnants of an older Tertiary erosion surface (the "Older Plateau" of Jutson, 1934).

Gently sloping pediments of rock fragments in loamy soil surround most outcrop areas. These pass downslope into extensive sheet-wash plains, where soils are thicker and rock fragments are less abundant. Broad, ill-defined drainages are filled with alluvium, and in their lower reaches, incised watercourses (typically lined with large eucalypts) have been cut into the alluvium.

All streams and lakes in the area are ephemeral, and flow occurs only after heavy rain. Some drainage lines are now inactive, and, together with the major salt-lake systems, they form part of an extensive palaeodrainage system which ceased significant flow in the middle Miocene (van de Graaff and others, 1977). Tributary drainages are preserved on the lateritized Tertiary erosion surface. Saline playa lakes represent the infilling of the trunk drainages of the palaeodrainage system, and receive most of the current drainage. A major playa-lake system, consisting of Lakes Gregory, Nabberu, Teague, Carnegie, and Wells, crosses the area from west-northwest to east-southeast. Although now broken into several internal-drainage basins, the entire system represents a major palaeoriver which once flowed into the Cretaceous sea in the Officer Basin. Following a marine regression, the river flowed via Lake Throssell into the Eocene sea in the Eucla Basin. Subsequently it was captured by the Lake Disappointment system and flowed northwards into the Indian Ocean (van de Graaff and others, 1977). In the far northeastern part of the Nabberu Basin, drainage is northeasterly into the Lake Burnside - Lake Disappointment system. Presently active streams, flowing southwards into Lake Carnegie (e.g. Sholl Creek), have captured part of this northeast-flowing system in the lake system south of Lee Steere Range.

In these trunk-drainage systems three physiographic units are present:

(a) flat, bare, salt lakes, which are covered with a few centimetres of water after heavy rain;

(b) dunes and sheets of eolian and alluvial material marginal to the salt lakes;

(c) calcreted valley floors, usually tributaries of, or marginal to the main salt lakes. In some areas the calcrete is being eroded by present-day drainage, leaving remnants of calcrete above the newly eroded valley floor.

A final physiographic unit is the desert sand plain. This usually occupies an intermediate position between the high-relief, outcrop areas and the major valley floors. The sand is eolian, and typical of the desert areas to the east, with longitudinal or seif dunes prominent in many areas. However, within the Nabberu Basin the amount of sand plain is small and limited to the fringes, where it is associated with Archaean granitoid rocks, the younger Bangemall Group sandstones, and Permian sediments.

PREVIOUS INVESTIGATIONS

The earliest reports on the area were by explorers. In 1874, John Forrest's route took him from south of the Kimberley Range, through the Frere Range to Weld Spring; then south towards the Timperley Range before turning east. Lawrence Wells traversed from south of the Von Treuer Tableland westwards towards Wiluna in 1892, returning in 1896 to travel from Wiluna around the northern shore of Lake Carnegie on his way to the Fitzroy River. Also in 1896, David Carnegie crossed the southeast corner of the area on his journey from Coolgardie to Halls Creek and return. Many of the most prominent geographical features in the area were named by these early explorers, the journeys of whom are summarized by Feeken and others (1970).

The first geological accounts of the area were by H. W. B. Talbot (1910, 1920), who accompanied A. W. Canning in the first preliminary survey of the stock route between Wiluna and Halls Creek, and subsequently covered the area in more detail. Talbot recognized the unconformity between the Bangemall Group ("Nullaginian" in Talbot's nomenclature) and the older Proterozoic rocks. However, he mistakenly identified another unconformity corresponding to the southern edge of the Stanley Fold Belt, correlating the undeformed rocks south of this line with the "Nullaginian" rocks to the north. He also correctly identified and mapped the unconformity between the Proterozoic rocks and the Yilgarn Block.

Talbot's interpretation was accepted until the mid 1960's when regional studies by Sofoulis and Mabbutt (1963) included all the Proterozoic rocks under one unit. Horwitz and Daniels (1967) included all the Proterozoic rocks in the Bangemall Group, an interpretation which first appeared on the State geological map of 1966. This interpretation remained in vogue (Daniels and Horwitz, 1969; McLeod, 1970; Sanders and Harley, 1971; Geological Survey of Western Australia, 1975) until the major unconformity between the Nabberu Basin and Bangemall Basin sequences was "rediscovered" by Hall and Goode (1975) and Horwitz (1975a and b).

The impetus for the recognition of the Nabberu Basin came with the publication by the Australian Bureau of Mineral Resources (1973a and b), of aeromagnetic maps in the eastern part, which clearly showed the iron-formations of the Frere Formation closing around the main synclinorium. This initiated a spate of work which culminated in the presentation of four papers on the newly rediscovered basin at the First Australian Geological Convention in May 1975 (Hall and Goode, 1975; Horwitz, 1975b; Preiss and others, 1975; Walter, 1975).

The most significant of these papers was that by Hall and Goode. They named the basin, and proposed stratigraphic subdivisions (subsequently modified slightly and defined by Hall and others, 1977) which have become generally accepted. A later paper (Hall and Goode, 1978) provided a comprehensive description of the basin. Barnett (1975) also made a notable contribution in recognizing that the sedimentary sequence in the western part of the basin was Proterozoic in age.

Horwitz (1975a and b) also recognized, but did not formally name, the broad subdivisions and age of the basin. However, he considered that the rocks (subsequently called the Glengarry Group by Gee, 1979b) in the western part of the Nabberu Basin unconformably overlie the lower part of the Earaheedy Group, and correlate with the upper part (a correlation repeated by Button, 1976, and Horwitz and Smith, 1978). In contrast, Hall and Goode (1978) correlated the Glengarry Group with the Yelma Formation, whereas Bunting and others (1977) considered the Earaheedy Group to be unconformable on the Glengarry Group.

Hall and Goode (1975, 1978) and Horwitz (1975a and b) recognized the significance of the iron-formations, in the eastern part of the basin, and compared them with the Superior-type iron-formations (Gross, 1965) of North America. This comparison was strengthened by the recognition of microfossils in the Frere Formation that are identical to forms in the Gunflint and Biwabick Iron Formations (Walter, 1975; Walter and others, 1976).

Mapping programs by the Geological Survey of Western Australia (GSWA) and Bureau of Mineral Resources (BMR) covered parts of the eastern and southern fringes of the basin adjacent to the Yilgarn Block and Officer Basin (Jackson, 1978; Bunting and others, 1978; Bunting and Chin, 1979). Following Jackson's work, glauconite from the basal Yelma Formation of the Earaheedy Group was radiometrically dated at about 1.7 Ga (Preiss and others, 1975), and Preiss (1976) described stromatolites from the Yelma and Windidda Formations.

Systematic mapping at 1:250 000 scale by GSWA was carried out between 1975 and 1977, and the results published as a series of explanatory notes (Bunting, 1980; Commander and others, 1982; Bunting and others, 1982; Elias and Bunting, 1982; Elias and others, 1982; Gee, 1983). Bunting and others (1977) presented a preliminary synthesis of the work, and proposed that the Earaheedy Group unconformably overlies the "Glengarry Axial Sequence" to the west. Gee (1979b) defined new formations in the Glengarry Group, and redefined the Padbury Group (Barnett, 1975) which is proposed as a correlate of the lower part of the Earaheedy Group. Grey (1979, 1981, and 1984) described the numerous and diverse stromatolites from the Nabberu Basin, and reclassified some of the forms described by Preiss (1976).

GENERAL GEOLOGY

The principal tectonic units in and around the Nabberu Basin are given in Figure 2, and stratigraphic units are listed in Figure 3.

Granitoid rocks, supracrustal "greenstone" belts, and gneisses form the basement to the Proterozoic sequences, and are presently exposed in the Yilgarn Block, Malmac Dome, Marymia Dome, and Goodin Dome. Most of these rocks are presumed to be Archaean in age (2.5 Ga) although some younger granites may be about 2.4 Ga old.

The Nabberu Basin contains two sub-basins, namely the Glengarry Sub-basin in the west and the Earaheedy Sub-basin in the east; it is the latter which forms the major subject of this bulletin. Most of the Glengarry Sub-basin is occupied by the Glengarry Group, which consists of a thin shelf facies (quartz arenite, shale, carbonate) in the south, and a thicker trough facies (wacke, shale, basalt, arkose) in the north. Unconformably overlying this, and infolded with it, is the Padbury Group, containing shale, iron-formation, minor carbonate, sandstone, and conglomerate. The Earaheedy Sub-basin is largely occupied by the Earaheedy Group, although there is a lower sequence of shale, wacke, and phyllite (the Troy Creek beds) which is correlated with the Glengarry Group. The Earaheedy Group consists of the Tooloo Subgroup (quartz arenite, iron-formation, chert, shale, and carbonate) which is correlated with the Padbury Group, and the overlying Miningarra Subgroup (sandstone, shale, and carbonate).

A major unconformity separates Nabberu Basin sequences from the younger Scorpion and Bangemall Groups. The Scorpion Group consists mainly of sandstone, shale and dolomite, and is deformed, but unmetamorphosed. The exact age of the Scorpion Group is unknown, but is probably between 1.2 Ga and 1.6 Ga. The Bangemall Group is almost certainly unconformable on the Scorpion Group, and its age is considered to be about 1.1 Ga (Gee and others, 1976).

The youngest rocks in the vicinity of the Nabberu Basin are sediments of the Officer Basin

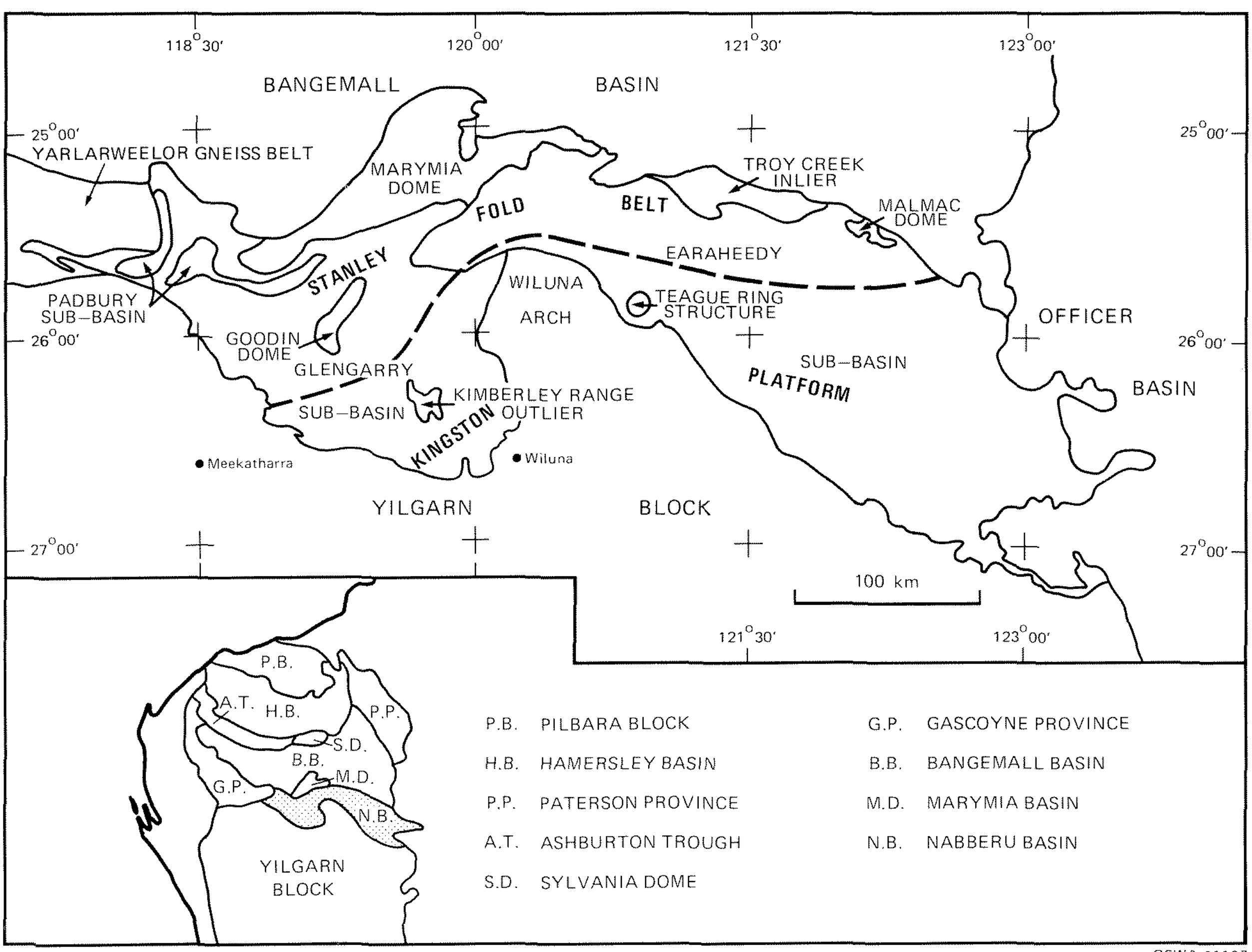

Figure 2. Regional setting and tectonic subdivisions of the Nabberu Basin.

which lap on to the eastern edge. The Early Permian Paterson Formation consists of glacigene rocks (tillite, lacustrine claystone, and fluvial sandstone). A few outcrops of the Cretaceous Bejah Claystone are present in the extreme east of the area.

Dolerite of at least two different ages intrudes the sediments of the Nabberu Basin. In the west, dolerite sills and a few dykes have been involved in the deformation of the Glengarry Group, and are probably comagmatic with basalt in the sequence. In the eastern part of the Earaheedy Sub-basin, dolerite sills intruding the Earaheedy Group have beeen dated at about 1.03 Ga (Table 1) and are probably part of the extensive suite that intrudes the adjacent Bangemall Group.

The post-Earaheedy Group rocks are not discussed further in this report. More comprehensive descriptions of the Scorpion Group are provided by Bunting and others (1982) and Commander and others (1982), of the Bangemall Group by Muhling and Brakel (1985), and of the Officer Basin by Jackson and van de Graaff (1981).

LIMITS OF THE BASIN

The Nabberu Basin is exposed over an area of approximately 60 000 km^2 and is 600 km long by 120 km wide. The total stratigraphic thickness is about 15 000 m, of which the top 4 000 m occurs mainly in the Earaheedy Sub-basin and the basal 7 000 m occurs mainly in the Glengarry Sub-basin.

None of the present boundaries represents the original depositional extent of the basin. To the south, the boundary is the unconformity with the underlying granitoid and metamorphic rocks of the Yilgarn Block, and a few outliers of Proterozoic sedimentary rocks on the Yilgarn Block were almost certainly once part of the basin.

On its western and northwestern sides, the Nabberu Basin becomes increasingly affected by the tectonism and plutonism of the Gascoyne Province. A useful boundary here is the tectonized contact with the Yarlarweelor Gneiss Belt and the Marymia Dome. Small areas of metasediments preserved within the Gascoyne Province are probably outliers of the Nabberu Basin, e.g. Mount James Formation (Williams and others, 1979).

In the north and northeast, sediments of the Nabberu Basin are unconformably overlain by, or are faulted against, those of the Bangemall Group and Scorpion Group, and the northerly extent of Nabberu Basin units beneath the Bangemall Group remains concealed.

In the east, the Nabberu Basin sequence is unconformably overlain by sub-horizontal sediments, principally the Early Permian Paterson Formation, of the western margin of the Phanerozoic Officer Basin. The surface trace of the Permian unconformity is very irregular, and, because the youngest known rocks in the Nabberu Basin are affected by this unconformity, it is reasonable to postulate that the Nabberu Basin continues for a considerable distance under the Officer Basin. Seismic evidence (Harrison and Zadoroynyj, 1978; Jackson and van de Graaff, 1981) suggests that over 10 000 m of early Proterozoic sediments may be present near the western edge of the Officer Basin.

There is still some doubt about the age of certain units near the eastern exposed margin of the Nabberu Basin; hence the 1:1 000 000 geological map (Plate 1) has been terminated arbitrarily at longitude 123° 30′E, despite the existence of undoubted Earaheedy Group further east. Stromatolitic dolomite in northeastern THROSSELL*, previously assigned to the Earaheedy Group, may be part of the Scorpion Group on the basis of stromatolite correlation (K. Grey, pers. comm.). This throws doubt on some tentative correlations in eastern ROBERT, particularly outcrops of unassigned Proterozoic sandstone east of Ida Range (Jackson, 1978), which could be either Scorpion Group or a facies change of the Windidda Formation of the Earaheedy Group.

AGE OF THE BASIN

It should be noted that Rb-Sr ages, as extracted from the cited literature, have been recalculated for this bulletin where necessary—in accordance with the revised value of $1.42 \times 10^{-11} a^{-1}$ for the decay constant of ^{87}Rb (Steiger and Jäger, 1977).

Definitive radiometric evidence for the age of the Nabberu Basin is not yet available, but numerous age determinations from rocks collected within and around the basin allow a reasonable estimate to be made (Fig. 4 and Table 1). The age of the basin must lie between the accepted age of the Bangemall Basin—about 1.1 Ga (Gee and others, 1976; Muhling and Brakel, 1985) and the youngest granitoids of the adjacent Yilgarn Block—about 2.4 Ga (Williams and others, 1978; Roddick, cited in Bunting and Chin, 1979; Stuckless and others, 1981).

Between these two extremes, several authors report radiometric ages in the range 1.57 to 1.71 Ga. In terms of depositional events, most of these results must be regarded as minimum ages, particularly the 1.57 Ga age on glauconite from the Mount Leake Sandstone (Butt and others, 1977) and the 1.63 Ga

* 1:250 000 sheet names are printed in capitals to distinguish them from place names.

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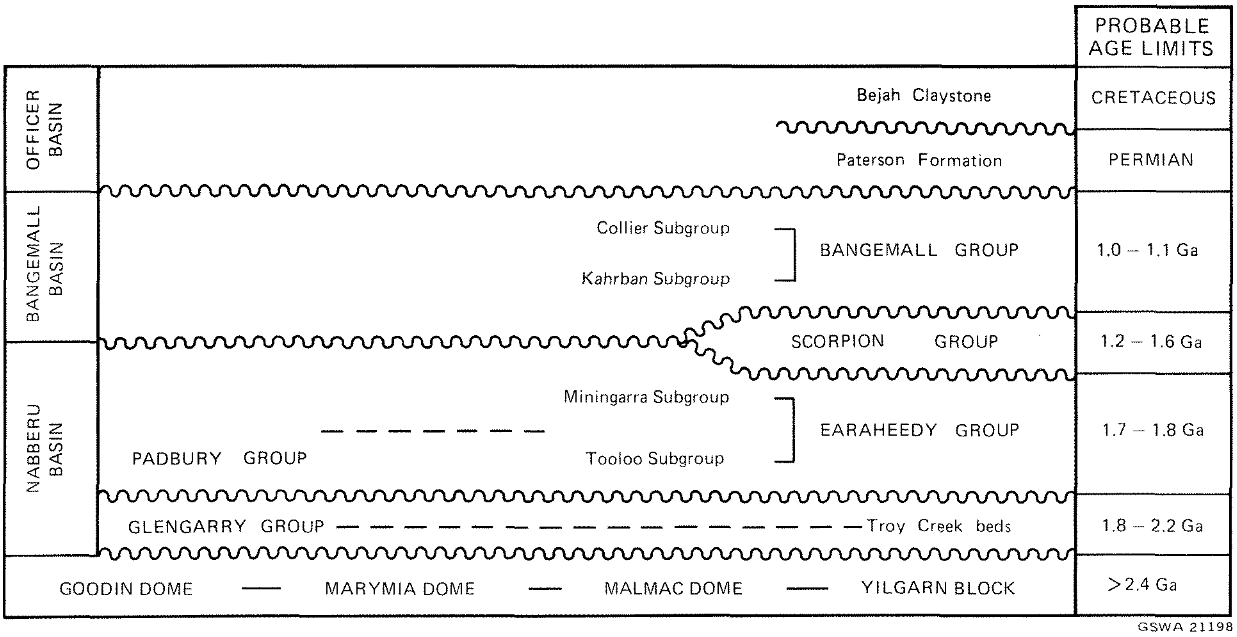

Figure 3. Summary of main stratigraphic units within and around the Nabberu Basin.

age for the quartz syenite in the Teague Ring Structure (Bunting and others, 1980). Glauconite from two localities in the Earaheedy Group gave reasonably consistent K-Ar results, ranging from 1.67 to 1.71 Ga (Preiss and others, 1975; Horwitz, 1975a), which are in broad agreement with studies on galena from the base of the Earaheedy Group (Johnston and Hall, 1980; Richards and Gee, 1985). Richards obtained an age of ca. 1.7 Ga on galena from dolomite in the Yelma Formation near Sweetwaters Well. The result falls on his revised crustal growth curve for lead, and is thus an indication of the time of crystallization of the galena. The galena occurs as fillings in fenestral fabrics, spaces between stromatolite columns, and gas bubbles, as well as veins; it is probably, therefore, of diagenetic origin, forming soon after deposition of the enclosing dolomite.

An age of 1.7 Ga appears again as a model age for metamorphic muscovite in the eastern Gascoyne Province, where it is interpreted as representing the waning phase of a period of tectonic and metamorphic activity that began about 1.8 Ga (Williams and others, 1978). This activity affected "pre-Padbury Group" supracrustals (equivalent to Glengarry Group), but not the Padbury Group. Intrusion of granitoid rocks at between 1.56 and 1.6 Ga is considered by Williams and others (1978) to have been responsible for the deformation of the Padbury Group.

The coincidence of 1.7 Ga for the maximum age of the Padbury Group and minimum age (from glauconite and galena) of the Earaheedy Group suggests that the age represents a real event, and it is regarded here as the best estimate currently available for the deposition of the Padbury and Earaheedy Groups. An age for the Glengarry Group is more difficult to determine. Deformation and intrusive events in the Gascoyne Province indicate a minimum depositional age of about 1.8 Ga but the only available maximum age is around 2.4 Ga from granite in the underlying Yilgarn Block (Williams and others, 1978).

In summary, the sequence of events in the Nabberu Basin is seen as follows:

1.7-1.56 Ga	Deformation of the Earaheedy and Padbury Groups, coincident with granitoid intrusion in the eastern Gascoyne Province.
ca. 1.7 Ga	Deposition of the Padbury and Earaheedy Groups.
1.8-1.7 Ga	Deformation of the Glengarry Group and equivalents, coincident with basement remobilization and intrusion in the Gascoyne Province.
2.0-1.8 Ga	Deposition of the Glengarry Group and equivalents.

REGIONAL CORRELATION

The Glengarry Group and Wyloo Group (in the Ashburton Trough) can be correlated. They are similar in gross lithology (Horwitz and Smith, 1978; Gee,

TABLE 1. RADIOMETRIC AGE DETERMINATIONS BEARING ON THE AGE OF THE NABBERU BASIN

Stratigraphic or Tectonic Unit	*Locality*	*Rock or Mineral*	*Age* (*x* 10^6 years)	*Method*	*Reference* (*Number refers to Fig. 4*)	
Bangemall Gp	Tangadee (COLLIER)	rhyolite	[a]1075 ± 42	Rb-Sr	Gee and others, 1976	(1)
Intrusion into Earaheedy Gp	Central ROBERT	dolerite	[a]1028 ± 50	Rb-Sr	Compston, 1974	(2)
Mt Leake Sst	SW PEAK HILL	glauconite	1573	K-Ar	Butt and others, 1977	(3)
Teague Ring Str.	SE NABBERU	qz syenite	1630 approx	Rb-Sr	Bunting and others, 1980	(4)
Earaheedy Gp (Wandiwarra Fm)	NW KINGSTON	glauconite	1685 ± 35	K-Ar	Horwitz, 1975a	(5)
Earaheedy Gp (Yelma Fm)	NE DUKETON	glauconite	[a]1556 - 1674 1670 - 1710	Rb-Sr K-Ar	Preiss and others, 1975	(6)
Earaheedy Gp (Yelma Fm)	Sweetwaters Well (NABBERU)	galena	1700 approx 1700 approx	Pb-Pb Pb-Pb	Richards and Gee, 1985 Johnston and Hall, 1980	(7) (8)
Gascoyne Prov.	Central ROBINSON RANGE	adamellite adamellite muscovite (in schist)	1560 approx 1600 approx 1700 model	Rb-Sr Rb-Sr Rb-Sr	Williams and others, 1978	(9)
Yilgarn Block	SE ROBINSON RANGE	granite granodiorite	2461 ± 92 2573 ± 52	Rb-Sr Rb-Sr	Williams and others, 1978	(9)
Yilgarn Block	E SIR SAMUEL	granite qz syenite	2370 ± 100 2760 ± 210	Pb-Pb Pb-Pb	Stuckless and others, 1981	(10)
Yilgarn Block	Woorana Well (SIR SAMUEL)	qz syenite	2520 ± 13	Rb-Sr	Libby and de Laeter, 1981	(11)
Yilgarn Block	Mt Boreas (LAVERTON)	granite	[a]2430 ± 30	Rb-Sr	Roddick (written comm. in Bunting and Chin, 1979)	(12)
Yilgarn Block	Mt Keith (SIR SAMUEL)	granodiorite	[a]2632 ± 17	Rb-Sr	Roddick and others, 1976	(13)

(a) Rb-Sr results recalculated for ^{87}Rb $\lambda = 1.42 \times 10^{-11} a^{-1}$

1979a), have comparable radiometric ages for the principal deformational events, and are almost physically continuous through a series of infolded keels of metamorphosed sediments in the Gascoyne Province (Williams and others, 1979; Williams, in press). The eugeosynclinal belt containing the Wyloo and Glengarry Groups has been named the Capricorn Orogen by Gee (1979a). Horwitz and Smith (1978), Hall and Goode (1978), and Goode (1981) correlate the iron-formations of the Frere Formation with those of the Hamersley Group. The reasons given are mainly an apparent regional-tectonic symmetry in the area between the Yilgarn and Pilbara Blocks, and a proposed sedimentational model in which the Frere iron-formations are regarded as shallow-water, shelf-facies equivalents of those in the Hamersley Group. Data from GSWA work and other sources indicate that such a correlation is no longer tenable, and that the Nabberu Basin is younger than the Hamersley Basin. The reasoning is as follows:

(a) Foremost amongst the evidence is the discrepancy between the radiometric ages of the two basins. Since the late 1970s, it has become apparent that the Mount Bruce Supergroup—as redefined by Trendall (1979) to include the Fortescue, Hamersley, and Turee Creek Groups, and to exclude the Wyloo Group—is much older than previously thought, probably in the vicinity of 2.5 Ga for the Hamersley Group (Compston and others, 1981) and 2.75 Ga for the Fortescue Group (Trendall, 1983). These ages are in marked contrast to the postulated 1.7 Ga age of sedimentation of the Earaheedy Group and indeed are significantly older than some of the basement granites underlying the Nabberu Basin (as young as 2.37 to 2.43 Ga).

(b) As discussed above, the Wyloo Group can be correlated with the Glengarry Group. The Wyloo Group overlies the Mount Bruce Supergroup with marked unconformity (Trendall, 1979; Gee, 1979a; Horwitz, 1981) and the Glengarry Group is unconformably overlain by the Padbury and Earaheedy Groups (Bunting and others, 1977; Gee, 1979a and b; and this bulletin). Thus, on regional correlations, the iron-formations of the Hamersley and Earaheedy Groups are separated spatially and temporally by at least two major unconformities and the Capricorn Orogen.

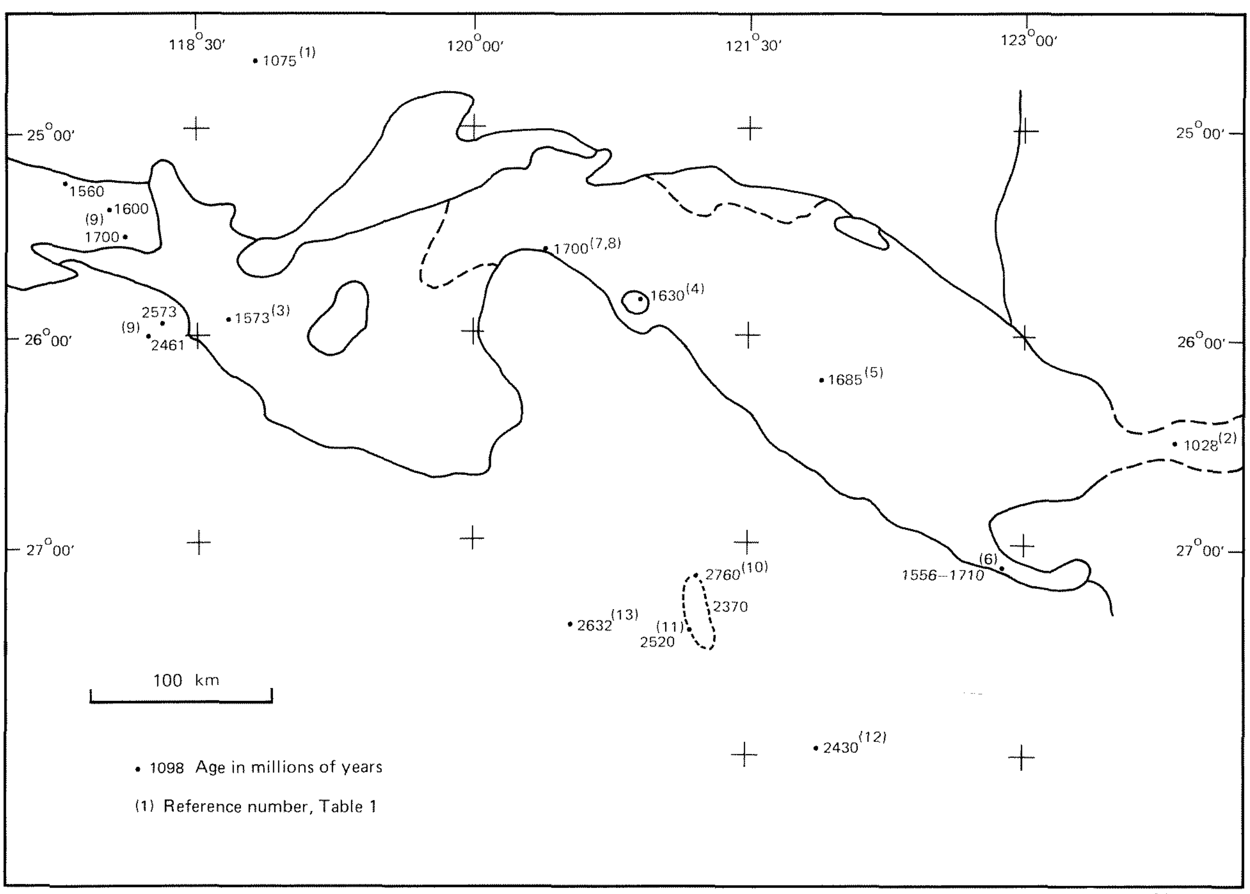

Figure 4. Distribution of radiometric age determinations in and around the Nabberu Basin.

CHAPTER 2

Pre-Earaheedy Group Rocks

CRYSTALLINE BASEMENT

YILGARN BLOCK

The Archaean Yilgarn Block forms the exposed basement to the Nabberu Basin along its entire southern margin (Fig. 2). In this area it can be described as a typical Archaean granitoid-greenstone terrain that has exhibited tectonic stability throughout the evolution of the Nabberu Basin. Its geological features need only be briefly summarized in order to establish the framework for subsequent discussions of basement influences on basin evolution.

The granitoid rocks vary in composition, texture, and grain size. Syntectonic granitoid rocks (adamellite to granodiorite in composition) display a metamorphic fabric and occur as irregular plutons within greenstone belts. Post-tectonic granitoid rocks generally lack a metamorphic fabric and commonly form extensive batholiths, in which the composition ranges from adamellite to granite. A suite of alkaline rocks, including syenite, quartz syenite, and Na-amphibole granite, features prominently in the northeastern part of the Yilgarn Block (Bunting and Williams, 1979; Libby, 1978; Stuckless and others, 1981). Radiometric age determinations reveal a range in ages for the granitoid rocks from 2.37 to 2.76 Ga as summarized in Table 1.

The greenstone belts contain a sequence of mafic and ultramafic volcanics and related intrusions, felsic volcanics and related sediments, banded iron-formation, chert, shale and coarser sediments. All are metamorphosed to varying degrees, ranging from prehnite-pumpellyite facies to amphibolite facies. Minor gneiss belts, consisting of orthogneiss, paragneiss and amphibolite, flank some of the greenstone belts, and may, in part, represent remnants of pre-greenstone basement that was remobilized during the emplacement of the granitoid rocks. The north to north-northwesterly trend of the greenstone belts is accentuated by regional folds and faults. Some of the complex major faults are termed "tectonic lineaments", and these had a major influence on deposition and structure throughout greenstone-belt development. Some show continued but reduced activity during and after the deposition of the Nabberu Basin sediments.

MARYMIA DOME

The Marymia Dome (Bunting and others, 1977) is the largest of the basement inliers affecting the Nabberu Basin. It is an elongate basement dome that forms the present northern margin of the western part of the Nabberu Basin. At its western end it becomes increasingly affected by the Proterozoic tectonic activity associated with the Gascoyne Province, and it also contains gneissic areas similar to the Western Gneiss Terrain (Gee, 1979a). However, the eastern end shows all the attributes of a segment of typical Archaean granitoid-greenstone basement. Both the Earaheedy Group and the Glengarry Group unconformably overlie the Marymia Dome, although contacts with the Glengarry Group are strongly tectonized in places.

Lithologically the rocks at the eastern end of the Marymia Dome are similar to those in the Yilgarn Block—various granitoid rocks, including both foliated and unfoliated types, and a small greenstone belt containing mafic, ultramafic and sedimentary rocks (Baumgarten area). Felsic volcanic rocks in the core of an anticline in northeast PEAK HILL are also Archaean.

Unlike the Yilgarn Block itself, structural trends in the eastern part of the Marymia Dome are predominantly northeasterly. This trend is apparent in foliations in granitoid rocks, fracture cleavage, and faulting, and is parallel to structures within the Proterozoic rocks.

MALMAC DOME

The Malmac Dome (Horwitz, 1976) is a small granitoid inlier within the eastern part of the Stanley Fold Belt. The granitoid rocks are overlain unconformably by basal sediments of the Earaheedy Group, and are presumed to be Archaean in age. They are poorly exposed and in most cases deeply weathered. Medium, even-grained, coarse-grained and porphyritic varieties can be recognized. In places the rock is strongly foliated.

GOODIN DOME

The Goodin Dome (Bunting and others, 1977) lies within the Glengarry Sub-basin, and its presence has had an influence on the structural development of nearby Earaheedy Group rocks. Basal Earaheedy Group is present only 10 km east of the dome.

The dome is a body of granitoid rock some 35 km across surrounded by highly faulted, outward-dipping sediments of the Glengarry Group. Most of the granitoid rocks are comparable with those in the Yilgarn Block to the south, although a leucocratic muscovite-biotite granite with rare fluorite is present at

Utahlarba Spring. A suite of mafic dykes cuts the granitoid rocks but not the overlying Glengarry Group.

Clear evidence for a pre-Glengarry age for the granitoid rocks includes an unconformity along the northwestern side of the dome, the termination of mafic dykes at the sediment contact, the presence of an arkose wedge in the adjacent Glengarry Group, and an absence of any intrusive effects.

TEAGUE RING STRUCTURE

The Teague Ring Structure consists of a circular core of pyroxene-quartz syenite and granite, some 10 km in diameter. It is surrounded by a ring syncline, 18-20 km across, in rocks of the Earaheedy Group. There is further discussion of this structure in Chapter 5.

GLENGARRY GROUP

Although this bulletin is concerned primarily with the Earaheedy Group, a summary of the Glengarry Group is necessary in order to place the Earaheedy Group in the context of the tectonic development of the basin. The following account is a summary of more detailed descriptions by Gee (1979b, 1983), Bunting and others (1977), Elias and Bunting (1982), and Elias and others (1982).

The Glengarry Group is exposed entirely within the Glengarry Sub-basin. It consists of a shelf facies, occupying the southern third of the sub-basin, a deeper water trough facies in the north, and a transitional fluvial to marine facies. Stratigraphic relationships are shown in Figure 5. The shelf facies has remained essentially undeformed, but deformation increases rapidly northwards in the direction of the trough.

SHELF FACIES

A basal unit (Finlayson Sandstone) of quartz arenite, with subordinate shale and silicified carbonate (chert), unconformably overlies Archaean basement along the entire southern boundary of the sub-basin. The arenite is typically mature in texture and mineralogy, and contains sedimentary structures indicative of a shallow-marine, tide-influenced environment along a transgressive shoreline.

The overlying Maraloou Formation is the lateral equivalent of much of the trough sequence to the north, and also perhaps the upper part of the Finlayson Sandstone. The dominant rock type is micaceous and thinly laminated shale, but dolomitic marl, limestone, and chert are also present. Deposition occurred in a shallow shelf sea with low carbonate banks.

TROUGH AND TRANSITION FACIES

These two facies represent the complex interrelationships between processes of erosion from rising basement domes, rapid transportation, and deposition in deep, flysch-filled troughs. The stratigraphic subdivisions are mainly those of Gee (1979b, 1983).

As with the shelf facies, the basal unit is a mature quartz arenite (Juderina Sandstone). Physical continuity between the Juderina and Finlayson Sandstones is assumed, but cannot be demonstrated, and together they are considered to have formed a thin, transgressive blanket over the entire sub-basin.

The Johnson Cairn Shale overlies the Juderina Sandstone on the northern margin of the Goodin Dome (Gee, 1983), and along the southern margin of the Marymia Dome. It is a lateral equivalent of part of the lower part of the Maraloou Formation.

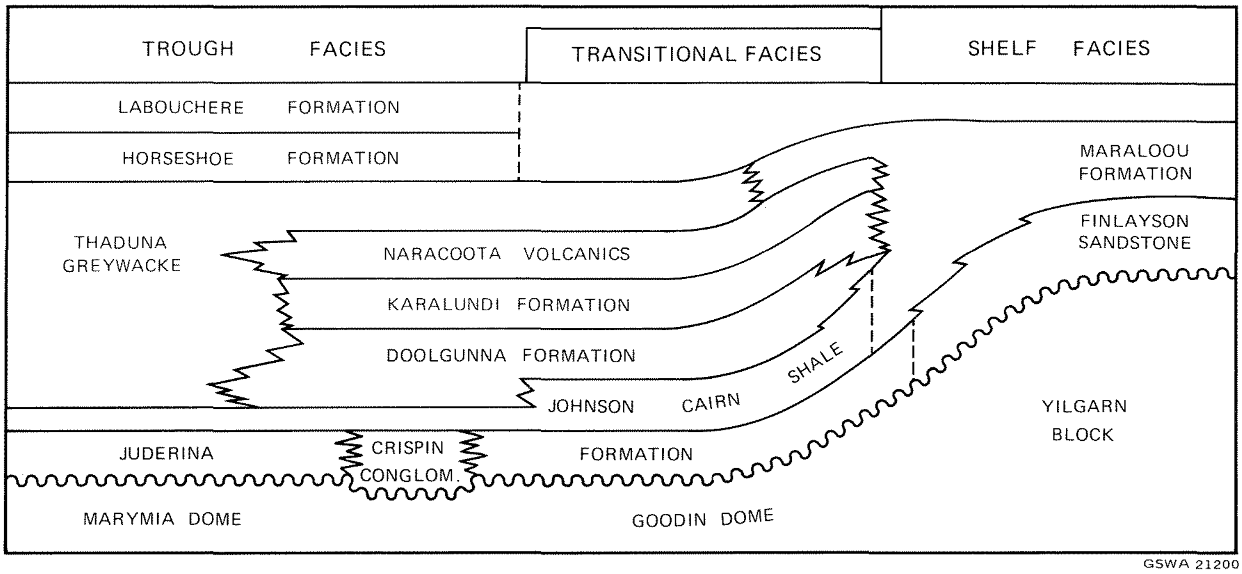

Figure 5. Diagrammatic relationships of stratigraphic units, Glengarry Group.

Around the Goodin Dome the Doolgunna Formation lies conformably on Johnson Cairn Shale. The Doolgunna Formation, which consists predominantly of arkose, and feldspathic and quartz sandstones, passes laterally into Thaduna Greywacke and thins towards the shelf to the south; it forms an immature clastic wedge around the basement domes, and encompasses a range of environments from fluvial, through a brief transitional environment (deltaic?), to deeper water where mass flow was the dominant process.

Overlying the Doolgunna Formation, the Karalundi Formation is a dominantly shaly sequence, with minor but definitive quartz wacke, chert, tuff, and carbonate. It is mainly confined to a zone running west-southwest from the northern flank of the Goodin Dome, although its equivalent (but without the distinctive tuff and chert) is present south of the dome. The tuffs are basaltic and clearly waterlaid. The chert has a colloform texture in places, contains jasper, magnetite, and micaceous hematite, and takes the form of pipes or concordant layers within shale and tuff, or veins within a thin basalt flow near the top of the formation. Gee (1979b) interprets the chert to be the result of hydrothermal activity in fumaroles associated with basaltic volcanism. Its distinctive appearance is important in correlating the Glengarry Group with the Troy Creek beds, and as evidence (boulders in conglomerate) for the relative youth of the Earaheedy Group.

Overlying the Karalundi Formation, the Narracoota Volcanics form a predominantly basaltic pile which is very thick west of the Goodin Dome, but which wedges out into the Thaduna Greywacke to the northeast. Most of the formation is massive basalt, but pillowed, vesicular and fragmental varieties are present. In the northwestern part of the sub-basin, tremolite-chlorite varieties form a high-magnesian suite.

The Thaduna Greywacke is a thick sequence of turbidite flows and shale beds which marks the deepest part of the trough. The thickest development is in the northeast, near the Thaduna copper mine, where it lies directly on quartzite correlated with the Juderina Sandstone. This thick part of the Thaduna Greywacke passes laterally into the whole Johnson Cairn-Doolgunna-Karalundi-Narracoota sequence, and its upper part laps on to and overlies the Narracoota Volcanics. Within the greywacke, lithic grains of basalt and jasper testify to a derivation from a volcanic source, such as the Narracoota Volcanics or Karalundi Formation. Thus the only difference between these greywacke turbidites and laterally equivalent arkosic turbidites of the Doolgunna Formation is the provenance of the clastic material.

The uppermost units of the Glengarry Group, the Horseshoe and Labouchere Formations, are present only in the extreme western part of the sub-basin. The Horseshoe Formation is predominantly chloritic shale with thin wacke interbeds and a distinctive banded iron-formation member. The Labouchere Formation consists of a basal quartz arenite overlain by feldspathic sandstone and phyllite.

TROY CREEK BEDS

Below the lowest continuous quartz arenite of the Yelma Formation, on the north side of the Earaheedy Sub-basin, is a belt of diverse sedimentary rocks, whose age and relationships are not clear, although similarities in lithology and stratigraphic position indicate correlation with the Glengarry Group. The rocks are tightly folded and cleaved, and in the northern part of the belt are schistose. The unit is absent over the Malmac and eastern Marymia Domes, due to onlap of the Yelma Formation, and it is not clear how the Troy Creek beds relate structurally to the domes. To the north the Troy Creek beds are faulted against the Scorpion Group and are unconformably overlain by the Bangemall Group. The southern contact is believed on structural grounds (an earlier period of deformation in some of the Troy Creek beds) to be an unconformity with the overlying Yelma Formation.

Rock types include shale, phyllite, quartz sandstone, chert, carbonate, felsic volcanics, and banded iron-formation, but due to the poor exposure and structural complexity, no stratigraphic sequence has been identified. The most common rock type is a brown, maroon or purple pelite, which may be either cleaved shale, phyllite or, in the extreme northwest, mica schist. The main cleavage trends 070°-090°, has a moderately steep northerly dip, and is parallel to that in the Earaheedy Group. An earlier cleavage in some areas gives a sub-horizontal intersection lineation with the main cleavage. A later crenulation cleavage is common.

Quartz arenite, usually with a ferruginous cement, forms single beds less than 1 m thick, or discontinuous lenses, whilst a thick unit of thick-bedded, slightly feldspathic quartz arenite and interbedded shale occurs near Imbin Rock Hole (western STANLEY). Tight parasitic folds, a regional fold closure to the west, bedding-cleavage relationships, and graded bedding, all indicate that the arenite lies in the core of an overturned, west-plunging anticline.

Figure 6. Embayed and partly rounded, subhedral quartz phenocryst in altered felsic tuff. Troy Creek beds, 15 km north of Sydney Heads Pass. GSWA thin section 46541.

In an area 15 km north of Sydney Heads Pass, felsic volcanic rocks are interbedded with shale and phyllite. The volcanics contain quartz phenocrysts which are commonly embayed and occasionally have rounded, bipyramidal terminations (Fig. 6). Some crystal tuffs contain embayed phenocrysts and fragments (several centimetres across) of fine-grained felsic and shaly material set in a quartz-mica or quartz-clay matrix.

Pink chert with colloform textures occurs as thin lenses in ferruginous pelite towards the northwest part of the belt. The chert is very similar to material forming fumarolic pipes and lenses in the Karalundi Formation of the Glengarry Group. The basaltic volcanism of the Glengarry Group is absent, although chlorite schist in the vicinity of Mount Davis could represent a mafic tuff.

Banded iron-formation (alternating microbands of chert and altered magnetite) and thin carbonate bands are minor constituents.

The stratigraphic thickness and depositional environment of the Troy Creek beds are difficult to determine; however, an order of several thousand metres in thickness is inferred, and the diversity of rock type indicates a variety of localized environments. The similarity of rock-type associations to the Glengarry Group, particularly the Karalundi Formation, suggests a similar regional environment, that is, an active and rapidly sinking basin margin that was transitional into a deeper trough.

CHAPTER 3

Earaheedy Group

STRATIGRAPHIC SUMMARY

Systematic stratigraphic nomenclature of rocks in the Earaheedy Sub-basin was first proposed by Hall and Goode (1975) and that nomenclature was formalized, with some modification, by Hall and others (1977), who also defined the Earaheedy Group and two subgroups. Principal modifications of the initial proposal were the downgrading of the Sholl Creek Formation to a member of the Wongawol Formation, and the addition of a new formation, the Mulgarra Sandstone, at the top of the sequence. The nomenclature of Hall and others (1977) is retained in this bulletin, although some further subdivision into members is made, and some formation boundaries are redefined.

The Earaheedy Group is subdivided into the Tooloo and Miningarra Subgroups on the basis of a disconformity at the top of the Windidda Formation on KINGSTON. Each subgroup represents a cycle comprising an initial transgressive phase and a terminal regressive phase, although minor subordinate transgressive/regressive cycles do occur within the subgroups. The disconformity, which is marked by a period of emergence from the dominantly marine sequence, does not occur in the northern and northwestern parts of the sub-basin. In these areas the stratigraphic position of the Windidda Formation is occupied by shale which is indistinguishable from that of the overlying Wandiwarra Formation.

The Tooloo Subgroup consists of the Yelma, Frere, and Windidda Formations. Total thickness is about 2 700 m in the southern part of the sub-basin where all three formations are present, but the Yelma Formation is very thin. In the western and northern parts of the sub-basin, the Yelma Formation is up to

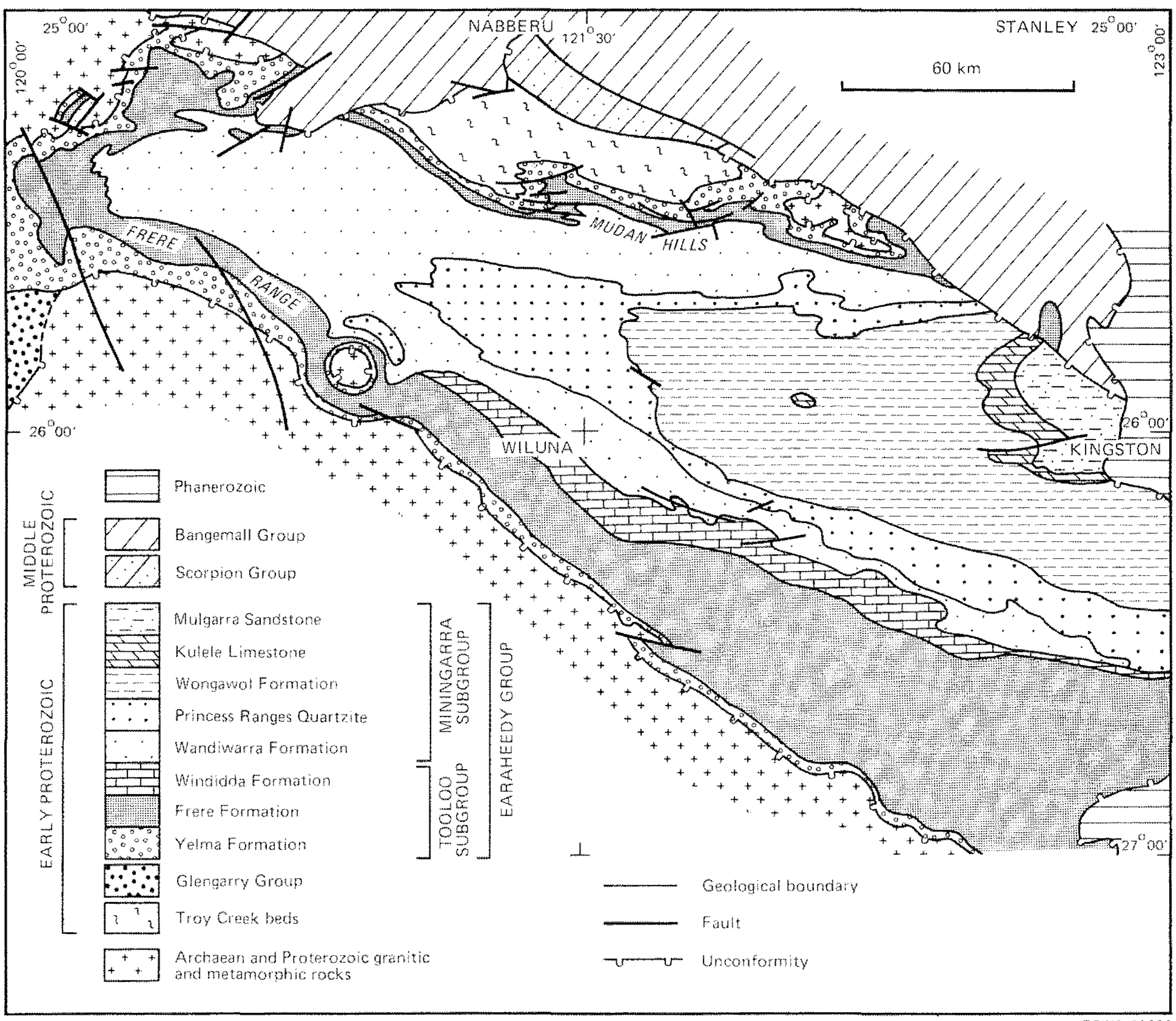

Figure 7. Generalized stratigraphic map of the Earaheedy Sub-basin.

TABLE 2. SUMMARY OF STRATIGRAPHY IN THE EARAHEEDY GROUP

Sub-group	*Formation and map symbol*	*Lithology*	*Approximate thickness*	*Stratigraphic relations*	*Deposition environment*	*Distribution*	*Remarks*
MININGARRA SUBGROUP	Mulgarra Sandstone PEm	Medium-grained quartz arenite; minor shale and limestone in middle section	100 m	Unconformably overlain by Bangemall Group. Probably disconformable on Kulele Limestone	Shallow marine	Restricted to SE STANLEY, NE KINGSTON and NW ROBERT	
	Kulele Limestone PEk	Limestone, which is variously oolitic, pisolitic, stromatolitic and brecciated, interbedded with calcarenite, sandstone, shale and rare dolomite	300 m	Base is gradational into Sholl Creek Member of Wongawol Formation	Cyclical, from shallow marine (subtidal) to tidal mudflats or lagoons	SE STANLEY, NE KINGSTON, NW ROBERT	Formerly Kulele Creek Limestone
	Wongawol Formation PEo	Arkosic and lithic sandstone, shale, with minor carbonate in upper part (Sholl Creek Member)	1500 m	Base is gradational into Princess Ranges Quartzite	Very shallow water, non-tidal (lagoon or partly enclosed sea)	Southern STANLEY, northern KINGSTON, west-central ROBERT	Includes Sholl Creek Member (Formerly Sholl Creek Formation of Hall and Goode, 1975)
	Princess Ranges Quartzite PEp	White orthoquartzite, kaolinitic siltstone and silty sandstone	280-600 m	Diachronous lower boundary with Wandiwarra Formation. Disconformable on Windidda Formation in SE	Shallow marine, tidal; localized deltaic and shoreline facies. Regressive	Follows synclinorium from central STANLEY to SE NABBERU to central ROBERT	Includes entire interval of Wandiwarra Formation in E KINGSTON and ROBERT
	Wandiwarra Formation PEw	Ferruginous quartz arenite and shale	0-1500 m	Disconformable on Windidda Formation in south; conformable on Frere Formation in west and north	Shallow marine, becoming deeper above stratigraphic level of Windidda disconformity. Transgressive	Extensive throughout NABBERU, central and NW KINGSTON and west-central STANLEY	Includes Windidda Formation equivalent where carbonates are absent
TOOLOO SUBGROUP	Windidda Formation PEd	Laminated or stromatolitic dolomite and limestone interbedded with mudstone and shale	300-800 m	Conformable on Frere Formation	Barred basin or lagoon. Regressive	Southern part of sub-basin between SE NABBERU and central ROBERT	Passes laterally into lower Wandiwarra Formation with disappearance of carbonate
	Frere Formation PEf	Granular and banded iron-formation, shale, chert, minor carbonate	1200 m	Conformable on Yelma Formation	Shallow marine. Transgressive	Continuous along N and S sides of sub-basin and around western closure	Correlates with Robinson Range Formation of Padbury Group
	Yelma Formation PEy	Quartz arenite, shale, minor carbonate, chert and conglomerate	5-1500 m	Unconformable on Archaean basement; probably unconformable on Glengarry Group and Troy Creek beds	Shallow marine, locally fluvial near base. Transgressive	Continuous around N, W and S edges of sub-basin	Includes Malmac Formation of Horwitz (1976). Probably correlates with unassigned units in Kimberley Range area on GLENGARRY and with Wilthorpe Conglomerate of Padbury Group

1 000 m thick, but the Windidda Formation is indistinguishable from, and included within, the overlying Miningarra Subgroup; consequently the total thickness in these areas is about 2 400 m. In some places, for example in the vicinity of Lake Teague, the Yelma Formation is very thin, the Windidda is absent, and the total thickness may be only about 1 100 m.

The Miningarra Subgroup contains all formations from the Wandiwarra Formation upwards, and therefore includes the Wandiwarra Formation, Princess Ranges Quartzite, Wongawol Formation, Kulele Limestone, and Mulgarra Sandstone. The subgroup ranges from about 2 500 m to 3 600 m in thickness.

A generalized map of the stratigraphy is given in Figure 7 and a stratigraphic summary in Table 2.

YELMA FORMATION

DEFINITION

The Yelma Formation is the unit of dominantly clastic sediments, at the base of the Earaheedy Group, which lies unconformably on Archaean or older Proterozoic rocks and is conformably overlain by the Frere Formation.

Derivation of name: Named after Yelma outcamp on KINGSTON (Lat. 26°31′50″S, Long. 121°41′20″E).

Type section: The type section is taken between the two points of Lat. 26°29′25″S, Long. 121°40′00″E and Lat. 26°28′20″S, Long. 121°40′00″E, a line which crosses a low escarpment approximately 6 km northwest of Yelma outcamp. The sequence in the type section is as follows:

*　*　*　*

	Thickness (m)
FRERE FORMATION, contact with Yelma Formation conformable	
YELMA FORMATION	
4. Siltstone and fine-grained sandstone, laminated; top contact not exposed at type section.	30
3. Fine-grained sandstone with chert intraclasts; rubble only.	5
2. Fine- to medium-grained quartz arenite; poorly exposed on dip slope.	110
1. Quartz arenite, medium- to very coarse-grained, pale grey, with minor feldspar; well sorted, subrounded, rare cross-bedding and ripple marks; rare, thin lenses of quartz-pebble conglomerate.	5
Total thickness of Yelma Formation	150
ARCHAEAN GRANITOID; contact with overlying Yelma Formation unconformable.	

Lithology. Mainly quartz arenite and shale, with a few lenses of stromatolitic carbonate near the top. Throughout much of NABBERU the carbonate is silicified to a cream, pale-green or grey, chalcedonic chert breccia. The quartz arenite is mature, medium to coarse grained, and may contain variable amounts of feldspar or magnetite.

Thickness and distribution: Forms an arcuate belt around the Earaheedy Sub-basin (open to the east) and varies in thickness from a few metres in the southeast, to about 1 500 m in western STANLEY, along the northern edge of the sub-basin, where a reference section is proposed in the vicinity of Mount Evelyn. The western extent of the formation, where it overlaps the Glengarry Group on PEAK HILL, is poorly defined.

Boundary criteria: The base is everywhere an unconformity, either with older Proterozoic sedimentary rocks of the Glengarry Group and Troy Creek beds or with Archaean granitoid and greenstone rocks of the Yilgarn Block, Marymia Dome, and Malmac Dome. The top is the conformable passage from shale to granular iron-formation and chert of the Frere Formation, and is taken at the base of the first extensive iron-formation. This is a modification of the original definition (Hall and others, 1977) which took the top of the Yelma Formation as the top of the uppermost arenite unit. The shale between the arenite and iron-formation has therefore changed from the Frere to the Yelma Formation.

Synonymy: The Malmac Formation (Horwitz, 1976), a sequence of clastic sediments unconformably overlying the Malmac Dome and underlying the Frere Formation, is here regarded as a continuation of, and is included within, the Yelma Formation.

LITHOLOGICAL DESCRIPTION

Because of changes in thickness and problems in defining its western limits, the Yelma Formation is described in relation to three geographically distinct areas within the Earaheedy Sub-basin.

Southern margin: The Yelma Formation can be traced intermittently for nearly 400 km from south of Lake Wells (western THROSSELL) to western NABBERU. At its type section in western KINGSTON it is about 150 m thick, including the shale member which had previously been included in the Frere Formation by Hall and others (1977). Southeast of the type section the formation is much thinner, and may be absent for a few kilometres in southern KINGSTON. In the vicinity of Lake Wells the formation is 25 to 30 m thick. It is between 30 and 50 m thick in and around the Teague Ring Structure, but thickens to the northwest where a thickness of about 400 m, of predominantly shale and carbonate with minor tuff beds, has been recorded from drill holes under Lake Nabberu (Johnston and Hall, 1980).

The typical quartz arenite of the Yelma Formation is medium to coarse grained, well sorted, subrounded to well rounded, with an authigenic silica cement. Feldspar and rounded glauconite pellets may each form up to 10% of the rock. Cross-bedding and ripple marks are locally abundant, and in northeast WILUNA the prevailing palaeocurrent direction is east-southeast.

The quartz arenite generally becomes finer grained upwards, and grades into shale and fine-grained sandstone at the top of the formation. This shaly unit was originally included in the overlying Frere Formation (Hall and others, 1977; Bunting, 1980) but is now included in the Yelma Formation because the overlying iron-formation of the Frere is a better marker than the top of the Yelma arenite. The shale/sandstone unit is thinly bedded and micaceous.

Carbonate bands and lenses occur within the Yelma Formation at a number of localities, and these commonly occur between the arenite unit and the overlying shale. In northeast DUKETON, a carbonate band of between 3 and 5 m thickness is richly stromatolitic. The predominant form was described by Preiss (1976) as *Tarioufetia yilgarnia* (Fig. 8) which is a group previously described only from younger Proterozoic rocks; however Grey (Appendix 1, and 1984) has reclassified the form into the group *Externia*. Total thickness of the Yelma Formation here is about 25 m.

Immediately east of Sweeney Creek on NABBERU, an isolated outcrop of Yelma Formation consists of fine- to medium-grained quartz arenite. Within the arenite is a band of grey limestone containing stromatolites which are of two types: small, closely packed columns each of 20 to 50 mm in diameter, growing vertically from a laminated algal mat; and domed forms, up to 300 mm across, which may have small columns growing from them. In this and other localities to the northwest and within the Teague Ring Structure, the limestone is partly silicified to a cream or pale-green chert and chert breccia.

Near Sweetwaters Well, a band of carbonate occurs which is interpreted to lie in the same stratigraphic position as those described above, being separated from the basal iron-formation of the Frere Formation by a thin (about 5 m thick) shale unit. Previously, this carbonate band has been included in the Frere Formation (Bunting and others, 1978; Hall

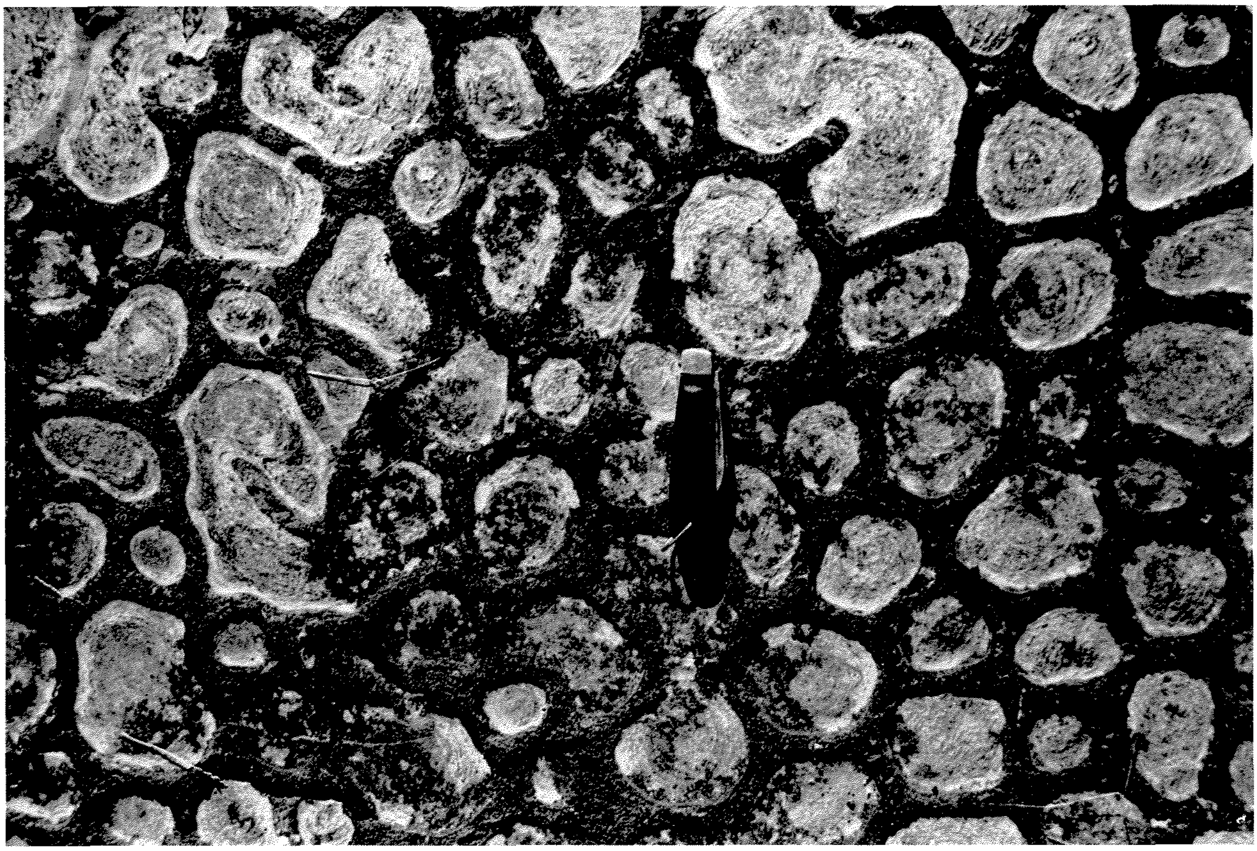

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Figure 8. Small columnar stromatolites (*Externia yilgarnia*), Yelma Formation, northeast DUKETON.

TABLE 3. PARTIAL CHEMICAL ANALYSIS OF CALCITIC DOLOMITE, YELMA FORMATION, SWEETWATERS WELL.

	Element (%)	*Oxide (%)*	*Carbonate (%)*
Ca	*19.8*	*27.7*	*49.5*
Mg	*11.4*	*18.9*	*39.5*
Fe	*1.31*	*1.87* (Fe_2O_3)	
CO_2	41.2		
	ppm		
Cu	*70*		
Pb	*460*		
Zn	*65*		

Sample: GSWA 25239
Location: 3 km southeast of Sweetwaters Well.
Analysis by Government Chemical Laboratories, Perth.
Sample collected and submitted by A. S. Harley (1970).

and Goode, 1978). The band is about 10 m thick, but the base is concealed. The carbonate is massive in the lower part and finely laminated in the upper part. Analysis of typical carbonate (Table 3) shows that the rock is a calcitic dolomite (18.9% MgO). Stromatolites are abundant, and consist of domed and stratiform types together with *Murgurra nabberuensis*, *Pilbaria perplexa* and *Yelma digitata*. The stromatolites appear to occur in a series of regressive cycles (see Appendix 1). Galena is a common accessory mineral in the stromatolitic dolomite (see Chapter 6, Economic Geology).

Northern margin: The Yelma Formation is much thicker along the northern margin of the sub-basin. Over the Marymia Dome, the thickness varies from a few tens of metres to several hundred. The basal unconformity is well exposed in northwest NABBERU, where the basal unit is cleaved shale (Fig. 9A). This is overlain by shale and interbedded sandstone which is variously feldspathic, glauconitic, and magnetitic. West of the Carnarvon Range the formation is up to 500 m thick, and consists of conglomerate, quartz arenite, phyllite, and minor arkose. The conglomerate is concentrated near the base and contains subrounded quartz pebbles in an arkosic matrix.

From the Carnarvon Range eastwards to the Malmac Dome the formation is about 1 500 m thick, and contains two sandstone members, each overlain by cleaved shale with rare, thin carbonate bands. The base is arbitrarily taken at the base of the lower quartzite member for reasons which were discussed in the section on the Troy Creek beds.

A reference section in the vicinity of Mount Evelyn is as follows:

* * * *

Member		Lithology	Thickness approx. (m)	Cumulative thickness (m)
FRERE FORMATION; granular iron formation, contact with Yelma Formation conformable				
YELMA FORMATION				
Upper shale	1.	Cleaved micaceous shale and phyllite with rare, thin (ca. 100 mm) carbonate bands.	400	1150-1550
Upper quartzite	7.	Medium-grained quartz arenite, 5% feldspar; cross-bedded.	100	
		Fine-grained, micaceous, feldspathic, quartz arenite, coarsening upwards.	30	
	6.	Very coarse sandstone to granule conglomerate with rounded clasts of quartz, quartzite, chert, and rare feldspar.	5	
	5.	Grey, medium-grained quartz arenite coarsening upwards, with thin, very fine-grained, feldspathic, micaceous sandstone beds; trough cross-beds.	100	

Member	Lithology	Thickness approx. (m)	Cumulative thickness (m)
Upper quartzite	4. Alternating 1 m thick beds of fine-grained quartz arenite and feldspathic sandstone; faint, low-angle cross-beds.	70	
	3. Very mature fine-grained quartz arenite; low-angle cross-beds.	5	
	2. Fine-grained feldspathic, micaceous sandstone and shale.	70	
	1. Fine- to medium-grained, well-sorted quartz arenite interbedded with feldspathic, micaceous shale; thinly bedded (50-150 mm) with small cross-beds.	120	650-1150
Lower shale	1. Cleaved micaceous shale, phyllite, fine-grained, micaceous, feldspathic sandstone, and rare, thin carbonate bands.	450	200-650
Lower quartzite	2. Medium-grained quartz arenite; cross-bedded, occasionally rich in magnetite.	100	
	1. Fine-grained, micaceous, feldspathic sandstone; rare grading and cross-beds.	100	0-200

TROY CREEK BEDS; cleaved shales; contact with overlying Yelma Formation not exposed, possibly unconformable.

* * * *

In thin section, the quartz arenites display a weak to strong tectonic fabric, defined by flattening of quartz grains, granulation along grain boundaries, and strain extinction. Only rarely is the outline of the original rounded detrital grain preserved with its authigenic envelope. Minor and accessory detrital minerals include feldspar, chert, tourmaline, zircon, and muscovite. A lamination (10-20 mm thick) within beds displays trough-style cross-stratification. However, the tectonic fabric has destroyed much of the primary lamination, and in the thicker arenite units even the bedding is difficult to determine.

The Yelma Formation is much thinner over the Malmac Dome, due largely to the wedging out of the arenaceous members. A thickness of approximately 500 m was measured in a section located on the southern side of the dome, 10 km southwest of Malmac Well. The section recorded below is approximately that described by Horwitz (1976) and designated the type section of the Malmac Formation. Horwitz's Malmac Formation is here included in the Yelma Formation.

FRERE FORMATION; ferruginous chert, wavy-bedded peloidal; contact with Yelma Formation conformable

YELMA FORMATION

Lithology	Thickness approx. (m)	Cumulative thickness (m)
9. Cleaved shale, buff to pale-purple, minor fine-grained feldspathic sandstone. Possibly some thin carbonate beds.	92	408-500
8. Fine-grained mature quartz arenite, cross-bedded.	8	400-408
7. Poorly exposed cleaved shale.	19	381-400
6. Breccia, containing angular intraclasts up to 200 mm long of chert and silicified shale, in a ferruginous, medium-grained, clayey sandstone matrix.	1	380-381
5. Cleaved shale.	5	375-380
4. Medium-grained ferruginous sandstone.	5	370-375
3. Poorly exposed, cleaved shale with rare, thin, fine-grained sandstone beds.	160	210-370
2. Interbedded fine- to medium-grained feldspathic quartz arenite and feldspathic shale, in beds 100-300 mm thick. Symmetrical ripples.	10	200-210
1. Cleaved cream to purple micaceous shale, with minor fine-grained feldspathic sandstone in lower part. At base, intermittent lenses up to 3 m thick of arkose and rare quartz-pebble conglomerate. Thin carbonate beds about 100 m from base.	200	0-200

ARCHAEAN GRANITOIDS (Malmac Dome); contact with overlying Yelma Formation unconformable

* * * *

The Yelma Formation in this section dips south at between 20° and 40°, averaging about 25°. The dip steepens in the overlying Frere Formation to the south. On the north side of the Malmac Dome, where the Earaheedy Group is very poorly exposed and gently dipping, the Yelma Formation must either be very thin, in order to fit between the granitoid rocks and the nearest iron-formation, or it has been faulted out. The unconformity between Yelma Formation and granitoid is exposed 4 km southwest of Malmac Well, where a few metres of lateritized shale and sandstone, with a 100 mm basal conglomerate band, are exposed.

Western margin: The western limits of Yelma Formation exposure are difficult to define, due to poor exposure in the narrow neck between the Earaheedy and Glengarry Sub-basins, and to similarities between the basal units of the two sub-basins. These similarities are such that Hall and Goode (1978) suggest that the basal formations are equivalent, a situation which necessitates a rapid wedging out of the thick trough sequence in the Glengarry Sub-basin as the Earaheedy Sub-basin is approached. However, evidence presented here and at the end of this chapter indicates that the Yelma Formation is unconformable on the Glengarry Sub-basin sequence.

An important outcrop occurs 10 km south-southeast of Baumgarten and consists of very coarse, poorly sorted conglomerate of at least 100 m thickness. The conglomerate is here termed the Yadgymurrin Conglomerate Member of the Yelma Formation, for which a type area is designated at Lat. 25°21′30″S, Long. 119°59′30″E. A strong cleavage and closely spaced jointing dip steeply to the west and give the appearance of bedding; however, sandstone lenses within the conglomerate indicate an easterly dip of 5° to 20° (Fig. 9B), reaching 40° on the eastern side of the outcrop. Clasts in the conglomerate range in size up to 1 m across, with the majority between 50 mm and 200 mm. The clasts are well rounded, and include quartzite, pink chert and chert breccia. Some of the chert and chert-breccia clasts are similar to rocks occurring as isolated lenses and "pipes" in the Karalundi Formation of the Glengarry Group. The quartzite clasts are consistent with a derivation from the basal quartzite of the Glengarry Group, which is exposed some 15 km to the west. No clasts were found which could definitely be attributed to a derivation from the resistant iron-formations of the Frere Formation. Thus the nature of the clasts and the easterly dip together indicate that the conglomerate is part of the Yelma Formation at the base of the Earaheedy Group. The well-rounded and indurated nature of the clasts suggests that an unconformity is present.

Along strike to the north-northeast, the clasts disappear and the Yelma Formation is represented by a few isolated outcrops of ferruginous quartz arenite. Similar conglomerate also occurs 37 km due south of the type area, 5 km east of Combine Well (Gee, 1983).

Mapping by Gee (1983) has established the extent of the Yelma Formation in eastern PEAK HILL, where it consists of scattered exposures of stromatolitic dolomite, chert breccia (silicified carbonate), silicified sandstone and quartz arenite. These rocks are invariably shallow dipping, and unconformably overlie the more steeply dipping shale and greywacke of the Glengarry Group. Similar rocks, including stromatolitic carbonate, which occur in the Kimberley Range area on GLENGARRY, may also correlate with the Yelma Formation.

DEPOSITIONAL ENVIRONMENT

The Yelma Formation marks an extensive marine transgression which probably began in the area underlain by the Troy Creek beds, where the thickest Yelma Formation occurs. The transgression then progressed westwards over the Marymia Dome and proto-Wiluna Arch, and eastwards over the Malmac Dome. Immature arkosic sediment in the basal Yelma Formation over the two domes indicates rapid erosion in the source areas during the early stages of the transgression.

The extension of the sea over the Yilgarn Block must have been very rapid over a flat, mature landscape. Although extensive, the Yelma Formation is seldom more than 100 m thick along the southern margin of the basin. With few exceptions, the basal sandstone is mature and quartz rich. Sedimentary structures are characteristic of shallow-marine, shoreline conditions, possibly with some tidal influence. These structures include shallow, trough-style, cross-stratification and ripple marks, including symmetrical, asymmetrical, and linguoid varieties. The gradual fining upwards of the sequence as in the type section, is probably due to a decrease in clastic input and a slight increase in water depth. The presence of stromatolitic carbonate lenses in the upper part of the formation along the southern edge of the basin also indicates a shallow shelf-sea environment.

Evidence of non-marine sedimentation exists in the conglomerate around the western end of the Earaheedy Sub-basin. Clasts in the conglomerate are well rounded, suggesting transportation in a fluviatile environment. The poor sorting of the sandstone matrix, associated with a lack of well-defined bedding, suggests that the transporting mechanism may have been a debris flow. The thin lenses of quartz-pebble conglomerate, at the base of the Yelma Formation along the southern margin of the sub-basin may also represent fluvial deposits immediately prior to the marine transgression.

FRERE FORMATION

DEFINITION

Derivation of name: Named after the Frere Range in central NABBERU.

Type section: The type section is taken in the Frere Range, between Lat. 25°43′00″S, Long. 120°40′00″E and Lat. 25°40′00″S, Long. 120°42′00″E, where the middle part of the formation is well exposed (about 200 m). Nowhere is the entire formation exposed in a single section.

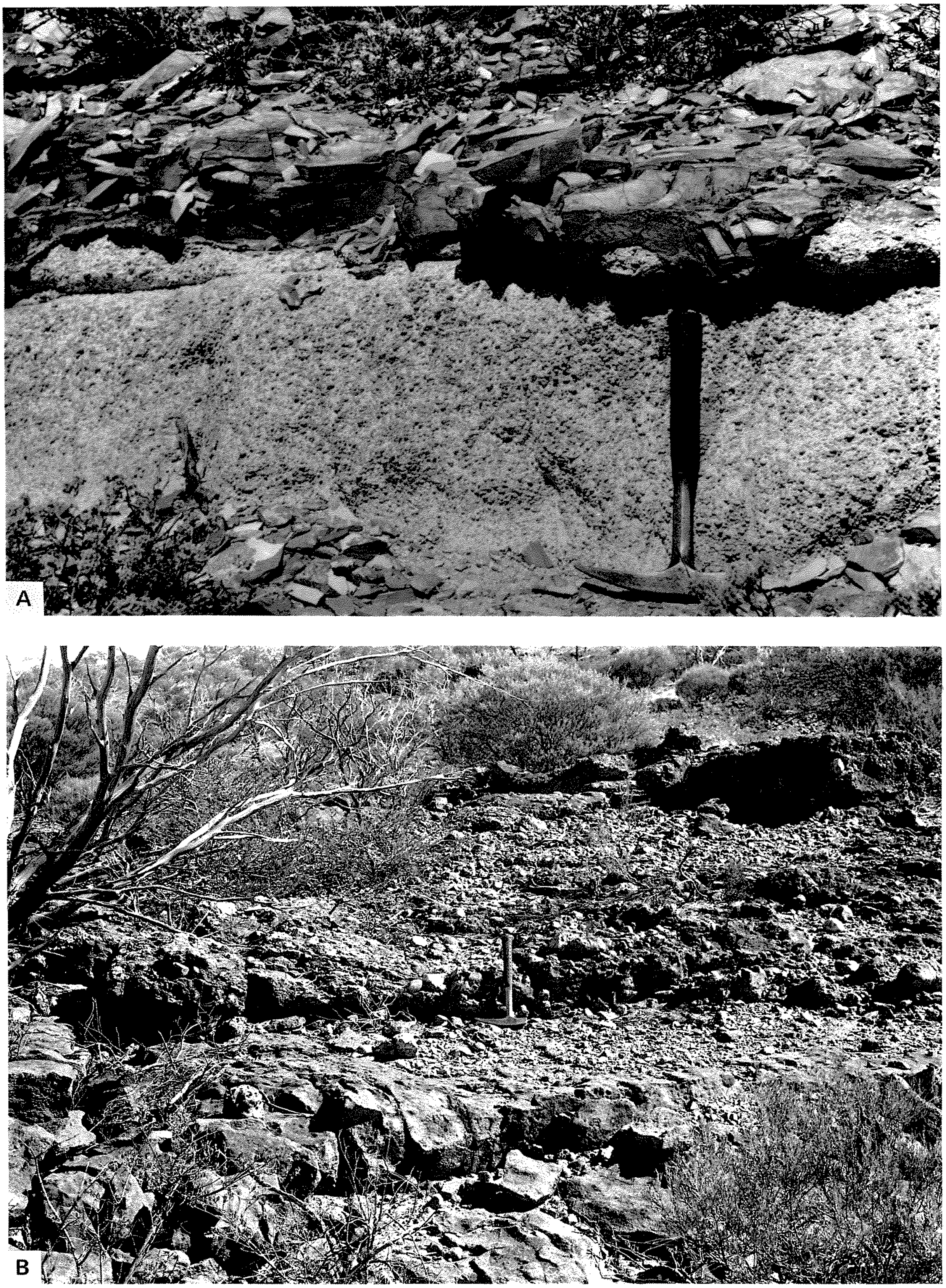

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Figure 9. A — Unconformity between coarse-grained granitoid of the Marymia Dome and cleaved siltstone of the Yelma Formation; 22 km southeast of Marymia homestead.

B — Yadgymurrin Conglomerate Member, Yelma Formation, showing shallow-dipping lenses of sandstone (foreground and top right) in boulder conglomerate. Type area, 10 km south-southeast of Baumgarten.

Lithology: Mainly buff, cream and purple shales interbedded with iron-formation and chert. Minor carbonate bands occur rarely. The iron-formation and chert are characteristically granular, although there is some microbanded iron-formation. The iron-formation:shale ratio increases steadily from southeast KINGSTON to northwest NABBERU.

Thickness and distribution: Shallow dips make accurate calculations difficult, but total thickness is probably in the order of 1 200 m, with little observable variation. The formation is exposed along both the northern and southern limbs of the Earaheedy Sub-basin and around the sub-basin closure in western NABBERU. The apparent surface width increases in an east-southeasterly direction on KINGSTON due to progressively shallower dips.

Boundary criteria: The Frere Formation conformably overlies the clastic rocks of the Yelma Formation. The base is taken at the base of the first continuous iron-formation/chert bed. This is a departure from the original definition by Hall and others (1977) who took the base at the top of the uppermost arenite of the Yelma Formation. The revision has been made because the basal iron-formation of the Frere Formation is a more prominent and continuous marker than the uppermost arenite of the Yelma Formation. Thus all clastic rocks and minor carbonates below the first continuous iron-formation/chert are placed in the Yelma Formation.

The top of the Frere Formation passes conformably into either the Windidda or Wandiwarra Formations, and is at the top of the uppermost iron-formation/chert bed. The contact with the Windidda Formation is taken at the base of a 5-10 m thick transition zone consisting of interbedded chert and carbonate.

LITHOLOGICAL DESCRIPTION

The Frere Formation typically consists of several units (informally termed "members") of iron-formation separated by shale. The iron-formation members range from less than 1 m to about 200 m thick (the latter named the Illagie Iron-Formation Member, by Hall and Goode, 1978). As a general rule they increase in thickness and abundance in a northwesterly direction; in other words, the ratio of iron-formation to shale is increasing to the northwest. The iron-formation members form resistant strike ridges and cuestas, whereas the intervening shales occupy valleys or broad, colluvium-covered flats. Thus it is difficult to calculate accurately the iron-formation:shale ratios, however figures of 1:50 for southeast KINGSTON, 1:10 for northeast WILUNA, 2:1 for the eastern Frere Range and 5:1 for the western Frere Range are not unreasonable. Along the northern edge of the basin, the ratio is more consistent at about 1:1, decreasing only slightly to the east.

Superimposed on the increasing iron-formation:shale ratio is an increase in the total iron content of both iron-formation and shale. In eastern KINGSTON, most iron-formation members are represented by ferruginous chert containing between 1 and 10% iron. In northeast WILUNA, it is estimated that most iron-formations contain between 10 and 20% iron, while in western and northern NABBERU they contain over 20% iron; secondary enrichment in many cases brings the iron content to over 50%.

Broadly speaking, the rocks of the Frere Formation can be subdivided into five lithological types. These are: shale; granular iron-formation; banded and shaly iron-formation; chert; and carbonate rocks. Many rocks are transitional within this broad grouping, and boundaries between the types are often gradational. In view of the importance of the iron-formations, these are discussed in more detail in Chapter 4, which includes a section on nomenclature.

Shale: The shale units of the Frere Formation contain various fine-grained, terrigenous clastic rocks, including siltstone, sandy siltstone, fissile shale and massive mudstone. These rock types are usually interbedded at outcrop scale. In the Stanley Fold Belt, deformation has produced cleaved shale, slate and phyllite.

In the southeast part of the basin, where the shale units form by far the greatest part of the Frere Formation (Fig. 10), the predominant rock type is a cream, buff, red, brown or purple siltstone with variable degrees of fissility. The shale is typically well laminated, on a scale of 1 to 10 mm, and commonly displays small-scale cross-lamination. The cross-lamination occurs either as single sets of ripples, or as multiple sets of climbing ripples (Fig. 11). Small scour structures are common, and both scours and single ripples show load deformation where silt layers have sunk into softer clay layers. Some sandy siltstone beds show poorly developed graded bedding. Thin beds and lenses of cryptocrystalline, greenish-grey chert occur within the shale.

The shale and siltstone in the southeast of the basin consist predominantly of quartz, clay minerals and white mica, with minor amounts of feldspar, biotite and chlorite. To the northwest, the rocks become more ferruginous, and fissility is more pronounced.

Granular iron-formations: Although shale is more abundant, granular iron-formation (GIF) is the most characteristic rock type of the Frere Formation (Fig. 12). Individual GIF beds attain a thickness of 1 m, but are generally between 50 and 300 mm thick. Beds

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Figure 10. Laminated shale of the Frere Formation, northeast DUKETON. Note small fault above the man left of centre. Hill is about 30 m high. Top 3 m is Permian fluvioglacial sandstone.

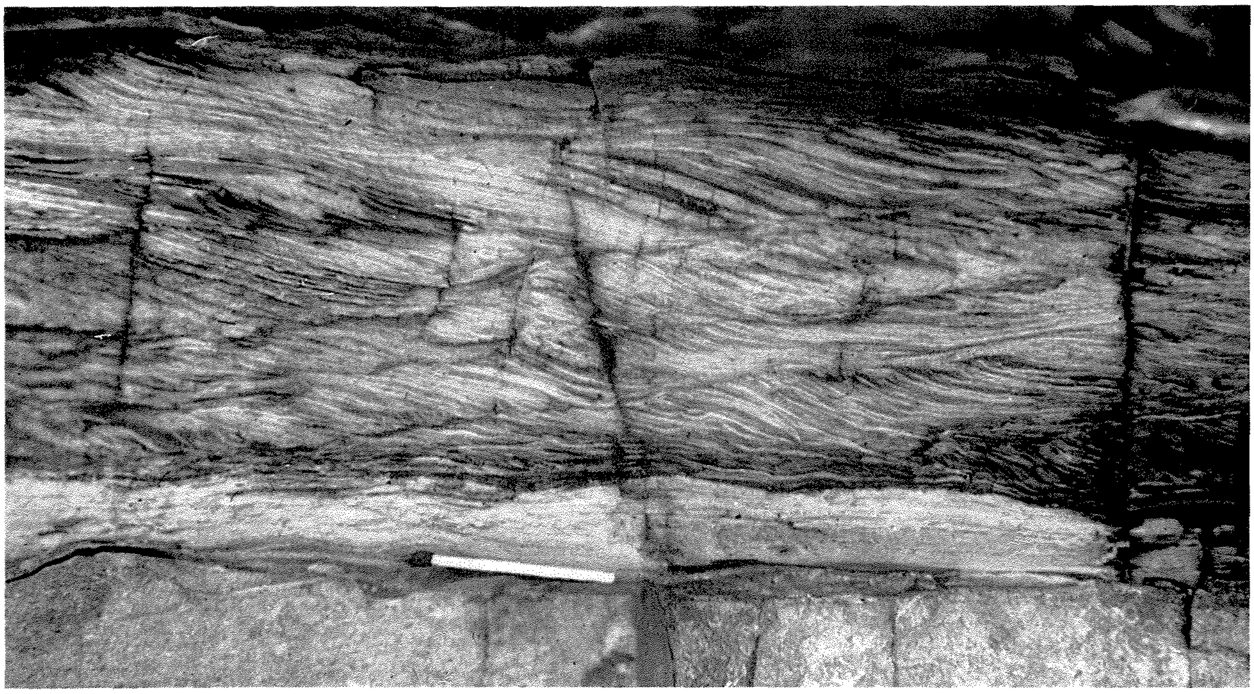

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Figure 11. Climbing-ripple cross-lamination, near top of Frere Formation, southeast NABBERU.

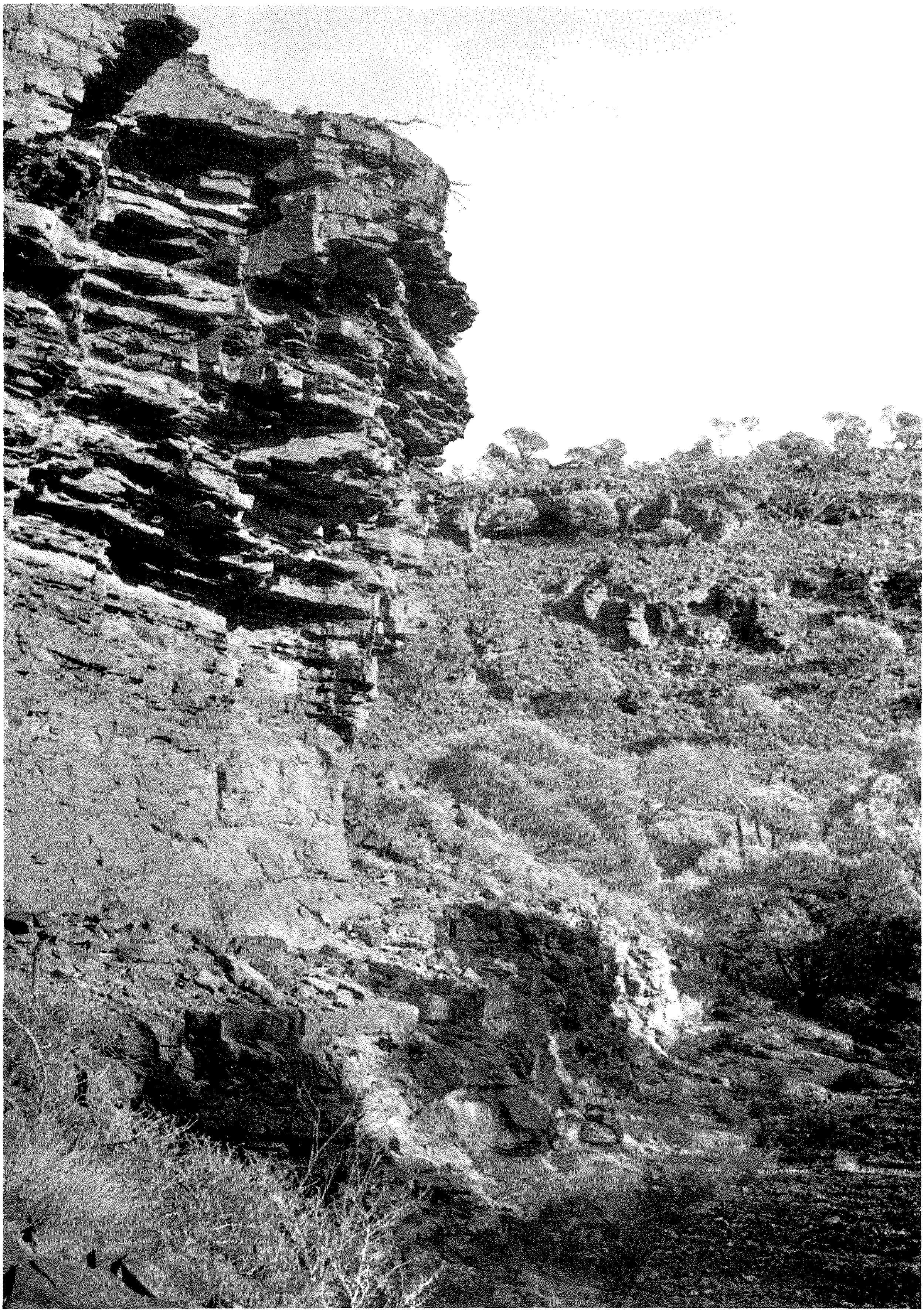


Figure 12. Granular iron-formation with thin shale interbeds. Frere Formation type section, 1 km southwest of Mount Teague.

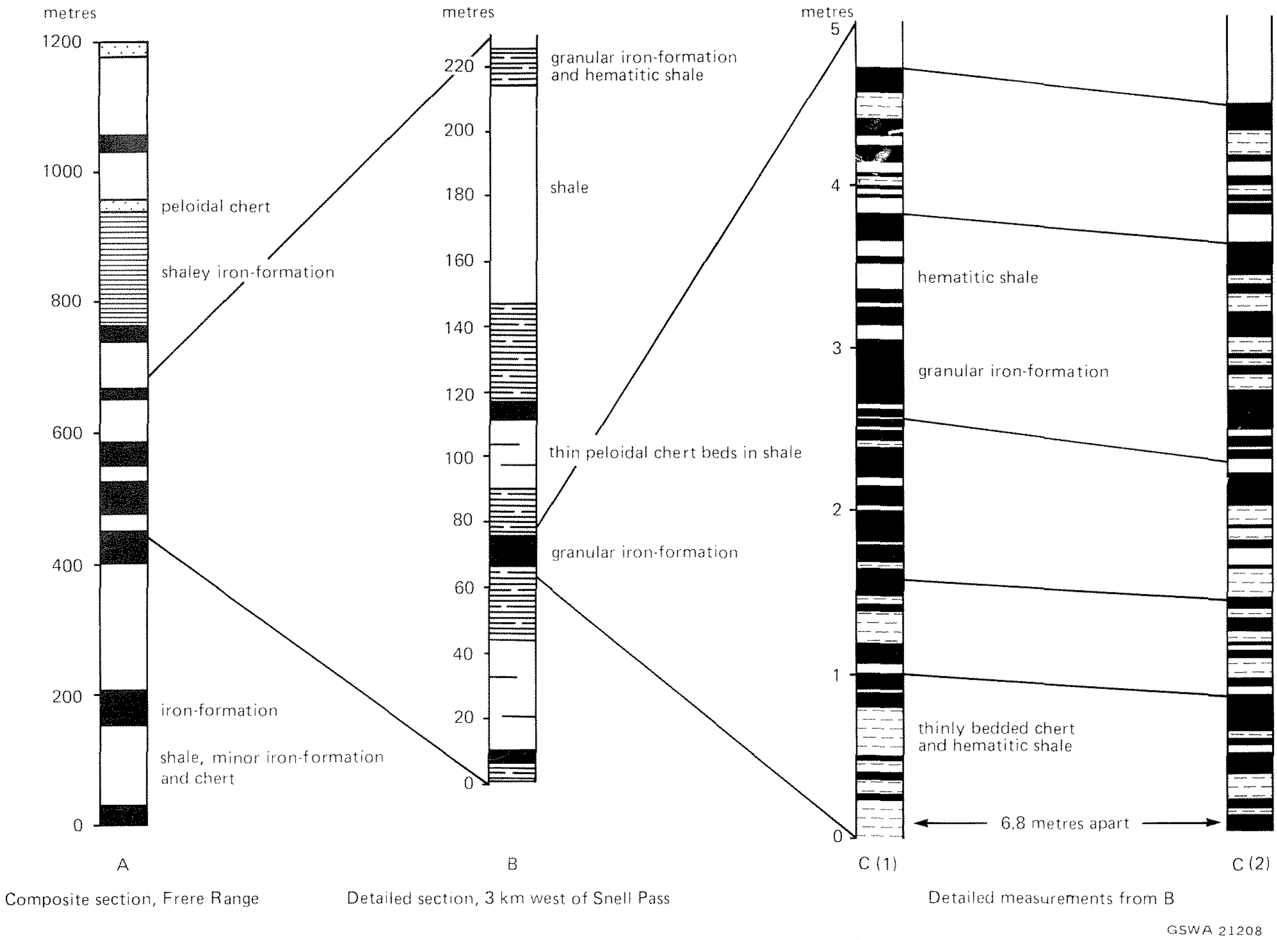


Figure 13. Composite and measured sections, Frere Formation.

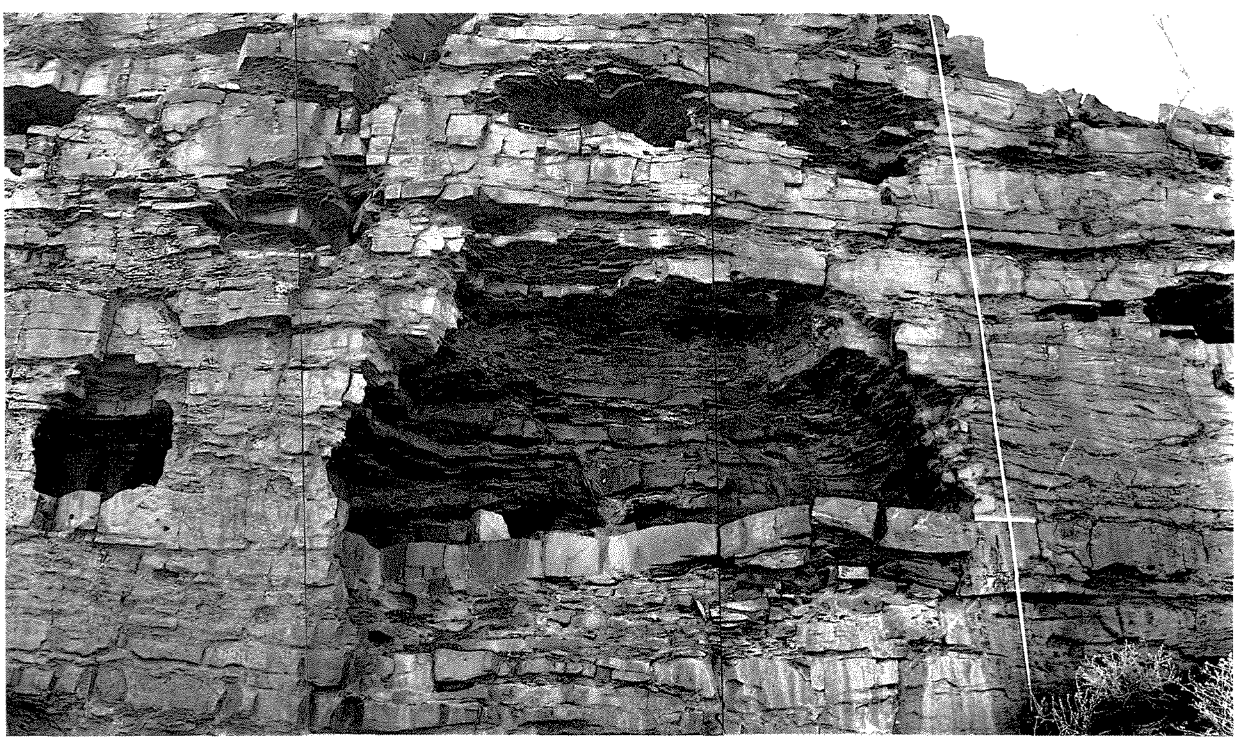


Figure 14. Site of detailed measured section C(2) from Figure 13. Section C(1) is just to the left of the field of view. Frere Formation, 3 km west of Snell Pass. Note resistant GIF beds and weaker shaly partings. Ruler on tape is 85 cm from base of section. The excavation is old, and probably man-made.

are typically wavy and lenticular, and seldom persist laterally for more than a few tens of metres. This is shown in Figures 13 and 14, where two detailed measured sections, less than 7 m apart, show little correlation of GIF beds. GIF beds are separated by hematitic shale which may be anything from a few millimetres to several metres thick.

It is clear, therefore, that at all scales greater than those of individual beds, shale is a feature of the iron-formation units. In Figure 13, iron-formation is shown at three different scales. At a regional scale (section A) iron-formation is concentrated into distinct units, which are informally termed "members", each of which in turn contains a significant amount of shale, much of it hematitic. At the scale of a measured traverse (section B) the iron-formation members are seen to contain variable amounts of shale, and thus thinner units of predominantly GIF can be distinguished. Some parts of a member may

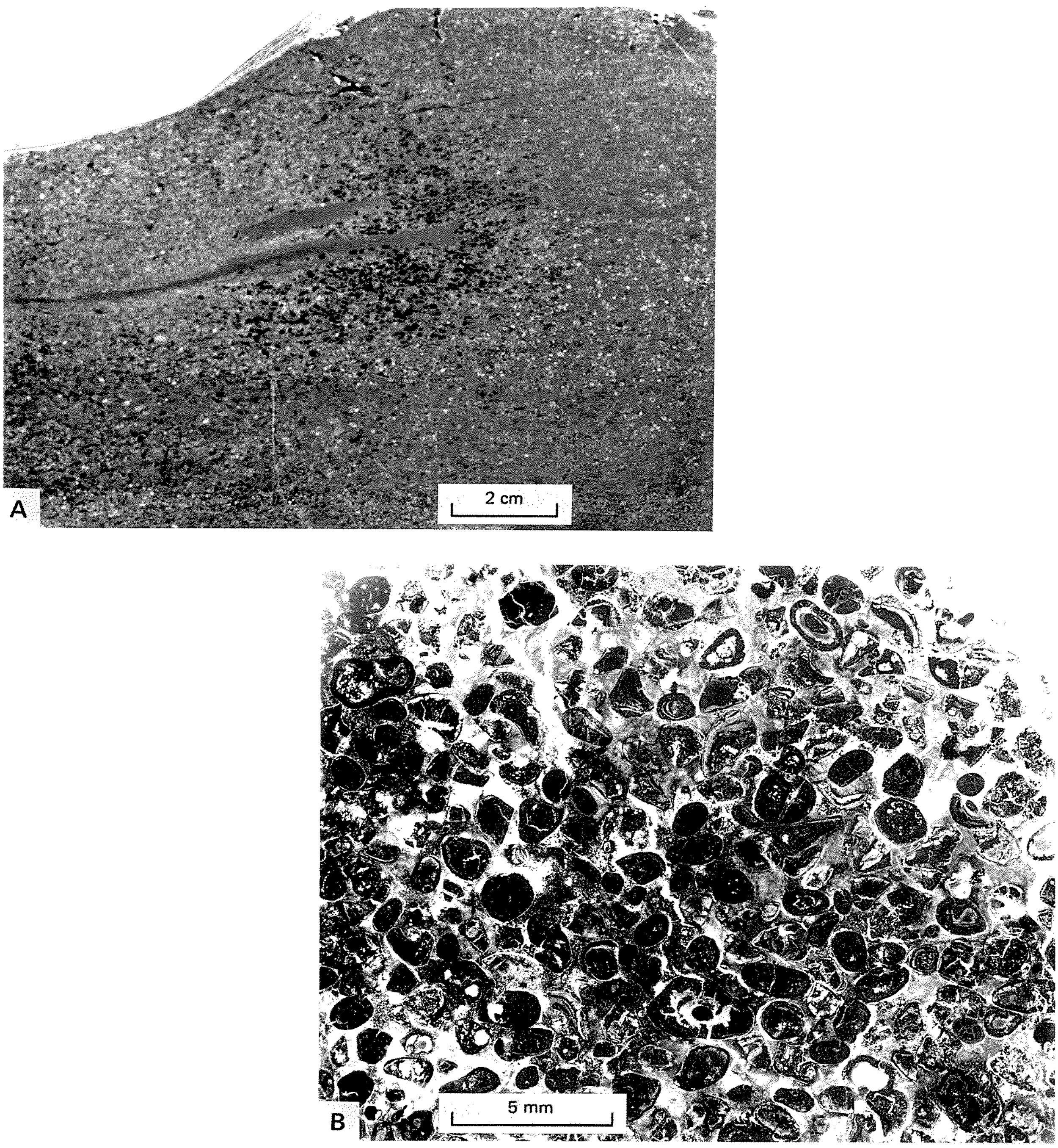


Figure 15. A — A single granular iron-formation (GIF) bed, containing both peloids and ooliths. Note that load deformation has created compaction on left side of bed and tension cracks on top surface. Frere Formation, 6 km east-northeast of Yelma outcamp. Polished slab.

B — Detail of A, showing typical peloids and ooliths. GSWA thin section 46401.

contain only 10% GIF; others can have up to 95% with the remainder being hematitic shale or shaly iron-formation. Thin, isolated beds of GIF or granular chert occur in the regional-scale shale members.

The term granular iron-formation is used to describe those rocks, containing chert (usually recrystallized) and an iron-bearing mineral, that have a granular texture to the unaided eye. No firm limit, in terms of iron content, is imposed, and thus the ferruginous cherts interbedded with the iron-rich GIF are included in the following discussion (see Chapter 4 for discussion of terminology).

Typically the GIF is purple, grey or red, occasionally greenish. The preponderance of chert (usually over 50%) produces a hard, flinty rock. The iron mineralogy is dominated by the oxides (hematite and magnetite), whilst iron phyllosilicate, siderite and pyrite are extremely rare. Jasper, consisting of hematite dust in chert, is responsible for the red colour of much of the GIF. Hematite also occurs as specular hematite and as a replacement (martite) of magnetite. Magnetite and martite are restricted to the more deformed and recrystallized rocks of the Stanley Fold Belt.

The granular texture is composed of particles or aggregates of particles (allochems) set in a cement and/or matrix of chert (orthochem). The allochem fraction is usually peloidal (Fig. 15), but may be

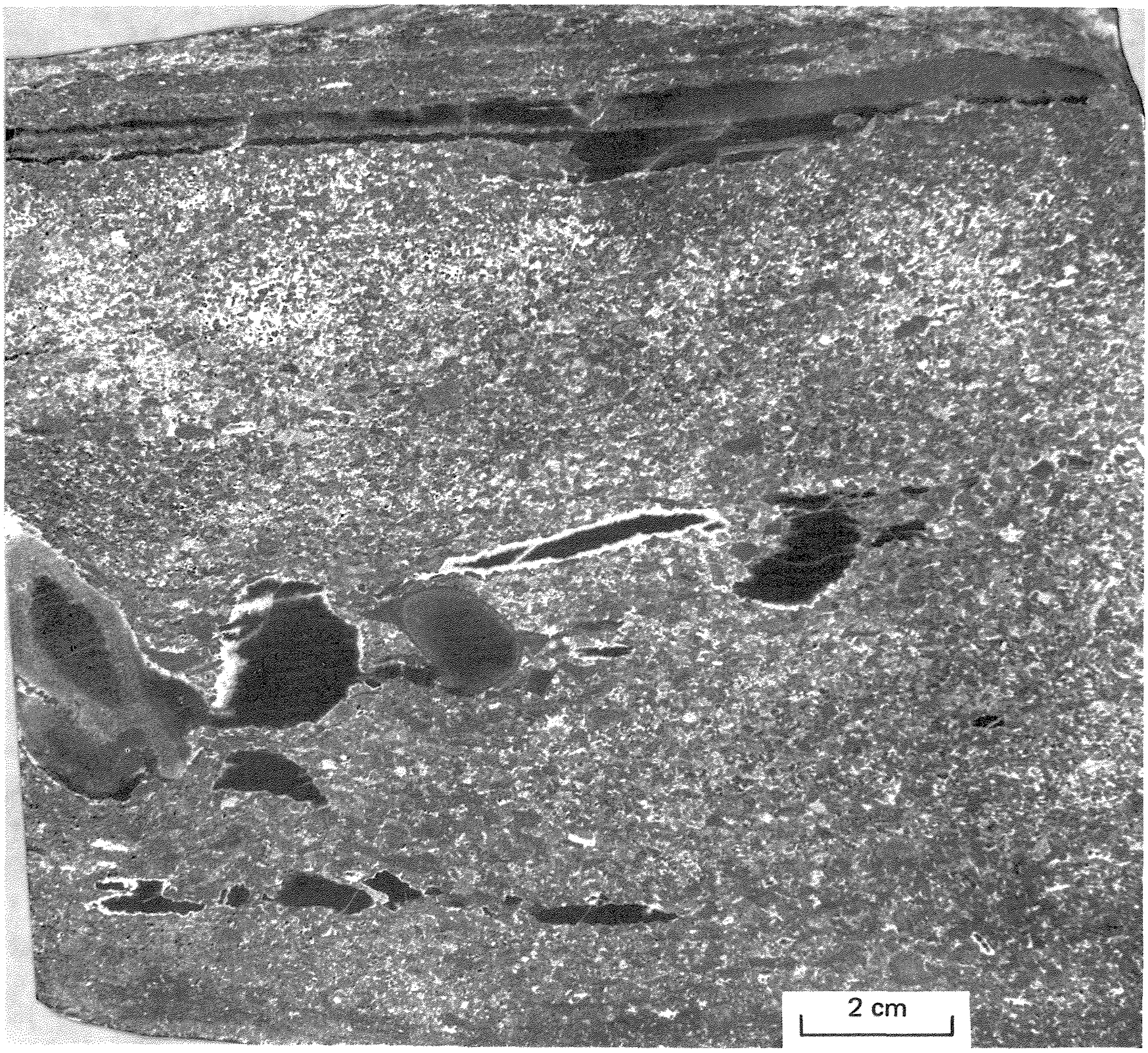
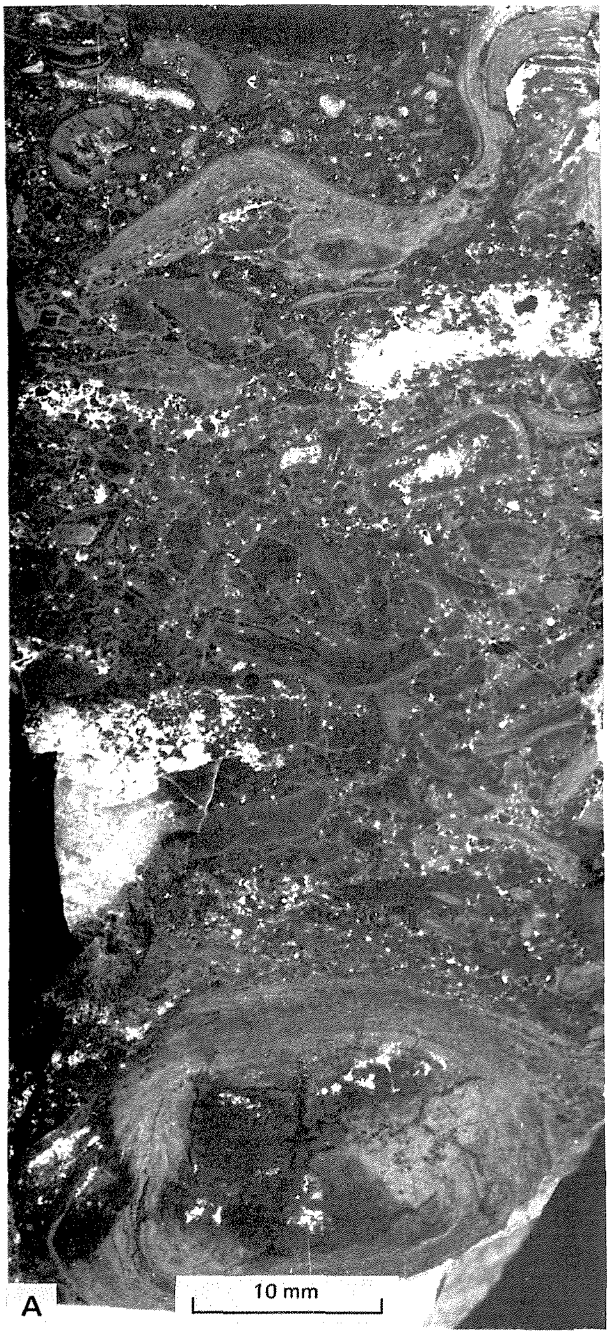


Figure 16. Single GIF bed, showing large intraclasts of ferruginous chert set in a mixture of rounded peloids and intraclastic microbreccia. Note the deformation of some of the larger intraclasts (remobilization of chert due to loading?) and the continuous layer of chert near the top of the bed. Frere Formation, 3 km west of Snell Pass. Polished slab.



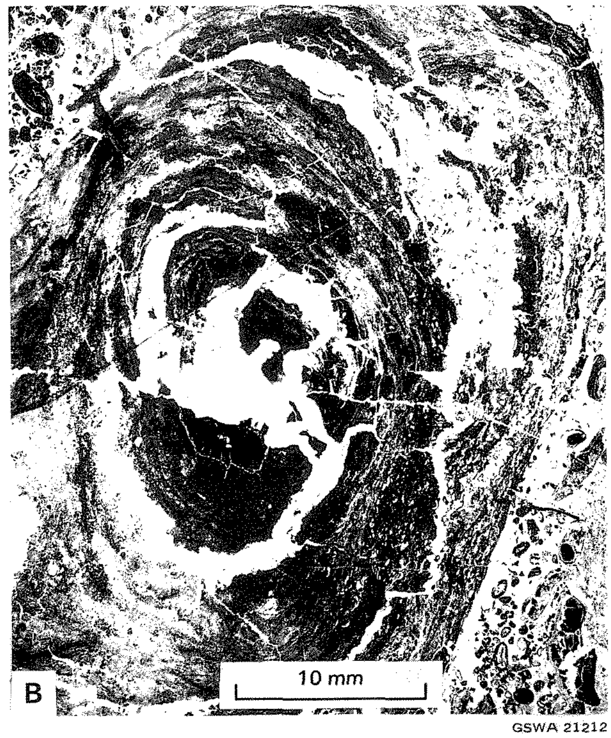


Figure 17. A — Oncolites and stromatolitic breccia. Frere Formation, Camel Well, northeast WILUNA.

B — Oncolite, showing greater detail of internal structure. GSWA thin section 46081.

oolitic, oncolitic, or an intraclastic breccia. The peloids are rounded, subspherical grains, consisting of either jasper ± specular hematite ± chert, or, in the more recrystallized areas, granular quartz and martite. The ooliths contain concentric layers of jasper and chert, in part altering to specular hematite. They show a complex history of fragmentation and regrowth. Well-formed ooliths are restricted to the southern margin of the basin. On NABBERU and STANLEY, some recrystallized peloids show a vague concentric layering. Both peloids and ooliths range from 1 to 2 mm in diameter. Radial shrinkage cracks are common, and are usually filled with quartz.

Breccia fragments include both iron-rich and cherty varieties; they are usually angular, and form tabular fragments several centimetres long, although smaller ones may be rounded and grade into peloids. Most peloidal iron-formation beds contain some breccia fragments, and may contain a 10 mm thick layer of cryptocrystalline chert near the top of the bed which is in part brecciated (Fig. 16). Some beds consist almost entirely of breccia fragments (in the 2 to 20 mm size range) and could be termed ferruginous intraclastic microbreccia.

Load deformation has produced flattened peloids and breccia fragments. In extreme cases, peloids form accommodation shards (Dimroth and Chauvel, 1973) in the vicinity of coarse fragments.

An unusual bed of oolitic and pisolitic (pisoliths up to 10 mm diameter) iron-formation, 3 km northeast of Camel Well on WILUNA, contains stratiform stromatolites, oncolites (unattached stromatolites) and intraclastic breccia derived from them. The stromatolites are not well formed, and have a thin, wavy lamination. The oncolites range from 10 to 50 mm and have a lamination which is concentric but irregular (Fig. 17). These structures contain a complex assemblage of microfossils (Walter and others, 1976; Hofmann and Schopf, 1983) which closely resemble microfossils from the Gunflint Iron Formation of Canada.

Surface weathering of the iron-formation takes the form of oxidation and iron enrichment in the western and northern part of the basin. The product is a ferruginous duricrust in which primary textures and structures have been partly destroyed, and the mineralogy now consists of hematite, hydrated iron oxides (goethite, limonite) and secondary quartz or chalcedony. In the northeast and southeast parts of the basin, some ferruginous granular cherts have a honeycomb texture due to leaching of the peloids.

Banded and shaly iron-formation: These rocks form a small but very varied percentage of the Frere Formation, and therefore have to be treated as a number of separate occurrences rather than as a unified group. Banded iron-formation (BIF) is virtually restricted to the northern side of the basin, although some thin, microbanded, jasperoidal chert beds occur in the hematitic shale portions of the granular iron-formation on the southern side.

The most conspicuous BIF occurs at the top of the formation, and is best exposed at two places—west-southwest of Mount Royal (Fig. 18) and south of Malmac Dome. The BIF is about 50 m thick, and contains alternating chert and iron oxide mesobands between 20 and 200 mm thick (average about 50 mm near Mount Royal, and 100 mm south of the Malmac Dome). Chert mesobands are white or pale pink, and show a faint 2 to 3 mm microbanding. Iron oxide mesobands display microbanding 0.5 to 2 mm thick. At the surface, the black iron oxide mesobands consist of hematite and goethite, and this secondary alteration obscures the microbanding in many instances.

BIF which crops out immediately east of Mount Royal is probably close to the base of the formation. This BIF contains microbanded hematite (mostly martite) mesobands, and peloidal-textured chert mesobands. Both are highly distorted by wavy bedding, tectonic folding and crenulation (see Chapter 4).

Shaly iron-formation along the southern side of the basin occurs both as thin beds interlayered with granular iron-formation, and as a thick member with no granular iron-formation near the top of the formation. In both cases the main iron mineral at the surface is hematite, although textural evidence in the interlayered type suggests it may, in part, be replacing pyrite or magnetite.

Bedding in the interlayered type is variable in scale, from less than 1 mm to 20 mm. Some red jasperoidal chert beds contain faint microbanding about 0.1 mm thick. The bedding is laterally continuous, except for thin lenses and scour-fills of intraclastic and peloidal iron-formation (Fig. 19). The intraclastic material consists of fragments of jasperoidal chert and shaly, hematitic iron-formation similar to the undisturbed beds. These granular beds are seldom more than a few centimetres thick.

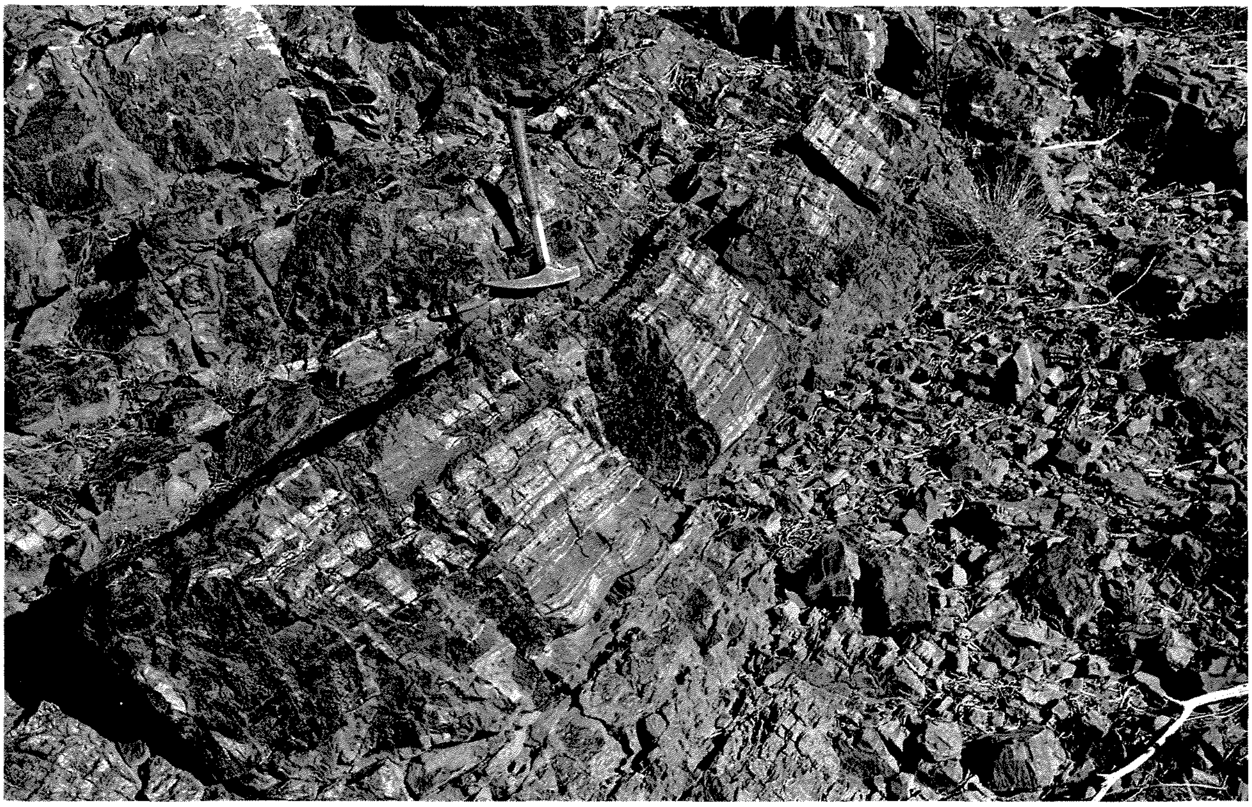


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Figure 18. Banded iron-formation. Near top of Frere Formation, west-southwest of Mount Royal.

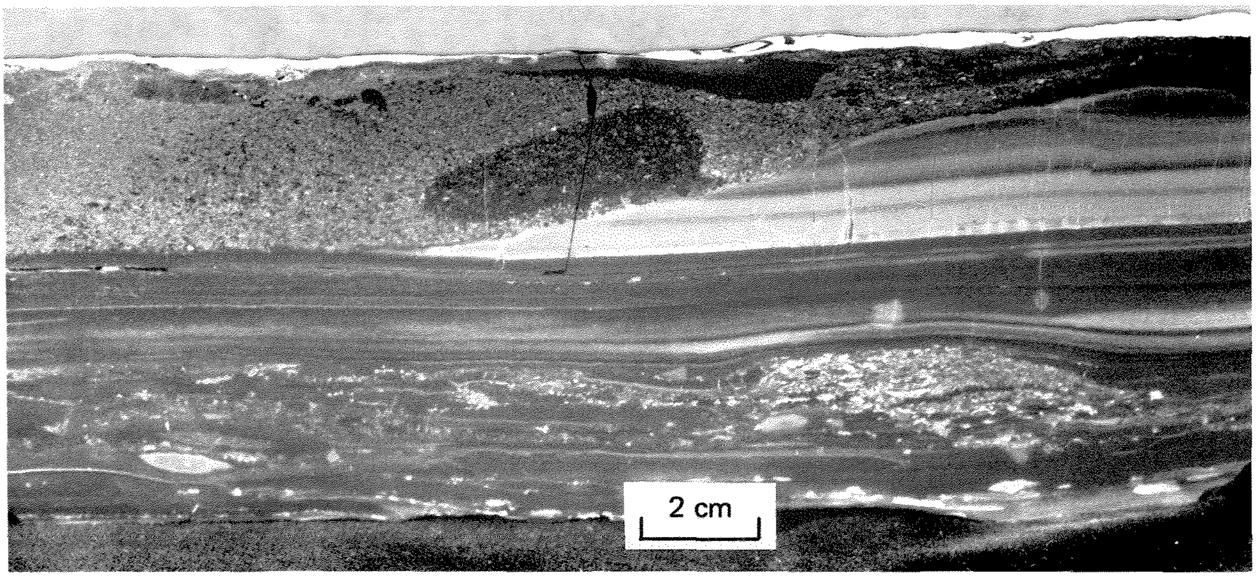

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Figure 19. Scour structure, filled with intraclastic microbreccia (the *large dark oval shape is an imperfection caused during* slab preparation), cutting into jasperoidal chert (pale, banded) but not affecting shaly iron-formation (dark-grey, banded). Bottom layer contains intraclasts of shaly iron-formation and ferruginous chert, and a small lens of intraclastic microbreccia. Near top of Frere Formation, Frere Range. Polished slab.

The shaly iron-formation member near the top of the formation has been named the Illagie Iron-Formation Member by Hall and Goode (1978). It is 180 m thick in its type area (at Lat. 25° 32′ 50″S, Long. 120° 22′ 50″E, near Murgurra Pool) and has been recognized for approximately 15 km to the east and west of this point. It is not clear whether the shaly iron-formation pinches out laterally, or passes by a facies change into the more typical GIF/shale association. The principal rock type is a dark-grey or purple, shaly iron-formation with bedding on a scale of 10 mm or less. Bedding is defined by variations in iron and clay content. A fine lamination or microbanding is present, and is laterally continuous over several metres. The scour structures of the interlayered type are absent, suggesting a quieter, deeper environment, however some unusual sedimentary structures are present. These include "ripples" on bedding surfaces (possibly due to differential loading), microslumps, and flattened nodules, about 10 to 20 mm across (possibly after manganese). Small "bubbles" on lamination surfaces are due to thin films of clay draping over small (less than 0.5 mm) detrital grains of jasperoidal chert.

Chert: Non-ferruginous chert is of two types—massive and peloidal. Massive chert is usually cryptocrystalline and shows no internal fabric or structure. It is pale green, cream or pale grey, and occurs mainly as thin (<20 mm) bands in granular iron-formation. The bands are commonly broken, and the fragments contribute to the intraclastic component of the iron-formation. These bands are often concentrated near the top or near the base of individual GIF beds.

Grey-green, cryptocrystalline chert forms rare beds and lenses, up to 100 mm thick, in non-ferruginous siltstone in the southeastern part of the basin (KINGSTON). Their presence is useful in distinguishing between flat-lying siltstone of the Frere Formation and the lacustrine facies of the Permian Paterson Formation.

Non-ferruginous peloidal chert in eastern KINGSTON and ROBERT forms beds 1 to 2 m thick, separated by great thicknesses of shale and siltstone. The chert layers form low, resistant cuestas, and are equivalent to the ferruginous chert and iron-formation of the rest of the Frere Formation. Because of large-scale pinching out of these units, they are seldom continuous for more than a few kilometres. Textures are similar to those in the peloidal iron-formation. Round or elliptical peloids up to 2 mm across consist of cryptocrystalline chert, and are set in a matrix/cement of similar material or chalcedonic silica. Consequently boundaries between peloid and matrix/cement are usually diffuse. Oolitic chert occurs in parts of ROBERT.

A distinctive green peloidal chert forms an important marker unit at the top of the Frere Formation on KINGSTON, and is here called the Wellesley Chert Member (after Mount Wellesley in northwest KINGSTON). A type area is taken at the western end of the Wellington Range (Lat. 26°17′20″S, Long. 121°40′20″E). The member extends intermittently from a point 3 km southeast of Lorna Glen homestead to near Tooloo Bluff, a distance of about 75 km (Fig. 20). It is about 50 m thick, and in the top few metres it is interbedded with grey-to-pink limestone and dolomite, in a zone which forms a tran-

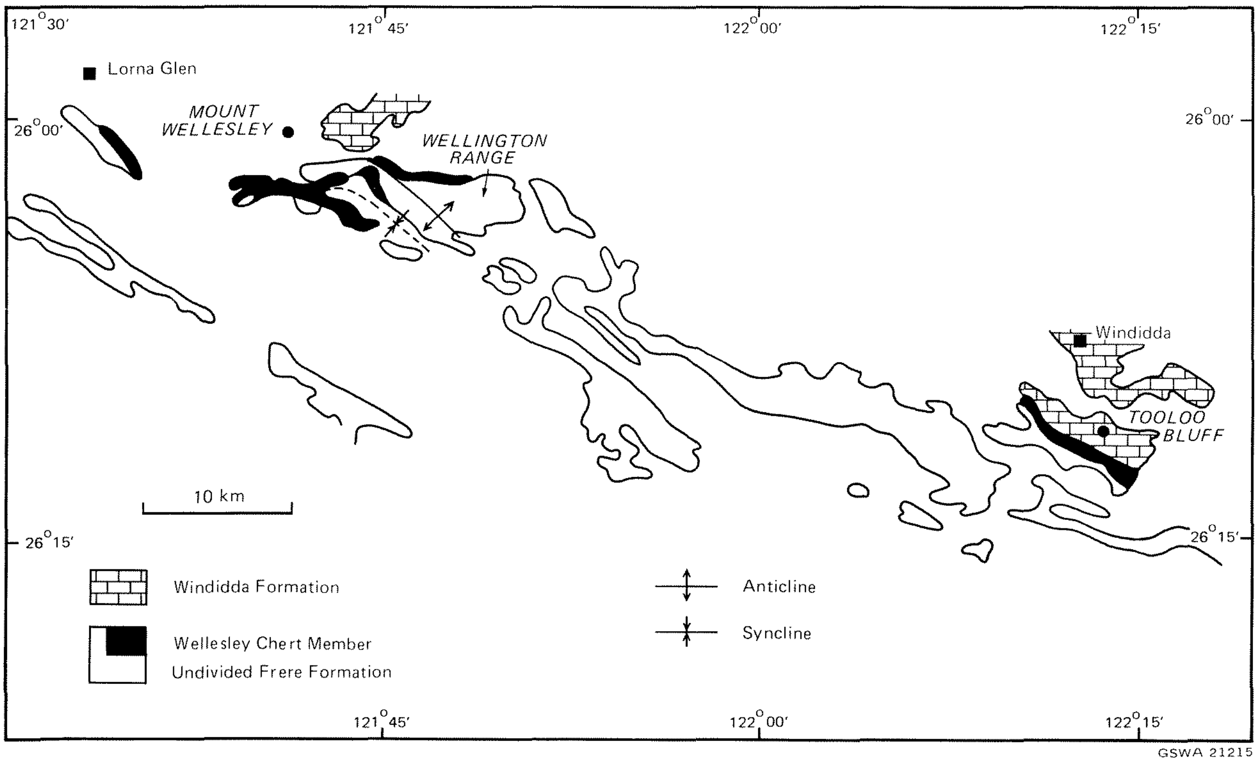

Figure 20. Sketch map of Wellesley Chert Member, Frere Formation.

sition to, and is arbitrarily included within, the overlying Windidda Formation. The member is underlain by a few metres of shale and siltstone which in turn overlie granular iron-formation. The chert displays an irregular thick bedding, which in places is lensoid. Texturally the chert is very similar to the peloidal chert described in the previous paragraph. A distinctive feature in outcrop is the presence of anastomosing white quartz veins, up to 20 mm thick, which exhibit a crudely radial pattern (Fig. 21). The veins are probably filling syneresis cracks formed during diagenesis.

The Wellesley Chert Member occurs only in the area described above; however rocks equivalent to it occur in a similar stratigraphic position at two other localities. On NABBERU, 4 km south of Oxbys Well, a bed of chert is exposed above the top iron-formation. It is greyish green to white and is variable in texture from massive to peloidal to brecciated (intraclastic). A crude banding on a 1 to 10 mm scale is present in places. Anastomosing quartz veins are rare. At Coonabildie Bluff on STANLEY, dark-green chert, similar to the Wellesley Chert Member, is interbedded with grey limestone. The stratigraphic position at Coonabildie Bluff is not clear, but on a lithological comparison it is probably near the top of the Frere Formation.

Carbonate rocks: In addition to those carbonate rocks now included in the Yelma Formation (e.g. Sweetwaters Well, Sweeney Creek and northeast DUKETON), the Frere Formation contains carbonate lenses in four known localities.

The largest of these is northeast of Simpson Well on NABBERU, where a lens 35 m thick occurs about 500 m above the base of the formation. The carbonate is pale grey-to-white, and is recrystallized to a sparry calcite. Fifteen metres above the base is a band containing fresh pyrite cubes, and faint, small, branching stromatolites. Twenty-five metres from the base, a band 2 m thick contains columnar stromatolites of *Pilbaria deverella* about 300 mm high. Both of these stromatolite forms are similar to forms at Sweetwaters Well (see Yelma Formation). Carbonate near the top of the lens has been partly replaced by chert nodules.

The other occurrences of carbonate are much smaller. A lens of pink-to-grey, poorly laminated limestone, approximately 10 m thick, occurs about 200 m above the base of the formation, 8 km west-southwest of Old Windidda homestead (KINGSTON). A band of stromatolitic limestone is interbedded with iron-formation just above the base of the formation northwest of Mount Deverell on NABBERU (W.D.M. Hall, pers. comm.) and this is probably related to other occurrences nearby in the Yelma Formation. A limestone lens 4 m thick occurs between the uppermost iron-formation and the Wellesley Chert Member at the western end of the Wellington Range. This limestone is pale pink, weathering to grey, and contains flat, domed stromatolites a few centimetres high and between 100 and 500 mm in diameter. Both carbonate and stromatolites are similar to those in the overlying Windidda Formation, and the lens is regarded as a precursor to that formation.

Stromatolites from the Frere Formation are described in Appendix 1 of this bulletin.

DEPOSITIONAL ENVIRONMENT

The Frere Formation represents a shallow, subaqueous environment (as suggested by ripple marks and scour structures) which was probably marine. Clastic material is either absent (in the cherts and iron-formations) or very fine grained, indicating either remoteness from a source, or a hinterland of very low relief, or a barred environment. The cherts and iron-formations represent a period of silica and iron precipitation punctuated by influxes of fine-grained sediment.

Regional lithological variation is a direct response to variation in the local environment of deposition. A higher clastic component (shale, siltstone and fine-grained sandstone) in the southeast part of the basin suggests proximity to the source area of the sediment, and therefore a decrease in water depth. This direction to the sediment-generating shoreline is consistent with directions deduced from the Windidda/Wandiwarra disconformity and palaeocurrent data from the Princess Ranges Quartzite. Detrital quartz grains in oolitic chert in eastern ROBERT perhaps indicate a nearshore environment in that area.

Palaeocurrent data from the Frere Formation are sparse indeed. Micro cross-laminae in shale in southeast NABBERU indicate currents to the northwest, and it is notable that this direction coincides with elongation of wavy and lensoid bedding in iron-formation near Mount Teague (Fig. 22). The cause of this elongation is not known, but the coincidence in direction remains.

Oolitic and pisolitic iron-formations are well developed in the region between northeast WILUNA and central KINGSTON, although poor examples occur elsewhere within the Frere Formation. They represent oolite shoals and banks which must have risen only slightly (less than 1m?) above the surrounding muds of the shelf. Biological activity may have helped to build up these shoals, particularly in the coarser pisolitic varieties where the pisoliths are associated with oncolites, stromatolites, and bacterial microfossils. Carbonate banks were also aided by biological activity in the form of stromatolites, and in places (for example the Simpson Well locality) the

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Figure 21. Anastomosing quartz veins filling syneresis cracks in pale-green chert, Wellesley Chert Member, 3 km south-southeast of Lorna Glen.

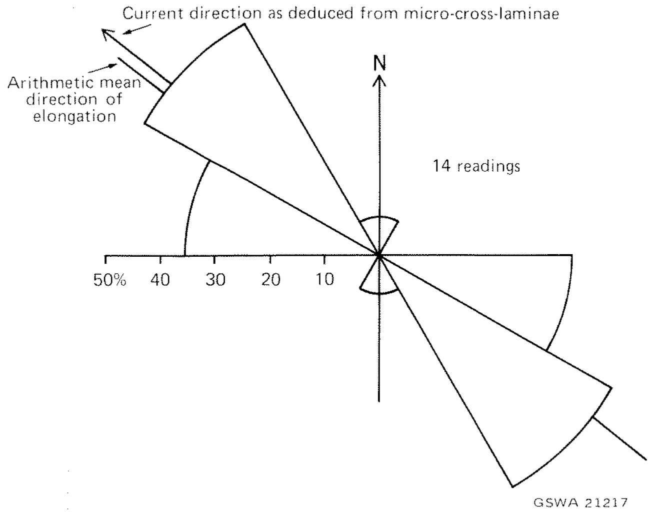

Figure 22. Elongation of wavy and lensoid bedding, Frere Formation iron-formation, 1 km northwest of Mount Teague.

bank may have been several metres above the surrounding shelf.

Facies changes in the iron-formation reflect changes in water depth. The banded iron-formation formed below wave base, where the delicate laminae were disturbed only by gentle bottom currents. The granular iron-formation formed in shallower water on an extensive shelf, where wave action was constantly breaking up recently deposited, laminated iron-formation and massive chert. Brief periods of quiescence allowed thin, continuous chert bands to form. The occurrence of BIF in the western and northern part of the basin and absence in the southeast is another indication of the shallowing of the basin to the southeast. BIF in the northern part of the basin is also concentrated near the top of the formation, suggesting that water depth was increasing towards the end of iron-formation deposition.

As well as the west to east decrease in the iron-formation: shale ratio, there is a decrease in the iron content of the iron-formations. This suggests a source of iron from the west, and the most obvious source is the widespread occurrence of mafic rocks of the Glengarry Sub-basin. Mechanisms for the introduction of the iron into the water of the basin are discussed in Chapter 4.

WINDIDDA FORMATION

DEFINITION

Derivation of name: Named after Windidda homestead in central KINGSTON, Lat. 26°23′S, Long. 122°13′E.

Type section: The type section is taken in central KINGSTON, from Lat. 26°26′30″S, Long. 122°12′00″E to Lat. 26°21′00″S, Long. 122°11′00″E. The sequence, approximately, is as follows.

	Thickness approx. (m)
WANDIWARRA FORMATION; contact with Windidda Formation conformable	
WINDIDDA FORMATION	
8. Unexposed (30 km northwest of type section, the top 2 m is limestone conglomerate, overlying mudstone with thin limestone beds).	20
7. Maroon, pink or greyish-green, micaceous mudstone with thin (20-300 mm) limestone and limestone breccia beds, and minor fine-grained micaceous sandstone. Poorly exposed.	500
6. Pinkish-grey limestone, abundant stromatolites, with interbeds of micaceous mudstone.	10
5. Maroon, pink or greyish-green, micaceous mudstone with thin (100-120 mm) carbonate beds. Very poorly exposed.	200
4. Grey limestone, well-bedded, with thin interbeds of micaceous mudstone and irregular layers of red, peloidal, jasperoidal chert. Stromatolitic in part.	50
3. Multicoloured carbonate in layers 20 to 200 mm thick. Includes greenish-grey limestone, brown dolomite and pink-white, laminated, stromatolitic limestone.	13
2. Interbedded grey limestone and brown ferroan dolomite, with some breccia bands.	1.5-2
1. Interbedded brown-yellow ferroan dolomite and grey-green to black sulphidic chert; transition from Frere Formation.	5
FRERE FORMATION; contact with overlying Windidda Formation conformable	

* * * * *

Lithology: Mainly grey-to-pink, laminated or stromatolitic carbonate (dolomite and limestone) interbedded with maroon or grey-green micaceous mudstone and shale. Some intraclastic breccia beds and jaspilitic chert beds.

Thickness and distribution: Maximum thickness is estimated to be about 800 m in the type section. This decreases to about 300 m in western ROBERT (Mount Elisabeth area) and southeast NABBERU (Oxbys Well area). The formation is restricted to the southern part of the Earaheedy Sub-basin, between southeast NABBERU and central ROBERT.

Boundary criteria: The base is taken as the first appearance of regionally continuous carbonate sedimentation above the Frere Formation. This is difficult to determine in detail because of the transitional nature of the contact, and in the type section it is taken at the base of a unit of interbedded chert and carbonate. Elsewhere, the actual contact is rarely exposed.

The top of the formation is defined by the appearance of exclusively clastic sedimentation of the Wandiwarra Formation. The contact is sharp and probably disconformable. The uppermost Windidda Formation is usually a conglomerate consisting of carbonate clasts set in a carbonate matrix. In places, this is overlain by conglomerate in which carbonate clasts are set in a glauconitic sandstone matrix; the latter is regarded as basal Wandiwarra Formation.

LITHOLOGICAL DESCRIPTION

With the exception of a few areas the Windidda Formation is not well exposed. It occurs in a broad tract of low-lying ground between the resistant iron-formation hills to the south (Wellington Range, Von Treuer Tableland) and the Princess Ranges to the north. Where exposed, the formation has a characteristic, striped air-photo pattern due to the style of exposure in which the gently dipping carbonate layers form low, resistant ridges (sometimes just a few centimetres high) separated by sub-exposure of small mudstone flakes.

The basal few metres of the formation, which are exposed in a gorge 6 km south of Windidda homestead in the type area, consist of interbedded chert and ankeritic dolomite (Fig. 23). The chert varies from pale grey-green to black and is sulphidic, the sulphides consisting mainly of pyrite and rare chalcopyrite. In addition, Hall and Goode (1978) report traces of sphalerite and locally abundant apatite from these cherts. The dolomite is yellow-brown and contains up to 14% iron (Table 4, samples 46312B and 46315B). Fe:Mg ratios are about 2:1, putting these rocks well within the ankerite range (Deer, Howie and Zussman, 1967). In thin section the dolomite is seen to be a microcrystalline sparite.

A notable feature of the ankeritic dolomite is the presence of chamositic chlorite, first described from this locality by Hall and Goode (1978). The chlorite occurs as small flecks up to 1 mm across, and as pellets up to 20 mm long (Fig. 23A), both of which occur within the bedding planes and are almost certainly detrital. One case of crude cross-bedding is

TABLE 4: FULL CHEMICAL ANALYSES, WINDIDDA FORMATION

GSWA Sample No.	*46312B*	*46315A*	*46315B*	*46317*
SiO_2 (%)	13.2	90.4	6.1	10.3
Al_2O_3	0.5	0.3	0.9	1.6
Fe_2O_3	1.1	1.0	1.3	0.4
FeO (a)	13.00	0.86	12.40	0.65
MgO (a)	7.18	0.41	8.76	0.37
CaO	23.91	2.59	25.69	44.58
Na_2O (a)	0.01	0.05	0.07	0.02
K_2O	0.1	0.1	0.2	0.8
H_2O+ (a)	0.62	0.58	0.57	0.31
H_2O- (a)	0.17	0.16	0.16	0.16
CO_2 (a)	37.00	2.60	40.30	37.30
TiO_2	0.00	0.00	0.01	0.06
P_2O_5	0.08	0.01	0.04	0.08
MnO	1.15	0.05	1.58	0.11
TOTAL	**98.0**	**99.1**	**98.1**	**96.8**
Ba (ppm)	20	20	40	180
Cu	30	80	90	105
Pb	40	20	20	20
Ag	5	5	5	5
Sr	30	15	45	100
Zn	2	4	9	10
	Mineral composition (%) (b)			
Quartz	12	90	5	9
Calcite	—	5	—	80
Dolomite	86	2	90	2
Kaolinite	2	—	3	—
Chlorite	—	3	2	1
Mica	—	—	—	8

All samples are from a gorge located 6 km bearing 190° from Windidda homestead.
46312B Brown, ankeritic dolomite
46315A Pale-grey chert
46315B Yellow, ankeritic dolomite
46317 Greenish-grey, fine-grained, sparry limestone
(a) analysis by chemical methods; other analyses by X-ray fluorescence techniques.
(b) derived from XRD and chemistry.
All analytical work by Government Chemical Laboratories, Perth.

defined by the presence of 'trails' of chlorite pellets. The chlorite is pale green, and in thin section shows marked pleochroism, a refractive index of 1.65-1.66, and low birefringence. The XRD pattern shows a chlorite structure with a strong 1.4 nm (14 Å) reflection. These properties are consistent with the mineral "being thuringite, an iron-rich chlorite which may be described in a loose sense, as chamosite" (Government Chemical Laboratories). Some samples contain fine-grained, iron-rich ankerite or siderite replacing the chlorite.

The alteration of chert and dolomite in this basal unit results in an irregular, wavy layering. There is some evidence in the field that dolomite has been replaced by silica to form the chert, giving a form of podding which has been subsequently modified by differential compaction (Fig. 23B). The wavy nature of the upper boundary of some chert layers suggests that they may have originated as microbial mats prior to silicification. Filamentous microfossils occur in apatitic and sulphidic cherts (Walter and others, 1976; Hofmann and Schopf, 1983).

Above the basal dolomite/chert unit, 5 m above the bottom of the gorge, limestone appears in the place of chert for a thickness of about 2 m. The limestone is thickly bedded, pale grey, and contains

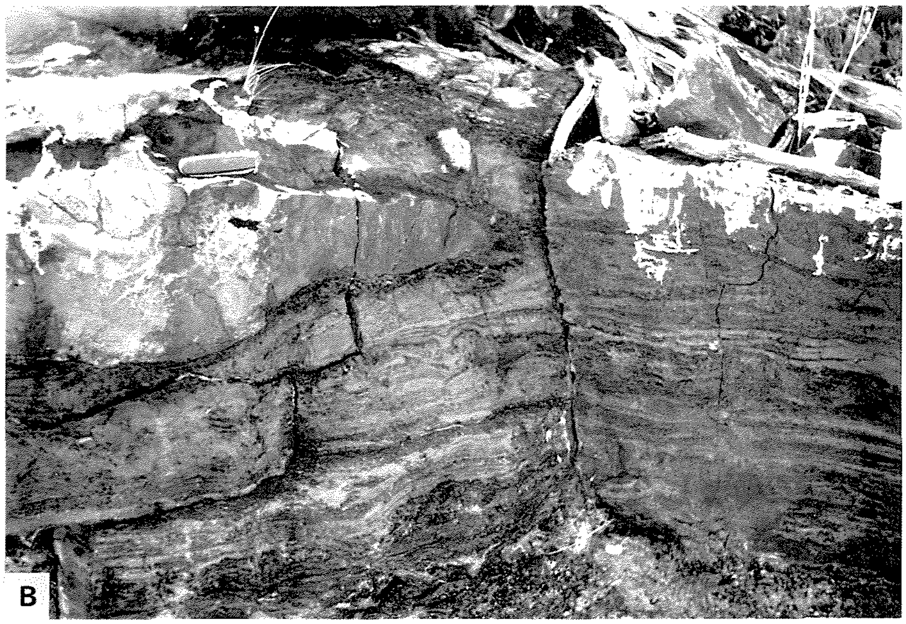

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Figure 23. Ankeritic dolomite and chert unit, base of Windidda Formation, near Tooloo Bluff.

A — Pellets of chamositic chlorite in ankeritic dolomite. Note wavy stromatolitic lamination in silicified carbonate below and to left of knife.

B — Termination of chert layer, replacing dolomite.

some poorly preserved, domed stromatolites up to 500 mm in height. Dolomite and intraclastic breccia bands are a minor component.

The next unit is best exposed two hundred metres north along the gorge. It is a striped facies, consisting of varicoloured carbonates in layers 20 to 200 mm thick (Figs. 24 and 25). Three main variations are present. The most distinctive is a pink and white limestone with a wavy, stromatolitic fabric of hematite and calcite. The limestone layers each have a flat base and a wavy or irregular top due to colloform stromatolitic growth of the laminae. In thin section, the hematite-rich laminae are seen to consist of very fine hematite crystals forming irregular patches set in sparry calcite similar to the white laminae.

Grey limestone layers which weather to a greenish tinge, consist essentially of micritic or very fine-grained sparry calcite. A weak lamination is usually present. In some samples, radiating fans of fine needles or thin plates diverge upwards from bedding planes (Fig. 25C). Individual crystals may be up to 20 mm long, less than 1 mm wide, and are superimposed on bedding laminae. They now consist of the same microcrystalline calcite mosaic as the enclosing limestone, but it is not clear whether the fans were originally calcite or gypsum. The original crystals were undoubtedly early diagenetic, and probably grew from supersaturated pore fluid in a carbonate ooze. As such, they constitute tenuous evidence for local evaporitic conditions.

The brown dolomite layers contain less ferrous iron (sample 46310C, Table 5) than the dolomite in the basal unit and are more properly regarded as ferroan dolomite rather than ankerite. The dolomite consists of a fine- to medium-grained, sparry mosaic of carbonate, with minor chamositic chlorite as small flakes.

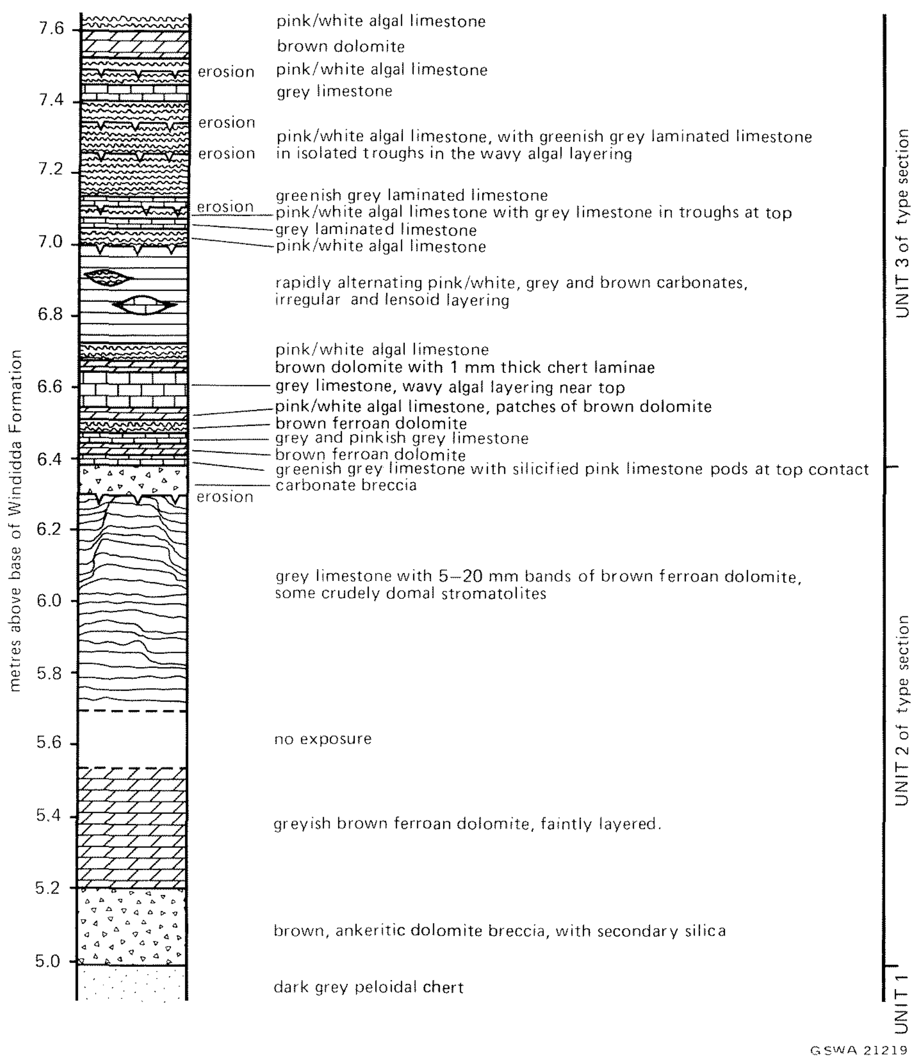


Figure 24. Detailed section in small cliff in gorge near Tooloo Bluff; Windidda Formation, showing style of variation in units 2 and 3 of type section.

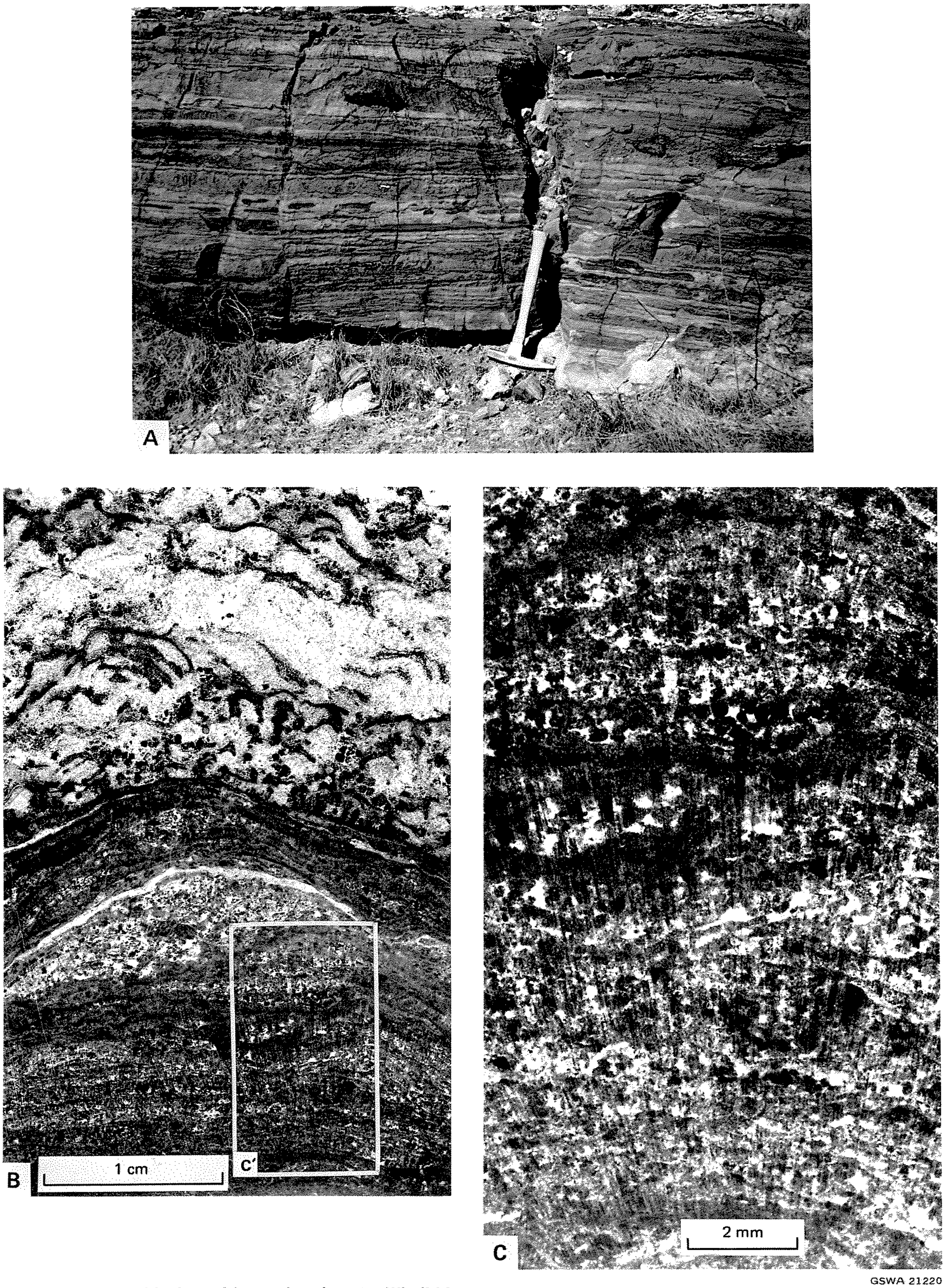


Figure 25. A — Multicoloured layered carbonate, Windidda Formation, Tooloo Bluff.

B — Detail of multicoloured layered carbonate, Windidda Formation, Tooloo Bluff; pink and white limestone, with wavy stromatolitic fabric, overlies grey limestone with relic crystal sheaf-fans superimposed on bedding. GSWA thin section 46422.

C — Detail of single radiating fan consisting of elongate relic crystals of possible evaporitic origin.

A crude cyclicity is displayed in the middle part of the unit (upper part of Fig. 24), where the pink-white stromatolitic limestone is commonly overlain by an erosion surface that marks a hiatus in sedimentation.

The multicoloured carbonate passes upwards into a thickly bedded limestone and shale unit (unit 4 of the type section) which forms low, rugged hills. The limestone is pale grey when weathered, but has a pinkish tinge on a fresh surface. Stromatolitic laminations and poorly developed domes are abundant. A notable feature here is the presence of red jasperoidal chert in irregular layers and lenses. The chert has a peloidal texture and has locally cross-cutting boundaries, suggesting that it is replacing carbonate. Shaly, micaceous mudstone forms thin interbeds with the limestone, and these become thicker and more abundant towards the top of the unit.

Units 1 to 4 are well exposed only within a few kilometres of the type section. Rocks equivalent to unit 4 occur east of Mount Wellesley and east of Fourteen Mile Creek, although in both cases the lower units are not exposed. A small outcrop in a gorge 5 km southeast of Mount Wellesley contains interbedded pyritic chert and limestone, and may equate with unit 1.

Units 5 and 7 of the formation consist mainly of micaceous mudstone with thin limestone beds. The only areas of good exposure are in the type area, in the vicinity of Windidda homestead, and in the headwaters of Wongawol Creek. The mudstone is grey-green or maroon, massive or thinly laminated, and has a weak fissility. Apart from the lamination, sedimentary structures are confined to a few examples of load deformation (on a scale of a few centimetres) and microripples.

The limestone beds in these units vary from 20 to 300 mm thick, and the limestone:mudstone ratio is usually less than 1:4. The limestone beds are usually continuous for several hundred metres, although many are lensoid. Concentrations of limestone beds occur at several localities, forming low ridges and hills. In these concentrations, which may be up to 20 m thick and several kilometres long, mudstone is reduced to a minor component. Unit 6 of the type section is one such concentration.

Stromatolites are abundant in some of the thin limestone beds, and are of a distinctive type described as *Carnegia wongawolensis* Grey 1984. They form domed, circular colonies of small, irregularly branching columns (Fig. 26A). The colonies range in diameter from about 50 mm to over 200 mm, and

TABLE 5. PARTIAL CHEMICAL ANALYSES, WINDIDDA FORMATION

GSWA Sample No.	*46310A*	*46310B*	*46310C*	*46311*
CaO (%)	49.0	46.2	28.0	51.8
MgO	0.73	0.80	13.4	0.66
FeO	0.65	2.59	4.53	0.29
MnO	0.41	0.27	1.39	0.34
Ignition loss	42.1	37.3	40.0	42.7
Ba (ppm)	20	60	80	20
Cu	50	40	130	60
Pb	10	10	10	10
Ag	5	5	5	5
Sr	85	85	85	65
Zn	22	22	72	6
Mineral composition (%) (a)				
Quartz	1	1	1	1
Feldspar	-	-	3	-
Calcite	98	90	-	98
Dolomite	-	-	86	-
Chlorite	2	3	3	2
Mica	-	7	8	-

All samples are from a gorge located 6 km bearing 190° from Windidda homestead.

46310 Multicoloured, layered carbonate: A— red and white limestone with wavy, algal laminae; B— greenish-grey, fine-grained micritic limestone; C— brown sparry dolomite.

46311 Grey, fine-grained limestone.

(a) derived from XRD and chemistry.

All analytical work by Government Chemical Laboratories, Perth.

have a synoptic relief* of up to 60 mm. They protrude into the overlying mudstone, which eroded easily leaving a limestone bedding surface studded with circular or coalesced domes (Fig. 26B).

In northeast WILUNA and southeast NABBERU, outcrops of shale previously placed in the Wandiwarra Formation (Elias and Bunting, 1982; Bunting and others, 1982) are here included in the Windidda Formation, because of a prominent limestone bed stratigraphically above them and marking the top of the Windidda. The shales lack the thin limestone beds of units 5 and 7 of the type section, but are stratigraphically equivalent. They show a pronounced parallel lamination, on a millimetre scale, and individual laminae can, in some instances, be traced for the length of the exposure (up to 50 m in a cliff 2 km northwest of Fergy Bore). Three features disturb the regularity of the laminae.

(a) Small lensoid nodules: these are up to 80 mm long and 15 mm thick and occur singly or at irregular intervals along a lamination. They are internally homogeneous, ferruginous and highly porous, and may originally have been calcareous.

(b) Manganese nodules: these are roughly spherical, up to 300 mm across, and contain cavernous deposits of black manganiferous material. They bow out the laminae above and below them, and occur at intervals along distinct horizons. They are probably the result of diagenetic remobilization of a manganiferous layer.

(c) Clay balls: these are spherical, a few centimetres across, and truncate rather than disrupt the bedding laminae.

These features, and the continuous lamination, indicate a quiet, deeper water environment than that of the stratigraphically equivalent mudstone-limestone facies.

The top of the formation is marked by an increase in the proportion of limestone and an abundance of intraformational conglomerate. The contact itself is usually sharp and disconformable.

The uppermost unit is not exposed in the type section, as it is covered by scree from hills of the overlying sandstone; however, excellent exposures occur near Cork Tree Well to the east, and between Lat. 26° 14′S, Long. 121° 59′E and Lat. 26° 11′S, Long. 121° 43′E, in northwest KINGSTON. Other outcrops of the top unit occur near Oxbys Well in southeast NABBERU. The conglomerate which characterizes the top of the formation (in addition to bedded limestone and mudstone which are similar to underlying units) consists of tabular clasts of bedded limestone set in a matrix of detrital and sparry calcite. At the top contact, the matrix also contains detrital quartz and pelletal glauconite—the precursors of the marine sandstones at the base of the Wandiwarra Formation.

In southeast NABBERU, a 10 m thick limestone band near the top of the formation extends intermittently for 10 km east-southeast from Oxbys Well. The limestone is pink-to-grey, sparry calcite and contains occasional domed stromatolites up to 100 mm high. The limestone is interbedded with, and overlies, jasperoidal granular iron-formation which is similar to that in the underlying Frere Formation. However, the iron-formation here is stratigraphically equivalent to the top of the Windidda Formation, and is considered to be a local facies variant. One variant of the iron-formation is a mixed rock consisting of jasperoidal chert peloids in a carbonate matrix. There is no evidence that the peloids have been silicified after deposition. Most of the peloids are matrix-supported, and it is concluded that they originated as rounded, allochemical jasper grains which were mixed on the sea floor with calcareous ooze, perhaps after they were washed from a shoal-like environment represented by the peloidal iron-formation. No Windidda Formation is exposed between these iron-formations and the Frere Formation 4 km to the southwest. However, the gap in exposure is at the same stratigraphic level as the shales in northeast WILUNA, described above.

The Oxbys Well locality is the most westerly outcrop assigned to the Windidda Formation. To the west, and along the northern edge of the sub-basin, the equivalent stratigraphic position is occupied by shale and fine-grained sandstone assigned to the lower part of the Wandiwarra Formation. A few scattered, and very thin, limestone beds occur in that part of the Wandiwarra Formation. The various facies changes are schematically illustrated in Figure 27.

In the southeast part of the sub-basin, the Windidda Formation extends as far as central ROBERT (Jackson, 1978) before disappearing under younger sediments of the Officer Basin. Near Mount Elisabeth, in western ROBERT, the lowest exposed beds (probably about the middle of the formation) are yellowish-brown to dark-grey stromatolitic dolomite. These are overlain by pale-grey laminated limestone and shale. The stromatolites have been described by Preiss (1976) and Grey (1984) and are *Windidda granulosa* (Preiss) and cf. *Kulparia* f. indet. Preiss. The total thickness of the formation here is only about 300m.

*Synoptic relief: ". . the relief of a stromatolite above its substrate at an instant of time . ." (Walter, 1976); usually indicated by the maximum height of a lamination in a stromatolite column above the corresponding horizon in the columnar interspace.

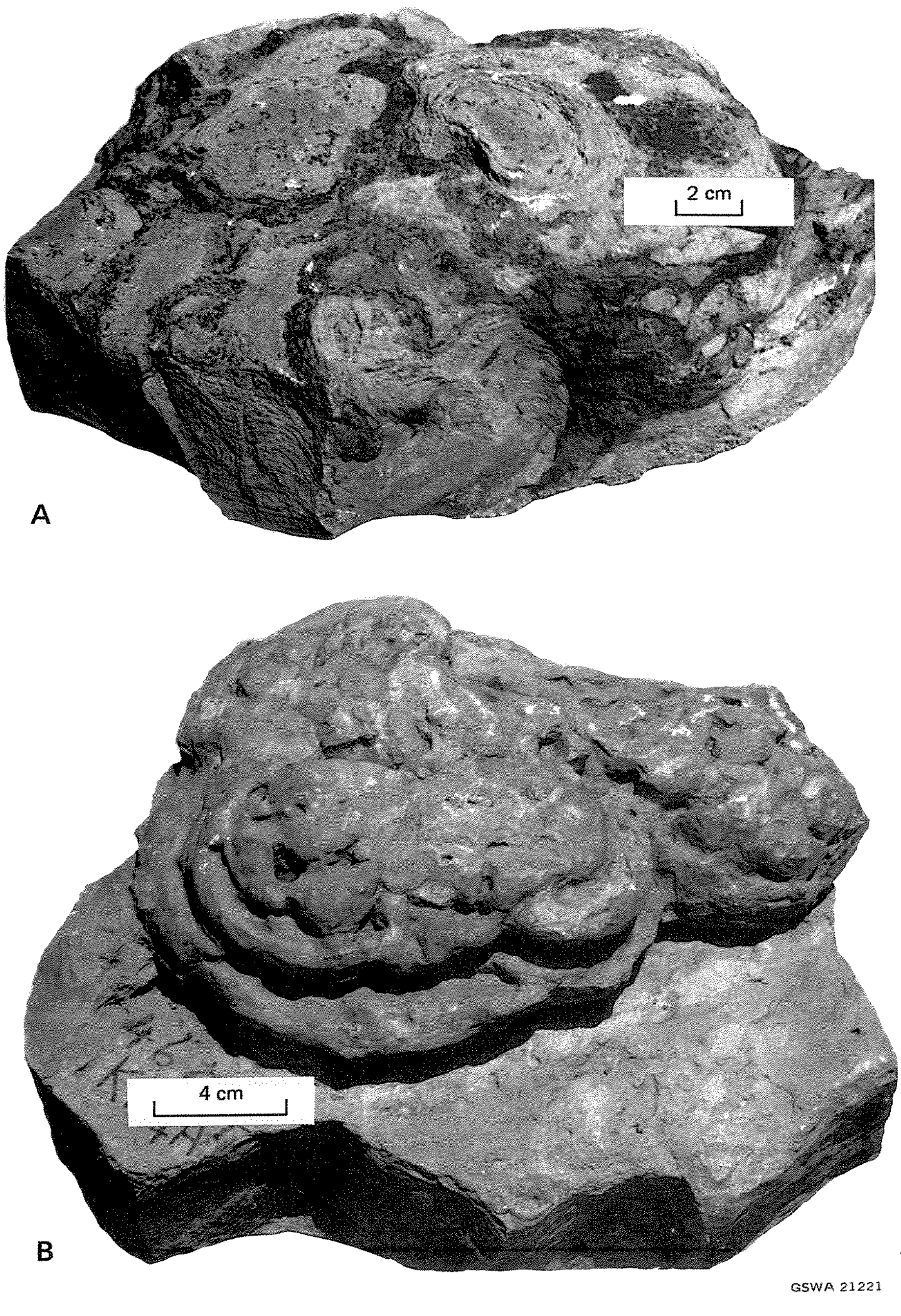


Figure 26. ***Carnegia wongawolensis*** **Grey 1984, columnar branching stromatolites forming small domes, from Windidda Formation.**

A — From limestone (unit 6 of type section) 2 km south of Windidda homestead.

B — Isolated stromatolite on thin limestone bed, from Wongawol Creek, north of Wellington Range.

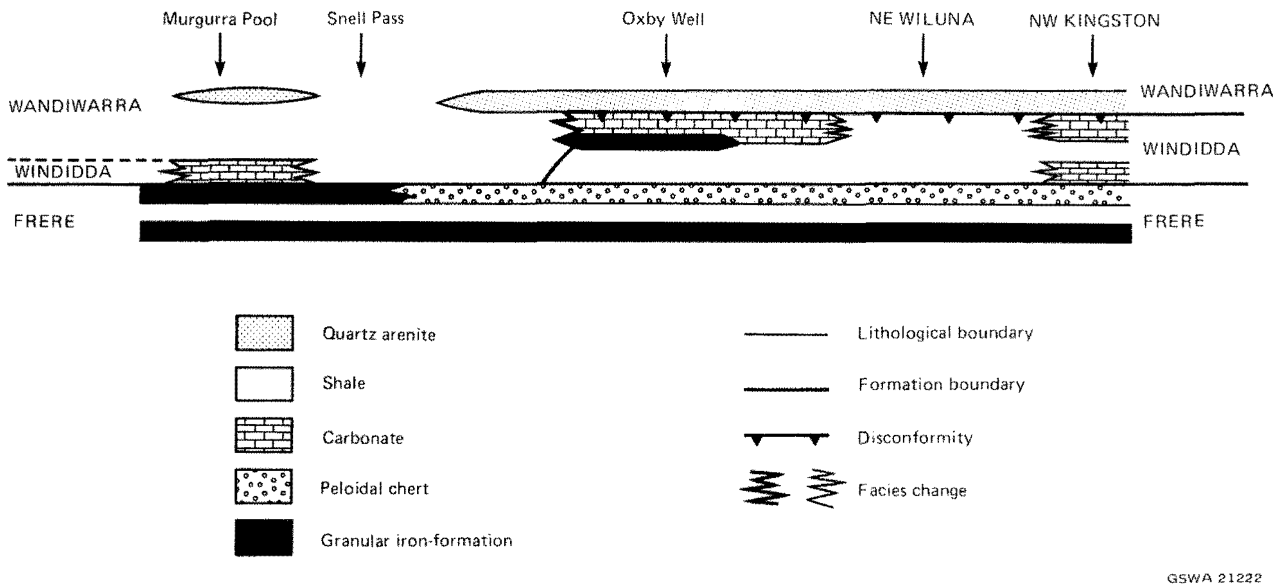


Figure 27. Diagrammatic relationships of stratigraphic units; Windidda Formation, from Murgurra Pool to northwest KINGSTON.

DEPOSITIONAL ENVIRONMENT

The depositional environment of the Windidda Formation is characterized by two main features—a lack of current structures and, as in the preceding Frere Formation, an absence of coarse terrigenous detritus. Only in the basal few metres in the type area is there evidence of weak currents—some poor cross-bedding in dolomite, and one example of stromatolite elongation. An environment away from tidal, shoreline and fluvial influences is indicated, probably in a barred basin or lagoon with a low-lying, mature hinterland. In this current-free environment, the intraclastic conglomerate beds are indicative of storm action and therefore shallow water—a feature also indicated by microripples in some mudstones, and by possible evaporitic textures in the multicoloured carbonate near the base. Planed tops to some stromatolites in unit 6 may indicate localized and temporary emergence, perhaps due to drying-out of the lagoon (K. Grey, pers. comm.).

Lithological features described here and in the following Wandiwarra Formation section, indicate that at the top of the Windidda Formation there was a period of emergence which marks a regional disconformity. Thus the overall pattern in the formation is one of regression in a shallow-water environment.

The start of the Windidda Formation saw a change from the silica and iron oxide precipitation of the Frere Formation, to dominantly carbonate precipitation in the early stages of the regressive environment—a situation also reported by Hall and Goode (1978) who envisage the carbonate rocks as having formed closer to the shore and therefore in shallower water than the underlying oxide-facies iron-formation. A complication is introduced by the presence of iron carbonate, iron silicate and sulphidic chert in the transition from Frere to Windidda Formation. This may indicate a brief increase in the water depth prior to the rapid regression.

The thickest known development of the formation is in the vicinity of Windidda homestead, and it is here that the regressive change to carbonate sedimentation began, with the development of an extensive carbonate bank which may have extended as far west as Mount Wellesley. The carbonate bank was only a few metres high, sufficient to create a barrier which cut off the area to the east from the open sea; the thick mudstone-limestone sequence, which forms the bulk of the formation, was deposited within the lagoonal environment behind this barrier. The shale facies in northeastern WILUNA and southeastern NABBERU was seaward of the carbonate banks-lagoonal environment, in slightly deeper water, below wave base on a shallow, flat, sea floor.

The regression culminated with partial emergence in the KINGSTON area. The western limit of this emergence is represented by a thick carbonate sequence in the headwaters of Wongawol Creek. This sequence contains the most westerly example of carbonate conglomerate with a glauconitic sandstone matrix-conglomerate which probably formed by erosion of a carbonate bank by the Wandiwarra transgression. In southeast NABBERU, the uppermost Windidda Formation shows evidence of a very shallow-water, shoaling environment, but no emergence.

WANDIWARRA FORMATION

DEFINITION

Derivation of name: Named after Wandiwarra Well (now abandoned) on KINGSTON, Lat. 26°22'30"S, Long. 122°19'30"E.

Type section: The type section is taken near Cork Tree Well on KINGSTON, between Lat. 26°23'20"S, Long. 122°20'50"E and Lat. 26°22'20"S, Long. 122°21'10"E. The formation here is 350 m thick, and is as follows.

* * * * *

	Thickness approx. (m)
WANDIWARRA FORMATION	
5. Quartz arenite, medium-grained, pale-brown, slightly ferruginous and clayey; patchy carbonate cement.	20
4. Shale, minor sandstone. Poorly exposed.	100
3. Quartz sandstone, fine- to medium-grained, pink-brown, thick-bedded; contains minor detrital muscovite, plagioclase and iron oxide. Some interbeds of coarse ferruginous quartz arenite and maroon-purple micaceous shale.	100
2. Shale, minor sandstone. Poorly exposed.	100
1. Quartz arenite, fine- to medium-grained, flaggy, laminated, glauconitic; contains shale intraclasts.	30
WINDIDDA FORMATION; contact with overlying Wandiwarra Formation disconformable.	

* * * * *

Lithology: Predominantly shale and sandstone, with minor conglomerate. The shale is laminated, micaceous, and grey, brown, green or maroon. The sandstone ranges from fine to coarse grained, is quartz rich, but can contain significant amounts of iron oxide (detritus or cement), mica, glauconite, feldspar, and carbonate cement. Ripple marks, flute casts, cross-bedding and intraclasts (shale and sandstone) are locally abundant.

Boundary criteria: Both the lower and upper boundaries of the Wandiwarra Formation are interpreted to be diachronous. Where the underlying Windidda Formation is present, as in the type section, the lower boundary is taken at the base of the first sandstone or quartz-matrix conglomerate bed above the carbonates of the Windidda Formation. In places, a shale unit a few metres thick is present between the carbonate or mudstone and the sandstone, and this shale is included in the Wandiwarra Formation. The boundary between Wandiwarra and Windidda Formations may represent a local disconformity (Hall and Goode, 1978). Where the Windidda carbonates are absent, in the western and northern parts of the sub-basin, the base of the Wandiwarra Formation is taken at the top of the uppermost chert or iron-formation of the Frere Formation. In this case, the contact is conformable.

The upper boundary is taken at the base of the first mature quartz arenite of the Princess Ranges Quartzite. This appears lower, and probably earlier in the sequence at the southeastern end of the sub-basin than at the northwestern end.

Distribution and thickness: The Wandiwarra Formation is an extensive unit throughout NABBERU, central and northwest KINGSTON and west-central STANLEY. Because of facies changes and diachronous boundaries, the thickness varies considerably from zero, in eastern KINGSTON (where Princess Ranges Quartzite rests directly on Windidda Formation) to perhaps 1 500 m in central NABBERU, where the lower 800 m or so is the lateral equivalent of the Windidda Formation. The formation is 350 m thick at its type section, and about 500 m thick in northwest KINGSTON.

LITHOLOGICAL DESCRIPTION

Typical rocks of the Wandiwarra Formation are green, green-brown or maroon siltstones, fine-grained lithic sandstone, and fine- to coarse-grained quartz arenite. The quartz arenites form cuestas, or isolated, cuesta-shaped hills, and occur as lenses up to 20 m thick and several kilometres long. On KINGSTON the arenites are abundant in the lower half of the formation and near the top, but a thick siltstone and fine-grained sandstone unit is conspicuous in the upper half of the formation, where it forms a broad belt of low-lying ground, e.g. in the valley of Two Mile Creek. The arenites in the lower half of the formation continue intermittently through NABBERU and STANLEY, where, because of the inclusion of Windidda equivalent in the Wandiwarra Formation, the arenites occupy a position in the central part of the formation.

The quartz arenite is typically a pale-grey, texturally mature rock in which the quartz grains are well sorted, rounded, and surrounded by authigenic silica which is in optical continuity with the host grains. Accessory minerals may include biotite, feldspar (generally altered to clay minerals), glauconite, and tourmaline. Nearly all the quartz arenite beds are ferruginous. Iron occurs either as a cement, or as hematite after detrital magnetite, or, in the basal parts of the formation, as hematite or jasper peloids derived from the underlying Frere and Windidda Formations. Some ferruginous grains may be after glauconite. In places, rhomboid brown spots, up to

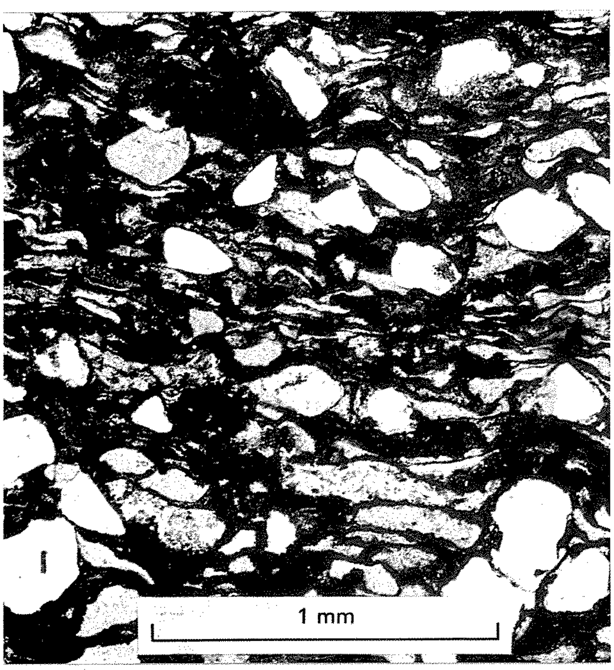
20 mm across, are apparent on the weathered surface. The spots contain normal quartz grains, with an iron oxide staining which is probably replacing carbonate cement. The grey colour, ferruginous nature and slight mineralogical immaturity distinguishes the Wandiwarra quartz arenites from those in the overlying Princess Ranges Quartzite.

The fine-grained sandstone and siltstone are generally chloritic and micaceous and contain subangular quartz. The chlorite is detrital, and occasionally forms aggregates presumably derived from chloritic schist. Lamination is well pronounced in the siltstone, but is less apparent in the sandstone. Glauconite pellets occur in some of the sandstones, and these may be streaked out and deformed in the vicinity of shaly intraclasts (Fig. 28).

On weathering, the shales and fine sandstones become white, cream, pink or red-brown, and are kaolinitic with some iron staining. Manganese surface staining is common, and in places (e.g. west of Earaheedy), manganese enrichment almost reaches ore grades (see Chapter 7). Manganese nodules a few centimetres across are scattered through some of the shales.

SEDIMENTARY STRUCTURES

Bedding varies from very thick to massive in some of the coarser sandstones, through medium and thin in the fine-grained sandstones, to a millimetre-scale lamination in the shales. In the sandstones, the bedding is occasionally lensoid, especially where the sequence is variable and contains alternating coarse and fine beds. Scour channels are locally abundant. Cross-bedding is evident in some of the sandstone, and is nearly always trough rather than planar in style. Sets are generally between 200 and 300 mm thick.



GSWA 21223

Figure 28. Deformed glauconite pellets (mottled) and subangular quartz grains in ferruginous cement. Basal Wandiwarra Formation, 11 km south of Wongawol homestead. GSWA thin section 46414.

Green shale and fine-grained sandstone near Skull Soak, on KINGSTON, show a fine lamination together with a rhythmic bedding on a 100 to 200 mm scale (Fig. 29). Some laminae are wavy and in places are disrupted by incipient convolutions. These rhythmites may represent the distal portions of turbidites.

Ripple marks, which are abundant in some fine-grained sandstones, include symmetrical (oscillation), asymmetrical and linguoid varieties. Wavelengths are generally between 80 and 150 mm, although 2 km west of Breakaway Bore (southwest STANLEY) symmetrical ripples with slightly sinuous crests have wavelengths up to 500 mm.

Flute and current-crescent moulds occur on the base of some arenite beds. The moulds usually occur singly, and are seldom more than 50 mm long. The obstacles causing the current crescents are never present, but were probably shaly intraclasts. The moulds have been subjected to weak load-casting. Loading is also responsible for the development of incipient ball-and-pillow structures, although these are rare.

Intraclasts of shale and ferruginous arenite are abundant in the quartz arenite, and they may account for up to 40% of the rock. The intraclasts are tabular, generally aligned parallel to bedding, and commonly weather out, leaving angular holes.

NATURE OF BASAL CONTACT

On NABBERU and STANLEY, where the Windidda Formation carbonates are absent and the dominantly shale sequence is included in the Wandiwarra Formation, the contact between the Wandiwarra and underlying Frere Formation is conformable. However, on KINGSTON, the basal contact of the Wandiwarra Formation with the Windidda Formation is more complex, and is worthy of description in some detail.

Nine stratigraphic sections through the contact zone are shown in Figure 30, which covers a 70 km strip from the type section to northwest KINGSTON. The exposures are variable in quality, but the contact is well exposed between K169 and K236, where the basal sandstones form a low escarpment.

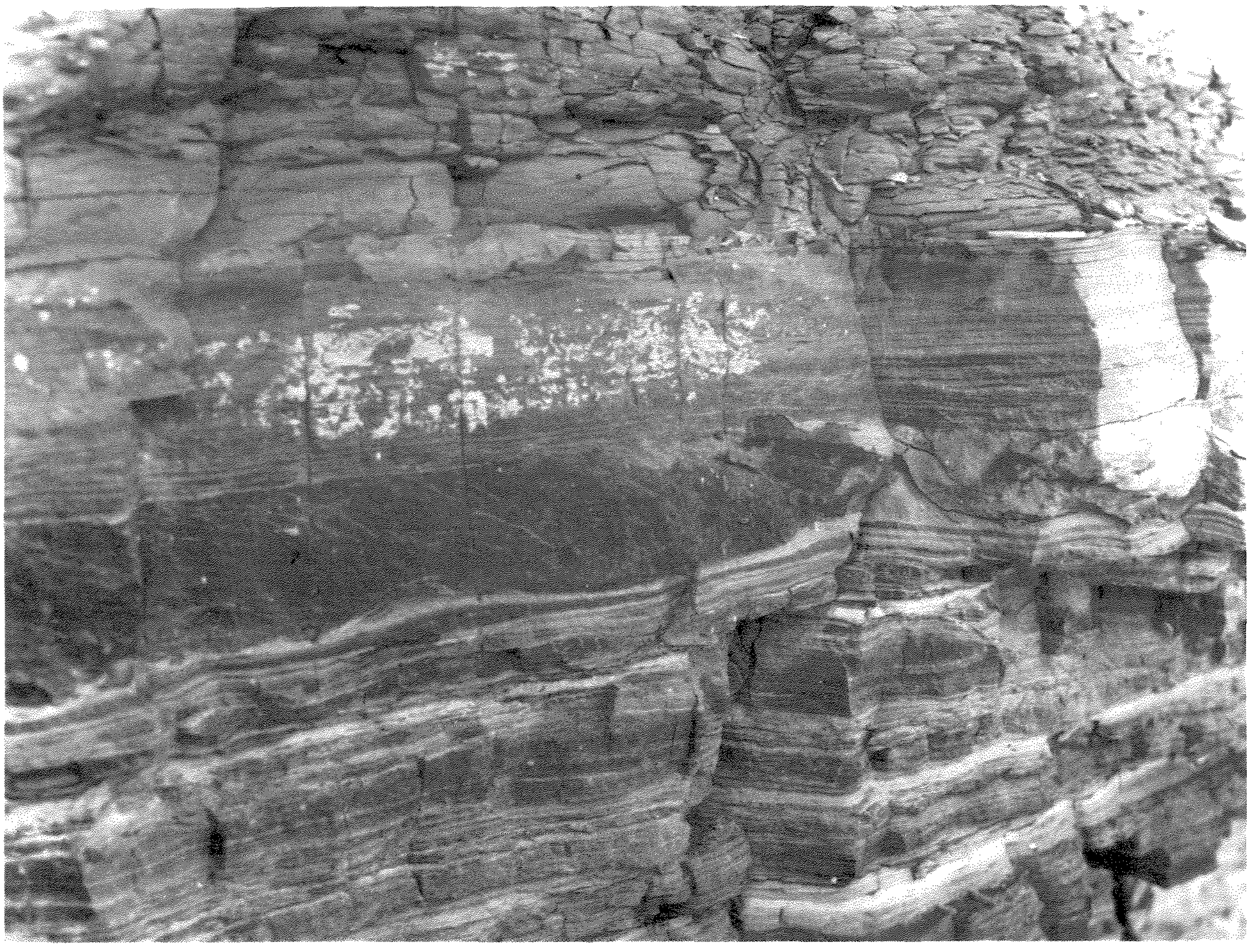


Figure 29. Thinly bedded, fine-grained sandstone and laminated green shale. Top of Wandiwarra Formation, Skull Soak.

The uppermost beds of the Windidda Formation are a mixture of grey-green or maroon micaceous mudstone and grey or pink carbonate. Some of the carbonate beds are intraclastic breccias. The basal beds of the Wandiwarra Formation are more variable, but fall into three lithological types (fine-grained, arenaceous, and conglomeratic), any one of which can form the basal bed on top of the carbonate-mudstone sequence.

The fine-grained basal facies, consisting of micaceous siltstone, shale and fine-grained silty sandstone, is easily distinguishable from the Windidda mudstones by the buff to cream colour and coarser overall grain size of the younger unit. The quartz arenite is well sorted and mature, but is variable in both grain size and mineralogy (particularly the amount of glauconite and iron oxide). Some arenite contains small peloids of hematitic chert, which may be derived from the underlying Windidda or Frere Formations. Glauconite pellets are commonly deformed by compaction against shale intraclasts, to form a streaky or wavy texture (Fig. 28). These deformed pellets are similar to the accommodation shards of Dimroth and Chauvel (1973).

Wherever it occurs along the contact zone, the conglomerate unit is always basal. The conglomerate consists of clasts up to boulder size in a matrix of medium-grained quartz arenite which is usually rich in glauconite. The matrix is commonly lustre-mottled due to sparry calcite cement. The clasts in the conglomerate are mainly limestone (Fig. 31), although in section K168 (Fig. 30) the conglomerate contains a large proportion of ferruginous shale clasts. The limestone clasts can be matched with rock types in the underlying Windidda Formation. Some clasts themselves consist of intraclastic breccia. At K102, some subangular fragments of banded limestone are surrounded by concentrically banded carbonate. The origin of these structures is not clear, as much of the microtextural evidence is obscured by recrystallization, but they are interpreted to be derived from the reworking of collapsed tepee structures. Assereto and Kendall (1977) describe similar features associated with tepee structures which are in the senile stage of development, and attribute them to a vadose origin during periods of partial emergence. At K102, pockets of red, ochrous material occur between some of the concentrically zoned structures, and these may represent the terra rossa found in

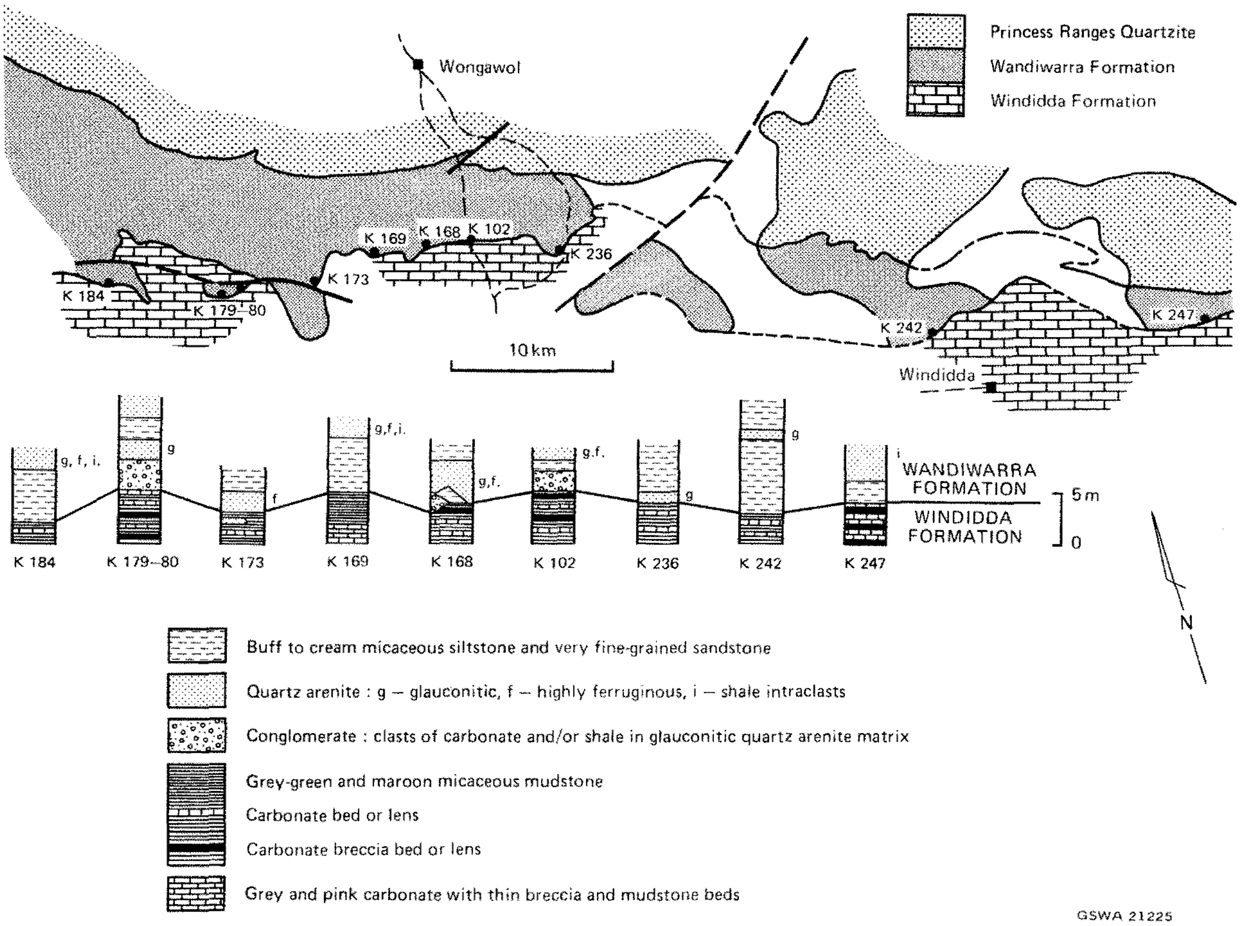


Figure 30. Basal contact of the Wandiwarra Formation, KINGSTON.

senile tepees (Assereto and Kendall, 1977). Hall and Goode (1978) have suggested a vadose origin for wavy pisolites with radial structure in the Wandiwarra Formation.

The conglomeratic basal facies is not laterally extensive, but forms elongate lenses; this is best displayed by the highly ferruginous conglomerate which fills an erosional depression in the underlying Windidda Formation at section K168 (Fig. 30).

Thus, for at least part of the contact there is evidence of subaerial exposure and erosion. The contact is therefore disconformable, a conclusion reached by previous workers (Hall and Goode, 1975, 1978; Bunting, 1980). Horwitz (1975a) proposed a major unconformity in this position, corresponding to a period of granite intrusion in the western part of the basin, but this event is not substantiated by the present work. Rather, the evidence points to the disconformity being a localized event. The disconformity can be advocated with some certainty where there is evidence, from the carbonate conglomerates, of subaerial exposure. Elsewhere, the sharp, sudden change in sedimentation type, with the incoming of coarser, mature clastic sedimentation, may be regarded as evidence of a major marine transgression, with or without subaerial exposure. However, in some sections the contact is not sharp, implying a brief transitional passage from carbonate to clastic sedimentation. This is particularly the case in section K173, where, in the banks of a large creek, a lens of glauconitic sandstone is seen within the green and maroon mudstones of the Windidda Formation.

The contact between these two formations is therefore complex. It marks a major marine transgression, following the regression that culminated in partial emergence at the top of the Windidda Formation. Even where there is no emergence (and therefore no disconformity), the sudden change in sedimentation makes the transgression readily recognizable.

DEPOSITIONAL ENVIRONMENT

The Wandiwarra Formation is totally a marine sequence. The early part of the sequence, on NABBERU and STANLEY, marks the disappearance of chemical sediments after the Frere Formation and an increase in the amount of clastic input. Deposition occurred in a shallow-marine environment, with some sand shoals, in slightly deeper water offshore from the lagoonal muds and carbonates of the Windidda Formation. The regression which affected the upper part of the Windidda probably also affected the laterally equivalent sequence in the Wandiwarra Formation, in that the water became shallower, but there was no major change in sedimentation.

Following the regression, a rapid transgression moved south and east across the partly emerged carbonate and mud shelf of the Windidda Formation. Carbonate and granular jasperoidal chert near Oxbys Well, on southeast NABBERU, mark the northwest extremity of this shelf. Here the earliest evidence of the succeeding transgression is a lens-shaped body of mature quartz arenite at the base of the Wandiwarra Formation. This quartz arenite, which is interbedded with white siltstone, displays ripple marks, cross-bedding, intraclasts, current lineation, and brown spots after carbonate cement. In all its features it is identical to the Princess Ranges Quartzite, and it probably represents a shoaling, shallow-water environment. It must represent a pause in the transgressive advance of the sea, during which quartz sand was constantly being reworked. Subsequently, the transgression proceeded rapidly to produce the less mature, but still well-sorted, glauconitic quartz arenites on KINGSTON.

After the transgression, the water deepened, and most of the formation was deposited in an open-marine environment. In general, the grain size of the sediments decreases upwards as the water depth increased. Towards the top of the formation the green shales exposed near Skull Soak are indicative of the greatest depth of water to be found anywhere in the Earaheedy Group. The even, rhythmic bedding and fine lamination in these shales suggests that they are possibly distal turbidites.

Near the top of the Wandiwarra Formation, lenses of white, mature quartz arenite are precursors of the Princess Ranges Quartzite which marks a major regression. The lenses coalesce to the east-southeast to form continuous arenite units. These appear progressively earlier towards the eastern side of the sub-basin, until at the eastern edge of KINGSTON the mature quartz arenite sits directly on Windidda Formation carbonates. Thus the regression began in the east and proceeded towards the west-northwest.

The source of the detritus in the Wandiwarra Formation was probably to the north and northwest. During the early transgressive phase of the sedimentation, the sand and silt of the Wandiwarra Formation would have been derived from offshore, rather

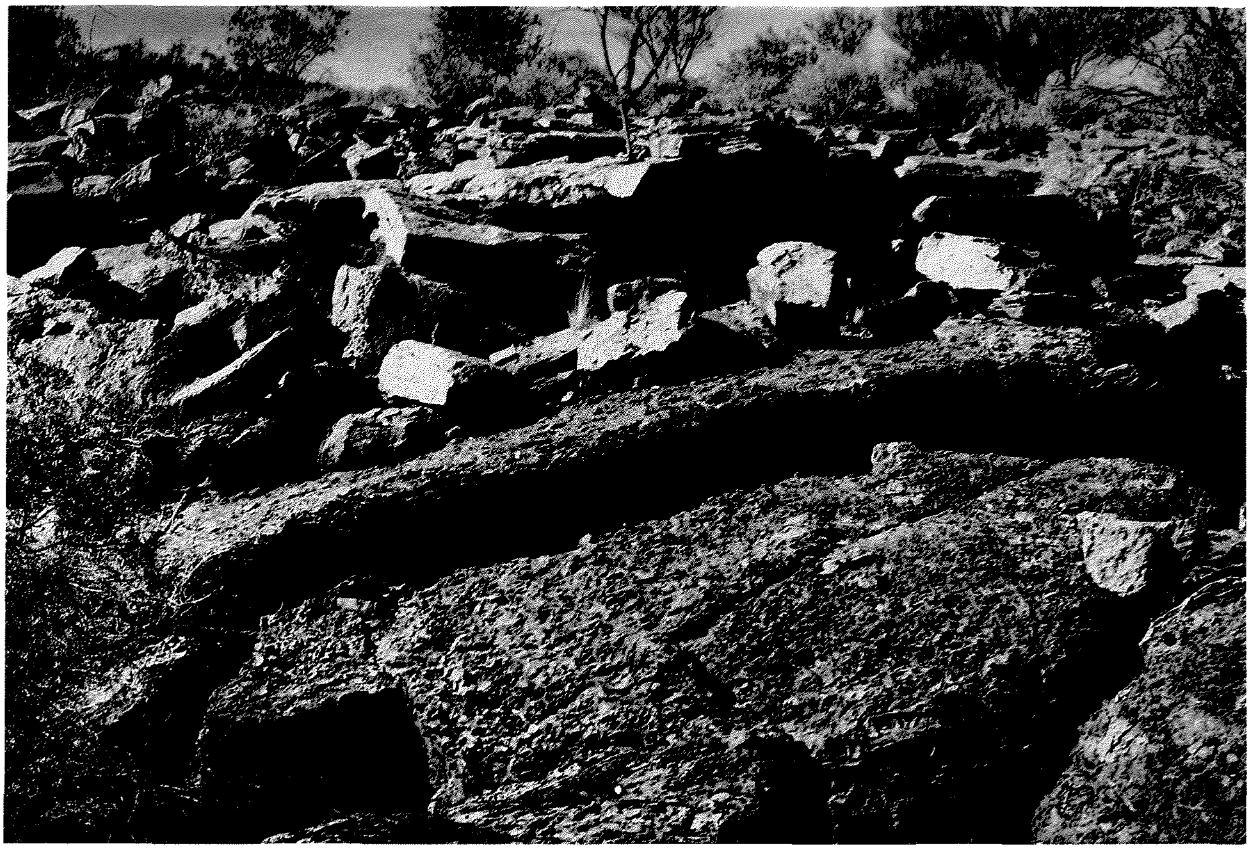


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Figure 31. Basal Wandiwarra Formation, locality K179 (Fig. 30). Glauconitic quartz arenite rests on conglomerate consisting of limestone clasts in a quartz-glauconite matrix.

than across the wide, partly emerged carbonate banks and lagoonal sediments of the Windidda Formation. Once the transgression was established, some material may have originated from sluggish rivers on the low-lying landmass to the south and east.

PRINCESS RANGES QUARTZITE

DEFINITION

Derivation of name: Named after the Princess Ranges in northwest KINGSTON.

Type section: Along an unnamed creek that crosses the Princess Ranges, between Lat. 26° 04′ 00″S, Long. 121° 56′ 40″E and Lat. 26° 02′ 40″S, Long. 121° 45′ 00″E.

Only the lower part of the type section is well exposed. A general description of the type section is as follows:

* * * * *

	Thickness approx. (m)
WONGAWOL FORMATION; contact with Princess Ranges Quartzite conformable	
PRINCESS RANGES QUARTZITE	
5. Mature, white quartz arenite with siltstone interbeds.	40
4. Siltstone with minor sandstone, largely unexposed.	65
3. Mature, white quartz arenite with siltstone interbeds.	20
2. Siltstone and minor sandstone, mostly unexposed.	90
1. Mature, white quartz arenite, with variable amounts of siltstone and fine-grained feldspathic sandstone.	65
Total thickness of Princess Ranges Quartzite	280
WANDIWARRA FORMATION; contact with overlying Princess Ranges Quartzite conformable.	

* * * * *

Lithology: The diagnostic lithology is a white, well-sorted, fine- to medium-grained quartz arenite (orthoquartzite) which is interbedded with white to buff, kaolinitic siltstone and silty sandstone. The arenite varies from thinly to thickly bedded and displays a wide range of sedimentary structures. The formation commonly forms rugged hills in which the finer grained rocks are obscured by quartzite rubble.

Thickness and distribution: In the type section, the formation is about 280 m thick. To the northwest, and along the northern margin of the sub-basin, this thickness is probably maintained, although poor exposure and folding make calculations impossible. Southeast from the type section, mature quartz arenite appears progressively lower in the sequence, and thus the Princess Ranges Quartzite, which is defined as the incoming of mature arenite, thickens at the expense of the underlying Wandiwarra Formation. At the eastern edge of KINGSTON, the Princess Ranges Quartzite rests directly on Windidda Formation and may be 600 m thick.

Boundary criteria: The lower boundary is taken at the base of the first regionally continuous, white, mature quartz arenite above the green shale and immature sandstone of the Wandiwarra Formation. This boundary is diachronous. The upper boundary is taken at the top of a similar, regionally continuous arenite which is followed by fine-grained feldspathic quartz sandstone and shale of the Wongawol Formation. Both boundaries are conformable and, at outcrop scale, gradational.

LITHOLOGICAL DESCRIPTION

Where the Princess Ranges Quartzite is well exposed, for example in the lower part of the type section, it can be seen that quartz arenite (quartzite) is very much subordinate to white or cream micaceous siltstone and fine-grained feldspathic sandstone. However, over much of the formation, the softer siltstone and sandstone are obscured by rubble and scree of quartz arenite. The quartz arenite is typically white, supermature, consisting almost entirely of fine- to medium-grained detrital quartz grains and authigenic silica cement, which gives a very hard, resistant, indurated rock. Rare accessory detrital minerals include tourmaline, zircon, muscovite, magnetite, rutile-bearing quartz, and glauconite pellets. In some beds, brown ferruginous spots several centimetres across give the rock a speckled or blotchy appearance ("dalmation texture" of Hall and Goode, 1978). The spots are roughly diamond-shaped, and are probably weathered-out rhombs of carbonate cement (Fig.32).

SEDIMENTARY STRUCTURES

The following descriptions of sedimentary structures principally come from the lower part of the type section, a lithological profile of which is given in Figure 33.

Cross-stratification: Medium-scale cross-bedding is common in most of the arenaceous units. Sets range in thickness from 30 mm to about 500 mm, and these can occur as either single or multiple sets. The foreset beds are generally between 10 and 30 mm thick.

Both trough and planar cross-bedding are present, but the trough cross-beds are by far the more common. These exhibit near-tangential bottom contacts, and usually, an erosional top. On a principal bedding surface, the cross-beds form large crescentic troughs up to 1 m across. Rarely, the orientations of successive sets are opposed, giving herringbone-style cross-beds (Fig. 32).

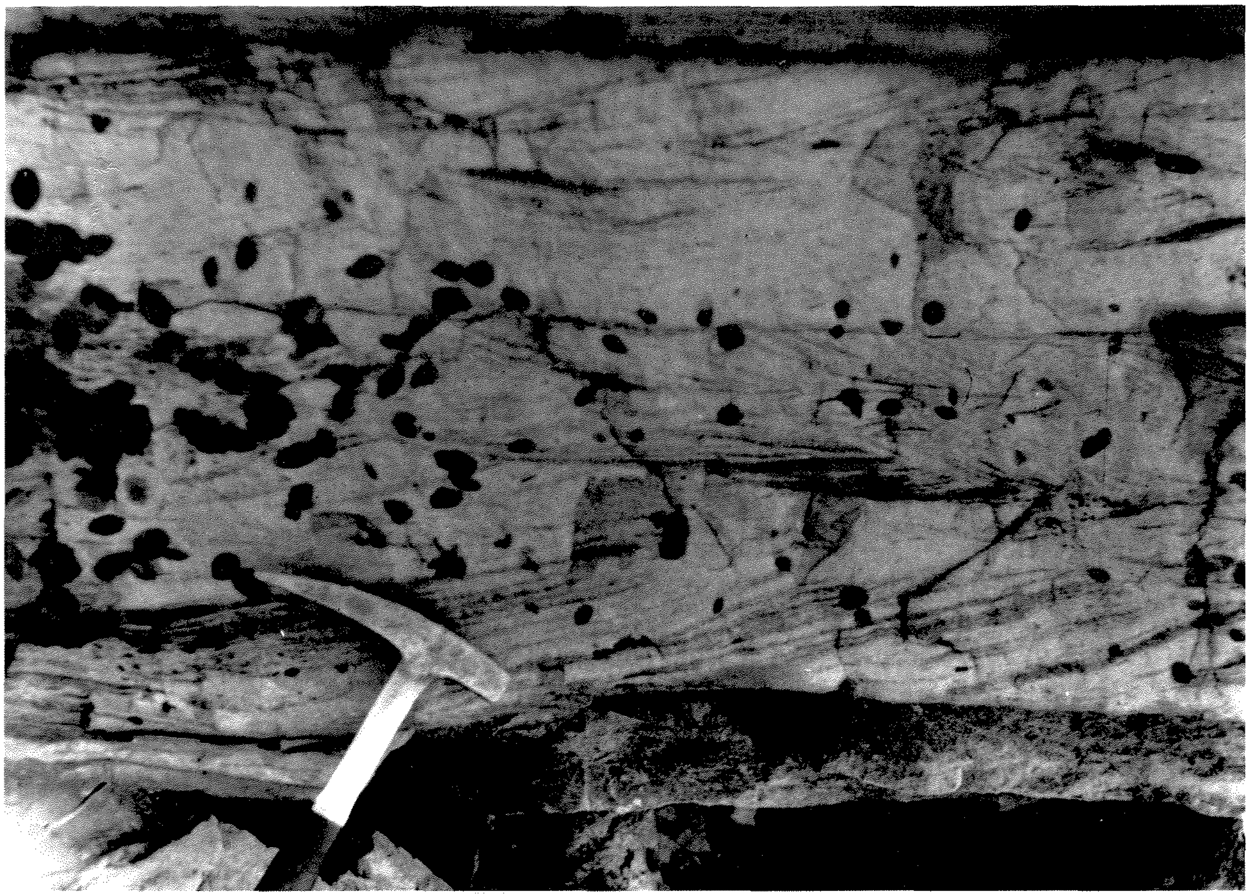

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Figure 32. Typical quartz arenite of the Princess Ranges Quartzite, showing ferruginous rhombs after carbonate, and herringbone cross-beds. Lower part of type section.

Cross-beds in which the foresets are subplanar are uncommon. They are usually associated with elongate lenses or wedges of sandstone which may represent advancing sand waves. In the arenite band Q2 (Fig. 33), planar cross-beds are capped by a flat erosion surface containing abundant shale and sandstone intraclasts.

Small-scale (ripple) cross-lamination takes several forms and is abundant in some silty and fine-grained sandstone units. These various types of ripple cross-lamination form complex vertical sequences in which "draped" cross-laminae are succeeded by in-phase climbing ripples, which in turn are followed by wavy lamination which passes into flat-laminated siltstone. Trains of isolated sand ripples are common in flat-laminated siltstone. The isolated ripples usually have a symmetrical shape, with a consistent dip direction on foreset laminae. Climbing ripples-in-drift occur rarely, and usually pass upwards into the in-phase type. Flaser bedding is common in the ripple-cross-laminated sequences, but is seldom as well developed as the example shown in Figure 34A.

Ripple marks: Small-scale ripple marks, with wavelengths under 150 mm, are well exposed at numerous localities in the type section. Four types are recognized:

(a) Symmetrical ripples: these are generally of the "in phase" type of ripple drift. Occasionally the symmetrical shape may disguise a preferred orientation of foreset laminae. The symmetrical ripples are considered to be of the wave-generated or oscillation type. The crests are usually straight or slightly sinuous.

(b) Asymmetrical ripples: these have a short lee side and a long stoss side, and are due to current action. They have straight or sinuous crests, and pass into linguoid ripples. Current directions vary from bed to bed.

(c) Linguoid ripples are a variety of asymmetrical ripples which have short crests that are convex in a down-current direction. They are associated with sinuous-crested asymmetric ripples.

(d) Interference ripples: these have a reticulate pattern formed by two intersecting sets of ripple crests. They occur when a change in current direction superimposes ripples on previously formed ripples. Where the early ripples are of linguoid type, the resultant interference ripples have the appearance of small, closely spaced but isolated, non-directional mounds.

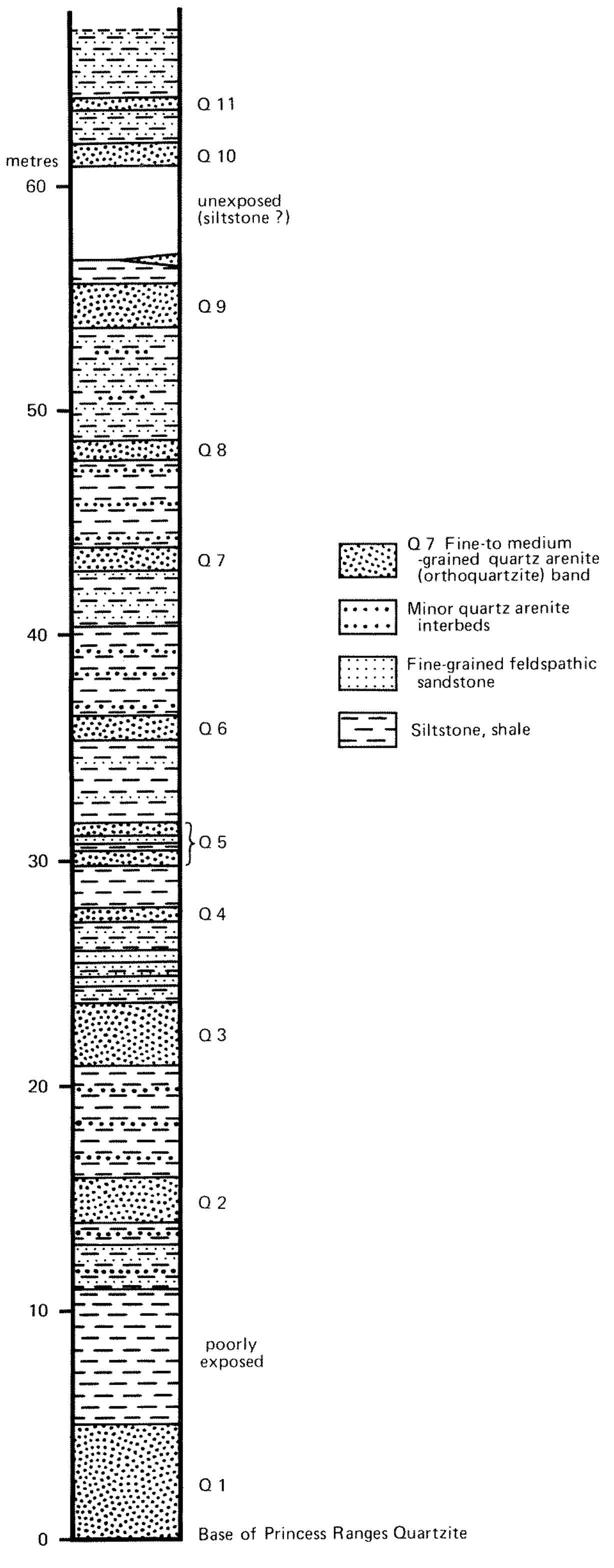


Figure 33. Measured section, lower part of Princess Ranges Quartzite type section.

Large-scale ripples, with wavelengths greater than 150 mm, are always symmetrical. The average wavelength for these ripples is about 800 mm, although examples up to 1.5 m have been found. Ripple crests are straight or slightly sinuous (Fig. 34B) and bifurcations are common. Internal bedding lamination is not usually visible, but in places it is seen to consist of several opposing sets of subplanar cross-beds. This may result from oscillatory waves in an environment also influenced by ebb and flood tides.

Hydroplastic load structures: These result from the foundering of more-or-less cohesive sands into a sub-layer of less-cohesive silt or silty sand. The foundered sands form ellipsoidal or irregular bodies within the underlying layer. In places, these sand bodies may still be attached to the parent layer, in which case they form bulbous "synclines" or lobes, separated by peaked anticlines in which the underlying unit has been injected upwards into the sand. Bedding in the detached portions is commonly pinched at the ends (Fig. 34C) but in places it has been rolled into concentric pseudonodules.

These structures are restricted to well-defined layers which may be from 300 mm to 800 mm thick. The best preserved example occurs within quartz arenite Q3 (Fig. 33); the layer is 800 mm thick and is overlain by 800 mm of trough-style cross-bedded arenite. Traces of cross-bedding are also preserved in both detached and attached lobes (Fig. 34C).

Several constraints on the origin of these structures are provided by field evidence:

(a) Preservation of folded current lineation within the detached lobes precludes a primary current origin.

(b) The intrusion of the fragmentary layer into the overlying cross-bedded arenite indicates that the cross-bedded unit was present during the formation of the flow structure.

(c) In both the detached and attached lobes, synclines prevail over anticlines. This is a similar situation to that described in experimental work on ball-and-pillow structures by Kuenen (1965).

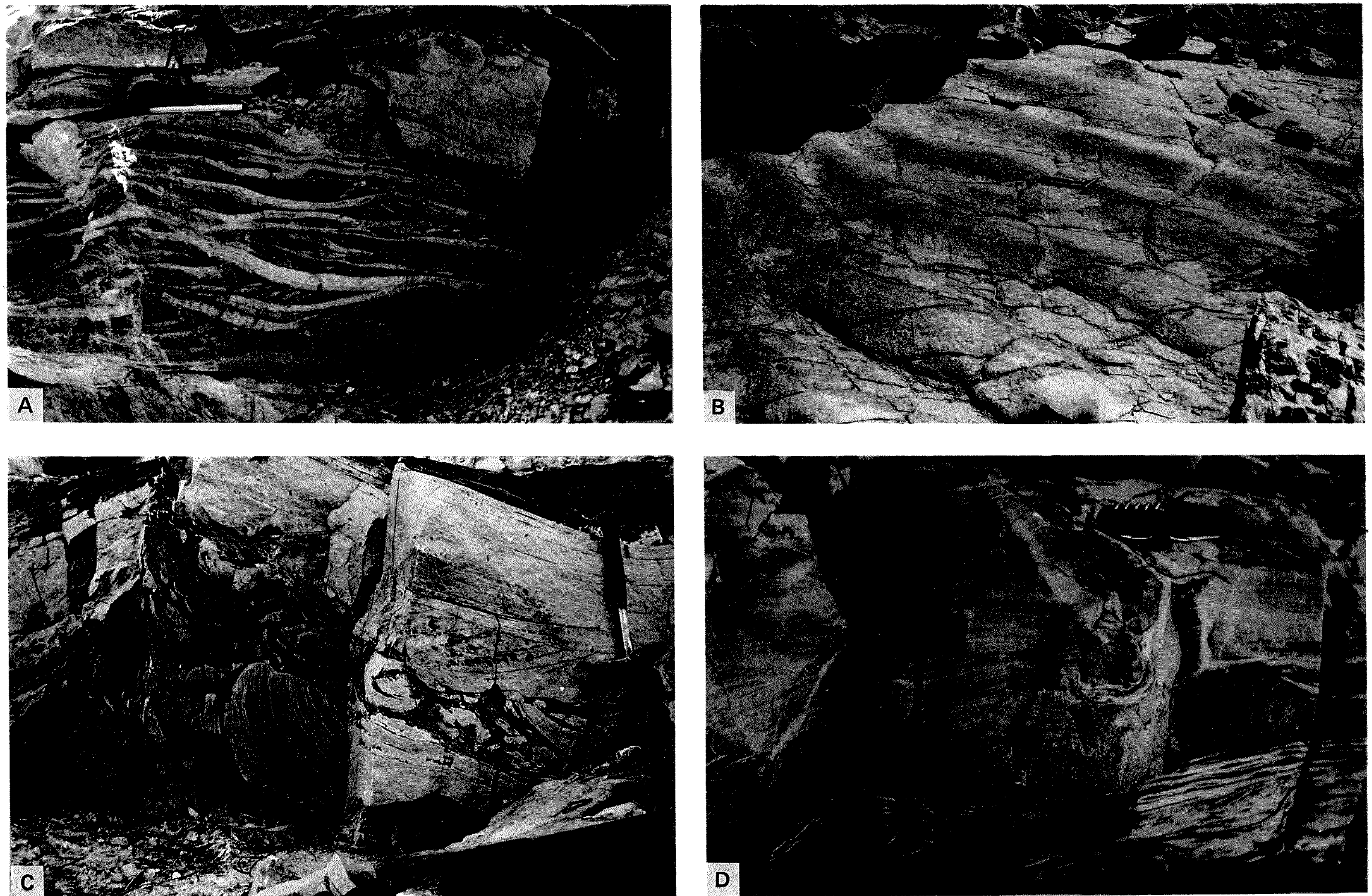
Figure 34. Bedding structures in Princess Ranges Quartzite. Lower part of type section.

A — Flaser bedding in siltstone and fine-grained sandstone.

B — Large-scale symmetrical ripples in quartz arenite.

C — Hydroplastic deformation structures. Note disrupted cross-bed to left of hammer.

D — Sandstone dyke (below penknife) injected upwards into siltstone from thin sandstone bed. Note lensoid laminae (incipient flaser bedding) in siltstone below sandstone bed.



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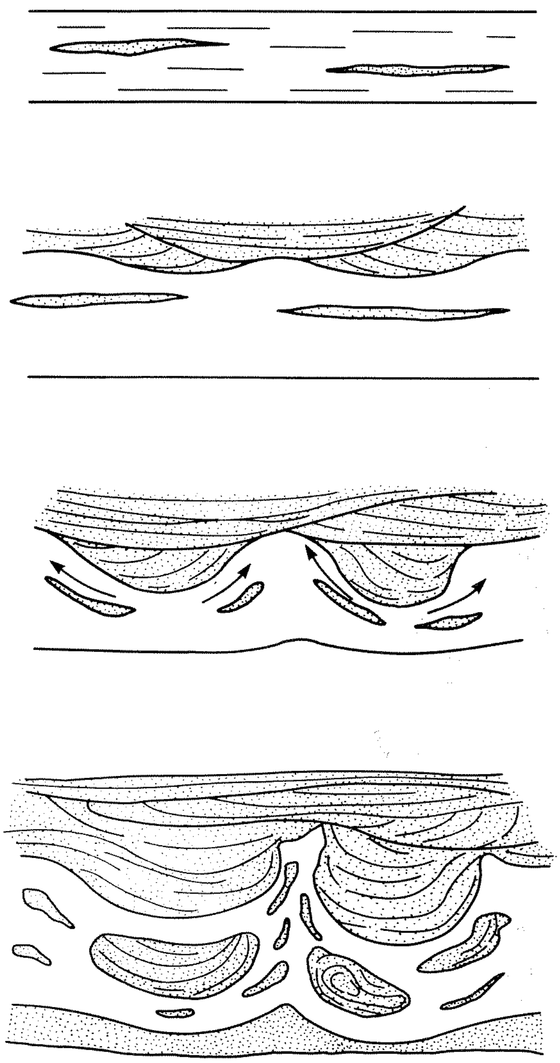

1 Rapid deposition of a sandy silt layer with thin sand lenses

2 Erosion of small scours and deposition of mature, trough-cross-bedded sand. Underlying sandy silt remains wet and hydroplastic

3 Further deposition of cross-bedded sand causes foundering of troughs under load, breaking up the underlying hydroplastic layer and forcing it into low pressure areas between troughs

4 Further foundering causes separation of some lobes and continued fragmentation of sand lenses in hydroplastic layer. Material flow upwards between the downward bulging lobes, and may penetrate the overlying beds. Continued deposition of non-cross-bedded sand

5 A final stage, not normally reached, is the complete reworking of the hydroplastic layer and sand lobe fragments to form a more-or-less homogeneous silt and sand layer, with a few remnant fragments of balls and pillows

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Figure 35. Sequence of formation of hydroplastic load-deformation structures, Princess Ranges Quartzite.

(d) Lack of overall asymmetry precludes a derivation by slumping down the palaeoslope.

(e) The direction of elongation of the detached and attached lobes (6 readings) approximately coincides with the axes of the trough cross-beds, both in the overlying layer and in the formation generally.

With these constraints in mind, an origin by vertical load deformation of trough crossbeds is proposed. The sequential development of the structure is illustrated in Figure 35. The structure has affinities with both the "ball-and-pillow" structure, which commonly develops at sand/mud interfaces, and load-casted ripples.

Small-scale load-deformation structures occur in some of the laminated siltstones and fine-grained sandstones. Small sandstone pseudonodules have become detached, while load-casting of sand on mud has produced small flame structures and, at one locality, a small sandstone dyke (Fig. 34D).

Sole markings: The most common sole structure is a wrinkling on the undersurface of some quartz arenite beds. It may show random orientation but in many

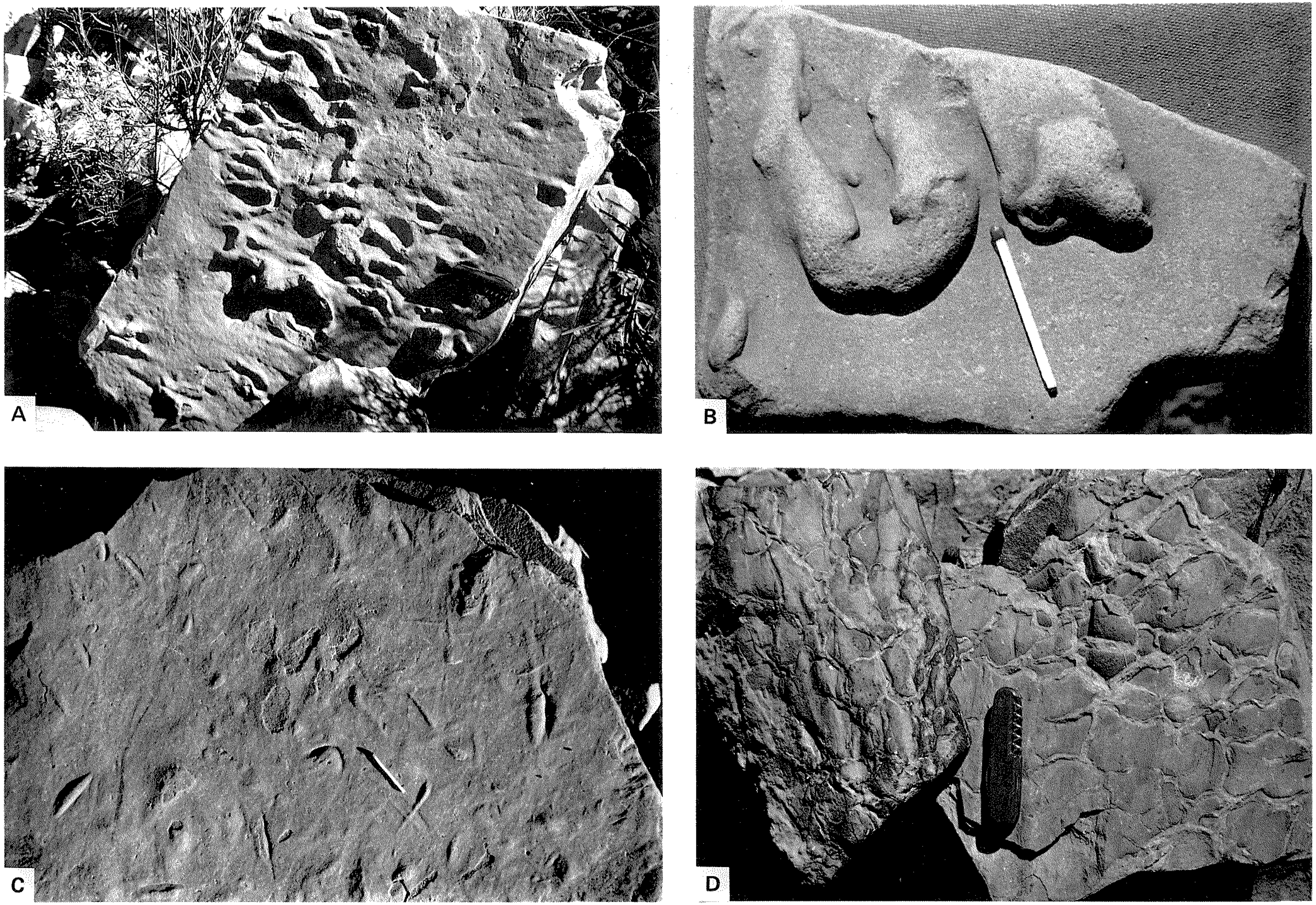
Figure 36. Sole structures and desiccation cracks, lower part of Princess Ranges Quartzite type section.

A — Wrinkled sole structures in quartz arenite.

B — Horseshoe-shaped sole structures (casts of current crescents?)

C — Skip or bounce casts on the underside of bedding plane.

D — Desiccation (mud) cracks.



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examples there is a pronounced linearity, in which cases the wrinkles usually have one side consistently steeper than the other (Figure 36A). The origin of these structures is not clear. Some are horseshoe-shaped (Fig. 36B) suggesting a primary current origin, perhaps as casts of current crescents formed by turbulence around small obstructions. Others are probably related to flute moulds. All are modified to varying degrees by load deformation.

Another type of sole marking is the skip or bounce cast.These occur as small ridges 1-2 mm high, and are the casts of marks made by objects such as small pebbles or intraclasts impinging on the sediment surface as they are carried along by the current. In Figure 36C, these ridges are oriented in several directions, indicating variable currents. Crescentic marks, near the centre of Fig. 36C, could be caused by curved flakes from desiccation cracks striking the sediments.

These various sole markings are rarely observed *in situ* and hence the significance of the directional fabric is not known.

Parting and current lineation: These features commonly occur together in the thinly laminated sandstone. Parting lineation is apparent as a linearity in the edges of successive laminae as they break off in the bedding plane. The current lineation is a faint striation on the surface of the laminae. Current and parting lineations on any one lamina both give the same current direction, however, on successive laminae, the orientation on the lineations may vary.

Parting and current lineations are diagnostic of the upper flow regime (Reineck and Singh, 1973), and in the Princess Ranges Quartzite they are not found in immediate association with ripple marks which form in a lower flow regime. The lineations are found on planar-bedded surfaces and on the foresets of some cross-beds.

Desiccation cracks: Desiccation cracks (Fig. 36D) are rare, but are important as evidence of brief periods of emergence when the surface sediment could dry out. The cracks occur in fine-grained silty sandstone and are filled with quartz arenite. The width of the cracks precludes an origin by subaqueous syneresis (Reineck and Singh, 1973, p. 51).

Intraclasts: Tabular intraclasts of shale and sandstone are abundant in some beds of quartz arenite. They are up to 100 mm across and are usually aligned parallel to bedding.They commonly weather out, leaving irregular angular holes in the rock. The intraclasts are formed by the breaking up of recently deposited (and partly consolidated) mud and sand. Mud flakes produced by desiccation cracking may contribute to the intraclasts, but most are probably produced by scouring and strong-current activity during the deposition of cross-bedded units.

Ex-carbonate rhombs: In many of the quartz-arenite beds, brown spots a few centimetres across are conspicuous. This is the "dalmation texture" of Hall and Goode (1978). The spots are rhomb-shaped, with rounded corners, and contain quartz arenite of the same grain size as the host rock but with ferruginous rather than siliceous cement. The shape of the spots suggests that they were originally diagenetic crystals of carbonate cement, which have been weathered and replaced by iron oxide. They commonly occur in a random scatter, and cut the primary-bedding lamination (Fig. 32). Normally 20 to 30 mm across, they can be up to 80 mm and can constitute up to 30% of the rock.

Pyrite moulds: A few rare beds of sandstone contain cubic holes up to 20 mm across. The sides of the holes have parallel striations, and therefore probably represent weathered pyrite cubes.

DEPOSITIONAL ENVIRONMENT

Sedimentary structures within the Princess Ranges Quartzite indicate a shallow-marine environment. A tidal influence is apparent from such structures as herringbone cross-bedding, flaser and wavy bedding. In the fine-grained sandstone and siltstone units, the abundance and variety of asymmetrical ripples indicate currents which were continuously changing in both direction and intensity. Flaser and wavy bedding indicate alternating suspension and bedload deposition, and reflect the low-velocity (slack water) and high-velocity (ebb and flood) currents of tidal cycles. Rare periods of emergence are indicated by mudcracks. Most of the sedimentation was probably located in the shallow subtidal zone occasionally reaching the lower intertidal zone.

Cross-beds in a tidal or subtidal domain can be due to migrating ebb-tide or flood-tide megaripples and sand waves, or point-bar deposits in meandering tidal creeks. Button and Vos (1977) record, in the subtidal zone, tabular cross-bed sets caused by migrating flood-tide sand waves, diametrically opposed to trough cross-beds caused by migrating ebb-tide megaripples. In the Princess Ranges Quartzite, there are insufficient data to comment on the significance of opposing orientations; however, both planar (tabular) and trough cross-beds are present.

Mud and sand intraclasts occur both on erosional surfaces of tabular cross-beds and in sand-filled scour structures. These intraclasts may represent channel-lag deposits in tidal creeks. The scours show a comparable current direction to trough-style cross-beds (see next section); therefore the trough cross-beds are more likely to represent an ebb-dominated regime, as described by Button and Vos (1977), rather than a flood-dominated regime.

Ball-and-pillow structures, pseudonodules, and other hydroplastic, soft-sediment, deformational structures, have been described from recent tidal-flat channels by Klein (1977), and Proterozoic tidal deposits by Button and Vos (1977). These structures imply rapid sedimentation to produce a water-saturated, hydroplastic layer which can deform, by liquefaction, under the load of the overlying layer. In the fine-grained sandstone, a rapid and continuous supply of sediment is implied by the presence of rare climbing ripples-in-drift, a phenomenon which is characteristic of fluviatile and floodplain environments (McKee, 1966), but which a number of workers have reported from tidal-channel and deltaic deposits.

The symmetrical megaripples (wavelengths of between 0.6 and 1.5 m) are considered to be typical of the subtidal, shallow, marine-shelf environment, in which ebb and flood currents are modified by the oscillatory action of waves. It is difficult to imagine such large, symmetrical ripples forming anywhere in the intertidal zones.

In summary, therefore, the Princess Ranges Quartzite formed under the influence of tidal currents, in the shallow subtidal and lower intertidal domains. Subtidal sand bodies occurred offshore from a mixture of intertidal flats and tidal channels. The clean mature nature of the arenite units is typical of a shallow-marine environment where there is constant reworking of the sediment. Typical features of deltas (coarsening-upwards sequences, large gravity-induced slumps and slides) are absent from the type area. However, in the northeastern part of the rim syncline, around the Teague Ring Structure, lensoid beds of quartz arenite in the sandy siltstone, accompanied by small slumps, may represent a minor delta complex.

The environment just described marks a regressive sequence, in which a prograding shoreline was advancing over the deeper water shales and muds of the top of the Wandiwarra Formation. This regressive phase, as mentioned in the section dealing with the Wandiwarra Formation, began earlier in the southeast part of the basin, indicating a coastline somewhere to the east or south. This will be discussed in the next section, on palaeocurrent data.

Finally, it should be noted that the palaeoenvironmental interpretation of the Princess Ranges Quartzite is based entirely on the more arenaceous members, three of which occur in the type section, and only the lowest is well exposed. The two intervening units are not exposed, but are presumed to be softer argillaceous sediments. It is therefore not possible to identify the environments of these two units. Two models are possible:

(a) Argillaceous units mark regressive phases, e.g., a fine-grained lagoonal type of environment; this would imply that the second and third arenaceous units mark short transgressive phases; or

(b) argillaceous units mark a short return to deeper water sedimentation similar to the top of the Wandiwarra Formation.

The two models are illustrated in Figure 37.

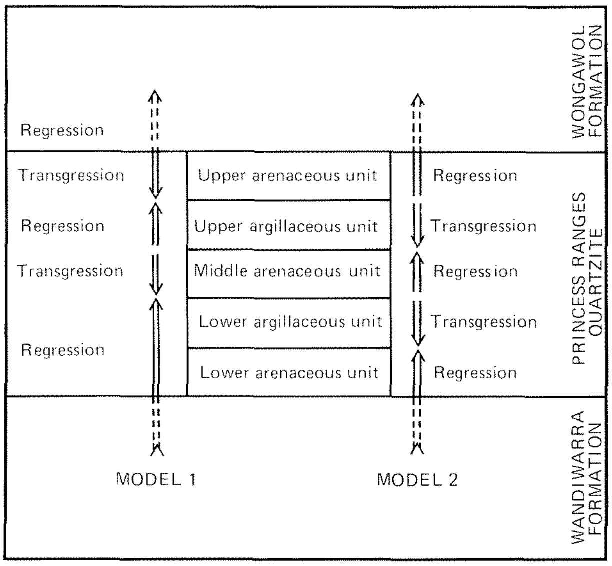

Figure 37. Alternative models for the sedimentational history of the Princess Ranges Quartzite.

PALAEOCURRENT DATA

Palaeocurrent directions were measured in the lower part of the type section, in southwest STANLEY (three localities) and in the syncline of the Teague Ring Structure (two localities). Elsewhere, only a few scattered single measurements could be taken. No data are available from the northern side of the sub-basin because of the strong deformation in the Stanley Fold Belt.

Palaeocurrent recordings indicate either specific or non-specific directions. The specific directional data give a unique palaeocurrent vector, and are provided by such structures as asymmetrical and linguoid ripple marks, cross-bedding, and troughs and scours. The non-specific data, which give only the trend of the palaeocurrents or waves, include symmetrical ripple marks, current lineation, and skip and bounce marks.

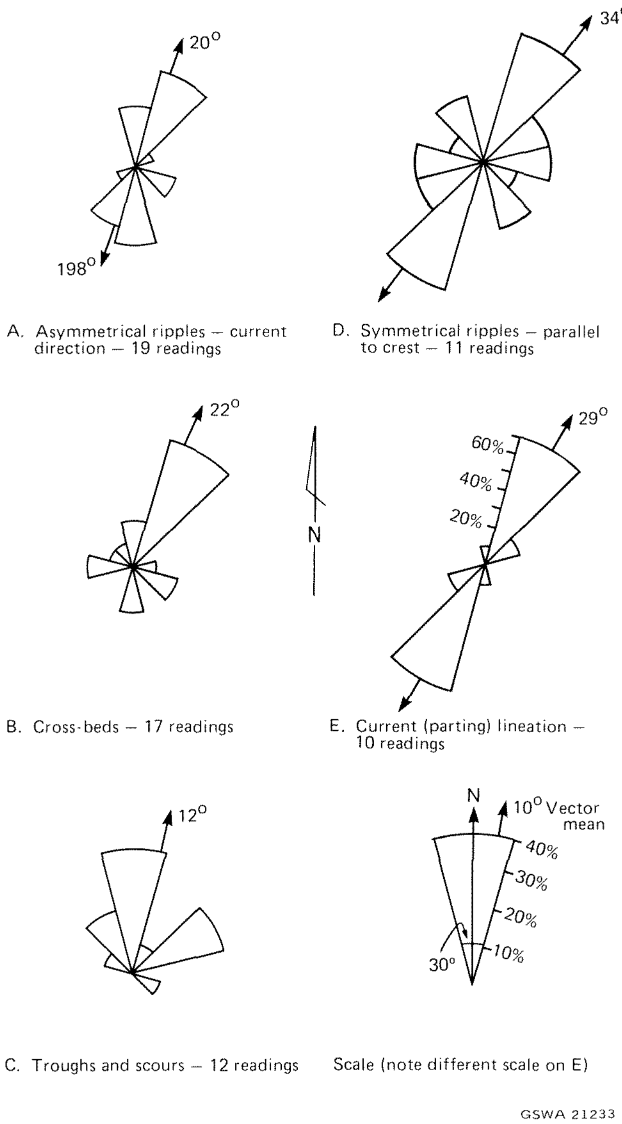

Figure 38. Palaeocurrent measurements, Princess Ranges Quartzite type section.

These categories are illustrated in Figure 38 which shows palaeocurrent data from the type section. The asymmetric ripples show a distinct bimodality (Fig. 38A), a pattern indicative of a tidally influenced environment. Larger structures, such as cross-beds, troughs and scours, show a broadly unimodal pattern, with a mean in a north-northeasterly direction. The scatter in the cross-bed pattern (Fig. 38B) reflects the presence of herring-bone, planar and trough cross-bedding. However, the data show a directional tendency which is in accord with both the asymmetrical ripple marks and the troughs and scours. This north to northeasterly direction can therefore be taken as the dominant current direction in the lower part of the type section.

Trough cross-bed measurements from two other localities in the Princess Ranges Quartzite—in south-western STANLEY and in the rim syncline of the Teague Ring Structure (NABBERU)—also indicate a general north to northeasterly palaeocurrent direction (Fig. 39).

The non-specific measurements from the type section are less reliable than the specific data. Ten measurements of current lineation are in broad agreement with the north-northeasterly current direction obtained from the specific current data (Fig. 38E). However, the symmetrical ripples (Fig. 38D)show a somewhat random pattern, with a slight preference for the crests to align north-northeast to south-southwest. The predominant direction of wave propagation would have been perpendicular to this.

The interpretation of the palaeocurrent data in terms of palaeoslope (and hence orientation and direction of shoreline) is beset with difficulties, due in no small way to the variety of current regimes in modern shallow tidal seas. Selley (1976) comments that whereas palaeocurrents are slope-controlled in fluvial, deltaic and most turbidite environments, they are not related to slope in marine-shoreline environments. In the latter, the current patterns can be consistently oriented either onshore, offshore, or longshore. Nevertheless, when combined with data from other sources, the palaeocurrent patterns can yield some indication of the palaeogeography.

The depositional environment in the area of the type section, as discussed in a previous section, is shallow marine, probably in the lower intertidal or

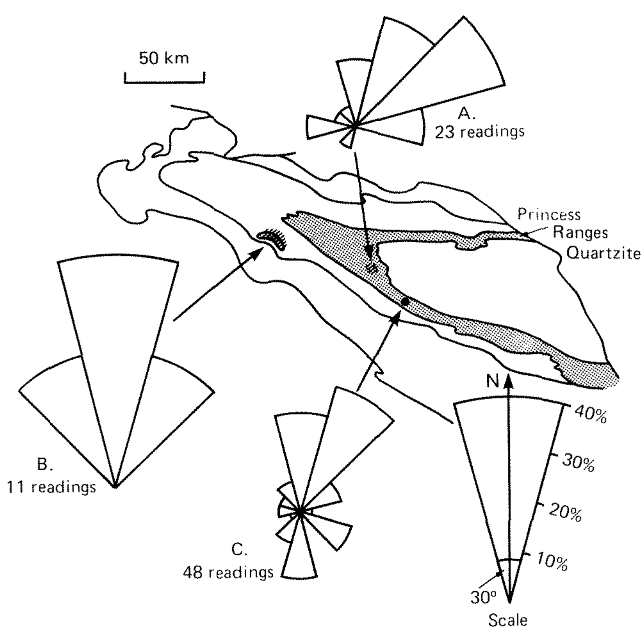

Figure 39. Palaeocurrent measurements, Princess Ranges Quartzite.

A — Southwest STANLEY.

B — Rim syncline, NABBERU.

C — Type section KINGSTON (TOTAL OF A, B and C of Figure 38).

shallow subtidal zones of a regressive barrier coastline. The regressive sequence began in the southeastern part of the basin, and moved northwesterly. Thus the coastline was somewhere in an arc, between due south and due east of the present Princess Ranges Quartzite exposure. Klein (1967, 1977) considers that in subtidal and some intertidal sand bodies, the bimodal or unimodal palaeocurrent pattern is oriented parallel to the depositional strike. Applied to the Princess Ranges Quartzite, the north-northeasterly trends of the asymmetrical ripples and cross-beds could represent the depositional strike, which would be consistent with a coastline somewhere to the east-southeast. This is also consistent with the orientation of symmetrical (oscillation) ripples, which in both modern and ancient deposits, have been shown to lie with crests approximately parallel to the shoreline (Potter and Pettijohn, 1963). A poorly developed 90° bimodal pattern is apparent in the trough and scour marks (Fig. 38C), with one mode approximately north-northwest and another mode northeast. Such a bimodal pattern is described by Swett and others (1971) from a similar tidal sandstone (the Cambrian Eriboll Sandstone of northwest Scotland). They suggest that the trend, which is at right angles to the main shoreline, may represent the filling of tidal-surge channels. In the Eriboll Sandstone, as in the Princess Ranges Quartzite, the secondary mode is directed away from the interpreted shoreline, and into the deeper water part of the basin.

The above palaeocurrent interpretation is highly speculative, and based on too few data to be definitive. It is consistent with the palaeogeography implications of the regressive shoreline proposed for the Wandiwarra Formation-Princess Ranges Quartzite transition, and also with the implied direction of marine transgression at the base of the Wandiwarra. However, such an interpretation places the depositional strike at a considerable angle to the present regional strike of the rocks, and has important implications for the age of the Wiluna Arch and the palaeogeography of the area to the east and southeast of the Nabberu Basin. These will be discussed in a later chapter.

An alternative and less preferred interpretation follows the tidal model of Button and Vos (1977), in which the currents are predominantly offshore and produced by a combination of ebbtides and palaeoslope runoff. Alternating ebb and flood currents predominate in the subtidal (non-emergent) facies. Applied to the Princess Ranges Quartzite, the cross-bed and scour roses of Figure 38 would have their modes pointing offshore, and the bimodal current ripples would be parallel to the ebb and flood currents and also the palaeoslope. The shoreline would therefore be to the south or southwest. While this is consistent with the present margin of the sub-basin, and with the Yilgarn Block as a source area, it is inconsistent with the direction of wave propagation as indicated by the symmetrical ripples. Furthermore, if the regressive model for the base of the Princess Ranges Quartzite is correct, then a shoreline to the southwest is inconsistent, but a shoreline to the south is feasible.

In summary, the palaeocurrent data are inconclusive, but do suggest, in conjuction with other lines of evidence, that the shoreline was somewhere in an arc between south and east of present exposure.

WONGAWOL FORMATION

DEFINITION

Derivation of name: Named after Wongawol homestead on KINGSTON, Lat. 26°07′30″S, Long. 121°56′30″E.

Type section: The type section is along the Wiluna-Carnegie Road, between Lat. 26°08′00″S, Long. 121°56′40″E (1 km south of Wongawol homestead) and Lat. 25°56′10″S, Long. 112°03′10″E (1 km east-southeast of Thurraguddy Bore on STANLEY).

The Wongawol Formation forms extensive low hills of patchy exposure and abundant scree; consequently, nowhere is a complete section well exposed. The type section illustrates the main rock types, and, being along the main road, is the only accessible section which crosses the whole thickness of the formation.

Lithology: The Wongawol Formation consists of arkosic and lithic sandstone, shale, mudstone and carbonate, and marks a slow, upwards-fining transition from the mature clastic sedimentation of the Princess Ranges Quartzite to carbonate sedimentation in the Kulele Limestone.

Distribution and thickness: Except for a small outcrop which occurs on the northeast edge of the Teague Ring Structure on NABBERU, in the core of the rim syncline, the Wongawol Formation is restricted to the southern part of STANLEY and northern half of KINGSTON. The large area of Wongawol Formation shown by Hall and Goode (1978) in central and eastern Nabberu is now considered to be in the Wandiwarra Formation, because of a closure of the main basin synclinorium in Princess Ranges Quartzite along the western edge of STANLEY.

Total thickness is probably about 1 500 m, of which about half comprises the Sholl Creek Member. An exact thickness cannot be calculated because of shallow dips, abundant gentle folds and numerous suspected faults (determined from air-photo interpretation) of unknown magnitude.

Boundary criteria: The base of the Wongawol Formation is taken at the top of the last continuous, mature, white quartz arenite of the Princess Ranges Quartzite. This contact is well exposed south of Wongawol homestead where the basal Wongawol Formation consists of interbedded fine-grained and coarse-grained immature sandstones, in part glauconitic. Near the base, this sequence contains thin lenses of a more mature arenite similar to the underlying Princess Ranges Quartzite; thus the contact is conformable and gradational over a thickness of 30 m.

The upper boundary of the formation is also gradational and conformable. It marks an increase in the proportion of carbonate bands, and is taken at the base of a continuous band of cross-bedded calcarenite and stromatolitic limestone which is exposed 2 km east of Thurraguddy Bore.

Previous nomenclature: The original nomenclature divided the Wongawol Formation into two formations: the Wongawol Sandstone and the overlying Sholl Creek Formation (Hall and Goode, 1975). These were combined and renamed the Wongawol Formation by Hall and others (1977), with the Sholl Creek unit being retained as a member thereof.

LITHOLOGICAL DESCRIPTION

The Wongawol Formation is a sequence of mineralogically immature sandstone and shale which shows an overall upwards fining, and contains thin carbonate beds in the upper part. The lower part of the formation is a monotonous sequence of grey-green to pink-brown, sub-feldspathic, micaceous sandstone. Grain sizes range from medium to very fine sand.

The sandstone generally has up to 10% clay and hematite matrix. Quartz constitutes between 30 and 50%, is moderately sorted and subangular to rounded. Feldspar content varies from a trace to about 20%, and in grain size and texture is similar to quartz. Plagioclase is dominant over alkali feldspar, although alteration of the latter makes it difficult to identify the true percentages. Small, thin flakes of detrital muscovite are common, and define a weak bedding fissility. Green chlorite with anomalous purple birefringence is ubiquitous. It occurs both as an alteration product of muscovite and as a detrital mineral, and is the main cause of the greenish tinge in much of the formation. Glauconite pellets are present in some sandstone beds. Calcite occurs both as a cement and as small, rounded pellets or intraclasts possibly replacing glauconite. The calcite cement occurs either as small discrete patches, or more commonly, as interlocking, poikilotopic grains which impart a lustre mottling to the rock. Accessory detrital minerals include tourmaline, zircon, and iron oxide. Iron oxide occurs either as cement or as replacement of other minerals.

Two unusual rocks deserve special mention. About one kilometre southwest of the point where Wannabooline Creek crosses the Wiluna-Carnegie Road, a band of ferruginous and manganiferous breccia is interbedded with shale. The breccia contains shale and chert intraclasts which are set in a matrix of quartz, glauconite, carbonate, rare plagioclase, and iron oxides. Adjacent to the larger intraclasts (10 to 30 mm across), glauconite and carbonate pellets have been deformed and moulded against each other and against quartz grains (Fig. 40), in a compaction texture similar to the "accommodation shards" described by Dimroth and Chauvel (1973). A similar texture has been described from the Frere and Wandiwarra Formations in this bulletin (see Fig. 28). Much of the original glauconite and carbonate have been replaced by coarsely crystalline calcite.

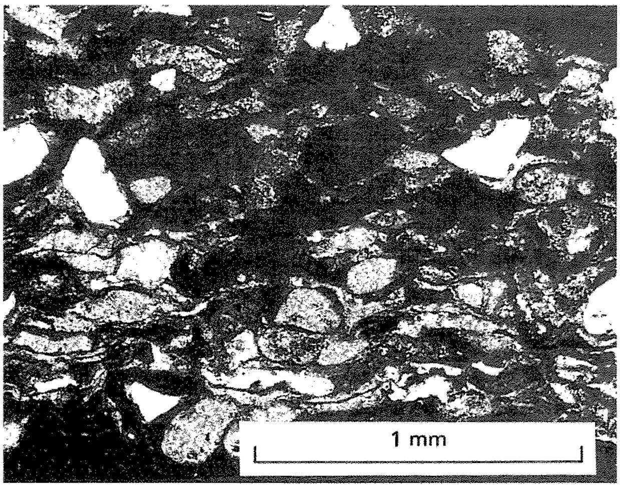

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Figure 40. Deformed glauconite pellets (mottled), and subangular quartz grains, in a ferruginous cement. Lower part of photomicrograph is part of a shale intraclast. Wongawol Formation, 5 km north of Wongawol homestead. Compare with Figure 28. GSWA thin section 46403.

The other unusual rock is a medium-grained sandstone from an outcrop 11.5 km northwest of Mount Hosken on STANLEY. It consists of hematite (about 40%), albite (25%), yellow phyllosilicate (20%), and the remainder is detrital quartz, glauconite, chlorite, and muscovite. Hematite occurs as more or less ovoid masses of fine-grained material, suggesting a peloidal or rounded detrital origin, and also as a replacement mineral in the matrix and along cleavage planes in the albite. Most albite grains contain between 20 and 40% hematite, giving a striped effect. The albite is subhedral to euhedral. The yellow phyllosilicate is probably stilpnomelane, and it occurs as irregular masses of very fine-grained radiating plates. The origin of this rock is not clear, but it may originally have contained a more calcic detrital plagioclase which, on being partly replaced by hematite, expelled calcium and redistributed sodium to form albite. During this process, authigenic albite formed at the margins of the grains, giving the euhedral shape.

To the north and west of the type section, the lower, sandy facies of the Wongawol Formation becomes finer grained, and the predominant lithology is pink-to-brown shaly siltstone, interbedded with thin, fine-grained feldspathic and glauconitic sandstone. The siltstone can be laminated or non-laminated, but is always micaceous and kaolinitic.

The upper part of Wongawol Formation shows a gradual upwards increase in the ratio of shale to sandstone, accompanied by the appearance of thin carbonate bands. This highly variable sequence was termed the Sholl Creek Formation by Hall and Goode (1975), but Hall and others (1977) reduced it to a member. The defining characteristic of the Sholl Creek Member is the presence of the carbonate bands, but as these make up only a small part of the total thickness and are seldom laterally extensive, it is difficult to map the exact boundary of the unit.

The type section of the Sholl Creek Member is along Sholl Creek between Lat. 25°51'30"S, Long. 122°10'00"E, and Lat. 25°55'00"S, Long. 122°12'00"E. The section has not been measured in detail, and this would be difficult due to the lateral variation in rock type and the repetition of strata due to folding and faulting. However, the typical rock types are well exposed.

The sandstone of the Sholl Creek Member is fine grained and brown, with a faint pink or green tinge. Muscovite and chlorite are detrital and show an alignment of flakes which defines a weak bedding fissility. Quartz is the most abundant mineral, and feldspar is present in amounts ranging from a trace to about 10%. Both quartz and feldspar are subangular to subrounded, and have an average grain size of between 0.06 and 0.2 mm. In one sample, from 5 km southwest of Thurraguddy Bore, some of the quartz shows pyramidal terminations and some resorption features, suggesting it may be volcanogenic. This rock also contains about 40% calcite cement, giving it a lustre-mottling. In the Sholl Creek area, some muscovite flakes are surrounded by secondary quartz growing perpendicularly to the flakes. The muscovite is partly altered to chlorite which also protrudes between some of the secondary quartz crystals. The quartz probably forms in tensional pressure shadows during the late stages of diagenesis or very early stages of deformation.

Other constituents in the sandstone include glauconite, iron oxide (both as hematite dusting in the matrix and as detrital grains of magnetite or martite), clay pellets, and lithic grains (mainly chlorite or muscovite schist). Most of these constituents are generally less than 5%. Clay matrix, which may be iron rich, is always under 10%. Cement can be siliceous, ferruginous, or calcareous, and manganese staining is apparent.

The shale is generally well laminated and coloured cream, maroon, brown, purple, and occasionally green; these are the weathering colours. The shale is micaceous, and grades into sandstone.

The carbonates which characterize the Sholl Creek Member are rarely more than 1 m thick. They are pink to grey, and each band may consist of oolitic limestone, intraclastic carbonate conglomerate, silty micritic limestone, and stromatolitic limestone. Ooliths are present in some of the conglomerate matrix. The intraclasts in the conglomerate are mainly pink-to-brown micritic carbonate, but a few oolitic intraclasts and laminated fragments occur. Small, poorly developed stromatolite domes occur within laminated micrite.

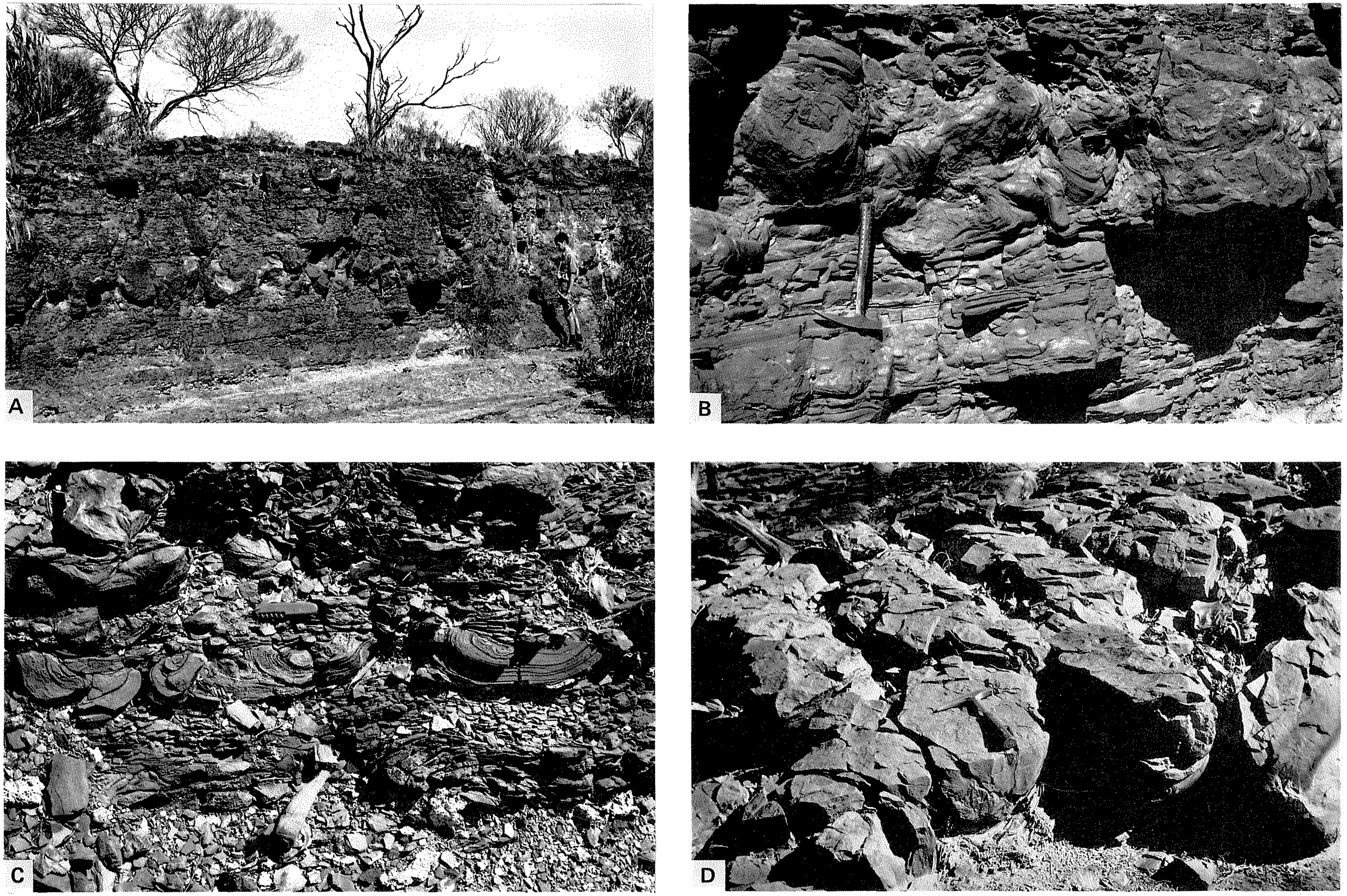
Sandstone and siltstone are about equal in abundance, whereas the carbonate beds represent only a small proportion, perhaps 5%, of the member.

SEDIMENTARY STRUCTURES

The Wongawol Formation abounds in small-scale sedimentary structures. However, the principal feature of the sandstone-dominated lower part of the formation is its monotonous, flaggy bedding, devoid of cross-bedding.

In general, the finer grained sandstone situated away from the monotonous zone contains faint ripple cross-lamination plus irregular and sinuous-crested asymmetrical ripples on the bedding surface. Small washout structures and siltstone intraclasts are rare, but indicate weak bottom currents. Microripples, with wavelengths up to 20 mm, occur in some flaggy siltstones. Less common are wrinkled bedding surfaces, the wrinkles are arcuate, and suggest an origin due to drag on a muddy surface, either due to gravity or wind. They are similar to wrinkle marks (Reineck and Singh, 1973, p. 56) which can be formed by wind blowing over a partly cohesive mud, covered by a thin (< 10 mm) film of water. Both the microripples and the wrinkles suggest that the sediments were deposited in very shallow, still water.

A common feature throughout the formation is the development of hydroplastic deformation structures (Fig. 41). These occur in the fine-grained sandstones, and are restricted to particular beds. In the



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shaly parts of the formation, most of the thin sandy interbeds (50-100 mm thick) contain load casts and ball-and-pillow structures. In the thick sandstone units, beds up to 1 m thick may be completely disrupted over distances of several hundred metres (Fig. 41, A and B).

The origin of these structures is similar to that of the ball-and-pillow structures in the Princess Ranges Quartzite *i.e.* hydroplastic deformation caused by density contrasts within the strata. The structures are gravity controlled, and movement is essentially vertical. Unlike the Princess Ranges Quartzite, there is no connection with cross-bedding in the Wongawol structures. A factor in the Wongawol sandstones may have been the presence of detrital muscovite. The platy alignment of this mineral imparts a reflective sheen to the contorted bedding surfaces, and the abundant mica flakes may have acted as a lubricant once the hydroplastic deformation was in progress.

In Sholl Creek, complex ball-and-pillow structures are elongate in plan (Fig. 41D) and their long axes have a preferred orientation in a 060° direction (Fig. 42). The origin and significance of the elongation are uncertain. Sorauf (1965, p. 562) suggests that orientation of long axes of ball-and-pillows may be inherited from pre-existing channels elongated in the current direction. However, the ball-and-pillow structures at Sholl Creek also show an asymmetry in which the northwest sides have more pronounced bulbous shapes than the southeast sides. This suggests an element of downslope slumping in conjunction with the dominantly vertical movements of load deformation. It also supports the interpretation of a palaeoslope to the northwest.

DEPOSITIONAL ENVIRONMENT

In the underlying Princess Ranges Quartzite, the mineralogical maturity indicates repeated reworking, and therefore the immediate source of sediment was probably either offshore or longshore. By contrast, the mineralogically immature sediment of the Wongawol Formation, with its detrital feldspar, mica, chlorite, and lithic grains, indicates relatively direct and rapid transport from onshore. Some environments which

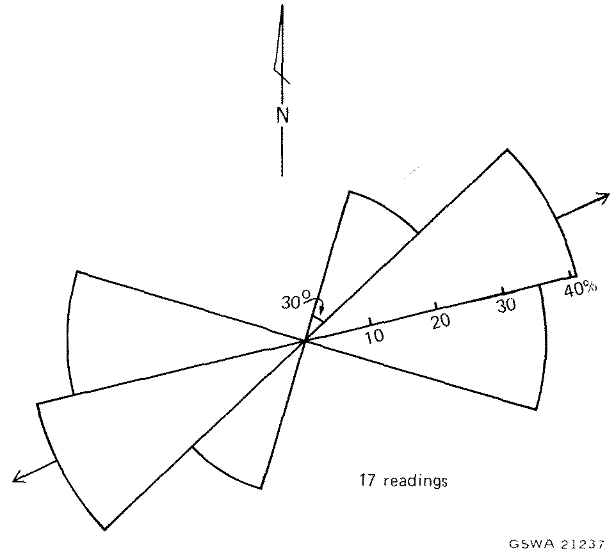

Figure 42. Elongation of ball-and-pillow structures, Wongawol Formation, Sholl Creek.

satisfy this criterion can be ruled out: there are no coarser sediments such as occur in a fluvial sequence; there are no shoestring sands, cross-bedding or slumping characteristic of deltaic deposits; and there are no graded bedding or sole structures typical of turbidite deposits.

Microripples and wrinkles (described above) are evidence of very shallow, still-water deposition. Weak current activity is indicated by the small scour marks and slightly asymmetrical ripples. However, with the exception of some cross-beds near the base of the formation (in the transition from the Princess Ranges Quartzite), there is little evidence of strong currents in the lower half of the formation. A few small thin lenses (up to 0.5 m thick), of intraclastic breccia containing siltstone fragments, may represent localized channel deposits, but these are uncommon. The presence of glauconite in many samples indicates a marine influence.

These sedimentary features indicate a very shallow marine environment in which tidal influence, current activity and turbulence were minimal, but which had a sufficient rate of sediment input to allow deformation of the wet sediments under load. Following the regressive Princess Ranges Quartzite, such an environment could exist in a barred lagoon between a shelf sea and a fluvial-dominated shoreline, or in a very shallow, enclosed or partly enclosed inland sea.

The lateral facies variation, from predominantly sandstone in the south to predominantly shale and siltstone in the northwest, reflects a gradual deepening of the sea in a northwesterly direction, and an associated decrease in detrital input. This is consistent with the palaeoslope deduced from other evidence.

Figure 41. Hydroplastic deformation structures, Wongawol Formation, Sholl Creek.

A — Two disrupted beds (at top of cliff and level with man's chest) in undisturbed micaceous sandstone.

B — Detail of A. Note sheen, caused by alignment of mica on the curved deformation surfaces.

C — Small-scale, ball-and-pillow structures. Note planed tops, in contrast to B.

D — Large, roll-type structures, similar to C, showing elongation (parallel to hammer).

The vertical variation from sandstone to shale and carbonate in the Sholl Creek Member may also indicate an increase in depth of sedimentation. Thus the culmination of the regressive trend is about the middle of the formation. The appearance of algal carbonate beds corresponds to an overall reduction of the sediment influx, but may also be related to a return to more turbulent conditions, perhaps with some tidal scour, indicated by the presence of abundant oolites and intraformational carbonate breccias.

KULELE LIMESTONE

DEFINITION

The Kulele Limestone is the formation of limestone, shale and fine-grained sandstone that lies between the Wongawol Formation and the Mulgarra Sandstone.

Derivation of name: The formation takes its name from Kulele Creek in central southern STANLEY. The name is a modification of the name Kulele Creek Limestone used by Hall and others (1977), but the definition remains essentially the same.

Type section: This is in the vicinity of Mount Throssell, between Lat. 26°01′00″S, Long. 122°39′00″E and Lat. 25°59′30″S, Long. 122°42′00″E. No measured section has been attempted.

Lithology: The main rock types are calcarenite (commonly cross-bedded), stromatolitic limestone, intraclastic carbonate breccia, oolitic and pisolitic limestone, micaceous shale and sandstone, dolomite (rare), and fine-grained quartzose or feldspathic sandstone. These various types are usually interbedded, in cyclic sequences of shale, breccia and limestone.

Distribution and thickness: The formation is about 300 m thick in its type section. It is restricted to the central-eastern edge of the exposed basin in northeast KINGSTON and southern STANLEY, except for one important outlier near Thurraguddy Bore, about 50 km west of the type section.

Boundary criteria: The base is gradational with the Sholl Creek Member of the Wongawol Formation, but it is conveniently taken at the base of a continuous, one metre-thick bed in stromatolitic limestone or cross-bedded calcarenite. The top is taken at the base of the continuous, thick quartz arenite which is the lowest unit of the Mulgarra Sandstone. This upper contact may be disconformable.

LITHOLOGICAL DESCRIPTION

In broad terms, the Kulele Limestone consists of various interbedded limestone types, which form crudely cyclic sequences, separated by thicker units of micaceous shale, mudstone and fine-grained sandstone. The carbonate units are up to 10 m thick, but more commonly are between 1 and 2 m thick. Some shale/sandstone units are up to 50 m thick, although this includes a few thin ($<$ 1 m) carbonate beds.

Chemical analyses of the two most common types of carbonate (pale-grey oolitic, and pink laminated) show them to be true limestones, containing minor amounts of quartz, feldspar, chlorite, and mica. Dolomite beds are rare and difficult to distinguish from the limestone.

The common type is pink, thickly bedded, laminated limestone. The laminae are on a scale of 1 to 5 mm, and are identical to the laminae in domed stromatolites; indeed, in places the laminae in flat-laminated limestone are physically continuous with those in the stromatolites. The limestone itself is a fine-grained mosaic of sparry calcite, possibly with a small micritic component.

Stromatolites are largely confined to this type of limestone. They range from small, flat domes a few centimetres across, through bulbous domes up to 0.5 m across, to large domes over 2 m high and 4 m across which may coalesce into bioherms up to 30 m long. Elongation and asymmetry of the domes is a common feature and is discussed later in this section. Columnar stromatolites are not common, although some of the smaller domes (under 500 mm) have a columnar internal structure. Small slender columns only a few centimetres across occur in a limestone bed in northeast KINGSTON near the top of the formation.

Oolitic and pisolitic limestone is typically pale grey. As pointed out by Hall and Goode (1978), the allochems in this rock are of two types. The large pisoliths are elongate, and have a core which may be an intraclast of either micritic or sparry pink limestone. The zone of concentric laminae surrounding the core may be up to several millimetres thick. The smaller ooliths have both a radial and a concentric zonation. Davies and others (1978) regard such radial textures as indicative of quiet-water rather than agitated conditions. Some pisoliths have cores of radial ooliths. Thus a two-stage origin is required, in which either quiet-water conditions were succeeded by a period of agitation, or ooliths formed in quiet water were transported into a more agitated environment.

Shallow trough-style cross-bedding, and upward-fining graded bedding are features of some oolite-rich layers. In some instances, grading occurs within individual sets of cross-beds.

Breccia layers are highly variable in lithology. Some are pinkish-brown and composed of tabular intraclasts, up to 200 mm long, of micritic and laminated limestone set in a micritic matrix. These are commonly interlayered with beds of domed stromatolites, and may form discrete beds at the base of limestone cycles. The intraclasts may be either flat-lying, randomly orientated, or subvertical (edgewise conglomerate).

Another type of breccia consists of small pink intraclasts (up to 20 mm) set in pale-grey, sparry calcite matrix. This commonly grades into the oolitic type via an intermediate "mixed" rock containing intraclasts, ooliths and pisoliths. The mixed rock is common in the depressions between large domed stromatolites.

The typical calcarenites are pale-grey, fine- to medium-grained rocks of granular calcite and, usually, some detrital quartz. Near the top of the formation the proportion of quartz increases and small amounts of glauconite may be present. Calcarenite layers are rarely more than 1 m thick. Internally they may be massive, but most display a 10 to 20 mm bedding lamination defined by iron oxide-rich layers. Cross-bedding in the latter type is very common (Fig. 43). Most cross-beds are trough style, in which the foresets have a high degree of curvature in both section and plan. However, some cross-beds are subplanar.

The shaly rocks are purple, maroon, grey or greyish-green. A weak fissility is defined by the platy alignment of very fine-grained white mica and chlorite. Rare small-scale ripples and mud cracks are present.

Sandstone, interbedded with the shale, is seldom more than 200 mm thick, except in the upper part of the formation where some layers of quartz sandstone are several metres thick. The thin sandstone beds in the lower part usually occur near the centre of shale units. The sandstone is fine grained and contains abundant feldspar. Calcareous or dolomitic cement is common. Muscovite and chlorite are concentrated in thin laminae a few centimetres apart, and give the rock a fissility. This is particularly apparent where the bedding is folded by small slump rolls and ball-and-pillow structures. This type of sandstone and the enveloping shale form an association similar to the underlying Sholl Creek Member of the Wongawol Formation. The sandstone in the upper part of the formation is thicker, coarser, and contains glauconite. Feldspar is a minor constituent. Sedimentary structures include intraclasts, symmetrical ripples and

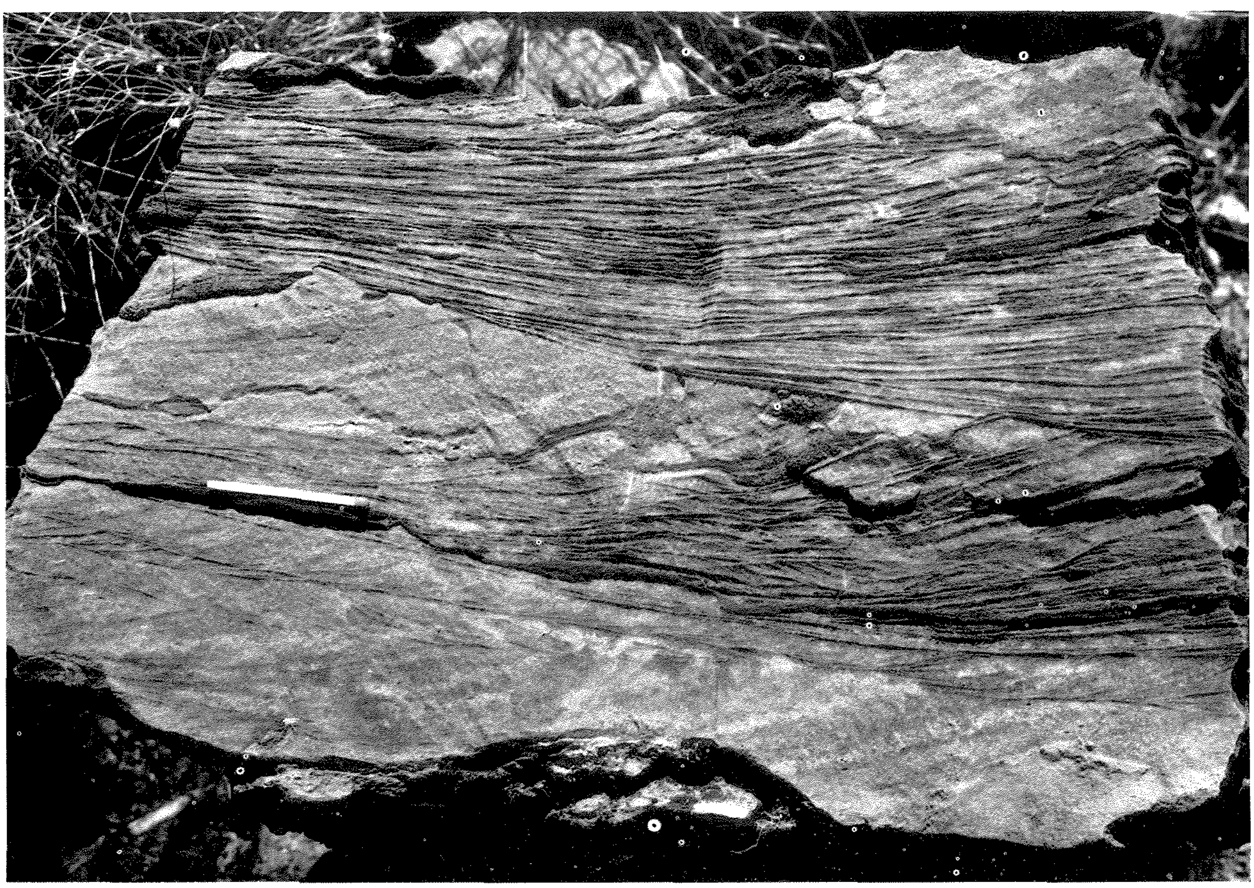

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Figure 43. Cross-bedded calcarenite, Kulele Limestone, Mount Hosken.

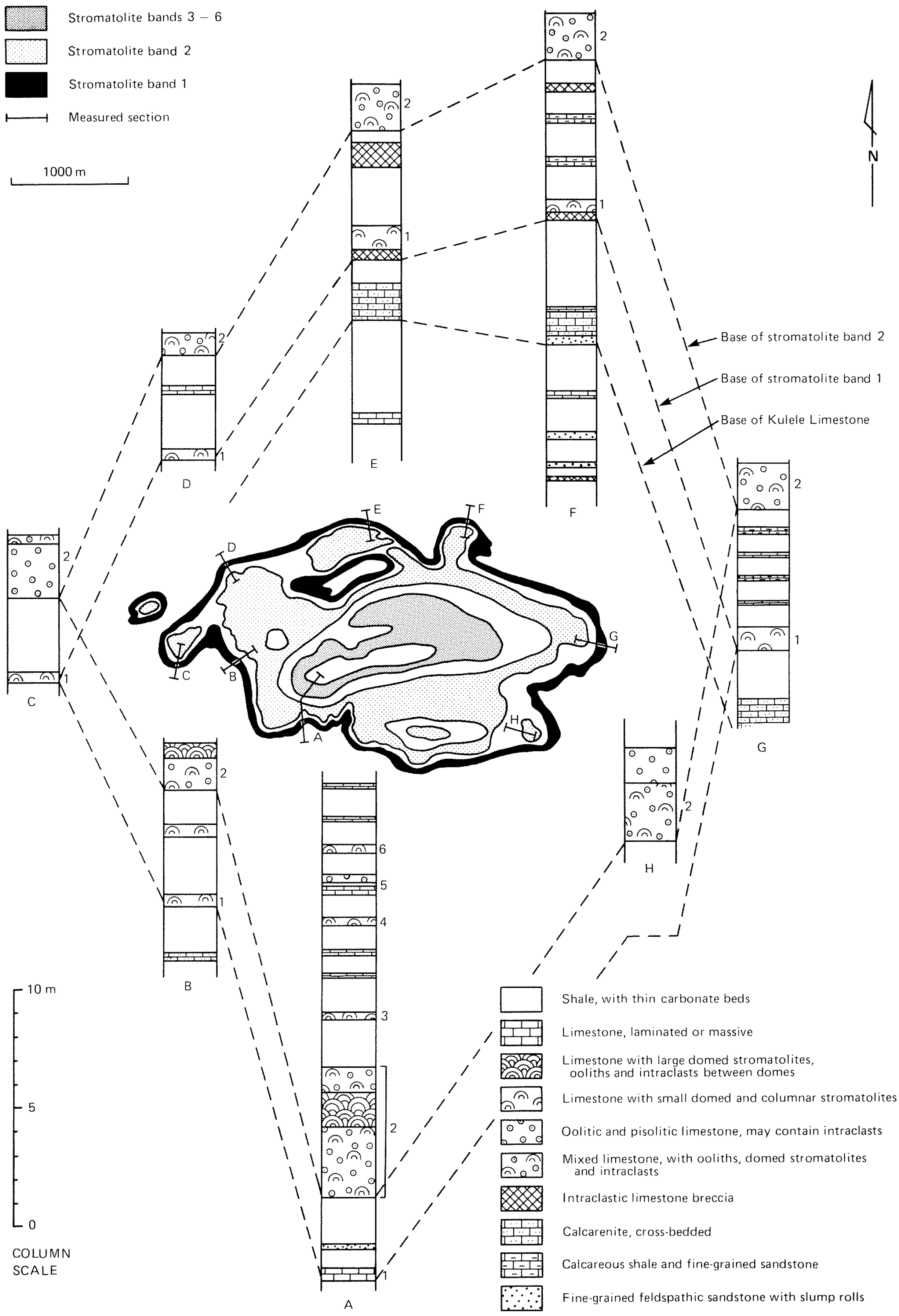

Figure 44. Measured sections, Thurraguddy outlier of the Kulele Limestone.

poorly developed trough-style cross-bedding. The sandstone is similar to that in the overlying Mulgarra Sandstone.

CYCLICITY

Hall and Goode (1978, p.170, Fig.22) described repeated (cyclic) sequences 5 to 10 m thick, and showed a schematic representation of a cycle from the basal Kulele Limestone. Sedimentary cycles undoubtedly exist, but variations in proportion and sequence of lithologies occur. The order in which rock types occur is not constant, and may be reversed, even within the same cycle, for instance with stromatolite band 2 (Fig. 44) compare sections C and H.

Figure 44 shows eight detailed sections from the outlier of Kulele Limestone near Thurraguddy Bore, where the basal 25 m are exposed. The cyclic pairs of shale and carbonate are immediately apparent, and, ignoring some of the thinner limestone bands, parts of seven such cycles are present in the exposed section. Observed cycles vary from 2.5 m to 7.5 m in thickness.

Each cycle represents a transgressive-regressive event. The beginning of a cycle is taken somewhere in the middle of the shale unit, where a thin sandstone may be present. This represents the most regressive phase of the cycle, perhaps in an intertidal mud flat, but more probably (in view of the general absence of tidal structures) in a shallow lagoon. Conditions in this phase of the cycle were very similar to the underlying Sholl Creek Member. The first carbonate band is usually brecciated, and marks an agitated zone in shoreline conditions. In some cases the first carbonate is not brecciated, and contains small flat-domed stromatolites suggesting a more passive environment. Continued deepening of the water resulted in the growth of larger stromatolites, the largest of which have a synoptic relief of over two metres, (suggesting a water depth in excess of this). Evidence from other parts of the world indicates that such large stromatolites grew in the subtidal zone (Truswell and Ericksson, 1975). The passage upwards into laminated, non-stromatolitic limestone, or oolitic and intraclastic limestone, marks a return to shallower conditions.

STROMATOLITES: STRUCTURES, ELONGATION AND ASYMMETRY

In the outlier near Thurraguddy Bore (Figs 44 and 45), resistant limestone units form a series of subhorizontal terraces which allow the stromatolites to be studied both in section and plan. Six stromatolite bands are recognized. Of these, the top four are very thin and they occur in only one of the measured sections. The stromatolites in these upper bands are low domes, circular in plan and up to 1 m across. These stromatolites are seldom more than

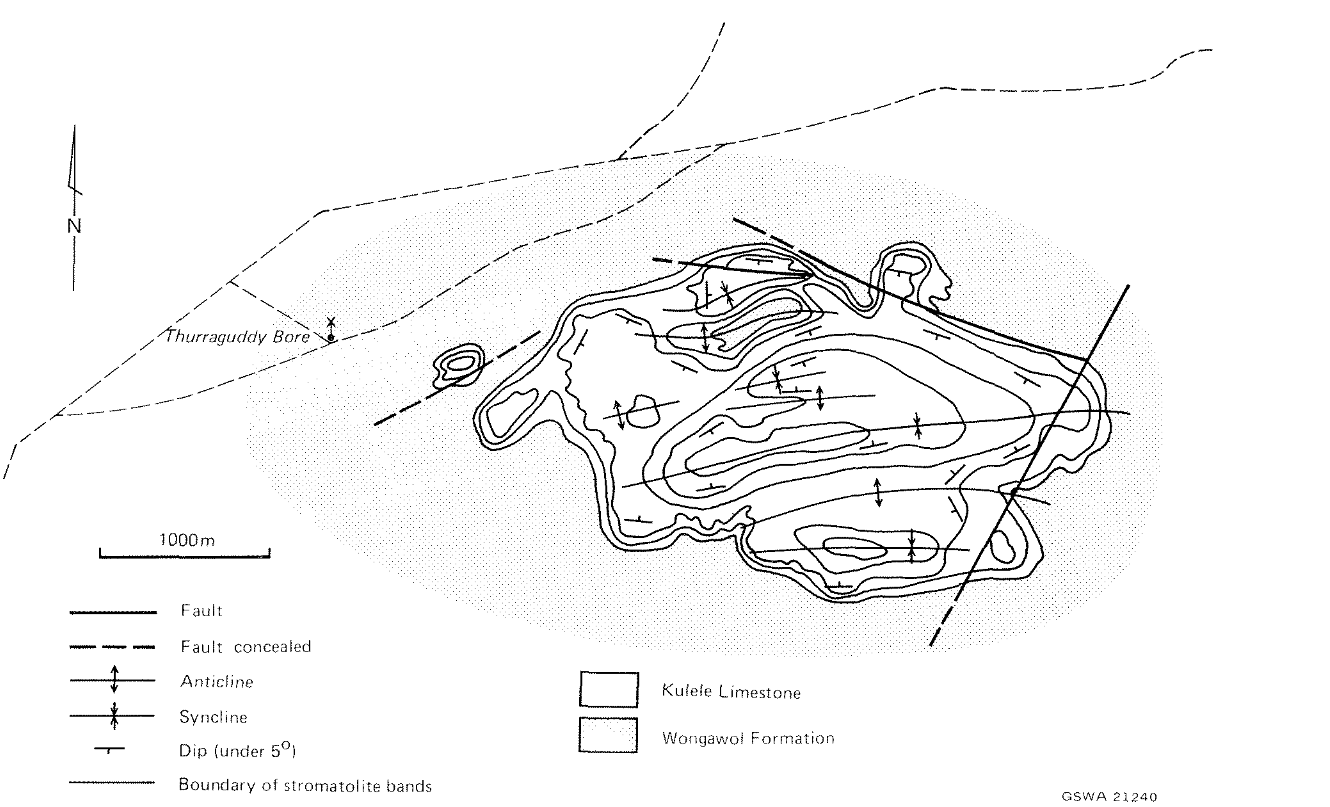

Figure 45. Structure of Kulele Limestone outlier near Thurraguddy Bore.

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Figure 46. Elongate stromatolites, stromatolite band 1, Kulele Limestone, Thurraguddy Bore.

100 mm high. Band 5 is itself composed of two smaller bands 300 mm thick, separated by 800 mm of shale. Bands 3 to 6 will not be discussed further.

Stromatolite band 1: This band is seldom more than 1 m thick and may be as thick as 300 mm. Stromatolites form low domes up to 100 mm high. Many are roughly circular, ranging from 200 to 500 mm across, but many are elongate with length:breadth ratios ranging up to 3:1 (Fig. 46). Orientation of the long axes of these domes is consistently 080 ± 10° (Fig. 47). Rarely, at the tops of some domes, incipient columns (20-30 mm across and a few tens of millimetres high) protrude from the normally smoothly convex laminae.

Asymmetry of the stromatolites in band 1 is not usually apparent, although in section F (Fig. 44), the stromatolites all lean to the north. Between E and F, the stromatolites in band 1 are bulbous rather than domed, and in places approach a columnar form up to 100 mm high and 50-100 mm wide.

Stromatolite band 2: Band 2 contains domed stromatolites up to 2 m high, and is the thickest band in the Thurraguddy outlier, ranging from 2 m to 5.5 m thick. The stromatolites vary from bulbous domes only 300 mm across, through single domes up to 3 m across, to compound domes or bioherms formed by coalescing of smaller domes. The bioherm may be up to 30 m long by 10 m wide. In general, there is a complete gradation in the style: from bulbous, through domed, to compound (Fig. 48) although there are some notable exceptions, such as the large bulbous form (Fig. 49.)

The laminae of the domed stromatolite form simple hemispheroids; only rarely are poorly developed columns (10 to 20 mm high) present on the top of the domes. In some places, particularly 200 m southeast of section E (Fig. 44), the laminae at the sides of domes have been truncated by erosion. In such cases, the eroded material provided the intraclastic debris, mixed with ooliths and pisoliths, in the interdomal areas.

An unusual feature of the laminae is a wavy or crinkled structure. This occurs in stromatolites of all sizes from 300 mm across to the largest domes. The crinkles show up on the surface of the stromatolite as a fine ribbing with a distance between crests of about 20 mm. In vertical section, the crinkling is seen to be formed by asymmetrical slippage planes which fall away at about 30-60° either side of the stromatolite (Fig. 50), effectively forming two series of small, normal faults. In the centre of the dome, interference of the two sets gives a symmetrical crinkle pattern.

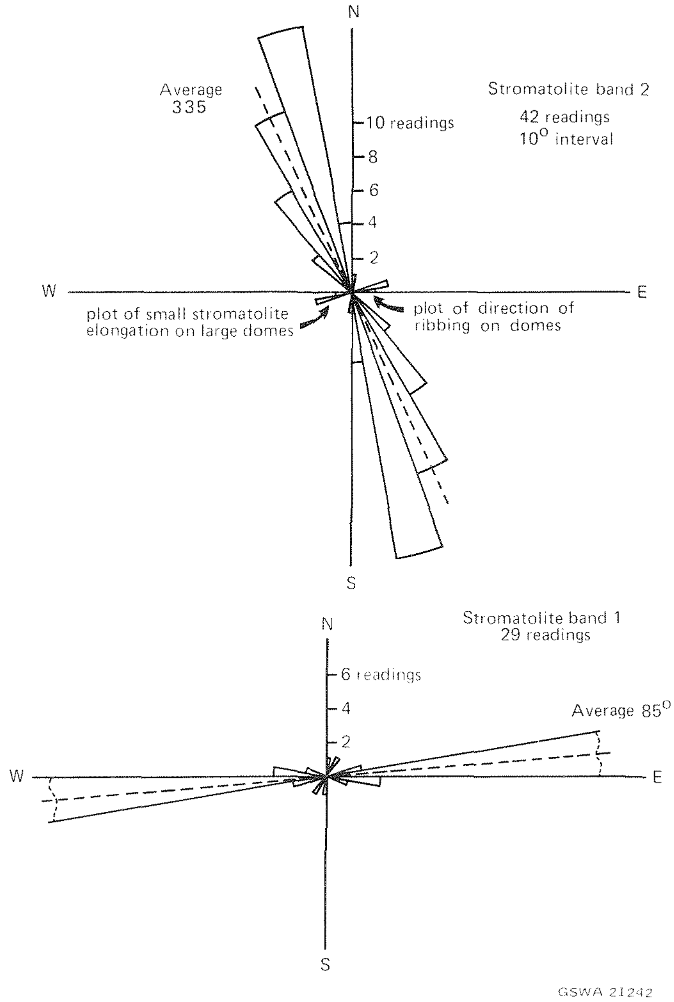

Figure 47. Elongation of stromatolite domes, Kulele Limestone, Thurraguddy Bore.

The ribbing on top of the domes (which is the surface expression of the crinkles) trends 080° ± 20°, parallel to the direction of stromatolite elongation in band 1 and also to a few smaller stromatolites in band 2. The reasons for the preferred trend of the ribbing are not clear, but are discussed in the following section (Origin of stromatolite elongation and asymmetry).

Possible mechanisms for the origin of the crinkles include: ripple marks during growth of the stromatolite; unusual morphological development of stromatolitic laminae; and tectonic deformation. The shallow, opposing dips and lack of tectonic effects in interbedded shales are evidence against a tectonic origin, and the mechanism now favoured is one of gravity sliding in the partly consolidated stromatolite. According to Walter (1976) "stromatolites are organosedimentary structures produced by sediment trapping, binding and/or precipitation as a result of the growth and metabolic activity of micro-organisms, principally cyanophytes." They originate as microbial mats which range in cohesiveness from loose, mucilaginous films to hard lithified crusts (Monty, 1965; Golubic, 1976). Mats in the intertidal and sublittoral zones rapidly acquire a lithified crust (Hoffman, 1976; Logan and others, 1974); however in a subtidal regime, stromatolites may retain a partly gelatinous consistency during growth. In such a case, and particularly in the enlarged simple domes, gravity-induced collapse may occur when the weight of material in the dome reaches a critical point. Collapse may be aided by sudden violent events such as storms or current surges which remove supporting sediment (ooliths and intraclasts) from the interdomal areas. The dip angles of the small fault planes are consistent with gravity-induced slippage (such as in landslides). The spacing of slippage planes, and timing of the event in relation to the size of a dome, would depend on the cohesiveness and degree of lithification of the stromatolite. Elongation of the stromatolite domes in band 2 is predominantly north-northwest (Fig 47), although a few smaller domes (< 500 mm across) show an elongation in the same orientation (080°) as those in band 1. Elongation of the domes seldom reaches more than a length:breadth ratio of 2:1, and most are under 1.5:1. Alignment of coalesced groups of domes is also predominantly north-northwest.

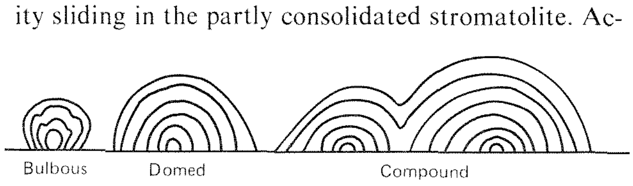

Figure 48. Generalized stromatolite types, band 2, Kulele Limestone.

Asymmetry of the domes takes the form of steepening and narrowing of the laminae on the northeasterly side. This is unusual, in that the plane of symmetry of the domes is approximately perpendicular to the direction of elongation; in most recorded cases of asymmetry in domed stromatolites, the plane of symmetry is parallel to the elongation direction (Hoffman, 1967; Vidal, 1971; Eriksson, 1977; Campbell and Cecile, 1975).

ORIGIN OF STROMATOLITE ELONGATION AND ASYMMETRY

Much has been written about the origin of elongation and asymmetry in stromatolite domes and mounds. Many workers report that elongation and asymmetry are directly related to current activity in both ancient (Hoffman, 1967; Campbell and Cecile, 1975; Haslett, 1976a) and modern (Gebelein, 1969) examples. Various types of current have been invoked, including tidal (Hoffman, 1974; Truswell and Eriksson, 1975; Young and Long, 1976; Button and Vos, 1977) and longshore drift (Young, 1974). In these examples, elongation is always parallel to the dominant current. At Shark Bay, recent stromatolites, from both subtidal and intertidal domains, show elongation in the direction of wave

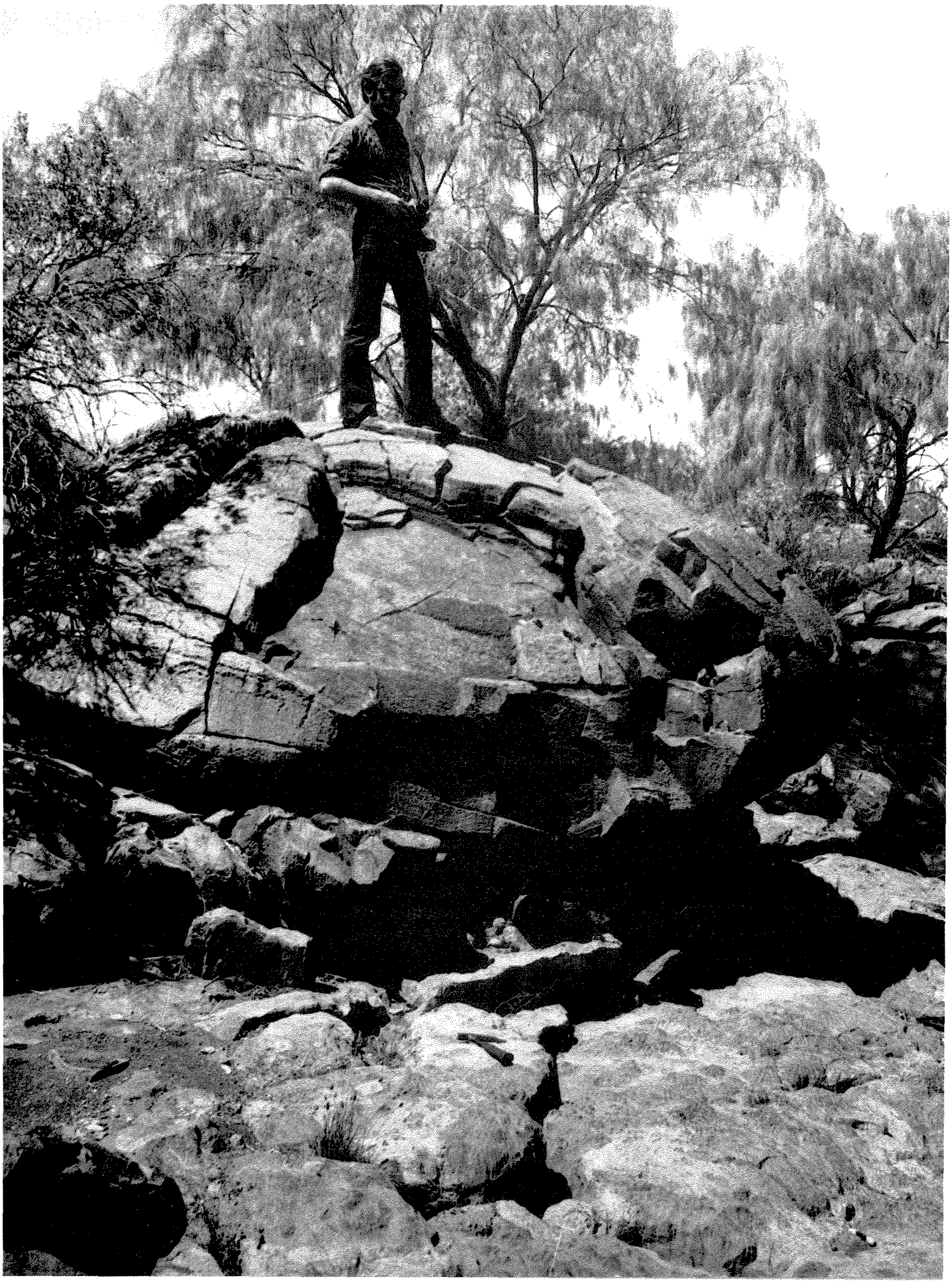

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Figure 49. Large bulbous stromatolite, Kulele Limestone, band 2, Thurraguddy Bore.

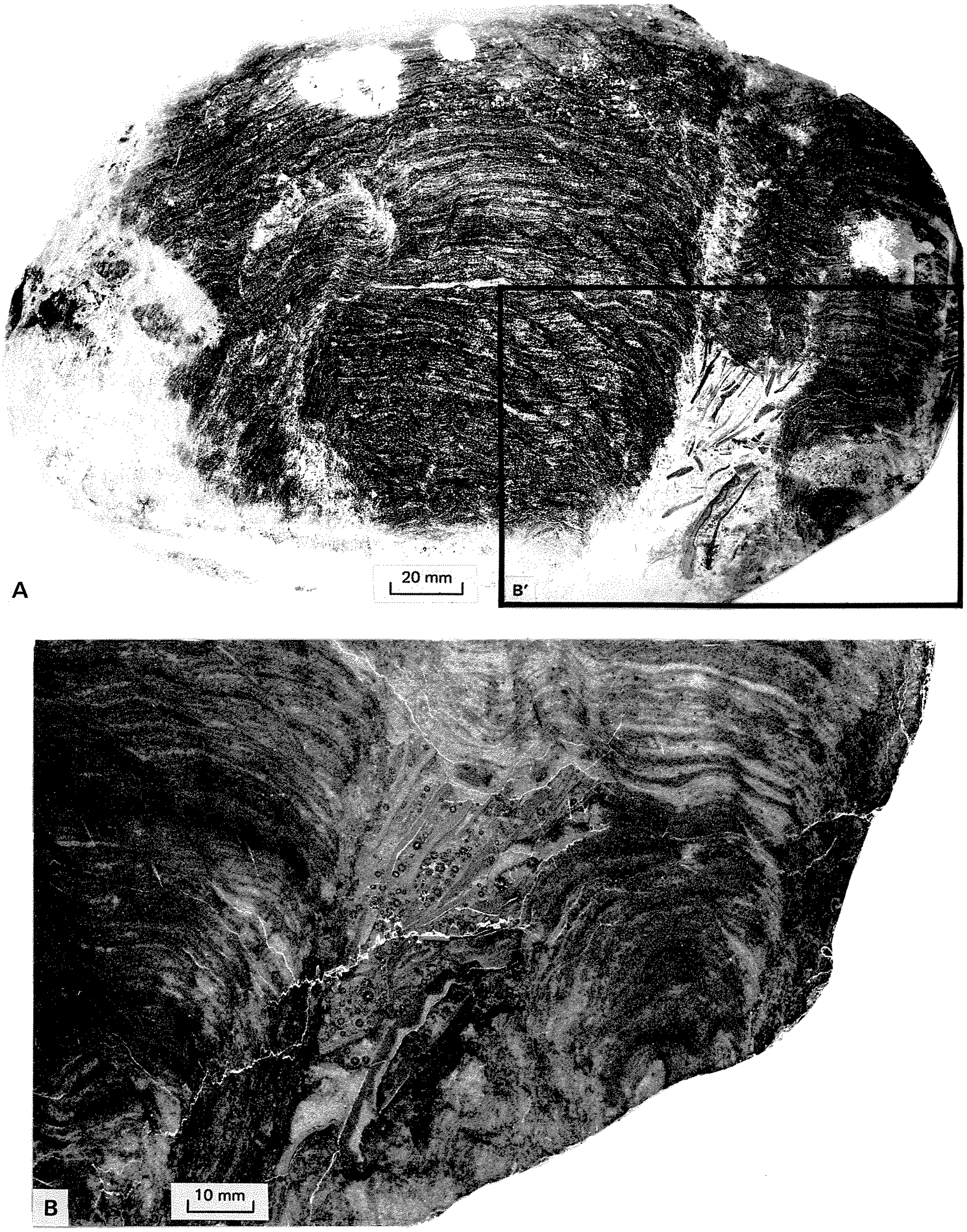

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Figure 50. A — Section through a small bulbous stromatolite, *Earaheedia kuleliensis* Grey 1984, Kulele Limestone, Thurraguddy Bore. Note the slippage planes, and the areas of intraclastic breccia. Acetate peel of cut slab.

B — Detail of intraclastic breccia; polished slab.

propagation, *i.e.* perpendicular to the shore (Logan and others, 1974; Playford and Cockbain, 1976). Truswell and Eriksson (1975) invoked a similar origin for the elongation of small, intertidal stromatolites in Proterozoic dolomite in South Africa. Playford (1980) also documented a wind influence in the Shark Bay stromatolites.

The mechanism invoked in most of these reports is one of scouring in the channels between stromatolite domes, and this mechanism is applicable to the Kulele Limestone. The presence of intraclastic breccias, ooliths, pisoliths and cross-bedding indicates sufficient current agitation to produce a scouring effect. The occurrence of two directions of elongation suggests different processes. The smaller domes, elongated at 080° (Fig. 47), closely resemble in size and form the wave-influenced domes described by Playford and Cockbain (1976) and Truswell and Eriksson (1975). This implies a shoreline to either the east or west.

The large domed stromatolites, occurring as they did in deeper water, were probably beyond the direct influence of wave activity at the shoreline itself. It is likely that their elongation is due to scouring by longshore tidal currents, acting at a high angle to the direction of wave propagation.

DEPOSITIONAL ENVIRONMENT

Much has been said in the preceding pages about the depositional environment of the Kulele Limestone. It remains only to summarize the interpretations.

The main feature of the Kulele Limestone is the cyclic development, albeit inconsistent, of the variable rock types. This suggests rapidly changing water depth, resulting in repeated transgression and regression. There is evidence of turbulence and agitation (ooliths, pisoliths, intraclastic breccias), current activity (cross-bedding in calcarenites, and ooliths) and periods of quiet but rapid sedimentation (fine-grained sandstones and shales with ball-and-pillow structures). Herringbone cross-bedding in some calcarenite also suggests a tidal influence. A repeated cycle from regressive lagoons or tidal mudflats, through an intertidal to a shallow subtidal regime is envisaged. The lack of coarse, terrigenous material indicates a low-lying, stable source area.

This postulated environment is in marked contrast to the quiet, probably lagoonal environment of the Wongawol Formation. It indicates the return of an open, shallow-marine influence and therefore marks a transgression. However, the upwards decrease in the amount of carbonate, and increase in clastic component of the cycles throughout the formation, show that the transgression was short-lived and that the Kulele Limestone ended with an overall regressive phase.

MULGARRA SANDSTONE

DEFINITION

Derivation of name: Named after Mulgarra Pool in the northeast corner of KINGSTON, Lat. 26°06′30″S, Long. 122°44′00″E.

Type section: The type section is taken in the Timperley Range area, between Lat. 26°00′40″S, Long. 122°43′40″E and Lat. 25°56′00″S, Long. 122°46′00″E. Only the lower part, which is predominantly sandstone, is well exposed. The remainder is too poorly exposed to allow accurate documentation of a type section.

Lithology: Mainly medium-grained quartz arenite in the lower 20 m of the formation. The middle part of the formation, in addition to thin arenite beds, contains some shale and pinkish limestone units. Sandstone again dominates the upper part of the formation.

Thickness and distribution: Because of poor exposure, total thickness is impossible to determine, but would approximate 100 m. The formation is confined to the eastern end of the exposed part of the main synclinorium, in southeast STANLEY and northeast KINGSTON.

Boundary criteria: The base is taken at the base of a 20 m thick unit of quartz arenite which forms a prominent scarp in the Timperley Range. The contact of the Mulgarra Sandstone with the shale and carbonate of the underlying Kulele Limestone is probably a disconformity. The top of the formation is not exposed, being unconformably overlain by either the Bangemall Group or the Permian Paterson Formation.

LITHOLOGICAL DESCRIPTION

The basal 20 m of the formation, which forms the main scarp of the Timperley Range, is a medium-grained, slightly ferruginous quartz arenite. The arenite is faintly bedded on a 100-200 mm scale, and has internal lamination on a scale of 5 to 15 mm, which in places, shows poorly developed, trough-style cross-stratification. The only minerals of note, other than quartz, are feldspar and glauconite, which together account for up to 5% of the total. Sedimentary structures are not abundant, but include ball-and-pillow structures, flute casts, load casts, current lineation, and current ripples.

The basal contact against the Kulele Limestone is sharp, although there is commonly a thin band of cream shale (1 to 2 m thick) between the top limestone band and the first quartz arenite.

The basal sandstone becomes finer grained in its upper few metres, and passes upwards into a mixed unit of interbedded, reddish-brown, fine-grained micaceous sandstone and shale. A few thin, lenticular-bedded limestones are present. This mixed unit is largely unexposed, and is probably between 20 and 40 m thick.

The remainder of the formation is mainly fine- to medium-grained, laminated micaceous sandstone with thin shaly interbeds. Cross-laminae, symmetrical ripple marks and intraclasts are features of the sandstone.

DEPOSITIONAL ENVIRONMENT

As with most of the rocks in the Earaheedy Group, the Mulgarra Sandstone is probably of shallow-marine origin, and may be either a transgressive or regressive sequence.

The transgressive model relies on the sharp basal contact with the Kulele Limestone, a contact similar to the Windidda-Wandiwarra Formation contact lower in the Earaheedy Group. In this case, the basal sandstone marks a sudden and rapid influx of sand during a marine incursion over an intertidal or lagoonal environment.

In the regressive model, the sandstone marks the appearance of shoreline conditions following the shallow subtidal deposition of stromatolitic carbonates. In view of the similarity between the basal quartz arenite and the thin arenite bands in the upper part of the Kulele Limestone, it seems likely that the Mulgarra Sandstone is a regressive sandstone phase of the Kulele Limestone cycles.

COMPARISON OF THE EARAHEEDY AND PADBURY GROUPS

DESCRIPTION OF THE PADBURY GROUP

Barnett (1975) defined the Padbury Group after modifying the original stratigraphy of McLeod (1970). Recognition of an unconformity in the Labouchere Formation, high in the Padbury Group, led Gee (1979b) to redefine the Labouchere Formation as the top unit of the Glengarry Group, and to define a new unit, the Wilthorpe Conglomerate, as the base of the Padbury Group.

The Padbury Group occurs in a complex synclinorium in the northwest part of the Glengarry Sub-basin (distinguished as the Padbury Sub-basin by Hall and Goode 1978). Interference between rising basement domes has produced three isolated, arcuate synclinal keels. This folding postdates an earlier phase of folding in the Glengarry Group; consequently there is a marked discordance at the base of the Padbury Group.

The following summary descriptions are based on those by Gee (1979b).

The Wilthorpe Conglomerate is a basal unit which has its maximum thickness of about 1 000 m near its type area around the westernmost syncline. Further east, it is only intermittently exposed beneath the scree of iron-formation from the Robinson Range Formation. The Wilthorpe Conglomerate typically consists of a coarse basal conglomerate up to 100 m thick, passing upwards into feldspathic granular sandstone with pebble beds, and then into siltstone and shale with thin layers of white chert. Clasts within the conglomerate consist predominantly of fine-grained silica- or hematite-cemented orthoquartzite: types that can be matched with the Karalundi Formation, the Juderina Sandstone and the Finlayson Sandstone. Clasts of vein quartz and lepidoblastic-textured metamorphic quartzite are also present. Lensoid bedding, imbrication of clasts, and shallow, trough-style cross-beds indicate a fluvial origin.

The Robinson Range Formation conformably overlies the Wilthorpe Conglomerate. It consists of banded and granular iron-formation interbedded with ferruginous shale. In the eastern syncline, two sub-parallel ridges of iron-formation are separated by magnetite-bearing hematitic shale. Whilst the lower unit is a true banded iron-formation, the upper unit is a granular iron-formation, characterized by lenses of granular and oolitic chert 10-20 mm thick, more continuous beds of the same thickness of jasper, and beds of intraclastic iron-formation up to 1 m thick. The intraclasts include spherical peloids of chert 0.5 mm in diameter, fragments up to 10 mm across of hematitic shale, green chloritic shale, chert and specular hematite, and larger fragments of disrupted jasper beds.

The uppermost unit, conformably overlying the Robinson Range Formation, is the Millidie Creek Formation which is best exposed in the central syncline. At least 1 500 m of feldspathic wacke, chert, dolomite, sericitic and hematitic shale, and granular iron-formation are exposed.

COMPARISON WITH THE EARAHEEDY GROUP

Comparisons can be made on regional and lithological grounds. Both the Earaheedy and Padbury Groups overlie the Glengarry Group with angular unconformity. Basal conglomerates in the

Wilthorpe Conglomerate and Yadgymurrin Conglomerate Member of the Yelma Formation each contain distinctive clasts which can be attributed to the Karalundi Formation.

The iron-formations of the Robinson Range Formation and Frere Formation display striking similarities in their distinctive granular textures, with peloidal, oolitic and intraclastic variations in each formation. A point of contrast is that, in the Robinson Range Formation, banded iron-formation dominates the lower part, whereas in the Frere, most of the banded iron-formation is in the upper part. It may have been that the Frere Formation was deposited on a more stable basement, so that the marine transgressive phase occurred as a gradual deepening, whereas the Robinson Range Formation was deposited in a part of the basin which was sinking more rapidly in the waning phases of the post-Glengarry Group deformation, but which filled up and became shallower as the area stabilized.

Although the iron-formations provide the most diagnostic lithological comparisons between the groups, other stratigraphic comparisons are possible. The rock types within the Wilthorpe Conglomerate (conglomerate, coarse sandstone, shale, and white chert) show a similar range of lithologies to those in the Yelma Formation around the western margin of the Earaheedy Sub-basin. The chert, dolomite and shale of the Millidie Creek Formation bear comparison with the Windidda Formation, and even minor granular iron-formation in the former could have its equivalent in the Windidda Formation in southeastern NABBERU.

From the foregoing, it is apparent that there is every reason to suppose that the Padbury Group can be correlated with the lower part (Tooloo Subgroup) of the Earaheedy Group. Whether there was ever a direct connection between the two is a matter for conjecture.

THE BASAL UNCONFORMITY OF THE EARAHEEDY AND PADBURY GROUPS

Following the deposition of the Glengarry Group, there was a period of deformation which was the consequence of continued upward and southward movement of the basement domes. The movement took place along the same lines of weakness that had influenced sedimentation in the Glengarry Group, and produced the Wilthorpe Conglomerate (of the Padbury Group) and the Yadgymurrin Conglomerate Member of the Yelma Formation.

Evidence for an unconformity between the Padbury and Glengarry Groups was presented by Gee (1979b) and a summary of this follows:

(a) A proven unconformity separates Wilthorpe Conglomerate from basement gneiss.

(b) Lenses of similar Wilthorpe Conglomerate occur immediately below the Robinson Range Formation in several widely scattered localities.

(c) There is an angular discordance, both local and regional, between bedding in the Wilthorpe Conglomerate and bedding in underlying rocks.

(d) Basal Padbury Group overlies a variety of rock types representing different stratigraphic levels of the Glengarry Group.

(e) The Wilthorpe Conglomerate contains clasts that can be matched with distinctive rock types in the Glengarry Group.

(f) The Padbury Group contains only one cleavage and is unmetamorphosed, whereas the Glengarry Group (in this part of the sub-basin) has been deformed and metamorphosed repeatedly.

An unconformity between the Earaheedy and Glengarry Groups is similarly inferred from the following evidence:

(a) The Yadgymurrin Conglomerate Member, which is considered to be basal Yelma Formation, dips gently east. It contains rounded clasts of quartzite that can be matched with Finlayson Sandstone, and it contains clasts of a distinctive pink chert, with brecciated or colloform textures, that can be matched with rocks of possible fumarolic origin in the Karalundi Formation. Such cherts do not occur within the Earaheedy Group.

(b) On the southern side of Lake Gregory on the extreme eastern margin of PEAK HILL (near Freshwater Well), conglomerate, which is probably equivalent to the Yadgymurrin Conglomerate Member, lies in contact with Glengarry Group sandstone and shale. There is marked angular discordance between the conglomerate and the underlying sandstone and shale, and the conglomerate contains pebbles of those underlying rock types.

(c) The conglomerate mentioned above in point (b) is overlain by chert breccia, sandstone and dolomite of the Yelma Formation. Extensive areas of these chert breccias, which are interpreted by R. D. Gee (pers. comm.) as a regolith at an unconformity, occur in the areas around the western side of Lake Gregory. These chert breccias in several places are seen to blanket the Glengarry Group on PEAK HILL.

(d) Regional correlation, on the grounds of lithology and stratigraphy, between the Earaheedy and Padbury Groups as discussed previously in this chapter, is accepted by most workers (Hall and Goode, 1978; Goode, 1981; Bunting and others, 1977; Gee, 1979a). Such correlation implies that the unconformity between the Padbury and Glengarry Groups must also be present between the Earaheedy and Glengarry.

Hall and Goode (1978) equate the Finlayson Sandstone and Yelma Formation, and appeal to a time-transgressive model to account for the vast thickness of Glengarry Group that must wedge out between the Yelma and Frere Formations. Goode (1981) cites the lithological similarity of the Yelma and Finlayson as evidence for their correlation. However, there are lithological differences between the Yelma and the basal Glengarry Group sandstones (Finlayson Sandstone and Juderina Formation) for the latter contain abundant ripple marks, are very fine grained and lack feldspar, in contrast to the Yelma Formation. It is suggested here that any apparent similarity of the Yelma and basal Glengarry sandstones is a reflection of similar depositional environments—as shoreline and shallow-marine deposits transgressing over the mature, peneplain landscape of the Yilgarn Block.

Following the erosion of the uplifted Glengarry Group, the whole area of the Nabberu Basin subsided and was inundated by an extensive marine transgression, during which the bulk of the Yelma Formation and the upper parts of the Wilthorpe Conglomerate were deposited. It is reasonable to postulate that the Padbury and Earaheedy Groups were physically connected during this marine transgression.

CHAPTER 4

Iron-formations

INTRODUCTION

The general description of the Frere Formation, given in Chapter 3, serves as an introduction to this chapter. The iron-formations of the Nabberu Basin are important, as they are one of the few representatives, outside of North America, of Superior-type granular iron-formations. The textures, structures and, to a limited degree, the mineralogy are identical to those of the North American rocks. They contrast strikingly with both the banded iron-formations of the Hamersley Basin and those of the Archaean granitoid-greenstone terrains of Western Australia.

This chapter documents the iron-formations of the Earaheedy Group in some detail, and is divided into three broad parts. Firstly, the problems of iron-formation nomenclature are discussed, and the terminology used herein is defined. Following this the mineralogy and texture of the granular iron-formations are described. Finally, the genesis and sedimentary environment are discussed in the light of constraints and conclusions provided by this data.

TERMINOLOGY OF IRON-FORMATIONS

DEFINITION

The term "iron-formation" has been defined in a number of different ways, each giving varying emphasis to certain characteristics of the rock. These characteristics include:

(a) chemical sedimentary nature;

(b) iron content;

(c) presence of chert; and

(d) presence of banding or lamination.

Each of these characteristics has a claim for inclusion in a definition, but each also needs to be qualified. For detailed discussion the reader is referred to James (1954), Brandt and others (1972), Trendall (1983) and other papers cited in this section.

A minimum iron content of 15% for iron-formations was advocated by James (1954) and Gross (1965), and this figure was subsequently adopted by Gary and others (1972). In this bulletin, a definition is preferred which is similar to that of Beukes (1973), in that the 15% iron boundary is not used rigorously, but it serves as an approximate limit between iron-rich sedimentary rocks (e.g. iron-formation) and ferruginous sedimentary rocks (e.g. ferruginous shale, ferruginous chert).

Iron-formation is a general name for dominantly chemically and/or biochemically precipitated iron-rich sedimentary rocks that consist mainly of chert and one or more iron-bearing minerals. Intraformational reworking of the precipitate may have occurred. Some fine-grained detrital material may be present, in which case the rocks are termed shaly iron-formation or iron-rich shale depending on the proportion of detrital material.

Ferruginous chert is a rock consisting predominantly of chert, but with a small amount of iron-bearing mineral. It grades into iron-formation, and where it is closely associated with iron-rich rocks it may be included with them, even though the iron content is below 15%.

Chert is cryptocrystalline silica which may have recrystallized to a microcrystalline, interlocking mosaic of quartz grains. It can contain minor amounts of dusty hematite (forming jasper), and occurs either as discrete beds, or as micro- or mesobands in banded iron-formation, or as allochems or matrix in granular iron-formation (see below).

TEXTURAL TERMINOLOGY OF GRANULAR IRON-FORMATION

Gross (1965) divided iron-formations into four types—Algoma, Superior, Clinton and Minette. The last two lie within the definition of "ironstone" (James, 1966). The Algoma type is predominantly banded iron-formation, and is largely restricted to the Archaean. The Superior type contains some microbanding, but is characterized by thicker layers in which the microfabric is granular rather than banded. These textures have been described in numerous papers dealing with the Lake Superior region (e.g. Aldrich, 1929; Moorhouse, 1960; Goodwin, 1960; Gross, 1965 and 1972; La Berge, 1964) and the Labrador Trough (Dimroth and Chauvel, 1973; Zajac, 1974; Klein and Fink, 1976). In the Nabberu Basin, the iron-formation is predominantly of the Superior type.

Those Superior-type iron-formations that lack bands typically contain grains that: are spherical to ovoid; are generally well-rounded; and are composed of iron oxide, chert, silicate mineral or, rarely, carbonate. The grains range in size from under 0.2 mm to 5 mm, with an average of about 1 mm. They commonly lack internal structure, but may be oolitic. In the Lake Superior region, these structureless grains have been called "granules", a term now widely used.

However, as pointed out by Dimroth and Chauvel (1973), use of the word "granule" is undesirable as it is generally used for the grain-size class between 2 to 4 mm in clastic sediments. In the Nabberu Basin, the structureless grains have been called "pellets" (Hall and Goode, 1975; Bunting, 1980) but this is an unfortunate choice for two reasons: (a) the word is used with a slightly different meaning in carbonate terminology (Bathurst, 1971), and with a more restrictive meaning in chert terminology (Dimroth and Chauvel, 1973), and, (b) in iron ore processing, pellets are the end product of a method of iron enrichment.

Dimroth and Chauvel (1973) introduced a new terminology for iron-formations, based on the similarity between textures in iron-formations and in limestones, and a slightly modified version is used here (Table 6). Subsequent to the completion of the present study, Beukes (1980) introduced a more detailed and comprehensive terminology, based on Dimroth and Chauvel's work, and more applicable to iron-formations generally.

Explanation of terms is given in a glossary (see Table 7), but further explanation of some terms is required here, particularly where usage differs slightly from that of Dimroth and Chauvel (1973). Where the words "iron-formation" or "chert" are prefixed by type of allochem (e.g. intraclastic, peloidal, oolitic) the rock consists predominantly of the allochem type. However, peloids (being a specific type of intraclast) may be present in rocks described as "intraclastic".

The term "granular iron-formation" is used to describe all cherty iron-formations containing allochems. The word "granular" here relates to the presence of discrete grains or aggregates of grains, and is not related to the word "granule" of particle size connotation. Thus "granular iron-formation" and "banded iron-formation" are two major subdivisions of iron-formation.

MINERALOGY

At the time of this study, all samples were from surface exposures, as no drill core was available from the iron-formations. This severely limits the amount of mineralogical and chemical work that can be applied to the iron-formations, because the amount and extent of surface weathering cannot be accurately determined. This problem is particularly severe when trying to determine the primary mineralogy of the rock. Oxide-facies assemblages dominate the mineralogy, the most common being hematite-quartz, and hematite-magnetite-quartz. Iron-rich carbonate and chamositic chlorite form a distinctive assemblage near Tooloo Bluff on KINGSTON, in the transition from Frere to Windidda Formation, but elsewhere, carbonates, silicates and sulphides of iron form only minor components of oxide-facies iron-formation. This is in marked contrast to the otherwise similar Superior-type iron-formations of North America, and it is not clear whether the difference is due to primary causes or to surface alteration in the Nabberu Basin.

QUARTZ

Most quartz is present as a very fine-grained mosaic in matrix chert or cherty allochems. Grain size is usually less than 0.01 mm, but increases with metamorphic grade. Silica cement has several forms, the most common being elongate crystals growing outwards from the surface of peloids. This was called "impingement" texture by Dimroth and Chauvel (1973). Voids between peloids may be filled with either radial chalcedony or a mosaic of coarse, euhedral quartz. Fine, granular quartz is present in syneresis cracks in peloids and ooliths.

TABLE 6. TEXTURAL ELEMENTS OF LIMESTONES AND IRON-FORMATIONS COMPARED (MODIFIED AFTER DIMROTH AND CHAUVEL, 1973)

Limestone (after Folk, 1962)	*Iron-formation*
Orthochems	
Micrite	Femicrite Matrix chert
Cement (spar)	Cement chert Cement quartz chalcedony Calcite (spar) Microbanded iron oxide-chert
Allochems	
Pellets	Pellets
Intraclasts	Intraclasts
Peloids	Peloids (equiv. to "granules")
Ooliths	Ooliths
Pisoliths	Pisoliths
Fossils (skeletal fragments)	Fossils (organic traces)
Shards (rare)	Accommodation shards

TABLE 7. GLOSSARY OF TEXTURAL TERMS (AS UTILIZED IN THIS BULLETIN)

Accommodation shards: Compaction of peloids to produce complex shapes with concave boundaries against other grains.

Allochem: A discrete particle, or aggregate of particles, that was transported before deposition, and which consists of material formed within the basin of deposition e.g. oolith, intraclast, peloid, fossil fragment.

Cement: Material precipitated in the pore space of a rock after deposition.

Femicrite: Matrix consisting predominantly of iron carbonate and iron silicate, probably deposited as a microcrystalline ooze.

Intraclast: Transported and redeposited fragment of penecontemporaneous sediment. Folk (1959) uses the term to describe carbonate rocks. Dimroth and Chauvel (1973) extend its use to iron-formations but it can be extended further to cover any sediment (e.g. penecontemporaneous shale fragments in a sandstone). It includes peloids, but the term usually refers to the coarser, more angular fragments in an intraformational breccia.

Matrix chert: Cryptocrystalline or microcrystalline silica filling the interstices between allochems, or supporting allochems in a more open fabric.

Oncolite: Unattached, subspherical to ovoid stromatolite with encapsulating laminae resembling an oolith, and formed by the accretion of successive layered masses of sheaths of micro-organisms. In the case of the Frere Formation oncolites, Hofmann and Schopf (1983) record a diverse assemblage of microfossils, with spheroidal, filamentous and bizarre forms.

Oolith: More or less spherical grain, under 2 mm in diameter, showing a concentric and/or radial internal structure. Oolite is a rock composed predominantly of ooliths.

Orthochem: Material directly precipitated at the site of formation of a rock.

Pellets: Elliptical or eye-shaped bodies of rather uniform size, generally about 0.2 mm long, loosely strewn in matrix chert against which they show gradational boundaries. In the Frere Formation, they are found only within intraclasts derived from matrix chert.

Peloid: Small rounded intraclast, devoid of internal textural variation, generally between 1 and 2 mm across.

Pisolith: Similar to oolith, but larger. Pisolite is a rock composed predominantly of pisoliths.

HEMATITE

In all the jasperoidal chert and iron-formation, hematite occurs as finely disseminated, submicroscopic dust, giving the characteristic red colour. In peloidal rocks, it forms dense aggregates of very small, almost opaque, red platelets, which are usually less than 3 μm across. These platelets are concentrated within peloids, and either fill the peloid, or form irregular patches, or display concentric zoning in peloids and ooliths (Fig. 51A). In the centres of some peloids, the hematite has recrystallized to a coarser type of specular hematite which is visible in hand specimen as small, metallic, grey clusters.

Hematite also occurs in the form of martite, as a secondary replacement of coarse-grained, commonly octahedral magnetite. In the iron-enriched weathering zone of the laterite profile, hematite occurs in association with hydrated iron oxides (limonite and goethite). The iron minerals have replaced silica which has been leached out.

MAGNETITE

Magnetite in its unaltered form seldom occurs in surface exposures of the Frere Formation. It is not present in the unmetamorphosed, flat-lying iron-formations in the southeast part of the basin, and elsewhere it is largely replaced by hematite to form martite. Despite this, the former presence of magnetite can be readily established by its octahedral habit. Prior to surface oxidation, magnetite was the main (and possibly the sole) iron oxide in those rocks which show some degree of metamorphic recrystallization, and as such it is therefore more commonly associated with coarsely recrystallized quartz than with microcrystalline chert.

Individual crystals of magnetite range from about 3 to 200 μm, most being in the range 20 to 50 μm. Much of the magnetite occurs as discrete octahedra, but most is concentrated in irregular aggregates (Figs 51B, C and D). The aggregates form patches within peloids, but occasionally transgress boundaries. The coarseness of the magnetite and associated recrystallized chert has the effect of obscuring the boundaries of peloids and coarser intraclasts, in places to a stage where the original shape of the fragment is barely discernible.

SULPHIDE

Traces of pyrite and chalcopyrite occur in some cherts associated with iron-formations, and in the cherts interbedded with ankerite and dolomite in the

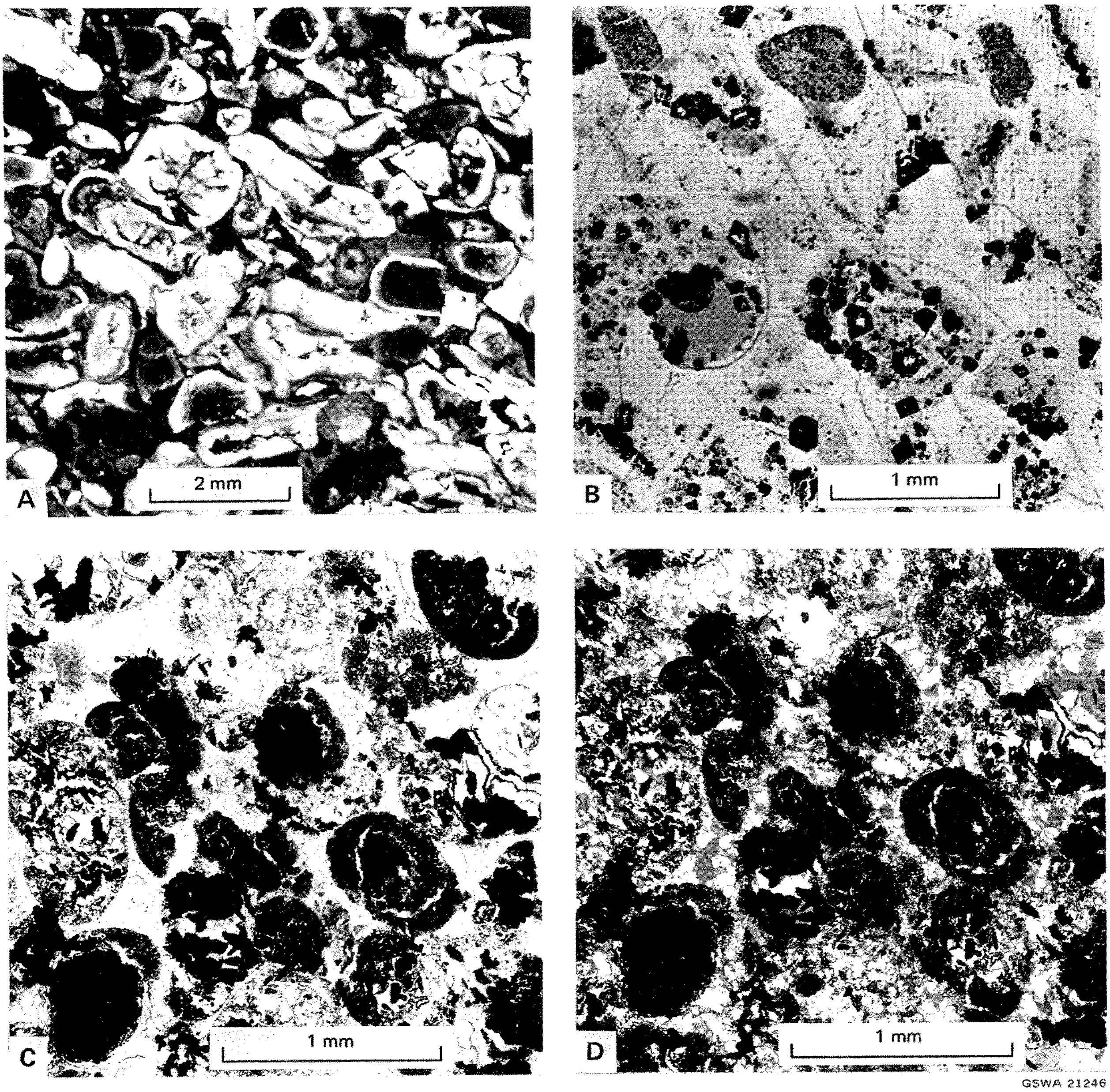

Figure 51. Hematite and magnetite in iron formations.

A — Zoned peloids with hematite dusting concentrated in cores. East of Camel Well. GSWA thin section 46550.

B — Disseminated octahedral magnetite in peloidal chert. Mount Lockeridge. GSWA thin section 46478.

C — Irregular patches of coarse magnetite replacing hematite in peloidal iron-formation. Some peloid outlines have become irregular. East of Simpson Well. GSWA thin section 46561.

D — Same as C, but crossed nicols. Note irregular recrystallization of quartz between and within peloids.

transition between the Frere Formation and the overlying Windidda Formation. Some limestone and dolomite units within the Frere Formation contain cubes up to 10 mm square of pyrite or limonitized pyrite. Pyritic shales, which account for much of the sulphide-facies iron-formations of North America, have not been found in the Frere Formation, although this may be due to surface-weathering effects.

Fresh pyrite was found at only one locality, 9.5 km north-northeast of Camel Well in northeast WILUNA, as small cubes in peloidal hematitic iron-formation.

CARBONATE

Iron-rich carbonate minerals are rare, and only at one locality is there a rock which approaches a carbonate-facies iron-formation. This is near Tooloo Bluff on KINGSTON, where a thin bed of ankeritic carbonate occurs in the transition zone between the Frere and Windidda Formations. The rock is buff coloured, and is speckled pale green by pellets and clots of chamositic chlorite. The carbonate minerals are a granular mixture of yellow-buff ferroan dolomite and brown ankerite. Varying proportions of brown ankerite define a crude 20 to 30 mm bedding. The bed is about 0.5 m thick and is poorly exposed, but good examples occur as rubble. Ankeritic carbonate forms a matrix/cement to jasperoidal peloids in the uppermost Windidda Formation in southeast NABBERU. Rhombs of iron carbonate occur rarely in granular iron-formation but more commonly in banded and shaly iron-formations, where they form isolated crystals up to 100 μm across. Nowhere do they occur in more than accessory amounts. In many cases, the original carbonate has been pseudomorphed by quartz, hematite or limonite.

IRON SILICATES

Iron silicates are extremely rare in the iron-formations of the Frere Formation. Near Tooloo Bluff, the ankeritic carbonate described in the preceding section contains elongate clots of a pale-green phyllosilicate. The mineral is pleochroic in shades of green, has low birefringence and has a refractive index in the range 1.65 to 1.66. The X-ray diffraction pattern shows a chlorite structure with a strong 1.4 nm (14Å) reflection, and thus has properties consistent with its being thuringite, an iron-rich chlorite which may be described, in a loose sense, as "chamosite" (Government Chemical Laboratories, Perth, Lab. No. 6118/78).

Elsewhere, phyllosilicate minerals occur only as rare accessories, usually as small flakes only a few micrometres long. Most are colourless, although one variety is pale yellow. All are too fine grained for positive identification. In wavy-bedded iron-formation east of Mount Royal (STANLEY), the iron-rich, microbanded portions contain sparse pseudomorphs of quartz after a phyllosilicate which, by its coarseness (up to 100 μm long) and habit was probably stilpnomelane.

Greenalite, an iron silicate which is abundant in the granular iron-formations of North America, has not been identified in any samples from the Frere Formation, although in some ferruginous chert the iron oxide peloids have a dark-green tinge which may be finely disseminated, submicroscopic greenalite.

Small rosettes of minnesotaite were recorded by Hall and Goode (1978) from the oncolitic iron-formation in northeast WILUNA. Nontronite has been observed in an iron-formation from an unknown locality (M. Gole, pers. comm.). These are the only reported occurrences of these minerals in the Earaheedy Group.

SPECULATIONS ON PRE-WEATHERING MINERALOGY

Iron-formations in the Frere Formation are remarkably similar to those of the Superior Province and Labrador Trough of North America, in terms of textures, structures, age, and tectonic and sedimentational setting. The only apparent difference is the abundance of iron silicates (greenalite, minnesotaite, stilpnomelane) and iron carbonates (siderite, ankerite) in the North American rocks (Dimroth and Chauvel, 1973; Floran and Papike, 1975) and their absence in the Frere Formation. The apparent absence may or may not reflect the primary mineralogy, as in Western Australia the rocks may have undergone severe surface oxidation and leaching due to prolonged weathering, and little trace of the original mineralogy remains.

Only drill-hole information will resolve this matter, and in the absence of such data, inferences have to be made from thin sections of surface samples. There is some direct evidence of carbonate and phyllosilicate at depth, such as the pseudomorphing of siderite rhombs and stilpnomelane by quartz and iron oxide. However, these pseudomorphs represent only accessory amounts of the replaced minerals. Because silicate and carbonate minerals in unmetamorphosed iron-formations are usually very fine grained, it may be that the process of replacement destroys original crystalline textures. Thus the main method of detecting their former presence, that is relic textures, is not available. In view of the similarities with other iron-formations, it therefore seems likely that, in the unmetamorphosed iron-formations of the Kingston Platform, some silicate- and carbonate-facies rocks are present below the surface weathering zone.

By contrast, the coarseness of the grain size in the recrystallized iron-formation in the Stanley Fold Belt means that silicate minerals, if of a similar grain size, should be visible as pseudomorphic replacement textures, even after weathering. The general absence of such textures in thin section suggests that silicates were not present prior to metamorphism.

TEXTURE

The terminology used here for describing the texture in the iron-formations was discussed earlier in this chapter. It is based on the terminology of Dimroth and Chauvel (1973) who made the textural analogy between iron-formation and limestone, and who therefore adapted the limestone terminology of Folk (1959) for use with iron-formation.

Adoption of this system of terminology in no way implies that the iron-formations were once carbonate rocks; rather it is an acceptance of the textural similarity of the two groups of rocks. It should also be noted that the advantages with the limestone analogy apply mainly to granular iron-formation—the analogy is inadequate to describe extensive banded iron-formations such as those in either the Hamersley Group or Archaean greenstone belts.

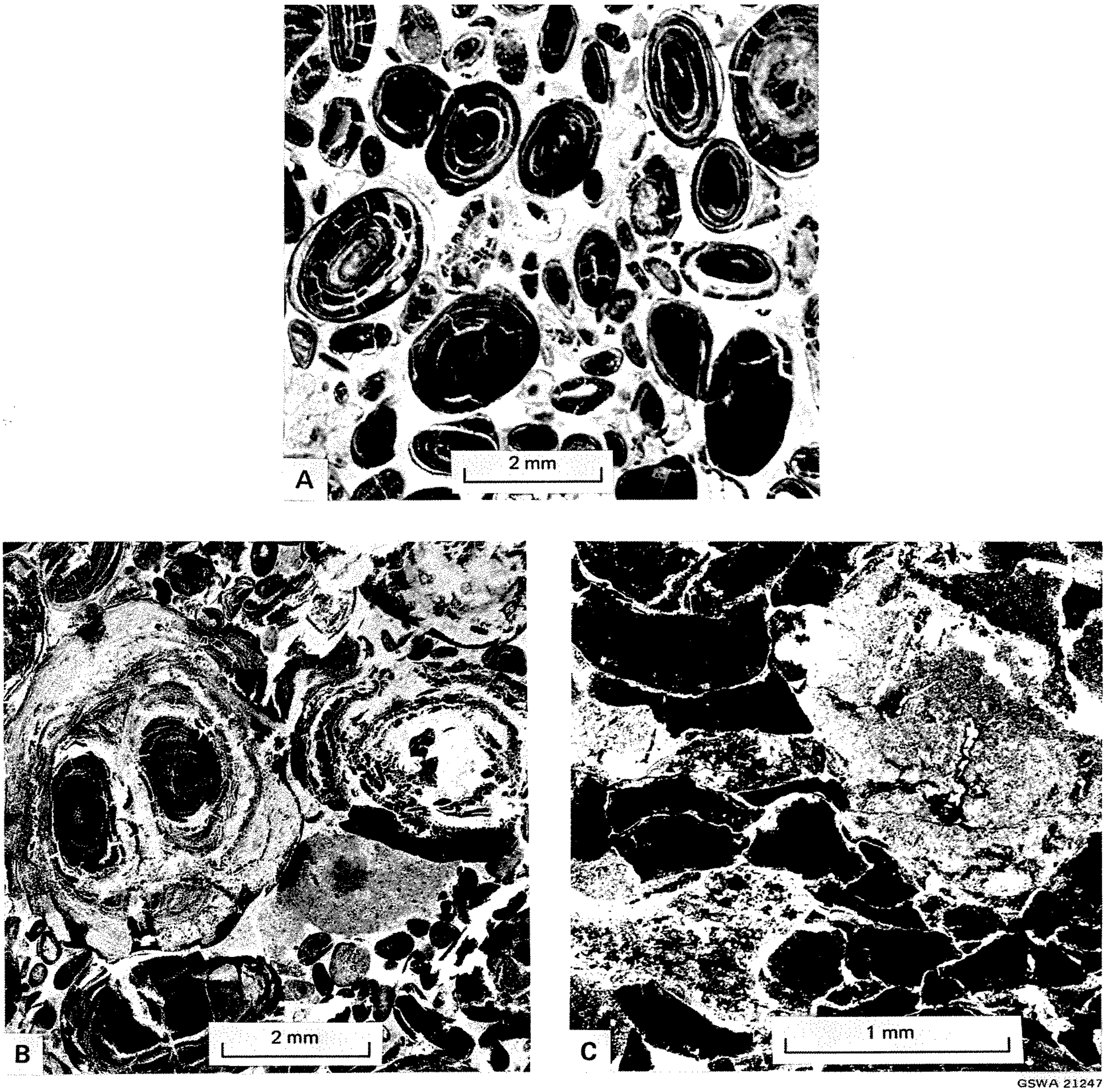

Figure 52. Allochems in granular iron-formation.

A — Ooliths, with smaller hematitic and cherty peloids. Note the radial and concentric syneresis cracks. Camel Well area. GSWA thin section 46547.

B — Complex pisoliths, the largest containing two ooliths in the core. Note numerous smaller peloids. Camel Well area. GSWA thin section 46546.

C — Small intraclasts or peloids deformed into "accommodation shards" between larger cherty intraclasts. Snell Pass. GSWA thin section 40085.

ALLOCHEMS

This term embraces all the granular fraction of iron-formations and cherts, and includes ooliths, pisoliths, peloids, intraclastic breccia fragments, oncolites and stromatolite fragments; that is, chemically or biochemically derived material formed within the basin of deposition but not at the site where it was finally incorporated into the rock. This excludes clastic detritus brought in from outside the basin. The allochems are usually visible in hand specimen, and are set in an orthochemical matrix or cement.

Ooliths and pisoliths: These are particularly common in the ferruginous cherts along the southern margin of the basin. Where best developed, they consist of multiple alterations of hematite and chert arranged as concentric laminae about a central nucleus (Fig. 52A). The chert is usually cryptocrystalline, and may have a dusting of hematite which give the rocks a reddish jasperoidal tinge. The formation of specular hematite has destroyed some of the fine laminae. Recrystallization to coarser chert and magnetite (now martite) also destroys the laminae, often to the extent that the oolitic form is barely discernible.

Many of the ooliths are compound and show a complex history of fragmentation and regrowth. The final stage of concentric precipitation in such cases commonly results in a skin of very thin laminae.

Pisoliths (diameter greater than 2 mm) are restricted to a single bed, 4 km east of Camel Well in northeast WILUNA. The bed is between 0.2 and 0.5 m thick, and is part of a sequence of interbedded hematitic shale and peloidal iron-formation. Apart from their size, and a greater number of concentric laminae (Fig. 52B), the lithology of the pisoliths is similar to that of the ooliths. The association of the pisoliths with oncolites and stromatolites suggests that the activity of micro-organisms may have played a part in their formation.

The origin of ooliths and pisoliths, in both iron-formation and carbonate rocks, has been discussed at length in the literature. Apart from a few distinctive types of quiet-water ooliths (Davies and others, 1978; Freeman, 1962), most researchers agree that concentric ooliths form in shallow, agitated water above wave-base.

Intraclasts: Intraclastic breccias in the Frere Formation form distinctive beds within the iron-formation members, particularly in the Frere Range. Microbreccias occur within the shaly iron-formations in northeast WILUNA, and isolated intraclasts are present in many peloidal beds. The term "intraclast" is used to describe a fragment of penecontemporaneous sediment, which has been broken up, transported, then redeposited in its present situation. Although the definition includes peloids, these are described separately in the next section. The intraclasts described here differ from peloids in being generally coarser and more angular—a consequence of being more lithified and therefore brittle when fragmentation occurred. However, there is complete gradation between the two types.

The intraclastic breccias form beds 100 to 400 mm thick, in much the same manner as the oolitic and peloidal iron-formations. The clasts range in size from less than 1 mm to over 100 mm long by 40 mm thick. Commonly there is a bimodality in the clast size. Most intraclasts are very irregular but roughly tabular in shape, and many show evidence of post-depositional plastic deformation by remobilization of silica at the edges. A common feature of many breccia beds is a partly continuous, partly disrupted chert layer near the top of the bed.

The breccias are characterized by a great diversity in clast lithology, within the constraints of the ubiquitous iron oxide-chert mineralogy. Predominant types are black chert, red jasperoidal chert, shale, shaly iron-formation, and specular hematite (only in intraclasts under 10 mm). Many of the large intraclasts show internal microbands, and a few are compounds of previously formed peloids. Some intraclasts contain small (<0.1 mm), irregular spheres, giving a globular texture. These spheres are similar to the "pellets" described by Dimroth and Chauvel (1973) from matrix chert in the Sokoman Iron Formation (Labrador). Dimroth and Chauvel suggest an origin either by aggradation of silica-gel drops in sea water, or by diffusion segregation in the sediment.

Peloids: Peloids are small, well-rounded, spherical or ellipsoidal grains, generally about 1 mm across although they can reach 2 mm. They are by far the most abundant allochem in the iron-formation. They are distinguished from ooliths by a lack of concentric layering, being either homogeneous or with a simple and gradual zonation (Figs 52A and B). The mineralogy, in unmetamorphosed rocks, is essentially simple—a mixture of chert and hematite. There is complete gradation in the proportions, however, ranging from pure chert, in some of the non-ferruginous peloidal cherts, through the jasperoidal cherts in which the peloids consist of hematite dusting in chert, to iron-rich rocks in which the peloids are entirely hematite (usually specular). As in the ooliths, metamorphic recrystallization of magnetite and quartz tends to destroy the simple, rounded outline.

Zonation takes the form of variation in percentage of iron mineral. There is no overall consistency of zonation; it can be from an iron-rich core to an iron-deficient rim or vice-versa. A mixed provenance is indicated in some samples by the occurrence of both types of zonation, and unzoned peloids, in the same rock. In all cases, diagenetic chert-filled syneresis cracks postdate the zoning; therefore the variable zoning probably occurred after the formation of the peloids but before final deposition as a sediment. In some rocks where the peloids are all zoned to iron-rich rims, the zoning was probably post-depositional and very early diagenetic.

As with the larger intraclasts, peloids originated by the break-up (by storm action) of penecontemporaneously precipitated chemical sediment. Their high degree of roundness and sphericity suggests that the break-up was prior to lithification of the silica gel. Although in many cases the peloids show evidence of plastic deformation after burial (Fig. 52C), they must have been fairly cohesive at the time of deposition.

An interesting relationship between intraclasts and peloids is present in breccia beds near the top of the Frere Formation at Mount Teague. The breccia intraclasts were brittle during fragmentation, and some have suffered slight mechanical rounding. The intraclasts are lithologically variable, and include ferruginous shale, jasperoidal chert, and peloidal iron-formation. The presence of these peloids indicates an earlier period of peloid formation, prior to lithification and brecciation. However, between the breccia intraclasts there is a later generation of individual peloids within matrix chert. These later peloids were deposited with the breccia intraclasts, and are moulded against the intraclasts and each other in a plastic rather than brittle fashion. Thus, peloids were still being formed while earlier lithified peloids were being brecciated.

Oncolites and stromatolites: These are restricted to the jasperoidal pisolitic bed east of Camel Well in northeast WILUNA. The oncolites form irregular bodies, several centimetres across, which display very fine concentric laminae (Fig. 17). Some have cores of jasperoidal, intraclastic breccia, and others are themselves brecciated. Stromatolitic beds a few centimetres thick are laterally continuous, and display a fine, wavy lamination. Both oncolites and stromatolites are rich in microfossils, some of which resemble modern iron bacteria (Walter and others, 1976). Both filamentous and spheroidal microfossils are present in the oncolites.

ORTHOCHEMS

The orthochemical portion of iron-formation is the material directly precipitated at the site of formation of the rock. A distinction is made between matrix (material deposited at the same time as the allochems) and cement (material precipitated in pore spaces during lithification). In some cases, this distinction is difficult to make, particularly where the orthochem is cryptocrystalline, or where it has been recrystallized to a coarse quartz mosaic during metamorphism. Recrystallization also obscures some boundaries between matrix chert and chert within allochems.

Matrix chert: Matrix chert, in the unmetamorphosed iron-formation and cherts, typically forms a very fine-grained quartz mosaic with an average grain size of between 1 and 5 μm (Fig. 53A). Recrystallization has coarsened the quartz to about 20 μm in the slightly metamorphosed rocks of the Frere Range, and up to 0.1 mm along much of the northern and western side of the sub-basin. The matrix chert is commonly similar in appearance to chert within allochems, and probably had a similar origin as a gel-like precipitate or ooze. Much of the matrix chert in the iron-formation is coloured red by hematite dust. Rarely, such as in the peloidal iron-formation at the top of the Windidda Formation in southeast NABBERU, the matrix is a ferroan carbonate (?ankerite). The carbonate is sparry, but may be replacing a micritic matrix similar to the "femicrite" of Dimroth and Chauvel (1973).

Cement: Most cement in the iron-formation and chert is siliceous, although in some of the more iron-rich varieties hematite or goethite is now the dominant cementing agent. Four principal types of siliceous cement are recognized, and commonly more than one is present in the same rock.

(a) Chalcedonic cement occurs as radiating fans, with undulose extinction, projecting outwards from the surface of allochems (Fig. 53C). During the early stages of metamorphism, the chalcedony breaks down to microcrystalline quartz, and the boundaries of the fans become sutured.

(b) Quartz with columnar impingement texture forms elongate crystals perpendicular to the surface of the allochems (Figs. 53B, C and D). Some terminate in poorly developed crystal faces.

(c) Quartz mosaic represents the final infilling of the pore space, and occurs with both the chalcedonic and columnar types. The quartz is usually equant and has an allotriomorphic texture (Fig. 53B).

(d) Microcrystalline quartz cement is identical to matrix chert in texture, and can be distinguished only when it fills desiccation cracks in allochems (Fig. 53D).

Microbanded iron oxide-chert: This is included with the orthochems, because it forms *in situ* as a precipitate and has not been transported. In the Earaheedy Group, it is a minor component of the iron-formations, but it is the dominant rock type in most of the major banded iron-formations such as those of the Hamersley Group. Unlike the granular iron-formations it is generally accepted that, for the microbanding to be preserved, it must be precipitated below wave base, and this is probably true of the genuine, thicker banded iron-formations within the Frere Formation.

However, an unusual variant occurs east of Mount Royal, where wavy-bedded iron-formation consists of alternating layers of microbanded iron-formation and peloidal chert (Fig. 54). The layers are very irregular and do not persist laterally for more than a metre or so. The microbands consist of irregular and crinkled alternations of quartz and hematite (after magnetite). The textural and compositional contrast between the peloidal and microbanded layers is so great that it is difficult to explain it by a rapidly alternating depth to wave base. Perhaps precipitation of the microbanded layers was aided by organic activity, such as the growth of iron bacteria. An alternative explanation might be found in Haslett's (1976b) model for the origin of similar features in Cambrian limestone. This would envisage the microbanded portions as being pathways of maximum chemical solution, along which soluble material (in this case silica) was removed, perhaps during early diagenesis, leaving behind the insoluble residue (iron oxides).

ORIGIN OF THE IRON-FORMATIONS

The genesis of iron-formations is a topic which has been debated for most of this century. The problem has been approached from many directions, dealing with the chemical, physical, biological, sedimentational, environmental, mineralogical, and lithological aspects of iron-formation development. The complex interplay of all of these factors has resulted in a plethora of genetic models, many of

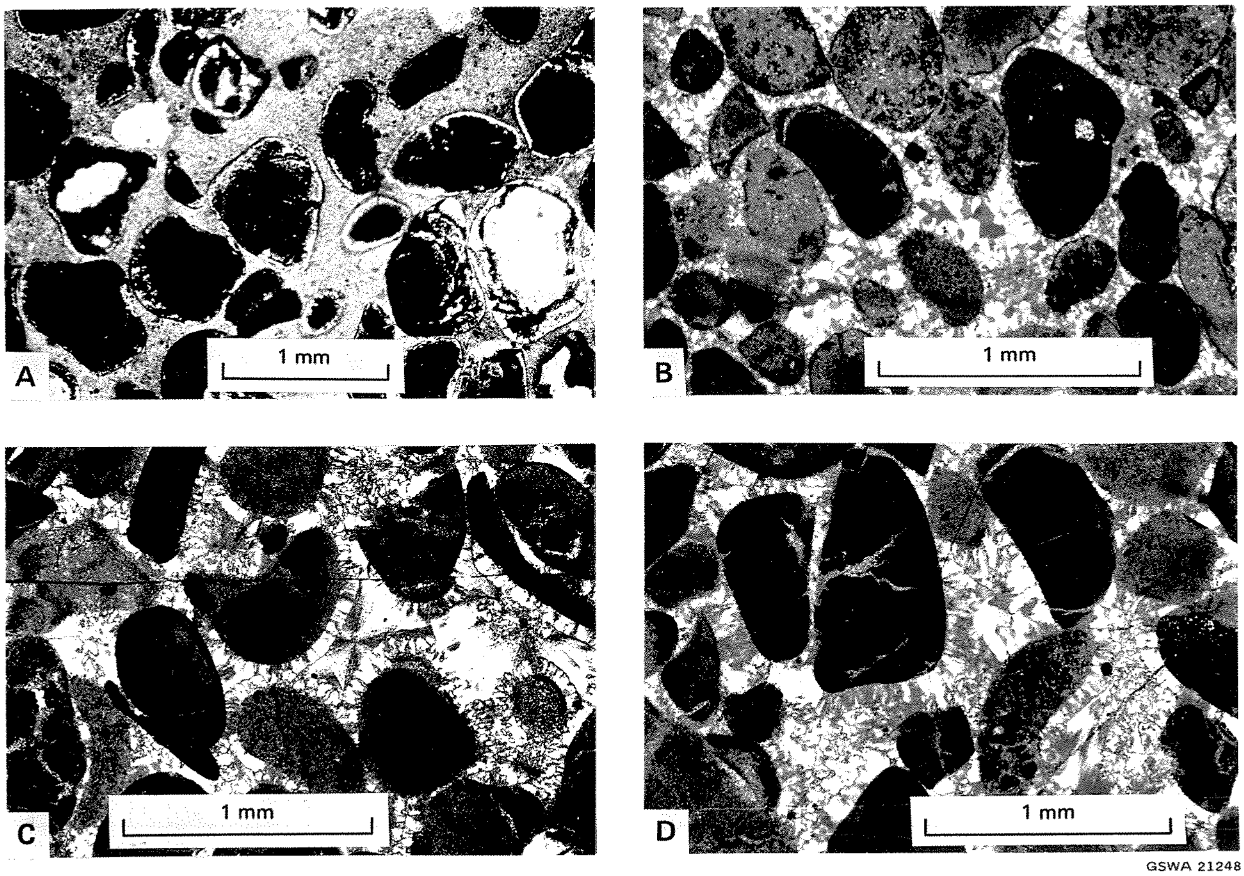

Figure 53. Orthochems in peloidal iron-formation.

A — Cryptocrystalline matrix chert, supporting zoned peloids. The Jump-Up. GSWA thin section 46419.

B — Granular quartz cement in voids, with weakly developed "impingement" texture surrounding peloids. Northeast WILUNA. GSWA thin section 46346 (crossed nicols).

C — Well-developed "impingement" texture, with chalcedonic cement filling the voids. East of Lake Teague. GSWA thin section 46482 (crossed nicols).

D — Same thin section as C. Void filling partly recrystallized. Note microcrystalline cement filling syneresis cracks. Crossed nicols.

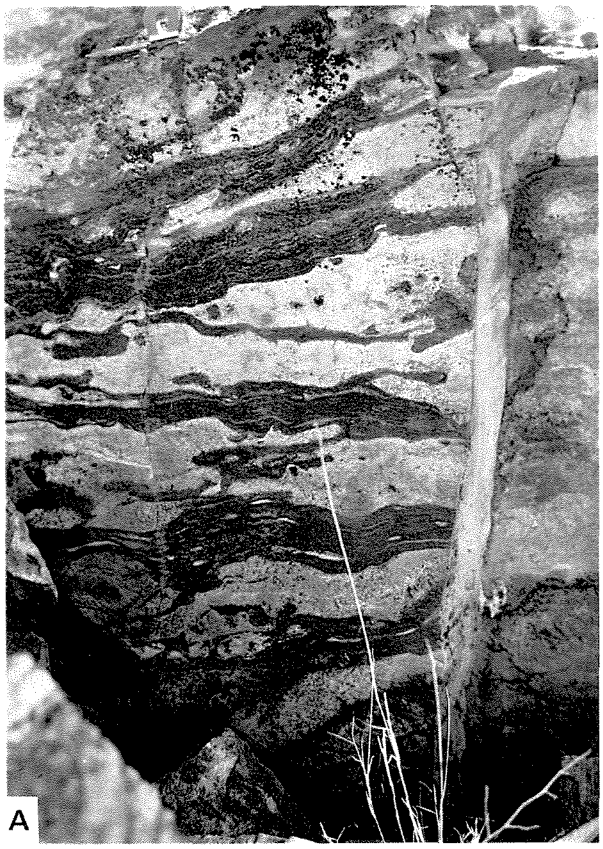

Figure 54. Microbanded iron oxide-chert.

A — Wavy-bedded iron-formation, with microbanded iron oxide-chert layers (dark) and peloidal chert (light). Roadside east of Mount Royal.

B — Thin section from A showing irregular and wavy microbanding, expressed as alternating iron oxide (hematite after magnetite) and recrystallized chert. GSWA thin section 42878.

them contradictory. For general discussions the reader is referred to papers by James (1954, 1966), Trendall and Blockley (1970), Gross (1972), Kimberley (1979) and Cloud (1983).

Most workers have suggested that the precursor material of iron-formation was probably a complex chemical precipitate containing iron hydroxides, iron-silica gels, and colloidal particles of iron silicates and carbonates (French, 1973). In the Nabberu Basin there is, at the present stage of knowledge, little evidence as to the nature of the original precipitate; however, it would seem that iron silicates and carbonates were of minor and localized importance. Early diagenesis resulted in lithification of the original precipitate by crystallization of quartz (as cryptocrystalline chert and interstitial cement) and iron oxides. There is no evidence in the Frere Formation that the primary oxide was other than hematite (the only iron oxide in the unmetamorphosed iron-formation), although in the metamorphosed iron-formations the coarsely crystalline magnetite may have been derived from a fine-grained primary magnetite phase. It is feasible that in the slightly deeper water in the northwest part of the Earaheedy Sub-basin (corresponding to the zone of higher metamorphic grade), conditions were less oxidizing than in the near-surface area to the southeast, and were therefore conducive to the formation of primary magnetite.

Unlike the true banded iron-formations, in which diagenesis and lithification took place in the undisturbed precipitate, the granular iron-formations suffered mechanical disruption prior to final lithification. In some parts of the Frere Formation, the plastic deformation of the peloids indicates that the precursor material was still gel-like when it was disrupted. The resulting peloids attained their well-rounded shapes while still gel-like, and were deformed against each other under depositional load pressure. However in many cases, the peloids were sufficiently lithified to maintain their shape and even rupture in a semi-brittle fashion prior to final deposition. The angularity of some intraclastic breccias indicates that some of the precipitated material was lithified *in situ* to the extent that the disruption was more or less brittle. Thus it appears that, in gross terms, early diagenesis and mechanical disruption overlapped.

SEDIMENTARY ENVIRONMENT AND WATER DEPTH

Granular iron-formations were formed in shallow water, where storm action could disrupt the layers of silica gel/iron hydroxide or chert which had been precipitated during quieter periods. The choice of storm activity rather than strong-current activity (such as tidal currents) as the mechanism of disruption is based on the general lack of current structures within the iron-formations. Some ripple cross-lamination is present in the interbedded shales, but the iron-formations contain only a few scours. The finely banded iron-formations require deeper water, and a very flat sea floor, in order to preserve the delicate and continuous microbanding from the disruptive effects of storms and bottom currents. Trendall and Blockley (1970) suggested that a water depth of between 150 and 250 m is reasonable for the banded iron-formations of the Hamersley Basin. The storm-disrupted granular iron-formations must have formed at a depth considerably less than this.

The storm waves disrupted not only the newly formed precipitate, but also earlier formed shale and iron-formation, and apparently mixed all of these to form the intraclastic breccias. Some of the finer material was transported into shallower water where it was winnowed by normal wave activity, thus producing the better sorted peloidal iron-formation and chert. The peloids were deposited in low, shoal-like deposits which resulted in the lensoid and wavy bedding that is characteristic of these rocks. Repeated agitation of the material in some of the shoals produced ooliths. Broken ooliths and peloids, some of them recemented into complex ooliths, indicate periods of storm activity within these shallower domains.

SOURCE OF IRON

Several different models for the source of iron in Precambrian iron-formations have been presented in the last few decades, and these have been discussed by Holland (1973) and Drever (1974). Most fall into three broad groups:

(a) surface weathering (James, 1954; Lepp and Goldich, 1964);

(b) volcanic exhalations (Goodwin, 1956; Oftedahl, 1958; Trendall and Blockley, 1970); and

(c) upwelled deep ocean water (Borchert, 1960; Holland, 1973; Drever, 1974).

With regard to the Earaheedy Group, three points are relevant.

(a) There is no convincing evidence of concurrent volcanic activity within the depositional area. Although a few shards of possible tuffaceous origin have been noted (Hall and Goode, 1978) most of these are very fine-grained and could have been transported by wind for a considerable distance from a volcanic source.

(b) Several observations from the Frere and other formations of the Earaheedy Group indicate that there was a low-lying land area to the south and/or east. The shale:iron-formation ratio increases dramatically in those directions, and in central ROBERT oolitic chert of the Frere Formation contains coarse detrital quartz. However, iron content of the iron-formations decreases towards the presumed shoreline, and thus weathering of that landmass is an unlikely source of the iron.

(c) The deeper water iron-formations of the western and northern Frere Formation are typically iron-rich.

These observations suggest that the source of iron for the Frere Formation was to the northwest, in the Glengarry Sub-basin or beyond. The Glengarry Group (which at this stage would have been deposited, lithified, deformed and eroded) contains abundant mafic volcanics, including some unusual exhalative cherts, deposited in a deep-water trough which was tectonically unstable. A source of iron could be available either from erosion of newly uplifted Glengarry Group, or from exhalations produced by waning volcanic activity related to the volcanics in the Glengarry Group. In both cases, the source would have been remote from the present limit of the iron-formations. The erosion model implies a trough of Earaheedy Group age which would admit iron into the waters of the basin, but would form a barrier to the movement of detritus.

Whatever the ultimate source of the iron, it appears to have come into the Earaheedy Sub-basin from deeper water to the northwest, and to have been deposited in water less than the depth to storm wave-base. Precipitation of iron in this zone limited the availability of iron in the area to the east and, combined with limited circulation of water in the shallower shoal-like environment, effected a progressive easterly decrease in iron content.

SUMMARY OF CONCLUSIONS ON IRON-FORMATION DEVELOPMENT

This largely regional investigation does not permit the formation of a complete model for the origin of iron-formations in the Nabberu Basin. However, certain conclusions can be made which limit the application of existing models.

(a) The original precipitation of iron and silica took place above storm-wave base, during periods between storm activity.

(b) The iron was probably precipitated from water upwelling from deeper parts of a trough to the northwest. The upwelling oceanic model of Drever (1974) could have applied to the Frere Formation.

(c) Storm activity was responsibile for the breaking up of both the original precipitate and previously deposited iron-formation.

(d) Some intraclastic breccia formed at or near the site of disruption. Smaller grains (peloids) were transported into shallower water.

(e) In this shallow water, matrix chert and siliceous cement were deposited between the transported grains. This material, generally speaking, was less ferruginous than the material transported from deeper water.

(f) The characteristic mineral assemblage hematite-chert was formed during early diagenesis—a process which spanned the time from pre-breakup to post-deposition.

(g) Some shaly clastic sediments which are interbedded with the iron-formations show features attributable to a tidal influence, indicating a shallow-marine environment. The iron-formations themselves show no tidal influence.

(h) Most of the peloidal and oolitic iron-formation formed as low shoals and banks in shallow, wave-agitated water. The environment was similar to that of the present day Bahamas, and to the environment, derived from a Bahaman model, described by Chauvel and Dimroth (1974) for the Sokoman Iron Formation of Canada.

CHAPTER 5

Structure

REGIONAL STRUCTURAL SETTING OF THE EARAHEEDY SUB-BASIN

Although the structure of the Earaheedy Sub-basin is simple, it is related to the complex and repeated deformations that occurred along the northern margin of the Yilgarn Block. This tectonic boundary also forms the southern margin of the Capricorn Orogen (Gee, 1979a) which is a major orogenic belt that evolved in the period circa 2.0 to 1.5 Ga ago. The Earaheedy Sub-basin, being the youngest of the major sedimentary sequences in this marginal zone, records the last major deformation of the orogen, and reflects the waning phases of basement reactivation.

Figure 55 shows the major structural features of this orogenic margin, and a brief discussion is warranted to place the deformation of the Earaheedy Group in its tectonic context. Three different expressions of the nature of this tectonic boundary are seen in a west-to-east section along the boundary. The most westerly segment is an *en echelon* line of mylonite zones, to the north of which the Archaean granitoids and gneiss are affected by metamorphism, migmatization and penetrative deformation that are related to basement reworking of Archaean rocks in the orogenic belt. Movement on the shear zones is inferred to be dextral and north-side-up because the Yarlarweelor Gneiss Belt seems to have pushed upwards and laterally into the central segment.

En echelon regional, near-isoclinal folds in the Glengarry Sub-basin mark the easterly extension of the major fracture zone. South of the fold belt, the Glengarry Group has a gentle northerly dip, and is interpreted to be the sedimentary cover over a stable craton of the Yilgarn Block. In addition to the tight regional folds, the mobile part of the tectonic zone is characterized by a horst block of basement (Goodin Dome) which was active during sedimentation, and a basement diapiric dome (Marymia Dome) which was

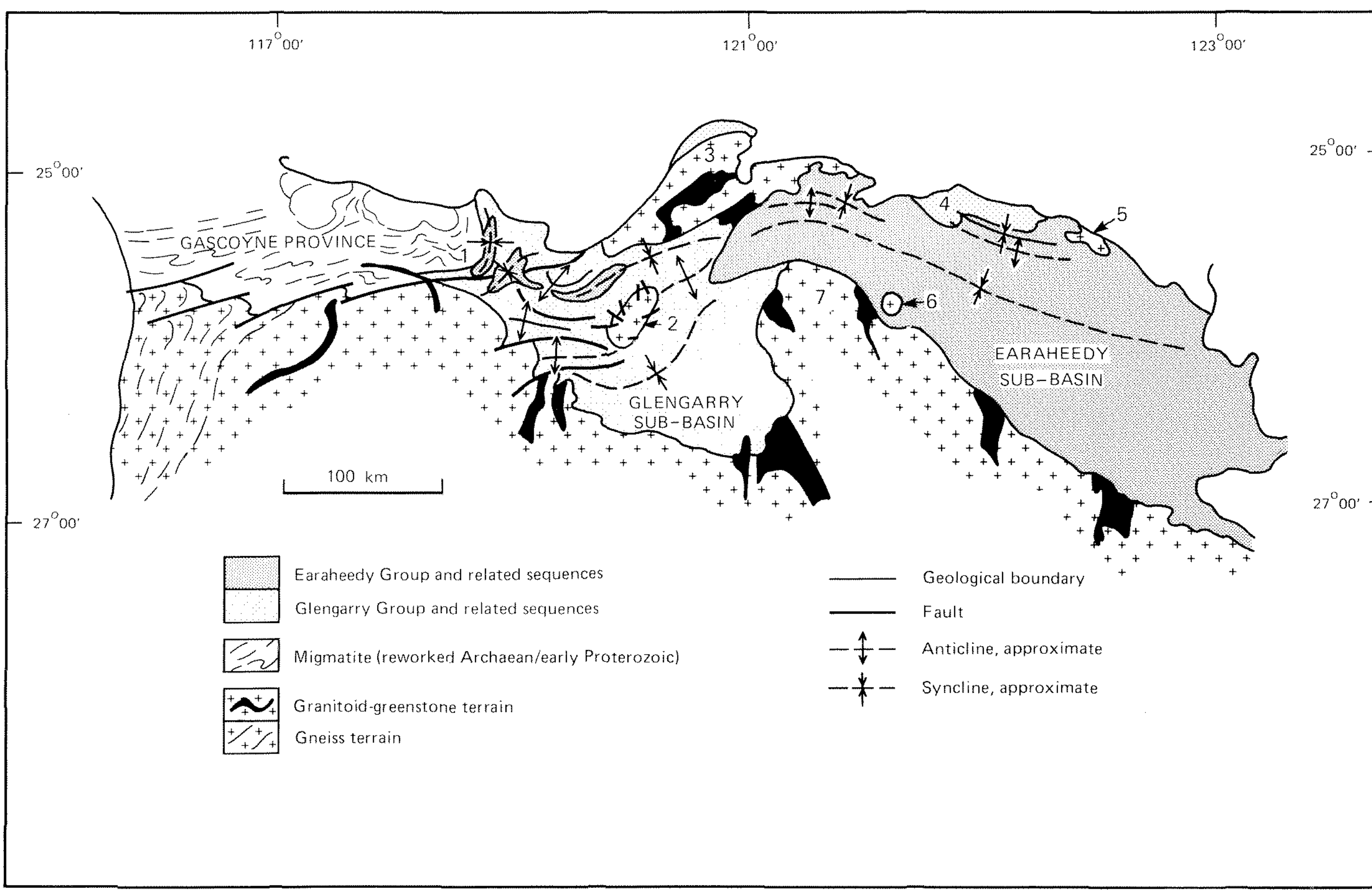

Figure 55. Regional structural setting of the Earaheedy Sub-basin in relation to the northern margin of the Yilgarn Block. The Kingston Platform lies as cover on the stable craton; the Stanley Fold Belt lies north of the first major synclinal axis in the Earaheedy and Glengarry Sub-basins. Points labelled by numbers: 1) Yarlarweelor Gneiss Belt, 2) Goodin Dome, 3) Marymia Dome, 4) Troy Creek Inlier, 5) Malmac Dome, 6) Teague Ring Structure, 7) Wiluna Arch.

active during the later stages of deformation. Tight synclinal keels of Padbury Group, which is correlated with the Earaheedy Group, occur in the fold belt, and indicate at least two main phases of deformation. The demarcation between platform cover and fold belt in the Glengarry Sub-basin is taken to reflect a sharp transition from cratonic basement to the south and mobile basement to the north. Deformation patterns in this segment result from upward and southward movement of basement beneath the orogenic zone, and the *en echelon* nature of the folds also indicates a dextral component of movement.

The most easterly segment of this tectonic zone is represented by the relatively simple synclinorium in the Earaheedy Group. The trace of the syncline aligns with the general east-west trend of the structures to the west, and collectively with those structures, defines the major east-west crustal fracture zone along the northern margin of the Yilgarn Block. Subsequent descriptions in this chapter deal with the details of structural patterns within the eastern part of the Nabberu Basin.

KINGSTON PLATFORM

Strike trends in the gently dipping eastern part of the Kingston Platform are west-northwest in NABBERU and western KINGSTON, but swing in a broad arc to east-west in western ROBERT. The width of the platform diminishes westward, and is only a few kilometres at the point of inflection on the Wiluna Arch.

FOLDS

Depite the absence of penetrative deformation, three phases of folds with different styles and orientations are recognized, and all postdate the earliest period of deformation in the Glengarry Sub-basin. Folds on the Kingston Platform take the form of gentle undulations and angular chevron folds, of wavelengths from a few metres to several kilometres. Folds commonly occur singly in otherwise uniformly dipping strata, although there is some clustering of folds, often of more than one phase, particularly in western KINGSTON. There is a tendency for the intensity and abundance of folds to increase to the north and northwest, towards the Stanley Fold Belt.

The earliest and dominant folding is the main deformation in the fold belt to the north that produced the regional synclinorium. The folds have subhorizontal axes, and axial trends subparallel to the regional strike. The folds are commonly asymmetrical, with short, steep south-dipping limbs and long, gentle north-dipping limbs; some are almost monoclinal. This style of folding is best seen in clusters of folds west of Banjo Well, in the Wellington Range (where the Wellesley Chert Member is repeated by the folding) and in northwest KINGSTON. The asymmetry of the folds and their spatial association in some cases with axial-plane reverse faults, indicate that the mechanism of folding may be by draping over basement faults.

The second generation of folding is visible only on outcrop rather than regional scale. The folds are open, generally symmetrical, and have axes which trend between 010° and 030°. These folds generally plunge to the north-northeast at between 5° and 10°. In northwest KINGSTON, interaction between these and the earlier asymmetric folds has resulted in complex strike patterns.

The last generation of folding was a gentle warping along north-trending axes. The age of these folds relative to the second phase is inferred and not proved. The warp folds create dome-and-basin interference patterns where they intersect the asymmetrical folds of the earliest period. This has produced the outlier of Kulele Limestone at Thurraguddy bore (Fig. 45), inliers of Princess Ranges Quartzite east of Wongawol, and inliers of Windidda Formation at Forked Creek. The complex-bedding trends in the Wongawol Formation and Mulgarra Sandstone, north of Lake Carnegie, also result from this interference.

FAULTS

Faulting is of minor importance in the Kingston Platform. Most faults trend within 20° of east-west, and observed planes of such faults dip steeply to the north. Most are reverse faults, with downthrows to the south, and some are associated with the early asymmetrical folds (Fig. 56). In some instances the presence of such faults, with this trend, is inferred in order to explain major offsets in otherwise continuous bedding trends, such as the basal unconformity near Yelma Outcamp, and the Windidda-Wandiwarra contact west of Mount Alexandra.

A group of small, normal, north-trending faults cuts the Windidda Formation near Windidda homestead.

The Lockeridge and Merrie Range Faults are two important north-northwest-trending structures that displace rocks of the Earaheedy Group. They are extensions of major faults in the Yilgarn Block, and have a history of movement and influence on regional tectonics going well back into the Archaean. The Perseverance Fault near Wiluna has a similar history of prolonged movement (Elias and Bunting, 1982).

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Figure 56. Small reverse fault associated with asymmetrical fold in shale of Frere Formation, northeast DUKETON. North is *to the right.*

TEAGUE RING STRUCTURE

The Teague Ring Structure is a near-perfect ring structure, of uncertain origin, that lies close to the southern margin of the Kingston Platform. Only a brief summary of the present state of knowledge of this structure is given in this bulletin. Full details are given in Bunting and others (1980).

It contains a circular core, 12 km in diameter, most of which seems to be composed of very poorly exposed foliated granitoid of the otherwise buried Yilgarn Block basement. In addition, a minor component is an alkalic syenite which gives a Rb-Sr isochron age of 1 630 Ma. The exact nature of the contact between the foliated granitoid and the Earaheedy Group is uncertain, due to poor exposure, but is probably an unconformity.

The rim is composed of upturned Yelma and Frere Formations, outside of which lies a near-perfect ring syncline approximately 21 km in diameter. Steeply dipping normal faults form a polygonal pattern around the northern half of the ring. These faults all indicate an upward movement of the core. In general, the structure is slightly asymmetrical in a northeast-southwest oriented profile, with steeper dips and stronger deformation to the northeast than to the southwest. A plane of symmetry therefore exists perpendicular to the regional trend of the basin margin.

Several lines of evidence indicate that the core and rim were subject to shock metamorphism. Shatter cones are sparsely distributed in the iron-formation beds of the rim. Fluid-filled, planar, chevron-patterned deformation lamellae occur in quartz of the granitoid rocks of the core, but have not been found in the syenite. Very small (less than 1 mm) veins of pseudotachylyte cut the granite and syenite. These veinlets consist of brecciated quartz and feldspar fragments in devitrified glass.

Two possible origins for the Teague Ring Structure were advanced by Butler (1974): a meteorite impact and a diapiric intrusion. Horwitz (1975b) suggested an origin by interference of folds. Bunting and others (1980) argued against an extraterrestrial origin, noting that it bore similarities with similar alkaline granitic plugs, of similar age, associated with evidence of very high-strain rates. They advocated a cryptoexplosive origin caused by a sudden discharge of magmatic gases related to the emplacement of syenite.

STANLEY FOLD BELT

FOLDING

In the Stanley Fold Belt east of the Wiluna Arch, there is only one major period of deformation and folding. This produced a slaty cleavage which increases in intensity northwards, away from the Kingston Platform. This increase is accompanied by an increasing development of metamorphic muscovite in the pelitic rocks, such that they grade from cleaved shale in the south to phyllite and muscovite schist in the north. At the same time, bedding lamination in both pelitic and arenaceous rocks is progressively destroyed.

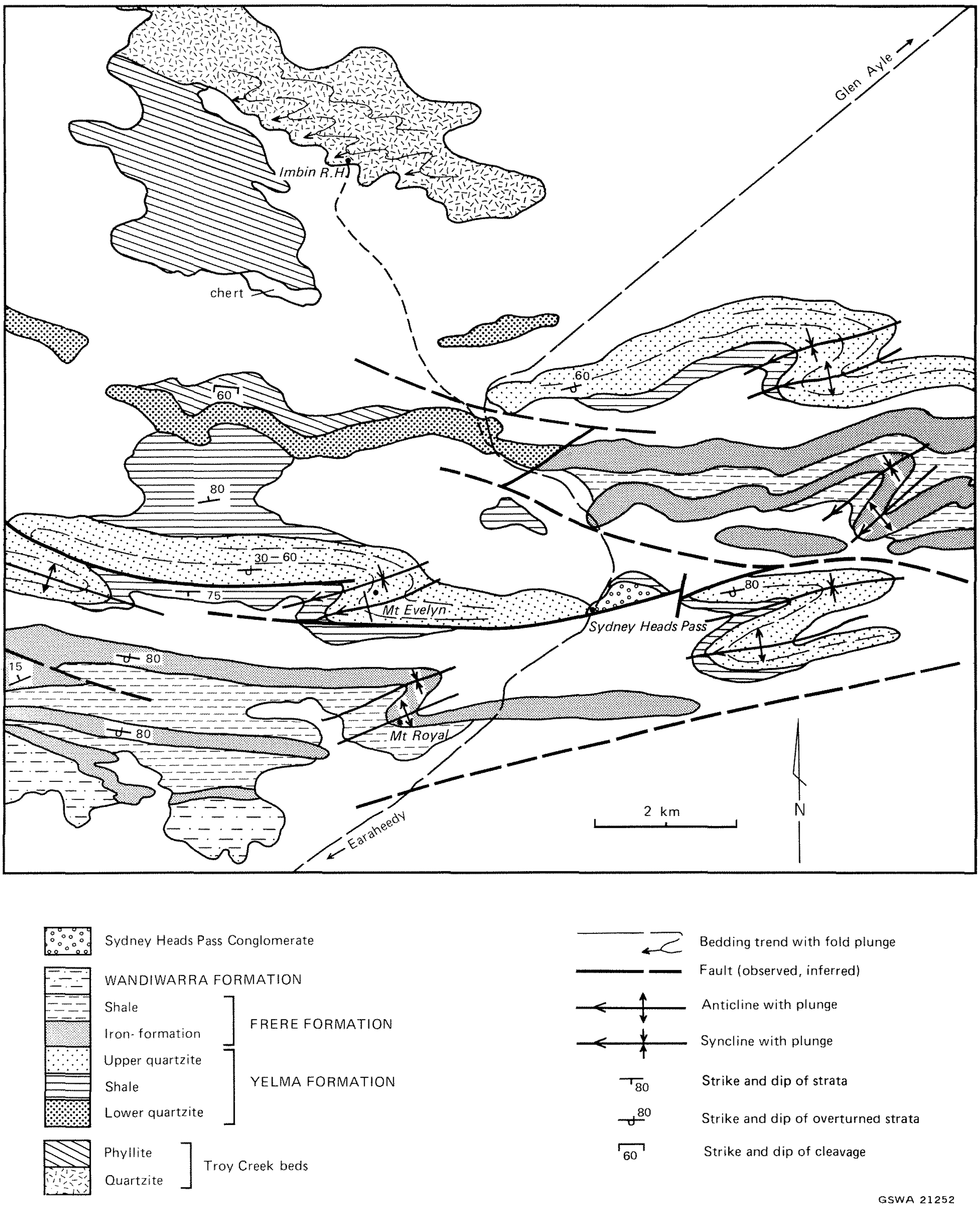

Figure 57. Outcrop geology and structure—Sydney Heads Pass area.

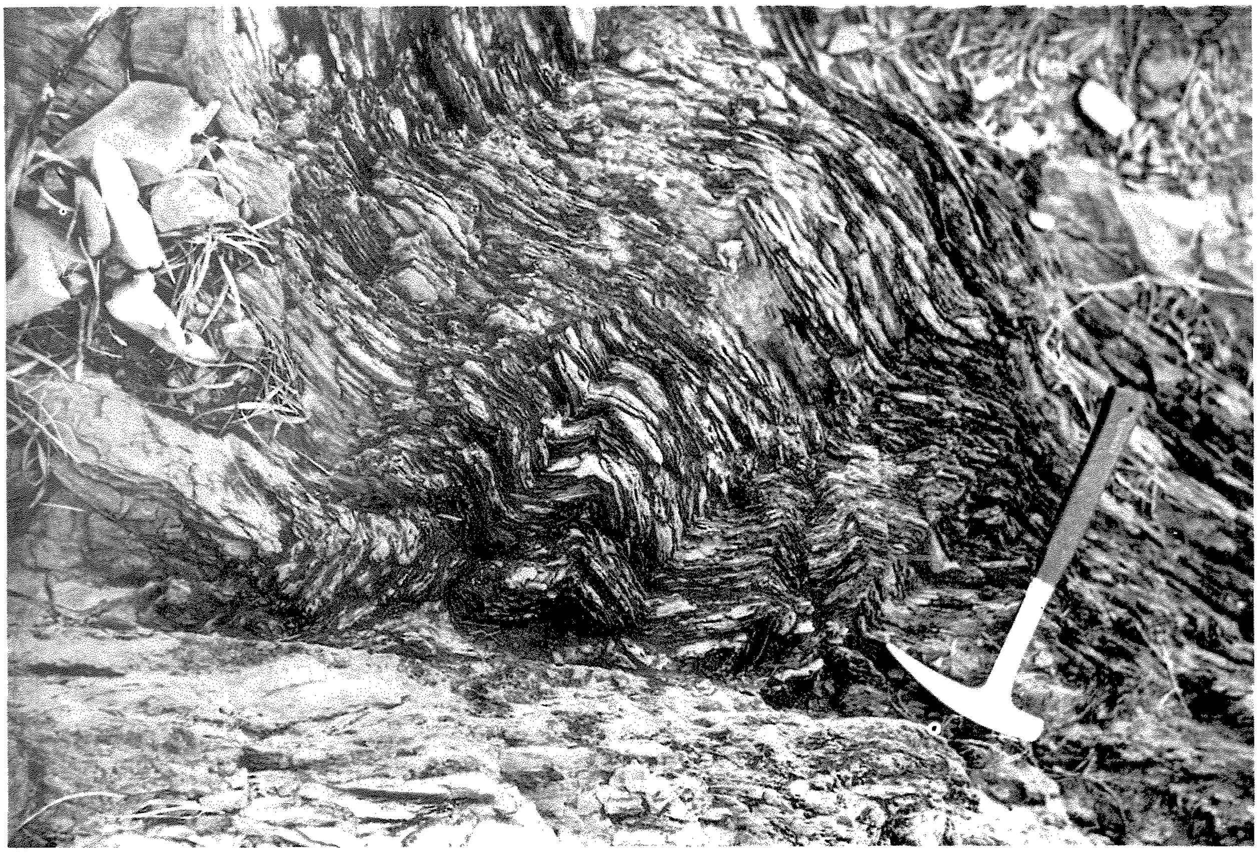

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Figure 58. Chevron folds in cleaved shale, Yelma Formation, east of Sydney Heads Pass.

Folding in this part of the belt is along approximately east-west axes. Axial planes dip steeply north and most folds are asymmetrical. The longer limbs are steeply dipping to the south and commonly overturned, whereas the short limbs dip gently north (up to 30°). The tightness of the folds and the degree of asymmetry increase northwards away from the Kingston Platform.

Between the Troy Creek Inlier and the Malmac Dome, these asymmetrical folds are clearly outlined by the parallel ridges of resistant iron-formation and quartzite. From Mount Evelyn to the Malmac Dome the folds display Z-asymmetry and plunge west at 10° to 20° (Fig. 57). Near Mount Cecil Rhodes they plunge east at up to 60° and display S-asymmetry. In the intervening area the axes are near horizontal. To the west of Mount Cecil Rhodes, in the vicinity of Troy Creek, they again plunge west and have Z-asymmetry. This pattern of plunge reversal is the result of open regional cross-folds on north-south trends which are probably related to folds of the same style in the Kingston Platform. Small-scale features, such as chevron folds, crenulation and kink bands, trend at a high angle to the regional slaty cleavage, and may also be related to the later cross-folding (Fig. 58).

The main asymmetrical folds are generally similar-style folds, often with sharp inflections across the hinges. Some smaller folds at outcrop scale have a tendency for the more resistant quartzite and iron-formation layers to be more concentric in style. Small north-dipping thrusts and reverse faults are associated with the asymmetrical folds, and are usually parallel to the axial planes.

THRUST FAULTING

In the area from the Malmac Dome to Mount Cecil Rhodes, the intense folding of the northern limb of the synclinorium was accompanied by steep thrusts subparallel to strike. In many cases these had the effect of cutting out the overturned limbs of asymmetrical folds. This is most apparent in the Sydney Heads Pass area (Fig. 57), where several steeply dipping shear zones and fault breccias can be seen in outcrop. Near Mount Cecil Rhodes, several asymmetrical folds within a regional fold have been faulted into a series of thrust slices. In the intervening area, and elsewhere along this limb of the synclinorium, such faults are probably more abundant than actually observed. Being parallel to strike and preferentially developed in the incompetent shale beds, thrust faults are difficult to identify.

The major synclinorial structure of the Stanley Fold Belt is therefore seen as distinctly asymmetrical, with the northern limb being strongly cleaved, and with abundant minor folds and thrusts indicating upthrusting and overturning from north to south.

TROY CREEK INLIER

The Troy Creek Inlier is a small structural high composed of rocks similar in lithology and deformational history to those of the Glengarry Group. It is interpreted to be part of the Earaheedy basement that was elevated to its present position during the main penetrative deformation of the Stanley Fold Belt. Together with the other structural highs, the Marymia Dome and the Malmac Dome, it may mark the approximate northern limit of the fold belt, although pre-Earaheedy Group rocks to the north are now largely covered by those of the Bangemall Group.

In the Troy Creek Inlier, two penetrative foliations are present: an earlier phyllitic schistosity, and a later crenulation cleavage. The later foliation is related to the main deformation of the Earaheedy Group, whereas the earlier one probably records the main deformation that affected the Glengarry Sub-basin. The later foliation is dominant throughout the inlier, but in the northern part the earlier foliation is clearly recognizable. Figure 59 shows the earlier foliation crenulated by the later foliation. Porphyroblasts of ?garnet (now totally weathered) formed during the earlier event.

INTERACTION OF FOLD PATTERNS ACROSS THE WILUNA ARCH

Over the Wiluna Arch, a zone of interference folds has resulted from the interaction of the north-easterly fold trends of the Glengarry Sub-basin and the west-northwesterly trends of the Earaheedy Sub-basin. These patterns are expressed by dome-and-basin deformations involving both the Marymia Dome and the Earaheedy Group.

It would appear that the earlier pre-Earaheedy Group structures have influenced the direction of the later deformations, perhaps by renewed movement on faults and tightening of pre-existing folds. Indeed

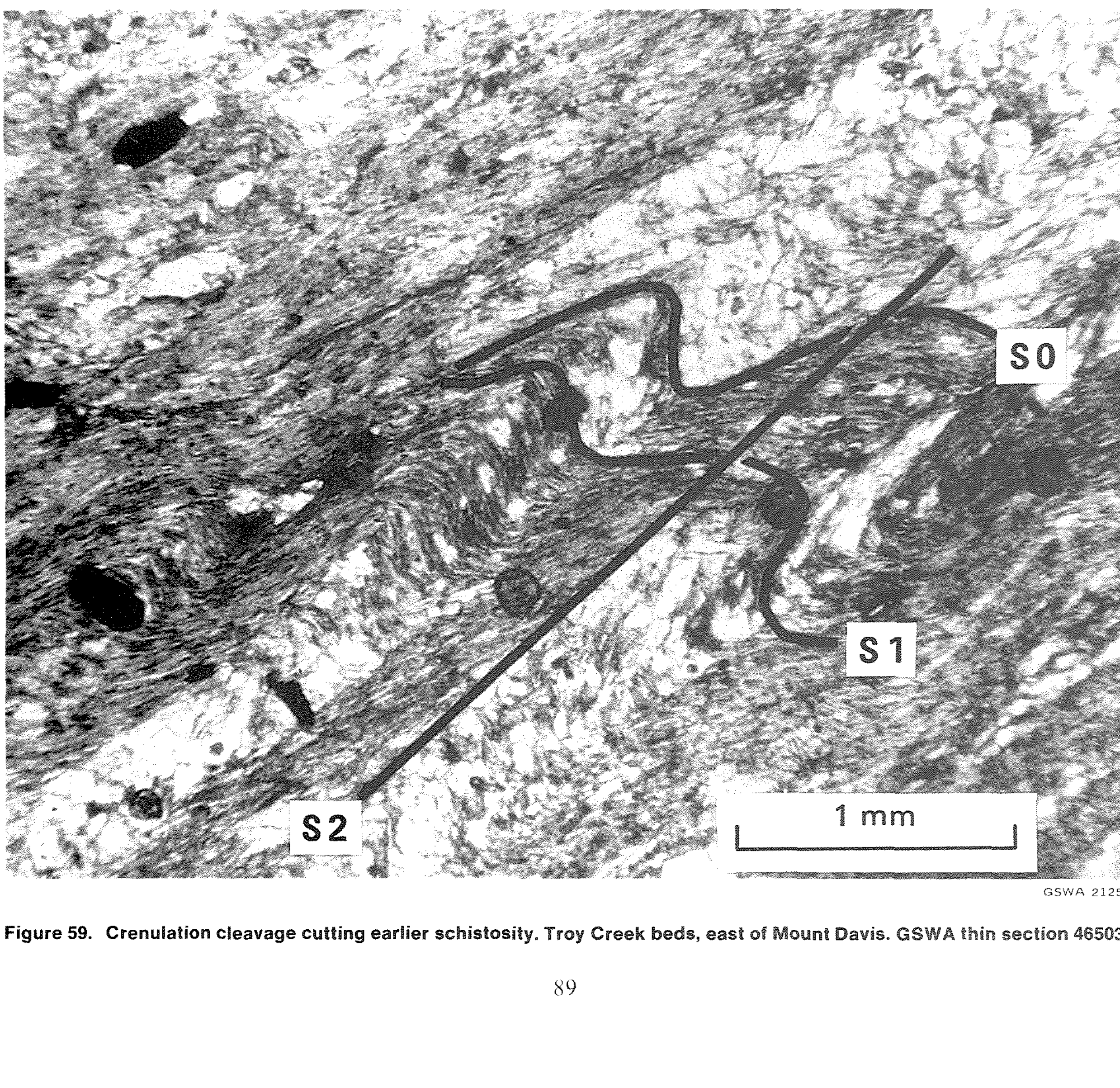


Figure 59. Crenulation cleavage cutting earlier schistosity. Troy Creek beds, east of Mount Davis. GSWA thin section 46503.

some of the northeast-trending faults in the Glengarry Sub-basin are reflected in both the Earaheedy and Bangemall Groups.

LATE FAULTING

The Salvation Fault, in the northeast corner of NABBERU, is a normal fault, downthrown to the north, which brings the Troy Creek beds into contact with the Scorpion Group. It is quite clearly unrelated to the general deformation pattern of the Stanley Fold Belt which involves tectonic transport from north to south. The Scorpion Group immediately adjacent to the fault is a conglomeratic unit which wedges out away from the fault, indicating that sedimentation was related to a growth fault. In places the conglomerate has been dragged into steeply north-dipping attitudes near the fault. As far as can be determined, the fault does not affect the Bangemall Group.

CHAPTER 6

Economic Geology

INTRODUCTION

There has been no recorded mineral production from the Earaheedy Sub-basin, in contrast to the Glengarry Sub-basin which has yielded significant amounts of copper (Thaduna and Horseshoe Lights), gold (Peak Hill and various smaller mines) and manganese (Horseshoe mine). There has, however, been extensive exploration in the Earaheedy Sub-basin, mainly for iron, base metals and uranium. The relevant areas covered by this exploration are shown in Figure 60, while the following account is a summary of that exploration and its findings, together with some speculation on ore-deposit models that may be applicable to the area.

A major hindrance to any potential mineral development in the Earaheedy Sub-basin is its remoteness. The western parts of the area are over 200 km from railheads at Newman, to the northwest, and Meekatharra (now closed) to the southwest, while the railhead at Leonora is over 250 km from the southern edge of the sub-basin. Consequently any mineral deposit would have to be of considerable dimensions and grade to be of economic interest.

IRON ORE

The impetus for modern investigation of the Nabberu Basin came with the recognition that anomalies shown on the regional aeromagnetic maps (compiled by the Australian Bureau of Mineral Resources) were caused by iron-formations (Hall and Goode, 1978). Subsequent exploration covered the Frere Formation from the Lee Steere Range westward to the Miss Fairbairn Hills, around the western closure of the synclinorium and east-southeastwards along the Frere Range.

Although numerous companies were involved in reconnaissance exploration for iron ore, only three—the Broken Hill Proprietary Company (BHP), Amax Exploration and Pacminex Pty Ltd—were granted occupancy rights to Temporary Reserves. Much of the following information is taken from the work of these companies. Only BHP and Amax undertook rotary-percussion drilling. The areas explored are shown in Figure 60.

BHP reported hematite enrichment in two main stratigraphic units—an upper banded iron-formation (the Illagie Iron Formation Member) and a lower peloidal iron-formation (BHP, 1978). The banded iron-formation, which mainly occurs in the area between Hawkins Knob and the western end of the Frere Range, contains patchy hematite-enriched zones containing up to 66% Fe, and generally low phosphorous (<0.06%). The peloidal iron-formation, which is present throughout the Frere Formation, is thickest at Mount Deverell (300 m, which includes interbedded shales), however, surface enrichment of hematite is generally less than in the banded iron-formation (up to 60% Fe). The best result from drilling showed 50% Fe over a nine-metre intersection width in peloidal iron-formation, 5 km south of Miss Fairbairn Hills. In most of the other areas, drilling results indicated minor hematite enrichment with between 25 and 47% Fe.

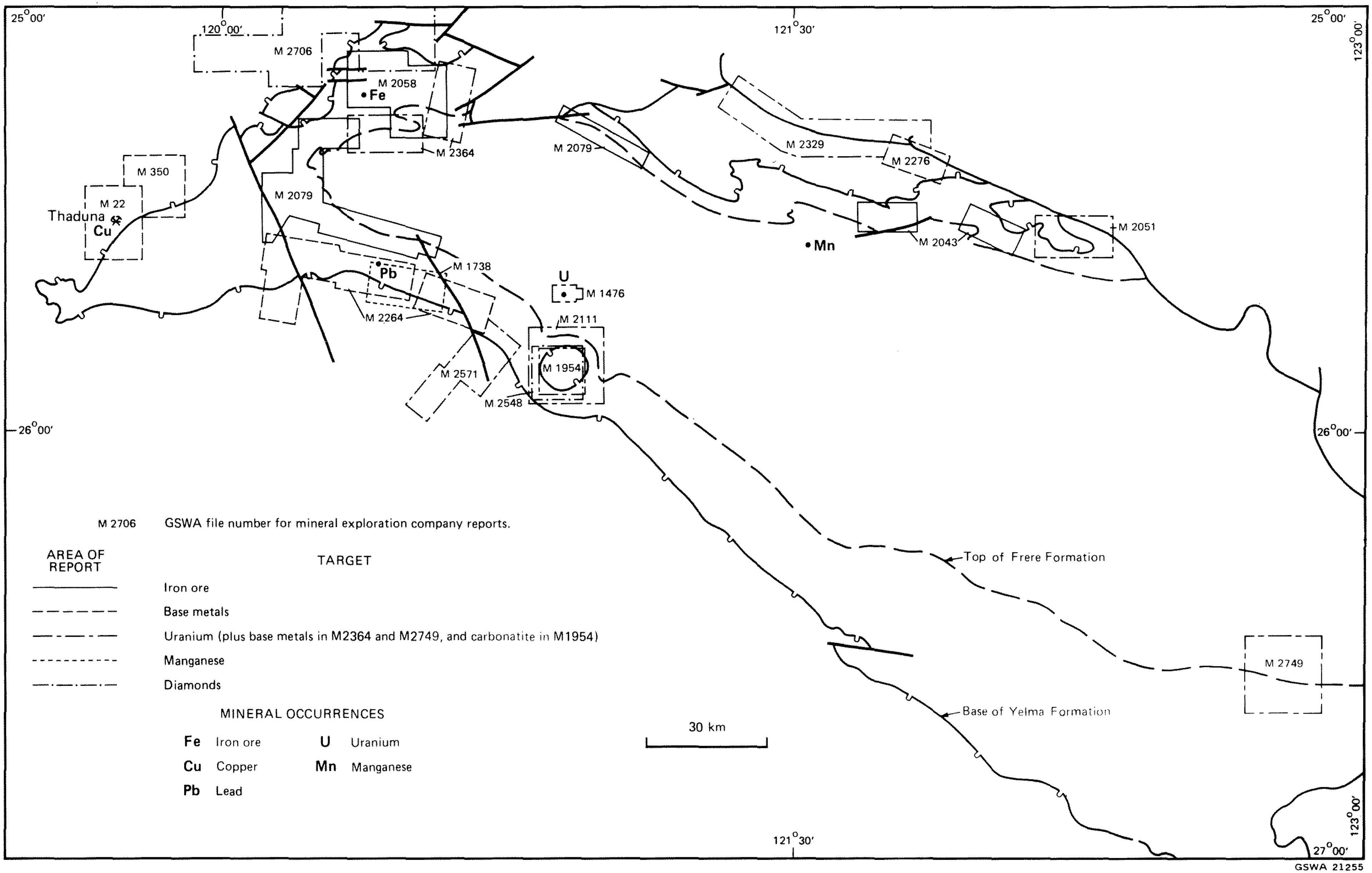
Amax recognized eight zones of hematite/goethite enrichment (Robinson and Gellatly, 1978), of which their Zones 1 and 2 (starting from stratigraphic base) are of limited extent and are not particularly rich. Zones 3 to 8 occur towards the top of the Frere Formation, and it is probable that Zones 5 to 8 correspond to the Illagie Iron Formation Member of BHP. Zone 4 consists of pelletal hematite. Ranging in thickness from 30 to 60 m, it is the thickest and richest of the eight recognized zones, with surface grades around 60% Fe. Drilling of Zone 4 in the Miss Fairbairn Hills revealed intersection widths of 14 m averaging 59% Fe and 8 m averaging 54% Fe.

Results from both BHP and Amax indicate that, with the exception of the Miss Fairbairn Hills area, iron content in zones of hematite/goethite enrichment decreases abruptly with depth. Enrichment is related to the present land surface, which has been subject to Tertiary lateritic weathering, and enrichment is therefore probably Tertiary in age. Parts of the present land surface may, however, be exhumed unconformities upon which are now exposed ancient regoliths. Some patches of primary enrichment in the Miss Fairbairn Hills, close to the Bangemall Unconformity, may be of this origin.

URANIUM

Uranium mineralization has been explored for in several different geological settings, but nothing significant has been found to date.

The most popular exploration target has been vein-type mineralization associated with palaeosurfaces and unconformities. The Nabberu Basin falls into the 1.5-2.0 Ga age bracket in which such deposits are typically found (Robertson and others, 1978). Characteristically these deposits consist of pitchblende, and its alteration products, emplaced


25°00'
120°00'
121°30'
25°00'
123°00'
M 2706
M 2058
Fe
M 2364
M 2079
M 2329
M 2276
M 350
M 22
Thaduna
Cu
M 2079
M 2043
M 2051
Mn
M 1738
Pb
U
M 1476
M 2264
M 2111
M 1954
M 2571
M 2548
26°00'
26°00'
M 2706 GSWA file number for mineral exploration company reports.
AREA OF REPORT
TARGET
Iron ore
Base metals
Uranium (plus base metals in M2364 and M2749, and carbonatite in M1954)
Manganese
Diamonds
MINERAL OCCURRENCES
Fe Iron ore
Cu Copper
Pb Lead
U Uranium
Mn Manganese
Top of Frere Formation
M 2749
Base of Yelma Formation
30 km
121°30'
123°00'
27°00'
GSWA 21255

in open, permeable structures close to or under a protective cover of more impermeable rock. The concentration probably results from supergene processes related to the unconformity.

In Australia, the most favourable comparison is with the uranium deposits of the Pine Creek Geosyncline (described in detail in Ferguson and Goleby, 1980). There, uranium mineralization is mainly contained within Proterozoic schists which have a sedimentational age of about 2.0 Ga and a metamorphic age of about 1.8 Ga, and which were subject to palaeoweathering prior to being overlain by younger Proterozoic sandstone of age about 1.65 Ga. The uranium mineralization is contained exclusively within the schists but is closely related spatially to the unconformity. Alteration including silicification and chloritization, some of which affected the overlying sandstone, was associated with the actual mineralizing event.

Applying these criteria to the eastern part of the Nabberu Basin, the most favourable areas for exploration are located where the Glengarry Group, or the chloritic schists of the Troy Creek beds, are unconformably overlain by either the Earaheedy or the Bangemall Groups. The unconformity between the Earaheedy and Glengarry Groups, west of the Wiluna Arch, is characterized by conspicuous chert breccias that developed upon the unconformity in carbonate-rich rocks, and may have uranium-mineral potential.

A second type of unconformity-related uranium deposit is that associated with basal quartz-pebble conglomerates, taking as a model the Blind River deposits in Canada (Roscoe, 1968). At Blind River, conglomerate, argillite and arenites of age about 2.3 Ga rest on an eroded granitoid basement. In the eastern Nabberu Basin, this model has been applied to exploration over the Malmac Dome (Magnet Metals, 1977), and may have influenced other company exploration along the unconformity between the Yelma Formation and basement west of the Carnarvon Range.

Anomalous uranium values (up to 55 ppm), from sediments overlying the Malmac Dome, were reported by Horwitz and Mann (1975); however, Butt and others (1977) indicate that most of this uranium was due to detrital zircon. Some mature, quartz-pebble conglomerate beds occur in the Earaheedy Group, but have not so far yielded uranium minerals.

Figure 60. Mineral localities and exploration activity, Earaheedy Sub-basin.

A possible difficulty in applying a Blind River model to the basal conglomerates of the Earaheedy Group is that the Earaheedy Group seems to be younger than the critical period in which the uraniferous quartz-pebble conglomerates were formed. No detrital pyrite has been observed, and clearly the Frere Formation was largely deposited in oxidizing conditions. It is interesting to note however that conglomerates at Kaluweerie Hill (Allchurch and Bunting, 1976), 140 km south of Wiluna, and near Mount Yagahong (Elias and others, 1982), 160 km west-southwest of Wiluna, have been tentatively correlated with rocks of the Nabberu Basin, although no specific correlations can be suggested. Both the Mount Yagahong and Kaluweerie Hill occurrences contain a mixture of basement-derived clasts in an immature arkosic or wacke matrix. The Kaluweerie Conglomerate is fluvial, and contains authigenic and possibly detrital pyrite indicating anoxic conditions. Allchurch and Bunting (1976) noted the presence of carnotite in weathered Kaluweerie Conglomerate, and suggested that it may represent Proterozoic placer-type mineralization. It may be, therefore, that conditions suitable for the development of Blind River-style mineralization were present along the southern margin of the Nabberu Basin, but that the resulting rocks are now represented only by a few scattered outliers.

A more specific type of unconformity-related uranium deposit is that associated with circular structures. Uranerz (1978) explored the Teague Ring Structure following a comparison between it and the Carswell Structure in Canada which contains the Cluff Lake uranium deposit. At Cluff Lake, pitchblende occurs in veins in Archaean basement, immediately beneath the unconformity with overlying Proterozoic sediments. The veins are located in a circular feature, the core of which is about 20 km across (Ruzicka, 1975). However, despite superficial similarities, there are differences between the Teague and Carswell structures, notably the presence of quartz syenite and the lack of a regolith and deep weathering beneath the unconformity at Teague. Uranerz reported no anomalous uranium from the Teague structure.

Carnotite has been reported from calcrete, in the Fyfe Bore area, by Esso Minerals (Lindeman, 1972). The carnotite occurs in small, irregular patches as thin coatings in cracks and cavities; an origin similar to that postulated for the Yeelirrie uranium deposit (Mann and Deutscher, 1978) may be invoked. Although the calcrete and mineralization are Tertiary in age, the modern and Tertiary drainage patterns would suggest that the immediate source of the uranium may have been sedimentary rocks of the Earaheedy Group rather than the granitoid rocks of the basement.

BASE METALS

Galena is present in outcrops of stromatolitic dolomite of the Yelma Formation 4 km southeast of Sweetwaters Well. The galena is coarsely crystalline and occurs as small veins, stringers, and cavity fillings within the dolomite. In places it occurs between the columns of small, digitate stromatolites (*Yelma digitata*) as in Figure 61A, and in the core of large domed forms (Fig. 61B). Thin pyrite stringers, now replaced by limonite, also cut the dolomite.

Exploration by the Broken Hill Proprietary Company for base metals covered the area from south of the Frere Range to Mount Clarence and the Verscher Range (Johnston and Hall, 1980). Drilling revealed a sequence of shale, sandstone and dolomite, with minor chert and conglomerate bands, beneath the dolomite outcrop at Sweetwaters Well. No significantly anomalous metal values were obtained. Anomalous lead values of up to 7 000 ppm are regarded as surface concentrations related to recent hydrochemical processes.

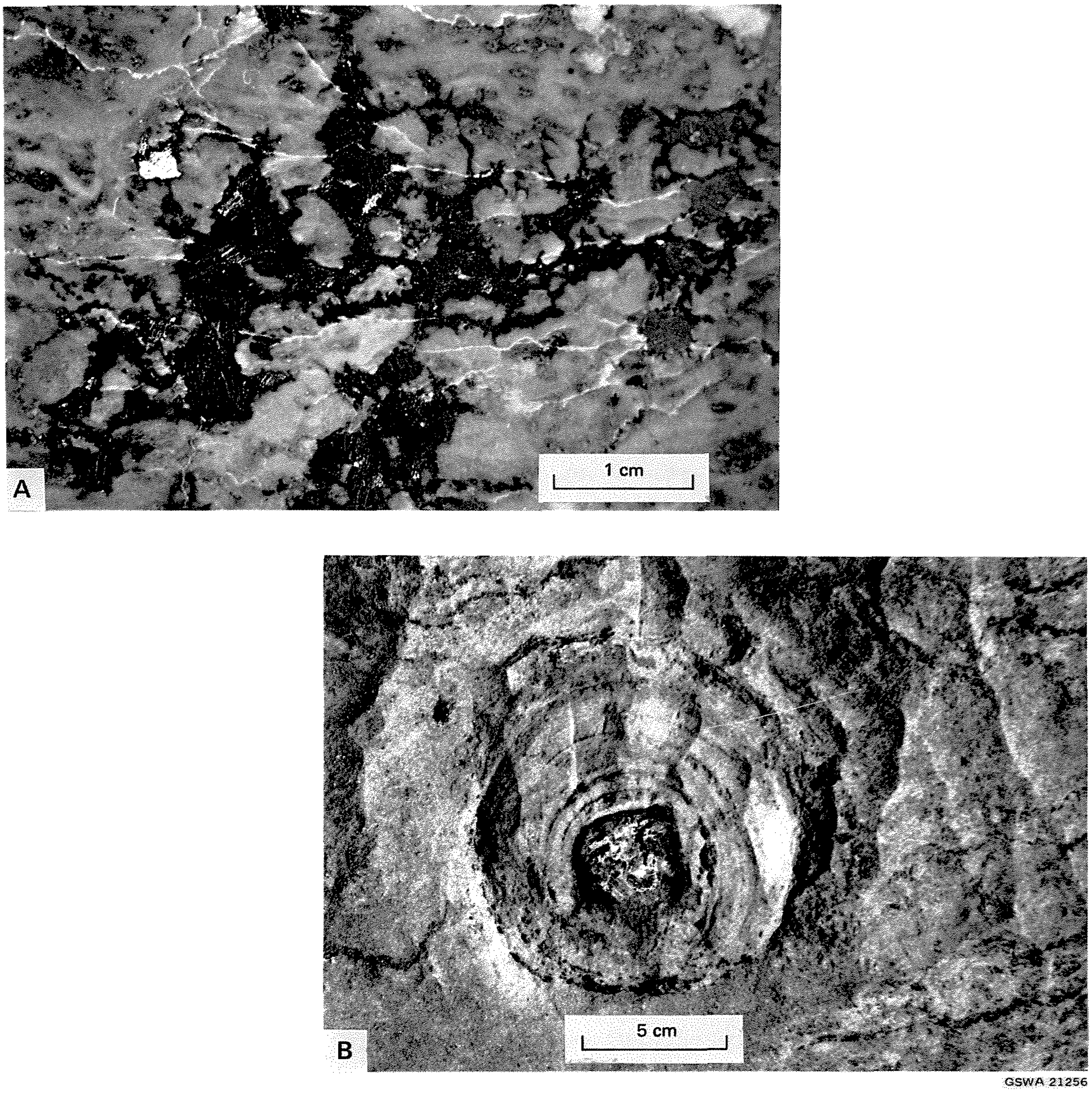


Figure 61. Galena mineralization associated with stromatolites.

A — Galena (dark) infilling interspaces between recrystallized columns of *Yelma digitata* Grey 1984.

B — Cross section of a domed stromatolite with central laminae replaced by galena (dark).

The occurrence of galena in veins and replacement patches in stromatolites suggests that the mineralization is not strictly syngenetic, but was introduced during later diagenesis. Galena occurs in solution cracks, and predates the stylolization of the dolomite. In this respect the galena occurrence resembles a Phanerozoic, carbonate-hosted, lead deposit. However, the association of black shale and dolomite beneath the mineralized outcrop provides a possible indication of stratiform base-metal mineralization, such as occurs at McArthur River (Williams, 1980). Three features of the setting at McArthur River are:

(a) evidence for evaporitic sedimentation to provide connate brine;

(b) presence of an extensive carbonaceous and pyritic shale unit, to provide a reducing environment;

(c) proximity to growth faults, along an active trough margin, which acted as conduits for circulating hydrothermal fluids.

To some extent these features are apparent at Sweetwaters Well. There is a possibility that some of the stromatolites, particularly the smaller, digitate forms, could have grown in an intertidal to supratidal, evaporitic environment (Grey, 1984). The area also lies close to the interpreted buried edge of the Yilgarn Block, which corresponds both to the basin axis, north of which the sediments were laid down in deeper water, and the edge of the Kingston Platform. There is no firm evidence of an active trough margin as at McArthur River.

Another structural feature of possible significance is the position of the Sweetwaters Well mineralization between the Lockeridge and Merrie Range Faults. These faults are reactivations of major tectonic lineaments in the Archaean. A parallel lineament to the west, the Keith-Kilkenny Lineament, exerts a strong control on nickel, and to a lesser extent gold and copper mineralization, in the Archaean rocks. The copper at Thaduna, which occurs in greywackes of the Glengarry Group, lies along the northwest projection of this line, and the mineralization is related to northwest-trending shears. Thus it is possible that the Merrie Range and Lockeridge Faults could also have influenced mineralization.

At other stratigraphic levels in the Earaheedy Group, the application of geological models for base-metal mineralization is less appealing. The Windidda Formation, within which there is evidence for localized evaporitic conditions in a restricted lagoonal environment, is one place where the setting may have been suitable for circulating groundwaters to precipitate base metals, especially near the base and the partly emergent top of the formation. The few samples of Windidda Formation carbonate that have been analyzed show no anomalous concentration of base metals. Hall and Goode (1978) record the presence of minor chalcopyrite and sphalerite from chert at the base of the type section.

North of Sydney Heads Pass, felsic volcanic rocks are interbedded with shale and phyllite of the Troy Creek beds. The volcanics are mainly cream to pale cream tuffs containing prominent quartz crystals. There is a slight potential here for volcanogenic massive sulphides.

MANGANESE

West of Earaheedy homestead, bands of superficial manganese enrichment occur in shale of the Wandiwarra Formation. The bands are controlled by joints and bedding-plane fractures in the steeply dipping sediments. The known deposits are too small to be of economic interest, the largest having a maximum thickness of 8 m and a strike length of some 80 m. Four pooled samples, taken at 2 m intervals across strike, assayed 27% Mn, 2.3% Fe and 36.8% SiO_2.

SUMMARY OF ECONOMIC POTENTIAL

Exploration in the Earaheedy Sub-basin so far has revealed no mineralization of economic significance, and the overall potential must be considered as low. However, a few areas may still warrant attention.

SWEETWATERS WELL AREA

Although parts of this area were explored for base metals in some detail, particularly near the galena mineralization, the area between Sweetwaters Well and the Merrie Range Fault has received only limited attention, especially where Lake Nabberu has created difficulties of access. The zone of interest should extend to the west (downthrow side) of the Merrie Range Fault.

TEAGUE RING STRUCTURE

Although already explored for uranium (see above) and as a potential carbonatite, many questions about the nature and origin of the rocks in the core of the structure remain unresolved. Although the uranium potential must now be regarded as low, the possibility remains that mineralization related to carbonatite, which would be very susceptible to weathering during the Tertiary, could be concealed beneath the numerous salt lakes. Over 90% of the core is not exposed.

INTRA-PROTEROZOIC UNCONFORMITIES

Some potential may still exist for uranium mineralization at the presumed unconformity between the Earaheedy Group and the underlying Troy Creek beds or Glengarry Group, and between the Troy Creek beds and the younger Proterozoic sediments of the Scorpion and Bangemall Groups.

SALVATION FAULT

Although not strictly part of the Nabberu Basin, the conglomerate-arenite facies of the Scorpion Group adjacent to the Salvation Fault (which marks the northern edge of the Troy Creek beds) may have some potential for placer gold/uranium deposits. The Salvation Fault was a growth fault, which was moving during the deposition of the Scorpion Group, and may have formed the southern margin of the depositional basin. Some of the conglomerates are almost certainly fluviatile (possibly alluvial fans), and a few quartz-pebble conglomerates with a matrix of quartz arenite may represent braided streams.

CHAPTER 7

Basin Development

EARLY HISTORY OF THE NABBERU BASIN

The primary location of the entire Nabberu Basin is the junction between the stable Archaean granitoid-greenstone terrain of the Yilgarn Block, and remobilized Archaean terrain to the north. In the western part of the basin, this junction is a continuation of the Murchison Lineament—an east-trending line which marks the boundary between granitoid-greenstone terrain and older gneissic terrain on ROBINSON RANGE (Elias and Williams, 1980). The Yarlarweelor Gneiss Belt and Marymia Dome preserve remnants of this gneissic terrain, now modified in places by deformation, metamorphism and migmatization. The junction between the two types of Archaean terrain passes beneath the Glengarry Sub-basin between the Marymia and Goodin Domes, then turns northeast and crosses the Marymia Dome in northwest NABBERU before disappearing under the Bangemall Group.

Under most of the Nabberu Basin, from the Goodin Dome eastwards, there is a zone in the basement to the south of the gneissic terrain, in which granitoid-greenstone terrain of the Yilgarn Block has been subjected to vertical fault-block movements (Goodin Dome, Malmac Dome and eastern end of Marymia Dome). Internally, these domes have not suffered the plastic deformation that affected the Yarlarweelor Gneiss Belt and western part of Marymia Dome. The southern margin of this zone of block movement is the northern boundary of the Kingston Platform which marks the extent of the underlying stable Yilgarn Block.

Sedimentation in the Nabberu Basin was probably initiated by instability in the Archaean gneissic terrain and the adjacent zone of block movement. It is becoming apparent, both in Western Australia and elsewhere, that many Proterozoic orogenic belts are sited on older Archaean gneissic, rather than granitoid-greenstone terrains (Kröner, 1977; Gee, 1979a), and that many Proterozoic marginal basins are centred at the boundaries of these terrains (Sims, 1976; Dimroth, 1970).

Upward movement of domes within the zone of block movements probably did not begin until after deposition in the basin had commenced. Nevertheless, the zone represents a region of crustal weakness in which a downwarp occurred at an early stage in the basin development.

The land surface over which the early Proterozoic seas transgressed was probably a peneplain (Allchurch and Bunting, 1976; Hall and Goode, 1978; Gee, 1979b), as indicated by the essentially flat nature of the unconformity surface, the lack of thick, coarse-detrital units, and evidence of deep weathering in the source areas (Hall and Goode, 1978). This applies equally well to the Glengarry and Earaheedy Sub-basins where they transgress over the Archaean cratons. The small lenses of conglomerate in the Finlayson Sandstone and Yelma Formation, where they overlie the Yilgarn Craton, are only a few metres thick, and contain only chemically resistant rock types such as vein quartz and chert. These thin conglomerates represent small fluvial channels, or in some cases, wedge-shaped scree deposits flanking low, resistant ridges of vein quartz or banded chert which protruded only a few metres above the plain. The record of the first cycle of erosion from the stabilizing Yilgarn Block is missing in this area, and no analogue of the Fortescue Group on the Pilbara Block is present.

The original extent of this widely distributed shallow-marine blanket is uncertain, and may have covered large areas of the Yilgarn Block, as well as the developing Capricorn Orogen. The appearance of Archaean basement rocks in the Wiluna Arch and the Malmac Dome suggests, however, that the Glengarry Group may not have extended entirely across the area that was later to become the Earaheedy Sub-basin. However, correlation of the Troy Creek beds with the Glengarry Group suggests the presence of a localized trough of Glengarry Group sediments in the area north of the Kingston Platform. The basement upon which the Earaheedy Group was deposited seems to have included both cratonic Archaean basement, with or without earlier sedimentary covers, and belts that already had undergone rifting, sedimentation and deformation.

SEDIMENTATION IN THE EARAHEEDY SUB-BASIN

THE YELMA TRANSGRESSION

The Yelma transgression began in the northern part of the Earaheedy Sub-basin, where the Yelma Formation is thickest over the Troy Creek beds. The sea spread over the basement domes and transgressed the fluviatile sediments that had developed in the area of the Glengarry Sub-basin. Finally it transgressed southeastwards over the extensive peneplain of the

Yilgarn Block, depositing the very thin cover of sand that now forms the Yelma Formation in THROSSELL, DUKETON and southeast KINGSTON.

It is interesting to note here the control that pre-existing structures had on Yelma deposition. The Yelma Formation is no more than 150 m thick along the southern boundary of the sub-basin, and is less than 20 m thick in southeast and central KINGSTON. However, in central NABBERU, south of the Frere Range, it thickens rapidly to the northwest. Hall and Goode (1978) relate this thickening to north-northwest-trending faults (now termed the Merrie Range and Lockeridge Faults) which are reactivations of Archaean structures. However, the formation continues to thicken around the northwestern side of the sub-basin, and eastwards along the northern side, before eventually thinning over the Malmac Dome. Thus it seems more likely that thickening is related to an east-west structure in the basement, rather than the north-northwest structures as proposed by Hall and Goode (1978). The thickening coincides approximately with the inferred position of the boundary between cratonic and mobile basement—a structure that had a profound and long-lasting influence on sedimentation and deformation in the whole Nabberu Basin. This apparent hinge line has implications for the control of possible stratabound lead-zinc mineralization, such as at Sweetwaters Well.

The Yelma Formation, then, began as a strandline deposit which blanketed the entire sub-basin floor. With continued transgression, the upper part of the formation developed as a mud-floored shelf sea, in which localized carbonate shoals and reefs provided some low relief. Some of these reefs may have occurred in a partially emergent, evaporitic environment (e.g. the stromatolitic dolomite at Sweetwaters Well).

The transgressive regime continued into the time of the Frere Formation, during which an extensive shelf sea formed. This received very little detrital input, circulation was restricted, chemical precipitates were able to form, and stromatolite reefs developed. The water depth was probably less than 100 m for most of the basin, and the sea bottom appears to have been disturbed by the occasional storm. Water depth was greatest in the north and northwest parts of the sub-basin, where banded, as distinct from granular, iron-formation occurs near the top of the formation, indicating deposition below wave base. The sub-basin, therefore, seems to have been open in this direction, and the increase in iron content of the Frere Formation in the northwest suggests that the iron was derived from the upwelling of deeper waters. The southern, shallower part of the sub-basin contained low banks and shoals of carbonate and ferruginous cherty ooliths.

THE WINDIDDA REGRESSION

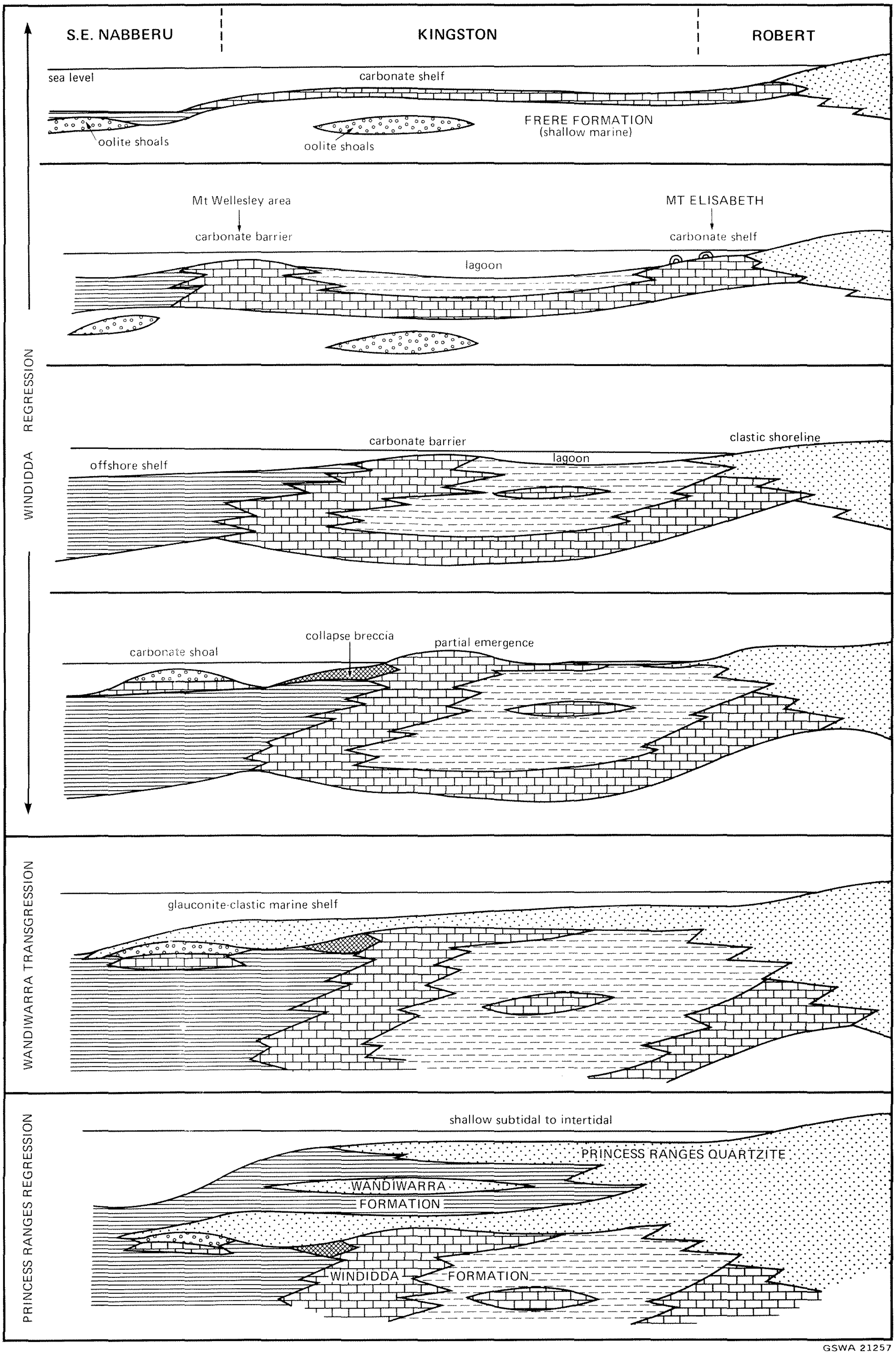
The regressive phase following the Frere Formation mainly affected the southern part of the sub-basin, where lagoonal conditions developed behind extensive carbonate banks (Fig. 62). An evaporitic influence is indicated locally near the base of the Windidda Formation. The lagoonal facies was deposited as a thick sequence of muds and thin carbonates, in very shallow, quiet water that was subjected to occasional storm activity. Seaward of the lagoons, possibly on the offshore side of a carbonate bank, the basin received a small but constant supply of fine-grained detritus, into water that was shallower than that during deposition of the banded iron-formation facies at the top of the Frere Formation. The environment in this northern and western part of the sub-basin (where the regressive phase is included in the Wandiwarra Formation) must have been similar in many respects to the clastic, shaly parts of the Frere Formation in southeast KINGSTON and ROBERT. The landward side of the lagoon was probably on ROBERT, where a sandstone facies may be laterally equivalent to the Windidda Formation and top of the Frere Formation. The sandstone is probably of shallow-marine origin (Jackson, 1978) and could be a nearshore or strandline deposit (the right-hand side of Fig. 62).

The culmination of the regressive phase of the Windidda Formation was partial emergence accompanied by subaerial brecciation and vadose encrustation. The emergence seems to have occurred only in the southeastern part of the sub-basin.

THE WANDIWARRA TRANSGRESSION

The Wandiwarra transgression marks the beginning of the second major transgressive-regressive cycle in the Earaheedy Group, and as such defines the base of the Miningarra Subgroup. The transgression is most apparent where the Wandiwarra Formation overlies the Windidda Formation. It marks the rapid change to coarse clastic sedimentation after the carbonate-mudstone facies of the Windidda Formation. Where the Windidda Formation is absent, the transgression is marked by sandstone lenses within shale of the Wandiwarra Formation.

Figure 62. Facies evolution and environments during the Windidda regression, Wandiwarra transgression and Princess Ranges regression.


S.E. NABBERU
KINGSTON
ROBERT
sea level
carbonate shelf
FRERE FORMATION
(shallow marine)
oolite shoals
oolite shoals
Mt Wellesley area
carbonate barrier
MT ELISABETH
carbonate shelf
lagoon
carbonate barrier
clastic shoreline
lagoon
offshore shelf
WINDIDDA REGRESSION
collapse breccia
partial emergence
carbonate shoal
glauconite-clastic marine shelf
WANDIWARRA TRANSGRESSION
shallow subtidal to intertidal
PRINCESS RANGES QUARTZITE
WANDIWARRA
FORMATION
WINDIDDA FORMATION
PRINCESS RANGES REGRESSION
GSWA 21257

The combination of a transgressive regime and an increased clastic input implies increased slopes in order to provide a means of transporting the clastic material from the source area. Figure 62 illustrates a possible scheme for achieving this, whereby the regressive arenite phase on ROBERT continues westwards over the newly inundated Windidda Formation, but becomes transgressive due to active tectonic sinking (epeirogenic?) of the basin. In effect, the sinking allowed an increased clastic input, but the rate of sinking was greater than the rate of sediment accumulation. Consequently the eastern limit of the transgressive phase was in eastern KINGSTON, where the mature sands of the Princess Ranges Quartzite form a shoreline deposit which is laterally equivalent to the basal transgressive Wandiwarra Formation.

The Princess Ranges Quartzite eventually spreads westwards and northwards as a prograding mature clastic sequence, overlying the deeper water shales at the top of the Wandiwarra. Whereas the Wandiwarra transgression was a rapid event, the partial regression of the Princess Ranges Quartzite was slow, and produced a markedly diachronous boundary. This contrast in timing of events may be due to the transgressive phase being of tectonic origin, whereas the regression was due to the slow filling of the basin.

The Princess Ranges Quartzite represents several shallow-water prograding environments, including possible deltaic deposits in the vicinity of the Teague Ring Structure, and barrier bar, shallow-subtidal and intertidal shelf deposits. Several local phases of regression and transgression may be present.

FILLING OF THE SUB-BASIN

The Wongawol Formation saw another temporary barred-basin environment, either in very extensive lagoons or a partly closed inland sea. Certainly there was a lack of tidal and strong-current influences. Sedimentation was rapid and in very shallow water.

Deeper water, possibly with some tidal influence (implying once again access to open sea) characterizes the lower part of the Kulele Limestone, but this was very short-lived, and shallow-water regressive environments in the upper part of the Kulele Limestone and in the Mulgarra Sandstone may herald the final filling of the sub-basin.

PALAEOGEOGRAPHY OF THE EARAHEEDY SUB-BASIN

The palaeogeographic evolution of the basin has been discussed both in the preceding sections on basin development and in the depositional environment sections of the various stratigraphic descriptions. One theme keeps recurring, and this concerns the direction of the palaeoslope and the position and direction of the shoreline. It is tempting to suggest that the Yilgarn Block, and in particular the Wiluna Arch, was the source of much of the sediment in the Earaheedy Sub-basin. However, several lines of evidence, encompassing almost the whole stratigraphic section, indicate that the coastline was to the southeast, not the southwest. This evidence, which is mostly drawn from Chapter 3 (Earaheedy Group stratigraphy) includes:

(a) the regional thickness variations of the Yelma Formation;

(b) the proportion of clastic material, in the Frere Formation, which increases to the southeast, presumably towards the source area;

(c) the distribution of deep-water iron-formation to the northwest and shoaling iron-formation to the southeast;

(d) facies distribution in the Windidda Formation (including the possible clastic shoreline equivalent on ROBERT);

(e) the presence of the Windidda-Wandiwarra disconformity only in the southeast part of the basin;

(f) the northwestwards progradation of the Princess Ranges Quartzite;

(g) palaeocurrent directions in the Princess Ranges Quartzite;

(h) lithofacies distribution in the Wongawol Formation, whereby the coarser, sandy, shallow-water facies is in the southeast, and deeper water shales predominate to the north and southwest;

(i) inferences from stromatolite elongation in the Kulele Limestone.

Thus there is a large body of evidence which indicates that, throughout the deposition of the Earaheedy Group, deeper water prevailed in the northwest and north and shallower water in the south and east (Fig. 63). The strike of the palaeoslope therefore would have been somewhere between east-west and northeast-southwest. This is at a high angle to the present regional strike of the southern side of the basin (the passive side) and suggests that the features controlling the present configuration, namely

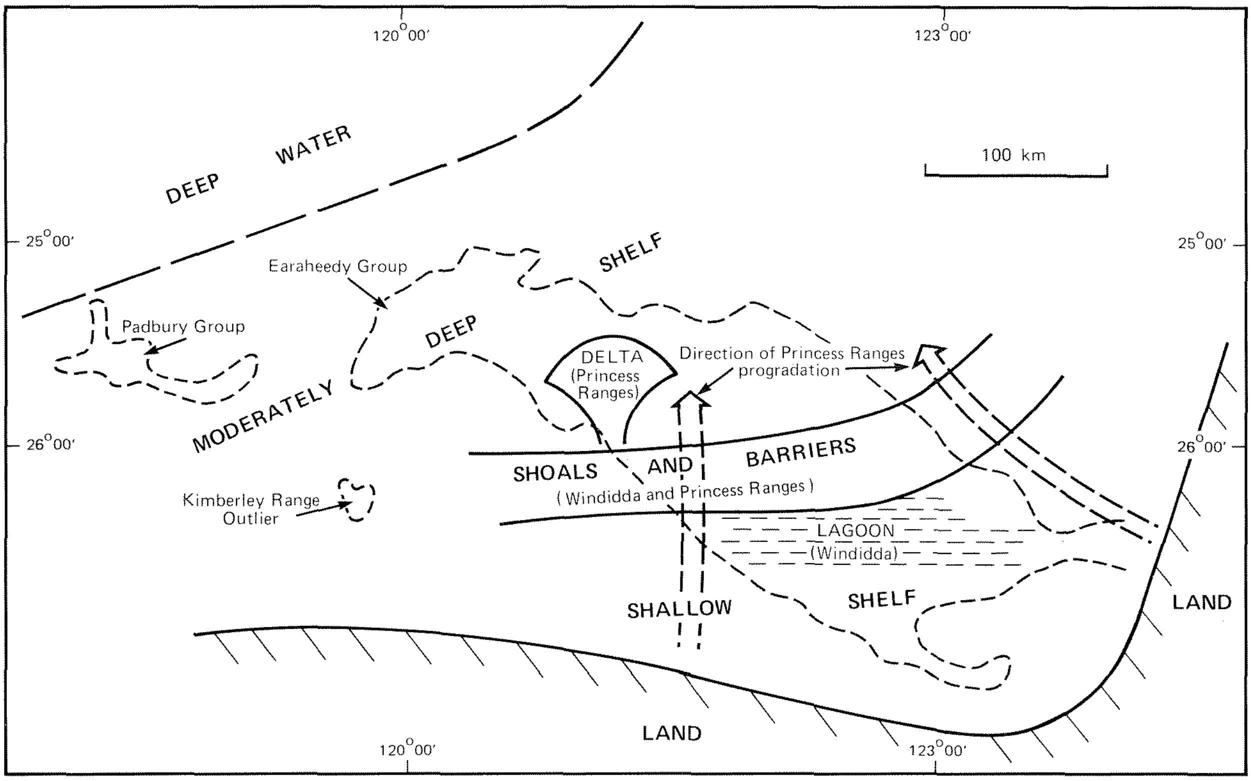


Figure 63. Generalized palaeogeography of the Earaheedy Sub-basin.

the Wiluna Arch and the main synclinorium, are tectonic features unrelated to the depositional shape of the basin.

A further significant conclusion is that there must have been land in the area now occupied by the Officer Basin, and that the shoreline was not too far from the eastern side of ROBERT.

REGIONAL TECTONIC IMPLICATIONS

The Capricorn Orogen is a major orogenic zone involving geosynclinal sedimentation, metamorphism, basement reworking, and granitoid emplacement, in the region between the Yilgarn Block and the Pilbara Craton (Gee, 1979a). In addition to the Nabberu Basin, it includes the geosynclinal and shelf sediments of the Ashburton Trough, the metamorphosed sediments, remobilized Archaean basement and granitoid rocks of the Gascoyne Province, and the Early Proterozoic basement rocks beneath the Bangemall Basin. Citing evidence for the correlation of the trough sequences in the Glengarry Sub-basin and Ashburton Trough, and for the nature of the basement beneath the western Bangemall Group, Gee (1979a) postulates that an elongate belt of thick greywacke and volcanic fill, of geosynclinal dimensions, occupies almost the entire area between the Pilbara and Yilgarn Cratons. The boundaries of this geosynclinal trough are hinge lines along the buried edges of the cratons.

From evidence presented earlier in this report, it is clear that the southern boundary of the Capricorn Orogen is not a simple suture, but a complex zone of elongate fault blocks in the basement to the Nabberu Basin. The early development of this southern boundary was tensional rifting which created a series of horsts and grabens, with an overall effect of subsidence towards the centre of the orogen to the northwest. The amount of spreading associated with these rifts could not have been great—McElhinny and Embleton (1976) present evidence, from polar-wander curves, which indicates that there has been no large-scale movement between the various parts of the Western Australian Shield for at least the last 2.5 Ga. There is no evidence in the Nabberu Basin in favour of large-scale spreading of oceanic crust with subsequent closure to form a collisional suture. There is evidence, from within the Nabberu Basin, and from elsewhere between the Yilgarn and Pilbara Blocks (Brakel and Muhling, 1976; Horwitz and Smith, 1978; Gee, 1979a) that the development of the geosyncline was essentially ensialic.

However, the statistical error limitations on the palaeomagnetic constraints of McElhinny and Embleton (1976) do allow for movement of up to about one hundred kilometres. It is feasible, therefore, that a limited amount of spreading followed the rifting, and that during this phase the basalts of the Glengarry Group were extruded to overlap onto the earlier sediments. The volume of basalt and associated dolerite decreases to the south (towards the Yilgarn Block) and east (towards the Earaheedy Sub-basin). Further east, the Troy Creek beds, which are correlated with the Glengarry Group, contain neither basalt nor dolerite. The implication is, therefore, that the amount of tensional opening increased from east to west along the axis of the Nabberu Basin.

The Earaheedy Group is notable for its lack of contemporaneous volcanism and paucity of dolerite intrusions. The few dolerite sills that are present towards the eastern edge of the sub-basin are probably related to voluminous intrusions in the overlying Bangemall Group. There is no evidence, therefore, of crustal tension, and the margins of this part of the basin were either passive (on the Yilgarn Block) or were related to vertical reactivation of pre-existing basement faults. Clear sedimentological evidence from the Earaheedy Group favours deepening of the basinal waters to the northwest (beyond the present outcrop limits) in the direction of the main area of the Capricorn Orogen. The presence of a more actively sinking trough in that region can also be inferred.

To the northeast of the Earaheedy Sub-basin, the eastern part of the Bangemall Basin has a sedimentational and structural pattern which implies a much more stable basement than does the pattern in the western Bangemall Basin (Brakel and Muhling, 1976; Muhling and Brakel, 1985). A possibility is that this basement consists of Archaean granitoid-greenstone terrain, similar to the eastern Marymia Dome and northeastern Yilgarn Block, rather than Archaean gneiss terrain which is more susceptible to reactivation in Proterozoic mobile belts as in the Gascoyne Province (Gee, 1979a). If this is the case, then the Earaheedy Group was deposited in an arm of the Capricorn Orogen which was bounded to the north and south by stable cratonic blocks.

In both the Glengarry and Earaheedy Sub-basins, the deformation that followed sedimentation was essentially compressional, with basement blocks in the north moving up and southwards.

The overall picture, then, is one of the incipient continental rift, open to the west, which failed to develop—an aulacogen, related to a more extensive and successful oceanic basin which perhaps existed to the west of the Gascoyne Province. The situation is comparable to that in aulacogens associated with the Wopmay Orogen in the Canadian Shield (Hoffman, 1973, 1980), where not only is the tectonic setting very similar to the Nabberu Basin, but the age, styles of sedimentation and interpreted environments are also comparable. Similar comparisons can be made between the Nabberu Basin and the Proterozoic basins surrounding the Superior Province in North America (Sims, 1976; Larue, 1981; Baragar and Scoates, 1981; Dimroth, 1981). In particular, the pattern of sedimentation in platform and basinal settings described by Larue (1981) from the southern Lake Superior area, bears a striking resemblance to the development of the Nabberu Basin described here.

It has become apparent that this style of incipient but unsuccessful cratonic breakup was a common feature of this period of crustal development, which marked an intermediate stage between the transient and constantly recycled crust of the Archaean, and the continental oceanic crustal decoupling that characterizes Phanerozoic plate tectonics (Hargraves, 1981; Kröner, 1981).

Appendices

APPENDIX I

STROMATOLITES AND BIOGENIC ACTIVITY IN THE NABBERU BASIN

by Kathleen Grey

INTRODUCTION

A study of fossil material from the Nabberu Basin was undertaken in conjunction with the program of regional mapping at 1:250 000 scale. A diverse assemblage is present in the Earaheedy Group, and particular emphasis was placed on the biostratigraphic potential of the stromatolites. Microfossils were first reported from oncolites in the Frere Formation (Walter and others, 1976) and have since been found in several other formations (M. R. Walter, 1979, pers. comm.), they are not preserved in the stromatolites examined in this study, but a few filaments were observed in oncolites from near Camel Well.

PREVIOUS STUDIES

The earliest record of stromatolites in the Nabberu Basin was a description by Edgell (1964) of material collected by K. H. Morgan. Edgell identified the specimens as *Collenia undosa* Walcott. During regional mapping of the Officer Basin, stromatolites were collected from outlying outcrops of the Earaheedy Group and four new taxa were described by Preiss (1976b)—*Minjaria granulosa*, cf. *?Kulparia* f. indet., *Tungussia heterostroma*, and *Tarioufetia yilgarnia*.

Geologists of the Broken Hill Proprietary Co. Ltd (B.H.P.) began mapping parts of the Nabberu Basin in 1973, and in 1975 the Geological Survey began systematic regional mapping of the whole basin. During the course of these studies, approximately 20 new stromatolite localities were discovered. Brief descriptions of some of these localities have been published by Hall and Goode (1978), and others are shown on the relevant 1:250 000 geological sheets. Bunting (in this bulletin) gives a detailed description of the large, elongate domes present in the Kulele Limestone, and presents a palaeogeographical reconstruction based on the direction of elongation. A review of stratigraphic occurrences throughout the Western Australian Proterozoic, including those from the Nabberu Basin, has recently been published (Grey, 1981) and systematic studies of forms present in the Earaheedy Group have now been completed (Grey, 1984).

METHODS OF STUDY

Stromatolite samples, other than those described by Edgell (1964) and Preiss (1976b), were collected by J. A. Bunting and R. D. Gee of the G.S.W.A. during regional mapping of the Nabberu Basin between 1976 and 1979, W. D. M. Hall of B.H.P., or by the author in conjunction with M. R. Walter and S. M. Awramik in 1977. Methods of study, both field and laboratory, are basically those outlined by Walter (1972) and Preiss (1976a). The literature on stromatolitic carbonates is extensive, but comprehensive bibliographies have been published by Awramik and others (1976), and Awramik and others (1979).

COMPOSITION OF THE STROMATOLITE ASSEMBLAGE

In the Earaheedy Group, stromatolites are present in several formations. Small, complexly branching, bushy forms are common; larger branching forms occur at some localities, and domed and stratiform stromatolites are present at many horizons. The following forms have been recognized:

(a) *Carnegia wongawolensis* Grey 1984 (Figs 26 and 65C) is a common form in the Windidda Formation, and usually occurs as small, weathered-out nodules formed by complexly branched, irregular columns with a filmy microstructure.

(b) *Earaheedia kuleliensis* Grey 1984 (Figs 50 and 66D) is the only form described so far from the Kulele Limestone. Columns are broad and tend to be vertical, branching is usually parallel, and the form is characterized by a distinctive wavy lamination. The columns are grouped into domes which are frequently elongate. The interstitial material is usually composed of intraclasts and large oolites.

(c) *Ephyaltes* form indet. (Fig. 65B) is known from only one locality in the Earaheedy Group, from the Yelma Formation. It has not been possible to identify the taxon to form level because only one small sample is available. There are major differences between the Yelma specimen and conical stromatolites, which occur in the overlying

Scorpion Group, and also with *Conophyton garganicum australe* Walter which occurs in the Bangemall Group.

(d) *Externia yilgarnia* (Preiss) (Figs 8 and 66C) was originally reported to occur in the Frere Formation (Preiss, 1976b), but more recently the locality has been placed in the Yelma Formation (Bunting, in this bulletin). The form is characterized by a very distinctive, tufted microstructure, which differs considerably from the type form of the group *Tarioufetia*, to which it was assigned by Preiss (1976b), but closely resembles that of *Externia* Semikhatov 1978, (see Grey, 1984).

(e) The specimen described as cf. *?Kulparia* f. indet. Preiss is known from only one locality in the Windidda Formation, and because of the small size of the sample and poor preservation, could only be tentatively identified by Preiss (1976b). No additional material has been obtained. The taxon is characterized by straight, narrow columns with numerous peaks.

(f) *Murgurra nabberuensis* Grey 1984 (Fig. 64A) is a small, bushy form with very complex branching and a filmy microstructure. It occurs in the Yelma Formation.

(g) *Nabberubia toolooensis* Grey 1984 is a microcolumnar or nodular form in the Windidda Formation. It consists of numerous small wart-like projections which encrust other stromatolite forms.

(h) *Omachtenia teagiana* Grey 1984 (Fig. 65A) is a fairly large, branching form with a pseudocolumnar base, a banded microstructure, and short patches of wall formed by a single lamina. This form occurs at only one locality in the Yelma Formation, and may be present at a second, in the Frere Formation.

(i) *Pilbaria deverella* Grey 1984 (Fig. 64C) occurs at several localities in both the Yelma and the Frere Formations. It can be readily distinguished by the presence of very large niches and an irregularly streaky microstructure. It is usually present at the same localities as *Yelma digitata* and forms part of a cyclic sequence.

(j) *Windidda granulosa* (Preiss) Grey 1984 includes both *Minjaria granulosa* Preiss and *Tungussia heterostroma* Preiss, which are the upper and basal parts respectively of a single stromatolite with granular microstructure and multilaminate walls. Additional material has indicated that smooth-walled, parallel-branching columns arise from irregular, divergent branches, which occur at the base of the structure, and the new group *Windidda* Grey 1984 has been erected for stromatolites with this particular branching habit.

(k) *Yandilla meekatharrensis* Grey 1984 (Fig. 66A) occurs at a single locality in what is probably Yelma Formation. Columns are medium sized and have irregular margins, and microstructure is vermiform. Preservation is poor as a result of secondary alteration.

(l) *Yelma digitata* Grey 1984 (Fig. 64D) is a common form in the Yelma Formation, and is probably also present in the Frere Formation. It is a small, bushy form with long, narrow, parallel columns and a banded microstructure. It is commonly found in the upper part of a cyclic sequence which includes *Pilbaria deverella*.

STRATIGRAPHIC DISTRIBUTION

The stratigraphic distribution of stromatolites in the Earaheedy Group of the Earaheedy Sub-basin is summarized in Table 8.

YELMA FORMATION

Carbonate lenses in the dominantly clastic Yelma Formation have yielded seven taxa—*Ephyaltes* f. indet., *Omachtenia teagiana, Pilbaria deverella, Yandilla meekatharrensis, Murgurra nabberuensis, Externia yilgarnia* and *Yelma digitata*.

The northeast DUKETON locality with *Externia yilgarnia* has been described in detail by Preiss (1976b). The stromatolites are abundant in a carbonate band 3-5 m thick. Near Sweeney Creek, *Omachtenia teagiana* occurs in a partially silicified outcrop of grey limestone. One of the most interesting localities is near Sweetwaters Well, where stromatolites are abundant in a laminated, pale-grey dolomite. Five types have been recognized (Grey, 1984): *Murgurra nabberuensis, Pilbaria deverella, Yelma digitata,* and domed and stratiform stromatolites.

Preliminary work indicates that four of these types occur in a series of regressive cycles in which the most probable sequence is: domed stromatolites (Fig. 64B) overlain by *Pilbaria deverella, Yelma digitata* and, finally, stratiform stromatolites. More detailed studies are required to determine the precise relationships of the various components of the cycles.

TABLE 8. STRATIGRAPHIC RANGES OF STROMATOLITE TAXA IN THE EARAHEEDY GROUP.

STROMATOLITE TAXA	STRATIGRAPHIC DISTRIBUTION							
	Yelma Formation	Frere Formation	Windidda Formation	Wandiwarra Formation	Princess Ranges Quartzite	Wongawol Formation	Kulele Limestone	Mulgarra Formation
Earaheedia kuleliensis Grey 1984							▬	
Carnegia wongawolensis Grey 1984			▬					
Nabberubia toolooensis Grey 1984			▬					
Windidda granulosa (Preiss 1976)			▬					
cf. ? *Kulparia* Preiss 1976			▬					
Pilbaria deverella Grey 1984	▬	▬						
Yelma digitata Grey 1984	▬	▬						
Ephyaltes Grey 1984	?▬							
Externia yilgarnia (Preiss 1976)	▬							
Murgurra nabberuensis Grey 1984	▬							
Omachtenia teagiana Grey 1984	▬							
Yandilla meekatharrensis Grey 1984	▬							

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Poorly preserved specimens of *Yelma digitata* occur at a locality near Combine Well, and *Pilbaria deverella* is probably present in a dark-grey limestone at a locality near Lake Gregory. Northwest of Edingunna Well, a dark-grey dolomite contains *Ephyaltes* f. indet. and *?Yelma digitata*. A thick sequence of stromatolitic dolomite, probably in the Yelma Formation, outcrops approximately 5 km east of Phar Lap Well, as a component of the Kimberley Range Outlier, and contains *Yandilla meekatharrensis*.

FRERE FORMATION

The Frere Formation consists predominantly of ferruginous shale and granular iron-formation with minor chert and carbonate. Branching columnar stromatolites occur in the carbonate horizons but have not been reported from the cherts or iron-formation. The latter, however, contains stratiform stromatolites and oncolites. Walter and others (1976) reported abundant, well-preserved microfossils, some of which resemble modern iron-bacteria, from an oncolite-rich horizon near Camel Well. A few poorly preserved filaments (Fig. 67) were found during the present study.

Pilbaria deverella occurs in a pale-grey to white carbonate near Simpson Well, and specimens tentatively identified as *Yelma digitata* are known from south of Lake Carnegie.

WINDIDDA FORMATION

The Windidda Formation is grey-to-pink, laminated dolomite or limestone, interbedded with maroon or grey mudstone, and contains numerous stromatolites. Preiss (1976b) described *Windidda granulosa* (as *Minjaria granulosa* and *Tungussia heterostroma*) and cf. *?Kulparia* f. indet. from Mount Elisabeth.

In the lower and middle part of the Windidda Formation, *Carnegia wongawolensis* is abundant and occurs as weathered-out nodules. It is particularly common from the gorge section near Tooloo Bluff, and in Wongawol Creek. *Nabberubia toolooensis* is a rare form, but is known from Mount Elisabeth and the gorge near Tooloo Bluff.

WONGAWOL FORMATION

Poorly preserved stromatolites occur in this formation (Bunting, in this bulletin).

KULELE LIMESTONE

Stromatolites are abundant in the Kulele Limestone, although the predominant type is a large, domed form which has not been named. Smaller domes are formed by the branching-columnar *Earaheedia kuleliensis*. This form is common near Thurraguddy Bore, and a detailed description of the locality is given by Bunting (in this bulletin). Unidentified stromatolites are also present at Mount Lancelot and Mount Hosken.

STROMATOLITE BIOSTRATIGRAPHY

The significance of the Earaheedy Group stromatolites to both Australian and world-wide Precambrian biostratigraphy has been discussed in some detail by Grey (1984). This study shows that a diverse assemblage, with restricted stratigraphic distribution, occurs in the Earaheedy Group (Table 8). The extent to which the distribution is controlled by major facies changes is uncertain. Nevertheless, the stromatolites provide a useful tool for the recognition of lithostratigraphic units within the basin.

Comparisons with other Australian and overseas assemblages have been made by Grey (1984) and show that the forms recognized in the Earaheedy Group have not been recorded from other parts of Australia. However, none of the stromatolite-bearing units are considered to be coeval with the sequence in the Earaheedy Sub-basin. Few studies of early and middle Proterozoic stromatolites from other parts of the world have been published, but the limited descriptions available suggest that similar forms may occur in other units of approximately the same age, particularly in Canada (Semikhatov, 1978).

ENVIRONMENT OF DEPOSITION

Stromatolites have frequently been used as environmental indicators in studies of sedimentation and basin analysis (for example, see papers by several authors in Ginsburg, 1975 and Walter, 1976). Direct comparisons between ancient and modern stromatolite environments must be approached with caution (Walter, 1977) because of the very specialized environments which support stromatolite growth today. In Proterozoic times, stromatolites were much more widespread and occupied a wider range of ecological niches. A major problem is the fact that complex, divergently branching columns have not been reported from recent environments, and there is some doubt about whether such forms represent subtidal or intertidal conditions. However, studies of the Malmani Dolomite (age 2 250 Ma) by Eriksson and others (1976) and of the Pethei Group (age 1 795-1 865 Ma) by Hoffman (1976) suggest that they are subtidal. Conical stromatolites almost certainly indicate a subaqueous environment Donaldson, 1976). The carbonate lenses in the Yelma and Frere Formations which contain *Ephyaltes*, and a variety of divergently branching stromatolites, probably formed as lagoonal phases in a predominantly clastic shelf environment. At Sweetwaters Well, *Pilbaria* and *Yelma* occur together in probable upward-shallowing sequences, with *Pilbaria* in lagoonal phases and *Yelma* in intertidal to supratidal phases.

Some of the stromatolites in the Windidda Formation appear similar to those occurring in recent freshwater algal-marshes (Monty and Hardie, 1976) but further work is necessary to confirm this. The domes and short columns which occur in the Kulele Limestone are indicative of an intertidal to shallow-subtidal regime.

STROMATOLITES AND MINERALIZATION

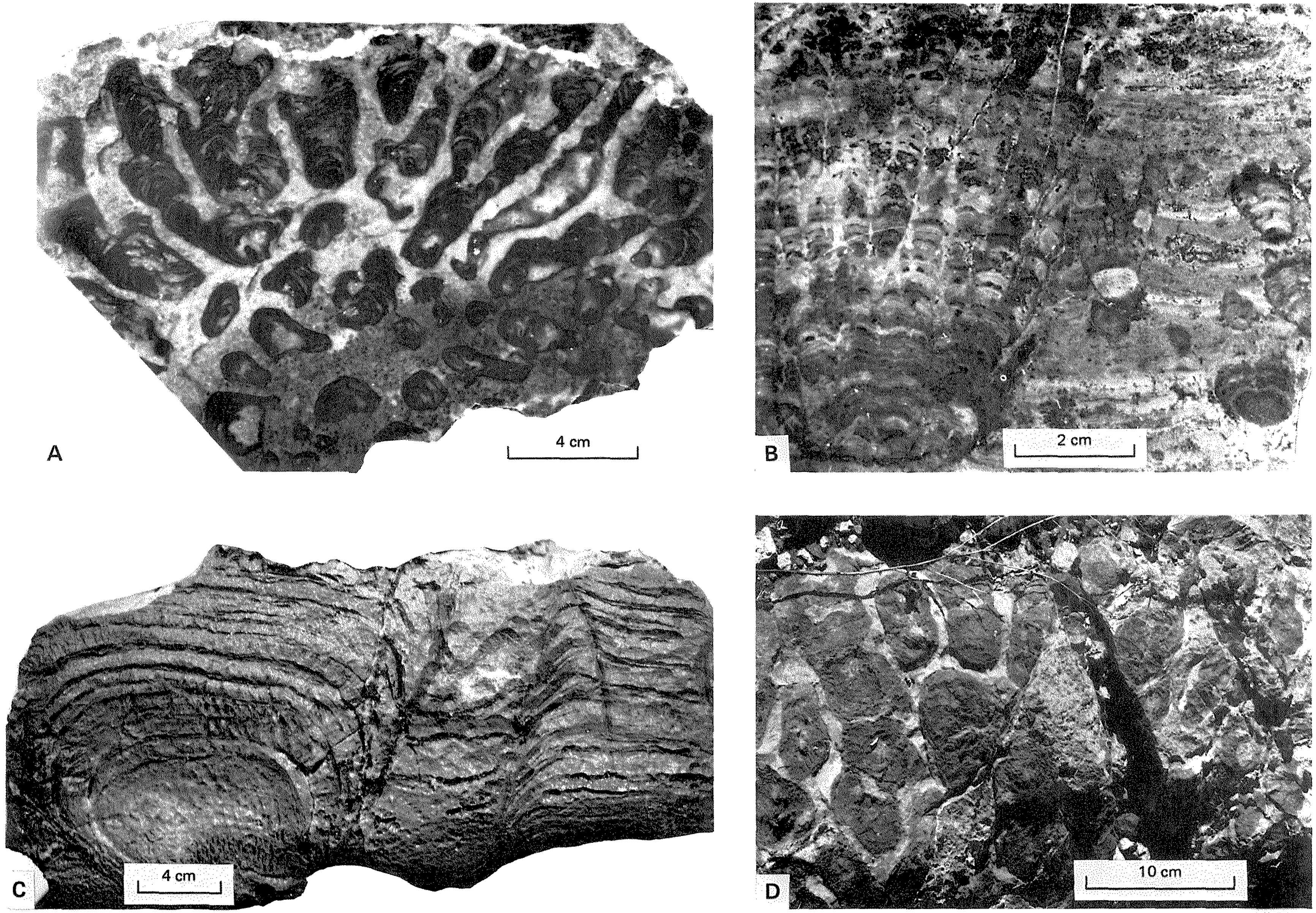
In recent years there has been increasing interest in the role of stromatolites, and associated microbial organisms, in the formation of stratabound ore deposits. A summary of associations recognized so far, and of the possible mechanisms involved, is given by Mendelsohn (1976) and Trudinger and Mendelsohn (1976). The trapping and binding function of a microbial mat may concentrate a mineral, which is present in the sediment supply, or some organisms may extract minerals from solution and precipitate them as sheaths. Furthermore, the decomposition of a thick microbial mat may provide a source of hydrogen sulphide. These mechanisms are the subject of much current investigation.

Detailed studies of the association between mineralization and biogenic activity have not been carried out in the Earaheedy Sub-basin, although several examples have been recognized which could form the basis of future studies. Of particular interest is the presence of bacteria in the Frere Formation (Walter and others, 1976). Some of these could have played an active role in the precipitation of iron-rich oncolites (Glaessner and Walter, 1981), and may even have controlled the deposition of much of the iron present in the formation. A second example is the association between stromatolites and galena near Sweetwaters Well, where the lead occurs in column interspaces and its distribution is apparently controlled by the higher porosity of the inter-columnar areas.

CONCLUSIONS

The diverse stromatolite assemblage present in the Earaheedy Group is significant for several reasons. Stromatolites and micro-organisms played a major role in the deposition of the various stratigraphic units, and possibly influenced mineralization in some parts of the sub-basin. The stromatolites provide useful lithostratigraphic markers, and are of significance for environmental interpretation.

Figure 64. Stromatolites occurring in cyclic sequences near Sweetwaters Well. A — *Murgurra nabberuensis* Grey 1984; polished face showing bushy branching. The relationship to the cyclic sequence is uncertain. B — Stromatolite dome, typical of the types occurring at the base of a cycle. C — *Pilbaria deverella* Grey 1984; cross sections of columns, usually occurring above domes (64B) and below *Yelma digitata* (64D). D — *Yelma digitata* Grey 1984; typical branching pattern and lamination development, usually occurring above *Pilbaria deverella* (64C) in each cycle.



GSWA 21259

Figure 65. A — *Omachtenia teagiana* Grey 1984; part of stromatolite showing pseudocolumns and numerous bridges.B — *Ephyaltes* form indet. Grey 1984; cone showing axial zone. C — *Carnegia wongawolensis* Grey 1984; weathered surface showing cross sections of concentric laminae.

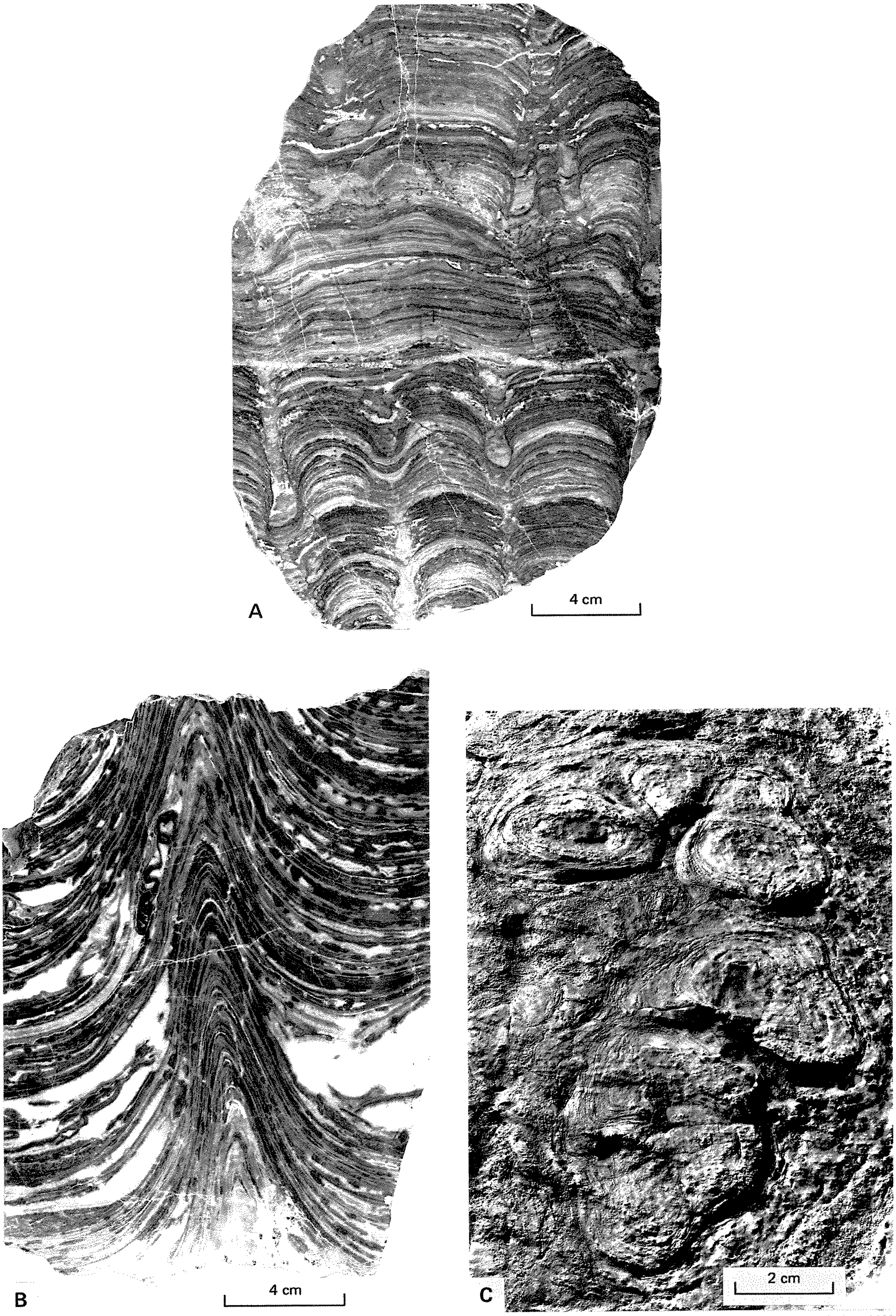
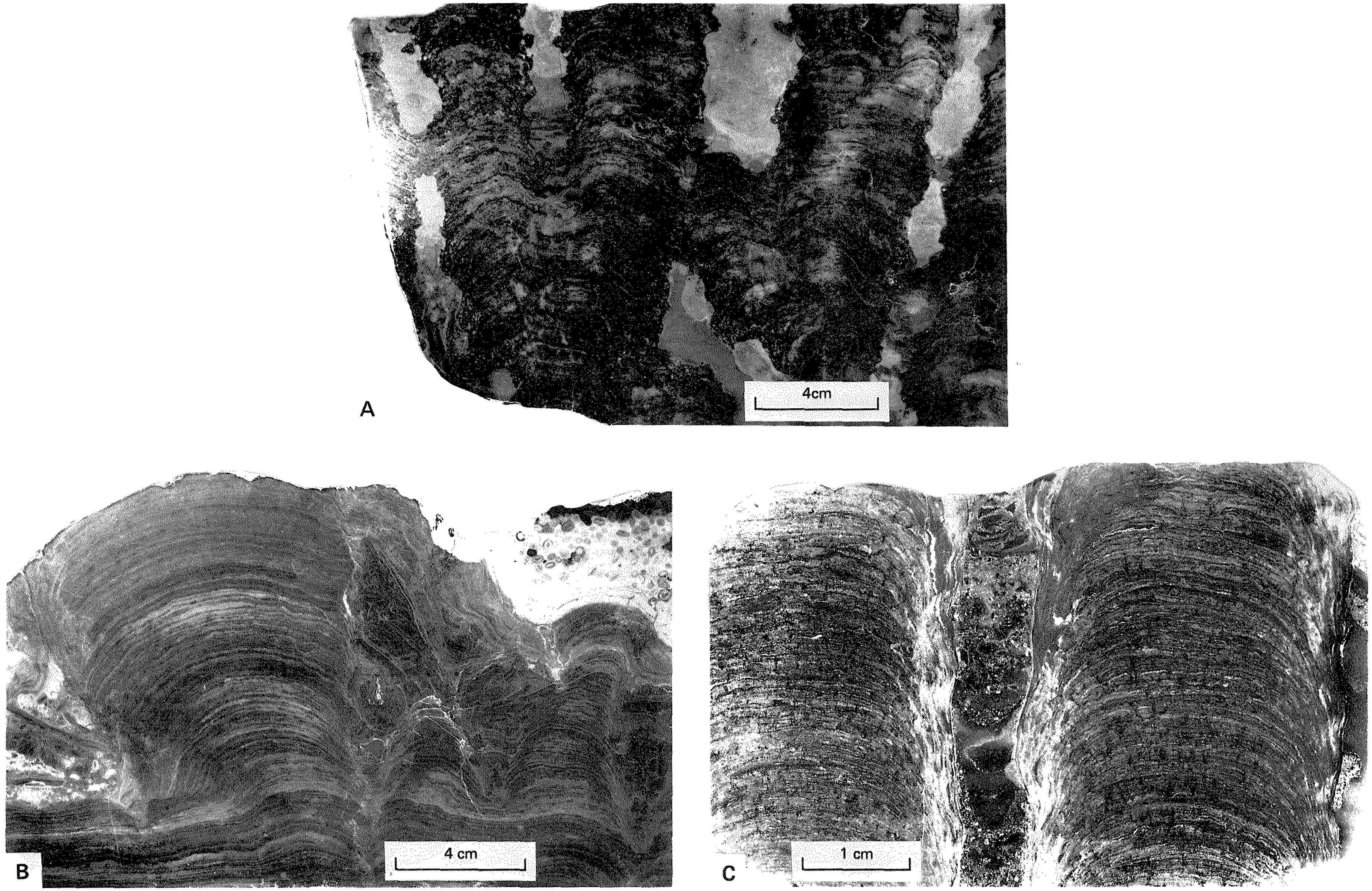
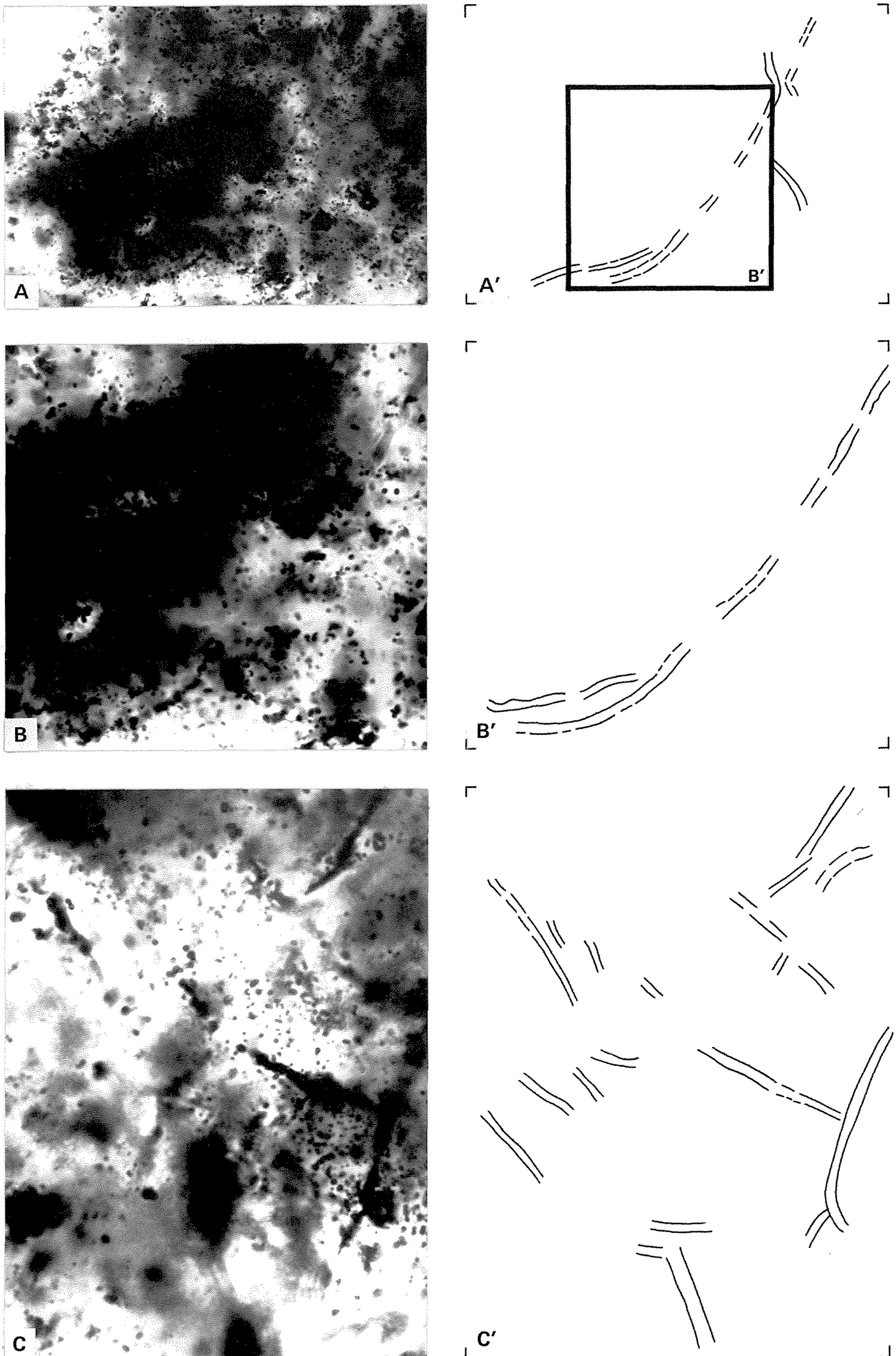

A
4 cm
B
4 cm
C
2 cm
GSWA 21260

Figure 66. A — *Yandilla meekatharrensis* Grey 1984; several columns showing branching pattern. B — *Earaheedia kuleliensis* Grey 1984; pseudocolumns, interspace clasts and later infill with oolites. C — *Externia yilgarnia* (Preiss 1976); showing vertical sections of columns. GSWA thin section F12360.



GSWA 21261

Figure 67. Traces of poorly preserved iron-encrusted bacterial filaments, from the Frere Formation, similar to specimens recorded by Walter and others, 1976. B is enlargement of A. C is another field of view. Filament diameters approximate 1 μm. GSWA thin section 46546.



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APPENDIX 2

GAZETTEER OF LOCALITIES MENTIONED IN TEXT

	Latitude (S)	Longitude (E)
Banjo Well	26°29′	121°47′
Baumgarten	25°16′	119°57′
Breakaway Bore	25°59′	121°34′
Camel Well	26°09′	121°17′
Combine Well	25°41′	119°57′
Carnarvon Range	25°15′	120°42′
Coonabildie Bluff	25°46′	122°41′
Cork Tree Well	26°23′	122°21′
Earaheedy homestead	25°36′	121°35′
Edingunna Well	25°41′	119°51′
Fergy Bore	26°06′	121°27′
Forked Creek	26°27′	122°34′
Fourteen Mile Creek	26°30′	122°25′
Frere Range	25°40′	120°28′
Fyfe Bore	25°40′	120°58′
Glengarry Range	26°12′	119°00′
Halls Creek	18°14′	127°40′
Hawkins Knob	25°27′	120°14′
Horseshoe	25°22′	118°35′
Horseshoe Lights	25°22′	118°37′
Imbin Rock Hole	25°27′	121°44′
Kaluweerie Hill	27°51′	120°10′
Kimberley Range	26°19′	119°48′
Kulele Creek	25°58′	122°10′
Lake Burnside	25°17′	122°58′
Lake Carnegie	26°10′	122°25′
Lake Disappointment	23°28′	122°48′
Lake Gregory	25°35′	119°55′
Lake Nabberu	25°43′	120°29′
Lake Teague	25°45′	120°53′
Lake Throssell	27°35′	124°10′
Lake Wells	26°40′	122°59′
Laverton	28°38′	122°24′
Lee Steere Range	25°35′	122°10′
Leonora	28°53′	121°20′
Lorna Glen homestead	26°14′	121°33′
McArthur River	16°26′	136°06′
Malmac Well	25°31′	122°17′
Meekatharra	26°36′	118°30′
Miss Fairbairn Hills	25°14′	120°21′
Mount Alexandra	26°16′	122°07′
Mount Boreas	28°03′	121°57′
Mount Cecil Rhodes	25°25′	121°26′
Mount Clarence	25°33′	120°06′
Mount Davis	25°13′	121°08′
Mount Deverell	25°35′	120°21′
Mount Elisabeth	26°33′	123°01′
Mount Evelyn	25°30′	121°45′
Mount Hosken	25°57′	122°32′
Mount Keith	27°11′	120°32′
Mount Lancelot	26°14′	123°10′

	Latitude (S)	Longitude (E)
Mount Lockridge	25°35′	120°31′
Mount Methwin	25°05′	120°41′
Mount Royal	25°31′	121°46′
Mount Teague	25°41′	120°41′
Mount Throssell	26°00′	122°40′
Mount Wellesley	26°16′	121°41′
Mount Yagahong	26°54′	118°40′
Mudan Hills	25°29′	121°38′
Mulgarra Pool	26°07′	122°44′
Murgurra Pool	25°33′	120°23′
Newman	23°16′	119°34′
Old Windidda homestead	26°38′	122°03′
Oxbys Well	25°53′	121°08′
Peak Hill	25°38′	118°43′
Phar Lap Well	26°17′	119°35′
Princess Ranges	26°07′	121°50′
Shark Bay	26°25′	114°05′
Sholl Creek	25°55′	122°12′
Simpson Well	25°24′	120°05′
Skull Soak	26°02′	121°40′
Snell Pass	25°43′	120°46′
Sweeney Creek	25°48′	120°39′
Sweetwaters Well	25°35′	120°22′
Sydney Heads Pass	25°31′	121°47′
Tangadee	24°25′	118°56′
Thaduna Mine	25°31′	119°43′
The Jump Up	26°37′	122°48′
Thurraguddy Bore	25°56′	122°02′
Timperley Range	25°56′	122°38′
Tooloo Bluff	26°26′	122°13′
Troy Creek	25°27′	121°12′
Two Mile Creek	26°12′	121°56′
Utahlarba Spring	26°04′	119°14′
Verscher Range	25°45′	120°05′
Von Treuer Tableland	26°37′	122°50′
Wandiwarra Well	26°23′	122°20′
Wannabooline Creek	26°00′	121°59′
Warburton	26°08′	126°35′
Weld Spring	25°01′	121°35′
Wellington Range	26°18′	121°47′
Wiluna	26°35′	120°14′
Windidda homestead	26°23′	122°13′
Wongawol Creek	26°14′	121°49′
Wongawol homestead	26°08′	121°57′
Woorana Well	27°30′	121°12′
Yelma Outcamp	26°32′	121°41′

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SUBJECT INDEX

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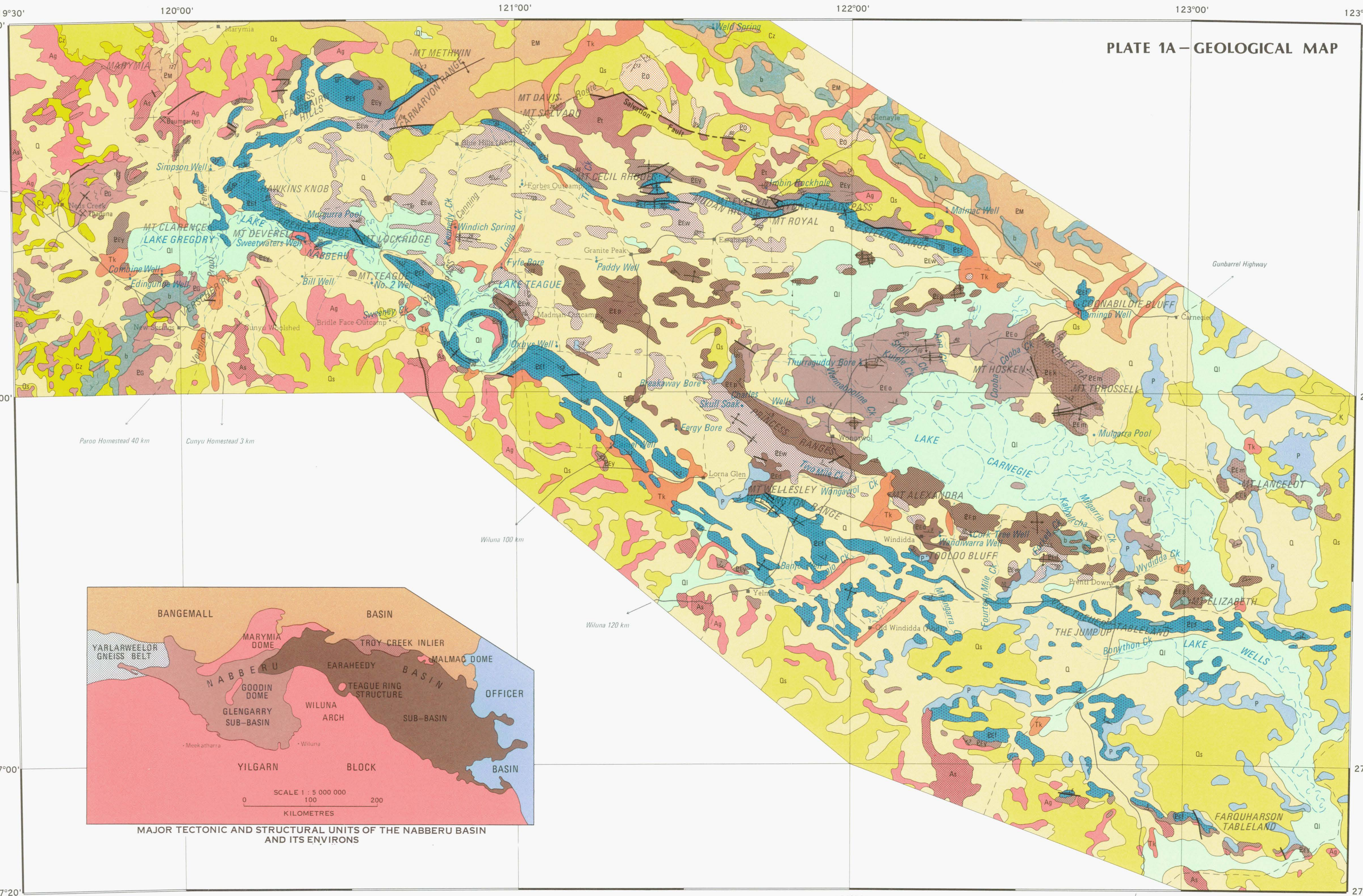
EASTERN NABBERU BASIN



MAJOR TECTONIC AND STRUCTURAL UNITS OF THE NABBERU BASIN AND ITS ENVIRONS

PHANEROZOIC	CAINOZOIC	Q	Unassigned superficial units, mainly colluvium and alluvium
		Ql	Playa lake sediments and associated marginal eolian quartz sand, gypsum and allvium
		Qs	Eolian sand
		Cz	Duricrust (laterite and silcrete) and associated superficial units
		Tk	Valley calcrete
	MESOZOIC – CRETACEOUS	K	BEJAH CLAYSTONE and LAMPE BEDS (mainly sandstone)
	PALAEOZOIC – PERMIAN	P	PATERSON FORMATION : sandstone, claystone and glacigene conglomerate
PROTEROZOIC	MIDDLE	b	Dolerite, minor gabbro and granophyre, in sills
		₽M	BANGEMALL GROUP : sandstone, shale, minor conglomerate and chert
		₽O	SCORPION GROUP : sandstone, shale, conglomerate, stromatolitic dolomite
	EARLY – Earaheedy Group – Miningarra Subgroup	₽Em	MULGARRA SANDSTONE : quartz sandstone (locally glauconitic), shale, minor carbonate
		₽Ek	KULELE LIMESTONE : stromatolitic limestone, calcarenite, micaceous mudstone, minor sandstone
		₽Eo	WONGAWOL FORMATION : arkosic sandstone and mudstone, minor carbonate
		₽Ep	PRINCESS RANGES QUARTZITE : white quartz arenite, clayey sandstone and siltstone
		₽Ew	WANDIWARRA FORMATION : quartz sandstone and shale, locally glauconitic
	EARLY – Earaheedy Group – Tooloo Subgroup	₽Ed	WINDIDDA FORMATION : micaceous mudstone and carbonate
		₽Ef	FRERE FORMATION : granular and banded iron-formation, chert, shale and carbonate
		₽Ey	YELMA FORMATION : quartz arenite, shale, minor conglomerate carbonate, chert and arkose
	EARLY	₽t	TROY CREEK BEDS : phyllite, shale, sandstone, minor carbonate, iron-formation, chert and felsic volcanics
		₽G	GLENGARRY GROUP : quartz arenite, greywacke and shale
ARCHAEAN		Ag	Granitoid and gneiss
		As	Metamorphosed sediments, basalt, dolerite, ultramafic and felsic volcanics

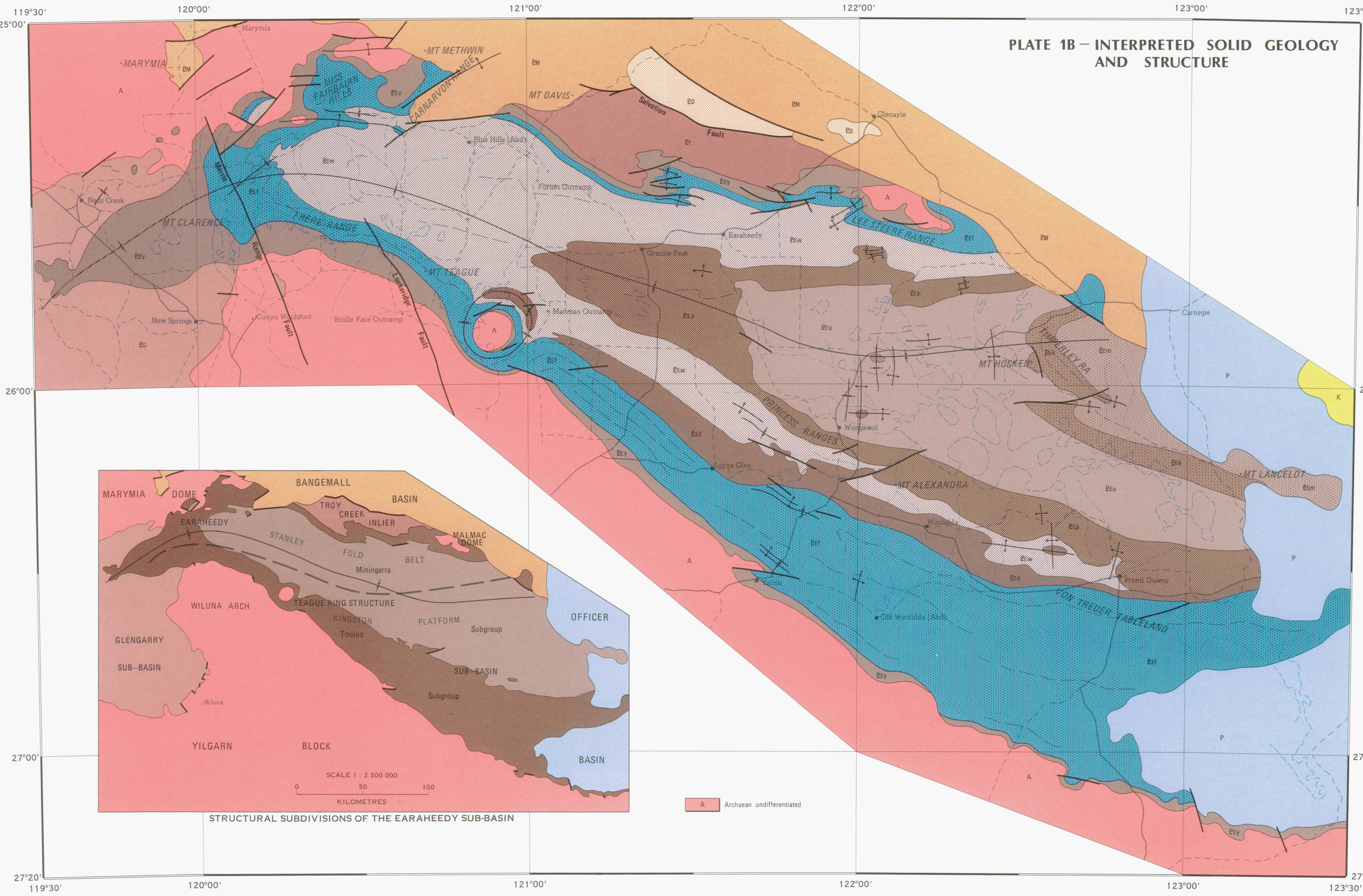
SYMBOLS

Geological boundary
Trend line
Fault
Anticline showing plunge
Syncline showing plunge
Bedding
Inclined
Overturned

Formed road
Track
Homestead — Carnegie

Watercourse (ephemeral)
Spring, soak, waterhole or pool
Well or bore

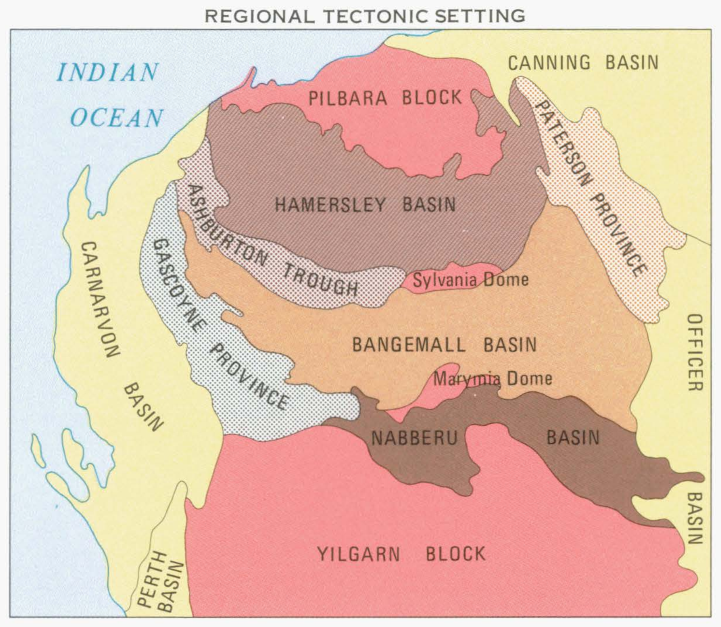
Mining centre — Thaduna



STRUCTURAL SUBDIVISIONS OF THE EARAHEEDY SUB-BASIN

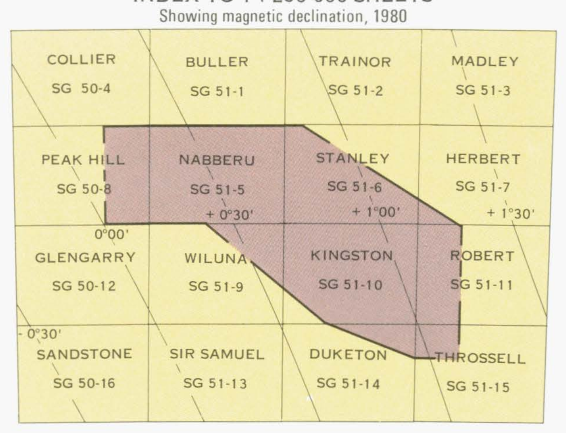
A Archaean undifferentiated

REGIONAL TECTONIC SETTING



LOCALITY DIAGRAM

INDEX TO 1 : 250 000 SHEETS
Showing magnetic declination, 1980



HON. DAVID PARKER, M.L.A.
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SCALE 1 : 1 000 000

20 0 20 40 60 80
KILOMETRES

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GEOLOGICAL MAP OF THE EASTERN NABBERU BASIN