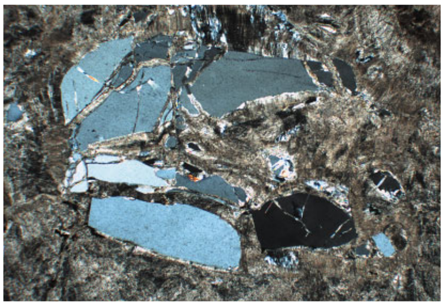
Government of **Western Australia**
Department of **Mines and Petroleum**

RECORD 2011/14

PERMIAN–CARBONIFEROUS GEOLOGY OF THE NORTHERN PERTH AND SOUTHERN CARNARVON BASINS, WESTERN AUSTRALIA — A FIELD GUIDE

by
AJ Mory and DW Haig



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PERMIAN–CARBONIFEROUS GEOLOGY OF THE NORTHERN PERTH AND SOUTHERN CARNARVON BASINS, WESTERN AUSTRALIA — A FIELD GUIDE

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Geological Survey of Western Australia

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Contents

Appendix

Figures

Tables

Permian–Carboniferous geology of the northern Perth and Southern Carnarvon Basins, Western Australia — a field guide

compiled by

AJ Mory and DW Haig[1]

Abstract

Lower Carboniferous strata are preserved only north of 24°S, but during the Middle Carboniferous to Permian the northern Perth and Southern Carnarvon Basins formed a single intracratonic depositional realm from 32°S to 23°S within the East Gondwana interior rift system. This succession is dominantly of shallow marine to fluvial origin, and is relatively undeformed (outcrop dips rarely exceed 15°) and has not been deeply buried due to a general lack of sedimentation on the eastern margin of the Perth Basin and across the Southern Carnarvon Basin throughout the Triassic–Jurassic and Cenozoic. This guide describes the most informative Carboniferous–Permian outcrops in the basins in terms of stratigraphy, sedimentology and paleontology, but also includes Pleistocene–Holocene localities that have been used as models for carbonate deposition, of which the carbonate banks and stromatolites at Shark Bay are the most renowned.

Early Carboniferous deposition in a broad interior sag basin produced two major carbonate–sandstone sequences: the Tournaisian Moogooree Limestone to Williambury Sandstone cycle; and the possible Visean Yindagindy Formation to Harris Sandstone cycle. Carbonates in the basal formations of each cycle, although poorly dated, include microbialites, ooid–peloidal and skeletal grainstones–packstones, indicating warm and occasionally hypersaline conditions.

Mid-Carboniferous–Permian deposition in a narrow interior rift, over 1000 km long, was dominated by alternating marine and terrestrial facies with marine conditions increasing to the north. A 5-km thick shallow-marine succession formed in the Merlinleigh Sub-basin in the north, and a somewhat thinner succession, including coal measures, accumulated to the south in the Perth Basin. Carboniferous to early Sakmarian glacially-influenced sediment rapidly infilled the initial basin that had an irregular topography (probably shaped by thick continental ice sheets, and affected by a later phase of the Alice Springs Orogeny). By the mid-Sakmarian, a shallow sea floor with a very low gradient developed throughout the region: water depths probably did not exceed 50 m during the Early–Middle Permian. Subsidence apparently just exceeded sediment influx, and variations in these parameters resulted in distinct depositional cyclicity of several orders. In the post-glacial successions, two major depositional cycles of late Sakmarian to mid-Artinskian and late Artinskian to at least Roadian indicate warming sea conditions during the Sakmarian, with a minor cooling phase during the middle or late Artinskian.

KEYWORDS: Permian, Carboniferous, stratigraphy, sedimentology, biostratigraphy, Perth Basin, Southern Carnarvon Basin, Shark Bay stromatolites, Western Australia.

Introduction

In Western Australia, outcrops of Permian strata south of the Tropic of Capricorn are confined to the eastern side of the Southern Carnarvon Basin and the eastern portion of the northern Perth Basin (Fig. 1). Carboniferous strata in this region are exposed only in a narrow strip along the eastern margin of the Southern Carnarvon Basin. This guide covers representative sections of the main Permian formations exposed along the Irwin River in the northern Perth Basin, and the main Carboniferous and Permian units outcropping in the vicinity of the Minilya River in the Southern Carnarvon Basin, in order to illustrate the Carboniferous–Permian lithostratigraphy, biostratigraphy, geological history, and paleogeography of this region.

Between Perth and Carnarvon, two areas of modern carbonate deposition, Carbla Point at Shark Bay and Lake Thetis in Nambung National Park, provide outstanding analogues for some facies present in the Lower Carboniferous of the Southern Carnarvon Basin. These sites are internationally known for the study of stromatolites and other types of microbialite deposition, and provide a contrast with the cool-water carbonates of the Permian.

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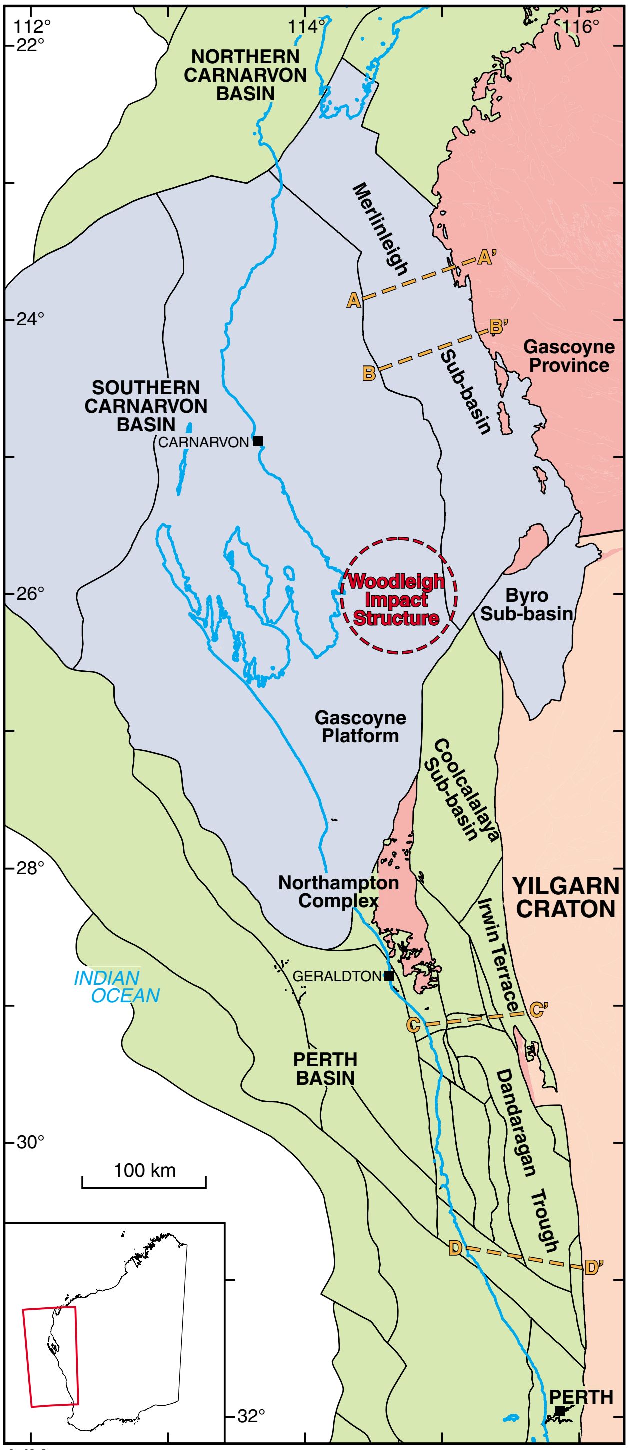

Figure 1. **Location and major tectonic subdivisions, northern Perth and Southern Carnarvon Basins, with locations of geological cross sections shown in Figure 5**

This guide has been compiled from existing guides to the northern Perth Basin (Mory et al., 2005), Southern Carnarvon Basin (Hocking, 2000), and coastal features (Haig, 2002), but also includes new observations. It also incorporates information from the review of Permian stratigraphy by Mory and Backhouse (1997); a review of the Carboniferous succession in the Southern Carnarvon Basin following Geological Survey of Western Australia (GSWA) mapping in the late 1970s by Hocking et al. (1987); a summary of more recent GSWA work in the Southern Carnarvon Basin by Mory et al. (2003); and unpublished work from The University of Western Australia (UWA) student theses.

Access to the northern Perth Basin is excellent with a network of roads branching off the Brand Highway, Indian Ocean Drive and Midlands Road to small towns and numerous farms (Fig. 2). Ownership and access details for the localities detailed in the text are provided in Appendix 1. Damage to roads and tracks after flooding, such as from December 2010 to February 2011 when rainfall in many areas exceeded annual averages, may require some flexibility with itineraries.

The area has been largely cleared of natural vegetation from just south of the Murchison River and the vermin-proof fence near latitude 28°S. The main industries in this region are agriculture (cereals, cattle, and sheep), oil and gas production (3×10^5 Kl and 1.5×10^5 m^3 in 2009, largely from Permian reservoirs), fishing, and tourism. Mining of heavy mineral sands has declined significantly in the last two years with the closure of Iluka Resources' mine at Eneabba in early 2010. Apart from small-scale mining for industrial minerals, all other mines in the region are in Precambrian terranes. By comparison, the Southern Carnarvon Basin is far less developed and has a more widely spaced network of minor roads, most of which are not sealed, connecting cattle stations to the Great Northern Highway (Fig. 2). The main industries are cattle and tourism, followed by plantations, mostly along the Gascoyne River near Carnarvon, and fishing. In addition, there is significant offshore gas and oil production farther north from the North West Shelf (33.9×10^9 m^3 and 9.3×10^6 Kl in 2009 from Mesozoic reservoirs, mostly out of Exmouth, 330 km north of Carnarvon, and Karratha/Dampier, 570 km north-northeast), and minor salt (~3 Mt/year) and gypsum (~1.2 Mt/year) production from Shark Bay and Lake McLeod.

Climate, topography and vegetation

The climate in the Carnarvon Basin is semi-arid to arid with mean annual rainfall less than 250 mm. Normally, almost 80% of annual rainfall falls in January–July due to cyclonic summer and normal winter weather patterns, whereas June and July have the highest average number of rainfall days per month(5). In Gascoyne Junction (population ~80) mean temperatures vary from 24 to 40°C in January–February to 9 to19°C in July, compared to 23 to 32°C and 11 to 23°C, respectively, for the coastal town of Carnarvon (population ~6000; 160 km west of Gascoyne Junction). In the northern Perth Basin the

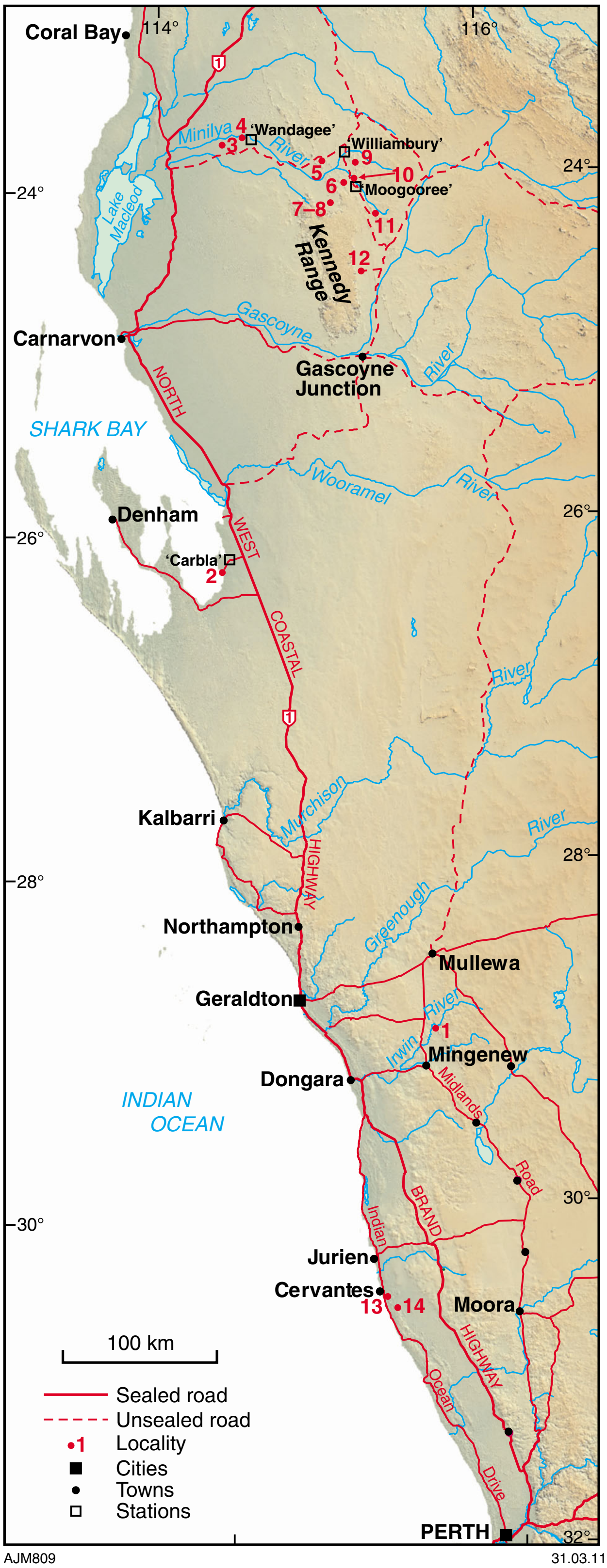

Figure 2. **Access and distribution of localities described in text, superimposed over digital elevation image**

climate is typically Mediterranean with warm dry summers and cool wet winters. About 75% of the mean annual rainfall (400–450 mm) is in May–August. At latitude 29°S in the coastal city of Geraldton (population ~32 000) mean temperatures vary from 19 to 32°C in January–February to 10 to 20°C in July–August, compared to 19 to 36°C and 7 to 19°C, respectively, 50 km inland at Mingenew (population ~300). This summary is based on data from the Bureau of Meteorology website (www.bom.gov.au/climate/data). Average annual evaporation in the region varies from 2675 to 3500 mm, with rates of 90 mm/month in June up to 490 mm in January (based on Carnarvon, Geraldton, Mullewa and Muggon data in Luke et al., 1987).

Most of the Carnarvon Basin is made up of plains with some seif dunes and low rocky exposures. The most prominent topographic feature is Kennedy Range, a tableland 80 km long and about 90 m higher than the surrounding plains, up to 370 m ASL. The plains are covered by open to dense shrub-steppe vegetation, dominated by *Acacia* scrub and hummock grassland. North of Carnarvon spinifex grasses become dominant, especially inland where rainfall is less reliable. South of the Gascoyne River, *Acacia* scrub is dominant, and is replaced south of Shark Bay by mallee heath. Little of the region has been cleared of native vegetation. The major rivers, the Wooramel, Gascoyne and Minilya, debouch into the south and north of Shark Bay and Lake Macleod, respectively, but rarely flow. All have large stands of *Eucalyptus* (gum trees) up to 20 m tall with *Acacia* along the channels. In the lower reaches, the rivers form broad *Acacia*-covered sandy flood plains with halophytic vegetation near the coast. On Kennedy Range, sand dunes overlie Oligocene duricrust and are covered with spinifex and *Acacia* scrub, as are the low coastal Cretaceous to Cenozoic limestone exposures. Both areas were uplifted in the Miocene, and are loosely associated with the collision of the Australian and Indonesian plates, as is the development of the large coastal playas of Lake Macleod and Shark Bay.

The northern Perth Basin contains a distinct coastal plain with significant heavy mineral sand deposits at the foot of a subdued Cenozoic erosional scarp (Gingin Scarp) to the east against dissected undulating hills, commonly capped by ferricrete. Farther north these hills merge into the Victoria Plateau, a largely undissected area covered with sand and ferricrete with an elevation of 250–280 m. To the east, the Darling Scarp represents the eroded position of the Darling Fault, and forms the edge of a low eroded plateau in Archean rocks of the Yilgarn Craton. The most significant rivers are the Irwin, Greenough, and Murchison, all of which originate east of the Perth Basin and reach the coast. Flow of all rivers is intermittent, and none north of the Swan River in Perth are navigable for any great distance. Along the coast, vegetation is dominated by scrub heath, grading inland to *Acacia–Casuarina* thickets and, on hard-setting loams, *Acacia* scrub with *Eucalyptus* trees. Large areas south of the Murchison River have been cleared of native vegetation, but deficiencies in the leached sandy soils and areas of lateritization and chemical depletion require fertilizers with trace elements and other strategies to maintain crop and animal health. Further information on native vegetation can be obtained from Beard (1990), and on soils and geomorphology from Bettenay (1983) from which this summary was made.

Previous studies

Southern Carnarvon Basin

Following the first European exploration of the region east of Carnarvon, Francis Gregory, an early colonial surveyor, published a remarkable geological cross-section (originally engraved in 1847) across what is now known as the Southern Carnarvon Basin. With the help of Sir Roderick Murchison and Dr Ferdinand von Hochstetter, Gregory (1861) depicted 'Devonian(?)', 'Carboniferous(?)', and 'Permian(?)' strata in stratigraphic order overlying metamorphic basement and dipping below the Kennedy Range. Later, Permian fossil material was incorrectly ascribed a Carboniferous age (Hudleston, 1883; Foord, 1890; Etheridge, 1903), and subsequently emended to Permo-Carboniferous as the Permian ages did not entirely convince many workers (e.g. Condit, 1935; Raggat, 1936). The matter was not fully resolved until the 1940s (Teichert, 1941, 1942). Skwarko (1993, p. 18–20) provides a brief summary of Permian studies in the State. Carboniferous strata were first positively identified in the basin by Teichert (1949).

The first comprehensive stratigraphic studies on the Permian of the Carnarvon Basin were based on the exploration activities of Oil Search Limited in the 1930s (e.g. Condit, 1935; Condit et al., 1936; Raggat, 1936). Teichert (1941, 1950, 1957) built on this company work and established a stratigraphic framework that forms the basis of modern studies. The Bureau of Mineral Resources (BMR, now Geoscience Australia) commenced a detailed study of the basin in 1948, with fieldwork continuing until 1956. The Permian and Carboniferous part of this work was presented by Condon (1954, 1962, 1965, 1967) and Konecki et al. (1958). West Australian Petroleum (WAPET) commenced its assessment of the Carboniferous–Permian of the Carnarvon Basin in 1953 (McWhae et al., 1954) and continued to have an active interest in the Permian of the basin until the late 1960s.

Much of the WAPET work was incorporated into the BMR reports and into later GSWA reports. GSWA mapping in the 1970s resulted in significant modifications, and a simplified nomenclature, compared to the earlier stratigraphy (Hocking et al., 1980, 1987). Some amendments to this work were made by Hocking (1990) as a result of reconnaissance trips in the late 1980s. The most complete review of previous stratigraphic nomenclature is by Skwarko (1993, table 2). The most recent revisions by Iasky et al. (1998) and Mory and Backhouse (1997) emphasize the subsurface data, especially the seismic data collected by Esso Exploration and Production Australia in 1982–84 and shallow drilling by various companies, but attempt correlation with outcrop.

Northern Perth Basin

Coal was first discovered on the South Branch of the Irwin River by the Gregory brothers on 9 September 1846 (Gregory and Gregory, 1884), but the area has yet to yield a mineable deposit. The first brief geological report on coal (von Sommer, 1848) located more seams in the North Branch which were tunnelled briefly (Woodward, 1888) but not drilled until 1921 (Maitland, 1922). Gregory (1861) identified the Permian succession as Carboniferous, a misconception that persisted until Dun and David (1923) correlated an ammonoid, *Juresanites jacksoni* (Etheridge), from below the coals with a Lower Permian fauna from Leti (about 60 km east of Timor Leste). In spite of paleontological studies indicating entirely Permian ages (e.g. Miller, 1932, 1936; Hill, 1937), the succession continued to be considered Carboniferous or Permo-Carboniferous by some workers until the early 1950s.

The Irwin River area was first mapped by Campbell (1910) as part of a larger study for GSWA; he was the first to confirm the glacial origin of the boulder beds. His work was followed by Woolnough and Sommerville (1924) as part of a search for salt deposits [!]. Their work was incorporated into that of Clarke et al. (1951), which was the culmination of numerous UWA excursions through the 1940s. The following re-evaluation of the area by GSWA (Johnson et al., 1954) gave a negative view of the coal prospects, so serious exploration did not start until the 1980s (Le Blanc Smith and Mory, 1995). To date a small deposit (Lockier) has been delineated 12 km south of the Irwin River.

West of the Permian outcrop belt, BMR gravity surveys in 1949 and 1951–52 prompted WAPET to explore for hydrocarbons (Playford et al., 1976) with the first commercial discoveries (Yardarino in 1964, and Dongara and Mount Horner in 1965) in Upper Permian reservoirs. Mory and Iasky (1996) provide a summary of that work, but exploration has continued to the present prompted by new discoveries — 12 small fields are currently operational.

Regional geology

Basins containing Carboniferous and Permian strata are part of the East Gondwana interior rift system (Fig. 3; Veevers, 1971; Harrowfield et al., 2005) that extended from western New Guinea, Ceram, and Timor in the north to southern Africa in the south. Final continental breakup along the rift system took place at the beginning of the Late Jurassic (~155 Ma) in the north and at about 130 Ma (intra-Valanginian, Early Cretaceous) off southwest Australia (Heine and Müller, 2005). Major facies transitions are present along the rift system, generally with more open-marine conditions towards the north. South of a level corresponding to present-day Perth, non-marine conditions prevailed during the Permian and persisted until Early Cretaceous breakup. A north-to-south climatic gradient is evident in the rift system with glacially influenced facies conspicuous in the Lower Permian of the Perth and Southern Carnarvon Basins, less so in the Canning Basin (Mory et al., 2008), and apparently absent from coeval strata in Timor to the north.

Within the rift system, the main Permian depocentres extend between the Merlinleigh Sub-basin in the Southern Carnarvon Basin and the Irwin Sub-basin in the northern Perth Basin. Because of their narrowly elongate rift structure and thick succession of strata, these basins are a conspicuous feature of the Australian gravity anomaly

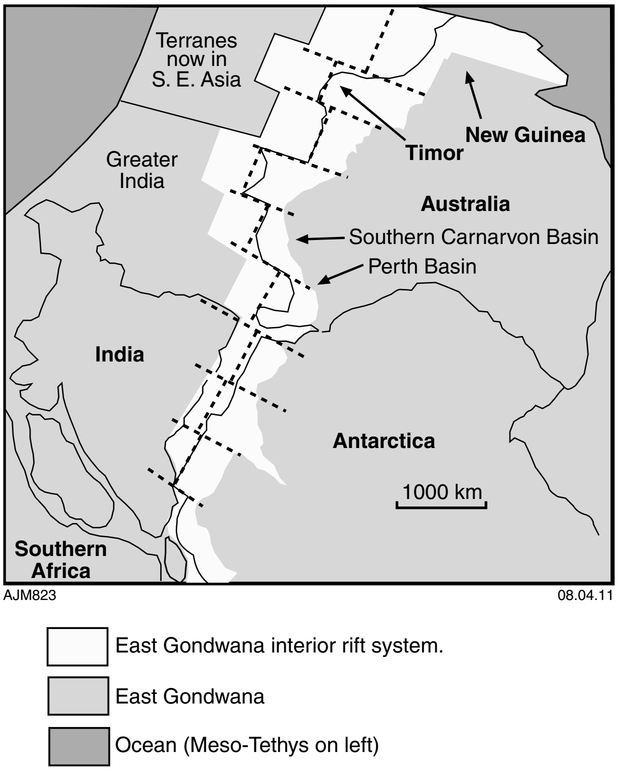

Figure 3. The East Gondwana interior rift system (after Harrowfield et al., 2005). Dashed lines indicate positions of rifts during final breakup (at 155 Ma in vivinity of Timor; 136 Ma off SW Australia). Ages of continental breakup follow Heine and Müller (2005); Eary Permian age of initial rifting along the Meso-Tethyan ocean boundary follows Metcalfe (2006)

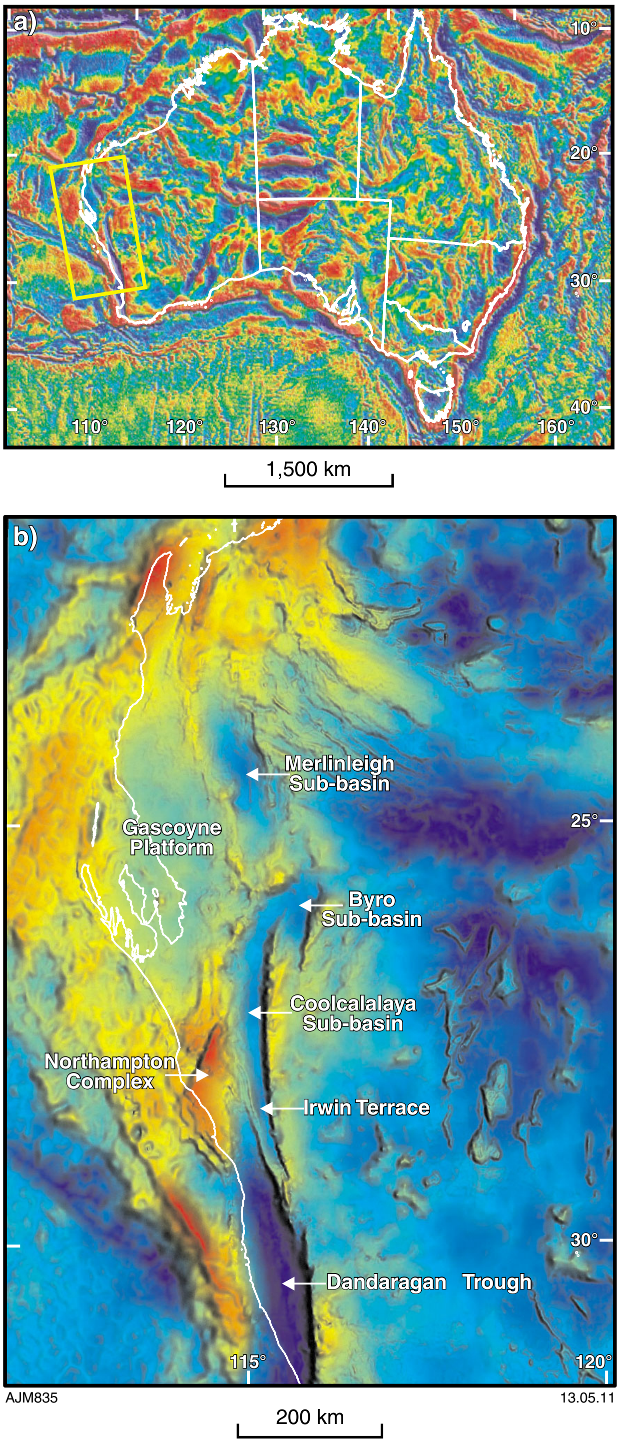

Figure 4. Gravity anomaly images of a) Australian region; and b) northern Perth – Southern Carnarvon Basins

map (Fig. 4). The Permian depocentres are superimposed on broader earlier Paleozoic basins or lie on Precambrian basement. In the onshore Southern Carnarvon Basin, a broad tectonostratigraphic succession of stacked depocentres is recognized (numbered in ascending stratigraphic order):

1. Ordovician to Lower Carboniferous broad interior sag basin with alternating sand-dominated and carbonate-dominated formations. The eastern boundary of this basin onlaps onto Precambrian metamorphic basement. The western part of the basin was truncated during Early Cretaceous continental breakup.
2. Late Carboniferous – Permian interior rift basin including 5 km of strata in the Merlinleigh Sub-basin (Fig. 1). The Wandagee Fault, active during deposition of the Lyons Group, defines the western side of the basin. Such movement in northern Australian basins has been placed within the later phases of the Alice Springs Orogeny (Haines et al., 2001). Although the eastern side of the basin has been removed in part by erosion, maturity data indicates substantial thinning of the succession to the east (Iasky and Mory, 1999). Permian strata may have extended across highly deformed and metamorphosed Proterozoic rocks of the Gascoyne Province to the east (see Veevers et al., 2005 for speculation on the extent of Permian cover on the Archean Yilgarn Craton to the south). By comparison, maturity data from the Lower Paleozoic succession west of the Merlinleigh Sub-basin (Ghori et al., 2005) indicates this region was a relative high throughout the Early Permian to earliest Cretaceous.
3. Triassic–Jurassic pre-rift and syn-rift deposition in the northern Perth Basin shows a rapid regression to fluvial conditions in the Early Triassic with a minor, but extensive, marine transgression in the Middle Jurassic marking the change from pre-rift to syn-rift facies. In most of the Southern Carnarvon Basin, only the Lower

Jurassic is preserved as lacustrine deposits infilling a confined depression over the Woodleigh impact structure (Mory et al., 2000, 2001). This 120 km-diameter, multi-ring circular structure was identified largely in gravity data (Iasky et al., 2001) and probably formed close to the Devonian–Carboniferous boundary (Uysal et al., 2001, 2005). These deposits point to terrestrial conditions during the Triassic–Jurassic, a period that encompasses a major regional unconformity in the remainder of the Southern Carnarvon Basin.

4. Cretaceous (Hauterivian and younger) to Cenozoic passive continental shelf with marine deposition at times extending across all of the Southern Carnarvon Basin (and possibly to the east over metamorphic rocks of the Gascoyne Province, where such deposits may have been removed by erosion). In the northern Perth Basin deposits of this age are thin onshore, where they are mostly preserved south of 32°S, but thicken dramatically offshore.

Because of the administration associated with resource exploration licensing, the sedimentary basins of Western Australia have not been classified following a stacked basin model, such as above, but by geography and the present structural configuration. The Phanerozoic basins have been portrayed as extending vertically down to the Precambrian, usually metamorphic or granitic, basement, totally ignoring changes in depositional style and major breaks. Some physiographic terms have been used in the current basin nomenclature (e.g. Gascoyne Platform; Irwin Terrace); these should not be taken to imply paleogeography.

The Southern Carnarvon Basin comprises three Palaeozoic sub-basins: the Gascoyne Platform to the west, and the Merlinleigh and Byro Sub-basins to the east (Fig. 1). Paleoproterozoic gneissic basement of the Carrandibby Inlier separates the Merlinleigh and Byro Sub-basins, which may have been a single depocentre extending south into the Coolcalalaya Sub-basin, and possibly farther south into the Irwin Terrace of the Perth Basin (Fig. 1). The Southern Carnarvon Basin covers most of the onshore part of the greater Carnarvon Basin, but includes an offshore portion south of the Rough Range Fault system and north of the Abrolhos Sub-basin (Fig. 1). The basin has largely Paleozoic fill, and on that basis is distinguished from the northern, mostly Mesozoic, offshore part of the basin (Hocking et al., 1994; Hocking, 1994). The Paleozoic succession probably extends throughout the Northern Carnarvon Basin, but at depths greater than 4 km.

The Perth Basin extends south of the Southern Carnarvon Basin to the south coast of the State and offshore to the continent–ocean boundary, and contains a thick Permian–Cretaceous succession. Separating the two basins are two transitional areas, the onshore Coolcalalaya Sub-basin (Mory et al., 1998) and the offshore Abrolhos Sub-basin (Crostella, 2001) that traditionally have been regarded as part of the Perth Basin. Between these two sub-basins is the Northampton Province, part of the Meso- to Neoproterozoic Pinjarra Orogen that has acted as a relative high throughout much of the Phanerozoic.

The northern part of the basin comprises an eastern faulted terrace between the Perth and Urella faults (Irwin Terrace), an onshore deep half graben (Dandaragan Trough) gradually shallowing westwards towards a coastal basement high (Dongara Saddle and Beagle Ridge; Fig. 5), deepening westwards into the offshore, dominantly Mesozoic, Abrolhos, and Houtman Sub-basins. The east–west Allanooka Fault marks the boundary with structurally shallow basement at the northern end of the Dandaragan Trough. South of 31°S the Dandaragan Trough grades into another set of deep troughs also dominated by a thick Mesozoic succession. Carboniferous strata in the Perth Basin are restricted to the Coolcalalaya Sub-basin (Mory et al., 1998), and mostly are part of the glacial succession. Most outcrop of the Permian succession is within the Irwin Terrace but extends west in the subsurface where it is a productive part of the Permian–Triassic Gondwanan petroleum system (Bradshaw et al., 1994) with 12 small fields still in production largely from Upper Permian reservoirs.

Isopach maps of Permian formations in the northern Perth Basin show distinct changes from a half-graben controlled by the Darling–Urella Fault in the Sakmarian, followed by a sag-phase centred on the northern Dandaragan Trough in the Artinskian (Eyles et al., 2006, fig. 4) and a return to a half graben in the Middle–Late Permian (Mory and Iasky, 1996, fig. 13). In the Southern Carnarvon Basin, a similar trend is evident, but the Early Permian depocentre lies along the western, rather than eastern margin of the Merlinleigh Sub-basin, and probably also the Byro Sub-basin (Iasky et al., 1998; Eyles et al., 2003). This reversal in asymmetry (Fig. 5) underpins the recognition of the Perth and Southern Carnarvon Basins as separate structural identities — although not spelt out specifically by Hocking (1994) — but is in contrast to the sedimentological similarities of the two regions throughout the Permian. Nevertheless, this difference in structure indirectly supports the contentions of Eyles et al. (2002, 2003, 2006), that the onset of glacial conditions, at least in these basins, had a strong structural control.

Biostratigraphy

To date, most outcrop paleontological studies have been at formation level; bed-by-bed sampling has only taken place for foraminifera (e.g. Haig, 2003; Dixon and Haig, 2004) and conodonts (Nicoll and Metcalfe, 1998; RS Nicoll, 2010, written comm.) implying that more detailed biostratigraphic (and paleoenvironmental) control could be achieved. However, in many Carboniferous and Permian formations only a few beds are noticeably fossiliferous. In addition, facies controls and provincialism of many of the faunas hinder long-range correlation. Whereas palynomorphs provide the most coherent means of correlation, both within and between basins, even though the zonal resolution is generally broad, there is a dichotomy between subsurface and outcrop sections as macrofauna are rarely recovered from the former and palynomorphs are seldom obtained from the latter, due to surface weathering/oxidization. Therefore, the relationship between the spore-pollen and brachiopod zonal schemes in the Permian, for example (Figs 6 and 7), depends largely on the correct correlation of formations from outcrop to the subsurface.

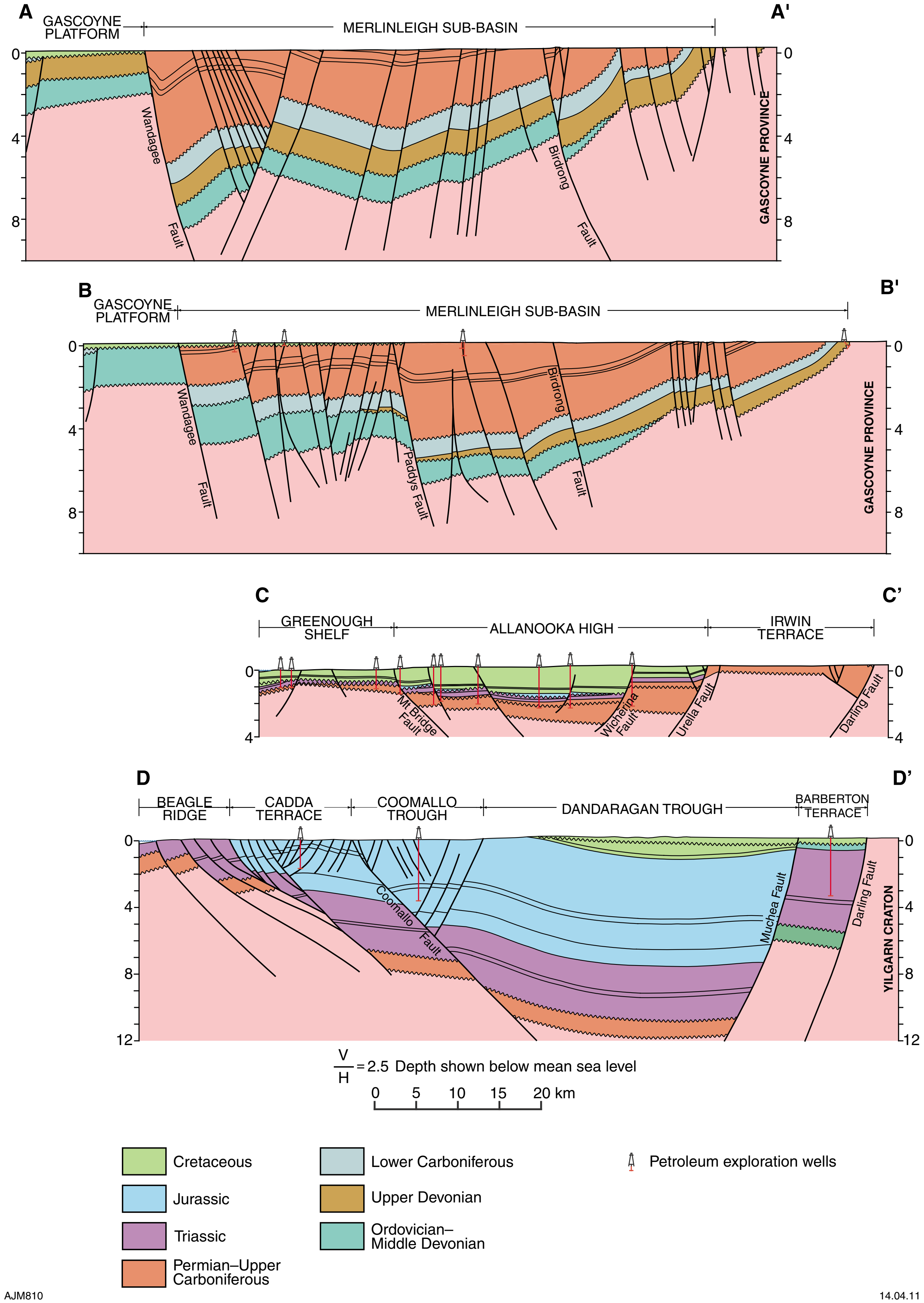

Figure 5. **Geological cross-sections across the Merlinleigh Sub-basin (from Iasky et al., 1998) and northern Perth Basin (after Mory and Iasky, 1996). Locations shown on Figure 1**

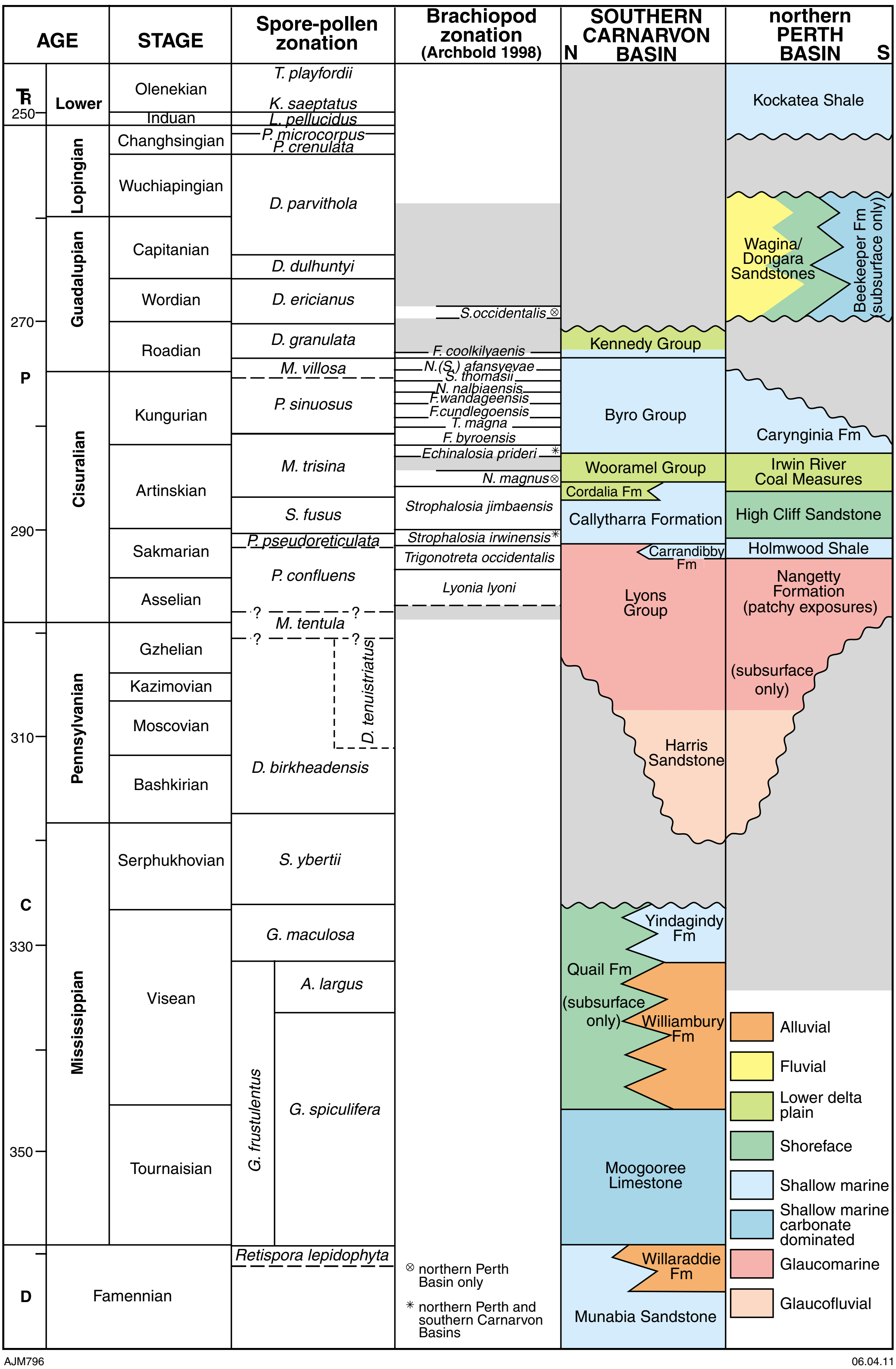

Figure 6. **Generalized Carboniferous and Permian stratigraphy and biostratigraphy, northern Perth and Southern Carnarvon Basins. Timescale from Gradstein et al. (2004), modified after Permophiles 55**

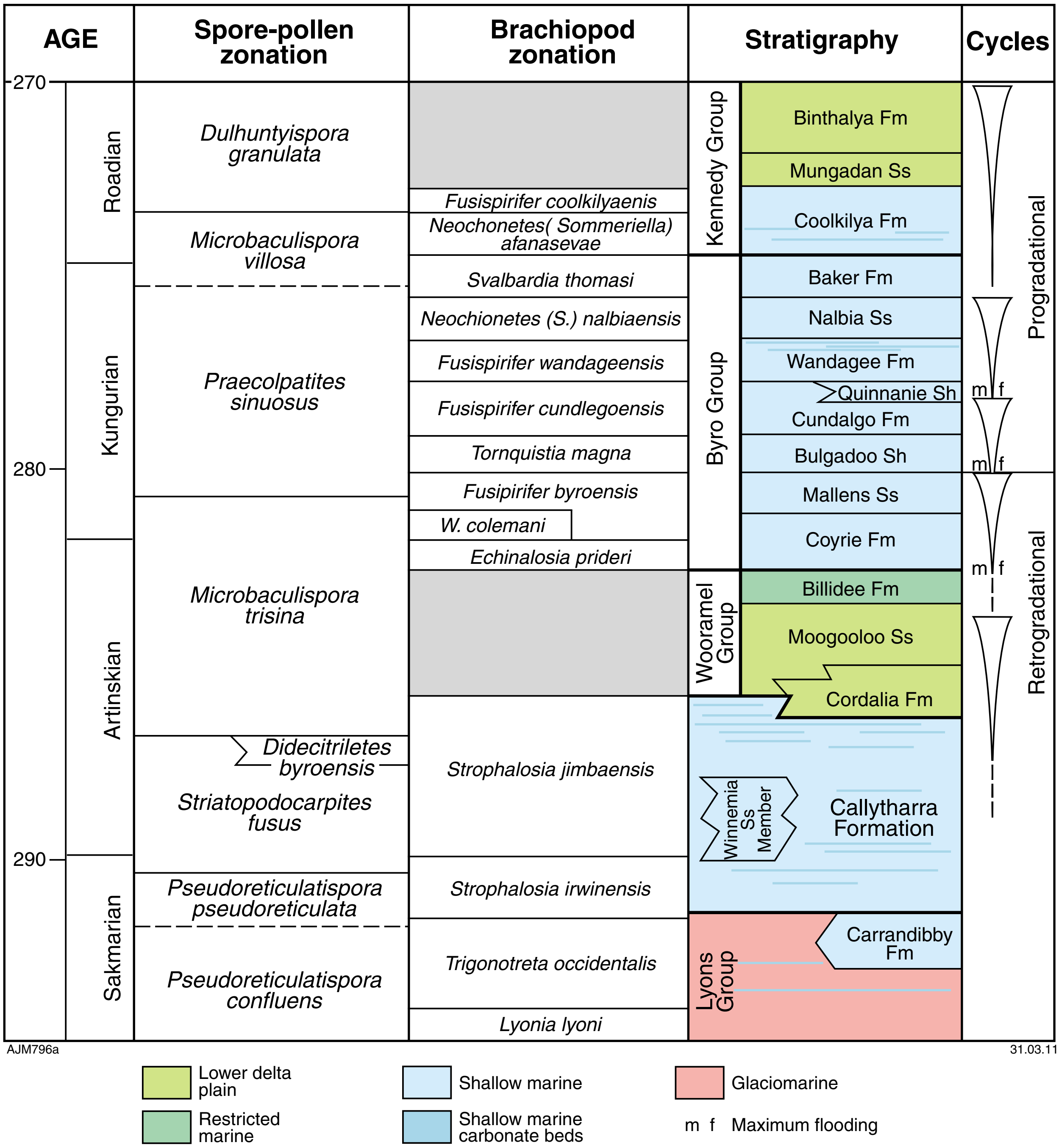

Figure 7. Lower–Middle Permian (Sakmarian–Roadian) stratigraphy, biostratigraphy and major cycles, Southern Carnarvon Basin

Carboniferous

Of the moderately diverse Lower Carboniferous marine faunas in the Southern Carnarvon Basin, only brachiopods have been studied in detail (Glenister, 1955; Thomas, 1971), but their biostratigraphic use is limited due to provincialism (Skwarko, 1988) — strata of this age are not known from the Perth Basin. Although the two Lower Carboniferous carbonate units have yielded conodont and ichthyoid faunas, they are yet to be described. The conodont faunas are dominated by long ranging, shallow-water forms, similar to British 'shelf' faunas, and also showing considerable facies control (RS Nicoll, 2009, written comm.). Therefore, although there are strong similarities to faunas from the Canning and Bonaparte Basins described by Nicoll and Druce (1979) and Druce (1969), detailed correlation with these basins is likely to be difficult. Preliminary work on the shark faunas has been reported in an abstract (Trinajstic and George, 2007); further work under way by K Trinajstic (Curtin University, Perth, written comm.) is expected to show good parallels with European and north American faunas described by Ginter (2009). The presence of foraminifera has been recorded in a number of general studies, but only limited work has been carried out on these faunas, mostly by DWH. Whereas palynomorphs provide useful, although broad, controls for correlation of subsurface sections based on the work of Kemp et al. (1977), this part of the Carboniferous has been intersected in just seven petroleum exploration wells in the basin.

The only fossils from the Serpukhovian–Pennsylvanian succession are palynomorphs of the *Spaeotriletes ybertii* zone known from just one petroleum exploration well (Warroora 1) on the Gascoyne Platform and the *Deusilites tenuistriatus* zone from shallow water bores on Lyons River station and within the Coolcalalaya Sub-basin (Mory et al., 1998). In this interval, ages are very poorly constrained as non-marine facies dominate across Australia with very few low-diversity marine faunas to provide a link to international stages.

Permian

The Permian marine facies of Western Australia, especially from the Southern Carnarvon and Canning Basins, contains diverse faunas, and as such have provided the main link to international correlations for eastern Australian Permian successions (e.g. Archbold, 1998a, 2002). Nevertheless, that correlation path is likely to be reversed as more work is done on dating ash beds in the east of Australia as such beds are rare in the west. The most biostratigraphically useful macrofossil groups are ammonoids (Glenister and Furnish, 1961; Glenister et al., 1973) and brachiopods (Archbold, 1998a, 2002), although there appears to be strong facies controls on the latter. Archbold (1998a, figs 3 and 5) indicates that 14 of the 18 Western Australian Permian brachiopod zones have been identified in the Southern Carnarvon Basin but just 4 of these zones in the Perth Basin. Other macrofossil groups of more limited use at present include bryozoa (Crockford, 1957) and bivalves (Dickins, 1963, 1964). This work, and others, is summarized in the volume on Permian groups edited by Skwarko (1993).

Amongst the microfossils, palynomorphs are virtually ubiquitous in the subsurface, therefore most significant biostratigraphically (e.g. Backhouse, 1993a; Mory and Backhouse, 1997). In addition, foraminifera provide some coarse age constraints, and valuable information on paleobathymetry (Crespin, 1958; Belford, 1960; Foster et al., 1985; Ferdinando, 2002; Haig, 2003; Dixon and Haig, 2004). DWH continues to work on this group. Conodont faunas (Nicoll and Metcalfe, 1998) are strongly endemic and have a limited vertical distribution due to the prevalence of cold-water facies. Fish, shark, and ostracod faunas have received scant attention (Foster et al., 1985; Turner, 1993; Fleming and Skwarko, 1993; Ferdinando, 2001).

Permian microfloras, which are common in the subsurface, and Permian brachiopod faunas, which are often abundant within a few beds in outcrop, are the most biostratigraphically useful groups in Western Australia, but provide only tentative and indirect correlations to the international time scale. Both zonations show poor resolution in the Asselian–Tastubian (Early Sakmarian) due to the dominance of glacial siliciclastic facies, and in the mid-Capitanian – Lopingian due to the stability of palynofloras and the paucity of marine facies. In addition, facies control of the brachiopod zonation is clearly illustrated by the correspondence of zones to individual formations, especially the eight zones within the Artinskian–Roadian Byro Group (Archbold, 1998a, fig. 5; 2002, fig. 2), which correspond to the *Praecolpatites sinuosus* and uppermost *Microbaculispora trisina* palynomorph zones (Mory and Backhouse, 1997).

Northern Perth Basin stratigraphy

Upper Carboniferous – Permian stratigraphy

Permian sedimentary rocks form well-known exposures in the northern part of the Perth Basin and along the Irwin Terrace between the Darling and Urella Faults, and are widespread in the subsurface throughout the remainder of the basin. Late Carboniferous ages based on palynological correlations have been recorded near the base of the succession, but often are based on low-diversity floras. The succession (Figs 6 and 7) comprises mixed marine and continental deposits that locally probably reach thicknesses in excess of 2600 m (Playford et al., 1976) and typically rest unconformably on Precambrian metamorphic and plutonic rocks, and on Ordovician and ?Devonian strata in the north. In the southern Perth Basin, by comparison, the Permian succession is represented entirely by continental deposits (Stockton and Sue Groups; Le Blanc Smith, 1993; Le Blanc Smith and Kristensen, 1998).

The best exposures of Permian rocks are in the Irwin River and Woolaga Creek areas but there are also scattered outcrops along the Lockier, Greenough, and Murchison river valleys. Coalseam Conservation Park (**locality 1**), near the junction of the north and south branches of the Irwin River contains the most accessible Permian exposures.

The Gregory brothers first discovered coal in the State along the South Branch of the Irwin River in 1846, and Gregory (1861) and Campbell (1910) made the first geological descriptions of the area. The Lower Permian succession consists of glacigene deposits (Nangetty Formation and Holmwood Shale), locally capped by cold-water, richly fossiliferous shallow-marine carbonates (Fossil Cliff Member of the Holmwood Shale), overlain in turn by siliciclastic paralic deposits (High Cliff Sandstone), fluvial–deltaic facies (Irwin River Coal Measures), and restricted marine facies (Carynginia Formation). Upper Permian fluvial deposits (Wagina Sandstone) overlie the Carynginia Formation with a markedly erosive base. There is little evidence of an angular unconformity at this level in outcrop, but seismic profiles show a distinctly angular relationship offshore (Smith and Cowley, 1987).

West of the Urella Fault, outcrops mostly belong to the Upper Jurassic Yarragadee Formation or the uppermost Jurassic to lowermost Cretaceous Parmelia Group. The low maturities (about 0.5% vitrinite reflectance) measured from the Irwin River Coal Measures on the Irwin Terrace imply the Permian succession was buried by no more than 500–1000 m of Mesozoic strata, whereas the Triassic to Cretaceous section west of the fault in the Dandaragan Trough is up to 3500 m thick.

Nangetty Formation

Lithology: The Nangetty Formation consists of shale, sandstone, conglomerate, and minor tillite. Erratic boulders (commonly faceted and striated) up to 6 m in diameter within the formation indicate glacial activity and ice rafting of dropstones. The largest glacial erratic is the quartzitic ‘White Horse’, on ‘Mungaterra’ at 29°02'03"S 115°28'53"E, Dongara, derived from the Proterozoic Coomberdale Chert near Moora (~160 km south).

Distribution and thickness: The unit extends through much of the northern Perth Basin with a thickness exceeding 1000 m adjacent to the Urella Fault, and pinching out against the Northampton Complex and Beagle Ridge to the northwest and west, respectively. The unit appears to be absent farther west. The unit unconformably overlies Precambrian metamorphic and plutonic rocks that resemble many of the erratic boulders in the formation.

Exposures are typically poor, even in Nangetty Hills, the type area of the unit. Although a specific type section has not been proposed there are reasonable, but discontinuous, nearby exposures along the Irwin River between 29°04'07"S 115°26'30"E, Dongara*, and 28°59'49"S 115°28'24"S, Geraldton.

Fossils and age: The unit typically contains palynomorphs belonging to the ‘Stage 2’ (Backhouse, 1998; equivalent to the *Microbaculispora tentula* zone discussed by Mory, 2010) and is considered to be of Asselian or early Sakmarian age. The unit has also yielded mid- to Late Carboniferous palynomorphs north of the Greenough River in the Coolcalalaya Sub-basin (Mory et al., 1998).

Environment of deposition: The formation was probably deposited in both marine and continental glacial environments, although the only marine fossils from the unit are foraminifera (Crespin, 1958).

Holmwood Shale

Lithology: The Holmwood Shale comprises grey–green shale and thin limestone beds in the lower part passing transitionally into grey–black micaceous, jarositic, and gypsiferous shale and siltstone, with minor discontinuous beds of cross-laminated fine-grained sandstone and coquinite in the upper part. In situ dropstones are rare within the formation.

The Holmwood Shale was originally proposed for the dark shale conformably overlying the Nangetty Formation and conformably underlying the ‘Fossil Cliff Formation’ (Clarke et al., 1951). Its type section is along Beckett Gully 8 km south of Coalseam Conservation Park.

Distribution and thickness: The formation thins from approximately 450 m in outcrop, and 625 m in Depot Hill 1 (10 km east of the Urella Fault), to less than 100 m near Dongara. Johnson et al. (1954) and Playford et al. (1976) noted that the richly fossiliferous calcareous facies attributed to the formation was thin, lenticular, and difficult to map, and redefined these beds as an uppermost member within the Holmwood Shale. Additional lenticular calcareous facies lower in the formation (Woolaga Limestone and Beckett Members) do not extend far beyond the areas after which they are named.

Fossils and age: The shale and siltstone facies are poorly fossiliferous, commonly only containing cryptostomate bryozoans. In contrast, the limestone facies are richly fossiliferous. Skwarko (1993) recorded foraminifera, bryozoans, bivalves, gastropods, ammonoids, nautiloids, crinoids, ostracods, annelids, and conulariids from the Holmwood Shale (excluding the Fossil Cliff Member). The spore-pollen *Pseudoreticulatispora confluens* and *P. pseudoreticulata* zones have been identified from the unit in the subsurface and from nodules 50 m and 20 m below the High Cliff Sandstone, respectively (Backhouse, 1993b, 1998). Ammonoids, including *Juresanites jacksoni* (Etheridge) from the Beckett Member and *J. jacksoni* and *Uraloceras irwinense* Teichert and Glenister from the Woolaga Limestone Member, are indicative of the Sakmarian (Miller, 1932, Teichert, 1942; Teichert and Glenister, 1952; Glenister and Furnish, 1961; Glenister et al., 1973; Glenister et al., 1990; Leonova, 1998).

Environment of deposition: Lithologies, sedimentary structures, and fossils representative of this formation indicate chiefly cold-water, low-energy marine depositional environments. Fossiliferous limestone lenses probably represent localized, well-aerated, shallow-marine banks.

Fossil Cliff Member

Lithology: The member consists of interbedded dark micaceous and gypsiferous siltstone, sandy siltstone, shale, and bioclastic calcarenite deposited in a series of

* Capitalized names refer to standard 1:250 000 map sheets

coarsening-upward parasequences (Fig. 8). The carbonate beds are markedly lenticular and the member outcrops sporadically from Fossil Cliff to about 16 km to the south. The unit is best exposed at its type section, Fossil Cliff on the North Branch of the Irwin River (at 28°56'39"S 115°32'52"E, PERENJORI) immediately upstream from High Cliff.

Fossils and age: Foraminifera, corals, bryozoans, brachiopods, bivalves, ammonoids, nautiloids, gastropods, ostracods, trilobites, and crinoids have been recorded from the member (Skwarko, 1993). The member is regarded as Sakmarian (either late Tastubian or Sterlitamakian) based on a single specimen of the ammonoid *Metalegoceras kayi* from the type section (Glenister et al., 1973). A spore-pollen assemblage from the type section described by Foster et al. (1985) belongs within the *Pseudoreticulatispora pseudoreticulata* zone of Mory and Backhouse (1997) and Backhouse (1998).

Environment of deposition: Changes in lithology and fossil content are attributed to sea-level fluctuations and changes in terrigenous input (Ferdinando, 2002). The skeletal component of the limestone beds is diverse though fragmentary, whereas siltstone beds host less-diverse macrofaunas preserved mostly as moulds.

High Cliff Sandstone

Lithology: Clarke et al. (1951) defined this formation for the interbedded sandstone, conglomerate, and siltstone transitional between the 'Fossil Cliff Formation' and the Irwin River Coal Measures. Both lower and upper contacts are conformable.

Distribution and thickness: The type section (Fig. 9) at High Cliff on the North Branch of the Irwin River is 24 m thick (Playford et al., 1976). Clarke et al. (1951) and Sanders and Ingram (1964) listed thicknesses of 37 m and 26 m, respectively, for this section. The discrepancies appear to relate to positioning of the formation's upper boundary, here taken to be at the base of a thick dark shale and siltstone bed in the upper part of High Cliff. No coal is exposed in High Cliff and the lenticular character of many beds inhibits ready correlation with the section 500 m upstream. The High Cliff Sandstone and Irwin River Coal Measures are often difficult to distinguish in the subsurface where core is not available, but collectively these formations are recognized throughout the northern Perth Basin.

Fossils and age: Body fossils are absent from the formation at High Cliff although both high energy (*Skolithos*-type) and low energy (*Planolites*- and *Rosselia*-type) burrow forms are abundant (Fig. 10a). At other localities, foraminifera, bryozoans, brachiopods, bivalves, and gastropods have been recorded from the formation (Skwarko, 1993; Archbold, 1997). On the basis of stratigraphic position and comparison of brachiopod faunas from other formations on the Irwin Terrace and in the Southern Carnarvon Basin, Archbold (1997) suggested an early Artinskian (Aktastinian) age for the High Cliff Sandstone.

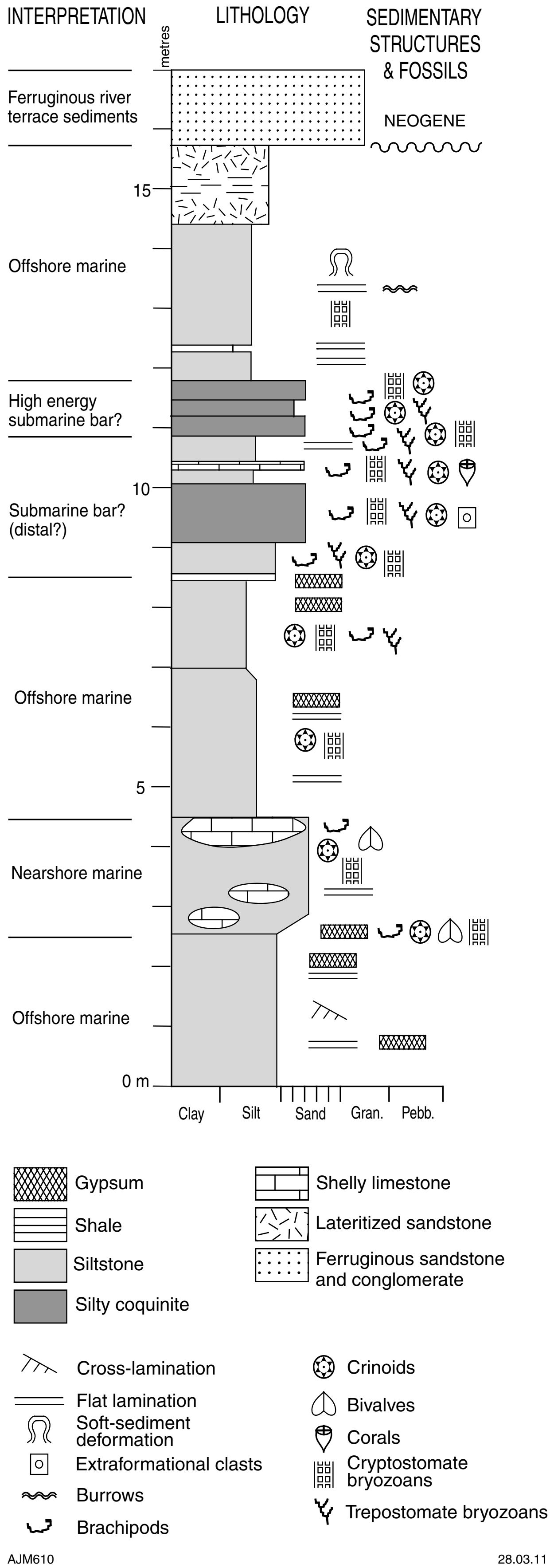

Figure 8. Measured section, Fossil Cliff, Coalseam Reserve

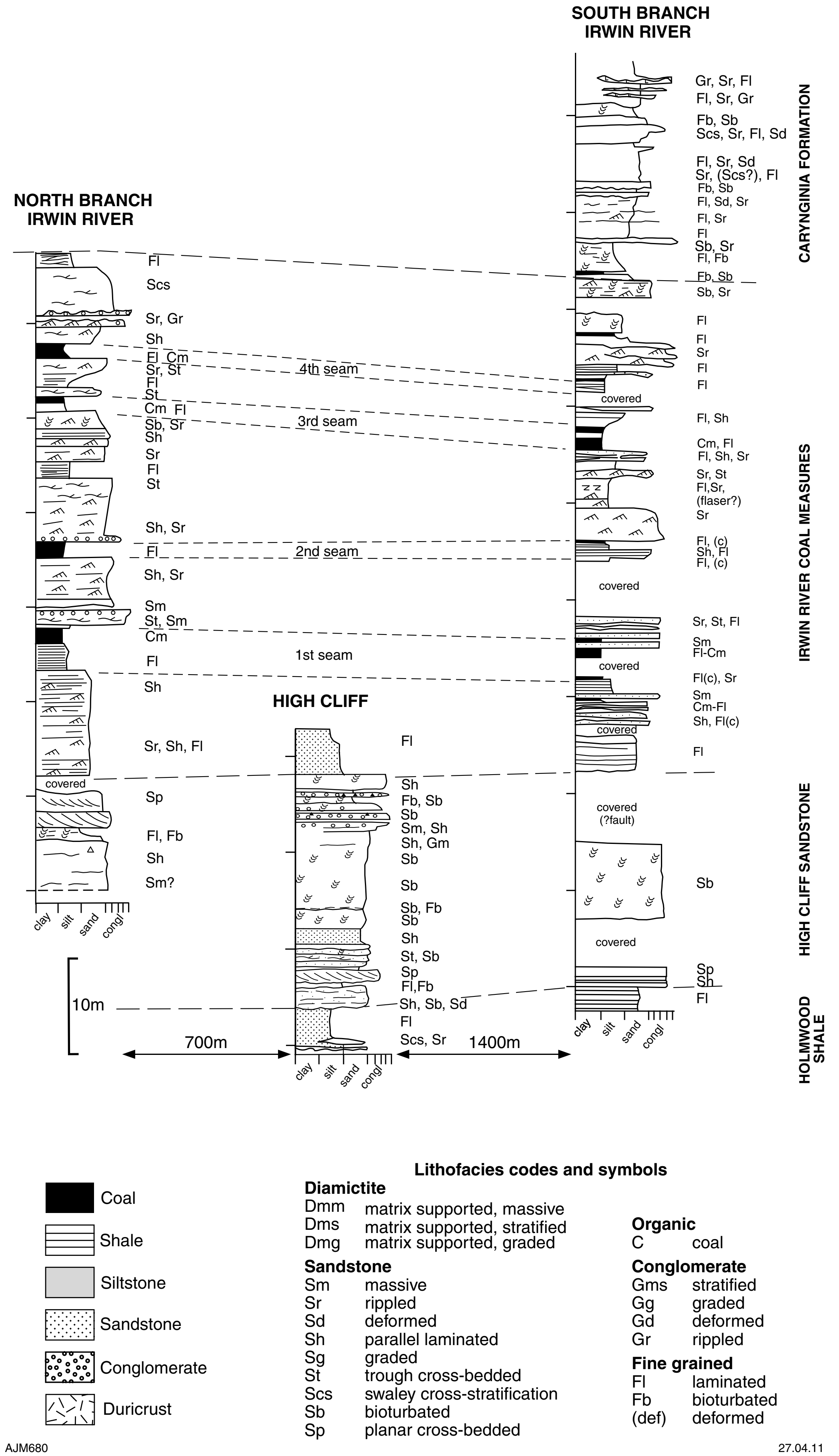

Figure 9. Measured sections, High Cliff Sandstone to Caryngina Formation, Coalseam Reserve

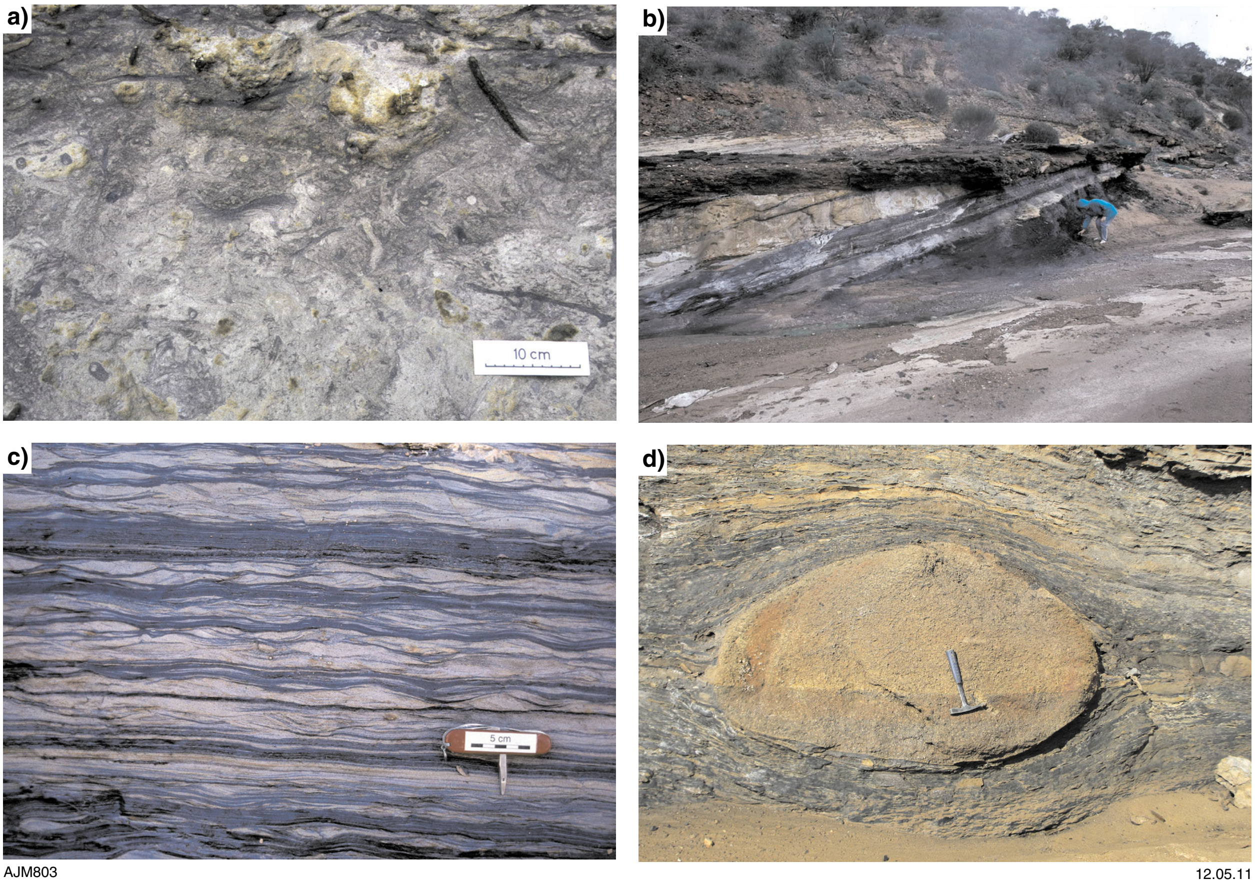

Figure 10. Lower Permian outcrops, Coalseam Reserve: a) bioturbation in the High Cliff Sandstone; b) coal seam in Irwin River Coal Measures, north branch Irwin River; c) wave ripples, Irwin River Coal Measures, south branch Irwin River; d) large dropstone, Carynginia Formation, south branch Irwin River

Environment of deposition: The invertebrate faunas, trace fossil assemblages, local hummocky cross stratification, wave ripples, and sporadic conglomeratic lenses (including dropstones) suggest deposition in shallow-marine to shoreface environments.

Irwin River Coal Measures

Lithology: The formation consists of a mixed succession of sandstone, siltstone, carbonaceous shale, and coal (Fig. 10b,c). Clarke et al. (1951) introduced this name for the coal-bearing section along the North Branch of the Irwin River (Fig. 10b) that lies conformably between 'paralic' strata of the High Cliff Sandstone (below) and marine siltstone of the Carynginia Formation (above).

Four principal coal seams are represented in the North Branch of the Irwin River. Several test drives were opened into the seams during the late 19th century and again in the 1940s but seam splits, discontinuities, and relatively high ash and sulfur contents discouraged further exploration until the 1980s when shallow drilling found nine seams south of the Irwin River exposures. The thickest known seam reaches 8 m in the Lockier Deposit, 12 km south of the outcrops on the South Branch of the Irwin River, where the cumulative coal thickness is 14 m (Le Blanc Smith and Mory, 1995). Rapid facies changes in the coals, and other lithologies, a complicate subsurface correlations and resource estimates.

Distribution and thickness: The unit is 66 m thick in the type section (Le Blanc Smith and Mory, 1995) and reaches about 300 m in the subsurface west of the Urella Fault

Fossils and age: Body fossils have not yet been recorded from the unit although invertebrate burrows are locally common. The unit contains an abundant, but typically low diversity, Early Permian Gondwanan flora incorporating species of *Glossopteris*, *Vertebraria*, *Gangamopteris*, *Sphenophyllum*, *Neomariopteris*, *Paracalamites*, *Lelstotheca*, and *Gondwanophyton* (in decreasing order of abundance; McLoughlin, 1992, 1993; McLoughlin and Hill, 1996). The floras contain a greater proportion of herbaceous plants compared to the gymnosperm dominated coeval floras of the Collie Basin coal measures south of Perth.

Palynological studies suggest an Artinskian age for this unit (Balme in McWhae et al., 1958; Segroves, 1971) based on stratigraphic position and correlation to marine successions in the Southern Carnarvon Basin. According to Backhouse (1993b, 1998), the Irwin River Coal Measures spans the *Striatopodocarpites fusus* and *Microbaculispora trisina* spore-pollen zones.

Environment of deposition: Although previously regarded as fluvial (McIntosh, 1980), the unit is interpreted to represent various delta plain depositional environments (Le Blanc Smith and Mory, 1995; Mory et al., 2005). Floral differences can be attributed to deltaic versus fluvial plain depositional settings for these respective coal measures. Plant fossils are most visible in the shale bed immediately above the fourth (highest) coal seam in the North Branch of the Irwin River.

Carynginia Formation

Lithology: The Carynginia Formation (amended from 'Carynginia Shale' of Clarke et al., 1951 by Playford and Willmott, in McWhae et al., 1958) consists of black to grey micaceous jarositic shale and siltstone with lesser interbedded sandstone and conglomerate. The type section is in Carynginia Gully, a tributary of the North Branch of the Irwin River. As these exposures are poor, Playford and Willmott (in McWhae et al., 1958) proposed a reference section at Woolaga Creek, 27 km south of Coalseam Conservation Park.

Sandstone and conglomerate intervals typically contain internal cross-laminae, commonly reworked by wave and storm action into symmetrical wave ripples and hummocky cross stratification. Erratic pebbles and boulders of granite and metamorphic rock are common within the unit (Fig. 10d). The lower part of the Carynginia Formation is well exposed in the South Branch of the Irwin River (Fig. 9). Much of the upper part of the formation in the Irwin River area is poorly exposed or concealed by Neogene duricrust or ferruginous river-terrace sandstone.

Distribution and thickness: The unit extends throughout the subsurface of much of the northern Perth Basin and is up to 300 m thick.

Fossils and age: Extensive bioturbation is common, with forms assignable to *Planolites*, *Rosselia*, *Teichichnus*, and *Phycodus*?, together with minor *Skolithos*. The only marine macrofossils recorded from outcrop of the Carynginia Formation are 'Aviculopectens and *Anthracosia*-like shells and occasional fish' reported by David and Sussmilch (1931), possibly from the fossil locality shown in 'Carynginia Creek' by Clarke et al. (1951). None of the rare marine invertebrate faunas that are present in petroleum exploration core have been described. A glossopterid fructification was found recently in the Carynginia Formation in the northern branch of the Irwin River by DWH.

Palynomorph assemblages from subsurface samples appear no younger than the *Praecolpatites sinuosus* zone (Segroves, 1971; Backhouse, 1993a). Correlation with more fossiliferous marine sections in the Southern Carnarvon Basin containing this palynomorph zone (Byro Group) indicates the formation is no younger than Kungurian.

Environment of deposition: The facies are similar to those of the Holmwood Shale suggesting a comparable environment of deposition. The siltstone-dominated lithology, wave reworking of sediments, dropstones, intense bioturbation, sporadic invertebrates, and abundant acritarchs (Segroves, 1971) indicate deposition within relatively low energy environments with restricted access to open-marine conditions. The unit correlates with the Byro Group of the Merlinleigh Basin but is of more-restricted marine facies.

Wagina Sandstone

Lithology: The Wagina Sandstone (Clarke et al., 1951) consists chiefly of fine- to medium-grained, cross-bedded, clayey sandstone with lesser amounts of conglomerate, siltstone, shale, and thin coal. Exposure along the type section in the South Branch of the Irwin River near Wagina Well is poor, so Playford and Willmott (in McWhae et al., 1958) proposed a reference section 25 km to the south, east of Woolaga Creek.

Distribution and thickness: The formation is up to 250 m thick, although the upper contact is not preserved in outcrop, and rests conformably or disconformably on the Carynginia Formation in outcrop. In the subsurface to the west a mild angular unconformity separates equivalent units from Lower Permian strata.

The Wagina Sandstone is the only Guadalupian formation that outcrops in the northern Perth Basin. It thins to the west of the outcrop belt into the coeval Dongara Sandstone which in turn interfingers with carbonates of the Beekeeper Formation to the south (Mory and Iasky, 1996). These units do not extend beyond the Beagle Ridge – Dongara Saddle, but the presence of over 650 m of coeval shale in Turtle Dove 1B, implies a second half graben was active west of the Dandaragan Trough during the Guadalupian.

Fossils and age: Fossil plants are scattered in these fluvial deposits. Palynological data indicate a Kungurian–Guadalupian age (Segroves, 1971; Kemp et al., 1977) based on indirect correlation to marine units elsewhere. Archbold (1995a, 1998b) described brachiopods from the Beekeeper Formation, to which he later (Archbold, 2002) ascribed a Wordian age, and suggested a slightly younger age (Wordian–Capitanian) for the Wagina Sandstone.

Southern Carnarvon Basin

Lower Carboniferous stratigraphy

Lower Carboniferous strata (Fig. 6) were first recognized in outcrop by Teichert (1949), where they are restricted to the eastern margin of the Carnarvon Basin between 'Williambury' to about 17 km south-southeast of 'Moogooree', a distance of just over 50 km. In the subsurface, Lower Carboniferous has been identified north of 24°S in Gnaraloo 1 and Quail 1 on the Gascoyne Platform and East Marrilla 1 in the northern Merlinleigh Sub-basin. Just four cores (totalling 6.6 m) were cut from the former two wells, and the available subsurface paleontological determinations are imprecise apart from late Visean palynomorphs (*Grandispora maculosa* zone) within the type section of the Quail Formation

(2101–2453 m in Quail 1), a unit entirely restricted to the subsurface west of the Merlinleigh Sub-basin.

Moogooree Limestone

Lithology: The Moogooree Limestone contains grainstones with ooids, cortoids, and oncolites, brachiopod and peloidal packstone, and laminated wackestone (Condon, 1965; Hocking et al., 1987). Most of the limestone is recrystallized, dolomitized, or dedolomitized (Radke and Nicoll, 1981), to the extent that the original microfabric is rarely preserved. Exposures are dominantly carbonates, but the little mineral exploration drilling into the unit reveals a repetitive pattern of friable units alternating with indurated carbonate beds. The facies interpretation of this cyclicity is difficult because of diagenetic alteration in the carbonate beds and lack of exposure of the friable units.

Most of the limestones have been dolomitized; some display 'birdseye' and other fenestral structures. The former presence of evaporites is inferred from quartz geodes formed by the replacement of early diagenetic anhydrite nodules, and patchy dedolomitization that has probably resulted in the light yellowish-brown colour of the carbonate beds (indicative of iron sesquioxide inclusions; Radke and Nicoll, 1981).

Distribution and thickness: The formation overlies the Willaraddie Formation, probably disconformably, and is overlain conformably by the Williambury Formation (or in some areas, by the Yindagindy Formation). The type section, about 5 km southeast of 'Williambury', is about 326 m thick and is structurally conformable with the Willaraddie Formation to the east (Hocking et al., 1987).

Fossils and age: The brachiopods *Rhipidomella michelina*? (Léveillé), *Shellwienella* (*Schellwienella*) *minilyensis* Thomas, *Septemirostellum amnicum* (Veevers), *Composita carnarvonensis* Thomas, *Unispirifer fluctuosus* (Glenister), *Unispirifer* sp. aff. *U. laurelensis* (Thomas), *Syringothyris spissus* Glenister, *Kitakamithyris moogooriensis* Thomas, *Punctospirifer mucronatus* Thomas, *Punctospirifer plicatosulcatus* Glenister, *Cleiothyridina minilya* Thomas are the only elements of the fossil assemblage from the Moogooree Limestone that have been described and illustrated (Glenister, 1955; Veevers, 1959; Thomas, 1971; Skwarko, 1988). Condon (1965) noted that 12 fossiliferous beds had been mapped in the formation and although these contain abundant specimens of a shelly fauna, specific diversity is low. Apart from brachiopods, bivalves, gastropods, crinoid columnal plates, the coral *Syringopora*, bryozoa, echinodermal debris, ostracods, microvertebrate remains, dasycladacean algae, and calcareous microbialites have been noted from the formation (Thomas, 1962; Condon, 1965; Lavaring, 1979; Hocking et al., 1987; Trinajstic and George, 2007).

An Early Carboniferous (Tournaisian) age of the unit is based on an indirect correlation by brachiopods that are present in few beds (Thomas, 1971). These belong to the *Unispirifer fluctuosus* Assemblage Zone of Thomas (1971) that is probably equivalent to the *Unispirifer laurelensis*, *Grammorhynchus eganensis*, and *Septemirostellum amnicum* zones that Roberts (1971) recognized in the upper member of the Laurel Formation in the Canning Basin, and in the upper part of the Burt Range Formation and the Enga Sandstone of the Bonaparte Basin. In these formations, the zones are associated with conodonts and foraminifera of probable late early to middle Tournaisian age based on the synthesis by Jones (2004) that draws especially from work by Mamet and Belford (1968) and Druce (1969, 1974). The conodont fauna of the Moogooree Limestone has yet to be described, but is a shallow-water association similar to that from coeval Tournaisian strata in the Canning and Bonaparte Basins and includes *Bispathodus aculeatus aculeatus*, *Clydagnathus cavusiformis*, *Polygnathus communis collinsoni*, *P. inornatus*, *Pseudopolygnathus dentilineatus*, and rare *Siphonodella* sp. (RS Nicoll, 2011, written comm.).

Environment of deposition: Hocking et al. (1987) indicate deposition was in peritidal to very shallow marine conditions with a transgressive–regressive cycle as shown by basal fenestral and cryptalgal carbonate mudstone, middle coarser grained deposits and a return to peritidal mudstone facies high in the unit.

Williambury Formation

The Williambury Formation consists of immature sandstone, conglomerate and siltstone lying conformably between the Moogooree Limestone and Yindagindy Formation. The type section, 4 km east-southeast of 'Williambury', is 235 m thick (Condon, 1965). The only fossils known from the unit are poorly preserved impressions of plant rootlets. An alluvial fan to fluvial environment of deposition is indicated (Hocking et al., 1987).

Yindagindy Formation

Lithology: The Yindagindy Formation consists of wackestone with lesser packstone and grainstone (some beds oolitic), and poorly exposed quartz and calcareous sandstone. Many of the limestone beds are dolomitized; some display 'birdseye' and other fenestral structures (Condon, 1967; Read et al., 1973; Hocking et al., 1987).

Distribution and thickness: The type section 4 km west of 'Williambury' is approximately 76 m thick (Hocking et al., 1987). The unit overlies the Williambury Formation and is overlain unconformably by the Lyons Group.

Fossils and age: An Early Carboniferous (Visean) age was suggested by rare brachiopods attributed to a new species of *Composita* that resembles a species from the Meramecian and Chesterian of North America and from the late Tournaisian to Visean in Kazakhstan (Thomas, 1971). Foraminifera (under study by DW Haig) and condonts support this determination. Other fossils include ostracods, gastropods, bryozoa, crinoids, serpulids, and algae (Thomas, 1962), none of which have been described.

Preliminary work on the Yindagindy Formation indicates the presence of the algae *Koninckopora* together with foraminifera including *Koktjubina* n. spp., *Palaeospiroplectammina* aff. *longula*, *Endothyra prisca*, *Eotuberitina reitingerae*, *Parathurammina* (*Suleimanovella*) *suleimanovi*, *Forschia*? spp., and *Haplophrgamina*? spp. (from 'unit B' at **locality 10**), and points to an Arundian to Brigantian age (Middle to Late Visean; Daniel Vachard, written comm., May 2011). By comparison, the limited conodont fauna from the formation includes *Synclydognathus* cf. *S. geminus* and *Clydagnathus cavusformis* (RS Nicoll, 2011, written comm.). Whereas *Synclydognathus geminus* has a Visean–Serpukhovian range (Medina-Varea et al., 2005), most species of *Clydagnathus* are Tournaisian (von Bitter and Plint, 1987). *S.* cf. *S. geminus* has been recorded previously in Australia (as *Spathognathodus scitulus*) from the uppermost Anderson Formation (late Visean) in the Canning Basin (865.6–880.9 m in Yulleroo 1) in association with *Cavusgnathus unicornis* (Jones et al., 1973; Nicoll and Druce, 1979) and from the Utting Calcarenite (early Visean) in the Bonaparte Basin (Druce, 1969).

Environment of deposition: Detailed facies analysis is not possible due to poor exposures, especially of the sandstone, but very shallow marine to peritidal conditions are deduced from the generally sparsely fossiliferous carbonate facies (Hocking et al., 1987).

Upper Carboniferous–Permian stratigraphy

Lyons Group

Lithology: Quartzitic, granitic and igneous dropstones, very poorly sorted sedimentary breccia and tillite, indurated contorted bedding often with micro-syndepositional faults, minor varved shale and great thicknesses of micaceous shale characterize the Lyons Group, and indicate glacial activity (Condon, 1967; van de Graaff, 1981). In places, indurated sandstone is present as well as some fossiliferous beds. Outcrop of the group is poor because of the predominance of friable units. The identification of stromatolites in the diamictite clasts suggests they originated from the Proterozoic Bangemall Basin to the northeast of the Southern Carnarvon Basin, as do orientations of striations on glacial pavements etched on the Precambrian rocks (Grey et al., 1977).

The type area of the group is near the Wyndham River (Condon, 1967) in the southern part of the Merlinleigh Sub-basin, and a reference area, designated by Condon (1954, 1967) is north of the Middalya–Williambury Road in the northern part of the basin. Condon (1967) divided the Lyons Group into, in ascending stratigraphic order, the Austin Formation (mainly quartzwacke), Coyango Greywacke (mainly greywacke with less quartzwacke), Dumbardo Siltstone, Koomberan Greywacke, Mundarie Siltstone, Thambrong Formation (alternating units of tillitic quartzwacke and siltstone) and Weedarra Shale. Later workers have been unable to use this nomenclature because the formations are poorly exposed, disrupted by faulting, and there are few marker beds for correlation. Following Hocking et al. (1987), the only formations recognized in the group are the shaley Carrandibby Formation, which has a restricted distribution at the top of the group near Callytharra Springs, and the Harris Sandstone at the base of the group in the Williambury–Moogooree area.

Distribution and thickness: The unit unconformably overlies Lower Carboniferous, Devonian, and Precambrian rocks. The thickness of the group penetrated in subsurface is about 1500 m. Seismic data, however, suggest that the unit may reach 3000 m adjacent to the Wandagee Fault on the western side of the Merlinleigh Sub-basin (Fig. 5, A–A').

Fossils and age: Marine fossils from the Lyons Group are present in a few outcrops within the Merlinleigh Sub-basin, and in BMR 8 and 9 borehole sections in the north of the Byro Sub-basin (Skwarko, 1993; Archbold, 1995b; Archbold and Hogeboom, 2000; Dixon and Haig, 2004). The fossils include foraminifera, unidentified corals, bryozoans, brachiopods, bivalves, gastropods, unidentified scaphopods, conulariids, crinoids, and plants (macrofossils, spores and pollen). Spinose acritarchs are apparently absent in palynological samples; this and the low diversity marine fossil assemblages indicate a restricted marine environment of deposition for at least parts of the group, very high sedimentation rates and water turbidity, or unfavourable climatic conditions.

The age of the Lyons Group depends on palynological correlation of the upper part of the Group (viz. Carrandibby Formation) in the Southern Carnarvon Basin with the Holmwood Shale (below the Fossil Cliff Member) in the Perth Basin, where ammonoids dated as Early Sakmarian (Tastubian) are present. Brachiopod and bivalve faunas in the upper Lyons Group support this age (Archbold et al., 1993). Because of the great thickness of Lyons Group below the Carrandibby Formation, the formation is considered to range into the Asselian, and perhaps older.

The Lyons Group in petroleum wells and shallow-water bores has yielded palynomorphs mainly from the *Pseudoreticulatispora confluens* zone (Mory and Backhouse, 1997). This zone is present also throughout most of the Holmwood Shale in the Perth Basin (Backhouse, 1998). The identification of the underlying 'Stage 2' zone (*sensu* Backhouse 1991; equivalent to the *Microbaculispora tentula* zone discussed by Mory, 2010) is unclear due to low yields and poor preservation, especially in deeper petroleum wells. Nevertheless, palynomorphs from the *Deusilites tenuistriatus* zone (Apak and Backhouse, 1998, 1999), tentatively placed within the upper Pennsylvanian (Mory, 2010), have been recovered from water bores near the base of the group on 'Lyons River', implying the presence of the younger 'Stage 2' zone within the group. Eyles et al. (2002) and Stephenson (2008) discussed the broad geographic extent of the Early Permian and Late Carboniferous zones and the tentative position of the Permian–Carboniferous boundary in Gondwana.

Environment of deposition: The initial uneven topography of the basin was rapidly infilled by strata of the Lyons Group. At the end of deposition of the group, the seafloor gradient was very low and the water depth throughout the basin was very shallow. The present-day Baltic Sea probably provides a good modern analogue for the initial glacially sculptured basin. Based on a facies analysis of borehole sections, Eyles et al. (2003) suggested that subaqueous gravity flows were responsible for much of the deposition of the Lyons Group, with alternating diamictite and shale recording successive intervals of fault-related uplift.

Harris Sandstone

The Harris Sandstone is composed mainly of quartz sandstone (Condon, 1954, 1967). Lepidodendroid plant fossils in the unit (White and Condon, 1959), dated rather tentatively as latest Carboniferous or earliest Permian (McLouglin and Hill, 1996), are considered indicative of an alluvial plain environment (McLoughlin, 1993).

According to Condon (1967), the formation is present only in its type area, a narrow 11-km long outcrop belt about 500 m west of Williambury homestead. Here it overlies the Visean Yindagindy Formation disconformably, and is overlain by poorly exposed undifferentiated Lyons Group. Condon (1967) interpreted the formation to be a lateral equivalent of the 'Austin Formation' at the base of the Lyons Group. Hocking et al. (1987) regarded the two formations as synonymous and abandoned use of the 'Austin Formation'. Later work in the Moogooree area (including **locality 9**) by Read et al. (1973) found a concordant sandstone unit lacking glacigene influence above the Yindagindy Formation, which they considered a lateral equivalent of the Harris Sandstone. Read et al. (1973) and van de Graaff (1981) suggested that an angular unconformity exists between the Lyons Group and older formations with progressive overstepping of units.

On the basis of underlying depositional sequences, it seems logical that a prograding sandstone unit could conformably cap the Yindagindy sequence. It is therefore possible that at least parts of what Hocking et al. (1987) took as the Harris Sandstone may belong within a Visean (perhaps to Serpukhovian) Yindagindy depositional sequence within the broad Ordovician to Middle Carboniferous interior sag basin. The Lyons Group, confined mainly to the Merlinleigh interior rift basin, would unconformably overlie this sequence. Such a hypothesis needs to be tested by subsurface correlations (using the spore-pollen zonation).

Carrandibby Formation

The Carrandibby Formation has proved impossible to differentiate from the Callytharra Formation in subsurface data. In outcrop the Carrandibby Formation is distinguished from the overlying Callytharra Formation as it 'locally contains scattered glacial erratics' and has a low faunal diversity (Hocking et al., 1987), but this definition is difficult to apply to core (and clearly impossible on electric logs alone). The only dropstone present in core above the undifferentiated Lyons Group is in Ballythanna 1 is a red and green chert pebble 30 cm above the base of the Callytharra Formation. The only other record of glacial influence in the Callytharra Formation is in the Gooch Range area north of 'Middalya' where a thin limestone, containing dropstones, was referred to as Carrandibby Formation equivalent by Hocking et al. (1987, fig. 51). The Carrandibby Formation is restricted to the Byro Sub-basin and an area south and east of Pells Range in the south of the Merlinleigh Sub-basin, but is less than 10 m thick in many sections and thus not practical to map. In the type section, immediately downstream from that of the Callytharra Formation, the unit is 57 m thick but the base is not exposed.

Callytharra Formation

Lithology: The Callytharra Formation consists of mudstone, sandy mudstone and minor sandstone with interbeds of limestone marking the top of depositional cycles. At the type section at Callytharra Springs in the southern part of the Merlinleigh Basin (redescribed by Dixon and Haig, 2004), the unit lies conformably above the Carrandibby Formation (uppermost part of Lyons Group) with the lowest limestone bed marking its base. The top of the formation in this section, although poorly exposed, appears conformable with the overlying 'Nunnery Sandstone' (an abandoned unit considered part of the Moogooloo Sandstone by van de Graaff et al. (1977), but more likely to be at least partially equivalent to the Winnemia Sandstone Member of the Callytharra Formation, Mory and Backhouse (1997)). Correlations south to the Fossil Cliff Member in the Irwin Sub-basin (**locality 1**) and north to sections in the Merlinleigh Sub-basin, including the Winnemia Sandstone Member and the Jimba Jimba Calcarenite (Mory and Backhouse, 1997), are discussed in the outline of **locality 5**. Dixon and Haig (2004) suggested that the type section of the Callytharra Formation includes only the lower part of the unit recognized further north in the Merlinleigh Sub-basin.

Fossils and age: The Callytharra Formation contains a diverse fossil assemblage including foraminifera, corals, bryozoans, brachiopods, bivalves, gastropods, ammonoids, nautiloids, conulariids, blastoids, crinoids, ostracods, annelids, conodonts, fish micro-remains, plants (including spores and pollen), and acritarchs (Skwarko, 1993; Webster and Jell, 1992, 1999; Archbold and Hogeboom, 2000; Dixon and Haig, 2004).

Age determination of the Callytharra Formation is difficult because of the endemic nature of the marine faunas and, in particular, the low diversity and scarcity of conodonts and ammonoids, and the absence of fusuline foraminifera. A late Sakmarian to early Artinskian age has been attributed to the Callytharra Formation as now recognized in the basin (Mory and Backhouse, 1997) based on meagre conodont evidence (Nicoll and Metcalfe, 1998) from the lower and upper part of the formation, and three ammonoid identifications (Glenister and Furnish, 1961; Cockbain, 1980; Leonova, 1998) from low in the formation. The evidence relies mainly on correlations between the Callytharra Formation and units in other basins of the East Gondwana rift system that contain more diverse ammonoid and conodont assemblages. The local conodonts and ammonoids have been compared to the more diverse, more open-marine, and warmer-water assemblages known from the Maubisse Formation in Timor nearer the northern margin of the rift system (see Charlton et al., 2002). Brachiopod zones (Archbold, 1993, 1998a,b, 1999; Archbold and Shi, 1995; Hogeboom and Archbold, 1999) and palynomorph zones (Mory and Backhouse, 1997; Backhouse, 1998) link the Callytharra Formation to the Fossil Cliff Member in the northern Perth Basin and the Nura Nura Limestone in the Canning Basin, which contain additional ammonoid species (Foster and Archbold, 2001). Brachiopod (Archbold in Charlton et al., 2002) and crinoid correlations (Webster in Charlton et al., 2002) to the Maubisse Formation in Timor have supported the age determinations with the crinoids suggesting a slightly younger (early Artinskian) age for the type section of Callytharra Formation.

Environment of deposition: The Callytharra Formation represents a period of deposition in a quiet marine environment (similar to that of the underlying Carrandibby Formation) but with a much reduced glacial influence. The presence of small amounts of pyrite throughout the carbonaceous siltstone in the lower part of the formation implies that, at times, conditions were anaerobic.

Wooramel Group

The Wooramel Group represents a succession of quartz sandstone with minor conglomerate and siltstone, 50 to 460 m thick, which lies conformably between the Callytharra Formation and the Byro Group. In the Merlinleigh Sub-basin, the Wooramel Group comprises the Cordalia Formation, Moogooloo Sandstone and Billidee Formation, in ascending order. South of the Merlinleigh Sub-basin the group is represented by a single formation (Keogh Formation). The Group includes the late progradational part of a major post-glacial Sakmarian to Artinskian depositional cycle and the early retrogradational part of an Artinskian to Kungurian depositional cycle. It provides evidence for the latter part of a warming trend followed by cooling of the paleoclimate. Paleocurrent directions throughout the group are predominantly to the north-northwest.

Condon (1954, 1967) and many subsequent workers including Hocking et al. (1987) considered the base of the group to be an unconformity based on: 1) a strongly karstified contact apparently between the Moogooloo Sandstone and underlying limestone of the Callytharra Formation, in the south of the Merlinleigh Sub-basin; 2) the assumption that the Cordalia Formation is a lateral equivalent of the Moogooloo Sandstone. The palynological evidence, however, indicates that the Cordalia Formation is a lateral equivalent of the upper Callytharra Formation suggesting that, if there is a break in sedimentation at the base of the Wooramel Group, it is too brief to detect with the palynological control available at present (Figs 6 and 7; Mory and Backhouse, 1997). Crostella (1995) suggested the karst is probably Cenozoic in age. In the excursion areas (**localities 5 and 8**), the contact appears to be conformable within a progradational depositional cycle.

The Billidee Formation (**locality 8**) is probably better placed in the Byro Group because it commences a retrogradational trend which extends through the conformably overlying Coyrie Formation (**locality 8**) and Mallens Sandstone to a maximum flooding level in the Bulgadoo Shale (**locality 4**).

Cordalia Formation

Lithology: The Cordalia Formation (Cordalia Sandstone of Condon, 1967; Hocking et al., 1987) outcrops along Gooch Range in the northern part of the Merlinleigh Sub-basin. In its type section next to the Lyndon River (at 23°20'58"S, 114°4-'08"E, WINNING POOL), it includes mainly laminated to thin-bedded, fine-grained quartzwacke. At **locality 5**, about 68 km southeast of the type locality, the Cordalia Sandstone is thicker bedded and slightly coarser grained than in the type section. In the Toby Bore area (**locality 8**) on the eastern side of the Kennedy Range, about 100 km southeast of the type section, the formation is thin to medium bedded. In the subsurface, carbonaceous siltstone that coarsens upwards towards the base of the overlying Moogooloo Sandstone is placed within the formation.

Distribution and thickness: The unit is 53 m thick at its type section and, in the outcrop belt, thins to the south to pinch out just south of the Toby Bore area (**locality 8**) where it is 1–2 m thick. West of outcrop of the unit, it extends at least as far south as Kennedy Range 1, about 132 km south of the type locality, where it is 62 m thick but appears to be separated from the underlying Callytharra Formation by a fault. The unit also has been identified as far north as Remarkable Hill 1 (68 km northwest of the type section the unit is 94 m thick (Mory and Backhouse, 1997)).

Fossils and age: The Cordalia Formation is sparsely fossiliferous with most fossils coming from near the base of the unit. Fenestellid bryozoans, brachiopods, an ammonoid, and trace fossils have been recorded from the formation (Skwarko, 1993; Archbold and Hogeboom, 2000). The ammonoid *Pseudoschistoceras simile* Teichert (1944, p. 87–89, pl. 17, figs 1–5, text-figure 4) has been critical in placing the Cordalia Formation as no older than the Baigendzhinian of the late Artinskian (Glenister and Furnish, 1961; Cockbain 1980). An Artinskian upper age limit is suggested by the presence of *Aricoceras* and *Bamyaniceras* in the stratigraphically higher Billidee Formation (Leonova, 1998). In the subsurface, the Cordalia Formation contains the upper part of the *Striatopodocarpites fusus* (*Didecitriletes byroensis* subzone) and the lower part of the *Microbaculispora trisina* spore–pollen zones (Mory and Backhouse, 1997). Both the palynoflora and the character of the gamma ray log indicate that the unit is laterally equivalent to the uppermost part of the Callytharra Formation farther south in the basin (compare Burna 1 with Gascoyne 1; Mory and Backhouse, 1997, plate 1).

Environment of deposition: From outcrop, Hocking et al. (1987) interpreted the Cordalia Formation as a prodeltaic unit, but the dominance of thin bedded, bioturbated, fine-grained sandstone, indicates that a more proximal position on the paleoslope, such as a lower delta-front environment, is also possible. A very low gradient seafloor is implied from its stratigraphic position below the Moogooloo Sandstone, its relatively fine grain size, the coarsening-upwards sequence into the Moogooloo Sandstone, its large areal extent and its almost horizontal inclination in subsurface. In the subsurface west of the outcrop belt the unit is dominated by carbonaceous siltstone devoid of macrofossils, and probably represents deposition in an anoxic environment.

Moogooloo Sandstone

Lithology: The Moogooloo Sandstone consists mainly of medium to thick bedded, well-sorted quartz arenite (Condon, 1967; Hocking et al., 1987). Trough cross-bedding is dominant with some deformed bedding (both parabolic recumbent and contorted) developed in places near the base of fining upward cycles (Hocking et al., 1987). Aspects of diagenesis and petroleum reservoir characteristics in subsurface sections of the formation are discussed by Havord (1997, 1998), Baker et al., (2000), and Ghori et al. (2005).

Distribution and thickness: The unit is restricted to the Merlinleigh Sub-basin, and ranges in thickness from 30 m in the southeast to 115 m in Giralia 1 in the northernmost part of the basin. At the type section in the Gooch Range (23°51'1"S, 114°49'24"E, Winning Pool) the formation is 52 m thick (Condon, 1967). The Moogooloo Sandstone conformably overlies the Callytharra Formation south of Gooch Range and Kennedy Range 1, whereas to the north, including at the type section, the Moogooloo Sandstone lies conformably above the Cordalia Formation. The Billidee Formation conformably overlies the Moogooloo Sandstone at the type section and at **locality 8**.

Fossils and age: The sparse marine fossils recorded by Condon (1967) and Skwarko (1993) come from poorly located samples that may even be from the basal Cordalia Sandstone or basal Billidee Formation. In the sub-surface the Moogooloo Sandstone contains palynomorphs of the Artinskian *Microbaculispora trisina* zone (Mory and Backhouse, 1997; Backhouse, 1998).

Environment of deposition: Hocking et al. (1987) interpreted much of the formation in outcrop as deposited under fluvial-dominated conditions with increasing marine influence to the north. A study of subsurface sections to the west of the outcrop belt concluded that the formation was deposited under storm-influenced shallow-marine conditions (Eyles et al., 2003). Paleocurrent directions indicate a northwest direction (Hocking et al., 1987). The lower and middle parts of the Moogooloo Sandstone represent progradation, probably in a northwest direction, of fluvial-deltaic facies over shoreface facies (Hocking et al., 1987; Eyles et al., 2003). Contorted soft-sediment deformation may have been associated with episodes of faulting (Eyles et al., 2003). The upper part of the Moogooloo Sandstone, below the contact with the type section of the Billidee Formation, shows a retrogradational trend (discussed further under **locality 8**). A boundary between two major depositional sequences is interpreted within the uppermost Moogooloo Sandstone (Fig. 7).

Billidee Formation

Lithology: The Billidee Formation contains carbonaceous shale, silty sandstone, fine to coarse-grained sandstone, and pebble conglomerate. At the type section near Toby Bore (**locality 8**), Condon et al. (1987, fig. 79) placed the lower boundary of the Billidee Formation at the base of the lower of two friable dark grey laminated mudstone units in the section. This unit is conspicuous on aerial photographs (see **locality 8**) but seems to be stratigraphically above the boundary mapped by Condon et al. (1987, fig. 78). In the type area, the second significant mudstone interval lies in the middle of the formation and forms the lower part of a major scarp. Underlying and overlying this unit are thin to medium bedded highly ferruginized quartz wackestone beds, some of which are fossiliferous, particularly near the top of the formation. The top of the formation in the type section was placed at the contact of a thick fossiliferous micaceous quartzwacke with overlying friable laminated mudstone (Coyrie Formation). Rare dropstones are present low in the formation.

Distribution and thickness: The formation appears to be restricted to the Merlinleigh Sub-basin and ranges in thickness from 50 m 130 km south of the type area to 209 m in Giralia 1 (160 km north-northwest of the type area). In the type area the thickness has been variably measured as 74 m by Condon et al. (1987), 90 m by Hocking et al. (1987), and about 60 m (this publication).

Fossils and age: The fossil assemblage from the Billidee Formation (as distinct from the Jimba Jimba Calcarenite, once placed as a member of the formation) includes foraminifera, bryozoa, brachiopods, bivalves, gastropods, ammonoids, nautiloids, crinoids, and fossil wood (Condon, 1967; Skwarko, 1993; Webster and Jell, 1999; Haig, in prep.), but most are undescribed. The ammonoids include *Neocrimites* sp. (Glenister and Furnish, 1961; Cockbain, 1980), which Leonova (1998) considered better placed in *Aricoceras*, and *Bamyaniceras* sp. (Cockbain, 1980). According to Leonova (1998) these are indicative of the Tethyan Yakhtashian and Bolorian (upper Artinskian and Kungurian, according to Ogg et al., 2008).

Environment of deposition: The Billidee Formation was deposited under cold-water conditions with some sea ice as indicated by occasional dropstones. The sands were deposited as shallow banks with salinity approaching normal-marine levels particularly toward the latter period of sedimentation. Mudstone facies are devoid of shelly macrofossils but contain an abundant assemblage of siliceous organic-cemented agglutinated foraminifera, probably indicative of a shallow silled basin under possibly brackish, low-oxygen, cold-water conditions.

Byro Group

Lithology: The Byro Group is characterized by carbonaceous mudstone alternating with fine-grained, hummocky cross-stratified and bioturbated quartz sandstone (Condon, 1967; Moore et al., 1980a; Hocking et al., 1997). Mudstone–sandstone successions define shoaling upward sequences at various scales in a hierarchy that may correspond to parasequences – parasequence sets – sequences. Mud-dominated formations alternate with sandstone-dominated units. An overall retrogradational trend is interpreted extending from the Billidee Formation (see above) through the Coyrie Formation – Mallens Sandstone to maximum flooding in the lower part of the Bulgadoo Shale. An overall progradational trend is interpreted from the Bulgadoo Shale – Cundlego Formation ascending through the Quinnanie Shale – Wandagee Formation – Nalbia Sandstone and terminating in the Baker Formation – Kennedy Group cycle (Fig. 7). Type areas of formations belonging to the Byro Group at **localities 8**

(Coyrie Formation, Mallens Sandstone), **3** (Bulgadoo Shale) and **2** (Quinnanie Shale, Wandagee Formation) are discussed under '**Excursion localities'**.

Distribution and thickness: The group is restricted to the Merlinleigh and Byro Sub-basins where it probably does not exceed 1500 m based on seismic data and the thicknesses of individual units in outcrop. Only one well (Kennedy Range 1) contains a complete thickness (1363 m) of the group. North of the Merlinleigh Sub-basin the group appears to be absent, with the Kennedy Group sitting disconformably on the Callytharra Formation.

Fossils and age: Throughout the Byro Group, shelly macrofossils are usually confined to relatively few sandstone beds. Mudstone facies are usually devoid of macrofossils but contain an abundant assemblage of siliceous organic-cemented agglutinated foraminifera. Only in a few localities are these foraminifera accompanied by calcareous hyaline and more rarely porcelaneous types and by ostracods.

The Byro Group is placed within the Kungurian (see discussions under **localities 2, 3** and **8**) and encompasses just the *P. sinuosus* and uppermost *M. trisina* spore-pollen zones (Mory and Backhouse, 1997; Backhouse, 1998).

Environment of deposition: Despite the frequent alternations in lithofacies signifying corresponding changes in bathymetry (from shallow-offshore below wave-base to above wave-base in the low gradient interior sea), and marked changes in macrofauna from formation to formation (Skwarko, 1993), the climate remained relatively uniform throughout deposition of the group. Thus, water-depth changes and different faunal migrations may be due to tectonic activity and changing marginal-sea configurations associated with breakup on the Tethyan margin of Gondwana.

Kennedy Group

Lithology: The Kennedy Group was originally defined for the dominantly sandy succession above the Byro Group in outcrop between the Gascoyne and Minilya rivers, and comprises the Coolkilya Sandstone, Mungadan Sandstone, and Binthalya Formation, in ascending order (Fig. 7; Condon, 1954, 1967). Subsequently the group was extended north into the subsurface based on the similarity of the palynoflora in wells such as Onslow 1 (Balme *in* Jones, 1967), Hope Island 1 (Balme *in* Bowering, 1968) and Cunaloo 1 (Wiseman and Dolby *in* Meath, 1972) to that of Merlinleigh 1 (next to Kennedy Range 1 in the centre of Kennedy Range). The subsurface intersections in the north of the basin were later referred to the Chinty Formation as they were thought to be younger than, and therefore distinct from, the Kennedy Group in the Merlinleigh Sub-basin (Hocking et al., 1987).

Distribution and thickness: The group is restricted to the Merlinleigh Sub-basin and is likely to be 500–1000 m thick based on the thicknesses of individual formations.

Fossils and age: In the vicinity of Kennedy Range the age of the group is constrained by palynomorphs of the uppermost *Praecolpatites sinuous* to *Dulhuntyispora granulata* zones (Mory and Backhouse, 1997; Mory et al., 2003), which indicate a late Kungurian to Roadian age (after Nicoll et al., 2009). Archbold (1998a) records brachiopods of the *Fusispirifer coolkilyaensis* and *Neochonites (Sommeriella) afanasyevae* Zones which he assigned an early to middle Ufimian age, but subsequently modified to Roadian (Archbold, 2002). The only conodont known from the group is *Vjalovognathus* sp. nov. A (Nicoll and Metcalfe, 1998) in the Coolkilya Sandstone, but the stratigraphic range of this species is uncertain. Farther north in the Peedamullah Shelf, Mory and Backhouse (1997) included Roadian to likely Lopingian strata in the group, but the overlap of just the *D. granulata* zone with strata in the Kennedy Range area suggests that such an extension is not justified.

Environment of deposition: Moore et al. (1980b) proposed a composite depositional model for the Kennedy Group: a moderate paleoslope model with a narrow transition between shoreface and offshore environments corresponding to the Coolkilya Sandstone, and a broad sandy 'shelf' with a low paleoslope corresponding to the Mungadan Sandstone and Binthalya Formation. Hocking (1990) suggested that the Mungadan Sandstone represented renewed deltaic progradation, in a subaqueous but fluvially-dominated setting. For the uppermost Permian preserved on the Peedamullah Shelf, Hocking et al. (1987) suggested deposition in a sandy 'marine shelf'.

Excursion localities

Locality 1: Irwin River Coalseam Conservation Park

Significance: Outcrop sections at this locality of the Holmwood Shale (includes the type section of the Fossil Cliff Member), High Cliff Sandstone (type section), Irwin River Coal Measures (type section), and Carynginia Formation, in ascending order, are the best Lower Permian exposures in the Perth Basin and allow comparison with coeval sections 550 km to the north in the Merlinleigh Sub-basin (**localities 3–5, 7**). The outcrops show a transition from glacial marine to deltaic to restricted marine facies.

Location: Riverbank and cliff exposures along Irwin River along the Irwin River about 28 km north-northeast of Mingenew; Fossil Cliff: 28°56'37"S 115°32'49"E; High Cliff: 28°56'46"S 115°32'55"E; North Branch: 28°56'22"S 115°32'52"E to 28°56'14"S 115°33'03"E; South Branch: 28°57'28"S 115°33'04"E to 28°57'36"S 115°33'26"E (YALGOO). Note that Coalseam Conservation Park is administered by the Department of Environment and Conservation who do not allow collecting of samples without a permit.

Geology: Coalseam Conservation Park contains the best-exposed Lower Permian sections in the basin along the North and South Branches of the Irwin River where the upper part of the Holmwood Shale, High Cliff Sandstone, Irwin River Coal Measures, and Carynginia Formation are exposed. These sections show a variety of sedimentary structures, and have been affected by minor normal faults with similar orientations and sense of movement to major faults in the region, and rare minor thrust faults.

Fossil Cliff, on the north side of the river, is the type section of the Fossil Cliff Member at the top of the Holmwood Shale (Fig. 8). The member consists of grey, fossiliferous siltstone with thin to medium beds of fossiliferous limestone. The southeastward steepening of dips in Fossil Cliff may be the result of rotation along a normal fault (possibly controlling the river course) because there is a minor normal fault dipping steeply northwest on the opposite side of the river. Note that extensive jointing is parallel to this structure.

At High Cliff on the south side of the river, the contact between grey mudstone of the Holmwood Shale and bioturbated muddy sandstone of the High Cliff Sandstone is clearly visible. The latter unit consists of a broadly upward-coarsening sequence of highly bioturbated silty sandstone (Figs 9 and 10a). The predominantly carbonaceous and highly bioturbated sandstone beds contain large angular to subangular granite, quartzite and chert erratics, up to 60 cm across. The contact with the overlying Irwin River Coal Measures is placed at the first appearance of carbonaceous siltstone about two-thirds of the way up the section. The oblique orientation of most of this exposure makes the discordance in bedding between the Permian section, which dips 10° east, and the flat lying Cenozoic section difficult to locate (Fig. 11) on the outcrop.

The type section of the Irwin River Coal Measures is exposed on the south side of the river upstream from High

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Figure 11. High Cliff showing contact between Lower Permian and Cenozoic. Panorama compressed 50% horizontally

Cliff and Fossil Cliff, and contains four low-rank coal seams interbedded with coarse- to fine-grained sandstone, siltstone, and claystone (Fig. 9).

Siltstone, mudstone, and thinly bedded sandstone of the overlying Carynginia Formation are well exposed in the South Branch of the Irwin River (Fig. 9). Bioturbation and abundant carbonaceous material indicate deposition in either a shallow-marine or lacustrine environment; however, trace fossils on the base of some sandstone beds are more suggestive of marine conditions. Scattered dropstones in the section (Fig. 10d) indicate glacial conditions continued into the Artinskian. Small scale slumps suggest some instability during deposition, and rippled poorly sorted coarse-grained to pebbly sandstone beds and possible hummocky crossbedding indicate this material was dumped from ice and further transported by wave activity. The contact with the underlying coal measures in this section is abrupt, and may be a third- or fourth-order sequence boundary. In the North Branch, the contact lies on the northern bank of the river, but it is often covered by sand. It is marked by an erosion surface with granules and some small pebbles at the base, and is immediately overlain by typical Carynginia Formation shale–sand facies.

Locality 2: Carbla Point, Shark Bay

Significance: Shark Bay is a World Heritage Area internationally renowned as a modern analogue for carbonate deposition under elevated salinity conditions. Carbla Point is a classic locality for the study of stromatolites and microbial mats.

Location: Carbla Point lies within Carbla station on the eastern shore of Hamelin Pool at 26°15'50"S 114°13'01"E, YARINGA. Note that Shark Bay is covered by a marine park administered by the Department of Environment and Conservation who do not allow samples to be collected without a permit.

Background*:* In Western Australia, extensive microbial mats were first described by Clarke and Teichert (1946) from the inland Lake Cowan about 600 km east of Perth. The microbial mats and stromatolites in Hamelin Pool were recognized first in 1954 when Western Australian Petroleum Pty Ltd (WAPET) was exploring the region (Playford and Cockbain, 1976). Brian Logan commenced PhD studies on Hamelin Pool and other areas of Shark Bay in 1956. Shark Bay was placed on the World Heritage list in 1991 for its natural heritage values, meeting criteria for natural beauty, biological diversity, ecological processes, and Earth's history. Of these criteria, the last three are strongly related to why Shark Bay is used as an analogue for many ancient depositional systems. Further information on the heritage values of Shark Bay can be obtained from www.sharkbay.org/.

Modern environments and the Quaternary stratigraphy of Shark Bay (Fig. 12) have been described in Logan et al. (1970, 1974b). Of particular interest are: Hagan and Logan's (1974) discussion on the development of carbonate banks and hypersaline basins; the description of microbial mats and related sedimentary fabrics by

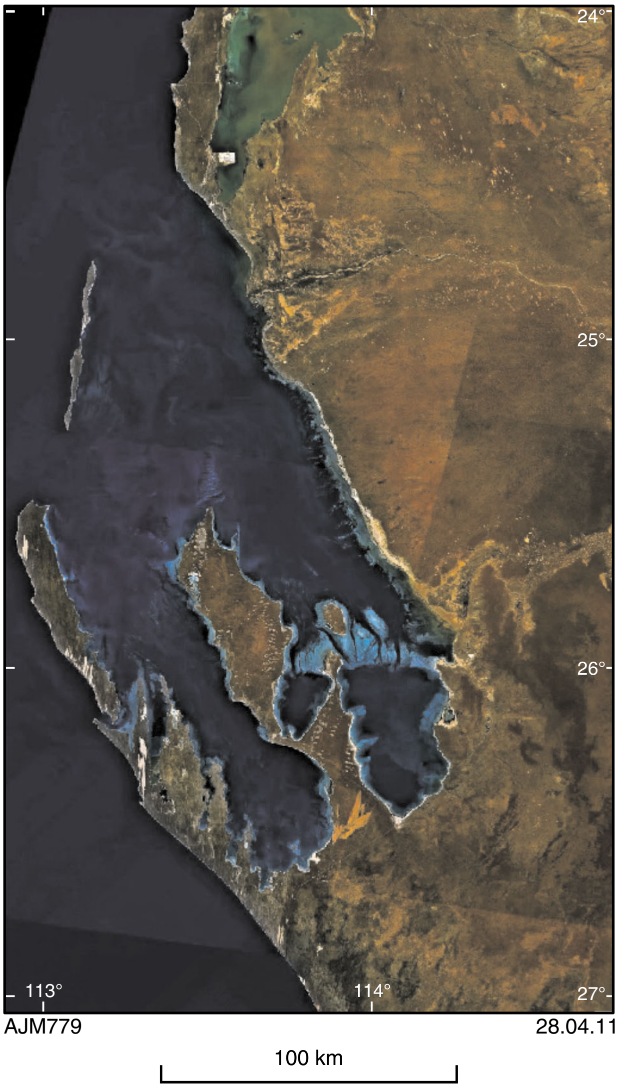

Figure 12. Satellite image of Shark Bay (image STS-67 Shark Bay Aus gif, NASA Flight 68 STS-67 accessible from http://spacelink.nasa.gov/)

Logan et al. (1974a); and the discussion of carbonate bank sedimentation along the eastern side of Shark Bay by Davies (1970).

The climate of Hamelin Pool (Shark Bay) is semi-arid, with annual evaporation of about 300 cm, (Luke et al., 1987) exceeding mean annual rainfall of 21 cm by an order of magnitude. Most rain falls in winter (May–July), but there are occasional heavy cyclonic rains in December–March. Strong southerly winds prevail throughout summer, and also during winter but with lower velocity especially in the southeastern part of Shark Bay. In the eastern part of Hamelin Pool (at Hamelin Pool station) winds commonly shift to the west in the afternoon (summarized from data for Hamelin Pool and Denham from the Bureau of Meteorology website www.bom.gov.au/climate/data).

Based on data presented by Logan et al. (1970, 1974a), Figure 13 shows the bathymetry of Shark Bay, Figure 14 shows temperature variations, and Figures 15 and 16 show the generalized salinity variation within the bay. Logan

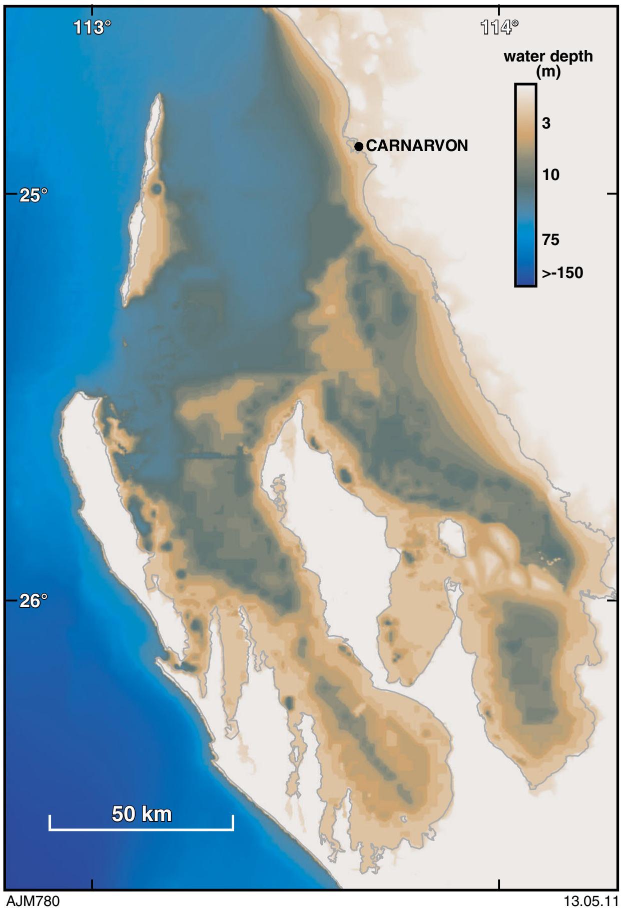

Figure 13. Bathymetry of Shark Bay

and Cebulski (1970) recognized three main divisions of the embayment waters based on salinity ranges: 1) oceanic with a range from 35–40‰; 2) metahaline with a range from 40–56‰; and 3) hypersaline with a range from 56–70‰. They emphasized that the salinity zones (Figs 15, 16) are maintained by: 1) the topography of the bay with barriers (such as the Faure Sill) partitioning various water bodies; 2) the north–south elongation of the embayments where prevailing southerly winds generate heavy seas and mix the water column; 3) the semi-arid climate and lack of freshwater inflow, where evaporation exceeds precipitation; and 4) the prevailing southerly winds which interact with tidal currents setting up an ebb-current dominance on the west-facing coasts and a flood-tide dominance on the east-facing coasts, creating a circulation system that involves slow rates of exchange.

Logan and Cebulski (1970) classified the Shark Bay environments into three groups: 1) intertidal–supratidal platform above the prevailing low-water level; 2) the sublittoral platform forming a 'shelf' around the margins of the bay and on the sills, with water depths to about 12 m; and 3) the embayment plain occupying the basinal areas of the bay at water depths below about 7–9 m. Table 1 summarizes environments of the sublittoral platform, and Table 2 summarizes environments of the embayment plain.

Shark Bay is a major site of biogenic carbonate sedimentation. Figure 17 shows the percentage of $CaCO_3$ in surface sediment samples studied by Logan (1959). These values encompass both modern biogenic and reworked carbonate grains. Logan and Cebulski (1970) noted that among shell-secreting organisms, the major contributors (in general order of abundance) are coralline algae, molluscs, foraminifera, echinoids, serpulids, and bryozoans. The biogenic grains are distributed according to the salinity gradient with corals and complex hyaline foraminifera confined to sediment deposited at salinities below 40 psu (practical salinity units: 1 psu ≈1‰); carbonate-cemented agglutinated foraminifera and echinoid plates and spines below 50 psu; and bryozoan and coralline algal debris below 60 psu (Mossadegh et al., 2009). At salinities above 60 psu, the biogenic component of the sediment is dominated by porcelaneous foraminifera, bivalve and gastropod shells of few species, and green algal debris. An attempt has been made to quantify the coralline algal contribution to the rate of carbonate sedimentation in the metahaline and normal-marine parts of the bay (Walker and Woelkerling, 1988), with a conclusion that $CaCO_3$ secreted by epiphytes on seagrass leaves (*Amphibolis antarctica*) could account for a carbonate-sediment depositional rate of 0.5 mm/yr.

The significance of seagrass stands as a habitat for carbonate-secreting organisms, a baffle for trapping sediment and an impediment for current flow, was discussed by Logan and Cebulski (1970) and Davies (1970). Hagan and Logan (1974) noted that the seagrass communities have been a major factor influencing the growth of carbonate banks in the bay. During the last 5 ka bank growth has led to progressive restriction of tidal exchange and increased salinity in the enclosed basins.

In their review of the seagrass communities of Shark Bay, Walker et al. (1988) and Walker (1989) indicate that the seagrass meadows are among the largest and most diverse in the world. Twelve species are present, with *Amphibolis antarctica* the most abundant type, occurring in monospecific meadows and occupying about 85% of the area covered by seagrass in the bay. The densest *Amphibolis* stands are within the 40–50‰ salinity range. *Posidonia australis* also is present in monospecific meadows, but is largely confined to channels particularly along Wooramel Bank (Fig. 13). In some areas, there are small patches of *P. australis* adjacent to *A. antarctica*, on the leading edge of stands, projecting into the prevailing current. Shark Bay is at the northern limits of the distributions of both *A. antarctica* and *P. australis*.

One hundred and sixty-one species of benthic macro-algae have been reported from Shark Bay with distributions controlled mainly by salinity tolerance and the presence of suitable substrates (Kendrick et al., 1990). The types and distributions of microbial mats are discussed below.

Seagrass, macroalgae, and microbial mats form a large nutrient store in the nutrient poor waters of the bay. Smith and Atkinson (1984) classified Shark Bay among the confined, phosphorus-limited ecosystems, and concluded that the net ecosystem production is limited by the

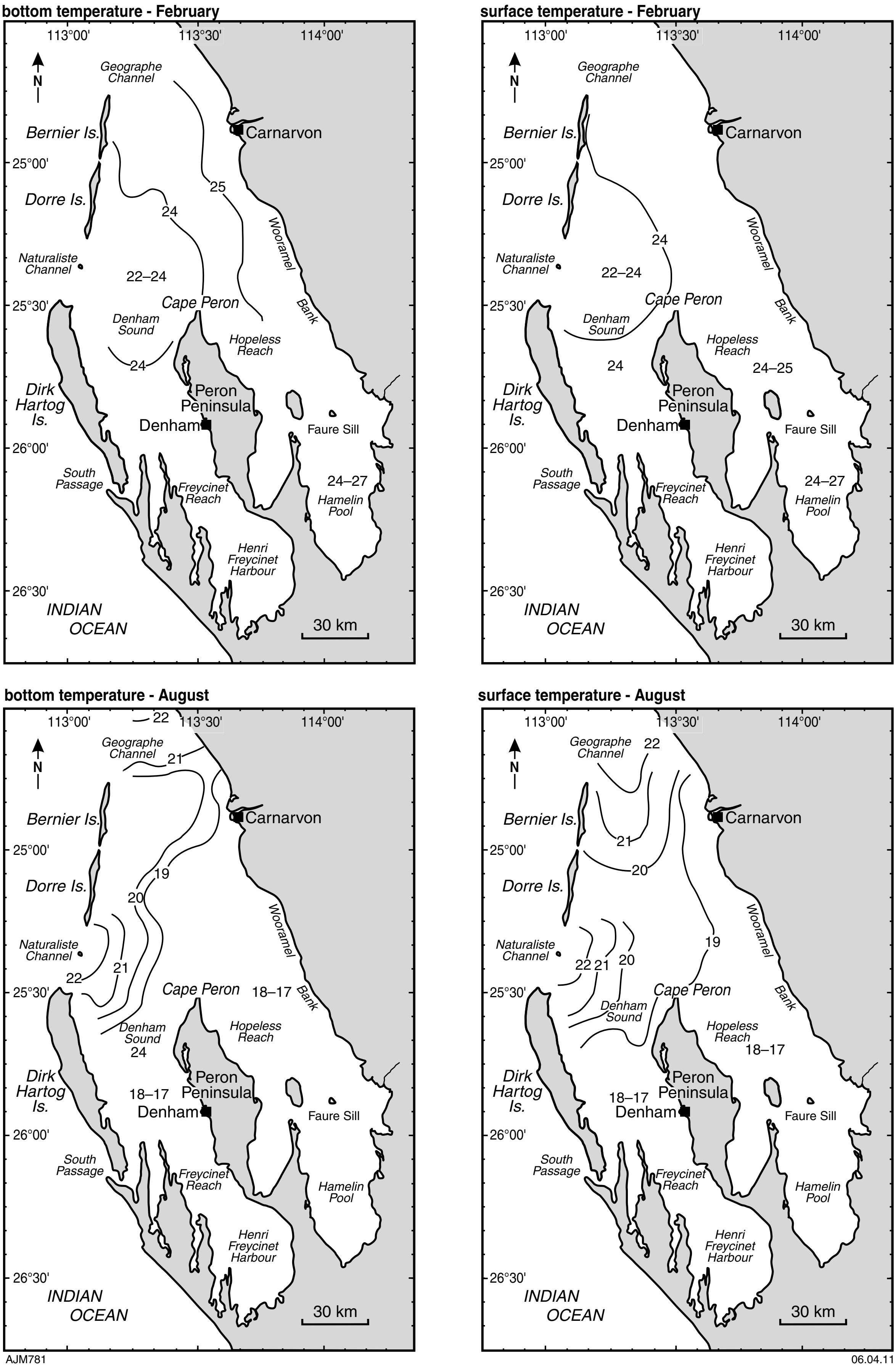

Figure 14. Bottom and surface temperature ranges (°C) in Shark Bay for a) and b) summer (February) and c) and d) winter (August), respectively. Based on 1965 data (after Logan and Cebulski, 1970)

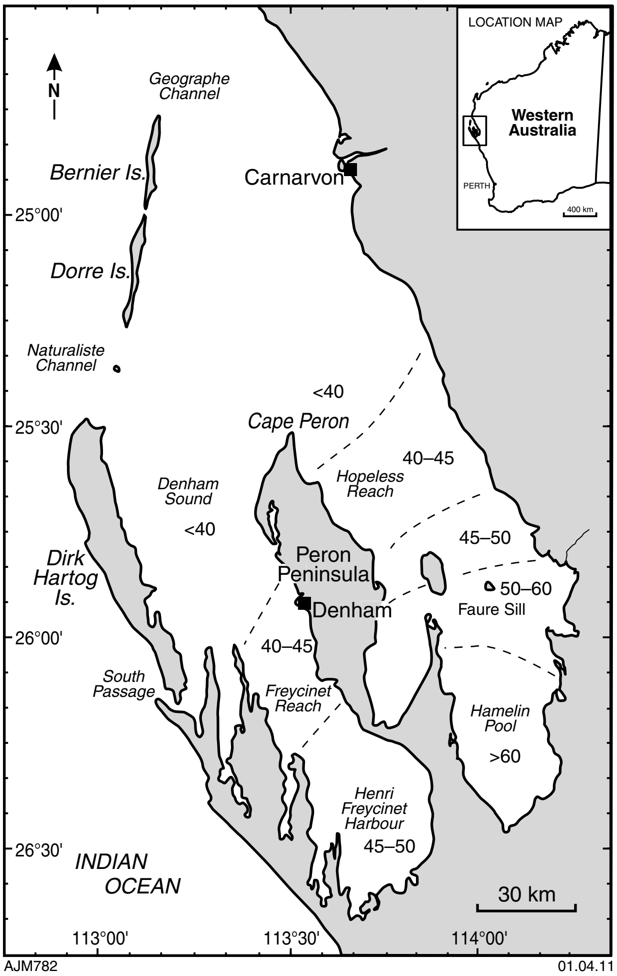

Figure 15. Summary of salinity in Shark Bay (from Kendrick et al., 1990; using, in part, data from Logan and Cebulski, 1970). Values in ‰

Figure 16. (below) Diagrammatic north–south cross sections showing salinity divisions in eastern and western gulfs of Shark Bay (from Logan and Cebulski, 1970). Values in ‰

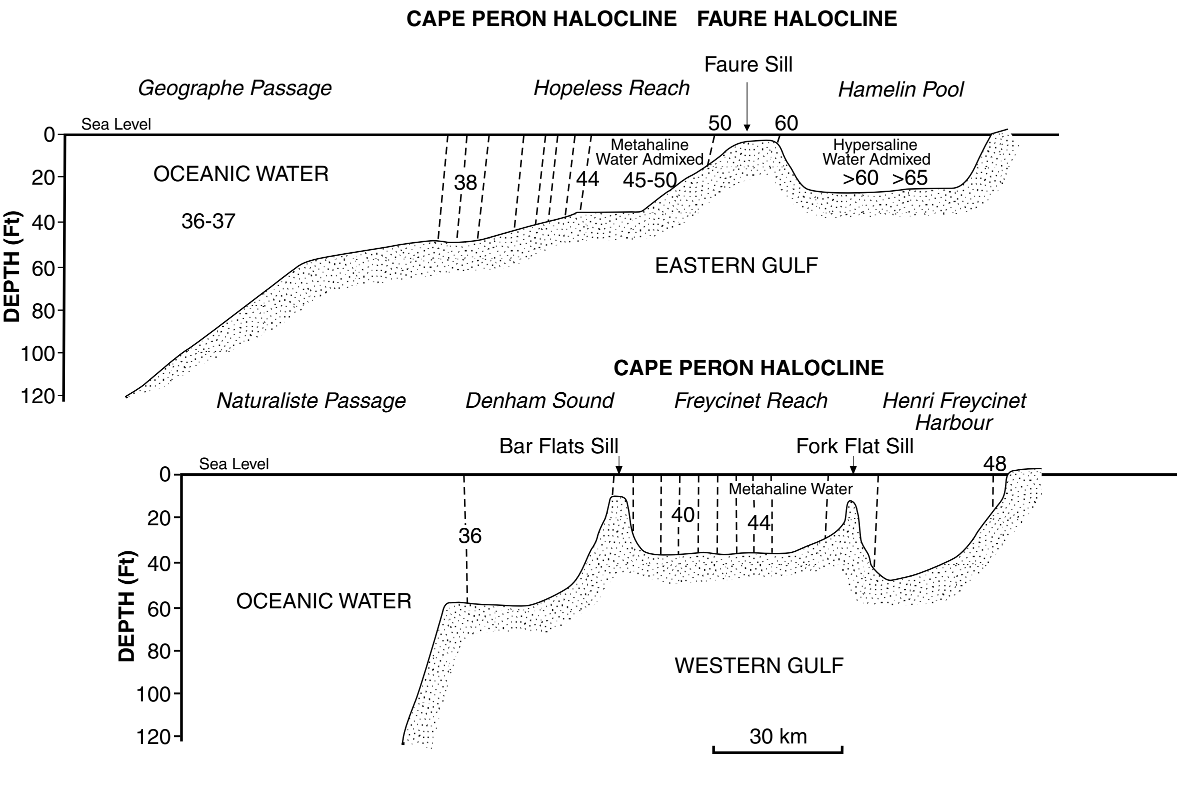

Table 1. Summary of the major parameters of the sublittoral platform environments in Shark Bay (after Logan and Cebulski, 1970)

	Seagrass meadows (oceanic and metahaline)	*Sand flats (oceanic and metahaline)*	*Sand flats (hypersaline)*
Area	Widespread; Uranie Strait, Uranie Bank, Cape Peron Bank, Bar Flats Sill, Freycinet Reach and Basin, Faure Sill	Widespread margins of Peron Peninsula, Faure Sill, Edel Peninsula, Freycinet Reach and Basin, inlets of Edel Region, east side of Bernier Island	Margins of Hamelin Pool and Lharidon Bight
Water depth	LWS level to about 12 m; densest grass development in metahaline areas of southerly aspect	Upper part exposed at spring tides and during southerly gales; average depth about 1.5 m, extends to depths of about 4.5 m	
Salinity	35–56‰, diurnal fluctuation variable depending on local gradient and movement of saline front; 1–3‰ fluctuation in open reaches; up to 10‰ in sill areas	35–56‰ extends through regional gradient; diurnal fluctuation variable depending on local gradient and movement of saline front; 1–3‰ fluctuation in open reaches; 5–10‰ in salinocline zones	56–70‰; diurnal fluctuation 1–10‰
Temperature (°C)			
Summer	22–30	22–30	22–30
Winter	15–18	15–18	15–18
Diurnal Range	1–2	1–2	1–2
Oxygen content	Large diurnal fluctuations, 2.0 – 6.0 ml/ll	Saturated, 4.0 – 5.5 ml/l	Saturated, 4.0 – 6.0 ml/
Turbidity	Often high as shown by fine particulate matter in suspension	Low	Low
Substrate nitrogen (‰)	0.4 – 1.5	0.08 – 0.2	0.8
Sediments	Variable depending on density of grasses; silty skeletal-fragment sands (dense growth), clean skeletalfragment sands (sparse growth); sediments commonly burrowed and skeletal grains fragmented	Medium-grained quartz sands, and lithoclastic sands; ripple marks and cross-bedding common	Coquinas, ooid sands, quartz sands, skeletal-fragment sands; ripple marks and cross-bedding common; sediments commonly lithified early
Wave action	Dependent on weather aspect and depth, as in sand-flat environments	Dependent on weather aspect; areas of southerly aspect subject to frequent attack; oscillatory currents 20–70 cm/sec	
Tidal currents	Dependent on location, up to 60 cm/sec	Dependent on location, usually moderate to strong up to about 45 cm/sec	

Table 2. Summary of the major parameters of the embayment-plain (basinal) environments in Shark Bay (after Logan and Cebulski, 1970)

	Oceanic Strait	*Metahaline Reach*	*Metahaline Basin*	*Hypersaline Basin*
Area	Uranie Strait, Denham Sound	Hopeless Reach, Freycinet Reach	Freycinet Basin – Lharidon Bight – Hamelin Pool	
Water depth	9–37 m	12–18 m	9–15 m	7.5 – 9 m
Salinity (‰)	35–38; diurnal 1–2	38–48; diurnal 2–4	45–48; diurnal 1–3	50–65; diurnal 5
Temp. (°C)				
Summer	24–25	24–25	24–25	25–27
Winter	19–22	17–18	17–18	17–18
Oxygen content	Saturated 4–5 ml/l; diurnal fluctuation 2–6 ml/l near seagrass stands			
Turbidity	Moderate	High	Moderate, fine particulate matter in suspension	Low
	Fine particulate matter in suspension			
Substrate nitrogen (‰)	0.15 – 0.77	0.3 – 0.8	0.08	1.4
Sediments	Variable, coarse to medium-grained skeletal fragment sands, silty in patches		Silty, fine-grained skeletal-fragment sands	Silty, coarse- to medium-grained microcoquinas
Wave action	Slight, areas near Naturaliste and Geographe Passages may come under influence of oceanic swell	Slight under prevailing conditions		
Tidal currents	9–30 cm/sec	15–45 cm/sec	0–3 cm/sec	
Biological factors	Flora sparse; seagrasses in scattered		Flora sparse; seagrasses Amphibolis in scattered stands in shallow marginal areas; shell-secreting benthos generally sparse, mainly foraminifera, bivalves, and irregular echinoids. Sediments are extensively burrowed and skeletal grains are fragmented	Flora sparse, mainly green algae. Fauna restricted to a few species not prolific in basin areas; sediments not burrowed; skeletal grains intact stands in shallow marginal areas; shell-secreting benthos generally sparse, mainly foraminifera, bivalves, and irregular echinoids. Sediments are extensively burrowed and skeletal grains are fragmented

oceanographic delivery of P, and that N requirements are met by nitrogen fixation within the system. There is little terrigenous nutrient input into the bay. Smith and Atkinson (1984) and Atkinson (1987) demonstrated that reactive (inorganic) P in Shark Bay waters decreases from about 0.2 µM P in oceanic regions to below 0.02 µM P in the hypersaline water, indicating net uptake and removal of dissolved P. They also showed that dissolved nitrate and ammonia are also low in the bay but increase with increasing salinity. This suggested that internal sources of N maintain the N concentration in the water column even through there is a net sink of N to the sediments. Walker and McComb (1988) found that the dominance of the seagrass *Amphibolis antarctica* in Shark Bay may be related to more efficient utilization of phosphorus compared to that in the other seagrass species present in the bay. Walker (1989) noted that while oceanic input of phosphorus is essential for seagrass production, recycling of nutrients was also important. Recycling may involve: 1) internal recycling within the plants by retranslocation of nutrients; and 2) decomposition of leaves and remineralization (Walker and McComb, 1985).

Microbial mat types and stromatolites: According to Logan et al. (1974a), microbial mats are present in the intertidal zone throughout Shark Bay. However, only where salinities are above 53‰, such as in Hamelin Pool, do the mats persist and trap and bind sediment. At lower salinities the mats are browsed and reduced to thin films. Features of the main microbial mat types recognized by Logan et al. (1974a; summarized in Table 3), and some of the main mat types visible in the intertidal zone are illustrated in Figure 18. Diatoms are abundant in the mats and have been recorded by John (1990) who listed 79 species and illustrated many of the types.

Awramik and Riding (1988) found that the intertidal stromatolitic columns were built mainly by fine-grained sediment (0.125 – 0.250 mm) trapping by two communities

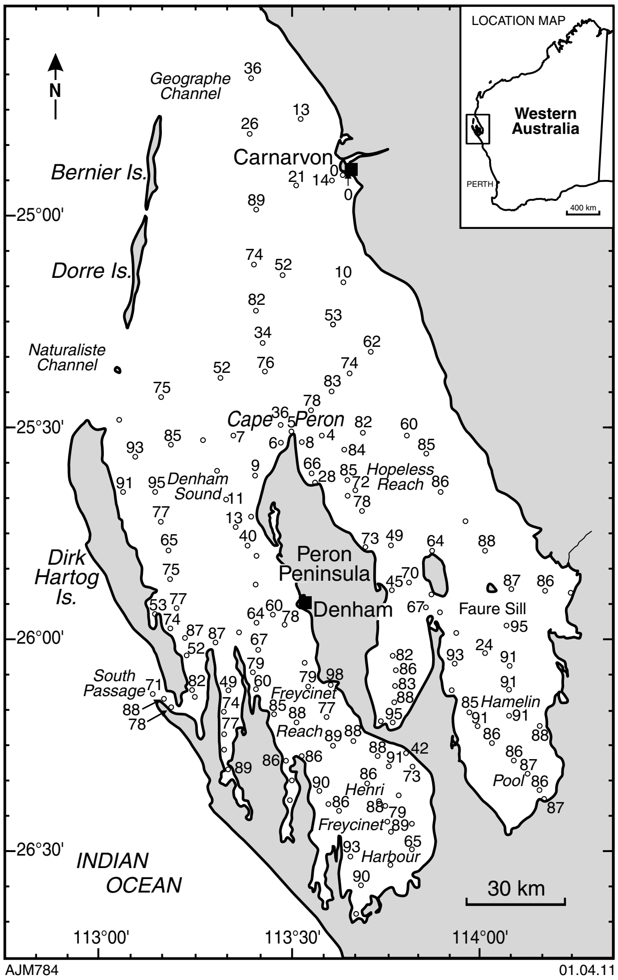

Figure 17. Logan's 1956–57 sediment sample sites in Shark Bay. Numbers refer to ‰ $CaCO_3$ in samples (after Logan, 1959, unpublished)

of cyanobacteria, one dominated by *Entophysalis* and the other by *Schizothrix*. Numerous other cyanobacteria are present. The incorporated sediment includes peloids and ooids, with some coarser bioclasts, cemented by cryptocrystalline aragonite. A complex community of eukaryotes (including diatoms some of which produce copious amounts of extracellular gel) and cyanobacteria is associated with the subtidal columns. In the subtidal stromatolites, Awramik and Riding (1988) found that the sediment mainly consists of coarse to very coarse (0.5–2.0 mm) bioclasts and ooids with some granule-sized mollusc fragments. The diatoms may play a significant role in trapping and binding the coarser sediment of the subtidal stromatolites.

According to Logan et al. (1974a), most of the microbial mats trap sediment and accretion rates vary with mat type: colloform mat, probably less than 1 mm/year; smooth mat, up to 10 mm/year; pustular mat, up to 2 mm/year; tufted mat, variable but up to 10 mm/year. The resulting sediment fabric also varies with mat type: for example, aligned large fenestrae formed by colloform mat; well-laminated sediment with fine fenestrae between layers formed by smooth mat; somewhat irregular fenestral fabric formed by pustular mat.

In contrast to the stromatolites at Lake Thetis (**locality 13**), which biochemically produce a calcareous 'skeleton', the Shark Bay stromatolites are composed of sediment agglutinated in the mats and later cemented by inorganically precipitated cryptocrystalline aragonite. Logan et al. (1974a) suggested that the microbial mats are shaped into stromatolitic structures by environmental factors such as waves, currents, substrate gradient, tidal behaviour, and long-term sea-level change (regression in Shark Bay).

The recent discovery of a new chlorophyll (chlorophyll f) within stromatolites at Shark Bay that can absorb light at a wavelength of approximately 720 nm — near the lower limit of the visible range (400–700 nm) — indicates that cyanobacteria within these stromatolites can photosynthesize and implies that the physical limits of photosynthesis are not well understood (Chen et al., 2010).

Marine and geological getting: The following features should be noted at Carbla Point.

1) The Hamelin Coquina (composed mainly of *Fragum erugatum*) forming beach ridges in the supratidal zone (Fig. 19a).
2) Beach rock formed by early cementation (with aragonite) of the coquina, in the lower supratidal zone with the upper surface sloping seaward (Fig. 19a).
3) Altered Upper Cretaceous Toolonga Calcilutite (with characteristic yellow-greenish colour) exposed on a relict wave-cut platform beneath a veneer of sand, microbial mats and stromatolites.
4) Ooid sands forming sheets between stromatolites and microbial mats. The sheets are often rippled at the surface (Fig. 19b).
5) Various mat types (see Fig. 19), and stromatolites in various stages of growth (Figs 19c,d, 20) and in places coalescing to form reef-like structures.
6) *Polyphysa peniculus* growing abundantly in the subtidal zone on the sides of stromatolite columns and on the hard substrate around stromatolites (Fig. 21). Rod-like fragments of this dasycladalean algae are a major contributor to sediment in Shark Bay.

Awramik and Vanyo (1986) recorded heliotropism in both subtidal stromatolites and intertidal microbial tufts at Carbla Point. The heliotropism is not ubiquitous, but northerly inclined stromatolites were found in about 2 m of water, 15 m off the northwest side of Carbla Point. According to Awramik and Vanyo (1986), inclination here is normal to water current and sediment supply. Inclined tufts near Carbla Point region are patchily distributed, with zenith angles averaging 33° to the north.

Age of the stromatolites*:* The stromatolites are Holocene based on radiocarbon dating by Chivas et al. (1990) of five stromatolites (34 samples). Chivas et al. (1990) also dated 11 samples from a sediment core, five samples from modern sediment and the tops of stromatolites, and two samples of the *Fragum* coquina. Errors in the Holocene ages are generally ± 60–80 years. The oldest stromatolite (1010–1250 years BP) from a terrace 0.3 m above present high tide on the west side of Hamelin Pool (at Booldah) is dead, whereas specimens with a live microbial coating are younger than 700 years BP (as are the ages from the

Table 3. Microbial mat types in Hamelin Pool (compiled from descriptions by Logan et al. 1974a)

Mat type	*Location*	*Gross morphology and colour*	*Microbial community*
(a) Mats that form stromatolites as well as flat sheets			
Film mat	Supertidal and upper intertidal zone	Thin coating on indurated substrates; black to dark greenish grey (on wetting, the mucilage absorbs water and swells to a thin pustular mat)	Layers of the coccoid microbes, mainly *Entophysalis major*
Pustular mat	Middle and upper intertidal zone	Rough mammillate surface composed of jelly-like pustules about 3 cm in diameter and 2 cm thick; dark reddish brown	Coccoid microbes, with *Entophysalis major* most abundant
Smooth mat	Lower intertidal zone	Smooth surface; contains much fine sand that adheres to the mat and/or are bound into a filament mesh; pale yellowish gray to brown	Fine filaments of *Schizothrix helva* and less abundant *Symploca laeteviridis*
Colloform mat	Subtidal zone	Colloform surface of hollow contiguous convexities 1–3 cm in diameter; pale yellowish-brown to grey	Biologically heterogeneous. At Carbla Point the mat is thin and leathery and is dominated by *Schizothrix helva*. Also present are filaments of *Oscillatoria latet-virens*, *O. foreavi*, and *Johannesbaptista pellucida*
(b) Mats that mainly form flat sheets			
Blister mat	Supratidal and upper intertidal	Smooth with a rubbery consistency. Where elevated well above the groundwater table, the mat is <1 mm thick and flat; where groundwater is nearer surface, the mat is thicker and characterized by hollow blisters; may detach from substrate; pale beige on top and green beneath	Intertwined *Microcoleus chtonoplastes* filaments with thick mucilaginous sheaths; with less abundant filaments of *Phormidium molle* var. which generally have no sheaths
Tufted mat	Upper and middle intertidal zone	Corrugated surfaces of sharpcrested ridges formed by vertically oriented tufts, and low, broad depressions with leathery mat; greenish-brown to black	Tufts composed of large filaments of *Lyngbya aestuarii* ; the leathery mat is composed of vertically oriented and intertwined *Microcoleus chthonoplastes* filaments
Gelatinous mat	Mat Middle and lower intertidal zone	Smooth slimy mats up to 1 cm thick; accumulate to form a laminated peat almost devoid of sediment; yellowish-brown to green	Gelantinous layers with vertical filamentous microstructure composed predominately of *Symploca laete-viridis* with *Johannesbaptista pellucida* and coccoid species, *Entophysalis*, *Aphanocapsa*, and *Aphanothece*; discontinuous pink mucilaginous layers of coccoid bacteria

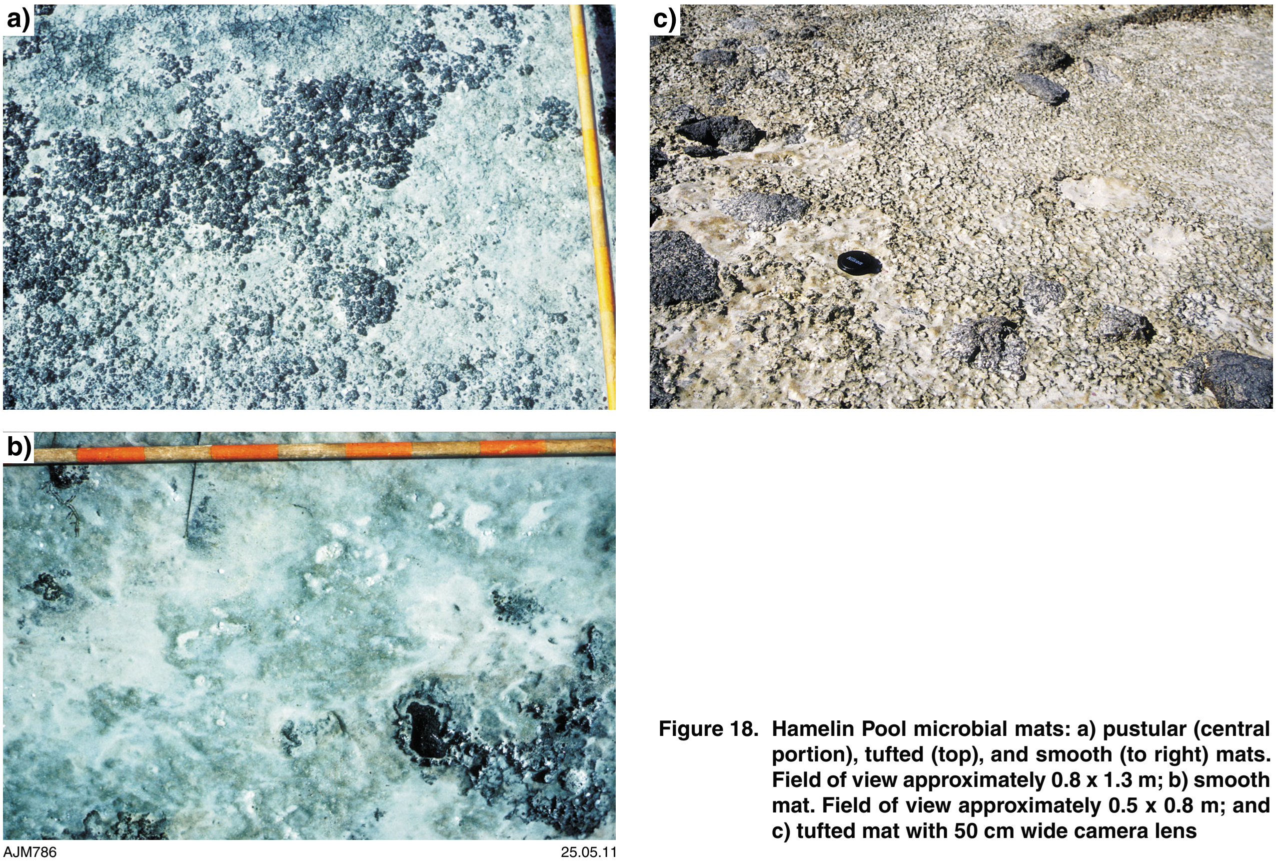

Figure 18. Hamelin Pool microbial mats: a) pustular (central portion), tufted (top), and smooth (to right) mats. Field of view approximately 0.8 x 1.3 m; b) smooth mat. Field of view approximately 0.5 x 0.8 m; and c) tufted mat with 50 cm wide camera lens

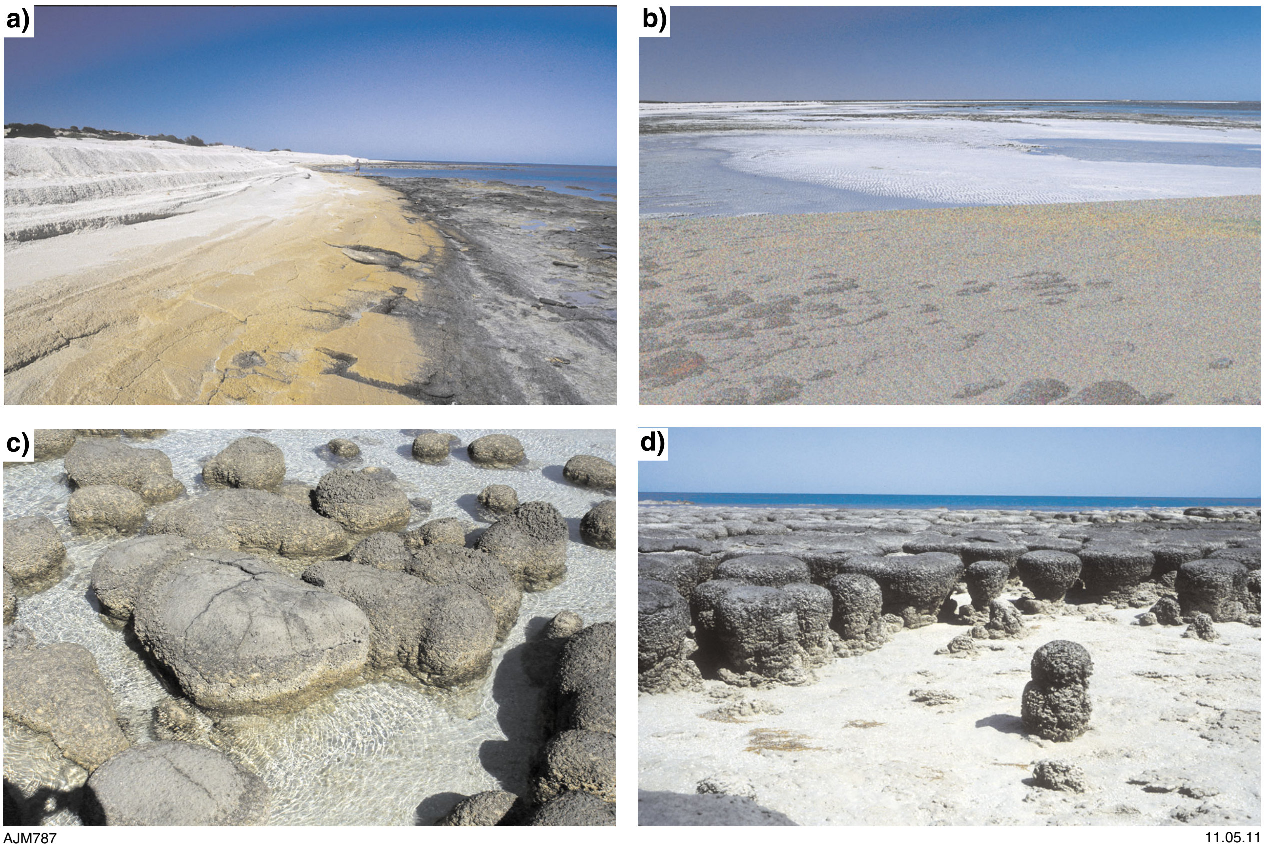

Figure 19. Views at Hamelin Pool at Carbla Point: a) Hamelin Coquina prograding over beach rock (also composed of coquina) with inactive stromatolites in the uppermost intertidal zone; b) intertidal zone just south of Carbla Point at low tide. The rippled sand sheets are composed mainly of ooids; c) live stromatolites in the intertidal zone; d) stromatolites in the intertidal zone at Carbla Point coalescing to form a 'reef' structure

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Figure 20. Hamelin Pool stromatolites with flat top covered with film mat, surrounded by pustular mat, and colloidal mat lower on the 0.5 m high column

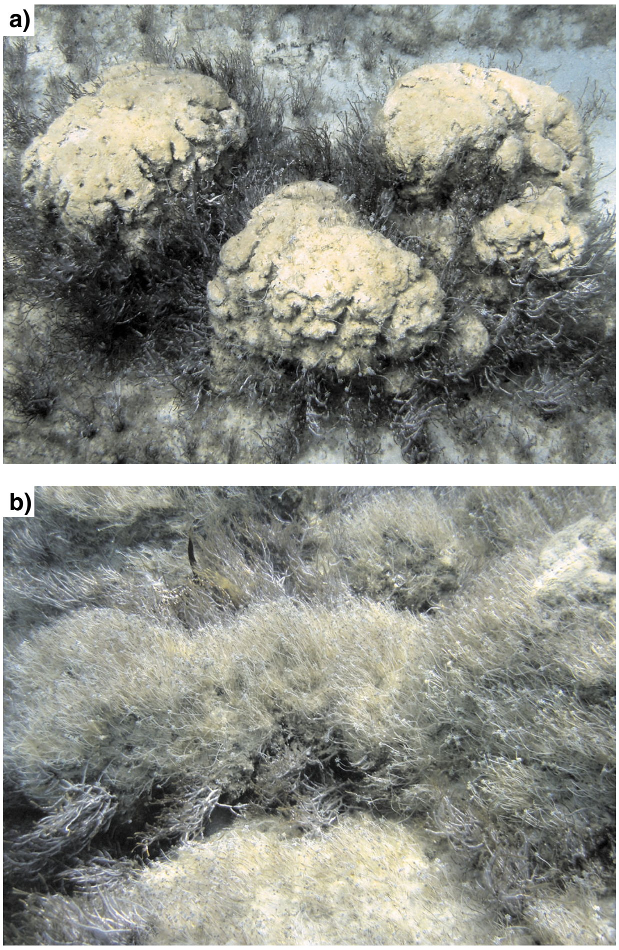

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Figure 21. Subtidal stromatolites off Carbla Point a) surrounded by and b) covered by the dasycladalean algae *Polyphysa peniculus*

sediment core). Eliminating out-of-sequence ages (clearly due to reworked sediment) leaves 22 stromatolite and six sediment ages that appear 'reliable'. These ages indicates breaks in deposition of up to 400 years in individual stromatolites, with rates of deposition up to 1.5 mm/year. When the ages are grouped these breaks appear not to be systematic, although there is a noticeable gap of about 100 years around 200 years BP, but that may be due to the relatively small number of analyses. Chivas et al. (1990) indicate net growth rates are ≤0.4 mm/year, and estimated that it takes about 1000 years for the stromatolites to reach heights of 35 cm. Interestingly, the ages from the *Fragum* coquina (730 and 1250 years BP) are close to the ages from the stromatolites, implying that reworking from these deposits contributed to the out-of-sequence ages. The older of the two ages from the Tamala Limestone substrate given by Chivas et al. (1990; 12 500 ± 245 and 148 850 ± 200 years BP) is somewhat older than the marine isotope stage 5e ages from coral specimens based on U-series dating by O'Leary et al. (2008; 120 000 – 131 000 ± 0.5 – 1.1 years BP).

Locality 3: Coolkilya Pool

Significance: The type sections of the Quinnanie Shale and Wandagee Formation (of the Quinnanie–Wandagee–Nalbia shoaling upward depositional 'sequence', Figs 22, 23) and the upper part of the Cundlego Formation (terminal part of the Bulgadoo–Cundlego shoaling upward depositional 'sequence', Fig. 24) show cycles progressively thinning up to the middle of the Wandagee Formation and again to the top of the unit, indicating decreases in accommodation into a highstand followed by major flooding events.

Location: The field sites are on banks of the Minilya River in the vicinity of Coolkilya Pool, Wandagee Station from 23°43'49"S 114°25'41"E to 23°44'25"S 114°24'50"E, WINNING POOL. The positions of Coolkilya Pool and other bodies of water that usually remain throughout the dry season change over the years because there is an enormous transport of sand (and even rock blocks) during periodic floods. Access to the Coolkilya Pool area is via a track

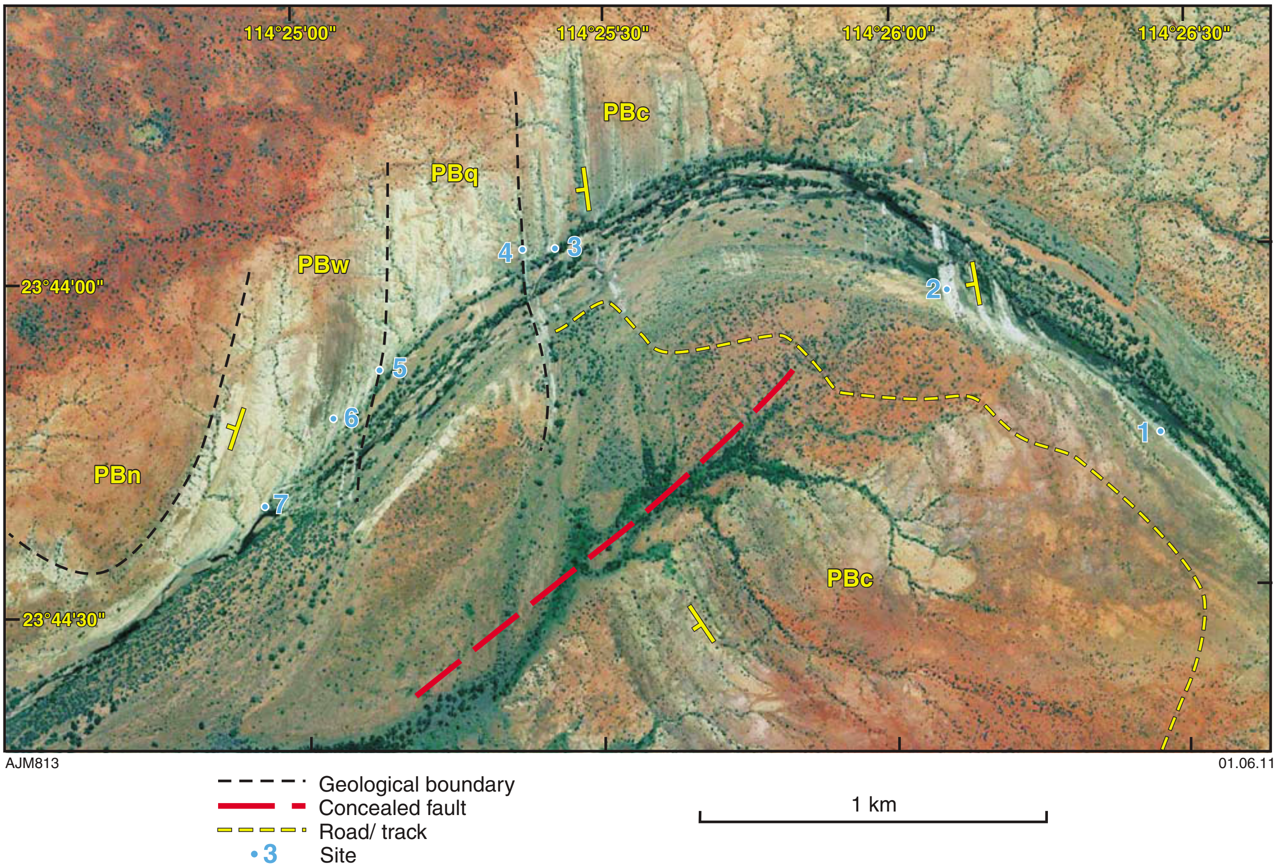


Figure 22. Simplified geological map superimposed on aerial image of Coolkilya Pool. PBn – Nalbia Sandstone, PBw – Wandagee Formation, PBq – Quinnanie Shale, PBc – Cundlego Formation

which leads off the main Minilya to Wandagee road at 23°45'16"S 114°26'33"E, Winning Pool.

Geology: The Cundlego Formation outcrops along the Minilya River upstream of 23°45'S 114°27'30"E (Fig. 24). In places the succession is disrupted by faulting and by incision of large channels containing Pleistocene or late Neogene fluvial deposits. Shale to sandstone parasequences of the Cundlego Formation are best exposed in river cliffs between sites 1 and 2 in the middle section of the formation, and between sites 3 and 4 in the upper part of the unit. Facies are interpreted according to a depositional model proposed by Moore et al. (1980b) and Hocking et al. (1987) based on the position of wave base (Fig. 25). In a confined interior sea, wave base was probably very shallow, so the boundary between 'offshore' and 'shoreface' facies may have been in just a few metres of water. The low gradient sea floor, suggested by the almost horizontal attitude of seismic horizons through the basin (except at faults) and the wide geographic extent of facies, implies that the 'shoreface' facies recognized by Moore et al. (1980a) and Hocking et al. (1987) may belong to sand banks within a shallow sea and does not necessarily indicate proximity to the shoreline. The parasequences of the Cundlego Formation may be grouped into sets with the upper parts of these defined by amalgamation of thick sandstone units (e.g. at sites 2 and 4 where sandstone 'bars' cross the river).

Only very few beds of the Cundlego Formation contain macrofossils. Some shale beds contain abundant siliceous organic-cemented agglutinated foraminifera. The fossil assemblage from the Minilya River, and other localities south of the river, includes foraminifera, corals, bryozoans, brachiopods, bivalves, nautiloids, gastropods, conulariids, crinoids, and trace fossils (Skwarko, 1993; Archbold, 1993; Webster and Jell, 1999). Palynomorphs of the *Praecolpatites sinuosus* zone have been reported from mudstone outcrop at 23°44.217'S, 114°26.409'E (Backhouse, 2007; Purcell, 2008), and in subsurface throughout the basin (Mory and Backhouse, 1997; Backhouse, 1998).

The boundary between the Quinnanie Shale and Cundlego Formation is at the last prominent sandstone bed at site 4. The first parasequence in the retrogradational trend that includes the lower Quinnanie Shale is placed at the top of the Cundlego Formation. The type section of the Quinnanie Shale (Fig. 26) becomes progressively sandier up section. Haig (2003) identified five shoaling upward cycles within the type section and many smaller scale cycles based on changes in foraminiferal biofacies (Fig. 26). The maximum

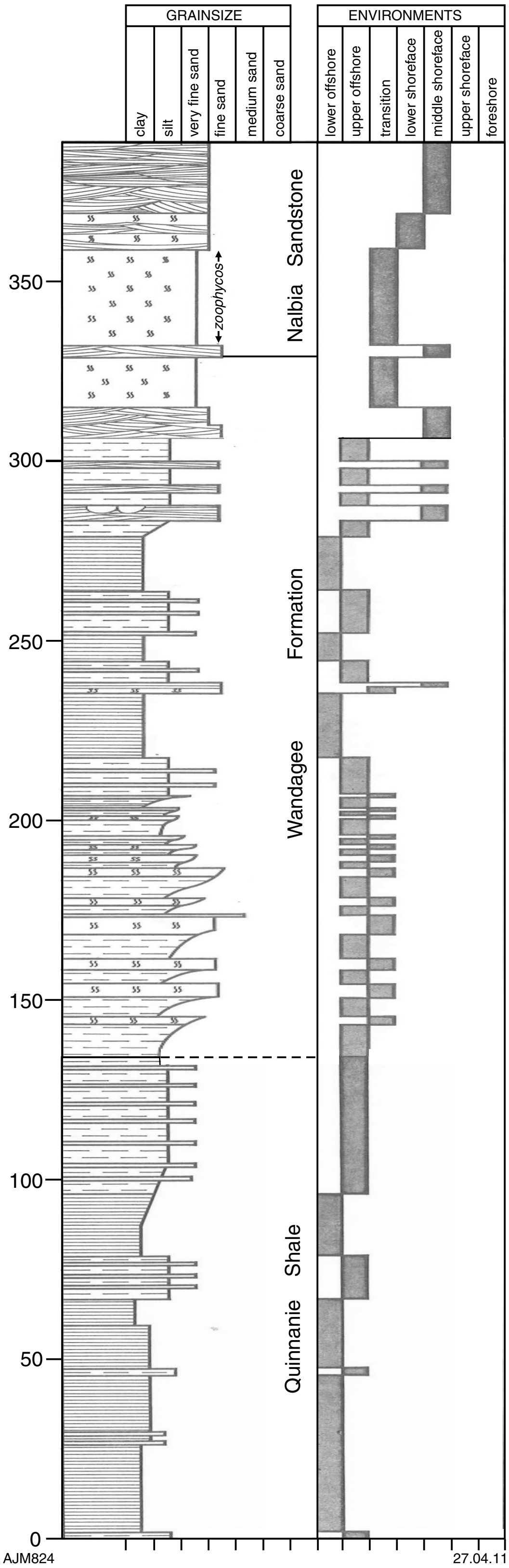


Figure 23. Measured section of Quinnanie Shale to Nalbia Sandstone, Coolkilya Pool (from Moore et al., 1980a). See Figure 25 for explanation of facies

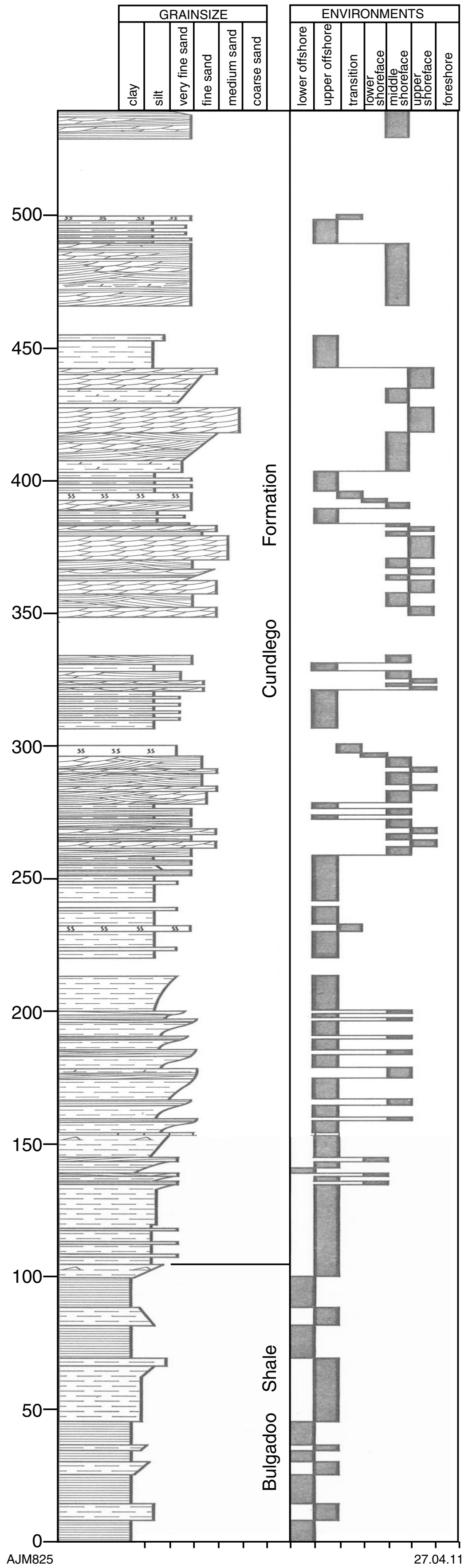


Figure 24. Measured section of Bulgadoo Shale and Cundlego Formation, Minilya River (from Moore et al., 1980a). See Figure 25 for explanation of facies

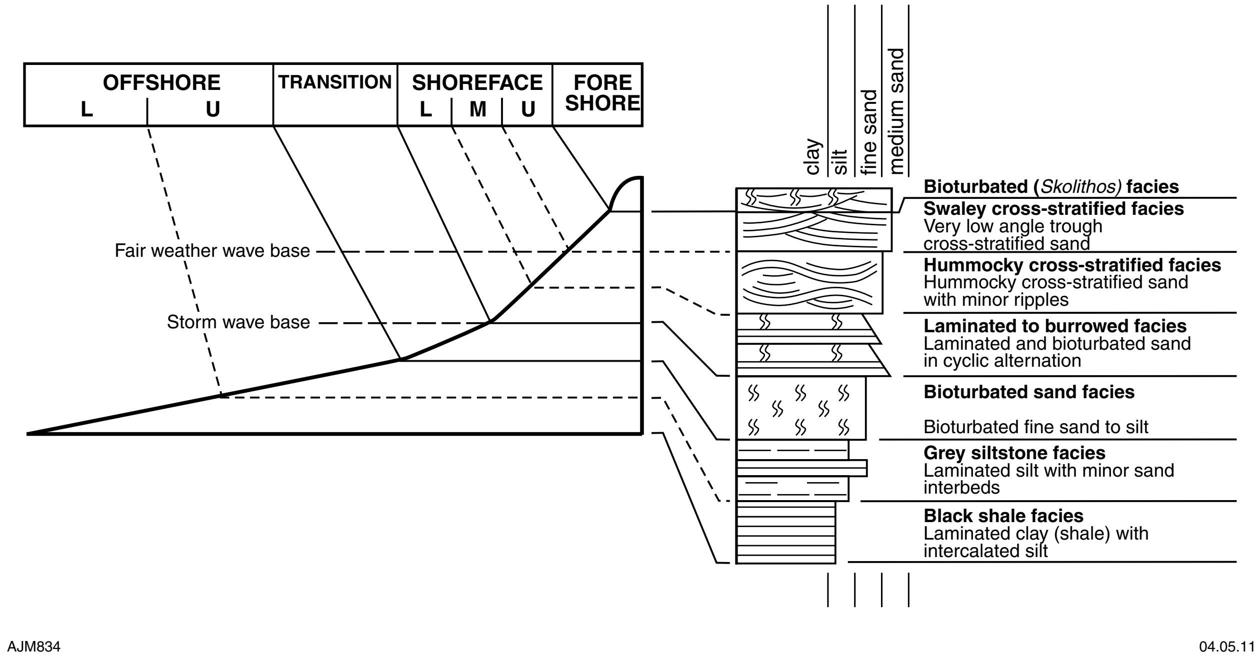


Figure 25. Depositional model for the Byro Group and lower Coolkilya Sandstone (after Hocking et al. 1987)

flooding level in the Quinnanie Shale succession was interpreted as being about 20 m above the base of the formation (Haig, 2003) at a concentration of calcareous mudstone nodules. The foraminiferal assemblages in the Quinnanie Shale are similar to those recovered from shale elsewhere in the Byro Group. A dominance of siliceous organic-cemented agglutinated foraminifera, with affinities to modern estuarine foraminifera (Haig, 2003) and to modern assemblages from the Baltic Sea, suggests that brackish conditions may have existed during deposition, in the shallow interior sea, of some of the mud units, and that bottom water conditions below a shallow wave base had low levels of dissolved oxygen. From the presence of lingulid brachiopods in siltstone to fine-grained sandstone of the upper Quinnanie Shale, and by analogy to the modern distribution of *Lingula*, Archbold (1981) concluded that deposition took place under fluctuating salinities. Apart from the foraminifera and brachiopods, corals, bivalves, ammonoids, nautiloids, gastropods, and crinoids have been reported from the Quinnanie Shale (Skwarko, 1993; Keane, 1999; Haig, 2003).

The lower parasequences of the Wandagee Formation (between sites 5 and 6, Fig. 22) form prominent 'bars' into the river channel and sometimes are exposed in central parts of the braided channel. The sands at the tops of the four lower parasequences are intensely bioturbated (*Zoophycos* association) and contain a diverse fossil assemblage including foraminifera, corals, bryozoans, brachiopods, bivalves, ammonoids, nautiloids, gastropods, conulariids, trilobites, scaphopods, asteroids, blastoids, crinoids, fish, annelids (Skwarko, 1993; Keene, 1999; Webster and Jell, 1999). Palynomorph assemblages belonging to the *Praecolpatites sinuosus* zone have been found from outcrop mudstone site 7 (Backhouse, 2007).

Above the Wandagee Formation in this area, the Nalbia Sandstone (Teichert, 1950, 1957; Hocking et al., 1987) is very poorly exposed. It includes the unit mapped by Condon (1967) as the 'Norton Greywacke'.

The age of the Byro Group succession in the vicinity of Coolkilya Pool is constrained by: 1) ages placed on lower formations in the Byro Group (Late Artinskian or Kungurian; see discussions for **localities 4 and 8**); 2) ages derived from ammonoids present in the Quinnanie Shale, Wandagee Formation and Nalbia Sandstone; (3) ages suggested by conodonts present in the Wandagee Formation and in the stratigraphically higher Kennedy Group.

The ammonoid *Bamyaniceras australe* (Teichert) is present in the underlying Bulgadoo Shale (see **locality 4**), and in the Quinnanie Shale, Wandagee Formation and Nalbia Sandstone (Teichert, 1942; Teichert and Glenister, 1952; Glenister and Furnish, 1961; Cockbain 1980). According to Leonova (1998), the presence of *Bamyaniceras* indicates a correlation to the Yaktashian and Bolorian (late Artinskian to Kungurian; following the time scale of Ogg et al., 2008). The Nalbia Sandstone contains the additional ammonoids *Daubichites goochi* (Teichert) and *Paragastrioceras wandageense* Teichert. Bogoslovskaya (1976), Archbold and Dickins (1989) and Leonova (1998) noted close similarities between *P. wandageense* and *P. kungurense* from the Upper Kungurian of the Urals. Leonova (1998) suggested that *D. goochi*, which came from the same locality as *P. wandageense*, marked the Roadian.

Nicoll and Metcalfe (1998) noted rare elements of an unnamed conodont species of the genus *Vjalovognathus*. These have morphological characteristics that suggest

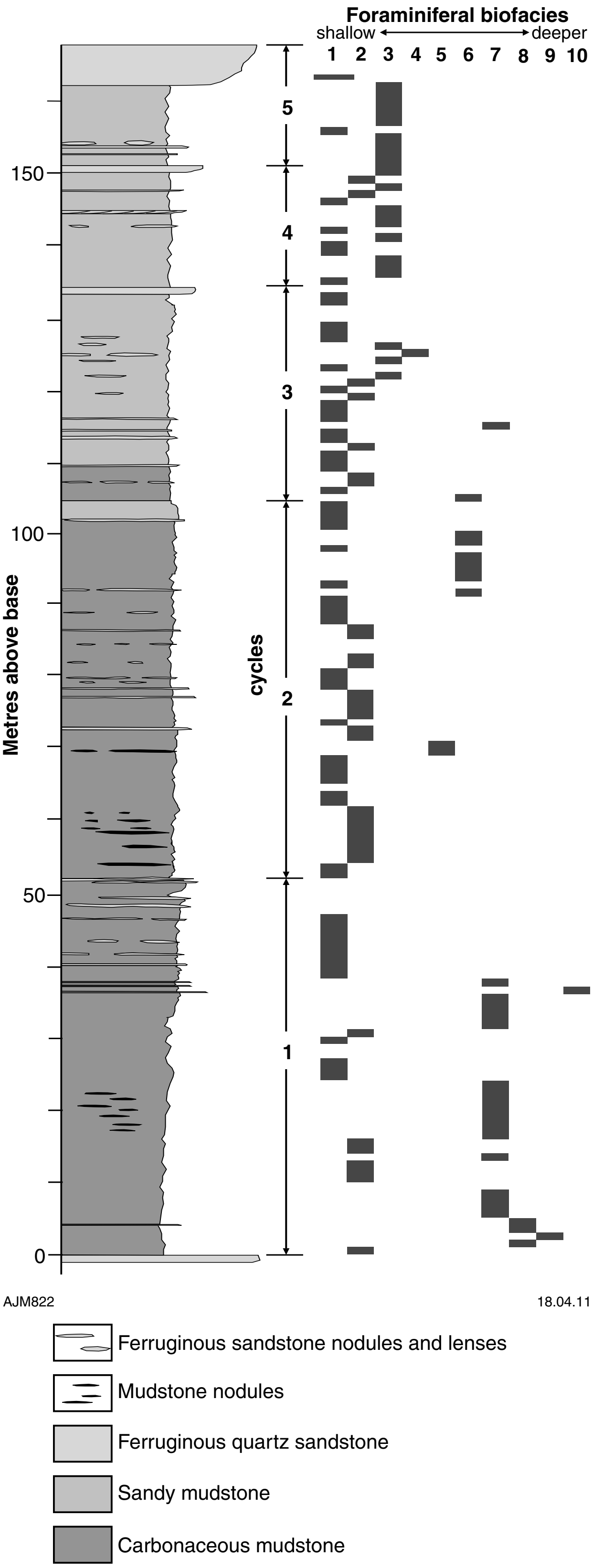


Figure 26. Measured section of the Quinnanie Shale with foraminiferal biofacies (from Haig, 2003). See Figure 23 for explanation of facies.

they are transitional between *V. shindyensis* (probably of Bolorian age; see discussion by Nicoll and Metcalfe, 1998, p. 427; now taken as Kungurian following Ogg et al., 2008) and unnamed *Vjalovognathus* known from the upper Changhsingian.

Based on these tentative ammonoid and conodont interpretations, at least part of the succession examined at locality 3 is probably Kungurian with the possibility of it ranging as high as Roadian.

Locality 4: Minilya River, 'Wandagee'

Significance: This locality includes part of the type area of the Bulgadoo Shale. Based on an interpretation of broad-scale palaeobathymetric trends, the lower Bulgadoo Shale is considered to represent the maximum flooding level for the major late Artinskian to Roadian depositional cycle in the basin. This cycle extends from the upper Wooramel Group through the Byro Group and terminates at the top of the Kennedy Group.

Location: The most accessible sections of Bulgadoo Shale in the type area, are cliffs along the west bank of the Minilya River: 1) about 300 m southeast of Wandagee Homestead (23°45'55"S, 114°38'16"E, WINNING POOL); and 2) extending downstream from about 600 m north-northwest of the homestead (Fig. 27). The best exposures are along the river about 1.5 km north-northwest of the homestead.

Geology: The Bulgadoo Shale at this locality is mainly black pyritic shale with very minor interbeds of fine-grained muddy sandstone. In subsurface total organic

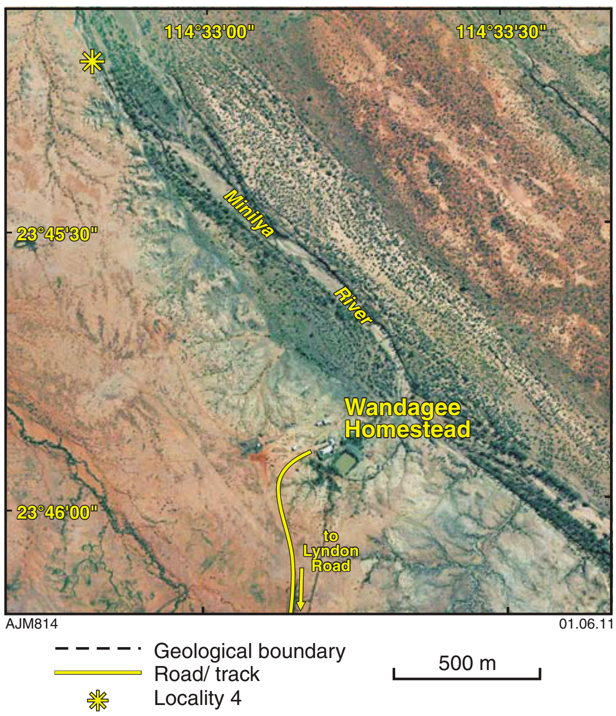


Figure 27. Locality map for Wandagee homestead area

carbon values of 2% or higher have been obtained from this formation (Ghori, 1996).

The fossil assemblage includes only sparse macrofossils but abundant microfossils at some levels. Foraminifera, corals, bryozoans, brachiopods, bivalves, ammonoids, nautiloids, gastropods, conulariids, ostracods, crinoids, fish, annelids, and trace fossils have been reported (Skwarko, 1993; Archbold, 1993; Webster and Jell, 1992). The foraminiferal fauna from beds at this locality, under study by DWH, includes a diverse assemblage of minute calcareous hyaline, procelaneous, and siliceous organic-cemented agglutinated species that suggest relatively deep water compared to that under which most shale beds in the Byro Group were deposited, normal-marine salinities, and low dissolved oxygen levels. Samples of dark grey mudstone collected from the river cliff at 23° belong to the *Praecolpatites sinuosus* zone (R Purcell, 2008, unpublished report), a spore-pollen zone that encompasses most of the Byro Group.

In terms of petroleum prospectivity the Bulgadoo Shale includes beds with some of the highest TOC levels recorded in the Permian of the basin. However, maturity levels suggest that this level lies above the oil window (Ghori et al., 2005).

Locality 5: Dead Man's Gully

Significance: This locality includes the oldest Permian succession free of glacial influence in the basin. Here sections of the Callytharra Formation contain typical cool-water carbonates as parts of sedimentary cycles (of two conspicuous orders) that form a progradational succession culminating in coarse-grained quartz sandstone of the overlying Wooramel Group (present, but better exposed at **locality 8**). The limestones, which are mainly bryozoan packstones and grainstones, are richly fossiliferous. The faunal succession is greatly influenced by facies, although this has not been documented in detail at this locality. Apart from at the type section (Dixon and Haig, 2004), no systematic stratigraphic studies have been published on outcrops of the formation. However, current work includes the bryozoan fauna (Andre Ernst, Universität Kiel, and Eckart Håkansson, UWA), foraminifera (DWH), and brachiopods (Zhong Qiang Chen, UWA). The progradational succession at locality 5 is contemporaneous with the Fossil Cliff – High Cliff – Irwin River Coal Measures succession of **locality 1**, about 560 km south in the interior rift system. These successions represent the final phase of the first major post-glacial marine depositional sequence in the East Gondwana rift system.

Location: The locality lies at the intersection of a double line of low ridges, trending at about 330°, with the Lyndon Road at 23°53'50"S, 114°56'49"E, KENNEDY RANGE (Fig. 28). The complete succession of the Callytharra Formation may be observed by walking out the two transects A–B and C–D marked on Figure 28. Section C–D is reached by the access road to an old seismic line. If time permits, a visit to the location marked X on Figure 28 may be made to better observe the contact between the Lyons Group and the overlying Callytharra Formation.

Geology: The Callytharra Formation and the overlying Wooramel Group outcrop here (Fig. 28) because of tilting, drag folding and uplift of strata along faults during breakup; some of these also were active in the late Neogene. Through most of the basin to the west of the strike ridges, the formations are flat lying in the subsurface (Fig. 29). Faulting, probably initiated by the separation of Australia from greater India, and reactivated by the collision of the Australian passive margin with the Banda Arc to the northwest of the basin, has caused repetition of strata perpendicular to strike and the offset of strata along strike at certain places. The series of strike ridges exposing the Callytharra Formation extend for over 200 km south of locality 5 and over 60 km north of this location.

Section A–B (Fig. 28) was measured on flat ground to the east of the eastern line of ridges. The upper part of this section stratigraphically overlaps the lower part of Section C–D measured from low ground in front of the western ridge up the ridge to the contact with the Wooramel Group. The Callytharra Formation is subdivided into three units (Fig. 30): units A and B form most of the low ground because of the dominance of friable mudstone/sandstone that is rarely exposed because of weathering, and unit C forms the ridges because of a greater proportion of indurated limestone.

The base of the formation is conformable on the Lyons Group. The weathered residue of a spectacular matrix-supported conglomerate is present at site X about 5 m below the lowest limestone bed that marks the base of the Callytharra Formation. Similar conglomerate beds, lower in the Lyons Group, will be observed at localities 6 and 7. At site X (Fig. 28), boulders to several metres diameter are present and include a large variety of metamorphic and igneous rock types, as wel as chert and stromatolitic dolostone/limestone derived from Precambrian basement to the east. Three distinct facies are recognized (Fig. 30): A: 5–10 cm thick packstone beds with a micrite matrix, but in some beds replaced by neomorphic calcite cement, alternating with much thicker friable units that generally are not exposed B: 5–10 cm thick packstone beds with a silty mud matrix, alternating with thicker friable units that are not generally exposed. Indurated calcareous fine-grained sandstone and siltstone beds are present at about 49 m and 57 m above base of formation C: 20–75 cm thick bryozoan grainstone beds alternating with thicker friable units (sandy blue-grey shale) except for the intervals 65–71 m and 101–109 m where grainstone beds are amalgamated. Many of the grainstone beds have laminae in which fine to medium quartz grains are concentrated. Low-angle cross bedding is present in some of the grainstone beds.

Bryozoans dominate the fossil assemblages in the limestone beds (Fig. 31a–d). In units A and B, they are mainly small colonies of fenestrate and arborescent species that apparently formed dense bryozoan thickets on the shallow sea floor. In unit C, large *Evactinoporella*, star-shaped in cross-section, and ramose ribbon-like *Hexagonella* are colonies adapted for life in mobile sediment that is now preserved as bryozoan grainstone.

The thin to medium bedded bryozoan packstone of Unit A is rich in attached porcelaneous foraminifera belonging to

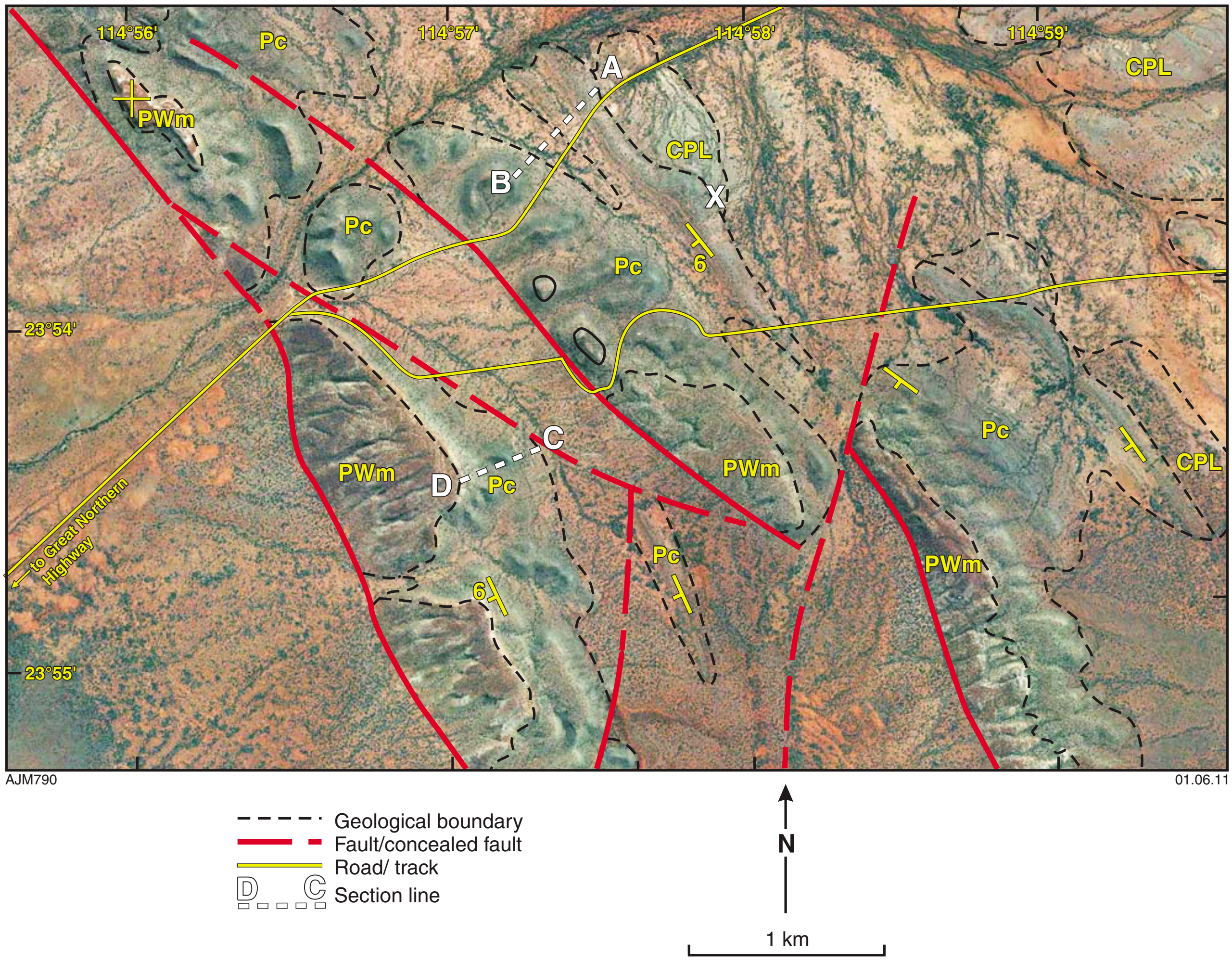


Figure 28. **Simplified geological map superimposed on aerial image of Dead Man's Gully. CPL Lyons Group, Pc Callytharra Formation, PWm Moogooloo Sandstone**

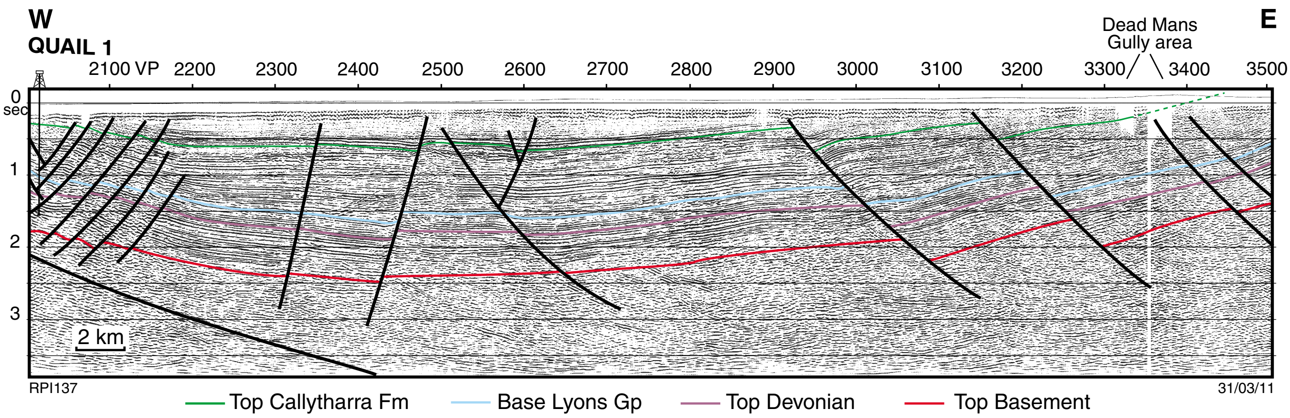


Figure 29. **Seismic section K82A-105 which crosses the Dead Man's Gully area (between VP 3300 and 3400; from Iasky et al., 1998)**

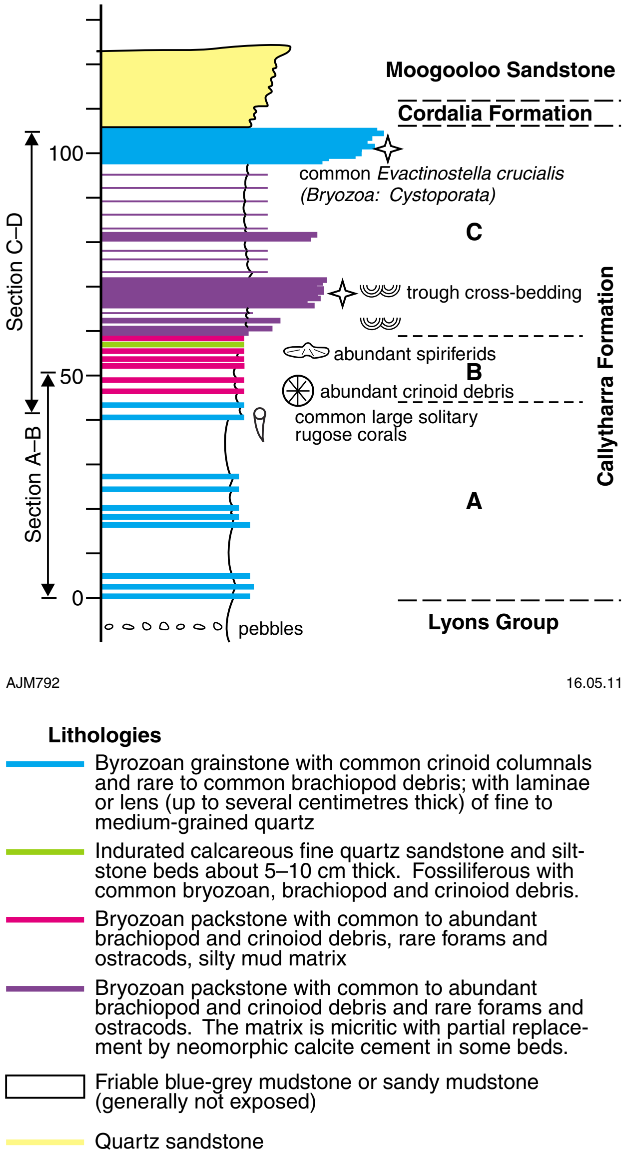


Figure 30. Composite measured section, Dead Man's Gully (see Fig. 28 for location)

Calcitornella spp. and *Trepeilopsis australiensis*. As in the type section of the Callytharra Formation (Dixon and Haig, 2004) the attachment surfaces of many *T. australiensis* are not preserved (Fig. 31a) and possibly included non-skeletal macroalgae. These foraminifera are rarer in unit B packstones but here are accompanied in some beds by *Hemigordius* sp. The most common foraminifera present in the bryozoan grainstones of unit C are nodosariids; porcelaneous foraminifera are very rare in this unit.

Although brachiopods are present in most of the limestone beds, they are particularly abundant in a few beds in units A and B. Zhong Qiang Chen (written comm.) provided the following brachiopod determinations. Bryozoan packstones in the lower 5 m of the formation contain an assemblage that includes *Neochonetes* (*Sommeriella*) *prattii* (Davidson), *Strophalosia irwinensis* Coleman, *Callytharrella callytharrensis* (Prendergast), *Cimmeriella foordi* (Etheridge), *Elivina hoskingae* Archbold and Thomas, *Cleiothyridina baracoodensis* Etheridge, *Permorthotetes* sp., *Stenoscima* sp., *Cyrtella* sp., and *Callispirina* sp. Beds higher in unit A lack *G. foordi* and *S. irwinensis* above about 10 m, but include the other species together with *Spirelytha fredericksi* Archbold and Thomas and *Neospirifer* sp. Near the top of unit A is a bed with abundant *Costatumulus irwinensis* (Archbold). The Unit B fauna includes shell concentrations of *Neospirifer* (*Neospirifer*) *foordi* Archbold and Thomas, *Neospirifer* (*Quadrospira*) *preplicatus* Hogeboom, *Latispirifer callytharrensis* Archbold and Thomas, *Crassispirifer* sp., *Trigonotreta neoaustralis* Archbold and Thomas, and *Neochonetes* (*Sommeriella*) *pratti* (Davidson). Brachiopods are not as prominent in the grainstones of unit C, but include some of the large spiriferids present in unit B.

Crinoids are abundant in many beds but are almost entirely disarticulated columnals or portions of stems. Cups are very rare. The holotypes of *Cosmetocrinus*(?) *middalyaensis* Webster and Jell (1992, fig. 14A), a relatively complete aphele crinoid, *Texacrinus goochensis* Webster and Jell (1992, fig. 16F), the cup and arms of a texacrinid, and *Glaukosocrinus middalyaensis* Webster and Jell (1999, fig. 10), preserved as cup and arms, were found on the surfaces of thick grainstone beds in unit C in this area. On the same rock slab containing the holotype of *C.*(?) *middalyaensis* debris from five other crinoids was recognized.

The precise age of the Callytharra Formation at locality 5 is uncertain. No conodont or ammonoid identifications have been published from these sections. Nicoll and Metcalfe (1998) recovered conodonts from a section of Callytharra Formation, equivalent to unit C, at Gooch Range about 14 km northwest of locality 5. The conodonts were attributed to their *Mesogondolella bisselli* – *Sweetobnathus inornatus* Zone known also from the Maubisse Formation in Timor and considered to be late Sakmarian (Sterlitamakian) to early Artinskian (Aktastinian) based on associated ammonoids in the Timor succession (see comment below). Nicoll and Metcalfe (1998) reported similar conodonts from the type section of the Jimba Jimba Calcarenite and from the lower section of Callytharra Formation beneath the Winnemia Sandstone Member at this locality (see below). UWA students found five very small ferruginized internal moulds of juvenile ammonoids from friable beds in unit C. Tatyana Leonova (Paleontological Institute, Moscow, written comm.) has identified these as *Pseudohistoceras*? spp. and *Metalegoceras* sp., broadly indicative of the Sakmarian or Artinskian. *Pseudoschistoceras simile* Teichert was found in the Cordalia Sandstone immediately overlying the Callytharra Formation about 70 km north of locality 5 (Cockbain, 1980). The age of this ammonoid depends on correlation to the Bitauni fauna in the Permian of Timor that includes taxa not stratigraphically constrained with any rigor, including forms that may range from late Sakmarian to Roadian (Owen in Charlton et al., 2002). Owen suggested that it was 'impractical' to correlate the Western Australian post-Sterlitamakian ammonoid succession with the Timor succession because of the very low diversity of the Australian fauna.

It is difficult to subdivide the locality 5 succession using the brachiopod zonation based on the assemblages outlined

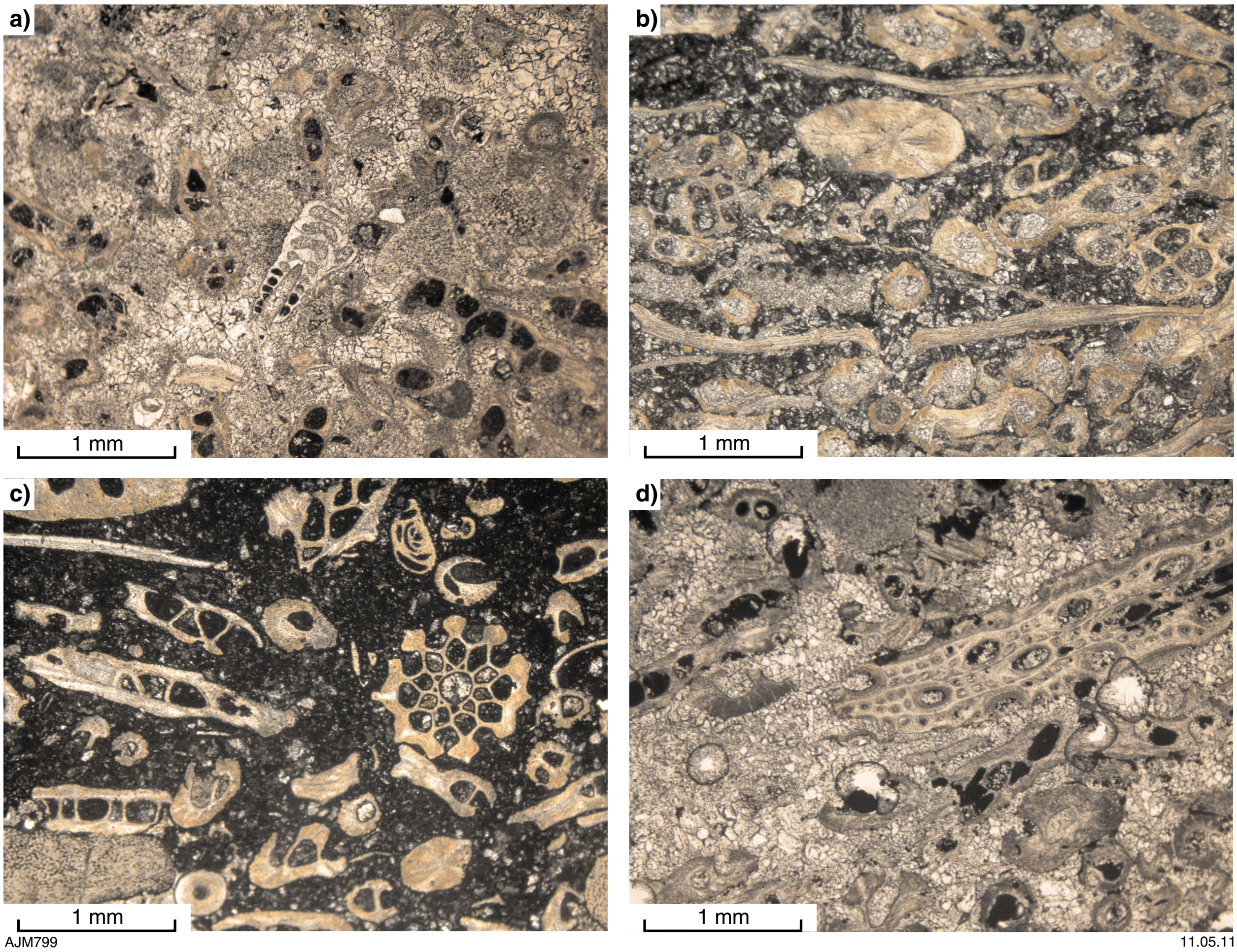


Figure 31. Acetate peel photographs of limestones from the Callytharra Formation, Dead Man's Gully: a) bryozoan packstone with recrystallized matrix from unit A, 3 m above base of formation; b) bryozoan–crinoidal packstone with silty-mud matrix from unit B, 52 m above base of formation; c) bryozoan packstone with silty-mud matrix and rare foraminifera (*Hemigordius* sp.) from unit B, 55 m above base of formation; d) bryozoan grainstone with sparry calcite cement from unit C, 60 m above base of formation. Bar scales are 1 mm

by Archbold (1993, 1998a) and Hogeboom and Archbold (1999). The assemblage zones are obviously facies influenced and these authors did not chart the species distributions within any outcrop sections. Characteristic species present at locality 5 that are apparently confined to the *Strophalosia irwinensis* Zone (Archbold and Dickins, 1989; Archbold 1998a) include *S. irwinensis*, present in the lower 10 m of the formation here, and *Cimmeriella foordi* (= *Globiella foordi*; see Archbold and Hogeboom, 2000) confined to the lower 5 m. According to detailed sampling by Dixon and Haig (2004) of the type section of the Callytharra Formation, the latter species is present throughout the lower 59 m but *S. irwinensis* is very rare and recorded from only one bed 14 m above base of that section. *Strophalosia jimbaensis* and *Cimmeriella flexuosa*, key species for identification of the *S. jimbaensis* Zone that overlies the *S. irwinensis* Zone (Archbold 1998a; Archbold and Hogeboom, 2000), have not been found at locality 5. In the type section of the Jimba Jimba Calcarenite, *S. jimbaensis* is common at 59.5 m above base of section and very rare at 69.8 m above base. It appears to be absent from other beds in this section (Dixon, 2000). This species has been recorded, at least in published literature, at one other site outside of the type locality (Archbold, 1991). According to Dixon's (2000) bed-by-bed study of the type Jimba Jimba Calcarenite, *Cimmeriella* is found abundantly in beds 7.2 m and 14.8 m above base of the type section but is sporadic and very rare above this level.

Blue-grey mudstone from the spoil heap of a shallow seismic shothole at 23°54.137'S, 114°56.926'E in unit C of the Callytharra Formation, yielded a spore-pollen assemblage, including very rare *Gondisporites ewingtonensis*, rare *Diatomozonotriletes townrowii*, *Brevitriletes cornutus, Horriditriletes spp.*, and *Pseudoreticulatispora pseudoreticulata,* and common *Microbaculispora trisina*, placed in Backhouse's (1991, 1998) *Microbaculispora trisina* zone (Purcell, 2008, unpublished). This zone correlates with the upper Callytharra Formation (Jimba Jimba Calcarenite equivalent) and overlying Wooramel Group in subsurface sections (Mory and Backhouse, 1997), well above the *Pseudoreticulatispora pseudoreticulata* zone in the lower Callytharra Formation in the surface of the Merlinleigh and Byro Sub-basins (Mory and Backhouse, 1997) and at a level above the type section of the Fossil Cliff Member in the Irwin Sub-Basin (Foster et al, 1985; Backhouse, 1991).

The bryozoan–crinoid–brachiopod association dominant in the skeletal debris forming the packstones and grainstones of the Callytharra Formation is indicative of a cool-water 'heterozoan' fauna (as distinct from a warm-water photozoan assemblage, following James, 1997). The faunal association suggests shallow-water conditions and, at least at the times of carbonate deposition, normal salinity and oxygen levels. The limestone beds may represent the tops of shoaling-upwards depositional cycles with shell debris concentrated, particularly in unit C, by wave and other current activity. The change from packstone with micritic matrix to packstone with silty mud matrix to grainstone containing laminae of fine to medium-grained quartz suggests an overall shoaling upward (progradational) trend for the succession. The transition from thin-bedded, fine-grained sandstone of the Cordalia Formation at the base of the Wooramel Group to overlying thick-bedded, coarse-grained sandstone of the Moogooloo Sandstone represents a conformable extension of the progradational trend. In early studies, the contact between the Callytharra Formation and the overlying Cordalia Formation was interpreted as a significant disconformity.

Compared to other sections of the Callytharra Formation to the south of Dead Man's Gully, Unit A corresponds to the lower part of Callytharra Formation below the Winnemia Sandstone Member in the Jimba Jimba Syncline, about 140 km south of this locality (Fig. 32). Other lateral equivalents include the type section of the Callytharra Formation at Callytharra Springs, about 220 km to the south, and the type section of the Fossil Cliff Member of the Holmwood Shale (possibly the uppermost part of the shale unit beneath this section) in Irwin River about 560 km to the south (Fig. 32). Unit B seems to correlate to the Winnemia Sandstone Member at Jimba Jimba, to the lower part of the 'Nunnery Sandstone' at Callytharra Springs, and to the High Cliff Sandstone at Irwin River. Unit C correlates to the type section of the Jimba Jimba Calcarenite. In some of the early studies of the Permian in the basin, the Jimba Jimba Calcarenite was considered a member of the Billidee Formation (at **locality 8**) and fossil assemblages from it were sometimes attributed to the latter formation. The Billidee Formation overlies the Moogooloo Sandstone at a much higher stratigraphic level than the Jimba Jimba Calcarenite (following Mory and Backhouse, 1997).

The broad facies similarities over large distances, the shallow-water nature of the facies, and the near horizontal bedding across the basin (as shown in seismic profiles) suggest that sea-floor gradients were very gentle throughout the interior rift basin (i.e. there was no shelf–basin topography). The effects of eustatic changes following melting of the Early Permian continental ice sheets and tectonic subsidence associated with rifting are difficult to separate, but by the time of deposition of the Callytharra Formation sediment influx probably keep pace with accommodation. Along the interior-rift basin, the northern part of the Merlinleigh Sub-basin was under the influence of more open-marine conditions. These conditions became more restricted farther south in the basin so that the Permian is completely non-marine about 900 km south of locality 5 (south of Perth).

The red-colour of sandstone in the Wooramel Group and the presence of iron-oxides in many parts of the Callytharra Formation succession at this locality are attributed to iron mobilization and precipitation from ground-waters under the present-day semi-arid intense weathering regime.

Locality 6: South Branch Well

Significance: An exposure of diamictite belonging to the Lyons Group at this locality contains clasts of a wide range of sizes (including boulders several metres in length) and rock types. These deposits have been interpreted as tillite or as glacially influenced subaqueous gravity flows.

Location: Gravel pit immediately east of the Williambury–Moogooree road, just north of the crossing of the South Branch of the Minilya River at 23°59'S, 115°10'12"E, Winning Pool.

Geology: A matrix-supported unsorted polymict conglomerate belonging to the Lyons Group is exposed at this locality. The largest boulder (a granite) stands about 2 m high and sizes of other clasts range down to granules (Fig. 33). The size of the granite boulder and the variety of other rock types incorporated in a rock 'flour' matrix suggests a glacial, ice-rafted origin. Condon (1967) interpreted the deposit as a tillite. A larger boulder of granite is present in a similar but stratigraphically higher conglomerate about 6 km upstream on the south side of the river (near 23°57'19"S, 115°06'49"E).

Condon (1967) designated these diamictitic exposures as the type area of the 'Thambrong Formation' (Fig. 34), below the 'Weedarra Shale' in the upper part of the Lyons Group. He characterized the formation as alternating tillitic siltstone and tillitic quartzwacke. Condon (1967) measured a formation thickness of 365 m (Fig. 35), but outcrop is poor and there may be structural complications in the area.

Locality 7: Track from Moogooree Homestead to Toby Bore

Significance: This locality has a series of laterally extensive boulders beds, each probably a metre to several metres thick, in the middle section of the Lyons Group (Fig. 36). Rare, sparsely fossiliferous beds of sandstone are present. The stratigraphic succession suggests a low gradient seafloor and sheet-like architecture of the boulders beds.

Location: The series of boulder beds is crossed along the Moogooree to Toby Bore track between about 24°05'14"S, 115°11'21"E and 24°05'27"S, 115°10'57"E, Kennedy Range. This stretch of track traverses a very low flat-topped ridge that trends approximately 330°.

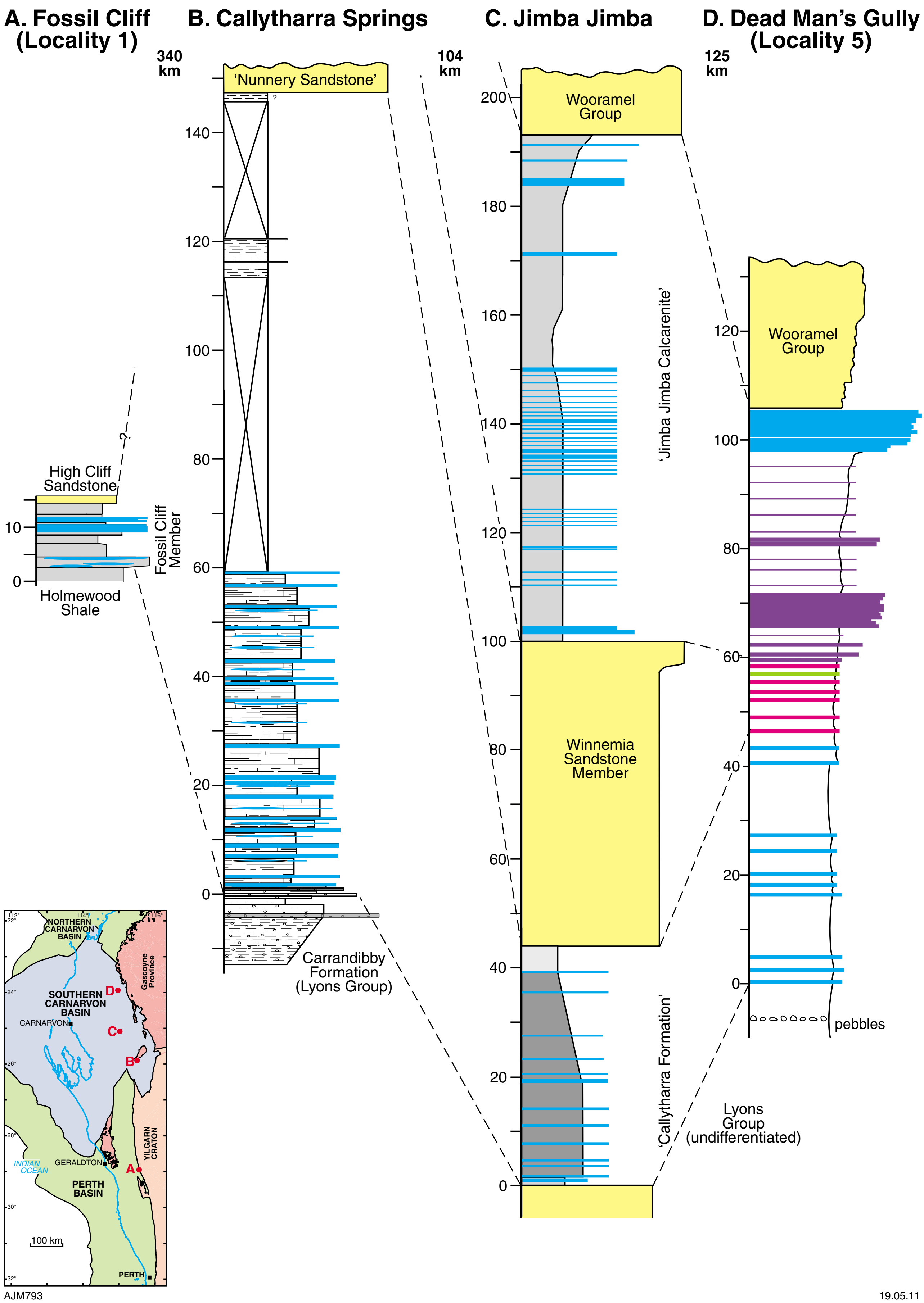


Figure 32. Correlation of Callytharra Formation, Dead Man's Gully, with type sections of coeval units

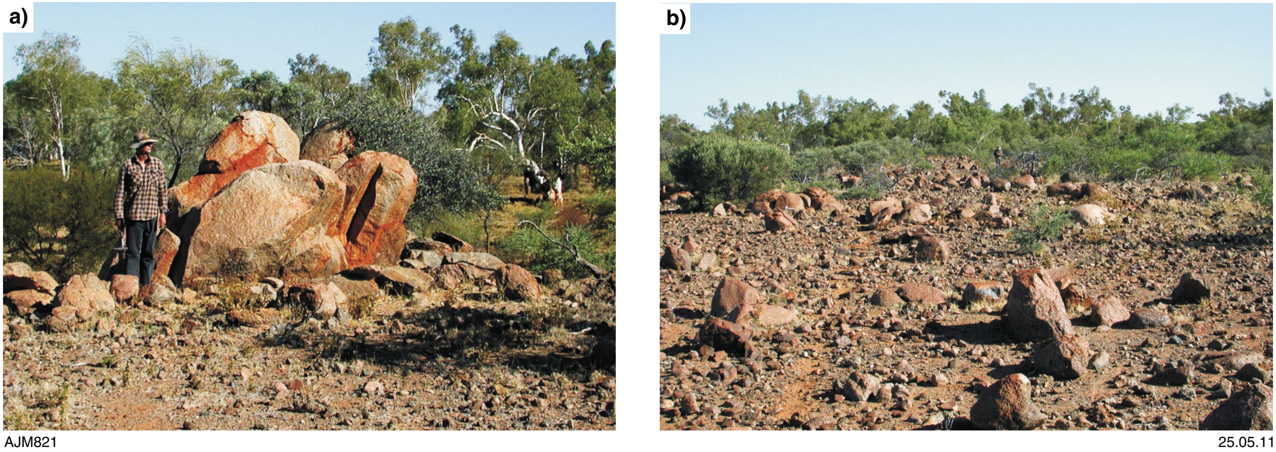


Figure 33. Boulder bed at South Branch Well locality: a) large granite boulder; b) general view along strike

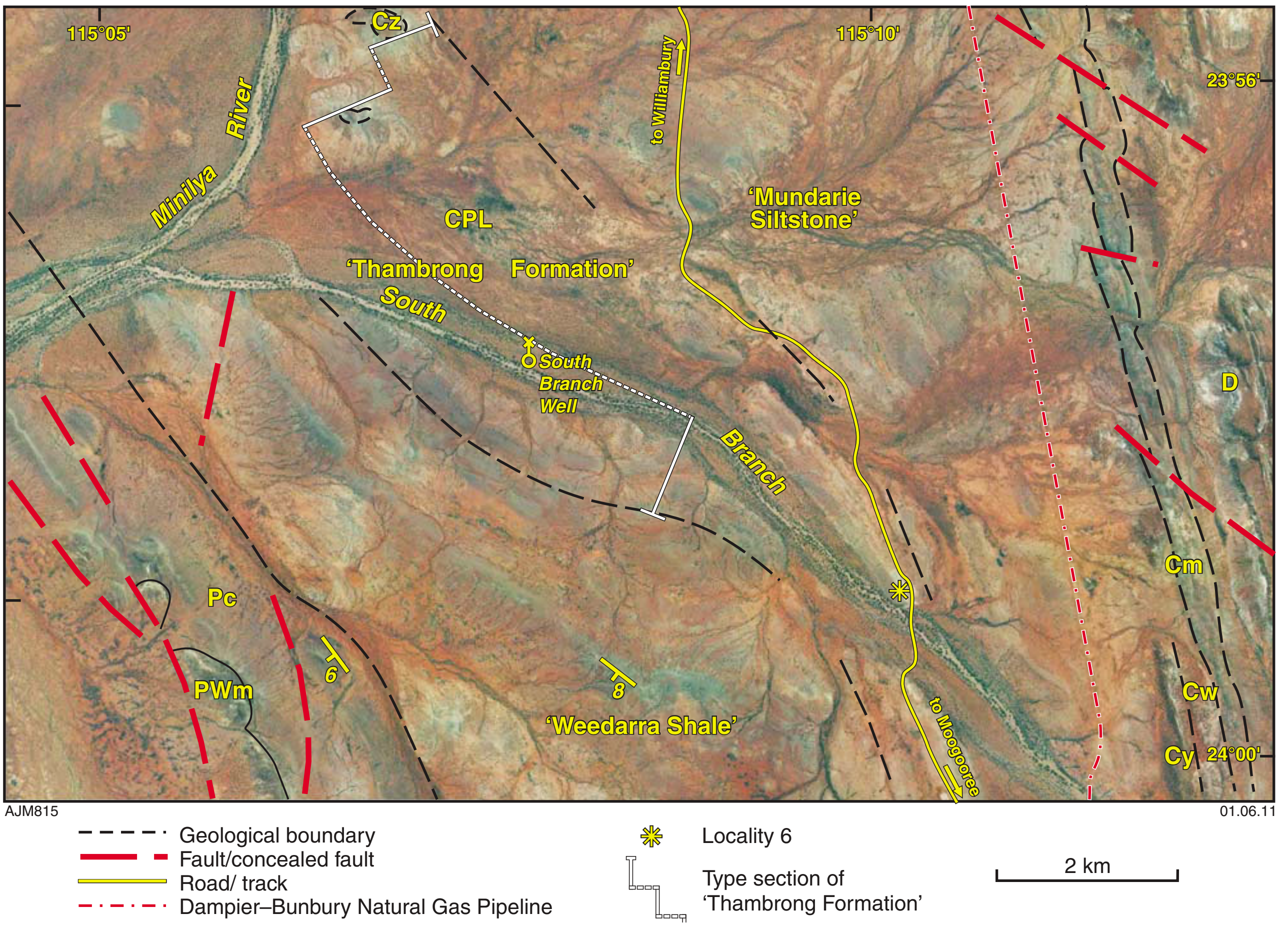


Figure 34. Simplified geological map superimposed on aerial image of South Branch Well area (after Condon, 1967). D – Devonian, Cm – Moogooree Limestone, Cw – Williambury Formation, Cy – Yindagindy Formation, Ch – Harris Sandstone, CPL – Lyons Group, Pc – Callytharra Formation, PWm – Moogooloo Sandstone, PWb – Billidee Formation, PBc – Coyrie Formation, Cz – Cenozoic

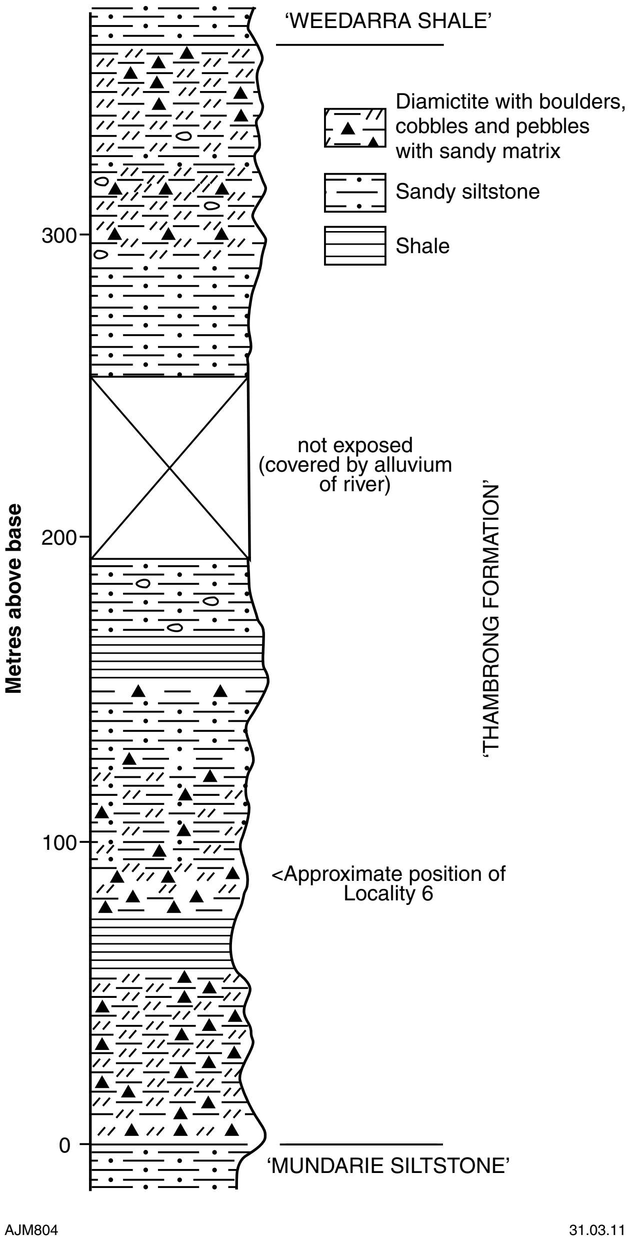


Figure 35. Measured section of the 'Thambrong Formation' (after Condon, 1967)

Geology: The dominant rock type in the area — friable shale or fine-grained sandstone — is not exposed. Weathered-out clasts from diamictites are left as boulders, cobbles, and pebbles strewn in linear fashion parallel to the axis of the low ridge. The lines of debris can be traced for up to 2 km before being covered by alluvium of broad braided drainage systems. A large variety of clast types is present, almost all derived from Precambrian areas to the east.

Rare outcrops of sandstone have yielded a meagre fossil assemblage. Dickins and Thomas (1959) record the following assemblage from a locality (F17203) 4.8 km west of Moogooree Homestead: brachiopods including Spiriferacea gen. et sp. indet., bivalves including *Nuculana lyonsensis* Dickins, *Astartila condoni* Dickins, *Cleobis* sp., *Parallelodon*? sp. indet., *Aviculopecten tenuicollis* (Dana); Scaphopoda and wood fragments.

Locality 8: Toby Bore

Significance: The stratigraphic succession west of Toby Bore includes an almost continuous exposure of the progradational transition through the Callytharra Formation, Cordalia Formation, and Moogooloo Sandstone (probably late Sakmarian through late Artinskian) followed by retrogradation from the upper Moogooloo Sandstone through the Billidee Formation to the Coyrie Formation (probably late Artinskian or early Kungurian). Deposits of the retrogradational phase include rare dropstones that are indicative of icebergs and suggest climatic cooling after a warmer phase represented in particular by the Callytharra Formation.

Location: Toby Bore lies on a track about 7 km west of Moogooree homestead (at 24°06'05"S, 115°10'15"E, KENNEDY RANGE). Sites that best illustrate the stratigraphic succession are shown in Figure 37.

Geology: The stratigraphic succession is outlined in Figure 38. The Callytharra Formation shows similar facies changes to those observed at **locality 5**. Note particularly the concentrations of crinoid debris and spiriferids in successive beds at site 2. Overlying the Callytharra Formation at the base of a low scarp (site 3) is a thin unit of thin-bedded fine-grained quartz sandstone that is correlated to the Cordalia Formation. This grades upward into thick-bedded coarse-grained quartz sandstone that is the base of the Moogooloo Sandstone (site 4). The red colour is due to recent iron staining by ground-water flow through the permeable sandstone, a product of the present-day arid climate. Bed forms include large-scale trough cross bedding, contorted bedding (site 5) and large-scale planar cross bedding (site 6).

A significant change occurs in the upper part of the Moogooloo Sandstone at site 7 (Fig. 39a) where a friable fine-grained sandy mudstone is overlain by a 20 cm thick medium- to coarse-grained sandstone with abundant *Skolithos* burrows followed by thin (5–10 cm) beds of coarse-grained quartz sandstone (Fig. 39a). About 15 m above the *Skolithos* bed, a three-metre thick mudstone unit (Fig. 39b) forms a conspicuous light-coloured marker horizon on aerial photographs (site 8, Fig. 37) and is the base of the Billidee Formation as charted on the stratigraphic log of the type section by Condon (1967). A granitic erratic, about 30 cm in diameter, is present at site 8 and is interpreted as a dropstone indicative of icebergs. The provenance of the rock lies in the Precambrian basement exposed to the east of the basin. A second, thicker mudstone unit about 30 m above the base of the Billidee Formation is best observed in the scarp of a high ridge at site 12. Below this mudstone is a unit of thin- to medium-bedded quartz sandstone that contains rare fossils including large elongate burrowing bivalves. A fossiliferous heavily ferruginized quartz sandstone forms the topmost unit of the Billidee Formation.

The lower Coyrie Formation (Figs 37, 38, 40) is dominated by mudstone. Some bioturbated sandstone and repetitive laminated to burrowed facies are present near the base of

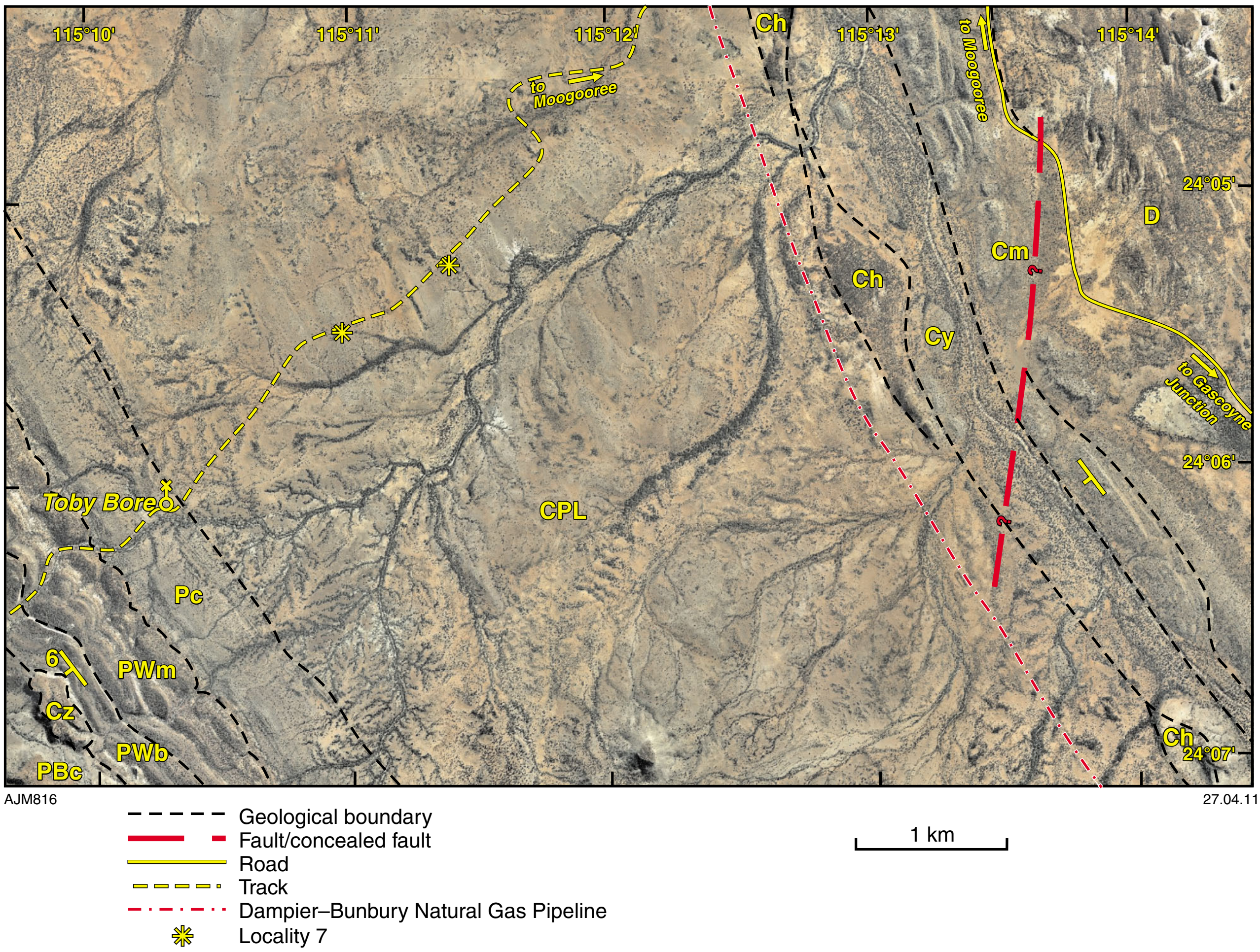


Figure 36. Simplified geological map superimposed on aerial image of the Grant Group between 'Moogooree' and Toby Bore. D – Devonian, Cm – Moogooree Limestone, Cy – Yindagindy Formation, CPL – Lyons Group, Ch – Harris Sandstone, Pc – Callytharra Formation, PWm – Moogooloo Sandstone, PWb – Billidee Formation, PBc – Coyrie Formation Cz – Cenozoic

the section. In places, erratics of various sizes are present in the mudstone and are interpreted as dropstones from ice bergs. To the west low hills of the sand-dominated Mallens Sandstone can be seen. This unit forms the top of the Coyrie Formation – Mallens Sandstone succession that is suggestive of an overall shoaling upward trend (Fig. 40).

A diverse fossil assemblage has been described from the Coyrie Formation; however, macrofossils are confined to relatively few beds. Groups identified include foraminifera, bryozoans, brachiopods, bivalves, ammonoids, nautiloids, gastropods, conulariids, trilobites, scaphopods, crinoids, annelids, fish, plants, and various trace fossils (Skwarko, 1993). A similar but less diverse association of foraminifera, brachiopods, bivalves, ammonoids, gastropods, conulariids, crinoids, trace fossils, and fossil wood has been recorded from the Mallens Sandstone (Skwarko, 1993).

The age of the succession is bounded by a late Artinskian determination for the Cordalia Sandstone (assuming synchronicity with the unit in its type section, 100 km to the north; see **Cordalia Formation**) and the late Artinskian or Kungurian age determined from ammonoids in the Billidee Formation at this locality (see **'Billidee Formation'**). *Neocrimites* sp. (= *Aricoceras* sp. according to Leonova, 1998), also recorded in the Billidee Formation, and *Bamyaniceras* sp. found in the Coyrie Formation in this area suggest a late Artinskian or Kungurian age (Glenister and Furnish, 1961; Cockbain, 1980), equivalent to the Yakhtashian and Bolorian stages indicated by Leonova (1998). Elsewhere in the basin specimens of *Pseudoschistoceras simile*, also known from the type section of the stratigraphically lower Cordalia Sandstone (Teichert, 1944; Glenister and Furnish, 1961; Cockbain, 1980), are present in the Coyrie Formation. From the Mallens Sandstone just to the northwest of the excursion site, *Bamyaniceras* sp. similar to the species recorded from the Billidee Formation has been identified (Cockbain, 1980).

Locality 9: Minilya River, 'Williambury'

Significance: Large scale slumping is attributed to glacial action, sediment loading, and/or tectonic influences.

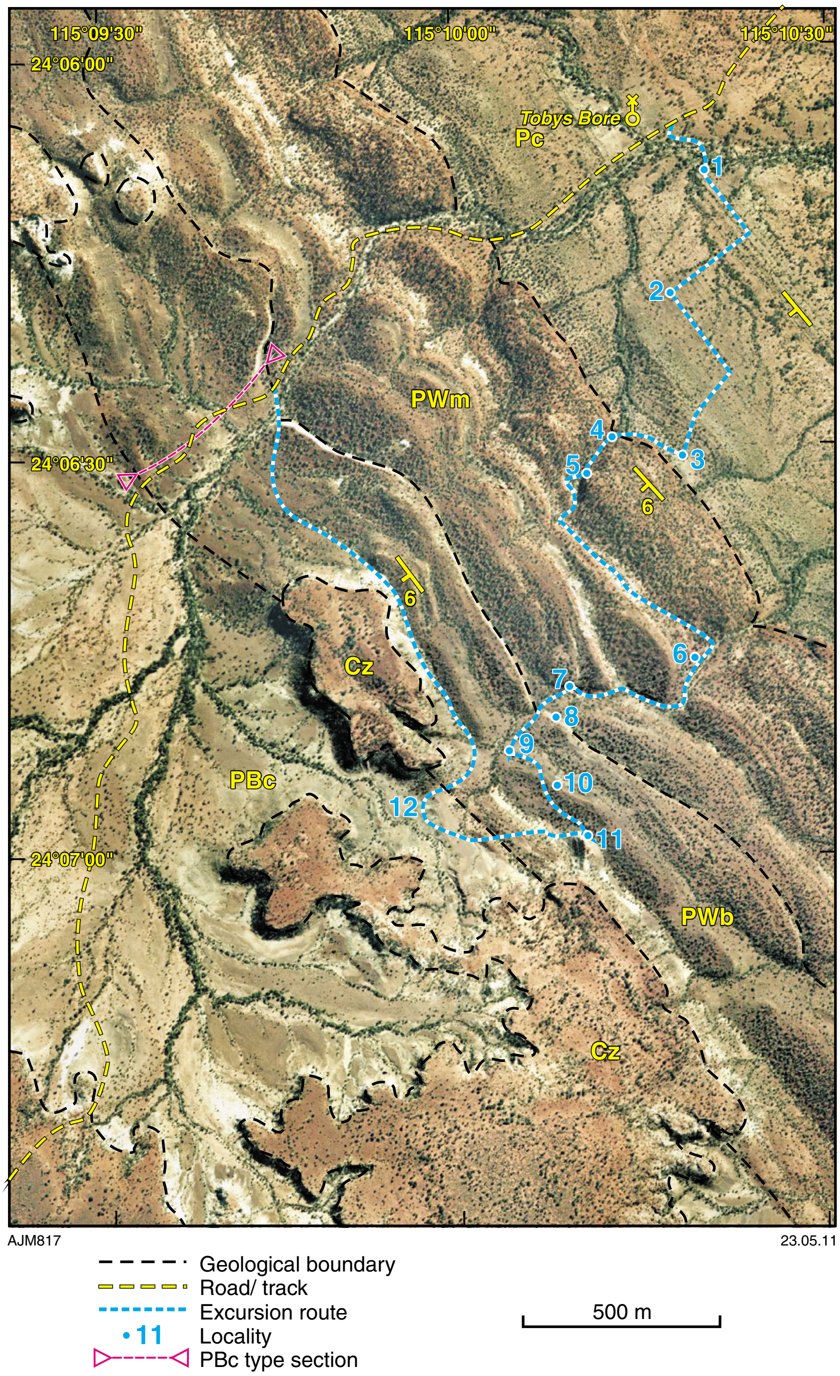


Figure 37. Simplified geological map superimposed on aerial image of the Toby Bore area (after Condon, 1967). Pc – Callytharra Formation, PWm – Moogooloo Sandstone, PWb – Billidee Formation, PBc – Coyrie Formation, Cz – Cenozoic. Numbers refer to sites shown on Figure 38

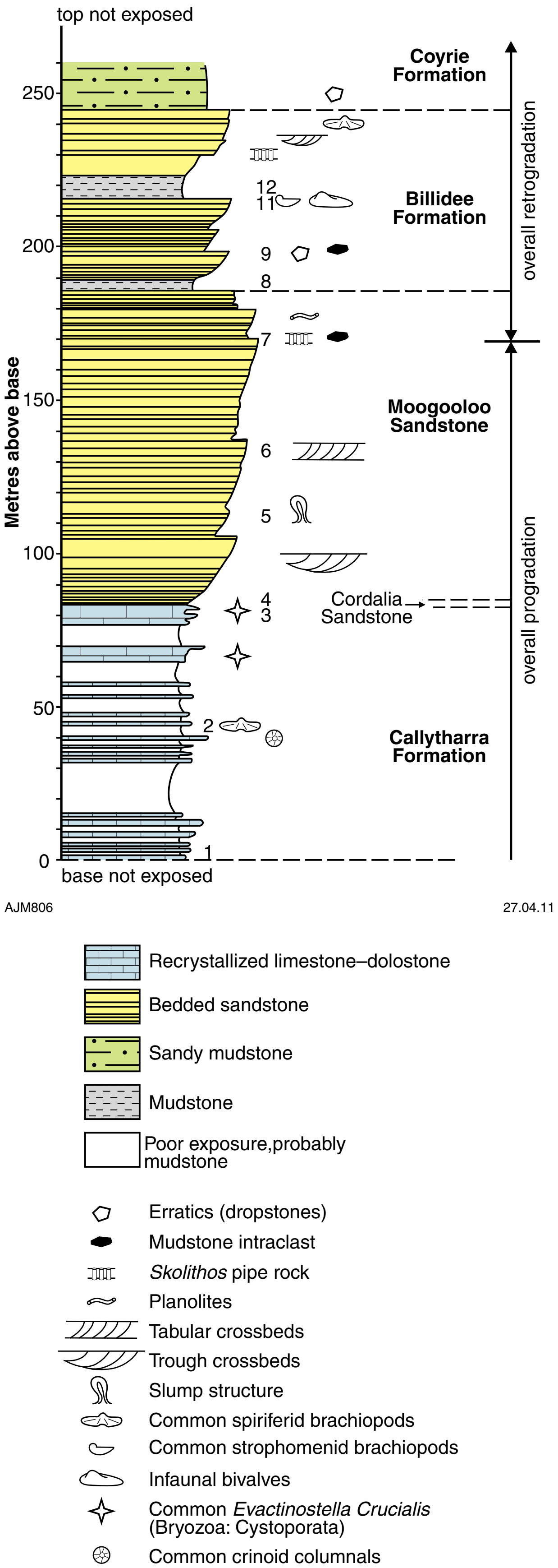


Figure 38. Measured section of the Callytharra to Coyrie Formations, Toby Bore area

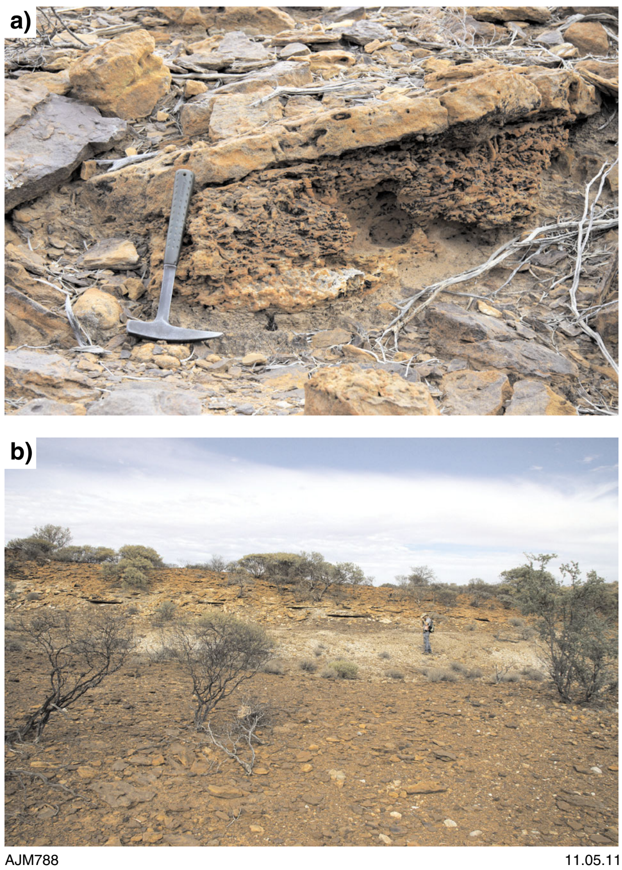


Figure 39. Outcrop images at Toby Bore: a) weathered sandy mudstone overlain by bioturbated sandstone (with *Skolithos*) in the upper Moogooloo Sandstone at site 7 marking the base of a major retrogradational succession; b) weathered mudstone unit (light coloured) that forms the base of the Billidee Formation at site 8 (sites shown on Fig. 37)

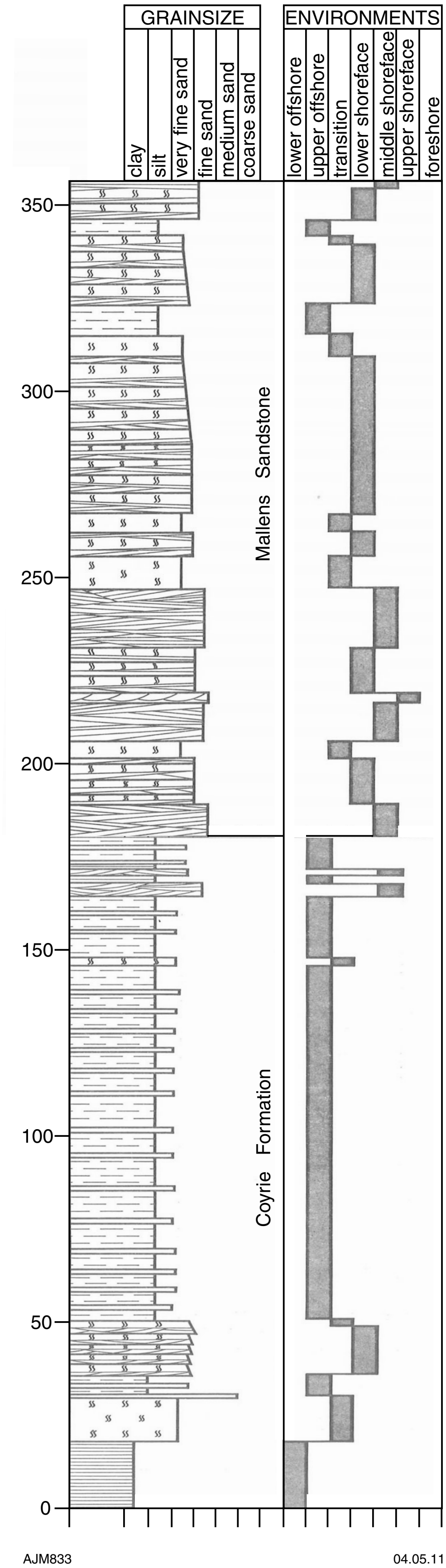


Figure 40. Stratigraphic log of the type sections of the Coyrie Formation and Mallens Sandstone, Toby Bore area (from Moore et al., 1980a). See Figure 25 for explanation of facies

Location: 800 m east of Williambury homestead at 23°51'34"S 115°09'18"E, Winning Pool.

Geology: The wide ridge of Harris Sandstone south of the Minilya River is fault bounded against Precambrian granite to the west, and overlies the Lower Carboniferous succession to the east. Dips within the ridge are about 20°W, and although there is some fault repetition, the Harris Sandstone is thicker compared to most other areas. See **Harris Sandstone** for a discussion of problems in recognizing and correlating this unit. Slumping of a massive sandstone unit between two siltstone horizons is likely to be due to fluidization of the sand and collapse of the overlying silts, possibly triggered by sediment loading, ice loading, or shock from an earthquake. The unit appears to be a nearshore marine deposit.

Locality 10: Minilya River south branch, 3 km NW of 'Moogooree'

Significance: This locality has the thickest, well-exposed section of Yindagindy Formation present in the region (Figs 41–43). It illustrates a major cycle of shallow-marine carbonate deposition during the Visean. The most conspicuous rock types are oolitic, peloidal, and aggregate-grain (grapestone) grainstone, and some skeletal wackestone and packstone.

Location: About 3 km northwest of Moogooree Homestead, and 1.3 km east of the Williambury to Moogooree Road at 24°02'17"S, 115°12'20"E, Kennedy Range.

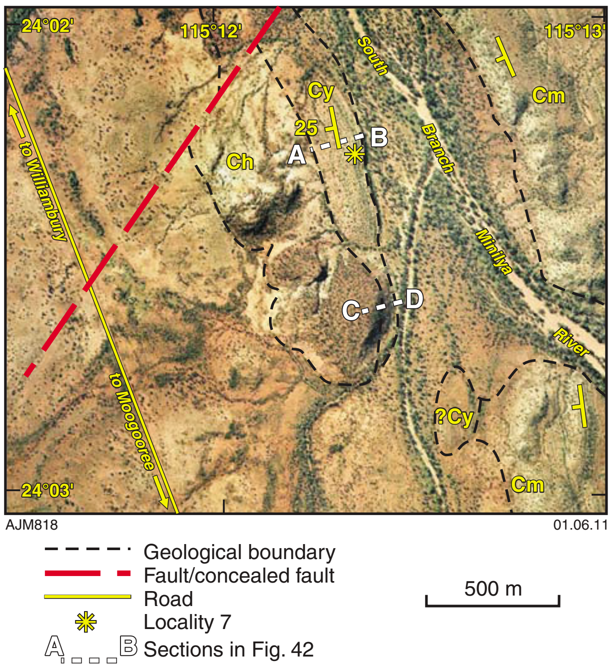


Figure 41. Simplified geological map superimposed on aerial image of the Yindagindy Formation, 3 km north of 'Moogoorie' (after Read et al., 1973). Cm – Moogooree Limestone, Cy – Yindagindy Formation, Ch – Harris Sandstone

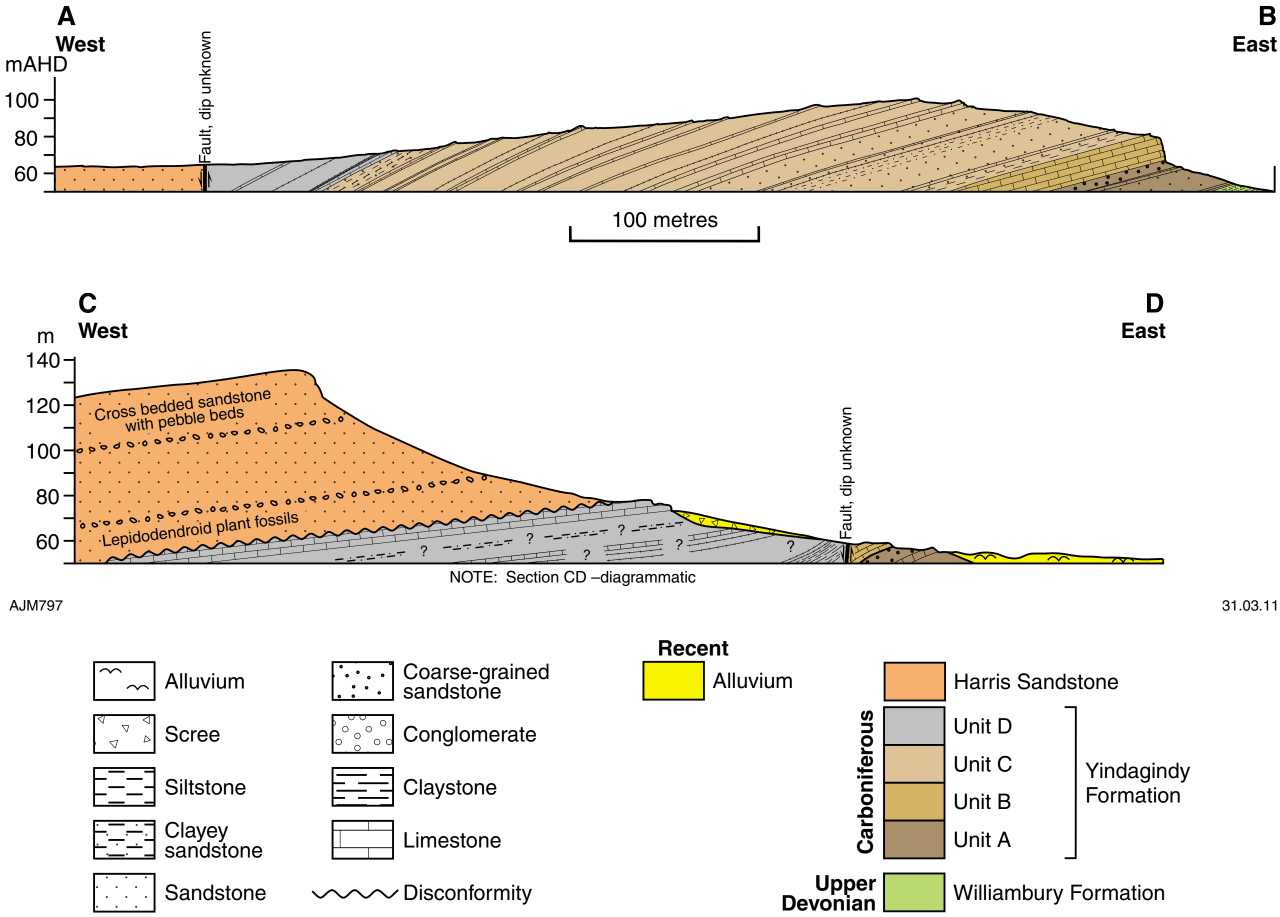


Figure 42. Geological cross-sections A–B and C–D (from Alcock et al., 1966; see Figure 40 for locations)

Geology: In the Moogooree region the Yindagindy Formation has been divided into four units all of which are exposed in section A–B (Fig. 43; Alcock et al., 1966; Read et al., 1973). The uppermost part of the Williambury Formation outcrops at the base of the section as friable, poorly sorted, medium- to coarse-grained sandstone. The Yindagindy Formation conformably overlies the Williambury Formation with the base of the formation marked by a thin limestone bed.

The lowest unit of the Yindagindy Formation (Unit A, Figs 42, 43) is about 7 m thick and is exposed beneath scree on the east side of the low limestone scarp near the start of section A–B. It consists mainly of poorly exposed fine- to coarse-grained, thinly bedded or cross-bedded sandstone. A second thin limestone bed is present about 5 m above the basal limestone bed, and is a medium-grained oolitic grainstone (Fig. 44a) with ooids developed mainly around angular quartz grains.

Unit B is about 5 m thick and forms the low scarp near the base of the section. It includes thick-bedded packstone with framework grains, mainly brachiopod fragments (Fig. 44b), and grainstone (Fig. 44c) composed mainly of varying proportions of peloids, ooids, and skeletal grains (mostly cortoids, including brachiopod, ostracod, crinoid, bryozoan, and possible green algal debris, recrystallized gastropods and rare endothryoid foraminifera). Minute 'calcispheres' are common in some of the beds; aggregate grains (grapestones) are rare.

Unit C, about 41 m thick, consists of units of poorly exposed friable calcareous sandstone and mudstone alternating with thin to medium beds of grainstone and packstone. Conspicuous grains forming the limestone include aggregate-grains (grapestone), peloids and ooids. In the aggregate-grain dominated facies (Fig. 44d), endothyriid foraminifera are common. Ostracods are common in many of the limestones of this unit and are the dominant grains in some of the wackestone beds high in the unit (Fig. 44e,f).

Unit D comprises the upper 15 m of the section and consists of very poorly exposed friable calcareous sandstone/mudstone alternating with thin beds of wackestone including some laminated units with fenestral fabric. Some of the beds have been extensively dolomitized.

The contact with the overlying Harris Sandstone may be observed on the C–D transect at the southern end of the locality. Here a vuggy limestone, the topmost bed of the Yindagindy Formation, is overlain by sandstone that in places contains abundant lepidodendroid stem impressions. A limestone-pebble conglomerate, up to 1 m thick, is present intermittently at this level.

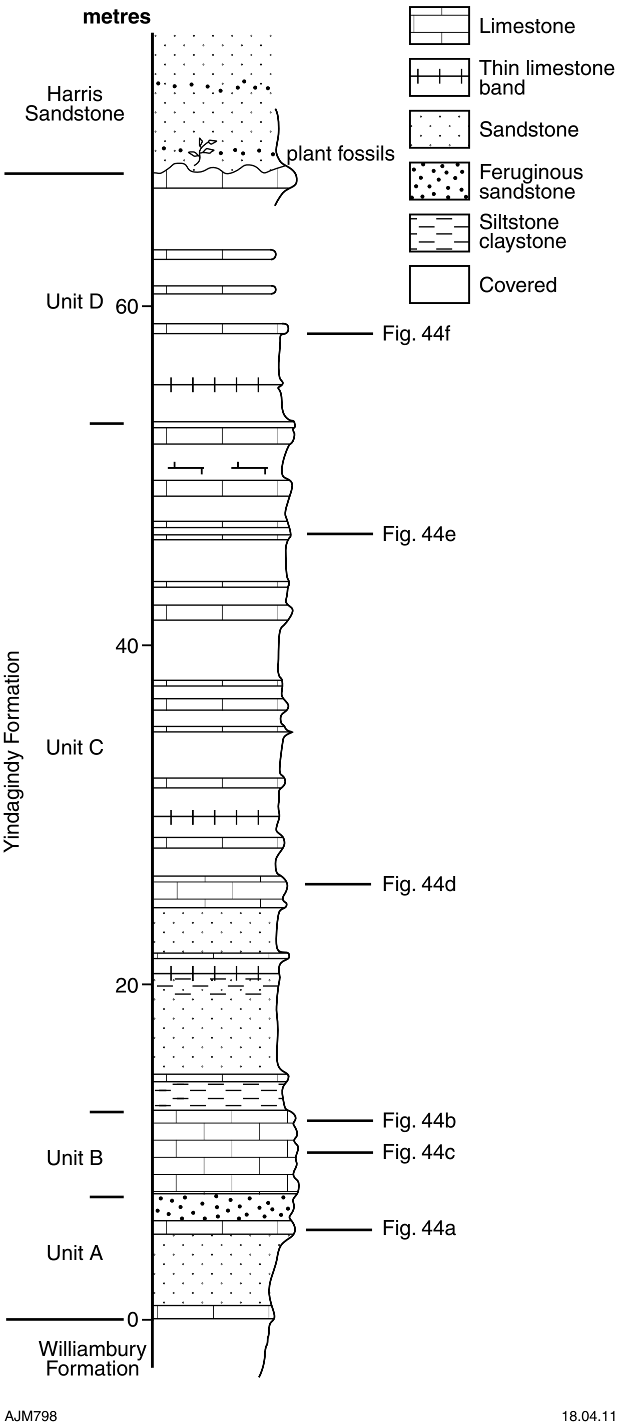


Figure 43. Composite stratigraphic log of the Yindagindy Formation 3 km north of 'Moogoorie' (after Read, 1966; Read et al., 1973)

The Yindagindy Formation is part of a major depositional cycle in a restricted marine basin. Maximum marine flooding to very shallow-water normal-marine to slightly metahaline conditions is evidenced in Unit B that has the most diverse biogenic grain component. Units C and D represent progradational conditions with salinities becoming higher and marine conditions much more restricted. As implied by Read et al. (1973), the Harris Sandstone probably represents the uppermost part of this progradational succession. The Williambury Sandstone below the Yindagindy Formation has not been studied sufficiently to determine the earliest part of the depositional cycle.

The only fossils previously identified to generic level from this locality are globular morphotypes of the calcified cyanobacteria *Ortonella* sp. from Unit D of the Yindagindy Formation and the ostracod *Cryptophyllus* sp. found particularly in wackestone from Unit D (Read, 1966).

During preparation for this guide, common foraminifera were located in an aggregate-grain dominated grainstone in Unit D by DWH. Tentative identifications include common biseriamminids, frequent paleospiroplectamminids, and rare archaesphaerids and tournayellids. The foraminifera are consistent with a Visean age for the Yindagindy Formation. From the Visean Windsor Group in Nova Scotia, Mamet (1970) recorded an abundance of *Biseriammina*?, which seems similar to the Yindagindy species, in facies interpreted to be 'oolitic and algal banks' and 'lagoonal supersaturated and intertidal-supratidal flats'. A similar ecological distribution is likely at this outcrop.

Locality 11: East side of Moogooree to Mount Sandiman Road

Significance: The earliest Carboniferous marine depositional cycle in the basin is represented by the Tournaisian Moogooree Limestone. This locality has the most accessible typical outcrop of the formation but includes a fossiliferous non-dolomitized unit. Most of the exposed beds of the formation are dolostone in which the original microfabric has been destroyed. No detailed stratigraphic or petrographic study has been made from this, or any other, section of Moogooree Limestone.

Location: About 10 km south-east of Moogooree homestead, low ridge on the southeast side of the Moogooree to Mount Sandiman Road at 24°08'29"S 115°15'48"E, Kennedy Range.

Geology: The Moogooree Limestone outcrops along a low ridge that intersects the road (Fig. 45). The base and top of the formation are not exposed. Only carbonate beds outcrop and friable interbeds, probably mudstone-sandstone, are too deeply weathered to be observed at this locality. A repetitive stratigraphic pattern of friable units alternating with indurated carbonate beds is present (Fig. 46) with the proportion of assumed mudstone-sandstone decreasing and the thickness of the carbonate beds increasing up the lower part of the section (0–27 m). The upper part of the formation at this locality appears to contain greater thicknesses of friable beds. In the lower part of the formation many of the carbonate units are fine grained and planar laminated, particularly in the lower beds of these units.

Extensive dolomitization has taken place with most carbonate beds now dolostone. Elsewhere in the basin, Radke and Nicoll (1981) noted the dominance of dolostone

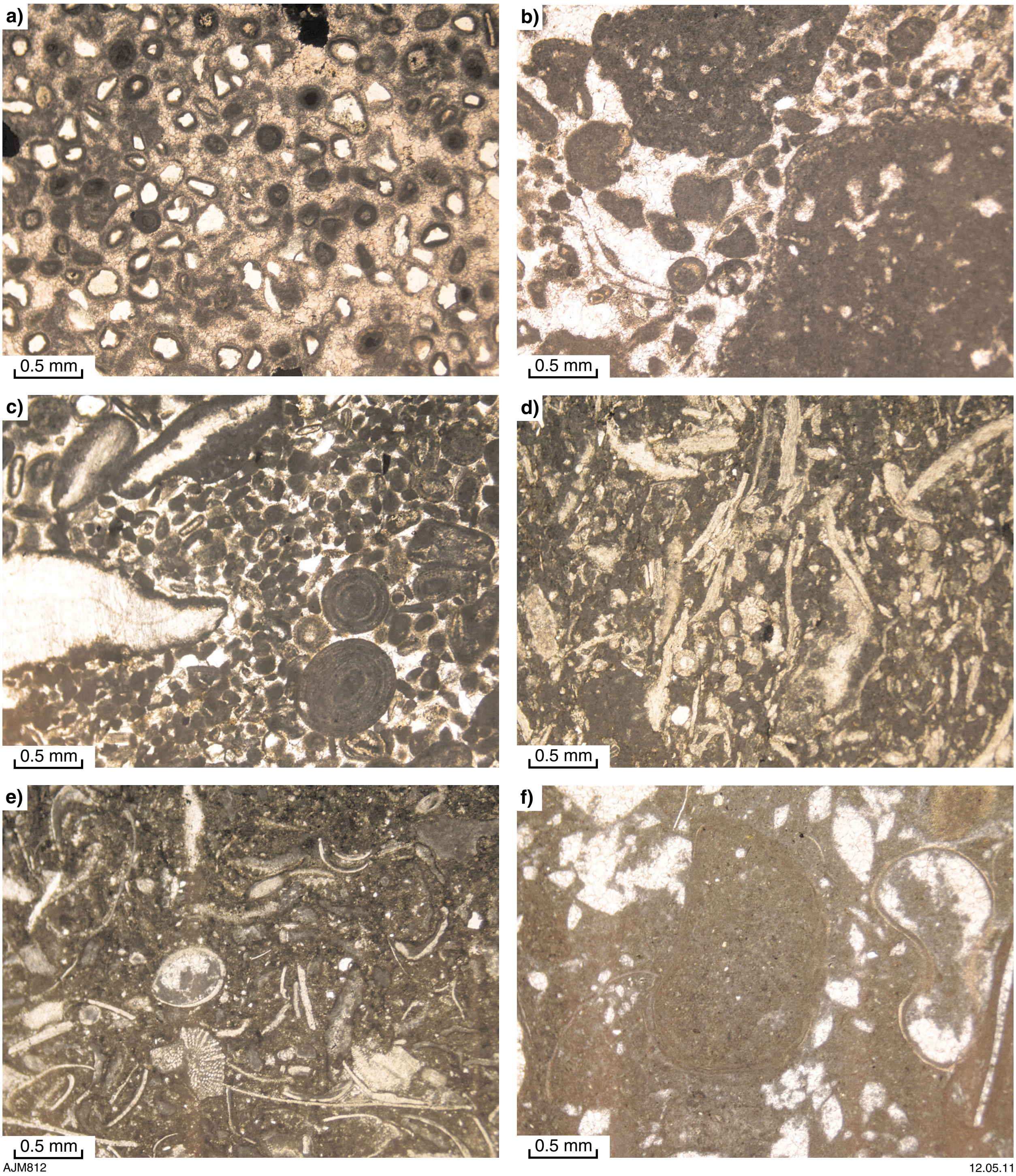


Figure 44. Thin section photographs of limestones from the Yindagindy Formation section A–B, in plain transmitted light: a) ooid grainstone (UWA 57537) from Unit A; b) brachiopod–peloid packstone (UWA 57541) from Unit B: c) peloid–ooid–brachiopod grainstone (UWA 57540) from Unit B; d) aggregate-grain–peloid–foram–ostracod grainstone (UWA 57545) from Unit C: e) ostracod wackestone (UWA 57549) from Unit C; and f) wackestone with gastropods, ostracods, laminated microbial concentrations, and calcite pseudomorphs of probable gypsum (UWA 57551) from Unit D. Stratigraphic positions of samples are shown on Figure 43

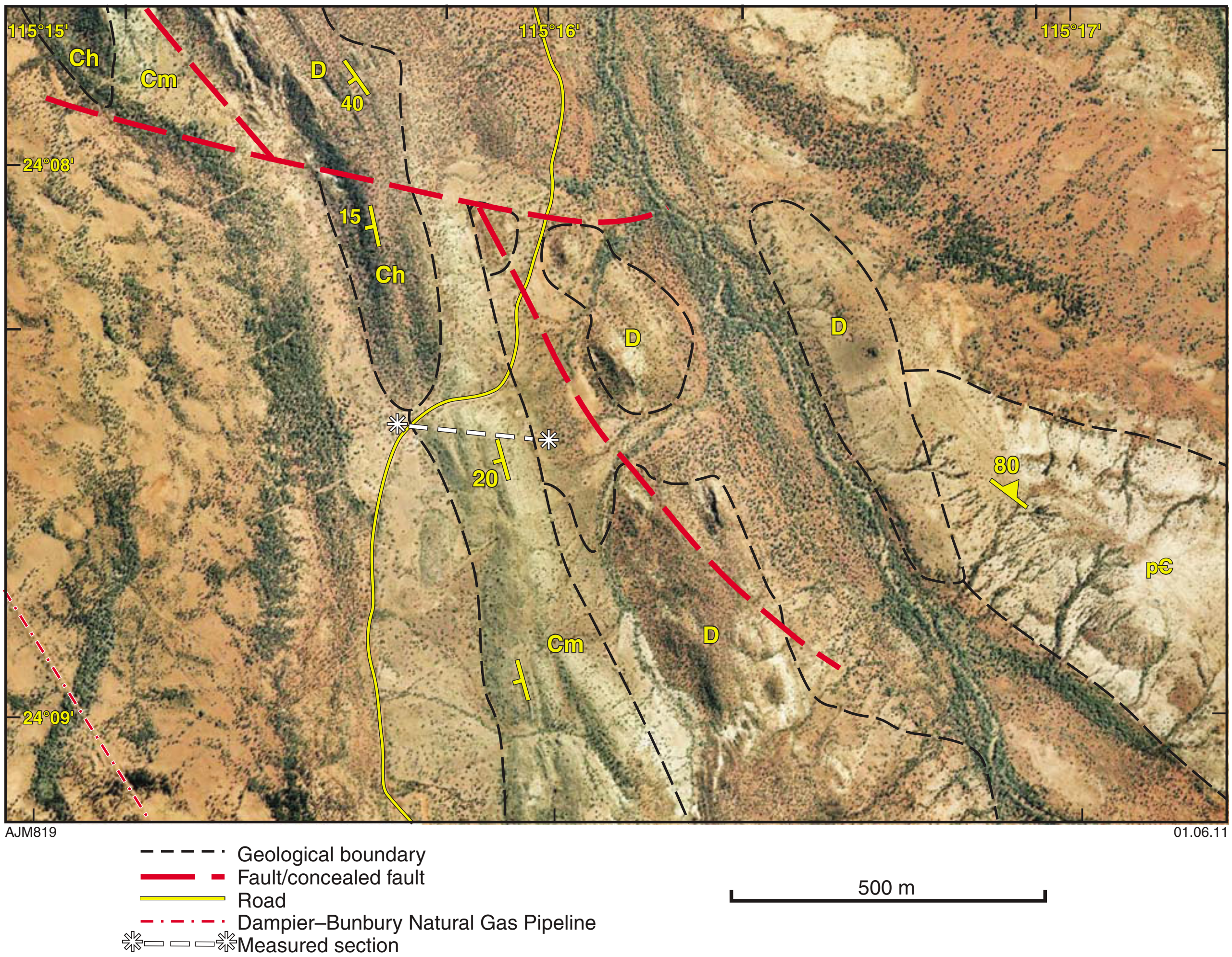


Figure 45. Simplified geological map of the Moogooree Limestone next to the Mount Sandiman Road. p€ – Precambrian, D – Devonian, Cm – Moogooree Limestone, Ch – Harris Sandstone

in the formation. Also present in the some of the carbonate beds are fenestral fabric and birdseye structures suggestive of peritidal conditions (Hocking et al., 1987). Radke and Nicoll (1981) noted that dolomitization was patchy and the light yellowish brown colour of the rock (from iron sesquioxide inclusions) suggested dedolomitization to calcite in some beds. At an outcrop nearer Moogooree Homestead, Hocking et al. (1987) described dedolomitized pipes of calcite, up to 10 cm across, and dedolomitization zones along fractures that resemble breccia with a 'matrix' of coarsely crystalline calcite.

At this locality, a fossiliferous limestone, with only minor patches of dolomite, is present about 50 m above the base of the exposure. This is a coarse-grained, crinoid–brachiopod grainstone that also contains minor bryozoan debris, echinoid spines, and calcareous algae (Fig. 47a–f). Disarticulated silicified valves of brachiopods are strewn on the upper bedding surface (Fig. 47a). Within the bed, only partial silicification is present in some of the brachiopod shells (Fig. 47b).

Because of the diagenetic alteration of the carbonate units in the Moogooree Limestone, very little can be described of the facies succession at this locality or at other localities in the basin. When compared to the Yindagindy Formation (at **locality 9**) a more diverse shelly fauna has been recovered from the Moogooree Limestone. Skwarko (1988) listed 11 brachiopod species from the Moogooree Limestone as against two species from the Yindagindy Formation. In the rare fossiliferous beds near Williambury Homestead, Lavaring (1979) recorded a *Syringopora* biofacies with scattered corals, some in growth positions, accompanied by bryozoan and brachiopod debris and a *Rhipidomella michelini*? biofacies consisting of brachiopod shells and crinoidal debris. The latter biofacies is present at **locality 10**. These biofacies suggest that more open marine conditions existed at least during intervals of Moogooree Limestone deposition than during deposition of the younger Yindagindy Formation. In other parts of the Moogooree Limestone, including in the lower part of this section, laminated carbonates with fenestral structures indicate restricted peritidal conditions (Hocking et al., 1987).

No fossils have been described or identified from this locality. Brachiopods previously described and illustrated from the formation are listed under **'Moogooree Limestone'**.

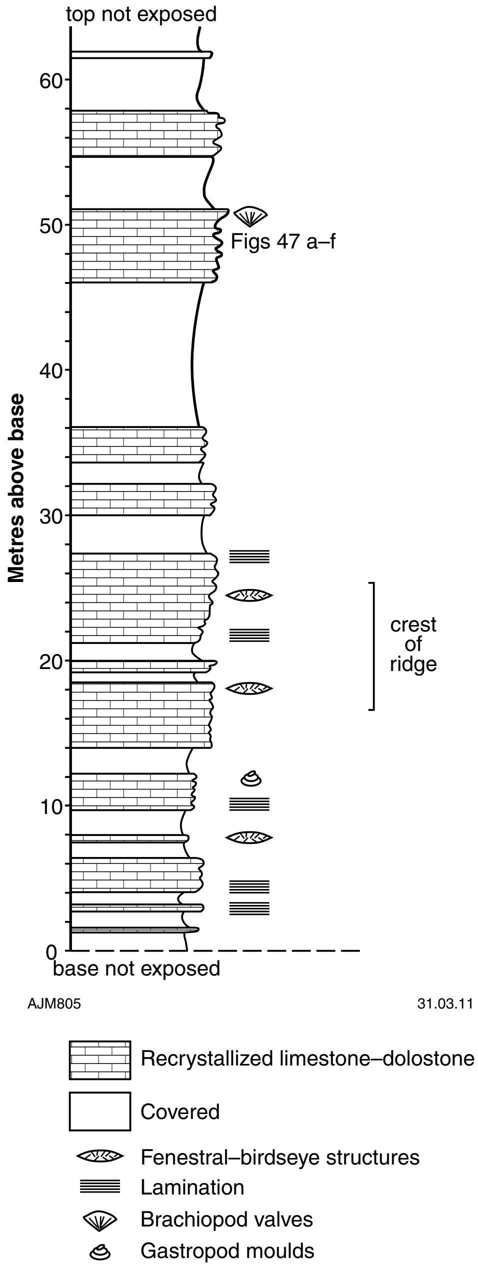


Figure 46. Schematic stratigraphic log of the Moogooree Limestone next to the Mount Sandiman Road. Thicknesses are approximate

Locality 12: Kennedy Range

Significance: The Kennedy Group, a mainly sandstone unit, forms the final Roadian phase of marine sediment progradation within the Merlinleigh Basin with a transition from marine-dominated processes within the Coolkilya Sandstone to fluvial-dominated processes in the overlying Mungadan Sandstone and Binthalya Formation.

Location: East side of Kennedy Range, next to Department of Environment camping area, west along gully starting at 24°40'00"S 115°10'42"E, KENNEDY RANGE.

Geology: A representative section of Coolkilya and Mungadan Sandstones is exposed in the face of Kennedy Range, in and along the face of the spur just south of the car park. Figure 48 shows a similar section about 5 km south of this area. When you reach a large, undercut face and vertical bluff in the Coolkilya Sandstone, you will need to go around to, and maybe beyond, the point of the spur. The section near Range Bore (Fig. 48) is sufficiently similar that it can be used as a rough guide to this section.

The Coolkilya Sandstone in the lower part of the section consists primarily of greenish-grey, fine to medium-grained silty sandstone (feldspathic to lithic wacke). At the base, the dominant facies is swaley cross-stratification, with lesser amounts of hummocky cross-stratification and hummocky to burrowed or laminated to burrowed facies. Bioturbated sandstone and laminated to burrowed facies become dominant higher in the succession. *Zoophycos* traces are abundant near the large undercut and vertical face.

The boundary between the Coolkilya and Mungadan Sandstone is marked by a pronounced change in colour. The Mungadan Sandstone is coarser grained than the Coolkilya Sandstone, and is dominated by bioturbated sandstone and thick, flat-bedded intervals. Some cross-bedded, channel-infill intervals are also present. The facies models proposed by Moore et al. (1980b) imply that the Mungadan Sandstone is, on the whole, a more distal deposit than the Coolkilya Sandstone, in that they considered hummocky and swaley cross-stratification to be shoreface deposits, and bioturbated and flat-bedded deposits to be shelf deposits, seaward of the shoreface. However, the Mungadan Sandstone is best interpreted as a more proximal deposit than the underlying Coolkilya Sandstone. Deposition of the Byro Group and Coolkilya Sandstone was dominated by marine processes, in that there is little indication of continental influence. Storms and waves reworked and distributed sediment that was brought into the basin. The overlying Mungadan Sandstone developed when that marine dominance ceased. According to Moore et al. (1980b) 'the abundance of long-crested oscillation ripples … seaward of the zone of megaripple migration suggests that the cross-stratification was produced … above storm wave base, and therefore possibly formed mainly in response to offshore directed, storm-generated currents'. In such an environment, sands prograded into the basin to form a nearshore, fluvial dominated, subaqueous channel complex. The decrease in marine influence may be due to tectonic silling of the basin, or limestone buildups to the northwest (e.g. Fennel 1, Hope Island 1).

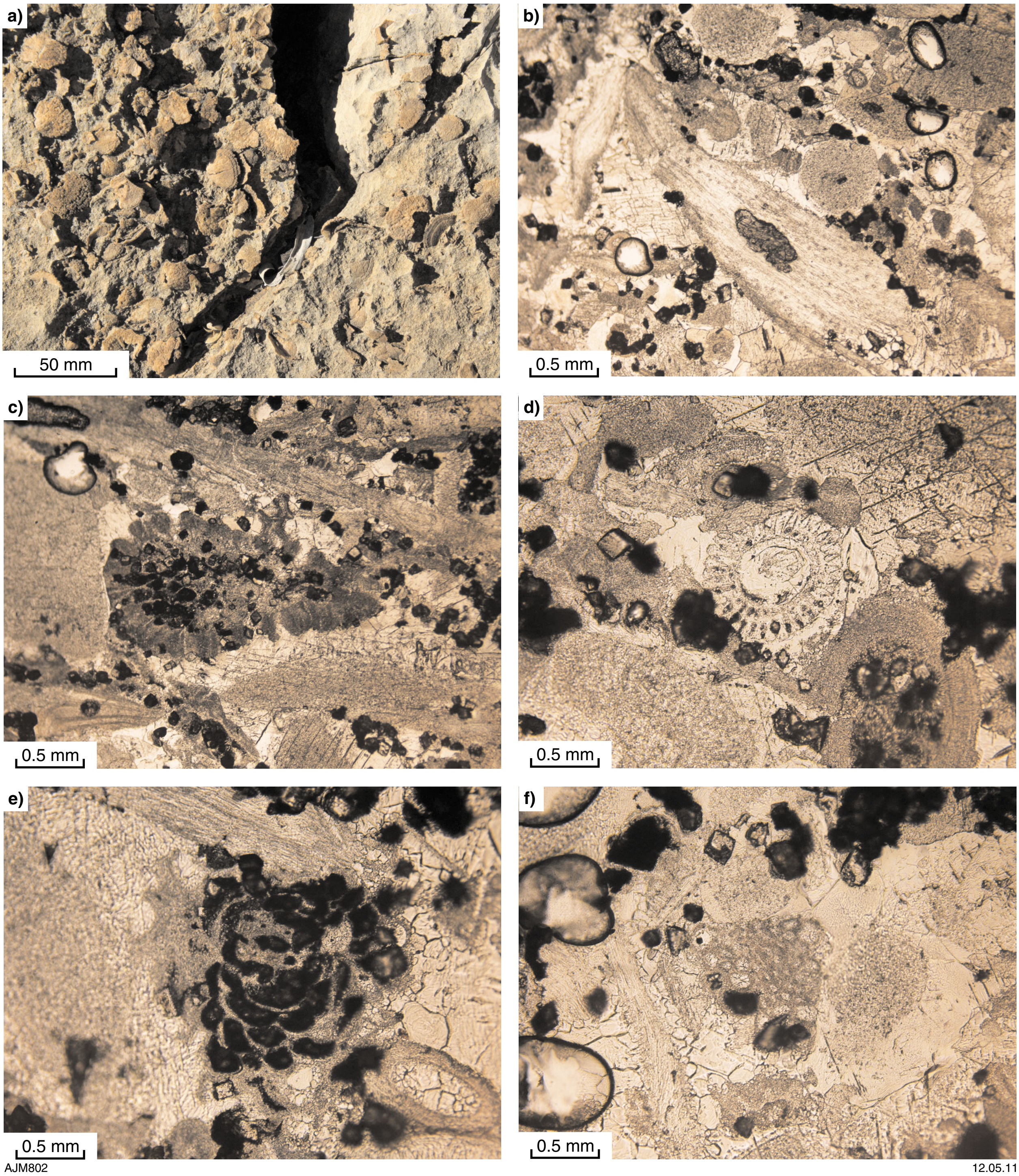
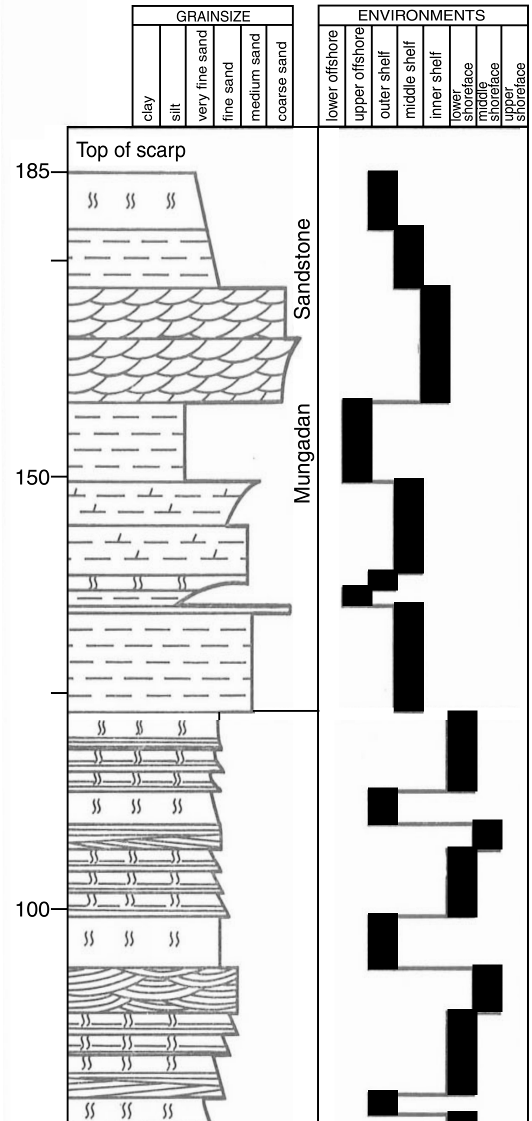


Figure 47. Outcrop and acetate peel photographs of fossiliferous brachiopod–crinoid grainstone exposed at 24°08'30.1"S 115°15'47.8"E: a) silicified casts of brachiopod valves on bedding surface; b) brachiopod debris (partly replaced by silica), crinoid columnals and scattered dolomite crystals; c) bryozoan colony, crinoidal columnals and scattered dolomite; d) echinoid spine, crinoidal columnals, scattered dolomite; e) and f) calcareous algae probably belonging to the *Wetheredella* Group (Riding, 1991) variously referred to encrusting calcified cyanobacteria, green algae, or dasycladalean algae. Crinoidal and brachiopod debris and scattered dolomite are also present. Acetate peels viewed in transmitted light

Figure 48. (facing page) Measured section of the Coolkilya and Mungadan Sandstones, Range Bore, east side of Kennedy Range about 5 km south of the picnic area. Densely patterned troughs are hummocky and swaley cross-beds, open troughs are normal trough cross-beds. From Moore et al. (1980b, fig. 3). See Figure 25 for explanation of facies.



Elsewhere in the Kennedy Group, Lever (2002, 2004a,b) described cyclicity at various scales observed in outcrop sections from around the Kennedy Range. She concluded that the presence of regular cycles of similar thicknesses along the range did not support a tectonic or autocyclic origin but suggested fluctuating eustatic sea levels with a possible contribution from changes in freshwater influx.

Skwarko (1993) listed foraminifera, bryozoans, brachiopods, bivalves, ammonoids, gastropods, conulariids, trilobites, crinoids, and trace fossils from the Coolkilya Sandstone, and a much reduced fauna of indeterminate bivalves and serpulids together with abundant trace fossils and problematica from the Mungadan Sandstone. The ammonoids *Agathiceras applanatum* (Teichert), *Daubichites goochi* (Teichert) and *Popanoceras* sp. recorded from the Coolkilya Sandstone (Teichert, 1942, 1944; Glenister and Furnish, 1961; Cockburn, 1980) suggest a Roadian age (Leonova, 1998). Conodonts of an unnamed species of *Vjalovognathus*, similar to those found in the underlying Wandagee Formation, were recorded by Nicoll and Metcalfe (1998), but offer no definitive age information.

Locality 13: Lake Thetis

Significance: Holocene stromatolites and microbial mats along the margin of a small lake formed in an interdunal depression.

Location: About 1.2 km southeast of Cervantes post office, 30°30'27"S 115°04'48"E, Wedge Island. The best time to see the stromatolites is in late summer and autumn when the level of the lake is lowest. The site is signposted as a local tourist attraction. The Department of Conservation and Environment does not allow the collection of samples without a permit at this locality.

Geology: Lake Thetis lies about 1.5 km inland from the present shoreline next to the small town of Cervantes, and occupies a collapsed doline at the northern end of an interdunal depression in the Holocene Quindalup Dune System (Grey et al., 1990; Grey and Planavsky, 2009). It is a small lake that contains permanent water to a maximum water depth of 2.25 m, although the water level shows seasonal variation in the order of 50 cm. There is no substantial surface drainage into the lake, which is apparently fed by direct rainfall and ground water. According to Grey et al. (1990), salinity readings vary seasonally from about 39 to 53 g/L, mainly as a result of high evaporation rates during summer months, and there is no evidence for a subterranean connection to the sea. Stromatolites are forming in this lake through microcrystalline carbonate precipitation mainly within cyanobacterial *Entophysalis* biofilms. The developing stromatolitic structures are crudely laminated and some exhibit digitate columnar branching. Grey et al. (1990) considered the age of the interdunal depression containing Lake Thetis to be about 3–4 ka, based mainly on a bivalve assemblage in a coquina exposed in the quarry adjacent to the northern edge of the lake. Carbon-14 dating of bivalves from this locality indicates an age of 5600 ± 260 years BP (Mory, 1995). The assemblage is similar to that in the middle Holocene strata of Rottnest Island (Playford, 1988).

Table 4. Physical and chemical factors at Lake Thetis (from Grey et al., 1990)

Mean rainfall	May to September	390 mm
	September to April	170 mm
Annual evaporation		1700 mm
Mean annual maximum temperature		24.6°C
Mean maximum temperature (February)		>30°C
Maximum temperature variation		<10°C to >37°C
Summer winds		southwesterly
Winter gales		northwesterly
Salinity		39–53 gL^{-1}
Alkalinity (carbonate plus bicarbonate)		0.5% meq L^{-1}
pH		8.28–8.6
Maximum water depth		2.25 m

Physical and chemical factors affecting the lake are listed in Table 4. As mapped by Grey et al. (1990), the substrate of the lake and adjacent foreshore is zoned in a concentric fashion based on different microbial mat types (Fig. 49a). Five types have been recognized (Table 5), and three of these may be visible from the shore. The crenulate mat can be seen in the seasonally flooded high foreshore areas where it grows in a reticulate pattern of ridges and blisters a few centimetres in diameter due to periods of desiccation. During February, at the height of summer, the mat will be desiccated and extremely friable. Nodular mat is best seen in the splash zones around the edges of stromatolite domes along the southwestern shoreline (Fig. 49b) where it is formed of aggregations of nodules (0.5–10 cm diameter) on the lower surfaces of the stromatolites. Diatomaceous mat forms an orange-brown gelatinous band in the shallows, commonly just below or coating the nodular mat.

The distribution of the diatom mat is probably tightly constrained by light penetration because it is nearly always about 25 cm below the surface and it migrates as lake levels change to maintain this position. The lithified surfaces of many of the stromatolites contain abundant diatom frustules. The floor of the lake beneath the flocculent mat is composed of fine-grained carbonate mud, with shell fragments, aragonite, and red-purple organic material composed mainly of purple sulfur bacteria (Grey et al., 1990). Silica is also being deposited inorganically as light brown organic particles containing traces of calcium, sulfur, and chlorine (MW Pryce, as cited by Grey et al., 1990). The purple mud of the lake floor includes irregular sandy laminae, and fine sand-sized irregular carbonate micronodules (Grey et al., 1990).

The margins of the lake have terraces of lithified carbonates and associated unconsolidated sediment. Grey et al. (1990) recognized three terraces of coalesced, planed and domal stromatolites. Fractured and weathered domes reveal that the centres of many of the stromatolites show a pattern of crude concentric upwardly convex laminae, and most include internal morphological variation. In some, there is a thrombolitic core (without layers) with an outer layer (up to 15 cm thick) of digitate branching columns. Grey et al. (1990) noted that branching seems to be confined to areas of low wave activity. Fenestrae

Table 5. Mat types in Lake Thetis (after Grey et al., 1990)

Mat type	*Location and substrate*	*Gross morphology and colour*	*Microbial community*
Crenulate mat	above high waterlevel on coarse, calcareous sand; position varies seasonally	reticulate ridges and blisters on surface producing alternating layers of organic rich sand and mud; black to olive green	predominantly filamentous but with some coccoid cyanobacteria. Genera include *Calothrix*, *Scytonem*a, *Gloeocapsa*
Nodular mat	littoral to mid foreshore zone on lithified stromatolite domes and reef; changes with water line position	on SW shore mat forms noduls in clusters with an irregular surface and abundant mucilage; on N shore mat is patchier with less surface relief; produces indistinct laminations; black to grey	coccoid cyanobacteria; including *Gloeocapsa*
Filamentous mat	low marginal shelf that is permanently submerged; on lithified plates and angular fragments; in cracks, on underside of plates and as a fragile benthic mat over flocculent mat with very little seasonal change	film and/or fragile coating; produces no laminations; bright green	filamentous cyanobacteria; including *Oscillatoria*
Diatomaceous mat	marginal shelf, at water depths below 1.5–2 m, on lithified stromatolites and plates; changes position with change in water depth in lake	mucilaginous coating; produces no laminations; beige/brown	diatoms
Flocculent mat	permanently submerged in centre of lake; surface approximates oxic/anoxic interface; sometimes is dispersed throughout water column and concentrated at water's edge	gently undulating mat up to about 50 cm thick, with distinct sediment/water interface where surface undisturbed; produces no laminations; purple/pink with blue-green patches on surface	filamentous and coccoid cyanobacteria, diatoms, purple sulfur bacteria; genera include *Oscillatoria*, ?*Synechocystis,* ?*Thiocystis*/*Thiocapsa*

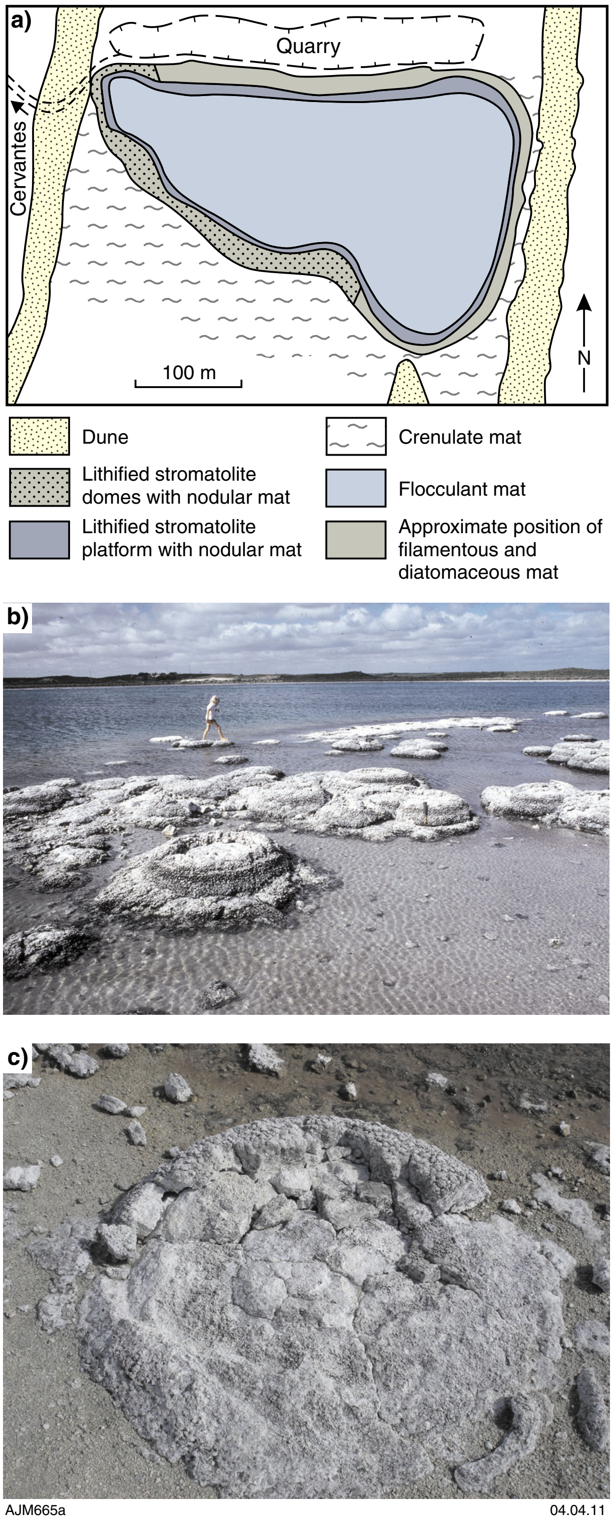


Figure 49. Lake Thetis: a) sketch map showing distribution of mat types (after Grey et al., 1990); b) Holocene stromatolites; c) close up of a stromatolite showing columns

(about 1 mm high and 10 mm or more in length) and larger elongate cavities may develop between the laminae.

Grey et al. (1990) noted that the lithified carbonate platform extends up to 10 m into the lake where there is an abrupt slope to the unconsolidated floor of the central part of the lake (at 2 – 2.5 m water depth). The surface of the platform consists of a crust (1–5 cm thick) including a massive white papillate to botryoidal surface layer (0.1 – 1.0 cm) and a fenestral cream lower layer (0.5 – 4.0 cm) with the basal section commonly coloured green due to associated micro-organisms.

Reitner et al. (1996) and Arp et al. (2001) discussed the method of calcification of the Lake Thetis stromatolites. According to Reitner et al. (1996): 'The recent growth results mainly from calcifying *Entophysalis* films which are forming a more or less laminated crust. Within the deeper parts of the *Entophysalis*-biofilm the outer basophilic polysaccharide envelopes contain abundant heterotrophic bacteria. Calcification events exactly start at these points. The older, subfossil portions of the microbialites are characterized by plumosely arranged *Scytonema*-filaments which are enclosed by fibrous aragonite. Within small cryptic primary and secondary cavities clearly laminated organomicrites are lining cavity walls. The formation of this type of 'microstromatolites' is related to organic films, which contain no active microbes. These organic films are composed of degraded organic material (polysaccharides, proteins etc.) acting as matrices and templates for nucleation and growth of organomicrites and fibrous aragonite crystals.'

Locality 14: The Pinnacles, Nambung National Park

Significance: This locality has limestone pinnacles and rootlets in Pleistocene Tamala Limestone formed by reprecipitation of calcium carbonate around tap roots.

Location: About 15 km south-southeast of Cervantes, 30°36'18"S 115°09'21"E, Wedge Island. The Department of Conservation and Environment does not allow samples to be collected without a permit in Nambung National Park.

Geology: The Pinnacles developed from deep differential weathering on the surface of the Tamala Limestone (McNamara, 1986). Weathering apparently took place preferentially along fissures in the limestone and residual columns of rock became covered by a residue of unconsolidated quartz sand. In the pinnacles desert, much of the residual sand has been blown clear of the limestone columns by persistent winds (Fig. 50a) commonly exposing abundant calcified fossil rhizoliths (plant roots; Fig. 50b). Some residual sands with fossil soil horizons are present nearby (Fig. 50c). The following model for the development of pinnacles is from McNamara (1986):

1. Large taproots penetrated the eolian dune deposits while they were stabilized by vegetation and lithification. Dissolution and reprecipitation of calcium carbonate around the taproots alternated between wet winters and dry summers, thereby preferentially lithifying these areas.

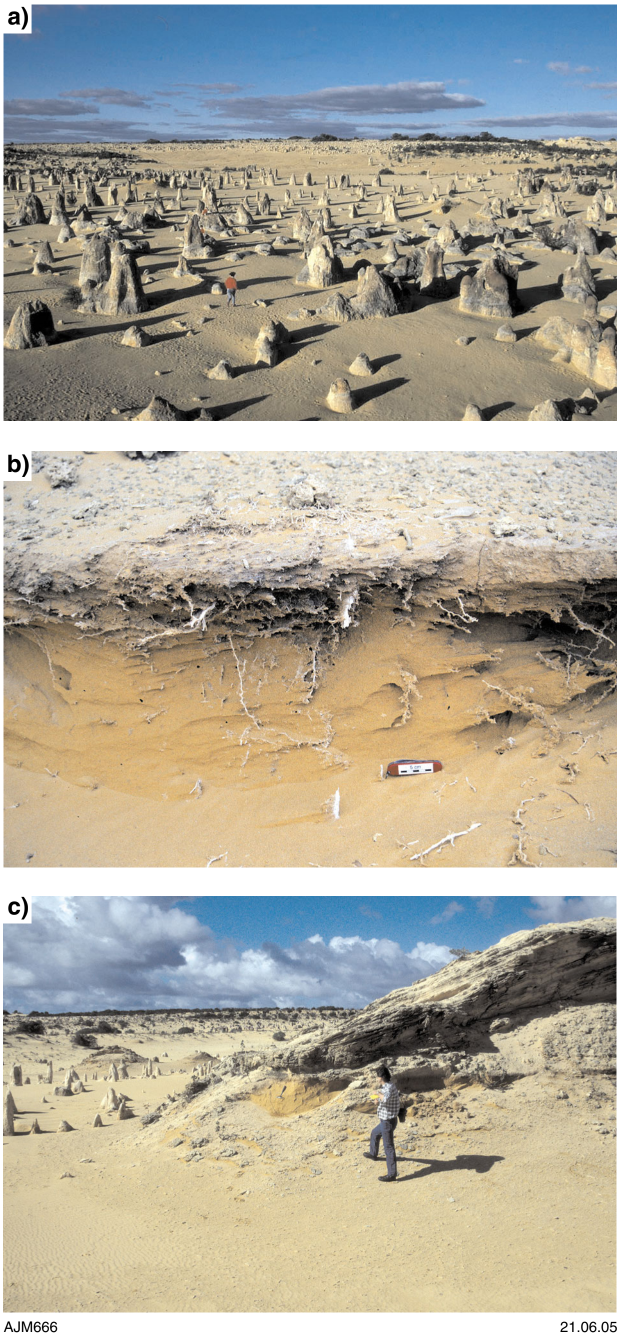


Figure 50. Pinnacles Desert: a) general view of pinnacles; b) fossil rhizoliths; c) silica and carbonate sand separated by a fossil soil horizon

2. A subsoil calcrete developed at the base of the thin humic layer on the surface of the dunes.

3. Cracking of the subsoil calcrete allowed preferential leaching of the underlying friable limestone by surface waters. After prolonged weathering only limestone pinnacles (originally lithified around tap roots) remain surrounded by residual quartz sand from the dune deposit.

Aboriginal artefacts (including flakes of chert) have been found in blowout depressions around the pinnacles and, in one instance, cemented to a pinnacle. Foraminifera in the chert indicate that the flakes are from an Eocene unit believed to lie on the now submerged continental shelf (Glover, 1975; Quilty, 1978). No exposures of this Eocene facies are known onshore in the Perth Basin.

Active coastal dunes of the Quindalup Dune System border the beach along the road into the Pinnacles. A 31.7 × 23 cm egg of the large, flightless, now extinct, Madagascan Elephant Bird (*Aepyornis maximus*) was found in 1992 buried in one of these Holocene dunes, and was dated at about 2000 years BP (Long et al., 1998). It is thought that the egg drifted on ocean currents from Madagascar rather than being transported by human intervention. In a discussion of alien vegetation, Rippey and Rowland (1995) mentioned a South African study in which drift cards took around 18 months to reach Western Australia from South Africa.

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Appendix

Locality access details

The following details are provided to clarify where permission to access private land or national parks needs to be sought beforehand. It is advisable to avoid the parts of the stations where mustering is underway. Stations may refuse access during wet periods to avoid damage to their tracks. Please leave all gates and fences as you find them, unless instructed otherwise.

A permit is required to collect samples from Department of Environment and Conservation (DEC) lands (contact the Licensing Officer, DEC, 17 Dick Perry Ave, Kensington, WA 6151, ph. 08 9334 0333).

For localities 3–11, the nearest fuel available is from the Minilya roadhouse, on the highway 5 km north of the Lyndon Road. Also, fuel may not be available at Gascoyne Junction.

1. **Irwin River Coalseam Conservation Park** is a small national park, which is well signposted from Mingenew (33 km south) and Mullewa (51 km north). The park is open to the pubic all year and there is no charge for entry. Within the park, High Cliff and Fossil Cliff (400 m upstream on opposite bank) are best accessed via the 'River Bend' area. The North Branch is upstream from 'Fossil picnic area' north of 'Irwin Lookout' (at the top of High Cliff), and the South Branch is next to the 'Miners Camp'. Several hours should be set aside to examine these exposures. Barbeques and toilets are available at 'River Bend' and 'Miners Camp', and there are also information panels at 'River Bend'. Camping is permitted at 'Miners Camp', and a ranger is in attendance during the winter months. The best time to visit this area is in winter or early spring because salt otherwise encrusts much of the sandstone facies along the river. The area is registered with the National Estate (Place ID: 9683).

2. **Carbla Point** lies 11 km southwest of Carbla homestead within Hamelin Pool Marine Nature Reserve. The nature reserve is managed by DEC, and all access must be arranged through them to ensure protection of the stromatolite and algal mat structures. Contact the DEC District Manager, 89 Knight Terrace, Denham 6537 or phone 08 99481 208, fax 99481024, email: sharkbayenquiries@dec.wa.gov.au. Note that DEC provides a staff member to accompany any group accessing the site, for which there may be a charge. Usually this charge is waived for scientific groups with no commercial interests. If a DEC officer is not available, they may delegate an honorary officer from Carbla Station for whom there will be a charge. Access is normally not granted to individuals because public access is available to view the stromatolites from the boardwalk adjacent to Hamelin Pool Caravan Park and Telegraph Station. Note that it is also necessary to contact Carbla Station beforehand, and that there is a small charge per vehicle for use of their tracks (present contact details: Carbla, PMB 45, Carnarvon 6701, phone 08 9942 5915 or email <rick@rickfennygroup.com>). There are no facilities at the site and a 4WD vehicle is recommended. The track is closed during wet weather. Nearby basic accommodation is available at the Carbla homestead, Woodleigh (40 km east, off the Byro Road), Overlander Roadhouse (37 km south-southeast by road), Hamelin Station Stay (30 km west of the roadhouse on the Denham Road), or Hamelin Pool Caravan Park and Telegraph Station (34 km west of the roadhouse on the Denham Road). Besides being a World Heritage area, Shark Bay also is registered with the National Estate (Place ID: 105686).

3 and 4. **Coolkilya Pool and the Minilya River** localities lie on Wandagee Station. The homestead is 60 km east of the Great Northern Highway via the Lyndon Road. Contact Graham and Gail Hopkinson, Wandagee Stn c/o PO Carnarvon, WA 6701, ph. 08 9943 0538.

5. **Dead Man's Gully** is on Middalya Station. The homestead is 84 km east of the Great Northern Highway via the Lyndon Road. Contact David Smith, Middalya Stn c/o PO Carnarvon, WA 6701, ph. 08 9943 0542.

6–11. These localities lie on **Williambury Station**. The homestead is 128 km east of the Great Northern Highway via the Lyndon Road. Contact John Percy, Williambury Stn, c/o PO Carnarvon, WA 6701, ph. 08 9943 0541. Note that Moogooree is now part of that station and the homestead is no longer occupied.

12. **Kennedy Range National Park** is administered by DEC. The only facilities are a basic camping ground with bush toilets. Water is not available. There is a small charge for staying overnight, but no entry charge. The nearest other designated campsite is at Gascoyne Junction, 60 km to the south by road. While the park is open to the pubic all year, in summer months temperatures can be high and cyclonic rains may close roads in the region.

13. **Lake Thetis**, a part of Nambung National Park, lies 1 km south-southeast of the small coastal town of Cervantes and is managed by DEC. The access track is on the east side of Hansen Bay Road, 850 m south of the turnoff from Cervantes Road. The best time to see the stromatolites is in late summer and autumn when the level of the lake is lowest. The site is signposted as a local tourist attraction. Visitors are requested to keep to the boardwalk or track around the lake. There are no other facilities at the lake.

14. **The Pinnacles, Nambung National Park**, lie 15 km south-southwest of Cervantes on Pinnacles Drive in Nambung National Park. Note that there is an entry charge to the park, but that it is accessible by two-wheel drive vehicles. Barbecues, information panels, tables and toilet facilities are provided by DEC, but camping is not permitted. Accommodation in nearby Cervantes during school holidays may be limited. The site is registered with the National Estate (Place ID: 10201).

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PERMIAN–CARBONIFEROUS GEOLOGY OF THE NORTHERN PERTH AND SOUTHERN CARNARVON BASINS, WESTERN AUSTRALIA — A FIELD GUIDE

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