

The geology of South Australia



Volume 2

The Phanerozoic

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Volume 2 The Phanerozoic

Edited by J.F. Drexel, W.V. Preiss

Issued under the authority of

The Hon. D.S. Baker M.P.
Minister for Mines and Energy

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PREFACE

The geology of South Australia — Volume 2, The Phanerozoic complements *Volume 1, The Precambrian*, released in December 1993. As with its companion, this volume is designed as a foundation upon which knowledge of South Australia's geology and mineral resources can be built, and exploration models for petroleum and minerals devised.

This volume describes the rocks that host the State's petroleum resources, and contain major economic mineral commodities such as gypsum, lime sand, silica sand and heavy-mineral concentrations. Groundwater, one of the State's vital resources, is discussed in terms of its geological context. Environmental geology and mining heritage, two relatively new terms to the community, are also introduced in this volume.

Included in the text are some of the interpretations derived from the SA Exploration Initiative, for which the State Government has now committed \$18.5 million towards aeromagnetic and radiometric surveys, data processing and interpretation, and drilling. Although the aeromagnetics are designed primarily for mineral exploration, the data are now being used in South Australia to assist in delineating likely petroleum prospects.

In preparing this volume, the Department recognises the contributions by numerous geoscientific and support staff within the Department of Mines and Energy, industry and tertiary institutions. It is regretted that Dr Reg Sprigg and Dr Nell Ludbrook, who contributed so much to the geosciences, particularly within South Australia, passed away recently. Reg was instrumental in the formation of Santos and ultimately in the discovery of Australia's largest onshore petroleum resource in the Cooper Basin. Nell, amongst her many achievements, pioneered the use of palaeontology in petroleum, mineral and groundwater exploration in South Australia.

This volume, along with its companion, provides an excellent overview of the State's geology. The volumes are major achievements by MESA, and continue the now century old tradition of promoting the geosciences and fostering mineral and petroleum development in South Australia.

ANDREW J. ANDREJEWSKIS
CHIEF EXECUTIVE
MINES AND ENERGY SOUTH AUSTRALIA

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ABSTRACT

The known Phanerozoic record of South Australia is preserved in widespread sedimentary basins, and igneous and metamorphic rocks; only the Silurian Period is not represented.

Early Cambrian marine, carbonate-dominated sediments, including algal, archaeocyathan and tabulate coral reefs, were deposited as three eustatically controlled sequences in the Arrowie and Stansbury Basins after an hiatus in sedimentation in the Adelaide Geosyncline at the end of the Neoproterozoic. In the late Early Cambrian, the Kangarooian Movements led to local uplift, and to renewed rifting, accompanied by mafic volcanism, in the eastern Stansbury Basin to form the Kanmantoo Trough. This was rapidly filled with thick, mainly siliciclastic marine sediments. In the Arrowie Basin, marginal marine redbed clastics were deposited, with one major marine incursion, and sedimentation continued into the Middle Cambrian.

In the Officer Basin, early Palaeozoic strata were deposited in epeiric sea and non-marine, partly evaporitic environments. To the northeast (western Warburton Basin), mainly marine conditions prevailed. In the northern Officer Basin, thick Neoproterozoic and early Palaeozoic sediments of the Munyarai Trough and Birksgate Sub-basin were deformed and overthrust by crustal shortening mainly during the mid-Palaeozoic Alice Springs Orogeny. In the eastern Warburton Basin, Cambro-Ordovician marine to deltaic deposition was more or less continuous. Thrusting and folding are related to the Alice Springs Orogeny and were followed by mid-Carboniferous granite intrusion.

In the Adelaide Geosyncline, mostly greenschist facies metamorphism and compressional deformation of the Cambro-Ordovician Delamerian Orogeny commenced in the Fleurieu Arc, with northwest-directed thrusts and folds and cleavage development. In the eastern Mount Lofty Ranges, the early structures were refolded by more upright, approximately meridional folds, accompanied by amphibolite facies metamorphism. These are continuous with arcuate, partly transpressional folds of the Nackara Arc, with axes swinging from north-south to east-northeast. The fold arc is separated from a similar arc in the North Flinders Zone by the less deformed Central Flinders Zone, with domes and basins due to fold interference. Diapirs that had risen during Adelaidean sedimentation were reactivated. Syntectonic granitoids intruded in the eastern Fleurieu and Nackara Arcs. Migmatite developed only within the Kanmantoo Trough. Undeformed felsic and mafic intrusives of Early Ordovician age, mostly near the western margin of the Murray Basin, are post-tectonic. Stratabound gold and base metals were concentrated during orogeny, and there is mineral potential associated with Cambrian volcanics and Delamerian intrusives.

Late Palaeozoic sediments are preserved in basins controlled by northeast and northwest-trending structures, glacially deepened valleys, and as thin mantles over bedrock. Australia was still part of Gondwana and lay near the South Pole. Widespread glaciation affected virtually all of South Australia in the Late Carboniferous and Early Permian. Ice moved north-northwest across the State, scouring bedrock, overdeepening valleys and structural troughs, and plastering lodgement till over parts of the landsurface.

Deglaciation, commencing in the Asselian and complete by the Sakmarian, was accompanied by marine transgression into tectonic basins and isostatically depressed lowlands marginal to the icesheet. Meltwaters deposited considerable thicknesses of complex ice-contact, fluvio-glacial and glaciolacustrine clastics. Glaciomarine sediments are preserved in the Denman, Arkaringa, Troubridge and Nadda Basins, and on the Padthaway Ridge.

Regression, probably due to isostatic recovery, terminated deposition in the southern basins but non-marine sedimentation continued in the north. With continuing climatic amelioration in the Early Permian, extensive peat swamps drained by rivers developed in the Arkaringa, Pedirka and Cooper Basins. In the latter they formed coally source rocks for a major oil and gas resource. Late Permian fluvial and flood-basin lake deposits are also gas-bearing in the Cooper Basin but are unknown elsewhere.

In the Mesozoic, a time of profound palaeogeographic change, global sea-level rise brought the last great marine transgression onto the Australian continent from the north. To the south, crustal extension and plate divergence created the Australian Southern Rift System, the

separation of Australia from Antarctica, and the concomitant seaway south of the Australian continent.

In the Triassic, subaerial weathering and pedogenesis were widespread. Low-energy terrigenous clastics were deposited locally, including coal swamp deposits, of which Leigh Creek Coalfield provides a major energy resource.

During the Jurassic, crustal instability increased. Epeirogenic downwarp in the north was accompanied by deposition of a very widespread fluvial and lacustrine clastic blanket. This forms the aquifer for the huge Great Artesian Basin water resource, and is a major hydrocarbon reservoir. In the south, the early stages of extension and rifting caused the extrusion of basalt and emplacement of kimberlitic rocks, the latter a target for diamond exploration. Thick terrigenous (including coal swamp) deposits also began accumulating in local rift valleys.

A major Early Cretaceous epicontinental marine transgression flooded the downwarped region in the north, depositing a blanket of fossiliferous mudstone and sandstone over the older non-marine clastics. The mudstones form the regional artesian aquiclude and locally host precious opal. In the south, conditions were still non-marine; thick terrigenous and volcanogenic sediments accumulated along the continental rift.

By the mid- to late Cretaceous, the retreating sea, and resumption of terrestrial conditions with low-energy fluvial to lacustrine sediments, marked the end of widespread deposition in the north. To the south, a marginal marine influence appeared in the developing rift system, where thick, fine to coarse clastics continued to accumulate as Australia and Antarctica separated. In the South-East these sediments, together with the older rift deposits, are now commercial gas producers.

Tertiary strata are widespread in South Australia, in gently downwarped cratonic basins, palaeochannels and as thin sheets in the interior. Thicker successions accumulated in passive continental margin basins, where they unconformably overlie the Mesozoic rift basins. Relatively thin, fluvial and lacustrine, carbonaceous and arenaceous sediments characterise Palaeogene inland depocentres, whereas thin argillaceous and carbonate mudstones formed in Neogene lakes. These continental sediments were deposited on a subdued landsurface, with uplands as presently located. With continental separation, Eocene marine transgressions into the southern basins deposited marginal marine clastics and extensive, temperate-water, platform to deep water carbonates.

The Tertiary climate changed from high-rainfall temperate to cool temperate in the Early Paleocene to warmer in the Late Paleocene and Early Eocene, but still with rainforest. Aridity increased in the late Tertiary, with open woodland and minor forest pockets. Prolonged weathering of a generally subdued landscape which began in the Mesozoic continued into the Tertiary. Ferricrete formed in the Early Mesozoic-Middle Eocene, mid-Tertiary and late Tertiary, and silcrete during the Late Eocene-Middle Miocene and Late Miocene-Pleistocene. Tertiary sediments and weathered rocks contain important groundwater, coal, construction material, sedimentary mineral and opal resources.

The Quaternary is characterised by marked climatic and sea-level oscillations, impacting on many geomorphological features. Barrier shorelines in the South-East represent successive Pleistocene sea-level highstands preserved by coastal plain uplift over the past 800 000 years. In the interior, cold arid phases (times of dune building and evaporative playa deposition) alternated with periods of moister climate, increased stream discharge and greatly extended lake systems. Subaerial weathering has formed carbonate palaeosols and a variety of karst features.

The Holocene sea transgressed the continental shelf, reaching its present level about 7 000 years ago. The rising sea was followed by variable coastal neotectonic warping, resulting in geographically variable apparent sea-level regression over the past 6 000 years. Holocene and Pleistocene shoreline deposits are used as datums from which neotectonic deformation is calculated. In the Port Adelaide estuary, significant land subsidence has been recognised as a result of human modification of the local environment.

Geological hazards in South Australia are minor. Quaternary volcanism is restricted (to the lower South-East) and seismic activity is relatively low.

INTRODUCTION

The geology of South Australia represents, in two volumes, the results of over 150 years of geoscientific investigations of South Australia. Whereas Volume 1 (Drexel *et al.*, 1993) is concerned primarily with the Precambrian geology of the State, Volume 2, The Phanerozoic, deals with the last 540 million years of the Earth's 4.5 billion year history. Aspects relevant to both volumes and summarised in Chapter 1 (Vol. 1) include: purpose and scope, economic overview, and an historical overview of geological mapping. Special topics treated in more detail are also to be found in Chapter 1: the geological time scale (Fig. 1.1), geochronology, sequence stratigraphy, nomenclature of mapping units, tectonic province nomenclature, and structural nomenclature. Of these topics, two are of special interest to Chapter 7 (this volume): sequence stratigraphy is central to the description of the early Palaeozoic history of South Australia, while the principles of structural nomenclature are applied in the discussion of the Delamerian Orogeny. Chapter 2 (Vol. 1) on the geological framework provides definitions and brief descriptions of the major tectonic provinces of the State, including sedimentary basins of both Precambrian and Phanerozoic age; these are defined and subdivided into their tectonic elements in Table 2.1. Chapter 2 also presents seismic sections of the Officer Basin, Kanmantoo Trough, Cooper Basin, and the offshore portion of the Otway Basin, which are relevant to Chapters 7, 8, 9 and 10 of this volume.

Volume 2, on the Phanerozoic evolution of South Australia, is divided into five chapters based on age: early and middle Palaeozoic, late Palaeozoic, Mesozoic, Tertiary and Quaternary. The geological record of these times in Earth history is of considerable scientific as well as economic and practical importance:

- Cambrian sediments in the Arrowie, Stansbury, Warburton and Officer Basins include known and potential hydrocarbon source rocks and reservoir rocks, as well as Mississippi Valley-type base-metal mineralisation (Ch. 7).
- Early Cambrian reefs of the Arrowie Basin contain the world's oldest tabulate-like corals (Ch. 7).
- Hydrocarbons found in drillcore of Cambrian evaporitic playa-lake sediments first drew attention to the oil and gas potential of the Officer Basin, which was then enhanced by later recognition of additional source rocks in the Neoproterozoic (Ch. 7).
- Stratigraphically controlled base-metal deposits occur in the Cambrian Kanmantoo Trough of the eastern Mount Lofty Ranges, while extensive volcanics of possible Cambrian age under the Murray Basin offer further exploration targets (Ch. 7).
- The Cambro-Ordovician Delamerian Orogeny (Ch. 7) further concentrated synsedimentary mineralisation, while some copper and gold deposits are associated with Delamerian igneous intrusives. Granitic and mafic intrusives supply valuable building and ornamental stones.
- Late Palaeozoic basins contain extensive, locally well-exposed Gondwanan glacial deposits (Ch. 8).
- The Permian Cooper Basin supplies natural gas to Adelaide, South Australian regional, Sydney, Canberra and New South Wales industrial markets. Oil and gas liquids from the Cooper and Mesozoic Eromanga Basins are piped to Port Bonython on Spencer Gulf for processing and export. In the near future, ethane will be piped from Moomba to a petrochemical plant in Sydney (Ch. 8).
- Other Permian and Mesozoic basins contain important coal deposits, of which Leigh Creek supplies the Northern Power Station at Port Augusta with fuel to produce ~45% of the State's electricity requirements (Ch. 9).
- Mesozoic basins are largely non-marine but with significant marine transgressions, and occupy contrasting intracratonic and rift settings. The intracratonic Eromanga Basin (Ch. 9) contains an immense resource of underground water, much of it artesian, which exits near the basin margins as spectacular mound springs. Eromanga Basin sediments are a major host to deposits of precious opal, South Australia's gemstone emblem.
- The Mesozoic basins of the Australian Southern Rift System (Ch. 9) contain extremely thick sedimentary successions with significant hydrocarbon potential. Of these, the Otway Basin is already a natural gas and carbon dioxide producer.
- Tertiary basins (Ch. 10) cover large areas of the State and contain numerous lignite deposits, underground water and construction materials. They also have significant potential for sedimentary minerals and gemstones.
- The Quaternary Period is the time in which the modern environments of South Australia were shaped. Economic deposits of Quaternary age include gypsum and construction materials. Chapter 11, as well as summarising the Quaternary stratigraphy of the State, discusses environmental geology and related topics: climatic change, sea levels, soils, landslides, karst, volcanic activity and seismic activity, and the potential hazards that these processes present.

Some terms, concepts, place names and tectonic elements cited in Volume 2 are defined or illustrated in Volume 1, and the reader is referred to that text.

The following abbreviations are used throughout this volume:

SADM – SA Department of Mines (1894-December 1977)

SADME – SA Department of Mines and Energy (January 1978-December 1993)

MESA – Mines and Energy South Australia (January 1994-).

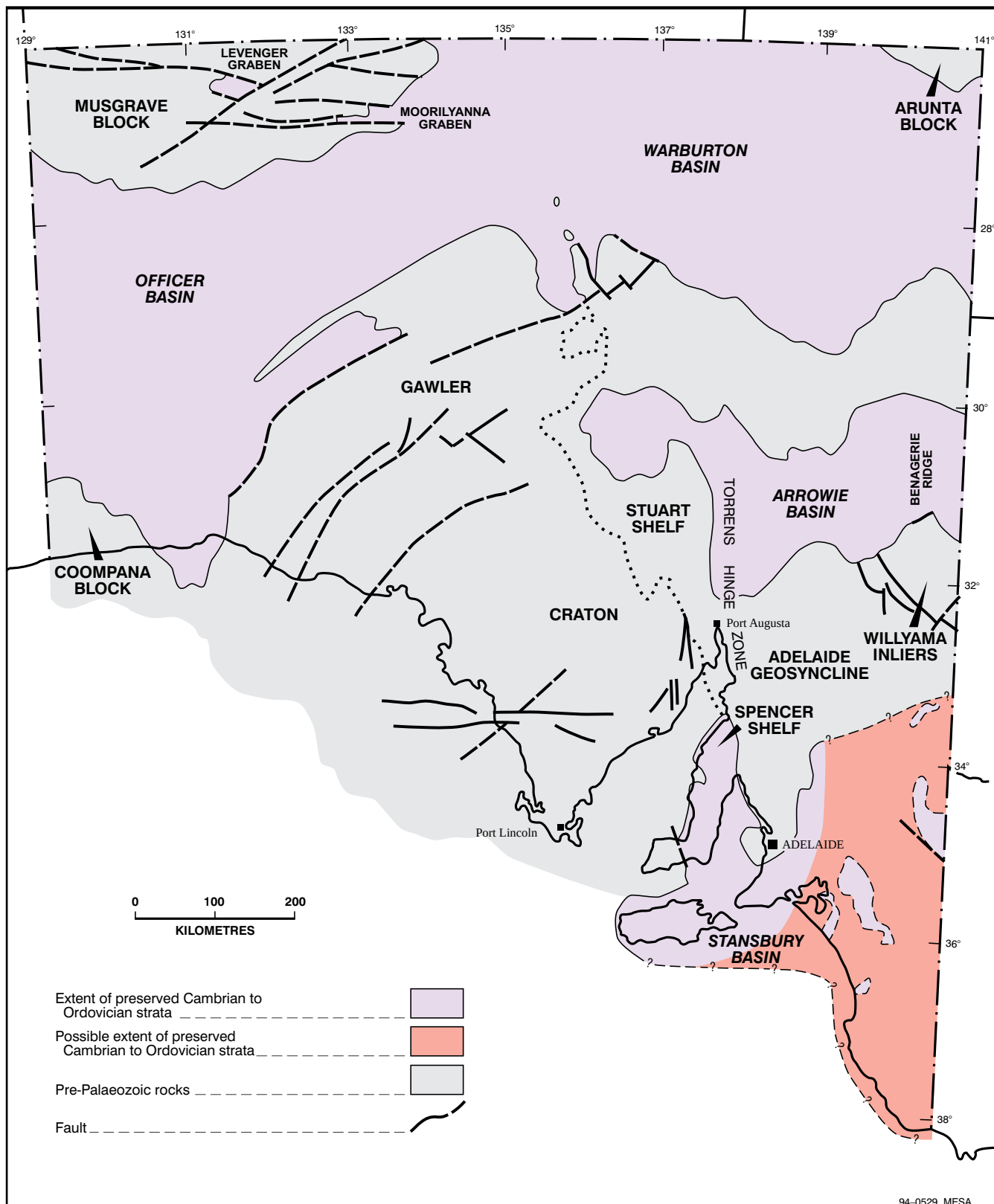


Fig. 7.1 Distribution of Cambro-Ordovician basins in South Australia.

Chapter 7

EARLY AND MIDDLE PALAEOZOIC

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with contributions from N.F. Alley, M.C. Benbow, W.M. Cowley,
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INTRODUCTION

Cambrian and Ordovician sediments were deposited in basins shown on Figure 7.1. These occupied tropical latitudes in two distinct settings. One was situated in present-day eastern South Australia on the palaeo-Pacific margin of Gondwana, while the other setting to the west was in shallow epeiric seaways which extended hundreds of kilometres into the supercontinent. The contrast between these settings is reflected by the nature, distribution and thickness of sedimentary facies which form the basis for discussing the eastern and western regions separately.

The western region embraces the South Australian portion of the Officer and western Warburton Basins. The latter is very poorly known but appears to span the transition from epeiric to more fully marine conditions. The eastern region comprises relatively thin, little-deformed strata on the Stuart and Spencer Shelves, the thicker folded Cambrian in the Arrowie and Stansbury Basins, and the Cambro-Ordovician of the eastern Warburton Basin. Fragmentation of Pangea into smaller continents, including Gondwana, has left its mark on the geology of these basins. Biostratigraphic correlation between the eastern and western regions is poor, but the available data, combined with sequence stratigraphic interpretations, provide a means of integrating Cambrian geology in the context of major depositional sequence sets (supersequences) and higher order sequences. Integration of Ordovician strata into a similar framework is at a preliminary stage.

EARLY PALAEOZOIC OF THE EASTERN REGION

Three major Cambrian depositional sequence sets (€ 1, € 2, € 3) are preserved in the Stansbury and Arrowie Basins. A fourth (€ 4/01), extending into the Ordovician, is suggested for the eastern Warburton Basin and may also include a suite of mafic igneous rocks beneath the Tertiary Murray Basin.

Sequence sets vary in thickness from less than 400 m to more than 1 800 m, and some sets can be subdivided into sequences *sensu* Vail (1987; see Vol. 1, Ch. 1). Each sequence comprises:

- a basal suite of relatively shallow-water sediments (lowstand deposits), conformable to disconformable on the underlying sequence
- an upward-deepening succession (transgressive deposits), sometimes with evidence of slope instability and turbidity current deposition
- deeper water, relatively organic-rich, laminated lime mud or shale (maximum highstand starved sediments)
- either an upward-shallowing succession (late highstand deposits) or a sharp conformable to disconformable (erosional) contact with relatively shallow marine sediments of the overlying sequence.

The transgressive-regressive sediment packages which comprise each sequence are best preserved in shelf margin and deeper basinal settings. Sediments on the shelf proper exhibit rapid facies changes in response to the combined effects of tectonism, eustasy

and sediment supply. Disconformities, which may be frequent on the shelf, diminish or disappear in a basinward direction.

Three sequences, designated € 1.1, € 1.2 and € 1.3 (Fig. 7.2) can be distinguished in sequence set € 1. They are composed of shelf and shelf margin slope or ramp carbonates, shallow-water and deeper water siliciclastics. Systems tracts can be identified in each sequence, and are described for the Arrowie and Stansbury Basins below.

Two sequences (€ 2.1, € 2.2) are recognised locally in sequence set € 2, but there is too little information at this stage to subdivide sets € 3 and € 4/01 meaningfully.

CORRELATION

Sequence sets € 1 and € 2 are Early Cambrian, set € 3 is late Early to late Middle Cambrian, while set € 4/01 ranges from the late Middle Cambrian to Early Ordovician. No formal stage subdivisions have been proposed for the Australian Early Cambrian, thus the Siberian Platform stage scheme of Rozanov and Sokolov (1984) is used here. The Ordian Stage (Öpik, 1968) is considered by Zhang and Jell (1987) to correlate with the Chinese Lungwangmiaoan Stage which is late Early and not Middle Cambrian in age. The Siberian Platform equivalent of the Ordian is the Toyonian Stage (Rozanov and Sokolov, 1984), with which the Ordian is replaced on Figure 7.2.

Daily (1972, 1976a) considered the basal formations of the South Australian Cambrian to correlate with the Tommotian Stage. He had earlier (Daily, 1956) assigned archaeocyaths, trilobites and other shelly fossils in overlying formations to 12 faunal assemblages. Gravestock's (1984) study of archaeocyaths led to subdivision of Faunal Assemblage 2 into upper and lower parts. He correlated Faunal Assemblages 1 and 2 with the middle and upper Atdabanian. Recent work on Yorke Peninsula has shown that the Atdabanian-Botomian boundary may be in the lower part of the Parara Limestone where it is conformable on the Kulpara Formation (Jell *et al.*, 1990; Zhuravlev and Gravestock, 1994). Trilobites and numerous shelly fossils collected from levels corresponding with Daily's assemblages have been described recently (Bengtson *et al.*, 1990). Trilobite zones are particularly useful in deeper marine facies where archaeocyaths are virtually absent (Jell *in* Bengtson *et al.*, 1990), and archaeocyathan zones are most suited to shelf or reef facies (Zhuravlev and Gravestock, 1994).

Middle to late Botomian archaeocyaths occur in the upper parts of the Parara, Wilkawillina and Ajax Limestones, in the Moorowie Formation and lower Oraparinna Shale, (Walter, 1967; Kruse, 1982; Gravestock, 1984; Lafuste *et al.*, 1991; Zhuravlev and Gravestock, 1994). With cessation of hitherto prolonged and widespread carbonate deposition at the top of set € 1 biostratigraphic correlation becomes restricted to a few key trilobite-bearing units in the Arrowie and Stansbury Basins, and to a problematic succession on the north coast of Kangaroo Island. In metamorphosed sections of the eastern Stansbury Basin (including the Kanmantoo Trough: see Vol. 1, p.21), lithostratigraphic correlation is used, which relies on the position within the younger depositional cycles of upward-shallowing or upward-deepening facies suites and on selected marker beds (Gatehouse, Jago and Cooper, 1990).

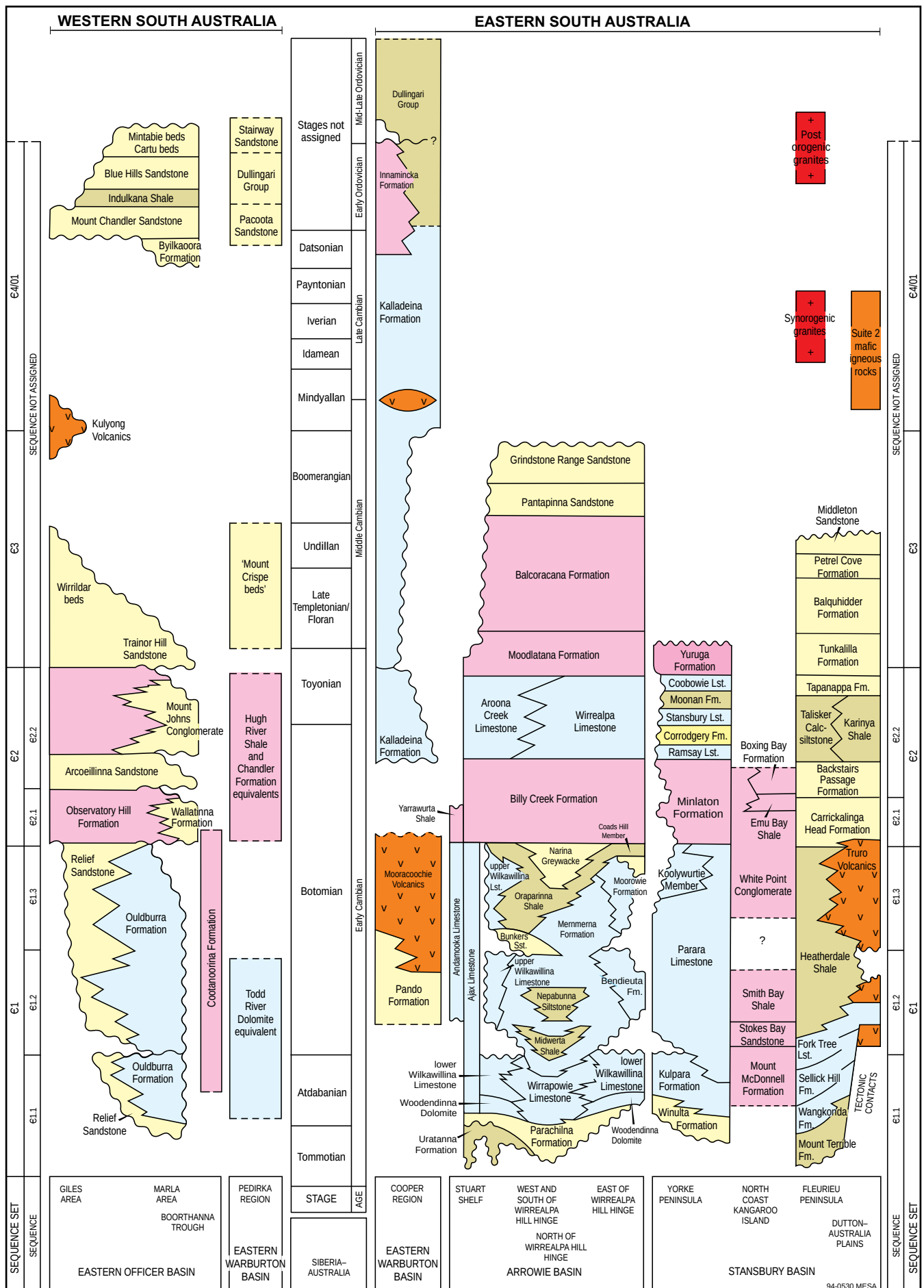


Fig. 7.2 Early Palaeozoic stratigraphy. Australian Middle and Late Cambrian stages after Shergold (1989, 1993).

The lower and upper boundaries of the Toyonian are tentatively placed within the Wirrealpa Limestone and Moodlatana Formation, respectively, based on the presence of the trilobite *Onaraspis rubra* (= *Anadoxides*) in the latter (Jell, 1983a; Jell in Bengtson *et al.*, 1990). Correlation of the Wirrealpa Limestone with the Toyonian is also supported by other skeletal fossils (Kruse, 1991; Brock and Cooper, 1993). A late Middle Cambrian age is reported for uppermost Balcoracana Formation based on a single effaced agnostid trilobite cf. *Leipyge* (Daily and Forbes, 1969; Daily, 1976b). Lack of other diagnostic fossils in these redbeds precludes assignment of Middle Cambrian stages. The latter, where shown on Figure 7.2, refer strictly to the eastern Warburton Basin, where carbonate and shale contain trilobites which permit correlation with Australian Middle and Late Cambrian stages. Ages assigned to the Ordovician strata are based on conodont and acritarch assemblages in the eastern Warburton Basin (Cooper, 1986; Zang, 1992).

STANSBURY BASIN

D.I. Gravestock and C.G. Gatehouse

Geological setting

Cambrian sediments assigned by Wopfner (1972) to the Stansbury Basin (Fig. 7.3) are contained in a triangular region of 10 000 km², bounded by Yorke and Fleurieu Peninsulas and Kangaroo Island. Unmetamorphosed outcrops are confined to coastal Fleurieu Peninsula (Sellick Hill-Carrickalinga Head), the central north coast of Kangaroo Island (Stokes Bay-Point Marsden) and inland along Freestone Creek, and central to northern Yorke Peninsula. The bulk of the sedimentary fill extends at depth beneath Gulf St Vincent.

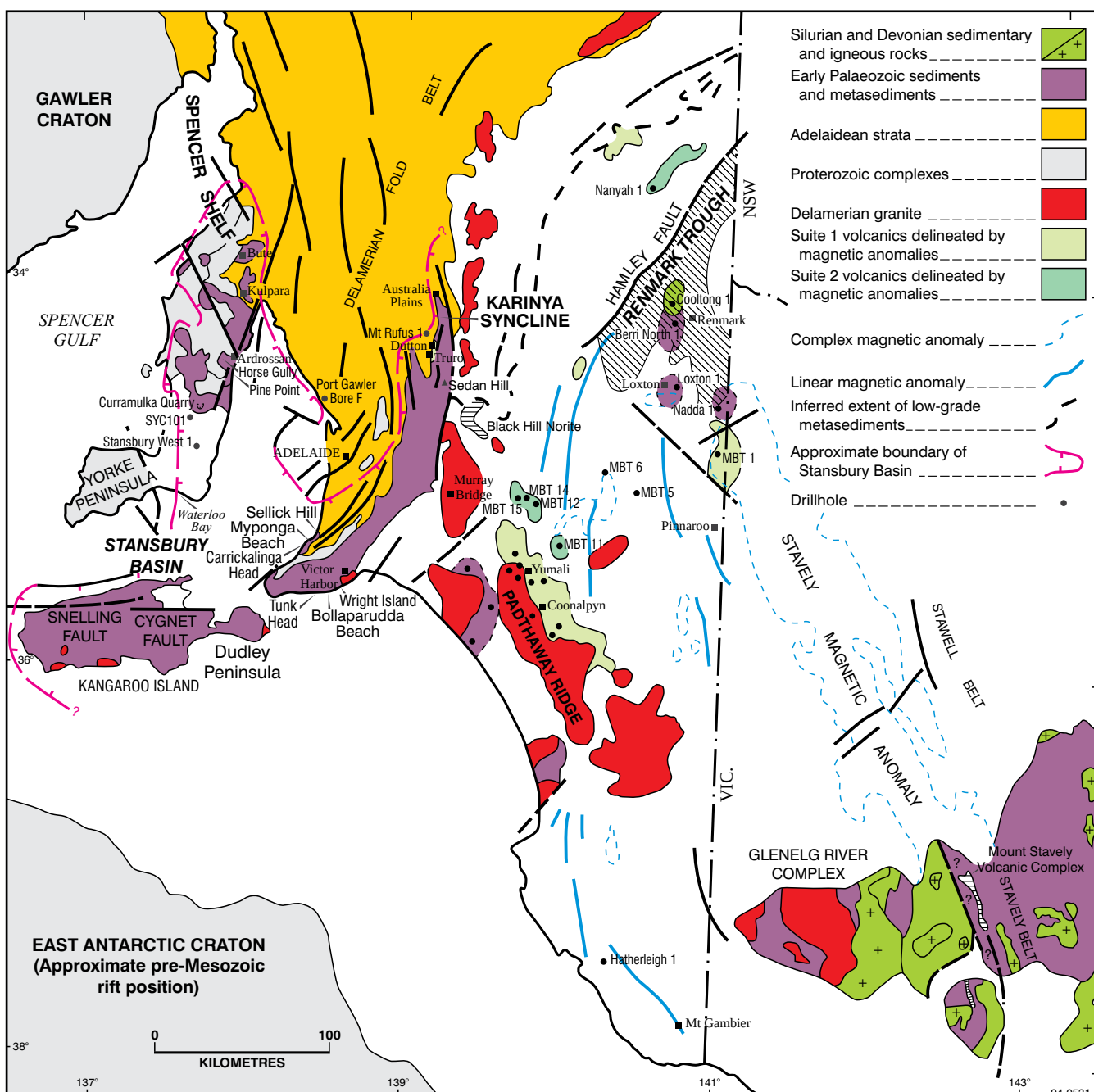


Fig. 7.3 Stansbury Basin location map (Victorian outcrops after VandenBerg and Wilkinson, 1982).

Geophysical surveying and drilling for minerals and petroleum on Yorke Peninsula have outlined subsurface geology from Bute to Edithburgh. Eastern Yorke Peninsula and the immediately adjacent gulf are in a north-south corridor of complex faults comprising the southern Torrens Hinge Zone. Relatively thin, flat-lying to gently folded strata, which onlap the Gawler Craton west of the Torrens Hinge Zone, form part of the Spencer Shelf (Sprigg, 1952a). Remnant Early Cambrian rocks occur on northern Yorke Peninsula but more complete sections are preserved to the south (e.g. 1 330 m in Stansbury West 1).

Marine seismic surveys have shown that Adelaidean and Cambrian strata beneath Gulf St Vincent are flat-lying to gently folded between the Gawler Craton and the Delamerian fold belt (Stuart and von Sanden, 1972), and high-quality data acquired in 1985 and 1992 revealed a thick section beneath southern gulf waters. However, an elevated block of Gawler Craton reaches shallow depth in Investigator Strait and recent geophysical interpretation has indicated that the southern craton margin coincides approximately with the Cygnet and Snelling Faults on Kangaroo Island (Belperio and Flint, 1993).

Cambrian metasediments and volcanics (Normanville and Kanmantoo Groups) on Fleurieu Peninsula and Kangaroo Island were deformed together with the underlying Adelaidean, and form part of the southern Delamerian fold belt (Preiss, 1987). Effects of the Late Cambrian to Ordovician Delamerian Orogeny, responsible for the current disposition of these rocks, are discussed later in the chapter.

The east-west Cygnet and Snelling Faults separate unmetamorphosed Cambrian on the central north coast from metasediments elsewhere on Kangaroo Island. This structure hosts several small gold and lead-zinc deposits.

Folded and faulted rocks on Fleurieu Peninsula have been exploited for small lead-silver deposits in sheared and jointed carbonates (Wright, 1969, 1970, 1972) but other lead, zinc and phosphate anomalies have not proved commercial. Shale and limestone are quarried for road metal. Copper and gold are mined at several stratigraphic levels within the metasediments and in shear zones associated with granitoid intrusives (Parker, 1984); pyrite, phosphate and high-quality building stone are obtained at several localities (Johns, 1988).

The exposed part of the fold belt does not coincide with the full eastern extent of Delamerian folding (Brown *et al.*, 1988), and certainly does not represent the eastern limit of the basin of sedimentation. Sparsely drilled metasediments and mafic igneous rocks to the east beneath the Tertiary Murray Basin are related to those in the Mount Lofty Ranges. Core from MBT 1 near Peebinga, close to the Victorian border, comprises basic volcanics, shale and limestone of Early Cambrian aspect (Rankin *et al.*, 1991a). The original definition of the Stansbury Basin is broadened here to include Cambrian sediments and volcanics in the Mount Lofty Ranges (including Kanmantoo Group of the Kanmantoo Trough) and their presumed equivalents in the subsurface beneath the Murray Basin, and southeast to include the Glenelg River beds in western Victoria (Fig. 7.3). There are also strong similarities in the nature of granitic intrusives and sequential development of granitic magmatism between the southern Delamerian fold belt and Glenelg River Complex in western Victoria which Foden *et al.* (1990) suggested were part of a single province. Correlation with Cambrian rocks in South Australia was first suggested by Wells (1956) but geochronological and geochemical studies by Turner *et al.* (1993) indicated that the maximum age of the Glenelg River Complex may be Neoproterozoic.

Sequence set € 1

Outcrops of 'Archaeocyathinae marble' from near Carrickalinga Head to Sellick Hill confirmed the Cambrian age of part of the succession on Fleurieu Peninsula (Howchin, 1897). Additional observations chiefly from coastal exposures at Myponga Beach were provided by Madigan (1925, 1926, 1927). Four conformable lower

Cambrian formations — Wangkonda and Sellick Hill Formations, Fork Tree Limestone and Heatherdale Shale — were mapped and described by Abele and McGowran (1959). These, together with the conformably underlying Mount Terrible Formation (Daily, 1963) and overlying or intertonguing Truro Volcanics (Forbes *et al.*, 1972), comprise the Normanville Group on Fleurieu Peninsula (Daily and Milnes, 1973) and constitute sequence set € 1 in the eastern Stansbury Basin (Fig. 7.2).

Cambrian trilobites and archaeocyaths (initially thought to be Silurian) were discovered in the Parara Limestone near Ardrossan on Yorke Peninsula (Tepper, 1879; Etheridge, 1890). Fossils from this unit, the underlying Kulpara Formation (Daily, 1956) and from Minlaton 1 stratigraphic drillhole, enabled correlation with Early Cambrian carbonates in the Flinders Ranges (Horwitz and Daily, 1958). Subsequently, Daily (1976a,b) distinguished the Winulta Formation as the lowermost Cambrian unit on Yorke Peninsula, and established that the shallow marine Koolyurtie Member forms part of the Parara Limestone. The Winulta and Kulpara Formations and Parara Limestone correlate with the Normanville Group and comprise € 1 in the western Stansbury Basin (Fig. 7.2).

Neither of the alternative correlations proposed by Daily and Milnes (1971) and Jenkins (1990) is adopted here for unmetamorphosed sediments between Stokes Bay and Point Marsden on Kangaroo Island. The lowest exposed unit, the Mount McDonnell Formation (Daily, 1969), bears no lithological resemblance to the Carrickalinga Head Formation of the Kanmantoo Group. Archaeocyaths recently discovered in the upper Mount McDonnell Formation near Cape Cassini are from upper Faunal Assemblage 2, or a slightly higher level, indicating correlation with the Fork Tree Limestone of the Normanville Group.

Daily (1956) defined the Kangaroo Island Group as 'a conformable sequence of strata ... between "Bald Rock" and White Point'. In ascending order, the strata are upper Smith Bay Shale (Horwitz and Daily, 1958), White Point Conglomerate, Emu Bay Shale (Sprigg *et al.*, 1954; Sprigg, 1955) and Boxing Bay Formation (Daily, 1956). An older section, comprising the Mount McDonnell Formation, Stokes Bay Sandstone and lower Smith Bay Shale, was recognised subsequently and, together with the Kangaroo Island Group, was correlated with the Kanmantoo Group on Fleurieu Peninsula (Daily, 1969).

The older section, possibly to the basal, sandstone-dominated levels of the White Point Conglomerate, is clearly not equivalent to the Kanmantoo Group. It will be necessary to establish the stratigraphic succession between the Stokes Bay Sandstone and Boxing Bay Formation before the Kangaroo Island Group can be satisfactorily redefined. The fossil evidence from outcrops and stratigraphic hole SADME Investigator 1 (Belperio and Hibburt, in prep.) shows that the older section correlates with the Normanville Group. A geological sketch and suggested correlation with the Normanville and Kanmantoo Groups are shown on Figure 7.4.

Distinctive conglomerate and sandstone in fault contact with the White Point Conglomerate at 'Bald Rock' were suggested by Daily *et al.* (1979) to comprise part of a younger Cambrian sequence possibly equivalent to Yuruga Formation on Yorke Peninsula. Dinnick (1985) described their fluvial character, and both he and Jenkins (1990) suggested a stratigraphic position in the upper part of a 'molassic apron' formed late in the Delamerian Orogeny. The beds contain clasts of granite and metamorphosed Kanmantoo Group, and may be as old as Ordovician or as young as Permian.

Eruption and emplacement of the Truro Volcanics near the top of sequence € 1.3 in the northeastern Stansbury Basin preceded tectonic activity which profoundly altered basin geometry during deposition of the ensuing sequence € 2.1. Onset of lithospheric extension, the Kangarooian Movements of Daily and Forbes (1969), was expressed as relative uplift on the margin of the Gawler Craton ('Cassinian Uplift' of Thomson, 1969), and as rapid subsidence across a hinge zone extending northeast from Kangaroo Island to northern Fleurieu Peninsula (Stuart and von Sanden,

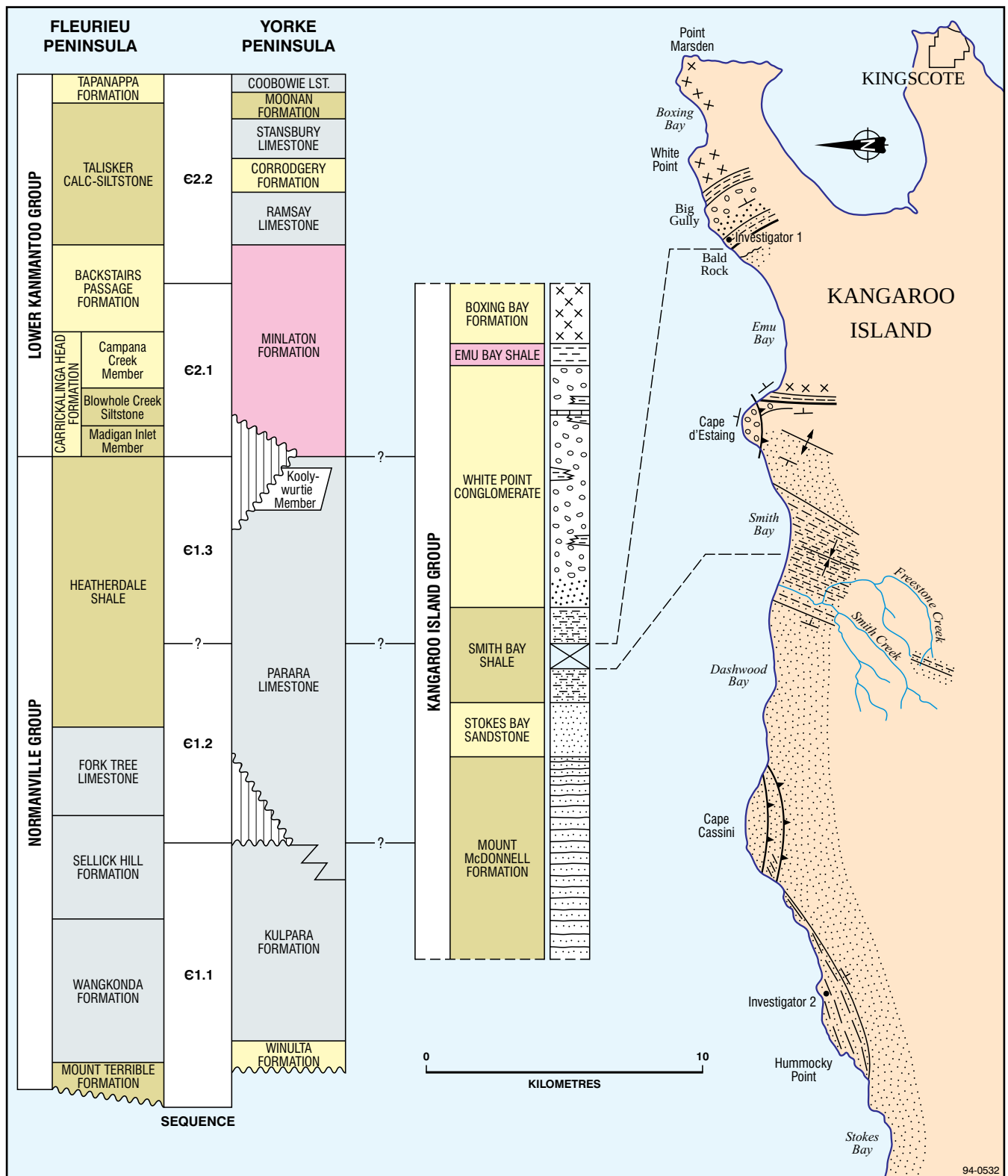


Fig. 7.4 Stratigraphy and suggested correlation, north coast of Kangaroo Island.

1972; 'Waitpingan Subsidence' of Thomson, 1969). Differential fault movement took place in the Torrens Hinge Zone along the east coast of Yorke Peninsula and immediately north of Kangaroo Island. Fault blocks were eroded locally to crystalline basement and debris accumulated on the flanks of elevated blocks and in structural depressions. These movements were restricted to the fault zones and there were no regional uplifts in the area of present day Gulf St Vincent as suggested by Thomson (1969) and Daily *et al.* (1979).

Sequence € 1.1 lowstand deposits

The palaeogeography of sequence € 1.1 is shown on Figure 7.5 (see legend on page 8).

The Mount Terrible and Winulta Formations discordantly overlie the lower Wilpena Group (Thomson and Horwitz, 1961). Where Winulta Formation rests non-conformably on crystalline rocks of the Gawler Craton on Yorke Peninsula, the basal unit is cross-bedded, coarse-grained to conglomeratic arkose. No fossils

have been found in the basal 2-3 m, which may be fluviatile in origin. Succeeding units are pyritic and glauconitic sandstone with minor shale, siltstone and dolomite.

Nearly 100 km further southeast on Fleurieu Peninsula, coarse-grained arkose of basal Mount Terrible Formation passes up rapidly into shallow marine, bioturbated siltstone and fossiliferous, silty calcareous sandstone (Daily, 1963). Sabelliditid tubes, small shelly fossils in the Mount Terrible Formation, and a similar shelly fauna and trace fossils in the Winulta Formation, were considered indicative of the Tommotian Stage by Daily (1976a).

The conformably overlying Kulpara and Wangkonda Formations are chiefly stromatolitic and fenestral micritic limestone, and ooid grainstone. Sandstone interbeds in the Wangkonda Formation are bioturbated; shelly fossils occur in mottled limestone near the base (Daily, 1963) and several phosphatised hardgrounds have been documented (Carroll, 1982). Both formations represent peritidal, carbonate-dominated mudflat to sandflat settings which extended from Yorke to Fleurieu Peninsula.

Sequence € 1.1 transgressive deposits

The Wangkonda Formation is locally scoured by shallow sub-tidal, cross-bedded, fine to coarse-grained, arkosic sandstone and burrowed siltstone of the basal Sellick Hill Formation (Abele and

McGowran, 1959; Daily, 1963; Alexander and Gravestock, 1990). Sediment starvation in the upper Wangkonda and lower Sellick Hill Formation is indicated by multiple phosphatised hardgrounds. Silty ribbon limestone of the lower Sellick Hill Formation is characterised by discordant surfaces, slumps and intraformational breccias. Alexander and Gravestock (1990) interpreted this succession as an upward-deepening carbonate ramp above initially shallow subtidal siliciclastics. The distinctive breccias with their unusually stacked platy clasts have been suggested to originate from intense storm activity on the ramp (Mount and Kidder, 1993).

On Yorke Peninsula, closest to exposed Gawler Craton, markedly different responses to sea-level change are evident in sequence € 1.1. South of a hinge line across central Yorke Peninsula, the Kulpara Formation and overlying Parara Limestone are conformable. In several cored drillholes near Curramulka (e.g. SADM CD 1 and 2 at Curramulka Quarry, Minlaton 1, Aquitaine SYC101), stromatolites, oolite and fossil wackestone of Kulpara Formation are overlain sharply by massive to nodular, glauconitic fossil wackestone of the Parara Limestone, signifying onset of transgression; in other holes the contact is transitional. Archaeocyaths from uppermost Kulpara and basal Parara in these drillholes are from lower Faunal Assemblage 2 (Jell *et al.*, 1990; Zhuravlev and Gravestock, 1994), and the succession marks an increase in water depth with no break in the fossil record.

CARBONATES AND EVAPORITES		SILICICLASTICS
Alkaline playa _____  Evaporative mudflat _____ 	NON-MARINE	Aeolian and fluvial deposits _____  Fluvio-lacustrine deposits _____  Alluvial fan conglomerate _____ 
Supratidal cyanobacterial boundstone (stromatolites, cryptalgal laminites, fenestral laminites) _____  Halite interbedded with sediments _____  Grainy sediments (oolite, bioclastic packstone) _____  ± sand ± tuff beds _____ 	MARGINAL-MARINE SETTINGS	Exposed redbeds _____  Conglomeratic deposits _____  Sandy tidal deposits _____  Shoreline sandstone _____ 
Calcified algal buildups, thrombolites _____  Archaeocyath and algal buildups _____ 	SHALLOW MARINE SETTINGS	Shelf sandstone _____  Channel sands _____  Fine-grained siliciclastics _____ 
Ribbon, mottled, nodular limestone _____  Slope limestone, turbidite _____  Basinal lime mudstone _____ 	OUTER SHELF, SLOPE AND DEEP BASINAL SETTINGS	Fine-grained siliciclastics _____  Shale, occasionally pyritic, phosphatic _____  Submarine fan deposits _____ 
Shoals, conjectural _____  Coast _____  Conjectural coast _____  Sequence thickness (m) determined from outcrop/drillhole _____  Strata exposed during ensuing lowstand _____ 		IGNEOUS ROCKS Volcanics, tuff _____  Syndepositional faults _____ 

94-1083 MESA

Legend to Figures 7.5-7.7, 7.10, 7.11 and 7.14.

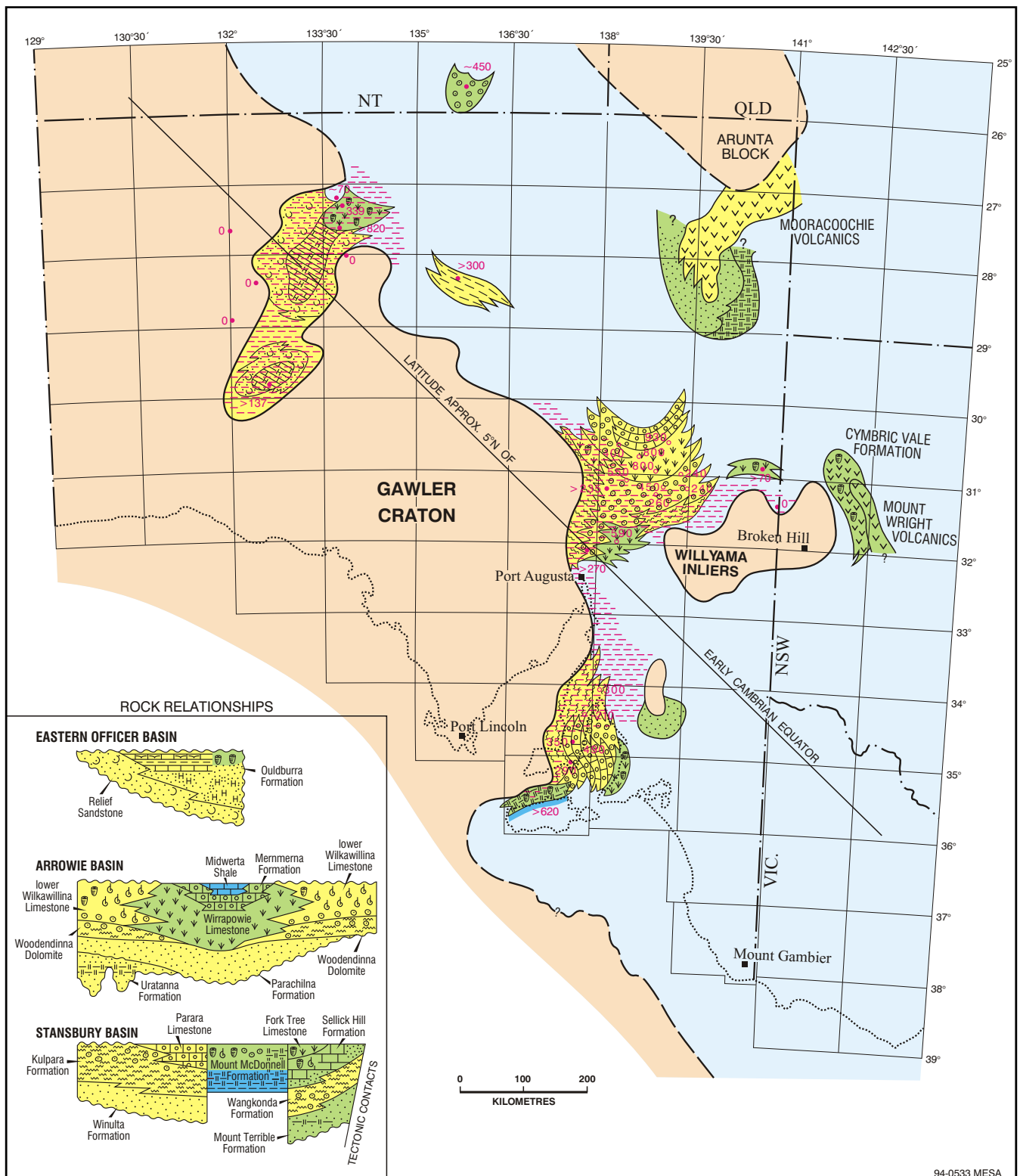


Fig. 7.5 Palaeogeography, sequence € 1.1. Basal clastic successions in the Arrowie and Stansbury Basins are not shown.

Sequence € 1.1 highstand deposits

Thin bioclastic packstone within the Sellick Hill Formation (facies association D, Alexander and Gravestock, 1990) represents multiple erosion and reworking of bioherms brought about by sea-level fall late in the highstand. It is the ramp response to shallowing and exposure of the adjacent carbonate shelf, and marks the top of sequence € 1.1 on Fleurieu Peninsula.

The Mount McDonnell Formation may represent the highstand systems tract on Kangaroo Island by virtue of its assumed

stratigraphic position (Figs 7.2, 7.4) and upward-shallowing nature. From its lowest intersection in drillhole Investigator 2, the 900+ m thick, gently dipping Mount McDonnell Formation comprises green-grey shale, siltstone and fine-grained sandstone. Linsen, flaser and ripple bedding are common but mudcracks are very rare. Burrows, possible mollusc trails, trilobite tracks and fragments of *Redlichia* have been recorded (Daily *et al.*, 1979; Moore, 1983). Within the upper 110 m, cross-sets of medium-grained sandstone up to 6 m thick are common, and oolite beds occur close to the conformable boundary with the Stokes Bay Sandstone. Moore (1983) interpreted a shallow subtidal environment with



Top: Upper Mount McDonnell Formation at Cape Cassini, Kangaroo Island. Photo 042414.

Bottom: Mounded storm breccia bed in flaggy carbonate ramp deposits, Sellick Hill Formation, Myponga Beach. Photo 042415.

development of migrating sand shoals at late stages of deposition. At Cape Cassini, a thin limestone interbed contains well-preserved archaeocyaths, trilobites and small shelly fossils. The problematic alga *Nuia* is also present in abundance.

Sequence € 1.2 lowstand deposits

The lowstand systems tract is recorded by a hiatus at the Kulpara-Parara boundary on northern Yorke Peninsula, by outer ramp biohermal limestones in the upper Sellick Hill Formation and Fork Tree Limestone on Fleurieu Peninsula, and by the Stokes Bay Sandstone on Kangaroo Island.

At Horse Gully, north of the central Yorke Peninsula hinge line, bioclastic packstone in the condensed upper few metres of the Kulpara Formation is overlain disconformably by Parara Limestone (Jell *et al.*, 1990; Bengtson *et al.*, 1990). Solution sculpture at the boundary and sediment-filled cavities beneath indicate subaerial exposure. A break in the fossil record at Horse Gully, where upper Faunal Assemblages 2 and 3 are condensed, is marked by a similar hiatus to that on the carbonate shelf in the Arrowie Basin. A sketch of regions affected by sea-level fall is shown on Figure 7.5.

On Fleurieu Peninsula, isolated bioherms in muddy limestone overlie sequence € 1.1 packstone in the upper Sellick Hill Formation. Similar bioherms coalesced as massive complexes in the conformably overlying Fork Tree Limestone. Assemblage 2 archaeocyaths (Debrenne and Gravestock, 1990) formed a rigid framework encrusted by calcified microbial microfossils. Recently discovered archaeocyaths in the Macclesfield Marble confirm earlier correlations with the Fork Tree Limestone (Sprigg, 1990).

Recrystallisation of much of the Fork Tree Limestone hinders facies interpretation. At least half the formation appears biohermal but there are also numerous calcarenite and occasional rudstone beds which, in places, resemble grain flows. These, and the underlying bioherms, represent moderately deep-water lowstand deposits on the outer part of a carbonate ramp on Fleurieu Peninsula, but facies metamorphosed to marble phases are indeterminable.

The Stokes Bay Sandstone, which transitionally overlies Mount McDonnell Formation, has an apparent thickness of 700 m on the north coast of Kangaroo Island. It consists of red, trough cross-bedded, planar-laminated and convolute-bedded, fine to medium-grained arkose. Minor interbeds 1-2 m thick of green-grey lithic siltstone contain reworked discoidal mud clasts.

Palaeocurrent data and the frequency of contorted and planar-laminated beds indicate rapid deposition by strong westerly currents (Daily *et al.*, 1979; Moore, 1983). Stokes Bay Sandstone is envisaged by Moore (1983) to have been deposited on a subtidal shelf swept by shore-parallel currents.

Sequence € 1.2 transgressive deposits

Transgressive deposits of sequence € 1.2 are shown on Figure 7.6.

The mottled upper member of the Fork Tree Limestone, as well as the lower Heatherdale Shale on Fleurieu Peninsula and Parara Limestone on Yorke Peninsula, represent renewed and widespread marine transgression (Fig. 7.2).

The passage from 'grainy' and biohermal to mottled, micritic Fork Tree Limestone is abrupt but conformable. Between Sellick Hill and Myponga Beach, the mottled member overlies bioherms with an upper Assemblage 2 fauna but is itself poorly fossiliferous. There is a rapid upward transition to Heatherdale Shale which is 300 m thick and comprises pyritic, calcareous shale and siltstone. Nodular calcareous shale predominates at Carrickalinga Head (Abele and McGowran, 1959; Daily, 1963) and may represent only the lower part of the Heatherdale Shale.

Thin, mottled and massive limestone associated with black shale and basic volcanics in drillhole MBT 1 near Peebinga possibly belong in sequence € 1.2 or € 1.3.

Systems tracts cannot be differentiated within the Heatherdale Shale which is regarded here as a transgressive to highstand basinal deposit encompassing both sequences € 1.2 and € 1.3. On Fleurieu Peninsula, it is either in fault contact with, or overlain erosively by, the Carrickalinga Head Formation, which locally contains reworked shale clasts. Up to 300 m of Heatherdale Shale may have been removed by erosion at Carrickalinga Head (Jago *et al.*, 1994a).

Most of the Heatherdale Shale at its type locality is massive to flat-laminated, rarely ripple-laminated, black, pyritic and phosphatic shale. Organic-walled microfossils occur in upper levels (Foster *et al.*, 1985) and conocoryphid trilobites have been discovered (Jago *et al.*, 1984; Jenkins and Hasenohr, 1989). From the presence of trilobites, their tracks, and other trace fossils, Jenkins and Hasenohr (1989) suggested a dysaerobic, deep-water environment of deposition. Of particular interest is their discovery of a thin tuff bed which has preserved some fossil traces. Zircons from the tuff have yielded a U-Pb age of 526 ± 4 Ma (Cooper *et al.*, 1992).

South of the hinge line on Yorke Peninsula, the Parara Limestone, up to 270 m thick, consists in the lower part of stylonodular and massive glauconitic skeletal wackestone. Archaeocyaths in the basal metre are followed by abundant specimens of the small pelecypod *Pojetaia*, common trilobites, hyoliths and *Chancelloria*. In several cored holes the nodules, initially with 'fitted fabric', are increasingly separated by black, laminated lime mud which eventually forms up to 80% of the sediment. Fossiliferous nodular wackestone and poorly fossiliferous, laminated lime mud alternate up-section on a centimetre scale. Several tuffaceous layers 0.1 to 0.4 m thick occur in SYC101 drillcore. The succession deepens



upward from a shallow to deep shelf with alternating aerobic and anaerobic bottom sediments. This part of the formation appears to span all of sequence € 1.2 and the lower part of sequence € 1.3. Tuffs in the Heatherdale Shale and Parara Limestone probably represent early eruptive phases of the Truro Volcanics (Forbes *et al.*, 1972) discussed below.

to have grown during a phase of rapid drowning of the shelf, and hence they represent initial transgression. The red layer is overlain by 25 m of Parara Limestone which crops out poorly at the type locality (Horse Gully). Beds comprise rubbly, nodular wackestone and metre-thick, massive, bioclastic packstone containing trilobites and numerous small shelly fossils from Faunal Assemblages 3 and 4. These are often concentrated in hollows on thin phosphate-coated hardgrounds (Carroll, 1982; Tucker, 1989). Near Kulpura, the 150 m thick nodular Parara Limestone contains Faunal Assemblages 5 to 7. *Pararaia tatei* (Assemblage 4) has recently been found at this locality (Jell *in* Bengtson *et al.*, 1990).

At Smith Bay and Freestone Creek on Kangaroo Island there is a conformable transition from Stokes Bay Sandstone to Smith Bay Shale. The latter comprises grey, green and red micaceous shaly siltstone with 0.3-0.5 m thick packages of convolute-bedded arkose of Stokes Bay type through most of the unit. Trilobite tracks and shallow, vertical burrows are abundant and well preserved in the fine-grained facies. Ripple laminations and synaeresis cracks are common. The large 'mudcracks' reported by Daily *et al.*, (1979) and Moore (1983) from Smith Bay are either sand sills and dykes or wedge-fill which may not be related to deposition or desiccation of the unit.

Thin intraclast lime mudstone near Freestone Creek (Madigan, 1928) contains fragments of *Redlichia* (Daily *et al.*, 1979). At 'Bald Rock', thin cross-beds and conglomerate lenses contain *Nuia*-like algal fossils.

Bipolar current ripples and upward-fining cycles suggest strong shore-parallel tidal influence (Moore, 1983) in a shallow subtidal setting. Upper Smith Bay Shale at 'Bald Rock' may be little more than 150 m thick, but the Smith Bay Shale could span sequences € 1.2 and € 1.3.

Sequence € 1.2 highstand deposits

The effects of the € 1.2 highstand, evident in shallow-water regions, are not discernible in the deep-marine Parara Limestone and Heatherdale Shale.

Sequence € 1.3 transgressive-highstand deposits

Renewed volcanism and tectonic activity recorded in sequence € 1.3 overprinted the effects of eustatic sea-level change. Lowstand deposits have not been recognised and inadequate knowledge of the sediments interbedded with the Truro Volcanics hinders sequence analysis. It is not yet possible to recognise the € 1.2 -€ 1.3 sequence boundary in the Stansbury Basin. Strata are described below with reference to locality, and systems tracts are assigned only where data permit. A sketch of € 1.2 highstand and all of sequence € 1.3 facies is shown on Figure 7.7.

From Horse Gully in the north to Stansbury West 1 drillhole on southern Yorke Peninsula, argillaceous, nodular carbonates are overlain by the Koolywurtie Member of the Parara Limestone. The member is 70 m thick and, judging by metre-scale boulders re-worked into White Point Conglomerate on Kangaroo Island (Daily *et al.*, 1979, 1980), it once extended as far south as Investigator Strait, a strike length of more than 120 km.

The Koolywurtie Member is a complex of stacked bioherms and interbiohermal beds capped by fenestral *Girvanella* boundstone, sometimes forming stromatolites. Rare desiccation cracks indicate local subaerial exposure of the uppermost bioherms. Late Botomian archaeocyaths (Jell *et al.*, 1990; Zhuravlev and Gravestock, 1994) and dense masses of *Epiphyton* form the bioherm frameworks. The bioherm complex fringed the southeastern shoreline of the Gawler Craton and represents a seaward-prograding highstand deposit in upper € 1.3. The reef was eroded from structurally elevated zones during the Kangarooian Movements.

Near the top of a fault-bounded exposure of Koolywurtie Member at Horse Gully, the first clastic influx from the Kangarooian Movements on Yorke Peninsula is represented by an interbed of coarse-grained, well-rounded, feldspathic sandstone (Tucker, 1989), some of which infiltrated growing bioherms (cf. Moorowie Formation in the Arrowie Basin).

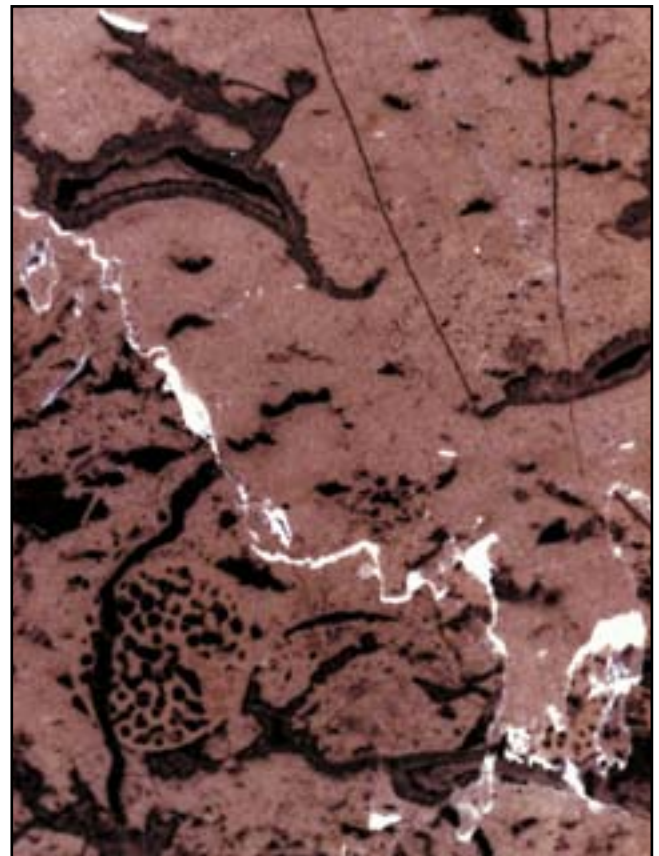
The contact with the overlying Minlaton Formation (Daily, 1976b) is conformable or erosional, depending on locality (contacts are known only in drillcores). Where conformable, the Koolywurtie Member is either overlain by black, fossil-rich calcareous shale, siliceous shale and then red siltstone, or is overlain transitionally by redbeds. Where erosional, much of the Koolywurtie Member is missing, incorporated as clasts in conglomerate of the Minlaton Formation. Stratigraphic relationships are shown on Figure 7.8.

White Point Conglomerate

The White Point Conglomerate, Emu Bay Shale and Boxing Bay Formation may span the sequences € 1.3 to lower € 2. Three over-generalised sedimentary facies associations have been described in the 1 300 m thick succession (Daily *et al.*, 1980). Dinnick (1985) mapped the complex intertonguing and erosional character of the White Point Conglomerate and Boxing Bay Formation, especially around Cape d'Estaing where the White Point Conglomerate is faulted against Stokes Bay Sandstone. The White Point Conglomerate and Boxing Bay Formation are probably both part of a single migrating facies complex. Only that part of the White Point Conglomerate beneath a burrow-mottled limestone marker is discussed below; the remainder is assigned to € 2.1.

Red sandstone, minor siltstone and shale in the basal 170 m of White Point Conglomerate may represent highstand deposits on Kangaroo Island. This succession comprises trough cross-stratified, fine-grained feldspathic sandstone in cosets 1-3 m thick. It is overlain erosively by conglomerate with a quite different sandstone matrix (Daily *et al.*, 1980).

Cobbles and boulders in the White Point Conglomerate *sensu stricto* include gneiss, chert, sandstone, dolomite and biohermal limestone, some identical to the Koolywurtie Member and others of bioherm types found hitherto only in the Arrowie Basin. Most of the coarse conglomerates are horizontally bedded and framework-supported, but trough cross-stratified, coarse-grained and granule sandstone forms a subordinate component (Daily *et al.*, 1979, 1980). The sandstone and fine-grained facies associations of Daily *et al.* (1980) are minor components of White Point Conglomerate above the basal 170 m. Ripple-bedded and flaser-bedded, very fine-grained sandstone and mudstone form units rarely more than 8 m thick. These are frequently eroded by succeeding conglomerate beds. One such mudstone bed has yielded trilobite fragments (Faunal Assemblage 11 of Daily, 1956).



Photomicrograph of bioherm framework composed of *Epiphyton* and archaeocyaths; growth cavities are rimmed with calcite cement and partly filled with sediment. Koolywurtie Member at 431 m in Minlaton 1. Width of view is 28 mm. Photo 042416.

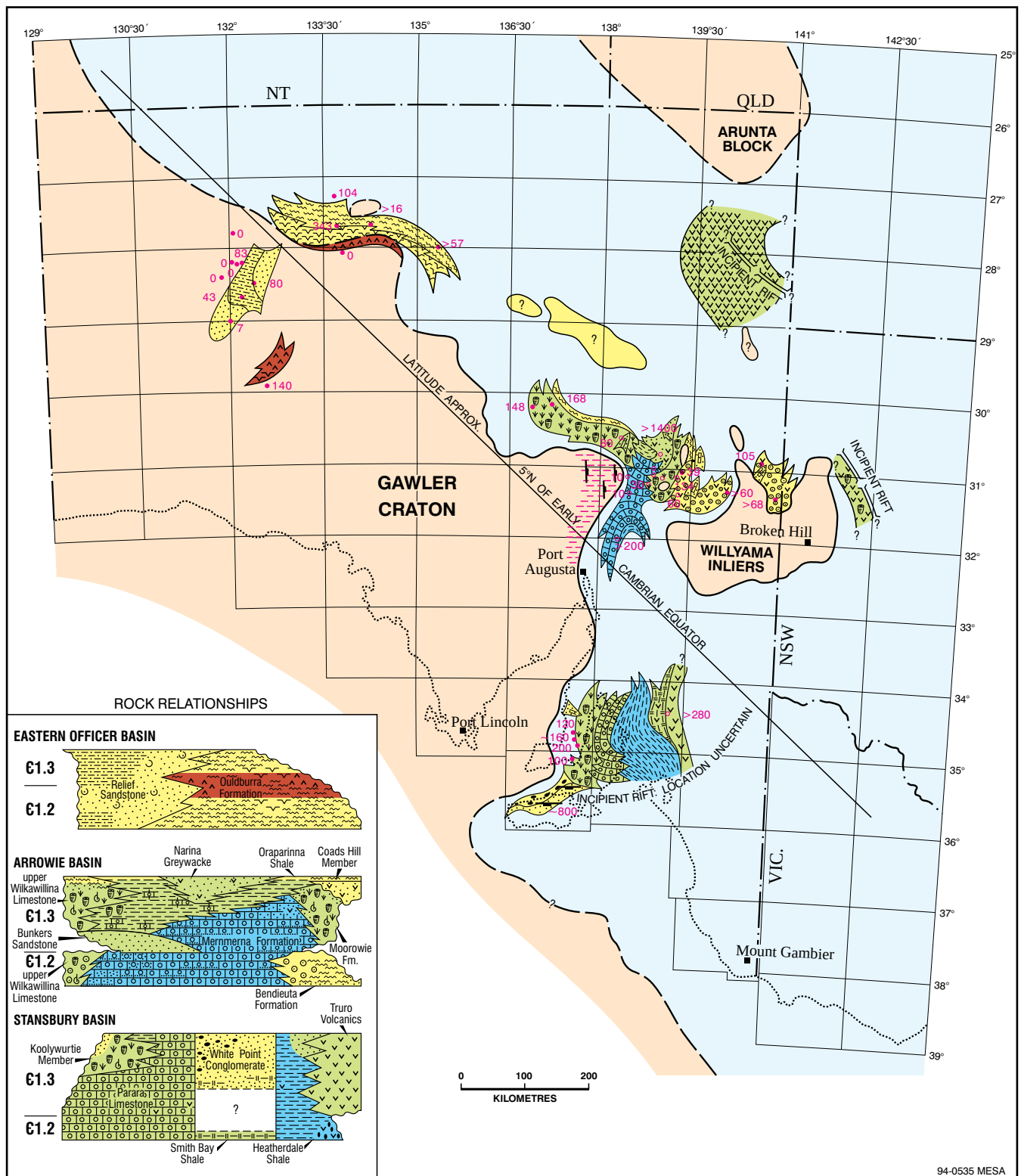


Fig. 7.7 Palaeogeography, sequence € 1.2 highstand and € 1.3.

Truro Volcanics

Forbes *et al.* (1972) suggested that the Truro Volcanics overlies Mount Terrible Formation, but available evidence suggests a faulted contact between Adelaidean siltstone and the volcanics.

The Truro Volcanics comprise 300+ m of altered vesicular and amygdaloidal andesite, trachyte and metabasalt of within-plate rift affinity (Rankin *et al.*, 1991b). Discrete beds, possibly individual flows, minor tuffs and intraformational conglomerates indicate extrusive volcanism. At the type section, apparently higher in the

section, 60 m of cream and mottled marble are correlated with upper Fork Tree Limestone (unit C4 of Forbes *et al.*, 1972). However, the contact between the volcanics and marble is faulted, so the stratigraphic relationship is uncertain. The Fork Tree Limestone is overlain by siltstone, volcanics and volcanoclastic sediments. A thin porphyritic andesite flow is capped by conglomerate 10 m or more thick with rounded clasts of scoria, porphyritic andesite and shale interbedded with black tuffaceous sandstone. Erosion of the underlying volcanics and sediments prior to this phase of eruption is implied, with the erosional break correlated tentatively

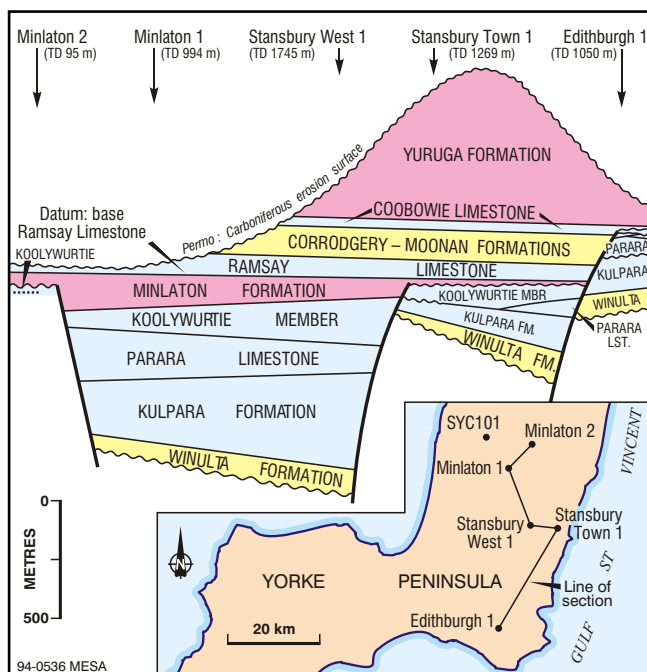


Fig. 7.8 Cross-section, Minlaton 2 to Edithburgh 1, Stansbury Basin, Yorke Peninsula.

with the base of € 1.3. Dark grey shale with phosphate nodules in nearby outcrop represents Heatherdale Shale overlying the conglomerate. The Heatherdale Shale passes conformably up into shallow marine clastics of the Madigan Inlet Member of the Carrickalinga Head Formation. *In situ* scoriaceous volcanics occur near Angaston (Cobb and Farrand, 1984). Approximately 260 m of interbedded Heatherdale Shale and Truro Volcanics crop out near Red Creek (Gatehouse, 1988a; Gatehouse *et al.*, 1993). Some beds are extrusive, and pillow lavas in one of these suggest sea-floor volcanism. The Heatherdale Shale here is predominantly siltstone with thin, cross-bedded sandstone interbeds, suggesting shallower water depth than at Sellick Hill. At Red Creek in the Karinya Syncline (Fig. 7.3), volcanics are also interbedded with the basal Madigan Inlet Member of the Carrickalinga Head Formation.

Tuffs and widespread volcanoclastic detritus in the Arrowie Basin, as well as the tuffs in Parara Limestone (Yorke Peninsula) and Heatherdale Shale (Jenkins and Hasenohr, 1989), are attributed to eruptive phases of Truro Volcanics. These are correlated with the Mooracoochie Volcanics in the eastern Warburton Basin, and with the upper Mount Wright Volcanics and Cymbric Vale Formation in western New South Wales.

Weakly foliated basalts, and fine-grained siliciclastics lacking penetrative deformation, have been drilled in the Yumali-Coonalpyn area (CSR Ltd, 1986). Contact relationships between the volcanics and sediments have not been established but the latter show alteration by hot, hydrous fluids. Approximately 130 km to the northeast, basalt was also intersected in MBT 1. Here, however, basalt and tuff occur with sheared and brecciated massive and mottled limestone indistinguishable lithologically from the Fork Tree Limestone, and with black pyritic shale reminiscent of the Heatherdale Shale (Clough and Rankin, 1990; Rankin *et al.*, 1991b). The succession dips at 50° and may be a series of fault slivers. The mafic igneous rocks at both localities constitute Suite 1 of Rankin *et al.* (1991b).

The Yumali-Coonalpyn drillholes are on complex magnetic anomalies parallel to the Padthaway Ridge which comprises mostly post-orogenic granitoid intrusives. MBT 1 was drilled on a magnetic anomaly at the northwestern termination of the geophysically interpreted extension of the Stavely Belt (Stavely magnetic anomaly on Fig. 7.3; domain V of Brown *et al.*, 1988). The location of the MBT 1 section adjacent to these aeromagnetic trends is likely to have resulted from tectonic emplacement, possibly



Top: Cobble-sized archaeocyath-bearing clasts in White Point Conglomerate, Cape D'Estaing, Kangaroo Island. Photo 042417.

Bottom: Pillow basalt in Truro Volcanics, Red Creek near Dutton. Photo 042418.

caused by convergence subsequent to rifting (model 1 of Rankin *et al.*, 1991b). The northwestern Stavely Belt, in particular, differs in magnetic intensity from the main belt (*sensu* Brown *et al.*, 1988, figs 3, 4). The main belt includes both the Mount Stavely Volcanic Complex and the Mount Dryden Sub-belt which Crawford (1988) pointed out to be compositionally distinct enough to have been derived from different igneous sources.

Geochemical composition and interbedded facies, where known, are considered here to be the best guides to the origin of the mafic suites. Basalts in the Yumali-Coonalpyn drillholes, in MBT 1, and in outcropping Truro Volcanics, all show within-plate affinities using the discrimination fields of Pearce and Cann (1973; Fig. 7.9). They are all considered to be Early Cambrian (sequence € 1.2 and/or € 1.3) correlatives despite their present wide separation and the lack of penetrative deformation in the drillcore examples. The suggestion of Rankin *et al.* (1991b) that the Truro Volcanics and correlatives represent a Gulf of Aden rift-type of volcanic suite dissected in the Middle-Late Cambrian by crustal extension is followed here. However, crustal extension probably culminated in the late Early Cambrian (late Botomian) with the onset of Kanmantoo Group sedimentation.

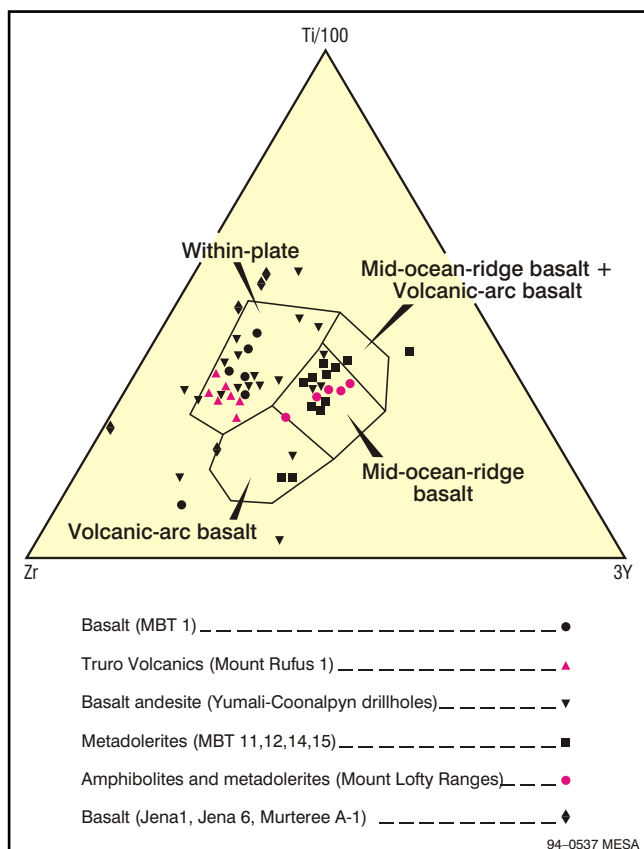


Fig. 7.9 Affinities of Cambrian volcanics in the Stansbury and eastern Warburton Basins. Discrimination fields are after Pearce and Cann (1973).

Sequence set € 2

Set € 2 in the eastern Stansbury Basin is represented by the lower Kanmantoo Group of Sprigg and Campana (1953), namely Carrickalinga Head Formation and Backstairs Passage Formation, Talisker Calc-siltstone and Tapanappa Formation (Daily and Milnes, 1971). This mainly siliciclastic succession is metamorphosed, folded and faulted and has not yet yielded sufficient fossils for correlation.

Formations in the western Stansbury Basin and on the north coast of Kangaroo Island have greater facies similarity to each other than to the lower Kanmantoo Group. Various correlations have been proposed, but that adopted here, in the absence of biostratigraphic constraint, is based on facies succession and tectono-sedimentary events. One particular event (recorded by the Milendella Limestone Member), recognised as a significant pause in clastic deposition, is used to correlate key beds.

The subgroup classification of Daily and Milnes (1972) is not followed here but the Nairne Pyrite Member (Thomson and Horwitz, 1962) and members of the Carrickalinga Head Formation are as proposed by Daily and Milnes (1971, 1972). Gatehouse, Jago, Gravestock *et al.* (1990) provided a preliminary sequence stratigraphic framework for the lower part of the Kanmantoo Group. Sequence stratigraphic subdivision is difficult to interpret in terms of sea-level change, since the Kanmantoo Trough was a tectonically active basin, probably with repeated extensional faulting, and is not attempted here. Sequence set € 2 is described below for separate regions of the Stansbury Basin. Palaeogeography and facies of part of sequence set € 2 are shown on Figure 7.10.

Lithic sandstone and mudstone couplets in the Madigan Inlet Member, Carrickalinga Head. Photo 042419.

Eastern Stansbury Basin

The base of the Madigan Inlet Member of the Carrickalinga Head Formation at Carrickalinga Head is erosional (Jago *et al.*, 1994a). Madigan Inlet Member comprises alternating couplets of lithic sandstone and mudstone, with sharp bases and gradational tops. Shale clasts occur in the fine-grained sandstone which shows evidence of rapid deposition, mass movement and dewatering. Thinner, more numerous couplets are exposed on southern Fleurieu Peninsula where the member is 550 m thick. In these southern areas, lowstand submarine fan facies are proposed (Gatehouse, Jago and Cooper, 1990). However, at Sedan Hill where the member is 1 000 m thick, flaser and linsen bedding and asymmetric ripples suggest a shallower water setting, and the basal contact is transitional.

The conformably overlying Blowhole Creek Siltstone Member is laminated phyllite with occasional thin fine-grained sandstone interbeds. These display ripples, linsen and worm burrows. In the Karinya Syncline a marble-calc-silicate marker horizon intervenes within the phyllites as the Milendella Limestone Member (White, 1959), representing an episode of arrested siliciclastic input with consequent carbonate sedimentation (Cooper, 1989). This member is chosen as the key to correlation with a unit in the Minlaton Formation on Yorke Peninsula and a marker bed in the White Point Conglomerate on the north coast of Kangaroo Island. However, the less widespread Tungkillio Marble Member of the Backstairs Passage Formation (White, 1959) should not be ignored as an alternative.

A sharp change from phyllite to metasandstone marks the appearance of Campana Creek Member which is 135 m thick at its type section and 535 m at Sedan Hill. It comprises locally slumped, cross-bedded, ripple and flat-laminated sandstone. Laminated siltstone and shale interbeds occur, and some sandy horizons are bioturbated. Hyolithid cones, the youngest body fossils in the Kanmantoo Group, occur near the top of the member. At both localities, upward-coarsening and upward-shallowing facies are indicated (Gatehouse and Jago, 1988).

Laminated siltstone is abundant in the lower part of the transitionally overlying Backstairs Passage Formation. It is bioturbated and interbedded with fine-grained sandstone, which increases upward to form planar laminated, metre-thick cross-beds. Shallow erosional channels filled with cross-bedded sand are common. Slumping of oversteepened foresets and upper flow regime planar beds indicate rapid deposition in a high-energy marine setting (Daily and Milnes, 1971).

An abrupt lithological change marks the appearance of Talisker Calc-siltstone, a phyllite with pale calc-phyllite or marble interbeds which give the formation a characteristically banded appearance. Sulphide-rich bands are common in the lower part of the formation and the Nairne Pyrite Member exemplifies exceptional sulphide enrichment. The calc-siltstone was deposited under strongly



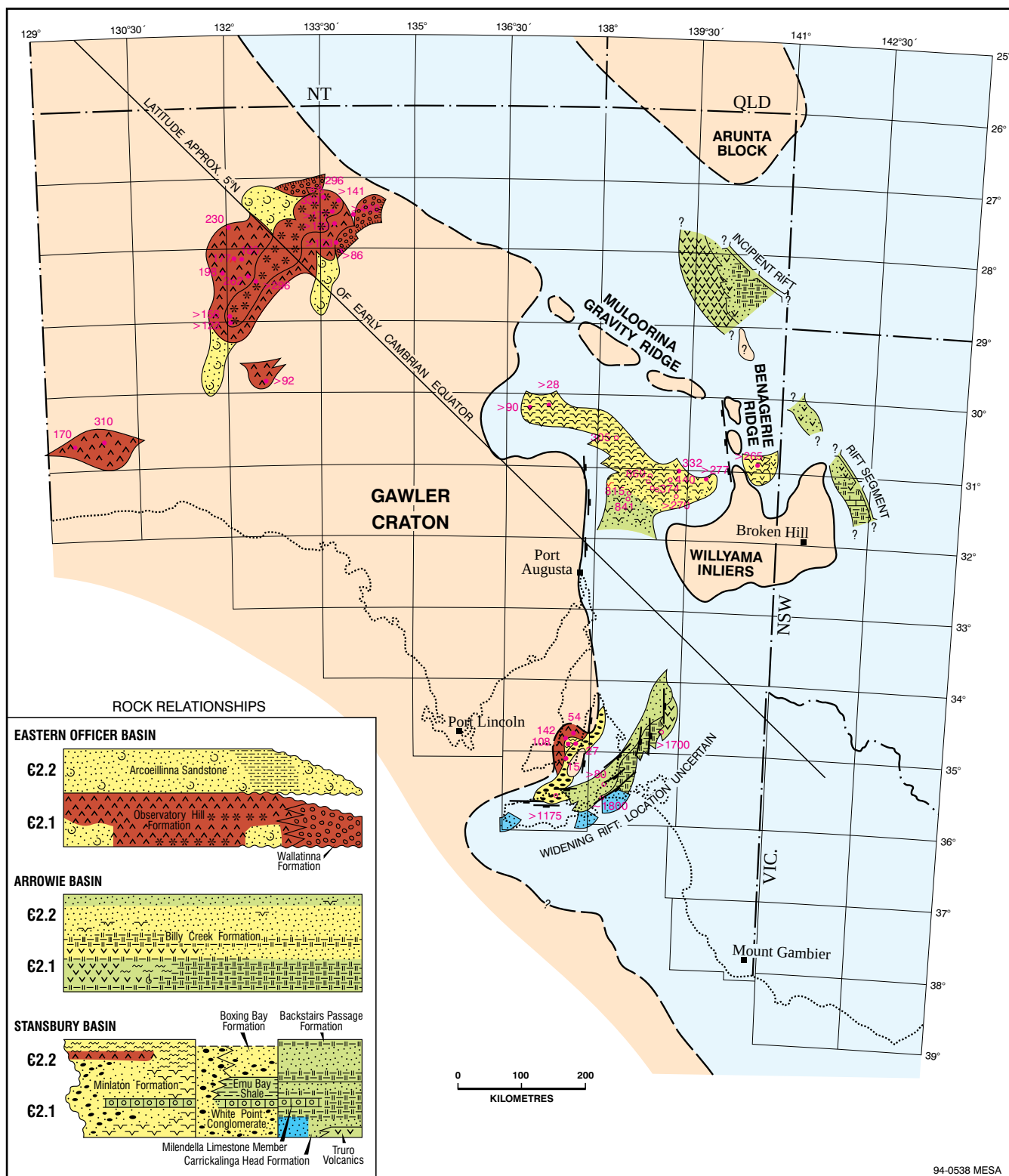


Fig. 7.10 Palaeogeography, sequence 62.1 and part 62.2.

reducing conditions (Daily and Milnes, 1972). A deep offshore environment with oscillating supply of terrigenous clay and lime mud is indicated. Uncommon lenticular sandstone interbeds may represent distal fan lobes (I.A. Dyson, Flinders University of South Australia, pers. comm., 1994).

The Karinya Shale (Thomson, 1969) is equivalent to Talisker Calc-siltstone and occupies a similar stratigraphic position in the Karinya Syncline. It comprises 200 m of dark grey, pyritic laminated siltstone deposited in a deep marine setting.

The abruptly overlying Tapanappa Formation at its type section is predominantly fine to coarse-grained metasandstone in beds 1 to 2 m thick, separated by very thin, laminated metasiltstone and phyllite. Cross-bedding is common, especially in channel-filling sandstone and conglomerate. Clasts include quartzite, rare phyllite, gneiss and marble (Daily and Milnes, 1971). Similar lenticular conglomerates up to 3 m thick crop out on Dudley Peninsula (Daily *et al.*, 1979). Numerous sulphide-rich bands occur in the upper part of the formation, and these contain significant copper sulphides at the Kanmantoo Mine (Seccombe *et al.*, 1985).

Turbidite facies were noted by Mancktelow (1979) at the type section, and by Flint (1978) in Tapanappa Formation equivalents on the west coast of Kangaroo Island. Daily *et al.* (1979) regarded the latter as a relatively shallow marine prograding sand sheet, which may be as young as Middleton Sandstone. Flint's interpretation is that of a submarine fan environment in which waning turbidity currents rapidly deposited sediments in water below storm wave base.

Western Stansbury Basin

Set € 2 in the western Stansbury Basin is less than 400 m thick. Apart from the poorly outcropping Ramsay Limestone (Crawford, 1965), the remaining formations are known only from drillholes. Uplift and erosion along the Torrens Hinge Zone during the Kangarooian Movements locally stripped units to basement. These were reworked as conglomerates in the Minlaton Formation (Daily, 1976b), which is disconformable on older units in Minlaton 2, Stansbury Town 1 and Edithburgh 1 drillholes (Fig. 7.8). Wireline log and core data show a conformable succession in Minlaton 1, SYC101 and Stansbury West 1.

Two contrasting facies associations are evident. The first is a redbed-conglomerate suite, the other is a succession of clastics, carbonate and evaporites, both found in Minlaton 1 drillcore (Crawford, 1965). There is a conformable passage from Parara Limestone to grey-green shale of the Minlaton Formation which becomes red-brown, silty and micaceous over a 20 m interval, before interbeds of conglomerate first appear. The conglomerate (10 m thick) is composed of clasts of Kulpara Formation and Parara Limestone and is crudely interbedded with muddy fine-grained sandstone and shale which then predominate over the following 65 m. A 20 m thick dolomitic limestone follows; this is stromatolitic, peloidal, bioturbated, and vuggy, corresponding to 'Unit D' of Watts and Gausden (1966) in Stansbury West 1. It is correlated here with Milendella Limestone Member of the Carrickalinga Head Formation (Fig. 7.10). Finally, 13 m of buff stromatolitic dolostone with numerous anhydrite and thin sandstone and shale interbeds intervenes before the Minlaton Formation is overlain transitionally by Ramsay Limestone.

In Minlaton 2, the Minlaton Formation is only 54 m thick (Blissett, 1968). Koolyurwurtie Member bioherms are overlain erosionally by debris-flow conglomerates 1 to 3 m thick, separated by muddy fine sandstone or upward-fining pebble conglomerate with clasts of reworked Kulpara Formation and Parara Limestone. The top of this 33 m sequence is marked by a corroded surface overlain by a 5 mm thick layer rich in glauconite. In the succeeding conglomerate, clasts are supported by a coarse arkosic sandstone matrix and basement gneiss pebbles are more abundant than carbonate. Fine-grained arkosic sandstone, siltstone, bioturbated mudstone and stromatolitic dolostone complete the succession which passes conformably into Ramsay Limestone. The thin glauconitic layer between the conglomerates points emphatically to a break in coarse clastic deposition and this is also correlated with the Milendella Limestone Member.

The transgressive marine Ramsay Limestone is richly fossiliferous, containing Faunal Assemblage 10 and other skeletal remains, found also in the Wirrealpa and Aroona Creek Limestones of the Arrowie Basin (Daily, 1956, 1957; Brock and Cooper, 1993). Basal beds of sandy ooid grainstone, oncolites and bioclastic packstone pass up into black, nodular lime mudstone which is also fossiliferous and comprises the bulk of the formation.

The remaining formations assigned to sequence set € 2 consist of alternating siliciclastics and carbonates, namely Corrodgery Formation (Jago and Daily, 1982; Daily, 1990), Stansbury Limestone (Daily, 1968), Moonan Formation (Jago and Daily, 1982; Daily, 1990) and Coobowie Limestone (Daily, 1972). The Corrodgery and Moonan Formations have not been cored. Brief cuttings descriptions (Daily, 1968, 1990) indicate that Corrodgery is calcareous, red and green-grey, micaceous, feldspathic sandstone, and Moonan is dark grey non-calcareous shale followed by thin

sandstone beds. The environment of deposition is presumed to have been shallow marine. The Stansbury Limestone is pyritic ooid grainstone and flat to ripple-laminated lime mudstone. Fragments of *Redlichia* were recorded by Daily (1968) from Stansbury West 1. Coobowie Limestone is pale grey oolitic limestone. Both formations were deposited in a shallow subtidal setting.

Wireline logs from Stansbury West 1 show gradual transitions between carbonates of the Ramsay and Stansbury Limestones and siliciclastics of the Corrodgery Formation. Cyclic packages (parasequences) 4-8 m thick are evident particularly in the Corrodgery Formation. They indicate balance between coarse and fine clastic influx and *in situ* carbonate production at maximum transgression. In contrast, the basal Moonan and top Coobowie contacts are abrupt transitions to siliciclastics. These may be related to downcutting associated with sea-level fall at the top of € 2. The transgressive-regressive Ramsay-Coobowie sediment package is assigned confidently to sequence € 2.2 (Fig. 7.2).

North coast of Kangaroo Island

A burrow-mottled limestone marker bed occurs near the top of the White Point Conglomerate at Cape d'Estaing (Fig. 7.4). This also marks a pause in the influx of coarse-grained clastics. It is equated with either a limestone or equivalent glauconitic layers in the Minlaton Formation on Yorke Peninsula and with the Milendella Limestone Member of the Carrickalinga Head Formation in the Mount Lofty Ranges. The marker bed, which contains trilobites allowing correlation with Billy Creek Formation and Oraparinna Shale in the Arrowie Basin (Pocock, 1970; Daily, 1976b; Moore, 1979a,b, 1980), and the succeeding Emu Bay Shale at its type section east of Cape d'Estaing, comprise mainly the fine-grained facies association of Daily *et al.* (1980). Faunal Assemblage 12, comprising several trilobite genera, the crustacean *Isoxys*, annelids and hyoliths, is from a shale at Big Gully, reportedly equivalent to Emu Bay Shale (Glaessner *in* Sprigg, 1955; Glaessner, 1979; Daily, 1956; Daily *et al.*, 1979). Remains of the large, invertebrate predator *Anomalocaris* have been described from this site (McHenry and Yates, 1993). The trilobites belong to the *Pararaia janeae* Zone (Jell *in* Bengtson *et al.*, 1990). Recent finds of juvenile *Hsuaspis bilobata* in the Emu Bay Shale have enabled ontogeny of the species to be established (C. Nedin, University of Adelaide, pers. comm., 1994). Desiccation cracks and black pyritic shale indicate both subaerial exposure and episodes of stagnant, anoxic deposition, but this unit is 75 m higher in the section if the mottled limestone present at Big Gully and Emu Bay is taken as a stratigraphic marker. Furthermore, type Emu Bay Shale is thinner and interbedded with sandstone and conglomerate (Daily, 1956); it is therefore unlikely that the two units correlate. Other, thinner shales occur throughout the succession; these are preserved or eroded depending on the vagaries of alluvial fan migration.

The Boxing Bay Formation is the highest unit in the succession. It is chiefly sandstone but also contains conglomerate beds with abundant gneissic and rare limestone clasts. The sandstone is initially fine to medium-grained with much higher feldspar and mica content than the coarse-grained, grey, calcareous sandstone of the White Point Conglomerate. However, on the eastern side of Point Marsden, conglomerate beds with carbonate clasts identical to those in the underlying formation reappear (A.P. Belperio, SADME, pers. comm., 1990). Palaeocurrent directions are bipolar, approximately east-west for both the White Point Conglomerate and Boxing Bay Formation. According to Daily *et al.* (1980), the Kangaroo Island Group was deposited on a shoreline, along which alluvial fan complexes interfingered with tidal deposits. The interplay between uplift to the north and basin subsidence is considered by those authors to have controlled shoreline migration.

Metasediments on Dudley Peninsula have been assigned to formations of the Kanmantoo Group by Daily and Milnes (1971). Although Flint (1978) suggested correlation of metasediments on the west coast of Kangaroo Island with Tapanappa Formation, the remaining bulk of Kanmantoo Group south and west of the Snelling-Cygnnet Faults (Fig. 7.3) has not been studied.

Sequence set € 3

Upper levels of sequence set € 3 have been eroded in the Stansbury Basin. Milnes *et al.* (1977) suggested that 5 to 10 km of additional sedimentary cover may have existed above the Kanmantoo Group, which hosts the Encounter Bay Granites, to account for the observed metamorphic grade. These authors suggested uplift and erosion soon after granite emplacement. This raises the possibility that equivalents of part of the missing sedimentary cover may be found at depth elsewhere, such as beneath the Permian Troubridge and Tertiary Murray Basins.

Generalised environments of deposition of sequence set € 3 are shown on Figure 7.11.

Upper Kanmantoo Group

The Tapanappa Formation is overlain conformably by the Tunkalilla Formation, at least 250 m thick at its type section, consisting chiefly of phyllite but with medium to coarse-grained metasandstone, 30 m thick, near the base, and thinner, finer grained sandstones higher in the section. Abundant sinuous worm casts are evident in the top bed of a channel filled with sandstone and shale-chip breccia. Blue-black carbonaceous and sulphide-rich phyllites mark the base and top of the formation (Daily and Milnes, 1973). Tunkalilla Formation does not crop out on Dudley Peninsula, and other possible correlatives on Kangaroo Island have not been studied.

The total thickness of the Balquhider Formation, exposed along 18 km of coastline, is not known (Daily and Milnes, 1973). The lower 300 m, however, comprise medium to coarse-grained metasandstone, metasilstone and sulphide-rich phyllite west of Tunk Head. At Bollaparudda Beach, coarse-grained metasandstone fills scour channels, and clay-chip breccias, load casts and climbing ripples have also been described (Daily and Milnes, 1973). Boord (1985) interpreted similar facies east of Tunk Head to have settled from non-channelised turbidity flows in lobe, lobe-fringe and basin-plain settings.

Fine-grained sandstone and siltstone of the Balquhider Formation are overlain by micaceous silty sandstone of the Petrel Cove Formation (Daily and Milnes, 1973). This unit is well laminated to cross-bedded with current and climbing ripples, channels and shale-chip breccias. Concordant granite sheets have been emplaced along bedding planes of this formation on Wright Island (Milnes *et al.*, 1977).

The Middleton Sandstone (Daily and Milnes, 1973) is the uppermost formation of the Kanmantoo Group. No contact has been observed with the underlying Petrel Cove Formation on the mainland or Kangaroo Island. On southern Fleurieu and Dudley Peninsulas it is a fine-grained, quartzose, commonly cross-bedded and current-rippled sandstone with rare, angular shale clasts. Palaeocurrents show consistent sand influx from the west and a relatively shallow marine environment of deposition is evident (Daily and Milnes, 1973; Daily *et al.*, 1979).

Yuruga Formation

The Yuruga Formation (Daily, 1976b, 1990) is the only western Stansbury basin unit assigned to € 3 (Fig. 7.2). A maximum preserved thickness of 548 m is recorded in Stansbury Town 1 drillhole where the formation is overlain disconformably by Permian sediments (Fig. 7.8). Drillcores consist mainly of red-brown, cross-bedded, fine-grained, feldspathic sandstone and siltstone. Desiccation cracks, mud-chip breccias and trilobite tracks occur at several levels (Daily, 1968). Outcrops at Pine Point consist of flat to cross-laminated, fine-grained, pebbly arkosic sandstone and stringers and lenses of subangular gravel and cobbles. Planar cross-beds and ripple laminations are common.

Sedimentary structures seen in drillcore suggest an intertidal origin while outcrops at Pine Point and nearby Rocky Point appear to represent alluvial stream deposits. Clasts include granite,

quartz-feldspar porphyry and laminated red-green mudstone. Subrounded cobbles of dense, buff-weathering dolomite are common at Rocky Point, but limestone and gneiss clasts, typical of Minlaton Formation conglomerates, are absent. The dolomite is not of Kulpara Formation origin as suggested by Crawford (1965) and may be Precambrian.

Sequence set € 4/01

Mafic igneous rocks in drillholes beneath the Murray Basin are presumed on present evidence to correlate with sequence set € 4/01 (Fig. 7.2). They comprise calc-alkaline dolerite and intrusive (or less likely, extrusive) basalt in Nanyah 1 (Preiss and Radke, 1989) and variably deformed metabasic rocks with primitive MORB-like geochemical characteristics in MBT 11, 12, 14 and 15 (Figs 7.3, 7.9). These occurrences comprise Suite 2 of Rankin *et al.* (1991b). Host sediments have not yet been drilled at these localities.

The Late Cambrian age assumed here implies that emplacement or eruption of Suite 2 post-dated Kanmantoo Group deposition. Certain dolerite and amphibolite dykes which intrude the Kanmantoo Group in the eastern Mount Lofty Ranges (Foden *et al.*, 1990; Liu and Fleming, 1990) are also geochemically similar to Suite 2, and recently exposed amphibolite dykes, sympathetically folded with the Angaston Marble (Normanville Group) in Penrice Quarry (J.L. Curtis, C.G. Gatehouse, MESA, pers. comm., 1994), may be related to the same suite. Altered basalt drilled in Hatherleigh 1 beneath the Otway Basin 15 km northwest of Millicent yielded an imprecise Rb-Sr age of 436 ± 150 Ma. An Ordovician age, comparable with post-kinematic granites, is preferred (Foden *in* Sagasco Resources Ltd, 1991).

Clough and Rankin (1990) and Rankin *et al.* (1991b) suggested that Suite 2 mafics follow an arcuate zone of magnetic anomalies into western New South Wales, with possible connection to



Metasandstone of the Balquhider Formation, deposited as storm beds, east of Parsons Beach, southern Fleurieu Peninsula. Photo 042420.

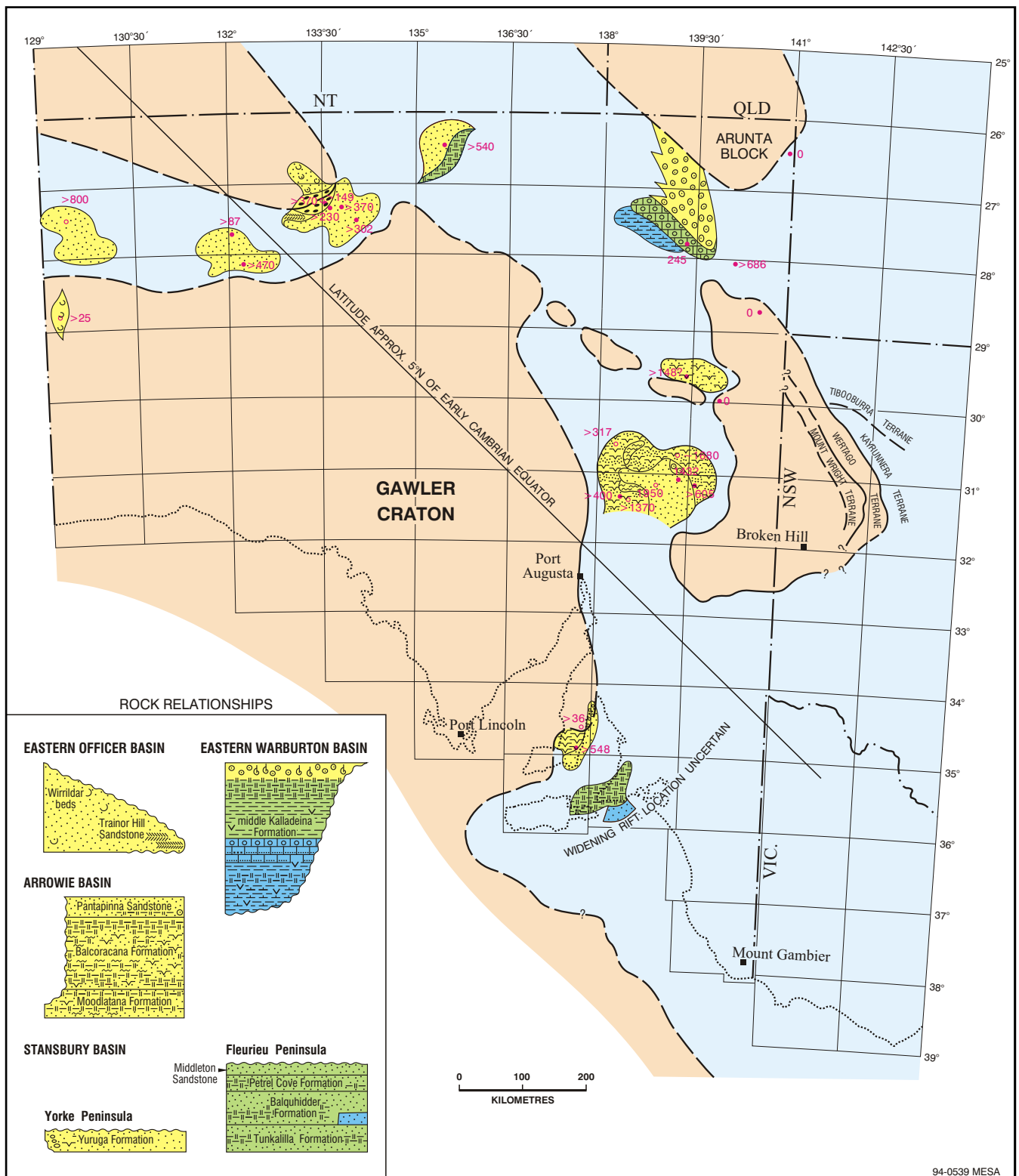


Fig. 7.11 Palaeogeography, sequence set € 3 (excluding Grindstone Range Sandstone).

the Mount Wright Volcanics. However, the latter volcanics are Early Cambrian and it is not clear whether the drilled anomalies are actually on the arcuate magnetic trend. The Mount Wright section is located on one of several 'terrane' which form the Wonominta Block (Leitch *et al.*, 1987; Figs 7.11, 7.12) and further work is required to better understand the structural complexity and relationships of the Murray Basin basement to potentially equivalent rocks in western New South Wales and Victoria.

Despite the uncertain rock relationships, model 1 of Rankin *et al.* (1991b) is favoured here. This involves intrusion or

extrusion of Suite 2 mafics at maximum crustal extension prior to reversal of movement which initiated the Delamerian Orogeny in the Stansbury Basin. However, contemporaneous sediments were not those now preserved as Kanmantoo Group, which are craton-derived and lack evidence of volcanic provenance. Instead, the host may be a younger succession not yet recognised (the 'missing section') above syn-orogenic granitoids (Milnes *et al.*, 1977) beneath the Murray Basin. These hypothetical sediments and the Suite 2 mafics are likely to have been the last deposits in the Adelaide Geosyncline.

ARROWIE BASIN

D.I. Gravestock and W.M. Cowley

Geological setting

Cambrian sedimentary rocks up to 4 km thick crop out in the Flinders Ranges, which intervene between flat-lying to gently folded Cambrian strata at depth to the east and west. These sediments were assigned by Wopfner (1972) to the Arrowie Basin, the western limit of which was placed at the Torrens Hinge Zone. However, Youngs and Moorcroft (1982) included thin, flat-lying Cambrian sediments on the Stuart Shelf. Their concept of the basin embraces three tectono-sedimentary regimes — the stable Stuart Shelf, mobile Adelaide Geosyncline and sub-stable Curnamona Craton. This extended region is adopted here as the Arrowie Basin (Fig. 7.12).

There was depositional continuity during the Cambrian between the Arrowie and Stansbury Basins and with the Gnalta Shelf in western New South Wales (Daily, 1956; Wopfner, 1966, 1972; Cook, 1982). Seaways also extended via the Warburton Basin into the Officer, Amadeus and Georgina Basins, and south to Antarctica.

The subsurface region (Torrens Hinge Zone) between the Flinders Ranges and Stuart Shelf is virtually unexplored. There have been no geophysical surveys of note and, apart from a number of closely spaced shallow petroleum exploration drillholes completed by Santos near Wilkatana, only four other holes have penetrated the Cambrian succession in the western region. East of the ranges, drillholes are widely spaced, but several geophysical surveys have outlined the Cambrian at depth (Callen, 1981; Adams, 1987; Santos Ltd, 1988). The succession deepens gently northward in the Moorowie and Yalkalpo 'Synclines', more correctly termed 'depressions' since they contain thick, unfolded strata. The depressions occur to the west and east respectively of the Benagerie Ridge which comprises a suite of Mesoproterozoic volcanics and sediments (Fig. 7.12). The Moorowie 'Syncline' is bisected by a northerly trending wrench fault complex, subparallel to the Benagerie Ridge, and referred to as the Poontana Fracture Zone (Adams, 1987; Poontana Structure of Coats *et al.*, 1969; Callen, 1981). The Moorowie 'Syncline' coincides loosely with Freeman's (1966) 'Lake Frome Gravity Depression' but it is terminated to the north by shallow basement which links the Benagerie Ridge and Mount Painter Inlier.

Mineral resources summarised by Johns (1968a) include Cambrian-hosted zinc, lead and silver, and low-grade copper and manganese which have been mined in the past. Since the 1957 Wilkatana drilling program which resulted in oil 'shows' in Early Cambrian limestone (Wopfner, 1970a), petroleum exploration has focused on the 12 300 km² Moorowie 'Syncline'. Beyond minor traces of oil and gas, no significant discoveries have been made.

Early Cambrian limestone on the Stuart Shelf has been investigated for lead, zinc and copper, but no economic prospects have been discovered, apart from copper mined at the turn of the century near the northern margin of Lake Torrens (Thomson, 1965; Johns, 1968a, 1972). Opal workings in bleached Cambrian shale near Yarra Wurta Cliff in the same vicinity have produced only potch (Barnes and Scott, 1979).

Fossil discoveries in the early part of the century (Etheridge, 1905, 1919; Howchin, 1907, 1922; Taylor, 1910) were followed by systematic description of the Cambrian in the Flinders Ranges (Mawson, 1939; Segnit, 1939; Daily, 1956; Dalgarno, 1964). Richly fossiliferous carbonates of the Hawker Group (Dalgarno, 1964) yielded archaeocyaths of exceptional variety which resulted in several monographs (Taylor, 1910; Bedford, R. and W.R., 1934, 1936; Bedford, R. and J., 1936, 1937, 1939; Debrenne, 1969, 1970, 1973, 1974a,b), but these fossils have been studied in stratigraphic order only comparatively recently (Walter, 1967; Gravestock, 1984). Daily's (1956) biostratigraphic correlation based on faunal assemblages in the carbonates has been greatly improved by detailed descriptions of trilobites and shelly fossils (Bengtson *et al.*, 1990).

The Arrowie Basin was viewed by Wopfner (1969a) and others as a central mobile zone of deeper water deposits flanked to the east and west by shallow marine sediments on the stable Stuart Shelf and Curnamona Craton. In an important departure, Haslett (1975) pointed out that deeper water facies predominated north of a hinge line connecting the Blinman, Wirrealpa and Chambers Diapirs. Following Haslett's example, Gravestock (1988) suggested a more arcuate hinge curving north into the Warburton Basin. In the Flinders Ranges it follows a line between the Lyndhurst, Beltana, Wirrealpa and Chambers Diapirs, and is named the Wirrealpa Hill Hinge (WHH) in this volume (Fig. 7.12).

Deeper water sediments probably extended north beyond the present Arrowie Basin limit (Mulloorina Gravity Ridge, Fig. 7.12) to the Warburton Basin and south to the Stansbury Basin. The palaeoslopes inferred from incised canyons in the Wilpena Group (Coats, 1964a; von der Borch *et al.*, 1982; Gehling, 1983) support this model and suggest that the saddle-shaped geometry of the Arrowie Basin was inherited from an older configuration (Gravestock and Hibburt, 1991).

Sequence set € 1

Three depositional sequences (€ 1.1 to € 1.3) constitute this set. The lowest sequence, € 1.1, was not deposited on the Stuart Shelf but is of substantial thickness north of the WHH and comprises both siliciclastic and carbonate sediments (Fig. 7.5). Mount and McDonald (1992) further subdivided € 1.1 into two parts, the Uratanna and Parachilna Sequences. They tentatively placed the Neoproterozoic-Cambrian boundary within the lower part of the Uratanna Sequence, based on the first appearance of organic-walled sabelliditids and the trace fossil *Phycodes coronatum*. This volume follows other workers who regard the base of the Uratanna Formation as the Precambrian-Cambrian boundary (Daily, 1972, 1973, 1976a,b; Gauld, 1976; Mount, 1990; McDonald and Mount, 1991; Nedin and Jenkins, 1991).

Sequence € 1.1 deposits represent shoreline shallow marine shelf and ramp environments. Local subsidence created intra-shelf depressions but there was no pronounced shelf-slope break. Sequence 1 of James and Gravestock (1990) corresponds to sequence € 1.1 which is represented by the Uratanna Formation (Daily, 1973) and the lower Hawker Group (Dalgarno, 1964).

The carbonate ramp evolved rapidly into distinct shelf, slope and basinal settings during the time represented by sequence € 1.2. The WHH more or less separated the shelf from slope and basinal carbonates further north. However, there are exceptions where the shelf-break was further south (Donkey Bore Syncline, Wilkawillina Gorge), suggesting the formation of local environments.

Platform carbonates were grouped by James and Gravestock (1990) into a single sequence, but Clarke (1990b) recognised two (sequences € 1.2 and € 1.3 herein); his 'lower platform sequence' corresponds to sequence € 1.2.

Sequence € 1.3 was deposited during an epoch of pronounced tectonic activity accompanied by increasing volcanism, which affected mainly the late highstand deposits. Sequence € 1.3 corresponds to upper sequence 2 of James and Gravestock (1990) and Clarke's (1990c) 'upper platform sequence'.

Sequence € 1.1 lowstand deposits

The Uratanna Formation disconformably overlies Rawnsley Quartzite as erosional channel fill up to 100 m thick southwest and west of the WHH. It comprises mudclast-rich, medium-grained to pebbly quartz sandstone with local evidence of channel reoccupation (Hull, 1973). More widespread green siltstone and shale with thin, mudcracked sandstone and limestone interbeds comprise middle and upper members up to 500 m thick north of the WHH (Coats, 1973a; Gauld, 1976), although only 16 m of the upper member are present in the eastern Nepabunna Syncline (Mann, 1981). *Phycodes*, *Skolithos*, *Planolites*, *Rusophycus* and *Didymaulichnus*

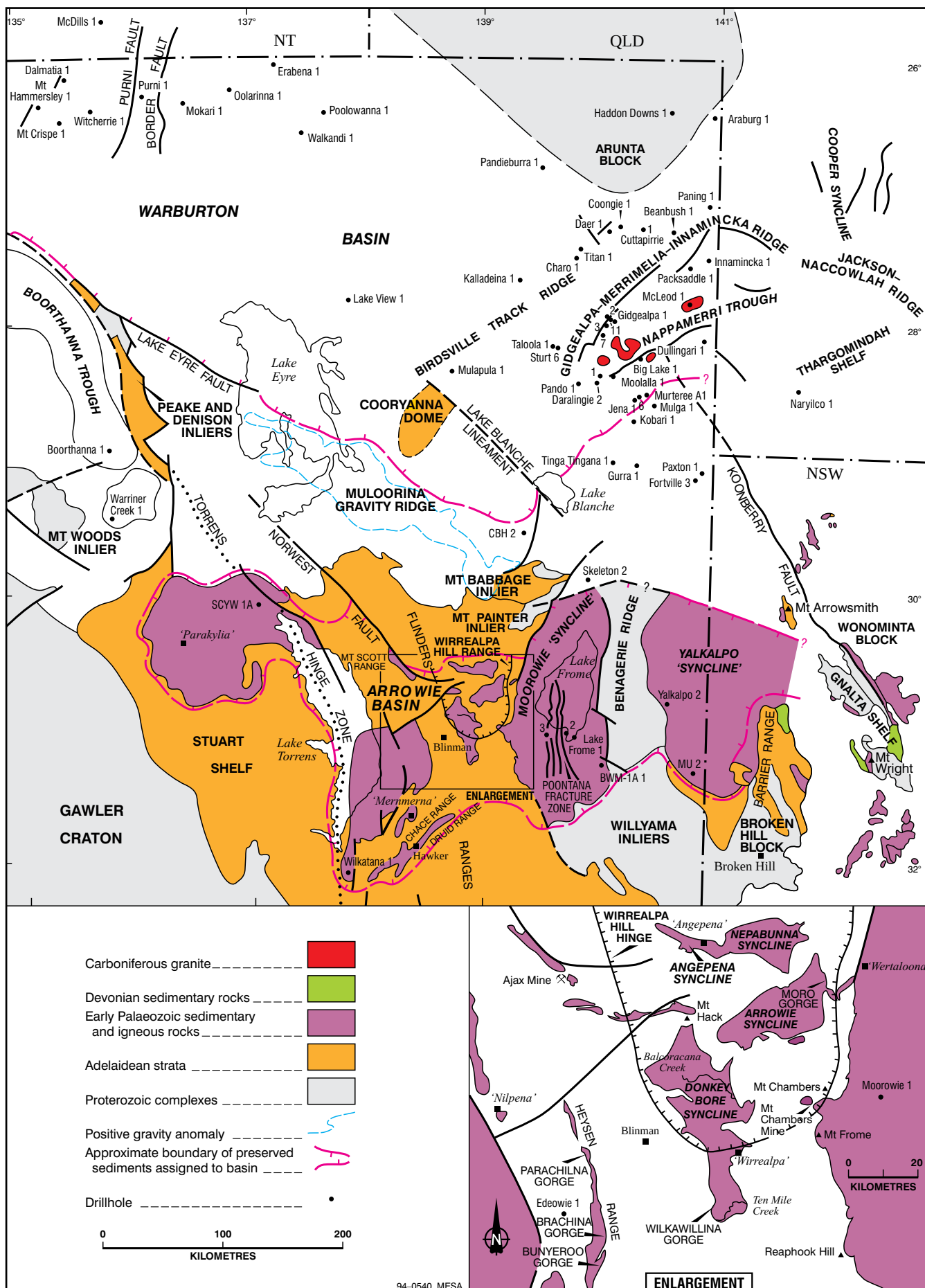


Fig. 7.12 Arrowie and eastern Warburton Basins location map.

are representative trace fossils (Daily, 1972, 1976c; Gauld, 1976). An upward-deepening marine succession from initially fluvial conditions has been interpreted by McDonald and Mount (1991), who also recorded the first 'shelly' fossil *Sabellidites cambriensis*.

Where Uratanna Formation is absent, siliciclastics of the Parachilna Formation (Dalgarno, 1962) disconformably overlie Rawnsley Quartzite and, in the absence of both units, younger carbonates onlap elevated Pound Subgroup locally, south of the WHH (e.g. Wilkawillina Gorge). Nowhere do carbonates rest directly on Uratanna Formation. The Parachilna Formation is widespread; it has been recorded in Wilkatana drillholes in the southwestern Arrowie Basin (Thomson, 1962) and has a maximum known thickness of 570 m in the Nepabunna Syncline (Mann, 1981), where its upper part may belong in the transgressive systems tract (Fig. 7.2). There is evidence of minor erosion between the Uratanna and Parachilna Formations (Daily, 1973; Hull, 1973; Gauld, 1976).

The Parachilna Formation is an upward-fining suite of sandstone and siltstone with minor carbonate interbeds (Ford, 1971; Gauld, 1976). 'Pipe rock' typifies basal units due to numerous U-shaped dwelling burrows of *Diplocraterion*. Ripple marks, thin brown shale laminae and desiccation cracks are common in the lower, bioturbated beds. Upper, fine-grained units lack mudcracks north of the WHH where shallow subtidal conditions prevailed. Sandy *Diplocraterion*-burrowed units were deposited in low to moderate energy environments. Other trace fossils and a single mould of the gastropod *Bemella* have been recorded (Daily, 1976b,c). Ford (1971) suggested the Stuart Shelf Adelaidean as a source of Parachilna Formation clastics.

The Ajax and Wilkawillina Limestones (Daily, 1956) conformably overlie Parachilna Formation south of the WHH, commencing a phase of extensive carbonate sedimentation. Basal beds, 25 to 100 m thick, are stromatolitic and oolitic with stringers of coarse-grained quartz sand. The first archaeocyaths occur at Wilkawillina Gorge (Faunal Assemblage 1) in bioclastic packstone, and in small bioherms with *Renalcis*. The equivalent Ajax Limestone is dolomitised but also comprises skeletal packstone with associated oolite.

Stromatolites, oolite, mudcracked dolomitic shale and abundant quartz-sand interbeds comprise the laterally equivalent Woodendinna Dolomite (Haslett, 1975). Lithoclast gravel trains occur near the Wirrealpa Diapir, which was active at this time. The Woodendinna Dolomite, which is usually 90 to 180 m thick, exceeds 400 m in actively subsiding grabens (e.g. Erina Waters), but thins to zero north of the WHH. Sediments were deposited on subtidal to subaerially exposed mudflats and formed ooid shoals on the marine shelf.

Sequence € 1.1 transgressive deposits

The basal transgressive surface of the overlying systems tract can be recognised locally. It coincides with the base of a mottled member of the Wilkawillina Limestone at Moro Gorge, and at the type locality (Hideaway Well Member of Clarke, 1986a) south of the WHH. Submarine erosion is evident at Wilkawillina Gorge where the basal transgressive surface is found within upper Faunal Assemblage 1. Archaeocyaths indicate a mid- to late Atdabanian age (Gravestock, 1984). To the north, transgression coincides with the base of the Wirrapowie Limestone (Haslett, 1976a), and is possibly in the upper Parachilna Formation in the Nepabunna Syncline.

The Woodendinna Dolomite is overlain north of the WHH by dark grey, laminated lime mudstone with thin tongues of cross-bedded oolite, and numerous stromatolite, thrombolite and *Renalcis*-archaeocyath bioherms of the Wirrapowie Limestone (Haslett, 1975, 1976a,b). Storm breccias are common but mudcracks and siliciclastics are rare. The Wirrapowie Limestone overlies Parachilna Formation erosively in the Puttapa Syncline (Hull, 1973, his 'Beltana Dolomite') and with apparent conformity in the Nepabunna Syncline (Mann, 1981) beyond the northern depositional limit of the Woodendinna Dolomite. In the vicinity of the hinge zone, Wirrapowie and Wilkawillina Limestones intertongue, while the Wilkawillina and Ajax Limestones dominate

on the shallow shelf. However, Wirrapowie Limestone is also predominant in the southwestern Arrowie Basin, for example in the Chace and Druid Ranges, at 'Nilpena' and near 'Mernmerna'.

The lower Ajax and Wilkawillina Limestone (Hideaway Well Member) transgressive deposits are lime mudstone, wackestone and isolated archaeocyath-*Renalcis* bioherms. Deeper water archaeocyath-sponge buildups are flanked by mottled lime mudstone in intra-shelf depressions (James and Gravestock, 1990). These buildups contain the world's oldest encrusting lithistid sponges (Reitner, 1991). In northern synclines the transgressive deposits also include archaeocyath-calcimicrobe bioherm complexes of the Wirrapowie Limestone and basal Mernmerna Formation (Haslett, 1975; Mann, 1981). The name Mernmerna Formation (Dalgarno and Johnson, 1962) replaces Parara Limestone in the Arrowie Basin, the Parara now being restricted to the western Stansbury Basin. The archaeocyaths are from lower Faunal Assemblage 2 and are late Atdabanian in age (Gravestock, 1984). The conformable basal Parara Limestone on southern Yorke Peninsula is also in lower Faunal Assemblage 2 and marks the transgressive phase of sequence € 1.1 in that region.

Sequence € 1.1 highstand deposits

In the vicinity of the WHH, highstand deposits are high-energy bioclastic grainstones of the Ajax and Wilkawillina Limestones. Archaeocyaths and other shelly fossils, including the oldest Australian trilobites (upper Faunal Assemblage 2) are in these beds. Bioherms up to a few tens of metres across form isolated buildups within the grainstones. Small bioherms in the Mount Scott Range are dominated by *Epiphyton* (James and Gravestock, 1990). Further south (Chace Range, 'Mernmerna'), however, mottled *Renalcis*-rich bioherms coalesce to form mound complexes (Toteff, 1971, his 'Chace Formation'). Distribution of this facies spectrum is shown on Figure 7.5.

At Old Wirrealpa Spring and along the Bunkers Range west of 'Wirrealpa', fault activity produced deep fractures which penetrated the lithified grainstones for 100 m or more, and were filled with marine fibrous calcite cement, fossil debris (usually *Micrina* sclerites) and quartz sand (Haslett, 1969; Pierce, 1969; Daily, 1976b; Clarke, 1986b, 1990b; James and Gravestock, 1990).

North of the WHH (Wirrealpa Hill, Puttapa Syncline, eastern Arrowie Syncline), thin tongues of archaeocyathan bioclastic 'hash' were transported into deeper water from the adjacent carbonate shelf where high-energy conditions prevailed. Deeper water conditions appear to have been rapidly established in the Yalkalpo 'Syncline' during early highstand. In Yalkalpo 2 drillcore, muddy archaeocyath bioherms with minor calcimicrobes are enclosed in rhythmically alternating nodular wackestone and black laminated lime mud. Sequence € 1.1 has not yet been drilled in the Moorowie 'Syncline' where the same deeper water conditions are assumed to have been present. Both synclines probably formed a single embayment at this time since the Benagerie Ridge had not formed a barrier to sedimentation (Fig. 7.5).



Bedding surface of *Diplocraterion* burrows, Parachilna Formation, ~7 km south of Mount Aleck. Photo 042421.

Sequence € 1.2 lowstand deposits

South of the WHH, much of the lower Wilkawillina Limestone was exposed as sea-level fall outstripped subsidence in all but a few localities. Bioclastic calcarenites with high sedimentation rate were particularly affected in a broad swathe from the Heysen Range to Reaphook Hill. The Stuart Shelf was still not covered at this stage. Exposure is recorded as far east as MU 2 drillhole in the southern Yalkalpo 'Syncline', and carbonates presumed to have been deposited marginally to the Torrens Hinge Zone would have been affected as well (Fig. 7.5).

Limestone at the exposure surface is solution sculptured. Penetrating the limestone at depths ranging from centimetres to tens of metres, solution-enlarged fractures, karst collapse breccias and possible calcrete profiles are variably developed (Daily, 1976b; Clarke, 1986a,c; Horn and Morris, 1988). The surface is capped by a conspicuous red crust, up to 0.1 m thick, composed of haematite-rich laminae which in places form small, digitate stromatolites (Haslett, 1969; Clarke, 1986a, 1988). These grew on the hard limy substrate at the top of the lower Wilkawillina Limestone. In MU 2 this surface is within a few centimetres of the disconformably underlying Precambrian Brachina Formation. Adelaidean strata are likely to have been exposed locally in the southern part of the Curnamona Craton at this time.

During the hiatus there was little or no siliciclastic influx into the Arrowie Basin. Lowstand deposits are thus virtually non-existent but, at Old Wirrealpa Spring, several large blocks of lower Wilkawillina Limestone rest on the disconformity surface and probably represent scarp debris (Haslett, 1976a).

The lower Wilkawillina Limestone is overlain disconformably by either lower Mernmerna Formation or upper Wilkawillina Limestone immediately south of the WHH (Flinders Unconformity of James and Gravestock, 1990). A break in the fossil record is evident here as on northern Yorke Peninsula (Fig. 7.2). From Reaphook Hill almost to Mount Frome, the lower Wilkawillina Limestone is overlain disconformably by the Bendieuta Formation of sequence € 1.2.

A much longer break is evident between Bunyerroo Gorge and Parachilna Gorge, and at Balcoracana Creek, where the Edeowie Limestone Member of the Oraparinna Shale (Moore, 1979a) onlaps exposed lower Wilkawillina Limestone. In such regions, the carbonate shelf remained exposed to meteoric diagenesis while sequences € 1.2 and € 1.3 were being deposited elsewhere.

Sequence € 1.2 transgressive deposits

Transgressive deposits representing this sequence are shown on Figure 7.6.

A thin wedge of Wilkawillina Limestone, known only from Wirrealpa Mine and Wilkawillina Gorge (Second Plain Creek Member of Clarke, 1986c), overlies the red crust which was formed during initial drowning of the shelf (J.L. Curtis, pers. comm., 1991). The unit comprises grey laminated lime mudstone up to 74 m thick, with thin bioclastic 'hash' beds and phosphate-coated omission surfaces. Large calcite-filled irregular fenestrae are common at Wirrealpa Mine. The depositional setting was initially shallow subtidal, deepening upward rapidly, with evidence of unstable slope conditions in upper beds at Wilkawillina Gorge.

Transgressive deposits on the carbonate shelf are condensed (10-20 m) sections in the mid-Ajax and Wilkawillina Limestones. Sediments are red and glauconitic (Ajax) or grey (Wilkawillina), thinly bedded, abundantly fossiliferous packstone and wackestone. Fossils are mainly small molluscs, hyoliths, brachiopods and *Chancelloria*, with less abundant archaeocyaths and trilobites of the *Abadiella huoi* and *Pararaia tatei* Zones (Jell in Bengtson *et al.*, 1990). Localised thicker Ajax Limestone in the Mount Scott Range is grey, sparsely fossiliferous wackestone.



Multiple reddened horizons of microstromatolites on disconformity at top of sequence € 1.1, lower Wilkawillina Limestone, ~7 km south of Mount Aleck. Photo 042422.

Proximal slope carbonates (lower Mernmerna Formation) intertongue with adjacent platform bioherm complexes at Ten Mile Creek (Clarke, 1988, 1990a). Distal slope carbonates onlap the sequence boundary at Wilkawillina Gorge, in Donkey Bore Syncline, and between Mount Chambers and Mount Frome. South of the WHH along the Heysen Range, condensed Mernmerna Formation partly onlaps the carbonate shelf.

In contrast, the lower Mernmerna Formation conformably overlies *Renalcis*-rich bioherms or mottled Wirrapowie Limestone at 'Mernmerna', near Mount Aleck and in the Chace Range. There is insufficient information at these localities to distinguish transgressive and highstand facies at present. In this part of the Arrowie Basin, the Mernmerna Formation is unusually thick (up to 600 m), and is suggested by Gravestock and Hibburt (1991) to indicate a palaeoslope towards the southeast (i.e. towards the Stansbury Basin). Dalgarno (1964) also recognised a depocentre in this region.

Flaggy lower Mernmerna Formation, 300-650 m thick north of the WHH and at Wilkawillina Gorge (Six Mile Bore Member and part of the Linns Springs Member of Clarke, 1986b), comprises turbidite couplets of lime silt and mud. Intraformational slumping is common; debris flows, peloid grain flows and platform-derived lithoclasts are rare. Trilobites in these slope carbonates are from the *Abadiella huoi* and *Pararaia tatei* Zones (including Daily's (1956) Assemblages 3 and 4); sponge spicules and echinoderm plates are common (Clarke, 1990a). Lower Mernmerna Formation turbidites in the northern synclines (Angepena, Nepabunna, Arrowie) conformably overlie Wirrapowie Limestone and intertongue with the Midwerta Shale and Nepabunna Siltstone (Leeson, 1967; Coats, 1973a; Mann, 1981). These latter units are green-grey to black, argillaceous, pyritic, manganiferous limestones which comprise the deepest water sediments of sequence € 1.2. Fossils are rare but Mann (1981) reported *Thambetolepis* (Faunal Assemblage 3?) from immediately below Midwerta Shale.

There is no evidence of shoaling at the sequence € 1.1 upper boundary in Yalkalpo 2. The Mernmerna Formation, 97 m thick in this drillhole, comprises nodular wackestone alternating with black laminated lime mud, representing a dysaerobic outer shelf environment. Upward increase in the proportion of lime mud may indicate passage from transgressive to early highstand conditions in this region.

Sequence € 1.2 highstand deposits

Early highstand deposits of sequence € 1.2 are thought to have spread to the Stuart Shelf for the first time (Fig. 7.6). Alternatively, Cowley (1990) considered basal units to represent part of sequence € 1.1. The base of the Cambrian on the Stuart Shelf is locally disconformable but is regionally a low-angle unconformity. Due to slight tilting and erosion of the Adelaidean succession, the

Cambrian Andamooka Limestone (Johns, 1968a) rests on Yarloo Shale in the north, then on successively older members of the Tent Hill Formation. Basal beds of the Andamooka Limestone are fine to coarse-grained, locally bouldery siliciclastics and carbonates which pass up rapidly into widespread ooid grainstones.

The upper Ajax and upper Wilkawillina Limestones around the WHH are shallow marine shelf deposits. At Wirrealpa Mine, sandy oncolite rudstone is overlain by quartz-sand-rich archaeocyath floatstone with bedding-parallel and solution-enlarged stromatactoid cavities (James and Gravestock, 1990). Ooid grainstone is common at Ten Mile Creek (Clarke, 1990c). In the eastern Flinders Ranges, highstand shelf deposits are represented by the Bendieuta Formation which conformably overlies lower Mernmerna Formation between Mount Frome and Mount Chambers Mine (Hatcher, 1970; Mount, 1970; Wigglesworth, 1970), but disconformably overlies lower Wilkawillina Limestone at Reaphook Hill to the south ('Reaphook Limestone' of Gehling, 1971). The Bendieuta Formation, up to 238 m thick, is planar to trough cross-bedded, coarse-grained, sparingly fossiliferous quartz sandstone, ooid-peloid grainstone and fenestral limestone deposited in a shallow subtidal setting. Intensively burrowed calcareous sandstones form distinctive marker beds in the middle part of the formation.

Fossiliferous ooid-peloid packstone of the Bendieuta Formation comprises the basal 16 m of the Cambrian succession in the southern Moorowie 'Syncline' (BWM1A 1 drillhole). In the Yalkalpo 'Syncline', muddy nodular Mernmerna Formation indicates continued deeper marine conditions. Several thin tuff bands occur in Yalkalpo 2 drillcore.

The middle Mernmerna Formation at Wilkawillina Gorge (Third Plain Creek Member of Clarke, 1986b), in Donkey Bore Syncline and north of the WHH, differs in detail from the transgressive lower Mernmerna Formation beneath. The background facies is still dark grey laminated lime silt and mud, but grain-flow deposits are common. These consist chiefly of ooids, calcified microbe peloids, abraded fossils and well-rounded coarse to very coarse-grained quartz sand. Intraformational slumps are frequent on the northward-deepening palaeoslope and blocks of lithified platform limestone up to 15 m across occur in the Donkey Bore Syncline. The transported boulders are composed of ooid grainstone and masses of *Renalcis*-archaeocyath 'reef rock'. The particulate allochthonous debris contains archaeocyaths contemporaneous with those on the platform. The *Pararaia bunyeroensis* Zone occurs in upper levels of the middle Mernmerna Formation at Wilkawillina Gorge (Jell in Bengtson *et al.*, 1990).

Sandy debris beds and clasts up to 50 mm in size persist as far north as the Arrowie Syncline. This widespread appearance of allochthonous shelf debris reflects northward progradation of the carbonate shelf during gradual sea-level fall.



Slump-folded flaggy Mernmerna Formation limestone; a north-deepening palaeoslope is indicated. Donkey well Syncline near 'Wirrealpa'. Photo 042423.

Sequence € 1.3 lowstand deposits

A second phase of subaerial exposure marks the sequence boundary on the carbonate shelf. Associated diagenetic effects are difficult to distinguish from older and younger phases of exposure which are superimposed at several localities. However, solution enlargement of stromatactis-like cavities probably occurred at this time in the Mount Scott Range and at Wirrealpa Mine. Ooid grainstone is truncated by an iron-oxide-stained karst horizon at Ten Mile Creek (Clarke's, 1990c, K2 horizon), while reddened pisolites at the top of the Bendieuta Formation at Reaphook Hill probably formed by subaerial weathering during the € 1.3 lowstand (Gehling, 1971). Dolomitisation has obscured the sequence boundary within the Andamooka Limestone on the Stuart Shelf.

Carbonate supply was effectively starved on the shelf but, in contrast to the sequence € 1.2 lowstand, quartz sand was transported over the shelf edge and extended basinward as far north as the Arrowie Syncline (Fig. 7.7). Clarke (1990b) first recognised the lowstand nature of the Bunkers Sandstone (Daily, 1956), which thins from 200 m adjacent to the shelf at Wilkawillina Gorge to 15 m in the Arrowie Syncline.

The basal Bunkers Sandstone is flat to ripple-laminated, silty, fine-grained quartz sandstone which overlies the Mernmerna Formation with a sharp, locally disconformable contact. The overlying unit, restricted to Wilkawillina Gorge and comprising the main part of the formation, is planar to trough cross-bedded quartz sandstone. Flat to cross-bedded silty, fine-grained sandstone predominates in the Donkey Bore and Arrowie Synclines. Trace fossils occur at several levels but body fossils are extremely rare. Bipolar current directions at Wilkawillina Gorge were interpreted by Clarke (1990b) to result from tidal circulation in an embayment opening to the north.

The Curdlawidny Siltstone Member of the Andamooka Limestone crops out near 'Parakylia' on the Stuart Shelf (Cowley, 1990). It is proposed tentatively here as a lowstand correlative of Bunkers Sandstone but could alternatively be a late highstand deposit of sequence € 1.2. The member is up to 22 m thick, comprising planar and ripple-bedded siltstone and fine-grained sandstone, locally with shale clasts and mudcracks; lower parts of the unit are dolomitic. Boundary relationships with the enclosing Andamooka Limestone are not clear.

Sequence € 1.3 transgressive deposits

Transgressive deposits on the Stuart Shelf are best represented by the Andamooka Limestone in SCYW 1A drillcore. These comprise peritidal oolite, stromatolites and shallow subtidal *Renalcis* boundstone, a middle succession of burrowed fossil wackestone, and upper levels comprising buildups of fenestral spiculitic wackestone with archaeocyaths and calcareous sponge-like organisms (James and Gravestock, 1990).

Platform carbonate of the upper Wilkawillina Limestone bordering the WHH is composed of archaeocyath-calcimicrobe buildups. These occur both as small isolated bioherms less than 1 m across and as coalescing bioherm complexes (reefs) up to hundreds of metres in length and tens of metres thick. The relative abundance of calcimicrobes (*Renalcis*, *Epiphyton*, *Girvanella*) and archaeocyaths depends on the location of buildups on the shelf and the energy of the environment (James and Gravestock, 1990). Archaeocyaths at Wirrealpa Mine, Ten Mile Creek (Clarke's, 1990c, 'lower bioherm complex') and the classic Ajax Mine locality flourished at this time.

The Ajax and Wirrealpa Mine reefs were of the flat-topped 'keep-up' type (terminology of Kendall and Schlager, 1981) and were not covered by deeper water deposits. The lower bioherm complex at Ten Mile Creek was a rimmed 'catch-up' type backed by a lagoon (Clarke, 1990c) and is interesting because its retreating rim managed to build to sea level. A series of 'reddened pockets' on the reef top represents Clarke's (1990c) K3 karst horizon. The thickness of this reef (181 m) provides an estimate of maximum

water depth attained during transgression. True water depth, taking subsidence into account, was probably much less; Clarke suggested somewhat less than 100 m.

Prior to growth of this reef at Ten Mile Creek, however, silty and calcareous facies of the upper Bunkers Sandstone lapped on to karsted sequence €1.2 grainstone. Sandy limestone breccia, channels near Wirrealpa Mine (Haslett, 1969), and sedimentary dykes along the Bunkers Range (Pierce, 1969), all formed during lowstand, were preserved beneath onlapping upper Bunkers or beneath reef sediments as sea level rose.

With continued transgression, the lower Oraparinna Shale (Daily, 1956) was deposited conformably on Bunkers Sandstone at Ten Mile Creek. To the north (Bunkers Range), it overstepped Bunkers Sandstone and lapped onto the lower Mernmerna Formation of sequence €1.2 (Dalgarno, 1964). To the southeast at Ten Mile Creek, the lower bioherm complex gradually retreated, but kept up with transgressing Oraparinna Shale, though little sediment was transported off-reef. Trilobites of the *Pararaia janeae* Zone occur in the Oraparinna Shale (Jell in Bengtson *et al.*, 1990). The lower Oraparinna Shale is khaki, sparingly fossiliferous and silty, with common limestone concretions. It is thickest at Wilkawillina Gorge (~200 m) but thins rapidly away from that locality.

The upper Mernmerna Formation is the main transgressive unit beyond the shelf edge. Thickness varies from 190 m in Donkey Bore Syncline to 700 m or more in the Arrowie Syncline, although some of the latter may be a highstand deposit. The upper Mernmerna Formation overlies Bunkers Sandstone conformably in these regions, but intertongues with, and is replaced by, Oraparinna Shale at Wilkawillina Gorge.

The growing reef at Wirrealpa Mine supplied the adjacent slope intermittently with bioclastic debris. This occurs as planar to cross-bedded calcarenite and rudstone, often with well-rounded coarse-grained quartz sand. Such beds are numerous in the upper Mernmerna Formation but are rarely more than 0.4 m thick, and contrast with the background muddy limestone deposits.

In the eastern Flinders Ranges, the upper Mernmerna Formation, up to 60 m thick, onlapped Bendieuta Formation from north to south with a sharp basal contact (Hatcher, 1970; Mount, 1970; Wigglesworth, 1970), but did not reach Reaphook Hill (Gehling, 1971). The unit is fossiliferous dark grey, flaggy to mottled argillaceous limestone with common coarse-grained quartz sand and occasional dolomitised pebbles near the base.

In the Moorowie and Yalkalpo 'Synclines', drillcores from BWM1A 1, Yalkalpo 2 and MU 2 all have sharp boundaries marked by conglomerate or cut-and-fill structures, interpreted here as submarine erosion surfaces at the boundary between sequences €1.2 and €1.3. The uppermost 68 m of Cambrian core in MU 2, 27 m in BWM1A 1, and a mere 6 m remnant in Yalkalpo 2, comprise the transgressive Mernmerna of sequence €1.3. Sediments are nodular lime mudstone, siliceous shale and thin oolite interbedded in all three drillholes with green tuffaceous sediments 0.05 to 0.15 m thick. In BWM1A 1, a further 19 m of ripple-laminated shale with thin calcarenite cross-beds may represent transgressive or highstand deposits of the Oraparinna Shale.

Sequence €1.3 highstand deposits

A final phase of reef growth around the WHH, followed by widespread regression ending carbonate sedimentation, resulted from this highstand. Contemporaneous tectonic activity and volcanism have left a distinctive stratigraphic record which is discussed separately below.

The upper Andamooka Limestone on the Stuart Shelf was initially shallow marine to supratidal, fenestral, mudcracked and bioclastic limestone overlain by bioherms which crop out on the north shore of Lake Torrens as low, flat-topped 'biscuits' 10 to 80 m diameter. The bioherms are a complex intergrowth of *Renalcis* and *Botomaella* with rare, dwarfed archaeocyaths and thrombolite-like

clotted fabric (James and Gravestock, 1990). They are overlain sharply by buff, dolomitised stromatolites at the top of the Andamooka Limestone.

In the Mount Scott Range, 60 m of dolomitised, peritidal, platy algal boundstone and stromatolites of the upper Ajax Limestone overlie syntectonic breccia and conglomerate (see below). The corresponding section at the Ajax Mine has been overthrust and altered to dolomite which hosts the Beltana willemite orebodies (Horn, 1975). Mineralised black shale intersected beneath the main orebody may be a tongue of Oraparinna Shale. The top 90 m of Wilkawillina Limestone at Wirrealpa Mine are oolite, fenestral limestone, a solitary bioherm of Andamooka type, and 10-20 m of buff, dolomitised stromatolites identical to those in the uppermost Andamooka and Ajax Limestones (James and Gravestock, 1990).

Small (1-2 m) buildups, surrounded by Oraparinna Shale, grew at Ten Mile Creek during the transgressive maximum between the lower and middle bioherm complexes. They are rich in archaeocyaths (Walter, 1967) and contain rare specimens of the oldest known tabulate corals (Lafuste *et al.*, 1988). Clarke's (1990c) middle and upper bioherm complexes at Ten Mile Creek on the platform margin prograded over Oraparinna Shale and were backed by quiet-water lagoonal wackestone and lime mudstone, the latter with evaporite pseudomorphs.

The Moorowie Formation (Mount, 1970) records a new phase of reef growth in the eastern Flinders Ranges at this time. Outcrops between Mount Frome and Chambers Gorge form a prograding facies complex of nearshore shale and siltstone, shelf margin oolite and reef limestone, all breached by high-energy erosional channels. Transported reef blocks which occur near Moorowie Mine were constructed by tabulate corals, archaeocyaths and calcimicrobes of late Middle Botomian age (Lafuste *et al.*, 1991; Savarese *et al.*, 1993; Fuller and Jenkins, 1994).

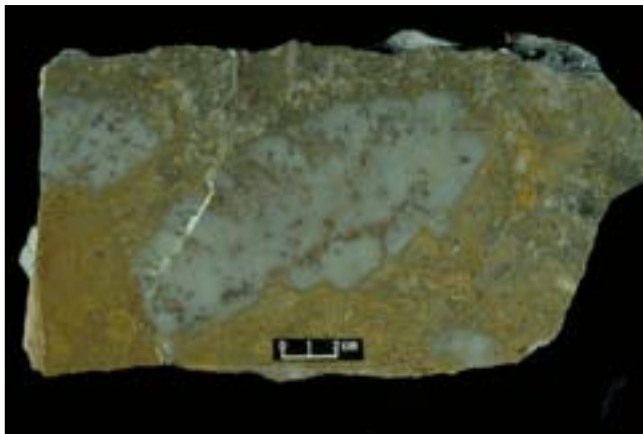
Early highstand carbonates of the uppermost Mernmerna Formation underlie the Moorowie Formation and overlie transgressive, flaggy, silty limestone at Chambers Gorge. Slump-folded, graded, silty limestone with abundant platform megaclasts up to 20 m across typifies these highstand slope sediments (Mount, 1970). A peloidal limestone immediately beneath may represent the basinward limit of Bendieuta Formation shelf-edge shoals which were rapidly drowned at maximum transgression and finally covered by debris from the advancing platform. A 0.6 m thick tuff bed has been preserved in the slope setting (Mount, 1970).

Beyond the platform margin, the Mernmerna Formation is replaced by calcareous upper Oraparinna Shale which becomes less limy and increasingly silty north of the WHH. The youngest unit in this region is Narina Greywacke (Dalgarno, 1964) which is 540 m thick at its type locality. Due to faulting, nowhere is its contact with overlying units seen in outcrop, and no subsurface occurrences are known. The Narina Greywacke intertongues with the Oraparinna Shale on decimetre scale over a passage of 100 m and comprises khaki, flat to ripple-laminated, silty to medium-grained feldspathic sandstone. Several distinct olive-green units contain coarse-grained quartz phenocrysts reworked from crystal tuff. Volcanic wacke was originally described by Chebotarev (1959), and Coats (*in* Dalgarno, 1964) mapped a tuff near 'Angepena'. Much of the detrital material in the Narina Greywacke is volcanoclastic. One coarse-grained tuffaceous bed at the type section contains several angular chips of peloid and archaeocyathan limestone, clearly derived from the carbonate shelf. Basal bedding contacts in the Narina Greywacke are sharp and erosional, or loaded into finer grained material beneath. Reworked shale clasts, limestone chips and coarse, well-rounded quartz granules appear sporadically, but carbonate interbeds are rare. The shallow marine environment was one of moderate energy, interrupted at times by high-energy influxes of terrigenous and volcanoclastic detritus.

The calcareous upper Oraparinna Shale is widespread south and southwest of the WHH, where the Narina Greywacke was not deposited, but is very thin (<20 m) or even absent in the Bunkers Range and near Brachina Gorge.

The Edeowie Limestone Member of the Oraparinna Shale is a thin marker horizon of platy carbonate south of the WHH (Dalgarno and Johnson, 1962; Moore, 1979a). Outcrops further north have been eroded to lower stratigraphic levels, thus the original extent of the member is unknown. The member overlies silty, calcareous beds of the Oraparinna Shale with a sharp, conformable contact, except at Brachina Gorge and Balcoracana Creek where it disconformably overlies the lower Wilkawillina Limestone of sequence €1.1. Facies ascribable to the member also occur above channel-fill deposits (described below) which were thus formed prior to final regression.

The Edeowie Limestone Member comprises planar to wavy laminated, platy dolomite mudstone with peloidal, sandy limestone and tuff beds (Moore, 1979a). At Ten Mile Creek, the member is overlain by Clarke's (1990c) upper bioherm complex and passes laterally into stromatolites and lime mudstone with evaporite pseudomorphs. Evaporite replacement is also evident in lenticular carbonate (?) collapse breccia at Bunyerroo Gorge, and as 'palisade' structures in the upper Moorowie Formation ('Book Limestone' ascribed by Mount (1970) to the Billy Creek Formation). The only drillhole to intersect uppermost sequence €1.3 in the Moorowie 'Syncline' is Moorowie 1 (Elliott, 1984), in which 31 m of sandy dolomitic limestone are overlain by 7 m of anhydrite which Gravestock and Hibburt (1991) correlated with the Edeowie Limestone Member.



Top: *Flindersipora bowmani* Lafuste, the world's oldest tabulate coral, Moorowie Formation, Moorowie Mine. Photo 042423.

Bottom: Palisade structure of calcite pseudomorphs after gypsum, upper Moorowie Formation, Moorowie Mine. Specimen width is 0.1 m. Photo 042425.

Volcaniclastic and syntectonic deposits

The first of two recognisable phases of tectonism, which represent the Kangarooian Movements (Daily and Forbes, 1969), affected upper sequence €1.3 prior to deposition of the Edeowie Limestone Member and its carbonate-evaporite lateral equivalents. Fault reactivation along the Torrens Hinge Zone, and uplift of the eastern Willyama Inliers and Benagerie Ridge, caused locally severe erosion of platform carbonates. Erosion products were transported towards the platform margin and redeposited either as channel fill or as upward-fining conglomerate beds. Well-rounded granules of quartz sand, released in large volume by tectonic activity, promoted further erosion and formed the matrix of channel breccias and lenticular to tabular sandstone cross-beds. A palaeogeographic reconstruction of this epoch is shown on Figure 7.7.

In the Mount Scott Range, channels in the upper Ajax Limestone cut down to levels containing the trilobite *Abadiella huoi* and were disconformably overlain by amalgamated conglomerate beds up to 9 m thick. Outcrops and drillhole intersections at Wirrealpa Mine comprise 'trough-fill' in the upper Wilkawillina Limestone (Roche, 1986). At Ten Mile Creek, cross-bedded sandstone and channel breccia occur in Clarke's (1990c) upper bioherm complex. From Mount Frome northward to Moorowie Mine, the Moorowie Formation contains breccia beds and lenses of sandy (quartz granule) limestone (Hatcher, 1970; Mount, 1970; Wigglesworth, 1970).

Volcanic activity is recorded in off-shelf environments by tuff in the upper Mernmerna Formation and volcaniclastic debris in the Narina Greywacke shortly before the erosional episode outlined above. On the carbonate shelf at Ten Mile Creek, bright olive-green, partly reworked ash fall tuff up to 2 m thick crops out several metres below the uppermost exposure of Wilkawillina Limestone. The tuff has a welded fabric of exquisitely preserved (undevitrified) glass shards, infilled vesicles and feldspar laths. Attempts to obtain an age of crystallisation from zircons have so far yielded only an imprecise age of 564 ± 61 Ma (Fanning, 1987).

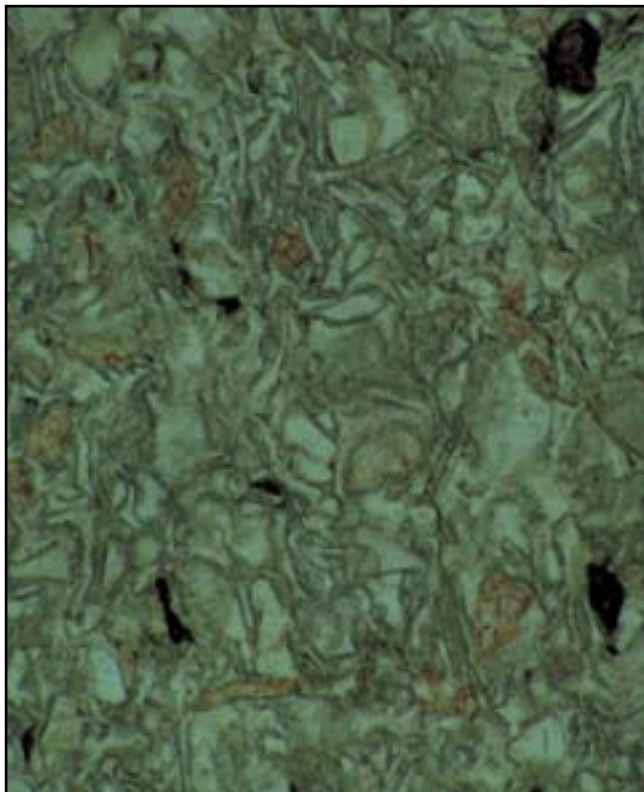
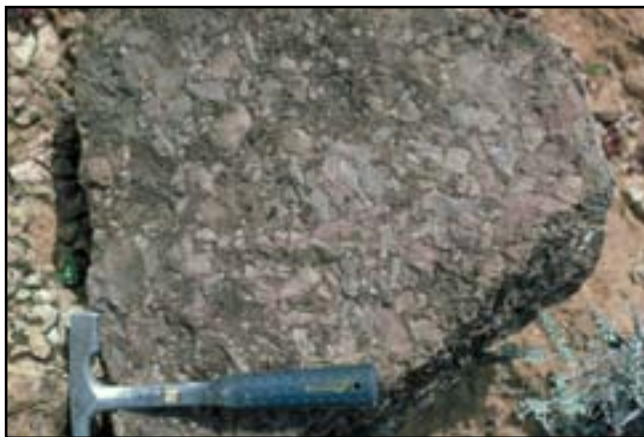
The best evidence of contemporaneous volcanic and tectonic activity is found at Reaphook Hill in the Coads Hill Member, assigned by Moore (1980) to the Billy Creek Formation and subdivided into informal units A-J. The Coads Hill Member is correlated here with upper Oraparinna Shale because unit H of Moore (1980) is a facies equivalent of the Edeowie Limestone Member, and because the Coads Hill Member is also represented in Yalkalpo 2 drillcore beneath the Billy Creek Formation. The basal Coads Hill Member at northern Reaphook Hill is a conglomerate 6 m thick with cobbles of eroded Bendieuta Formation in a very coarse-grained quartz-sand matrix. Elsewhere at Reaphook Hill, conglomerate is absent and feldspathic sandstone disconformably overlies calcrete-capped Bendieuta Formation (Gehling, 1971; Moore, 1980). The Coads Hill Member below unit H is ripple-laminated sandstone, siltstone, shale, burrow-mottled limestone and stromatolites. Shallow marine to intertidal, muddy alluvial plain and poorly oxygenated lagoon environments are represented in complex succession (Moore, 1980).

Unit H is nodular limestone with shaly to stromatolitic lower and upper beds, desiccation cracks and halite imprints. Trilobites (*Balcoracania dailyi*) are uncommon in unit H and abundant in units G and J (Moore, 1980). Of particular note are the dozen or more tuff bands 0.05 to 1.4 m thick in the southward thickening (140-200 m) Coads Hill Member.

Similar facies variation is found in Yalkalpo 2 drillcore but in a shallow subtidal setting. An unknown thickness of upper Mernmerna Formation has been removed by erosion, leaving a thin remnant beneath a scour surface. Clasts in the overlying conglomerate are platform limestone in quartz-feldspar granule and red volcanic porphyry matrix. Succeeding lithologies are silty to granule sandstone, burrowed fenestral mudstone, nodular oncolite limestone and oolite; all riddled with tuff beds 0.05 to 0.3 m thick. Porphyry granules which persist to several metres above the base of Billy Creek Formation imply erosion of newly exposed Benagerie Ridge.

Channels have been identified seismically in the Hawker Group at depth in the Moorowie 'Syncline' (Adams, 1987). They form a complex up to 4 km wide and appear to head north from an origin near Reaphook Hill. Although not yet drilled, they are attributed here to the same erosional events represented by the basal Coads Hill Member and other units beneath the Edeowie Limestone Member in the Flinders Ranges.

The Edeowie Limestone Member and equivalent evaporitic facies are thin but widespread, representing a return to tectonically quiescent conditions. Although the first phase of the Kangarooian Movements was short-lived in the Arrowie Basin, volcanic activity continued, since tuff beds occur in the Edeowie Limestone Member (Moore, 1979a), lower Billy Creek Formation (Dalgarno, 1964; Moore, 1979b, 1980) and Yarrowurta Shale on the Stuart Shelf (Daily, 1976b). The most intense extrusive volcanism, suggested by tuff bed frequency and distribution, is recorded in the eastern Flinders Ranges in upper sequence € 1.3 (late Botomian), and most likely correlates with eruptive phases of the Truro Volcanics in the Stansbury Basin to the south.



Top: Sandstone-supported limestone clast breccia in a channel near the top of the Ajax Limestone, Mount Scott Range. Photo 042426.

Bottom: Photomicrograph (plane light) of vitric tuff near the top of the Wilkawillina Limestone, Ten Mile Creek. Width of view is 2mm. Photo 042434..

Sequence set € 2

Two depositional sequences (€ 2.1, € 2.2) are distinguished in set € 2 (Fig. 7.2). The base of sequence € 2.1 at the top of the Hawker Group marks a regional lowstand, but with tectonic overprint. In the Flinders Ranges, sequence € 2.2 is represented by the Nildottie Siltstone and Eregunda Sandstone Members of the Billy Creek Formation (Moore, 1979b), and the Wirrealpa and Aroona Creek Limestones (Daily, 1956). Sequence € 2.2 is not preserved on the Stuart Shelf and has not been recorded in the Yalkalpo 'Syncline'. The palaeogeography of sequence € 2.1 and lower € 2.2 (excluding Wirrealpa and Aroona Creek Limestones) is shown on Figure 7.10.

Sequence € 2.1 deposits

Sequence € 2.1 comprises the Yarrowurta Shale on the Stuart Shelf (Johns, 1968a), Billy Creek Formation (Warragee and Erudina Siltstone Members of Moore, 1979b, 1980), and possibly unit J of the Coads Hill Member at Reaphook Hill (Moore, 1980). Undifferentiated Billy Creek Formation at depth in the Moorowie and Yalkalpo 'Synclines' may also belong in sequence € 2.1. Systems tracts have not been identified because of the primarily tectonic influence on these very shallow marine clastics.

Yarrowurta Shale, the youngest Cambrian unit preserved on the Stuart Shelf, abruptly overlies upper Andamooka Limestone stromatolites. Maximum preserved thickness is 90 m but higher levels have been eroded. The unit comprises flat to ripple-laminated red-green shale and siltstone, with tuff recorded near Andamooka (Daily, 1976b). The Warragee Member of the Billy Creek Formation in the Mount Scott Range and at Wirrealpa Mine likewise rests sharply on upper Ajax and upper Wilkawillina Limestone stromatolites. In other outcrops the basal Billy Creek Formation is conformable on the Edeowie Limestone Member of the Oraparinna Shale but at Ten Mile Creek it overlies a tongue of archaeocyathan Wilkawillina Limestone. Unfortunately, the contact is obscured by scree.

Moore (1979b) described the Warragee Member as evenly to ripple-laminated shale and siltstone up to 370 m thick. Desiccation cracks and salt casts are uncommon except in carbonate beds which cap poorly defined cycles above red, green and grey shale. Seven tuff bands, 0.04-0.5 m thick, occur near Wirrealpa Mine but are very rare further west; only one is known from the Mount Scott Range. Emuellid trilobites are abundant at several levels (Dalgarno, 1964; Pocock, 1970; Moore, 1979b; Bengtson *et al.*, 1990) and allow correlation with the upper White Point Conglomerate in the southern Stansbury Basin.

The Warragee Member and Yarrowurta Shale were deposited in a shallow, epicontinental sea. Shale-carbonate cycles 'developed in response to minor imbalance between the rates of subsidence and sedimentation' (Moore, 1979b, p.205).

At Reaphook Hill, unit J of the Coads Hill Member is thin (3 m), evenly laminated, khaki shale, siltstone and minor limestone. Brachiopods and the trilobite *Balcoracania dailyi* are abundant near the base and a mottled limestone is prominent at the top (Gehling, 1971; Moore, 1980). The unit is overlain conformably by the Erudina Siltstone Member of the Billy Creek Formation (Moore, 1980). This member and unit J are considered here to correlate with the Warragee Member and not with Nildottie Siltstone Member further west (cf. Moore, 1980, 1982a). Maximum preserved thickness of the Erudina Siltstone Member at Reaphook Hill is 270 m, but upper levels have been eroded.

The member comprises cyclic red-green shale and dolomite near the base, followed by red shaly siltstone. Several tuff bands and upward-coarsening shale-sandstone cycles occur in the middle part, while desiccation cracks, salt casts and ripple marks are present throughout the exposed succession. Moore (1980, 1982a) interpreted the Erudina Siltstone Member as a shallow epeiric sea deposit, again with cyclicity developed in response to pulses of tectonism associated with the Kangarooian Movements.

The 265 m of Billy Creek Formation in Yalkalpo 2 drillcore are red and green shale and siltstone with common sandstone interbeds, rare tuff bands near the base, and no carbonate (Youngs, 1978; Moore, 1980). Cyclic sediments occur in the middle; wavy and flaser bedding are common and desiccation cracks are abundant in red shale intervals. The formation is incomplete, upper levels having been eroded in the Yalkalpo 'Syncline'. The only drillhole to have fully penetrated the Billy Creek Formation in the Moorowie 'Syncline' is Moorowie 1 where 332 m are recorded (Elliott, 1984).

Sequence € 2.2 'lowstand deposits'

Because of the continued influx of clastics resulting from rejuvenation of uplifted sediment sources, the term 'lowstand' is applied with caution. There is a conformable passage from the Warragee to the Nildottie Siltstone Member, which reaches 460 m thickness and is composed mainly of ripple-laminated, coarse siltstone. Desiccation cracks and salt casts are abundant locally while bioturbation is rare (Moore, 1979b). At Ten Mile Creek, the Nildottie Siltstone Member disconformably overlies lower Wilkawillina Limestone (sequence € 1.1). This is the longest known stratigraphic break within the Early Cambrian with no record of three whole sequences.

The Eregunda Sandstone Member has a sharp basal contact and reaches 166 m in thickness near Bunyerroo Gorge. Moore (1979b) distinguished a fine-grained shale-siltstone association of shallow subtidal to intertidal origin in the thinner northern outcrops, and a current-laid deltaic sandstone association further south. Wopfner (1969a, 1970a) and Moore (1979b) suggested the region of the Willyama Inliers as a sediment source.

Vincent (1980) noted a southeasterly thickening trend for the entire Billy Creek Formation which parallels that indicated by Stock (1974) for the Lake Frome Group. Compared with sequence set € 1 of the Hawker Group, the Billy Creek Formation depocentre is shifted to the south, i.e. close to its inferred source area. An alternative pattern of sediment distribution, with the thickest sediments in the centre of the Arrowie Basin, is shown on Figure 7.10. Barriers are envisaged to the east (Benagerie Ridge), north (Mulloorina Gravity Ridge) and west (Stuart Shelf). Possible connection with the Stansbury Basin to the south is consistent with Moore's (1982a) suggestion of a southeasterly palaeoslope.

Sequence € 2.2 transgressive and highstand deposits

These are treated together since they are represented entirely by the Aroona Creek and Wirrealpa Limestones (Daily, 1956). Youngs (1977) documented the transgressive-regressive nature of these units in detail. She recognised two major facies associations composed of ooid banks advancing to the southwest over lagoonal deposits during transgression, then retreating to the northeast during regression (Fig. 7.13).

The Aroona Creek Limestone crops out in the Mount Scott Range and only the basal 10 m of the 70 m thick unit are subtidal skeletal-peloid packstone. The remainder is principally fenestral and cryptalgal-laminated dolomite mudstone deposited in the upper intertidal to supratidal zone during regression (Daily, 1976b).

The Wirrealpa Limestone is remarkably consistent in thickness (105-135 m) across the Flinders Ranges and into the Moorowie 'Syncline', but is not known from the Yalkalpo 'Syncline'. No age-equivalent carbonates occur at Mount Arrowsmith in western New South Wales but the lower Coonigan Formation near Mount Wright is a correlative. Trilobites and brachiopods in the basal Aroona Creek and lower Wirrealpa Limestone define Faunal Assemblage 10 (Daily, 1956) and numerous other skeletal remains, including cambroclaves and palaeoscolecids (worm-like organisms with scaly armour), have been described recently (Brock and Cooper, 1993).

Youngs' (1977) lagoonal facies association is dominated by lime mudstone and wavy-bedded to nodular limestone in a calcareous, silty matrix. Skeletal, peloid and oolitic beds are subordinate but calcimicrobe buildups are prominent at a number of localities.

These are considered by Youngs to have grown on the leeward margins of ooid shoals, either as columnar thrombolite complexes 40-50 m wide and 7 m thick, or as roughly domal isolated bioherms 0.2-3.0 m across. They are often grouped on single bedding surfaces and are encased in lime mud. A large bioherm of calcimicrobes, rare archaeocyaths and radiocyaths occurs at the type section near Ten Mile Creek (Daily, 1956; Kruse, 1991).

The ooid bank facies association 'is characterised by deposition of oolitic and oncolitic lithologies on the turbulent crests and flanks of ooid shoals' (Youngs, 1977, p.39). Oolite forms the bulk of the formation in the eastern Flinders Ranges (Youngs, 1977; Vincent, 1980) and in Moorowie 1 drillhole (Elliott, 1984) but is not recorded in Lake Frome 1 and 2 drillholes in the southern Moorowie 'Syncline' (Youngs, 1977).

Sequence € 2.2 transgressive and highstand deposits of the Wirrealpa and Aroona Creek Limestones were deposited in a tectonically quiescent phase as the last major carbonates in the Arrowie Basin. The suite including the Ramsay, Stansbury and Coobowie Limestones on Yorke Peninsula likewise represent the last carbonates in the Stansbury Basin. Relative sea-level changes in this sequence have Australia-wide and global significance. The transgressive deposits herald a major flooding event in the Daly, Bonaparte, Amadeus and Georgina Basins. The top of sequence € 2.2 or the overlying lowstand deposits also mark the onset of a break in deposition in these basins (Cook, 1982). In western New South Wales the onset of this break is somewhat younger, since the Coonigan Formation beneath the unconformity is Templetonian in age (Wang *et al.*, 1989).

Regressive facies are known from a number of localities in North America, Siberia and Europe (Zhuravlev, 1986; Rowland and Gangloff, 1988). The 'Toyonian regression' documented by Rowland and Gangloff is now thought to be represented by the upper Wirrealpa Limestone and Moodlatana Formation rather than by the upper Hawker Group and Billy Creek Formation. The latter regression, though recorded in the Arrowie, Stansbury and eastern Officer Basins, was influenced by the Kangarooian Movements which probably signalled renewed rifting in the Australian sector of the Gondwana supercontinent.

Sequence set € 3

Lake Frome Group redbeds and sandstones constitute set € 3 (Fig. 7.11). The apparently conformable succession is locally up to 2 700 m thick but erosion has removed younger units, leaving progressively less outcrop towards the northern Flinders Ranges for study. Drillholes have intersected Lake Frome Group west of the ranges and in the Moorowie 'Syncline' but not in the 'Yalkalpo Syncline'. Drillhole CBH 2 intersected 148 m of flat-lying possible Lake Frome Group sediments 20 km north of the Mount Babbage Inlier (Figs 7.11, 7.12).

Lowstand and transgressive deposits are represented by the Moodlatana and Balcoracana Formations (Daily, 1956) but evidence of renewed tectonism renders the distinction of highstand deposits impossible with available data. The two youngest formations, Pantapinna Sandstone (Dalgarno and Johnson, 1965) and Grindstone Range Sandstone (Mawson, 1939), are discussed separately and sequence stratigraphic terminology is not applied to them.

Lowstand deposits

These constitute the Moodlatana Formation which conformably overlies the Wirrealpa and Aroona Creek Limestones. The formation boundary is marked by abrupt appearance of red-brown micaceous siltstone or a passage through thin, mudcracked, calcareous siltstone and fenestral to laminated limestone (Stock, 1974; Vincent, 1980). The Moodlatana Formation consists of 170-590 m of micaceous siltstone, shale, arkosic fine to medium-grained sandstone and minor carbonate. Western outcrops are more sandy, usually tabular cross-bedded with occasional scour channels.

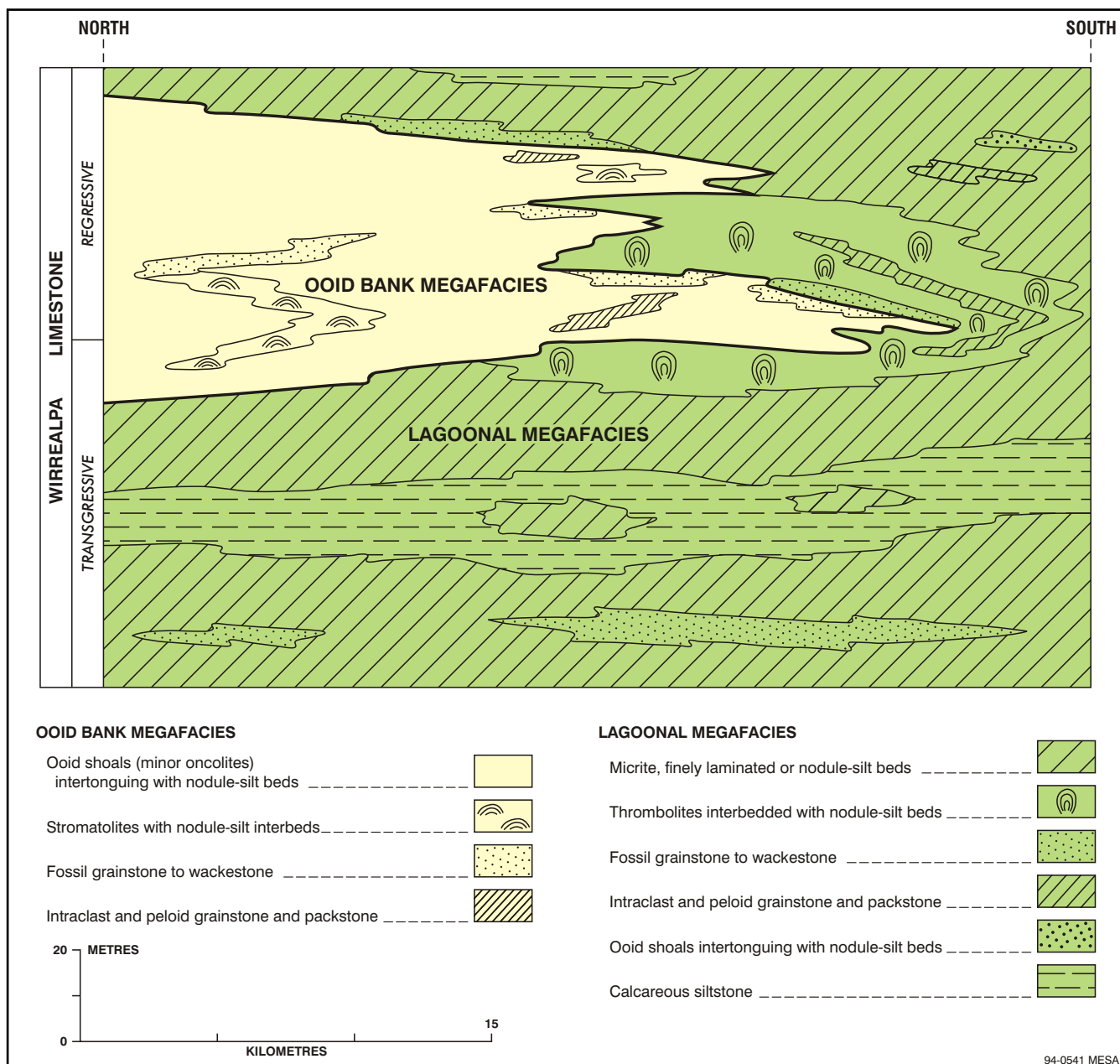


Fig. 7.13 Transgressive-regressive sequence, Wirrealpa Limestone, 'Wirrealpa' area (after Youngs, 1977).

The dominant siltstone lithology is planar to ripple-bedded with mudcracks, evaporite casts and trilobite tracks. Thin carbonate beds are cryptalgal laminated. Metadoxidid trilobites (*Onaraspis rubra*) occur near the top of the formation at 'Wirrealpa', Ten Mile Creek and Mount Frome (Daily and Forbes, 1969; Stock, 1974; Jell, 1983a; Jell in Bengtson *et al.*, 1990).

Lower Moodlatana Formation carbonates have a high proportion of acid volcanic clasts and, significantly, abundant fresh olivine (forsterite) near 'Wertaloona' and Mount Frome (Stock, 1974).

The depositional environment was shallow subtidal to intertidal with sand transport generally towards the east (Stock, 1974; Vincent, 1980).

Columnar thrombolite bioherm in Wirrealpa Limestone near 'Wirrealpa'.
Photo 042428.



Transgressive deposits

The Balcoracana Formation (80-280 m thick) is repetitively bedded, very micaceous siltstone, shale and carbonate, deposited conformably above the Moodlatana Formation (Daily, 1956; Dalgarno, 1964). Arkosic, fine-grained, trough cross-bedded sandstone is relatively common near Brachina Gorge but is a minor facies elsewhere. Red-green shale-carbonate cycles contain pencil-thin chalky limestone bands and laminated or stromatolitic carbonates 0.2-1 m thick. Ripples, mudcracks and halite casts are common; anhydrite occurs in patches in the Lake Frome drillholes.

Trilobite tracks occur commonly but the only known body fossil is a single effaced agnostid trilobite from Lake Frome 3 drillcore, from which a late Middle Cambrian age is assumed (Daily and Forbes, 1969). A large, bivalved arthropod in Edeowie 1 drillcore may be from the upper Balcoracana Formation or lower Pantapinna Sandstone (Jell, 1983b).

Stock (1974) and Vincent (1980) differ on whether the carbonates in each cycle represent subtidal or supratidal environments. Daily (1976b) envisaged multiple transgressions and regressions but the view held here is that, as in the lower Billy Creek Formation, cyclicity represents alternating supply and starvation of sediment from a tectonically active source area.

Pantapinna Sandstone

Analysis of this thick (950-1 300 m) feldspathic sandstone is difficult due to restricted outcrop and subcrop occurrences of the full unit. The basal beds lying conformably above the Balcoracana Formation are marine as they contain trilobite tracks, ooid grainstone and tidal channels. Bioturbated beds and large ripple bedforms at higher levels also indicate subtidal conditions (Vincent, 1980; Jago and Daily, 1982).

The Pantapinna Sandstone mainly comprises 0.5-1 m planar and trough cross-beds of red and white fine to medium-grained sandstone with minor micaceous siltstone and shale. Acid volcanic and metamorphic lithic clasts (schist, phyllite) are abundant at certain levels, but apart from common heavy-mineral laminations, sandstone mineralogy is uniform (Stock, 1974).

Following Vincent (1980), the environment of deposition is interpreted as low intertidal to shallow subtidal with bedload transport in tidal channels and migrating megaripples on a shallow seafloor. No age-diagnostic fossils have been found.

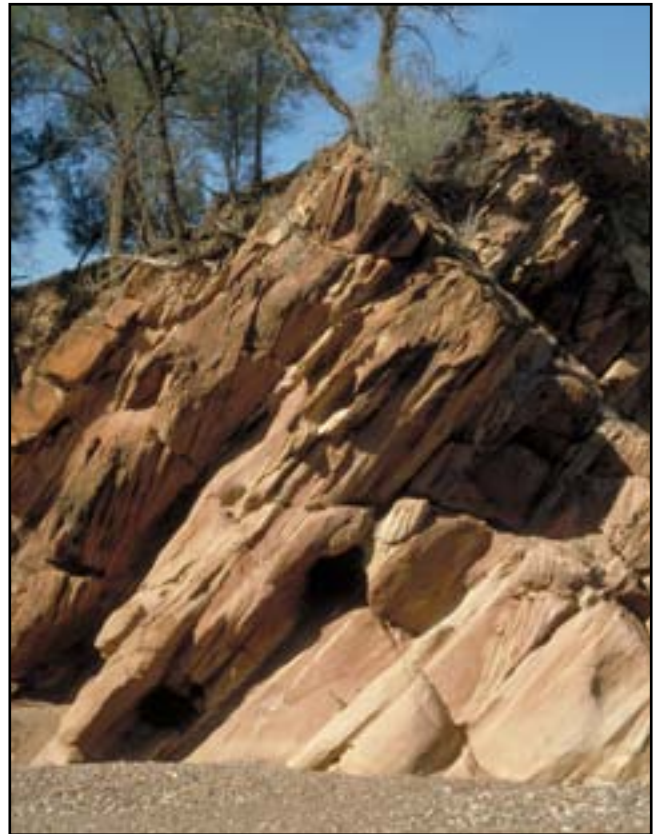
Grindstone Range Sandstone

The uppermost formation of the Lake Frome Group is thickest (750 m) near 'Wirrealpa', but nowhere is its top exposed. The only subsurface intersection is incomplete (68 m in Moorowie 1). The Grindstone Range Sandstone is an unfossiliferous mature quartz arenite with minor feldspar and upward-decreasing traces of volcanic lithic clasts (Stock, 1974). Well-rounded quartz-arenite pebbles and cobbles are concentrated in channels and down planar foresets in upper exposures near 'Wirrealpa'. Planar cross-beds, often slumped, are the dominant bedforms, but trough and herringbone cross-beds and planar laminations also occur (Vincent, 1980). Trilobite tracks and mudcracks are found in basal sandstones. Environments of deposition range from low intertidal at the base to bedload-dominated, fluvial, sandy braidplain in upper exposures.

Palaeogeography and age

While the Lake Frome Group was being deposited, the Stansbury and eastern Warburton Basins were also depocentres. As noted above, the exposed region of the Willyama Inliers may have supplied sediment to the Billy Creek Formation, while Stock (1974) and others have suggested a Gawler Craton provenance for the Lake Frome Group.

Like the Billy Creek Formation, the Lake Frome Group may also have been thickest in the central Arrowie Basin, suggesting that the WHH no longer controlled deposition. With rapid subsidence taking place in the Stansbury Basin at this time, the



Cross-bedded Pantapinna Sandstone on the bank of Balcoracana Creek. Photo 042429.

southeasterly palaeoslope suggested for the upper Wilpena Group and Hawker Group in the Chace Range region, may have assumed brief dominance. Thus, depositional continuity of the Billy Creek Formation, Wirrealpa Limestone and Lake Frome Group with the Kanmantoo Group is implied. However, a north-dipping palaeoslope is evident in fluvial sands in the upper Grindstone Range Sandstone, indicating uplift of the southern Arrowie Basin and, ultimately, severing of the connection with the Stansbury Basin (Powell and Gatehouse, Macquarie University and SADME, pers. comm., 1991). A sketch of the suggested palaeogeography for sequence set € 3 is shown on Figure 7.11, but Grindstone Range Sandstone is omitted for lack of data.

If a Middle Cambrian age is accepted for the upper Balcoracana Formation, the overlying Pantapinna and Grindstone Range Sandstones are younger, but by how much has been a source of conjecture. Assuming the Lake Frome Group represents a single sequence, a minimum age of middle Late Cambrian (Iverian) can be reasonably inferred, as explained below.

On the Wonominta Block of western New South Wales (Fig. 7.12), the late Middle Cambrian is represented by a substantial gap in the stratigraphic record. Further south towards Mount Wright, much of the Late Cambrian record is missing as well (Webby, 1978, 1982; Webby *et al.*, 1988; Wang *et al.*, 1989). Beneath the erosion surface, Early to Middle Cambrian marine slope sediments are tightly to isoclinally folded. They are immature lithic, feldspathic sandstones locally intruded by gabbroic and other mafic bodies. The angular unconformity and associated deformation were identified by Webby (1978) as the Mootwingee Movement, which he regarded as part of the Delamerian Orogeny.

Given the mineralogical immaturity of the fine-grained sediments and carbonates in the Moodlatana and Balcoracana Formations, their lowstand to transgressive positions in € 3, and their derivation from some source undergoing renewed tectonic stress, it is suggested that the deforming Wonominta Block, Willyama Inliers and Benagerie Ridge provided these sediments. The thick, increasingly mature Pantapinna and Grindstone Range Sandstones may have been highstand deposits. Well-rounded quartzite pebbles, in particular, may have been derived from uplifted Pound Subgroup.

The Kayrunnera and Mootwingee Groups, which diachronously overlie the unconformity in western New South Wales, represent a new transgressive phase extending into the Ordovician (Wang *et al.*, 1989). If the Grindstone Range Sandstone contains no breaks, it is not likely to correlate with this younger depositional sequence, and hence may be no younger than Iverian (=pre-Payntonian). If, on the other hand, fluvial deposits of the upper Grindstone Range Sandstone are lowstand deposits of a new sequence, then this uppermost section could be Payntonian or younger. It now appears certain that at least part of the Lake Frome Group was deposited during the Mootwingee Movement in western New South Wales.

EASTERN WARBURTON BASIN

D.I. Gravestock and C.G. Gatehouse

Geological setting

The entirely concealed Warburton Basin in northern and northeastern South Australia was considered by Wopfner (1972) to comprise part of a Cambro-Devonian depositional phase. He also linked the depositional areas of the Officer and Warburton Basins as his 'Circum-Denison Arc, the palaeo-extension of which also included the Bancannia Trough in northwest New South Wales' (Wopfner, 1972, p.36).

Here, the Warburton Basin is considered to contain only Cambrian to Ordovician strata (including the Innamincka Formation, see below). As pointed out by Wopfner (1972), the history of the eastern Warburton Basin is closely linked with that of the Gnalta Shelf in western New South Wales (the Bancannia Trough is a Devonian feature: Webby, 1982). Preserved limits of the Warburton Basin (Fig. 7.12) are as shown by Gatehouse (1986) but modified by possible southeast extension of the Arunta Block inferred from mylonitised pegmatitic gneiss in Haddon Downs 1 (Rankin and Gatehouse, 1990). Pre-Mesozoic basement in nearby Araburg 1, however, is grey siliceous siltstone; thus the gneiss may have been emplaced by late Palaeozoic or Mesozoic fault movements.

Eastern Warburton Basin sediments were deposited on and east of the Birdsville Track Ridge (Fig. 7.12) where drillhole and seismic data permit a reasonably coherent picture of the sedimentary record. Western Warburton Basin sediments are discussed briefly after those of the eastern Officer Basin.

Elements of five depositional sequence sets are preserved in the eastern Warburton Basin. Sets € 1 to € 3 are Early to Middle Cambrian, corresponding to those in the Arrowie and Stansbury Basins. The latest Middle Cambrian to Early Ordovician set € 4/01 has no known counterpart in other basins of the eastern region, except perhaps for mafic igneous rocks beneath the Murray Basin (Fig. 7.2). The incomplete fifth Middle to Late Ordovician sequence set is the youngest known in the region. Roberts *et al.* (1990) carried out a detailed seismic and sedimentary facies study of € 2-€ 4/01 (Kalladeina Formation) west of the Gidgealpa area.

Palaeogeography of the eastern Warburton Basin during the Early Cambrian (Figs 7.5-7.7, 7.10) is poorly known, since the oldest rocks drilled to date are bioturbated sandstone, siltstone and shale lacking fossils diagnostic of age. They overlie steeply dipping Neoproterozoic quartzite in some drillholes (e.g. Daralingie 1). Palaeoproterozoic to Mesoproterozoic metasediments occur beneath the southern Warburton Basin (Mulga 1, Gurra 1, Fortville 3, Paxton 1).

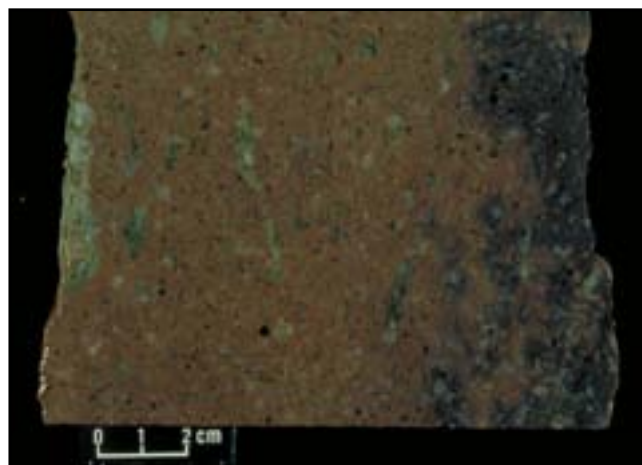
Sequence set € 1

Siliciclastics of the Pando Formation (Gravestock *et al.*, 1995) occur beneath and intertongue with the Mooracoochie Volcanics (Gatehouse, 1983). Almost 200 m were intersected in the type drillhole (Pando 1) but the section has not been fully penetrated. The formation is characterised by pale green-grey, irregularly laminated, bioturbated silty sandstone with common glauconite and detrital zircon. Slightly pyritic siltstone forms minor interbeds while olive-green shale with disseminated glauconite and thin sandstone interbeds predominates in some drillholes. The unit is a reservoir for Permian gas in Moolalla 1 (Taylor *et al.*, 1991). The formation was deposited on a shallow marine shelf of unknown extent.

Two units can be distinguished in the Mooracoochie Volcanics. The lower unit, which exceeds 370 m in Gidgealpa 2, consists of massive, orange to red-brown porphyritic trachyte with zoned albite phenocrysts and olivine pseudomorphs. In Gidgealpa 2, a hybrid trachytic lava appears to have been shattered and penetrated by dacitic lava, possibly from a vent or feeder (Fander *in* Harrison and Higginbotham, 1964).

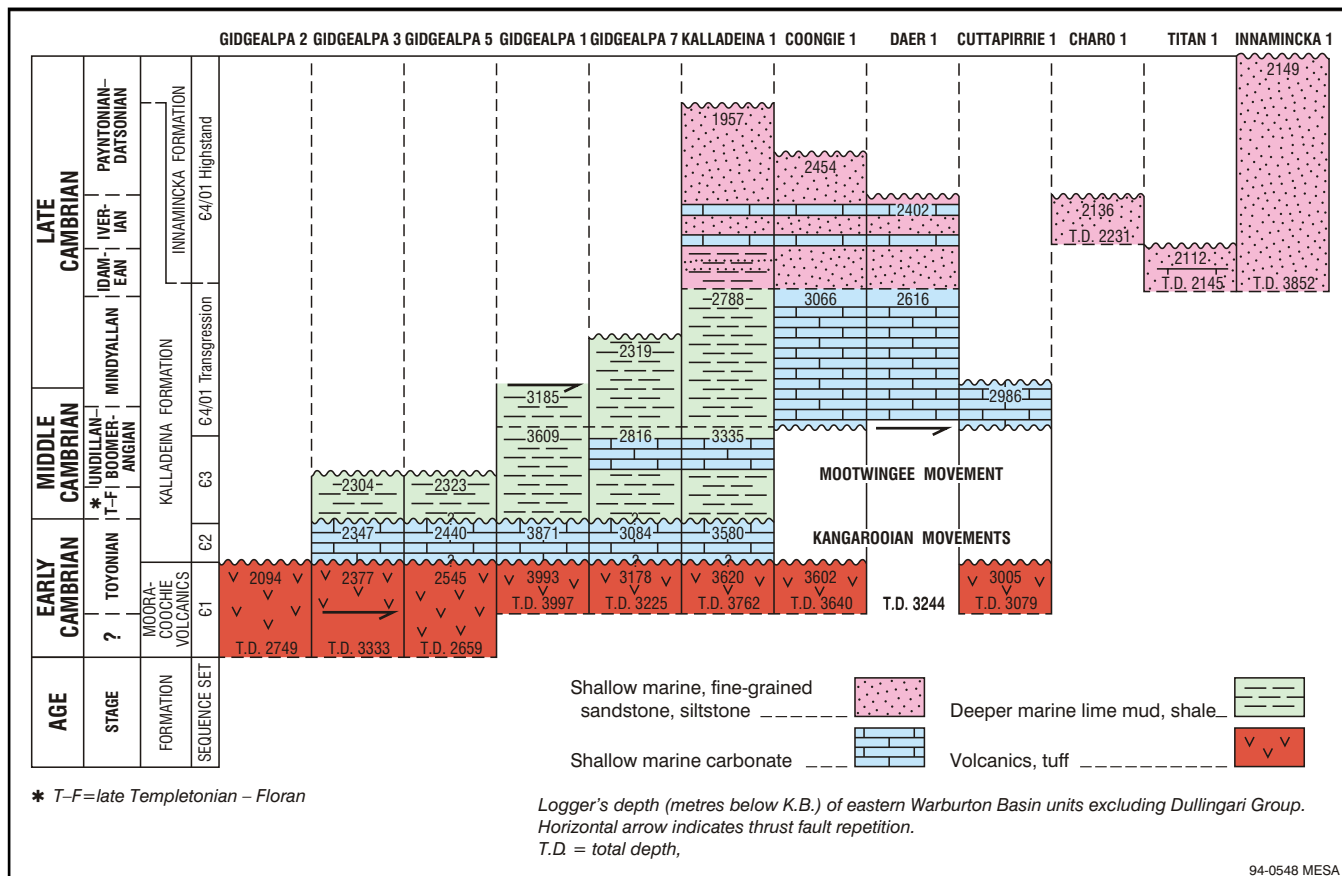
The contact between the upper and lower unit is sharp on wireline logs but has not been cored. The upper unit varies in thickness from 88 to 385 m in Gidgealpa drillholes, but has not been fully penetrated elsewhere. This unit is widespread from Taloola 1 to Paning 1, over 175 km to the northeast (Fig. 7.12). It consists of numerous rhyolitic to dacitic tuff beds (welded and lapilli tuff and ignimbrite), minor interbeds of coarse tuffaceous calcareous sandstone and thinly bedded siltstone. Poorly bedded agglomerate in Gidgealpa 1 includes scarce subangular pebbles of black, pyritic limestone and black chert with numerous cross-sections of hyoliths and polymerid trilobites (Daily *in* Harrison and Higginbotham, 1963). A Cambrian age is thus established for the upper Mooracoochie Volcanics, and they must also be older than late Templetonian (determined for € 3, Table 7.1). The Cymbric Vale Formation in western New South Wales likewise consists of tuff, chert and minor fossiliferous limestone on the Gnalta Shelf, where it is up to 1 900 m thick (Kruse, 1982).

Rb-Sr age determinations range from 700 to 360 Ma (Forbes, 1969) but the younger age was probably reset by the Alice Springs Orogeny (discussed below) and the older age is unreliable. Both units are considered here to be genetically related, and to correlate with the Mount Wright Volcanics and Cymbric Vale Formation on the Gnalta Shelf, for which an Early Cambrian age has been established in upper levels (Kruse, 1982). There are doubts as to the island arc affinities of the Mount Wright Volcanics (Edwards, 1978), and the Gidgealpa Volcanic Arc of Gatehouse (1986) may alternatively be interpreted as a continental rift. Aeromagnetic data suggest that the rift extended north-northwest from the Murteree A1 vicinity, passing the east flank of the Gidgealpa oil and gas field (Boucher, 1991). Commercial flows of Permian oil have been recorded from up-dip traps in the Mooracoochie Volcanics in Sturt 6.



Crystal tuff, Mooracoochie Volcanics at 1921 m in Taloola 1. Photo 042430.

Table 7.1 Eastern Warburton Basin, depths to principal unit boundaries from wireline logs.



Sequence set € 2

Sequence set € 2 is represented by 30-100 m of limestone with occasional granules and pebbles of acid tuff, suggesting a relative lull in volcanism. This constitutes the basal Kalladeina Formation in Kalladeina 1 and Gidgealpa drillholes (Gatehouse, 1983), but it was either not deposited or has been eroded in the Coongie-Cuttapirrie region (Table 7.1). The limestone is dolomitised and fractured but was originally an ooid grainstone with minor bioclastic beds. Trilobite and hyolithid fragments have not yielded a precise age but the unit is pre-late Templetonian. A Toyonian age is assumed, suggesting correlation with the Wirrealpa Limestone in the Arrowie Basin and lower Coonigan Formation on the Gnalta Shelf.

The unit has petroleum reservoir potential and was tested in Gidgealpa 1, resulting in recovery of heavily gas-cut salt water, but with no commercial gas flow.

Sequence set € 3

Lower Kalladeina Formation above the basal unit comprises sequence set € 3. Transgressive and highstand deposits can be distinguished in Kalladeina 1 and Gidgealpa 7, but moderately deep marine conditions apparently persisted in Gidgealpa 1. In contrast, a hiatus recognised by Roberts *et al.* (1990) is recorded through € 2 and € 3 above the Mooracoochie Volcanics in Coongie 1 and Cuttapirrie 1 (Table 7.1).

Lowstand deposits

Lowstand deposits have not been identified, but subaerial exposure of € 2 limestone caused extensive recrystallisation and dissolution at this time. Extreme brecciation of the upper tuffaceous unit of the Mooracoochie Volcanics in Gidgealpa 11 is thought here to be a product of penetrative fracturing during this lowstand.

Transgressive deposits

These sediments comprise mainly chert, grey micaceous shale, bioturbated occasionally ripple-laminated siltstone, and flat to wavy and lenticular laminated limestone. Tuffaceous interbeds contain pebbles of acid igneous rocks and rare pink dolomite probably reworked from the basal Kalladeina Formation. Transgressive deposits in Gidgealpa 3 overlie silicified limestone and 'chert breccia' suggesting a disconformable contact with the basal Kalladeina Formation.

Ripple lamination, trilobite tracks and current-sorted hyoliths indicate well-oxygenated, agitated bottom waters. Age-diagnostic agnostid trilobites, numerous polymerid trilobites and very abundant sponge spicules occur with brachiopods and gastropods in these sediments.

Highstand deposits

A return to shallow marine conditions is recorded by pale grey shale and siltstone, and white recrystallised limestone in Gidgealpa 1 and Kalladeina 1. In Gidgealpa 7, shoaling is represented by 80 m (partly cored) of ooid-oncolite packstone with abundant fossil fragments. Subaerial exposure is suggested by mottling and brecciation in the lower part of this unit.

How widespread the strata of sequence set € 3 may have been cannot be judged from available data. Hiatus continued in the Coongie-Cuttapirrie region to the north and, in view of the major depositional break (Mootwingee Movement) on the Wonominta Block (Wang *et al.*, 1989), there may have been substantial hiatus towards the southeast. An east-facing gulf is suggested, but marine conditions could have extended considerably further west of Kalladeina 1 than shown on Figure 7.11.

Sequence set € 4/01

The upper Kalladeina Formation and Innamincka Formation (=Innamincka Red-Beds) form transgressive and highstand deposits of set € 4/01 but lowstand deposits have not yet been identified. The lower part of the Ordovician Dullingari Group (Gatehouse, 1983) correlates with the uppermost Kalladeina Formation and all of the Innamincka Formation. The group is discussed separately as facies relationships are not sufficiently clear to place it in a sequence stratigraphic context.

Transgressive deposits

Differentiation into shallow shelf, slope, and moderately deep-water basinal deposits is possible from cores and wireline logs. Shelf limestone disconformably overlies the Mooracoochie Volcanics in the Coongie-Cuttapirrie region and comprises more than 530 m of ooid grainstone, partly dolomitised, with thin interbeds of grey and red siltstone. Fenestral and cryptalgal laminated limestone occur in the lower 100 m; these may turn out to represent lowstand deposits. Fossils are sufficient to indicate a Mindyallan age in the mid-part of the unit (Coongie 1, core 9).

Slope deposits are represented in Gidgealpa 1 by dark grey, wavy-laminated limestone and shale, interrupted by chaotic slump breccias of slope-derived ribbon limestone clasts and irregular blocks of shelf-derived ooid grainstone. Tuffaceous interbeds are common in basal units, indicating an early Late Cambrian resumption of volcanic activity.

Altered vesicular basalt in Murteree A1 was assigned questionably to the Mooracoochie Volcanics by Gatehouse (1983, 1986). Trace-element analysis of alkaline basalt from neighbouring drillholes Jena 1 and 6 indicated a within-plate origin (Boucher, 1991; Fig. 7.9). An age spanning the Middle-Late Cambrian boundary is evident from trilobites in limestone interbedded with tuff and vesicular agglomerate in Gidgealpa 1, hence the basalt is assigned to the Kalladeina Formation. It is compositionally very similar to the older Suite 1 basalt in the Stansbury Basin (Rankin *et al.*, 1991b).

Deeper, basinal deposits are recorded in Kalladeina 1 by 550 m of dark grey laminated limestone, pyritic shale, siltstone and thin chert interbeds composed entirely of sponge spicules (Daily in Delhi Australian Petroleum Ltd, 1967). Other fossils include age-diagnostic trilobites, brachiopods and hyoliths.

Palaeogeography at this time was similar to that suggested by Cook (1982). Ooid shoals persisted on the northern shelf throughout the transgressive epoch, indicating very high carbonate productivity. Slope facies in Gidgealpa 1 imply a shelf margin between there and Coongie 1. This shelf edge may be coincident with the rift shoulder and the site of volcanic activity. A basalt-associated aeromagnetic anomaly, which can be traced as far south as Kobari 1 drillhole, is presumed to outline in part the orientation of the rift. A deeper water setting is evident in Kalladeina 1, the westernmost source of data for the Late Cambrian (Mindyallan). Webby *et al.* (1988) and Wang *et al.* (1989) indicated contemporaneous deeper water conditions in the Kayrunnera-Koonenberry region, shallowing to marginal marine conditions at Mount Wright.

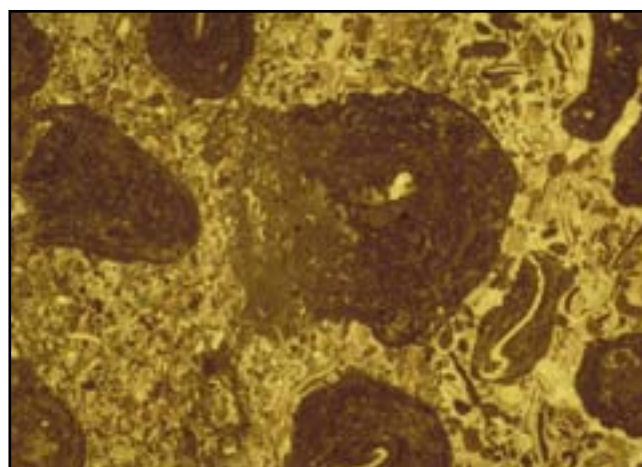
Kennard and Lindsay (1990) recorded marine transgression and persistently subtidal conditions in the upper Shannon Formation in the northeastern Amadeus Basin (their sequence 4). Marine conditions are also known from the Georgina Basin (Öpik, 1967) and Bonaparte Basin (Öpik, 1969), and it is likely that a seaway extended from the Gnalta Shelf across the Warburton Basin to these northern basins in Mindyallan time. Deeper marine conditions prevailed in the Kalladeina area; to the north and northeast, oolite shoals flanked the Arunta Block while braided fluvial and delta complexes transported siliciclastics to a shallow marine shelf from the south and southwest (Fig. 7.14).

Highstand deposits

Although there is no further record in Gidgealpa drillholes, seven cores from Kalladeina 1 give a good account of the highstand facies succession. Cores and/or wireline logs from Coongie 1, Daer 1, Titan 1, Charo 1 and Innamincka 1 provide additional coverage, and also establish that the Innamincka Formation forms part of the sequence € 4/01 highstand.

Deep-water, laminated grey-green shale, with numerous bedding-plane hydroid impressions, in Kalladeina 1 is overlain by silty micaceous shale with thin interbeds of ooid grainstone and fossil packstone. Following higher in the section are pyritic, glauconitic, micaceous, ripple-laminated siltstone and very fine-grained sandstone. Clay drapes contain horizontal burrows, and shelly fossils are common. These include lingulid brachiopods (cf. Innamincka 1, core 18; Ludbrook in Ryan, 1961), hyoliths, gastropods, and trilobites of latest Cambrian (Payntonian-Datsonian) age. Age-diagnostic trilobites also occur in Coongie 1 (Daily in Pemberton, 1970).

There is an upward increase in glauconitic siliciclastics to the exclusion of carbonate. A similar upward change occurs on the carbonate shelf where micaceous shale, glauconitic siltstone and fine-grained sandstone eventually replace carbonate, which is reduced to thin interbeds of ooid grainstone and bioclastic oncolite packstone. Abundant remains of the calcified alga *Nuia* form marker beds in this transition (X. Sun, National Centre for Petroleum Geology and Geophysics, pers. comm., 1992). Fossiliferous, glauconitic and *Nuia*-rich limestone with Early Ordovician (Arenigian) conodonts in Packsaddle 1 (Cooper, 1986;



Top: Slope breccia in upper Kalladeina Formation at 3 563 m in Gidgealpa 1. Top is to the right. Photo 042431.

Bottom: Bioclastic oncolite packstone (trilobite fragments commonly in oncolite nuclei) in upper Kalladeina Formation at 2 230 m in Charo 1. Width of view is 20 mm. Photo 042432.

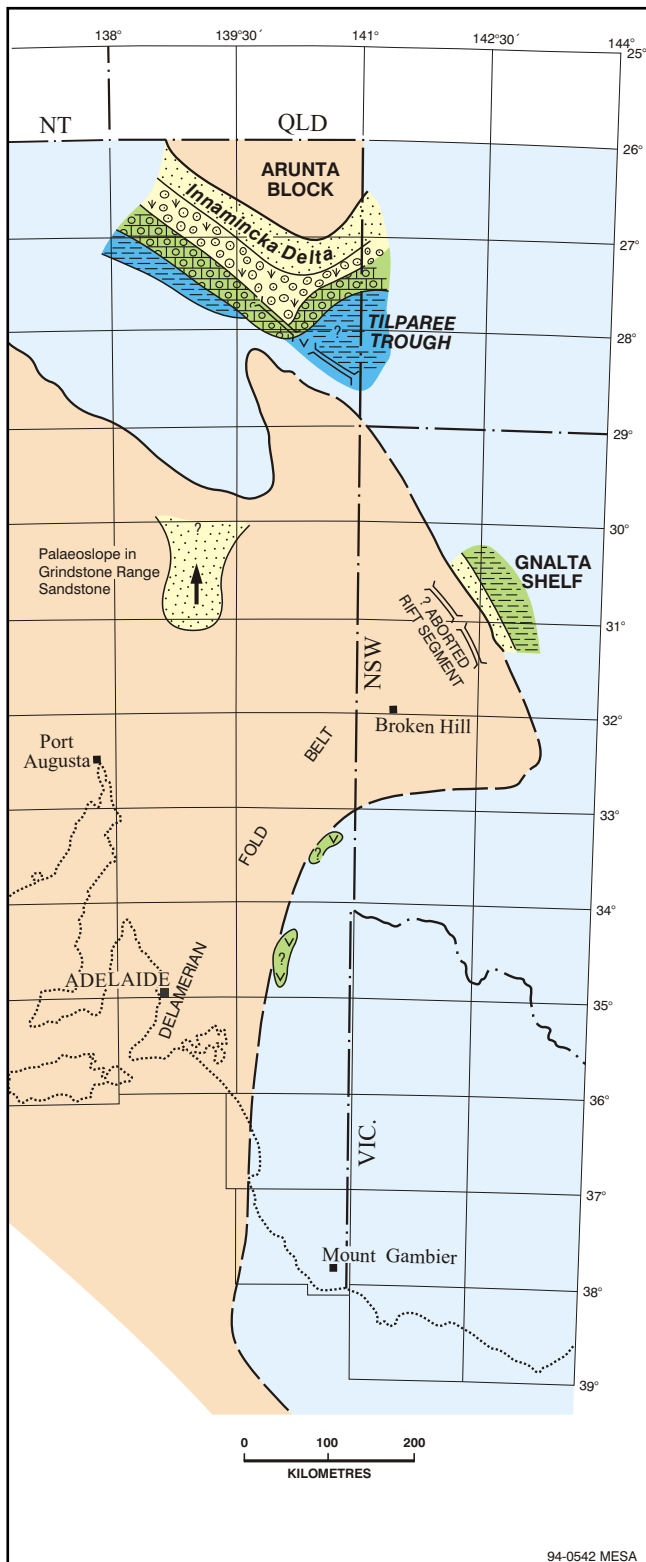


Fig. 7.14 Palaeogeography, sequence set € 4/01.

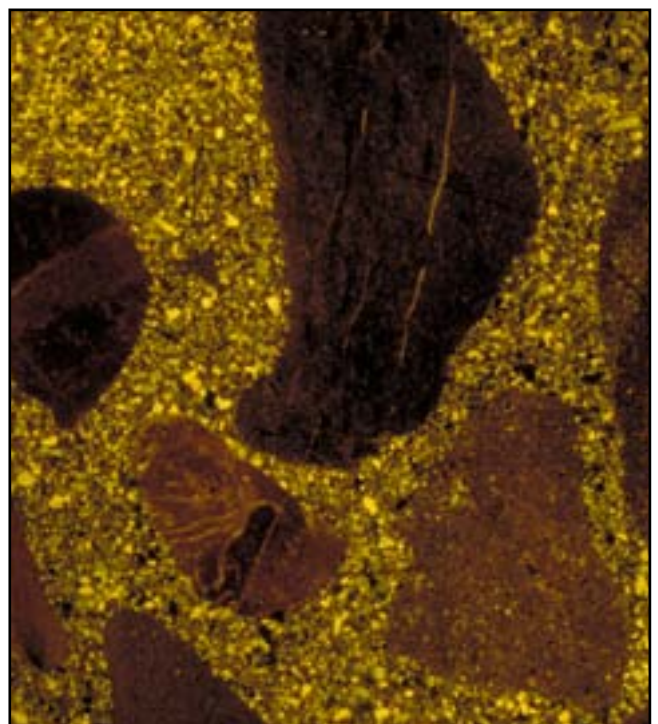
Gatehouse, 1986) is in the same transition from the Kalladeina to the Innamincka Formation. Preliminary acritarch studies (Zang, 1992) indicate a probable Late Cambrian to Early Ordovician age for the Innamincka Formation. Although carbonate is very rare in the 19 drillcores from Innamincka 1, the presence of ripple-laminated, fine-grained glauconitic sandstone and siltstone with worm trails on muddy bedding surfaces suggests a tenuous facies correlation with the other drillholes. Extensive shallow subtidal conditions are envisaged since evidence of subaerial exposure is very localised.

The abundance of red-brown shale and lack of carbonate suggest proximity of Innamincka 1 to a northern detrital source, possibly the Arunta Block. Granite pebbles in the Gnalta Shelf deltaic succession imply possible unroofing of Delamerian granites by earliest Ordovician time (Webby, 1978).

Dullingari Group

The Innamincka Formation passes south into deeper marine equivalents represented by the lower Dullingari Group (Gatehouse, 1983, 1986). The Dullingari Group in the type drillhole (Dullingari 1) is black, laminated pyritic shale in lower drillcores, and dark grey, flat to ripple-laminated silty shale and siltstone with small load or flute casts in upper cores. These may represent turbidite flows (Gatehouse and Cooper, 1986). Between the lower and upper levels is a breccia composed of well-rounded shale clasts, fossiliferous siltstone and carbonate pebbles, and pebbles of trachytic basalt. Uniserial graptolites beneath the breccia suggest an Early Ordovician age, while diplograptids above indicate Middle to early Late Ordovician (Öpik and Jones in Harrison and Greer, 1963). Webby (1978) suggested that the breccia may reflect uplift associated with regression of the Larapintine Sea in the Middle Ordovician (his Dullingari Movement). The pebbles were reworked from exposed Kalladeina Formation and may represent lowstand deposits, thus separating the Dullingari Group into two sequences (X. Sun, National Centre for Petroleum Geology and Geophysics, pers. comm., 1992).

The only other drillhole in which graptolites have been recorded is Pandieburra 1 (Daily in Harrison *et al.*, 1963). There, fragmentary graptolites of Early to Middle Ordovician age were found in dark grey, fine to medium-grained quartz sandstone with thin interbeds of black pyritic, micaceous shale, again of turbiditic aspect. The graptolitic shales were deposited in the deep marine Tilpatee Trough, which intervened between the Gnalta Shelf to the south and Innamincka Delta to the north (Webby, 1978, 1987; Gatehouse, 1986; Zang, 1992). There is some evidence of a major depositional break between the Early Cambrian Pando Formation and black shale of the Dullingari Group in southern Warburton Basin drillholes (e.g. Tinga Tingana 1, Daralingie 2). Non-deposition may have arisen from uplift during the Mootwingee Movement, with subsequent onlap resulting from major transgression of the Larapintine Sea during the Ordovician.



Photomicrograph of pebble conglomerate (some clasts are basaltic or fossiliferous) in the Dullingari Group at 2 988 m in Dullingari 1. Width of view is 16 mm. Photo 042433.

EARLY PALAEOZOIC OF THE WESTERN REGION

EASTERN OFFICER BASIN

D.I. Gravestock, M.C. Benbow, C.G. Gatehouse and G.W. Krieg

Geological setting

Three depositional sequence sets are recorded in the Cambrian of the Officer Basin. The Ordovician 'Munda Sequence' of Krieg (1973) disconformably overlies sequence set € 3, but its relationship to sequence set € 4/01 has not yet been established.

In Western Australia, four sub-basins (Gibson, Yowalga, Lennis and Waigen; Fig. 7.15) contain Neoproterozoic to early Palaeozoic sediments 8 km thick in places (Jackson and van de Graaff, 1981; Phillips *et al.*, 1985; Townson, 1985). In South Australia, depth to magnetic basement may be up to 6 km in the deeper axial zone of the Munyarai Trough, itself partly overthrust by a portion of the Musgrave Block (Milton and Parker, 1973).

The Munyarai Trough is flanked on the west by the north-trending Nurrai Ridge, a zone of nearly coincident positive magnetic and gravity anomalies. The gravity low west of the ridge is named the Birksgate Sub-basin (Fig. 7.15).

The Waigen Sub-basin is undrilled and poorly known but early Palaeozoic sediments in the remaining Western Australian sub-basins differ from those in the Munyarai Trough. The latter also contains Devonian sediments belonging to a much younger depositional cycle and separated by an unconformity from Ordovician strata.

The southeastern Officer Basin between the Munyarai Trough and Gawler Craton is a structurally complex region of northeast-trending major faults, thrust-wrench fault complexes and basement ridges (Karari Fault, Nawa Ridge, and Middle Bore and Ammaroodinna Ridges adjacent to the Manya Trough) which separate the Gawler Craton from intervening depressions (Tallaringa and Manya Troughs). Compressional deformation which affected this region during the Alice Springs Orogeny is discussed in a later section.

Thinner and shallower sediments shelve south beneath Cretaceous-Tertiary cover on the Murnaroo Platform, and extend to the south coast locally beneath Permian sediments and between the Coompana Block and Gawler Craton.

Sediments beneath the Permian in the northwest-trending Boor-thanna Trough of the Arckaringa Basin (Fig. 7.15) are regarded here tentatively as Early Cambrian. They probably linked the eastern Officer Basin with the Stuart Shelf but much of the sedimentary record has been lost in this region through Palaeozoic erosion. To the north, a former connection with the Amadeus Basin via the Eringa Trough is complicated by the east-west trending Bitchera Ridge which, with the Musgrave Block, was probably elevated during the Neoproterozoic and again in the Late Devonian to Early Carboniferous (see below).

Sequence set € 1

The Relief Sandstone and Ouldburra Formation (Brewer *et al.*, 1987) of the Marla Group (Benbow, 1982) comprise sequence set € 1 in the Manya and Tallaringa Troughs. Their presence in the Munyarai Trough is not yet established.

Contact relationships between the Relief Sandstone and younger units on the Murnaroo Platform and in the Manya Trough, transgressive-regressive facies in the Ouldburra Formation (Dunster, 1987), and alternating marine and non-marine facies in the Relief Sandstone (Gaughan, 1989; Gaughan and Warren, 1990), all suggest a gross cyclicity in which elements of sequences € 1.1 to € 1.3 can be recognised. They also imply that the Relief Sandstone is transgressive and diachronous over Precambrian strata (Fig. 7.2). Designation of lowstand, transgressive and highstand deposits, however, can only be applied in certain instances. These are

identified where possible, but otherwise each sequence is described below in general terms.

Sequence € 1.1 deposits

Sequence € 1.1 comprises Relief Sandstone and lower Ouldburra Formation in parts of the Marla area and the poorly known Tallaringa Trough. The Ouldburra Formation attains a maximum thickness of 1 114 m at its type section (Manya 6), of which 744 m represent sequence € 1.1. In the GILES map area, the Neoproterozoic Rodda beds remained exposed at this time. Since drillholes in the Marla area terminated in or above Relief Sandstone, the base of sequence € 1.1 has not been reached. The top of sequence € 1.1 corresponds to an upward passage from marine to intermittently then persistently exposed conditions (941 and 393 m depth in Manya 6 and Marla 6, respectively).

The Relief Sandstone is at least 79 m thick in Manya 6, where it is overlain conformably by the Ouldburra Formation. Gaughan (1989) described two facies suites in the Relief Sandstone in this drillhole. The lower suite is flat to planar cross-bedded, fine to medium-grained sandstone of predominantly aeolian origin. Similar sandstone in the upper suite has mud drapes, mud-chip intraclasts and occasional scour surfaces, and a fluvial origin is suggested.

Conformably above the Relief Sandstone and unique in the Cambrian of South Australia is a suite of halite-siliciclastic and halite-carbonate-siliciclastic cycles in the lower Ouldburra Formation. Dunster (1987) counted up to 78 halite beds, and provided detailed descriptions of the succession. A typical halite-siliciclastic cycle comprises, in ascending order, upward-fining sandstone to mudstone sometimes ripple cross-bedded with rare sulphate evaporites and displace halite hoppers, fine to very coarse-grained feldspathic sandstone, and grey to dark orange-brown bedded halite with bottom-nucleated texture. Carbonate in the more complex cycles occurs as dolomite mudstone and algal boundstone with anhydrite nodules, mudcracks and tepees, or as in thin caps on the halite, often eroded before the next sedimentation event.

The environment of deposition was a mosaic of isolated salinas on a shallow marine to exposed sandy mudflat bordered by aeolian dunes. Persistent subsidence in the Manya 6 area led to accumulation of 400 m of stacked cycles in an arid, low-relief coastal setting. Contemporaneous Relief Sandstone formed fluvial and aeolian deposits on the landward margin of the sandflat-salina complex. Both units in this area probably represent lowstand and early transgressive phases of sequence € 1.1.

Continued transgression led to the growth of subtidal algal boundstones, thrombolites and archaeocyath-calcimicrobe bioherms. The poorly preserved archaeocyaths suggest a Botomian age but greater precision is possible from trilobites in this part of the succession. Fragments from Marla 1 were assigned to *Eoredlichia*, *Wutingaspis* or allied redlichiid genera (Jago and Youngs, 1980). Well-preserved material from Manya 6 was subsequently confirmed as *Abadiella* sp. (Jago *et al.*, 1994b) which ranges from upper Faunal Assemblage 2 to lower Assemblage 3 and is thus close to the Atdabanian-Botomian boundary. These trilobites provide the only proof of age in the entire pre-Devonian Phanerozoic succession of the Officer Basin, and occur beneath the € 1.2 sequence boundary at a depth of 967 m.

Dunster (1987) also recognised sponge-like body fossils, sponge spicules and other skeletal debris. Archaeocyaths occur sporadically in bioherms and as reworked material over a 300 m interval in Marla 6.

Meteoric solution breccia appears abruptly in Marla 6, 30 m above the last archaeocyaths. Mudcracks, evaporite and fenestral dolostone at a similar stratigraphic position in Manya 6 point to subaerial exposure (Dunster, 1987). J. Dunster (Comalco Ltd, pers. comm., 1990) also suggested correlation with Marla 7 and Manya 3, and has indicated that these drillholes can all be tied to a widespread seismic reflector, a time line corresponding to the top of sequence € 1.1. Environments of deposition are shown on Figure 7.5.

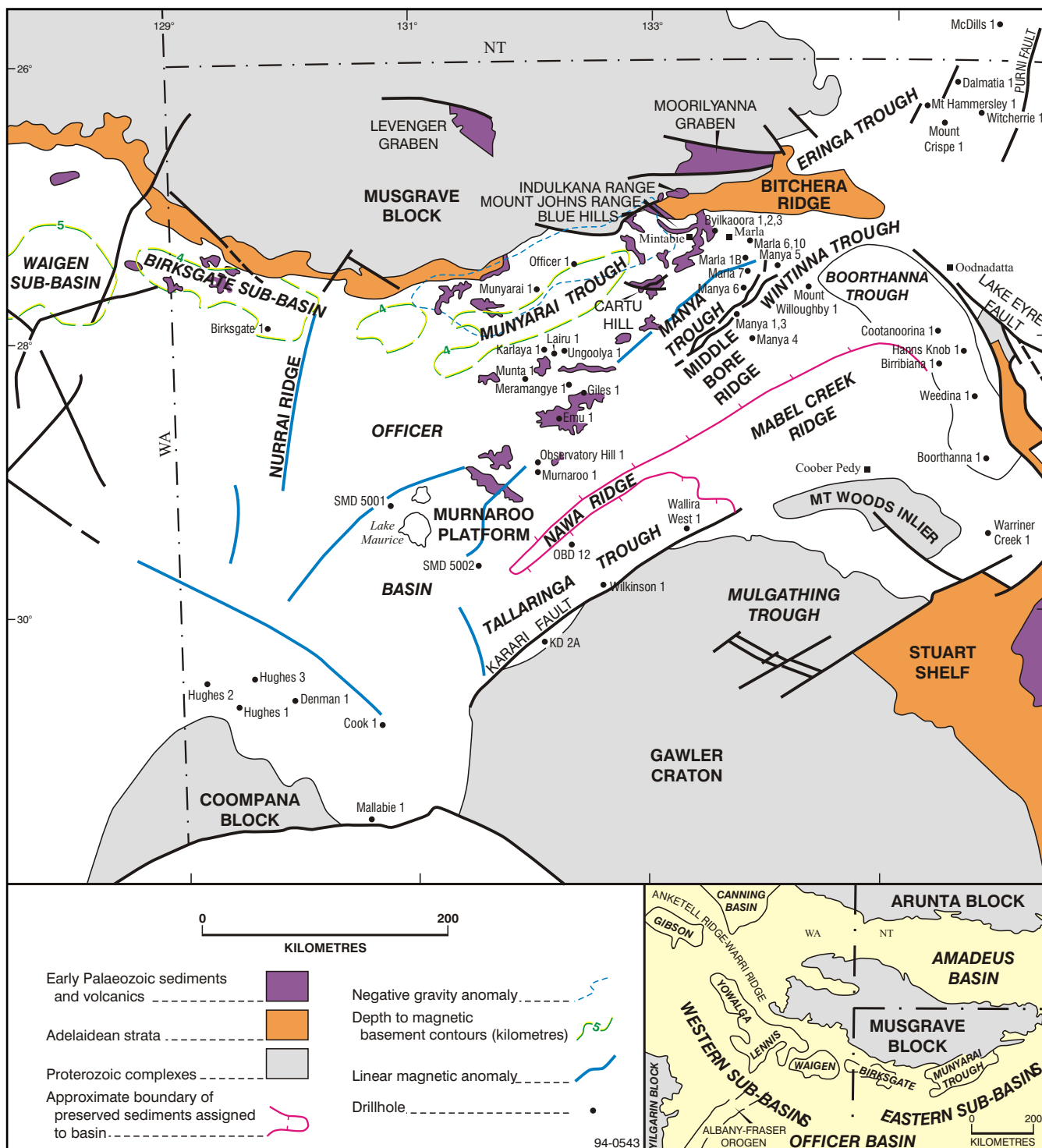


Fig. 7.15 Officer Basin location map.

Sequence € 1.2 deposits

Renewed transgression is marked in the middle of the Ouldburra Formation by subtidal fossiliferous nodular wackestone in Manya 6 and ooid-peloid-algal limestone in Marla 6. Although no more than 20 m of these facies are recorded in the GILES map area, and perhaps 140 m in Wilkinson 1, marine conditions are thought to have reached their maximum southwestern extent at this time (Fig. 7.6). Conglomerate and ooids in the basal Relief Sandstone in Observatory Hill 1 drillcore indicate marine reworking (Gatehouse and Hibburt, 1987; Gaughan, 1989). Gaughan also interpreted double mud drapes and somewhat cyclic 'bundle sequences' as tidal sand deposits above the basal Relief Sandstone in the GILES map area (Gaughan, 1989; Gaughan and Warren, 1990).

is absent from the GILES map area, thus this is the first sign of marine influence on the Murnaroo Platform and is considered here to represent the sequence € 1.2 transgression. Whether the sea penetrated as far as the Coompana Block is conjectural but it is noteworthy that Hughes 2 drillcore includes carbonates similar to those in the Ouldburra Formation.

A 6 m interval above the subtidal phase in Manya 6 comprises dolomitised ooid-peloid limestone, disrupted algal laminite, possible mudcracks and massive to 'chicken wire' anhydrite. These facies are overlain by sandstone and carbonate breccia marking subaerial exposure at the top of sequence € 1.2 in the vicinity of the Manya drillholes. Braided fluvial and aeolian facies characterise the Relief Sandstone in the GILES map area (Gaughan, 1989; Gaughan and Warren, 1990).

Sequence € 1.3 deposits

The upper Ouldburra Formation is chiefly a succession of redbed and carbonate sabkha deposits with evidence of repeated exposure and numerous 4-5 m thick upward-shoaling cycles or parasequences (Dunster, 1987). Typical carbonate sabkha parasequences comprise upward-fining couplets of carbonate mudstone with ripple laminations and rare marine fossils, stromatolite boundstone and evaporite nodules, gypsum-anhydrite beds, and mudcracked evaporitic, silty carbonate mudstone. The Relief Sandstone passed from shallow marine to fluvial facies in the GILES map area at this time (Gaughan, 1989). Dunster (1987) recognised that upper parts of the Ouldburra Formation were associated with regional regression which marks the top of sequence € 1.3, recorded also from platform areas of the Arrowie and Stansbury Basins (Figs 7.2, 7.10).

Cootanoorina Formation

The Cootanoorina Formation (Townsend and Ludbrook, 1975) is concealed beneath Permo-Carboniferous glacial sediments in the Boorthanna Trough (Fig. 7.15; see also Fig. 8.13). Cootanoorina 1, which investigated a sub-Permian high-speed seismic refractor, drilled 57 m of dolomitic sediments (Allchurch and Wopfner, 1973). A Devonian microflora recovered from clay shale cuttings between two coring runs in the dolomite (Harris and McGowran, 1973) appears to have caved from the overlying succession. While clasts in the disconformably overlying diamictite may include locally deposited but reworked Devonian sediments, there is no lithological connection with the dolomite beneath. The possible Devonian source for these clasts is discussed in a later section.

Other drillholes intersecting the Cootanoorina Formation in the Boorthanna Trough are Weedina 1 (type section) and Hanns Knob 1. Mount Willoughby 1 was located in the northern Wintinna Trough to intersect the dolomite refractor at shallow depth (Milton and Thornton, 1970). It encountered stromatolitic and porous intraclastic carbonates correlated with the Ouldburra Formation (Brewer *et al.*, 1987; Dunster, 1987).

The Cootanoorina Formation (Fig. 7.2) is underlain by silica-cemented quartz sandstone in Weedina 1 (Papalia, 1970) which may be Relief Sandstone equivalent. In Weedina 1, the 880 m thick Cootanoorina Formation comprises sandy peloidal dolomite, interbedded anhydrite and wavy algal laminite. Higher in the succession, slightly bioturbated carbonate mud and algal boundstone with reworked clasts predominate. The upper 300 m of the succession in Weedina 1 comprise pyritic shale, fine-grained sandstone and sandy ooid grainstone which may record the sequence € 1.2 transgression preserved beneath Late Carboniferous glacial sediments.

The lower part of the formation cored in Cootanoorina 1 comprises dark grey-green dolomitic, occasionally pyritic and magnesian lime mudstone, with cryptalgal lamination, disrupted bedding and reworked intraclast breccias. Halite casts are rare, but there are common interbeds and remobilised veins of anhydrite, possibly associated with movement of the nearby Mount Toondina 'diapir' (Allchurch and Wopfner, 1973). The upper part is flat to cross-bedded sandy dolomite with bedding rendered indistinct from growth of evaporite concretions. The gently dipping (25°) succession in Hanns Knob 1 is predominantly wavy and ripple-bedded red siltstone and very fine-grained sandstone. Pale grey-green clay drapes, tepees and mudcracks are common, and displacive evaporite concretions 2-3 mm in diameter are abundant. Thin beds, patches and seed crystals of anhydrite are present.

The carbonate sabkha to exposed sandflat environments in Cootanoorina 1 and Hanns Knob 1 are interpreted here as representing sequence € 1.3, but it is equally possible that the entire succession is Adelaidean (J.F. Allender, pers. comm., 1992).

Sequence set € 2

Continental interior environments in the lower part, and lack of diagnostic fossils in the marine upper part of sequence set € 2, hamper correlation with eastern seaboard basins. However, Gravestock and Hibburt (1991) have suggested correlation of non-marine deposits in the eastern Officer Basin with redbeds of the Billy Creek Formation in the Arrowie Basin (sequence € 2.1) as a result of changing coastline configuration caused by the Kangarooian Movements. The coastline is thought to have retreated north of the Officer Basin at this time (Fig. 7.10).

Sequence € 2.1 deposits

Sequence € 2.1, represented by the Observatory Hill Formation (Wopfner, 1969b), Wallatinna Formation and Arcoellinna Sandstone of the Marla Group (Krieg, 1973; Benbow, 1982), is entirely non-marine. The Observatory Hill Formation contains the world's only known Cambrian alkaline playa hydrocarbon occurrences (McKirdy and Kantsler, 1980; White and Youngs, 1980).

Wallatinna Formation

In the Marla area, the Wallatinna Formation (Benbow, 1982) disconformably overlies Proterozoic rocks. It underlies and intertongues with the Observatory Hill Formation and, on the northwestern margin of Mount Johns Range, it replaces the latter completely. However, the Wallatinna Formation is of restricted extent and is absent from most of the area of Cambrian deposition. Outcropping Wallatinna Formation, 260 m or more thick, consists of red-brown to green mudstone, siltstone, coarse-grained arkose and conglomerate. Rounded cobble and boulder conglomerate with sandy or carbonate matrix represents alluvial fan deposits sourced from active fault scarps northwest and southeast of the Manya Trough (Benbow, 1982).

Observatory Hill Formation

The Observatory Hill Formation crops out sporadically (Wopfner, 1969b) but has received detailed study in the subsurface since the discovery of oil and pseudomorphs after trona in playa sediments in Byilkaora 1 (Benbow and Pitt, 1979; Pitt *et al.*, 1980; White and Youngs, 1980). McKirdy and Kantsler (1980) and McKirdy *et al.* (1984) confirmed from oil and source rock geochemistry that the oil was formed *in situ* from bacterial precursors.

The Observatory Hill Formation is an alternation of redbed and minor fluvial sediments with three alkaline playa complexes. The oldest playa underlies and is partly eroded by cross-bedded arkosic fluvial sandstone at the type locality near Observatory Hill (units 18-21 of Wopfner, 1969b), where only the lower part of the formation is represented. The middle playa (Parakeelya Alkali Member) and upper playa (including Moyles Chert Marker Bed) are best preserved in drillcores. However, one or more of the three extend 250 km southwesterly from the Mount Johns Range, showing remarkable persistence of facies (Krieg, 1969, Brewer *et al.*, 1987). The Observatory Hill Formation can be mapped seismically over a considerably larger area (Stainton *et al.*, 1988). Except for the Oolarinna Member (Benbow, 1982), the intervening redbeds are unnamed and relatively little studied. The middle and upper playas are recorded from Munyarai 1, their most northwestern known occurrence (Womer *et al.*, 1987).

Cyclic sediments of the Parakeelya Alkali Member have received detailed study (White and Youngs, 1980; Southgate and Henry, 1986; Southgate *et al.*, 1989). Gross alternation between redbed and playa facies associations probably represents gradual rise and fall of the watertable whereas centimetre to decimetre scale cycles are interpreted as minor fluctuations in lake water level superimposed on longer term lake shore migration (Southgate *et al.*, 1989). These authors recognised four lithofacies; lake, saline mudflat, dry mudflat and sandflat.



Top: Flaggy micritic limestone from the alkaline playa facies, Observatory Hill Formation type section, ~10 km west of Observatory Hill. Photo 042434.

Bottom: Calcite pseudomorphs after trona in the Parakeelya Alkali Member at 320 m in Byilkaora 1. Photo 042435.

Lake sediments are laminated, grey-black dolomite mudstone relatively rich in organic carbon and with numerous calcite pseudomorphs after trona and shortite (sodium carbonate evaporites). Lack of evidence of exposure implies precipitation in shallow lakes. Alkaline conditions and high evaporation rates favoured growth of trona and shortite.

Nodular to laminated chert and dolomite mudstone with 'spotty' pseudomorphs are found in the saline mudflat where desiccation cracks, stromatolites and sinter crusts are characteristic. Crocodile skin texture, shrinkage cracks and penetrative evaporite crystal casts show that silica was precipitated initially as magadiite. Evaporative pumping favoured growth of surficial efflorescent crusts and evaporite precipitation in crack systems.

The dry mudflat facies is composed of graded siltstone to mudstone laminae with common erosion surfaces covered by stromatolite and sinter crusts and thin intraclast breccias. Mudcracks filled with quartz and dolomite silt and sand, some only one depositional unit thick, attest to frequent exposure and subsequent infill during sheet floods. Larger, more complex cracks may have facilitated brine

transport since 'spotty' evaporite pseudomorphs are confined to the muddy infills. Stromatolites and sinter crusts developed along spring watercourses.

Somewhat coarser, ripple-laminated sandstone, and calcareous and terrigenous mudstone dominate the sandflat facies. Lumpy stromatolites and sinter crusts are common; complex encrustations are sometimes found above mottled sand and mud-filled pipe-like structures which Southgate *et al.* (1989) interpreted as passageways for the migration of spring waters. Mudcracks indicate subaerial exposure.

Progress through the above facies associations from lake to sandflat documents gradual contraction of the lake (Lake Parakeelya of Southgate *et al.*, 1989; Fig. 7.10). Numerous facies oscillations recorded by these authors were presumably typical for the lower and upper playa systems as well. Eventually these passed into continental redbeds.

Benbow (1982) described the Oolarinna Member as a 22-45 m succession of red-brown, thinly bedded, sometimes conglomeratic calcareous claystone and siltstone coarsening upwards to very fine sandstone. Contact with the overlying formation varies from gradational to erosional. In the GILES map area, redbeds beneath the lower playa rest conformably or disconformably on Relief Sandstone and, further southwest, disconformably on Adelaidean Rodda beds (e.g. Murnaroo 1, Brewer *et al.*, 1987).

Arcoellinna Sandstone

This unit, like the underlying Observatory Hill Formation, extends beyond the Manya Trough to the Munyarai Trough, where it can be traced on seismic sections and wireline logs (Stainton *et al.*, 1988; Womer *et al.*, 1987). In outcrop in the Mount Johns Range, it is very fine to medium-grained, cross-bedded, red-brown sandstone with minor siltstone and claystone. On the northern margin of the range, however, fine siliciclastics are rare and basal granule beds disconformably overlie Wallatinna Formation. Benbow (1982) interpreted a mainly fluviolacustrine environment of deposition with some aeolian influence for this unit.

Sequence € 2.2 deposits

The Mount Johns Conglomerate (Krieg, 1973, redefined by Benbow, 1982) and Apamurra Formation (Benbow, 1982) comprise sequence € 2.2 in the Marla area and Mount Johns Range. There are, in addition, two isolated, unfaulted sedimentary successions in the Moorilyanna and Levenger Grabens on the Musgrave Block, which are north of the preserved limits of the Officer Basin.

The Moorilyanna Formation has been regarded as Adelaidean because of possible association with the Chambers Bluff Tillite (Wilson, 1952; Coats, 1963; Major, 1973a) and a similar correlation has been suggested for the Levenger Formation (Major, 1973a). A conglomerate mapped as tillitic by Coats (1963) overlies or forms the upper part of the Moorilyanna Formation, but shows no definite evidence of a glacial origin and hence no indication of a Sturtian age. There remains the possibility of a latest Neoproterozoic to Early Cambrian age as suggested by Thomson (1975), or even a Devonian age. For the present, the Moorilyanna and Levenger Formations are tentatively correlated with the Mount Johns Conglomerate and are therefore assigned to sequence € 2.2.

Mount Johns Conglomerate

The unit reaches 78 m in thickness and overlies Arcoellinna Sandstone sharply in all outcrops except in the northern Mount Johns Range, where it overlies Wallatinna Formation erosively (Benbow, 1982, fig. 4). Lithologies are poorly sorted, red-brown, pebble-boulder conglomerate and flat to cross-bedded feldspathic sandstone with conglomerate layers. Pebbles are mainly fine to coarse-grained siliciclastics different from those in the underlying Wallatinna Formation. An alluvial fan to fluvial setting, locally with northwesterly provenance, was suggested by Benbow (1982).

Apamurra Formation

Seismic surveys indicate that this unit can be mapped from the Murnaroo Platform into the Munyarai and Manya Troughs (Womer *et al.*, 1987; Stainton *et al.*, 1988). In contrast, the Mount Johns Conglomerate only occurs at its type locality (Benbow, 1982) and hence the Apamurra Member is elevated here to formation status. The Apamurra Formation comprises calcareous siltstone to fine-grained sandstone up to 70 m thick, thinly bedded, with minor limestone and medium-grained, well-rounded sandstone with herringbone cross-beds at middle levels. Trilobite traces, including *Rusophycus*, indicate a marine environment of deposition (Benbow, 1982).

Sequences € 2.1 and € 2.2 in the Officer Basin probably correspond with Kennard and Lindsay's (1990) sequence 3 in the Amadeus Basin.

Moorilyanna Formation

As recorded by Coats (1963), the Moorilyanna Conglomerate (herein Moorilyanna Formation) consists of thin, lenticular, basal slate and dolomite, then quartzite overlain by ~500 m of pebble conglomerate and coarse arkosic grit and siltstone. Conglomerate at the type section is well rounded and entirely basement derived. Active faulting during deposition may have controlled sediment distribution and thickness. The unit is fault-bounded where it abuts basement and appears to thicken westwards across the faults (Coats, 1963).

Associated with the conglomerate are widespread but discontinuous exposures of other strata mapped as 'Sturtian Series Undifferentiated' (Sprigg *et al.*, 1959). These are composed of purple, green and grey slate and shale with minor quartzite, feldspathic quartzite and dolomite. Their stratigraphic relationship to the conglomerate is uncertain, but outcrops along the Stuart Highway east of the Moorilyanna Formation type area are lithologically indistinguishable although they are locally very highly deformed and mylonitised.

Levenger Formation

Like the Moorilyanna Formation, the Levenger Formation is probably of syntectonic origin. Major (1973a) reported a thickness of 15 500 m but the likelihood that this is an over-estimate due to deformation cannot be assessed at present. The formation comprises red shale and laminated siltstone, arkosic sandstone, pebble conglomerate which is locally epidotised, and dolomitic beds near the exposed top. Planar beds of varying thickness and cross-beds are characteristic. Pebbles are derived mainly from the Musgrave Block but acid and basic extrusive clasts of uncertain provenance are also present.

Sequence set € 3

The Trainor Hill Sandstone (Krieg, 1973), uppermost formation of the Marla Group, and the partly correlative Wirrildar beds (Major, 1973b) represent sequence set € 3 in the eastern Officer Basin, but the Kulyong Volcanics (Major and Teluk, 1967) are also assigned tentatively to upper € 3 or lower € 4/01 (Fig. 7.2). Lateral and vertical facies changes evident in outcrop and on wireline logs suggest that the Trainor Hill Sandstone may yield to sequence stratigraphic interpretation, but the formation has an erosional top. Its relationship with the Wirrildar beds is poorly known, and no age diagnostic fossils have been found. No subdivision is proposed for these reasons, and each formation is discussed separately below. The distribution of € 3 sandstones is shown on Figure 7.11.

Trainor Hill Sandstone

Exposed thickness in the Mount Johns Range is 420 m, but widespread erosion has progressively removed the Trainor Hill Sandstone, in some places entirely (Benbow, 1982). The formation occurs widely from Marla 10 in the northeast to Ungoolya 1 in the south (Fig. 7.11). Even in the subsurface, its upper contact is erosional; Womer *et al.* (1987) interpreted 370 m of lost section in Munyarai 1 compared to Ungoolya 1.

Trainor Hill Sandstone overlies the Apamurra Formation sharply and erosively at its type section. Lower units are cross-bedded, fine-grained sandstone. Red micaceous siltstone to claystone, ~60 m above the base, contains interbedded very fine sandstone with clay galls, mudcracks and horizontal 'worm burrows'. Poorly sorted claystone and sandstone 80-90 m above the base contain granules and pebbles of white quartz and claystone. The section to 300 m is fine-grained, well-sorted feldspathic sandstone with minor interbeds of coarser sandstone and red micaceous, trace-fossil-bearing shale. Carbonate is present in the top 70 m, sometimes forming thin, herringbone cross-bedded sandy dolostone.

North of the type section where the Trainor Hill Sandstone is only 120 m thick, coarse pebbly horizons are more frequent and bedding is thicker. The lower contact with Mount Johns Conglomerate may be transitional (Benbow, 1982); alternatively there may be a break in deposition as shown on Figure 7.2. The Trainor Hill Sandstone-Apamurra Formation contact in the Munyarai Trough is conformable (Stainton *et al.*, 1988).

Northern outcrops suggest fluvial influence whereas further south the succession may pass up from occasionally exposed tidal sandflats to a marine, tide-dominated deltaic complex.

Wirrildar beds

The Wirrildar beds overlie the Neoproterozoic Punkerri Sandstone unconformably in the BIRKSGATE and LINDSAY map areas although the contact is concealed. The poorly exposed Wirrildar beds, which may reach 2 700 m in thickness, are gently folded and underlie flat-lying Kulyong Volcanics with a slight (5°) angular unconformity (Major, 1973b). This important folding event, recognised by Major (1968, 1972, 1973b), was regarded by Thomson (1969) as an early phase of Delamerian deformation. Adjacent to the Musgrave Block, Adelaidean strata are more intensely folded and overlain unconformably by very gently dipping to flat-lying Ordovician rocks and the age of folding is not as clearly constrained.

Lower exposures of Wirrildar beds are coarse-grained to granule arkose and feldspathic sandstone. Most of the section is fine-grained, micaceous, feldspathic sandstone and sandy siltstone with flaggy dolomite in middle levels. Isolated outcrops of shale, red siltstone, feldspathic sandstone and silicified conglomerate are also assigned tentatively to the Wirrildar beds (Major, 1972).

Assuming true thickness to be only one-third the apparent stratigraphic thickness (measured on the basis of a single 25° dip: Major, 1973b), a considerable volume of sediment nevertheless accumulated prior to folding. General lithological similarity suggests correlation with Trainor Hill Sandstone and, in the absence of sedimentological data, a similar fluvio-deltaic to shallow marine setting is assumed.

Kulyong Volcanics

Sparse outcrops of 3 m thick, flat-lying Kulyong Volcanics occur on western BIRKSGATE (Major and Teluk, 1967; Major, 1968) 50 km northwest of Birksgate 1 drillhole. In contrast, the Table Hill Volcanics, 100 m or more thick, occur in outcrop and at depth over more than 20 000 km² in the western Officer Basin (Townson, 1985). K-Ar geochronology suggests an Early Ordovician minimum age of 485±20 to 475±20 Ma for the Kulyong Volcanics (Major and Teluk, 1967), compared to a poorly constrained Rb-Sr age of ~570 Ma for the Table Hill Volcanics (Compston, 1974;

Jackson and van de Graaff, 1981). The Table Hill Volcanics age determination included two samples from WAIGEN, adjacent to the BIRKSGATE map area (Jackson, 1978). Outcrop and geophysical evidence, and petrographic similarity, confirm that the Kulyong and Table Hill Volcanics comprise a single tholeiitic basalt suite (Jackson and van de Graaff, 1981; Townson, 1985). However, the accuracy of both age determinations is questionable.

The lower 2 m of Kulyong Volcanics are relatively soft, grey and vesicular whereas the upper 1 m is hard, red-brown, fine-grained rock with no recognisable layering. Petrographically it is a subophitic mesh of plagioclase and augite, the brown colour being due to films of haematite. There are no overlying rocks in contact with the volcanics but they are underlain conformably by 7 m of thinly bedded silicified ferruginous sandstone, silicified greywacke (possibly tuffaceous) and fine-grained micaceous sandstone with a thin (50 mm) limestone interbed. This section is distinct from the underlying folded Wirrildar beds although no contact is seen (Major, 1968).

Stratigraphic and event correlation

In the BIRKSGATE and LINDSAY map areas, outcrops of flat-lying Kulyong Volcanics and gently folded Wirrildar beds unconformably overlie Punkerri Sandstone. From field evidence, the flat-lying Mount Chandler Sandstone (and the equivalent Boongar beds of Major, 1974) overlies the Wirrildar beds-Kulyong Volcanics section. Since the Wirrildar beds differ from the Apamurra and Observatory Hill Formations but resemble Trainor Hill Sandstone lithologically, they are considered correlative. The Wirrildar beds are carbonate-rich in middle levels and apparently much thicker than the Trainor Hill Sandstone, which is carbonate-rich only in upper levels. The Wirrildar beds may preserve a younger section than Trainor Hill Sandstone.

The folding event is correlated here with the Mootwingee Movement in western New South Wales. The Kulyong and Table Hill Volcanics may be no older than late Middle Cambrian, and hence may not correlate with the Antrim Plateau Volcanics. The latter are overlain by late Early Cambrian fossil limestone (€ 2.2) in the Daly, Bonaparte and other northern basins (Cook, 1982). The volcanics in the Officer Basin result from stress-release after folding of the underlying section. Feeder dykes for the Table Hill Volcanics follow major northeast and northwest fault trends in the western Officer Basin (Townson, 1985). Renewed volcanism in the late Middle Cambrian is evident in the eastern Warburton Basin, parts of eastern Australia and Tasmania.

`Munda sequence'

The term `Munda Sequence' was introduced by Krieg (1973) to comprise Mount Chandler Sandstone, Indulkana Shale, Blue Hills Sandstone and Cartu beds (Fig. 7.2). Mount Chandler Sandstone was modified by Benbow (1982) to exclude the Byilkaora Formation which is locally the basal unit. The Mintabie beds, which overlie Cartu beds, have similar lithology and sedimentary structures to the Blue Hills Sandstone and may comprise part of the same major depositional unit. Because of uncertain stratigraphic relationships above the Indulkana Shale, no attempt is made at sequence subdivision. Possible correlations are discussed after separate descriptions of the constituent units.

The unconformity between the `Munda Sequence' and older units probably reflects a distant echo of the Delamerian Orogeny, as do similar breaks in the Amadeus Basin succession.

Byilkaora Formation

The Byilkaora Formation overlies erosionally thinned Trainor Hill Sandstone and, in the northern Mount Johns Range, it disconformably overlies Mount Johns Conglomerate. It is overlain gradationally by Mount Chandler Sandstone (Benbow, 1982). The unit, ~15 m thick, is confined to the flank of the Mount Johns Range and has not been recognised in the subsurface.

The formation comprises poorly sorted kaolinitic sandstone with a basal conglomerate and minor siltstone and claystone beds. The conglomerate has a sandy to silty matrix with clasts of sandstone and quartzite, lacking igneous and metamorphic lithologies that characterise clast assemblages in Marla Group conglomerates. At higher levels the unit is finer grained, cleaner and more thinly bedded. Minor cross-bedding is evident. The top of the formation is placed at the base of very well-sorted quartzose sandstone.

The unit represents a distal alluvial fan deposit at the base, passing up to a marine tidal deposit. A north to northwest provenance is suggested by southerly decrease in clast size of the alluvial deposits (Benbow, 1982).

Mount Chandler Sandstone

The Mount Chandler Sandstone overlies the Byilkaora Formation conformably in the Mount Johns Range but elsewhere is either disconformable on Trainor Hill Sandstone or older units of the Marla Group, or unconformable on folded Adelaidean strata. In outcrop, the Mount Chandler Sandstone approaches 300 m in thickness and extends at depth across the Munyarai Trough to the Birksgate Sub-basin. Outcrops of the Boongar beds straddling the Nurrai Ridge have the same distinctive lithology as Mount Chandler Sandstone with which they are correlated (Krieg *et al.*, 1976).

Lower beds consist of very well-rounded and sorted fine to medium-grained, porous quartz sandstone, though commonly silicified in outcrop. Planar cross-beds are outlined by heavy-mineral laminae. The upper section is more feldspathic; some beds have abundant vertical burrows of *Skolithos* and U-shaped burrows of *Diplocraterion*, including large forms up to 0.3 m across. In some outcrops 'pipe rock' bioturbation has completely disrupted finer sedimentary structures (Packham and Webby, 1969; Benbow, 1982). 'Pipe-rock' bioturbation and quartzose lithologies are common to the Mount Chandler Sandstone and the Pacoota and Stairway Sandstones in the Amadeus Basin, the latter formations ranging from latest Cambrian to Early Ordovician age. However, the very large, distinctive *Diplocraterion* burrows in the Mount Chandler Sandstone do not appear to be matched by a similar form in the Amadeus Basin (B.D. Webby, Sydney University, written communication, 1990).

The Mount Chandler Sandstone is interpreted as a tide-dominated shoreface sand with burrowed beds representing high-energy, shallow subtidal sandflats. Frequent reactivation surfaces and red, laminated sandy mudstone interbeds in Munyarai 1 suggest a fluvial component.



Bioturbated Mount Chandler Sandstone, Purndu Hills. Photo 042436.

Indulkana Shale

This unit is best developed in the Indulkana Range and along the northern flank of Mount Johns Range where it is up to 60 m thick and overlies Mount Chandler Sandstone quite sharply. It was not intersected in Birksgate 1 and only 10 m are present in Munyarai 1. The unit lenses out east of Cartu Hill and is not readily distinguishable from other thin shales in drillholes in the GILES map area.

The Indulkana Shale in its type area is maroon, occasionally green along bedding and joint planes, and is pale green, slightly calcareous shale at Blue Hills. White, flaggy siliceous sandstone beds 0.05–0.15 m thick with pale green clay galls are locally present near the base and top. Limestone lenses ~1 m thick occur in the Indulkana Range, and red sandy siltstone and micaceous silty sandstone beds may be present. Ripple marks and desiccation cracks have been recorded in some upper sandstone beds by Packham and Webby (1969) who considered the limestone lenses to be surficial calcrete. Some beds are bioturbated by *Skolithos* and a small form of *Diplocraterion*.

The environment of deposition is uncertain due to lack of data in the Munyarai Trough. In outcrop, the shale appears abruptly and shoaling is indicated by desiccation cracks in upper levels. It has a transgressive-regressive aspect and may represent either an inner shelf or deltaic environment of deposition.

The Indulkana Shale gives Rb–Sr whole rock ages of 460 ± 15 Ma (Webb, 1978) and 438 ± 10 Ma (Womer *et al.*, 1987); this part of the succession is therefore Ordovician.

Blue Hills Sandstone

The Indulkana Shale is overlain apparently conformably by the Blue Hills Sandstone which reaches 800 m thickness in outcrop and has a similar distribution to the Mount Chandler Sandstone. In Munyarai 1, 162 m of Blue Hills Sandstone are overlain unconformably by Devonian sediments.

Unlike the Mount Chandler Sandstone, the Blue Hills Sandstone is red-brown, fine to medium-grained, kaolinitic quartz sandstone with gritty and pebbly bands and well-sorted, clean intervals. Cross-beds and ripples are well developed, while convolute structures, desiccation cracks, *Skolithos* burrows and trilobite tracks occur sporadically. In lower parts of the Blue Hills Sandstone, upward-coarsening cycles 7–15 m thick consist of basal fine-grained clay-rich sandstone with shale interbeds, passing up to thinly bedded then cross-bedded quartz sandstone. Upper parts of the section at Blue Hills are characterised by large trough cross-sets up to 3 m thick, similar to those in the Mintabie beds. The Blue Hills Sandstone represents a return to the shallow marine and occasionally exposed conditions of a deltaic environment, possibly with some fluvial influence.

Cartu beds and Mintabie beds

When first mapped, the boundary between the Cartu beds and Mintabie beds at separate localities was thought to be structurally discordant (Krieg, 1973). Other workers (Packham and Webby, 1969; Pitt *et al.*, 1980; Townsend, 1981) suggested a conformable boundary, the discordance being explained by subsequent deformation. Detailed mapping by Townsend (1981, 1991) and drilling at Mintabie Opal Field suggested continuity from Blue Hills Sandstone to Mintabie beds. The Cartu beds thus probably represent the lower, and the opal-bearing Mintabie beds the upper part of the same unit.

The Cartu beds, which conformably overlie Blue Hills Sandstone, are thinly bedded to laminated, fine to medium-grained kaolinitic sandstone with interbeds of green, biotitic sandstone, siltstone, and red and white shale. The Mintabie section, which is 25–30 m thick, is predominantly cross-bedded, fine to coarse-grained, feldspathic, micaceous sandstone with numerous claystone interbeds. Upward-coarsening cycles occur, with prominent trough cross-bedding at the top of each cycle. The Cartu and Mintabie beds represent continued deltaic deposition supplied from a more proximal detrital source than either Mount Chandler or Blue Hills Sandstone.

Stratigraphic correlation

There are three possible stratigraphic interpretations of the 'Munda sequence' and Mintabie beds. Krieg (1973) considered that the sequence from Mount Chandler Sandstone to Cartu beds is continuous and of Ordovician age, with a break between the Cartu beds and interpreted Devonian Mintabie beds. However, there are lithological and sedimentological similarities between the Blue Hills Sandstone, Cartu beds and Mintabie beds, and the whole succession could be regarded as Ordovician. Alternatively, the textural and sedimentological contrast between the super-mature Mount Chandler Sandstone and the increasingly less mature Blue Hills Sandstone to Cartu beds succession may suggest that all units above the Indulkana Shale are Silurian–Devonian and related to inception of the Alice Springs Orogeny.

If the base of the 'Munda sequence' is Early Ordovician, then there is no known record of the Late Cambrian to Early Ordovician sequence set ϵ 4/01. Mount Chandler Sandstone may correspond roughly to the lower Dullingari Group, which is poorly age constrained on sparse graptolite evidence. The Indulkana Shale has been correlated with Horn Valley Siltstone in the Amadeus Basin (Pitt *et al.*, 1980) but there is no compelling evidence from the sedimentology of the latter unit (Elphinstone, 1989). The radiometric age suggests correlation with the upper Dullingari Group.

WESTERN Warburton BASIN

D.I. Gravestock

Geological setting

The western Warburton Basin covers an area of 150 000 km² in northern South Australia (Fig. 7.16). It is situated east of the Officer Basin and north of the Peake and Denison Inliers, and extends east to the Birdsville Track Ridge in depositional continuity with the Cambro-Ordovician in that region. Thus, by analogy with the better known eastern Warburton Basin, there was also continuous sedimentation at least in part with the Amadeus and Georgina Basins in the Northern Territory during the Cambrian and much of the Ordovician.

Cambro-Ordovician sediments occupy a region more or less at the junction between western 'epeiric' or intracratonic basins and those to the east with more persistently well-developed marine shelf-slope characteristics.

Elements of the three main depositional sequence sets which typify the South Australian Cambrian can be recognised, the oldest sequence set, ϵ 1, being intersected in McDills 1 and probably Thomas 1 drillholes in the Northern Territory.

The stratigraphic order of units shown on Figure 7.2 is based on sparse fossil evidence, lithofacies and sequence stratigraphic considerations, and on the perceived geometry of strata in thrust slices. Only a single drillhole in the Cambrian (McDills 1) and one in the Ordovician (Colson 1) have penetrated two successive units where the boundary between them is not complicated by faulting. Further drilling may show the assumed order of strata to be incorrect and may also reveal new units in the early Palaeozoic succession.

The region is concealed by late Palaeozoic and younger sediments, and meagre information is available only from geophysical surveys and drillholes. As a result it is the least well known area dealt with in this chapter. Almost all of the geology described below is from beneath the Late Carboniferous–Early Permian Pedirka Basin where glaciogenic rocks are underlain by Cambrian–Devonian strata (see Fig. 8.27) caught up locally in thrust zones. Cumulative stratigraphic thicknesses drilled to date exceed 1 000 m for Cambrian, and approach 400 m for Ordovician strata; neither succession has been fully penetrated. The Devonian units and the style and timing of deformation are characteristic of the Amadeus Basin discussed in a later section.

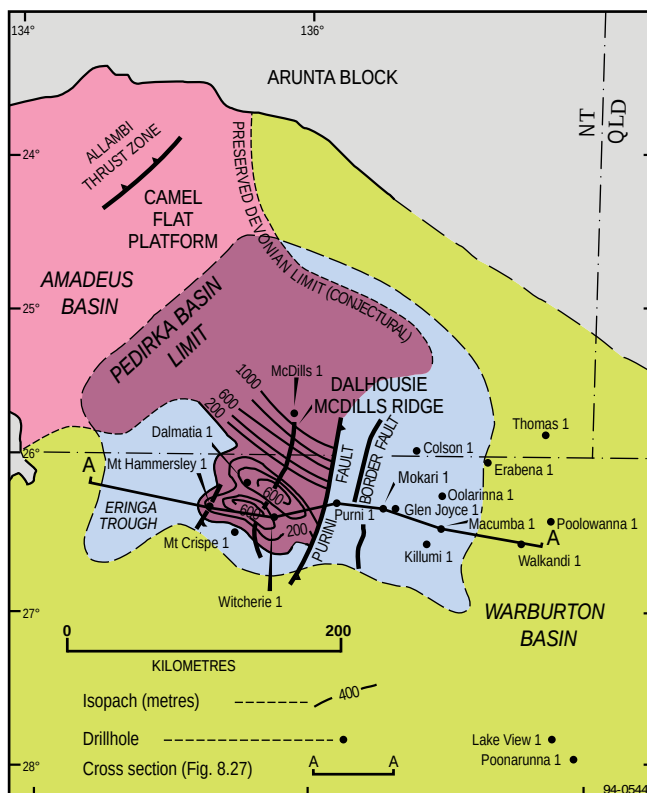


Fig. 7.16 Southeastern Amadeus and western Warburton Basins location map. Isopach is of Silurian-Devonian Mereenie Sandstone.

Sequence set € 1

This set is represented in McDiels 1 by at least 425 m of carbonate assigned to Todd River Dolomite on the basis of diagnostic shelly remains (Gilbert-Tomlinson in Amerada Petroleum Corp. of Australia Ltd, 1965). The fossils *Micrina etheridgei* and a brachiopod are characteristic of Faunal Assemblage 2 which elsewhere in the Amadeus Basin includes archaeocyaths and other shallow marine shelly remains (Daily, 1972; Kruse and West, 1980; Laurie and Shergold, 1985; Laurie, 1986).

In McDiels 1, the fauna is scattered within a 3 m core of irregularly banded sandy and shaly limestone ~180 m below the top of the carbonate. This carbonate succession alone is more than half the thickness of the combined upper Arumbera Sandstone and Todd River Dolomite in the Ooraminna Sub-basin (sequence 2 of Kennard and Lindsay, 1990). Wireline logs, however, suggest that a tripartite subdivision is possible in McDiels 1.

A lower unit 147 m thick comprises 'micaceous dolomite' with local shale laminae and vertical fractures filled with anhydrite. A sawtooth gamma ray log pattern suggests gross cyclicity on a 20 m scale. The fossil-bearing limestone forms part of a middle unit 222 m thick which is overlain by 83 m of grey, featureless dolomite. The boundary between the middle and upper units is marked by a sonic low-velocity 'spike' and the upper unit has a slightly slower sonic log response.

The Assemblage 2 fauna in the middle unit indicates sequence € 1.1 or lower € 1.2, which occur in the Arrowie, Stansbury and Officer Basins, in each of which there is widespread subaerial exposure at the boundary between these two sequences. Exposure is also evident at the top of the Todd River Dolomite in the Amadeus Basin (Kennard and Lindsay, 1990), thus the upper unit in McDiels 1 may disconformably overlie the fossil unit with the sonic excursion marking the exposure phase.

In the western Georgina Basin the Errarra Formation, which contains a more diverse Assemblage 2 fauna, is also separated from the overlying Arthur Creek Formation by a substantial hiatus (Laurie and Shergold, 1985; Southgate and Shergold, 1991), implying a lowstand of regional extent. The upper dolomite in McDiels 1 may be Chandler Formation (not to be confused with Mount Chandler Sandstone in the Officer Basin) or Giles Creek Dolomite. The cyclic lower unit may correlate with Namatjira Formation (Bradshaw, 1988) or may be a new carbonate equivalent to Woodendinna Dolomite in the Arrowie Basin. Tectonic thickening is also possible, however. The total carbonate package is regarded at present as Todd River Dolomite equivalent (Fig. 7.2).

Sequence set € 2

Sequence set € 2 is represented principally by mudrocks in McDiels 1, Dalmatia 1 and Glen Joyce 1 (Fig. 7.16). The succession is presumed equivalent to Chandler Formation or Hugh River Shale and has a drilled stratigraphic thickness of at least 112 m. The top is faulted against Ordovician Pacoota or Stairway Sandstone in McDiels 1 and is overlain disconformably by Mereenie Sandstone and Crown Point Formation (Silurian-Devonian, Carboniferous, respectively) in the remaining two drillholes.

The mudrocks are brick-red, haematitic and slightly micaceous with dark green reduction patches. Mud-flake intraclasts are common and desiccation cracks sparse. Thin linsen and starved ripples of fine-grained calcareous sandstone are locally abundant. Anhydritic mudrocks are underlain by 32 m of poorly sorted, micaceous sandstone with pebble beds above the Todd River Dolomite equivalent in McDiels 1. There are no fossils and material is not sufficient to choose between Chandler Formation, Hugh River Shale or even Giles Creek Dolomite as possible correlatives. Similar lithologies occur in all three units in the Amadeus Basin (Wells *et al.*, 1970; Kennard and Lindsay, 1990).

Sequence set € 3

Sequence set € 3 is represented by a distinctive bioturbated sandstone intersected only in Mount Crispe 1 and designated informally here as 'Mount Crispe beds' (Figs 7.2, 7.11). The steeply dipping (55°) unit is 540 m in stratigraphic thickness and is in interpreted fault contact with nearly flat-lying orthoquartzite above. Alternatively, this angular discordance could represent a Delamerian unconformity. The upper and lower depositional contacts are unknown.

The dominant facies prior to bioturbation originally comprised thickly laminated to low-angle cross-bedded, slightly feldspathic sand; very fine to fine-grained sand beds alternated with sheets of well-rounded very coarse quartz granules one depositional unit thick. Active burrowing by organisms, possibly crustaceans, excavated tubes up to 10 mm wide and at least 0.2 m long. Tubes are perpendicular to bedding and filled with packed quartz granules. Burrows are commonest in red-purple sandstone.

Buff-green, fine-grained sandstone with fewer, scattered granules forms a less common facies. Granule-packed burrows are smaller and less numerous but the sediment has been strongly bioturbated by constructors of oblique to horizontal fine sand-filled burrows resembling *Chondrites*. The least common facies is buff-green-grey very fine-grained sandstone with wispy shale laminae, virtually devoid of burrows. This passes up through the cf. *Chondrites* facies into the burrowed granule sands in vertical succession, suggesting an upward-shoaling package in part of the cored interval. The 'Mount Crispe beds' are correlated tentatively with Trainor Hill Sandstone and Wirrildar beds of the Officer Basin, and with middle Pertaoorrt Group of the Amadeus Basin, the latter being the original correlation (Jacque, 1966).

Ordovician sedimentary rocks

Pacoota or Stairway Sandstone

Quartzitic sandstone with horizontal to gentle dips was intersected in McDills 1, Mount Crispe 1, Witcherie 1 and Colson 1 (unnamed quartzitic sandstone, Krieg, 1985). It is in fault contact with the underlying Cambrian in two drillholes but its contact with younger units is unknown. Maximum drilled thickness to date is 293 m.

The unit is white-pale green-pink, fine to coarse-grained quartz sandstone with rare heavy minerals. Low-angle cross-beds with tangential toesets, and frequent reactivation surfaces, are typical in Witcherie 1. Cores from Mount Crispe 1 display in addition rare green shale partings, and reworked shale clasts, ripple-laminated sandstone and grey, fossiliferous siltstone. Fossils and traces include brachiopod fragments, an orthocone impression, small vertical burrows and rare sinuous trails, all pointing to a marine environment of deposition (Fisher *in* Jacque, 1966).

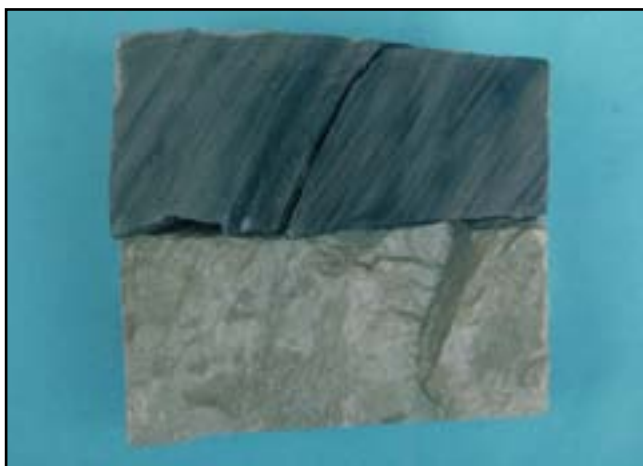
The fossils are Ordovician but no greater accuracy is possible. The unit is generally assigned to the Middle-Late Ordovician Stairway Sandstone (Jacque, 1966; Wells *et al.*, 1970; Youngs, 1976) of the Larapinta Group. However, a shaly siltstone described below may overlie the orthoquartzite. Dullingari Group in the eastern Warburton Basin also includes an orthoquartzitic-pyritic siltstone lithotype with Early-Middle Ordovician graptolites (Gatehouse, 1986). The orthoquartzite described above may correlate with this lithotype and hence the possibility remains of equivalence with Pacoota Sandstone, also of Early Ordovician age.

Dullingari Group

A moderately to more often steeply dipping fine-grained siliciclastic unit is widespread in the western Warburton Basin. Purni 1, Mokari 1, Oolarinna 1, Erabena 1, Walkandi 1, Poolowanna 1 and other drillholes have intersected this unit beneath the Permian or Mesozoic. Stratigraphic thickness is unknown but exceeds 85 m. Where the rock is steeply dipping it has a pervasive tectonic fabric, but there is no evidence of metamorphism beyond chlorite grade. Gently dipping intersections are not metamorphosed. The unit is tentatively correlated with the Dullingari Group (Gatehouse, 1986) but no fossils have been found to date.

The rock is dark grey-black, thinly bedded to laminated siltstone and shale. Shale-siltstone couplets of turbiditic aspect and slump-folded intervals suggest slope deposition in northern drillhole intersections. Further south in the vicinity of Lake View 1 and Mulapula 1, fine-grained, ripple-laminated, linsen-bedded and thixotropically loaded sandstone is common. A clastic source may lie in this southerly region.

Upper and lower sedimentary contacts are unknown. The unit is in fault contact with underlying sheared green phyllitic rock of presumed Proterozoic age in Walkandi 1.



Steeply dipping Dullingari Group shale unconformably above Proterozoic phyllite, at 3122 m in Walkandi 1. Width of core is 95 mm. Photo 042437.

MIDDLE PALAEOZOIC (DEVONIAN)

N.F. Alley and D.I. Gravestock

GEOLOGICAL SETTING

Adelaidean and early Palaeozoic strata in the Delamerian fold belt had already been welded to the stable Gawler Craton by the Silurian and Devonian while major mobile belts to the north and east were receiving sediments.

East of the craton, terrestrial to marine clastics filled major troughs or rifts in western Victoria and western New South Wales in the latest Silurian and Early Devonian. By Late Devonian time, alluvial-lacustrine clastics began to accumulate in the Darling Basin in response to foreland loading further east (Evans, 1977; Evans and Remus, 1990). The Tararra Trough, a southwestern extension of the Darling Basin, probably extended into South Australia where Devonian sediments are now known.

North of the Gawler Craton, an Early Devonian seaway spread from the Canning through the Amadeus Basin to link with basins to the east. In the Amadeus Basin, aeolian and fluvial systems supplied sediment to this seaway but, by the Late Devonian, thick molasse deposits were filling foreland depressions south of the Arunta Block, and alluvial-lacustrine sediments of the Finke Group were shed from the Musgrave Block. Compressive deformation of the Alice Springs Orogeny accompanied and eventually brought to a close this era of mainly terrestrial deposition.

In situ Devonian sediments are confirmed in South Australia from fossil evidence in the Munyarai Trough in the eastern Officer Basin and beneath the Nadda Basin. Some unfossiliferous sediments found at depth in northern South Australia have been assigned to the Devonian on lithological grounds and stratigraphic position. These include:

- siliciclastics beneath the Pedirka Basin (Wopfner, 1969a; Youngs, 1976), regarded here as a southeastern extension of the Amadeus Basin
- redbeds beneath the Triassic Walkandi Formation in the Simpson Basin (Moore, 1986a)
- a thin, acoustically transparent unit mapped seismically in the Boorthanna Trough (Allender *et al.*, 1987).

Devonian rocks and fossils have also been recycled into younger deposits; all of these occurrences are discussed below.

EASTERN OFFICER BASIN

'Munyarai sequence'

Munyarai 1 was thought at the time of drilling to have intersected Devonian sediments to a total depth of 2 899 m (Continental Oil Co. of Australia, 1969). This view persisted until recent work confirmed that the succession from 1 018 m to total depth is Cambro-Ordovician and Neoproterozoic. In spite of the most scrupulous sampling techniques, however, rare poorly preserved spores regarded as Middle Devonian (*Ancryospora*) have been extracted from drillcores (Womer *et al.*, 1987). These are now known to be giant Neoproterozoic acritarchs (Jenkins *et al.*, 1992).

Vertebrate scales in cuttings from 817 to 823 m were thought to have been from Middle Silurian to Early Devonian fish (Gilbert-Tomlinson *in* Continental Oil Co. of Australia Ltd, 1969). The fossil-bearing section has been extended to 738-860 m (J. Hibbert, SADME, pers. comm., 1987). Thelodonts, an acanthodian, teeth, plates and bone fragments were identified by Long *et al.* (1988) who regarded the fish fauna as probably Middle Devonian (Eifelian) in age.

The richest zones are in olive-green-grey, micaceous, calcareous mudstone with minor siltstone and fine-grained sandstone. Similar fish faunas are found elsewhere in Australia in both marine and freshwater environments but no conclusion favouring one or the other can be reached for the Munyarai 1 specimens. In the

Amadeus Basin, a comparable fauna is found in the basal Pertnjarra Group (Parke Siltstone) for which a lacustrine environment was suggested by Wells *et al.* (1970).

A Late Devonian (Frasnian) age is indicated, however, from palynology over the interval 634–870 m (Womer *et al.*, 1987), thus no greater accuracy than Eifelian to Frasnian can be determined from available fossils. Wireline logs suggest that the Devonian may extend in Munyarai 1 down to 1 018 m, the basal 106 m being planar to ripple-laminated, fine to medium-grained quartz sandstone. This disconformably overlies the Blue Hills Sandstone, but the contact has not been cored.

Munyarai 1 strata above 600 m depth consist chiefly of red-brown gypsiferous mudrock and sandstone. No age determination is possible, and the sediments could be Parke Siltstone equivalents.

SOUTHEASTERN AMADEUS BASIN

Mereenie Sandstone and Finke Group

Drillholes Witcherrie 1 in South Australia (Magnier, 1964) and McDills 1 in the Northern Territory (Amerada Petroleum Corp. of Australia Ltd, 1965) first penetrated Silurian-Devonian strata beneath glaciogene sediments of the Pedirka Basin (Youngs, 1976). Later, Mount Hammersley 1 and Dalmatia 1 (New, 1988a,b) confirmed Mereenie Sandstone from its distinctive though unfossiliferous lithology. Formations of the overlying Finke Group were also intersected in McDills 1, and possibly Macumba 1 and Erabena 1 further southeast (Moore, 1986a; Fig. 7.16; see also Fig. 8.27).

Subcrop area of the Mereenie Sandstone in this region is at least 8 000 km², including 5 000 km² in South Australia (Fig. 7.16). It underlies an area of 256 000 km² elsewhere in the Amadeus Basin (Wells *et al.*, 1970). The formation is bounded to the east by the Purni Fault and has a southern erosional limit near Mount Crispe 1. To the west the formation is terminated by the Mount Hammersley and Dalmatia Faults (here interpreted as thrusts) but may resume at depth further west in the Eringa Trough. The thickest intersection drilled to date is 459 m dipping at 35° in Dalmatia 1. It may exceed 600 m in depressions which cross the younger Dalhousie-McDills Ridge but is thin or absent on intervening ridges.

The formation consists of pink-white, flat to cross-bedded, fine-medium-grained quartz sandstone with traces of microcline, zircon and tourmaline. Coarse gritty bands are common while laminae and thin beds of red-brown micaceous shale are rare. Wells *et al.* (1970) suggested a complex of aeolian-deltaic-shallow marine environments with sediments sourced from eroding Ordovician sediments uplifted during the Rodingan Movement.

In McDills 1, the Mereenie Sandstone is overlain conformably by the Polly Conglomerate, Langra Formation and Horseshoe Bend Shale of the Finke Group. An interval originally believed to be Idracowra Sandstone (Amerada Petroleum Corp. of Australia Ltd, 1965) is assigned here to the Pedirka Basin following the drilling of Mount Hammersley 1.

The Polly Conglomerate, an alluvial fan deposit sourced from the Musgrave Block, crops out on the northwestern margin of the Pedirka Basin (Wells *et al.*, 1970). Clasts comprise pebbles and cobbles of granitoid rocks, siliciclastics and marble in a coarse sandstone matrix. The Langra Formation is cross-bedded sandstone, conglomerate, micaceous siltstone and shale overlain conformably by the evaporitic lacustrine Horseshoe Bend Shale. This red-brown to green shale may reach a considerable thickness in the undrilled Eringa Trough.

The Finke Group is a lateral equivalent of the Pertnjarra Group. Both were deposited during the Late Devonian although the upper age limit is unknown and continuation into the Carboniferous is possible.

Probable Devonian units in Macumba 1 (Wiltshire, 1978a) and Erabena 1 (Delhi Petroleum Pty Ltd, 1981) comprise brick-red silty shale, siltstone and thin interbeds of fine-grained rippled sandstone. They are assigned to the Finke Group (Moore, 1986a) but formation-level correlation is not possible with available data.



Mereenie Sandstone at 1211 m in Dalmatia 1. Width of core is 95 mm. Photo 042438.

BOORTHANNA TROUGH REGION

A seismically mappable unit underlies the late Palaeozoic sediments of the Boorthanna Trough and unconformably overlies folded Cambrian and older rocks (Allender *et al.*, 1987). This unit, possibly of Devonian age, fills topographic hollows on the unconformity surface. Allender *et al.* suggested correlation with the Cootanoorina Formation, (but the latter is regarded here as Cambrian), and also pointed to a possible intersection in Boorthanna 1. The intersection in this drillhole is currently regarded as lower Boorthanna Formation (Townsend and Ludbrook, 1975; see Fig. 8.13), although the basal 29 m may alternatively be Devonian. These sediments comprise pale brown, slightly to strongly calcareous shale with traces of gypsum and minor calcareous fine to medium-grained sandstone. Cobbles of reworked volcanic basement occur in the cuttings. Whether or not this is indeed a Devonian formation remains to be seen, but it does point to the possibility of a local provenance for the Devonian microflora reworked into the Boorthanna Formation in Cootanoorina 1 (Harris and McGowran, 1973).

DARLING BASIN

Underlying the Permian Urana Formation and overlying Adelaidean strata or Kanmantoo Group equivalents in the Renmark Trough area is a sequence of clastics believed to be of Devonian age (Thornton, 1974). It is probable that the sediments extend southwestwards from the Darling Basin in New South Wales. Such an inference gains support from the presence of over 1 000 m of current-bedded sandstone, grey-brown to green shale and slightly carbonaceous, micaceous siltstone tentatively dated as Devonian in Tararra 1 drillhole, in the Tararra Trough in western New South Wales.

Recent drilling in the Renmark Trough, which is on the same trend (Fig. 7.3), confirms the presence of Devonian strata at depth. Sediments intersected in Cooltong 1 are 191 m of fluvial and fluvio-lacustrine sandstone, siltstone and shale similar in many respects to the sediments in Tararra 1. Total thickness is unknown because drilling ceased within the unit. The unnamed basal clastics contain sparse palynofloras of Devonian (probably Late Devonian) age.

Although no other drillholes in the South Australian portion of the Darling Basin have intersected Devonian sediments, seismic information suggests that strata of this age may exceed 1 500 m in thickness in the north, decreasing to ~600 m in the south (Thornton, 1974). It is likely that sedimentation was continuous from the Tararra Trough. Seismic data also indicate that Devonian sediments may extend westwards across the Hamley Fault to the Canegrass Lobe area.



Photomicrograph of a Devonian Springsore (Ancyrospora Springs.) at 1379 m in Cooltong 1. The Springsore is 80 microns across. Photo 042439.

RECYCLED ROCKS AND FOSSILS

Evidence of recycled Devonian rocks and fossils in some Permian, Cretaceous and Tertiary basins has long intrigued geologists. In the absence of known Devonian strata in those areas the source was believed to be in undiscovered Devonian basins within South Australia or elsewhere in southeastern Australia.

Well-preserved Middle to Late Devonian spores have been recovered from Permian glaciogene sediments at Waterloo Bay on southern Yorke Peninsula (Harris and McGowran, 1971). Because of the preservation of fine ornamentation on some spores they were believed to have a source nearby, perhaps an intramontane lens (now eroded) in the Mount Lofty Ranges. However, palynomorph-bearing rocks can be transported long distances by glaciers and thus a source much further afield is possible.

Rare Devonian spores are also present in Cretaceous sediments of the Berri Basin, but these have probably been derived from Devonian rocks underlying Permian strata, or from extensive Devonian outcrops in Victoria. These rocks may have been a source for the recycled spores at Waterloo Bay.

Fossiliferous Devonian boulders occur in the Early Cretaceous marine Bulldog Shale around the southern margins of the Eromanga Basin. The provenance of the boulders has long been debated, although they are believed to originate from the Devonian Amphitheatre and Mulga Downs Groups cropping out in the Cobar area of New South Wales (Campbell *et al.*, 1977; Flint *et al.*, 1980). The boulders may have been transported northwestwards to South Australia during the Permian glaciation and reworked into the basal Bulldog Shale during the Early Cretaceous. The alternative possibility of reworking from formerly more extensive Devonian strata in the eastern Officer-Amadeus Basin region and from undiscovered strata at depth in the southern Warburton Basin region cannot be discounted.



Fish plate of Wuttagoonaspis Springs, in quartzite limestone in Bulldog Shale, 50 km north of 'Billalakina'. The plate is 22 mm long. Photo 042440.

EARLY AND MIDDLE PALAEOZOIC OROGENESIS

DELAMERIAN OROGENY

W.V. Preiss

Introduction

Sedimentation in the Adelaide Geosyncline (including Arrowie and Stansbury Basins, and Kanmantoo Trough) was terminated towards the end of the Cambrian by a succession of major tectonic events, named the Delamerian Orogeny and described by Thomson (1969) as 'fully developed in the southeast of the State where the fold movements continued with mounting intensity into the early Ordovician'. Tectonic subdivisions of the Delamerian fold belt are defined in Volume 1, page 24.

The regional structural and metamorphic synthesis by Offler and Fleming (1968) established three major phases of deformation in the Mount Lofty Ranges, based on studies of tectonic style, orientation and overprinting relationships. Subsequent, more detailed investigation by Mancktelow (1979, 1990) revised the metamorphic boundaries and reinterpreted parts of the deformational history. Structural analysis by Berry *et al.* (1978) in the Olary region concentrated on the basement, but recognised that the last two deformation phases affected the Adelaidean rocks as well and therefore record the Delamerian Orogeny.

Regional interpretations of the Delamerian fold belt by other authors stress different aspects of the orogeny. Campana (1958) viewed the fold belt as framed between rigid cratonic blocks; Thomson (1969, 1970) stressed basement control and vertical movements; Bell (1978a) traced the development of slaty cleavage across the Nackara Arc; Coward (1976) cited evidence for a major, north-northwest-trending basement shear zone. From descriptions of various Delamerian structures by previous authors, Preiss (1987) selected those that could be interpreted as reactivation of structures that had previously been active during Adelaidean and Cambrian sedimentation. Mancktelow (1979, 1990), amplifying the interpretations found in much of the previous literature, regarded the orogen as a relatively simple intracratonic fold belt, in which folding is mainly upright and due in large part to the first deformation. In contrast, the interpretations of Richert (1976), Clarke and Powell (1989), Steinhardt (1989) and Jenkins (1986, 1990) all stress the importance of thrusting, a view foreshadowed in the early work by Sprigg (1946) in the western Mount Lofty Ranges. Sprigg (1990) emphasised the role of megashears in controlling Adelaidean rifting and Delamerian deformation.

Extent of the Delamerian Orogeny

Delamerian folding affected Adelaidean and Cambrian sediments of the Adelaide Geosyncline, at least as far north as the Peake and Denison Inliers, but possibly extending into the northeastern corner of the Officer Basin. The western boundary of deformation is regionally well defined by the western edge of the Torrens Hinge Zone. The extent of the fold belt to the southeast is less certain. In the northeast, the Curnamona Craton remained undeformed, but is surrounded by belts of Delamerian deformation to the west (Flinders Ranges), south (Olary region) and east (western New South Wales).

In western Victoria, the Glenelg River beds, frequently compared to the Kanmantoo Group (Wells, 1956; Richards and Singleton, 1981), are intruded by granite plutons of Delamerian age (Turner, 1986; Foden *et al.*, 1990), and Delamerian deformation probably extends eastwards at least up to the Stavely Belt. A study of aeromagnetic anomalies in basement beneath the Murray Basin (Brown *et al.*, 1988) demonstrated a wide belt of curved anomalies which outline folds approximately parallel to the tectonic trends of the Nackara Arc; these may reflect a further eastward extension of Delamerian folding.

In western New South Wales, Wang *et al.* (1989) showed that the Late Cambrian to Early Ordovician Kayrunnera Group unconformably overlies tightly to isoclinally folded turbidites of possible Early Cambrian age. This folding, ascribed to the Mootwingee Movement, is therefore pre-Late Cambrian and may predate folding of the Middle to ?Late Cambrian Lake Frome Group in the central Flinders Ranges. It is uncertain whether folding of Adelaidean sediments in the Barrier Ranges took place in the Mootwingee Movement, or later in the Delamerian Orogeny. It is also difficult to exclude the possibility that the earliest Palaeozoic deformation in the Olary region reflects the Mootwingee Movement, especially as fold trends are sub-parallel to those of the Mootwingee area. Limited geochronological data suggest that pegmatite intrusion and retrograde shearing occurred in the Willyama Inliers near Broken Hill at about the time of Mootwingee and Delamerian events (Pidgeon, 1967; Etheridge and Cooper, 1981; Stevens *et al.*, 1990).

In the western part of the Warburton Basin, Williams (1975) interpreted Ordovician K-Ar age determinations on very low-grade, slaty metasediments as reflecting the Delamerian Orogeny. The age of sedimentation of these rocks has not been established; some may be Adelaidean, some Ordovician (see above). The angular discordance between the ?Cambrian 'Mount Crispe beds' and Ordovician Pacoota or Stairway Sandstone in Mount Crispe 1 may have resulted from Delamerian deformation.

A Proterozoic and Cambrian sedimentary and volcanic succession in Tasmania was folded by the Jukesian Movement, which may correspond to the Delamerian Orogeny, and is unconformably overlain by Ordovician clastic sediments of possible molasse facies (Webby, 1978).

Description of deformational structures

Previous literature on deformation phases in the Delamerian fold belt has used informal nomenclature (D_1 , D_2 etc.), which applies within each individual study area. The scheme proposed here attempts to place locally derived interpretations into a regional synthesis with some chronological significance, based on interpreted overprinting relationships. The symbols for all Delamerian structural elements are prefixed by **D**. Characteristics of the deformational phases in the various domains are summarised in Table 7.2.

DD₁

The oldest penetrative deformation of Adelaidean and Cambrian rocks in the Mount Lofty Ranges is recorded by slaty cleavage, folding and thrusting (Figs 7.17, 7.18). The cleavage here is mostly gently east and southeast-dipping although there are significant local variations. Cleavage is best developed in slate and fine-grained siltstone; folds are tight to isoclinal, reclined to recumbent, and there are numerous east and southeast-dipping thrust faults and ductile shear zones. The general parallelism of thrusts to the slaty cleavage suggests that thrusting is closely associated with **DD₁**, but some thrusts probably post-date significant early folding, since inverted fold limbs are commonly cut by thrusts, and some thrusts cut down-sequence into the cores of incipient anticlines. Variations in the orientation of folds, cleavages and thrusts are in part due to overprinting by younger fold phases and some thrusts have clearly been folded. In addition, cleavage is commonly refracted into a steeper orientation in coarse siltstone, sandstone and carbonate beds; quartzite frequently does not display true cleavage, except in shear zones, but is strongly jointed. Detailed descriptions of **DF₁** fold styles in the Mount Lofty arm are to be found in Sprigg (1946), Talbot (1964), Offler and Fleming (1968), Mancktelow (1979, 1981) and Jenkins (1990).

As pointed out by Jenkins (1986, 1990), relatively narrow, gently dipping zones of highly foliated phyllite within the Burra Group of the Mount Lofty Ranges represent ductile thrusts with a generally northwesterly tectonic transport direction. Besides the thrust zones noted by Jenkins (e.g. at Clarendon), many other examples are being

documented by more recent work. Thrust faults probably originated in shear zones in the basement and propagated along shaly horizons in the Burra Group, where they appear as bedding-parallel shears, many of which were not recognised in early mapping. They also ramped over more competent arenaceous or carbonate units, for example the Stonyfell Quartzite and, at such locations, the faults are steeper and more obvious. The stratigraphic separation on individual thrusts in the Burra Group is not large, perhaps in the order of 0.5 km, but the total amount of lateral shortening may be much greater. One major thrust exposed in Black Hill Conservation Park near the western entrance to the Torrens Gorge (see Preiss, 1987, p.77) is likely to involve significant lateral transport of a gently folded sheet of Woolshed Flat Shale and Stonyfell Quartzite over more tightly folded Stonyfell Quartzite and Saddleworth Formation.

The change in tectonic style from thrust and recumbent-fold-dominated structures in the lower parts of the sequence to much more upright structures in the upper part (Umberatana and Wilpena Groups) appears to be a gradual one in the western Mount Lofty Ranges. The thrusts may be mostly blind, movement being largely taken up by asymmetric folds in the upper structural levels. On southern Fleurieu Peninsula, recumbent structures affect not only the Burra Group but also upper Adelaidean and Cambrian units.

In the southern extension of the Torrens Hinge Zone on Kangaroo Island, Cambrian sedimentary rocks are gently folded and overthrust, but essentially unmetamorphosed and uncleaved.

In the outer Fleurieu Arc, macroscopic **DF₁** folds are not commonly seen and the regional outcrop pattern is different from the imbricate thrust style of the western Mount Lofty Ranges. The dominant large-scale fold structures are more upright and were related by Offler and Fleming (1968) to younger fold phases, based on overprinting relationships. The amphibolite-facies rocks of the eastern Mount Lofty Ranges (Mount Barker arm) are characterised by strong schistosity sub-parallel to bedding. Bedding is commonly well preserved in arenaceous metasediments such as the Aldgate Sandstone and much of the Kanmantoo Group, despite high-grade metamorphism. The development of bedding-parallel schistosity may reflect recumbent isoclinal folding, especially in the relatively incompetent pelitic rocks of the Burra Group. Such folding would be difficult to recognise from mapped outcrop patterns in such a region of relatively poor exposure, since the effects of later deformations dominate the outcrop patterns and younging indicators are rarely preserved in the pelites.

Within the more arenaceous Kanmantoo Group, consistent younging directions have been determined from cross-bedding in many sections, suggesting that isoclinal **DF₁** folds are rare. Bedding-parallel thrusts may exist, as suggested by Jenkins (1990) and Steinhardt (1989), but the recognition of consistent stratigraphy over wide areas probably places constraints on the scale of such thrusting within the Kanmantoo Group; it is likely that thrust repetitions are restricted within formations.

The possible effects of **DD₁** outside the Mount Lofty Ranges are uncertain. It is tentatively suggested here that décollement may have taken place along particular horizons in the lower Adelaidean succession without involving penetrative deformation and folding of the whole sedimentary pile. The east-dipping thrust fault at the western margin of the southern Flinders Ranges (Preiss and Faulkner, 1984) may represent one of the major décollement zones associated with **DD₁**, where it intersects the present land surface. A westerly transport direction for this thrust is implied by the placement of Callanna Group over Wilpena Group.

Nairne Fault

This is an important structure forming the main boundary between Adelaidean rocks and the Kanmantoo Group between Tanunda and Mount Magnificent (Fig. 7.17). Originally mapped as a fault by Sprigg *et al.* (1951), Campana and Whittle (1953) and Sprigg and Wilson (1954), this boundary was reinterpreted as a normal sedimentary contact by Kleeman and Skinner (1959) and

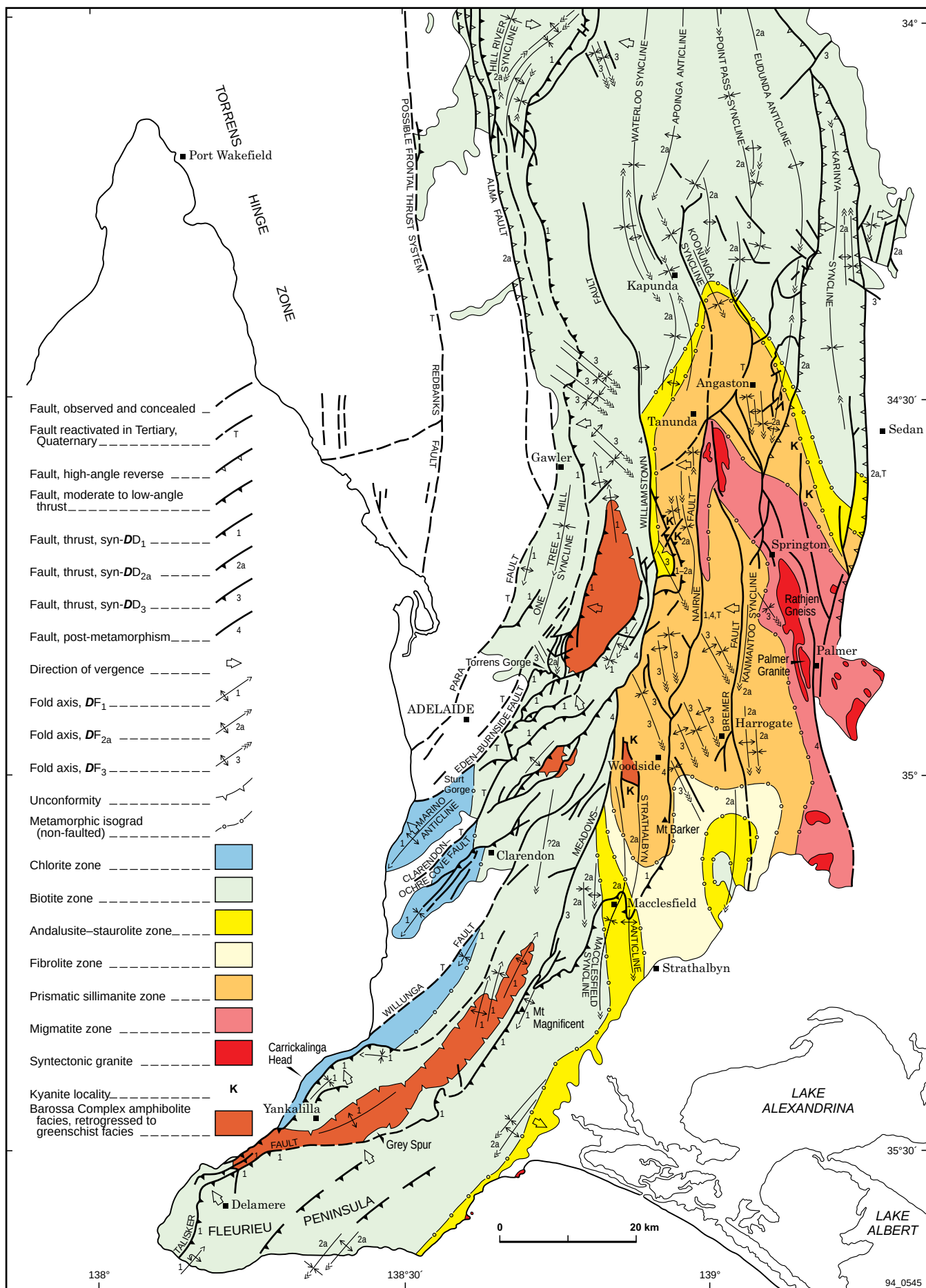


Fig. 7.17 Delamerian structural and metamorphic map of the Mount Lofty Ranges.

Table 7.2 Folding, faulting, fabric development, metamorphism, igneous activity and diapirism associated with the major phases of deformation in the tectonic subdivisions of

Tectonic domains	Stuart Shelf	Spencer Shelf	Curnamona Craton	Torrens Hinge Zone	Central Flinders Zone	North Flinders Zone Willouran arm
DD₁ (?Late Cambrian)						
<i>Folding</i>	None recorded	None recorded	None recorded	May be associated with thrusting at E margin	Recumbent folding in Callanna Group; timing uncertain	Recumbent folding in Callanna Group; timing uncertain
<i>Faulting</i>	Minor, steep	Local at E margin	None recorded	E margin overthrust	Décollement in Callanna Group	Décollement in Callanna Group
<i>Fabric</i>	None	None	None	None	None	None
<i>Metamorphism</i>	None	None	None	Burial only	Burial only	Uncertain
<i>Igneous activity</i>	None	None	None	None	None	?Intrusion of Bungadillina Monzonite
<i>Diapirism</i>	None	None	None	None	Possibly active	Possibly active
<i>Comments</i>	Stable, cratonic	Relatively Stable	Relatively Stable	Transition to fold-thrust belt	Bedding-parallel detachment	Age of décollement uncertain
DD_{2a} (?Earliest Ordovician)						
<i>Folding</i>	Minor, very gentle tilting: drag on some faults	Minor warping	Minor warping	Open folds with N-S axes, slightly asymmetrical to W	Upright, open, concentric folds with NNW-SSE axes	Upright, open to tight folds with NW-SE axes
<i>Faulting</i>	Possible faulting of Pernatty Upwarp	Steep faults along E margin	Possible steep faults, e.g. along Benagerie Ridge	Minor N-S faults	Minor	Major, steep NW-SE faults, in part strike-slip, e.g. Norwest Fault
<i>Fabric</i>	None	None	None	Minor radial fractures	Jointing only	Weak to moderate slaty cleavage DS_{2a} strikes NW-SE
<i>Metamorphism</i>	None	None	None	Burial only	Burial only	Lower to mid-greenschist facies
<i>Igneous activity</i>	None	None	None	None	None	Uncertain
<i>Diapirism</i>	None	None	None	None	Possibly active	Possibly active
<i>Comments</i>	Stable, cratonic	Relatively stable	Relatively stable	Transition to fold belt	Initiation of folding	Major folding
DD_{2b} (?Early Ordovician)						
<i>Folding</i>	None recorded	None recorded	None recorded	None recorded	Upright, open, concentric cross-folds with ENE-WSW axes	E-W cross-folds in Willouran Ranges
<i>Faulting</i>	Rare	Rare	Rare	Uncertain	ENE-WSW trending cross-faults	Minor
<i>Fabric</i>	None	None	None	None	None	Rare cleavage
<i>Metamorphism</i>	None	None	None	Burial only	Burial only	Mid-greenschist facies
<i>Igneous activity</i>	None	None	None	None	None	Uncertain
<i>Diapirism</i>	None	None	None	None	Possibly active	Possibly active
<i>Comments</i>	Stable, cratonic	Stable, cratonic	Relatively stable	Transition to fold belt	Fold interference patterns generated	Local fold interference
DD_{3a} (?Ordovician)						
<i>Folding</i>	None	None	None	Local kinks, timing uncertain	None recorded	None recorded
<i>Faulting</i>	None	None	None	Local faults	Possible ENE-WSW wrench faults link diapirs	Possible movement on Norwest Fault
<i>Fabric</i>	None	None	None	None	None	None
<i>Metamorphism</i>	None	None	None	Burial only	Burial only	Probably waning
<i>Igneous activity</i>	None	None	None	None	None	None
<i>Diapirism</i>	None	None	None	None	?Major intrusive phase	Possible minor activity
<i>Comments</i>	Stable, cratonic	Stable, cratonic				Late orogenic effects only
Post-Orogenic (?Ordovician to Quaternary uplift and unroofing of the Delamerian Orogen)						
<i>Folding</i>	None	None	None	None	?Cainozoic upwarping	?Cainozoic upwarping
<i>Faulting</i>	None	None	None	None	Some faults reactivated during Cainozoic	Some faults reactivated during Cainozoic
<i>Fabric</i>	None	None	None	None	Fault breccia and gouge	Fault breccia and gouge
<i>Metamorphism</i>	None	None	None	None	None	None
<i>Igneous activity</i>	Lamprophyre intrusions	None	None	None	None	None
<i>Diapirism</i>	None	None	None	None	?Late-stage intrusions	?Late-stage intrusions

the Delamerian fold belt.

North Flinders Zone Arkaroola arm	Nackara Arc		Inner Fleurieu Arc		Outer Fleurieu Arc	
	Mid-North arm	Yunta arm	Penneshaw arm	Mount Lofty arm	Kangaroo Is. arm	Mount Barker arm
Uncertain	None recorded	None recorded	Tight folds with E-W to NE-SW axes	Tight to isoclinal reclined folds with N-S to NE-SW axes	Tight folds with E-W to NE-SW axes	Rare isoclinal folds of uncertain orientation
Shearing at cover-basement contact	Overthrust W margin	?Décollement in W in Callanna Group	N to NW-directed thrusts	N-W to W-directed thrusts	Possible thrust repetitions	
Local foliation	None	None	S to SE gently dipping cleavage	E to SE gently dipping cleavage	Schistosity DS_1 sub-parallel to bedding	
Uncertain	Uncertain	Uncertain	Greenschist facies	Greenschist facies	Upper greenschist to upper amphibolite facies; NNW-trending migmatite zone	
Uncertain	None recorded	None recorded	None recorded	None recorded	Intrusion of earliest granitoids and mafic dykes, e.g. Rathjen Gneiss	
Possibly active	Minor	Minor	None	None	None	None
—	No penetrative deformation		Major compressive deformation		Major fabric-forming deformation	
Upright, open to locally very tight folds with N-S to NNE-SSW axes	Upright, open to moderately tight <i>en echelon</i> folds with N-S axes slightly arcuate; main asymmetry to E	Local upright, open to tight folds with N-S to NW-SE axes controlled by early Adelaidean extensional faults	Uncertain	Open to moderately tight folds with N-S axes; upright or asymmetrical to W	Open to moderately tight folds with E-W to NE-SW axes; upright or asymmetrical to N	Open to very tight, rarely isoclinal folds with N-S axes; upright or asymmetrical to W or E
Movement on Paralana Fault, with reverse and strike-slip components	Steep N-S reverse and strike-slip faults	Local reactivation of early Adelaidean extensional faults	None recorded	Steep N-S reverse faults and shear zones	None recorded	Steep N-S reverse and strike-slip faults and shear zones
Moderate to strong slaty cleavage DS_{2a} strikes N-S to NNE-SSW	Moderate to strong slaty cleavage DS_{2a}	NW-SE cleavage only near basement inliers	None recorded	Local crenulation cleavage DS_{2a}	Crenulation cleavage DS_{2a} dips steeply WNW to NNW	Crenulation cleavage DS_{2a} strikes N-S, dips steeply E or W
Mid-greenschist to lower amphibolite facies	Mid-greenschist facies	Mid-greenschist facies	Lower to upper	greenschist facies	Upper greenschist to upper amphibolite facies	
Intrusion of British Empire Granite, Mudnawatana Tonalite	?Intrusion of 'Bendigo' Granite	Uncertain	None recorded	None recorded	?Intrusion of some granitoids	Intrusion of Palmer Granite, Summerfield Granodiorite
Possibly active	Possibly active	Possibly active	None	None	None	None
Major folding	Major folding	Folding local only	Minor refolding	Minor refolding	Major folding	Major folding
Dominant upright E-W folds	Rare NE-SW cross-folds	Dominant NE-SW to ENE-WSW upright, mainly open folds	None recorded	None recorded	None recorded	None recorded
E-W faults associated with folding	Rare NE-SW cross-faults	ENE-WSW faults associated with folding; ductile deformation along Anabama Lineament	Possible renewal of thrusting		Possible renewal of thrusting	
Strong E-W slaty cleavage	Rare weak second slaty cleavage, NE-SW striking	Dominant upright, ENE-WSW striking slaty cleavage	None recorded	None recorded	None recorded	None recorded
Mid-greenschist facies	Mid-greenschist facies	Mid-greenschist facies	Lower to upper greenschist facies		Upper greenschist to upper amphibolite facies	
Uncertain	None recorded	Possible intrusion	Uncertain	None recorded	None recorded	None recorded
Possibly active	Possibly active locally	Possibly active locally	None	None	None	None
Local fold interference	Apparently smooth transition from DF_{2a} to DF_{2b} axes but with local interference		Tectonic effects uncertain		Tectonic effects uncertain	
None recorded	Kinks with NW-SE axes; possible conjugate kink sets	Local kinks	Effects uncertain	Broad folds and local kinks; NW-SE to NNW-SSE axes	Effects uncertain	Local to regional open to tight folds with NW-SE to NNW-SSE axes
Possible movement on Paralana Fault	Local NW-SE strike-slip faults	NNW-ESE fracture zones	Possible local faults	Minor NW-SE faults	Possible local faults	Minor NW-SE to NNW-SSE faults
None	Local kink bands	Local kink bands	Uncertain	Local crenulation of DS_1 , DS_{2a}	NE-dipping crenulation cleavage DS_3	
Probably waning	Probably waning	Probably waning	Probably waning	Probably waning	Upper greenschist to upper amphibolite facies	
Uncertain	None	None	None	None	Possible local pegmatites	
Possible minor activity	Possible minor activity	None	None	None	None	None
Late orogenic effects only	Late orogenic effects only		Late orogenic effects only		Continuing deformation under high temperatures	
?Cainozoic upwarping	Possible upwarping during Cainozoic		Possible upwarping during Cainozoic		Possible upwarping during Cainozoic	
Some faults reactivated during Cainozoic	Some faults reactivated during Cainozoic		Some faults reactivated during Cainozoic		N-S faults displace isograds near Springton; some faults reactivated during Cainozoic	
Fault breccia and gouge	Fault breccia and gouge		Fault breccia and gouge		Fault breccia and gouge	
None	Contact zones around post-tectonic intrusions		None	None	None	Contact zones around post-tectonic intrusions
Hydrothermal activity at Mount Painter, granitoid plugs near Arkaroola	Post-tectonic granitoids in western Murray Basin; lamprophyre intrusions	Lamprophyre intrusions	None	None	None	Post-tectonic granitic and mafic intrusions of western Murray Basin
?Late-stage intrusions	None recorded	None recorded	None	None	None	None

as a transgressive unconformity by Campana and Horwitz (1955) and Horwitz *et al.* (1959). However, more detailed subdivision of the Kanmantoo Group by Daily and Milnes (1971, 1973) and mapping by Marlow (1975) and Toteff (1977, 1990) re-established the existence of the Nairne Fault. Drag indicates that the east block is downfaulted at Mount Barker (Toteff, 1990).

The continuation of the Nairne Fault south of Mount Barker is less certain. The contact between Adelaidean and Kanmantoo Group units is mostly concordant with bedding in adjacent formations, but lenticular limestone slivers (fault slices of ?Normanville Group) in Barker Creek, the cutting out of stratigraphic units both above and below the contact, and the presence of mylonite at Mount Magnificent (A.E. Grady, Flinders University, pers. comm., 1990) indicate a bedding-parallel fault. The fault is folded (Marlow, 1975), together with bedding and DS_1 schistosity, around the Macclesfield Syncline and Strathalbyn Anticline (DF_2 folds).

Near Grey Spur, Daily and Milnes (1971) interpreted Kanmantoo Group faulted against Tapley Hill Formation and Brighton Limestone. At Delamere, Daily (1963) described a continuous stratigraphic section below the Kanmantoo Group, through Normanville Group, Wilpena Group and Umberatana Group rocks. A reverse fault at the base of this succession has brought Tapley Hill Formation over Backstairs Passage Formation occurring to the west. On the south coast of Fleurieu Peninsula, the steep reverse Talisker Fault separates different units of the Kanmantoo Group on either side; Harvey (1989) demonstrated that the Kanmantoo Group is sheared against Ulupa Siltstone near the Talisker Mine. It is likely that all these faults are part of a continuous system that is, on a regional scale, structurally concordant with bedding.

These relationships can be partly explained as due to bedding-parallel thrusting during DD_1 . However, because the Nairne Fault mostly places younger rocks over older, and because there are marked changes in the thickness of stratigraphic units in the Kanmantoo Group across the Talisker Fault (Mancktelow, 1979), it is suggested that this fault system had a pre- DD_1 history as an extensional listric fault, active during deposition of the Kanmantoo Group. Strong compression in the inner and outer Fleurieu Arcs from the southeast during DD_1 caused bedding-parallel shortening along this listric fault, bringing Kanmantoo Group rocks back up against stratigraphically older but structurally higher Adelaidean units. It is also possible that, as with some other smaller scale thrust zones which cut down-sequence, thrusting took place after initiation of DF_1 folding.

The Nairne Fault has often been shown as the western boundary of the Kanmantoo Trough, and even as the expression of the 'Tasman Line' (Hill, 1951) or western boundary of the Tasman mobile belt within the Mount Lofty Ranges (Harrington, 1974; Scheibner, 1974). While the structurally concordant section of the fault is here interpreted as a Delamerian compressional reactivation of an original Cambrian extensional fault, it is not the western limit of Kanmantoo Group sedimentation, since the lower parts of the group are preserved to the west of the fault. This also precludes consideration of the Kanmantoo Group as totally allochthonous, since its stratigraphic relationship above the fossiliferous Normanville Group is preserved at Carrickalinga Head, Delamere, and on the south coast of Fleurieu Peninsula. The Nairne Fault is thus considered as a feature within the Kanmantoo Trough.

DD_2

Previous interpretations have regarded the long, slightly arcuate, relatively upright folds of the Mid-North arm of the Nackara Arc (Fig. 7.18) as first generation folds (Offler and Fleming, 1968; Mancktelow, 1979). Preiss (1987) pointed out the change in dominant fold vergence between the southern Mount Lofty Ranges and the Mid-North region, and related this to Coward's (1976) sinistral shear, which approximately coincides with O'Driscoll's (1973) structural corridor G2. Over most of the Mid-North arm, these folds do appear to be the earliest recorded structures, and the associated cleavage that is well developed in suitable lithologies is

a normal slaty cleavage. However, the following evidence suggests that the north-south folds actually represent an early phase of the second deformation, DD_{2a} :

- Within the western Mount Lofty Ranges, slaty cleavage DS_1 is commonly warped or crenulated on north-south axes, and there are some macroscopic north-south folds that have affected the cleavage and thrusts as well as bedding.
- In the eastern Mount Lofty Ranges (Mount Barker arm), major folds such as the Macclesfield Syncline, Strathalbyn Anticline and Kanmantoo Syncline, shown as F_3 folds by Offler and Fleming (1968), have been reinterpreted as F_2 (P.D. Fleming, Latrobe University, pers. comm., 1988). These folds all affect DS_1 which is parallel, or at a low angle, to bedding.
- The Kanmantoo Syncline is comparable in style and orientation to the Karinya Syncline in the northern outcrop area of the Kanmantoo Group, especially as expressed in tracing stratiform aeromagnetic anomalies (S. Rajagopalan, University of Adelaide, pers. comm., 1989).
- The Karinya Syncline is the southernmost of the long, north-south folds of the Mid-North arm; as with other folds of this belt, its rocks display a single, axial-plane slaty cleavage.
- On the east-dipping limb of a smaller scale syncline 4 km south of Angaston, west of and parallel to the Karinya Syncline, DS_1 sub-parallel to bedding is affected by two sets of crenulations (DF_{2a} and DF_3 folds). The stronger of these is associated with a crenulation cleavage which is axial-planar to the syncline, and dips to the west.

Although DF_{2a} folds are generally more upright than DF_1 folds of the Mount Lofty arm, there is a tendency for asymmetry in opposite directions across the G2 corridor, as was originally suggested for DF_1 . Most folds in the Mid-North arm of the Nackara Arc have west-dipping axial planes. Variations in the dip of cleavage, from very steeply east to relatively gently west, can mostly be ascribed to fanning or to rotation by tightening of folds after cleavage formation. West of the G2 corridor, in the Mount Lofty arm of the inner Fleurieu Arc, DF_{2a} folds commonly have steeply east-dipping axial planes. This feature is consistent with Odonne and Vialon's (1983) model for folding above a sinistral basement wrench fault, as is the left-stepping *en echelon* arrangement of successive synclinal closures within the G2 corridor near Kapunda, Clare, Gulnare and Murraytown.

The contrast between westward-directed thrusting and macroscopic folds inclined to the east at Depot Creek, at the western margin of the Nackara Arc, was suggested by Preiss (1987) to indicate more than one phase of deformation. It is here proposed that the folds are DF_{2a} , whereas the thrusting occurred during DD_1 . Consistent with this interpretation, A.E. Grady (Flinders University, pers. comm., 1988) has reported smaller scale, low-angle faults that have been folded by the regional north-south folds in the Quorn-Buckaringa Gorge area.

Most of the north-south faults of the Mount Lofty arm and Mid-North arm were probably active during DD_{2a} , although many may have had a previous extensional history during sedimentation, especially in the Torrensian and Cambrian. This may apply to the steeply dipping, cross-cutting section of the Nairne Fault north of Mount Barker, where it has a normal sense of movement and considerable brecciation. Late, high-level reactivation of this ancient feature possibly took place partly during uplift of the Delamerian fold belt, when the fault may have once again become extensional. No change in metamorphic grade has been determined across this fault, hence the main displacement may be due to DD_1 thrusting, predating high-grade metamorphism, while later movements involved only small displacements. The Meadows-Williamstown Fault, on the other hand, although possibly initiated in DD_{2a} or even during deposition, suffered its main displacement after metamorphism, as it sharply truncates the metamorphic isograds (Offler and Fleming, 1968; Mancktelow, 1979).

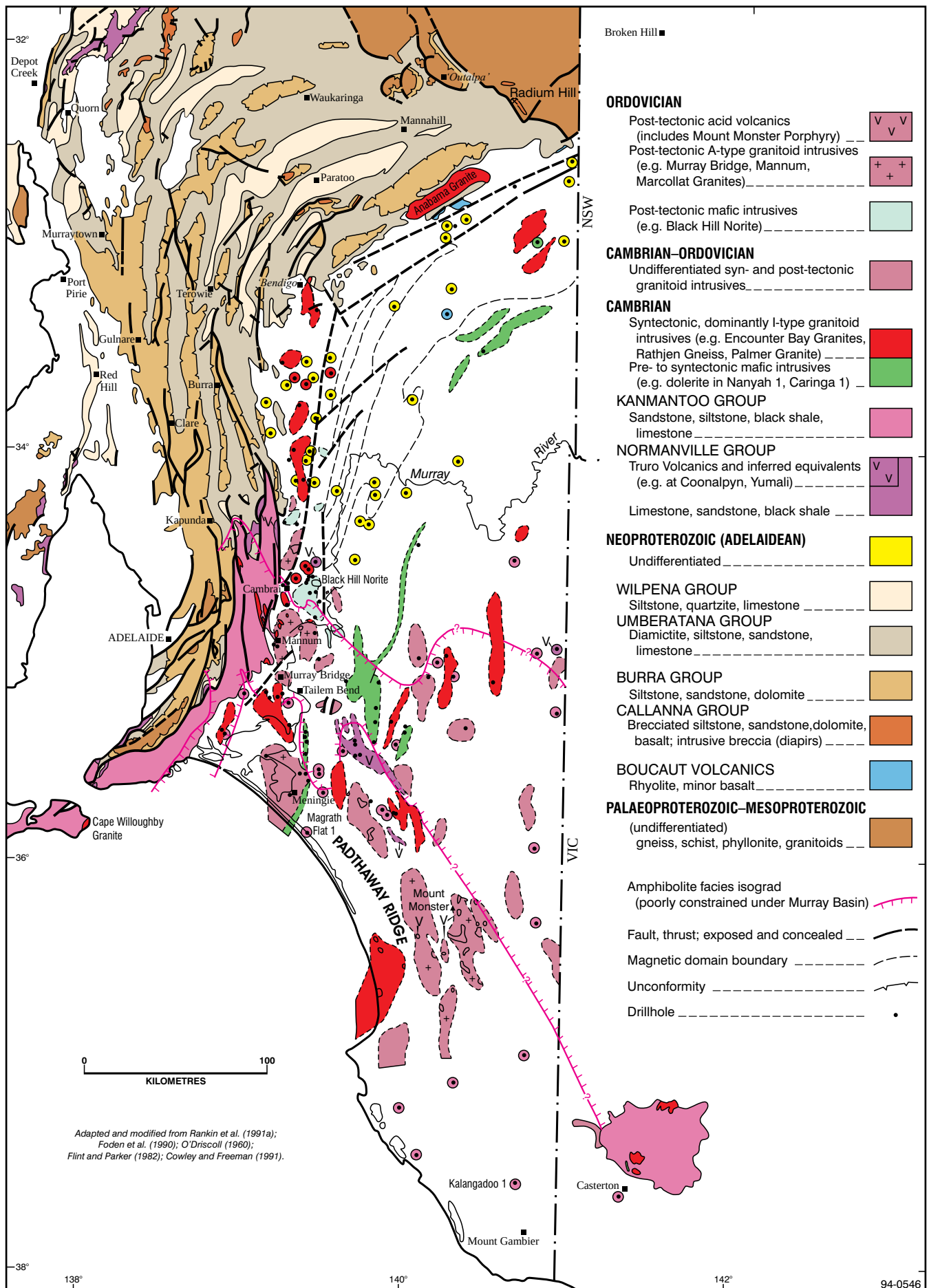


Fig. 7.18 Tectonic interpretation of the Nackara Arc, Fleurieu Arc and basement to the Murray Basin.

At least some of the north-south faults are likely to have undergone strike-slip displacement, possibly initiated during DD_{2a} (a transpressional origin is favoured for the north-south folds of the Nackara Arc), but later reactivation is also likely. A mylonite zone exposed in the Marne River near Sedan has gently north-plunging lineation and steeply plunging folds suggesting a strong dextral strike-slip component.

Continued displacement on some thrusts in the western Mount Lofty Ranges, during and probably after DD_{2a} , is indicated for the thrust in Black Hill Conservation Park, which displays intense, gently east-dipping crenulation cleavage within the shear zone, and has overridden a steep fault (? DD_2) and vertical crenulation cleavage.

In the North Flinders Zone, the northeast-trending folds of the Arkaroola arm, which Richert (1976) regarded as the older phase, are here assigned to DF_{2a} . The Paralana Fault has a long history of movement probably dating back to the Mesoproterozoic. During early Adelaidean rifting, it was active as an extensional or transtensional fault. It may have been reactivated as a steep reverse fault during DF_{2a} .

The dominant east-northeast-trending folds in the Yunta arm of the Nackara Arc are considered to represent a later phase of second-generation folding (DF_{2b} ; Fig. 7.18). In the 'Outalpa' region, Berry *et al.* (1978) showed that folds of northeasterly orientation overprint earlier folds with north to northwest-trending axes (see also Vol. 1, Ch. 4, which discusses Delamerian deformation of the basement). These latter folds may be DF_{2a} , by analogy with the Mid-North arm, while the younger folds are DF_{2b} . Alternatively, the north-northwest-trending folds could have been generated by the very early (or pre-) Delamerian Mootwingee Movement.

Preiss (1985, 1987) interpreted the mapped outcrop patterns around the Mount Grainger Anticline (Binks, 1968) as fold interference patterns; however, the relative ages of the east-west and north-south components have not been determined. In the north-south-trending section of the Worumba Anticline, both limbs are transected by east-northeast-striking cleavage (Preiss, 1985). These relationships suggest that DF_{2b} folding and DS_{2b} cleavage overprint an earlier meridional DF_{2a} fold.

Although DF_{2b} folds are dominant in the Yunta arm, there is some evidence of fold interference, especially near the boundary between the Yunta and Mid-North arms, but there is also a suggestion that some fold axes have been rotated. Interference of DF_{2a} and DF_{2b} folds is probably responsible for the dome-and-basin patterns of the Central Flinders Zone, as suggested by Richert (1976). However, the general absence of tectonic fabrics in this zone and the lack of unambiguous refolding of fold axes makes the relative timing difficult to establish independently.

Timing of fold events in the North Flinders Zone may become clearer when detailed structural studies are carried out, since cleavage is well developed over much of this zone. Meanwhile, the suggestion of Richert (1976) that the northeast-southwest folds (DF_{2a}) predate the east-west folds (DF_{2b}) is adopted. In the Willouran arm, weak east-northeast-trending folds (possible DF_{2b}) interfere with northwest-trending folds (DF_{2a}), but the relative timing is uncertain.

Although DD_{2a} and DD_{2b} are distinguished here, they may not have been widely separated in time. This is evident in the transition between the Mid-North and Yunta arms, where some fold axes appear as relatively smooth arcs continuous between the two arms, e.g. the Chace Range Syncline at the northwestern boundary of the Nackara Arc. This may indicate a continuous dragging of DF_{2a} folds into the DF_{2b} direction, whereas some adjacent folds show interference structures.

DD_3

This phase corresponds to folds labelled as F_2 in the Mount Lofty Ranges by Offler and Fleming (1968) but reinterpreted by P.D. Fleming (La Trobe University, pers. comm., 1988) as F_3 .

These folds, dominantly of northwesterly or north-northwesterly orientation, are here correlated with northwest-trending zones of kinking and fracturing that affect DS_{2a} in the Mid-North arm and DS_{2b} in Yunta arm. Such zones include the Teetulpa Fracture Zone (Forbes, 1991) and its northerly extension near the Waukaringa Goldfield, which are important in localising gold mineralisation. Morris and Horn (1990) have identified several sets of lineaments from a study of satellite images in the Nackara Arc; additional northwest-trending features include lineaments through Terowie, Mount Grainger and Paratoo.

In the Mount Barker arm, DF_3 folds are prominently developed in a north-northwest-trending belt between the Barossa Ranges and Woodside. This zone separates the relatively simply folded Karinya Syncline (DF_{2a}) from the Kanmantoo Syncline in which DD_{2a} structures affect DS_1 schistosity. An axial planar crenulation cleavage (DS_3) is generally steep, although some folds in the Barossa Ranges are quite strongly asymmetrical and have gently northeast-dipping cleavage. The zone of very strong DD_3 overprint is marked by the highest grades of metamorphism and by the intrusion of syntectonic granitoids, one of which, the granitic Rathjen Gneiss, has been intensely deformed by at least three events (DD_1 , DD_{2a} , DD_3). The high-grade zone probably remained at an elevated temperature during DD_3 , allowing ductile folding to take place, while the low-grade regions to the north were affected only by brittle deformation.

Mills (1973) mapped a number of thrusts, interpreted as syn- DD_{2a} , in the Warren Reservoir area. These shear zones are now mostly steeply dipping, and have been folded around the southern margin of the Warren Inlier of pre-Adelaidean basement by DF_3 folds of north-northwesterly orientation.

Late (or post-) Delamerian structures

Late tectonic effects associated with the Delamerian Orogeny, mostly faults and kinks, do not necessarily refer to a single deformation event. They occurred at high structural levels, possibly after considerable erosion of the original Delamerian mountains, and may relate to isostatic adjustments. For example, in the Mount Lofty Ranges, Offler and Fleming (1968) recognised rare kinks (their D4) that could not be related to earlier phases.

A number of north-south faults suffered at least their latest movements after metamorphism, since isograds are displaced, e.g. a fault west of Sedan (Mills, 1964; Sandiford *et al.*, 1990), and the Meadows-Williamstown Fault (Mills, 1973). Although the latter is generally poorly exposed, a road cutting west of Tanunda displays part of the brecciated fault zone. Between Tanunda and Woodside, this fault brings sillimanite-grade rocks in contact with the biotite zone. At its southern extremity, the Meadows-Williamstown Fault may merge with the southern extension of the Nairne Fault, but timing relationships have not been established.

Diapirism

Diapirs composed of carbonate-cemented breccia are concentrated in the Flinders Ranges, the largest bodies being in the Central Flinders Zone. It is likely that the evaporite-associated breccias were derived by disruption of mainly the Curdimurka Subgroup, but evidence for the timing is equivocal (see also Vol. 1, Ch. 6, and above for evidence of diapiric intrusion during the Adelaidean and Cambrian). Mount (1975) demonstrated that at least the later phases of breccia intrusion truncate Delamerian folds.

In the Worumba Anticline, Preiss (1985) showed that at least some of the deformation of the Callanna and Burra Groups, involving recumbent folding and thrusting truncated by a Sturtian unconformity, was pre-Sturtian; hence these events cannot be regarded as Delamerian. Parker (1983a) suggested that thrusting and isoclinal folding in the Callanna Group in the Rischbieth Well area of the Willouran Ranges resulted from deformation before deposition of the Burra Group. Another possible interpretation of this area is that the Callanna Group was deformed as a thrust duplex within a décollement zone during DD_1 and refolded, together with the Burra Group, by DD_2 .

Evidence for strike-slip faulting

Beside the evidence from the mylonite zone in the Marne River, which records early (possibly DD_{2a}) ductile deformation, there is local evidence for strike-slip faulting at higher structural levels in late stages of the Delamerian Orogeny.

In the Central Flinders and North Flinders Zones, a number of east-northeast-trending faults cut across the major fold structures (Coats, 1973a). N.M. Lemon (University of Adelaide, pers. comm., 1989) has suggested that these are sinistral strike-slip faults linking separate diapiric bodies. There is also evidence of strike-slip faulting in the Torrens Hinge Zone at Redhill (Reed, 1977), where a slickensided fault plane dips at 80–85° to the north-northwest, with two sets of fault striations plunging at 25°W and 8°E.

In the Burra Mine, gently north-plunging striations on the Kingston Fault (a north-northwest-trending fault at or about the eastern margin of the G2 corridor) indicate a strike-slip component. This fault juxtaposes mineralised and hydrothermally altered Skilloogalee Dolomite against unmineralised diapiric breccia (Drexel and McCallum, 1986). Because of associated volcanic-vent features (see also 'Delamerian igneous rocks'), these authors argued that mineralisation post-dated uplift and erosion of the Delamerian fold belt; hence the last strike-slip movement recorded by the Kingston Fault must be post-Delamerian. However, this may well be a reactivation of an old wrench fault which was important in transpressional DD_{2a} deformation as well as in localising the intrusion of the Burra Diapir and later volcanic-hydrothermal activity.

At the southwestern edge of the Willouran arm of the North Flinders Zone is the Norwest Fault, a strikingly linear, oblique-slip feature with marked lateral variations in amount and sense of stratigraphic separation across it (Coats, 1973a). Drag folds developed adjacent to the fault were first described by Sprigg (1949), and possibly indicate a dextral strike-slip component of movement on the fault during the Delamerian Orogeny. Associated south-directed thrusting was recorded by Scotford (1984) in the Mount Hack area.

Metamorphism

W.V. Preiss

Regional metamorphism associated with the Delamerian Orogeny has long been recognised to be of the low-pressure, intermediate to high-temperature type (Offler and Fleming, 1968; Daily and Milnes, 1971, 1973; Mancktelow, 1979, 1990; Sandiford *et al.*, 1990). Offler and Fleming mapped five metamorphic zones in the Mount Lofty Ranges, based on first appearances of index minerals:

- chlorite zone along parts of northern Kangaroo Island and the western edge of the Mount Lofty Ranges
- biotite zone in the remainder of the western half of the Mount Lofty Ranges and in the Mid-North
- andalusite zone over most of the outer Fleurieu Arc, except for the sillimanite and migmatite zones
- sillimanite zone in the eastern Mount Lofty Ranges between Murray Bridge and Angaston
- migmatite zone in the Palmer-Springton area.

Mancktelow (1979), with further sampling, readjusted some of the isograds and concluded that metamorphic biotite is present throughout Offler and Fleming's chlorite zone, including on Kangaroo Island. Nevertheless, the presence of translucent organic matter with low hydrocarbon maturity in a limestone sample from the Mount McDonnell Formation on the northern shore of Kangaroo Island (D.I. Gravestock, SADME, pers. comm., 1990) suggests less than greenschist grade alteration in this region, which is here regarded as part of the Torrens Hinge Zone, overlying shallow Gawler Craton basement intersected in Investigator 1 (Belperio and Hibbert, in prep.).

In the Mount Lofty arm, Cambrian and late Adelaidean rocks adjacent to the coastline between Sturt Gorge and Yankalilla have

a weaker cleavage and distinctly less metamorphosed appearance than biotite-grade rocks to the east and south, which are commonly phyllitic (see also comments by Daily, 1963). For these reasons, Mancktelow's metamorphic isograds have been modified by inclusion of a chlorite zone in the western Mount Lofty Ranges (Fig. 7.17).

Data for the basement to the Murray Basin are very sparse, but some generalisation on metamorphic grade is possible (Fig. 7.18). Most drillhole intersections north of the Murray River and south of the Padthaway Ridge suggest metamorphism to no more than greenschist facies (fine-grained rocks variously described as siltstone, slate or phyllite). Mafic rocks in Nanyah 1 may be of sub-greenschist grade, as they contain zeolite and prehnite (Preiss and Radke, 1989). There is evidence of amphibolite-facies metamorphism in a central belt extending from the amphibolite zone of the Mount Lofty Ranges in an easterly direction to Pinnaroo then south-southeasterly to the Glenelg River region of western Victoria. Within this zone are outcrops of amphibolite near Tailem Bend (Morris and Nichol, 1974) and migmatite northeast of Mannum; drillholes intersected garnet-staurolite schist west of Coonalpyn, and coarse-grained mica-schist at Bordertown. Turner (1986) described amphibolite-facies rocks in the Glenelg River region while, to the west, very low-grade metasediments occur beneath the Otway Basin (e.g. Kalangadoo 1) and along The Coorong (e.g. Magrath Flat 1).

Offler and Fleming (1968), Fleming and Offler (1968), Milnes *et al.* (1977) and Fleming and White (1984) presented evidence for early heating in the Mount Lofty Ranges, in the form of pre-tectonic porphyroblasts, early (pre- to syn- DD_1) granite intrusion, and intensely folded migmatite veins. Fleming and White (1984) suggested that migmatite in the Palmer area resulted from partial melting of Kanmantoo Group metasediments and that this process commenced prior to DD_1 . Mancktelow (1979, 1990) argued that porphyroblast relationships and the timing of granite intrusion are equivocal, and that most mineral growth occurred post- DD_1 and syn- DD_2 . Moreover, it is not clear that the schistosity that post-dates the porphyroblasts recorded by Offler and Fleming in the Kapunda area necessarily records the earliest deformation, since the dominant folds of that area are here interpreted as DF_{2a} .

Clarke and Powell (1989) and Jenkins (1990) suggested that the Kanmantoo Group was tectonically emplaced as a 'hot slab' which metamorphosed Adelaidean rocks beneath the overthrust contact. However, mapping of isograds by Mills (1964), Offler and Fleming (1968) and Mancktelow (1990) showed that isograds cut across both stratigraphic and tectonic boundaries and that, therefore, Adelaidean and Kanmantoo Group rocks were metamorphosed together. Moreover, almost half of the area of exposed Kanmantoo Group rocks is metamorphosed only to biotite grade. This suggests that the peak of metamorphism probably post-dated early Delamerian thrusting. Mancktelow's mapping in the Harrogate-Macclesfield area also shows that isograds have been folded by regional structures, here interpreted as DF_{2a} and DF_3 folds. This is the only area of the Mount Lofty Ranges with any correlation between metamorphic grade and stratigraphic level, the youngest rocks at the southern extreme of the Kanmantoo Syncline being at only biotite grade (Fig. 7.17). These observations are consistent with metamorphism persisting from DD_1 to DD_3 in the highest grade areas, but being mainly syn- DD_1 to the southwest.

Metamorphic conditions have been estimated from mineral assemblages (Offler and Fleming, 1968; Daily and Milnes, 1971, 1973; Mancktelow, 1979). Pressure conditions of around 4 kbars* were suggested by Offler and Fleming for the amphibolite facies rocks, while the highest temperatures attained in the migmatite zone were in the vicinity of 650°C. Mancktelow (1979) quoted pressure determinations of similar order (3.6–3.8 kbars and 4.4±0.5 kbars) and estimated temperatures greater than 450°C for the biotite zone, 540°C for staurolite, and ~640°C for migmatite.

* 1 kbar=0.1 Gpa

From a study of a cordierite-staurolite-muscovite association near Springton, Sandiford *et al.* (1990) concluded that prograde reactions occurred syn DD_2 at a minimum of 600°C and at pressures in the range 3-5 kbars. Metamorphism and intrusion of granite appear to be related to deep-seated heat sources, possibly controlled by basement structures. The greatest thermal alteration in the outer Fleurieu Arc (up to migmatite grade) is mainly confined within a linear zone which cuts obliquely across the dominant Delamerian tectonic trends (Coward, 1976) and which coincides with the G2 structural and gravity corridor of O'Driscoll (1973).

Outside the Mount Lofty Ranges, much of the Delamerian fold belt appears to have been metamorphosed to mid-greenschist facies (Fig. 7.19). Scattered samples of slaty, fine-grained rocks from various stratigraphic units throughout the Nackara Arc examined by Offler and Fleming (1968) and by the present author contain green biotite aligned along the dominant cleavage (here interpreted as DS_{2a}). Spry (1976) noted that despite variations in the degree of recrystallisation, metamorphic grade is constant along a transect across the Yunta arm from near Anabama Hill in the south to 'Outalpa' in the north, with quartz-biotite-muscovite assemblages, sometimes with carbonates and scapolite.

McKirdy *et al.* (1975) mapped a boundary between graphitic and sub-graphitic kerogen in organic-rich siltstone samples from the Tapley Hill Formation. This boundary coincides approximately with the first appearance of linear aeromagnetic patterns following certain iron-rich Adelaidean strata such as the Benda Siltstone and Ulupa Siltstone, due to the growth of metamorphic magnetite, and with a marked increase in the degree of illite crystallinity. The boundary also approximates Offler and Fleming's (1968) biotite isograd (Fig. 7.19).

The Central Flinders Zone generally lacks cleavage and obvious recrystallisation, although the sedimentary rocks of this domain are strongly lithified. They may not have attained greenschist facies metamorphism, although the absence of suitable volcanogenic components in these sediments makes recognition of very low-grade metamorphism difficult.

Metamorphism in the North Flinders Zone has been studied little, although McKirdy *et al.* (1975) showed this region as having graphitic kerogen. A strong cleavage is developed in parts of this domain, and metamorphism to biotite grade is likely. In an aureole surrounding the Mount Painter Inlier, metamorphic grade reaches amphibolite facies, with the development of diopside calc-silicate rocks in the Wywyana Formation and andalusite schist in the Woodnamoka Phyllite. Coats and Blissett (1971) illustrated a slightly wider zone of scapolitisation in Adelaidean rocks around the Mount Painter Inlier, the margin of this zone being displaced by the Paralana Fault (Fig. 7.19). Much of the metamorphism in the Mount Painter Inlier probably took place during the Delamerian Orogeny in association with the intrusion of granitic plutons; the pronounced gravity low coincident with the Mount Painter Inlier may be due to a hidden granulite at depth. Unlike the Palaeoproterozoic rocks of the inlier, the Mesoproterozoic sedimentary and volcanic rocks apparently had not been subjected to an earlier high-grade metamorphic episode.

Narrow contact metamorphic aureoles occur around some Delamerian granitoids which intruded low-grade country rocks, e.g. 'Bendigo' and Anabama Granite. Contact effects around the Bungadillina Monzonite of the Peake and Denison Inliers appear to be very limited (Morrison and Foden, 1990).

Delamerian igneous rocks

M.G. Farrand and W.V. Preiss

Compositionally bimodal magmatic activity accompanied the Delamerian Orogeny and persisted until after the last penetrative deformation, but emplacement of plutons and minor intrusives was confined to discrete areas of the fold belt. Magmatism associated with pre-orogeny extension (e.g. Early Cambrian volcanism) is not

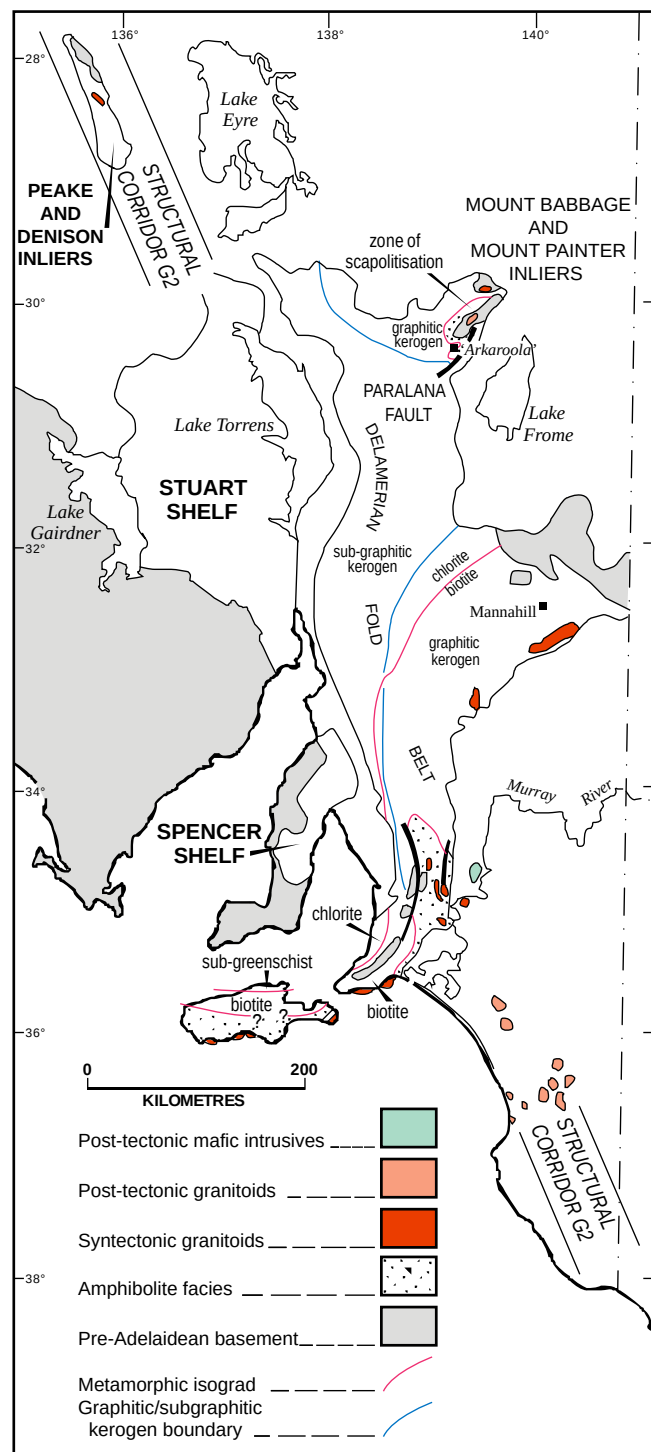


Fig. 7.19 Metamorphism and igneous intrusives of the Delamerian fold belt.

included in this discussion. Some granites and porphyritic acid volcanics in the Padthaway Ridge area post-date deformation; they have not been observed in contact with deformed rocks, although metasediments are known from nearby drillholes.

Distribution and structural control

Igneous intrusives known from outcrop and subsurface are shown on Figures 7.17 to 7.19. A sigmoidal belt of syn- and post-tectonic granitic and, less frequently, mafic intrusions extends along the eastern exposed edges of the Nackara and outer Fleurieu Arcs. Interpretation of scattered outcrops, drillholes and geophysical data indicates that a major granitic province extends eastwards beneath part of the Murray Basin. Igneous rocks beyond this province include the Delamerian granitoids and numerous small pegmatites

of the Mount Painter and Mount Babbage Inliers, the Bungadillina Monzonite and mafic dykes of the Peake and Denison Inliers and, possibly, a gabbro plug of uncertain age in the Willouran Ranges.

The syntectonic granite intrusions of the eastern Mount Lofty Ranges are aligned in a north-northwesterly direction along the G2 corridor, at the core of the zone of highest grade regional metamorphism (Figs 7.17, 7.19). Pegmatite veins and migmatitic metasediments, which may result from either partial melting (Fleming and White, 1984) or metamorphic segregation (Mancktelow, 1979), are widespread in the zone of highest grade metamorphism. At least some of the granite plutons along the sigmoidal belt are also syntectonic (Encounter Bay Granites, Cape Willoughby and other granites on Kangaroo Island and, possibly, the 'Bendigo' and Anabama Granites).

The absence of contact aureoles around some syntectonic granitoids has been attributed by Milnes *et al.* (1977) to regional heating prior to intrusion, estimated at a depth of 5 to 10 km. Granites that intruded metasediments at lower metamorphic grade ('Bendigo' and Anabama Granites) have discernible metamorphic aureoles.

In the Peake and Denison Inliers, the Bungadillina Monzonite (Ambrose *et al.*, 1981; Morrison and Foden, 1990) occurs around the intersection of G2 and a continuation of the Karari Fault, another major basement lineament (Rankin *et al.*, 1989). The British Empire Granite and Gordon Springs Granodiorite in the Mount Painter Inlier and Mudnawatana Tonalite of the Mount Babbage Inlier (Coats and Blissett, 1971; Teale, 1979), like intrusives of the Mount Lofty Ranges, are associated with a region of high heat flow. Post-Delamerian granitoids (e.g. The Pinnacles, 3 km northwest of Arkaroola) range in age from 450 ± 4 to 372 ± 2 Ma (G.S. Teale, University of Adelaide, pers. comm., 1992).

Within the sigmoidal granitic belt, the Anabama Granite is the largest pluton, elongated along the dominant east-northeasterly tectonic grain of the Yunta arm and parallel to the Anabama Lineament that lies just to the south. These features, with the Darling Lineament of New South Wales, lie within O'Driscoll's G7 corridor. Other plutons in the sigmoidal belt are less elongated and tend to be aligned mainly meridionally, following the dominant DF_{2a} fold axes of the Mid-North arm. The syntectonic granites of Encounter Bay and Kangaroo Island are aligned along the east-west to northeasterly trends of the outer Fleurieu Arc.

Granitoid plutons account for the greatest volume of magma intruded during and after the Delamerian Orogeny. Minor mafic dykes are common and geophysical data suggest that mafic or ultramafic intrusions of substantial volume may exist at depth. Small serpentinite outcrops occur in the Glenelg River region of western Victoria. Compositions intermediate between the granitic and mafic-ultramafic end members are rare and result from hybridisation.

Tectonic and age relationships of igneous activity

Although the earliest granitoids (~500-520 Ma) may have intruded before the onset of deformation (Milnes *et al.*, 1977; Morrison and Foden, 1990), the evidence for pre-tectonic intrusion is not conclusive. Table 7.3 summarises geochronological data and structural relationships for Delamerian igneous rocks.

The syntectonic granitoids are mainly of I-type, with rare-earth signatures suggesting a significant mafic source component, possibly derived from a mafic underplate developed during early Adelaidean rifting (Foden *et al.*, 1990). Some samples of the Encounter Bay Granites (Milnes, 1973) show marginally S-type characteristics; the crustal component may have been derived at the point of emplacement by digestion of xenoliths or at the point of magma generation by partial melting of deep crust (Milnes, 1990). Turner and Foden (1990) reported an I-type granodiorite with an S-type migmatitic aureole at Vivonne Bay, Kangaroo Island. On tectonic discriminant diagrams, the syntectonic granites fall in the volcanic-arc granite field.

Unlike the strongly foliated granitoids (e.g. Rathjen Gneiss) in

regions of highest metamorphic grade, the Encounter Bay Granites are only weakly foliated along the margins and massive in the inner zone. As Milnes *et al.* (1977) suggested, they may have behaved as relatively rigid bodies during deformation at less elevated temperatures than granitoids in the high-grade zone.

Syntectonic granitoids have been intersected in the Murray Basin region in MBT 14 (amphibolite and protomylonitic granodiorite), MBT 5 and 6 drillholes (Rankin *et al.*, 1991a), and occur in the South-East region of South Australia (Taratap Granite) and Glenelg River region of western Victoria (the I-type Wando Granodiorite and the migmatitic, S-type Harrow Granite; Foden *et al.*, 1990).

Liu and Fleming (1990) argued that the earliest mafic dykes in the southern part of the Delamerian fold belt intruded medium to high-grade metamorphics before or during the first deformation. The geochemistry of tholeiitic dykes suggests a mid-ocean-ridge basalt affinity, implying mantle derivation and intrusion under at least a strongly extensional, if not actually oceanic, regime. Others were interpreted as pre- to syn-DD₂ or post-DD₂. Rankin *et al.* (1991a) reported metadolerite and amphibolite from drillholes in the Murray Basin (MBT 11, 12, 14 and 15). Coarse-grained amphibolite with a foliated texture and igneous chemical characteristics occurs in limited outcrop and shallow drillholes east of Tailem Bend (Morris and Nichol, 1974) and near Meningie.

Post-orogenic granitoids occur mainly in the region of the western Murray Basin, both along the Padthaway Ridge and immediately east of the Mount Lofty Ranges. Most of this area is under Tertiary cover and no contact relationships are exposed. Foden *et al.* (1990) concluded that these granitoids are of A-type, distinguished by high silica content, low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and a rare-earth signature characteristic of mantle origin. The intrusives are commonly homogeneous, with interlocking, xenomorphic granular textures, and tend towards hypersolvus mineralogy. The magmas were highly fractionated, hot and dry, and intruded into high levels in the crust. Subsolvus mineralogy and rapakivi textures developed with later build-up of volatiles.

The A-type granites have higher $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratios than the syntectonic I-types and are characterised by smoky quartz. Rb-Sr whole rock dates range between 480 and 460 Ma with low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Webb, 1976), while K-Ar dates as young as 440 Ma probably reflect cooling ages. Foden *et al.* (1990) suggested that these magmas originated under a tensional regime as a second melt of refractory residue, after removal of I-type magmas during orogeny. The source is interpreted as the same mafic lithosphere which had underplated the base of the crust in the early Adelaidean, as produced the I-type syntectonic granites. On tectonic discriminant diagrams, the post-tectonic granites fall in the within-plate field.

The syntectonic Summerfield Granodiorite is cut by a diorite dyke with only a very weak fabric (Turner and Foden, 1990); this may be the earliest of the post-tectonic suite. Nearby, sub-horizontal, undeformed granite veins cut the migmatized and multiply deformed Kanmantoo Group. The Mannum and Murray Bridge Granites are relatively coarse grained, while the granite at Long Ridge near Sedan is finer grained.

Post-tectonic magmatic rocks of the Padthaway Ridge include the Marcollat Granite and porphyritic rhyolite at Mount Monster, which contain relict pyroxene and olivine. Some granites associated with the porphyry may be sub-volcanic intrusions. In western Victoria, the Dergholm Granite is also post-tectonic, and texturally and compositionally similar to plutons around the western margin of the Murray Basin.

The Black Hill Norite (Farrand *et al.*, 1989) is a large maficultramafic complex of mantle origin with a rare-earth signature almost identical to that of the Marcollat Granite (Foden *et al.*, 1990). Gravity and magnetic anomalies indicate that the complex may extend for 25 by 15 km at shallow depth. Drilling near Cambrai and Coonalpyn has revealed smaller, similar plutons concealed by Tertiary sediments.

The refractory composition of the Black Hill Norite may result from more complete melting and less fractionation than in the

Table 7.3 Summary of lithology, geochronology, geochemical type and structural relationships of major Delamerian syn-and post-tectonic igneous rocks.

Rock unit	Lithology	Geochronology (Ma)			Geochemical type	Structural relationships
		K-Ar	Rb-Sr	U-Pb		
Anabama Granite ¹	Coarse-grained biotite granite to granodiorite	455±7 (biotite)	463±3 (TR-K-feldspar Mu; IR=0.707) ² 458±62 (IR=0.707) ³		S-type to marginally I-type	Possibly syn-DD _{2b}
Pegmatite, Arkaroola Creek	Albite-microcline-muscovite pegmatite		450 (TR-Mu) ²		—	—
'Bendigo' Granite	Medium-grained granodiorite		454±84 (IR=0.707) ³	564±57 ⁴ (imprecise)	I-type	Possibly syn-DD _{2a}
Black Hill Norite	Coarse-grained gabbro, norite	486 ⁵ (biotite)	476±5 (IR=0.707) ⁵ 490±39 (IR=0.707) ⁶		Related to A-type granites ⁷	Post-tectonic
British Empire Granite ⁸	Fine to medium-grained muscovite granite ⁸		442±6 (IR=0.7592) ⁹		S-type ⁸	Post-tectonic
Bungadillina Monzonite ¹⁰	Quartz monzonite to syenogabbro ¹¹		521±35 (IR=0.7065) ¹¹	527±27 ¹²	A-type	Possibly pre-tectonic ¹³
Pegmatite, Cape Coultts	Pegmatite		505 (TR-Mu; IR=0.69) ¹⁴		I-type to marginally S-type	Possibly pre- to syntectonic
Granite, Cape Kersaint	Granite		467±3 (IR=0.718) ⁵		—	—
Rhyolite, Cape Gantheaume	Rhyolite dyke ¹⁵		—	488±6 ¹⁶	Marginally I-type	Possibly late to post-tectonic
Cape Willoughby Granite	Megacrystic granite	472-475 ⁵	—	508±7 ¹⁷	I-type	Possibly syntectonic
Encounter Bay Granites	Megacrystic and even-grained granite	417-474 ⁵	504±8 (TR-Mu-K-feldspar; IR=0.717) ⁵		I-type to marginally S-type	Possibly syn-DD ₁ , pre-DD ₂
Granite, Florieton	Granite; hornblende diorite ¹⁸ ; granitic gneiss		—		—	Possibly syntectonic
Gordon Sprins Granodiorite ⁸	Fine-grained granodiorite		IR 'low' ¹⁸		I-type ⁸	Syntectonic ⁸
Mannum Granite	Medium to coarse-grained biotite granite		481±9 (IR=0.7057) ⁶		A-type ⁷	Post-tectonic
Monarto Granite	Foliated fine-grained adamellite		—		—	Possibly syn-DD ₂
Mount Crawford Granite Gneiss ¹⁹	Highly foliated granitic gneiss		—		—	Pre- to syntectonic
Mount Monster Porphyry	Porphyritic rhyolite		478±23 (IR=0.7042) ³		A-type ⁷	Post-tectonic ⁷
Mudnawatana Tonalite ⁸	Oligoclase-microcline-biotite tonalite		427±133 (IR=0.7056) ⁹		I-type ⁸	Syntectonic
Murray Bridge Granite	Massive coarse-grained granite		471±12 (IR=0.7053) ³		A-type ⁷	Post-tectonic ⁷
Pegmatite, Penneshaw	Pegmatite		511±3 (TR-Mu; IR=0.7014) ⁵		—	?Pre- to syntectonic
Palmer Granite ²⁰	Weakly lineated granite		479±15 (IR=0.7086) ²⁰		I-type	Pre- to syn-DD ₂ ²¹
Rathjen Gneiss	Highly foliated granitic gneiss		—	516±4 ⁶	I-type	Pre- to syn-DD ₁
Granite, Long Ridge	Fine to medium-grained granite ²² ; leucocratic adamellite ²³		—		I-type ²² or A-type ⁷	Post-tectonic ⁷
Soda leucogranite, The Pinnacles, The Needles and Tourmaline Hill	Albite-quartz-muscovite granite		450±4	372±2 ²⁴	—	Post-tectonic ²⁴
Diorite, Reedy Creek	Mafic quartz diorite		—	497±10 ⁶	—	?Post-tectonic
Summerfield Granodiorite ²⁵	Granodiorite, monzonite, microgranite		—	486±6 ⁶	I-type ⁶	Syn-DD ₂
Tanunda Creek Gneiss ²⁶	Metamorphic and metasomatic alteration of Kanmantoo Group ²⁶		—		Alternatively, may be S-type granitoid	Syntectonic
Taratap Granite	Biotite granodiorite with mica foliation ⁷		475±65 (TR-K-feldspar; IR=0.7160) ³		I-type ⁷	?Syntectonic
Granite, Vivonne Bay	Pegmatite, granite, leucogranite, migmatite.		491±3 (IR=0.7184) ⁵ 523±6 (IR=0.7188) ⁶		I-type core, with S-type aureole ²⁷	Syntectonic

¹ Forbes (1991).

² Compston *et al.* (1966).

³ Webb (1976).

⁴ Fanning (1989).

⁵ Milnes *et al.* (1977).

⁶ Foden *et al.* (in prep).

⁷ Foden *et al.* (1990).

⁸ Teale (1979).

⁹ G.S. Teale and G.E. Mortimer, University of Adelaide, pers. comm., 1992.

¹⁰ Ambrose *et al.* (1981).

¹¹ Morrison and Foden (1990).

¹² Fanning (1987).

¹³ Xenoclasts of Bungadillina Monzonite are included in diapiiric breccia.

¹⁴ Fanning (1990a).

¹⁵ R.B. Flint and A.P. Belperio, SADME, pers. comm., 1992.

¹⁶ Fanning (1990b).

¹⁷ Fanning (1990c).

¹⁸ BHP Ltd (1978).

¹⁹ Mills (1973).

²⁰ White *et al.* (1967).

²¹ Mancktelow (1990).

²² Farrand (1986).

²³ Radke (1985).

²⁴ G.S. Teale, University of Adelaide, pers. comm., 1992.

²⁵ Gatehouse (1988b).

²⁶ Offler (1966).

²⁷ Turner and Foden (1990).

granitic rocks. Olivine gabbro, troctolite, pyroxenite and peridotite have been identified (Forwood, 1977; Wegmann, 1980) and a sulphide band was found to be enriched in platinum and palladium (Billiton, 1986). Drilling in the Black Hill area has revealed intense thermal aureoles in which country rock was hybridised and melted.

No metamorphic imprint has been identified within the Black Hill Norite. A patchy preferred orientation of plagioclase phenocrysts and ferromagnesian assemblages seen in outcrop has been ascribed to magma flow (Farrand, 1986). Evidence of high viscosity and a concentration of hydrous and alkaline minerals suggests that the exposed level is close to the original roof of the pluton.

Time scale of orogeny and igneous activity

Isotopic age determinations in general confirm the relative ages of igneous rocks inferred from their structural and geochemical characteristics. The age of the Rathjen Gneiss suggests that DD_1 may have commenced after ~515 Ma, but the weak fabric (DS_{2a}) of the Summerfield Granodiorite suggests that this deformation phase could have persisted as late as 490–480 Ma. Geochronology on post-tectonic granites suggests that penetrative deformation had ceased by 470 Ma. Milnes *et al.* (1977) argued from Rb–Sr and K–Ar mineral dates that low-temperature thermal effects persisted until ~420 Ma.

Other post-Delamerian igneous and hydrothermal activity

Other post-tectonic igneous and hydrothermal effects of uncertain age are recorded in breccia and associated hydrothermal deposits of the Mount Painter region, mineralised porphyry at Burra, and widespread lamprophyric intrusions of early Palaeozoic and Mesozoic age.

In the Mount Painter Inlier, Drexel and Major (1987) mapped the extent of the uraniferous Radium Ridge Breccias and their relationships to the Palaeoproterozoic and Mesoproterozoic basement. Clasts derived from basement gneiss and granite are set in a matrix containing comminuted granitic material and uraniferous haematite, in places with layers of haematitic breccia and chloritic breccia. They also contain local lenticular bodies of internal sediment, such as layered and graded-bedded siltstone and sandstone. The breccia body at the East Painter prospect is cone-shaped, but the form of other bodies is not known. Extensive potash metasomatism has resulted in the partial to almost complete feldspathisation of parts of the breccias. Near Mount Gee, several small dyke-like bodies of diamicite contain rounded clasts of porphyry, and superficially resemble Sturtian tillite. Unlike Adelaidean rocks surrounding the inlier, they are unmetamorphosed. Although of problematic origin, the diamicite appears to grade into chloritic breccia and may occur as pebble dykes.

On Mount Gee, Radium Ridge and Mount Painter, the 20–200 m thick Mount Gee Sinter is a rhythmically layered, vughy, quartz-haematite unit containing pseudomorphs after fluorite, stilbite, calcite, barite and pyrite. Late-stage quartz-encrusted prismatic laumontite crystals grew in large cavities. At the summit of Mount Painter, this unit forms a gently dipping capping over deeply kaolinised basement. Haematitic breccia locally intrudes the sinter sequence, suggesting that the breccias and sinter are related and broadly contemporaneous. Lambert *et al.* (1982) and Drexel and Major (1987) interpreted them to have resulted from an eruptive hydrothermal event after the Delamerian Orogeny: a breccia dyke cuts a pegmatite of possible Delamerian age, and the breccias are generally unfolded and unmetamorphosed, unlike the andalusite-grade Adelaidean rocks adjacent to the basement. The vughy nature of the sinter suggests emplacement near the present land surface, and hence that substantial erosion of the Delamerian mountain belt had taken place at the time of the hydrothermal event. Pidgeon (1979) obtained a very poorly constrained and probably inconclusive early Palaeozoic U–Pb date on monazite from the breccia. Recent palaeomagnetic determinations suggest an age as young as Permo–Carboniferous, at least for the Mount Gee Sinter (Idnurm and Heinrich, 1993).

The porphyry dyke at the centre of the Burra copper deposit, like the Radium Ridge Breccias, is associated with hydrothermal alteration and potash metasomatism (Drexel and McCallum, 1986). The presence of welded rhyolitic tuff (Scriven, 1977), flow-banding, trails of gas bubbles, devitrified glass shards, and one sample of agglomerate, suggest a very high-level intrusion (less than 1 km below surface: C.D. Branch, SADME, pers. comm., 1982). As the porphyry is in intrusive contact with the lower Adelaidean Skilloalee Dolomite, many kilometres of overburden must have been removed by erosion before the time of intrusion.

Lamprophyre (used in the broad sense of Rock (1986) to include kimberlite and lamproite) intrusions are common though poorly exposed in the Delamerian fold belt (Colchester, 1972; Stracke *et al.*, 1979). Those of the kimberlitic suite are Jurassic (see Ch. 9). Many of the age determinations are K–Ar and Rb–Sr dates on strongly altered rocks, but the Jurassic age is confirmed by one U–Pb date on zircon (Black *et al.*, 1993). On the other hand, Jurassic kimberlites cannot be the source of diamonds in a Late Permian conglomerate in the Springfield Basin (see Ch. 8), suggesting the presence of Palaeozoic (or older) lamprophyric intrusions. Morris (1991) recorded a suite of lamproite intrusions from the Truro–Frankton region which are petrologically distinct from the kimberlites and have yielded late Delamerian ages of 450 Ma (Webb and Lowder, 1971) and 481 to 478 Ma (Webb, 1990). Webb and Lowder also obtained K–Ar ages between 426 and 424 Ma on biotite from minette dykes in the Radium Hill area. Intrusion of the lamproites and minettes may therefore have been approximately synchronous with other post-orogenic igneous activity in the Delamerian fold belt.

Mineral deposits associated with the Delamerian Orogeny

W.V. Preiss and M.G. Farrand

Mineralisation related to igneous activity

The mineral potential of the Delamerian intrusives is partly within the plutons themselves and partly in country rock affected by convection systems driven by the heat associated with the intrusions. Granitoid plutons have potential for porphyry and vein-type deposits of gold, copper, molybdenum, bismuth, tin and tungsten but, apart from gold at the Kitticoola Mine in the Palmer Granite (Drexel, 1979) and minor copper in the 'Bendigo' and Anabama Granites (Morris, 1981; Langsford, 1971, respectively), little mineralisation has been discovered within granitoid plutons.

The copper orebody at Burra appears to be unique in that it is associated with a small, high-level porphyry intrusion and concomitant hydrothermal alteration.

Gold, silver and base-metal sulphides hosted by metasediments in the Mount Lofty Ranges may be derived in part from Delamerian hydrothermal processes. Stratabound copper sulphides in Kanmantoo Group metasediments are products of thermally driven, migrating solutions reacting with pyritic sediments, apparently prior to metamorphism (Both, 1990). Anderson (1993) reported the discovery of a base-metal deposit in the Kanmantoo Group at Strathalbyn, for which an exhalative origin is indicated by the presence of garnet-andalusite-chlorite rocks and laminated, garnetiferous, quartz-gahnite chert.

The potential for magmatic concentrations of chromium and platinum group elements in the Black Hill Norite remains to be investigated. Drilling in the pluton at Cambrai has demonstrated compositional layering and elevated platinum and palladium concentrations at depths of less than 300 m; more effective concentration of platinoid elements is likely at greater depths.

Beneath the Murray Basin, extensive volcanics of probable Cambrian age and mafic plutonic rocks intersected in drillholes near Coonalpyn, Karoonda (MBT 12 and 14) and northwest of Renmark (Nanyah 1) are prospective by analogy with the mineralised Mount Read Volcanics of Tasmania (Clough and Rankin, 1991).





Small-scale steep reverse fault within the ?Backstairs Passage Formation (Kanmantoo Group), 5 km southeast of Cape Jervis. Similar reverse faults have been imaged on the Backstairs Passage seismic line (see Vol.1, p.28), which shows that they flatten into gently dipping thrusts at depth. Photo 042564.

1. Sub-horizontal thrust zone at 'The Rocks' on River Wakefield, 10 km E of Balaklava. The mylonitic foliation, developed in dolomitic siltstone of uncertain stratigraphic position within the Burra Group, is ascribed to DD_1 and has been openly folded and crenulated on gently plunging north-south axes (DF_2). Photo 042484.
2. Stretched pebbles in conglomerate near the top of the Tapanappa Formation (Kanmantoo Group), Penneshaw, Kangaroo Island. The down-dip elongation of the pebbles is parallel to the northwesterly transport direction inferred for DD_1 in the southern part of the Delamerian fold belt. Photo 042491.
3. Outcrop of Rathjen Gneiss near Palmer, eastern Mount Lofty Ranges. The gneiss was probably intruded as a sill-like granitic pluton either in the very early stages of DD_1 or even prior to deformation, in the Middle to Late Cambrian (U-Pb zircon age 516 ± 3 Ma). Photo 042494.
4. Evidence of at least two phases of deformation in the Tindelpina Shale Member near 'Four Brothers', Olary region. The main cleavage is tentatively assigned to DS_{2a} whereas the gently plunging crenulations are DF_{2b} . Photo 042559.
5. Close-up of the 'Bendigo' Granite, a poorly exposed granodiorite of I-type geochemistry in the Nackara Arc. Photo 042560.
6. Migmatite showing several generations of granite veining, near Rocky Point on the Murray River, 15 km northeast of Mannum. Photo 042561.
7. Z-style fold in steeply dipping mylonitic foliation (with gently north-plunging lineation), indicating dextral strike-slip displacement, at least in the last phase of movement, on a mylonite zone exposed in the Marne River, 2 km east of Black Hill. Photo 042562.
8. Lamprophyre dyke with chilled margins intruded along a joint plane in Backstairs Passage Formation (Kanmantoo Group) in an abandoned quarry 6 km east of Truro. Photo 042563.

Structurally controlled mineralisation

Most of the ore deposits in sediments of the Adelaide Geosyncline have, to some extent, been concentrated by tectonic processes, even where the original sediments were enriched in metals. Numerous copper, gold, lead, zinc and silver deposits have such an origin, the ore occurring in shear zones, breccia zones and quartz or carbonate veins. These include copper at Kapunda, Princess Royal, Paratoo, Carrieton, Blinman, Mount Coffin, Mountain of Light, Lorna Doone and West Mount, and silver-lead at Glen Osmond, Talisker, Aclare, Mount Rhine, Baratta and Avondale. Gold mineralisation is most widespread in zones of tectonic fracturing in the Nackara Arc (Morris and Horn, 1990), for example Mongolata, Ulooloo, Mount Grainger, Waukaringa, Nillinghoo, Kings Bluff, Mannahill and Wadnamanga. Despite the tectonic ore-concentrating processes, there is an underlying stratigraphic association of these metallic deposits (i.e. copper in the Callanna Group, Tapley Hill Formation and Bunyeroo Formation, lead-zinc in Cambrian carbonates, and gold in glaciogenic sediments of the Umberatana Group).

Conclusions

W.V. Preiss

The Delamerian Orogeny was a major tectonic episode affecting the eastern portion of South Australia and an as yet undefined portion of the Palaeozoic mobile belt in the eastern States of Australia. Although the western boundary of the Delamerian fold belt with the Gawler Craton is generally defined by the Torrens Hinge Zone, the eastern limit of Delamerian deformation has not been established. The easternmost granites of ~500 Ma occur in the Glenelg River region of western Victoria, but deformation may have extended further east.

The causes and mechanisms of the Delamerian Orogeny are unclear and it is uncertain whether collision of continental blocks occurred after an episode of plate convergence, or whether the deformation was entirely ensialic. The following tentative conclusions are drawn from a synthesis of available structural and tectonic analyses:

- The complex system of arcuate and intersecting folds results from a succession of compressive and transpressive deformational phases. Pre-existing tectonic controls include the primary configuration of basin sediments and basement blocks, and structures within the basement, such as extensional faults.
- Sedimentation in the Adelaide Geosyncline, accompanied by repeated epeirogenic and extensional movements, ceased at some time in the Middle to Late Cambrian, but the age of the youngest sediments is uncertain. The Late Cambrian Mootwingee Movement could have affected the Barrier Ranges of New South Wales and Olary region of South Australia while the Lake Frome Group was still being deposited in the Arrowie Basin.
- Deformation commenced in the southern part of the Adelaide Geosyncline; a strongly compressive first phase was accompanied by generally northwest-directed thrusting, reclined to recumbent folding, and low-angle cleavage development in the Fleurieu Arcs. The southeastern promontory of the Gawler Craton was a rigid buttress against which the sediments and their underlying basement were imbricated during DD_1 .
- I-type granitoid and lesser mafic magmas were generated from an early Adelaidean mafic, mantle-derived underplate during, after, and perhaps prior to DD_1 , and intruded into relatively high levels in the deforming sedimentary pile. In part the intrusions follow the sigmoidal trend of the Nackara and Fleurieu Arcs, but a north-northwest-trending set was associated with a zone of regional high heat flow, controlled by an important linear crustal weakness, the G2 corridor. The resultant zone of amphibolite-facies metamorphism extends from the eastern Mount Lofty Ranges to the Glenelg River region of western Victoria and cuts across the tectonic trends

of the fold arcs. The A-type plutons of the Peake and Denison Inliers may also have intruded at this time, but are imprecisely dated.

- Basement involvement and penetrative deformation of cover and basement during the Delamerian Orogeny was limited to the regions southwest of the G2 corridor, northeast of Mannahill, and north of Arkaroola. All these regions were on the Willouran rift shoulders and little or no Callanna Group had been deposited on them.
- Compression during DD_1 affected areas as far north as the thrust fault at the western margin of the southern Flinders Ranges. No penetrative deformation occurred at this time in the region between the G2 corridor and Mannahill, where Callanna Group with a variable evaporitic component is inferred from the existence of diapir-cored rather than basement-cored anticlines. Here, DD_1 compression may have been taken up by sliding on salt-lubricated décollement surfaces.
- In response to continued northwest-directed compression against the Gawler Craton, the G2 corridor became reactivated as a sinistral basement shear zone. A meridional zone of transpression associated with this shear produced the more upright structures in the cover sequence of the Mid-North arm during DD_{2a} . Intrusion of syntectonic I-type granitoids continued and metamorphism peaked during this phase in the Nackara and Fleurieu Arcs and near Mount Painter.
- As northwest-directed stresses approached the more stable basement of the Central Flinders Zone, the cover sequence in the Yunta arm, previously little affected by deformation, began to buckle along northeast-trending fold axes (DF_{2b}). Such folding took advantage of the dominant structural grain of the basement and caused fold interference in the transition zone between the Mid-North and Yunta arms, although some DF_{2a} axes were simply rotated into near parallelism with DF_{2b} .
- In the Central Flinders Zone, DF_{2a} folds interfered with DF_{2b} folds to produce the prominent dome-and-basin style of deformation.
- In the North Flinders Zone, north-south compression during DD_{2b} utilised the pre-existing east-west grain of the basement to impose east-west-trending DF_{2b} folds on the cover. These are dominant in the Arkaroola arm but occur only as minor cross-folds in the Willouran arm where thick Callanna Group provided opportunities for uncoupling from the basement.
- During DD_3 , only the outer Fleurieu Arc suffered penetrative deformation, probably because of continuing elevated temperatures in this region. Deformation may relate to a zone of transpression associated with limited east-west compression. Brittle fractures and kinks of variable orientation affected parts of the Nackara Arc.
- After the main orogenic deformation, brittle fracturing and minor kinking affected the outer Fleurieu Arc, while isostatic adjustments took place along mainly meridional faults, displacing metamorphic isograds. Strike-slip faulting in parts of the Nackara Arc and Flinders Ranges may have continued at this time.
- Perhaps at this time or a little later, A-type granitoids and mafic plutons were intruded in the region of the Murray Basin. Some of the lamprophyric intrusions of the Nackara Arc have similar radiometric ages (others, of kimberlitic composition, are Jurassic). Post-tectonic igneous and hydrothermal mineralising events such as at Burra and Mount Painter are possibly younger, post-dating major uplift and exhumation of the Delamerian mountains.

POST-DELAMERIAN COMPRESSIVE DEFORMATION

D.I. Gravestock and R.B. Flint

Breaks in the Cambrian to Ordovician stratigraphic record from the eastern Officer Basin and western Warburton Basin are related in part to Delamerian orogenesis on the northern and eastern margins of the Gawler Craton. The Middle to Late Cambrian Mootwingee Movement and post-Early Ordovician Dullingari Movement (pre- and post-Delamerian) are also responsible for unconformities in the eastern Warburton Basin. Strata in these northern basins, though apparently free from severe effects of this Delamerian deformation, did undergo later compression and, locally, granitic intrusion prior to Gondwanan glaciogene deposition.

In the eastern Officer and southeastern Amadeus Basins this younger tectonic phase is an expression of the Alice Springs Orogeny and related movements (Wells *et al.*, 1970; Bradshaw and Evans, 1988). Ordovician K-Ar ages in two drillcores from the western Warburton and southeastern Amadeus Basins (Poona-runna 1, Mokari 1, Fig. 7.16) were ascribed by Williams (1975) to Delamerian overprint on Proterozoic or early Palaeozoic rocks. However, their shale-siltstone lithologies are consistent with those of the Ordovician Dullingari Group. The absence of a Devonian Carboniferous radiometric overprint from Mokari 1 and from Indulkana Shale samples from the eastern Officer Basin shows that thermal events associated with the Alice Springs Orogeny are not always evident.

Granitic intrusives of the Big Lake Suite in the eastern Warburton Basin are Carboniferous (Gatehouse *et al.*, 1995). However, now that the Innamincka Formation has been demonstrated to be Late Cambrian to Ordovician and not Devonian (Zang, 1992), the age of thrust-faulting in the region is not tightly constrained.

Compressive deformation is discussed below for each sedimentary basin. There is no evidence that the Darling Basin in South Australia was affected by post-Delamerian deformation.

Southeastern Amadeus Basin

Interpretations differ as to the intensity and duration of the Alice Springs Orogeny in the Amadeus Basin, as reviewed by Bradshaw and Evans (1988). These authors recognised two tectonic regimes, each influenced by the operation of a separate major stress vector. The first, following a pre-orogenic stage of crustal extension, involved Late Ordovician to Carboniferous northeast-southwest compression, and the second, Late Carboniferous to Early (?) Permian northwest-southeast compression. Bradshaw and Evans (1988) incorporated the 'Movements' of earlier workers as subdivisions of the Alice Springs Orogeny but lack of data led them to suggest that their 'southern province' (beneath the Pedirka Basin) may have been unaffected by Phanerozoic orogenesis.

Stratigraphic relationships beneath Permo-Carboniferous sediments in the Pedirka Basin and Triassic in the Simpson Basin are not yet clear enough to separate movements of the Alice Springs Orogeny from those of the Delamerian. Both may have contributed to deformation and to breaks in the succession. The distribution and structural relationships between Devonian and older rocks, however, point at least to Alice Springs deformation since the Mereenie Sandstone and Finke Group have themselves been folded locally.

As far as can be determined, the Mereenie Sandstone and Finke Group occupy parallel zones of gently folded strata. The axes of these zones are interpreted to be northwest-southeast, parallel to magnetic, gravity and major Proterozoic structural trends (Youngs, 1976). The region was probably mildly deformed during a Late Devonian-Early Carboniferous phase of northwest-southeast compression.

Permo-Carboniferous, Mesozoic and Tertiary structural reactivation has superimposed north-northeast trends, almost

perpendicular to the Devonian zones. Of particular interest is the southwest-northeast trending Eringa Trough which was initially not considered part of the Pedirka Basin (e.g. Youngs, 1976). It is now recognised as the major depocentre of the western Pedirka Basin (see Ch. 8). The Eringa Trough is parallel to the Allambi thrust zone and Camel Flat Platform in the Northern Territory, which Bradshaw and Evans (1988) suggested to have developed in a final northwest-southeast compressional movement of the Alice Springs Orogeny (Fig. 7.16).

Possible Chandler Formation equivalents were drilled in McDills 1, raising the question that salt withdrawal may have contributed to genesis of the Eringa Trough as it did in the Camel Flat Platform.

Musgrave Block

Part of the southern margin of the Musgrave Block (Vol. 1, Fig. 5.30) is a north-dipping fault, along which basement was thrust over Neoproterozoic and early to mid-Palaeozoic sediments of the Officer Basin (Milton and Parker, 1973). Like other thrust faults in the basin that have been shown on seismic sections to affect sediments as young as Devonian, it is related to the Alice Springs Orogeny. However, isotopic data from some of the numerous east-west trending faults and shear zones within the Musgrave Block suggest earlier phases of movement. Many of these structures formed during the Mesoproterozoic (see Vol. 1, Ch. 5) but some have been reactivated and others initiated during the late Neoproterozoic and early Palaeozoic.

The Woodroffe Thrust zone, up to 1 km wide, dips generally southwards at $\sim 30^\circ$ and contains mylonitised gneiss, mylonite, ultramylonite and veins of pseudotachylite (Major, 1970; Collerson *et al.*, 1972; Bell, 1973, 1978b, 1979). The age of this structure is uncertain, and radiometric estimates are likely to reflect only the youngest phase of movement. Forman (1966, 1972) proposed an age of 600-570 Ma based on Rb-Sr analyses of biotite and feldspar, whereas Camacho (1990) suggested 730 Ma from K-Ar analyses of biotite within a sheared granite in the Northern Territory. Maboko *et al.* (1991) determined a thrusting age of 550-530 Ma from ^{40}Ar - ^{39}Ar analyses of muscovite from a genetically related shear. Collectively, the data imply the last major tectonism at ~ 550 Ma, that is, the Petermann Ranges Orogeny.

The Woodroffe Thrust represents a major boundary separating two distinct basement terrains. Teleseismic and gravity data suggest that at depth the boundary occurs south of the Woodroffe Thrust and is probably associated with the Mann Fault (Lambeck and Burgess, 1992). It is probable that during tectonism at ~ 550 Ma, the Mann Fault was reactivated as a ramp up which a 15 km thick slab of granulitic basement rocks of the Fregon Subdomain was uplifted from mid-crustal levels and thrust northwards for possibly 24 km along the Woodroffe Thrust over amphibolite-facies gneisses of the Mulga Park Subdomain (Maboko *et al.*, 1991; Vol. 1, Fig. 5.32).

Eastern Officer Basin

Overthrusts and strike-slip faults have locally deformed the sedimentary fill of the northeastern Officer Basin. In the Munyarai Trough, such faulting has affected Middle to Late Devonian sediments. Further south, however, erosion or non-deposition of sediments of this age precludes accurate determination of the timing of major fault activity.

In the Indulkana Range, the Ordovician 'Munda sequence' is folded into a gentle, east-west trending syncline while, in the Mount Johns Range, the whole of the exposed Officer Basin succession is gently folded adjacent to a major fault complex known as the Marla Overthrust Zone (Mackie and Gravestock, 1993).

Seismic, gravity and magnetic data show that part of the southern margin of the Musgrave Block is a north-dipping overthrust (Milton and Parker, 1973). Thrust faults also approach the surface towards Cartu Hill on seismic sections (Womer *et al.*, 1987) and a large surface fault trace in the EVERARD map area has been confirmed

seismically as a major thrust. Seismic sections in the Marla area show a series of thrust faults with a cumulative horizontal displacement of 18 km or more towards the southeast. A significant wrench component is recognised from seismic data in the Middle Bore Ridge (Wintinna Fault zone, Dunster, 1987).

Although proof is lacking, there may be a common origin for the Nawa Ridge, Eringa Trough and the northeasterly thrust fault trends (Fig. 7.15). It is conjectured here that these structures may have all developed or been reactivated at a late stage in the Alice Springs Orogeny.

Eastern Warburton Basin

Petroleum exploration drillholes and seismic surveys show that Cambrian rocks in certain regions beneath the Cooper Basin have been overthrust, mainly to the northwest. The thrust zones and their Permian and Mesozoic cover form spectacular arcuate domal trends with right and left-lateral strike-slip offsets (Kuang, 1985). Best known is the Gidgealpa-Merrimelia-Innaminka Ridge (Fig. 7.12, and see Ch. 8) but the Birdsville Track Ridge is also an arcuate zone of faulted domes with major strike-slip offsets and is probably of the same origin. Some of the southern domes, however, are devoid of Cambrian rocks on their crests (e.g. Cooryanna Dome).

Roberts *et al.* (1990) illustrated thrust-associated stratal repetition in the Cambrian succession on seismic sections and wireline logs, while Daily (in Harrison and Higginbotham, 1963) demonstrated a repeated section in Gidgealpa 1 from palaeontological evidence. Roberts *et al.*, (1990) ascribed overthrusting to possible micro-continent collision accompanied by westward jump of a postulated subduction zone during the Ordovician to explain greater severity of deformation in the Birdsville Track Ridge area than in the Coongie-Gidgealpa vicinity. They attributed this to a younger phase of the Delamerian Orogeny than is evident elsewhere. Gatehouse (1986) considered the Benambran Orogeny to have affected these rocks.

Although it is difficult to determine the time of onset of faulting, it post-dates the Ordovician since Cambrian and Ordovician rocks are structurally concordant in drillcores. It is assumed here to have occurred as part of the same series of events that caused granitic intrusion beneath the Nappamerri Trough (Fig. 7.12). A Carboniferous rather than Ordovician age of deformation is preferred here, related to the intracontinental Alice Springs Orogeny rather than the 'subduction zone jump' model of Roberts *et al.* (1990).

Granites of the Big Lake Suite have yielded whole-rock Rb-Sr ages of 362-333 Ma (Webb, 1974) and single zircon U-Pb ages of 342 ± 28 and 310 ± 17 Ma (Fanning, 1986; Gatehouse *et al.*, 1995), indicating a Carboniferous age of crystallisation. Rb-Sr overprint at 330 Ma in the Cambrian Mooracoochie Volcanics is consistent with a Carboniferous thermal event (Wopfner, 1972).

It is concluded that compressive deformation in the eastern Warburton Basin is not due to the Delamerian Orogeny but to some younger event contemporaneous with the Alice Springs Orogeny. However, more work is required to clarify the link that the Warburton Basin provides between mid-Palaeozoic orogenesis in central Australia and that in the Tasman mobile belt.

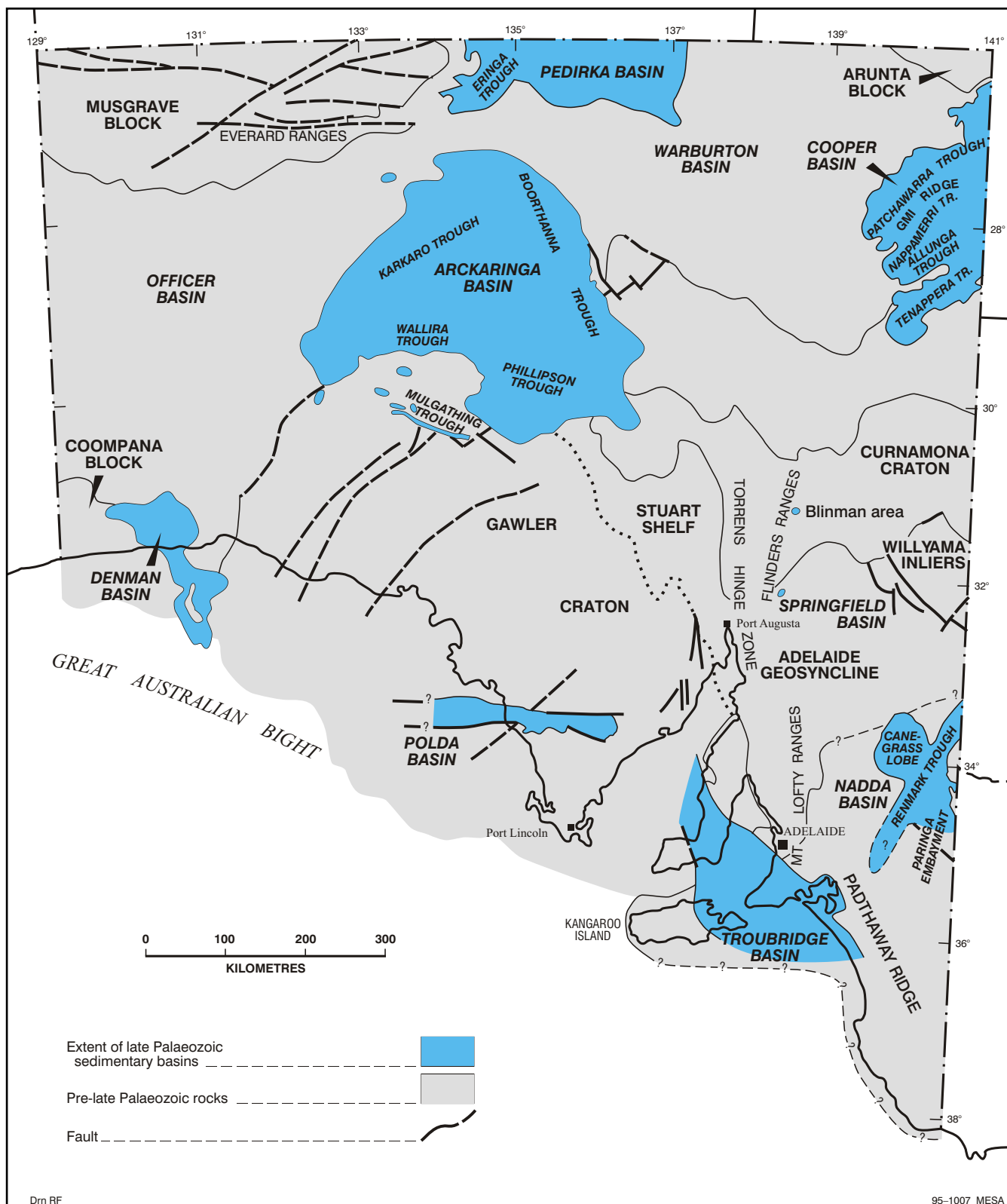


Fig. 8.1 Distribution of late Palaeozoic basins in South Australia.

Chapter 8

LATE PALAEOZOIC

Written and compiled by N.F. Alley

with contributions from R.P. Bourman¹, D.I. Gravestock, J.E. Hibburt, A.J. Hill and L.R. Rankin

INTRODUCTION

N.F. Alley

Apart from evidence of a few local thermal events in the Carboniferous, the record for the late Palaeozoic in South Australia is largely one of glacial erosion and deposition. During this interval, widespread glaciation affected Gondwana and, as a consequence, glaciogene sediments are widespread throughout the Australian continent (Figs 8.1, 8.2). Evidence for this was first recorded in the Inman Valley on Fleurieu Peninsula by Selwyn in 1859 (Selwyn, 1860). This site is marked by a glacially scoured and striated bedrock surface, now a geological monument referred to as 'Selwyn Rock' or, more commonly, 'Glacier Rock'.

Although the glaciogene sediments on Fleurieu Peninsula occupy glacially eroded valleys, major intracratonic Permian sedimentary basins are largely structural features (Wopfner, 1970b, 1980; Veevers and Evans, 1975; Gravenor, 1979; Veevers *et al.*, 1982). Deep South Australian basins are of structural origin but thin mantles of glaciogene sediment (e.g. on the Padthaway Ridge) and thicker valley fills (e.g. Inman Valley) are also common. Basin shape was controlled by northwesterly and northeasterly trending structures that originated in the Late Carboniferous to Early Permian, with intermittent movements continuing until the end of the Permian (Paten, 1969; Wopfner, 1970b, 1980; Kapel, 1972; Thornton, 1978). The structures were probably the result of compression produced by plate convergence along the eastern margin of the craton and rifting on the western side of the continent (Wopfner, 1980). Initiation of some of the faults in the northern part of the State, however, may have occurred during the Alice Springs Orogeny.

The nature, extent and timing of the glaciation that affected Gondwana continue to be debated but evidence indicates that Gondwana began to move across the south rotational pole in the Early Carboniferous, and an intermittent glacial imprint was imposed (Kemp, 1973; Crowell and Frakes, 1975; Kemp *et al.*, 1977; Crowell, 1978; Truswell, 1980; Martin, 1981; Caputo and Crowell, 1985; Veevers and Powell, 1987). Glaciation is considered to have begun in western South America in the Middle Carboniferous and to have ceased in eastern Australia and Antarctica during the Kazanian (early Late Permian; Figs 8.2, 8.3). The age of glaciation



Striated, grooved and polished surface on Kanmantoo Group metasediments at Glacier Rock, Inman Valley. Photo 042441.

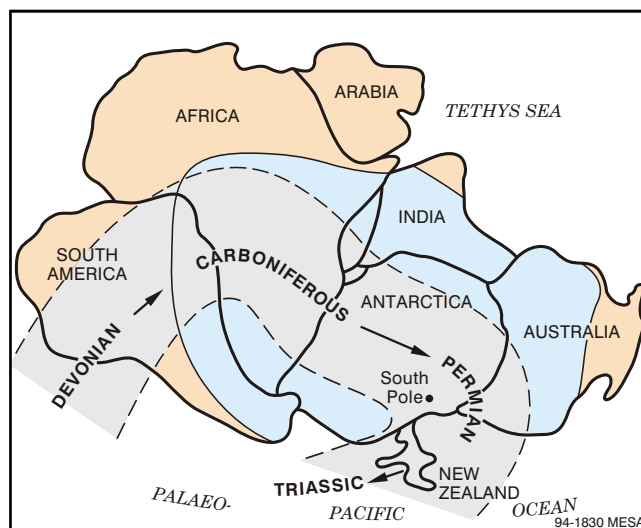


Fig. 8.2 Approximate path of major ice centres as they migrated across Gondwana from Devonian to Permian times. The limit of the area affected by late Palaeozoic glaciation is indicated by the blue area. The position of the South Pole at the Permian-Carboniferous boundary is shown. (After Caputo and Crowell, 1985; Veevers and Powell, 1987).

in South Australia has not been determined with any certainty because palaeontological evidence is restricted to the glaciomarine facies deposited during deglaciation. This information indicates that the glacial phase is at least as old as Asselian, and possibly Late Carboniferous, and that glacial conditions no longer pertained in the Sakmarian.

Precise dating of glacial events in Australia is complicated by lack of agreement on the Permian-Carboniferous boundary as determined by palaeontological evidence (Kemp *et al.*, 1977; Balme, 1980; Cooper, 1981, 1983; Foster, 1983). Palynological and marine faunal evidence from the Canning Basin and correlation with other parts of Australia (including the Troubridge Basin) indicates that Stage 2 palynofloras are Asselian in age (Foster and Waterhouse, 1988); such a determination is followed in this Chapter (Fig. 8.4).

In Australia, as seems to have been the case elsewhere in Gondwana, there are important differences in the nature and timing of glacial events. Montane glaciers probably commenced to grow in Queensland, northern New South Wales and Tasmania in the latest Namurian (Middle Carboniferous), and reached their greatest extent in the Westphalian and Stephanian (Bowen, 1969; Crowell and Frakes, 1970, 1971a,b, 1975; Crowell, 1978; Veevers and Powell, 1987). It is possible that local centres of ice grew in upfaulted highlands to the north and northwest of the Pedirka Basin and southeast of the Arkaringa Basin (Wopfner, 1970b, 1980). As Gondwana drifted across the rotational pole, large continental-scale ice-masses grew in Antarctica and spread northeastwards across Tasmania and through basins in continental Australia. Such an icesheet influenced Australia at the end of the Carboniferous and reached its maximum extent in the Early Permian (Sakmarian), then rapidly decayed in the Artinskian.

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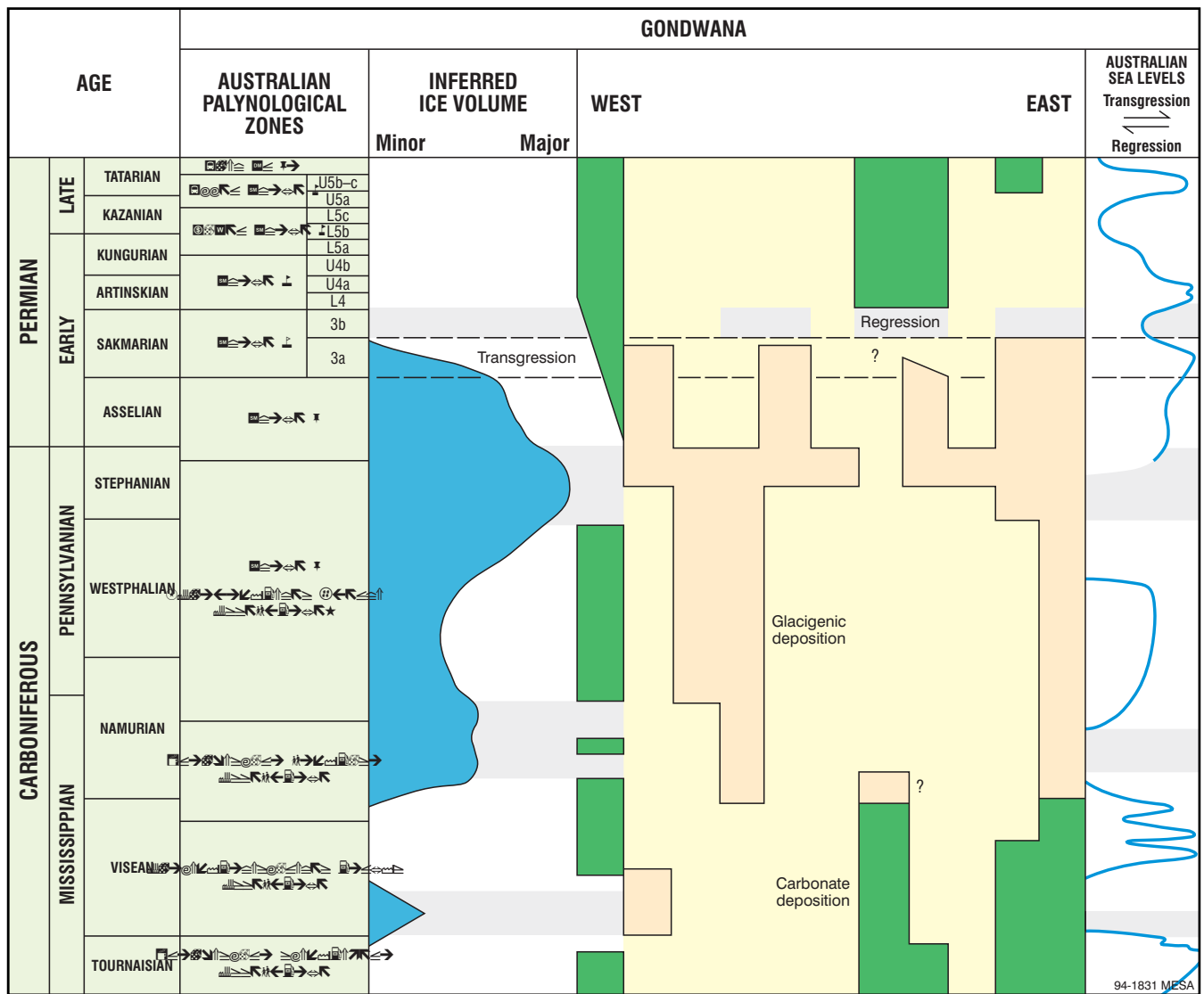


Fig. 8.3 Time distribution chart for the Carboniferous and Permian showing interval and magnitude of glaciation, and the nature of sedimentation in Gondwana. The Sakmarian marine transgression is highlighted on the Australian sea-level curve. (After Veevers and Powell, 1987).

Decay of the ice was accompanied by transgression of the sea into tectonic basins and isostatically depressed lowlands, along the margin of the icesheet (Crowell and Frakes, 1971a). Glaciomarine sedimentation commenced in Asselian to Sakmarian time and continued into the Kazanian (Late Permian) along eastern Australia. In South Australia, Early Permian marine sediments are preserved in the Denman, Arckaringa, Troubridge and Nadda Basins. This distribution led McGowran (1973a) to suggest that the sea entered from the west (around the present southern Australian margin) along an incipient rift, which became more pronounced in the Mesozoic. Glacial influence in South Australia had disappeared by the end of the Sakmarian (Ludbrook, 1967, 1969a; Foster, 1974; Foster and Waterhouse, 1988), but non-marine sedimentation continued in a number of the northern basins in which considerable thicknesses of coal and other carbonaceous sediments were deposited.

The direction of ice movement across South Australia has been constructed from a variety of indicators including the distribution of erratics, striae, chatter marks and other gouge and friction features, *roches moutonnées* and till fabrics. Early workers considered that ice moved from the south towards the north and northeast (Crowell and Frakes, 1971a,b, 1975). More detailed analyses have since demonstrated that the direction of movement in southern South Australia was from the southeast to the northwest (Milnes and Bourman, 1972; Bourman and Milnes, 1976; Bourman *et al.*, 1976; Alley and Bourman, 1984; Bourman, 1987). Local differences in direction of ice movement reflect divergences of a wet-based

icesheet around topographic irregularities (Alley and Bourman, 1984; Bourman, 1987).

Evidence for multiple glaciation in Gondwana, where tillites are separated by major disconformities, has been reported from South America, Antarctica, South Africa and Australia (Tasmania, New South Wales, South Australia, and with at least 51 phases recorded at Bacchus Marsh in Victoria). Such evidence, however, is dubious since the multiple tills at many localities are flowtills or similar proglacial sediments deposited during the deglacial phase of a single glacial event, as discussed in more detail below.

Worldwide climatic changes related to plate tectonic events had a considerable influence on climate in Gondwana. By the beginning of the Middle Carboniferous, the circum-global equatorial seaway had been closed by the connection of Euramerica and Gondwana, with disastrous effect on warm, shallow-water marine faunas. Cooling occurred in Australia; carbonate banks became increasingly rare, shelly faunas were greatly reduced in number and diversity, conodonts disappeared, and strata became dominantly clastic (Ross and Ross, 1985). The increasing cold had a marked effect on plants, and the giant clubmoss flora of the Early Carboniferous disappeared (White, 1986). Vegetation became dominated by cold-adapted seed-bearing ferns, a few lycopods and perhaps conifers.

By the Early Permian, restricted cold-water faunas associated with glaciogene sediments were common over southern Australia (Ludbrook, 1967, 1969a; Dickins, 1977, 1978, 1985). At the end of the main glaciation in the Sakmarian, milder conditions prevailed

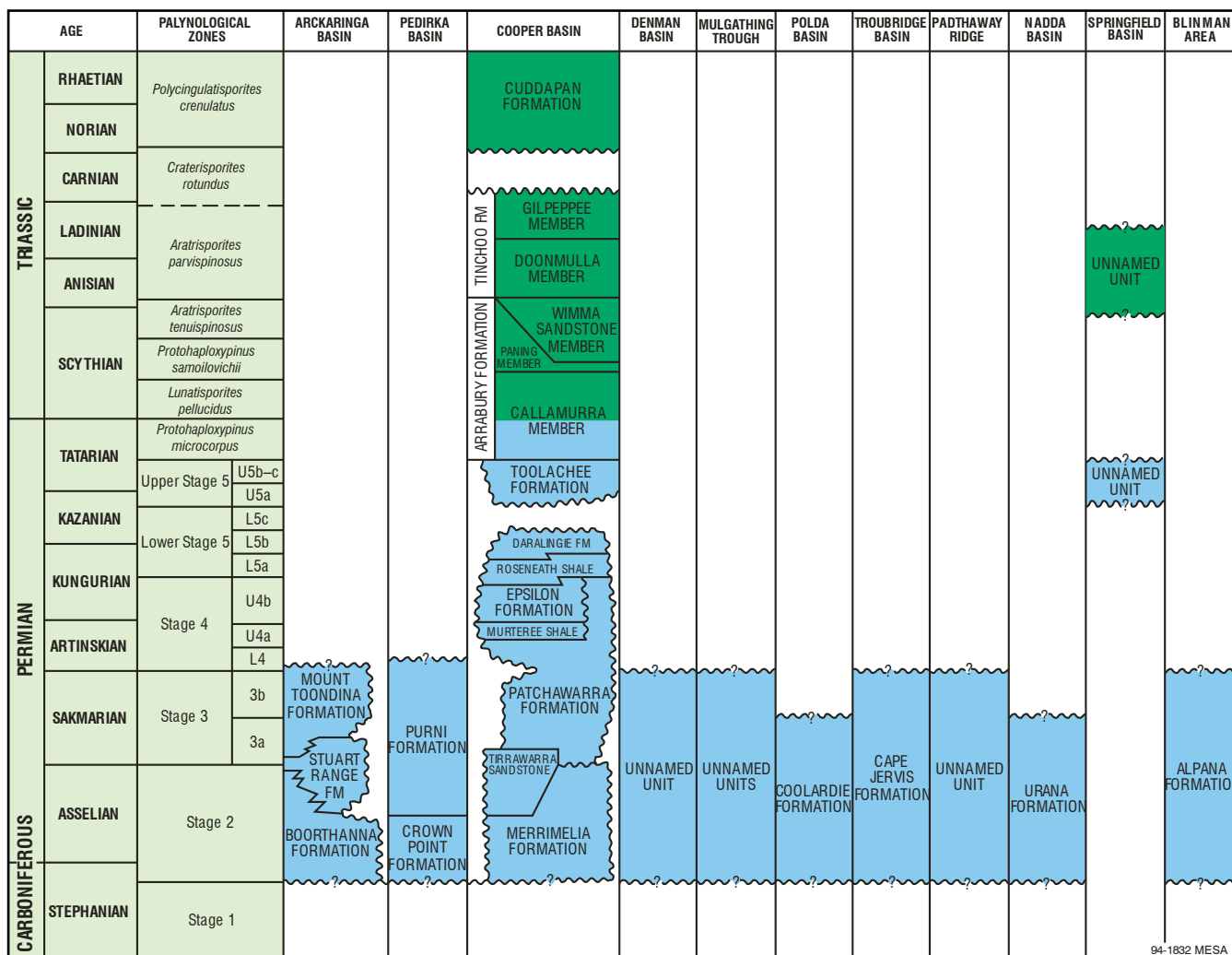


Fig. 8.4 Correlation chart for late Palaeozoic sediments in South Australia.

in the western part of Australia although cold-water faunas persisted in southeastern Australia. Following the eustatic rise in sea level and continuing climatic amelioration, extensive coal swamps developed in the eastern part of the continent with smaller swamps in South Australia and Western Australia. These beds formed extensive coal seams in the Cooper, Arckaringa and Pedirka Basins, and the organic matter provided the source for production of natural gas and petroleum in the Cooper Basin. Unlike the more tropical vegetation which produced the Carboniferous coals of the Northern Hemisphere, Australian coal deposits were the product of cold swampy areas in which dense stands of horsetail (*Equisetum*) dominated. Other swamp vegetation comprised ferns and seed-ferns, mosses and herbaceous lycopods like *Selaginella*. Perhaps wet-adapted trees and shrubs of the glossopterids grew in swamps and adjacent drier areas (White, 1986).

Later in the Artinskian and Kungurian a general warming of ocean waters around Australia occurred and faunas diversified, but cooler conditions lingered on in parts of New South Wales and Tasmania. By Late Permian times, drier terrain provided ideal habitats for ancestors of the conifers, ginkgos and cycads. However, seasonal climate was still severe enough that many plants, such as the ginkgos, were probably deciduous.

Correlation of events and strata between sedimentary basins is often difficult, reflecting the discontinuous nature of glaciogene sediments and paucity of faunas and floras in cold, climatically severe environments. The inferred age and stratigraphic relationships between sediments in late Palaeozoic basins in South Australia is shown on Figure 8.4.

TROUBRIDGE BASIN

N.F. Alley and R.P. Bourman

SETTING

The term 'Troubridge Basin' is here used in the sense of Wopfner (1972) to denote that area of late Palaeozoic sediments stretching from the northern Coorong across Fleurieu Peninsula and Kangaroo Island to Yorke Peninsula and on the continental shelf east of Kangaroo Island (Fig. 8.5). Its boundaries are not tectonically controlled but, rather, reflect remnants of a more extensive blanket of sediment although some of the thickest successions are now preserved in erosional and tectonic troughs.

In 1859, A.R.C. Selwyn, a Victorian Government Geologist under contract to the South Australian Government, first discovered evidence of late Palaeozoic glaciation in the floor of the Inman River (Selwyn, 1860) where well-preserved, glacially scoured Cambrian rocks are overlain by glaciogene sediments; this area continues to provide considerable tourist and scientific interest. Selwyn's discovery promoted further geological research, and evidence of the influence of glaciers was found elsewhere on Fleurieu Peninsula and at Hallett Cove (Tate, 1879a; David and Howchin, 1897). The work of Howchin, particularly, showed the widespread occurrence of glaciogene deposits in the Troubridge Basin, extending knowledge to Kangaroo Island and Yorke Peninsula. Following the observations of these early workers, great scientific debate ensued about the age and style of the glaciation and, indeed, whether the features were truly glacial (for a comprehensive review of these controversies see Bourman, 1987).

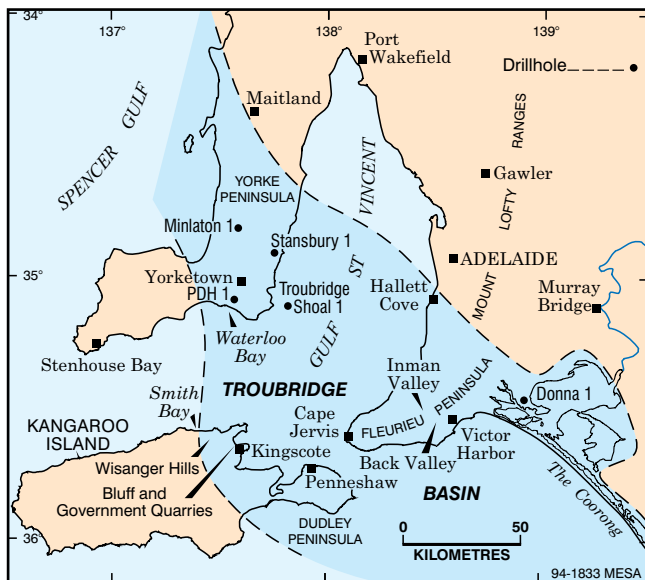


Fig. 8.5 Approximate limits of the Troubridge Basin, and localities of important sites and drillholes.

STRATIGRAPHY

Cape Jervis area

Glacigene sediments in the Troubridge Basin were referred to as Cape Jervis Beds by Ludbrook (1967), but later named Cape Jervis Formation (Bourman and Alley, 1990). At the type section the formation is in part plastered against steeply sloping bedrock surfaces and elsewhere fills small north-south-trending bedrock depressions developed along the strike of Kanmantoo Group metasediments. The glacigene beds comprise five broad units within which there is considerable lithological variability (Fig. 8.6; Alley and Bourman, 1984), and are overlain by the Early Pleistocene Point Ellen Formation (Ludbrook, 1983). Underlying the lodgement till are 0.5 m of frost-shattered metasediments (gelifractions) overlain by several metres of proglacial cross-bedded medium sand and finely bedded clay containing thin lenses of diamicton and striated dropstones. The lodgement till is a 3 m thick clayey diamicton with a strong clast fabric indicating north-northwesterly ice movement (Figs 8.6, 8.7). The clasts are dominated by local lithologies and low frequencies of the Encounter Bay Granites and gneiss, contrasting with the overlying diamictons (Fig. 8.8). Rare clasts of porphyritic volcanics may have been derived from Mount Monster in the South-East, whereas other volcanics may be of Antarctic provenance.

Overlying the lodgement till are 15-20 m of interstratified sand, silt and clay containing gravel lenses and striated dropstones. Graded bedding, intraformational slumping and cyclopels in the finer fraction, and the moderate dip of the beds to the east, indicate that the unit was deposited subaqueously, probably as a fan against a stagnating ice-mass. These sediments are overlain by beds and lenses of poorly stratified pebbly diamicton with a sandy matrix, intercalated with sand and minor silt and clay, together measuring 8-10 m in thickness. This unit is characterised by large granite erratics and a high concentration of coarser clasts. The diamictons possess chaotic larger clast fabrics and are regarded as flowtills shed as supraglacial moraine from stagnating ice into an adjacent aqueous basin.

The uppermost unit at the type section comprises at least 14 m of fine silt and clay containing dropstones and thin grit lenses. A restricted arenaceous marine foraminiferal assemblage recovered from the clay is Sakmarian in age and indicative of low temperatures and/or relatively low salinity (Ludbrook, 1967, 1969a). The faunas are correlative with assemblages from other parts of the Troubridge Basin and from other basins.

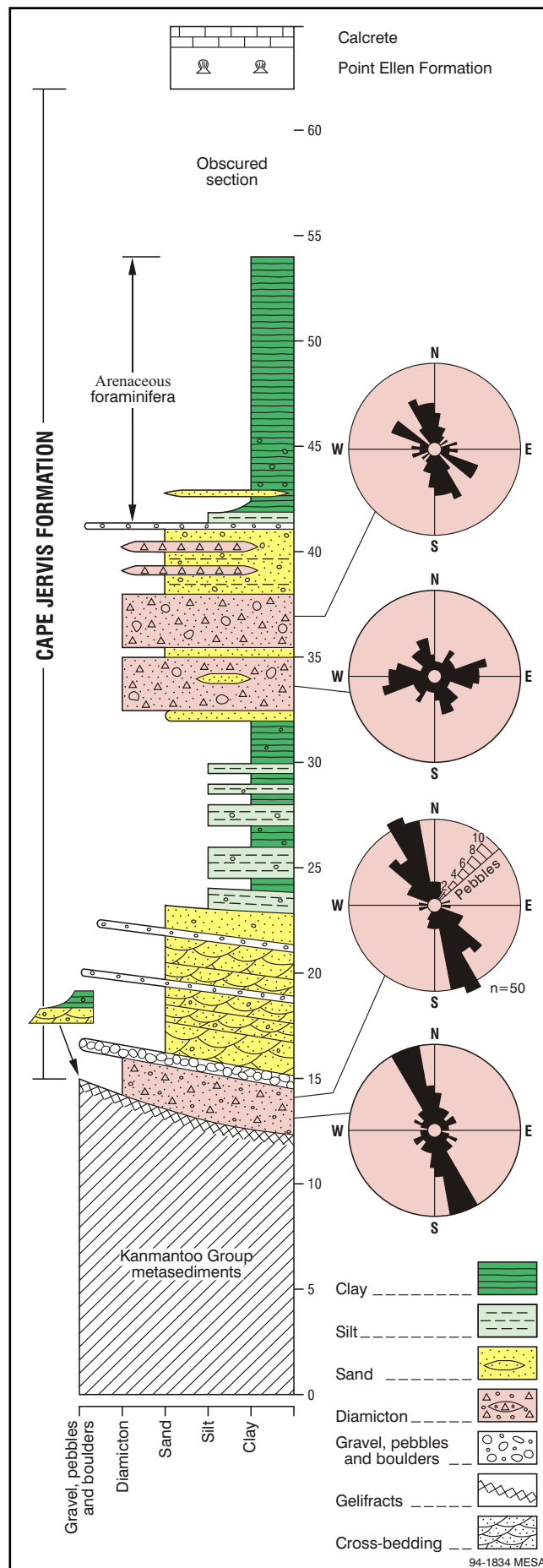


Fig. 8.6 Litholog for the Cape Jervis Formation at the type section. Till fabrics are based on 50 pebbles. (After Alley and Bourman, 1984).

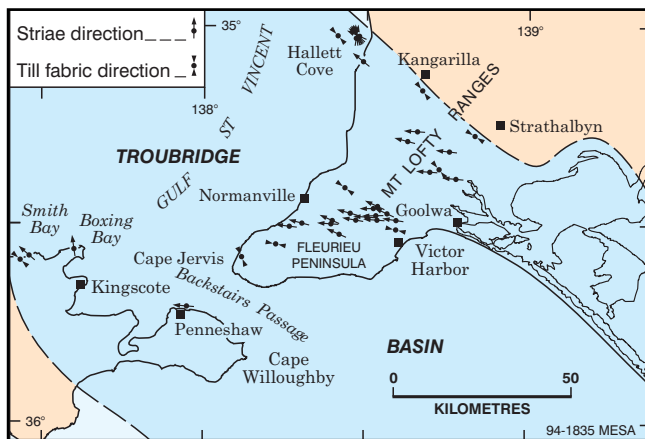


Fig. 8.7 Ice movement directions in the Troubridge Basin, based on till fabrics and striae orientations (after Bourman and Alley, 1990).



Top: Granite erratics in glacial deposits, west bank of the Inman River at Glacier Rock. Photo 042442.

Bottom: Three lenses of flow till, highlighted by calcium carbonate, interbedded with proglacial sand and gravel, Cape Jervis Formation, Cape Jervis type section. Photo 042443.

The model developed to explain the deposition of the Cape Jervis Formation (Alley and Bourman, 1984) at the type section can be generally applied throughout the Troubridge Basin. Probably the most important conclusion is that the sediments were deposited during the passage and decay of a single temperate (wet-based) continental ice-mass.

Hallett Cove area

At Hallett Cove, the Cape Jervis Formation overlies Neoproterozoic bedrock and is overlain by Pliocene Hallett Cove Sandstone or Quaternary clay (Bourman and Alley, 1990). Of great interest at this site are numerous, well-preserved glacial erosional features which are being exhumed from beneath the glaciogene sediments, particularly on the bedrock surface extending from Black Cliff to just beyond the northern limit of Hallett Cove Conservation Park. Diagnostic features include grooves, striae, friction cracks, polished bedrock, small rock crag-and-tail forms, a variety of small stoss-and-lee features (including *roches moutonnées* and 'p-forms' (irregular, polished and smoothed bedrock surfaces eroded by ice and subglacial meltwater).

Cross-cutting striae are common and have been used by Crowell and Frakes (1971a) as evidence for multiple glaciation. However, variations in the local direction of ice movement are the norm under wet-based glaciers and reflect only the plastic nature of the base of the ice as it diverged around minor rock irregularities or as ice centres shifted during the course of glaciation. The plastic nature of the ice is supported by the presence of curved 'p-forms' which are formed by the erosive action of subglacial meltwater charged with rock debris. This occurs only at the base of a glacier where ice is at pressure melting point.

The lowermost unit at Hallett Cove is a blue-grey, sandy lodgement till of unknown thickness cropping out on the shore platform at low tide. The till possesses a strong fabric defined by larger clasts, indicating ice flow from southeast to northwest. Overlying this are a few metres of contorted sandy and pebbly to bouldery flowtills and other ice-contact deposits. Next in the sequence are up to 30 m of bedded glaciolacustrine sand, silt and clay; thickness varies because of the irregular underlying bedrock floor (Fig. 8.9). Thin gravel lenses and flowtills, striated dropstones, graded bedding and intraformational slumping are common (Bourman and Alley, 1990). At a few localities, such as along Waterfall Creek, the beds dip 5-15°E, suggesting that the unit was probably deposited as a subaquatic fan complex in close proximity to ice, in a fashion similar to the glaciolacustrine unit at Cape Jervis.

Exposed in the coastal slope and preserved in bedrock depressions south of the Cove proper are glaciogene sediments dominated by thin lenses of sandy and pebbly diamicton. At one site, 20 diamictons dipping steeply to the east, interbedded with gravel and sand and possessing poorly orientated fabrics, overlie an irregular bedrock floor (Fig. 8.9). These are regarded as subaquatic fan deposits that may include subaerial flowtill derived from nearby ice. Although the outcrop cannot be traced northwards, these fan deposits are probably a proximal ice-contact facies while the glaciolacustrine sequences in Sandison Reserve are a more distal facies.

No marine facies has been detected at Hallett Cove; if such a unit exists it may be in the Lonsdale railway cutting where at least 4 m of laminated fine sand and silt containing minor grit lenses, granules and dropstones are exposed (Bourman and Alley, 1990). These sediments are not fossiliferous and may correlate with glaciolacustrine sediments cropping out along the coast and 30-40 m lower in elevation.

Elsewhere on Fleurieu Peninsula, bedded fine to medium-grained sands predominate; they are often indistinguishable from Tertiary sands that were probably reworked from the glacial units. Some coarser sand facies are useful for building materials and locally for groundwater sources. The sediments are easily eroded and are often gullied, causing significant slope stability problems. Thick sequences are preserved in glacially eroded troughs, 300 m being recorded in Back Valley. At least 167 m of late Palaeozoic sand are present in Donna 1 in the southeastern part of the Troubridge Basin (Ludbrook, 1965a). Thin diamictites and rare carbonaceous clay overlie basement of the northern Padthaway Ridge and in the Coorong area (Beecroft *et al.*, in prep.).

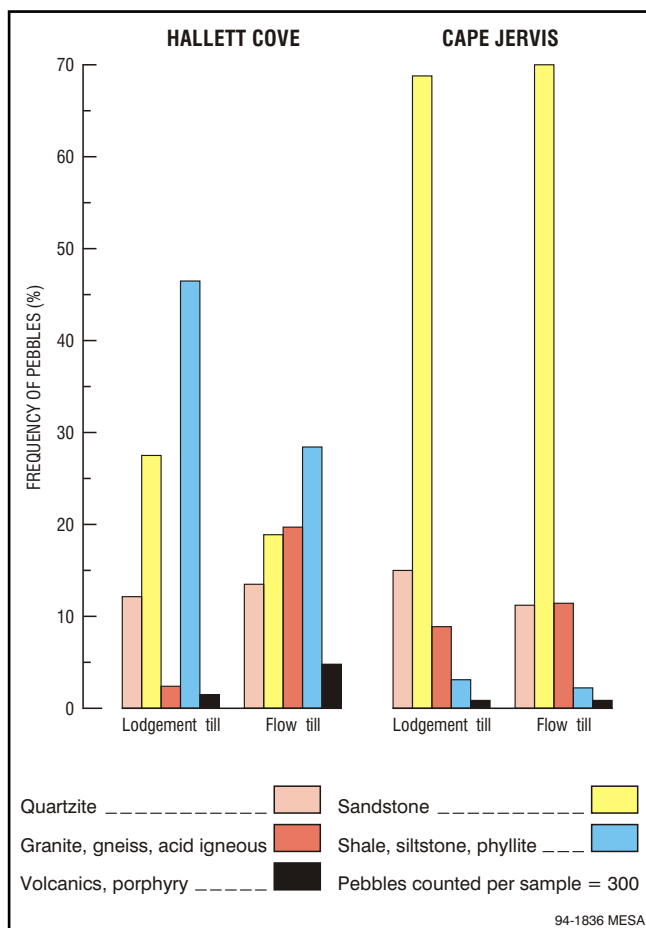


Fig. 8.8 Pebble lithologies of lodgement and flow tills for the Cape Jervis Formation at Cape Jervis and Hallett Cove. Graphs represent averages from multiple samples.



Kangaroo Island

The Cape Jervis Formation crops out on Kangaroo Island where it is overlain by either Tertiary sediments or the Jurassic Wisanger Basalt. Glacigene sediments up to 289 m thick (and perhaps as much as 318 m) were intersected in Kingscote Bore (Howchin, 1929; Ludbrook, 1967). Up to 450 m of the formation capped by ~200 m of Cainozoic strata are believed to fill a deep north-south trending trough eroded in bedrock along the western side of Dudley Peninsula (van der Stelt *et al.*, 1992). A similar trough west of Kingscote contains ~250 m of glacigene sediments overlain by Wisanger Basalt and Cainozoic strata. Seismic information suggests that ~2 000 m of Cape Jervis Formation overlie early Palaeozoic crystalline and metamorphic rocks on the continental shelf east of Kangaroo Island (see Vol. I, Fig. 2.14). The formation is in turn overlain by up to 1 000 m of Cainozoic sand and limestone. Limited drilling in the Troubridge Basin has failed to indicate hydrocarbon occurrences, but the recently defined thick Permian succession east of Kangaroo Island has potential.

Scattered outcrops of Cape Jervis Formation occur along the northern coast of the island between Dudley Peninsula and Smith Bay, and in the Wisanger Hills. At Christmas Cove, in Penneshaw, a few metres of fluvioglacial and glaciolacustrine clay, silt, sand and diamicton overlie an irregular, glacially moulded surface eroded across steeply dipping Kanmantoo Group rocks.

Other significant outcrops of glacigene sediments are in the coastal cliffs at Kingscote and to the west of the town in road cuttings and The Bluff basalt quarry. A considerable thickness of sometimes contorted clay, silt and sand containing rip-up clasts, erratics and dropstones is exposed along the cliff west of Kingscote jetty. Turbidites and cross-bedding are common in the sediments which appear to have been deposited on the distal part of a subaquatic fan. In the now disused Government Quarry, at least 8 m of level to cross-bedded medium to coarse quartz sand and fine gravel containing a lower 1 m thick sandy diamicton are overlain by the Wisanger Basalt. Belperio and Flint (1992) suggested that the sand may not be part of the Cape Jervis Formation but a Middle Jurassic fluvial deposit. However, at The Bluff Quarry, 23 m of interbedded clay, silt and sand containing large erratics, rip-up clasts and thin diamictons grade upwards into 4 m of cross-bedded medium to coarse quartz sand identical to that in the Government Quarry. The sand is part of a proglacial succession, probably the distal sandy part of an outwash plain (sandur).

The only other significant exposure of Cape Jervis Formation is at Smith Bay where 18 m of compact lodgement till overlie glaciated Cambrian sedimentary rocks, in turn overlain by a transgressive shelly boulder beach lag and an unknown thickness of Point Ellen Formation or Burnham Limestone.

Evidence from striae, grooves, *roches moutonnées*, chatter marks and till fabrics indicates that the direction of ice movement was more westerly near Penneshaw and north-northwesterly at Boxing Bay and Smith Bay. Perhaps the most spectacular glaciated rock surface to be found anywhere in Australia occurs at the latter site. The evidence here not only shows very clearly the ice flow direction but also demonstrates that the base of the ice was at pressure melting point, typical of a temperate glacier, agreeing with data from Fleurieu Peninsula and Hallett Cove.

Top: Cross-cutting striae on glacially scoured Neoproterozoic Brachina Formation at Black Cliff, Hallett Cove. View to the northwest in the general direction of ice movement. Photo 038744.

Bottom: Glacially scoured and plastically moulded surface with winding channels (p-forms) developed on Neoproterozoic Seacliff Sandstone, northern end of Sandison Reserve, Hallett Cove. Photo 038745.

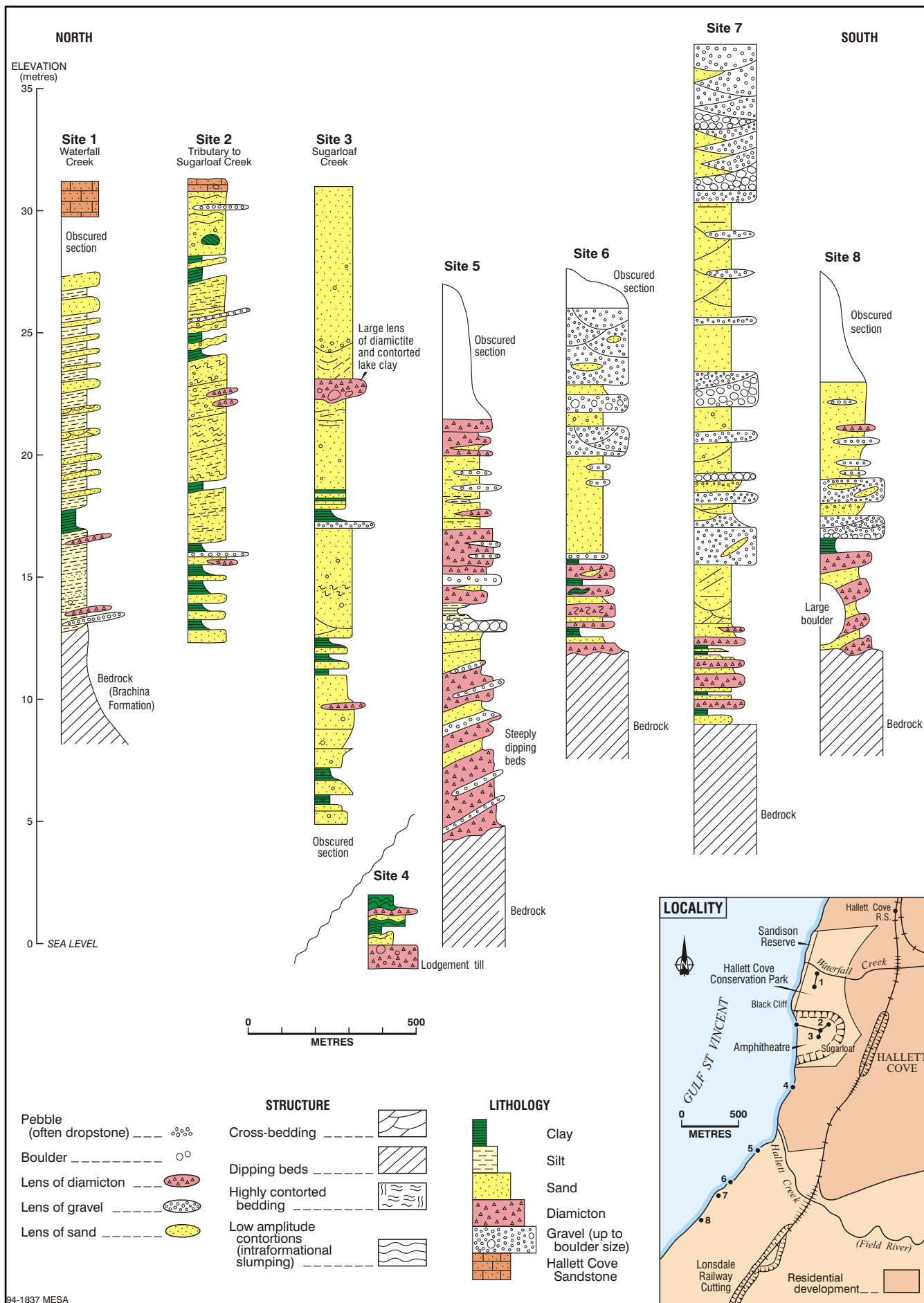


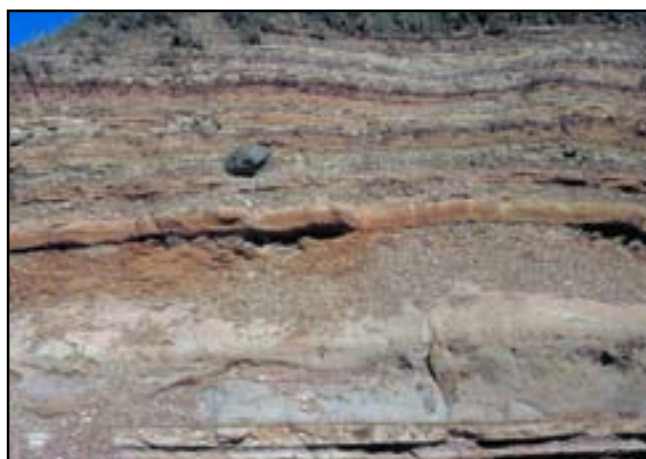
Fig. 8.9 Lithologs for the Cape Jervis Formation at Hallett Cove (after Bourman and Alley, 1990). Sections are located on the map insert.

Yorke Peninsula

The Cape Jervis Formation is widespread on southern Yorke Peninsula and forms the base of coastal cliffs (as at Waterloo Bay) and the floor of many salt lakes. At Waterloo Bay, it includes a carbonaceous, sandy micaceous clay unit containing microfossils and is overlain by Late Eocene sediments. Although ~50 m of Cape Jervis Formation are exposed in outcrop, greater thicknesses have been disclosed by drilling; for example, 236 and 181 m of late Palaeozoic sediments were intersected in Stansbury 1 and Minlaton 1, respectively (Crawford, 1965; Ludbrook, 1965b).

AGE

Glaciomarine sediments representing the final phase of deglaciation in the basin have been dated by foraminiferal and palynological evidence. Arenaceous foraminifera from Cape Jervis, Minlaton 1, Stansbury 1 and Troubridge Shoal 1 are assigned an Early Permian (Sakmarian) age (Ludbrook, 1957a, 1965b, 1967, 1969a). Palynofloras from Waterloo Bay and the nearby PDH 1 are correlated with the late Palaeozoic Stage 2 palynofloras of Evans (1969), believed to be Sakmarian in age (Harris and McGowran, 1971; Foster, 1974). Based on a correlation with other Australian and Gondwanan palynofloras, Foster and Waterhouse (1988) assigned the Waterloo Bay assemblages to the *Granulatisporites confluens* Oppel-Zone of earliest Permian (Asselian) age. Since all of the fossil evidence comes from the youngest part of the Cape Jervis Formation, the age range of the glaciation is unknown and may extend back into the Carboniferous.



Top: Turbidites containing dropstones and erratics in a small west-facing exposure of Cape Jervis Formation near the southern limit of Hallet Cove Conservation Park. Photo 042444.

Bottom: Slumped cliff-face at Waterloo Bay, Yorke Peninsula, exposing glaciomarine clay (lower left) and steeply dipping sand and silt of the Cape Jervis Formation, overlain here by slumped Quaternary clay, but elsewhere by Tertiary limestone and Quaternary Bridgewater Formation. Photo 042445.

NADDA BASIN

N.F. Alley

STRUCTURE AND SETTING

Late Palaeozoic sediments have been encountered in many drillholes beneath strata forming the Mesozoic Berri Basin and the Tertiary Murray Basin. The sediments are preserved and were partly deposited in structurally controlled depressions referred to as Renmark Trough, Canegrass Lobe and Paringa Embayment (Fig. 8.10; Derrington and Anderson, 1970; Thornton, 1974). These sediments are contiguous with the more extensive sequence of glaciogene sediments named Urana Formation in New South Wales and Victoria, and this name is also used for the formation in South Australia (O'Brien, 1986). Thornton (1974) considered that Devonian and Cambrian rocks form the basement but it is likely that bedrock also comprises Adelaidean strata.

The infrabasins in South Australia are here combined into a single entity named the Nadda Basin after Nadda 1 (Paringa Embayment) in which the greatest thickness of Permian sediments has been intersected, although seismic surveys indicate a thicker succession in the Renmark Trough (Fig. 8.10). Drilling on the Padthaway Ridge to the south of the basin disclosed glaciogene sediments underlying Tertiary deposits and filling small depressions in the basement.

The Renmark Trough is a northeasterly trending feature deepening to the north from ~2 700 to 3 500 m (Thornton, 1974). Although the southern part is poorly known, it appears to be a half graben with the northwestern side delineated by the Hamley Fault. The latter extends up into the overlying Early Cretaceous sediments and has a throw of up to 1 500 m. In the northern half, the trough is a graben with its eastern boundary formed by the north-south-trending Chowilla Fault, with a similar throw to the Hamley Fault. Drilling has shown that the Urana Formation extends southwards from the Renmark Trough along a structurally controlled depression towards the Padthaway Ridge.

The Paringa Embayment lies southeast of the Renmark Trough and comprises a series of erosional valleys scoured into the basement, each with a pronounced gradient towards the trough. The Canegrass Lobe is a large, gently westward-shallowing depression northwest of the Renmark Trough, with depth to basement reaching ~1 200 m. Geophysical evidence (R.A. Gerdes, SADME, pers. comm., 1991) indicates that the northern and northwestern margins of the lobe are fault controlled, the latter fault having a significant throw down to the east. The Urana Formation may onlap, or be pierced by, a horst of basement rock in the northern part of the lobe (Fig. 8.10).

Although limited data suggest potential for the occurrence of hydrocarbons, only minor shows have been reported.

STRATIGRAPHY

Variable thicknesses of Urana Formation have been encountered, the thickest being 395 m in Nadda 1 in the Paringa Embayment. Seismic evidence, however, suggests that the late Palaeozoic sediments attain 900 m in the Renmark Trough (Thornton, 1974). The formation comprises shale, siltstone, sandstone, diamictite and occasional conglomeratic and carbonaceous intervals. Dropstones are common in the finer clastics and probably reflect ice-rafting in glaciomarine to glaciolacustrine conditions. Although diamictites are frequently encountered, these are not lodgement tills but rather proglacial sediments as suggested by their close association with sandstone and other bedded units. The Urana Formation in South Australia appears to represent a proglacial to post- or non-glacial sedimentary sequence, lacking evidence of the direct action of glacial ice.

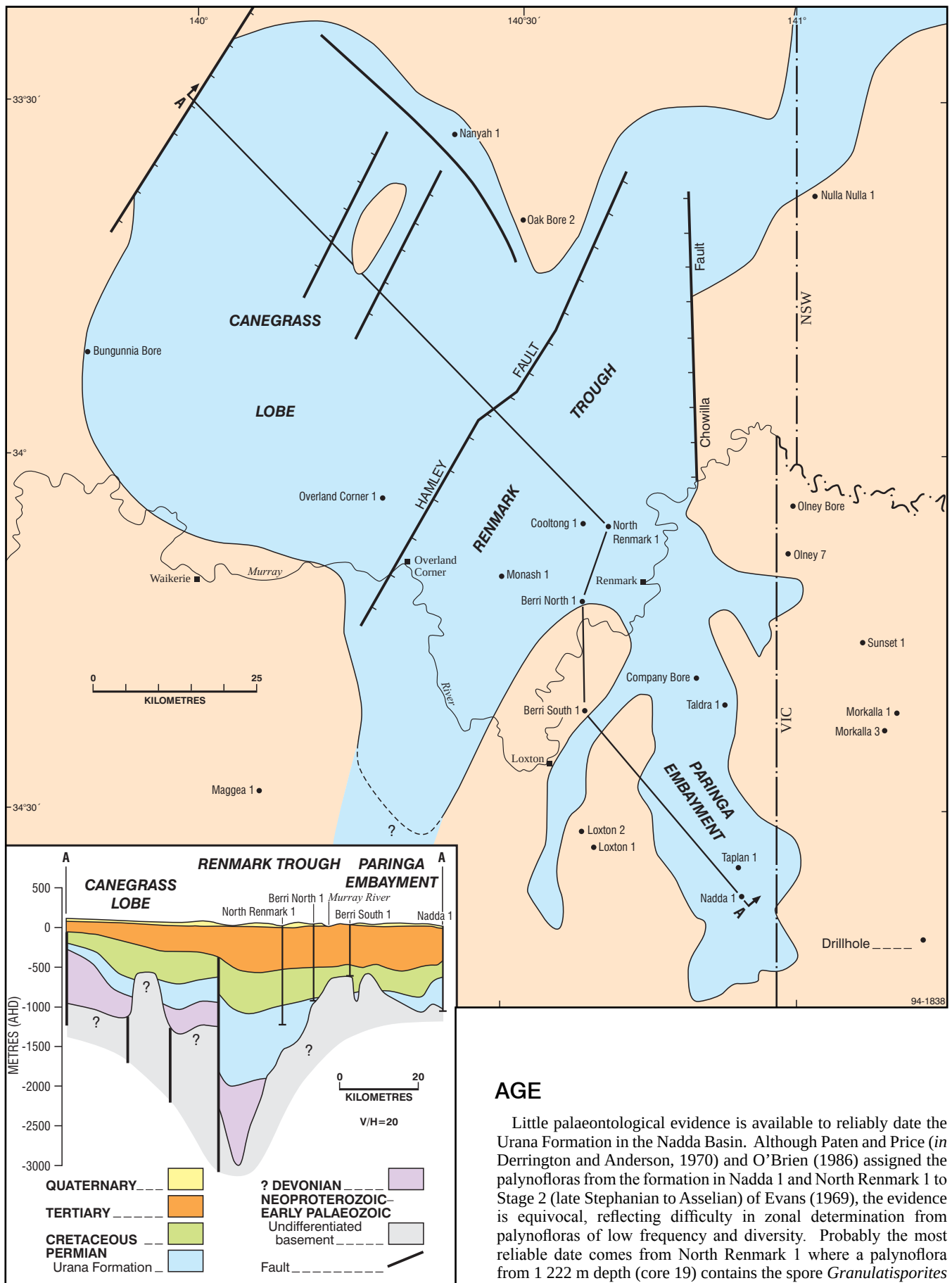


Fig. 8.10 Approximate limits of the Nadda Basin, major structural elements and location of drillholes that indicate distribution and thickness of the Urana Formation (after Thornton, 1974).

AGE

Little palaeontological evidence is available to reliably date the Urana Formation in the Nadda Basin. Although Paten and Price (*in* Derrington and Anderson, 1970) and O'Brien (1986) assigned the palynofloras from the formation in Nadda 1 and North Renmark 1 to Stage 2 (late Stephanian to Asselian) of Evans (1969), the evidence is equivocal, reflecting difficulty in zonal determination from palynofloras of low frequency and diversity. Probably the most reliable date comes from North Renmark 1 where a palynoflora from 1 222 m depth (core 19) contains the spore *Granulatisporites trisinus* (Harris *in* Ludbrook, 1963a). This, along with the absence of *Phaselisporites cicatricosus*, indicates that the palynoflora should be assigned to Stage 3b (Sakmarian) of Evans (1969). The latter age would be more consistent with the age of similar facies in the Troubridge Basin.

POLDA BASIN

L.R. Rankin

SETTING

During the late Palaeozoic, a sequence of diamictite, sandstone, siltstone and mudstone was deposited in the Polda Basin (Fig. 8.11) and probably on the surrounding Gawler Craton. Subsequent erosion has left the sediments restricted to the subsurface of the Polda Basin. These sediments, named the Coolardie Formation (Cooper *et al.*, 1982), have been intersected in several drillholes in the KIMBA map area (Flint and Rankin, 1989), and offshore in Mercury 1, Gemini 1 and Columbia 1 (McClure, 1982a,b; Flint, 1992). The sequence unconformably overlies the Archaean to Mesoproterozoic Gawler Craton and Itileedoo Basin in the shallow sections, and the Neoproterozoic Kilroo Formation in deeper sections of the Polda Basin. It is unconformably overlain by the Late Jurassic Polda Formation (Fig. 8.11).

STRATIGRAPHY

Within SADME Lock 1, the type section of the Coolardie Formation comprises 181 m of brown, green and white diamictite, mudstone, sandstone and siltstone, with abundant matrix-supported pebbles and cobbles of various lithologies (Kwitko, 1982a; Cooper *et al.*, 1982). This is the thickest section intersected; the base was not disclosed. Seismic refraction surveys conducted at Lock 1 indicated 480 m of sediments that may be late Palaeozoic (McInerny, 1977), but could include the underlying Kilroo Formation.

The only drillhole to fully penetrate the Coolardie Formation is CRA 83 KD 1A, in the eastern section of the Polda Basin, where the unit is 22 m thick and unconformably overlies Kilroo Formation. The Coolardie Formation here comprises an upper and a lower unit. The lower, 12 m thick unit is dark grey, matrix-supported massive diamictite with angular to subrounded pebbles up to 60 mm across. Pebble lithologies include pink granite, white and biotite-rich quartzite, biotite schist, granite gneiss, granodiorite, feldspathic

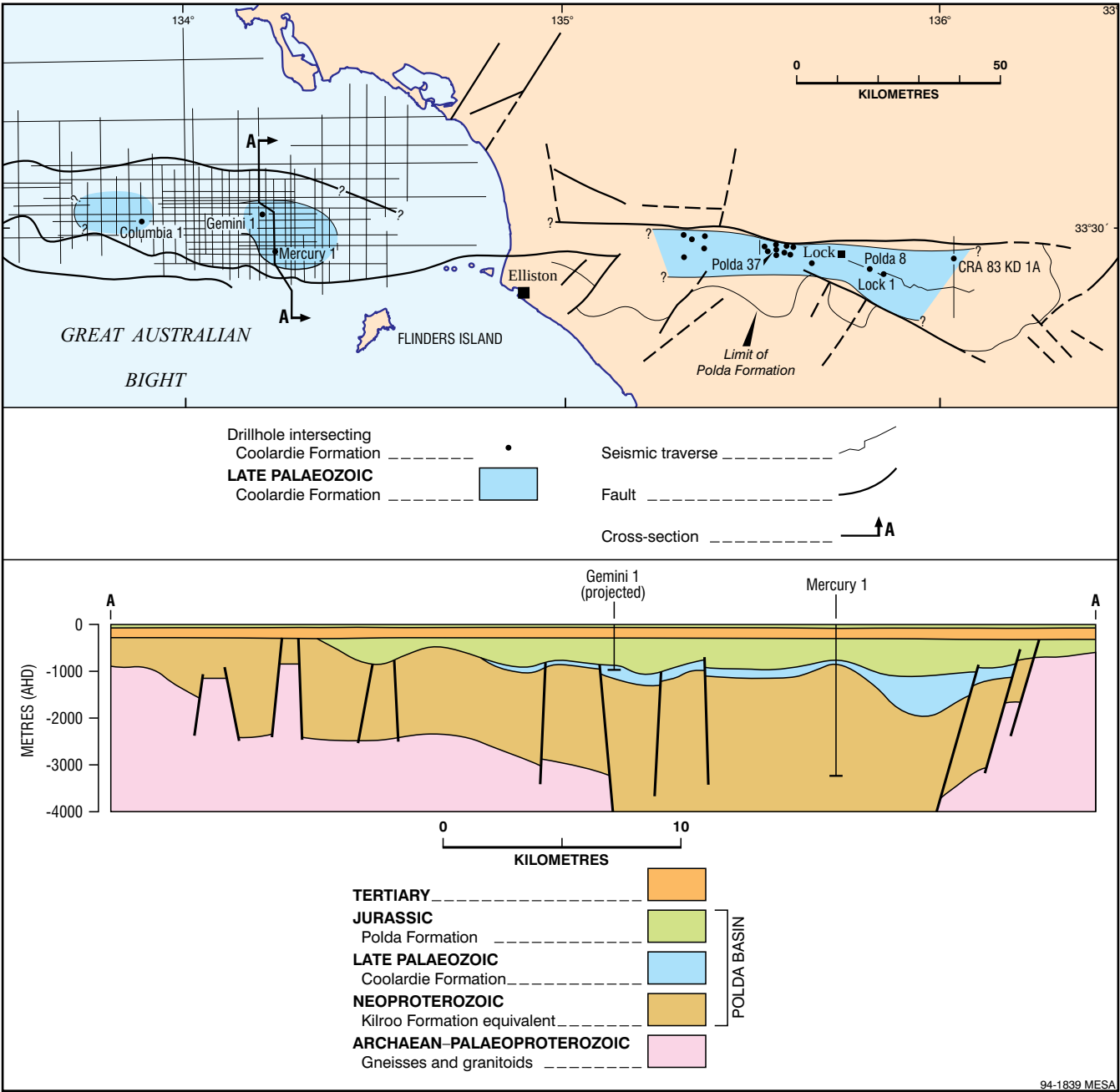


Fig. 8.11 Approximate limits of the Polda Basin, major structural elements, drillholes, seismic lines and cross-section (after Flint, 1989; Flint and Rankin, 1989).



Finely developed foreset beds in silty sandstone with minor pebbles, overlying diamictite containing clasts of igneous rocks, Coolardie Formation at 243.14–243.28 m in Lock 1. Top is to the right. Photo 042446.

granofels, black siliceous slate, grey slate, tourmaline-bearing schist and red-brown siltstone (CRA Exploration Pty Ltd, 1984). The upper 10 m thick unit comprises white, well-sorted, fine-grained, thinly bedded sandstone with minor pebble and diamictite bands; pyrite framboids are common. The upper 9 m of the sequence is intensely kaolinised, evidence of a weathering event which predated deposition of the overlying Late Jurassic Poldia Formation.

Offshore, Gemini 1 was abandoned after intersecting a supposed basalt (below the Poldia Formation) which was tentatively interpreted as a sill and correlated with Jurassic basalts on Kangaroo Island (Carter and Scott, 1975; McInerney, 1978; Daily *et al.*, 1979; Milnes *et al.*, 1982). However, the presence of granite and gneiss fragments in drill cuttings suggests that the basalt may, likewise, be erratic boulders within the Coolardie Formation.

The formation is interpreted as a sequence of glaciogene sediments (Cooper *et al.*, 1982; Parker *et al.*, 1985). A single foraminifer retrieved from contaminated cuttings in Mercury 1, believed to have been derived from the Coolardie Formation, suggests that there was some marine influence in the western portion of the basin (Lindsay and Cooper, 1982; Parker *et al.*, 1985).

AGE

A Carboniferous-Permian age for the unit was first assigned on the basis of palynological evidence from Poldia 8 (Gatehouse, 1980) and Poldia 37 (Harris, 1979a). Further work by Cooper *et al.* (1982) on samples from Lock 1 allowed correlation with the lower Tamarian Stage (equivalent to palynological Stage 2) of the Permo-Carboniferous of Tasmania (Clark and Farmer, 1976; Truswell, 1978), and local correlation with the Cape Jervis Formation of the Troubridge Basin and Boorthanna and basal Stuart Range Formations of the Arckaringa Basin.

ARCKARINGA BASIN

J.E. Hibburt

STRUCTURE AND SETTING

The Arckaringa Basin covers an area of 80 000 km² and is mostly restricted to the subsurface (Fig. 8.12). Outcropping Officer Basin, Gawler Craton and Stuart Shelf rocks mark approximately its western, southern and southeastern limits. However, samples from drillholes west of the Everard Ranges indicate that scattered subcrops of Permian sediment extend almost 100 km beyond the western margin of the basin (N.F. Alley, MESA, pers. comm., 1994). The eastern margin is formed in part by the Peake and Denison Inliers, whereas the northern and northeastern margins are not readily defined due to lack of outcrop and subsurface data. Geophysical surveys, together with information from petroleum exploration, stratigraphic, mineral and coal drillholes, and limited outcrop data, have partly outlined the structure and stratigraphy of the basin (Wopfner, 1970b; Townsend, 1976; Moore, 1982b,c; Hibburt, 1984).

Permian sediments in the basin include (from the oldest) Boorthanna, Stuart Range and Mount Toondina Formations as a subhorizontal succession unconformably overlying Cambrian sediments or Precambrian sediments, metasediments and crystalline rocks (Fig. 8.13). In the Boorthanna Trough, the shallow marine clastic-carbonate Cootanoorina Formation is correlated with the Early Cambrian Ouldburra Formation in the Officer Basin. The Permian formations may crop out or be overlain unconformably by up to 300 m of Late Jurassic to Early Cretaceous Eromanga Basin sediments and generally less than 10 m of Tertiary cover (Fig. 8.13).

Structurally, the Arckaringa Basin is an intracratonic feature comprising a central platform of gently undulating shallow crystalline basement surrounded by depressions. Major depocentres such as the Boorthanna and Phillipson Troughs, developed principally along northwest trends, contain more than 1 000 m of Permian sediments. Sedimentation spread from the troughs across the less disturbed areas of basement highs. Wilmot (1987, based on Veevers, 1984) considered that initiation of the Boorthanna Trough was related to an extension of the Adelaide Geosyncline or a crustal weakness around the northeastern edge of the Gawler Craton. Wopfner (1972) suggested that faulting was synchronous with deposition, an assumption supported by fission track dates on the early Palaeozoic Bungadillina Monzonite (Ambrose *et al.*, 1981). Basin development commenced during the Early Permian or earlier, as evidenced by thinning of Stuart Range Formation over structural crests (Wilmot, 1987).

The Early Permian depositional phase (Boorthanna Formation) in the Arckaringa Basin commenced (as elsewhere) with widespread glaciation. Following deglaciation, marine transgressive conditions prevailed (Stuart Range Formation), with eustatic rise in sea level and flooding of glacio-isostatically depressed lowlands. Palaeontological data indicate that the sea transgressed from the west (Wopfner, 1980), northwest (Jones, 1987), or the southwest (McGowran, 1973a); a southeastern entrance is also inferred from the presence of a marine facies in basal Mount Toondina Formation which is restricted to the Phillipson and southern Boorthanna Troughs. It is possible that the sea entered from several directions. Termination of marine sedimentation was brought about by crustal isostatic recovery or eustatic fall and return to freshwater conditions with development of coal swamps (Mount Toondina Formation). Sedimentation ceased as a consequence of positive epeirogenic movements (Wopfner, 1972).

Diapirism has affected the Arckaringa Basin, and Permian sediments at Mount Toondina are seen to form a piercement structure (Freytag, 1965; Wopfner, 1977). Rejuvenation of late Precambrian-Cambrian diapirism in the late Tertiary is related to tectonics (Jones, 1988).

STRATIGRAPHY

Three major sedimentary units have been defined on the basis of drilling and mapping at Mount Toondina and in the Peake and Denison Inliers (Ludbrook, 1961a, 1967; Freytag, 1963, 1965; Allchurch and Wopfner, 1967, 1973; Townsend and Ludbrook, 1975; Townsend, 1976). However, cored sections in drillholes are sparse, petrophysical log quality is poor and outcrops do not represent the complete succession, thus formation definitions are inadequate.

Boorthanna Formation

Boorthanna Formation comprises a glaciogene sequence with marine facies towards the top. The glaciogene suite is thought to be similar to the fluvio-glacial system of the Merrimelia Formation and Tirrawarra Sandstone of the southern Cooper Basin (see below). Two units distinguished at the type section in Boorthanna 1 include a basal facies dominated by diamictite with shale intercalations and an upper unit of rhythmically bedded coarse and fine clastics. The diamictite facies is a sandy to bouldery claystone, which is often calcareous and sometimes dolomitic, interbedded with shale and occasionally thin sandstone or carbonate. This unit appears to be restricted to the Wallira and Phillipson Troughs and the southern part of the Boorthanna Trough. Wopfner (1970b) suggested that

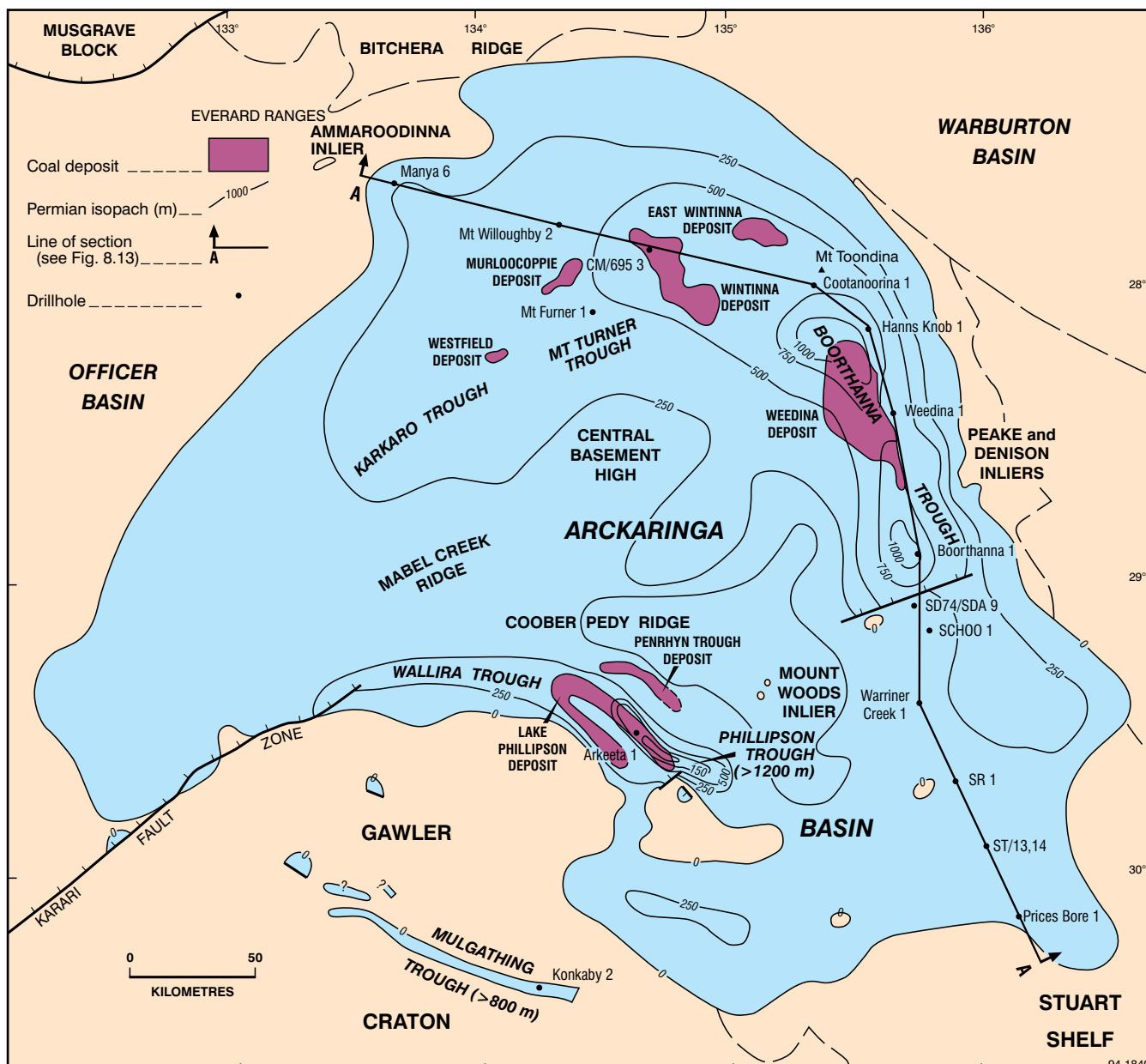


Fig. 8.12 Approximate limits of the Arckaringa Basin, major structural elements, isopachs and drillholes.

the diamictite represents subaqueously redeposited glacial debris transported downslope by mudflows. Scattered outcrops of the Boorthanna Formation, mainly diamictite, occur along the eastern margin of the Arckaringa Basin (Ambrose *et al.*, 1981) and may be part of a moraine and esker complex (Hibburt, 1984). Tillites have not been recognised in the subsurface. The rhythmically bedded facies is generally medium to coarse-grained sandstone but ranges from siltstone to boulder beds. It was probably deposited by turbidity currents from detrital material originating during and after the glacial phase (Wopfner, 1970b). In the southwestern portion of the Arckaringa Basin, Benbow (1993) noted that diamictite and clastics are not confined to one stratigraphic level.

Although other subunits have been recognised within the Boorthanna Formation, these are rarely consistent with those of the type section (e.g. Moore, 1982b; Allender *et al.*, 1987; Jones, 1987; McBain, 1987; Benbow, 1993). The lowermost 29 m of the type section may alternatively be Devonian (see Ch. 7).

On palaeontological and sedimentological evidence, Jones (1987) postulated an Asselian/Sakmarian marine transgression from the northwest with restricted marine conditions in the southeast and apparent younging of sediments in this direction. Palynological evidence shows that the uppermost part of the Boorthanna Formation is absent in localised areas (e.g. Hanns Knob 1) due

either to non-deposition because of emergence, or uplift and erosion prior to deposition of the Stuart Range Formation (Jones, 1987).

Stuart Range Formation

Boorthanna Formation is conformably (occasionally disconformably) overlain by Stuart Range Formation, although seismic evidence indicates that the relationship is unconformable in the Boorthanna Trough (Allender *et al.*, 1987; Wilmot, 1987). The two units intertongue locally in the southwestern Arckaringa Basin (Benbow, 1993). The Stuart Range Formation, which directly overlies pre-Permian rocks locally, is generally an homogenous marine shale with minor siltstone and sandstone, deposited under quiet-water, restricted marine conditions (Wopfner, 1970b; Harris and McGowran, 1973; Barnes *et al.*, 1977; Gilby and Foster, 1988). Brief episodes of non-marine deposition are indicated in the Cootanoorina 1 and Mount Furner 1 successions (Gilby and Foster, 1988) and in the upper part of the formation in Arkeeta 1, suggesting alternating lacustrine and marine conditions (McBain, 1987). Deposition of the Stuart Range Formation probably occurred under maximum transgressive conditions (Jones, 1987). The formation may have oil-shale potential (Meyer, 1981; Moore, 1982d).

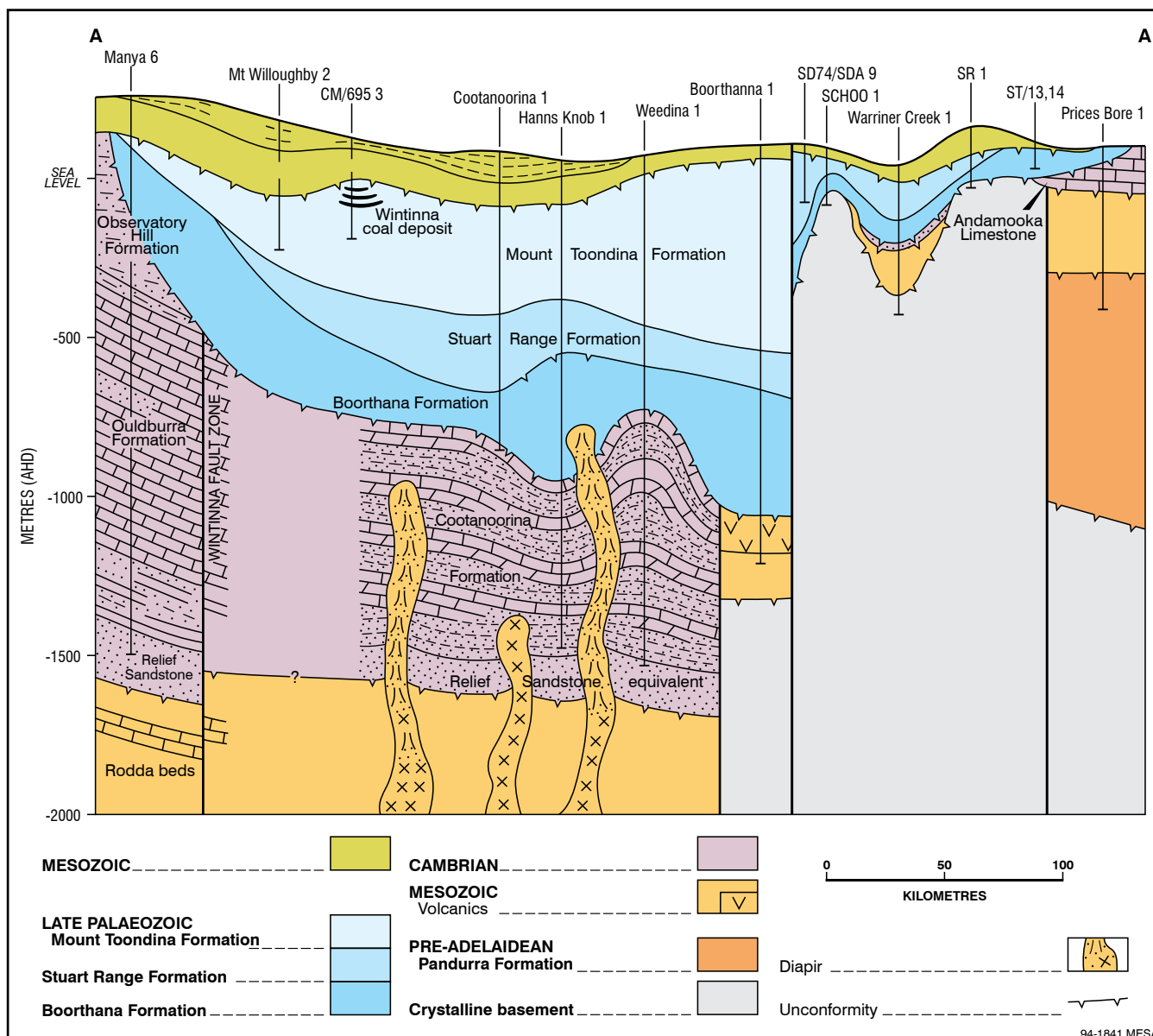


Fig. 8.13 Schematic cross-section through the Arkaringa Basin.

Mount Toondina Formation

Mount Toondina Formation overlies, usually conformably but occasionally disconformably, the Stuart Range Formation; in some parts of the basin it may unconformably overlie Boorthanna Formation. A vitrinite reflectance profile from Arkeeta 1 in the Phillipson Trough indicates that up to 170 m of Stuart Range Formation was eroded prior to deposition of Mount Toondina Formation (McBain, 1987). The latter comprises siltstone and sandstone interbedded with coal, shale and rare carbonate; the coal is generally restricted to the upper part of the formation. Three units have been tentatively recognised and are mappable by seismic methods (Wilmot, 1987). The basal unit (not recognised by Townsend, 1976) has been recorded in two drillholes, with marine shales in the lowermost 30 m of Boorthanna 1 (Cook, 1981) and interbedded anoxic marine lacustrine shales in the basal section of Arkeeta 1 (McBain, 1987). The middle unit is interpreted from seismic information and comprises westward-prograding deltaic sandstone and siltstone (Wilmot, 1987). The high frequency of reworked vitrinite in Arkeeta 1 supports a deltaic environment (McBain, 1987). The uppermost unit is dominantly lacustrine shale and minor fluvial clastics with intervals of coal formed in peat swamps.

On the basis of vitrinite reflectance data from three drillholes in the Boorthanna Trough, Moore (1982b) suggested that the Permian sequence has been subjected to pre-Jurassic uplift and erosion in the order of 0.5 to 1 km. Wilmot (1987) placed this uplift in the Middle to Late Triassic and extending into the earliest Jurassic. In contrast to the above, vitrinite reflectance data from Arkeeta 1 in the Phillipson Trough suggest that there was no significant post-Permian uplift or erosion (McBain, 1987). Late Miocene tectonism caused minor late-stage structuring of the Permian (Wilmot, 1987), uplift of blocks of Precambrian rocks (e.g. the Peake and Denison Inliers) and stripped off some of the Mesozoic sequence.

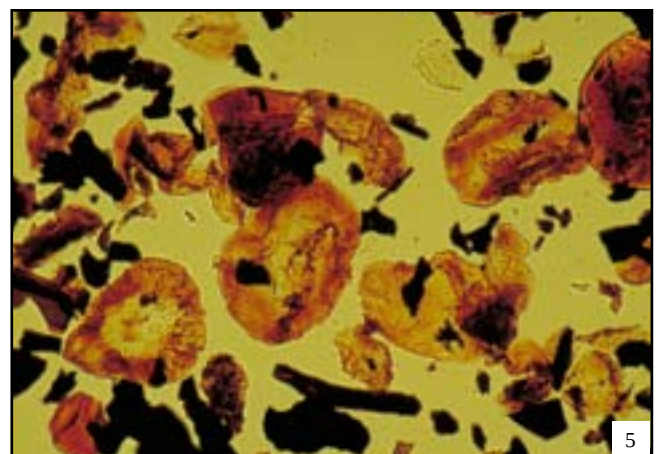
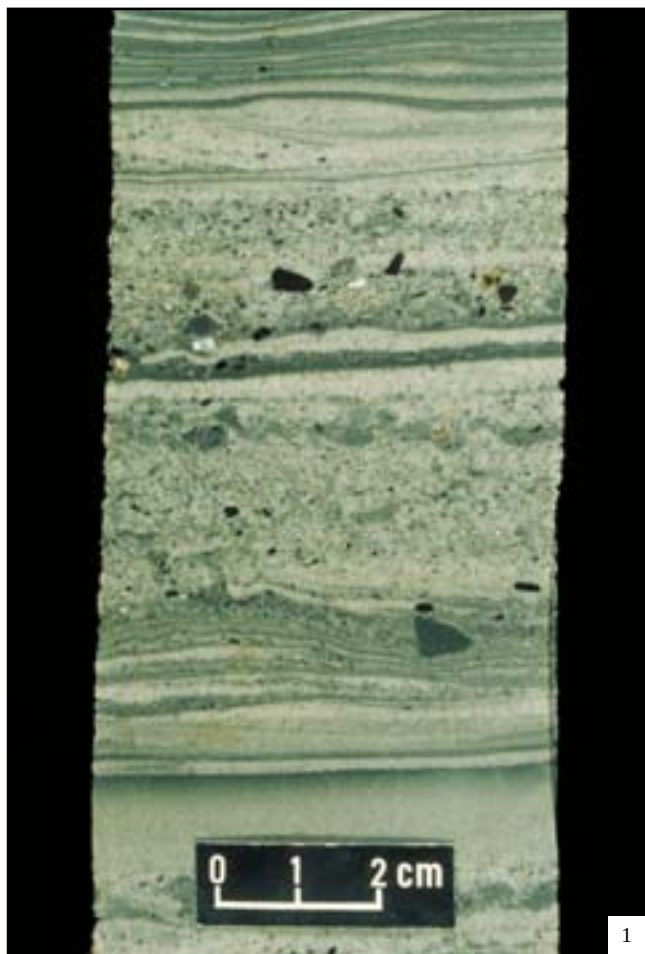
AGE

The age of the formations has been determined by palynology to be Early Permian (Harris and McGowran, 1973; Jones, 1987; Gilby and Foster, 1988) but could be interpreted to extend into the Late Carboniferous, depending on placement of the Carboniferous-Permian boundary. Boorthanna Formation and lower Stuart Range Formation are assigned to Stage 2 palynofloral zone (Asselian); the upper part of the Stuart Range Formation and the Mount Toondina Formation are Stage 3 (Late Asselian to Sakmarian).

ECONOMIC GEOLOGY

Six deposits of lignite A/Sub-bituminous C rank coal (American Society for Testing and Materials classification) aggregating more than 20 Gt of measured, indicated and inferred resource have been identified in the upper part of the Mount Toondina Formation (SADME, 1991; Fig. 8.12). These are multi-seam deposits with individual seams ranging up to 10 m, with a cumulative thickness of up to 30 m. MESA drilling in early 1995 discovered a new coal-bearing trough (Penrhyn Trough) north of Lake Phillipson, which is being evaluated.

Five exploration drillholes to date have failed to find any petroleum despite the occurrences of potential source, reservoir and seal rocks. In many areas depth of burial has not been sufficient for generation of hydrocarbons. Moore (1982b) suggested that Permian sediments may be thermally mature in the southern, deep parts of the Boorthanna Trough. McBain (1987) noted possible higher heat flow in the Phillipson Trough at ~150 Ma. Hydrocarbon potential of the trough has been evaluated by Moore (1982b) and Delhi Petroleum Pty Ltd and Santos Ltd (1989a,b).



1. Interbedded diamictite and sandstone (probably turbidites), Boorthanna Formation at 941 m in Cootanoorina 1. Photo 042447.
2. Erratic of Adelaidean glacial diamictite from the Boorthanna Formation in the Arckaringa Basin, southern Peake and Denison Ranges. Photo 041929.
3. Mount Toondina Formation exposed in a probable piercement structure at Mount Toondina; northerly view. Photo 042448.
4. Vertical view of bedding plane in carbonaceous shale containing well-preserved plant fossils in Mount Toondina Formation, 145 m in Mount Furner 1. Photo 042449.
5. *Plicatipollenites densus* and other conifer pollen from outcrop of the Mount Toondina Formation, southwestern Peake and Denison Ranges. Width of view is 11 mm. Photo 042450.

DENMAN BASIN

J.E. Hibburt

STRUCTURE AND SETTING

The name 'Denman Basin' was introduced to include all Mesozoic, Permian and early Palaeozoic sediments beneath the Eucla Basin (see below) and interpreted to extend offshore (Wopfner, 1972). In Western Australia, Permian sediments have been included within the Eucla Basin (Playford *et al.*, 1975). However, the definition now used for the Denman Basin is restricted to Permian sediments only (Foster, 1974; Youngs, 1974; Kemp *et al.*, 1977; Harris, W.K., 1980; Veevers, 1984).

The Denman Basin is an intracratonic, north-northwest-trending trough ~200 km long and varying from 20 to 70 km in width (Fig. 8.14). A narrow basement high (~10x50 km) divides the southern part of the basin (Cockshell, 1991). Permian sediments are >400 m thick to the north of the high. Seismic evidence indicates that initiation of the basin structure was by normal faulting (Fig. 8.15) with subsequent modification by glacial erosion (Cockshell, 1991). Numerous faults with north-northwest trends are evident on seismic sections; movement along some of these probably began in the Precambrian and early Palaeozoic. The Denman Basin was most likely a major depocentre and thus represents an erosional remnant of a larger basin of deposition.

The offshore extent of the basin (5 000 km²) is based on seismic data (Cockshell, 1991) and drilling results from Apollo 1 (Fig. 8.15). The onshore extent is difficult to define, and is interpreted from limited drilling information and sparse seismic coverage of very poor quality due to overlying karst-modified Tertiary limestone. Isolated pockets of Permian sediments are probably locally preserved, for example along the northern margin of the Eucla Basin in Western Australia where the Permian Paterson Formation is up to 15 m thick (Playford *et al.*, 1975).

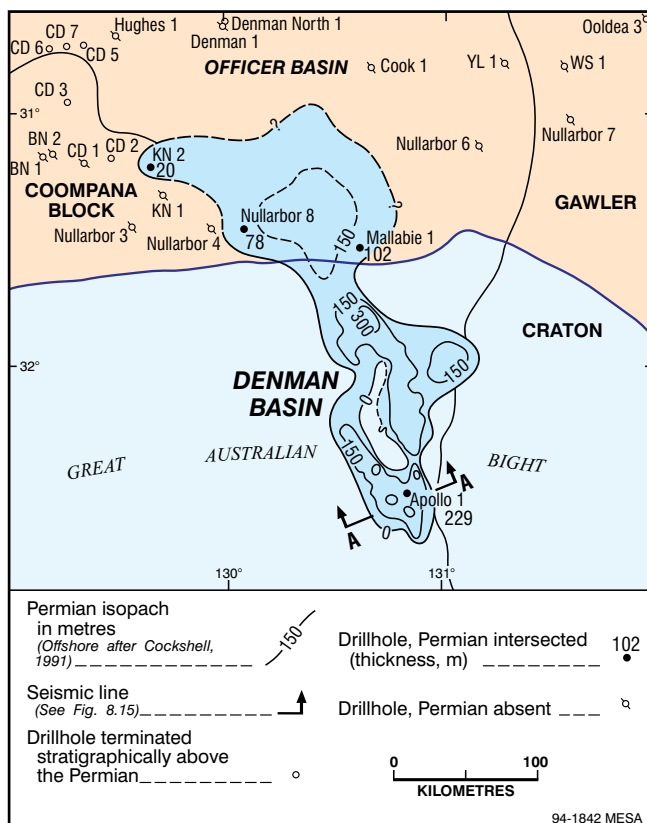


Fig. 8.14 Approximate limits of the Denman Basin.

Permian sediments have been identified in stratigraphic drillholes Nullarbor 8 and Mallabie 1, and Carpentaria mineral exploration drillhole KN 2 (N.F. Alley, SADME, pers. comm., 1991); 18 m of possible tillite and sandstone in Shell Co. of Australia Ltd CD 1 (Weeden, 1982) may be equivalent to the Paterson Formation. However, sediments originally thought to be Permian (Laws, 1966) in Hughes 2 and 3 and Denman 1 have been identified as Cretaceous (N.F. Alley, SADME, pers. comm., 1991).

STRATIGRAPHY

Permian sediments of the Denman Basin occur as a subhorizontal succession unconformably overlying early Palaeozoic or Proterozoic sediments and Precambrian crystalline rocks. The basin sediments are unconformably overlain by Cretaceous and Tertiary strata which increase in thickness from 300 m in the north to 1 000 m in the south.

Deposition in the Denman Basin commenced with a glacial phase, followed by deglaciation and glacio-eustatic marine transgression from the west into the isostatically depressed lowlands (McGowran, 1973a; Wopfner, 1980). Direction of ice movement was towards the north to northwest, from an ice source in Antarctica (Cockshell, 1991). As in the Arkaringa Basin, freshwater conditions were established following tectonically induced marine regression; sediments of Middle and Late Permian age have not been recorded.

The Denman Basin was modified during the Cretaceous by severe channelling concentrated along the eastern side (Fig. 8.15), followed by gentle tilting from east to west, resulting from a broad basement sag (Cockshell, 1991).

No detailed studies of the unnamed Permian strata have been undertaken and thus the succession is poorly understood and correlation between drillholes is difficult. In Nullarbor 8 and KN 2 the sequence comprises uniform grey claystone which, together with its depauperate assemblage of foraminifera, was used to interpret a lagoonal environment with restricted access to open marine conditions (Harris and Ludbrook, 1966). W.K. Harris (1980) described the basal sequence, which is sandy to pebbly, as possible diamictite. Apollo 1 intersected 229 m of Permian sediment (log depths 633.4 to 862.6 m) comprising three units: a basal claystone interbedded with shale and containing sandstone towards the base; a middle unit of sandstone interbedded with thin siltstone, shale and claystone; and an upper, poorly consolidated, predominantly sandstone unit with occasional siltstone interbeds. The basal claystone contains a restricted algal assemblage regarded as lacustrine rather than marine in origin (Partridge, 1976).

The 102 m thick succession in Mallabie 1 (log depths 335.3 to 437.4 m) comprises three units identifiable from cuttings but not distinguishable on logs. At the base is a thick sequence of interbedded sandstone, siltstone and claystone, overlain by grey claystone with minor siltstone and sandstone and, finally, sandstone and siltstone (Scott and Speer, 1969).

The hydrocarbon potential of the Denman Basin is regarded as poor as the sequence is thermally immature and favourable source rocks are absent.

AGE

Although Denman Basin sediments are expected to range in age from Asselian to Sakmarian, only Sakmarian sediments have been identified. A reinterpretation (N.F. Alley, SADME, pers. comm., 1990) of palynofloral lists given in well completion reports (Harris and Ludbrook, 1966; Partridge, 1976) indicated that the species list in Apollo 1 is restricted, but the sediments are no older than Stage 3a (Sakmarian); a Sakmarian age is interpreted for the Permian sequence in Nullarbor 8, which is therefore older than Artinskian as determined by Harris and Ludbrook. In support of the reinterpretation, restricted foraminiferal assemblages are similar to Sakmarian assemblages from sediments overlying

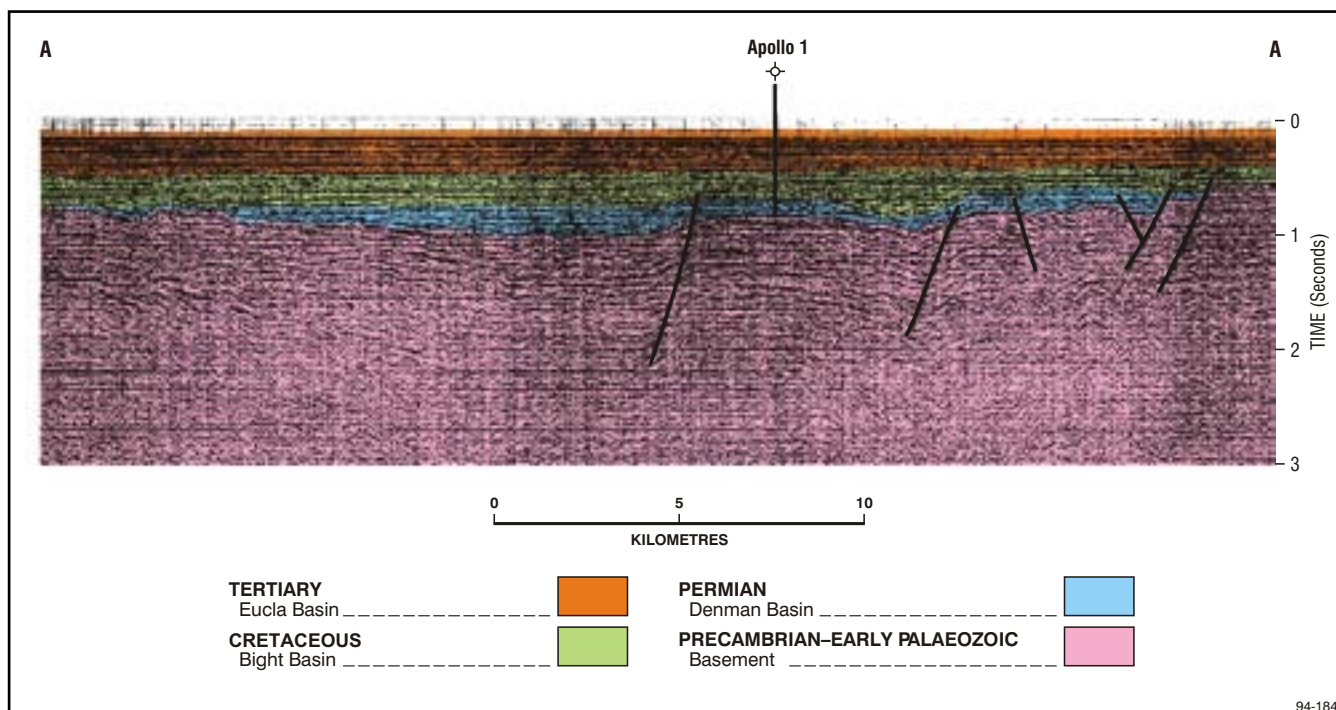


Fig. 8.15 Interpreted seismic line across the Denman Basin. The line of section is located on Figure 8.14.

diamictite elsewhere in South Australia. Examination of Mallabie 1 and KN 2 core and cuttings indicated that the sediments are Stage 3b (Sakmarian) in age (N.F. Alley, SADME, pers. comm., 1991).

Tentative correlations with the Arckaringa Basin are based on lithological and palaeontological information. Nullarbor 8 and KN 2 sediments, and the basal unit in Apollo 1, may be Stuart Range Formation equivalents, whereas Mallabie 1 sediments and the upper two units in Apollo 1 may correlate with the Mount Toondina Formation. Asselian glaciogene sediments, equivalent to the Boorthanna Formation, have not been positively identified.

MULGATHING TROUGH

J.E. Hibburt

The Mulgathing Trough is a northwesterly orientated, narrow, in part fault-bounded structure measuring 4x90 km within the Gawler Craton, 100 km south of the Arckaringa Basin. Permian sediments are preserved here and in several other small depressions to the northwest (Fig. 8.12). Geophysical surveys and mineral and stratigraphic drilling have partly delineated the geology of the trough. Seismic refraction data indicate that the trough bifurcates and shallows at its northwestern end, with the southern branch terminating abruptly against shallow basement (Hall, 1980). The southern margin is fault controlled (Nelson, 1976); faulting was probably contemporaneous with sedimentation, as in the Arckaringa Basin. Seismic data indicate that at least 800 m of Permian sediments overlie crystalline basement and, in turn, are overlain by up to 100 m of Mesozoic and Cainozoic sediments (Nelson, 1976). The thickest intersection of Permian is 214 m in SADME Konkaby 2, although the total sequence was not penetrated (Daly, 1985).

Permian sediments are correlated with the Boorthanna and Stuart Range Formations of the Arckaringa Basin, but differ in that they appear to be non-marine (SADM, 1973, 1975). The lower unit generally comprises a pale grey gritty claystone and sandstone, whereas the upper consists of grey to black mudstone, shale and siltstone. A correlation with Stage 2 and 3 palynofloras (SADM, 1975; Cooper, 1982) indicates an Asselian to Sakmarian age consistent with that in the Arckaringa Basin.

COOPER BASIN

A.J. Hill and D.I. Gravestock

STRUCTURE AND SETTING

The intracratonic Cooper Basin contains Late Carboniferous to Middle Triassic sedimentary rocks. Deposition commenced with the late Palaeozoic glaciation and ceased at the end of the Middle Triassic due to widespread compressional or transpressional deformation, regional uplift and erosion. The basin sediments overlie Cambrian-Ordovician Warburton Basin strata and are overlain by Eromanga Basin sediments. It covers an area exceeding 130 000 km², with one-third in northeastern South Australia and the remainder in southwestern Queensland (Fig. 8.16).

The stratigraphic succession comprises three upward-fining, non-marine sequences restricted to the subsurface, 1 250-3 670 m below sea level. They make up the Late Carboniferous to Late Permian Gidgealpa Group and Late Permian to Middle Triassic Nappamerri Group (Fig. 8.17). Stratigraphic successions which span the Permian and Triassic without a major break are also recorded in the Pedirka, Bowen, Galilee and Sydney Basins (Wiltshire, 1989; Channon and Wood, 1990). The ages of the various groups and their formations have been derived from palynofloral studies (Fig. 8.4).

The Cooper Basin is approximately rectangular (570x160-220 km) with its long axis trending northeast-southwest (Fig. 8.16). A prominent east-southeast striking series of faults and fault-bounded anticlines separates the basin into northern and southern parts. Various names have been applied to this feature which lies on the continental-scale G3 structural corridor (Campbell and O'Driscoll, 1989); these include Arrabury-Naccowlah Trend (Kantsler et al., 1983), Pepita-Wackett-Nockatunga Trend (Kuang, 1985) and Jackson-Naccowlah-Pepita Trend (Heath, 1989).

The G3 corridor (Fig. 8.16) separates arcuate northeast-trending structures in the southern Cooper Basin from approximately meridional trends in the northern sector (e.g. Battersby, 1976; Kuang, 1985) and corresponds to changes in pattern and trend of geothermal gradients (Pitt, 1986). The southern Cooper Basin contains the thickest (1 600 m) and most complete Permian

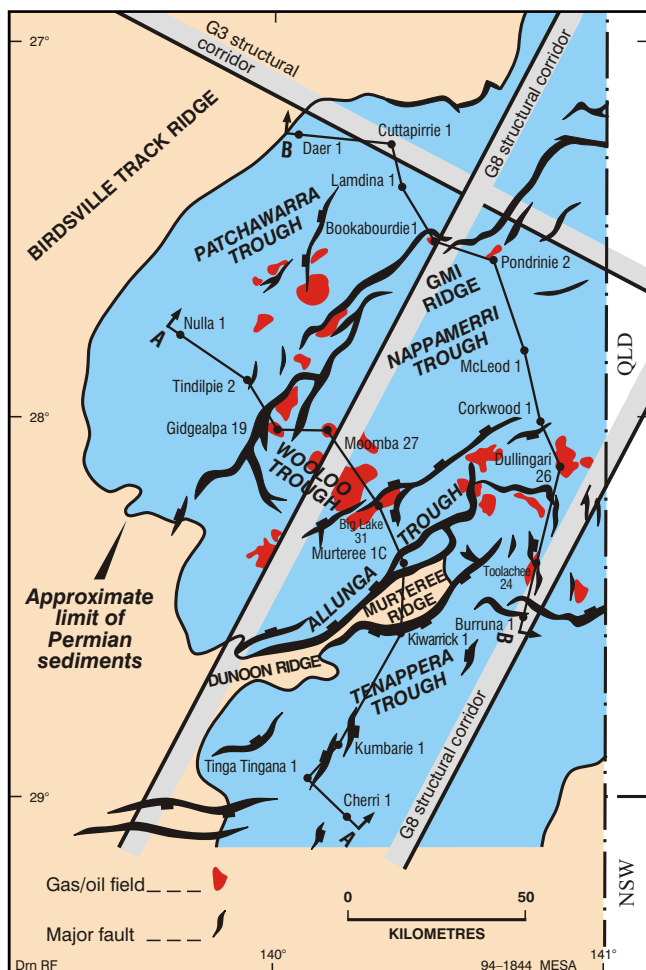


Fig. 8.16 Approximate limits and structural elements of the Cooper Basin.

succession, whereas the Permian is thin (300 m), and the Triassic sediments considerably thicker, in the northern Cooper Basin. Most hydrocarbon discoveries lie within the G8 corridor (Fig. 8.16; Campbell and O'Driscoll, 1989).

The southern Cooper Basin is divided into the Patchawarra, Nappamerri and Tenappera Troughs separated by the structurally elevated Gidgealpa-Merrimelia-Innaminka (GMI) and Murteree Ridges (Fig. 8.16). In cross-section these ridges of faulted anticlines are markedly asymmetrical (Figs 8.18, 8.19), with thick Permian sediments faulted against their northwestern margins and thinner sediments onlapping their southeastern flanks (Stuart, 1976; Heath, 1989; Apak *et al.*, 1993). Anticlinal crests are frequently devoid of Early Permian sediments, although Late Permian strata are locally continuous.

Uplift of the GMI and Murteree Ridges has long been recognised as resulting from late Early Permian diastrophism (Kapel, 1966) but the tectonic history of the basin is still much debated. Battersby (1976) and others regarded the Cooper Basin as an intracontinental downwarp, punctuated by uplift of the GMI and Murteree Ridges. Reverse faults along the GMI Ridge and offsets of possible strike-slip origin were believed to have resulted from Permian reactivation of faults penetrating the underlying Warburton Basin (Stuart, 1976). Wopfner (1985) envisaged a generally compressional tectonic regime for the Cooper Basin, a view shared by Kuang (1985) who interpreted three phases of crustal shortening by fault reactivation. Others interpret the tectonic history of the basin as extensional (Channon, 1984; Heath, 1989; Stanmore, 1989).

The model followed here is that of an intracratonic sag basin as proposed by Kapel (1966) and Battersby (1976). Growth faulting culminating in uplift at the end of the Early Permian (Daralingie unconformity) did not, however, elevate the GMI and Murteree

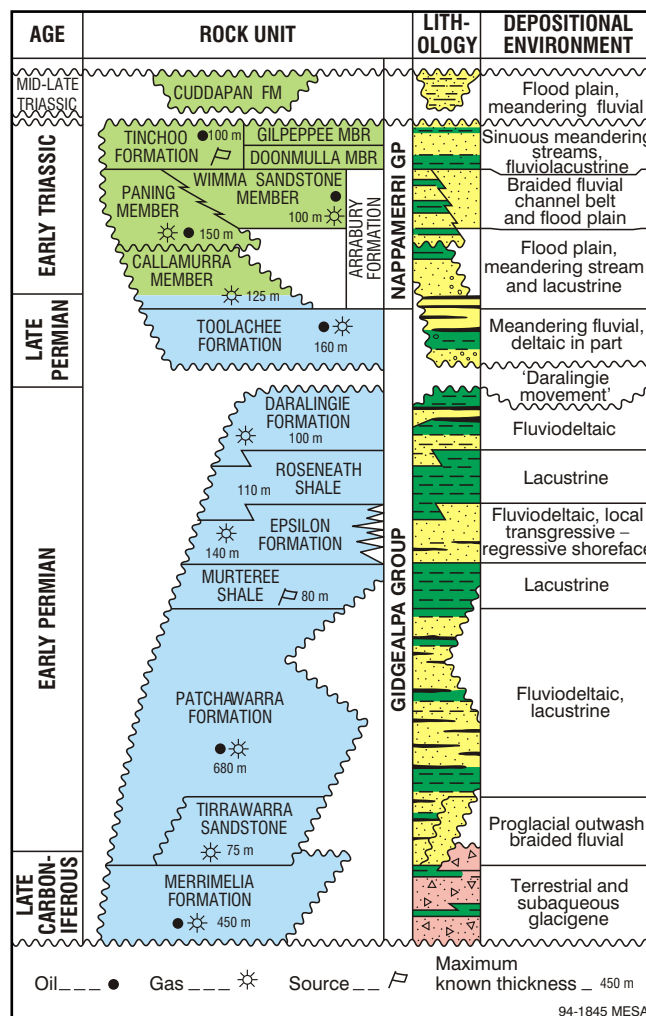


Fig. 8.17 Geological summary of the Cooper Basin.

Ridges in their present form. An alternative model may be that uplift followed older Palaeozoic or even Proterozoic northwest, meridional and northeast trends, as indicated for the GMI ridge by Apak *et al.* (1993). Early Permian depositional facies were also governed by these older structural trends as suggested by Stanmore and Johnstone (1988). Elevation of the GMI and Murteree Ridges took place later, at the close of Cooper Basin deposition, marked by the Middle Triassic unconformity.

STRATIGRAPHY

Facies analyses of the Late Carboniferous, Permian and Triassic sediments have either concentrated on individual oil and gas fields, on particular structural domains, or on widespread but stratigraphically narrow, genetically related depositional sequences. Emphasis has necessarily been on sandstone petroleum reservoirs, with less attention to widespread lacustrine units. In general, interpretation has relied upon core and cuttings samples and wireline logs, but attribute-processed seismic sections now provide valuable data in areas of sparse drillhole control (Stanmore and Johnstone, 1988; Taylor and Thomas, 1989).

The stratigraphy of the Carboniferous-Permian to Triassic southern Cooper Basin was first outlined by Kapel (1966) and formally defined by Martin (1967). The Carboniferous-Permian succession was subsequently modified and named the Gidgealpa Group (Kapel, 1972); further refinements to the group were made by Gatehouse (1972) and Morton and Gatehouse (1985). It is conformably overlain by the Nappamerri Group (Powis, 1989; Channon and Wood, 1990). The succession is summarised on Figure 8.17.

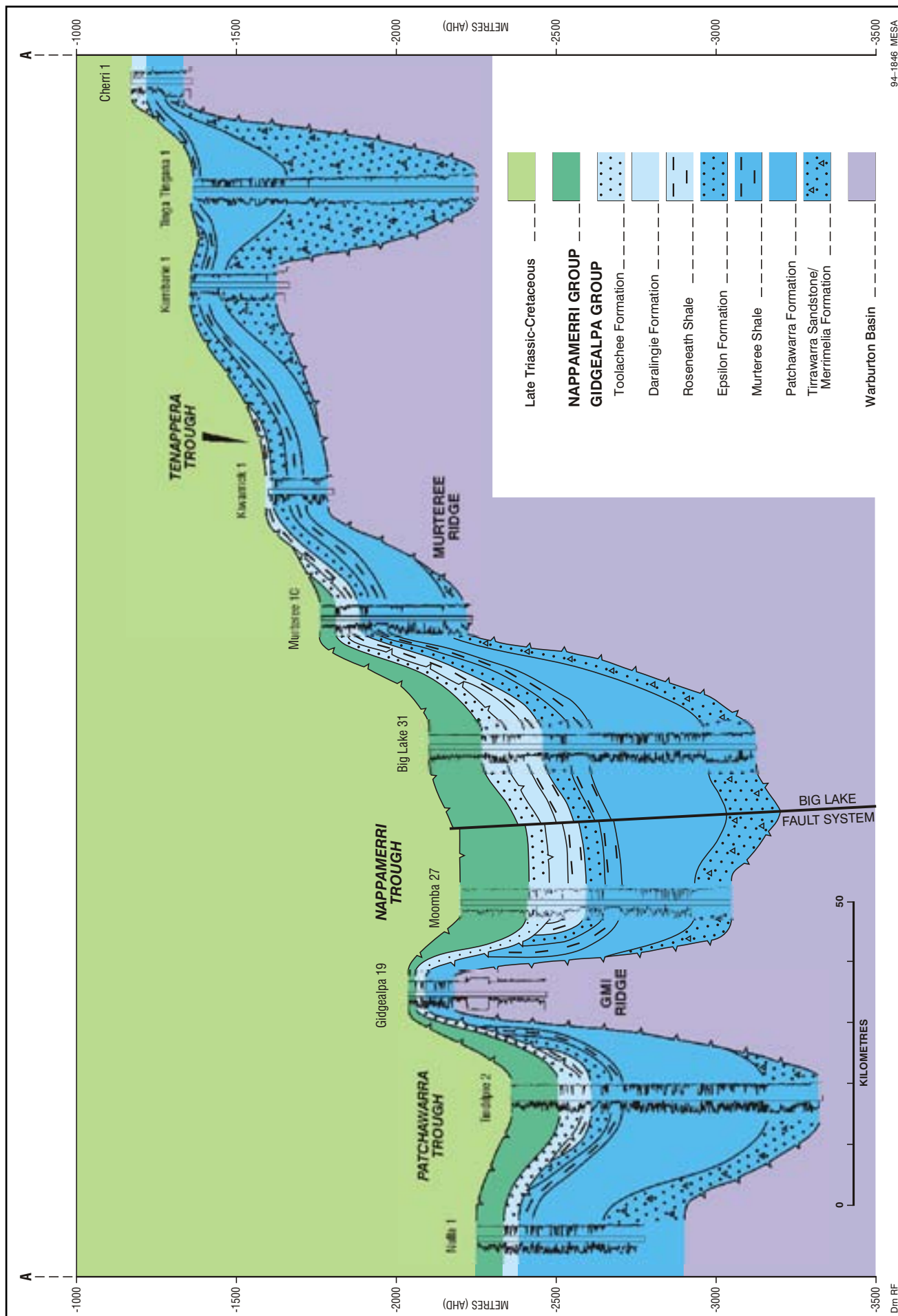


Fig. 8.18 Cross-section from Nulla 1 to Cherri 1, Cooper Basin.

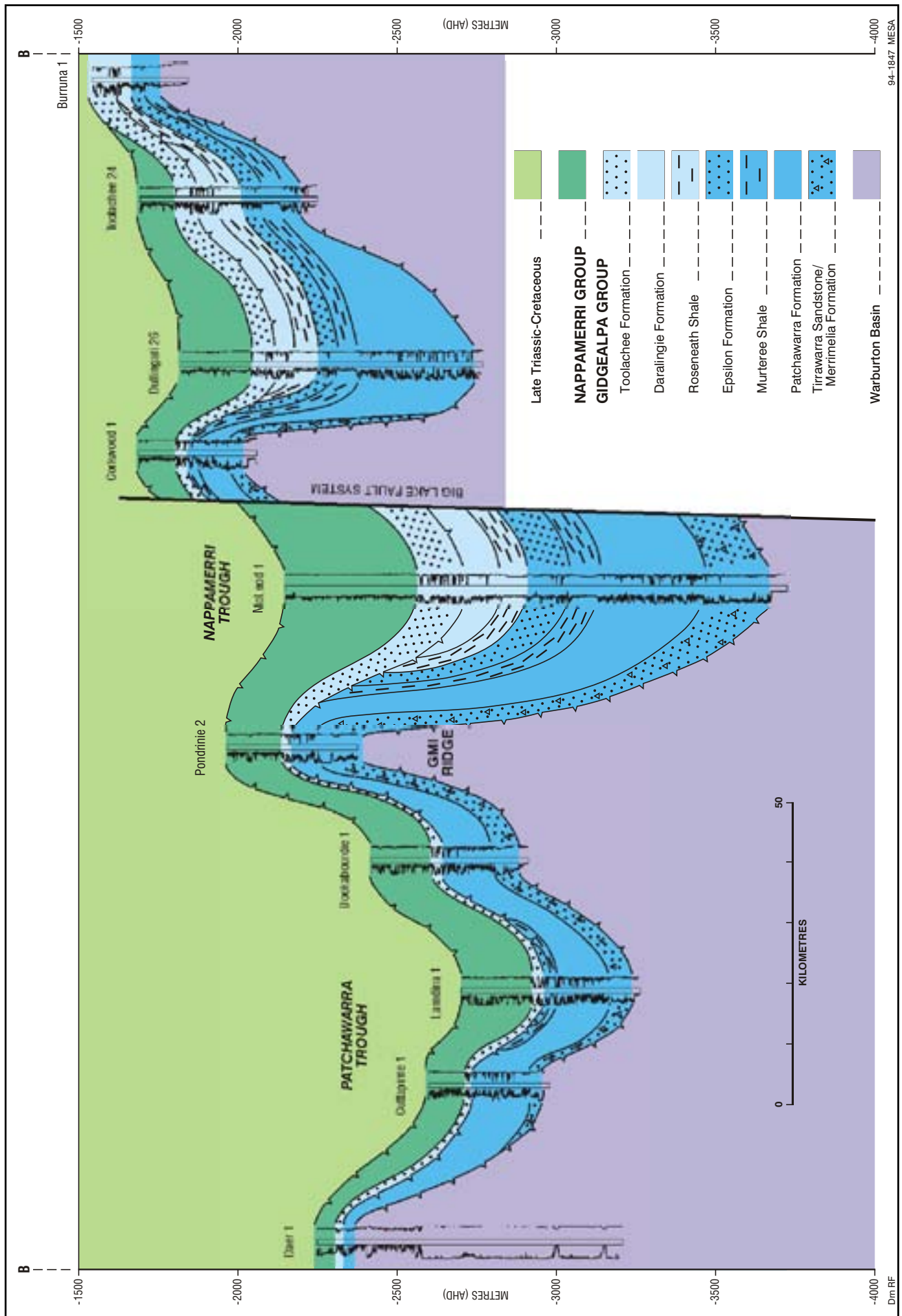


Fig. 8.19 Cross-section from Daer 1 to Burruna 1, Cooper Basin.

Gidgealpa Group

The Gidgealpa Group, deposited on Warburton Basin sediments or granitic basement, comprises, in ascending order, the fluvio-glacial Merrimelia Formation and Tirrawarra Sandstone overlain by peat swamp and flood-plain facies of the Patchawarra Formation. The lacustrine Murteree and Roseneath Shales, with intervening fluviodeltaic sediments of the Epsilon and Daralingie Formations, were deposited during a phase of continued subsidence. Early Permian uplift led to erosion from basement highs of the Daralingie and older formations. The Toolachee Formation was deposited during an interval of renewed subsidence, and is overlain by the Nappamerri Group.

The Gidgealpa Group is thickest in the Nappamerri Trough where in excess of 1 200 m are known (Fig. 8.20).

Merrimelia Formation

The basal part of the Cooper Basin comprises the Merrimelia Formation, a fluvio-glacial system (Williams and Wild, 1984) that forms part of a depositional continuum with the Tirrawarra Sandstone. The glaciogene suite forms a complex facies mosaic including relatively deep-water lacustrine rhythmites, fluvio-glacial outwash, shoreline deposits, subaerial and subaqueous diamictite and aeolianite (Williams *et al.*, 1987), passing transitionally into low-sinuosity, bedload-dominated fluvial sandstone. Syndepositional tectonism is reflected in diachroneity, onlap and erosion of clastic wedges, especially in the vicinity of growth structures and residual inliers. Fault reactivation gave rise to major escarpments (Kuang, 1985) and development of extensive alluvial fans. Fanglomerates are observed in Pinna 1, close to the Murteree Ridge and in Tinga Tingana 1 (Fig. 8.18), where the basal Merrimelia Formation facies assemblage consists of thick, polymict, framework-supported massive conglomerate containing many acid igneous clasts (Williams and Wild, 1984). The

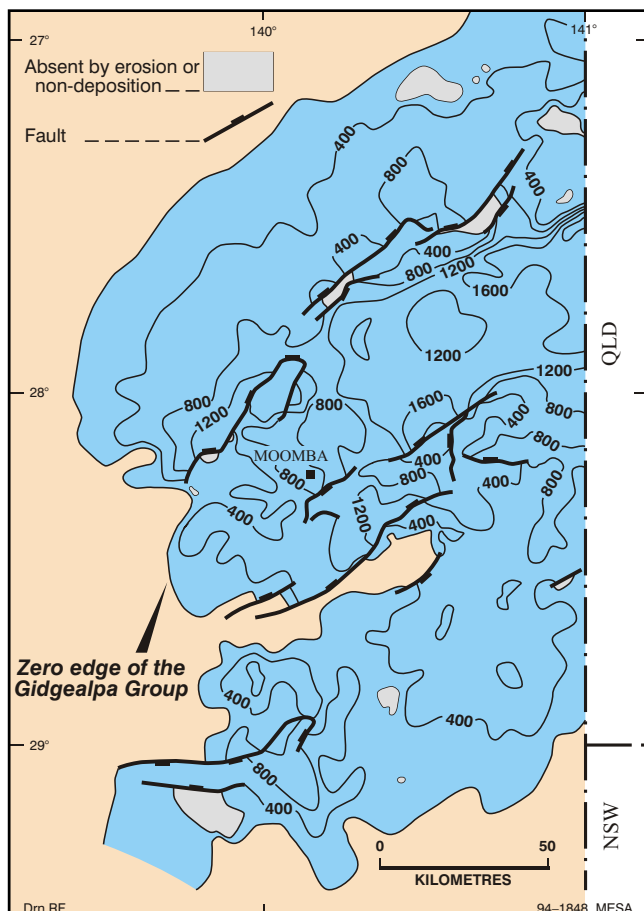
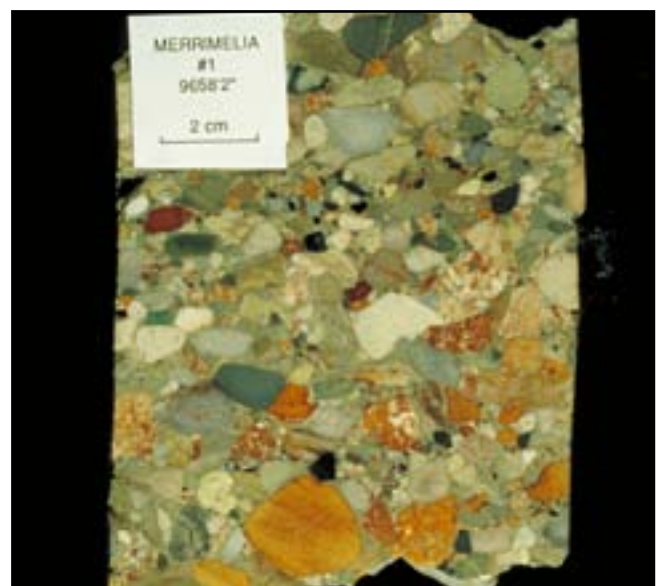


Fig. 8.20 Isopach map of the Gidgealpa Group, Cooper Basin (400 m contours).



Top: Diamictite in Merrimelia Formation at 2 925 m in Merrimelia 3. Photo 042451.
Centre: Paraglacial aeolianite in Merrimelia Formation at 2626.5 m in Merrimelia 5. Photo 042452.
Bottom: Fanglomerate in Merrimelia Formation at 2944 m in Merrimelia 1. Photo 042453.

	PATCHAWARRA TROUGH	GIDGEALPA AREA	LITHOLOGY	DEPOSITIONAL ENVIRONMENT
TIRRAWARRA SANDSTONE	Facies Association 4	Facies Association 3	Facies Association 4 Bedload dominant, low to moderate sinuosity channel system. Possible tectonic uplift.	Facies Association 3 Mixed load channel system with prominent overbank flood events. Coaly bar tops.
'INTERFACE SEQUENCE'	Facies Association 2	Facies Association 1	Facies Association 2 Shallow, braided low sinuosity fluvial channel system. Shallow water lacustrine shoreline deposition.	Facies Association 1 Progradational sheet-flood gravel and sands of outwash fans.
MERRIMELIA FORMATION			Glaciolacustrine deposition with ?turbiditic influence. Deep water lacustrine rhythmites. Shallow water lacustrine shoreline deposition. Braided outwash fan deposits. Terrestrial tillite deposit (ablation till).	94-1849 MESA

Fig. 8.21 Facies associations of the Merrimelia Formation and Tirrawarra Sandstone, Cooper Basin.

Merrimelia Formation attains a maximum thickness exceeding 290 m, but with irregular distribution.

Relatively deep-water lacustrine rhythmites and shallower water lacustrine deposits tend to be concentrated in the major troughs (e.g. the succession in Packsaddle 1), whilst fluvio-glacial conglomerate, diamictite and sandstone are common along the basement highs. Characteristically, the lithologies include lodgement till, ablation till, dropstones and wave and current-rippled, fine-grained sandstone (Williams and Wild, 1984; Fig. 8.21). An aeolian suite which occurs in several Merrimelia drillholes is characterised by 'pin-striped' sand in close association with fluvial sediments (Williams *et al.*, 1987), and can be mapped seismically.

Tirrawarra Sandstone

The Tirrawarra Sandstone of the southern Cooper Basin represents a depositional continuum with the Merrimelia Formation (Williams and Wild, 1984). A tripartite sedimentological subdivision of the Tirrawarra Sandstone has been made within the Tirrawarra Field (Gostin, 1973; Battersby, 1976; Thornton, 1979). B.P.J. Williams (1982) divided the sequence into four major facies associations of which Associations 1 and 2 demonstrate interfingering with the Merrimelia lithotypes (Fig. 8.21).

Association 1 comprises conglomerate and medium to coarse-grained sandstone characterised in some parts by flat and trough cross-bedding with low-angle plane beds and in others by planar cross-bedding. The fluvial processes appear to have been

bedload-dominated, low-sinuosity channel systems which produced longitudinal gravel bars, typical of a proximal, low-slope, braided outwash fan environment (Miall, 1977; Rust, 1978; Williams and Wild, 1984). Sheet-flood and debris-flow gravel, suggestive of high-energy unconfined flood events (Williams and Wild, 1984), are also present and are commonly related to alluvial fan deposition.

Association 2 represents a fluvial regime with more consistent discharge and channel pattern, leading to the deposition of multi-laterally stacked sandstone bodies in low-sinuosity channels. This facies may have been laid down as alluvial fans, perhaps in a distal proglacial outwash setting (Williams and Wild, 1984).

Association 3, comprising sandstone and siltstone representing a mixed-load fluvial environment, with occasionally preserved bar-top coal and backswamp mudstone, passes laterally into the lower Patchawarra Formation. A return to coarse-grained facies is represented by Association 4 which is probably a lateral equivalent of Association 3, but originating from localised tectonic activity (Williams, B.P.J., 1982). Typical facies are upward-fining exotic conglomerate and sandstone beds with clasts reworked from the Merrimelia Formation. Conglomerate up to 60 m thick in the upper Tirrawarra Sandstone in the Big Lake Field probably belongs in this association which is well developed in the Gidgealpa Field (cf. Steveson and Gravestock, 1984). Fault reactivation was the likely cause of deposition of coarse clastics (Williams, B.P.J., 1982).

Tirrawarra Sandstone is quite variable in thickness (average ~70 m) and its distribution mirrors that of the Merrimelia Formation. The sandstone most commonly underlies, but locally passes laterally into, the lower Patchawarra Formation. Minor coal and mudstone units in the upper Tirrawarra Sandstone indicate gradual waning of the depositional slope. A combined isopach map of the Merrimelia Formation and Tirrawarra Sandstone is shown on Figure 8.22.

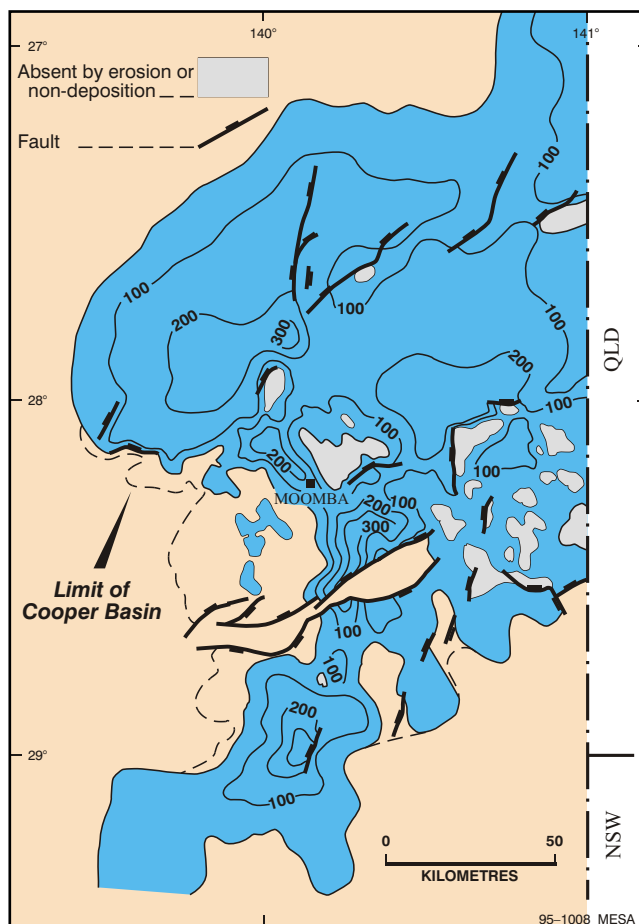


Fig. 8.22 Isopach map of the Merrimelia Formation and Tirrawarra Sandstone, Cooper Basin (100 m contours).

Patchawarra Formation

The Patchawarra Formation conformably overlies Tirrawarra Sandstone in deeper portions of the basin but, on structural ridges and near the basin margin, the Patchawarra may unconformably overlie Merrimelia Formation and/or pre-Permian rocks. Along the southeastern margin of the Cooper Basin, the Tirrawarra Sandstone grades laterally into thick basal sandstone of the Patchawarra Formation. The Patchawarra comprises fluviolacustrine, flood-plain and minor deltaic deposits up to 680 m thick deposited by meandering river systems flowing in a north to northeasterly direction (Thornton, 1979; Stanmore and Johnstone, 1988). Fault movement early in deposition influenced rapid development of thick sand sheets with significant onlap against structurally positive areas. Several local mid-Patchawarra unconformities are present in the southern Cooper Basin (Apak *et al.*, 1993). Extensive peat swamps (now preserved as 30 m thick coal) present during deposition of the middle part of the formation in the Patchawarra Trough, and widespread flood-plain sediments at the top of the formation, reflect periods of tectonic quiescence.

The Patchawarra Formation can be divided into three broad facies assemblages within the Patchawarra Trough region (Kapel, 1972) although, in the more tectonically active southeastern and southwestern portions of the Cooper Basin, these assemblages are not developed. The lowermost assemblage is dominated by interbedded fluviolacustrine to coal-swamp siltstone, sandstone and minor coal. The middle assemblage contains areally extensive meandering stream point-bar facies up to 15-20 m thick, reflecting episodic fluvial incursions into a coal-swamp to deltaic environment. Sandstone predominates in upward-fining cycles which are often abruptly overlain by channel abandonment plugs of mudstone (oxbow lake sediments) that provide seals for the porous reservoir sandstone. Gradual thickening of the middle and upper assemblages reflects regionally stable tectonic conditions. Flood-plain lacustrine deposits predominate over delta-plain sediments in the upper assemblage. In the western part of the Cooper Basin, broad deltaic complexes up to 30 km wide are crossed by a number of channel systems that flowed eastwards into extensive lakes (Stanmore and Johnstone, 1988).

The Patchawarra Formation varies greatly in depth of burial, ranging from 1 500 to 3 500 m below sea level, with the greatest depth being in the Nappamerri Trough (Fig. 8.23).



Point-bar sandstone with collapsed overbank deposits in Patchawarra Formation at 2773 m in Tirrawarra 2. Photo 042454.

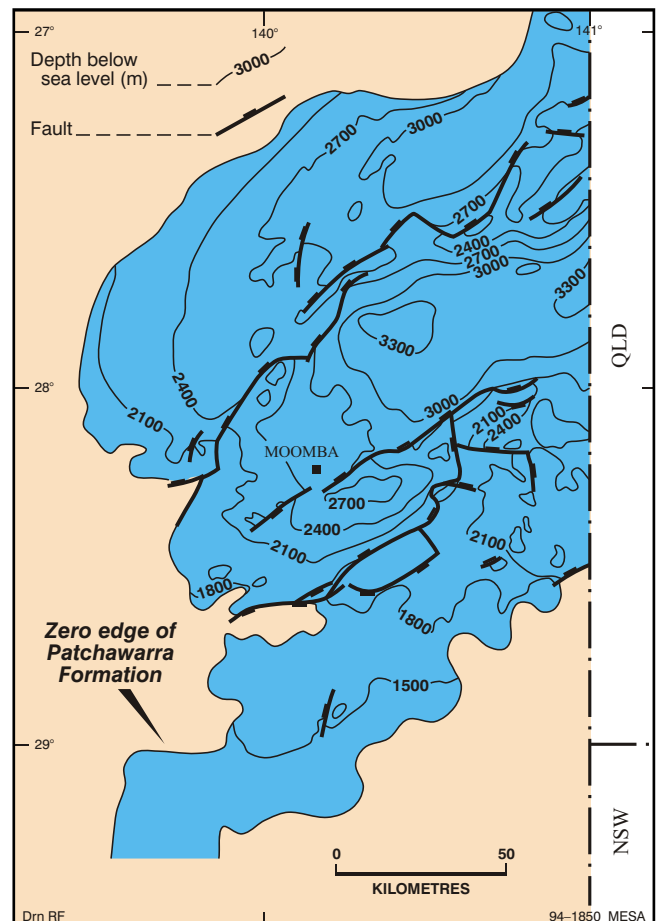


Fig. 8.23 Structure map of the top of the Patchawarra Formation, Cooper Basin (300 m contours).

Murteree Shale

The Murteree Shale is up to 80 m thick, conformably overlies the Patchawarra Formation and varies only slightly in thickness away from the structural highs, indicating a low-relief depositional surface (Thornton, 1979). The Murteree Shale consists of a homogeneous succession of grey-green shaly siltstone containing thin lenses of fine sandstone, deposited in a large lake ('Murteree Lake') which spread southwestwards across the basin. This succession forms a regional seal above the Patchawarra Formation, especially where it is thicker in the Nappamerri Trough.

Epsilon Formation

A series of fluviodeltaic to lacustrine shale, siltstone, sandstone and coal deposits up to 140 m thick comprises the Epsilon Formation (Fairburn, 1992). This unit has its depocentre in the Nappamerri Trough where it conformably overlies Murteree Shale and represents a major reduction in lake levels across the basin. The formation is absent along the northern flank of the Patchawarra Trough and western flank of the Cooper Basin, and has been eroded from the major structural ridges. Deposition of deltaic sediments in the basal Epsilon Formation followed the easterly retreat of the 'Murteree Lake' (Thornton, 1979). It is quite possible that the eastern portion of the basin in South Australia has undergone prolonged submergence as suggested by the marked decrease in coal. The presence of a series of transgressive barrier-bar sands in the Munkarie-Toolachee-Kidman areas reflects waxing and waning lacustrine conditions (Fairburn, 1992). Shoreface sand, reworked into barrier bars, although vertically restricted, is laterally extensive (Morton, 1983) and may have locally excellent reservoir potential. Two transgressive sequences punctuated by regression occur in the Moomba-Big Lake region where distributary mouth-bar deposits formed on fluvial dominated deltas (Taylor *et al.*, 1991).

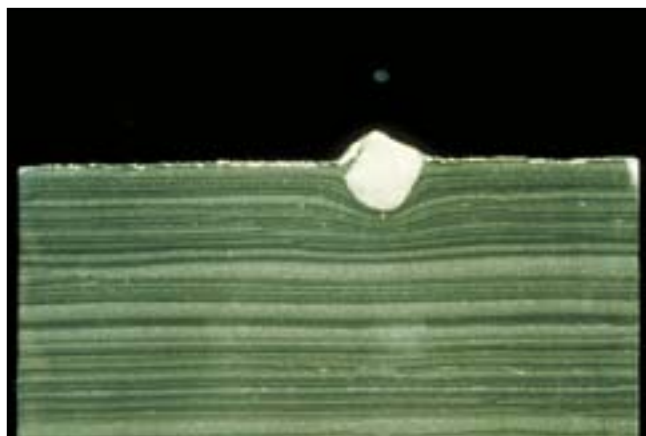
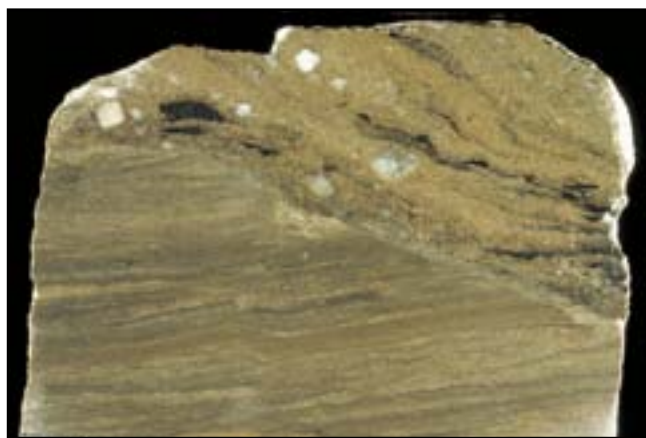
Roseneath Shale

The Roseneath Shale is a diachronous unit overlying Epsilon Formation and, like the Murteree Shale, comprises shaly siltstone with minor sandstone lenses deposited in lacustrine conditions which extended from the east. It is restricted to the Nappamerri and Tenappera Troughs, with a thin veneer preserved in the southwestern Patchawarra Trough.

Daralingie Formation

Easterly regression of the lake ('Roseneath Lake') in which the Roseneath Shale was laid down led to deposition of the Daralingie Formation, a fluviodeltaic sandstone, siltstone and coal sequence conformably overlying the Roseneath (Gatehouse, 1972; Thornton, 1979; Morton and Gatehouse, 1985). The formation is absent on the structural ridges and is thickest (up to 100 m) in the Nappamerri Trough. Sandstone and coal interbeds tend to be thin and exhibit an upward-coarsening gamma ray log signature (Morton and Gatehouse, 1985); they are interpreted to be deltaic in origin (Stuart, 1976; Thornton, 1979) and represent distributary mouth bars.

Deformation in the late Early Permian resulted in regional truncation of the Daralingie Formation and underlying units from the GMI and Murteree Ridges, often exposing basement on structural culminations. Uplift led to gradual removal of older Permian sediments over much of the Patchawarra Trough, defining the western margin of the Cooper Basin and resulting in angular discordance with the Late Permian Toolachee Formation (Figs 8.18, 8.19). Any connections that may have existed with the Pedirka and Arckaringa Basins to the west would have been severed at this time. The time break at the Daralingie-Toolachee Formation boundary is a disconformity or angular unconformity of possible regional significance, as noted by Morton and Gatehouse (1985). There remains a possibility, however, that in the deeper parts of the basin the two formations are conformable.



Top: Toolachee Formation unconformably overlying Patchawarra Formation at 2027 m in Della 2. Width of view is 80 mm. Photo 042455.

Bottom: Dropstone in finely laminated mudstone of the Toolachee Formation at 1837 m in Tilpatee A1. Width of view is 0.1 m. Photo 042456.

Toolachee Formation

The Toolachee Formation comprises sandstone deposited in mixed-load fluvial channels of moderate to high sinuosity, overbank and flood-plain lacustrine mudstone, and peat-swamp coal (Stuart *et al.*, 1988; Fairburn, 1989). The unit is, with minor exceptions on ridge crests (e.g. Murteree Ridge), ubiquitous across the basin, reaching a maximum thickness of 160 m. A basal conglomerate containing reworked clasts is locally evident. Upper parts of the Toolachee Formation were deposited under deltaic conditions, locally in flood-basin lakes (e.g. Dullingari North 4; Williams, 1984). The contact between the Toolachee Formation and overlying Arrabury Formation is conformable across the basin and is marginally diachronous on palynological evidence (Channon and Wood, 1990).

Nappamerri Group

The Nappamerri Group consists of the Late Permian to Middle Triassic Arrabury Formation, comprising the Callamurra, Paning and Wimma Sandstone Members, and the Middle Triassic Tinchoo Formation. The group is unconformably overlain by the Late Triassic Cuddapan Formation (see Ch. 9), which has limited areal extent and is unconformably overlain by Eromanga Basin sediments. Facies of the Nappamerri Group have been described by Papalia (1969), Youngs and Boothby (1985), Powis (1989), and Channon and Wood (1990).

The depositional geometry of the Late Permian-Triassic sequence is related to orogenic movements along the eastern edge of the continent, which diminished during the Early Triassic (Wopfner, 1985). These movements produced upward and westward tilting of the southeastern Cooper Basin along the Naryilco Lineament (Queensland), resulting in erosion of Early Triassic sediments from that basin margin and the gradual northerly shift in depocentres from the Middle Triassic onwards. An increase in stream gradients led to reorganisation of the drainage system and deposition of dominantly sandstone in the Paning and Wimma Sandstone Members.

Middle Triassic sedimentation was concentrated to the north of the GMI Ridge as a result of uplift along the Naryilco Lineament. In the Late Triassic the whole basin was strongly tilted towards the northeast (Kuang, 1985). Reactivation of pre-existing structures led to widespread erosion and the hiatus provides an obvious boundary between the Cooper Basin and younger Mesozoic sediments (Channon and Wood, 1990).

The Nappamerri Group is thickest in the Nappamerri and Patchawarra Troughs where in excess of 550 m are known (Fig. 8.24).

Arrabury Formation

Redbeds of the Arrabury Formation were deposited in a broad flood-plain setting that frequently supported lakes or was crossed by either east to southeasterly (Channon and Wood, 1990) or east to northeasterly (Youngs and Boothby, 1985) fluvial channel systems. The abrupt change from a coal-forming to an oxidising redbed environment was part of a continental-scale response to a eustatic fall in sea level and probably a change in climate as reflected in a major palynofloral change (Powis, 1989). Triassic climate is interpreted to have been temperate with, at times, marked seasonality in rainfall, in contrast to the cold to cool temperate Permian climate.

The Callamurra Member includes a Late Permian basal carbonaceous facies which grades rapidly into slightly carbonaceous, grey to reddish purple mudstone and siltstone containing thin interbeds of fine to medium-grained sandstone and lithic sandstone. Core from this facies shows the sediment to be finely cross-bedded and locally bioturbated, perhaps indicating deposition in a fluviolacustrine environment. On the basis of limited drillhole information, Youngs and Boothby (1985) concluded that a river system existed to the northwest of the basin, possibly discharging eastwards.

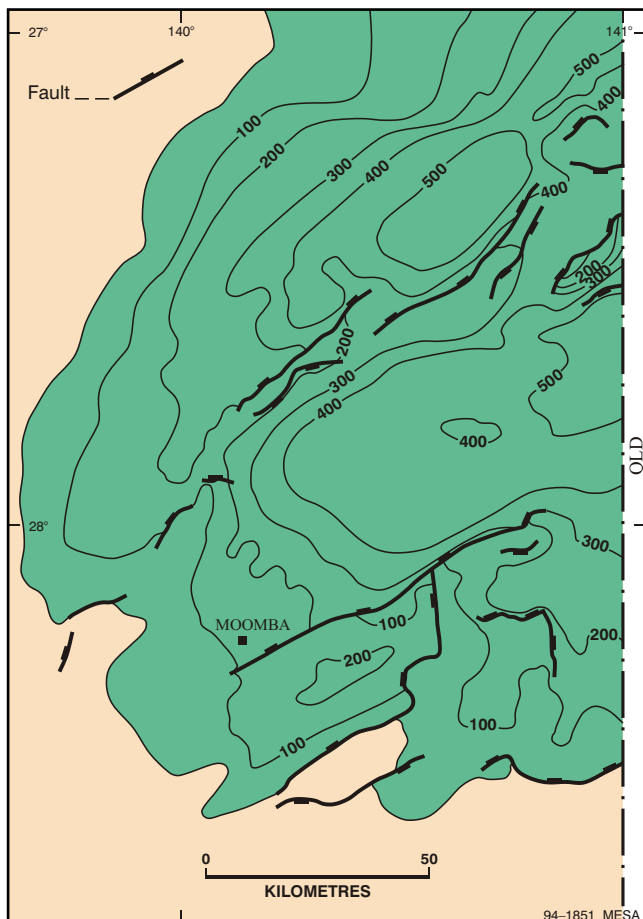


Fig. 8.24 Isopach map of the Nappamerri Group, Cooper Basin (100 m contours).

Alternatively, flow may have been in a southeasterly direction with meandering tributary systems occurring on the flood-plain to the north and south of the main channel (Channon and Wood, 1990). The Callamurra Member is thickest in the Nappamerri Trough but has an erosional edge to the north of the Patchawarra Trough.

The Paning Member comprises upward-fining sandstone, with minor mudstone deposited in a fluviolacustrine environment. The contact between this member and the underlying Callamurra Member is unconformable in the northern Patchawarra Trough and Arrabury Trough, and is taken to lie at the base of the first significant sandstone (Channon and Wood, 1990). The fluvial system in which deposition occurred is believed to have been located along the flanks of the GMI Ridge and Patchawarra Trough (fig. 7 of Youngs and Boothby, 1985). Away from the major channel system, the member consists of siltstone and interbeds of fine sandstone, consistent with deposition on a broad flood-plain. The Paning Member reaches its maximum thickness in the central Patchawarra Trough and Arrabury Trough, whereas to the south of the GMI Ridge the member has suffered erosion.

The Wimma Sandstone Member is a clean sandstone interbedded with minor mudstone which conformably overlies Paning Member except along portions of the GMI Ridge and in Cuttipirrie 1 (Youngs and Boothby, 1985). Deposition is believed to have occurred in a southeasterly flowing, low-sinuosity, braided river system parallel to that of the Paning Member; away from the main channel belt, deposition may have been by low-energy streams meandering across the flood-plain. The Wimma Sandstone Member is distributed widely across the basin, with the main depocentre in the Patchawarra Trough. Erosion has removed the member along portions of the GMI Ridge and along the southern and western margins of the basin. In general, reservoir potential in this member is high.



Synaeresis cracks and load casts in fine-grained sandstone of the Arrabury Formation at 2364 m in Merrimelia 3. Photo 042457.

Tinchoo Formation

The Tinchoo Formation conformably overlies the Wimma Sandstone Member, except possibly in the 'Innamincka' area, and in South Australia is restricted to the Nappamerri and Patchawarra Troughs. It consists of interbedded sandstone, siltstone and mudstone deposited by highly sinuous meandering streams. Upward-fining sandstone beds dominate the basal part of the formation, grading upwards into laterally extensive fluviolacustrine siltstone and mudstone which provide a regional seal across the basin, except where interrupted by faulting (Channon and Wood, 1990). The marked similarity in facies and age between the Tinchoo Formation and Moolayember Formation (of the Galilee and Bowen Basins) suggests that they are lateral equivalents, although continuity across the Canaway Ridge has yet to be proven.

Although the Tinchoo Formation has been divided into Doonmulla and Gilpeppie Members (Powis, 1989), the subdivision in the southern Cooper Basin can only be vaguely determined and becomes even less clear in southwestern Queensland, where minor coal seams occur in a heterolithic suite of upward-fining flood-plain deposits. The top of the Tinchoo Formation is marked by a sharp erosional surface which is overlain by the Cuddapan Formation or Eromanga Basin sediments.

ECONOMIC GEOLOGY

The South Australian portion of the Cooper Basin represents part of the largest onshore hydrocarbon province in Australia. Natural gas was first encountered in Gidgealpa 2 in 1963, and oil in Tirrawarra 1 in 1970. Gas reserves as at 1 January 1993 were assigned to 121 fields. Permian to Middle Triassic oil reserves were assigned to 25 fields (Fig. 8.25). Annual production in 1992 for the 68 gas and 11 oil fields that are currently exploited was $\sim 4.5 \times 10^9$ m³ of sales gas, 0.8 MkL of oil and condensate, and 474 000 t of liquefied petroleum gas. The major producing horizons are highlighted on Figure 8.17. Petroleum currently provides 71% (41% oil, 30% natural gas) of the State's primary energy requirements for electricity generation and transport fuel. Estimated remaining reserves are 85×10^9 m³ (2 860 petajoules) of sales gas, 5.9×10^6 m³ of oil, 4.1×10^6 m³ of condensate and 2.9×10^9 m³ of ethane. Pipelines supply gas from the Moomba processing plant to Adelaide and Sydney, whilst a liquids pipeline carries oil, condensate and LPG, supplemented by Eromanga Basin oil, to Port Bonython.

Enhanced oil recovery (EOR) methods are used in the basin. In 1986, fracture stimulation and ethane injection commenced in the Tirrawarra Sandstone reservoir of the Tirrawarra and Moorari Fields. Combined, these methods have lifted production by 300% in some wells, and the potential exists for further EOR developments. Significant gas reserves are contained in low permeability sands and may be recoverable by massive hydraulic fracturing techniques.

PEDIRKA BASIN

J.E. Hibburt and D.I. Gravestock

STRUCTURE AND SETTING

The Pedirka Basin is an intracratonic basin straddling the South Australia-Northern Territory border and perhaps extending eastwards into Queensland (Fig. 8.26). The South Australian portion represents about one-fifth of the total basin area and is confined to the subsurface. However, sediments crop out along the northwestern basin margin in the Northern Territory.

All late Palaeozoic sediments in the area are included in the Pedirka Basin (Wopfner, 1972; Youngs, 1976). Seismic surveys and petroleum exploration drillholes completed in the late 1970s and 1980s have helped refine the boundaries of the basin as early limits were conjectural. In South Australia, the interpreted area of preserved Permian sediments has been considerably reduced (Fig. 8.26). Initially the name 'Pedirka' was used for the western portion only and the term 'Simpson Desert Sub-basin' (Moore, 1982e) was applied either to the eastern portion or to overlying Mesozoic sediments in the Simpson Desert-Lake Eyre region (Sprigg, 1962, 1963; Canaple and Smith, 1965; Williams, 1973a). Much of the eastern area has been shown to contain Triassic (but not late Palaeozoic) strata, now assigned to the Simpson Basin (Moore, 1982e, 1986a). A major north-northeasterly trending structural ridge, previously referred to as the McDills Trend or McDills-Mayhew Trend (Youngs, 1976; Giuliano, 1988), is here defined as the Dalhousie-McDills Ridge. This ridge comprises shallow folded and faulted Devonian and older rocks that separate

the Pedirka Basin into eastern and western sectors (Fig. 8.27). A thin cover of late Palaeozoic sediments is preserved across the ridge linking the two depocentres.

The western boundary of the basin is extended here to include the Eringa Trough. Youngs (1976) introduced this term for a southwest-northeast-trending Palaeozoic structural depression inferred to contain largely pre-Permian strata. Rogers (1986), however, using Bouguer gravity anomaly data, interpreted that the trough could contain as much as 2 500 m of Carboniferous-Permian and older Palaeozoic sediments. Mount Hammersley 1 (New, 1988b), near the eastern upfaulted margin of the trough, intersected 990 m of Carboniferous-Permian strata, suggesting that the Eringa Trough is a Late Carboniferous to Early Permian feature. This conclusion is strengthened by isopach geometry of underlying Devonian sediments which indicate alternating belts of thick and thin strata aligned in a northwest-southeast direction, perpendicular to the trough axis and thus structurally unrelated to it (see Ch. 7).

In South Australia, the Musgrave Block and Bitchera Ridge mark the approximate western and southern boundaries of the Pedirka Basin. The late Palaeozoic strata unconformably overlie southeastern Amadeus Basin and western Warburton Basin rocks which were deformed during the Alice Springs and possibly Delamerian Orogenies. A final northwest-southeast compressional phase of the Alice Springs Orogeny in the Middle to Late Carboniferous created thrust faults such as occur at Mount Hammersley. This phase initiated subsidence of the Eringa Trough and probably of similar troughs in the western Pedirka Basin in the Northern Territory. Early Permian sediments were subsequently deposited in a tectonically quiescent sag phase.

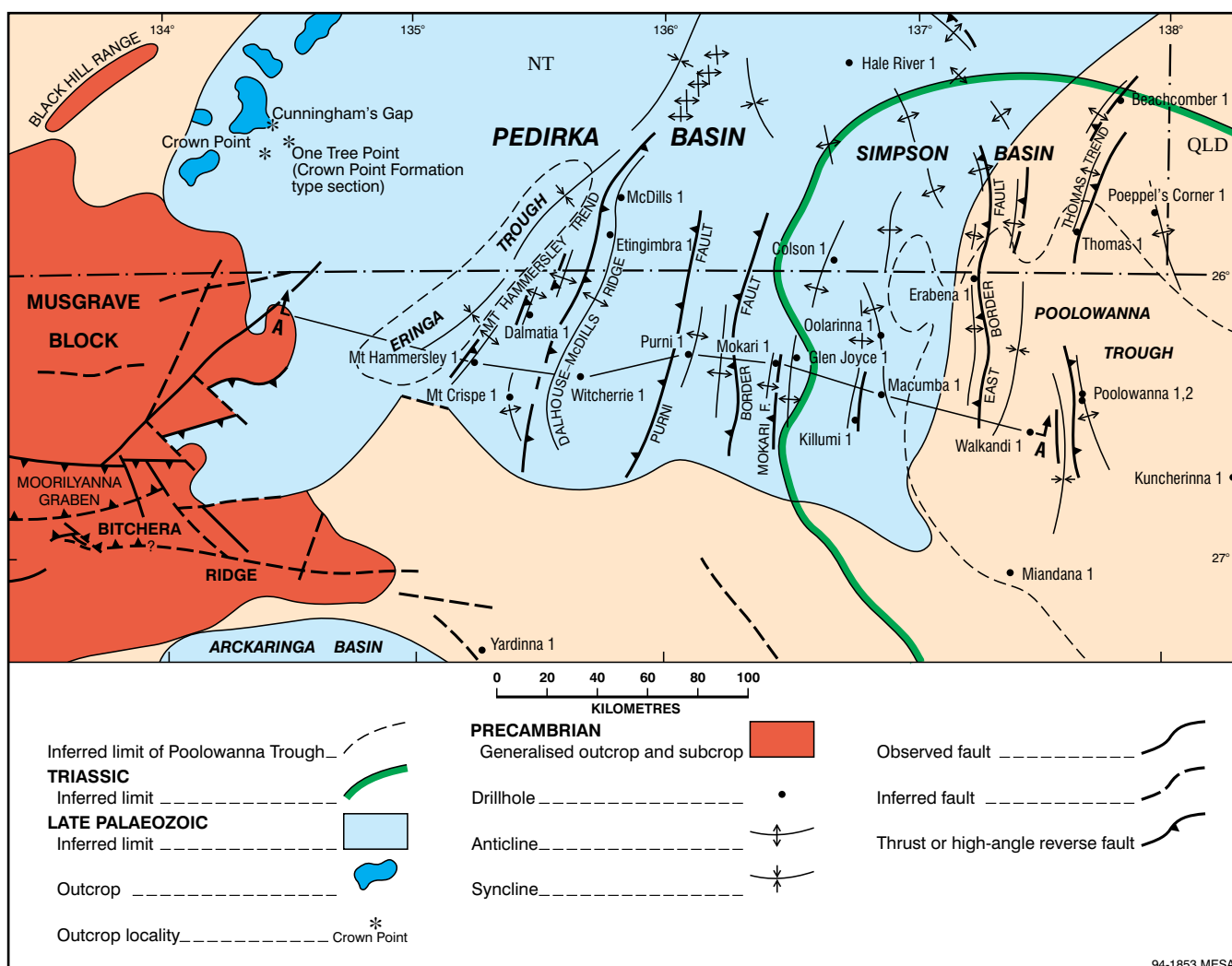


Fig. 8.26 Approximate limits and structural elements of the Pedirka Basin.

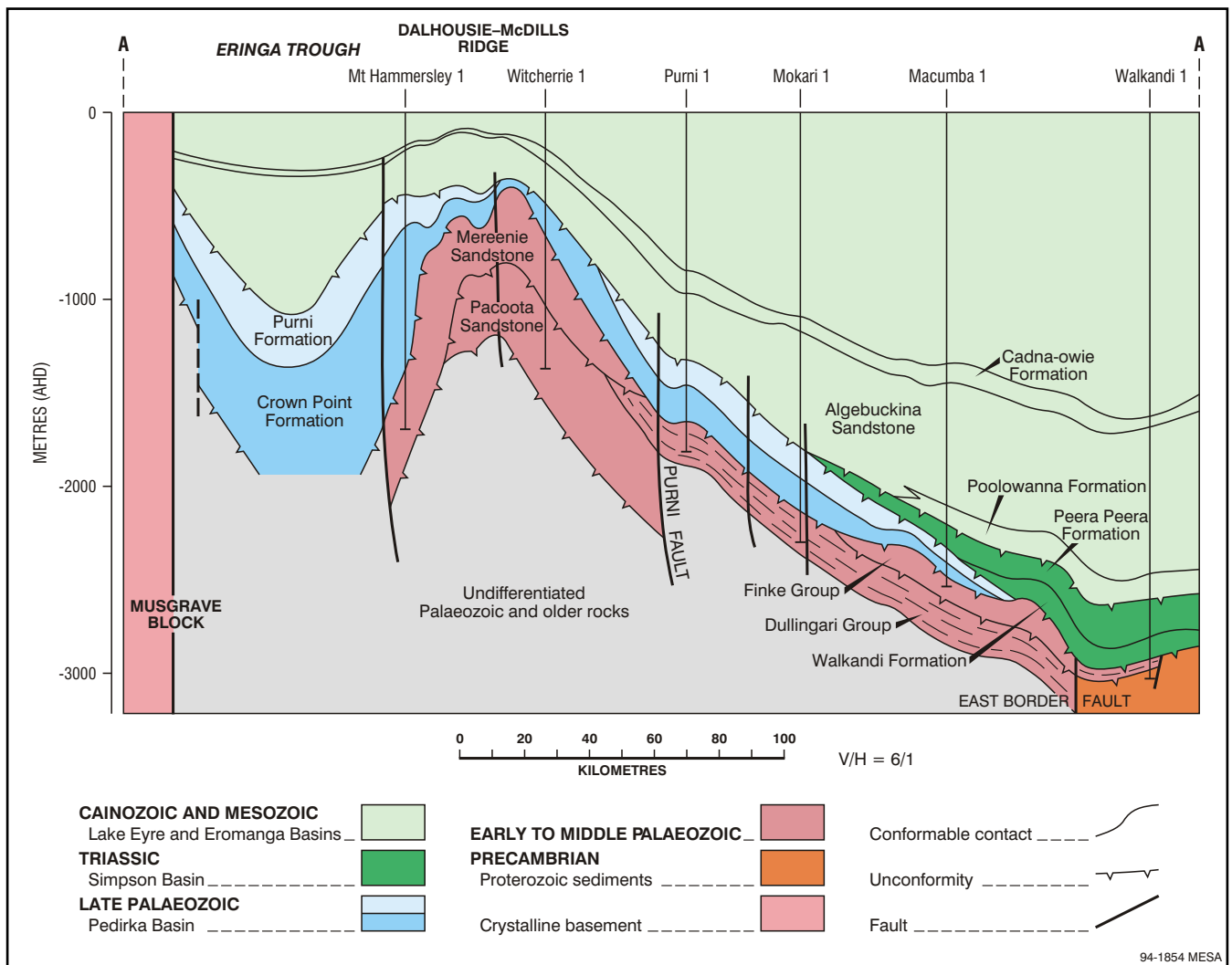


Fig. 8.27 Schematic cross-section through the Pedirka Basin.

North-northeasterly striking high-angle reverse faults separate north-northwest to north-northeast-trending anticlinal complexes, the most prominent being the Dalhousie-McDills Ridge. The Eringa Trough west of the ridge may contain up to 1 525 m of Carboniferous-Permian sediments (Giuliano, 1988) whereas on the eastern side a maximum thickness of ~600 m is interpreted near Mokari 1 (Fig. 8.28). Sediments of the Pedirka Basin may once have extended southwards across the Bitchera Ridge to the Arckaringa Basin. It is unlikely that the Pedirka and Cooper Basins were connected since Early Permian deposition was largely restricted to localised structural troughs in the latter basin.

Two major compressional episodes which deformed Pedirka Basin sediments during the late Early to Late Permian and then during the Tertiary resulted from reactivation of older fault systems (Wopfner, 1985; Giuliano, 1988). Tectonic uplift during the first phase led to basin-wide erosion, particularly on the Dalhousie-McDills Ridge. In the western portion of the basin, uplift and erosion continued through the Triassic whereas regional tilting and subsidence of the eastern portion initiated deposition of the onlapping Simpson Basin sediments.

STRATIGRAPHY

The Pedirka Basin comprises two major sedimentary sequences, the basal Crown Point Formation (Wells *et al.*, 1966) and the Purni Formation (Youngs, 1975). Both formations are disconformably overlain by Triassic strata near the eastern limit of the basin (Moore, 1986a), but elsewhere the Purni Formation is disconformably overlain by Jurassic sediments of the Eromanga Basin. However, in Mount Crispe 1 and Witcherrie 1 on the southern part of the Dalhousie-McDills Ridge, the Purni Formation has been eroded and the Crown Point Formation is disconformably overlain by Jurassic sediments.

Crown Point Formation

The Crown Point Formation consists of fluviglacial and glaciolacustrine sediments. Along the western margin of the basin in the Northern Territory, such as at Cunningham's Gap and Crown Point (Fig. 8.26), sandstone and sandy and pebbly diamictite are interpreted to have been deposited rapidly by lacustrine outflow following breaking of retaining ice dams (Crowell and Frakes,

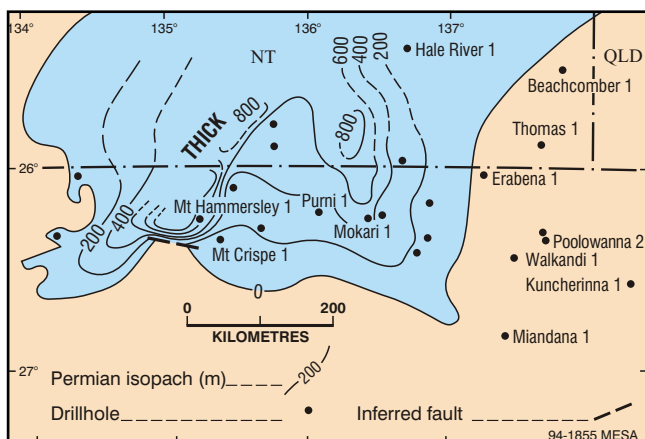


Fig. 8.28 Isopach map for the total Permian succession in the Pedirka Basin.



Cross-bedded fluvio-glacial sandstone in the Crown Point Formation at 1400 m in Mount Hammersley 1. Photo 042458.

1971a). Distribution and lithology of erratics suggest that glacial transport was locally to the south to southeast (Wopfner, 1980). Data from petroleum drillholes indicate that the formation includes extensive diamictite, fluvio-glacial sandstone, glaciolacustrine ripple-laminated sandstone and siltstone, and clay-dominated rhythmites (Giuliano, 1988). The coarser sediments are common at structurally elevated localities whereas glaciolacustrine facies are restricted to the troughs. The formation generally terminates with a clean sand which is regarded as marking the end of glaciation (Giuliano, 1988). The Crown Point Formation is probably equivalent to the Merrimelia Formation and Tirrawarra Sandstone in the Cooper Basin (Giuliano, 1988).

Purni Formation

The Purni Formation contains three facies suites distinguishable on the basis of relative proportions of sandstone, shale and coal (Youngs, 1975). Carbonaceous shale predominates over lenticular sandstone and coal stringers in the lowest suite. Upward-fining cross-bedded sandstone with minor shale and coal characterises the middle suite, while fine-grained, silty, cross-bedded sandstone passes rapidly upward into carbonaceous shale with abundant coal seams in the uppermost suite. These suites vary in relative thickness across the basin. The Purni Formation is equivalent to the Patchawarra Formation in the Cooper Basin and was deposited in a flood-plain environment containing meandering river systems and extensive coal-forming peat swamps. It is likely that the Purni Formation conformably overlies Crown Point Formation, forming a depositional continuum that records the transition from a glaciogenic environment to fluvial and/or swamp conditions through the Late Carboniferous and Early Permian, as does the lower Gidgealpa Group in the Cooper Basin (Williams and Wild, 1984).

The age of the Crown Point Formation has been determined on palynofloral evidence to be Asselian (Stage 2) and the Purni Formation to be ?Late Asselian to Sakmarian (Stage 3; Youngs, 1975, 1976).



Finely laminated, carbonaceous uppermost member of the Purni Formation at 2001 m in Mokari 1. Photo 042459.

ECONOMIC GEOLOGY

The Purni Formation contains oil-prone and gas-prone organic matter but, despite 13 exploration drillholes being completed, no significant hydrocarbon shows have been encountered. In many areas vitrinite reflectance and temperature data suggest that heat flow has been too low for the generation of significant quantities of hydrocarbons from Permian source rocks. Sandstone of excellent reservoir quality is locally developed in the Crown Point and Purni Formations. Intraformational seals occur, and Triassic and Jurassic rocks provide a regional seal over the eastern part of the basin.

Sub-bituminous coal occurs throughout the Purni Formation, mostly in the upper part. Ten seams varying in thickness from 1 to 8 m are present between 620 and 760 m in Mount Hammersley 1 (New, 1988b). Since this drillhole is located on the margin of the Eringa Trough there is the possibility of thicker but deeper coal in the central part of the trough. Coal seams in the eastern portion of the basin range up to 3 m in thickness and occur at depths between 550 and 2 000 m. The coal is uneconomical to mine because of depth and remote location.

OTHER OCCURRENCES OF PERMIAN SEDIMENTS

N.F. Alley

Blinman area

A small isolated outlier of glaciogenic sediments (Alpana Formation) unconformably overlies Precambrian Callanna Group rocks 3 km south of Blinman (Coats, 1964b; Morton *et al.*, 1984). The sediments are preserved in a palaeovalley eroded in bedrock and extend for 200-500 m along two west-bank tributaries to Blinman Creek. The Alpana Formation is ~20 m thick and comprises a 3 m thick lower lodgement till overlain by probable proglacial sand. The till contains faceted, polished and striated clasts with a strong north-south fabric which, when corrected for local tectonic

deformation, indicates that the direction of ice movement was northwesterly. Overlying the till is a very fine to medium-grained sandstone with small pockets of gravel and limestones. Although the origin of the sand is unknown, the lithology of the larger clasts is similar to those of the lodgement till and thus the sand is interpreted as a proglacial deposit (Morton *et al.*, 1984).

The age of the Alpana Formation cannot be determined by palaeontology because the unit is deeply weathered but, based on the presence of a lodgement till, it is correlated with late Palaeozoic glaciogene sediments elsewhere in South Australia. The preservation of this isolated Permian outlier within the Flinders Ranges suggests that the sediments were once more extensive but have been largely removed by erosion.

Springfield Basin

An unnamed sequence of Permian sediments is preserved in the small synclinal Springfield Basin in the southern Flinders Ranges. The sediments overlie Adelaidean strata and are overlain by Middle Triassic coal-bearing clastics. The Permian section comprises at least 132 m of interbedded quartz conglomerate, sandstone, siltstone, shale and coal. The coal seams vary from very thin beds a few centimetres in thickness to over 7 m thick. The steep dip of the seams, depth of burial, high ash content and limited tonnage suggest that the coal is of little economic interest. The basal conglomerate has yielded small diamonds (SADME, 1990).

The age and depositional environment of the Springfield Basin sequence is poorly known. The sediments were formerly regarded as Triassic and deposited as piedmont fans built out into local lakes and swamps (Johnson, 1960a). Preliminary palynological analysis of samples at 106 and 224 m in Freeport-McMoran DDH 4 indicated probable Middle Triassic and Late Permian (Upper Stage 5) ages, respectively. The age of the sequence between the two samples is unknown although conglomerate similar in lithology to the Permian section extends up to ~204 m. Whether the section from 224 m to the bottom of the sequence at 356 m is Late or Early Permian is unknown, but the lack of diamictite or clasts of glacial origin suggests that an Early Permian glaciogene facies is not present.

Padthaway Ridge

Drillholes on the Padthaway Ridge have disclosed glaciogene sediments resting on an irregular bedrock floor. They comprise a few metres of diamictite overlain by a similar thickness of bedded silt and clay which, in turn, is overlain by Tertiary lignite and clay. The glaciogene sediments probably have a wide but sporadic distribution across the Padthaway Ridge and may represent a link between the Troubridge and Nadda Basins. Restricted Early Permian palynofloras containing low frequencies of acritarchs in the diamictite facies suggest a correlation with glaciomarine facies in the Troubridge and Nadda Basins.

POSSIBLE LATE PALAEOZOIC IGNEOUS EVENTS

N.F. Alley

Stratigraphic studies and dating of a number of rocks of igneous and hydrothermal origin suggest possible late Palaeozoic thermal events in northern South Australia. The Radium Ridge Breccias and associated Mount Gee Sinter of the Mount Painter region are discussed in Chapter 7; recent palaeomagnetic evidence suggests a Carboniferous age for these hydrothermal events (Idnurm and Heinrich, 1993).

Granites beneath the Permian in the Cooper Basin, sampled from Moomba 1 and McLeod 1, have produced U-Pb zircon dates of Carboniferous age (342 ± 28 and 310 ± 17 Ma, respectively; Krieg, Callen *et al.*, 1990). The intrusion of granite appears to be concurrent with the Alice Springs Orogeny of central Australia and the Kanimblan Orogeny of southeastern Australia.



Bouldery and weathered lodgement till facies of Alpana Formation exposed in a west bank tributary to Blinman Creek near Blinman. Photo 042460.

Whether these thermal events are related to some more general igneous activity is unknown. However, studies of Permian sediments in the Springfield and Arckaringa Basins have revealed the presence of kimberlitic indicator minerals. This evidence, together with the indications of heating in the Cooper Basin and northeastern Flinders Ranges, implies some pre-Permian (?Carboniferous or ?Ordovician; see Ch. 7) heating event or events that were widespread in the northern part of South Australia.

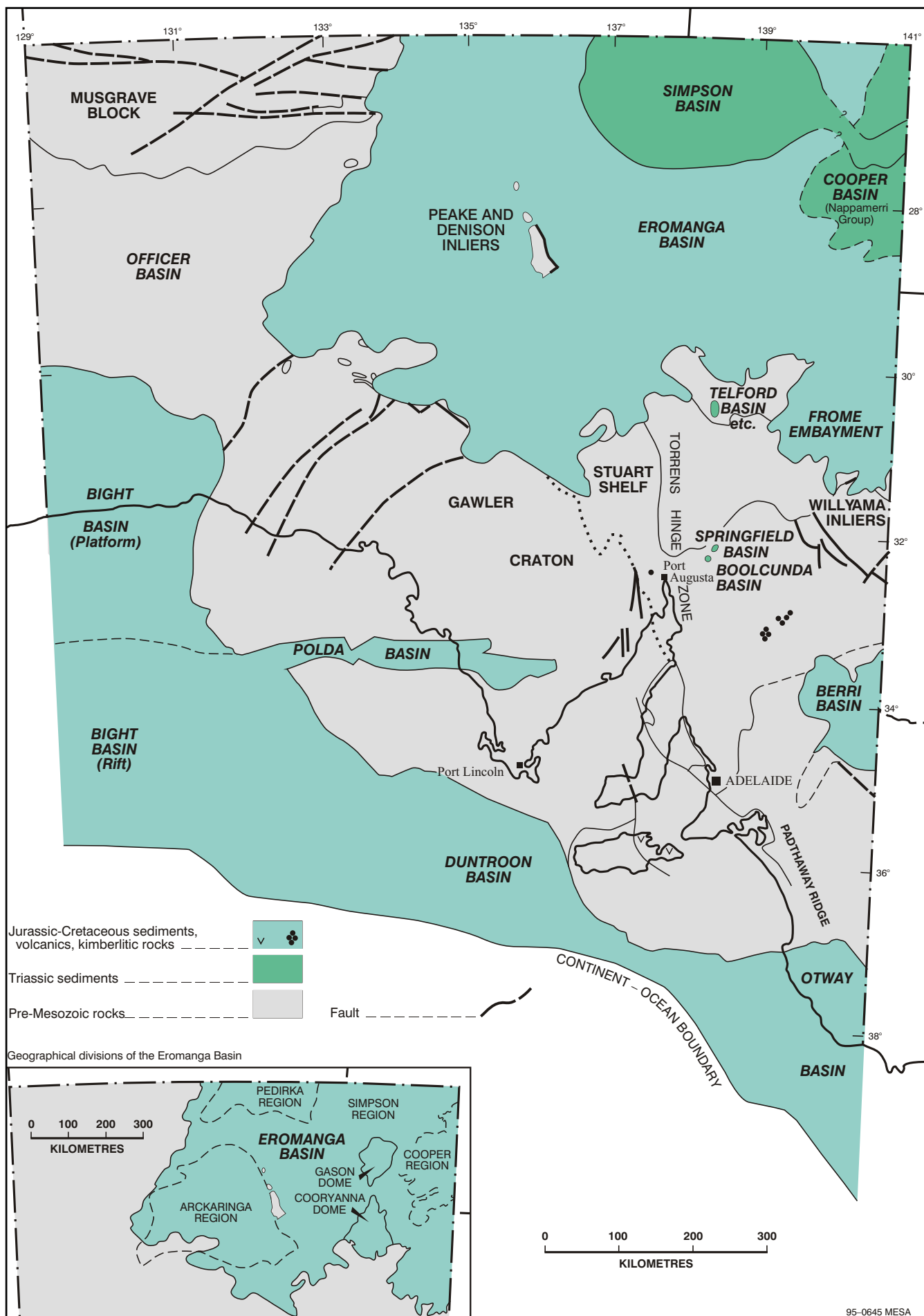


Fig. 9.1 Distribution of Mesozoic basins in South Australia.

Chapter 9

MESOZOIC

Compiled by G.W. Krieg

*with contributions from E.M. Alexander, N.F. Alley, D. Armstrong, M.G. Farrand,
C.G. Gatehouse, D.I. Gravestock, A.J. Hill, G.W. Krieg, G.W. Kwitko, J.G.G. Morton and P.A. Rogers*

INTRODUCTION

The Mesozoic rocks of South Australia consist of Triassic sediments of limited extent, Jurassic-Cretaceous sediments of regional extent, and Jurassic volcanics kimberlitic and lamprophyres which occur only locally (Figs 9.1, 9.2). Reflecting global changes of the Mesozoic Era, these rocks record the break-up of Gondwana and dispersion of the plates, a shift towards cooler climates, and the change from a landmass-dominated to an oceanic regime with sea-level oscillation exerting a major influence on sedimentary environments. A reconstruction of some palaeogeographic changes through the Mesozoic is shown on Figure 9.3.

In the Triassic Period, South Australia was landlocked, tectonically stable and of low relief. The climate varied from wet to dry and was, perhaps, warmer than in later Mesozoic times. Subaerial weathering and pedogenesis were widespread and deposition took place in low-energy, terrestrial freshwater environments of low gradient streams, flood plains, shallow lakes and swamps. These deposits, consisting of either fine carbonaceous clastics or redbeds, are now preserved in northeastern South Australia in the Cooper and Simpson Basins and in small intramontane basins within the Flinders Ranges.

Early in the Jurassic Period there emerged a new geotectonic regime that produced a fundamental palaeogeographic change. Plate movements induced crustal instability and created two tectonically distinct basin types of different style and age. Epicratonic basins, located inland and typified by the Eromanga Basin, are broad, continental downwarps which developed in response to distant plate-boundary interactions north and east of the Australian continent. Their sedimentary successions are thin, blanket-like and widespread; they are only gently deformed, and contain mainly quartzose to sub-labile lithologies. Deposition began early in the Jurassic (Sinemurian) and ended in the mid-Cretaceous (?Turonian).

In contrast, rift basins such as the Otway Basin are located along the southern coastal margin where continental extension and rifting led to the eventual separation of Australia from Antarctica. These basins are elongate, narrow and deep, structured with down-to-basin and back-to-coast fault systems and contain a significant mudstone-greywacke-volcanogenic assemblage. Deposition in this region did not start until late in the Middle Jurassic (Callovia) and continued through the Cretaceous and into the Tertiary.

Sedimentation during the Jurassic and Cretaceous took place in a generally moist, cool to cold climate with possible aridity early in the Jurassic. Terrestrial conditions prevailed at first but gave way gradually to marine conditions as warping and rifting progressed, and sea level rose. In the epicratonic basins, terrestrial conditions returned after only a relatively brief marine transgression when continental uplift in that region outpaced rising sea level. In the rift basins, however, marine or marginal marine conditions prevailed and continued into the Tertiary. The Jurassic-Cretaceous deposits, and those of the Triassic, are pre-eminently clastic; it was not until the Tertiary that major carbonate deposition occurred. The deposits include:

- coarse to fine-grained sandstone, siltstone, shale and coal of medium to low-energy fluvial, lacustrine and coal-swamp environments
- chemically immature range-front fan and braid-plain clastics and basic extrusives of the rift basin margin
- grey, fossiliferous, marine mudstone, siltstone and glauconitic sandstone.

In this chapter, the palynological scheme of Helby *et al.* (1987) is used throughout as the biostratigraphic framework.

The Mesozoic rocks of South Australia are of major economic importance. Coal from the Triassic Leigh Creek Coal Measures in the Telford Basin is used as fuel at Port Augusta to generate almost 60% of the State's electricity requirements. Oil is produced from Eromanga Basin reservoirs, and gas is recovered from Cretaceous sequences of the Otway Basin; the Bight and Duntroon Basins are considered hydrocarbon prospective. The Eromanga Basin comprises part of the Great Artesian Basin, a huge groundwater reservoir of vital importance to the pastoral and mining industries. Opal from Coober Pedy and Andamooka, the world's largest suppliers of this precious stone, occurs in the lower part of the Cretaceous marine interval of the Eromanga Basin.

TRIASSIC INLAND BASINS

COOPER BASIN

Triassic deposits of the Cooper Basin region comprise the Nappamerri Group and Cuddapan Formation. The Nappamerri Group is discussed in Chapter 8.

Cuddapan Formation

The Cuddapan Formation, defined by Powis (1989), is a thin, areally restricted sandstone-siltstone-minor coal unit which, in South Australia, is confined to the Patchawarra Trough (see Ch. 8). The maximum known thickness of 52 m is in Cuddapan 1 (Queensland). Palynomorph assemblages indicate a Late Triassic age. The unit unconformably overlies the Nappamerri Group and is unconformably overlain by the Poolowanna Formation of the Eromanga Basin.

SIMPSON BASIN

D.I. Gravestock

A seismic survey across part of the Simpson Desert in northern South Australia revealed an eastward-thickening sediment package between Early Permian strata of the eastern Pedirka Basin, and Jurassic strata of the Eromanga Basin. This sediment package was thought to be of either Late Permian or Triassic age (Williams 1973), the latter being confirmed when Poolowanna 1 penetrated the succession (Wiltshire, 1978b).

Poolowanna 1 recovered oil on drillstem testing, the first in the Mesozoic of South Australia, but further seismic surveys and exploration drilling have failed to discover commercial petroleum. They have, however, added to geological knowledge of the Simpson Desert region. The Triassic sequence is now known to be separated unconformably from the underlying Pedirka

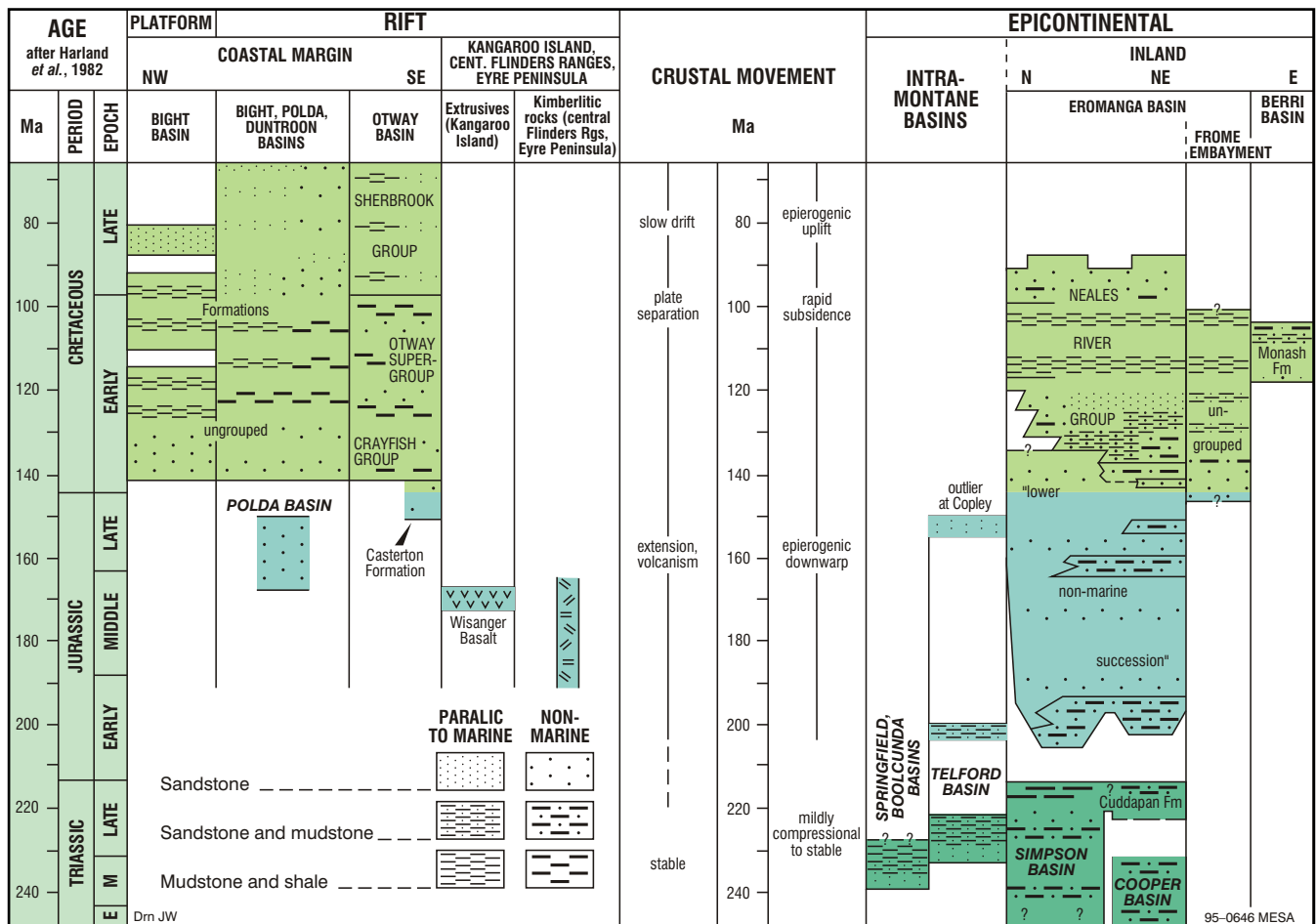


Fig. 9.2 Nomenclature and correlation of the Mesozoic basins in South Australia.

Basin and overlying Eromanga Basin. Definition of the Pedirka Basin by Wopfner (1972) to include only Permo-Carboniferous strata led to other proposals for placement of the Triassic rocks (Wiltshire, 1982a,b; Moore, 1982, 1986a; Powis, 1989). The Triassic succession was thus defined as the Simpson Desert Basin by Moore (1986a), but has been renamed in this volume as the Simpson Basin. The subsurface extent of the Triassic in South Australia, Northern Territory and western Queensland is still not well known but approaches 60 000 km²; the thickest deposits are in the Poolowanna Trough.

Structural setting

Episodic northeasterly tilting of the Cooper Basin region at the end of the Early Triassic and again in the early Late Triassic has been ascribed by Kuang (1985) to wrench-induced compressional reactivation of older fault systems. Triassic sediments west of the Birdsville Track Ridge were unaffected, reaching their greatest thickness in the Poolowanna Trough. During the Triassic, the western Pedirka Basin remained elevated while the eastern portion subsided to form the lowest sedimentary sequence within the developing Poolowanna Trough. There appears to have been little structural deformation in the Poolowanna Trough prior to Early Tertiary compressional deformation which created major, meridional, faulted anticlines (Moore and Pitt, 1984; Kuang, 1985).

Stratigraphy

Two Triassic formations have been defined by Moore (1986a), the older Walkandi Formation restricted to the Poolowanna Trough, and the overlying Peera Peera Formation which is more extensive (Fig. 9.4).

Walkandi Formation

The Walkandi Formation is a redbed unit which unconformably overlies either steeply dipping strata of presumed Ordovician age or flat-lying to gently dipping, possibly Devonian, sediments. It has been penetrated by exploration drillholes over the Poolowanna Trough and its maximum known thickness is 134 m in Poolowanna 1 (Fig. 9.5).

The unit consists of interbedded red-brown shale, green siltstone and fine-grained sandstone. Clay-filled cracks in mottled shale are regarded as desiccation and pedogenic features; oxidising continental redbed and shallow ephemeral lacustrine environments are proposed (Moore, 1986). Palynofloras indicate a Triassic age but are too poorly preserved for greater precision.

Peera Peera Formation

The Peera Peera Formation is a fine-grained carbonaceous unit which overlaps and rests conformably on Walkandi Formation. It extends westward onto the Early Permian Purni Formation of the Pedirka Basin, and to the east it overlaps the Birdsville Track Ridge (Fig. 9.6). A maximum thickness of 190 m has been recorded in Walkandi 1.

Three lithofacies have been distinguished in the deepest parts of the Poolowanna Trough (Moore, 1986a). Basal grey shale, siltstone, minor sandstone and coal are overlain by commonly upward-fining sandstone, while the upper unit is black, silty, highly carbonaceous shale. Moore (1986a) proposed meandering fluvial, flood plain and lacustrine environments. The Peera Peera Formation has been dated as Late Triassic.

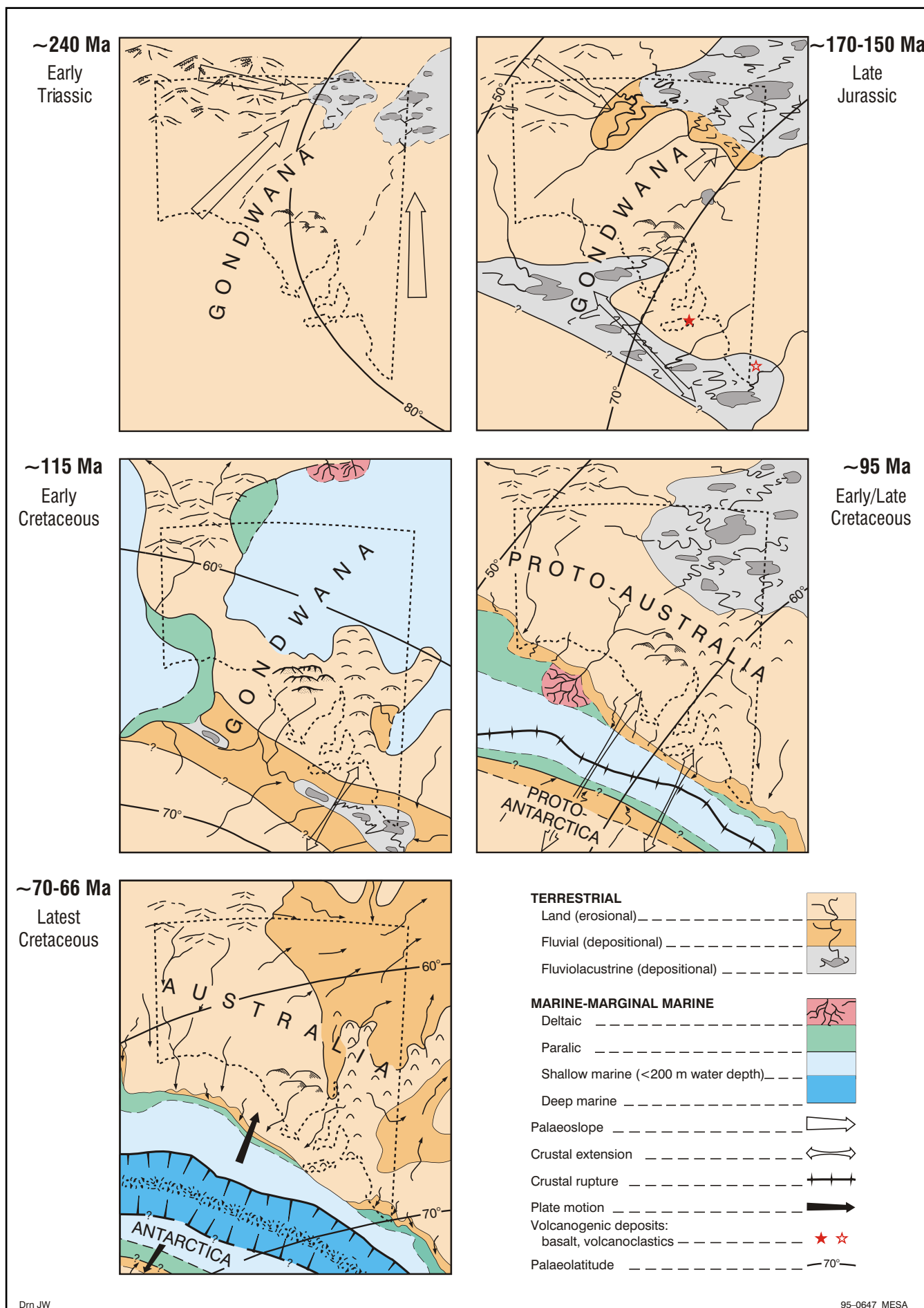


Fig. 9.3 Palaeogeographic change through the Mesozoic (after BMR Palaeogeographic Group, 1992; Frakes *et al*, 1987; Veevers, 1984).

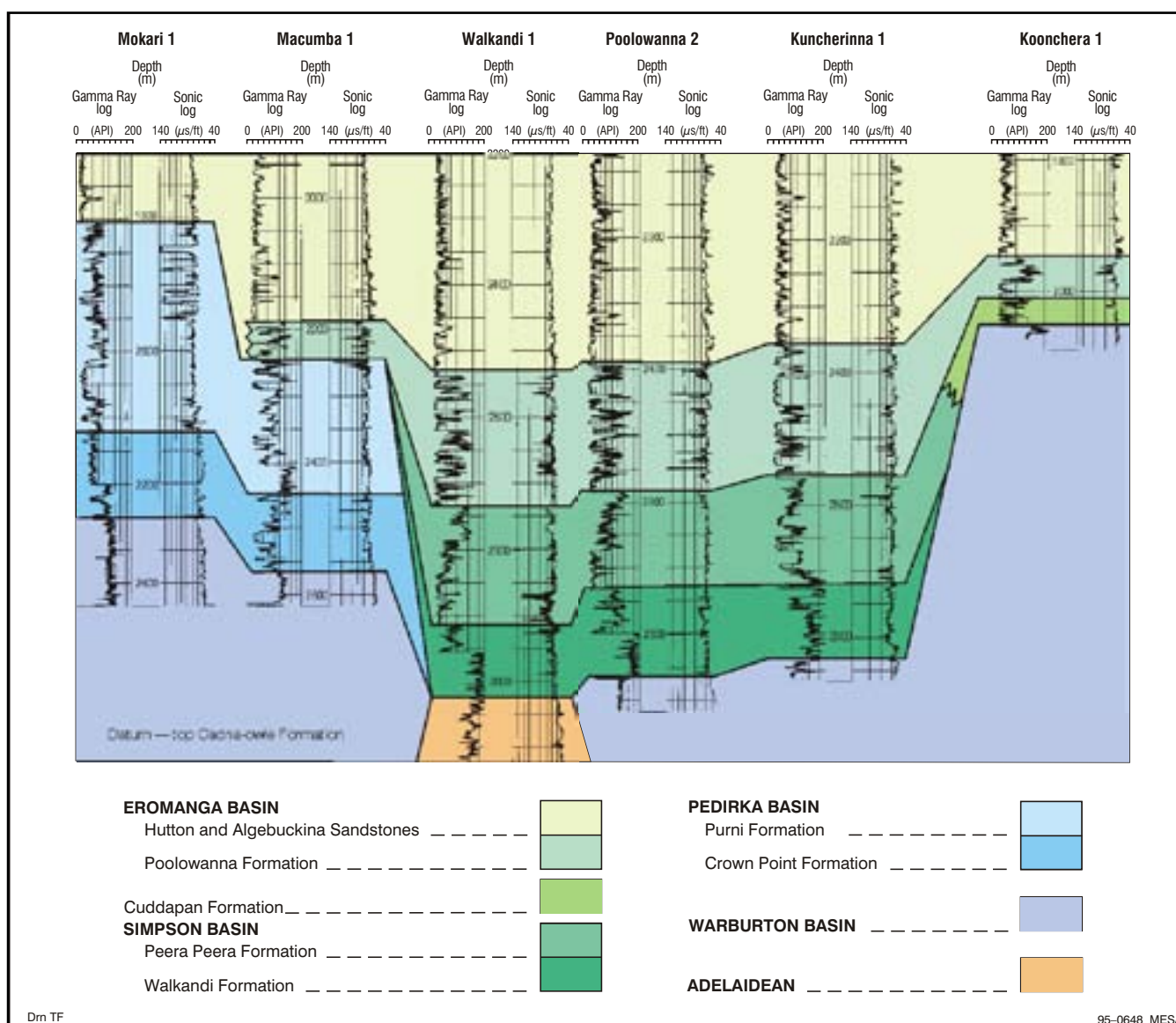


Fig. 9.4 Drill-log correlation, Simpson Region. Line of section is located on Figure 9.5.

Basin relationships

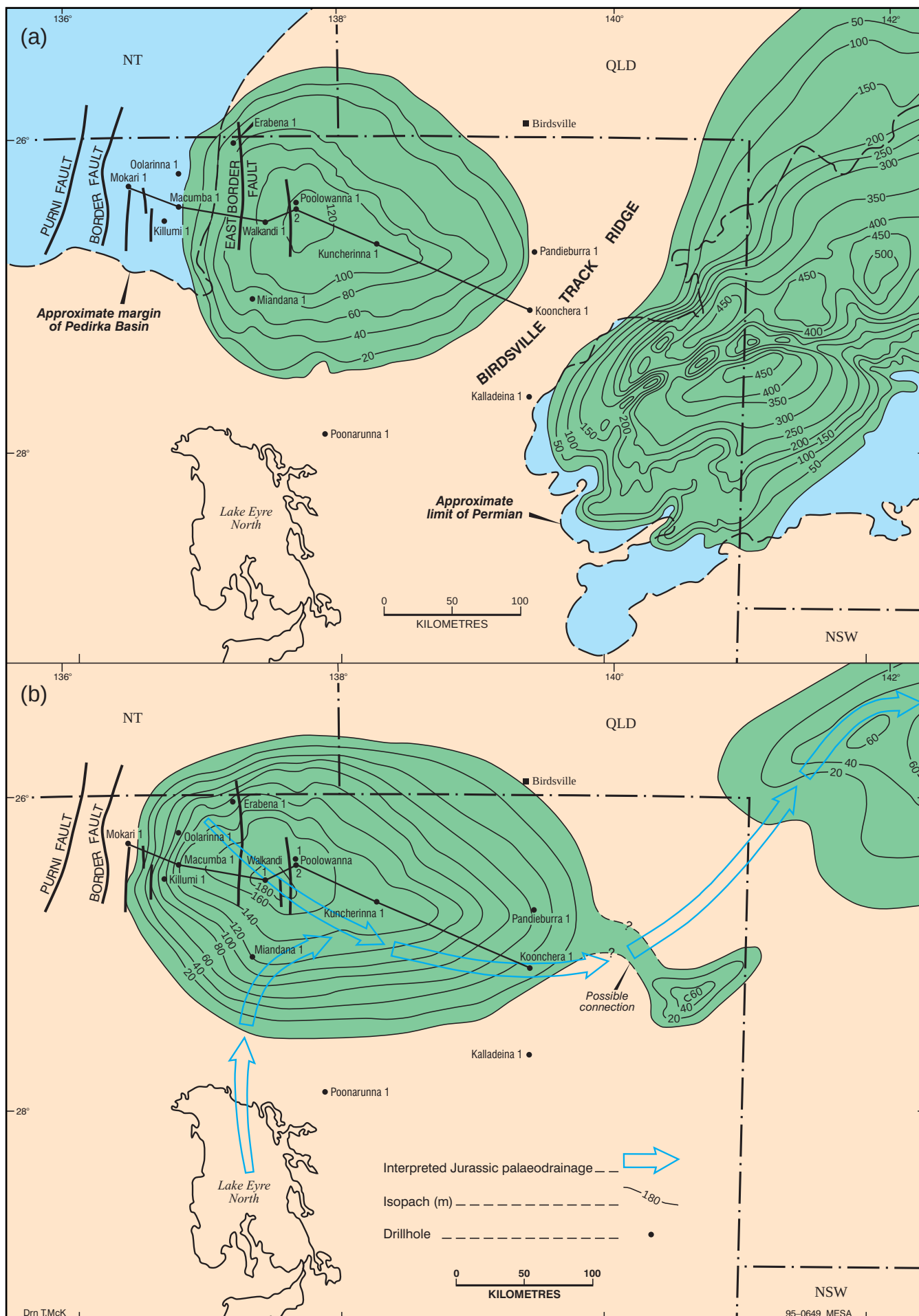
The Triassic stratigraphy of the northeastern region of the State differs east and west of the Birdsville Track Ridge. The apparent eastward and upward increase in areal extent of the Simpson Basin sequence contrasts with the Triassic east of the Birdsville Track Ridge in the Cooper Basin region. There the lower unit, the Late Permian-Middle Triassic Nappamerri Group, is far more widespread than the overlying, areally restricted Cuddapan Formation (Powis, 1989) which correlates with the Peera Peera Formation. Furthermore, the base of the Cuddapan is disconformable.

The differing stratigraphies have led to various basin nomenclatures as attempts to refine geological interpretation continue. Wiltshire (1982a,b), having noted the occurrence of isolated but widespread Upper Triassic erosional remnants in the region, including that of the Leigh Creek Coal Measures, placed the Upper Triassic in the overlying Eromanga Basin. An interpretation of Powis (1989) includes not only the Upper Triassic Cuddapan Formation but also the underlying Middle Triassic Tinchoo Formation (Nappamerri Group) in the Eromanga Basin, as both these formations were considered to be precursors to Eromanga Basin deposition. Palaeogeographic interpretations

further imply a genetic link between Triassic and subsequent depositional episodes. Powis (1989) recognised essentially fluvial environments in remnants of the Cuddapan Formation. Taken together with the coal-swamp facies of the Leigh Creek Coal Measures, this was interpreted by Wiltshire (1982a) as indicating a drainage connection to the Surat Basin during the Late Triassic, ancestral to that developed in the Early Jurassic (Fig. 9.5).

In this volume, regional unconformities are used to differentiate basins. Triassic deposits are thus excluded from the Eromanga Basin as the Upper Triassic units are everywhere separated by unconformity from the overlying Jurassic-Cretaceous succession. The Upper Triassic succession west of the Birdsville Track Ridge is here assigned to the Simpson Basin while that to the east (Cuddapan Formation) is not assigned to either the Cooper or Eromanga Basin and may prove to be an eastern extension of the Simpson Basin.

Fig. 9.5 (a) Isopach map for the Walkandi Formation and Nappamerri Group. (b) Isopach map for the Peera-Peera and Cuddapan Formations.



TRIASSIC INTRAMONTANE BASINS

G. Kwitko

A number of isolated structural depressions or intramontane basins in the Flinders Ranges contain Late Triassic, fluvialite and freshwater lacustrine deposits that formed concurrently with sedimentation in the Simpson Basin (Figs 9.1, 9.7, 9.8). As a result of tectonism, the sediments are now preserved in synclines developed within Adelaidean rocks in the Leigh Creek area, and 180 km to the south in the Springfield and Boolcunda Basins. All are small, asymmetrical, closed basinal structures 3-25 km² in area, with centripetal dips of 10-60°. Post-depositional tectonic movements have caused faulting, tilting and folding of the sediments, particularly along basin margins. The basins were eroded during the late Mesozoic and Cainozoic, and are now predominantly covered by a veneer of Quaternary alluvium. The occurrence of Triassic sediments in topographic depressions with only sporadic outcrop of weathered and ferruginous shale, sandstone and resistant ironstone contributed to their accidental or 'late' discovery.

TELFORD AND ASSOCIATED BASINS

The Leigh Creek Coalfield comprises four separate intramontane basins, the geology and economic aspects of which have been discussed by Dickinson (1944), Parkin (1953), Johns (1970, 1975a), Coats (1973b), Johns and Townsend (1975), and Springbett *et al.* (1995). From north to south they are designated North Field (Lobe D and Lobe C), Telford Basin (Lobe B) and Copley Basin (Lobe A), the largest of which is the northwest-trending, elliptical Telford Basin. MESA and ETSA drilling in late 1994 revealed another basin (Lobe E) to the southeast of Lobe A, and evaluation is continuing. The basins are aligned north-south over ~20 km, and their close proximity has led to dual nomenclature comprising basin name and 'lobe' identification (Fig. 9.8). The Late Triassic sequence designated as the Leigh Creek Coal Measures by Parkin (1953) occurs in all basins, ranging in thickness from 120 m in Lobe D to >1 000 m in the Telford Basin. Grey carbonaceous mudstone is dominant, with coal, sandstone and siltstone present in lesser and varying proportions.

Stratigraphy for the Telford Basin (Fig. 9.8) is based on Johns (1970) and Springbett *et al.* (1995). The coal measures rest with marked unconformity on folded and faulted Adelaidean rocks and consist of a thin basal grit overlain by an alternating sequence of grey carbonaceous mudstone and coal seams with interbedded siltstone and sandstone. The coal occurs in three distinct zones named informally, in ascending stratigraphic order, the 'Lower', 'Main' and 'Upper Series'. The 'Lower Series' contains two coal seams in a zone ~60 m thick separated from bedrock by up to 50 m of mudstone and a thin basal gravelly sandstone. The 'Main Series' is represented by a zone up to 20 m thick consisting predominantly of coal separated from the 'Lower Series' by 50-115 m of grey carbonaceous mudstone containing thin, hard sideritic siltstone interbeds locally termed 'hardbars'. The interval between the 'Main' and 'Upper Series' is composed of similar mudstone and varies considerably in thickness from 650 m in the northeast to <200 m in the southwest. The 'Upper Series' contains interbedded mudstone, siltstone and numerous coal seams with minor fine-grained sandstone, together attaining a maximum thickness of ~100 m. Thickness trends are similar to those in the 'Main Series', with the 'Upper Series' coal seams and clastic sediments thickening to the northeast. A predominantly sandstone sequence with some siltstone and mudstone and minor thin coal overlies the 'Upper Series' and is up to 150 m thick. Individual sandstone beds are 5-15 m thick and are generally saturated with saline groundwater.

The overlying Early Jurassic sequence is represented by ~200 m of interbedded mudstone, siltstone and fine-grained sandstone. The precise location and nature of the contact with underlying Triassic strata have not been established.

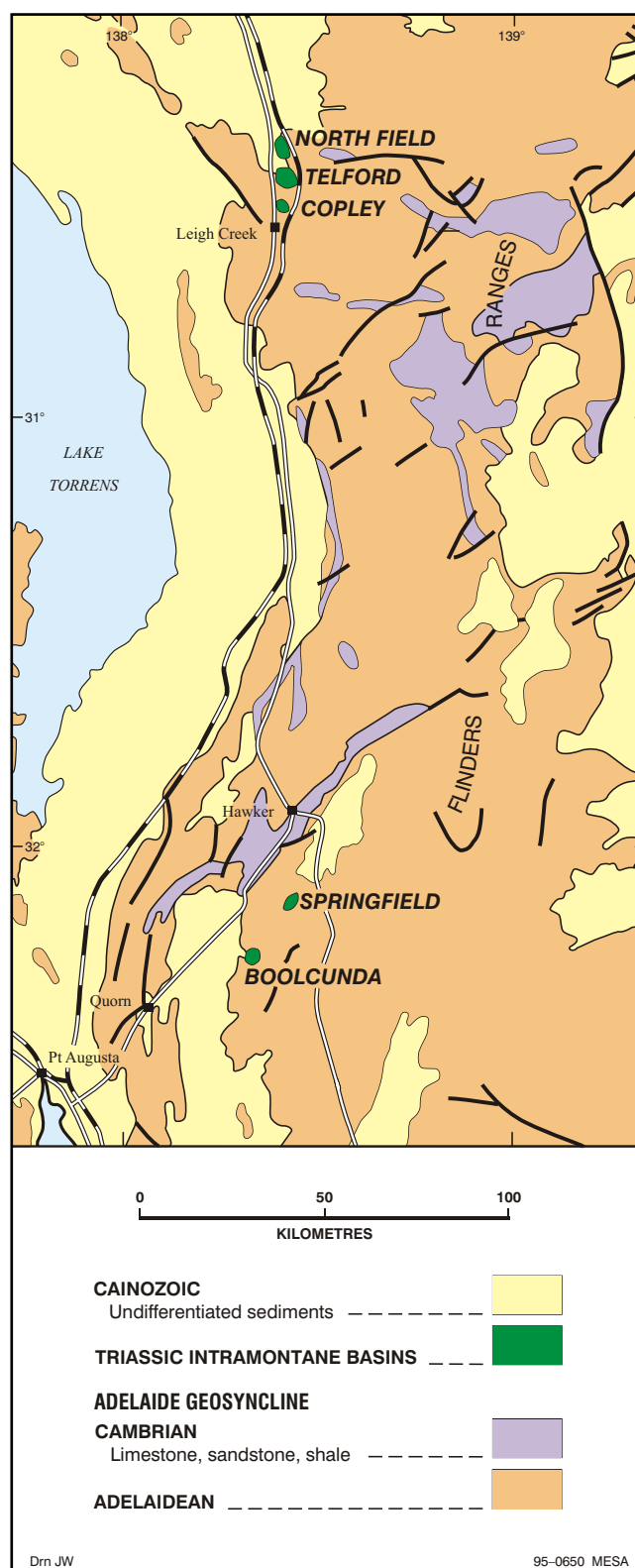


Fig. 9.6 Location and setting of Triassic intramontane basins.

The Triassic sediments contain numerous plant remains (dominantly the ferns *Thinnfeldia* and *Dicroidium*), freshwater mussels (*Unio eyrensis*) and rare fish (*Leighiscus hillsii*; Chapman and Cookson, 1926; Ludbrook, 1961b; Wade, 1953). The Triassic age of the measures was originally based on macrofaunal and macrofloral assemblages identified by Chapman and Cookson (1926). Subsequent microfloral investigations by Playford and Dettmann (1965) and Hos (1977, 1978) were principally concerned with the upper, non-coal-bearing part of the succession, and these studies determined a Rhaetian-Liassic age. A recent study by

AGE			SPORE-POLLEN ZONES				LEIGH CREEK AREA	CENTRAL FLINDERS RANGES	SIMPSON REGION	COOPER REGION	
after Harland <i>et al.</i> , 1982			Helby <i>et al.</i> , 1987		Price <i>et al.</i> , 1985						
Ma	JURASSIC			<i>Corollina torosa</i>			Kimmeridgian and Sinemurian components not shown				
210		HETTANGIAN		<i>Polycingulatisporites crenulatus</i>	PT5	5.2					
		213				5.1					
220	LATE	RHAETIAN			?						
			219								
		NORIAN			4.2						
			225	<i>Craterisporites rotundus</i>	PT4	4.1					
230	TRIASSIC	CARNIAN					COPLEY BASIN (Lobe A)	TELFORD BASIN (Lobe B)	NORTH FIELD (Lobes C,D)	SIMPSON	
			231	<i>Samaropollenites spectosus</i>		?					
		LADINIAN									
			238	<i>Samaropollenites quadrifidus</i>	PT3						
			ANISIAN	<i>Aratrisporites parvispinosus</i>							
240	MIDDLE		<i>Aratrisporites tenuispinosus</i>	PT2	2.2		SPRINGFIELD AND BOOLCUNDA BASINS	Wakandi Formation	COOPER BASIN		
			<i>Protohaploxypinus samoilovichii</i>		2.1						
			<i>Lunatisporites pellucidus</i>	PT1							
			EARLY	SCYTHIAN							
250	PRE-MESOZOIC										
Dm RF											

95-0651 MESA

95-0651 MESA

Fig. 9.7 Correlation of Triassic Basins.

Parker (1987) reviewed and extended the earlier work and proposed a subdivision of the Telford Basin succession into biostratigraphic stages spanning Carnian (Late Triassic) to Sinemurian (Early Jurassic) with the coal measures in the Carnian and Norian. In the Leigh Creek Coalfield the Early Jurassic sediments are restricted to the central part of the Telford Basin.

Correlations between coal measures in the various basins have been proposed by Townsend (1979) and Parker (1987). On the basis of palynofloral evidence, Parker correlated the section at North Field with the lower portion of the Telford Basin sequence, whilst the Copley Basin was correlated with the upper portion of the Telford Basin sequence. This evidence supports an earlier concept (Parkin, 1953) that the separate basins represent remnants of a more widespread but not necessarily very extensive sequence.

Sedimentological and palaeontological aspects indicate deposition in and around freshwater lakes and river systems in a mainly lacustrine environment with fluvial influx, under conditions of sluggish drainage and subdued topography. A warm and moist climate during coal formation is indicated by fossil fern types. Sedimentation in these basins was brought to a close by uplift and erosion during the Middle to Late Jurassic.

Structural studies of the Leigh Creek Coalfield, in particular of the Telford Basin, show that the flexure-fault tectonic boundaries to the south and west of Telford Basin seem to be connected with synsedimentary basement fault movements (Bogacz and Kivior, 1988; Bogacz, 1989). However, the northern and northeastern

Top: Open-cut operations at Lobe B, Telford Basin. In the middle distance is the original site of Leigh Creek township, now used as a garage facility by the Electricity Trust of South Australia. Photo courtesy of ETSA. Photo 042461.

Bottom: Plant fragments (*Dicroidium odontopteroides*) in white claystone from Leigh Creek Coalfield. Photo 042462.



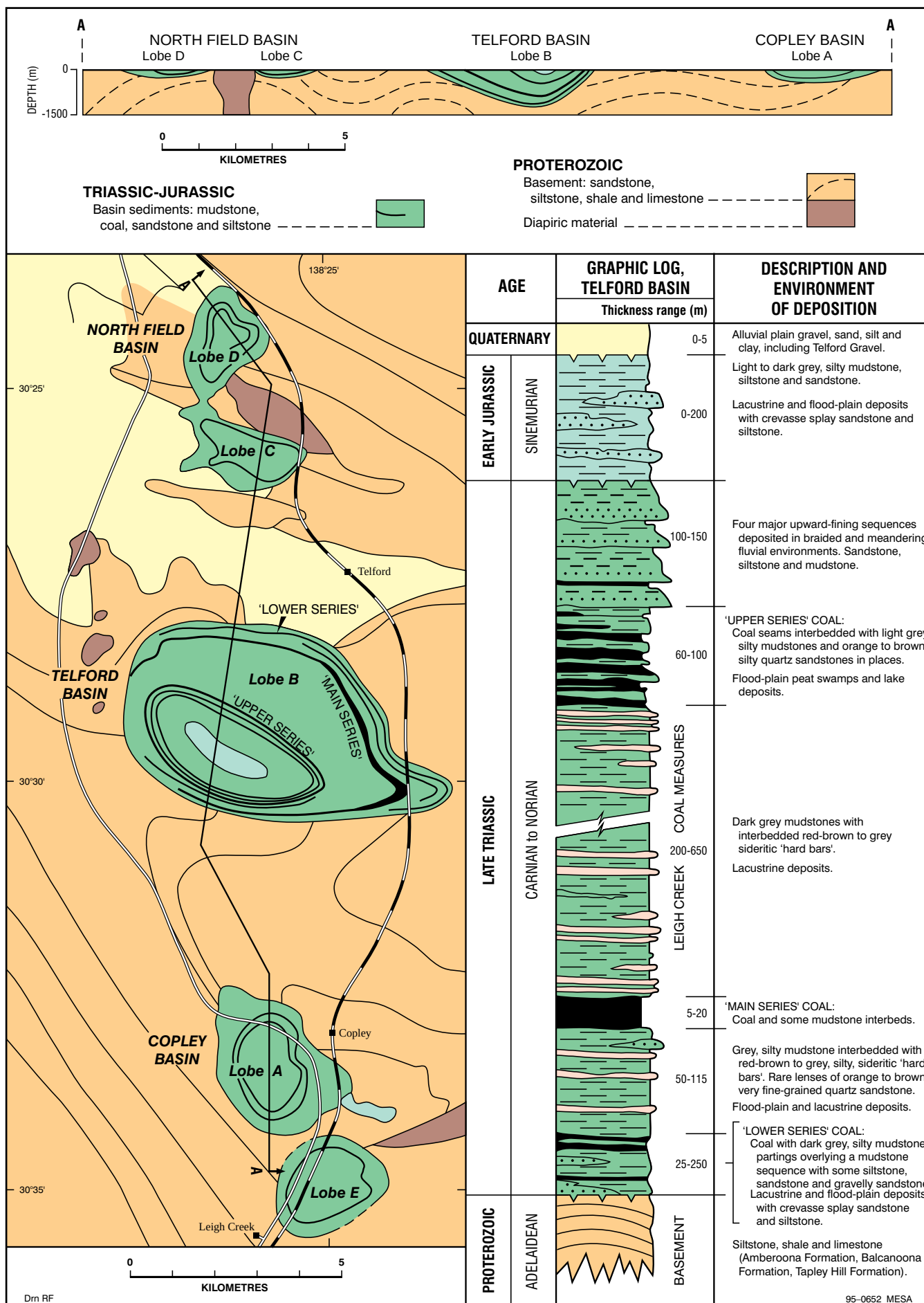


Fig. 9.8 Geology of the Leigh Creek coalfield (after Johns, 1970; Springbett et al., 1995).

basin margins exhibit primary, non-tectonic features, and only later peripheral faulting has given rise to structural deformation. These peripheral faults are characterised by steep dips (80-90°) and generally follow the elliptical shape of the basin. Fault throws throughout the basin are generally 0.5-10 m but may attain 35 m. Tectonic structures seen to be more prevalent at the basin margins than centres and decrease up the stratigraphic sequence. Bogacz (1989) confirmed that the preservation and structure of the basins resulted from local structural movements predominantly controlled by a gravitational normal stress field and the vertical movements of basement faults. Reactivation of faults has long been suggested as the driving force for basin formation although more recently a diapiric mechanism for basin subsidence has been postulated (N.M. Lemon in Springbett *et al.*, in press).

Economic geology

All basins in the Leigh Creek Coalfield contain coal seams. Reserves in Lobe D have been mined out while thin seams of poor quality coal remain in Lobe C. Open-cut mining in the Telford Basin (Lobe B) took place between 1944 and 1958, and resumed in 1976 after depletion of North Field (Lobes C and D). Coal from Leigh Creek is railed to the Playford and Northern Power Stations at Port Augusta. The economic aspects of these deposits are detailed in Parkin (1953), Johns (1970, 1975) and Springbett *et al.* (1995). Coal reserves within Lobe B are estimated to exceed 500 Mt of which ~100 Mt are considered economically recoverable by current mining methods, indicating a 25-30 year life for the deposit.

Economic coal seams occur in the 'Lower', 'Main' and 'Upper Series' (Fig. 9.9). The 'Upper' and 'Lower Series' contain numerous seams up to 5 m thick which are interbedded with mudstone and siltstone, the 'Lower Series' seams being the poorest and characterised by a number of thin and relatively high-ash coal seams. The 'Main Series' is essentially a single seam which varies in thickness from 6 to 18 m and contains some thin lenticular mudstone partings. All the seams are characterised by moderate dips (10-30°), variable seam thickness and seam splitting. The coal is low rank and classified as Sub-bituminous C (American Society for Testing and Materials) with typical mined coal containing ~30% moisture, 15-20% ash, 0.4% sulphur and heat values of up to 15 MJ/kg.

SPRINGFIELD BASIN

Triassic rocks overlie diapiric Callanna Group other Adelaidean units in the Springfield Basin. The sediments have been infolded and infaulted with the Adelaidean rocks into an asymmetric, 9 km² basin with a curved, northeast-trending axis. Sediments dip at 20-60° towards the centre of the basin, the steeper dips being adjacent to faulted margins. The succession, interpreted by Johnson (1960a) as Triassic age, attains a maximum thickness of ~600 m and consists of conglomerate, carbonaceous mudstone, siltstone and sandstone with thin coal seams. Such a thickness in so small a basin suggests a much larger former extent.

The succession begins in most parts of the basin with a conglomerate up to 180 m thick, probably deposited in a piedmont environment (Johnson, 1960a). Along the poorly defined eastern part of the basin the lowest unit appears to consist of grey to yellow shale and argillite with carbonaceous bands. Approximately 100 m of red argillite and sandstone overlie the conglomerate, which in turn are overlain by coal measures up to 250 m thick. These comprise carbonaceous mudstone with thin coal seams and minor fine-grained sandstone.

The top unit, preserved only in a single mesa (Fossil Hill) in the centre of the basin, is an argillite containing abundant leaf impressions and casts of freshwater molluscs including *Unio springfieldensis*, *Protovirgus jaenschii* and *Unio eyrensis* of the family Unionidae (Ludbrook in Johnson, 1960a; Ludbrook, 1961b). The leaf impressions were identified by Ludbrook (1961b) as *Dicroidium feistmanteli*. Further investigation of the macroflora by Amtsberg (1969) revealed similarities with Leigh Creek flora and determined a Late Triassic age for the sediments.

The Upper Triassic unit represented at Fossil Hill has been described by Johnson (1960a) as a 'baked argillite' because of its bleached and baked appearance and the presence of reddened shale, porcellanite and pseudo-igneous rock fragments. A plausible explanation for this phenomenon is heating by the subterranean combustion of coal seams as suggested by Johnson and Bucknell (1958). Similar phenomena have been recorded at North Field in the Leigh Creek Coalfield (Baker, 1952).

The coal potential of the Springfield Basin was investigated by drilling in the late 1950s (Johnson, 1960a) but the deposit was regarded as uneconomic. Stratigraphic and exploration drilling discovered coal of low grade, sub-bituminous rank, with high ash and water content. Coal seams seldom exceed 0.3 m; the thickest seam, 3.6 m at a depth of 77 m, has an ash content of ~30% and does not persist over sufficient area to be of interest.

Small alluvial diamonds have been disclosed in boulder horizons within the basal conglomerate. Exploration drilling intersected thin coal seams within the conglomerate. Exploration drilling intersected thin Late Permian coal seams within the conglomerate.

BOOLCUNDA BASIN

The Boolcunda Basin, 15 km south-southwest of the Springfield Basin, is 2.4x1.6 km, elongated in a northeast-southwest direction, and contains a Triassic sequence infolded and faulted with the underlying Adelaidean rocks. Investigation by von der Borch (1960) indicated a maximum thickness of Triassic sediments of ~300 m, with a thin basal conglomerate in the southeastern part of the basin. Elsewhere, basal sandstone and mudstone are succeeded by a carbonaceous mudstone-coal sequence, followed by mudstone and siltstone. Palynofloras from the coal were determined to be of Triassic age. The sediments dip steeply (up to 60° around the margins) towards the centre of the basin. The coal was investigated by drilling but seams proved to be thin, discontinuous and of poor quality.

JURASSIC-CRETACEOUS EPICRATONIC BASINS

EROMANGA BASIN

G.W. Krieg, E.M. Alexander and P.A. Rogers

The Eromanga Basin is the largest and most central of three epicontinental depressions that together comprise the Great Artesian Basin, a non-marine and marine Jurassic-Cretaceous superbasin which covers approximately one-fifth of the Australian continent (Fig. 9.9). Although the term 'Great Artesian Basin' has obvious hydrological connotations, it is entrenched in geological literature (e.g. Whitehouse, 1955) and is useful as a geological term even though the artesian and sedimentary limits are not coincident. This duality of meaning is widely understood; retention of the term obviates the introduction of a new name and acknowledges the historical and economic importance of this superbasin. Useful reviews of basin nomenclature and definition are given by Habermehl (1980, 1986) and Wiltshire (1982a).

Only the southwestern third of the Eromanga Basin occurs in South Australia. It extends as an almost continuous blanket over the northeast of the State from where it laps onto the older basement blocks and basins to the west and southwest. It partly onlaps and is partly in fault contact with sharply elevated Adelaidean rocks of the northern Flinders Ranges and the Peake and Denison Inliers. The southwestern extremity of the basin may have been connected to the Bight Basin. East of the northern Flinders Ranges, a small lobe extending beneath Lake Frome is known as the Frome Embayment (Fig. 9.1). The now-separate Berri Basin, underlying the Tertiary Murray Basin, is regarded as a former extension of the Eromanga Basin.

Structurally, the Eromanga Basin in South Australia is divided

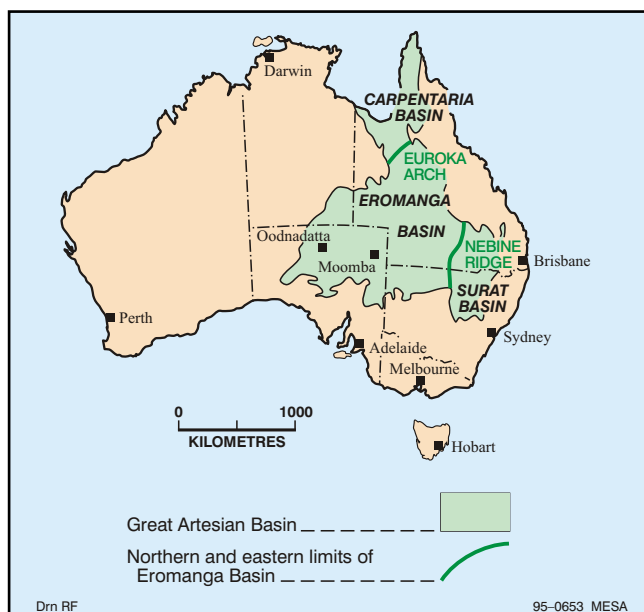


Fig. 9.9 The Great Artesian Basin showing component Eromanga, Surat and Carpentaria Basins.

into northeastern and southwestern parts by a northwest-trending zone, here referred to informally as the Denison-Willouran divide (Fig. 9.10). The sedimentary succession thickens rapidly eastwards across this divide, attaining 3 000 m in the northeast of the State, within the basin depocentres. West and southwest of the divide, the contiguous basin succession is much thinner because of depositional thinning and erosional stripping of the higher stratigraphic units (Fig. 9.11). Transecting the thicker part of the basin another major structural zone extends northeast through Haddon Corner into Queensland. The broadest and westernmost structure in this zone is the Birdsville Track Ridge, expressed as a line of gentle surface anticlines which separates a western region overlying the Simpson Basin from an eastern region overlying the Cooper Basin. For convenience, these regions are herein referred to informally as the Simpson region and Cooper region, respectively. The Cooper region, itself composed of parallel southwest-northeast arcs of fault-bounded domes and troughs, is the focus for hydrocarbon exploration and production.

The sedimentary record of the Eromanga Basin extends without major break from early Jurassic to mid-Cretaceous (Figs 9.2, 9.12) and consists of a cyclic succession of sandstone and mudstone units thought to reflect oscillating sea level (Burger, 1986). During this time, terrestrial freshwater conditions prevailed, except for an Early Cretaceous marine transgressive phase in the latter part of the depositional period. The resulting sedimentary succession is thus composed of lower and upper terrestrial sequences and an intervening marine sequence. Much of the lower sequence is a medium to moderately high-energy, cross-bedded fluvial sandstone, whilst the upper is mainly a lower energy, finer grained, flood-plain and lacustrine sandstone, siltstone and coal-swamp deposit. A ten-fold increase in average rate of accumulation of the upper compared to the lower sequence represents a contrast between steady basin subsidence (the upper) and deposition with frequent interruption (the lower). The intervening marine succession, recording the early stages of a global sea-level rise (Morgan, 1980a), is a grey, fossiliferous mudstone-siltstone sequence with regressive marine sandstone intervals and a basal, non-marine to marginal marine sandstone-siltstone unit. Braided fluvial sandstone units of the lower sequence form the principal aquifers and contain economically significant oil reserves sourced from the underlying Cooper Basin and from *in situ* strata. The marine shale sequence, particularly the muddy lower part, forms the aquitard to the artesian groundwater. Coal in the upper sequence forms a considerable resource.

The stratigraphy and nomenclature for the Eromanga Basin is derived essentially from Forbes (1966), Freytag (1966), Nugent (1969), Wopfner *et al.* (1970) and Moore (1986b). Different stratigraphic and nomenclatural schemes have been developed by various investigators working in widely separated parts of the basin. A regional integration of these schemes was devised by Wopfner (1969c) and this was further generalised and developed by Moore and Pitt (1982), following studies of petroleum drillholes in the Poolowanna Trough. In this volume a synthesis accommodating both the exposed marginal stratigraphy and that of the Simpson and Cooper regions derived from petroleum exploration is used (Fig. 9.12). The palaeontology incorporated in this scheme represents a vast body of data collected from over 800 exploration drillholes and numerous outcrop samples.

The only record of events possibly attributable to the remainder of the Cretaceous is the regional deep weathering profile below early Tertiary cover. There is evidence for a Maastrichtian component to this weathering event (Senior and Mabbutt, 1979) although Wopfner (1978) preferred a younger age for the main development. The profile, which forms a zone up to 40 m thick, comprises conspicuous, pale-coloured bluffs of thoroughly kaolinised Mesozoic sediments in the tableland regions of the basin, and represents a time of stable tectonic conditions when chemically active groundwater altered near-surface strata. The Late Cretaceous in South Australia was also a time of erosion when detritus was transported to the developing Australian Southern Rift System.

Basin development

After general uplift in the Triassic involving local movements and changing depocentres, South Australia, by the beginning of the Jurassic, had become elevated and erosional. The crustal events that followed caused a continental downwarp or syncline (Wopfner, 1985), termed the Great Artesian Basin and its component Eromanga Basin where contiguous deposition resumed over a very wide area.

The mechanism that initiated and developed the syncline is poorly understood although a state of overall distension may have existed (Wopfner, 1985). Attempts to explain the stress field (Veevers, 1984, 1987; Austin and Williams, 1978; Middleton, 1980; Finlayson, 1983; Wopfner, 1985; Zhou, 1989) have involved:

- the transmission of extensional, collisional or rotational plate boundary stresses to the plate interior
- changes of mantle heat flux beneath the plate interior with cooling and thermal contraction, or heating with phase changes and volume reduction
- eustatic loading
- isostatic adjustments
- various combinations of the above.

To explain synchronicity of basin subsidence on different continents, Sloss and Speed (1974) postulated episodic changes in the proportion of sub-crustal melt through global heat flux variation. Subsidence history of the Eromanga Basin, derived from sediment accumulation rates in a number of exploration drillholes (e.g. Moore and Pitt, 1984), shows initial steady subsidence followed by very rapid subsidence from about mid-Aptian to Cenomanian and then a cessation of subsidence when the basin filled. The 'collapse' of the basin floor at about 100 Ma may have been triggered by tectonic compression, as rifting with stress release replaced lithospheric stretching along the southern and eastern plate margins (Zhou, 1989).

Whatever the mechanism for initial formation and shaping of the basin, the subsequent tectonic development and eventual structural design of the basin were controlled by two major factors — the inherited pattern of pre-Jurassic structures and post-depositional (mainly Tertiary) continental compression. By comparison, syndepositional structuring was minor.

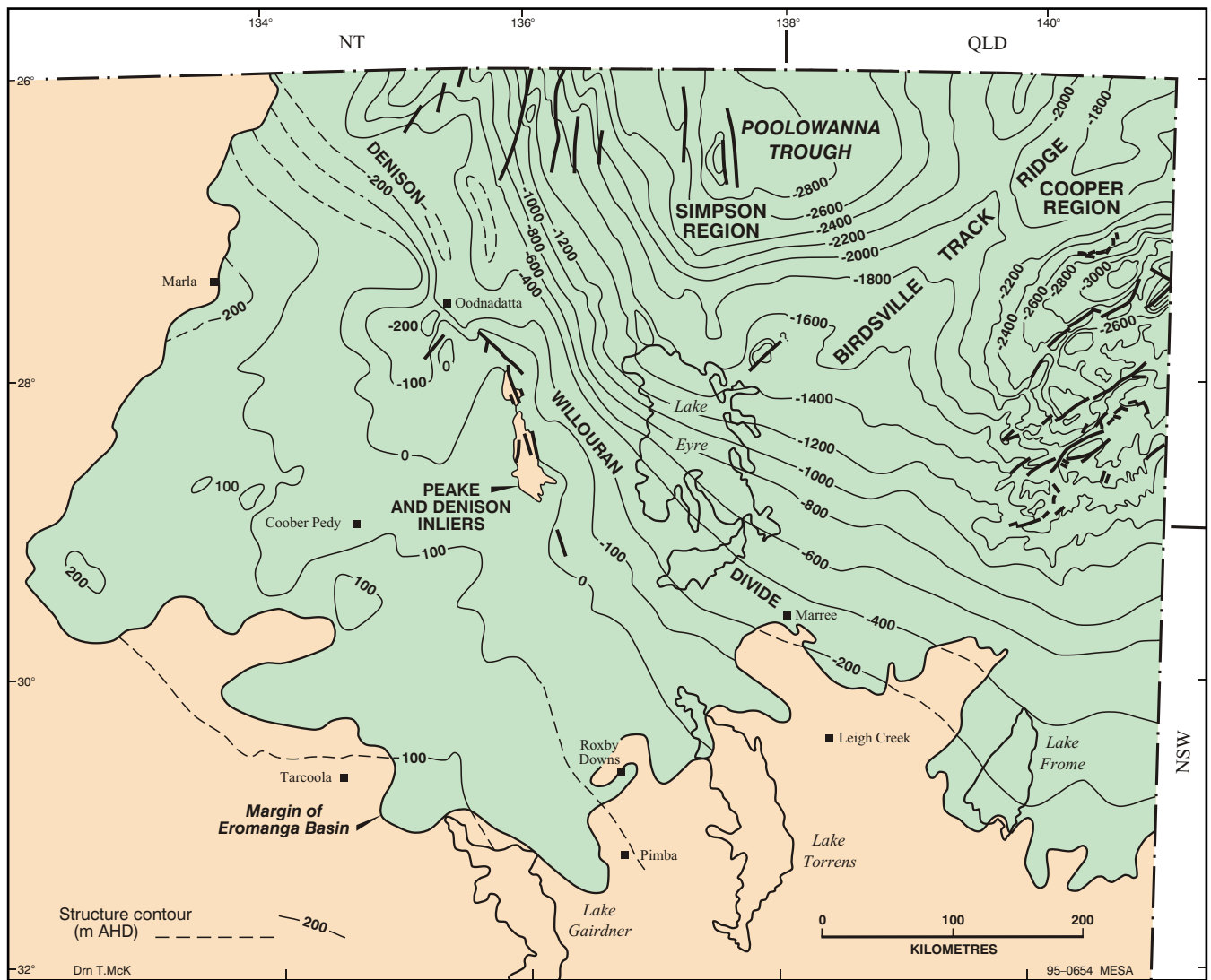


Fig. 9.10 Structure map of the base of the Mesozoic.

Pre-Jurassic orogenic folding and faulting, development of basins, and epeirogenic movements created many separate geological provinces and structural elements which form the basement of the Eromanga Basin, each event imposing its own structural grain. In South Australia, components of Palaeoproterozoic (Kimban) tectonism on the Gawler Craton and Cambro-Ordovician (Delamerian) folding northwest of the Flinders Ranges are dominant structural controls. Effects of the mid-Carboniferous Alice Springs Orogeny in the centre of the continent extended into the Warburton Basin in the north of the State. Other important effects may include Adelaidean tectonism along the Torrens Hinge Zone (Wopfner, 1985), possibly compressive Permian structuring in the northeast of the State (e.g. Kuang, 1985) and widespread post-Early Permian (post-glacial) isostatic adjustment following ice melting.

By the beginning of the Jurassic this tectonic history had produced a pre-Eromanga terrain of very varied topography constructed as a mosaic of quasi-rigid blocks which could move differentially along the lines of dislocation that separated them (Anfilloff, 1982; Wopfner, 1985; Hoffmann, 1989). Under post-depositional stress fields the crust yielded preferentially according to stress direction, along certain of these pre-existing lines of weakness. This transferred the linear structural network of the basin floor into the overlying sediments and, with the associated local folding effects, established the structural trends and styles of the Eromanga Basin. The much lesser effects of syndepositional structuring (Jurassic to Early Cretaceous) are documented by drape and compaction folds

over Permian-Triassic highs, the subdued nature of growth faulting (Fairburn, 1989), and the remarkable uniformity of thickness and lateral persistence of the various stratigraphic units (Wiltshire, 1989).

Present basin geometry is reflected in the regional lineament pattern (Fig. 9.13) expressed by basin outline, outcrop patterns, surface lineaments, structure contour trends, and gravity and magnetic features. A rhombic pattern is clearly evident with dominant northeast and northwest trends and subsidiary sub-meridional to sub-latitudinal trends. Some lineament sets broadly align with the continent-wide structural corridors G2, G3 and G8 (Campbell and O'Driscoll, 1989), and reflect continent-scale linear zones of crustal weakness where flexuring and faulting have recurred.

Particular attention is drawn to the Lake Eyre Lineament and Muloorina Gravity Ridge, which represent a northwest-trending, pre-Warburton Basin structure (Wopfner, 1972, 1985) that acted as a hinge line during Eromanga deposition, dividing a southwestern shelf environment from a northeastern basinal environment. Main features of the southwestern region are:

- On the Gawler Craton, Mesozoic deposits are bounded by regionally linear, northeast and northwest-trending outcrop terminations, and are disrupted by late Cainozoic reactivation of the ancient, northeast-trending Karari Fault Zone (Benbow, 1986; Rankin *et al.*, 1989) and by the northwest-trending fault-bounded Mulgathing Trough (Daly, 1985).

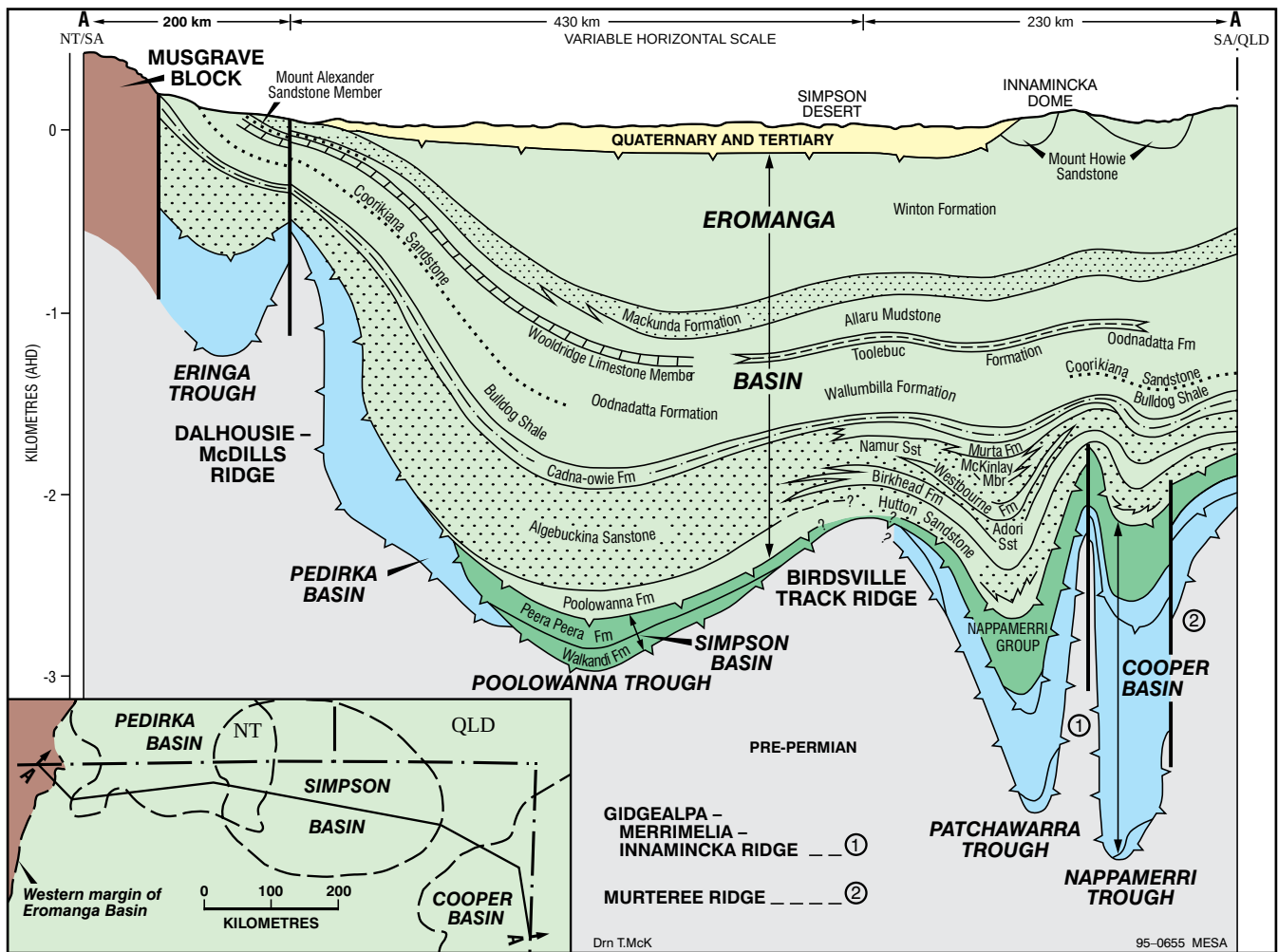


Fig. 9.11 Cross-section through the Eromanga Basin.

- Very broad, low-amplitude folding, probably reflecting the Late Cretaceous and Tertiary compression, formed regional synclines along which Cainozoic trunkline rivers flowed (Tallaringa and Mirackina Palaeochannels of Barnes and Pitt, 1976). The Stuart Range between the two channels lies on a basement upwarp.
- The Norwest Fault and related parallel faults are Delamerian, and possibly older, northwest-trending structures. With the associated Delamerian fold trends of the Willouran Ranges area, and like the Gawler Craton fault pattern, they produced a structural grain in the Mesozoic sediments by controlling sedimentation patterns. These patterns are reflected by palaeocurrent directions and channel alignments of fluvial deposits. The fault system was reactivated in Cainozoic times, disrupting Mesozoic sediments and forming a partial barrier to the flow of artesian water.

Main features of the northeastern region are:

- The Birdsville Track Ridge is a major northeast-trending concealed basement high. It is expressed in the Cooryanna and Gason Domes where uppermost Eromanga deposits and Tertiary units are exposed as low, broad surface anticlines rising above the surrounding dunefields. The feature divides the Eromanga Basin into Cooper and Simpson regions where structural patterns and sedimentary history differ. The Birdsville Track Ridge influenced deposition of the Poolowanna Formation, and Hutton and Algebuckina Sandstones, but had little influence on deposition of the Birkhead Formation and younger Eromanga Basin units.
- The structure of the Cooper region consists of a northeast-trending arcuate set of alternating troughs and ridges. One of the ridges, the Gidgealpa-Merrimelia-Innamincka Ridge, is expressed at the surface by the Innamincka Dome. Right-lateral fault offsets, producing an *en echelon* pattern of anticlines, have been observed along this trend. Major northwest to west-northwest structures appear to truncate the arcuate fold pattern; in the south, the Lake Blanche Lineament is expressed as a pronounced physiographic feature parallel to the Lake Eyre Lineament, and in the north is the Arrabury-Naccowlah Structure. The relatively complex Eromanga Basin structure in this region strongly reflects a pre-existing pattern established particularly in the Permian and Triassic, and reactivated in the Tertiary. Structural contour patterns of various horizons such as the 'Z' (pre-Permian unconformity), 'P' (near top of the Permian) and 'C' (near top of the Cadna-owie Formation) seismic horizons are thus broadly coincident in plan and show increasing relief with depth.
- The structure of the Simpson region (Moore and Pitt, 1984; Moore, 1986b) consists of an array of sub-meridional faults and associated elongate anticlines which are commonly steeper and fault bounded on their western flanks. Major faults step down the Eromanga Basin strata from the west and east into the Poolowanna Trough with vertical displacements of 100-150 m on the base of the Jurassic. In the west the Dalhousie-McDills Ridge is a major pre-Mesozoic structure which was active particularly during Late Permian and Tertiary compression (Alexander, 1990; Moore and Pitt, 1984) while, in the east, structuring occurred mainly during the mid-Tertiary (Moore, 1986b).

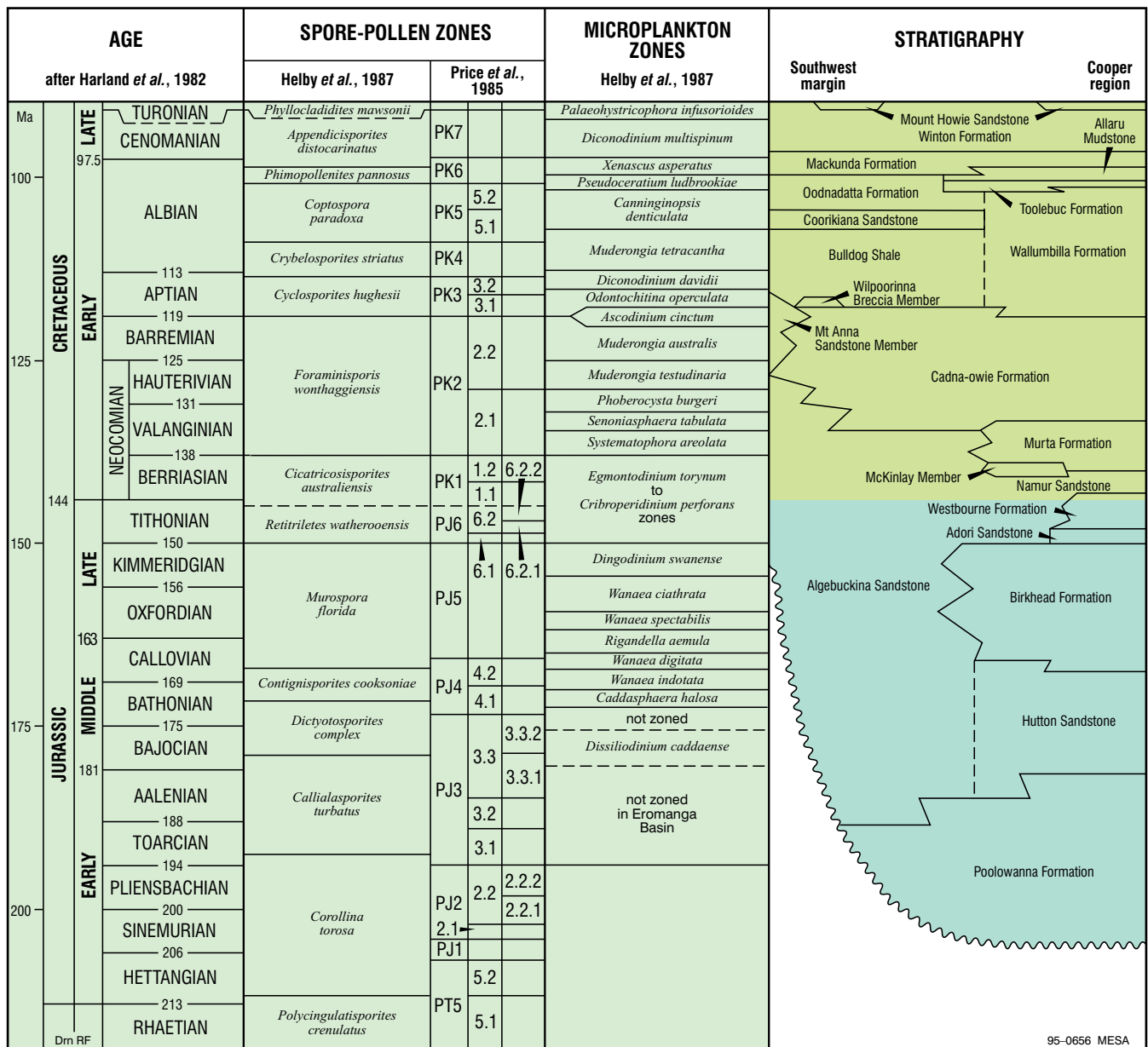


Fig. 9.12 Stratigraphic and palynological terminology, southwestern Eromanga Basin.

Stratigraphy — lower non-marine succession

E.M. Alexander and G.W. Krieg

Regional weathering and regolith development occurred at least around the margins of the Eromanga Basin prior to sedimentation. This is expressed as deep kaolinisation underlying Jurassic-Cretaceous deposits (Wopfner *et al.*, 1970; Krieg *et al.*, 1991) and as remnants of an ancient erosional surface and waste mantle accumulation beneath Cretaceous deposits (Rogers and Freeman, 1994; Krieg *et al.*, 1991). A pre-Jurassic and post-Early Permian age is tentatively suggested for this event.

Sedimentation began with a terrestrial succession up to 1 000 m thick spanning ~65 million years from Early Jurassic to earliest Cretaceous (Fig. 9.11, 9.12). Large fluvial systems drained centripetally (Wiltshire, 1989) into a central lowland of lakes and swamps, through a landscape vegetated by conifers, ferns and cycads. In the Cooper region, the succession consists of alternating mudstone and sandstone units. To the west and southwest, the mudstone pinches out and the section gradually thins, merging into a single unit, the Algebuckina Sandstone, which is exposed extensively around the basin margin. The lower non-marine part of

the Eromanga Basin forms the main target for oil exploration in the region, and sandstones form the Great Artesian aquifers.

Poolowanna Formation

The Poolowanna Formation (Moore, 1986a) occurs entirely in the subsurface in both the Poolowanna Trough and Cooper region (Fig. 9.14). The unit has previously been termed 'basal Hutton (Sandstone)' (e.g. Barr and Youngs, 1981), 'basal Jurassic' and 'Cuttapirrie beds' in the Cooper region. The primary depocentre is in the Poolowanna Trough west of the Birdsville Track Ridge (Fig. 9.15), where the maximum thickness intersected, through the type section in Poolowanna 1, is 206 m. The formation lies unconformably on Triassic sediments of the Simpson Basin and Permo-Triassic sediments of the Cooper Basin. In the Poolowanna Trough it intertongues with and is overlain by the Algebuckina Sandstone. Towards the basin margin the Algebuckina Sandstone overlaps the Poolowanna Formation. In the Cooper region, the Poolowanna Formation intertongues with and is overlain by the Hutton Sandstone. Distribution of the formation is controlled by facies changes into the Hutton Sandstone and by palaeotopographical features on the Triassic unconformity surface.

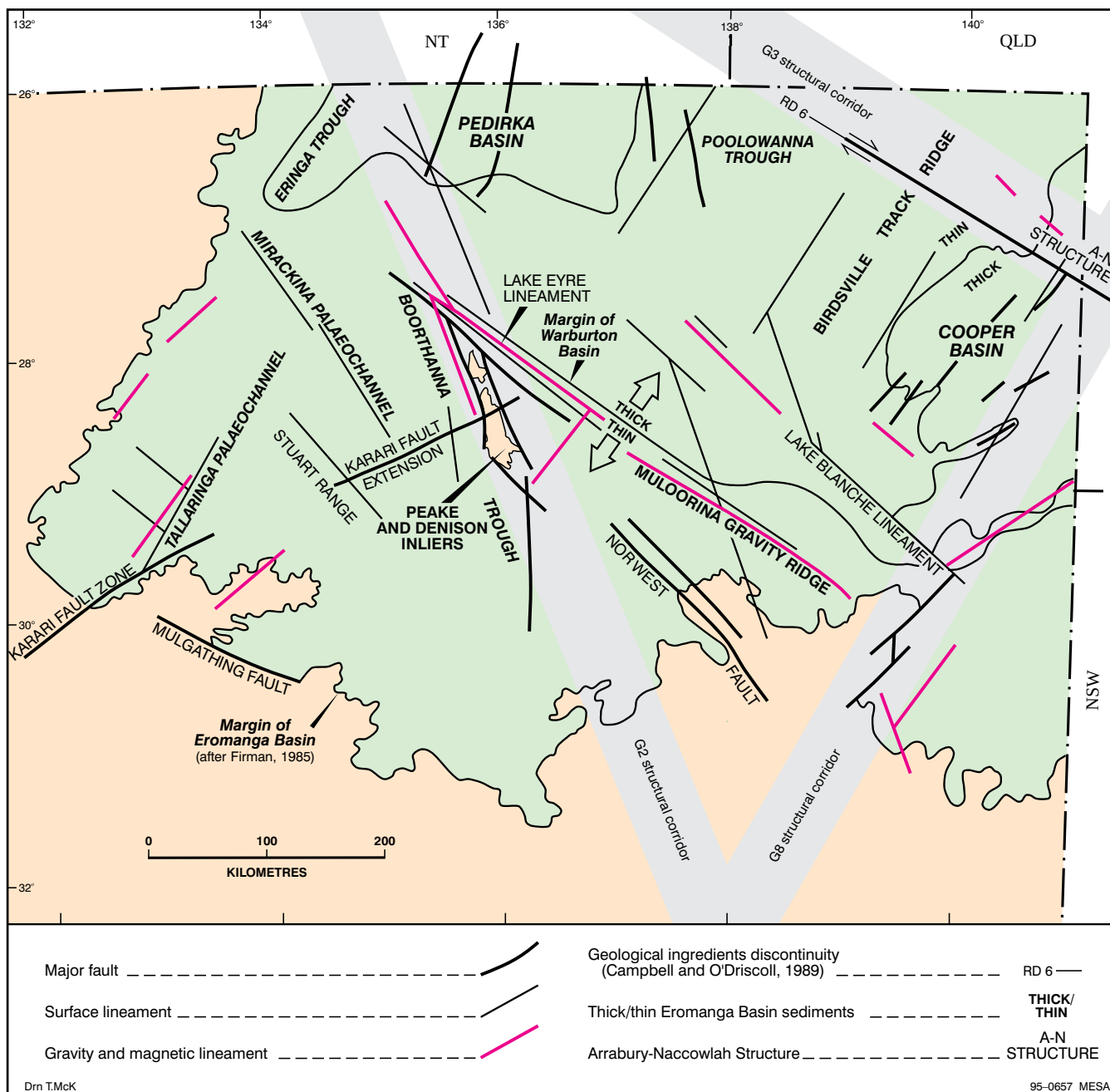


Fig. 9.13 Structural framework of the Eromanga Basin.

The Poolowanna Formation is a sequence of white to pale grey, fine to medium-grained, planar cross-bedded, horizontally bedded and ripple cross-laminated sandstone, interbedded with laminated and massive dark grey and brown siltstone and shale, and thin, discontinuous coal seams and intraclast breccias. A meandering or anastomosing fluvial environment of deposition with minor flood-plain sediments has been interpreted (Moore, 1986a). Plant remains (carbonised leaves, stems and wood fragments) and rare sinuous trace fossils occur in core. Palynomorph assemblages indicate an Early Jurassic age.

Algebuckina Sandstone

The Algebuckina Sandstone (Wopfner *et al.*, 1970) was defined from thin, stratigraphically incomplete exposures along the southwestern margin of the basin where it rests unconformably on pre-Mesozoic units. Only the upper parts of the unit are exposed. It extends subsurface eastwards into the Poolowanna Trough depocentre, where a maximum recorded thickness of 750 m occurs (Moore, 1986a), and across to the southern Cooper region. Other thin, discontinuous occurrences of Algebuckina Sandstone and

correlative clayey units, such as Birkhead Formation and Namur Sandstone equivalents (see below), extend southwest of the type section onto the Gawler Craton, southwards to Stuart Creek, and along the western margin of, and as outliers within, the northern Flinders Ranges as far south as Mount Bayley.

Nugent (1969) assigned the names Hutton Sandstone, Birkhead Formation, Adori Sandstone, Westbourne Formation, Namur Sandstone and Murta Formation to the laterally equivalent sandstone-mudstone succession in the Cooper region where up to 850 m have been intersected in petroleum drillholes (Figs 9.11, 9.12).

The Algebuckina Sandstone is a fine to coarse-grained quartzose sandstone with granule and pebble layers; shale intraclasts are common in the coarser beds. Clasts of reworked Adelaidean units occur on the basin margin (e.g. Krieg *et al.*, 1991). Minor lenses of siltstone and shale are locally developed. Noteworthy regional features displayed in outcrop are ubiquitous large and medium-scale, almost unidirectional cross-bedding, and the widely observed two-fold division separated by an erosional boundary (Wopfner *et al.*, 1970) into a lower kaolinitic, rather poorly sorted part and an upper, clean and well-sorted part. The occurrence of such unconformities

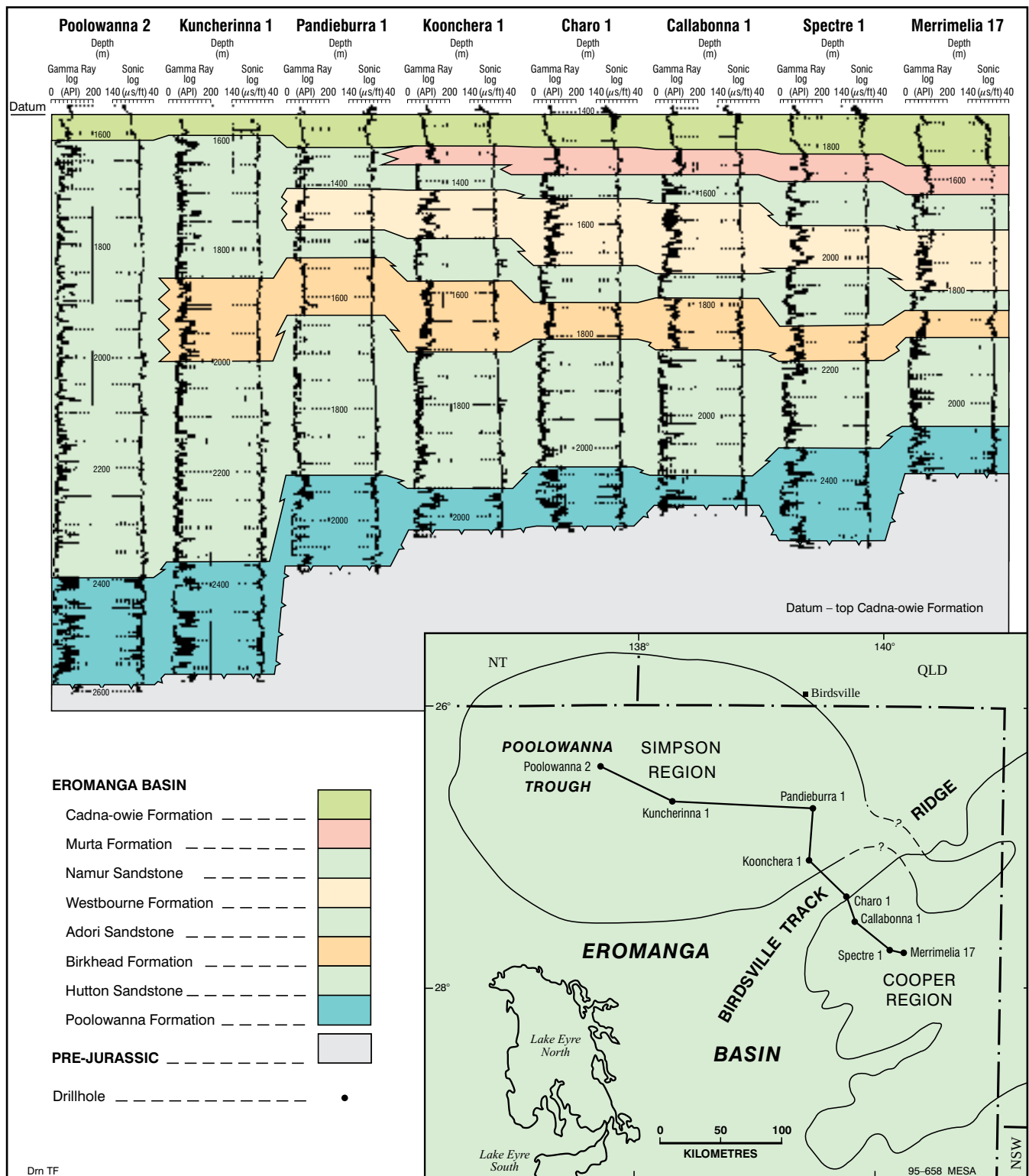


Fig. 9.14 Drill-log correlation and stratigraphic subdivision, Hutton Sandstone to Murta Formation.

near basin margins is common and this particular break is not observed in the central part of the basin. Cross-bedding in the lower part is typically either tabular-planar or large-scale, unidirectional scour-and-fill. In the upper part, large-scale trough cross-bedding with upward-fining intervals is prevalent. Similar cross-bedded facies occur in drillholes in the Poolowanna Trough. In outcrop at the top of the unit, secondary silicification, ferruginisation and carbonate cementation may be present, in some places preserving abundant plant remains as casts and moulds.

Macrofossils along the exposed margin of the basin consist of leaves, stems, reproductive structures and pieces of wood up

to 4 m long. Ferns, cycads, bennettites, seed ferns and conifers are represented (Harris, 1962; Hopgood, 1987). These plant fossils indicate that the oldest parts of the Algebuckina Sandstone in outcrop are of Late Jurassic age (Harris, 1970a; Wopfner *et al.*, 1970; Krieg *et al.*, 1991). Of particular importance for age determination is the occurrence of the seed fern *Rienitsia variabilis* which is restricted to megafloral zone B (*Ptilophyllum-Pachypteris austropapilloso* Zone) of Neocomian age in Victoria (Hopgood, 1987). As these specimens were recovered from the silicified top of the formation, the youngest likely age of the unit is Neocomian (Valanginian).

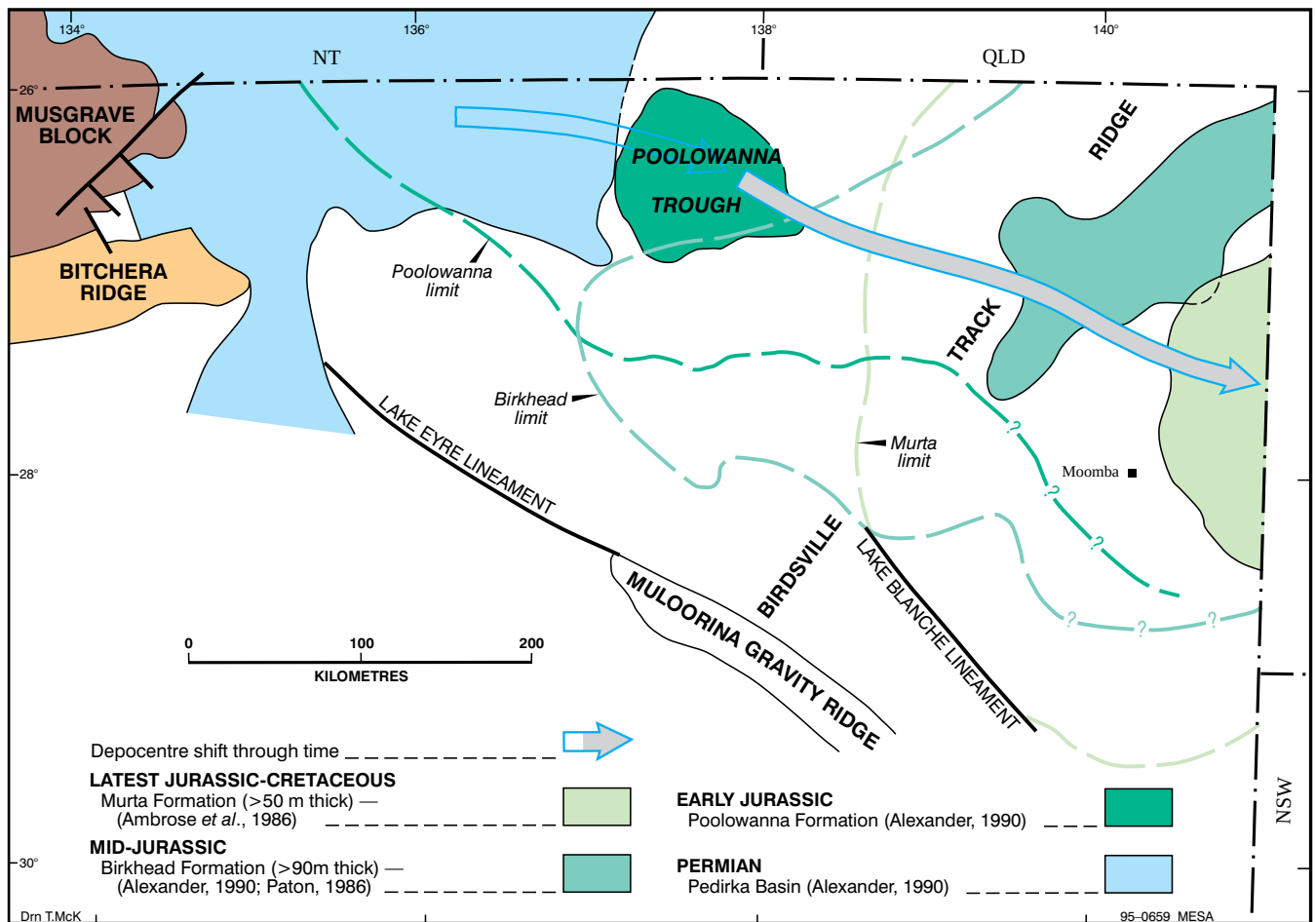


Fig. 9.15 Distribution of mudstone units in the lower non-marine sequence of the Eromanga Basin. Note the east southeasterly shift of depocentres through time.



In the subsurface, spore-pollen microfossils enable palynological zonation of the sequence. Algebuckina Sandstone ranges in age from Early Jurassic to Early Cretaceous (Neocomian). In the Poolowanna Trough the unit ranges from Middle Jurassic to Early Cretaceous (Moore, 1986a). Further west in Witcherie 1, Purni 1 and Mokari 1, palynological work by Price (1978) indicated that the Algebuckina Sandstone is of earliest Jurassic age.

The palaeogeography interpreted for the latter part of deposition by Wopfner *et al.* (1970) is that of a senile, deeply weathered landscape with low-gradient rivers, episodic sheet flooding and an arid to seasonally arid climate; this was followed by wetter conditions with stronger, more uniform flow, and lacustrine-bar as well as fluvial deposition. The cross-bedding style and coarse-grained sandy lithology suggested to Moore (1986a) a braided fluvial environment of deposition with possible development of alluvial facies near the cratonic source areas in South Australia and the Northern Territory. Wiltshire (1989) interpreted a different palaeogeography, ranging from mixed aeolian and receding scarp fluvial sedimentation in the hinterland to a quartz-sand dominated shallow lacustrine regime, a view consistent with the variously arid and wet conditions of Wopfner *et al.* (1970). A wet, cool to temperate, conifer-forested landscape interpreted from plant fossil studies, and marked seasonal snow melt causing variable river flow, have also been suggested by Krieg *et al.* (1991).

Top: Angular cross-bedding in kaolinitic sandstone of the Algebuckina Sandstone, southeast of 'Finniss Springs'. A bounding, near-horizontal pebble layer occurs near the hammer head. Photo 042463.

Bottom: Algebuckina Sandstone, large scale low angle cross bedding and layers of kaolinised clasts, south east of Finniss Springs homestead. Photo 039077.



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Hutton Sandstone

The name Hutton Sandstone was first used by Reeves (1947) for outcrop in the Surat Basin in Queensland. Kapel (1966) applied the name to sandstone of 'Lower to Middle Jurassic age' in the Cooper region. The Hutton Sandstone intertongues with the Poolowanna Formation in the Cooper Basin region and, where Poolowanna Formation is absent, the Hutton lies unconformably on Cooper and Warburton Basin units. The unit intertongues with and is overlain by Birkhead Formation (Watts, 1987), and is restricted to the subsurface in South Australia.

The Hutton Sandstone consists of fine to coarse-grained, white sandstone with minor dark grey lenticular carbonaceous siltstone and shale interbeds and pebble conglomerate layers. The unit is a mineralogically mature quartzarenite with minor feldspar and muscovite, and trace amounts of zircon, garnet and tourmaline (Gravestock *et al.*, 1983). Planar cross-bedding, often with upward-fining foresets and horizontal bedding, is common in cores from the upper parts of the unit. As Wiltshire (1989) observed, little is known about the lower Hutton Sandstone, which is not generally cored and is mostly known from cuttings and wireline logs. Gravestock *et al.* (1983) described thin duricrusts of calcrete and silcrete within the upper Hutton Sandstone and Birkhead Formation from thin sections of cuttings and wireline log responses.

Earlier models have assumed a braided fluvial model with transport to the east and northeast from westerly sediment sources (e.g. Moore, Hobday *et al.*, 1986; Watts, 1987), reaching the sea via the Surat Basin in Queensland. Wiltshire (1989) pointed out that such a high-energy fluvial regime 'would certainly have required a substantial drainage exit and there is no evidence of such anywhere in the (eastern Eromanga) Basin'. As an alternative to the wholly braided fluvial model, Wiltshire (1989) proposed that braided streams transported sand into the basin — the sand was then distributed across the basin by aeolian and lacustrine processes. The Hutton Sandstone is of Middle Jurassic age (Fig. 9.12).

1. Soft-sediment injection features with gravelly infill, Algebuckina Sandstone, southeast of 'Finniss Springs'. Photo 039582.
2. Climbing ripples in Algebuckina Sandstone, southeast of 'Finniss Springs'. Photo 039078.
3. Reverse graded bedding in Algebuckina Sandstone, southeast of Finniss Spring'. Photo 039079.
4. Fern frond impressions (*Cladophlebis cf. australis*) from silicified top of the Algebuckina Sandstone, Mount Anna. Photo 041930.
5. Medium to coarse-grained, porous Hutton Sandstone showing cross-stratification and reverse graded bedding, from 1848.6 m in Gidgealpa 17. Width of view is 80 mm. Photo 042464.



5

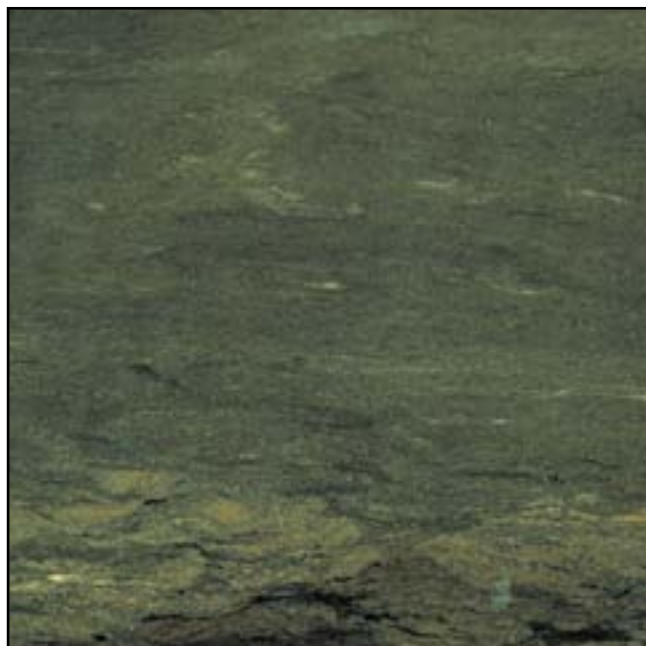
Birkhead Formation

The Birkhead Formation is poorly exposed on the eastern margin of the Eromanga Basin near Tambo in Queensland and was defined by Exon (1966). The Birkhead Formation intertongues with the upper Hutton Sandstone in the Cooper region (Watts, 1987) and with Algebuckina Sandstone west of the Birdsville Track Ridge (Figs 9.11, 9.12, 9.14). The unit is conformably overlain by Namur or Adori Sandstone in the eastern Cooper region in South Australia. Paton (1986) studied the Birkhead Formation in detail in South Australia, Watts (1987) studied the transition between the Hutton Sandstone and Birkhead Formation in Queensland, and source-rock potential of the unit was examined by Smyth *et al.* (1984).

The Birkhead Formation consists of interbedded dark grey and brown siltstone, mudstone and buff, fine to medium-grained sandstone with thin, lenticular coal seams (less than 0.3 m thick), and thin silcretes and calcretes (Gravestock *et al.*, 1983). Siltstone and mudstone are carbonaceous, finely laminated to massive and slumped, and are often disrupted by root mottling and weak bioturbation. Fine-grained sandstones range from massive with water escape structures to ripple cross-laminated; horizontally laminated clay drapes are common. Medium-grained sandstones are commonly trough and planar cross-bedded, with graded foresets, coal fragments, pebble lags and shale intraclast breccias. Birkhead Formation silt, mud and coal were deposited in a lacustrine ('Lake Birkhead') and coal-swamp environment, cut by meandering fluvial channels (Paton, 1986). Paton proposed blockage or restriction of the Hutton Sandstone fluvial system producing lower energy Birkhead environments. In contrast, Wiltshire (1989) regarded climate change and/or changes in the rate of silt and shale supply into the basin as the origin of Birkhead deposition.

The transition from clean, quartzose Hutton Sandstone to the Birkhead Formation is marked by an influx of volcanoclastic sandstones in Queensland (Watts, 1987), reflecting a change in provenance from stable cratons in the south and west to an active mobile belt off the east coast of Queensland (Moore, Hobday *et al.*, 1986). South Australia was distal to any volcanoclastic source, and lithics, ferromagnesian minerals and plagioclase feldspar are present in trace amounts only (Gravestock, 1982).

The Birkhead Formation ranges in age from Middle to Late Jurassic (Fig. 9.12). Of interest is a biostratigraphically constrained carbonaceous clay of Late Jurassic age, described from a Department of Mines and Energy seismic shothole in the Arkaringa Basin region by Harris (1970a). The clay is identified herein with the PJ5 Zone (Fig. 9.12), equivalent to the Birkhead Formation.



Adori Sandstone

The Adori Sandstone was defined by Woolley (1941) from outcrop in the Tambo area in Queensland and was first published by Hill and Denmead (1960). Exon (1966) remeasured the type section and provided a new definition of the unit. Williams (1989) conducted a sedimentological study of the Adori Sandstone in Queensland. The unit occurs in the northeastern part of the Cooper region, where it conformably overlies the Birkhead Formation. Adori Sandstone can be picked on wireline logs only if Westbourne Formation shales are present — both units are equivalents of the Namur Sandstone and Algebuckina Sandstone (Nugent, 1969).

The Adori Sandstone consists of well-sorted, subrounded, cross-bedded, fine to coarse-grained sandstone. Diagenetic calcite cement, distinct from the duricrusts described by Gravestock *et al.* (1983) in underlying formations, is developed locally in the basal Adori and Namur Sandstones (Gravestock, 1982). These calcite cemented zones are up to 45 m thick (Schultz-Rojahn, 1993) and cause problems in seismic interpretation when mapping structures. The origin of the calcite has been related to vertical migration of CO₂ from the underlying Cooper Basin, via faults into the Adori or Namur Sandstones and reacting with artesian water, to precipitate calcite (Schultz-Rojahn, 1993), or to the oxidation of hydrocarbons trapped in the unit and oxidation of organic compounds in siltstone of the Birkhead Formation (Townsend, 1993).

A low-sinuosity braided fluvial system dominated by transverse bars has been interpreted for the Adori Sandstone from outcrop and cored drillholes in Queensland (Williams, 1989). In the subsurface in South Australia, a braided fluvial environment of deposition is widely accepted, although Wiltshire (1989) proposed an alternative model involving wave reworking of a pulse of sand supplied into Lake Birkhead. The Adori Sandstone is Late Jurassic in age (Fig. 9.12).

Westbourne Formation

Exon (1966) defined the Westbourne Formation from a drillhole near Tambo in Queensland where it also crops out. The unit is restricted to the northern Cooper region (Nugent, 1969) in South Australia, where it lies conformably on the Adori Sandstone and is overlain by and intertongues with Namur Sandstone. The formation consists of interbedded dark grey shale and siltstone with minor sandstone. Williams (1989) described hummocky cross-bedding, ripple cross-lamination, plant fossils and vertical burrows from outcrop near Tambo, and interpreted a lacustrine depositional environment. Wiltshire (1989) envisaged similar palaeogeography for the Birkhead and Westbourne Formations, although the latter is not as extensive. The Westbourne Formation ranges from Late Jurassic to Early Cretaceous in age (Fig. 9.12).

Namur Sandstone

Mount (1981) named the Namur Sandstone Member after Namur 1 petroleum exploration drillhole which flowed gas from the unit. Previously the unit was referred to as Mooga Formation (Nugent, 1969), 'Mooga Porosity' or as Hooray Sandstone. Following the work of Martin (1983), Gravestock *et al.* (1995) elevated the Namur Sandstone Member of the Mooga Formation to formational status as Namur Sandstone, and discarded Mooga Formation (Fig 9.16). The Namur Sandstone conformably overlies Birkhead Formation and intertongues with Westbourne Formation. The unit is overlain by Murta Formation or McKinlay Member, is distributed throughout the Cooper region, and is a lateral equivalent of the Algebuckina Sandstone.

Namur Sandstone consists of white to pale grey, fine to coarse-grained sandstone with minor interbedded siltstone and mudstone deposited in a braided fluvial environment. Channel lag conglomerates consist of lithic and quartz pebbles, carbonaceous mudstone intraclasts and carbonised plant fragments. The basal Namur Sandstone, like the Adori Sandstone, has been strongly

Birkhead Formation carbonaceous siltstone with irregular thin bedding, from 1564.5 m in Wancoocha 4. Width of view is 85 mm. Photo 042465.

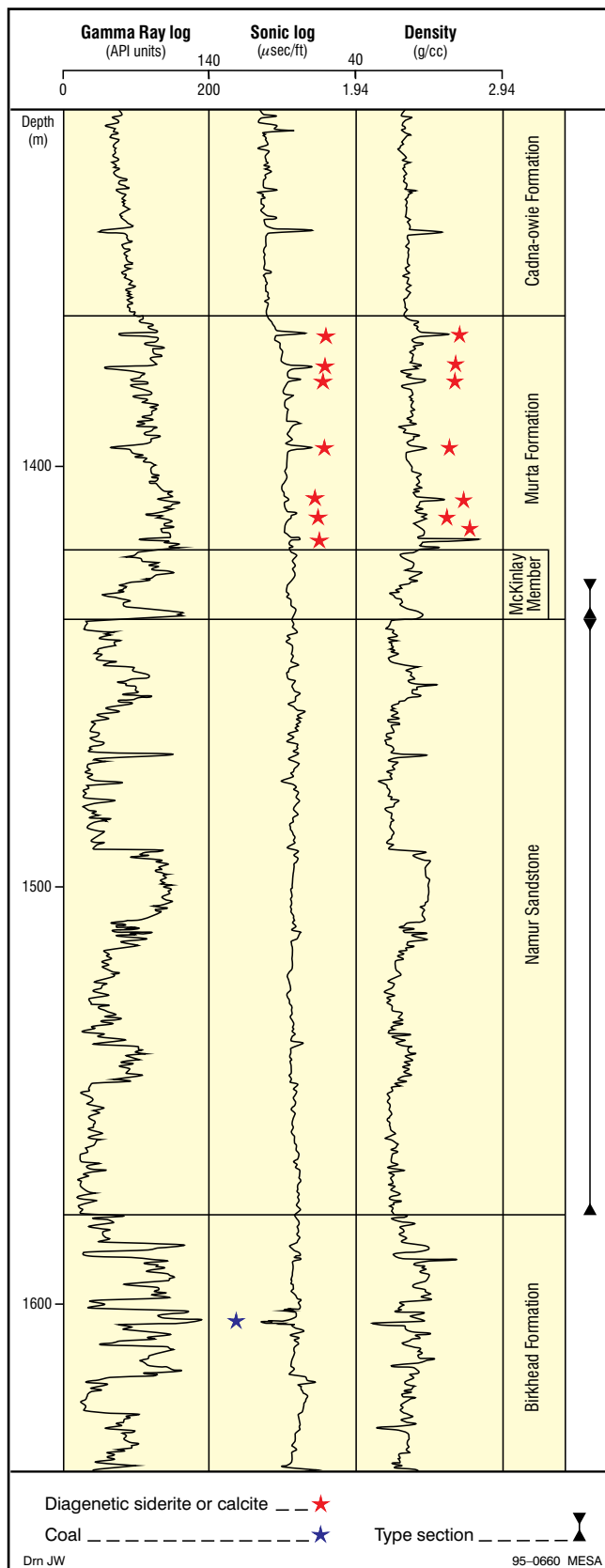


Fig. 9.16 Stratigraphy subdivision and lithology features from Strzelecki 4 drill-log interpretation.

cemented with diagenetic calcite in some drillholes. The formation ranges from Late Jurassic to Early Cretaceous (Neocomian) in age; Alley (1987a) determined that a mudstone on the basin margin at Davenport Springs falls into the PJ6.2.2 Zone (Fig. 9.12), equivalent to the Namur Sandstone.

Murta Formation

Nugent (1969) defined siltstone and shale in the upper part of the Mooga Formation as the Murta Member. A type section in Dullingari North 1 was defined by Mount (1981). Gravestock *et al.* (1995) elevated the name to Murta Formation. The unit overlies and intertongues with the McKinlay Member and Namur Sandstone, and is overlain transitionally by the Cadna-owie Formation. The Murta Formation occurs throughout the Cooper region and intertongues with Algebuckina Sandstone west of the Birdsville Track Ridge (Alexander and Hough, 1990, fig. 5). The sedimentology of the Murta Formation has been studied in detail by Mount (1982) and Ambrose *et al.* (1982, 1986), who proposed a reference section in Dullingari 9. Michaelson and McKirdy (1989) and Powell *et al.* (1989) examined the petroleum geochemistry of the Murta Formation.

The unit consists of finely interbedded dark grey siltstone, shale and pale grey fine to very fine-grained sandstone, and minor medium to coarse-grained sandstone. Rusty diagenetic siderite nodules and layers have cemented sandstone and siltstone beds, and produce distinctive spikes on sonic and density wireline logs (Fig 9.16). A basal laminated siltstone unit is widespread in the Cooper region. The siltstone coarsens upwards as the frequency of fine-grained sandstone beds increases. These finely interbedded siltstone and sandstone beds are often slumped; water-escape structures, ball and pillow structures, and bioturbation also disrupt bedding (Ambrose *et al.*, 1986). Sharp-based sandstone interbeds typically progress from massive to laminated to ripple cross-laminated (Bouma ABE sequence), and fine upwards into laminated, rippled, then quieter water laminated, and massive siltstone beds (Bouma BCE and BCDE sequences), indicating deposition on a slope by density currents (Ambrose *et al.*, 1986). In the middle and upper parts of the Murta Formation, massive and flat-bedded medium to coarse-grained sandstones up to 0.6 m thick form important oil reservoirs in the Murteree Ridge and Dullingari areas. The Murta Formation gradually fines upwards into thin graded siltstone beds of the Cadna-owie Formation.



Fine-grained lacustrine turbidites with microfaulting and bioturbation, Murta Formation from 1206.7 m in Limestone Creek 3. Width of core is 97 mm. Photo 042466.



Heavily bioturbated micaceous and carbonaceous shoreface sandstone of the Murta Formation (McKinlay Member), from 1423 m in Strzelecki 4. Width of view is 85 mm. Photo 042467.

Mount (1981, 1982) and Ambrose *et al.* (1986) interpreted the following facies: open lacustrine, lacustrine fan delta, distributary channel and lacustrine shoreline. The presence of ubiquitous *Botryococcus* (a planktonic alga) and land plant spores and pollen is also evidence of a lacustrine to paralic environment (Michaelsen and McKirdy, 1989). Ambrose *et al.* (1986) found no evidence of marine influence in the Murta Formation; geochemical work by Zoellner (1988) suggested some evidence of a brackish to marine environment for the upper Murta Formation. The formation is of Neocomian age (Fig. 9.12).

The silty sandstone transitional between clean, braided, fluvial Namur Sandstone and lacustrine siltstone and mudstone of the overlying Murta Formation was informally named 'McKinlay Member' of the Mooga Formation by Mount (1982). Gravestock *et al.* (1995) defined the unit as a basal member of the Murta Formation and assigned a type section from Strzelecki 4, where the entire unit has been cored. The McKinlay Member consists of buff, micaceous, fine to medium-grained sandstone interbedded with dark grey carbonaceous siltstone. Sandstones are horizontally bedded and ripple cross-bedded. Strongly bioturbated intervals are common and carbonised plant rootlets are present. The McKinlay Member has a high gamma ray log response (a so called 'hot sand') due to relatively high mica, feldspar and zircon contents. Ambrose *et al.* (1986) interpreted a low-energy fluvial and flood-plain environment; Gravestock *et al.* (1995) proposed a lacustrine shoreface environment. The McKinlay Member is of Early Cretaceous age (Fig. 9.12).

Stratigraphy — marine succession

G.W. Krieg and P.A. Rogers

The marine succession, which includes some non-marine deposits near the base, represents the last great marine transgression onto the Australian continent; subsequent transgressions are peri-continental. The marine sediments reach ~1 000 m thickness and span most of the Early Cretaceous (Valanginian to latest Albian), some 40 million years. In ascending order, the units are Cadna-owie Formation, Marree Subgroup and equivalents, and Mackunda Formation (Figs 9.12). It is noteworthy that the time of maximum transgression, probably during the Aptian, does not coincide with the maximum global sea-level highstand of Haq *et al.* (1987) which occurred later, in the mid-Cretaceous.

Cadna-owie Formation

The Cadna-owie Formation is a thin, mainly fine-grained unit that extends throughout the Eromanga Basin. First named 'Transition beds' by Whitehouse (1955) and later defined as Cadna-owie Formation by Wopfner *et al.* (1970), it records the transition from terrestrial-freshwater to marine conditions. The unit thus represents the onset of the Cretaceous marine transgression as illustrated by the upward passage from non-marine to marginal marine deposition in the central part of the basin (Moore and Pitt, 1985). However, the fluvial-deltaic Mount Anna Sandstone Member (Wopfner *et al.*, 1970) and the fluvial-tidal Trinity Well Sandstone Member (Forbes, 1966; Lemon, 1988) have been identified near the top of the unit along the southwestern and southern margins of the basin, respectively. Two other named stratigraphic units equivalent to the Cadna-owie Formation are the Parabarana Sandstone (Ludbrook, 1966) and Pelican Well Formation (Forbes, 1966) on the southern margin.

The Cadna-owie Formation is typically 10-20 m thick in exposures around the basin margin, increasing to 75-100 m in the deeper parts of the basin (Moore and Pitt, 1985). The lower boundary may be either erosional as in the type section on the basin margin or gradational into the Murta Formation (Ambrose *et al.*, 1986) in the central part of the basin. The upper boundary is sharp, being marked by a sudden upward change from calcareous sandstone to shaly mudstone. Thus, the top of the Cadna-owie approximates a pronounced and distinctive seismic reflector, the 'C' seismic horizon, traceable throughout the Eromanga Basin (Fig. 9.17). Along the southwestern margin the Cadna-owie is more continuous than, and in places overlaps, the Algebuckina Sandstone. Here, the thinness of the unit compared to its depositional time-span, and the abrupt lithological changes, suggest breaks in sedimentation, a view supported by palynological evidence (Alley, 1987; Dettmann and Williams, 1985). These may reflect tectonic instability influencing depositional margins of the basin, or changes in subsidence rate and/or eustatic sea level.

Lithology and sedimentary features

Excluding the named sandstone members, the Cadna-owie Formation is a pale grey siltstone and very fine to fine-grained sandstone with laterally extensive or locally developed medium to very coarse sandstone interbeds and minor carbonaceous claystone intervals. Pebbly layers, diamictites and coarse breccia layers occur locally and large, rounded limestones up to boulder size are widely distributed around the basin margin. These are concentrated in layers, as in the type section, or are randomly scattered in the fine-grained beds, and consist mainly of quartzite and, occasionally, igneous and gneissic lithologies. Apart from quartz, the main mineral constituents are mica, fresh feldspar, heavy minerals, pyrite and glauconite (green smectite). Hard, coarser sandstone beds occurring regionally at or near the top of the unit are cemented by extremely coarse calcite crystals enclosing quartz grains. Intense ferruginisation in thin, distinct layers is also common and calcareous sandstone beds containing ooids form a minor constituent of the unit.

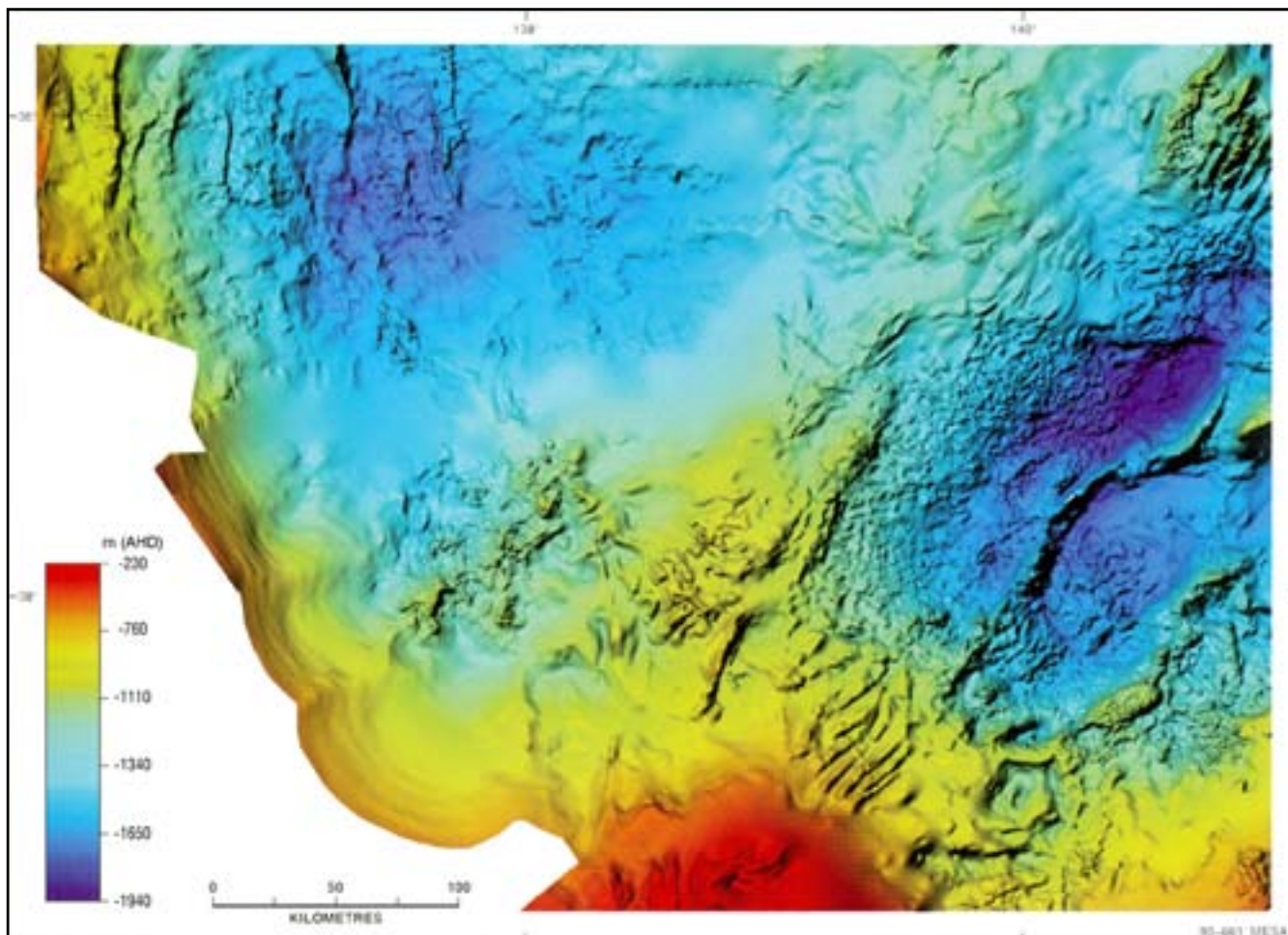


Fig.9.17 Computer-generated plot of seismic 'C' horizon showing structural features of the Eromanga Basin.

The Cadna-owie Formation is mainly a thick-bedded sequence with parallel bedding and upward-coarsening intervals. Lamination and minor cross-lamination occur in the finer grained beds while small to medium-scale trough cross-bedding is sometimes present in the coarser layers. Other rarely observed sedimentary structures include pipe-like dewatering structures, ripple-marked bedding planes, hummocky cross-stratification, tidal bundle sequences and minor bioturbation. In the central parts of the basin, subsurface Cadna-owie Formation has been recorded as a single, uniform lithological unit coarsening upward from grey siltstone at the base into fine to occasionally medium-grained calcareous sandstone at the top (Moore and Pitt, 1984, 1985).

The Mount Anna Sandstone Member (Wopfner *et al.*, 1970) is a well-sorted, medium to coarse-grained feldspathic sandstone with large-scale trough cross-bedding and is characterised by abundant rounded pebbles of red porphyritic rhyolite. Calcitic and ferruginous cements are common. At its most northerly observed exposure in the Cadna-owie Formation type section this member is only 0.3 m thick but thickness and proportion of total Cadna-owie Formation increases southwards to 30-40 m and 100% of the unit.

The Trinity Well Sandstone Member, originally placed in the Marree Formation (now Subgroup) of Forbes (1966), is regarded as part of the Cadna-owie Formation. It consists of pebbly, ferruginous and carbonate-cemented sandstone, and medium to coarse micaceous and feldspathic sandstone with large-scale cross-bedding. Vertical tubular structures have been observed locally, and plant stem impressions and fossil shells have been reported.

The Parabarana Sandstone is a similar, calcareous unit with pebbly interbeds containing quartz, quartzite and porphyry clasts. It includes grey, carbonaceous sand, gravel and shale near the base, and laminated medium to coarse-grained micaceous quartzite

higher in the section. Large U-shaped burrows occur in the unit, and plant stem impressions and fossil shells have been reported.

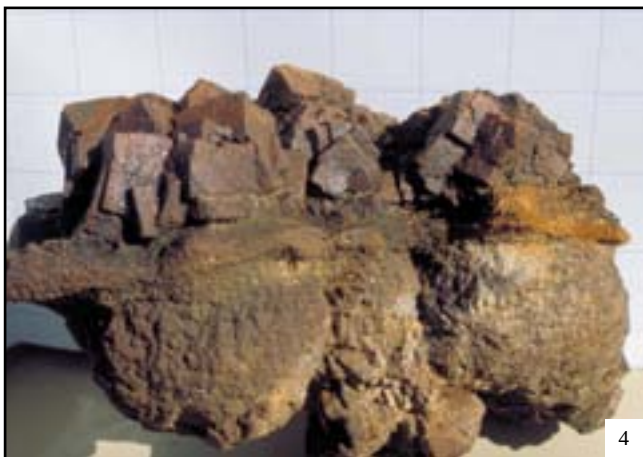
The Pelican Well Formation, an equivalent of the Cadna-owie, is a grey, carbonaceous claystone-siltstone-fine sandstone unit with boulder beds and scattered, large limestones. A basal cross-bedded sandstone with a pebble bed, and calcareous concretions higher up, occur in the type section.

Fossil content and age

Palaeontological data for the Cadna-owie Formation derive largely from palynomorph assemblages, with lesser microplankton, algae, dinoflagellates and rare foraminifera. These indicate a mid-Neocomian (Valanginian) to early Aptian age (Fig. 9.12). Macrofossils consist of occasional plant and rare molluscan remains, and a shark tooth has been reported from the type section. A freshwater mussel *Protovirgus coatsi* is known from the Trinity Well Sandstone Member and various invertebrates such as the brachiopod *Lingula subovalis* occur in the Parabarana Sandstone.

Microfossils include conifer pollen such as *Microcachrydites antarcticus* and *Podocarpidites ellipticus*, the Neocomian dinoflagellate *Fusiformacysta salasii*, algae including *Schizosporis reticulatus*, and a foraminiferal assemblage containing *Gandryinella permacra*, diagnostic of the early Aptian *Textularia annacooraensis* Zone.

Palynological dating shows that the major part of the Cadna-owie Formation is within the *Foraminisporis wonthaggiensis* Zone (Helby *et al.*, 1987) of mid-Neocomian to Barremian age. In the central part of the basin and in the Trinity Well section, the unit ranges down into the top of the early Neocomian *Cicatricosisporites australiensis* Zone and, in the southwest margin, as at the type



section, it ranges up into the base of the *Cyclosporites hughesii* Zone of Aptian age (Fig. 9.12). This suggests that the upper boundary may be diachronous, younging towards the basin margin, but the exact relationship of the top defined in outcrop to the continuous, seismic 'C' horizon in the centre of the basin is still uncertain. Studies by Alley (1987a) and Dettmann and Williams (1985) suggest locally diachronous boundaries and hiatuses within the unit. Evidence for a hiatus at the base of the upper sandstone member of the unit (Wyandra Sandstone Member in Queensland) is suggested by the incompleteness of the *C. hughesii* Zone.

Interpretation

The Cadna-owie Formation was deposited in the coastal regions of the Early Cretaceous sea and represents a variety of marginal-marine to non-marine environments associated with the transgressive front and adjacent coastland. The consistent thickness and lithological uniformity of the unit in the central part of the basin suggest a steady advance of the sea across a fairly uniform landscape, but the hiatuses (Dettmann and Williams, 1985; Alley, 1985a, 1987a) and rapid lithological changes around the basin margin suggest an irregularly advancing, perhaps oscillating shoreline possibly with extended periods of stillstand. Mild tectonism (Wopfner *et al.*, 1970) superimposed on the eustatically induced transgression may have had a major effect on coastline configuration and local shoreline history. The vertical distribution of marineness and the particular stratigraphic interval locally preserved may, thus, vary from place to place (Fig. 9.18). For example, a succession with paralic, shoreface and tidal facies commencing in the Valanginian occurs at Trinity Well, whereas a mainly paralic sequence no older than early Aptian represents Cadna-owie Formation in Toodla 1 (Alley, 1988).

Particular environments interpreted for the Cadna-owie Formation around the basin margin include near offshore or outer shoreface (hummocky cross-stratification), inner shoreface to foreshore (ripple marks, fine cross-lamination, minor bioturbation), shallow agitated waters of a marine shoal or lagoonal shoreline (oolitic sandstone), intertidal (tidal bundle sequences), high-gradient gravelly beach (talus and swash zone lag deposits representing the depositional edge of the unit), back barrier lagoon or coastal marsh (fine-grained, pyritic and carbonaceous deposits) and fluvial (Mount Anna Sandstone). For the deeper part of the basin, Moore and Pitt (1985) suggested a lacustrine environment, passing up to paralic then to shoreface and beach, for the Cadna-owie Formation.

Mega-sparry calcite cement, commonly occurring in the top of upward-coarsening cycles, extends as a regional characteristic hundreds of kilometres around the southwestern margin of the basin (Wopfner *et al.*, 1970). Such an unusual yet widespread fabric would require a singular set of conditions controlled by regional



1. Basal Cadna-owie Formation regolith developed on steeply dipping Neoproterozoic Burra Group near Davenport Springs. Width of view is 2 m. Photo 042468.
2. Massive carbonate-cemented sandstone with angular, locally derived Burra Group clasts, Cadna-owie Formation near Davenport Springs. Photo 039083.
3. Water-escape structures within fine-grained sandstone of the Cadna-owie Formation near Davenport Springs. Photo 039084.
4. Ferruginous pseudomorphs after pyrite, and downward-extending lobes in growth orientation, form uppermost part of the Cadna-owie Formation, near Cockatoo Bore. Specimen width is 0.18m. Photo 039081.
5. Vertical burrows in carbonate-cemented massive sandstone of the Cadna-owie Formation, southeast of the disused Edwards Creek railway siding. Photo 042469.
6. Lens of hummocky cross-stratified, very fine sandstone in a siltstone facies of the Pelican Well Formation near Trinity Well. Photo 042470.
7. Anticline in siliceous medium-grained Parabarana Sandstone near Parabarana Hill. Bulldog Shale lies under the plain to the left of view. Photo 042471.
8. U-shaped burrow (*Rhizocorallium Springs.*) in Parabarana Sandstone near Parabarana Hill. Photo 042472.

Top: Cast of *Panopea maccoyi* (Moore) in siliceous medium-grained Parabarana Sandstone near Mount Babbage. Photo 042473.

Centre: Mount Anna Sandstone Member at the type section near Mount Anna. Pebbles of red porphyritic rhyolite and various quartzose lithologies occur in coarse-grained, cross-bedded sandstone. Photo 042474.

Bottom: Ferruginous grit layer with rhyolite, quartz and quartzite pebbles in the Mount Anna Sandstone Member at the type section near Mount Anna. Photo 042475.

influence. The hydrodynamic and hydrochemical processes of a temporarily retreating shoreline (Krieg, Rogers *et al.*, 1990), or later diagenesis when calcite was precipitated preferentially within regionally extensive, porous sand layers are possible causes.

The climate during Cadna-owie time was cooler than during Algebuckina Sandstone deposition. A decrease in the more temperate plant species such as the *Araucaria* group and podocarps, and an increase in the more cold-tolerant conifer *Microcachrys*

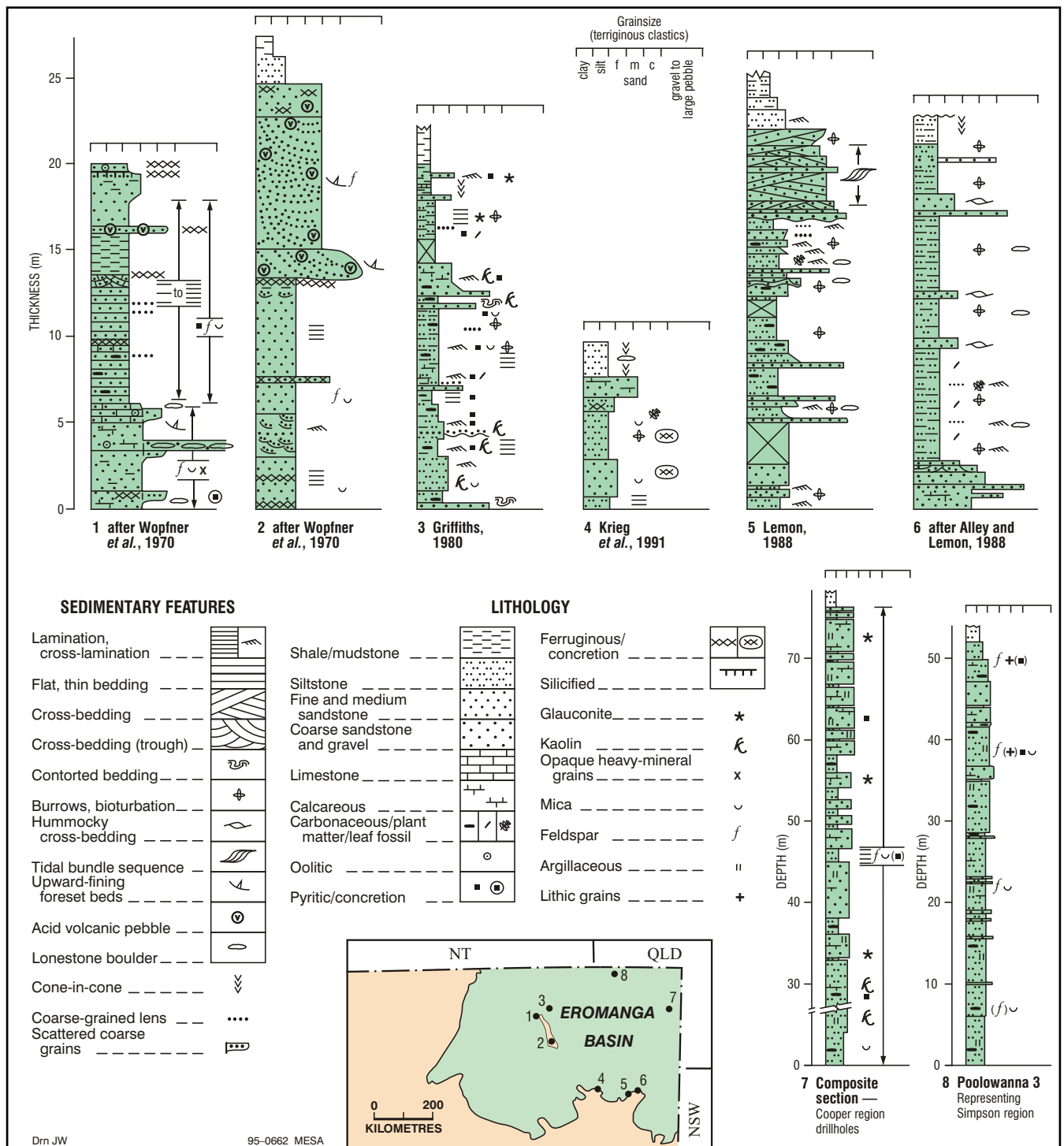


Fig. 9.18 Exposed and subsurface sections of Cadnaowie Formation showing lithological and thickness variations.

(Krieg *et al.*, 1991), is consistent with the proposal of Frakes and Francis (1988) that the climate was very seasonal and the winters cold enough to allow rivers and basin shorelines to freeze. Oxygen isotope data for nearby regions of Gondwana (e.g. Gregory *et al.*, 1989) also support cool to cold temperatures at this time. Such an environment may explain the presence of very large boulders, the likely sources of which were Permian boulder clay and bouldery river beds. Seasonal ice may have carried the boulders to the coast from where they were ice-rafted basinwards and emplaced on the seafloor as dropstones (Frakes and Francis, 1988) or transported basinwards by tectonically driven seafloor creep (Wopfner *et al.*, 1970).

The Mount Anna Sandstone Member is a fluvial to deltaic sandstone whose source was in part the Gawler Ranges Volcanic

Province. A major tectonic event late in the Neocomian or earliest Aptian which uplifted the province to form the Gawler Range Massif would account for the sudden incoming of the unit within the upper part of the Cadna-owie Formation (Wopfner *et al.*, 1970; Ambrose and Flint, 1981). Large, high-energy streams transported abundant coarse detritus, notably porphyritic rhyolite and dacite clasts, northeastwards to the coastline and built deltas into the sea. Near the source area, deltas were river-dominated as indicated by the abundance of plant remains (e.g. Ambrose and Flint, 1981) but deltaic tide-dominated environments may be represented by the Trinity Well Sandstone Member and Parabarana Sandstone.

The depositional environments described above were replaced in the early Aptian by a major marine transgression that set the conditions for the next cycle of sedimentation.

Marree Subgroup

The Marree Subgroup, comprising Bulldog Shale, Coorikiana Sandstone and Oodnadatta Formation (Fig. 9.12) in ascending stratigraphic order, is essentially the marine shaly interval of the Eromanga Basin in South Australia. Formerly named Marree Formation (Forbes, 1966), this lithostratigraphic unit was elevated to Subgroup status (of the Neales River Group, Wopfner *et al.*, 1970) when the regional stratigraphic relationships of its component formations were better understood (Thomson, 1980; Moore, 1986b). The interval thickens rapidly from the margin to the basin centre where the Queensland terms Wallumbilla Formation and Allaru Mudstone (Vine *et al.*, 1967), corresponding approximately to Bulldog Shale and Oodnadatta Formation, respectively, are used (Moore, 1986b).

Bulldog Shale

The Bulldog Shale (Freitag, 1966) is a grey marine mudstone which locally includes the basal pebbly Wilpoorinna Breccia Member (Forbes, 1966) around the basin margin. Its maximum known thickness is ~340 m in northeastern South Australia (Moore, 1986b), but it thins to <200 m in the Oodnadatta and Marree regions in the exposed reference sections. Further westwards the unit thins stratigraphically and by erosional stripping; outlying remnants of Bulldog Shale have been recorded in the Andamooka and Woomera areas and in coal exploration drilling by ETSA west of the northern Flinders Ranges. Outcropping Bulldog Shale has commonly undergone chemical alteration by weathering events that produced the multi-hued, picturesque colouration of the Arkaringa-Stuart Range landscape. Top and bottom boundaries are conformable, the lower usually sharp and the upper gradational.

Lithology and sedimentary features

The Bulldog Shale consists of dark grey, bioturbated and fossiliferous shaly mudstone, with pale grey micaceous silt to very fine sand intervals that commonly show fine cross-lamination or irregular interlamination with mudstone. Carbonaceous matter and pyrite are also present. Detrital quartz, glauconite and feldspar are the main constituents of the sandy fraction (Radke and Brown, 1985). The unit has thin, parallel bedding that produces broadly horizontal shaly parting in outcrop. An organic-rich dark shale forms the basal 10–25 m of the unit and, because of its low sonic velocity, accentuates the 'C' seismic reflector (Freitag, 1966; Moore and Pitt, 1985).

In marginal areas the basal portion of the Bulldog Shale contains numerous rounded limestones up to very large boulder size like those in the Cadna-owie Formation, sometimes associated with sandy pebble lenses, sand stringers and occasional lenses of khaki-yellow cone-in-cone limestone. Large pieces of fossil wood are common. Higher in the unit, and most commonly in the middle part, dark grey ellipsoidal limestone concretions up to 1 m thick and 2–3 m across occur in layers at various stratigraphic levels (Krieg, Rogers *et al.*, 1990). Some of the concretions contain abundant shelly remains which locally form coquinoid layers and are commonly rich in microfossils. Glendonite nodules (crystal aggregates pseudomorphing the calcium carbonate hexahydrate mineral ikaite) have been discovered in Bulldog Shale (Sheard, 1990).

Along the western basin margin, west of Oodnadatta, the Bulldog Shale includes several horizons, generally up to 5 m thick, of very fine to fine-grained sandstone with thin, planar bedding, small-scale cross-bedding, hummocky cross-stratification and bioturbation. Also present are numerous thin beds of coarse feldspathic quartz sandstone and grit with mega-ripple structures. These sandy units have previously been misidentified and mapped as Coorikiana Sandstone, a slightly younger unit.

The Wilpoorinna Breccia Member, described by Forbes (1966) as a basal 'bouldery slate-pebble breccia', is 3–5 m thick and rests directly on Adelaidean rocks south of Marree. Besides the predominant, locally derived slate clasts it also contains pebbles, cobbles and boulders of quartzite, laminated siltstone, and red

and grey porphyry. The breccia has a dark grey, partly sandy and calcareous clay matrix; it is partly ferruginised and contains ironstone concretions. Fossil wood is present at some localities. The member is thought to be represented by a pebbly unit in SADME Finnis 2 and also in a small exposure near 'Callanna' on the flank of the Willouran Ranges.

Fossil content and age

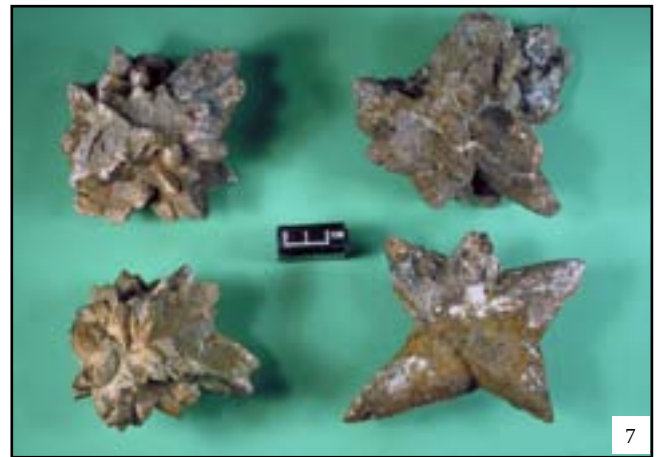
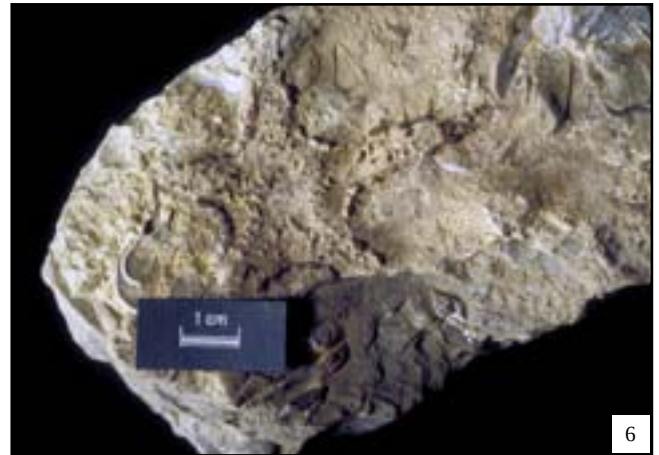
A rich and diverse fossil assemblage occurs in the Bulldog Shale (Ludbrook, 1966). Macrofaunas are dominated by molluscs, especially bivalves, notable amongst which is the species *Maccoyella barklyi*. Gastropods (e.g. *Euspira reflecta*), ammonites and belemnites, bryozoa, a toothplate from a chimaerid fish (*Edaphodon eyrensis*), and remains of marine reptiles (ichthyosaurs and plesiosaurs) are also recorded (Ludbrook, 1966).

Micropalaeontological studies and refinement of biostratigraphic zonation based on spore-pollen, dinoflagellate, microplankton and foraminiferal assemblages, including those of Ludbrook (1966), Scheibnerova (1980), Morgan (1978, 1980b), Dettmann and Williams (1985), Alley (1987, 1988) and Rogers *et al.* (1989), have established the age as early Aptian to mid-Albian (Fig. 9.12), the unit spanning the *C. hughesii* and *Crybelosporites striatus* Zones and extending into the *Coptospora paradoxa* Zone (Helby *et al.*, 1987a). Micro-plankton assemblages indicate *Odontochitina operculata*, *Diconodinium davidii* and *Muderongia tetracantha* Zones (Helby *et al.*, 1987). The equivalent Wallumbilla Formation includes the *C. paradoxa* Zone, and the *Canninginopsis denticulata* and lower part of the *Pseudoceratium ludbrookiae* Zones indicating an age extending to late Albian. Studies on microfossil assemblages (e.g. Dettmann and Williams, 1985; Alley, 1985a,b) suggest the presence of local hiatuses of various ages and duration, diachronous boundaries, and depositional rates that varied from place to place (Fig. 9.19).

Interpretation

The Bulldog Shale is a shelf mud deposited from suspension in an epicontinental sea. Dysaerobic or restricted marine conditions prevailed generally in the early transgressive stage as represented by the basal dark organic shale with its very low diversity of microplankton assemblages (Krieg *et al.*, 1991). However, the presence of locally abundant burrowing suggests aeration for parts of the basin floor. The Wilpoorinna Breccia Member may be a locally developed transgressive deposit representing this early stage. As the sea advanced, conditions changed to an open, oxygenated marine environment, as indicated by the great diversity of fauna and abundance of macrofossils. Coquinoid layers might represent swell lags developed during major storms (Krieg, Rogers *et al.*, 1990), and the sandy units along the western basin margin are also interpreted as storm-related deposits. Possible evidence of an early Aptian sea-level highstand is seen in seismic records (N.M. Lemon, University of Adelaide, pers. comm., 1988) where a weak reflector in about the middle of the unit has been identified. The cause of the change in acoustic impedance is presumed to be a thin layer of incipient compaction, chemical alteration, or increased bioturbation.

The climate during deposition of the Bulldog Shale is thought to have been cooler than for the Cadna-owie Formation. The presence of glendonite nodules indicates very cold water as the parent mineral ikaite is stable only at near-freezing temperatures. An increase in size and abundance of the large limestones near the base could represent a greater development of seasonal river and shore ice able to capture and transport the larger boulders. Diminution of the *Araucaria* plant group (Krieg *et al.*, 1991) at this time may also be consistent with cooler conditions. Slight warming later in the Aptian and early Albian is possibly indicated by the smaller and much less frequent limestones and, as speculatively suggested by Krieg *et al.* (1991), by a mixed plant assemblage with abundant ferns and podocarps.



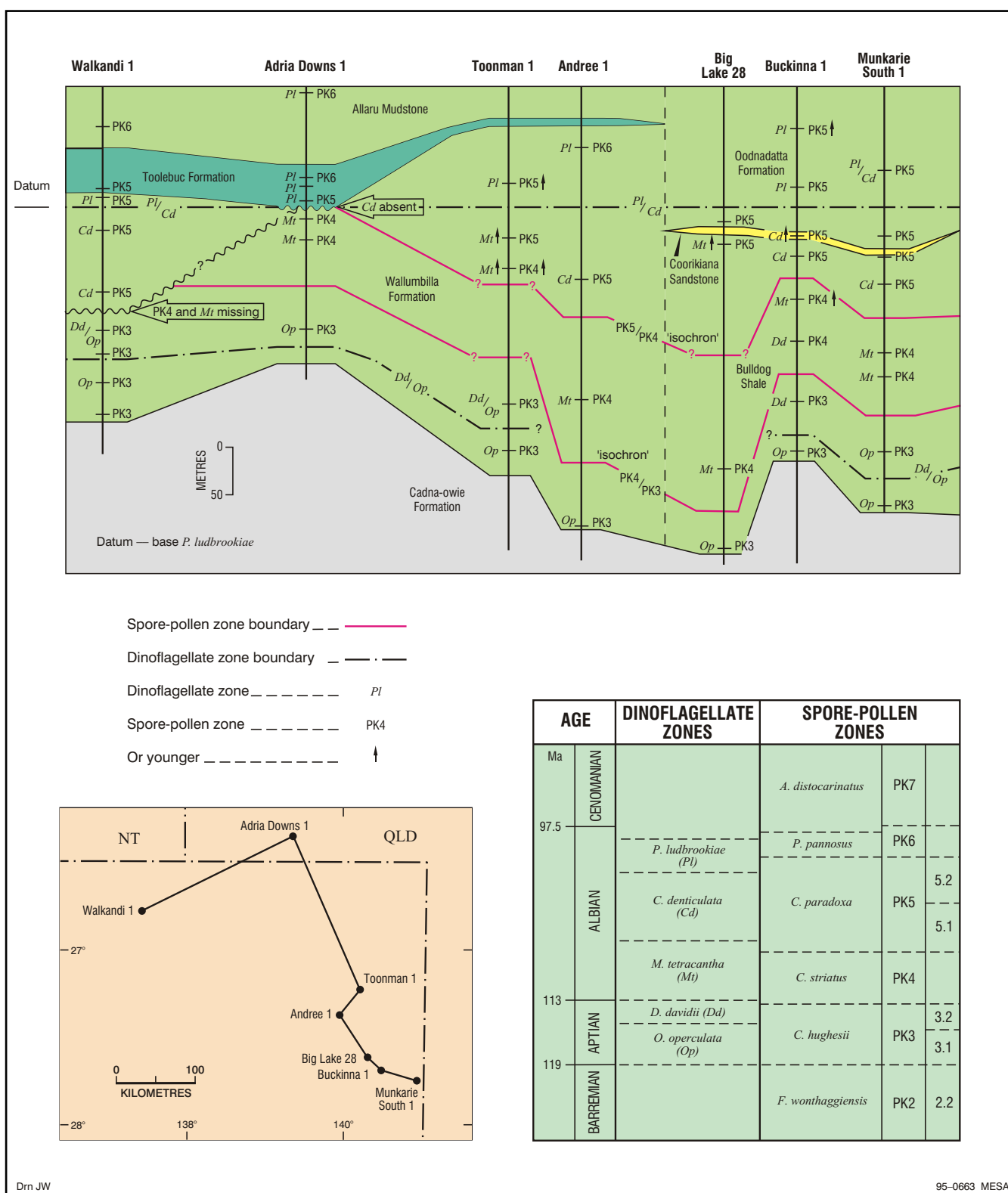


Fig. 9.19 Drillhole correlation of Bulldog Shale and Wallumbilla Formation showing biostratigraphic intervals, hiatuses and variation in sedimentation rate (after Dettmann and Williams, 1985).

1. Layer of ellipsoidal carbonate concretions in dark grey shaley mudstone of the Bulldog Shale, southeast of 'Finniss Springs'. Photo 042476.
2. Large, waterworn Adelaidean quartzite lonestone in the Bulldog Shale. Photo 041931.
3. Boulder beach deposit in Bulldog Shale near 'Moolawatana'. Photo 042477.
4. Silicified, burrowed wood from an area of lag deposits on Bulldog Shale near Lake Phillipson. Photo 042478.
5. Shell hash in carbonate concretion of the Bulldog Shale near Prospect Hill. Photo 042486.
6. Fossil starfish in the Bulldog Shale, Gregory Creek. Photo 042480.
7. Glendonites (calcite clusters pseudomorphing the calcium carbonate hexahydrate mineral ikaite) in the Bulldog Shale near Petermorra Creek. Photo 042481.
8. Cretaceous wave-smoothed surface developed on Palaeoproterozoic granitic gneiss, Lagoon Hill. Photo 041934.

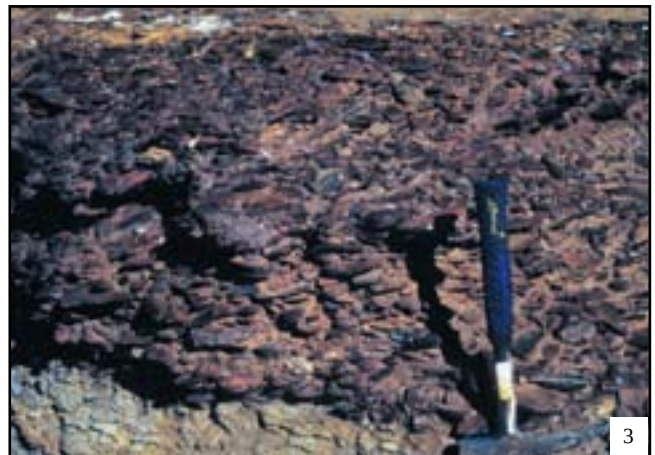
Coorikiana Sandstone

Along the southwestern margin of the Eromanga Basin, the Bulldog Shale is conformably overlain by Coorikiana Sandstone (Thomson, 1980; Moore and Pitt, 1982), formerly a member of the Oodnadatta Formation, with a type locality at Coorikiana Creek (Freytag, 1966; Fig. 9.20). The Coorikiana Sandstone is exposed discontinuously between the South Australia-Northern Territory border and the Marree area, where Forbes (1966) named the unit the Attraction Hill Sandstone Member of the Marree Formation. The unit also crops out, and has been traced seismically in the subsurface, near the northeastern margin of the Flinders Ranges (Sheard, 1993; Sheard and Cockshell *et al.*, 1992). In the north, it can be traced from the basin margin eastwards to the Dalhousie Anticline (Krieg, 1986) but not beyond. From the southern margin, the formation extends northwards in the subsurface to the Cooper region where it attains a maximum thickness of 18 m. In this region Moore *et al.* (1986) designated a subsurface reference section 8 m thick in Dullingari 11, and mapped by gamma-ray log correlation a broad east-west-orientated zone extending at least 200 km through the Moomba area and into Queensland. The Coorikiana Sandstone passes laterally basinwards into marine shale of the upper Wallumbilla Formation. Drillhole sections of Coorikiana Sandstone show an upward-coarsening profile with a sharp, conformable upper boundary and a gradational lower boundary.

The Coorikiana Sandstone is predominantly a calcareous, clayey, very fine to fine-grained sandstone and sandy siltstone with minor siltstone interbeds (Moore and Pitt, 1985). Most outcrops consist of a distinctive yellowish brown, calcareous, very fine to medium-grained sandstone which is micaceous and slightly carbonaceous. Sedimentary structures include small to medium-scale, low-angle cross-bedding, ripple marks, burrows and trails. Petrographic studies of the sandstone revealed pellets of clay and glauconite, and angular grains of quartz and feldspar in a clayey matrix.

On the southern shore of Lake Eyre South, sections of Coorikiana Sandstone display a regressive sequence of subtidal, irregularly

1. Calcareous, bioturbated, argillaceous siltstone, fine-grained cross-laminated sandstone, and thicker bedded blocky sandstone of the Coorikiana Sandstone near 'Umbum' ruins. Photo 042482.
2. Tracks on a rippled bedding surface in calcareous siltstone of the Coorikiana Sandstone near 'Umbum' ruins. Photo 042483.
3. Basal imbricated pebble conglomerate of a gravel bar or beach unit in the Coorikiana Sandstone near Lake Eyre South. Most pebbles are ferruginised mudstone reworked from the Bulldog Shale. Photo 039088.
4. Shark teeth in coarse sandy conglomerate of the Coorikiana Sandstone near Lake Eyre South. Photo 039090.
5. Ammonite (*Myloceras* ?sp.) cast in blocky, calcareous Coorikiana Sandstone northwest of Dalhousie Springs. Photo T024559.



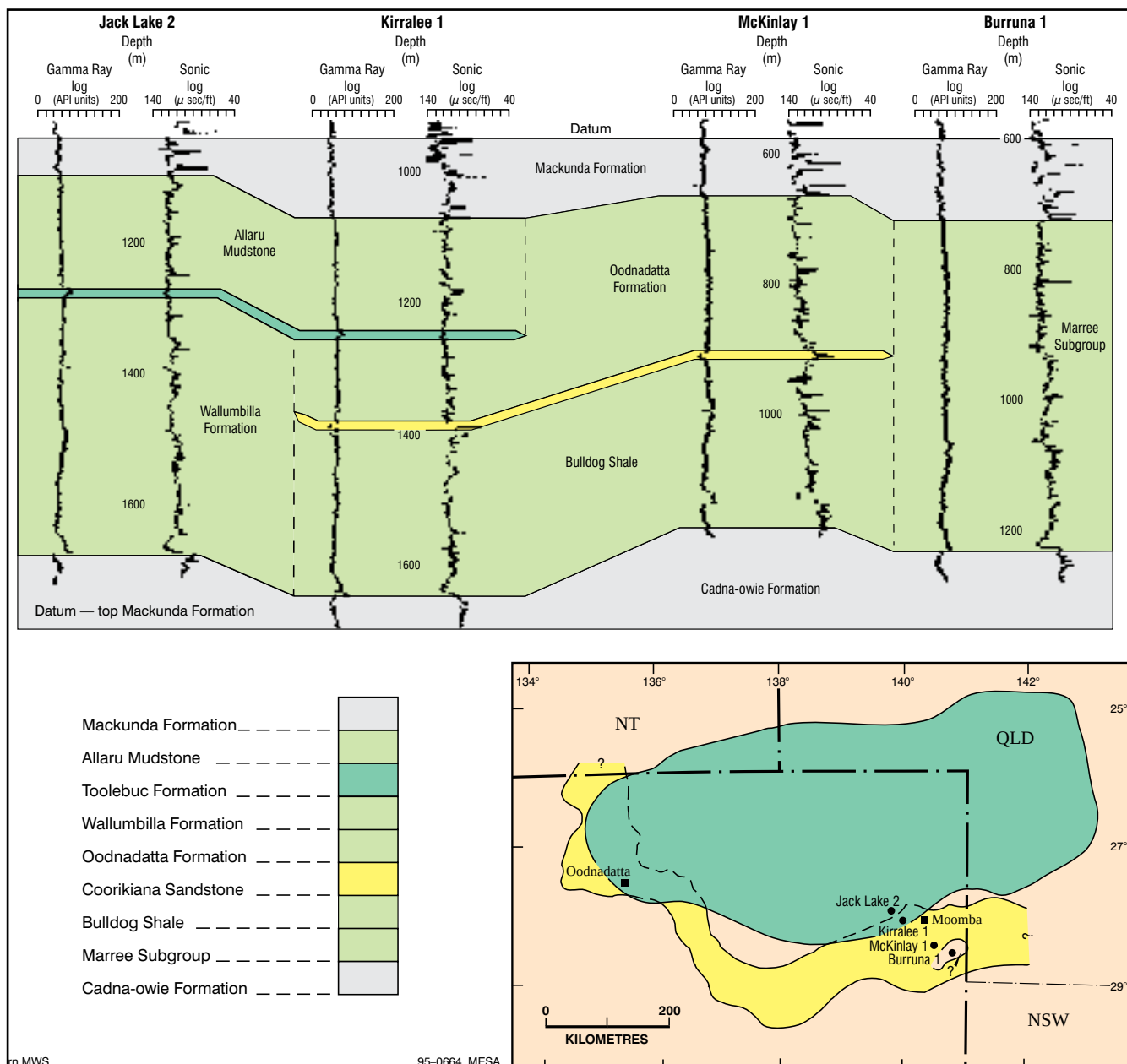


Fig. 9.20 Distribution and drill-log correlation of Coorikiana Sandstone and Toolebuc Formation.

interlayered and bioturbated sand and mud overlain by tidal and supratidal sands with large-scale, low-angle cross-bedding, bi-directional cross-bedding, and planar bedding. Thin conglomerate layers which interfinger with the shoreface sands are interpreted as gravel bars, beaches, and channel deposits which formed at river outlets. The conglomerates are composed mainly of flat, rounded pebbles of concretionary calcareous mudstone (ferruginised in part), fossil wood with *Teredolites* borings, quartz and quartzite clasts, and rare shark teeth in a coarse sandy matrix. Much of the coarse sediment, including mudstone pebbles, was reworked from areas of Bulldog Shale exposed along the basin margin by a fall in sea level. Similar ferruginised conglomerates occur at Attraction Hill near Marree, the type section of the Attraction Hill Sandstone Member.

The Coorikiana Sandstone contains no significant macrofauna. Ludbrook (1966) assigned an earliest Albian age to the unit, on the basis of ammonite faunas in the Oodnadatta Formation and Bulldog Shale. Ludbrook (1966) and Scheibnerova (1980) have

studied foraminiferal microfaunas from Coorikiana Sandstone in Oodnadatta 1 drillhole; Scheibnerova suggested a late Albian age.

Palynofloras from the formation in Oodnadatta 1 and drillholes in the Cooper region can be correlated with the *C. paradoxa* and *C. denticulata* Zones (Moore, Pitt and Dettmann, 1986). Coorikiana Sandstone in Toodla 1 and Alford 1 is placed in the upper *C. paradoxa* Zone (Alley, 1985; Rogers *et al.*, 1989). The dinoflagellate assemblage in Toodla 1 correlates with the *P. ludbrookiae* Zone (Alley, 1985b), whereas the formation in Alford 1 is placed in the *M. tetracantha* Zone (Rogers *et al.*, 1989). The palynofloras of the Coorikiana Sandstone indicate a middle Albian age.

The regressive shoreface deposits of the Coorikiana Sandstone mark a eustatic sea-level fall in the Eromanga Basin in middle Albian time (Morgan, 1980a). The global sea-level curve of Haq *et al.* (1987a) shows a rising trend during the Albian with several minor regressions, but it is not clear which, if any, of these corresponds with the Coorikiana Sandstone.

Oodnadatta Formation and Toolebuc Formation

The Oodnadatta Formation was defined by Freytag (1966), with a type section at Mount Arthur, east of Oodnadatta. A fossiliferous calcareous unit, the Wooldridge Limestone Member, is present in the lower part. The formation originally included the Coorikiana Sandstone and Mount Alexander Sandstone as members at the base and in the upper part of the formation respectively, but these are now regarded as separate formations.

The formation consists of laminated and thin-bedded claystone and siltstone with interbeds of fine-grained sandstone, calcareous and ferruginous concretions, cone-in-cone limestone, and celestite-barite veins. It was deposited in a low-energy, shallow marine environment. Sedimentary structures include planar and lenticular bedding, ripple cross-lamination and bioturbation. The formation reaches a maximum thickness of ~300 m in the Moomba area.

The Oodnadatta Formation contains a varied fauna, including ammonites of middle and late Albian age (Ludbrook, 1966; McNamara, 1980, 1985), bivalves (dominated by the genus *Inoceramus*), belemnites and foraminifera (Ludbrook, 1966). Palynological studies of drillholes in the Cooper region (Moore and Pitt, 1985), and from Toodla 1 (Alley, 1984), Alford 1 and Crowsnest 2 (Rogers *et al.*, 1989), and Muloorina 1 (Alley, 1984), place the unit in the upper *C. paradoxa* and *Phimopollenites pannosus* Zones. The dinoflagellate assemblages correlate with the *C. denticulata* and *P. ludbrookiae* Zones, indicating a middle to late Albian age.

The Wooldridge Limestone Member (Freytag, 1966) at its type section (Wooldridge Creek, northwest of Oodnadatta) comprises ~7.5 m of celestite-bearing, calcareous and sandy siltstone with numerous ellipsoidal limestone concretions which contain a rich fauna of late Albian ammonites (Reyment, 1964; Ludbrook, 1966). The type section is separated from the underlying Coorikiana Sandstone by 30 m of lower Oodnadatta Formation. The member is also recorded in the Oodnadatta Formation type section at Mount Arthur, and possibly in Santos Oodnadatta 1 (Freytag, 1966), although Moore, Pitt and Dettmann (1986) cast doubt on correlation of the interval in Oodnadatta 1 with Wooldridge Limestone Member.

Several authors (e.g. Scheibnerova, 1980; Ozimic, 1986) have correlated the Wooldridge Limestone Member with the Toolebuc Formation (Senior *et al.*, 1975), a widespread unit in the central Eromanga Basin. In South Australia, the Toolebuc occurs in the subsurface east of the Dalhousie Anticline and north of Moomba, reaching a maximum thickness of ~60 m near Poeppel Corner. The Toolebuc Formation usually occurs where the Coorikiana Sandstone is absent, although the two units coexist in a few drillholes, where they are separated vertically by 50–120 m of marine shale (Moore, Pitt and Dettmann 1986). In the central Eromanga Basin, the Toolebuc Formation forms the boundary

between the Wallumbilla Formation and Allaru Mudstone. These two units are near equivalents of Bulldog Shale and Oodnadatta Formation respectively, and differ from them only in that their boundary is at a higher stratigraphic level (Fig. 9.12). The passage from the Wallumbilla-Toolebuc-Allaru sequence into the Bulldog-Coorikiana-Oodnadatta sequence through to Marree Subgroup is illustrated on Figure 9.20.

The Toolebuc Formation consists of laminated black, carbonaceous, clayey mudstone with fish remains, dark grey, calcareous, clayey mudstone or coquinite with abundant *Inoceramus* shells and laminated dark grey, clayey mudstone (Moore, Pitt and Dettmann 1986). The unit also includes oil shale containing organic matter derived mainly from planktonic algae and cyanobacterial mats (Sherwood and Cook, 1986; Glikson and Taylor, 1986). The Toolebuc is characterised by a positive gamma-ray anomaly in drillholes, caused by the uranium content of carbonaceous material and phosphatic fish remains (Ramsden *et al.*, 1982).

In the southern Eromanga Basin, palynological studies place the Toolebuc Formation in the *P. ludbrookiae* and upper *C. paradoxa* Zones (Moore, Pitt and Dettmann 1986). The unit may be slightly diachronous, as it appears to enter the *P. pannosus* Zone further north (McMinn and Burger, 1986). These results indicate a latest middle to late Albian age.

The Toolebuc Formation was deposited at the peak of a marine transgression that followed the middle Albian (Coorikiana) regression (Morgan, 1980a,b). During the highstand, clay and organic matter accumulated in the central Eromanga Basin, beyond the limit of clastic sedimentation. Regular fine lamination of the sediment indicates deposition in quiet, possibly stratified, water. The abundance of organic matter, lack of bioturbation, and general absence of benthic faunas indicate conditions of low oxygenation probably resulting from high phytoplankton productivity. Occasional shell beds developed when conditions were suitable for establishment of benthic faunas of bivalves such as *Inoceramus* and *Aucellina*, that are tolerant of low oxygen concentrations, (Ozimic, 1986). In contrast, the Wooldridge Limestone Member was deposited in shallower and more oxygenated water nearer the basin margin.

The sea-level curve of Morgan (1980a) indicates a relative fall in sea level after the Toolebuc transgression, and this is reflected in microplankton frequencies which show an upward decline in the Oodnadatta Formation (e.g. in Toodla 1). Apart from the Toolebuc transgression, it is likely that relative sea levels during Oodnadatta time were generally lower than for the earlier transgressive phase of Bulldog Shale. This is supported by microplankton frequencies in palynofloras from Oodnadatta Formation in Alford 1 and Crowsnest 2, which are lower than those in the lower Bulldog Shale in Finniss 2 (Rogers *et al.*, 1989). Further evidence of a shallower sea in the middle to late Albian is provided by palaeogeographic studies (Frakes *et al.*, 1987), and by the higher proportion of silt and sand in the Oodnadatta Formation. According to Haq *et al.* (1987a), global sea level during the middle and late Albian was rising, and was higher than in Aptian-early Albian time. This is not reflected in the Eromanga Basin sequence, suggesting that mild tectonic uplift, a reduction in subsidence rate or increased sediment supply may have counteracted a eustatic rise in sea level.

Mackunda Formation

The Mackunda Formation (Vine and Day, 1965) is a sandy, regressive marine unit conformably overlying the Oodnadatta Formation and Allaru Mudstone. The Mackunda includes sediments previously placed in the Mount Alexander Sandstone Member (Freytag, 1966) and the unnamed basal, fossiliferous member of the Blanchewater Formation (Forbes, 1966). Both these names are now superseded.

In the southwestern Eromanga Basin, the Mackunda Formation comprises ~100 m of interbedded, partly calcareous sandstone, siltstone and shale (Moore and Pitt, 1985). On the basin margin, Crowsnest 2 intersected the lower Mackunda Formation consisting of subtidal, bioturbated and irregularly layered, very fine-grained



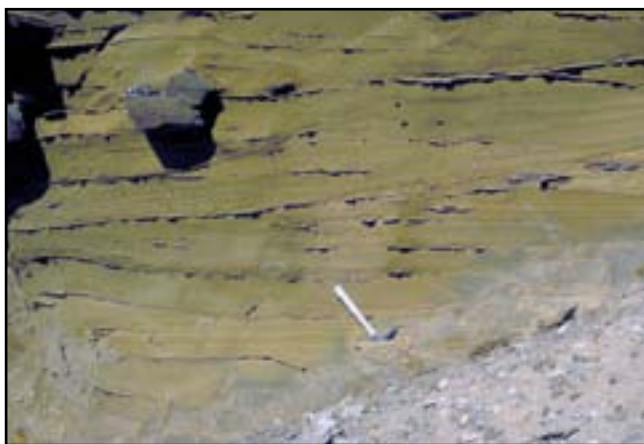
Thinly bedded siltstone and very fine-grained sandstone with tabular carbonate layers of the Oodnadatta Formation, southwest of Dalhousie Springs. Photo 042485.

sand and mud, and shoreface bodies of very fine to fine-grained sand. A thin transgressive unit of fossiliferous marine mudstone and several thin limestone beds are also present (Rogers *et al.*, 1989). A complete section of Mackunda Formation in Clayton 3 (Kwitko, 1986) includes three shoreface sand units with basal erosive contacts, and two transgressive mudstone units.

Shoreface sands in Crowsnest 2 and Clayton 3, and outcrops at Belt Bay, Lake Eyre North, closely resemble the Coorikiana Sandstone. These are micaceous and carbonaceous, very fine to medium-grained sandstones with thin planar and low-angle cross-bedding, clay intraclasts, wood fragments, minor thin clay layers, and tabular calcareous concretions. The sand fraction is composed of angular quartz and feldspar grains, and clay pellets (including smectite and clinoptilolite).

The Mackunda Formation generally contains only rare macrofossils, but bivalves such as *Inoceramus* are locally common, for example in the transgressive mudstone unit in Crowsnest 2. Palynofloras from the Mackunda Formation in Crowsnest 2 and Clayton 3 are correlated with the *P. pannosus* and lower *Appendicisporites distocarinatus* Zones (Alley, 1986a; Rogers *et al.*, 1989). The marine microplankton components of the palynofloras are restricted, but can be placed in the interval spanning the *P. ludbrookiae*, *Xenascus asperatus*, and lower *Diconodinium multispinum* Zones (Alley, 1984, 1986b; Rogers *et al.*, 1989). These correlations indicate a late Albian to earliest Cenomanian age.

The Mackunda Formation is generally regarded as a marginal marine to paralic unit deposited during a single regression, as shown in the sea-level curve of Morgan (1980a). However, alternating shoreface sand, subtidal sand and mud, and marine mudstone indicate that several cycles of regression and transgression occurred during deposition. The regressive-transgressive cycles are superimposed on a general falling trend in sea level during Mackunda time, represented by the transition from marine Oodnadatta Formation and Allaru Mudstone to freshwater Winton Formation. The late Albian-earliest Cenomanian regression in the Eromanga Basin was probably a result of tectonic uplift, as global sea level was slowly rising at that time (Haq *et al.*, 1987a).



Stratigraphy — upper non-marine succession

P.A. Rogers

The Winton Formation and Mount Howie Sandstone represent a final, non-marine phase of deposition in the Eromanga Basin. Up to 1 200 m thick, and with a depositional time span of ~8 million years, these units indicate a much faster average sedimentation rate than that of the underlying sequence.

Winton Formation

The Winton Formation (Whitehouse, 1955) is a sequence of non-marine shale, siltstone and sandstone, with minor coal layers, which reaches a maximum thickness of ~1 200 m in the Patchawarra Trough (Moore and Pitt, 1985). The unit is usually overlain by thick Cainozoic sediments, except near 'Cordillo Downs', and in the Gason and Innamincka Domes, where it is exposed and deeply weathered. Much thinner sections of lower Winton Formation (top eroded) are found along the southwestern margin of the Eromanga Basin, including the equivalent upper Blanchewater Formation of Forbes (1966) in the Marree area.

The Winton Formation in Clayton 3 (Kwitko, 1986) and the Lake Eyre bores (Johns and Ludbrook, 1963) consists of laminated to medium-bedded claystone, siltstone and very fine to fine-grained sandstone, with varying amounts of carbonaceous material. Clay intraclasts are common, and the sandstone layers show small-scale (ripple) cross-bedding. Lamination is irregular and wavy, and is disturbed by penecontemporaneous slump structures, microfaults and bioturbation. Minor thin layers of coal and dolomitic mudstone are present. The sand fraction of the Winton Formation contains a significant proportion of feldspar and ferromagnesian minerals, which were probably derived from a volcanic source.

A sequence, up to 12 m thick, of thin sandy and muddy limestone layers and lenses interbedded with medium-grained sandstone in the upper Winton Formation was referred to informally as the 'Betoota limestone horizon' by Wopfner (1960). This marker horizon is recorded in drillholes near 'Haddon Downs', and is exposed in the core of the Innamincka Dome.

The Winton Formation contains plant fragments, fossil wood and freshwater bivalves, including *Pledgia eyrensis* from Babbage Peninsula, Lake Eyre North (Ludbrook, 1985). Palynofloras are correlated with the *A. distocarinatus* Zone, indicating a Cenomanian age, (Fig. 9.12.), and include megaspores (*Arcellites*, *Balmeisporites*) which were probably derived from heterosporous aquatic ferns such as *Azolla*.

The Winton Formation was deposited in a low-energy fluvial to lacustrine environment consisting of sluggish streams meandering through broad flood plains dominated by lakes and marshes. Coal seams and associated highly carbonaceous clays resulted from the accumulation of plant matter in swamps and lakes. Carbonate layers may have been deposited in response to lacustrine plant growth and dolomitised subsequently by magnesium-rich groundwater. Active volcanoes in eastern Australia are thought to have contributed the volcanic component of the Winton Formation.

Uplift of the Eromanga Basin region, combined with a rapid influx of sediment, resulted in withdrawal of the sea at the beginning of the Cenomanian at a time when global sea level was rising to a maximum (Haq *et al.*, 1987a). However, in central parts of the basin, rates of subsidence and sediment accumulation appear to have increased during deposition of Winton Formation, and considerable thicknesses occur in troughs which were actively subsiding during Late Cretaceous time.

Top: Possibly bi-directional (?tidal influence) large-scale cross-bedding in calcareous fine-grained argillaceous sandstone of the Mackunda Formation, Belt Bay. Photo 042486.

Bottom: Tabular calcareous concretion within sandstone for the Mackunda Formation, Belt Bay. Photo 039091.

Mount Howie Sandstone and equivalents

The Winton Formation marks the final phase of sedimentation in the Eromanga Basin, apart from localised younger, fluvial channel infills. Wopfner (1963) named channel deposits from the 'Cordillo Downs' area the Mount Howie Sandstone, which consists of up to 45 m of large-scale cross-bedded, medium to coarse-grained quartz sandstone interbedded with shale and shale conglomerate. The contact with Winton Formation is a sharp and irregular erosional disconformity. At its type section, the Mount Howie Sandstone contains a fossil flora (*Brachyphyllum* and cycads) which Wopfner (1963) regarded as having Mesozoic affinities, suggesting a Turonian or younger age.

Outcrops tentatively assigned to the Mount Howie Sandstone occur in the Innamincka and Gason Domes, in the Dalhousie Anticline (Krieg, 1986), and in the Marree area where Forbes (1972) noted that they lie stratigraphically between the Winton Formation and Early Tertiary Eyre Formation.

Fine to very coarse-grained sand at the top of the Cretaceous succession in Muloorina 1 contains a palynoflora that is tentatively assigned to the *Phyllocladites mawsonii* Zone of Turonian to Coniacian age (Fig. 9.12, Alley, 1984). Balme (*in* Johns and Ludbrook, 1963) suggested a Turonian to Senonian age, on scanty palynological evidence, for clay, silt and fine to medium-grained sand conformably overlying Winton Formation in Lake Eyre Bore 20.

Uplift at the end of the Cenomanian caused a change in depositional regime from a widespread, low-energy, fluvial-lacustrine environment (Winton Formation) to localised, high-energy, fluvial channels (Mount Howie Sandstone and equivalents). Erosion of the Winton Formation provided a source for fine-grained sediment and shale clasts in the Mount Howie Sandstone, while the coarser sand fraction was probably reworked from older Mesozoic units (e.g. Algebuckina Sandstone) exposed on the basin margin. During the Late Cretaceous, the now-upwarped Eromanga Basin underwent further erosion and weathering (Senior and Mabbutt, 1979). Erosion products are interpreted to have been carried southwest (Veevers, 1984) towards the central part of the Australian Southern Rift System where a thick, Late Cretaceous sedimentary succession occurs.

Economic geology

Petroleum

E.M. Alexander

Petroleum exploration commenced in the 1950s when licences covering the Cooper and Eromanga Basins were acquired by Santos, (Sprigg, 1958). ignoring conventional wisdom that commercial accumulations of oil would not be found in Mesozoic formations within the Great Artesian Basin. The first petroleum drillhole was completed in 1959, and Cooper Basin gas was discovered in 1963. The petroleum exploration history of the area has been presented by Sprigg (1986), Moore (1986b) and Hollingsworth (1989).

In 1976, Namur 1 flowed gas from the Namur Sandstone at $3.9 \times 10^5 \text{ m}^3$ (13.9 million cubic feet) per day, the first significant flow of hydrocarbons from a Mesozoic reservoir in South Australia (it was found later that the gas was from a Permian source). The first oil flow from the Eromanga Basin was recorded in 1977 from the Poolowanna Formation in Poolowanna 1. This proved uneconomic but commercial oil was discovered in 1978 in Strzelecki 3 with a flow of 382 000 L (2 400 barrels) of oil per day from the Hutton Sandstone.

Prior to these discoveries it was widely believed that any Jurassic oil and gas had been flushed out by movement of artesian water. Bowering (1982) postulated that the hydraulic gradient during the Plio-Pleistocene may have been stronger and partially flushed some hydrocarbons from accumulations, but concluded that the modern hydraulic gradient is insufficient to flush hydrocarbons.

Since the discoveries in the late 1970s, oil has been recovered from almost every unit in the lower non-marine sequence. The Hutton Sandstone has proved to be the most productive unit (Fig. 9.21) with good to excellent reservoir characteristics (Gravestock and Alexander, 1988); approximately two-thirds of oil discovered in the Eromanga Basin is reservoir in this (Watts, 1987). Other significant reservoirs include the Namur Sandstone and lenticular sands within the Poolowanna, Birkhead and Murta Formations, and McKinlay Member. Oil and gas have been recovered from the Coorikiana Sandstone, and economic gas reserves have been discovered in the Marabooka Field. The Algebuckina Sandstone has good to excellent reservoir properties.

Siltstone and shale of the Poolowanna, Birkhead and Murta Formations act as seals in the Cooper region. In the Poolowanna Trough, siltstone and shale in the Poolowanna and Cadna-owie Formations may also form seals. Potential seals elsewhere in the basin include the Cadna-owie Formation and Bulldog Shale.

Exploration activity in the Eromanga Basin has concentrated on the Cooper region (Fig. 9.22), as the origin of oil in Eromanga reservoirs is now widely regarded as the result of vertical migration from Permian source rocks (Heath *et al.*, 1989). However, long-distance lateral migration towards the basin margin has also been proposed (McKirdy and Willink, 1988) and units within the Eromanga Basin have some source potential (Smyth *et al.*, 1984; Jenkins, 1989). Lower non-marine sequence source rocks are sufficiently organic rich and oil mature in the Cooper region to generate hydrocarbons, but suffer from low yield (Jenkins, 1989). Michaelson and McKirdy (1989) concluded that Murta Formation

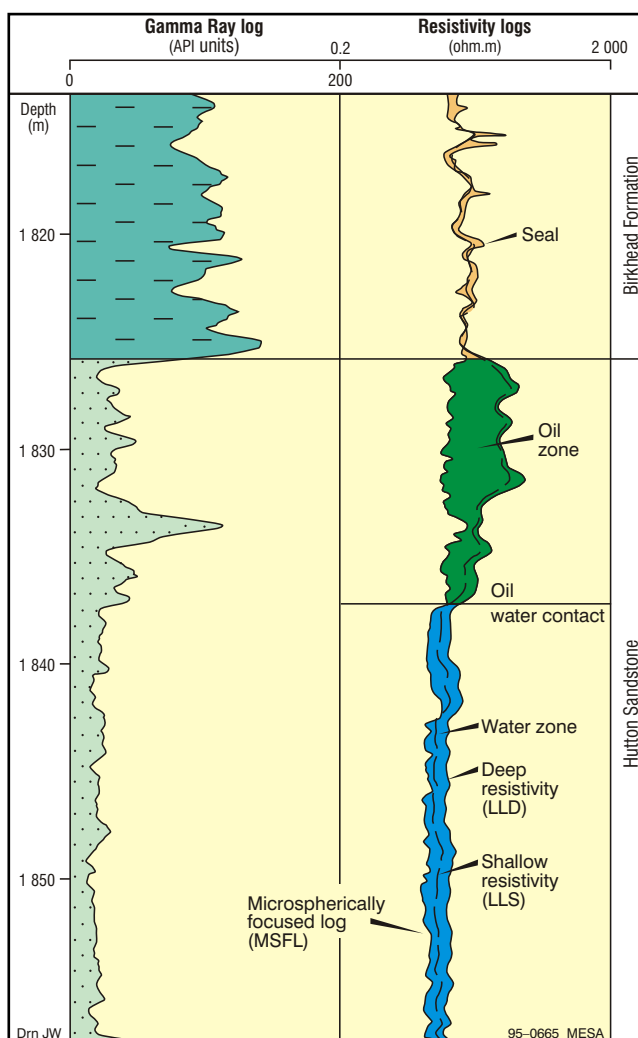


Fig. 9.21 Drill-log anomaly indicating oil in Gidgealpa 17 Hutton Sandstone reservoir.

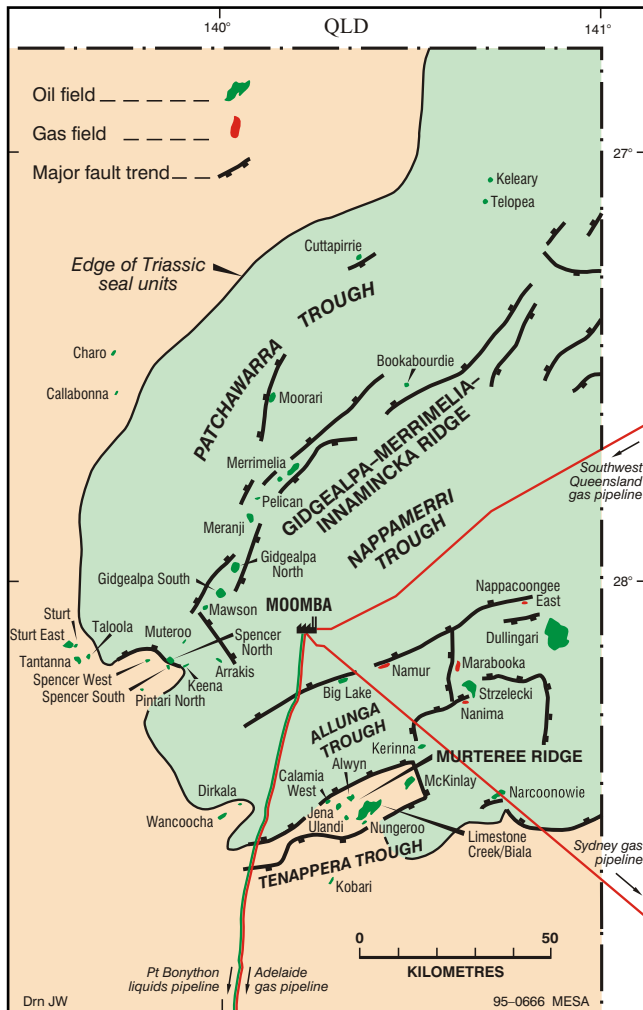


Fig. 9.22 Eromanga Basin oil and gas fields in relation to major faults.

oil was generated from shale rich in organic matter within the formation. Oil-prone source rocks are also reported from the Birkhead Formation (Cook, 1986). Rich source rocks from the Poolowanna Formation in the Poolowanna Trough have also been reported (Moore, 1986b). The marine sequence has some source potential but is of low organic maturity (Kanstler *et al.*, 1986).

Trapping mechanisms for Eromanga Basin oilfields are dominantly structural (anticlines with four-way dip closure or



Moomba petroleum processing facility. The plant receives raw gas, condensate and oil from producing wells, and separates gas from liquids. Purified gas is piped to Adelaide (782 km) and Sydney (1 299 km). The liquids are combined and piped 659 km to Port Bonython on Spencer Gulf for shipping. Courtesy of Santos Ltd. Photo 038701.

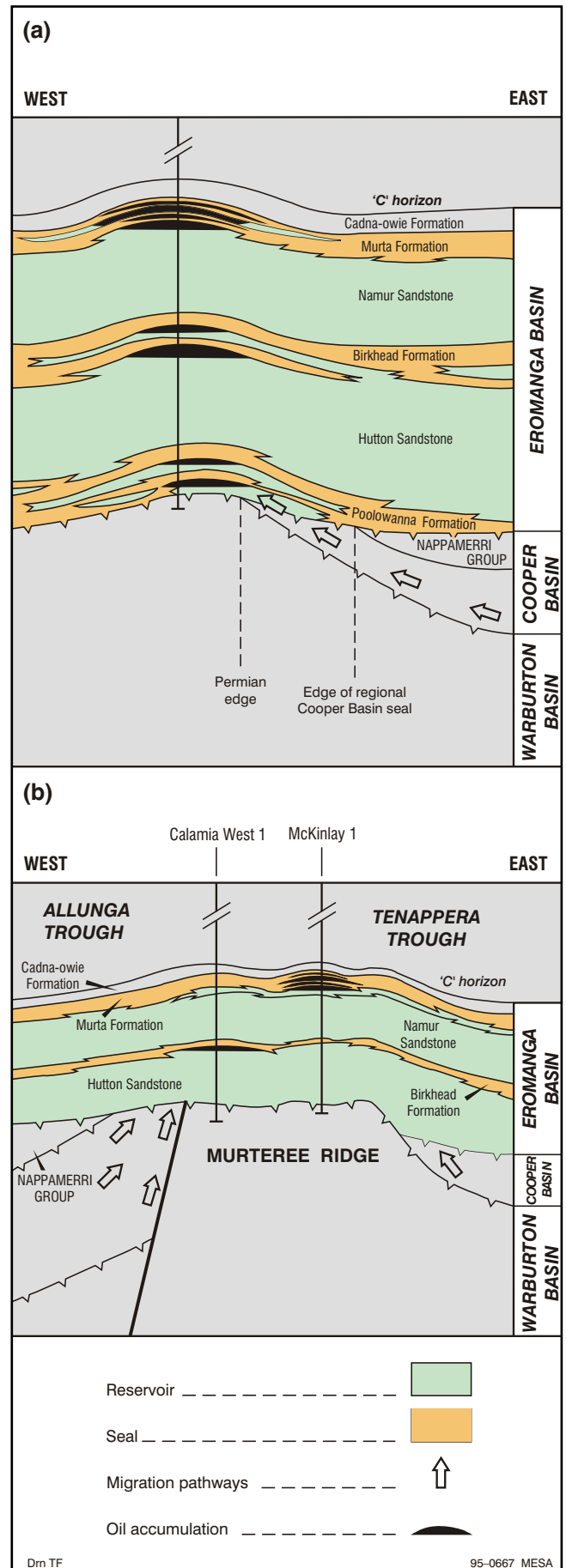


Fig. 9.23 Schematic Eromanga Basin sections showing (a) hypothetical petroleum trap sourced primarily from the Cooper Basin and (b) petroleum traps over the Murteree Ridge.

drapes over pre-existing highs) with a stratigraphic component (at the Hutton-Birkhead and Murta-McKinlay-Namur contacts, and within the Poolowanna and Murta Formations). Onlap and pinchout traps against Warburton Basin highs are possible (Fig. 9.23). Eromanga structures in South Australia are not filled with oil to spill — net oil columns are relatively thin compared to the area under closure.

Since 1959, over 1 000 drillholes have penetrated the Eromanga Basin sequence and over 80 000 km of seismic acquired. As at 1 January 1993, 39 oil and 72 gas fields were on-line from the Cooper and Eromanga Basins (Hibburt, 1993).

Groundwater

D. Armstrong and R. Aldam

The Great Artesian Basin (GAB) is one of the world’s largest groundwater basins, covering an area of ~1.7x10⁶ km². It is a multi-aquifer system comprising Jurassic and Cretaceous sediments of the Eromanga, Surat and Carpentaria Basins, and occupies large portions of Queensland, New South Wales, South Australia and the Northern Territory.

In South Australia, the GAB sequence is composed entirely of Eromanga Basin sediments forming two major confined aquifers (Fig. 9.24). The most important is the lower confined aquifer, which consists of sand, silt and gravel of the Algebuckina Sandstone and Cadna-owie Formation. To the east of the Birdsville Track Ridge this aquifer thickens considerably and includes sediments of the Namur Sandstone, Murta Formation, Birkhead Formation and Hutton Sandstone. Aquifer thickness ranges from <50 m around the western margin to >500 m in the deeper parts of the basin (Fig. 9.25a). The upper confined aquifer consists of sediments of the Winton and Mackunda Formations, and is overlain and confined by Tertiary sediments of the Lake Eyre Basin. It is separated from the lower confined aquifer by marine shale of the Oodnadatta Formation and Bulldog Shale. The Coorikiana Sandstone occurs between these shaly units and forms a discrete aquifer of high salinity and low yield. The upper confined aquifer is not as important in resource terms as the lower aquifer because water quality is poor and supplies are less easily obtained.

The lower confined aquifer is artesian throughout much of the basin, including large areas of South Australia. Recharge occurs

GREAT ARTESIAN BASIN (South Australia) HYDROSTRATIGRAPHY	
Quaternary sediments	Watertable (unconfined aquifer)
Tertiary of the Lake Eyre Basin	
Winton Formation	Upper confining bed
Mackunda Formation	Upper confined aquifer
Oodnadatta Formation Coorikiana Sandstone Bulldog Shale	Lower confining bed
Cadna-owie Formation Algebuckina Sandstone Murta Formation Namur Sandstone Westbourne Formation Adori Sandstone Birkhead Formation Hutton Sandstone	Main artesian aquifer

Fig. 9.24 Hydrostratigraphy of the Great Artesian Basin in South Australia.

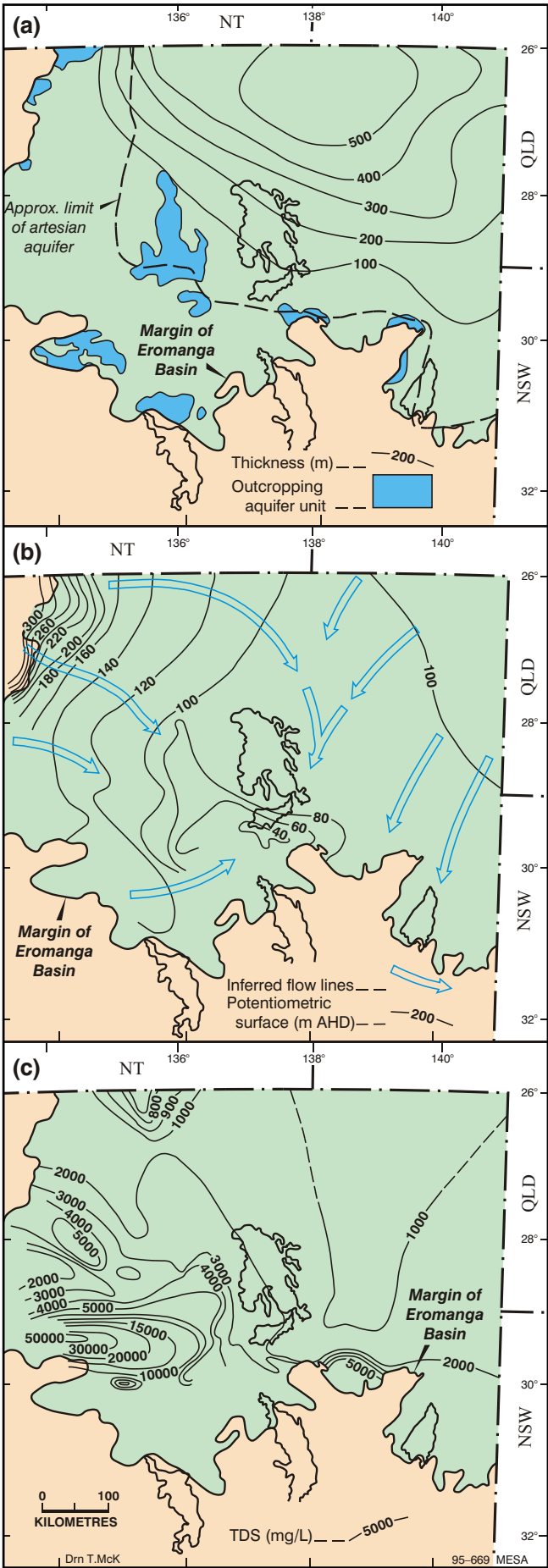


Fig. 9.25 Aquifer properties of the Great Artesian Basin in South Australia: (a) effective thickness of the aquifer (b) potentiometric surface corrected to 15°C, and inferred flow lines (c) total dissolved solids.

by direct infiltration of rainfall through outcropping aquifer units in The Great Dividing Range in central Queensland. Minor recharge also occurs around the southern and western margins (Fig. 9.25b). Flow velocities are very low, with time taken for water to travel ~1 000 km from the recharge to discharge zone estimated in the range of 10^5 – 10^6 years on the basis of hydraulic gradient, conductivity and porosity, and 2.5×10^6 years from isotopic studies (Habermehl, 1980). Recharge to the GAB is estimated to be ~36 m³/sec or 2.5×10^6 ML/year (Habermehl, 1980), of which ~14% is discharged in South Australia. The temperature of waters in the deeper parts of the aquifer is often at or near 100°C.

Discharge of aquifer waters occurs by abstraction through flowing and pumped drillholes, flow from mound springs, and upward-vertical leakage through the confining beds. Measured or estimated discharge figures for the South Australian portion of the GAB are as follows:

	ML/year
Flowing drillholes	47 450
Springs	
Dalhousie	20 500
Others	4 380
Cooper Basin	
product and process water	7 300
Olympic Dam Project	
township and industrial water	5475
Total	85 105

The volume of water lost from the aquifer by upward leakage in South Australia is difficult to quantify, but water balance calculations indicate it is likely to be ~2.3 m³/sec (72 500 ML/year).

Chemistry of groundwater in the South Australian portion of the GAB differs markedly depending on the area of origin. Waters originating from the Great Dividing Range in Queensland are typically fresh, and characterised by a predominance of sodium chloride/bicarbonate, whilst those originating in the west typically contain sodium chloride/bicarbonate/sulphate, and are very corrosive. Total dissolved solids (TDS) show a consistent pattern in the artesian part of the basin, ranging from ~1 000 mg/L near the State borders to >2 000 mg/L near the discharge area (Fig. 9.25c). A broad plume of <1 000 mg/L TDS water extends south along the Birdsville Track, suggesting shorter residence time or variation in upward leakage and mixing from deeper aquifers. The sub-artesian area (western margin) water quality is variable, with salinity ranging from ~2 000 to >20 000 mg/L, which may result from contact with saline formations on infiltration, evaporitic discharge and slow movement.

Groundwater of the GAB is a vital resource used by pastoralists, mining industry and tourism. It is fundamental in sustaining natural ecosystems of the mound springs and has strong associations with Aboriginal and early colonial heritage. It is a resource of major significance for the future of South Australia.



Uncontrolled flow of hot artesian water at Purni Bore prior to rehabilitation. Photo 042487.

BERRI BASIN

P.A. Rogers

The Early Cretaceous basin underlying the Tertiary Murray Basin is here named the Berri Basin and regarded as a former extension of the Eromanga Basin. Early Cretaceous sediments occur in the Tararra and Wentworth Troughs, and it is likely that a connection existed between these troughs and the Eromanga Basin through the Menindee and Blantyre Troughs and the Wilcannia area, although later uplift and erosion has removed Mesozoic sediments from these areas.

Sediments of Aptian-middle Albian age disclosed by drilling in the Murray Basin region (Fig. 9.26) have been named Monash Formation, divided into the Pyap, Merreti and Coombool Members, with a subsurface type section in North Renmark 1 (Thornton, 1972; Fig. 9.27). Similar sediments are known from adjacent parts of Victoria and New South Wales. Lawrence (1972) named the succession in northwestern Victoria the Millewa Group, comprising the Taparoo Sandstone (equivalent to Pyap Member) and Morkalla Formation (equivalent to Merreti and Coombool Members).

This account of Early Cretaceous units is based largely on detailed descriptions by Thornton (1974), but seven additional holes have been drilled in the Berri Basin in South Australia; Cooltong 1 (Archer, 1988), Maggea 1, Nanyah 1, Oak Bore 2, Overland Corner 1 (Furr, 1984), Taladra 1 and Taplan 1. Nanyah 1, Oak Bore 2 and Overland Corner 1 disclosed Monash Formation in the eastern Canegrass Lobe.

The Monash Formation unconformably overlies late Palaeozoic Urana Formation (O'Brien, 1986), Neoproterozoic to early Palaeozoic basement and possible Devonian sandstone, and is unconformably overlain by the Late Paleocene to Early Oligocene Renmark Group. The Early Cretaceous sequence overlies ?middle to late Palaeozoic depositional structures, namely the Renmark Trough, Canegrass Lobe and Paringa Embayment. The distribution of the Pyap Member appears to broadly follow these structures. The two younger members of the Monash Formation, however, are more widespread and overlap older sediments and intervening basement ridges onto adjacent basement highs (Fig. 9.26). The Early Cretaceous section is thickest in the Renmark Trough (up to 600 m on seismic evidence).

Monash Formation

The Pyap Member consists of medium to very coarse-grained pebbly quartz sandstone and quartz-lithic sandstone, with minor interbeds of claystone, siltstone and clayey fine sandstone. The sandstones are commonly kaolinitic and micaceous, and more carbonaceous with thin coal layers in the lower part of the unit; pyrite nodules are present in the upper part. The sandstone in Cooltong 1 contains angular to subrounded clasts to 70 mm, mainly of quartz (Archer, 1988). A core from Monash 1 showed the sandstone to be extensively cross-bedded. The fine-grained interbeds are frequently micaceous and carbonaceous, pebbly in places, and are glauconitic and bioturbated in the upper part of the unit. Dinoflagellates (Nadda 1), acritarchs (Cooltong 1), abundant agglutinated foraminifera (North Renmark 1) and rare shell fragments (Berri North 1) have been recorded in the upper Pyap Member (Thornton, 1974). The Pyap Member is interpreted as a transgressive unit passing up from non-marine fluvial sands to near-shore marine deposits dominated by sandbars and other sand bodies (Thornton, 1974).

Paten and Price (*in* Derrington and Anderson, 1970) correlated microfossil assemblages from the Pyap Member in Nadda 1 and Sunset 1 with the *C. hughesii* Zone of Aptian age. Owen (*in* Archer, 1988) obtained palynofloras from lower Merreti Member and Pyap Member in Cooltong 1, which were correlated with the upper *C. striatus* Zone, implying an early Albian age; however, this is anomalously young when compared to age determinations from units in other drillholes.

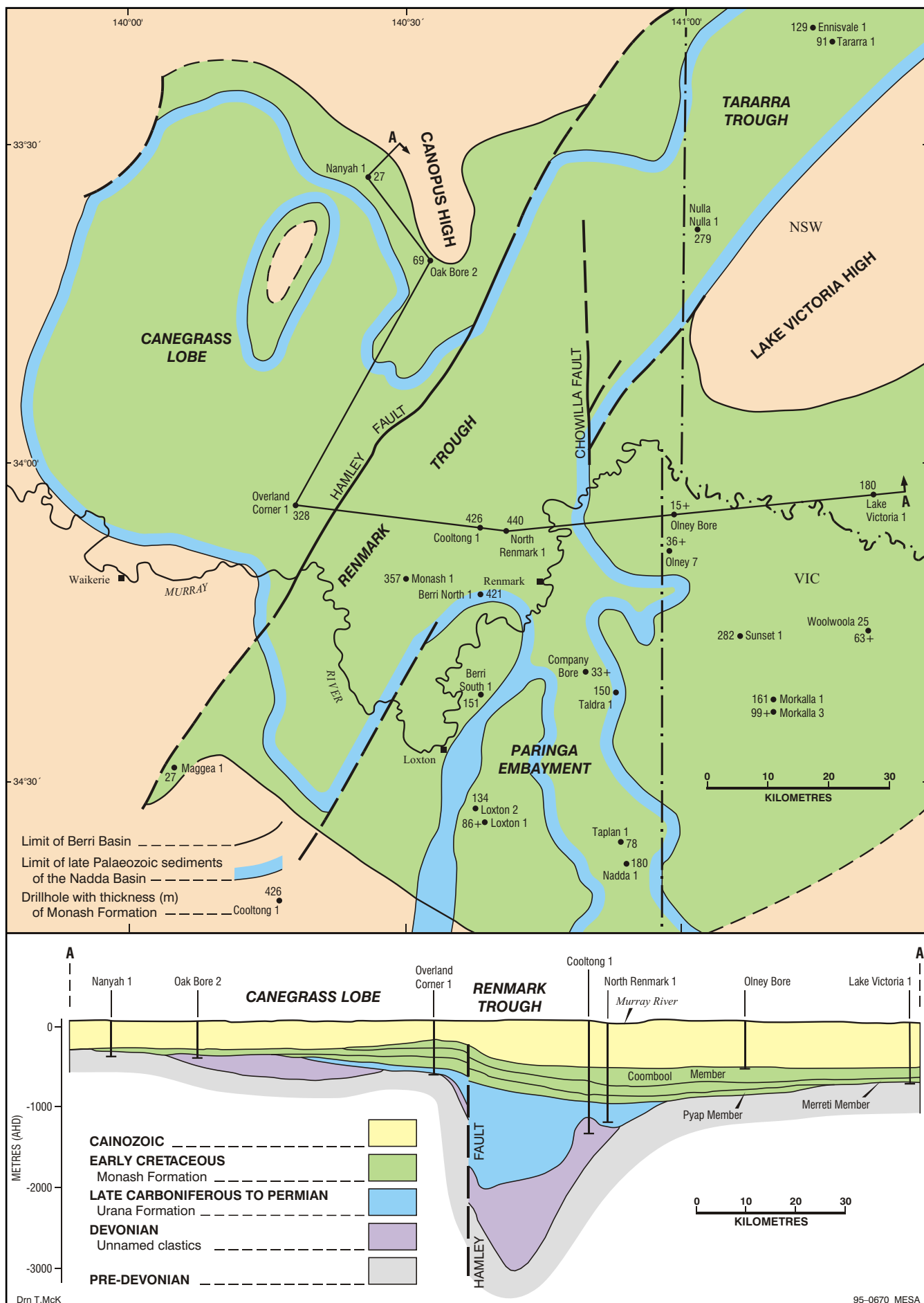


Fig. 9.26 Location, structure and geological cross-section of the Berri Basin.

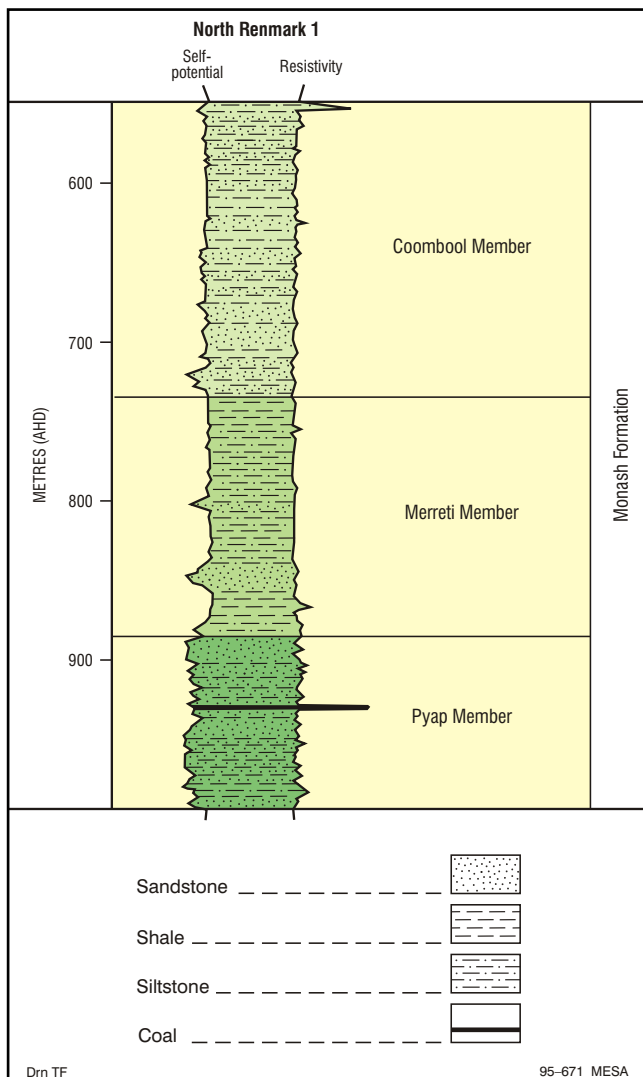


Fig. 9.27 Subsurface type section of the Monash Formation (after Thornton, 1972).

The Merreti Member consists predominantly of claystone and siltstone, with interbeds of very fine to coarse-grained sandstone. The claystone and siltstone are sandy in places, and contain varying amounts of glauconite, pyrite, carbonaceous fragments, mica and clay intraclasts; pebbles and cobbles are rare. Disturbed bedding and bioturbation (including vertical burrows) are features of the fine-grained units. The sandstone interbeds are mainly clayey quartz and quartz-lithic varieties, locally cross-bedded, with glauconite, mica, carbonaceous fragments and clay intraclasts; pebbles are rare in the coarse layers.

The Merreti Member is a marine unit, as indicated by the presence of glauconite, foraminifera and dinoflagellates. The fauna also includes molluscs, fish and *Lingula*. Bivalves are common in the upper part of the unit in Overland Corner 1, and include a specimen referable to '*Yoldia*' *freytagi* (J.G.G. Morton, SADME, pers. comm., 1984). A depositional environment of quiet, moderately deep water is indicated.

Microfloral assemblages from the Merreti Member have been placed in the upper part of the *C. hughesii* and *C. striatus* Zones indicating a late Aptian to early Albian age (Thornton, 1974). The Merreti Member in Maggea 1 yielded an assemblage of spores and pollen correlative with the *C. hughesii* Zone, and dinoflagellates that indicate a correlation with the *O. operculata* Zone (Alley, 1989a). This suggests that only the lower (late Aptian) part of the Merreti Member is present in Maggea 1.

The Coombool Member in most holes is predominantly claystone and siltstone, with minor interbeds of sandstone. In Nadda 1, claystone, siltstone and sandstone are in equal proportions and, in North Renmark 1, the unit is mainly siltstone and sandstone with minor claystone interbeds. The fine-grained sediments are pyritic in places, and contain traces of glauconite and carbonaceous material. The sandstone is generally very fine to fine-grained, although medium to very coarse-grained layers were encountered in Overland Corner 1. Cross-bedding was observed in core from Monash 1. Petrographic examination of sandstone samples revealed a large lithic component (50-60%), mainly of chert with some volcanic grains. Many of the lithic grains are chloritised. The Coombool Member sandstones are largely derived from a volcanic source, unlike the quartzose sandstones of the Pyap and Merreti Members (Evans and Hawkins, 1967).

The Coombool Member is regarded as a fluvial-lacustrine unit, although the presence of glauconite may indicate a marginal marine influence in the lower part. The lack of foraminifera and dinoflagellates, and the presence of *Arcellites* spp. megaspores, are evidence of freshwater conditions (Thornton, 1974). Microfloral assemblages from the Coombool Member have been placed in the uppermost *C. striatus* and *C. paradoxa* Zones (Thornton, 1974), indicating an early to middle Albian age.

Correlation and interpretation

Equivalents of the lower parts of the Eromanga Basin succession (Algebuckina Sandstone and Cadna-owie Formation) are not present in the Berri Basin. However, the lower fluvial part of the Pyap Member may be equivalent to the fluvial Mount Anna Sandstone Member of the Cadna-owie Formation.

The marine upper Pyap Member can be correlated with the lowermost part of the Bulldog Shale. At this time, the sea in the Berri Basin was considerably shallower than in the Eromanga Basin as indicated by the much sandier lithofacies. The Merreti Member represents the peak of marine transgression in the Berri Basin, and is equivalent to the middle part of lithologically similar Bulldog Shale; these two units together represent a vast area of epicontinental marine mud sedimentation.

Uplift or cessation of subsidence in early Albian time caused the sea to withdraw from the Berri Basin, and a freshwater environment was established in which the Coombool Member was deposited. Further uplift in the middle Albian ended Cretaceous sedimentation in the Berri Basin. Uplift and erosion later removed a significant proportion of the Early Cretaceous sequence from the margins; the Coombool Member is either reduced in thickness (Nadda 1, Loxton 2) or missing (Maggea 1, Nanyah 1, Oak Bore 2).



Mollusc referable to '*Yoldia*' *freytagi* from the Merreti Member at 358.1 m in Overland Corner 1. The fossil is 20 mm across. Photo 042488.

JURASSIC-CRETACEOUS RIFT BASINS

Although a thick Mesozoic sedimentary succession in the coastal region of southeastern South Australia had been known for many years (e.g. Glaessner and Parkin, 1958), its significance for the entire southern coastal margin became clear only following offshore seismic and drilling programs, and research on continental rifting. These revealed very thick sedimentary successions with fault-controlled geometry which are now interpreted as a rift-valley fill of continental dimensions. This continental rift is here named the Australian Southern Rift System (ASRS), the central portion of which is occupied by the South Australian rift basins. A useful synthesis, on which this summary is based, is provided by Stagg *et al.* (1990).

In the Jurassic and Cretaceous, tensional stress across the Gondwana supercontinent led to the divergence and eventual parting of the Australian and Antarctic plates. Crustal strain prior to breakup occurred in two phases, a 300 km northwest-southeast extension which developed large transform fault zones and lasted until the Neocomian, and a later northeast-southwest extension from the Neocomian to the Cenomanian (Stagg *et al.*, 1990). At this latter time, ~95 Ma, the Australian and Antarctic plates physically parted along a major crustal rupture, the conjugate edges of which became the continent-ocean boundaries of Australia and Antarctica. Mantle melt emerging from the rupture continuously created new oceanic crust as the gap between the diverging plates widened and the area of new crust expanded by seafloor spreading. While the details of plate kinematics are unresolved, plate separation is commonly assumed to have been synchronous along the rift (e.g. Veevers and Eittreim, 1988), but progressive rupture between 110 and 86 Ma, 'unzipping' the two continents from west to east, has also been suggested (e.g. Mutter *et al.*, 1985). Evidence from South Australia suggests major subsidence in the central portion of the ASRS (i.e. Ceduna Depocentre of Veevers, 1984) to accommodate up to 11 km of sedimentary succession. Palaeocurrent data from the Otway Basin (Medwell, 1977; Gravestock *et al.*, 1986) indicate a westerly transport direction during the Early Cretaceous into a central lacustrine to restricted marine setting in the Eyre Sub-basin (Western Australia). Establishment of this depocentre may have resulted from initial sea-floor spreading in the central portion of the rift system, with diffuse extension at the eastern and western extremes.

Plate detachment may have been a two-layer process in which a lower, Australian layer was 'pulled out' from beneath an upper, Antarctic layer. This model, based on magnetic basement studies, may explain the presence of a broad 'magnetically quiet zone' that extends along much of the rift in a band 100-200 km wide, landwards from the continent-ocean boundary. It is thought that the crust here may be of hybrid oceanic-continental material, perhaps with a more oceanic composition in the west.

The variation in stress history along the 1 200 km South Australian rift interval has resulted in a rift system highly complex in detail and varied along its length. The outer margin is taken to be the continent-ocean boundary. The inner margin (or 'headwall') is a major monocline or fault zone extending from the offshore area in the northwest, co-linear with the northern boundary fault of the Poldas Basin, then offset by a series of transform faults to the Otway Basin onshore in the southeast. Seismic records show this margin as a pronounced escarpment along the down-thrown Gawler Craton. Landward of the headwall boundary, pre-Mesozoic basement is shallow and the sedimentary cover platformal.

Initial extension may have begun before the Jurassic (Stagg *et al.*, 1990) but the oldest direct evidence is in rift-related rocks of mid- to Late Jurassic age; these rocks are associated with latitudinally trending faults aligned *en echelon* in a northwest-southeast direction along the South Australian coastline (Fig. 9.28). From the northwest, these structures are:

- the mid- to Late Jurassic fluvial and lacustrine syn-rift sediments of the Poldas Basin
- the mid-Jurassic Wisanger Basalt north of the Cygnet and Snelling Faults on Kangaroo Island
- the Late Jurassic to Early Cretaceous Casterton Formation of the Otway Basin.

It is possible that these three systems are early tensional effects adjacent to the large, northwest-trending transform movements that accommodated the first phase of extension. Analogous syn-rift deposits are thought to be represented in the Bight and Duntroon Basins by intervals of pre-Cretaceous sediments imaged on seismic sections. The Poldas Basin and Cygnet and Snelling Faults are likely to be reactivated structures; the Poldas Basin was initiated as an intracratonic rift in the Neoproterozoic (see Vol. 1 Ch. 6) while the Cygnet Fault was probably active during the Cambrian (see Ch. 7).

Rift-system history subsequent to initial extension is reflected in the two successive tectono-sedimentary regimes that correspond to Early Cretaceous extension and Late Cretaceous breakup and drift. Extension established the broad geometry of the basins, controlling their locations between major accommodation zones. It also caused basement-involved, down-to-basin normal faulting, block rotation, and associated rapid subsidence that produced graben and half-graben structures where thick, partly volcanogenic, under-compacted sediments accumulated rapidly. As well, a discontinuous horst system was formed, dividing the basins into inner and outer zones. Within the inner zones, back-to-coast fault systems extend landwards from the dividing horsts.

Cenomanian breakup was accompanied by the formation of an outer continental-margin ridge and associated rim basin. As drift set in, thermal subsidence became significant although faulting continued and sediment accumulation was still rapid. Clastic supply became more quartzose as volcanism diminished, and depocentres shifted away from the continent towards the rim basin, the inner margin of the rift system becoming a clastic bypass area. The accumulation of thick Late Cretaceous sediments caused sediment loading of the under-compacted, overpressured, Early Cretaceous deposits. Assisted by fault reactivation, sediment loading led to compaction and shale mobilisation with the formation of domes, shale diapirs and growth faults. Major faulting of the outer margin rim in the northwest and uplift with tilting of the inner-outer dividing ridge in the Otway Basin led to deposition of a condensed Late Cretaceous section.

The ASRS in South Australia is divided into four basins. The Jurassic Poldas Basin is the oldest, and may be regarded as a failed rift valley. Forming the main part of the ASRS are the Bight, Duntroon and Otway Basins that extend, respectively, from northwest to southeast around the coastal margin; they are separated arbitrarily by linear zones of structures termed 'accommodation zones'. The Bight Basin extends onshore under the Nullarbor Plain (Eucla Basin) as a thin platformal equivalent. The Otway Basin also extends onshore beneath the Gambier Basin but maintains its rift geometry.

POLDA BASIN

C.G. Gatehouse

The Poldas Basin (Wopfner, 1969c) is an elongate graben first established in the Neoproterozoic, extending from eastern Eyre Peninsula to the continental shelf edge offshore (Fig. 9.28). This feature has been variously termed Elliston Trough (Smith and Kamerling, 1969) and Poldas Trough (Cooper and Gatehouse, 1983), but the name 'Poldas' was first used for the Poldas Freshwater Basin with Cainozoic aquifers (Jack, 1912). The onshore portion of the Poldas Basin has been subdivided for convenience into the Lock Sub-basin, Kilroo Sub-basin (Nelson *et al.*, 1986; Flint and Rankin, 1991; Fig. 9.29). Nelson *et al.* (1986) reviewed geophysical exploration in the basin and produced a map of interpreted depth to magnetic basement.

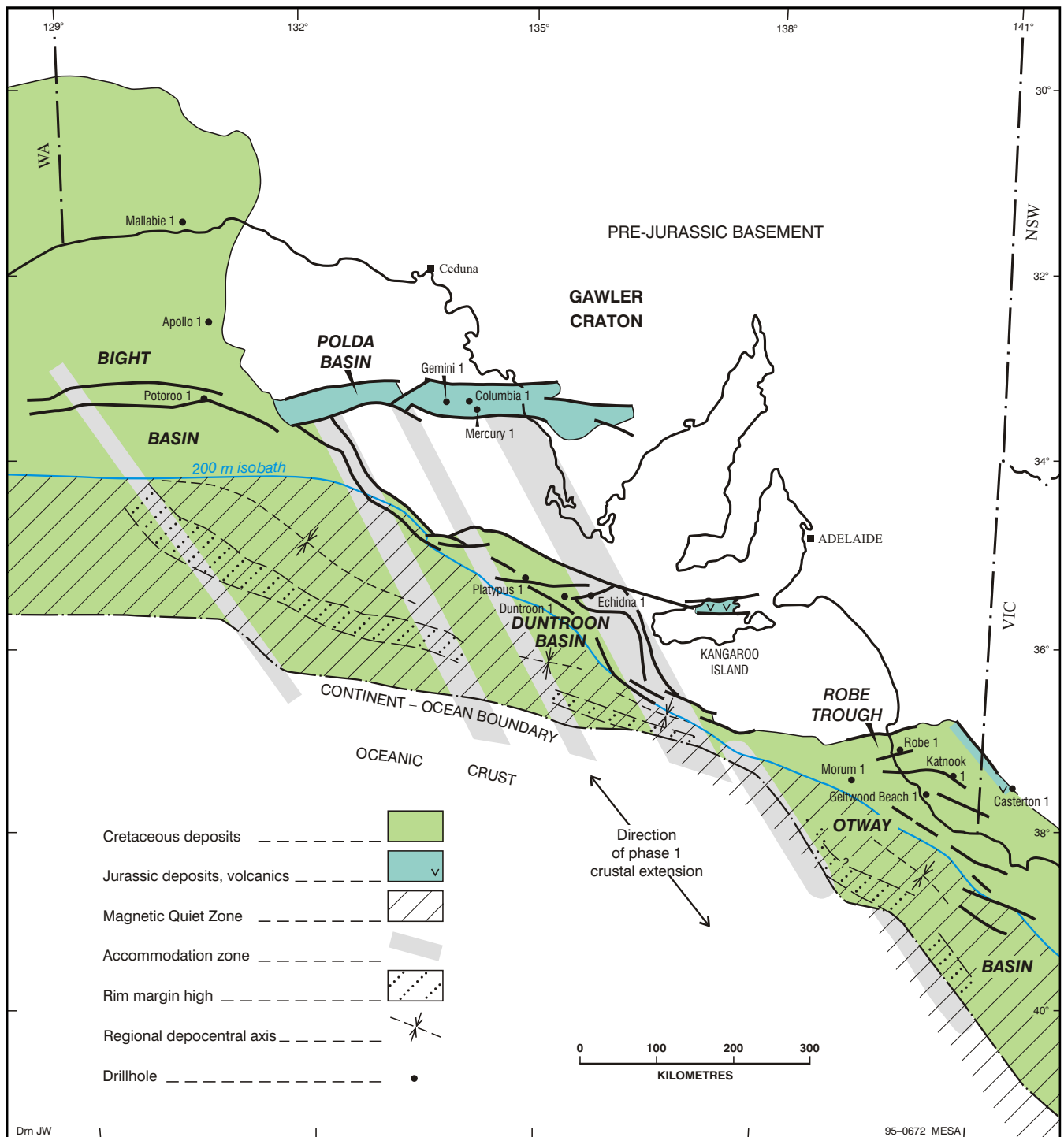


Fig. 9.28 Main structural elements of the Australian Southern Rift System in South Australia.

Jurassic sedimentation is represented by the Poldas Formation (Harris and Foster, 1974; Gatehouse and Cooper, 1982) which extends the full length of the graben. The formation is underlain unconformably by the Permian Coolardie Formation (Cooper *et al.*, 1982) and overlain disconformably by the Tertiary Poelpena Formation (Harris, 1966). Griffiths (1971) suggested that sedimentation was a response to the initial extensional tectonism that led to the breakup of Gondwana, with reactivation of the faults bounding the basin.

In its type section the Poldas Formation comprises sandstone and carbonaceous and lignitic claystone ~86 m thick (Harris and Foster, 1974). In the Lock Sub-basin the unit is divided into three unnamed members: a basal dark grey claystone with subangular 'floating' sandgrains, a middle dark grey to black carbonaceous claystone

with lignite, siltstone and fine-grained sandstone interbeds, and an upper dark brown and grey, medium to coarse-grained, angular to subangular and pyritic sandstone. Within the Lock Sub-basin, sub-bituminous coal seams up to 15 m thick are present. Similar coal seams and other lithotypes were intersected in drilling offshore and in the Kilroo Sub-basin. Coal seams pinch out abruptly towards basin margins to the north and south (Gatehouse, 1981).

Sedimentary structures in the Poldas Formation include cross-bedding, upward-fining sequences, and channelling. Poor sorting in coarser grained sandstone and conglomerate, with better sorting in finer grained intervals, and clay units beneath coal seams and within fine-grained sandstone beds, indicate a fluvial environment. The age of the Poldas Formation is late Middle to Late Jurassic based on the presence of palynoflora from the *Murospora florida* Zone.

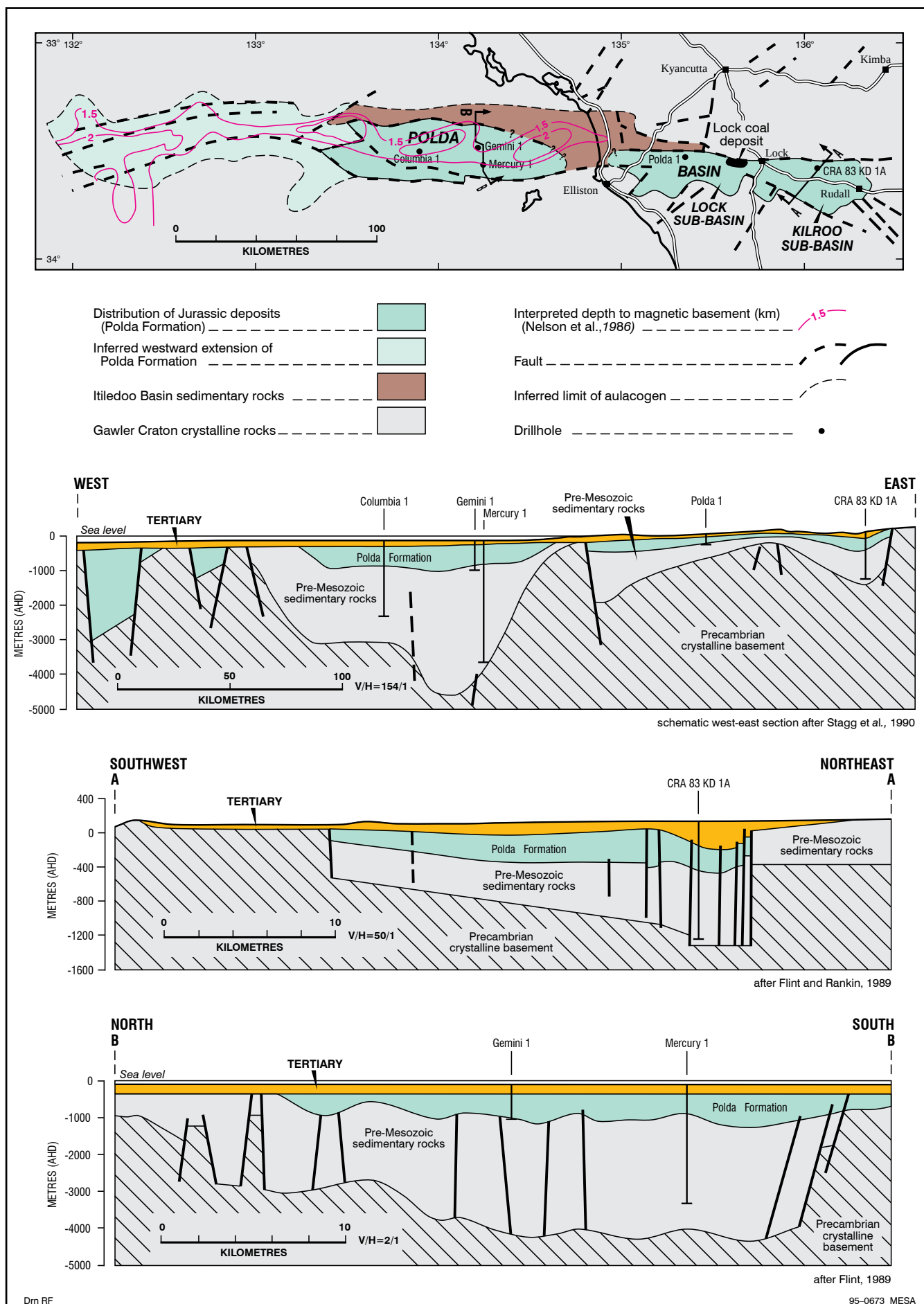


Fig. 9.29 Location, structure and geological cross-sections of the Polda Basin.

The depositional environment is interpreted as a broad valley with relatively low relief in which meandering streams provided lateral mudflats, oxbow lakes and swamps in which coal accumulated. Lateral streams from the valley sides may have provided coarser grained sediments. The known distribution of the Polda Formation along the length of the graben (400 km) contrasts with a width of only 20-60 km.

The basin has been explored for coal (Gatehouse, 1980), uranium and petroleum. Extensive offshore seismic investigations in the Great Australian Bight led to the drilling of Columbia 1, Gemini 1 and Mercury 1 for petroleum. Coal exploration, centred in the Lock Sub-basin, delineated 260 Mt of sub-bituminous coal in the measured-indicated category, with a proximate analysis of 26% moisture, 23% ash, 30% volatile matter, and 21% a fixed carbon. The average heat value of 14.6 MJ/kg is typical of South Australian coals.

BIGHT BASIN

A.J. Hill

The Early to Late Cretaceous Bight Basin occupies the central portion of the ASRS and is situated along the western part of the continental margin of South Australia, both onshore and offshore in water depths up to 2 500 m. The northern limit was originally taken as the regional monocline separating the platform from the deeper basin. However, the thin but laterally persistent Cretaceous succession on the platform was deposited at a bypass margin as the rift basin subsided rapidly to the south, and is considered to be genetically related. The Bight Basin as now defined includes Cretaceous sediments of the platform (formerly included in the Eucla Basin), which extends at least 200 km onshore from the coastline (Figs 9.28, 9.30). The South Australian portion of

the Bight Basin is flanked to the west by the Eyre Sub-basin of Western Australia (Bein and Taylor, 1981) and is separated from the Duntroon Basin by a north-northwest-south-southeast trending accommodation zone (Fig. 9.30). The southern limit is the continent-ocean boundary. In South Australia, the basin covers an area of 183 000 km² and contains up to 11 km of Mesozoic rift and post-rift sediments.

The Bight Basin has been sparsely explored. Numerous shallow drillholes have been completed onshore whilst only Apollo 1 and Potoroo 1 have been drilled offshore, all to Precambrian crystalline basement. The basin geology has been studied in detail by Pattinson *et al.* (1976), von Sanden and Barten (1977), Fraser and Tilbury (1979), Forbes *et al.* (1984), Wilcox *et al.* (1988), Stagg *et al.* (1990) and Hill (1991).

Structure

Three main sub-basins have been identified within the Bight Basin (Fig. 9.30):

- The Ceduna Sub-basin (Stagg *et al.*, 1990), previously referred to as the 'Ceduna Depocentre' (Veevers, *et al.*, 1990) and the Great Australian Bight Basin (Willcox, 1978; Fraser and Tilbury, 1979), comprises the principal sediment accumulation underlying the bathymetric Ceduna Terrace.
- The Eyre Sub-basin is a discrete extensional basin located in offshore Western Australian waters. It is separated from the continental rise to the south by shallow basement but it appears to be contiguous with the Ceduna Sub-basin to the east.
- The Recherche Sub-basin is also restricted to Western Australian waters and is separated from the Ceduna Sub-basin to the northeast by a northwesterly trending accommodation zone (Fig. 9.30).

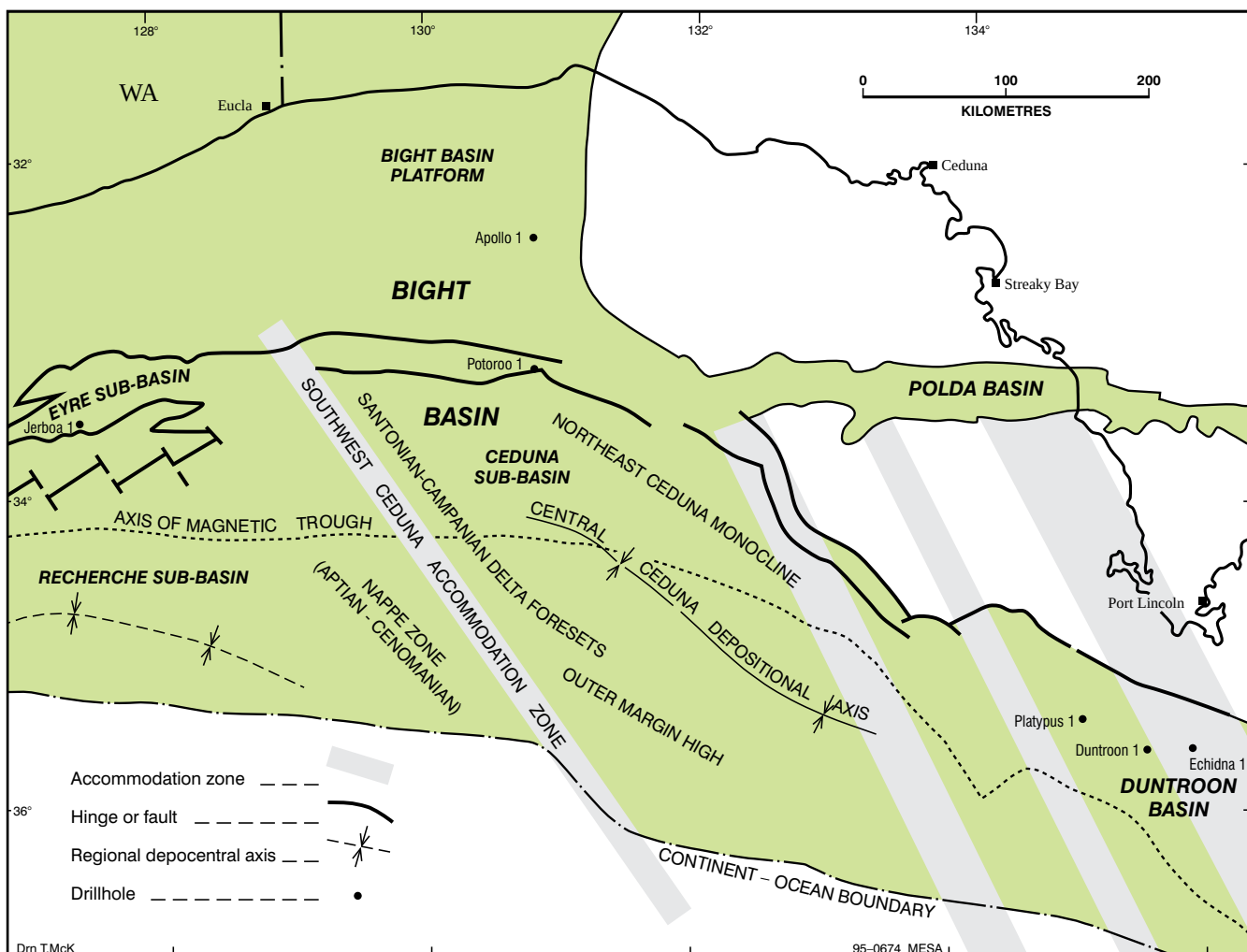


Fig. 9.30 Tectonic elements of the Great Australian Bight region.

The Ceduna Sub-basin is divided into three major structural elements. From north to south these are:

- Northeast Ceduna Monocline — this comprises the most rapidly thickening part of the basin between the edge of the continental shelf and the Ceduna Terrace. It varies between 40 and 50 km in width, thickens from <2 000 to >10 000 m, and appears to have minimal faulting within the sedimentary section.
- Central Ceduna Depositional Axis — a broad depositional trough in both lower and upper Cretaceous sections, extending ~350 km along the central Ceduna Terrace and up to 100 km in width. The area is characterised by high-angle, low-extension block faulting in the centre and northwest, and growth faulting in the southeast.
- Outer Margin High — an uplifted zone up to 40 km in width extending along the southwestern flank of the Ceduna Sub-basin beneath the outer flank of the Ceduna Terrace. Late Cretaceous sediments appear to be strongly eroded and the entire zone has been subject to complex structuring, including wrenching and possible overthrusting. Extensive Early Cretaceous shale movements, including gravity assisted sliding (Boeuf and Doust, 1975), resulted in compressional structures and toe thrusts.

The three structural elements of the Ceduna Sub-basin described here all have a general west-northwest-east-southeast orientation that appears to be the result of a large sediment influx from the northeast and deposited rapidly onto thinned crust.

Stratigraphy

Stratigraphic nomenclature for the Bight and Duntroon Basins (Fig. 9.31) has evolved largely from tentative correlations by von Sanden and Barten (1977) and Templeton and Peattie (1986), with formations previously defined by other authors in the onshore platform areas (Fairbridge, 1953; Lowry, 1970; Cockbain and Hocking, 1989; Hocking, 1990).

Stagg *et al.* (1990) used the terms Duntroon Group and Bight Group to describe the major Early and Late Cretaceous successions, respectively, in the Bight and Duntroon Basins. They are broadly coeval with the Otway Supergroup and Sherbrook Group of the Otway Basin to the southeast. However, due to uncertainty in correlating units between basins, interbasin facies changes, and the limited number of drillholes, use of the group names is not recommended until correlations can be resolved, perhaps by application of sequence stratigraphic concepts. The provisional stratigraphic synopsis presented below combines the stratigraphy of Potoroo 1 (Fig. 9.32) with detailed seismic mapping by Fraser and Tilbury (1979) of key marker horizons in the Ceduna Terrace region, and the work of Cockshell (1990a,b), Thomas (1990) and Hill (1991).

Loongana Formation

The Loongana Formation is a sandy to shaly succession of Valanginian to Hauterivian age that unconformably overlies either early Palaeozoic sediments or Precambrian basement and is conformably overlain by the Early to Late Cretaceous Madura Formation. Originally named 'Loongana Conglomerate' from Loongana Bore in Western Australia (Fairbridge, 1953), and formally defined from Transcontinental No. 3 Bore (McWhae *et al.*, 1958), it was renamed Loongana Sandstone by Lowry (1970). Hill (1991) redefined it as Loongana Formation, using the much thicker 324 m interval (units JF, JG and JH of Huebner, 1980) in Jerboa 1 and forms the basis of the following description, as reference section. An unnamed basal succession in Jerboa 1 (units JI and JJ of Huebner, 1980), comprising interbedded dark brown to black claystone and sandstone, was originally dated as Late Jurassic to Neocomian (Powis and Partridge, 1980). However, Morgan (1990) assigned a Berriasian (upper *C. australiensis* Zone) to Hauterivian (*F. wonthaggiensis* Zone) age to this non-

marine succession, which may have affinities with the Loongana Formation. Units JF to JJ inclusive in Jerboa 1 comprise sequence MZ3 (Fig. 9.33) in the sequence-stratigraphic scheme proposed by Hocking (1990).

The Loongana Formation in Jerboa 1 consists of interbedded sandstone, siltstone and shale. Sandstone is commonly white to light grey, quartzose, feldspathic in part, lithic, anhydritic, dolomitic in part, medium to coarse-grained, subangular to subrounded, poorly sorted, and commonly poorly cemented. It is locally conglomeratic, with granule-sized granite and other lithic fragments. Shale is grey, dark brown or black, pyritic in part, micaceous and occasionally contains carbonised wood fragments. The formation is present in numerous onshore drillholes within the Bight Basin and extends offshore to Jerboa 1 and Potoroo 1 (Fig. 9.34). It is a time correlative of the Echidna Formation in the Duntroon Basin.

Palynological age determinations range from Valanginian (upper *C. australiensis* Zone) to Hauterivian (*F. wonthaggiensis* Zone; Morgan, 1990).

The Loongana Formation represents sediments of the initial rift phase. The absence of microplankton suggests a non-marine environment. The lenticular and partly conglomeratic nature, and presence of carbonised wood and pyrite, possibly indicate deposition in a low-sinuosity fluvial environment within isolated grabens of the developing rift valley.

Madura Formation

Fairbridge (1953) introduced the term 'Madura Shale' for sandstone, siltstone, claystone and shale overlying the Loongana Formation and disconformably underlying the Tertiary Hampton Sandstone. Lowry (1970) modified the name to Madura Formation because of the heterogeneous lithology. Cockbain and Hocking (1989) recognised persistent time breaks and lithological variations within the formation, and separated the upper part which they divided into the Toondi and Nurina Formations. The redefined Madura Formation has a type section in Transcontinental Railway No. 1 Bore (Madura Bore) at Madura, Western Australia, between 614 and 640 m, and is restricted to the Bight Basin. In Apollo 1 it occurs as part of a condensed Early and Late Cretaceous succession deposited under starved clastic conditions. It lies conformably on the Loongana Formation and is overlain disconformably by the Albian to Cenomanian Toondi Formation in the Eyre Sub-basin and the platformal portion of the Bight Basin. In the Ceduna Sub-basin, however, it is overlain disconformably by two unnamed units, referred to informally here as 'unit a' and 'unit b', collectively equivalent in age to the Toondi Formation. The Madura Formation ranges in thickness from 26 m in Madura Bore to 474 m in Jerboa 1 (formerly unit JE of Huebner, 1980).

The type section consists of glauconitic and carbonaceous sandstone, siltstone, claystone and shale. Sandstone is micaceous, clayey, poorly sorted and interbedded with claystone; shale is dark grey, silty, sandy, micaceous and pyritic. Thin microcrystalline dolomite bands occur sporadically within the claystone and shale beds.

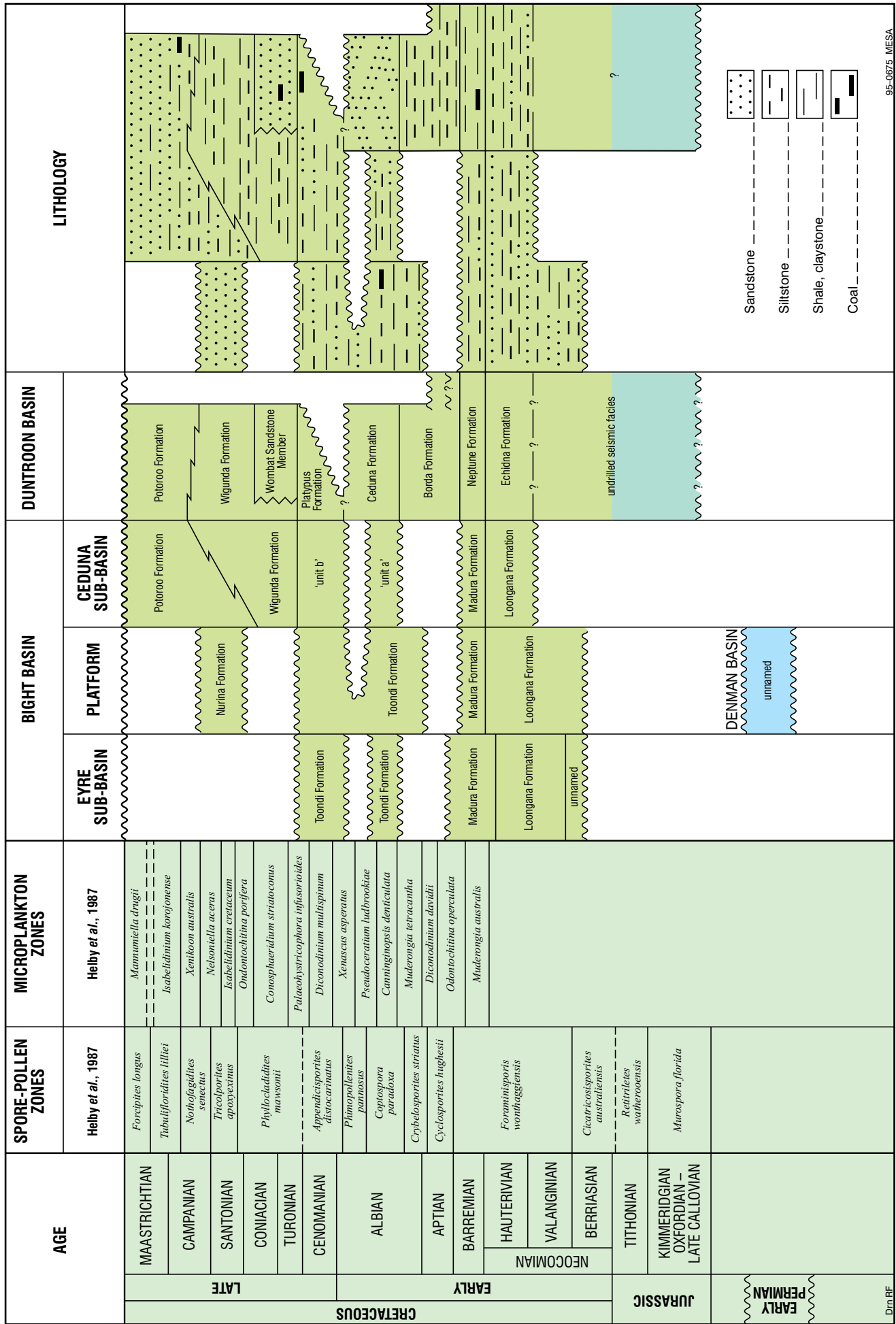
The age determined by palynological dating is Barremian at the base to Aptian at the top (Hocking, 1990).

The inferred depositional environment is marine, commonly with restricted circulation, although Hocking (1990) recorded low-energy lacustrine conditions in Jerboa 1.

Toondi Formation

The Albian to Cenomanian Toondi Formation is a siltstone, claystone and shale succession with minor interbeds of sandstone. It disconformably overlies the Madura Formation and is disconformably overlain by the Santonian to early Campanian Nurina Formation, and represents a fine-grained facies originally included in the Madura Formation.

In the type section (Madura Bore, 336-614 m) the formation



95-0675 MESA

Fig. 9.31 Stratigraphic summary of the Bight and Duntroon Basins, South Australia.

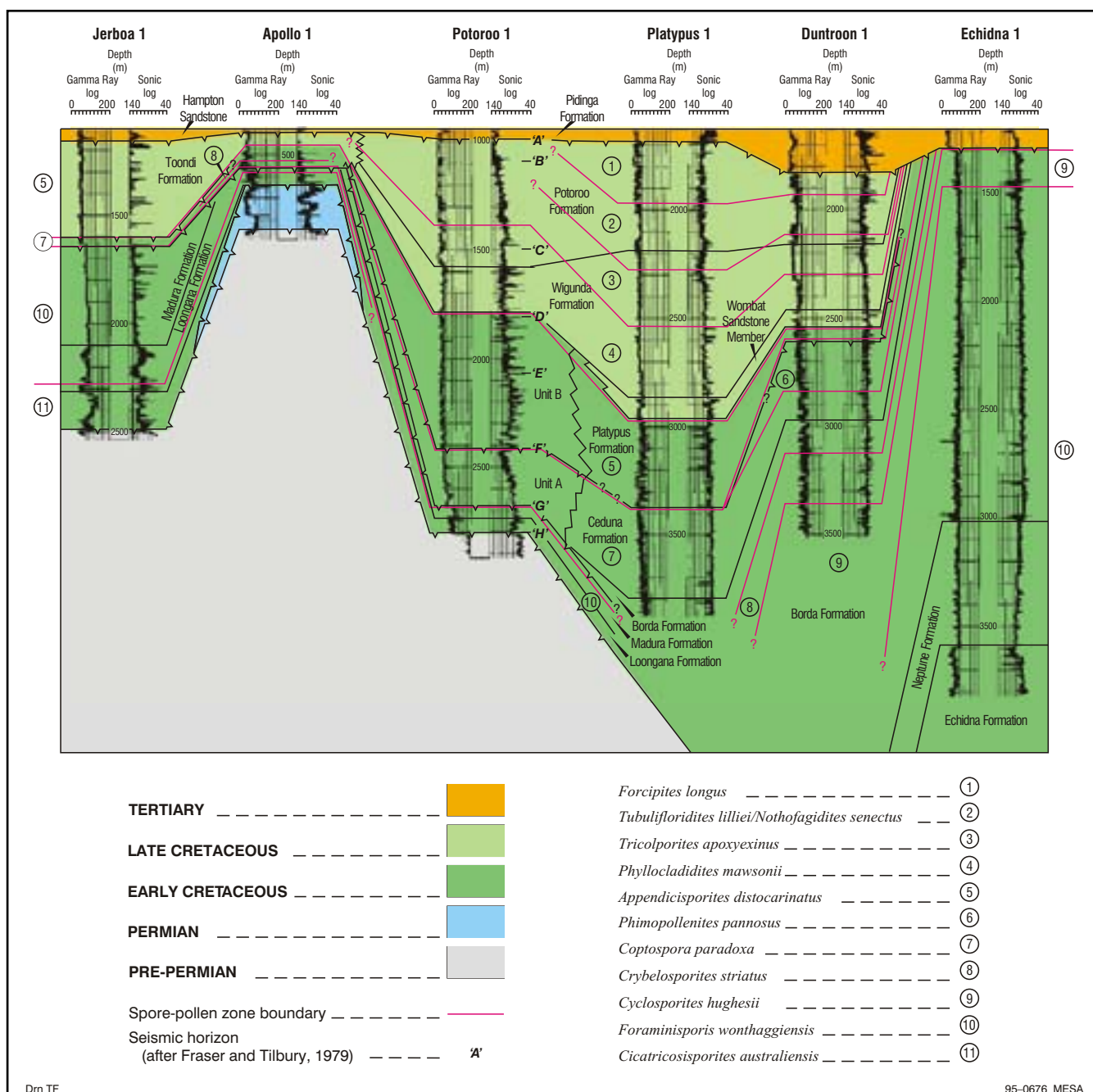


Fig. 9.32 Drillhole Correlation, Bight and Duntroon Basins.

is monotonous and continuous with no recognisable time breaks (Cockbain and Hocking, 1989) although this may be a consequence of sparse palynological sampling. Cookson and Eisenack (1970, 1971), however, recognised an intraformational hiatus spanning the Albian (*C. paradoxa* and *P. pannosus* Zones) in Madura Bore, and this may be linked to a similar break in Jerboa 1 between the informal units JA to JC and JD (Huebner, 1980; Fig. 9.33). Further east in Apollo 1 and Potoroo 1 there is likewise an Albian hiatus (Morgan, 1990).

Whilst there appears to be no demonstrable lithological variation in the Toondi Formation in the type section, there is a gradual variation above and below the disconformity in the offshore regions of the basin, in both Jerboa 1 and Potoroo 1. Unit JD (1 600-1 535 m) in Jerboa 1 comprises calcareous, dark grey, black and brown claystone with some thin quartz sand interbeds near the top, possibly representing deposits of a prograding delta system. Dinoflagellates are more abundant than spores and pollen, and a marine depositional environment is inferred. Units JA-JC comprise light to greenish dark grey argillaceous sandstone interbedded

with reddish to dark grey carbonaceous claystone and siltstone, with varying amounts of glauconite. A nearshore environment of deposition is assumed. To the east in Potoroo 1 there is an even greater variation above and below the disconformity and for this reason it would be erroneous to assign this thick Albian to Cenomanian succession to age-equivalent formations in the Duntroon Basin, the Eyre Sub-basin or the platformal portion of the Bight Basin. Consequently these thick unnamed units in Potoroo 1 are referred to informally here as 'unit b' (1 776-2 420 m) and 'unit a' (2 420-2 690 m), which are time equivalents of the Ceduna and Platypus Formations, respectively (Fig. 9.31).

'Unit a' comprises fine to medium-grained, micaceous, glauconitic sandstone with minor coal bands, grading up to silty shale. The top of this unit correlates with seismic horizon 'F' (Fraser and Tilbury, 1979) and is a time correlative of the Ceduna Formation of the Duntroon Basin.

'Unit b' comprises interbedded silty mudstone and dolomite with minor coal bands near the top, and corresponds to horizon 'D' of Fraser and Tilbury (1979). It was originally assigned to the

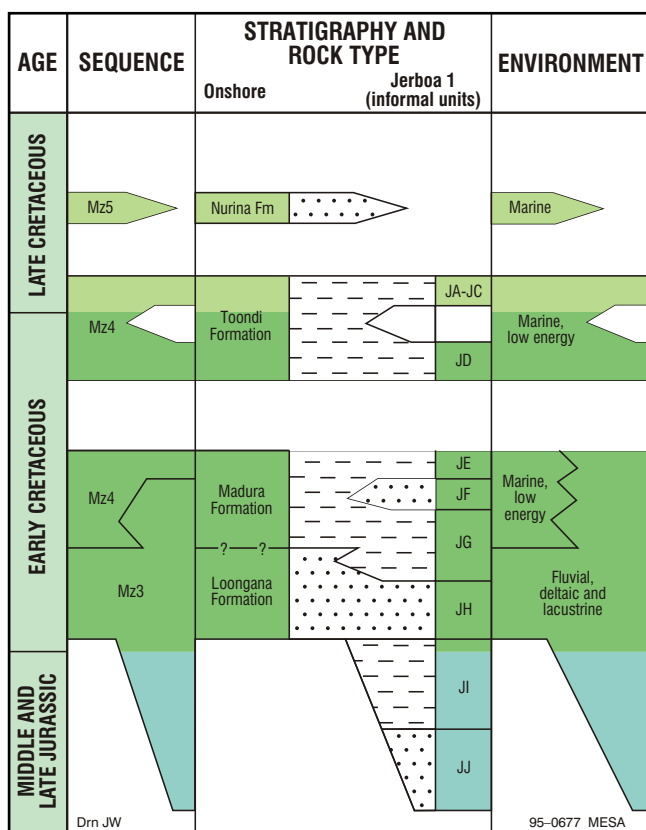


Fig. 9.33 Correlation between Jerboa 1 (Eyre Sub-basin) and onshore Bight Basin, Western Australia.

Platypus Formation by von Sanden and Barten (1977), but differs from the type section of that formation and also from the Toondi Formation of the northern Bight Basin.

A pronounced truncation surface evident in seismic sections, suggestive of a major sequence boundary, is coincident with this regional hiatus in the late Albian (*P. pannosus* Zone; Morgan, 1990).

Wigunda Formation

The Wigunda Formation, defined in the Duntroon Basin, extends to Potoroo 1 in the Ceduna Sub-basin where it lies unconformably between unnamed 'unit b' and the Potoroo Formation. The type section is the interval between 2 206 and 2 871 m in Platypus 1 (Hill, 1991), which also represents the maximum known thickness (665 m). It is absent to the west in the Eyre Sub-basin and is considered to be a correlative of the Belfast Mudstone of the Otway Basin.

In its type section, the formation consists of dark grey to olive, carbonaceous and glauconitic mudstone with rare bands of muddy siltstone and sandstone, and becomes increasingly silty towards the base. Minor bands of microdolospirite with rare coal bands near the top of the formation occur in Potoroo 1. The top of the Wigunda Formation is strongly diachronous from the Bight to the Duntroon Basin, although this may be an artefact of varying palynological interpretations. A Turonian to Coniacian age (*P. mawsonii* Zone) is indicated in Potoroo 1 (Morgan, 1990). The dominance of dark grey to olive mudstone with only minor interbeds of siltstone and the incidence of microdolospirite is characteristic of prodelta deposits in a restricted coastal bay environment.

Nurina Formation

The Nurina Formation is confined to the northern onshore portion of the Bight Basin in the vicinity of Madura in Western Australia. The type section is in Madura Bore (283-336 m) where the maximum thickness of 53 m is attained. Formerly assigned to the Madura Formation, it was named by Cockbain and Hocking (1989) to describe a glauconitic sandstone and glauconitic sandy siltstone. The formation is absent in Jerboa 1 and may represent

a more proximal facies of the Wigunda Formation. A marine environment with low to moderate terrigenous supply is inferred (Hocking, 1990).

Potoroo Formation

The Potoroo Formation conformably overlies the Wigunda Formation and is unconformably overlain by the Tertiary Pidinga Formation (Fraser and Tilbury, 1979). It was defined in Potoroo 1 for the sandstone, siltstone, mudstone and minor coal succession between 990 and 1 570 m (Hill, 1991) which also represents the maximum penetrated thickness (580 m). In the type section, the basal part of the formation consists of glauconitic siltstone and sandstone with minor coal, and is expressed seismically as strong, continuous reflectors in which large-scale foreset beds can be distinguished (Fraser and Tilbury, 1979). The upper 298 m consist of clean, unconsolidated, clear to milky quartzose sand with minor bands of silicified siltstone, mudstone and sub-bituminous coal.

The formation is laterally and vertically extensive across the Ceduna Sub-basin and the Duntroon Basin, and correlates with seismic horizon interval C to A (Fraser and Tilbury, 1979). It is absent in Jerboa 1 in the Eyre Sub-basin and has not been penetrated in the platformal portion of the Bight Basin.

Palynological dating of the Potoroo Formation indicates a Coniacian to Maastrichtian age (*P. mawsonii* to *Forcipites* (*ex Tricolpites*) *longus* Zones) in the west (Morgan, 1986, 1990).

Previous workers (Boeuf and Doust, 1975; Pattinson *et al.*, 1976) recognised large-scale foreset beds with a thick, prograding sequence, which possibly indicates development of a fully marine basin. However, Fraser and Tilbury (1979) suggested that conditions were not fully marine. The pronounced downlap of reflectors onto the underlying Wigunda Formation ('C' horizon of Fraser and Tilbury) in the west of the Ceduna Sub-basin suggests prograding of foreset beds from shallow to deep water to the south of the basin. Depositional environments were paralic in the north, possibly bathyal to the south and nearshore marine to continental to the east in the Duntroon Basin where discontinuous sand and coal are thought to have been deposited in a delta-plain environment.



AGSO research vessel *Rig Seismic* operating in the Bight Basin. Photo 042489.

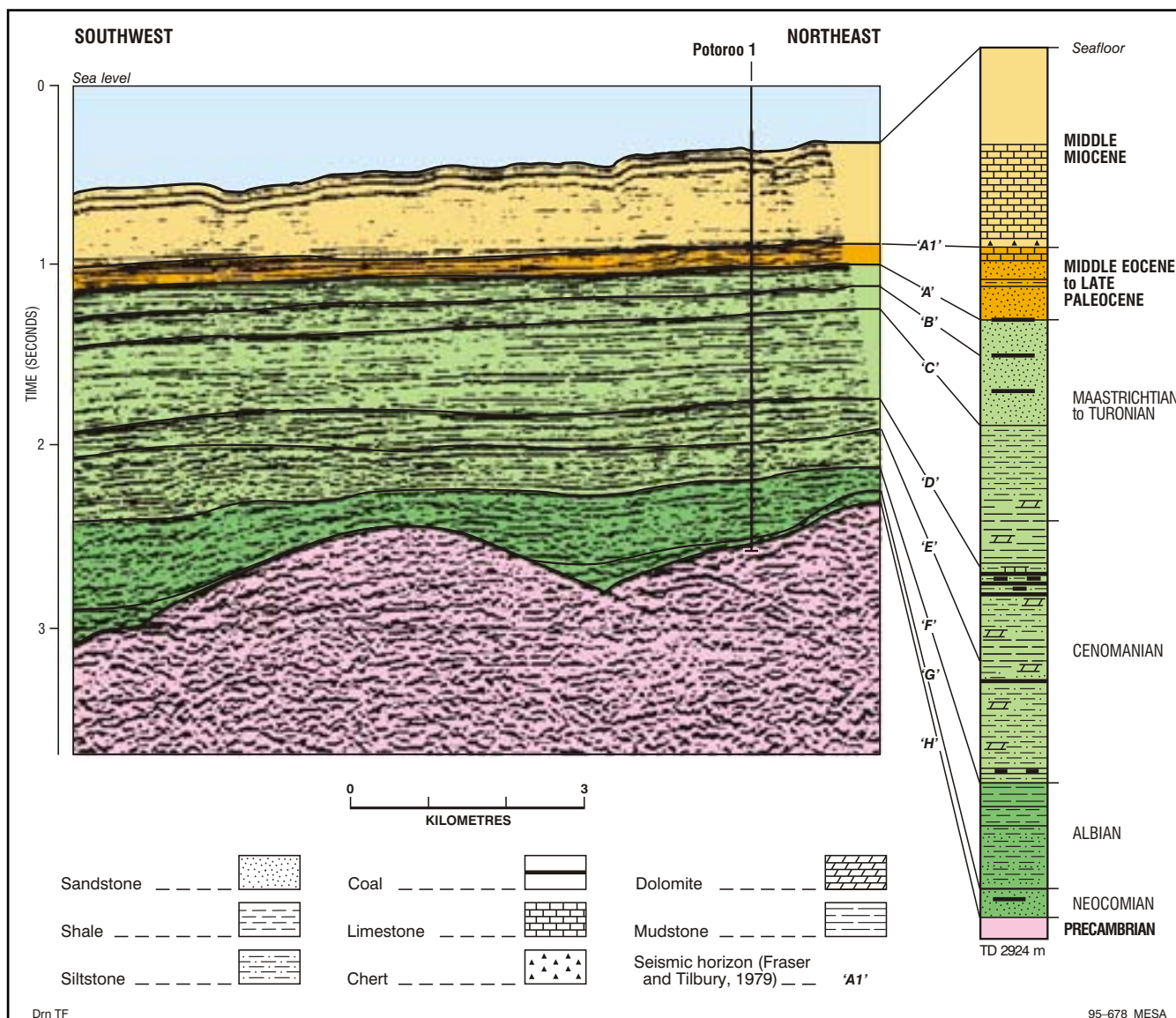


Fig. 9.34 Interpretation of seismic line 65-3P1 from the northern margin of the Ceduna Sub-basin showing location and summary of Potoroo 1.

DUNTROON BASIN

A.J. Hill

The Duntroon Basin, covering more than 57 000 km², comprises the central part of the ASRS below water depths of up to 2 500 m and contains up to 11 km of Mesozoic rift and post-rift sediments. It is separated from the Bight Basin to the west and the Otway Basin to the east by prominent north-northwest-south-southeast fault systems (accommodation zones). The northern limit is expressed in seismic records as a pronounced escarpment against the downfaulted Gawler Craton (Willcox *et al.*, 1988; Figs 9.28, 9.35); the southern limit is the continent-ocean boundary.

Six petroleum drillholes have been completed (Fig. 9.35), largely on intensely faulted broad anticlines, without disclosing economic accumulations of hydrocarbons, although Greenly 1 was plugged and abandoned with significant oil and gas shows. Source-rock analyses indicate that drilled portions of the basin have high gas and moderate oil potential although source maturity is variable. Of these drillholes, Borda 1, Greenly 1 and Vivonne 1, all completed by BHP Petroleum in 1993, were still confidential at the time of publication and are not included in this review.

The basin geology has been studied in detail by von Sanden and Barten (1977), Fraser and Tilbury (1979), Forbes *et al.* (1984), Wilcox *et al.* (1988), Stagg *et al.* (1990), Cockshell (1990a) and

Thomas (1990). Regional studies of the tectonic evolution of the southern margin of Australia have been conducted by Weissel and Hayes (1977), and Falvey and Mutter (1981). Cande and Mutter (1982) revised the interpreted age of continental breakup as late Albian-Cenomanian (110-86 Ma); the age has been further refined by Veevers (1988) as the Early-Late Cretaceous boundary (95±5 Ma).

Structure

A prominent series of fault-bounded northwest-trending basement blocks divide the Duntroon Basin into inner and outer zones, separated by distinct but discontinuous basement horst blocks (Fig. 9.35). The inner zone consists of relatively undisturbed landward-dipping half grabens, characterised by east-southeast-trending, landward-dipping tilted blocks that appear to have been active much later in the eastern portion of the Duntroon Basin. Within this inner faulted zone, broad anticlines with areal closures up to 450 km² are coincident with reactivated basement highs (Fig. 9.36). The anticlines are intensely faulted and often collapsed at their crests above a major ?Jurassic low-angle detachment fault below which sediments are relatively undisturbed (Etheridge *et al.*, 1988).

In contrast, the outer zone extends south and west towards the continental slope and is intensely faulted. Extensive Early Cretaceous shale movements are also apparent along large down-

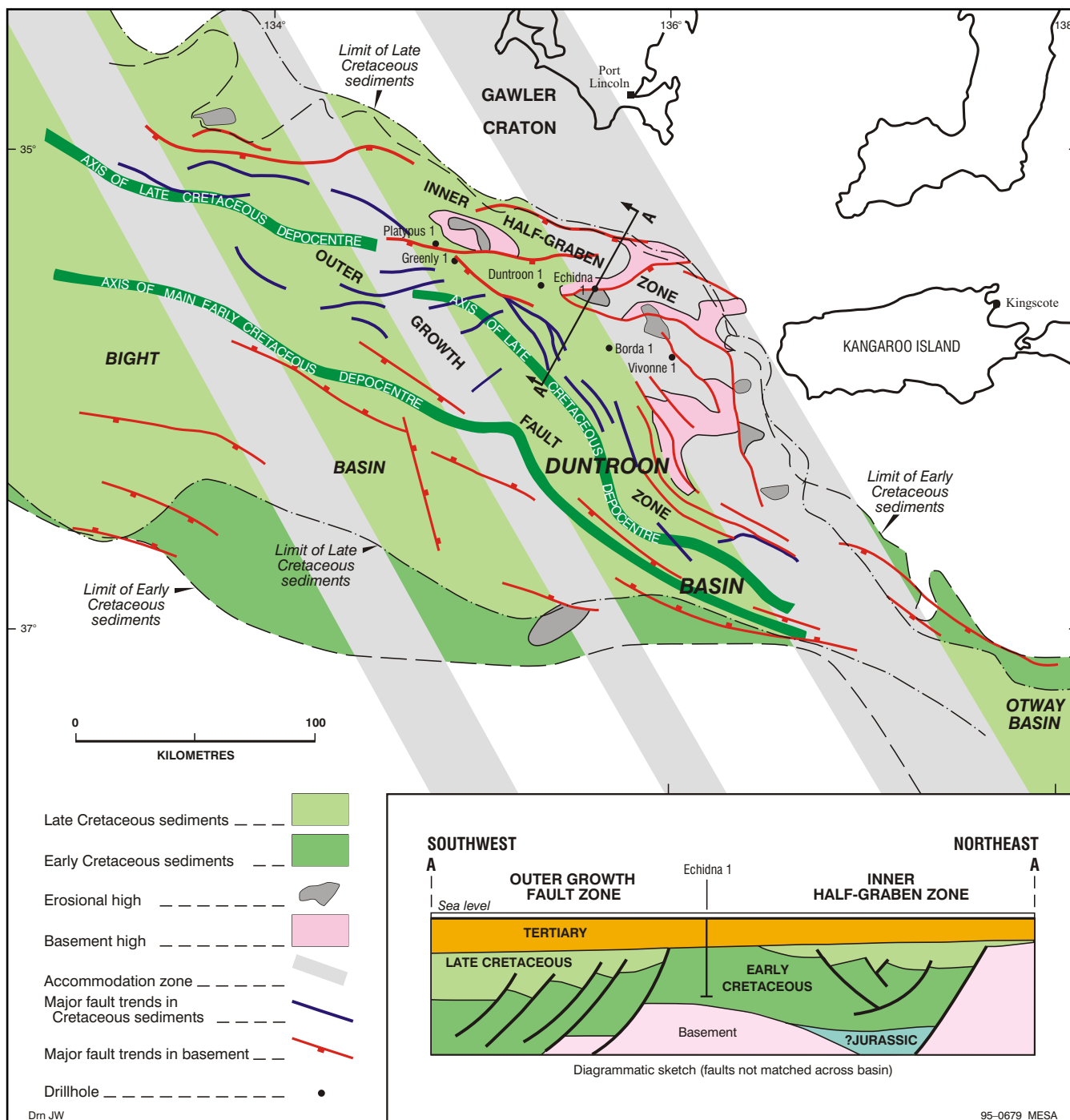


Fig. 9.35 Structural elements of the Duntroon Basin.

to-basin listric faults. Associated erosion of anticlinal crests allowed continuous sedimentation on their flanks whilst creating major unconformities at the crests (e.g. Echidna 1; Fig. 9.35). Only minor tectonism is recorded in the Tertiary by local adjustment along fractures within faulted anticlinal complexes.

Stratigraphy

Early Cretaceous

The Early Cretaceous succession has not been fully penetrated in the Duntroon Basin. Echidna 1, located on a faulted, central basin high, reached a total depth of 3 839 m in non-marine sediments of Hauterivian age (*F. wonthaggiensis* Zone). Seismic data indicate Precambrian crystalline basement at ~4 200 m. However, away from basement highs, the top of basement may be as deep as 12 500 m below sea floor, suggesting that a significant ?Jurassic to Neocomian section may be present. The Early Cretaceous succession resulted from syn-rift deposition.

Echidna Formation

The name 'Echidna beds' was used informally by von Sanden and Barten (1977) for dark shale and mudstone grading down to siltstone and shaly sandstone, intersected in Echidna 1 between 3 595 and 3 839 m. The unit is now defined as Echidna Formation (Hill, 1991) and presumed to be vertically and laterally extensive from seismic interpretations (Thomas, 1990). The increase in sand content towards the base of the unit suggests correlation with the Loongana Formation of the Bight Basin and possibly with a distal equivalent of the Pretty Hill Formation in the Otway Basin.

The formation is characterised by interbedded silty, slightly gypsiferous, dark grey shale and very fine-grained muddy sandstone, with sandy beds more frequent towards the base. The top of the unit is marked by a prominent decrease in the gamma ray log at the base of the overlying thick, monotonous shale. Palynological data indicate a Neocomian age (*F. wonthaggiensis* Zone; Morgan, 1990).

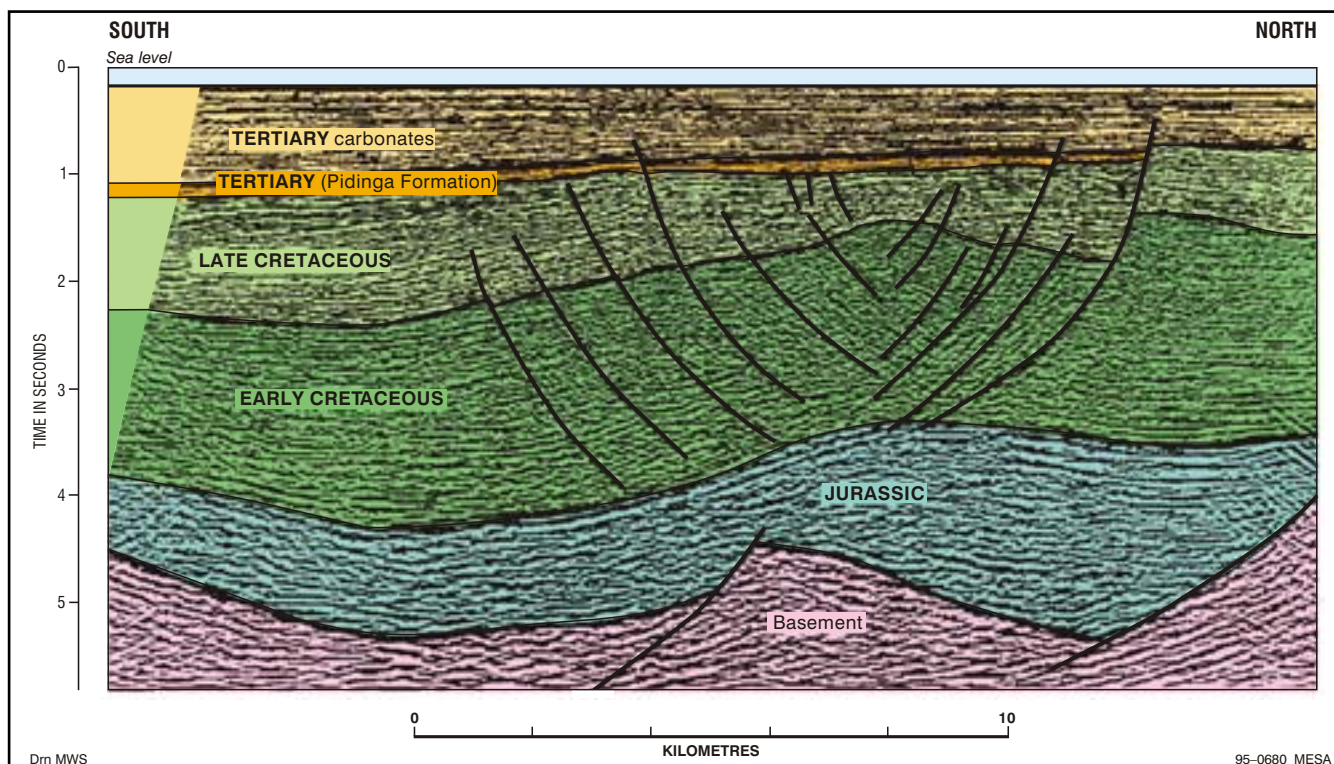


Fig. 9.36 Interpretation of seismic line 83-64, Duntroon Basin.

Poor sorting and immature grain shapes of detrital quartz sand suggest a shallow, lacustrine environment of deposition with sediment supply from a proximal landmass.

Neptune Formation

The Neptune Formation is a dark brown to black, carbonaceous shale lying conformably between the Echidna and Borda Formations. The name is derived from Neptune Island northeast of Echidna 1, in which the type section is the interval from 3 016 to 3 595 m. The formation has been penetrated only in Echidna 1, but is inferred from seismic data to be laterally extensive in deeper portions of the basin. A dark shale intersected in Jerboa 1 (Western Australia) between 1 635 and 2 099 m (unit JE of Huebner, 1980) bears strong similarities to the Neptune Formation, but is assigned to the Madura Formation until additional drilling or seismic data confirm continuity.

The Neptune Formation is characterised by a uniformly high gamma ray signature and a relatively quiet sonic trace when compared to the overlying Borda Formation, reflecting the uniform lithology of dark brown to black shale with minor interbedded siltstone and light to medium grey, fine to very fine-grained argillaceous sandstone. The unit is variably micaceous and slightly gypsiferous, calcareous and carbonaceous. Palynological data indicate a Neocomian to Barremian age (*F. wonthaggiensis* Zone; Morgan, 1990).

The persistence of uniform, dark, carbonaceous shale with interspersed very fine-grained silty sand and the absence of microplankton suggest deposition under anoxic conditions in a quiet, non-marine environment, possibly a stratified lake similar to that proposed for the Otway Basin by McKirdy *et al.* (1988).

Borda Formation

The Borda Formation, previously assigned to the Madura Formation, consists predominantly of light grey siltstone and claystone. The formation has not been fully penetrated by any single drillhole, but the upper conformable contact with the Ceduna Formation was intersected by Duntroon 1 and possibly Platypus 1,

and the lower conformable contact with the Neptune Formation in Echidna 1. Together, these drillholes provide a reference section in the interval 2 983 to 3 515.6 m (Duntroon 1) and 1 304 to 3 016 m (Echidna 1), giving thicknesses of 532.6 and 1 712 m respectively, but it is by no means certain that the combined intervals represent the entire unit.

The unit consists of interbedded light to medium grey, feldspathic siltstone, shale and minor claystone, and grades transitionally up into the more sandy Ceduna Formation, the top being placed at the last major shale unit. The formation is distinguished petrophysically by an irregular gamma ray and sonic trace reflecting its interbedded nature compared to the uniform log signature of the underlying Neptune Formation.

Palynological data indicate a Barremian to mid-Albian age (*F. wonthaggiensis* to lower *C. paradoxa* Zones; Morgan, 1990). The dominance of shale and siltstone suggests a low-energy environment with a poorly developed drainage system. The increase in siltstone and fine sandstone interbeds compared to the underlying Neptune Formation, and the abrupt change from dark to light grey shale, becoming partly lignitic towards the base and top, possibly indicate the development of a more regionally extensive fluvial-lacustrine system as rifting accelerated.

The gradual increase in microplankton to the west suggests that conditions were paralic with a westerly connection to the open sea, whilst continental conditions are apparent to the east. Abundant dinoflagellates between 3 254 and 3 279 m in Duntroon 1 (Morgan, 1986) provide evidence of a minor marine incursion in the early Albian.

Ceduna Formation

The Ceduna Formation is an upward-coarsening sandstone gradationally overlying the Borda Formation and conformably to disconformably overlain by the Platypus Formation. Previously assigned to the Madura Formation in Platypus 1 and Potoroo 1 by von Sanden and Barten (1977), it represents a distinct facies variation from the shale-dominated Madura Formation in the onshore Bight Basin. Templeton and Peattie (1986) assigned the

unit to the 'basal Platypus Formation' in Duntroon 1. The type section is 2 590 to 2 983 m in Duntroon 1, core 1 (2 751.3 to 2 769.75 m) being the reference core. The penetrated thickness ranges from 393 m in Duntroon 1 to 513 m in Platypus 1 showing an apparent increase in thickness to the southwest. The age is Albian (upper *C. paradoxa* to *P. pannosus* Zones; Morgan, 1986).

The unit appears as upward-coarsening on the gamma ray log, with a corresponding positive sonic shift towards the top, reflecting an increase in porosity. In core, the unit consists of an assemblage of stacked cross-bedded, light grey, medium-grained, well-sorted, partly feldspathic sandstone beds and massive, parallel-laminated sandstone with small-scale wave and current ripples, silt intraclasts and siltstone interbeds. The Ceduna Formation is interpreted to be a regressive progradational sequence that was deposited in a lower delta-plain environment. Fluvial delta-top deposits are overlain by delta distributary mouth deposits (i.e. channel fill within distributary mouth channels). The similarity between the successions in Platypus 1 and Duntroon 1 suggests that the delta system extended for at least 50 km.

Late Cretaceous

The Late Cretaceous succession, conformably overlying Early Cretaceous deposits, is paralic to marine and represents the transition from syn-rift to post-breakup deposition resulting from the separation of Australia and Antarctica (Veevers, 1988; Mutter *et al.*, 1985). Uplift of the continent between 95 and 90 Ma and the associated retreat of the sea (Veevers, 1984) led to the development of an extensive drainage system centred in the Ceduna Sub-basin. Provenance shifted from the highlands of eastern Australia and volcanic complexes along the margins of the rift valley in the Early Cretaceous to multiple sources in the Late Cretaceous. Fission track dating on detrital apatite from Duntroon 1 (Gleadow and Duddy, 1986) indicates two distinct sources for the sediments. Apatite from Early Cretaceous sandstone yielded corrected ages of 131 Ma and displayed etched dislocation swarms typical of volcanogenic apatite prevalent in the Otway Basin. Detrital apatite from the Late Cretaceous was derived from the Gawler Craton to the north and volcanogenic terrains to the east or the contemporaneous east-west rift valley. This implies that a significant seaway did not develop on the southern margin until at least the Tertiary (Gleadow and Duddy, 1986). The depositional model for the Duntroon and Bight Basins in the Late Cretaceous is similar to that for the Otway Basin.

Evidence for a 'breakup unconformity' at the boundary between the Early and Late Cretaceous in the Bight Basin may be supported by the absence of the Albian *P. pannosus* Zone (Morgan, 1990) in Potoroo 1, Apollo 1 and Jerboa 1. In the Duntroon Basin, however, this break is of minor importance. The *P. pannosus* Zone is present in Duntroon 1 and may have been faulted out in Platypus 1. Complex faulting and erosion due to movements along basement highs, at least within the Duntroon Basin, negate confidence in assigning a breakup unconformity. Significant reworking of late Albian palynomorphs, including *P. pannosus*, in the Duntroon Basin suggests considerable erosion of the upper Ceduna Formation during deposition of the Platypus Formation.

Platypus Formation

The Platypus Formation, intersected between 2 974 and 3 368 m in Platypus 1, consists of interbedded calcareous sandstone, siltstone, carbonaceous shale and coal. It ranges in thickness from 49 m in Duntroon 1, where it occurs as an erosional remnant, to 394 m in Platypus 1. The unit rests conformably or disconformably on the Ceduna Formation and is conformably overlain by the Wombat Sandstone Member of the Wigunda Formation.

The formation is widespread and mappable although it is locally absent over some basement highs. In the more distal margins of the Duntroon Basin it may be indistinguishable from the Potoroo

Formation as the delta-front facies of the Wigunda Formation pass laterally into delta-plain deposits of the Platypus and Potoroo Formations.

The Platypus Formation is of late Albian to Cenomanian age (*A. distocarinatus* Zone; Morgan, 1990).

The general increase in sand and coal compared to the underlying Ceduna Formation is thought to indicate deposition within an upper delta plain, possibly in response to tectonic uplift of horst blocks on the basin margin which provided sediment for the prograding delta complex. Coal seams in the upper part of the formation extend further seaward than those below, suggesting maximum delta progradation towards the top.

Wigunda Formation

The Wigunda Formation is a mudstone unit lying conformably between the Platypus and Potoroo Formations. The type section is the interval between 2 206 and 2 871 m in Platypus 1 (Hill, 1991), which also represents the maximum known thickness (665 m). The formation consists of dark grey to olive, carbonaceous and glauconitic mudstone with rare bands of muddy siltstone and sandstone, characteristic of prodelta deposits. It occurs in Duntroon 1 and Platypus 1 but is absent on basement highs (e.g. in Echidna 1) due to erosion.

The top of the formation is strongly diachronous from the Bight to the Duntroon Basin, although this is probably a result of varying palynological interpretations. The age is Turonian to Santonian (*P. mawsonii* to *Nothofagidites senectus* Zones; von Sanden and Barten, 1977; Morgan, 1990) in the Duntroon Basin.

The Wombat Sandstone Member is a thin but laterally persistent sandstone deposited as a transgressive shoreface facies at the base of the Wigunda Formation, conformably overlying the Platypus Formation in deeper portions of the Duntroon Basin. Originally referred to informally as 'basal Wigunda Formation' (Morgan, 1986) it is representative of delta-front deposits equivalent to the Flaxman Formation of the Otway Basin. The type section is in Platypus 1 between 2 871 and 2 974 m, which provides the maximum known thickness of 103 m.

The member is characterised by light to medium grey, fine-grained, moderately sorted, quartzose, slightly carbonaceous sandstone with thin coal bands. The unit is well developed in Duntroon 1 and Platypus 1 but becomes distinctly silty and less defined in Potoroo 1 and thus does not extend to the Bight Basin.

The lithology is unsuitable for palynological age determination and only a maximum and minimum age based on the palynological zones of adjacent formations can be suggested. A Turonian to Coniacian age (*P. mawsonii* Zones) is tentatively assigned (von Sanden and Barten, 1977; Morgan, 1986).

Potoroo Formation

Within the Duntroon Basin, the Potoroo Formation varies slightly from the type section in the Ceduna Sub-basin (Potoroo 1) and can be subdivided, in ascending stratigraphic order, into four mappable units, possibly reflecting transgressive-regressive phases in a nearshore marine to delta-plain environment — an interbedded sequence of mudstone and sandstone, a monotonous mudstone unit analogous to the Wigunda Formation, a transitional interbedded mudstone and sandstone unit, and a sandstone unit with minor shale interbeds.

The formation conformably overlies the Wigunda Formation and is conformably overlain by the Tertiary Pidinga Formation of the Eucla Basin.

Palynological dating of the Potoroo Formation indicates a late Santonian to Maastrichtian age (*Tricolporites appoxyxinus* to *F. longus* Zones) in the Duntroon Basin, and Coniacian to Maastrichtian age (*P. mawsonii* to *F. longus* Zones) in the Ceduna Sub-basin (Morgan, 1986, 1990). This variation in age is possibly due to inaccuracies in palynological dating between basins.

OTWAY BASIN

J.G.G. Morton

The Otway Basin, together with the Bass and Gippsland Basins to the east in Victoria, forms part of the southeastern coastal-margin rift zone of the Australian continent. Only the western portion (>25 000 km²) of the Otway Basin occurs in South Australia, straddling, roughly equally, the present day coastline. It is the best known of the State's rift basins, being partly onshore (though entirely subsurface in South Australia) and therefore the earliest and most readily explored, with more than 30 onshore and eight offshore drillholes.

Structural and tectonic setting

The Otway Basin is bounded to the north by the Padthaway Ridge, consisting of a partly exhumed Ordovician igneous complex, which rises steeply from the south and marks the limit of thick basin fill (Fig. 9.37). The basin is divisible into northern and southern parts separated by the northwest-trending Tartwaup Hingeline. Sedimentary sequences in the two areas express contrasting depositional settings between the Early and Late Cretaceous: the northern area is dominated by the syn-rift Otway Supergroup and the southern by the post-rift Sherbrook Group. The main structural features are shown on Figure 9.37.

The Red Sea Rift System is considered as a modern analogue of the Early Cretaceous portion of the basin (Blake, 1985), although the East African Rift System displays certain tectonic similarities (Carey, 1969; Veevers, 1984). The precise location of the original continent-ocean boundary between Australia and Antarctica is ill-defined and the unique nature of the Magnetic Quiet Zone with its complex 'rift crust' consisting of hybrid oceanic and continental material further complicates the obscurity of the boundary. Seismic data acquired by BMR (Willcox *et al.*, 1985) along the offshore portions of the Otway Basin from South Australia to western Tasmania indicate an extensive rift divergence zone. Offshore seismic line 48/043 (Willcox *et al.*, 1985), which extends 175 km southwards from Argonaut A1 with no clear indication of the continent-ocean boundary, together with the onshore portion of the basin, indicates a total basin width of at least 300 km.

Rift phase — Early Cretaceous Otway Supergroup

Along the onshore portion of the Otway Basin, thick accumulations of terrestrial Otway Supergroup sediments are preserved within the Robe, Penola and St Clair Troughs, in turn overlain by condensed Late Cretaceous Sherbrook Group. To the south of the Beachport and Kalangadoo Highs, the Tartwaup Hingeline, coincident with the Tartwaup Fault and a series of down-to-basin faults marks the northern extremity of the main rift valley within the rifted arch (Figs 9.37). The area to the north of the hingeline became a clastic bypass margin in the Late Cretaceous.

Northeast-southwest extension between the Australian and Antarctic plates in the Neocomian may have begun as a northward-dipping detachment that became southward-dipping once a critical degree of extension was reached (Hill and Durrand, 1993). The early phase of extension initiated development of half-grabens (Penola, Robe and St Clair Troughs) and basement horsts.

The unconformity at the top of the Crayfish Group was probably formed as a result of regional thermal uplift, prior to the locus of extension shifting southward and the formation of a south-dipping detachment zone near the Tartwaup Hingeline (Hill and Durrand, 1993). This rather more narrow zone controlled deposition of the Eumeralla Formation until breakup near the Early/Late Cretaceous boundary.

Post-rift phase — Late Cretaceous Sherbrook Group

Estimates vary as to the exact time and mechanism of plate separation. Whether it was a multi-stage, multi-rate process or a single event (Willcox *et al.*, 1985) remains unresolved. However, the marked contrast in sedimentation patterns between the volcanoclastic, fluviolacustrine Otway Supergroup and the quartzose, marginal-marine to fluviodeltaic Sherbrook Group would support a breakup at the boundary of these groups (~95 Ma; Veevers, 1986). Breakup occurred earliest to the west in the Bight Basin, and only slow spreading occurred in the Otway Basin until the Eocene.

The development of an 'Outer Continental Margin Ridge' (Boeuf and Doust, 1975) at the inception of breakup created a rim basin (Veevers *et al.*, 1982) that dominated for up to 40 million years. Onlap of prograding Late Cretaceous deltaic sediments against this outer rim within a restricted sea is suggested (Veevers, 1984).

Major northwest-southeast-trending synthetic growth faults formed in delta depocentres as a consequence of high rates of sedimentation. Delta-front and prodelta deposits >3 km thick prograded into a juvenile ocean south of the Tartwaup Hingeline. Down-to-basin slumping and gravity sliding due to synsedimentary deformation resulted from rapid burial.

Individual faults, often concave to the basin, are of limited lateral persistence and are considered to have evolved independently from the Early Cretaceous back-to-basin normal fault blocks. Early Cretaceous faults are unlikely to have persisted into the Late Cretaceous south of the Beachport and Kalangadoo Highs.

Development of the northwest-trending highs in the Voluta Trough resulted from rotation of the downthrown strata to form broad rollover anticlinal complexes as the depocentre of the delta complex rapidly migrated basinward. Regional warping has dominated over faulting since the close of deposition of the Sherbrook Group.

Stratigraphy

Casterton Formation

The earliest unit deposited in the Otway Basin is the poorly known Casterton Formation, a succession of interbedded carbonaceous shale and minor feldspathic sandstone and siltstone, characterised by interbedded olivine basalt and volcanoclastics in Casterton 1 (Victoria). This unit has not been formally defined but the name is in common usage, although other names include 'Unit T' of Reynolds *et al.* (1966), 'Basal Unit' of Ellenor (1976), and 'Casterton Group undifferentiated' of Kopsen and Scholefield (1990). K-Ar dates of the basalt range from 153±5 Ma (*in* Harding, 1969) to 120±10 Ma (Reynolds, 1971), corresponding to a Late Jurassic to Early Cretaceous age. The Jurassic Wisanger Basalt from Kangaroo Island (Milnes *et al.*, 1982) may be a correlative. Only one drillhole in South Australia has intersected the unit, but it is recognisable in seismic sections in deeper areas. The unit has been dated palynologically as latest Jurassic to earliest Cretaceous (*Retitriletes watherooensis* to lower *C. australiensis* Zones; Kopsen and Scholefield, 1990).

Lower Otway Supergroup — Crayfish Group

Kopsen and Scholefield (1990) introduced the name Crayfish Group for the early syn-rift, graben-fill megasequence above the Casterton Formation, but below the major angular unconformity at the base of the Eumeralla Formation. The Crayfish Group consists of the Pretty Hill Formation, Laira Formation and Katnook Sandstone, but the stratigraphy is complex and the relationships between these units are not always clear (Figs 9.38, 9.39). Together, the Crayfish Group and Eumeralla Formation comprise the Otway Supergroup, a term that now has limited geological meaning.

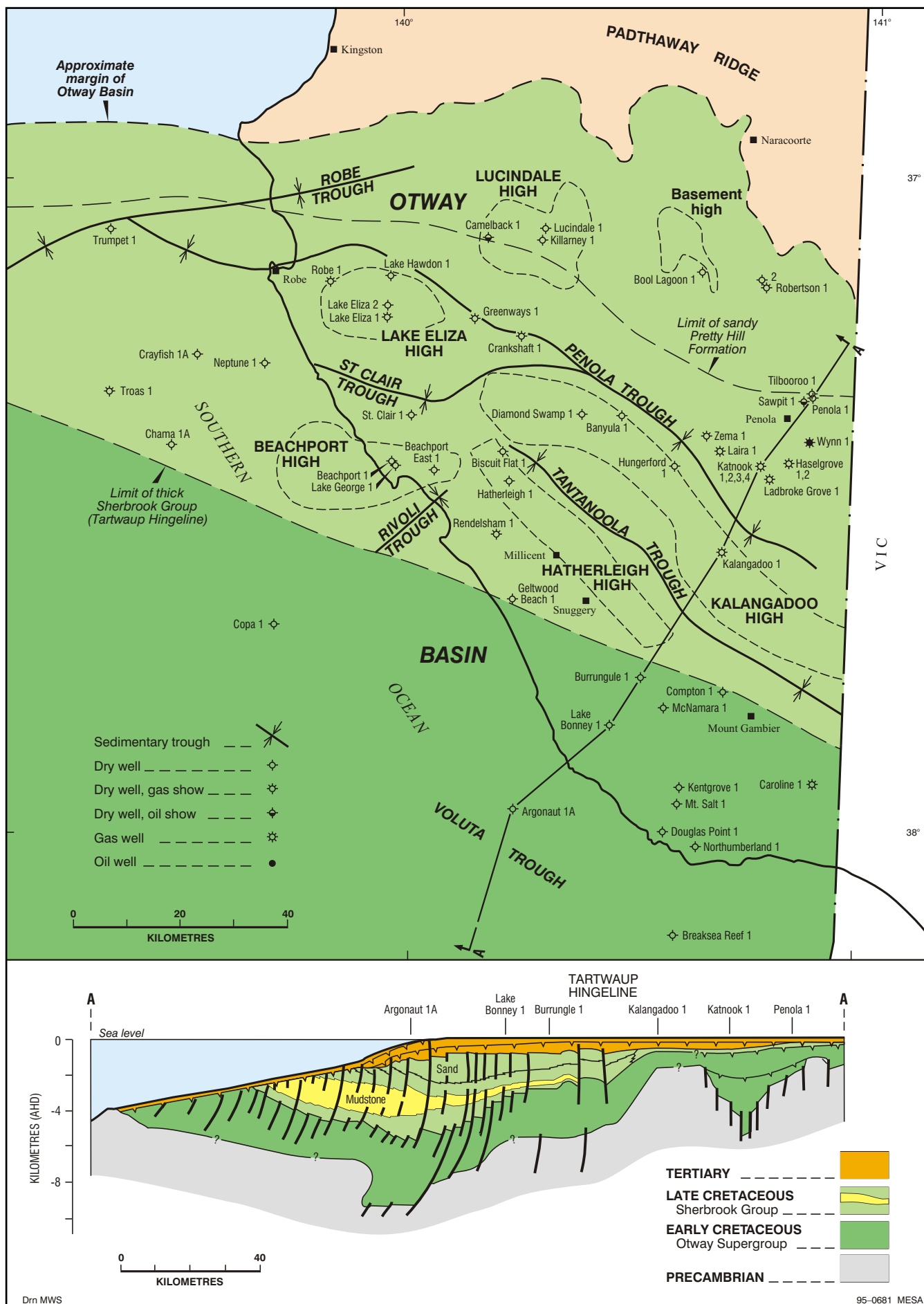


Fig. 9.37 Structural elements and schematic section across the Otway Basin.

Pretty Hill Formation

The ‘unnamed sandstone’ below the Eumeralla Formation recognised by Bain (1962) in Pretty Hill 1 (Victoria) was informally named Pretty Hill Sandstone by Edworthy (1963), and subsequent authors have used the name without formal definition (Reynolds *et al.*, 1966; Wopfner *et al.*, 1971; Reynolds, 1971; Ellenor, 1976; Abele *et al.*, 1976). Reynolds *et al.* (1966) maintained that the Pretty Hill Sandstone in Pretty Hill 1 is unique, and introduced the name ‘Geltwood Beach Formation’ for the more lithic and shaly sandstone intersected in other drillholes, but this name was rejected (Morton, 1990) since Geltwood Beach 1 did not penetrate the top of the Crayfish Group. ‘Crayfish Formation’ (from Crayfish 1A) has also been used informally for Pretty Hill Formation equivalents in South Australia. Kopsen and Scholefield (1990) subdivided the Crayfish Group into four units (A to D) of which A and B correspond to the Pretty Hill Formation. These names are unnecessary if reasonable lithological variability is allowed, reflecting different source areas but not different palaeoenvironments. The name Pretty Hill Formation is now applied to the predominantly arenaceous succession of the lower Crayfish Group (Morton *et al.*, 1994).

The unit generally consists of interbedded feldspathic litharenite and carbonaceous siltstone and mudstone, which are confined to structurally low areas such as the St Clair and Penola Troughs. Lithics include felsic volcanic grains exhibiting trachytic texture and devitrified glass, metamorphic grains (schist), and sedimentary grains (reworked siltstone, sandstone and chert). Accessory minerals typically include garnet, zircon, tourmaline, apatite and sphene. Authigenic chlorite is common, forming rims on grains and producing the characteristic green colour of the Pretty Hill Formation. Authigenic calcite and kaolinite are common cements, whereas laumontite is less common. Sedimentary structures in cores include high-angle cross-bedding (some with graded foresets), massive beds with water-escape structures, and climbing ripples. Slump features (interpreted to be bank slumps) are common throughout, and minor shale intraclast conglomerate is present. The unit is a gas reservoir in Katnook and Ladbroke Grove Fields.

The formation thickens westwards along the axis of the Robe-Penola Trough to over 1 585 m in Crayfish 1A. Pretty Hill Formation is possibly also present south of the Beachport and Kalangadoo Highs but no drillhole, with the possible exception of Copa 1, has penetrated the Crayfish Group. The formation ranges from upper *C. australiensis* to *F. wonthaggiensis* Zones of Morgan (1993), indicating an Early Cretaceous age.

The lithology and sedimentary structures suggest a low-sinuosity, fluvial palaeoenvironment similar to that proposed by Medwell (1977) for slightly younger sediments, equivalent to the Eumeralla Formation, in the Otway Ranges of Victoria. The unit is in some cases age-equivalent to the Laira Formation, and a complex palaeoenvironment of high gradient, local rivers flowing into restricted lakes is envisaged.

Laira Formation

This name was proposed by Morton (1990) for the predominantly shale unit above the Pretty Hill Formation and below the mid-Otway Supergroup unconformity. The unit was recognised by Kopsen and Scholefield (1990) as ‘Crayfish Group, Unit C’. The type section is in Katnook 2 between 1 979 and 2 842 m. It consists of monotonous shale and siltstone of slightly higher natural gamma radiation than Eumeralla Formation shales, and ranges in thickness up to 863 m. Algal material is common. The unit is wholly within the *F. wonthaggiensis* Zone of Valanginian to Hauterivian age (Kopsen and Scholefield, 1990). It is generally restricted to the Penola Trough area, and is lacustrine in origin.

Block of slumped channel-bank shale (lower part of photo) incorporated within point-bar sandstone which includes carbonised wood fragments, Pretty Hill Formation at 2 400 m in Banyula 1. Photo 039490.

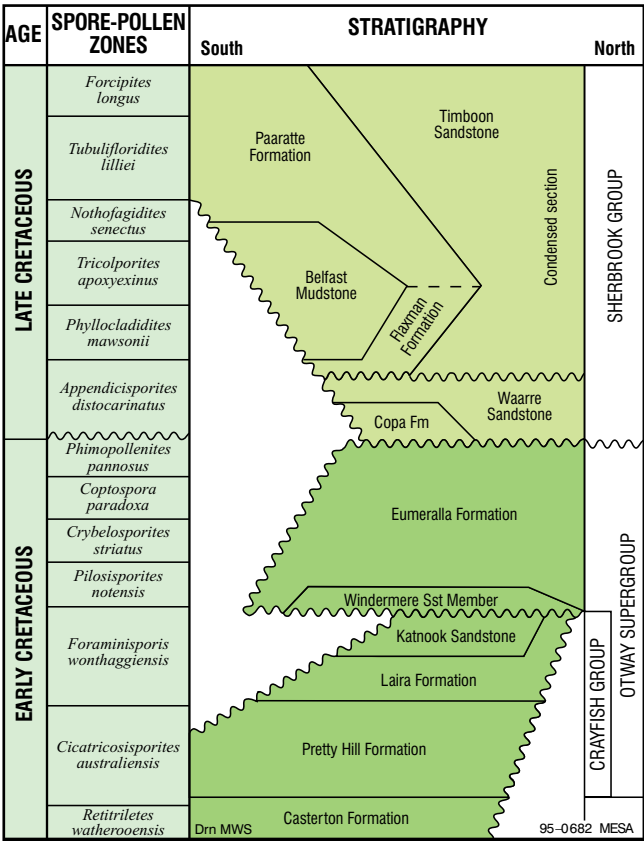


Fig. 9.38 Stratigraphic nomenclature of the Otway Basin.



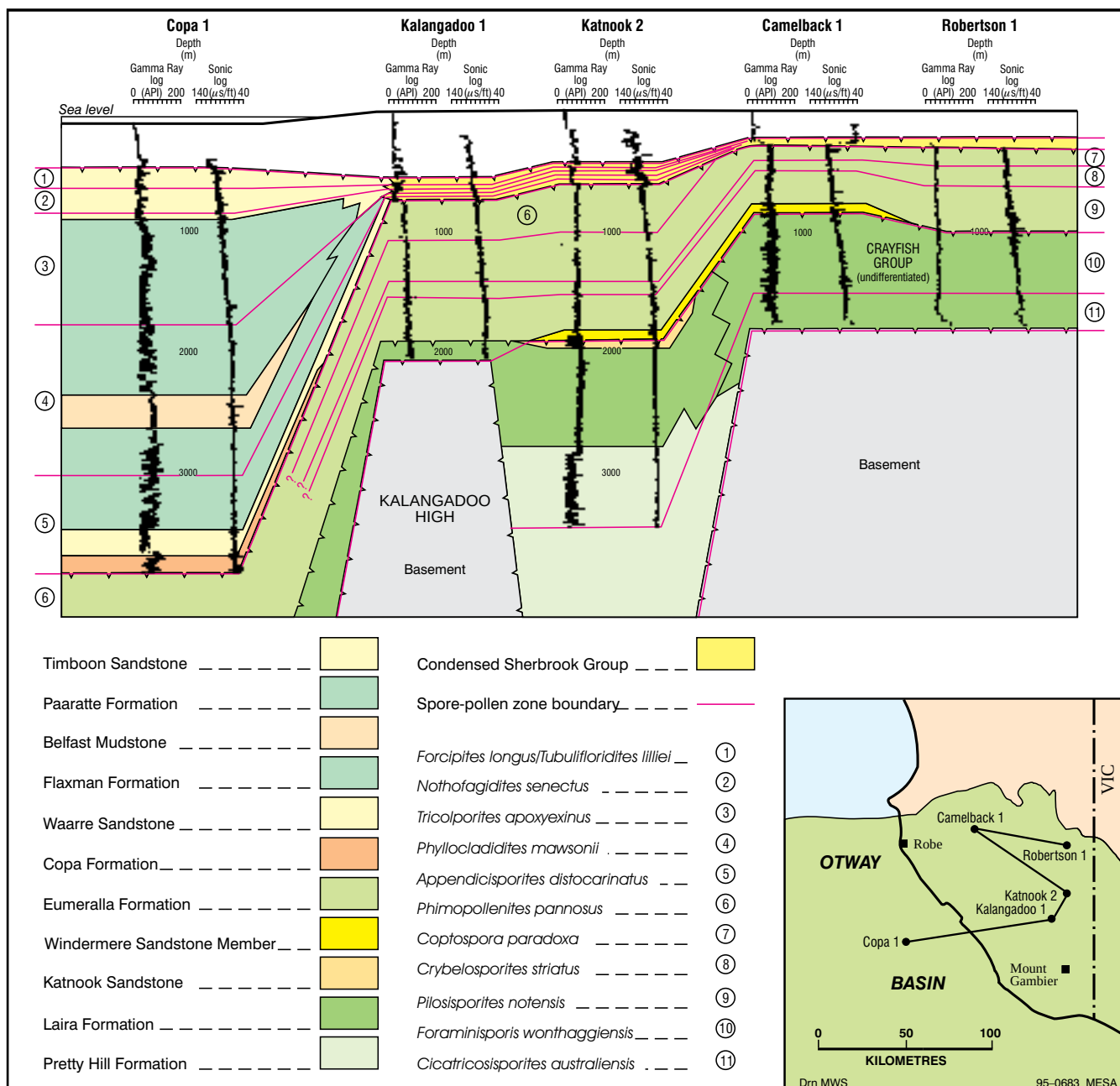


Fig. 9.39 Correlation of Otway Basin petroleum drillhole logs.

Katnook Sandstone

The name Katnook Sandstone was proposed by Morton (1990) for sandstone above the Laira Formation and below the basal Eumeralla Formation unconformity. The unit was recognised by Kopsen and Scholefield (1990) as 'Crayfish Group, Unit D' and its type section is in Katnook 2 between 1 892 and 1 979 m. It consists of sand and interbedded shale, with a maximum thickness of 530 m in Trumpet 1 (Kopsen and Scholefield, 1990). The unit is better developed to the west and northwest (St Clair and Robe Troughs), and in the Penola Trough. It appears to be restricted to the deeper parts of the troughs, but these occurrences may be erosional remnants only. The unit is within the upper part of *F. wonthaggiensis* Zone of Hauterivian age.

Upper Otway Supergroup

Eumeralla Formation

Reynolds *et al.* (1966) proposed the name Eumeralla Formation for the monotonous sequence of chloritic siltstone, shale and

subordinate sandstone of the upper Otway Supergroup in Eumeralla 1 (Victoria). The Eumeralla Formation, which forms the major portion of upper Otway Supergroup sediments, either unconformably overlies pre-Mesozoic basement or conformably overlies Windermere Sandstone Member. The formation ranges in thickness from less than 500 m in Bool Lagoon 1 to 2 350 m in Geltwood Beach 1, and consists of laminated greenish grey, micaceous, carbonaceous silty claystone, with interbeds of very fine-grained sandstone showing climbing ripples. Minor associated features seen in cores include well-preserved plant fragments, slumped and contorted beds, rare bioturbation, massive medium-grained sandstone, shale intraclast conglomerate, and thin coal beds. The formation ranges in age from Barremian to Albian (*P. notesis* to *P. pannosus* Zones).

Kopsen and Scholefield (1990) proposed the term 'Windermere Sandstone Member' for basal sandstone of the Eumeralla Formation. The member generally occurs only in the main axis of the Robe and Penola Troughs, and unconformably overlies lower Otway Supergroup units. The unit is within the *C. hughesii* Zone of Aptian age and its type section is in Windermere 2 (Victoria)

between 3 187 and 3 292 m. It is generally less than 10 m thick in South Australia but ranges up to 100 m in western Victoria. The Windermere Sandstone Member is the reservoir unit of the Katnook 2 gas discovery.

The dominance of fine-grained (mudstone and siltstone) over coarser grained lithologies suggests a relatively low-energy palaeoenvironment compared to the Pretty Hill Formation. The fine-grained sandstones with climbing ripples and rare bank-slump features suggest the presence of minor high-sinuosity streams flowing over a wide flood plain. Lacustrine and back-swamp environments were probably also present. To the east, sandstone in the Otway Ranges, equivalent in age to the basal Eumeralla Formation, has palaeocurrent directions indicating a westward-flowing river system (Medwell, 1977). The presence of dinoflagellates and foraminifera (Reynolds, 1971) in the western portion of the basin also suggests that parts of the formation may be marginally marine, and may have had connection to fully marine environments to the west in the Bight and Duntroon Basins.

Sherbrook Group

For most of the Late Cretaceous, Australia and Antarctica were separated by a narrow, rapidly subsiding basin. The depositional model for the Sherbrook Group is one of a large, deep, saline lake, probably connected with a restricted sea further west and east, with very large prograding deltas feeding into the basin at right angles to the main axis of rifting. Areas to the north that were not underlain by thinned continental crust or new oceanic basement, and hence underwent much less subsidence, have a condensed Sherbrook Group sequence that is at least an order of magnitude thinner, and which is probably mostly of fluvial origin. Thus, sediment sourced in areas to the north was 'by-passing' this relatively stable marginal area to be 'dumped' in the rapidly deepening depocentre to the south. There is also an apparent east to west change in sedimentation style comparable to, but in an opposite sense from, that found in the upper Otway Supergroup. There are clear indications of fully marine conditions in the eastern part of the basin in Victoria, which decrease westwards; the glauconitic Nullawarre Greensand Member is not found in South Australia and marine molluscs and glauconite found in the Belfast Mudstone in Victoria are not found in the South Australian portion of the basin. This suggests that the deep east-west elongate basin was open to the east as well as to the west. The black, pyritic nature of the Belfast mudstone strongly suggests that anoxic conditions existed within the Sherbrook Group, indicating a stratified and non-circulating lake or restricted sea.

Copa Formation

This name was introduced by Morton (1990) for a predominantly shaly unit underlying the Waarre Sandstone and overlying the Otway Supergroup-Sherbrook Group unconformity. The unit consists of medium grey-brown claystone and dark grey to green carbonaceous siltstone, with minor fine to medium-grained sandstone. The unit has been recorded in only three drillholes, Copa 1, Caroline 1 and Burrungule 1, and no cores have been cut in South Australia. The thickness ranges from 51 m (Caroline 1) to 167 m (Copa 1). Palynological age determinations are limited and indicate a Cenomanian age (*A. distocarinatus* Zone).

The limited number of drillhole intersections and lack of cores in this unit preclude any definite statement as to palaeoenvironment but the scattered occurrence south of the Beachport and Kalangadoo Highs suggests meandering fluvial to lacustrine, possibly upper delta plain.

Waarre Sandstone

McQueen (1961) introduced the name 'Waarre Formation' for a unit of mixed lithology overlying the Eumeralla Formation and underlying the Belfast Mudstone. Bain and McQueen (1964) restricted the definition to the dominantly coarse-grained sandstone section above the Copa or Eumeralla Formations and below the Flaxman Formation. The unit consists of light grey, medium to very

coarse-grained sandstone, with subordinate grey, carbonaceous siltstone and mudstone interbeds. The unit is somewhat patchy in distribution (similar to that of the Copa Formation) in areas south of the Beachport and Kalangadoo Highs. Thickness ranges from 52 m (Caroline 1) to 108 m (Burrungule 1). Palynological age determinations are limited and indicate a Cenomanian age (*A. distocarinatus* Zone).

The Waarre Sandstone, deposited in an upper delta plain to low-sinuosity fluvial environment, marks the beginning of a prograding delta complex that culminated in the Belfast Mudstone. This sudden change of depositional style from the Early Cretaceous is evidence for more rapid rifting at this time, and the creation of a deep rift basin to the south.

Flaxman Formation

Bain and McQueen (1964) first published the name 'Flaxmans Beds', nominating Port Campbell 2 (Victoria) as the type drillhole, but no interval was specified as the type section. Bock and Glenie (1965) applied the term Flaxman Formation to a unit transitional between the Waarre Sandstone and Belfast Mudstone. The name is used in South Australia for a unit of mixed lithology below the Belfast Mudstone and above the Copa and Eumeralla Formations or Waarre Sandstone. The formation is restricted to the southern portion of the Otway Basin but, because of the great thickness of the Sherbrook Group, has been fully penetrated in relatively few drillholes. To the north it probably passes laterally into a condensed Sherbrook Group sequence. On petrophysical logs, the formation appears as interbedded coarsening-upward sandstone and fine-grained units. In cores, the unit consists of ubiquitously bioturbated fine-grained sandstone interbedded with carbonaceous, micaceous, muddy siltstone. Associated sedimentary structures include slumped beds, climbing ripples, rare pillar structures and some red oxidation of sandstone. Where the Belfast Mudstone is absent, the Flaxman Formation cannot be distinguished from the Paaratte Formation. Thickness ranges from 33 m (Burrungule 1) to 619 m (Copa 1). The unit is entirely of Turonian to Santonian age (*P. mawsonii* to *T. apoxyxinus* Zone).

The Flaxman Formation is the first unit of a major episode of deltaic deposition which dominates the Sherbrook Group. Coarsening-upward sandstone beds were deposited by delta distributaries on the lower delta plain, and the ubiquitous bioturbation in this unit implies deposition near a large, permanent body of water, possibly marine to restricted marine (Reynolds, 1971; Glenie, 1971; Ellenor, 1976).

Belfast Mudstone

The name Belfast Mudstone was first published by McQueen (1961) for black glauconitic mudstone overlying the Waarre Formation and comprising the basal member of the Paaratte Formation. He noted that it had been discovered in Belfast 4, (Victoria) but did not nominate a type section. Bain and McQueen (1964) nominated Port Campbell 1 (Victoria) as type section, which was subsequently redefined by Glenie (1971). The unit is restricted to the southern area of thick Sherbrook Group deposition; to the north it pinches out and the Paaratte and Flaxman Formations become a single unit. On petrophysical logs, the Belfast Mudstone has a uniform high gamma ray signature with some sandstone interbeds of Flaxman-Waarre-Paaratte Formation aspect. Core recovery is generally low but, where recovered, the Belfast Mudstone consists of black, massive, very pyritic and carbonaceous mudstone. Sedimentary structures in the very fine-grained sandstone to siltstone beds include climbing ripples and ubiquitous bioturbation. The upper and lower contacts are conformable and strongly diachronous, younging away from the basin margin. Penetrated thickness ranges to over 1 355 m in Breaksea Reef 1. The unit thickens to the south and ranges in age from Turonian to Cenomanian (*P. mawsonii* to *N. senectus* Zones).

The Belfast Mudstone is a complex of upper prodelta, slope and interdistributary deposits, and upper delta-front deposits. The highly carbonaceous and pyritic lithology suggests deposition in either a marine or brackish body of standing water. The most landward occurrence may correspond to a transgressive maximum, due either to eustatic sea-level rise, or a phase of maximum subsidence within the main rift. However, unequivocal evidence of marine sediments is lacking in South Australia.

Paaratte Formation

The name 'Paaratte Formation' was first published by McQueen (1961) for the formation above the Belfast Mudstone 'consisting of interbedded sandstone, siltstone and mudstone, dolomitic, pyritic and in parts containing marine fossils'. Glenie (1971) redefined the type section as the interval from 1 295 to 1 501 m in Port Campbell 1. The formation thickens to the south; to the north it passes laterally into the Timboon Sandstone and further north into a condensed Sherbrook Group sequence. Both the upper and lower contacts of the formation are strongly diachronous.

In cores, the formation consists of an interbedded sequence of black carbonaceous shale and bioturbated, fine to coarse-grained, quartz-rich sandstone. The sandstone beds have a coarsening-upward gamma ray log signature, similar to the Flaxman and Waarre Formations; the upper contact is either conformable, passing into the Timboon Sandstone, or unconformable with the Tertiary Dilwyn Formation. Thickness ranges from 421 m (Caroline 1) to 1 836 m (Breaksea Reef 1), and age range is Santonian to Maastrichtian, based on the *P. mawsonii* to *Tubulifloridites* (ex *Tricolporites*) longus Zones.

The Paaratte Formation was deposited in a lower delta-plain environment; distributaries prograded south over the delta-front and interdistributary sediments of the Belfast Mudstone. Lagoonal and marginal shoreface environments are also probably represented by the Paaratte Formation (Reynolds, 1971).

Timboon Sandstone

The 'Timboon Sand' was initially proposed as a member of the Paaratte Formation by Bock and Glenie (1965) for a largely non-marine quartz sandstone unit above the Paaratte Formation *sensu stricto* and below the Tertiary Pebble Point Formation; the unit is synonymous with Curdies Formation of the earlier literature. Glenie (1971) further defined the unit, and suggested Port Campbell 1 (856 to 1 295 m) as the type section. Morton (1990) defined the unit as Timboon Sandstone.

The Timboon Sand is distributed as a relatively narrow band of thick sand over the nearshore area of thick Sherbrook Group deposits. To the south the unit passes laterally into the Paaratte Formation, and to the north passes into part of the condensed Sherbrook Group (Fig. 9.39). The gamma ray log signature through the Timboon Sandstone is typical of a clean sandstone, with relatively minor shale interbeds.

In cores the unit consists of medium to coarse-grained massive quartz sandstone commonly with ferruginous cement and oolites, with minor ripple cross-laminated fine sandstone and siltstone. This is interbedded with black to brown, very micaceous silty mudstone, commonly bioturbated. The upper contact with the Dilwyn Formation is unconformable. Thickness ranges from 656 m (Burrungule 1) to 1 590 m (Argonaut 1A). Age range is Late Cretaceous (Santonian to Maastrichtian), based on *T. apoxyxenus* to *F. longus* Zones.

The Timboon Sandstone is regarded as a regressive unit deposited on the upper delta plain during the last phase of the prograding deltaic sequence of the upper Sherbrook Group. It was presumably formed in a non-marine environment, but bioturbation present in cores suggests an intermittent marginal-marine influence.

Katnook 3 being drilled in January 1990. The 3105 m drillhole flowed gas at 716 000 m³/day and condensate at 11 300 L/day during drill stem testing of the Pretty Hill Formation, and is now in production. Photo 042490.

Economic geology

Petroleum

There are three main hydrocarbon reservoirs containing economic gasfields in the Otway Basin. These are the Pretty Hill Formation, Windermere Sandstone Member, and Flaxman-Waarre Formations. Oil has also been recovered from fractured basement in Sawpit 1, and from the Pretty Hill Formation in Wynn 1.

The main source rocks of the Otway Supergroup are the Casterton, Laira, Pretty Hill and Eumeralla Formations. Rock Eval pyrolysis data indicate a mainly gas-prone, fair source rock potential (Felton and Jackson, 1987) for these formations, and vitrinite reflectance (V_r) data for the Laira Formation indicate peak levels of gas generation in the deepest parts of the St Clair and Penola Troughs ($V_r = 0.8$ to 1.2). Peak oil generation ($V_r = 0.9$ to 1.4) occurs slightly lower in the succession.

Commercial gas deposits occur in the Katnook, Hazelgrove and Ladbroke Grove Fields, where the majority of reserves are in the Pretty Hill Formation, sealed by the Laira Formation, in faulted anticlinal structures. Reservoir quality within the Pretty Hill Formation is variable, but core data have demonstrated porosities in excess of 25% and permeabilities over 1 darcy. Gas flow rates up to 4.6×10^5 m³ (16.4 million cubic feet) per day and condensate flows of up to 5.7×10^4 L (362 barrels) per day were recorded from the Pretty Hill Formation in Katnook 2. A smaller gas reservoir was also discovered in the Windermere Sandstone Member in the Katnook Field. Gas compositions vary; the Katnook Field has a CO₂ content of less than 0.1% but in the Ladbroke Grove Field it is over 50%.

The Sherbrook Group contains Caroline CO₂ field, reservoired in the Waarre and Flaxman Formations, and sealed by the Belfast Mudstone. The CO₂ is presumed to have originated from the Holocene Mount Gambier volcanic chain. Structures are generally tilted step-faulted blocks, but stratigraphic targets may exist in sandstone interbeds within the Belfast Mudstone. The Waarre and Flaxman Formations are the main target for exploration in the Victorian portion of the basin, with gas tested at flow rates up to 2.7×10^5 m³ (9.6 million cubic feet) per day in the North Paaratte Gasfield. Carbon dioxide content is variable but the Caroline field in South Australia and the Pine Lodge 1 discovery near the Victorian-South Australian border contain nearly 100% CO₂. Porosities and permeabilities are very high, but prospectivity for hydrocarbons is restricted onshore due to the CO₂ risk and lack of Belfast Mudstone seal north of Lake Bonney 1. Source rocks of the Sherbrook Group are marginally mature (up to $V_r = 0.6$) and gas prone (Felton and Jackson, 1987).

There is a reasonable chance (50% probability) of finding more than 930 PJ of sales gas or 30×10^6 kL (190 million barrels) of recoverable oil from the three main hydrocarbon plays in the onshore portion of the Otway Basin in South Australia (Morton, 1992).



MESOZOIC IGNEOUS ROCKS

M.G. Farrand

Igneous activity in South Australia during the Mesozoic consisted of the intrusion of kimberlitic lamprophyres, extrusion of the enstatite-bearing Wisanger Basalt on eastern Kangaroo Island, and incorporation of volcanic detritus in Otway Supergroup sediments. Sedimentary evidence of Mesozoic volcanism is described in the Pretty Hill Formation of the Otway Basin. Known kimberlitic lamprophyres occur in an area extending from Elliston on western Eyre Peninsula to Blanchetown on the Murray River in the east, and to Eurelia, possibly as far as the Springfield Basin, in the north. Lamprophyres outside this roughly triangular area consist of Proterozoic kimberlites, Ordovician lamproites (see Ch. 7), and lamprophyres of indeterminate age.

Kimberlitic lamprophyres

Kimberlitic rocks were first identified in South Australia near Terowie in 1969 (Fig. 9.42; Colchester, 1972), and the greatest concentration of pipes and dykes occurs in this vicinity. Over 70 kimberlites are now known throughout the State although the only locality where they are known to be the primary host of diamonds is Eurelia. There is a high probability that a primary kimberlitic source no younger than Permian underlies the diamondiferous placer of the Springfield Basin, 80 km northeast of Port Augusta, but this has not been confirmed. In addition, a suite of lamproite intrusions has been recognised in the Truro region, but geochronology indicates that these are late Delamerian (see Ch. 7).

The Eurelia group of kimberlitic rocks, comprising more than 25 dykes up to 30 m wide and 1 km long, forms a northwest-trending zone 20 km long and 5 km wide (Scott Smith *et al.*, 1982, 1984). The dykes intrude Neoproterozoic (Adelaidean) sediments and almost all strike northwest-southeast. Fresh rock from drillcore consists of carbonate-bearing micaceous kimberlite. Dykes, and two possible plugs, in the southernmost cluster contain up to 80% calcite and intrude breccia of the Walloway Diapir. U-Pb dating on zircon provides an age of 170 Ma for two of the dykes near Eurelia (Scott Smith *et al.*, 1984; Black *et al.*, 1994), confirming Jurassic ages obtained previously by K-Ar and Rb-Sr dating of kimberlitic lamprophyres. Seven of the dykes yielded 140 small diamonds, the majority from three dykes (Scott Smith *et al.*, 1984). Three rock types were distinguished, of which only two are diamondiferous. The barren type also lacks garnet.

The most abundant lamprophyric material in South Australia makes up a group of intrusions around Terowie and Whyte Yarcowie (Fig. 9.42). The group includes five clusters of dykes and pipes intruding middle Adelaidean sediments in a zone 45x25 km with a north-northeast-trending axis. Some dykes strike north-northeast or east-west, but the majority strike northwest. The northernmost cluster consists of five small, highly weathered micaceous dykes at Nackara. About 15 km to the south, a cluster of four pipes, four dykes and a dyke swarm of phlogopite kimberlite is known as the Calcutteroo or Pandappa cluster. Five kilometres to the east, the Pine Creek cluster contains three dykes and the largest pipe in the province, measuring 650x150 m; no diamonds were recovered from 1 100 t of kimberlite, mainly from the Pine Creek pipe (Sas *et al.*, 1981). A cluster of 12 pipes and dykes of mica-poor basaltic kimberlite is sited 15 km southwest of Calcutteroo near Wonna Creek; this group is also known also as the Mittopitta or Mungibbie cluster. A further 10 km to the south, two dykes of mica peridotite constitute the Ulooloo cluster.

A group of three sills up to 2 m thick intrudes Wilpena Group (Adelaidean) sandstone over 25 km along a northwest-southeast trend between 'Corraberra' and a point 8 km southeast of El Alamein Camp near Port Augusta (Stracke, 1974). Outcrops are deeply weathered and consist of ferruginous clay with mica and serpentinite relics. The rock is identified as lamprophyric from xenoliths of eclogite and granulite.

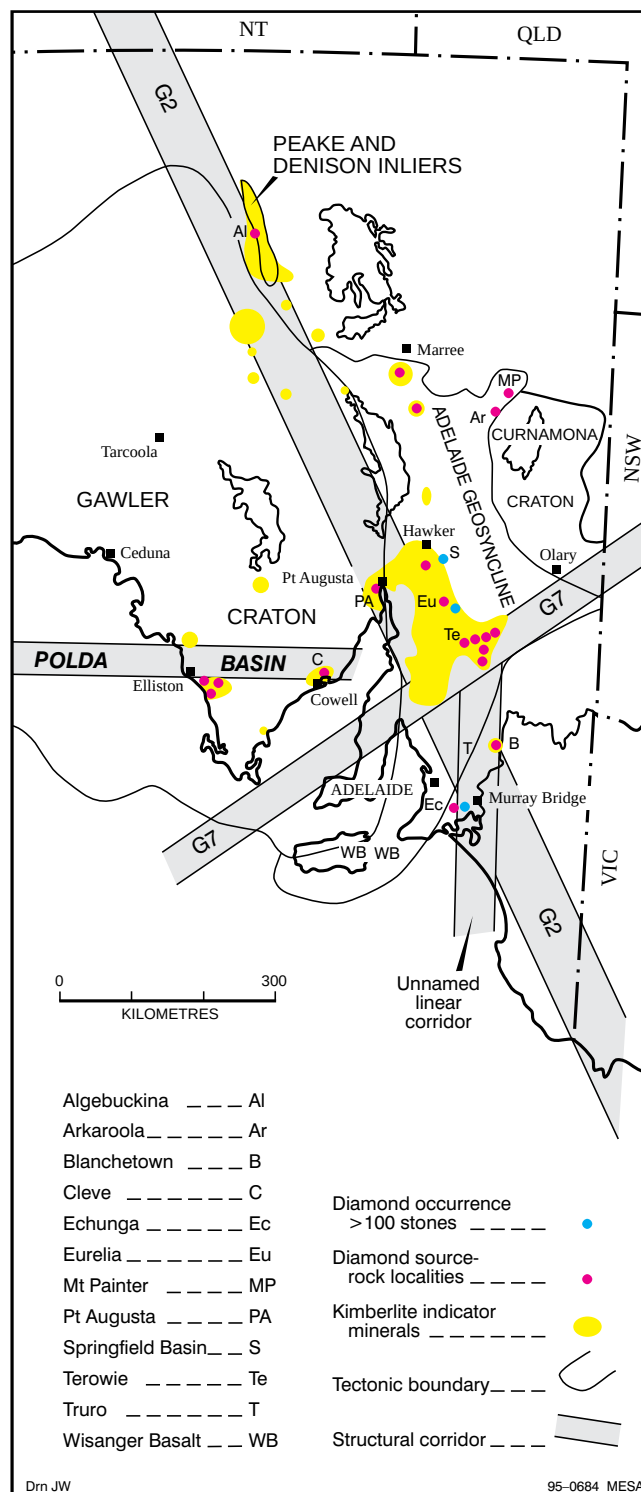


Fig. 9.40 Mesozoic diamond occurrences, sources, indicators and tectonic environment.

More recent exploration has delineated a group of eight subcropping kimberlites of crater and subvolcanic facies southeast of Elliston on the west coast of Eyre Peninsula (Mitchell, 1992a,b; Wilson *et al.*, 1992). Dates of between 212 and 177 Ma were obtained, with a preferred average of 190 ± 17 Ma (U/Pb method on perovskite). Abundant indicator minerals but no diamonds were recovered.

On the eastern side of Eyre Peninsula a kimberlitic pipe and dyke system has been located at Mount Desperate, 14 km north-northeast of Cleve (H.R. Robison, Stockdale Prospecting Ltd, pers. comm., 1989). An age of 180 ± 3 Ma was obtained using the U/Pb method on groundmass perovskite. No diamonds were recovered.



Sample of kimberlite from Calcutteroo, near Terowie. Phlogopite laths 0.5–3 mm across are set in a groundmass of fine-grained phlogopite, clay and serpentine. An ovoid autolithic xenolith is at the base of the specimen. Photo 024578.

(Wyatt *et al.*, 1991). In the same region, a porphyritic amphibole lamprophyre (vogesite) was intersected 27 km northwest of Cowell.

Less well-authenticated kimberlites have been reported. A kimberlitic plug may underlie decomposed micaceous rock in 'Vermiculite Valley', 30 km northeast of Arkaroola (Crawford, 1963). Brecciated amphibole-phlogopite-carbonate rocks at Wywyana Creek, 10 km south of Mount Painter may be part of a carbonatite diatreme (Lottermoser, 1988). Host rocks are early Adelaidean but the age of the breccia is not known. Brecciated olivine-bearing rocks intersected in two drillholes to the west of the Murray River, 10 km north of Blanchetown, are probably lamprophyric (Lewis, 1985).

Geochemical compositions of the Australian kimberlitic suite plot towards the more evolved alkaline end of a differentiation trend encountered in similar rocks of southern Africa (Ferguson and Sheraton, 1979). Estimated pressures and temperatures of nodule assemblages indicate a steeper than average P-T gradient leading to a high degree of partial melting of parent peridotite at relatively shallow levels. The depth of origin suggested for most South Australian lamprophyres is about 70 km and diamond prospectivity is correspondingly poor (Ferguson *et al.*, 1979). However, intrusions of the Eurelia cluster are exceptional and formed by <1% melting of garnet lherzolite at depths in excess of 125 km (Ferguson and Sheraton, 1979) with a consequently enhanced diamond potential. That the stones are microdiamonds may be the result of only marginal diamond stability at this depth. Magma forming one of the barren sills in the Port Augusta group is estimated to have equilibrated at a depth of 120 km. The diamonds may have originated at a depth in the order of 800 km on the evidence of syngenetic magnesio-wüstite and enstatite inclusions in stones from Eurelia (Scott Smith *et al.*, 1984).

Many lamprophyric intrusions are located in areas of high seismicity; a compilation of seismic centres and structural corridors identified by O'Driscoll (1981) is provided by Preiss (1987, p.261). Ferguson *et al.* (1979) suggested that contemporary seismic activity marks the continental projection of an oceanic fracture zone active in the widening of the Southern Ocean. This trends north-northwest from the coast near the Victorian border to the eastern end of the Musgrave Block as a broad structural and gravity corridor without sharply defined margins (the G2 Corridor of O'Driscoll, 1981). The groups of intrusions making up the nucleus of kimberlite magmatism are located within or marginal to G2 where it is intersected by G7, a projection of the Darling Lineament of New South Wales. Dyke trends and the elongation of clusters usually reflect one or more of the major structures. The diamondiferous Eurelia cluster is close to the northeastern margin of G2 where a major gravity low may indicate downthrown (cool)

upper crust which possibly promoted the stability of diamonds above the normal limit of 140 km. In addition, escape of heat along mantle-tapping structures may have led to a cooling effect at depth, a downwarp of the geotherm, which could enlarge the diamond window.

Structural associations and the Middle Jurassic age of most Mesozoic lamprophyres suggest that these rocks are products of thermal activity in the mantle related to rifting which was the precursor to the opening of the Southern Ocean.

Wisanger Basalt

The Wisanger Basalt, which crops out as a fragmented sheet extending 20 km westwards from Kingscote on Kangaroo Island, flowed over fluvial sediments of probable Permian age. A slightly differentiated variant forms an isolated outcrop near Penneshaw (Milnes *et al.*, 1982). The main outcrop is up to 30 m thick, with columnar jointing in places. The rock is amygdaloidal and characterised by porphyritic labradorite, augite, pigeonite and enstatite but some glass remains in the fine-grained groundmass. An age of 170 Ma has been obtained by the K-Ar method (McDougall and Wellman, 1976). Chemistry and high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios indicate an affinity with Tasmanian and Antarctic tholeiitic dolerite and basalt (Compston *et al.*, 1968). These tholeiites are attributed to the tensional regime which resulted in the separation of Australia and Antarctica (Milnes *et al.*, 1982). In this sense, and in the congruent ages, the Wisanger Basalt is associated with the lamprophyric suite. A mineralogical similarity is that enstatite is a mantle pyroxene and enstatite inclusions have been found in South Australian diamonds.

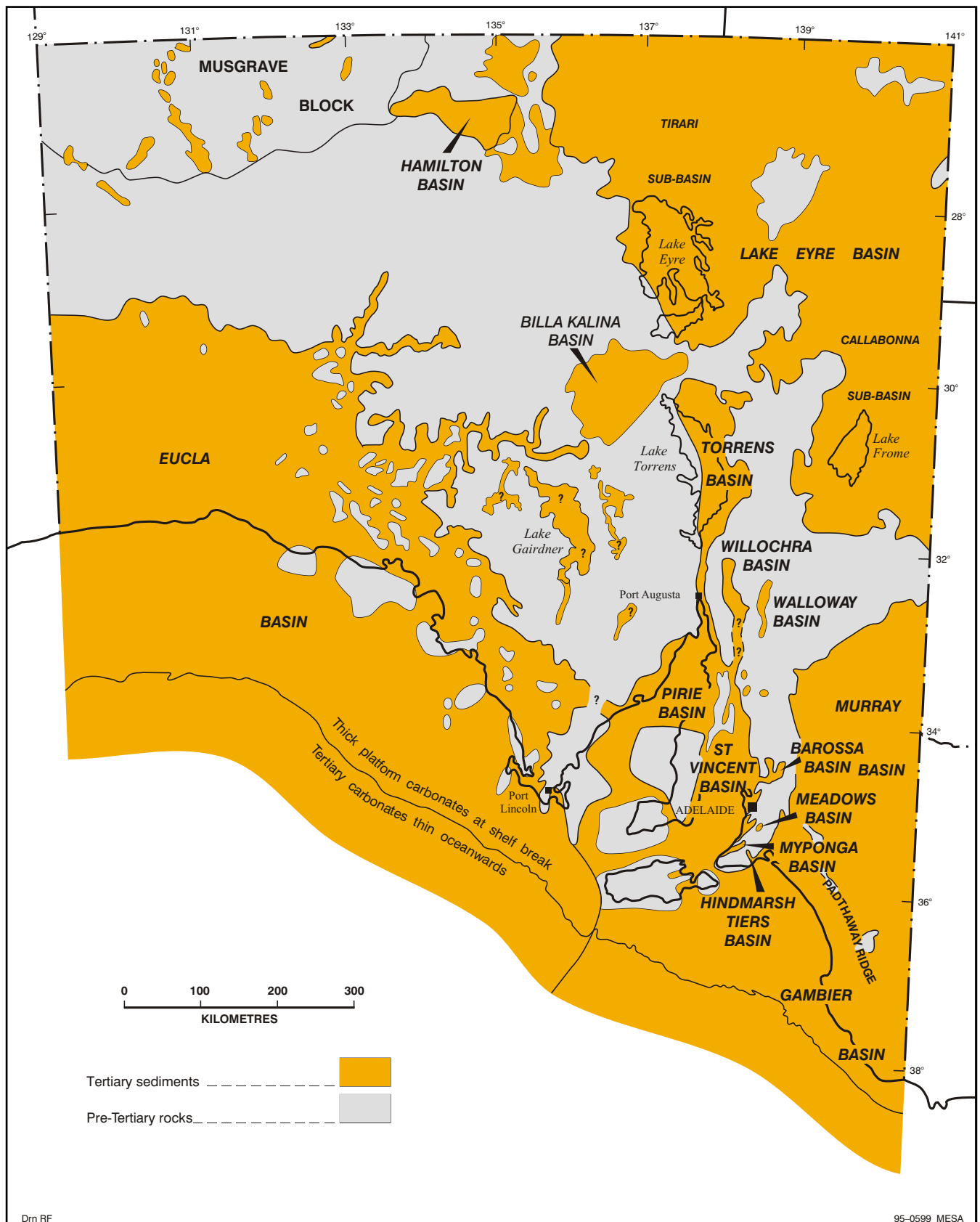


Fig. 10.1 Distribution of Tertiary basins in South Australia.

Chapter 10

TERTIARY

Compiled by N.F. Alley and J.M. Lindsay

with contributions from N.F. Alley, S.R. Barnett, M.C. Benbow, R.A. Callen, W.M. Cowley,
D. Greenwood, G. Kwitko, K.L. Lablack, J.M. Lindsay, P.A. Rogers, P.C. Smith and M.R. White

INTRODUCTION

M.C. Benbow, N.F. Alley and J.M. Lindsay

Tertiary strata are widespread in South Australia, occupying gently downwarped cratonic basins, palaeochannels, and occurring as extensive thin sheets in the interior (Fig. 10.1). Thicker successions accumulated in passive continental margin basins, where they unconformably overlie Mesozoic rift-basin sediments. Exposed sediments in the inland are extensively weathered and often capped by siliceous and ferruginous duricrusts. Although there are excellent coastal sections in the St Vincent Basin, along the coast of the Great Australian Bight and inland along the Murray River, understanding of the Tertiary stratigraphic record relies heavily on drillhole data. The stratigraphic record in the southern basins is similar and they are therefore grouped here as 'southern marine basins', while the inland depocentres beyond the limits of marine influence are grouped as 'interior non-marine basins'. A few small 'intramontane basins' record relatively brief intervals of sedimentation.

Deposition of relatively thin, fluvial and lacustrine, carbonaceous and arenaceous sediments characterised the Palaeogene inland depocentres. Thin but extensive lacustrine argillaceous and carbonate mudstone were laid down during the Neogene in a subdued landscape. Uplands were located in similar areas as now and, together with the Great Dividing Range of Eastern Australia, were major sources of terrigenous sediment.

A consequence of the separation of Australia from Antarctica was the initiation of marine transgressions into the southern marine basins and deposition of considerable thicknesses of marine sediments. Thus, the Tertiary succession here passes upwards and seawards from early Tertiary non-marine clastic sediments into marine temperate water limestones. The carbonates, up to 1.5 km thick, which were laid down in extensive platform and deep water environments at the continental margin, now occur beneath the continental shelf and slope.

Although Tertiary sediments often contain a rich array of macro- and microfossils, the foraminiferal faunas and palynofloras provide the most valuable biostratigraphic framework for dating and correlation between basins and with the international time scale. There are, however, significant biostratigraphic problems. Because of the extratropical nature of the local foraminifera, few of the biostratigraphic events on which the P-zones were based can be recognised (McGowran, 1979). Dating of inland sediments is difficult because of weathering and the condensed stratigraphic record.

Dating and correlation of non-marine sediments rely heavily on palynological schemes developed for marine sequences of the Gippsland, Gambier and Murray Basins (Harris, 1971, 1985; Stover and Partridge, 1973, 1982). Modified versions of these systems can be applied with reasonable confidence to the early Tertiary in South Australia, but not reliably to the middle and late Tertiary. The latter is largely related to the increasing regionalisation of vegetation through the Tertiary (see 'Geological history and palaeoclimate'). Dating of the early Tertiary is also complicated by earlier first appearances of some species in inland successions than in the southern basins (Alley and Benbow, 1989; Macphail and Truswell, 1989; Alley and Beecroft, 1993; Alley and Sluiter, in prep.; Beecroft *et al.*, in prep.). The palynostratigraphic scheme followed in this chapter is largely that of Stover and Partridge (1973, 1982; Fig. 10.2).

SOUTHERN MARINE BASINS

GAMBIER BASIN

P.C. Smith, P.A. Rogers, J.M. Lindsay, M.R. White and G. Kwitko

Structure and setting

Cainozoic tectonic movements divided the onshore Tertiary sediments in the southeast of the State and adjacent Victoria into separate basins, the westernmost of which is here referred to as the Gambier Basin (Fig. 10.3). The basin is synonymous with the 'Gambier Embayment of the Otway Basin' (Spencer-Jones *et al.*, 1971) but in this volume is restricted to Tertiary strata whereas the term 'Otway Basin' includes only the Mesozoic succession (see Ch. 9). The meridional trends of these southeastern basins and intervening highs are thought to have been controlled by reactivation of Palaeozoic structures of the Tasman Fold Belt System (Gravestock *et al.*, 1986).

The Tertiary sediments reach a maximum onshore thickness of ~1 000 m in the Port MacDonnell area, thinning southwards offshore. They overlie the Late Cretaceous Sherbrook Group, generally unconformably, except on the northern margin of the basin where they overlie the Early Cretaceous Otway Group and early Palaeozoic basement of the Padthaway Ridge (Fig. 10.4). Although the Padthaway Ridge forms a barrier between the Gambier and Murray Basins, there was interbasinal connection especially during the Late Oligocene transgressive peak. The boundary between the two basins is therefore diffuse and generally taken to lie along the basement high of the Padthaway Ridge where the Tertiary sediments are thinnest.

Sedimentation in the Gambier Basin commenced in the Late Paleocene to Middle Eocene with the paralic Wangerrip Group (Figs 10.2, 10.4). Increasing marine influence led to deposition of the Middle to Late Eocene marginal-marine Nirranda Group, and the open-water marine Gambier Limestone of latest Eocene to early Middle Miocene age.

A period of eustatic sea-level fall and tectonism during the remainder of the Miocene resulted in subaerial exposure, karstic weathering and erosion of the Gambier Limestone. Apart from Quaternary volcanism and associated uplift, the Gambier Basin has been relatively stable since this time. Deposition of a shallow-water marine sand, equivalent to Loxton Sand (formerly Loxton Sands) of the Murray Basin, occurred during an Early Pliocene transgression.

Regional warping has been dominant over faulting since the Late Cretaceous, although seismic sections indicate that many faults displace the Gambier Limestone. Sprigg (1952b) described gentle folding in Gambier Limestone, which he related to differential movement of fault-bounded basement blocks, and noted zones of dolomitisation in the same unit which appear to be associated with faults. Major structures in the Cretaceous Otway Basin produced only minor warping or faulting in the Tertiary units. In Allen's Quarry, for example, a gentle anticline in Tertiary sediments is related to the nearby Tartwaup Fault (Fig. 10.5). Minor Tertiary movement has also occurred along the Kanawinka Fault, which controls the eastern margin of the Gambier Basin (Figs 10.3, 10.4).

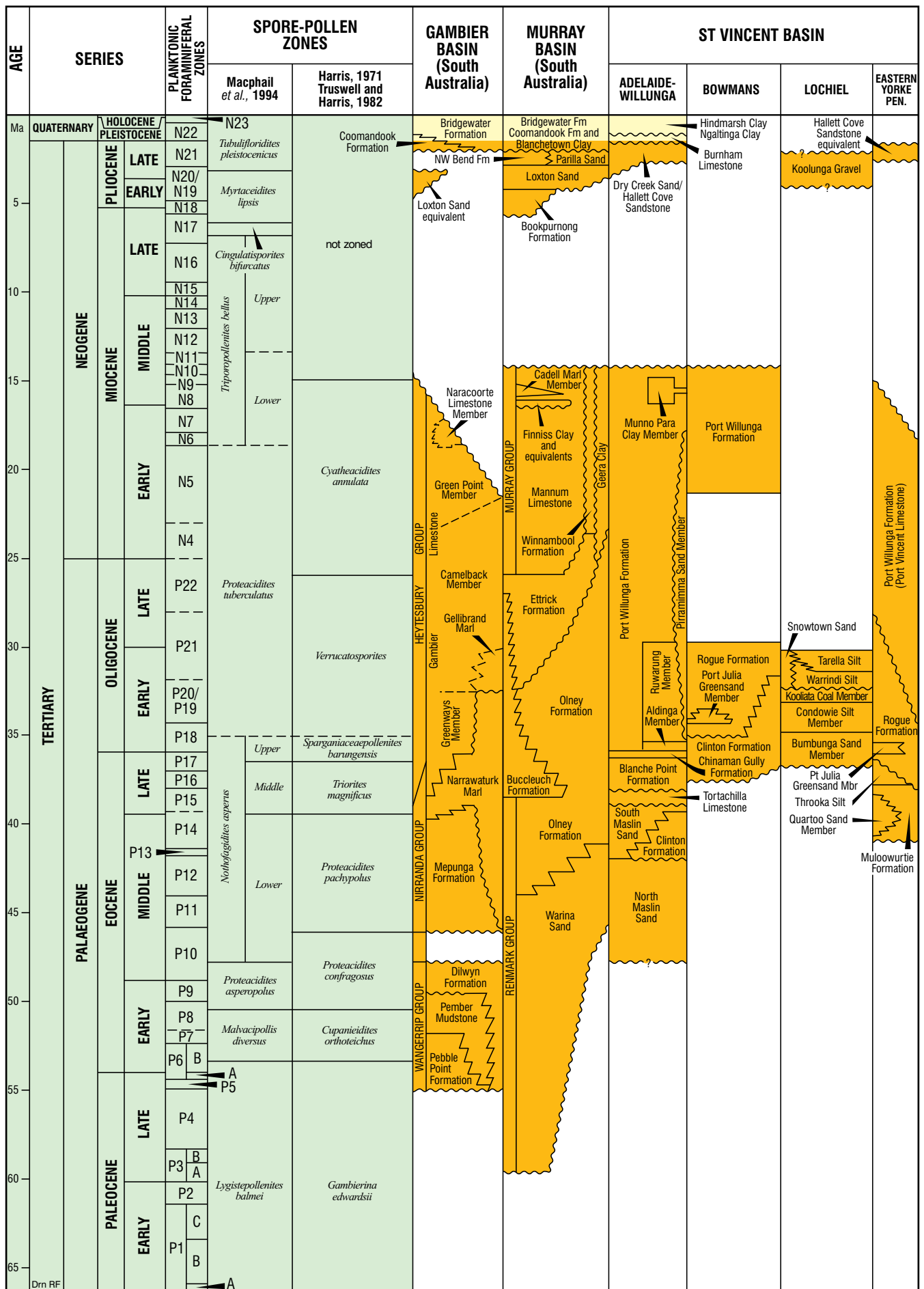
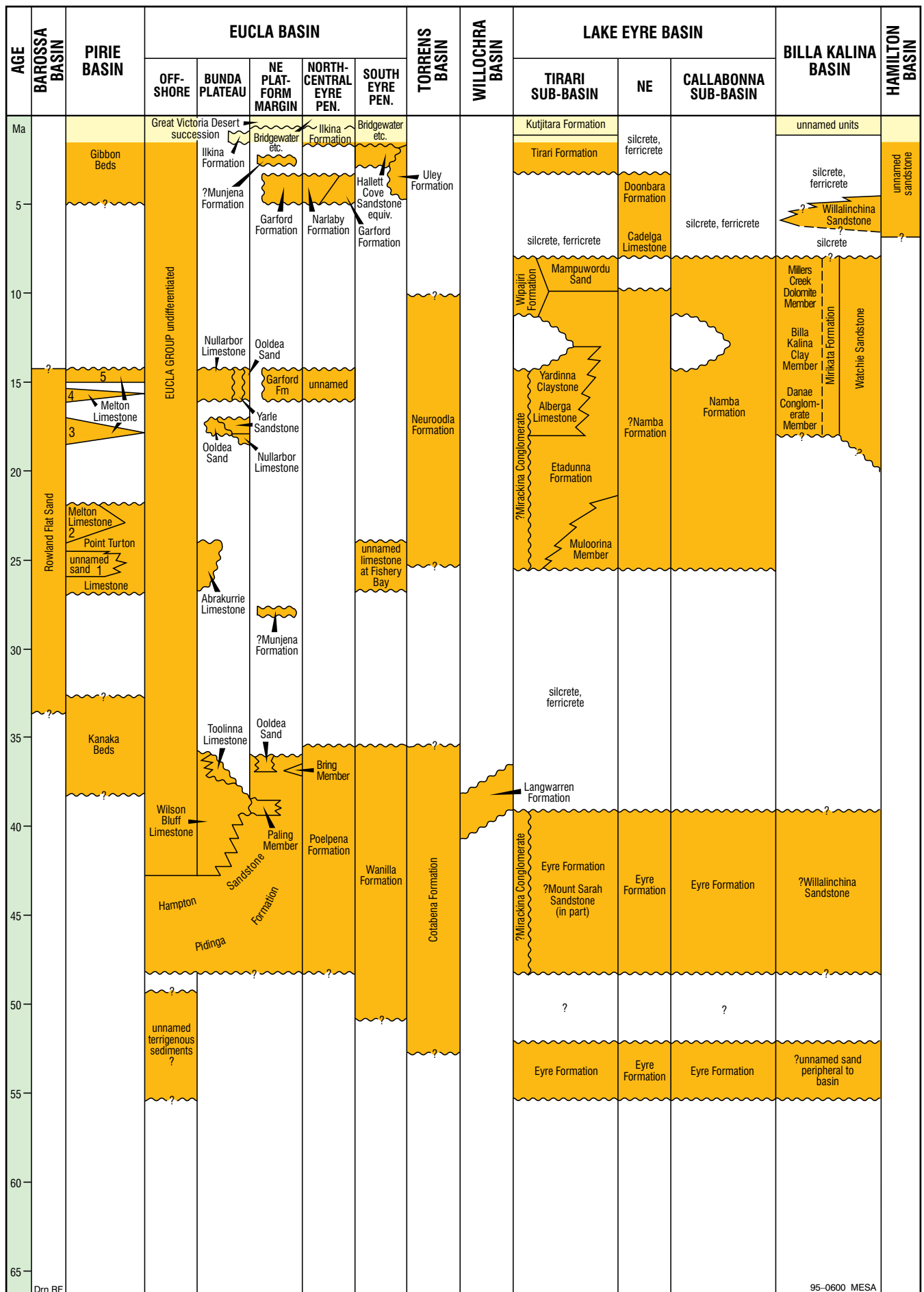
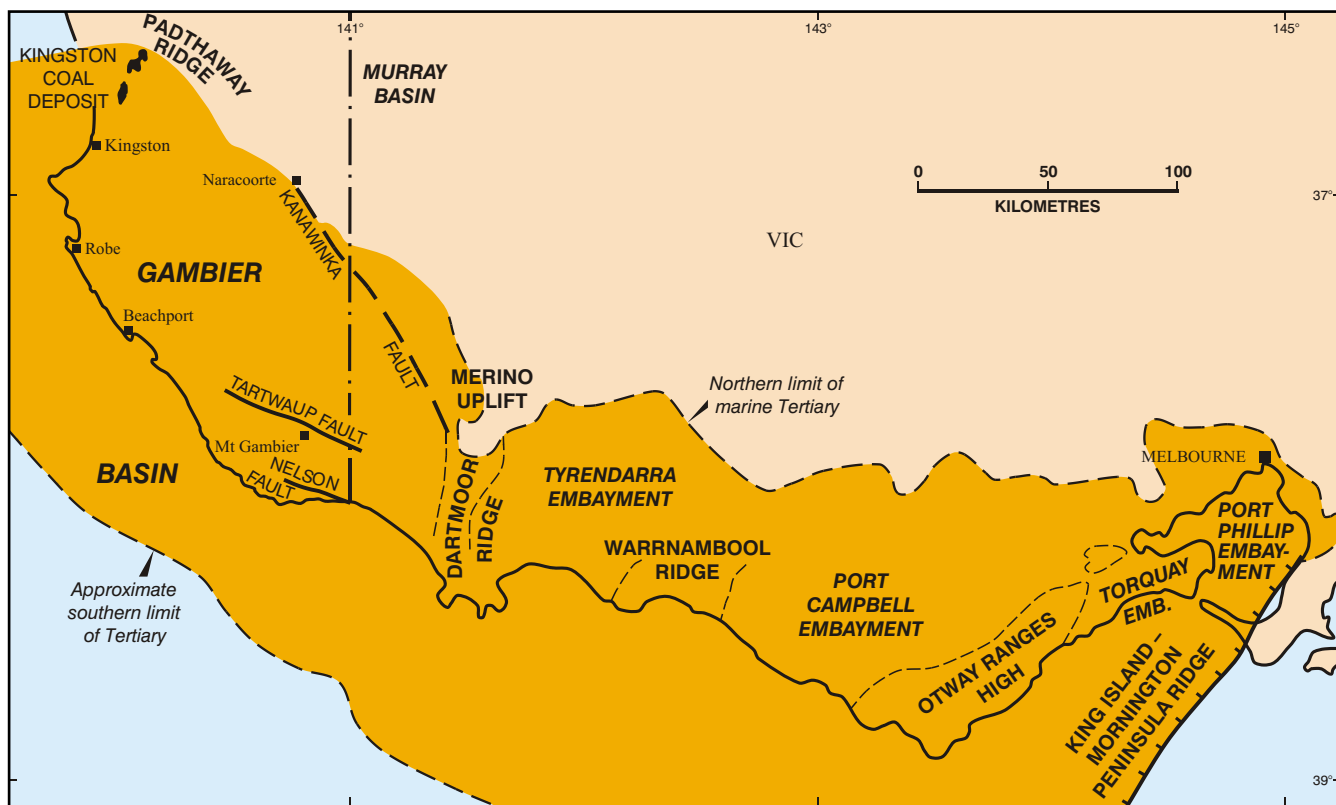


Fig. 10.2 Correlation chart for sediments in South Australian Tertiary basins.





Stratigraphy

Late Paleocene-Middle Eocene paralic sequence set (Wangerrip Group)

Pebble Point Formation

In South Australia, the Pebble Point Formation is discontinuous and known only from drillholes. In Mount Salt 1 it unconformably overlies the Late Cretaceous Sherbrook Group and consists of 40 m of dark green and brown, chamositic, oolitic grit with a restricted-marine *Cyclammina* microfauna. The presence of large, chamosite-rich ooids and poorly sorted quartz grains in a muddy matrix of ferruginous spherulites is regarded as evidence for deposition in a low-energy, interdistributary bay environment on a lower delta plain during an episode of reduced detrital sedimentation (Gravestock *et al.*, 1986).

In Victoria, the formation yields Late Paleocene planktonic foraminiferal faunas correlative with upper Zone P4 and P5 (McGowran, 1991). Palynofloras from South Australia indicate that the formation ranges in age from Late Paleocene to Early Eocene (*Lygistipollenites balmei* to *Malvacipollis diversus* Zones; e.g. White, 1995a; Fig. 10.2).

Pember Mudstone

The Pember Mudstone interfingers with lower Dilwyn Formation and is a homogeneous, well-laminated, carbonaceous and mica-ceous silty mudstone with thin, fine sand interbeds. The formation, which ranges in age from Late Paleocene to Early Eocene (*L. balmei* to early *Proteacidites asperopolus* Zones; e.g. White, 1995a), is interpreted as originally interdistributary bay muds.

Dilwyn Formation

The overlying Dilwyn Formation comprises two broad facies associations: a landward assemblage consisting of sandy sediments and a seaward succession of more mixed lithology (Fig. 10.4). These can be further subdivided into deltaic cycles. The sandy

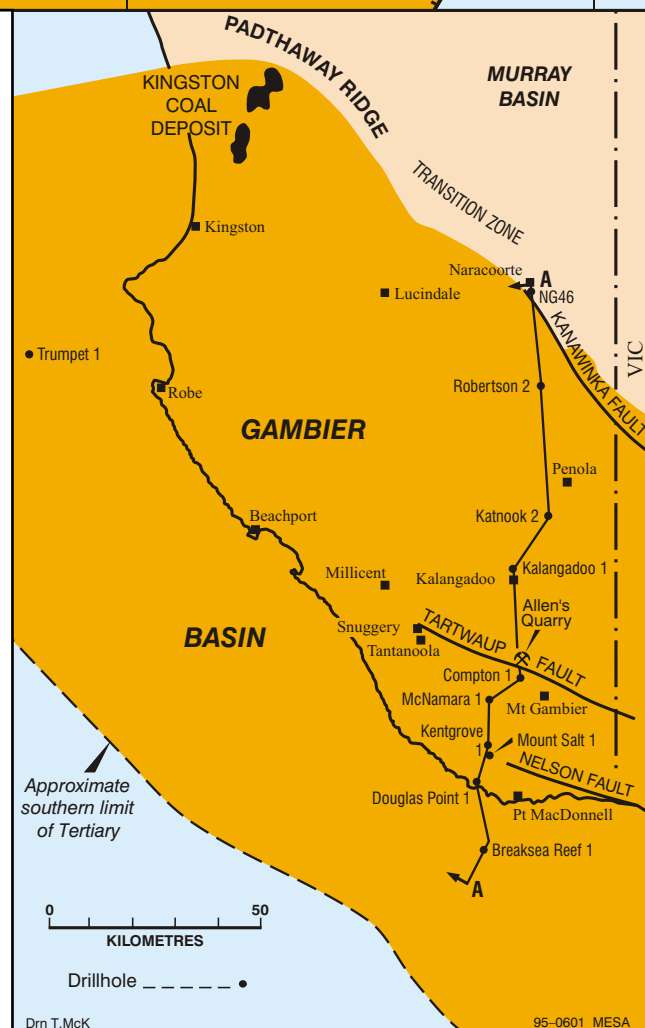


Fig. 10.3 Locality map of the Gambier Basin showing basin limits, related embayments, major structures, petroleum drillholes, and location of the geological section shown on Figure 10.4.

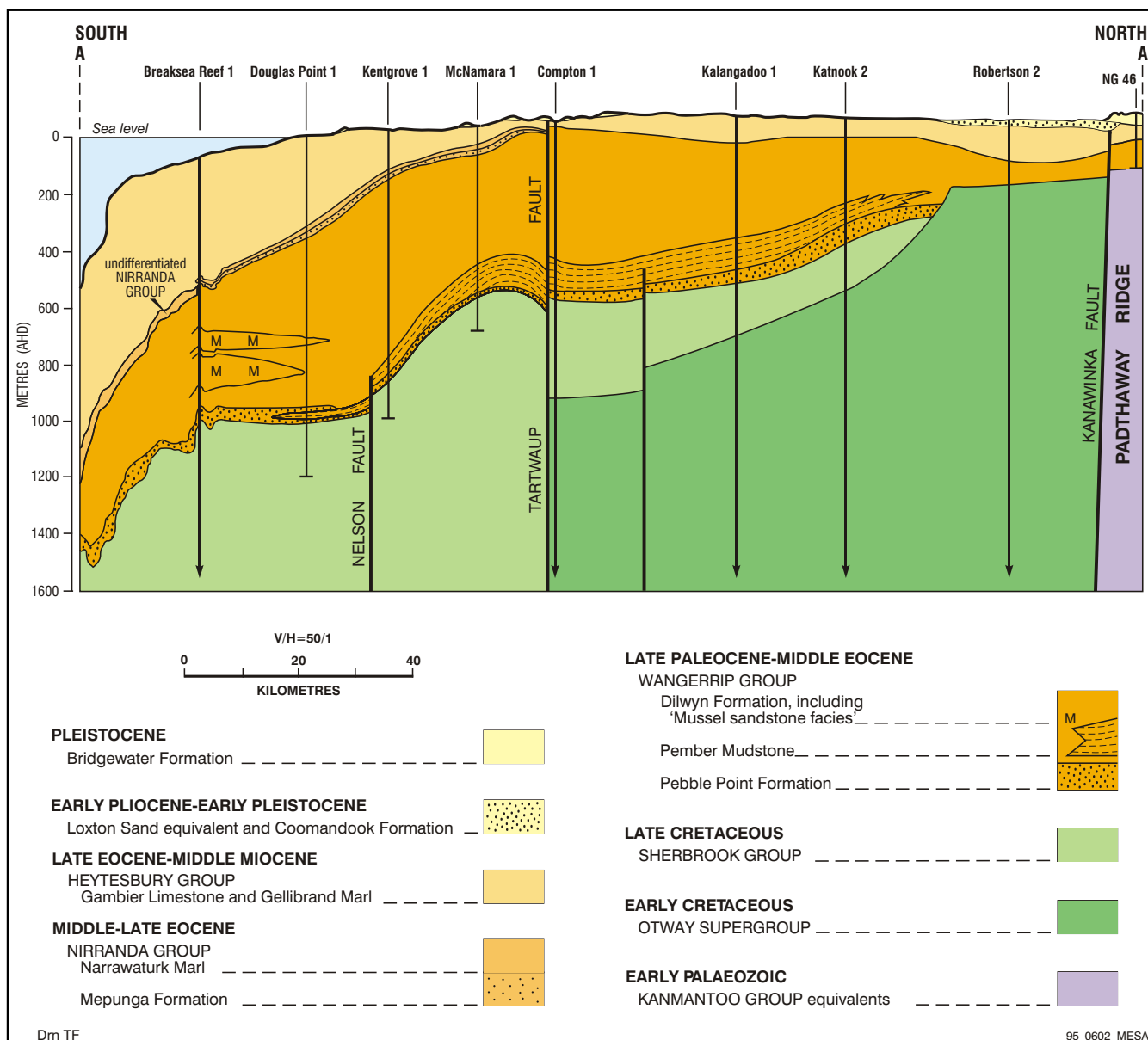


Fig. 10.4 North-south-cross section through the eastern Gambier Basin. Line of section is located on Figure 10.3.

facies of the formation comprises ferruginous and partly pyritic, silty, fine to coarse-grained sandstone, with carbonaceous and micaceous mudstone and siltstone interbeds, glauconitic pellets and sparse shelly fossils. Extensive coal seams in the Kingston area are provisionally included in this formation. The weathered and eroded top of the Dilwyn is exposed only in Allen's Quarry, on a structural high related to uplift along the Tartwaup Fault (Fig. 10.5). Elsewhere, the formation is known only from drillhole intersections where it ranges in thickness from 101 to 740 m.

The Dilwyn Formation is generally characterised by restricted-marine *Cyclammina* facies. Blake (1980) and Holdgate (1982) regarded the depositional environment of the landward sandy facies as fluvial and delta distributary channels. The seaward facies were deposited in prodelta and interdistributary bay environments, with sandy intervals ('Mussel sandstone facies') representing offshore barrier bars (Fig. 10.4).

In accord with the restricted marine environment of deposition, marine microfaunas in this formation are sparse. Palynofloras indicate an age range from Late Paleocene to earliest Middle Eocene (*L. balmei* to *P. asperopolus* Zones; e.g. White, 1995a). The Kingston coal deposit contains palynofloras believed to be correlative with the *P. asperopolus* Zone (Wood, 1981), but plant cuticle data indicate a Late Eocene age (Rowett and Sparrow, 1994).

Middle to Late Eocene marginal marine sequence set (Nirrandra Group)

The Nirrandra Group marks the onset of a major marine transgression and consists of a succession of ferruginous sand (Mepunga Formation) and glauconitic marl (Narrawaturk Marl).

Mepunga Formation

The Mepunga Formation is a sparsely fossiliferous, brownish, goethitic quartz grit, 9 to 12 m thick over much of the southern Gambier Basin. Blake (1980) interpreted the unit as a beach barrier sand. Foraminiferal faunas correlate approximately with Zone P14 (McGowran, 1978) of late Middle Eocene age and palynofloras are correlative with the Middle Eocene *Proteacidites pachypolus* Zone (Harris, 1971).

Narrawaturk Marl

The overlying Narrawaturk Marl comprises carbonaceous silt and sandy clay with abundant glauconite pellets, shell fragments and foraminifera, overlain by greenish brown shelly marl with glauconite and goethite pellets, mollusc fragments, foraminifera and other calcareous fossils. A middle shelf environment of deposition has been recorded for this formation in South Australia

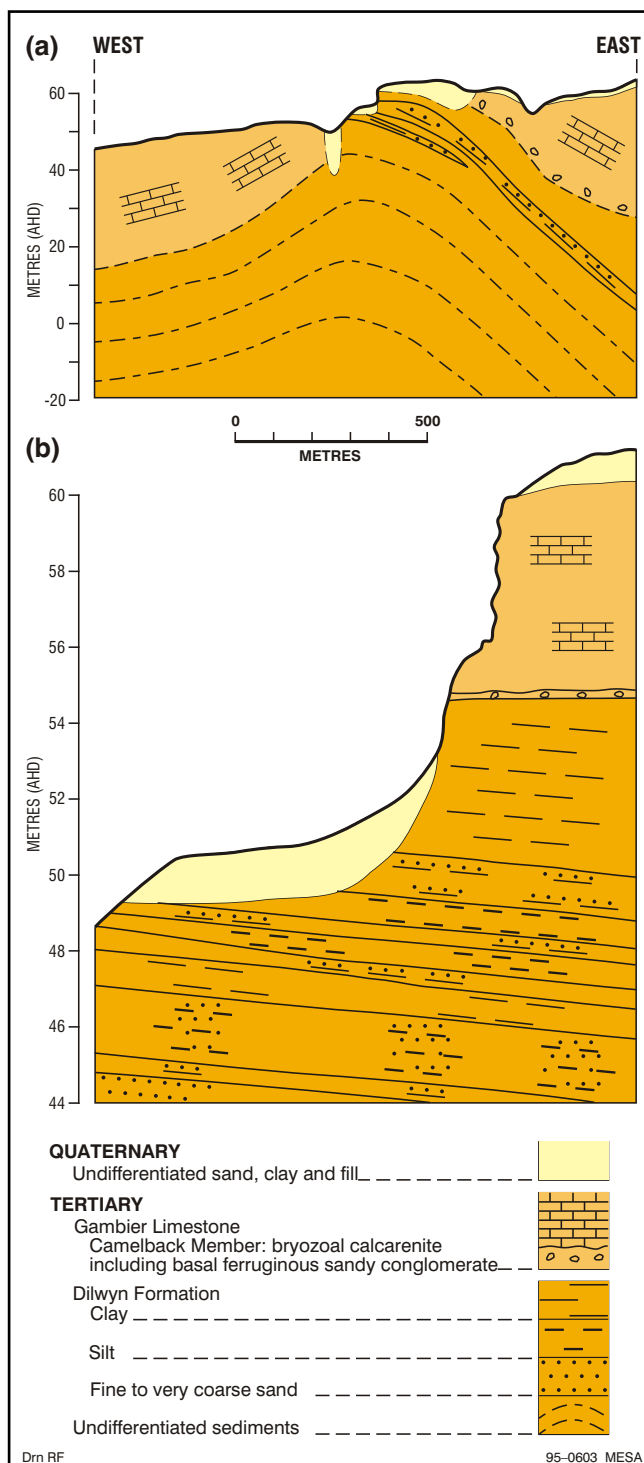


Fig. 10.5 Stratigraphy of Tertiary sediments in the Allen's Quarry area (after Keeling *et al.*, 1985): (a) generalised geology of the quarry area, (b) composite stratigraphy of the quarry.

(White, 1995a). This interpretation is consistent with the open marine, mostly below storm-wave base palaeoenvironment recorded in Victoria (Tickell *et al.*, 1992). Foraminiferal faunas from the marl are generally correlative with latest P14 to P17 Zones (e.g. White, 1995a), and palynofloras with the Middle and Upper *Nothofagidites asperus* Zones, indicating a Late Eocene age.

The formation is best developed in the northern and western Gambier Basin where it reaches 15 m in thickness and conformably overlies Mepunga Formation. The boundary with the Mepunga may be at least partly diachronous as suggested in Victoria (Tickell *et al.*, 1992). The Narrawaturk grades upwards into the lithologically similar basal member of the Gambier Limestone.

Late Eocene-Middle Miocene marine sequence set (Heytesbury Group)

The Nirranda Group is succeeded by the Heytesbury Group which in South Australia consists of the Gambier Limestone and Gellibrand Marl (Morton and Drexel, 1995). The Glenelg Group (Sprigg and Boutakoff, 1953; Ludbrook, 1957b, 1961c) is regarded as a junior synonym of the Heytesbury Group (Abele *et al.*, 1976, 1988).

Gambier Limestone

The Gambier Limestone is a major fossiliferous carbonate, varying in thickness up to 400 m offshore, deposited on an open-marine shelf. The Gambier Limestone in the southern part of the basin can be subdivided into three members, described formally by White (1995b), based on work by Ludbrook (1962, 1971), Lindsay (1967, 1985), McGowran (1973b) and White (1995a):

- The Greenways Member typically comprises grey limestone, marly limestone and marl, and is glauconitic at its base. Although this member is easily distinguished from the Narrawaturk Marl by its colour, its lower part is thought to grade laterally into the Narrawaturk. The Greenways Member includes the 'sponge spicule zone' noted by Ludbrook (1962) in Mount Salt SH 1 and 3-5, where the member is up to 70 m thick. It is of latest Eocene to Early Oligocene age (P15 to ~P20).
- The Camelback Member, 10-250 m thick, consists of bryozoal calcarenite with rare chert, and now includes the former Compton Conglomerate (White, 1995b). It is partly recrystallised and often saccharoidal with abundant carbonate rhombs. It ranges in age from Early Oligocene to Early Miocene (~P20 to ~N5). The member is quarried extensively for building stone.
- The Green Point Member, at least 193 m thick, is characterised by its grey colour and abundant chert, and is of Late Oligocene to earliest Middle Miocene age (~N4 to at least N9). It is commonly absent north of Mount Gambier due to erosion.

A thin basalt flow in lower Gambier Limestone offshore in Trumpet 1 is probably related to basalt near the top of the Dilwyn Formation in drillholes in western Victoria; it is dated there as latest Eocene (37 Ma, K-Ar; Abele *et al.*, 1988).

In the northern part of the basin, the upper Gambier Limestone contains the Naracoorte Limestone Member, a 6 m thick, coarse-grained, rubbly, highly fossiliferous, leached limestone with abundant casts and moulds of molluscs and echinoids. In the Naracoorte area, the member is Early Miocene, as is the underlying Gambier Limestone. Its rubbly appearance results from solution processes which commenced in the Middle to Late Miocene.

Dolomitised zones in the Gambier Limestone are associated with the Tartwaup and Nelson Faults, and a deposit near Tantanoola is quarried for high-grade dolomite used in the glass industry. Intervals of dolomite at the base of the Gambier Limestone could be a source of magnesium for the zones of secondary dolomitisation.

Deposition of the Gambier Limestone ended in the early Middle Miocene during a global eustatic fall in sea level (Haq *et al.*, 1987a) and uplift of the southern Australian continental margin. During the remainder of the Miocene, the formation was exposed to extensive karstic weathering and erosion which has continued intermittently to the present day (see Ch. 11).

Gellibrand Marl

In South Australia, the Gellibrand Marl has been identified only in Katnook 2, where it comprises 7 m of Early Oligocene (~P20; White, 1995a) brown, shelly marl, and immediately underlies the Camelback Member of the Gambier Limestone.

Late Tertiary units

Marine sand containing an Early Pliocene foraminiferal microfauna, consisting of *Fabularia howchini*, *Crespinella umbonifera* and *Globorotalia puncticulata*, rests unconformably on Gambier Limestone west of Naracoorte and the Kanawinka Fault (Fig. 10.4; Cook *et al.*, 1977). This sand is correlated with Loxton Sand of the Murray Basin.

Tertiary units of the coastal plain are overlain in places by fossiliferous sand, sandy limestone and calcarenite with oyster beds. These sediments are correlative with the Coomandook Formation of the Murray Basin and Werrikoo Limestone of western Victoria. Both of these units are predominantly Early Pleistocene in age, but the Werrikoo extends into the latest Pliocene (Singleton *et al.*, 1976).

Economic geology

Coal

Although carbonaceous and lignitic intervals in the Dilwyn Formation have been intersected in many drillholes, the most significant is the Kingston coal deposit (Fig. 10.3). This resource comprises 985 Mt of lignite in a northeast-trending depression onlapping the Padthaway Ridge near the northern margin of the basin (Meyer, 1982). Most of the coal is in a single seam up to 12 m thick, with overburden 28–75 m thick. It is low rank Lignite B (ASTM), similar in quality to other Tertiary coals in South Australia.

Dimension stone

Building blocks (ashlars) have been produced from Gambier Limestone since 1844. The quarries are located in distinctive pale cream or white porous bryozoal calcarenite of the Camelback Member which crops out at Marte, 10 km west of Mount Gambier. The quarries, with current output of ~11 000 t/year, are among the largest dimension stone producers in Australia.

Groundwater

The Dilwyn Formation contains the regional confined aquifer which has been developed for pasture irrigation in the Lucindale-Beachport-Kingston area where artesian conditions exist. Elsewhere, this aquifer is used if the overlying unconfined aquifers are unproductive or contaminated. The Mepunga Formation also forms a confined aquifer which has been exploited in the Snuggery area and for the Kingston town water supply, although the high iron content is a limiting factor. Gambier Limestone is an important high-yielding aquifer, particularly in the Mount Gambier area. Groundwater of generally low salinity is used for industrial, domestic, agricultural and recreational purposes. Its susceptibility to pollution, however, causes problems for both local and regional groundwater management.



Gambier Limestone with solution pipes at Steetley building stone quarry near Mount Gambier. Photo 042492.

MURRAY BASIN

P.A. Rogers, J.M. Lindsay, N.F. Alley,
S.R. Barnett, K.L. Lablack and G. Kwitko

Structure and setting

The epicratonic Murray Basin, of Paleocene to Quaternary age, covers an area of 300 000 km² in southeastern South Australia, southwestern New South Wales and northwestern Victoria (Fig. 10.6). The basin succession is up to 600 m thick and comprises freshwater, marine, coastal and continental sediments, with the thickest part occurring between Renmark and northwestern Victoria.

In South Australia, the western and northern basin margins are marked by uplifted Neoproterozoic (Adelaidean) and Cambrian metasediments, and Cambro-Ordovician granites. To the

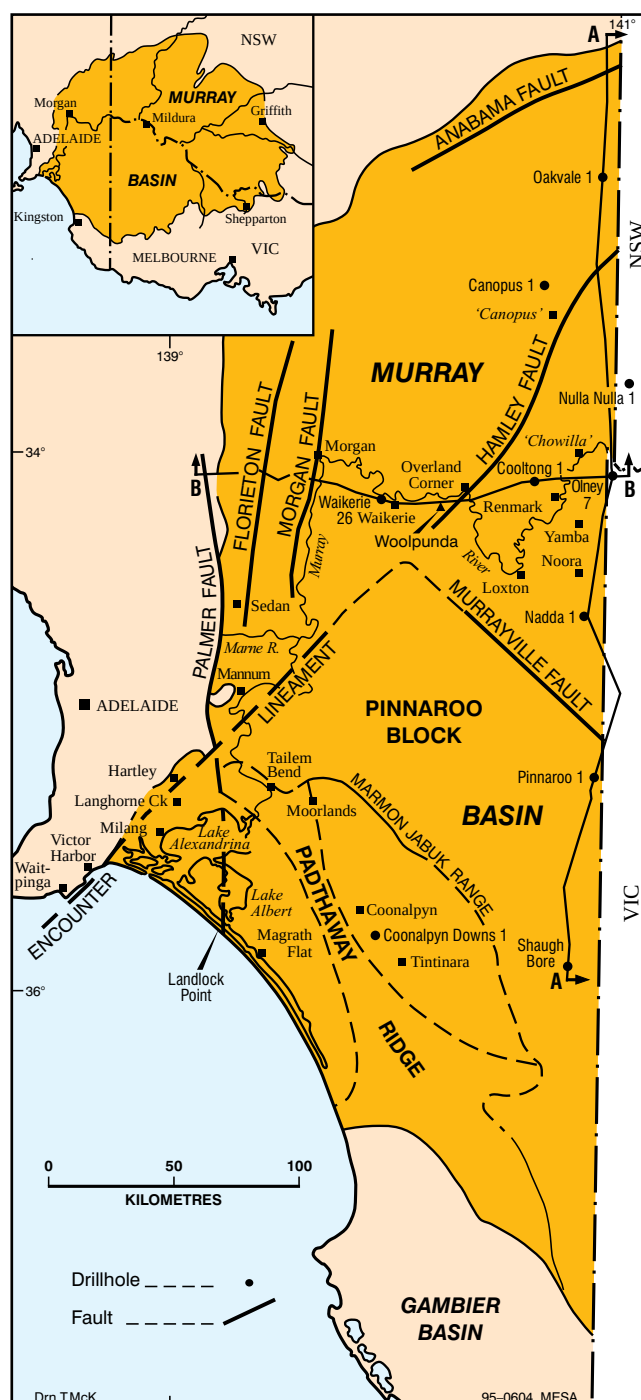


Fig. 10.6 Locality map of the Murray Basin in South Australia showing basin limits and major structures. Sections are shown on Figure 10.8.

southwest, the basin is partly separated from the thicker succession in the Gambier Basin by the Padthaway Ridge, formed of shallowly buried early Palaeozoic basement rocks. The Tertiary succession in South Australia rests unconformably on Neoproterozoic to early Palaeozoic basement, late Palaeozoic glauconitic sediments of the Troubridge and Nadda Basins, and Early Cretaceous Monash Formation of the Berri Basin.

Stratigraphy

The Tertiary succession can be divided into three sequence sets:

- Late Paleocene to Early Oligocene non-marine and marginal marine sediments — includes the Renmark Group
- Late Eocene to Middle Miocene transgressive marine sediments — includes the Murray Group
- latest Miocene to Late Pliocene marine, coastal and non-marine sediments (Figs 10.2, 10.7).

Late Paleocene-Early Oligocene non-marine sequence set (Renmark Group)

Warina Sand

The oldest sediments are the Renmark Group, comprising the Warina Sand and Olney Formation, together exceeding 330 m in thickness near Renmark (Fig. 10.8). Warina Sand is restricted to the deeper parts of the basin and consists of medium to coarse-grained quartz sand with minor thin, lenticular interbeds of locally carbonaceous silt and clay. The sandy facies is interpreted as a fluvial, braided-channel deposit and the finer grained interbeds as lacustrine and flood-plain deposits. Absence of marine fossils and an abundance of freshwater dinoflagellates indicate non-marine conditions. The oldest date for the Warina Sand is Late Paleocene, based on correlation of palynofloras from the base of the unit with the *L. balmei* Zone (Harris, 1970b; Lindsay and Bonnett, 1973).

Olney Formation

The overlying Olney Formation is more widespread than the Warina Sand, and extends into marginal areas of the basin where it rests on pre-Tertiary basement (Fig. 10.8). Thinly bedded carbonaceous sand, silt, clay and lignite (including Moorlands Lignite Member, and lignite in the Sedan and Anna coal deposits) were deposited in fluvial, lacustrine and swamp environments (Fig. 10.9a). The upper part of the Olney Formation is often marginal marine in the western Murray Basin.

Although the Olney Formation ranges in age from Middle Eocene to Early Oligocene in the western part of the basin, deposition of lithologically identical sediments continued, probably with interruptions, into Middle Miocene time further east (Figs 10.2, 10.7).

Late Eocene-Middle Miocene marine sequence set (includes Murray Group)

This marine sequence set comprises the Buccleuch Formation, Ettrick Formation and overlying Murray Group. The Murray Group consists of predominantly shallow marine fossiliferous limestone and sandstone with minor clay and silt, well exposed in cliffs of the Murray River between Overland Corner and Tailem Bend. In the past, subdivision of the group was based mainly on age differences. The limestone units (Mannum Formation, Morgan Limestone and Pata Limestone) are difficult to distinguish lithologically and probably form part of a continuous succession; they are here interpreted in the same fashion as Brown and Stephenson (1991). All of the limestones are therefore incorporated into a single unit (Mannum Limestone), with the type section as described by Ludbrook (1961c) at Mannum pumping station and reference sections in the river cliffs 5 km south-southeast of Morgan and in drainage shaft no. 18 near Loxton.

Buccleuch Formation

Deposition of the Buccleuch Formation (formerly Buccleuch Beds, Buccleuch Group) began in the Late Eocene with a marine transgression into the southwestern Murray Basin (Figs 10.7, 10.9a). These marine sediments grade laterally into the marginal marine upper Olney Formation. Ludbrook (1961c) divided the Buccleuch in the Coonapyn Town Bore into three informal units (from the base):

- unit A, bryozoal limestone and glauconitic limestone with basal glauconitic marl
- unit B, bryozoal clayey sand and black pyritic clay
- unit C, a thin layer of black ferruginous clay.

This series of units can be recognised only along a narrow strip of the northeastern Padthaway Ridge, with the dark brown to black clay and bryozoal sand facies distributed over a wider area. Buccleuch Formation black clay is distinguished from a similar facies in the Olney Formation by the lack of quartz sand and mica, and the presence of shells and bryozoal calcarenite. Palaeontological evidence from the Buccleuch along the margins of the Padthaway Ridge indicates that it ranges in age from Late Eocene to possibly early Late Oligocene (Beecroft *et al.*, in prep.).

Limestone intercalated with marl, sand and clay in the Milang and Langhorne Creek area are regarded as Buccleuch Formation (Lablack, 1991). In the Waitpinga drainage basin near Victor Harbor, a thin bed of fossiliferous and ferruginous sandy limestone at 70 m AHD contains Late Eocene foraminiferal faunas correlative with limestone in the Buccleuch (Bourman and Lindsay, 1973) and may represent a southwestern extension of the Murray Basin. Pale green limestone cropping out at the base of the Murray River cliffs south of Tailem Bend is regarded as the Early Oligocene part of the Buccleuch Formation (Lablack, 1991).

Transgressions in the Early and Late Oligocene led to the deposition of widespread thin conglomerate, ferruginised sandstone, sandy limestone and sandy clay referred to as Compton Conglomerate equivalents (Lindsay and Williams, 1977). An exposure at Cross's Bridge near Hartley, and fossiliferous sand at depth in the Waikerie area (Lindsay and Barnett, 1989; Brown and Stephenson, 1991), contain the foraminifera *Subbotina angiporoides* and *Chiloguembelina cubensis*, indicating an Early Oligocene age. A coarse conglomerate resting on early Palaeozoic basement on the southwestern margin of the basin contains foraminifera of Late Oligocene age (Lindsay and Williams, 1977). However, these transgressive sediments in the Murray Basin are here regarded as littoral facies in the basal part of the overlying units. The different ages may reflect either two transgressive phases in the Oligocene separated by a stillstand, or different stages of the same transgression.

Ettrick Formation

The Ettrick Formation, comprised predominantly of glauconitic and fossiliferous marl, calcareous clay and mudstone, with silt and fine quartz sand, represents a Late Oligocene transgressive phase found extending northwards from the margin of the Padthaway Ridge.

Mannum Limestone

The Mannum Limestone ranges in age from Early to Middle Miocene. In the lower part it consists of echinoidal and bryozoal calcareous sandstone and sandy limestone. A thin bed of crinoidal limestone overlying granite near Mannum (Ludbrook, 1961c) contains a foraminiferal microfauna including *Bolivinopsis cubensis*, thus indicating that the age here may extend back to the latest Oligocene (Lindsay, 1982). The middle part of the formation near Morgan comprises an upper and lower bryozoal limestone, separated by the Cadell Marl Member (see below). A thin zone in limestone of this area contains foraminiferal faunas correlative with the *Lepidocyclina* horizon in the early Middle Miocene *Praeorbulina glomerosa* Zone (Lindsay and Giles, 1973). The youngest part of the Mannum Limestone occurring in the Loxton

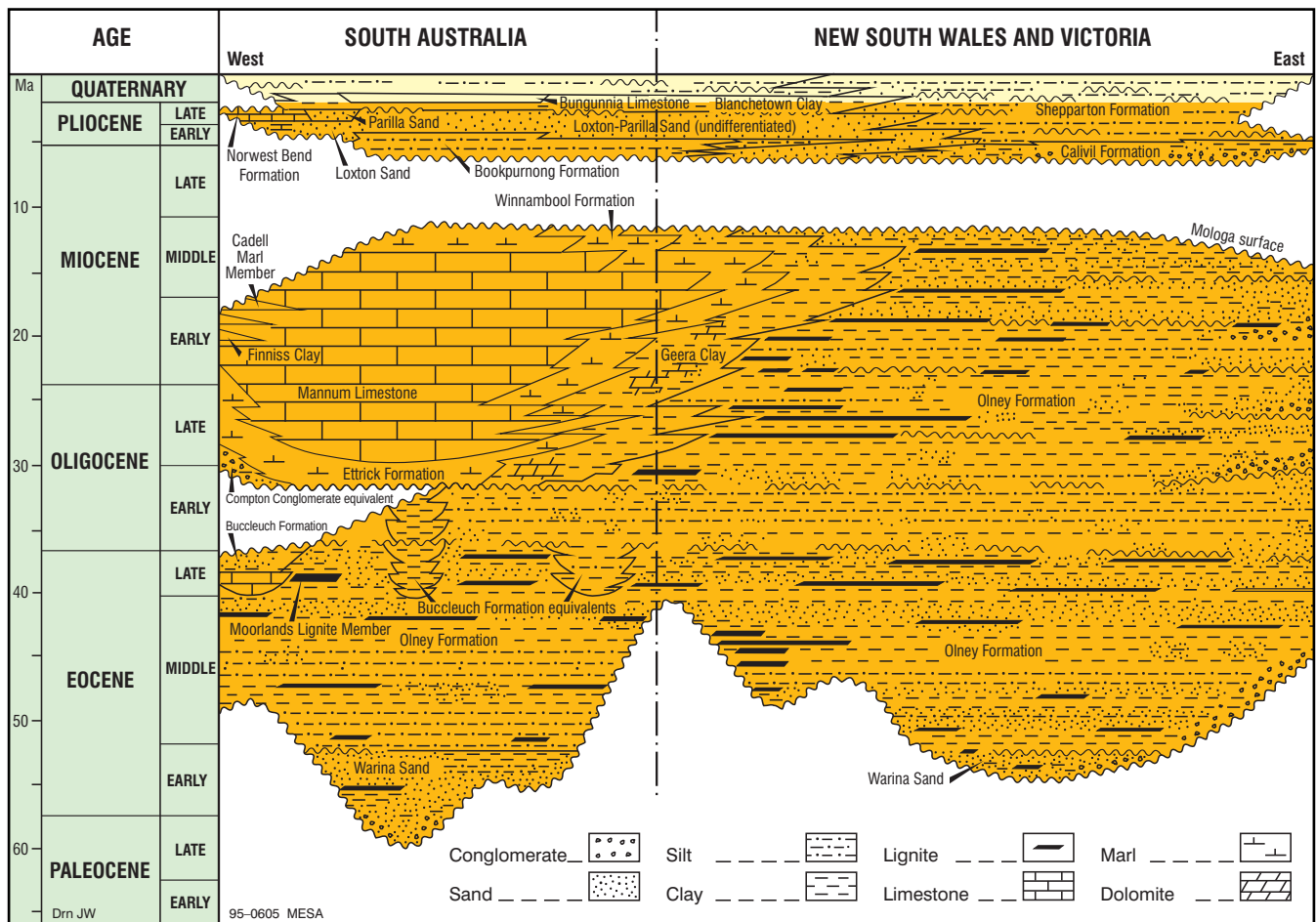


Fig. 10.7 Stratigraphic units in the Murray Basin (after Brown and Stephenson, 1991).

area, exposed only at Sunlands Pumping Station and near Ross Lagoon, comprises bryozoal limestone and marl with abundant tubes of the annelid *Ditrupa*. The limestone here contains the benthonic foraminifer *Heterolepa victoriensis*, indicating a Middle Miocene age (Ludbrook, 1961c; Carter, 1964).

The Cadell Marl Member occurs in the subsurface north of Morgan, and in cliff sections and drillholes along the Murray River valley (Lindsay and Barnett, 1989). The unit contains a rich molluscan fauna with tropical affinities of early Middle Miocene age (Lindsay and Giles, 1973; Ludbrook, 1973a).



Succession from Cadell Marl Member (base) to upper Mannum Limestone then Norwest Bend Formation, near Morgan. This is the type section of the former Morgan Limestone. Photo 042493.

Finniss Clay

Several lithologically distinct lenses occur within the Murray Group, and at some localities are intercalated with Mannum Limestone. In some areas the lower part of the formation passes up sharply into the late Early Miocene Finnis Clay, a thin regressive marine clay lens which in turn is overlain conformably and gradationally by the middle part of the Mannum Limestone. The Finnis is patchily distributed along the southwestern margin of the basin, although fine-grained fossiliferous correlatives can be recognised in subsurface and outcrop throughout the Waikerie area and eastwards as far as Woolpunda and Overland Corner (Lindsay and Bonnett, 1973; Lindsay and Giles, 1973).

Winnambool Formation and Geera Clay

The shallow marine carbonates of the Murray Group grade northwards to the progressively more marginal Winnambool Formation and Geera Clay (Figs 10.7-10.9b). Winnambool Formation, comprising fossiliferous marl, glauconitic marly limestone and marly clay, is a shallow-water restricted-marine and lagoonal facies. Apart from being generally younger (Early to Middle Miocene), it is difficult to distinguish from the Ettrick Formation.

Late Oligocene to Middle Miocene marginal marine and tidal sediments (Geera Clay) in the northwestern Murray Basin are predominantly black to grey-green, carbonaceous, pyritic clay with minor glauconite and sparse marine fossils. Palynofloras from the unit contain dinoflagellates, freshwater algae, spores and pollen correlated largely with the Upper *Proteacidites tuberculatus* Zone (Truswell *et al.*, 1985).

At the northern margin of the Murray Basin, non-marine carbonaceous and pyritic sand, silt and clay with lignite overlie Geera Clay and Winnambool Formation (Fig. 10.8). The non-marine

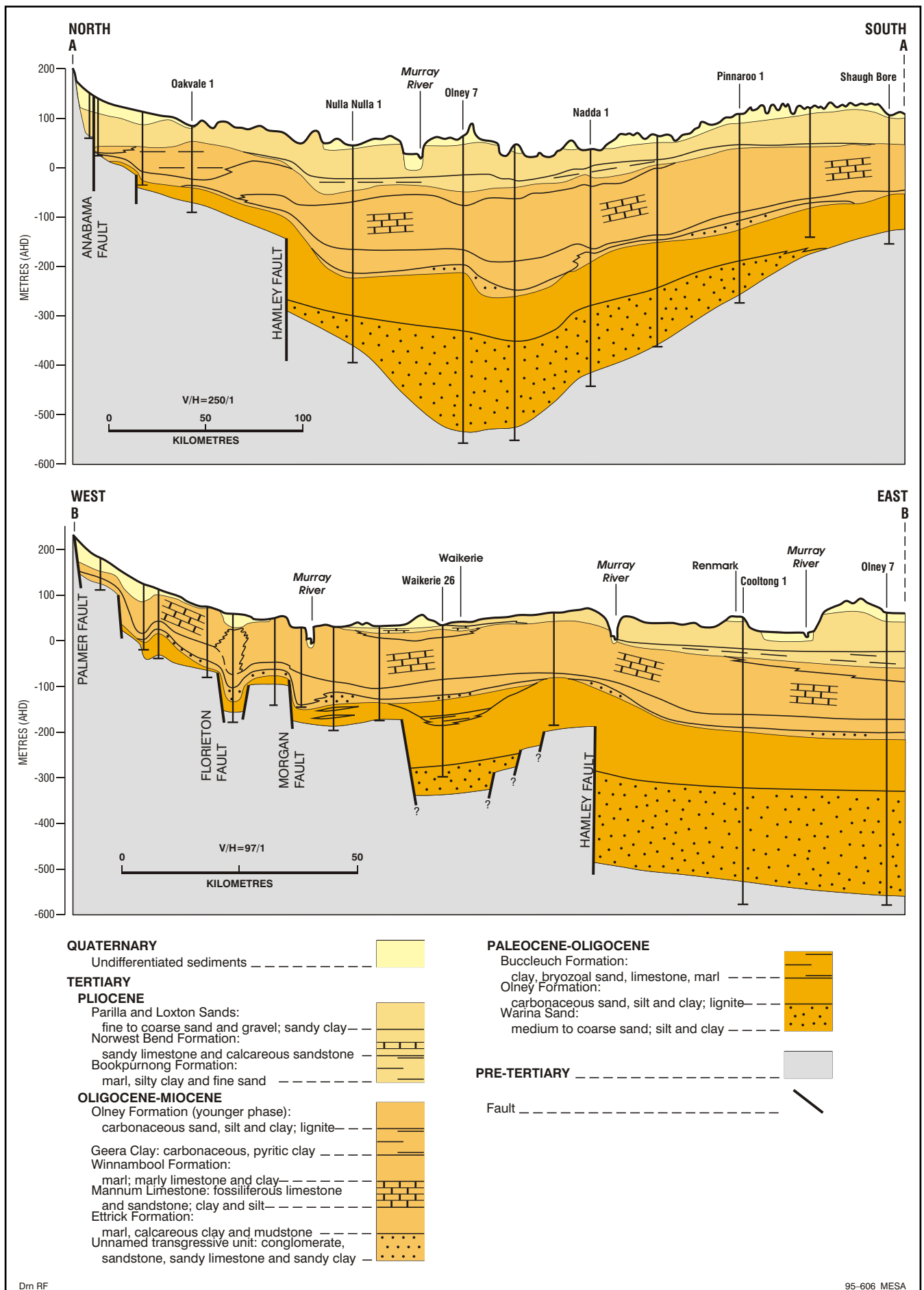


Fig. 10.8 North-south (A-A) and east-west (B-B) cross sections through the Murray Basin (after Barnett, 1989). Lines of section are located on Figure 10.6.

unit may be correlative with the Late Oligocene-Middle Miocene component of the Olney Formation, although alternatively it could be a carbonaceous facies of the Early Pliocene Loxton Sand.

Latest Miocene-Late Pliocene sequence set

The unconformity at the top of the Murray Group and its equivalents resulted from both a marked fall in sea level and from mild tectonic activity and erosion in Middle to Late Miocene time. In the eastern part of the Murray Basin, exposure led to development of the Mologa weathering surface during the Late Miocene (Macphail *et al.*, 1993). Reactivation of basement structures caused gentle but significant warping in the Murray Group and erosion of Mannum Limestone from areas west of Waikerie (Lindsay and Giles, 1973; Lindsay and Barnett, 1989). Upfaulted remnants of Miocene marine limestone indicate post-Miocene uplift of the eastern Mount Lofty Ranges in the order of 60 to >100 m, with some of this movement occurring during the Pleistocene (Lindsay, 1986; Bourman and Lindsay, 1989).

Bookpurnong Formation

Deposition of the third major sedimentary association in the Murray Basin commenced in the latest Miocene and continued through to the Late Pliocene. The lower part of this sequence is the Bookpurnong Formation, a shallow-water marine deposit comprising marl, silty clay and minor fine sand, all variably shelly, glauconitic and micaceous (Ludbrook, 1961c; Carter, 1985). Although widely distributed in the basin east of Loxton, it occurs largely in the subsurface with only limited outcrop near the town. The age of the formation was considered Late Miocene or Early Pliocene, but it is now regarded as straddling the Miocene-Pliocene boundary (Ludbrook, 1961c, 1969b, 1980; Lawrence and Abele, 1988; Macphail and Truswell, 1993). The molluscan faunas have a warm-water aspect, and include the pearl oyster *Pinctada crassicaudia* (Ludbrook, 1973a).

Loxton Sand

The Bookpurnong Formation is conformably overlain by the Early Pliocene Loxton Sand (formerly Loxton Sands). Around the western margin of the basin, the basal part of the unit may be equivalent to the Bookpurnong Formation (Brown, 1985). Loxton Sand comprises glauconitic, micaceous and shelly fine sand, overlain by planar and cross-bedded fine to coarse-grained sand and fine gravel, and planar-bedded, calcareous, micaceous, medium to coarse-grained sandstone with abundant shell debris. It is interpreted as an upward-coarsening regressive sequence of shallow-water marine and marginal marine sediments passing up into beach and coastal barrier deposits (Fig. 10.9c).

The upper part of the Loxton Sand forms an extensive regressive strand plain composed of gently arcuate beach ridges. The latter extend from their southernmost limit, the Marmon Jabuk Range, northwards to the vicinity of 'Canopus', but have been eroded and reworked into Late Pliocene to Early Pleistocene fluvial and lacustrine deposits in the vicinity of the Murray River. The beach sand is a potential source of heavy minerals, including ilmenite, zircon, rutile and tourmaline (Belperio and Bluck, 1990).

The Early Pliocene (Kaiman) foraminifer *Elphidium pseudonodosum* is often present in the Loxton Sand. Fossiliferous calcareous sandstone and sandy limestone bearing Early Pliocene microfaunas occur at several localities on southern Kangaroo Island (Milnes *et al.*, 1983). These are correlated with basal Loxton Sand and are regarded as the southwestern extremity of the Murray Basin.

Parilla Sand

Loxton Sand is overlain by Parilla Sand, composed of unfossiliferous, non-marine, fine to medium-grained clayey quartz sand with thin beds of sandy clay. The formation was derived mainly from Loxton Sand by aeolian and fluvial reworking, and may include large transgressive dune complexes (Belperio and Bluck, 1990),

as well as lacustrine and fluvial deposits laid down in depressions between stranded coastal ridges. A Late Pliocene age is suggested by super position on Loxton Sand, lateral gradation into Norwest Bend Formation, and palaeomagnetic dating of a widespread horizon of weathering and silicification (the Karoonda Surface; Firman, 1973), developed at the top of the Parilla Sand (Bowler, 1980; An *et al.*, 1986).

Norwest Bend Formation

Fossiliferous sandy limestone, calcareous sandstone and oyster beds comprising the Norwest Bend Formation were deposited in a narrow estuarine zone extending northwards from Tailm Bend along the western margin of the Murray Basin, with an arm extending eastwards as far as Overland Corner (Fig. 10.9d; Ludbrook, 1957b, 1973a).

Pliocene-Pleistocene units

Deposition of older Quaternary units (Blanchetown Clay and Chowilla Sand) may have commenced during the Late Pliocene, but these units are discussed in Chapter 11.

Economic geology

Coal

Low rank coal (lignite) in the Olney Formation has been reported from many localities, particularly along the western margin of the Murray Basin. The main coal deposits occur in the Sedan area (Sedan and Anna deposits) and near Moorlands. The Sedan deposit, which contains 200 Mt of coal, is the largest and has been investigated as a potential fuel source for electricity generation. Two coal seams have been identified: a continuous upper seam up to 8 m thick and a lenticular, discontinuous lower seam, averaging 5 m in thickness. Overburden ranges from 32 to 55 m in thickness. The coal is classified as Lignite B (ASTM), with high moisture, low heating value, moderate ash, and relatively high sodium, chlorine and sulphur contents. The other deposits are of similar quality, but are considerably smaller, deeper, and contain coal reserves of <50 Mt (Johns, 1975b).

Groundwater

Extensive investigations into the hydrogeology of the Murray Basin have greatly increased knowledge of the three main aquifer systems: the Renmark Group confined aquifer, the Mannum Limestone aquifer and the Pliocene sands aquifer (Barnett, in prep.).

The Renmark Group sand aquifer occurs at the base of the Tertiary succession over most of the basin and is separated from the overlying Mannum Limestone aquifer by glauconitic marl of the Ettrick Formation confining layer. The aquifer has been developed for stock and domestic supplies only around the basin margin, where it is relatively shallow and contains groundwater that is usually fresher than the sometimes saline water in the overlying limestone aquifer.

The Mannum Limestone is the most widely used aquifer. It is generally unconfined and has been developed for stock, domestic, irrigation and town water supplies because it contains low-salinity groundwater over large areas. Recharge occurs from higher rainfall areas around the basin margin, such as the Mount Lofty Ranges to the west and Dundas Plateau in southwestern Victoria. Ground water flows under low hydraulic gradients to the Murray River which acts as a drain for all three aquifer systems. Because salinity increases downgradient, from <1 000 mg/L in the recharge areas to >20 000 mg/L adjacent to the river, groundwater discharge to the river has a significant impact on its salinity, which increases markedly downstream.

To the northeast, the Mannum Limestone aquifer is downwarped over the Murrayville Monocline and is overlain and confined by the Bookpurnong Formation. These units are blanketed by the Loxton and Parilla Sands which form the Pliocene sand aquifer in

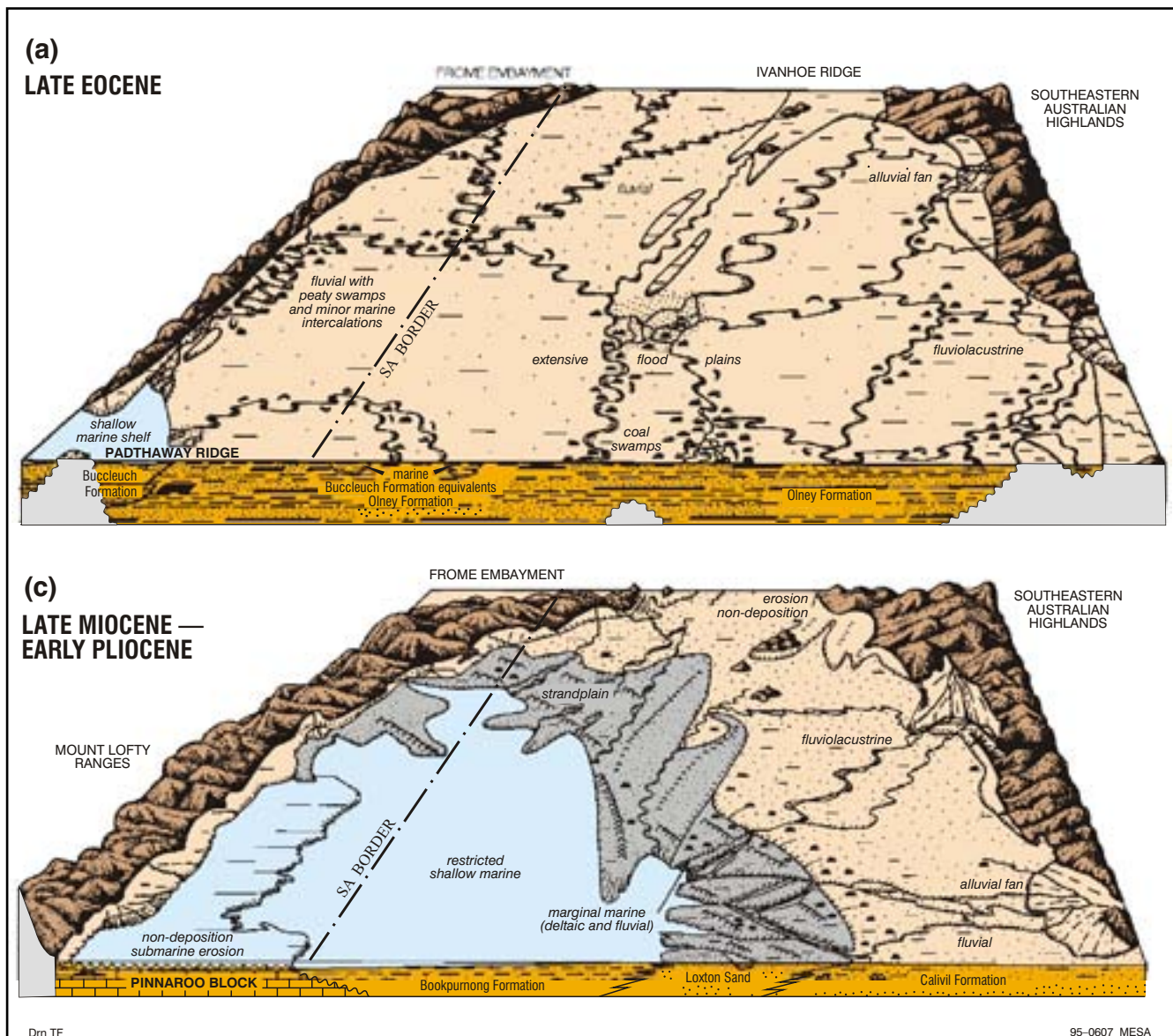


Fig. 10.9 Palaeofacies and palaeoenvironments during selected intervals of the Tertiary in the Murray Basin (after Brown and Stephenson, 1991).

this area. This aquifer contains highly saline groundwater which flows slowly towards the Murray River.

Basement structure, which is reflected in the modern surface topography by resurgent tectonics, exerts a strong influence on groundwater flow patterns and discharge mechanisms. Evaporative discharge occurs in the low-lying Noora-Yamba area which lies above the Renmark Trough, the deepest part of the Murray Basin, as the watertable is only 1 m below the ground surface. Basement faulting has caused thinning of confined aquifers which promotes discharge by upward leakage, as in the Woolpunda area (Lindsay and Barnett, 1989), and can also act as a barrier to groundwater flow in watertable aquifers.

Widespread clearing of native vegetation has led to a substantial increase of two orders of magnitude in annual recharge to the aquifers. The resultant rising watertable will cause two major problems. The more immediate is dryland salinisation on the low-lying coastal plain around Lakes Alexandrina and Albert, and also to the west of Tintinara. Here, the watertable is rising an average of 0.1 m/year and, if current trends continue, all low-lying land below 20 m AHD will be at risk to salinisation within 50 years. It is estimated that ~40 000 ha are affected, with an increase of

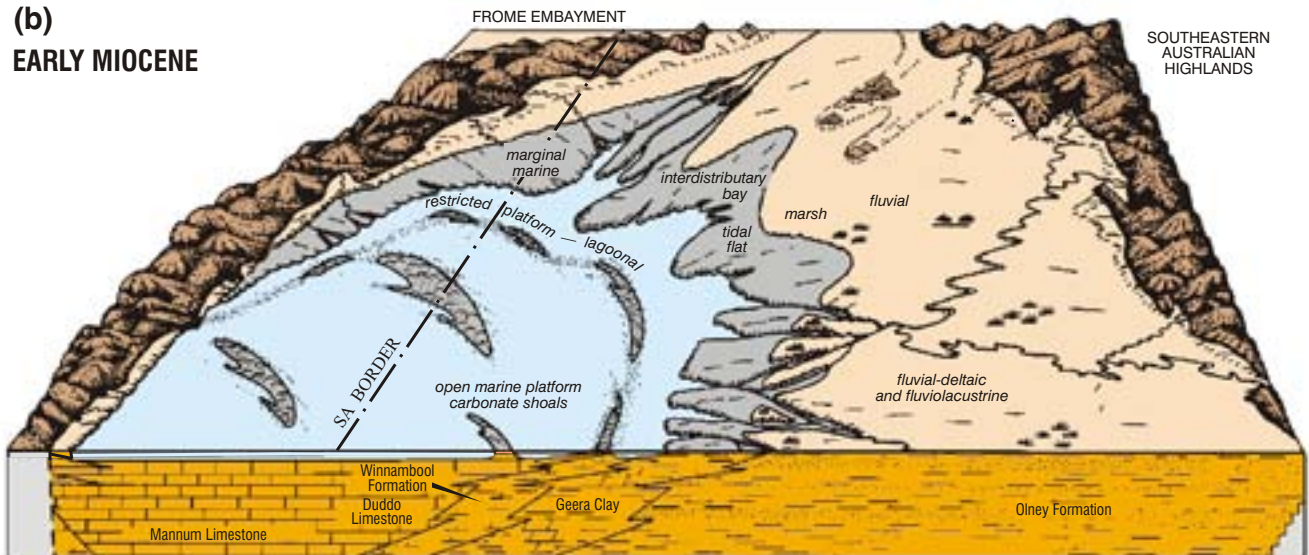
25% expected in the next 25 years. The cost in lost agricultural production and land value will be in millions of dollars.

The watertable over the remainder of the Murray Basin has not yet started to rise because of its greater depth (40-50 m), and is not expected to do so for another 20-30 years. Computer modelling indicates that in 25 years time, the increase in saline groundwater flow into the Murray River will have increased its salinity at Morgan by ~50 EC units, resulting in an annual cost of \$4.7 million in corrosion effects, treatment costs and decreased agricultural yields.

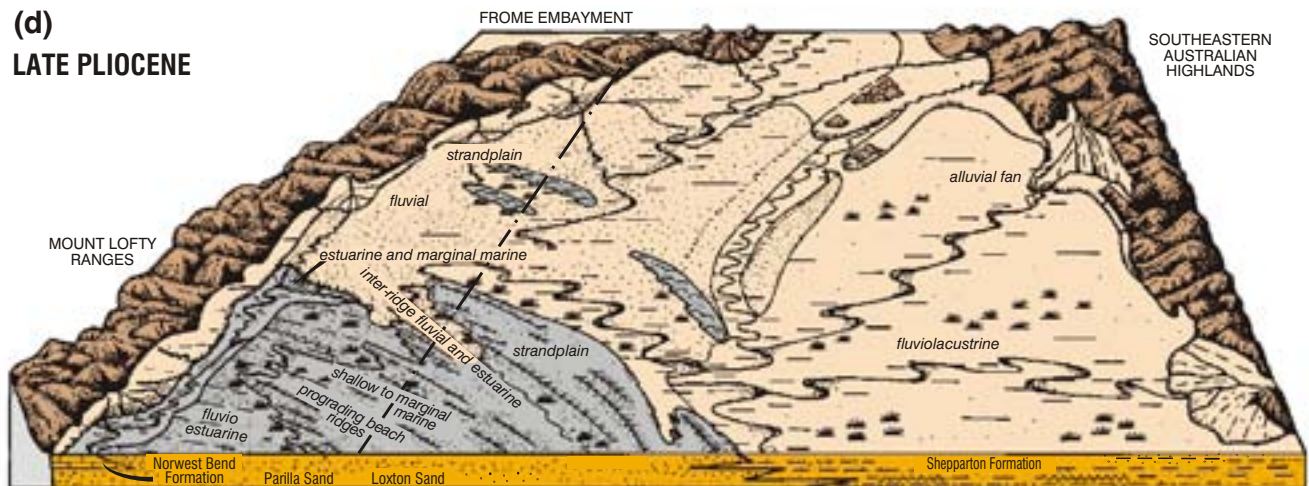
The establishment of extensive irrigation areas adjacent to the river has led to the formation of large watertable mounds due to the percolation of drainage water. These mounds displace saline groundwater into the river, adding an estimated 570 t/day of salt.

Groundwater irrigation in the Mallee Proclaimed Region uses about 8 000 ML/year. A management study has recommended a maximum theoretical allotment of 47 500 ML/year which incorporates an allowance for controlled mining of the large amount of groundwater stored in the limestone aquifer. Smaller areas of irrigation are found at Tintinara and along the Marne River where some concern has been expressed about farm dams in the catchment reducing stream flow and aquifer recharge.

(b)
EARLY MIOCENE



(d)
LATE PLIOCENE



Dm TF

95-0608 MESA

ST VINCENT BASIN

J.M. Lindsay and N.F. Alley

Structure and setting

The intracratonic St Vincent Basin dates back to the Middle Eocene (Figs 10.10, 10.11). It is connected to the Pirie Basin by a narrow corridor in the Crystal Brook area containing over 100 m of Tertiary carbonaceous quartz sand (Kremor, 1986), and a palaeochannel trending northwards from Ardrossan containing up to 90 m of Tertiary sand, silt and limestone (Pain *et al.*, 1992). In the south the basin is largely separated from the southern continental margin and Eucla Basin by Kangaroo Island, and by a geophysically defined basement high with little Cainozoic cover across Investigator Strait.

Gulf St Vincent has limited access to the Southern Ocean through Backstairs Passage and Investigator Strait. The nature of the sedimentary and fossil record suggests that the basin has had similar restricted marine access through much of its history. Basin strata are up to 700 m thick and were laid down in a shallow graben bounded by folded and block-faulted Proterozoic and Palaeozoic rocks (Sprigg and Stackler, 1965; Cooper, 1985).

The basin has been subdivided into a number of sub-basins concordant with the grain of the Delamerian fold belt. The largest is the Adelaide Plains Sub-basin which has been drilled extensively for groundwater supplies, yielding much information on subsurface stratigraphy. The Golden Grove, Noarlunga and Willunga Embayments are asymmetric tectonic valleys in which the wedge of sediments dips gently southwards and thickens towards presumed faults near their southeastern margins.

Stratigraphy and age

Unlike other basins near the South Australian continental margin, the St Vincent Basin does not appear to be underlain by Mesozoic sediments, although rare recycled Cretaceous palynomorphs of unknown provenance are encountered here and in the Barossa Basin. Some of the Tertiary sediments rest on Permian Cape Jervis Formation.

Generally, highly weathered bedrock underlies the early Tertiary sediments in most parts of the basin (Daily *et al.*, 1976; Selby and Lindsay, 1982). Drillholes intersecting the Tertiary-Precambrian contact in the northern Adelaide Plains have revealed plant roots preserved at the surface of a palaeosol developed on weathered

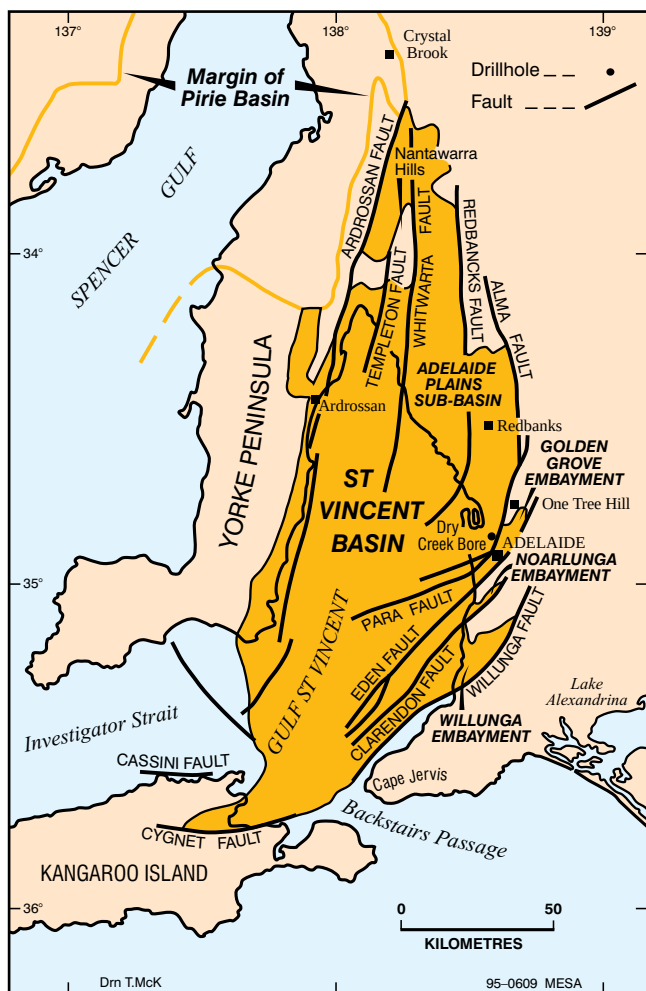


Fig. 10.10 Map showing limits and major structures of the St Vincent Basin (after Cooper, 1985).

basement. At One Tree Hill a similar palaeosol underlying Middle Eocene North Maslin Sand equivalent is exposed in PGH sandpit (Alley and Broadbridge, 1992). Late Eocene sediments overlie weathered and ferruginised bedrock in the northernmost St Vincent Basin. No Paleocene or Early Eocene deposits are known, indicating that the basin did not begin to develop until the Middle Eocene.

Middle Eocene non-marine sequence

North Maslin Sand

The Tertiary succession commences with the North Maslin Sand (formerly North Maslin Sands) which contains palynofloras correlated with the Middle Eocene Lower *N. asperus* Zone (Fig. 10.11; McGowran *et al.*, 1970; Harris, 1985; Alley, 1987b). The formation comprises generally upward-fining units of cross-bedded quartz sand with poorly sorted quartz gravel, suggesting deposition in a braided-river system. It contains occasional plant-bearing clay lenses deposited in fluvial-lacustrine conditions. These lenses are usually thin, oxidised and bleached, but an unweathered bed in the Readymix Maslin Beach quarry yielded well-preserved plant macro- and microfossils of Middle Eocene age (Lange, 1970; McGowran *et al.*, 1970; Christophel and Blackburn, 1978; Harris, 1985; Alley, 1987b; Pain, 1988). Similar plant-bearing clays have been uncovered and intersected by drilling at Golden Grove (Christophel and Greenwood, 1987; Alley, 1987b; McCallum, 1990). North Maslin Sand equivalent consisting of fluvial-lacustrine sand, silt, gravel and carbonaceous clay is exposed

in quarries in the One Tree Hill area (Pain and Scott, 1984; Alley and Broadbridge, 1992). The Maslin Bay, Golden Grove and One Tree Hill macrofloras contain abundant and diverse leaves, fruits, flowers and plant cuticle (see 'Geological history and palaeoclimate'). Remnants of fluvial-lacustrine sand and gravel are widespread at high levels within the Mount Lofty Ranges. Some of these, because of their geomorphic position, may be equivalent to North Maslin Sand.

A palaeochannel eroded in Cambrian and Precambrian bedrock and extending for more than 60 km northwards from Ardrossan contains up to 90 m of Tertiary sediments (Fig. 10.12; Pain *et al.*, 1992). The basal part of the succession is a fluvial sand with minor silt and clay. Although unfossiliferous, these clastics are tentatively correlated with North Maslin Sand. Bedding indicates they were deposited in fluvial to deltaic environments by flow from the north. Valley sides of the palaeochannel are probably largely controlled by faulting (Pain *et al.*, 1992). The southeastern margin is marked by the Ardrossan Fault, where the sediments may be displaced by ~100 m with downthrow to the east. In some parts of the palaeochannel the sand is overlain by Miocene Melton Limestone.

Middle Eocene-Middle Miocene marine sequence set

Clinton Formation

Brown coal and carbonaceous clay, silt and sand comprising the Clinton Formation is widespread in the northern and eastern St Vincent Basin (Harris, 1966, 1970b, 1985). In the Gawler area the formation overlies North Maslin Sand with sharp disconformity (Lindsay, 1985, fig. 8). Here and beneath the City of Adelaide and in the Noarlunga Embayment the formation is overlain conformably by South Maslin Sand, but is not present in the Maslin Beach succession.

Coal deposits within the Clinton Formation are restricted to the northern St Vincent Basin (see below). In this area the formation is largely paralic to marine, but terrestrial intercalations occur throughout and may be more common north of the Nantawarra Hills; marine limestones are restricted to south of the hills (Alley, 1973). Along the eastern part of Yorke Peninsula the Muloowurtie Formation (containing the Quartoo Sand Member) and probably Throoka Silt are correlatives of the Clinton Formation. South of the Nantawarra Hills, the Clinton Formation is generally ~70 m thick and comprises interbedded sand, silt, clay and lignite with abundant wood and other plant matter (Fig. 10.16). However, greater thicknesses occur north of Clinton and below Gulf St Vincent (Johnson, 1960b).

Palynofloras from the Clinton Formation have been correlated with the Middle and Upper *N. asperus* Zone, indicating an age range from late Middle Eocene to Early Oligocene (Harris, 1966, 1970b; Meakin, 1985). However, palynofloras from the type and reference sections in SADME drillholes Clinton 23/23A and Inkerman 32, and in Sawax drillhole Clinton 20, are now regarded as being largely correlative with the Upper *N. asperus* Zone, with a few lower samples extending down into the Middle *N. asperus* Zone. This implies that Clinton Formation in the northern St Vincent Basin is Late Eocene (probably latest Eocene) to Early Oligocene in age. Microplankton are rare to very infrequent throughout the formation, indicating deposition in estuarine to deltaic and fluvial-lacustrine environments.

A thick succession is preserved in small, tectonically controlled basins north of the Nantawarra Hills, the most extensive being near Lochiel (see below). The formation here has been subdivided into three units (Kremor *et al.*, 1995), from the base:

- Bumbunga Sand Member, 30 m of fluvial sand, silty sand and coal
- Condowie Silt Member, 20 m of partly carbonaceous lacustrine silt
- Kooliata Coal Member, 20 m of coal and lacustrine sand and silt.

These units are unconformably overlain by 5-20 m of intertonguing, marine silty sand (Warrindi Silt) and 5-20 m of

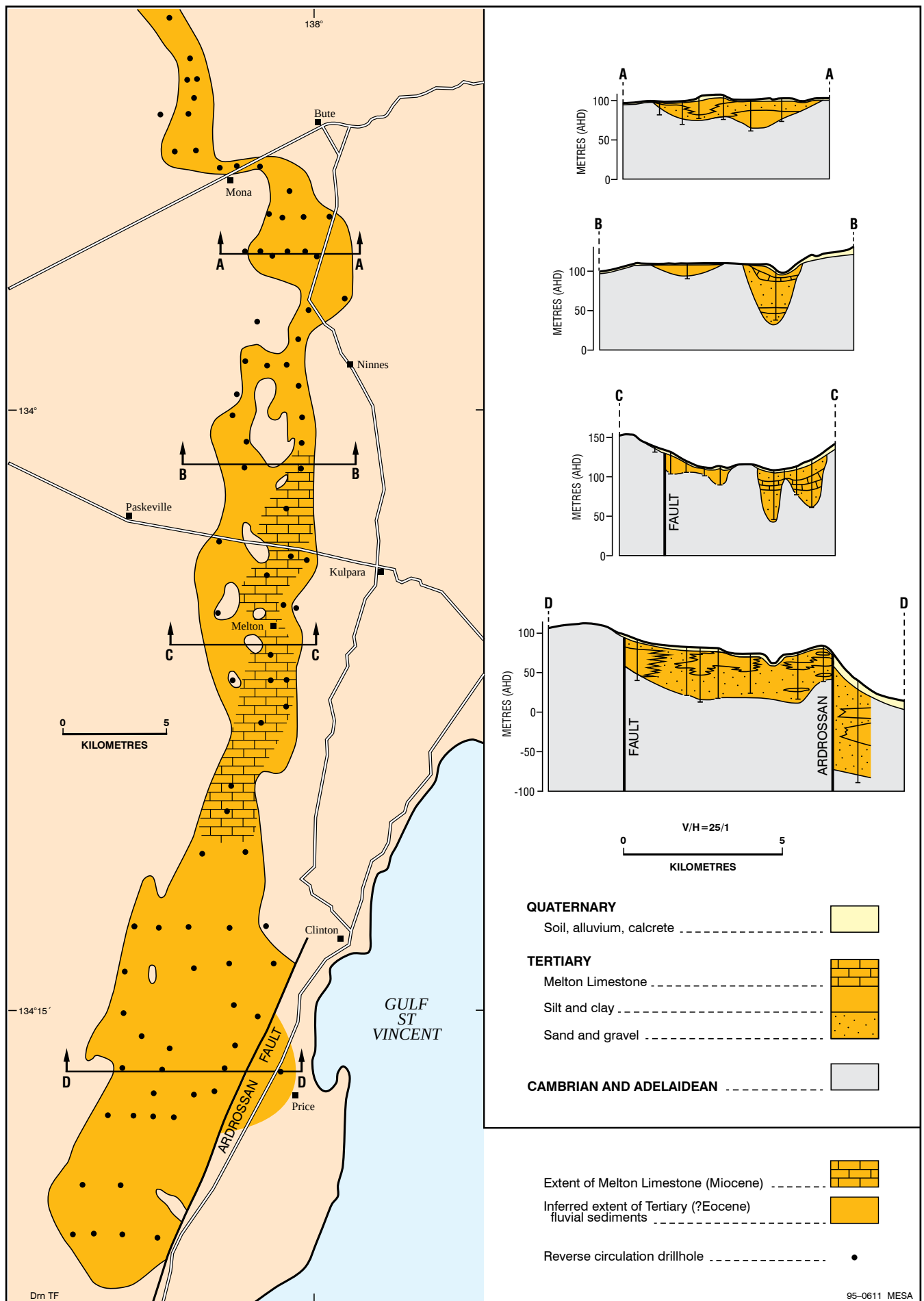


Fig. 10.12 Tertiary palaeochannel sediments, northeastern Yorke Peninsula (after Pain *et al.*, 1992).



estuarine carbonaceous silt (Tarella Silt), both dated as Oligocene.

In Barunga Coal Bore 4, east of Snowtown, the Clinton Formation was divided into Eocene and Miocene units on palynological evidence (Harris, 1966, 1970b). Sediments below 74 m depth were assigned to a Middle to Late Eocene phase of estuarine deposition and the interval 58-74 m to a Miocene lacustrine phase. On this basis, Alley (1973) defined the subsurface Miocene Snowtown Sand (formerly Snowtown Sands), believed to have been deposited in a lake formed by uplift of the Nantawarra Hills which caused damming of southerly drainage. However, palynofloras from below 58 m in Barunga Coal Bore 4 are now regarded as being correlative with the Oligocene-Early Miocene *P. tuberculatus* Zone. Sediments are non-marine below 74 m, and estuarine from 58 to 74 m. A sample from 52 m depth in the nearby Boucaut Coal Bore 4 contains a non-marine palynoflora also correlative with, but significantly different from, those in the *P. tuberculatus* Zone. This palynoflora bears some similarity to the later Oligocene and Early Miocene palynofloras from the Barossa Basin (see below). The dating and lithology of sediments from the Snowtown area strongly suggests that the Snowtown Sand extends to the total depth of Barunga Coal Bore 4, and that it correlates with the Warrindi and Tarella Silt of the Lochiel area.

Dating in the northern part of the basin and the stratigraphic position of the Clinton Formation between North Maslin Sand and South Maslin Sand in the Adelaide area indicates that it becomes progressively younger northwards. The marginal marine and diachronous character of the unit implies deposition of Clinton Formation under conditions of gradual marine transgression from the south.

South Maslin Sand

At Maslin Bay, the first sign of marine influence appears at the bioturbated top of the North Maslin Sand, after which the transgression develops slowly in the marginal marine and glauconitic South Maslin Sand (formerly South Maslin Sands). The South Maslin Sand is oxidised in outcrop, but carbonaceous and pyritic at depth (Lindsay, 1969; Selby and Lindsay, 1982). The unit contains a sparse marine fauna of bryozoal, bivalve, gastropod and echinoid remains, foraminifera and sponge spicules. Goethite pellets are abundant in places; sand-sized faecal pellets and aggregates of glauconitic and goethitic clay commonly comprise 20% of the sediment. The South Maslin Sand in the Willunga Embayment is at least 41 m thick (Pain, 1988; Pain and Keeling, 1990), and the contact with North Maslin Sand varies from conformable interfingering to erosional disconformity. In the Adelaide Plains Sub-basin, South Maslin Sand interfingers with Clinton Formation and ranges in age from Middle to earliest Late Eocene.

Tortachilla Limestone

Deposition of the Tortachilla Limestone in the southern part of the St Vincent Basin resulted from the Tortachilla Transgression and contains foraminiferal faunas correlative with Zone P15 of Late Eocene age (McGowran, 1989, 1991). The unit contains warm-water benthonic foraminifera and represents the peak of the first of three Late Eocene regional transgressive warm episodes (Lindsay, 1976; McGowran, 1989). At Maslin Bay, the Tortachilla Limestone comprises a basal bryozoal ferruginous sand ~1 m thick, grading up into 1 m of richly fossiliferous limestone, which

Top: Southerly view of Maslin Bay, with South Maslin Sand in the foreground. The wave-cut platform below the cliffs at top right is Tortachilla Limestone, overlain by Blanche Point Formation, thin Hallett Cove Sandstone, and reddish Pleistocene units. Photo 032120.

Centre: Middle Eocene leaf fossils from rainforest plants in a carbonaceous clay interval of North Maslin Sand at Golden Grove. Photo 042495.

Bottom: Cainozoic succession in the Bowmans trial coal pit. From the base: Clinton Formation lignite, carbonaceous clay and then white sand; yellow and brown glauconitic sand and silt of the Rogue Formation; red-brown Hindmarsh Clay (Quaternary). November 1980. Photo 042496.

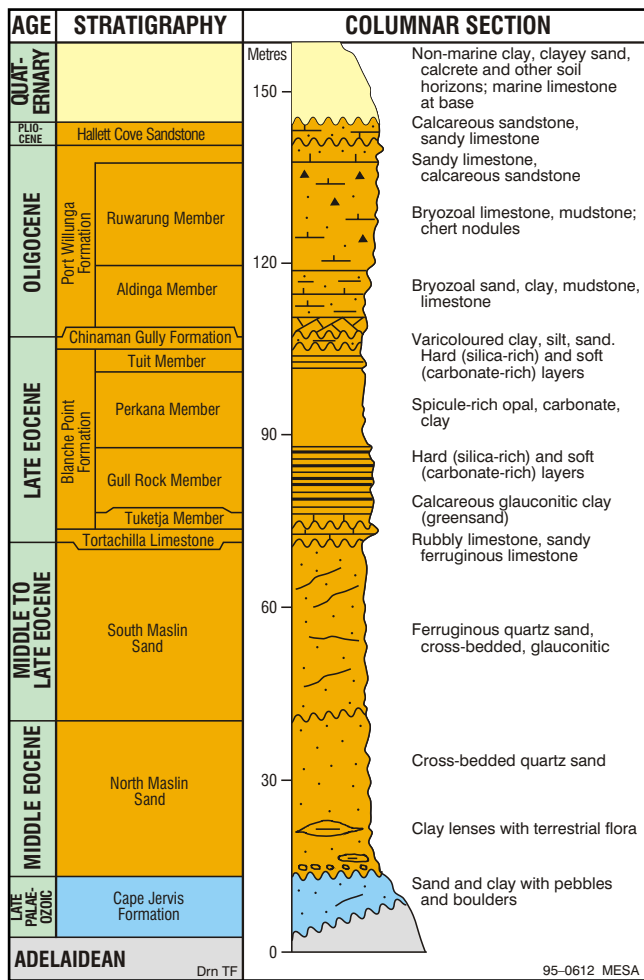


Fig. 10.13 Composite columnar section of Tertiary units, Maslin-Aldinga area (after Lindsay and McGowran, 1986).

is overlain irregularly by 1 m of glauconitic fossiliferous limestone (Fig. 10.13). This unit crops out near Witton Bluff and the Onkaparinga River estuary, and occurs subsurface in the Adelaide Plains Sub-basin and along the Golden Grove Embayment as far as Campbelltown.

The lowest unit of the Kingscote Limestone on Kangaroo Island correlates in part with Tortachilla Limestone, as does the lower part of the Muloowurtie Formation along eastern Yorke Peninsula.

Blanche Point Formation

The Tortachilla Limestone is succeeded by the Late Eocene Blanche Point Formation which has been subdivided into four members. The lowest, the Tuketja Member, is 2.3 m thick at the type section and consists of green and grey, very glauconitic and fossiliferous calcareous mudstone with a few thin limestone lenses. It contains the planktonic foraminifer *Hantkenina alabamensis primitiva*, which is valuable in regional correlation and as an indicator of warm transgressive conditions (Tuketja Transgression of McGowran, 1989, 1991). Overlying the Tuketja at the type section is the 12.2 m thick Gull Rock Member, characterised by numerous bands of spicular chert (Cooper, 1979; Jenkins *et al.*, 1982). It is widespread in the Willunga and Noarlunga Embayments, beneath the City of Adelaide, and in the Adelaide Plains Sub-basin (Fig. 10.14). The overlying Perkan Member is a massive calcareous mudstone and spongolite that is not present beneath the City of Adelaide (partly due to erosion), whereas in the Adelaide Plains Sub-basin it is difficult to distinguish from

mudstone of the Aldinga Member of the Port Willunga Formation. The Blanche Point Formation occurs extensively subsurface in the Willunga Embayment, grading in places to marginal sandy facies (Cooper, 1979). The unit crops out along the lower Onkaparinga River and at Witton Bluff and Blanche Point.

The top of the formation at the type section and at Port Noarlunga is marked by the Tuit Member, containing several chert-bearing bands similar to the Gull Rock Member. The silica-rich nature and layering of Blanche Point Formation and abundance of smectite have been ascribed to volcanic ash generated by explosive volcanism during separation of Australia from Antarctica (Jones and Fitzgerald, 1986, 1987). However, the source of the silica is problematic because no volcanic glass shards have been found.

Chinaman Gully Formation

At Port Willunga, the marine Blanche Point and Port Willunga Formations are separated by 2 m of regressive, varicoloured sand, silt and clay of the Chinaman Gully Formation. This unit is usually dark, carbonaceous and pyritic where unoxidised at depth further inland (Lindsay, 1969; Cooper, 1979). The formation is overlain disconformably by the basal gravelly sand of the Port Willunga Formation (Ludbrook, 1956; Lindsay, 1967). Beneath the City of Adelaide, where it is about 10 m thick, the formation comprises an upper mudstone unit and a lower sand, the latter being the Tandanya Sand Member. In the Adelaide Plains Sub-basin, Chinaman Gully Formation has been intersected in only a few drillholes; here it forms a regressive intercalation of dark lignitic silt and clay (Lindsay, 1985, fig. 7).

The Eocene-Oligocene boundary was originally placed near the top of the Aldinga Member of the succeeding Port Willunga Formation, this also marking the top of the Aldingan Stage (Ludbrook and Lindsay, 1966; Lindsay, 1985; Lindsay and McGowran, 1986). However, in view of the major faunal turnover at the base of the Chinaman Gully Formation, concurrent with a major global sea-level fall and cooling, the Eocene-Oligocene boundary is now placed at the base of the Chinaman Gully Formation (McGowran *et al.*, 1992).

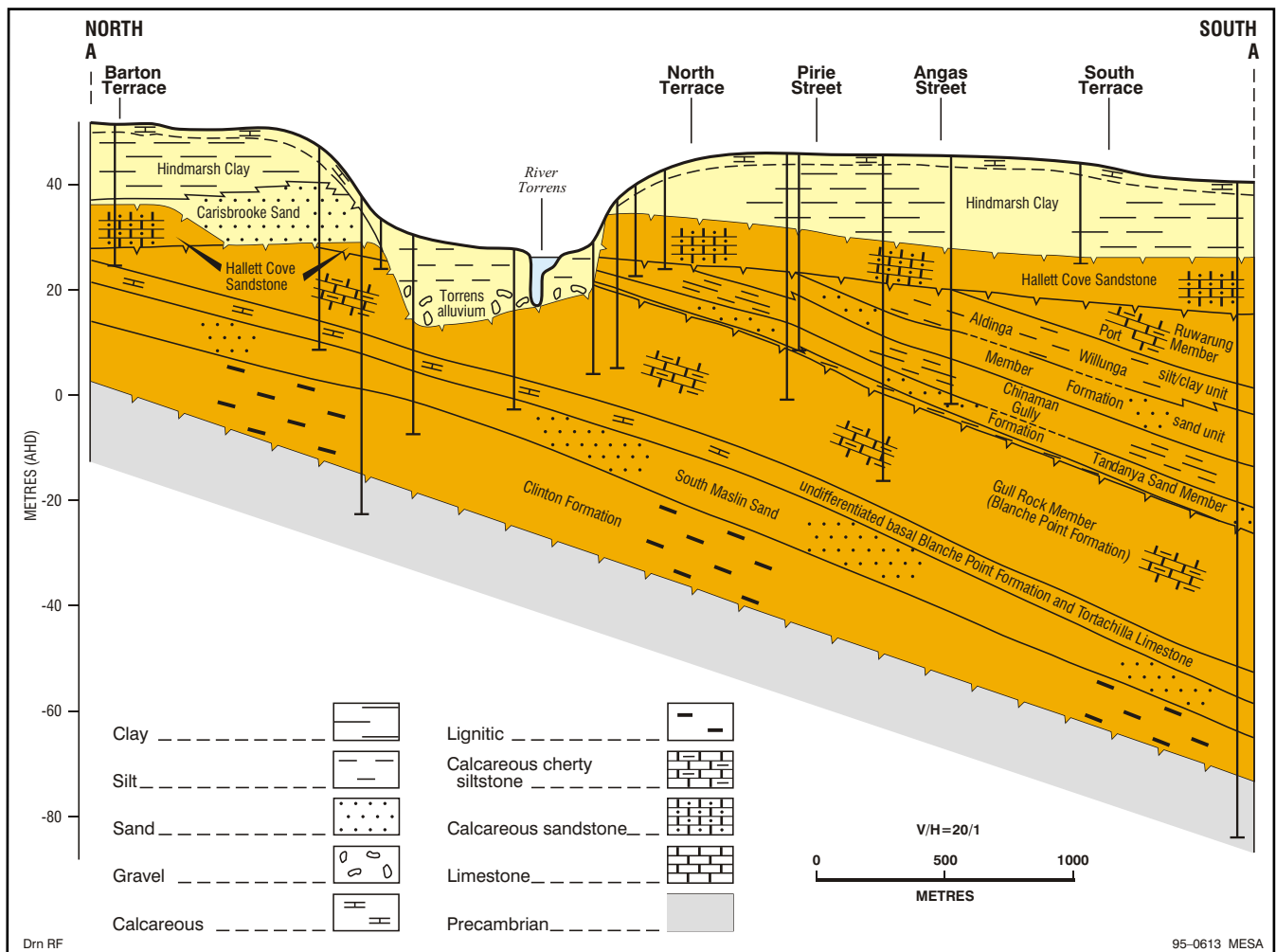
Port Willunga Formation

The Oligocene Port Willunga Formation is present through much of the St Vincent Basin as bryozoal marly limestone up to 33 m thick. At the type section the formation is divided into three units:

- the lowest, the Aldinga Member, comprises fossiliferous, partly cross-bedded sand and sandstone, muddy limestone, mudstone, and a few thin bands of green clay
- the Ruwarung Member is a fossiliferous limestone, marl and silt containing chert bands and Early Oligocene faunas
- the uppermost unnamed unit near Snapper Point consists of ~4 m of fossiliferous limestone, silty sand, and cross-bedded sandstone overlain by Pliocene sediments.

The global planktonic foraminiferal extinction datum of *C. cubensis* (Zone P21a/b boundary marker), in the presence of continuing *Guembelitra*, occurs ~2 m above the base of the unnamed member. This provides a biostratigraphic tie with the basal part of the Janjukian Stage stratotype defined from western Victoria (Lindsay, 1967). Hence, the uppermost part of the Port Willunga Formation at the type section has been termed informally the 'Janjukian unit' (Lindsay, 1985; Lindsay and McGowran, 1986). The same unit crops out in the Noarlunga Embayment and is widespread subsurface to the Adelaide Plains.

On Kangaroo Island the 1 m thick, non-cherty, disconformity bounded middle unit of the Kingscote Limestone probably correlates with part of the Ruwarung Member, whereas the uppermost 3 m of the limestone corresponds to the 'Janjukian unit' of the Port Willunga Formation. Late Oligocene fossiliferous chert, known only from beach boulders west of Cape Hart on southeastern Kangaroo Island, represents either Ruwarung Member



in the southernmost part of the St Vincent Basin or a facies of Ettrick Formation in the westernmost Murray Basin.

Erosion between the Middle Miocene and Late Pliocene has removed Miocene beds from the type section of the Port Willunga Formation. At Sellicks Beach, however, coarsely bryozoal limestone of the Port Willunga Formation yields foraminiferal ages as young as Early Miocene. In coastal cliffs and the wave-cut platform extending southwest from Sellicks Beach, the formation is steeply upturned, presumably due to movements along the Willunga Fault. Elsewhere in the Willunga Embayment, Port Willunga Formation of the same age occurs from 30 m below to 60 m above sea level (Cooper, 1979). Similar bryozoal limestone is encountered subsurface but some 102-122 m above sea level in the Myponga Valley, suggesting significant post-Middle Miocene uplift (see 'Various intramontane basins').

In the Adelaide Plains Sub-basin, more complete and thicker subsurface sections of the formation are encountered, including a Middle Miocene facies. The upper part of the formation includes the Munno Para Clay Member, an intercalation of fossiliferous, restricted marine blue-grey clay (Fig. 10.11; Lindsay and Shepherd, 1966; Lindsay, 1969). The member reaches a thickness of 23 m but is often much less and tapers out near the Gawler River. Where thicker, it is an aquiclude between the two main groundwater aquifers of the Adelaide Plains.

The Pirramimma Sand Member is a quartzose, sparsely fossiliferous and glauconitic, marginal sandy facies up to 50 m thick. The member was defined primarily in relation to the Willunga Embayment, but probably has wide application in the eastern St Vincent Basin. Sparsely fossiliferous marginal marine sandstone on the Para Fault scarp near Gawler is Late Oligocene (Janjukian) and correlates with the Pirramimma Sand Member (Lindsay, 1988).

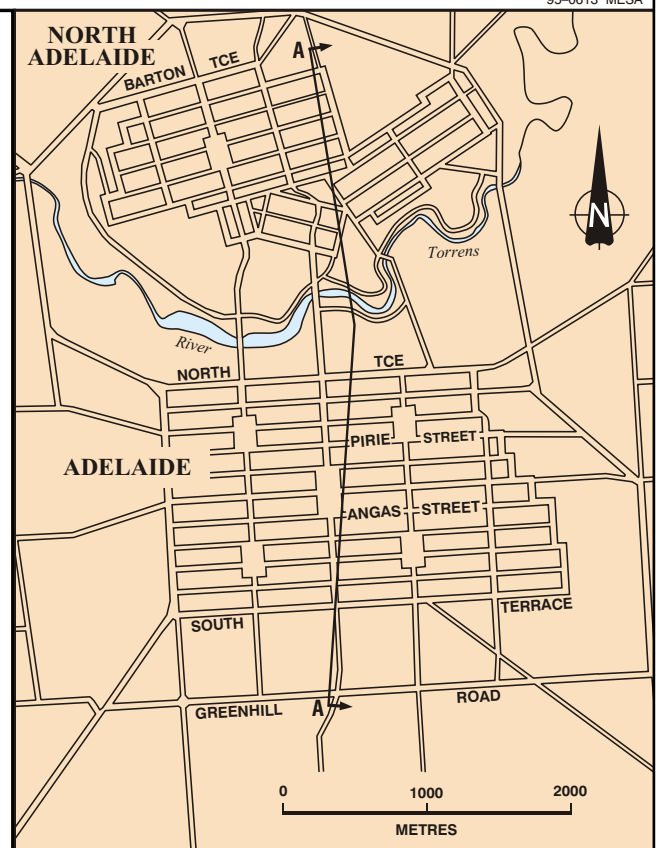


Fig. 10.11 North-south cross-section through the Adelaide area (after Selby and Lindsay, 1982).

Rogue Formation

On eastern Yorke Peninsula, the Rogue Formation comprises sparsely fossiliferous, marginal marine quartz sand and sandstone with subordinate limestone, mudstone and clay (Stuart, 1970). Cherty bands and nodules occur at two levels, the lower apparently correlating with part of the Blanche Point Formation and the upper with the Ruwarung Member of the Port Willunga Formation. Foraminifera show that the thin but distinctive Port Julia Greensand Member of the Rogue Formation correlates with part of the Aldinga Member of the Port Willunga Formation (Fig. 10.11). The bryozoal Port Vincent Limestone (Stuart, 1970) and Melton Limestone (see Pirie Basin) overlying ?North Maslin Sand equivalent in the palaeochannel on northeastern Yorke Peninsula are correlative with the Port Willunga Formation.

Miocene-Pliocene units

Koolunga Gravel

Around the northern fringes of the St Vincent Basin, and in the major valleys of the Broughton River drainage basin, numerous deposits of fluviolacustrine siliceous sand and gravel overlie bedrock or older Cainozoic strata. These clastics are referred to as the Koolunga Gravel (formerly Koolunga Gravels), and are probably of late Tertiary age (Alley, 1973).

Hallett Cove Sandstone and Dry Creek Sand

The Port Willunga Formation is truncated with a low-angle unconformity by the Late Pliocene Hallett Cove Sandstone, a 1 m thick, transgressive, shallow marginal marine calcareous sandstone to sandy limestone at the type section near Sandison Reserve Monument. Near Port Willunga, the formation comprises 4 m of interbedded flaggy sandstone, limestone and sand, with a thin band of fossiliferous shelly sand near the base. Outcrops of Hallett Cove Sandstone also occur in the City of Adelaide area, where it has been intersected by drilling and excavation (Figs 10.11, 10.14). At Redbanks it forms a small and thin remnant near the eastern margin of the Adelaide Plains Sub-basin; thin remnants also crop out on Yorke Peninsula and Kangaroo Island.

The shelly Dry Creek Sand (formerly Dry Creek Sands) was the first marine bed to be dated as Pliocene in Australia (Tate, 1890). In the Adelaide Plains Sub-basin, where it is part of an important aquifer, the unit underlies and intertongues with Hallett Cove Sandstone facies. The shelly sand which often occurs near the base of the Hallett Cove Sandstone in the coastal section around Port Willunga and beneath the City of Adelaide is Dry Creek Sand facies. Molluscan faunas of the Hallett Cove Sandstone and central part of the Dry Creek Sand (the Yatalan Fauna), and the associated Yatalan Stage, have been dated as Late Pliocene (Ludbrook, 1963b, 1973a).

Drilling near the original Dry Creek Bore has shown that Dry Creek Sand continues below the depth reached by the old bore, becoming finer grained and silty. This is the basal silty member of the Pliocene sequence referred to as 'Croydon facies' (Lindsay, 1969). It is bryozoal, glauconitic, has common *Ditrupa* tubes, abundant *Elphidium* (*Cellanthus*) *adelaidense* and contains some foraminiferal indication of an Early Pliocene age. Including this member, the total thickness of Dry Creek Sand at the type locality is 43 m but may exceed 60 m elsewhere.

Burnham Limestone

At Port Willunga and Blanche Point, 1-3 m of sparsely fossiliferous, clayey, micritic Burnham Limestone overlies a karstic surface on Hallett Cove Sandstone. In the City of Adelaide area, the formation is often difficult to distinguish from weathered and eroded Hallett Cove Sandstone. The Burnham Limestone, and the correlative Point Ellen Formation at Cape Jervis and on Kangaroo Island (Ludbrook, 1983), contain the pelagic gastropod *Hartungia* (May and Bourman, 1984) which is not known above the Pliocene in New Zealand (Beu and Maxwell, 1990).

Economic geology

Construction sand

The North Maslin Sand extends near-surface for at least 5 km across the northern margin of the Willunga Embayment where it forms a major resource of construction sand for metropolitan Adelaide (Olliver and Weir, 1967; Pain, 1988). Annual production of sand and gravel from the Maslin Beach quarries between 1990 and 1993 averaged 500 000 t. The deposits have been mined since 1927, and ~30 Mt of construction sand and 20 Mt of filling sand remain (A.M. Pain, MESA, pers. comm., 1994). Reserve estimates are being revised.

North Maslin Sand is also quarried in the Golden Grove and Highbury districts of the Golden Grove Embayment where reserves of 54 Mt of construction sand and 24 Mt of filling sand have been proven by drilling (McCallum, 1990). Although clay is quarried from North Maslin Sand and the weathered bedrock at One Tree Hill, there are no significant resources of construction sand in this area (Pain and Scott, 1984). A significant resource of construction sand, currently being mined, occupies a palaeochannel extending northwards from Ardrossan (Pain *et al.*, 1992).

Coal

Carbonaceous sediments of the Clinton Formation and Kooliata Coal Member in the northern St Vincent Basin contain reserves of brown coal exceeding 3 Gt. Five deposits have been delineated, namely Bowmans (previously referred to as either Inkerman-Balalava or Wakefield), Lochiel, Clinton, Whitwarta and Beaufort (Fig. 10.15). The Bowmans and Lochiel deposits are the largest and have been investigated as potential fuel sources for power generation. The coals are low rank (classified under ATSM as Lignite B), possessing high moisture, low heating value, moderate ash, and relatively high sodium, chlorine and sulphur contents.

The Bowmans resource is 1.25 Gt and occurs as a number of flat-lying seams 55-100 m below the surface. Individual seam thicknesses vary from 0.5 to 15 m, with a maximum cumulative thickness of 25 m, although rapid thinning and splitting are common (Fig. 10.16). The Lochiel resource is 629 Mt, situated in a fault-angle depression in which both the basin floor and Tertiary sediments dip at ~1° northwestwards towards the Ardrossan Fault (Fig. 10.16). The coal occurs largely in three seams within the Kooliata Coal Member, averaging 1.8, 6.5 and 2.6 m in thickness. It is overlain by unconsolidated clay, silt and water-saturated sand, ranging from 20 to 70 m in thickness.

Groundwater

Groundwater occurs throughout the St Vincent Basin but the better quality resources are contained in the Adelaide Plains Sub-basin, and Willunga and Noarlunga Embayments. Important aquifers (where confined) are the North and South Maslin Sands, Tortachilla Limestone, Chinaman Gully Formation, and sandy facies of the Aldinga Member, lower and upper Port Willunga Formation, Hallett Cove Sandstone and Dry Creek Sand. The following summary is taken from Shepherd (1978), Selby and Lindsay (1982) and Gerges (1982, in prep.).

Aquifers within the Adelaide Plains Sub-basin contain good quality groundwater but north of Two Wells the salinity increases significantly. Two important pressure aquifers within the sub-basin include the lower part of the Port Willunga Formation and the Dry Creek Sand. Salinity of both aquifers is strongly influenced by intake from the Torrens, Little Para and Gawler Rivers. Salinity adjacent to the rivers is low but increases markedly to more than 3 000 mg/L in the intervening areas.

The northern part of the Adelaide Plains is a Proclaimed Region under the Water Resources Act, 1976, and water use is controlled mainly for irrigation purposes. Water consumption in this area greatly exceeds recharge, causing reduction in the potentiometric surface of the aquifers and the possibility of increasing salinity by contamination from adjacent or overlying higher salinity groundwater. In the City of Adelaide area there are no controls

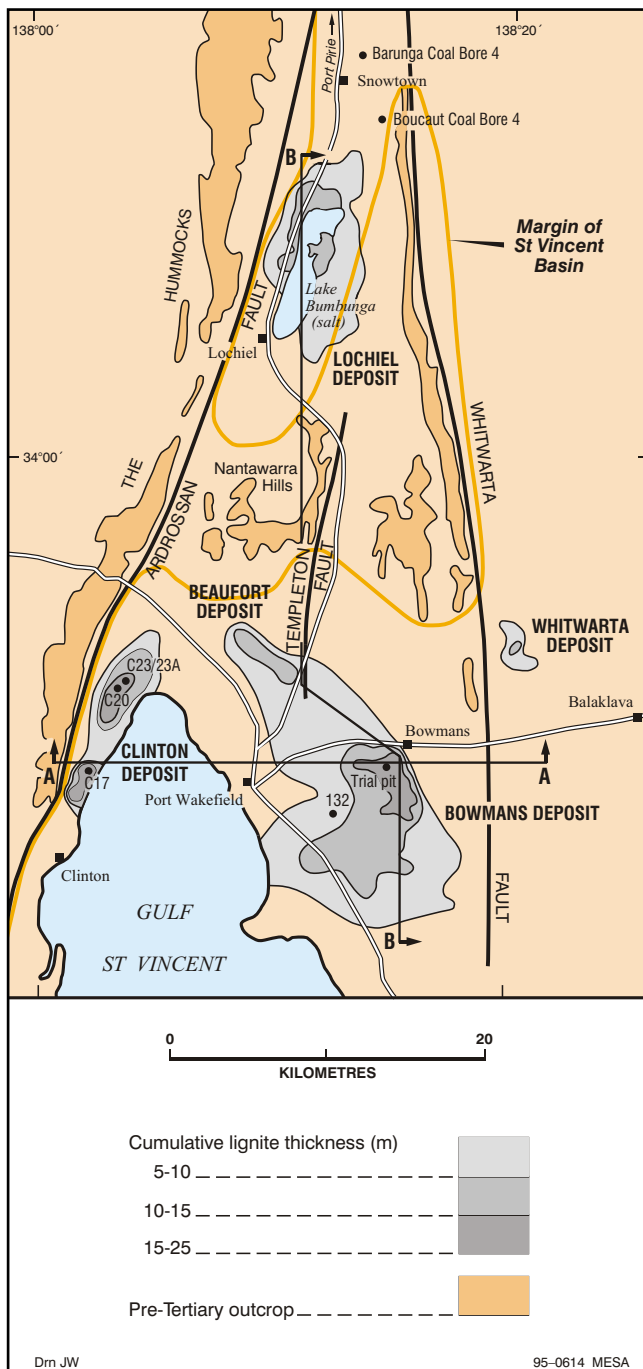


Fig. 10.15 Coalfields of the northern St Vincent Basin.

over the volume of water pumped and substantial quantities are used in industry and for watering golf courses.

The North and South Maslin Sands are the most important aquifers in the Willunga Embayment, providing irrigation for mixed farming, in particular vineyards. Groundwater quality is good, with increasing salinity up to 2 000 mg/L towards the coast. In the Noarlunga Embayment the groundwater resource is relatively small, of poorer quality and used mainly for stock and irrigation.

1. Blanche Point at the southern end of Maslin Beach. Tortachilla Limestone (in the water) is overlain by cliff-forming Blanche Point Formation, Hallett Cove Sandstone and Pleistocene sediments. Photo 042497.
2. Noarlunga and Willunga Embayments, with the Clarendon Fault scarp (foreground) and Willunga Fault scarp on the horizon. Photo 042498.
3. Contact (sloping up to the right from figures) between North and South Maslin Sands in the Maslin Bay area. Photo 042499.
4. North Maslin Sand in the Maslin Bay area. Photo 042500.



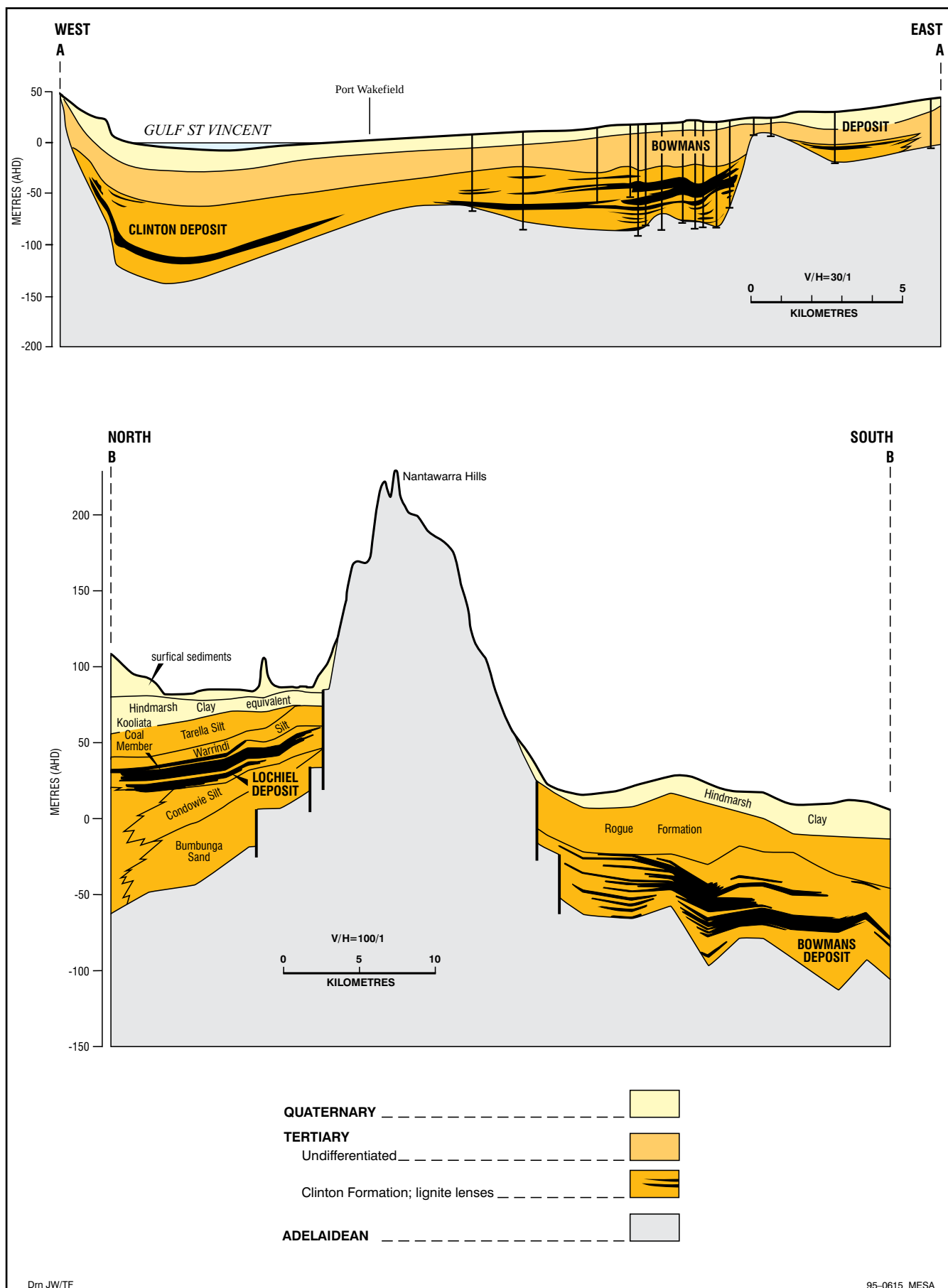


Fig. 10.16 East-west (A-A) and north-south (B-B) cross-sections through the northern St Vincent Basin. A-A shows the distribution of lignites in the Clinton Formation, from Clinton to Bowmans (after Springbett, 1980). B-B shows the relationship between lignites in the Bowmans and Lochiel deposits (after Kremor et al., 1995). Lines of section are located on Figure 10.15.

BAROSSA BASIN

N.F. Alley

Structure and setting

Tertiary fluviolacustrine sand, silt, clay, and lignite in the Barossa Valley, known as the Rowland Flat Sand (formerly Rowland Flat Sands; Fig. 10.17), reach a maximum thickness of ~110 m and are commonly overlain by up to 30 m of Quaternary sediments (Fig. 10.18). The sediments thicken eastwards towards the Barossa Ranges and thin southwards, terminating on bedrock near Rowland Flat and Lyndoch. The valley has been described as a fault-angle depression with a fault along the eastern margin, or as a shallow, asymmetrical syncline with a fault or flexure inferred along the steep southeastern flank (Chugg, 1955; Dalgarno, 1961; Olliver, 1967a). The bedrock floor is irregular, with local relief as great as 20 m along the western side of the valley. Weathering of the floor is highly variable, and Rowland Flat Sand overlies fresh or highly weathered bedrock which, in some localities, is heavily ferruginised (Alley, 1973, 1977).

The Tertiary sediments extend westwards from the Barossa Valley to the Gawler area, and northwards to Freeling and Stockport. These sediments are regarded as Rowland Flat Sand and are included in the Barossa Basin (see below).

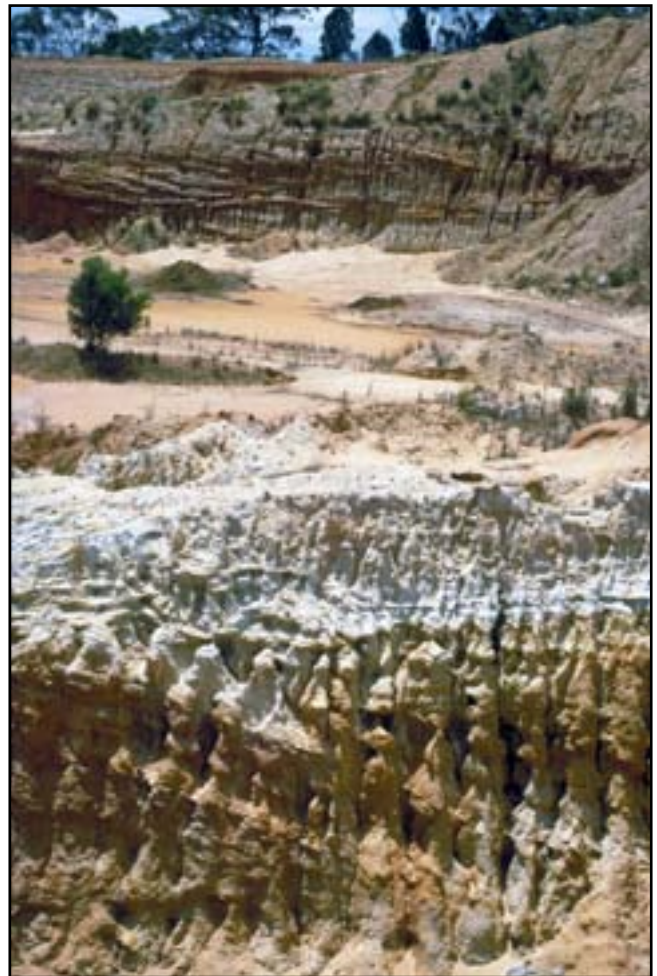
Stratigraphy

Rowland Flat Sand is exposed along the western flank of the Barossa Valley where it is mined extensively for construction material. At the base in some localities is a pebbly sand or conglomerate up to 4 m thick in which the larger clasts consist predominantly of rounded to subrounded milky quartz and quartzite. Overlying the basal conglomerate, and reaching 50 m thickness in the Rowland Flat area, is a succession of sand and gravel characterised by medium and large-scale deltaic bedding. These beds contain 'coal balls' which have been used for palynological dating.

Heavy-mineral assemblages are dominated by staurolite, with lesser amounts of rutile and tourmaline. The shape and surface texture of these minerals suggest a short but violent transport history, whereas their abundance indicates that the source area contained weathered bedrock in which the more resistant minerals were concentrated. Staurolite is derived from Marinoan schist at the northern end of the valley (Dalgarno, 1961; Alley, 1973). Palaeocurrent directions indicate that the sediments were deposited from the north, which is consistent with the geomorphic evolution of the area.

A lignitic facies occurs at depths of 30–100 m in deeper parts of the basin to the north of Rowland Flat (Fig. 10.18). Associated with this facies east of Nuriootpa and northeast of Tanunda is a succession of coarse-grained fluviolacustrine sand and gravel fining upwards to bedded silt and clay (Cobb, 1986). Along the eastern margin of the basin, thin lenses of coarse angular gravel dominated by milky quartz are present, and may be associated with the lignitic sediments. The position of these lenses in the basin, and the angularity of clasts, suggest that they may have been deposited as fans built westwards from the Barossa Ranges. The fluviolacustrine, alluvial fan and lignitic sediments are here included as part of the Rowland Flat Sand.

Scattered outcrops of sand and gravel of the Rowland Flat Sand extend westwards across the Meadows-Williamstown (Kitchener) Fault to the Gawler area, and northwards to Freeling and Stockport (Fig. 10.17). The Lyndoch-Sandy Creek deposits contain the same heavy-mineral assemblage as Rowland Flat Sand (Olliver and Weir, 1967) and, together with sediments as far north as Stockport, occur within the same range of elevation, and palynological evidence indicates deposition contemporaneously with the type Rowland Flat Sand (Alley, 1973). This correlation is confirmed by palynological dating of sediments in the Freeling area (Alley, 1989b).



Cross-bedded Rowland Flat Sand west of Rowland Flat. Photo 042501

Gravel and sand capping the plateau south of Lyndoch and west of Williamstown in the One Tree Hill area are at a similar elevation as Rowland Flat Sand, and were previously interpreted to be correlatives (Campana, 1953). However, palynological dating shows that the One Tree Hill sediments are the Middle Eocene North Maslin Sand (Alley and Broadbridge, 1992; see 'St Vincent Basin').

Age and correlation

There has been considerable debate about the age of the Rowland Flat Sand. On the basis of fossil leaves from near Lyndoch, Scouler (1882) suggested a Miocene age, whereas Hossfeld (1949) assigned a Pliocene age to fossil fruit recovered from a drillhole near Bethany. Due to similarities between Rowland Flat Sand and North Maslin Sand, the Barossa sequences were regarded as Eocene and probably equivalent to North Maslin Sand at Golden Grove (Campana, 1955; Glaessner and Wade, 1958; Dalgarno, 1961). Preliminary palynological evidence suggested possible ages ranging from Late Oligocene to Early Pliocene (Olliver, 1967a; Harris and Olliver, 1965; Olliver and Weir, 1967; Alley, 1973). The carbonaceous facies were inferred to be either Eocene, and thus equivalent to Clinton Formation in the St Vincent Basin, or Miocene-Pliocene in age (Cobb, 1986).

Palynological dating of subsurface carbonaceous sediments, and of 'coal balls' from exposed fluviolacustrine sediments west of Rowland Flat (Alley, in prep.), indicates the following. Generally the older, deeper basinal sediments correlate with the *P. tuberculatus* Zone (Early Oligocene to Early Miocene), while the upper, coarser facies, including the deposits cropping out west of Rowland Flat, correlate with the *Triplopollenites bellus* Zone (late Early to Late Miocene). Rare marine dinoflagellates are present in the older facies, indicating very weak marine influence, possibly of the kind expected at the uppermost part of a delta or estuary.

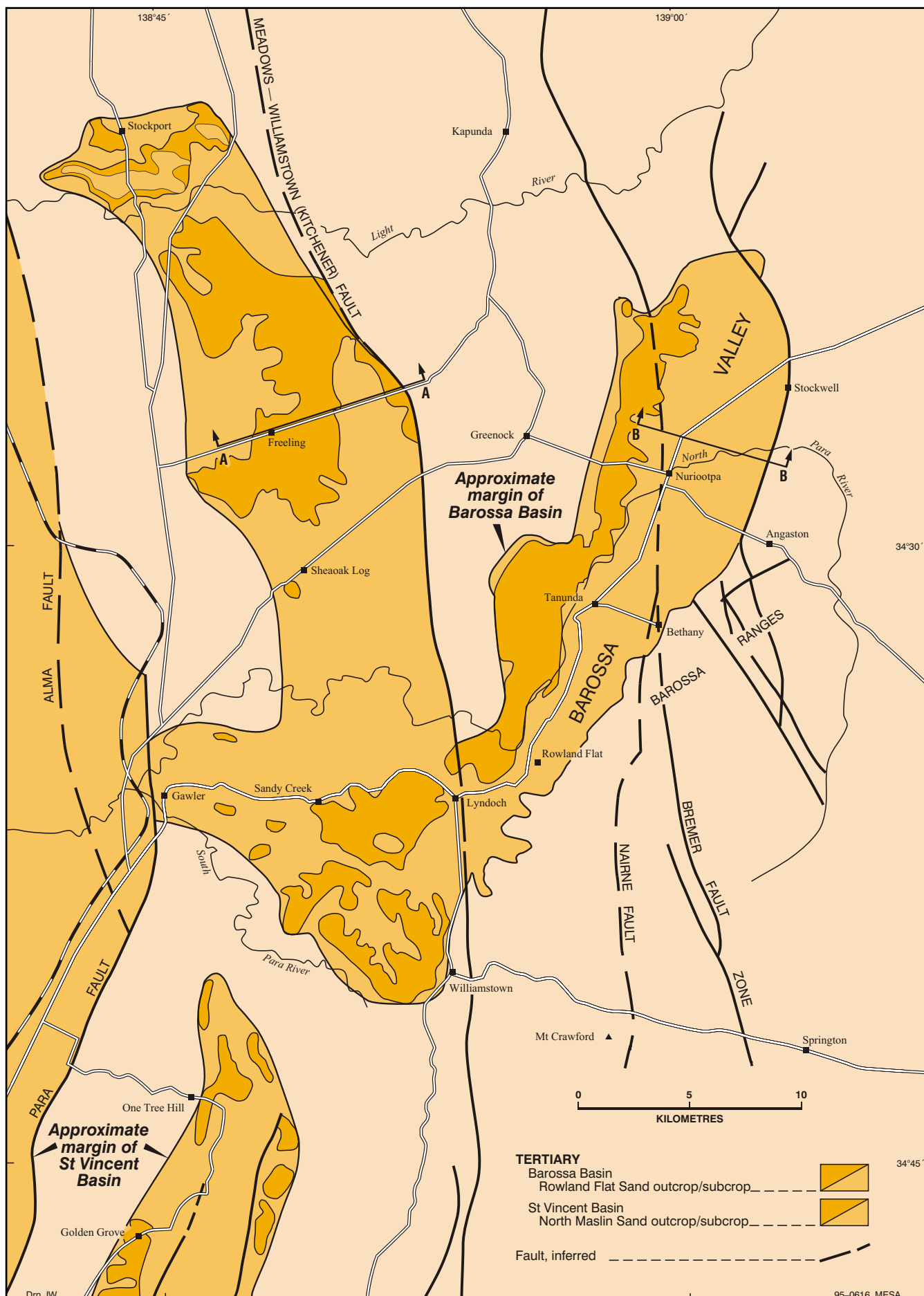


Fig. 10.17 Locality map and structural setting of the Barossa Basin.

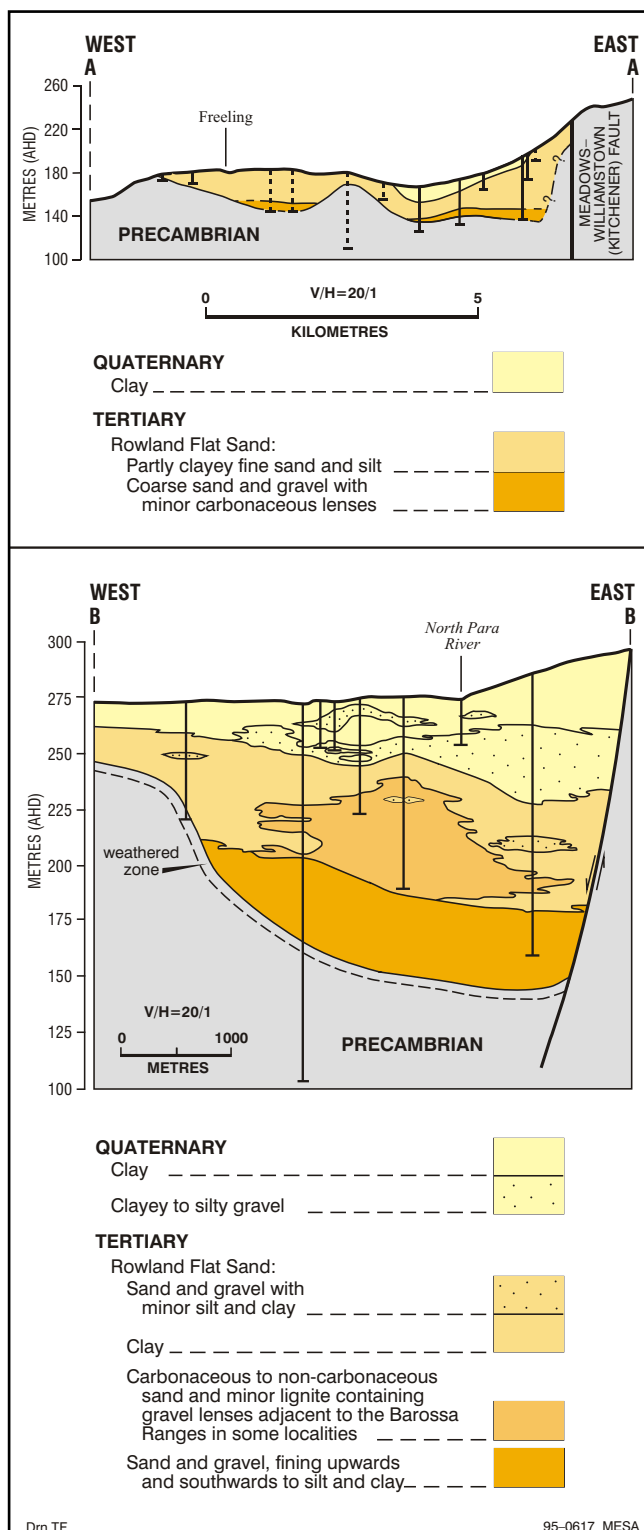


Fig. 10.18 East-west cross-section (A-A) through Tertiary sediments in the Freeling area (after Scott, 1989). East-west cross-section (B-B) through Tertiary sediments of the Barossa Basin north of Nuriootpa (P. Tewkesbury, MESA, pers. comm., 1983). Lines of section are located on Figure 10.17.

Reworked Early Cretaceous palynomorphs which occur rarely in both the older and younger beds suggest that sediments of this age were present somewhere in the former drainage basin. The age of the Rowland Flat Sand implies that weathering of the underlying bedrock occurred no later than the Oligocene.

Drillholes east of Freeling intersected carbonaceous intervals which correlate with the *P. tuberculatus* Zone (Alley, 1989b), indicating equivalence with the lower Rowland Flat Sand. Very rare marine dinoflagellates are also present.

Sandstones which infill palaeovalleys eroded in the Para Fault Scarp at Gawler are dated Early Miocene to latest Oligocene (Ludbrook, 1987; Lindsay, 1988). Scattered outcrops of Tertiary sand extend eastwards from here into the Barossa Basin. The distribution of all of these sediments, and their lithological similarity, suggest that they were deposited contemporaneously. This correlation is compatible with the age of the Rowland Flat Sand and the paralic nature of the Oligocene to Early Miocene beds in the Barossa Valley and Freeling area. Thus, the older part of the Rowland Flat Sand may have been deposited in an estuary at the end of the Oligocene.

Economic geology

Construction sand

The Rowland Flat Sand, particularly in the Rowland Flat-Sandy Creek area, forms an important resource of construction sand, with inferred resources totalling at least 50 Mt of construction sand and 56 Mt of filling sand (Pain and Keeling, 1990). Investigations along the western side of the Barossa Valley between Rowland Flat and Nuriootpa have determined the presence of an economically significant resource of sand and gravel (Keeling, 1984). The extensive deposits of Rowland Flat Sand between Freeling and Stockport, however, are predominantly of fine sand and are unsuitable for construction purposes (Scott, 1989). The Gawler deposits contain ~6 Mt of construction sand and 2 Mt of filling sand (Pain and Keeling, 1990).

Groundwater

Groundwater from the Rowland Flat Sand is used extensively in the irrigation of vines and, to a lesser extent, for washing and cooling in a few wineries. The extremely variable character of the fluvio-lacustrine sediments has led to a complex aquifer system, which can be divided into lower, middle and upper aquifers, the latter probably confined to Quaternary sediments (Cobb, 1986). Groundwater quality is extremely variable but is generally close to that recommended for the irrigation of vines. There is some evidence for rising salinity levels on a regional scale and gradual decline of water storage levels, especially in the Tanunda region.

PIRIE BASIN

N.F. Alley and J.M. Lindsay

Structure and setting

Tertiary sediments in this basin were once thought to be continuous with those of the Lake Torrens lowland, hence the two areas were referred to as the Pirie-Torrens Basin, extending 400 km from Port Broughton to the northern end of Lake Torrens. However, Tertiary sediments beneath the Lake Torrens lowland are non-marine, with deposition commencing in the Early Eocene, whereas those of the Pirie Basin contain marine facies, with deposition commencing in the Late Eocene. For this reason, the Pirie Basin and Torrens Basin are here recognised as two discrete basins (Figs 10.2, 10.19). The extent of any connection is unclear.

There are striking similarities between the Tertiary stratigraphy of the Pirie Basin on the eastern side of Spencer Gulf and the Cowell Basin on the western side. Hence, these basins are regarded as one and referred to as the Pirie Basin. Its southeastern boundary encompasses northern Yorke Peninsula, extending from Port Broughton south to Melton, then west to Wallaroo, in order to include the distribution of Melton Limestone. However, in several aspects of age, lithology and marine influence, sediments in the Pirie Basin show similarities to early and middle Tertiary sediments of the adjacent St Vincent Basin. Drilling by the Electricity Trust of South Australia has demonstrated a connection between the northernmost St Vincent Basin and the Pirie Basin via a narrow trough containing Clinton Formation facies in the Crystal Brook area.

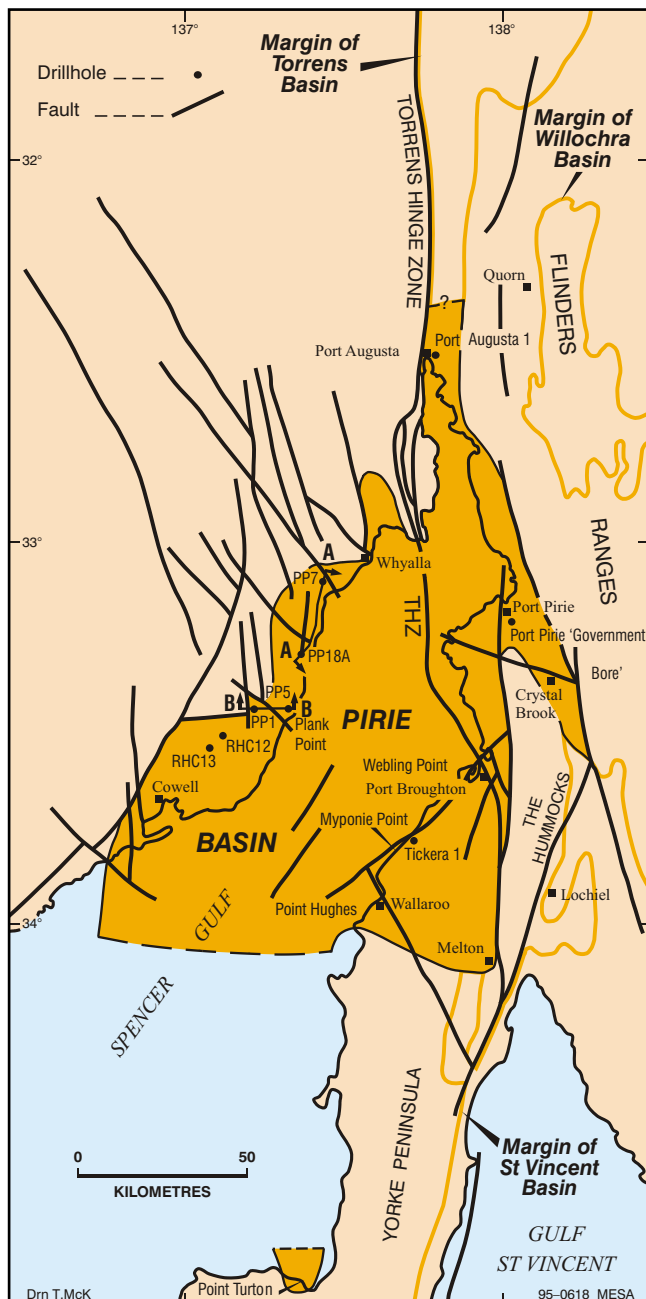


Fig. 10.19 Locality map and structural setting of the Pirie Basin.

The structure of the Pirie Basin is poorly known, although faults along the western side of The Hummocks and Flinders Ranges define the eastern basin margin, and north-south and northeast-southwest-trending faults mark the western margin (Fig. 10.19). These latter faults have vertical displacements of 100-150 m and are often delineated by low scarps up to 70 km in length facing towards Spencer Gulf (Miles, 1952; Parker, 1980).

Stratigraphy

The basin fill comprises three major stratigraphic units including the early Tertiary Kanaka Beds, unconformably overlain by Late Oligocene to Early Miocene unnamed marine sand and Miocene Melton Limestone and then Gibbon Beds which comprise a sand-clay succession (Figs 10.2, 10.20). Very little of the early Tertiary succession crops out at the surface, although it is intersected in many drillholes, particularly along the western basin margin. No information is available on the lithology and thickness of Tertiary units which may underlie Spencer Gulf.

Kanaka Beds

The name 'Kanaka Beds' was introduced for the succession of carbonaceous siltstone, shale and sand, with minor lignite, underlying Melton Limestone (and a thin, unnamed basal sand) in SADME Tickera 1 (Parker, 1980). The Kanaka Beds extend discontinuously from Port Augusta to northern Yorke Peninsula along the eastern basin margin and from north of Whyalla to south of Cowell in the west. In the latter area the unit is up to 60 m thick, with considerable local thicknesses of lignite (Fig. 10.20). However, the lignite is too deep, too poor in quality and too discontinuous (due either to faulting or lensing) to be of economic significance. The Kanaka Beds unconformably overlie fresh and highly weathered Adelaidean strata, or Gawler Craton basement. In the Cowell-Whyalla area, the basal part of the beds comprises 10-15 m of carbonaceous sand and minor conglomerate.

Only scant palynological information is available for the age of the Kanaka Beds. A sample from Port Augusta 1 contains a palynoflora correlative with the Upper *N. asperus* Zone indicating that the beds here may be as young as latest Eocene to Early Oligocene. Several samples from Port Pirie 'Government Bore' contain palynofloras of Eocene aspect. A sludge sample from 148 to 160 m contains an assemblage correlative with the Middle *N. asperus* Zone of Late Eocene age, although no zonal species could be found (Harris, 1972). Palynofloras between 71 and 150 m in Tickera 1 are difficult to interpret and could correlate with any of the Lower, Middle or Upper *N. asperus* Zones. However, the spore-pollen composition and low numbers of dinoflagellates are similar to assemblages from the lithologically similar Clinton Formation in the St Vincent Basin and are here tentatively assigned a Late Eocene to Early Oligocene age. Palynofloras from the Cowell succession (including drillhole PP18A) correlate with the Middle *N. asperus* Zone. The presence of low frequencies of dinoflagellates at some localities indicates that the unit was deposited in a marginal marine to fluvial environment. The thickest section examined (Tickera 1) shows that marine influence may have been restricted to the upper part.

In Tickera 1, an unnamed, poorly sorted quartzose sandstone between the Kanaka Beds and Melton Limestone may correspond to a thin, lenticular, sparsely fossiliferous, glauconitic sand and ferruginous sandstone underlying the Melton Limestone along the Tickera-Myponie Point coast (Lindsay, 1970). Here the sand is up to 3 m thick, rests transgressively on Tickera Granite and contains a Late Oligocene to earliest Miocene foraminiferal microfauna.

Between Whyalla and Cowell, up to 20 m of fossiliferous glauconitic gravelly sand and sandy marl may unconformably overlie the Kanaka Beds and underlie Melton Limestone. In marginal situations, the unit becomes less fossiliferous, sandier and more gravelly laterally, and overlies basement rocks. The unit yields foraminiferal assemblages similar in age to those on the Tickera coast.

Melton Limestone

Nomenclature and stratigraphy of the Melton Limestone are outlined by Lindsay (1970), who distinguished four unnamed members on the basis of lithology and foraminiferal biostratigraphy. Including the underlying fossiliferous sand, the succession represents five Late Oligocene to early Middle Miocene transgressive intervals. The maximum exposed thickness is 8 m at Myponie Point, but up to 32 m were intersected in Tickera 1.

At the type section on northeastern Yorke Peninsula, only the younger units 4 and 5 are exposed. Unit 4 contains the large warm-water foraminifer *Lepidocyclina howchini*, indicating an early Middle Miocene age. At least 4 m of the unit are exposed in the area, the lower part consisting of quartzose, bryozoal, coarse calcarenite and calcirudite; the upper part is finer grained and densely recrystallised. Unit 5 is a recrystallised, quartzose calcarenitic limestone containing low frequencies of the large warm-water foraminifer *Flosculinella bontangensis* and *Marginopora vertebralis* whereas miliolid foraminifera, including

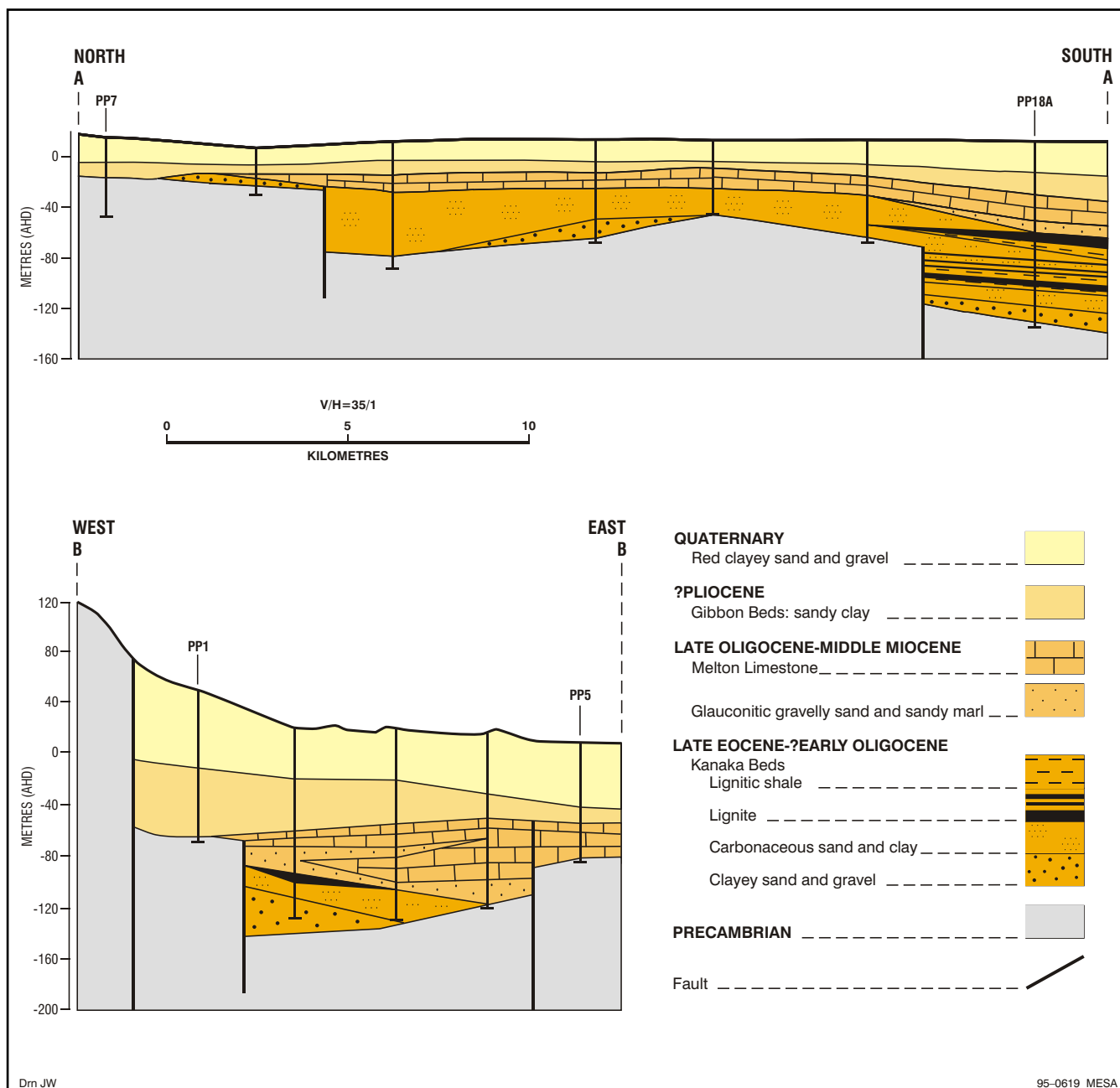
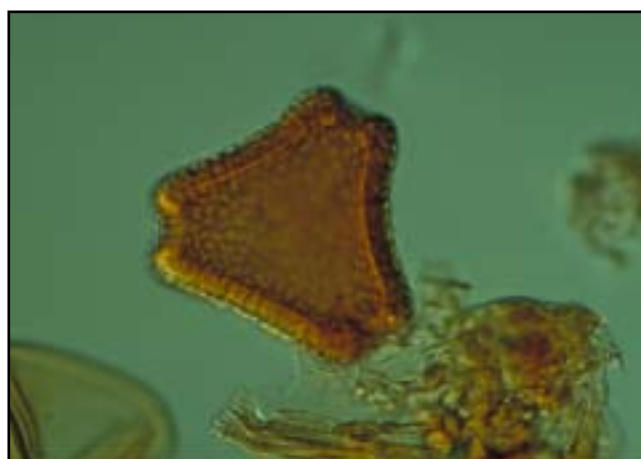


Fig. 10.20 Cross-sections along the western margin of the Pirie Basin (after Lemon, 1982). Lines of section are located on Figure 10.19.



Melton Limestone overlying Tickera Granite, Myponie Point, northern Yorke Peninsula. Photo 042567.



Triorites magnificus, a key zonal pollen for the Middle N. asperus Zone, from the western Pirie Basin. Width of the pollen is 60 micro μ m. Photo 042568.

Austrotrillina howchini, are common. This youngest and most widespread unit crops out intermittently as far north as Webling Point, and includes limestones near Wallaroo and Whyalla formerly thought to be Pliocene.

Unit 3 is a pink, quartzose, glauconitic recrystallised limestone containing miliolid foraminifera with lesser bryozoal and echinoid remains and brachiopods. At Point Hughes, the unit is at least 1.5 m thick and is overlain by unit 5. The foraminiferal assemblages are probably Early Miocene. Unit 2 is a gravelly quartzose, bryozoal, horizontally bedded to cross-bedded calcarenite, calcirudite and calcareous sandstone deposited in a near-shore, shallow marine environment. Southwest of Tickera, 3-5 m of the unit rest on Oligo-Miocene sand or basement. The foraminiferal evidence indicates a very early Miocene age (Lindsay, 1970).

Bioclastic limestone of Oligo-Miocene age (Crawford, 1965) at Point Turton may represent a southern extension of the Pirie Basin and connection with the St Vincent Basin, as does Melton Limestone on northeastern Yorke Peninsula.

Gibbon Beds

Mottled clayey sand, sandy clay and conglomerate 8-10 m thick, comprising the Gibbon Beds, overlie the Melton Limestone in sea cliffs, creek banks and drillholes. Although unfossiliferous, the beds are probably non-marine and of Pliocene-Pleistocene age (Parker, 1983b), and are correlated with the Uley Formation on southern Eyre Peninsula.

EUCLA BASIN AND PALAEODRAINAGE

M.C. Benbow, J.M. Lindsay and N.F. Alley

Structure and setting

The Eucla Basin includes Tertiary sediments deposited in marine and terrestrial settings in the southwestern part of the State (Figs 10.2, 10.21, 10.22). This is a much greater extension of the basin limits of Ludbrook (1958) and Lowry (1970) as shown by Benbow (1990a, in prep.). The Tertiary sediments of the onshore Poldia Basin and the Uley and Wanilla groundwater basins were probably once contiguous and are here included in the Eucla Basin. Mesozoic sediments formerly included in the basin are now regarded as part of the Mesozoic Bight and Duntroon Basins (see Ch. 9).

Substantial neritic carbonates and inner platform non-marine to marine terrigenous sediments were laid down on an adjacent shallow-water platform, a region of generally shallow basement (Fig. 10.23). The onshore limit of the marine Tertiary limestones largely coincides with the margin of the Bunda Plateau. The Tertiary sediments are relatively thin on the platform but thicken up to 1 500 m towards the margin and thin towards the continental rise. They are largely unaffected by the normal faults that characterise the structural style of the underlying Cretaceous rift sediments.

An extensive region of palaeodrainage fringing the Eucla Basin, which has preserved alluvial, colluvial and paralic sediments, is not included in the basin definition. The palaeochannels, although partly obscured by a mantle of Quaternary sediments, are remarkably intact. In South Australia, the palaeorivers drained the Musgrave Block, Stuart Range and Gawler Ranges.

A unique assemblage of Late Eocene aeolian landforms occurs along the inland margin of the Eucla Basin and includes the 650 km long Ooldea Range in the region of the basin extension (Figs 10.21, 10.23).

The platform underwent epeirogenic movement during the Tertiary, with the greatest uplift being in the northwest. The Tertiary sediments are flat lying to very gently dipping offshore and onshore, and overlie Mesozoic cover sediments and basement of the Gawler Craton and Coompana Block. There has been syndepositional subsidence towards the platform margin, where over 0.5 km of mostly carbonates has accumulated. A number of gentle 'sags' occur within the platform succession and epeirogenic movement

has given rise to some significant structures. In the central southern part of the emergent platform, Precambrian crystalline basement separates major regions of subsidence, the eastern region being coincident with the Permian Denman Basin (see Ch. 8).

Stratigraphy

The Tertiary succession is divided into the Eucla Group of marine limestones, and the Immarna Group of predominantly terrigenous sediments (Fig. 10.22). The Immarna Group commonly forms the base of the succession and extends around the northeastern margin, where it makes up the greater part of the section. The name Burdunga Subgroup is used for the Middle to Late Eocene part of the succession.

The base of the Tertiary succession along the southern margin is marked by a conspicuous and extensive seismic reflector (horizon 'A' of Fraser and Tilbury, 1979; Stagg *et al.*, 1990). An angular unconformity is marked by erosion of Late Cretaceous sediments and by the upper limit of significant syndepositional faulting. On the submerged platform to the north, the unconformity with the Late Cretaceous is difficult to interpret from seismic records and can be difficult to discern on lithological grounds. On the emergent platform the Immarna Group unconformably overlies Officer Basin and Bight Basin sediments and basement of the Gawler Craton.

Undifferentiated Paleocene-Early Eocene sediments

Sedimentation may have commenced in the Paleocene-Early Eocene with undifferentiated fluvial to marginal marine clastics of the Immarna Group along part of the southern basin (Figs 10.22, 10.23; Taylor, 1975; Barten, 1975; Davies *et al.*, 1989). In the far west, ?Paleocene-Early Eocene sediments form a prograding wedge (Bein and Taylor, 1981).

Small mounds are interpreted from seismic evidence within these early Tertiary terrigenous sediments and are associated with strong local reflectors in the Late Cretaceous section, which are interpreted as sills (Fraser and Tilbury, 1979).

Middle to Late Eocene sequence set

Sedimentation during the Middle to Late Eocene was widespread, leading to deposition of five major units: the Pidinga Formation, Hampton Sandstone, Ooldea Sand, Wilson Bluff Limestone and Toolinna Limestone (Fig. 10.22). Correlatives of the Pidinga Formation in the Poldia and Cummins-Uley-Wanilla groundwater basins are named the Poelpena and Wanilla Formations, respectively.

Pidinga Formation and equivalents

These terrigenous clastics, averaging 30-60 m in thickness, are generally carbonaceous and include poor quality lignite. Onshore they are mostly confined to topographically low settings such as palaeochannels and broader depressions marginal to the Bunda Plateau. The Poelpena Formation commonly exceeds 100 m in thickness in the eastern part of the Poldia Basin, suggesting syndepositional faulting, but is also widespread over central Eyre Peninsula (Fig. 10.24).

The Pidinga Formation overlies the Mesozoic Bight Basin, Palaeozoic Officer Basin sediments, and Precambrian basement. The underlying rocks northeast of the Bunda Plateau are commonly deeply weathered and the base of the Tertiary succession is generally well defined. Locally, however, the basal Pidinga Formation may not be easily distinguished from the underlying weathered rocks, due to reworking and/or weathering. The basal sediments in the palaeochannels are commonly coarse-grained sand and grit.

The Middle to Late Eocene terrigenous formations of the basin and palaeodrainage are a composite of transgressive and regressive sediments, which helps to explain rapid vertical and lateral lithological variations.



Fig. 10.21 Locality map of the Eucla Basin and related palaeodrainage in South Australia. Geological sections are shown on Figs 10.23-10.25.

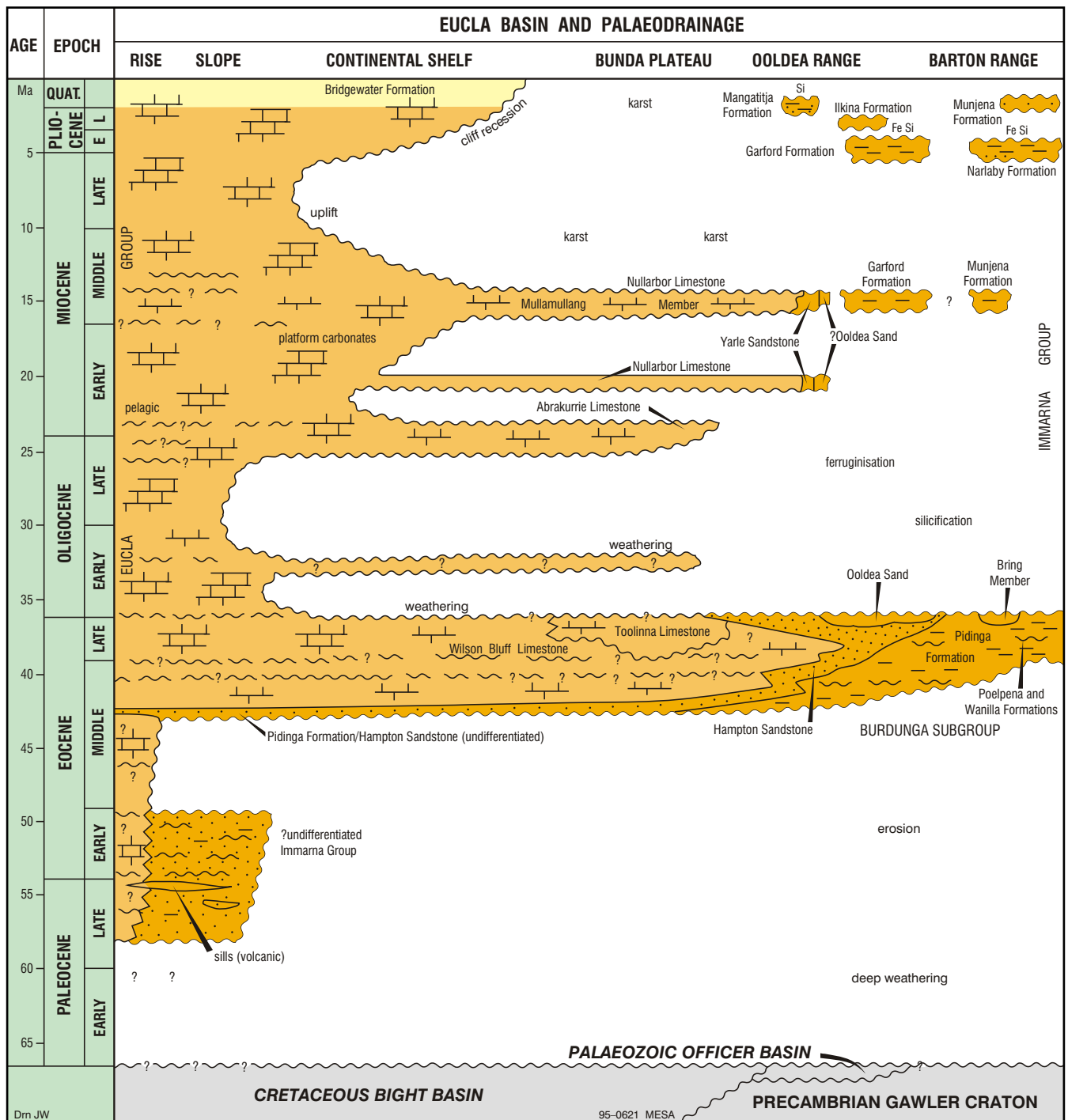


Fig. 10.22 Stratigraphic units of the Eucla Basin (Fe, ferruginisation; Si, silicification).

The oldest onshore part of the Pidinga Formation is correlated with the Middle Eocene Lower *N. asperus* Zone (Lindsay and Harris, 1975; McGowran, 1989; Alley and Beecroft, 1993). Much of the formation around the northeastern margin of the Eucla Basin contains palynofloras correlative with the Middle-Late Eocene Middle *N. asperus* Zone (Alley and Benbow, 1989; Alley and Beecroft, 1993). This youngest imprecisely dated phase includes the sponge-spicule-bearing Bring Member, deposited during the last Tertiary transgression to extend along the palaeochannels beyond the Bunda Plateau (Figs 10.24, 10.25).

The Poelpena and Wanilla Formations have a similar age range. Palynofloras from the type section of the Poelpena Formation in SADME Pold 1 are correlative with the Lower *N. asperus* Zone. Jurassic palynomorphs reworked from the Pold 1 Formation into the Poelpena Formation indicate that Jurassic sediments formed

the source for some of the Tertiary unit. Palynofloras in the type Wanilla Formation in Cummins School Bore between 58 and 116 m are probably correlative with the Middle Eocene Lower *N. asperus* Zone. The deepest sample contains a palynoflora more suggestive of the *Proteacidites asperopolis* Zone of Early Eocene age. Assemblages above 58 m are tentatively correlated with the Upper *N. asperus* Zone of Late Eocene-Early Oligocene age. Between 43 and 51 m in the nearby Cummins Police Station Bore, planktonic microfaunas suggest an age close to the Eocene-Oligocene boundary.

Sedimentary environments range from fluvial, shorewards through to estuary, delta and interdistributary bay. Facies in the Middle to Late Eocene paralic sediments north of the Barton Range include fringing-swamp carbonaceous mud containing palm tree remains and fragments of wood, and mixed mud and sand tidal-flat

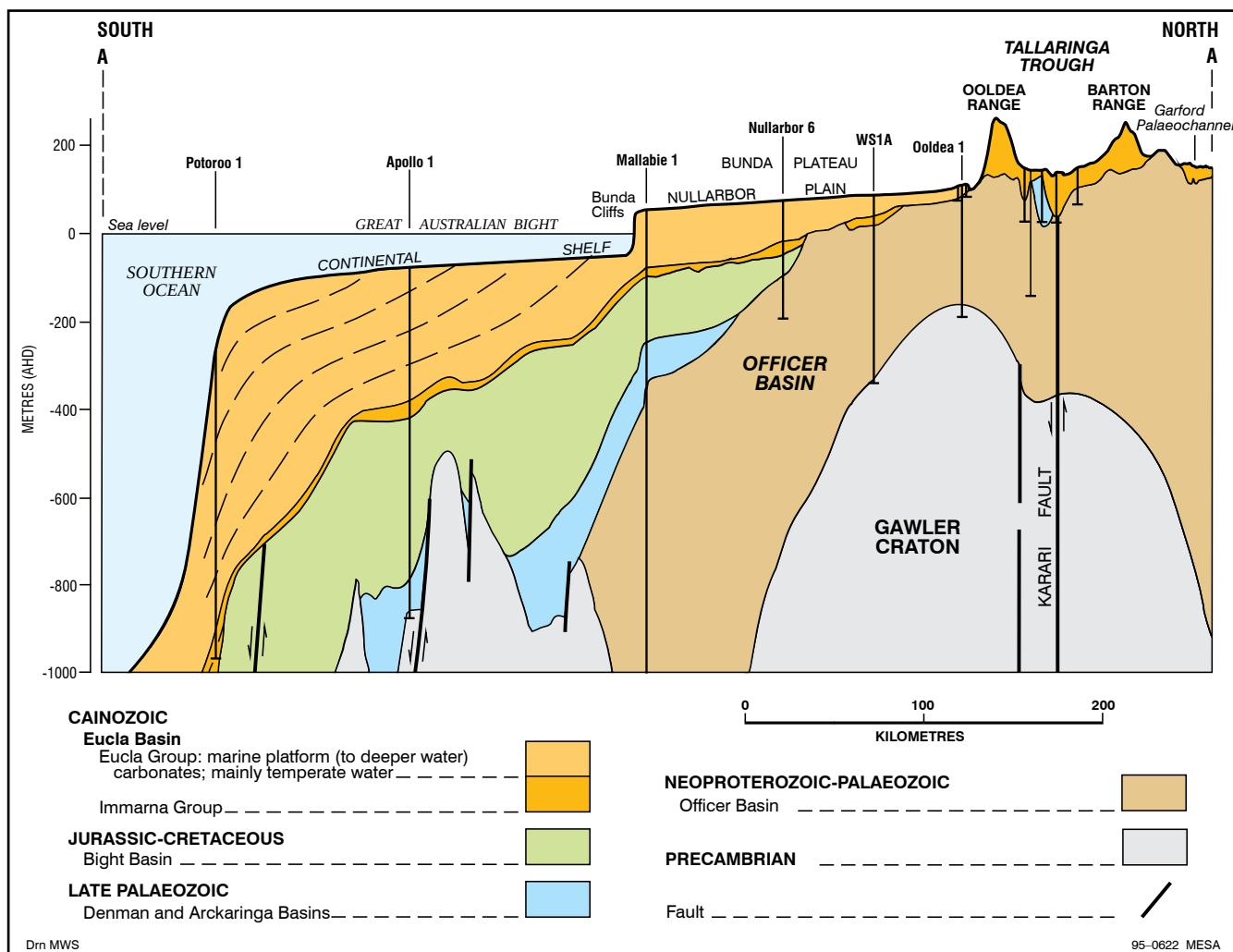


Fig. 10.23 North-south cross-section through Tertiary sediments in the eastern Eucla Basin. Line of section A-A is located on Figure 10.21.

sediments. Tidal influence is indicated by common reactivation surfaces, bipolar current directions in ripple-laminated mud and sand, and large-scale festoon cross-beds indicating up-slope palaeocurrents. Intraclastic horizons are common with some containing body fossils of sponges.

Widespread marine influence is found north of Mount Wedge adjacent to the Poldia Basin and in the Yaninee Palaeochannel. Spicule-bearing sands and black, glauconitic, sapropelic muds contain abundant marine dinoflagellates, spicules, foraminifera,

and bryozoal and echinoid fragments (Flint and Rankin, 1991). Spicule-bearing sediments also occur along the margin of the Gawler Ranges in the Yaninee Palaeochannel and are widespread in the Lake Maurice area north of the Ooldea Range (Figs 10.21, 10.24; Binks and Hooper, 1984; Rankin and Flint, 1991; Benbow, in prep.).

Early Cretaceous palynomorphs have been recycled into the Pidinga Formation. The source was probably Bulldog Shale in the upper reaches of the palaeochannels (Alley and Beecroft, 1993).

Hampton Sandstone

The lensoid to sheet-like marine, estuarine and fluvial Hampton Sandstone overlies the Pidinga Formation. Originally described from the central and western part of the Eucla Basin (Fairbridge, 1953; Lowry, 1970), these quartz-rich sands are widespread around the inner margin, particularly beyond the Bunda Plateau (Figs 10.22, 10.25; Benbow, 1990a). The formation is partly clayey at the base and glauconitic and fossiliferous at the top, where it is overlain by Wilson Bluff Limestone.

Over the central part of the Eucla Basin, the Hampton Sandstone is late Middle Eocene (foraminiferal Zone P12). However, a widespread younger, possibly Late Eocene phase beyond the Bunda Plateau is up to 50 m thick and in part bears siliceous sponge spicules where it overlies the Bring Member (Benbow, 1990a). Very thin Wilson Bluff Limestone is interbedded in the middle of the Hampton Sandstone beneath the Ooldea Range east of Ooldea.



Cross bedded, marine sponge spicule bearing carbonaceous sand of the Bring Member Pidinga Formation, Lake Anthony area. Width of view is 0.3 m.. Photo 039356.

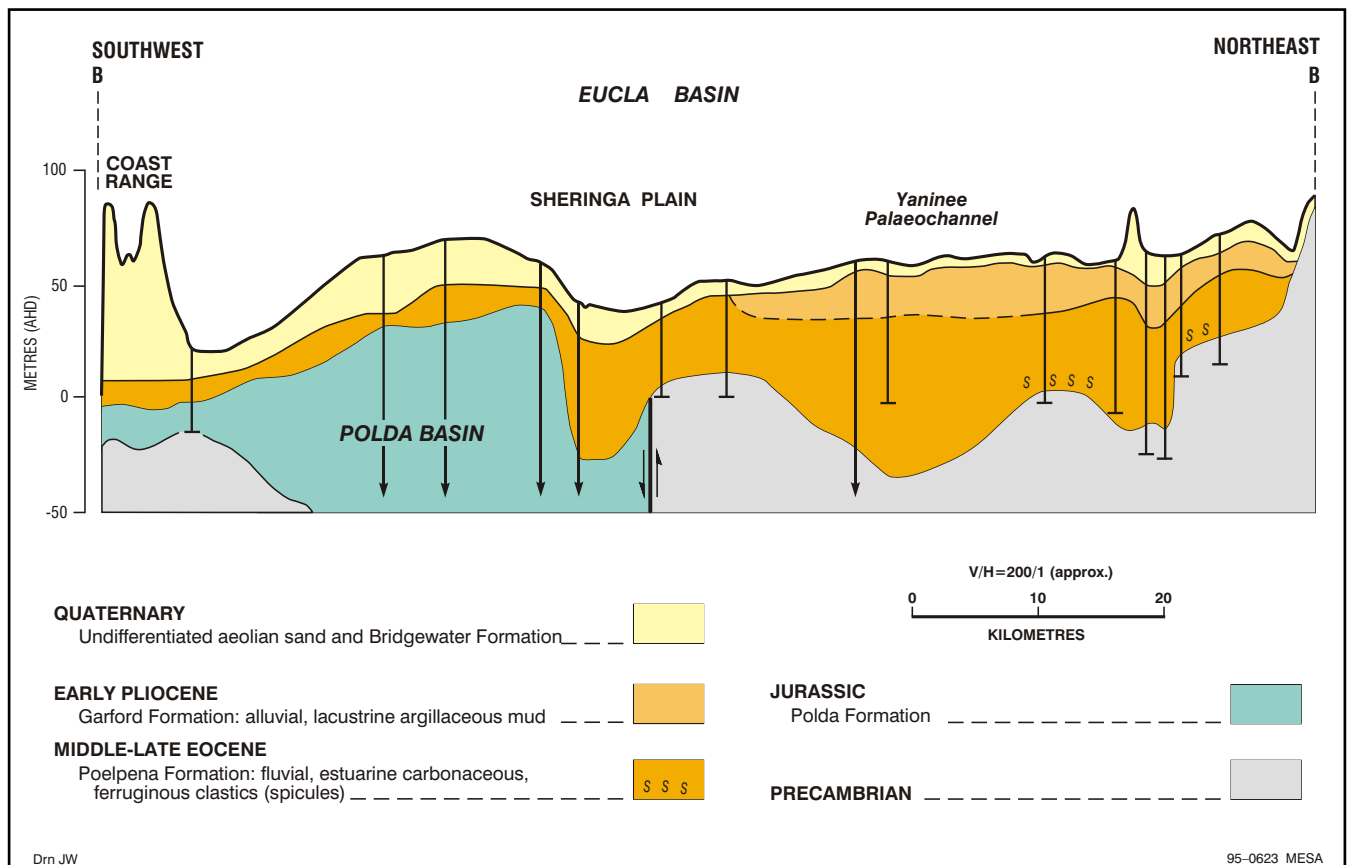


Fig. 10.24 Cross-section through Eucla Basin sediments of the southwestern Eyre Peninsula (after Benbow, 1993). Line of section B-B is located on Figure 10.21.

Ooldea Sand

The large dunes of the Ooldea and Barton Ranges, formed during regression probably in the Late Eocene (Figs 10.21-10.23), are composed of Ooldea Sand, a predominantly medium-grained quartz sand up to at least 112 m thick, conformably overlying Hampton Sandstone. The sand varies from clean to slightly clayey, although some horizons are clay-rich; sand grains are well rounded and some show aeolian frosting. Trace amounts of heavy minerals are present, particularly east of Ooldea.

Oligocene-Miocene equivalents are interpreted to be present as relatively low coastal dunes on the southern flank of the Ooldea Range (M.C. Benbow, MESA, pers. comm., 1994).

Wilson Bluff and Toolinna Limestones

Extensive deposition of biogenic platform carbonates in the Eucla Basin commenced during the Middle Eocene. Conformably overlying the initial transgressive deposits of the Pidinga Formation and Hampton Sandstone is the Wilson Bluff Limestone, which averages <150 m thickness east of the Coompana Block but increases westwards to ~300 m in Western Australia (Fig. 10.22). The limestone was deposited in temperate-water conditions during the Middle to Late Eocene marine transgressions, on a drowned carbonate platform or ramp.

The formation is mostly white to grey wackestone, skeletal mudstone (mud is micritic and microbioclastic) and rudstone with minor packstone. It is very thinly to thickly bedded, but without well-developed partings; beds have sharp or transitional boundaries and are defined by differences in grain size, texture or induration. At Head of Bight, the limestone displays discontinuous



Bunda Cliffs, Great Australian Bight. From the base: whitish cross-bedded and parallel-bedded Wilson Bluff Limestone; a thin dark brown band of ?Abakurrie Limestone; light brown parallel-bedded Nullarbor Limestone. Cliff height is ~50 m. Photo 042502.

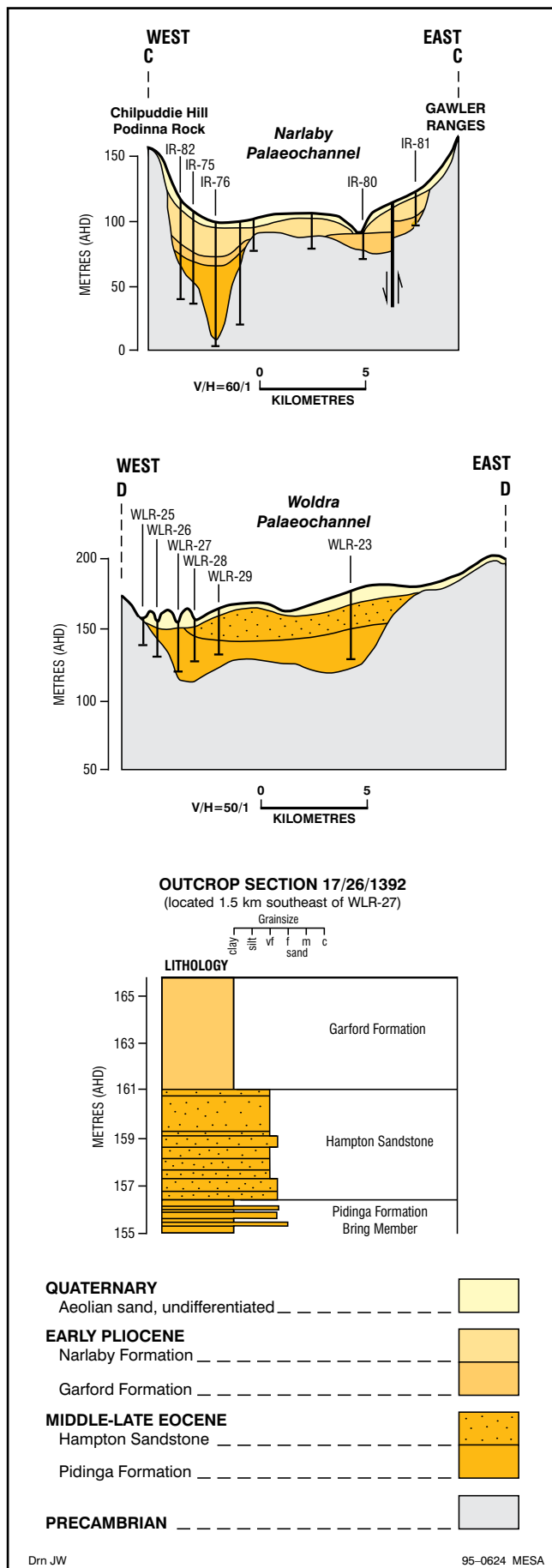


Fig. 10.25 Cross-sections through Tertiary sediments in the Narlaby Palaeochannel (C-C) and Woldra Palaeochannel (D-D) (after Benbow, 1993). Lines of section are located on Figure 10.21.

lamination and scour channels, possibly storm generated, which are infilled with coarser grained, less muddy sediment. The basal part of the formation beneath the Bunda Plateau is locally a sandy, glauconitic marl. This facies, deposited during the Wilson Bluff Transgression, was first recognised by Ludbrook (1960a) and contains a well-preserved Middle Eocene planktonic foraminiferal fauna correlated with upper Zone P12 (McGowran and Lindsay, 1969; McGowran, 1989). An upper, regressive quartzose facies with sponge spicules in the 'Colona'-Nundroo' area may be of similar age (Lindsay and Harris, 1975).

The Toolinna Limestone is a grainy bryozoal limestone which intertongues with and overlies Wilson Bluff Limestone in the southwestern part of the basin (Lowry, 1968, 1970). Similar facies occur within the Wilson Bluff Limestone, such as the grainy limestone exposed in the Bunda Cliffs southeast of 'Koonalda' where it contains large-scale cross-bedding, with tabular foresets up to 6 m thick. On the Bunda Plateau, the Wilson Bluff and Toolinna Limestones range in age from Middle to Late Eocene. In Outback Oil-Pickrell Mallabie 1, possibly all of the Wilson Bluff Limestone section is Middle Eocene (Lindsay and Harris, 1975).

Isolated occurrences of marginal Wilson Bluff Limestone deposited during the early Late Eocene (Zone P14/15) Tortachilla Transgression are found beyond the Bunda Plateau. Sediments of the transgression are carbonaceous in CRAE RCH 2 adjacent to the Paling Range where the limestone is intercalated with Pidinga Formation (Alley and Beecroft, 1993; Benbow, in prep.).

Very little is known of the outer platform margin and deeper basin limestones deposited during the Middle to Late Eocene. Samples of Cainozoic limestone dredged from the Ceduna Terrace in the southern part of the Eucla Basin are of oceanic bathyal facies, containing planktonic microfaunas (Davies *et al.*, 1989; Alley and Clarke, 1992). The most common lithology is fine-grained pelagic mud or chalk (mudstone to packstone). Some muds show little lithification and yield abundant, well-preserved foraminifera, while other chalks are more lithified. Skeletal grains are mostly foraminifera and sponge spicules; glauconite, and less commonly peloids, may be present. Silicified facies contain abundant siliceous sponge spicules and radiolarians deposited in slope and platform-margin environments where there appears to have been considerable reworking of microfaunas.

The unconformity between the Wilson Bluff Limestone and overlying Abrakurrie and Nullarbor Limestones on the Bunda Plateau is regarded as a complex, multi-generation contact (James and Bone, 1991). Part of the upper 2 m was indurated on the seafloor and during subaerial exposure, and extensively perforated by burrowing crustaceans. Infilling the tubes is a red to brown, bryozoan-echinoid floatstone, with gastropod and bivalve moulds and casts, together with phosphate-rich clasts of fine-grained fossil wackestone. This sediment is interpreted to be a condensed deep-water carbonate, deposited during the Early Oligocene high sea level.

Late Oligocene-Pliocene units

Abrakurrie Limestone

The Abrakurrie Limestone is the least extensive of the limestones on the emergent platform, and is up to 100 m thick in Western Australia where syndepositional subsidence is likely. In South Australia, however, the limestone is generally <10 m thick. A distinct yellow-brown band extending across the length of the Bunda Cliffs includes the Abrakurrie Limestone and grainy facies of the Wilson Bluff Limestone because both Late Eocene and Late Oligocene to earliest Miocene ages have been obtained from drillhole intersections (Lindsay and Harris, 1975).

The Abrakurrie Limestone consists predominantly of bryozoan-rich skeletal limestone ranging in texture from wackestone to rudstone deposited in cool to temperate water on a partly to completely drowned platform (James and Bone, 1991).

At Fishery Bay, southwest of Port Lincoln, an erosional remnant of brown and pink bryozoal sandy limestone, a few metres thick and gritty and pebbly near the base, may be an equivalent of the Abrakurrie Limestone. It contains abundant *Victoriella conoidea* in a microfauna considered to be upper Janjukian (latest Oligocene) in age. The unit can be traced south to Cape Wiles and has been recognised on Thistle Island 40 km to the east.

The disconformity between the Abrakurrie and Nullarbor Limestones probably represents much of the Early Miocene. In contrast to the heavily modified top of the Nullarbor Limestone, there is a less complex diagenetic zone at the top of the Abrakurrie, the upper part of which is lithified and weathered due to subaerial exposure.

Nullarbor Limestone

The Nullarbor Limestone is mostly a bioclastic and micritic limestone deposited over much of the emergent Eucla Basin (Figs 10.22, 10.23), with the Ooldea Range acting as a barrier, separating marine deposition on the platform from extensive lacustrine sedimentation inland (Benbow, 1986b). In contrast to the underlying carbonates, the Nullarbor Limestone is remarkably uniform in thickness, averaging 20–35 m with a maximum of 45 m. There was little or no local subsidence during deposition, which took place in part by vertical accumulation of subtidal limestone, mostly above fair-weather wave base with low to moderate wave and current energies. Nevertheless, energy levels were sufficient to turn algal nodules on the seafloor. In higher energy regimes, the formation is less muddy and better sorted, locally containing reworked pebbles. Bipolar cross-bedding exposed in the Cook railway quarry indicates tidally influenced shoals well onto the platform.

The Nullarbor Limestone exhibits lateral and vertical facies changes. The lower part of the formation is dominantly algal, bryozoal and echinoidal, whereas the upper part has abundant miliolid foraminifera and common agglutinated and encrusting forms. This upper part of the formation is locally rich in corals (e.g. Watson Quarry), although no reefs have been observed.

An unnamed transgressive lag, 0.1–0.35 m thick and comprising poorly to better sorted clay, sand and grit, was deposited at the base of the Nullarbor Limestone in some parts of the basin. Carbonate mudstone, suggesting low-energy, shallow-water deposition, perhaps in intertidal to supratidal environments, marks the top and most proximal part of the formation north of Cook and local base in the Ifould Lake area. Micritic and texturally variable sediments in the Ifould Lake–Nundroo area may indicate the presence of current-baffling meadows such as seagrass. At several localities on the basin margin, extending from Lake Tallacootra to ‘Charra’ near Ceduna, such sediments contain the large benthonic, warm-water foraminifer *Lepidocyclus* (Benbow and Lindsay, 1988).

The Mullamullang Member (Lowry, 1970) was deposited across much of the southern emergent part of the platform. The lithology generally comprises nodules of calcareous algae, including the warm-water genus *Archaeolithothamnion*, in a wackestone to packstone matrix. The nodules vary in size (<0.1 m) and form, and contain a major bryozoal component. Similar facies occur at different stratigraphic positions in the region of Ifould Lake.

The occurrence of the *Lepidocyclus* Zone near the local base of the Nullarbor Limestone suggests an early Middle Miocene age (Benbow and Lindsay, 1988). However, a thin muddy to mud-free bryozoal limestone in the subsurface 7 km west of Seven Mile Swamps yields *Sherbornina cuneimarginata* and *Astrononion centropax*, foraminiferal species of Longfordian age whose ranges in southern Australia do not extend up to the *Lepidocyclus* Zone or into the Middle Miocene (Lindsay, 1987a). This limestone is petrologically similar to the *Lepidocyclus*-bearing Nullarbor Limestone nearby and, for the present, is included in that formation. It may represent a short-lived transgression which preceded the main Nullarbor Limestone transgression.

The Nullarbor Limestone contains a negligible terrigenous clastic component, except at the base and in near-shore settings, indicating



Algal limestone of the Middle Miocene Nullarbor Limestone showing contrasting form and size of algal nodules. Clockwise from right, the rock slabs are from Madura, Head of Bight (two samples) and Ifould Lake. Photo 042503.

that river systems carried little debris to the coast. It was deposited in warm temperate water in a shallow platform setting.

Yarle Sandstone

Around the eastern limit of the Nullarbor Limestone there is a narrow coastal terrigenous clastic unit named the Yarle Sandstone (Fig. 10.22; Benbow, 1990b). At Yarle Lakes, this silicified and ferruginous sandstone contains rare clasts of syndepositionally deformed Nullarbor Limestone and fossils. Similar sandstone conformably overlies Longfordian Nullarbor Limestone west of Seven Mile Swamps, where it is 10 m thick. Other fossiliferous calcareous quartz sandstones along the eastern margin of the platform at Chundie Swamps and Kidnippy Well may be Yarle Sandstone (Lindsay and Harris, 1975). There is no evidence along the eastern and northeastern edge of the Eucla Basin for widespread terrigenous Early-Middle Miocene sediments such as the Colville Formation delineated in the northwest (Lowry, 1968, 1970).

Unnamed late Tertiary offshore platform limestones

Up to 600 m of grainy limestone is widespread over the seaward margin of the Eucla Basin, such as over the Eyre Sub-basin (Bein and Taylor, 1981; James and von der Borch, 1991). The limestone contains large-scale prograding clinoforms indicating deposition on an energetic open platform during lower sea level. High levels of shelf carbonate production and/or significant erosion of the inner platform carbonate are implied. A Late Miocene age has been obtained from grainy limestone in the upper part of Platypus 1 on the southeastern margin of the basin south of Eyre Peninsula (Apthorpe, 1972).

Garford and Narlaby Formations

The Garford Formation comprises extensive lacustrine carbonate, 10-20 m thick, between the Ooldea and Barton Ranges and in the Wilkinson Lakes region. These sediments include stromatolitic, oncolitic and oolitic carbonate and argillaceous mudstone. Part of the formation possibly accumulated landward of the Ooldea Range during deposition of the Nullarbor Limestone, implying an early Middle Miocene age (Figs 10.22, 10.25). The carbonate is similar to Etadunna Formation of the Lake Eyre Basin which may range in age from latest Oligocene to Early Pliocene (see 'Lake Eyre Basin').

In the Poldia Basin area there are possible lateral equivalents disconformably overlying the Poelpena Formation (Fig. 10.24). They differ, however, in that carbonate lithologies are either sparse or absent. Palynological data indicate a Miocene to Pliocene age (Harris, 1973; Truswell and Harris, 1982). On central Eyre Peninsula, the Garford Formation contains rare marine microplankton while further westwards the frequency is higher, indicating probable estuarine conditions. Sediments up to 20 m thick exposed along the coast near Port Kenny include black carbonaceous grit and sand, clean sand and green to grey clay, and unconformably overlie weathered basement.

Uplift and southeastward tilting of the Eucla Basin sediments and regression followed in the Middle-Late Miocene. The Bunda Plateau marks that part of the carbonate platform that has not undergone subsequent widespread inundation by the sea.

Deposition of a possibly younger phase of the Garford Formation resumed in the palaeochannels around the basin, such as in the Tarcoola region and northern Eyre Peninsula. This part of the succession is up to 60 m thick and dominated by grey, green and cream to black mud, minor carbonate and local basal and interbedded carbonaceous sand and grit. These sediments were deposited in flood-plain, lacustrine, lagoonal swamp and locally evaporative environments. Near the Gawler Ranges, the Garford Formation becomes sandy and locally contains clasts of acid volcanic rocks. Likely equivalent sediments form part of the large fan in the Serpentine Lakes Palaeochannel on the Bunda Plateau (M.C. Benbow, MESA, pers. comm., 1994).

Palynological examination of carbonaceous sediments from the Narlaby and Kingoonya Palaeochannels suggests an Early Pliocene age for the younger phase of the Garford Formation (Harris, 1979b; Truswell and Harris, 1982; N.F. Alley, MESA, pers. comm., 1994). The presence of rare microplankton implies a very weak marine influence in the Narlaby Palaeochannel. Palynofloras from this part of the Garford Formation are very similar to Early Pliocene assemblages from palaeochannels on the western margin of the basin (Bint, 1981), indicating that the phase of sedimentation was widespread.

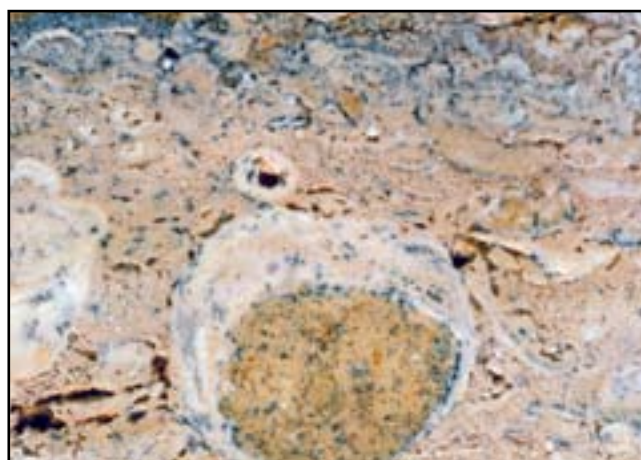
Overlying the Garford Formation in the Narlaby Palaeochannel is the Narlaby Formation, a fluvial to (?)estuarine fine to medium-grained, moderately to well-sorted sand with little clay, up to 60 m thick (Fig. 10.25; Benbow, in prep.). The sediments are partly silicified, iron-stained and varicoloured.

The Nullarbor Plain was probably incised locally during deposition of the Garford and Narlaby Formations. The Serpentine Lakes Palaeochannel continues southwards across the plain in the west whilst, south of Maralinga, LANDSAT imagery suggests outlines of former rivers 50 to 100 km in length. It is likely that this was a time of significant karst development in Eucla Group limestones. Blue and green-grey clay, similar to that of the Garford Formation, infills numerous interconnected channels and cavities in the Nullarbor Limestone at Haig Cave (Western Australia) and was probably carried in by river systems originating around the Eucla Basin. The limestone was extensively eroded or not deposited at the base of the large fan of the Serpentine Lakes Palaeochannel.

Munjena Formation

Poorly sorted conglomeratic sand, clayey sand and breccia of the Munjena Formation, up to 3 m thick, blanket wide areas of interfluvial and palaeochannels around the margin of the Eucla Basin (Benbow, 1986a, 1993). Where the unit overlies pre-Tertiary rocks it commonly contains rounded quartz pebbles, which may be scattered throughout or form discrete, flat-lying horizons. In the palaeochannels the sediments are not as pebbly and unconformably overlie the Hampton Sandstone. The pebble-bearing horizons were derived from lags reworked from underlying Palaeozoic and Mesozoic sediments. The formation has been silicified to pedogenic silcrete; rare polished and rounded silcrete clasts indicate reworking of older silcretes.

The formation is believed to be mostly Oligocene or younger, since it overlies Hampton Sandstone of the palaeochannels. Similar sediments that cap Aptian mudstone northwest of the Eucla Basin in the Gibson Desert in Western Australia were believed to have been deposited in the Late Paleocene to Early Eocene (Jackson and Van de Graaff, 1981). In the Billa Kalina Basin, presumed Munjena Formation equivalents have been suggested to be early Tertiary (Cowley and Martin, 1991).



Shallow-water lacustrine carbonate of the Miocene-Pliocene Garford Formation bearing ooids, algal nodules and nodule fragments. Width of view is 65 mm. Photo 042504.

Silicified Munjena Formation overlying bleached and ferruginised Hampton Sandstone. Photo 042505.

Pliocene units and landforms

Unnamed Late Pliocene onshore marine unit

A thin, unnamed sandy and pebbly limestone at Fishery Bay has casts and moulds of a molluscan fauna similar to that of the Dry Creek Sand and Hallett Cove Sandstone, and may indicate a Late Pliocene marine transgression.

Ilkina Formation

Deposition of the Ilkina Formation occurred in the palaeochannels during the Late Pliocene or Early Pleistocene. The formation is a conspicuously laminated to thinly bedded clay, silt and sand up to 5 m thick in the Narlaby Palaeochannel, and also forms the floors of playas such as Ifould Lake where it contains clasts of silcrete and ferricrete. The upper part of the formation is commonly gypseous and capped by a gypsum crust, suggesting that these sediments were laid down largely in saline playa lakes during the final stages of desiccation.

Mangatitja Formation

This formation (formerly Mangatitja Limestone of Major, 1973c) is up to 30 m thick and was deposited in palaeochannels adjacent to the Musgrave Ranges, including the fan of the Serpentine Lakes Palaeochannel on the margin of the Nullarbor Plain where it unconformably overlies probable Garford Formation.

The extensive limestone facies described by Major (1973c) contains oogonia of Characeae, ostracodes and gastropods. Major alluvial redbed facies occur, including sand, grit and flood-plain to fan clay (M.C. Benbow, MESA, pers. comm., 1994). Although the age of the formation is uncertain, it may be Plio-Pleistocene.



Ilkina Formation in the Narlaby Palaeochannel type section. Laminated mud is overlain by non-bedded mud and laminated gypsum. Photo 042506.

Karst of the Nullarbor Plain

The Nullarbor Limestone is well known for its caves, including some of the world's largest (Lowry and Jennings, 1974). A wide range of products of reworking and weathering, each characterised by its own signature, is developed in the upper part of the limestone. There is a complex succession of karst breccias, ferricrete, silcrete, calcite and halite speleothems, calcrete and gypcrete, all providing a valuable history of climatic change for this part of Australia since the early Middle Miocene. These products of carbonate diagenesis are largely undated. A minimum age of 400 000 years, beyond the range of ^{230}Th - ^{234}U dating, has been determined for at least three distinct phases of calcite speleothem formation in the western Eucla Basin (Goede, 1988). Calcite speleothems developed during a major episode in which karst breccia with red-brown matrix was formed. These speleothems may have formed contemporaneously with deposition of Mangatitja Formation in the palaeochannels.

Economic geology

Middle to Late Eocene carbonaceous sediments have been explored extensively for coal and uranium. Significant quantities of poor quality coal in the Pidinga Formation have been delineated in the Chundie Swamps, Tallacootra, Pidinga and 'Burdunga' areas, the thickest being in Chundie Swamps (Kwitko, 1983). Other smaller deposits occur in the Wilkinson Lakes, Garford Palaeochannel, Lake Anthony and Lake Bring areas. Although the lignitic sediments are generally thick and widespread, coal seams with <45% ash on a dry basis are only 0.5-2 m thick (rarely exceeding 5 m). The Poelpena Formation in the Polda Basin contains a 5-15 m thick lignite seam at depths of 200-220 m (Scott, 1984; Flint and Rankin, 1991).

The Ooldea Sand and Hampton Sandstone of the coastal aeolian landforms are potential targets for heavy-mineral exploration (Benbow, 1990b). Up to several percent of silt to fine sand-sized minerals, including rutile, ilmenite, zircon and tourmaline, occur in sand east of Ooldea on the southwestern margin of the Ooldea Range.

Exploration for sediment-hosted uranium has been carried out, particularly in the Narlaby Palaeochannel region southwest of the Gawler Ranges (Binks and Hooper, 1984; Flint and Rankin, 1991). Anomalous radioactivity is associated with carbonaceous horizons. Surface anomalies associated with saline lacustrine sediments are attributed to radium and/or radon in the groundwater, and/or minor uranium within the saline sediments. In the Narlaby Palaeochannel, mineralisation is associated with redox fronts near the contact of the carbonaceous, variably oxidised Pidinga Formation and overlying oxidised sand of the Pliocene Garford and Narlaby Formations. Mineralisation appears greatest in areas where the Pidinga Formation occurs as low erosional remnants surrounded by Garford Formation (Flint and Rankin, 1991). Highest values over 1 m intervals at the Yarranna Prospect are 3 550 and 1 275 ppm U^{308} .



Murrawijine No.1 Cave developed in Nullarbor Limestone, Nullarbor Plain. Photo 042507.

INTERIOR NON-MARINE BASINS

TORRENS BASIN

N.F. Alley and M.C. Benbow

Structure and setting

The Torrens Basin is a north-south-trending structural depression coincident with the Torrens Hinge Zone, with the greatest accumulation of sediments along the flank of the Flinders and Willouran Ranges (Figs 10.26, 10.27). Refraction seismic studies and drilling indicate locally complex structure and fault-controlled horsts and troughs along the eastern basin margin (ETSA, 1989). The western basin margin is defined by the Torrens Hinge Zone. Subsidence in the basin occurred at least as early as the Jurassic, producing a link with the Eromanga Basin to the north. Erosion and subsidence have obscured any connection that may have existed with Tertiary sediments of the Lake Eyre and Billa Kalina Basins. Although a narrow corridor containing Cainozoic sediments north of Port Augusta appears to form a link with the Pirie Basin, there is no evidence for marine transgression entering the Torrens Basin from the south.

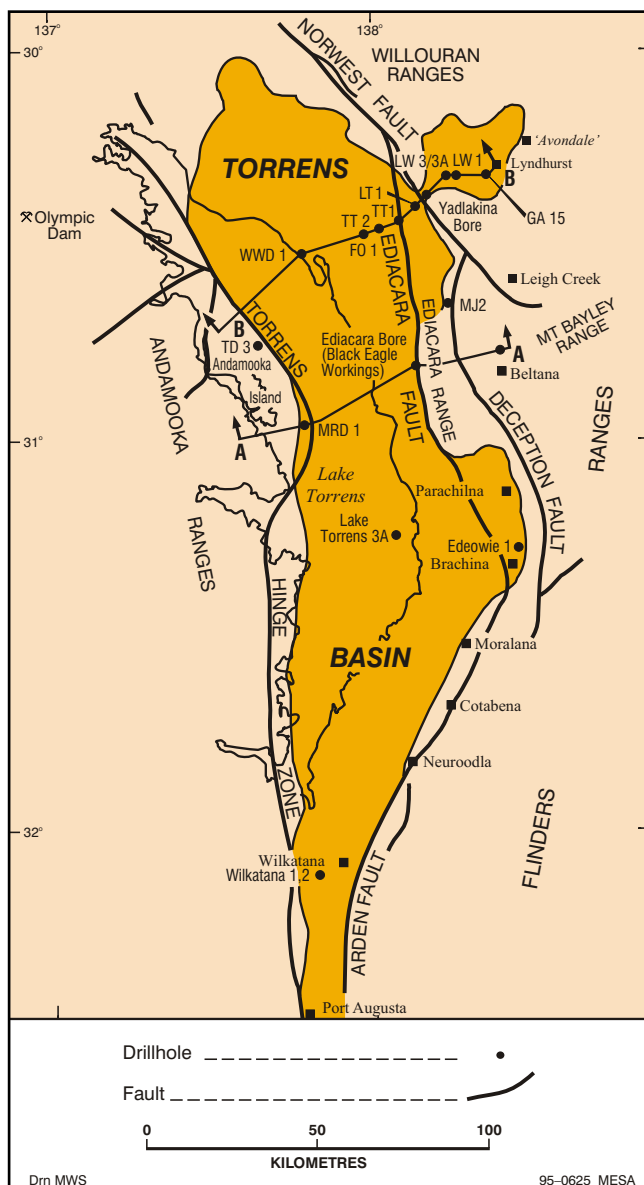


Fig. 10.26 Locality map and structural elements of the Torrens Basin. Sections are shown on Figure 10.27.

Stratigraphy and age

Cotabena Formation

Harris (1970b) referred informally to an early Tertiary unit, the 'Wilkatana Formation', from a type section between depths of 90 and 140 m in Santos Wilkatana 1, with a reference section between depths of 84 and 143 m in Wilkatana 2. This name, however, is invalid since it has been used elsewhere and the rock unit is herein named Cotabena Formation, with the type section between 173.4 and 270.35 m in Lake Torrens 3A, and reference sections as above in Wilkatana 1 and 2. The Cotabena Formation comprises fluviolacustrine, partly carbonaceous, fine to coarse-grained sand, silt, clay and lignite. Finely laminated silt and carbonaceous mudstone between 173.4 and 185 m in Lake Torrens 3A are included in the formation. White clay horizons near the base of the succession in this well indicate intraformational exposure and weathering. The formation overlies Precambrian and Cambrian rocks and, in deeper parts of the basin adjacent to the Flinders Ranges, sediments of Mesozoic age; it is absent in the shallower western and northern parts of the basin. Thin, partly conglomeratic silicified sand that is widespread at a higher elevation in the northeastern part of the basin may also be Cotabena Formation, as may be other sediments described as Eyre Formation at a similar elevation within the ranges (Coats, 1973b).

Palynofloras recovered from various drillholes indicate that the Cotabena Formation ranges in age from at least Early Eocene (*M. diversus* Zone) to late Late Eocene (Upper *N. asperus* Zone; Harris, 1964, 1970b, 1972; Alley and Sluiter, in prep.).

Neuroodla Formation

Overlying the Cotabena Formation is a green, grey to black argillaceous and white calcareous mudstone correlated with the Etadunna Formation of the Lake Eyre Basin (Johns, 1968b; Johns *et al.*, 1981). This succession is here named the Neuroodla Formation with a type section between 30.2 and 173.4 m in Lake Torrens 3A. The upper boundary of the formation cannot be determined because of weathering and erosion. The Neuroodla Formation is commonly ~100 m thick although ~250 m may be present in SADME Edeowie 1 (Cockshell, 1983). It is thickest along the western flank of the Flinders Ranges and similar sediments extend eastwards into the ranges, where the upper part may be equivalent to the Avondale Clay. The southern extent of the formation along the piedmont zone of the ranges is unknown.

Carbonaceous mud is locally present in lower parts of the Neuroodla Formation and the basal 10 m of Lake Torrens 3A are rich in very small shell fragments and ostracodes. The fauna includes mainly impoverished planorbid gastropods, one hydrobiid gastropod shell and a fish otolith, indicating a freshwater environment. In the central part of the basin, the lower clay beds contain montmorillonite, kaolin, illite and palygorskite (Johns, 1968b). The mudstone is dolomicrite and the associated argillaceous clay is predominantly palygorskite and halloysite, the latter being the most common in the upper weathered part. In Lake Torrens 3A, the mudstone grades up into red-brown silty to fine-sand-bearing argillaceous sediments between 30.2 and 79.5 m. Since this transition suggests a conformable relationship, the latter sediments are included in the Neuroodla Formation. Quartz gravel and pebble layers are locally major components near the ranges. Beneath Lake Torrens, the formation is overlain sharply and probably unconformably by undated gypsum and halite-bearing sediments. Eastwards toward the ranges the formation is overlain by Quaternary aeolian sand and conglomeratic alluvial sediments.

Palynofloras from the base of the sequence in WMC WWD 1 on the eastern margin of Lake Torrens are correlated with the Early Oligocene to Early Miocene *P. tuberculatus* Zone (Harris, 1986). The age of the top of the formation is, however, unknown.

Fine-grained sediment dominated by argillaceous mud may have been deposited in low-energy flood-plain environments and localised sand in channels, whereas the calcareous mudstones indicate deposition in evaporative playa lakes. The increase in

detrital quartz towards the eastern margin of the basin indicates that the Flinders Ranges were in existence during deposition.

The difference in elevation between the succession in Lake Torrens and correlative sediments within the Flinders Ranges implies maximum relative post-depositional uplift of the ranges in the order of 100 to 250 m. Greatest subsidence appears to have occurred in the southern portion of Lake Torrens. The distribution of silcretes along the northern and northeastern margins of the basin indicates no relative movement along the Torrens and Norwest Faults since silicification. Binks (1972) interpreted uplift of the Ediacara Range to have taken place in the Quaternary.

LAKE EYRE BASIN

R.A. Callen, N.F. Alley and D.R. Greenwood

There is very little record of sedimentation over most of inland South Australia between the Cenomanian and Late Paleocene, when weathering and erosion appear to have prevailed. During the Late Paleocene, tectonic subsidence in northeastern South Australia produced a large, shallow basin, here referred to as the Lake Eyre Basin, in which episodic fluvial and lacustrine sedimentation has taken place until the present (Figs. 10.28, 10.29). Although only the Tertiary sediments are discussed in this section, deposition of

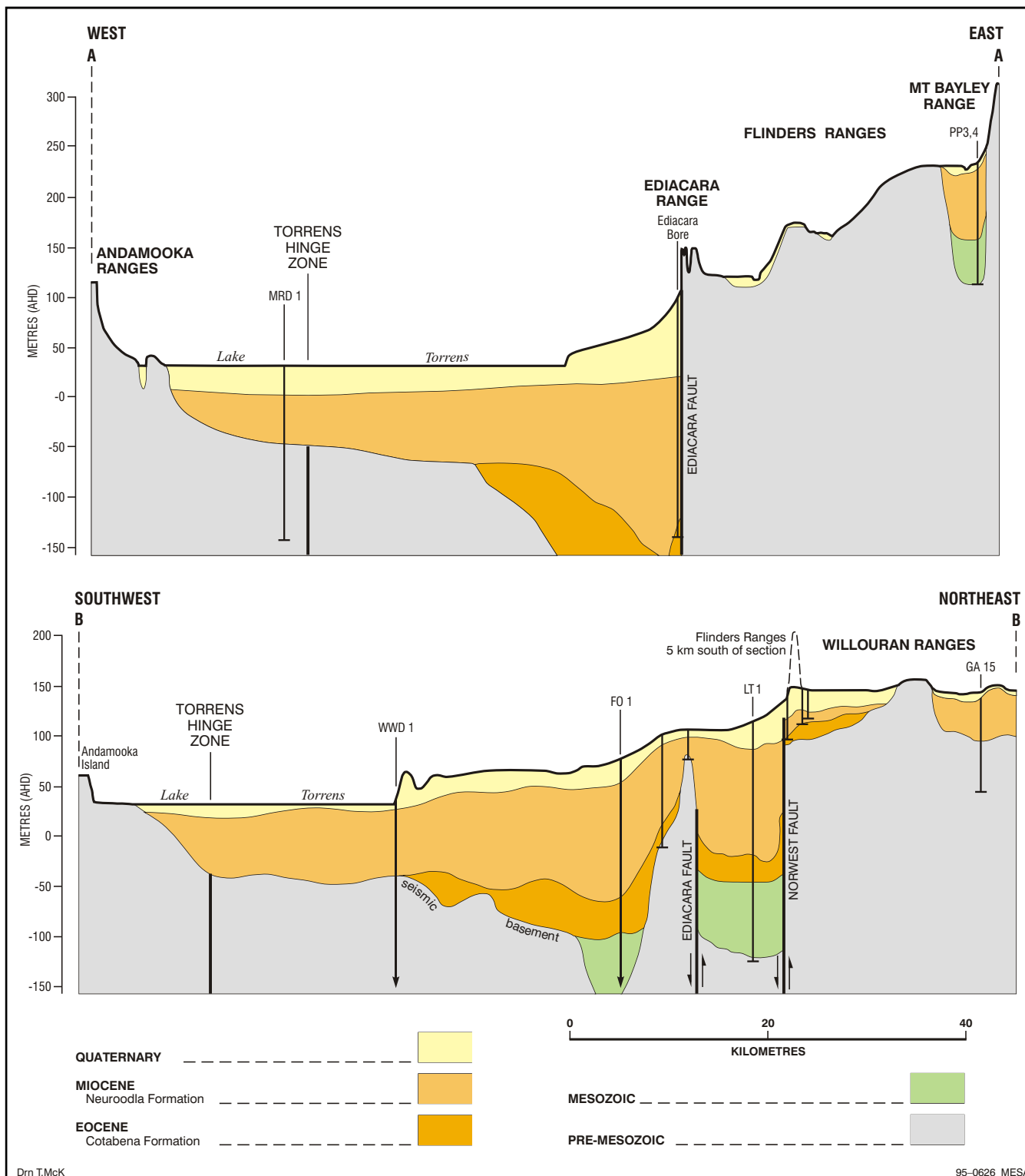


Fig. 10.27 Cross-sections through Tertiary sediments in the Torrens Basin. Lines of section are located on Figure 10.26. The profile of B-B is also shown on Figure 10.33 to demonstrate the height relationship between the Torrens and Billa Kalina Basins.

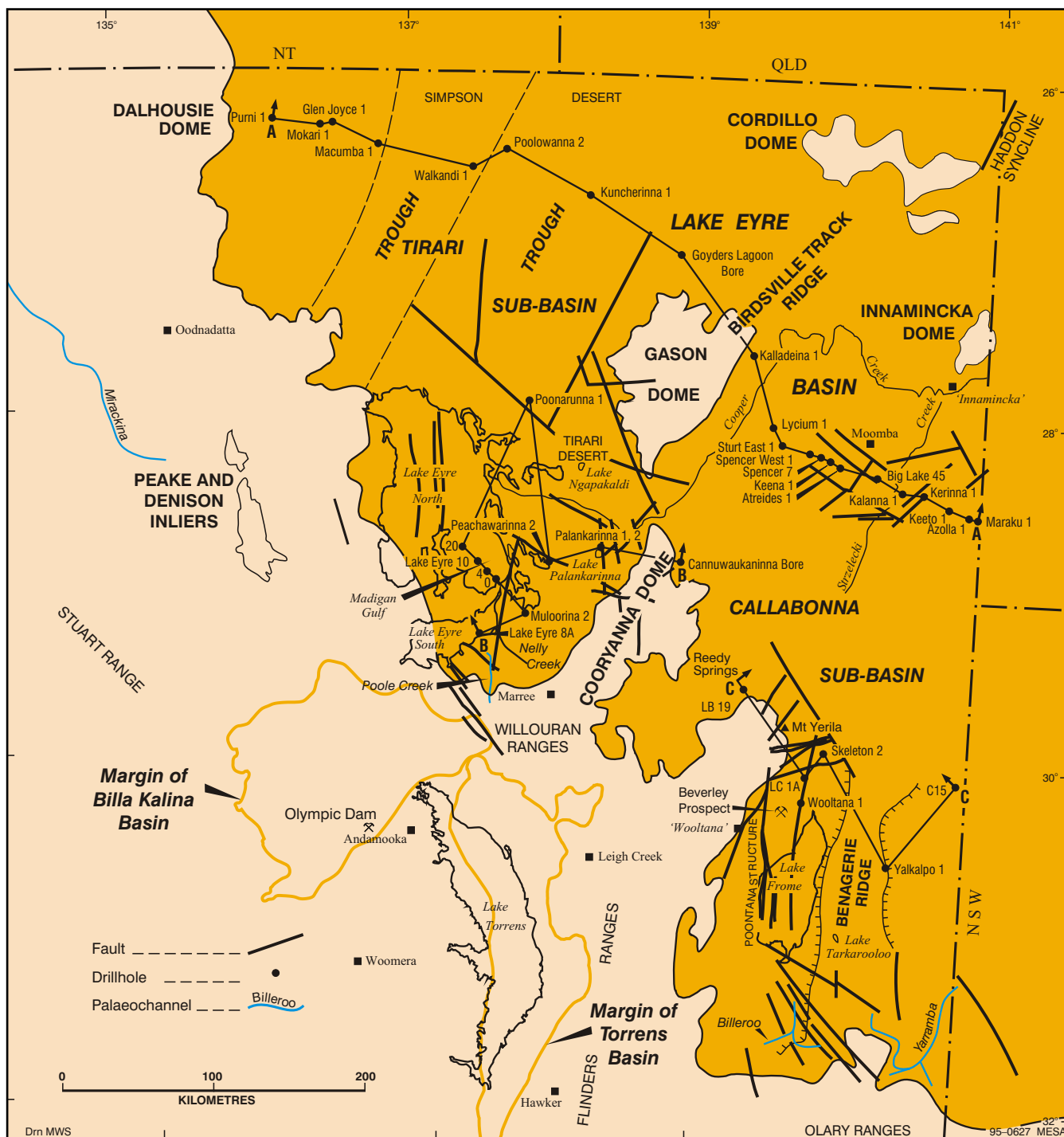


Fig. 10.28 Locality map and structural elements of the Lake Eyre Basin. Sections are shown on Figures 10.30 and 10.31.

some units commenced in the late Tertiary and continued into the Quaternary.

Structure and setting

Like the underlying Eromanga Basin, the Lake Eyre Basin consists of a thick sedimentary succession in a northeastern depocentre and a southwestern part where the sediments are significantly thinner and discontinuous. The basin is bordered on the west by the Peake and Denison Inliers and western margin of the Dalhousie Dome, in the south by the Wilhouran, Flinders and Olary Ranges, and in the east by the Barrier Ranges. In the north the basin extends into the Northern Territory, Queensland and New South Wales, and is bordered by domes of Mesozoic rocks. The southern margin is sharply defined and faulted as a result of late

Cainozoic uplift.

The basin is separated into the Tirari Sub-basin (Krieg *et al.*, 1991) in the west and Callabonna Sub-basin in the east by the Birdsville Track Ridge, expressed in the Tertiary as a series of faulted, duricrust-capped domes (Figs 10.28, 10.30). The stratigraphy of the Tirari Sub-basin (previously named the 'Lake Eyre Basin'; Callen and Tedford, 1976; Callen, 1990) is illustrated on Figure 10.31. Stratigraphic units used in the Callabonna Sub-basin (previously known as Tarkarooloo Basin and then Callabonna Basin; Callen and Tedford, 1976; Callen, 1981, 1990) are also shown on Figure 10.31.

Much of the Lake Eyre Basin is close to sea level, with subdued topography and poor outcrop in low cliffs, around salt lakes, or as discontinuous silicified channel remnants between dunes. Tertiary

AGE		LAKE EYRE BASIN				BILLA KALINA BASIN	ADJACENT INTERSTATE TERMINOLOGY					
		TIRARI SUB-BASIN (South Australia)	CALLABONNA SUB-BASIN		WAITE BASIN (Northern Territory)		(Western Queensland)					
			(Northern)	(Southern)								
QUATERNARY		Kutjirata Formation		undifferentiated								
TERTIARY	PLIOCENE	LATE	Tirari Formation		silcrete, ferricrete		silcrete, ferricrete		'laterite'		'laterite'	
		EARLY	silcrete, ferricrete		Cadelga Limestone & Doonbara Formation		silcrete, ferricrete		Willalinchina Ss		Austral Downs L/S Noranside L/S Horse Creek Fm	
	MIOCENE	LATE	Mampuworlu Sand, Wipajiri Formation Etadunna Fm (upper)				Namba Formation		Willalinchina Ss		ferruginous gravels	
		MIDDLE	Yardinna Claystone		Namba Formation		?		Millers Creek Dolomite Member		Waite Formation	
		EARLY	Alberga Limestone				Namba Formation		Billa Kalina Clay Member		Etadunna Formation	
		EARLY	Etadunna Formation (lower)				Namba Formation		Danae Conglomerate Member		Springvale Formation	
	OLIGOCENE	LATE	?Mirackina Conglomerate		silcrete		silcrete		silcrete		Arcoota Beds	
		EARLY	Muloorina Member									
		LATE	?Mirackina Conglomerate									
	PALEOCENE	MIDDLE	Mount Sarah Sandstone		Eyre Formation		Eyre Formation (Murnpeowie Formation)		?Willalinchina Sandstone		Glendower Formation	
		EARLY	Eyre Formation (Macintia Sandstone)								Marion Formation	
		LATE					unnamed sandstone peripheral to Billa Kalina Basin					
		EARLY	Dm JW								95-0628 MESA	

Fig. 10.29 Stratigraphic units in the Lake Eyre Basin, and correlatives in adjacent areas.

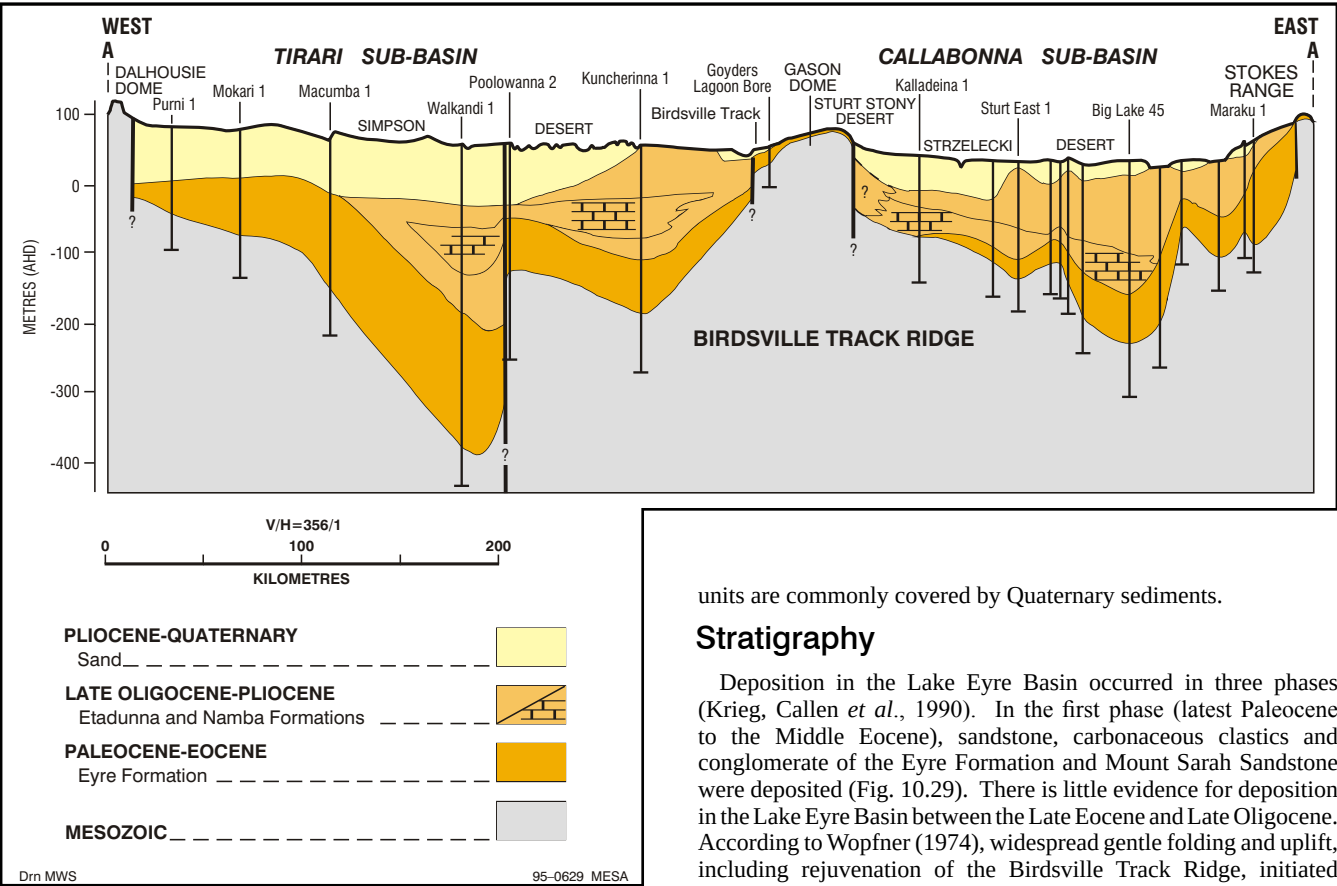


Fig. 10.30 Cross-section through the Tirari and Callabonna Sub-basins. Line of section is located on Figure 10.28.

units are commonly covered by Quaternary sediments.

Stratigraphy

Deposition in the Lake Eyre Basin occurred in three phases (Krieg, Callen *et al.*, 1990). In the first phase (latest Paleocene to the Middle Eocene), sandstone, carbonaceous clastics and conglomerate of the Eyre Formation and Mount Sarah Sandstone were deposited (Fig. 10.29). There is little evidence for deposition in the Lake Eyre Basin between the Late Eocene and Late Oligocene. According to Wopfner (1974), widespread gentle folding and uplift, including rejuvenation of the Birdsville Track Ridge, initiated the division of the basin into two sectors and produced such structures as the Cooryanna and Gason Domes. The second phase

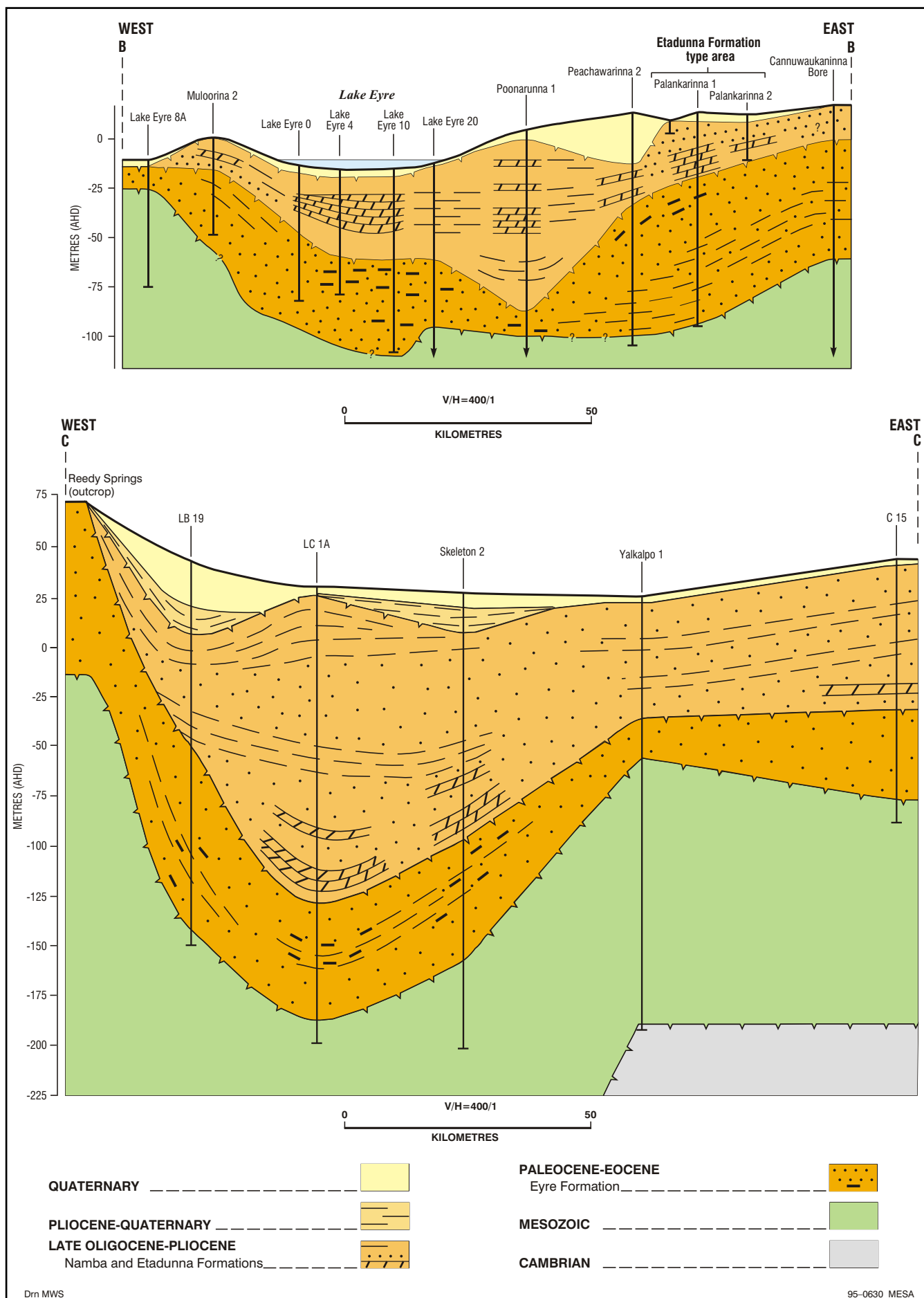


Fig. 10.31 East-west correlations across the southern Tirari Sub-basin (B-B) and Callabonna Sub-basin (C-C). Lines of section are located on Figure 10.28.

may have commenced in the Oligocene and extended through to the Pliocene; sediments deposited include grey, green and white clay, fine-grained sand and carbonate, with minor conglomerate, of the Etadunna Formation and its correlative Namba Formation, Doonbara Formation and Cadelga Limestone. The third phase, during the Pliocene to Quaternary, was characterised by deposition of red and yellow-brown sand and sandy clay, and development of gypsum and carbonate palaeosols.

Late Paleocene-Middle Eocene units

Eyre Formation

The Eyre Formation (Fig. 10.32; Wopfner *et al.*, 1974), basal unit of the Tertiary succession and widespread throughout the Lake Eyre Basin, incorporates the Macumba Sandstone (Freytag *et al.*, 1967) and Murnpeowie Formation (Forbes, 1966). The Eyre Formation mostly disconformably overlies Mesozoic sediments, but locally oversteps onto Cambrian and Precambrian rocks; it is overlain unconformably by the middle to late Tertiary Etadunna and Namba Formations and Quaternary sediments. Upper exposed horizons are commonly silicified.

The Eyre Formation consists of mature, pyritic, carbonaceous sand, although the carbonaceous lithologies are often leached in outcrop. Grain size varies from silt to gravel, with clasts up to small cobble size, sand grains usually being subangular to subrounded and polished. Beds of lignite and clay composed of montmorillonite and kaolinite with some illite, sometimes with root horizons, are common; a clay matrix is often present in sand around the basin margins. Polished gravel composed of resistant siliceous lithologies such as yellow, grey and milky quartz, black chert, red jasper, agate, fossil wood and buff and grey silcrete, are a diagnostic feature of the basal horizons, with pebbles of basement rock in some localities. These are a lag of reworked second or third-cycle clasts, derived from Permian and Mesozoic rocks. Pebbles of white and coloured quartz are dominant in palaeochannels on the northern margin of the Olary Ranges.

Wopfner *et al.* (1974) correlated palynofloras from the formation with the *L. balmei* to Lower *N. asperus* Zones, indicating an age range from early Late Paleocene to Middle Eocene. An hiatus encompassing the Early Eocene was believed to be represented as an erosional disconformity separating older and younger parts of the formation. Alley and Sluiter (in prep.), however, indicate that palynofloras range from latest Paleocene (Upper *L. balmei* Zone) to Middle Eocene with a relatively short hiatus, perhaps in the late Early Eocene to early Middle Eocene.

Other plant fossils are well preserved in silicified sandstone ('silcrete floras') of the Poole Creek Palaeochannel and include cuticle, algae, fungi, wood, twigs, leaves, shoots, seeds, fruits and flowers. A range of coniferous taxa and capsules of possible *Eucalyptus* and various other Myrtaceae have been described (Greenwood *et al.*, 1990; Christophel *et al.*, 1992). The only



Cross-bedded sand unit in the Eyre Formation north of Mount Yerila. Photo 042508.

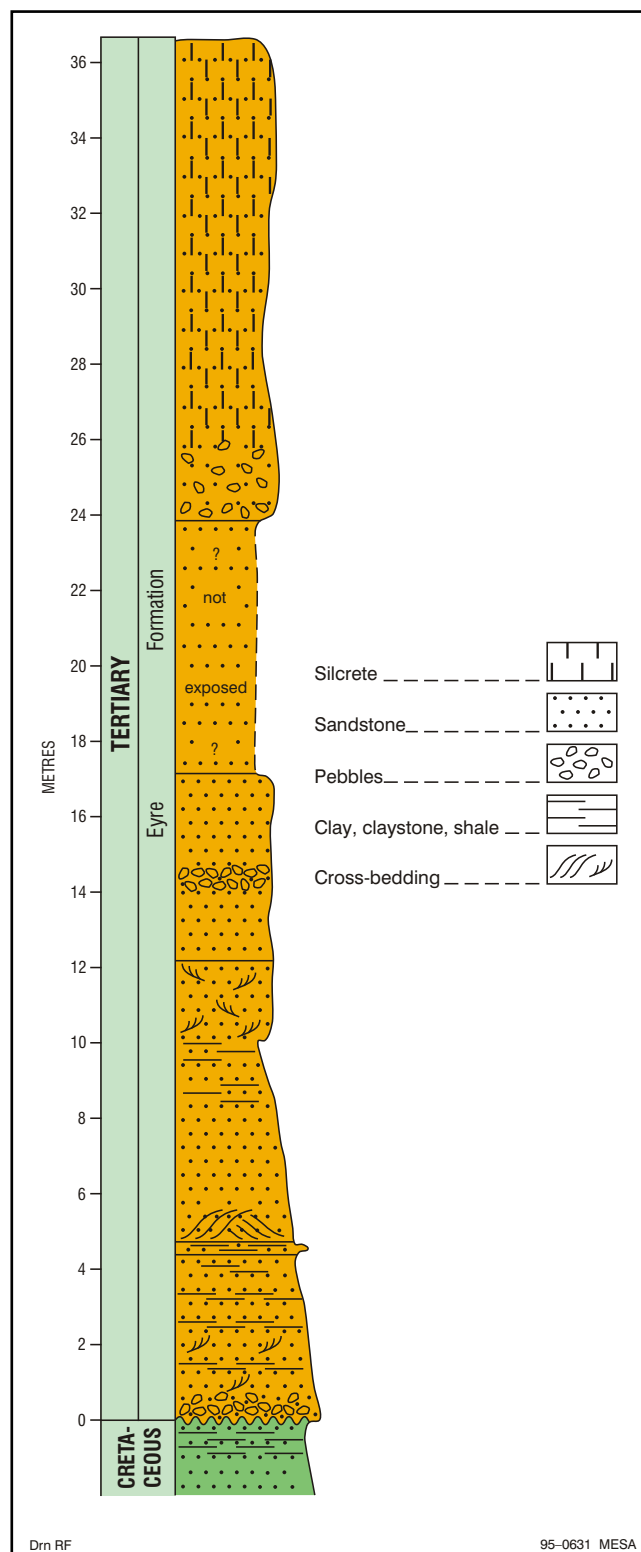


Fig. 10.32 Stratigraphy of the Eyre Formation at the type section, 4.8 km east of 'Innaminka'.

exposure of sediments in which palynofloras and macrofloras occur together is in the lower reaches of Nelly Creek near the southeastern margin of Lake Eyre South. Here macrofossils are preserved in carbonaceous clay lenses as mummified remains and impressions dated palynologically as Middle Eocene (Alley, 1989b), indicating that this is a younger part of the Eyre Formation.

The Eyre Formation was deposited largely by braided streams during epeirogenic uplift of the Olary, Barrier and northern Flinders Ranges, accompanied by subsidence in the Lake Eyre Basin (Krieg,

Callen *et al.*, 1990). Extensive erosion of the Eyre Formation during the early Middle Eocene was followed by fluvial deposition later in the Middle Eocene.

The formation has been correlated with the Glendower Formation of Queensland and Marion Formation of New South Wales (Fig. 10.29), although Whitehouse (1954) and Wopfner *et al.* (1974) believed the Glendower Formation to be younger.

Mount Sarah Sandstone

Barnes and Pitt (1977) defined the Mount Sarah Sandstone from an area west of the main depocentre of the Lake Eyre Basin in the Dalhousie Dome and Stuart Range. Silcrete-capped, massive to cross-bedded sandstone, siltstone and basal conglomerate form a sinuous series of mesas east of the Stuart Range. The basal channel facies containing silcrete, quartz and shale clasts was correlated with the Mirackina Conglomerate by Barnes and Pitt (1976, 1977).

The Mount Sarah Sandstone unconformably overlies weathered Cretaceous sediments and is considered to be disconformable on, and/or in part equivalent to, the upper Eyre Formation (Krieg, 1985). Although similar to Eyre Formation, the Mount Sarah Sandstone was thought to be distinguished by its silcrete clasts (Barnes and Pitt, 1977). However, since silcrete clasts are widespread in the Eyre Formation, the two units are indistinguishable. Alternatively, parts of the Mount Sarah Sandstone may be a shoreline facies of the younger Etadunna and Doonbara Formations (see below).

Late Oligocene-Miocene units

Mirackina Conglomerate

The Mirackina Conglomerate, a cross-bedded fluvial conglomerate with silcrete clasts and medium to coarse-grained sandstone, occupies the Mirackina Palaeochannel, an eroded Tertiary palaeodrainage system on the western margin of the Lake Eyre Basin. Fine-grained sandy siltstone and shale occur in the upper and distal parts; the top is generally ferruginised and silicified. A Miocene age was inferred by Barnes and Pitt (1976) from the abundance of silcrete clasts. Alternatively, these fluvial sediments may correlate in part with the Eyre Formation. Probable equivalents have been recognised in the ABMINGA and WARRINA map areas.

Etadunna Formation

The Etadunna Formation of Stirton *et al.* (1961), with its type section along the western side of Lake Palankarinna, is restricted to the Tirari Sub-basin and usually rests on Eyre Formation or, locally, on Cretaceous strata. The Mampuworudu Sand (Stirton *et al.*, 1961), with a type section at the same locality, is now regarded as a channel facies of the Etadunna Formation. In the southern Lake Eyre region, a basal fine clastic unit has been named the Muloorina Member (Callen and Plane, 1985; Krieg *et al.*, 1991). The Yardina Claystone and Alberga Limestone (Freytag *et al.*, 1967; Wopfner, 1974) may also be members of the Etadunna Formation.

In the type area and westwards to Lake Eyre, the Etadunna Formation comprises white dolomite and limestone overlain by green and grey magnesium-rich claystone and fine sand (Callen and Plane, 1985; Callen, *et al.*, 1986). Thickness varies from 25 m in the type section to 35 m beneath Lake Eyre and 80 m in Poonarunna 1 (Fig. 10.31). Intraformational conglomerate and bioturbation are widespread.

The 5-15 m thick Muloorina Member is prominent beneath the Madigan Gulf and Tirari Desert regions (Callen and Plane, 1985) and consists of greenish grey, very fine-grained sand and silt with calcareous intervals. A basal ferruginised sand contains clasts of

silcrete and sand reworked from the Eyre Formation.

Namba Formation

The Namba Formation was defined by Callen and Tedford (1976) from a type section in Yalkalpo 1, a supplementary section in Wooltana 1, and a supplementary outcrop section on the western side of Lake Tarkarooloo, southeast of Lake Frome. In the southwestern Callabonna Sub-basin, it disconformably overlies Eyre Formation and averages 90 m in thickness but thickens towards the Barrier and Flinders Ranges (e.g. 170 m in Wooltana 1), and thins over the Benagerie Ridge. The unit is similar lithologically to the Etadunna Formation and comprises alternating fine to medium-grained, poorly sorted, angular sand, silt and clay, with thin dolomite and limy, often oolitic, dolomite interbeds. The clay is black due to ferrous iron and manganese, tough and bears slickensided fracture surfaces. Horizontal lamination and small to medium-scale cross-bedding are common in silt and very fine-grained sand. Dolomite frequently



Top: Etadunna Formation along the western margin of Lake Eyre. Photo 042509.

Bottom: Fossils from the Etadunna Formation, including possum, marsupial cat, potoroo, *Diprotodon* (giant wombat), platypus, koala and flamingo. Width of view is 0.3 m. (photo courtesy N. Pledge). Photo 042510.

contains rod-shaped oolites and is affected by widespread bioturbation, intraformational brecciation and slump structures. The Namba Formation is informally divided into a lower member characterised by smectite-dominated clay and cyclic deposition, and an upper member with illite and kaolinite (Callen, 1977).

Carbonate facies of the Etadunna and Namba Formations contain abundant gastropods, ostracodes, algae (*Chara*, *Botryococcus*), stromatolites and rare foraminifera. In the upper clayey parts of the succession, remains of fish, reptiles, birds and mammals are locally abundant. These have aided in reconstructing palaeoenvironments and in understanding the evolution of Australian faunas (see 'Geological history and palaeoclimate'). The uppermost grey clay contains leaf impressions and sparse palynofloras. Occasional leaf impressions have been found in older parts of the Etadunna but none have been observed in the lower Namba Formation. The widespread distribution of the Etadunna and Namba Formations containing abundant fossils of aquatic animals indicates the development of shallow, brackish to freshwater lakes throughout much of the Lake Eyre Basin.

The age of the Etadunna and Namba Formations may range from (?)Late Oligocene to Pliocene. A Rb-Sr age of 25 Ma determined on green diagenetic illite from the upper Etadunna Formation approximates the Oligocene-Miocene boundary (Norrish and Pickering, 1983). A monospecific foraminiferal microfauna of *Buliminoides* cf. *B. chattonensis* in Palankarina 1 tentatively suggests a Late Oligocene age (Lindsay, 1987b). This occurrence is enigmatic since *Buliminoides*, unlike *Ammonia* or *Elphidium*, is not recorded as an immigrant or survivor in inland lakes. Age estimates of fossil vertebrates range from Oligocene to Early-Middle Miocene (Stirton *et al.*, 1961; Woodburne and Clemens, 1986; Woodburne *et al.*, 1985). Palynofloras from the formations are sparse but suggest an age range from Miocene to Pliocene. A possible late Tertiary palynoflora was obtained from the uppermost Etadunna Formation in Palankarina 1 (E.M. Truswell, BMR, pers. comm., 1977), whereas Callen and Tedford (1976) suggested an Early to Middle Miocene age for palynofloras from the Etadunna and Namba Formations. Palynofloras from the Namba Formation in Wooltana 1 indicate an age range from Late Oligocene to Pliocene (Martin, 1990). Late Miocene palynofloras have been reported from Namba Formation equivalents in the southeastern Callabonna Sub-basin in New South Wales (Morgan, 1977; McMinn, 1981).

Doonbara Formation and Cadelga Limestone

Wopfner (1974) defined the Doonbara Formation in the northeastern extremity of South Australia, where it comprises pisolitic and ferruginous fine to medium-grained sand overlying partly silicified Eyre Formation (partly silcreted). The formation is generally 7-10 m thick, but reaches 40 m in the core of the Haddon Syncline. Deposition followed, or was partly contemporaneous with, folding of the underlying silcrete surface into broad domes. Although there is no biostratigraphic evidence for its age, Wopfner (1974) interpreted the Doonbara Formation as a fluvial facies variant of the basal Etadunna Formation. Callen (1983a), however, suggested that fluvial sand and clay at the top of the Namba Formation east of Strzelecki Creek become increasingly ferruginised to the north, and are equivalents of the Doonbara Formation. It is possible, however, that these clastics are a regolith and beach equivalent of the Cadelga Limestone.

The Cadelga Limestone (Wopfner, 1974) is a pale cherty and dolomitic lacustrine limestone up to 5 m thick, which overlies and incorporates fragments of Doonbara Formation. Sparse gastropods, possible diatoms and algae are present but are not diagnostic of age. On the basis of their similar lithologies, Wopfner (1974) correlated the Cadelga Limestone with the Etadunna Formation and with the Alberga Limestone of the Oodnadatta area.

The weathered nature of the sediments and lack of biostratigraphic evidence make correlations with other similar units difficult. However, the Austral Downs Limestone and Noranside Limestone of the Boulia region (Queensland) closely resemble the Cadelga Limestone and Doonbara Formation, and are probably of Late Miocene to Quaternary age (Paten, 1964).

Pliocene-Quaternary units

Wipajiri, Tirari and Kutjitara Formations

The final phase of Cainozoic sedimentation in the Lake Eyre Basin is characterised by red-brown arenites and dark, fine-grained lacustrine sediments. Aeolian and evaporitic facies are common and dense horizons of calcrete and gypsum developed in soils. Although many of these deposits may range in age up into the Quaternary, the deposits of possible Pliocene age are discussed below. There is no direct evidence of a Pliocene age for any of the rock units in the Lake Eyre Basin.

The Wipajiri Formation, defined by Stirton *et al.* (1967) from Lake Ngapakaldi in the central Tirari Desert, consists of over 1 m of fine-grained sand and silt channel fill cut into the Etadunna Formation. It contains coarser conglomeratic lenses up to 0.3 m thick with intraclasts including siliceous and ferruginous nodules, vertebrate remains and *Unionid* shell impressions, but their stratigraphic implications are uncertain (Archer and Hand, 1984). The upper part of the section consists of laminated claystone with leaf impressions.

Unnamed black clay and white sand in the Tirari Desert is either Wipajiri Formation or the younger Kutjitara Formation. The clay contains well-preserved palynofloras, including *Nothofagidites* spp., *Podocarpidites ellipticus* and *Phyllocladidites mawsonii*, which is suggestive of a late Tertiary rather than Quaternary age. This sequence, characterised by shallow sand-filled channels, is interpreted as a meandering fluvial system or lacustrine delta-fan.

The Tirari Formation, a widespread unit with a type section at Lake Palankarina, consists of 4-5 m of bright red-brown silt with sandy channel deposits and fossil vertebrates, including *Diprotodon* and short-faced kangaroos. The formation is extensively exposed over the central Tirari Desert and northwards into the Simpson Desert, and is usually cemented with massive gypsum crusts. The depositional environment was probably one of intermittent stream flow under semi-arid conditions. The unit overlies possible Wipajiri Formation at a small lake near Peachawarina 2 and has been dated there as >0.5 Ma by thermoluminescence methods (Callen and Nanson, 1992).

Economic geology

The Eyre Formation has been explored extensively for sedimentary uranium, particularly in the Lake Frome area where several deposits have been defined in the Billeroo and Yarramba Palaeochannels. Mining by *in situ* leaching is feasible (Brunt, 1978; Ellis, 1980; Callen, 1990; Curtis *et al.*, 1990). Although lignite has been disclosed by drilling, the beds are thin and of low quality. Sandstone beds form useful aquifers in the Callabonna Sub-basin and are locally pressurised along the western margin of Lake Frome. Zircon and topaz occur locally in significant concentrations in palaeochannels along the southern part of the Callabonna Sub-basin; similar channels elsewhere are potential sources of heavy minerals.

Prospecting for dolomite as a blast furnace flux was undertaken in the Etadunna Formation around the margins of Lake Eyre, but the silica content proved to be too high (Whitehead, 1970).

The Namba Formation has been explored for palygorskite, a swelling clay used as a catalyst, an ingredient of drilling mud (attapulgitite), a dye absorbent, and a sager earth in pottery. However, these investigations have failed to disclose marketable material.

Uranium in a sandy palaeochannel of Namba Formation near Lake Namba (Curtis *et al.*, 1990) is potentially recoverable by *in situ* leaching at the Beverley Prospect near 'Wooltana'.

Celestite, a mineral with growing demand in the television and computing industries, is widespread throughout Cainozoic sediments. Exploration programs are currently assessing potential economic sources.

BILLA KALINA BASIN

R.A. Callen and W.M. Cowley

Structure and setting

The Billa Kalina Basin (Ambrose and Flint, 1981; Callen *et al.*, 1986; Cowley and Martin, 1991; Krieg *et al.*, 1991) is separated by the Denison-Willouran divide and Norwest Fault from the Lake Eyre Basin to the northeast (Fig. 10.33). The western margin is bounded by high country along an extension of the Stuart Range. Northwest-trending faults along the southwestern margin (Cowley and Martin, 1991) are interpreted to post-date sedimentation within the Billa Kalina Basin, but may have been active during deposition. Other margins have uncertain tectonic controls and may be largely erosional. Billa Kalina Basin sandstones are commonly silicified; eroded remnants in the east form an arc around dolomite and clay exposed in a plateau to the west. The base of the Tertiary succession varies in elevation from ~40 to 100 m above sea level.

Stratigraphy

The Billa Kalina Basin contains a thin, flat-lying succession of early Tertiary to possibly Pliocene age, including the Willalinchina Sandstone, Watchie Sandstone and Mirikata Formation (Figs 10.2, 10.34).

?Late Paleocene-Eocene units

Silicified sandstone capping the Stuart Range

Although more correctly assigned to the older part of the Lake Eyre Basin, these sediments are described here because of their relevance to understanding the development of the Billa Kalina Basin. Referred to as Munjena Formation equivalent by Cowley and Martin (1991), they crop out on elevated areas in the western part of the basin at the southern end of the Stuart Range (Ambrose and Flint, 1981; Cowley and Martin, 1991). The poorly sorted sandstone contains rounded pebbles of quartz, chert, quartzite and silcrete, and overlies Mesozoic sediments and occasionally Precambrian rocks. Cowley and Martin (1991) interpreted the sandstone as largely a residual to colluvial deposit with only minor fluvial influence. Because of its position as isolated mesas on the highest parts of the topography, the sandstone is considered as the oldest Tertiary unit in the area and a colluvial-regolithic equivalent of the lower Eyre Formation in the Lake Eyre Basin, and hence ascribed a tentative Paleocene to Eocene age (Cowley and Martin, 1991).

Uplift and dissection before and during the Middle Eocene is implied by the present distribution of the sandstone on the Stuart Range and by a possible reversal of flow along channels containing Willalinchina Sandstone, from southwesterly to northeasterly towards the Lake Eyre Basin (Krieg *et al.*, 1991). This divide, and the Stuart Range to the north, thus became a watershed for drainage flowing towards the Eucla Basin in the south and Lake Eyre Basin to the north.

Willalinchina Sandstone

The Willalinchina Sandstone consists of a pebble conglomerate less than 0.5 m thick, overlain by 7-9 m of strongly silicified quartz sandstone. The unit has been cut by a series of 2-3 m thick brown-weathering, sandstone-filled palaeochannels, without basal pebble layers. The top of the succession comprises ~2 m of fine-grained sandstone silicified by 'reed-mould' silcrete (Fig. 10.35). Medium-scale cross-bedding and rippled surfaces are common in the channel facies; other sandstones are thinly bedded. The basal conglomerate contains clasts of Wilpena Group quartzite, milky quartz, ferruginised Bulldog Shale and other Cretaceous sediments, silcrete (including 'reed-mould' type) and sandstone intraclasts. Being a channel fill, the Willalinchina Sandstone is restricted to linear outcrops in central parts of palaeovalleys in the Stuart Creek Valley and Olympic Dam areas, where it disconformably overlies

Adelaidean, Cambrian and Cretaceous sediments. The formation was deposited in meandering channels probably flowing to the northeast in the Stuart Creek area and to the southwest near Wirragilpina Dam.

The channel sediments preserve abundant leaf impressions and casts, the uppermost sandstone with reed-like plant impressions, sometimes with tuberous root stocks, being referred to as the 'reed-mould' silcrete (Ambrose *et al.*, 1979; Callen *et al.*, 1986). The fossils include the oldest confirmed *Eucalyptus* in Australia and contain macrofossils with some affinities to those from the Eyre Formation in the Poole Creek Palaeochannel and Nelly Creek (Ambrose *et al.*, 1979; Greenwood *et al.*, 1990). These were considered to be consistent with a Middle Eocene age for the Willalinchina Sandstone (Greenwood *et al.*, 1990).

Scattered outcrops of sandstone assumed to be Willalinchina equivalent in the Woomera and Andamooka areas also contain remarkably well-preserved silcrete floras (Offer, 1969, 1984; Lange, 1978, 1982). The sand resembles the lower Eyre Formation from the Lake Eyre Basin and was deposited along an east-northeast-trending fault zone in a palaeochannel apparently draining towards the Torrens Basin.

Miocene units

Watchie Sandstone

The Watchie Sandstone overlies Cretaceous or Cambrian rocks and, in the Stuart Creek area, is interpreted to overlie Willalinchina Sandstone although stratigraphic relationships here are ambiguous (Krieg *et al.*, 1991). The sandstone is 5-7 m thick and comprises a thin basal conglomerate with occasional sand lenses, a lower silt to fine-grained sand 2-3 m thick and an upper, upward-coarsening sand. Pebbles at the base of the formation are similar to those of the basal Eyre Formation, but also include Precambrian quartzite clasts derived directly from bedrock or via the Bulldog Shale, whereas pebbles near the top of the unit are mainly polished silcrete (Krieg *et al.*, 1991). Pebble beds often mark the sharp contact between the fine-grained unit and the upper sand. Cross-bedding and cross-lamination are ubiquitous, but horizontal lamination is rare. Alternating silcrete and ferricrete horizons exhibit pedogenic structures such as nodules, rhizomorphs and burrows, the morphology of which vary laterally.

The upper sand is formed into a series of concentric, arcuate beach ridges which developed on the eastern shoreline of a desiccating lake under the influence of a strong westerly airstream (Ambrose and Flint, 1981; Cowley and Martin, 1991; Krieg *et al.*, 1991). Preferential erosion of the ridges has left the intervening flats, which were more strongly silicified, as a series of arcuate mesas (beach mounds) up to 130 km long (Figs 10.33, 10.35, 10.44).

The Watchie Sandstone is regarded as a regressive lacustrine sequence; the lower fine-grained arenite and the lacustrine clay-dolomite sediments of the Mirikata Formation in the western part of the basin represent the lake facies, and the upper coarse-grained sand the shoreline facies. The arenaceous sediments have not been shown to intertongue with the clay-dolomite and the stratigraphic relationship between the two facies is enigmatic. All workers agree, however, that the Watchie Sandstone was deposited more or less concurrently with the lacustrine Mirikata Formation (see below). The dolomite units contain mollusca interpreted to be of Miocene age (Ludbrook, 1980) and, thus, the Watchie Sandstone is also Miocene. Other fossils include burrows, some related to termites and ants, and plant impressions, but none are useful in dating. The lower laminated sandstone and siltstone contain ferruginised plant impressions including forms not seen in either the Willalinchina Sandstone or Eyre Formation.

The Willalinchina Sandstone, which crops out in palaeovalleys topographically lower than the Watchie, has been interpreted as representing fluvial incision immediately preceding the main lacustrine phase (Ambrose and Flint, 1981). Thus, there are

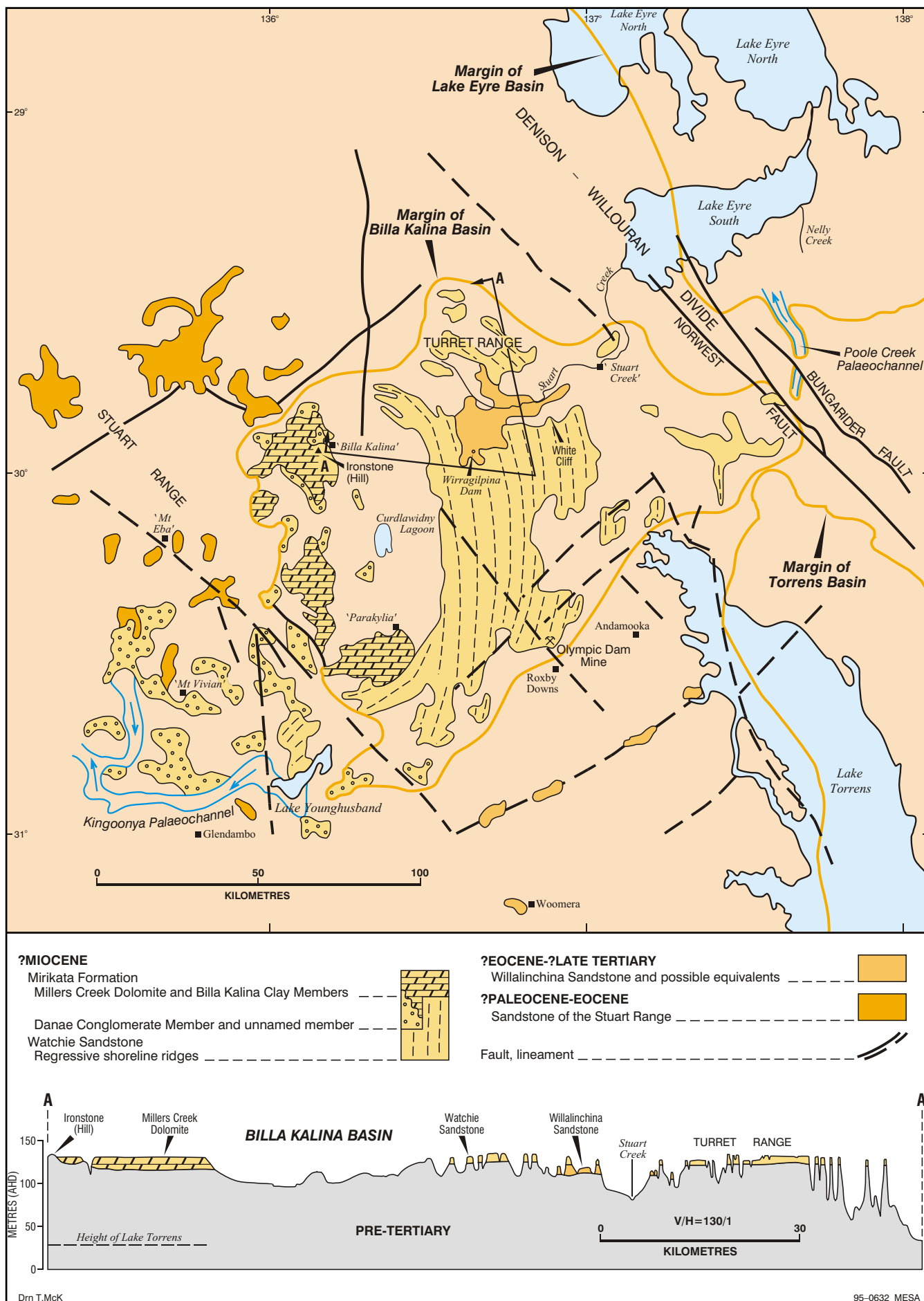


Fig. 10.33 Locality map and structural elements of the Billa Kalina Basin.

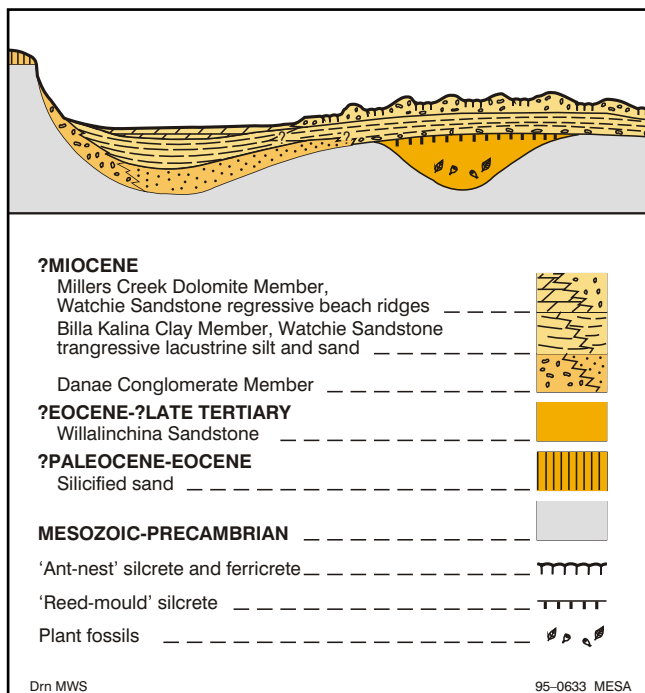


Fig. 10.34 Rock relation chart for Tertiary sediments in the Billa Kalina Basin.

alternative models to explain the above apparent stratigraphic relationships. However, the palaeochannels may incise, and their sediments post-date, the Watchie Sandstone, or reflect lowering of base level and incision of earlier formed shorelines (Ambrose and Flint, 1981).

Mirikata Formation

The Mirikata Formation, as redefined by Krieg *et al.* (1991), is a clay-dolomite succession up to 15 m thick and restricted to the central western part of the basin. It is divided into the thin basal Danae Conglomerate Member, Billa Kalina Clay Member, and Millers Creek Dolomite Member (Figs 10.2, 10.34; Ambrose and Flint, 1981; Cowley and Martin, 1991).

Top: Leaf fossils contained in Willalinchina Sandstone, Stuart Creek. Photo 042132.

Bottom: Watchie Sandstone, closer view of unconformable contact with Bulldog Shale (at top of geologist's head). Ant nest silcrete is at top of section, southern side of Turret Range. Photo 039098.

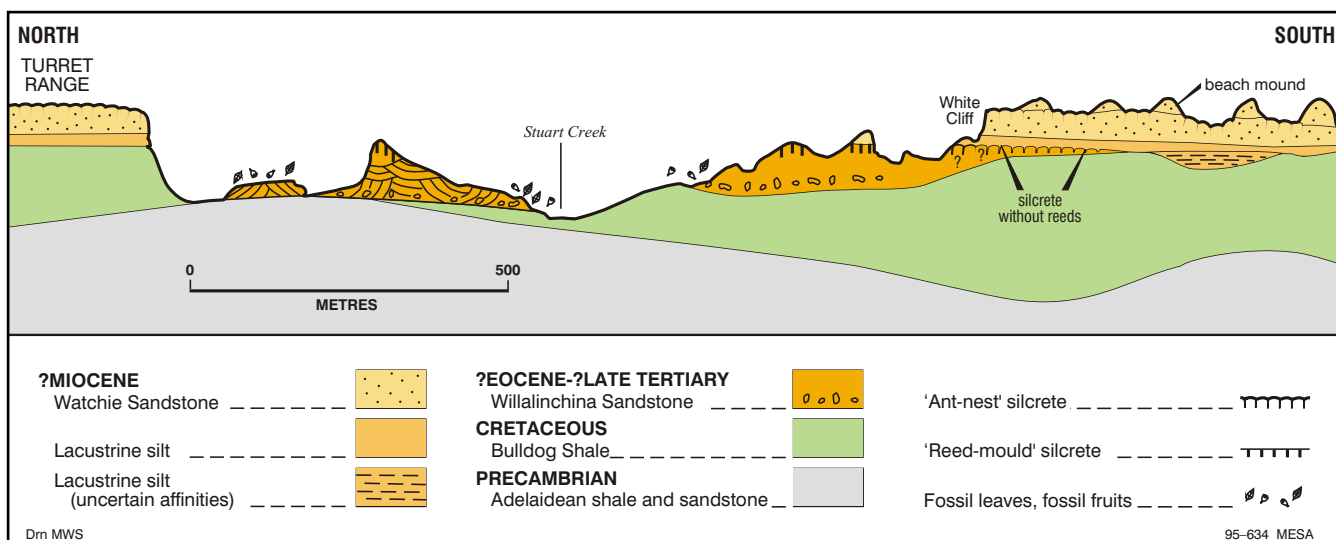
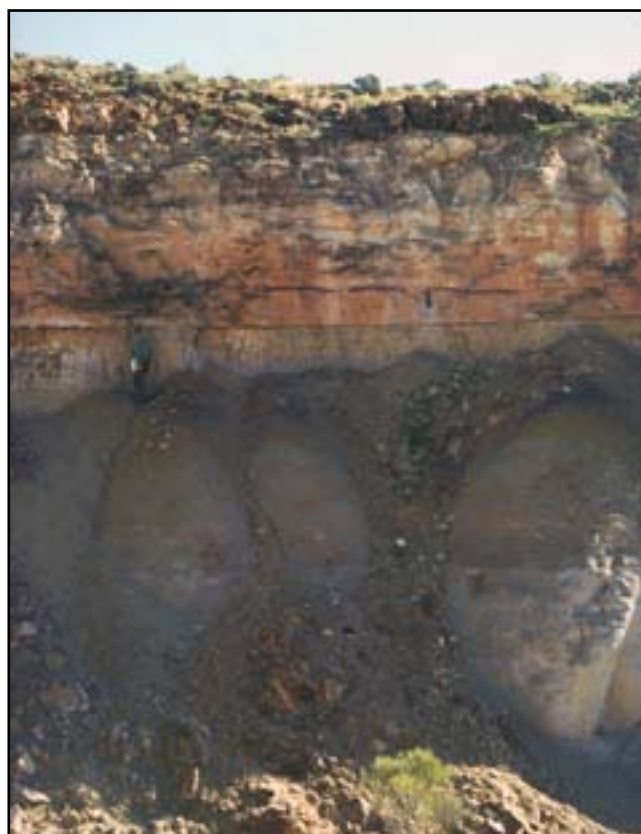


Fig. 10.35 Relationships between silcretes and rock units in the Billa Kalina Basin.

The Danae Conglomerate Member, up to 3 m thick, comprises resistant clasts, some of silcrete and others reworked from Permian and Cretaceous sediments, in a matrix of fine to coarse-grained sand. The unit is usually ferruginised and silicified, forming thin sheets and lenses deposited in channels cut into Mesozoic sediments. The contact with the overlying Billa Kalina Clay Member is gradational. Reed impressions are present west of Curdlawidny Lagoon and north of 'Parakylia' (Cowley and Martin, 1991).

The Billa Kalina Clay Member is a 4-6 m thick, yellow to olive-green and grey clay including magnesium-rich clay, palygorskite, and a mixture of montmorillonite and illite.

The Millers Creek Dolomite Member near 'Billa Kalina' is up to 10 m thick and comprises two dolomite or dolomitic limestone layers separated by dolomitic palygorskite clay. To the south, thicknesses of 1-5 m are more usual. Contact with the underlying Billa Kalina Clay Member is gradational. The dolomite is white or cream, aphanitic, commonly intraformationally brecciated, and locally includes sandy dolomite or calcareous sandstone. The Danae Conglomerate Member either intertongues with or underlies this unit (Ambrose and Flint, 1981). Outcrops are commonly silcreted or calcreted. The Millers Creek Dolomite Member contains mollusca dated as Miocene (Ludbrook, 1980). The Mirikata Formation is therefore correlated with the middle to late Tertiary Etadunna Formation of the Lake Eyre Basin, and was deposited in an alkaline lacustrine environment at the centre of the Billa Kalina Basin either during the regression represented by the Watchie Sandstone or later.

Silicified, moderately to poorly sorted, coarse-grained to granule and pebbly sandstone south and southwest of the Billa Kalina Basin was referred to as an unnamed member of the Mirikata Formation by Cowley and Martin (1991). It is interpreted as the remnant of a fluvial blanket bordering the Billa Kalina Basin and Kingoonya Palaeochannel in a palaeogeographic setting analogous to that of the Danae Conglomerate Member to the north. Silicified ridges possibly representing dunes and fluvial channels occur northwest and east of Lake Younghusband (Cowley and Martin, 1991).

The Mirikata Formation and Watchie Sandstone occur ~200 m topographically higher than correlatives in the adjacent Torrens Basin (Fig. 10.33). This may be due to tectonic uplift along the western margin of the Torrens Basin or southern margin of the Billa Kalina Basin, or to different base levels.

HAMILTON BASIN

P.A. Rogers

This relatively small late Tertiary basin lies between Hamilton Creek and Alberga River and extends eastwards from the Musgrave Block to the western flank of the Mount Sarah Anticline of Krieg (1985), which separates it from the Lake Eyre Basin (Fig. 10.1). The sediments thicken gradually to 79 m at the western edge of the basin.

The succession comprises poorly to well-sorted, silty and clayey, very fine to very coarse, feldspathic quartz sandstone with angular granules and pebbles of quartz, silcrete and claystone, and very fine sandy siltstone and claystone. The basal part of the succession contains abundant kaolinised, porcellanised and ferruginised clasts reworked from weathered Mesozoic rocks. The sediments are interpreted as fluvial deposits laid down in flood-plain and piedmont fan environments.

Smaller sub-basins and a palaeochannel containing carbonate-rich sediments border the northern edge of the Hamilton Basin. Crocodile teeth of probable Pliocene age have been found in this area (N. Pledge, SA Museum, pers. comm., 1985). Sediments of the basin are correlated with the lithologically similar lower clastic part of the Neogene Mangatitja Formation which crops out extensively through the Musgrave Block.

VARIOUS INTRAMONTANE BASINS

Tertiary sediments are found in a number of intramontane basins in the Flinders and Mount Lofty Ranges, the two largest being the Willochra and Walloway Basins. The age and stratigraphy of their sediments and their tectonic evolution are poorly known.

WILLOCHRA BASIN

N.F. Alley and J.M. Lindsay

The Willochra Basin, which occupies a series of bedrock depressions between Booleroo Centre in the south and the Simmonston area in the north, is an important source of underground and artesian water (Fig. 10.36; O'Driscoll, 1956). The southern part of the basin is a graben delineated by approximately north-south-trending faults (Milton and Twidale, 1978), but the origin of the remainder of the depression is unknown. Three stratigraphic drillholes penetrate the Tertiary succession which is locally more than 115 m thick. The basal unit, the Langwarren Formation, comprises sand and clay with minor gravel beds and carbonaceous intervals, believed to have been deposited under fluvial piedmont fan conditions (Shepherd and Thatcher, 1959) or in a lacustrine environment (Twidale, 1966). Since the Langwarren Formation was defined from outcrop around the fringes of the basin, the stratigraphic relationship with deeper basinal sediments is unknown.

Approximately 10 m of brown, carbonaceous silt and sand occur near the base of the Tertiary succession in Willochra 3, northeast of Melrose (Ludbrook, 1954; Shepherd and Thatcher, 1959). An early palynological assessment indicated a post-Eocene but pre-Pliocene age. Palynofloras from carbonaceous silt between 24 and 31 m in Willochra 2, northeast of Wilmington, were regarded as Middle to Late Eocene (Twidale, 1966) and correlated with the Middle Eocene Lower *N. asperus* Zone (Harris, 1970b). However, assemblages from this drillhole, at least at 24 m depth, are correlative with the Middle *N. asperus* Zone, indicating that a latest Middle Eocene to Late Eocene age is more likely.

Rare dinoflagellates in palynofloras from 24 and 30 m in Willochra 2 and low numbers of benthonic forams in outcrop samples (N.H. Ludbrook, SADME, pers. comm., 1972) imply paralic conditions during deposition of at least the upper part of the Tertiary succession. The direction from which this weak marine influence originated is uncertain but the sea may have transgressed from the St Vincent or Pirie Basins into the Gladstone and Booleroo Centre areas. Subsequent uplift of this part of the ranges could have led to stripping of Eocene sediments; extensive erosion of early and middle Tertiary landsurfaces in the Mid-North has been attributed to tectonically induced stream rejuvenation (Alley, 1973, 1977).

WALLOWAY BASIN

N.F. Alley and J.M. Lindsay

The Tertiary succession in this intramontane basin occurs entirely subsurface, mostly in the lowland north of Orroroo (Fig. 10.36). The Walloway Basin has been interpreted to extend ~35 km south of Orroroo to Mannanarie (Hillwood, 1969; Shepherd, 1978), but little geological information is available for the southern part and it is uncertain whether Tertiary sediments exist beneath the Quaternary. Over 300 water wells have been drilled in the basin; the artesian aquifer in the lower part of the Tertiary sequence yields the best quality groundwater near Orroroo (Shepherd, 1978).

Geophysical surveys indicate that the basin is probably a fault-angle depression, with a fault along the southwestern margin (Finlayson, 1980; Kwitko, 1982b). The unnamed Tertiary succession thickens southwestwards and also west of a probable major basement fault extending north-northeast from Orroroo (ETSA, 1983). The greatest thickness of Tertiary sediments (250 m) was penetrated in SADME Walloway 1, 10 km north of Orroroo. Two lithological units have been recognised, a lower interbedded fluvial sand, silt and clay, and an upper, more extensive and continuous lacustrine unit of grey, brown and black clay (Kwitko, 1982b).

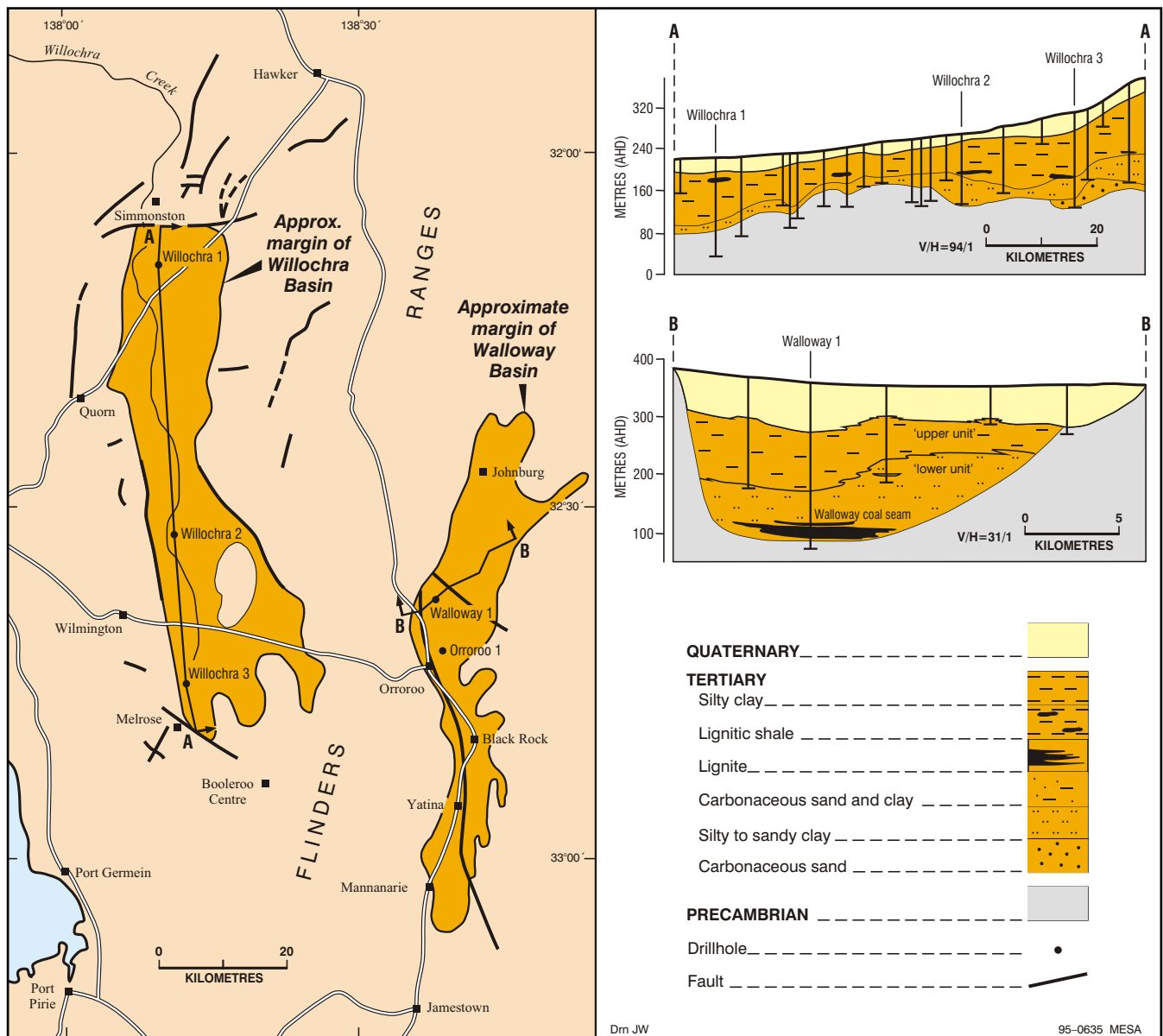


Fig. 10.36 Locality map, structural elements and cross-section through Tertiary sediments in the Willochra and Walloway Basins (after Kwtiko, 1982b).

Although occasional lignitic horizons occur within the Tertiary succession, only Walloway 1 has encountered a significant thickness of coal, comprising two seams, one 20 m thick (250-270 m) and the other 3 m thick (244-247 m). However, the thick seam is not extensive and too deep to be prospective (ETSA, 1983).

Palynofloras from several samples at a depth of ~173 m in SADME Orroroo 1 are correlated with the Lower *N. asperus* Zone of Middle Eocene age (Harris, 1970b; N.F. Alley, unpublished information).

MEADOWS BASIN

N.F. Alley and J.M. Lindsay

In the Meadows-Kuitpo valley, 30 km south of Adelaide, Tertiary sediments may extend to a depth of 90 m. A drillhole 3 km south-southwest of Meadows penetrated a succession of white, pink, black and blue-grey sand between 30 m and total depth of 87 m. The facies varies rapidly laterally and, only 350 m to the north, another drillhole intersected mainly sandy clay to a total depth of 79 m, with dark grey, carbonaceous sandy silt at 20-27 m. Poorly preserved palynofloras indicate a middle to late Tertiary age. Shallow marine (possibly estuarine) to fluvial conditions are implied by very rare acritarchs, and a weak marine connection with the Myponga Basin is possible.

MYPONGA AND HINDMARSH TIERS BASINS

J.M. Lindsay and N.F. Alley

Like the Meadows Basin, the adjacent Myponga and Hindmarsh Tiers Basins are peripheral to the St Vincent Basin (Fig. 10.37). However, despite their narrow width, they contain more than 170 and 119 m of mid-Tertiary marine bryozoal limestone respectively (Furness *et al.*, 1981), largely overlying Permian sand in pre-existing valleys. Despite its limited outcrop, the limestone is locally an important aquifer.

In the Hindmarsh Tiers Basin, the base of the limestone succession is 80 m above present mean sea level. The terrigenous content increases towards the base. The Myponga Basin has a thicker and more complex Tertiary succession and drillhole data suggest a base near present sea level. Marine interconnection with the St Vincent Basin is more likely than with the Murray Basin, although a throughgoing marine passage across Fleurieu Peninsula at eustatic high peaks is possible. The divide was less significant than at present because of uplift of the Mount Lofty Ranges in the order of 100 m since the Middle Miocene (Lindsay, 1986; Bourman and Lindsay, 1989).

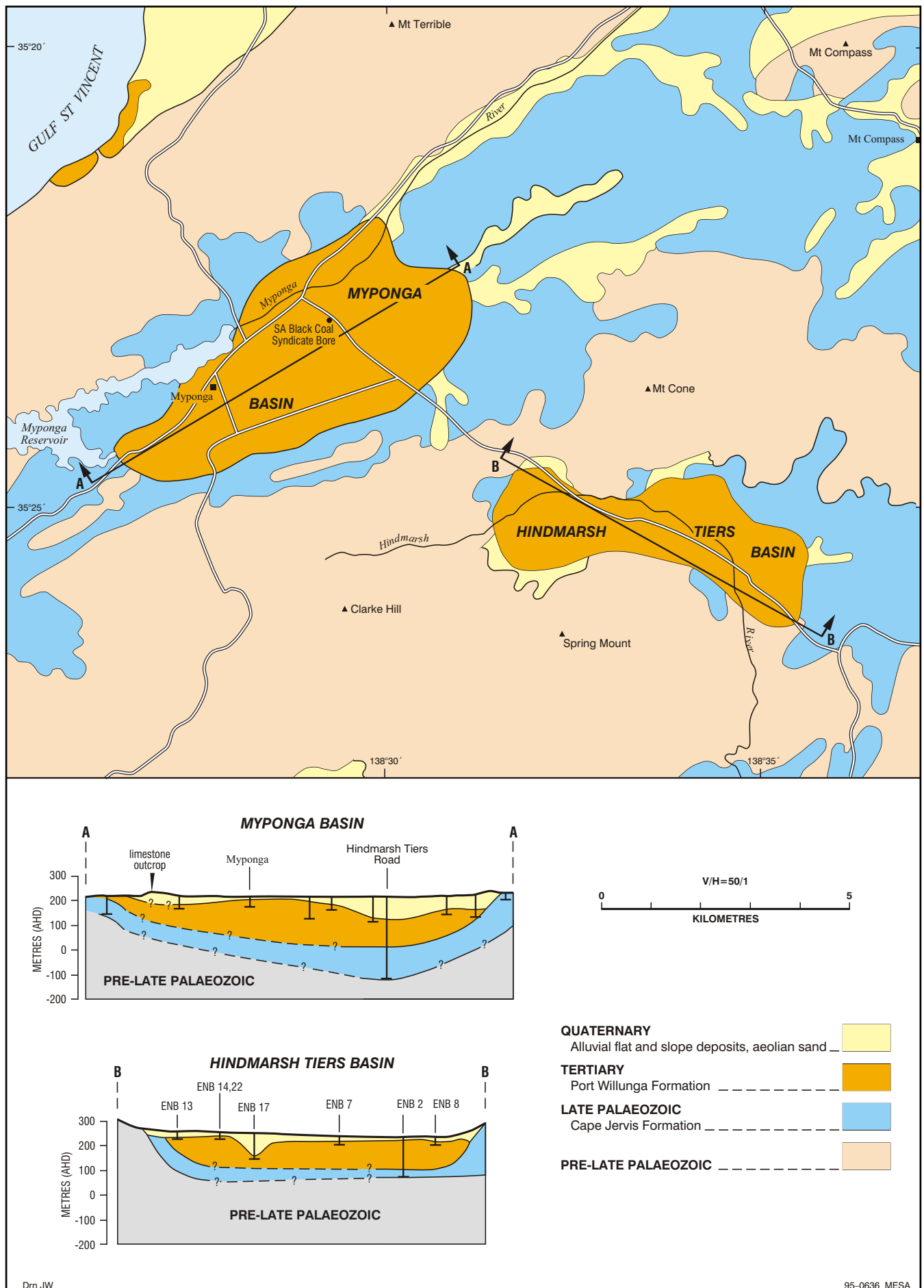


Fig. 10.37 Locality map and cross-section through the Myponga and Hindmarsh Tiers Basins (after Furness *et al.*, 1981).

The initial transgression in the Hindmarsh Tiers Basin has been dated by a foraminiferal fauna as Late Oligocene (early Janjukian) age on a core sample from near the base of the limestone in ENB 2 (Furness, 1975). Core samples of skeletal calcarenite from between 10 and 66 m yielded benthonic foraminiferal assemblages indicative of a Longfordian (Early Miocene) age.

A younger Miocene episode of bryozoal limestone deposition has been recognised only in the Myponga Basin, where the early Middle Miocene *L. howchini* Zone was penetrated within the interval 107.6-127.7 m in SA Black Coal Syndicate Bore, 2.5 km east of Myponga (Glaessner and Wade, 1958; Lindsay, 1986). The Tertiary succession between 84 and 201 m in this drillhole comprises predominantly bryozoal limestone divided by at least four regressive intervals of carbonaceous clay or quartz sand, representing regressive episodes.

Although this is a striking record of transgressive and regressive episodes, there are insufficient data to correlate with the sequence cycles and eustatic curves of Haq *et al.* (1987b), except that the *Lepidocyclina* Zone almost certainly correlates with third order cycle and eustatic peak 2.3, as in the Murray Basin.

DEEP WEATHERING, FERRICRETE AND SILCRETE

M.C. Benbow, R.A. Callen, R.P. Bourman and N.F. Alley

Much of Australia is deeply weathered and mantled by silcrete and ferricrete duricrusts. Although Woolnough (1927) visualised the various products of weathering as related to the formation of a vast 'peneplain' in the Miocene, landform and duricrust evolution in the Tertiary is now known to be more complex. Duricrusts formed a number of times and cap a variety of palaeosurfaces of different ages and origins; some weathering profiles and palaeosurfaces are regarded as polygenetic (Senior and Mabbutt, 1979; Firman, 1981; Callen, 1983a; Bourman *et al.*, 1987; Ollier *et al.*, 1988).

The observation that particular duricrusts are extensive over large areas led to the conclusion that they are morphostratigraphic markers formed during discrete intervals of time under certain climatic conditions (Wopfner, 1974, 1978; McGowran, 1979; Firman, 1981). Whilst this may be true on a local scale, caution is required as there are problems in correlating duricrusts on a broad regional scale. Weathering is an ongoing process that, in time, has operated through a range of climatic conditions (Bird and Chivas, 1993; Bourman, 1993a). A duricrust or a weathering profile currently exposed at the surface may or may not be related to any one particular climatic phase (McGowran, 1979; Milnes *et al.*, 1985; Bourman *et al.*, 1987). The occurrence of duricrust within a sedimentary sequence may instead reflect the geological event leading to its burial and preservation, rather than indicating the cessation of weathering due to climatic change, although there is no doubt that a new phase of erosion, weathering and deposition may also be caused by climatic change.

DEEP WEATHERING

Deeply weathered rocks are widespread in South Australia, commonly reaching to depths of 30 to 60 m, and with or without duricrust cappings. Differentiating discrete events is problematic because weathering may occur in vague or discrete zones and some profiles are composite.

The deep weathering of Eromanga Basin sediments is often characterised (from the surface) by pallid, mottled or varicoloured, and ferruginous zones (Senior and Mabbutt, 1979). The upper, white, bleached zone is usually siliceous, particularly at the top and commonly capped by silcrete. Kaolinite predominates throughout the profile, with lesser illite and/or montmorillonite (Jessup and Norris, 1971; Barker, 1980). Iron occurs in the lower ferruginous zone as disseminated oxides, concretions or accumulations parallel to bedding.



Top: White, bleached weathered zone in Bulldog Shale capped by silcrete, Stuart Creek area. Photo 042511.



Bottom: Mottled and ferruginous zones in Bulldog Shale below silcrete, Painted Hills area east of 'Arckaringa'. Photo 042512.

On the margin of the Eucla Basin and in the region dissected by the palaeodrainage, widespread deeply weathered Precambrian and Palaeozoic rocks are exposed along the escarpments of the duricrust-capped plateaus and younger surfaces adjacent to low-lying playas. Although the weathering profiles are often capped by silcrete and/or ferricrete, there is not necessarily a genetic relationship between them. Deep weathering appears to be absent beneath much of the carbonate-capped part of the Eucla Basin.

Rocks of various ages on Eyre Peninsula and Kangaroo Island and in the Mount Lofty and Flinders Ranges are deeply weathered, probably caused by continuous weathering and reworking of the profiles over long intervals of time (Milnes *et al.*, 1985; Bourman

et al., 1987). Many weathered rocks in the Mount Lofty Ranges still contain variable amounts of feldspars, smectite, micas, chlorite and vermiculite, indicating that the degree of weathering is not intensive. Weathering profiles in the southern part of the State commonly have iron accumulations in their upper parts (see below).

Much of the deep weathering in Eromanga Basin and older sediments may have occurred prior to the Middle to Late Eocene. In the Lake Eyre Basin the Eyre Formation is incised into, and contains clasts of, the Morney profile (Senior and Mabbutt, 1979). Palaeomagnetic measurements on ironstone in the lower part of the profile have given a Maastrichtian to Eocene age (Idnurm and Senior, 1978), and limited oxygen isotope determinations suggest an age greater than 30-40 Ma (Bird and Chivas, 1989). Kaolinised Adelaidean rocks overlain by Late Jurassic to Early Cretaceous Algebuckina Sandstone equivalent near Copley produced oxygen isotope ratios consistent with a pre-late Mesozoic age (Bird and Chivas, 1993).

The Middle to Late Eocene Pidinga Formation of palaeodrainage channels in the Eucla Basin commonly overlies deeply weathered Permian and older rocks. There are also occurrences of deeply weathered Pidinga Formation capped by silcrete, but these are uncommon. In the St Vincent Basin, unweathered Middle to Late Eocene North Maslin Sand and/or Clinton Formation overlie at least 20 m of deeply weathered Adelaidean bedrock; unweathered plant roots occur at the top of the profile in a few localities (see 'St Vincent Basin').

Oxygen isotope dating of kaolinite samples from the summit surface on Fleurieu Peninsula suggests that kaolinisation here is younger than mid- to late Tertiary (Bird and Chivas, 1989). However, analyses of kaolinite samples from successively lower levels on the Parawa Plateau on southern Fleurieu Peninsula suggest that the age of kaolinite formation decreases with depth. This trend supports the conclusion that deep weathering under the summit surface on the peninsula may have continued after, and have been facilitated by, uplift of the Mount Lofty Ranges (Bourman, 1973). Oxygen isotope dating on weathered pegmatite from Williamstown produced a middle Tertiary age and, on kaolinite from Birdwood, a pre-late Mesozoic age. Oxygen isotope dating of weathered sediments underlying Jurassic basalt on Kangaroo Island suggests that weathering there may have occurred prior to the late Mesozoic. However, isotope ages from weathered rock on the summit surface of the island are similar to those from post-middle Tertiary kaolinites elsewhere in Australia (Bird and Chivas, 1989). Thus, erosional remnants preserving elements of pre-late Mesozoic and mid- to late Tertiary weathering are present in the Mount Lofty Ranges-Kangaroo Island area.

Deep weathering of the inland rocks is believed to be associated with development of the Cordillo Silcrete over the Eyre Formation, implying a post-Middle Eocene age (Wopfner, 1978). This would appear to be only partly true since silcrete formation need not be associated with deep weathering. For example, unweathered Pidinga Formation occurs only several metres below pedogenic and groundwater silcretes at Lake Anthony on the margin of the Eucla Basin. Similarly, unweathered Eyre Formation occurs immediately below silcrete in the southern Lake Eyre Basin. This suggests that the association of some silcretes with weathered profiles is fortuitous.

The evidence thus indicates that extensive weathering occurred prior to the Middle Eocene and may be as old as early Mesozoic. Deep weathering after the Late Eocene-Early Oligocene appears to be related to the formation of duricrusts.

FERRICRETE

Many ferruginous, mottled and pallid zones of deep weathering profiles in South Australia have been referred to as 'laterite', and the processes responsible for their formation as 'lateritisation'. Most of these occurrences, however, do not fit the definition of laterite (Bourman *et al.*, 1987) and, to avoid the genetic connotation in 'laterite', the term used here is 'ferricrete'.

Lithology, chemistry and mineralogy

Three broad categories of ferricrete are recognised, largely differentiated by bulk chemical composition and strongly influenced by the nature of the parent material (Fig. 10.38; Milnes *et al.*, 1987; Bourman *et al.*, 1987). These include ferricreted bedrock, ferricreted sediments and complex sedimentary and pedogenic ferricretes.

Ferricreted bedrock commonly can be identified by metamorphic and tectonic structures and relict texture and minerals inherited from the host rock. Ferricreted sediments display features which reflect the nature of the original sediment, such as skeleton grains, clasts of quartz and other rock fragments, and sedimentary structures. Ferricreted sediments may also include detrital ferricrete reworked from pre-existing ferricrete, sediment and weathered bedrock.

Complex ferricretes are largely composed of iron oxide-cemented pisoliths. Non-cemented pisoliths are common as ironstone gravel or as stonelines in soils. Nodular ferricretes contain rounded bodies larger than pisoliths, without multiple surface rinds, which are surrounded by a matrix similar in composition to the nodules. Slabby ferricretes, subhorizontal slabs of ferricrete up to several centimetres thick, consist of indurated masses of iron and aluminium oxides separated by softer clay-rich material.

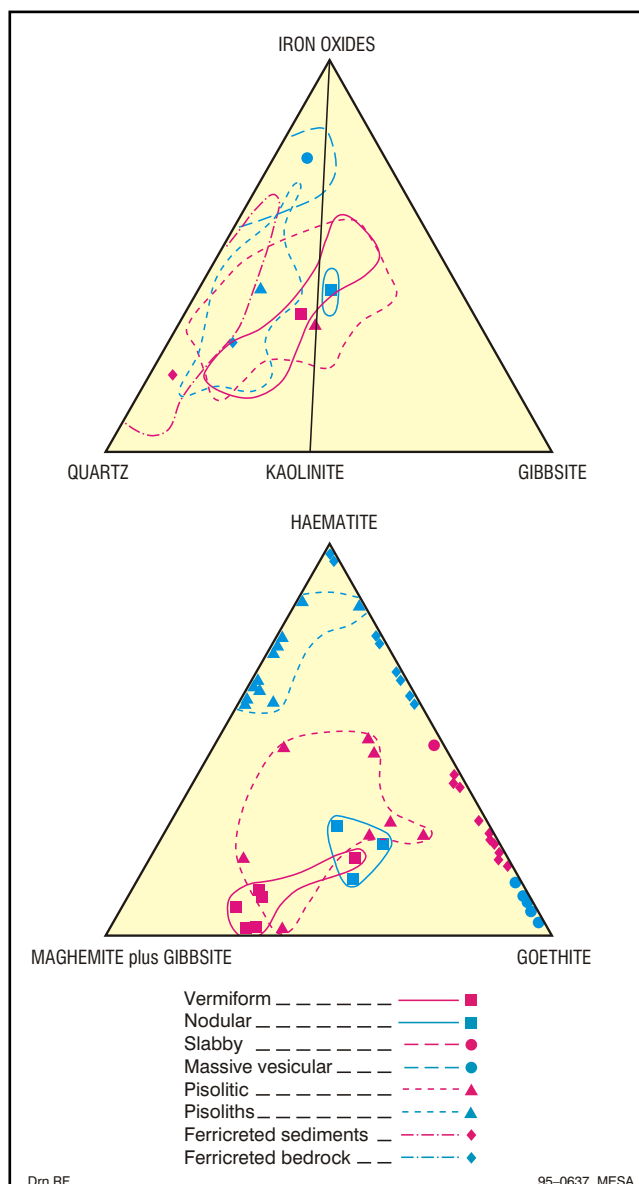


Fig. 10.38 Ternary diagrams showing composition of ferricretes.

Vermiform ferricretes display very distinctive light-coloured soft clay filling tubule structures penetrating the crusts.

Ferricreted bedrock has a chemical composition intermediate between Fe_2O_3 and SiO_2 . Ferricreted sediments have a broad range of chemical compositions varying from silica-rich to silica-poor types but, in contrast to ferricreted bedrock, are generally depleted in alumina. In marked contrast, complex ferricretes have a high alumina content, with roughly equal concentrations of SiO_2 and Fe_2O_3 . Pisolitic ferricretes have a wide compositional range, reflecting variations in chemical compositions of constituent pisoliths and the matrix material, which invariably contains clay and often gibbsite.

The main mineral constituents of ferricretes are goethite, haematite, maghemite and gibbsite (Fig. 10.38; Bourman *et al.*, 1987). There is a clear demarcation between haematite-rich ferruginised bedrock and goethite-rich ferruginised sediments; both are also free of gibbsite and contain only rare traces of maghemite. Differences in the relative proportions of haematite and goethite distinguish iron-impregnated clastic sediments from massive vesicular ferricrete, which contains high amounts of goethite.

Complex ferricretes contain greater oxide diversity that includes occasional boemite and corundum. Varieties can be distinguished on the basis of mineralogy and texture: vermiform and nodular types are significantly depleted in haematite relative to pisolitic ferricrete, which contains high amounts of haematite and maghemite. Formation of these oxides by surface heating of goethite by fire is probably responsible for the magnetic character of many surface pisoliths.

Geomorphic setting and distribution

Nodular, pisolitic, cellular, vermiform and massive ferricrete caps the plateau surface around the northeastern margin of the Eucla Basin and ferruginous gravel mantles lower subdued rounded rises. Similar ferricretes have been described to the north and northwest of the Eucla Basin (Jutson, 1934; Jackson and Van de Graaff, 1981). A weathering profile beneath degraded ferricrete near Observatory Hill shows deeply weathered Cambrian Observatory Hill Formation with iron-oxide mottling. Ferricrete is also extensive around the dissected margin of the Nullarbor Plain, along palaeochannels of the Eucla Basin and in and around playa lakes where it stands up to 5 m above lake bottoms and is commonly capped by pedogenic silcrete.

In northeastern South Australia, pisolitic ferricrete is extensively developed in the Doonbara Formation (Wopfner, 1967, 1974). Ferruginisation has occurred in the broad subdued synclines, where sediment accumulated adjacent to the exposed silcrete-capped Cordillo Surface (see below). Ferricreted sediment, widespread in the Mirackina Conglomerate (Barnes and Pitt, 1976; Pitt, 1976), allows the Mirackina Palaeochannel to be detected by LANDSAT imagery.

A widespread weathering surface (Karoonda Surface and Karoonda Palaeosol of Firman, 1966, 1973) is developed at the top of the late Tertiary coastal strand plain in the Murray Basin. The flat-lying to gently undulating surface is characterised by nodular ferruginisation and patchy silicification.

Ferricreted bedrock is preserved in the Mid-North on dissected remnants of a landsurface high in the relief between Gladstone and the northern Barossa Valley area. Weathering is greater and ferruginisation more extensive in the south (Alley, 1973, 1977). Fragments of ferricrete have been eroded and redeposited lower in the landscape, often cementing slope deposits and gravel above present valley bottoms.

A range of ferricrete types occurs on and below the level of the summit surfaces in southern South Australia (Bourman *et al.*, 1987; Bourman, 1993b). A landsurface of some relief in the ancestral Mount Lofty Ranges led to considerable areal variations in conditions of ferricrete formation. Bedrock in higher parts of the landscape was leached of iron, which was transported laterally to accumulate in lower lying areas such as in organic-rich swamps, at breaks of slopes on valley sides and in valley bottoms (Fig. 10.39).

Slabby ferricrete occurs at plateau margins where laterally moving groundwater neared the surface to cause precipitation and oxidation of iron oxides in near-horizontal layers. Elsewhere, ferrous iron was redistributed within weathered and partly kaolinised rock to accumulate as haematite-rich mottles. Progressive weathering and erosion led to exposure of the mottles at the surface where they hardened, disintegrated and formed lags. During downslope transport, surface rinds of goethite formed on haematite-cored clasts, including pisoliths and nodules.



Top: Ferricrete developed in Cambrian Observatory Hill Formation, capped by dark, ferruginous buckshot gravel, Observatory Hill area. Palaeomagnetic dating here indicates an Early to Middle Miocene age for the weathering. Photo 042513.

Centre: Palaeomagnetic dating here indicates an Early to Middle Miocene age for the weathering. Photo 042514.

Bottom: Slabby ferricrete at Chandler's Hill, Fleurieu Peninsula. Photo 042515.

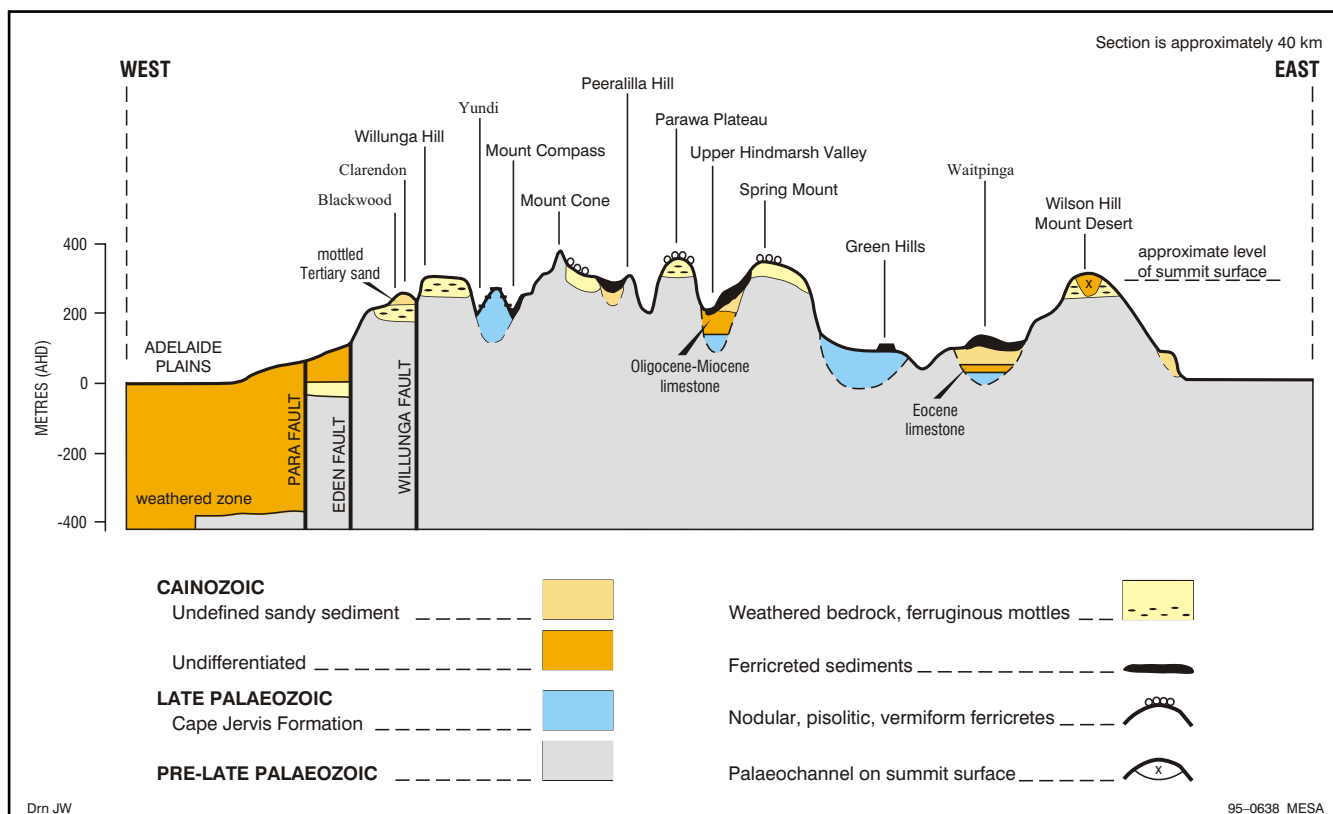


Fig. 10.39 Schematic cross-section across the Mount Lofty Ranges showing the relationships of different types of ferricretes to topography and geology (after Milnes et al., 1985).

Stratigraphy and age

In western South Australia, ferricrete developed on sediments of the Officer, Arkaringa and Eromanga Basins, and is locally present on extensive silcrete that partly formed in the Munjena Formation. The weathering profile near Observatory Hill has a minimum late Early-early Late Miocene or Early-Late Oligocene age based on palaeomagnetic dating, depending on which pole path is used (M. Idnurm, BMR, pers. comm., 1989). The early Middle Miocene Nullarbor Limestone unconformably overlies a ferricrete and iron-mottled Precambrian crystalline basement and Proterozoic or Palaeozoic sediments on the margin of the Nullarbor Plain; the ferricrete may predate the limestone, or may be a subsequent groundwater phenomenon.

Other ferricretes of the Eucla Basin region have formed in the Early Oligocene Hampton Sandstone and overlying Munjena Formation along the palaeochannels and in the Early-Middle Miocene Yarle Sandstone before a major phase of silcrete formation. Palaeomagnetic dating of ferricrete and iron mottles in the Lake Anthony region suggests a Late Miocene or latest Miocene-Pleistocene age (M. Idnurm, BMR, pers. comm., 1989).

In the northern Lake Eyre Basin, the ferricreted Doonbara Formation is overlain by the late Tertiary Cadelga Limestone, thus providing a minimum age for ferricrete development here. The Late Oligocene part of the compound Cannaway Profile in adjacent southwestern Queensland may be equivalent.

Ferricrete of the summit surface in the Mid-North has been assigned a pre- to early Tertiary age on the basis of ferruginous pisoliths in basal Tertiary sediments of the northern St Vincent Basin (Horwitz, 1961). This age was also favoured by Alley (1973, 1977), who regarded weathered and ferruginised bedrock buried beneath Clinton Formation in the Snowtown and Balaklava areas to be equivalent to portions of the downfaulted and eroded ferricreted summit surface. In the North Mount Lofty Ranges the ferricrete predates a silicified ?middle Tertiary landsurface.

The age of ferricrete in the South Mount Lofty Ranges has been long debated, partly reflecting the difficulties in demonstrating correlation with what are believed to be equivalent weathering profiles in bedrock buried in adjacent sedimentary basins. Some workers in this region have argued that there was once a continuous weathering profile over an extensive peneplain surface that has remained intact since the Mesozoic (Daily *et al.*, 1974; Maud, 1972; Twidale, 1976, 1983; Twidale and Bourne, 1975). However, some evidence suggests that a uniform blanket of ferricrete may not have formed over large areas, but rather that the different types of ferricretes and weathering zones developed in specific sites in response to local environmental conditions (Bourman *et al.*, 1987). Moreover, it is not necessary to presume the existence of a peneplain prior to ferricrete development because some ferricretes developed during landscape downwasting.

In some localities on Fleurieu Peninsula, such as at Spring Mount, Mount Cone and the Parawa area, the summit surface stood above the highest level reached by maximum marine transgression during the Tertiary, so potentially the surface may have been subjected to weathering and erosion during and since the Mesozoic. Some portions of the summit surface have been buried by Tertiary sediments that in turn were ferruginised; at a few of these localities the older weathered surface has been exhumed from beneath Tertiary sediments. Consequently the interpreted age of the summit surface could be extremely variable. It is probable that true relics of Mesozoic weathering persist only where they have been protected by burial.

Rocks of various ages within and marginal to the Mount Lofty Ranges have also been mottled and ferricreted. Ferruginisation of Eocene sediments has been regarded as evidence of post-Eocene 'lateritisation' (Rix and Hutton, 1953; Glaessner and Wade, 1958). Sediments of pre- and post-middle Tertiary age in the Barossa Valley (Harris and Olliver, 1965; Alley, 1973, 1977) and Pliocene age in the Happy Valley area (Olliver, 1967b), Hindmarsh Valley (Horwitz, 1960) and on the Green Hills Surface near Victor Harbor (Bourman, 1973) have been ferruginised. Late Pliocene marine

sediments at Ochre Point have been iron-impregnated (Ward, 1966), indicating post-Pliocene ferruginisation, whereas various Pleistocene deposits in the Willunga and Noarlunga Embayments have been weathered and mottled (May and Bourman, 1984).

In the Victor Harbor area, some late Quaternary alluvial sediments contain thin lenses and bands of iron oxide, indicating iron mobility in relatively recent times. In the present environment, iron oxides (largely as ferrihydrite) occur in creeks and seepage zones in many topographic settings in the region, including valley sides, groundwater salinas and peritidal environments (Ferguson *et al.*, 1984).

In the Murray Basin, the widespread Karoonda Surface and Karoonda Palaeosol are developed at the top of the Late Pliocene Parilla Sand (Firman, 1966, 1973). This is consistent with palaeomagnetic dating at 'Chowilla', which provides a Late Pliocene (2.5-3.4 Ma) age for ferricrete development (Bowler, 1980; An *et al.*, 1986).

In summary, the range in age of ferricretes may indicate either continuous weathering over extended periods of time, or repeated episodes of formation. Long periods of landscape stability may have provided the conditions suitable for extensive weathering and ferricrete formation. There have been at least three major phases of iron accumulation, two of them being possibly related to deep weathering. The oldest may have taken place in the early Mesozoic and certainly before the Middle Eocene. A second major phase appears to have occurred in the middle Tertiary, possibly between the Late Oligocene and Middle Miocene. The youngest is largely a phase of ferruginisation of sediments during the Late Miocene to Pleistocene. The last two occurred under very different climatic conditions and thus climate was probably not a major factor in ferricrete development.

SILCRETE

Silcrete is here regarded as secondary silica cementing sediments and relict soils or replacing skeletal or carbonate grains and forming sheets and layers more or less paralleling the original landsurface (Callen, 1983a).

Lithology

Although many forms of silica are present in silcrete, the fine-grained type dominates and is usually associated with anatase which can locally form up to 70% of the rock. Micro- to cryptocrystalline quartz and quartz overgrowth silcretes are better developed in porous and permeable rocks with low clay content, such as Eyre Formation of the Lake Eyre Basin. Chalcedonic and opaline varieties dominate less permeable rocks with abundant clay, especially montmorillonite (such as Bulldog Shale of the Eromanga Basin).

The silica is highly variable, but there are two main types of silcrete: groundwater silcretes which formed at depth in the sediments and retain sedimentary structures, and pedogenic silcretes which display vertical differences in structure and silica mineralogy (Wopfner, 1978; Callen, 1983a; Callen *et al.*, 1986; Thiry and Milnes, 1991). Both types are present in some profiles, with pedogenic forms occurring at the top. The thickness of the profiles and silcrete horizons are commonly 3-10 and 1-3 m, respectively (Fig. 10.40).

'Ant-nest' silcrete is one variety of pedogenic silcrete that is characterised by burrows resembling those of grass termites and meat ants (Callen *et al.*, 1986). Dissolution along burrows has resulted in variable enlargement, infilling by sediment, and deposition of silica and anatase. This silcrete passes laterally into narrow-columnar and massive forms (Callen *et al.*, 1986). 'Reed-mould' silcrete is characterised by numerous vertical tubes and strap-like impressions with longitudinal ribbing or striations. The moulds and impressions are believed to be of aquatic monocotyledons (Nichol, 1979), an interpretation supported by the recognition of tuberous root stocks (D. Greenwood, Smithsonian Institute, pers. comm., 1989), but in other areas the origin of the structures is uncertain.

A peculiar but common form of probable groundwater silicification (Krieg *et al.*, 1991) produces large botryoidal, ropy and lava-like masses. Its distinctive botryoidal form is believed to reflect internal growth layering and is characterised by microcrystalline quartz cement and quartz overgrowth on detrital grains.

Geomorphic setting and distribution

Silcretes have been variously named in different areas, regardless of whether they are correlative, and nomenclature tends to confuse discussion of areal distribution.

Pedogenic silcretes are very extensive over the central and northern parts of the State, whereas groundwater forms are generally confined to localised, topographically lower settings (Callen, 1983a). Pedogenic silcretes cap a variety of surfaces, generally of low relief and slope, but form distinctive tableland topography where dissected.

In the northern Lake Eyre Basin, the pedogenic Cordillo Silcrete rests on the Cordillo Surface, a palaeosurface believed to be a peneplain around the margins of the basin, passing basinwards into an extensive sediplain largely coincident with the top of the Eyre Formation (Wopfner, 1974, 1978). There is evidence, however, that the surface within the basin formed after erosion of the Eyre Formation and during a younger episode of deposition (Krieg *et al.*, 1991). The silcrete surface has been folded into broad synclines and anticlines (e.g. Cordillo and Gason Domes); erosion of these structures led to local relief inversion and the formation of plateaus,



Top: Thin, columnar pedogenic silcrete of the Carnadinna Surface developed in the Munjena Formation, Wilkinson Lakes area, Eucla Basin. Photo 042516.
Bottom: Botryoidal and ropy groundwater silcrete in the southern Lake Eyre Basin. Photo 042517.

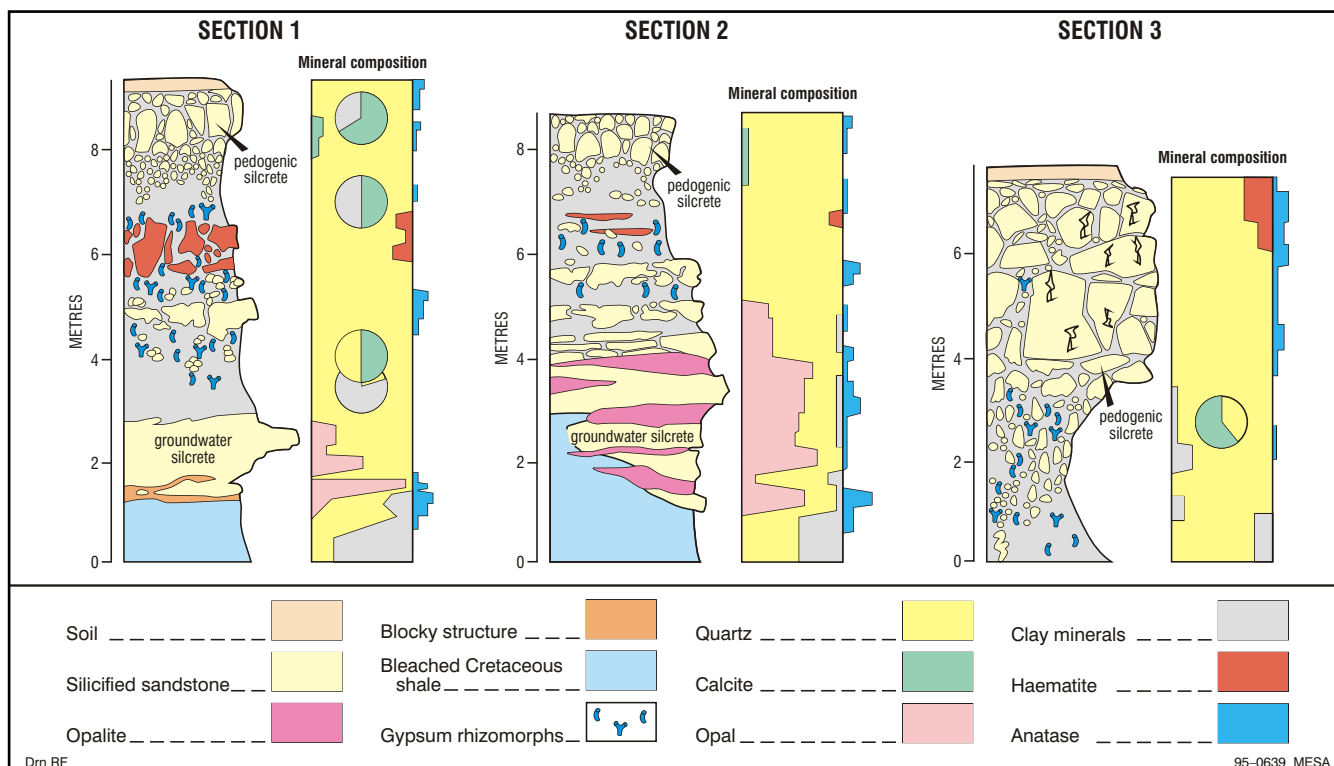


Fig. 10.40 Sections through silcrete profiles at Stuart Creek Opal Field, showing lithological and mineralogical variations (after Thiry and Milnes, 1991).

mesas, buttes and cuestas. Extensive pedogenic silcrete caps the strandplain of the Billa Kalina Basin but, unlike that of the Cordillo Surface, is unaffected by folding but a similar tableland topography has resulted from dissection. Around the margins of uplands such as the Flinders Ranges the silcrete is locally faulted or more strongly folded. On the western margin of the Lake Eyre Basin, pedogenic silcrete caps the Mirackina Palaeochannel now exposed high in the landscape as a chain of mesas and buttes. Silcrete also caps the Poole Creek Palaeochannel on the southern margin of the basin.

Silcretes occupying similar topographic positions to the Cordillo Silcrete occur around the northeastern margin of the Eucla Basin (Benbow, 1993). They mantle the dissected plateau, including the Carnadinna Surface, younger adjacent pediments, and aggradational and erosional surfaces of the palaeorivers; they appear not to have been folded and, on the broad scale, have not been topographically inverted. Silcretes, particularly the groundwater types, occur within palaeochannels and around the margin of the Nullarbor Plain; these have undergone little topographic inversion since their formation.

Pedogenic silcrete forms a prominent capping on dissected pediment surfaces flanking the Flinders Ranges and in the Arcoona Plateau and Tent Hill areas southwest of Lake Torrens (Twidale *et al.*, 1970; Hutton *et al.*, 1972). Silcrete is thicker and largely composed of silica cementing quartz grains on the lower parts of the pediments. In the scarp-foot zone, silcrete occurs as thin skins around quartzite blocks and infilling joints; here the silcrete matrix contains significant amounts of titania.

Pedogenic silcrete caps a dissected landscape in the Mid-North (Alley, 1973, 1977) and represents the most southerly occurrence in South Australia. Higher parts of the surface support silicified bedrock or regolith whereas lower portions are capped by much thicker and finer grained silcrete. Abutting the ridges, such as the Mount Misery Range, silcrete may occur as a capping on pediment surfaces or as silicified slope deposits. The silcrete surface occurs amongst, but topographically lower than, remnants of the dissected ferricrete surface. Silicified gravel, containing boulders of silcrete, occupies valleys cut in the silcrete surface but stands well above present valley bottoms.

Stratigraphy and age

Tertiary silcretes have formed in sediments ranging in age from Middle Eocene to Pliocene, in older cover strata, particularly of Jurassic to Early Cretaceous age, and on Precambrian rocks. Precise dating is difficult because the silicified sediments and overlying sediments commonly lack age control.

Extensive pedogenic Cordillo Silcrete formed in the Late Paleocene-Middle Eocene Eyre Formation of the Lake Eyre Basin and in the post-Eocene Munjuna Formation around the margin of the Eucla Basin. An Oligocene-Miocene age has been proposed for silicification, the Cordillo Silcrete having been folded prior to deposition of the Etadunna Formation and equivalents (Wopfner, 1974, 1978). This is consistent with a possible minimum age of Late Oligocene to Middle Miocene, based on dating of younger iron accumulations, for silcrete on the margin of the Eucla Basin. However, no widespread silcrete marks the unconformity between early and middle Tertiary sediments around the margins of these or other basins. There are a few exceptions such as the massive 'grey billy' nodular silcrete in Eyre Formation at 130 m depth in Poolowanna 1 from the northern Lake Eyre Basin and the silcrete developed beneath probable Namba Formation in the Callabonna Sub-basin.

The absence of silcrete at the unconformity in the northern Lake Eyre Basin has been interpreted to be due to silicification around the margins of the basin and over emerging domes concurrent with deposition of the upper Eyre Formation in the subsiding basin centre (Wopfner, 1974, 1978). Such an interpretation implies either a significantly younger age for the top of the Eyre Formation than is currently known, or that silicification occurred at the end of the Eocene. The latter interpretation would be consistent with the fact that in the Eucla Basin region pedogenic silcretes appear to be not associated with any of the major episodes of sedimentation, but rather with very thin contemporaneous deposits (Munjuna Formation).

'Reed-mould' silcrete in the Billa Kalina Basin is developed in the top of the ?Eocene Willalinchina Sandstone. However, the age of this silcrete is uncertain because it has not been shown to



Top: Moderately dipping silcrete developed on Early Cretaceous Parabarana Sandstone and early Tertiary Eyre Formation east of Parabarana Hill, northern Flinders Ranges. Photo 042418.



Steeply dipping silcrete developed in the Eyre Formation east of Mount Terila, northern Flinders Ranges. Photo 042419.

underlie the ?Miocene Watchie Sandstone which itself is capped by silcrete (see 'Billa Kalina Basin'). In this basin, the possibility exists that all silcretes formed after the Miocene.

In the Mid-North, pedogenic silcrete post-dates dissection of the ferricrete surface. There is no stratigraphic evidence for the age of the silcrete except that probable late Tertiary gravel contains silcrete boulders and overlies the etch surface developed by erosion of the pedogenic silcrete surface. The gravel itself contains layers of groundwater silcrete. Based on this evidence, and the distribution and geomorphic relationships of the pedogenic silcrete to early Tertiary strata in the northernmost St Vincent Basin, the pedogenic silcrete may have formed between Middle-Late Eocene and Miocene (Alley, 1973, 1977). As in the Lake Eyre and Eucla Basins, however, no silcrete is developed within the basin sediments, except for a localised thin sheet of groundwater silcrete within the Late Eocene to Early Oligocene Clinton Formation in the Bowmans area.

Middle to late Tertiary sediments that are locally capped by pedogenic silcrete include the Etadunna and Namba Formations of the Lake Eyre Basin. The 'ant-nest' silcrete capping the Watchie Sandstone of the Billa Kalina Basin is regarded as syndepositional, implying a ?Miocene age (Ambrose and Flint, 1981). Silcretes

in palaeochannels of the Eucla Basin that cap the dissected Late Eocene part of the Hampton Sandstone and overlying Munjerna Formation post-date iron-enriched weathering profiles dated palaeomagnetically as probably Late Miocene to Pleistocene. In the Murray Basin, pedogenic silcrete formed at the top of the Late Pliocene Parilla Sand as part of the Karoonda Surface and Karoonda Palaeosol (Firman, 1966, 1973). A Late Pliocene age for the silcrete is confirmed by palaeomagnetic studies (An *et al.*, 1986).

Groundwater silcretes are widely developed in middle and late Tertiary sediments. In the Eucla Basin they occur within and post-date ferricrete in the Early-Middle Miocene Yarle Sandstone and have affected the ?Early Pliocene Garford and Narlabby Formations of the adjacent palaeochannels.

The less common 'puddingstone' variety of silcrete (Wopfner, 1978) is developed in Miocene Namba Formation (Callen, 1983a). In the Lake Eyre Basin this silcrete has been related to the gypcreted Warrina Surface, regarded as probably Late Pliocene (Wopfner and Twidale, 1967; Wopfner, 1978). On the uplands of the Stuart Range, such silcrete is associated with the lower gypseous unit of the Russo Beds (Barker *et al.*, 1979; Barker, 1980) and in the Doonbara Formation (Pitt, 1978).

Although the timing of opal formation is not clear, it is commonly correlated with deep weathering (Hiern, 1967; Carr *et al.*, 1979; Barnes *et al.*, 1992), the formation of 'grey billy' silcrete (Barnes and Scott, 1979; Nichol, 1979) or with 'puddingstone' (opaline) silcrete (Wopfner, 1978). Alternatively, Barker (1980) suggested that opal formed penecontemporaneously with deposition of the Early Cretaceous Bulldog Shale.

A simple stratigraphic approach to determining the age of silcretes, however, may have problems. Detailed analyses of silcretes in the Stuart Creek opal field by Thiry and Milnes (1991) indicated that pedogenic and groundwater silcretes may or may not relate to one broad period of silicification. The superposition of silcrete layers at any one site need not imply different periods of silcrete formation but rather the varying position of groundwater table levels during landsurface evolution. Stacked groundwater silcretes are believed to have formed at different stratigraphic levels in response to successively deeper levels of watertable during landsurface rejuvenation. Thus, in such situations, deeper groundwater silcrete layers may actually become progressively younger than the overlying pedogenic silcrete.

In summary, there appear to be two significant phases of pedogenic silcrete development. The older phase may have occurred between the Late Eocene and Middle Miocene, an interval during which widespread silcrete formed in the interior of the continent. The second major phase occurred in the Late Miocene to Pleistocene and may also have involved significant groundwater silicification.

Economic geology

Duricrusts and weathered rocks are economically important. Silcrete has been used as a marker horizon to map structures in the Cooper Basin area (Wopfner, 1974). Opal is the basis of a multi-million dollar industry in this State, with the estimated value of production of \$41.3 million/year averaged over the period 1978-90 (Barnes *et al.*, 1992).

Industrial clay, another product largely of Tertiary weathering, has a wide range of important applications particularly in brickmaking and refractories, with the main demand being in the Adelaide metropolitan area. The two important sources of clay are weathered bedrock within and along the western margin of the Mount Lofty Ranges, and clay lenses in the North Maslin Sand (Pain and Keeling, 1990). Weathered bedrock is the most important source of raw materials for the brick industry, principally from Precambrian Saddleworth Formation and, to a lesser extent, Woolshed Flat Shale. Clay is mined from weathered bedrock in the southern Barossa Basin and from North Maslin Sand in the Golden Grove and Willunga Embayments of the St Vincent Basin. Current mining tenements near Adelaide contain resources totalling 32 Mt of weathered bedrock and 15 Mt of Tertiary clay. Golden Grove is the most important source of both types of material.

GEOLOGICAL HISTORY AND PALAEOCLIMATE

M.C. Benbow, N.F. Alley, R.A. Callen and D.R. Greenwood

This section of the chapter is a synthesis of the Tertiary geological record presented in a palaeogeographic and palaeoclimatic framework that incorporates the broader events affecting Australia, Antarctica and the Southern Ocean.

The general character of the early Tertiary Australian vegetation and the evidence it provides for palaeoclimate can be inferred from palynofloral and, to a lesser extent, macrofloral evidence. However, interpretations of past vegetation from fossil palynofloras suffer from a number of problems including uncertainties about affinities with modern species, the environmental tolerances of fossil taxa and the lack of quantitative data on fossil and modern pollen-spore spectra. In general, all the palaeoclimatic interpretations suffer from the lack of a complete record or from inadequate dating of the terrestrial record.

CRETACEOUS-TERTIARY BOUNDARY: EARLY PALEOCENE (65-60 MA)

At the beginning of the Tertiary, southern Australia lay in high latitudes (65-66°S) adjacent to a nascent, narrow Southern Ocean. There is no record in South Australia of marine transgressions during the earliest Tertiary (McGowran, 1991) and the Cretaceous-Tertiary boundary is an unconformity but the hiatus is of shortest duration in the offshore basins (Figs 10.2, 10.41).

The lack of marine sedimentation is in part a reflection of the huge eustatic fall in sea level from its maximum in the Maastrichtian. However, because of problems in defining the boundaries of palynostratigraphic zones, sediments assigned to the largely Late Cretaceous *Forcipites* (ex *Tricolpites*) *longus* Zone in the offshore basins may range into the Danian (earliest Paleocene). Thus, the age of the upper part of the Late Cretaceous Potoroo Formation in the Bight and Duntroon Basins and the upper part of the Sherbrook Group in the Otway Basin may extend into the earliest Paleocene.

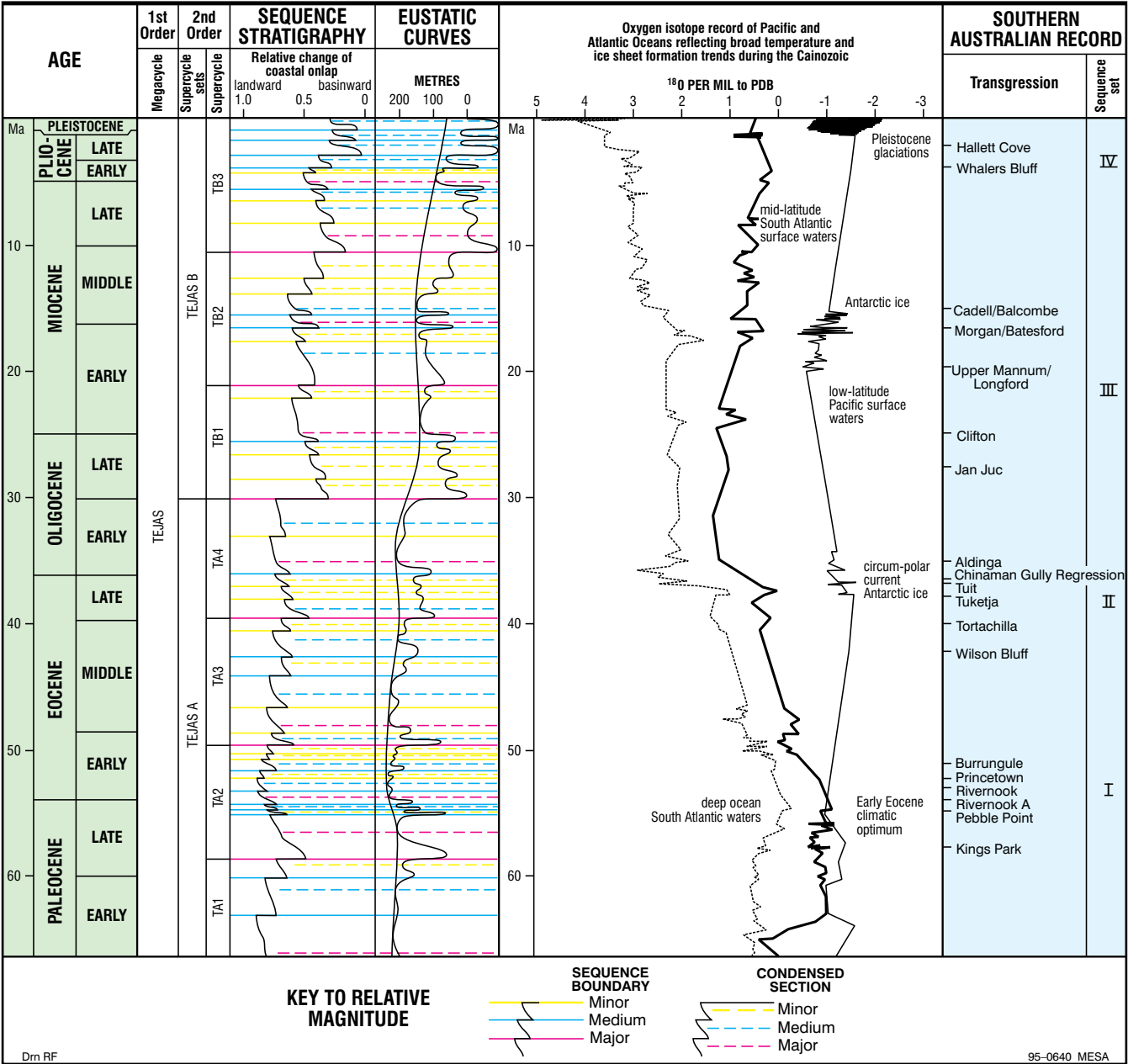


Fig. 10.41 Tertiary sea-level curves, sequence stratigraphy (after Haq et al., 1987b), oxygen isotope curves (after Shackleton, 1984) and South Australian sea-level events (McGowran, 1989; McGowran in Feary et al., 1994).

Little is known about the topography at this time although the ancestral Mount Lofty and Flinders Ranges appear to have been subdued. Widespread plains existed in the interior and extensive deep weathering that commenced during the Mesozoic may have continued during the early Tertiary.

There is no onshore record in South Australia for the nature of the vegetation. However, palynological evidence from dredge samples in the Great Australian Bight suggests that vegetation in the south-central part of the continent was dominated by temperate (perhaps cool temperate) coniferous rainforests (Alley and Clarke, 1992). Like the great fall in global sea level, temperatures also had dropped from their high in the Cretaceous. This factor, along with southern Australia's position in high latitudes at this time, promoted the expansion of (cool) temperate rainforests. Podocarps were common, particularly the cool temperate species *Lagarostrobos* (= *Dacrydium*) cf. *L. franklinii* or huon pine, *Dacrydium* Group B spp., *Microcachrys* and *Podocarpus* spp., along with the araucarians. The forest association contained a well-developed understorey of ferns and mosses. Tree ferns were the most common and probably included species of Cyatheaceae and Dicksoniaceae along with *Osmunda*, Polypodiaceae, clubmoss, *Gleichenia* and peat moss (*Sphagnum*).

Flowering plants included a number of species with affinities to extant genera in montane and/or lowland rainforest in temperate to tropical areas, such as *Gevuina*, and/or *Hicksbeachia*, ?*Knightsia* and *Macadamia*. The pollen of the rainforest taxa *Beauprea*, *Ilex*, *Trimenia* and *Gunnera* are recorded in very low frequencies. Rare pollen of *Nothofagus* are from the *Fuscaspora* group which is currently confined to cool temperate rainforests in Tasmania, New Zealand and Chile.

Thus, although evidence indicates that cool temperate rainforest was widespread in the southern part of South Australia, the flora includes minor frequency taxa of more tropical aspect. This mixture of apparently tropical through to subalpine plants growing in a lowland plain environment poses a dilemma in palaeoecological interpretation. However, the ecological tolerances of some or all of the species may have changed with time, and there may be no modern analogue for the fossils and therefore the fossil pollen with affinities to extant plants of megathermal aspect may have occupied different ecological niches during the early Tertiary. This problem is also true of palynofloral interpretations from the Paleocene up into the middle Tertiary.

LATE PALEOCENE-EARLY EOCENE (60-53 MA)

During the Late Paleocene and Early Eocene, fluvial and marginal marine, quartz-rich, terrigenous sediments were deposited in some interior basins and along the continental margin (Fig. 10.42). The volcanogenic zone which was active along the eastern part of the continent during the Mesozoic was no longer a major source of volcanic detritus (Veevers, 1984). Late Paleocene terrestrial sedimentation (Warina Sand) in the Murray Basin suggests blocking of westerly drainage from the eastern Australian highlands to the Eucla Basin by incipient uplift of the Flinders and Mount Lofty Ranges. This interpretation is supported by the deposition of very thin early Tertiary terrigenous sediments over thick Cretaceous sediments in the Eucla Basin, suggesting reduced sediment supply.

Although much of South Australia was undergoing erosion, the possible commencement of downwarping in the Lake Eyre Basin led to deposition of an extensive sand sheet in braided stream systems and finer grained, carbonaceous sediments in restricted swamps and lagoons. Similar sedimentary environments were confined to deeper parts of the Murray Basin. The absence of thick, extensive, fluvial conglomeratic facies and the lack of piedmont deposits flanking the ranges indicate subdued terrain. Recycled resistant quartz-rich lithologies at the base of the Eyre Formation in the Lake Eyre Basin indicate erosion of pebble lags from a stable, weathered landsurface. Colluvial to residual pebble-bearing

sandstone may have been deposited on the Stuart Range and in the Billa Kalina Basin area.

At least five successive marine transgressions from the Indian Ocean penetrated along the opening southern continental margin (Fig. 10.41; Deighton *et al.*, 1976; Davies *et al.*, 1989; McGowran, 1991). The transgressions were characterised by restricted, marginal marine planktonic and arenaceous foraminifera (*Cyclammina* biofacies). The oldest transgression recognised in South Australia (Pebble Point Transgression) is within the Pebble Point Formation in the Gambier Basin (McGowran, 1991).

Palynofloral evidence indicates widespread rainforest over much of the continent which, together with depositional environments in the Lake Eyre Basin, implies a very wet climate (Truswell and Harris, 1982; Sluiter, 1991; Macphail *et al.*, 1994; Alley and Sluiter, in prep.).

Data interpreted from Harris (1965) suggest that Late Paleocene vegetation along the south-central margin of the continent, like that of the earliest Paleocene, was dominated by coniferous rainforests in which podocarpaceous types were common, including high-rainfall, cool-temperate species. The araucarians and diverse proteaceous forms were also well represented but *Nothofagus* was uncommon, although a number of species of the *Brassospora* and *Fuscaspora* groups were present. The forest contained a well-developed understorey of moisture-loving ferns and mosses.

Quantitative data (Sluiter, 1991; Alley and Sluiter, in prep.) and general descriptions of palynofloras (Wopfner *et al.*, 1974) from the lower part of the Eyre Formation in the Lake Eyre Basin indicate that vegetation in the northern part of the State was significantly different from the south. The vegetation mosaic within the basin varied from angiosperm-dominated (usually Cunoniaceae) with modern temperate rainforest affinities to gymnosperm-dominated and mixed angiosperm-gymnosperm rainforest assemblages, all with a diverse fern and moss understorey. This implies that northern South Australia was warmer than the south, an interpretation supported by the presence of pollen with affinities to a number of modern subtropical to tropical (megathermal) plants, including *Anacalosa*, Cupanieae, *Beauprea*, *Santalum*, *Ilex* and common Trimeniaceae.

Since the Eyre Formation was deposited on an extensive flood plain containing swamps, shallow lakes and peat bogs, the environment was ideal for the spread of *Sparganium* (bur-reed), the first Australian record of which is from the Eyre Formation, and the herb *Gunnera* which currently grows in wet, boggy conditions. Areas of open water are indicated by abundant freshwater microplankton.

Similar rainforests may have persisted through the Early Eocene, although vegetation is interpreted to have become a Myrtaceae-dominated complex notophyll vine forest in the Lake Eyre Basin (Sluiter, 1991). However, some drillhole data in the Lake Eyre and Torrens Basins indicate continuing dominance of Cunoniaceae in some localities, but Myrtaceae in others (Alley and Sluiter, in prep.). Likewise, the Casuarinaceae (probably *Gymnostoma*), araucarians and podocarps, although well represented in some parts of the basins, were less common elsewhere, suggesting a variable rainforest mosaic.

Plants of megathermal aspect were still present along with palms, Euphorbiaceae and *Gevuina/Hicksbeachia* of the Proteaceae. The proteaceous component of the vegetation is diverse and includes the sclerophyllous *Xylomelum occidentale* which, along with the sun-loving Malvaceae, suggests drier and/or more open parts of the forest existed. This interpretation is supported by a great decline in the importance of fern spores which are locally common at some sites in the Lake Eyre Basin but rare in the Torrens Basin. Data interpreted from Harris (1965) in the Gambier Basin indicate a high frequency of gymnosperm pollen with temperate to cool-temperate rainforest affinities which is in agreement with the more cool-temperate aspect of the vegetation.

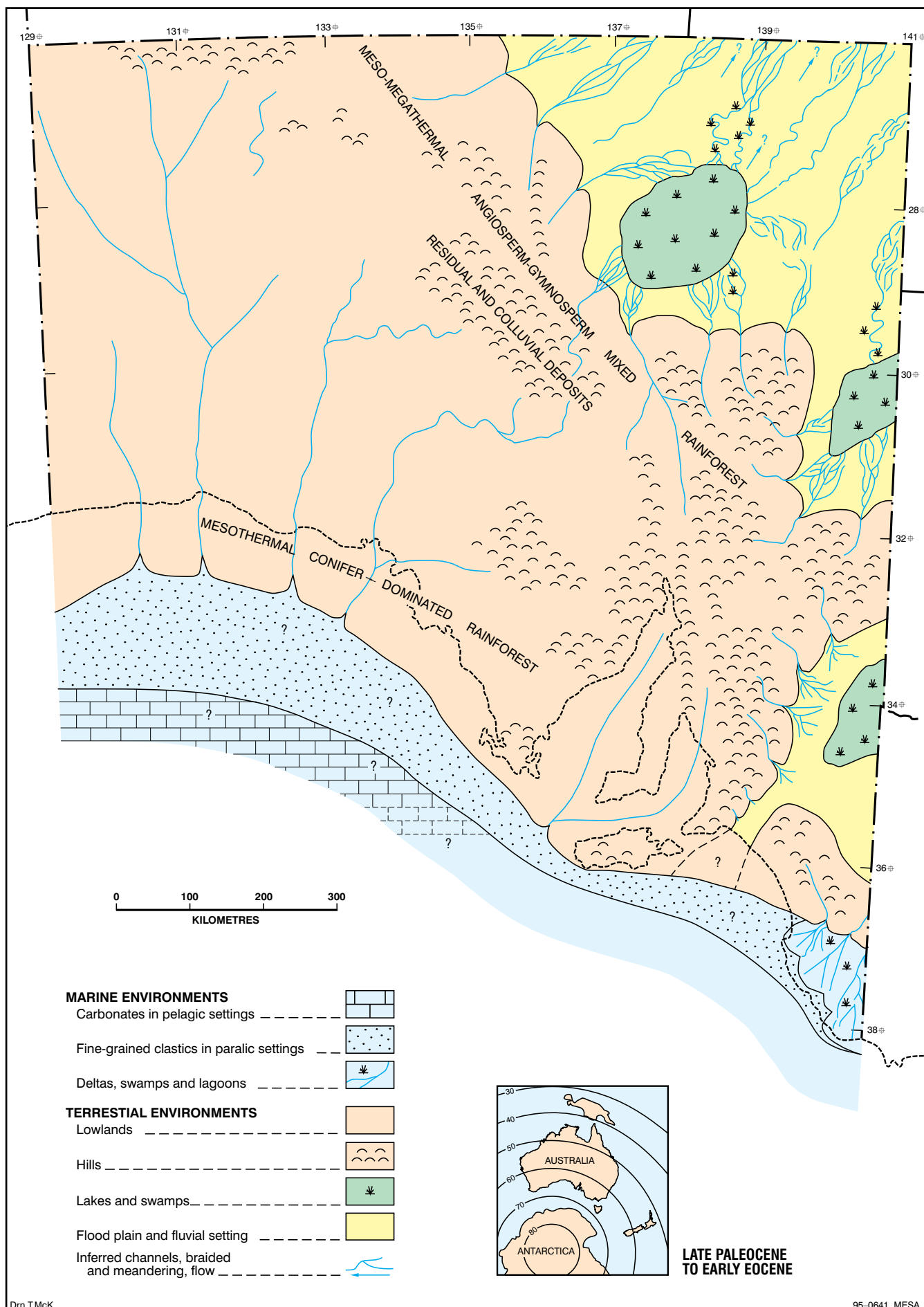


Fig. 10.42 Late Paleocene-Early Eocene palaeogeography of South Australia.

EARLY TO MIDDLE EOCENE (53-44 MA)

There is almost no record of sedimentation during the late Early to early Middle Eocene (McGowran, 1989), the time of the first major cooling in the Tertiary as indicated by an increase in oceanic $\delta^{18}\text{O}$ (Fig. 10.41; Shackleton and Kennett, 1975). Cooling is also inferred from oceanic micropalaeontological evidence (McGowran and Beecroft, 1985).

The only palynofloral evidence for the character of Early Eocene vegetation and climate comes from the Gambier, Lake Eyre and Torrens Basins. Although sedimentation continued in the Murray Basin, the facies are not suitable for the preservation of palynomorphs. Data from the other areas, however, show that conditions established in the earliest Eocene persisted through the Early Eocene (Macphail *et al.*, 1994).

MIDDLE TO LATE EOCENE (44-36 MA)

Southern Australia was at relatively high latitudes (50-55°S) when widespread deposition began in the late Middle Eocene. The Southern Ocean had widened significantly, allowing extensive flooding of the continental margin during at least three transgressions correlated with eustatic sea-level rises (Johnson and Veevers, 1984; McGowran, 1989, 1991; McGowran *et al.*, 1992).

Uplift of the ancestral Mount Lofty and Flinders Ranges and downfaulting within and adjacent to the ranges probably resulted from compressional reactivation of meridional to northeast-trending Delamerian structures. Deposition of terrigenous, non-marine and sometimes paralic sediments began in the St Vincent, Pirie and Torrens Basins adjacent to the uplands and in the intramontane Willochra and Walloway Basins.

Deposition which commenced in the Paleocene in the Lake Eyre Basin, and perhaps in the Billa Kalina Basin, continued in the Middle Eocene. The later phase, however, was interrupted by an hiatus in the Early Eocene in the Lake Eyre Basin and by uplift and erosion along the Stuart Range-Billa Kalina Basin axis. This was probably the cause of a reversal of drainage direction in the Billa Kalina Basin from channels flowing towards the Eucla Basin to those flowing northwards to the Lake Eyre Basin. The appearance of recycled Cretaceous palynomorphs in the Middle Eocene part of the Pidinga Formation in the Eucla Basin may be coincident with uplift and erosion in the headwaters of the palaeochannels (Alley and Beecroft, 1993). This phase of deposition in some areas and erosion in others suggests widespread tectonic movements, especially along the Torrens Hinge Zone.

At commencement of the Wilson Bluff Transgression in the late Middle Eocene, deposition of extensive neritic carbonates began in platform settings around the continental margin (Fig. 10.43). Bryozoal meadows thrived on the platform while pelagic limestone accumulated in deeper water; clastic sediments were deposited in coastal and paralic environments.

Terrigenous sedimentation in the west was confined to the margin of the Eucla Basin and its palaeodrainage system, and estuarine conditions migrated up into the channels. Seaward migration of a high-energy, siliciclastic shoreline and the formation of an extensive coastal dune complex marked the end of deposition on the emergent part of the Eucla Basin, possibly in the Late Eocene (Benbow, 1990a). This may be part of the Aldinga Transgression which may correspond to a warming of oceanic water along southern Australia (Benbow, 1990a; McGowran, 1986). Moderate intensity prevailing westerly winds are inferred from the formation and orientation of coastal dunes in the Eucla Basin (Benbow, 1990a).

Terrigenous sediment throughout the eastern coastal basins was mostly derived from the eastern Australian highlands. In the Murray Basin, deposition of bryozoal limestone, confined to the southwestern region, commenced in the Late Eocene. The Padthaway Archipelago acted as a barrier to open marine conditions in the eastern Murray Basin (Brown and Stephenson, 1991). In the

Gambier Basin, neritic bryozoal limestone was not deposited until the latest Eocene or earliest Oligocene.

Widespread rainforest persisted over much of the continent. A major floral change occurred with *Nothofagus* and conifers becoming important, whereas proteaceous types dwindled in frequency (Macphail *et al.*, 1994). However, the relatively high frequencies of pollen with affinities to temperate to mesothermal rainforest types, along with megathermal rainforest elements, still represent a dilemma to palaeoclimatic interpretation. This palynofloral association occurs throughout the South Australian Middle to Late Eocene sedimentary basins, even in areas where there is no evidence for any uplands high enough to have supported cool temperate rainforest. Thus, the vegetation may have been a rainforest mosaic with no modern analogue. In southern South Australia, the rainforest in many localities included a dense understorey of ferns and mosses.

Middle Eocene palynofloras from the Torrens and Lake Eyre Basins indicate that the structure of the vegetation continued to differ from that of the southern continental margin. Pollen of rainforest types such as *Nothofagus* and the Podocarpaceae generally form only a minor part of the spectra. Because *Nothofagus* produces vast quantities of wind-transported pollen, it is commonly over-represented in modern and fossil palynofloras. The low frequency of *Nothofagus* pollen may imply that the genus formed only a very minor part of the vegetation or, more likely, that the pollen was transported considerable distances from the ancestral Flinders and Willouran Ranges.

The Myrtaceae are relatively common with up to 2% *Eucalyptus* pollen and, in the Torrens Basin, up to 10% *Myrtaceidites parvus*. The Cunoniaceae are very patchily distributed, with the highest recordings in a few drillholes in the southwestern Lake Eyre Basin but are virtually absent from the Torrens Basin. Although pollen of the Proteaceae are numerically low, species diversity is relatively high. The subaquatic families Cyperaceae, Sparganiaceae and Restionaceae are locally common, indicating the presence of wet areas in the basins. Although grass was established in the interior its pollen was very minor.

The low frequency of pollen from rainforest plants, in association with high Casuarinaceae (highest in the Torrens Basin) and diverse and reasonably well-represented proteaceous pollen, suggests restriction of rainforest to moister valley floors while more sclerophyllous vegetation dominated the drier hinterland.

Middle Eocene macrofloras in the North Maslin Sand (St Vincent Basin) contain abundant and diverse leaves, fruits and flowers including *Lygodium* (a climbing fern), the conifers *Podocarpus* and *Retrophyllum* (Podocarpaceae), *Brachychiton* (Sterculiaceae; bottle trees or kurrajong), *Banksiaephyllum* (affinities to *Banksia* and/or *Musgravea*), *Myrtaciphyllum* (Myrtaceae; affinities to *Syzygium* or lillypilly), and the families Araliaceae (umbrella trees), Elaeocarpaceae (quandong), Lauraceae and Arecaceae (palms; Maslin Bay only). The majority of leaves have affinities with extant plants in mesothermal-megathermal rainforests (Lange, 1970, 1982; Christophel and Blackburn, 1978; Blackburn, 1981; Christophel and Greenwood, 1989; Greenwood, 1994), consistent with the presence of epiphyllous fungi on leaf cuticles (Lange, 1969). *Nothofagus* pollen, a high-rainfall indicator, is abundant in Middle to Late Eocene sediments, but macrofloral remains are absent.

Macrofloras of the Lake Eyre and Billa Kalina Basins contain abundant conifer shoots with affinities to the families Araucariaceae, Cupressaceae (*Papuacedrus*) and Podocarpaceae (*Dacrycarpus* and *Dacrydium*; Offler, 1969, 1984; Callen *et al.*, 1986). Rare fruits of *Eucalyptus* are probably the earliest record of this genus (Lange, 1978, 1982). *Banksiaaeformis* (Proteaceae), *Brachychiton* (Steruliaceae), abundant leaves of Myrtaceae (*Eucalyptus* and/or *Melaleuca*), and cones of *Gymnostoma* and *Casuarina* are also present (Greenwood *et al.*, 1990; Christophel *et al.*, 1992).

The mean size and morphology of leaves of macrofloras in the North Maslin Sand are characteristic of mesothermal-megathermal

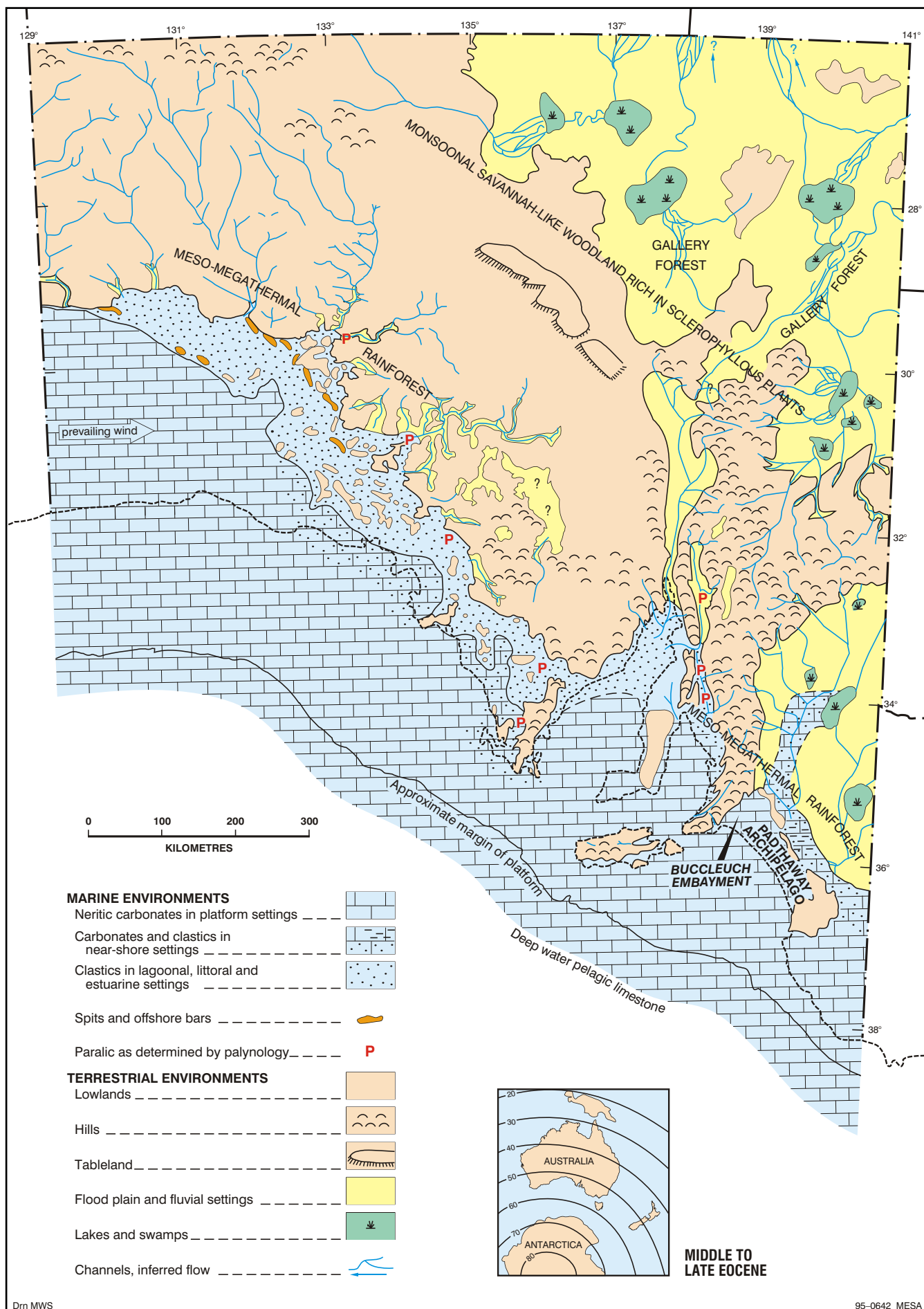


Fig. 10.43 Middle to Late Eocene palaeogeography of South Australia.

rainforests (Christophel and Greenwood, 1989; Greenwood, 1994), whereas the same characteristics of the silcrete floras suggest the presence of rainforest and sclerophyllous vegetation (Greenwood *et al.*, 1990; Greenwood, 1994). This association might imply a monsoonal inland climate, with rainforest plants forming gallery forests along permanent water courses. Alternatively, the presence of sclerophyllous leaves may indicate areas of low soil fertility in a humid climate (Lange, 1980, 1982; Christophel and Greenwood, 1988, 1989; Greenwood *et al.*, 1990). Thus, by the Middle Eocene there is some evidence from the palynofloras and macrofloras for latitudinal climatic zonation in southern Australia. Based on the correlations between leaf morphology and climate in modern vegetation, the Middle Eocene macrofloras indicate mean annual temperatures of ~18°C in the south (e.g. Golden Grove) and ~21°C in the southern Lake Eyre Basin (Greenwood, 1994).

Additional evidence for palaeotemperature during the Middle-Late Eocene is provided by the marine platform limestones typical of modern temperate-water limestones being deposited south of the 18° isotherm at about 30°S latitude (Lees and Buller, 1972). This apparent contradiction with the botanical record may reflect lowering of ocean salinity in a region of high rainfall. For example, extensive temperate-water carbonate (foramol) assemblages occur offshore from wet, tropical southeast Asia (Lees, 1975; Nelson, 1988). However, similar limestones continued to be deposited in the southern Australian marine basins after a major cooling and relatively arid phase around the Eocene-Oligocene boundary.

The presence of the foraminifera *Halkyardia*, *Linderina* and *Hantkenina* is evidence of the southward expansion of warm ocean waters (McGowran, 1979). The rare occurrence of *Truncorotaloides spinulosus* in the unnamed basal glauconitic member of the Wilson Bluff Limestone is interpreted as an important indicator of warm-water influence (McGowran and Lindsay, 1969). On the other hand, the environment was not tropical because of the rarity of the keeled *Globorotalia*-like tests, the relative abundance of spinose, acarininid tests, and the absence of true tropical indicators. The very low diversity of larger foraminifera, compared to their high diversity in lower latitudes during the Middle to Late Eocene, indicate marginally warm or temperate water along the southern Australian coast.

Oceanic temperatures are interpreted as significantly lower later in the Eocene, although short-lived warmer episodes were superimposed on the overall cooling trend (Fig. 10.41; Shackleton and Kennett, 1975; Shackleton, 1984, 1986a) and can be matched with several of the later Eocene transgressions (McGowran, 1989; McGowran *et al.*, 1992). The cooling trend is reflected in the oxygen isotope record and probably indicates growth of the Antarctic icesheet during this time (Barrett *et al.*, 1987; Mathews and Poore, 1980).

EOCENE-OLIGOCENE BOUNDARY AND THE EARLY OLIGOCENE (36-30 MA)

Major changes in oceanic biota around the Eocene-Oligocene boundary have been referred to as the 'Terminal Eocene Event'. However, the boundary does not seem to mark an abrupt major evolutionary break but is part of an episodic or gradual change which started in the Middle Eocene and continued into the earliest Oligocene (Pomeroy and Premoli-Silva, 1986). This conclusion is supported by accelerated overturn of much of the marine biota, a significant decrease in temperature, and development of the modern two-layered structure of the oceans. Oxygen isotope studies show that this climatic change followed rather than preceded the biostratigraphic events, thus precluding a causal relationship (Shackleton, 1986a).

In the St Vincent Basin, foraminiferal studies show that there was no catastrophic change in the local inner-neritic community (Lindsay and McGowran, 1986). The rate of arrival of new species appears to have increased during each of the Eocene transgressions whereas the greatest increase in disappearances occurred at the Chinaman Gully Regression.

Although widespread deposition continued into the Early Oligocene in the Murray, Gambier, St Vincent, Pirie and Eucla Basins, the later Early Oligocene stratigraphic record in South Australia is relatively restricted. Deposition of carbonates continued in the St Vincent and offshore basins; in the Murray Basin, carbonates passed landwards to fine-grained fluviolacustrine sediments.

Palynological evidence from the Murray, St Vincent and Pirie Basins shows continued decline in angiosperm diversity matched by increasing dominance of *Nothofagus* and *Gymnostoma*. *Nothofagus* pollen was dominated by the *Brassospora* group with minor *Fuscaspora* and rare *Lophozonia*. Rainforest gymnosperms, largely of cool temperate aspect, occur with lower frequency than earlier in the Tertiary, although a single taxon may be relatively high in some palynofloras. The Myrtaceae and Proteaceae are greatly diminished and Cunoniaceae is very rare or absent in many palynofloras. Ferns and mosses formed a minor part of the vegetation. The semiaquatic Cyperaceae, Sparganiaceae and Restionaceae are consistently present, being relatively common in the northern St Vincent Basin and northwestern Murray Basin; they indicate extensive marsh conditions.

Thus, in the southern part of the State at least, latest Eocene to Early Oligocene vegetation was largely *Nothofagus*-dominated, cool temperate rainforest. This floral change was probably related to a global cooling trend (see above).

The end of deposition in the late Early Oligocene coincided with a fall in global sea level of ~125 m and a phase of icesheet development in Antarctica (Haq *et al.*, 1987a; Shackleton, 1986a). In the deeper Gambier Basin, however, deposition was continuous across the Early-Late Oligocene boundary (Lindsay, 1985).

Extensive silcrete on the Cordillo Surface and in the southwestern part of the State may have formed during the Oligocene.

LATE OLIGOCENE-MIDDLE MIOCENE (30-14 MA)

The Early Oligocene regression was followed by a rapid transgression which initiated deposition of widespread neritic limestone on the platforms, in newly formed embayments in the St Vincent and western Murray Basins, and in a number of small intramontane basins of the southern Mount Lofty Ranges area (Fig. 10.44). At least six transgressive episodes are marked by Late Oligocene intercalations of glauconitic Ettrick Formation within the marginal marine Geera Clay of the Murray Basin and by Early to Middle Miocene horizons containing warm-water, extra-tropical, large foraminifera, *Lepidocyclina* and *Austrorillina*-*Marginopora*-*Flosculinella*.

Significant warming of the sea is interpreted from the decrease in oceanic and shelf $\delta^{18}\text{O}$ (Dorman, 1966; Shackleton and Kennett, 1975; Shackleton, 1986a). The skeletal composition of the Nullarbor Limestone and the more widely spread and diverse larger foraminifera along southern Australia indicate that early Middle Miocene shelf seas were warmer than at any time in the Tertiary (McGowran, 1979; Adams *et al.*, 1990). Warm-water coralline algal *Archaeolithothamnion* facies were widespread in the Eucla, Pirie and Murray Basins. However, the absence of extensive coral reefs, calcareous cyanobacteria and chemically precipitated ooids indicates that true tropical conditions were not attained.

In the Murray Basin, shallow platform limestones passed shorewards into terrigenous sediments deposited in restricted marine lagoons, intertributary bays and tidal flats. The coastal plain merged eastwards with alluvial plain, fluvial and lake environments. In contrast, carbonates of the Eucla and St Vincent Basins were rimmed by very narrow terrigenous clastic shorelines.

Relief amplitude of part of the ancestral highlands was generally similar to the present. However, in the southern Mount Lofty Ranges, the difference in height between the base of Oligocene-Early Miocene limestone in the Myponga and Hindmarsh Tiers Basins, and the top of the adjacent hills, implies local relief of

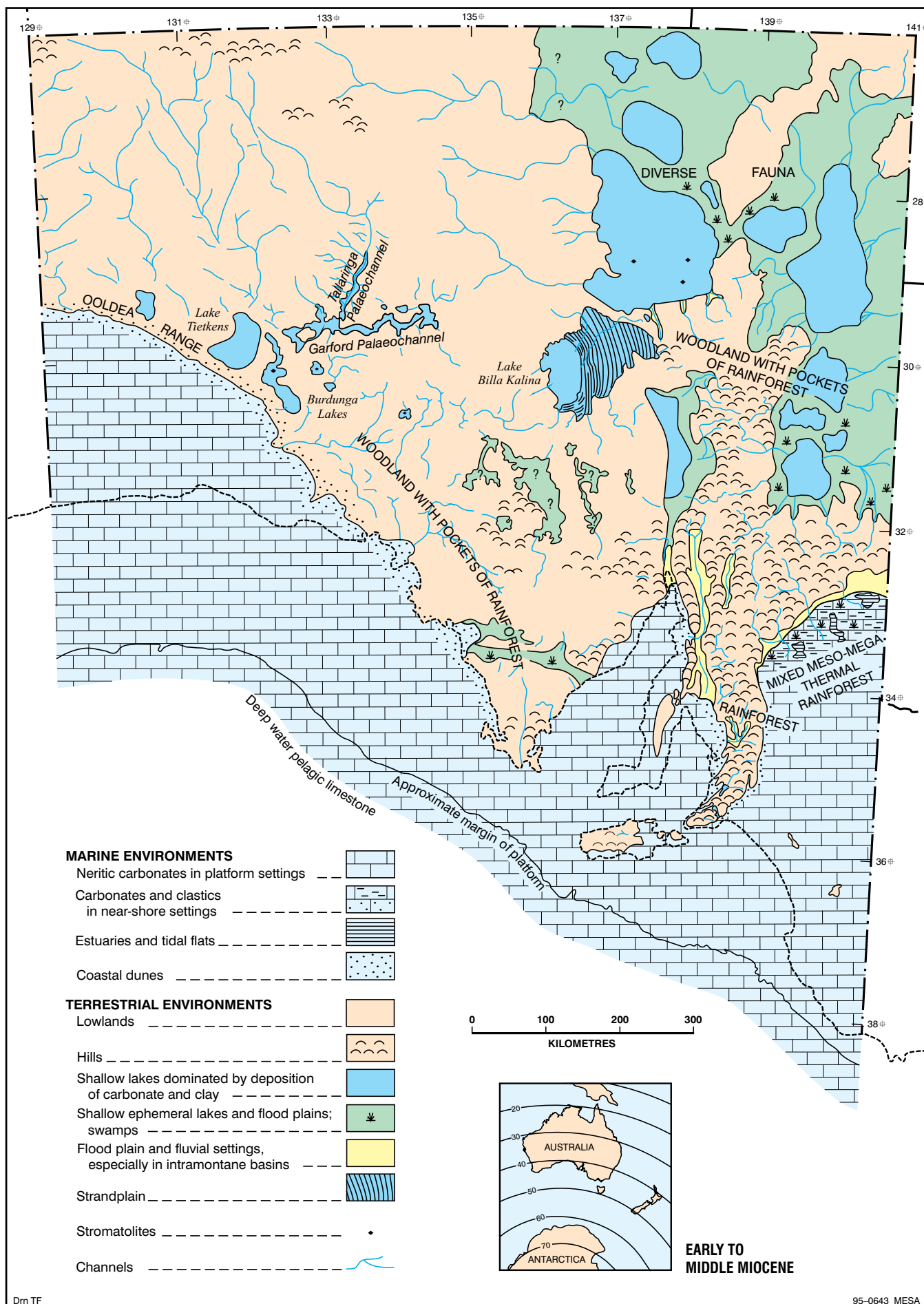


Fig. 10.44 Early to Middle Miocene palaeogeography of South Australia.

250-430 m, which is greater than at present. Similarly, in the ranges south of Leigh Creek, the difference in height between the base of the middle Tertiary Neuroodla Formation and the ridge tops indicates local relief amplitude of ~350-450 m, considerably more than the modern landscape. Piedmont fans along the eastern margin of the intramontane Barossa Basin provide evidence of uplift of this part of the Mount Lofty Ranges during the Oligocene-Middle Miocene.

The inland non-marine basins and palaeodrainage systems contained extensive, shallow, alkaline lakes in which argillaceous and calcareous and dolomitic mudstones were deposited (Krieg, Callen *et al.*, 1990; Benbow, in prep.). Centres of deposition were controlled by local epeirogenic movements in the Tirari Sub-basin (Wopfner, 1974) and by already established palaeochannels in the Callabonna Sub-basin and Eucla Basin. These inland sediments remain poorly dated, but maximum deposition may have occurred around the Early-Middle Miocene boundary, more or less contemporaneously with maximum deposition over the marine platforms.

Although the inland climate was significantly wetter during the Early-Middle Miocene than today, the presence of extensive dolomite indicates evaporative conditions, and the local presence of ooids and stromatolites implies warmth as well as shallow water (Krieg, Callen *et al.*, 1990; Benbow, in prep.). Decreasing carbonaceous sediments and increasing dolomite towards the northwest suggest increasing relative aridity in that direction.

Vertebrate fossils, particularly in the upper green clay and channel sands of the Etadunna Formation, have aided reconstruction of palaeoenvironments and understanding of the evolution of Australian faunas (see Callen, 1990, for a summary of the main groups). The faunas contain numerous birds, including many aquatic species, land-dwelling marsupials characterised by browsing types, fish, lungfish, dolphins, large crocodiles and turtles, indicating warm conditions and the existence of a substantial body of permanent water (Archer and Clayton, 1984). The presence of dolphins indicates an exorheic drainage system while koala-like animals, phalangers, possums and native cats suggest that trees grew in the area.

Although trees grew around lakes and rivers, the vegetation record is poor. The few sparse palynofloras from the Lake Eyre Basin contain relatively common *Podocarpus* and *Nothofagus* (Callen and Tedford, 1976) implying rainforest elements. A palynoflora from near the base of the Namba Formation in Wooltana 1 is dominated by Restionaceae, growing as swamp plants, with lesser conifers and *Casuarina*, minor *Nothofagus* (*Brassospora*) and very rare grass pollen (Martin, 1990). Thus, widespread interfluvial grasslands had not yet developed in central Australia, an interpretation supported by the preponderance of browsing marsupials.

Around the Murray Basin, mixed rainforest prevailed, dominated by myrtaceous genera (possibly *Syzygium*, *Acmena* and *Tristania*) and *Nothofagus* (*Brassospora*), although the latter are much lower in frequency than in the Gippsland area (Victoria; Truswell *et al.*, 1985). These rainforest trees are associated with significant *Gymnostoma* and lesser Cunoniaceae and gymnosperms, including consistently present *Araucaria*, *Dacrycarpus*, *Dacrydium* Section B and *Podocarpus*, and very minor *Lagarostrobos* and *Microcachrys*. There is uncertainty about the palaeoclimatic significance of this rainforest assemblage, but the preponderance of Myrtaceae may be in response to seasonally drier conditions (Truswell *et al.*, 1985).

Palynofloras from the intramontane Barossa Basin contain high frequencies of *Nothofagus* pollen, mainly *Brassospora* but with significant *Fuscaspora*, in association with common *Casuarina*, very variable Myrtaceae and podocarps. These spectra indicate rainforest similar to that of southeastern Australia, perhaps growing in locally suitable sites such as uplands surrounding the basin.

Palynofloras from the eastern Eucla Basin imply drier conditions on Eyre Peninsula. *Casuarina* dominated the vegetation with reduced *Nothofagus* (*Brassospora* the most common) and podocarps. The Proteaceae were very poorly represented and

Acacia was present. The lack of grass indicates that woodland and/or forest, rather than grassland, extended across the peninsula.

Palynofloral evidence suggests significant regional and perhaps even local variations in vegetation during the Oligocene to Early Miocene in response to a shift to drier conditions, with greater impact on inland areas.

MIDDLE TO LATE MIOCENE (14-6 MA)

The major increase in $\delta^{18}\text{O}$ values in the South Tasman Rise area during the Middle Miocene reflects an increase in the size of the Antarctic Ice Sheet and not so much a decrease in ocean temperature (Shackleton and Kennett, 1975; Mathews and Poore, 1980). This icesheet greatly influenced atmospheric and oceanic circulation patterns in the Australasian region and patterns were most likely similar to those of today (Kemp and Barrett, 1975; Kemp, 1978).

The regression that followed deposition of extensive marine limestones in the Middle Miocene was accentuated by uplift of the southern trailing plate margin. Despite falling sea level, deposition of neritic carbonates continued in the Gambier Basin and outer Eucla Basin platform well into the Late Miocene (Fig. 10.45; Abele *et al.*, 1976, 1988; Apthorpe, 1972).

Mild epeirogeny during the Late Miocene resulted in gentle warping, small-scale faulting, weathering, erosion and formation of low-angle unconformities in the St Vincent and Murray Basins. Uplift of the Eucla Basin platform and Gambier Basin resulted in cessation of deposition and a change to subaerial exposure and development of karstic features. Limestone of the Eucla Basin may have been reworked and thus been a major source of bioclastic sediment, perhaps being eroded from the newly formed Bunda Cliffs. Further doming of the Cordillo Surface with major subsidence in the Lake Eyre and Torrens Basins may have occurred at this time.

The change in atmospheric circulation brought about by the Antarctic Ice Sheet and the lack of widespread Late Miocene deposits inland suggest increasing aridity at this time (McGowran, 1987; Truswell and Harris, 1982). In the eastern Murray Basin, cessation of deposition of carbonaceous sediments and the decreasing proportion of *Nothofagus* pollen indicate diminishing precipitation by the end of the Middle Miocene (Martin, 1977). In the southeastern part of the continent wet conditions with rainforest persisted into the Late Miocene (Kemp, 1978). Similar conditions prevailed in the Barossa Basin area where *Nothofagus*-dominated rainforest (mainly *Brassospora*) varied in time and space with vegetation assemblages dominated by podocarps and occasionally Casuarinaceae.

The immaturity of karst development on the Nullarbor Plain and the intact preservation of palaeodrainage and coastal dunes of the Eucla Basin suggest aridity here for much of post-Miocene time (Jennings, 1967; Van de Graaff *et al.*, 1977; Benbow, 1990a).

PLIOCENE (6-1.6 MA)

The Pliocene was an interval of widespread deposition of largely terrigenous sediments in non-marine and paralic to shoreline settings. The record of marine transgressions is best known in the St Vincent and Murray Basins but dating is not precise enough to distinguish closely spaced global eustatic peaks. Little is known of the marine carbonate record of the continental shelf.

In the Murray Basin, deposition began with latest Miocene to Early Pliocene extensive, shallow marine platform marl in the western part; eastwards, these sediments passed into sand deposited in shoreface, deltaic and alluvial plain environments (Fig. 10.45). During subsequent progradation, an extensive strandplain with arcuate beach ridges formed under the influence of a westerly wind system which produced a southwesterly swell and longshore current regime (Belperio and Bluck, 1990). Large quantities of sand were supplied from the Eastern Highlands. In contrast, the Early Pliocene transgression(s) left a restricted clastic record in the

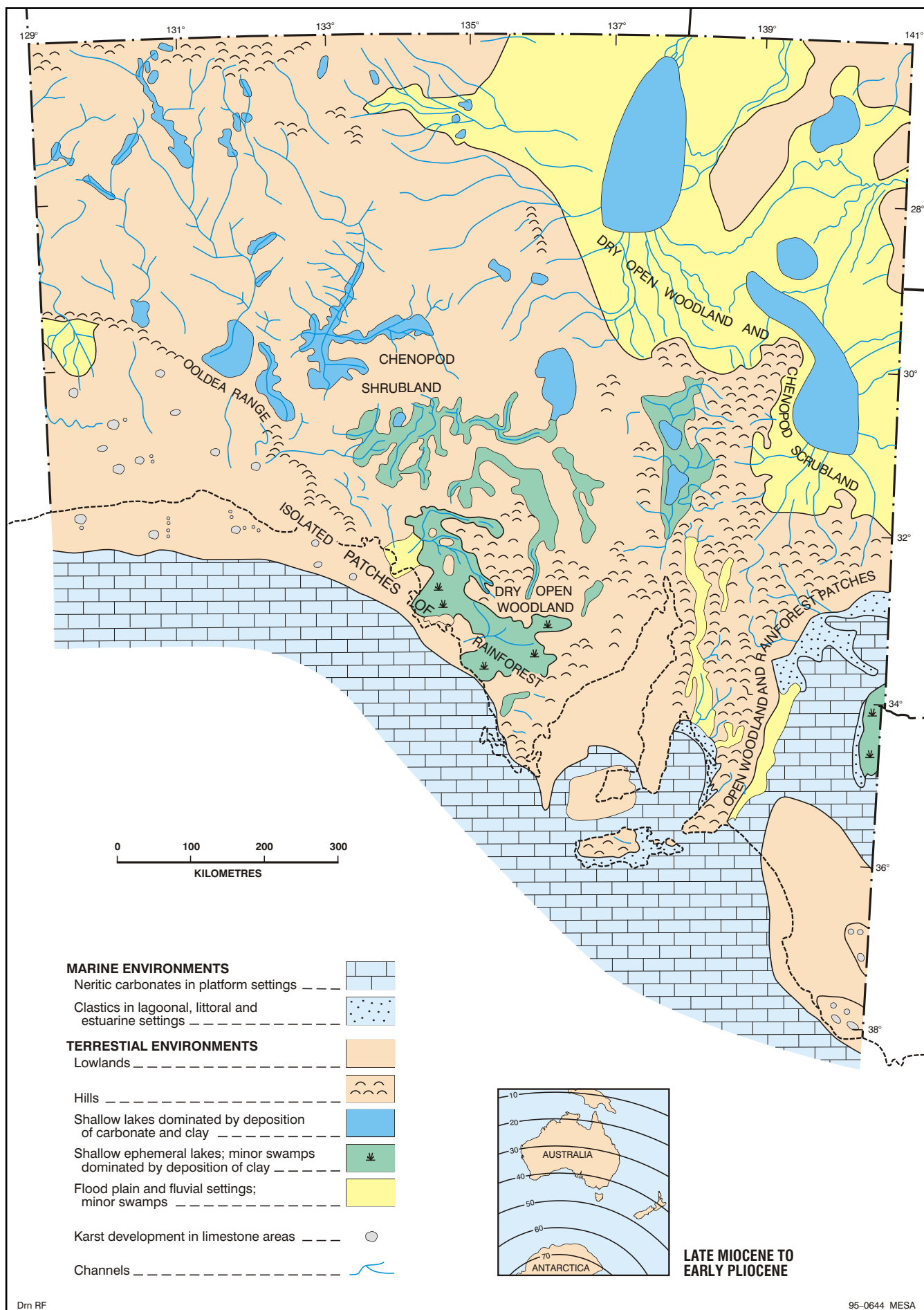


Fig. 10.45 Late Miocene to Early Pliocene palaeogeography of South Australia.

St Vincent Basin, probably as a result of limited sediment supply from adjacent uplands.

During the Early Pliocene (and/or possibly Late Miocene) there was very extensive fluvial and lacustrine sedimentation along palaeochannels around the Eucla Basin, in the Lake Eyre Basin and elsewhere. Thick and extensive argillaceous and minor calcareous Early Pliocene mudstone on northern Eyre Peninsula possibly correlates with the uppermost part of the vertebrate-bearing Etadunna and Namba Formations, which may extend into the Early Pliocene (Martin, 1990). The Serpentine Lakes Palaeochannel on the Bunda Plateau was reactivated, leading to deposition of a broad alluvial fan (M.C. Benbow, MESA, pers. comm., 1994). Clastic sediments in the Hamilton Basin were deposited in flood-plain and piedmont settings marginal to the Musgrave Ranges. The presence of crocodile remains suggests that conditions were warm and wet in this area.

The record of late Cainozoic development of aridity is unfortunately very meagre. Superimposed on this overall trend of increasing aridity were warm, wet episodes in the Pliocene. Open vegetation growing in a climate more arid than that of the Middle Miocene has been inferred from the presence of grass, composites and Chenopodiaceae pollen on northern Eyre Peninsula (Truswell and Harris, 1982). *Casuarina* is abundant and *Eucalyptus* is present in these assemblages. The presence of *Dacrydium*, *Podocarpus*, *Phyllocladus* and very rare *Nothofagus* (*Brassospora*) pollen indicate rainforest in edaphically suitable sites (Truswell and Harris, 1982; N.F. Alley, unpublished information). These assemblages are also present in the Early Pliocene of the Narlabay Palaeochannel, along with significant frequencies of Restionaceae (rather than grass) and *Micrantheum*, in association with *Acacia*, Cupressaceae (probably *Callitris*) and sedges, both the latter being locally common. The occurrence of open woodland containing pockets of rainforest accords with the interpretation of Late Miocene to Early Pliocene vegetation in the eastern Murray Basin by Martin (1973a, 1978).

Ferricrete, and subsequently extensive silcrete, formed across the inland during the Late Pliocene. Epeirogenic uplift of the Flinders and Mount Lofty Ranges to near their current altitude culminated at about this time, and flanking fans (Willawortina Formation and Hindmarsh Clay) were formed. These movements may have been responsible for the folding of silcrete marginal to the northeastern Flinders Ranges although no major displacement occurred across faults.

In the closing stages of the Tertiary, clay was deposited in extensive shallow lake systems. In the Murray Basin this resulted from tectonic damming and/or damming by coastal barriers of streams following regression of the sea. Lake Bungunnia occupied much of the western Murray Basin between 2.5 and 0.7 Ma, with its northern and eastern margins surrounded by wide alluvial plains (Bowler, 1980; Stephenson, 1986; Stephenson and Brown, 1989).

Widespread carbonate deposition took place in coastal and offshore environments during a period of higher sea level near the Pliocene-Pleistocene boundary. Sheltered estuaries and bays of Gulf St Vincent passed to more energetic sandy and rocky bays and headlands around Kangaroo Island (Milnes *et al.*, 1983; Ludbrook, 1983). Coastal dunes formed from shelf-derived calcarenites, which became a feature of southern Australia during the Quaternary, were established along part of the coastline in the Late Pliocene, as indicated by dated basalt interbedded with aeolianite near Warnambool in Victoria. Saline and carbonate facies in the Murray Basin, other inland basins and palaeodrainage of the Eucla Basin indicate the drying of lakes and general climatic warming and/or drying in the late Tertiary and Quaternary.

The processes of erosion and deposition that shaped the late Tertiary landscape continued during the Quaternary; these are described in the following chapter. The lack of later widespread tectonic uplift and absence of Quaternary glaciation, such as characterised Europe and North America, has resulted in the preservation of many Tertiary sediments, landforms and weathering horizons in the modern landscape.

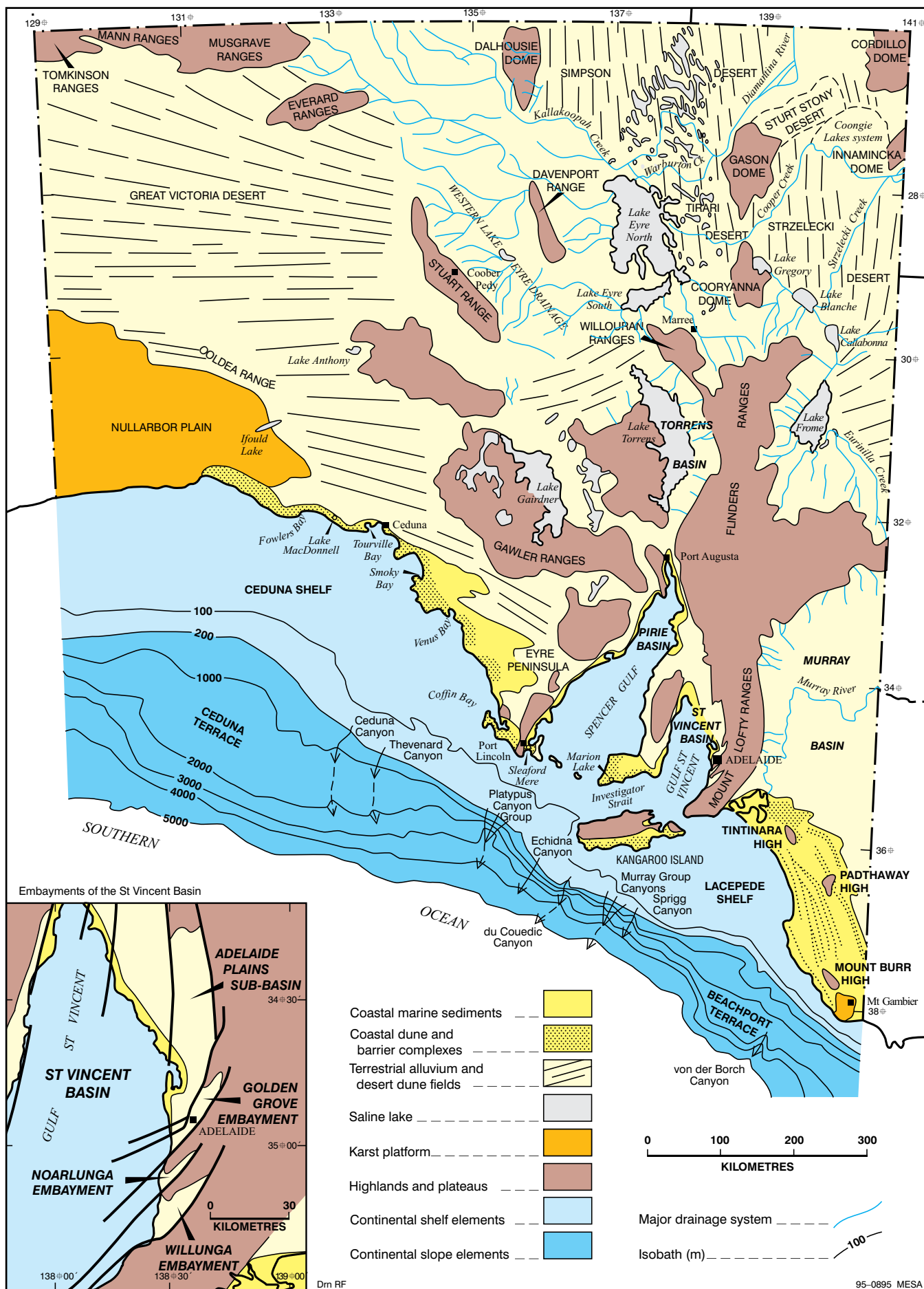


Fig. 11.1 Quaternary morphological elements of South Australia.

Chapter 11

QUATERNARY

Compiled by A.P. Belperio

with contributions from A.P. Belperio, M.C. Benbow, R.A. Callen, G.J. Drew,
D.N. Love, P.A. Rogers, W.V. Preiss, M.J. Sheard, P.C. Smith, M. Stone and S. Walker

INTRODUCTION

A.P. Belperio

The Quaternary Period covers the last 1.6 million years of Earth history and is characterised by marked global climate oscillations, the emergence of modern faunas including humans, and development of many of the present-day geomorphological features.

Global climate changes caused the build-up and decay of continental-scale icesheets, resulting in transgressions and regressions of the oceans and advances and retreats of deserts. Emiliani (1955) was able to outline the major features of the glacial-interglacial fluctuations by measuring oxygen isotope ($^{18}\text{O}/^{16}\text{O}$) ratios of planktonic foraminifera extracted from deep ocean sediments. Glacial episodes coincide with core intervals of ^{18}O enrichment in foraminifera as a result of changes in oceanic isotope compositions coincident with the growth of continental icesheets. Interglacial episodes coincide with depletion of ^{18}O in microfossils as a result of higher ocean temperatures and addition of meltwater to the oceans. This work established a quasi-periodic cycling of climate that had recurred many more times than the four glaciations recognised from previous continental European stratigraphic studies. In South Australia, these global climatic changes were manifested as repeated transgressions and regressions of the continental shelf, leaving an intermittent sedimentary record along the coast and in the gulfs (Fig. 11.1). Inland, repetitive climatic changes resulted in alternating periods of lake expansion and desert expansion.

A refined and detailed chronology of palaeoclimatic fluctuations is now established from several continuous deep-sea cores (Broecker and van Donk, 1970; Shackleton and Opdyke, 1973, 1976; Shackleton, 1986b; Williams *et al.*, 1988). Oxygen isotope stage boundaries correspond to times of rapid isotope change in the oceans, equated with global continental glaciation and deglaciation events (Fig. 11.2). Spectral analyses of the oxygen isotope records indicate a close (but not necessarily linear) relationship with periodicities of Milankovitch orbital parameters of Earth precession (19 000 and 23 000 years), obliquity (41 000 years) and eccentricity (100 000 years; Hays *et al.*, 1976; Imbrie *et al.*, 1984). Age calibration of oxygen isotope stages is based on a combination of palaeomagnetic reversal stratigraphy and radiometric dating (Berggren *et al.*, 1980, 1985a,b). The pattern of geomagnetic reversals for the Cainozoic is well established, although some revision of the chronology will be necessary as the radiometric data on which it depends are updated (Fig. 11.2).

Twenty-seven glacial-interglacial cycles since the Olduvai Subchron can be recognised in the deep ocean isotope record (Fig. 11.2), with nine of these occurring during the Brunhes Chron. The Brunhes-Matuyama palaeomagnetic reversal occurs at stage 19/20 and, with an estimated age of $780\,000 \pm 20\,000$ years (Spell and McDougall, 1992), is used as a major control point for dating oxygen isotope stage boundaries.

The distinction between Tertiary and Quaternary Periods is rather artificial, and the age of this boundary has varied in the past (Berggren and van Couvering, 1974). Because onset of continental glaciation is now believed to extend back to the Miocene, climatic

deterioration can no longer be considered a valid or useful datum for defining the beginning of the Quaternary.

Despite advances in the fields of magneto- and oxygen isotope stratigraphy, the base of the Pleistocene Series was formally defined in a stratotype section at Vrica, southern Italy (Aguirre and Pasini, 1985), following a recommendation that the boundary be placed at the physical horizon below the first appearance of the 'cold water ostracode' *Cytheropteron testudo*. This has resulted in the base of the Pleistocene being located a few metres above (i.e. slightly younger than) the top of the Olduvai Subchron (1.66 Ma).

The Quaternary Period is subdivided into Pleistocene and Holocene Epochs, with a somewhat artificial boundary fixed at 10 000 years BP¹. Informal subdivisions within the Pleistocene correspond to the Brunhes-Matuyama reversal (Early-Middle Pleistocene) and the beginning of the last interglaciation (Middle-Late Pleistocene). These correspond to the base of oxygen isotope stages 19 and 5, respectively. Following Berggren *et al.* (1985b) and Spell and McDougall (1992), boundary age estimates of these informal divisions of the Quaternary are:

Early Pleistocene	1.6-0.78 Ma
Middle Pleistocene	780 000-128 000 years BP
Late Pleistocene	128 000-10 000 years BP
Holocene	10 000 years BP-present.

The ocean basins provide an excellent, largely complete record of Quaternary sedimentation and climatic change. In regions proximal to continental margins, the cyclic climatic changes are reflected in pelagic-hemipelagic or carbonate-terrigenous alternations that are in phase with the oxygen-isotope variations (e.g. Nelson *et al.*, 1985). However, continental shelf, coastal and terrestrial deposits present a much more irregular record, with more complex response and time lag to climate change.

Radiometric data and the magnetic polarity time scale provide a means of correlating coastal and terrestrial deposits with the deep-sea record. Oxygen isotope stratigraphy, however, is not as useful in the punctuated shallow marine and terrestrial sequences. The stratigraphic record contained in inland Australia and in the shallow gulfs and coastal zone reflects, to varying degrees, the changing 'global' climates and fluctuating sea levels. This record, however, is only a fragmentary portion of the total Quaternary record that has accumulated in deeper ocean basins of the globe and to which it must necessarily be referred. Not surprisingly, the history of classical land-based investigations of this fragmentary Quaternary record in Australia is plagued with changing age interpretations and correlations, and this will undoubtedly continue as geochronometric techniques applicable to the Quaternary improve. In this Chapter, the Quaternary environments and sediments of South Australia are described in terms of two major provinces — the marine-dominated shelf and coastal zone, and the continental interior. The economic significance of these morphostratigraphic units is summarised, and a final section discusses Quaternary geological processes as they relate to our natural and modified environment.

¹ Holocene dates are given in radiocarbon years 'Before Present' which, by convention, is defined as 1950 AD.

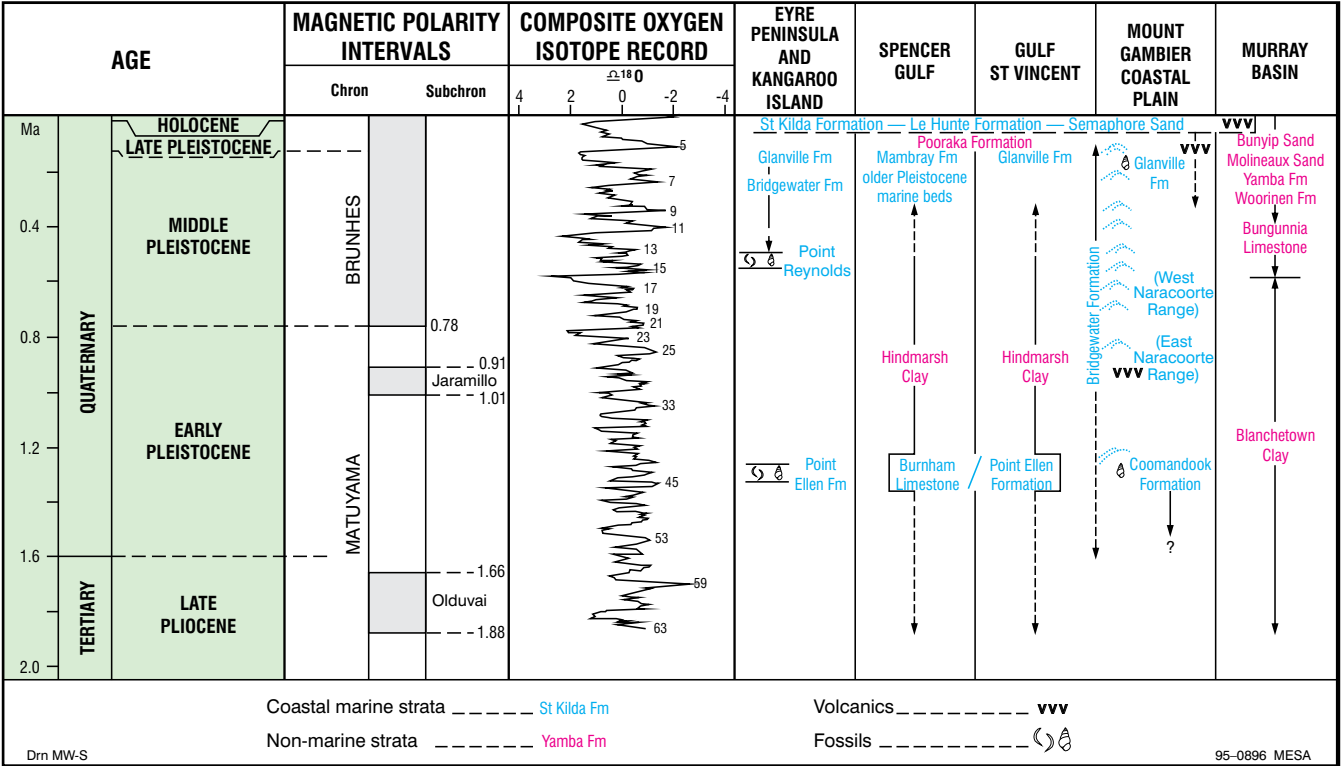


Fig. 11.2 Palaeomagnetic and oxygen isotope stratigraphy of the Quaternary (after Berggren *et al.*, 1985a,b; Haq and van Eysinga, 1987; Williams *et al.*, 1988; Spell and McDougall, 1992). Black is normal polarity. Also shown are stratigraphic correlations of major coastal marine and associated non-marine strata.

COASTAL AND MARINE SEQUENCES

A.P. Belperio

The global deterioration of climate that commenced in the Pliocene accelerated in the Pleistocene, but with increasingly marked climatic fluctuations. In the Early Pleistocene, the Earth's climate was dominated by the 42 000 year obliquity cycle as seen in the oxygen isotope signal (Fig. 11.2). In the Middle and Late Pleistocene, the climatic fluctuations became increasingly regular about the 100 000 year cycle with greater and more dramatic changes in temperature and sea level. Consequently, Early Pleistocene marine sequences are poorly represented on land compared with later transgressions. In addition to these global sea-level cycles, the South Australian Quaternary coastal and marine record also reflects local climatic, oceanographic and tectonic factors.

COASTAL PALAEOGEOGRAPHY

The continental shelf, the exposed coastline of southern and western South Australia, and southern and western Kangaroo Island, have consistently been dominated by high-energy swell from the Southern Ocean, with its prevailing onshore southwesterly winds. Pleistocene and Holocene coastal sediments from these regions consist of comminuted bioclastic beach, barrier and transgressive dune complexes (Fig. 11.1). Repeated marine transgressions have resulted in construction, destruction and reworking of a number of coastal barrier complexes. Along western and southern Eyre Peninsula, many of the earlier coastal barriers were destroyed as the coast regressed inland with each subsequent highstand. In the South-East, similar coastal facies have been physically separated as a result of uplift of the Mount Gambier coastal plain. Within the protected, tidally influenced central gulfs, Quaternary sediments are characterised by shallow marine muddy limestones that pass laterally into marginal peritidal and intertidal shelly facies and interdigitating terrigenous clastics.

In the Murray Basin, a marine coastal plain merges with an inland basin that was dominated by fluvio-lacustrine and aeolian sedimentation. Epeirogenic basin subsidence was complicated by

geographically variable warping and tilting associated with Pliocene and Quaternary volcanism in western Victoria and Mount Gambier as well as uplift of the Pinnaroo Block. The resulting tilting of the Mount Gambier coastal plain, in addition to separating successive coastal barriers, also assisted in the formation and maintenance of Lake Bungunna, and in the constriction of the palaeo-Murray River against the eastern Mount Lofty Ranges. During the Pleistocene, the Australian continent became progressively more arid with a gradual change from summer to winter-dominated rainfall (Bowler, 1982). Increasing aridity is reflected by a change from fluvial to aeolian and saline gypseous sediment at ~500 000 years BP in the Murray Basin. Increasing Quaternary aridity is also reflected in the composition of preserved coastal and shallow marine sediments, with a change from mixed quartz-carbonate sediments in the Early Pleistocene to increasingly carbonate-dominated sediments with cool-temperate affinities.

EARLY PLEISTOCENE DEPOSITS

Fossiliferous, shallow marine limestone of inferred Early Pleistocene age underlies much of the Mount Gambier coastal plain (Coomandook Formation; Firman, 1967a, 1973; Rochow, 1971; Ludbrook, 1978; Rogers, 1980). Equivalent deposits occur in the St Vincent Basin (Burnham Limestone), on southern Fleurieu Peninsula and on Kangaroo Island (Point Ellen Formation), and in the Eucla Basin (Roe Calcarenite).

Coomandook Formation

Sediments of the Coomandook Formation reach a maximum thickness of 75 m on the Mount Gambier coastal plain. The marine transgression extended to the Kanawinka Fault at Naracoorte and transgressed the Padthaway Ridge north of Coonalpyn (Fig. 11.3). Reworking of underlying Pliocene sands and Miocene limestones was extensive, and the main lithologies are sandy limestone, calcareous sandstone and shelly sandstone and clay. Fossils of the pelagic marine gastropod *Hartungia dennanti chavani* allow correlation with the earliest quartzose calcarenite dunes of the Bridgewater Formation east of Naracoorte and Keith (Ludbrook, 1978).

During the transgression, granite hills of the Padthaway Ridge formed a series of islands and shoals that influenced coastal sedimentation and coastal palaeogeography (Fig. 11.3). Major coastal barriers developed (Coonalpyn Range, East Naracoorte Range), the first of a series of coastal barriers collectively referred to as the Bridgewater Formation, that were constructed intermittently throughout the Pleistocene (see below). The Coomandook Formation thins to the south so that, by Robe, the younger Bridgewater Formation coastal barriers rest directly on Miocene limestone. Calcareous sand sheets which may be related are developed south of Penola.

Palaeomagnetic studies by Idnurm and Cook (1980) showed that the coastal barriers east of Naracoorte are of reversed polarity, indicating a minimum age of 780 000 years BP. Based upon elevation and uplift rate of the Mount Gambier coastal plain (see below), the East Naracoorte Range may be correlated with the oxygen isotope stage 25 sea-level highstand with an inferred age of 860 000 years BP. Alternatively, it may be an older structure associated with an earlier highstand such as stage 59 in the Late Pliocene. The Coonalpyn Range is a similar, more complex, Early Pleistocene coastal facies, probably equivalent in age to the Coomandook Formation.

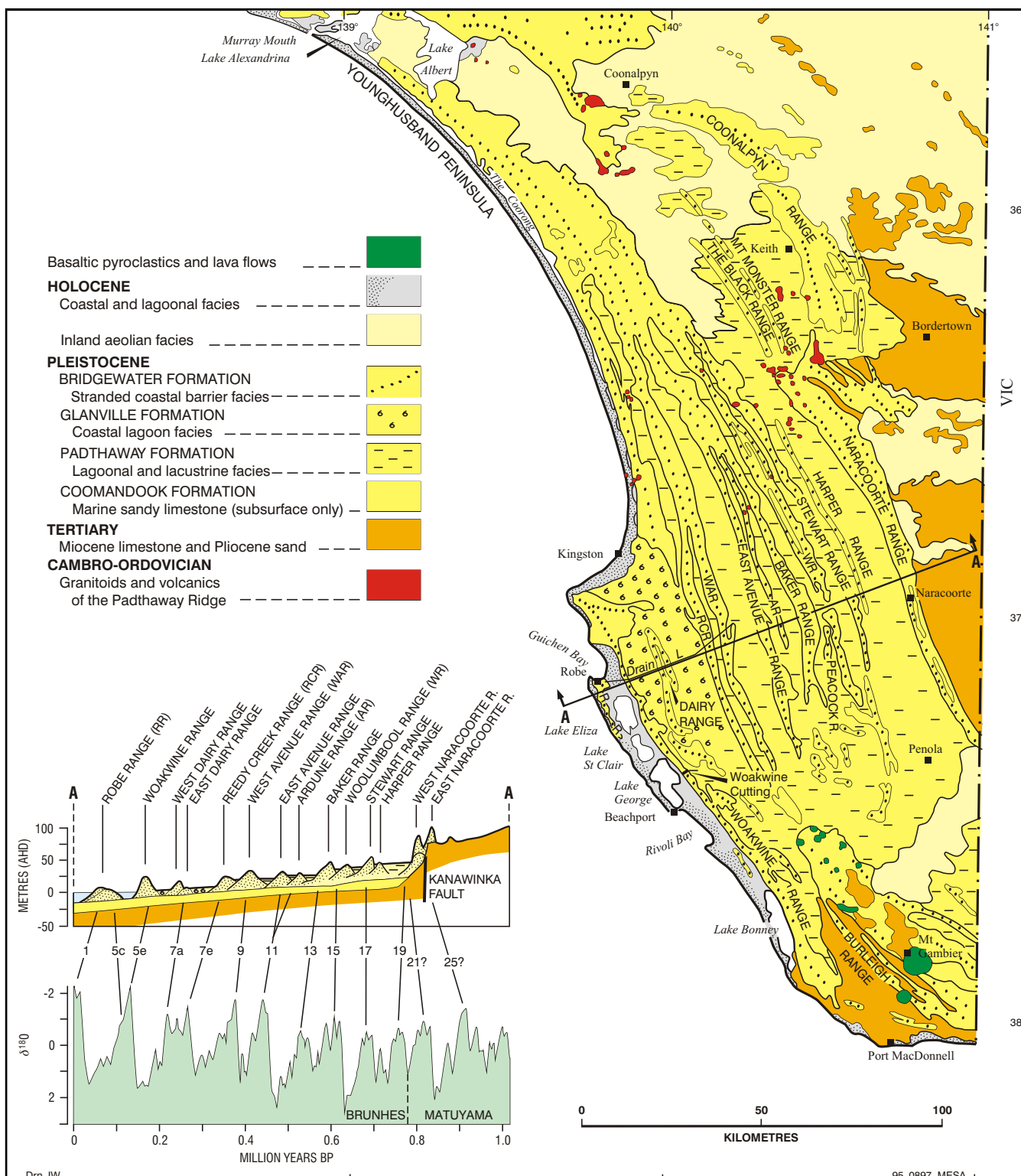


Fig. 11.3 Geology of the South-East and correlation of shoreline sequences with oxygen isotope stages.

Coeval deposits that support an Early Pleistocene age for the Coomandook Formation include the Nelson Bay Formation and Werrikoo Limestone of western Victoria (Boutakoff, 1963; Lawrence *et al.*, 1976; Singleton *et al.*, 1976; MacFadden *et al.*, 1987). At Portland, the shallow marine Nelson Bay Formation and overlying Bridgewater Formation aeolianite are of reversed magnetic polarity. The Nelson Bay Formation is important for its wealth of vertebrate and invertebrate fauna and its flora. It also contains the planktonic foraminifera *Globorotalia truncatulinoides* which first appeared in the Southern Hemisphere ~1.9 million years ago. The Portland sequence is therefore constrained within the reversed magnetic chron of 1.66-0.78 Ma. A similar framework has been established for the sequence in the Glenelg River valley, where subaerial basalts yielding K-Ar ages of 2.5 to 2.2 Ma are overlain by marine sand and littoral calcarenite of the Werrikoo Limestone. *G. truncatulinoides* appears some distance above the base, indicating that this unit spans the Pliocene-Pleistocene boundary.

Burnham Limestone

In the St Vincent Basin, the Early Pleistocene Burnham Limestone crops out as eroded remnants on uplifted fault blocks between Kingston Park and Sellicks Beach south of Adelaide (Firman, 1976; May and Bourman, 1984). Elevation differences between basin and upfaulted remnants imply ~80-90 m of relative subsidence of the St Vincent Basin during the Quaternary. Burnham Limestone is a pale grey, micritic limestone with poorly defined bedding and a fauna indicative of an estuarine or lagoonal environment. Clasts of underlying Hallett Cove Sandstone are common along a basal unconformity with karst joints and solution holes. Correlatives within the St Vincent Basin are the fossiliferous sand at Lockleys (Ludbrook, 1963b) and thin marly limestone at the base of the Hindmarsh Clay from Port Gawler southwards (Lindsay, 1969).

Point Ellen Formation

Further south, at Cape Jervis and on Kangaroo Island, equivalent Early Pleistocene fossiliferous limestones are termed Point Ellen Formation (Milnes *et al.*, 1983; Ludbrook, 1983). At these more exposed localities, the fossil molluscan fauna is a mixed assemblage of rock-dwelling and shallow sand-dwelling species. Bryozoa, echinoids and foraminifera are also abundant. Shells of the pelagic gastropod *Hartungia dennanti chavani* permit correlation with the Roe Calcarene of the Eucla Basin, and Coomandook and Bridgewater Formations of the Naracoorte Range. The foraminiferal assemblage, although not diagnostic, includes *Marginopora vertebralis*, *Ammonia beccarii*, *Elphidium rotatum* and *Cribrobulimina mixta*. At Cape Willoughby, beach and aeolianite facies are unconformable on a weathered granite surface ~10 m above present sea level. The aeolianite facies marks the first appearance of a repetitive coastal transgressive dune facies (Bridgewater Formation) that is widely distributed along the coastal margin. At Pennington Bay on southern Kangaroo Island, fossiliferous shell beds of the Point Ellen Formation, 3 m above sea level and dated at 1.2 Ma by the Sr isotope method, are overlain by a 50+ m sequence of Bridgewater Formation calcarenites that contain numerous carbonate palaeosols.



TIME-TRANSGRESSIVE COASTAL AEOLIANITES

Bridgewater Formation

The Bridgewater Formation was formally defined by Boutakoff (1963) at Bridgewater Bay, Victoria, on the eastern extremity of the Mount Gambier coastal plain. The name applies to bioclastic barrier shoreline deposits preserved as conspicuous subparallel topographic ridges across this coastal plain and on the continental shelf (Fig. 11.3; Sprigg, 1952, 1959a, 1979; Sprigg and Boutakoff, 1953). The surface on which the ridges rest rises to 60 m above sea level at the Kanawinka escarpment and around the Mount Gambier-Mount Burr volcanic centre. The stranded coastal ridges on the coastal plain average 15 m in height, and their altitude and degree of separation decrease to the northwest (Fig. 11.3). Thirteen ridges related to specific Pleistocene sea-level highstands are preserved between Robe and Naracoorte, but these coalesce into a single composite aeolianite complex towards Lake Alexandrina (Belperio and Cann, 1990). Southwards, the older ridges lose their expression and only those seaward of the Avenue Range continue clearly into western Victoria. Older, inland coastal complexes are more silica-rich than later deposits and contain trace quantities of heavy minerals reworked from underlying Pliocene sands (Colwell, 1979; Belperio and Bluck, 1990).

Five sediment facies associated with shoreline deposits of the Bridgewater Formation are exposed in various drains and cuttings through the ranges (Fig. 11.4). These comprise skeletal calcarenite and marl deposited in shallow subtidal, littoral, dune, lagoon and



Top: Stranded coastal barrier (Burleigh Range) near Mount Gambier. Photo 042521.

Bottom: Woakwine Cutting, bulldozed through a stranded coastal barrier near Beachport. Photo 042522.

Left: Marine gastropod *Hartungia dennanti chavani* from the East Naracoorte Range. Photo 042520.

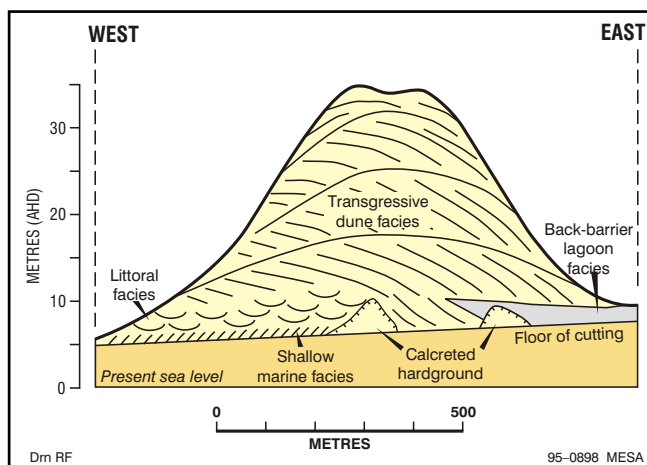


Fig. 11.4 Cross-section exposed by the Woakwine Cutting through the Woakwine Range.

ephemeral lacustrine environments associated with the former shorelines. The spatial association of facies resembles that of the contemporary Coorong barrier-lagoon system. The first three facies make up the Bridgewater Formation *per se*, whilst the last has been defined separately as the Padthaway Formation. The lagoon facies has no formal nomenclature apart from the Glanville Formation behind the Woakwine Range.

Shallow subtidal facies of Bridgewater Formation comprise seaward-dipping, coarse carbonate sands with abundant broken shells of marine gastropods and bivalves together with foraminifera and coralline algae. They occur as a basal transgressive facies and, in the South-East, have often been incorrectly referred to the more time-specific Coomandook Formation. Sprigg (1952) recognised a 'reef' or rocky shore fauna of *Haliotis*, *Turbo* and *Brachiodontes* adjacent to calcreted promontories within the Woakwine Range.

Sediments of the littoral zone are well-sorted, medium to coarse-grained, quartzose and skeletal carbonate sands deposited as shoestring sand bodies subparallel to the present coast. The beach deposits, up to 3 m thick, extend into the core of the ranges as distinctive bidirectional trough cross-bedded strata.

Transgressive dune or aeolianite facies form the bulk of the visible ridges. They comprise weakly cemented, fine to medium-grained skeletal sand composed of well-sorted and rounded molluscan, foraminiferal and algal grains. Reactivation surfaces, marked by weakly developed palaeosols and rhizolith concentrations, are common. These separate the cosets of very large-scale cross-beds that dip consistently landward at 3 to 30°. The dune facies toe-out in their lee where they interfinger with, or overlie, sediments of the lagoon facies.

Poorly sorted sediments of back-barrier lagoon facies are preserved in interdune corridors between the stranded ranges. These sediments are typically very shelly and muddy calcarenite with numerous gastropods and large articulated intertidal to subtidal bivalves including *Anadara*, *Katylsia* and *Ostrea*. The intertidal fauna provide a useful datum for measuring shoreline elevation and reconstructing former sea levels (Table 11.1).

Lacustrine sediments of the Padthaway Formation were deposited in interdune corridors, but are not necessarily coeval with formation of the adjacent Bridgewater Formation barriers. They consist of dense, white calcitic and dolomitic mudstone typically 1 to 2 m thick. Interbedded greenish clay and clayey quartz sand is also common. The calcilutite is amorphous to indistinctly bedded, often pelletal, and contains remains of a variety of fresh to brackish water molluscs, foraminifera, ostracodes and charophytes. Holocene dolomite has formed in similar settings, in contemporary continental groundwater-dominated ephemeral lakes behind the modern barrier, as well as in inland ephemeral lakes (Alderman, 1965; von der Borch, 1981). The ephemeral lake deposits

commonly overlie lagoonal facies of the Bridgewater Formation and together comprise upward-fining regressive sequences. Where thickly developed, they have been mapped separately and referred to as Padthaway Formation.

Chronological control is scattered for the aeolianite and calcarenite deposits of the Bridgewater Formation on the Mount Gambier coastal plain. At Warrnambool in western Victoria, aeolian phases of the Bridgewater occur stratigraphically above and below basalts dated at 1.8 Ma (Orth, 1988). At Portland, the preserved Bridgewater sequence is entirely of Early Pleistocene age (reversed magnetic polarity) as are the aeolianite ranges east of Naracoorte. West of Naracoorte, the barrier sediments are of normal magnetic polarity (Cook *et al.*, 1977) and therefore largely of Middle and Late Pleistocene age ($\leq 780\,000$ years BP). Quartz grains from the barriers show steadily increasing thermoluminescence (TL) with distance inland and increasing ages in that direction (Huntley *et al.*, 1985, 1993); ages of 120–800 000 years have been derived. Uranium series data for the Woakwine Range (Schwebel, 1983) and aminostratigraphic correlations (Murray-Wallace and Belperio, 1991) indicate a Late Pleistocene (125 000 years BP) age for this feature. Clearly, the Bridgewater Formation of the Mount Gambier coastal plain is a punctuated, time-transgressive sequence associated with successive sea-level highstands that span the Early, Middle and Late Pleistocene.

Since the earliest studies of Sprigg (1952), it has been recognised that there is good agreement between the uplifted coastal plain sequence and *de facto* global climatic and Milankovitch insolation signatures (Cook *et al.*, 1977; Sprigg, 1979; Idnurm and Cook, 1980; Schwebel, 1983). In the absence of more complete chronological data, reasonable age estimates can be made by matching individual barriers to the oxygen isotope record and utilising the indicated palaeomagnetic and geochronological constraints (Fig. 11.3). A relatively complete sequence of stranded high sea-level deposits from stage 1 to stage 19 is suggested, indicating coastal sedimentation in phase with global climatic events. Former sea-level highstands may be estimated by correcting shoreline elevations for tectonic uplift. A rate of uplift of 0.07 mm/year for the past 800 000 years is indicated by the highly coherent plot of shoreline elevation versus inferred age for a Robe–Naracoorte line of section (Fig. 11.5). The last interglacial shoreline, with an elevation of +8 to +10 m near Robe (Murray-Wallace and Belperio, 1991), corroborates this uplift rate. Uplift-corrected palaeoshoreline data indicate a repetition of sea-level highstands close to or just below contemporary sea level (Table 11.1). Hence uplift has been a fundamental factor in the preservation of these strandlines above present sea level.

Calcarenites similarly referred to the Bridgewater Formation occur extensively along the coast of Kangaroo Island, Yorke Peninsula and western Eyre Peninsula. These have a similar coastal origin (barrier shoreline complexes), but also include transgressive dune sands and aeolian sand sheets that form extensive veneers

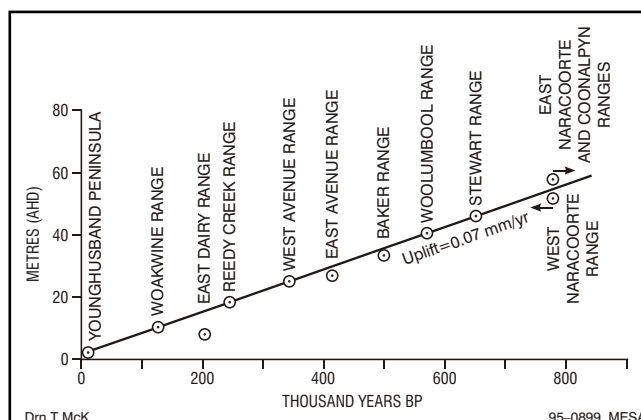


Fig. 11.5 Plot of shoreline elevation versus inferred age (from Table 11.1) for coastal barriers of the Mount Gambier coastal plain.

Table 11.1 Estimates of tectonically corrected sea levels and ages for individual barrier shorelines of the Mount Gambier coastal plain.

Range	Isotope stage ¹	Estimated age ¹ (years BP)	TL age ²	Shoreline altitude ³ (m)	Uplift corrected sea level ⁴ (m)
East Naracoorte-Coonalpyn	25-?	>780 000	720 000	+58	<+3
West Naracoorte-Black	19	≤ 780 000		+55	0
Harper-Stewart	17	680 000	585 000	+45	-3
Peacock-Woolumbool	15	570 000		+40	0
Baker	13	500 000	456 000	+34	-1
Ardune-East Avenue	11	420 000	414 000	+26	-3
West Avenue	9	350 000	342 000	+24	-1
Reedy Creek	7e	250 000	258 000	+18	0
Dairy	7a	200 000		+8	-6
Woakwine	5e	120 000	132 000	+10	+2
Robe III	5c	100 000		-2	-9
Robe II	5a	80 000		-12	-17
Younghusband Peninsula	1	7 000		+0.5	0

1. From Figure 11.2.

2. TL ages after Huntley *et al.* (1993).

3. For a Robe-Naracoorte section from the elevation of lagoon or littoral facies and including modified data from Sprigg (1952b), Schwebel (1983) and Keeling *et al.* (1985).

4. Assuming a +2 m local reference level for the last interglacial and consistent uplift of 0.07 mm/year.



Bridgewater Formation showing stacked calcarenite sequences separated by palaeosols, Cape Spencer, southwestern Yorke Peninsula. Photo 042523.

extending well inland. Without the consistent uplift experienced by the Mount Gambier coastal plain, and because of extensive aeolian reworking, these coastal complexes lack the same degree of physical separation between (and therefore resolution of) successive sea-level highstands. On southern Kangaroo Island and at Cape Spencer on southern Yorke Peninsula, up to 12 successive stacked aeolianite sequences, separated by calcretes and carbonate palaeosols, are present in the coastal cliffs. The aeolianites are typically cross-bedded on a large scale, composed of up to 90% fine to medium-grained skeletal carbonate, and are weakly cemented internally by sparry calcite or dolomite. With stratigraphic drilling and careful mapping, successive coastal sequences separated by pedogenic horizons can be distinguished, such as the four shoreline complexes at Fowlers Bay (Fig. 11.6).

Although masked by extensive surficial calcretes, littoral, lagoonal and aeolian facies were identified within these barrier sequences (Belperio, 1988a). Bioclasts were derived largely from the fragmentation of molluscs, coralline algae, bryozoa and echinoids that thrived on the continental shelf, and the assemblage also contains abundant foraminifera. Finer bioclasts and carbonate silt were blown inland by onshore winds to form extensive loess deposits. Remains of the land snails *Bothriembryon barretti*, *B. mastersi* and *?Exilibadistes sutilosa*, and fossilised pupal



Land snail *Bothriembryon barretti* within calcarenite of the Bridgewater Formation, west of Point Sinclair. Photo t024552.

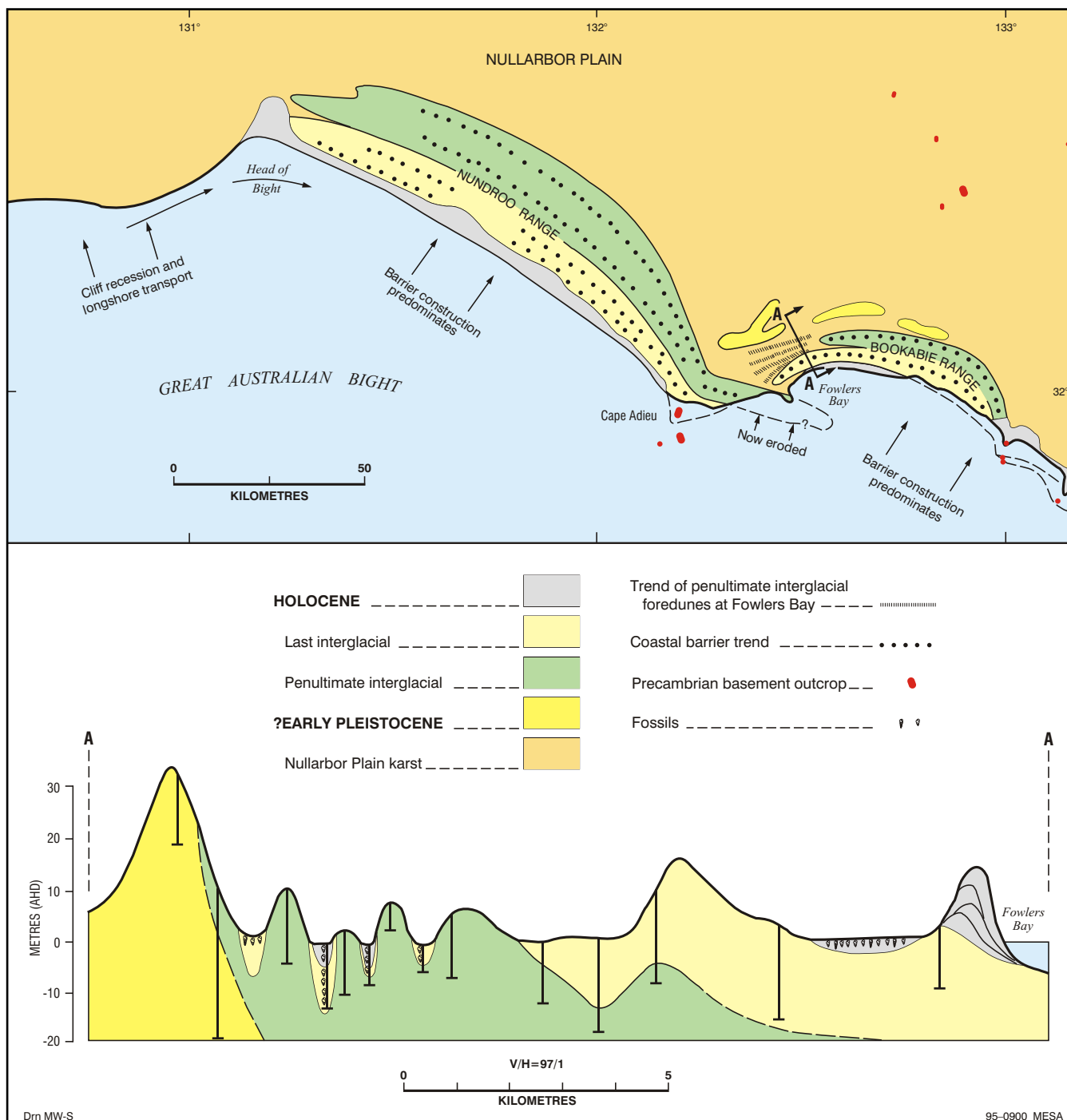


Fig. 11.6 Palaeogeographic reconstruction of Quaternary coastal barriers between Head of Bight and Fowlers Bay. Cross-section A-A through Fowlers Bay shows the development of four successive coastal sand barriers. (After Belperio, 1988a).

cases of the weevil *Leptopius duponti*, are common within these inland deposits (Ludbrook, 1973b, 1984; Flint, 1987). A small percentage of the microfossil assemblage is of Miocene origin (Milnes and Ludbrook, 1986) reflecting multiple cycling of these skeletal components. Carbonate palaeosols and calcretes formed on and within the Bridgewater Formation vary greatly in nature, thickness and induration. They include chalky, peloidal, nodular and laminar forms that reflect variable degrees of soil development and pedogenic alteration (Warren, 1983a; Bateman, 1987; Farrand and Belperio, 1987; Belperio, 1988a). Also common are reworked lithoclast horizons and truncated profiles including residual hardpans. In general, grain alteration, micritisation, cementation and matrix occlusion decrease over a few metres down a calcrete profile to reveal original textures and bedforms associated with littoral or aeolian sedimentation. Rhizoconcretions are common to much greater depths.

The stacked coastal calcarenite sequences of central and western South Australia present particular problems for chronological discrimination and intra-unit correlation. At Fowlers Bay, Belperio (1988a) was able to discriminate the youngest phase (Late Pleistocene, 125 000 years BP) of the Bridgewater Formation and its associated back-barrier lagoon facies (Glanville Formation). Earlier phases of coastal barrier construction were equated with Early and Middle Pleistocene sea-level highstands (Fig. 11.6). Flint (1987) also demonstrated time equivalence of the Glanville and this 'upper member' of the Bridgewater Formation from field relationships at Lake MacDonnell. Magnetostratigraphic measurements undertaken by Bateman (1987) on southern Yorke Peninsula indicate a Middle to Late Pleistocene age for the entire Bridgewater Formation complex preserved in the coastal cliffs. Wilson (1991) obtained a range of TL age estimates of 630 000 to 180 000 years BP for the stacked aeolianite sequences on southern and western Eyre Peninsula. At Point Reynolds on southern



Glanville Formation overlain by calcarenite of the Bridgewater Formation, Lake MacDonnell. Photo 024549.

Kangaroo Island, shells within the basal part of a 50 m thick aeolianite sequence have been dated at 520 000 years BP by the Sr isotope method. Clearly, as for the South-East, the Bridgewater Formation at these localities is also a punctuated time-transgressive unit that spans much of the Pleistocene.

THE PENULTIMATE INTERGLACIAL

Shallow marine and coastal sediments ascribed to a penultimate interglacial sea-level highstand (isotope stage 7) are known from sporadic localities around the coastline, but no formal nomenclature has been applied. When corrected for tectonic uplift, they indicate a stage 7e highstand very close to present sea level (Table 11.1).

On the Mount Gambier coastal plain, the Dairy, Burleigh and Reedy Creek Ranges (Fig. 11.3) are partly buried coastal correlatives of the stage 7 multiple highstand, with estimated ages of between 250 000 and 200 000 years BP (Table 11.1). The East Dairy Range records a sea level between 6 and 10 m above present (uncorrected for uplift). A higher elevation (+18 m) is recorded by back-barrier facies of the Reedy Creek Range but the fossil molluscs are too degraded to date positively by amino acid racemisation. Huntley *et al.* (1993) obtained an age of ~258 000 years BP from TL measurements on quartz grains from this coastal structure. Quartz-carbonate littoral sand facies at this same level on the seaward side of the barrier are 2-3 m thick and overlain by up to 20 m of large-scale trough cross-bedded dune sand. Further south near the volcanic edifice of Mount Schank, the Burleigh Range palaeoshoreline is elevated to +34 m, reflecting a greater degree of uplift.

Marine sediments of similar age encountered around the margins and in the subsurface of upper Spencer Gulf were termed 'Older Pleistocene marine beds' by Hails *et al.* (1984a,b). Although the strata do not crop out, they were intersected by several coastal and submarine cores beneath younger marine strata, from which they were clearly separated by a well-developed palaeosol. Pedogenic

alteration of the older shelly marine sediment has, in places, largely obliterated original textures and framework grains, and carbonate dissolution and illuviation of clay are characteristic. The lateral extent of this unit in the subsurface is poorly known. It occurs to levels of up to 4 m above low water datum beneath the modern supratidal zone at Redcliff where the presence of large *Anadara* sp. indicates a protected, shallow intertidal environment. Murray-Wallace *et al.* (1988b) used amino acid racemisation of *Anadara trapezia* to confirm a penultimate interglacial age (oxygen isotope stage 7) with a mean of 200 000±50 000 years BP at this site.

On western Eyre Peninsula, coastal aeolianites and associated littoral sediment facies of this age are also present. These are generally included within the time-transgressive Bridgewater Formation, and their delineation from older and younger coastal strata of similar origin is difficult. Wilson (1991) obtained TL age estimates of 180 000±40 and 190 000±40 years BP for an upper aeolian phase at Whalers Way (near Port Lincoln) and at Streaky Bay, respectively. Stratigraphic drilling at Fowlers Bay has revealed several shoreline structures equated with a penultimate interglacial sea-level highstand (Fig. 11.6). This shoreline complex also includes six distinct foredunes inland of Fowlers Bay, and associated major coastal aeolian barriers that make up the Nundroo Range and Bookabie Range. Fore-dune sediments are moderately sorted, fine-grained biocalcarene dominated by foraminifera and molluscs. They are moderately cemented and capped by multiple-generation, massive, laminar to nodular calcrete. Shelly shoreface sediments indicate a sea level 5-10 m below present. The Nundroo and Bookabie Ranges are bedrock-attached, coastal aeolian structures that define two southwest-facing, swell-dominated arcuate shorelines. The preserved fore-dune complex within Fowlers Bay was deposited in a semi-protected, southeast-facing, crenulate embayment between the coastal barriers.

THE LAST INTERGLACIAL

Shallow marine and coastal sediments deposited during the last interglacial sea-level highstand (isotope substage 5e) are common around much of the world's coastline. Globally, the best studied sites are from settings affected by high rates of tectonic uplift, such as the Huon Peninsula of Papua New Guinea (Chappell, 1974; Bloom *et al.*, 1974), and Timor and the adjacent Atauro Island (Chappell and Veeh, 1978). At these localities, uranium series dating of the last interglacial terraces has provided ages ranging from 160 000 to 111 000 years BP. On the basis of internal relationships and dates within the uplifted terraces, these authors interpreted a compound interglacial highstand with a double sea-level peak. More recently, Aharon and Chappell (1986) have refined estimates of this double sea-level peak to 138 000±5 000 and 118 000±2 000 years BP.

In Australia, uranium series age estimates of 143 000 to 111 000 years BP have been reported by Reeckmann and Gill (1981) from western Victoria, by Szabo (1979) from Rottnest Island, by Veeh *et al.* (1979) from Lake MacLeod, and by Marshall and Thom (1976) from northern New South Wales. In southern Australia, the lack of corals suitable for uranium series dating has resulted in greater reliance on amino acid racemisation dating and aminostratigraphic correlation of this interval and, more recently, TL dating of quartz grains. Although well exposed around coastal southern Australia, chronologic and stratigraphic studies have been unable to delineate multiple from single transgressive events or to further reduce the inherent errors in age estimates (Murray-Wallace and Belperio, 1991).

In South Australia, the widespread peritidal and shallow marine sediments are referred to as 'Glanville Formation' (Firman, 1966b; Ludbrook, 1976, 1984; Cann, 1978), although similar deposits in upper Spencer Gulf were termed 'Mambray Formation' by Hails *et al.* (1984a,b). Equivalent sandy coastal aeolian and coastal barrier facies are referred to the 'upper member of the Bridgewater Formation'. Initial radiocarbon dating of molluscs from the Glanville Formation provided only minimum ages of 45 000 to 35 000 years BP. Schwebel (1983) arrived at uranium series ages

of $125\,000 \pm 20\,000$ and $100\,000 \pm 30\,000$ years BP for molluscs and aragonite mud from lagoonal facies associated with the Woakwine Range near Robe. Belperio, Smith *et al.* (1984) calculated an age of $110\,000^{+19\,000}_{-17\,000}$ years BP from amino acid racemisation of *Anadara* in the Mambray Formation of upper Spencer Gulf. Additional TL age estimates of $96\,000 \pm 24\,000$ and $90\,000 \pm 15\,000$ years BP for coarse and fine quartz grains from the same strata confirmed the last interglacial age, allowing correlation with global substage 5e marine deposits. Subsequently, more refined aminostratigraphic correlation between different coastal deposits within and outside South Australia were undertaken by Belperio and Murray-Wallace (1984), Murray-Wallace and Kimber (1987, 1990), Murray-Wallace *et al.* (1988a, 1991) and Murray-Wallace and Belperio (1991). Huntley *et al.* (1993) obtained a model age of $132\,000 \pm 6\,000$ years BP from TL data for the Woakwine Range.

The Glanville Formation is characterised by some elements of a warmer water fauna that no longer live in South Australian waters. These include the Sydney blood cockle *Anadara trapezia*, Shark Bay pearl oyster *Pinctada carcharias*, conical-fusiform gastropod *Euplia bidentata*, and the large foraminifer *Marginopora vertebralis* (Howchin, 1888, 1909, 1924, 1935; Ludbrook, 1976). However, the bulk of the foram-mollusc detritus is similar to that found in modern peritidal environments. Typically, the sediments range from mottled shelly clay to moderately cemented skeletal calcarenite and coquina with a microfauna dominated by *Elphidium*, *Peneroplis* and *Marginopora* (Lindsay and Harris, 1975). In former protected lagoonal and intertidal environments, the fossil fauna include abundant cockles of the genus *Katelysia* (*K. rhytiphora*, *K. peronii* and *K. scalarina*), and numerous gastropods such as *Batillaria* (*Zeacumantus*) *diemenensis*, *Diala lauta*, *Niotha pyrrhus*, *Turritella* sp. and *Bulla botanica*. Other common large bivalves include *Anadara trapezia*, *Ostrea angasi*, *Chlamys* (*Equichlamys*) *bifrons*, *Fulvia tenuicostata*, *Brachiodontes* (*Austromytilus*) *rostratus* and *B. erosus*.

Equivalent calcarenites and aeolianites (upper member of the Bridgewater Formation) of more exposed coastlines reflect a mixed and comminuted shallow continental shelf source. The skeletal constituents are similar to those that comprise the contemporary oceanic beaches and barriers. Typically they include abraded and well-sorted tests of foraminifera, calcareous red algae, and comminuted molluscan, bryozoan and echinoid debris. Reworked skeletal aggregates, microids and pisoids are also common and quartz is the dominant silicate in a relatively small non-carbonate fraction. A fine micritic carbonate matrix is common, usually a result of illuviation of carbonate within calcrete profiles. Grain micritisation and surficial carbonate hardpan development are ubiquitous. Below the carbonate hardpan, the sediment is typically grain-supported and very weakly cemented by a thin isopachous or meniscus calcite cement or by pervasive dolomitisation. Timing and extent of pedogenic alteration, and the range of vadose and phreatic diagenetic processes, are geographically highly variable (Hutton and Dixon, 1981; Billing, 1984; Farrand and Belperio, 1987; Belperio, 1988a).

The last interglacial shoreline on the Mount Gambier coastal plain is represented by the Woakwine Range complex and associated back-barrier facies. The coastal barriers consist predominantly of sand-sized biogenic carbonate with occasional thin conglomerate and shell beds. They interfinger on their lee-side with cockle, scallop and oyster-rich lagoonal to intertidal shell beds (Glanville Formation) and dolomitic lacustrine facies that are extensively developed within interdune corridors south of Kingston, as far inland as the base of the Reedy Creek Range. The Woakwine Range complex is largely a single linear form about 30 m high, rising to ~60 m near Cape Jaffa where it bifurcates to form an easterly trending bar or 'hook'. Internal complexities of the Woakwine Range are revealed by good exposures in the walls of deep drains (Fig. 11.4). In the Woakwine Cutting near Beachport, littoral and back-barrier facies indicate a single major depositional event with

a sea level that regressed from 10 to 8 m (relative to present sea level, and uncorrected for uplift). In Drain L, east of Robe, three successive depositional events are recognisable at between +5 and +10 m above present sea level, and a rocky 'reef' fauna associated with former promontories is present on the seaward side of the range (Sprigg, 1952).

Within and beneath Spencer Gulf and Gulf St Vincent, marine sediments of the last interglacial occur extensively as a blanket deposit in the subsurface and subcrop at low elevations around the gulf margins. A variety of lithofacies are preserved including poorly sorted subtidal, intertidal and supratidal facies, and low beach ridge and shoreface deposits. All contain a rich molluscan assemblage. Accurate mapping of the transgressive feather-edge of the Glanville Formation is difficult where it is masked by younger alluvial and marine sediments. In northern Gulf St Vincent, the calcreted and karstified surface of the Glanville Formation crops out discontinuously within the contemporary supratidal zone. The Glanville lithofacies here are intertidal, and the present supratidal zone is largely inherited from the earlier morphostratigraphy. The Glanville shoreline is represented by gravelly beach accretion ridges beneath and to the southeast of Port Wakefield. These record a sea level 3 m above present. In the Port Gawler region, Belperio and Bateman (1986) mapped the subsurface thickness, extent and landward limit of the Glanville Formation from borehole and vibrocore information. They delineated a coastal wedge up to 8 m thick beneath the modern intertidal zone.

Apart from limited exposure in drains and excavations, the Glanville Formation does not crop out in the Adelaide region, though it is extensively developed in the subsurface of low-lying northwestern and northern suburbs (Ludbrook, 1976; Johnson, 1977; Cann, 1978; Belperio, 1985a,b,c; Belperio *et al.*, 1986; Belperio and Rice, 1989). Belperio (1985c) ascribed the decrease in elevation of the intertidal facies of the Glanville Formation from Port Wakefield to Port Adelaide to differential subsidence between these sites.

On Kangaroo Island, a seaway existed for at least part of the last interglacial through the lowland area now occupied by Pelican Lagoon, separating Dudley Peninsula from the remainder of the island. This seaway was progressively choked by northward migrating transgressive dunes. Richly fossiliferous shelly limestones (Glanville Formation) deposited in intertidal flat and sandy, littoral environments crop out extensively around the margins of Pelican Lagoon and New Lake, throughout the low-lying former marine seaway, and around the margins of the Nepean Bay and D'Estrees Bay lowlands. Major transgressive dune complexes associated with the last interglacial highstand are developed along the south coast of Dudley Peninsula, west from Cape Gantheaume to Vivonne Bay, and between Cape du Couedic and Rocky River. Smaller sand bodies were deposited around the northeast coast. The large-scale cross-bedded, weakly cemented calcarenites cannot be easily distinguished from earlier multiple episodes of



Glanville Formation exposed near Robe. Photo 042524.

Pleistocene dune transgression, but a coeval relationship with the littoral Glanville Formation is evident at several locations.

Along western Eyre Peninsula, Late Pleistocene aeolian and littoral calcarenites are similarly well developed, but only locally has chronostratigraphic separation of earlier aeolianite complexes been attempted (Fig. 11.6). These high wave energy coastal complexes are veneered by thin, partly developed calcretes and, internally, are very weakly cemented and contain abundant rhizomorphs. Also present in this upper member of the Bridgewater Formation are abundant and unusual elliptical calcareous cases or cocoons (fossil pupal cases of the weevil *Leptopius duponti*) and fossil bee cells (Flint, 1987). In addition to the dominant aeolianites, littoral and foredune facies have also been identified, and low-energy lagoonal to protected intertidal facies (Glanville Formation) are common. The latter are correlated with the type Glanville Formation at Port Adelaide by aminostratigraphy (Murray-Wallace *et al.*, 1988a).

Drillholes at Fowlers Bay clearly demonstrate lateral and vertical changes from coastal aeolian to back-barrier lagoon and high wave energy littoral facies (Fig. 11.6). The former extensive lagoons and sheltered embayment systems at Fowlers Bay, Lake MacDonnell, Tourville Bay, Ceduna and Smoky Bay hosted a rich cockle, oyster and scallop assemblage, and large *Anadara* specimens are common. Macropodid footprints are preserved in beach facies sediments at Clare Bay. At its landward limit, the Glanville Formation hosts a typical intertidal sand-flat facies dominated by *Katylsia* sp. Along the west coast of Eyre Peninsula, the calcreted surface of this relatively flat intertidal facies consistently indicates a last interglacial sea level ~2 m higher than present. This level is an important datum for coastal deformation studies in southern Australia (Murray-Wallace and Belperio, 1991).

INTERSTADIAL MARINE SEQUENCES

Several marine transgressions inundated the South Australian gulfs after withdrawal of the Glanville sea, but did not reach the height of the present transgression (Fig. 11.7). The preserved deposits of three of these interstadial marine highstands (oxygen isotope substages 5c and 5a, and stage 3) have been intersected in the gulfs, but at levels well below present sea level. Because tectonic corrections are small, the onlapping strata pinch out at water depths that relate closely to the former sea levels (Fig. 11.8). They are separated by variably developed palaeosols.

The False Bay Formation (Hails *et al.*, 1984b) occurs in the subsurface of Spencer Gulf in water depths >8 m. Preserved lithofacies include poorly consolidated quartzose beach sand, calcareous estuarine to lagoonal sandy clay, and finer grained sabkha-like carbonate mud. Foraminifera, juvenile bivalves, bryozoa and sponge spicules are common skeletal components. Palaeosols in the formation developed carbonate-enriched zones with minimal clay (Billing, 1984) and were subject to aeolian reworking. False Bay Formation strata up to 10 m thick were deposited in a gulf of more restricted extent than present-day Spencer Gulf (Fig. 11.9a). Based on stratigraphic superposition, Hails *et al.* (1984b) concluded that the False Bay Formation was deposited during the substage 5c highstand (~105 000 years BP) that reached a maximum height of 8 m below present sea level. A similar uplift-corrected level is recorded for the Robe III barrier on the Mount Gambier coastal plain (Table 11.1).

The Lowly Point Formation (Hails *et al.*, 1984b) consists of laminated, calcareous clay of lagoonal to sabkha origin, with thin intercalated quartz sand layers of aeolian origin. Dolomite, high-magnesian calcite and gypsum are all common, often coexisting with aragonite. Bivalve fragments dominate the skeletal fraction, in addition to foraminifera, turreted gastropods, sponge spicules and rare corals. Pedogenic alteration is limited, with weak calcareous horizon formation, slight clay enrichment and minimal weathering of shell fragments. The strata are restricted to the deeper channels of Spencer Gulf in water depths >14 m, and thickness rarely exceeds 4 m (Fig. 11.9b). The presence of abundant dolomite and gypsum in association with aragonitic

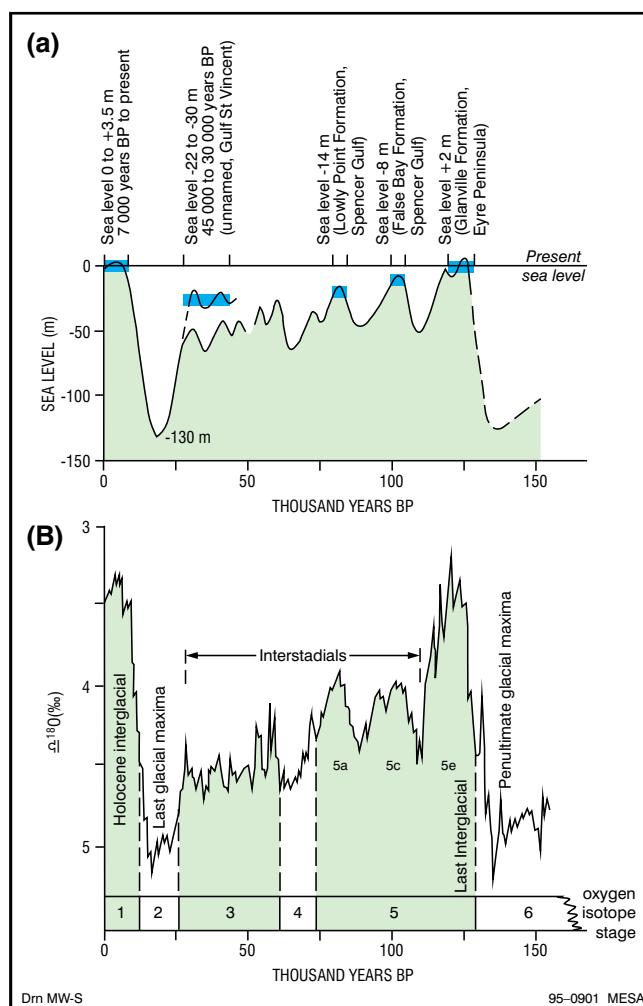


Fig. 11.7 Sedimentary record of Late Pleistocene high sea-level events compared to the orbitally tuned sea-level record for Australia of Chappell and Shackleton (1986). Higher sea levels are recorded in Gulf St Vincent for the period 45 000–35 000 years BP (a). For comparison, the high-resolution benthic oxygen isotope record (b) is from equatorial Pacific core V19-30 (Shackleton, 1986b).

lagoonal mud, and the restricted extent of the strata, indicates deposition in a narrow, protected, ephemeral lagoon system. Hails *et al.* (1984b) concluded that the Lowly Point Formation was deposited during the substage 5a highstand (~80 000 years BP) that reached a maximum height of -14 m relative to present sea level.

Gulf St Vincent has a maximum water depth in excess of 40 m, considerably deeper than the shallow (20–25 m) waters that characterise northern Spencer Gulf. As a consequence, even younger (stage 3) interstadial sea-level highstands that did not enter Spencer Gulf inundated the deeper central parts of Gulf St Vincent. These sediments were deposited over the time interval 45 000 to 30 000 years BP, during which sea level exceeded -30 m and reached as high as -22 m (Figs 11.10, 11.11). Amino acid racemisation dating of fossil bivalves has confirmed the radiocarbon age ranges. Sediments comprise un lithified, fossiliferous, calcitic mud with variable proportions of shallow marine bivalves, gastropods and foraminifera. Well-preserved aragonitic fossils of *Katylsia*, *Chlamys* and *Batillaria* indicate that no significant pedogenic alteration has occurred. Cann *et al.* (1988, 1993) used the down-core variation in foraminiferal species abundance to calculate palaeo-water depths of 10–18 m in central Gulf St Vincent at that time. The stage 3 marine sediments are overlain by Holocene marine sediments and, in places, by saline lacustrine sediments associated with the Holocene transgression.

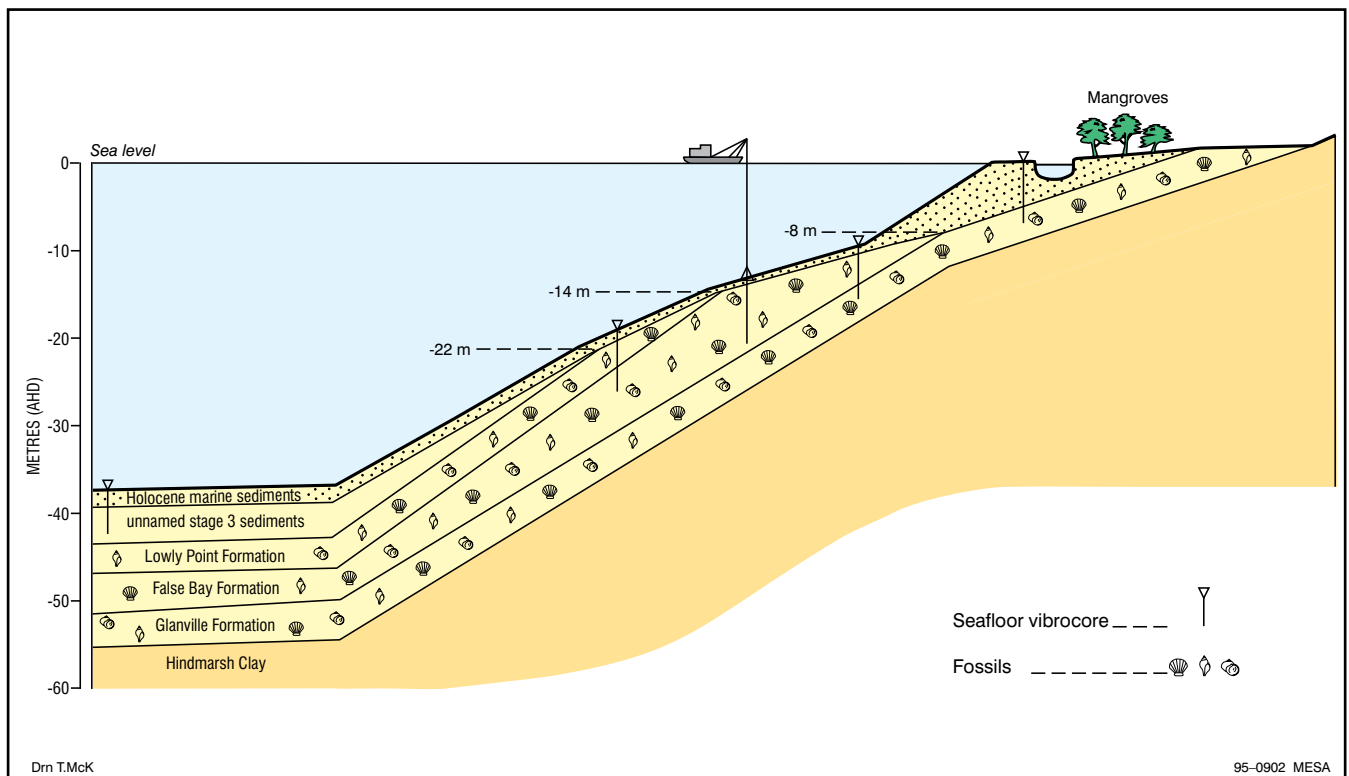


Fig. 11.8 Schematic representation of Late Pleistocene stratigraphic units beneath Spencer Gulf and Gulf St Vincent showing critical pinch-out depths and superposition of marine strata in seafloor cores. Each marine unit is separated by a palaeosol.

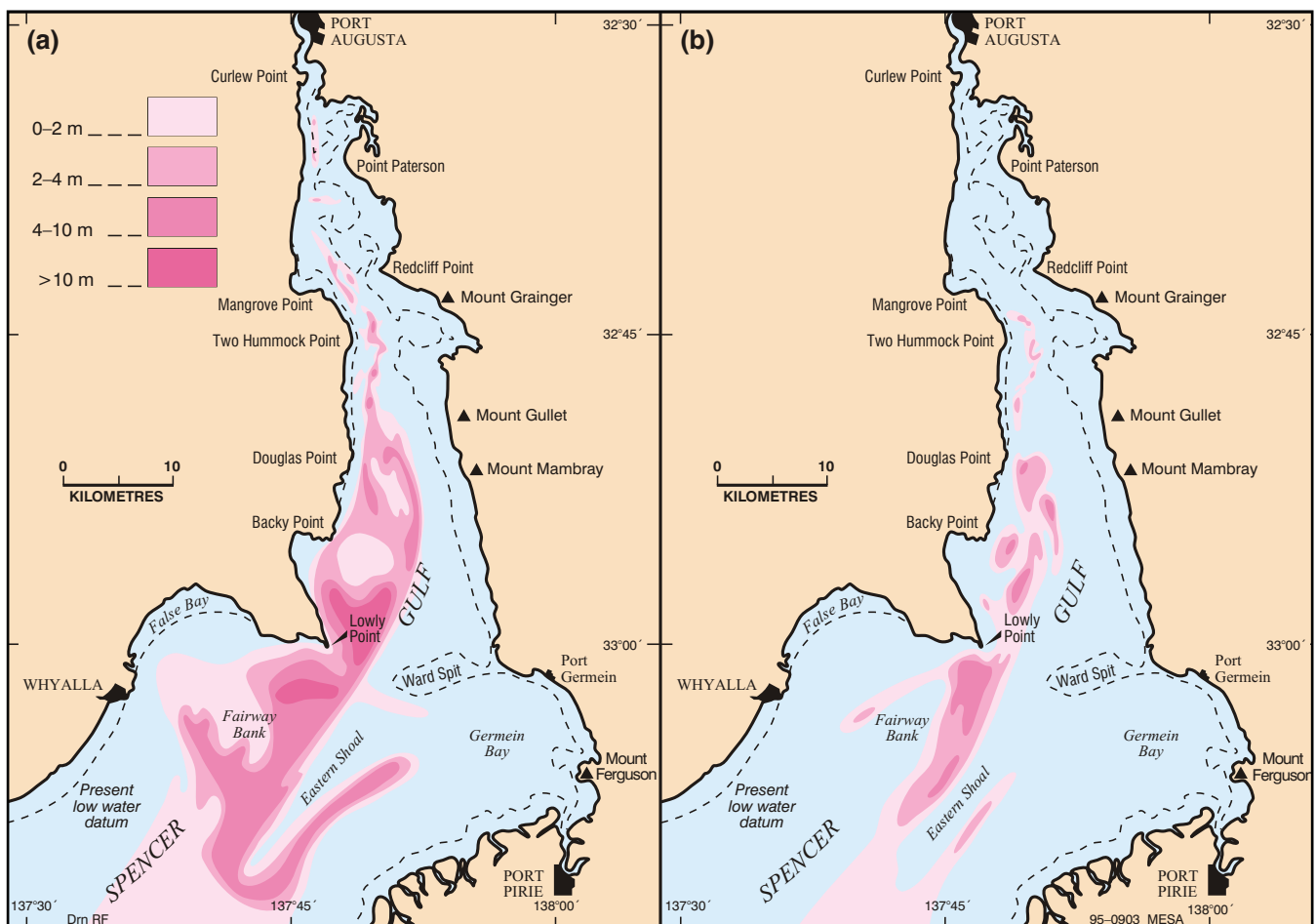


Fig. 11.9 Isopachs (m) of (a) False Bay Formation, and (b) Lowly Point Formation in northern Spencer Gulf (after Hails *et al.*, 1984a,b).

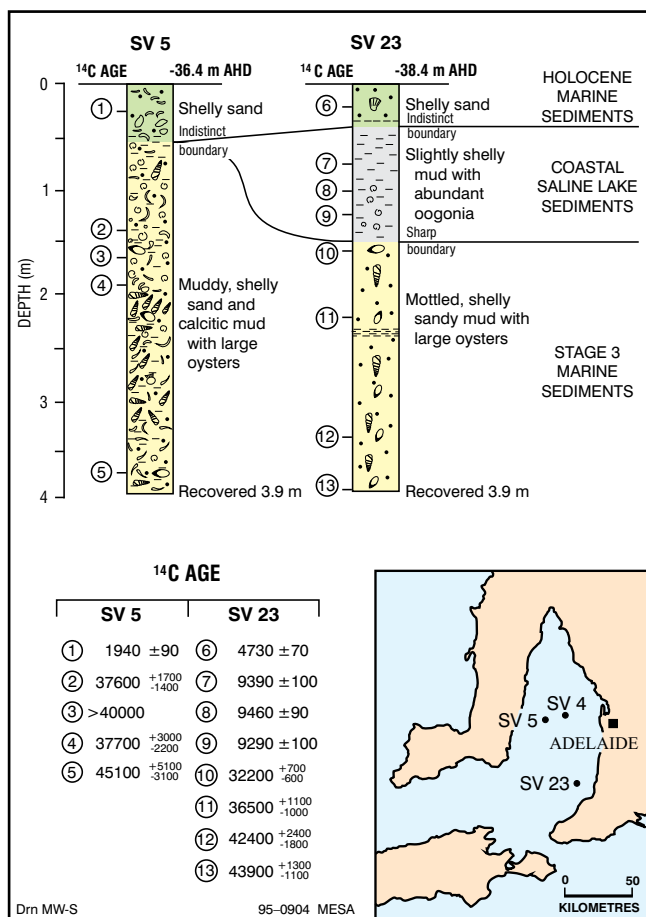


Fig. 11.10 Two vibrocores from Gulf St Vincent with Holocene (stage 1) and Late Pleistocene (stage 3) marine sediments. Radiocarbon dates from molluscs are indicated.

ASSOCIATED NON-MARINE SEDIMENTS

In meridional marine basins adjacent to the Mount Lofty and southern Flinders Ranges (St Vincent and Pirie Basins, respectively), alluvial and associated continental sedimentation and aeolian redistribution occurred between, and in part contemporaneously with, the Quaternary sea-level highstands (Fig. 11.12). Within the Adelaide Plains Sub-basin, these siliciclastic deposits include Carisbrooke Sand (Lindsay, 1969), Hindmarsh Clay (Firman, 1966b; Sheard and Bowman, 1987b), Keswick Clay (Sheard and Bowman, 1987a), Pooraka Formation (Firman, 1969; Daily *et al.*, 1976), Fulham Sand (Bowman and Sheard, 1988a) and Kilburn Sand (Bowman and Sheard, 1988b).

Extensive inland dunefields of aeolian quartz sand (erroneously equated by Firman (1966b) with coastal-derived Fulham Sand) and various palaeosols and pedogenic mantles developed sporadically throughout the Quaternary, although their deposits are often difficult to put in regional stratigraphic context. Fulham Sand, derived from reworking of Late Pleistocene coastal sand dunes and other inland dunefields, was formed or remobilised during a period of aeolian landscape instability in the last glacial maximum ~18 000 years BP.

Marine seismic and coring investigations have confirmed that the Hindmarsh Clay, Pooraka Formation and aeolian sands are widespread beneath Gulf St Vincent and Spencer Gulf (Hails *et al.*, 1984b; Belperio *et al.*, 1990). The intercalated marine sediments provide useful chronostratigraphic control on some of the continental stratigraphy. Thus, the Pooraka Formation is marker-defined in coastal regions where it is separated from Hindmarsh Clay by the marine Glanville Formation.

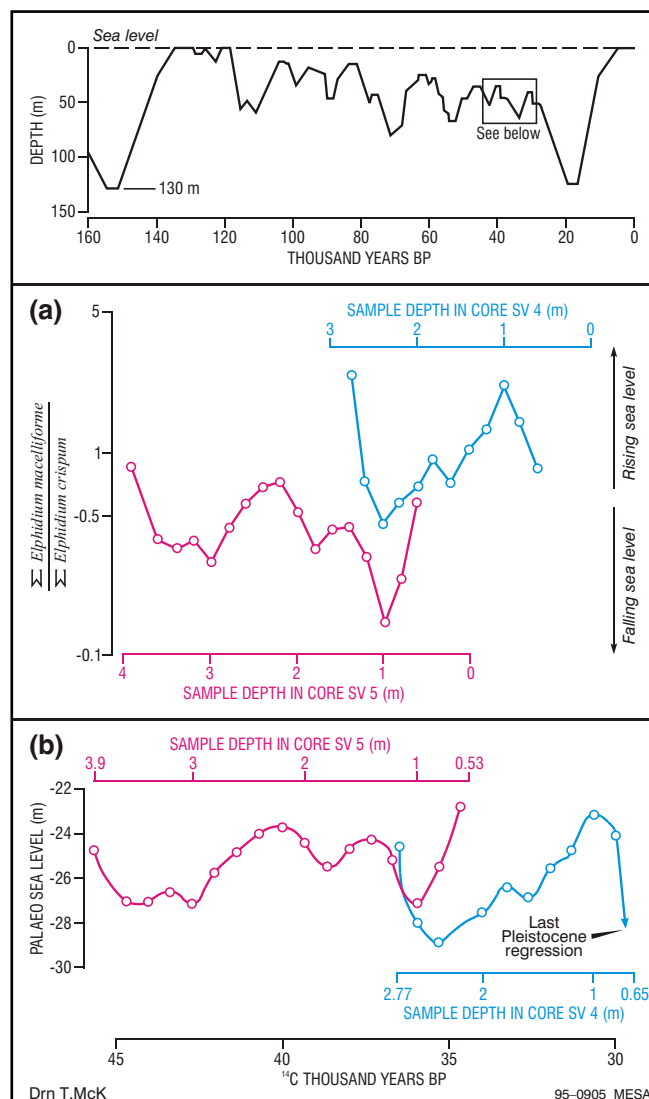


Fig. 11.11 Late Pleistocene sea levels in Gulf St Vincent for the period 45 000–30 000 years BP as interpreted from (a) the point count ratio of *Elphidium* species, and (b) ^{14}C data from vibrocores SV 4 and SV 5 (after Cann *et al.*, 1988); the holes are located on Fig. 11.10.

HOLOCENE COASTAL AND MARINE SEDIMENTATION

The sea level rose rapidly after the last glacial maximum, transgressing the continental shelf and entering the gulfs by 9 000 years BP, and reaching the present coastline ~7 000 years BP (Fig. 11.13). Since then, coastal and shelf processes and coastal sedimentation have progressively equilibrated, producing a range of depositional styles that relate increasingly to the contemporary hydrodynamic regime. Loading of the continental shelf by the rising sea also resulted in different degrees of coastal warping and a geographically variable apparent sea-level regression around the State's coastline over the past 6 000 years (Fig. 11.13).

The St Kilda Formation refers to all Holocene coastal and shelf sediments, and comprises various sedimentary facies deposited by a variety of marine processes. Its base is clearly defined, as it disconformably overlies Pleistocene marine or continental sediments, and its upper surface can be one of active sedimentation. Coastal seagrass bank and coastal aeolian facies (Lipson Formation and Semaphore Sand of Firman, 1966b), and coastal gypsum facies (Le Hunte Formation of Firman, 1974), are all members of the St Kilda Formation. Cann and Gostin (1985) and Belperio (1988b,c) provide a practical breakdown of the various facies of the St Kilda Formation as encountered in the coastal zone and shallow gulf waters north of Adelaide.

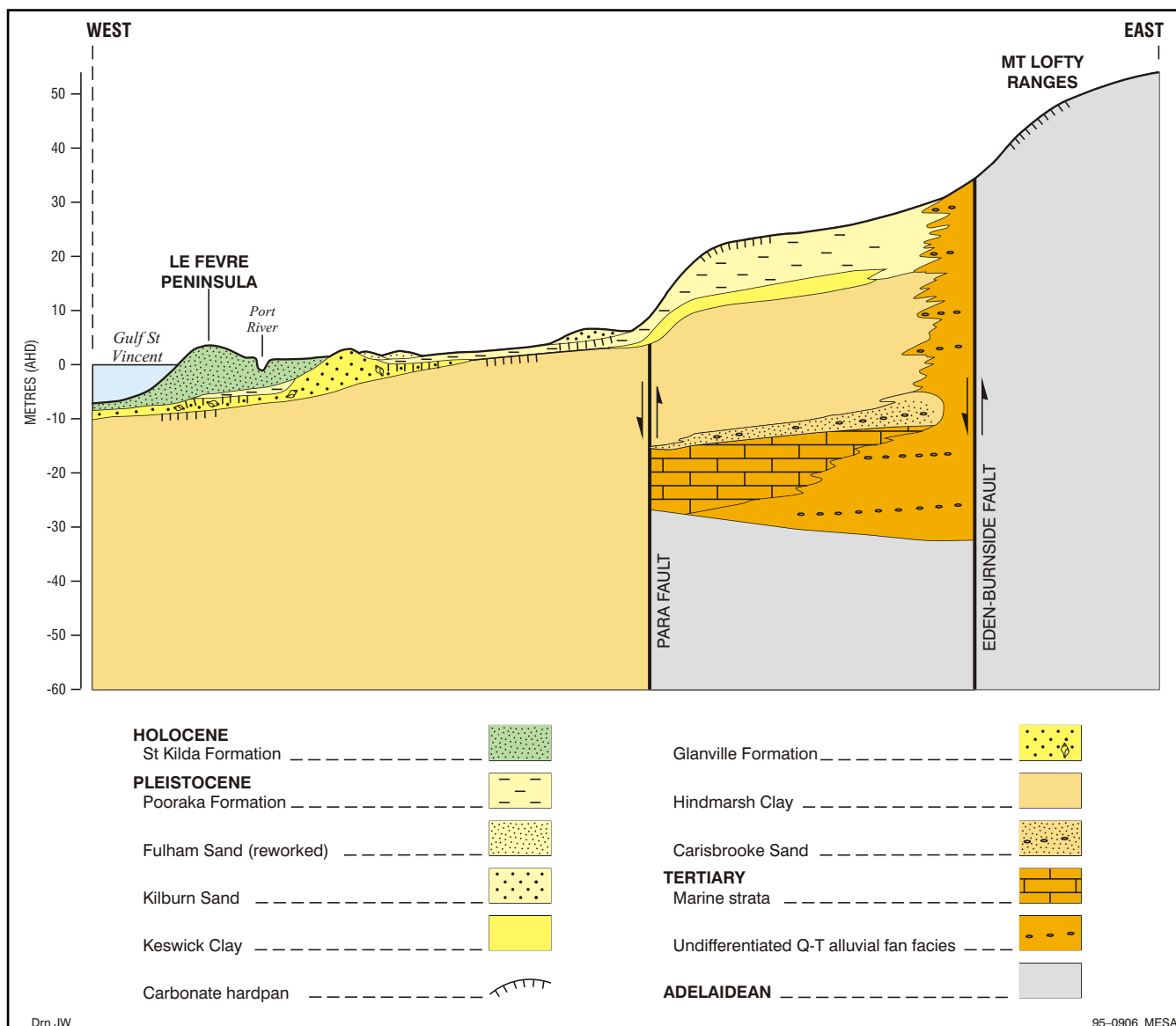


Fig. 11.12 Schematic cross-section from Le Fevre Peninsula to the Mount Lofty Ranges, showing relationships between Quaternary coastal marine and continental facies of the St Vincent Basin (after Sheard and Bowman, in prep.; Belperio and Rice, 1989). Not to scale.

Carbonate of biogenic origin dominates both shelf and coastal sediments around much of South Australia (Fig. 11.1). The paucity of continental detritus results from low relief of the interior and a predominantly arid climate. The prevailing coastal climate is cool-temperate in the south and semiarid in the north, with a concentration of rainfall in winter that is greatly exceeded by evaporation during the rest of the year. Consequently, coastal evaporites, particularly sulphates, are common. The oceanic coastlines of western Eyre Peninsula, southern Yorke Peninsula, Kangaroo Island and the South-East are affected by persistent southwesterly swells and dominant southeasterly to northwesterly winds, and tidal range is low. Within Gulf St Vincent and Spencer Gulf, the effect of swell decreases and local wind waves and tides become more important. Sea-surface temperatures on the ocean coastline have a seasonal range of ~14°C in winter to 19°C in summer. This increases significantly within the gulfs, ranging from 13 to 26°C near Adelaide and 13 to 28°C at the head of Spencer Gulf (Bye, 1976; Womersley, 1984). High summer evaporation raises the salinity of gulf waters, creating a seasonally variable lateral salinity gradient.

The gulf and shallow shelf waters have an outstanding richness of marine plants. Larger algae and seagrasses provide shelter and a substrate for smaller plants and animals including carbonate-

secreting biota such as coralline algae, foraminifera, molluscs and bryozoa. The general subtidal ecology is described by King (1981), Womersley (1981a,b) and Shepherd and Sprigg (1976), molluscan fauna are described by Ludbrook (1984), and relationships between the ecology and sediment components are discussed by Cann and Gostin (1985), Belperio *et al.* (1986, 1988) and Gostin *et al.* (1988). Site-specific investigations are referred to in the following syntheses of major depositional provinces, viz. the continental shelf, South-East coast, Lake Alexandrina, West Coast and Kangaroo Island, Spencer Gulf, and Gulf St Vincent.

Continental shelf

The modern shelf is an intrinsic part of the southern margin of Australia, a passive continental margin that was initiated by continental separation from Antarctica in the Cretaceous. Little detailed work has been undertaken on the contemporary shelf sediments and processes, outside Spencer Gulf and Gulf St Vincent. The pioneering work of Sprigg (1947), Conolly and von der Borch (1967), von der Borch *et al.* (1970) and Wass *et al.* (1970) has been supplemented by recent shallow and deep-penetration seismic studies, imaging and grab sampling by AGSO, MESA, and Flinders and Adelaide Universities (James and von der Borch, 1991; von der

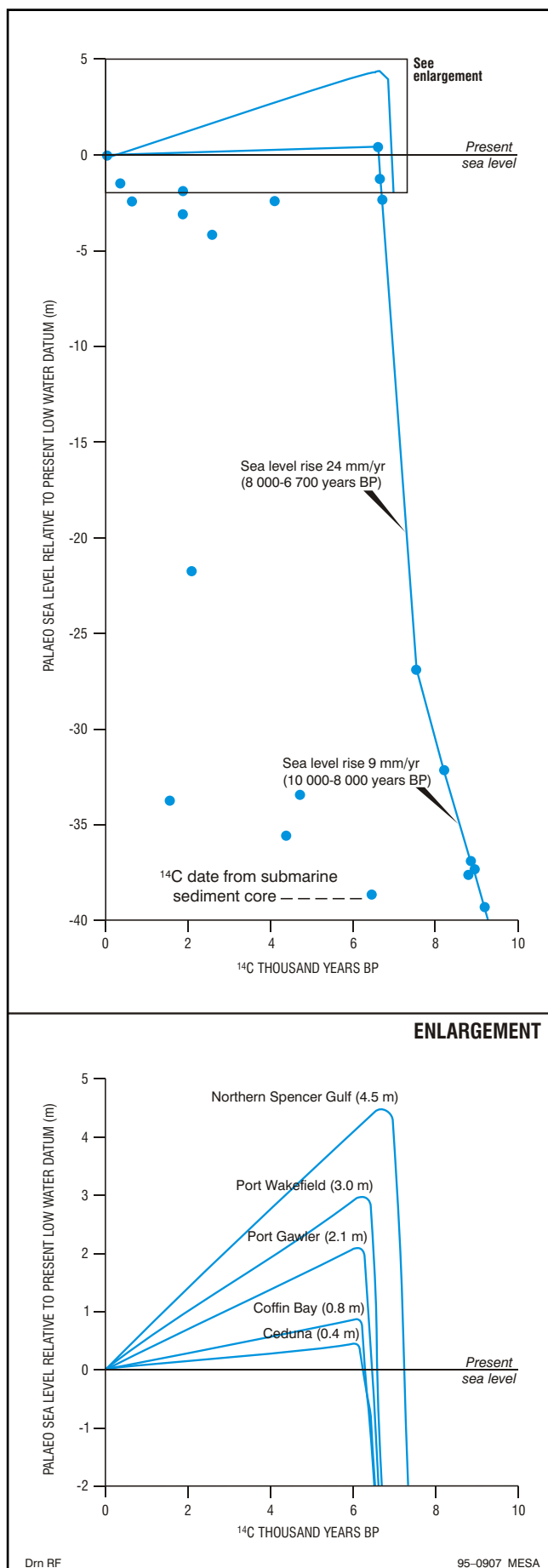


Fig. 11.13 Holocene sea-level rise for South Australia. The enlargement shows that variable but predictable sea-level regression has occurred around the State due to differential isostatic response.



Top: Coring Holocene coastal sediments – intertidal sand flats near Port Pirie. Photo 042525.

Centre: Holocene coastal facies – Semaphore Sand at Fowlers Bay. Photo 042526.

Bottom: Holocene coastal facies – mangrove woodland in Smoky Bay. Photo 042527.

Borch and Hughes Clarke, 1993; Davies *et al.*, 1988; James *et al.*, 1992; Bone and James, 1993; Boreen *et al.*, 1993).

The contemporary shelf is a storm-influenced, temperate, bioclastic to biolithic, palimpsest carbonate ramp with bryozoans providing the dominant component. It is broadest in the west (~200 km) and narrows to <20 km off the South-East coast. The wider shelf regions in general coincide with areas of shallow Palaeozoic and Precambrian basement (Lacepede Shelf and Ceduna Shelf, Fig. 11.1). Cainozoic sedimentary basins are partly masked, or are expressed in the contemporary margin morphology by such features as the Ceduna and Beachport Terraces.

The inner shelf floor (<50 m water depth) and entrances to Spencer Gulf and Gulf St Vincent are affected by persistent swell and winter storms. This region is characterised by receding residual islands and an eroded and sculptured seafloor composed of cemented Pleistocene and Tertiary carbonates and scoured Permian clay with gravel and boulder lag deposits. Hardgrounds and rocky surfaces host rich encrusting communities of sponges, algae, ascidians, bryozoans, crinoids, molluscs and worms. Bioclastic debris from these local carbonate factories, together with winnowed lithoclasts, are generally redistributed as patchy, coarse to medium-grained bioclastic grainstone sheets or as larger isolated sand banks in regions of flow separation or deceleration.

The wider expanses of continental shelf below fair weather wave base are largely dominated by a thin and patchy, rippled sand blanket derived from comminuted bryozoan, sponge and coralline algal detritus, and benthic and planktonic foraminifera. Offshore from the Murray (River) Mouth, fine quartz sand is an important component. Along the outer shelf, thicker deposits of coarse-grained bioclastic sands result from the increased productivity associated with oceanic upwelling. Dense bryozoan colonies, sponges and associated filter-feeding biota thrive in these nutrient rich waters. Below storm wave base, sediments are muddy but rich in bryozoa and foraminifera, and seaward-dipping clinoforms result from off-shelf sediment transport.

The continental slope is characterised by fine bioclastic sands of turbidite or contourite origin (Davies *et al.*, 1988; Wilson, 1988) and large-scale downslope slumping of shelf sediments. Foraminiferal and nannofossil pelagic ooze dominates the Ceduna and Beachport Terraces and the deeper ocean floor. Major submarine canyons are entrenched into the slope and provide conduits for shelf to ocean floor sediment transport (Sprigg, 1947; von der Borch, 1968, 1979) as well as windows into underlying strata exposed in the canyon walls. The Murray Group canyons, the locus for discharge of the Murray River during low sea-level stands, are deeply incised and topographically complex. Echidna and Du Couedic Canyons host seasonal saline outflows from Spencer Gulf that travel across the shelf and cascade over the shelf break (Lennon *et al.*, 1987). More westerly canyons, such as Ceduna and Thevenard Canyons, appear to be currently inactive (Davies *et al.*, 1988). Beneath a veneer of pelagic ooze, older canyon-floor sediments comprise mixed terrigenous and carbonate, turbidite-derived sand and debris flow deposits. The carbonate component includes shelf-derived biogenic detritus, whilst the terrigenous component includes well-rounded quartz sand and gravel with heavy minerals. Rocks of Precambrian to Tertiary age, including granodiorite, alkali basalt and carbonates, have been dredged from the walls and floors of these canyons.

South-East coast

Holocene deposits along this coastal plain display a style of depositional facies and hydrodynamic and hydrochemical processes similar to those inferred for the older Pleistocene coastal barrier complexes (Fig. 11.3). The coast receives the full force of the dominant southwesterly waves, with swell at Port MacDonnell exceeding 2 m for 68% of the year (Short and Hesp, 1984). Semidiurnal tides are microtidal with range consistently <1 m. Storm surges and longer period oscillations of sea level have amplitudes of up to 1.5 m.

The barrier complex that makes up Younghusband Peninsula is ~180 km long and is backed by a shallow lagoon (The Coorong). Offshore gradients increase from north to south along The Coorong beach, giving rise to a pronounced longshore gradient in near-shore wave energy and beach morphology. In turn, this results in longshore variations in beach width, sediment grainsize, foredune stability and transgressive dune activity. Away from the Murray Mouth, carbonate content of both the beach and dune sands and The Coorong lagoon sediments progressively increases and grainsize decreases. Transgressive dune systems are characterised by unvegetated active dune sheets, large-scale parabolic and transverse dunes reaching 30-40 m above sea level, and deflation basins or blowouts that extend down to sea level. Much of the Holocene barrier complex that makes up Younghusband Peninsula consists of former transgressive dune sheets now stabilised by vegetation. Fossil soil horizons exposed in contemporary blowouts indicate periods of former dune stability, vegetation and soil development. Aboriginal remains, including middens composed largely of the cockle *Plebidonax deltoides*, are commonly associated with the palaeosols. Radiocarbon dating of humus extracted from the palaeosols, or of associated bioclastic or cultural remains, indicate repeated episodes of dune transgression and stabilisation dating from the present back to 6 000 years BP (Ohmori *et al.*, 1987).

The Coorong is a 130 km long, back-barrier saline lagoon, created ~6 000 years BP behind the evolving Younghusband Peninsula. It consists of a number of interconnected basins up to 4 km wide and 4 m deep that become increasingly restricted, hypersaline and ephemeral towards their southern end (Figs 11.14, 11.15). Waters are derived from the ocean via the Murray Mouth, from the Murray River, and from seasonal rainwater and unconfined groundwater. Distinct seasonal fluctuations are superimposed on the marked salinity gradient. There is also a pronounced seasonal oscillation of water level of ~1 m that results in part from higher winter sea



Top: Southeasterly view of The Coorong lagoon and Holocene barrier shoreline. Photo 042528.

Bottom: Mudcracked dolomicrite of The Coorong lagoon. Photo 042529.

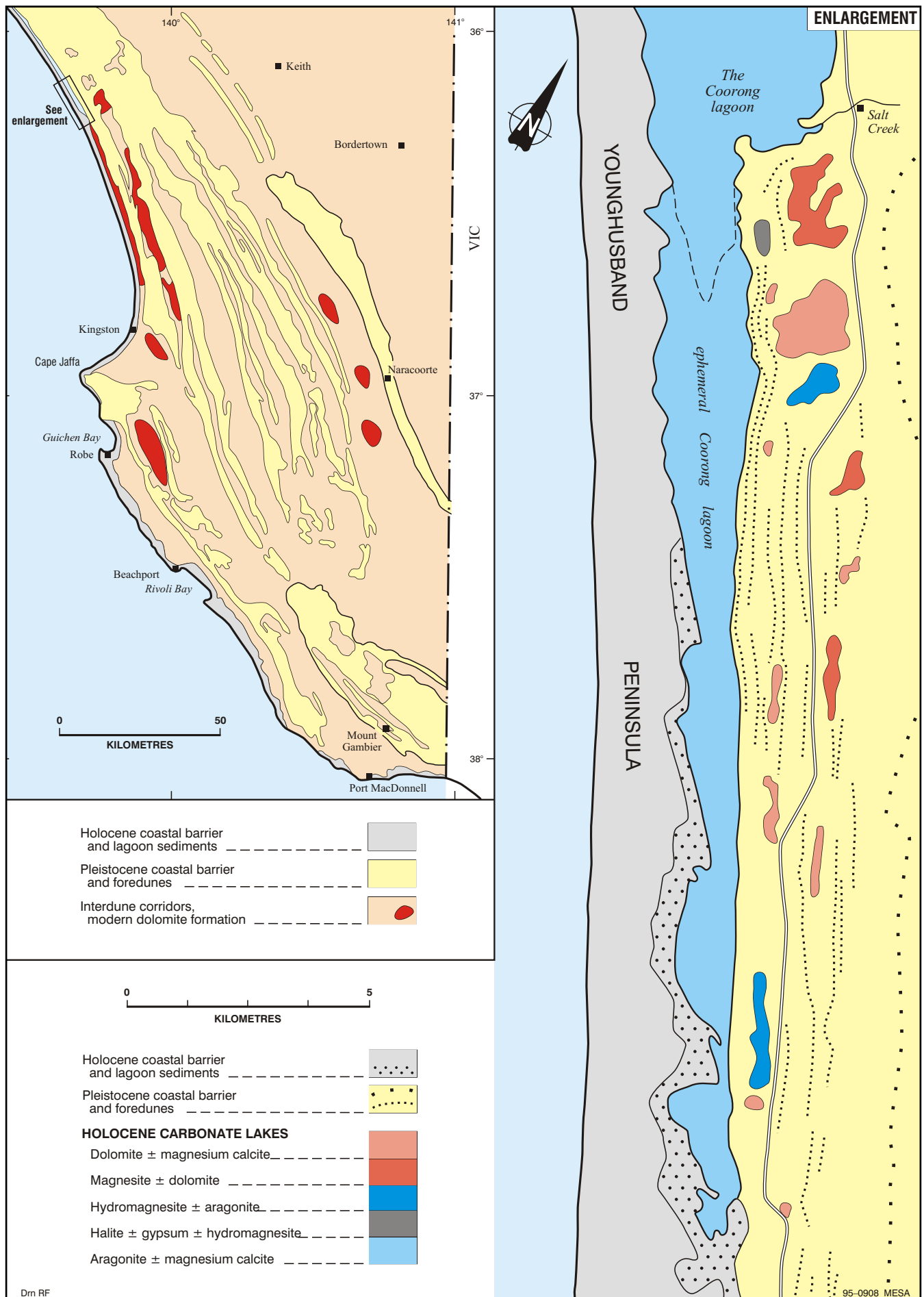


Fig. 11.14 Coastal and inland carbonate precipitating lakes and swamps of the South-East (after von der Borch, 1965; von der Borch and Lock, 1979).

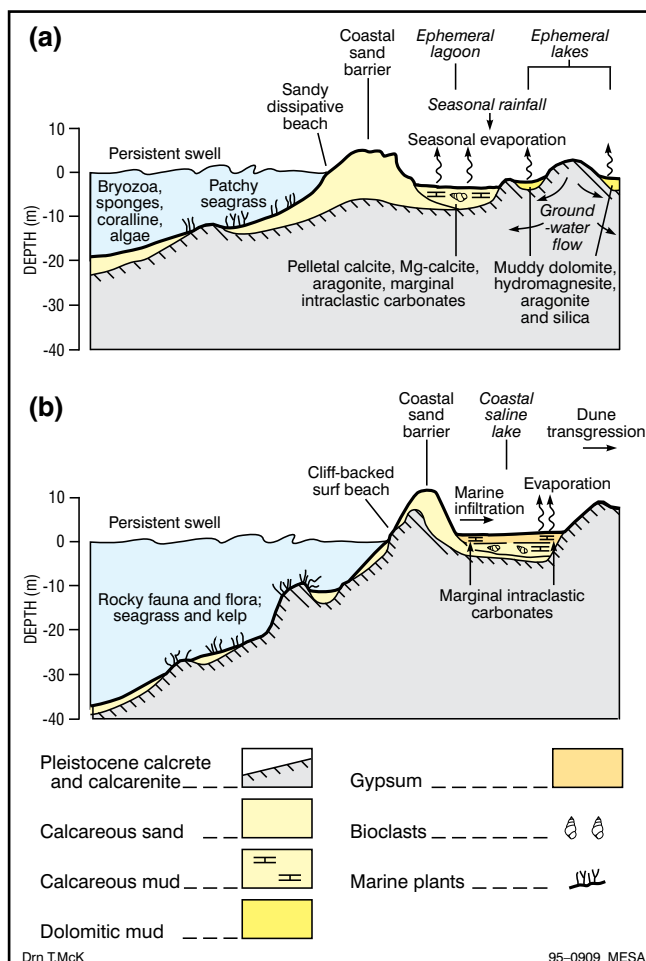


Fig. 11.15 Schematic representation of high wave energy coastal facies of (a) The Coorong region (swell environment, cool temperature climate), and (b) Eyre Peninsula coastline (swell environment, semi-arid climate).

levels, as well as the alternating dominance of winter rainfall and summer evaporation. Sediments of The Coorong lagoon consist of bioclastic magnesian calcite and aragonite sand and pelletal mud derived from molluscs, ostracodes and foraminifera (Brown, 1965). The gastropod *Coxiella confusa* and the foraminifera *Ammonia beccarii* are characteristic. Cores indicate that the protected marine lagoon became increasingly restricted as sedimentation progressed, and hypersaline conditions subsequently developed in the southern half of the lagoon. Aragonitic mud characterises the seasonally exposed ephemeral lagoon floor. Drainage channels constructed to the south of The Coorong have stopped the overflow from swamps reaching the lagoon in years of heavy winter rainfall and may have accentuated the trend towards hypersalinity.

Holocene coastal uplift has resulted in ~1 m of relative fall of sea level, a factor that has compounded the restrictive effects of regressive sedimentation. As a result, at its southern end, the ephemeral Coorong lagoon has further developed into a number of isolated, ephemeral lacustrine depocentres that are now largely removed from marine influence and dominated by the groundwater regime. These ephemeral lakes have a similar biota to the ephemeral lagoon facies, but also precipitate a variety of fine-grained carbonate minerals during their annual evaporative phase as discussed by Alderman and Skinner (1957), von der Borch (1976), Warren (1988) and Rosen *et al.* (1989). These include partially ordered dolomite and poorly crystallised protodolomite, magnesite and hydromagnesite, magnesian calcite, aragonite and monohydrocalcite.

Dolomite also precipitates in inland lakes and swamps within broad interdune corridors (Fig. 11.14) but much less is known of hydrogeochemical parameters controlling this process. Most of the dolomite-bearing lakes are situated where mean annual rainfall is

between 500 and 700 mm. They support a prolific flora of algae and higher aquatic plants (*Ruppia* sp.) during their less saline winter stage, but summertime evaporation causes extreme salinity variations from 10 to >180 per mil. With progressive evaporation and desiccation in early summer, distinct carbonate assemblages precipitate or form around the lake margins. The assemblages reflect meso- and macro-scale variations in geography, topography and water level, and the relative roles of marine to meteoric groundwater mixing, plant abundance, alkalinity and water chemistry. Algal stromatolites, including stratiform, crenulate and globular forms, are associated with hydromagnesite-bearing lakes. Monohydrocalcite is common in sapropelic, more permanent lakes closer to Robe, whilst in higher rainfall areas closer to Mount Gambier, only freshwater lakes and swamps are found.

Burne *et al.* (1980), Cann and De Deckker (1981) and Burne and Ferguson (1983) described the distinctive biota and sediments of the seasonally hypersaline lakes, including oncolites, stromatolites, ostracodes, foraminifera and charophyte oogonia. Warren (1988) and Rosen *et al.* (1988) described the vertical sequences generated by sediment infilling of the lakes. These include a basal marine or estuarine skeletal grainstone, an organic-rich unit and a laminated lacustrine unit. Desiccation of the ephemeral lake margins produces fenestral and laminar crusts, polygonal cracks, tepee structures and mud chip breccias that have analogues in ancient carbonates (Muir *et al.*, 1980).

South of Robe, a highly irregular coastal morphology has resulted from exposure of Pleistocene calcarenites and Tertiary limestone to the high wave energy of the Southern Ocean (Fig. 11.3). This has resulted in pocket or repressed beaches alternating with a rocky coastline partially blanketed by Holocene dune sands. Flint pebbles and cobbles reworked out of the Tertiary Gambier Limestone are concentrated in pebble beach deposits between Lake Bonney and the Victorian border (Flint *et al.*, 1989). More extensive transgressive dunefields are developed from Lake George to Lake Bonney. Guichen and Rivoli Bays occupy two major breaches in the Robe Range that have largely been infilled by onshore transport of sand and seaward progradation of foredune ridges. Within Guichen Bay, more than 80 foredune ridges up to 10 m high extend back 3.5 km and date to 7 000 years BP (Thom *et al.*, 1981). The large coastal lakes are major groundwater discharge points with no direct connection to the sea. Water quality varies from fresh to hypersaline depending on the balance between marine and continental groundwater infiltration, evaporation, rainfall and artificial drainage measures. The lakes contain abundant marine fossils in their older sediments indicating a marine origin and the existence of a major marine lagoon complex behind the Robe Range in the early to middle Holocene. Aboriginal middens, marked by food remains such as shells and bones, flint tools and charcoal, are commonly associated with these environments.

Lake Alexandrina

This lake into which the Murray River flows has been extensively modified by human intervention and a system of barrages now restricts saltwater intrusion from the sea. Prior to construction of the barrages, the lake was a brackish estuary and, at times of low river flow, marine water was able to penetrate far up the lake system and the Murray River. The margins of the lake contain extensive low-lying areas which are former embayments but are now infilled with sediments. Studies by Altmann (1976), Spagnuolo (1978) and von der Borch and Altmann (1979) have confirmed that these marginal deposits are of Holocene age. They contain a consistent lithology, characterised by a basal eutrophic lagoon facies, a main lacustrine facies and a regressive lake facies. The eutrophic facies (sapropel and diatomite), dated at $6\,930 \pm 150$ years BP, was attributed to the initial Holocene development of Lake Alexandrina as a result of rising sea levels and lake damming. A sea level ~1 m above present level ~6 500 years ago was responsible for significant lake expansion and marginal lacustrine deposition. Gypsum is associated with marginal deposits at Cooke Plains.

West Coast and Kangaroo Island

The 500 km sweep of southwest-facing coastline from Head of Bight to Kangaroo Island and the gulf entrances is dominated by persistent south to southwest ocean swell and strong onshore winds. Coastal geometry and morphology are controlled by the undulating surface of calcreted Pleistocene aeolianites and underlying Precambrian basement. Within protected embayments and back-barrier lagoons, a relatively complete range of peritidal and intertidal to supratidal facies is preserved. Unprotected coastal reaches are characterised by eroding cliffs, islands and shoals, headland-attached barriers, large transgressive dunefields and coastal saline lakes (Fig. 11.15). The largest active dunefields are near Coffin Bay.

Equally extensive transgressive dunes reaching 10 km inland occur on the south and west coasts of Kangaroo Island (Fig. 11.1). The fine, well-sorted dune sands consist of fragmented remains of bivalves, gastropods, coralline algae, echinoids and bryozoa, and abraded foraminifera. Short (1988), Short and Fotheringham (1986) and Short *et al.* (1986) provided detailed accounts of the variety of sandy barriers and dune complexes on Kangaroo Island and Eyre Peninsula, and the morphodynamic processes responsible for their formation. The most extensive dunefields were initiated by the post-glacial marine transgression, but many of these systems have now largely exhausted their sediment supply. Extreme cases are stranded cliff-top dunes that have had their sand ramp completely removed. Carbonate dominates the mineralogy of most beach, barrier and dune deposits, though siliciclastics can be locally important near bedrock cliffs and platforms.

Stratigraphic studies of several protected embayments (e.g. Tourville Bay, Smoky Bay) demonstrate the influence of local bioclastic productivity (Freeman, 1985; Lablack, 1985; Belperio *et al.*, 1988; Ludbrook, 1988). Intertidal sand flats, mangroves and cyanobacterial-halophyte marshes have developed and are prograding across shoaling peritidal banks dominated by the seagrass *Posidonia australis*. This biotic control of sediment facies is similar to that described for the major gulfs. Benthic foraminifera are a major component of the sediment, as are a host of molluscan species. The relative abundance of different species reflects the degree of wave protection, elevation, inundation frequency and tidal zonation of environments. Tourville Bay originally covered an area of 100 km² but has been infilled largely with bioclastic sediment. The vertical succession of sedimentary facies (Fig. 11.16) reflects the regressive sedimentation and mimics the lateral zonation of environments visible today across the embayment. Embayments with restricted entrances, such as Venus Bay and Coffin Bay, are partially infilled by flood-tide deltas and associated sand shoals. These sediments are largely transported from the adjacent inner shelf and eroding cliffed coastline, and contain a high portion of reworked carbonate grains.

Where embayments have been completely cut off by transgressive dunes, such as Fowlers Bay and Marion Lake, or where sea water was able to seep into pre-existing coastal depressions such as Sleaford Mere and Lake MacDonnell, well-developed gypsum and marginal carbonate precipitates occur as a result of the very high summer evaporation. The variety of carbonate and gypsum sediments and structures that fill these depressions are illustrated by Forbes (1960), von der Borch *et al.* (1977), Warren (1982a,b, 1983b), De Deckker *et al.* (1982) and Gostin *et al.* (1988). Most of these coastal saline lakes precipitate gypsum and aragonite because of continual brine renewal by marine groundwater. Crystalline gypsum precipitated freely and continuously in the young (6 000 years BP) brine ponds, but progressive shallowing, seasonal freshening and water-level variations have created aragonite pelletoid layering and a planated selenite surface that approximates mean sea level. Simultaneously, gypsarenite lunettes and a veneer of supratidal algal-gypsite silt developed, whilst various carbonate structures formed around the saline lake margins. These marginal carbonates include aragonitic boxwork boundstones and fenestral to intraclastic crusts with tepees and mudcracks, calcitic and aragonitic stromatolites, oncolites and algal mats, and dolomitic and calcitic marls. In general, this western coastal province is dominated by gypsum

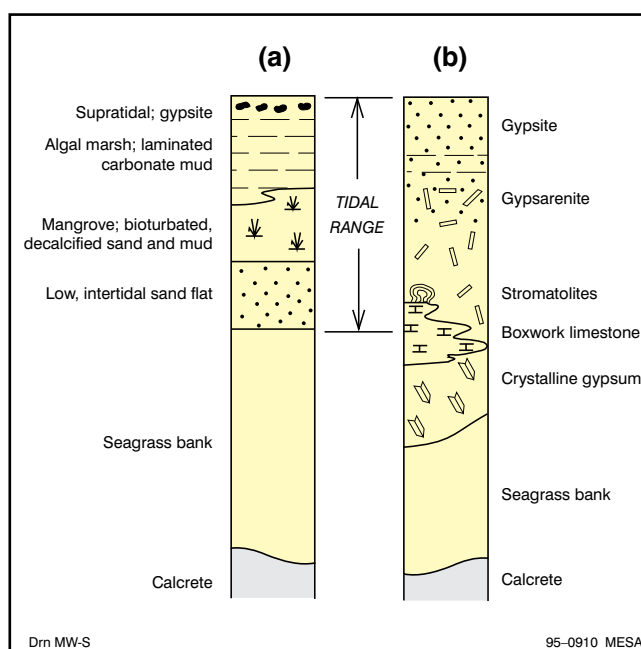


Fig. 11.16(a) Vertical sequence generated in protected embayments such as Tourville Bay and Smoky Bay, compared to **(b)** a sequence generated in embayments that degenerate into saline lakes through inlet blockage, such as Marion Lake and Fowlers Bay. Coastal saline lakes that were never directly connected to the sea (e.g. Lake MacDonnell, Head of Bight) lack basal marine strata. For gypsum terminology see Warren (1982b). Tidal range varies from 2 to 4 m, and the subtidal seagrass facies may be up to 10 m thick.

and aragonite precipitation rather than dolomite as recorded in the more meteoric groundwater-dominated ephemeral lakes of the South-East. Lake MacDonnell is a particularly significant gypsum resource with reserves in excess of 500 Mt (Fig. 11.17). Holocene gypsum derived from marine waters is termed Le Hunte Formation (a facies of the St Kilda Formation).

At Clare Bay, west of Ceduna, human footprints, together with tracks of emus, kangaroos and wallabies, are preserved in lithified dolomitic and calcitic carbonate mud around the margin of a small saline lake (Belperio and Fotheringham, 1990). The case-hardened carbonate crust consists of a clotted fenestral fabric of intermixed dolomitic and calcite with reworked pisoliths and a few altered bioclastic grains. The prints were formed ~5 000 years ago as humans and other animals walked over soft carbonate mud marginal to the salt lake; lithification and progressive case-hardening occurred soon after.



Crenulate algal mats, Tourville Bay. Photo 042530

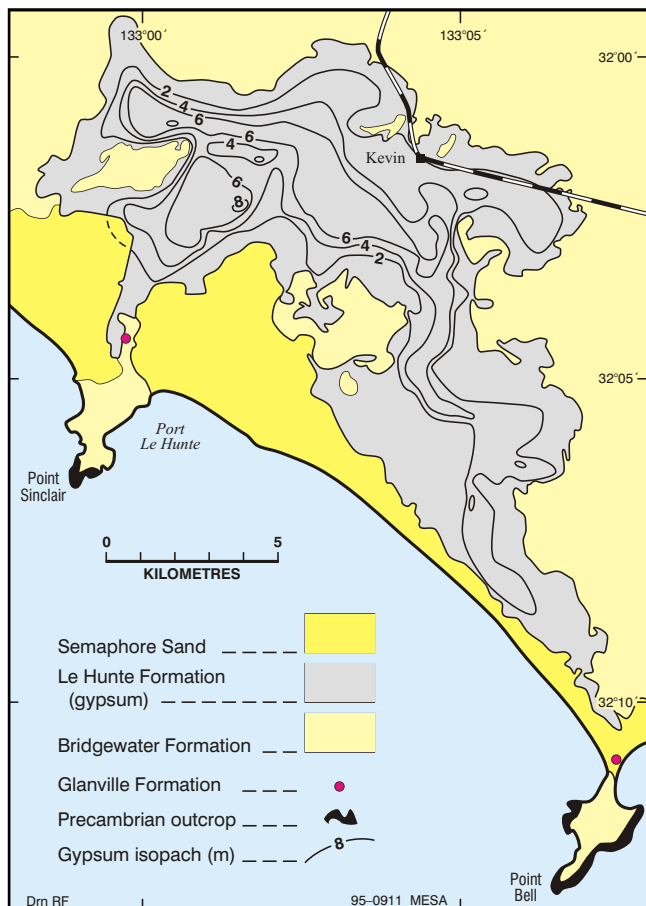


Fig. 11.17 Lake MacDonnell gypsum deposit showing simplified geology and gypsum isopachs (after Warren, 1983b; Flint, 1987).



Spencer Gulf

Bounded by Yorke and Eyre Peninsulas, this gulf forms a long, shallow marine re-entrant into the semiarid continental interior. Limited freshwater inflow and high evaporation have resulted in a hydrologically inverse estuary, with a positive salinity gradient from south to north. Swell is largely excluded and tides enhanced, with spring tide ranges reaching 3.9 m at Port Augusta. Holocene sedimentation is dominantly biogenic carbonate and reflects temperate (13–28°C), high salinity (35–49 per mil) waters. The skeletal detritus consists largely of bivalves, gastropods, foraminifera, echinoids, coralline algae and bryozoa (foramol skeletal association).

Distinct ecological and morphological zonations are determined by water depth, degree of subaerial exposure (tide level), and local wave climate. Ten major sedimentary environments in subtidal, intertidal and supratidal zones were identified by Burne and Colwell (1982), Gostin *et al.* (1984), Belperio, Hails *et al.* (1984) and Belperio *et al.* (1988) in northern Spencer Gulf.

In water deeper than 15 m, the sea floor consists of a smooth thin sheet (0.5–2 m thick) of Holocene calcareous sediment (Fig. 11.18). Megaripples and sandbanks are locally present and cores indicate that the sediment is a poorly sorted mixture of calcareous mud, sand and shells. Filter-feeding benthic organisms (pelecypods, bryozoa, sponges and soft corals) are the main contributors to the substrate. In the narrow reach north of Stony Point, enhanced tidal flows have generated linear banks and broad areas of megaripled bioclastic sand dispersed over a scoured channel floor.

Shallow subtidal areas (<15 m) support extensive seagrass meadows, particularly *Posidonia australis* and *P. sinuosa*, with subordinate growth of *Heterozostera tasmanica*, *Amphibolis antarctica* and *Halophila ovalis*. Growth of *Posidonia* species is particularly vigorous, with high leaf blade production and turnover (West and Larkum, 1979). The seagrass meadows support a rich and diverse fauna and flora characterised by coralline algae (*Goniolithon*, *Lithothamnium*, *Jania*), benthic and epiphytic foraminifera, encrusting bryozoa, gastropods, hydroids, echinoderms and ascidians. Epibenthos and infauna include crustaceans, the large bivalves *Pinna*, *Pecten*, *Ostrea* and *Katelysia*, and numerous other molluscs (Shepherd, 1983).

Sedimentologically, the seagrass meadows act as major carbonate factories, with high rates of *in situ* carbonate production and bioclastic skeletal accumulation. Also trapped are terrigenous and aeolian detrital grains and terrigenous mud. Decaying leaves, roots and rhizomes contribute to a strongly reducing environment, and the resulting sediment is a very poorly sorted, organic-rich skeletal packstone. Decaying leaf sheaths of *P. australis* leave behind a residue of fine, pale, lignin-impregnated cellulose fibre. Sediments accumulating in *P. australis* seagrass meadows are therefore characteristically bound in the subsurface by the fibres and create organically bound carbonate banks. In the Redcliff region (Fig. 11.19), many of the seagrass banks have accumulated to low water level, and are now mantled by thin intertidal sand sheets.

Environments regularly inundated by neap-to-spring tides include bare or *Zostera*-colonised intertidal sand flats, and mangrove-algal and halophyte-algal marshes. Areas of bare, saline flats with isolated stranded beach ridges characterise the supratidal zone. Small beaches with a wide frontal sand flat are developed along shorelines exposed to southwesterly winds and wind-generated waves. There is a well-developed surficial zonation of environments and sediments in the northern gulf. Monospecific stands of the mangrove *Avicennia marina* form a low woodland from about mean sea level to spring high tide level. Progressively higher vegetation zones are a *Halosarcia-Sarcocornia* samphire

Top: Tepee disrupting the fenestral carbonate crust on Marion Lake. Photo 042531.

Bottom: Holocene stromatolites on Marion Lake. Photo 042532.

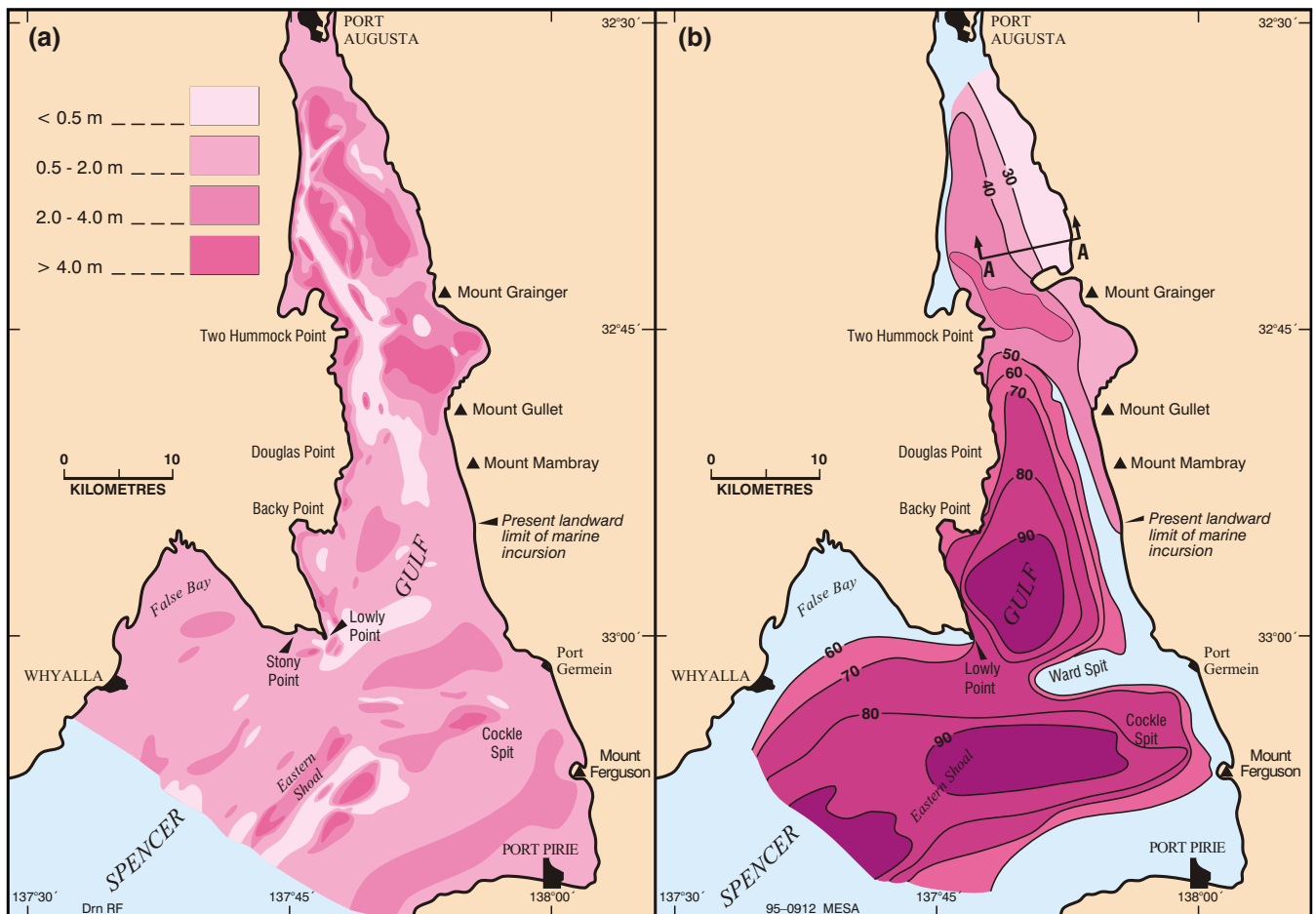


Fig. 11.18(a) Isopachs of Holocene sediments in northern Spencer Gulf, and **(b)** calcium carbonate content of surface sediments, contoured at 10% intervals (after Hails *et al.*, 1984b). Section A-A is illustrated on Figure 11.19.



marsh, a *Sclerostegia-Halosarcia* samphire marsh, and an *Atriplex* sp. saltbush community. Cyanobacterial mats occur throughout the mangrove and halophyte zones, with best growth occurring 2.5-3.5 m above low water. The evaporative supratidal flats are gypsiferous and locally dolomite nucleates within a fenestral calcitic mud. Desiccation and deflation generally restricts sediment accumulation on the flats, although small depressions act as ephemeral saline lakes and localised centres of evaporite deposition. The surface of the flats represents the upper supratidal limit of sediment accumulation in the prograding tide-dominated coastal environment.

Belperio, Hails *et al.* (1984) concluded that the production of skeletal bioclastic detritus within seagrass meadows and rapid vertical accumulation of coastal seagrass banks are the dominant mechanisms controlling coastal evolution in the protected Spencer Gulf. Coastal progradation and shoreline regression in the northern gulf has been further enhanced by a relative sea level fall of up to 4.5 m, attributed to hydro-isostatic upwarping of the distal parts of the gulf relative to the continental shelf (Nakada and Lambeck, 1989). This has resulted in stranded or inactive supratidal flats along the eastern shoreline and stranded shingle and cobble beach ridges along the western shore. Bioclastic beach ridges are scattered over the inactive supratidal flats and comprise a significant shellgrit resource (Dubowski, 1989).

Top: Tidal zonation of supratidal flats, samphire-algal marsh, mangrove woodland and intertidal sand flats, Point Paterson, northern Spencer Gulf. Photo 042533.

Bottom: Subtidal seagrass *Posidonia australis*, with a diverse encrusting calcareous fauna and flora. Photo 042534.

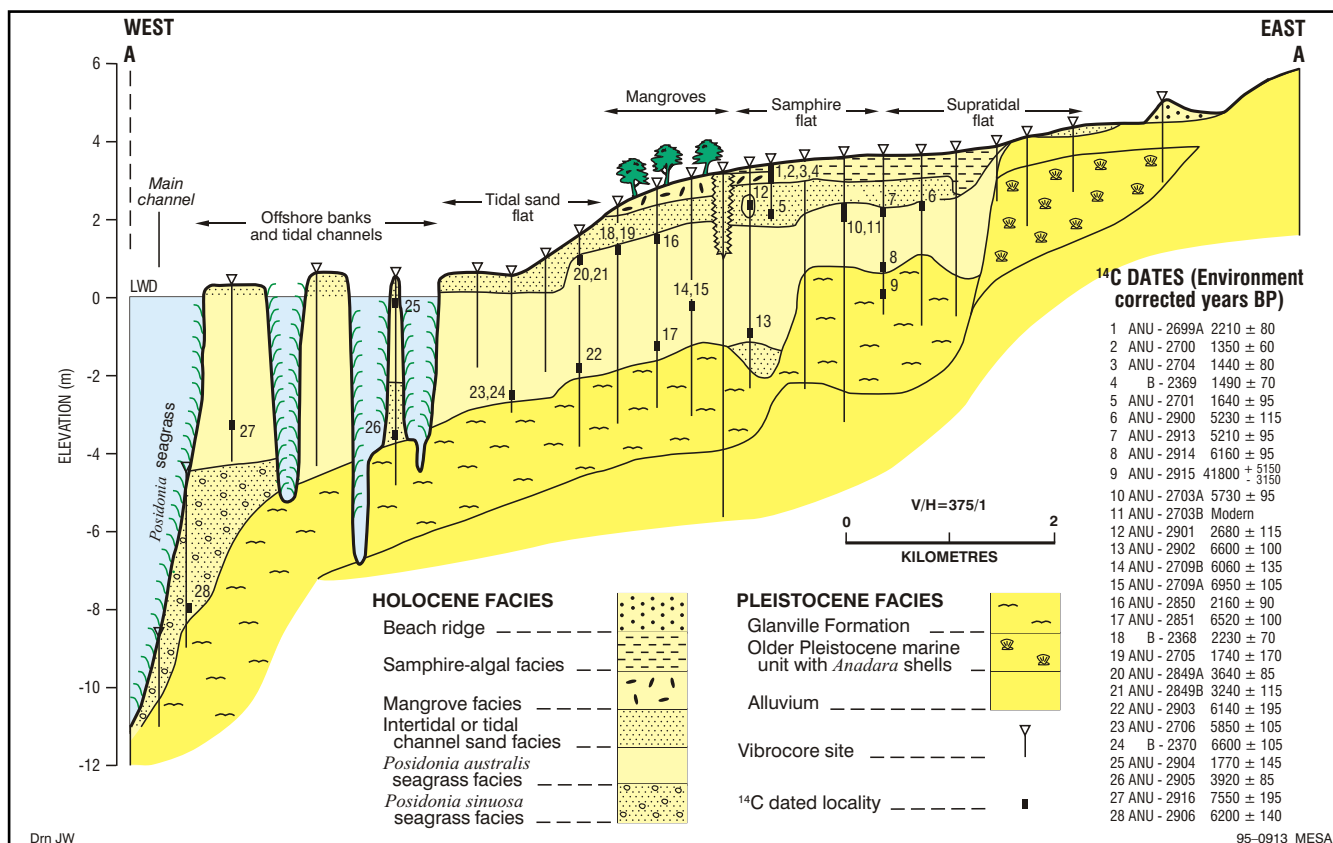


Fig. 11.19 Coastal cross-section at Redcliff, northern Spencer Gulf, showing surficial environments, subsurface facies and radiocarbon dates relative to low water datum (LWD; after Belperio, Hails *et al.*, 1984); the section is located on Figure 11.18.

Gulf St Vincent

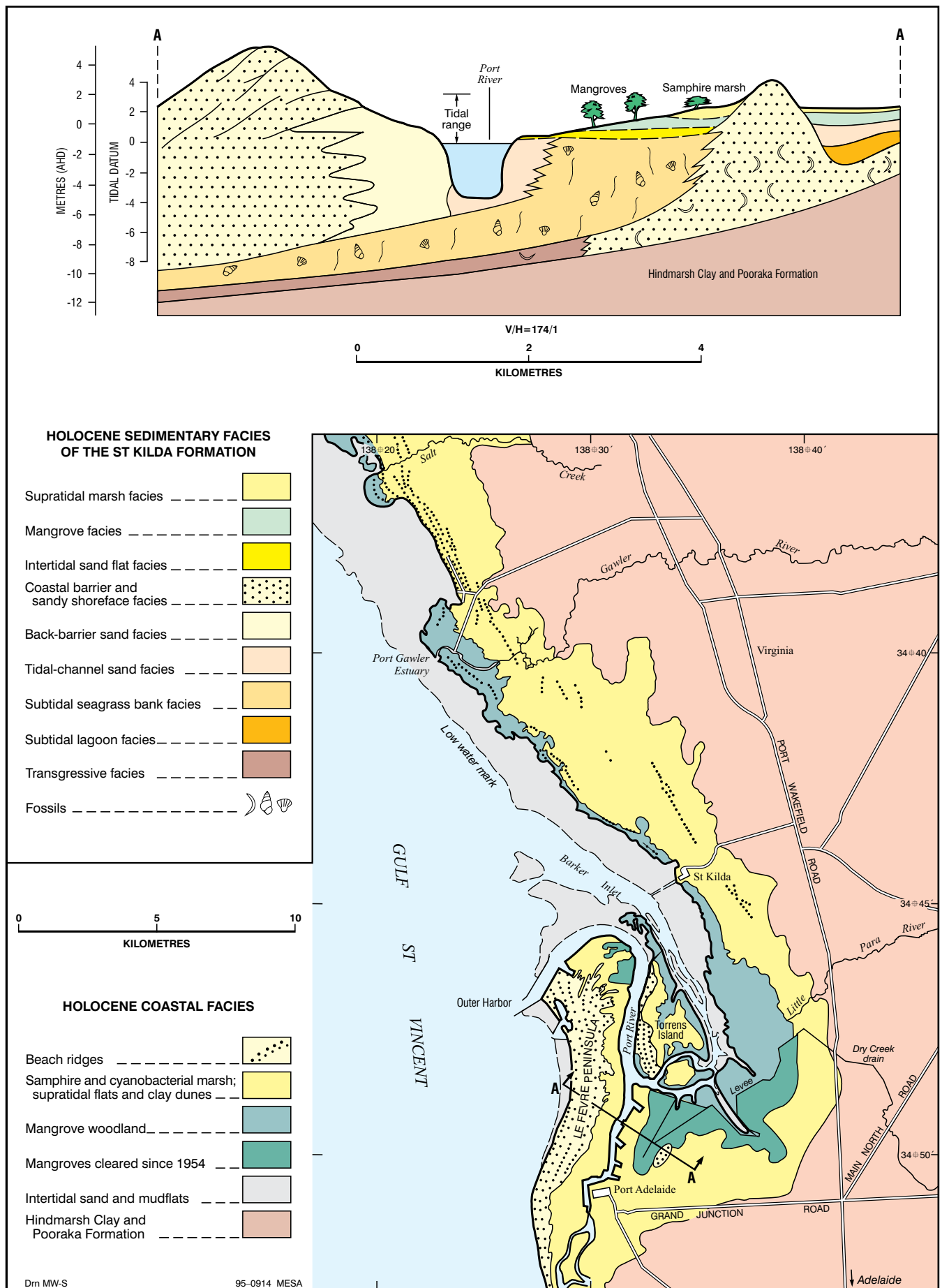
Gulf St Vincent is a protected marine re-entrant similar to Spencer Gulf, although it has a reduced north-south salinity gradient and less enhancement of tidal levels (3.0 m range). Holocene sediments are similarly dominantly biogenic carbonates, with quartzose sands restricted to the littoral zone south of Adelaide. The 30-70 m deep entrances to the gulf are scoured by strong tidal currents, and the sea floor generally consists of eroded and sculptured Pleistocene and Tertiary calcarenite. Sponges, ascidians, algae, bryozoans, crinoids and gastropods colonise the rocky surfaces, providing skeletal debris that is swept into the gulf. The central floor of the gulf, a flat plain 30-40 m below sea level, retains its last glacial lacustrine morphological elements beneath a thin (0.5 m) blanket of Holocene marine muddy, shelly sand. Benthic foraminifera are particularly abundant with the distribution of species closely related to water depth (Cann *et al.*, 1988, 1993).

In the shallow gulf margins (0-20 m), patchy to thick seagrass meadows (dominantly *Posidonia*) with associated epiphytes and epibenthos dominate the sea floor. The sediment is a coarse, poorly sorted, shelly sand, with carbonate content typically >90%. Intertidal and supratidal flats display a zonation of plants, sediments and environments similar to that described for Spencer Gulf. As sediment accumulates, *Posidonia* seagrasses grow upwards, binding the sediments in the subsurface, but are constrained by their inability to survive intertidal exposure. Growth of seagrass banks to low water cause the outer margin of the intertidal zone to migrate seawards. Very wide peritidal flats have thus been created north of Adelaide, across which a zonation of intertidal environments has subsequently been established. The prograding intertidal sand flats north of Adelaide are colonised by other species of seagrass (*Zostera*, *Heterozostera*) that tolerate subaerial exposure. Like *Posidonia*, *Zostera* and *Heterozostera* trap sediment, bind the substrate and contribute organic matter and microscopic epiphytic bioclasts to the sediment. Mangrove woodlands are well developed

in the mid-intertidal zone from Port Adelaide to the Light River. Finer sediment accumulates in this environment, and intensive burrowing by small crabs thoroughly homogenises and oxidises the sediment. Aggradation proceeds to levels where the mangrove woodland is inundated for only brief intervals during high tides. At these levels, where mangrove growth is stunted and die-back occurs, is a transition to samphire and cyanobacterial vegetation. Supratidal flats represent the upper limit of accumulation and marine flooding. On the floors of these flats or sabkhas, cyanobacterial mats intermittently flourish and photosynthesis favours chemical nucleation of carbonates and sulphates.

Overall, the Holocene sediments form a coastal wedge that is thickest (up to 10 m) beneath the contemporary intertidal zone in the Port River estuary. Here, Holocene morphological development resulted from coastal seagrass bank growth and estuarine infilling in the lee of the northward migrating Le Fevre Peninsula. Significant accumulation (>10 m) has occurred beneath Outer Harbor, Torrens Island and the Barker Inlet intertidal marshes (Fig. 11.20). South of Adelaide, much of the shallow sea floor is devoid of Holocene sediment cover and Tertiary strata crop out in a coast-parallel band from Port Stanvac to Sellicks Beach (Belperio *et al.*, 1990).

The earliest observations of Holocene coastal sediments were made at Port Adelaide and Dry Creek by Tate (1879b) and Howchin (1888, 1912). The 'low-lying swampy tract' around the gulf shoreline to the north of Port Adelaide (Fig. 11.20) was subsequently termed St Kilda Formation, and attributed to a 'highstand of the Flandrian Sea' by Daily *et al.* (1976). Belperio *et al.* (1983) refuted the concept of a higher Holocene sea level for the Adelaide region, based on the preserved lithostratigraphy, and this was confirmed by subsequent biostratigraphic studies by Cann and Gostin (1985) and Belperio *et al.* (1986). Other coastal investigations north of Adelaide have been undertaken by Johnson (1977), Belperio (1985a,b,c), Belperio and Bateman (1986), Bowman and Harvey (1986), and Belperio and Rice (1989).



CONTINENTAL SEQUENCES AND SOILS

The most characteristic feature of the State's interior is the widespread extent of arid and semiarid landforms and sediments (Fig. 11.1). More than 80% of the State receives <250 mm of annual rainfall, and evaporation exceeds precipitation by more than 10 to 1. However, geological evidence indicates that aridity is a relatively youthful feature, and that climatic conditions fluctuated throughout the Quaternary between arid and moister phases (Bowler, 1976). Whilst Pleistocene climatic cyclicity has been well documented in glaciated regions of the Northern Hemisphere and in the oxygen isotope stratigraphy of the deep oceans, the effects of these changes on the Australian continental interior are not as well documented. Here, sedimentation was episodic and the sedimentary record discontinuous. Thus, short periods of deposition in a wide variety of environments were separated by long intervals of erosion, reworking and pedogenesis. During periods of lowered sea level, the continental area was larger, air and sea temperatures were lower, and atmospheric circulation and wind speeds were stronger. These cold, arid phases were times of dune building and evaporative playa deposition in the inland desert regions. At other times, periods of moister climate and increased stream discharge created extensive lacustrine environments, and alluvial and colluvial fan aprons were constructed intermittently around the highland regions. Changing climate is also reflected in the secondary cements, soils and palaeosols of these regions.

ALLUVIAL FANS AND PIEDMONT SLOPE DEPOSITS

R.A. Callen, M.J. Sheard, M.C. Benbow and A.P. Belperio

Coalescing alluvial fan aprons flanking the uplifted Flinders and Mount Lofty Ranges, and the residual highlands of the Gawler, Stuart, Davenport, Musgrave, Mann, Tomkinson and Everard Ranges (Fig. 11.1), formed thick depositional sequences during the Quaternary (Callen, 1977, 1981; Wasson, 1979). These consist of coarse gravel and sand with minor silt and clay near the ranges, while distal equivalents are dominated by sand, silt and clay. The alluvial fan aprons merge downslope into more gentle piedmont slopes or plains.

The proximal alluvial fans exhibit various lithotypes and structures which indicate several modes of clastic transport and deposition. Conventional fluvial transport has yielded 'shoe-string gravels' and cobble conglomerates of channel deposits, lag deposits of boulders, overbank flood deposits of sand, silt and clay, and various minor terrace units. However, transport of essentially thixotropic mixtures of clasts and mud has resulted in debris flow, or slurry deposits. These deposits are characterised by massive structure with angular to rounded cobbles and pebbles supported by a muddy matrix. D.L.G. Williams (1982a) described repeated debris flows, and isolated debris flow units within fan and valley-fill sequences. In a semiarid environment, localised, intense rainfall events of short duration on slopes mantled with unstable, mechanically weathered debris, are typical causes of such flows.

Correlation between similar units along the length of the Flinders and Mount Lofty Ranges and elsewhere around South Australia has been problematic due to lateral facies variations, inadequate exposure and limited chronological control. Consequently, a large number of unit names, both formal and informal, have appeared in the literature. Table 11.2 sets out the major stratigraphic units which are of alluvial fan-piedmont origin. Only general correlations are indicated, together with the stratigraphic order for each region.

Mount Lofty and southern Flinders Ranges

The colluvial-alluvial stratigraphy is best studied in regions flanking the ranges that coincide with subsiding marginal basins, that is the Murray Basin, St Vincent Basin (including Adelaide Plains Sub-basin, and Willunga and Noarlunga Embayments) and Pirie Basin. Within the fault-bounded St Vincent Basin, alluvial fan sediments (and associated facies) include Hindmarsh Clay, Keswick Clay and Pooraka Formation (Table 11.2). In the Pirie Basin, Pleistocene alluvial fans comprise Hindmarsh Clay, Telford Gravel and Pooraka Formation. In the Noarlunga Embayment, the Quaternary is dominated by a punctuated sequence of alluvial, fluvial and related lacustrine and aeolian sediments, the stratigraphic nomenclature of which is undergoing revision (Phillips and Milnes, 1988).

Included within the ranges are several small intermontane basins that were formed by faulting during the Tertiary to early Quaternary. These basins also contain alluvial fan and piedmont slope deposits similar to those flanking the ranges. The largest of these are the Myponga, Hindmarsh Tiers, Barossa, Willochra and Walloway Basins. In addition, there are numerous valleys containing a variety of colluvial and alluvial sedimentary facies.

Hindmarsh Clay

The Hindmarsh Clay (Firman, 1966b) is a consolidated, mottled clay and sandy clay with sand and gravel lenses, resting in places on Pliocene sediments. Firman assigned a Pleistocene age based on the recorded occurrence of *Diprotodon* and *Macropus* remains, and on the presence of basal fossiliferous sand at Lockleys which was related by Ludbrook (1963b) to the Early Pleistocene Calabrian Stage. Deposition occurred in alluvial fan, valley flat and playa lake environments. Aeolian sand, gypseous and pelletal clay dunes, and calcareous palaeosols are also present. Good exposures of these deposits occur in the low, red cliffs at Port Augusta, Redcliff, Redbanks, Hallett Cove and Nepean Bay. The sequence exceeds



Top: Alluvial fans shed off the western flank of the Mount Lofty Ranges near Aldinga Bay. Photo 042535.

Bottom: Westerly view of Hindmarsh Clay equivalent alluvial sediments at Redbanks, Kangaroo Island. Photo 042536.

Table 11.12 Major Quaternary alluvial fan and piedmont slope units in South Australia.

Eyre Peninsula	St Vincent Basin	Flinders Ranges	Murray Basin	Southern Lake Eyre Basin	Stuart Range	Dalhousie region
Callabonna Clay ¹	Callabonna Clay ¹	Callabonna Clay ¹	Callabonna Clay ¹	Callabonna Clay ¹	Callabonna Clay ¹	Callabonna Clay ¹
Pooraka Formation ²	Pooraka Formation ²	Pooraka Formation ²	Pooraka Formation ²	—	Benitos Clay ³	—
—	—	—	—	—	Oolgelima Gravel ³	Woodgate Gravel ⁴
Coomunga Formation ⁵	Keswick Clay ⁶	Telford Gravel ⁵	Telford Gravel ⁵	—	—	—
—	—	—	—	—	Giddinna Formation ³	—
—	Ngaltinga Clay ⁷	—	—	—	—	—
Gibbon Beds ⁸	Hindmarsh Clay ^{2, 9, 10}	Avondale Clay ⁵	Blanchetown Clay ⁵	Willawortina Formation ¹¹	Russo Beds ¹²	Alinerta Gravel ⁴

1. Firman (1970); 2. Firman (1966b); 3. Benbow (1983); 4. Krieg (1985); 5. Firman (1967a); 6. Sheard and Bowman (1987a); 7. Ward (1966); 8. Parker (1983b); 9. Lindsay (1969); 10. Sheard and Bowman (1987b); 11. Callen (1977); 12. Barker *et al.* (1979).

70 m in thickness in the Adelaide Plains Sub-basin, where its upper part is predominantly clay and exhibits relict soil profiles similar to contemporary red-brown earths (Sheard and Bowman, 1987b). A discontinuous calcrete mantle with complex internal structure occurs on the Hindmarsh Clay. East of the Para Fault, Hindmarsh Clay is overlain by Keswick Clay or Pooraka Formation.

Keswick Clay

The Keswick Clay, consisting of smectite-rich grey-green clay, often with red or yellow mottling, represents a mid-Pleistocene cycle of terrestrial alluvial fan and fluvial overbank sedimentation (Sheard and Bowman, 1987a). The clay contains rare sand lenses and occasional basal colluvium, and is extensively developed in the Golden Grove Embayment and on the Para Fault Block, but is generally <6 m thick.

Pooraka Formation

The Pooraka Formation consists of silty clay, sand and carbonate earth with occasional gravel lenses. The unit is unconsolidated and represents a more time-specific Late Pleistocene (~35 000 years BP, Williams, 1973b; 30 000–20 000 years BP, Belperio and Bateman, 1986) resurgence of alluvial and fluvial sedimentation. In some parts of the Flinders Ranges, this unit may be up to 50 000 years old (Williams, D.L.G., 1982b).

In northern Spencer Gulf, the continental sequences overlie an irregular bedrock topography and total thickness is highly variable: 40 m at Port Augusta, 150 m at Port Pirie and 450 m near Redcliff. Within the St Vincent Basin, the pre-Quaternary surface is more regular, and the continental sequences generally thin in a seaward direction, where alluvial fan sediments merge imperceptibly with fluvial and overbank alluvial deposits.

Northern Flinders Ranges and adjacent basins

Fans began building around the northern Flinders Ranges and Olary Ranges during the Late Miocene and Pliocene. To the east and west of the northern Flinders, fine sandy mud and coarse clastics of debris flow and stream flood origins are referred to the Willawortina Formation (Callen and Tedford, 1976; Callen, 1977) and Avondale Clay (Firman, 1967a), respectively.

Willawortina Formation and equivalents

The Willawortina Formation southeast of Lake Eyre comprises two facies, one fine and one coarse, along the edge of the Flinders Ranges. The finer sequence of sandy mud and silty dolomite is present in Woollana 1, where nine upward-fining cycles are recognisable. These sediments are of partly lacustrine and partly mudflow and flood-plain origin, and may be equivalent to the Avondale Clay. The top of the finer unit can be observed in creek exposures along the margins of the northern Flinders Ranges, where gravely stream-channel sediments are interbedded with

clay. Thick calcareous palaeosols are present. Extensive outcrop of green clay in the CALLABONNA map area is also part of this unit, representing distal fan, mudflow and playa-lacustrine complexes.

The coarser facies is exemplified by WC 2 and other drillholes around the Beverley uranium deposit, in which there are coarse, framework-supported gravels of braided fan origin and matrix-supported clayey, poorly sorted gravels interpreted as debris flows. The Poontana Fault Complex is a graben-like structure which contains a thick sequence of this alluvial detritus. The Willawortina Formation has also been recognised along the northern margin of the Olary Ranges.

Drilling by ANU (Herczeg and Chapman, 1991; Magee *et al.*, 1988) and Comalco Exploration beneath Lake Frome has demonstrated intertonguing of fine and coarse facies. The western boundary of the lake fluctuated widely, at times being within a few kilometres of the Flinders Ranges, and at other times well to the east of the present Lake Frome shoreline. The shoreline fluctuations and cyclicity of the finer lacustrine facies were probably largely climatically controlled. Palaeomagnetic dating on sediments and palaeosols from the Lake Frome area and correlation with the northern Murray Basin indicate that much of the Willawortina Formation is older than 780 000 years (reversed polarity).

The Avondale Clay, defined on the western side of the Flinders Ranges, may correlate with similar mudflow deposits on the eastern side of the ranges. Just west of the Ediacara Range, in drillholes E50 and T1, Callen and Tedford (1976) equated ~100 m of gravel with the Willawortina Formation. A finer grained, probably equivalent unit is found in Old Motpena 1 on the eastern edge of Lake Torrens.

Around the northern margins of the Flinders and Willouran Ranges, there are extensive fan and colluvial deposits which include equivalents of the Willawortina Formation and the coarser Telford Gravel. These are frequently preserved in dissected 'high plains' and are generally cemented with carbonate or gypsum, derived from saline sulphate and/or carbonate groundwaters percolating through the fans. Towards Port Augusta to the south, the Willawortina Formation equivalent merges with the Hindmarsh Clay.

Pooraka Formation

Younger fans on both sides of the Flinders Ranges examined by Williams (1973b), Callen (1982) and D.L.G. Williams (1982a,b) are referred to the Pooraka Formation. The Eurinilla Formation (Callen and Tedford, 1976) in the southeastern Lake Eyre Basin is probably equivalent. Younger fans along the western margin of the Murray Basin produced Late Pleistocene radiocarbon dates similar to those at Hookina Creek. Alluviation took place in valleys cut into older debris flows beginning ~35 000 years BP. Lacustrine deposits with *Chara* are also present in the Burra Creek region and in Pekina Creek. Their age is unknown, although Callen (1977) suggested that they may equate with the playa-lacustrine Millyera Formation of the Lake Eyre Basin (see below). A thin, surficial sandy clay deposited on the Pooraka Formation has been termed Callabonna Clay (Firman, 1970).

Stuart Range

Russo Beds

The subdued, 170 km long escarpment of the Stuart Range is capped by silcrete formed on weathered Cretaceous sediments during the Tertiary when it was an important palaeodrainage divide between the Eucla and Lake Eyre Basins (Fig. 11.1). Much of the range, and the gently undulating plains to the west and southwest, are overlain by a complex mantle up to 12 m thick of undated late Cainozoic sediments and weathering products. Barker *et al.* (1979) applied the name Russo Beds to most of this post-Cretaceous mantle in the region of the Coober Pedy opal fields. Here, bulldozer cuts provide excellent sections where stratigraphic relationships have been studied (e.g. Benbow, 1983).

The lower complex unit of the Russo Beds consists of colluvial sediments, silcrete, gypcrete, gypsum powder and satinspar veins. The matrix is a red-brown gypseous silt which infills termite burrows in the top of the bleached Bulldog Shale. Locally overlying this lower unit are gypseous and silty conglomeratic sediments which may be a fluvial facies.

Giddinna Formation

The basal colluvial sediments are mostly overlain by brick-red to red-brown, clay-rich colluvial and alluvial outwash and flood-plain sediments, named the Giddinna Formation. It is unclear whether this formation represents single or multiple depositional episodes. The absence of the Giddinna Formation and underlying sediments east of the escarpment is likely to be the result of major and prolonged erosion as the escarpment has retreated westwards. Apart from adjacent to the Stuart Range, this region is underlain by soft, relatively unweathered Cretaceous Bulldog Shale. It is possible that, in this contrasting environment, conditions were unsuitable for formation of part of the regolith that is preserved on the range.

Developed within the Giddinna Formation is the Coober Pedy Palaeosol of inferred Late Pleistocene age (Benbow, 1983), a clay-rich hardpan characterised by its red-brown colour, platy or biscuity habit, the development of vesicular texture and black manganese coatings. Extensive but variably developed calcrete is also superimposed within the formation, including sheet, nodular and earthy forms, and chalcedony veins.

Oolgelima Gravel

Alluvial, talus, channel and flood-plain sediments up to 5 m thick, including the Oolgelima Gravel, are widespread east of and flanking the Stuart Range. These sediments are red-brown to brown, matrix (clay, silt) to framework supported, and are largely weakly consolidated. Gravel and coarser clasts were derived from weathered Bulldog Shale and Tertiary silcrete. A locally developed hardpan may be equivalent in part to the Coober Pedy Palaeosol.

Benitos Clay

Overlying these sediments, and extensive over the Stuart Range, is a thin (~1 m) finer grained blanket of red-brown to brown silty clay to clayey silt named the Benitos Clay. This unit is locally gravelly and is mantled by silcrete gibber or ironstone gravel. It is alluvial and possibly aeolian in origin, and both syn- and post-depositional pedogenic processes are superimposed. Gilgai, outlined by silcrete gibber, are prominent and form a distinctive patterned terrain, in contrast to the smooth indurated surface of the Giddinna Formation. Within the lower part of the Benitos Clay there is locally a vertically structured layer of white, very fine gypsum. In the upper part of the formation there are isolated hardpans which are locally weakly cemented by calcium carbonate and the top of the formation is reddened.

During the Holocene, grey to brown unconsolidated gravelly clay, silt and sand were deposited, mostly along contemporary drainage systems. These sediments have been variously reworked and form the modern bed load.



Top: Stacked alluvial fan deposits along Hamilton Creek, 0.5 km north of Terrapinna Waterhole, northern Flinders Ranges. Photo 042537.

Centre: Small intermontane basin in the central Willouran Ranges infilled with Pleistocene and Holocene alluvial debris. Photo 042538.

Bottom: Russo Beds exposed in an opal excavation on Larkins Folly Field, Coober Pedy.. Photo 035794.

THE DESERTS — PLAYAS, DUNEFIELDS AND WATERCOURSES

R.A. Callen and M.C. Benbow

The present-day landscape of much of the desert regions of South Australia comprises dunefields, saline playas and a network of claypans and ephemeral watercourses. Rising out of these extensive plains are gentle structural domes in which Cretaceous and Tertiary rocks are locally exposed, and with which mound springs, stony deserts and tablelands are associated. The deserts can be divided broadly into the complex and diverse Great Victoria Desert in the northwest, the geomorphologically coherent northeast deserts region, and a central zone of eroding Mesozoic landforms that make up a western Lake Eyre drainage system (Fig. 11.1). The coherence of the northeast deserts region results from its relationship with the Tertiary Lake Eyre Basin and underlying Mesozoic Eromanga Basin.

Northeast deserts region

This region incorporates the Simpson, Tirari, Strzelecki and Sturt Stony Deserts, along with the large playas of Lakes Eyre, Frome and Callabonna (Krieg, Callen *et al.*, 1990). Lake Eyre is the terminal part of the vast Lake Eyre drainage basin, which extends into Queensland, New South Wales and the Northern Territory. Strzelecki Creek, a tributary of Cooper Creek, terminates in the Lake Frome-Callabonna playas.

Claypans and small playas form prominent patterns — parallel curvilinear chains of pans — which are clearly visible in the infrared spectrum of Landsat imagery. These playa patterns are controlled by Pleistocene and late Tertiary shoreline features (e.g. shorelines of the Billa Kalina Basin amongst the dunes north of Olympic Dam), the intersection of longitudinal and weak transverse dunes, and trends of former watercourses, the Coongie Lakes, and salina chains parallel to the Warburton-Kallakoopah Creek system north of Lake Eyre.

Lake Frome, Lake Eyre South and the other larger playas are located at the boundary between fans and dunefields, and their shorelines fluctuated according to past run-off and the amount of detritus carried into the fans by local drainage from the Flinders and Willouran Ranges. Lake Eyre North differs in being related to a large drainage system from the north and northeast, and its modern depositional facies are described in a separate section.

In the Channel Country of southwestern Queensland and the Coongie Lakes area of South Australia, muddy, anastomosing channels occupy broad stretches of the desert, along with the sandy meandering trunk stream courses. In the lower Cooper and

Warburton Creek system, the meander belts are entrenched into prior stream deposits, sometimes down to the top of underlying Tertiary rocks. These older fluvial and lacustrine deposits preserve a relative stratigraphy that is, as yet, poorly dated (Table 11.3). Enhanced fluvial conditions coincided with the last two interglacials (oxygen isotope stages 5 and 7), with aridity and dune building characterising the glacial periods (Nanson *et al.*, 1992).

Quaternary fluvial sequences assigned to the Tirari and Kutjitarra Formations are entrenched into Tertiary units, including Etadunna Formation and Mampurdu Sand (Fig. 11.21a). A younger, more widespread fluvial-lacustrine cycle is known as the Katipiri Formation. A similar fluvial-lacustrine sequence has been described separately from the Strzelecki Desert and around the margins of Lakes Frome and Callabonna (Fig. 11.21b).

Pliocene-Early Pleistocene

This interval contains several rock units, some of which may span the Pliocene-Pleistocene boundary — the Tirari Formation in the Tirari Desert, and Willawortina Formation and Avondale Clay of fans marginal to the Flinders Ranges. The marginal fans are discussed in another section, but include flood-plain and lacustrine facies related to the playas and their attendant watercourses. Along Balcanoona Creek which runs into Lake Frome, laminated carbonates of lacustrine origin are exposed. These have been strongly altered by soil processes and occur quite close to the ranges; they indicate high lake levels coincident with Willawortina Formation deposition. From TL and palaeomagnetic dating, the age of these deposits is believed to be between 1.66 and 0.78 Ma.

Tirari Formation

The Tirari Formation (Fig. 11.21a), with a type section at Lake Palankarina, is a fluvial redbed unit of fine silt and medium sand deposited in palaeovalleys in the Tirari Desert. The stream systems had broad, shallow channels and extensive flood plains. TL dating of quartz grains from a lake west of Peachawarina 2 has yielded minimum age estimates of 400 000-290 000 years BP for the top of the unit. Tedford *et al.* (1986) suggested from vertebrate evidence that much of the unit may be of Pliocene age. Tirari Formation vertebrates belong to the Kanunka Local Fauna (Stirton *et al.*, 1961, 1968; Tedford *et al.*, 1986) and include thylacines, wombats, diprotodontids, potoroos and short-faced kangaroos. The Tirari Formation is strongly impregnated with gypsum, suggesting a high, saline watertable, and the reddening of quartz grain surfaces suggests seasonally arid conditions. These, and other sedimentary features, indicate a wetter climate and greater streamflow, partly as a consequence of greater monsoonal rainfall than at present in Queensland.

Table 11.3 Quaternary lithostratigraphy of the northeast deserts region.

Formation	Age	Maximum thickness (m)	Depositional environment	References
Simpson and Tirari Deserts				
Simpson Sand	Pleistocene-Holocene	50	Dunefield	Firman, 1970.
Katipiri Formation	Mid-Late Pleistocene	5	Fluvial channel, lacustrine shoreline	Stirton <i>et al.</i> , 1961; Wells and Callen, 1986.
Kutjitarra Formation	Pleistocene	10	Fluvial channel, playa-lacustrine	Well and Callen, 1986.
Tirari Formation	Plio-Pleistocene	12	Alluvial, fluvial and aeolian	Stirton <i>et al.</i> , 1961; Firman, 1970.
Strzelecki Desert, Lake Frome, Lake Callabonna				
Coonabine Formation	Pleistocene-Holocene	50	Dunefield	Callen and Tedford, 1976; Callen <i>et al.</i> , 1983.
Eurinilla Formation	Late Pleistocene	10	Alluvial	Callen and Tedford, 1976; Callen <i>et al.</i> , 1983.
Katipiri Formation equivalent	Mid-Late Pleistocene	45	Fluvial channel, lacustrine shoreline	
Coomb Spring Formation	Mid-Late Pleistocene	8	Lacustrine shoreline	Callen, 1984.
Millyera Formation	Mid-Late Pleistocene	9	Playa-lacustrine	Callen and Tedford, 1976; Callen 1984.
Willawortina Formation	Plio-Pleistocene	100	Alluvial and lacustrine	Callen and Tedford, 1976; Callen 1977.

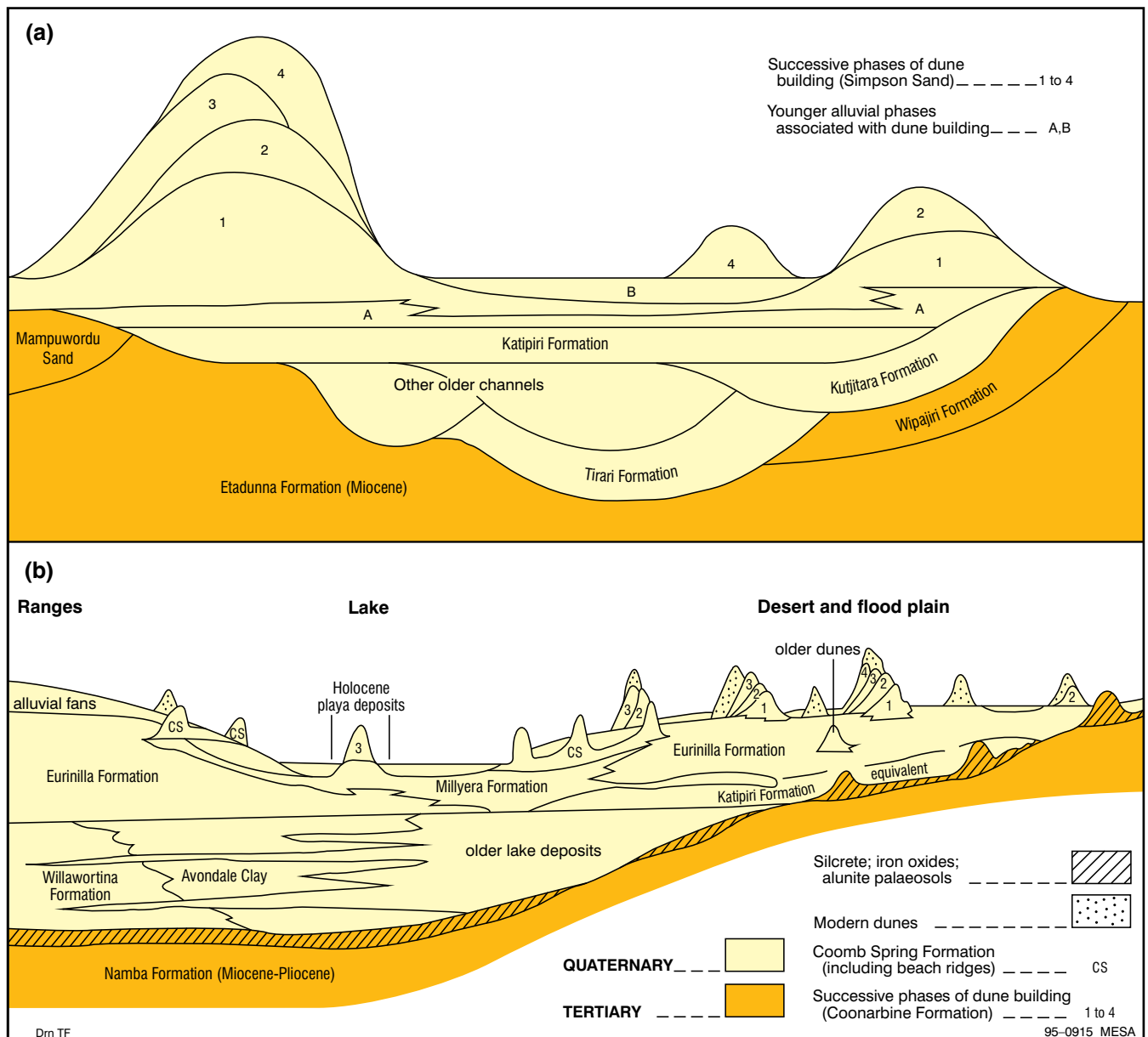


Fig. 11.21 Rock relation diagrams for (a) the Simpson and Tirari Deserts, and (b) the Strzelecki Desert.

Middle Pleistocene

Sediments from the extreme northeast of the State provide evidence of fluvial sedimentation during interglacial stage 7, and similar deposits occur extensively through the Channel Country of southwestern Queensland (Nanson *et al.*, 1988). TL dates of ~240 000 to 167 000 years BP at Moomba (Gardner *et al.*, 1987), and an electron spin resonance date of ~220 000 years BP interpreted by Wopfner and Grun (1984) as a maximum age, on lacustrine limestone at Cordillo Dome provide limited chronological data. Some of the channel deposits between the Tirari Formation and Katipiri Formation in the Tirari Desert may also be of this age.

Kutjitarra Formation

The Kutjitarra Formation in the lower Cooper Creek comprises fluvial deposits associated with the penultimate glacial period (Tedford and Wells, 1990). It is difficult to differentiate from underlying and overlying units and, although poorly dated, contains *Diprotodon* species not found in the overlying Katipiri Formation. The Kutjitarra and Tirari Formations indicate a distributary system draining to the north, in agreement with a depocentre to the north of Lake Eyre. Geomorphological evidence of excursions

of the Kallakoopah Creek into the Simpson Desert, as revealed by the claypan belts, may be related to this northward gradient. Subsidence in this region is additionally indicated by the presence of Tertiary pedogenic silcrete at over 100 m depth in Poolowanna water bore (Callen, 1983b).

Around Lake Eyre North and South, alternating red and green gypsiferous clay and fine white to yellow sand, overlain by beach deposits containing vertebrates, may be equivalent to the Kutjitarra Formation. Sediments similar to the Tirari and Kutjitarra Formations have not been recorded in outcrop in the Strzelecki Desert, but Yandruwantha Sand, described by Callen (1992) from drillholes in the Moomba region, is a probable correlative of undifferentiated Pleistocene age.

Late Pleistocene

Katipiri Formation

The Katipiri Formation comprises similar channel deposits of meandering streams with palaeocourses closely associated with their present-day descendants (Tedford and Wells, 1990). The unit generally consists of white to yellow and orange fine-grained

sand, becoming coarser at the base, with local clasts, and is clearly channel confined. Concordant TL and uranium series dates of 132 000-108 000 years BP have been obtained from quartz, gypsum and calcrete from Cooper Creek (Callen and Nanson, 1992) indicating a last interglacial age (Fig. 11.22). Equivalent deposits, affected by pervasive red iron staining, are present in the Simpson Desert, and meanders have been mapped on the surface of some of the salinas north of Lake Eyre. The Katipiri Formation contains abundant vertebrates, gastropods, bivalves and microfossils. The vertebrates include similar genera to those of the Tirari Formation, such as *Thylacoleo*, dasyurids and koalas. Deposits similar to the Katipiri Formation occur near Moomba and have probable

equivalents in the Queensland Channel Country, where Nanson *et al.* (1988) identified extensive sand deposits tentatively dated at 120 000-100 000 years BP.

Extensive beach deposits of well-sorted shelly sand and gravel with clay clasts have been mapped on CURDIMURKA around Lake Eyre South at the -2, 9-10 and 14-16 m contour levels. These are in part coeval with the Katipiri Formation. Radiocarbon dating on shell, TL dating on quartz sand and amino acid racemisation dating of bivalves and gastropods indicate lake-full stages at 150 000, 110 000, 60 000-45 000, and ~25 000 years BP. Good correlation is shown with lacustrine shoreline deposits around Lake Frome.

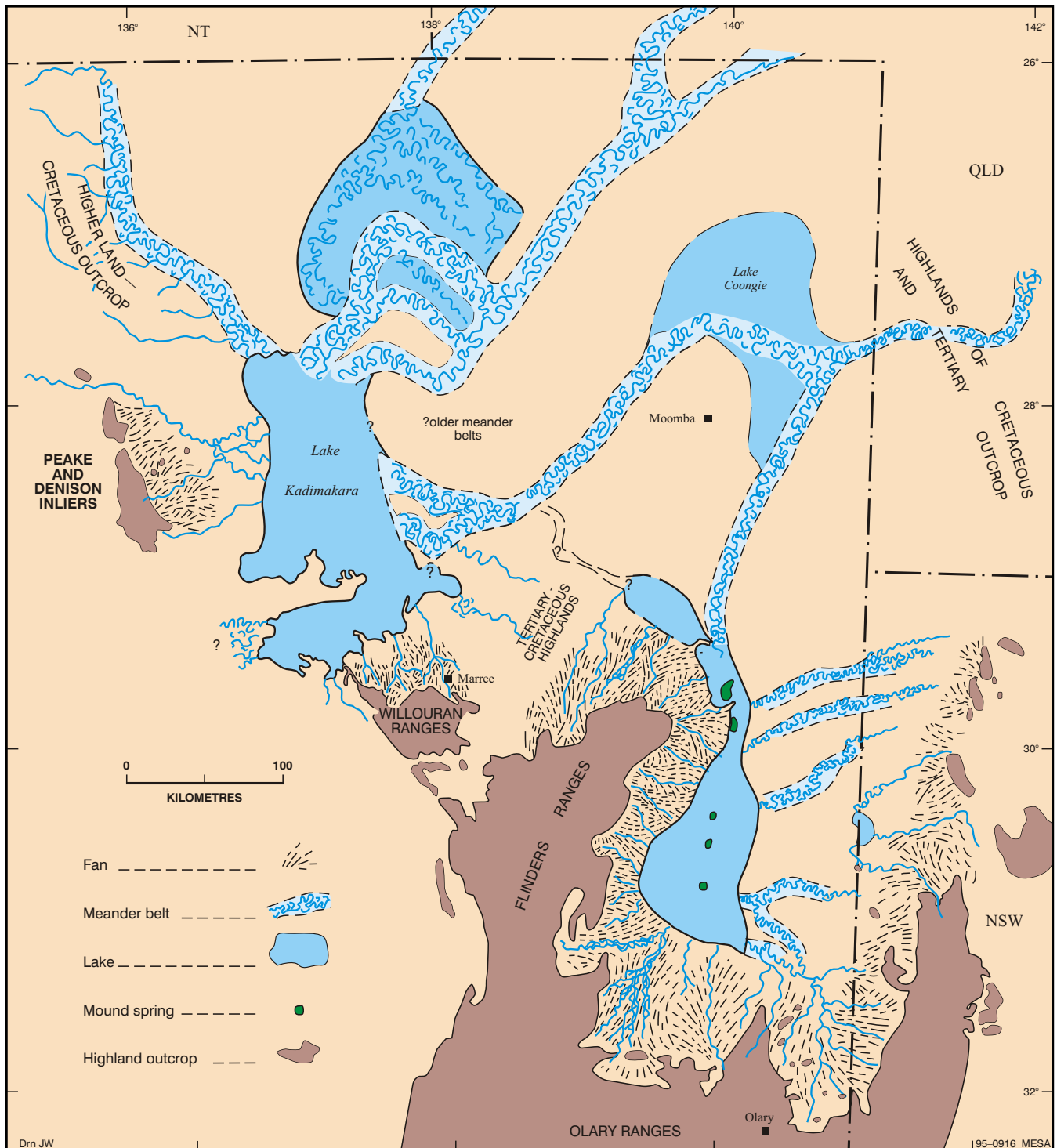


Fig. 11.22 Palaeogeography of the Lake Eyre region in the last interglacial.

Eurinilla Formation

A more complete succession of Late Pleistocene facies (alluvial, lacustrine, lacustrine shoreline) is preserved to the south in the Strzelecki Desert and near Lake Frome (Fig. 11.23). The Eurinilla Formation (Callen and Tedford, 1976), in part equivalent to the Katipiri Formation, has a type section at Lake Moko to the east of Lake Frome. It consists of bright red-brown to yellow-brown sand and gravel of intermittent streams and their overbank deposits. The unit intertongues with playa-lacustrine deposits (Millyera Formation) and was deposited partly in palaeovalleys formed within the Willawortina Formation. West of Strzelecki Creek, coarse, poorly sorted brown sand with angular grains forms a blanket deposit up to 50 m thick. Often cemented with gypcrete, this sand is regarded as equivalent in part to the Katipiri and Eurinilla Formations. These deposits indicate that, in this region, fluvial activity was extensive and not restricted to palaeovalleys.

Millyera Formation

The Millyera Formation (Callen and Tedford, 1976) comprises greenish clay, thin algal limestone (mainly charophytes) and fine-grained white to greenish sand with fish remains and gastropods. These are lacustrine and flood-plain deposits at their type section at Lake Millyera (Callen, 1984). The white sands intertonguing with the unit on the east resemble the Katipiri Formation but the fossils suggest fluctuating salinity from fresh to very salty. The famous Lake Callabonna vertebrate deposits occur in this unit, and contain *Diprotodon*, *Genyornis* and other fossils. Bowler and Magee (1988), using uranium series and TL dates, suggested major lacustrine phases at ~145 000 and 60-30 000 years BP.

Coomb Spring Formation

Around Lake Frome, beach deposits similar to those around Lake Eyre South are named the Coomb Spring Formation, with a type section in Eurinilla Creek. They consist of white to yellow, often coarse-grained, well-sorted sand deposited at levels of 8-9 and 16 m AHD. These elevations correspond to the higher levels at Lake Eyre. Present topographic contours indicate that the two lakes may have been linked around the northern edge of the Cooryanna Dome during times of higher lake level. Dates on the beach ridges indicate accumulation mainly before 40 000 years BP, and possibly as far back as 140 000 years BP (Herczeg and Chapman, 1991). These deposits intertongue with the Millyera Formation. Celestite pods associated with these sediments formed as a result of early Holocene groundwater movement.

Aeolian sediments

Coonarbine Formation and Simpson Sand

Aeolian deposits of Late Pleistocene to Holocene age are represented by the Coonarbine Formation in the Strzelecki Desert and Simpson Sand in the Simpson Desert (Table 11.3). The Coonarbine Formation includes older dune deposits in addition to the last glacial age dunes, whereas the Simpson Sand is largely of last glacial to Holocene age. Radiocarbon dating of calcareous palaeosols and algal material has provided only minimum ages (Callen *et al.*, 1983; Wasson, 1983a,b, 1984; Wasson *et al.*, 1988; Bowler and Wasson, 1984). Uranium series dates (Nanson *et al.*, 1988) suggest that soil carbonate radiocarbon dates could be several orders of magnitude too young.

The aeolian sands are formed into linear dunes typically 15 m high, with subdued transverse features in some areas, and various 'tuning fork' junctions. They trend roughly north-south, oblique to the dominant transport direction, and have migrated laterally to the east whilst extending their linear form northwards (Rubin, 1990). The sand is disconformable on the Late Pleistocene alluvial and lacustrine sediments and consists of red-coated, frosted quartz grains, yellow to orange grains, clay pellets and gypsum flakes; aeolian cross-bedding is prominent. The older phases of the linear

dunes are marked by calcareous palaeosols, and are often paler than the overlying younger phases, as a result of the greater proportion of clay. Claypans, aligned obliquely to the linear dunes, also developed as a result of Late Pleistocene aeolian erosion (Stevens, 1991).

Large transverse features also form along the downwind margins of playas and pans. The southeastern shore of Lake Frome, for example, has gypsum dunes up to 20 m high along the downwind shore. The well known 'Cobbler' dune along the north shore of Lake Callabonna is another such feature. On the bed of Lake Frome, gypsum dunes have built upon clay dunes indicating that after the lake floor clays were deflated, the groundwater gypsum horizon was exposed and deflated. Along the north shore of Lake Eyre South, huge transverse dunes of quartz sand have been derived from Eocene sands exposed in the lake bed.

A major phase of Coonarbine Formation deposition occurred during the last glacial maximum 20 000-15 000 years BP (Figs 11.23, 11.24), when tree vegetation was minimal (Singh and Luly, 1991). During these more arid times, the former lakes dried up, large vertebrates became extinct, and dunefields extended throughout the State's interior. From 17 000 to 14 500 years BP, *Callitris*, *Casuarina*, *Acacia*, *Dodonaea* and myrtaceous shrubs were common around Lake Frome, with well-developed woodlands. There was increased winter rainfall, but summer rainfall was about the same as the present and temperatures were less extreme. Between 14 500 and 13 000 years BP there was a sudden decline in *Callitris*, and a more gradual decline in chenopods and *Dodonaea*, whereas ephemeral vegetation increased, suggesting increasing aridity (Singh and Luly, 1991).

From 13 000 years BP, grasslands developed and chenopod shrubland extended as a result of increasing summer rain due to summer monsoon incursions. During another wetter interval from 6 000 to 4 000 years BP, *Callitris* again increased. Since this time, *Callitris* has become restricted to the southeastern corner of the Strzelecki Desert and isolated remnants along defunct watercourses. Chenopod shrublands and ephemerals have increased and *Casuarina* has declined. During the last few thousand years, playa environments were re-established and dunes re-activated as the incidence of summer monsoonal rainfall declined. Lakes Eyre and Frome rarely fill and the present arid climate marks the final demise of permanent brackish lakes of the northeast deserts.

During the last 10 000 years, grazing marsupials increased at the expense of browsing forms, and a reduction in body size took place, in keeping with warmer conditions (Archer and Clayton, 1984; Tedford *et al.*, 1986; Woodburne *et al.*, 1985). Today, only small crepuscular and burrowing marsupials are to be found in the dunefields, together with emus and red kangaroos in the more open areas.



Eurinilla Formation overlain by Coonarbine Formation near 'Curnamona'. Photo 042539.

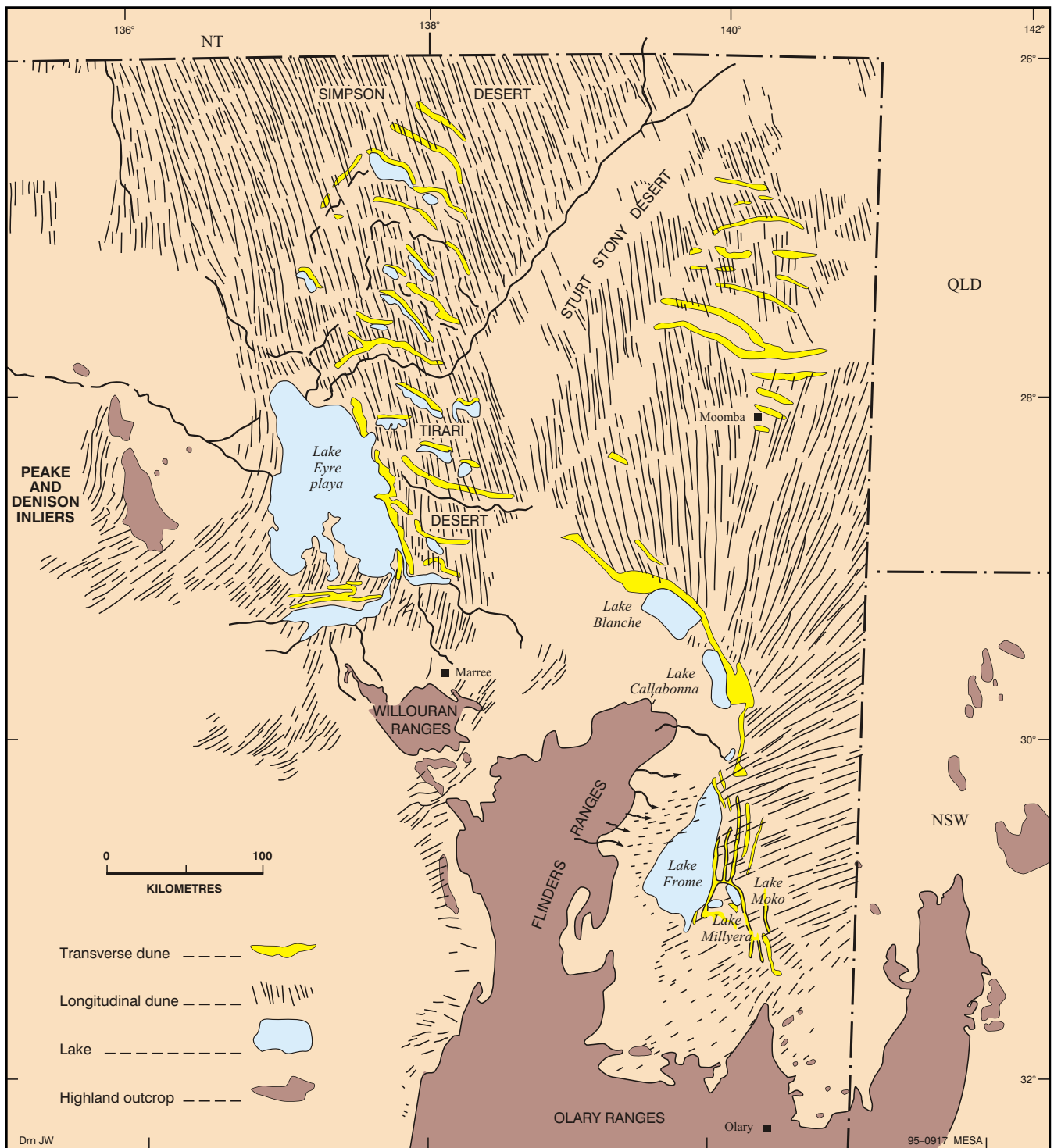


Fig. 11.23 Palaeogeography of the Lake Eyre region during the last glacial.

Holocene sediments of Lake Eyre

Lake Eyre, normally a dry salina covering 9 300 km², is the terminus of an inland drainage system that extends over more than one million square kilometres of the Australian continent. Most of the water reaching Lake Eyre comes from the northeastern side of its drainage basin in Queensland, which is situated within the summer monsoon zone with annual rainfalls of 370-500 mm. Monsoon floodwaters flow south into semiarid, very low gradient terrain, known as the Channel Country. There they spread out into innumerable channels and distributaries, and much of the water is lost to evaporation. Any water passing through the Channel Country and Coongie Lakes enters the arid deserts around Lake Eyre, where it fills distributaries, interdune corridors and small

playas, and soaks into sand dunes. Water which passes through the dunefields finally enters Lake Eyre. Minor flooding of the salina usually requires less than a year to dry up and redeposit a salt crust.

Less frequently, monsoonal rain depressions move further south, bringing appreciable falls of rain to the Channel Country and areas around the lake. At the same time, the heavy falls send down large volumes of floodwater through the Channel Country, the deserts and into Lake Eyre. These may cover the whole lake bed to depths up to 6 m, as in 1974. Low-level wave-built shingle terraces dating back to ~3 000 years BP suggest major fillings up to 8.5 m deep in the recent past (Dulhunty, 1975).

A gentle slope of 24 mm/km to the south within Lake Eyre North has resulted in development of three different environments of arid ephemeral lacustrine sedimentation within its shores (Fig. 11.25).

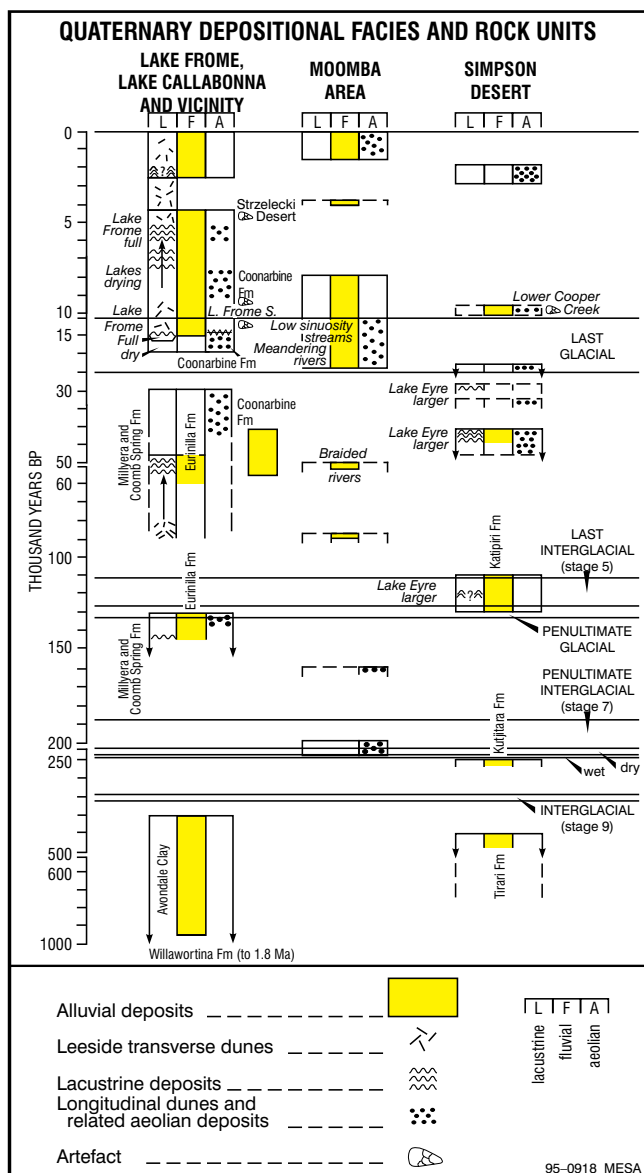


Fig. 11.24 Schematic illustration of episodic fluvial, lacustrine and aeolian sedimentation in the northeast deserts region.

The northern area of the lake is an arid playa floored by at least 3 m of saline Holocene clay with silt and some very fine-grained sand, but no crystalline gypsum. In places there are surface layers of black organic mud. Powdery salt forms on the lake bed between floods, but there are no hard salt crusts. To the south, mainly within Madigan Gulf and Jack Boot and Belt Bays, an arid terminal salina exists. A surficial salt crust overlies ~4 m of gypseous Holocene and Late Pleistocene sediments, consisting of seed gypsum crystals, fine silt, clay and buried halite (Johns and Ludbrook, 1963), with interbeds of black gypseous mud. The deposit as a whole represents a mixture of introduced sediment and indigenous evaporites.

The northern playa and southern salina are separated by a narrow east-west zone of fine clay and organic mud ('slush zone') extending across the lake, described as an arid saline flocculation environment. Salt crusts are either absent or very thin, and the fine sediments are a result of flocculation and deposition of suspended clay and colloidal organic matter within a saline-freshwater mixing zone.

The principal locus of salt precipitation in Lake Eyre is in the southern bays of the north lake (Fig. 11.25) where the crust is

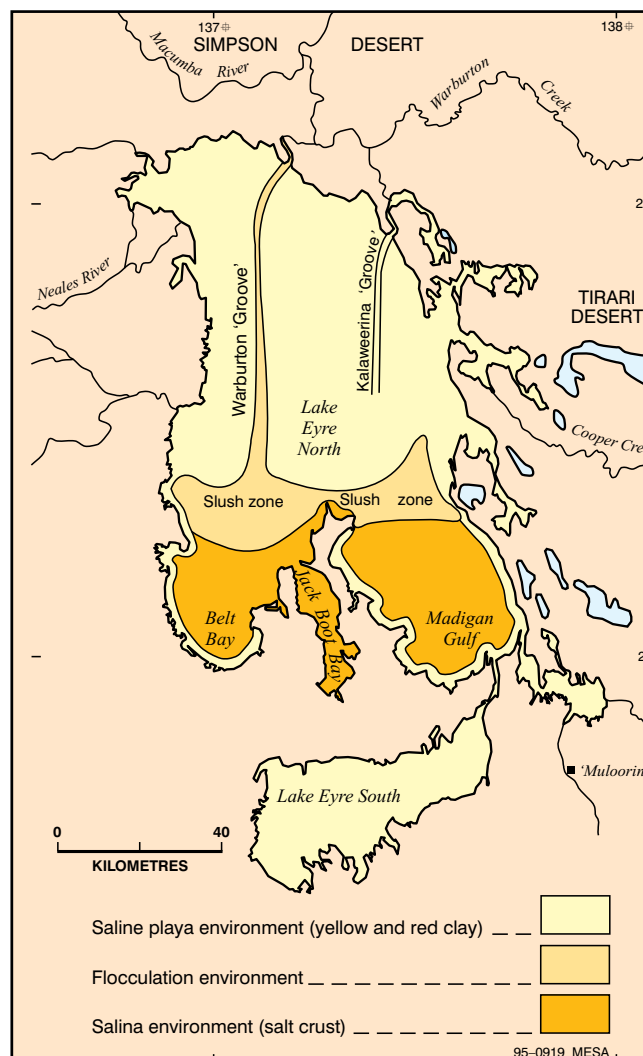


Fig. 11.25 Modern environments of Lake Eyre.

up to 0.5 m thick. The salt crust dissolves completely at times of lake filling, and is redeposited as the waters evaporate. Little salt occurs in Lake Eyre South as its bed is higher than in the north lake (Bonython, 1956; Dulhunty, 1974, 1975). Estimates of the total amount of salt in Lake Eyre vary from ~400 to 700 Mt. The brines consist principally of sodium chloride (90-95%), magnesium sulphate (5-7%), magnesium chloride (up to 4%) and calcium sulphate (up to 2.5%). Potassium content (0.11-0.34%) is lower than in sea water.

Coongie Lakes and the Cooper-Strzelecki system

Cooper Creek enters South Australia across the Innamincka Dome, cutting through silicified Tertiary and Cretaceous rocks. Where it debouches onto the vast plain of the Moomba area, the Cooper has built a very large but extremely low-angle fan (Callen and Bradford, 1992). Subtle changes in topography reflect major changes in the course of distributaries and channel avulsion. A major division into northerly and southerly directed distributaries occurs at 'Innamincka'. Floodwaters spreading to the north cover vast areas of the surrounding plains. Coongie Lakes is one such area between the fan and the Gascon Dome to the west. Some of these floodwaters eventually reach Lake Eyre. Southerly directed floodwaters follow the course of Strzelecki Creek, eventually reaching Lake Blanche where another extensive fan has developed during the Holocene. The pattern of drainage and flow of floodwaters is also strongly influenced by north-trending dunes and east-trending transverse dunes and lunettes.



1



2



3



4

Gibber and Sturt Stony Desert

Gibber spreads are extensive areas of angular to rounded cobbles forming an armoured surface in desert areas. They are residual weathering horizons, usually with gypsum and carbonate cements beneath the surface armour. Such areas are common around Marree and in the Sturt Stony Desert and, to a lesser extent, on the tablelands. They may originate from *in situ* breakdown of silicified Tertiary sediments or by alluvial and colluvial transport. The alluvially derived gibber often contains clasts from basement areas, and gibber formed on Mesozoic sediments commonly includes a concentration of well-rounded and polished clasts, some of basement origin, derived by reworking of these sediments. Gibber also forms from alluvial materials without a Tertiary silcrete source, in which case the clasts are better rounded. Gibber clasts frequently have a desert varnish on the exposed surfaces, resulting from weathering and aeolian processes, that can be used as a relative dating tool.

Gibber armour forms a largely impermeable barrier, which increases run-off and effectively prevents vegetation growth. Gibber often forms a thin layer over subcropping expansive Cretaceous clays or their alluvial derivatives and is generally cemented by gypsum. Cracks and gilgai structures are common, and may form the locus of stunted (shrub-steppe) vegetation growth. On slopes, stone-walled terraces may form as a result of soil creep combined with gilgai processes.

Sturt Stony Desert occupies a region of upwarp within the Lake Eyre Basin. Silicified and ferruginised Tertiary sediments from former outcrop have broken down and been redistributed in colluvial-alluvial deposits from which the gibber is derived.

Western Lake Eyre drainage system

This name is applied herein to the diverse region between the Simpson Desert and Great Victoria Desert (Fig. 11.1). The area is poorly defined, being roughly bounded to the east by the Davenport-Willouran Ranges divide and the Stuart Range divide in the west. The region corresponds to a prominent zone of dissected Eromanga Basin marginal sediments with associated breakaways, mesas, gibber and gypcrete regolith. Drainage is directed mostly into Lake Eyre, but also into Lakes Torrens and Gairdner. Lack of geochronological and detailed stratigraphic data hinder interpretation of the landforms and associated Quaternary sediments.

Playas within and north of the Gawler Ranges form a distinctive pattern, probably reflecting former Tertiary watercourses which drained either into the Billa Kalina Basin or westwards to the Eucla Basin. They contain only thin deposits of Quaternary alluvium described by Krinsley *et al.* (1968) and Johns (1968). Lake Acraman has well-preserved clay-gypsum lunettes on its surface, suggesting similar groundwater conditions to Lake Frome. Sulphur isotope measurements on surficial gypsum in these playas indicate a dominantly marine source from aerosols driven inland by contemporary southerly winds (Chivas *et al.*, 1991).

Much of the area is characterised by extensive, flat gibber plains and gypsiferous gravels. Alluvial deposits and eroded pediments developed on a wide range of older rock types have been variably cemented and infused by crystalline gypsum, forming caprock surfaces termed 'gypsite'. Several cycles of gypsite development, uplift, dissection and redistribution of gypsum to topographically lower surfaces have been documented on WARRINA (Wopfner and Twidale, 1967), DALHOUSIE (Krieg, 1985) and CURDIMURKA (Krieg *et al.*, 1992).

1. Southerly view over the flooded Warburton Creek and Tirari Desert, 1971. Photo 042540.
2. Northeasterly view of Douglas Creek and sand dunes, with Lake Eyre in the background. Photo 042541.
3. Coongie Lakes wetlands, northwest of 'Innaminka'. Photo 042542.
4. Quartzite gibber near Stuart Creek. Photo t015029.

Small dunefields exist in the northern part of the area, including the Pedirka Desert south of Hamilton Creek, and in the Roxby Downs-Stuart Creek area. Indurated dune cores with carbonate palaeosols have been recorded on northeastern BILLA KALINA and CURDIMURKA. These features suggest that construction of the dunefields, as elsewhere in northern South Australia, commenced in the Late Pleistocene.

Lake Torrens, with a small catchment in comparison to its size, rarely fills, hence its sediments are essentially distal alluvial fan facies. The lake is underlain in the west and north by Cambrian and Neoproterozoic rocks with a very thin Cainozoic sedimentary cover, but the eastern portion has a thicker (300+ m) succession, including dune and lacustrine facies (Johns, 1968).

Lake Gairdner overlies Gawler Range Volcanics and has a thin sedimentary cover veneered by salt and epsomite crusts. Brines are of the Na-Cl-SO₄ type, with potassium content lower than that of sea water. From limited drilling, the gypseous and saline sediments are known to be up to 18 m thick (Johns, 1968).

Great Victoria Desert region

Of the major dunefields of Australia, the Great Victoria Desert in western South Australia is the least studied and its exact extent has never been clearly defined (Fig. 11.1). To the south, it borders the Nullarbor Plain, although the longitudinal dunes west of the Gawler Ranges also form a continuum.

This desert forms the southern part of the large anticlockwise whorl of dunes that covers much of central Australia (King, 1960; Jennings, 1968a). The dominant dunes are narrow-crested linear types which have an average spacing of <300 m (Wasson *et al.*, 1988). Overall direction of the dune crests in the Great Victoria Desert is east-west but there are local variations. One example of this occurs along the southwest margin of the Late Eocene Ooldea Range which deflects the prevailing wind, giving rise to dunes with contrasting orientation along its flanks. The migration of linear dunes over the northern Nullarbor Plain suggests a dominant transport direction from the north.

The substrate of the Great Victoria Desert varies from Precambrian crystalline basement to duricrusted cover rocks as young as Tertiary. Around the eastern margin, part of the dunefield directly overlies Middle Miocene Nullarbor Limestone and Garford Formation. There is another broad region around the margin of the Eucla Basin where the dunes overlie Middle to Late Eocene marine to non-marine sand, which may have been a major source for dune construction.

Delisser Formation

Weakly indurated, brick-red sand of the Delisser Formation (Benbow, 1986a) is exposed around some playa lakes such as those to the west of Lake Anthony. This appears to be the oldest widespread Quaternary aeolian sands, although dune forms have not been recognised. At Lake Anthony there is equivalent, and possibly older, bioturbated yellow to orange-brown aeolian sand.

Wintrena Formation

The Wintrena Formation of the Great Victoria Desert dunefield (Benbow, 1986a) is a widespread sand unit exposed around the margins of playa lakes near the Ooldea Range. The sand varies in colour but is commonly red-brown and orange-brown, and is locally up to 20 m thick. It is capped by a major, variable carbonate palaeosol which is also very widespread and includes mottled, indurated, massive and nodular calcrete up to 4 m thick. The palaeosol can be clearly traced from swales up into the cores of dunes. Palaeosol development appears to be weakest on the east side of the dunes over the Ooldea Range where it includes a complex of intermeshing rhizomorphs. The similarity of this carbonate palaeosol with those that cap the coastal Bridgewater Formation suggests that these dunes probably formed before 125 000 years BP. There were several subsequent major episodes of dune construction. Sand was remobilised to form parabolic,

crescentic and network dunes in discrete areas within the desert, including a notable discontinuous zone along the southwestern margin of the Ooldea and Gawler Ranges. Formation of these dunes was at the expense of longitudinal dunes which were locally completely destroyed. Weakly indurated sand 6 m or more thick overlies the carbonate palaeosol of the longitudinal dunes and may have been deposited at the same time.

The area of parabolic dunes adjacent to the Gawler Ranges was named the Kwaterski Dune Field by Bourne *et al.* (1974). Several episodes of dune construction are evident in exposures around the margins of associated playas. The oldest dunes are weakly to moderately indurated and locally contain red-brown carbonate palaeosols. The youngest overlying sands are mobile and currently active. Bourne *et al.* (1974) recorded radiocarbon ages of 27 000-10 300 and 22 000-16 000 years BP for secondary carbonate accumulations in the longitudinal and parabolic dunes, respectively, of the Kwaterski Dune Field. They suggested that there was no significant age difference between dune formation and secondary carbonate accumulation, but this is debatable.

Gypsum and quartz sand lunettes are commonly present around playa lakes within the Great Victoria Desert region, including those around Ifould Lake on the margin of the Nullarbor Plain. Minor longitudinal dune formation is known to post-date lunette development in the Lake Phillipson region, southwest of Coober Pedy.

Wiabuna Formation

The eastern Nullarbor Plain is mantled in part by an orange-brown calcareous silt to fine-grained sand. It locally thickens and is conglomeratic along the flanks of playas such as Ifould Lake. Firman (1974, 1978) named this the Wiabuna Formation, which he considered as underlying the dunefield, although there is no evidence to support this interpretation. A weakly developed carbonate palaeosol has formed locally.

Tektites

Tektites (natural glass spheroids) occur throughout much of inland Australia, but are particularly common around playa lake margins and palaeodrainage systems of the Great Victoria Desert. The Australian tektites form part of a large Southeast Asian 'strewn field' for which consistent ages of 800 000-700 000 years BP have been obtained (Lovering *et al.*, 1972). A number of astronomical theories have been advanced, with the most widely accepted being a genetic association with impact craters (McNamara, 1985). The size and close association of the Southeast Asian tektite event with the Brunhes-Matuyama palaeomagnetic reversal boundary has led some researchers to propose a causal relationship (Burns, 1989; c.f. Schneider *et al.*, 1992). Concentration of tektites within the contemporary landscape, and their enigmatic occurrence in younger Quaternary strata, is attributed to surficial geomorphic processes (Johnson, 1965; Annells, 1989).



Southern margin of the Quaternary Great Victorian Desert encroaching over the Middle Miocene Nullarbor Plain, near Ooldea. Photo 042543..

CONTINENTAL SEDIMENTS OF THE MURRAY BASIN

P.A. Rogers

Late Pliocene-Middle Pleistocene sediments of Lake Bungunnia

Tertiary sediments of the Murray Basin are disconformably overlain by Late Pliocene-Middle Pleistocene fluviolacustrine deposits comprising Chowilla Sand, Blanchetown Clay and Bungunnia Limestone.

Chowilla Sand

The fluvial Chowilla Sand (Firman, 1966b), a lensing unit up to 10 m thick of fine to medium-grained quartz sand reworked from the underlying Parilla Sand, occurs at the base of the Blanchetown Clay in some Murray River cliff sections upstream from Berri; it is equivalent to the Irymple Member (Victoria) of Lawrence (1975). Gill (1973) described the Chowilla Sand as interfingering with, and in places entirely replacing, Blanchetown Clay. Both units contain a fauna of ostracodes, bivalves, tortoise and fish remains, and marsupial bones (Gill, 1973; Marshall, 1973).

Blanchetown Clay

The Blanchetown Clay (Firman, 1965), up to 20 m thick, was deposited in a large body of freshwater known as Lake Bungunnia which covered a large part of the western Murray Basin in the Late Pliocene to Early Pleistocene (Fig. 11.26). The sediment consists of greenish grey sandy clay with thin layers of limestone and quartz sand overlying red-brown and greenish grey mottled sandy clay. The unit is commonly gypsiferous at the top, and often contains calcareous septarian nodules.

Palaeomagnetic studies of Blanchetown Clay sections at 'Chowilla' (Fig. 11.27) and Lake Tyrrell (Bowler, 1980; An *et al.*, 1986) place the upper part of the unit in the Matuyama Chron (2.47-0.78 Ma) and the lower part in the Gauss Chron (3.41-2.47 Ma). The base of the Blanchetown Clay at Lake Tyrrell in Victoria may be older, as it appears to extend into the Gilbert Chron (5.05-3.41 Ma). An *et al.* (1986) placed the Brunhes-Matuyama boundary (0.78 Ma) in the weathered zone at the top of the Blanchetown Clay. Hence, deposition of the formation commenced between ~3.5 and 2.5 Ma, and ended at ~0.7 Ma. These studies showed that Blanchetown Clay includes sediments of both Late Pliocene and Early Pleistocene age. At 'Chowilla', the Pliocene-Pleistocene boundary lies a short distance above the identified Olduvai Subchron in the middle of the unit (Fig. 11.27).

Stephenson (1986) related the filling of Lake Bungunnia to a tectonic dam that formed in the area south of Swan Reach. However, filling of the lake was also influenced by the construction of large coastal barriers in the southwestern Murray Basin in Pliocene and Early Pleistocene time (Belperio and Bluck, 1990). Climatic fluctuations (wet and dry periods) caused large shoreline migrations in the shallow lake, resulting in interfingering of lacustrine Blanchetown Clay and fluvial Chowilla Sand near the lake margin. The Pinnaroo Block, a late Cainozoic structural high, formed the southern limit of Lake Bungunnia in South Australia.

Deposition of Blanchetown Clay ended early in the Brunhes Chron when much of Lake Bungunnia was drained. This occurred during a period of low sea level when fluvial erosion and downcutting breached the coastal barriers, allowing the release of the lake waters, and cut an irregular surface in the Blanchetown Clay. This major sea-level lowstand is most likely the one represented by stage 16 (600 000 years BP) in the composite oxygen isotope record (Fig. 11.2). This agrees closely with the age of 620 000 years BP estimated for the top of the Blanchetown Clay at Lake Tyrrell (An *et al.*, 1986).

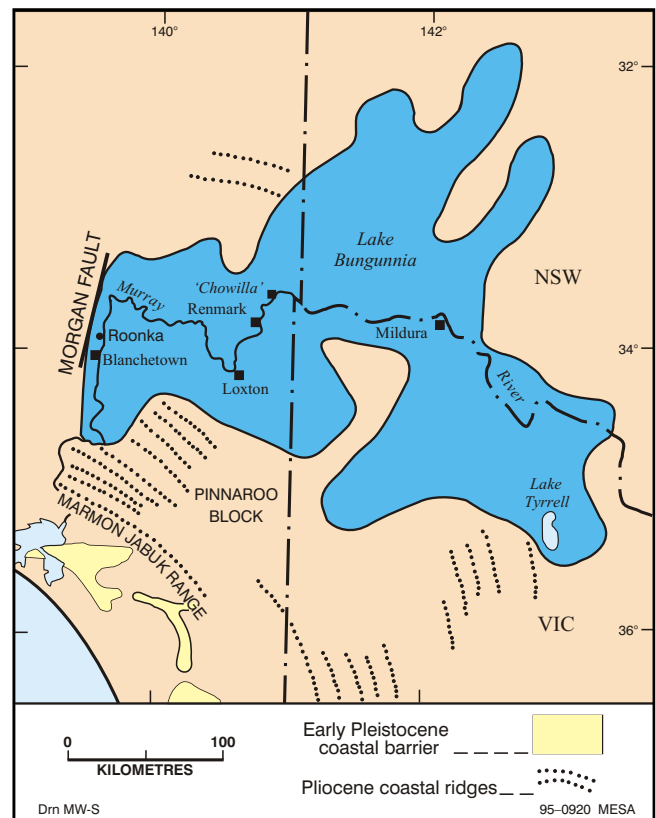


Fig. 11.26 Interpreted maximum extent of Lake Bungunnia (after Stephenson, 1986).

Bungunnia Limestone

As Lake Bungunnia drained, it fragmented into separate, smaller, clastic-starved basins in which the Bungunnia Limestone was deposited (Firman, 1965). At other places, gypsum was precipitated at the top of the Blanchetown Clay and Chowilla Sand as a result of lake drainage and increasing aridity. The Bungunnia Limestone is a thin (up to 3 m), patchy, lacustrine carbonate unit of thinly bedded micritic limestone and dolomite, and includes algal and oolitic varieties with ostracodes and gastropods.

Middle Pleistocene-Holocene playa deposits

Yamba Formation

Low-lying areas in the bed of the former Lake Bungunnia have evolved into modern saline playas, fed partly by groundwater discharge. Deposits of these playas in South Australia are termed Yamba Formation (Firman, 1966a, 1973) and consist of lacustrine gypsiferous clay and gypsum-quartz sand mixtures, associated with dunes of seed and flour gypsum. The lacustrine component of the Yamba Formation is equivalent to the Tyrrell Beds (Bowler and Teller, 1986), a 4-5 m thick sequence of interlayered clay and silt, gypsum and dolomite described from Lake Tyrrell.

Radiocarbon dating of organic carbon from the top part of the Tyrrell Beds has produced Holocene ages (Luly *et al.*, 1986), and dates of $33\,250 \pm 2\,100$ to $28\,100 \pm 1\,250$ years BP have been obtained from underlying fine-grained carbonate (Bowler and Teller, 1986). Bowler (1980) interpreted the Lake Tyrrell sequence as a record of wet and dry climatic cycles related to interglacial and glacial periods, which may extend back as far as 450 000 years.

At Noora, east of Loxton, a linear dune of gypsum, clay and carbonate pellets and quartz sand (aeolian component of the Yamba Formation) has yielded TL dates of $20\,100 \pm 3\,500$ and $8\,500 \pm 1\,800$ years BP (Gardner *et al.*, 1987).

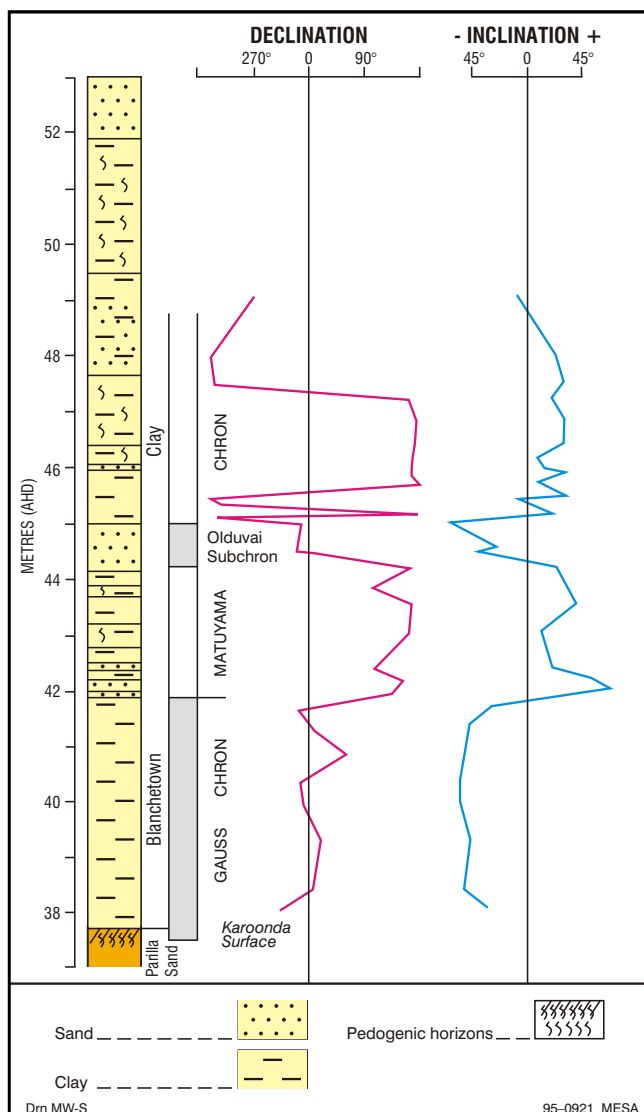


Fig. 11.27 Stratigraphic section, magnetic stratigraphy and magnetic inclination and declination for the Blanchetown Clay at 'Chowilla' (after An *et al.*, 1986).

Middle Pleistocene-Holocene dunes

Much of the surface of the Murray Basin in South Australia is composed of aeolian landforms dominated by linear and parabolic dunes of quartz sand comprising the Woorinen Formation, Molineaux Sand and Bunyip Sand.

Woorinen Formation

The Woorinen Formation comprises east-west linear dunes in the northern part of the basin. These are composed of pale reddish brown silty and clayey quartz sand with layers of pedogenic carbonate. The unit was named by Lawrence (1966), with a type section near Swan Hill in Victoria, and divided into five members separated by palaeosols: Piangil Member (youngest), Kyalite Member, Speewa Member, Bymue Member and Miralie Member (oldest).

No dates have been obtained from the Woorinen Formation in South Australia but, in Victoria, Bowler and Polach (1971) recorded radiocarbon dates of $29\,750^{+1200}_{-1450}$ and $24\,000 \pm 900$ years BP from pedogenic carbonate in the Bymue and Speewa Members, respectively. However, Readhead (1988) obtained much older TL dates of $\sim 130\,000$ years BP from quartz grains of the Speewa Member. The Kyalite Member has yielded radiocarbon dates of $16\,400^{+450}_{-560}$ (Gill, 1973) and $15\,500 \pm 230$ years BP (Bowler and Polach, 1971) from pedogenic carbonate, and a TL date of $26\,300 \pm 3\,300$ years BP (Readhead, 1988). The Piangil Member consists of modern drift sand.

Molineaux Sand

East-west linear and irregular parabolic dunes covering a large area south of Loxton are composed of the Molineaux Sand (Firman, 1973), a pale yellow, fine to medium-grained quartz sand with some pedogenic carbonate in the lower part. Molineaux Sand is equivalent to Lowan Sand (Lawrence, 1966) of the Victorian dunefields. Ohmori and Wasson (1985) obtained radiocarbon dates of $27\,900^{+3100}_{-5000}$ and $19\,400^{+740}_{-620}$ years BP from charcoal near Mindarie. TL dates ranging from $34\,500 \pm 5\,000$ to $8\,000 \pm 900$ years BP were recorded at Mindarie and two other locations (Gardner *et al.*, 1987).

Bunyip Sand

The Bunyip Sand (Firman, 1966a) occurs mainly as large tongues of irregular linear and parabolic dunes trending east-northeast from the Murray River valley between Morgan and Swan Reach. The unit also includes source-bordering dunes within the Valley (see below). Bunyip Sand is a pale red-brown quartz sand with some pedogenic carbonate in the lower part of the unit. TL studies on the west bank of the Murray River at Roonka have produced dates ranging from $\sim 27\,000$ to $2\,700 \pm 300$ years BP (Prescott, 1983).

Most of the dates obtained from aeolian deposits in the Murray Basin fall within the period from $30\,000$ to $15\,000$ years BP. The present aeolian landforms and most of the aeolian sequences therefore formed during the last glacial period which reached its maximum at $\sim 18\,000$ years BP. Younger dates indicate periods of reworking during the Holocene. TL dates of $\sim 130\,000$ years BP from the Speewa Member of the Woorinen Formation suggest that this unit formed prior to or during the last interglacial. The Bymue and Miralie Members of the Woorinen Formation represent older periods of dune building which may extend back to $\sim 400\,000$ years BP, comparable with the interpreted lower age limit for the evaporitic Tyrrell Beds.

Late Pleistocene-Holocene sequence of the Murray River valley

The contemporary Murray River valley was incised during a period of low sea level at $\sim 600\,000$ years BP when coastal barriers of Pliocene and Early Pleistocene age were breached, allowing drainage of Lake Bungunnia. The Murray River valley has subsequently been infilled with fluvial sediments largely reworked from Pliocene sands. The valley fill can be divided into two sequences: Late Pleistocene upper terrace deposits, and the latest Pleistocene to Holocene Monoman and Coonambidgal Formations (Fig. 11.28), both laid down during post-glacial periods. Coarse-grained, high-energy fluvial deposits pass up into finer grained, low-energy flood-plain deposits as a result of river gradient adjusting to rising sea level.

The upper terrace deposits are preserved mainly in the wide flood-plain section of the Murray River valley upstream from Renmark. They occur as broad eroded remnants of clayey fine sand, up to 7 m above the modern flood plain. A palaeosol characterised by reddening and pedogenic carbonate is developed at the top of the terrace remnants. The unit has been termed Rufus Formation by Gill (1973) and similar deposits occur in stratigraphic drillhole PD 2 at the Chowilla dam site (Ludbrook, 1960b; Firman, 1973, fig. 22).

Small eroded remnants of upper terrace deposits, elevated $2\text{--}4\text{ m}$ above the modern flood plain, occur in the gorge section of the Murray River valley in the Blanchetown-Roonka area (Rogers, 1990). These form an upward-fining succession up to 16 m thick, consisting of a basal unit of medium to coarse-grained sand, a lower fluvial unit of silt and very fine to medium-grained sand, and a lagoonal to flood-plain unit of silty and clayey, very fine to fine-grained sand and sandy clay. The lagoonal and flood-plain unit is partly overlain and cut into by an upper fluvial unit of silty very fine to medium sand (Fig. 11.28). The pedogenically altered top of the upper terrace deposits has yielded a TL date of $65\,000 \pm 12\,000$ years BP (Prescott, 1983). This date suggests that

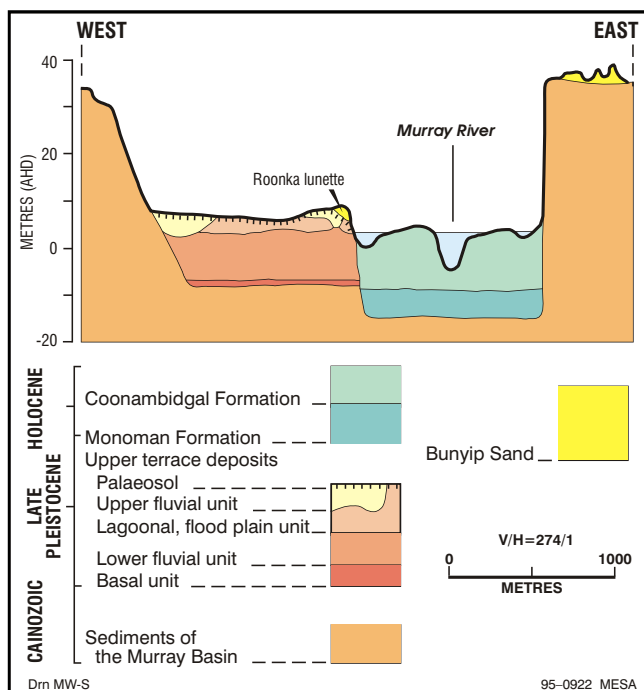


Fig. 11.28 Geological section of the Murray River valley at Roonka.

the upper terrace deposits at Roonka can be correlated with a period of rising sea level between ~65 000 and 59 000 years BP as seen in the Late Pleistocene sea-level curve of Chappell and Shackleton (1986).

In South Australia, the younger fluvial sequence of latest Pleistocene to Holocene age that occupies most of the Murray River valley is divided into Monoman Formation and overlying Coonambidgal Formation (Firman, 1966a, 1973). The Coonambidgal Formation was defined originally by Lawrence (1966) for all the alluvial deposits of the Murray River valley, including equivalents of Monoman Formation. Combined thickness of Monoman and Coonambidgal Formations ranges from 16.5 m at Pelican Point (11 km south of Morgan) to 31 m at the Chowilla dam site, upstream from Renmark.

Downcutting by the Murray River during the sea-level lowstand of the last glacial maximum extended to 15 m below present sea level at Blanchetown (compared to -8 m at the ~65 000 years BP sea-level lowstand), and to -65 m near Murray Bridge. The Monoman Formation, consisting of medium to coarse-grained sand with basal fine gravel, was laid down during the following post-glacial transgression. Fossil marsupial bones from the lower part of the unit at 'Chowilla' have been identified as *Phascolonius cf. gigas* and *Macropus ferragus*, and may indicate a Late Pleistocene age (Marshall, 1973).

By ~7 000 years BP, as the sea approached its modern level around the South Australian coast (Belperio *et al.*, 1983), depositional surfaces in the Murray River valley became more stable and forests developed. At 'Chowilla', the Monoman Formation is separated from overlying Coonambidgal Formation by an horizon of logs and stumps. Radiocarbon dates of 7 200±140 and 4 040±100 years BP have been obtained from the wood, the younger age coming from the upper part of the horizon (Firman, 1971; Gill, 1973).

Deposition of Coonambidgal Formation commenced at ~7 000 years BP, and continues to the present day. The formation consists mainly of fine-grained sand and sandy clay, with some medium to coarse-grained sand and gravel, and includes modern point bar, lagoonal and flood-plain deposits.

Aeolian dunes of Bunyip Sand (Firman, 1966a) at several locations in the Murray River valley, including a lunette at Roonka, were formed by aeolian reworking of upper terrace deposits during

the last glacial period. The Tartangan Beds, described by Hale and Tindale (1930) from Tartanga Island 14 km south of Swan Reach, may be an eroded remnant of a similar aeolian sequence. Tindale (1957) reported a radiocarbon date of 6 020±150 years BP from shells in this unit.

In post-glacial times, the Murray River valley supported a large Aboriginal population. Aboriginal remains, including middens, fireplaces and skeletons, are found at many locations including Roonka and Tartanga Island. These remains occur mainly in aeolian sands associated with upper terrace deposits. Radiocarbon dating of bone and charcoal from the Roonka lunette has produced a series of dates extending from 18 050±340 years BP to the modern period (Pretty, 1977).

Hydrogeological investigations in alluvium of the Murray River valley have revealed that the flood plain is a zone of discharge, through the process of evapotranspiration, for the regional aquifer systems of the Murray Basin (Barnett, 1990). This has led to concentration of groundwater salinities to well over 35 000 mg/L, with the subsequent storage of large quantities of salt in the alluvium and salinisation of low-lying areas on the flood plain. The salt is easily mobilised into the river during a flood recession, resulting in a temporary increase in river salinity.

SOILS, PALAEOSOLS AND CALCRETES

M.J. Sheard

Australian soils have developed over a long period of time and it has been argued that the continent has some of the oldest surviving soils in the world (Beckmann, 1983; Stace *et al.*, 1968; CSIRO Division of Soils, 1983). The age is, however, problematic in many areas. Stratigraphic evidence for a maximum age and microfossil evidence or charcoal dating for a minimum age have been employed. Lithosols in arid areas of the State may extend back to the Early Pleistocene, while some desert loams may date back to the Middle Pleistocene. Soils forming in areas of active sedimentation, especially on piedmont slopes of the Mount Lofty and Flinders Ranges, are <20 000 years old.

In addition to parent material, soils also reflect local vegetation and climatic conditions and, because of climatic fluctuations that have affected the landscape, many are not in equilibrium with their modern environments. Residual hardpans, including calcretes, silcretes and ferricretes, are prominent features of much of the surficial mantle. Additionally, many Australian soils have formed from transported pre-weathered materials rather than only from *in situ* weathering of parent materials.

South Australia has a broad range of soils that are dominated by sand and calcium carbonate, with lesser areas of loam and clay (Fig. 11.29). Many soils were described by Northcote and Skene (1972) as having highly saline-sodic properties. Layers developed through soil forming processes are referred to as soil 'horizons'. They are distinguished from each other by variations in colour, texture, structure and mineralogy, and form the basis for soil classification and mapping.

Principal Great Soil Groups of South Australia

The summary soil map (Fig. 11.29) uses Great Soil Groups (after Stace *et al.*, 1968) to show landscape areas where a particular soil group is dominant. Such map units necessarily include a wide variety of soil types and associations. To assist in correlation with more recent work, the more important principal profile forms from Northcote (1979) are included in the legend.

Lithosols are essentially thin, stony or gravelly mantles, lacking horizon development other than organic accumulations in an A1 horizon. Usually they are shallow sands, loams or clay loams with a large proportion of fragmentary rock material. Lithosols occur where erosion has been sufficient to prevent development of thick soil cover, the parent material is resistant to weathering, or the soils are very young. This group contains many variants and further

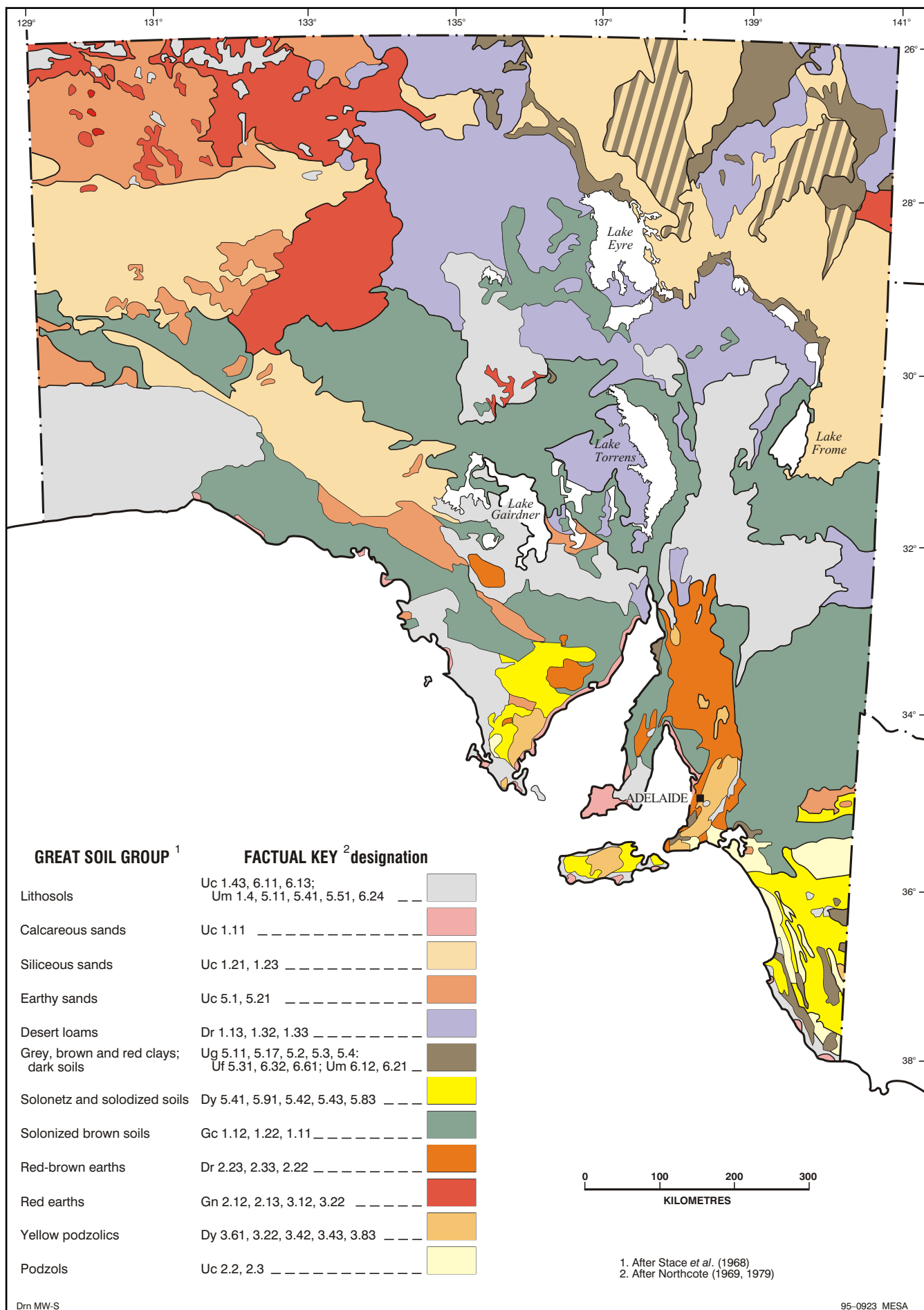


Fig. 11.29 Great Soil Groups of South Australia.

work is required to fully classify all of these. Soils of this group are developed over the Flinders and Olary Ranges, Nullarbor Plain, and coastal calcarenite dunes of Eyre Peninsula, Kangaroo Island, southern Yorke Peninsula and the South-East.

Calcareous sands show no profile development beyond a little organic matter in the surface horizon when sufficiently stabilised by vegetation. They occur on southern Yorke Peninsula, as narrow coastal strips around Eyre Peninsula, Spencer Gulf, Gulf St Vincent and along the South-East coast. These are recent calcareous sands of littoral and coastal aeolian origin, with the calcareous fraction being mostly skeletal fragments. Erosion and remobilisation by wind is a characteristic problem where these soils are denuded of vegetation. Older forms tend to be more siliceous due to carbonate leaching and range from pale yellow to pale brown in colour.

Siliceous sands form a broad group of permeable and excessively drained soils with considerable colour variation. They are dominantly quartzose, sand to clayey sand, with deep profiles and little or no horizon differentiation. There are several subgroups, based on grain size and colour. Siliceous sands are well represented in the arid interior where they constitute dunes and swales of the sandy deserts.

Earthy sands are characterised by uniform profiles of dominantly red, coherent, siliceous sand, clayey sand or light sandy loam. The earthy appearance is due to coating and bridging of sand grains by clay and iron oxides. Earthy sands are the coarser textured equivalents of the red earths (see below). There is weak development of horizons with gradational boundaries and the profiles may contain mineral nodules (glaebules) and pisoliths. Earthy sands are principally associated with low tablelands in the northwest and on Eyre Peninsula.

Desert loams have moderate horizon texture contrast with thin loamy A horizons clearly separated from structured clay B horizons, are brown to red in colour, and have an alkaline reaction trend. These soils occur on wide alluvial plains and associated stony uplands — the bevelled tablelands and cuevas in the arid areas (generally <250 mm annual rainfall) of central and eastern South Australia. Desert loams have formed on a variety of parent materials, believed to include moderately clayey colluvium, alluvium and weathered precursors or derived by aeolian reworking as proposed by Jessup (1960).

Grey, brown and red clays and dark soils form a broad group whose common properties are determined by their high clay contents (50-80%). Typically, they are moderately deep to very deep soils with uniform colour and texture profiles, weak horizon development, and some carbonate and/or gypsum in their lower profile. They crack deeply on drying and many members are described as self-mulching by Northcote (1969). These soils are widespread and occur in the South-East and northeastern parts of the State. For convenience, the small occurrences of dark soils, including black earths, rendzinas, chernozems and wiesenboden are included here.

Solonetz and solodized solonetz have prominent texture differentiation with an abrupt boundary between loamy A and clayey B horizons, neutral to alkaline surface soil and strongly alkaline sub-soil. Sodium and magnesium dominate the exchangeable cations in the lower B horizon. The solodized solonetz have a similar texture contrast between A and B horizons but with a very abrupt wavy boundary. The B horizon has a blocky or strongly coarse columnar structure. A and B horizons are strongly acidic, grading to strongly alkaline in the B-C transition. These soils are distributed in the sub-humid areas of the South-East and on Eyre Peninsula. They often occur on medium-textured alluvium-colluvium as well as on a wide range of weathered rocks.

Solonized brown soils, formerly known as malleisols or mallee soils, are characterised by large amounts of calcareous material present as fine earth and soft to hard segregations in the profile. Generally these soils are grey-brown, brown or red-brown, and sandy to loamy in texture. Maximum carbonate concentrations occur 0.3-1.0 m below the surface and this usually produces a pale

colouration with depth. Most changes in the soil profile are gradual. Solonized brown soils are widely distributed in the semiarid to arid (<400 mm annual rainfall) areas of South Australia. They are derived from highly calcareous, light to medium-textured materials where some salt and carbonate leaching in the upper profile has taken place.

Red-brown earths characteristically have grey-brown to red-brown, weakly structured to massive, loamy A horizons and an abrupt texture boundary between A and B horizons; the brighter brown to red-brown B horizons have a well-developed medium prismatic to blocky structure. The surface soil is usually moderately thick and mildly acid to neutral, while the B horizons are alkaline. These soils are found in the temperate areas of the State with annual precipitation of 375-625 mm, in particular the southern Flinders and Mount Lofty Ranges, and parts of Yorke and Eyre Peninsulas. Differential weathering of alluvial-colluvial deposits and a wide range of sedimentary, metamorphic and igneous rocks, in combination with some salt and mineral leaching and moderate rainfall, have given rise to these soils.

Red earths are distinguished by massive, predominantly sandy textured, porous profiles, red-brown to red colour, and weak profile differentiation. Horizon boundaries are gradational and there is a gradual increase in clay content with depth. They have a characteristic 'earthy fabric' which is evident as a dull matt or dusty appearance on the face of freshly broken soil, in contrast to the glossy or polished appearance of more clayey soils. These soils occur predominantly in the northwest of the State (Major, 1973d). They form by a process of weathering and leaching of siliceous parent materials with some differentiation of clay accompanied by mobilisation of iron and manganese oxides.

Yellow podzolics are distinguished by strongly differentiated profiles in which a light to medium-textured A horizon is clearly separated from a predominantly yellow-brown to brown, friable, more clayey B horizon. A distinct pale A2 horizon is also present, as is an acid reaction throughout the profile. These soils are the yellowish equivalent of red podzolic soils and intermediate forms, often with brownish mottling, also occur. The main occurrences are in the southern Mount Lofty Ranges, lower Eyre Peninsula and parts of Kangaroo Island. Strong weathering of siliceous parent materials under humid or seasonally humid climatic conditions and slow leaching combine to form these soils. Salts and some silica are lost and the clays are acid kaolins which are considerably undersaturated with bases. Prolonged periods of waterlogging apparently allow hydration of oxides that yield the yellowish colours.

Podzols are acid sandy soils with strongly differentiated profiles distinguished by grey to black A1 horizons (due to organic accumulation), thick white (bleached) A2 horizons, and B horizons of illuviated organic matter and sesquioxides. They are slightly clayey in texture and vary from reddish to yellowish brown and black in colour. Podzols occur sporadically throughout the South-East, mostly confined to the siliceous sands (see above; Fig 11.29) and sandy parent materials. They require free-draining conditions, high rainfall and strong leaching, and abundant organic materials, derived from woodlands and shrublands.

Palaeosols and calcretes

Palaeosols are soils of the geological past that have been buried and/or exhumed by sedimentation and other processes. Well-developed palaeosols are indicative of periods of landscape stability or cessation of sedimentation. Clay illuviation features may be indicators of humid periods, whereas calcretes are considered indicators of semiarid conditions with a marked seasonality of precipitation. In South Australia during the Quaternary, soils have been buried by marine transgressions (Billing, 1984; Belperio and Bateman, 1986), by lacustrine and aeolian mantles and, in the South-East, by volcanic ash and lava flows (Sheard, 1978, 1983, 1986). Specific palaeosols have been described from Kangaroo Island (Northcote, 1946), south of Adelaide (Ward, 1966), near Port Wakefield (Firman and Chugg, 1963), Whyalla area (Jessup

and Wright, 1971), Port Augusta (Billing, 1984), Stuart Range (Benbow, 1983), Lake Eyre Basin (Jessup and Norris, 1971), Officer Basin (Wopfner, 1967) and the South-East (Tiller, 1958; Dodson, 1977; Blackburn, 1983).

Calcrete formation involves leaching and reprecipitation of calcium carbonate in the vadose zone, with increasing alteration of original sedimentary characteristics. The lithification of differentiated calcium carbonate horizons to form hard pans is a common polygenetic end process of soil development in semiarid environments. Hence, calcretes are largely fossil soil profiles with a high preservation potential. South Australia has ubiquitous Quaternary calcretes and related earthy carbonate palaeosols. These comprise sheet-like bodies and sedimentary 'mantling' units that occur within or on Quaternary strata. The palaeosols include chalky, peloidal, nodular and laminar forms that reflect various stages of soil development and truncation (Hutton and Dixon, 1981; Warren, 1983a; Farrand and Belperio, 1987; Phillips and Milnes, 1988).

The concept of palaeosols proposed by Firman (1986) for the Adelaide region has proven to be conjectural, as the stratigraphic connotation ascribed to them cannot always be sustained (Bowman and Sheard, 1988a,b). A stratigraphic framework applied to the calcretes and related materials resulted in palaeosols being referred to as Ripon Calcrete (Firman, 1967a,b), Bakara Soil (Firman, 1963, 1964) and Loveday Soil (Firman, 1966a). However, Belperio (1988a) and Phillips and Milnes (1988) have demonstrated a cyclic repetition of calcrete formation in coastal sequences that make this calcrete stratigraphy problematic and unworkable, given that specific calcretes cannot be distinguished lithologically. The calcretes have a complex internal structure and geochemistry, can be discordant or conformable with neighbouring units, and reflect multiple sedimentary and pedogenic events. Laterally, facies vary from thick, indurated hardpans to unconsolidated intraclast breccias and calcareous silts. In addition, pedological processes have been modified further by gravitational and aeolian processes, with aeolian carbonate dust blown well inland (Crocker, 1946; Milnes and Ludbrook, 1986; Phillips and Milnes, 1988). Climatic changes and fluctuating watertables and vegetation cover have resulted in multiple calcrete horizons, coalescing profiles, and calcretes that transect stratigraphic boundaries. This is particularly evident within calcarenites of the Bridgewater Formation.

KARST AND MOUND SPRING DEPOSITS

M.J. Sheard and P.C. Smith

Subaerial exposure of carbonate rocks, in addition to forming carbonate and calcrete weathering profiles, also results in a variety of karst features associated with meteoric vadose and phreatic dissolution and leaching. Unlike calcrete formation, where calcium carbonate is dissolved and reprecipitated within a profile, karst processes involve a net removal of carbonate in solution. The term 'karst' is used generally to designate such solution-affected limestone terrains.

South Australian karst features occur in three main carbonate assemblages: limestones, dolomites and marbles of Precambrian to Cambrian age, marine Tertiary limestones, and aeolian Pleistocene calcarenites of the Bridgewater Formation (Fig. 11.30). There are many karst features present with names derived from type areas world wide. Figure 11.31 illustrates some of the more common forms.

Karst in Precambrian to Cambrian carbonates

Exposures of indurated or recrystallised limestone or dolomite of Precambrian or Cambrian age have varying degrees of karst development from karren to major cave systems. Rock jointing has provided the main control on development and style because these rocks have no remaining primary porosity (Jennings, 1971). Bed partings and/or solubility differences, topographic drainage sinks, and the presence of soils with carbonic and humic acids which help



Carbonate-calcrete mantle in a northern suburb of Adelaide. Photo 042544.

to dissolve the carbonate are of secondary importance.

The largest cave systems in these rocks are located on Yorke Peninsula around Curramulka, where they consist of strongly joint-controlled shaft and horizontal passage mazes. They are centred on a large uvala or karst drainage area and have doline or shaft entrances (Lewis, 1976). Karst development in the carbonate rocks of the Flinders and Mount Lofty Ranges is more isolated and confined to small dolines, potholes, karren and small horizontal caves or shafts. These usually reflect strong joint control although solubility differences within the rocks contribute to local variations (Jennings, 1971). Officer Basin karst is of more limited extent, being restricted to Cambrian carbonate rocks towards the eastern margin, although little is known about this karst area or its extent.

Dating karst developments in these rocks is problematic, but there is evidence to indicate that appreciable development took place in the late Tertiary and has continued throughout the Quaternary. Some areas do not demonstrate any breaks in karst development, while others reveal evidence of several hiatuses due to cave infilling or blockages.

Karst in Tertiary limestones

Marine Tertiary limestones are located in the Eucla, St Vincent, Murray and Gambier Basins. They exhibit karstic features in surface and subsurface forms. Water movement is controlled by both primary and secondary porosity. The sediments display marked permeability variations due to lithology, bedding, jointing, faulting, facies changes, and existing karst solution features. More detailed lithological accounts of these host rocks are provided in Chapter 10. Karst location, extent and form has been strongly controlled by variation in host rock permeability and structure.

Two basins where karst on Tertiary strata is well developed and well documented are the arid Nullarbor Plain and the much wetter Mount Gambier coastal plain. Karst development in the St Vincent and Murray Basins is not as extensive. Examples from these basins are given by Allchurch (1967), Weatherby (1974), Lewis (1976) and Waterhouse and Gerges (1979).

Nullarbor Plain karst

The Nullarbor Plain covers an area of 200 000 km² and has remained exposed as a land surface for the past 14-16 million years. Karst features are formed within Tertiary limestones of the Eucla Group. Jennings (1963) described Nullarbor karst as being 'retarded and immature and on the whole there is a poverty of caves'. However, the work of the Cave Exploration Group of South Australia (CEGSA) and others has documented a wide variety of karst forms on and beneath these plains (Pilkington and Mott, 1986; Lewis, 1976). Surface karst features include uvalas, dolines, sinkholes, potholes and karren, with subsurface passages, mazes, chambers and lakes. Many of the known cave systems occur in the

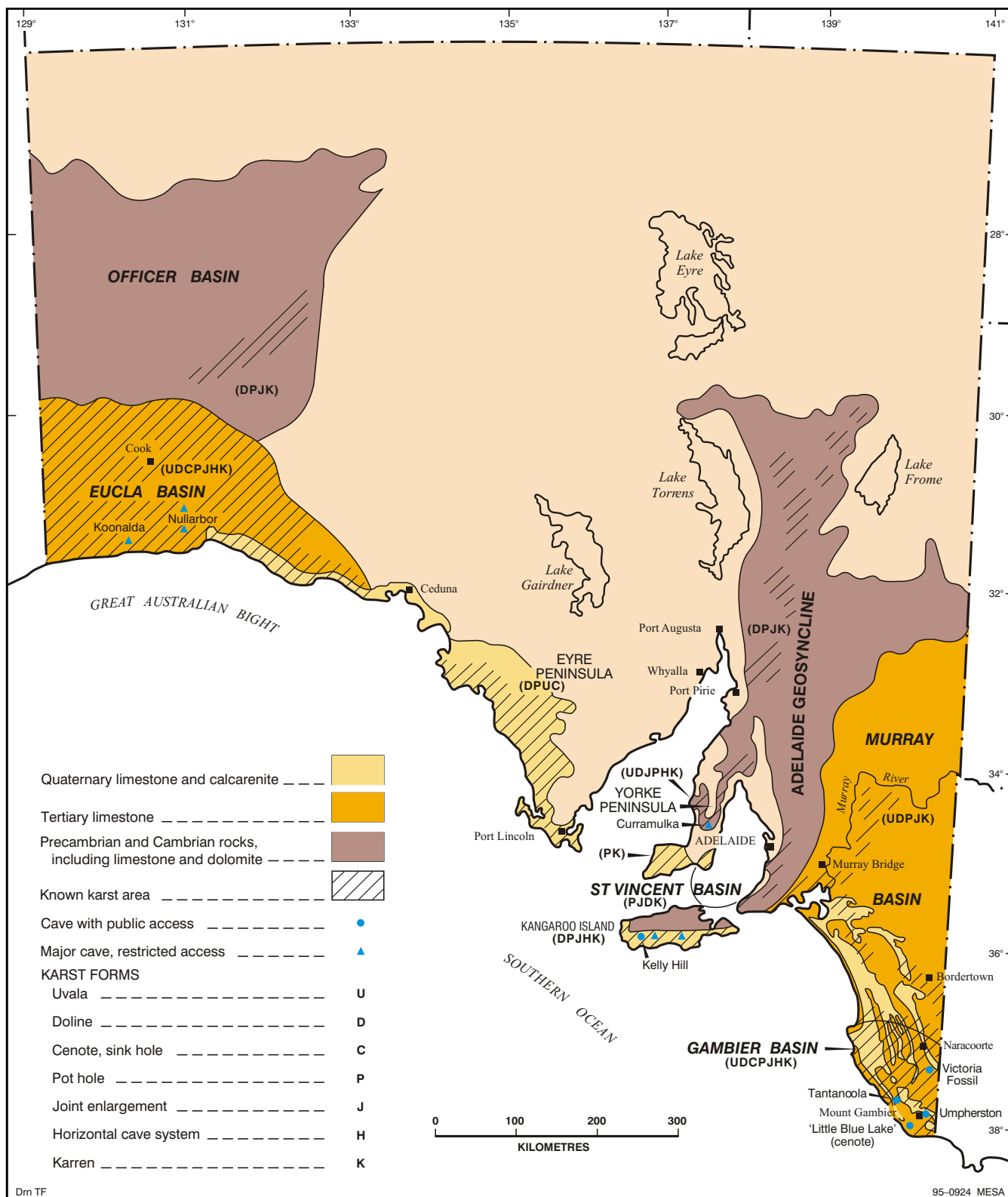


Fig. 11.30 Major regions of karst formation in South Australia.

Western Australian part of the plain. Within the South Australian portion there are 165 designated karst features, the vast majority of which are located within 25 km of the present coastal cliffs and the higher rainfall zone of the plain (Pilkington and Mott, 1986). Koonalda Cave is arguably the best known of these and consists of an entrance at the base of a doline, extensive passage ways, several collapse chambers, four exposed watertable lakes, and some

submerged tunnels; its overall length is 1 640 m. This cave was an important Aboriginal flint mine and workshop during the Late Pleistocene to early Holocene, and contains Aboriginal markings from the same period (Lowry and Jennings, 1974; Martin, 1973b).

Ages for the cave systems are problematic and dependent on radiometric dating of speleothems, cave earths, bones and charcoal, none of which necessarily date the timing or initiation of solution



activity. Karstic processes most likely began in the Late Miocene to Early Pliocene and continue through to the present day.

Karst in the Gambier Limestone

In contrast to the arid Nullarbor Plain, the South-East experiences a seasonally wet, humid to sub-humid climate. Spectacular karst features are developed throughout the region, with notable concentrations near Naracoorte and to the southwest of Mount Gambier (Lewis, 1976; Grimes, 1994). Overall, this karst region extends as a broad band southwards from Naracoorte towards and beyond the present coast, reaching a maximum width of >60 km near Nangwarry (Marker, 1975). The region hosts most forms of temperate climate karst in two rock units, the Gambier Limestone (Miocene) and the Bridgewater Formation calcarenites (Pleistocene). Karst in the latter unit is described in the next section. Karst in the Gambier Limestone is more spectacular and extensive, and includes large cenotes, complex joint-controlled cave systems and dolines.

In the Mount Gambier area there are numerous accessible caves, some of which are open to the public. These include Tantanoola Cave, with well-developed speleothems (stalactites, stalagmites, helictites), Umpherston Cave, Engelbrecht Cave and Mitchell Cave (Cave Gardens, Mount Gambier). These caves commonly intersect the watertable. There are dozens of cenotes in this area, with two fine examples being 'Little Blue Lake' and 'Hells Hole', while 'The Shaft' forms a spectacular water-filled collapse chamber with a narrow entrance. The depth to which karst features have developed increases towards the coast due to an increasing thickness of the Gambier Limestone and lower sea levels and watertables during Pleistocene glacial periods.

Within the Naracoorte area there are also numerous caves, some of which are open to the public. Most of these include doline to horizontal passage solution and collapse complexes. These caves are usually hosted by Tertiary limestones, although some are partly within Pleistocene aeolianites. The best known are around the Victoria and Blanche Caves, where significant bone-bed fossil deposits are also located. 'Victoria Fossil Cave' contains the largest such deposits in the world, with estimates exceeding $2 \times 10^5 \text{ m}^3$ of fossiliferous mud, sand and clay. This cave complex has acted as a repository for animal (including bird) skeletal remains derived from doline 'pit traps' and animal lairs (Wells, 1975; Wells and Pledge, 1983). Blockage of doline entrances to subterranean chambers by swamp mud and sandy infill, with repeated flushing due to swamp breakthrough or heavy rain, provided a unique environment for bone retention, burial and preservation. Remains range from delicate bones of frogs, small marsupials, birds and bats to more robust and now extinct browsing kangaroos, diprotodontids, carnivorous thylacines, emus and snakes. More than 90 species have been recognised and thousands of partial or complete skeletons exhumed. Dates for these bones have yielded ages ranging from >150 000 to 18 000 years BP (Wells, 1975; Williams, 1980; Wells and Pledge, 1983). This cave and its bone deposits are of international significance and have been placed on the Register of the National Estate.

Karst in Pleistocene aeolianites

The Pleistocene aeolian calcarenites of the Bridgewater Formation, occurring over much of western and southern coastal South Australia (Fig. 11.30), are variably cemented by pedogenic and vadose diagenetic processes and affected by karstic solution.

Karstic processes were thought by Jennings (1968b) to have been partially syngenetic, with caves well developed as a result of lateral solution at the watertable. However, Jennings considered that vadose action dominates karst processes within these terrestrial to marginal marine calcarenites. Karst occurs within these sediments on the west coast of Eyre Peninsula, the southwestern end of Yorke Top: Karst surface developed on Gambier Limestone near Mount Gambier. Photo 010615.

Centre: Cave speleothems in Tantanoola Cave. Photo 036490.

Bottom: Clay pot karst developed in Bridgewater Formation, 15 km south of Penola. Photo 042545.

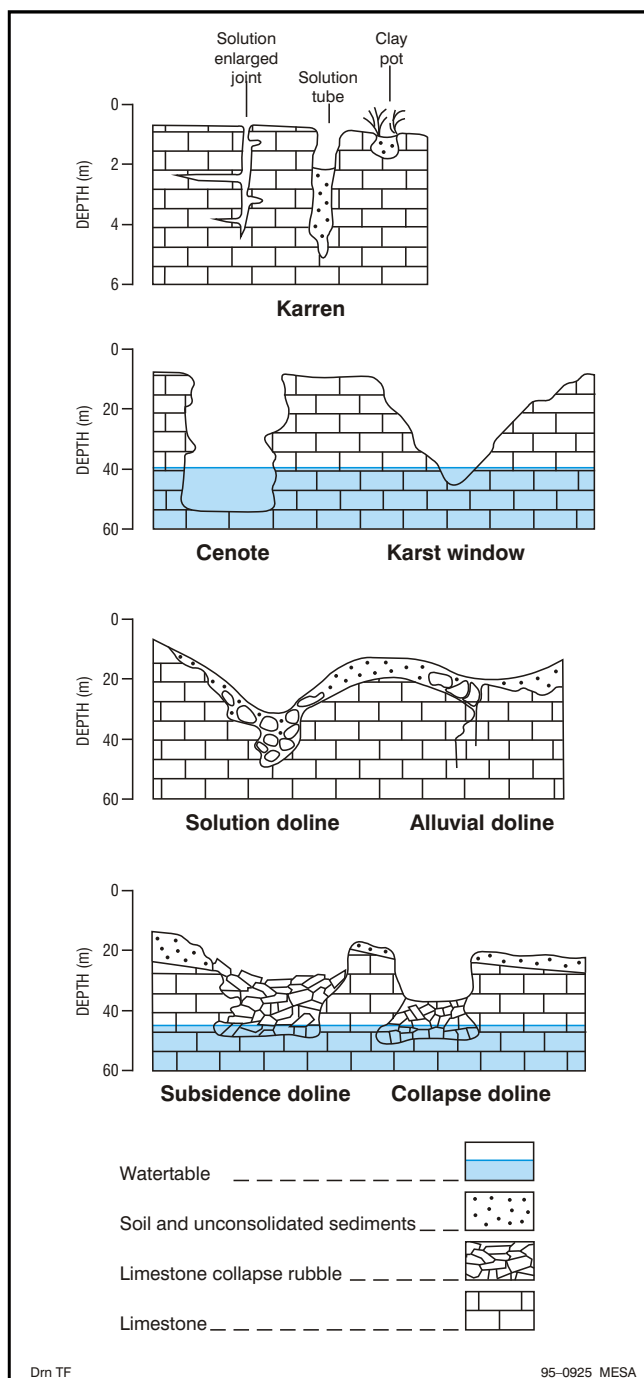


Fig. 11.31 Some common karst forms (after Bögle, 1980).

Peninsula, southern Kangaroo Island, and on the Mount Gambier coastal plain.

Eyre Peninsula karst is characterised by large, dry sinkholes (e.g. in the vicinity of Talia), and sea caves. Karren on calcreted surfaces are ubiquitous and solution-enlarged joints and potholes are common. That karst processes are still active is demonstrated by 'comparatively recent subsidence ... in the Cape Wiles National Park near Port Lincoln' and other areas (Lewis, 1976). More than 25 caves have been documented in this region. Where topographic depressions (?solution dolines) occur within subcropping Bridgewater Formation calcarenites with little or no vegetation, low salinity groundwater can often be found. These water resources are vital to many of Eyre Peninsula's coastal towns and are strongly related to the occurrence of karst (e.g. the groundwater basins of Counties Flinders and Musgrave). Yorke Peninsula karst is poorly developed, being primarily restricted to meso-scale features such as potholes and solution enlarged joints (grikes and karren). Caves

may be present but are not documented. As with Eyre Peninsula, karst is partially responsible for shallow, low salinity groundwater lenses (e.g. Carribe and Para Wurlie 'Basins').

On Kangaroo Island the best documented karst feature is the Kelly Hill Caves developed in Pleistocene calcarenites. This is an extensive collapse system covering an area of 200x180 m and extending to 30 m depth, with excellent decoration and bone deposits (Pledge, 1979). Numerous caves punctuate the coastline, some eroded by wave action and others exposed by cliff collapse. Lewis (1976) documented 36 caves on Kangaroo Island, mostly within the Bridgewater Formation. Lagoons have developed in sinkholes (dolines) near Vivonne Bay.

In the South-East, syngenetic karst features are developed in calcareous dunes of the Bridgewater Formation (Grimes, 1994). These are generally restricted to 'clay pots' (soil-filled undercut karren), karst pavements, 'flatteners' (horizontal solution features) and some cave development along the coast.

Artesian (mound) spring deposits

The Pleistocene to Holocene artesian springs of South Australia occur along the southwestern margin of the Great Artesian Basin (Fig. 11.32). They provide one of the mechanisms of natural discharge from the basin, with a total calculated flow in the State of 23 500 ML/year from ~90 springs or groups of springs (Boucaut *et al.*, 1986). About 20 500 ML/year or 90% of the total is derived from a group of ~60 discrete sources in the Dalhousie Springs area.

The line of mound springs extending from Marree to near the Northern Territory border has played an important part in the State's European development, as well as providing a focus for Aboriginal occupation (Harris, C.R., 1980). The early explorers, Overland Telegraph, Ghan railway between Port Augusta and Alice Springs, and pastoralists have all used the springs and their associated ecosystems. Krieg (*in* Zeidler and Ponder, 1989) gave an account of the natural history of Dalhousie Springs, the State's largest artesian spring group, and Krieg (1985) outlined its geological evolution.

Mound springs occur under some or all of the following conditions (Habermehl, 1980; Aldam and Kuang, 1988):

- uplift of basement causing erosion of the confining bed
- reactivated basement faults that breach aquifer(s)
- topographic lows associated with eroding stream channels.

Individual discharge from springs ranges from a low rate barely able to support associated vegetation (commonly cutting grass, reeds, rushes, sedge, ti-tree and acacias), to a measured flow of between 140 and 170 L/sec at Dalhousie (Williams, 1974; Boucaut *et al.*, 1986). Water salinity varies between 700 and 8 000 mg/L, and source temperature is in the range 20 to 46°C. Height of associated mounds varies from a few centimetres to 40 m at Beresford Hill.

The springs are part of a dynamic system controlled by the interplay between erosion, deflation, seismic activity, and hydraulic pressure of the artesian aquifer(s). As a result the springs and associated mounds constantly change their morphology, discharge rate and vegetation. Some mounds are 'extinct' (e.g. Mount Hamilton) whilst others are nascent (the so-called 'mud heaps' at Dalhousie).

The interpreted structure (schematic) of a typical mound spring is shown on Figure 11.33. Groundwater from a generally shallow artesian aquifer (in South Australia either the Algebuckina Sandstone and/or Cadna-owie Formation) moves via conduits (often fault controlled) to the surface. The upward-moving groundwater often has sufficient velocity and the conduits are wide enough for the transfer of aquifer and/or confining bed sediment. At the surface, particles (often with aeolian sand) are cemented to varying degrees by precipitated minerals from the evaporating groundwater.

There are few records documenting the internal composition of mound spring deposits. Limited drilling (Forbes, 1959; Ludbrook, 1959) has been undertaken near Coward Springs (Fig. 11.32) to locate fluorine deposits, postulated to be concentrated from

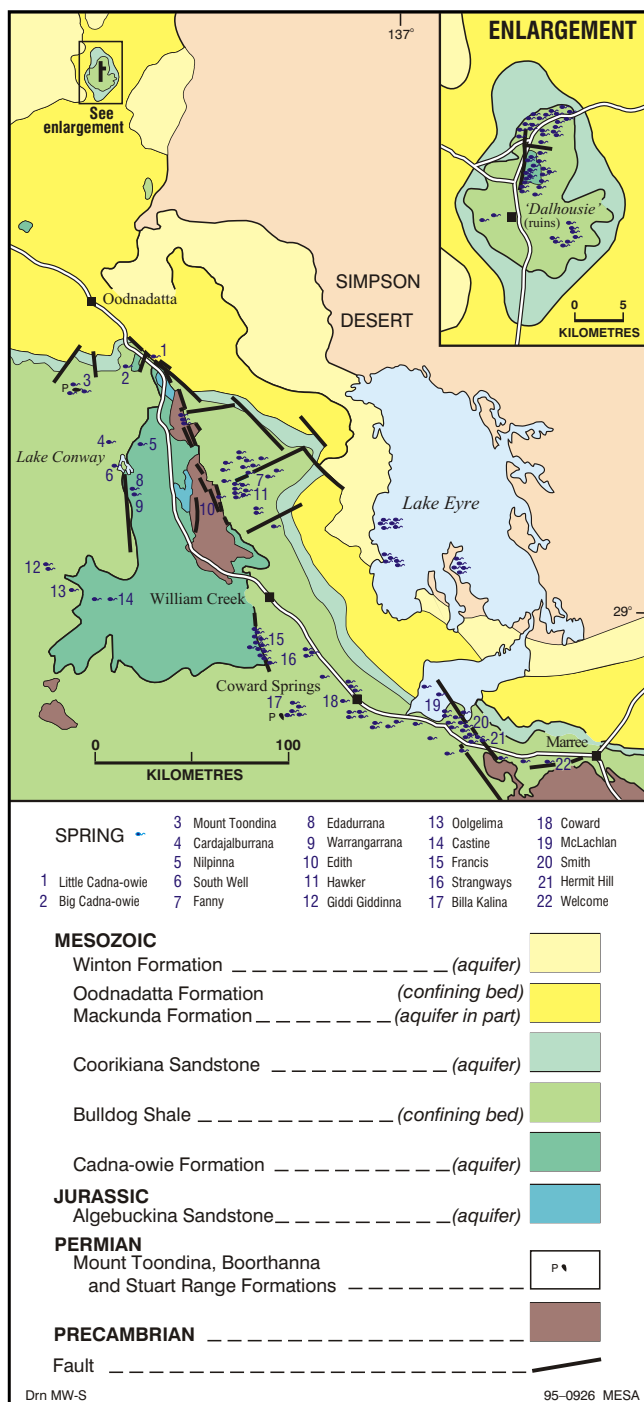


Fig. 11.32 Springs and regional geology of the southwestern margin of the Great Artesian Basin.

evaporating groundwater. Fluorine anomalies were not identified but the composition of the mound was found to be '... tufa, sand and clay deposited by the spring on an old freshwater limestone or travertine' (Ludbrook, 1959). Early Cretaceous (Aptian) foraminifera were identified which were interpreted as being carried upward from the aquifer(s) and confining beds. One hole penetrated possible Cadna-owie Formation at 16.8 m, which confirms mound spring occurrence associated with shallow artesian aquifers.

Thomson and Barnett (*in* Greenslade *et al.*, 1985) classified mound springs into four categories (active, waning, non-active, extinct) according to relative age and morphological characteristics. Petrographic examination of limestone samples from Beresford, an extinct mound, indicate a higher iron oxide (goethite) content and a generally greater degree of crystallisation than carbonates from an adjacent younger spring. Algal layering and brecciated

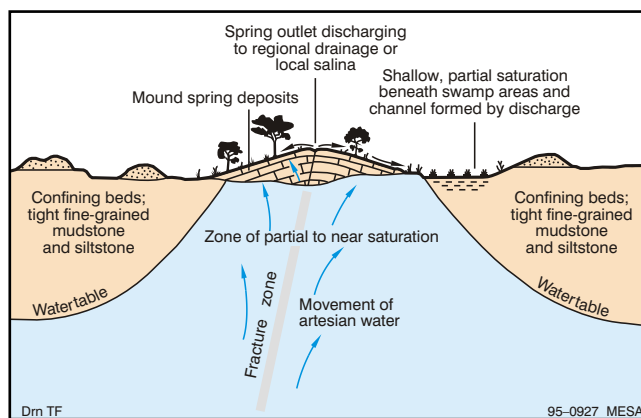


Fig. 11.33 Section of the inferred structure of a typical active mound spring.



Top: Units Czm and Qm, aerial view of Beresford Hill area, looking south east along alignment of large extinct mound springs and smaller active mound springs. Photo 039109.

Bottom: 'The Bubbler' spring near Coward Springs. Photo 042546.

and intraclastic textures are recognised, and an age of no older than Pleistocene is suggested.

The mound spring deposits generally contain some or all of the following: 'spring limestone' (travertine/calcrete) with reed casts, gypsiferous, calcareous and carbonaceous silt; aeolian sand and layers of organic silt. A salt crust often forms on the surface. Minor sulphur accretions have been reported from the Strangways area.

The mound springs and their associated deposits, flora and fauna are ecologically significant 'oases' in the northeast deserts area (C.R. Harris, 1980; Greenslade *et al.*, 1985; Zeidler and Ponder, 1989). Dalhousie Springs is protected by incorporation into Witjira National Park; it is classified as a geological monument (Toteff and McBriar, 1979) and is listed under the National Estate. Springs

likely to be affected by large-scale groundwater withdrawal in the Hermit Hill area are being monitored as an aid to management of this unique, dynamic environment.

ECONOMIC SUMMARY

SILICA SAND

Sand and gravel of the coastal zone and inland drainage systems is an important source of aggregate, construction and filter sand for Adelaide and other South Australian population centres. Much of Adelaide's requirements are now met by Tertiary rather than Quaternary sediments. In the South-East, major deposits (Comaum, Naracoorte) have developed as coarse Bridgewater Formation beach facies and finer aeolian facies in the Naracoorte Range (Keeling *et al.*, 1985). The bulk of Port Lincoln's construction sand requirements are mined from siliceous beach sands at Louth Bay (St Kilda Formation). In the Port Pirie region, 83% of construction sand is produced from Holocene and Pleistocene stream channels. Specification filter sand for the treatment of Adelaide's water is won from the Coonambidgal Formation in abandoned channels of the Murray River at Murbko (Scott, 1985). Another source of specialised, well-sorted filter sand is from a modern beach at Baird Bay (Johnson, 1980).

LIME SAND

Active Holocene dunefields facing the Southern Ocean represent major reserves of high-grade carbonate sand. West of Port Lincoln, an active dunefield covers more than 300 km² with reserves of ~4x10⁹ m³ (Johns, 1963). Only intermittent mining has taken place for flux in the smelting of iron and lead ores.

On Kangaroo Island, soil acidity on farmland is being remedied by the application of lime sand mined from a similar active Holocene dunefield. The unvegetated, mobile seif dunes allow mining with no deleterious effects.

Around Gulf St Vincent and Spencer Gulf, small beach ridges (coastal cheniers) form a significant shellgrit resource (Pain and South, 1987). Production prior to 1936 was largely from the St Kilda area, but since then leases at Port Prime and Parham have supplied 83% of output. The main use is in cement and glass manufacture, with ample reserves at current rates of extraction.

HEAVY-MINERAL SAND AND OFFSHORE AGGREGATE

Holocene beaches and barriers are largely dominated by calcareous sediment, but siliciclastics can be important near bedrock cliffs. Mixed quartz-carbonate sands suitable for beach replenishment and aggregate occur as isolated banks and sand sheets on the seafloor in Gulf St Vincent, Backstairs Passage and Investigator Strait (Belperio *et al.*, 1990). Heavy-mineral sand concentrations of zircon, ilmenite and garnet occur at Morrisons Beach on Kangaroo Island and at several localities along the Eyre Peninsula coastline (Morris, 1978); all are subeconomic.

GYPSUM

Coastal saline lakes around South Australia are an important source of high-quality gypsum. Gypsum is similarly widespread on the floors and adjacent lunettes of inland playa lakes. Holocene gypsum formed from evaporation of marine water is termed Le Hunte Formation, whereas gypsum formed in continental playas is termed Yamba Formation. South Australia accounts for ~75% of Australia's gypsum production. In 1993, 1.3 Mt were produced with a commercial value of ~\$10 million. Principal sources for high-grade gypsum for cement and plaster manufacture are Lake MacDonnell on Eyre Peninsula, Blanchetown in the Riverland, and Spider Lake, Marion Lake and Lake Fowler on Yorke Peninsula.

In addition, numerous agricultural grade gypsum deposits supply farmers, horticulturists and fertiliser distributors. Drilling at Lake MacDonnell has outlined reserves in excess of 500 Mt of crystalline gypsum (selenite) with an average grade of 91% CaSO₄.

SALT

The coastal zone is also a preferred location for large-scale solar salt production, with both sea water and groundwater being used as evaporative feedstock. Concentrating ponds and crystallisation pans occupy large tracts of former supratidal marshes north of Adelaide, from Dry Creek to Middle Beach. The Penrice Soda Products operations include the only soda ash (sodium carbonate) facility in Australia, supplying 90% of Australia's requirements. The brine is also used in manufacturing a range of sodium and chlorine chemicals. Other coastal salt operations are at Price, Whyalla and Lake MacDonnell. Solar evaporation in inland playas is carried out at Lake Bumbunga near Lochiel, Peesey Swamp near Warooka, and Mulgondawa near Wellington.

GOLD AND DIAMONDS

Alluvial gold reworked from Tertiary leads occurs in Quaternary fluvial and outwash deposits in the Mount Lofty Ranges (Echunga, Barossa), Flinders Ranges (Angepena, Waukaringa, Clare) and on Kangaroo Island (Daws Diggings). Numerous diamonds have been recovered by gold panners, particularly near Echunga, south of Adelaide, where up to 250 diamonds have been reported. With the widespread recognition of diamondiferous lamprophyres, there exists a high potential for placer diamond plays. Large Quaternary and Tertiary palaeosystems which drained source terrains into the Eucla, Lake Eyre, Pirie, St Vincent and Murray Basins are largely unexplored exploration targets.

FLINT

Fossiliferous chert pebbles reworked from Tertiary limestone are concentrated on modern beaches of the Lower South-East. These have been worked intermittently since 1909 as a source of flint for ball mills and high-grade silica for ceramics. Current production is minimal (<<10 t/year) from a coastal resource estimated at 330 000 t (Flint *et al.*, 1989).

DOLOMITE

Quaternary lacustrine dolomite (Padthaway Formation) is common in the lee of stranded coastal barriers between Keith and Mount Gambier. The dense, fine-grained dolomite is used as road aggregate throughout the Lower South-East. Lacustrine dolomite is also common in the northern part of the State but there has been little commercial exploitation.

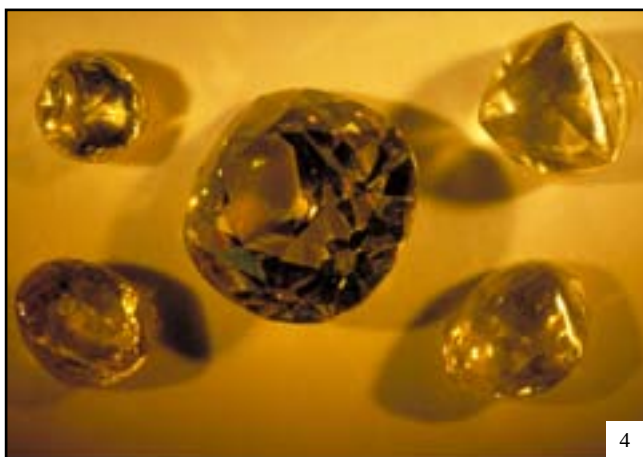
ALUNITE AND CELESTITE

There is currently no commercial production of alunite (potassium sulphate) or celestite (strontium sulphate) in South Australia, although both are widely reported in Cambrian to Quaternary sediments. Late Pliocene–Quaternary weathering and concentration in the regolith is a feature of both these minerals. A significant concentration of alunite clay occurs in the beds of saline lakes of the Pidinga palaeodrainage. Here, drilling has outlined reserves of 380 000 t averaging 6.3% potash (Rankin *et al.*, in prep.).

Celestite is common throughout the Lake Eyre Basin. It occurs as concretionary nodules and cement within Quaternary fluvial sand and gravel, as nodules within the gypsite regolith, and as cement and nodules in lacustrine beach ridges (Williams, 1972; Callen and Kennedy, 1993). The celestite originates from vein deposits in underlying Cretaceous sediments. It is commonly associated with gypsum, indicating a similar evaporative concentration process.

CLAY

The red, plastic alluvial clays (Hindmarsh Clay, Pooraka



Formation) of the Adelaide Plains are ideal for bricks and pottery, and brickmaking was one of Adelaide's earliest industries. Over 30 clay pits once operated in the suburbs of Thebarton, Brompton and Beverley. Pressure from urban development has forced these to close, and most supplies now come from Tertiary deposits and weathered shale in the Mount Lofty Ranges.

WATER

Groundwater is a vital resource for much of the State, and accounts for 35% of total water consumption. Although much is sourced from Tertiary strata, Quaternary sequences are locally important. On Eyre Peninsula, shallow groundwater is extracted from coastal aeolianites (Bridgewater Formation, Semaphore Sand). Recharge is closely related to local rainfall and over-exploitation quickly leads to saline groundwater intrusion. Along the flanks of, and within, the Flinders, Gawler and Musgrave Ranges, groundwater occurs in sand and gravel beds of modern and abandoned stream courses and alluvial fans. Small basins such as the Willochra and Walloway Basins provide enhanced quantities of water from Quaternary sediments.

TOURISM

The varied landscape of South Australia is inherited from underlying geology, modified by Quaternary weathering, erosion and deposition. Landscape and landform are the key to many of the State's most important tourist attractions and hence are of major economic significance. The rugged features of the Flinders Ranges, the flat expanse of the Nullarbor Plain and the karstic terrain of the Lower South-East, at the broadest scale, owe their existence to geological processes. Specific features such as the Blue Lake, The Coorong, Naracoorte Caves, Wilpena Pound, Coongie Lakes, Dalhousie Springs, Remarkable Rocks, to name a few, similarly have been shaped by, and owe their existence to, geological processes operating over millennia. Many of these scenic geological features are protected as geological monuments and heritage sites under the National Estate. The State's mining heritage, such as the Burra, Moonta, Kapunda and Talisker Mines, is another distinctive feature of South Australia's tourism industry (see 'Mining heritage').

ENVIRONMENTAL GEOLOGY AND GEOLOGICAL HAZARDS

INTRODUCTION

A.P. Belperio

Geological processes cause both catastrophic and slow, long-term changes to the physical environment, thereby providing the engine for ecological change and evolution. More recently, human adaptation of the environment has, in many cases, become dominant over natural change. Environmental geology covers all of these fields. In particular, our standard of living necessarily requires the use of land, water and mineral resources of the Earth, but in using those resources we can unduly affect our environment. Understanding the physical landscape of rock, sediment, soil and groundwater, associated geological processes, and how they may be modified by our activities, is vital to successful and complementary adaptation to both ecologically and economically sustainable development and effective land-use planning. The variability and extremes of natural processes, often ascertained from the geological record, must also be understood as a precursor

1. Crystalline gypsum at Lake MacDonnell. Photo 042547.
2. Southerly view towards Adelaide of the Penrice Soda Products solar evaporation salt ponds at Dry Creek. Coastal mangroves are in the foreground. Photo 042548.
3. Alunite patches in weathered Bulldog Shale, Greek Gully, Coober Pedy. Photo 039430.
4. Diamonds from the Echunga alluvial gold workings. The large faceted diamond weighs 2.84 ct. Photo 039282.

to the confident identification of subtle human-induced changes.

The aim of environmental geology is to adapt human activities to the constraints of nature, thereby avoiding both 'catastrophic' events as well as long-term unsustainable change that can impact on human existence. The undue delay in incorporating geological information and understanding into an essentially political land-use planning process is a matter of regret that has been detrimental to its effectiveness. Land-use issues related to geological processes fall into three major groupings; geological hazards (volcanic activity, earthquakes, landslides, problem soils, coastal erosion), long-term environmental change (climate and sea-level change, land subsidence, soil and groundwater contamination, dryland salinity), and transient disturbances caused by exploration and mining activity. The most important of these as they relate to geological phenomena in this State are discussed briefly below.

By world standards, geological hazards in South Australia are not excessive because we are not subject to undue extremes of relief, rainfall or seismicity. Nevertheless, a level of risk exists associated with seismic and volcanic activity. Unique hazards also result from a range of problem soils, particularly around Adelaide, although these are not necessarily 'life threatening' in the usual sense of a hazard. In contrast, environmental degradation is a serious long-term problem that necessitates increasing attention. Understanding the local geology and contemporary geological processes are fundamental to solving many of these problems.

QUATERNARY VOLCANIC ACTIVITY AND VOLCANIC HAZARDS

M.J. Sheard

Quaternary volcanism is restricted to the Lower South-East of the State and this is the only region in which a current volcanic hazard may be considered to exist. Prediction of the type of volcanic eruption and the likely area affected can be determined from studies of past eruptions and their remains. The volcanoes comprise two distinct groups that constitute a western extension of the late Tertiary to Quaternary 'Newer Volcanics' of western Victoria (Fig. 11.34).

The older (northern) Mount Burr group includes 14 fissure-controlled centres with lava flows, composite domes to cones, and tuff rings or maars (Sheard, 1983). These volcanoes represent alkaline basaltic varieties, four having been further classified by Irving and Green (1976) as ranging from K-rich nepheline hawaiite at Mount Burr, to olivine analcinite at Mount Watch, Mount McIntyre and The Bluff. Further subdivision according to basalt type and the degree of magma fractionation within each volcanic pile has so far not been attempted. Some of the magmas carried mantle-derived xenoliths (usually lherzolite) expressed as green, granular to 'sugary' textured inclusions within lava flows or as bombs, blocks and lapilli (20 to 500 mm diameter) within the tephra piles. These xenoliths have been recorded at Mounts Edward, Lyon and Watch (Irving and Green, 1976; Sheard, 1983; Nixon, 1987; O'Reilly *et al.*, 1988). They are considered to represent remnants from partial melting of upper mantle lithotypes. In addition, the tephra piles contain abundant non-igneous country rock fragments, mostly Cretaceous and Tertiary sedimentary rocks through which the magma conduits passed. There are also rare examples of deeper rocks including upper crustal basement lithotypes (granodiorite, etc.; Sheard, 1978, 1985).

Most of the volcanoes in the Mount Burr group exhibit erosional modification by Pleistocene interglacial high sea levels and mantling to some degree by Bridgewater Formation calcarenites. Radiometric dating of the Mount Burr group has so far been unsuccessful due to rock alteration and weathering, and the presence of ubiquitous calcium carbonate. However, palynological work by Dodson (1974, 1977) and other geomorphic and stratigraphic evidence suggests volcanic activity in a time span of 1 000 000-20 000 years BP (Fig. 11.35).

Southeast of Mount Burr lies the second, younger group of



Fig. 11.34 Mount Gambier and Mount Burr groups of Quaternary volcanic eruptive sites.

volcanoes consisting of Mount Schank and Mount Gambier. Detailed eruptive histories for these volcanic complexes are presented by Sheard (1978, 1983, 1986). Because both members of this southern group overlie the Bridgewater Formation (Fig. 11.35), they are stratigraphically younger than the Mount Burr group and are considered to be mid-Holocene. Mounts Gambier and Schank are dominantly constructional features produced by phreatomagmatic explosive interactions, rather than collapse structures analogous to calderas. Basaltic lithologies determined by Irving and Green (1976) are nepheline hawaiite for Mount Gambier and K-rich nepheline hawaiite or nepheline trachyandesite for Mount Schank. Both magmas have been derived by partial melting of basanite parent materials in either deep crustal or upper mantle environments.

Mount Gambier

Initial eruptions commenced at two sites, one at the western end of the present-day complex (Tenison College oval) and the other near the Leg of Mutton Lake crater (Fig. 11.36). According to radiocarbon dating by Blackburn *et al.* (1982) and palaeomagnetic data in Barbeti and Sheard (1981), these eruptions commenced ~5 000 to 4 300 years BP. Small, low maars were quickly formed, blanketing the surrounding area in up to 2 m of ash and lapilli. Lava then flowed from a fissure (western part of the complex) and a vent (Leg of Mutton Lake crater area), but the two relatively small flows did not coalesce. A scoria cone developed adjacent to, and partly over, the western lava flow. A section through this cone is visible in the Brownes Lake western crater wall due to subsequent exhumation by volcanic explosions. During this first period, the eruptions passed from mildly phreatic and passive Icelandic-style flows to weakly explosive Stromboli-style cone building.

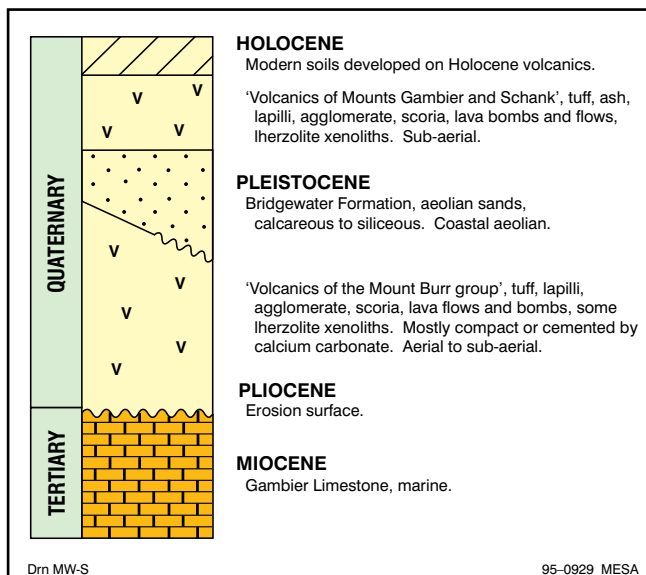


Fig. 11.35 Stratigraphic relationships of Quaternary volcanics in the South-East of South Australia.

Activity ceased for up to 300 years according to the palaeomagnetic evidence of Barbetti and Sheard (1981), allowing complete lava crystallisation and cessation of flows to occur (Sheard, 1978). A second stage of activity was more dramatic and involved a much larger area. Groundwater interaction with several ascending magma columns resulted in a series of phreatomagmatic explosion vents excavating maar craters. Faulting in the Otway Basin region controlled the spatial arrangement of these craters in a northwest-trending alignment of nested maars (Fig. 11.36). Ash, lapilli and bombs made up the bulk of ejecta that form layered deposits up to 8 km from source vents, exhibiting ample evidence of base surge blasts. The dimensions and volcano-sedimentary structures of the Blue Lake crater indicate that it is an example of an extreme end member of the phreatomagmatic explosion 'maar-diatreme' sequences of Moore (1967), Ollier (1967, 1974), Kokelaar (1986) and Lorenz (1986), with a moderately high explosive index. During this second stage of volcanism, a tephra volume of $\sim 1.3 \text{ km}^3$ was ejected. Mantle-derived Cr-diopside

1. Southeasterly aerial view of Mount Gambier volcanic complex, a series of fissure-controlled nested maars. Photo 014277.
2. Stacked lava flows and feeder dyke, Mount McIntyre Quarry. Photo 042549.
3. Crudely bedded basalt agglomerate and ballistic bombs, Mount Watch. Photo 014278.
4. Bedded phreatomagmatic surge and airfall tuff, Mount Gambier. Photo 042550.
5. Lherzolite xenoclast from Mount Gambier tephra. Photo 042551.



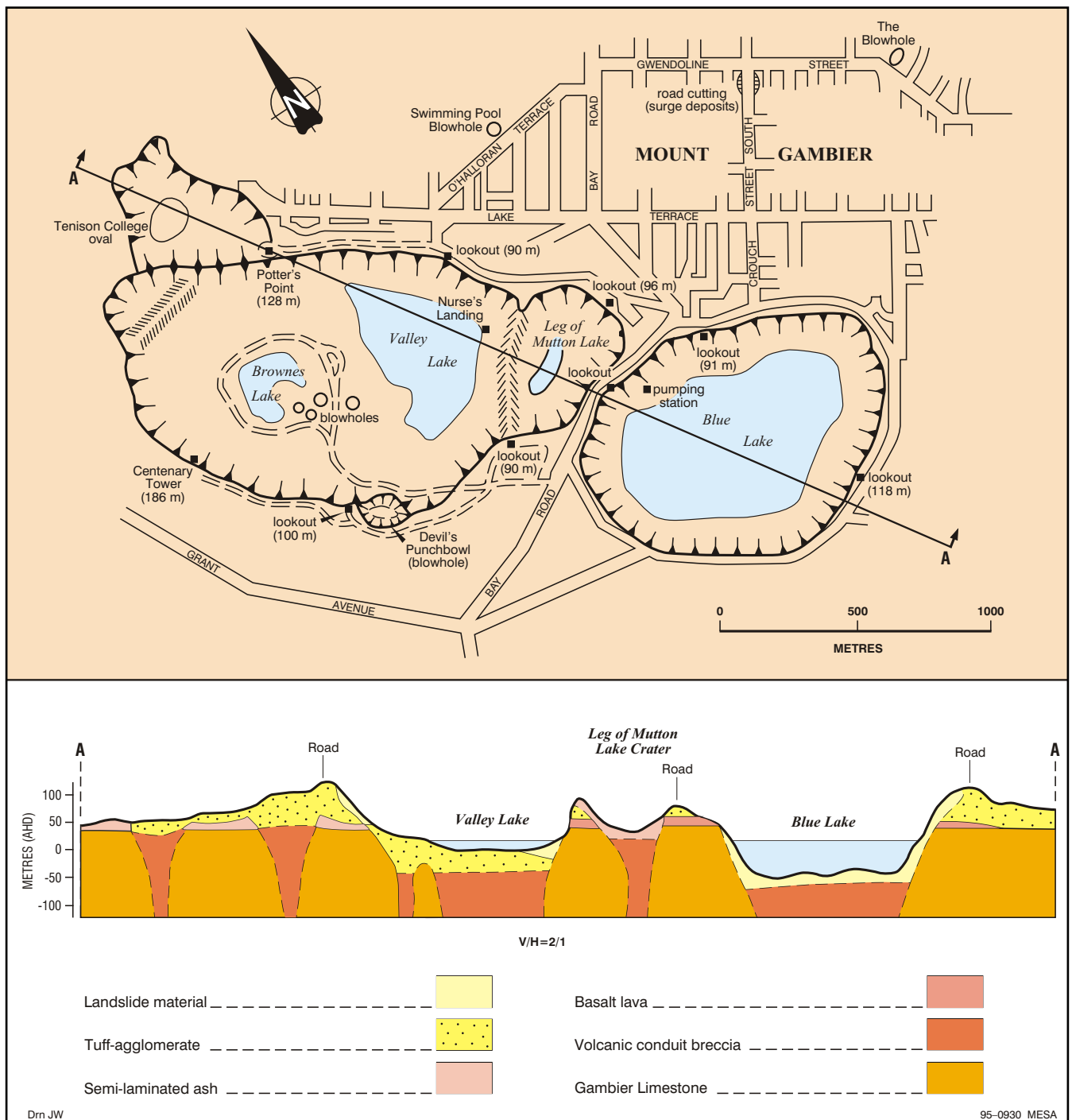


Fig. 11.36 Mount Gambier craters and diagrammatic cross-section (after Sheard, 1983).

Iherzolite xenoliths are relatively common in some horizons, especially those representing late phases of activity; they have coarse to fine-grained textures, some with gneissic fabrics. Non-igneous country rock fragments abound in the tephra, accounting for up to 20% of the volume. Activity climaxed with lava fountaining in the Brownes Lake crater and minor steam emissions from blowholes both within and outside the complex.

Mount Schank

This volcanic centre, ~9 km south of Mount Gambier, is an order of magnitude smaller in volume than Mount Gambier and would have rated a correspondingly smaller explosive index (Fig. 11.37).

Westerly aerial view of Mount Schank, a hybrid maar-cone complex. Photo 014282.



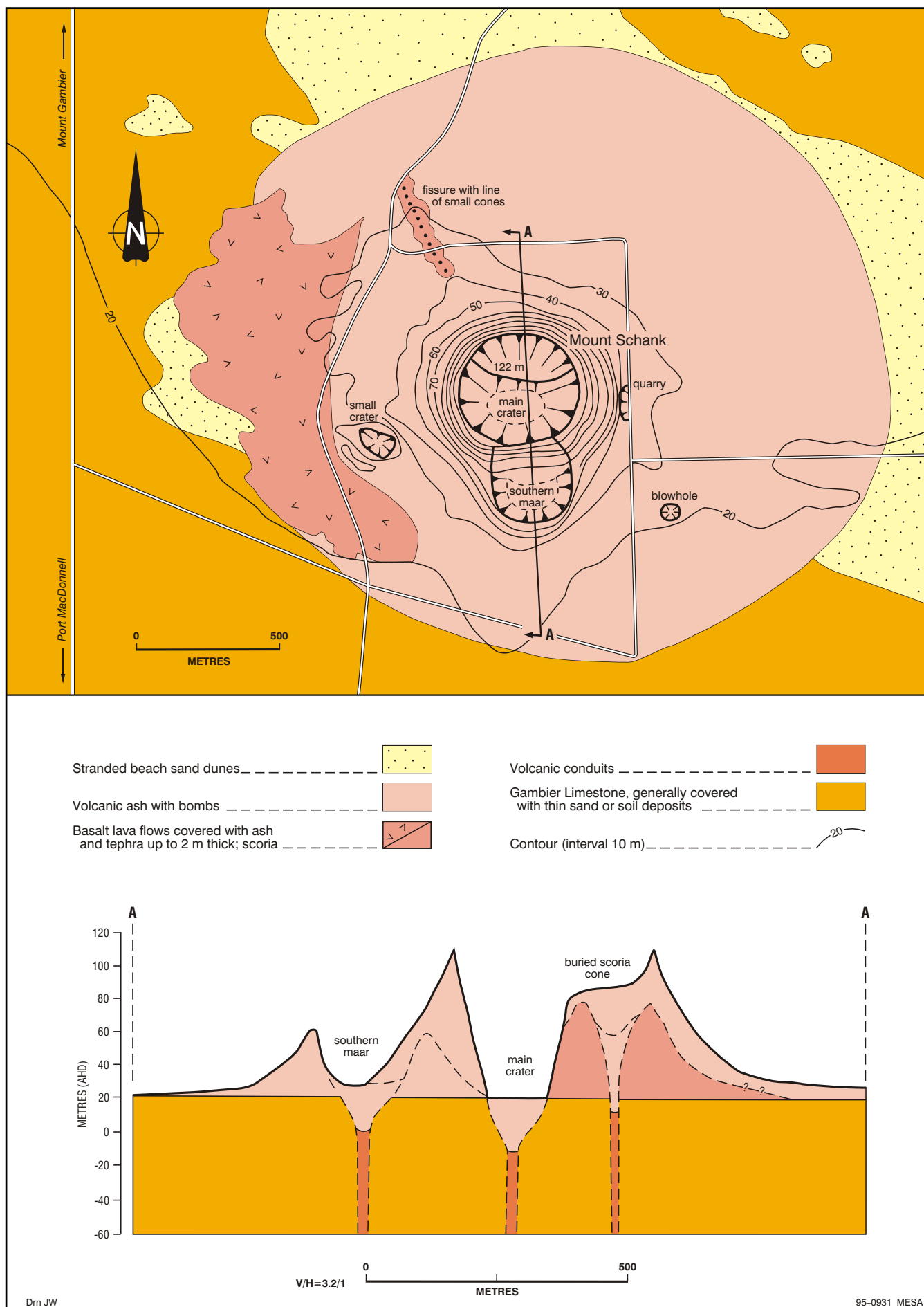


Fig. 11.37 Mount Schank geological map and diagrammatic cross-section.

Joyce (1975, 1985) classified this volcano as an hybrid maar-cone structure, formed chiefly by phreatomagmatic explosions. While palaeomagnetic measurements by Barbetti and Sheard (1981) indicated a range of possible ages from >7 000 to 1 000 years BP, they demonstrated that Mount Schank was not contemporaneous with Mount Gambier. TL dating by Smith and Prescott (1987) yielded an age of 4 930±540 years BP.

Eruptions at Mount Schank commenced from a 1.2 km long fissure in the Gambier Limestone. Ash was first vented as the ascending magma dyke degassed and possibly interacted briefly with groundwater. The magma supply rapidly increased, allowing lava to flow freely from the fissure while small scoria cones built up along its length (Fig. 11.37). This flow moved westwards and southwards to cover an area of ~300 000 m² with an average lava thickness of 3.5 m. Eruptive activity contracted to the centre and southern end of the fissure, where first a small scoria cone and then a small maar were constructed. These indicate more groundwater access to the magma at the southern end of the fissure. When activity from these two sites ceased, a final vent opened up between them, leading to the formation of the large main cone seen today. Major vent-clearing phreatomagmatic explosions during this eruption resulted in a broad cone with both maar and cone crater characteristics.

Volcanic gases

In 1966, two deep wells, Kalangadoo 1 and Caroline 1, were drilled within the volcanic province (Fig. 11.34). Both intersected CO₂ fluid at depths of 2.1 and 2.5 km, respectively. Caroline 1 remains a commercial producer of liquid CO₂.

Wopfner and Thornton (1971) considered that the CO₂ has a volcanic origin, based on the proximity of Holocene volcanism and absence of carbonate rocks at and below the reservoirs. More recent isotope work by Chivas *et al.* (1987), Caffee *et al.* (1988) and Giggenbach *et al.* (1991) has reinforced this view. Their analyses of $\delta^{13}\text{C}$, $^3\text{He}/^4\text{He}$ ratios and the concentrations of trace noble gases ^{20}Ne , ^{22}Ne and excess ^{129}Xe from Caroline 1 collectively indicate a distinctly primitive mantle source rather than crustal or organic sources. Caroline 1 yields the most primitive Ne and Xe isotopic gas signatures so far tested worldwide.

VOLCANIC ACTIVITY

No modern eruptions have been observed although Aboriginal legend indicates human witness to such events in this province (Smith, 1880). Sprigg (1959b) linked two strong earthquakes centred on Kingston in 1897 and Robe in 1948 with presumed 'submarine lava flows' offshore from Beachport. Such a link has yet to be confirmed by sampling and palaeomagnetic measurements.

The volcanic province is seismically active and has undergone geologically recent upwarp (Murray-Wallace and Belperio, 1991). Thus, the province can be considered to be dormant rather than extinct. Using available age data and eruption probability equations developed in New Zealand (Dibble *et al.*, 1985), the volcanic hazard is <2% for an eruption in the next 100 years.

Monitoring of seismicity is via a single seismometer near Mount Gambier with three further stations located at Naracoorte and Willalooka, 100 and 160 km to the north, respectively, and within the Grampians in Victoria, 165 km to the east. This current level of monitoring is inadequate to detect small tremors emanating from a volcanic source and, at best, only magnitude and approximate location of moderate seismic events can be calculated.

Prior warning of an eruption may range from hours to days and possibly weeks and, given an appropriate monitoring network, would consist of on-going local seismic activity and localised ground deformation. A short time before an eruption, hot gases, smoke and/or an increase in groundwater temperature may be observed. Future eruptions are expected to be rare, isolated and

new features rather than the re-activation of existing volcanoes. A random probability as to location is expected within the province and major population centres cannot be excluded.

Examples of volcanic hazard potential for the various eruptive types known from a 5 000 km² area of the South-East are presented in Table 11.4 using criteria and methods developed by Dibble *et al.* (1985).

SEISMICITY, NEOTECTONICS AND EARTHQUAKE RISK

D.N. Love, W.V. Preiss and A.P. Belperio

Earthquakes occur throughout the world on a daily basis, clearly demonstrating the dynamic nature of the Earth's crust. Larger earthquakes cause significant ground vibration and, if they strike built-up areas, constitute a major geological hazard. Earthquakes may cause physical rupture at the surface, or movement along pre-existing faults. As for volcanic eruptions, historical records provide the chief means of predicting seismic characteristics and risk but, as with volcanoes, it is difficult to determine if faults are active, inactive or dormant.

Because of its intraplate setting, South Australia is relatively stable in terms of global seismicity, with lower maximum magnitudes (maximum reported is 6.5) and attenuation of vibration, broader distribution, and much shallower foci (at depths of 0–40 km, mostly <10 km) than for plate margins. As with earthquakes of other stable regions, the causes are not well understood. In general, the older more stable basement of Western Australia appears able to accumulate more stress than the Palaeozoic mobile belts of eastern Australia.

Historical seismicity

Numerous but mostly minor seismic events have been recorded in South Australia since the first earthquake (28 July 1837) after European settlement (McCue, 1975). In 1897, South Australia's largest known earthquake, centred offshore from Beachport and with a magnitude of ~6.5 and an estimated Modified Mercalli intensity of VIII, caused considerable damage in coastal towns in the South-East, with some reports of liquefaction of poorly consolidated sand. A few chimneys were damaged in Adelaide and the event was felt as far away as Port Augusta and Melbourne. Further events over magnitude 5 were recorded offshore in 1899 and 1948, but damage was minimal. Another large earthquake in 1902 (magnitude estimated at ~6) has traditionally been placed at Warooka but considerable damage also occurred at Yankalilla, Willunga and Yorketown; Malpas (1991) suggested an epicentre in Gulf St Vincent. Variable damage may be attributed to variable thickness of unconsolidated sediments. On the basis of damage reports, the epicentre of the Adelaide earthquake of 1954 (magnitude 5.5) was estimated as being on the Eden-Burnside Fault system at Darlington but, with no local instruments in operation at the time, this is uncertain.

Instrumental monitoring

The first seismograph in South Australia, installed at the Adelaide Astronomical Observatory in 1909, was a Milne horizontal pendulum, a long-period, low-gain instrument not well suited to the study of local seismicity although many distant events were recorded. In 1923, a more sensitive Milne-Shaw device was installed, but this was still not well suited to recording local seismicity. In 1958, a short-period three-axis instrument with photographic recording was established at Mount Bonython. By 1963, two further instruments (photographic recording, short period, vertical) were established at Hallett and Cleve so that local events down to magnitude 1.5 could sometimes be located.

The State seismograph network was developed by the University of Adelaide from 1958 to 1980 and is now maintained by MESA. Details of the development of monitoring in South Australia are

Table 11.4 Volcanic hazard potential, South-East volcanic province.

	Phreatic explosion	Phreatic tephra	Composite cone	Scoria cone	Lava flow
Type example	Mt Gambier	Mt Schank (see also lava)	Mt Graham	Mt Watch	Mt McIntyre
Active area (km ²)	200	200	600	60	60
Ejected volume (km ³)	1.4	0.2	0.7	<0.2	0.8
Area destroyed (km ²)	50	1	10	<0.5	2-5
Range of damage (km)	1-5	<1	2	<0.5	1-5
Velocity of destruction (m/sec)	>100*	100*	50	<1	<1
Area damaged (km ²)	65	2	15	1.56	10
Range of damage (km)	4-8	1-2	2-4	1	1-5
Velocity of damage	ballistic	ballistic	ballistic	<1 m/sec	<1 m/sec
Deposit temperature	<100° to incandescent	<100° to incandescent	incandescent	incandescent	incandescent
Incidence	Rare	Rare	— Rare and pre-date Aboriginal occupation —		
Pre-settlement	5	2	7	3	Many
1840-present day	None	None	None	None	None
Number of events	5	2	7	3	Many
Return period (years)	4 000-10 000	4 000-10 000	>10 000	>10 000	>10 000
Last event (years BP)	4 400	4 500-8 000	>10 000	>100 000	>100 000
Probability in 100 years	2.2%	2.2%	<1%	1%	1%
Value at risk					
Population**	5 000-25 000	10-1 000	100-3 000	0-1 000	0-500
Property** (\$million)	<1 to hundreds	<1 to tens	<1 to hundreds	<1 to tens	<1 to tens
Stock numbers	Low	Low	Low	Low	Low
Utilities					
The cities of Mount Gambier and Millicent; the towns of Mount Burr, Kalangadoo, Tarpeena and Glencoe; State forests, saw mills, paper mills and tourist facilities.					
Mitigation					
Surveillance	— Two seismometers near Mount Gambier and Naracoorte; telemetered to Mount Gambier; visual inspection of record —				
Warning signs	— No historical experience; seismicity and ?deformation may precede an eruption —				
Zoning	Not possible	Not possible	Not possible	Not possible	Not possible
Evacuation	Prior essential	Prior essential	Progressive	Progressive	Progressive
Rescue	Dig	Dig	Dig	As for fire	As for fire
Medical	Violence, suffocation, burns	Violence, suffocation, burns	Violence, burns	Burns	Burns
Property removal	Prior only	Prior only	Prior only	Progressive	Progressive
Prevention	None	None	None	None	Dams, water cooling, diversion

* Base surge blasts

** The lesser figures have no chance; the greater figures require action to survive.

covered by Parham *et al.* (1988), McCue (1975) and Doyle and Underwood (1965). By 1980, there were 12 stations, giving improved coverage of the northern Flinders Ranges and the South-East. The first permanent, digitally recording, three-axis seismograph was installed near Naracoorte in 1989. By 1990, the network of 17 stations included four digital recorders, situated mostly south and east of Adelaide. Installations outside the State but important for locating earthquakes within South Australia include Alice Springs, Northern Territory (1970), Bellfield, Victoria (1972), Stephens Creek, New South Wales (1974) and Warburton, Western Australia (1978).

Although in the central Flinders Ranges events down to magnitude 0.5 can occasionally be located, not all events up to magnitude 3 can be located in the Far West of the State. Similarly, the spatial accuracy of located events is quite good in the central area of the Flinders and Mount Lofty Ranges, sometimes down to <2 km in the horizontal direction, but far western and northwestern events may have location errors of 50 km, depending on the distribution of recording stations.

The magnitudes assigned to some large early events are calculated from intensity reports and attenuation relations, resulting in variations between authors (McCue, 1975; McDougall, 1990; Malpas, 1991). For most of the Flinders and Mount Lofty Ranges

the recorded data are complete down to magnitude 3.5 since 1969 and down to magnitude 2.5 since 1973. From 1986 to 1991, ~300 events per year have been located by the network. Of these, an average of only 10 per year exceed magnitude 3 and one exceeds magnitude 4. Only five earthquakes over magnitude 5 are known from the Flinders and Mount Lofty Ranges and environs — Quorn (1887) 5.2, ?Warooka (1902) 6.0, Jamestown (1921) 5.1, Motpena (1939) 5.7, Adelaide (1954) 5.5 and Wilpena (1972) 5.3.

With the large distances between some events and seismographs in Australia, and the low attenuation, the Richter magnitude formula has been shown to cause systematic errors in South Australia (Greenhalgh and Singh, 1986).

The depth-velocity model used for most epicentre determinations is a single layer with the Moho at 38 km. One-way refraction measurements throughout the State, summarised by Greenhalgh *et al.* (1989), show Moho depths varying from 34 to 40 km, and a mid-crustal interface at 14 to 25 km. Shackleford and Sutton (1981) suggested a possible mid-crustal interface on Eyre Peninsula at <10 km.

Seismicity and neotectonics

Regions of seismicity are related to the tectonic framework of

South Australia, the greatest concentrations and magnitudes of seismic events occurring in tectonic provinces such as the Delamerian fold belt, the Palaeoproterozoic mobile belt (Cleve Submain) of eastern Eyre Peninsula, and the Musgrave Block. More than 75% of earthquakes have been recorded in the Flinders and Mount Lofty Ranges, but activity is quite variable and there is a gap in activity between Eyre Peninsula and the Flinders Ranges. Wellman and Greenhalgh (1988) suggested a partial correlation with elevation, but elevation and seismicity probably both result from reactivation of ancient tectonic structures under the stress regime that has been operating on the Australian continent during the Cainozoic. Focal mechanism studies (e.g. McCue and Sutton, 1979; Denham *et al.*, 1981; Greenhalgh and Singh, 1988) indicate that southeastern Australia is undergoing compression along axes of variable orientation, but mostly with an east-west component.

Although no comprehensive neotectonic analysis of the State has been carried out, observations on structural features that have been active in the recent past are consistent with such a compressive stress regime. The most youthful faults are believed to be those that bound the western and eastern flanks of the Flinders and Mount Lofty Ranges. There are very few sites at which such faults are exposed, most having been buried by Quaternary alluvial fans showing little if any evidence of internal displacement. Much of the uplift may therefore have occurred during the Tertiary. However, at Wilkatana, north of Port Augusta, Williams (1973b) recorded Adelaidean bedrock thrust westwards over Pooraka Formation alluvial sediments on an east-dipping reverse fault, suggesting movement at about 30 000 years BP. Near Cambrai on the eastern margin of the Mount Lofty Ranges, Kanmantoo Group metasediments have been thrust eastwards over Pleistocene conglomerate, indicating reactivation of the Delamerian Milendella Fault during the late Cainozoic, involving 60-90 m of uplift on the ranges since the Miocene (Bourman and Lindsay, 1989). A similar, west-dipping reverse fault between Burra Group siltstone and late Cainozoic gravel is exposed at the Mittopitta Goldfield in the Mid-North. In the northern Flinders Ranges, the east-dipping Mount Babbage Thrust has displaced Mesoproterozoic granitic rocks over Late Pleistocene to early Holocene colluvium. High-resolution seismic profiling north of Mount Hopeless revealed the Blanche Lineament to be a south-dipping thrust fault with displacements recorded in Tertiary and Quaternary sediments (Sheard and Cockshell, 1992).

Near Adelaide, there is no obvious displacement of Quaternary sediments on the Para and Eden-Burnside Faults, and Belperio (1992) used the absence of displacement in surficial strata offshore to infer a lack of significant fault movement over the



Top: Aerial view of the Eden-Burnside Fault scarp and the eastern suburbs of Adelaide in 1987. Stonyfell and Greenhill Quarries are adjacent to the scarp in the lower centre of view. North is to the top. Photo 034225.

Bottom: The Eden-Burnside Fault exposed at Highbury. Photo 015834.

Below: Steep reverse fault near Parabarana Hill. Photo 042552.



past 30 000 years. Nevertheless, inland from Sellicks Beach, the Cambrian Heatherdale Shale is faulted against Quaternary clay and fanglomerate at the only known exposure of the Willunga Fault (Campana *et al.*, 1953; May and Bourman, 1984). At other sites on the western side of the Mount Lofty Ranges, relationships are less clear. For example, at the Highbury sand quarry, Moana Beach and Sellicks Beach, Tertiary St Vincent Basin sediments are preserved in unconformable relationship with bedrock; the unconformities are folded and locally steeply dipping to the west or northwest (e.g. Glaessner, 1953). These relationships may indicate the existence of east or southeast-dipping reverse faults in the Adelaidean-Cambrian bedrock, perhaps expressed largely as monoclines in Tertiary sediments. Together with the known current stress regime, they argue against an extensional origin for the St Vincent Basin; rather, the depression of the basin and concomitant uplift of the Mount Lofty Ranges during the Cainozoic are best explained by compressional reactivation of Delamerian reverse faults. Many of these dip eastwards near the western margin of the ranges, and westwards near the eastern margin (see Ch. 7). A horst-and-graben extensional model for the region could not be accommodated by reactivation of ancient structures, and would require the development of new, basinward-dipping faults during the Cainozoic, for which there is no evidence.

Flinders and Mount Lofty Ranges

Most seismic events have been recorded between latitudes 30 and 34°S, but there is also a correlation with lineament, gravity and aeromagnetic patterns (Fig. 11.38), and with the region of the Flinders Ranges in which diapirs are known. South and west of the G2 structural corridor there is a marked drop in activity although some events are recorded as far as the western end of Kangaroo Island. It is suggested that this relatively more stable region corresponds to the belt of shallower basement (as indicated by higher total magnetic intensity, Fig. 11.38). In contrast, the area east and north of the G2 corridor was deeply subsident during early Adelaidean rifting, suggesting significant extension of the pre-Adelaidean basement, and accumulated the thickest Callanna Group (associated with evaporite deposition and subsequent diapirism; see Vol. 1, Ch. 6) and Burra Group. It is suggested that modern compressive stresses are being preferentially relieved within this zone of attenuated and weakened continental crust.

Events occur as far northwest as Oodnadatta, although the capability of the network is limited in this area, and there may be undetected events extending as far as the Musgrave Block. Few events have been recorded in the Arkaroola area or Olary Ranges; their scarcity in the northeastern Flinders is surprising in view of the elevated and youthful topography of the region.

Seven focal mechanisms have been calculated within the Flinders and Mount Lofty Ranges region (Greenhalgh *et al.*, 1986, 1994; Fig. 11.38). Most show strike-slip motion on steeply dipping faults, the predominant direction of stress being east-northeast-west-northwest compression, except for the Carrieton earthquake with a northwest-southeast stress direction. However, the Moralana earthquakes of 1992 indicated thrusting along moderately dipping faults. Swarms have occasionally been noted in the region but, in general, events are broadly spread and temporally isolated, with larger events sometimes having foreshocks or aftershocks.

The area east of Port Pirie and Port Augusta has the most accurately located events, but the distribution cannot be directly related to the main mapped surface faults. This may be a result of problems with velocity model errors, single (vertical) component recording, and inaccuracies inherent in analogue instruments. Alternatively, the epicentres may be real and reflect stress release throughout a large volume, not principally in the planes of mapped faults. The gap in activity between this area and Eyre Peninsula corresponds to the Torrens Hinge Zone, an area of only slight crustal attenuation during early Adelaidean rifting (see Vol. 1, Ch. 6).

Simpson Desert

Large earthquakes in the Simpson Desert between 1937 and 1941, and in 1972, were mostly north of the South Australia-Northern Territory border, although the location threshold in this area is very poor (around magnitude 3), as is the location accuracy. The larger events appear to be sporadic, unlike those of the ranges; no fault scarps related to these events have been positively identified.

Musgrave Block

In 1986, a surprisingly large earthquake of magnitude 6.0 occurred in the Musgrave Block, an area previously considered inactive, with a small number of aftershocks, one of magnitude 5.5 about four months later (McCue *et al.*, 1987). The main shock caused a fault scarp ~13 km long, the first recorded since settlement in the State and one of only five in Australia.

The focal mechanism revealed nearly horizontal compression in a northeast-southwest direction. The nodal planes (east-west and north-south) closely match the directions of the two arms of the arcuate scarp. The east-west arm falls within the Marryat Creek Fault zone, and the north-south arm coincides with other local trends, although no pre-existing fault had been mapped at the site. Maximum displacements were 0.8 m left-lateral movement on the east-west arm and 0.6 m vertically.

Eyre Peninsula

There is a diffuse band of regular activity, though not as great as that of the Flinders Ranges, along eastern Eyre Peninsula, coinciding with arcuate structural trends. The area includes a number of swarms, some of which have been felt but barely recorded on the nearest instrument at Cleve, indicating extremely shallow foci.

Continental shelf

A number of large events close offshore in the South-East where the continental shelf is narrowest may represent an isolated zone of activity or may be associated with the continental margin. A broad band of epicentres from the Great Australian Bight through to western Victoria includes another notable magnitude 5.3 event near Warrnambool in 1903 (McCue, 1978). Sprigg (1959b) postulated a connection between the two events (1897 and 1948) recorded near Kingston and Robe and the offshore occurrence of presumed submarine lava flows. Other authors have indicated a possible link between seismicity in the South-East and the volcanic province centred on Mount Burr and Mount Gambier (Sheard, 1983).

Earthquake risk

The maximum magnitude of earthquake that could occur in South Australia is uncertain. The largest known magnitude is 6.5, although events up to magnitude 7 have been recorded from elsewhere in Australia. The largest known intraplate events, with magnitudes as high as 8.3, have occurred in rifts (Johnston and Kanter, 1990), but the State does not appear to contain any active rift zones. In the Flinders Ranges, the most seismically active region, the presence of diapiric material at depth may limit the amount of stress that can be accumulated to produce a large event. The expected return period for a magnitude 6 event is 112 years (Greenhalgh *et al.*, 1986), although three events of this magnitude have been recorded in South Australia during the last 100 years.

The first seismic risk study focusing particularly on South Australia used data to 1973 (McCue, 1975). This study, in conjunction with McEwin *et al.* (1975), was the basis for the risk map in the first earthquake code (AS 2121-1979; Standards Association of Australia, 1979). A subsequent study by Stewart (1984) used data up to 1979. Gaull *et al.* (1990) published a nationwide risk assessment based on pre-1985 data. Greenhalgh and McDougall (1990), McDougall (1990) and Malpas (1991) used

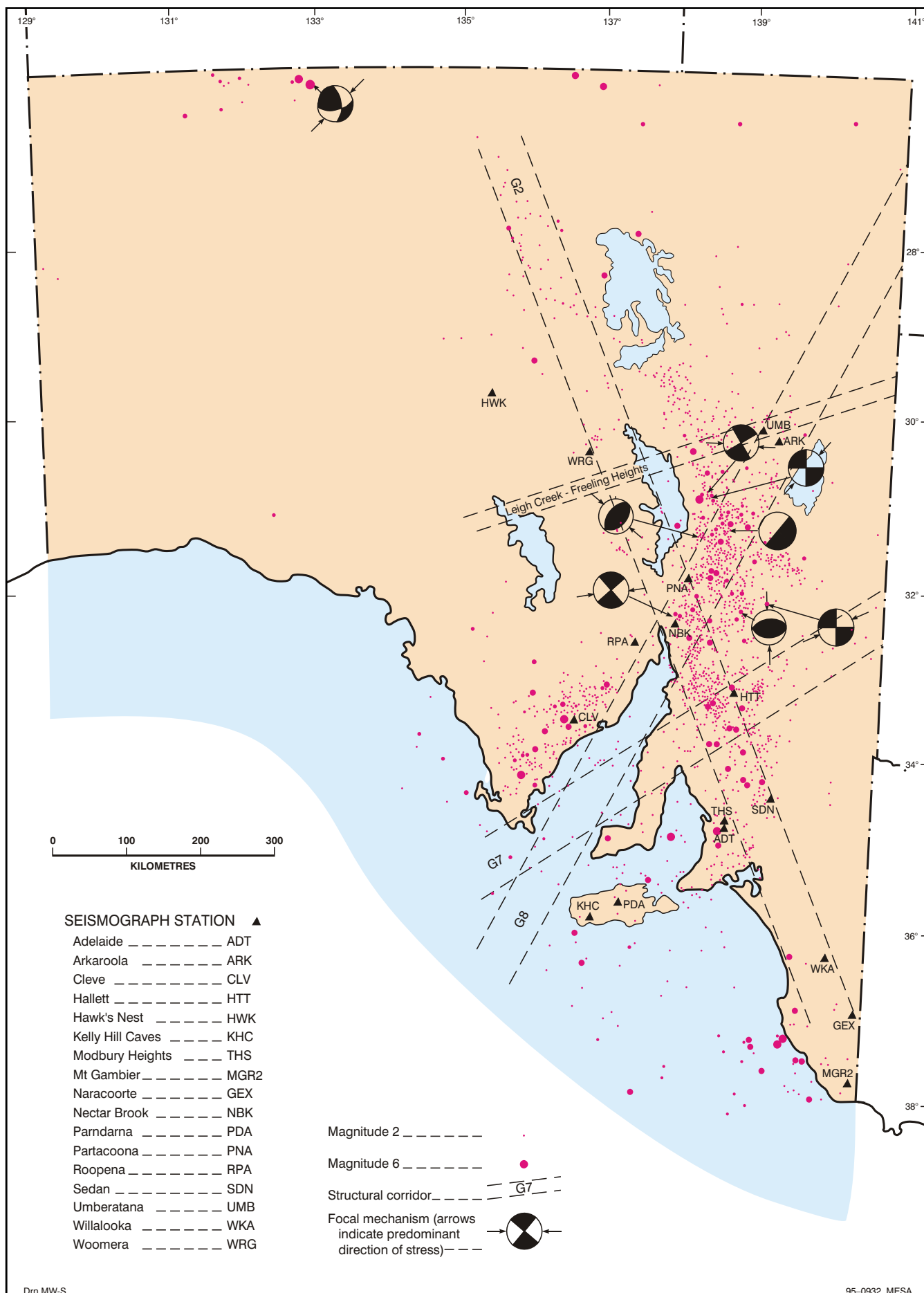


Fig. 11.38 South Australian seismicity and relationship to known structural corridors.

more recent data, but the latter two studies also used rediscovered records of some old events not previously considered.

Without a suitable predictive mechanism, the only methods available to produce risk estimates are based on the assumption that the next few decades are likely to be similar to the last century or two. This has been shown to be a fair assumption, but does not handle the occurrence of a few large earthquakes in unexpected places. Risk studies also rely on other factors which require some subjective judgement.

The main risk methodology used throughout the world, the Cornell-McGuire method, tends to produce a regionally smoothed map; Gaull *et al.* (1990) applied it to encompass an area from Kangaroo Island to Oodnadatta in a single zone. On the other hand, Greenhalgh and McDougall (1990) used the seismic moment method, which tends to produce bullseyes around previous large, isolated events and is therefore less widely favoured. However, for a long zone of variable activity such as the Flinders and Mount Lofty Ranges, it shows possibly significant variations. Stewart (1984) provided a useful comparison of the two methods. The region between latitudes 32°30' and 34°S has a greater ratio of small to large earthquakes than other parts of the State. This distinction is lost with the regional smoothing by Gaull *et al.* (1990). This results in a lower risk than might be expected, given the large number of events in the region.

It is instructive to compare the effect of various factors on the risk in population centres such as Adelaide and Port Augusta. Adelaide lies on the boundary of one of the zones of Gaull *et al.* (1990), with a return period of 200 years for an intensity VI earthquake. In contrast, the McDougall (1990) study shows Adelaide on a bullseye caused by the 1954 event but, if the 1902 Warooka event was actually centred closer to Adelaide and the two are related in some way, then this is particularly relevant to the bullseye. McDougall estimated a return period of 70 years for intensity VI using an attenuation function derived from only South Australian data, or 200 years using attenuation from all Australian data. The function used by Gaull *et al.* is similar to that of McDougall for South Australian attenuation near the epicentre, but decreases more rapidly with distance. The difficulty with using the South Australian function is that the data are limited to ~30 events, mostly small, so that the reliability is poor. Collection of more strong-motion data will improve the attenuation function and risk estimates for Adelaide. Port Augusta is just outside a zone on the map of Gaull *et al.*, and use of acceleration and acceleration attenuation data would suggest that Port Augusta is outside the area of risk. However, use of velocity and velocity attenuation data indicates that the city lies within an area of reasonable risk, but that this risk decreases rapidly to the west. Although acceleration is traditionally used by the engineering profession, velocity appears to be more closely related to degree of damage.

It can be concluded that the risk of a large earthquake in any part of South Australia is not high, but cannot be ignored. The State's history of seismological observations is too short in comparison to the expected return period for large events.

LANDSLIDES

S. Walker

The term landslide encompasses almost all forms of mass movement of geological materials. The many types and classifications of landslides vary from rapid to slow moving, rock to soil, wet to dry, and in many other characteristics (Varnes, 1978; Selby, 1980). Although landslides have been occurring naturally throughout geological history, those which are of significance to humankind are generally exacerbated, accelerated, or indeed caused by human activity. These activities include changes to land use, alteration to drainage, water infiltration rates and erosion. In particular, many modern landslides can be directly related to vegetation clearance since settlement (Bourman and Harvey, 1986).

Often, heavy rainfall is the trigger that initiates mass movement.

Most of the main types of landslides which have occurred in South Australia have been previously described (e.g. Kerr-Grant, 1956; Van Deur, 1978; Selby, 1980; May and Bourman, 1984; Proudman and Morris, 1990). They include:

- debris slides associated with steeper slopes of the Flinders and Mount Lofty Ranges
- earth flows associated with late Palaeozoic glaciogene sediments of Fleurieu Peninsula
- rotational slumping, particularly in coastal cliffs at Aldinga and Kingscote
- rockslides, rockfalls and rock topples
- mudslides.

Debris flows, the most common type of landslide, are generally restricted to slopes steeper than 1 in 4 (14°). Earth flows occur on flatter slopes, but in South Australia these are generally restricted to regions of exposed glaciogene sediments on Fleurieu Peninsula. The steeper parts of the Flinders and Mount Lofty Ranges consist of strongly bedded and jointed or cleaved rock, and afford many examples of rockslides due to slab and wedge failure.

The significance of landslides in South Australia, both economically and in terms of safety hazard, has been small, and there are no known related instances of death or serious injury. However, several residential dwellings have been destroyed and a number seriously damaged, and there have been a number of recorded instances of damage to roads. One large mudslide was possibly initiated by the 1954 earthquake in Adelaide and finally triggered by heavy rain one month later, resulting in a 2 km flow down Brown Hill Creek valley (Kerr-Grant, 1956).

The overall trend is towards greater activity, and this is likely to continue. There are a number of reasons for this increase. As areas



Little Para Dam. Talus slide at (20 degrees slope). Photo 012731

close to Adelaide become more built up, there is increasing pressure for development on or adjacent to steeper, more landslide-prone slopes. This is accompanied by greater need for roads, facilities and infrastructure, all of which gradually encroach on more susceptible land. The risk posed by landslides is recognised by the State Planning Authority which requests a geological appraisal by MESA of residential subdivisions on slopes steeper than 1 in 4. With proper landslide risk assessment incorporated into planning and development policies, these risks can continue to be kept to an acceptably low level.

PROBLEM SOILS

M.J. Sheard and A.P. Belperio

Land suitable for building and construction is a resource that is constrained by its geological and geotechnical properties. In South Australia there are a number of unique problems related to the physical and chemical properties of the soils. These include saline or gypseous soils, and expansive, collapsing and compressible soils. Ideally, these soils should be removed and replaced with more suitable compacted fill, or alternative suitable founding horizons at depth may be utilised.

Many South Australian soils are sodic or gypseous (Northcote and Skene, 1972) and these promote salt damp reactions in masonry. Unchecked and with sufficient time, this condition can lead to wall failure. Salt damp can be prevented by installing a damp proof barrier between walls and footings.

Expansive clays and soils are common around Adelaide, including Keswick Clay, black earth and some red-brown earth soils. These materials typically expand 15-25% upon wetting from a dry condition (Sheard and Bowman, in prep.). Gilgai are surface expressions of unequal differential movements in these reactive soils. Light-weight buildings such as houses often suffer cracking as a result of such soil movement and present a significant maintenance cost to home owners. Methods used to minimise damage include removing the expansive soil or constructing deep pile footings to below the reactive zone.

Collapsing soils are common on fine calcareous wind-blown silts. Because of their low density, these soils lose strength and may fail dramatically when loaded and made wet. Soft and compressible peaty soils are associated with coastal wetlands and reclaimed swamps. These soils undergo long-term settlement under loading, or can result in acid drainage where disturbed or the watertable is lowered. Many coastal peats, organic muds and organic sands contain colloidal pyrite, sulphate-reducing bacteria and organic acids that can lead to production of acid sulphate soils which are corrosive to concrete and steel (Bowman, 1993; Sheard and Bowman, in prep.). The West Lakes development in metropolitan Adelaide is one example where remedial engineering works were necessary to remove such hazards prior to building. Where thin

and shallow, compressible estuarine sediments were removed and replaced by compacted sand fill. In areas where the material was too thick to remove, an overcompaction and settlement monitoring technique was used. Large structures on coastal sediments, such as the Torrens Island Power Station, rest on piles driven into suitable material below more than 10 m of unconsolidated sediments.

Much of the low-lying land around Port Adelaide is underlain by unconsolidated and compressible estuarine sediments similar to those of West Lakes that similarly require remedial engineering work (Belperio and Rice, 1989). The subsurface sediments comprise a range of sand, mud and peat peritidal facies that reflect progressive shoaling and coastal progradation of a former marine embayment (Fig. 11.20). Specific investigations are necessary to identify zones prone to compaction, acid sulphate soil generation, resonant amplification of seismic shock waves and liquefaction, and appropriate remedial techniques applied to reduce these hazards.

Contaminated land is largely a legacy of past industrial activity and waste disposal practices. Where localised, contaminated soils may be remedied by removal or by suitable isolation. However, the contamination may become more extensive when, for example, pollutants or leachates mix with groundwater. This is a particular problem in karst areas such as the Mount Gambier region, where the regional watertable is a sink for most of the surficial runoff.

Regions of karst development also present significant engineering problems in addition to potential pollution. Engineering complications in foundation design have been reported by Allchurch (1967) and Selby and Lindsay (1982) for the calcareous Hallett Cove Sandstone in the City of Adelaide. Other examples are grain silo siting at Millicent (Selby and Smith, 1978) and building siting at Mount Gambier (Pilkington, 1978). Of a more anecdotal nature is a collapse in the Glencoe West Oval during a football match (Nelson, 1971). Beal and Coates (1987) investigated the roof stability of the Victoria Fossil Cave at Naracoorte and remedial work was performed to prevent roof collapse, thereby allowing continued tourist access.

THE COASTAL ZONE

A.P. Belperio

Land versus sea-level change

The sediments of the coastal zone contain a record of past sea levels, climate and rates of change over time scales of hundreds to thousands of years. Understanding and quantifying these changes is important for understanding present and predicted patterns of sea-level behaviour, inundation and coastal erosion. From age and height attributes of palaeoshorelines, it is possible to determine how much the sea level has risen or fallen relative to the adjacent land surface, and how much of this is due to global climatic change, crustal deformation or local subsidence of the land surface. Neotectonic deformation calculated from the geological record, and human-induced land-level changes derived from geodetic surveys, are necessary correction factors for the secular sea-level trends recorded by tide gauges (Belperio, 1989, 1993).

The local geological record contains abundant evidence of repetitive marine transgressions and regressions across the continental shelf and into the gulfs (as discussed previously). The oldest Quaternary marine sediments, the Point Ellen Formation and Burnham Limestone, indicate contrasting degrees of uplift of Fleurieu Peninsula and Kangaroo Island. Thin marine remnants occur at an elevation of 5-10 m above present sea level from Kingston Park to Sellicks Beach, and on northern and southern Kangaroo Island. This contrasts with an elevation of 50 m above present sea level at Cape Jervis and 70 m below present sea level within the St Vincent Basin.

The last interglacial shoreline facies (Glanville Formation) provides a laterally more persistent record that has been used as a benchmark measure of neotectonic deformation around Australia



Several forms of gilgai exposed in an excavation at Burnside, an eastern suburb of Adelaide. Photo 042553.

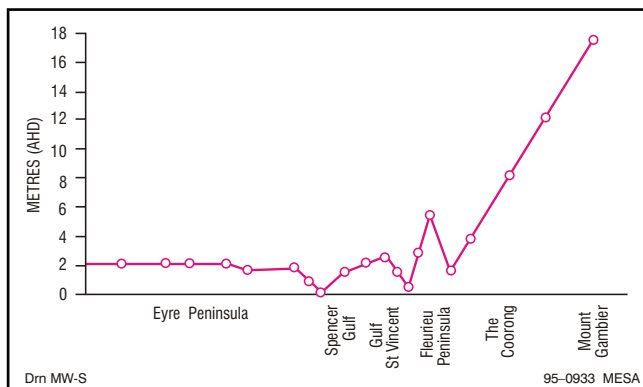


Fig. 11.39 Elevation of the 125 000 years BP (last interglacial) shoreline around the South Australian coast relative to present sea level.

(Murray-Wallace and Belperio, 1991). In South Australia, this shoreline varies systematically in elevation between geological provinces (Fig. 11.39). Along the Eyre Peninsula coastline, a consistent elevation of +2 m above present sea level for this shoreline is ubiquitous, reflecting tectonic stability of this Precambrian craton. At the other extreme, last interglacial shoreline sediments increase in elevation from +3 to +18 m above present sea level along the South-East coastal plain, reflecting uplift and tilting associated with Pleistocene and Holocene volcanic activity in the vicinity of Mount Gambier.

The greater resolution afforded by radiocarbon dating allows a more detailed record of sea-level change to be extracted from Holocene coastal sediments. In addition to global sea-level changes, it has been possible to deduce significant regional and local differences attributable to isostatic and neotectonic processes (Belperio *et al.*, 1983; Nakada and Lambeck, 1989). For South Australia, this has resulted in coastal regression over the past 6 000 years that is most pronounced (3-5 m) in the northern reaches of Gulf St Vincent and Spencer Gulf (Fig. 11.13). The magnitude of this apparent sea-level highstand decreases progressively down the gulfs towards the continental shelf.

In contrast to the rest of the State's coastline, the coastal facies of the Port River estuary indicate a recent rise of sea level in this region. This is corroborated by tide gauge data that indicate a relative sea-level rise of ~100 mm over the past 40 years. Human interference is a major factor that must be taken into account in explaining the rise. Much of Adelaide's western and northwestern suburbs have been constructed on estuarine and coastal sediments (Aitchison *et al.*, 1954). Over the past 150 years, large areas of former tidal wetlands have been drained and reclaimed, and levees protect low-lying suburban and industrial areas from marine inundation. Drainage and loading have led inexorably to compaction of unconsolidated coastal strata, with a subsidiary contribution from excess groundwater withdrawal.

Thus, three quarters of the secular rise of sea level of 2.8 mm/year indicated by the tide gauge at Port Adelaide is attributable to local land subsidence (Belperio, 1993). Many of the world's cities and their attendant ports developed on coastal lowlands similarly suffer from ground subsidence and rising relative sea levels. In most cases, drainage and reclamation of wetlands and groundwater withdrawal are the principal contributing factors (Carbognin, 1985).

Greenhouse effect

The 'greenhouse effect' is a climatic scenario that predicts a sea-level rise as a result of global warming caused by increasing pollution of the atmosphere. Although increasing atmospheric pollution is unequivocal, its effect on the climate and sea level is still not clear. A variety of global predictions have been made based largely on computer models of atmospheric circulation and climate. Because sea level has been rising at many coastal

locations around the world, it is argued that the greenhouse effect is already operating. A commonly quoted estimate of a 100-150 mm global rise over the past century based on historical measurements of tide heights from ports and harbours is now believed to be an over-estimate, possibly by as much as two or three times (Pirazzoli, 1989). Tide gauges cannot distinguish unambiguously between real and apparent sea-level rise because the measurements from them are biased by natural and anthropogenic land-level changes. Coastal regions which suffer from existing subsidence already have problems which will be exacerbated if the greenhouse predictions eventuate. Different parts of the State will not necessarily suffer the same degree of inundation and estimates of relative sea-level rise from one locality cannot be applied uniformly around the coast (Belperio, 1989).

High-resolution geodetic and tidal networks are planned for key areas of the State to compile local data on rates of relative land and sea-level change. To extract a greenhouse signature from tide gauge records requires that geological, isostatic, anthropogenic and other signals in the data be recognised and eliminated (Harvey and Belperio, 1994). This type of factual local data will provide for much improved and realistic coastal management, rather than the vague 'global' predictions that we have had to rely on in the past.

Coastal erosion

Erosion of sandy shorelines is a common problem around the world, resulting primarily from the diminution of natural sand sources to active beach systems. Coastal erosion is also exacerbated by human activities that interrupt or intercept littoral sediment pathways through urban development or coastal engineering works. Erosion of Adelaide's beaches is a typical case. The metropolitan beaches were constructed 6 000-7 000 years ago as the sea reworked unconsolidated sediments in the shallow gulf margins. Le Fevre Peninsula evolved progressively by littoral transport of sands northwards along the beaches (Bowman and Harvey, 1986). With slow coastal regression over the past 6 000 years, sediment accumulation has progressively decreased as the littoral drift system has exhausted its finite sediment sources to the point where the northward drift of sand exceeds the natural input.

Suburban development has further sterilised large areas of stranded foredune and dunes, thereby removing the buffer that existed to counteract occasional severe storms. The foreshore has suffered a number of major erosional events that have required further engineering works to protect property. Natural input into the littoral system is now essentially zero, and artificial nourishment is required to maintain the beach in its present condition. Several offshore sand prospects have been delineated for dredging for



Former mangrove woodlands drained and reclaimed by levees at Gillman, a western suburb of Adelaide. Photo 042554.



Top: Sand drifts at Brighton, a coastal suburb of Adelaide, in the 1880s. Photo 022611.

Centre: Coastal erosion at Glenelg after 1964 storms. Photo 022609.

Bottom: West-Lakes-Developmt Placing & Compact Sand Fill At West Lakes, By Kinnaid & Co. Photo 024399.

beach replenishment purposes, which is the best option to combat the erosion problem (Belperio *et al.*, 1990).

The coastline south of Marino is uplifted and rocky, with wave-cut platforms and embayments or pocket beaches. Here, headland erosion is a problem, particularly where soft Tertiary strata are exposed such as at Witton Bluff and Blanche Point. In the pocket beaches such as Hallett Cove, the limited quantities of beach sand are often removed by winter storms, and move back onshore with appropriate conditions in summer.

GROUNDWATER CONTAMINATION AND DRYLAND SALINITY

P.C. Smith and A.P. Belperio

Over-use of groundwater

Under State legislation (Water Resources Act, 1990), most intensive irrigation or high water use areas are proclaimed, that is, limits have been placed on the amount of groundwater which can be extracted so as to sustain the State's resources. Figure 11.40 shows the current Proclaimed Wells Areas. The outcomes of uncontrolled groundwater extraction are increased groundwater salinity and decline in potentiometric head. Committees with local representation have been set up in all Proclaimed Wells Areas to provide strategies and policies for managing the groundwater resources.

In the Northern Adelaide Plains Proclaimed Wells Area for example, wells which used to flow early this century now have potentiometric levels tens of metres below sea level at the end of the irrigation season. Salinity is increasing in some aquifers because of pumping-induced leakage from aquifers with higher salinity and through direct leakage of saline groundwater via corroded well casings.

Within the South Australian portion of the Great Artesian Basin, uncontrolled flowing wells have led to potentiometric head declines to the extent that some wells could require pumping if the pressure surface continues to decline. A well rehabilitation program is in progress to control free flowing wells. A similar situation of uncontrolled wells exists in the artesian area of the Dilwyn Formation in the Beachport-Lucindale-Kingston area of the South-East.

Problems of local over-exploitation also occur outside the proclaimed areas. For example, mutual interference between adjacent pumping wells is a problem in the Mount Lofty Ranges because of small allotment sizes. Although not a threat to the groundwater resource on a regional scale, the impacts locally can be profound, possibly leading to landholder disputes and litigation. Increasing effort is now being directed towards artificial recharge of aquifers, particularly utilising stormwater runoff.

Urban stormwater

Stormwater from major urban centres has traditionally been disposed of into creeks and out to sea. The management approach taken has been directed at disposal as quickly and cheaply as possible while preventing flooding as much as possible. Urban runoff is a major contaminant source, containing high levels of heavy metals, grease, oils, nutrients and bacteria. In addition, there is frequent accidental and deliberate disposal of industrial wastes. It has long been known that natural wetlands, especially coastal mangrove swamps, have an extraordinary capacity to accumulate sediments and pollutants (Harbison, 1986; Belperio and Harbison, 1992). However, mangroves are also an important shelter, food and nutrient source for a variety of marine organisms, so their use as a pollutant 'sink' is not desirable. Stormwater retention basins, although primarily constructed to manage large run-off events, produce a similar effect through ponding and settling of particulate-associated pollutants. In recent years, the concept of constructing artificial wetlands to mimic natural biogeochemical processes to further enhance the quality of urban stormwater has evolved.

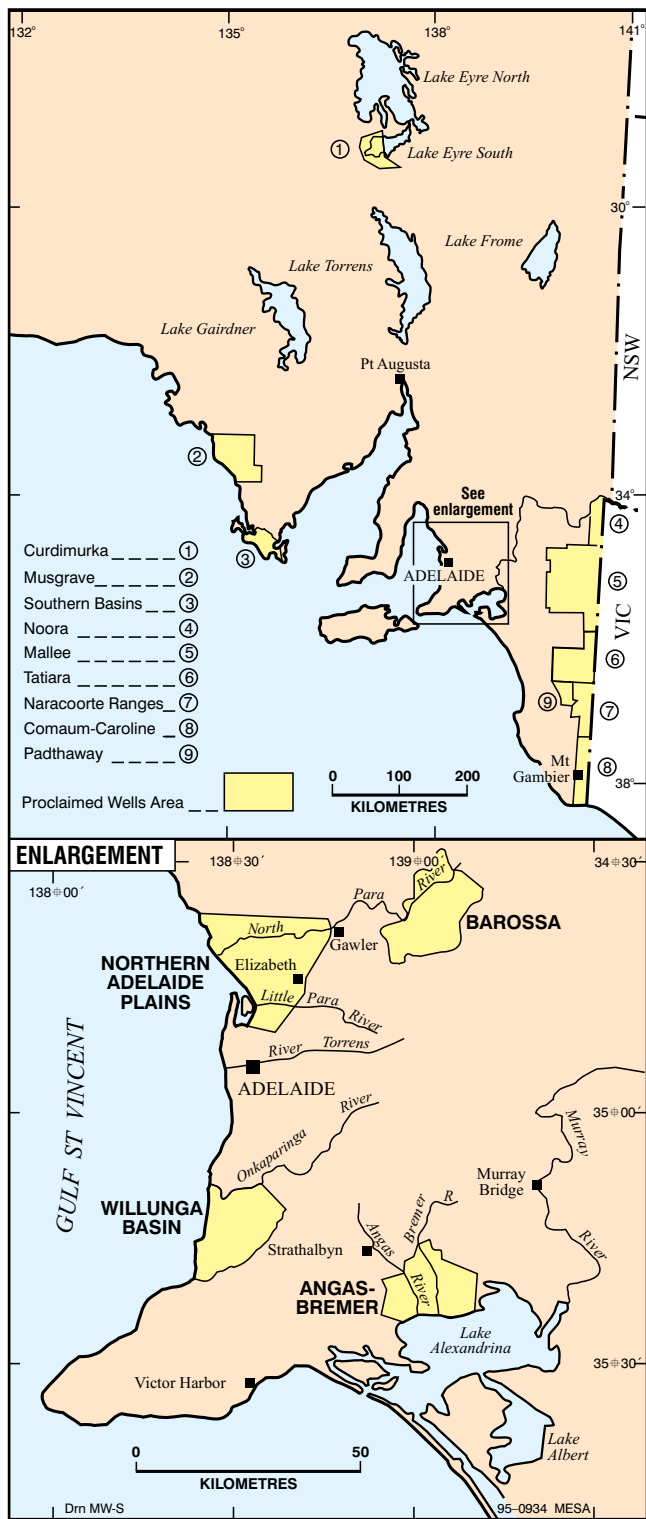


Fig. 11.40 Proclaimed Wells Areas of South Australia.

The Greenfield Wetlands is one such project conceived and undertaken by Salisbury Council to improve the quality of local stormwater prior to discharging into Dry Creek and the sea. The artificial wetlands consist of a number of shallow ponds and meandering channels designed to maximise water flowpath and retention time. Ponding and low flow rates allow the settlement of particulates. Aquatic vegetation assists sediment trapping and creates elevated pH in the water, enhancing metal precipitation and adsorption onto clays. Most pollutants, particularly metals and phosphorus, are removed from the water column by sedimentation and coprecipitation within a few days. Removal of nitrogen relies

on plant uptake and regular harvesting of algal and other growth is required to remove the nutrient repository as will the ultimate removal of contaminated sediments.

Similar wetland systems are being designed as part of the River Torrens Linear Park and for stormwater management around the Gillman Wetlands. As the quality of the runoff is improved, the potential for its re-use also increases. This is particularly relevant when considering that Adelaide's annual runoff of ~180 000 ML equals the total annual water use in the city. A critical factor to consider is the displacement time between maximum stormwater yield in winter and maximum demand in summer. Because of the limited surface storage available, the potential for redirecting runoff to temporary storage in Quaternary and Tertiary aquifers is being investigated. In addition to better resource utilisation, the twin approaches of benefiting and re-using stormwater has the added advantages of reducing marine pollution, improving flood mitigation and aesthetic enhancement of waterways.

Groundwater contamination

In the Lower South-East, shallow unconfined aquifers are at particular risk of contamination because of the extensive development of karst. Prior to development of modern sewerage, septic and household wastes were disposed directly to the ground throughout Mount Gambier and other South-East towns. Point sources of pollution such as stock yards, food processing facilities, fuel oil storage facilities, timber mills, stormwater disposal points and abattoirs have been documented (Waterhouse, 1977; Smith, 1983; Smith and Schrale, 1982; Schrale *et al.*, 1984; Emmett and Telfer, 1994). Diffuse pollution also occurs from animal husbandry activities, and fertiliser, herbicide and pesticide application. Karst regions also have a positive effect on groundwater in that meso-scale solution features aid rapid infiltration and recharge. Similar problems of diffuse pollution sources associated with urbanisation and farming activities occur in parts of the Mount Lofty Ranges where eutrophication of surface water and elevated nitrate levels have been reported (Edwards, 1979).

Land salinisation

Dryland salinity is a major land degradation problem in South Australia, with over 250 000 ha of agricultural land currently affected by saline seepage. The cost in lost agricultural production in 1992 was estimated at ~\$30 million and the problem is worsening.

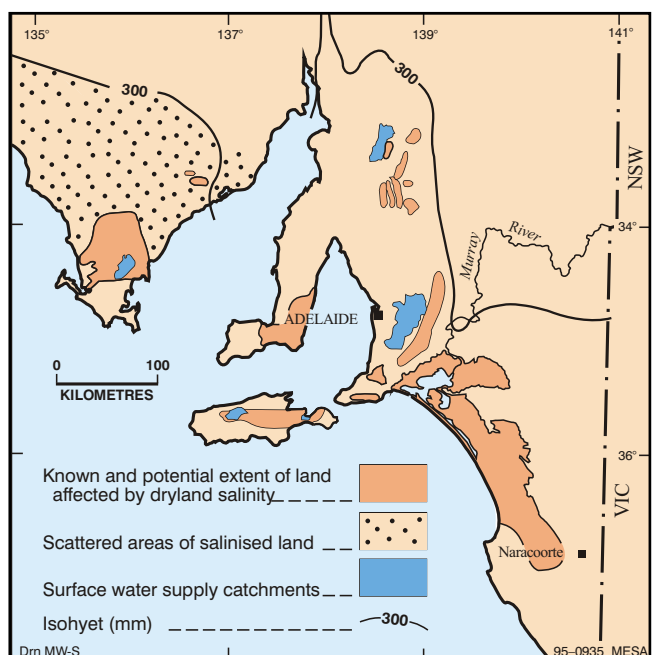


Fig. 11.41 Known and potential extent of land affected by dryland salinity.

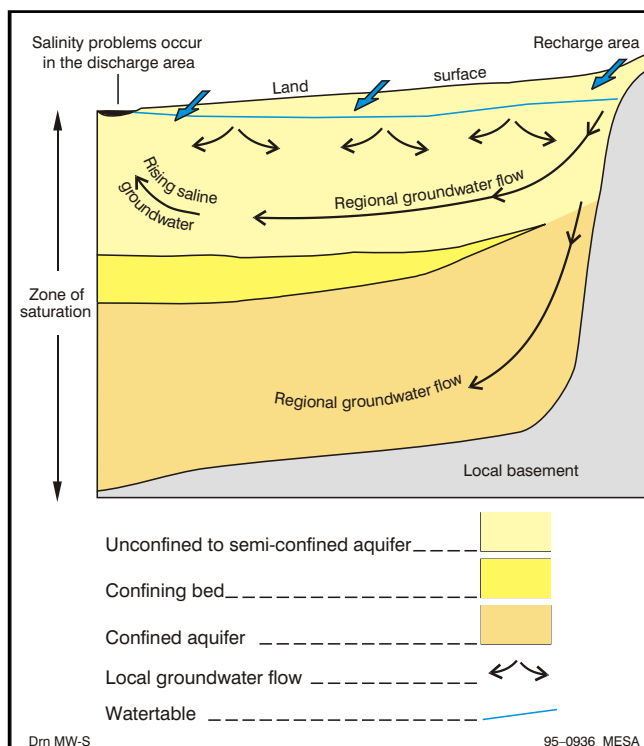


Fig. 11.42 Mechanism for development of dryland salinity.

Figure 11.41 shows the regions of the State affected or potentially liable to land salinisation.

Dryland salinity is caused by the historical widespread clearing of native vegetation which has upset a delicate hydrologic balance. Replacement of deep-rooted perennial vegetation with shallow-rooted annual pastures has allowed watertables to rise, with salts concentrated at the surface by evaporation (Fig. 11.42). In the Upper South-East, for example, groundwater levels are rising at a rate of 0.5-1.0 m every 10 years. The increasing salt levels near the ground surface, manifested as saline seepages and scalds in low-lying depressions, have a devastating effect on crops and vegetation. Because dryland salinity is essentially a groundwater problem, measuring the long-term trends in groundwater levels is the only way to predict where and when salinisation will occur and at what rate it will spread. Remedial measures include revegetating recharge areas with high water use plants, improved land-use practices and, in some areas, enhancing discharge and drainage. MESA is increasing its groundwater observation network in affected areas and areas under threat, and is contributing to integrated catchment studies.

MINING AND THE ENVIRONMENT

G.J. Drew and M. Stone

Mining is an intrinsic and necessary part of our society, providing the raw materials needed to build houses through to the rare metals needed for high technology components. It is also a major component of the economy, providing employment, infrastructure, royalties and export income. The resource industries spent \$65.2 million on exploration in South Australia in 1994, produced minerals and petroleum products worth \$1 165 million and paid royalties of \$49.6 million to the State. Direct employment is estimated at 5 000, with a further 10 000 employed in related processing industries. In addition, the mining sector makes large direct contributions to government revenue through rents and royalties, and has contributed significantly to South Australia's infrastructure and mining heritage. Whyalla, Port Pirie, Leigh Creek, Coober Pedy, Moonta, Kadina, Wallaroo, Burra and Roxby Downs are examples of towns that exist largely because of current or past mining and mining-related activities.

Because specific mineral deposits have a finite life, there is a

consequential need to discover more reserves, and to prevent known or potential areas of mineralisation from being sterilised by other land uses. Access to land for exploration is vital for maintaining a requisite level of exploration, as is the marketing of exploration concepts, models and geological data specific to South Australia and its strategic mineral requirements. Exploration, especially at the reconnaissance stage, requires access to large areas of land. At this stage, the techniques used (remote sensing, airborne surveys, etc.) typically have little or no impact on the environment. Once more prospective areas are identified, some land disturbance may be required to accommodate more detailed sampling or drilling. The industry is increasingly adopting techniques that minimise associated environmental impacts. Compared with exploration, the area used for mining is very small. Nevertheless, both exploration and mining are temporary forms of land use and consequently any disturbance needs to be minimised or rectified, and rehabilitation of mine sites is now an intrinsic part of the planning process. The aim of rehabilitation is to ensure that disturbed land is brought back to a form suitable for requisite use or in which a sustainable ecosystem of soil and plants can develop. Examples of successful rehabilitation include:

- housing estate developments (e.g. Glen Osmond and Magill)
- a golf course built on landscaped quarry waste (e.g. Linwood)
- return of land to an aesthetic natural setting (e.g. Rowland Flat)
- transfer of land to a national park (e.g. Tea Tree Gully)
- creation of a waste transfer and landfill site (e.g. Highbury)
- heritage conservation (e.g. Burra, Kapunda, Talisker Mines).

Codes of environmental practice emphasise the concept of environmental management for 'focused' protection. This concept recognises, firstly, that some sites and areas are more environmentally sensitive than others and that some, if damaged, may be lost forever (e.g. cultural heritage sites). Secondly, it recognises that operations must be managed according to the sensitivity of such areas, including the avoidance of extremely sensitive areas. Voluntary codes of environmental practice were initiated in 1984 for both seismic and drilling activities in the Cooper Basin. These became a statutory requirement in 1988. Codes have also been produced for petroleum companies operating in the South-East. Guidelines have been developed with the National Parks and Wildlife Service for mineral exploration in the Conservation Reserves of the Far West (including the Nullarbor and Yellabinnia Regional Reserves) and it is intended to extend these guidelines to cover other areas of the State. Sensitivity maps which identify sites of environmental significance are being developed by MESA to complement the codes and to assist in the planning stages of both mineral and petroleum activities.

Mining heritage

South Australia's economic development has relied heavily on the mineral industry. In the late 1840s, the rich copper deposits at Kapunda and Burra saved the Colony's economy when South Australia became a world copper producer. Copper lodes developed at Moonta-Wallaroo during the 1860s brought further prosperity when mining activities at Kapunda and Burra were in decline. In addition to these large mines, there were many smaller workings scattered through the Flinders and Mount Lofty Ranges producing copper, gold, silver, lead and zinc.

In recent years, a number of historic mine sites have been conserved, interpreted and opened to the public by way of museums and self-guided walking trails. A series of heritage drive trails have also been established in the major copper mining towns. Mining heritage is now one of the most distinctive and significant features of the State's cultural tourism industry. Various mining heritage sites are visited by ~100 000 people per year, the largest numbers being at Moonta and Burra. The most significant mining heritage sites are listed below (Fig. 11.43).

Glen Osmond Mines

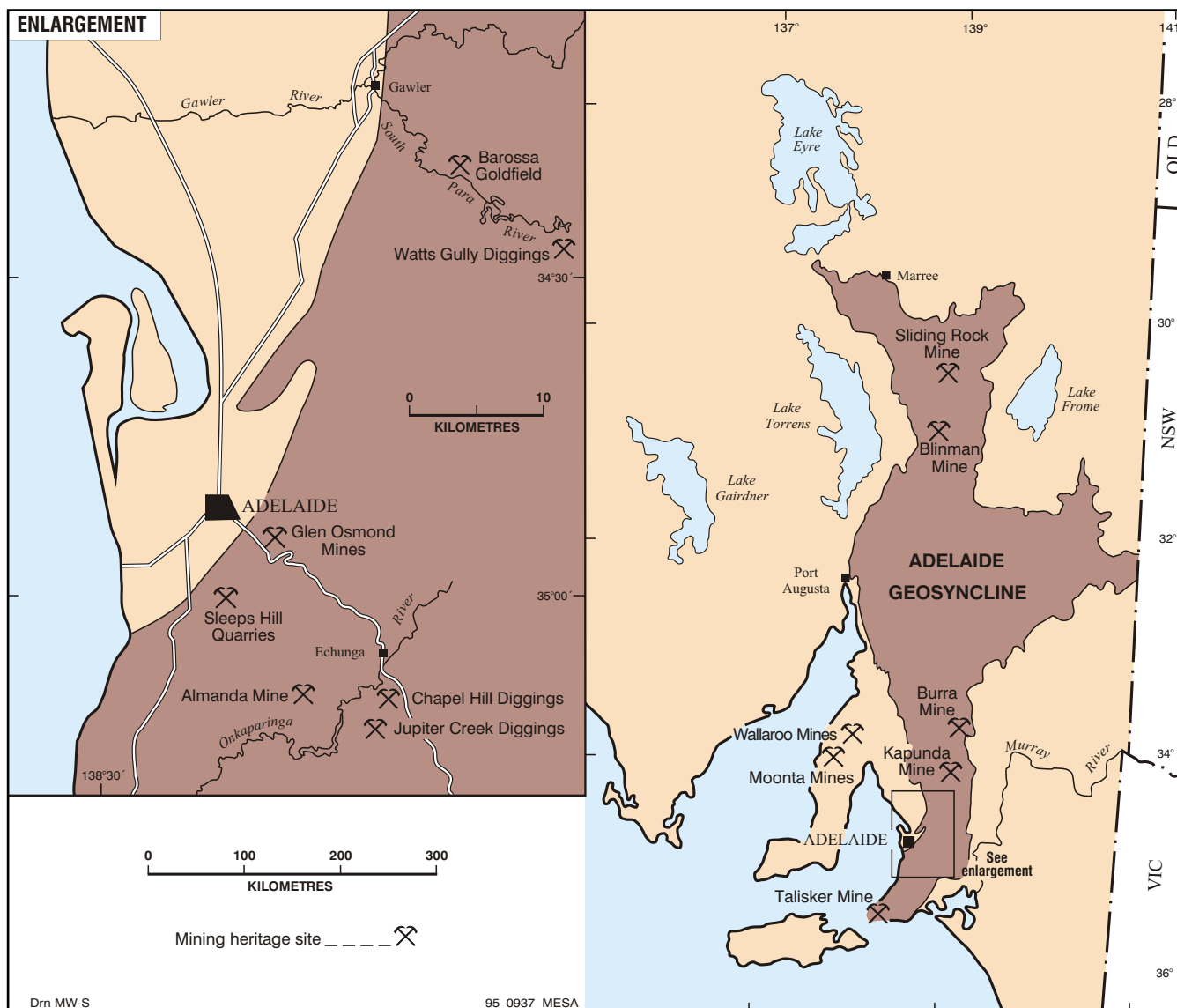


Fig. 11.43 Location of significant mining heritage sites in South Australia.

A number of small silver-lead veins were discovered at Glen Osmond between 1840 and 1842, which produced ~2 500 t of hand-picked ore up to 1850. Historically, this group of mines is of national significance. Wheal Gawler was the first metal mine in Australia and produced the first mineral exports from this country. The area is also the site of one of the earliest smelters in Australia. Two tunnels on the Wheal Watkins Reserve have been preserved and interpretive signs erected. Tours can be arranged through the Burnside City Council.

Kapunda Mine

The discovery of copper at Kapunda in 1842, together with Burra, contributed enormously to the Colony's recovery from economic crisis in the 1840s. Total production from the mine between 1844 and 1879 was 64 000 t of dressed ore, which yielded ~14 000 t of copper metal. A self-guided trail is available through the mine area and a local mining interpretive centre has been established in the town.

Burra Mine

Burra was famous for the richness of its copper ores and for the first 10 years of its life (1845-55) was the largest mine in Australia. Between 1845 and 1877, the mine produced ~50 000 t of copper metal. It was reworked between 1971 and 1981 by a modern open

cut to a depth of 100 m, producing a further 24 000 t of copper.

The mine ruins and township of Burra provide a comprehensive record of the methods and lifestyle of 19th century Cornish miners. The Burra Mine Museum allows access to the mine drive trail and interpretive walking trails. Within the mine is Morphet's Enginehouse, the only reconstructed Cornish beam enginehouse and shaft in the world. Nearby is the Bon Accord Mine where the mine offices and workshops (1859) have been developed into a mining museum.

Talisker Mine

Situated in Talisker Conservation Park, the Talisker Mine produced more than 1 000 t of lead between 1862 and 1872. The mine ruins and workings are part of an interpretive trail that includes details of Cornish mining and ore processing techniques.

Almanda Mine

This small silver mine is located in Scott Creek Conservation Park near Cherry Gardens. About 10 000 oz of silver were produced between 1868 and 1870. The site is now marked by a round chimney and the remains of an enginehouse. Interpretive signs have been installed, and an underground tunnel and open cut can be inspected.



1



4



2



3

1. Interpretive signposting at Sleeps Hill Quarries Trail. Photo 042555.
2. Remains of the calcining furnace (foreground) and crusherhouse at the Talisker Mine. Photo 042556.
3. Hughes Enginehouse after conservation, Moonta Mines State Heritage Area. Photo 042557.
4. Reconstructed Morphet's Enginehouse, Burra Mine Museum. Photo 042555.

Barossa Goldfield

Alluvial gold was discovered in 1868, resulting in a rush of up to 4 000 people, which lasted up to 1870. About 25 000 oz of gold were produced during this rush period. Reef mining took place in the 1890s but resulted in little production. Part of the goldfield is located on the northern section of Para Wirra Recreation Park where an interpretive trail has been established.

Jupiter Creek Diggings

Between 25 000 and 50 000 oz of gold were produced from these diggings after discovery in 1868. The site is now an Historic Reserve managed by MESA, and an interpretive walking trail has been established.

Blinman Mine

Blinman was the largest mine in the Flinders Ranges and produced ~10 000 tons of copper metal between 1862 and 1908. The mine is now protected by an Historic Reserve managed by MESA, and an interpretive trail has been established.

Moonta Mines

During the late 19th century, the Moonta and Wallaroo Mines formed one of the largest copper producing areas of the world. Their combined total production was ~330 000 t of copper metal. A driving trail now visits key sites of interest including Hughes and Richmans Enginehouses. A mining museum has been established in the old Moonta Mines School and a model railway allows access to the remains of the Precipitation Works. Hughes Shaft was reconstructed in 1992 and interpretive trails have been established.

Sleeps Hill Quarries

These quartzite quarries, which were worked between the two World Wars, are located on a council reserve. A self-guided historical/geological trail has been developed that links a series of quarries.

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- All entries refer to South Australia except where other States, countries or continents are indicated in brackets. The Australian States and Northern Territory are shown by the following abbreviations:

NSW	New South Wales	Tas.	Tasmania
NT	Northern Territory	Vic.	Victoria
Qld	Queensland	WA	Western Australia

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