Eucla Basin and peripheral paleovalleys

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SOUTH AUSTRALIA

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Acknowledgement of Country
The Department for Energy and Mining acknowledges Aboriginal people as the First Nations Peoples of South Australia. We recognise and respect the cultural connections as the traditional owners and occupants of the land and waters of South Australia, and that they continue to make a unique and irreplaceable contribution to the state.

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Eucla Basin and peripheral paleovalleys

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ABSTRACT

The Cenozoic Eucla Basin, located on the southern margin of the Australian continent with an onshore margin extending over 2,000 km from Western Australia into South Australia, comprises a thin passive margin succession that extends from onshore to more than 500 km offshore, to the approximate foot-of-slope of the Australia's continental margin. The basin contains a large onshore province of up to 300 m thick marine and coastal sediments of Cenozoic age, linked to an extensive network of peripheral paleovalleys that drained the Precambrian Yilgarn Block, Gawler Craton, Musgrave Province and Officer Basin.

Understanding the geology and sedimentary evolution of the Eucla Basin and peripheral paleovalleys has relevance to the exploration for placer deposits (e.g., gold, heavy minerals), secondary geochemical deposits (e.g., uranium) and for saline and rarely potable groundwater resources in the basin and channel sediments. Knowledge of the basin and paleovalley architecture and any concentration of minerals in the channels is also of interest as guides to the location of both paleochannel and bedrock lode deposits in the surrounding cratons (e.g., Yilgarn and Gawler). Geoscientific datasets have been integrated in an investigation of this Cenozoic basin and peripheral paleovalleys that have significance for mineral exploration. The objective of the study was to understand the basin characteristics and history, and develop a comprehensive spatial-depositional model to assist exploration in such huge basin-paleodrainage terrains. This was achieved through the combination of results from various geographical, geological and geophysical datasets. These include interpretations drawn from field observations, a compendium of geological and drilling data, computer modelling of ancient landscapes, topographic and evaluated digital elevation models, remote sensing imagery, geophysical data (e.g., magnetics, seismic, gravity, airborne and transient electromagnetics and radiometrics, where available), all of which have contributed to a systematic investigation of both shape and depth of the basin-paleodrainage terrains. Physical property contrasts that exist between the basin/channel sediments and the underlying bedrocks, for instance, can be differentiated by geophysical methods to locate the basin framework and paleoshorelines/paleovalleys.

Evidence from sedimentology was combined with other geological, geomorphological and geophysical characteristics to arrive at a general reconstruction of basinal and paleovalley architectures and depositional environments. The paleovalleys were incised originally into the pre-Cenozoic landscape, mostly weathered basement and Paleozoic and Mesozoic sediments, and became the sites where fluvial, lacustrine, estuarine and marine sediments accumulated during the Paleogene and Neogene. The application of sequence stratigraphy and facies analysis across the basin and adjacent paleodrainage network were integrated to establish the changes experienced in the basin and paleovalleys as conditions, notably sea level and sediment supply, fluctuated.

This study is a review and synthesis of geoscientific research undertaken in the Eucla Basin, southern Australia during last two decades. Over that time, various investigations have been made of the geophysical and geological characteristics of the Eucla Basin and paleovalleys, and related mineralisation. These projects, particularly in the eastern basin, have assisted exploration, and provide fundamental data for increasing knowledge of geological processes and landscape evolution within this important region. This report largely reviews previous results to develop a better understanding of the characteristics, geometry, geomorphology, and geological/depositional environment of the whole basin and adjacent areas, particularly on mineralised sediments associated with placers and uranium deposits on the margins of the Eucla Basin.

1. INTRODUCTION

1.1 PROJECT RATIONALE

The Eucla Basin and its margin, extending over 2,000 km from Western Australia into South Australia, contain a large offshore, nearshore and onshore province of up to 300 m-thick marine and coastal sediments (Fig. 1.1). Integral to the Eucla Basin was the development of Cenozoic radial paleodrainage systems that mostly discharged into the preserved sedimentary basin. These comprised at least 16 major river networks that drained large tracts of the Precambrian Yilgarn Block, Gawler Craton and Musgrave Province, as well as Officer Basin (e.g., Alley et al., 1999). Thus, the development of paleodrainage systems and the Eucla Basin are genetically related and can be unified to form part of a greater depositional system. Owing to its size and remoteness, however, the Eucla Basin is comparatively understudied. The sedimentary records of both basin and peripheral paleovalleys can provide an understanding of the geological history of the region that is greater than that which can be established by a study of the basin sediments alone. The presence of the offshore basin, marginal basin (e.g., paleoshorelines) and onshore basin (peripheral paleovalleys) presents numerous exploration difficulties and opportunities (e.g., within and beneath these sediments), and challenges traditional usage of geological, geophysical and geochemical exploration techniques (Hou et al., 2000). The processes of the basin and paleovalley formation can, to some extent, both highlight or disguise exploration targets and, when understood, can be applied to advantage in mineral exploration. Studies of the Eucla Basin and of the peripheral paleovalleys were generally carried out in isolation from one another (e.g., local geological mapping and notes/reports) and prior to commencement of this project in 2002, there were few concerted attempts at synthesis or of linking the various records.

Another major motivating factor behind evaluating the whole Eucla Basin region was the opportunity in such a study to provide correlation of the stratigraphic framework and sedimentary evolution between the western and eastern basins and their adjacent paleovalleys, and thereby integrate evidence on the timing of episodes of marginal marine and non-marine sedimentation and heavy mineral (HM)/uranium sources in the Eucla Basin and paleovalleys that drained inland Australia. Interpreting the evolution of inland paleovalley systems in particular, presents several problems which can extend also to the nearshore and offshore settings of the basin. Problems arise, from the lack of suitable fossils for precision dating (e.g., compound weathering, particularly of onshore sediments, is a serious factor that can destroy part, or all, of what fossil record existed), from the difficulty in calibration of the reconstructed sedimentary record against regional and/or local sea-level events, and thus placing the records within the context of an international time scale. Uncertainty around depositional timeframes makes sequence recognition and correlation difficult. For the Eucla Basin, some of these issues are magnified due to tectonic influences leading to thin depositional sequences and subsequent poor exposure, making the recognition and correlation difficult, with too few suitable geological sections available for detailed study. However, the margin of the Eucla Basin and peripheral paleovalleys were considered to offer a rare insight into the evolution of the southern Australian margin during early Cenozoic time, with the opportunity for linking onshore and offshore events through age and sequence constraints extrapolated across the wider basin by relating non-marine sediments with dateable marine sediments.

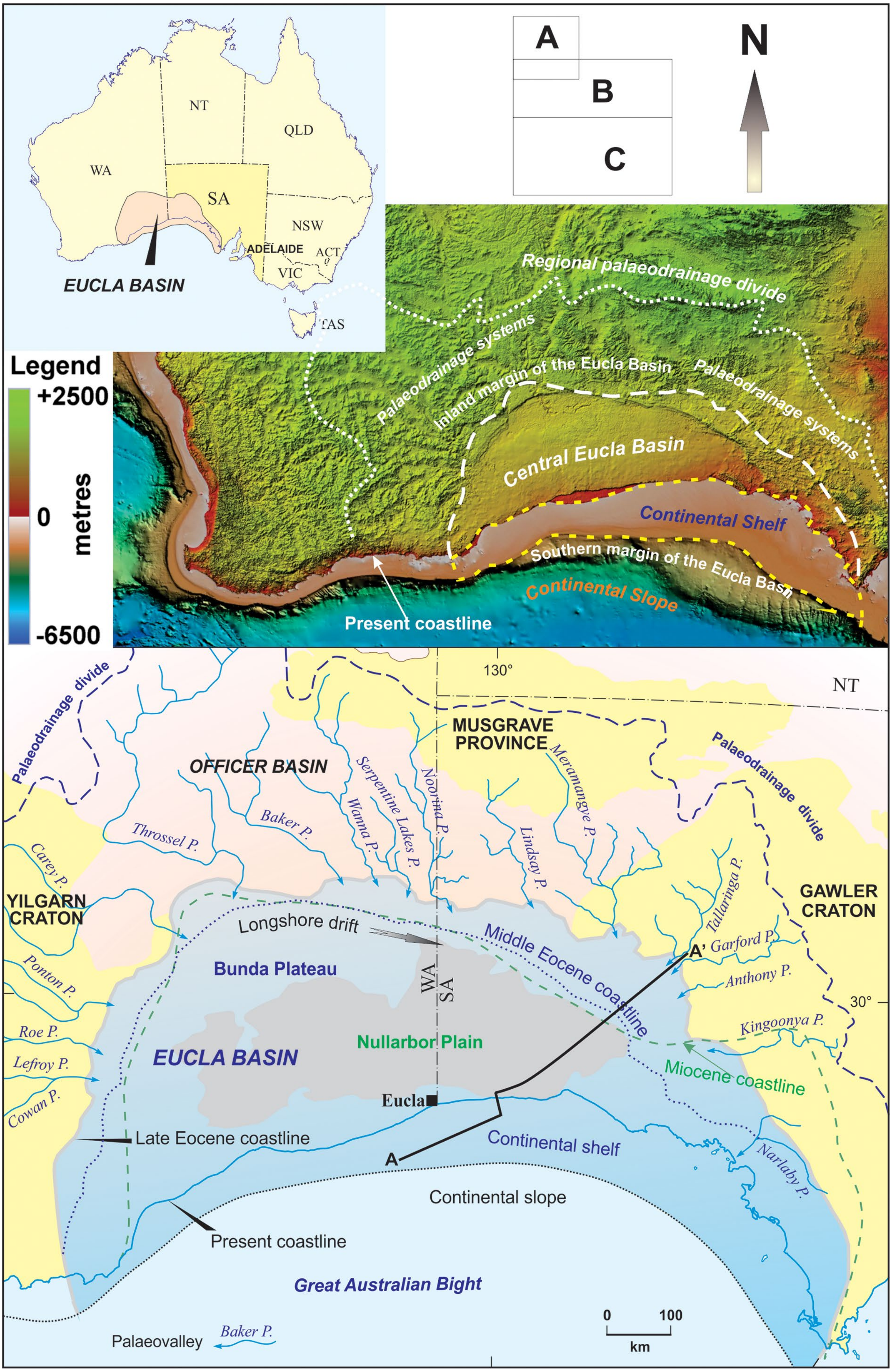


Figure 1.1 Eucla Basin of southern Australia (from Hou et al., 2021). (a) Location of the Eucla Basin. **(b)** Interpreted Eucla Basin frameworks draped over modern digital elevation

model (Imagery courtesy of Geoscience Australia) for the region, showing the main paleodrainage divide, onshore valleys, present coastline, continental shelf and platform margins. **(c)** Detail of the Eucla Basin showing geologic relationships with surrounding highlands (cratons and older basin), Cenozoic coastlines and their approximate ages, paleovalley systems and adjacent major features referred to in the text.

The Eucla Basin and peripheral paleovalleys that are preserved well in some regions of the basin, generally provide a more complete geological history of the basin that is not available from other regions of the basin. There can be various reasons for this. For instance, the information about paleoshorelines, paleoclimate and tectonic history may have only formed or be best developed in a marginal setting, in which the records were formed in certain depositional environments. During highstands, terrigenous sediment may be trapped in nearshore environments of the basin whilst the onshore and offshore basins were starved of such sediment. Due to the sedimentary, paleoclimatic and tectonic impacts, the coastal barriers may be best developed in parts of paleoshoreline whilst the remainder of the paleoshoreline were absence of such barriers. Also, the coastal barriers, which were well-developed in the eastern basin margin might protect and allow preservation of sediment behind them. Subsequent marine transgressions could also result in the preservation of an earlier depositional record in the eastern basin margin; the depositional results of which might be different from those developed in the western margin.

Placer and other (e.g., uranium) deposits associated with sediments in basin and paleovalleys have been known in many places around the world since the last century (Hou et al., 2014). During the last two decades, the analogous Cenozoic paleodrainage systems of Western Australia have yielded substantial quantities of gold derived from the chemical or physical dispersion of lode deposits on the Yilgarn Craton (Morgan, 1993). Also, the discovery of primary gold deposits/prospects (e.g., Challenger, Campfire Bore, Golf Bore, and Birthday) beneath cover in the Gawler Craton Gold Province of South Australia, and heavy mineral (HM) deposits (e.g., Jacinth and Ambrosia) in the marginal sediments, and uranium (e.g., Narlarby and Warrior) in the channel sediments, the Eucla Basin and its paleodrainage networks have become priority exploration targets. Over the Cenozoic period, the weathered sediments eroded from mineral-bearing surface rocks of the Yilgarn, Musgrave and Gawler Cratons and Officer Basin were carried by paleovalley rivers that discharged into the Eucla Basin. Gold, diamond and heavy minerals may occur as placers in coastal barriers or within channels. Sandstone-hosted uranium (e.g., Binks and Hooper, 1984; Curtis et al., 1990) and palygorskite (Keeling et al., 1995) may be present in economic concentrations. Widespread coal deposits may be found to reach commercial significance in associated floodplains, while groundwater resources are known from the sediments within the paleochannels. Therefore, a systematic and detailed study of such a huge basin and adjacent paleovalleys can provide the possibility of locating economic minerals in channel deposits and even give indication of nearby bedrock lode deposits.

In exploring and studying unknown buried basin and channel deposits, one of the major difficulties is how to explore for such deposits beneath many metres of overburden over quite large areas without resorting to systematic, broad-spaced drilling. Carefully defined targets tested by close-spaced drilling has been shown to be more effective in this environment. In exploration for mineral deposits in basin and channel sediments, precise geometric definition of the paleoshorelines and paleovalleys becomes important. In recent years, many attempts at using geophysical techniques to successfully delineate paleochannels have been reported; these have included magnetics, seismic, gravity, electromagnetics, and down-hole gamma logs (e.g., Smyth and Barrett, 1994; Deen et al., 2000; Leslie et al., 2000; Mackey et al., 2000; Hou et al., 2001a; 2003a). Additionally, remote sensing and digital topography assisted by computer techniques, have proven to be useful and important in delineation of paleochannels (e.g., Tapley and Wilson, 1985; Tapley, 1988; Statham-Lee, 1994; Hou et al., 2000; Hou and Mauger, 2005).

In the present study, there are considerable uncertainties due to the unknown characteristics of much of the subsurface geology, that is, subsurface sediments and bedrock beneath the surficial cover across this huge area. Recognition of important paleogeographic setting and boundaries within the sedimentary sequences is readily achieved in some cases but is very difficult in others. Even when guided by prior experience, the accuracy of geophysical models (imagery) and the estimation of the parameters of marine and stream dynamics from both field and borehole observations is sometimes problematic. Since the late-1980s, the Eucla Basin has been a target for heavy mineral sand exploration, with several significant discoveries along the eastern margin of the basin, at various stratigraphic levels, but with less success in the western basin. Although there exists a body of literature describing the geomorphology and origin of the basin, there have been few works studying its sedimentary history based on a whole of basin view, particularly paleoclimatic and tectonic impacts on the paleogeography, and economic significance. It is necessary and essential now to build a model of the entire basin and its resources.

1.2 AIMS

During last 20 years, the Eucla Basin and paleovalley studies in South Australia investigated the geophysical and geological expression of the eastern basin and adjacent paleovalleys and related mineralisation in both transported and residual regolith and its relationship with paleodrainage landforms. These projects have assisted exploration in the eastern Eucla Basin and provide fundamental data for building knowledge of geological processes and landscape evolution within this important region. The interest is extended to the west because it has strong affinities with other well-known commodity (e.g., heavy mineral sands and uranium) producing regions, but has been under-explored due to the widespread regolith cover, including paleovalley sediments.

The project’s primary objective was to develop a better understanding of the dynamic evolution, geometry, geomorphology, and geological/ depositional environment of the whole basin and adjacent areas, particularly on mineralised sediments associated with heavy mineral and uranium deposits discovered recently on the margins of the Eucla Basin. The project’s secondary objective was to facilitate prospectivity analysis of the basin and peripheral paleovalleys by reconstructing prospective paleoshorelines and mapping paleovalleys, and to develop geoscientifically and technically efficient procedures for mineral exploration through an integrated understanding of the geological processes in the Eucla Basin and paleovalleys. This challenging task required research into the history and post depositional deformation of the whole basin. Consideration and integration were required of detailed investigations of the basement, tectonic, sedimentary, transported and residual regolith development, and landscape evolution and their effects on the surface expression of concealed mineralisation. This research, together with new information about the provenance of the commodities, such as heavy minerals and uranium, has led to the identification of new exploration targets, as well as a better understanding of the history and development of the basin and peripheral paleovalleys.

The principal purpose of this report is to provide an overview of the whole Eucla Basin and adjacent paleovalleys, as well as associated mineral resources and their exploration characteristics. A review of the geological and geomorphological features that are common to and may control the deposition of the basin and associated paleovalleys is presented to assist with understanding the basin evolution history. Studies of the sedimentology of the basin and associated channel sediments, particularly their correlation across the whole basin, are described in this report where sufficient regional data are available. Aims and expected outcomes / benefits of the project include:

- to test various methods/techniques for successfully delineating the Eucla Basin and paleovalleys.
- to reconstruct the paleogeographic settings of the Eucla Basin and paleovalleys.
- to identify the provenance of the infill sediments of the Eucla Basin and paleovalleys.
- to study and correlate the stratigraphic framework of the whole basin and adjacent paleovalleys.
- to apply sedimentary geology to exploration for sediment-hosted mineral deposits.
- to trace changes in Cenozoic environmental conditions affecting the surrounding cratons by establishing the history of the basin and channel sedimentation.
- to devise models for locating possible mineralisation related to the basin and paleovalleys.

1.3 OVERVIEW OF EUCLA BASIN AND PERIPHERAL PALEOVALLEYS

Eucla Basin and peripheral paleovalleys are characterised as a Cenozoic drainage basin (i.e., primary- and incised-valley networks; Fig. 1.2) bounded by continental drainage divides and occupied by drainage networks (Bates and Jackson, 1987), ranging from individual hillslopes surrounding small headwater tributaries, to large tracts of continents drained by major river networks. These major paleovalleys, such as Raeside, Cowan, Lefroy, Roe, Rebecca, Ponton, Carey, Throssel, Baker, Wanna, Serpentine Lake, Noorina, Lindsay, Meramangye, Tallaringa, Garford, Kingoonya and Narlaby drainage basins, are bounded by the regional divides in their headward tracts and then form a roughly radial pattern focussed on the Eucla Basin margin (Fig. 1.1).

The term 'Eucla Basin' was first used by H.Y.L. Brown on a geological map of South Australia published in 1900 (Brown, 1900), and later extended to the limestone area in Western Australia (Lowry, 1970) for describing a large, shallow, arcuate sedimentary basin. The Eucla Basin is located in south central Australia, north of and adjacent to the Great Australian Bight (roughly south of the 26°S and between the 121°E and 136°E; Fig. 1.1). The offshore margin of the Eucla Basin is taken as the limit of the present distribution of the Eucla Group, which corresponds with the limit of the Wilson Bluff and Nullarbor Limestones; whereas the onshore margin is taken here as the limit of the present distribution of the onshore sediments deposited in the Eucla paleovalleys, which corresponds to the limit of the marine-influenced sediments deposited in the paleovalleys. The basin also includes the continental shelf in the Great Australian Bight, i.e., the southern limit of the basin defined at about the edge of the continental shelf (Fig. 1.1b). Thus, the present study has been so oriented as to include the Eucla Basin and peripheral paleovalleys on the Yilgarn and Gawler Cratons and the Officer Basin adjacent to the continental margin of the Eucla Basin. Accordingly, the investigation of the coastal area of the eastern Eucla Basin (e.g., Hou et al., 2008; 2011b) has been used to build a generalised sedimentary picture, but the main focus is extended to the whole basin and peripheral paleovalley networks. Detailed new data were collected within this area.

Principal means of access are the Transcontinental Railway that centrally crosses the Eucla Basin being located in the Nullarbor Plain, and the Eyre Highway (Fig. 1.3). There are several small settlements and motels along the railway and highway and sparsely distributed homesteads on sheep or cattle stations. Access for 4WD is good in the central and northern parts of the basin where there are numerous tracks and open grassy plains or sparse scrub. Further west, north and east of the Eucla margin and south of the Eyre Highway, there are numerous tracks in the sand-dune and dense mallee scrub areas, with poor driving conditions.

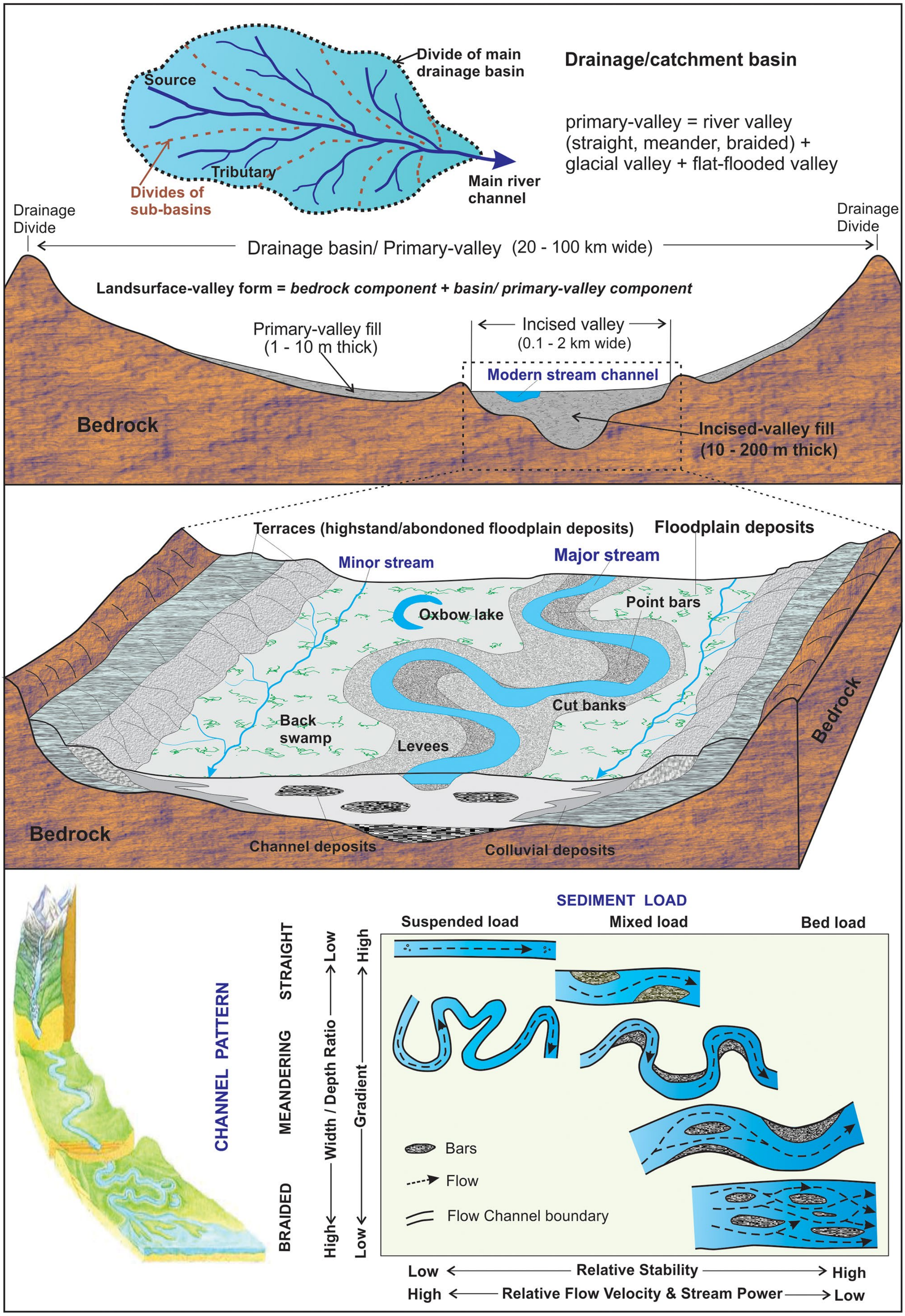


Figure 1.2 Major geomorphic and sedimentary components of the drainage basin developed in the Eucla Basin and peripheral paleovalleys, southern Australia

(after Schumm, 1981; Dalrymple et al., 1994; de Brockert, 2002; Hou et al., 2001a; 2014, 2017b).

The climate ranges from warm semi-arid along the present coast to warm arid inland. The most of central and inland margin of the basin have a desert climate, with short, cool winters and long, hot summers. Both diurnal and seasonal temperature variations are high. In January, for instance, the normal daily minimum and maximum temperatures are about 18°C and 35°C in the central and inland margin of the basin, and 15°C and 24°C along the coast. In July, however, the respective temperatures are 4°C and 18°C in the central and inland margin of the basin, and 7°C and 17°C along the coast. Rainfall is very low and variable, with a mean annual rainfall of 150–500 mm. The average annual rainfall, for instance, along the coast, ranges from 500 mm at Israelite Bay to 280 mm at Eucla and diminishes inland to about 150–250 mm. There is no distinct seasonal pattern, with summer rainfall often due to thunderstorm activity inland, but along the coast most of the rain falls in the winter. The mean annual evaporation rate ranges from about 1500 mm in the south to over 3,000 mm in the north (e.g., about 2600 mm is the mean annual potential evaporation at Kalgoorlie).

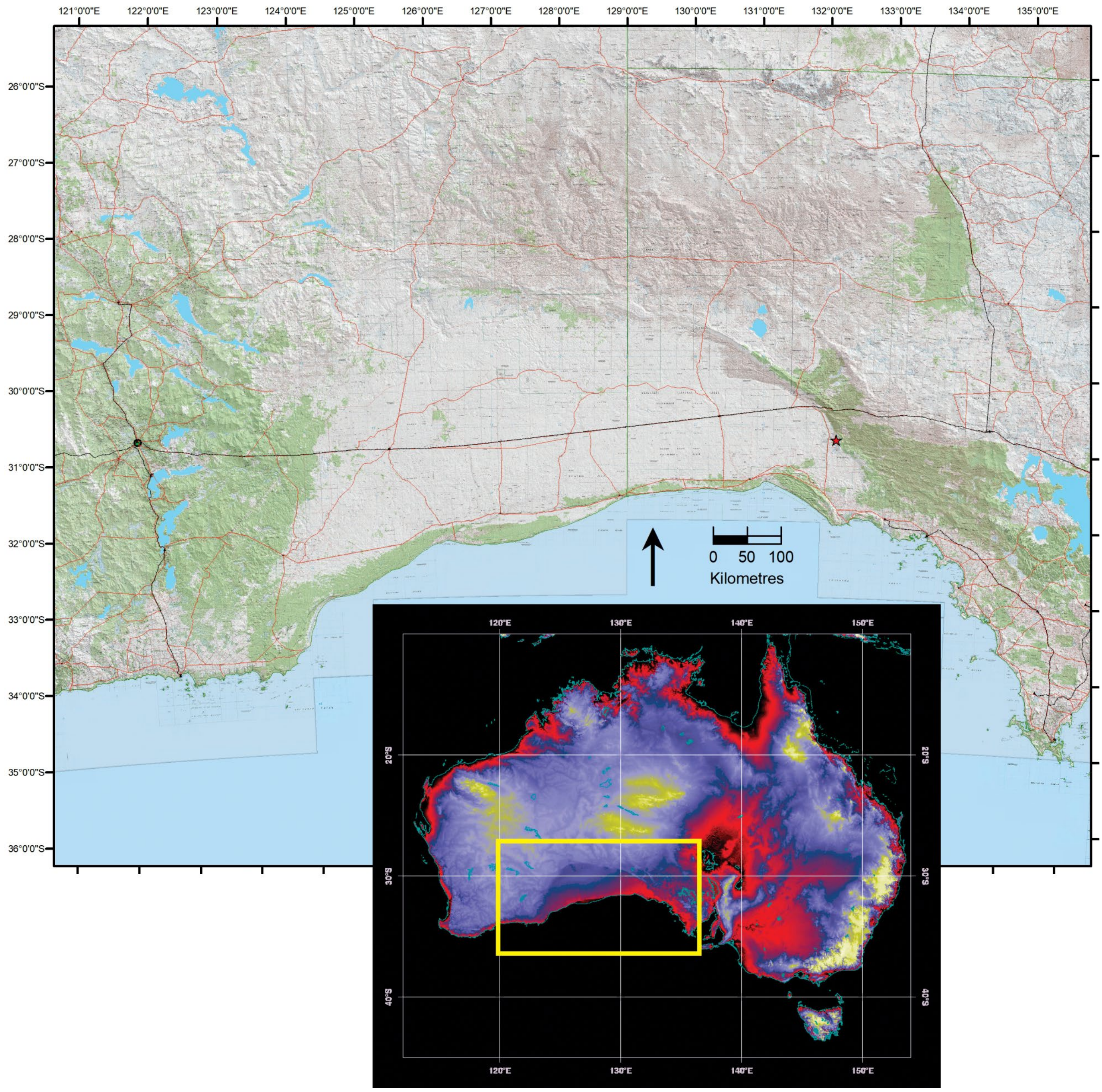


Figure 1.3 Major geographic and access road and rail features across the Eucla Basin and peripheral paleovalleys, southern Australia.

1.4 PREVIOUS WORK

The initial interest was started from the scattered observations on the geology of the coastal part of the Eucla Basin by the early explorers, Matthew Flinders and Edward J. Eyre, but the first important study was made in 1879 in the area between Fowlers Bay of South Australia and Eucla of Western Australia (Lorry, 1970). The description (Tate, 1879) of three limestone units and their correlation, including physiographic remarks, was considered as the best description of the surface geology of the Eucla Basin during the following 70 years. The first geological map was made by Gibson (1909) in the western part of the basin in early 1900s, with little stratigraphic information. The stratigraphy of the basin was relatively unknown until some authors (e.g., Glaessner, 1953; Singleton, 1954; Crespin, 1956) worked on the ages of the limestones, although the water bore drilling along the Transcontinental Railway provided information on the structure of the basin and the underlying Cretaceous beds (Maitland, 1904, 1911, 1915; Fairbridge, 1953). Significant works progressed during late 1950s – 1960s, which contributed knowledge of the fossil faunas and ages (Ludbrook, 1958, 1960, 1963, 1967a, 1967b; Lowry, 1968a; Ingram, 1968), stratigraphic drilling (Shiels, 1960a, 1960b; Stach, 1964), the geomorphology of the region (Jennings, 1961, 1962, 1963, 1967a, 1967b), cave development (e,g., Lowry, 1964, 1968b) and water bore information for understanding the stratigraphy and structure of the basin. Minor geophysical surveys were carried out, including marine seismic surveys (Tenneco Australia Inc., 1967, 1968), aeromagnetic traverses (Quilty and Goodeve, 1958), a magnetometer survey (Grasso and Blumer, 1964) and gravity surveys (Gunson and van der Linden, 1956; Blumer and Webb, 1965a, 1965b). Lowry (1970) produced a relatively complete geological description of Western Australian part of the Eucla Basin, which became a baseline reference for subsequent studies.

Since 1970s, the geology of the Eucla Basin and paleovalleys has been studied in geological review/mapping and exploring for significant resources of alluvial and beach placers, uranium, coal, clay minerals and groundwater, mostly at a local-scale, by various workers (e.g., Lowry, 1970; Bunting et al., 1974; Benbow, 1990a, b, c, 1991, 1993; Barnes and Pitt, 1976; van de Graaff et al., 1977; Benbow and Pitt, 1978; Benbow et al., 1982; Pitt et al., 1978; Smyth and Button, 1989; Jones, 1990; Hocking, 1990; Devlin and Crimeen, 1990; Fulwood and Barwick, 1990; Cowley and Martin, 1991; Commander et al., 1991; Clarke, 1993, 1994a, 1994b; Morgan, 1993; Kern and Commander, 1993; Dusci, 1994; Benbow et al., 1995a, b, c, d; Keeling et al., 1995; Alley and Lindsay, 1995; Rankin et al., 1996; Dodds, 1996, 1997; Lintern et al., 1997; Alley et al., 1999; Carey and Dusci, 1999; Johnson et al., 1999; Benbow et al., 2000; Clarke and Hou, 2000; Rogers, 2000; Hou, 2008; Hou et al., 2000; 2001a, b, 2003a, b, c; 2006a, b; 2007; 2008; 2010; 2011a, b; 2012; 2021; Johnson and McQueen, 2001; de Broekert, 2002; Hou and Alley, 2003; Hou and Mauger, 2005; Li et al., 1996a, b, 2003, 2004; Clarke et al., 1996, 2003; de Broekert and Sandiford, 2005; Johnson, 2015; Gartmair et al, 2021; Gartmair, 2022), in which available drilling data have provided stratigraphic information for the Cenozoic sediments of the Eucla Basin and associated paleovalleys.

The eastern Eucla Basin initially experienced mineral exploration during the 1970s to early 1980s for sedimentary uranium and brown coal (lignite), but in the late 1980s – early 1990s heavy mineral exploration became widespread, which resulted in identification of occurrences of anomalous heavy minerals (Ferris, 1994). In exploring for heavy mineral deposits, of particular note was the recognition of Eocene coastal barrier systems. The recognition of Paleogene – Neogene coastal features included the Ooldea Range (Benbow, 1983a, b, 1986a, b; Benbow and Crooks, 1988). The recognition that these were associated with a former coastal barrier system was a breakthrough (Benbow, 1990a, b) but focused attention on the topographically elevated Ooldea, Barton and Paling Ranges that have substantial components of reworked younger aeolian deposits with low HM content. Heavy mineral exploration was carried out from the middle 1980s until the early 1990s in some areas of the Ooldea Range. Previous company exploration had shown the Ooldea Range contained several anomalous areas for heavy minerals, such as the discovery of HM sands (8.27% HM) by CRA Exploration Pty Ltd (Close, 1973), in the eastern basin (the Lake Anthony-Lake Bring area).

Detailed work by Benbow (e.g., 1983a, b; 1990a, b; 1991; 1993) and Benbow et al. (1995a) on unravelling the Cenozoic stratigraphy of the eastern basin and associated paleodrainage and initial paleogeographic reconstructions, substantially raised awareness of the exploration potential of the area. The Ooldea Range became the main target for company exploration drilling with several zones of anomalous HM mineralisation outlined (Ferris, 1994). This phase of exploration confirmed the presence of HM in sands within the Ooldea Range but the remoteness, difficult drilling conditions, thick overburden, and typically low content of rutile in the HM suite were regarded as impediments to further exploration (e.g., Jurica and Rothnie, 1990). A highlight of the work was the intersection near Lake Ifould of sands at 38–40 m depth grading 27% HM comprised of 51% zircon, 43% ilmenite and altered ilmenite, and 3% rutile (Jurica and Rothnie, 1990). Subsequently, recommendations were made that further exploration for heavy minerals within the eastern basin would require additional geological studies to provide a greater understanding of the stratigraphic framework and a greater emphasis on geomorphic factors including possible provenance areas and detailed analysis of sediments within paleodrainage channels to determine flow rates and other factors which influence the transport and concentration of heavy minerals (Benbow, 1990b; Ferris, 1994). Although hypsometric and Landsat topographic data together with detailed work on the Cenozoic stratigraphy of the Eucla Basin and surrounding paleochannels and paleogeographic setting has greatly enhanced the search for heavy minerals before 2000, the age, origin and stratigraphic relationships of the sediments, particularly basin-wide factors such as the role of sea-level events and neotectonics on the paleoshoreline development and basin evolution were not well established. Afterwards palynology, along with facies and sequence-stratigraphic analyses, were used to identify distinctive key-surfaces that corresponded to changes in relative sea-level (e.g., Clarke et al., 2003; Hou et al., 2003a, b). Further subdivision was proposed that included the terms Ooldea Sand and Barton Sand to denote barrier sands deposited in the respective shorelines (Hou et al., 2006b; 2008). The correlation of sediments east to west across the basin and from offshore to onshore made use of many earlier studies, often on specific sites within and marginal to the basin (e.g., Lowry, 1970; Bunting et al., 1974; Barnes, and Pitt, 1976; Alley, 1985; Alley and Benbow, 1993; Benbow et al., 1982; Hocking, 1990; Jones, 1990; Cowley and Martin, 1991; Clarke, 1993, 1994a, 1994b; Kern and Commander, 1993; Keeling et al., 1995; Alley and Lindsay, 1995; Clarke and Hou, 2000; Rogers, 2000; Hou et al., 2001a, b, 2003c; 2007; Johnson and McQueen, 2001; de Broekert, 2002; Hou and Alley, 2003; Li et al., 2003; Hou, 2008; de Broekert and Sandiford, 2005).

1.5 METHODOLOGY AND SCOPE

While the landscape of the Eucla Basin and the surrounding region contains remarkably well-preserved and varied paleolandforms of Cenozoic age (e.g., Benbow, 1990a), pervasive aeolian sand and clay cover and widespread duricrusts in the present-day landscape hamper thorough mapping of the Eucla Basin and associated paleovalley sediment fills. Mapping information presented here is based largely on techniques and procedures that are typically used in delineation and evaluation of the Eucla Basin and adjacent paleovalley deposits. Reconstructing the distribution, thickness and sedimentary facies of the Eucla Basin and paleovalley sediments therefore challenges the application of geological and geophysical exploration techniques and presents both difficulties and opportunities for mineral exploration (Hou et al., 2000). Topography, digital elevation models (DEMs), Landsat, NOAA, ASTER and radiometric images, magnetic, seismic, gravity, and transient electromagnetics (TEM), and airborne electromagnetics (AEM) methods, where available, are integrated into this phase. Experience and knowledge in geomorphology and sedimentology therefore make the task of interpretation of paleocoastal landform features from remotely sensed images, and in the field, much easier.

Evidence from existing geological maps and drillholes was used together with geomorphologic and field studies to improve knowledge of the dimensions, trends and continuity of the Eucla Basin and paleovalleys. Materials used during the previous projects include samples collected during field work and others taken from mining open-pits, cores and cuttings mainly stored in the GSSA Core Library in Adelaide and the GSWA Core Library in Perth. These samples were examined for determination of sedimentary type and facies, environment of deposition, conditions of weathering, and some were subjected to petrological, mineralogical, and geochemical analysis of the sediments. The stratigraphic framework for the basin that underpins our interpretations relies to a large extent on palynological age determinations (e.g., Smyth and Button, 1989; Waterhouse et al., 1994; Benbow et al., 1995a; Hou, 2008; Hou et al., 2003a, 2003b; Stoian 2003a, 2003b, 2003c, 2004) from an extensive drillhole dataset as discussed in detail elsewhere (Hou et al., 2001a, b, 2003a, b, c; 2006b; 2008). Combining this with sea-level history, sedimentological analysis has been summarised for the interpretation of the formation of the basin, including the paleoshorelines and paleovalleys. To understand the sedimentary source for the HM accumulation, which has importance for exploration, detrital zircons and rutiles were dated using laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) at the University of Adelaide. Therefore, the methods of study on the basin and paleovalleys can be summarised below:

- Overview of the region and previous work.
- Province-scale mapping of the Eucla Basin and paleovalleys by using various geological and geophysical techniques.
- Relationship of the basin geography and provenance.
- Correlation of the stratigraphy basin-wide.
- Petrological, mineralogical and geochemical analyses of the basin and channel fills.
- Sedimentological analysis of the basin and channel sediments.
- Reviewing the Cenozoic paleogeography adjacent to the nearshore Eucla Basin.
- Interpretation of the source rocks.
- Implication of the effects of weathering and erosion to the Eucla Basin and paleovalleys.
- Developing conceptual and exploration models for the Eucla Basin and peripheral paleovalleys.
- Overview of economic significance.

2. REGIONAL SETTING

The Cenozoic Eucla Basin is a large onshore-offshore basin on the southern Australian passive margin. The geological history of the Eucla Basin and beneath and surrounding basement/bedrock is complex but reasonably well known (e.g., Lorry, 1970; Benbow et al., 1995a). On the pre-Paleogene landform, marine, marginal marine and terrigenous sediments constitute a partially dissected platform over which the paleovalleys extended from the crystalline cratons into the Cenozoic marine basin proper. Overlying basement and other bedrocks, major phases of marine, marginal marine and fluvial sediments were developed in the Cenozoic across the basin, with the marine influence extending more than one hundred kilometres up the paleovalleys during numerous major transgressions (e.g., Clarke and Hou, 2000; Benbow et al., 1995b; Alley et al., 1999). Geographically and geologically associated basement and bedrock with the Eucla Basin and peripheral paleovalleys mainly include the Yilgarn and Gawler cratons, Musgrave, Madura and Coompana provinces, Albany-Fraser Orogen, Madura Shelf, Officer Basin, Bight Basin, Eromanga Basin, Denman Basin and Polda Basin (Fig. 2.1). In the northern and northeastern parts of the Eucla Basin the bedrocks of pre-Cenozoic stratigraphy are more complex, including Mesozoic and Permian sediments. Therefore, the Eucla Basin onlaps the Madura Shelf, Officer and Eromanga Basins to the north and northeast, Precambrian cratonic rocks to the west and east, and extends south to the edge of continental crust. Early seismic surveys in the Great Australian Bight, interpreted by Tenneco Australia Inc. (1968), show flatlying Cenozoic and Mesozoic beds on a very irregular surface of granitic basement (Lowry, 1970). Recent drilling project in the Coompana Province and seismic surveys across the Eucla Basin along the railway line indicate flatlying Cenozoic and Mesozoic beds on a very irregular surface of granitic basement (Wise et al., 2015; Dutch et al., 2016; 2018a, b; Pawley et al., 2018).

2.1 BASEMENT – PRECAMBRIAN

Precambrian rocks form the basement beneath and around the Eucla Basin, including the crystalline cratonic rocks of the Yilgarn and Gawler cratons, and Albany-Fraser Orogen occurring beneath and around the Eucla Basin western and eastern margins respectively (Fig. 2.1a). A number of bores in the Eucla Basin have reached Precambrian basement rocks, and samples of basement from some of them are stored in the core libraries of the Geological Survey of Western Australia (GSWA) and the Geological Survey of South Australia (GSSA), such as: Transcontinental

Railway No .3 Bore (also known as Loongana Bore) – mylonitic gneiss, Transcontinental Railway No. 4 Bore-biotite granite, Gambanga No. 1-pyroxene-bearing granite, Eyre No. 1-gneissic granite, Eucla No. 1-granite, and a few drillholes from recent 2013/14 Eucla Basin basement drilling program of GSWA-Tech (western Eucla Basin), and more recent 2016/17 Eucla Basin basement drilling program of GSSA (central Eucla Basin).

The Precambrian crystalline basement underlying the majority of the Madura Shelf (beneath the central Eucla Basin) largely comprises granitic gneiss with up to 20 m of weathering horizon (quartz rich mottled saprolite) (Reynolds, 2016). Cratonic rocks, of the Coompana Block, also occur beneath the basin in the central southern region (straddling the WA/SA border). Musgrave Province in the north was genetically linked with the basin by several paleovalleys via the Officer Basin. The basement rocks of Yilgarn Craton, Albany-Fraser Orogen, Musgrave Province, and Gawler Craton have been identified as contributing heavy minerals to the Eucla Basin (Reid and Hou, 2006). Because of the relationships of known uranium and heavy mineral resources within the Eucla Basin and potential source rocks, some of the significant basement provinces are briefly described here.

YILGARN CRATON

The Yilgarn Craton (Groenewald and Tyler, 2006) is a highly mineralised granite-greenstone terrain that is the major crystalline basement region in the western hinterland of the Eucla Basin (Reid et al., 2013). The major rock-forming timelines in the Yilgarn Craton occur over the interval c. 3000–2600 Ma, with c. 2750–2650 Ma being a dominant timeline for magmatism within the eastern Yilgarn in particular (e.g. Cassidy and Champion, 2004; Griffin et al., 2004; Cassidy et al., 2006; Krapež and Hand, 2008; Pawley et al., 2012; Wyche et al., 2012). Assembly of the Yilgarn Craton was completed by c. 2600 Ma (Czarnota et al., 2010) although the craton margins have been reworked by Paleo- to Mesoproterozoic tectonism in the Albany-Fraser Orogen (Kirkland et al., 2011).

Much of the Yilgarn Craton in the Eastern Goldfields region, where a number of paleodrainage systems drained and discharged into the Eucla Basin (Figs 1.1 and 2.1), comprises linear belts of deformed and metamorphosed, sedimentary, felsic volcanic, and mafic-ultramafic volcanic and intrusive rocks (collectively termed greenstones) that are intruded by unmetamorphosed and weakly deformed granitoids (Myers, 1993; Swager, et al., 1995).

The geologic-geomorphic evolution of the Yilgarn Craton since Permian glaciation is summarised as follows (Lorry, 1970; Hocking, 1990, 1994; Alley et al., 1999; de Broekert, 2002):

- Permian glaciation of already planate bedrock.
- Mesozoic deep weathering, with minor planation.
- Establishment of a drainage pattern with major valleys several kilometres wide (playa lakes paleodrainage); regolith terrain may represent inverted NW flowing paleodrainage which would be presumably older than playa lakes drainage.
- The break-up of Gondwanaland.
- Incision of valleys and deposition of Eocene terrestrial sediments.
- Upper Eocene marine incursion penetrating 300 km up valleys, and associated marine sediments.
- Tectonic activity largely resulting in an epiorogenic uplift of about 300 m.
- Cenozoic landscape evolution involving renewed weathering, and duricrust formation; progressive landscape inversion.
- Climatic change from a warm and moist climate in the Mesozoic and Cenozoic, through increasing dryness in the late Cenozoic, to a Quaternary period of aridity involving salt weathering, aeolian deposits, limited drainage and the formation of remnant lake systems.

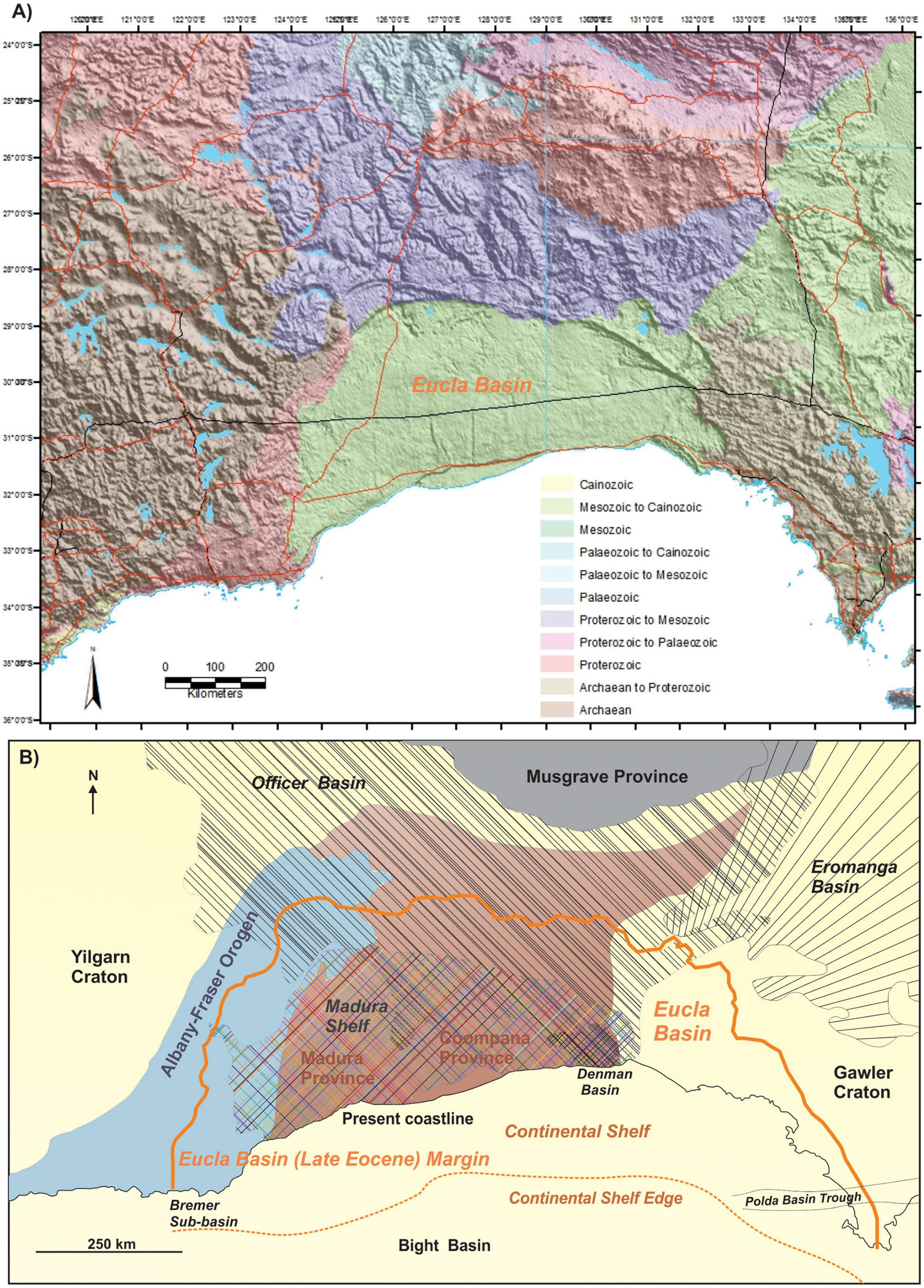


Figure 2.1 Geological framework of the Eucla Basin. (a) Time-slice geology draped over SRTM DEM. **(b)** Schematic geological relation diagram of the major geological domains and basement blocks of southern Australia. After Lowry (I970); Jackson and Van de Graaff (1981); Benbow (1990a); James et al. (1996); Hou et al. (2008, 2011); Gartmair (2022).

ALBANY-FRASER OROGEN

Precambrian basement rocks of the Albany-Fraser Orogen occur in the southwestern hinterland of the Eucla Basin with a high relief when the basin developed (Fig. 2.1), and the tops of scores of old hills now protrude through the Paleogene–Neogene limestone of the Mardabilla Plain and the Quaternary deposits on the adjoining shield (Lowry, 1970). The Albany-Fraser Orogen encompasses Paleoproterozoic to Mesoproterozoic rocks formed close to the Yilgarn Craton to the north and northwest, with major rock forming phases over the interval c. 1800–1650 Ma and between c. 1350 and 1150 Ma (Kirkland et al., 2011; Spaggiari et al., 2011). In part the Albany-Fraser Orogen developed as an active margin to the Yilgarn Craton during the Paleoproterozoic and was reworked extensively during the Albany-Fraser Orogeny over the interval c. 1350–1150 Ma, which resulted in the emplacement of abundant granitic melts associated with high-grade metamorphism (Kirkland et al., 2011).

GAWLER CRATON

The basement rocks of the Gawler Craton consist of Precambrian (episodes over the interval c. 3200–1500 Ma) metamorphic complexes which are the products of three (late Archean – early Proterozoic - c. 2550–2420 Ma), Paleoproterozoic - c. 1200–1620 Ma, and Mesoproterozoic - c. 1600–1500 Ma) megacycles of orogenic development (e.g., Daly, 1985; Daly and Fanning, 1993; Daly et al., 1979; Fanning, 1988). Major magmatic and mineralising events occurred at the beginning of the Mesoproterozoic (Hand et al., 2007; Reid and Hand, 2012). The Gawler Craton occurs in the eastern hinterland of the Eucla Basin (Fig. 2.1), and the western and northwestern Gawler Craton is mainly characterised by the Mulgathing Complex (c. 2550–2440 Ma) that encompasses all Archean to Paleoproterozoic rocks, and rhyolite, dacite and granitic rocks associated with the c. 1592 Ma Gawler Range Volcanics and Hiltaba Suite (1585 Ma felsic intrusive rocks) (Daly et al., 1979; Daly, 1986; Daly and Fanning, 1993). The Archean to early Paleoproterozoic Sleaford and Mulgathing complexes, forming an older basement in the western and northwestern portions, include gneiss and granulite formed during the Sleafordian Orogeny from sediments (including iron formation) and felsic to mafic igneous rocks. These rocks contain the Challenger gold deposit and numerous other gold prospects.

The Gawler Craton is overlain by Paleozoic, Mesozoic and Cenozoic rocks in places and locally by paleochannel fills of Mesozoic and Cenozoic sediments. Quaternary sands, clay and regolith blanket most of the area. Pre-Cenozoic sedimentary sequences at the surface and beneath Cenozoic sediments in the area consist mainly of Cambrian Observatory Hill Formation, Carboniferous-Permian Boorthanna and Stuart Range Formations, and Jurassic-Cretaceous Algebuckina Sandstone. All are unconformably overlain by the Cenozoic fluvial-marine sediments in the Cenozoic depressions.

MUSGRAVE PROVINCE

The Musgrave Province occurs in the northernmost portion of the hinterland to the Eucla Basin (Fig. 2.1). The southern boundary of the Musgrave Province is defined by aeromagnetic data (Major and Conor, 1993). This boundary is partly coincident with the unconformable contact with overlying Neoproterozoic sediments of the Officer Basin, and partly with thrusts along which the basement rocks have been transported over Officer Basin sediments during the Alice Springs Orogeny. The eastern boundary of the Musgrave Block is poorly defined where the basement is concealed beneath Phanerozoic cover of the Pedirka and Eromanga Basins.

The Musgrave Province, comprising Mesoproterozoic very high-grade metamorphics, with protoliths of still uncertain age and origin, and major mafic and felsic intrusives, developed over two major intervals of magmatism, sedimentation and metamorphism. An early c. 1600–1550 Ma phase of magmatism (Wade et al., 2006) was followed by sedimentation in the western Musgrave Province over the interval c. 1450–1350 Ma, (Smithies et al., 2009) and by a prolonged phase of magmatism and deformation over the interval c. 1345–1100 Ma known as the Musgravian Orogeny (Wade et al., 2008; Smithies et al., 2009, 2010). The final phase of significant magmatism within the Musgrave Province occurred c. 1080–1000 Ma associated with the Giles magmatic event (Wingate et al., 2004; Evins et al., 2010a, b). The Birksgate Complex and Wataru Gneiss of the Musgrave Province are high-grade metamorphic rocks produced during the Mesoproterozoic

Musgravian Orogeny (1200 Ma), which acted upon sedimentary, volcanic and older metamorphic rocks of uncertain age, but partly at least 1605 Ma old (latest Paleoproterozoic). During this orogeny intrusive rocks of the felsic Pitjantjatjara Supersuite (formerly Kulgera Suite) (1225–1190 Ma) were introduced. Large intrusions of the mafic to ultramafic Giles Complex, with potential for nickel and platinum mineralisation, were emplaced about 1070 Ma. During the Petermann Orogeny (540 Ma), a huge fault zone which filled with mylonite and partly melted rock (now solidified) allowed uplift of this part of the Musgrave Block from very deep in the Earth's crust and thrust it over the Olia Gneiss, which had formed at about 1600–1540 Ma in shallower parts of the crust.

MADURA–COOMPANA PROVINCES

The Madura–Coompana provinces are the regions of Precambrian crystalline basement that lie beneath the Eucla Basin and between the Gawler Craton to the east, the Musgrave Province to the north and the Albany-Fraser Orogen to the west (Fig. 2.1) (Spaggiari, et al., 2012; Dutch et al., 2018; Gartmair, 2022). Crystalline basement rocks of these regions exhibit a complex multi-phase history resulting in numerous mafic and felsic intrusive events as well as bimodal volcanics that have been metamorphosed to amphibolite facies (Spaggiari and Smithies, 2015). Rock-forming events of the Madura–Coompana provinces are very similar to those in the Musgrave Province and Albany-Fraser Orogen, and include Mesoproterozoic Moodini Supersuite (~1181–1125 Ma) and metamorphic episodes of the Toolgana and Undawidgi supersuites (~1179, ~1167, ~1150 and ~1174 Ma) and the c. 1070 Ma Giants Head Suite being a mafic magmatic event of identical timing to the Giles event of the Musgrave Province (Neumann and Korsch, 2014; Wingate et al., 2015; de Gromard et al., 2016; Jagodzinski et al., 2018).

2.2 BEDROCK–PRE-CENOZOIC

The bedrock sedimentary sequences at the surface and beneath the sediments in the Eucla Basin and associated paleovalleys consist mainly of Neoproterozoic-Paleozoic (e.g., Officer Basin) and Mesozoic sediments (e.g., Eromanga Basin), such as Ilma Beds on the northern margin of the basin (Lowry, 1970); and Cambrian Observatory Hill Formation, Carboniferous-Permian Boorthanna and Stuart Range Formations and Jurassic-Cretaceous Algebuckina Sandstone on the northeastern margin of the basin (e.g., Benbow et al., 1995a). All are unconformably overlain by the Cenozoic fluvial-marine sediments in the Cenozoic depressions. The Observatory Hill Formation is a sequence of micro-micaceous, argillaceous and carbonaceous mudstones with characteristic chert layers (Benbow, 1993). The lithology of the Boorthanna Formation is predominantly blue-grey to grey sandy clay/claystone and sandstone, which have been interpreted as of glacial, fluvial and lacustrine origin with some possible marine influence (Benbow, 1993). The Stuart Range Formation (e.g., drillholes TPS 7, 8 and 16) is characterised by grey-green massive compact mudstone with local silts and sand, believed to have been deposited in low energy, restricted marine environments as attested to by lithology and microfauna (Pitt et al., 1978). The dominant lithology of Algebuckina Sandstone is kaolinitic sandstone, which is partly gritty and well sorted, and interbedded with (sandy) claystones (Benbow, 1993).

Sedimentary bedrocks can be intermediate hosts of the heavy minerals that were reworked and deposited in younger heavy mineral sands (Reid et al., 2013). Sedimentary rocks in coastal regions can also contain enrichments in heavy minerals derived from local erosion of older igneous and metamorphic rocks. Erosion of the sedimentary rocks ('intermediate host rocks') by fluvial processes, storms, waves and currents along the coast can liberate the detrital heavy minerals from the consolidated sedimentary rocks; in this way the heavy minerals are remobilised and re-deposited, this time in coastal sands. Examples of this type of relationship are described from the Murray Basin of southern Australia (Roy et al., 2000; Roy and Whitehouse, 2003), the northern part of the Perth basin in West Australia (Shepherd, 1990), the central coast of Brazil (Leonardos, 1974), and southern India (Ali et al., 2001).

NEOPROTEROZOIC–PALEOZOIC ERA

Officer Basin

The Officer Basin, onlapped the Musgrave Province, Gawler and Yilgarn Cratons and was overlain by the Eucla Basin (Fig. 2.1), is an intracratonic Neoproterozoic to Devonian basin in northwestern South Australia and Western Australia (e.g., Preiss et al., 1993; Gravestock et al., 1995). The northern margin is partly unconformable on the Musgrave Block, but is largely overthrust by the basement, as defined by magnetic, gravity and seismic data. The southeastern margin is defined by the Karari Fault, delineated by magnetic data and forms the western margin of the Gawler Craton. The eastern part of the Officer Basin is overlain by Arckaringa Basin sediments, and the southern part by Denman, Bight and Eucla Basin sediments; the margin in these areas has been mapped with sparse drillhole and seismic data. The Officer Basin was linked to the Warburton Basin in the east and Amadeus Basin in the northeast, and probably to the Stuart Shelf in the southeast (Morton, 1997; Press et al., 2002).

The Officer Basin has been described by a number of authors, such as Jackson and van de Graaff (1981), Townson (1985), Phillips et al. (1985), Brewer et al. (1987), Gravestock and Hibburt (1991), Hocking, (1994), Drexel and Preiss (1995), Harvey and Hibburt (1999), Carlsen et al. (1999), and Ghori (1998, 2002). Thick terrigenous clastic and carbonate sediments of the Neoproterozoic–Paleozoic Officer Basin underlie the Eucla Basin and thicken northwards beyond the Eucla Basin. To the northeast lie the margins of the Carboniferous–Permian Arckaringa Basin and the overlying and younger Mesozoic Eromanga Basin. The Carboniferous–Permian sediments outcrop in the region dissected by paleovalleys (e.g., Tallaringa) and include glaciogenic sediments of the Boorthanna Formation, marine fine clastics of the overlying Stuart Range Formation and coal bearing terrestrial sediments of the capping Mount Toondina equivalents occur around the northeast margin of the basin (as near Maralinga), which were also described around the northwest margin by Lowry (1970).

The Ilma Formation on the northern margin of the basin is a medium-grained sandy oolite with a matrix of fine sparry calcite of probably Neoproterozoic age and thickens northwards into the Officer Basin (Lowry, 1970). The flat-lying Lower Miocene Colville Sandstone of the Eucla margin rests on the Ilma Formation with angular unconformity, which in turn rest unconformably on the basement crystalline rocks of the Eucla Basin further south (Lowry, 1970). Tillite, together with claystone and conglomeratic sandstone of probable Lower Permian age, occur as inliers near the northern margin of the Eucla Basin, and are correlated with the Wilkinson Range Beds of the Officer Basin (Fairbridge, 1953; Wells, 1959; Lowry, 1970), which was replaced by the Paterson Formation (Wales and Forman, 1981).

The Paterson Formation of the Officer Basin (Wales and Forman, 1981) is characterised by poorly sorted sandstone, claystone, conglomerate, tillite, siltstone, basal diamictite; varves and erratics in places — glacigene, lacustrine, to fluvioglacial. The sandstone and claystone exposed at a prominent scarp around the northern margin of the Eucla Basin, north of Lake Gidgi and creeks (at lat. 28°50'S, long. 125°50'E), are tentatively correlated with the Wilkinson Range Beds (Lowry, 1970).

The intracratonic Officer Basin contains abundant clastic sediments and minor, mostly mafic volcanism (e.g. Table Hill Volcanics; Grey et al., 2005). The sediments include tillite sequences of probable Sturtian age. It is likely that the basin was at times contiguous with sediments of the Adelaide Rift Complex to the east and the Gunbarrel Basin in the west (Preiss, 1987; Carlsen et al., 2003). The provenance of clastic sediments within the Officer Basin includes the Musgrave Province and Gawler Craton, with lesser input from the Yilgarn Craton and possibly the Albany-Fraser Orogen (Carlsen et al., 2003; Wade et al., 2005; Bodorkos et al., 2006). Equivalents of the upper Heysen Supergroup of the Adelaide Rift Complex are also widespread in the Officer Basin, and include thick clastic rocks deposited as a result of rapid erosion of the adjacent Musgrave Block, which was then uplifted by intraplate compression during the Neoproterozoic to Cambrian Petermann Orogeny.

The post-Cambrian geologic-geomorphic evolution of the Officer Basin is summarised below:

- Permian glaciation of bedrock.
- Mesozoic deep weathering, with minor planation.
- Establishment of drainage patterns with major valleys several kilometres wide (playa lakes, paleodrainage).
- The break-up of Gondwanaland.
- Cenozoic landscape evolution involving renewed weathering, and duricrust formation.
- Climatic change from a warm and moist climate in the Mesozoic and Cenozoic to a period of aridity involving salt weathering, aeolian deposits, limited drainage and the formation of remnant lake systems.

Permian sediments are also deposited locally in the Denman Basin beneath the southern Eucla Basin (Fig. 2.1; Bendall et al., 2016).

MESOZOIC ERA

Bight Basin

The Bight Basin, including the Ceduna, Duntroon, Eyre, Bremer and Recherche sub-basins, situated along the southern parts of the Eucla Basin (Fig. 2.1), developed during the break-up of eastern Gondwana in the Jurassic and Cretaceous (e.g., Totterdell and Bradshaw, 2004). As one of a series of Mesozoic to Cenozoic depocentres that developed along the southern margin of Australia during the breakup of eastern Gondwana, the Bight Basin is partly overlain by the Cenozoic Eucla Basin and was initiated during a period of Middle–Late Jurassic to Early Cretaceous upper crustal extension (e.g., Willcox and Stagg, 1990; Stagg et al, 1990; Hill, 1995; Totterdell et al, 2000; Bradshaw et al. 2003; Teasdale et al, 2003; Totterdell and Bradshaw, 2004). In the southern part of the Eucla Basin, the Bight Basin, at depths ranging from 200 m (on the shelfal parts) to 4,000 m (in the distant offshore), is overlain unconformably by the dominantly cool-water carbonates of the Cenozoic Eucla Basin. To the south, the sequences of the Bight Basin onlap extended continental crust and rocks of the continent–ocean transition on the abyssal plain between Australia and Antarctica (Sayers et al, 2001). To the north, underlying the Eucla Basin, a thin Bight Basin succession (e.g., Madura Shelf) overlies Precambrian basement, including the Gawler Craton and Albany-Fraser Orogen, which have had a profound influence on the early structural development of the Bight Basin, with basement trends controlling the location and orientation of early basin-forming structures (Stagg et al, 1990; Totterdell et al, 2000; Teasdale et al, 2003; Totterdell and Bradshaw, 2004).

Bremer Sub-basin

The Bremer Sub-basin of the Bight Basin (previously Bremer Basin; Playford et al., 1975), containing the Mesozoic- Cenozoic succession (Cockbain, 1968b), lies to the southwest of the Eucla Basin and south of the Yilgarn Craton and Albany-Fraser Orogen (south of continental shelf edge; Fig. 2.1), where siliciclastic rocks of the Bremer Basin grade laterally into carbonates of the Eucla Basin (Hocking, 1994). A recent reassessment of basin terminology along the southern margin of Western Australia groups the Mesozoic succession into the Bight Basin (i.e., Bremer sub-basin) and the Cenozoic succession into the Eucla Basin (e.g., Clarke et al., 2003; Bradshaw et al., 2003; Totterdell and Bradshaw, 2004).

The Bremer Sub-basin, a structurally complex series of perched half-graben depocentres, contain up to 11 km of predominantly Jurassic and Cretaceous sedimentary rocks overlain by a thin Cenozoic cover from the Eucla Basin (Monteil et al., 2005). The sub-basin sits above Proterozoic rocks of the Albany-Fraser Orogen, extending seawards across the continental slope off the southern coast of Western Australia in water depths of 100 to 4500 m (Totterdell and Bradshaw, 2004). The Eucla Basin extends landward from the northern margin of the Bremer Sub-basin as a thin veneer (<500 m) of Cenozoic cool-water carbonates and siliciclastics overlying the Albany-Fraser Orogen (Hocking, 1994; Bradshaw et al., 2003; Clarke et al., 2003).

Madura Shelf

The Madura Shelf is an onshore component of the expansive Bight Basin that is completely obscured by the Cenozoic sediments of the Eucla Basin, and only known from a small number of drillholes, which provide a fragmentary record and understanding of the Madura Shelf (Fig. 2.1; Lowry, 1970; Reynolds, 2016). The Early to Late Cretaceous sedimentary rocks on the Madura Shelf were deposited in areally restricted non-marine environments in the Early Cretaceous, through to increasingly marine environments followed by a small delta complex in the Late Cretaceous (Totterdell and Krassay, 2003). The clastic sediments of the Madura Shelf, comprise the Madura Formation and the Loongana Formation, which were deposited during the Cretaceous, following the Gondwanan breakup of southern Australia and Antarctica. A 25–60 Ma hiatus followed before sedimentation recommenced in the Eocene with the deposition of the Hampton Sandstone and Eucla Group carbonates of the Eucla Basin (Lowry 1970; Totterdell et al. 2000; Bradshaw et al. 2003; Totterdell and Bradshaw, 2004). Some pre-rift sediments, including the Pre-Cretaceous Decoration Sandstone (claystone overlying sandstone) and Shanes Dam Conglomerate, occur as irregular unconformable deposits between the Madura/Loongana Formations and Precambrian basement (Lowry, 1970; Reynolds, 2016). Cretaceous strata of the eastern Madura Shelf extend over older Neoproterozoic, Permian and Jurassic strata of the Polda Basin, and a thin erosional remnant of Permian glacial deposits, previously mapped as the Denman Basin (Totterdell and Krassay, 2003).

Recently, Barham et al. (2015) interpreted that the Mesozoic sediments of the onshore extension of the Bight Basin (Madura Shelf) are overlain by the expansive Cenozoic carbonates of the Nullarbor Plain of the Eucla Basin. Although these sediments are more accessible than those of contemporaneous offshore sub-basins within the Bight, they remain poorly studied. The Mesozoic sequence of the Madura Shelf underlying the central Eucla Basin may testify to pre-existing depocentres along the southern margin of Australia that influenced development of the Eucla Basin.

Lowry (1968a, 1970) divided the Mesozoic sediments beneath the Eucla Basin into the Loonganna Sandstone and the overlying Madura Formation. These sediments are mostly confined to the more central parts of the basin and overlie Precambrian basement (e.g., in Transcontinental Railway No. 3 Bore/ Loongana Bore; lat. 31°02'S, long. 127°03'E). (Ludbrook, 1969; Lowry, 1970). The Loonganna Sandstone is a feldspathic arenite of up to 30 m thick, and possibly equivalent sediments in the eastern part of the basin are ‘grey gravels, quartz grits and pyritic sandstones with thin mudstone bands’ (Ludbrook, 1958). The Madura Formation (up to 300 m thick) is of mostly carbonaceous and glauconitic mud, shale and sand, which is more feldspathic in the lower sands and distinctly more glauconitic in the uppermost beds locally (Lowry, 1970). These sediments are of Cretaceous age (Ingram, 1968), and the distinctiveness of three polynomorph assemblages, suggests two major hiatuses, in the Aptian, Albian and Onomanian to Semonian (Lowry, 1970). Frakes et al. (1987) have provided postulated sea level maxima for a number of time slices in the Cretaceous. As shown in several wells (e.g., Transcontinental Railway No. 1, Ganibanga No. 1, and Eyre No. 1), the Madura Formation is overlain by the Hampton Sandstone of the Eucla Basin but in areas where the Hampton Sandstone is not developed, the Madura Formation is overlain by the Wilson Bluff Limestone of the Eucla Basin (e.g., Transcontinental Railway No, 3 Bore and Eucla No.1 Bore; Lowry, 1970). The Madura Formation is underlain either by the Loongana Sandstone (e.g., the Transcontinental Railway No. 3 Bore) or by Precambrian basement (e.g., Eyre No. 1 Well; Lowry, 1970).

Polda Basin

The Polda Basin is a narrow, elongate, east-west trending trough that underlies the continental shelf of the Great Australian Bight in water depths of up to 500 m and extends onshore to the eastern margin of the Eucla Basin in the central Eyre Peninsula, South Australia, containing Jurassic fluvial sandstone, overlying Permian and Neoproterozoic sediments (Fig. 2.1). The Polda Basin has formed during several phases of tectonism up to and including the Mesozoic rifting of Australia and Antarctica, although it originated as a Neoproterozoic intracratonic feature (Drexel and Preiss, 1995). Three distinct depocentres, separated by faulted shallow basement, occur in the Polda Basin: the eastern depocentre (1.5 to 2 km), the central depocentre (Neoproterozoic to Jurassic section, ~5 km), and the western depocentre (poorly defined but may be a rift-splay from the Bight Basin containing only Mesozoic rocks; Drexel and Preiss, 1995). Sediments deposited in the Polda Basin include Neoproterozoic redbeds interbedded with evaporite deposited in arid fluvial to playa lake environments, Permo-Carboniferous glaciogenic sediments, and Upper Jurassic fluvial sandstone and coal (Cooper et al., 1982; Parker et al., 1985; Flint, 1989; Flint and Rankin, 1989).

Eromanga Basin

In the Triassic, most of southern Australia was subjected to weathering and erosion, and sedimentation only occurred in a few places. In the Jurassic and Cretaceous, a time of profound geographic change for Australia, crustal rifting beginning in the Jurassic led to the separation of Australia from Antarctica, and the resulting seaway south of the Australian continent widened as Australia drifted northwards. This represented the latter stage of the disintegration of Gondwana. Meanwhile, global sea-level rise brought the last great marine flooding onto the Australian continent from the north, and this shallow sea covered much of southern Australia's inland.

Due to the extensive burial of the Eromanga Basin, uncertainty remains as to where to draw the boundary between the marginal deposits of the Eucla Basin and those of the Eromanga Basin to the northeast (Fig. 2.1). In the area now partly covered by the inland extent of the northeastern Eucla Basin, during the Jurassic and Cretaceous, crustal downwarping was accompanied initially by the deposition of a very widespread fluvial and lacustrine to marginal marine clastic blanket. This includes outcrop of largely Algebuckina Sandstone and Cadna-owie Formation of the Eromanga Basin, which are potential host units for heavy mineral sands (Hou et al., 2021).

2.3 EUCLA BASIN SETTING AND PALEOVALLEYS: PALEOGENE–NEOGENE

The Cenozoic Eucla Basin is a large onshore-offshore basin on the southern Australian passive margin covering a vast area stretching ~2,000 km from the southwestern margin to the southeastern margin, and northward through to the boundary of the Nullarbor Plain with the Great Victoria Desert (Figs 1.1 and 2.1). It adjoins the Yilgarn and Gawler cratons and Albany-Fraser province and is separated from the Musgrave Province to the north by the Officer Basin, including offshore extensions to the platform edge at the continental shelf break that was regarded as being a supracratonic extension of the Great Australian Bight Basin to the south (Lowry, 1970). This large marine depocentre, along the southern continental margin of Australia during Paleogene–Neogene time (Alley et al., 1999) excluded the Paleozoic–Mesozoic records from the Eucla Basin and named new basins (e.g., the Denman and Bremer basins; Fig. 2.1), extends some 350 km inland from the present coastline of southwestern Australia and extends seaward approximately 150 km to the approximate edge of slope, ~500 km wide from north to south.

The large areal extent of the Eucla Basin and adjacent paleovalley system encompasses a complex succession of marine and non-marine strata, making correlation of stratigraphic, sedimentary and tectonic events non-trivial (see Chapter 4). Offshore, Cenozoic cool-water carbonates (Feary and James, 1995) of the Eucla Group overlie Cretaceous passive margin sequences (Bight Basin), the latter being deposited within half grabens developed as a result of the separation of Australia and Antarctica (Veevers and Eittreim, 1988). The onshore paleovalley sediments unconformably overlie Permian and Mesozoic sediments, Neoproterozoic–Cambrian Officer Basin sandstones, and Precambrian basements of the Yilgarn Craton, Gawler Craton and Albany-Fraser Orogen (Clarke et al., 2003; Hou et al., 2008, 2011b). The present landscape of the Eucla Basin and adjacent paleovalley system is dominated by extensive tracts of Quaternary deposits largely superimposed on the Paleogene to Neogene morphology, but the excellent preservation of the paleomorphology is evident from regional topographic data and from satellite images (Figs 2.2, 2.3 and 2.4). Features include large-scale paleovalleys, lagoons, estuaries, and coastal barriers. The pattern of paleovalleys in the onshore basin margin is dominantly sub-dendritic, reflecting both the pre-Cenozoic land surface gradient, and bedrock lithology and structure (Alley et al., 1999; Hou et al., 2001a; 2003a; de Broekert and Sandiford, 2005). On the eastern Yilgarn Craton, many trunk paleovalleys are parallel to the northeast orientation of a regional fracture field (Johnson and McQueen, 2001) while on the western Gawler Craton, paleovalleys preferentially cut into and follow weakly resistant, deeply weathered bedrock (Hou et al., 2003a).

The Eucla Basin, comprising a thin passive margin succession, was deposited during the final stages of separation between the Australia and Antarctica plates. Sediments of the Eucla Basin were deposited unconformably over pre-Paleogene landscapes, including the Precambrian basement, Paleozoic (e.g., Officer Basin) and Mesozoic (e.g., Eromanga, Polda, Bremer, and Bight Basins) bedrocks (Figs 2.1 and 2.5). The sequence boundary at the base of the Eucla Basin succession is associated with considerable weathering, erosion, incision and, in some cases, canyoning. The Wobbegong Supersequence was deposited on the shelf during an overall transgressive phase of sedimentation. This was followed by the accumulation of the cool-water carbonates of the Dugong Supersequence, which were deposited in rapidly increasing water depth during the fast-spreading episode that commenced in the Late Eocene (Geoscience Australia, web). Rapid northward drift and increasing Southern Ocean circulation during the Eocene resulted in the accretion of about 1 km-thick, largely cool-water, carbonate shelf (see Chapter 4). During this period the margin was essentially tectonically stable.

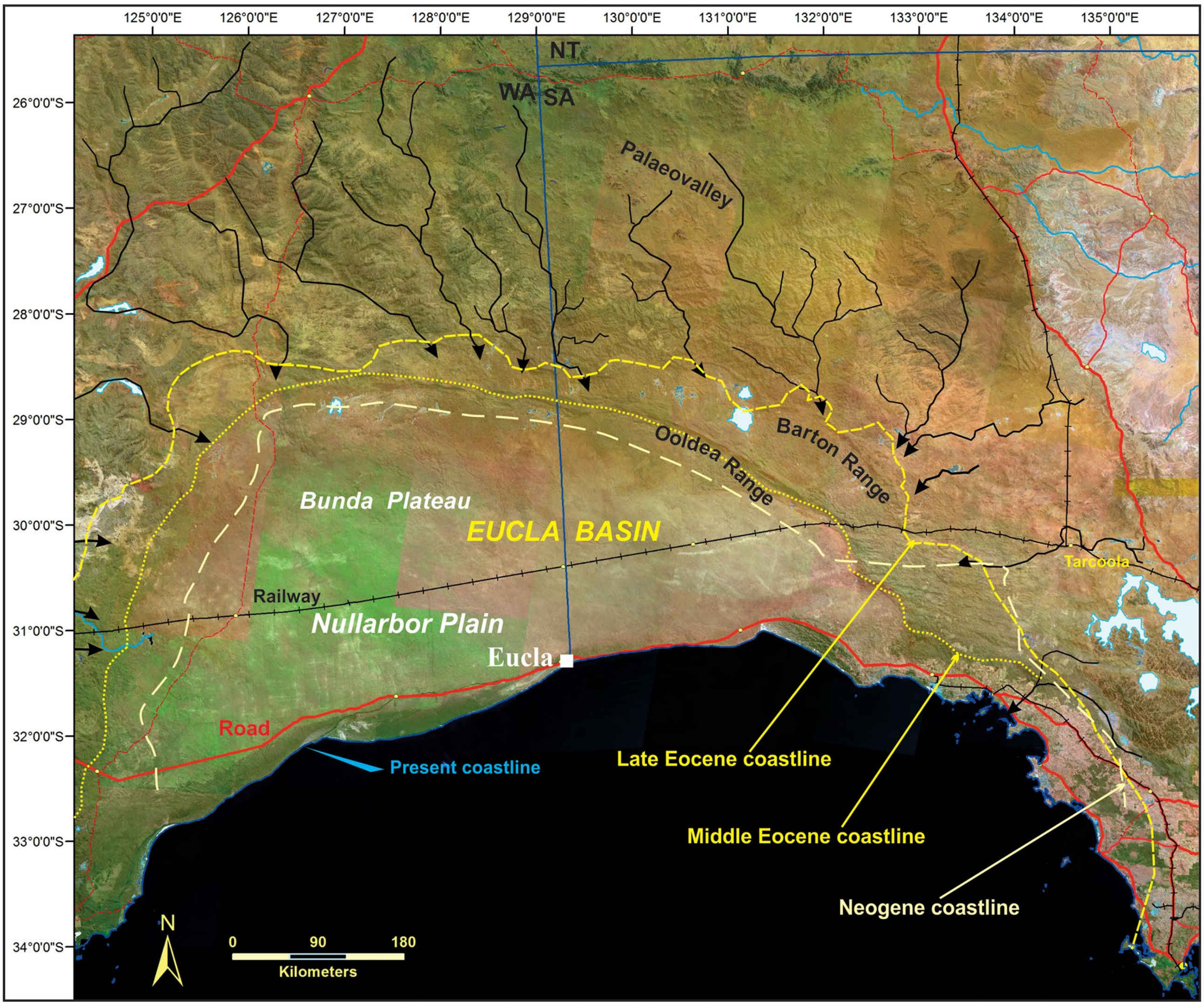


Figure 2.2 Landsat imagery of the Eucla Basin showing central basin, major paleovalleys and shorelines of varying age (modified from Hou et al., 2021).

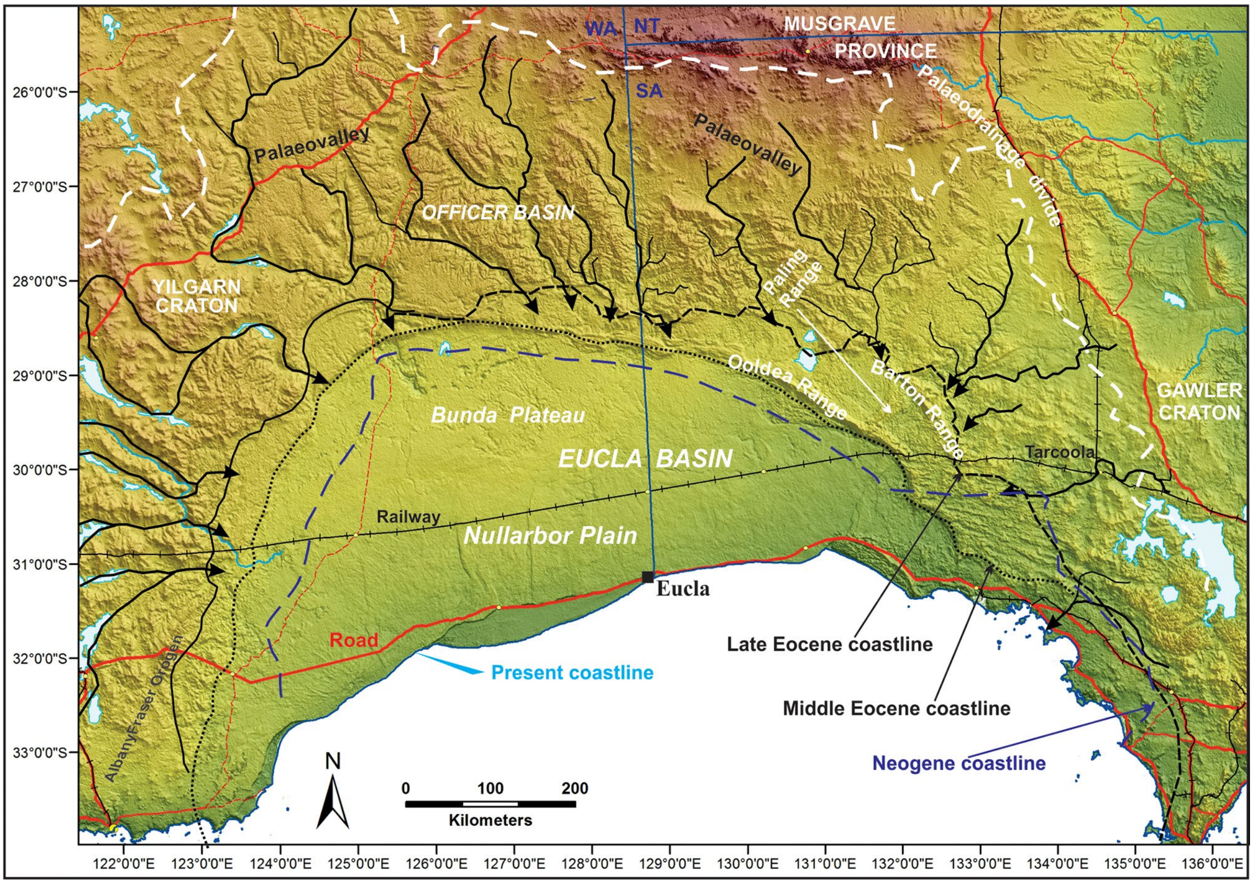


Figure 2.3 SRTM DEM imagery of the Eucla Basin showing central basin, major paleovalleys and shorelines of varying age (modified from Hou et al., 2021).

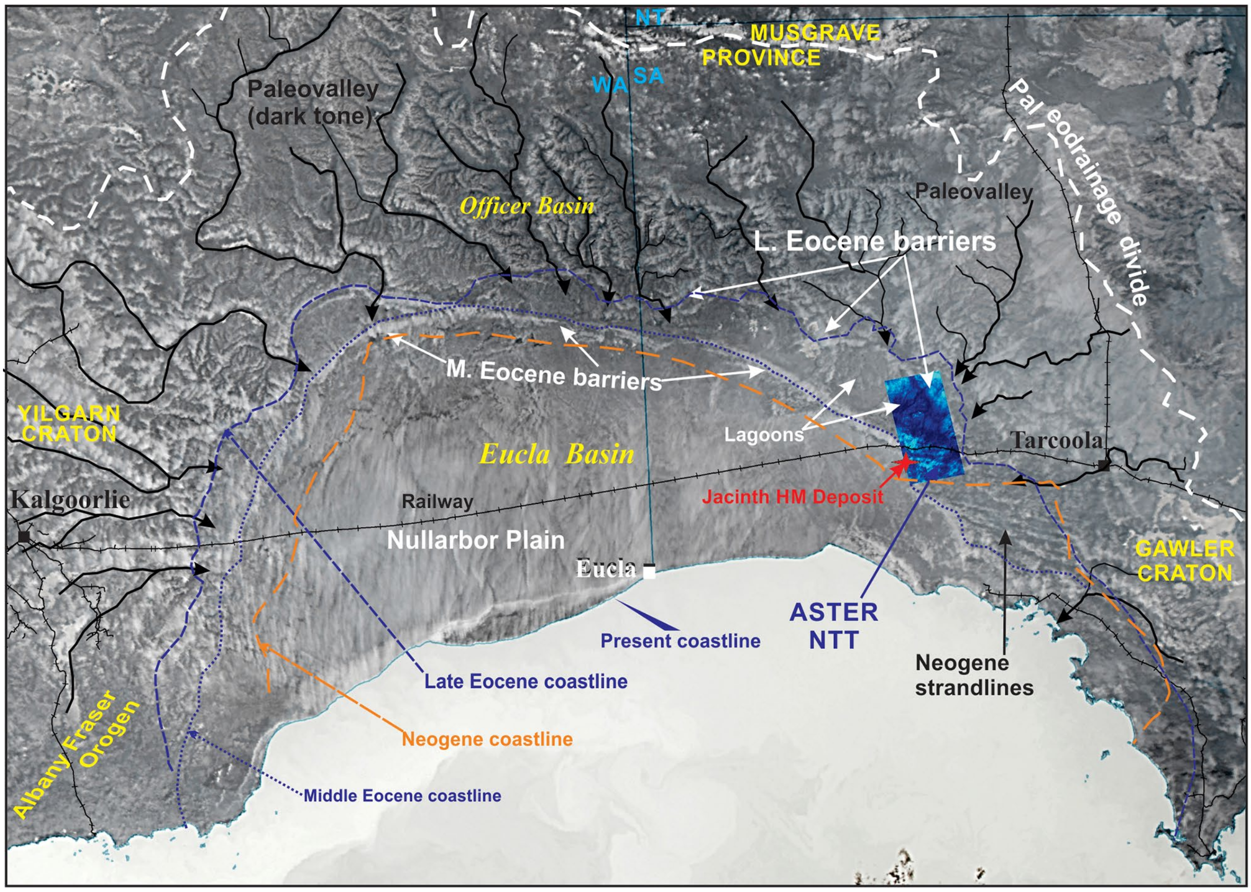


Figure 2.4 NOAA-AVHRR (1.1 km pixel) and ASTER (90 m pixel) night-time thermal images of the Eucla Basin showing major paleovalleys and shorelines of varying age;

the textural difference between paleovalleys (dark tone) and the sand barriers (light tone) is apparent (modified from Hou et al., 2011b and 2021).

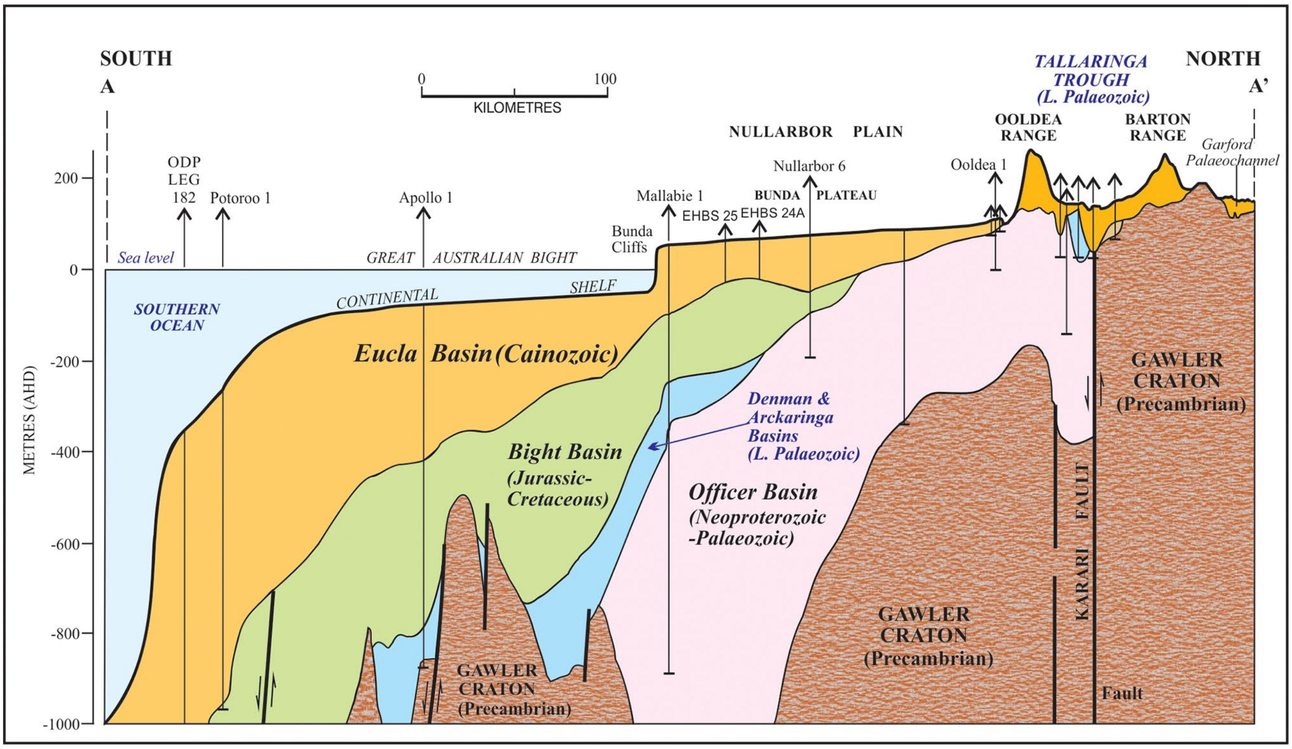


Figure 2.5 Cross-section through the eastern Eucla Basin, showing geologic relationships with underlain cratonic and bedrock geologies; location of the cross-section indicated in Figure 1.1 (from Hou et al., 2021).

The relationship of the Eucla Basin to the Great Australian Bight Basin, is that of ramp or platform to basin (Benbow et al., 1995a). Previously, the landward extent of the basin was believed to coincide with that of the Bunda Plateau or Nullarbor Plain (Tate, 1879; Brown, 1900; Ludbrook, 1958, 1969; Lowry, 1970; Lindsay and Harris, 1975; Firman, 1975, 1978), so that the Eucla Basin was defined on the extent of Paleogene marine limestones of the Eucla Group (Singleton, 1954), and later defined as the margin at the limit of Miocene terrigenous sediments as marked by a prominent scarp along the basins northern margin (Lowry, 1970). However, later studies in the eastern basin showed there are very extensive Paleogene–Neogene sediments beyond the generally accepted margin of the Eucla Basin, that is beyond the Bunda Plateau and that the basins margin can be extended for a further at least 50–100 km to the northeast and east (e.g., Benbow, 1982, 1990a, b; Benbow et al., 1995a).

Around the margin of the Bunda Plateau the basin margin should be also extended further inland to include the margin settings such as Neales Plateau and Barton Range at least (see Chapter 3). Around its landward margin, the Eucla Basin contains remarkably preserved Cenozoic coastal barrier island features and peripheral paleovalley systems. The central part of the basin is characterised by a carbonate platform, named the Nullarbor Plain, a distinctive, remarkably flat, and treeless landscape extending some 1,400 km along the southern Australian margin (see Chapter 4).

Paleogene and Neogene climate changed from high-rainfall temperate or cool temperate to warmer in the Paleogene, but rainforest persisted, and arid conditions increased in the Neogene, with the development of open woodland with minor forest pockets (based on palynological data and interpretation; see Chapter 4). Prolonged weathering of a generally subdued and tectonically stable landscape during much of the arid Neogene concentrated iron oxides and silica in hard layers of ferricrete and silcrete within the near-surface portions of exposed rocks at various times.

It is the extreme resistance to erosion of these weathering materials that has given rise to the plateaus, mesas and breakaways typical of our outback landscape. Paleogene and Neogene sediments and weathered rocks contain important groundwater, coal, uranium, construction material, and industrial mineral resources (see Chapter 5 Economic geology).

The Eucla Basin owes its distinctive landscape to a unique set of interactions between eustatic, climatic and tectonic processes over the last ~50 Ma (James and Bone, 1994; Benbow et al., 1995b; Hou et al., 2006b, 2008). The result is one of the largest onshore areas of Cenozoic marine sediments anywhere in the world (Benbow, 1990a; Clarke et al., 2003), with a remarkably preserved paleoshoreline sequence along the northeastern basin margin that is highly prospective for heavy mineral sand deposits (Benbow, 1990b; Hou et al., 2003b). Open marine conditions were not established over the whole of the Eucla Basin until accelerated spreading in the Southern Ocean in the Middle Eocene (McGowran et al., 1997). At this time, predominantly terrigenous to shallow marine sediments such as the Hampton Sandstone and Pidinga Formation were covered by transgressive carbonates of the Wilson Bluff Limestone (see Chapter 4). It is this Middle Eocene transgression that pushed the shorelines far into the continental realm and resulted in the development of subsequent transgressive shorelines, including major coastal barrier complexes, in which significant deposits of heavy mineral sands were formed. Understanding of the history of the basin and the paleovalleys that drained from the surrounding cratons is important because they contain major mineral deposits, and the sediments derived from them contain remobilised gold, uranium, and heavy minerals. In particular, a remarkably preserved paleoshoreline sequence along the margin of the Eucla Basin is highly prospective for heavy mineral sand deposits. The extent and complexity of the paleoshoreline deposits hampered early heavy mineral exploration.

The presence of Eucla paleovalley features has been previously documented along the Eucla Basin margin (e.g., Bunting et al., 1974; Barnes & Pitt, 1976; Van de Graaff et al., 1977; Pitt et al., 1978; Jones, 1990; Clarke, 1993, 1994a, b; Benbow et al., 1995a; Alley et al., 1999). The Eucla paleodrainages have their headwaters in the Yilgarn and Musgrave Blocks, Officer Basin and the Gawler Craton. Late studies have suggested that the western and eastern Eucla paleovalley systems are comparable in history and lithology (e.g., Alley et al., 1999; Clarke and Hou, 2000; Clarke et al., 2003; Hou et al., 2008). The Eucla paleovalleys are defined by the connection of sedimentary and geomorphological evidence that existed between the Eucla Basin and inland paleovalleys. These paleodrainage systems are widespread in the onshore Eucla Basin and are now present in the form of a subtle, branching depression largely obliterated by the Quaternary aeolian sands of the Great Victoria Desert. Their best-preserved sedimentary infillings are found in the Eucla Basin and peripheral paleovalleys. The roughly mapped paleodrainages in the onshore basin (e.g., Benbow, 1983a, 1993; Cowley & Martin, 1991; Rankin et al., 1996; Clarke, 1993, 1994a, b; de Broekert, 2002; de Broekert and Sandiford, 2005) and on the 1:1,000,000 map sheet of South Australian paleochannels (Rogers, 2000), comprise channels, wetlands and in later stages, shallow alkaline lakes. The dimensions of the channels vary greatly, with widths of the river valleys ranging from a few tens of metres to more than 30 km and depths of up to 100 m. The principal direction of drainage is towards the central basin but there is considerable variation (see Chapter 3). Locally, some channels show irregular traces on the map because of the low gradients. The paleochannel mouths are characterised by a series of estuaries (e.g., Wilkinson and Anthony). Between the coastal barriers (the Ooldea and Barton Ranges) lies NW-SE trending lagoons. The Paleogene–Neogene sediments in the paleochannels and adjacent marginal Eucla Basin infill the sites of paleorivers and adjacent depressions carved into much older sedimentary rocks and basement of the interfluves.

The geologic-geomorphic evolution of the Eucla Basin is summarised as below (for details see Chapters 3 and 4):

- Pre-Cretaceous erosion (possibly fluvial) causing development of irregular topography in onshore and offshore areas of the Eucla Basin, which was subsequently buried by Cretaceous sediments.
- The subsidence of irregular basin areas probably commenced during Late Cretaceous time with terrestrial and shallow marine sediments accumulated, the deposition and downwarping recommenced in the Eocene.

- Eocene downwarping and sedimentation within the Eucla Basin and peripheral paleovalleys developed until the Eocene seas retreated and weathering and erosion recommenced.
- Early Miocene marine transgression followed, due probably to further downwarping of the central Eucla Basin.
- Middle to Late Miocene downwarping and sedimentation within the Eucla Basin and peripheral paleovalleys developed until the Miocene seas retreated and weathering and erosion recommenced.
- Pliocene sedimentation in places of the Eucla margin and peripheral paleovalleys developed until the Pliocene seas retreated and weathering and erosion recommenced.
- Late Cenozoic weathering, erosion, regolith development and landscape evolution.

2.4 COVER – QUATERNARY/GEOMORPHOLOGY

Most of the area is today covered by Quaternary sand plains and dunes, which are superimposed on and obscure the paleo-features, indicating that the landscape of the Eucla Basin and surrounding region contains remarkably well preserved and varied paleolandforms of Paleogene-Neogene ages (Benbow, 1990a). One of the significant features in the area is the geomorphology of a string of playa lakes and lunettes developed in topographic lows. These generally reflect underlying Eucla paleodrainage systems, although relief is subdued and drainage is generally disorganised. Other obvious features in the eastern Eucla margin include the parallel Ooldea and Barton Ranges, and the interconnecting Paling Range, well preserved coastal dunes or barriers of late Eocene or early Oligocene age (Benbow, 1989, 1990a; Clarke and Hou, 2000). These remarkable landforms can be interpreted by considering their topographic relief, by applying the digital elevation models, and by interpreting the Landsat and NOAA images. A history of their recognition and delineation is described in Benbow (1989), and their origin is discussed in Benbow (1990a) and Clarke and Hou (2000). The large and extensive ranges have been interpreted as coastal dunes that formed along the northeastern margin of the Eucla Basin during the highstands of two Eocene transgressions. Other morphological components such as preserved headlands and embayments are also recognisable in the Eucla coastal area (see Chapter 3).

The Quaternary is characterised by marked climatic and sea-level oscillations, which have impacted on many landscape features. In the interior, cold, arid phases were times of sand dune building and evaporitic playa deposition, which alternated with periods of moister climate, increased stream discharge and greatly extended lake systems, when extensive alluvial and fluvial sand, clay and gravel were deposited. Periods of weathering have formed hard carbonate soil profiles (calcrete) and a variety of solution features such as caves in carbonate rocks. In the most recent of the Quaternary sea-level oscillations, the sea advanced across the continental shelf, reaching its present level about 7,000 years ago, since which time it has remained relatively stationary.

2.5 TECTONIC AND STRUCTURAL FRAMEWORKS

Although the Nullarbor Plain has traditionally been considered a tectonically quiescent domain, very few early studies considered the onshore evidence of Cenozoic tectonic activity (Hillis et al., 2008). The sedimentation of the Eucla Basin began after the Late Cretaceous and there were periods of gentle downwarping mainly in the Middle to Late Eocene, and Miocene, followed by uplift with continental tilting and faulting mainly in Pliocene (see Chapter 4). Faulting during the sedimentation of the basin seems unlikely because of the absence of any sign of subsequent faulting affecting the strata of the Eucla Basin and of the relatively gentle slopes on the basin margins. Despite of the lack of tectonic disturbance of the strata on a large scale in the Eucla Basin, there is evidence for the existence of numerous faults up to 120 km long with displacements of up to about 60 m. SRTM DEM images reveal numerous linear scarps on the Nullarbor Plain, ranging up to 20 m in height, indicating fault scarps that have significant amount of vertical displacements (Fig. 2.3; Lowry, 1970; Sandiford, 2007; Hou et al., 2008). For example, it is evident that two scarps are obviously associated with tectonic deformation (Lowry, 1970): a scarp (35 km long and up to 20 m high) trending northward and northwestward from the Madura landing ground; and other scarp lying about 25 km north-northwest of Caiguna (at about lat. 32°03'S, long.

125°22'E). A gravity survey by Blumer and Webb (1965a, b) also indicated a northerly trending anomaly in this region, suggesting a trough of sedimentary rocks.

Based on digital elevation model, Clark et al. (2011) recognised a number of previously unidentified faults in the Nullarbor region and considered that the maximum vertical displacements had not exceeded a few tens of metres over the last 15 million years, indicating the continuing tectonic stability of the local Proterozoic basement. In terms of the shortage of local stress field data from the Nullarbor region and the roughly N-S orientation of numerous fault scarps, Hillis et al. (2008) proposed that the deformation occurred in response to the regional-scale, E-W oriented maximum horizontal stress that has been inferred from plate boundary force modelling. The Mundrabilla Fault occurring in the south-central region of the Nullarbor Plain, is thought to be lithospheric-scale and it may be sustaining a large differential stress (Mounsher, 2016) that was corresponding on large-scale Precambrian motions on the continent-scale N-S oriented Lasseter Shear Zone (LSZ, the deep shear crossing Australian landform and underlying the Mundrabilla Fault; Braun et al., 1991; D'Ercole and Lockwood, 2004).

The Eucla Basin has yielded important evidence of continent-scale Cenozoic crustal deformation. Jennings (1963) suggested that the present surface of the Bunda Plateau represents the virtually unmodified Early Miocene sea floor and concluded that the slope of the plateau (about 1 m per 3 km) corresponds to the amount of post-Early Miocene tilting. However, this calculation ignores the differential erosion of the plateau and the differences in depth of deposition in different parts of the basin (Lowry, 1970). Sandiford (2007) presented a distinct latitudinal asymmetry in the geomorphology of the Australian landmass by showing a number of irregularities in the sedimentary record, implying that significant continent-scale tilting transpired throughout the Cenozoic (SW up, NNE down) generated a total differential vertical displacement of as much as 300 m over the last 15 million years. This apparent tilting uplift of the southwestern margin of the Eucla Basin is also evidenced by variability in the present-day elevations of coeval early Neogene paleoshorelines, which decrease eastward by approximately 150 m over a 760 km interval in the onshore Eucla Basin (Quigley et al., 2010; Sandiford, 2007). Cenozoic marine sediments of the Eucla Basin have been preserved up to 250 m above the present-day sea level. Based on the comparison of the Late Eocene marginal marine spicule facies deposition between the western and eastern onshore basin, it is indicated that the western Late Eocene Eucla margin has been tilted up at least 130 m compared to the eastern Late Eocene Eucla margin (Ruperto et al., 2006; Hou et al.,2008; see Chapter 4).

3. PALEO-PHYSIOGRAPHIC SIGNATURES

The landscape of the Eucla Basin and paleovalley region, extending more than 150 km offshore and about 350 km inland from the coastline, contains remarkably well preserved and varied paleoforms of Cretaceous–Neogene age. The Eucla Basin and surrounding country contain an assemblage of landforms of different origins and all of some paleo-features, such as the marine Bunda Plateau, Nullarbor Plain, scarps/ cliffs/ sand ranges (e.g., Hampton, Nundroo, Ooldea, Paling and Barton Ranges), coastal/ estuarine plains/ embayment/ lagoons (e.g., the Roe Plains, Israelite Plain, Immarna Lagoon, Tietkens Plain, Wilkinson-Anthony Estuary, Neales Plateau/ Embayment), continental shelf (the Eucla Shelf), and a number of paleoshorelines and paleovalleys (Fig. 3.1). They mostly consist of Paleogene–Neogene marine and/or non-marine sediments. These sediments have a much greater extent but are largely covered by aeolian sand plains of Quaternary age including the dune field of the Great Victoria Desert.

3.1 BUNDA PLATEAU – NULLARBOR PLAIN

The Bunda Plateau (Tate, 1879), comprising Quaternary sand-covered plains around the Nullarbor Plain (see below), was commonly recognised as defining the emergent part of the Eucla Basin (Fig. 3.2; Ludbrook, 1958; Lowry, 1970). It is essentially an emergent carbonate ramp or platform that dips very gently (0.6 m/km) south and southeastwards and was formed by flat-lying Paleogene–Neogene marine strata (Lowry, 1970). The Bunda Plateau is almost entirely covered by a thin and continuous layer of horizontal Nullarbor Limestone. This covers a huge area, including the Nullarbor Plain, which extends from an elevation of about 245 m at its northern margin to between 60 and 120 m at the southern boundary, marked mostly by vertical cliffs 40–90 m high, which face the Southern Ocean (Benbow, 1990a). Of the Nullarbor Plain, some have viewed this surface as being essentially a sediplain or exposed Neogene sea floor, which is believed to be largely true, whilst an alternate view has been that it is an eroded surface (Lowry, 1970 and references therein). On the north margin of the Bunda and Neale plateau (see below) there is a 25 m high scarp of possible marine origin (Benbow, 1988). As one of the most featureless large areas of the Earth's land surface, the extreme flatness of the Bunda Plateau is due to characteristics of the Neogene limestone (i.e., Nullarbor Limestone) that forms the entire surface of the Bunda Plateau (Lowry, 1970).

NULLARBOR PLAIN

The Nullarbor Plain (Delisser, 1867), one of the world's large arid to semi-arid karst terrains (<150–400 mm rainfall; 2,000–3,000 mm evaporation), has been long renowned as a vast and remarkably flat tree-less surface (Lowry and Jennings, 1974 and references therein; Fig. 3.2). It forms part of the Bunda Plateau that includes Quaternary sand covered plains around the Nullarbor Plain and has been commonly envisaged as defining the extent of the Eucla Basin (Lowry, 1970). The surface the Nullarbor Plain slopes very gently seawards from 240 m in the northwest and terminates abruptly at the Great Australian Bight in a cliff-line (its southern margin) mostly marked by vertical cliffs 40–90 m high that extends more or less continuously for nearly 900 km, facing the Southern Ocean (Fig. 3.2). The cliffs fall sheer into the sea except in two areas in the centre and west, where there are coastal plains (Roe and Israelite Plains respectively, see below).

The Nullarbor Plain is the surface outcrop expression of the Eucla Group marine limestones of the Eucla Basin (see Chapter 4), in particular of the Nullarbor Limestone. These sediments have a much greater extent, however, being covered around the margin of the Nullarbor Plain, by aeolian sand plains of Quaternary age including the dune field of the Great Victoria Desert. Thus, the Nullarbor Plain and adjacent sand plains are part of a larger physiographic entity (>200,000 km^2 in area) in the Bunda Plateau of Tate (1879). The surface of the Nullarbor Plain, largely covered with a layer of calcrete, is not completely flat in places, rising and falling several metres between clay-floored depressions, up to 1,000 m wide, separated by stony ridges of the same width that are frequently aligned parallel to jointing in the underlying limestone (Lowry and Jennings. 1974; Benbow and Hayball, 1992). The karst terrain in the Nullarbor Plain is one of the largest arid, semi-arid karsts of the world.

On the surface of the Nullarbor Plain there are numerous subdued circular depressions called dongas, ridges, and in the southern part dolines, many of which connect with underlying caves (Lowry and Jennings, 1974). The rises have a thin clay soil with abundant limestone fragments up to boulder size, and rare pavements of the underlying Nullarbor Limestone. Depressions in the plain are occupied by clay flats, and in some areas these are all elongated in a particular direction, while in other areas they are circular but form parallel chains. The caves have been variously categorised and those, such as Cocklebiddy Cave on the central southern part of the Bunda Plateau, are some of the longest in the world. Several degraded fault scarps with a relief of 6 to 9 m occur across the Nullarbor Plain (Fig. 3.2). The vegetation, often with abundant grasses after rain, consists mainly of low shrubs of bluebush (*Kochia seclifolin*) and saltbush (*Atriplex* spp.) (Lowry, 1970).

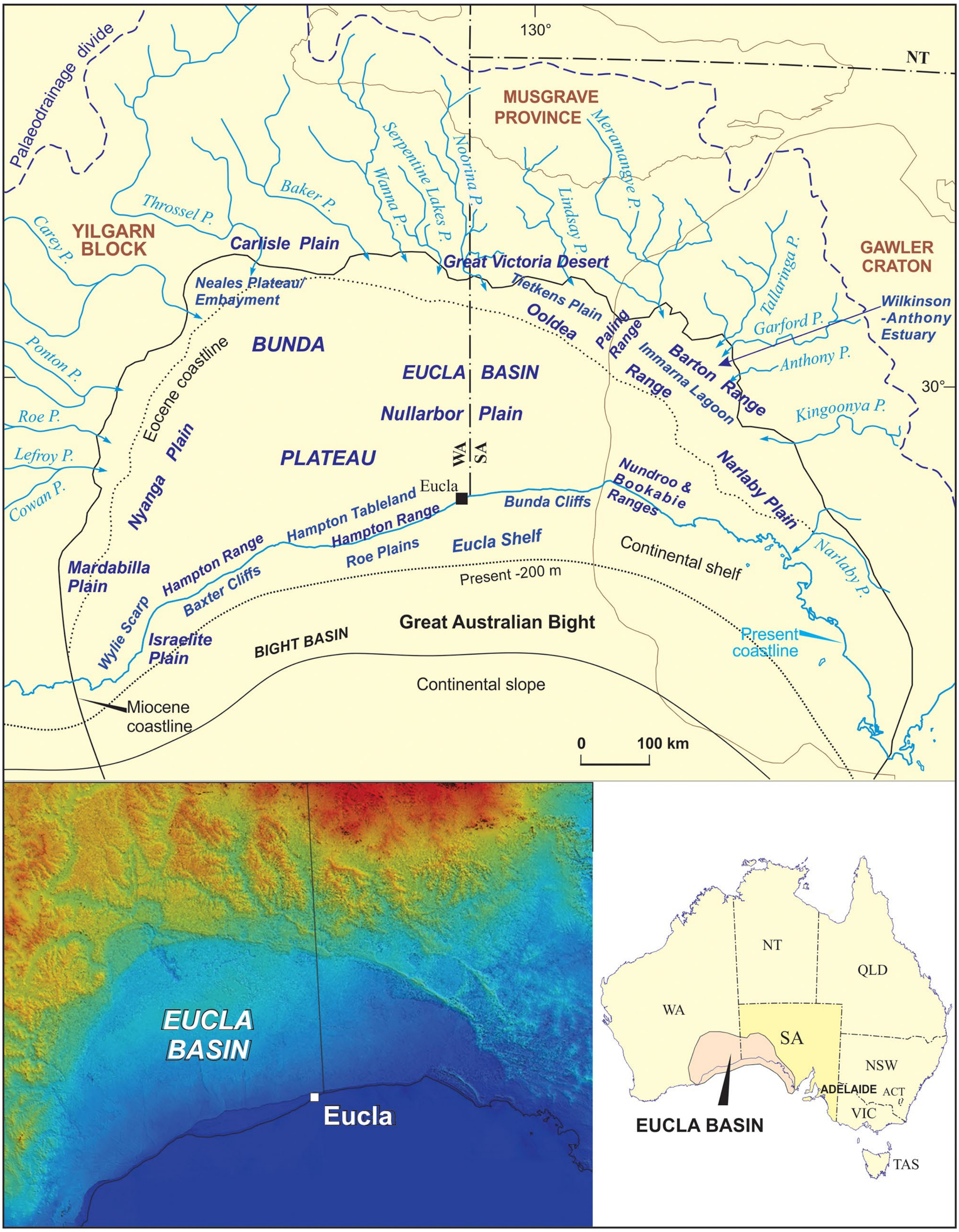


Figure 3.1 Physiographic regions of the Eucla Basin (after Lowry 1970; Jackson and Van de Graaff 1981; Benbow 1990a; James et al. 1996; Hou et al. 2008, 2011).

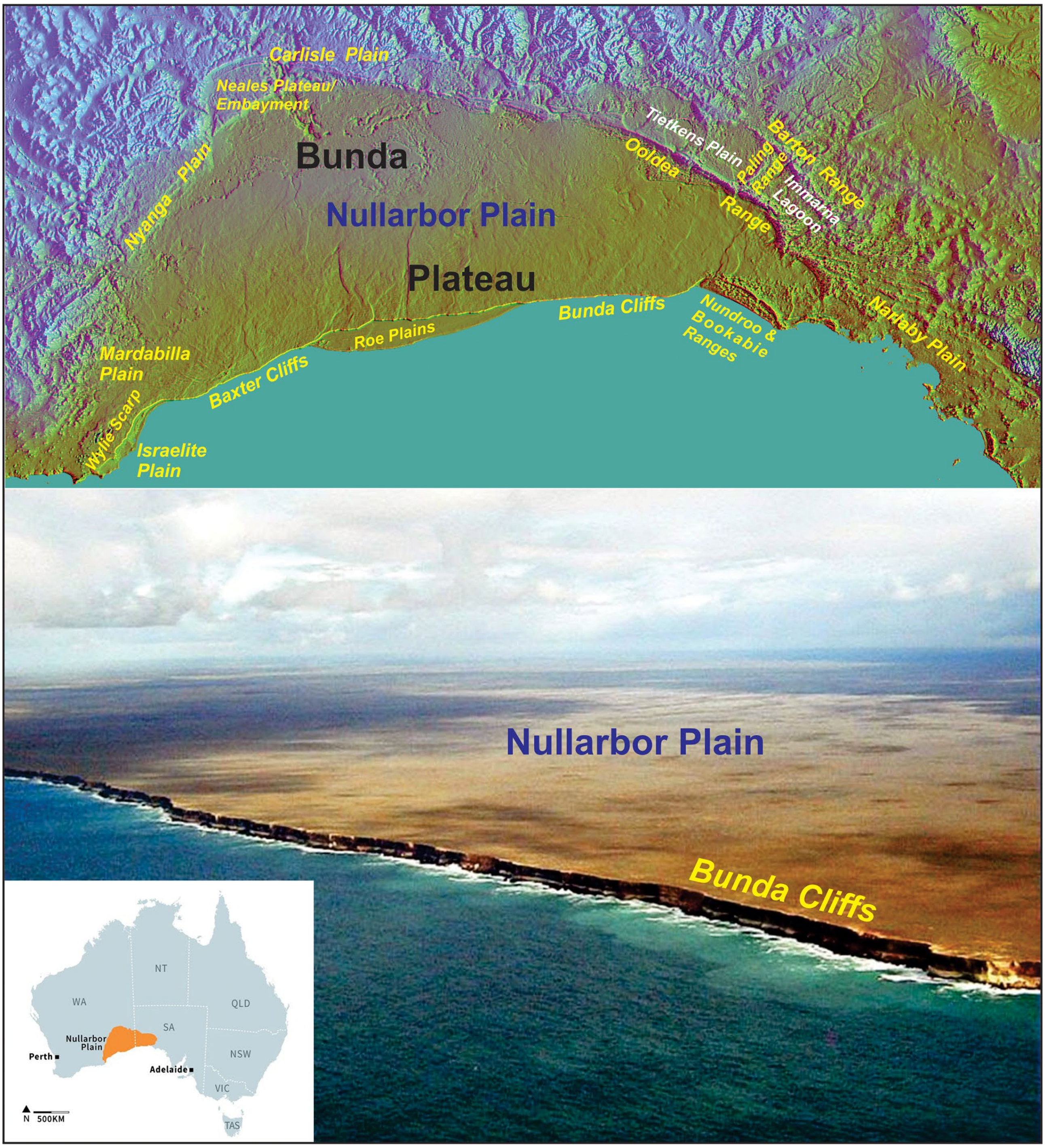


Figure 3.2 DEM imagery (M. Sandiford) of the Eucla Basin showing major physiographic features and photograph (M. Sandiford) of the Bunda Plateau and Nullarbor Plain showing the vast, flat and treeless nature.

Spectacular features of the Nullarbor Plain include its dolines and caves, such as Abrakurrie, Mullamallang, Thampanna, Old Homestead, Koonalda, Cocklebiddy, Weebubbie, Warbla, Bunabie Blowhole, Bildoolja, Koomooloobooka, Koonalda, Weekes, Clay Dam, Wigunda, The Catacombs, Knowles, Murrawijinie, Ivy, New, White Wells, Jimmies, Disappointment and 3 Diprose Caves (Lowry and Jennings, 1974). More than 150 collapse dolines occur mostly within 60 km of the present coast, as steep-sided, closed depressions 2–35 m deep and 10–240 m across (Lowry and Jennings, 1974). Many have been degraded and eroded since the original collapses, resulting in weathered sides and partial infill with sediments. Over 100 caves, also mostly restricted to the present coastal belt, have been recorded. These vary in depth and form and have significant passage lengths, of which the biggest, Abrakurrie Cave, is 300 m long, 30 m wide and 15 m high

(Lowry and Jennings, 1974). The deep caves that extend 50–120 m below the surface of the plain are almost all formed in the Wilson Bluff Limestone, whereas some shallow caves (<30 m deep) formed either entirely within the Nullarbor Limestone or extend downwards into underlying limestones (Lowry and Jennings, 1974).

HAMPTON TABLELAND

The Hampton Tableland is a feature that lies between the Nullarbor Plain and the Hampton Range (Fig. 3.1). The Hampton Tableland is about 98 m above sea level and is characterised by stony tree-covered ridges (usually standing 4.5 to 9 m above the flats with a northeasterly trend) and enclosed grassy clay flats (Lowry, 1970). The crests of the ridges are commonly spaced 180–915 m apart. Pavements of limestone that contain abundant fragments of limestone and kankar are common on the ridges, with thin soil. On the ridges there are also some vegetation, including myall (*Acacia* spp.) in the northern part and mallee-type eucalypts in the south (Lowry, 1970). Grasses occur on the flats, but samphire (*Arthrocnemum* spp.) is present on the flats where the clay is saline. Low shrubs (including bluebush and saltbush) are present on both the ridges and the flats.

The Bunda Plateau is broken by steep-walled dolines formed by the collapse of cave roofs, and most of these lie on the Hampton Tableland that thus has a significant number of caves and rock holes, some with notable features. The rock holes (commonly 0.3 to 1.8 m across and 0.3 to 1.3 m deep) are developed in many of the rock pavements, and large rock holes hold water for several months after rain (Lowry, 1970). Several degraded fault scarps occur on the Hampton Tableland, mostly trending north from Madura for 35 km with a relief of about 20 m.

NYANGA PLAIN

The Nyanga Plain (the name derived from Lake Nyanga) lies to the north and west of the Nullarbor Plain (Fig. 3.1). The surface of the Nyanga Plain is featureless, except for depressions (6 to 15 m deep and 0.8 to 8 km across such as Lake Nyanga) that are believed to have formed partly by solution of limestone and partly by deflation of clay (Lowry, 1970). An abrupt margin to these depressions is characterised by a low scarp, which, in part, also separates the Nyanga Plain from the Carlisle Plain. The Nyanga Plain is underlain by a thick layer of clay and kankar and is covered by myall scrub (Lowry, 1970).

MARDABILLA PLAIN

The Mardabilla Plain (the name derived from Mardabilla Rock) is in the southwestern part of the Bunda Plateau (Fig. 3.1). Its flat land surface, mainly consisting of clay and kankar covered by dense mallee scrub 3 to 6 m high, is broken by numerous inliers of crystalline basement rock which project above the level of the plateau (Lowry, 1970). The inliers are generally surrounded by depressions 5 to 10 m deep and 180 to 550 m across, with the most prominent inlier, Mount Ragged, at an elevation of 593 m and stands about 460 m above the plain. East of Mount Ragged, the surface of the plateau is broken by numerous craters 1.5 to 6 m deep and 4.5 to 90 m across, caused by the collapse of the kankar crust into shallow caves (Lowry, 1970). The relief in the basement rocks is greatest in the south, and several of the low granite rocks do not project above the general level of the limestone surface. Saline lakes, associated with gypsiferous dunes on their eastern (lee) sides, occur on the western margin of the plain.

CARLISLE PLAIN

The Carlisle Plain (named after the Carlisle Lakes) lies in the northwestern part of the Bunda Plateau, north of the Nyanga Plain (Fig. 3.1). The plain is characterised by sandy soil with sparse myall and mulga scrub and enclose depressions up to 30 m deep and 10 km across and having centripetal drainage, with clay pans occurring in the centres of most depressions (Lowry, 1970). Sandstone and claystone (~20 m) and sandy limestone (~2 m) are exposed in the creeks on the eastern wall of the 3.2 km-wide depression (at lat. 29°22'S, long. 127°07'E). Spinifex (*Triodiu*) grows in abundance on aeolian sand but is virtually absent from the more clayey residual soils in the Eucla Basin. Gypsiferous dunes occur on the sides of salt lakes, such as the Carlisle Lakes.

3.2 EUCLA MARGIN

Superimposed on the margin of the Bunda Plateau is an assemblage of paleocoastal landforms, including coastal dunes, dune-ridges and possible beach ridges, embayment, lagoons, estuarine plains, cliffs, scarp and shelf.

SOUTHERN PART OF THE GREAT VICTORIA DESERT

Superimposed on the northern-northeastern region of the Eucla paleovalleys and the northern margin of the Bunda Plateau is the Quaternary dune field of the Great Victoria Desert that lies in the central southern part of the continental anticlockwise whorl of dunefields (Brookfield, 1970; Wasson et al., 1988; Fig. 3.1). Sedimentary rocks in the extreme north of the Eucla Basin are overlain by easterly-trending longitudinal dunes that extend northwards into the Officer Basin (Fig. 2.1). The principal landforms are longitudinal dunes (5–30 m high, 1.5 to 5 km long, and spaced 230 to 730 m apart) and playa lakes (e.g., Wilkinson Lakes, Lakes Maurice, Dey Dey, Jubilee Lake, Yarle Lakes, Lakes Ifould, Tallacootra and Chundie Swamps; Fig. 3.3). The aeolian sands of the Great Victoria Desert, are typically red-brown but vary in colour from orange, brown, to pale brown and are mostly weakly consolidated or unconsolidated. The dunes vary in thickness from 5–40 m and were deposited during development of the longitudinal dunes in cyclic arid periods of the Quaternary, possibly extending back to 220 ka (Sheard et al., 2006). An extensive calcrete paleosol is present, which can be traced from the swales up into the core of the dunes (Benbow, 1988, 1990a).

Figure 3.3 Landform of northern Eucla margin (Jubilee Lake near the border of Western Australia and South Australia) area. (Courtesy Diatreme Resources 2009).

OOLDEA RANGE

The Ooldea Range and parallel Barton Range (40–70 km to the northeast) as well as Paling Range (as a third feature connected the Ooldea and Barton Ranges), are a preserved assemblage of Eocene coastal aeolian landforms that include coastal dune, dune ridge and possible beach ridge (Benbow, 1990a). The ranges formed on the northeast margin of the Nullarbor Plain. Dune crests reach 300 m AHD, which is 40 to 180 m above the Nullarbor Plain. The width of dunes ranges from 10–25 km (Fig. 3.4). The linear formation of up to 175 m-thick quartz sand (e.g., Ludbrook, 1961), together with their morphology and position in the landscape, indicates that these ranges are paleo-coastal dunes (Benbow, 1990a). The longest of the linear landforms, which have a morphology similar to the Pleistocene and Holocene coastal dunes of southeastern Australia, is the Ooldea Range, which marks the limit of the Bunda Plateau and lies 25–300 km from the present coast (Benbow, 1988; Benbow and Crooks, 1988). The northwestern limit of this range is marked by a prominent scarp, described by Lowry (1970) as marking the northern margin of the Eucla Basin. Lowry (1970) believed this feature may have been of wave and wind origin. In each of the Ooldea, Barton and Paling ranges, marine and fluvial Middle – Late Eocene Pidinga Formation occurs in the lower part and is overlain by the aeolian Ooldea and Barton Sands in the upper part (see Chapter 4). The flanks and crests and the adjacent country of these ranges are blanketed by Quaternary aeolian sand and calcrete of the longitudinal dunes of the Great Victoria Desert (Benbow, 1988).

The Ooldea Range is oriented northwest–southeast, over 650 km long and 10–25 km wide, standing 40–180 m above the Bunda Plateau (Benbow, 1988). The Ooldea Range probably developed along a structurally controlled region of elevated crystalline basement or lithified pre-Paleogene cover rocks such as those of the Late Proterozoic – Paleozoic Officer Basin (e.g., Benbow and Crooks, 1988). The range is composed of coastal dune ridges that mark the edge of the sea during the Middle Eocene (e.g., Benbow, 1990a).

A number of segments occur long the Ooldea Range, which are separated by low saddles or breaches or tidal inlets that are the termination of, or through which flowed, paleorivers such as those associated with Serpentine Lakes, Karari, Woldra and Wynbring paleochannels. The northwestern major segment consists of several gently curved small segments that are characterised by broad flat crests on which are superimposed numerous distinct dune ridges that are usually separated by 2–5 km and have an amplitude of 10–15 m. These may also include less distinct smaller ridges that can be recognised from LANDSAT and DEM images. The ridge elevation decreases toward the basin and against the Serpentine Lakes Paleovalley to the northeast, and the ridges curve around toward the southwest. In contrast, the middle major segments of the Ooldea Range are straight and have a well-defined crest. Closely spaced beach ridges occur on the northeastern margin of the Ooldea Range adjacent to this middle major segment. Here, parallel to the range, there are strings of circular depressions, generally less than 1 km across. Topographic data over this region suggests that the series of depressions are separated by ill-defined ridges which stand from 5 to 10 m higher (Benbow, 1988).

The southeastern major segment of the Ooldea Range gradually lowers in elevation toward the southeast and is composed of a few arcuate segments with pronounced headlands separating them. It is arguable whether the very subdued area of ‘fragmented’ positive relief in the Narlaby Plain (Hou et al., 2012) should be considered part of the Ooldea Range (Benbow, 1988). Hou et al. (2008) considered that the southeastern segment of the Ooldea Range in the Narlaby Plain may have been modified by flooding and reworking as a result of relative downtilting during neotectonic overall uplift of the Neogene Eucla Basin, as the Australian continent was tilted towards the northeast.

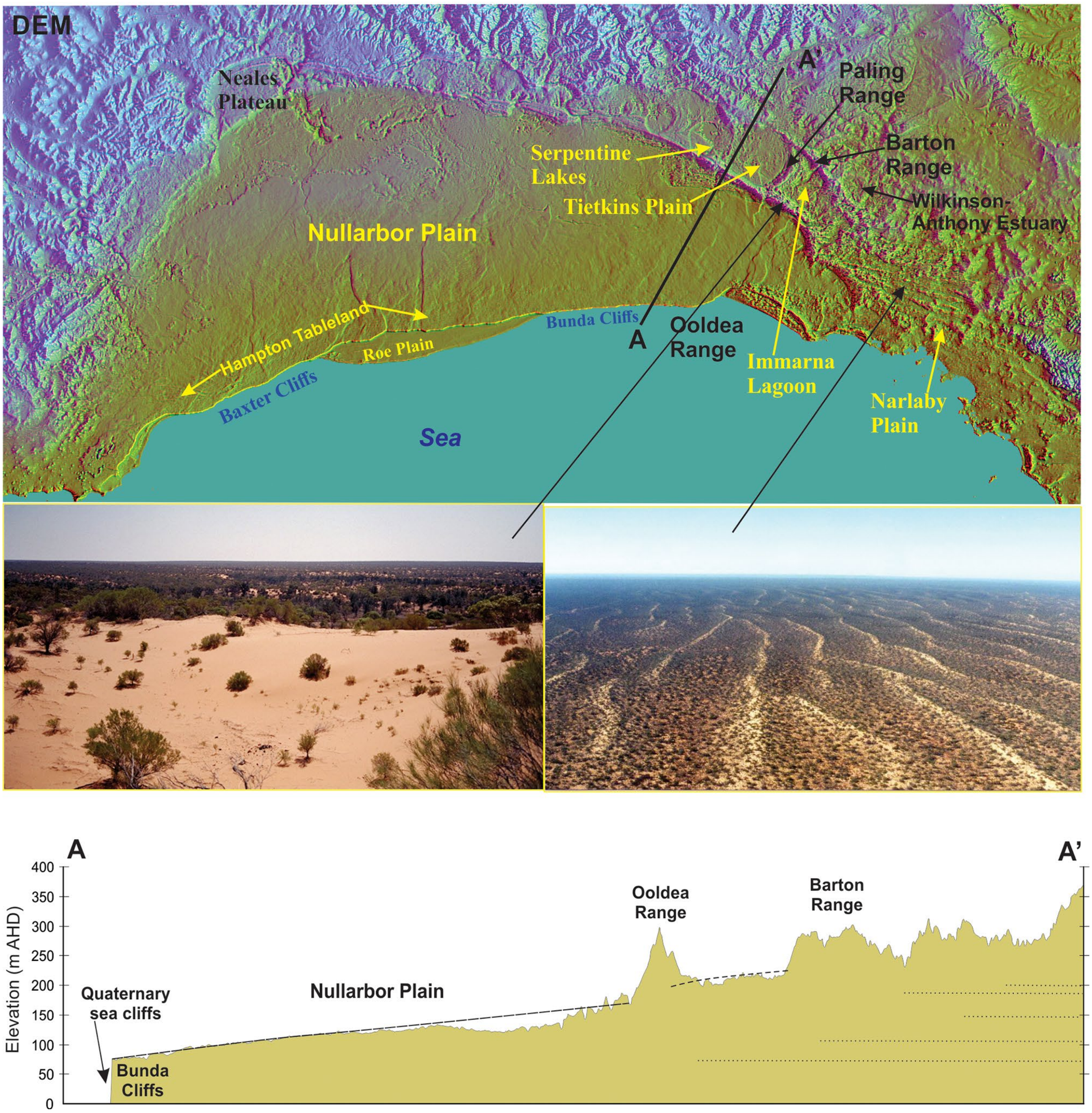


Figure 3.4 Landforms of northeastern Eucla margin (Ooldea Range) area. (DEM imagery by M. Sandiford; photos courtesy Geological Survey of South Australia).

BARTON RANGE

Parallel to the Ooldea Range and some 40–70 km northeastwards lies the shorter Barton Range that marks the edge of the sea during the Late Eocene. The Barton Range is a broadly sinuous coastal dune, which is about 250 km long, and 5–7.5 km wide at its northwestern end and up to 25 km wide over a significant part of its length (Fig. 3.4). The Barton Range is 20–120 m above the surrounding country, with the elevation varying from 200–300 m (A.H.D.; Benbow, 1990a). The northwest 78 km-long segment of the Barton Range is similar to the middle segment of the Ooldea Range, but to the southeast appears to broaden in width to about 35 km (middle segment), occurring together with elevated shallow crystalline basement and lithified cover rocks in the eastern part (Benbow, 1988). Further to the southeast the Barton Range is broad but low, lying only 20–40 m above the surrounding country (Benbow, 1990a). The less well-defined southeastern segment of the Barton Range appears to divide into two major branches of similar broad elevation, where there are many isolated to partly connected areas of topographically low relief within it (Benbow and Crooks, 1988). The southern branch is irregular, and drilling indicates the presence of substantial sands at its northwest end, suggesting this feature is a continuation of the coastal dune (Benbow, 1988).

The northwest-southeast oriented and parallel features of both the Ooldea Range and Barton Range, which are parallel to significant segments of the present coastline, continental shelf and plate margins, are very likely related to the pre-Cenozoic morphologic fractures which were active during the separation of Antarctica and Australia (Benbow et al., 1995b). The southeastern parts of both the Ooldea Range and Barton Range, of arcuate and sinuous form (respectively) are inferred to have resulted from wave refraction around headland promontories. These are underlain by crystalline basement rocks of the Gawler Craton. In this area, the Ooldea Range has three distinct projections or headlands where crystalline basement either crops out and/or is at shallow depth (Benbow, 1988).

PALING RANGE

In the northeastern Eucla Basin margin, both Ooldea Range and Barton Range are joined by the Paling Range that is 71 km long and arcuate shaped, connecting the Ooldea Range near Maralinga to the northwest end of the Barton Range (Fig. 3.4; Benbow, 1988). A detailed account of the geomorphology of these landforms is provided by Benbow (1989). The Paling Range averages 20 km width along most of its length and narrows to 5–7.5 km at its northern end, where it is less well defined (Benbow, 1988). In the Paling Range there are parallel dune ridges and a number of small, disconnected depressions which may indicate the existence of a former paleochannel (Benbow, 1989; Roger, 2000; Hou et al., 2007a). Largely, the Paling Range has an orientation that parallels the major fabric of the crystalline basement of the western part of the Gawler Craton, laying along the boundary between aeromagnetic provinces of contrasting character (Benbow, 1988).

IMMARNA LAGOON

Northeast of the Bunda Plateau (southeast of the Paling Range), situated between the parallel Ooldea and Barton ranges, Immarna Lagoon is a mostly buried coastal terrain of lagoonal origin, and to the northwest, beyond Lake Maurice there is evidence for a similar landform (Fig. 3.4).

TIETKENS PLAIN

Tietkins Plain, standing at ~180 m AHD and is distinctly higher than the Nullarbor Plain, is a playa lake carbonate karst, located in the northeast of the Bunda Plateau (north of Maralinga), situated between the parallel middle major segment of the Ooldea Range and northwestern part of the Barton Ranges. Tietkens Plain is an extensive and partly buried coastal terrain of lagoonal origin, being mantled in part by the dunefield. Its surface formed on Miocene lacustrine carbonates of the Garford Formation. From the Paling Range to the northwest, there are strings of circular depressions that occur across strike for up to 25 km northeast of the Ooldea Range and are prominent on LANDSAT and SRTM DEM images (Fig. 3.4). These discontinuous strings of depressions have variable lengths ranging from 5–30 km and the width of the ridges is 1.5–2 km, oriented northwest but change in direction and became arcuate shaped at the northwestern limit (Benbow, 1988).

NARLABY PLAIN

In the northern Narlaby Plain, east of the southeastern segment of the Ooldea Range, there are a number of parallel linear ridges which are connected to the Ooldea Range with a small angle, and are probable beach ridges (Fig. 3.4; Benbow and Crooks, 1988) that mainly formed during Neogene time (Hou et al., 2008). Some of these ridges stand about 20–60 m above the adjacent country and are about 50 km long and 4.5–10 km wide (Benbow, 1988).

NEALES PLATEAU/EMBAYMENT

The Neale Plateau is believed to be of marine origin, lying 50 m or so above the Bunda Plateau (Benbow, 1988). The marine origin of the Neale Plateau is evident from its smooth unruled shape and the occurrences to the south of the Neale Plateau of marine Late Eocene sediments that occur at a similar elevation (Lowry, 1970; Bunting et al., 1974; Van de Graaff et al., 1977). This scarp terminates near the known western limit of the Ooldea Range (Fig. 3.4).

WILKINSON-ANTHONY ESTUARY

The Anthony Paleovalley mouth is characterised as a wide estuarine funnel with islands or basement highs, suggesting a persistent swamp environment during the Eocene. The channel was probably blocked by basement highs or beach barriers built up as the sea advanced up the Anthony Paleovalley, where together with Wilkinson, Karari and Woldra paleovalleys are recognised as Wilkinson-Anthony Estuary (Fig. 3.4; Hou et al., 2001a). The Karari Fault trends through the Wilkinson-Anthony Estuary and probably influenced the development of the central part of the funnel. The well-developed carbonaceous facies in the Pidinga Formation indicates that the floodplain in the estuarine funnel area formed in a climate suitable for the growth of abundant vegetation. Silicification at the top of the Khasta and Pidinga formations indicates post-Eocene weathering. Wilkinson, Karari and Woldra paleochannels in the estuary are characterised by tidal inlet deposition with a strong marine influence as indicated by tidal structures and relatively abundant siliceous spiculate facies (Hou et al., 2001a). The facies evolution from channel to estuarine to marginal marine can be recognised in drillhole logs. Locally, tidal current erosion prior to deposition of the Khasta Formation can be observed.

BUNDA CLIFFS

The Bunda Plateau’s southern margin is abrupt, marked by a long line of cliffs that have various local names such as the Wylie Scarp, Baxter Cliffs, Hampton Range and the Bunda Cliffs, from the west to the east, which face the Southern Ocean, having heights of 40–90 m (Fig. 3.5). As a geographical feature, they form part of the longest uninterrupted line of sea cliffs in the world. Standing at the edge of the Australian continent, west of the Head of the Bight, rise the majestic Bunda Cliffs. The Bunda Cliffs (an aboriginal name used in South Australia for the name of the Nullarbor coastal cliffs) are the eastern part of a larger scarp of the Bunda Plateau that spreads from the western part of South Australia across to the southeastern corner of Western Australia, extending for around 100 km along the Great Australian Bight (Fig. 3.5).

The Bunda Cliffs, extending for 200 km to Wilson Bluff from the west of South Australia, range in height from 40–70 m above sea level, are composed of Eucla Group (see Chapter 4) fossiliferous limestones of Paleogene–Neogene age. The combination of changing sea levels and the uplift of the Nullarbor Plain culminated in the creation of these cliffs over several million years. The cliff front also stretches to the east of the Bight but now lies buried under the sand of the present Holocene dune system.

HAMPTON RANGE

The Western Australian scarp at the southern edge of the Bunda Plateau is divided into three segments from the east to west: Hampton Range, Baxter Cliffs and Wylie Scarp (Fig. 3.5). The Hampton Range (Tate, 1879) is named for the eastern segment between Wilson Bluff and Twilight Cove where it separates the Hampton Tableland from the Roe Plains. A steep ~15 m-high cliff of limestone forms the upper part of the scarp, and below it is covered by clay, kankar and limestone debris. The elevation of the top of the Hampton Range scarp ranges from 70 to 120 m above sea level and the base from sea level to 30 m. In plan view, the scarp edge forms a series of cusps ranging from 1.5 to 35 km apart. Like the Bunda Cliffs, the Eucla Group limestones can be seen from Eucla to Madura, where they form a scarp separating the Roe Plain from the Hampton Tableland, but in this section the coastline has moved away from the cliffs. The scarp in this area runs parallel to, and within sight of, the Eyre Highway.

In the early stage, the morphology of the Hampton Range was interpreted as an emerged sea cliff (Tate, 1879), or a fault scarp (e.g., Woolnough, 1933; Gentilli and Fairbridge, 1951; Frost, 1958), which was not supported by later study of the stratigraphy (e.g., Ludbrook, 1958b), i.e., the stratigraphic evidence being against any displacement. Later, marine erosion was suggested for the origin of the morphology of the scarp (e.g., Jennings. 1961), showing that the relative straightness of the cliff does not necessarily demonstrate a fault origin as it could equally have formed from marine erosion of beds with extremely uniform structure and lithology (Lowry, 1970).

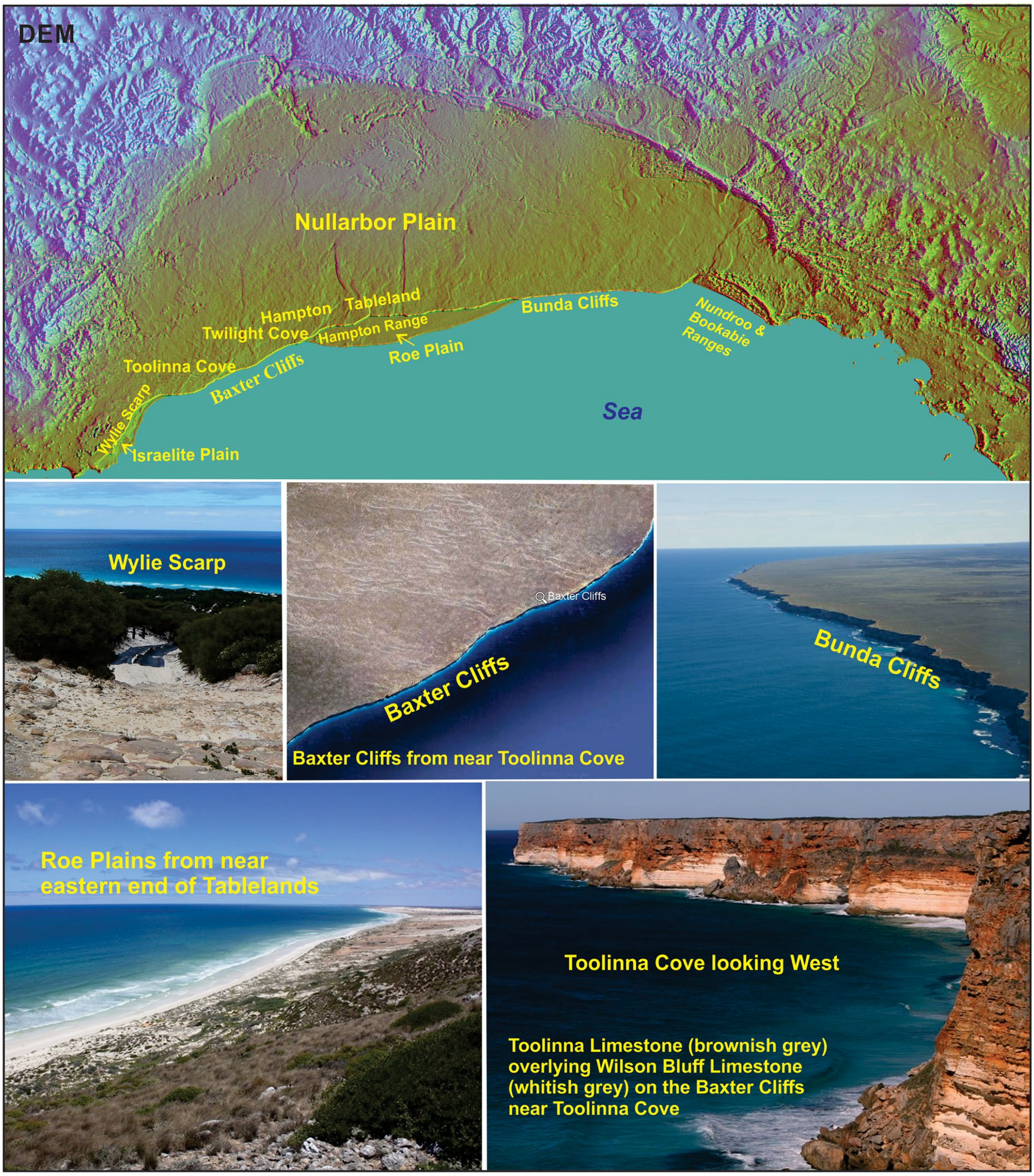


Figure 3.5 Landforms of southwestern Eucla margin. (DEM imagery and photos courtesy M. Sandiford and website).

BAXTER CLIFFS

The Baxter Cliffs (named after J. Baxter and the site marked by the Baxter Memorial; Lowry, 1970), the middle segment of the scarp at the southern edge of the Bunda Plateau, extends for 158 km southwest from Twilight Cove to the eastern end of Bilbunya beach near Point Culver (Fig. 3.5) and the beginning of the Wylie Scarp. The elevation of the cliffs average about 75 m and range from 60 to 90 m according to spot heights shown on the CULVER 1:250,000 topographical sheet. The cliffs drop vertically into the sea in most places, with masses of recently collapsed rock at their base in a few places, and a small sandy bay (Fig. 3.5) accessible with difficulty from the cliff top at one place. A composite profile occurring at the cliff in places, shows the lower half vertical and freshly eroded, and the upper half forming a talus slope surmounted by a 15 m bluff like the

Hampton Range (Lowry, 1970). A similar composite profile also occurs at Wilson Bluff, indicating two periods of marine erosion separated by a period of subaerial erosion (Jennings, 1963).

Geographical change occurs at Twilight Cove with the Baxter Cliffs turning inward to become the Hampton Tablelands that overlook the Roe Plains from the north and go inland more than 40 km.

WYLIE SCARP

The Wylie Scarp (named after Wylie) joins the Baxter Cliffs at its eastern end near Point Culver and extends southwestwards along the margin of the Eucla Basin to about Point Malcolm where it becomes indistinct (Fig. 3.5). The Wylie Scarp is about 85 m above sea level near Point Culver and stands more than 90 m and 75 m respectively above the coastal plain at latitudes 33°06'S and 33°33'S (Lowry, 1970). Between these points, a spot height of 142 m above sea level is shown on the MALCOLM 1:250,000 topographical sheet. The top of the scarp is characterised by a 1.5–3 m high rocky kunkar bluff, and below it is a steep slope with rubbly exposures of Wilson Bluff Limestone over a vertical distance of 15 to 30 m, with a 1.5 km wide apron of debris in the lowest part of the scarp. In plan view the scarp is smooth and regular on a large scale, but intricately indented on a small scale because of headward erosion of gullies along joints. The Wylie Scarp is believed to have been formed by marine erosion during a highstand of sea level in the Pleistocene, in the same manner as the Hampton Range (Lowry, 1970).

ROE PLAINS

The Roe Plains is situated between modern seacliffs that extend 200 km to the west (Baxter Cliffs) and 220 km to the east (Nullarbor Cliffs), extending from Twilight Cove to Wilson Bluff, reaching a width of ~40 km opposite Madura, and reaches an altitude of 30 m at Madura (Fig. 3.5; Lowry, 1970). The Roe Plains, a crescent-shaped scrubland on Australia’s southern shore (Fig. 3.1), is a coastal plain backed by a wave-cut scarp and partly covered with coastal sand dunes. The Plains are bounded on the north by the Hampton Tableland escarpment rising to the Nullarbor Plain and to the south by the Great Australian Bight. Most of the Roe Plains has a smooth gently sloping surface and is at an elevation of approximately 6 m above sea level. The Roe Plains, considered geologically to be late Neogene, was created by the scouring action of waves that, over millions of years, planed down a thick layer of limestone (an earlier era would be Mesozoic??; Wikipedia — this refers to Late Pliocene fossils in the calcarenite, presumably exposed in the caves?).

The Roe Plains are predominantly marine dunes on a coastal plain, particularly such as the large accumulation of sand on the western end of the plain where dunes reaching 90 m in height and have blown approximately100 km inland. Most of the plain is characterised by clay soil and is covered in grass or tea-tree or mallee scrub, while the sand dunes on the fringes are either mobile or are fixed by mallee scrub (Lowry, 1970). The plain is underlain by a thin (2–3 m-thick) layer of Late Pliocene – Early Pleistocene marine limestone (the Roe Calcarenite), deposited in a shallow illuminated shoreface to the inner shelf with the seafloor probably covered by seagrass, much like the modern seafloor offshore the Roe Plains today (James et al., 2006), and it is believed that the plain was carved in the Neogene limestones by marine erosion during the Pleistocene. The process, which led to ~85 km of cliff retreat in ~3 million years, is interpreted to be a variant on the shaved shelf process operating on the shelf today (James et al., 2006).

ISRAELITE PLAIN

Like the Roe Plains, the Israelite Plain is backed on the landward side by a scarp cut in Paleogene–Neogene limestone and is believed to have been formed also by marine erosion in the Pleistocene (Lowry, 1970). The Israelite Plain is characterised as the low-lying area of salt lakes and Quaternary sand dunes between Point Culver and Israelite Bay, with a few granite hills projected through the dunes on the southern part of the plain. (Fig. 3.5). The older sand dunes further inland have a cover of dense mallee scrub, whereas the younger sand dunes are mobile or have low vegetation (Lowry, 1970).

NUNDROO AND BOOKABIE RANGES

Along the southeast margin of the Bunda Plateau, there are the discontinuous Pleistocene coastal dunes, the Nundroo and Bookabie ranges (Fig. 3.5). Nundroo Range is at an altitude of about 90 m above sea level Bookabie Range is at an altitude of about 101 m above sea level.

EUCLA SHELF

The Eucla Shelf is the offshore extension of the Eucla Basin. As part of the shallow continental shelf of the Great Australian Bight, the Cenozoic Eucla Shelf is characterised as a very large, arcuate, relatively flat submarine plain (average gradient of 0.2–0.5 degrees) that is ~80 km wide at either end (e.g., near Israelite Bay of Western Australia and near Lincoln of South Australia), and 195 km wide at Eucla and 260 km wide at the Head of Mouth (Figs 3.1 and 3.6; Harris et al., 2005; Willcox et al., 1988). The Eucla Shelf is not easily studied. Towards the southern edge of the shelf, the continental slope has a gradient of 2 degrees to 5 degrees extending down to about 3,000 m (Lowry, 1970). The Bunda Plateau forms the northern edge of the Bight, formed of ocean floor that has been uplifted. The Eucla Shelf is commonly 30 m deep within 20 km of the coast, and it slopes gently to the outer edge of the shelf at a depth of 140 m to 183 m (Phizackerley, 1967). The continental shelf comprises the inner shelf (<50 m deep water), the middle shelf (50–120 m deep water) and the outer shelf (10–30 km wide and 125–170 m deep), extending to the shelf break (James et al., 2001).

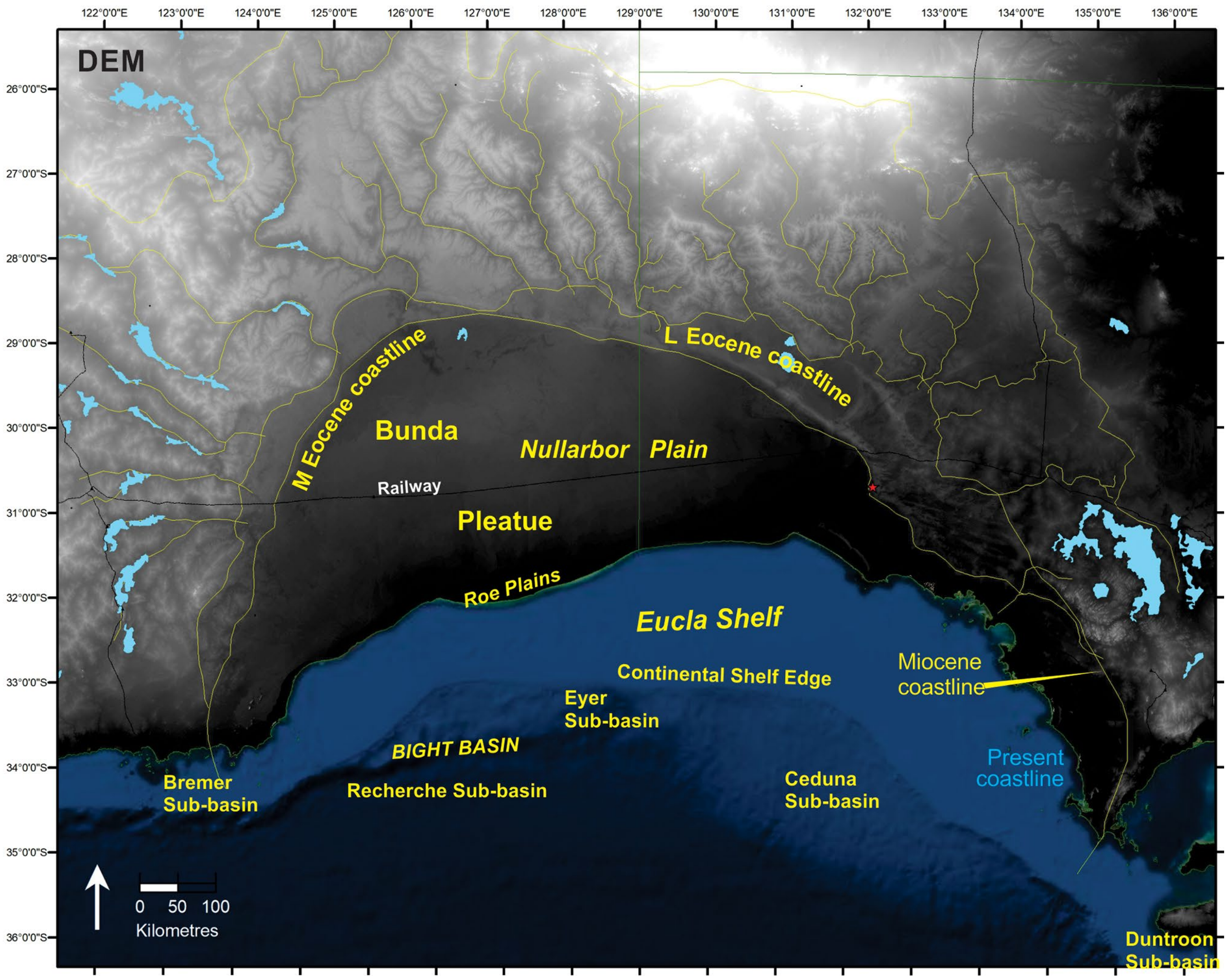


Figure 3.6 Landforms of southern Eucla Basin with paleoshorelines.

The Eucla Shelf is a cool-water carbonate sedimentary province located in a high-energy, swell-dominated oceanographic setting (James et al., 1994). Much of the seafloor is shallower than 70 m water depth, 100 km-wide, is bare Cenozoic coral-rich limestone (?Late Pleistocene in some places), or is covered by patches of Holocene sediment, up to 1.5 m thick (James et al., 1994). On the outer shelf, Holocene sediment (below the base of wave abrasion but inboard of the shelf edge) is fine, microbioclastic muddy sand with minor delicate bryozoans overlying a 9–13 ka rhodolith gravel (James et al., 1994). Due to the lack of inflow from the land, with the associated lack of sediments being deposited, the Pleistocene sand on the sea floor of the Eucla Shelf is not buried by later sediment, essentially unchanged, for 20 million years. Features preserved include evidence of lagoon environments (Li et al., 1996b) and bryozoan reef mounds (Holbourne et al., 2002).

Underlying the Eucla Shelf, at depths ranging from 200–4,000 m, is the Bight Basin, including the Eyre, Ceduna and Duntroon sub-basins and Polda Basin to the east, Recherche and Bremer sub-basins to the west (Figs 3.1 and 3.6).

3.3 PALEOSHORELINES

The Eucla basin and its adjacent paleovalley system have a large areal extent that contains a complicated succession of marine and non-marine strata spanning a wide range of depositional environments. Marine transgressions of the Eucla Basin, established from regional biostratigraphic studies (e.g., McGowran, 1979; McGowran et al., 1997, 2004), make up at least five third-order transgressions (see Chapter 4). Cenozoic sediments in the eastern Eucla Basin are bounded by major erosion sequence boundaries corresponding to relative sea-level changes (Hou et al., 2006b). The sedimentary units recovered across the basin correspond to the established Cenozoic transgressions from neritic components of the chronostratigraphic record, i.e., Wilson Bluff, Tortachilla, Tuketja-Tuit, Bairsdale, and Jemmys Point-Hallett Cove (McGowran et al., 1997, 2004).

The general orientation and parallelism of the Eucla Basin paleoshorelines (Figs 1.1 and 3.1) suggests that the primary control on shoreline position was the gentle topographic slope established along the southern Australian margin as a consequence of earlier Mesozoic rifting (Fig. 2.5) associated with the separation of Antarctica and Australia (Benbow, 1990a, Alley et al., 1999). However, significant disparities in the elevations of correlative nearshore sequences across the Eucla have long been evident (Jones 1990; Benbow et al., 1995b; Hou et al., 2001a, 2003a; de Broekert and Sandiford, 2005), to the extent that the relative position of shorelines formed during successive transgressions shows systematic variations across the basin, which is obvious that the paleo-landscapes had been deformed or tilted by neotectonic movements (Fig. 3.7 and 3.8). This should be adjusted for terminal Neogene tilting and flooded to referred paleoshorelines (see Chapter 4, Figs 4.7, 4.8 and 4.9).

The Cenozoic shorelines, defined as bodies of coastal sand, here comprise beach, shoreface, barrier, dune, tidal inlet, washover and lagoonal facies (Hou et al., 2003b), regionally representing third-order highstands (Clarke et al., 2003; Hou et al., 2003c). Along the Eucla margin, the paleoshorelines and paleoriver mouths are characterised by deposits that reflect a series of former estuaries (e.g., Neales, Wilkinson and Anthony; Fig. 3.1), as well as lagoons and coastal barriers (e.g., Ooldea and Barton; Fig. 3.4). Four distinct constructional phases for the development of shorelines can be recognised and correlated with major third-order sea-level events, established by others from the marine depositional record as occurring during the middle Eocene (~42.5 Ma), late Middle Eocene (39–36 Ma), late Eocene (36–34 Ma), and Miocene-Pliocene (15–2.6 Ma) (see Chapter 4).

Prior to the Wilson Bluff transgression in the Middle Eocene (~42.5 Ma) paleovalleys probably delivered a substantial volume of siliciclastic sediment to a shoreline close to or beyond the present-day coastline (Quilty, 1994). The position of the shoreline at commencement of the Wilson Bluff transgression is placed at the most landward occurrence of Hampton Sandstone in the central Eucla Basin (see Chapter 4), now overlain by lower Wilson Bluff Limestone (Clarke and Hou, 2000; Hou et al., 2003b, 2006a, b). Progression of the shoreline inland during marine transgression between 42 and 41 Ma, is uncertain but may be marked locally by the presence of sandy, sponge spicule facies (Hou et al, 2003b). The shoreline received fluvial sediment input but the distribution of paleoriver systems beneath the cover of marine carbonates is poorly known. Early Middle Eocene paleovalley sediments beneath the Wilson Bluff Limestone, west of Jacinth, were identified from recent airborne electromagnetic surveys, and confirmed subsequently by drilling and palynological investigation of drill cuttings (see Appendix 1, unpublished data). The paleovalley is 3–5 km across with sandy sediment fill up to 65 m thick, overlain by 25–30 m of marine carbonates. At around 45–50 m below ground level the sediments are saturated with saline groundwater, which is used now as the principal source of water for beneficiation of mineral sands at the Jacinth-Ambrosia HM mine.

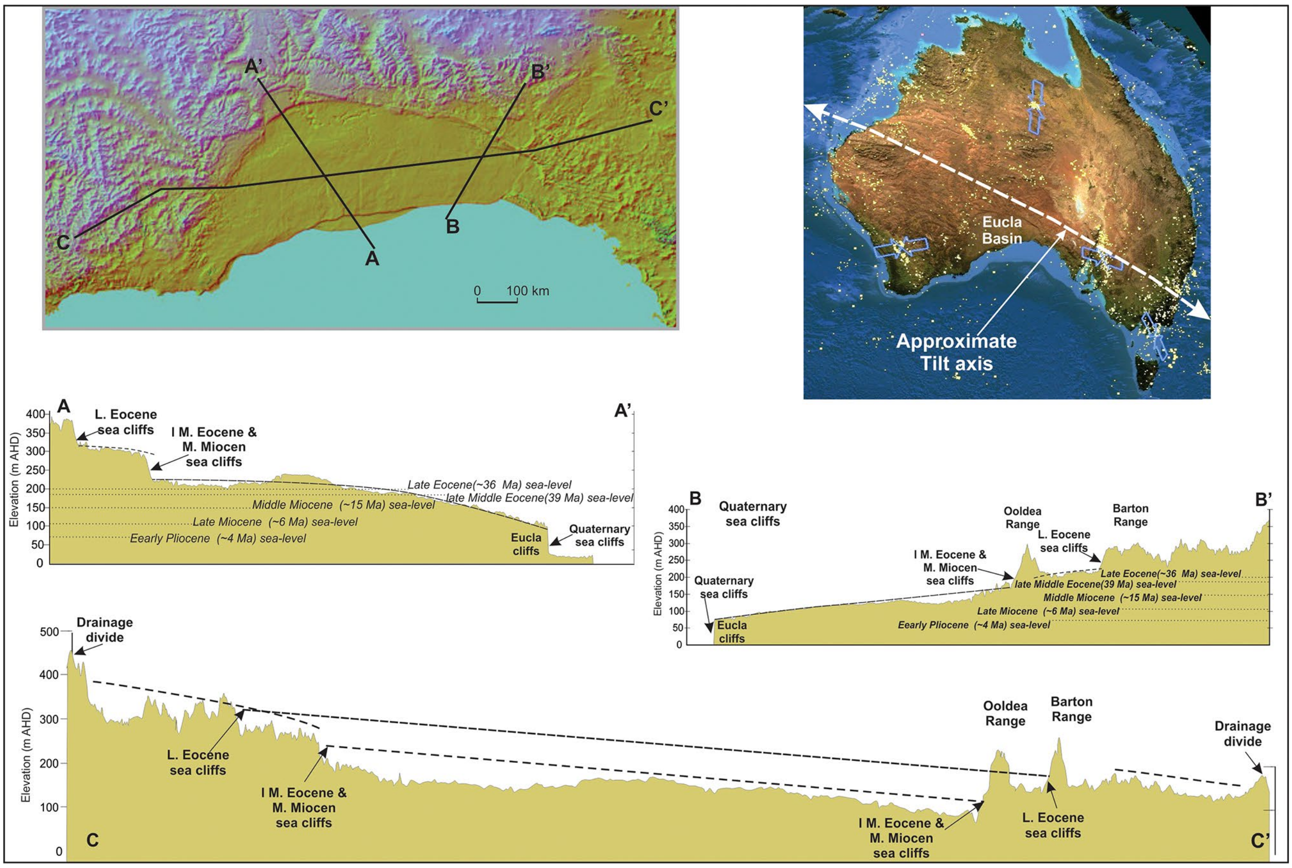


Figure 3.7 The topographic difference between the western and eastern margins of the Eucla Basin showing the downtilting to the east and interpreted tilt axis across the continent (Hou et al., 2011b).

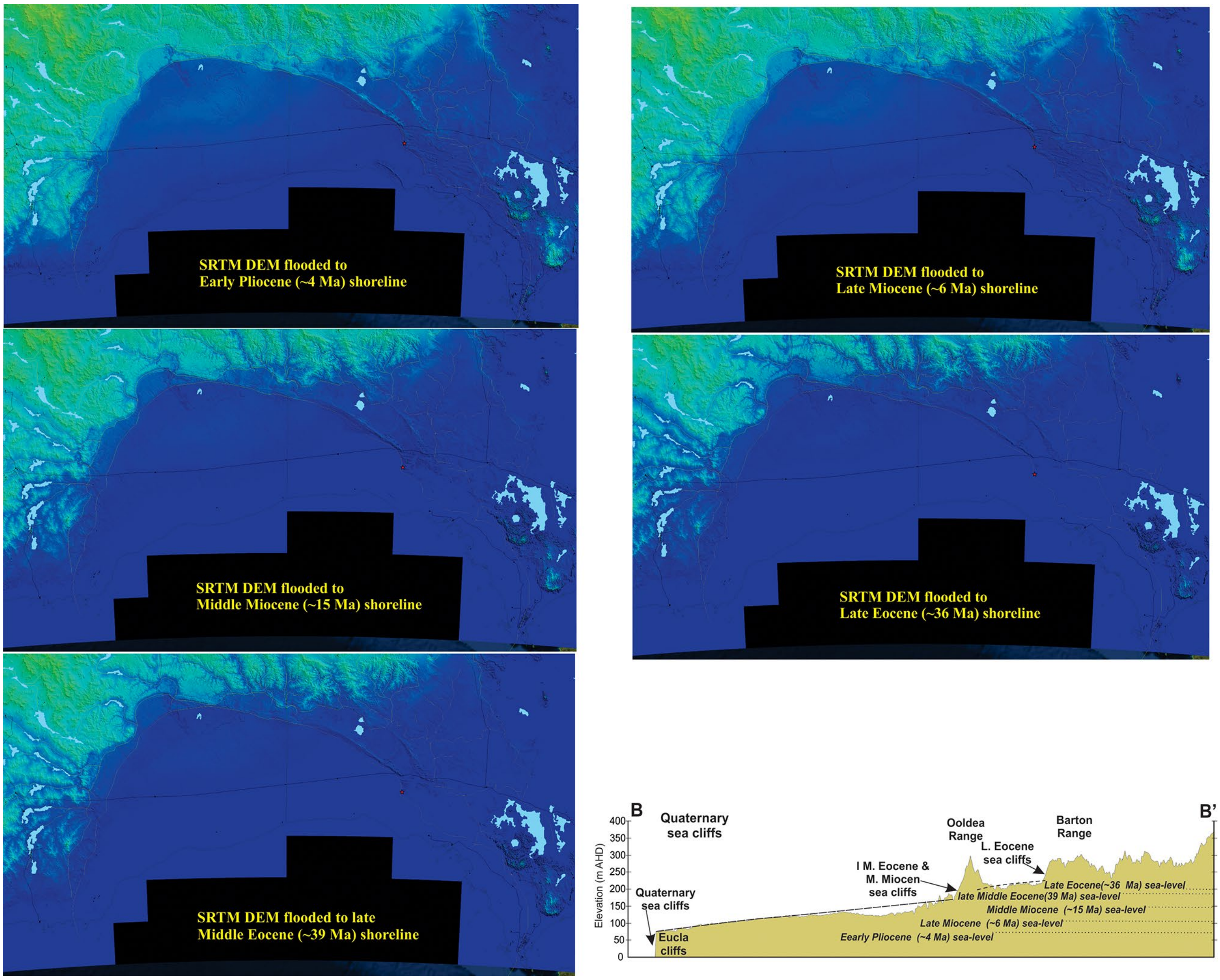


Figure 3.8 SRTM DEM flooded to referred paleoshorelines of the Eucla Basin.

The influence of late Middle Eocene Tortachilla and Late Eocene Tuketja-Tuit transgressions extended some several hundred kilometres up paleovalleys that drain into the Eucla Basin (see Chapter 4; e.g., Alley et al, 1999; Clarke et al., 2003). During these stages, input of terrigenous flux resulted in extensive aggradations, first as non-marine to marginal marine sediments and then, as highstand deposition of biogenic sediments (Alley et al., 1999). Initial development of the late Middle Eocene shoreline during the Tortachilla Transgression is expressed as marginal marine and estuarine (carbonaceous) sand, silt and mud onlapping an erosional surface in the coastal plain and adjacent paleovalleys along the Eucla margin (see Chapter 4). The highstand deposits of this sequence are locally intercalated with carbonaceous, calcareous and glauconitic limestone (e.g., Paling and Norseman formations), and with barrier sands (Benbow, 1990a) of the Ooldea Sand (Hou et al., 2003a; Zone P14-15, see Chapter 4) in the eastern part of the basin. The Tortachilla shoreline is characterised by the Ooldea Barrier along the northern and eastern margin of the basin. Shortly afterwards, a yet higher sea level led to further marine transgression during the Tuketja-Tuit transgressions and resulted in the shoreline migrating farther inland (see Chapter 4). The shoreline position extended to the inland margin of the Neales Plateau in the northwest (Clarke and Hou, 2000) and to the Barton barrier – Wilkinson estuary in the northeast (Hou et al., 2003b, 2006b; Fig. 3.2), as characterised by the Barton Range in the eastern basin. The absence of Tortachilla and Tuketja-Tuit coastal barriers in the western Eucla margin may reflect coastal processes of longshore drift towards the east under the influence of prevailing westerlies (Kemp, 1978).

The Barton and Paling barrier systems are characterised by a series of Late Eocene (36–34 Ma) coastal sand barriers, oriented SW-NE, while the Ooldea barrier was partly submerged to form a string of offshore barrier islands (Clarke and Hou, 2000; Hou et al., 2003a, c; 2006b). In contrast to previous models (e.g., Benbow, 1990a; Rogers, 2000), recent interpretations (Hou et al., 2006b;

2008) suggest that the distribution pattern of the Late Eocene coastal barriers is neither parallel to that of the Ooldea Barrier, nor aligned along the present Barton Range (see Chapter 4). Thus the ‘Ooldea and Barton barriers’ (characterising early Cenozoic morphology) are not strictly equivalent to the present day ‘Ooldea and Barton ranges’, which reflect additional aeolian reworking following several marine regressions and later dry and windy conditions that accompanied cycles of Southern Hemisphere glacial maxima during the Quaternary (Sheard et al., 2006). Landward of the Ooldea Barrier, parallel beach ridges developed in the Late Eocene that reflect the interaction of localised sediment supply along paleodrainage networks and continuous, westerly long-shore drift processes operating behind the Ooldea Barrier ‘islands’ (see Chapter 4). Sediments comprise both lagoonal facies and barrier facies with beach placer potential (Reid and Hou, 2006), although wave energy was likely much less than that for the seaward Ooldea Barrier. The elevation of the Barton–Paling barrier/ridge decreases toward the southeast, following the long-shore drift direction, reflecting the combined effects of a decrease in sand supply from the northwest and a broader-scale, relative east-down vertical movement (Hou et al., 2008). Paleoclimate modeling (Kemp, 1978) indicates westerly winds predominated across southern Australia at latitudes between 60° and 80°S in the Paleocene and Eocene. This is consistent with the occurrence of coastal dune barriers along the northeastern Eucla Basin margin and the apparent absence of substantial dunes along the western margin. J-shape barriers behind the Ooldea Barrier that formed during the Late Eocene may indicate a more north-westerly component to the prevailing winds at that time (Hou et al., 2011b).

The Neogene shorelines mainly followed those of the late Middle Eocene in the western, northern and northeastern margins of the Eucla Basin (Fig. 3.1). Analysis of the geomorphology of Cenozoic coastal deposits and sequences in selected drillholes across the southeastern basin, however, provide a record of progressive Mio-Pliocene coastal progradation, followed by a series of regressive strandlines (Hou et al., 2008; 2011b). This pattern of sedimentation, reflecting Mio-Pliocene sea-level change, is not apparent along the western margin of the basin because of continental scale tilting and relative uplift of the region (see Chapter 4). Relatively high cliffs developed along this margin during the Miocene, mimicking earlier, slightly more landward, Eocene cliffs (see Chapter 4). However, in the southeast Eucla margin, parallel linear ridges at an angle to the Late Eocene Ooldea Barrier represent Neogene strandlines (A. Crooks, pers. comm., 2002) of the Narlaby and Ilkina Formations. These are fine- to medium-grained, moderately to well-sorted sandstone that formed from reworking of the Middle Eocene Ooldea Barrier sands (Benbow, 1990a; Hou et al., 2003a, b, c, and 2006b). More extensive inundation of this area was due to downward tilting of the basin toward the east. The prolonged nature of the dynamic tilting is indicated by a relative northeastward decline in elevation of the key Eocene surfaces and ‘reverse flow’ features in Neogene valleys in the western Gawler Craton (Hou et al., 2008). Thus, the Neogene shorelines that developed in the southwestern Gawler Craton following maximum marine transgression extend farther inland than those of the Eocene (see Chapter 4; Hou, 2008). These may prove also to be prospective for heavy minerals.

3.4 EUCLA PALEOVALLEY SYSTEMS

A series of a dendritic system of paleovalleys is a prominent feature of the onshore Eucla Basin. Around and beyond the Bunda Plateau, there are a number of extensive and largely intact paleovalleys with some paleorivers that once carried water eastwards to the Eucla Basin (Fig. 3.1). Some paleorivers had their source toward the continents centre in the region of the Musgrave Province, Officer Basin, Albany-Fraser Orogen, Yilgarn and Gawler cratons and some up to 750 km from the plateau’s margin. These ancient rivers have incised broadly undulating terrains that are now capped by silcrete and ferricrete (Ollier et al., 1988; Benbow, 1986a). Many of these paleovalleys had their origins before the Late Eocene and after the Early Cretaceous, such as those on the southwest margin of the basin being beheaded in the Late Cretaceous, during separation of Australia from Antarctica (Ollier et al., 1988). For instance, the Tallaringa Paleovalley has dissected sediments of Early Cretaceous age and is infilled with Eocene sediments (Benbow, 1993). Also, the paleovalleys clearly defined in the regions of dissected crystalline basement of the Gawler and Yilgarn cratons are likely to be considerably older than Late Eocene (de Broekert, 2002; Hou, 2008).

The major Eucla paleovalley system, with watersheds covering a larger area than the Bunda Plateau, have been recognised and delineated using SRTM DEM, Landsat and NTT images, and geological data (e.g., maps and drillholes; Figs 2.2, 2.3 and 2.4). The modern topography has been shown to largely reflect the pattern of the paleovalleys. These paleovalley systems were developed in the onshore extensions of the Eucla Basin and are covered by Quaternary sand plains and colluvial regolith. They are largely superimposed on and locally obscure the older Paleogene and/or Neogene landforms. Other subsurface features such as those interpreted from NOAA and geophysical datasets (where available) have also been valuable in delineating the paleochannels. The Eucla paleovalleys have been delineated by regional and more detailed hypsometric maps in a number of places around the Eucla margin, such as the studies of Jutson (1934), Morgan (1966), Bunting et al. (1974), Van de Graaff et al. (1977), Barnes and Pitt (1976), Benbow (1986a, b, 1993), Benbow and Crooks (1988), Rogers (2000), Clarke (1993, 1994a, b), Kern and Commander (1993); de Broekert (2002), Hou (2008), Hou et al (2001a, 2003a, 2007, 2012).

The pattern of paleovalleys in the onshore Eucla Basin is dominantly sub-dendritic, reflecting both the pre-Paleogene land surface gradient and bedrock lithology and structure (Alley et al., 1999; Hou et al., 2001a; 2003a; de Broekert and Sandiford, 2005). In the eastern Yilgarn Craton, many trunk paleovalleys parallel the northeast orientation of a regional fracture field (Johnson and McQueen, 2001) while in the western Gawler Craton paleovalleys preferentially follow the topographic lows within weakly resistant, deeply weathered bedrocks (Benbow et al., 1995a; Hou et al., 2003a). The dimensions of the Eucla paleovalley systems vary greatly. River valley widths range from a few tens of metres to more than 30 km, depths can be up to 100 m, and main channel lengths extend typically 100 to 700 km into the hinterland (Alley et al., 1999). The principal direction of the paleodrainage flow was directly towards the basin from the continent, but, locally, some paleochannels feeding the eastern Eucla Basin show apparent reversed gradients at key stratigraphic horizons (Fig. 3.1). The lower reaches of the major paleovalleys are characterised by a series of estuaries (e.g., Wilkinson), coastal barriers and lagoons. Borehole controlled transects in eastern Yilgarn (Mt Morgan) and western Gawler cratons (Garford Paleovalley) indicate that the irregularities along the paleovalley floors dominantly occur as bedrock highs with an elevation of about 10–20 m above the regional gradient (Hou et al., 2003a; de Broekert and Sandiford, 2005). Changing sediment patterns and flow reversals observed within the eastern Eucla paleovalley fills suggest that the paleovalleys were affected by post-Eocene tectonic movements (Hou et al., 2008). The paleovalleys form a well-integrated contributory pattern. They typically form ‘V-’ (mostly upper reaches), ‘U-’ (mostly middle reaches) or ‘W-shaped’ (mostly lower reaches) cross-sections (Hou et al., 2001b; de Broekert, 2002).

The Eucla paleovalley fills have palynological and lithological correlatives in most parts of the paleodrainage systems, characterised as Paleogene fluvial-dominant sediments and Neogene lagoonal clay/ carbonate-dominate sediments. For instance, the Mangatitja Limestone (consisting of dolomitic limestone, limestone; Major, 1973) occurring in the Lindsay, Noorina and tributary paleochannels adjacent to the Musgrave Block can be correlated to its equivalents occurring in the paleovalleys that drained the Yilgarn Craton (e.g., Lefroy and Cowan Paleovalleys); and the sediments (Neogene carbonaceous sand and overlying green, grey white and varicoloured mottled mud) occurring in the paleochannels southwest of the Cowan Paleovalley (upper reaches, e.g., the Lake Tay and Three Star Lake region) can be directly equated with those in the Narlaby Paleovalley that drained the western Gawler Craton, based on similarity of lithology and palynoflora (e.g., Bint 1981). Also, the sediments (Neogene dolomitic and palygorskite-bearing clays overlying Paleogene carbonaceous Eocene sediments) in the Lefroy and adjacent paleovalleys (Clarke, 1993, 1994a, b) are equivalent to the Pidinga Formation and overlying Garford Formation (e.g., Benbow,1993; Benbow et al., 1995a).

The paleodrainages that drained northwest and north of the Eucla Basin contain valley calcrete (in contrast to ‘plains calcrete’), locally extensively silicified (Jackson and Van de Graaff, 1981). This valley calcrete is mostly laminated and/or nodular at the surface but is mostly structureless with minor nodules and is locally dolomitic at the subsurface. It is likely that part of the thicker occurrences of these valley calcretes are altered lacustrine carbonates. In the Throssell Paleovalley (in drillhole Throssell -1), for instance, these valley calcretes (27.5 m thick) are intensely silicified up to the basal 3.5 m and overlie about 70 m of grey and red gypsiferous clay (locally dark grey and contain black carbonised wood fragments, whilst at the base there is a very fine sand; Jackson et al., 1981). These clays are very similar in appearance to part of the Garford Formation deposited in the eastern and northeastern Eucla paleovalleys (e.g., Tallaringa, Garford and Kingoonya).

There are a number of exit points of Paleogene to Neogene major paleovalleys through the paleoshorelines, some of the most notable being those marked in the Figure 3.1. Major Eucla paleovalley systems (e.g., Raeside, Cowan, Lefroy, Roe, Rebecca, Ponton, Carey, Throssel, Baker, Wanna, Serpentine Lake, Noorina, Lindsay, Meramangye, Tallaringa, Garford, Kingoonya and Narlaby paleovalleys) are bounded by regional drainage divides and occupied by drainage networks (Bates and Jackson, 1987), ranging from individual hillslopes surrounding small headwater tributaries, to large tracts of continents drained by major river networks. These major paleovalleys form a roughly radial pattern focussed on the Eucla Basin margin (Fig. 3.1). The characteristics of these drainage basins offers important clues as to the origin of the Eucla paleovalleys. Within some drainage basins (e.g., Cowan, Lefroy and Kingoonya paleovalleys), locally, the stream tributaries are left free to branch in almost any direction and to follow irregular courses with many random bends.

Due to an absence of drillhole data in most parts of these major paleovalleys, such as northwestern-northern-northeastern Eucla paleovalleys (e.g., Meramangye, Lindsay, Noorina, Serpentine Lake, Wanna, Baker, Throssel, Carey, Ponton and Rebecca paleovalleys), the distribution and sedimentation of the channel sediments in these paleovalley areas are poorly understood.

Recent groundwater drilling programs conducted within Lindsay East Paleovalley, followed by palynology analysis and dating (Krapf et al, 2019), indicate inland marine conditions during the Late Miocene to Early Pliocene. The marine influence reaches as far as the foothills of the Musgrave Ranges, where alternating freshwater, lacustrine and marginal marine to estuarine conditions during the Late Miocene – Early Pliocene are documented in drillhole DH1A (Krapf et al, 2019). Palynological evidence for similar marine influence during Late Miocene – Early Pliocene are also identified in adjacent areas in drillholes Officer 1 and MAN1 (Zang and Stoian, 2006).

A few major Eucla paleovalleys, such as those which drained the Yilgarn and Gawler cratons, have been studied in relative detail by various authors (e.g., Bunting et al., 1974; Van de Graaff et al., 1977; Smyth and Button, 1989; Jones, 1990; Clarke, 1993; Kern and Commander, 1993; Rogers, 2000; Hou, 2008; Hou et al., 2001a, 2003a, 2007, 2012; Krapf et al, 2019). These studies can be used to interpret the probable development of other less well-studied Eucla paleovalleys. It should be noted that besides the major named Eucla paleovalleys, there exist other, less obvious, and unnamed paleovalleys across the onshore Eucla Basin.

KINGOONYA PALEOVALLEY

Kingoonya Paleovalley System (KPS) is one of several incised-valley systems that contain Paleogene-Neogene marine-influenced fluvial sediments in the Gawler Craton of South Australia, which joins Wynbring Paleovalley to the west (Hou, 2008). Paleochannel mapping and sampling with test drilling has indicated that the KPS drained across the western Gawler Craton from east to west; significant gold mineralisation is known in the area (e.g., Tarcoola and Glenloth) and there is potential for channel mineralisation (e.g. gold, uranium, groundwater; Hou, 2008). The vast blankets of surficial cover have masked much of the geology of the KPS drained region, including the history of the paleochannels. Older suites of rocks, in particular those more prone to weathering, such as regolith derived from various basement rocks, are similarly largely hidden.

The present surface of the KPS indicates little of the complex paleovalley architectures beneath. Drilling reveals Quaternary sediments concealing a complex pattern of Paleogene–Neogene paleochannels commonly incising and overlying weathered bedrock, with channel sediments ranging in thickness from a few metres to 145 m (Hou, 2008). Complex relationships between the Quaternary sediments, Paleogene–Neogene channel fills, and older (e.g., Mesozoic) sediments and deeply weathered basement can make paleochannel identification difficult. Paleogene–

Neogene fluvial, lacustrine and even estuarine to marginal marine sediments accumulated in the KPS during the Middle to Late Eocene and Middle to Late Miocene and Pliocene.

Anthony Paleovalley occurs at ~200 km northwest of KPS and Narleby Paleovalley is located at ~200 km south of KPS (Hou, 2008). Due to an absence of drillhole data in most parts of these major paleovalleys, the distribution and sedimentation of the channel sediments in these paleovalley areas are poorly understood. In the Anthony Paleovalley, the relative thinness of the poorly sorted and silicified gravelly alluvial sands in the upper reaches of the channel is probably related to the high elevation of this area; clay-rich carbonaceous deposits do not occur because marine transgression did not reach to this area (Hou et al., 2001a). The lack of Miocene lacustrine clays in the Anthony Paleovalley indicates either a depositional environmental difference from the Garford/Tallaringa Paleovalley that occur to the north of the Anthony Paleovalley, or post-Miocene erosion in the Anthony Paleovalley (Hou et al., 2001a).

GARFORD PALEOVALLEY

The Garford Paleovalley was studied in detail by Hou et al. (2001a) using sedimentary analysis of numerous cross- and longitudinal sections, depth-to-basement maps, and TEM and gravity surveys in places across the paleovalley (see Figs 3.26 and 3.27 in Hou et al., 2001a). It is noted that the general trend is for the paleovalley base to rise from ~80 m a.s.l. at the bayline (Posamentier et al., 1988) to 120 m a.s.l. at drillhole 5638-253 in the middle reach of the Garford Paleovalley, giving a gradient of ~0.1%. This negative gradient in the lower channel reach (-0.16‰) may be due to post-Eocene uplift along the northeastern margin of the Eucla Basin (Hou et al., 2008). The current expression of the drainage lows and channel bases therefore rarely coincides with the deepest parts of the paleovalley, particularly in their lower reaches.

TALLARINGA PALEOVALLEY

As a large and subtle valley, Tallaringa Paleovalley system occurs in the northeastern margin of the Eucla Basin, showing pale-toned and branching features on satellite images (Fig. 3.1). Although it can be identified by numerous methods, such as topography, geological maps, DEM, Landsat TM and NOAA images (Figs 2.2, 2.3 and 2.4), the information provided by sparse drilling is limited, and profiles and sections for interpreting the sedimentation therefore are affected. Same as the Garford Paleovalley, the Miocene lacustrine clay facies (Garford Formation) fills a depression eroded in the Eocene carbonaceous sandy facies (Pidinga Formation) (Hou et al., 2001a). The top of the channel-fills has been silicified and ferruginised, forming a duricrust along the valley (Benbow et al., 1995a). The drainage channel is prominent as a dark linear feature on the NTT imagery due to the cooling response of the channel fill (Fig. 2.4). The Tallaringa paleovalley system has a maximum channel length of over 280 km upstream of the present-day Wilkinson Lakes.

LINDSAY PALEOVALLEY

The Lindsay Paleovalley system occurs in the northern margin of the Eucla Basin (Fig 3.1) and generally drains southwards into the Eucla Basin from the Musgrave Province (Krapf et al, 2019). The use of Airborne Electromagnetic (AEM) geophysical survey data has allowed detailed mapping and identification of paleochannels and their distributaries, hidden beneath recent cover (Soerensen et al, 2018). The recent drilling in the Lindsay East Paleovalley has further defined two paleovalley units, the Eocene fluvial Pidinga Formation and the Miocene to Pliocene lacustrine Garford Formation identified by Zang and Stoian (2006). These units are well known from other paleovalleys located along the northern margin of the Eucla Basin (Hou et al. 2003, 2006, 2008). In the Lindsay Paleovalley, the Pidinga Formation is represented fluvial deposits of paleovalley fill, while the Garford Formation shows marine influences at the beginning and end of the Late Miocene to Early Pliocene deposition of fluvial-lacustrine, brackish to estuarine/marginal marine sediments. During the Early Pliocene to Pleistocene, the Garford Formation transitions from marginal marine to estuarine depositional environments through to fluviolacustrine, freshwater depositional environments (Krapf et al 2019).
The paleochannel fill is overlain by a pedogenic calcrete horizon up to 50 cm thick (Krapf et al 2019). Periods of evaporation and drying up of the paleovalley led to the deposition of a 1.25 m thick gypsum layer (Krapf et al 2019).

ROE PALEOVALLEY

The Roe Paleovalley system, bounded to the north and to the south by the Rebecca and Lefroy Paleovalleys respectively, drained easterly into the Eucla basin and flowed through the Lake Roe area (Fig. 3.1). A variety of Middle – Late Eocene sediments was deposited in the Roe Paleovalley system, including its tributaries such as Black Flag, Kunanalling, Bonnie Vale, Hannan and Wollubar paleochannels, which are now concealed by a surficial cover of Quaternary age, and the broad present-day valleys are occupied by discontinuous playa-lakes (Smyth and Button, 1989; Hocking, 1990; Clarke, 1993). The paleorivers within Roe Paleovalley system incised deep, narrow valleys into the Archean bedrock, and the thalwegs of the paleovalleys have low gradients that are steepest (greater than 1 m per km) in the upper parts of the catchments, especially where the drainage lines traverse relatively resistant greenstone belts (Kern and Commander, 1993).

The paleochannels in the Roe Paleovalley system (catchments) are generally V-shaped with few U-shaped and W-shaped in the cross sections, ranging about 400 to 700 m in width, and 25 to 40 m in depth, but the corresponding widths and depths are about 1,000–1,500 m and 55–75 m respectively in the east of Kalgoorlie (Kern and Commander, 1993). The lower reaches of the main paleochannels (towards to the Eucla margin, i.e., east of Lake Rivers) are infilled with Middle Eocene marine sandstone overlain by spongelite of Late Eocene age (Jones, 1990). These sediments were described by Kern and Commander (1993) as Middle Eocene Wollubar Sandstone (grey to buff sandstone with minor clay and lignite, equivalent to Werillup Formation in the Lefroy Paleovalley, see below) and overlying Late Eocene Perkolilli Shale (grey and mottled clay with minor sandy clay, equivalent to Pallinup Siltstone in the Cowan Paleovalley, see below).

LEFROY PALEOVALLEY

The Lefroy Paleovalley, including Lake Lefroy, is in the southeastern part of the Archean Yilgarn Craton (Fig. 3.1), and within the paleodrainage network defined by Bunting et al. (1974) and Van de Graaff et al. (1977). The bedrock surface beneath the Lake Lefroy surface defines a V-shaped paleovalley (Clarke, 1993). In the Lefroy paleovalley, the channel sediments of the Werrilup Formation, which consists of laminated red-brown to green silts, white, grey or black clays and silts, and lignite (Clarke, 1993), rests unconformably on Archean basement, and interfingers with the marine sediments of the Hampton Sandstone (Clarke et al., 2003). In Lake Lefroy the Werrilup Formation, which is equivalent to the Pidinga Formation in the eastern Eucla paleovalleys, reaches a maximum known thickness of 60 m in CD 1916 (Clarke, 1993; Clarke et al., 2003). The presence of the Hampton Sandstone, a widespread unit at the base of the Paleogene-Neogene succession of the Eucla Basin (Lowry 1970), in the Lefroy Paleoriver to the west of Kambalda was recognised by Jones (1990), reaching a maximum thickness of 24 m in KD 3001 (Clarke, 1993).

The Princess Royal Spongelite Member is common along the margins of Lefroy and Cowan paleovalleys, having been removed by erosion from much of the central portions, but is more continuous further east (Clarke, 1993). The Princess Royal Spongelite Member, lithologically consisting of over 50% siliceous sponge spicules with lesser amounts of silt and clay overlying or interfingering the Werillup Formations, reaches a thickness of 12 m beneath Lake Lefroy in drillhole CD 1916 (Clarke, 1993; Lyneham, Knutsen and Dirks, 2019). The Princess Royal Spongelite onlaps directly onto Archean basement along the shores of Lake Lefroy near Loves Find (Clarke, 1993).

The Paleogene sediments in the Lefroy Paleovalley are fluvial to shallow marine in origin, contrasting with the predominantly lacustrine sediments of the Neogene succession. In Lake Lefroy, red sands and ferruginous sandstones predominate along the margins of the lake and small lenses of fine sand and ferruginous sandstone occur within the Revenge Formation beneath the lake. The Revenge Formation reaches a thickness of 17 m in Lake Lefroy (Clarke, 1993). The Miocene Gamma Island Dolomite occurs in Lake Lefroy, which is equivalent to the Cowan Dolomite in Lake Cowan (Clarke et al., 2003). Resting disconformably on the erosional top of the Revenge Formation is the Pliocene to Holocene Roysalt Formation, which is lithologically and stratigraphically equivalent Polar Bear Formation in Lake Cowan and occurs across the floor of Lake Lefroy and on its margins (Clarke, 1993).

COWAN PALEOVALLEY

The Cowan Paleovalley, in which Lakes Cowan and Dundas occur, is in the southeastern part of the Archean Yilgarn Craton (Fig. 3.1) and within the paleodrainage network defined by Bunting et al. (1974) and Van de Graaff et al. (1977). The bedrock surface beneath the lakes defines a V-shaped paleovalley (Clarke, 1993). In the Cowan Paleovalley, the channel sediments of the Werrilup Formation, rests unconformably on Archean and Proterozoic basement, but the Hampton Sandstone is absent (Clarke, 1993). The Werillup Formation is disconformably overlain in the paleodrainages by the Princess Royal Spongolite, and is overlain by the Pallinup Siltstone along the south coast (Clarke, 1993). The Norseman Limestone, a fossiliferous marine carbonate, crops out along the southeastern margins of Lake Cowan, and occurs widespread beneath Lake Cowan, reaching a maximum known thickness of 37 m (e.g., in drillhole ET 120R; Clarke, 1993). The Princess Royal Spongolite is also common along the margin of the Cowan Paleovalley and is more continuous further east (Clarke, 1993). The Princess Royal Spongolite reaches a thickness of 21 m on the northern shores of Lake Cowan, near Bingeringie (Hooper, 1959).

The southern extension of the Cowan Paleovalley (previously into the Bremer Basin) at Esperance contains calcareous marine sandstones in the lower Werillup Formation (Cockbain, 1967; Morgan and Peers, 1973), which are almost certainly equivalent to the Norseman Limestone (Clarke, 1993). The Neogene Revenge Formation in Lake Cowan is composed of similar lithologies to those found in Lake Lefroy. The dark green-grey sediments of presumed Neogene age at Lake Cowan can be correlated to the Garford Formation, which is widely distributed in the eastern Eucla paleovalleys. Also, in Lake Cowan the Miocene Cowan Dolomite crops out along the lake shores, and Pliocene–Holocene Polar Bear Formation extends across the lake floor and lake margins, and at some sites occurs together with low relief gypsum dunes (generally less than 2 m high; Clarke, 1993).

4. SEDIMENTOLOGY OF THE BASIN SEDIMENTS

The sediments of the Eucla Basin and peripheral paleovalleys provide an excellent record of the evolving marine environments of the Southern Ocean and the terrestrial hinterland of the Australian continent. This overview of the sedimentology of the Eucla Basin is mainly based on studies of the eastern basin (e.g., Pitt et al., 1978; Benbow, 1983b, 1990a, 1991; Alley and Lindsay, 1995; Benbow, et al., 1995a; Rankin et al., 1996; Alley et al., 1999; Hou et al., 2001a, b, c, 2006, 2008, 2011b, 2021; Krapf et al., 2019; Zang and Stoian, 2006) and is correlated and compared to the western basin (e.g., Fairbridge, 1953; Lorry, 1970; Quilty, 1974; Hocking, 1990; Jones, 1990; Clarke, 1993, 1994a, b; Clarke et al., 2003), and offshore basin including the continental shelf (e.g., James and Bone, 1991, 1992, 1994; Li et al., 2003, 1996a, b). As one of the main objectives of this study is to gain a regional understanding of the depositional, environmental and paleogeographic frameworks, a grasp of the stratigraphic and geographic evolution of the paleovalleys is required. For this, the sequence stratigraphic method, paleoclimatic and tectonic impacts, supplemented by studies in mineralogy, petrology, geochemistry and facies analysis of the sedimentary fills, have proved invaluable.

4.1 STRATIGRAPHY AND LITHOLOGY

4.1.1 Introduction

The successions of the Eucla Basin show a remarkably consistent stratigraphy across the entire basin, including its inshore and offshore extensions (Clarke et al., 1996; James and Bone, 2000; Hou et al., 2008). The major phases of deposition are present in the basin and almost all adjacent paleovalleys, and probably are equivalent to third-order cycles in the marine record (Benbow, 1990a; Clarke et al., 2003; Hou et al., 2003a, 2006b). Such lateral continuity in inshore and marginal marine facies across over 2,000 km between the west and east of the basin highlights the environmental controls across the whole basin, including eustacy, tectonics, climate, hydrology and oceanography. Clarke et al. (2003) considered that the Eocene sediments that infill paleovalleys of the Bremer Basin in southwest Western Australia should be included in the margin of the Eucla

Basin; and the Eocene sediments of the Polda Basin are likewise a marginal extension of the eastern Eucla Basin.

Previously, the Cenozoic succession of the Eucla Basin was divided into the Eucla Group of marine limestones (Singleton, 1954), and later included the Hampton Sandstone, Toolinna Limestone, Abrakurrie Limestone and Colville Sandstone (e.g., Lowry, 1968a, b, 1970), and the Immarna Group of predominantly terrigenous sediments (e.g., Benbow et al., 1995a). There exist a few hiatus between the largely continuous (conformable) depositional sequence sets within the Cenozoic succession of the Eucla Basin (Li et al., 2003; Clarke et al., 2003; Hou et al., 2003c, 2006b, 2008) but the time-bounded description of the stratigraphic sequences better reflects the evolution and development of the paleoshorelines that are genetically related to HM accumulation along the Eucla margin (Hou et al., 2003b; 2011b). The correlation of sediments east to west across the basin and from offshore to onshore made use of many earlier studies, often on specific sites within and marginal to the basin (Hou et al., 2008). The record of marine, marginal marine, estuarine, fluvial, and lacustrine environments is characterised by an extensive borehole dataset and spans five major depositional phases during the Paleocene – Early Eocene, Middle – Late Eocene, Oligocene – Early Miocene, Middle Miocene – Early Pliocene, and Pliocene–Quaternary (see Appendix 1). These phases identify the key role of eustatic change, during which highstands inundated the craton margins and flooded paleovalleys as far as 400 km inboard of the present coastline (Hou et al., 2006b; 2011b).

Many stratigraphic units occur continuously across the entire Eucla Basin. The paleovalley successions in both western and eastern onshore basins have the same stratigraphic architecture. This continuity illustrates the very strong allostratigraphic control on sedimentation (Clarke et al., 2003). Due to conflicting and overlapping nomenclature in places across the whole basin, it has become necessary to revise some parts of the stratigraphy (Clarke et al., 2003; Hou et al., 2008) in contrast to previous works (e.g., Pitt et al., 1978; Benbow, 1983b, 1990a, 1991, 1993; Benbow et al., 1995a; Alley et al., 1999). For example, some stratigraphic unit names, such as Eucla Group, Immarna Group and Redmine Group, Hampton Sandstone, Toolinna Limestone (e.g., Figure 10.22, p180; Drexel and Preiss, 1995), have not been used much or revised because not only do they cross genetic (sequence) stratigraphy units across multi depositional settings but also they are somewhat superfluous or conflicting when describing the formation-level units in chronological order relating to sea-level events (Clarke et al., 2003). A revision of stratigraphic correlations across the basin provided new correlations relating marine-coastal sediments to the stratigraphic record of offshore sediments (Clarke et al., 2003; Hou et al., 2003b). This included a reconstruction for sequences of coastal deposition in the context of well-constrained chronology of major sea-level events (e.g., McGowran et al., 1997). The clarified and revised aspects of the nomenclature are summarised as follows:

1. Eucla Group includes all units of offshore sediments deposited in the marine environment, i.e., Hampton Sandstone, Wilson Bluff Limestone, Nullarbor Limestone, Abrakurrie Limestone; and abandoned Mullamalang Member, Toolinna Limestone (Li et al., 2003).
2. Immarna Group contains all units of nearshore-onshore sediments deposited in the marginal marine – fluvial channel environments across eastern margin of the basin, i.e., Pidinga Formation, Ooldea Sand, Khasta Formation, Barton Sand, Yarle Sandstone, Colville Sandstone, Garford Formation, Narlaby Formation, Ilkina Formation, Munjena Formation, abandonding Burdunga Subgroup (eastern basin).
3. Eundynie Group (Cockbain, 1968b) contains all units of Paleogene nearshore-onshore sediments deposited in the marginal marine – fluvial channel environments across the western margin of the basin, i.e., Werrilup Formation, Norseman Formation, Pallinup Siltstone, Princess Royal Spongolite.
4. The Redmine Group, used for Neogene successions of non-fossiliferous, yellow, brown, and red alluvial and lacustrine sediments deposited in the western Eucla margin (e.g., Smyth and Button, 1989; Jones, 1990), includes all units of nearshore-onshore sediments deposited in the Neogene marginal marine – fluvial channel environments across western margin of the basin, i.e., Colville Sandstone , Revenge Formation, Cowan Dolomite, Gamma Island Dolomite, Polar Bar Formation, Roysalt Formation,?Poelpena Formation, ?Wanilla Formation (Clarke, 1993, 1994a, b).

5. The name Hampton Sandstone is restricted to its original usage, i.e., Middle Eocene calcareous marine sand underlying the Wilson Bluff Limestone in the central Eucla Basin; and abandonded the usage in the eastern nearshore-onshore settings (Clarke et al., 2003).
6. Pidinga Formation, correlated to Werrilup Formation in the west, is redefined to include all carbonaceous Eocene sediments found along the margins of the Eucla basin, including those in the paleochannels, and non-carbonaceous sediments genetically related to or interbedded with carbonaceous sediments.
7. Khasta Formation, correlated to Pallinup Formation and Princess Royal Spongolite in the west, consisting of Late Eocene estuarine sand with siliceous sponge spicule facies in the eastern Eucla coastal plain, is elevated to formational status; the previous designations, Khasta and Bring Members, are abandoned. Thus, it is recommended that all Late Eocene estuarine sediments with siliceous sponge spicule facies are named by Pallinup and Khasta Formation in the western and eastern extremities of the Eucla Basin, respectively (Clarke et al., 2003).
8. The previous Ooldea Sand is divided into Ooldea Sand and Barton Sand (Hou et al., 2006b), owing to their development in different paleoshorelines and at different times (see below).

It should be noted that not all features are continuous over the entire region due to some factors, such as erosion, facies changes and tectonic impacts (Hou et al., 2006b, 2008). For example, major sources of discontinuity formed by subaerial erosion between the depositions of the marine successions resulted in the erosion and removal of some sediments deposited during the previous transgression, so that the sediments of the late transgression were deposited directly on bedrock in places. The discontinuities caused by facies changes are well developed in profiles from onshore to marine. The most striking is the very abrupt lateral transition from inshore biosiliceous sedimentation of the Khasta and Pallinup formations to the central basin carbonate of the Wilson Bluff Limestone and the development of the barriers and lagoons of the eastern Eucla Basin (Clarke and Hou, 2000). Within individual units, there are also abrupt facies changes caused by local features, such as paleochannel geometry, the role of basement/bedrock highs in the estuarine and channel settings. Despite these changes, the overall impression is of remarkable lateral continuity of stratigraphy.

4.1.2 Principal stratigraphic units

The principal stratigraphic units of the Eucla Basin and their relationship and correlation are shown on Figures 4.1, 4.2 and 4.3. During the Paleogene–Neogene, the Eucla Basin was largely marine carbonate platforms with inner margins dominated by terrigenous clastics. The marine sediments (mainly carbonates) of the Eucla Group are predominant over the Bunda Plateau, while the mainly terrigenous sediments of the Immarna, Eundynie and Redmine Groups are extensive around the coastal plain and extend landward into paleovalleys through estuaries.

Eucla Group (offshore)

The Eucla Group is defined here for all sedimentary units deposited in the offshore setting of the Eucla Basin, including marine carbonates and clastics. Apart from the basal marine clastics (e.g., Hampton Sandstone), three major carbonate units were defined from bottom to top: Wilson Bluff Limestone, Abrakurrie Limestone, Nullarbor Limestone and a more restricted younger carbonate unit Roe Calcarenite (Fig. 4.1). The contacts between the units of the Eucla Group are erosional and mostly irregular, indicating unconformities (Figs 4.2 and 4.3; Feary et al., 1994; James and Bone 1991, 1994, 2000).

UNDIFFERENTIATED PALEOCENE – EARLY EOCENE SEQUENCES

The existence of undifferentiated Paleocene – Early Eocene sequences is interpreted from seismic evidence within these early Paleogene terrigenous sediments and are associated with strong local reflectors in the Late Cretaceous section, which are interpreted as sills (Fraser and Tilbury, 1979). During the Paleocene – Early Eocene, undifferentiated marine clastics of and/or beneath the Hampton Sandstone and undifferentiated fluvial clastics of the Pidinga Formation were probably deposited along parts of the southern basin (Taylor, 1975; Barten, 1975; Davies et al., 1989; Benbow et al., 1995a; Clarke et al., 2003, Hou et al., 2003c, 2006b). In the far west, Paleocene – Early Eocene sediments form a prograding wedge (Bein and Taylor, 1981).

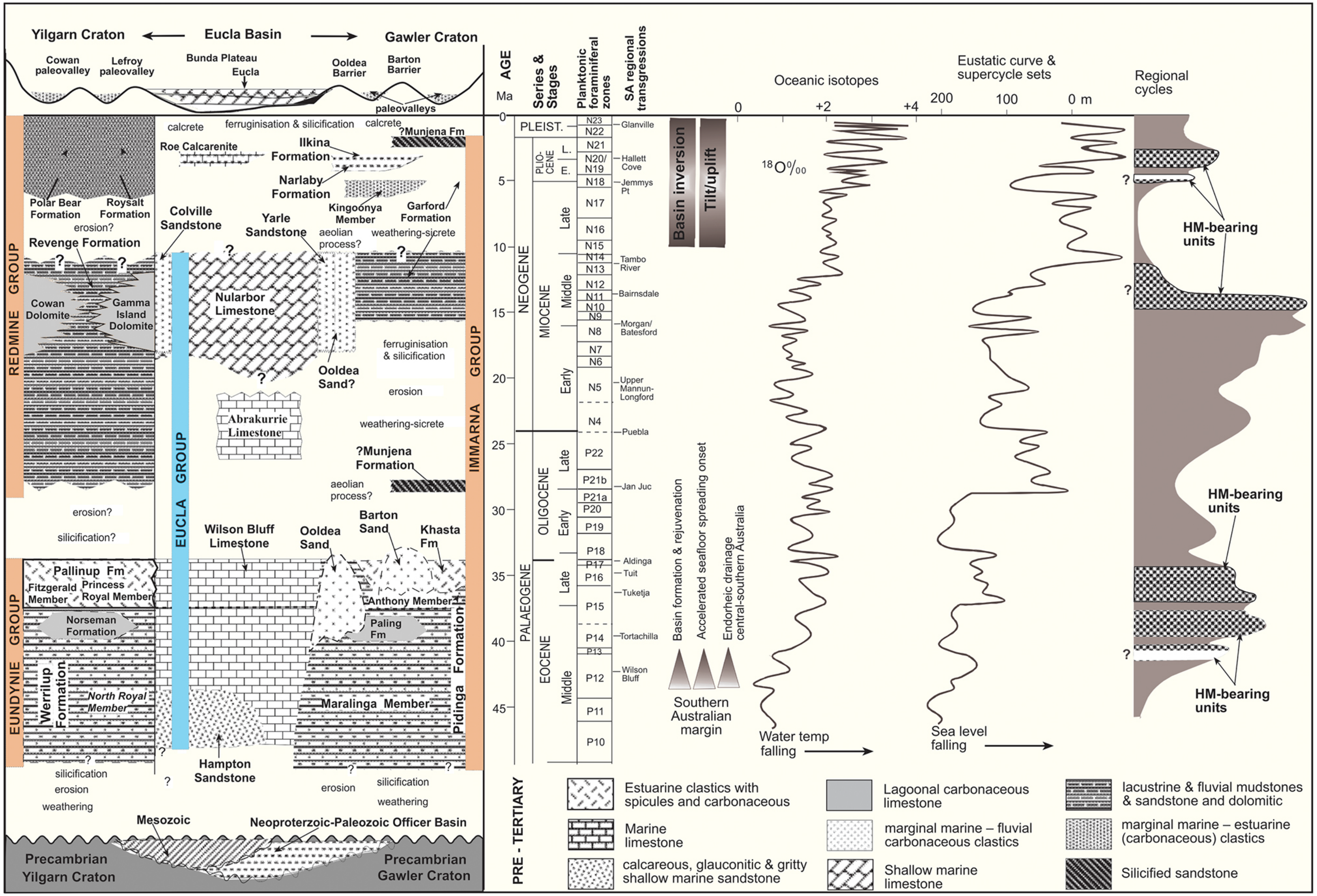


Figure 4.1 Stratigraphic correlation chart and inferred depositional environments in the Eucla basin and paleovalleys (from Hou et al., 2011b). Planktonic foraminiferal zones (Taylor, 1983, 1986; McGowran et al.,1997), regional transgressions (McGowran et al.,1997), tectonic events in southern Australia (Li et al., 2004), oceanic oxygen-isotopic cycles (Abreu et al., 1998), eustatic supercycle sets (Haq et al., 1987, 1988), and local cycles (McGowran et al., 2004) in relation to age of HM-bearing units. Modified from Hou et al. (2008, 2011b).

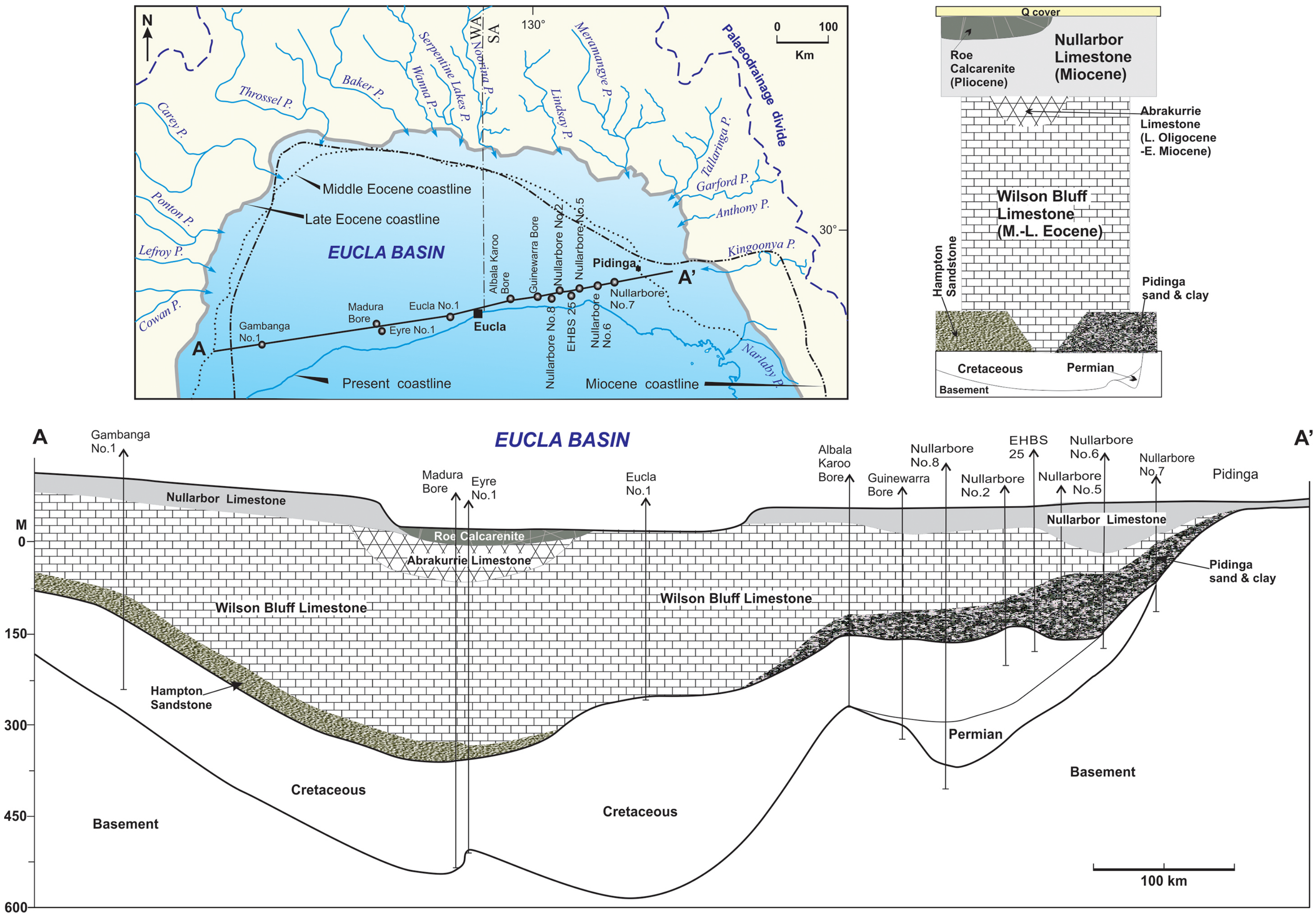


Figure 4.2 West-east cross-section of the Eucla Basin showing stratigraphic relations (after Lorry, 1970; Li et al., 2004; Hou et al., 2003c, 2008, 2011b).

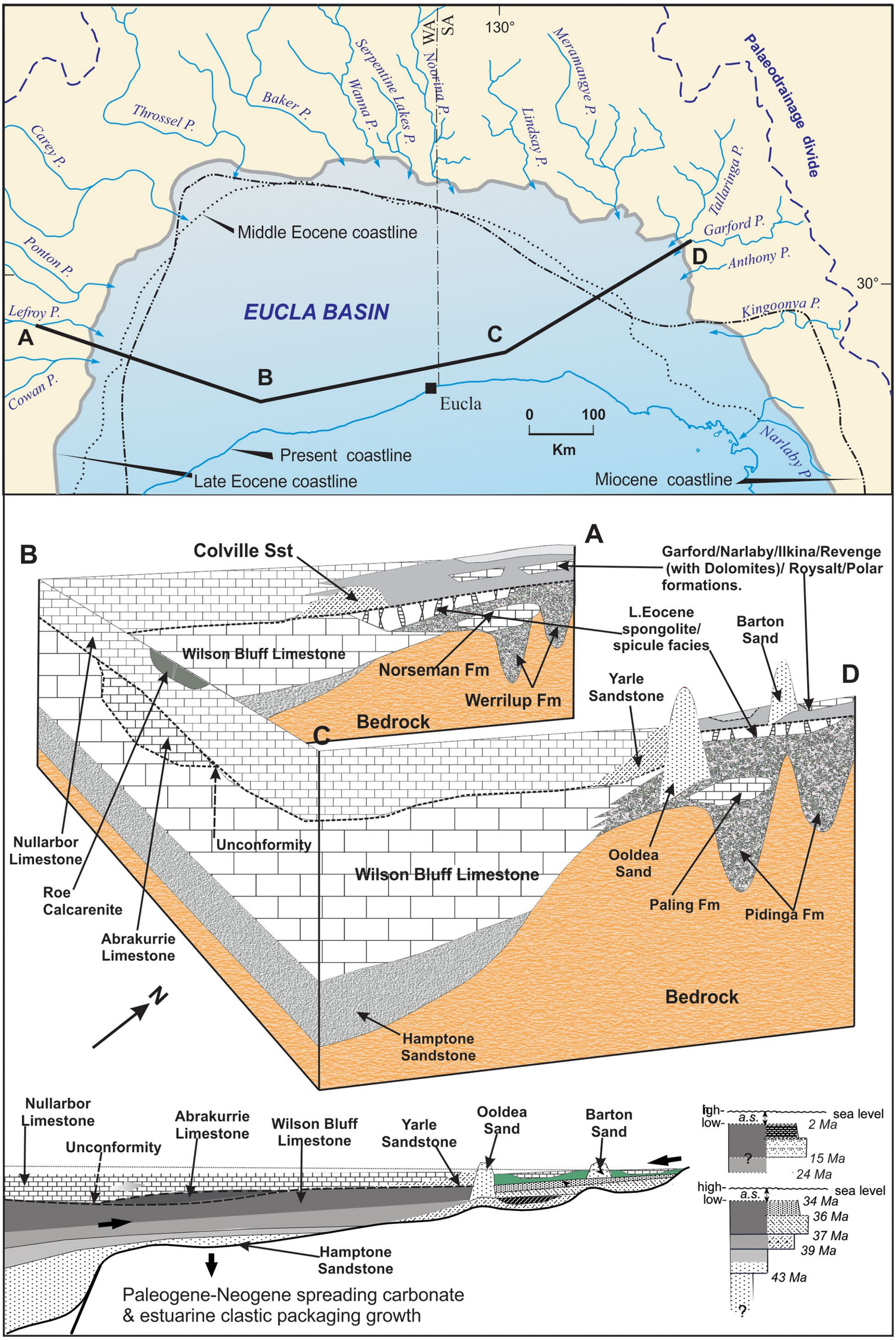


Figure 4.3 Schematic fence diagram showing stratigraphical relationships across the Eucla Basin and adjacent onshore sediments in marginal marine and paleovalley

settings. A schematic interpreted relationship between marine carbonate deposition and the extent of marginal marine carbonate and clastic equivalents over time is summarised in the inset diagram (from Hou et al., 2008).

MIDDLE TO LATE EOCENE MARINE SEQUENCES

Hampton Sandstone

As a discontinuous sand unit, Hampton Conglomerate (Fairbridge, 1953) was found at the base of the Wilson Bluff Limestone in Transcontinental Railway Bore No. 1. The Hampton Sandstone (name derived from the adjacent Hampton Range; the type section: the Transcontinental Railway No. 1 Bore at Madura: lat. 31°54'35"S, long. 127°00'20"E) was formally defined by Lowry (1968a) for the base of the Tertiary succession in the central Eucla Basin, where it consists of sand and sandstone, often calcareous and gritty, containing marine fossils, that underlies the Wilson Bluff Limestone. Subsequent to its original definition, the term Hampton Sandstone was used by other authors more loosely for other stratigraphic units. For example, Jones (1990) and Clarke (1993) correlated sand interbedded and underlying carbonaceous clay in the Lefroy Paleovalley in Western Australia as Hampton Sandstone, while Alley and Lindsay (1975) and Benbow (1990a, b, 1993) correlated the sand interbedded and underlying the Pidinga Formation in the eastern margin of the Eucla Basin as Hampton Sandstone. However, Clarke et al. (2003) believed that the term Hampton Sandstone should be restricted to lithologies occurring in this stratigraphic position at the base of the Wilson Bluff Limestone (Fig. 4.1).

The Hampton Sandstone comprises very fine-coarse grained sand and grit and is partly clayey at the base and glauconitic and fossiliferous at the top, where it is overlain by Wilson Bluff Limestone. The lensoid to sheet-like marine Hampton Sandstone is widespread in the central and southwestern basin and mostly overlain by the Wilson Bluff Limestone (Benbow, 1990a). In the western and central parts of the Eucla Basin, the basal Hampton Sandstone of Middle Eocene age (foraminiferal Zone P12; Fig. 4.1) comprises up to 35 m (in the Zanthus-Kitchener area) of shallow marine (greenish grey medium-grained) sands overlain by up to 300 m of mostly cool-water carbonates of the Wilson Bluff, Abrakurrie and Nullarbor Limestones (Lowry, 1970; Hocking, 1990; Jones, 1990; Clarke et al., 2003).

Palynological data, such as the dinoflagellates *Deflandrea phosphoritica* Eisenack, several species of *Hystrichosphaeridium*, and pollen grains of *Nothofagidites* spp. and *Proteacidites* sp. obtained from drillhole samples of the 160-mile-peg bore on the Eyre Highway and Kanandah No. 12A Bore (Lowry, 1970), suggested a middle Eocene age (Fig. 4.1). An early Middle Eocene age of the Hampton Sandstone is also supported by the fact that it apparently conformably underlies the Eocene Wilson Bluff Limestone (Ludbrook, 1958) and a Middle Eocene age was established for the base of the Wilson Bluff Limestone (Ludbrook, 1963, 1967b), based on the occurrence of both *Pseudogloboquadrina primitiva* and *Globigerapsis* index in Core 4 from Gambanga No. 1 Well, and in the lower part of the formation in Eyre No. 1, of *Pseudogloboquadrina primitiva* and *Globigerina linaperta* (Lowry, 1970).

The Hampton Sandstone probably does not crop out and is known only from boreholes (Lowry, 1970). The formation occurs in some places across the central basin in the Naretha and Madura areas (Fig. 3.2 correlation chart) but it is apparently absent in bores along the railway east of Naretha and east of Madura (e.g., Transcontinental Railway No, 3 Bore and Eucla No.1 Bore). The formation is very variable in thickness and in places discontinuous, such as reaching a thickness of ~14 m in Kanandoh 12A Well, 10 m in Transcontinental Railway No. 1 Well, 11 m in Eyre No. 1 Well, 28 m in Gambanga No. 1 Well and probably exceeds 85 m in the 160-mile-peg bore on the Eyre Highway, east of Balladonia (Lowry, 1970), suggesting an irregular base of the Eucla Basin and an initial depocentre located in the southwestern part of the Eucla Basin.

Wilson Bluff Limestone

The Wilson Bluff Limestone, covering most of the Bunda Plateau, is a temperate bryozoan water limestone that formed during the Middle and Late Eocene (Singleton, 1954; McWhae et al., 1958). The type section is located at Wilson Bluff (at the west end of the Bunda Cliffs; McWhae et al.

1958; Ludbrook, 1958; Lowry, 1970). Initially, Tate (1879) referred to this formation at Wilson Bluff as the 'White polyzoal limestone', but the formation excludes the 'yellow polyzoal bed' of Tate (1879) (Ludbrook, 1958; McWhae et al., 1958), which later was named by Lowry (1968a, 1970) as the Abrakurrie Limestone (see below).

Typically, the Wilson Bluff Limestone consists of a fine-grained to chalky limestone, medium-to-thick-bedded, commonly rich in bryozoans and echinoids, and locally rich in molluscs and brachiopods (Lowry, 1970; Clarke at al., 2003). To the southwest, the uppermost part of the Wilson Bluff Limestone consists mainly of cross-bedded bryozoan grainstone and rudstone, which was distinguished as the former Toolinna Limestone of Lowry (1968a, 1970) who described Eocene bryozoan grainstone along the western margin of the Eucla Basin, in contrast to the chalkier lithology of the Wilson Bluff Limestone. The term Toolinna Limestone has been suggested to be abandoned by Clarke et al. (2003) as the type locality is of Oligocene Abrakurrie Formation (see below) and other occurrences mapped as this unit are of Eocene Wilson Bluff Limestone (James and Bone, 1994; Li et al., 1996a). The Wilson Bluff Limestone, averaging <150 m thickness in the eastern basin (e.g., east of the Coompana Block) but increasing westwards to ~300 m (Eyre-1) in the central southern part of the Basin (Fig. 4.2), is mostly white to grey wackestone, skeletal mudstone (the mud is micritic and microbioclastic) and rudstone with minor packstone (Ludbrook, 1960; Lowry, 1970). The basal part of the formation beneath the Bunda Plateau is locally sandy and glauconitic marl. Bryozoan-derived skeletal clasts are the major framework component of the Wilson Bluff Limestone, and glauconite occurs throughout the formation in trace or minor amounts and locally up to 25–30% with marl as at the base (e.g., Singleton, 1954; Ludbrook, 1958, 1960, 1961; McGowran and Lindsay, 1969; Lowry, 1970; Benbow et al., 1995a). The glauconitic facies with a higher terrigenous component than is typical for most of the formation at the base, appears to be widespread across the basin (e.g., Ludbrook, 1960; Lowry, 1970).

The Wilson Bluff Limestone ranges in age from Middle Eocene to latest Eocene (Fig. 4.1). Exact correlation of the Wilson Bluff Limestone is difficult, despite its chalky nature, because of the rarity of planktonic foraminifers as a result of extensive winnowing or an oligotrophic environment (Clarke et al., 2003). Outcrop samples from near the type locality of the Wilson Bluff Limestone indicate that it contains *Wadella* cf. *hamiltonensis*, suggesting a Middle to Late Eocene (P14–P16) age (Li et al., 1996a). The basal part of the formation in the region of the Bunda Plateau was recognised as middle Middle Eocene (Ludbrook, 1963, 1969; McGowran and Lindsay, 1969), which later was put in Zone P12 by McGowran (1989). Late Middle Eocene ages were also determined for the formation in the east part of the basin in Outback Oil - Pickrell Mallabie Water Well and to the west of SADME NRD-13 (near Yalata; Lindsay and Harris, 1975). Conflicting with these, the evidence was also presented by Lowry (1970) and Philip (in Lowry, 1970) for a late Late Eocene age for the top of the formation at the type section and Point Culver. Therefore, this extensive deposition of biogenic platform carbonates commenced during the Wilson Bluff Transgression, containing a well-preserved Middle Eocene planktonic foraminiferal fauna correlated with upper Zone P12 (McGowran and Lindsay, 1969; McGowran, 1989). Although the Tortachilla Transgression (McGowran 1989) has not been recognised, to date, from the Wilson Bluff Limestone, it occurs in the Paling Formation (see below) that is the equivalent of the Wilson Bluff Limestone (Clarke at al., 2003). The recorded occurrence in the Wilson Bluff Limestone at Point Culver of the foraminifers *Truncorotaloides collactea*, *Globigerina linaperta* and *Globorotaloides turgidus* (Clarke et al., 2003), together with the data in Lowry (1970) from Point Culver, Booanya Rock and Balladonia, suggest a P14–P17 zonal age range, with P15–P17 as the most likely. Consequently, the Wilson Bluff Limestone at Point Culver is most probably of Tuketja age, as is the spicular Pallinup Formation along its western margin (Clarke et al., 2003).

The Wilson Bluff Limestone, as the basal carbonate unit of the Eucla Basin, disconformably rests on black shale of the Cretaceous Madura Formation in many areas of the basin, unconformably rests on the Precambrian basemen rocks in places and on Permian or Proterozoic rocks in the northern part of the basin, and conformably overlies the Hampton Sandstone in the southwestern central basin (Lowry 1970; Rankin et al., 1996; Hou et al., 2008), and is interbedded with the Pidinga Formation along the eastern Eucla margin (Benbow et al., 1995a; Hou et al., 2003a, b, c; 2006b). In the southwestern part (e.g., Toolinna Cove) of the Eucla Basin, the lower part of the Wilson Bluff Limestone is overlain and also passes laterally into the coarser-grained upper part (former Toolinna Limestone of Lowry, 1968a, 1970). The unconformity between the Wilson Bluff Limestone and overlying Abrakurrie (in the Madura area) and Nullarbor Limestones (see below) on the Bunda Plateau is regarded as a complex, multi-generation contact (James and Bone, 1991). In the southwestern and central parts of the basin, Wilson Bluff Limestone is overlain unconformably by the Late Oligocene – Early Miocene Abrakurrie Limestone (Lowry, 1970; James and Bone, 1994; Li et al., 1996a), and farther to the north and northeast, by Middle Miocene Nullarbor Limestone (Lowry, 1970), suggesting a northeast migration of the Eucla Basin depocentre from Eocene to Miocene times (Hou et al., 2008).

The Wilson Bluff Limestone, deposited on a wide continental shelf, is developed throughout most of the Eucla Basin (Lowry, 1970). Beyond the Bunda Plateau, there are also isolated occurrences of bryozoan limestones. About 150 km west of the Bunda Plateau, Norseman Limestone (Gregory, 1916) occurs in the Norseman area and is confined to a tributary of the Lefroy paleoriver that formerly flowed into the Basin (Clarke, 1993, 1994a, b). Also, beyond the Bunda Plateau, to the northeast, there are stratigraphically equivalent occurrences, such as Paling Formation (Fig. 4.1). Several of these are grainy bryozoan limestone and were included previously in the Wilson Bluff Limestone. The stratigraphy of grainy bryozoan facies in the Eucla Basin is not well understood. Lindsay and Harris (1975) regarded the limestone (and associated calcareous sandstone) north of Maralinga in the eastern basin, as likely correlatives of the distinctly coarser grained Toolinna Limestone (Lowry, 1968a). Similar facies also occur at various levels and in both proximal and distal positions within the Wilson Bluff Limestone. However, extensive limestone readily referable to Toolinna Limestone does not appear to occur extensively about the basin's east margin, contrary to the expectation of Lindsay and Harris (1975). Consequently, Clarke et al (2003) proposed that the term of the Toolinna Limestone be abandoned as the type locality is of Abrakurrie Formation and other occurrences mapped as this unit are of Wilson Bluff Limestone.

LATE OLIGOCENE – EARLY PLIOCENE MARINE CARBONATES

Abrakurrie Limestone

The Abrakurrie Limestone, with exposures limited to the subsurface, in caves below and along the southern margin of the Nullarbor Plain, was first recognised by Tate (1879) as the 'yellow friable polyzoal bed' at Wilson Bluff. It was grouped it variously with part or all of the Tertiary limestones of the basin (Brown, 1885; Maitland, 1901; McWae et al., 1958; Ludbrook, 1958) but later given separate formation status by Lowry (1968a) (the 32.9 m thick type section in Abrakurrie Cave, ~60 km west of Wilson Bluff, lat. 31°39'20"S, long. 128°29'20"E). The Abrakurrie Limestone has maximum thickness of 90 m observed at Mullamullang Cave (Fig. 4.4), and 100–120 m at the coast (James and Bone, 1994).

The Abrakurrie Limestone, known only from the central part of the basin and known limits occurring approximately 150–230 km from the Early/Middle Miocene shoreline, is a shallow shelf bryozoal limestone of Late Oligocene – Early Miocene age (Benbow et al., 1995a; Li et al., 1996a). The position of the shoreline at this time is unknown. It may have been much further inland (?50–100 km) as post-depositional erosion prior to and during initial deposition of the Nullarbor Limestone is likely. The relatively thick and localised occurrence of Abrakurrie Limestone near Madura compared with the overlying thinner and extensive Nullarbor Limestone suggests local tectonic subsidence in this part of the basin. Subsidence would have occurred adjacent to the major elevated crystalline basement ridge in the east (the Coompana Block of Flint and Parker, 1982).

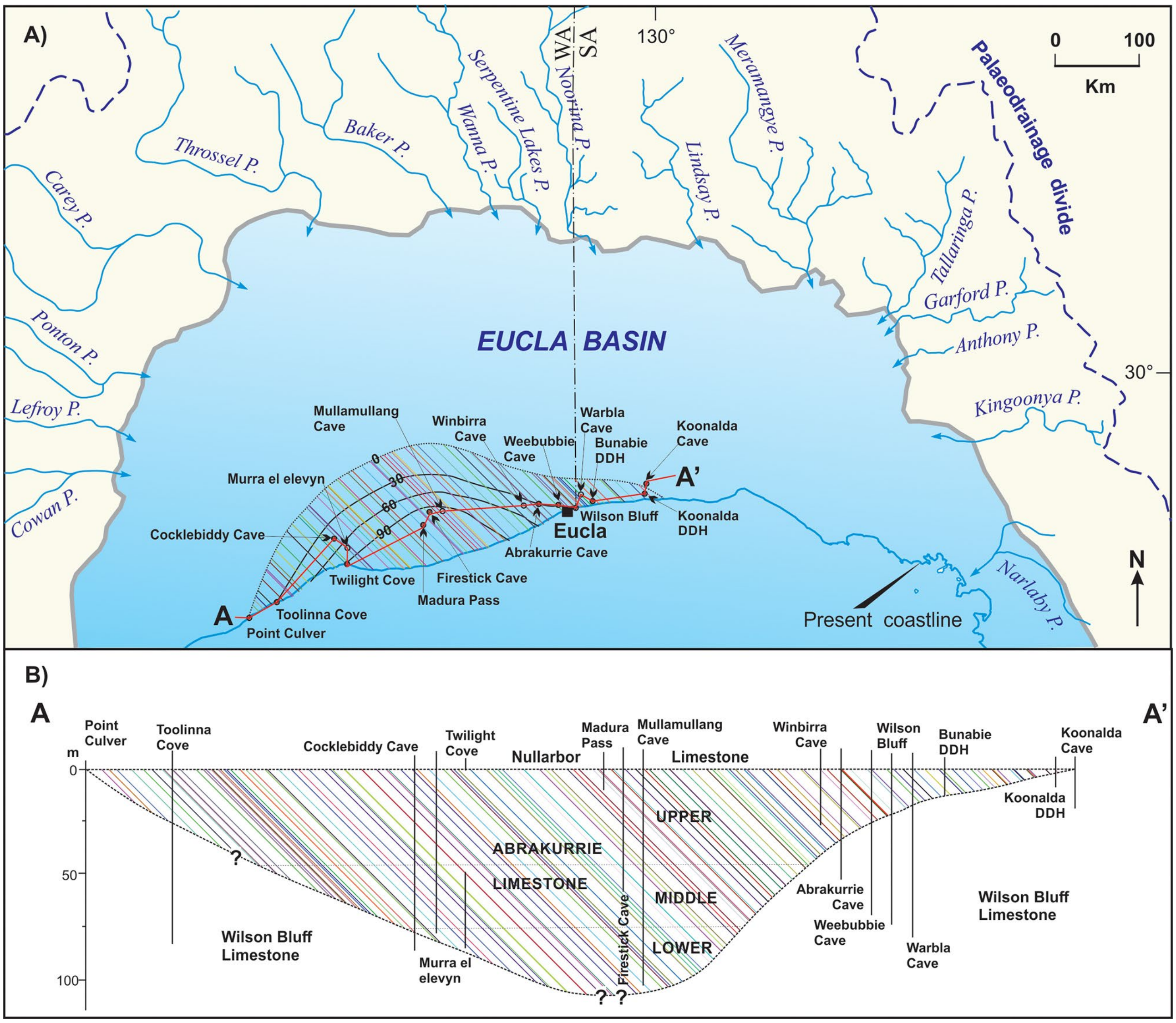


Figure 4.4 **(a)** Distribution and variations in thickness of the Abrakurrie Limestone. **(b)** West-east cross-section (A-A') showing the generalised stratigraphy at 16 localities (studied by James and Bone, 1991, 1992, 1994). Modified from Li et al. (1996a).

The Abrakurrie Limestone, containing abundant *Victoriella conoidea* in a microfauna considered to be upper Janjukian (latest Oligocene) in age (Fig. 4.1), consists predominantly of bryozoan-rich skeletal limestone ranging in texture from wackestone to rudstone deposited in cool to temperate water (mainly in water depths around 100 m; Fig. 4.4; Li et al. 1996a) on a partly to completely drowned platform, which did not contain terrigenous clastic grains (James and Bone, 1991). Three types of hardground-bounded cycles were recognised by James and Bone (1992, 1994) in the sequence: cross-bedded, thin-bedded and thick-bedded cycles. The Late Oligocene to Early Miocene (c. 21–28Ma, approximately equivalent to the planktonic foraminiferal zones P22 to low N4; Li et al. 1996a). The limestone is composed of coarse-grained bryozoan calcarenite (James and Bone 1991; Li et al. 1996a), is distinctly cyclic, contains numerous hardgrounds (James and Bone 1992), and is locally dolomitised (James et al., 2006).

In the central basin, the Abrakurrie Limestone occurs stratigraphically and disconformably between the underlying Middle–Late Eocene Wilson Bluff Limestone and the overlying Early–Middle Miocene Nullarbor Limestone (Figs 4.2 and 4.3) (Lowry, 1968a, 1970; Li et al., 1996a; James et al., 2006). The Abrakurrie Limestone is not readily distinguishable from former Toolinna Limestone of Late Eocene based on lithology alone (Lowry, 1970; Lindsay and Harris, 1975). The disconformity between the Abrakurrie and overlying Nullarbor Limestones probably represents much of the Early Miocene (Benbow et al., 1995a). 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 (Ludbrook, 1958; Lowry, 1970) is mostly a bioclastic and micritic limestone (hard grey or yellow indurated calcarenite, algal limestone) deposited over much of the emergent Eucla Basin, with the Ooldea Range acting as a barrier, separating marine deposition on the platform from extensive lacustrine sedimentation inland in the eastern basin (Fig. 4.3; Benbow, et al., 1995a). The lower part of the Nullarbor Limestone shows predominantly floatstone and rudstone textures and is an algal limestone composed of dominantly algal, bryozoal and echinoidal fragments, whereas the upper part shows mudstones to rudstones textures and has abundant miliolid foraminifera and common agglutinated and encrusting forms, including *Austrotrillina howchini* and *Marginopora vertebralis* (Li et al., 1996a), and also contains bivalves, gastropods, serpulid worm tubes, bryozoans and scleractinian corals (Lowry, 1967; O'Connell et al., 2012). 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 locally the base in the Ifould Lake area (Benbow, et al., 1995a). In contrast to the underlying carbonates of the Wilson Bluff Limestone and Abrakurrie Limestone, the Nullarbor Limestone is remarkably uniform in thickness, averaging 20–35 m with a known maximum of 45 m or >60 m towards the coast. The Nullarbor Limestone was deposited in warm temperate water in a shallow platform setting (Benbow, et al., 1995a). The Nullarbor Limestone grades laterally into the calcareous Colville Sandstone inboard, and contains a negligible terrigenous clastic component, except at the base and in nearshore settings, representing widespread shallow marine deposition (O'Connell, 2011) and indicating that river systems carried little debris to the coast.

Lowry (1970) described a variety of fossils commonly occurring in the Nullarbor Limestone. Foraminifera results from the Nullarbor Limestone, such as *Lepidocyclina howchini* from the basal part of the limestone, suggest an early Middle Miocene (Benbow and Lindsay, 1988) and Middle Miocene age (Lindsay, 1987), and a species restricted to Zones N8–N9 equivalents (early Middle Miocene) in southern Australia (Lindsay and Harris, 1975; McGowran, 1979; Li and McGowran, 1995).

In most areas of the basin the Nullarbor Limestone disconformably overlies the Eocene Wilson Bluff Limestone but overlies the Abrakurrie Limestone in the central basin (Lowry, 1970; Benbow, et al., 1995a; Li et al., 1996a). The limit of the Nullarbor Limestone along the southern edge of the plateau is relatively well established (e.g., the cliffs in Fig. 3.2), but in most other areas there are few outcrops, and the limit has been mapped on the basis of topography and soil. In the northern part of the Eucla Basin, beds of the same age as the upper Nullarbor Limestone are dominated by sandstone of the Colville Sandstone (see below) that grades southwards into the Nullarbor Limestone (Jones, 1990). Around the margin of the Nullarbor Plain, the Nullarbor Limestone unconformably overlies the Eocene sediments (Pidinga, Khasta and Wilson Bluff Formations respectively) of the paleovalleys and of the margin of the Eucla Basin, but also older bedrock and crystalline basement (Fig. 4.2). The Nullarbor Limestone is the youngest of the marine limestones of the Eucla Basin, and it is overlain by a variety of soils, dunes, and other Quaternary deposits.

The absence of the Nullarbor Limestone over much of the southern edge of the plateau (Fig. 4.2) is believed to be due to post-depositional erosion (Lowry, 1970), but its absence from the southwestern part of the basin may be due to non-deposition because of a continental tilting movement during the Neogene (Sandiford, 2007; Hou et al., 2008). The basal member of the Nullarbor Limestone thickens southwestwards in a manner similar to that of the underlying Eocene formations (Lowry, 1970), indicating the depocentre continued from Eocene deposition. The Nullarbor Limestone is thickest beneath the Nullarbor Plain (it reaches 31.1 m in thickness on Old Homestead Cave; Lowry, 1970), and is absent from parts of the Balladonia Hampton Tableland, and thus it is reasonable to assume that the upper part of the Nullarbor Limestone thickened eastwards or at least maintained a constant thickness. This eastward thickening is therefore believed to be due to post-Eocene continental tilting movement and erosion that increased deposition towards the east, probably as a result of migration of the depocentre eastwards (Hou et al., 2008).

LATE PLIOCENE – EARLY PLEISTOCENE MARINE CARBONATES

Roe Calcarenite

Roe Calcarenite is the youngest marine carbonate deposited in the Eucla offshore basin veneering the Roe Plain (Fig. 3.5), covering approximately 8,000 km^2 at a thickness of 2–3 m, with a marine erosional surface cutting into the centre of the Great Australian Bight during the Late Pliocene to Early Pleistocene (Lowry, 1970; James and Bone, 2007). The Roe Calcarenite disconformably overlies the Wilson Bluff Limestone in the eastern parts of the Roe Plain, and the Abrakurrie Limestone elsewhere (Fig. 4.2; Lowry, 1970).

The Roe Calcarenite consists predominantly of calcarenite that is usually weakly lithified, coarse-grained, poorly-bedded, and highly fossiliferous (Lowry, 1970), comprising molluscs (a diverse range of bivalves and gastropods), echinoids, serpulid worm tubes, large benthic foraminifera and coralline algae (James and Bone, 2007). The rock contains 10–15% detrital clasts including quartz grains, intraclasts and lithic fragments, and shows common textures of rudstones, grainstones and floatstones (James and Bone, 2007). The faunal assemblage suggests a shallow marine setting, in which the accumulated calcarenite was at least periodically influenced by tidal currents and/or open ocean swell (James et al., 2006). Thus, the calcarenite was interpreted as a product of shallow marine, subtidal sedimentation in a seagrass-dominated environment (James and Bone, 2007).

Immarna Group (eastern nearshore-onshore)

The Immarna and Redmine groups are defined here for all sedimentary units deposited in the nearshore and onshore settings of both eastern and western margins of Eucla Basin, including marginal and non-marine costal and channel sediments.

MIDDLE TO LATE EOCENE MARGINAL-MARINE SEQUENCES

Pidinga Formation

Pidinga Formation (the type section in the old coal bore SADM P5 between 0.77–14.63 m, located on the floor of Ifould Lake, South Australia; ‘Pidinga Clay and Sands’ of Ludbrook (1958) was originally defined by Harris (1966) for the Middle to Late Eocene terrigenous sediments that are mostly carbonaceous in the eastern Bunda Plateau region of the basin, and subsequently applied to some of the paleovalley occurrences (e.g., Lindsay and Harris, 1975; Barnes and Pitt, 1976; Bembow et al., 1995a; Hou et al., 2000). The Pidinga Formation has very limited outcrop or subcrop (generally <2 m exposure) is restricted to playa lakes, in the northeast and east margin of the basin (e.g., Ifould Lake and Lake Tallacootra) and along the paleovalleys (e.g., Lakes Anthony and Bring, eastern Eucla margin).

The Pidinga Formation, averaging 30–60 m in thickness, comprises marginal marine and non-marine lignitic and pyritic sands, clays, silts and lignite deposited during reactivation of paleorivers. The basal and lower part of the Pidinga Formation, particularly in the paleovalleys, consists of gravels and sands which fine upward into carbonaceous sediments. Colour varies from pale grey to white where clean, to pale to dark brown and black where they are carbonaceous, but carbonaceous sediments are bleached white occasionally. The reduced carbonaceous sediments are interbedded with oxidised sediments in places (e.g., in the Barton region and Ooldea Range to the west, as in cored SADME Ooldea Range-6, eastern Eucla margin; Figs 4.2 and 4.3). Carbonaceous sediments and lignite may make up a minor but significant part to all of the sequence, occurring as single thin beds to stacked beds over 40 m thick in places. Oxidation is evidenced by orange to yellow-brown iron oxide staining on quartz and clays. Quartz is the major detrital component (generally >90%) and other minor minerals include muscovite, feldspar and polymineralic grains with carbonised wood and leaf remains, particularly in the carbonaceous sediments.

The Pidinga Formation displays a range of common sedimentary structures, including bi-directional crossbedding, wavy cross-bedding and lamination in the tidal influenced channel sediments, but massive structure dominates fluvial sediments (Hou et al., 2001a). The degree of rounding varies generally from poor, particularly in areas of crystalline basement and at the base, to good, towards up through the sequence. Upward fining cycles can be recognised and these may be capped by carbonaceous mud (common) or clay (less common). Boundaries of sand overlying clay (e.g., the outcrop in the Ifould Lake area, eastern Eucla margin) and carbonaceous muds (e.g., in drillholes) are usually sharp, indicating clearly erosional conditions.

The Pidinga Formation is widespread in the eastern Eucla margin (e.g., Wilkinson Lakes region and the regions between the Ooldea and Barton ranges and between Lake Tallacootra and Chundie Swamps, up to 100 m thick and 40 km wide), and infills most paleovalleys in the eastern onshore basin (e.g., Narleby, Kingoonya, Anthony, Garford, Tallaringa, Meramangye paleovalleys) occurring upstream at least about 350 km inland from the present coast. The channel width of the Pidinga Formation fill averages between 2.5–5 km (Hou et al., 2001a).

The Pidinga Formation unconformably overlies Archean to Middle Proterozoic crystalline basement and locally on sedimentary rocks of Neoproterozoic–Paleozoic or Mesozoic age, such as Officer Basin and Eromanga Basin sediments over much of the eastern Eucla Basin (Figs 2.1 and 4.2). Due to the limited drillhole controls, however, the relationship of Pidinga Formation with the Hampton Sandstone in the nearshore-offshore setting (beneath the Wilson Bluff Limestone) was uncertain (Ludbrook, 1969), where Lindsay and Harris (1975) believed the relationship with the Wilson Bluff Limestone was an unconformable one. Beneath the Bunda Plateau, the Pidinga Formation is interpreted to conformably be overlain by and intertonges with the Wilson Bluff Limestone, as demonstrated on the east margin as in the Colona–Nundroo region and in Endeavour Resources W5-WB-1 (Hou et al., 2006b). Towards the northeastern margin of the basin, the formation intertongues with the basal part of the overlying Wilson Bluff Limestone (zone P14/P15 age). The youngest part of the Pidinga Formation, of latest Eocene age, intertongues with the Wilson Bluff Limestone as in the western areas of the Ooldea barrier and is overlain uncomformably by the Nullarbor Limestone and is overlain conformably and may be interbedded with the Khasta Formation or Paling Formation (see below) towards the landside of the Ooldea and Barton barriers. In the onshore setting, the Pidinga Formation is overlain by the Garford Formation (estuarine and fluvial settings). The upper Pidinga overlies the highstand flooding sediments of the lower Pidinga or the Paling Formation (if present) and is in continuous succession with the overlying Khasta Formation. Along the eastern margin of the Eucla Basin, marginal marine sands of the Pidinga Formation were found to contain concentrations of heavy minerals (Benbow, 1990b).

Age of the Pidinga sediments is based largely on palynological determinations by Alley and Benbow (1989), who assigned the Pidinga above the Paling Formation to the middle *Nothofagidites asperus* Zone of the late Eocene. Regionally, however, the age of the Pidinga Formation is suggested to range from Middle–Late Eocene to possibly early Oligocene (Lindsay and Harris, 1975; Alley, 1985; Alley and Benbow, 1989; Alley and Beecroft, 1993; Zang and Stoian, 2006). A Middle to Late Eocene age is indicated by terrestrial palynomorphs in numerous drillholes (Pitt et al., 1978; Zang and Stoian, 2006) but the base of the Pidinga Formation is not well dated in the paleovalleys. Initial sedimentation in the paleochannels is not always datable and at any rate may not be preserved due to the dynamic processes of stream incisement and seasonal flows. The lignitic sediments deposited in the paleovalley incised in the western side of the Ooldea Range in CRAE RCH2 drillhole, underneath the Nullarbor Limestone, was palynologically dated as Early Eocene age and correlated with *Proteacidites asperopolus* Zone (Zang and Stoian, 2006), indicating the existence of early Eocene paleovalleys on the Eocene margin of the basin.

Maralinga Member

Between the Ooldea and Barton Ranges, the Pidinga Formation is divided into upper and lower members by the Ooldea Sand and the correlative Paling Formation (if present; Figs 4.1 and 4.3). The lower Pidinga Formation, and its correlative Maralinga Member of the Pidinga Formation (Zone P14) along the eastern margin, is the principal basal Paleogene sequence, unconformably overlying and onlapping Cretaceous and older sediments locally, and also Precambrian crystalline basement (Fig. 4.1). At Tietkins Well, to the north of Maralinga, the lower Pidinga Formation intertongues with the basal part of the overlying Wilson Bluff Limestone (Zone P14/15) and passes laterally (northwards) into the basal sand equivalents of paleovalley fills (Benbow et al., 1995a). Massive to locally laminated, poorly sorted carbonaceous (gravelly) sands, clays, and (sandy/clayey) lignites, generally make up much of the carbonaceous sediments of the lower Pidinga Formation.

Maralinga Member occurs entirely in the subsurface, and the type section is from the borehole CRAE RCH 2 (228637mE, 6658008mN, BARTON 1:250,000 map sheet, 98 m thick, extending from 35 to 133 m below surface). It is the direct equivalent of North Royal Member (see below), the lower part of Werrilup Formation in Western Australia. The upper surface of the Maralinga Member is gradational into the calcareous Paling Formation (see below), where present, or else truncated by a disconformity, and the basal surface of the unit unconformably overlies older sediments or crystalline basement (Clarke et al., 2003).

Anthony Member

The upper Pidinga Formation (Zone P16/17), and its correlative Anthony Member of the Pidinga Formation on the eastern margin, intertongues with the associated facies (Paling Formation) of the Wilson Bluff Limestone. It is overlain disconformably and may be interbedded with the Khasta Formation (Figs 4.1 and 4.3; Clarke et al., 2003; Hou et al., 2003a, b, c, and 2006b). Anthony Member lacks fluvial gravels and coarser sands and consists of estuarine channel sediments. Relatively well-sorted carbonaceous sand units with clays, and (sandy/clayey) lignites, with black, carbonised wood and leaf fragments, are dominant in the upper Pidinga Formation (Hou et al., 2003a).

Paling Formation

In the eastern Eucla Basin, isolated occurrences of a thin sandy glauconitic and carbonaceous limestone, a marginal equivalent of the Wilson Bluff Limestone, named the Paling Formation conformably overlies and is intercalated with the Pidinga Formation (e.g., in Tietkens Well; Hou et al., 2003a). Like the Pidinga Formation, the unit is also highly carbonaceous with abundant plant fragments. The Paling Formation was deposited during the late Middle Eocene (Zone P14/15, e.g., Tietkens section north of Maralinga; Fig. 4.1) Tortachilla Transgression (McGowran, 1989; Alley and Beecroft, 1993; Clarke et al., 2003), in the lagoons along the northeastern margin of the Eucla Basin (landward of the Ooldea Range; Fig. 4.3; Benbow, 1990a). The Paling Formation, consisting of 75% carbonate, 20% quartz and 5% glaucony as sand- to silt-sized grains (Benbow et al., 1982), is not in stratigraphic continuity with the Wilson Bluff Limestone (of which it was formerly a member), although it is laterally equivalent to it. The limestone of the Paling Formation passes laterally into calcareous sand (see cross-sections in Rankin et al., 1996 and Alley and Benbow, 1989). The Paling Formation contains a diverse marine biota, together with abundant glaucony, which indicates normal marine salinities (Clarke et al., 2003). It is found landward of the Immarna and Pidinga passages through the Ooldea Range (Clarke and Hou, 2000). Thus, either exchange between the ocean and the lagoon behind the barrier of the Ooldea Range was sufficient for normal salinity to be established and with it a normal marine biota, or carbonate sand from the shelf was swept through the passes and deposited as flood-tide delta lobes.

The sandy, glauconitic limestone was intersected in the subsurface in drillholes CRAE RCH 2 (the type locality, at depths between 28 and 35 m), KD 1 and OR 7, and Tietkens Well – Maralinga region; previously referred to as Paling Member of the Wilson Bluff Limestone in the correlation chart of Alley and Beecroft (1993; Rankin et al., 1996). The Tietkens well provides a complete and best studied section (with 11.4 m of complete section, over the interval 28.4–16.7 m below surface) with lower and upper boundaries clearly visible, making this an important section of the formation for this part of the Basin. These occurrences are indicative of a flood-delta in an inter-distributary bay setting (Hou et al., 2006b), and comprise very fine-grained skeletal (mainly bryozoal) grainstone or skeletal limestone or lime sand, with minor glauconite and quartz (Rankin et al., 1996). 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). This is interpreted as an equivalent of the Wilson Bluff Limestone deposited in the area of flood-delta adjacent tidal channels in the landward side of the Ooldea Barrier (Hou et al., 2011b).

Ooldea Sand

Benbow (1990a) recognised the Ooldea, Barton and Paling Ranges as an Eocene barrier island complex and named the unit comprising them the Ooldea Sand. Clarke and Hou (2000) suggested that the Ooldea Sand in the Ooldea Range was deposited during the late Middle Eocene (i.e., Tortachilla Transgression), which was informally referred by Clarke et al. (2003) as the ‘lower’ Ooldea Sand that can be differentiated from the ‘upper’ Ooldea Sand deposited during the Late Eocene (i.e., Tuketja Transgression) where they occur in different dune systems. Hou et al., (2006a) redefined these two units as Ooldea Sand for the Ooldea barrier dominantly deposited in the Ooldea Range area and Barton Sand (see below) for the Barton barrier mainly deposited in the areas of the Barton and Paling Ranges respectively, although no physical or boundary criteria are presently known that would allow them to be differentiated should they occur in the same stratigraphic section.

The large dunes of the Ooldea barrier, formed during the late Middle Eocene, are respectively composed of Ooldea Sand (Figs 4.1 and 4.3). The only exposure of the Ooldea Sand is in the Immarna cutting on the Trans-Australian railway line, to the east of Immarna Siding. The Ooldea Sand is predominantly barrier sands up to 112 m thick (at Maralinga; Benbow, 1988), conformably overlying Pidinga Formation (Hou et al., 2006a). The western Eucla margin lacks the late Middle Eocene costal barrier facies of the Ooldea Sand. The Ooldea Sand contains heavy minerals (Benbow, 1990b), and is composed predominantly of white to pale grey medium-fine barrier/dune quartz sands that vary from clean to slightly clayey. The sand grains are well rounded and some show aeolian frosting. The top of the Ooldea Sand is usually overlain by red-brown sands of Quaternary age (Rankin et al., 1996; Hou et al., 2008, 2011b). It is almost impossible to directly determine the age of Ooldea Sand due to the absence of fossils, but the sequence relationship of the lower stratigraphic unit suggests it is of late Middle Eocene age.

Khasta Formation

Spicule-bearing sands and black, glauconitic, sapropelic muds containing abundant marine dinoflagellates, spicules, foraminifera, and bryozoal and echinoid fragments (Flint and Rankin, 1991) are mainly deposited in the eastern margin of the Eucla Basin during Late Eocene. This estuarine facies was revised as the Khasta Formation (Figs 4.1 and 4.3), mainly deposited on the landward side of the Ooldea and Barton Ranges and is partly the equivalent of the former Bring Member of the Pidinga Formation (Benbow, 1986a; 1993), which represents the highstand deposition of the estuarine environment during the Late Eocene Tuit-Tuketja Transgression (Clarke et al., 2003; Hou et al., 2006b, 2008). The Khasta Formation (Clarke et al., 2003) is established to encompass the Late Eocene estuarine sand and two spicule-bearing units [former Bring Member (Benbow, 1986a) of the Pidinga Formation, and former Khasta Member (Benbow, 1993) of the Hampton Sandstone (Fairbridge, 1953; Lowry, 1970)]. In areas where deep post-depositional incision has not affected the Pidinga Formation, the Khasta Formation conformably overlies the Pidinga Formation.

Lithologically, this formation consists of fine to medium grained sand and silt, with tidal-estuarine structures including parallel and ripple lamination, cross-bedding, reactivation surfaces, compound bedding, wavy and lenticular bedding, rip-up clasts, trace fossils in the form of escape burrows and an abundance of siliceous sponge spicules. Sorting is commonly moderately good to good. These spicule-bearing fine-grained sediments occur mostly in the areas of more confined deposition, such as the estuary and estuarine channels particularly landward of the Ooldea and Barton barriers over the eastern basin, and extended landward to the eastern margin of the basin in the paleovalleys, such as Anthony, Kingoonya, Cummins, Wanilla and Yaninee paleovalleys and are widespread in the Lake Maurice and Lakes-Anthony-Bring regions (Binks and Hooper, 1984; Rankin and Flint, 1991; Benbow et al., 1995a). They also extend landward as far inland as 15–20 km upstream of the paleovalleys (e.g., Garford and Narlaby Paleovalleys; Rankin et al., 1996). These sediments also have equivalents on the western margin of the Eucla Basin in the Norseman–Kalgoorlie region (Figs 4.1 and 4.3; Princess Royal Spongolite of Glauert, 1926; Clarke et al., 2003).

The trace-fossil assemblage inland, dominated by *Skolithos*, *Thallasinoides* and *Ophiomorpha*, with body fossils of lithistid sponges, are present in some localities of the eastern basin margin, such as paleovalleys near Lake Bring and Arthurs Lake, northeast of Ceduna (e.g., Narlaby Paleovalley; Rankin and Flint, 1991), although not in the abundance characteristic of the Fitzgerald Member of the Pallinup Formation of the western basin (Clarke et al., 2003). In the Cummins and Wanilla paleovalleys in the eastern basin margin (Fig. 1.1), also, there are such sediments composed of carbonaceous spicular clay and spicular sand containing dinoflagellates assignable to the Corriudinium incompostum Zone (Harris and Foster, 1974; Harris, 1985) and a sparse fauna of siliceous spicules, foraminifers and very fragmentary bryozoans in glauconitic and carbonaceous sediments (Lindsay, 1974).

Barton Sand

The large dunes of the Barton barrier, formed during Late Eocene (Fig. 4.1), are respectively composed of Barton Sand. Clarke and Hou (2000) interpreted the Ooldea Sand of Benbow (1990a) in the Barton and Paling Ranges as being the barrier complexes developed during Late Eocene (i.e., the Tuketja Transgression; Fig. 4.1), which was informally referred by Clarke et al. (2003) as the ‘upper’ Ooldea Sand, and then redefined as the Barton Sand by Hou et al. (2006a). Lithologically, there are no discernible differences between the Ooldea Sand and Barton Sand; heavy minerals are also present in Barton Sand. The Barton Sand is conformably overlying Khasta Formation (Clarke et al., 2003; Hou et al., 2003a, b, c) and usually is overlain by red-brown sands of Quaternary age (Rankin et al., 1996; Hou et al., 2006a). As is the case with the Ooldea Sand, the Barton Sand is unfossiliferous, but the sequence relationship of Barton Sand suggest it is of Late Eocene to Early Oligocene age (Benbow, 1990a). The western Eucla margin lacks the Late Eocene costal barrier facies of the Barton Sand.

LATE OLIGOCENE – EARLY PLIOCENE SEQUENCES

Yarle Sandstone

During the Middle Miocene, a narrow coastal terrigenous belt (<10 km wide) of sandstones was deposited along the northeastern margin of the Nullarbor Plain, adjacent to the Ooldea Range. The sandstones, at the Yarle Lakes type area, are referred to as the Yarle Sandstone (Benbow, 1990c), and are interpreted as of Middle Miocene age, based partly on the occurrence of *Marginopora vertebralis*-bearing clasts of Nullarbor Limestone (Benbow, 1989). At Yarle Lakes, this silicified and ferruginous sandstone contains rare clasts of syndepositionally deformed Nullarbor Limestone and fossils. Equivalent sediments of the Colville Sandstone occur in a belt up to 120 km wide around the western and northern margin of the Nullarbor Plain (Figs 4.1 and 4.3; Lorry, 1970).

The Yarle Sandstone consists of fine to medium grained quartz sand that is moderately well-rounded and well-sorted and may contain minor bioclastic sediments (Benbow, 1990c; Rankin et al., 1996). The Yarle Sandstone, mostly silicified and ferruginised, belongs to an upper shoreface facies that was deposited in a moderate energy strandline (Rankin et al., 1996). It is correlated basinward with the Nullarbor Limestone and landward with the Ooldea Sand (Benbow et al., 1995a). Deposition occurred along a northwest-trending Middle Miocene shoreline comprising mainly reworked Ooldea Sand of the Ooldea barrier/dune system (Hou et al., 2008). In the Ooldea-Yarle Lakes region where the formation is known to be most extensive, a strandline of moderate energy is indicated for this part of the Eucla margin, at least during Middle Miocene deposition (Rankin et al., 1996).

Garford Formation

The Garford Formation (type section: drillhole SADME TPS-4, 10 m thick, including the nearby outcrop of around 3 m on the margin of a playa, the Garford Paleovalley;) comprises fluviolacustrine (sandy, illitic, and dolomitic) clays (palygorskitic in places) and dolomitic carbonate (Fig. 4.1; Benbow and Pitt, 1978). Sand facies, carbonaceous, cross bedding and parallel lamination are rare in the northeastern (e.g., Tallaringa, Garford, Anthony) paleovalleys (Benbow, 1993; Hou et al., 2001a), but are present in the southwestern (e.g., Kingoonya and Narlaby) paleovalleys (Hou, 2008). A marked and abrupt colour change from greenish grey to greyish white is generally associated with the lithological change from illite-dominated clays to dolomitic clays. The base is commonly sandy clay to rarely clayey sand, changing upward from argillaceous units to dolomitic clay or to dolomitic carbonates. In drillholes the base is generally difficult to recognise because of apparent reworking of older sediments and infiltration of the basal sediments into the underlying sequence (Hou et al., 2001a).

The Garford Formation is restricted in occurrence to paleochannels and lakes along the eastern margin of the basin and associated paleovalleys, such as Tallaringa, Garford, Anthony, Kingoonya and Narlaby paleodrainage systems. In the northeastern margin of the basin, the Garford Formation comprises argillaceous green to grey muds, white-grey to green-grey lacustrine carbonate muds and mudstones deposited in depressed areas between the Ooldea and Barton Ranges, in the Wilkinson Lakes area, and within paleovalleys (Fig. 4.3; Rankin et al., 1996; Benbow et al., 1995a). In the southeast Eucla margin (e.g., Kingoonya and Narlaby paleovalleys on the west and southwest margin of the Gawler Ranges), a marginal marine and estuarine channel sequence of the Garford Formation (Kingoonya Member, Hou et al., 2003a) and Narlaby (Benbow et al., 1995a) were deposited during the Middle Miocene–Pliocene (Fig. 4.1).

The sediments of the Garford Formation, unconformably overlie the Pidinga and Khasta formations (e.g., in the marginal basin) and older bedrocks as well as crystalline basement (e.g., in the paleovalley areas), include stromatolitic, oncolitic and oolitic carbonate and argillaceous mudstones. It is possible that very marginal Garford Formation may intertongue with the youngest part of the Ooldea Sand. Upstream in the northeastern onshore basin (e.g., Tallaringa Paleovalley and in the Lake Phillipson area), Garford Formation unconformably overlies Mesozoic sediments of the Eromanga Basin. Within the Narlaby Paleochannel, the Garford Formation is overlain by the Narlaby and Ilkina Formations (e.g., in the vicinity of Scrubby Peak Homestead), a sequence up to 60 m thick comprising fluvial to estuarine, fine to medium grained, moderately to well sorted sand with minor clay (Benbow et al., 1995a). The Garford Formation is most commonly overlain by Quaternary aeolian and fluvial sand (e.g., of the Great Victoria Desert), but prior to deposition of these sands, there was significant erosion of up to 20–25 m in places, such as in the Narlaby Paleochannel, and thus the upper boundary can be irregular and is clearly unconformable.

The exact age of these sediments is uncertain. The formation was probably deposited during the Middle Miocene to Early Pliocene interval (Benbow, 1986a, 1990a; Benbow et al., 1995a; Stoian, 2004b). Lithologically correlative sediments to the south of Tarcoola and to the southwest in the Narlaby Paleovalley are dated palynologically as Early Pliocene (Alley and Lindsay, 1995), while in the Immarna region they are dated on palynomorphs as Early – Middle Miocene. Palynological examination of carbonaceous sediments of the Kingoonya Member from the Narlaby and Kingoonya Paleovalleys suggests an Early Pliocene age for the younger phase of the Garford Formation (Harris, 1979; Truswell and Harris, 1982; Stoian, 2004b). The presence of rare microplankton implies a very weak marine influence in the Narlaby Paleochannel. Palynofloras from this part of the Garford Formation are very similar to Early Pliocene assemblages from paleochannels on the western margin of the basin (Bint, 1981), indicating that the phase of sedimentation was widespread.

Kingoonya Member

The Kingoonya Member 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. The Kingoonya Member, the younger phase of the Garford Formation deposited in the southeast Eucla margin (e.g., Kingoonya and Narlaby paleovalleys on the west and southwest margin of the Gawler Ranges), is characterised by a marginal marine and estuarine channel sequence (Hou et al., 2003a) that were deposited during the Middle Miocene–Pliocene (Fig. 4.1). Palynological analysis and dating of samples from selected drillholes in Kingoonya paleovalley (Stoian, 2004a, 2004b) indicates the presence of Late Miocene – Early Pliocene interval and correlates with the *Monotocides galeatus* Zone as defined by Macphail, 1999. The palynomorphs assemblages indicate a gradation of marine influence ranging from an open water marine through estuarine to lacustrine environments, following a northeastern arcing trend from NALARA NR 3-4 drillholes in the south through KINPC 1-2 to KIN45, and KIN20-21 and KIN 22 in the east and then back to CAR 37 in the north (Stoian, 2003b, 2003c, 2003d, 2004a, 2004b).

Narlaby Formation

Overlying the Garford Formation in the Narlaby Paleochannel area 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 (Figs 4.1 and 4.3; Benbow et al., 1995a). These sediments are partly silicified, iron-stained and varicoloured.

LATE PLIOCENE UNITS

Ilkina Formation

Following deposition of the Garford Formation, the Ilkina Formation (type section, of 4.2 m, on the north margin of a lake, 6 km southeast of Ilkina Hill on the northeast margin of the Narlaby Paleovalley) was laid down in the paleovalleys of the southeast Eucla margin Basin (on the southwest margin of the Gawler Ranges) during the Late Pliocene or Early Pleistocene (Benbow et al., 1995a). Extensive occurrences on the eastern margin of the Nullarbor Plain also occur along the northwest margin of Ifould Lake, where discontinuous outcrops occur as low-lying benches up to 2.5 m above the lake floor and as thin relatively indurated sheets on the lake floor. Possible occurrences have also been observed elsewhere, such as a ~2 m thick, horizontal, indurated and laminated gypsum bed localised on the southern margin of Lake Woorong east of the Garford Paleovalley (on COOBER PEDY; Benbow, 1982).

The Ilkina Formation is a thin sequence of conspicuously laminated clays, silts, sands and minor carbonate up to 5 m thick in the Narlaby Paleochannel (Figs 4.1 and 4.3), and also forms the floors of playas such as Ifould Lake where it contains clasts of silcrete and ferricrete (Benbow et al., 1995a). At Ifould Lake the formation sharply and unconformably overlies weathered Precambrian basement and carbonaceous sediments of the Pidinga Formation and near Ooldea ferruginised Yarle Sandstone. In the Narlaby Paleovalley the formation is comprised of pale to dark grey, very fine to medium grained sands (clayey in part, mostly weakly indurated to unconsolidated, moderately to well sorted), pale to dark grey and white clay (in part sandy) and ~1 m thick laminated gypsum cap. These sediments are laminated in part and very thin to medium, horizontally bedded. In the Narlaby Paleovalley, the formation unconformably overlies green to grey clays and white dolomite of the Garford Formation. Actual contacts with the Garford Formation can be sharp and clear, however, in many instances the contact can be difficult to pin-point precisely, presumably because of reworking. In both regions where the formation occurs it is unconformably overlain by aeolian sands of Quaternary age.

Munjena Formation

The Munjena Formation, commonly formed within a thin alluvial to colluvial unit around the eastern margin of the Nullarbor Plain and along the paleovalleys (Rankin et al., 1996), comprises poorly sorted conglomeratic sand, clayey sand and breccia, up to 3 m thick, blanket wide areas of interfluves and paleochannels (Benbow, 1993). These sand-rich sediments may contain local reworked quartz pebbles from underlying Paleozoic and Mesozoic sediments (Benbow et al., 1995a). 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 Oligocene and possibly younger, since it overlies Khasta and Pidinga formations in the paleochannels (Figs 4.1 and 4.3).

Eundynie Group (western nearshore-onshore)

The Eundynie Group, unconformably overlying the Precambrian basement and pre-Paleogene bedrocks in the western margin of the Eucla Basin, comprises Eocene nearshore and onshore facies.

MIDDLE TO LATE EOCENE MARGINAL-MARINE SEQUENCES

Werrillup Formation

The Werillup Formation in the type section (the Werillup 3 borehole near Albany prison; Cockbain, 1968b) is a succession of grey and black carbonaceous sands, silt and clay with local lignite found in rare outcrops along the southern coastal exposures of southwestern Western Australia and quite commonly in the subsurface (Cockbain and van de Graaff 1973; Elms et al., 1982). Clarke (1993) divided the Werillup Formation into informal upper and lower members separated by the Norseman Formation. The Werillup Formation is usually underlain by Precambrian rocks, such as at its type section where it overlies weathered Precambrian granite (Cockbain, 1968b). As an equivalent of the Pidinga Formation of the eastern basin, both the upper and lower members consist of upward-fining successions of basal sand and gravel grading into silt and then clay and with minor lignite in the upper part, which were mainly deposited in non-marine to marine environments along the western margin of the Eucla Basin, especially in the paleovalleys (Clarke et al., 2003).

Extended further west onshore paleovalleys of the Kalgoorlie area, two paleovalley filling units were defined as the Wollubar Sandstone and the Perkolilli Shale (Kern and Commander, 1993). The Wollubar Sandstone, filled in the deepest part of the paleovalleys, is interbedded with clays, and is often carbonaceous where unweathered, and is overlain by the Perkolilli Shale, which is always oxidised (Dusci, 1994). Clarke (1993) and Clarke et al. (2003) recommend that both these terms be replaced by the Werillup Formation and Revenge Formation respectively, because both units are laterally equivalent and probably laterally continuous to the Werillup and Revenge formations (Fig. 4.1).

North Royal Member

Clarke et al. (2003) proposed the North Royal Formation for all sediments previously placed in the lower Werillup Formation, named after the North Royal open pit at Norseman (387750E and 6443255N on the NORSEMAN 1:250,000 map sheet (SI51-02)). Approximately 35 m of Cenozoic sediment was exposed in the pit (Watchorn, 1980), of which the lower 13 m consisted of the North Royal Formation. North Royal Member used here, as the direct equivalent of the Maralinga Member in the eastern margin of the Eucla Basin, is a succession of Middle Eocene clastics and lignites, together with their lithified and weathered equivalents, which were deposited in non-marine to marginal marine environments along the margins of the Eucla Basin in Western Australia, especially in the paleovalleys (Fig. 4.1). For example, these Eocene sediments fill a north-trending paleovalley that was a tributary to the trunk Cowan paleovalley (Clarke, 1993). The unit consisted of an upward-fining succession with lithic cobbles at the base, fining upwards into sand and then clay (Clarke et al., 2003).

The basal surface of the North Royal Member was deposited in the North Royal paleovalley incised in weathered Archean basement. Its sharp contact with the overlying Norseman Member suggests that here there was a sudden change in depositional environment from non-marine or marginal marine to shallow marine, indicating a rapid marine transgression. This contact dates the North Royal Member as being older than the Tortachilla Transgression (Clarke et al., 2003).

Norseman Formation

A calcareous sand to grainstone, dominated by bryozoans, coralline algae, echinoids and molluscs in the western Eucla Basin, was defined by Cockbain (1968a) as Norseman Limestone that was redefined as the Norseman Formation by Clarke et al. (1996) to recognise the extent of poorly calcareous lithologies, chiefly sand and gravel, intimately associated with it in the subsurface. The type locality is just to the north of the Norseman township and the unit extends as scattered outcrops along the shores of Lake Cowan (Clarke et al. 1948) and also in the subsurface (Clarke, 1993). The Norseman Formation consists of glauconitic skeletal wackestone (silicified outcrops at the type section) and trough cross-bedded skeletal grainstone with varying proportions of calcareous fossils and quartz grains (Clarke, 1994a, b).

The marginal marine Norseman Formation, interbedded with the Werillup Formation, is an equivalent of the Paling Formation of the eastern Eucla Basin, developed in the western basin and deposited during the Tortachilla transgression (Fig. 4.1; McGowran 1989; Clarke 1993, 1994a, b; Clarke et al., 2003). In addition, local lenses of limestone such as occurring in bores at Neridup, near Esperance (Cockbain, 1967), and in the Nanarup quarry near Albany (the 'Nanarup Limestone' of Quilty 1969, 1981), were suggested to be equivalent in stratigraphic position with the Norseman Formation and the Paling Formation (Clarke et al., 2003).

The Norseman Limestone was deposited in the drowned estuary of the paleovalleys (e.g., Cowan Paleovalley), with dimensions of 200 km long and 100 km wide at the entrance, which is similar to that of the present-day Spencer Gulf in South Australia (300 km long and 150 km wide at the entrance), a close modern analogue (Clarke, 1994a, b; Clarke et al., 1996). The skeletal wackestones were deposited in low energy environments within the embayment, and skeletal grainstones were deposited in high energy environments such as submarine barriers. Water depths ranged from shallow, possibly in the lower shore-face, to moderately deep, with carbonate facies preserved as relict sediments on the floor of the drowned estuary during the peak of transgression. Locally, higher input of clastic sediment from tributary paleodrainage channels gave rise to the calcareous sand facies (Clarke, 1994a, b).

Pallinup Formation

Cockbain (1968b) defined spicular siltstone, sandstone and mudstone grading to spiculite of the Bremer Basin as the Pallinup Siltstone (the type locality at Beaufort Inlet). Gammon et al. (2000) redefined this unit as the Pallinup Formation. The same lithologies are laterally continuous in the subsurface between the coastal outcrops of southwest Western Australia and the former Princess Royal Spongolite (see below) in the Cowan Paleovalley. Scattered outcrops of these lithologies occur almost continuously across the watershed between the Cowan and Lefroy paleovalleys (Clarke, 1994a, b). These lithologies also occur semicontinuously around the paleoshoreline from Esperance to north of Balladonia (Doepel and Lowry, 1970a, b; Morgan and Peers, 1973) where they were also assigned to the Pallinup Siltstone. Similar siliceous sponge-rich limestone occurs near Balladonia, separated from equivalent Pallinup Formation by outcrop chains of crystalline basement. Interfingering of the Wilson Bluff Limestone and Pallinup Formation was found in a drillhole at North Rocks, northeast of Balladonia (Clarke et al., 2003). Therefore, the Pallinup Formation was extended to include all such lithologies along the western margin of the Eucla Basin throughout this area (Clarke et al., 2003), including all Late Eocene spicule-rich marine sediments, and the spicular sandy sediments formerly referred to as upper Hampton Sandstone by Clarke (1993).

The Pallinup Formation at the type locality (Gammon et al., 2000) and along the length of southwest Western Australia consists of four informal members fining-upwards: conglomeratic sand in the lowermost, sand in the second and spicular muds in the third and fourth (Clarke et al., 2003). Uppermost, a fifth member was formally defined by Gammon et al. (2000) as the Fitzgerald Member (see below), with its type locality in the Fitzgerald River National Park. Calcareous biota are normally rare but can be moderately common in some beds. Molluscs, with less abundant bryozoans, brachiopods and echinoids dominate the calcareous biota (Clarke et al., 2003). Cyclic bedding is common through much of the unit, as is glaucony. The trace-fossil assemblage is dominated by *Thalassinoides* and *Chondrites*. The unusual opal-dominated biota occurs on the landward side of paleotopographic barriers formed by bedrock highs that would have been rocky reefs and islands in the Eocene.

The Pallinup Formation is the direct equivalent of the Khasta Formation in South Australia (Fig. 4.1). These similar and equivalent lithologies were deposited in similar lagoonal environments. It is possible that minor occurrences of these units may be present in the subsurface of the lower reaches of the Carey, Baker, Throssel, Wanna and Serpentine Lakes paleovalleys. The Neales Embayment might provide an analogous Eocene lagoonal environment to the Immarna lagoon further east, and Pallinup lithologies may be common there (Clarke et al., 2003). The digital elevation model and NOAA ntt images show a low but prominent ridge in this area, paralleling the margin of the embayment and another less prominent ridge along its outer edge and separating it from the Nullarbor Plain (Figs 2.3 and 2.4). These might represent Eocene barriers analogous to the Ooldea and Barton ranges, further to the east.

Fitzgerald Member

The Fitzgerald Member, defined by Gammon et al. (2000) as the uppermost member of the Pallinup Formation (type section in the Fitzgerald River National Park), consists mainly of spiculite to muddy spiculite, in which body fossils of lithistid sponges are locally abundant, forming sponge conglomerate or rudstone in a spiculite matrix. The member occurs in former sheltered localities behind barriers of bedrock highs that formed rocky reefs and islands during the Eocene (Clarke et al., 2003). Calcareous biota can be common in the Fitzgerald Member, with large fenestrate bryozoans, epifaunal bivalves and echinoids in a spiculite matrix, as exposed at Bremer Bay, Thomas River east of Esperance, and near Balladonia. Foraminifers found in samples from Blue Dam north of Balladonia contain the semiendemic benthic rotaliid taxon *Maslinella chapmani*, suggesting a Middle to Late Eocene (P14–P15) age, with the possibility that it may be as young as P16–P17 (Clarke et al., 2003). This assemblage indicates a Tuketja age for the Pallinup Formation (Fig. 4.1). Unlike the Princess Royal Member (see below), where the biota is composed almost entirely of siliceous sponge spicules, the Fitzgerald Member has beds rich in sponge body fossils and calcareous biota (Clarke et al., 2003).

The Fitzgerald Member lithology of large carbonate bioclasts floating in a spicular matrix may represent a transitional facies between the sponge-rich Pallinup Formation and the Wilson Bluff Limestone, as interpreted by Clarke et al. (2003) who noted that the Wilson Bluff Limestone exposed at Point Culver and in drillholes near Balladonia contains common body fossils of siliceous sponges and may represent a further aspect of this transition. While interfingering, the transition from biosiliceous to calcareous lithologies is rapid (Jones, 1990), and probably occurs over a distance of a few hundreds of metres, explaining the scarcity of transitional facies.

Princess Royal Member

Units mapped as Princess Royal Spongolite (the original type locality at the Princess Royal townsite near Norseman; Glauert, 1926) occur in the Cowan (Cockbain, 1968a), Lefroy (Jones, 1990; Clarke, 1993) and Ponton Creek (Doepel and Lowry, 1970b) Paleovalleys. Because these units comprise only spiculite (spicular sands, silt and clay and lenses), rather than spongolites *sensu stricto* (sediments comprised largely of sponge body fossils), Clarke et al. (2003) suggested changing the status of the Princess Royal Spongolite from a separate formation to the Princess Royal Member of the Pallinup Formation (type section: 34 m thick in the drillhole CD1916, location 375800E, 533400 N, Zone 51, WIDGIEMOOLTHA 1:250,000 map sheet SH51-14; Fig. 4.3). Lithologically, thus, the Princess Royal Member consists of spicular sands, silt and clay and lenses, with siliceous sponges (commonly carbonaceous fragments followed by calcareous spicules) and phytoliths, but contains rare fossils.

The Princess Royal Member is lithologically equivalent to the Fitzgerald Member but can be differentiated from it by its restriction to paleovalleys, rather than along margins of the more open part of the basin (Clarke et al., 2003). In the Lefroy and Cowan Paleovalley, for example, the Princess Royal Member rests in part on benches eroded into weathered bedrock and terminated by increase in slopes (Clarke, 1993, 1994a, b). The Princess Royal Member was deposited during the Aldinga transgression Tuketja-Tuit Transgression (Fig. 4.1), and the bench occurring in the Lefroy Paleovalley at 270–280 m represents a ravinement surface cut during the transgression (Clarke et al., 2003).

Redmine Group (western nearshore-onshore)

The Redmine Group, unconformably overlying the Eocene succession and Precambrian basement in the western margin of the Eucla Basin, often with a ferruginous erosional relief of up to 20 m (de Broekert, 2002), comprises Neogene nearshore facies (e.g., Colville Sandstone) and onshore facies (e.g., Revenge, Cowan, Gamma Island, Roysalt, and Polar Bear formations; Fig. 4.1).

LATE OLIGOCENE – EARLY PLIOCENE SEQUENCES

Colville Sandstone

Colville Sandstone for lateral equivalents of the Nullarbor Limestone (Lorry, 1968a), named after Lake Colville 17 miles east of the type section (lat. 29°31'S, long. 126°23'E; about 177 km northwest of Loongana), is dominantly composed of nearshore calcareous sandstone that shows a variety of lithologies, such as yellow, fine-grained, micaceous, thin-bedded sandstone; fine- to coarse-grained sandy calcarenite; interbedded ferruginous claystone, siltstone, sandstone and granule conglomerate (Lorry, 1970). The Colville Sandstone, more than 20 m in thickness at the type section, thins towards the north-northwestern margin of the basin, and is ~2 m thick where it overlies Wilkinson Range Beds near Lake Gidgi (Lorry, 1970).

The Colville Sandstone, dated Early to Middle Miocene, ~18.5–14.5 Ma (Geoscience Australia, Australian Stratigraphic Units Database), partly overlain by and interfingers with Nullarbor Limestone. The Colville Sandstone is believed to grade laterally into the Nullarbor Limestone because of the structural continuity, gradation in lithology, and faunal similarity between the formations (Lorry, 1970). As the lateral equivalents of the Yarle Sandstone deposited along the northeastern margin of the Nullarbor Plain, the Colville Sandstone deposited in the northern and northwestern margin of the Nullarbor Plain passes laterally basinwards into the Nullarbor Limestone (Figs 4.1 and 4.3). The boundary between the Colville Sandstone and the Nullarbor Limestone is difficult to map, partly because of the buried feature and partly because of the interfingering nature of the boundary, but the distribution of sand is easily recognised both in the field and on air-photographs because the clay soil over the Nullarbor Limestone supports myall scrub whereas the sand or clayey sand over the Colville Sandstone supports mulga (Lorry, 1970).

The Colville Sandstone disconformably overlies the Wilkinson Range Beds around the northern edge of the basin (e.g., near Lake Gidgi, lat. 29°06'S, long. 126°03'E), and surrounds the inlier of Ilma Beds and appears to overlie Precambrian metamorphic rocks on the edge of the basin near Plumridge Lakes (Lorry, 1970). The Colville Sandstone is overlain by a variety of Quaternary deposits.

Revenge Formation

As the basal unit of the Redmine Group, the Revenge Formation (type locality exposed in the walls of the Revenge open pit, shown on the Lefroy map sheet at GR 765412, but in reality occurring 500 m further east at GR 770412), occurs widely in the western Eucla paleovalleys, such as beneath the Lakes Lefroy and Cowan (Clarke, 1993). In the lower reaches of the paleovalleys (e.g., Lefroy and Cowan), western Eucla Basin, there is up to 30 m of sediment equivalent to the Garford Formation in the eastern basin, where it overlies carbonaceous Eocene sediments equivalent to the Pidinga Formation (Fig. 4.1; Clarke et al., 2003). Similar to the Garford Formation of the eastern basin, the white clays in the Revenge Formation are in part dolomitic and palygorskite-bearing.

The Revenge Formation consists of a fine-grained 'central' facies that comprises massive to faintly bedded red-brown silty clays with varying amounts of siderite cement, and red sands locally cemented by iron oxides to form ferruginous sandstone and a coarse-grained 'marginal' facies that comprises iron oxide cemented conglomerate and sandstone (de Broekert, 2002), such as those predominate in the Cowan and Lefroy paleovalleys (Clarke, 1993, 1994a, b). The Revenge Formation reaches known thickness of 17 m in Lake Lefroy and 25 m in Lake Cowan (Clarke, 1993). Dark grey-green coloured sediments, similar to those of the Garford Formation of the eastern basin, are locally present.

The base of the Revenge Formation, commonly presented by pebble/gravel beds rests unconformably on Archean bedrock, as displayed in the Revenge open pit, or disconformably on Eocene sediments. The top of the sequence is the disconformable base of the overlying Roysalt Formation in Lake Lefroy and the Polar Bear Formation in Lake Cowan (Clarke, 1993). Although the Revenge Formation was unable to be dated (Clarke, 1993; de Broekert, 2002), its maximum and minimum ages set by the Princess Royal Spongolite (early Late Eocene) and Roysalt Formation (latest Miocene; see below), respectively leave a wide time range of Late Oligocene-Miocene during which the Revenge Formation could have been deposited (Clarke, 1993). Given the widespread preservation of lacustrine sediments from Miocene elsewhere in Australia (De Deckker, 1988), a Miocene age of the Revenge Formation is set by the underlying Late Eocene sediments, and the overlying Pliocene pollen assemblage in the Roysalt Formation (Parker, 1988), which also coincides with the last major transgression on the Eucla Basin that deposited the Nullarbor Limestone (Lowry, 1970).

The lithologically very similar Perkolilli Shale (described by Kern and Commander (1993) in the Kalgoorlie area) in the Roe Paleodrainage is probably correlative with the Revenge Formation (Clarke, 1993), and the Miocene fluvial to marginal marine Plumbridge Formation (Hocking, 1990), as well as the marine Yarle Sandstone (Benbow, 1990c) along the eastern margin of the Eucla Basin are likely to be distal equivalents. Some of the numerous Cenozoic formations defined by Glassford (1987) from Yeelirie may be partly equivalent to the Revenge Formation (Clarke, 1993).

Cowan Dolomite

The Cowan Dolomite (Fairbridge, 1953), that crops out along the shores of lakes Cowan and Brazier, consists of white to buff-coloured dolomite to dolomitic carbonate mudstone (Clarke, 1993). The maximum thickness of the Cowan Dolomite is unknown, but in SN 78R it is 2 m thick. The age of the Cowan Dolomite is unknown but is probably Miocene as carbonate facies are common in Miocene lacustrine sediments elsewhere in Australia (De Deckker, 1988), and may be partly equivalent to the Miocene–Pliocene Garford Formation in the paleovalleys of the eastern Eucla Basin (Benbow et al., 1995a).

The Cowan Dolomite and Gamma Island Dolomite (see below) have similarity of sedimentary facies (compositions range from dolomite to limestone), interpreted depositional environment and probable equivalence, which consist of two rock associations: cryptocrystalline dolomite mudstone (with local peloids present) and oolitic packstone towards the depocentres; and fenestral dolomitic carbonate mudstone, intraclastic rudstone breccias and oncolitic wackestone to grainstone along the margins of channels (Clarke, 1994a).

Gamma Island Dolomite

Locally overlying the Revenge Formation (e.g., at Lake Lefroy), but more usually occurring as lenses within it (Fig. 4.1), is the Gamma Island Formation that comprises up to 2.5 m of massive to oolitic, peloidal, fenestral, and oncolitic dolomite (de Broekert, 2002). The Gamma Island Dolomite in Lake Lefroy is equivalent to the Cowan Dolomite in Lake Cowan (Clarke, 1993). The Gamma Island Dolomite (named for the Lake Lefroy Gamma Island shown at GR 775539 on the Lefroy map sheet) occurs patchily beneath Lake Lefroy, and along its northern and western margins. The Gamma Island Dolomite occurs within the Revenge Formation in KD 3006, with a thickness varying from at least 2 m in the type locality, to 2.5 m in KD 3006, and overlies it along the western shores of Lake Lefroy (Clarke, 1993). The lithology of lenses of the Cowan Dolomite is demonstrated in drillhole KD 3006 as oolitic and fenestral dolomite and on the margins and estuaries of Newtown, Merougil and Muldolia creeks as white oncolitic, fenestral, peloidal, and massive dolomites to dolomitic carbonate mudstones (Clarke, 1993).

The stratigraphic relationships of the Gamma Island Formation suggest a Miocene age (Clarke, 1993). As for the Cowan Dolomite, a Miocene age is favoured for the lacustrine facies, given the widespread preservation of lacustrine sediments, especially carbonates, from this time elsewhere in Australia (De Deckker, 1988), and Miocene–Pliocene age in the paleodrainage channels along the margin of the Eucla Basin (Benbow et al., 1995a).

LATE PLIOCENE UNITS

Roysalt Formation

The Roysalt Formation (named after Roysalt Siding, GR 688271, near the old Lake Lefroy salt works) consists of flat-bedded gypsum facies and cross-bedded gypsarenite, with some beds containing up to 22% mixed carbonates (dolomite, calcite and siderite; Clarke, 1994b). On the margins of Lake Lefroy, the flat-bedded gypsum facies of the Roysalt Formation consist of sandy silts and clays, passing into bedded gypsum crystals in a silty, carbonaceous, and pyritic matrix towards the centre of the lake, with known thickness of 9 m (KD 3010; Clarke, 1993). The gypsarenite facies is composed of gypsum-cemented steeply cross-bedded gypsum sand (Clarke, 1994b).

The Roysalt Formation rests disconformably on the erosional top of the Revenge Formation (Fig. 4.1). It is demonstrated that the current lake floor is an erosional surface, as preservation of the sediments above the current lake floor in gypsum dune islands indicates a minimum thickness of 1 m for eroded material (Clarke, 1993).

In Lake Lefroy, the Roysalt Formation occurs right across the floor of Lake Lefroy and on its margins. Exposures of the formation in walls of the Revenge open pit, attains 1.5–2 m thickness (Clarke, 1993). Given a Pliocene age of palynomorphs from the base of the formation at Revenge (Parker, 1988), the deposition of the Roysalt Formation may have commenced in the Pliocene and continued into the Holocene (Clarke, 1993). The equivalent unit in Lake Cowan is Polar Bear Formation. Other time equivalents of the Roysalt Formation are Pliocene sediments at Lake Tay, west of Lake Dundas (Bint, 1981), the Darlot Formation at Yeellrie (Glassford, 1987), and of the Pliocene Narlaby Formation (Benbow et al., 1995a) in the eastern Eucla Basin. Along the margins of Lake Lefroy, the formation passes laterally into, and interfingers with fluvial and aeolian clastic sediments (Clarke, 1993).

Polar Bear Formation

Lithologically, most of the Polar Bear Formation (named after Polar Bear Peninsula that occurs on the Cowan topographic sheet at GR 870680 and extends out into Lake Cowan north of Norseman) appears to be similar to the flat-bedded gypsum facies of the Roysalt Formation (Clarke, 1994a). The Polar Bear Formation consists of sandy silts and clays along the edges of Lake Cowan, and bedded gypsum crystals in a silty matrix towards the centre of the lake (Clarke, 1993). In Lake Cowan, Polar Bear Formation occurs across the floor of the lake and on its margins and reaches 3 m thick in CNG drillhole ET 120R (type section; Clarke, 1993). The earliest Pliocene–Holocene Polar Bear Formation rests disconformably on the erosional top of the Miocene Revenge Formation, and its top is defined by the present land surface (Fig. 4.1; Clarke, 1993).

Quaternary cover

The present landscape is dominated by extensive tracts of Quaternary sand plains and colluvial regolith, largely superimposed on, and locally obscuring, the older Paleogene–Neogene landforms. These veneers of unconsolidated sediments of Quaternary age occur throughout the region of the Eucla Basin and peripheral paleovalleys. The sediments comprise colluvium, alluvium and playa lake deposits as well as variety of regolith.

In the northern onshore basin, aeolian sand of the Wintrena Formation is the dominant dune forming unit of the Great Victoria Desert (Lowry, 1970). The Wintrena Formation is comprised of very fine to medium, red brown to orange, brown, moderately sorted calcareous sand. During periods of low sea level, fine calcareous material was blown inland forming a series of coastal foredunes and sief dunes as well as being incorporated into the soil over much of the southeastern portion of the basin. Calcareous sediments belong to the Bridgewater Formation, mostly along the present coastline.

Sheet kunkar or calcrete is widely distributed across the surface of the Eucla Basin, and is typically 1–3 m thick, composed mostly of a mosaic of calcite crystals in the range 1 to 3 microns with a scattering of crystals up to 10 microns. Surficial calcrete is extensive across the Nullarbor Plain, Nyanga Plain, Mardabill Plain, Carlisle Piain and Hampton Tableland, and is developed close to the surface of many coastal sand dunes, (e.g., Hampton Range) and gypsiferous dunes associated with salt lakes (e.g., the Plumridge Lakes; Lowry, 1970).

In the north and east margin of the Bunda Plateau and the coastal dunes, and superimposed across the paleovalley region, is the southeastern extent of dunefields of the Great Victoria Desert of Quaternary age. The principal landform are siliceous longitudinal dunes, 10 to 30 m high and 1–15 km long (Pell et al., 1999).

4.2 CORRELATION OF STRATIGRAPHIC NOMENCLATURE

Previous work has established correlations of the sediments marginal to the Eucla Basin with transgressive-regressive sequences elsewhere in South Australia (Alley and Beecroft, 1993; Alley et al., 1999; Clarke and Hou, 2000). Marine-influenced intervals throughout the basin are correlated with the Middle – Late Eocene Wilson Bluff (Benbow et al., 1995b; Alley et al., 1999), Tortachilla and Tuketja transgressions, as identified elsewhere in southern Australian basins by McGowran (1989), and McGowran et al. (1992, 1997). Another recognised Eocene transgression, Tuit, is represented in the central part of the Eucla Basin and probably also in the margin of the basin. Dating of both marine and non-marine paleovalley fills is thus possible through these correlations with well-dated marine sequences. The revisions in nomenclature and succession in Figure 4.1 result in correlation of Figure 4.3. Such a correlation can be evaluated by applying facies and sequence stratigraphy (see below). Studies of the main sedimentary units identified throughout the basin have informed our current knowledge of the stratigraphic relationships, age ranges and environments of deposition, and allow ties to regional geological events. The age ranges at various times and the eustatic and regional depositional cycles are well correlated throughout the Cenozoic Eucla Basin.

In the western and central parts of the basin the basal unit, the Hampton Sandstone, comprises up to 30 m of shallow-marine sands. This is overlain by up to 300 m of carbonates of the Wilson Bluff, Abrakurrie and Nullarbor limestones (Lowry, 1970; Hocking, 1990; Jones, 1990; Clarke et al., 2003). In the southwestern and central parts of the basin, Wilson Bluff Limestone is overlain unconformably by the Late Oligocene – Early Miocene Abrakurrie Limestone (Lowry, 1970; James and Bone, 1994; Li et al., 1996a), and farther to the north and northeast, by Middle Miocene Nullarbor Limestone (Lowry, 1970). The facies distribution over time suggests a northeasterly migration of the Eucla Basin depocentre from Eocene to Miocene times (Fig. 4.3; Hou et al., 2008). The relatively thick and localised occurrence of Abrakurrie Limestone compared with the overlying thinner but more aerially extensive Nullarbor Limestone is consistent with local tectonic subsidence in the central part of the basin in the early Neogene. Further inland (?50–100 km), post-Abrakurrie erosion prior to and during initial deposition of the Nullarbor Limestone has apparently removed any record of laterally equivalent Late Oligocene – Early Miocene terrigenous facies on the Eucla margin and paleovalleys (Benbow et al., 1995a; Alley et al., 1999; Hou et al., 2003a).

A veneer of Paleogene clastic-dominated sediments onlaps craton margins and older sedimentary deposits along the Eucla Basin margin (Clarke, 1994a; Alley and Lindsay, 1995; Hou et al. 2003a). These are principally Middle Eocene sediments that in places unconformably overlie and onlap Cretaceous, Permian and Neoproterozoic–Cambrian sedimentary rocks as well as Precambrian crystalline basement. Locally, these intertongue with the basal Wilson Bluff Limestone and pass laterally into the sand equivalents of estuarine and paleovalley fills (Benbow et al, 1995a; Clarke et al., 2003) through tidal inlets along the Ooldea Barrier (Hou et al., 2003a; 2006b). Late Eocene sediments (Zone P16/17) are overlain and intertongue with the sponge-spicule facies of the Khasta Formation and Princess Royal Member (Clarke et al., 2003; Hou et al., 2003a, b, c, 2006b).

The Yarle Sandstone and Colville Sandstone correlate basinward with the Nullarbor Limestone, and landward with the youngest aeolian facies of the Ooldea Sand, and were deposited along the Middle Miocene shoreline. They consist mainly of sediment reworked from beach/barrier sands of the Eocene beach/barrier/dune system. In the southeast Eucla margin, marginal marine and estuarine channel sediments of the Garford (Kingoonya Member and Narlaby Formations were deposited during the Middle Miocene and Pliocene as a thin sequence of laminated clay, silt and sand. Locally, these are capped by a minor carbonate unit up to 5 m thick (Ilkina Formation of Pliocene age). Along the western Eucla margin, however, this marginal marine and estuarine sequence is absent, consistent with uplift on this side of the basin, which limited the extent of Miocene marine transgression and development of backshore lagoons (Hou et al., 2008).

4.3 FACIES AND SEQUENCE STRATIGRAPHIC INTERPRETATION

FACIES

In the central Eucla Basin, the basal sequence of shallow marine sandstone is overlain by the marine limestones of the Wilson Bluff Limestone, indicating a change from marine clastic facies to marine carbonate facies. In the Eucla paleovalleys, as in most cases of marine-influenced fluvial lithologies elsewhere around the world, the channel fill lithologies are highly variable and facies relationships are complex. Both vertically and laterally, two facies transitions are of particular importance within the paleovalley fills: the transition from fluvial gravel, sand and clay to tidal-estuarine sand and clay in the alluvial-estuary plain, and the transition from tidal-estuarine sand and clay to estuarine/tidal-flat sand in the coastal plain (Hou et al., 2003c; 2006a). Posamentier et al. (1988) has found that the fluvial to estuarine facies transition at the landward limit of the estuary is situated in the vicinity of the bayline, which represents the junction between the low gradient estuarine plain and the more steeply sloping alluvial plain. However, the tidal currents during low river discharge can extend landward of the bayline, resulted in tidal mud preserved in the channel sand; alternatively, high river discharge during lowstand extends seaward of the bayline resulting in fluvial sand intercalated with tidal sediments (Allen and Posamentier, 1993). The latter can be observed in vertical sections in places of the eastern Eucla paleovalleys (Hou et al., 2001a). These vertical and lateral facies changes within the Paleogene channels are due to the complexity of fluvial sedimentation and erosion, but in the younger Neogene channels, lacustrine facies are dominant.

SEQUENCE BOUNDARIES (SB)

Sequence stratigraphic models have been used to describe the idealised stratigraphic architecture of sediments deposited at a continental margin during a single ‘sea-level cycle’, which embraced the concept of widespread stream rejuvenation following a fall in eustatic sea-level (Jervey, 1988; Van Wagoner et al., 1988; Posamentier et al., 1988). A rapid fall in eustatic sea-level associated with unconformity (a type of sequence boundary), for instance, could cause streams to incise as they extend themselves across the newly exposed continental shelf and the hinterland tracts of their drainage basins to be rejuvenated, and so that ‘incised-valleys’ are formed leading from the lowstand shoreline up into the hinterland drainage basins (Van Wagoner et al., 1988). The sequence stratigraphic method can help in reconstructing the relative chronology of deposits in biostratigraphically poorly dated successions and finds particularly important application in the incised fluvial succession of the Eucla Basin (Fig. 4.5; Hou et al., 2001a; 2003c; 2006a).

A change in relative sea-level influences sediment deposition to produce recognisable discontinuities such as sequence boundaries and ravinement surfaces. Several important stratigraphic surfaces punctuate the Eucla Basin sediments and paleovalley fills: the third-order sequence boundary, the transgressive surface, the tidal ravinement surface, the wave ravinement surface, and the maximum flooding surface (Hou et al., 2003c; 2006a).

The basal sequence boundary is expressed as an unconformity at the basin and channel floors. The bounding unconformity located between the Paleogene and Neogene sequences indicates the termination of Paleogene sedimentation. The top sequence boundary overlain by the Quaternary sediments is difficult to identify exactly in places, because post-Neogene erosion has removed part of the Neogene lacustrine clayey facies. In paleovalleys, the stratigraphic expression of the basal sequence boundary depends on its position within the channel (Allen and Posamentier, 1993). In the thalweg it separates lowstand fluvial deposits from underlying basement or bedrocks in places, but it is directly overlain by transgressive sediments on the channel walls. As the interfluves in the estuarine area (e.g., Wilkinson-Anthony estuarine area of the eastern Eucla Basin, the mouths of Tallaringa, Garford, and Anthony paleovalleys) are progressively transgressed by the eroding shoreline, the sequence boundary is expressed as a wave ravinement surface with transgressive marine-influenced sediments unconformably overlying pre-Paleogene rocks (Fig. 4.5). The stratigraphic expression of the transgressive surface in the onshore basin is characterised by onlap of transgressive estuarine channel sediments onto the fluvial deposits, whereas on the paleovalley walls the transgressive surface merges with the sequence boundary. Tidal scour in the coastal plain forms an erosional tidal ravinement surface overlain by estuarine / tidal-flat sands. Locally, the tidal-inlet sands were eroded by waves associated with the passage of the transgressing shoreline to produce a wave ravinement surface (Hou et al., 2003c). For the maximum flooding surface, the expression is as a downlap surface where the highstand (clayey or sandy) lignites and/or (carbonaceous) clay prograde over transgressive sands. This surface is difficult to identify in the upper reach of the paleovalleys.

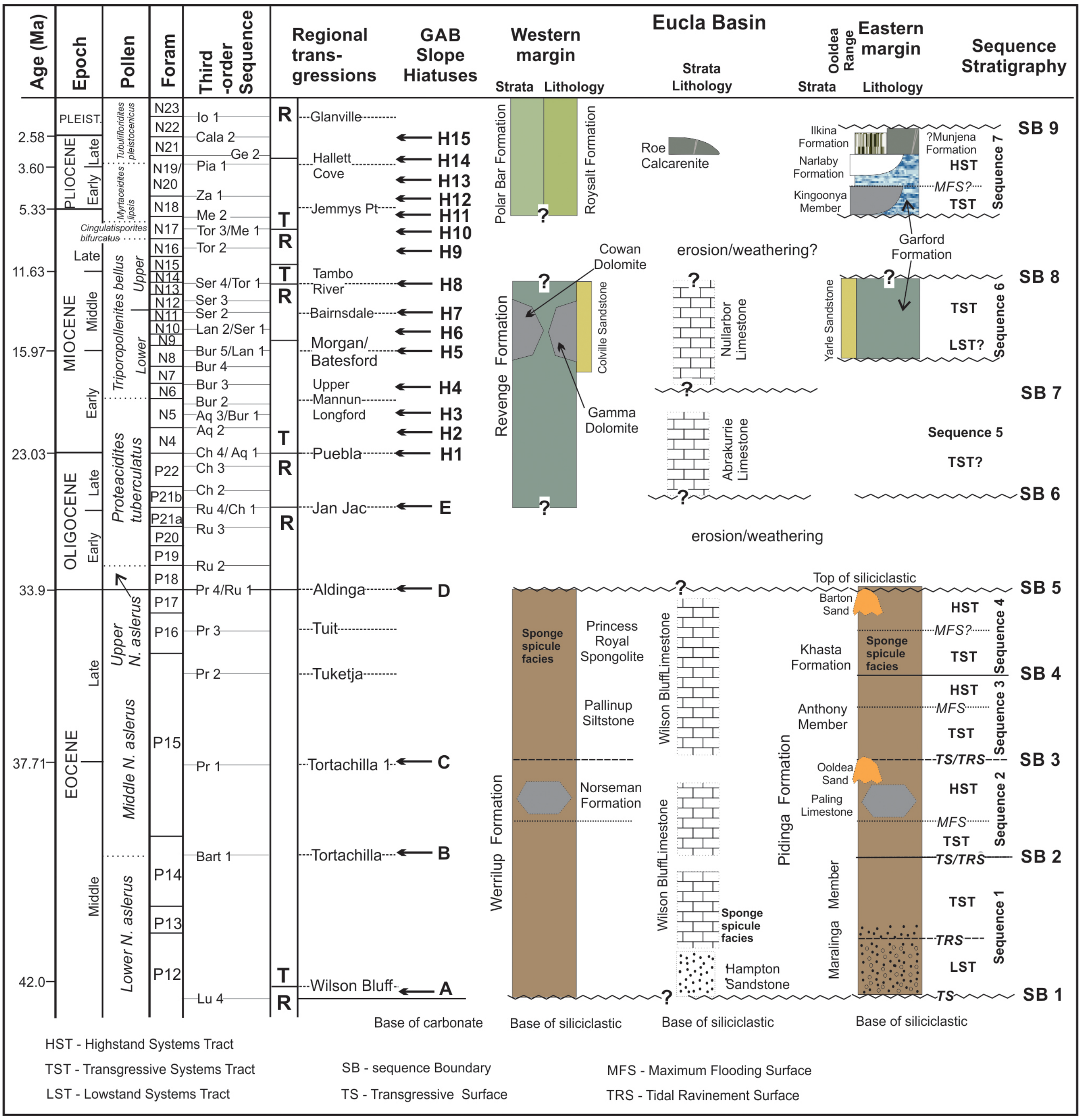


Figure 4.5 Lithostratigraphic and sequence-stratigraphic correlation of the Eucla Basin sediments (modified from Hou, 2008; Hou et al., 2003c, 2006a, 2008).

Sequences

Sequence 1 (S1): S1 erosively overlies weathered Pre-Paleogene–Neogene rocks and comprises shallow marine clastic with carbonates in the offshore basin, and fluvial gravelly sand and clay units as well as carbonaceous materials in the onshore basin (Fig. 4.5). Deposition of this sequence commenced in shallow basin and low sinuosity channels that occupied the sea and paleovalley floors, and the basal sediments of the sequence indicate deposition during a lowstand of relative sea or base levels. The widespread coal and related clay beds in the upper part of S1 indicate a regional rise of base level, possibly corresponding to a maximum flooding during the Wilson Bluff Transgression.

Sequence 2 (S2): In the offshore basin, the S1 is overlain conformably by transgressive marine carbonate facies (S2; Fig. 4.5). In the coastal plain, the top surface of S1 is overlain apparently conformably by transgressive estuarine channel facies (S2), the base of which rarely contains gravels. Inland, towards the paleovalleys, S2 lithologies are similarly developed progressively from transgressive to highstand deposition in the estuarine-alluvial plain during the Tortachilla Transgression; meanwhile, Ooldea Sand developed along the Ooldea (Tortachilla) coastal barrier. Marine waters penetrated the lagoonal area through barrier passes during a sea-level rise, which is stratigraphically correlated to the Paling/Norsman Formations and the marine Wilson Bluff Limestone throughout the Eucla Basin (Clarke and Hou, 2000).

Sequence 3 (S3): In the offshore basin, the S2 is overlain conformably by highstand carbonate facies (S3; Fig. 4.5). In the coastal plain, the top surface of S2 is overlain conformably by highstand estuarine sand facies of S3. In the estuarine-alluvial plain, the top surface of S2 is overlain apparently conformably by transgressive estuarine channel facies, formed during the Late Eocene Tuketja Transgression, and the base of S3 contains rare basal gravels. The retrogradational to progradational turnaround at the base of S3 is a composite boundary (Reynaud et al., 1999), as this surface appears less erosional than those at the bases of S1 and S2 (Posamentier and Vail, 1988). This explains why here there is no lowstand system tract at the base of this sequence. In this interpretation there is no need for a significant sea-level fall at this time and the paleovalleys are believed to have filled with sediment during a continuous overall relative sea-level rise.

Sequence 4 (S4): In the coastal-estuarine plain, the top of S3 (upper Pidinga and Werrilup formations) is in turn overlain by the estuarine sand of the Khasta and Princess Royal formations (S4) that probably formed during Tuketja-Tuit Transgressions (Fig. 4.5). This sequence is most recognisable in the coastal-estuarine plain and is commonly separated by a tidal/wave ravinement surface from underlying S3 and even pre-Paleogene bedrocks. An erosional break between the S4 and S3 provides evidence for a regression. Spicule facies of the Khasta Princess Royal formations is locally formed where the estuary was less influenced by tides. Like S2, in the eastern Eucla margin, the S4 was widely succeeded by the Barton Sand along the Barton (Tuketja) coastal barrier.

Sequence 5 (S5): In the offshore basin, the S4 is overlain unconformably by transgressive marine carbonate facies (S5; James and Bone, 1994; Li et al., 1996b), while in the eastern coastal plain, the S5 is absent with an erosion base towards the paleovalleys (Fig. 4.5; Hou et al., 2003c, 2008). In the western coastal plain, however, the S5 is probably represented by Revenge Formation towards the paleovalleys (e.g., Clarke, 1993; de Broekert, 2002). S5 lithologies are likely developed progressively during a small marine transgression (i.e., Upper Mannum-Longford Transgression) in the western Eucla Basin with a depositional centre located between Eucla and Lefroy Paleovalley mouth (Fig. 1.1); meanwhile, erosion occurred in the eastern Eucla margin. This erosional break between the S4 and S6 provides evidence for a non-deposition/ unconformity in the eastern basin.

Sequences 6 (S6): S6 comprises shallow marine carbonates of the Nullarbor Limestone unconformably overlying the Wilson Bluff Limestone and Abrakurrie Limestone locally in the central basin, and clayey fluvial and lacustrine units of the Garford Formation unconformably overlying the Eocene marginal marine and fluvial sediments in the eastern margin, as well as clayey fluvial and lacustrine units of the Revenge Formation and Cowan/Gamma Island Dolomites that unconformably overlies the Eocene marginal marine and fluvial sediments in the western margin (Fig. 4.5). Costal sandy sediments (i.e., Coville Sandstone and Yarle Sandstone) developed in the Eucla margins. In the onshore settings, the basal sandy clay unit includes interpreted transgressive lake or channel facies. The illitic clay and dolomitic clay units mainly accumulated in the lakes or channels during highstand of relative base level. S6, comprising dolomitic carbonate, is indicative of local deposition on the top of S5 during a lowstand of relative base level.

Sequences 7 (S7): S7 comprises four Pliocene genetically related shallow and marginal marine/estuarine and fluvial/lacustrine units of the Roe Calcarenite, Ilkina, Narlaby and Garford formations disconformably/unconformably overlying the Miocene sediments in the southern margin, and eastern margin while fluvial and lacustrine sediments of the Polar and Roysalt formations, unconformably overlying the Miocene sediments, developed in the western Eucla margin (Fig. 4.5). the sedimentation in the central basin was absent during Pliocene, and the top part of the Nullarbor Limestone (S6) of is characterised erosional surface.

Systems tracts

LOWSTAND SYSTEMS TRACT (LST)

When sea or base level was at its lowset, erosion was the dominant geologic process and lowstand systems tract deposits dominated deposition throughout the basal basin and channel thalweg. The sequence boundary in the basal basin and channel thalweg is overlain by a discontinuous veneer of basal gravel and gravelly coarse sand units of S1, locally with carbonaceous materials. The LST comprises the basal gravels and sands and their lithified and weathered equivalents deposited during the initial stages of the marine transgression, such as basal sediments of the Hampton Sandstone in the central basin (Wilson Bluff Transgression, Early–Middle Eocene), basal sediments of the Pidinga and Werrilup formations along the margins of the Eucla Basin (Tortachilla Transgression, late Middle Eocene; Fig. 4.5). This LST is representative of shallow marine deposition in the central basin and fluvial deposition on the floors of incised valleys. The channels adjacent to estuaries at this time probably played a more important role as a zone of sediment bypass than as a depositional setting, and these discontinuous lowstand fluvial sediments were preserved when they were buried by the onlapping TST (Hou et al., 2001a). Some were locally coeval with the TST. In sequence S6, however, the LST may occur in places but commonly lacking, or probably played a much less important role than TST. LST sediments is absent during other marine Transgression (e.g., Late Eocene and Neogene) probably due to the rapid sea-level rise (Fig. 4.5). The apparent absence of significant fluvial aggradation during lowstand in most parts along the estuarine channel may be explained by the time lag between coastal progradation and subsequent upstream fluvial aggradation (Allen and Posamentier, 1993). Locally, little or no lowstand fluvial aggradation would be expected in the thalwegs of the upper reaches of the channels unless lowstand conditions lasted a sufficiently long time.

TRANSGRESSIVE SYSTEMS TRACT (TST)

The transitions from low to high sea-level with the onset of the rapid Paleogene and Neogene eustatic rises are mostly characterised by transgressive system deposits, which resulted in accumulation of several facies associations – shallow marine, marginal marine, tidal-estuarine channel, fluvial and lacustrine deposits, with landward onlapping of estuarine plain over alluvial plain deposits. The bulk of lower S1 and lower S6 and other sequences in the nearshore and onshore settings is placed within the TST rather than within the LST because the shift from fluvial to estuarine sedimentation clearly indicates that marine transgression commenced within the channels (Allen and Posamentier, 1993). As the sea transgressed landward, the craton-derived sediments were trapped within the valleys. The elimination of fluvial sediment contribution to the littoral drift on the adjacent coast resulted in accelerated shoreline transgression. The fluvial facies association of TST within S1 and S2 is bounded by the transgressive surface and a distinct maximum flooding surface, while the tidal-estuarine facies association of TST within S1, S2 and S3 is bounded by the transgressive surface or tidal ravinement surface and a maximum flooding surface. The surface between the S4 and underlying S3 is represented by a tidal ravinement surface, which normally originated at the thalweg of the landward-migrating tidal inlet. This erosional surface can be regionally continuous in the distal part of the estuarine funnel fill and could be misinterpreted as another sequence-bounding unconformity within the channels (Fig. 4.5) (Hou et al., 2003c; 2006b). Wave erosion associated with retreat of the shoreface could result in removal of several metres of the Pidinga sediments (e.g., Demarest and Kraft; 1987).

HIGHSTAND SYSTEMS TRACT (HST)

When sea or base level was at its highest, highstand systems tract deposits were dominant. The HST deposits, developed from transgression to regression in the onshore setting, mainly consist of prograding overbank, flood plain and tidal flat sediments. The top of the HST is marked by an erosional surface, corresponding to the next eustatic fall. In the coastal plain, coeval with this filling of the estuary, the adjacent coastline was starved of sediments and eroded by waves in places, indicating that TST and HST locally can be synchronous (Allen and Posamentier, 1993). In the later phases of the HST, as the accommodation space reduced and approached zero, and the channel's grade reached the final equilibrium profile, vertical stacking graded upwards into lateral migration, resulting in the deposition of well interconnected sheet sands and/or clays in the latest HST (Aitken and Flint, 1995; Fig. 4.5).

The transition from the TST to the overlying HST is represented by a downlap surface, indicating the maximum flooding surface (MFS, Fig. 4.5; Posamentier and Vail, 1988), but lithologically this contact can be difficult to recognise (Allen and Posamentier, 1993), particularly in the upper reaches of the paleovalleys. In the channels, the MFS lies between transgressive phase transverse/point bars and regressive phase condensed intervals, or can be recognised by the coal-bearing facies. In the estuary and further seaward, the MFS may be at the downlap surface between the massive estuarine or marsh (sandy) clays or coal-bearing facies of the HST and underlying carbonaceous sands of the TST. It may even be absent owing to local scour by highstand tidal inlet channels (Hou et al., 2001a; 2003c).

The highstand systems tract of the Paleogene and Neogene transgressions consists of marine carbonate rocks in the main part of the basin and lenses of limestone and calcareous sandstone within fine-grained siliciclastic rocks found in former protected settings such as lagoons, archipelagos and estuaries (Clarke et al., 2003). The continuous outboard lithologies are separated from the more discontinuous inboard units by a gap and, in the northern and eastern Eucla Basin, by barrier sands. The gap is probably due to development of a ravinement surface in shallow water during the highstand (Clarke et al., 2003). The Wilson Bluff Limestone records the highstand systems tract in the central part of the Eucla Basin, whereas along the margins the highstand is represented by a number of units including the Norseman Formation, Nanarup Limestone Member and the Paling Formation (Fig. 4.5). The Ooldea Sand preserves the shoreline facies of this highstand. The Norseman Formation forms the highstand systems tract overlying the transgressive systems tract of the North Royal Formation (Clarke et al., 2003). The Norseman Formation onlaps onto an erosional surface at an elevation of ~260 m in the Cowan Paleovalley (Clarke, 1993). This represents a ravinement surface cut during the highstand.

4.4 PALEOCLIMATIC REGIMES

4.4.1 Introduction

Any attempt to understand the evolution of the Eucla Basin and peripheral paleovalleys will be enhanced if basic knowledge of the climate regime is available. Among the climatic elements of importance is that of rainfall, in terms of both annual and seasonal amounts, as the volumes of water transiting the drainage as runoff will have governed the extent and distribution of erosion and deposition. Basically, there are two ways to determine precipitation/humidity in the study area. Firstly, the work of palynologists in recognising plant types among fossil assemblages that serve as guides to the conditions. Secondly, some further clues to climate can be gained from numerical models of climate for certain time slices. The results are qualitative in the first method and quantitative, and provisional, in the second. Both approaches are used here.

The separation of Australia and Antarctica by sea-floor spreading has been the driving force for Cenozoic climate change since the latest Early Cretaceous (Macphail, 2007), due to the migration of Australia northwards into wetter and warmer middle-low latitude positions (Embleton, 1984; Veevers et al., 1991; Wilford and Brown, 1994). Paleolatitudes of the Eucla Basin from Paleocene through to the Miocene are estimated as follows: Early Paleocene (60 Ma) -56 to 60°S; Late Eocene (40 Ma) -53 to 57°S; Early Miocene (20 Ma) -37 to 41°S (Benbow et al., 1995b). This northward motion into lower latitudes meant that the continent entered progressively warmer climate zones, as against cooling on a global scale (Zachos et al., 1994). McGowran et al. (1997) suggested that the overall decline in temperature and rainfall during the Paleogene–Neogene was temporarily reversed during periods of high sea-level, and infer that the four phases of high sea-level (I–IV) in southern Australia represented intervals of high temperature and precipitation, each terminated by a rapid period of cooling or 'chill' (Fig. 4.1). On a continent-wide basis the balance of these opposed trends favoured slight warming (Frakes 1999). McGowran et al. (1997) considered that during individual warm phases, the Leeuwin Current brought warm, low salinity water from the north of Australia (South Equatorial Current) to the western and southern coasts of Australia and thereby created warm and wet conditions onshore, but during third-order and fourth-order (Milankovitch) cool times, the Subtropical Convergence Zone along the south coast of Australia moved north and shut the Leeuwin Current down, leading to cooler and drier onshore conditions. For the post-Paleocene interval, therefore, the strongest warming seems to have taken place in the middle Eocene; other warmings occurred in the early Eocene and the middle Oligocene. Prominent cooling took place in the earliest and the latest Eocene. It might be expected that rainfall, and the resulting runoff, would have increased during times of warming, owing to the increased capacity of warm air to accept (and precipitate) water, relative to cool air.

4.4.2 Paleoclimate indicators

At the time the Eucla Basin and peripheral paleovalleys first formed, and for a time thereafter, the climate of southern Australia was generally warm and wet (Alley and Beecroft, 1993; Benbow et al., 1995b; Alley et al., 1999; Frakes, 1999). Key events in the history of the Eucla Basin onshore margin were the apparently episodic duricrust-forming episodes in the Early and Late Eocene, the Middle Miocene and the Late Pliocene (McGowran, 1979). The fossil evidence from Eocene megaflora and microflora suggests at least local warm-temperate rainforest conditions (Alley et al., 1999; Zang and Stoian, 2006); this is supported by the widespread accumulation of lignites along the southern continental margin. Aridity in the region increased markedly beginning in the Late Miocene.

Physical, chemical and biochemical indicators of paleoclimate can be present in the non-marine sedimentary record, such as coal, evaporite, bauxite, laterite, and paleosols, as well as colours in channel sediments (Miall, 1996). They may yield limited information about paleoclimate, particularly rainfall (Frakes, 1979). Extensive deposition of carbonaceous sediments and lignite in the marginal Eucla Basin and related paleovalleys implies wet and probably warm conditions, but relatively low evaporation, for a lengthy period of Paleogene. Neogene dolomitic carbonates intercalated with gypsum and halite, deposited in extensive lakes and paleovalleys during the evaporation of waterbodies from sheet flooding during Late Miocene, are indicators of warm and arid climates. The carbonate cappings and paleosol dolomitic carbonates, probably formed during the Early Pliocene, suggest arid conditions. The development of silcrete may be correlated with times of major cooling in the Early Oligocene and the Late Miocene – Early Pliocene. This was preceded by ferruginisation which may mark peaks of warmth (Benbow, 1982). Associated weathering resulted in widespread kaolinisation at the sequence boundaries, indicating warm and humid climates where extensive leaching could take place (Folk, 1968). These thus systems suggest a change from humid toward arid environments since the Eocene.

There are known to have been relatively warm peaks immediately prior to several significant temperature drops near the boundaries between the Early–Middle Eocene, the Eocene–Oligocene (the terminal Eocene event), Middle–Late Miocene and Miocene–Pliocene (Fig. 4.1). The earliest cooling, and accompanying slight drying, began in the late Early Eocene after a considerable period of fluctuating temperatures. The large temperature drop at the Eocene/Oligocene boundary has been linked with ice cap formation on Antarctica (Shackleton and Kennett, 1975; Barrett et al., 1987). More frequent variation followed, with a major drop in temperature at the Middle–Late Miocene and Miocene–Pliocene boundaries. By the Late Miocene, aridity began to develop in conjunction with cooling and this culminated eventually in the formation of widespread aeolian sediments and calcrete during the Quaternary.

The paleoclimatic model of Kemp (1978) suggests that in the Paleocene and Eocene, westerly winds affected southern Australia (between 60–80°), while for the inland, circulation would have been weak and variable. Such conditions support the idea of coastal dune barriers in the northeastern Eucla Basin margin and their absence in the western margin of the basin; westerly onshore winds of at least moderate intensity are implied. The J-shaped coastal barrier (the Paling Range) which connected the Ooldea Range to the Barton Range implies a more northwesterly component in the prevailing winds (Benbow, 1990a; Clarke and Hou, 2000).

The spread of rainforests across southern Australia from the southeast to the southwest occurred in the Eocene, the best documented period of the Paleogene–Neogene (Kemp, 1978). The Middle–Late Eocene sedimentary and palynological records of the Eucla Basin and adjacent paleovalleys (Pitt et al., 1978; Hos, 1978; Alley, 1985; Benbow et al., 1995b, Alley et al., 1999; Zang and Stoian, 2006) attest to widespread rainforest and a humid climate for the region. High rainfall was postulated by Alley (1985) based on the presence of a significant percentage of pollen from coniferous taxa which have modern affinities in tropical areas and/or the high rainfall, temperate areas of Australia and New Zealand. Furthermore, he indicated that a landscape dominated by rainforest but with more open areas along streams or swamps is evident from the presence of pollen with affinities to grass (Gramineae), sedges (Cyperaceae) and reeds (*Sparganiaceaepollenites barungensis* and *Aglaoreidia qualumis*). The warm and humid conditions during the late Eocene were suitable for the growth of subtropical rainforests (Hos, 1978). Rainforest thus covered much of the continent at a time when it occupied higher southern latitudes. Much of the Eocene was a time of equable global climate (e.g., McGowran in Frakes et al., 1987) and when carbonate platforms and shelves formed along the southern margin of Australia.

4.4.3 Computer modelling of the paleoclimate

To refine understanding of the climates over Cretaceous–Cenozoic time period in the eastern Eucla Basin, an analysis of computer-generated general circulation models (GCM) of the paleoclimates made by Chris Poulsen of Pennsylvania State University was utilised by Hou et al. (2001a). The result covers the available modelling for the interval from the early Cretaceous to the early Miocene in 7 time-slices: 1) Early Cretaceous (130 Ma); 2) Early Cretaceous (116 Ma); 3) 'Middle' Cretaceous (100 Ma); 4) Late Cretaceous (80 Ma); 5) Early Paleocene (60 Ma); 6) Early Eocene (40 Ma); and 7) Early Miocene (20 Ma). The study examines the GCM results on a seasonal basis, that is, the predictions from the models are for each of the four seasons. The variables selected for display in coloured maps include mean temperature, precipitation, evaporation, precipitation minus evaporation, runoff, sea level pressure and wind stress.

The following points can be made about GCM simulations of Cretaceous to Miocene climates in the eastern Eucla Basin, which can be applied also in the western Eucla Basin as both eastern and western parts of the Eucla basin were located in the same latitude zones:

- **Temperature:** 1) Cretaceous climates were cool and included freezing in the spring, winter and autumn mainly in the early Cretaceous. Freezing took place in Early Paleocene time but only in winter and autumn. There is a slight indication of winter freezing in the early Eocene but none thereafter. 2) The 'Middle' Cretaceous to Miocene interval showed substantial warming.
- **Precipitation:** 1) Although variable, rainfall over the craton during the Cretaceous to Paleocene was generally less than 750 mm annually. 2) Spring and autumn rains characterised the region in the early Eocene and attained totals of about 1,000 mm annually. 3) Dry summers and autumns prevailed in the early Miocene, but annual rainfall increased markedly to more than 1200 mm.
- **Precipitation minus evaporation = moisture balance:** 1) Early to mid-Cretaceous rainfall exceeded evaporation by some 500 mm/year. 2) Late Cretaceous balance was more than 750 mm/year in favour of precipitation. 3) Paleocene–Eocene precipitation was greater than evaporation by only about 400 mm/year. 4) Early Miocene precipitation exceeded evaporation by only about 200 mm/year.
- **Runoff:** 1) Runoff was generally low and mostly isolated from the Gawler Craton area, except in the spring (Early Cretaceous through Paleocene). 2) Moderate winter and spring runoff

characterised the early Eocene. 3) Early Miocene runoff was very low and took place only in the spring.

In summary, climate modelling suggests that paleovalley incision most likely took place by means of moderate spring runoff in the early Eocene and was coincident with lowstands of sea level. Any earlier drainage networks were probably ephemeral and replaced by Eocene ones. Variations in Eocene runoff may have contributed to local, possibly even more extensive, erosion as well as subsequent deposition. The evidence of high rainfall and rainforest growth from palynological studies is indicative of much higher rainfall and runoff during the Eocene. Erosion of the channel fill may possibly have been caused by Oligocene runoff, but this interval has not been analysed by GCM. The Miocene was near the end of a long period of drying out in the region. Low rainfall led to significant runoff only in the spring, a likely reason for the development of lake systems on the craton

4.4.4 Implication

The preservation of a Cenozoic marine record within the continental Eucla Basin is a result of widespread aridification and the overall fall in eustatic sea level. Superimposed on these climate related variations are anomalous topographic effects due to mantle processes interacting with a northwards-drifting Australian continent. This interaction is argued to be a significant control on the extent of onshore flooding of the Eucla Basin during the Paleogene, and a factor in sediment preservation post-Miocene (DiCaprio et al., 2009; Heine et al., 2010).

The distribution of paleoshoreline features both informs and is informed by paleoclimatic interpretations (Hou et al., 2008). In the last 43 Ma, since the onset of fast spreading in the Southern Ocean, Australia has drifted ~3,000 km northward. The climatic effects of Australia's motion into lower latitudes and progressively warmer climate zones (Fig. 4.6) were countered to a significant extent by cooling on the global scale (Benbow et al., 1995b). On a continent-wide basis, the balance of these opposed trends favoured slight warming (Frakes, 1999). For the post-Paleocene interval, the strongest warming seems to have taken place in the Middle Eocene; other warmings occurred in the Early Eocene and the Middle Oligocene. The evidence from Eocene megaflora and microflora, as well as widespread accumulation of lignites along the southern continental margin, suggests that at least local warm temperate rainforest conditions prevailed during the Middle Eocene (Macphail et al., 1994; Quilty, 1994; Benbow et al., 1995b; Alley et al., 1999; Zang and Stoian, 2006). Paleoclimate modelling (Kemp, 1978) suggests westerly winds dominated southern Australia between 60–80°S in the Paleocene and Eocene consistent with the occurrence of coastal dune barriers in the north-eastern Eucla Basin margin but not along the western margin. Late Eocene J-shape coastal barriers behind the Ooldea Barrier imply a more northwesterly component in the prevailing winds. Large river systems up to tens of kilometres in width and 700 km in length, flowed through paleovalley systems that drained the exposed Precambrian basement terranes of the Yilgarn Craton in the west, the Gawler Craton in the east, together with the Albany-Fraser Orogen, Musgrave Province, and Officer Basin.

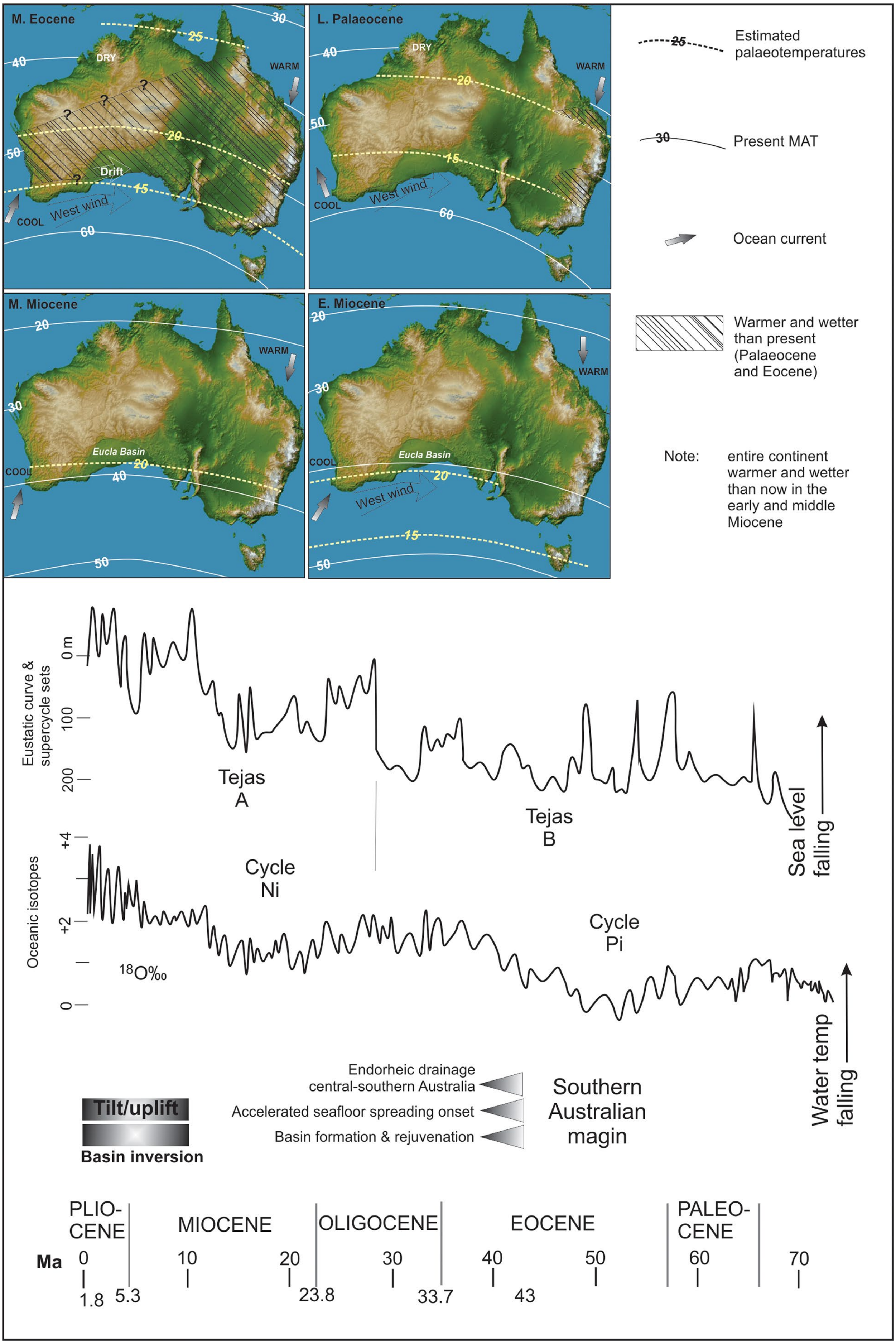


Figure 4.6 Cenozoic climatic interpretations: (a) Australia's migration on a time-latitude grid, with the continent shown in its Late Paleogene, Middle Eocene, Early Miocene and

Middle Miocene latitudes (modified from Frakes, 1999). The West Wind Drift is from Kemp (1978). **(b)** Cenozoic eustatic super cycle sets Tejas A and Tejas B (Haq et al., 1987, 1988) and water temperature curves approximating the Paleogene (Pi) and Neogene (Ni) oceanic oxygen-isotopic cycles (Abreu et al., 1998). Summary of tectonic events in southern Australia modified from Li et al. (2004). From Hou et al. (2008).

Prominent cooling took place in the earliest and the latest Eocene. Key events in the history of the Yilgarn and Gawler cratons and the Eucla Basin margin were the apparently episodic duricrust developments in the Early and Late Eocene, the Middle Miocene and the Late Pliocene (McGowran, 1979). Variable and wetter conditions, which were coeval with a major cooling of oceanic waters (Frakes, 1997), occurred at the Early/Middle Eocene boundary This may therefore have temporarily resulted in a substantial decrease in evapo-transpiration and an increase in discharge, which would be expected to trigger the deep and widespread fluvial incision evident in the 'inset valleys' of the paleodrainage systems (de Broekert and Sandiford, 2005).

Aridity in the region increased markedly beginning in the Late Miocene. Indicators of warm and arid climates include dolomitic carbonates intercalated with gypsum, halite and sheet flood deposits in extensive paleo-lakes and-channels. Paleosol carbonates probably started forming during the early Pliocene, and mark increasing aridity in conjunction with cooling, culminating in widespread formation of aeolian deposits and calcrete in the Quaternary.

4.5 TECTONO-EUSTATIC LANDSCAPE EVOLUTION MODEL

Relative uplift of the southern margin of the Australian continent since Early Miocene time is the combined effect of southern Australia moving away from a mantle low, aligned with the Australia-Antarctica Discordance, and interacting with the South-East Asian subduction system resulting in an overall tilt of the continent to the northeast (Sandiford, 2007; Sandiford et al., 2009). The continental tilting of the Australian landform has resulted in a vertical motion of elevations of Cenozoic shorelines since the mid Miocene, indicating some 250–300 m north down and south-southwest up, and yielded implications to the development of the Eucla Basin (e.g., Sandiford 2007). In the Eucla Basin, the eustatic sea-level variation during highstands inundated the craton margins, flooded paleovalleys to up to 400 km inboard of the present coastline.

A difficulty in correlation of the paleoshorelines between the western and eastern margin of the basin indicated that interpreted syn-depositional sediments had been influenced by post-depositional deformation of basin sediments. The present elevations of the Eocene marginal marine sediments on the western and eastern margins of the basin indicates that significant post-depositional deformation of basin sediments has occurred. By using the displacement of same facies sediments deposited during the same transgression on both western and eastern margins of the Eucla Basin (~138 m), Ruperto et al. (2006) studied the marginal marine (lignitic) carbonates in the western margin (Norseman Limestone currently sits at 274 m asl in Lake Cowan) and eastern margin (Paling Limestone currently sits at 136 m asl in Immarna Lagoon, drillhole CRAE RCH4) of the Eucla Basin, both were deposited during the Middle Eocene Tortachilla transgression (Fig. 4.1) and suggest that the western portion of the basin was uplifted by approximately 138 m, relative to the eastern portion of the basin.

McGowran (1989) correlated the late Middle Eocene (Tortachilla) and early Late Eocene (Tuketja) transgressions across the southern Australian Cenozoic basins with the third-order eustatic cycles TA3.6 (40.5-39.5 Ma) and TA4.I (39.5-38 Ma) of Haq et al. (1987) who indicated that these reached elevations of 140–150 m above present sea-level. The outcrops of the Late Eocene Princess Royal Spongolite at Cundeelee (Bunting and van de Graaff, 1977) and Mulga Rock are now at 325 m AHD in the western margin of the Eucla Basin (e.g., the region of Lake Lefroy and Lake Cowan), but the sediments are inferred to form during the Tuketja Transgression (Clarke, 1993, 1994a; De Broekert, 2002). Thus, Clarke (1994a) considered that an additional 110–150 m of post-Eocene uplift was required to bring the Eocene sediments at Lake Lefroy and Lake Cowan to their current elevations (up to 325 m AHD). The ravinement surfaces and the spicular facies beneath these surfaces occurring at same times throughout both western and eastern margins of the Eucla Basin (e.g., Pallinup and Khasta formations) show that they are now at different elevations, i.e., 260–280 m in the western margin of the basin but ~140 m in the eastern margin of the basin (Clarke, 1993, 1994a; Benbow, 1990a; Ruperto et al.,2006; Hou et al., 2008), of which, coincidentally or not, the latter is very close to the relative sea-level estimate made by Haq et al. (1987) in their 'global' sea-level curve. It is suggested that these features are genetically related to the scarps formed during the same marine transgression, i.e., Tuketja-Tuit highstand. Assuming minimal post-Eocene tilting of the southwestern Australia along an east-west axis (Sandiford, 2007) and hence preservation of elevation differences between the western and eastern margins of the Eucla Basin, the Late Eocene Tuketja highstand is of equal extent (Hou et al., 2008). Gentle uplift of the Yilgarn Craton occurred after the Miocene, to judge from the ~150 m differential in elevation of the Tuketja shoreline between the eastern and western Eucla Basin (Clarke et al., 2003).

The resulting SRTM DEM was flooded to 136 m (Tortachilla highstand; Fig. 4.5) in Figure 4.7. Hou et al. (2008) used marginal marine spicule profiles of Late Eocene to further demonstrate that uplift has occurred in the western margin of the Eucla Basin since this spicule facies was deposited. The tilt model allows for a good fit between the extent of Middle Eocene flooding and distribution of Middle Eocene coastal features on the western and eastern margin of the Eucla however, there is a poor fit in the central part of the Basin where Miocene limestone crops out at elevations greater than 136 m. A similar good fit between the extent of Late Eocene flooding during the Tuketja Transgression (based on an elevation on 200 m for the Tuketja highstand in South Australia on the eastern margin of the Basin: Hou et al, 2001a) and the distribution of Late Eocene coastal features (Fig. 4.8; Ruperto et al., 2006). The Ooldea barrier is flooded to form an offshore barrier system and a lagoonal environment (Immarna Lagoon) has been recreated behind the barrier.

When the SRTM DEM adjusted for terminal Miocene tilting is flooded to 115 m, the elevation of the Late Miocene highstand on the eastern margin of the Eucla Basin in South Australia (Fig. 4.1; Hou, 2008), the Lake Cowan area had been uplifted but the areas south of Jacinth were still flooded by marine transgression, which resulted in deposition of Kingoonya Member of the Garford Formation. The anomalous high elevations of Miocene limestone in the central part of the basin are even more apparent (Fig. 4.9). These anomalous elevations may be a result of the second mode of deformation described in the Sandiford et al. (2009); long wavelength (100+ km) compressional folding. Regionally, the Nullarbor Plain of the Eucla Basin has traditionally been considered a tectonically quiescent domain (Sandiford, 2007; Hillis et al., 2008). Exhumation of a large portion of the Eucla Basin across southern Australia, forming the low relief, limestone-dominated Nullarbor Plain (Fig. 4.10 from Hou et al., 2011b) is regarded as evidence of this tilting, which shows an ~150 m elevation differential of Neogene shorelines from the west to the east of the Eucla Basin (Sandiford 2007; Quigley et al. 2010). As the inferred tilt axis is to the north of the basin and at an angle to the present coastline (Sandiford, 2007) this offers an explanation for differential uplift across the basin resulting in Eocene sediments on the western side being at ~130 m higher elevations than stratigraphically-equivalent units on the eastern side, and the alignment of regressive beach strandlines of Late Miocene – Early Pliocene age being discordant with the dominant Eocene beach ridges along the eastern basin margin (Hou et al., 2008).

A systematic eastward migration of the depocentre across the Eucla Basin during the Neogene, together with apparent flow reversals in a number of paleovalley systems draining the Gawler Craton, suggest that the Eucla Basin has also been subject to differential vertical movements, expressed as a west-side up, east-side down tilting of ~100–200 m (Hou et al., 2008). This differential movement forms part of a broader north-down – southwest-up dynamic topographic tilting of the Australian continent associated with relatively fast (6–7 cm/yr) northward plate motion since fast spreading commenced in the Southern Ocean at ~43 Ma (Sandiford, 2007; Sandiford et al., 2009). Therefore, the region of the Eucla Basin provides a model that illustrates possible sedimentary responses to gentle tectonic uplift, where the uplift is marked mainly by a hiatus followed by an extremely thin sedimentary sequence (James et al., 2006; Hou et al., 2008).

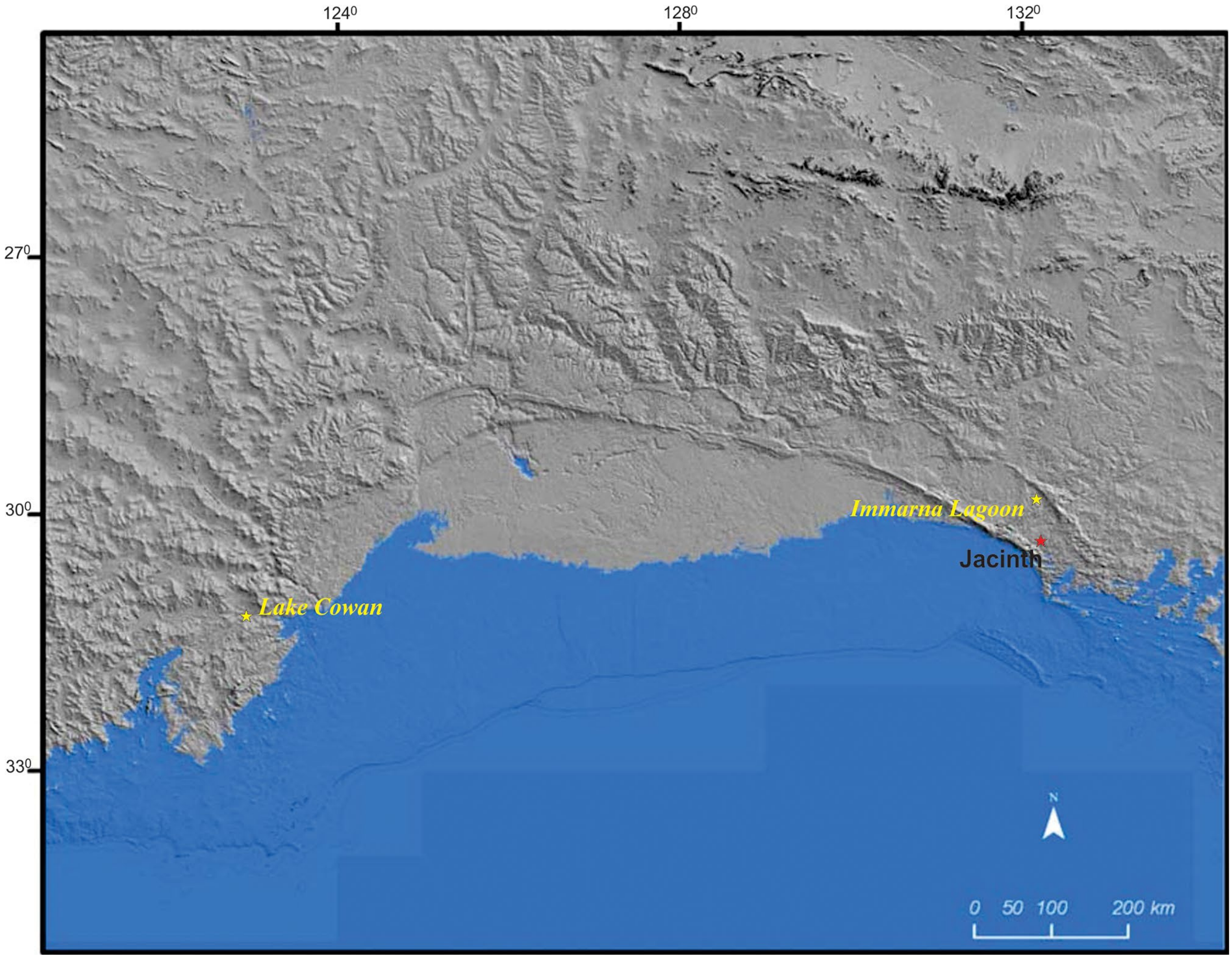


Figure 4.7 SRTM DEM adjusted for terminal late Neogene tilting and flooded to 136 m (Tortachilla transgression; Fig. 4.1), showing the locations of Lake Cowan, Immarna Lagoon and the Jacinth HM Deposit (modified from Ruperto et al., 2006).

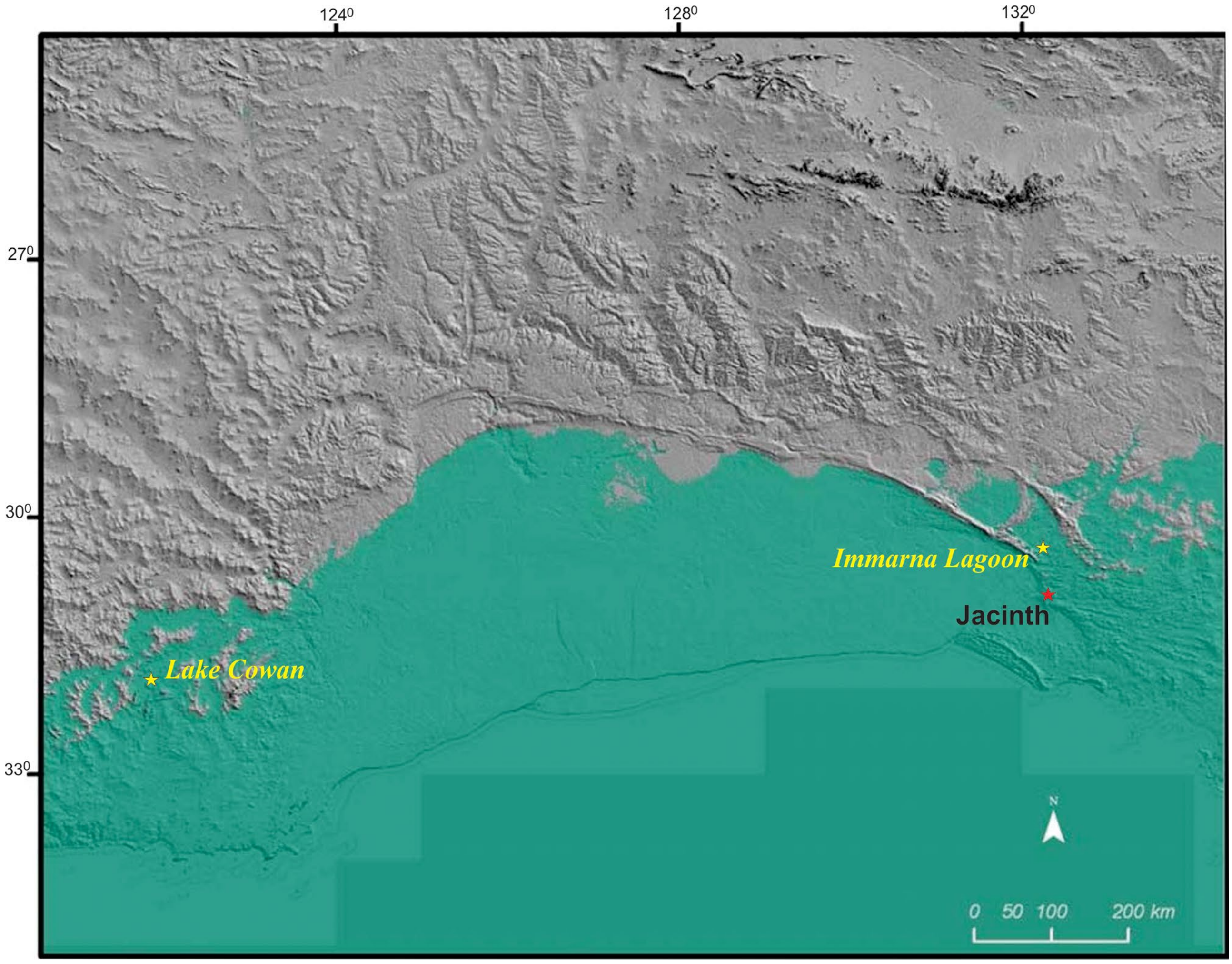


Figure 4.8 SRTM DEM adjusted for terminal late Neogene tilting and flooded to 200 m (Tuketja Transgression; Fig. 4.1), showing the locations of Lake Cowan, Immarna Lagoon and the Jacinth HM Deposit (modified from Ruperto et al., 2006).

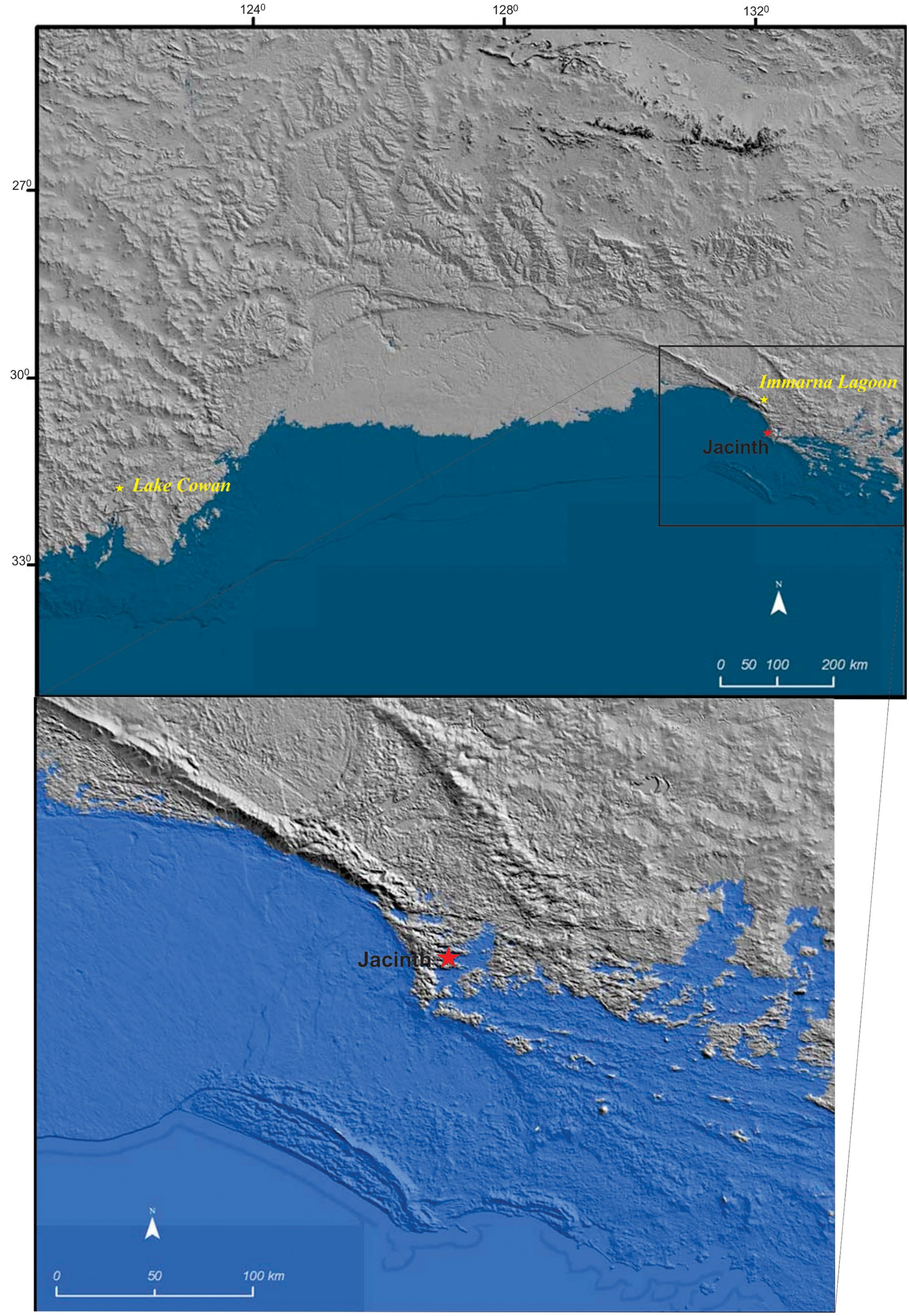


Figure 4.9 SRTM DEM adjusted for terminal late Neogene tilting and flooded to 115 m (Jemmys Pt transgression; Fig. 4.1), showing the locations of Lake Cowan, Immarna Lagoon and the Jacinth HM Deposit (modified from Ruperto et al., 2006).

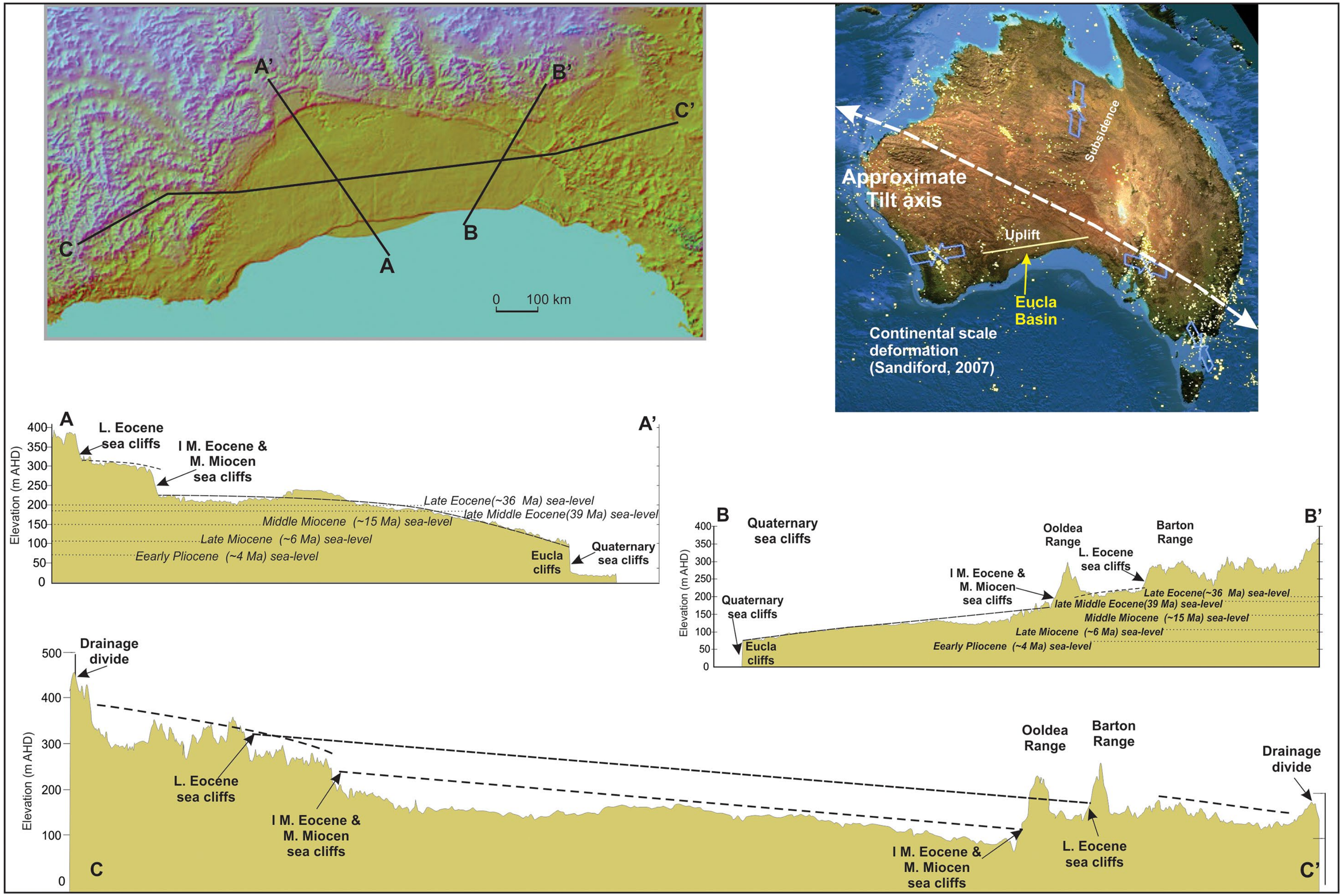


Figure 4.10 **The topographic difference between the western and eastern margins of the Eucla Basin, showing the evidence of dynamic movement, which resulted in down tilting to the east, and the interpreted tilt axis across the continent (modified from Sandiford, 2007; Hou et al., 2008).**

4.6 EVOLUTION OF THE BASIN AND PALEOVALLEYS

4.6.1 Environments of deposition

The southern continental margin of Australia, including the Eucla Basin, has been the site of cool-water carbonate deposition throughout the Cenozoic (McGowran et al. 1997), where the inboard and shallow-water carbonates of the Eucla Group have shown variably subtropical characteristics at times during the Cenozoic (Fig. 4.1; Lowry 1970; James and Bone 1989; Boreen and James 1993; Alley and Lindsay 1995; Lukasik et al. 2000).

The sparsity of boreholes in the Eucla Basin and peripheral paleovalleys makes it difficult to fully assess their environments of deposition across such a huge area, including broad and shallow paleovalleys, developed on the ancient Eucla Basin landsurface. However, schematic cross-sections of the Eucla Basin and peripheral paleovalleys show interpretable depositional environments developed on the pre-Cenozoic landsurface, now buried beneath Paleogene–Neogene sediments (Figs 4.2 and 4.3).

UNDIFFERENTIATED PALEOCENE–EARLY EOCENE SEQUENCE

Although the Paleocene–Early Eocene marine transgressions along the southern margin of Australia are not represented in the stratigraphic record of the onshore Eucla Basin (Lowry, 1970; Hocking, 1990), sedimentation of the Eucla paleovalleys may have commenced in the Paleocene-Early Eocene with undifferentiated fluvial to marginal marine clastics of the Immarna and Eundynie groups along part of the southern basin (Taylor, 1975; Barten, 1975; Davies et al., 1989). In the far west, Paleocene–Early Eocene sediments probably form a prograding wedge (Bein and Taylor, 1981), while in the eastern basin (west of the Ooldea Range) the fluvial-estuarine sediments were deposited in the Early Eocene paleovalleys (Hou et al., 2006a, 2008; Zang and Stoian, 2006). Hence, the shoreline remained on or below the present continental shelf during the Paleocene–Early Eocene when sea-levels were relatively high, probably because of relatively low early Paleogene sea-levels and low subsiding rate of the Eucla Basin. These low early Paleogene sea-levels for the region also feature in paleogeographic reconstructions of the Australian continent (Deighton et al., 1976; Veevers et al., 1991; Quilty, 1994) and of southwestern Australia (Johnstone et al., 1973).

MIDDLE–LATE EOCENE SEQUENCE

The shallow marine occurrence of the Hampton Sandstone with highly variable thickness, mainly occurring in some places across the central basin (e.g., in the Naretha and Madura areas) and its apparent absence to the east of Madura suggest an irregular base of the Eucla Basin and an initial depocentre located in the southwestern part of the Eucla Basin. To the west, the Hampton Sandstone fills the base of incised-valleys near Lake Harris (Jones, 1990), which provides supporting evidence that the incised-valleys extend from the Yilgarn Craton beneath the Wilson Bluff Limestone cover of the Eucla Basin. The overlying Wilson Bluff Limestone, as a temperate water carbonate, was deposited on a very broad and gently inclined platform of the Eucla Basin, characterised by dipping gently basinwards and by having higher energy facies predominantly at or near the Middle Eocene shoreline (Benbow et al., 1995b). The platform over which deposition took place, was around 1,000 km wide across a line marked by the present coastline, and distance between the maximum Late Eocene shoreline and platform margin was around 500 km. The evidence of the Wilson Bluff Limestone being a temperate water carbonate is supported by the existence of the skeletal composition that best fits the foramol assemblage of Lees and Buller (1972) which today occurs in temperate and polar waters. The skeletal assemblage lacks coral and calcareous green algae (e.g., Halimeda) which Lees and Buller (1972) found typically occur in warm water sediments. McGowran and Lindsay (1969) examined the Wilson Bluff Limestone in Outback Oil - Pickrell Mallabie Water well and found that the record of poor and rare *Truncorotaloides spinulosus* is an important indication of warm-water influence.

The Wilson Bluff Limestone was deposited during at least three transgressive cycles that have been recognised by McGowran (1989) in southern Australia during the Middle–Late Eocene (Fig. 4.1). Deposition occurred in a subtidal environment, mostly below wave base as indicated by its generally muddy (micritic) nature and by its poor sorting. The cleaner and coarser-grained beds, with evidence for local scouring and both upward coarsening and fining, as occurs at Head of the Bight, indicate periodic storm generated currents and deposition above storm wave base. Higher energy conditions are indicated by the existence of the current bedded and better sorted skeletal limestones, as occurs near Koonalda in a more distal setting, which were probably deposited during lowstands, and represent shoaling and prograding deposition. The middle part of the formation in the basin mainly presents transgressive deposition. Upward shallowing is suggested in both western and eastern parts of the basin where the formation has the higher energy limestone (former Toolina Limestone; Lowry, 1970) and a higher terrigenous component. This upward shallowing associated with highstands is indicated where the formation is overlain by barrier facies in the eastern margin of the basin, such as Ooldea and Barton sands. Deposition of bryozoan limestone with a little terriginous content, in nearshore settings, indicates at least local minor input of terrigenous sediment, such as the Paling Limestone (a marginal equivalent of the Wilson Bluff Limestone) north of Ooldea (e.g., CRAE RCH-2) and the Norseman Limestone in the west side of the basin (e.g., Lefroy Paleovalley).

The Pidinga and Werrilup formations were deposited in a range of environments, from alluvial/fluvial (e.g., channels), through floodplain (overbank, swamp), estuarine to marginal marine under terrestrial conditions characterised by plant associations growing in high rainfall tropical/subtropical environments (Alley, 1985; Rankin et al., 1996). Specific environments can be expected to be very varied, particularly for estuarine, deltaic and shoreface facies (Hou et al., 2006b, 2008). The fluvial medium to very coarse sands and grits, on upward fining cycles and capping and interbedded carbonaceous muds, were mainly deposited in paleovalleys (e.g., Lefroy, Tallaringa and Garford). The sands represent point bar and perhaps channel bar deposits, whilst comparatively rare grits or conglomeratic facies indicate base of stream, channel lag deposits. Carbonaceous sands indicate likely reworking of adjacent flood plain environments. The fossil wood and leaf bearing carbonaceous silts and muds were deposited in overbanks, flood plain swamp and lagoonal environments. Laminated and cross laminated carbonaceous muds and very fine to fine sands (e.g., drillhole TFS-4 in the Garford Paleovalley) were deposited in channels. Energy level is indicated by relative proportion of the matrix and various framework constituents, and the nature of the lamination. Stream flood plain environments can be interpreted for the similar sediments deposited in the nearshore basin proper, particularly in the estuarine channels.

The Khasta Formation is interpreted to have been deposited in the full range and complexity of estuarine to marginal marine environments, which are similar to those (e.g., The Princess Royal and Fitzgerald members) described by Clarke (1993, 1994a) in the Lefroy and Cowan paleodrainage channels of Western Australia. Relatively low energy and restricted marine conditions must have been present locally in the estuary and adjacent paleorivers to account for the fine grain size, good sorting and the presence of marine fossils and spicular facies with less than 25% siliceous spicules. For spongolite deposition, a nearshore environment within a coastal embayment during a highstand of sea level was suggested by Cavaroc and Frem (1968), whereas Lane (1981) preferred a shallow marine environment of less than 27 m depth. The spicular facies of the Khasta Formation are interpreted as having been deposited in the numerous relatively low energy, tidal-influenced channels, estuaries and coastal embayments during highstands (Clarke and Hou, 2000; Clarke et al., 2003), of which the laminated and cross-bedded carbonaceous fine-grained sediments (as near Lake Bring) are tidal-influenced estuarine plain/swamp deposits, indicating repetition of changing current conditions and reactivation surfaces. Deeper water lagoonal or estuarine environments are also likely to be represented by part of the Khasta Formation, suggesting back-barrier overwash flat deposits, such as the dinoflagellate non-spicule bearing laminated carbonaceous sediments that were deposited in such environments, and which occur elsewhere, such as beneath Barton Range and the Ooldea Range.

The Ooldea Sands in the Ooldea barrier (late Middle Eocene) and the Barton Sands in the Paling and Barton barriers (Late Eocene) were deposited in energetic upper shoreface, beach and coastal dune environments during development of the Eocene coastal barriers (Benbow, 1990a; Clarke and Hou, 2000; Hou et al., 2006b; 2008). Under dominant offshore winds, the sands were redistributed by waves, longshore and tidal currents, and wind, and gave rise to seaward prograding sandy coastal belts (Einsele, 1992). The interplay of aeolian processes, river flow, tidal oscillations, wave action, and variations in relative sea level provided energy for the ongoing dynamism in the coastal areas. The large quantities of sand required for barrier construction were supplied to the coast by the paleorivers draining the Yilgarn and Gawler Cratons as well as Officer Basin. The growth of coastal dunes in the estuarine plains may have begun as chains of islands, which amalgamated to form large coast-parallel dunes during progradation.

LATE OLIGOCENE – EARLY MIOCENE SEQUENCE

Lowry (1970) concluded that the depositional environment of the Abrakurrie Limestone was a shallow water open shelf. Normal salinity is indicated by the diverse fauna including such stenohaline forms as echinoids. Strong bottom currents are indicated by the degree of sorting and crossbedding. The shelf may have been both tide and wave influenced, with no terrigenous input (James et al., 2006). A detailed study of the Abrakurrie Limestone by Li et al (1996a) demonstrates that the infauna-rich assemblages from the middle and lower members of the Abrakurrie Limestone indicate relatively cooler, mesotrophic conditions, while a well-lit, warmer, oligotrophic environment is inferred for the upper Abrakurrie Limestone, indicating that the Eucla Basin was a swell-dominated, embayed shelf during the deposition of the Abrakurrie Limestone.

MIOCENE–PLIOCENE SEQUENCE

The sediments the Garford and Revenge formations were deposited in fluvial channel, flood-plain, lacustrine, lagoonal swamp and locally evaporative environments, with the Ooldea Range and scarps acting as barriers separating lacustrine deposition from marine deposition (Nullarbor Limestone; Benbow et al., 1995b). The sediments deposited in lacustrine settings without direct connection to the sea unfortunately are correlated with difficulty to marine eustatic changes. The fine-grained facies of these formations are interpreted as a shallow lake deposit with the sandy lenses possibly representing sub-aquatic distributary channel fills, whereas coarse-grained facies are probably the result of deposition in fluvial and fan systems feeding into the lakes (Clarke, 1994a, b). The local presence of dark grey-green sediment may indicate deposition in slightly deeper, and possibly stratified, lake environments. Cowan Dolomite and Gamma Island Dolomite that are intercalated in the Revenge Formation in the western Eucla Paleovalley, are similar to the dolomitic facies of the Garford Formation of the eastern Eucla margin, deposited in local evaporative environments such as lacustrine and lagoonal swamp. The dominance of extremely fine clastic and chemical sediments of the Garford and Revenge Formations suggests that, although water was at times plentiful, the flow was not sufficiently vigorous to flush clay out of the river system, and water was often ponded for extended periods. Subsequent drying of these lakes or channels led to deposition of regressive dolomitic clastic and chemical sediments.

Uplift and southeastward tilting of the Eucla Basin sediments and regression followed during the Middle Miocene – Early Pliocene. Deposition of a carbonaceous-dominant phase (i.e., the Kingoonya Member; Hou et al., 2003a) of the Garford Formation and Narlaby Formation took place in the paleovalleys around the southeastern basin, such as in the Tarcoola region and the northern Eyre Peninsula.

The Ilkina Formation, predominantly a saline lacustrine sequence deposited in the southeastern Eucla margin and adjacent paleovalleys, was deposited in arid playa lakes and streams after and during marked deterioration of climate in the latter part of the Cenozoic (Rankin et al., 1996). At the Type Section of the Ilkina Formation, the upward fining beds and the interlaminae of clay, silt and sand, indicate oscillating lake levels. Carbonaceous clays accumulated locally, where and when uninterrupted by oscillating lake level. 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, indicating drying of the lakes and the concentration of saline brines.

The Roysalt and Polar Bear formations of the western Eucla margin are interpreted as the deposits of an evaporitic environment not dissimilar to those in the present-day lakes Cowan and Lefroy, with palynomorphs indicating the presence of arid vegetation in the area of Lake Lefroy since the latest Miocene (Clarke, 1994a, b).

4.6.2 Geological history

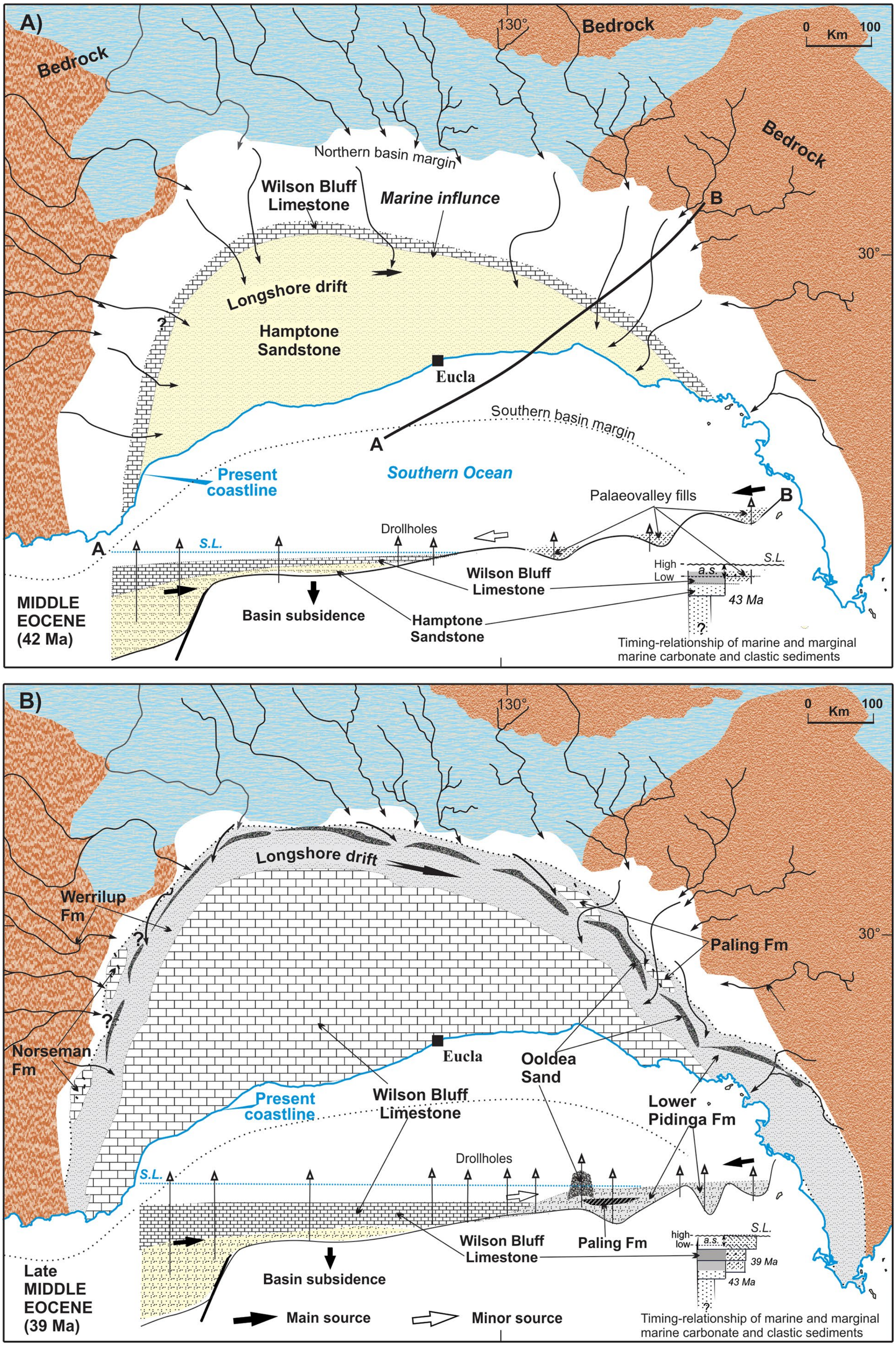
Summarised below in a 'time-slice' format are the major phases of erosion and deposition leading to construction of the sedimentary succession in the Eucla Basin and peripheral paleovalley network (Fig. 4.11). The paleoshorelines that lie between the Paleogene–Neogene onshore coastal zones and the modern shoreline or the shelf-break, was determined from drillhole data in the Eucla Basin and on the Gawler and Yilgarn cratons, and Officer Basin. The main stages and implications of the evolution with respect to the present discussion are as below.

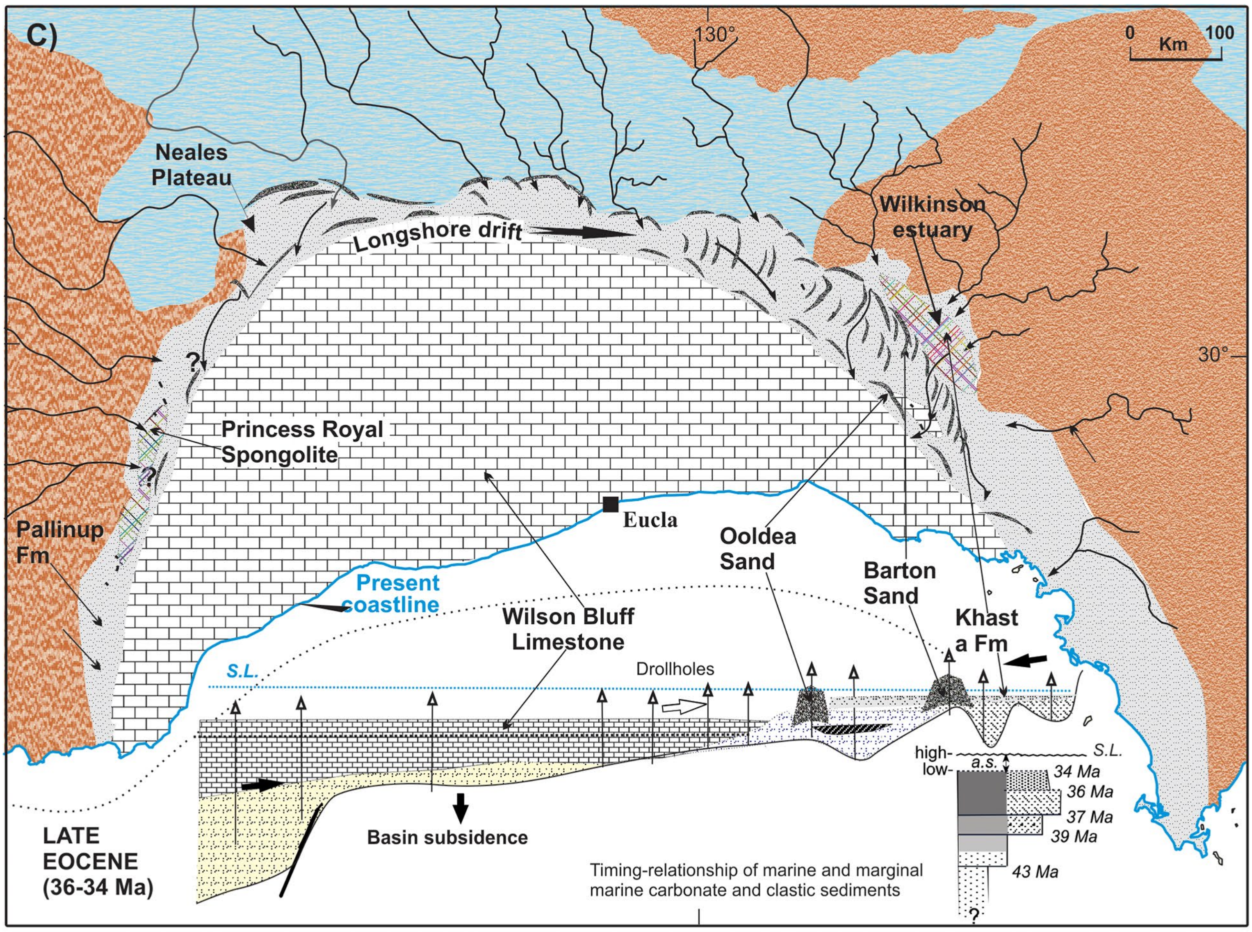
PRE-CENOZOIC

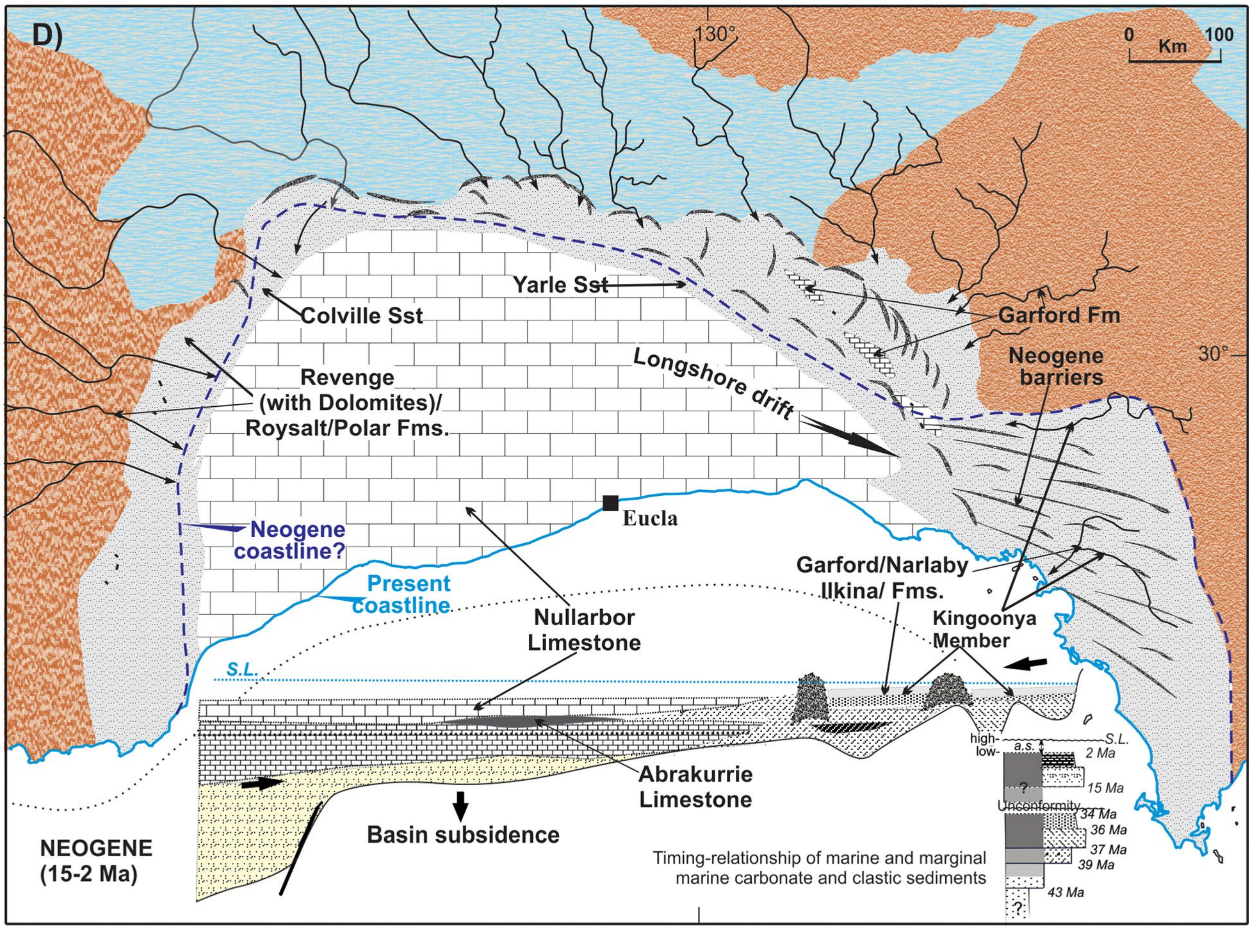
The Eucla Basin, as part of the southern Australian passive margin formed by Jurassic rifting and Early Cretaceous separation (Veevers et al. 1991), occupies and extends the landward part of the Great Australian Bight (Fig. 1.1). The basement and bedrock of the Eucla Basin is formed from Precambrian granite, schist, gneiss, and Paleozoic sedimentary rocks, and the area was a stable landmass during the Cretaceous, the final stages of separation between the Australia and Antarctica plates, while the Australian landmass spanned temperate and polar latitudes with significant inundation of the continental interior during marine highstands (Barham et al., 2016). During the Cretaceous, the Bight Basin developed following rifting from Antarctica and thermal subsidence combined with eustatic sea-level rise in the Cretaceous led to flooding of the Madura Shelf in the Eucla platform and deposition of the Loongana and Madura formations (Barham et al., 2016). With the continued thermal subsidence, the high energy fluvial regime that deposited the Loongana Formation gave way to more widespread low energy lacustrine regime in which the Madura Formation started to accumulate conformably over the Loongana Formation in places, and then the sedimentation switched to marine conditions as indicated by the presence of glauconitic and offshore siltstones (Totterdell et al. 2000).

The Bight Basin underwent a period of regional uplift at the end of the Cretaceous, which marked the depositional end of the Madura Shelf. Subsequently the onshore portion of the Bight Basin experienced a prolonged period of non-deposition of between 25–60 Ma duration (Lowry 1970; Totterdell and Krassay 2003; Hou et al. 2011b). The significant hiatus between the Late Cretaceous and Eocene in the region suggests that a long interval of non-deposition and/or erosion prevailed before significant deposition of Hampton Sandstone in offshore and its equivalents (e.g., Werrilup and Pidinga Formations) in onshore settings commenced (e.g., Alley and Beecroft, 1993; Alley et al., 1999). There is a strong contrast between the environments of deposition of the Cretaceous and the Cenozoic sediments. The former was one of low current activity and reducing bottom conditions whereas most of the Cenozoic beds of the Eucla Basin were deposited on a broad open shelf with good circulation and strong bottom currents (Lowry, 1970). The paleovalley systems surrounding the Eucla Basin that were formed before and during the Cretaceous, have been largely inactive since (Lowry, 1970).

Figure 4.11 *(next page)*. **Simplified maps and cross-sections along line A-B, showing successive stages in the depositional history of the Eucla Basin (modified from Hou et al., 2011b; 2021). (a)** Middle Eocene (Wilson Bluff Transgression and lowstand): shallow marine Hampton Sandstone (overlain by lower Wilson Bluff Limestone), onshore lower Pidinga / Werrilup fluvial. **(b)** late Middle Eocene (Tortachilla Transgression and highstand): lower Wilson Bluff Limestone (shelf), nearshore Pidinga/Werrilup marginal marine and estuarine (including transgressive barriers and lagoonal Paling and Norseman limestones), and onshore Pidinga/Werrilup fluvial. **(c)** Late Eocene (Tuketja-Tuit Transgressions and highstand): upper Wilson Bluff Limestone (shelf), nearshore upper Pidinga/Pallinup marginal marine and estuarine (including transgressive-regressive barriers and estuarine spongolite or sponge-spicule facies), and onshore Pidinga/Pallinup fluvial. **(d)** Neogene (Transgression-Regression): shelf Nullarbor Limestone, shoreface Yarle/Colville sandstones, Garford/Revenge fluvial (including lacustrine dolomites); fluvial dominant in the western margin of the basin, fluvial and lacustrine in the northeastern margin of the basin (Garford Formation), and marginal-estuarine in the eastern margin of the basin (Kingoonya Member, Narlaby and Ilkina formations).


A)
Bedrock
130°
Bedrock
0 Km 100
Bedrock
Northern basin margin
Wilson Bluff Limestone
Marine influnce
B
30°
Longshore drift
?
Hamptone Sandstone
Eucla
Southern basin margin
A
Present coastline
Southern Ocean
Palaeovalley fills
B
Drollholes
A
S.L.
S.L.
High
Low
a.s.
Wilson Bluff Limestone
43 Ma
MIDDLE EOCENE (42 Ma)
Basin subsidence
Hamptone Sandstone
?
Timing-relationship of marine and marginal marine carbonate and clastic sediments
B)
130°
0 Km 100
Longshore drift
Werrilup Fm
?
Paling Fm
30°
?
Norseman Fm
Eucla
Ooldea Sand
Present coastline
Wilson Bluff Limestone
Lower Pidinga Fm
Drollholes
S.L.
high-
low-
a.s.
S.L.
39 Ma
Wilson Bluff Limestone
Paling Fm
43 Ma
Late MIDDLE EOCENE (39 Ma)
Basin subsidence
?
Main source
Minor source
Timing-relationship of marine and marginal marine carbonate and clastic sediments


C)
130°
0 Km 100
Neales Plateau
Longshore drift
Wilkinson estuary
30°
Princess Royal Spongolite
Pallinup Fm
Eucla
Ooldea Sand
Barton Sand
Present coastline
Wilson Bluff Limestone
Khasta Fm
S.L.
Drollholes
high-
low-
a.s.
S.L.
34 Ma
36 Ma
37 Ma
39 Ma
43 Ma
Basin subsidence
LATE EOCENE (36-34 Ma)
Timing-relationship of marine and marginal marine carbonate and clastic sediments


D)
130°
0 Km 100
Yarle Sst
Colville Sst
Garford Fm
Neogene barriers
30°
Revenge (with Dolomites)/ Roysalt/Polar Fms.
Longshore drift
Eucla
Neogene coastline?
Present coastline
Garford/Narlaby Ilkina/ Fms.
Nullarbor Limestone
Kingoonya Member
S.L.
high-
low-
a.s.
S.L.
2 Ma
15 Ma
Unconformity
34 Ma
36 Ma
37 Ma
39 Ma
43 Ma
Abrakurrie Limestone
Basin subsidence
NEOGENE (15-2 Ma)
Timing-relationship of marine and marginal marine carbonate and clastic sediments

The Jurassic was a period of intense erosion in the Eucla paleovalley areas, probably promoted by the development of the rift basins along the present southern continental margin of Australia (Krieg, 1995), while much of the erosion of the weathered bedrock and inferred Permian cover (Clarke and Alley, 1993) would have occurred (Alley et al., 1999). Pre-Cenozoic erosion, possibly caused by paleorivers, produced in the areas of the Eucla Basin an irregular topography which was later buried by Cenozoic sediments. The area began subsiding in the Early Cretaceous (~Aptian) with the accumulation of lenses of non-marine conglomeratic sandstone (the Loongana Sandstone) and of siltstone and shale (the basal part of the Madura Formation). The sea entered the basin area from the south and deposition of glauconitic, carbonaceous and pyritic sediments continued into the Late Cretaceous, with possible periods of non-deposition in the Aptian–Albian and Turonian (Lowry, 1970). Then, deposition ceased towards the end of the Late Cretaceous (Santonian) and did not recommence until the Eocene.

The incision of major Eucla paleovalleys were probably initiated in the pre-Jurassic, as it is evidenced in the southern Yilgarn region where the Cowan and Lefroy paleodrainage channels developed (Clarke, 1993, 1994a, b; Alley et al., 1999; De Brockert, 2002). On the pre-Cenozoic landscapes, the Eucla paleovalleys could have originated by faulting and river incision/erosion, draining from surrounding onshore settings and discharging into the central basin (Hou et al., 2011b). This is supported by evidence of Early Eocene paleochannels beneath the Nullarbor Limestone between the central basin and Ooldea Range (Hou et al., 2008). The geometry of the paleovalley network on the southern Yilgarn Craton suggests that the paleodrainages formed a unified system with the headwaters of some streams in Antarctica prior to the break-up of Gondwana (Ollier, 1988). Most of the paleodrainage systems here flowed eastwards into the Eucla Basin area, but a few, such as the Cowan Paleovalley, drained southwards as a result of drainage reversal, probably during the Jurassic when rifting commenced in the Bremer Basin (Clarke, 1994a). This is supported by the evidence that the lower reaches of paleovalleys in the south-western margin of the Eucla Basin contain the Madura Formation of Hauterivian–Barremian age, indicating the Early Cretaceous (Jones, 1990) and presumably the paleodrainages flowed down the proximal slope of the pre-rift bulge (Alley et al., 1999). Based on geophysical evidence from offshore sub-Cenozoic strata, Clarke and Alley (1993) interpreted that two major paleodrainage systems (Mallabie and Twilight Cove paleovalleys) flowed into the Eucla Basin depocentre. The acute angles formed by convergence of the Cowan and Lefroy paleovalleys and their tributaries indicate the combined Cowan and Lefroy paleodrainage system originally flowed north and east. The width of the southern end of the Cowan Paleovalley, however, truncated by the modern coastline at Esperance, is poorly defined and probably exceeds 10 km, suggesting that the original headwaters were located in Antarctica (Oilier, 1988). Thus, the Cowan paleodrainage channel has been truncated by the onset of rifting along the southern margin of Western Australia (Clarke, 1993, 1994a, b; Alley et al., 1999). Tallaringa Paleovalley was interpreted to drained north-eastwards during Jurassic–Cretaceous to discharge into the Eromanga Basin, but as a result of drainage reversal then drained southwestwards during the Paleogene–Neogene when subsidence commenced in the Eucla Basin (Hou et al., 2007a). This is supported by the evidence that the upper reaches of Tallaringa Paleovalley in the northeastern margin of the Eucla Basin contain fluvial-estuarine sediments of Mesozoic age (Benbow et al., 1995a, b) and presumably the paleodrainages initially flowed down the proximal slope of the Eromanga Basin (Hou et al., 2012).

PALEOCENE–EARLY EOCENE

At the beginning of the Cenozoic, about 65 Ma ago, the southern margin of Australia was located at about latitude 60–65°S (compared to 40–55°S today), while Australia had begun to split from Antarctica (Bernecker et al., 1997). Downwarping and subsidence of the Eucla Basin probably recommenced during the Paleocene–Early Eocene and persisted until about the end of the Pliocene, with some depositional hiatus mainly during Oligocene throughout the basin, Early and Late Miocene in places. During early Paleocene the Australian continent lay at much higher latitudes than present and temperate rainforest existed along the southern continental margin of Australia (Alley et al., 1999). By the late Paleocene to early Eocene interval, rainforest of megathermal aspect existed in central Australia, indicating that conditions there were warmer than along the southern continental margin. Global temperatures were still relatively elevated (Frakes et al., 1987), Antarctica was ice-free, and a zone of westerly winds prevailed at 60–80°S, although weak and chaotic wind patterns may have existed in the central part of the continent (Kemp, 1978). Sedimentation in the Eucla paleovalleys largely commenced in middle Eocene times, with a few channels on the Yilgarn and Galwer cratons perhaps developing in early Eocene times (Benbow et al., 1995b; de Brockert, 2002; Hou et al., 2008; Zang and Stoian, 2006). With the subsidence of the central Eucla Basin, incision in the upper reaches of the paleovalleys is believed to have continued into late Eocene times on the basis of similar palynomorphs in the youngest part of the Pidinga Formation (Alley et al., 1999).

The paleoshoreline of Paleocene–Early Eocene would lie between the modern shoreline and the shelf-break. Given the existence of Early Paleogene terrigenous sediments from the continental slope in the vicinity of the Eyre Sub-basin (Clarke and Alley, 1993), their high-water depth (>500 m) and seismic profiles (Bein and Taylor, 1981), these most likely represent the marine component of deposition during the Late Paleocene–Early Eocene. Farther to the north, remnants of the shoreline and continental components of Late Paleocene – Early Eocene deposition presumably lie buried beneath the sediments of the Hampton Sandstone and/or Wilson Bluff Limestone deposited during the Middle Eocene marine transgressions. The 'formative streams' (yet to incise and form the incised valleys) are therefore considerably lengthened, with at least 300 km being added to the lower reaches of major Eucla paleovalley networks (de Brockert, 2002; Hou et al., 2008). Increased sediment supply resulting from incised-valley incision would have enlarged the deltas and caused the shoreline to migrate seaward.

In the central Eucla Basin, because of the low southward slope of landsurface, the fine-grained nature of the bedrock (Cretaceous Madura Formation) and a perennially high discharge (Lowry, 1970), it is likely that the formative streams were meandering and therefore able to adjust to a fall in base-level by altering their sinuosity (de Brockert, 2002). On the Gawler Craton to the east and Yilgarn Craton to the west as well as Officer Basin to the north, the formative streams flowed within the base of broad and shallow paleovalleys, (primary-valleys; Fig. 1.2) that were in existence by the earliest Cretaceous, transferring large quantities of terrigenous sediment to the evolving rift valley (Bight Basin and Eyre Sub-basin) between Australia and Antarctica, which began to form in the latest Middle Jurassic (Benbow et al., 1995a, b; Alley et al., 1999; de Brockert, 2002; Hou et al., 2008; 2012).

MIDDLE–LATE EOCENE

In middle Eocene times, monsoonal-like conditions prevailed in central Australia and moister conditions in the south, where rainforest of meso- to megathermal aspect grew, here extending late into the Eocene Epoch (Alley et al., 1999; Zang and Stoian, 2006). In the Middle Eocene deposition began in the central part of the Eucla Basin with lenticular sandstone (the Hampton Sandstone) followed by marl (the lower part of the Wilson Bluff Limestone) (Lowry, 1970). Significant downwarping was evidently restricted because marine sediments did not spread throughout the basin until the late Middle Eocene (P14; Fig. 4.1) when chalky bryozoan limestone (the middle part of the Wilson Bluff Limestone) was deposited in quiet water in most parts of the basin, and well sorted bryozoan limestone (the former Toolinna Limestone) was deposited under higher energy conditions in the southwest (Lowry, 1970). The influence of Middle–Late Eocene Wilson Bluff, Tortachilla and Tuketja transgressions extended several hundred kilometres up along the Eucla paleovalleys (Alley and Beecroft, 1993; Benbow et al., 1995d; Alley et al., 1999), while fully marine environments in the Eucla Basin were established in the later Eocene when the separation of Australia from Antarctica accelerated (McGowran, 1989; Davies et al., 1989).

WILSON BLUFF TRANSGRESSION (41–43 MA)

The Eucla Basin was still at relatively high latitudes (50–55°S, Benbow et al., 1995b) at the commencement of the Wilson Bluff (Zone P12) transgression (correlated with eustatic sea-level rise: Johnson and Veevers, 1984; McGowran, 1989, 1991; McGowran et al., 1992; Alley et al., 1999). Following the downwarping and subsidence of the Eucla Basin, the middle Middle Eocene incised-valleys were formed by stream rejuvenation in the onshore Eucla Basin, beyond the shallow marine deposition of Hampton Sandstone, which may have been influenced by a major marine transgression at 41–43 Ma (Figs 4.1 and 4.11). During this period, the alluvial/fluvial sediments of lower Werrilup and lower Pidinga formations were deposited in the middle-upper reaches of the Eucla paleovalley networks, the sea was beginning to invade the Eucla Basin during the Wilson Bluff Transgression. Deposition of the coarse-grained silici- and bio-clastic sandstone facies of the Hampton Sandstone occurred at the advancing shoreface, followed (conformably) by deposition of the shallow-water Wilson Bluff Limestone (Lowry, 1970; Hocking, 1990). An average thinness of the Hampton Sandstone in the western part of the Eucla Basin attests to low rates of sediment delivered to the advancing shoreline, probably in large part due to high rates of fluvial aggradation and sediment storage within the paleovalleys developed on the Yilgarn and Gawler Cratons, as well as the Officer Basin.

Subsidence of the Eucla Basin may have contributed to the landward migration of the shoreline, but marine incursions into the lower-middle reaches of the paleovalley networks suggest that the Eocene transgressions were dominantly eustatically controlled. The paleoshoreline was between the modern shoreline and the Ooldea Range, but at a substantially higher elevation than the marine transgressions during the Late Paleocene – Early Eocene (Fig. 4.11). Supporting evidence for a low relative sea-level (in contrast to late Middle Eocene – Tortachilla Transgression) during this period comes from the Albany region (Fig 1.1), where the Werillup Formation infills an irregular basement topography eroded to some 50 m below present sea-level (Hos, 1975), and also from the eastern region of the Nullarbor Plain, where the Pidinga Formation infills Early Eocene paleovalley (Hou et al., 2006b, 2008). Although much of the cratonic regions were undergoing erosion, possible accelerated gorge erosion combined with subsidence of the Eucla Basin probably led to deposition of basal coarser sands in paleoriver systems and finer grained, carbonaceous sediments in restricted swamps and lakes. Extensive aggradation occurred during the transgression as non-marine to marginal marine sediments in onshore setting and as highstand deposition of biogenic sediments in offshore basin where the terrigenous flux permitted. Drowning of a large part of the paleovalley area occurred promoting deposition of carbonate facies well into the basin from the limit of the Bunda Plateau.

Within the lower reaches of the paleovalley network, aggradation of the fluviatile lower Werillup and Pidinga Formations was probably also taking place in response to a climatically induced excess of sediment supply. Thus, lower stream gradients and greater incised valley widths would have favoured the development of high-sinuosity suspended-load streams with larger proportions of fine-grained and organic-rich overbank sediments being deposited in the estuarine plain where high rates of aggradation may also have been promoted by an increase in base-level associated with the Wilson Bluff Transgression.

TORTACHILLA TRANSGRESSION (39.5–41 MA)

The middle Middle Eocene sea-levels (i.e., the Wilson Bluff Transgression) remained high within the Eucla Basin for the remainder of the Eocene, with deposition of the Wilson Bluff Limestone and its facies variant (e.g., the former Toolinna Limestone), continuing unabated throughout Eocene. During the Tortachilla Transgression (late Middle Eocene, Zone P14), estuarine conditions migrated up into the paleovalleys while freshwater sedimentation developed elsewhere in channels, swamps and lakes, resulting in partial submergence of the lower-middle reaches of the Eucla paleovalley networks and deposition of the marine Wilson Bluff Limestone in the central basin with its equivalent lagoonal facies of the Norsemen and Paling Formation, and the estuarine-deltaic facies of the Werillup and Pidinga Formations in the onshore margins (Fig. 4.1). Landward migration of a high-energy, siliciclastic shoreline and the development of an extensive coastal (Ooldea) barrier in the north-eastern margin of the basin marked the highstand of deposition during the Tortachilla Transgression. A northwest–southeast-trending basement high to landward of the barrier may have provided a site on which the barrier originally nucleated (Clarke and Hou, 2000). Generally, during lowstands, tidal inlet channels behind the Ooldea barrier existed as paleorivers extending out across the highstand sea floor. There were large estuaries drowned during this transgression, and large marginal marine areas (e.g., lagoons) during highstands.

The ravinement surfaces (presented as the Princess Royal Member resting on benches eroded into weathered bedrock) closely resemble the surface approximately 20 m lower in elevation that may have been cut during the Tortachilla Transgression (Clarke et al., 2003). Similar surfaces occur throughout southwest Western Australia but are generally less well defined though lack of exposure.

TUKETJA-TUIT TRANSGRESSION (36.5–39.5 MA)

A minor marine regression followed the Tortachilla Transgression, resulting in removal of partial Werillup and Pidinga Formations from the Eucla margins, and the preservation of these units within the upper reaches of the Lefroy inset-valley network indicates that erosion decreased fairly rapidly upstream. (Clarke, 1993, 1994a, b; de Brockert, 2002; Hou et al., 2006b). In the western margin of the Eucla Basin (e.g., at Lake Lefroy), parts of the lower Werillup Formation and the overlying shallow-marine lower Hampton Sandstone were eroded by a sea-level fall which took place between the Tortachilla and Tuketja transgressions (de Brockert, 2002). After this marine regression, the Eucla Basin margins were flooded more fully by the Tuketja Transgression (early Late Eocene), resulting in deposition of the upper Pidinga Formation (Anthony Member) and upper Werillup Formation followed by the Pallinup Siltstone (Fitzgerald and Princess Royal members).

During the Late Eocene (Zones P15–16), a new barrier system, the Barton barrier, was developed in the eastern coast of the basin (e.g., the areas of the Wilkinson and Anthony estuaries), and appears to have been truncated to the south by the Kingoonya paleodrainage systems, with the Ooldea barrier forming a chain of offshore sand islands. Some former tidal inlet valleys cutting across the Ooldea Range, such as Pidinga and Ifould passes, appear to have been kept open by tidal currents (Clarke and Hou, 2000). The shoreline of the Tuketja Transgression in both west (Clarke, 1994a, b) and east (Clarke and Hou, 2000) extended much further inland at this time, as indicated by the elevation difference relative to the Tortachilla Transgression (Clarke and Hou, 2000). The maximum level of the Late Eocene transgression was at least 20 m higher than that of the late Middle Eocene transgression (assuming any post-Eocene uplift in this area is homogeneous) and led to widespread deposition of spicular facies in the marginal marine and estuarine environments and even in lower reaches of the paleochannels (Clarke et al., 2003; Hou et al., 2003a).

Indirect evidence (Clarke et al., 1996) shows that the Princess Royal Member (and thus the laterally equivalent upper part of the Wilson Bluff Formation) is the same age as the Tuketja Transgression. With Late Eocene sediments thickening towards the centre of the basin, a general relative sea-level rise occurred around the Precambrian areas of the basin margins to an elevation of about 300 m above present sea level (Hou et al., 2008).

On the western margin of the basin, Late Eocene sediments were deposited on an irregular surface of Precambrian rocks in places, where these basement rocks stood above the Late Eocene sea as the islands, such as Dean, Esmond (Russell Range) and Mount Ragged (Lowry, 1970). Farther west, north and east, the Late Eocene sea flooded the lower reaches of the paleovalleys that drained the Precambrian (e.g., Albany-Fraser, Yilgarn and Gawler) shields and old sedimentary (e.g., Officer and Eromanga) basins, such as Lefroy, Cowan, Ponton, Wanna, Lindsay, Tallaringa, and Kingoonya paleovalleys, respectively (Fig. 1.1). If at present the country were flooded to ~200 m contour in elevation (Fig. 4.8), the remaining land would have very little relief, and there may not have been much more in the Late Eocene Eucla Basin. Presumably most of the detritus carried by the paleorivers was dropped in the drowned valleys some distance from surrounding inlands of the basin.

EOCENE–OLIGOCENE BOUNDARY (35.5–36.5 MA)

Towards the end of the latest Eocene the sea retreated beyond the present shoreline and did not return until the Early Miocene. The Late Eocene depositional episode was followed by regression, judging by the erosional break between the Paleogene and Neogene depositions across the whole basin, resulting in creation of the bounding unconformity which separates the Paleogene sequence from the Neogene sequence (Fig. 4.1). The regression is probably equivalent to the major sea level lowstand of the Chinaman Gully regression (Alley et al, 1999; McGowran, 1989). Although widespread deposition continued into the Early Oligocene in most of the Tertiary basins of South Australia (e.g., Zone P18; Aldinga Transgression), the late Early Oligocene stratigraphic record in the Eucla Basin is relatively restricted (Benbow et al., 1995a, b).

OLIGOCENE HIATUS

In the Early Oligocene the sea withdrew from the Eucla Basin. This major marine regression within the Eucla Basin is recorded by a hiatus between the Wilson Bluff and Abrakurrie limestones (Fig. 4.1), and evidence for subaerial exposure of the Wilson Bluff Limestone (Lowry, 1970). A major expansion of the Antarctic ice sheet following the 'Terminal Eocene Event' at about 33.2 Ma further suggests that the regression was glacio-eustatic and took place in the earliest Oligocene. Stratigraphic evidence from the Australian Northwest Shelf, indicated the earliest Oligocene sea-level fall to have been in the order of 250 m (Quilty, 1977). Based on calculation of the Antarctic ice volume from the deep-sea oxygen isotope record, Miller (1992) suggested the eustatic fall was more likely to have been in the order of 30–90 m, perhaps reaching a maximum of 180 m.

Closely following the earliest Oligocene marine regression, a major cooling and drying of climate onshore is likely to have persisted until the mid-Late Oligocene (Fig. 4.1), which marks the onset of the third cycle of marine transgressions along the southern margin of Australia, and also the return of warm and wet conditions onshore (McGowran and Li, 1998). Sediments from Maralinga 6 drillhole (long. 131.5904961°, lat. -30.1622114°; depth interval 144.78 m to 137.46 m) have been palynologically analysed and indicate Oligocene–Early Miocene age, correlating with *Spiniferites ramosus* Zone, *Proteacidites tuberculatus* Zone and *Canthiumidites bellus* Zone respectively (Zang and Stoian, 2006). Dry rainforest was developed on land and high salinity lagoons are indicated by the abundance of specific dinoflagellate cysts such as *Polysphaeridium zoharyi*. To date this is the only record of Oligocene–Miocene marine sediments in the eastern Eucla Basin.

Oligocene conditions in the Eucla Basin area are represented largely by a phase of non-deposition, erosion and weathering, during which the extensive silcrete probably formed. During the Oligocene weathering, erosion, induration, ferruginisation, and silicification developed in the Eucla Basin, including offshore, nearshore and onshore areas. Extensive silicified sands and silcrete on the top of Late Eocene (e.g., Werrilup, Pallinup, Pidinga and Khasta formations) sediments in the onshore areas of the Eucla Basin may have formed during this time. The extensive hardening of the top 1–3 m of Wilson Bluff Limestone developed in the central basin also reflects Oligocene erosion and weathering because the Wilson Bluff Limestone was exposed during the Oligocene (Lowry, 1970), a period of about 11 million years (Harland et al., 1964).

MIOCENE–PLIOCENE

Due to lack of the accuracy of the age correlations, a time interval of Oligocene is proposed between the deposition of the Wilson Bluff Limestone and the Abrakurrie Limestone by comparing the visible effects of Oligocene erosion with those at the contact of Wilson Bluff Limestone and the Abrakurrie Limestone (Fig. 4.1). It is probable that the Early Miocene transgression eroded a weathered layer at the top of the Wilson Bluff Limestone before deposition of the Abrakurrie Limestone (Lowry, 1970; Dickinson et al., 2002). In the Eucla Basin, a series of transgressions occurred from early Oligocene times through to the Middle Miocene (Benbow et al., 1995b; d), extended across the shelf to the inland margin of the Bunda Plateau, but do not appear to have breached the Tortachilla shoreline (e.g., the Ooldea barrier) and thus had little influence on sedimentation in the paleovalleys north of the Tortachilla shoreline. In the onshore basin, the Miocene-Pliocene interval was a time of development of extensive shallow, alkaline lakes in parts of the paleovalleys. Lakes in the inland area supported a diverse fauna, including crocodiles (Alley et al., 1999). Vegetation had changed to dry, open woodland throughout the paleovalley areas, with rainforest-like vegetation confined to wetter valley bottoms.

UPPER MANNUM-LONGFORD TRANSGRESSION (20–24 MA)

Although widespread marine conditions did not return the Eucla Basin until deposition of the Nullarbor Limestone in the late Early Miocene, the Abrakurrie Limestone was deposited by a minor marine transgression (Upper Mannum-Longford) which did not extend very far inland from the present coastline (Lowry, 1970). The marine transgression commenced at the beginning of the Early Miocene and resulted in deposition of the Abrakurrie Limestone in the central part of the Eucla Basin (Figs 4.1 and 4.11). The limestone is rich in bryozoans but lacks a muddy calcite matrix, because it accumulated under high enough energy conditions to wash away any fine-grained materials, resulting in high porosity (~40%) and permeability of the Abrakurrie Limestone. There is no evidence that other parts of the Eucla Basin were submerged in the Early Miocene, so this marine transgression was probably restricted in the central basin area largely due to downwarping of the Eucla Basin (Lowry, 1970).

BAIRNSDALE/TAMBO RIVER/CADELL/BALCOMBE? TRANSGRESSION (14–15.5 MA)

After a brief retreat, the sea spread over a much wider area, with deposition of the Nullarbor Limestone in most parts of the basin and the Colville Sandstone in the northern part of the basin and the Yarle Sandstone in the northeastern margin of the basin. The Colville Sandstone and Yarle Sandstone were deposited around the inner margin of the Eucla Basin. Marine currents were not frequent or strong enough to develop common current bedding and coarse-grained beds, and the lack of terriginous material in the Nullarbor Limestone indicates that there was no source of terriginous material from offshore, in the Miocene. The abundant terriginous material in the Colville Sandstone and Yarle Sandstone was transported essentially by paleorivers flowing into the Eucla Basin from the Precambrian areas of Yilgarn and Gawler cratons and Officer Basin to the western-northwestern and eastern-northeastern margins of the basin.

Cadell/Balcombe transgression-initiated deposition of widespread neritic (Nullarbor) limestone on the Bunda Plateau, seaward of the Ooldea Range, while fluvial and lacustrine sediments dominated the margin and paleovalley settings of the Eucla Basin. Several phases of rejuvenation of the Ooldea Range are indicated by younger phases of Ooldea Sand during highstands in the: (i) Early Miocene Upper Mannum Longford Transgression and (ii) Middle Miocene Cadell/Balcombe Transgression (Benbow et al., 1995b). During the early Middle Miocene the sea probably reached the Ooldea Range in the northeast, where the Ooldea Range was in part recycled as the Miocene coastline, and reached the prominent scarp (e.g., limit of the Colville Sandstone) in the north and northwest. In the onshore settings, the centres of deposition were controlled by local epeirogenesis and by earlier Eocene paleochannels on the Eucla Basin landward margins. The inland sediments remain poorly dated, but maximum deposition may have occurred during the Cadell/Balcombe Transgression (Zones N9–11), around the Early-Middle Miocene boundary, more or less contemporaneously with maximum deposition over the marine platforms (Benbow et al., 1995b). The sea withdrew for the last time about 10 Ma ago, and soon afterwards the region of the Eucla Basin was gently uplifted (Fig. 4.1). Later sea-level rises eroded the coastal cliffs and formed the Roe and Israelite plains. During the past 10 Ma the surface of the Nullarbor Limestone has been exposed and eroded.

LATE MIOCENE UNCONFORMITY

Sediments deposited above the unconformity marking the upper surface of the Middle Miocene sequences dominantly reflect deposition under an arid to semi-arid climate, resulting primarily from major expansions of the Antarctic ice sheet in the Middle Miocene (Fig. 4.1), coupled with the continued northward migration of Australia into warmer latitudes (Kemp, 1978). The carbonates of the Eucla Group have been largely subaerially exposed since the middle Late Miocene (Lowry and Jennings, 1974). The mid-Miocene Nullarbor Limestone forms the upper surface of the Nullarbor Plain. Typically overlying the fills of paleovalleys are sequences of playa and playa-lake sediments, such as gypseous sands and clays of the Polar Bear/Roysalt and Garford formations. The base of the Nullarbor Limestone is today tilted gently southward (~0.03°) and when projected oceanward intersects the outer part of the eroded inner shelf (James et al., 2006). The inner part of the modern seafloor is confirmed by seismic images and shallow coring to be bevelled, cut into progressively older Eucla Group carbonates landward (James et al., 1994; Feary and James 1998). The Roe Plains are contiguous with, and part of, this larger erosional inner Great Australian Bight shelf sector (James et al., 2006).

With a major time of globally low sea-levels (Haq et al. 1987; Miller et al. 1998), the Late Miocene tectonic event occurred throughout southern Australia (McGowran et al., 1997) is generally interpreted to record early stages of basin inversion (Dickinson et al. 2002; Sandiford, 2003, 2007). This interval, designated as Megahiatus 2 (9–8 Ma; Fig. 4.1), is interpreted to reflect mass wasting and erosion due to a series of strong, rapid uplift and subsidence events (Li et al., 2004), resulting in the uplifting and seaward tilting of Eucla Group strata in the central Eucla Basin. Then, the tilted Eucla Group limestones were likely subject to subaerial exposure and karstification, with deposition confined to the outer accretionary (Eucla) shelf sector (Feary and James, 1998), which was the start of the prolonged period of karsting that characterises the Nullarbor Plain today (James et al., 2006). Along the (Great Australian Bight) shelf, basin inversion has dramatically changed the nature of deposition from the Eucla Group passive margin accumulation to a partitioned shelf system in the Neogene (James et al., 2006).

JEMMYS PT – HALLETT COVE / WHALERS BLUFF TRANSGRESSION (2.5–5.5 MA)

An interval of high sea-level and warm conditions occurs in the Pliocene both globally (Zachos et al., 2001) and throughout the Southern Ocean (Hodell and Warnke, 1991). The Pliocene units across southern Australia are transgressive and usually interpreted as consistent with regionally high sea-levels (McGowran, 1979; Alley and Lindsay, 1995) that are generally consistent with the warm Pliocene climate (Tiedemann et al., 1994). The regional Pliocene transgressions are the most likely times for shoreline and marine erosion to have begun during the latest Miocene–Early Pliocene (Zones N18–20; Fig. 4.1), while uplift of the Eucla Basin platform resulted in cessation of deposition and a change to subaerial exposure and development of karstic features (Benbow et al., 1995b). The tilting of the Eucla Basin towards northeast resulted in the extensive deposition in the Narlaby Plain, while the western margin of the Eucla Basin was uplifted (Hou et al., 2008). The uplift on the western Eucla margin resulted in essentially no accommodation space, and the record is therefore one of marine planation and little or no sediment accumulation. The major periods of uplift, while abruptly truncating deposition in the western and southwestern parts of the basin, are mostly followed by marine erosion and planation (James et al., 2006). In the eastern Eucla Basin, however, the presence of marine dinoflagellates south of the Ooldea barrier indicates marine influence in the lower reaches of the paleovalleys, extending at least into the central Eyre Peninsula (Alley et al., 1999; Hou, 2008; Zang and Stoian, 2009; Krapf et al., 2019), due to the uplift and tilting of the Eucla Basin (Hou et al., 2008).

Along the paleovalley systems, lacustrine dolomitic carbonates were deposited on older Miocene sediments in places. The climate became increasingly dry around 5 Ma ago, at the beginning of the Pliocene (Fig. 4.1). The limited/isolated dolomitic carbonates with gypsum at the top of the Garford Formation suggest increasing aridity. By the Pliocene Epoch further drying had produced a chenopod shrub to open woodland environment, containing isolated pockets of forest in edaphically suitable sites (Alley et al., 1999; Stoian, 2002a, 2002b, 2002c, 2002d; Zang and Stoian, 2006; Krapf et al., 2019). This increasing aridity was interrupted by a warm and wet episode from about 5–3 Ma ago, when there was substantial deposition of muds and sands in the paleovalleys (Clarke, 1994a, b; Alley et al., 1999).

LATE PLIOCENE–QUATERNARY

The dominance of evaporites and the lack of clastic material in the extensive fluvio-lacustrine sediments of the Eucla paleovalleys formed during Pliocene times indicate very reduced, intermittent flow along paleochannels and a marked drying and/or warming (Alley et al., 1999). Part of the Serpentine Lakes paleochannel, however, was reactivated, leading to the deposition of a broad alluvial fan south of the Ooldea Range (Benbow et al., 1995d). Increasing aridity during the Pliocene was punctuated by warm, wet episodes (Benbow et al., 1995d). These intervals may have facilitated the widespread weathering, silicification and ferruginisation so characteristic of Neogene sediments in the basins (Alley et al., 1999). Regionally, the Holocene veneer over Neogene bedrock confirms that most or all previous Pleistocene deposits have been erased by marine/shoreline ravinement associated with transgression and regression (James et al., 2006).

4.6.3 Summary

Subsidence of the Eucla Basin probably commenced in the Late Cretaceous – Middle Eocene shortly before the earliest paleovalleys formed, in concert with on-going slight epeiorogenic uplift of the cratonic margin and incision of the valleys (Frakes and White, 1997). Increase in sediment supply following tectonic uplift and rejuvenation of relief (accelerated stream incision) in headwater drainage basins has long been regarded as a major cause for aggradation within stream networks and the construction of thick alluvial/fluvial sequences (Miall, 1996). Sedimentation of the Eucla Basin may have commenced in the Paleocene–Early Eocene with undifferentiated marine to fluvial clastics of the Hampton Sandstone and Pidinga Formation along part of the southern basin, and terminated in the Pliocene with marine carbonates of the Nullarbor Limestone and fluvial clastics and carbonates in the onshore and nearshore settings (Taylor, 1975; Barten, 1975; Davies et al., 1989; Clarke et al., 2003, Hou et al., 2003c, 2006b, 2008, 2011b; Zang and Stoian, 2006). In the far west, Paleocene–Early Eocene sediments form a prograding wedge (Bein and Taylor, 1981).

Prior to the Wilson Bluff marine flooding, deposition of fluvial gravel and coarse sand was limited to the valley thalwegs near the coastline. This interpretation implies that unless the fluvial deposits were originally thicker and subsequently eroded, there was no significant fluvial aggradation in the landward extremities of the incised valleys during either lowstand or the subsequent Wilson Bluff sea-level rise (Hou et al., 2001a; 2003c). The later basin and channel sediments thicken and are inclined gently in a seaward direction, and the Middle Eocene sediments developed in the coastal-estuarines plain during the Tortachilla Transgression. The landward migration of the estuary and coastaline during the Tuketja-Tuit Transgressions led to the formation of the Late Eocene sediments. The sea did partially invade the central Eucla Basin during the Late Oligocene – Early Miocene leading to deposition of the Abrakurrie Limestone (Li et al., 1996a), and then more fully during the Early–Middle Miocene leading to deposition of the Nullarbor Limestone (McGowran et al., 1997). The Abrakurrie Limestone is up to 100 m thick in Western Australia, where syndepositional subsidence is likely, and is generally <10 m thick in South Australia. The relatively thick and localised occurrence of Abrakurrie Limestone compared with the overlying thinner but more aerially extensive Nullarbor Limestone suggests local tectonic subsidence in the central part of the basin in the Early Neogene. Farther inland (?50–100 km), post-Abrakurrie erosion prior to and during initial deposition of the Nullarbor Limestone has removed any identifiable record of laterally equivalent Late Oligocene – Early Miocene terrigenous facies on the Eucla margin and paleovalleys (Benbow et al., 1995a; Alley et al., 1999; Hou et al., 2003a).

During the Middle Miocene and Early Pliocene transgression and highstand, controlled by base-level change in onshore setting, there was reactive erosion and incision of the Eocene channel deposits while lacustrine sedimentation dominated in the lakes and channels. In Late Miocene-Early Pliocene time, the paleochannels were fragmented into chains of relatively small lakes, where dolomitic carbonates accumulated in the late arid stages.

5. ECONOMIC GEOLOGY

5.1 INTRODUCTION

The Eucla Basin and peripheral paleovalleys have had a long but intermittent history of investigation for placers and uranium as well as groundwater, such as the paleoshorelines and paleovalleys that drained the eastern Yilgarn Craton (e.g., Baxter, 1977; Uranerz Australia Pty Ltd, 1986; Smyth and Button, 1989; Jones, 1990; Devlin and Crimeen, 1990; Fulwood and Barwick, 1990; Commander et al., 1991; Morgan, 1993; Lintern et al., 1997; Carey and Dusci, 1999; Johnson et al., 1999; de Broekert, 2002; Bronzewing Gold Ltd, 2006) and western Gawler Craton (e.g., Close, 1973; Benbow, 1990b, 1993; Ferris, 1994; Benbow et al., 1995a; Hou and Alley, 2003; Hou and Warland, 2005; Hou and Keeling, 2008; Hou et al., 2001a, 2003b, 2006a; 2010; 2011a, b; 2014; 2017a, b; 2021; Krapf et al., 2019). Due to the huge area and extensive surficial cover, further drillholes in the Eucla Basin and peripheral paleovalleys would be desirable to extend understanding of the stratigraphy of the basin and adjacent paleovalleys, to relate this to the surrounding geology, and to evaluate the economic potential of the basin and channel-confined sediments. Details of individual deposits/prospects have been described previously, including in references listed above. This chapter briefly reviews the economic resources and potential for further mineral discovery hosted in the Eucla Basin and peripheral paleovalleys.

5.2 PLACER DEPOSITS

The onshore and nearshore settings of the Eucla Basin are favourable for the formation of channel and beach placer deposits. So far, the channel placers have been found mainly in the western Eucla paleovalleys and the most significant beach placers have been discovered in the eastern Eucla margin (Fig. 5.1).

5.2.1 Channel placer

Some heavy minerals (HMs) such as gold may have been carried in streams as bedload or suspended/saltating load in the case of fine particles. A change in flow velocity would have resulted as the paleoriver narrowed and then widened again, or changed direction of flow at point bars, or joined with other paleorivers. Velocities could also change where fast-moving narrow tributaries connected with slower moving and relatively wide tributaries. The heavier minerals in a fast-moving stream would drop out because of loss of velocity and mineral accumulation would result (Evans, 1987).

Several large paleorivers along the western margin of the Eucla Basin drained through the Goldfields Province of the Yilgarn Craton, which is particularly rich in gold associated with the Norseman-Wiluna Belt and has a long history of gold mining (e.g., de Broekert, 2002). Some of the paleovalleys host alluvial/fluvial gold placers (including deep leads) and supergene enrichments derived from the primary bedrock mineralisation (e.g., Zuleika Sands and Lady Bountiful Extended, in the upper reaches of the Roe Paleovaley system, northwest of Kalgoorlie, e.g., Lintern and Butt, 1998; Devlin and Crimeen, 1990; Commander et al., 1991; de Broekert, 2002) (Fig. 5.1). Like placer gold, heavy minerals also possibly accumulated in the areas where fast moving high-order tributaries connected with slower moving low-order tributaries or in turn with major rivers.

In the eastern Eucla Paleovalleys, some favourable sites for the concentration of placer gold and other heavy mineral deposits were predicted where the head waters could be traced to areas of known basement gold deposits (Hou, 2008; Hou et al., 2001a). Mineral exploration drilling has identified variable grades of gold mineralisation in basement rocks in areas of the Garford, Anthony and Kingoonya paleodrainage networks. These may have provided sources for placer mineralisation in the paleochannels and their tributaries. Traces of placer gold were found in channel sediments located in South Hilga and nearby Tarcoola gold deposits (S. Daly, pers. comm., 1998). These remain to be further investigated. Although no placer gold deposits have been located in the eastern Cenozoic channels, their presence is likely, particularly in tributaries spatially located nearby basement gold mineralisation.

5.2.2 Beach placer

The vast majority of the heavy mineral (HM) resources of the Eucla Basin are hosted by ancient beach and sand dune deposits that formed along Middle Eocene to Pliocene shorelines (Hou et al., 2011a, b). Since 2004, several large and high-grade HM deposits have been discovered along the Paleogene and Neogene coasts of the Eucla Basin (Fig. 5.1).

Exploration for HMs was initially carried out in areas of the eastern Eucla Basin during the mid-1980s to the early 1990s, with results of 1–10% HMs reported in some drill samples (Ferris, 1994). Early samples of Eocene sands from Lake Anthony – Lake Bring area of the eastern Eucla margin showed little enrichment of HMs (hematite, degraded ilmenite, rutile, zircon, xenotime, kyanite, garnet and other silicates (up to 8.27 wt%), but the mineral assemblage comprised predominantly ilmenite (37%) and goethite (41%), with lesser amounts of zircon and rutile (Close, 1973). Trace amounts of HMs were reported also over the Ooldea, Paling and Barton ranges (Benbow, 1990b), which were confirmed by further investigations during the 1990s, with HM concentrations of 0.16–1.59% comprised of 65–85% altered ilmenite, 10–20% zircon, and trace rutile and leucoxene (Ferris, 1994).

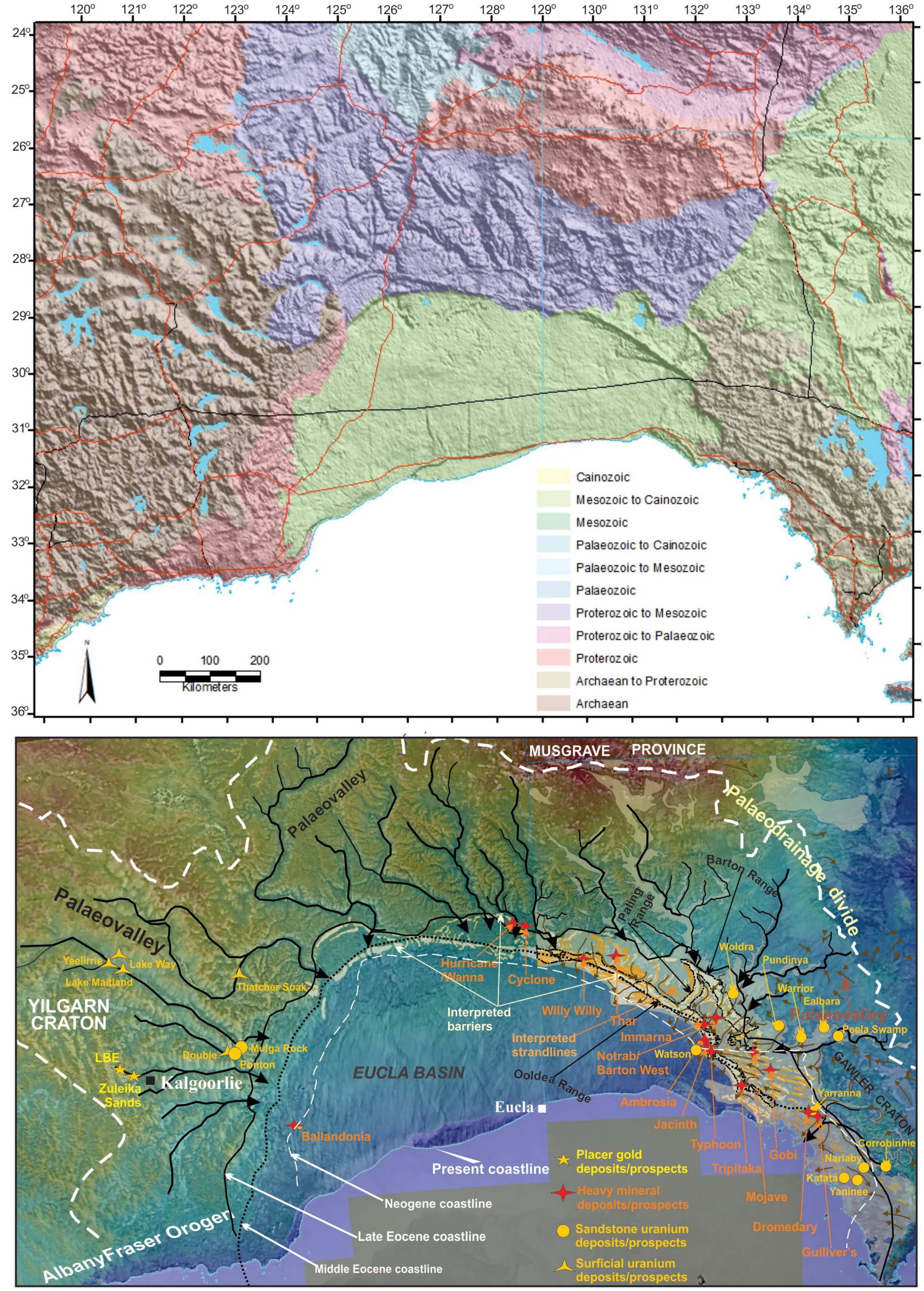


Figure 5.1 SRTM DEM imagery draped over NOAA-AVHRR night-time thermal imagery; the textural difference between paleovalleys and the sand barriers and surrounding basement terrain is apparent; showing orientation of the barriers, strandlines, paleovalley systems, dune ranges, and adjacent major features, including heavy mineral deposits and prospects, different age shorelines, and the relative high relief of the Musgrave province and the Yilgarn Craton / Albany-Fraser Orogen when compared to the Gawler craton to the east (modified from Hou et al., 2011b).

The recognition that Ooldea and Barton sand ranges were remnants of coastal deposits formed along the eastern Eucla Basin margin (Benbow, 1990a) stimulated resource companies to explore the region for HMs. However, the extent and complexity of the paleoshoreline deposits hampered early HM exploration. Almost all of the early exploration for HMs in the eastern Eucla Basin focused on the previously recognised, one-generation prospective Hampton Sandstone and Ooldea Sand along the Ooldea Range (Ferris, 1994). During 2000s, based on review of the borehole information and reconstruction of the paleoshorelines, the interpretation of a stepwise evolution across the eastern basin suggested various models for HM concentration (Hou et al., 2003a, b), which were incorporated into the stages of coastal barrier formation, being used to predict sites with higher prospectivity. Aided by this perspective and results of earlier drilling, renewed HM exploration by Iluka Resources Ltd in June 2004, had quick successes with greenfield discoveries of the Jacinth (November 2004) and Ambrosia (December 2004) HM deposits (Hou and Warland, 2005). The discovery of the first economic deposits (at Jacinth and Ambrosia) in the Eucla Basin alerted the world to a vast new greenfields HM province that contained deposits with exceptionally high zircon concentrations that made up ~50% of the HM suite (Hou and Keeling, 2008). The world-class zircon-dominated HM deposits of Jacinth and Ambrosia contain a combined resource of over 9.5 Mt HMs. HM average content amounts to 4.7%, up to about 50% of which is zircon (Hou et al., 2011a, b). Ongoing exploration identified further significant HM prospects on the eastern and northern margins of the basin (Table 5.1) while anomalous HMs have been reported in Cenozoic sediments near Balladonia on the western margin. Some deposits (e.g. Jacinth, Ambrosia, Tripitaka) are characterised by zircon-dominant HM assemblages throughout, and in this respect they differ from most HM deposits worldwide. The discoveries of high-grade, zircon-dominant, HM deposits in the onshore paleoshoreline systems of the Eucla Basin (Fig. 5.1), southern Australia, established the region as a new paleobeach placer province of major economic significance (Hou et al.,2011a, b).

Table 5.1a Eucla Basin heavy mineral (HM) deposits with resource estimate.

Deposit	**Overall dimensions**	**Resource**			**Mineralogy**		
		Tonnage (Mt)	**HM grade** (%)	**Contained HM** (Mt)	**Zircon** (%)	**Ilmenite*** (%)	**Rutile** (%)
Jacinth	3.2 km length, up to 0.9 km wide, ore thickness avg. 20 m, overburden thickness avg. 7 m.	124.3	5.2	6.5	47	30	5
Ambrosia	2.2 km length, up to 0.7 km wide, ore thickness avg. 12 m, overburden thickness avg. 8 m.	114.8	2.7	3.0	50	21	5
Typhoon	3.0 km length, up to 0.4 km wide, ore thickness 1–21 m, overburden thickness 5–27 m.	22.0	6.1	1.3	14	76	1
Atacama	7.0 km length, up to 3.0 km wide, ore thickness 1.5–18 m, overburden thickness 5–42 m.	29.2	11.3	3.3	15	75	c
Tripitaka	3.7 km length, up to 1 km wide, ore thickness avg. 10 m, overburden thickness avg. 9 m.	42.0	2.4	1.0	65	9	5
Cyclone	5.0 km length, up to 2.5 km wide – comprised of up to 5 coalescing beach/dune strandlines, ore thickness avg. 18 m, overburden thickness avg. 11 m.	98.4	2.9	2.8	33	44	12
TOTAL		**430.7**		**17.9**			

Notes

* Ilmenite includes altered ilmentite minerals - hydrated ilmenite (Ti/Ti+Fe = 0.5-0.6), pseudorutile (Ti/Ti+Fe = 0.6-0.7), and leucoxene (Ti/Ti+Fe = 0.7-0.9)

rounding may generate differences in the last decimal place.

Source: Company information releases to the Australian Stock Exchange prior to January 2011.

Table 5.1b Eucla Basin heavy mineral prospects.

Prospect	Indicative size	Indicative grade and mineralogy
Barton West	Variable HM intersections over an area 25 km by up to 12 km.	Discontinuous intersections, avg. 1–3% HM over a max. 16 m. HM content range 6–21% zircon, 2–10% rutile, 50–67% ilmenite, 6-24% leucoxene.
Barton	Over a strike length of ~12 km, and from 1.3–2.4 km wide, and from 4.5–9 m thick.	Inferred resource is ~285 Mt at 1.9% HM. Assemblage is approximately 47% ilmenite, 8.3% total zircon, 2.6% total leucoxene, and 2.2% total rutile.
Immarna	Three strandlines 4–8 km long and 150–500 m wide.	Upper and lower HM intervals of avg. 2 m thickness below 18 m overburden. Upper zone grade 2.7% HM with 86% ilmenite/leucoxene, 11% zircon.
Willy Willy	At least 900 m long and ~300 m wide.	Grades up to 4.93% HM over 6 m. Zircon content high at 25–40% VHM fraction.
Mojave	~ 8 km long and 1–3.5 km wide.	Grade range 1–22% HM over avg. thickness 10 m, below avg. 15 m thickness overburden. Preliminary mineralogy 13% zircon, 30% leucoxene and rutile, 2% ilmenite.
Dromedary	~ 1 km long by up to 500 m wide.	Up to 4.5 m at 8.4% HM. Indicative content 17% zircon, 57% ilmenite, 4% leucoxene. Overburden avg. 25 m thickness.
Gulliver	~ 7 km long and up to 2.5 km wide.	Ave 1–3% HM. Indicative HM content 60% altered ilmenite, 21% zircon, 5% leucoxene, 2% rutile. Thickness 3–7 m below avg. 25 m thickness overburden.
Cyclone Extended	2 km extension of Cyclone deposit that is up to 800 m wide. Two narrower standlines to the east have been traced for 4.5 km.	Grades up to 6.8% HM over 8 m. Maximum thickness around 17 m, below ave. 15 m overburden. Zircon content above 40% VHM fraction in some samples. Fines content avg. 4.2%.
Balladonia	Lefroy paleochannel outlet strandline ~1.5 km by 300 m wide.	Grades to 5% HM. Indicative HM content 82% ilmenite, 1.4% zircon, 3% leucoxene, 2% rutile. Thickness 6–10 m below avg. 10 m thickness overburden.
Plumridge	Minor strandlines, preliminary test drilling only.	Up to 4.2% HM of dominantly primary ilmenite with low rutile and zircon content.

HM – heavy mineral, VHM – valuable HM (includes zircon, ilmenite, rutile, leucoxene).

Based on the relationships of beach placer with the host barriers and associated shoreline facies, four different types of HM deposit are recognised in the Eucla Basin: lag deposits, transgressive deposits, regressive deposits, and aeolian deposits (Fig. 5.2), which is similar to those in the important deposits of southeastern Australia (Roy, 1999). The unique combination of eustatic, dynamic topographic, climatic and source factors provide the basis for the integrated models for HM beach placer accumulation in the Eucla Basin (Fig. 5.3). Several types of HM deposits probably exist in the eastern Eucla Basin (Fig. 5.4): (i) lag deposits along erosional disconformities (e.g., contacts between barriers of different ages) and/or unconformities (e.g, contact with bedrock); (ii) transgressive deposits landward of highstand (swash-aligned) barriers, including those trapped near paleovalley passes; (iii) regressive deposits at the front of prograded barriers (particularly in the Neogene barriers of the southeastern basin); and (iv) aeolian deposits, as low-grade disseminated concentrations in transgressive dunes. Exploration models can be further improved by greater understanding of the broader controls on sand movement and reworking across the basin.

PRINCIPAL UNITS AND SHORELINES RELATED TO HMS

Historically the Paleogene deposits of the Eucla Basin were recognised as a major transgressive – regressive cycle (e.g., Benbow et al., 1995a), with mapping in the eastern basin recognising Hampton Sandstone, Wilson Bluff Limestone, Pidinga Formation, Ooldea Sand. Of these, the Hampton Sandstone and Ooldea Sand were considered as prospective for heavy minerals across the eastern Eucla Basin (Benbow et al., 1995a; Rankin et al., 1996; review by Ferris, 1994; Hou et al., 2003a, b). Thus, HM exploration in the eastern Eucla Basin primarily involved exploring along the Ooldea Range (Ferris, 1994). Benbow (1990a) recognised the Ooldea, Barton, and Paling ranges as Eocene shoreline features and named this unit the Ooldea Sand. Subsequently, the terms 'lower' Ooldea Sand and 'upper' Ooldea Sand were used informally to separate the Eocene sand units in different age dune systems, but no physical or boundary criteria are presently known (Clarke et al., 2003). More recently, the terms Ooldea and Barton sands have been introduced.

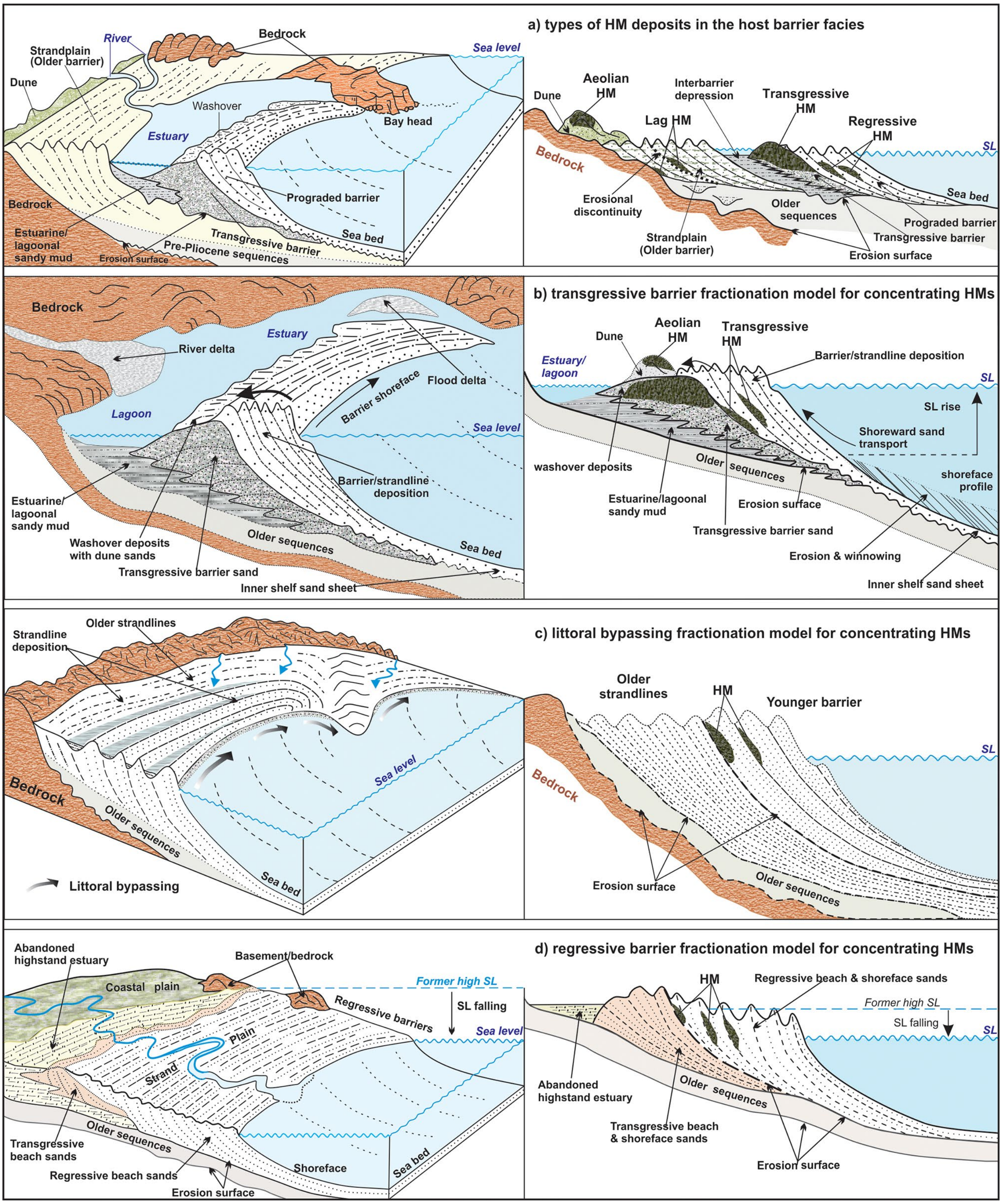


Figure 5.2 Genesis of South Australia's heavy mineral deposits based on the HM studies of Murray Basin and Eucla Basin (after Roy, 1999; Hou et al., 2003b, 2011b). (a) Four types of HM deposits defined by their location in the host barriers (lag deposits occurring along erosional discontinuities at the contact between different aged barriers along drift-aligned coasts bypassing and winnowing; transgressive deposits formed in swash-aligned, compartmented embayment at the rear of prograded barriers in their transgressive facies, immediately landward of the strandplain; regressive deposits located in compartmented embayments within prograded barriers of strandplain; aeolian deposits, relatively low grade disseminated HM deposits occurring in transgressive dunes downwind from HM-rich beaches). **(b)** Transgressive barrier fractionation model for concentrating HMs. **(c)** Littoral bypassing fractionation model for concentrating HMs. **(d)** regressive barrier model for concentrating HMs. From Hou et al. (2021).

This classification allows different-age barrier systems to be independently characterised (Hou et al., 2006a), but the distinction between barrier systems is not everywhere apparent. These marine-coastal barrier island sand complexes (including shoreface, aeolian and lagoonal components) are recognised as the main HM-bearing units. They are extensive along the eastern and northern basin margin and cover an area that is up to 30 km wide and 1,000 km long. Therefore, the revised time-bounded stratigraphic units and recognised paleoshorelines (Clarke et al., 2003; Hou et al., 2003b and 2006a) assist with targeting the heavy mineral sands (HMS) accumulated in the key facies zones along the Eucla margin (see Chapter 4; Hou et al., 2003b and 2011a, b). The paleoshoreline evolution indicates that deposition and concentration of the HM-bearing sediments resulted from multiple episodes of sea-level rise and attendant wave-wind action, multiple cycles of deposition, erosion, transport, and redeposition prior to final accumulation (Fig. 4.11).

There have been five major marine incursions into the Eucla Basin during the last 50 million years, of which four sets of shorelines, ranging in age from middle Eocene to Pliocene, have potential to host HM deposits (Fig. 4.1). The Middle–Late Eocene paleoshorelines of the northern and eastern Eucla basin are highly prospective for beach sand-hosted HMs related to sea-level changes (Hou and Warland, 2005; Hou and Keeling, 2008). Prevailing westerly winds built extensive barrier/dune systems by longshore drift. The geomorphic features of the northern and eastern Eucla Basin are very favourable for the formation of beach placers by long-shore drift (Figs 5.1 and 5.3). Effective delivery of sand by the longshore drift was the transport agent instrumental in the formation of extensive coastal barriers. Sediment movement was from west to east. Detrital zircon grains from the paleoshorelines show a distribution of zircon age that is consistent with the Proterozoic Musgrave, Coompana, Madura provinces as the dominant primary source areas of the heavy minerals, with a contribution from the Albany-Fraser Orogen, Yilgarn and Gawler cratons (Hou et al., 2006a; Reid et al., 2006; Hou et al., 2011a, b; Gartmair, 2022; Gartmair et al., 2021). The likelihood is that these HMs have been recycled via sedimentary basins that flank the Musgrave Province, including the Neoproterozoic to Cambrian Officer Basin and Permian to Mesozoic deposits of the Bight Basin (see below).

The most prospective strata include the barrier and associated sandstones of the paleoshorelines that were buried by voluminous sand dunes over 40 Ma. Quaternary sediments have blanketed the land surface and cover the eastern Eucla Basin with a series of longitudinal dunes, known geographically as the Great Victoria Desert. The geographic and stratigraphic distributions of HM-bearing sandstone in Paleogene and Neogene sediments suggest contemporaneous transport through paleovalleys predominantly draining Precambrian settings. Ideas on the formation and reworking of paleostrandlines of the Eucla Basin continue to evolve. Earlier models that identified dominant sediment input from large paleovalley networks that drained across the Precambrian Gawler Craton were revised after detrital zircon ages clearly identified rocks of the same crystalline age in the Musgrave Province as an important source area for barrier sands (Reid and Hou, 2006; Reid et al., 2013). Recent work of Gartmair (2022) suggests that the main sources of the HMs derived from Musgrave, Coompana, Madura provinces. Continental-scale tilting during the Late Cenozoic in response to northward drift of the Australian continent (Veevers, 2000; Sandiford, 2007) is supported by Eucla Basin studies (e.g., Hou et al., 2008, 2011b). The influence of such a dynamic topography is evident in the changing pattern of sedimentation and explains the greater extent of marine inundation on the eastern margin and the discordant alignment of younger paleobarrier deposits formed during Miocene–Pliocene sea level highstands (Hou et al., 2008).

MIDDLE EOCENE SHORELINE FACIES

In the central Eucla Basin, offshore marine sand and some of the terrigenous sand in the paleovalleys was recycled during the Middle Eocene (Wilson Bluff) transgression and transported to the onshore margin where it was deposited as Hampton Sandstone (Fig. 4.11a), unconformably on cratonic basement or older sediments (e.g., Hocking, 1990; Benbow et al., 1995a). These early beach deposits were later buried by marine limestone. A poorly defined Wilson Bluff shoreline of inferred beach facies of the Hampton Sandstone (Hou et al., 2003b, c, 2006a), may express placer characteristics in the thin beach sand sheet and basal lag, although limited drilling data indicate that this lithofacies contains no anomalous levels of heavy mineral sands (Hou et al., 2011a, b). This shoreline remains prospective for Middle Eocene beach placers (zones P12 and P13; Fig. 4.1). The economics of exploring and mining such deposits beneath marine limestone has not been evaluated.

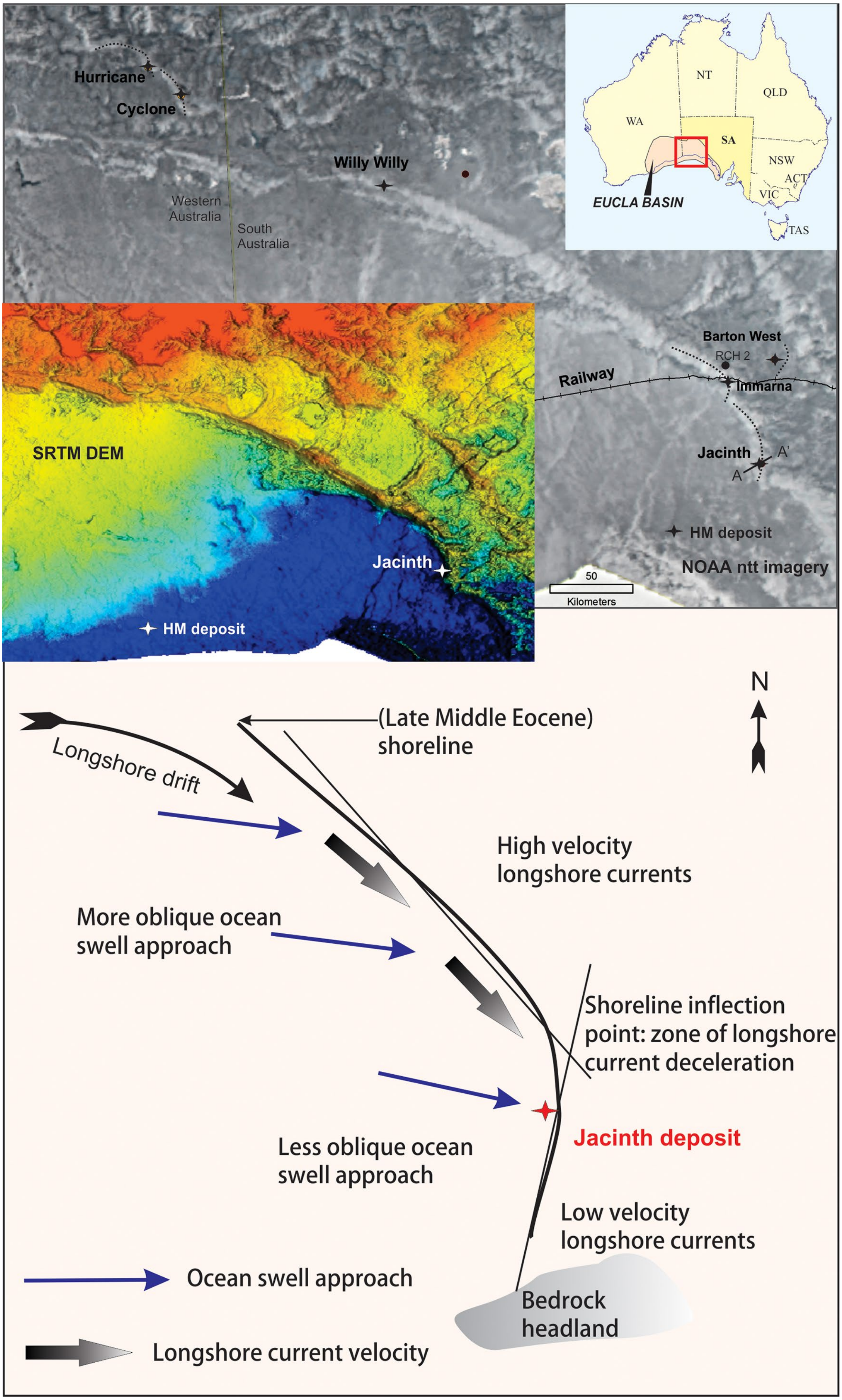


Figure 5.3 Schematic model of longshore current impact on the paleoshorelines, extracted from the case study of the Eucla Basin HMS. The longshore current velocity of

south of the inflection point is less than that of north of the inflection point due to relative differences in angle of oblique wave approach (Komar, 1979; from Hou et al., 2011b, 2021).

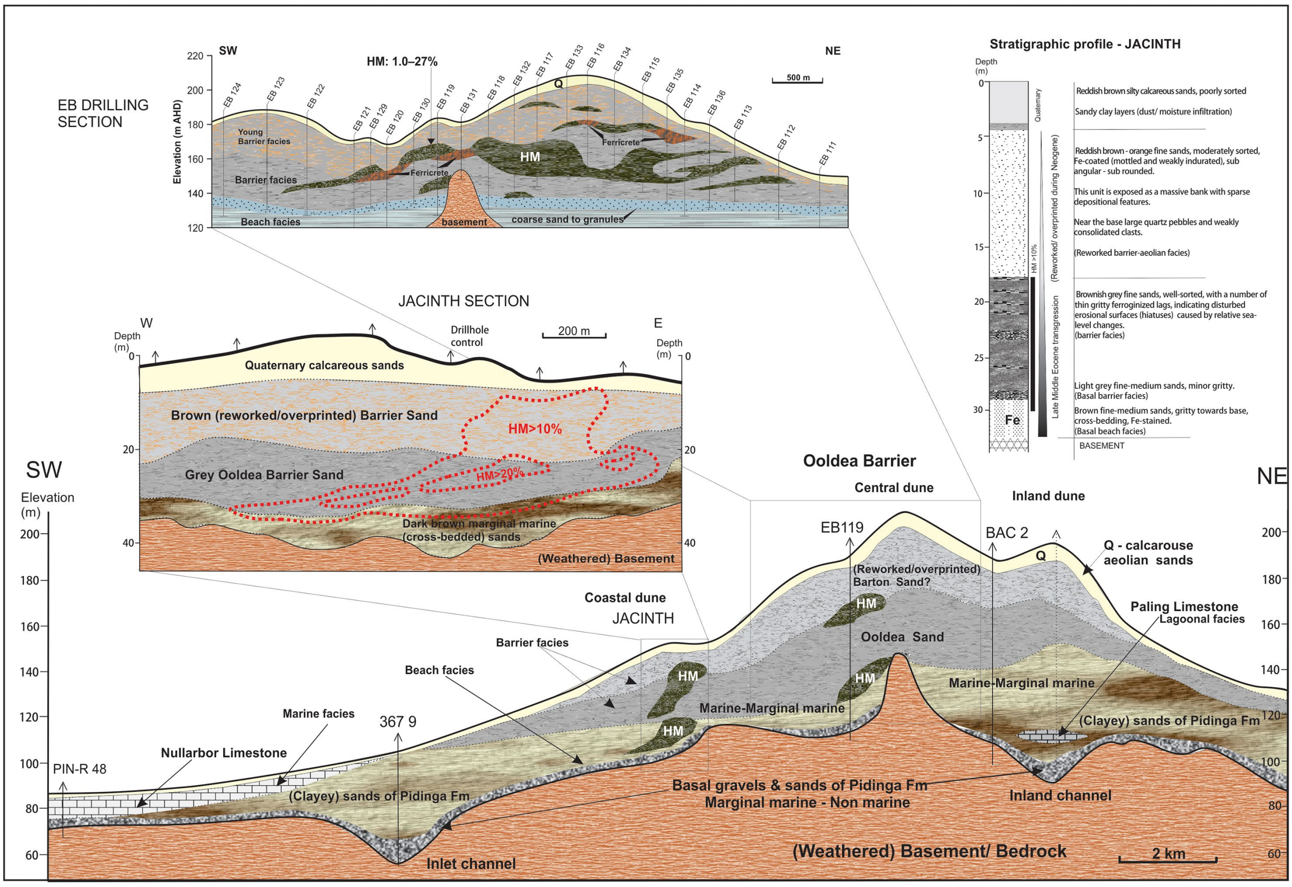


Figure 5.4 Geological cross sections through the Ooldea Barrier and mineralised zones from representative drillhole data, showing association of heavy mineral anomalies with their host barrier facies and topographic features. See Figure 5.3 for location of section A-A' (from Hou et al., 2011b).

LATE MIDDLE EOCENE SHORELINE FACIES

This paleoshoreline is highly prospective for late Middle Eocene beach placers (zones P14 and P15; Fig. 4.1). At present, the highest grade heavy mineral concentrations identified essentially follow the late Middle Eocene shoreline and correspond to the beach deposits of the Tortachilla Transgression (Hou et al., 2003b; 2011a, b). Marine transgression in the late Middle Eocene (eastern) coastal plain is marked by deposition of shelf, barrier-island, lagoonal, and locally flooding deltaic carbonaceous limestone sediments that in places blanket lower Pidinga Formation, Paling Formation and Ooldea Sands (Fig. 4.11b). The fine-grained, well-sorted sands of the shoreface and barrier island units contain a mature assemblage of HMs formed as water-lain beach sands or barrier-dune complex sands. Zircon, a very resistant and locally concentrated HM, is rounded, as are most of the other HMs. The HM assemblage is probably of multicyclic origin. In the eastern basin, HM accumulation is hosted mainly by highstand beach-barrier sands of the Ooldea Sand and occurs in zones ranging from tens to hundreds of metres wide and 2 to 20 m thick (Fig. 5.4). The Tortachilla transgressive (Ooldea) beach was characteristically reflective, displaying a steep beach face composed of fine to medium sand with sediment sourced from erosion of Musgrave cratonic rocks first transported down incised valleys and then by longshore drift toward the east (Fig. 4.11b). Multiple swash-aligned foredune ridges within the Ooldea Range provide evidence of prolonged accumulation of coastal sediments during the Tortachilla transgression (Clarke and Hou, 2000).

LATE EOCENE SHORELINE FACIES

The HM concentrations identified in the Late Eocene (Barton) barrier in the eastern basin are hosted by highstand beach-barrier sands of the Barton Sand and partly by Khasta Formation. These were deposited along the Late Eocene (Tuketja-Tuit) coast and on the top of the Ooldea barrier. As marine transgression progressed during the Late Eocene, reaching ~20 m higher than the earlier marine transgression (Clarke and Hou, 2000), deposition of the barrier island unit occurred at successively higher elevations. This transgression was followed by a rapid regression that resulted in erosion of a wave cut tidal plain (between the Pidinga and Khasta formations; Fig. 4.3; Hou et al., 2003c). The coast in the eastern basin was characterised by a series of coastal barrier island sands, oriented southwest-northeast (at an angle to the Ooldea barrier; Fig. 4.11c). The Ooldea barrier remained in the form of a series of offshore barrier islands. Paleorivers cutting through the Late Eocene (Barton) barrier island complex also contributed to placer deposition, particularly where local factors reduced stream velocity. The Tuketja-Tuit transgressive (Barton) beach was also typical of beach facies composed of fine-medium sand transported via incised valleys (Hou et al., 2003b). Protection afforded by offshore (Ooldea) islands in the Late Eocene coast created a wave energy gradient decreasing to the northeast in this coastal sector (Fig. 4.11c). This shoreline is also highly prospective for Late Eocene beach placers (zones P16 and P17; Fig. 4.1).

Upgrading of HM concentrations in some parts of the Ooldea and Barton barriers may have occurred during sea-level changes, by means of the transgressive barrier island and longshore current (littoral bypassing) fractionation model (Hou et al., 2011b). Identification of the western extension of the Ooldea barrier and later beach-barrier island complex facies in the northern and western margins of the basin is important and essential for advancing further exploration efforts to find HM deposits in these areas. High-resolution night-time thermal imagery (e.g., ASTER) and detailed elevation data combined with geologic and drillhole data and interpreted in a GIS environment have been shown to combine as an effective approach for evaluating Cenozoic geomorphology (Hou et al., 2011b; 2021). This method has allowed recognition of sequences of beach ridges interpreted as numerous strandlines <50 to 100 m apart within these coastal barrier island sands in the remarkably wide (up to 30 km) Ooldea and Barton barriers.

NEOGENE SHORELINE FACIES

The Miocene–Pliocene period of shoreline deposition is represented by reworked Eocene coastal sediments in the southeastern Eucla coastal plain. These sand units typically bear heavy minerals. They are fine-grained, moderately to well sorted, have a distinct enrichment in heavy mineral sands, and probably represent a period of beach, barrier, and/or dune complex deposition. Numerous low grade (>1%) heavy mineral deposits were discovered within the Middle Miocene – Early Pliocene sediments in the coastal plain (Ferris, 1994). Information from this poorly defined Mio–Pliocene coastal plain around the Kingoonya and Narlaby paleoriver mouths has been updated by the recent study of basin evolution, particularly the neotectonic impact on the HM-bearing strandlines (Hou et al., 2008, and 2011b). These Mio–Pliocene transgressive-regressive shorelines are very prospective for HM concentrations in the southeastern margin of the basin (Fig. 4.11d).

Two interdependent mechanisms are likely to produce localised concentrations of HMS in the southeastern Eucla basin during Mio–Pliocene time. The first involves transgressive-regressive barrier island fractionation processes operating in parts of the former Ooldea barrier, where old (Eocene) barrier island sands containing concentrations of HM were exposed to marine reworking while the Mio–Pliocene barriers were forming. The second involves longshore current fractionation processes operating in an environment where downwarping was active. Reworking older HM-bearing strandlines was the most likely source of HMS, as sediment supply from the flow in rivers draining the nearby Gawler Craton was reduced during Mio–Pliocene time. Continued exploration for HM deposits in the Mio–Pliocene coastal zones of the eastern basin is an indication that results are encouraging and that discovery of further HM-bearing shorelines beneath relatively thin cover is likely. In contrast, the uplift and the condition of beach-sediment starvation that apparently existed in the western margin of the basin in Late Miocene – Early Pliocene times would have been less favourable for the formation of HM placers.

HM DEPOSITS AND PROSPECTS

Weathering, erosion, paleoriver activity, sea level change and wind process, and tectonic movement over 40 million years have influenced the concentration of HMS along over 2,000 km arc of paleoshorelines in the Eucla Basin (Hou et al., 2010). Several cycles of marine transgression have left generations of beach placer concentrations buried beneath the present landscape. Barrier complexes, cumulatively more than 1,000 km in length were formed by the action of long-period marine waves and longshore drift. Concentration of HM sands in the Eucla Basin was influenced largely by changes in paleoclimate and geomorphology. Prevailing westerly winds built extensive dune systems by longshore drift with sediment movement from west to east. The dunes were reworked during highstand events in the late Middle Eocene and Late Eocene. J-shape bay heads along the 2,000 km arc of the coasts were ideal sites for HM concentration (Figs 5.3 and 5.4). They provided protected beaches to which HMs might be transported and concentrated via the combined interaction of westerly longshore drift and diffraction of waves around headland promontories. Dynamic regional tilting of the basin toward the northeast combined with longshore drift resulted in reworking of Eocene barriers, particularly along the eastern basin margin where large movements of sand were associated with extensive development of Neogene barriers.

Several HM deposits/prospects have been found in these favourable settings, which have been described in detail in Chapter 7 of Hou et al. (2021), including:

- Late Middle Eocene Shoreline System:
    - Jacinth, Ambrosia, Tripitaka, Typhoon and Atacama deposits
    - Willy Willy, Thar, Immarna, Sonoran and Mojave prospects
    - and a few other occurrences
- Late Eocene Shoreline System:
    - Barton, Barton West, Israelite Bay, and Cyclone deposits
    - Balladonia, Wanna and Gobi HM prospects, Esperance Mineral Field
- Neogene Shoreline System:
    - Gullivers deposit
    - Dromedary prospect
    - and a few other occurrences.

HM PROVENANCE

Sediments deposited in the paleoshorelines and adjacent paleovalleys of the Eucla Basin were largely derived from associated hinterland paleodrainages. Therefore, knowledge of where and what bedrocks the paleovalleys have incised and drained provides a useful guide to related mineral resources. Extensive coastal barriers of the eastern Eucla margin must have formed where large volumes of sand were supplied by the adjacent Eucla paleorivers or were reworked shorewards from the platform by marine processes (e.g., Benbow, 1990b). Knowledge of the source of the zircon and associated heavy minerals is a vital component in understanding the sedimentary architecture and transport processes that led to these deposits and is an important first order tool for prospectivity models in the region (Reid et al., 2013). In some settings, particularly in ancient orogenic belts, interpretation of the source region for detrital zircon data can be difficult due to the natural bias of the detrital zircon record by, for example, processes such as source region zircon fertility (Moecher and Samson, 2006) and sedimentary transport processes (Sircombe, 1999; Cawood et al., 2003). However, in cases where the sedimentological link between the source and the depositional sink can be established or inferred, analysis of detrital zircons can be integrated with the sedimentological framework to give a picture of the overall basin evolution (e.g., Morton and Hallsworth, 1999; Fedo et al., 2003). The data and information summarised here are mainly from the results of Reid and Hou (2006), Reid et al. (2013), Hou et al. (2011b; 2021) Gartmair et al. (2021) and Gartmair (2022).

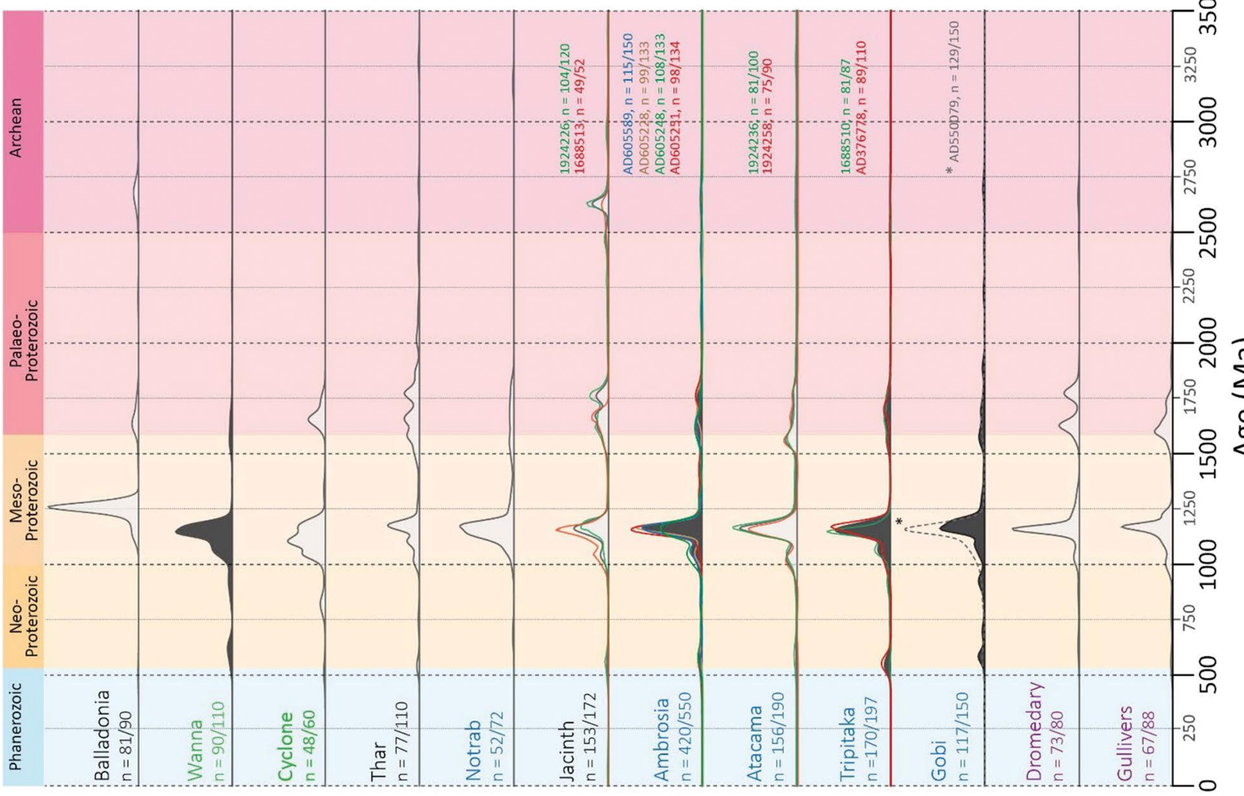
Most samples from the Eucla Basin paleoshoreline systems have statistical age populations within the Mesoproterozoic (Fig. 5.5). The likely source regions for zircons of this age are the Albany-Fraser Orogen, Madura, Coompana and Musgrave provinces (Fig. 5.6). There is broad similarity in the pattern of zircon provenance between deposits formed in different paleoshoreline complexes, such as the Middle–Late Eocene Jacinth deposit and the Late Eocene Thar prospect, which lie on the Ooldea and Barton barriers respectively (Fig. 5.5). The zircon provenance for the Miocene prospects such as Dromedary or Gullivers is likewise essentially identical to that for the Eocene deposits. Furthermore, this consistency in zircon source also extends more broadly to the Cambrian sediment analysed in Reid et al. (2013). The lack of major change in primary zircon source regions across basins in southern Australia implies there has been a consistent drainage and sediment supply through time, and also highlights the likely process of recycling of these zircons from one basin sediment to another. The influence of westerly longshore drift in sediment transport to the eastern Eucla Basin is strongly supported by the major sources of these Mesoproterozoic grains. In the case of most grains such as this, it is not possible to argue a main Gawler Craton source for the eastern HM deposits/prospects, since the influence of the Albany-Fraser Orogeny, Madura or Musgravian is absent from the bulk of the Gawler Craton (Reid et al., 2013).

Recent study (Gartmair, 2022; Gartmair et al., 2021) on the Eucla Basin HM compositional and detrital zircon U-Pb and Hf isotope data with novel integration of zircon grain shape analysis indicates that polymodal detrital zircon age spectra express a dominant Mesoproterozoic sediment provenance with secondary contributions from late Archean to the early Phanerozoic-aged sources, which is compared to the previous results in Figure 5.5. The results of Hf isotope data and distinct zircon shape characteristics of Mesoproterozoic-aged detrital zircon grains of northern and eastern prospects highlight a dominant juvenile magmatic source (northern lying Musgrave Province or underlying Madura and Coompana provinces), indicating the Madura and Coompana provinces as the dominant sediment sources for the northern and eastern Eucla Basin beach placers. Thus, Gartmair (2022) concludes that the sediment supply to the western Eucla Basin is restricted (Albany-Fraser Orogen and Yilgarn Craton), while northern and eastern basin margins evidence detritus derived from several Mesoproterozoic sources (Musgrave, Madura and Coompana provinces and Albany-Fraser Orogen) and secondary Archean sources (Yilgarn Craton and Gawler Craton) likely via intermediate sedimentary basins, which is compared to the previous results in Figure 5.6.

Models for the formation and reworking of paleostrandlines of the Eucla Basin have continued to evolve. Earlier models that identified dominant sediment input from large paleovalley networks that drained across the Precambrian Gawler Craton have been revised because the age distribution of detrital zircon from the heavy mineral deposits correlates with crystalline ages of the Musgrave Province, identifying the latter as an important primary source area for the barrier sands (Reid and Hou, 2006). Detrital zircon grains from the Ooldea and Barton barriers show a distribution of zircon age that is consistent with the Proterozoic Musgrave Province to the north of the basin as the dominant primary source area of the heavy minerals, with a contribution from the Albany-Fraser Orogen and Yilgarn Craton to the west (Reid et al., 2013). The likelihood is that these heavy minerals were recycled via sedimentary basins that flank the Musgrave, Madura and Coompana provinces (Fig. 5.6; Gartmair, 2022; Gartmair et al., 2021). The significance of eastward longshore drift in building beach barriers was emphasised. Sediment movement was from west to east. This is consistent with the Eocene climatic regime of dominantly westerly weather systems (e.g., Kemp, 1978) and the extensive development of lagoonal and estuarine facies at the terminal end of paleorivers draining from the Gawler Craton (Hou et al., 2011b).

Information provided on deformation of the Eucla Basin and HM provenance suggests that HM accumulations may be found farther to the west than the current batch of discoveries in the eastern Eucla basin because the dominant source of zircon was not in the immediate vicinity but was significantly to the north and/or west of the eastern HM discoveries (Fig. 5.6). Whereas the role of local control on sediment dynamics leading to localised HM concentrations in specific regions is still poorly understood owing to lack of detailed data, the predominant eastward longshore drift in the Eocene beach system makes knowledge of the cratonic provenance for the HM an important predictive exploration tool.

Zircon age spectra from Gartmair (2022). Deposit names in bold with grey fill spectra represent newly dated samples of Gartmair, (2022). Data from the prospects Balladonia, Cyclone, Thar, Jacinth (sample 1924226), Atacama, Tripitaka (sample 1688510), Dromedary and Gullivers (Reid et al., 2013a), Jacinth (sample 1688513; Hou et al., 2011b), and Notrab (Reid & Hou, 2006) are presented with white fill colour. Prospects with several samples (coloured age spectra lines) are combined to a single age spectrum (black age spectra lines).



Zircon age data (from Reid and Hou, 2006; Hou et al., 2011b; Reid et al., 2013).

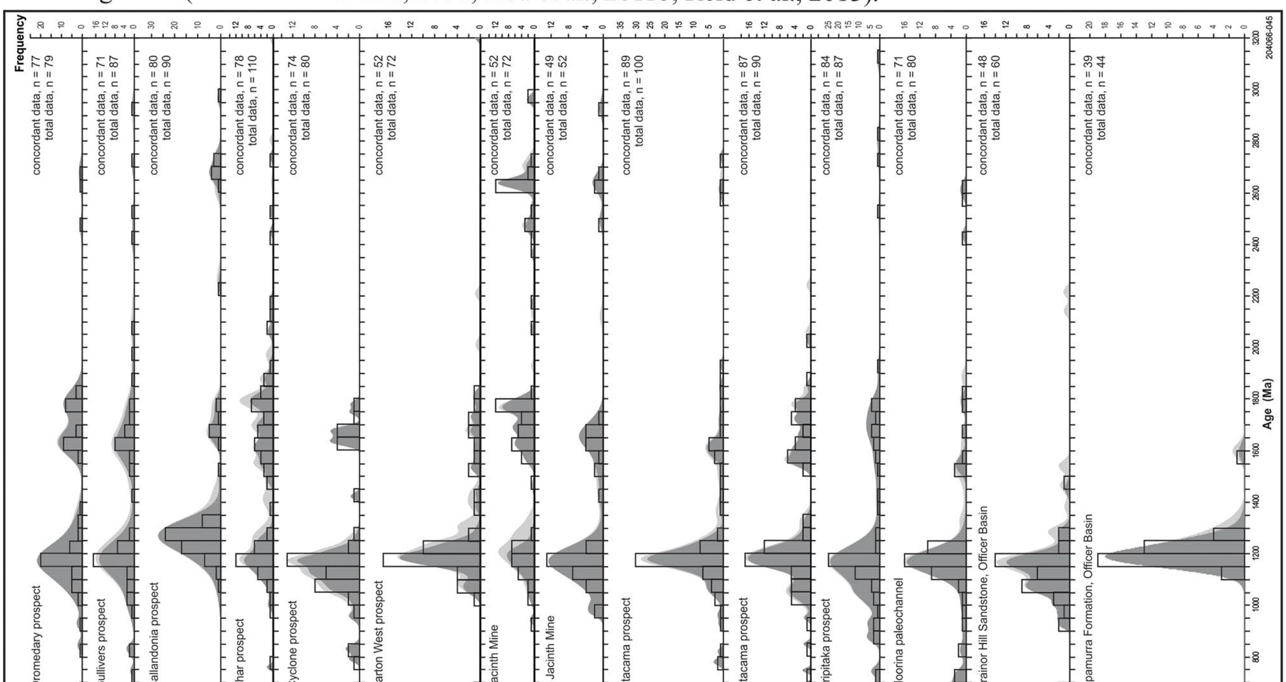


Figure 5.5 Zircon age spectra of the Eucla Basin HMS deposits (from Reid and Hou, 2006; Hou et al., 2011b; Reid et al., 2013; Gartmair, 2022).

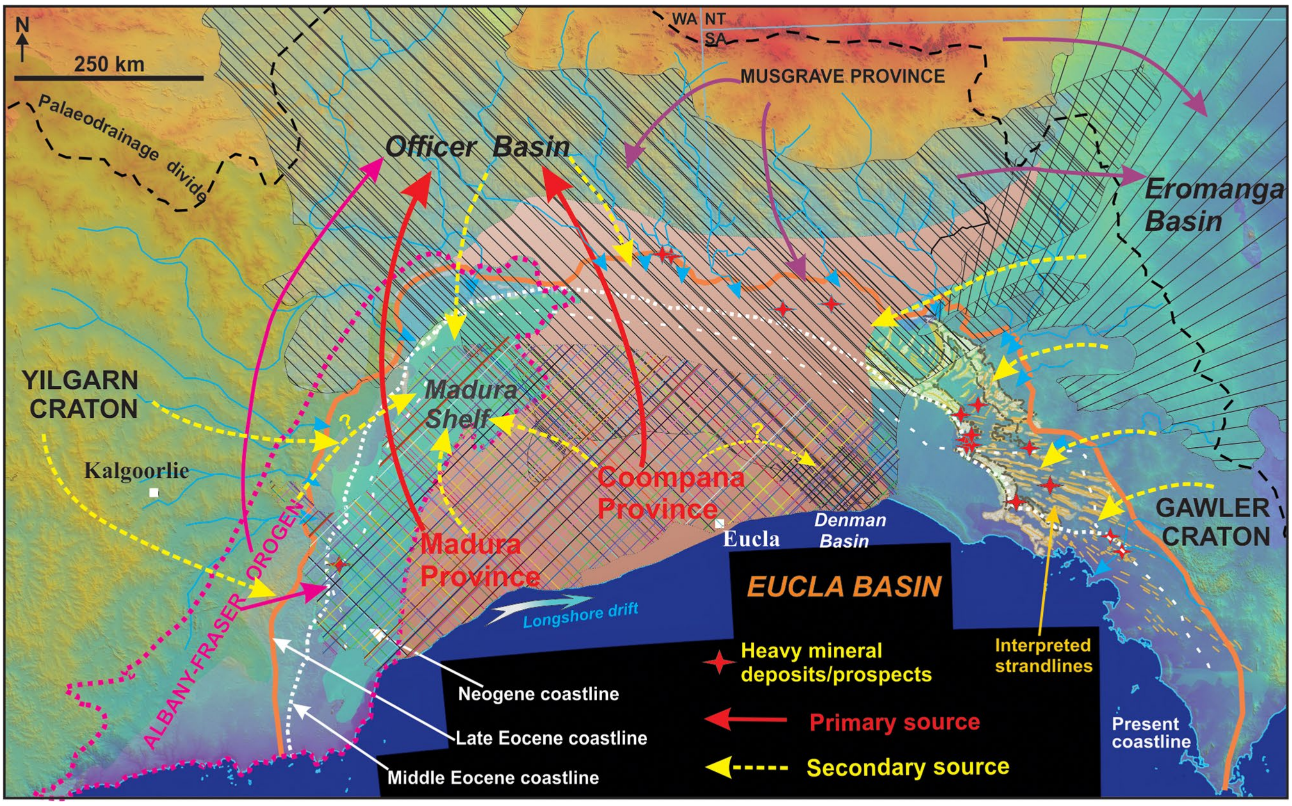


Figure 5.6 Interpretation of HM sources showing relationship of Eucla basin and paleovalley network and HM deposits and prospects to the bedrock source areas for the sediments (after Hou et al.,2011b; Gartmair, 2022).

The evolution of paleoshorelines in the Eucla Basin is characterised by a landward stepwise migration of the coastline. The paleoshorelines, formed during third-order sea-level rises, contain beach placers, mostly formed at sites where there was supply of HMS and conditions for concentration and preservation favourable for HM deposits. In aggregate, the accumulation of barrier island sands is up to 30 km wide and their complexity conceals numerous individual strandlines, deposited during high order events of sea-level rise that are mostly undocumented and remain highly prospective for HMS. Around the peak of transgression and highstand, HM deposits formed in beach or barrier island and dune sand complexes. Lengthy stillstands were likely associated with late Middle Eocene deposition as this shoreline includes some of the largest and richest deposits. The HM deposits are most likely to have arisen from multicyclic origins.

Apart from the relatively well-known Ooldea and Barton barriers, other shorelines are prospective for hosting HM deposits. These include the middle Eocene Hampton Sandstone in the central part of the Eucla Basin and Neogene strandlines in the eastern basin. Reconstructing early shorelines of the Middle Eocene Wilson Bluff shoreline in the central basin is hampered by marine limestone cover and limited drillhole data. Similarly, tracing the extent of Neogene shorelines, especially in the more prospective areas near paleoriver outlets and across older coastal barrier islands, is difficult due to extensive surficial cover of younger aeolian dunes. Further research to delineate these systems may provide additional heavy mineral targets.

Eucla Basin findings have significance for worldwide exploration for older (Eocene) HM deposits, in that concealed deposits of the type described here may exist in areas that have received little or no previous exploration, in particular where they lie adjacent to Precambrian shield areas. Given the example here, such deposits may be located over 120 km (Jacinth-Ambrosia deposits) and even 320 km (Cyclone) inland from modern shorelines. Major points arising from the present study are as follows:

1. The Eucla basin has potential to become a major HM province in Australia, and probably a world-class zircon source. At the time of writing, over a dozen separate HM prospects have

been reported from paleobeach and barrier island sands in the Eucla Basin, with mining at the Jacinth deposit commissioned in 2009.

2. The deposits are in mostly unconsolidated sands above the water table and generally have a favourable ore thickness - to - overburden ratio. Consequently, mining and processing appear relatively feasible. The remote location and general lack of infrastructure mean that large or high-grade deposits with a high content of the more valuable heavy minerals are more likely to be developed in the near future, as driven by economic conditions.
3. Recent stratigraphic investigations and new data from mineral exploration suggest that more than four generations of Cenozoic shorelines comprise many HM-bearing strandlines.
4. The regional distribution of the HM-bearing strata and the shoreline evolution is consistent with landward terrace migration of the shoreface complexes from the central basin to the onshore margin by over 300 km as a result of sea-level change. The predominance of zircon in high-grade HM deposits mainly originating from the Musgrave, Madura and Coompana provinces enhances the prospectivity of the northern and eastern margins of the Eucla Basin.
5. The valuable HM assemblage consists mostly of zircon and altered ilmenite with lesser proportions of rutile and leucoxene. Zircon is the dominant HM in parts of the eastern basin, while ilmenite predominates in prospects along the northern and western basin. The highest concentration of zircon (up to 65% of the HM fraction) is found along the Ooldea barrier in the eastern Eucla margin.
6. Similar to the situation in the southwestern margin of the Murray Basin, Quaternary strandlines may exist in the area of the Nundroo and Bookabie ranges that are located close to the present coastline and were largely formed during Pleistocene time. Both the Nundroo Range and Bookabie Range have similar geographical characteristics to the Ooldea Range. The Pleistocene dune sands of the Nundroo and Bookabie ranges could also possibly host Pleistocene HMS, but the heavy minerals would have been mainly sourced from the sea, rather than inland through paleovalleys, because no such age paleovalleys are known in the central Eucla Basin.
7. The Hampton Sandstone exists in the deeper part of the basin, which is known to host HMS as well. Early Eocene paleochannels have been confirmed to occur beneath the Wilson Bluff Limestone between the central Eucla Basin and the Ooldea Range (Hou and Keeling, 2008; Hou et al., 2008), but it is unknown how far these paleochannels extended from the Ooldea Range (towards the Nundroo and Bookabie Ranges), due to limited data. Presuming the Early Eocene paleochannels were continuing past the Ooldea Range and into the Nundroo and Bookabie Range regions, it would be hopeful to explore for Early Eocene mineral sands, under both the Nundroo Range and Bookabie Range.

5.3 URANIUM

The Eucla Basin and its peripheral paleovalleys host numerous uranium prospects (Hou et al., 2014). Known uranium occurrences are presently confined to the western and eastern margins of the Eucla Basin where crystalline basement rocks are potential source areas for uranium (Fig. 5.1). Exploration for these paleovalley-hosted uranium commenced in the mid-1970s, initially focusing on the eastern Eucla paleodrainage, specifically the Kingoonya, Narlaby and Yaninee paleovalleys (Binks and Hooper, 1984; Curtis et al., 1990; Flint and Rankin, 1991). This resulted in discovery of the Warrior prospect (discovered in the 1970s in Warrior Paleovalley), and Ealbara prospect (discovered in the early 1980s), both situated in tributaries of the Kingoonya paleovalley, and four Yarranna prospects (up to 3550 ppm U_3O_8) in the Narlaby Paleovalley and Yaninee prospect in the Yaninee Paleovalley.

Uranium present in the eastern Eucla paleovalleys may have been sourced from Hiltaba Suite granites that are widespread on the Gawler Craton, variably weathered, and typically contain uranium contents 2–3 times background levels. The eastern Eucla paleodrainages actively incised weathered basement during Early to Middle Eocene times with flow rates decreasing over time such that sandy valley fill and lignitic clays gave way, during Miocene time, to largely chemical carbonate sedimentation in a series of discontinuous lakes strung out along the major drainage lines (e.g., Hou et al., 2007b). Anomalous levels of uranium have been reported also in other paleodrainage networks, including the Woldra paleovalley and Lake Bring estuary (16 to 643 ppm

U_3O_8; Curtis et al., 1990) and the Thurlga paleovalley (8 m at 106 ppm U_3O_8; Adelaide Resources, 2009).

Paleovalleys along the western margin of the Eucla Basin drain potential uranium source rocks of the Yilgarn Craton. Uranium discoveries within this drainage network include the Mulga Rock and Ponton uranium deposits where extensive drilling has been undertaken in recent years (Penney, 2012). These uranium deposits lie within the Raeside Paleovalley system, which was incised in deeply weathered crystalline rocks of the Yilgarn Craton and Albany-Fraser province, and can be traced for more than 100 km across the western Eucla margin and eastern Yilgarn Craton. The Mulga Rock uranium deposits are characterised by numerous mineralised zones within interbedded sand and lignite, which are hosted by middle Eocene sediments within a sequence of up to 100 m of Paleogene fluvial, estuarine and lacustrine sediments (Hou et al., 2014). These overlie preserved remnants of Cretaceous and Permian sediments (Macfarlane and Inwood, 2010). The Mulga Rock deposits, comprising three separate zones of mineralisation, Shogun, Emperor and Ambassador, are situated in the western margin of the Eucla Basin, approximately 250 km east-northeast of Kalgoorlie in Western Australia (Fig. 5.1). Uranium mineralisation occurs in lignite and in clay and sandstone immediately adjacent to the lignite of the Mulga Rock uranium deposits (Inwood, 2009; Macfarlane and Inwood, 2010). The deposits are distributed along the outer margin of a broad bend in a paleovalley (Inwood, 2009). The uranium mineralisation is hosted by lignites, carbonaceous clays and sands. Highest grades, average 0.06% U_3O_8, are in lignitic clays immediately below the redox boundary at the base of the weathered zone (Macfarlane and Inwood, 2010). The mineralised zones, averaging about 2 m thick, are flat-lying and are from 20 to 50 m below surface. The redox boundary (commonly >30 m deep) presents as a sharp front between kaolinitic paludal clay and lignite and commonly approximates the present-day water-table (Douglas et al., 2003). The channel sediments were deposited during middle Eocene time and are composed of three broad units: (1) basal fluvial sands and gravels; (2) lacustrine to paludal sediments, including lignite, clay-rich lignite and carbonaceous sands; and (3) fluvial sands and interbedded lacustrine sediments, which are generally oxidised, ferruginised and silicified (Douglas et al., 2003). Uranium, sourced from granitoids and metamorphics of the Yilgarn and Albany-Fraser Provinces, was transported by groundwater flowing in the paleovalley sediments and adsorbed onto the organic matter within the lignite/carbonaceous clay (Fulwood and Barwick,1990). The adsorbed uranium was remobilised by oxidising groundwater and subsequently re-adsorbed onto lignitic layers near the base of the oxidised zone. This had the effect of upgrading the initial low concentration of uranium. The grade of mineralisation and thickness are controlled by permeability and organic matter-content of the host sediments, with the highest grades and thickest zones of mineralisation developed within the more organic-rich and more permeable sediments (Macfarlane and Inwood, 2010). Oxidised sands and silts above the redox boundary are depleted in uranium relative to daughter products, whereas reduced sediments immediately below the redox boundary are enriched in uranium relative to daughter products (Fulwood and Barwick, 1990). Three major styles of uranium mineralisation are recognised at the Ambassador deposit, Mulga Rock (Macfarlane and Inwood, 2010): (1) Upper Lignite - enriched in U, REE, Ni, Co, Sc, V and locally Cu and Ag; (2) Lower Lignite and associated sandstone; also enriched in base metals; and (3) a basal sandstone hosted mineralisation. Macfarlane and Inwood (2010) considered that the metals probably migrated upwards along fault zones, through the sandstone and lignite, finally being trapped by an impervious layer lying directly above the lignite.

Surficial uranium deposits have been discovered mainly in the western Eucla paleovalleys (Fig. 5.1). This type of uranium deposit is broadly defined as young, near-surface uranium concentrations that overlie and are adjacent to weathered Archean granite and greenstone basement rocks, which provide a source of soluble uranium and vanadium necessary to form carnotite (Mann and Deutscher, 1978). The calcrete formed in the paleovalley sediments is relatively young, post Pleistocene (Mann and Horowitz, 1979) and reflects the change to more arid conditions. Carbonate was precipitated from groundwater usually in the central and lower regions of the paleovalley where the watertable was close to the surface and evaporation and evapotranspiration rates were sufficiently high to concentrate calcium and magnesium ions. According to Mann and Horowitz (1979), carbonate deposits formed initially within the zone of saturated sediments. Continued deposition and crystallisation of carbonate caused mounding of older carbonate, which displaced overlying sediments and was pushed upwards above the watertable where the carbonate was recrystallised and hardened. Uranium in groundwater, as uranyl carbonate complexes, combined with vanadium and potassium to precipitate carnotite in areas of valley calcretes either through groundwater mixing or changed chemical conditions resulting in the solubility product of carnotite being exceeded (Gaskin et al., 1981). A typical model for the origin and formation of surficial calcreted hosted uranium deposits is summarised based on common key features identified in Australian deposits (e.g., Yeelirrie and Lake Way; Fig. 5.1). The uranium sourced from uraniferous granites in the region is transported in mildly oxidizing saline alkaline groundwater solutions in constricted drainages to semi-closed basins with variable evaporative conditions. General characteristic of calcrete-hosted deposits, with reference to Bowell et al (2009) and McGeough, pers comm. (Toro Energy Ltd., 2009) include:

- significant areas of uranium enriched bedrock, which are obvious in airborne radiometric imagery at both regional and continental scales, generally derived from Archean and Proterozoic weathered granites.
- mostly Cenozoic paleovalley systems drained these uranium-enriched bedrock areas and discharged into lakes, as well as hosting channel calcrete formations.
- a periodic wet and dry climatic region whereby there were long periods of wet developing drainage systems and dry allowing evaporation (e.g., Wiluna of WA).
- carnotite mineralisation in calcrete usually increases in the 'deltaic' segments of fluvial discharge into broader lake systems adjacent to paleovalleys.
- almost all deposits show remarkably similar lithologies, i.e., deeply weathered zone of saprolitic clay overlying a granite basement, and/or alluvial clay overlying highly weathered basement with anomalous vanadium content.
- uranium host materials formed in low-energy conditions dominated by calcrete and silcrete (e.g., approximately 25% of the carnotite mineralisation of the Wiluna and Napperby deposits of Australia is hosted in calcrete/silcrete layers), fine sand/silt and clay (e.g., approximately 70% of the carnotite mineralisation in these Australian deposits is hosted in silt and clay).
- the carnotite mineralisation is associated with secondary deposition of calcite, dolomite, gypsum, salt, celestine and barite, indicative of evaporative brine concentration.
- as a common feature, the calcrete uranium mineralisation is associated with highly saline water (1–3 times sea water salinity), even though the salinity of both upstream and downstream groundwaters is significantly lower.
- carnotite mineralisation occurs in fractures and vughs within a calcrete layer (often silicified or dolomitic) and as disseminated coatings on quartz grains in fine sands and on clay.
- the mineralisation depth is generally shallow, occurring between the surface and 15 m (at Yeelirrie, mineralisation extends down to 32 m below surface).
- vanadium, mainly sourced from greenstone and/or granite is necessary in forming carnotite mineralisation (e.g., Wiluna, Napperby and Yeelirrie deposits), although its origin is often debated.

5.4 OTHER RESOURCES

Besides placers and sedimentary uranium, other resources were also of interest in previous exploration activities in the Eucla Basin and peripheral paleovalleys, such as limestone, silica sands, clay minerals, coal, groundwater and petroleum (e.g., Lorry, 1970; Benbow et al., 1995a).

Limestone

The limestone resources of the Eucla Basin are mainly from Wilson Bluff Limestone and Nullarbor Limestone. Limestone from near Naretha has been used for the production of calcium hydroxide for use in gold extraction. Production began in about 1929 and reached 100 tons per week, but it later dwindled, and ceased in 1966. Surface slabs of Nullarbor Limestone were hand-picked and burnt with myall, a local timber with a low ash content. The resulting product had a high purity, reputed to reach 98.8 per cent. Recrystallised Nullarbor Limestone, despite its close physical resemblance to the original calcarenite, was rejected as unsuitable because it tended to disintegrate during burning and choke the fire. Large-scale mining of limestone from the Eucla Basin is a possibility, and the few scattered chemical analyses reported indicate that large quantities of limestone are available with about 97% calcium carbonate, and 1 or 2% magnesium carbonate. The Trans-Australian Railway could be used for transport but a quarry situated near the railway would encounter seams of clay and kankar in the top 15 m of limestone and this could lower the grade. A coastal quarry with sea transport, for example in the Abrakurrie Limestone at Twilight Cove, would probably obtain purer limestone, but the coast is exposed to storms and heavy swell from the south, so that the prospects seem poor for regularly loading ships anywhere along the coast of the Great Australian Bight.

Silica sand

The coastal dunes in the southwest of the basin might be a source of silica sand. Visual inspection of grab samples indicates that there is always at least a small proportion of calcium carbonate grains, but higher grade sand might be found through detailed prospecting.

Clay minerals

Palygorskite has been identified within the palygorskitic and dolomitic clays of the Garford Formation in the middle and upper reaches of the Garford Paleovalley (e.g., Pitt et al., 1978), and is likely to be also present within equivalent sediments of the Tallaringa Paleovalley system. Following up related exploration carried out by Steetley Industries Limited during 1983–1984, numerous aspects of the deposits with some economic potential were investigated in the Garford Formation in the Garford Paleovalley (Robertson, 1984, 1988; Warne, 1987; Churchman et al., 1994; Raven et al., 1995; Keeling et al., 1995; Self et al., 1996). Highest concentrations of palygorskite are generally confined to the zone of transition from illite/smectite clay to dolomitic clay, where palygorskite constitutes >50% (the best result 70–90%, ~1 m thick; Raven et al., 1995) of the sediment but is mostly less than 2 m in thickness (Keeling et al., 1995). The potential for palygorskite clay in other Eucla paleovalleys has not been evaluated or explored.

Coal

It was noted that the Eucla Basin sequence contains lignite deposits within lignitic facies strata. Lignite occurrences recorded in the eastern basin occur within the Eocene Pidinga formation of the marginal Eucla Basin and paleovalleys. Review of exploration for lignite in the eastern Eucla Basin is provided in Rankin et al. (1996). In the basin marginal area, lignites appear to be widespread but lenticular and thin (0.5–7m). In the low-grade lignites of the Pidinga Formation, ash content is generally well above 30% and the moisture content ranges between 12% and 60% (Pitt et al., 1978).

Drilling data obtained since the 1970s by companies, MESA and ETSA indicate that vertical and lateral changes in both coal thickness and quality are rapid, and in particular in the gradation of carbonaceous clay to clayey lignite or lignite. This, together with partial removal of a few metres of Pidinga Formation by tidal/wave events, make seam correlation and coal-value calculations difficult. Although the deposits are at shallow depth, with only 10–15 m overburden, lignite seams (including clayey or sandy lignite) are scattered throughout (carbonaceous) clay and sand units up to 40 m thick. The lithofacies analyses and correlations seem to suggest that most of the Wilkinson-Anthony estuarine plains may be prospective within fluvial and estuarine geometries tracing the highstand condensed intervals. But results of drilling to date, together with the high variability of the fluvial-related facies suggest the lignite deposits in the area are presently negligible and uneconomic. This possibly could change with improved quality analyses, with further drilling, and with development of future technology enabling the burning of high-ash coals using fluidised bed techniques.

Groundwater

Water supply is crucial to any mineral development and local residency, but particularly in the Eucla Basin area because of the lack of surface water resources due to the low and unreliable rainfall and high evaporation rate. Groundwater is therefore the only viable source for mineral processing and human consumption, and it is vital to assess resources.

Beneath the Eucla Basin, groundwater has been obtained from sandstones (e.g., the Madura Formation and the Loongana Sandstone) and the weathered surface of Precambrian basement (Lowry, 1970). The upper part of the Madura Formation in most places consists of impermeable siltstone and claystone, and although some sandstones are present they are almost always too clayey and carbonaceous to yield adequate supplies of groundwater. The impermeable beds of the upper part of the formation confine the groundwater in the lower part, giving sub-artesian supplies beneath the Bunda Plateau and an artesian flow from a bore on the low-lying Roe Plains.

The salinity, depth, and yield of supplies from the sedimentary beds of the Eucla Basin are largely controlled by conditions of local intake, the permeability of the sediments, and the distribution of the sandstone, which vary from one area to another. The Hampton Sandstone, which proved to be an aquifer in the Transcontinental Railway No. 1 Bore (Lowry, 1970), is developed in some parts of the western Eucla Basin and appears to have yielded an adequate stock supply in places (e.g., bore Kanandah No. 92) where the formation is probably thicken. However, water in the Hampton Sandstone is likely to be highly saline (Lowry, 1970).

The extensive network of Cenozoic paleovalleys represents a potential source of saline groundwater suitable for mineral processing (Martin, 1998; Martin et al., 1998). The limited previous investigations indicate that aquifers are localised and of highly variable quality and quantity (e.g., Read, 1987; Weber, 1980). Potential aquifers with both high quality and quantity probably exist locally in the Eocene Pidinga formation, as these formations generally show vertical and lateral rapid changes in lithology, particularly in the channel areas. Potential sources from the paleovalleys and inland paleoswamps are still targeted by recent investigations. Although drilled for stratigraphy (Pitt et al., 1978) and palygorskite (Keeling et al., 1995), for instance, the deeper channel sand sequences have not been fully and broadly tested for groundwater prospects, nor have the more prospective parts of channels, not affected by marine sedimentation.

Available drilling has indicated a 10–15 m thick aquifer containing up to 300,000 ML of saline groundwater in storage in the Garford paleovalleys and also a considerable thickness of sand with ~900,000 ML of water in storage (Russell, 1998). Other paleovalleys (e.g., Anthony, Kingoonya, Narlaby) in the eastern onshore basin show similar features, but no estimates of storage have been made so far. Aquifer tests need to be carried out to determine potential yields and drawdowns from pumping wells in the paleovalley area, including a number of tidal inlet channels. Other paleotopographic lows (including major tributaries and paleoswamps) can be investigated on a lower priority basis due to their less well-known features. Since the controls that determine the occurrence of groundwater in the area are not clear, it is expected that further geophysical and drilling investigations would help to clarify the pattern of groundwater and assist in mapping paleovalleys.

Water in Quaternary beds may be useful in places (Lorry, 1970). Eyre (1845) obtained potable water from coastal sand dunes near the site of the now-abandoned Eyre telegraph station. At the present time, potable water can be obtained from dunes at Israelite Bay and Twilight Cove, and limited supplies are obtained for stock from dunes on the coast south of Mundrabilla Station and at Eucla. It is possible that supplies could be obtained from other dunes, for example from the large dune masses near Point Dover (Bilbunya Dunes) and Eyre (Wurrengoodyea Hills). Gibson (1909) recorded a spring of potable water at the foot of the cliffs at Twilight Cove, and he and other writers have interpreted this as indicating that supplies of potable water could be obtained from the Tertiary limestones. However, the spring is probably fed from the nearby sand dunes that separate the cliffs from the coast at that point. Wind blowing along the cliff face appears to have scoured a channel in the sand almost down to sea level. At times water seeping out of the dunes forms a shallow pool, and at other times water can be obtained by digging (Gibson, 1909).

Petroleum

The prospects of finding reserves of petroleum in the Eucla Basin are poor, probably due to the lack of proper source beds. No signs of significant hydrocarbons have been reported in the known stratigraphic drillholes or numerous water bores in the basin. Although it is conceivable that the basal clayey part of the Wilson Bluff Limestone is also sufficiently impermeable to trap oil in the underlying Hampton Sandstone, closed anticlinal structures are absent (Lowry, 1970). Stratigraphic traps are possible where there are abutments of sandstone near the base of the Hampton Sandstone against the basement highs. It is quite possible, however, that the sandstones are connected by a continuous basal sandstone or by a surface zone of permeable weathered basement. If so, the sandstone has probably been flushed by the flow of confined groundwater.

The sediment-filled paleovalleys suggested by the interpretation of paleovalleys, such as the seismic surveys, may contain a sequence not yet encountered by drilling, and future efforts in petroleum exploration might be best spent trying to trace this valley onshore and offshore and drilling it to examine its stratigraphy and the possibility of stratigraphic traps being developed against its walls.

6. CONCLUSIONS

The study of the Eucla Basin now considers all aspects of the basin and associated paleovalleys, particularly the basin signatures and economic significance, and how they are formed. Various geoscientific information, models and exploration history are presented, and summarising figures, tables, and useful reference lists, are included. The two main objectives of this study are, firstly to develop a better understanding of the characteristics, geometry and geological/depositional environment of the Eucla Basin, but particularly on mineralised sediments and mineral deposits found in the certain depositional environments; and secondly to facilitate prospectivity analysis of the basin and its peripheral paleovalleys by reconstructing prospective paleoshorelines/paleovalleys and mapping paleoshorelines and paleovalleys, and to develop geoscientifically and technically efficient procedures for mineral exploration through a comprehensive understanding of the geological processes in the Eucla Basin and associated paleovalleys. This research has led to a better understanding of the history and development of the Eucla Basin and peripheral paleovalleys.

- This is a review report of the Eucla Basin and peripheral paleovalleys through the studies of two decades. These studies have assisted exploration in the basin and paleovalleys and have provided fundamental data for increasing our knowledge of geological processes and landscape evolution within the basin and paleovalleys.
- Studies of the Eucla Basin and peripheral paleovalleys have investigated the geophysical and geological expressions of the basin and adjacent paleovalleys together with the resources such as HM sands and uranium formed in the paleocoastal and paleochannel environments.
- The Eucla Basin, together with its peripheral paleovalleys, is one the largest Cenozoic sedimentary basins in the world. Evidence from sedimentology is combined with that of other geological and geophysical characteristics to arrive at a general reconstruction of basinal and paleovalley architectures and depositional environments.
- The paleodrainage in the onshore Eucla Basin represents all of the drainage that was active during Paleogene–Neogene periods because of wet or humid paleoclimate, with episodes of reactivation and modification recognisable.
- Sedimentation in the Eucla Basin took place in environments ranging from fluvial through estuary and marginal marine to shallow marine during Paleogene–Neogene marine transgressions and/or regressions.
- Evolution of the paleoshorelines of the Eucla Basin is characterised by a landward stepwise migration of the coastline. These paleoshorelines formed during several third-order sea level rises contain beach placers.
- Post-Eocene uplift in the western Eucla Basin led to a tectonic tilting influence in the basin evolution and Neogene strandlines being developed in the southeastern basin from the Middle Miocene onwards.
- Detailed study of the Eucla Basin and peripheral paleovalleys which drained cratonic provinces can provide important information useful in the exploration for HMS and uranium resources. The paleovalleys were originally incised into the pre-Paleogene landscapes, consisting mostly of weathered basement and bedrocks. They transported the HM/U-bearing

sediments to the paleocoasts through paleovalleys, where the heavy minerals and/or uranium concentrated mostly during the Paleogene and Neogene in the Eucla Basin and paleovalleys.

- After being transported by the paleorivers to the paleocoasts, the HM/U-bearing clastics were deposited in a variety of environments, including alluvial/fluvial channels, beach barriers (e.g., bay barrier, prograded barrier, and barrier Island), dunes (e.g., dune and spits), estuaries, deltas (e.g., flood delta, ebb delta, estuarine delta, and sandy delta), embayments and lagoons (including lagoon and embayment), tidal flats, cliffs and bluffs (e.g., cliff and headland) and offshore basin.
- The Eucla Basin is an important province of HMS in Australia and worldwide.
- The basis for the geologic model for placers and uranium deposition in the Eucla Basin is the integration of paleogeography (particularly of paleoshorelines and paleovalleys) with regional and global evidence for paleoclimate and relative sea-level change and major eustatic events, together with the identification of HM and uranium source and dispersion, and any tectonic activity, particularly during the time of deposition of the HM/U-bearing sediments.
- Knowledge of the basin and paleovalley architecture and of any sources of minerals deposited in the basin and paleovalleys is also of interest as guides to the location of potential deposits in the greenfield exploration areas.
- The Eucla Basin and peripheral paleovalleys are highly prospective for mineral resources, such as placers and uranium deposits as well as groundwater resources. Due to the buried features, some areas of the basin and peripheral paleovalleys are relatively unexplored and thus there is significant potential for further economic mineralisation.
- Complex paleogeography of the paleoshorelines and paleochannels determine the sites of placers and uranium accumulation. For example, the high specific gravity of placers tends to concentrate HM components in lag and transgressive/regressive barriers/dunes or J-shape bays, particularly during storms when lighter components, such as quartz, are carried downstream, offshore or along shore by strong littoral drift; and uranium tends to precipitate at redox front in the channel environments.
- In the Eucla Basin, in aggregate, the accumulation of Paleogene–Neogene channel and costal sands complexity conceals numerous individual strandlines and channel lags, many of which are mostly undocumented and remain highly prospective for placer deposits.
- The predominance of zircon in high-grade HM deposits (e.g., Jacinth) in the eastern Eucla Basin, originating mainly from the Madura, Coompana and Musgrave provinces provinces via Officer Basin, enhances the prospectivity of the northern and northeastern margins of the Eucla Basin.
- Any trace of valuable minerals in the paleoshorelines (e.g., beach placers) and paleochannels (e.g., placer gold and diamond) is of interest as a guide to the location of basement/bedrock mineralisation.
- Exploration for primary bedrock/basement mineralisation in the regions of the Eucla Basin and peripheral paleovalleys, e.g., beneath and surrounding the basin, can benefit from knowing the location and thickness of undesirable basin and channel sediments in a study area.

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9. APPENDIX

DRILLHOLE PALYNOLOGY

Eucla Basin drillholes palynology (XLSX 43 KB)