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Environmental character and history of the Lake Eyre Basin, one seventh of the Australian continent

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One of the world's largest internally drained (endorheic) basins, the 1.14 million km2 Lake Eyre (hydrological) Basin (LEB), covers nearly 15% of the Australian continent. Palaeoclimatic and associated flow regime variations are recorded in an archive of fluvial, aeolian and lacustrine sediments that are particularly accessible for deposits from the past ~ 300 ka, and are especially relevant for the past ~ 50 ka during which humans have inhabited Australia. Due to its great size, economic resources and diverse latitudinal extent, it has for over six decades been the focus of environmental, scientific and resource-based studies across numerous disciplines. With an emphasis on assessing the scientific and environmental research over this period, this paper is divided into four parts: Part A provides an introduction and background. Part B provides the geological history, including evidence of substantial Tertiary and Quaternary climate change. Part C covers the contemporary environmental conditions, and Part D presents a conclusion and summary. This not only represents the first comprehensive review of current knowledge of any of the world's truly large dryland drainage basins, but also highlights how more multi-disciplinary research is required. Key remaining questions revolve around the impact of global weather systems on the LEB; the lake-level story in relation to a more precise picture of Late Quaternary climate change; the role of humans and climate in the demise of the megafauna; modern hydrological changes; the role of vegetation in altering channel morphology and flow efficiencies; and nature of flood transmissions.

Keywords

Lake Eyre Basin, Geology, Hydrology, Geomorphology, Palaeoclimate, GeoQuest

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Key words: Lake Eyre basin; geology; hydrology; geomorphology; palaeoclimate

Purpose and structure of this review

This is the first comprehensive review of the current state of knowledge of the earth sciences in the Lake Eyre (hydrological) basin (LEB) (Fig. 1), one of the world's largest

dryland systems, or indeed in any of the world's large dryland drainage basins. The review focuses on the environmental character and history of the LEB, by presenting an integrated summary of the past and present physical environments of the basin. The extremely diverse nature of Earth science research in the LEB precludes the possibility of an overall synthesis. Instead, we have attempted to provide a number of syntheses under a variety of headings, ranging from the ancient (e.g. geological history, tectonics, palaeovegetation, megafauna, archaeology, palaeoclimatology) to the contemporary aspects (e.g. hydrology, fluvial systems, lacustrine systems, aeolian systems, gibber surfaces). This review is divided into four parts. Part A provides the introduction and preliminary background. Part B presents the geological history, including evidence for substantial Tertiary and Quaternary climate change. Part C reviews evidence for the contemporary environmental conditions and Part D concludes with a summary.

PART A: INTRODUCTION AND BACKGROUND

1 Introduction

More than one seventh of the Australian continent (1.14 million km^2) is composed of the semi-arid to arid LEB (Fig. 1), one of the world's largest internally drained (endorheic) basins. The basin has been the focus of numerous detailed scientific investigations in geology, geomorphology, hydrology and ecology; the environmental character of the LEB. It seems timely to assess and draw together this large body of diverse but often interrelated research. This manuscript does not mention some of the older pieces of research as they have either been superseded or their work is limited and acknowledged in publications we have referred to.

With current efforts to understand Australia's past climate and to predict future climate, over recent years considerable emphasis for research in the basin has been given to unravel the palaeoclimatic story recorded in the extensive fluvial, aeolian and lacustrine deposits of the LEB. Research into dune and dust deposits, channel morphology, floodplain geomorphology, alluvial sedimentology, lake-shoreline elevations, tectonism and palaeontology have all contributed to our present understanding of the changes that have taken place to the climate and hydrology of the basin. Actual research into these various aspects over the years has varied in terms of detail and sophistication, and this is, to some extent, reflected in our ability to treat them in the manuscript.

Palaeohydrology is one of the topics that has been investigated more intensely than others, and readers are directed to Nanson et al. (2008) for a review of this research, particularly that relating to the palaeoriver systems. Our intention here is not only to summarise this palaeohydrological story, but also to be much more inclusive and explore more broadly the environmental character and history of the LEB, starting with the topography and drainage.

2 Topography and drainage

The LEB is a shallow topographic depression that extends from the monsoon tropics at latitude 19° to the temperate zone at 32°. It is highest in the MacDonnell Ranges at ~1500 m in the central west and in the Flinders Ranges at ~1200 m in the far southwest, but neither feature contributes significant flows to Lake Eyre (McMahon et al., 2005). Approximately 70% of the basin lies below 250 m AHD (Australian Height Datum) (Kotwicki, 1986), much of it covered with aeolian dunes of the Simpson, Tirari and Strzelecki Deserts, which have partially obstructed and deranged drainage systems over these low gradient slopes. The remainder of the basin is formed of silcrete tablelands, vast gibber plains and rocky and often dissected interfluves separating wide floodplains flanking spatially ephemeral tree-lined rivers that can carry exceptionally large floods. Much of the LEB falls into the Central Lowlands Province of Jennings and Mabbutt (1986).

The Lake Eyre salt playa (9330 km^2) is -15 m AHD at Australia's lowest point and lies at the southwestern portion of the basin (Fig. 1). It is fed by ephemeral streams rising in mostly low-elevation ranges (~400 m AHD) at the northern and northeastern margins of the basin before crossing vast low-gradient (10 - 20 cm km^{-1}) plains. There are seven drainage catchments in Lake Eyre basin (Table 1); the three main river systems feeding Lake Eyre being Cooper Creek, the Diamantina River and the Georgina River (Fig. 1) and it is Cooper Creek at ~1500 km (including its primary upstream tributary, the Thomson River) which is the longest. It drains a 306,000 km^2 basin with extensive muddy floodplains up to 60 km wide. The Warburton River (sometimes called Creek), which is essentially the downstream end of the Diamantina River draining into Lake Eyre and includes the Georgina River upstream, has the largest basin area at ~360,000 km^2.
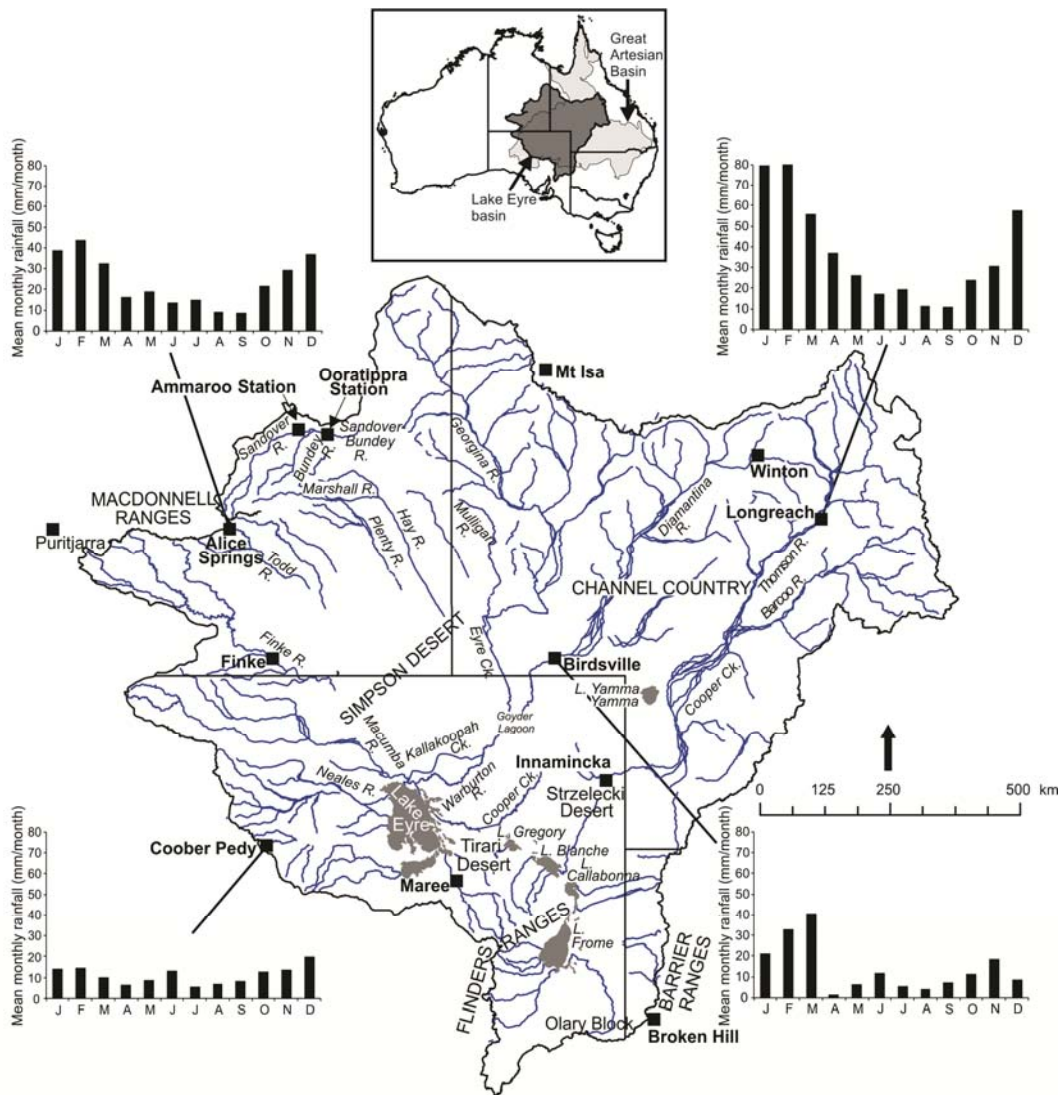
Figure 1: Overview map of the Lake Eyre basin, central Australia and mean monthly rainfall for Alice Springs airport between 1941 and 2013 (Australian Bureau of Meteorology (BOM) site number 015590), Longreach aero between 1949 and 2013 (BOM site number 036031), Birdsville airport between 2000 and 2013 (BOM site number 038026) and Coober Pedy airport between 1994 and 2013 (BOM site number 016090). Rainfall data from the BOM, http://www.bom.gov.au/climate/data/index.shtml. The extent of the Great Artesian Basin is also shown in the inset.

The largest tributary of the Diamantina River, the Georgina River, has an ~205,000 km^2 catchment, drains the central north of the basin before becoming Eyre Creek for a short distance and then joining the Diamantina at Goyder Lagoon. This 'lagoon' is an ~80 km long and ~30 km wide muddy 'swamp' that is dry most of the time. From Goyder Lagoon the river is known as the Warburton, traversing a distance of ~300 km to Lake Eyre. When the lagoon fills some of the flow diverges to Kallakoopah Creek, essentially an anabranch of the Warburton River that flows well north into the Simpson Desert before turning south to also enter Lake Eyre (Fig. 1). Such a deranged downstream

drainage system is probably a function of low valley gradients, interaction with the adjacent dunefield and the rise and fall of the Lake Eyre (drainage base level) during the Quaternary. Neotectonics may also play a role here, as it has elsewhere in the basin (Section 4.3).

Northeast of the MacDonnell Ranges the Georgina is joined by the Sandover and Bundey Rivers with a combined catchment area of ~20,000 km^2 and which are much sandier than the rest of the Georgina system. The Todd, Plenty and Hay Rivers, with basin areas between ~60,000 km^2 (Todd River basin) and ~100,000 km^2 (Hay River basin, of which Hay and Plenty Rivers belong to) are also sandy. These systems drain the southern and eastern slopes of the MacDonnell Ranges and flood into the Simpson Desert, their flows probably never reaching Lake Eyre under today's hydrological regime. The Finke River (basin area ~100,000 km^2) is also a sand-load system that floods into the Simpson Desert.

The low-lying western margin of the basin drains to Lake Eyre mostly via the Macumba and Neales Rivers (area ~35,000 km^2). In the far south lies the Frome basin, a second large endorheic terminus for rivers in the LEB (Fig. 1). In large floods, Cooper Creek overtops a natural spillway on its left bank immediately downstream of Innamincka and a portion of its flow passes down the Strzelecki Creek through the Tirari Desert to a sometimes interconnected system of lakes (Cohen et al., 2011) including Blanche, Frome, Callabonna and Gregory (Fig. 1). These flank the eastern margin of the Flinders Ranges. In the Pleistocene the Lake Frome system has overflowed to Lake Eyre via a natural spillway at about 15 m AHD (Nanson et al., 1998), but there is no connection under the present hydrological regime.

Table 1: The seven Lake Eyre basin drainage catchments, their primary rivers, maximum recorded daily discharge, mean monthly streamflow, rainfall and temperatures, and annual discharges, and Lake Eyre specifics.

Catchment and area (km²)	Primary rivers	Length (km) [a]	Indicative streamflow						Indicative weather [d]			
			Station	Period of record	Max. daily discharge (m³ s⁻¹) (date)	Mean monthly discharge (ML)	Mean annual total discharge (km³) (no. years with data)	Max. annual discharge (km³) (year)	Station	Years of operation	Mean rainfall (mm yr⁻¹)	Mean temp. (max / min) (°C)
Cooper Ck. 300,000	Thomson R.	450	Longreach (003202A) [c]	1969 - 2013	11,695 (28/01/74)	103,605	1.46 (22)	No data	Moomba airport (017123)	1995 - present	185	29 / 16
			Stonehenge (003203A) [c]	1966 - 2013	14,541 (31/01/74)	200,190	2.19 (21)	No data				
	Barcoo R.	400	Retreat (003301B) [c]	1999 - 2013	2147 (18/01/74)	102,983	1.46 (11)	No data				
			Blackall (003303A) [c]	1969 - 2013	1797 (21/04/90)	8617	0.11 (35)	No data				
	Cooper Ck.	900	Currareva (003101A) [c]	1939 - 1988	25,468 (02/02/74)	274,000 [f]	4.74 (9)	15.04 (1974) [b]				
			Nappa Merrie (003103A) [c]	1949 - 2013	6066 (16/02/74)	136	1.83 (16)	No data				
			Cullyamurra Waterhole (A0030501) [d]	1973 - 2012	6340 (19/02/74)	101,662	2.06 (11) [b]	11.47 (1974) [b]				
Diamantina R. 160,000	Diamantina R.	729	Diamantina Lakes (002104A) [c]	1966 - 2013	6823 (30/01/74)	158,068	2.19 (20)	No data	Birdsville airport (038026)	2000 - present	168	31 / 17
	Warburton R.	300	No data									
Georgina R. 200,000	Georgina R.	1130	Roxborough Downs (001203A) [c]	1967 - 2013	3833 (24/02/77)	98,570	1.46 (20)	5.71 (1974) [b]	Boulia airport (038003)	1886 - present	262	32 / 17
			Camooweal (001204A) [c]	1968 - 1988	1974 (21/02/77)	11,900 [f]	No data	No data				
	Sandover R.	300	No data									
	Burke R.	200	Boulia (00102A) [c]	1966 - 2013	2885 (02/02/74)	32,559	0.25 (7)	No data				
	Hamilton R.	180	No data									
	Eyre Ck.	450	No data									

Catchment and area (km²)	Primary rivers	Length (km) [a]	Station	Period of record	Indicative streamflow				Station	Indicative weather [d]		
					Max. daily discharge (m³ s⁻¹) (date)	Mean monthly discharge (ML)	Mean annual total discharge (km³) (no. years with data)	Max. annual discharge (km³) (year)		Years of operation	Mean rainfall (mm yr⁻¹)	Mean temp. (max / min) (°C)
Todd R.	Todd R.	50	Wigley Gorge (G0060046) [e]	1962 - 2013	1045 (01/04/88)	1060 [f]	0.01 (39)	No data	Alice Springs airport (015590)	1940 - present	284	29 / 13
60,000			Anzac oval (G0060009) [e]	1972 - 2013	1195 (01/04/88)	1150 [f]	0.01 (19)	No data				
	Hale R.	250			No data							

Catchment and area (km²)	Primary rivers	Length (km) [a]	Station	Period of record	Indicative streamflow				Station	Indicative weather [d]		
					Max. daily discharge (m³ s⁻¹) (date)	Mean monthly discharge (ML)	Mean annual total discharge (km³) (no. years with data)	Max. annual discharge (km³) (year)		Years of operation	Mean rainfall (mm yr⁻¹)	Mean temp. (max / min) (°C)
Finke R.	Finke R.	730	Railway bridge (G0050140 [e]	2010 - 2013	456 (11/01/10)	No data	0.12 (3)	No data	Finke post office (015526)	1938 - 1980	189	29 / 15
100,000	Macumba R.	220			No data							
Hay R.	Hay R.	200			No data				Ringwood (015526)	1964 - present	292	29 / 13
	Plenty R.	410			No data							
100,000	Marshall R.	430			No data							
L. Frome	Neales R.	430			No data				Coober Pedy (016007)	1921 - present	157	29 / 14
200,000	Frome R.	250			No data							

			Lake Eyre						
Catchment area (km²)	Max. depth (m)	Max. volume (m³)	1974 flood [b]		@ 5 m beach ridge [g]		@ 10 - 12 m beach ridge [g]		
			Deepest point (m AHD)	Shoreline (m AHD)	Palaeo area (km³)	Palaeo volume (km³)	Palaeo area (km³)	Palae ovolume (km³)	
1,140,000	5.7	30.1	-15.2	-9.5	16,000	214	21,000	318	

Source:

[a] Variability exists in reported length, here lengths measured from Google Earth

[b] Kotwicki (1986)

[c] Queensland government (http://watermonitoring.dnrm.qld.gov.au)

[d] Bureau of Meteorology (http://www.bom.gov.au/water/hrs)

[e] Northern Territory government (http://lrm.nt.gov.au/water/water-data-portal)

[f] McMahon et al. (2005), years of record up to 2004

[g] Leon and Cohen (2012)

3 Climate

Synoptically, the present rainfall patterns are most influenced by the moist tropical air passing over the surrounding highlands (i.e. the northern tablelands and the eastern highlands) in the warmer months (Allan, 1990). These incursions are a function of a subtropical high pressure ridge associated with the southern fringe of the Hadley Cell (McMahon et al., 2008a), which results in a strong summer-dominant rainfall regime in the north and northeastern LEB and weak winter dominant rainfall regime in the south of the basin (Gentilli, 1986). During the summer months, a southern migration of the subtropical high pressure ridge allows the northern and eastern LEB to be more influenced by tropical flows, as well as the southern limit of the monsoon and what are relatively uncommon tropical rainfall events (McMahon et al., 2008a). During summer the southern section of the basin faces dry conditions. Conversely, in winter as the high pressure ridge drifts north, the north/northeastern LEB is dry while the southern parts are affected by the northern limit of westerly cold fronts (McMahon et al., 2008a).

Data from the Australian Bureau of Meteorology (BOM)[1] indicate that mean annual rainfall in the basin varies significantly, from ~700 mm in the northeast to approximately 125 mm close to Lake Eyre itself. Following heavy rainfall events the water flows into the lake; Kotwicki (1986) suggests that this has occurred to depths of several metres in the lake every 5 - 10 years last century. Mean annual pan evaporation across the basin varies from 2400 mm in the northeast to > 3600 mm at the lake, greatly exceeding annual rainfall in all locations.

Mean monthly maximum temperatures at Birdsville airport between 2000 and 2013 (BOM site number 038026) are 40.2°C in January and 21.1°C in June, with mean monthly minimum temperatures ranging from 26.4 °C in January to 7.0 °C in July.

PART B: GEOLOGICAL HISTORY AND TERTIARY AND QUATERNARY CLIMATE CHANGE

The LEB is not just a hydrological basin; it is composed of nested geological basins that have strongly influenced the current landscape. Part B presents a summary of the geological and geo-environmental history of the LEB including tectonics, palaeovegetation, megafauna, archaeology and palaeoclimatology. Our understanding of many of the modern processes, reviewed in Part C, arises from an understanding of the geological past.

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[1] Available from http://www.bom.gov.au/climate/data/index.shtml

4 Geology and geo-environmental history

The geology of the LEB is dominated by Mesozoic and Cenozoic rocks, with unconsolidated Late Tertiary and Quaternary sediments at the surface (Fig. 2a). Substantial areas of sometimes intensely altered and mineralised Palaeozoic and Proterozoic basement rocks outcrop in an arc around the basin periphery, from the north near Mt Isa to the northwest in the central MacDonnell Ranges near Alice Springs, and again in the south in the Flinders and Barrier Ranges. Subsequent to the regression of shallow seas at the end of the Early Cretaceous (Late Albian), deposition within the region of the LEB has been terrestrial in nature. Subaerial weathering dominated inland Australia from the Late Cretaceous to the Late Palaeocene when tectonic subsidence in northeastern South Australia produced the wide and shallow intracratonic Lake Eyre (geological) Basin (Krieg et al., 1990). This consists of a thick (> 400 m) sedimentary succession that is separated into the Tirari and Callabonna Sub-basins by the Birdsville Track Ridge, which is expressed at the surface by silcrete-capped (Cordillo, Gason and Cooryanna) domes of Mesozoic sediments (Callen et al., 1995a,b) (Fig. 2b). Nested within a series of older basins, the surface of the LEB now extends to ~15 m below sea level at its depocentre in Lake Eyre itself. The alluvial, lacustrine and aeolian processes that have occurred within it since the Late Palaeocene are well recorded in its deposits, providing an exceptionally long record of past climate and flow regime changes. Three main phases of sedimentation are recognised: P1 occurred from the Late Palaeocene to the Middle Eocene (Fig. 3a); P2 occurred from the Late Oligocene to the Early Miocene (Fig. 3b); P3 occurred in the ?Pliocene and Quaternary (Fig. 3c) (Krieg et al., 1990; Callen et al., 1995a,b; Alley, 1998). These phases are summarised below. Broadly speaking, this sequence involves the Lake Eyre Basin shifting progressively from a much wetter environment in the Palaeogene to the arid conditions of present.
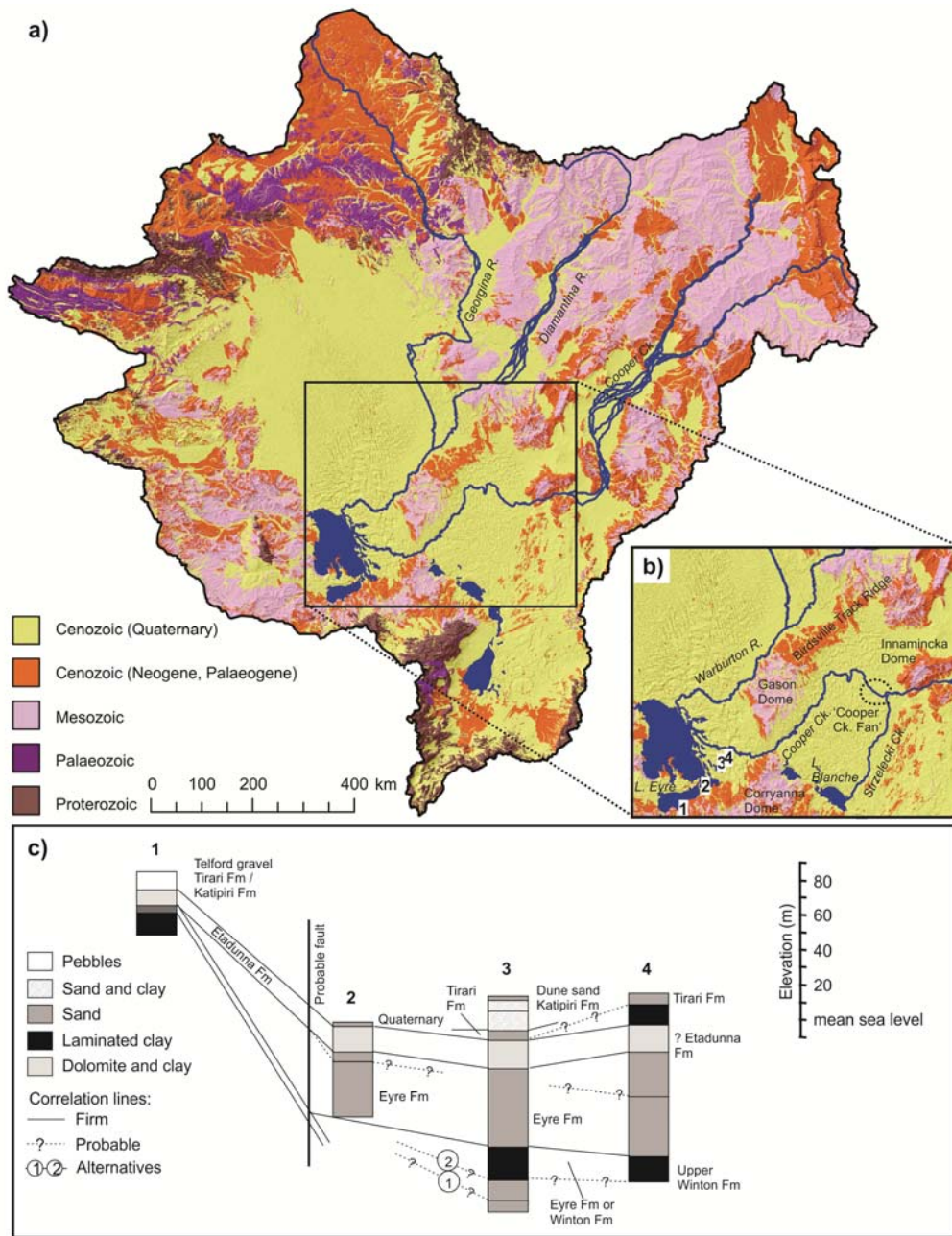
Figure 2: a) Geology and b) domes of the southeastern portion of the LEB, and c) simplified stratigraphy (modified from Wells and Callen, 1986). Location of cores highlighted on b).
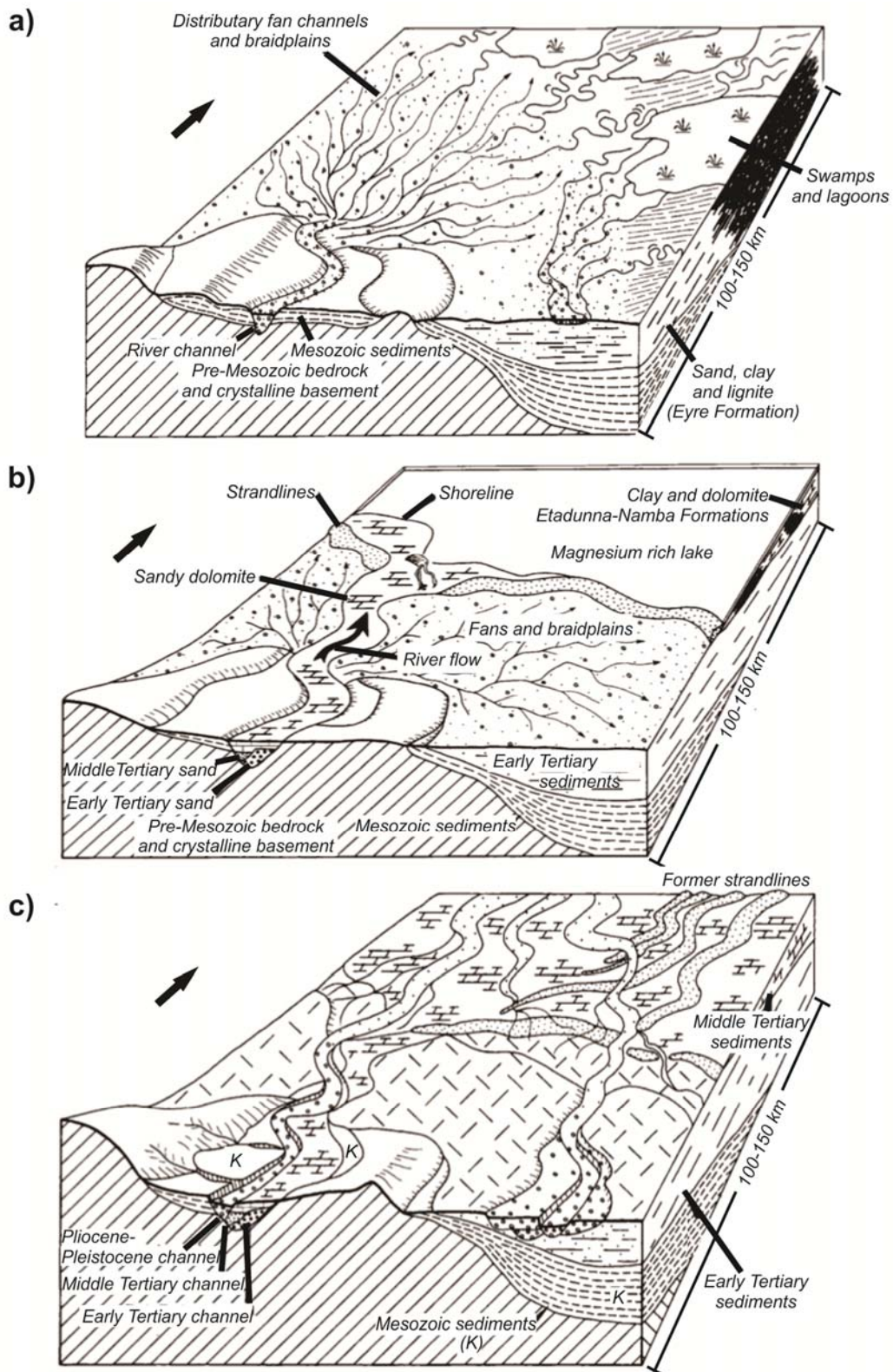
Figure 3: The three main phases of sedimentation. a) Phase 1 (P1) from the Late Palaeocene - Early Eocene, b) P2 from Late Oligocene - Miocene, and c) P3 from ?Pliocene - Quaternary palaeogeography of the LEB (from Krieg et al., 1990 and Alley, 1998).

4.1 The Mesozoic

A major unconformity separates the Lake Eyre Basin from the underlying Eromanga Basin (Early Jurassic - Late Cretaceous), which is also a broad intracratonic basin dominantly comprising terrestrial strata. The structure of the Eromanga Basin influences the characteristics of the modern LEB (Senior et al., 1978) in that some of the pronounced northwest and northeast trending anticlines within it control current drainage patterns. Upper Cretaceous Winton Formation (Senior, 1968) is exposed in the structural highs. This is the youngest unit of the Eromanga Basin and outcrops in southwestern Queensland, northwestern New South Wales and northeastern South Australia (Dettmann et al., 2009). The formation consists of terrestrial mudstone and sandstone derived from fluvial and lacustrine environments interpreted by Senior (1968) to have been deposited in association with northward regional drainage. Thicknesses of the formation range from < 100 m along the basin margin to > 1200 m in the centre (approximately at the Queensland/South Australian border).

4.2 The Cenozoic

The Cenozoic in inland Australia has been generally associated with episodic fluvial, lacustrine and aeolian sedimentation. This period has been the subject of considerable research, especially the latter stages, because it explains much about the present character of the LEB. A convincing record has been constructed for an overall drying trend through to the Late Quaternary (Alley, 1998; Nanson et al., 2008).

4.2.1 Palaeogene - Neogene

Phase one of sedimentation (P1) in the Lake Eyre Basin is represented by the Eyre Formation. It is Late Palaeocene to Middle Eocene in age and represents a generally warm and wet phase of deposition when active river systems dominated and numerous freshwater swamps and lagoons were present (Fig. 3a) (Alley, 1998). Wopfner et al. (1974) provide a comprehensive description of the type section of this formation, near Innamincka, as well as a number of other reference sections. It is a partly silicified, fluvially-derived arenite with minor conglomerate. The base of the formation is characterised by polished gravel comprising milky quartz, chert, jasper, agate and fossil wood and silcrete (Wopfner et al., 1974; Wopfner, 1978; Alley, 1998). The top of the formation is a silicified palaeosol in surface exposures, or the contact with a Miocene carbonate-shale sequence (the Etadunna Formation in the Tirari Sub-basin; Wopfner et al., 1974). The arenite becomes finer

grained, richer in organic material and more laminated towards the south (Senior et al., 1978). Lignite and carbonaceous shale interbeds are common, and microflora (e.g. pollen, spores, microplankton, cuticle, algae, fungi) and macroflora fossils (e.g. wood, twigs, leaves, shoots, seeds, fruits and flowers) are widespread (Harris, 1973; Wopfner et al., 1974; Alley, 1998). The significance of these fossils is discussed in Section 4.4.

The thickness of the formation was influenced by syndepositional epeirogenic movements of the Olary Block and Barrier Ranges (Fig. 1) (Wopfner, 1974), as well as by subsidence of the Lake Eyre Basin (Krieg et al., 1990), and locally exceeds 100 m (Wopfner et al., 1974; Alley, 1998). It is thought that in the Oligocene and subsequently the basin has also experienced slow tectonic uplift to form the Innamincka and Corryanna - Gason Domes through which Cooper Creek has incised and possibly is still incising (Fig. 2b) (Senior et al., 1978; Wells and Callen, 1986; Nanson et al., 2008; Jansen et al., 2013).

It has been suggested that the extensive silcretes across the basin may have formed from frequently recharged shallow groundwater in the wetter climate (Alexandre et al., 2004). Wopfner et al. (1974) were the first to assert that the silcrete formation probably ended in the Late Oligocene because the subsurface layers were dissected before the overlying Miocene beds were deposited. Earlier work indicated the silcrete primarily formed in the Pliocene (Stephens, 1971), and concomitantly with the laterites (Wopfner, 1960; Wopfner and Twidale, 1967). The fluctuating water table (key to cement formation) was interpreted by Alexandre et al. (2004) to probably have been as a result of monsoon-like climatic conditions and consequent seasonal wetting and drying, coupled with epeirogenic uplift. This eventually shifted conditions favouring fluvial to those favouring lacustrine processes in P2. Importantly, during this early Tertiary period moist warm conditions and associated vegetation induced tropical weathering and the formation of iron-rich weathering products that are believed to have given rise to the characteristic red-orange colour of the soils and dunes derived from this regolith in the LEB and other parts of Australia's 'Red Centre' (Pillans, 2005; Pain et al., 2012).

The second phase of deposition (P2) commenced in the Late Oligocene and lasted until the Early Miocene, and while it corresponds to a drier phase than P1, evidence indicates shallow freshwater lakes, sometimes magnesium rich, were still present (Fig. 3b) (Krieg et al., 1990; Alley, 1998). The Etadunna Formation was first described by Stirton et al. (1961) from an outcrop < 60 km from the east of Lake Eyre. The contact between it and the underlying Eyre Formation is typically ferruginous (Wopfner et al., 1974).

The Etadunna Formation in the Tirari Sub-basin, and its correlative the Namba Formation in the Callabonna Sub-basin, belong to P2 and generally comprise clay, fine sand and carbonate, with lesser conglomerate (Alley, 1998). Two more units, the Doonbarra Formation and Cadelga Limestone, also occur in the Callabonna Sub-basin, with the former a ferruginous fine to medium sand underlain by the latter, a cherty and dolomitic lacustrine limestone. There exist multiple interpretations of the assignment of the Doonbarra Formation and Cadelga Limestone (Wopfner, 1974; Callen, 1983), however, it is generally understood that its deposition ensued and/or was concurrent with development of an extensive Late Eocene to Early Oligocene 'Cordillo silcrete' surface (Wopfner, 1978). It appears that P2 ended with the deposition of the Etadunna Formation and its equivalents in the Miocene (Wopfner et al., 1974).

The third and final phase of sedimentation (P3) probably commenced in the Pliocene associated with a further increase in aridity, and continued through the Quaternary to present (Fig. 3c). P3 is represented by numerous near-surface formations throughout the basin and these are discussed here in stratigraphic order.

The Tirari Formation, identified by Stirton et al. (1967), is thought to represent a shift towards marked aridity. Late Pliocene to ?Early Pleistocene in age, it is characterised by red-brown mudstone and sandstone with gypcrete (Alley, 1998; Nanson et al., 2008). It is 4 - 5 m thick, widespread and exposed in the central Tirari Desert and north into the Simpson Desert (Stirton et al., 1967; Alley, 1998). The depositional environment is interpreted as a fan-delta system of channels related to intermittent stream flow in semi-arid conditions. Megafauna fossils (*Diprotodon* and short faced kangaroo, *Procoptodon*, see Section 4.5 and Fig. 5) are known from this formation. It too is representative of a moister environment than today (but drier than previously), with a larger Lake Eyre and more widespread alluvial activity.

4.2.2 Quaternary

Following deposition of the Tirari Formation, the Willawortina Formation of fine sandy mud with coarser piedmont-derived clastics was laid down on the flanks of the northern Flinders Ranges (Callen and Tedford, 1976). In the Callabonna Sub-basin it is represented by sandy mud, silty dolomite, clast-supported gravel and clayey gravel (Callen and Tedford, 1976; Alley, 1998) and, while it is understood that deposition commenced in the Early Pleistocene (Callen et al., 1995b), the deposition of comparable clastics has probably continued to the present (Alley, 1998).

With the exception of the Willawortina Formation, Lower to Middle Pleistocene sediments in the Lake Eyre Basin are collectively referred to as the Kutjitara Formation, and comprise fine sands and red mudstone deposited under fluviolacustrine conditions in the Tirari Desert (Callen et al., 1986; Krieg et al., 1990). It is less intensely gypcreted than the Tirari Formation (Nanson et al., 2008) and is well exposed on the lower reaches of Cooper Creek where it has been observed directly overlying the Etadunna Formation, and on the Warburton River, where its channels are cut into the Tirari Formation (Alley, 1998). Along these rivers, the Kutjitara Formation is predominantly fluvial, however, it is closer to a lacustrine facies near Lake Eyre (Alley, 1998). It has not been recognised as yet in the middle and upper reaches of the Cooper and Diamantina rivers due to a lack of exposed alluvium, but it presumably exists in the sub-surface.

Overlying the Kutjitara Formation is a similar but less weathered mostly fluvial unit, the Middle to Upper Pleistocene Katipiri Formation (Fig. 4). It was first described as the Katipiri Sands by Stirton et al. (1961) and described in more detail by Callen et al. (1986) and Nanson et al. (2008), and is a largely unconsolidated white fine to coarse river sand with trough cross bedding, mud drapes and a thin basal conglomerate of gypsum and carbonate-cemented lithic fragments (Tedford et al., 1986). Interbedded lacustrine units are at present proximal to Lake Eyre. The Katipiri is akin to the Eurinilla Formation in the Strzelecki Desert and the Millyera Formation near Lake Callabonna (Callen and Tedford, 1976), all evidence of periods of widespread fluviolacustrine conditions in the Lake Eyre Basin in the Middle to Late Pleistocene (Alley, 1998). Along much of Cooper Creek and the Diamantina River the Katipiri Formation is extensive, as fluvial sands underlying most of the floodplain; the uppermost unit is capped with 1 - 2 m of overbank mud (Nanson et al., 1986, 1988, 1992a, 2008). It tends to follow a fining downstream trend towards Lake Eyre. Close to the lake there are broad palaeochannels containing fluviolacustrine facies and shoreline dunes (Callen et al., 1986; Magee, 1997), the latter marking the extent of the former lake in which the lacustrine sediments accumulated (Alley, 1998).

The final part of the third phase of sedimentation (P3) is represented by mostly Late Pleistocene to Holocene linear dunes - the Coonarbine Formation (Strzelecki Desert) and Simpson Sand (Simpson Desert), which possibly overlie older aeolian deposits that now have a calcareous palaeosol crust (Alley, 1998; Fujioka et al., 2009). The lee side of many playas and pans in the basin, including Lakes Eyre, Frome and Callabonna, are now bordered by transverse dunes, the product of sediment deflation (Fitzsimmons et al., 2007a; Maroulis et al., 2007; Cohen et al., 2010). Intermittent sedimentation in and around the Lake Eyre Basin continued throughout the Late Pleistocene (Alley,

1998), leaving a record of when it was wet and dry. This record is examined in Section 5 where a detailed record of Quaternary palaeoenvironmental change for the LEB is interpreted.
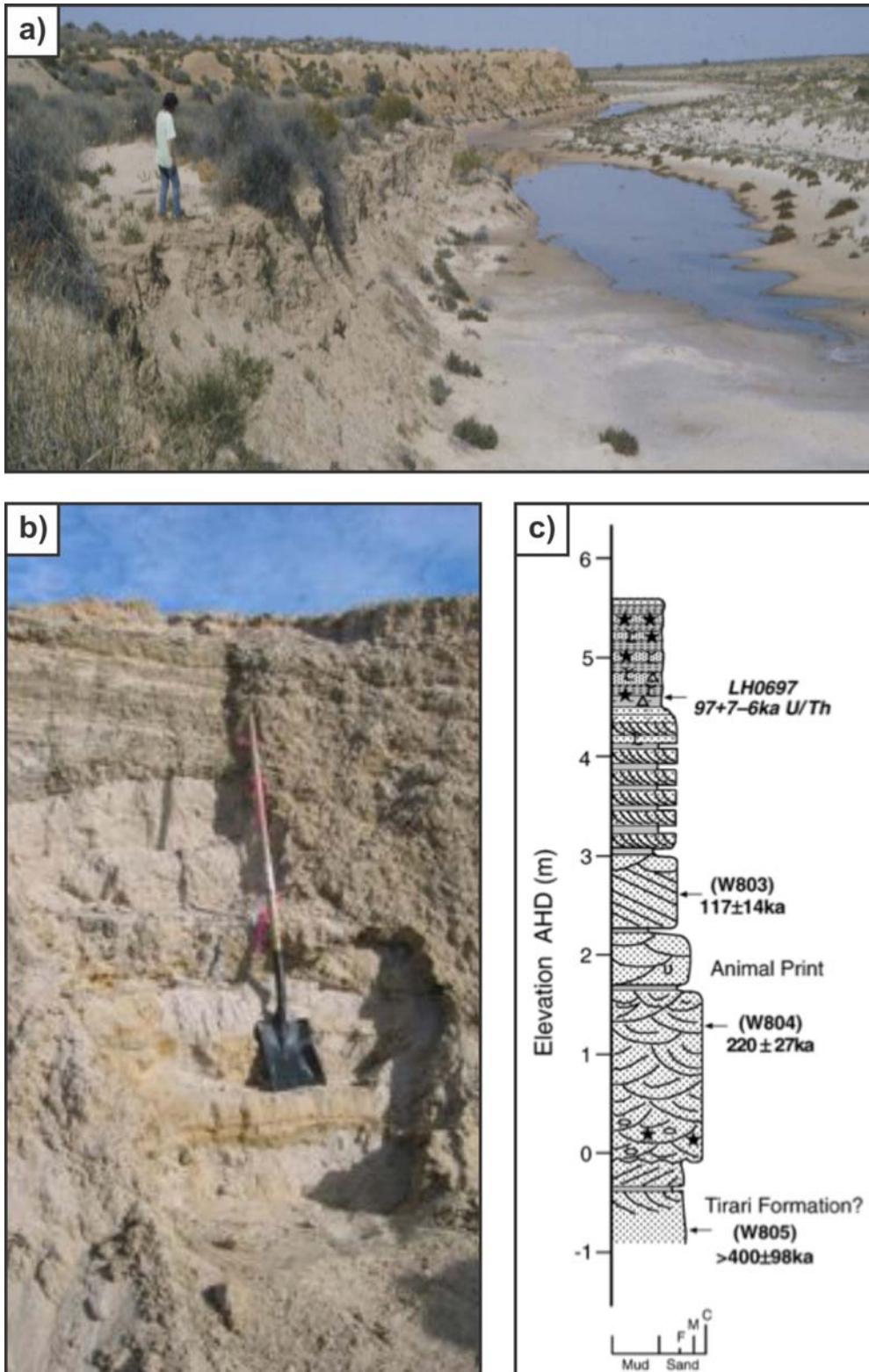

Figure 4: Katipiri Formation: a) distant and b) close up view of the exposed Katipiri Formation on the lower Cooper Creek at Etadunna Station. c) Stratigraphic section at South Tilla Tilla Waterhole, approximately 5.5 ± 1.5 m AHD, where the Katipiri Formation is possibly underlain by the Tirari Formation (from Nanson et al., 2008).

4.3 Tectonics

Much of central Australia is nearly flat, as evidenced by most alluvial areas within the Channel Country lying below 150 m AHD. The low-gradient rivers of this region are sensitive to subtle tectonic movements, and where this has occurred the effects on sedimentation and channel pattern can be profound.

Recent research indicates that Neogene-Quaternary tectonically induced landscape change in central Australia has been widespread but generally of low magnitude (Sandiford et al., 2009; Quigley et al., 2010). For example, the Lake Eyre and Torrens basins are separated by an ~80 m AHD divide. This divide is associated with a palaeolake of probable Late Miocene age, Billa Kalina (Callen and Cowley, 1995), centred approximately 150 km west of Marree with a maximum size of ~15,000 km^2 and an estimated catchment size equivalent to Lakes Eyre and Torrens combined (Sandiford et al., 2009; Quigley et al., 2010). The existence of such a Late Miocene lake indicates major drainage re-organisation in the region of the Flinders Ranges (Quigley et al., 2007) and the regional drainage patterns within LEB can be largely attributed to this Neogene-Quaternary tectonism (Senior et al., 1978; Wasson, 1983a; Wells and Callen, 1986; Alley, 1998).

Some authors (e.g. Magee et al., 1995) indicate that geomorphic processes rather than tectonics can explain many of the characteristics of Lake Eyre itself, including its shape and negative height datum. Whether tectonics have exerted a major (e.g. Quigley et al., 2010) or minor influence (e.g. Magee et al., 1995) on the geomorphic expression of the LEB at a regional level will not be explored here, however, the role of tectonics at the local level is more apparent (below). Readers are referred to Quigley et al. (2010, and references therein) and the earlier works contained in Veevers (1984) for a comprehensive review on the broader influence of tectonics on geomorphology in Australia, including the LEB.

At the local and more detailed level, Jansen et al. (2013) report the rate of sediment accretion upstream of the Innamincka Dome due to tectonic alluvial impoundment to be 48 ± 21 mm ka^{-1} during the past 270 ka. Based on a numerical model of episodic erosion calibrated with Beryllium-10 (^{10}Be) measurements, it has been suggested that Cooper Creek contemporaneously incised the Dome at a minimum long-term bedrock incision rate of 17.4 ± 6.5 mm ka^{-1} (Jansen et al., 2013). Background bedrock denudation rates for central Australia also founded on ^{10}Be are estimated to be significantly slower than this, at 0.2 - 5 mm ka^{-1} (e.g. Bierman and Caffee, 2002; Belton et al., 2004; Heimsath et al., 2010; Fujioka and Chappell, 2011).

The rising Innamincka Dome has resulted in the formation of an extensive Cooper Creek floodplain with associated anabranching channels extending hundreds of kilometres upstream (Nanson et al., 2008; Jansen et al., 2013) (Fig. 2b). Similar areas of extensive floodplain and swamp caused by possible neotectonics occur at Goyder Lagoon and above several bedrock structures distributed several hundred kilometres upstream of Birdsville (Fig. 1). The lower Cooper in the vicinity of the Corryanna and Gason Domes (Fig. 2b) also appears to be affected by movement of these structures, with the formation of a series of smaller swamps or lagoons in their vicinity. Drainage diversions resulting from faulting in the Neales River region further indicate that major structural lineaments appear to have controlled stream direction (Waclawik et al., 2008).

4.4 Palaeovegetation

At the start of the Cenozoic, warm temperate rainforests were extensive across Australia, including in the now arid centre (White, 1994; Blewett et al., 2012). By the end of the Eocene rainforest became generally restricted to the valleys while open sclerophyllous vegetation evolved to occupy the hinterlands. By the Early Miocene, rainforests existed only as pockets within woodlands and by the Pliocene vegetation was dominated by the greatly diversified Casuarinaceae and Chenopodiaceae families (Martin, 2006).

The micro and macroflora fossils identified in the basal unit of the Eyre Formation indicate that vegetation between the Late Palaeocene to Middle Eocene varied from angiosperm-dominated to gymnosperm-dominated (Alley, 1998), which supports the notion of a warm and wet climate. Cunoniaceae forests characterised central Australia during the Late Palaeocene to Early Eocene; Cunoniaceae floras were joined by Myrtaceae floras in terms of dominance and diversity towards the end of this time (Martin, 2006).

Mid-Eocene floras in the basin consisted of predominantly rainforest floras - Proteaceae, Sterculiaceae, Myrtaceae and Casuarinaceae (Christophel et al., 1992), while *Casuarina* and *Eucalyptus* forests and woodlands were prominent during the Early to Mid-Miocene (Benbow et al., 1995a).

The Mid-Miocene represents the first major shift towards aridity in central Australia, with abundant dry open woodlands and chenopod shrublands (Martin, 2006). Early Pliocene and Pliocene-Pleistocene palynofloras from near Lake Frome are similar to modern pollens, except for an

increased abundance of Cyperaceae, indicating dominance of largely arid scrubland vegetation (Martin, 1990). However, more wetlands were present at that time than what exist today.

4.5 Megafauna and possible human interaction

Fossils of twenty megafaunal species have been identified around Lake Eyre, Cooper Creek and the Diamantina/Warburton River (Webb, 2008), in the Tirari, Kutjitara, Katipiri Formations (Krieg et al., 1990; Alley, 1998). These include the 'giant wombat' (*Diprotodon*, Fig. 5a), 'giant short-faced kangaroo' (*Procoptodon*, Fig. 5b), a large flightless bird (*Genyornis*, Fig. 5c), 'giant goanna' (*Megalania*, Fig. 5d) and 'marsupial lion' (*Thylacoleo*, Fig. 5e). Similarities between species in the LEB and area below the southern margin of the basin (c.f. western sub-region of Webb, 2008) indicate that the megafauna probably migrated between the two (Webb, 2008).

Nanson et al. (2008) speculate that the abundance of fossils in the Tirari, Kutjitara and Katipiri Formations indicates that the marsupials and reptiles took advantage of the pluvial conditions associated with the mid to late interglacials and interstadials of the Quaternary recognised to have occurred in presently-arid inland Australia. The variety of carnivores attests to the species richness of the regional trophic pyramid consisting of terrestrial and aquatic fauna (Webb, 2009).
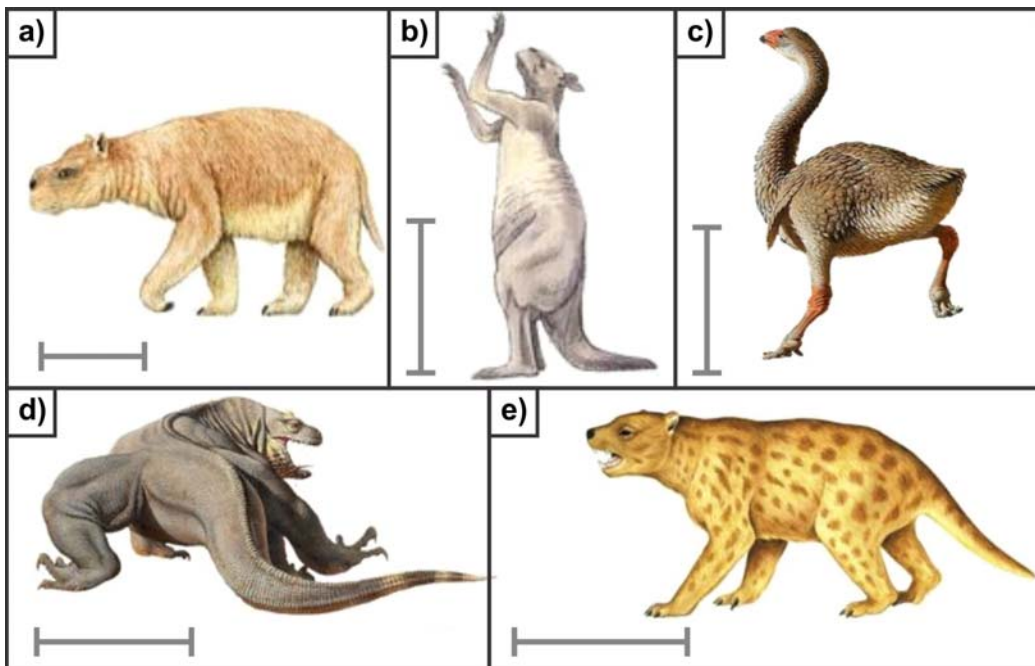

Figure 5: Artistic impressions of a) *Diprotodon* (3 m long), b) *Procoptodon* (2 m high), c) *Genyornis* (2 m high), d) *Megalania* (5.5 m long), and e) *Thylacoleo* (2 m long). Drawings a) and b) from Queensland Museum, c) and d) from Victorian Museum, and e) from Australian Museum.

By the Late Quaternary, nearly 96% of Australian Pleistocene land animal genera weighing more than 45 kg had become extinct (Flannery and Roberts, 1999; Roberts et al., 2001). This timing is broadly concurrent with human colonisation (60 - 46 ka, Roberts et al., 1990; Bowler et al., 2003) from the Indo-Malaysian region (Roberts et al., 1990). There is debate as to the primary cause of the megafaunal extinctions but it has been widely asserted that the extinctions were in some way related to humans (Miller at al., 1999, 2005a; Roberts et al., 2001). For example, an abrupt reduction in plant diversity consequent of the imposition of human fire regimes is one causal hypothesis (Miller et al., 2005a). Rule et al. (2012) suggest that human arrival was ultimately responsible for the extinctions, and further, that it triggered a pronounced ecosystem shift from mixed rainforest to sclerophyll vegetation by reducing herbivore pressure and increasing fire. Some recent interpretations, however, have argued that changing climate may have played a major role (Wroe and Field, 2006; Webb, 2008, 2009).

This ambiguity arises from three sources. Firstly, no kill-sites have been identified in Australia, unlike in North America and Europe where kill-sites have been well documented and large numbers of megafauna (e.g. mammoths) were slaughtered, mostly for food (e.g. Haury et al., 1959; Demay et al., 2012). Secondly, it appears that the period of extinction was also a period of significant climate change; 50 - 47 ka was the last time Lake Mega-Frome was connected to Lake Eyre (Cohen et al., 2011, 2012a) and Lake Eyre appears to have never been above -10 m AHD after this time (Leon and Cohen, 2012). The ensuing drop in water levels in both Eyre and Frome would appear to represent a major shift to more arid conditions coincident with both the arrival of humans and extinction of most of the megafauna (Cohen et al., 2011).

The third concern in relation to the implication that humans caused the extinction of the megafauna in the LEB is that there is no record of human occupation of the Australian arid zone until about 40 - 35 ka (Smith et al., 2008), somewhat after when humans first arrived in northern Australia between 60 and 50 ka (Roberts et al., 1990). The evidence of human occupation in the LEB itself is only for the period after 30 ka and is very limited in abundance until ~15 ka, indicating that occupation must have been sparse during this time and that there was possibly no human occupation of the basin prior to 30 ka. Surface collections of stone artefacts interpreted to date from the Mid to Late Holocene (i.e. last 5 ka) is the most common evidence of habitation found in arid northeast South Australia including the LEB (Hughes and Lampert, 1980), and the first European explorers some 100 - 150 years ago encountered sparse but widely distributed human habitation in the basin.

Smith (1987) described desert dwellers living in the MacDonnell Ranges, specifically the Puritjarra rock shelter just outside the basin (Fig. 1), at about 22 ka. These people would have relied on the resources from the surrounding spinifex sand hill habitat (Smith, 1993). Central Australian spinifex habitats have the greatest variety of plant food species compared to woodland and watercourse habitats (Latz and Griffin, 1978).

Remarkably, people of the Wangkangurru tribe did occupy the vast dunefields of the southern Simpson Desert (Fig. 1). Their only reliable sources of water were small soaks that they had to excavate below the dunes to access (Hercus, 1985; Hercus and Clark, 1986). Aborigines were present throughout much of the LEB when Europeans first explored the region in the nineteenth century (Hercus, 1985; Hercus and Clark, 1986).

4.6 Section 4 summary

Cenozoic and Mesozoic rocks and unconsolidated sediments extend over large areas of the LEB, however, intensely altered and mineralised Palaeozoic and Proterozoic rocks outcrop from the north to the northwest and again along the southern margins. Following the reversal of northward drainage in the Cretaceous, the LEB has known slow subsidence and associated geological changes during the Cenozoic. This resulted in a series of nested geological basins with endorheic drainage that have accumulated a remarkable record of sedimentary, climatic and hydrological history. Structural interfluves within the basin interior are predominantly Late Cretaceous. The Winton Formation and the subsequent sedimentary and fossil record indicated a progressive drying from tropical moist conditions during the Palaeogene. Until the Mid-Eocene, the basin was characterised by predominantly rainforest flora.

With progressive drying from Late Oligocene to Early Miocene, shallow freshwater lakes, sometimes magnesium-rich, remained to form cherty and dolomitic lacustrine limestones and clays of the Etadunna Formation. The Mid-Miocene represents the first major shift towards aridity in the LEB and to open woodlands and shrublands as Australia's iconic *Casuarina* and *Eucalyptus* genera greatly diversified. The Pliocene and Quaternary were associated with further drying and the development of playa lakes and deposition of the fluviatile but strongly gypsiferous Tirari Formation. The Quaternary largely fluvial units of the Kutjitara and Katipiri Formation followed and retain an extensive record of oscillating river conditions associated with wet-dry episodes. They harbour a remarkable fossil record of marsupial, reptilian and avian megafauna that occupied at least the river corridors before becoming extinct at ~45 - 40 ka, soon after human arrival on the

continent at ~60 - 50ka. This coincidence in timing is often attributed to human hunting and associated environmental impact, however, with no evidence of humans present in the basin at this time, and this being the last time that the megalakes of the LEB supported a significant volume of fresh water, it is possible that hydroclimatic changes may have had an exclusive or at least a significant additional role to play.

5 Palaeoclimatic interpretations

Cosmogenic exposure dating indicates that the stony deserts adjacent to the Simpson Desert in the LEB started to form at the beginning of the Quaternary with aridity commencing at about 2 - 4 Ma (Fujioka et al., 2005). The alternation between wetter and drier episodes in response to major climatic oscillations (Nanson et al., 1992a, 2008) resulted in evaporites dominating Quaternary lake sequences (Magee et al., 1995). During wetter episodes, Lake Eyre expanded to about 35,000 km^2 or more than 3 times its current maximum area of ~9,300 km^2 (DeVogel et al., 2004), and was more than 25 m deep. At this depth it is likely to have merged with Lake Mega-Frome (Leon and Cohen, 2012; see also Section 5.2.2). Contemporary climate is variable and oscillates between flood periods and droughts, each lasting several years, however, it is clear that periods of Quaternary climate were much more so, oscillating between dry episodes and those much wetter than today and lasting hundreds or possibly thousands of years and loosely linked to global glacial cycles (e.g. Nanson et al., 1992a; Croke et al., 1996; Hesse et al., 2004; Magee et al., 2004; Nanson et al., 2008; Webb, 2008, 2009; Reeves et al., 2013b). In terms of Quaternary climate, it is not well understood whether the dramatic deviations were a result of southern - northern hemisphere insolation differences (Allan, 1985; Suppiah, 1992) although Magee et al. (2004) and Miller et al. (2005b) have presented an interpretation that the wet phases were more closely associated with the northern hemisphere winter insolation minima rather than with the southern hemisphere summer insolation maxima. This assumes that northern hemisphere forcing has the primary control over the Australian monsoon and, consequently, the northern Australia moisture balance, an idea that will be reassessed below.

5.1 Palaeofluvial evidence

Until the advent of luminescence dating, an understanding of climate change during the Quaternary in the LEB was restricted to the past ~30 ka with carbon-14 (^{14}C) dating. Even then, suitable sites and dating material were very limited due to a lack of fossil carbon in this arid environment. Using luminescence-based sedimentary dating, Nanson et al. (1986, 1988) and Rust and Nanson (1986),

published the first alluvial record for the LEB that extended the period of observed climate and flow-regime change beyond the radiocarbon record to about 300 ka. They showed that the Channel Country had undergone prolonged periods of wetness, as evidenced by much larger sand-load rivers than the present small mud-dominated anabranching systems.

This early luminescence work for the LEB was summarised by Nanson et al. (1992a), showing that climate had fluctuated between wet and dry conditions through at least two full glacial cycles. In the more recent compilation of alluvial luminescence dates from both the LEB and catchments draining to seas north of Australia, Nanson et al. (2008) (Fig. 6) show a clearly defined period of enhanced fluvial activity during to mid to late marine isotope stage (MIS) 5 (~120 - 80 ka) and another period of lesser activity MIS 4 and early MIS 3 (~65 - 40 ka). After that, fluvial activity in the LEB seems to have declined markedly. There was some renewed activity at about the time of the LGM (~28 - 20 ka) and the Early to Mid Holocene (~12 - 4 ka), but these were minor episodes compared to that in MIS 5, and in MIS 4 - 3. In summary, the rivers of the LEB suggest pronounced wetness in mid to late MIS 5, considerable drying since then but with the last period of significant sand transport in the Channel Country having been in MIS 4 - 3, with the basin becoming particularly arid after about 50 - 40 ka.
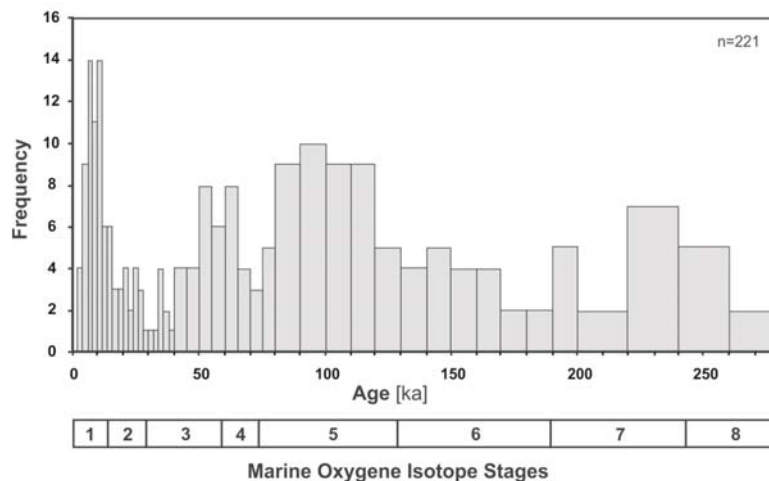

Figure 6: Combined age histogram showing the frequency of TL samples dated from the LEb and northern Australia (from Nanson et al., 2008). Note: given age uncertainties increase with age, preservation of material and probability of sampling material declines with age.

In a study looking at the palaeochannels, source-bordering transverse dunes and superimposed linear dunes at Gidgealpa on the Cooper Creek Fan (Fig. 2b), Cohen et al. (2010) found a similar story for the last full glacial cycle. Source-bordering dunes formed there during a period of enhanced river flow and sand supply from ~120 to ~100 ka, with another short episode of the same at ~85 - 80 ka and from ~68 to ~50 ka. At Chookoo on Cooper Creek just south of the Shire Road

(Fig. 7), river and associated source-bordering dune activity continued until ~40 ka (Maroulis et al., 2007). At Gidgealpa in the LGM the fan was laterally active from 28 to 18 ka and supplying sediment from channels to dunes. There is no sign of dunes being supplied with sand from active Cooper Creek channels since the LGM.
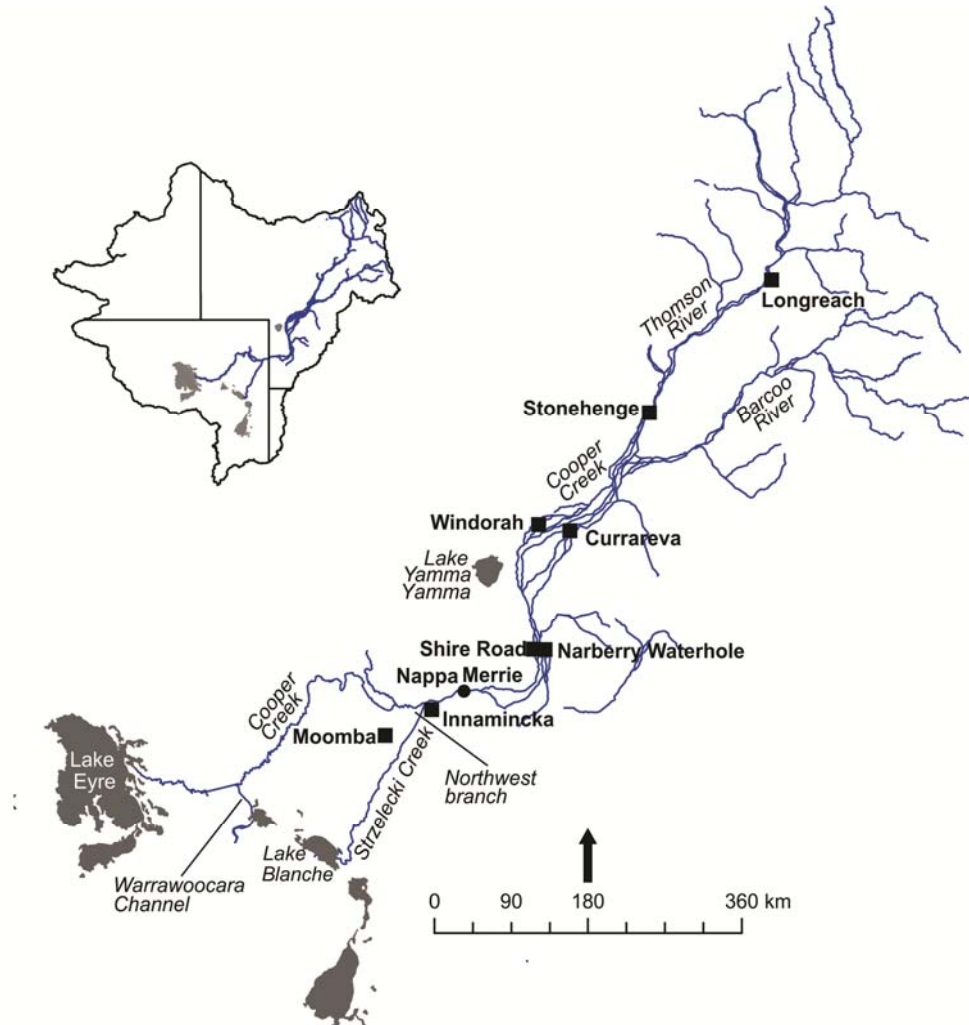

Figure 7: Cooper Creek extends from the confluence of the Thompson and Barcoo Rivers to Lake Eyre in the eastern Lake Eyre basin.

Maroulis et al. (2007) have identified pronounced periods of fluvial activity at Chookoo, represented by abundant sandy alluvium (Katipiri Formation) from MIS 8 to MIS 3. These sandy fluvial episodes on Cooper Creek were much more powerful than anything subsequent, and Maroulis et al. (2007) and Nanson et al. (2008) suggest that they appear to be ranked in order of declining activity, as if the LEB was becoming progressively drier with each flow-regime cycle. MIS 8 - 6 saw reworking of almost the *entire* floodplain at the Shire Road (Fig. 8), whereas subsequent phases of reworking were less extensive at the same location, even in MIS 5. The bedload sands are indicators of much more competent rivers than today during the Mid to Late

Pleistocene, systems capable of transporting abundant bedload now buried under extensive muddy floodplains. But such rivers must have been strongly seasonal, with a component of their bedload blown northward from periodically dry channels to create transverse-source-bordering dunes which have acted as the sediment source for the formation of adjacent linear dunes.

Source-bordering dunes are evidence for seasonal fluvial activity rather than aridity, as they form when rivers are sufficiently active to supply abundant sand as bedload, something not occurring in the LEB today. They were blown from active sandy channels between late MIS 5 (85 - 80 ka) and mid MIS 3 (50 - 40 ka), with some activity of the Cooper Fan at Gidgealpa at about the LGM (Cohen at al., 2010). The floodplain dunes at Chookoo show remarkable resilience, having survived without significant fluvial erosion or any downwind migration for at least 40 ka (Maroulis et al., 2007).

Research by Maroulis et al. (2007) and Cohen et al. (2010) indicates that the main periods of source-bordering dune activity correspond to periods of cool and probably seasonally wet-dry conditions between MIS 8 and 2. Page et al. (2001) found much the same for source-bordering dunes deposited during the last full glacial cycle beside the Murrumbidgee River on the Riverine Plain in NSW. It is likely that there were older episodes of source-bordering dune activity in the LEB but this evidence has been largely eroded or buried.
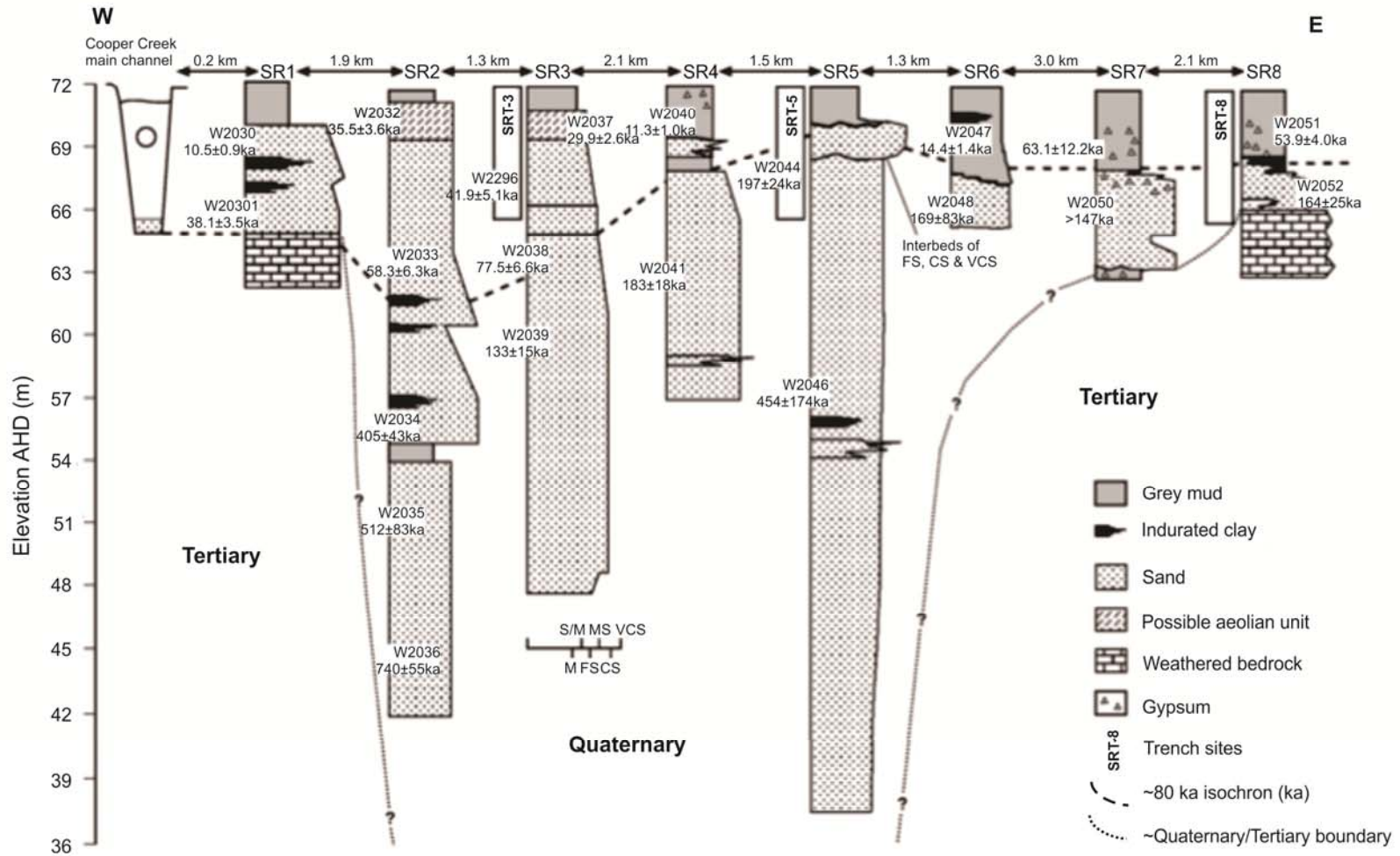
Figure 8: Cooper Creek floodplain stratigraphy in proximity to the Shire Road (from Nanson et al., 2008).

5.2 Palaeolacustrine evidence

5.2.1 Lake Eyre

Evaporites dominate the Quaternary sequences of Lake Eyre, with those in Madigan Gulf, the largest of three embayments located at the southern end of Lake Eyre North, the subject of a comprehensive study by Magee et al. (1995). These authors inferred the depositional environments, including deep- and shallow-water lacustrine environments and dry or ephemerally flooded playa environments, from eight cores and a cliff exposure. Coupled with numerous geochronological techniques, they reconstructed the last 130 ka of Lake Eyre palaeohydrology.

Magee et al. (1995) and Magee (1997) studied four sites in detail along the shoreline of Madigan Gulf concentrating on Williams Point (the most thoroughly described and dated stratigraphic section on the lake), a 15 m high cliff section ~10 km west of Goyder Channel in the southern portion of the gulf (Fig. 9). From a suite of ^{14}C, U/Th, AAR and TL ages on the cliff section of Williams Point and approximately 8 m into the lake floor (core 83/6), Magee et al. (1995) chronologically constrained their lake sequence depositional environment interpretations (Fig. 9). It is important to note Williams Point is on the upwind margin of the lake and the associated dune does not show typical lunette morphology, so it could not be used to define deflation events on the lake (Magee et al., 1995).

The Magee et al. (1995) reconstruction of lake levels has since been revisited and refined (DeVogel et al., 2004; Magee and Miller, 1998; Nanson et al., 1998, 2008; Magee et al., 2004; Cohen et al., 2012a) and over the past decade there have been varied proposals explaining Late Quaternary climate change in the LEB. Magee et al. (2004) produced a lake-level curve for Lake Eyre for parts of the past 120 ka (Fig. 10a). They suggest that the wettest periods within the past 120 ka occurred within MIS 5e and 5d (~128 - 110 ka) when the lake was permanently full (about 25 m above the present lowest point of lake floor), with another somewhat less but prolonged wet phase during MIS 5b and 5a (~100 - 85 ka; about 20 m) and a much less wet phase in mid to late MIS 4 (~68 - 60 ka; about 12 m).
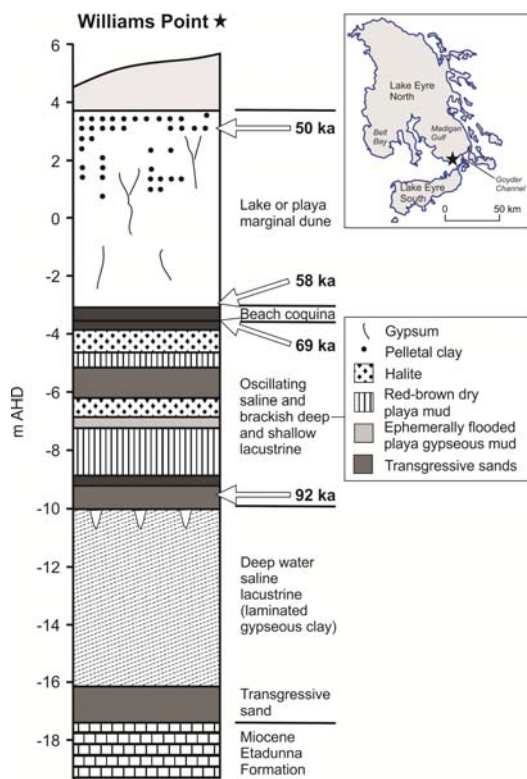
Figure 9: Williams Point cliff section and core 83/6 stratigraphy (Magee et al., 1995). This assisted in developing a palaeo lake-level curve that has since been revised and refined (Magee et al., 2004; Fig. 10a of this paper) and recently compared to lake-level phases obtained from Lake Mega-Frome (Cohen et al., 2012a; Figure 10b of this paper).
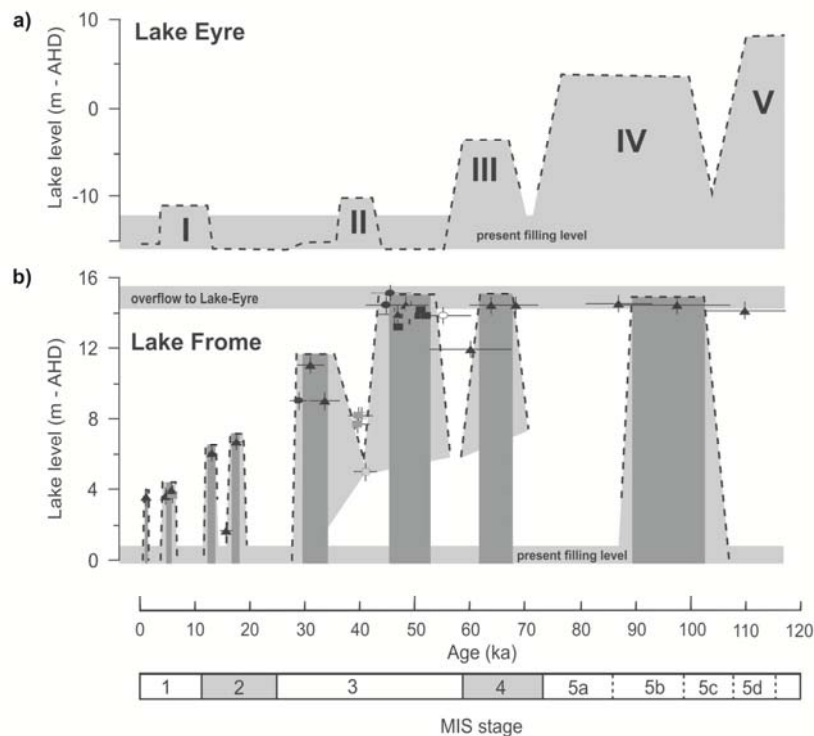

Figure 10: From Cohen et al. (2012a). a) The five Lake Eyre filling phases proposed by Magee et al. (2004). b) Lake Mega-Frome lake level phases (adapted from Cohen et al., 2011). These phases are based on single-grain OSL determinations on palaeoshoreline deposits (solid black triangles), multi-grain TL (solid black circles) and calibrated freshwater molluscs AMS ^{14}C ages (solid grey squares, black solid squares beyond calibration).

For comparison, maximum fillings today reach just 5 m above the present lowest point (i.e. reach ~ -10 m AHD), with the largest filling of modern times occurring in 1974. All major watercourses

draining to Lake Eyre flooded, and the lake subsequently reached its highest recorded level of -9.1 m AHD or ~6 m maximum depth (Dulhunty, 1978).

During MIS 6, and between each of the wet phases listed above, the lake was dry and has remained largely so or at levels below 5 m since deflation lowered the lake floor by more than 4 m below the present floor (Magee et al., 1995; Magee and Miller, 1998). It is argued that during early stages of MIS 5, the playa was 25 m above the modern playa (Magee et al., 2004). This record by Magee et al. (1995, 2004) is supported to some extent by his Williams Point evidence (Fig. 9) that suggested that a full lake evidenced by undated deep water clays deposited prior to 92 ka (but not as late as 85 ka), then a period of oscillating saline and brackish shallow and deep water phases to 70 - 60 ka. However, more recent work has identified problems in terms of when the lake was last at the height of the 5 m shoreline. A recent review of current literature on Late Quaternary climate change in central Australia by Couri (2011) reinterprets the evidence of extensive lake floor clays in Madigan Gulf used by Magee et al. (2004) to represent a high lake stand coincident with periods of a strong interglacial monsoon. Couri (2011) suggests that they may have instead resulted from distal sediment deposition in quite shallow water (e.g. from variable flood inputs from the nearest rivers such as Frome Creek). Cohen et al. (*in press*) are currently revisiting the question of Late Quaternary Lake Eyre levels with a team of researchers and their work suggests that the lake was last at ~10 m AHD periodically throughout MIS 5 and 4 and again at 5 m AHD between 68 - 47 ka, whereas the Magee et al. (2004) model has the lake last this full at about 80 - 75 ka. Lake Eyre periodically throughout Marine Isotope Stages 5 and 4 and again at 5 m AHD between 68 - 47 ka, whereas the Magee et al. (2004) model has the lake last this full at about 80 - 75 ka. Lake Eyre at +5 m at 68 - 47 ka would have been a vast water body at about the time humans arrived on the continent, with the lake presumably having dried out almost immediately afterwards. From this most recent work it is understood that Lake Eyre has not approached anywhere near these ~10 m or even ~5 m levels since ~50 ka.

5.2.2 Other terminal playas

Palaeoshorelines indicate clearly that Lakes Frome, Callabonna, Blanche and Gregory were formerly one megalake (Nanson et al., 1998; Cohen et al., 2011, 2012a), Lake Mega-Frome. This has palaeoclimatic implications and as such contributes to the ongoing debate of why the megafauna became extinct (see Section 4.5). Cohen et al. (2011) present a palaeoshoreline chronology for the past 100 ka and beyond for Lake Mega-Frome, which was subsequently compared to the lake level curve of Magee et al. (2004) by Cohen et al. 2012a) (Fig. 10b). The plot

of OSL/TL and AMS ^{14}C ages according to absolute elevation and previous Eyre lake-level record indicates that multiple filling episodes of differing magnitude occurred over the last glacial cycle (Cohen et al., 2012b). The periodicity of flood events is further indicated by the combined age histogram for northern Australia and LEB (Nanson et al., 2008) (Fig. 6).

Using OSL dates obtained from the 15 m shoreline, during mid to late MIS 5, the megalake was up to 15 m deep and covered ~6500 km^2 (OSL ages with a pooled mean of 96 ± 7 ka, Cohen et al. 2011) and was connected to Lake Eyre by the Warrawoocara Channel (Fig. 7) (Nanson et al., 1998). Evidence from Lakes Callabonna and Frome suggests that the megalake again reached this 15 m level in MIS 4 and MIS 3. OSL/TL ages range from 68.2 ± 4.2 ka to 60.1 ± 7.6 ka and again at 48.4 ± 4.2 ka to 45.4 ± 3.6 ka (Cohen et al., 2011). This is apparently the last time Lake Mega-Frome filled sufficiently to overflow to Lake Eyre.

The next dated high-stands of the megalake are shown on Figure 10b and reveal episodic but progressively lower high-stands until the Late Holocene. It also appears that the lake held a large volume of water (a remarkable 10 - 12 times that provided by the 1974 flood) during the Medieval Climatic Anomaly (MCA) at 0.96 ± 0.07 ka (Cohen et al., 2012b; Fig. 10b). The evidence from Mega-Eyre and Mega-Frome suggests that the last time there was sufficient water to fill them substantially was ~50 ka for the former and ~45 ka for the latter. Since Mega-Eyre appears to have not exceeded about -9 m (not very different to current fillings) and Mega-Frome has progressively declined. This would suggest a period of increasing aridity within the catchments of these lakes after about 55 - 45 ka.

5.3 Palaeoaeolian evidence

5.3.1 Dunes

Significant aridity and dune initiation in the LEB appears to have been at ~1 Ma when the duration of glacial cycle periods increased from 41 ka to 100 ka (Fujioka et al. 2009). These arid phases were almost certainly associated with significant aeolian activity (Nanson et al., 1992a, 1995; Hesse et al., 2004; Maroulis et al., 2007; Fitzsimmons et al., 2007a,b). While Fujioka et al. (2009) obtained minimum OSL ages of > 175 ka and cosmogenic age-estimates of about 1 Ma for dunes in the western Simpson Desert, to date, no dunes dated older than 116 ka, or MIS 5 have been found in the Strzelecki or Tirari deserts. Fitzsimmons et al. (2007b) state dunefields almost certainly existed prior to MIS 5, however, reworking in the last glacial cycle resulted in a lack of preservation.

Nanson et al. (1988) dated dunes older than 250 ka in the Diamantina valley, for the old dunes have probably survived relatively unaltered near the desert margins where their aeolian reworking is less likely. Based on OSL dates from five sites across the Strzelecki and Tirari deserts, Fitzsimmons et al. (2007b) suggest that there were four main linear dune building periods: 73 - 66 ka, 35 - 32 ka, 22 - 18 ka and 14 - 10 ka, which may represent enhanced aeolian activity as a function of intensified aridity.

Nanson et al. (1992b, 1995) and Hollands et al. (2006) have shown these linear dunes to be remarkably stable in terms of lateral or down-wind migration. In recently completed but as-yet unpublished work by one of us (Nanson) in the centre of the Simpson dunefield, there is no decline in TL ages from the upper 5 m of dunes collected at seven locations along an ~ 240 km down-wind transect to the north, as might have occurred if the linear dunes had been migrating and prograding northwards. Furthermore, there is no significant change in TL age in the upper 10 m along an ~300 km long east-west transect, as might have occurred if the dunes on one side or the centre of the dunefield were more susceptible to reworking and therefore younger than elsewhere. While the linear dunes of the Simpson Desert have crests commonly between 10 and 20 m and occasionally over 30 m above the swales, the dunes themselves are usually not indurated and are mostly little more than Holocene in age in their upper 10 m. They are commonly < 100 ka near their base. However, the depth of sand in the swales is in places over 8 m, usually indurated with carbonate and sometimes with mud, and yields TL ages over 150 ka, and from the work of Fujioka et al. (2009) it appears that the basal sand units beneath both swales and dunes may well be over 1 Ma. The Simpson Desert represents a very large and remarkably stable body of sand despite the upper parts of the dunes clearly being geologically recently reworked by wind. Sand extends well below the floor of the swales suggesting that the total depth of sand is surprisingly thick.

When the best available aeolian chronologies for the central deserts of Australia are compared, Cohen et al. (2010) demonstrate considerable intra-dunefield variability. Substantial variation in morphological responses of linear dunes to the same climatic periods of the last full glacial cycle suggests that Australian dune chronostratigraphy is a poor indicator of palaeoclimatic conditions. This may be because some dunes have their sand supplied by rivers during wetter albeit strongly seasonal periods, whereas others have their sand reworked during arid periods, and hence it is difficult to separate these two quite different climatic signals. Furthermore, even for dunes well removed from a river or lake, it is difficult to know if an abundance of aeolian sand deposition represented a period of enhanced aridity and sand reworking, or the period following when the re-

establishment of vegetation would encourage sand entrapment and dune stabilisation. In summary, it is unlikely that dunes alone provide a reliable palaeoclimate proxy.

5.3.2 Aeolian dust

Petherick et al. (2008) presents a high resolution multi-proxy record of environmental conditions between 33 and 18 ka from a lake core collected from North Stradbroke Island, off southeast Queensland. This is extended to 0 cal. ka by Petherick et al. (2009). Their dust transport pathway models suggest much of the dust was sourced from within the LEB. Such determinations are possible using trace element fingerprinting; for example, the LEB is characterised by low quantities of the highly to moderately incompatible elements and the Murray River region a relatively high abundance of these same elements (Marx et al., 2005).

The North Stradbroke Island lake-core lends support to the belief that the LGM in the LEB was associated with considerable aridity. Another record of Australian dust deposition extracted from a core from a peat bog in Central Otago, New Zealand, indicates that < 900 cal. BP dust to this location was again predominantly supplied by the LEB (Marx et al., 2009). Based on available information, McTainsh (1989) suggests that the seven recorded dust transport events to New Zealand that occurred from 1900 onwards were entrained from the lower LEB and/or the Murray-Darling Basin. Dust storms in the Channel Country take place more than three times a year (McTainsh and Pitblado, 1987). Moderate Resolution Imaging Spectroradiometer (MODIS) data offers a mechanism to determine the source of these dust plumes. Between 2003 and 2006, such plumes within the LEB itself originated from aeolian deposits (37%), alluvial deposits and floodplains (30%) and ephemeral lakes or playas (29%) (Bullard et al., 2008). Using this and similar data from other major dust source regions, Bullard et al. (2011) have proposed a scheme that classifies geomorphic types and their behaviour as dust sources. Should it prove more widely applicable, it represents an improved characterisation of preferential dust sources in global dust-cycle models.

5.4 Section 5 summary

At the end of the Pliocene, Australia became particularly arid with the stony deserts of the LEB starting to form at about 2 - 4 Ma. While there was progressively increasing aridity overall, the LEB has oscillated from wet to dry in response to climate changes linked broadly to the global glacial cycles. Rivers and lakes became more active only to have repeatedly dried again, and the dunefields

oscillated between more then less dynamic in response to these changes in moisture regime. Source-bordering dunes were fed when the rivers were particularly seasonally active. While there were prolonged climatically-induced periods prior to about 50 ka when the LEB rivers flowed and the lakes were full, it does appear that the basin has generally been much more arid since about 45 ka, albeit with possibly some enhanced fluvial activity during the LGM. Evidence for the alternation between wetter and drier episodes from 40 to 0 ka presented in the sections above is summarised in Figure 11. Fitzsimmons et al. (2013) compare this tropically-influenced northern Australia data to that from southern Australia, which instead experiences winter rainfall dominance connected to the westerly cold fronts. The compilation indicates that while wetter and drier episodes were experienced continent-wide, their timing and extent requires considerable additional research, including in the LEB.
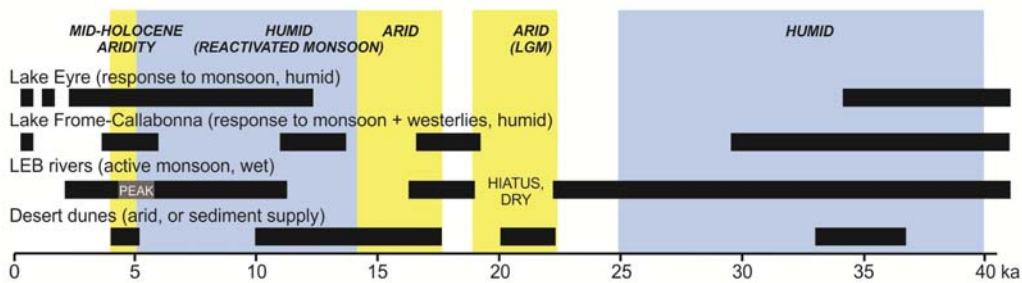

Figure 11: Palaeoenvironmental change in the Lake Eyre basin over the last 40 ka (modified from Fitzsimmons et al., 2013).

PART C: CONTEMPORARY ENVIRONMENTAL CONDITIONS

6 Hydrology

6.1 Rainfall

Hydrometric instrumentation in the LEB is minimal. Of the 103 Australia-wide Reference Climate Stations (BOM, 2007), only six are in the LEB. Consequently, rainfall records for inland Australia are imperfect (Kotwicki and Allan, 1998). However, Larsen (2011) indicates that more than 200 mm of localised rain in a 72-hour period is required to fall before large flows will connect many of the channels and floodplains in the Strzelecki, Tirari and Simpson Desert dunefields.

The great filling of Lake Eyre in 1974 is attributed to annual rainfalls of 700 mm in the eastern and 500 mm in the western regions of the basin (Kotwicki, 1986). Ten years later in 1984, the lake was

apparently filled to one-third of its 1974 storage in just few days following a rainfall event of 180 - 360 mm in the western part of the basin (Kotwicki, 1986).

Nearly half of the LEB receives an average annual rainfall of only 100 - 150 mm, with the highest annual averages of 400 mm occurring in the upper Georgina and Diamantina rivers and Cooper Creek (Kotwicki, 1986). Using 103 years of BOM data (1901 - 2003), McMahon et al. (2008a) found that mean annual rainfall ranges from less than 200 mm in the centre of the basin to 700 mm in the north and northeast. This variability of annual rainfall is the highest in Australia (Kotwicki, 1986), and the annual coefficient of variation is high (Knighton and Nanson, 1994a; McMahon et al., 2008a). Due to this high inter-annual variability of rainfall, drought intensity in the LEB during the dry periods is higher than expected for arid and semi-arid locations generally (McMahon et al., 2008a) and must be important for reactivating aeolian processes in the basin on decadal or longer time-scales. As yet, however, such data has not been analysed.

Large-scale ocean-atmosphere interactions (such as ENSO) appear to strongly influence rainfall patterns (McMahon et al., 2008a). Interestingly, and at a higher occurrence than in other arid locations around the world, La Niñas and El Niños often come in clusters: a high rainfall year in the LEB tends to be followed by another, and a low rainfall year tends to be followed by a second year of low rainfall (McMahon et al., 2008a). Analyses suggest more complex tropical and southern interactive synoptic conditions (Goodwin et al., 2013; Cohen et al., 2012b).

At times a well developed trough-like structure develops in the form of a northwest cloud band that extends southeastwards across central Australia, from a high pressure ridge in the northern Indian Ocean to a blocking high in the southern Tasman Sea. This produces high-intensity frontal rain over the central and southern part of the LEB. Rain can occur broadly across this region resulting from the combination of a tropically-fed and frontal Southern Ocean low pressure system over southern central Australia.

6.2 Runoff and streamflow

Rivers within the LEB experience extreme variations in discharge and flow duration, rendering them as some of the most variable rivers in the world (Knighton and Nanson, 1994a, 2001; Puckridge et al., 1998). Figure 12a shows that the flow duration curves for four rivers (Diamantina River at Birdsville, Thomson River at Longreach, Cooper Creek at Callamurra Waterhole near Innamincka (Fig. 7) and Todd River at Wigley Gorge northeast of Alice Springs) are steep

(McMahon et al., 2008b). The Cooper, Diamantina and Georgina systems originate in relatively humid headwaters with elevations of mostly < 400 m and they have long travel paths over gradients that rarely exceed 0.0002 (Nanson et al., 1988).
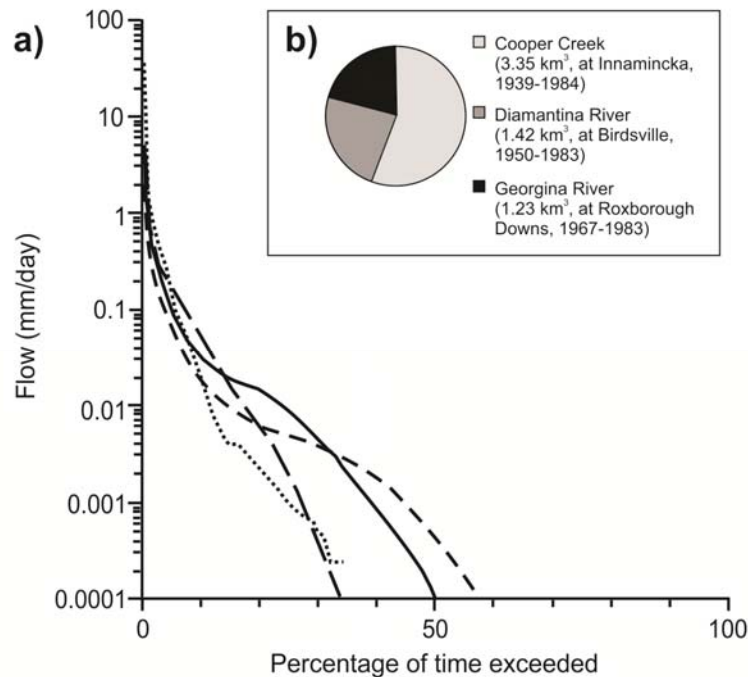

Figure 12: a) Flow duration curves for selected rivers (from McMahon et al., 2008b). Solid line represents the Diamantina River at Birdsville, long dashed line the Thomson River at Longreach, short dashed line Cooper Creek at Callamurra Waterhole and dotted line Todd River at Wigley Gorge. b) Estimated average annual flow that entered Lake Eyre between January 1974 and June 1976. Data from Tetzlaff and Bye (1978).

The rivers supplying Lake Eyre are capable of exceptionally large flows, but only very infrequently. They are not gauged as they enter the lake, however, Kotwicki (1986) has estimated that on average each year, Cooper Creek yielded about 3.35 km^3 between 1939 and 1984 (at Innamincka), the Diamantina 1.42 km^3 between 1950 and 1983 (at Birdsville) and the Georgina about 1.23 km^3 between 1967 and 1983 (at Roxborough Downs). Tetzlaff and Bye (1978) estimated that from numerous sources 56 km^3 entered Lake Eyre between January 1974 and June 1976 (Fig. 12b). Maximum daily discharges for Cooper Creek were recorded in 1974 with 25,000 m^3 s^{-1} and 5800 m^3 s^{-1}, Currareva and Nappa Merrie, respectively (Knighton and Nanson, 1994a). However, while inundating vast areas of interdune swales in the LEB, it is noted even the largest floods do not reach much above the base of the floodplain dunes (Maroulis et al., 2007).

It is suggested that over a three day period during 1984, and again in 1989, the western tributaries contributed ~8 km^3 to Lake Eyre (Kotwicki and Isdale, 1991). Knighton and Nanson (2001)

provided somewhat more up-to-date flow records for the basin than those by Kotwicki (1986) listed above, but of much the same order.

While Cooper Creek has a higher discharge relative to the Diamantina, in terms of volume and frequency of flow events recorded at Innamincka, midway along the river, most of the runoff into Lake Eyre is from the Diamantina River. This has largely to do with transmission losses. After Innamincka, the flow path of the Cooper comprises two major distributaries, a northwest branch and Strzelecki Creek (Fig. 7). Water diverted to the latter never returns to the Cooper, and thus never reaches Lake Eyre. A large proportion of the remaining flow following the northwest branch is lost into the adjacent dunefield, and therefore does not reach the lake. In contrast, the wetland and lake systems on the Diamantina do not divert or capture as much flow, but rather store it until being released again downstream. Akin to the Cooper, some flow is lost to groundwater and evaporation, however, the amount of discharge that remains in the channels causes floodwater to reach Lake Eyre on a more regular basis than for the Cooper. The Diamantina floods were also found to move up to four times as fast as those in the Cooper (Tetzlaff and Bye, 1978).

Approximately one flow every eight to ten years from the Cooper and Diamantina systems reaches the lake. Knighton and Nanson (1994a) have proposed three main causes for the transmission losses: (1) evaporation as the floods slowly head downstream; (2) infiltration to the soil, replenishing the soil moisture store depleted by evapotranspiration and possible minor percolation to shallow unconfined aquifers; and (3) drainage diffusion (pooling of floodwater), where floodwaters disconnect from the channels in low-lying areas and, therefore, further contribute to (1) and (2). However, as discussed below in Section 6.4, recent research (Cendón et al., 2010; Larsen, 2011) has recognised that a considerable amount is lost to the near surface aquifer due to scouring of the waterholes during floods. Hence direct evaporation in this arid region may not be as important as originally thought. The waterholes are self-sealing after floods and can hold water for several years without replenishment, even during severe droughts (Knighton and Nanson, 1994b; Hamilton et al., 2005), making them a vital source of water for wildlife and cattle farming. If evaporation were to be large, the solute load of Cooper Creek would increase significantly between Currareva and Nappa Merrie, but these values remain largely unchanged (Larsen, 2011). The bulk of the losses are more likely to be attributable to a combination of shallow aquifer recharge (e.g. Cendón et al., 2010) and flood routing into terminal wetlands (e.g. Costelloe et al., 2009). Evapotranspiration is another possibility to explain the losses, however, arid-zone riparian trees have low transpiration rates as an adaptation to limit water use (Costelloe et al., 2008), and probably

rely on shallow groundwater rather than floodwaters as their primary water source (Mensforth et al., 1994; Thorburn et al., 1994).

Transmission losses mean that the average annual runoff in Cooper Creek decreases from 96.7 $m^3 s^{-1}$ (3.05 km^3) at Currareva (Fig. 7) to 40.0 $m^3 s^{-1}$ (1.26 km^3) at Nappa Merrie (Fig. 7), some 420 km downstream, which equates to about 60% (Knighton and Nanson, 1994a). For flows above about 25% duration the average increases to about 75%, presumably due to losses resulting from overbank flows. Interestingly, the coefficient of variation for the daily data is greater than for the monthly data, which is greater than for the annual data, indicating that there is more variation within shorter time periods (McMahon et al., 2005). The highest mean monthly and daily flows occur on Cooper Creek at Currareva near Innamincka (McMahon et al., 2005) but, importantly, the Cooper for more than 400 km downstream of Innamincka to where it enters Lake Eyre is entirely ungauged and transmission losses through this section of abundant aeolian dunes and associated swales must be considerable.

The hydrology of the Diamantina system is less well understood than that of Cooper Creek. The mean annual flow at Birdsville, > 400 km upstream of Lake Eyre, is 1.45 km^3, whereas that in the upper reaches of the Georgina, at Roxborough Downs, which joins the Diamantina-Warburton downstream of Birdsville, it is 1.09 km^3 yr^{-1} (Knighton and Nanson, 2001). Nowhere downstream of Birdsville is the river gauged and, as with the lower reaches of Cooper Creek, the Warburton River and Kallakoopah Creek are greatly impacted by adjacent aeolian dunefields. Costelloe et al. (2003, 2006) have modelled flow regime and streamflow, respectively, on a 330 km reach between Diamantina Lakes and Birdsville. Transmission losses here are between 70 and 98% for floods with a total discharge at Diamantina Lakes of less than 2.3 km^3 yr^{-1} (total volumes > 2.3 m^3 are considered to reflect large input downstream of Diamantina Lakes due to heavier rainfall in the lower reaches during large floods), while non-linear relationships between flood size, timing and transmission losses at the reach scale are a function of variations in roughness and flow path length (Costelloe et al., 2006).

6.3 Groundwater

The LEB has two almost independent hydrological systems: a surface water system linked to shallow alluvial aquifers and a deep artesian groundwater system largely isolated from the surface, the latter termed the Great Artesian Basin (GAB) (Fig. 1 inset) (e.g. Armstrong, 1990). The GAB comprises three constituent sedimentary basins - Eromanga, Surat and Carpentaria, with the

Eromanga Basin underlying much of the LEB, with aquifers in the Triassic, Jurassic and Cretaceous sandstones. Recharge occurs largely along the eastern margin of the Lake Eyre Basin and groundwater flows towards the southern, southwestern, western and northern margins where mound springs and diffuse discharge occur (McMahon et al., 2005). The springs are associated with faults and fracture zones which enable the water to flow to the surface (Habermehl, 1980, 1982; Boyd, 1990).

Using stable isotope data ($\delta^{18}O$ and $\delta^{2}H$), Tweed et al. (2011) found that most groundwater recharge occurs during the larger and more intense rainfall events with groundwater moving laterally at rates of ~5 m yr^{-1} near recharge zones and < 1 m yr^{-1} towards the centre of the GAB (McMahon et al., 2005). In other words, water discharging near the centre of the GAB has taken up to 2 Ma to arrive there from the eastern recharge zones.

By surface area, pastoralism is the largest user of the LEB (Ponder, 1986; McMahon et al., 2005), with some interspersed mining operations (e.g. numerous base and precious metals, natural gas, oil, petroleum, opals, coal, phosphate, gypsum and uranium). Great Artesian Bain water is the primary (and often the only) source of stock water (Habermehl, 1980, 1982; Ponder, 1986; Mudd, 1998) and water bore activities from the early Twentieth Century onward have resulted in a drop in the piezometric head (Habermehl, 1980). It is estimated that since the commencement of pastoralism and modern extractive industries in the basin flow from the springs has declined by ~30% (Herczeg and Love, 2007), and some are reduced to a trickle (Harris, 2002) or have ceased to operate (Ponder, 1986). This problem is now being addressed with the state, territory and federally funded Great Artesian Basin Sustainability Initiative, whereby most bores are capped and inefficient bore drains are replaced with pipeline reticulation systems (Department of Sustainability, Environment, Water, Population and Communities (DSEWPaC), 2012).

6.4 Salinity

Kotwicki (1986) estimated a median total dissolved salts concentration of 100 - 110 ppm in the Diamantina River at Birdsville and Cooper Creek at Innamincka, which is relatively low. Water that reaches Lake Eyre is commonly quite fresh, becoming salty once in the lake and interacting with existing brines (Larsen, 2011). When the lake surface water subsequently evaporates, enormous volumes of dissolved salts can be left behind (Armstrong, 1990). For example, in the seven months of 1974 that water flowed through the Goyder Channel between Lake Eyre North and Lake Eyre South, salinity concentrations ranged from 10,000 mg/L to 50,000 mg/L (Dulhunty, 1978). From

our observations it is readily apparent that groundwater recharge supplies considerable salt to reaches of the large river systems proximal to the lake. However, these locations are downstream of any regular water monitoring program. On this basis, saline groundwater must also be entering through the lake bed as well.

Mean TDS concentrations decrease and then increase heading down Cooper Creek towards Lake Eyre: 119 ± 8 to 124 ± 10 mg/L in the headwaters at Longreach; 90 ± 12 mg/L at Stonehenge; 103 ± 8 mg/L at Currareva; and 146 ± 6 mg/L at Nappa Merrie, with HCO_3 responsible for much of this change (45%, 73% and 73% between Longreach and Stonehenge (Fig. 7), Stonehenge and Currareva, and Currareva and Nappa Merrie, respectively, Larsen, 2011). While it may seem intuitive that evaporative processes would dominate the surface waters in such a dryland basin, as shown above, this appears not to be the case here (Larsen, 2011).

Recently, fresh shallow groundwater proximal to and supplied from Cooper Creek waterholes in its middle reaches upstream of the Innamincka Dome has been found to interact, most prominently during recharge episodes, with saline regional groundwater (Cendón et al., 2010). These authors discovered that a well-defined salinity gradient, with increased near-surface (~10 - 15 m) regional groundwater increasing in solutes with distance from waterholes that appear to supply fresh water to them. The groundwater recharge events probably occur at the base of the waterhole when scoured by floodwaters and, as confirmed by stable water isotopes (δ^2H and $\delta^{18}O$), are consistent with water supplied by monsoonal flooding events. Compared to less significant and rainfall events, water sourced by monsoonal flooding generally shows a more depleted isotopic signature (Cendón et al., 2010). Once discharges decline, settling mud reseals the waterholes and further leakage ceases.

6.5 Section 6 summary

Through geological time, the variability of water has played a crucial role in determining environmental change in the LEB, with water the region's most critically limited resource today. With rainfall commonly only 100 to 400 mm yr^{-1} and evaporation rates often more than an order of magnitude greater, water is clearly a key limitation for agricultural and pastoral development. The impact of large scale ocean-atmosphere interactions such as ENSO on the basin are poorly understood, although changes in both the Pacific and the Indian Oceans are increasingly realised to have a profound influence, often in in the form of successive years of wetness or drought (Reeves et al., 2013a,b). The near-surface groundwater is commonly too saline for stock use but in places there

is input from the fresh water of the rivers when they flow and, as discussed below, waterholes along the major river systems remain virtually the only long term stores of water for ecological and pastoral sustainability. The geomorphology of these rivers and associated waterholes is key to understanding their pivotal role in the hydrology of the basin of today.

7 Contemporary geomorphology

The land surface of the LEB today is a palimpsest of past and present processes, with many of the geomorphological features influenced by global climatic changes experienced throughout the Quaternary and earlier (Nanson et al., 1988, 2008; Alley, 1998). Rivers drain centripetally from the margins to a system of salty playas at or below sea level and as they do so they pass through extensive dunefields. Today, and more so in the past, sediment is and was exchanged between these fluvial and aeolian systems. Interestingly, relatively little sediment accumulates in the terminal lakes that could otherwise be considered the depocentres of this landlocked basin. It is worth noting, however, that the availability of fine sediment for such exchanges is reduced because approximately 15% of the LEB surface area comprises stony desert surfaces ('gibber'; see Section 7.4).

Early physiographic descriptions on parts of the LEB by Gregory (1906), Carroll (1944), Madigan (1946), Whitehouse (1948), King (1960), Jennings (1968), Mabbutt (1967, 1977), Brookfield (1970), Folk (1971), Twidale (1972), Wasson (1983a,b) and Wells and Callen (1986) to name a few paved the way to current understanding of the basin's contemporary geomorphology. It is this and later work that is summarised below.

7.1 Fluvial systems

7.1.1 Cooper Creek

Cooper Creek has been the focus of numerous geomorphological studies (e.g. Nanson et al., 1986, 1988; Rust and Nanson, 1986; Knighton and Nanson, 1994a, 1994b; Fagan and Nanson, 2004; Maroulis et al., 2007; Cohen et al., 2010), rendering it the most intensively described of the Channel Country rivers (Nanson et al., 2008) and one of the most studied in Australia. Its floodplain comprises a complex network of channels that transport mud and a minor sand load from its headwaters in central Queensland towards Lake Eyre. Anabranching channels, waterholes, splays, braided flood channels, palaeochannels and aeolian dunes characterise the mud-covered floodplain surface of the middle reaches of the Cooper Creek floodplain (Fig. 13) (Nanson and Tooth, 1999).
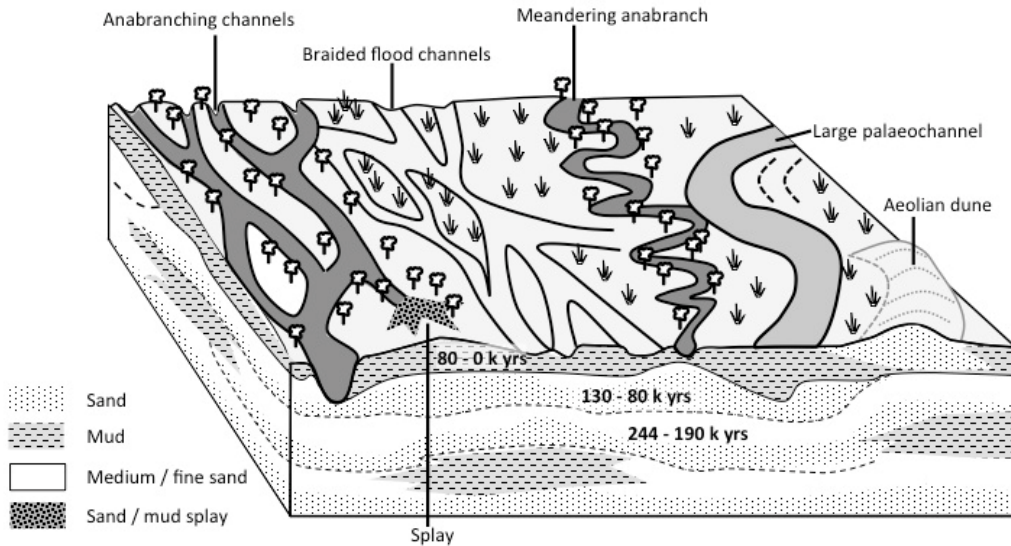
Figure 13: A schematic illustration of the landforms, palaeochannels, stratigraphic architecture and Quaternary chronology of the middle reaches of the Cooper Creek floodplain (modified from Nanson and Tooth, 1999).

Early research suggested that the complex surficial channel patterns may have been relict from the 'last glacial' (Whitehouse, 1948; Rundle, 1977) but Nanson et al. (1986) demonstrated that its two main channel patterns, low sinuous surficial braided floodplain channels and the more entrenched anastomosing channels (e.g. around Meringhina Waterhole, Fig. 14a), are contemporary and coexist, the latter active at low flow and both active during high flows. Sand, probably sourced from tributaries as well as scoured in waterholes from the sand sheet underlying the floodplain, is transported by the anastomosing channels during moderate to high flows, but not in abundance (Rust and Nanson, 1986). Rather surprisingly, copious mud is transported during overbank flow as *bedload* in the form of sand-sized mud pellets over extensive floodplains instead of being solely in suspension as it would be in most rivers (Nanson et al., 1986; Maroulis and Nanson, 1996).
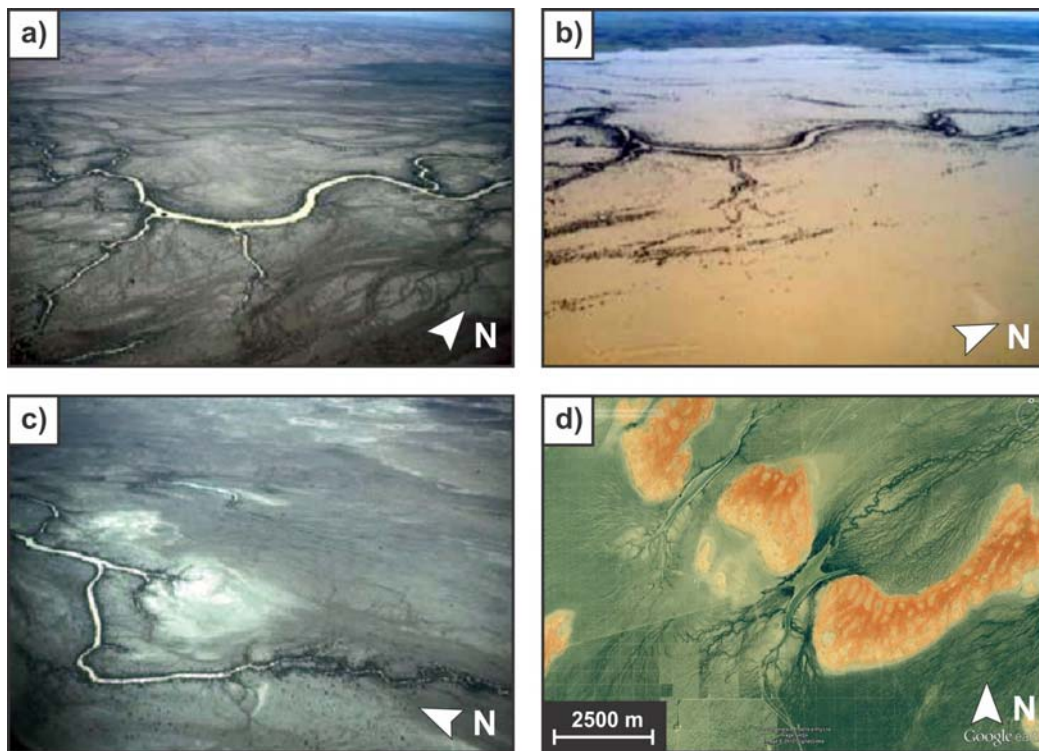
Figure 14: Waterholes on Cooper Creek (a), b), and c) are oblique aerial views, and d) a screen shot from Google Earth). a) Meringhina Waterhole (27°15'20"S, 141°58'30"E), with a dry floodplain. b) Meringhina Waterhole during the 1990 flood. c) Pritchilla Waterhole (27°09'02"S, 141°59'9"E) which shows two sediment splays, and how the bed of the waterholes can be scoured and basal sediment deposited on the floodplain. d) The two Tooley Wooley Waterholes (27°52'26"S, 141°50'29"E), showing the formation of waterholes when flood flows are confined between aeolian dunes on the floodplain.

Further evidence lending support to braided patterns of the floodplain surface being contemporary (Nanson et al., 1986) rather than relict (Whitehouse, 1948; Rundle, 1977; Rust, 1981; Rust and Legun, 1983) is provided by the relationship that exists between the three floodplain-surface flow patterns recognised along the middle reach of the Cooper between Windorah (Fig. 15) and Nappa Merrie (Fagan and Nanson, 2004): reticulate, braided and unchannelled. As floodplain width increases, so too does the extent of reticulate and braided patterns (however, the latter has a weaker correlation), while unchannelled areas appear to be confined to the widest areas of floodplain, which may be a function of low overbank flow power and reduced inundation frequency (Fagan and Nanson, 2004). Channel patterns on the floodplain surface and concomitant pedological characteristics are fundamentally linked. Where braided patterns dominate, the most common floodplain surface pattern type along the middle reach of the Cooper Creek, flood frequency is relatively high and gilgai development is minimal; the resulting channels are wide, shallow and only slightly inset (Fagan and Nanson, 2004). Reticulate floodplain surface patterns are generally associated with marked gilgai development and flood frequency is less. The gilgai encourages the

development of complex networks of ~right-angled confluences and bifurcated floodplain channels between gilgai mounds (Fagan and Nanson, 2004). Unchannelled areas of the floodplain occur where flood frequencies and surface irregularities are lowest; these are effectively restricted to the higher elevated floodplain surfaces (Fagan and Nanson, 2004).
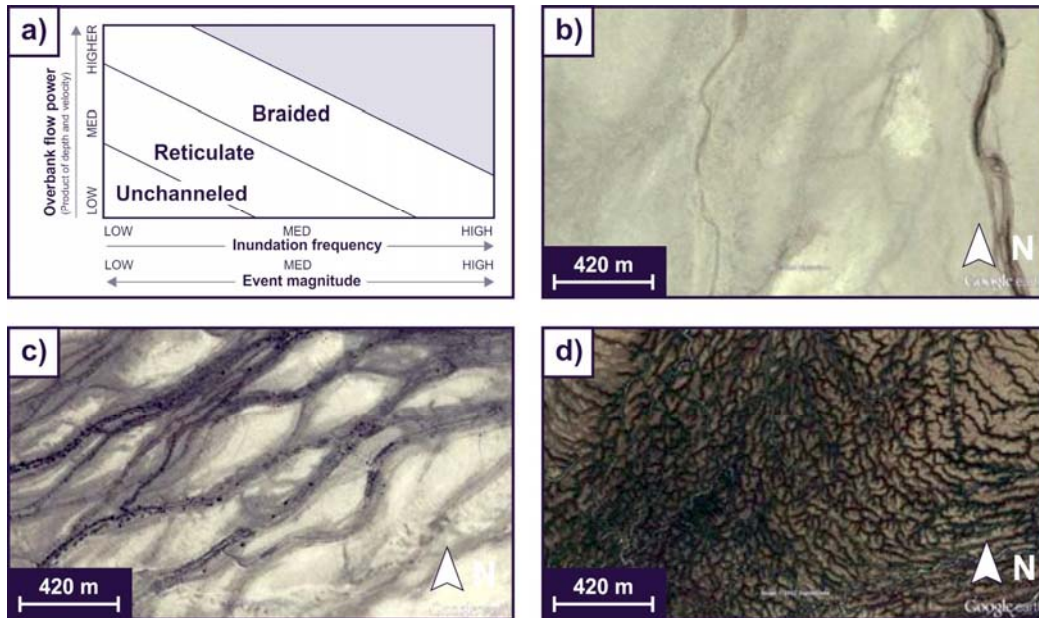

Figure 15: a) Continuum model of floodplain-surface channel pattern distribution models. The grey area occupying the top right corner designates that combinations of overbank flow power and event magnitude and inundation frequency were not found (from Fagan and Nanson, 2004). Examples of the patterns on the Cooper Creek floodplain at Wilsons Swamp, south of Ballera (screen shot from Google Earth): b) unchannelled floodplain (27°40'15"S, 141°59'00"E), c) braided flood channels (27°42'52"S, 141°43'00"E), and d) reticulate flood channels (27°50'32"S, 141°57'16"E).

Anabranching channels are often but not always sinuous and have a canal-like cross section, width-depth ratios of 4 - 60, low levees and with steep banks formed of cohesive mud commonly lined with coolabah trees (*Eucalyptus coolabah*) and lignum (*Muehlenbeckia* spp.). While there can be many such channels across the floodplain as a whole, there tends to be one dominant or main anabranch and numerous smaller ones.

The waterholes along Cooper Creek are an enigma; however, they occur in large numbers and are vital for the natural ecology and for pastoralism. They sometimes form at points where several anabranching channels converge to produce sufficient scour to form and maintain a waterhole within the muddy floodplain (e.g. Fig. 14a, c). They also occur where flow converges and scours between aeolian dunes on the floodplain (e.g. Fig. 14d), or where flow is concentrated along the bedrock valley side, for in these locations they have sufficient flow-energy to be self-maintaining

(Knighton and Nanson, 1994b, 2000). Some also occur at locations where there is no readily apparent explanation.

Waterholes are a remarkably prominent geomorphological feature of the Cooper Creek system; over 300 of them have been identified in the lower reaches between Windorah and Nappa Merrie (Knighton and Nanson, 1994b) and some examples are shown in Figure 14. They range in length from a few hundred metres to over 20 km, with typical widths of 20 - 100 m and depths of 6 - 10 m. Generally up to five times wider and three times deeper than their feeder channels, they provide foci for erosional energy at numerous discharges and are important in maintaining the stability of the multi-channel pattern (Knighton and Nanson, 2000). It is understood the majority of waterholes along the Cooper are a function of localised scouring processes during flood events (c.f. Knighton and Nanson, 1994b, 2000), however some, for example Kyearing Waterhole on the Paroo River, store water partly because of blockage by river sediments (Timms, 1999). On Cooper Creek, once scouring reaches the underlying sand sheet, erodibility increases (Knighton and Nanson, 1994b, 2000) and the waterhole deepens, often becoming an essentially permanent water body.

The permanence of water in the waterholes is largely a function of four factors: (1) depth; (2) frequency of inundation/inflow; (3) water-loss processes; and (4) groundwater interactions. The depth of the waterhole when the river stops flowing is the key determinant of its longevity (Costelloe et al., 2007). A cease-to-flow depth of 4 m has been suggested as the minimum for LEB waterholes to retain that water for a two-year period (Silcock, 2009).

Using fractional water loss and stage-volume relationships, Hamilton et al. (2005) arrived at a mean evaporative loss rate in depth of 2.1 m yr^{-1} for Cooper Creek waterholes, further extrapolating that these waterholes could dry to 10% of their bankfull volumes in 6 - 23 months. However, it should be noted that estimates were derived from sampling undertaken in 2002, when pan evaporation rates were more than 15% greater than the long term mean (Hamilton et al., 2005). Clearly, occasional and irregular flow pulses are important in sustaining aquatic refugia as well as stock.

The Cooper Creek waterholes are more frequent on the western side of the floodplain, which Knighton and Nanson (1994b) suggest may relate to the long-term migration of the channel system westward. Their evidence for this migration is fivefold: (1) the small to intermediate waterholes where the rate of excavation is surpassed by the rate of deposition (e.g. Narberry Waterhole, Fig. 7) are generally located on the eastern side of the floodplain; (2) signs of meandering palaeochannels, that Rust and Nanson (1986) took to indicate previous sand-dominated flow-regime phases, are

more clearly visible on the eastern side of the floodplain implying a shift of flow westwards and less reworking and greater preservation of these older channel-forms in the east; (3) the main flow path effectively lies on the western side of the floodplain; (4) the dunefield on the western side of the floodplain south of Windorah appears to have been partially invaded by anastomosing channels; (5) the tributaries on the eastern side of the valley between Windorah and the Shire Road (Fig. 7) are forming small fans onto the floodplain surface whereas the Cooper floodplain appears in contrast to be invading the tributary valleys on the western side; and (6) the Shire Road section (Fig. 8) shows the western side of the floodplain to be younger. In total, these suggest tectonic tilting of the Cooper valley westward but, given the apparently stable nature of the existing anastomosing channel pattern, any tectonism is occurring at a slow rate.

7.1.2 Diamantina River

The Diamantina River is even more remote from population centres than is Cooper Creek and consequently its geomorphology has not been studied as intensively. As part of a broad investigation of the Channel Country, Nanson et al. (1988) described the chronostratigraphy of the Western River, a headwater tributary, and the main channel near Diamantina Lakes, as well as some source-bordering dunes. As with Cooper Creek, a fairly consistent cover of 2 - 3 m of floodplain mud with a gilgai soil is underlain by a sand sheet of Late Pleistocene age. Typical of the Channel Country, braided/anastomosing channels and wide floodplains characterise the lower and middle reaches of the Diamantina (Nanson et al., 1988; Costelloe et al., 2003; Habeck-Fardy, 2014).

7.1.3 Georgina River

The eastern portion of the Georgina catchment is essentially part of the Channel Country with mud-load rivers and many waterholes (Fig. 1) (Box et al., 2008). It is similar in alluvial style to that of the Cooper and Diamantina, with intensely anabranching narrow and deep channels on extensive muddy floodplains. In contrast, the rivers in the western portion, such as the Sandover, Bundey, Marshall and Plenty (Fig. 1), have much sandier sediment loads and are consequently very different in form. Their channels are commonly straight, wide and shallow, with numerous reaches divided into ridge-form anabranches (Tooth, 1999; Tooth and Nanson, 1999; Tooth, 2000a,b; Habeck-Fardy, 2014).

Upstream of the confluence with the Bundey River, the Sandover River is largely a wide single-thread channel transporting medium to coarse sand bedload. In contrast, much of the Sandover-

Bundey River (as it is named below their confluence) transports a coarse sand and granule bedload and is anabranched in the form of long ridges that divide the overall channel into a system of long, relatively narrow and linear sub-channels (Tooth, 2000b). River red gums occupy and act to stabilise the ridges. This difference in bedload texture appears to have created two very different river styles, especially downstream of tributaries where more frequent irrigation of the main channel by tributary flows has encouraged the growth of trees on the commonly-dry channel bed. In the Sandover-Bundey, ridge-form anabranching enables the movement of a coarse sediment load down a low gradient system partially obstructed by trees (Tooth and Nanson, 1999; Habeck-Fardy, 2014).

In this arid, low gradient and sandy part of the LEB, maintenance of continuous channels along each river can be a challenge. The Sandover River has developed a series of floodouts. The first one at Ammaroo Station (Fig. 1) spans 200 km^2 and is followed by a network of channels that carry any floodwaters to the confluence (Tooth, 2000b, 2005). Below the confluence, the distributary channels of the Sandover-Bundey disappear again at Ooratippra Station (Fig. 1), and are replaced by an ~800 km^2 terminal floodout. It is understood flows reach the intermediate and terminal floodouts on average every 3 - 4 and 1 - 2 years, respectively (Tooth, 2000b, 2005).

7.1.4 Marshall, Plenty and Todd Rivers

Anabranching is a common river planform in the LEB, and Tooth and Nanson (2000) have demonstrated how a small indigenous shrub, the inland teatree (*Melaleuca glomerata*), has enabled the formation and maintenance of anabranching channels in a reach of the ephemeral Marshall River. They suggest that the development of the ridges and islands appears to represent a continuum of channel change that is influenced by teatree growth in association with alluvial sedimentation. Initially, the channel is obstructed with abundant yet irregular stands of teatrees, which develop into an organised system of ridge-form anabranches.

Tooth and Nanson (2004) further described juxtaposed reaches of the Plenty and Marshall Rivers (Fig. 1), demonstrating that although channel-bed gradient, discharge and bank strength are essentially similar, bed-material calibre and the occurrence of periodic tributary flows and a coarser load have led the Marshall River to anabranch in numerous places while the finer sediment load in the Plenty has enabled that river to remain in a single channel. The result is a remarkable contrast in channel cross-sectional geometry and planform between these two adjacent rivers and illustrates the

important role of anabranching aided by riparian vegetation for altering channel form and enhancing the throughput of an imposed sediment load.

Bourke and Pickup (1999) in a study of the Todd River (Fig. 1) have shown that downstream variation in channel morphology, planform and sediments here is not necessarily gradual, but can instead be a series of step-wise changes. This is a function of the variable discharge regime, which leads to episodic landform development, and particularly abrupt morphological change at river confluences. Unlike many of the LEB rivers that terminate at the lake itself, the Todd River, like the Plenty and the Hay, dissipates amongst the longitudinal dunes and associated swales of the northern Simpson Desert (Bourke and Pickup, 1999).

7.1.5 Section 7.1 summary

The large rivers in the northeastern part of the LEB, although fed from low rocky uplands, have floodplains and channel morphologies dominated by cohesive clays derived from the weathering of sedimentary strata and small areas of volcanic intrusions. This fine-textured material is often transported as sand-sized mud aggregates rather than in suspension, creating a range of fluvial features largely restricted globally to this remarkable region known as the Channel Country. The rivers, dominantly anabranching and lined with coolibah trees, are characterised by waterholes that scour into Pleistocene sands at their base to release fresh water into the adjacent water table, before self-sealing with mud to retain their remaining water for several years of drought. The floodplains here exhibit complex channel patterns ranging from straight and meandering anabranches set within the floodplain, to surficial braided and reticulate channels, resulting in an intricate system in equilibrium with prevailing flow in different parts of these extensive surfaces.

In the northwest of the basin the rivers are dominated by sandy bedload and the channels are either single-thread or they exhibit ridge-form anabranching where riparian trees and shrubs play an integral role in forming and maintaining the ridges, systems very different in character to the mud-load rivers of the Channel Country.

7.2 Lacustrine systems

7.2.1 Lake Eyre

The modern Lake Eyre playa comprises two basins, the larger Lake Eyre North of 8030 km^2, and Lake Eyre South of 1300 km^2, connected by the narrow Goyder Channel. Once the water level at either end of the channel exceeds the highest point in the channel, water moves freely between them (Dulhunty, 1978). Together with Lakes Blanche, Gregory, Callabonna and Frome, it has been suggested that Lake Eyre is the remnants of a Pleistocene lake, Lake Dieri (David, 1932) approximately ten times larger in area than the modern Lake Eyre (David, 1932; David and Browne, 1950). Lake Dieri most likely drained via Lake Torrens into the sea at Spencer Gulf (David, 1932; David and Browne, 1950). The analysis of satellite imagery by Löffler and Sullivan (1979) lends support to a former large inland lake. They noticed that a series of pans are roughly parallel to the arcuate shape of the chain of lakes, which they interpreted to be a reflection of the shrinking of Lake Dieri.

Lake Eyre is normally dry over most of its area (Kotwicki, 1986) although Dulhunty (1990) indicates surface water (from rivers and/or surfacing groundwater) may exist in the deepest areas of the lake (Belt Bay or Madigan Gulf) for ~75% of the time. When floodwaters do enter the lake, they first travel in two long, straight and narrow grooves that extend from north to south across the northern half of the bed of Lake Eyre North (Warburton and Kalaweerina Grooves, Dulhunty, 1990).

Magee et al. (1995) and Couri (2011) found that the lake floor at Williams Point comprises Etadunna Formation overlain by transgressive sands, lacustrine gypseous clay, interbedded muds, halite and sands (Fig. 9). Lake Eyre is sodium chloride-rich, like most other salt lakes in Australia (De Deckker, 1983). Displacive groundwater gypsum sampled by Chivas et al. (1991) from Lake Eyre North and South registered sulphur isotopic compositions of +14.4 and +11.9 %$_o$, respectively, indicating airborne sulphate of marine origin dominates.

While Lake Eyre's Late Quaternary history has been intensively studied (Magee et al., 1995, 2004; Magee, 1997; Magee and Miller, 1998; Nanson et al., 1998) and was discussed in Section 5.2.1, its contemporary geomorphology has not been. Three ephemeral lacustrine sedimentation environments within Lake Eyre North are recognised (Dulhunty, 1982): in the north, a saline playa environment where salt crusts are absent; in the south, a terminal salina environment where salt crusts up to 0.5 m thick overlie gypseous sediments; and in between the two a saline flocculation environment or 'slush-zone' where muddy floodwaters coming from the north interact with the saline water originating from the southern salina.

Magee (1997) has argued that the lake floor has accumulated fine sediment at times during the Pleistocene, and this has periodically deflated, but as yet there are few specific details to support this. What is interesting is that only 6 m of Quaternary mud overlies the Miocene Etadunna Formation (Magee, 1997), indicating either that relatively little fine material has arrived in the lake since then, or that it has accumulated and periodically been deflated. Jansen et al. (2013) have suggested neotoectonics along the contributing rivers may have trapped alluvial sediment in sub-basins, such as that upstream of the Innamincka Dome, before it reached the lake. Nevertheless, the lake is certainly a source of dust under suitable weather conditions today (see Section 5.3.2). During large floods in the basin, such as occurred in 1974, beach ridges can form rapidly, with the one from that particular year still widely visible today and creating an interesting comparison to the substantially larger ones formed at higher lake levels during the Late Quaternary.

7.2.2 *Other terminal playas*

Lakes Frome (~2700 km^2), Callabonna (~500 km^2), Blanche (~760 km^2) and Gregory (~280 km^2) form a ~300 km long arc of playas around the northeastern margin of the Flinders Ranges (Fig. 1), and the relationships amongst these lakes during the Quaternary, as well as with Lake Eyre at that time, are complicated (Nanson et al., 1998; Cohen et al., 2011, 2012a). Similar to Lake Eyre, these smaller playas are dry most of the time, only filling during flood events, with most of the water believed to come down Strzelecki Creek rather than from the Flinders Ranges, although this has not been verified (Fig. 7).

A core collected from the floor of Lake Frome indicates the top 1.2 m comprises sands, followed by muds to 3.5 m depth, before sandy mud dominates, and all sedimentary units are associated with bands of gypsum, of various thicknesses (Bowler et al., 1986; Cohen et al., 2011). Sodium chloride ions dominate the brines extracted from Lake Frome (Draper and Jensen, 1976), with the concentration of chloride in the salt crust showing minimal variation from that of pure sodium chloride (Ullman, 1995), although minor, bromide concentrations also increase with depth.

During flows that exceed a stage height of ~9 m at Innamincka, water from Cooper Creek travels down Strzelecki Creek to Lake Blanche, which is also connected to Lake Callabonna by the Moppa-Collina Channel (Nanson et al., 1998). Lake Blanche can also overflow north to Lake Gregory (DeVogel et al., 2004; Leon and Cohen, 2012); during the 1974 flood, Lake Blanche significantly overflowed to Lake Callabonna with possibly a minor overflow to Lake Gregory (Kotwicki, 1986).

Interestingly, the salt crust on Lake Frome is only about 0.2 m thick and covers only ~6% of the lake surface (Allison and Barnes, 1985) whereas that on Eyre is about 0.5 m and very extensive (Dulhunty, 1982). The difference may suggest that Eyre is an older lake and has been accumulating salt for longer, or that the rivers feeding it were more saline. However, a more likely explanation is that Eyre has a larger catchment and has over its life received much larger total inputs of water. Furthermore, higher river discharges at various times during the Pleistocene (Section 5.1) would have dissolved the salts that accumulated in a greatly expanded form of Lake Frome and then flushed through to Eyre via the elevated Warrawoocara Channel (Fig. 7) to the ultimate depocentre, Lake Eyre.

7.2.3 Section 7.2 summary

Lake Eyre is the lowest point in Australia and, in combination with Lakes Frome, Callabonna, Blanche and Gregory, all at about 0 m AHD, they represent the termini of this enormous drainage system. Extensive sandy dunefields lie to the east of Lake Eyre and what was Mega-Frome, and sand-dominated lunettes lie north and northeast of Eyre, Frome, Callabonna, Blanche and Gregory, evidence of sand barriers associated with beach deflation at various time in the past. Drilling of the lake floors suggests not much more than 6 m of mud lies above a Tertiary base, begging the question as to whether a larger accumulation has at some time been deflated. However, no part of the Australian landmass north or east of these lakes contains significant thicknesses of loess or other fine grained aeolian material so, in contrast say to the deserts of central eastern Asia, it is unlikely that the lakes and rivers of the LEB have deflated substantial amounts of fine sediment.

The greater thickness of the salt crust in Lake Eyre, compared to that in the Frome system is probably due to Eyre being the recipient of a much larger volume of water over the long period of salt accumulation, but also to the periodic flushing of water through the Frome system to Lake Eyre via the Warrawoocara Channel during times of much higher discharges in the Pleistocene. It may also be due to Lake Eyre, as the deeper system, receiving a great input from saline groundwater.

7.3 Aeolian systems

7.3.1 Morphology

The Australian continent is characterised by a vast anticlockwise whorl of mostly linear siliceous sand dunes (King, 1960; Jennings, 1968), with ~40% of Australia's surface occupied by these (Fig.

16), making them the continent's most common landform assemblage (Mabbutt, 1977; Wasson et al., 1988). Although sometimes linked to the synoptic anticyclonic high pressure that commonly overlies the continent, Brookfield (1970) suggested the continuous whorl may rather be a coincidence of wind movements; the westerlies in the south and the trade winds in the north with connecting vortices to the east and west. However, the exact cause for such a regular pattern remains unclear. The processes responsible for the evolution of the Australian dunefields do appear though to have more in common with humid climates rather than those associated with hyper-arid deserts (Hesse, 2011).

The dunes in the LEB occur essentially in three large fields; the 130,000 km^2 Simpson Desert dunefield north of Lake Eyre, the 70,000 km^2 Tirari Desert dunefield east of Lake Eyre, and the 100,000 km^2 Strzelecki dunefield north and east of Lake Frome (Fig. 1, 16) (Hesse, 2011). Wasson et al. (1988) found that narrow crested linear dunes are the most common type in the LEB. These mostly siliceous and slightly clayey linear dunes are continuous for up to 250 km, 8 to 50 m high and, reflecting wind direction, usually have steeper eastern than western slopes (Wright et al., 1990).
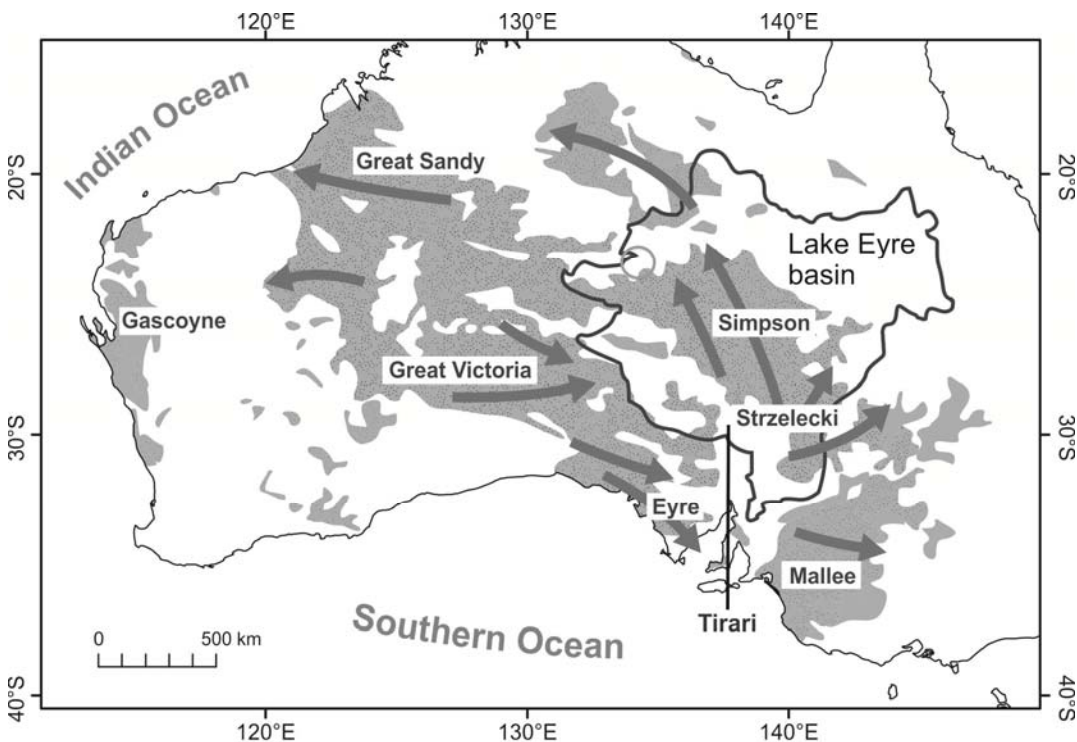

Figure 16: Australian dunefields and dune orientation mapped from Landsat TM imagery and location of Camel Flat basin (circle; adapted from Hesse, 2011).

Hesse (2010, 2011) has updated the mapping and classification of Australian dunes proposed by McKee (1979), Wasson (1986), Wasson et al. (1988) and Cooke et al. (1993), suggesting that none

of the previous work sufficiently represented the extensive morphological variations. Hesse's qualitative and morphological mapping from satellite imagery is based upon the orientation, continuity, connectedness, crest planform, crest sharpness, spacing and setting of the dunes. The resulting distribution pattern (Fig. 16) indicates wind directional variability is a dominant control; the centre of the dune whorl is largely associated with non-oriented mounds, which are then bounded by oriented network dunes, and longitudinal dunes occupy regions where dune orientation and sand transport direction is more constant (Hesse, 2011).

7.3.2 Sedimentology

It is recognised that the dunes of the LEB are derived from a large number of sources by different mechanisms, with the two primary colours of the Strzelecki and Simpson dunefields being pale brown and red-brown. In proximity to the major lakes and to the Diamantina and Warburton Rivers and Cooper and Strzelecki Creeks, the dune sands are derived from these water sources and are typically water-buffed and pale in colour. Distal to the lakes and major drainage lines (i.e. western Simpson Desert and southeastern Strzelecki Desert) the dunes are a red brown to orange colour and appear to be largely derived from the underlying weathered bedrock (Wasson, 1983a,b). Folk (1971) found that dune crests (2.5 ϕ) are coarser and more sorted than flanks (2.75 ϕ) and desert lag deposits (2.85 ϕ), largely because the crests comprise the most easily saltated fine sand materials. Studying the longitudinal dunes near Birdsville in the Simpson Desert, Nanson et al. (1992b) found no relationship between dune age and colour, at least in the upper half to one-third of the dune profiles, concluding colour is more likely to be a function of localised weathering and sediment transport history. Over two-thirds of Simpson Desert dune sand Folk (1971) investigated fell between 2.0 and 3.0 ϕ (fine sand), with a consistent very strong dark reddish-orange colour.

7.3.3 Sediment transport

Numerous researchers have assumed that most linear dunes grow linearly by downwind elongation (Tsoar, 1978; Livingstone, 1988; Rubin, 1990; Thomas, 1997; Tsoar et al., 2004), and in Australia, Twidale (1981), Wopfner and Twidale (1988, 2001) have proposed that in some cases, such as in the Simpson Desert, linear dunes may have extended many tens or even hundreds of kilometres downwind. However, using isotopes, Pell et al. (2000) have shown that the sands on the Simpson, Strzelecki and Tirari Deserts are derived mainly from local bedrock, with very little subsequent aeolian transport. Sediment transport from a wide range of proto-sources across the several hundred kilometres to their incorporation in the bedrock of various sedimentary basins was dominantly by

fluvial, not aeolian, means. Quaternary aeolian transport or reworking has also been minimal. In a detailed study of dunes at Camel Flat southeast of Alice Springs, Hollands et al. (2006) have shown that the field of linear dunes, located in a topographic basin between two ranges just 25 km apart, presented no evidence over the past 75 - 65 ka for sand having been moved any significant distance downwind; there has been no substantial ramping of sand against the windward side of the downwind ranges or thickening of the dunes in this direction. Rather, they appear to have accreted largely vertically, numerous small relatively closely spaced dunes being cannibalised over time to form larger more widely spaced dunes.

At Finke in the Simpson Desert (Fig. 1), Nanson et al. (1995) have shown pale-coloured source-bordering dunes blown northward from the Finke River have not encroached more than a few kilometres into the adjacent red-coloured regional dunefield over the past ~17 ka. Cohen et al. (2010) found the same for linear dunes near Cooper Creek, which almost certainly were supplied with sand from earlier source-bordering dunes, but the linear forms grew largely by vertical accretion with little northward migration over certainly the past 100 ka and possibly the past ~250 ka. Nevertheless, some linear extension has certainly occurred in places, perhaps by as much as a few kilometres, as is evidenced by linear dunes extending onto clay-pans and lake margins. Indeed, localised dune extension may have impeded the course of Cooper Creek in the Pleistocene, with palaeochannels winding through a combination of transverse and linear dunes on and downstream of the Cooper Fan (Cohen et al., 2010).

Most of the linear dunes in the LEB are not aligned with current wind directions (Hesse 2010, 2011) and despite this, they are not showing any sign of changing their orientation. Asymmetry of longitudinal dunes is an indication that recent wind directions have not exactly paralleled the dune trend, but have favoured a vector across the dune toward the steeper slope. It was suggested by Rubin (1990) that such asymmetry indicated that the linear dunes of the Simpson Desert could migrate laterally obliquely downwind, en echelon, over great distances. However, with luminescence dating, Nanson et al. (1992b) showed that there is no evidence such dunes have shifted laterally more than 100 m or so from their Pleistocene cores, and indeed not necessarily downwind. Upwind migration is probably characteristic of areas with relatively dense vegetation or periods of relatively humid climate when vegetation will trap mobile sand on the stoss faces, whereas downwind migration would be favoured on poorly vegetated dunes or during arid periods when sand will saltate up a largely unvegetated stoss slope, cross the dune crest and avalanche and accumulate on the lee slope (Nanson et al., 1992b). Just how dunes may be modified post-formation, or why they may resist modification, was recently explored by Hesse (2011), who

surmised that there is a range of modes in which a dune can modify following their initial formation: dune realignment to dominant sand transport direction with little lag; lagged transition between equilibrium states; superimposed generations of dunes; crest modification; and dune degradation by wind or water erosion.

Hesse (2010, 2011) suggests the stability of the Australian dunes may be a function of the low erodibility of the sand due to vegetation cover. Indeed, aeolian processes are most active during periods of drought when soil moisture is reduced and this vegetative cover is disturbed or removed and the potential for sand drift is higher (Burgess et al., 1989; McTainsh et al., 1990; Hesse, 2010, 2011). The majority of dunes in the Simpson and Strzelecki Deserts are associated with a 40% or greater coverage of vascular plants, including forbs, perennial grasses, shrubs and small trees (Hesse and Simpson, 2006). In an experimental quantification of the effect of sparse plant cover on dune sand transport, Buckley (1987) found that when wind velocity equals 10 m s^{-1}, a plant coverage of just 17% reduces the sand transport rate by 74%. Secondly, and less well understood, pedogensis cannot be discounted as another factor partially responsible for the stability of the dunes (Hesse, 2010, 2011).

7.3.4 Section 7.3 summary

Australia is dominated by an anticlockwise whorl of mostly linear siliceous and slightly clayey dunes less than 50 m in height and covering ~40% of the continent. In the LEB, this whorl is represented by the Simpson, Tirari and Strzelecki Deserts where the dunes trend mostly northward and have more gentle western and steeper eastern slopes. Across nearly all scales it appears that after initial formation, be it 1 Ma, 100 ka or 5 ka, the Australian dunefields are relatively stable. Dune migration in the LEB, linear or lateral, is minimal due to limited total sand being stored as the low and well-spaced dunes partially stabilised by vegetation and their alignment with the wind being such as to minimise down-wind transport. During the Pleistocene many dunes were formed as source-bordering along the rivers and as lunettes adjacent to the lakes, but there is little or no such source-related aeolian activity today.

7.4 Gibber surfaces

'Gibber' is an Aboriginal word now widely adopted in geomorphology as a term for a pebble-covered desert surface, also known in the Sahara as a 'reg', or as a 'gobi' in Mongolia and China. The Arabic term 'hamada' is now normally restricted to residual bedrock or boulder surfaces (Laity,

2011). Gibbers cover about 10% of Australia and are typically composed of a single layer of varnished pebbles (Mabbutt, 1977; Callen and Benbow, 1995). It is a common land-surface type covering near 15% of the LEB (Bullard et al., 2008), especially affecting the interfluves. Most gibbers in the basin are derived from silcrete, a resistant rock formed by pedogenic or groundwater silicification of sediment or regolith (see Section 4.2.1) (Thiry and Milnes, 1991; Benbow et al., 1995b; Simon-Coicon et al., 1996), and it occurs as a deposit on extensive pediments and fans flanking extensive linear escarpments (as well as isolated buttes and mesas etc.) and ranges. Not uncommonly, they overlie older alluvial surfaces and, regardless of their deeper substrate, often appear to 'float' as a pebble cover over stone-free clayey aeolian silt (Mabbutt, 1977). Once formed, the stony monolayer persists at the ground surface.

Laity (2011) thoroughly reviews the four main theories that have been proposed to explain gibber formation: (1) deflation leaving the surface stone layer as a lag; (2) concentration as a lag by heavy rainfall and sheetwash; (3) upward migration of stones to the surface by soil heave; and (4) upward displacement of stones by aeolian aggradation of fine material below the surface stones. In the Lake Eyre basin it has been deflation of silt and clay from gravelly sediment that has been argued to cause such pavements to accumulate (Litchfield and Mabbutt, 1962). However, a study in the Mojave Desert of California (Wells et al., 1995) favours theory (4) above. Using cosmogenic ^{3}He surface-exposure ages they indicated that clasts are continuously maintained at the surface in response to the deposition and pedogenic modification of windblown physically active dust between them. This expands with each wetting and repeatedly lifts the clasts, maintaining the surface pebble layer at all stages. This is very likely the process in the LEB as many of the gibbers sit on up to 2 m of what appears to be fine self-mulching aeolian sediment, completely free of pebbles. Nevertheless, lag surfaces formed by sheetwash or deflation also probably exist in particular locations where the gibber is not underlain by aeolian fines.

Regardless of their mode of formation, such surfaces are clearly persistent, having withstood the Pleistocene wetter episodes. Fujioka and Chappell (2010) have shown that older gibbers in the western LEB date from ~4 Ma with extensive areas dating at ~2 Ma. They have suggested that the presence of gibbers with post-exposure ages of ca. 2 Ma on a fan surface < 1 km from their valley-head sources suggests that dissection of the silcrete tableland to provide gibber pebbles has slowed in the past 2 Ma. The presence of such lags has profound implications for hydrology because it can prevent soil regeneration and increase net runoff (see review by Laity, 2011), however, the specifics of their hydrology have not yet been studied in the LEB.

PART D: CONCLUSION AND SUMMARY

The arid and semi-arid LEB of Australia has a long geological history as a series of basins extending in age well back into the Tertiary. This history, as well as the basin's contemporary features and agricultural potential, have intrigued explorers and scientists since soon after European settlement. The LEB offers a remarkable opportunity to understand continental climate change for it has a long and comprehensive alluvial, lacustrine and aeolian record of environmental change. The low gradients and various sub-basin repositories of retained depositional history have shown a progressive and cyclical shift from a substantially wetter environment, that must have been characteristic of much of Australia at times during the Tertiary and Quaternary, to the markedly arid conditions of today.

Approximately 70% of the LEB lies below 250 m AHD and while there is abundant evidence for significantly wetter conditions in the past, much of the basin is now covered with the aeolian dunes of the Simpson, Tirari and Strzelecki Deserts. In addition to these dunefields are extensive, dry rocky interfluves and wide infrequently-inundated muddy floodplains along the river systems in the north and northeast, and the sandier but equally ephemeral rivers in the west, north and northwest. Lake Eyre, one of the world's largest salt playas, with its deepest point at -15 m AHD, is infrequently fed by these rivers. The largest of these, Cooper Creek, the Diamantina River and the Georgina River, are capable of exceptional flows, but only infrequently, and although they typically run each year along at least part of their length, transmission losses are very high, commonly between 75 and 100%. Water variability has played a crucial role in determining environmental change in the LEB through geological time. The impact of global weather systems on the basin is still poorly understood, and while the monsoon has been usually named as the primary influence, changes in the western Pacific and eastern Indian Oceans are increasingly realised to have had, and still have, a profound influence.

The large muddy rivers in the northeastern LEB have floodplains and channel morphologies dominated by cohesive clays derived from the weathering of sedimentary and small areas of volcanic rocks. This fine-textured, self-mulching sediment is often transported in the rivers as sand-sized mud aggregates rather than in suspension. The remarkable mud-dominated sedimentology of these rivers and their seasonal hydrology sustain a major pastoral cattle industry dependent on a system of natural irrigation and fine-grained modestly-productive alluvial soils.

Waterholes on Cooper Creek are highly distinctive. They form at points in the mostly anastomosing channel system where sufficient flow-energy is concentrated for them to become self-maintaining, and they offer vital refugia for many of this arid environment's plants, fish, invertebrates and birds. Today they also offer a vital stock-water resource for the basin's pastoral industry.

While Lakes Eyre, Frome, Callabonna, Blanche and Gregory, represent the termini of this enormous drainage system, less than 10 m of mud overlies the Tertiary base of these lake floors. A lack of aeolian dust leeward of these lakes suggests substantial amounts of fine sediment have not been deflated. The very low relief of the basin, neotectonic activity partially damming the major rivers and trapping sediments in ephemeral 'swamps', the pelleted nature of the mud being transported as bedload rather than in suspension and extensive areas of dune swales flanking the eastern rivers and trapping floodwater in transit appear to have, in combination, led to relatively little mud actually reaching these terminal lakes. The floors of the lakes are mostly mud, so what sand has been deposited by the rivers appears to have been blown from the river beds before reaching the lakes, or on reaching the lakes is reworked as beach deposits and then into lunettes, with some of this transported by wind into the immediately adjacent dunefields.

Australia became particularly arid in the Late Pliocene and Early in the Pleistocene and its oldest dunes date from at least this time. The dunes of the Simpson, Tirari and Strzelecki Deserts trend mostly northward and, irrespective of their time of formation, it appears the dunefields of the LEB are relatively stable and hence not migrating, with their sand being locally derived either from adjacent lakes or dunes, or from the underlying weathered bedrock.

There is currently a lively debate as to the role of humans in modifying the environment of the LEB. The prevailing argument of late is that human impact, not climate change, was the cause of the extinction of the Pleistocene megafauna continent-wide, with an abrupt reduction in plant diversity (an ecological collapse) driven by a human-induced fire regime being the likely mechanism. However, convincing evidence of a direct human cause remains elusive. Firstly, no megafauna kill-sites have so far been discovered in Australia. Secondly, it appears that the period of extinction was also a period of very significant climate change, so climate cannot be so readily excluded. Thirdly, the oldest archaeological evidence for human habitation in the LEB is only for the period after 30 ka and evidence for their presence in reasonable abundance is only for the period after ~15 ka, both episodes well after the extinctions at 45 - 40 ka.

While our understanding of the geological evolution and related natural history of the LEB has been greatly advanced in the past several decades, it will almost certainly occupy the minds of scientists for many years to come as they strive to understand changes to one of the globe's great and unusual drainage basins. Important research is ongoing, for example, on the lake-level story in relation to a more precise picture of climate change, the role of humans and climate in the demise of the megafauna, modern hydrological changes, the role of vegetation in altering channel morphology and flow efficiencies, and nature of flood transmissions. Both government and industry are addressing land use and environmental issues associated with pastoralism, mining and tourism. The LEB remains one of the major regions of global drylands research.

Acknowledgements

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