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Supersequences, superbasins, supercontinents – evidence from the Neoproterozoic–Early Palaeozoic basins of central Australia

John F. Lindsay

JSC Astrobiology Institute, NASA-JSC (SA-13), Houston, Texas 77058, USA

ABSTRACT

Neoproterozoic sedimentary basins cover a large area of central Australia. They rest upon rigid continental crust that varies from c. 40–50 km in thickness. Whilst the crust was in part formed during the Archaean and early Palaeoproterozoic, its final assembly occurred at approximately 1.1 Ga as the Neoproterozoic supercontinent, Rodinia, came into being. The assembly process left an indelible imprint on the region producing a strong crustal fabric in the form of a series of north dipping thrusts that pervade much of the thick craton and extend almost to the Moho. Following a period of stability (1.1–0.8 Ga), a large area of central Australia, in excess of 2.5×10^6 km^2, began to subside in synchroneity. This major event was due to mantle instability resulting from the insulating effect of Rodinia. Initially, beginning c. 900 Ma, a rising superplume uplifted much of central Australia leading to peneplanation of the uplifted region and the generation of large volumes of sand-sized clastic materials. Ultimately, the decline of the superplume led to thermal recovery and the development of a sag basin (beginning at c. 800 Ma), which in turn resulted in the redistribution of the clastic sediments and the development of a vast sand sheet at the base of the Neoproterozoic succession.

The superbasin generated by the thermal recovery was short lived (c. 20 M.y.) but, in conjunction with the crustal fabric developed during supercontinent assembly, it set the stage for further long-term basin development that extended for half a billion years well into the Late Palaeozoic. Following the sag phase at least five major tectonic episodes influenced the central Australian region. Compressional tectonics reactivated earlier thrust faults that had remained dormant within the crust, disrupting the superbasin, causing uplift of basement blocks and breaking the superbasin into the four basins now identified within the central Australian Neoproterozoic succession (Officer, Amadeus, Ngalia and Georgina Basin). These subsequent tectonic events produced the distinctive foreland architecture associated with the basins and were perhaps the trigger for the Neoproterozoic ice ages. The reactivated basins became asymmetric with major thrust faults along one margin paralleled by deep narrow troughs that formed the main depocentres for the remaining life of the basins. The final major tectonic event to influence the central Australian basins, the Alice Springs Orogeny, effectively terminated sedimentation in the region in the Late Palaeozoic (c. 290 Ma). Of the six tectonic episodes recorded in the basinal succession only one provides evidence of extension, suggesting the breakup of east Gondwana at the end of the Rodinian supercontinent cycle may have occurred at close to the time of the Precambrian–Cambrian boundary. The central Australian basins are thus the products of events surrounding the assembly and dispersal of Rodinia.

INTRODUCTION

The present Australian continent consists of two major crustal components separated by the Tasman Line (Fig. 1) (see Murray *et al.*, 1989). East of the Tasman Line the crust is Cambrian or younger, the product of active margin processes during the Palaeozoic. West of the Tasman Line the crust is much older consisting of a complex of terranes of Archaean to Mesoproterozoic age (Rutland, 1976). In central Australia, a series of extensive intracratonic basins (Officer, Amadeus, Ngalia and Georgina Basins and possibly the Wolfe and Birrindudu Basins) of Neoproterozoic and early Palaeozoic age rests on this older crust. Regional evidence suggests that these basins are related through

Correspondence: John F. Lindsay, JSC Astrobiology Institute, NASA-JSC (SA-13), Houston, Texas 77058, USA. E-mail: john.f.lindsay@jsc.nasa.gov

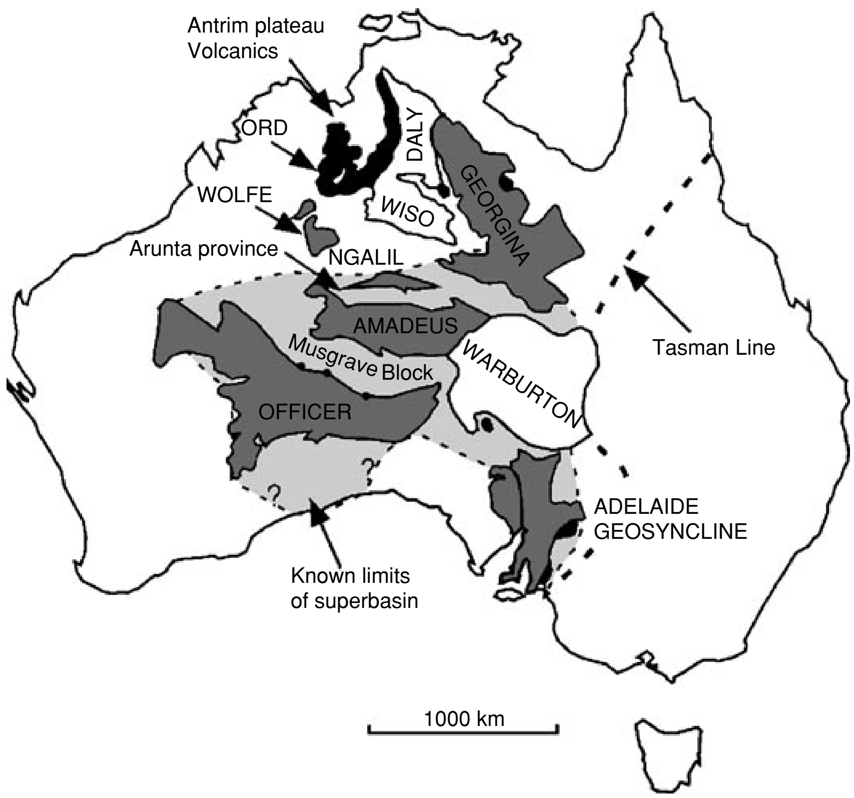

Fig. 1. Distribution of Neoproterozoic and early Palaeozoic basins on the Australian craton. Basins initiated in the Neoproterozoic are shown as dark grey, those initiated in the early Palaeozoic are white while the black areas indicate the distribution of basic volcanics (after Lindsay *et al.*, 1987). Light grey areas indicate the possible extent of the Neoproterozoic superbasin. Note the position of the Tasman Line. Recent work has suggested that the Wolfe and possibly the Birrindudu Basins may also be remnants of a Neoproterozoic basin (Grey & Blake, 1999; Grey, pers. Comm., 2001).

some large-scale mechanism, perhaps resulting from the accretion and ultimate breakup of the Proterozoic supercontinent Rodinia (e.g. Veevers & McElhinny, 1976; Lindsay *et al.*, 1987; Lindsay, 1993). However, while the basins appear to be related through some common causal mechanism, there is, to date, little agreement as to its nature. Suggestions have ranged from extensional settings (Korsch & Lindsay, 1989; Lindsay *et al.*, 1987; Lindsay & Korsch, 1989, 1991) to compressional and flexural settings (Lambeck, 1983; Ding *et al.*, 1992; Lindsay & Leven, 1996; Haddad *et al.*, 2001).

In this paper, I re-evaluate the central Australian basins in the light of recent improvements in the understanding of their internal architecture and underlying crustal structure.

CRATONIC ARCHITECTURE

The ancient Australian craton is complex and consists of a mosaic of crustal fragments with a broad range in age and degree of deformation. Crustal thickness beneath the central Australian basins is generally somewhat thicker than normal varying from approximately 40–50 km (Lindsay & Leven, 1996; Lambeck & Penny, 1984; Lambeck *et al.*, 1988; Korsch *et al.*, 1998). The craton has been divided into eight crustal mega-elements (Shaw *et al.*, 1996), four of which underlie the central Australian basins; the Southern Australian, Western Australian, Central Australian and Northern Australian Mega-elements (Fig. 2). The development of these mega-elements occurred largely in the Palaeoproterozoic but amalgamation of the mega-elements to form the Australian craton extended into the Mesoproterozoic.

Crustal evolution

The mega-elements aggregated in two major stages, both apparently associated with supercontinent assembly. There is a considerable body of evidence to suggest that by 2.0 Ga a supercontinent, Columbia, similar in significance to Pangea had come into existence (Hoffman, 1988, 1989, 1991; Rogers & Santosh, 2002). Crustal accretion in west Africa began at approximately 2.1 Ga and culminated in a series of major tectonic episodes at 1.9 Ga (Abouchami *et al.*, 1990; Boher *et al.*, 1992; Davies, 1995). Equivalent accretionary events can be identified in Laurentia and Baltica (Gower, 1985; Gaal & Gorbatschev, 1987). In particular, there appears to have been a rapid assembly of Archaean crustal blocks to form northern Laurentia between 1.95 and 1.8 Ga (Hoffman, 1988). Regional data thus suggest that the supercontinent began to assemble at some time close to 2.0 Ga and then began to disperse again at approximately 1.8 Ga, probably as a result of mantle instability (cf. Gurnis, 1988).

The Northern and Western Australian Mega-elements evolved together during the Barramundi and Capricorn Orogenies, respectively, as part of this putative supercontinent assembly. The Barramundi Orogeny, a significant event across much of northern Australia (Page & Williams, 1988; Le Messurier *et al.*, 1990; Needham & De Ross, 1990; Plumb *et al.*, 1990; O'Dea *et al.*, 1997), appears to be associated with the final phase of the

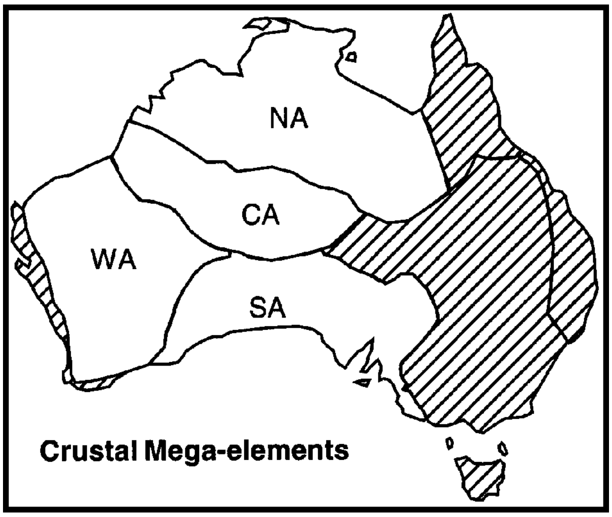

Fig. 2. Crustal mega-elements of the Australian Craton (SA, CA, WA and NA are the Southern, Central, Western and Northern Mega-elements, respectively) (after Shaw *et al.*, 1996).

assembly of this early Palaeoproterozoic supercontinent. Crustal shortening, voluminous igneous activity (Wyborn, 1988) and low-pressure metamorphism (Etheridge *et al.*, 1987) during this event produced the basement rocks underlying much of northern Australia (Plumb *et al.*, 1980). The crust, which probably evolved on earlier Archaean continental crust, is at least 43–53 km thick (Collins, 1983), and may well have been thicker in the past.

Beginning at approximately 1.8 Ga, large areas of the Northern Australian Mega-element began to subside, possibly as a response to mantle instability and the intrusion of anorogenic granites at the time of the breakup of the putative Palaeoproterozoic supercontinent (cf. Gurnis, 1988; Idnurm & Giddings, 1988; Wyborn, 1988; Pysklywec & Mitrovica, 1998). Subsidence in response to these events led to the development of a series of intracratonic basins including the McArthur, Mount Isa, Victoria and Kimberley Basins, which blanket the Northern Australian Mega-element (Lindsay, 1998). These basins are complex polyphase structures which continued to subside for more than 200 M.y. preserving in excess of 10 km of sediment, all with similar basin-fill architectures (Lindsay & Brasier, 2000).

The Western Australian Mega-element (Fig. 2) consists of two well-defined Archaean blocks, the Yilgarn and Pilbara Cratons which were sutured along the Capricorn Orogen at the same time as the Northern Australian Mega-element was evolving. Ocean closure was underway by c. 2.3 Ga and the Pilbara and Yilgarn cratonic margins became active with ocean floor possibly being subducted beneath the Yilgarn Craton leading to suturing of the two cratons between 2.0 and 1.8 Ga (Tyler & Thorne, 1990; Thorne & Seymour, 1991; Occhipinti *et al.*, 1998; Pirajno *et al.*, 1998). A series of basins (Yerrida, Bryah, Padbury and Earaheedy Basins) formed along the cratonic suture recording the convergence and collision of the two cratons. The later Bangemall Basin developed on the newly formed mega-element as much the same time as the northern Australian basins developed on the Northern Australian Mega-element (Muhling & Brakel, 1985).

The Central and Southern Australian Mega-elements were amalgamated somewhat later than their northern and western counterparts. However, the process was probably completed in the Late Mesoproterozoic by approximately 1.1 Ga (Myers *et al.*, 1994, 1996; Camacho & Fanning, 1995; Clarke *et al.*, 1995). The final amalgamation occurred as part of the aggregation of the Rodinian supercontinent. The development and later dispersal of Rodinia (McMenamin & McMenamin, 1990), beginning at around 1 Ga, has been broadly outlined (e.g. Bond *et al.*, 1984; Dalziel, 1991, 1992; Li *et al.*, 1996). More specifically the connection between these Australian basins and Rodinia is discussed in Lindsay *et al.* (1987) and Powell *et al.* (1994).

The Neoproterozoic basins of central Australia thus rest directly upon this thick rigid older craton (Haddad *et al.*, 2001) over most of the area, except for northern Australia where the Georgina Basin rests, in part, on the younger (1.7–1.5 Ga), relatively gently deformed rocks of the McArthur and Mount Isa Basins (Lodwick & Lindsay, 1990; Lindsay, 1998) which are in turn underlain by the Northern Australian Mega-element. On the Western Australian Mega-element the Officer Basin rests on Mesoproterozoic sediments of the Bentley Group and locally overlaps the Palaeoproterozoic Bangemall and Earaheedy Basins.

Crustal fabric

Deep seismic profiling of the crust has been carried out across parts of two of the Australian crustal mega-elements (see Goleby *et al.*, 1989; Wright *et al.*, 1991; Leven & Lindsay, 1995; Lindsay & Leven, 1996). The profiles across parts of the Central and Southern Australian Mega-elements providing insights into the crust and the fill of the Amadeus and Officer Basins to depths of 60 km.

Seismic reflection profiling across the Officer Basin and the southern Musgrave Block (Fig. 3) shows that the crust of the Southern Australian Mega-element is pervaded by north dipping surfaces (Lindsay & Leven, 1996). The surfaces are clearly imaged on seismic profiles and can be seen to extend almost to the base of the crust, which at this location is approximately 42 km thick. Many of these dipping reflections are imaged on shallow seismic profiles at depths as shallow as 4 s (TWT) where they can be seen to terminate against the sediments at the base of the Officer Basin (Fig. 4). A few show small displacement of the basin floor but do not extend any significant distance into the overlying sedimentary units. The surfaces appear to be thrust faults, probably generated in the crust by major compressional events during aggregation of the mega-elements earlier in the Proterozoic. The best indication that the surfaces are faults comes from the fact that some were reactivated during later basin forming events and form the major bounding faults associated with the basins (Lindsay & Leven, 1996).

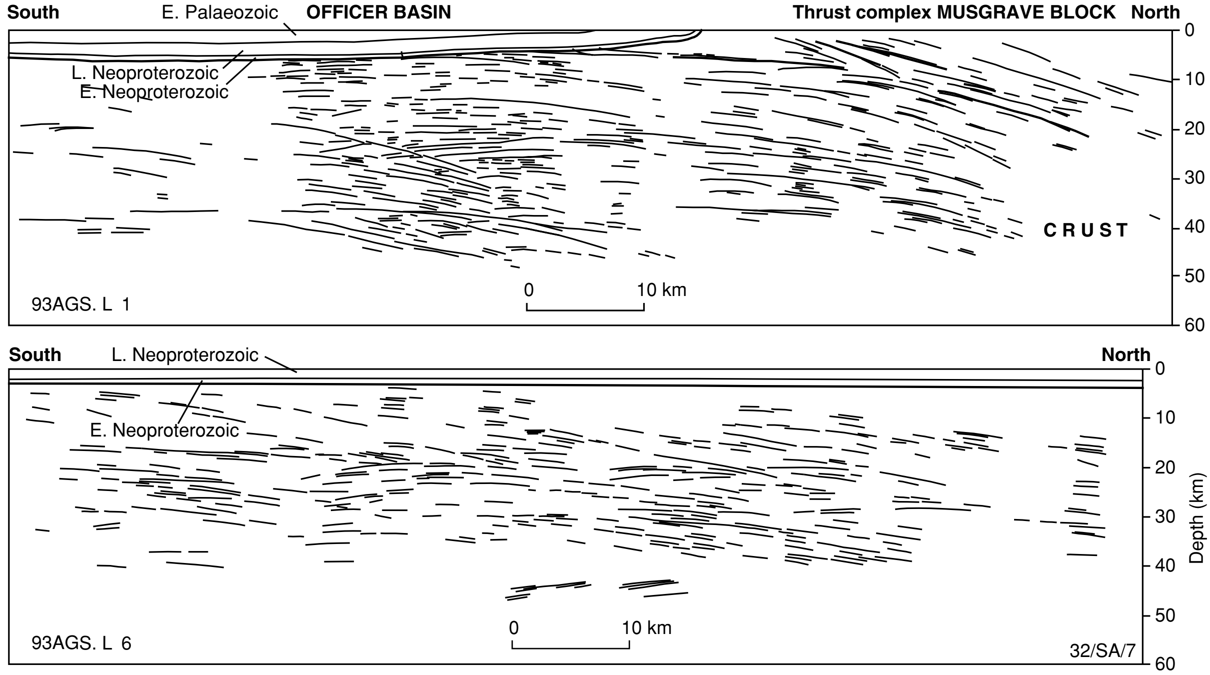

Fig. 3. Line drawing from deep seismic reflection profiles across the Officer Basin. Note the thrust complex within the Musgrave Block to the north of the basin and its relation to the monocline (Fig. 7) (after Lindsay & Leven, 1996). The crust is approximately 42 km thick across the width of the basin. See Fig. 6 for location.

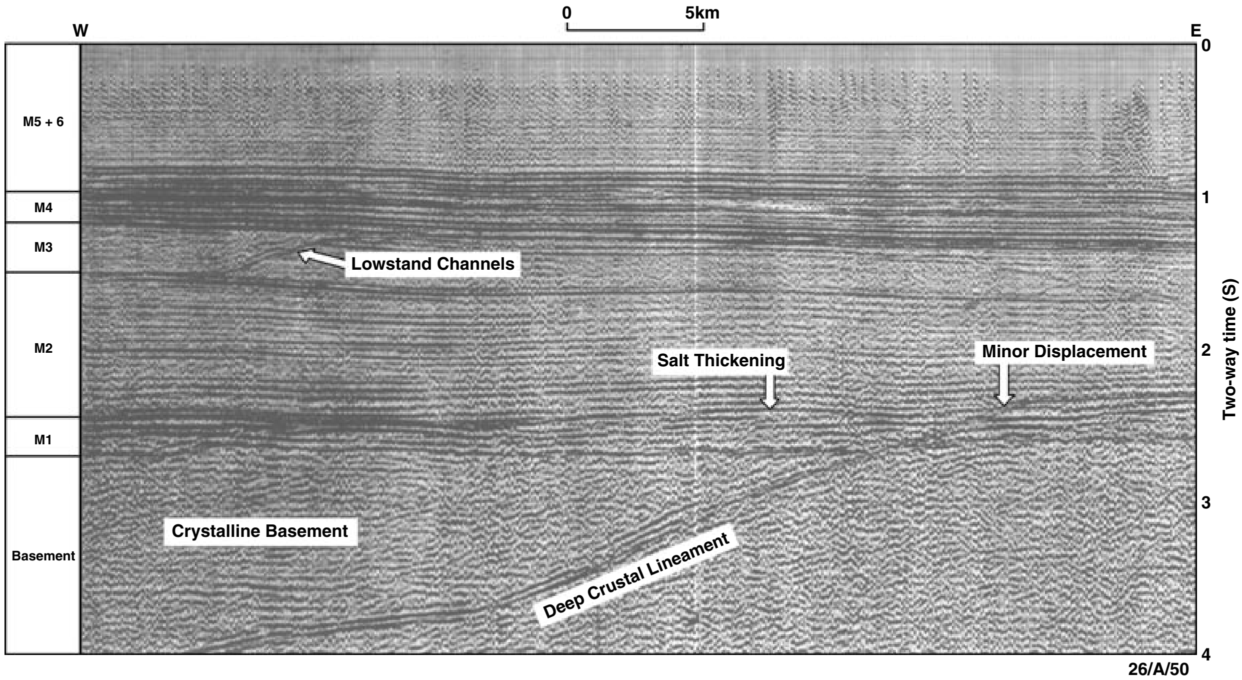

Fig. 4. Shallow seismic reflection profile (OF74-E109) showing planar crustal structure terminating against the basal surface beneath the Officer Basin (after Lindsay & Leven, 1996). Minor movement on the surface has resulted in displacement of the basin's basal contact and the localized movement of salt in the Neoproterozoic Alinya Formation. Impedance contrast across the planar surface is strong. See Fig. 6 for location.

Seismic reflection profiles across the Central Australian Mega-element and the overlying Amadeus Basin are not as clearly defined (Goleby *et al*., 1989; Wright *et al*., 1991). Thick Neoproterozoic halite units in the Bitter Springs Formation provide a strong impedance contrast such that much of the acoustic energy is reflected at the upper contact and does not penetrate the underlying deep crust (Lindsay, 1987b). However, enough of the crust has been imaged to the north of the basin on the Arunta Province to show that the gently dipping thrust surfaces also permeate the Central Australian Mega-element (Goleby *et al.*, 1989; Wright *et al.*, 1991). It is clear that these shallow dipping surfaces are incipient thrust faults that were developed earlier than the Neoproterozoic basins and probably related to the assembly of Rodinia in the Mesoproterozoic (Lindsay *et al.*, 1987; Lindsay & Leven, 1996). As discussed in the following section, this crustal fabric is critical in defining later basin architecture.

Isotopic dates from several basic intrusives (e.g. Stuart and Kulgera Dyke Swarm and the Lakeview Dolerites) (Zhao *et al.*, 1992; Page, 1983; Camacho & Fanning, 1995) suggest that the assembly of the crust beneath the central Australian basins was essentially complete by 1.1 Ga. Tectonic activity appears to have ceased for most of the following 200 M.y. as the next events recorded all have ages of approximately 800 Ma (see Zhao *et al.*, 1992; Sun & Sheraton, 1992). These dates mostly come from basic dyke swarms that intruded immediately before the initiation of the central Australian basins. The mafic dyke swarms and volcanics from central Australia have uniform geochemistries and isotopic signatures over distances as great as 400 km (Zhao *et al.*, 1994). It is likely that the dykes were intruded in a tensional regime during a superplume event (centred near the Adelaide Geosyncline) that uplifted the whole region, heating the crust in the process. Following the decay of the superplume subsidence commenced as a response to thermal recovery (Zhao *et al.*, 1994; Lindsay, 1999a).

BASIN ARCHITECTURE

The central Australian Neoproterozoic to Early Palaeozoic intracratonic basins (Fig. 1) share many similarities in their overall architecture. Most are asymmetric in cross-section – they have deep subbasins connected by troughs along one margin and a broad shallow platform along the opposite margin (Fig. 5). The basins were all initiated at approximately 800 Ma in the Neoproterozoic and sedimentation ceased in most of the basins in the late Palaeozoic in response to a major compressional event generally referred to as the Alice Springs Orogeny (Lindsay & Leven, 1996). The basin margins closest to the deep subbasins parallel major thrust zones (Fig. 6). This marked asymmetry is typical of foreland basin architecture (cf. DeCelles & Giles, 1996; Haddad *et al.*, 2001).

Deep seismic profiling has been undertaken across the margins of three of the central Australian basins, the Ngalia, Amadeus and Officer Basins. The profile across the Officer Basin offers the clearest image of the relationship between the basin and the underlying crust (Fig. 3) (Lindsay, 1995; Lindsay & Leven, 1996). The sedimentary rocks at the basin's northern margin form a steeply dipping monoclinal upturn (Fig. 7). Within the Musgrave Block, a series of shallow north-dipping thrust faults appears at depths of 30 km extending for more than 40 km towards the surface. At approximately 17 km from the erosional margin of the basin, the faults bifurcate and at depths less than 9 km steepen slightly, intersecting the surface north

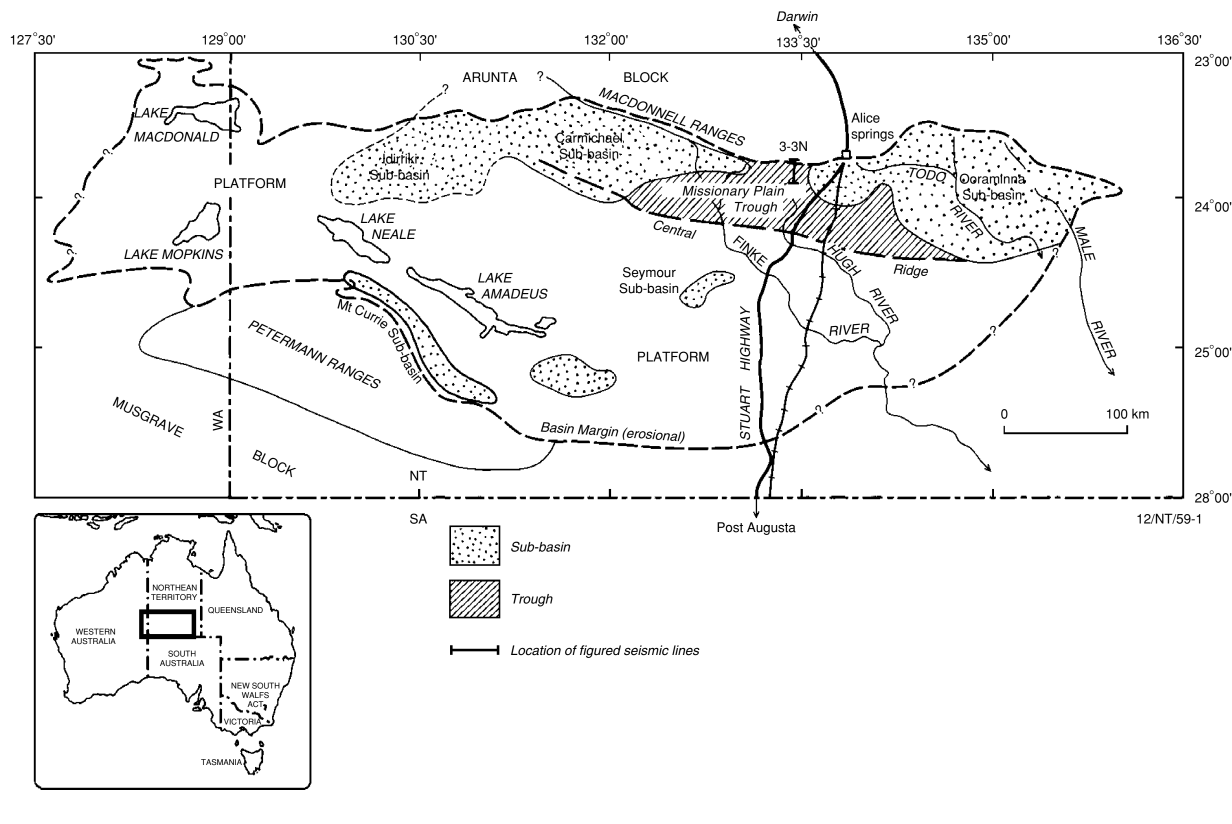

Fig. 5. Major morphological features of the Amadeus Basin (after Lindsay & Korsch, 1989). Note the location of the Petermann Ranges which formed a major sediment source during deposition of Megasequence M3. Seismic line 3–3 N is shown in Fig. 9.

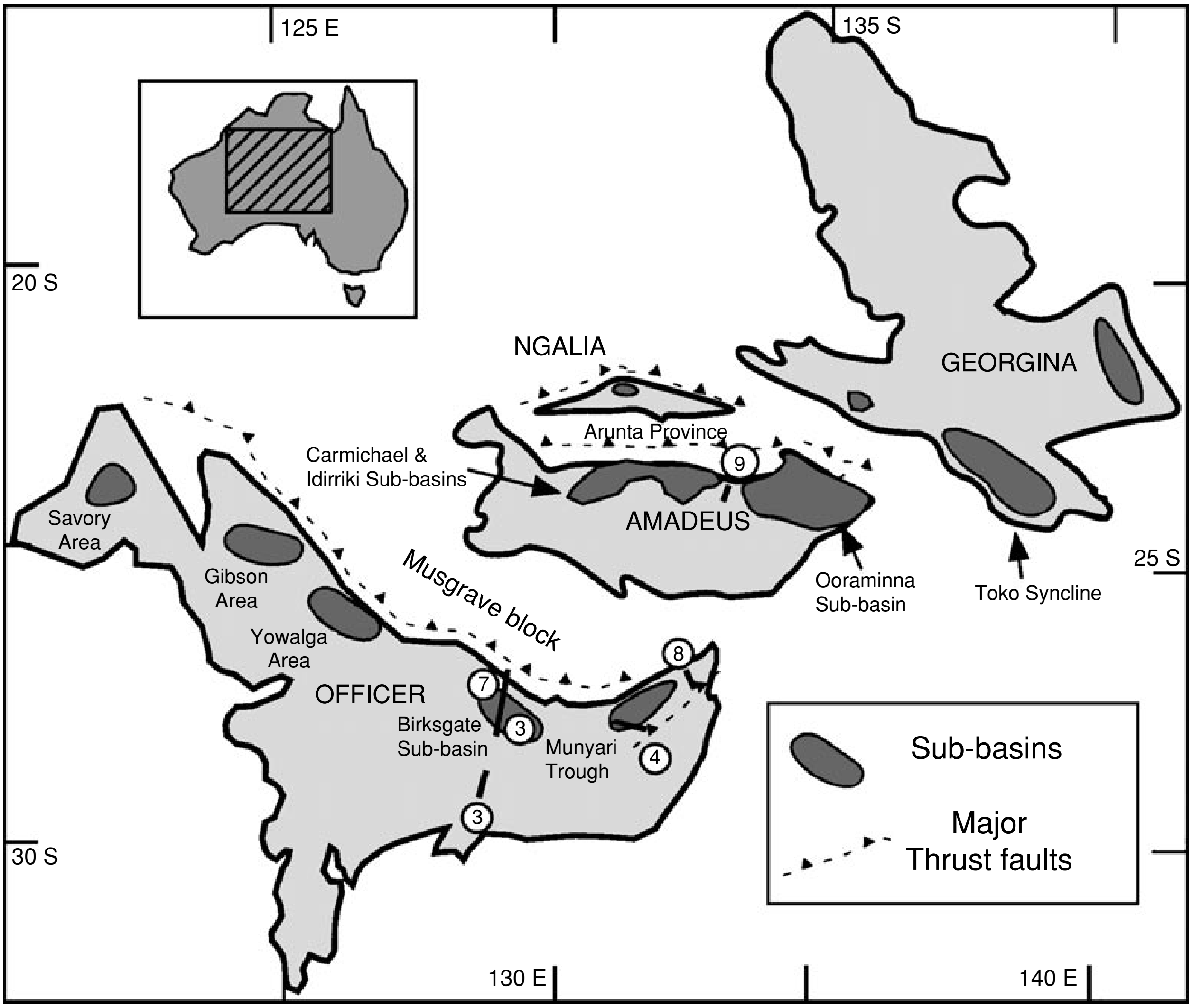

Fig. 6. Generalized architecture of the four central Australian Neoproterozoic basins. Note the general asymmetry of the basins with subbasins paralleling one thrust-faulted margin. Circled numbers indicate location of seismic sections appearing in Figs 3, 7, 8 and 9.

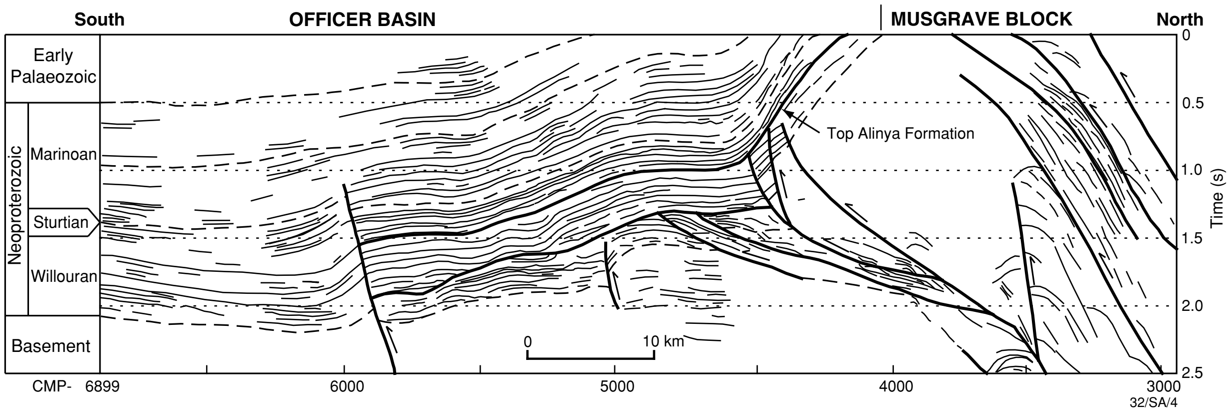

Fig. 7. Section across the Musgrave Monocline at the northern margin of the Officer Basin (after Lindsay & Leven, 1996). Note the location of the evaporite units in the Neoproterozoic Alinya Formation and its relation to faulting. See Fig. 6 for location.

of the basin margin. One major thrust fault complex disrupts the basal layers of the succession slightly before paralleling the strata where some of the strain was released along halite layers in the Alinya Formation. South of the monocline, Devonian and Ordovician units are structurally concordant and were deposited as synorogenic sediments in a shallow foreland basin setting. The northern margin of the Officer Basin is thus a monocline formed on

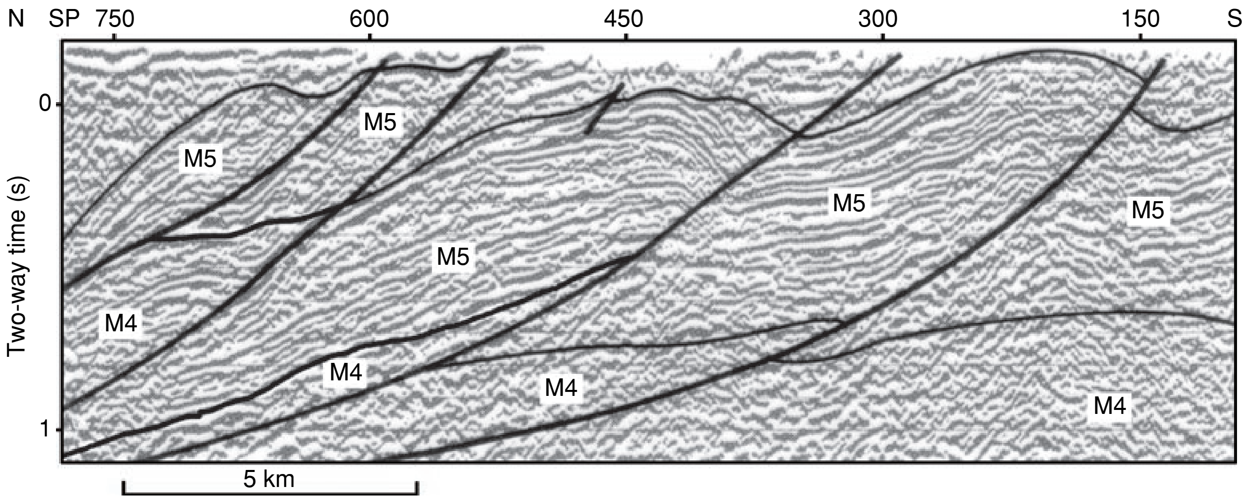

Fig. 8. Seismic reflection profile (85.0093) showing thrust faulting of Palaeozoic rocks along the northern margin at the eastern end of the Officer Basin (after Lindsay & Leven, 1996). See Fig. 6 for location.

a reactivated complex of north-dipping Mesoproterozoic thrust faults. The present geometry of the monocline is in large part due to the Late Cambrian Delamerian and Ordovician–Devonian Alice Springs Orogenies (Fig. 8). However, there is strong evidence from the basin-fill architecture to indicate that a foreland basin came into existence along the basin's northern margin in the Neoproterozoic as indicated by the major erosional surface at the top of Megasequence M1 (= Supersequence) (Lindsay & Leven, 1996). Even though the data are not as clear, deep seismic profiles across the northern margin of the Amadeus Basin indicate a very similar geometry (Goleby *et al.*, 1989; Wright *et al.*, 1991).

Immediately south of the monoclines, along the northern margins of both the Officer and Amadeus Basins (Figs 6 and 9), a series of deep troughs form the main depocentres of the basins. In the case of the Amadeus Basin the troughs contain as much as 15 km of sediments while the Officer Basin reaches depths of 10 km (Lindsay & Korsch, 1989; Lindsay & Leven, 1996). A deep trough, now the Toko Syncline, is also present along the southwestern margin of the Georgina Basin, while a less well defined depocentre containing up to 5 km of sediment is present along the northern thrust margin of the Ngalia Basin (Wells & Moss, 1983; Deckelman, 1995). The southern margins of the main troughs or subbasins in both the Officer and Amadeus Basins are defined by pronounced ridges. The ridges were probably produced initially as peripheral bulges resulting from sediment loading of the trough, but were later exaggerated by movement of Proterozoic salt into the rising structures (Lindsay, 1993; Lindsay, 1987b; Lindsay & Korsch, 1989; Lindsay & Leven, 1996). South from the deep troughs the Officer Basin extends into a very broad, gentle platform, the Murnaroo Platform, gradually shallowing to the south with a regional dip of approximately 0.3° (Fig. 3). Structuring within the basin fill is much less pronounced when compared to the Amadeus Basin, probably because the Neoproterozoic evaporite units in the Officer Basin are thinner and less mobile. In the Amadeus Basin, the southern platform is very similar to the Murnaroo Platform but its sediment fill is disrupted by thin-skinned tectonics most of which sole out in the evaporitic Neoproterozoic Bitter Springs Formation near the base of the basin succession (Lindsay, 1993, 1999b).

BASIN-FILL ARCHITECTURE

Superficially, at least, the central Australian intracratonic basins share many similarities in the architecture of their basin-fills (Lindsay *et al.*, 1987). The basins have comparable stratigraphies developed as a response to the common regional tectonic controls and superimposed effects of eustasy (Lindsay *et al.*, 1987; Lindsay & Korsch, 1989; Walter *et al.*, 1995). The similarities are clearly shown in subsidence curves (Fig. 10). However, like most intracratonic basins, the central Australian basins are complex and should be regarded as stacked polyphase basins (Lindsay & Leven, 1996). The basin fill, whilst giving the overall impression of conformity, in fact includes large hiatuses that define major basin phases or tectonic interludes during which depositional space was created by subsidence. The basin fill can thus be regarded as consisting of a number of depositional packages that have been referred to as megasequences or supersequences (Fig. 11) (Lindsay & Leven, 1996). These packages are not eustatically controlled but are tectonostratigraphic entities that were, however, internally modulated by eustasy, i.e. they can be subdivided into depositional sequences that formed in response to eustatic cycles.

The megasequences, especially in the Neoproterozoic, are most clearly defined in the main depocentre of the Officer Basin. Here six megasequences have been recognized using shallow (4 s TWT) reflection seismic data and well logs (Lindsay & Leven, 1996). Similar depositional packages can be identified in each of the basins. In spite of

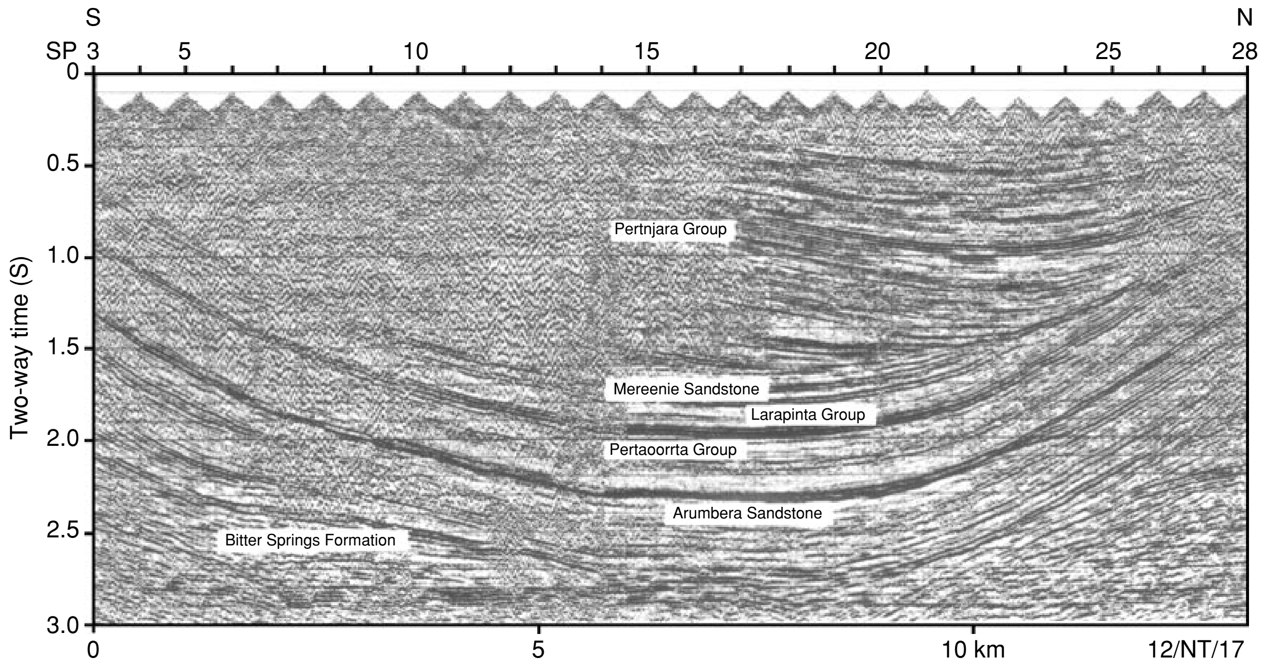

Fig. 9. North-south seismic line 3-3N across the Missionary Plain Trough in the Amadeus Basin (after Lindsay & Korsch, 1989). As much as 60% of the section in the middle of the trough was deposited during Megasequence M6 in response to the Alice Springs Orogeny. See Figs 6 and 7 for location.

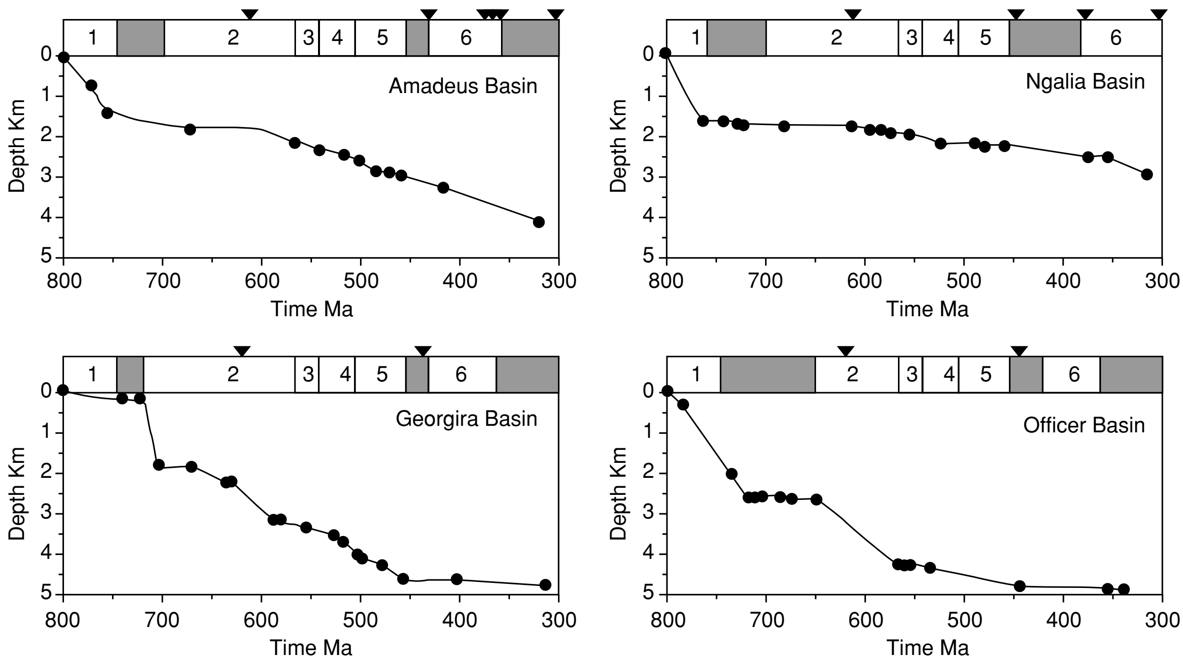

Fig. 10. Tectonic subsidence curves for the four main central Australian basins (updated from Lindsay *et al*., 1987). Note the broad similarities in the curves indicating regional tectonic controls whilst in detail the megasequences vary somewhat in duration and intensity from basin to basin. Backstripping was carried out following the method of Bond & Kominz (1984) assuming an average mantle density of 3.40. Densities and porosities of stratigraphic intervals are based on well-log analysis. Sealevel estimates were made from well-log analysis, regional facies evaluation and seismic analysis. The updated timescale is based on Young & Laurie (1996).

the similarities there are also significant differences between the architectures of the basins, indicating that tectonism and hence subsidence was not entirely synchronous throughout the region (Haddad *et al*., 2001). For example, whilst the Neoproterozoic megasequences are best displayed in the Officer Basin, the Palaeozoic megasequences are more clearly defined in the Amadeus Basin (Lindsay & Korsch, 1989). Throughout the paper

S ... N

Age	Officer	Amadeus	Ngalia	Georgina	Megasequences	Ma
Carbon-iferous Devon-ian	Lennis/ Wanna Unnamed	Brewer Park Mereenie	Mount Eclipse Kerridy	Dulcie	6 Alice Springs Orogeny	470–290
Ordovician	Cartu Blue Hills Mt. Chandler Byilkaoora Table Hill/Kulyong	Carmichael Stokes/Stairway Horn valley Paccota upper Goyder	Djagamara	Ethabuka Mithaka Carlo Nora Coolibah Kelly creek Tomahawk	5 Delamerian Orogeny	505–470 Volcanism
Cambrian	Wirrildar/ Trainor hill Observatory hill Relief/Ouldburra	Lower Goyder Shannon Giles Creek upper Arumbera	Bloodwood Walbiri upper Yuendumu	Arrinthrunga Arthur Ck/marqua Redheart/ Thorntonia and equivalent Mt Baldwin/Adam	4	Volcanism 545–505
Neoproterozoic	upper Ungoolya Gp.	lower arumbera	lower yuendumu	Central Mt stuary/	3	580–545
	lower Ungoolya Gp Murnaroo Tarlina Lupton/ Turkey Hill Chambers Bluff	Julie pertatataka olympic/pioneer Boord? Araika Inindia Areyonga winnall?	Mt. Doreen Rinkabeena Naburula	Eikera Grant biuff Eiyuah Gnalian-a-gea Oorabra Mt cornish/yardida	Petermann Orogeny Souths Ra movement 2 Areyonga event	Glaciation 740–580 Glaciation
	Browne/Alinya Towwsend/Pindyin	Bitter Springs Heavitree	Albinia Vaughan Spring	Yackah	1 Sag Basin	Volcanism 800–780
	Upilift and Peneplanation				Mantle Activity	900–800

Fig. 11. Basin-fill architecture of the central Australian basins in relation to a simplified lithostratigraphy. The vertical scale is time based and not indicative of relative basin-fill thicknesses which are quite variable. Not all lithostratigraphic units are shown.

the time scale has been updated to that proposed by Young & Laurie (1996).

M1 is the only megasequence that is consistent across the region. For the most part it consists of a thick basal sandstone unit followed by an equally thick carbonate and evaporite unit that averages 600–1000 m in thickness (Fig. 11). The basal sandstone is in particular enigmatic in that it is difficult to explain the deposition of such a thick and extensive sheet of immature sand. Recent work on the Heavitree Quartzite in the Amadeus Basin suggests, however, sands were deposited as a direct response to the events surrounding the assembly of Rodinia, in particular peneplanation during regional uplift in response to a rising mantle plume followed by broad regional subsidence as the plume decayed (Lindsay, 1999a). Once the sand was redistributed by fluvial and tidal processes, the sag basin was starved of clastic sediment and sedimentation switched to evaporites and carbonate deposition. Megasequence M1 thus records deposition in a sag basin which subsided synchronously across much of the Australia Craton over an area of at least 2.5×10^6 km^2 and, on the basis of subsidence modelling, a time interval of c. 20 M.y. (Gravestock & Lindsay, 1994; Lindsay, 1999a).

The five subsequent megasequences (Fig. 11) are, however, much less predictable. Megasequence M2 (c. 740–580 Ma) is well represented in the Officer and Amadeus Basins but relatively restricted in the Ngalia and Georgina Basins. Following thermal subsidence during the basin's initial sag phase, M1, there appears to be a pronounced hiatus in the sedimentary record. The time gap appears to be especially large in the Officer Basin, perhaps as large as 100 M.y., as only remnant glacial sediments stratigraphically consistent with a Sturtian age, are to be found (Chambers Bluff Tillite) suggesting that as much as 500 m of section may have been removed by erosion at the start of M3 time (Gravestock & Lindsay, 1994) and that there may have been a long period of nondeposition. However, while deposition was delayed in the Officer Basin, later subsidence led to the deposition of a thick succession of turbidites (lower

Ungoolya Group). In the Ngalia Basin a series of smaller blocks were uplifted resulting in considerable erosion of M1 (Vaughan Springs Quartzite and Albinia Formation) (Wells & Moss, 1983).

In contrast, in the Amadeus Basin, M2 is well represented and deposition began early with evidence of both Sturtian (Areyonga Formation) and Marinoan (Olympic Formation and Pioneer Sandstone) glaciations preserved (Preiss *et al.*, 1978; Preiss & Forbes, 1981; Lindsay, 1989). The lower contact is markedly erosional with as much as 200 m of relief cut into the underlying Bitter Springs Formation (Lindsay, 1989). Subsidence appears to have been triggered by a major compressional event, the Areyonga Movement (Wells *et al.*, 1970). It is likely that the Areyonga Movement was the first basinal event to reactivate the older thrusts that pervade the crust, thereby disrupting the sag-generated superbasin and initiating a series of smaller foreland basins. In both the Officer and Amadeus Basins, significant volumes of sediment were fed from the southern margin of the basin derived from newly exhumed source areas. Some units in the southern Amadeus Basin are relatively coarse grained, in places conglomeratic (Wells *et al.*, 1970). Along the northern margin of the basin the only coarse materials are diamictites (Areyonga Formation) deposited by relatively localized ice sheets that developed on highlands to the north between the Amadeus and Ngalia Basins (Lindsay, 1989). Following the Sturtian and Marinoan glaciations, Amadeus Basin water depths increased significantly leading to much finer grained sedimentation in the northern depocentres (Pertatataka Formation) while to the south coarser sediments were being shed from the uplifted Musgrave Block (Winnall beds). Later post-Sturtian uplift along the southern margin of the basin produced a localized unconformity and increased the supply of coarse clastic materials (Wells *et al.*, 1970). This shift in sedimentation appears to be the result either of minor tectonism (Souths Range Movement) or perhaps glacial sea level drawdown followed by postglacial uplift. A rapid increase in water depth across the region indicates that a major sea level rise followed both glaciations (Lindsay, 1989).

In the Ngalia Basin the lower, Sturtian part of M2 is represented by the diamictites of the Nabarula Formation and the overlying Rinkabeena Shale while the upper, Marinoan part of the megasequence is represented by the Mount Doreen Formation which has a basal diamictite unit (Wells & Moss, 1983). M2 is not well represented in the Georgina Basin and is generally restricted to the deep subbasins along its southwestern margin. Evidence of the Souths Range Movement is also present in the Ngalia and Georgina Basins (Wells & Moss, 1983). There is uncertainty as to whether the Souths Range Movement represents the beginnings of a separate megasequence or whether it is best regarded as a reactivation of Megasequence M2 (cf. Walter *et al.*, 1995). Either way it is significant that the glaciations are associated with each reactivation suggesting that tectonics and topography may have been more important than global climate in triggering the Neoproterozoic ice ages.

The depositional architecture of Megasequence M2 is very different from Megasequence M1 suggesting that the sag basin had ceased to be a single continuous depositional entity. Instead, major thrust faults that had remained dormant since the Mesoproterozoic were reactivated, elevating crustal blocks between (e.g. Arunta Province and Musgrave Block) and, at times, within basins such that the four intracratonic basins became independent but interconnected depositional settings (Lindsay & Leven, 1996).

The Petermann Orogeny, which resulted in the deposition of Megasequence M3 (c. 580–545 Ma), had its greatest effect along the southern margin of the Amadeus Basin and on the northern margin of the Officer Basin. It has been suggested that this orogeny was due to a shift from a dominantly extensional regime to a dextral shear regime (Veevers, 1984) with the Musgrave Block being a zone of contractional deformation. Certainly, the Musgrave Block underwent intense compression but there is little to suggest extension during the preceding megasequence, M2. As a result of the compression, the onset of subsidence in the Officer Basin was abrupt and rapid, producing a well-defined package of deeper water turbidites (upper Ungoolya Group) in the main basinal depocentres of the Officer Basin. At the same time, in the northern Amadeus Basin, subsidence was reactivated developing a narrow trough into which the lower depositional sequence of the Arumbera Formation prograded. Sediments deposited in the narrow elongate trough were derived from the uplifted Petermann Ranges and Musgrave Block and transported across the southern platform of the Amadeus Basin as part of a braided stream complex (Lindsay, 1987a). The sediments thus bypassed the southern Amadeus Basin and were all deposited in the major depocentres to the north. A similar but thinner package of sediment (lower Yuendumu Sandstone) is present in the Ngalia Basin and also in the Georgina Basin (Elkera and Central Mount Stuart Formations) (Walter & McIlroy, 1997) suggesting that the effects of the Petermann Orogeny extended that far north (Wells & Moss, 1983).

Megasequence M4 (c. 545–505 Ma) was initiated at or near the Precambrian–Cambrian boundary as the established Neoproterozoic basins began to subside again (Lindsay *et al.*, 1987; Lindsay, 1987a). The central Australian basins abruptly changed their depositional style at this time. Whereas compression had abruptly narrowed the basins in the Marinoan during the Petermann Orogeny and limited sedimentation to a narrow deep foreland trough, in the Cambrian the basins again began to subside regionally. For example, Lower Cambrian units first onlapped the southern margin of the Officer Basin subbasins then topped the Ammaroodinna Ridge onto the Gawler Craton (Lindsay & Leven, 1996). This broad regional subsidence, which involved the basin's southern margin as well as the deep subbasins, may simply relate to stress release at the end of the Petermann Orogeny; however, evidence from the Amadeus Basin to the north suggests that the event had regional effects quite apart from the associated outpourings of basalt. The Amadeus

Basin followed a similar depositional pattern with sedimentation initially restricted to the northern troughs, but then quickly topping the Central Ridge and transgressing southward onto the platform during the middle and later Cambrian (Lindsay & Korsch, 1989; Lindsay *et al.*, 1993; Lindsay, 1993). At the same time sand-rich sediments were transported into the southern Amadeus Basin to form conglomeratic fans. In the Ngalia Basin, to the north, Megasequence M4 is present, but is more restricted in its distribution (Wells & Moss, 1983). Perhaps the most striking change in basin architecture occurred in the Georgina Basin which almost doubled its areal extent during Megasequence M4.

As Megasequence M4 was being deposited in the older basins, a number of new intracratonic basins (Bonaparte, Ord, Daly, Warburton and Wiso Basins) were initiated following major outpouring of Lower Cambrian tholeiitic basalts (Antrim Plateau Volcanics) at c. 513 ± 12 Ma (Hanley & Wingate, 2000). The Antrim Plateau Volcanics and correlative units form part of a flood basalt province (Fig. 1) that originally covered 300 000–400 000 km^2 both beneath and adjacent to the surrounding basins (Bultitude, 1976). Similar but less well documented basic volcanics in the Officer Basin have been dated at c. 480 Ma (Compston, 1974; Stevens & Apak, 1999). These events all occurred at about the same time as the breakup unconformity in the Adelaide Geosyncline (von der Borch, 1980; Veevers, 1984). This suggests that M4 was triggered by extension (Lindsay & Leven, 1996) associated with the breakup of east Gondwana at the end of the Rodinian supercontinent cycle (Veevers, 1984; Lindsay *et al.*, 1987; Rogers, 1996) (Fig. 1).

Regionally, it can be shown that the eastern margin of the Australian plate became an active margin in the early Palaeozoic (Lindsay, 1990). Allochthonous Cambrian limestone blocks in the associated accretionary wedge units (Chappell, 1961; Fitzpatrick, 1975; Cawood, 1976; Hall, 1978; Pickett, 1982; Aitchison, 1988; Lindsay, 1990) suggest that prior to the opening of the forearc basin relatively shallow-marine conditions prevailed. This implies that eastern Australia broke free of Rodinia some time later in the Cambrian as Megasequence M4 was deposited in the Neoproterozic basins, but prior to the development of the forearc basin along the eastern margin of the continent.

The sediments associated with Megasequence M5 (c. 505–470 Ma) were deposited in response to the Late Cambrian Delamerian Orogeny. This orogeny affected the southern limits of the Neoproterozoic basins much more than in the north. In the eastern Officer Basin it resulted in reactivation of the major thrust faults along the basin's northern margin and to contractional deformation and the development of intense thin-skinned thrusting in the eastern part of the basin (Fig. 8) (Lindsay, 1995; Hoskins & Lemon, 1995). Two thick sequences were deposited in the eastern Officer Basin in a shallow foreland basin (Fig. 5). Initially, streams flowing along the axis of the basin resulted in intense erosion removing perhaps as much as 1500 m of sediment (Moussavi-Harami & Gravestock, 1995; Lindsay & Leven, 1996). Erosion was ultimately followed by shallow marine and deltaic sedimentation. Overall, sedimentation resulting from the Delamerian Orogeny is thin and much of the time interval, especially in the early stages, is represented by nondeposition and erosion. In the western Officer Basin the interval is almost entirely nondepositional or erosional. In the Amadeus, Ngalia and Georgina Basins the Delamerian Orogeny is not as clearly defined and the boundary between M4 and M5 disappears.

Palaeozoic Megasequence M6 is associated with the Alice Springs Orogeny, perhaps the most intense and widespread of the tectonic events recognized in the central Australian basins. Reactivation of the major crustal thrust faults during this episode masks much of the earlier activity on the faults. The orogeny persisted for a long period of time, as much as 160 M.y., reaching a peak in central Australia at c. 320 Ma (Shaw *et al.*, 1984; Mawby *et al.*, 1999) and uplifting the Arunta Province by at least 5 km (Jones, 1972, 1991). Whilst the Alice Springs Orogeny was involved in crustal scale, thick-skinned deformation, its most obvious effects are to be seen in thin-skinned tectonics in the Amadeus Basin where a number of fault ruptured anticlines are conspicuously exposed (Lindsay, 1993). The faults associated with these structures and nappe structures along the northern margins of the basin sole out in the underlying Proterozoic evaporites of the Bitter Springs Formation (Lindsay, 1989). However, even these late-stage structures show evidence of a long history involving salt tectonics (Lindsay, 1987b).

The Alice Springs Orogeny is not a simple event but consisted of at least six separate tectonic pulses (Bradshaw & Evans, 1988) that reactivated the major thrust faults along basin margins developing a synorogenic depositional system in the adjacent foreland basin (Lindsay & Korsch, 1989). In the Amadeus Basin as much as 1–2 km of sediment shed from the advancing thrust sheets produced a major depositional package, Megasequence M6 (Pertnjara Group). In the Missionary Plain Trough, where the megasequence varies locally from about 1000 m to more than 5000 m in thickness, it can form more than 60% of the basin fill (Fig. 9) (Lindsay & Korsch, 1989). Truncational and onlapping relationships evident in seismic reflection data show the complexity of the depositional setting, the locus of which shifted through time. The resulting sedimentary succession consists of a series of lacustrine, braided and meandering stream and alluvial fan deposits that accumulated as a southward thinning wedge during successive tectonic pulses (Jones, 1972, 1991). The intensity of uplift increased with time as the tectonic activity moved closer to the northern margin of the basin and, as a consequence, the clastic materials shed from the advancing front became coarser. Sediment accumulation ceased by 360 Ma at the end of the fourth major tectonic pulse. Subsequent deformational episodes involved a long period of erosion and nondeposition extending to perhaps as late as 290 Ma.

Whilst the Alice Springs Orogeny was responsible for folding of earlier sediments over much of the Officer Basin, sediments comprising M6 are confined to the main depocentre at the eastern end of the basin where Devonian lacustrine sediments and evaporitic redbeds overlie the Ordovician. The sediments accumulated in a narrow east–west-trending foreland basin that developed due to crustal loading ahead of the advancing thrust sheets, much as seen in the Amadeus Basin, but on a smaller scale. From the Ngalia Basin to the north of the Amadeus, the Alice Springs Orogeny also had a significant effect but the timing was somewhat different. The initial tectonic pulses began in the Ngalia Basin at the same time as in the Amadeus Basin, uplifting both the northern and southern margins of the basin. Initially the intensity of the event was somewhat subdued leading to limited sedimentation along the northern margin (Kerridy Sandstone). As sedimentation ceased in the Amadeus Basin at the end of the Devonian, deformational intensity increased in the Ngalia Basin with significant thrusting on the main faults along the basin's northern margin leading to as much as 10 km of uplift (Wells & Moss, 1983), extending M6 sedimentation well into the Carboniferous (Mount Eclipse Sandstone) and depositing more than 2 km of sediment in the main depocentres. In the Georgina Basin the depositional pattern paralleled that of the Amadeus Basin. Sedimentation was largely restricted to the rapidly subsiding foreland trough in the present Toko Syncline (Fig. 1). Initially, as in the Amadeus Basin, sedimentation was largely shallow marine, but as the thrust sheets advanced, widespread conglomeratic fans developed (Shergold & Druce, 1980).

The Alice Springs Orogeny terminated sedimentation in the central Australian basins marking the end of almost 500 M.y. of punctuated sedimentation.

DISCUSSION

Similarity in depositional patterns and subsidence curves between the central Australian basins led Lindsay *et al.* (1987) to first consider the possibility that the central Australian Neoproterozoic basins were in some broader sense related to supercontinent evolution. The architecture of the central Australian crust suggests that, while some significant components of the crustal mega-elements were assembled during the Archaean and Palaeoproterozoic (e.g. the Pilbara and Yilgarn Cratons), the fabric of the crust was not fully established until the Mesoproterozoic as a direct result of the assembly of the Neoproterozoic supercontinent Rodinia. The crust developed during this period was pervaded by north dipping thick-skinned thrust faults: it was thick (40–50 km) (Lindsay & Leven, 1996) and strong and able to support stress over long geological time periods (Haddad *et al.*, 2001). Thus, by 1.1 Ga, the crustal substrate was in place setting the stage for the evolution of the central Australian basins.

Considerable progress has been made in understanding the origins of the central Australian basins as new data, especially deep seismic data, have become available (Lindsay & Leven, 1996). The data eliminated the possibility that the basins had been initiated by extensional tectonics as earlier suggested (Lindsay & Korsch, 1989) implying, by default, that deep mantle processes, probably mantle plumes (Zhao *et al.*, 1994), were involved. There is a growing body of evidence to suggest that there is a periodic cycle of supercontinent coalescence and dispersal (Worsley *et al.*, 1984; Murphy & Nance, 1992; Duncan & Turcotte, 1994; Veevers *et al.*, 1997), driven by large scale mantle convection (Gurnis, 1988; Anderson, 1982; Kominz & Bond, 1991; Tackley, 2000). The development and dispersal of a Neoproterozoic supercontinent, Rodinia (McMenamin & McMenamin, 1990), beginning at around 1 Ga, whilst not as well documented as Pangea (cf. Veevers, 1988, 1989), has been broadly outlined (e.g. Bond *et al.*, 1984; Lindsay *et al.*, 1987; Dalziel, 1991, 1992; Li *et al.*, 1996). It is generally agreed that supercontinents assemble over geoid lows, mantle downwellings, and disperse over the geoid highs at mantle upwellings. This cycle is likely to be continuous because the same forces that fragment the old supercontinent over the geoid high are effectively assembling the next supercontinent over the associated low (Condie, 1998).

Recent work on the deep structure of the crust beneath the Officer Basin in particular supports arguments for a mantle superplume (Lindsay, 1999b) suggesting that the sediments of Megasequence M1 (Fig. 11) must have been locally derived and generated during regional uplift and peneplanation as the superplume rose. Upwelling of the plume beginning at c. 900 Ma initially caused domal uplifting of the continental lithosphere, regional peneplanation and generation of large volumes of clastic materials (cf. Dam *et al.*, 1998). Subsequent decline of plume activity led to thermal relaxation and widespread subsidence beginning at c. 800 Ma (Zhao *et al.*, 1994). In part this explains the hiatus between the time of assembly of the crust beneath the basins and the onset of sedimentation in the basins (see Sun & Sheraton, 1992; Zhao *et al.*, 1992). Ultimately, the sediments generated by erosion were simply redistributed as thermal recovery began and the sag basin formed. Beginning at c. 800 Ma subsidence was both uniform and slow across central Australia over an area of $\sim 2.5 \times 10^6$ km^2 (Fig. 1) (Lindsay & Leven, 1996). The broad scope of the early stages of basin subsidence led Walter *et al.* (1995) to refer to this region as the 'Centralian Superbasin'. The sag basin generated by plume activity was, however, short lived (cf. Dam *et al.*, 1998). Following a major hiatus, which locally varies in magnitude from perhaps 50–100 M.y., the region was affected by a major compressional event (Areyonga Movement) which isolated the intracratonic basins, thereby forming a series of smaller foreland basins, each delimited by a major thrust complex and by uplifted intermediate basement blocks such as the Musgrave Block and Arunta Province. This event established the asymmetry of the basins with the initiation of a foreland setting and the deposition of Megasequence M2 (Lindsay & Leven, 1996).

Once established the subsidence within the basin was repeatedly renewed as the deep-seated thrust faults were reactivated by major tectonic events affecting the Australian plate (Walter *et al.*, 1995; Lindsay & Leven, 1996). The superbasin (Walter *et al.*, 1995) thus persisted only briefly in response to regional thermal recovery following plume activity but the crustal architecture, once established, allowed the polyphase central Australian basins to persist for almost half a billion years as a series of thrust-controlled, related but independent, asymmetric basins (Lindsay *et al.*, 1987; Haddad *et al.*, 2001). These later basin phases were thus the product of thick-skinned tectonics (Korsch *et al.*, 1998; Haddad *et al.*, 2001). The marked asymmetry of the postsag-phase is typical of foreland basins (e.g. DeCelles & Giles, 1996). The main depocentres within the basins developed as elongate troughs paralleling the thrust-faulted margins as is typical of foreland basins (Dickinson, 1974; Beaumont, 1981; Jordan, 1981). Their length was determined by the morphology of the orogenic load (DeCelles & Giles, 1996) and their width was determined by the flexural rigidity of the supporting lithosphere (Turcotte & Schubert, 1982; Watts, 1992; Haddad *et al.*, 2001). Subsidence within the foreland basins was not entirely synchronous implying that the central Australian basins evolved as discrete geological features once the superbasin was dismembered at the beginning of Megasequence M2 (c. 740–580 Ma) (Haddad *et al.*, 2001; Lindsay & Leven, 1996).

The formation of the central Australian basins was thus dependent upon supercontinent assembly, first for the development of a thick and extensive craton, then for the development of a heat flow imbalance to allow superplume activity. Without these basic evolutionary elements, the basins could not have formed. Finally, the development of an internal fabric during supercontinent assembly defined basin morphology by controlling thrust fault reactivation. The development of these basins is thus not unique but is tied to the supercontinent cycle. Similar basins developed during the Early Palaeoproterozoic cycle (c. 1.7 Ga) (Lindsay, 1998; Lindsay & Brasier, 2000) and during the assembly and dispersal of Pangea in the Palaeozoic (Veevers, 1988, 1989; Lindsay, 2000).

CONCLUSIONS

The Neoproterozoic–Palaeozoic basins of central Australia developed as a direct result of the assembly and dispersal of the supercontinent Rodinia. Once initiated the basins persisted for almost half a billion years leaving behind a comprehensive but discontinuous record of regional events in the form of stacked megasequences or supersequences. The basins are thus complex stacked polyphase basins that rest upon a thick, strong, rigid crust (Lindsay & Korsch, 1989; Haddad *et al.*, 2001). The basins are not unique; similar basins could potentially evolve during each supercontinent cycle. The main processes involved in the evolution of these basins were:

1 Crustal nuclei formed during the Archaean and early Palaeoproterozoic evolved to form several crustal mega-elements (Shaw *et al.*, 1996).

2 The Australian crustal mega-elements were assembled during the Mesoproterozoic to form part of the supercontinent Rodinia. This assembly process produced a crustal architecture in the form of a pervading suite of thrust faults that almost reach the base of the crust (Lindsay & Leven, 1996). The process was complete by 1.1 Ga (Shaw *et al.*, 1996) and was followed by a depositional hiatus of c. 200 M.y.

3 Heat flow disequilibrium resulting from the assembled supercontinent led to mantle instability and superplume formation (cf. Anderson, 1982; Gurnis, 1988).

4 The rising superplume caused the doming and uplifting of a vast area of central Australia which was accompanied by peneplanation and the generation of large volumes of clastic sediment (Lindsay, 1999a).

5 Thermal recovery following the decay of the superplume led to regional subsidence and the onset of sediment accumulation at c. 800 Ma. Reworking of the clastic sediments formed during uplift led to their redispersal as a thick, widespread basal sandstone (Lindsay, 1999a). This primary sag or superbasin phase of basin formation was short lived, perhaps as brief at 20 M.y. and terminated at c. 780 Ma.

6 Following a hiatus of perhaps as much as 50–100 M.y., the first of a series of extrinsic tectonic events led to a reactivation of deep-seated thrust faults and the uplift of basement blocks dismembering the superbasin to develop a distinct foreland architecture that would persist for the life of the basins (Haddad *et al.*, 2001). At this point, the Centralian Superbasin (Walter *et al.*, 1995) ceased to exist and the individual basins recognized today came into existence. The subsequent evolution of these basins paralleled each other, their architectures showing many similarities, but they became discrete geological features (Lindsay & Leven, 1996; Haddad *et al.*, 2001).

7 Onset of the two identified Neoproterozoic glaciations appears to be related to the uplift of basin thrust margins during foreland basin development. It is likely that uplifted interbasinal basement blocks acted as nucleation sites for ice sheets.

8 Evidence of extension indicates that the breakup of Rodinia most likely occurred during the latest Neoproterozoic and earliest Cambrian.

9 The Alice Springs Orogeny, a major compressive event that influenced much of central Australia (Jones, 1972, 1991), led to the cessation of sedimentation and finally, basin closure at c. 290 Ma.

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