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Geodynamic Synthesis of the Gawler Craton and Curnamona Province

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2010/27
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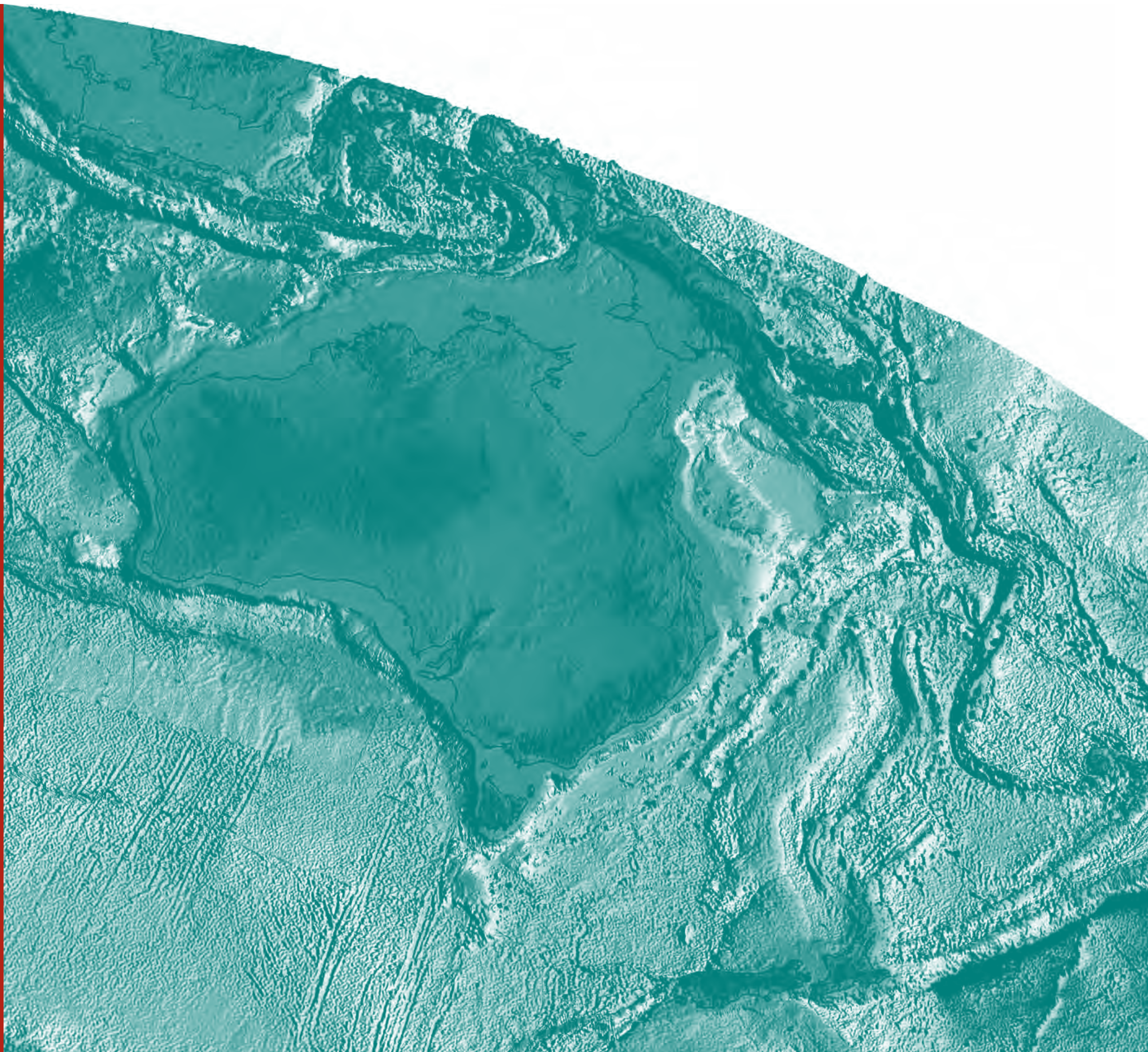
70544



APPLYING GEOSCIENCE TO AUSTRALIA'S MOST IMPORTANT CHALLENGES

Geodynamic Synthesis of the Gawler Craton and Curnamona Province

GEOSCIENCE AUSTRALIA
RECORD 2010/27

Edited by

N. Kositcin[1]

Australian Government
Geoscience Australia

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ISSN: 1448-2177
ISBN Print: 978-1-921781-24-7
ISBN Web: 978-1-921781-23-0

GeoCat # 70544

Bibliographic reference: Kositcin, N., 2010. Geodynamic Synthesis of the Gawler Craton and Curnamona Province. *Geoscience Australia, Record,* **2010/27**, 113p.

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Introduction

N. Kositcin

This report presents the results of a geodynamic synthesis of South Australia, focusing predominantly on the Archean to Mesoproterozoic of the Gawler Craton and Curnamona Province in terms of geodynamic setting, architecture, and age, using results of a geological synthesis, seismic interpretation, sequence stratigraphy, geochronology and geochemistry. This was undertaken with the dual aims: (1) to better understand the tectonic and geodynamic setting of the Gawler Craton and Curnamona Province, and (2) to accompany the interpretation of recently-acquired seismic reflection transects (Korsch and Kositcin, 2010), and to highlight new geochemical and geochronological data collected from South Australia.

Geological data are synthesised on a regional basis. Data compilation involved merging several pre-existing data compilations, and revising these by including more up-to-date data. This was undertaken to identify significant geological events, and to produce regional geological syntheses and accompanying time-space-event plots. These regional syntheses were used to produce an interpreted geological and geodynamic synthesis of the Archean to Mesoproterozoic Gawler Craton and Curnamona Province. Outputs are delivered in three parts. Geological summaries and time-space-event plots are presented in Section 1. An interpreted geological and geodynamic synthesis is presented in Section 2. Section 3 presents a preferred set of models for the geodynamic evolution of the Gawler Craton and Curnamona Province from the Archean to early Mesoproterozoic.

This report uses, to some extent, information synthesised as part of Geoscience Australia's Proterozoic Synthesis of Australia Project (Neumann and Fraser, 2007), which provided time-space-event plots for each of the Proterozoic provinces of Australia, for the interval 1900-1400 Ma, together with explanatory notes describing the geological interpretation of available data and highlighting major knowledge gaps. A companion report (Fraser et al., 2007b) presented an interpreted geological and geodynamic synthesis for Proterozoic Australia, and provided a geodynamic framework to both constrain known mineralisation and allow a predictive capability for potential new mineralisation. This report follows the same time-space-event plot/tectonic interpretation methodology as that used in the geodynamic syntheses of North Queensland (Kositcin et al., 2009), and the Phanerozoic of Eastern Australia (Champion et al., 2009a). The current report focuses specifically on the Gawler and Curnamona regions during the Archean to Mesoproterozoic period. The synthesis of the Gawler Craton and Curnamona Province is presented as a series of time-space plots, based on information captured within GA's PROVINCES and EVENTS databases.

The Gawler Craton (Figures 1, 2, 3, 4) is an ancient crystalline shield consisting of variably deformed and metamorphosed sedimentary, volcanic and plutonic rocks, which span the Archean to Mesoproterozoic, from ~3150 Ma to ~1450 Ma. Owing to sparse outcrop, the geology of the Gawler Craton is relatively poorly understood, and its boundaries are entirely subsurface, being interpreted from total magnetic intensity and gravity data combined with outcrop and drillhole information (Schwarz et al., 2006). The eastern boundary is defined as the transitional Torrens Hinge Zone, the western limit of the Delamerian Orogen, and the southern boundary is defined by the southern edge of shallow Precambrian cratonic basement of the southern continental shelf (Parker et al., 1993; Daly et al., 1998). Its northern, northwestern and western boundaries correspond to faulted or onlapping margins of Neoproterozoic and Phanerozoic successions and are less clearly defined (Daly et al., 1998; Figure 3).

The Archean to Mesoproterozoic rocks of the Gawler Craton have been subdivided into a series of domains (e.g., Thomson, 1970; Parker, 1990; Teasdale, 1997; Ferris, 2001), with each domain encompassing an area of crust which has similar lithological and structural associations, but which differs broadly from the surrounding domains (see Ferris et al., 2002; Figure 2). Given the paucity of outcrop within the Gawler Craton, many of the domain boundaries are defined by geophysical data (Figures 2, 4), particularly contrasting magnetic character and the presence of prominent curvilinear magnetic anomalies, interpreted as bounding faults and shear zones (Ferris et al., 2002):

Figure 1. Simplified geology of South Australia, showing the Gawler Craton and Curnamona Province (modified from www.pir.sa.gov.au/minerals/geology/geological_provinces).

The Nawa Domain, situated northwest of the Karari Shear Zone (Figure 4), is entirely covered by Phanerozoic sedimentary rocks, and geological knowledge of the crystalline basement is limited to sparse drillhole intersections and interpretation of potential field data. The northwest domain boundary is obscured by the Murnaroo Platform; the Karari Shear Zone bounds the domain to the south and the Torrens Hinge Zone to the northeast.

The Coober Pedy Domain is situated north of the Christie Domain, apparently bounded to the north by splays of the Karari Shear Zone, and to the south by the Tallacootra Shear Zone (Figure 4). Outcrop of basement rock is extremely rare and geological knowledge is limited to observation from sparse drillholes.

The Mount Woods Domain is situated to the east of the Coober Pedy Domain, and northwest of the Olympic Domain, in the far north of the Gawler Craton. The geology is relatively poorly known due to sparse outcrop. The northern boundary of the Mount Woods Domain is defined by the Tallacootra Shear Zone (Figure 4), the southern boundary is fault controlled, with its eastern boundary being obscured by a large increase in thickness of overlying sediments.

The Christie Domain is a large, northeast-trending, arcuate region bounded to the north by the Karari Shear Zone, and to the southeast by the Coorabie Shear Zone. This region is characterised by latest Archean to earliest Paleoproterozoic metasedimentary rocks (Daly and Fanning, 1993).

The Fowler Domain is a north-northeast trending zone of high magnetic intensity, bounded and dissected by anastomosing northeast-striking shear zones, including the Tallacootra and Coorabie Shear Zones. Teasdale (1997) divided the Fowler Domain into four blocks bounded by major shear zones (the Nundroo, Central, Barton and Colona Blocks; Figure 4). The geological characters of these blocks differ considerably, and have different metamorphic and geochronological histories. The domain is bounded by the Coorabie Shear Zone to the east.

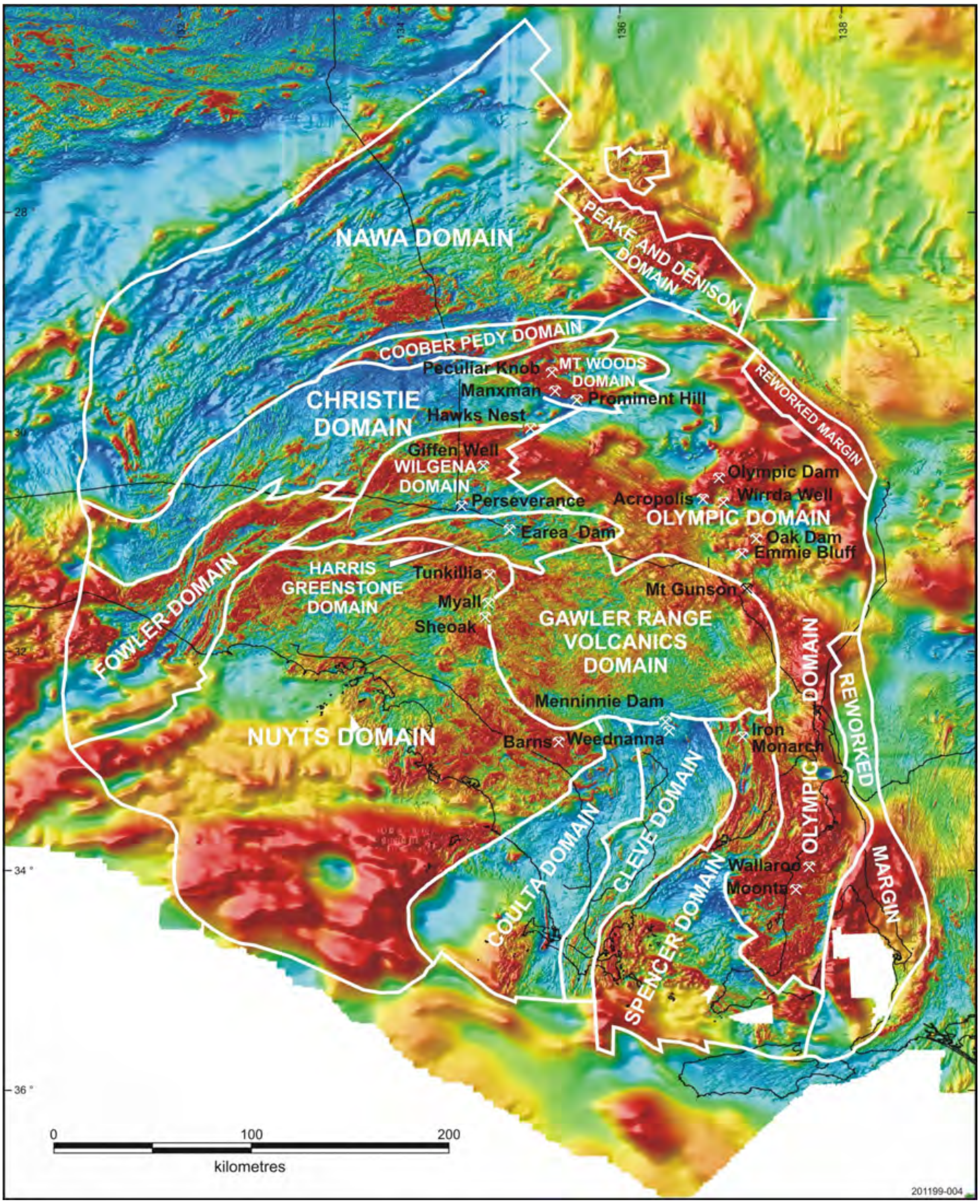

Figure 2. Location and province divisions within the Gawler Craton and mine prospects on a Total Magnetic Intensity image (after Ferris et al., 2002).

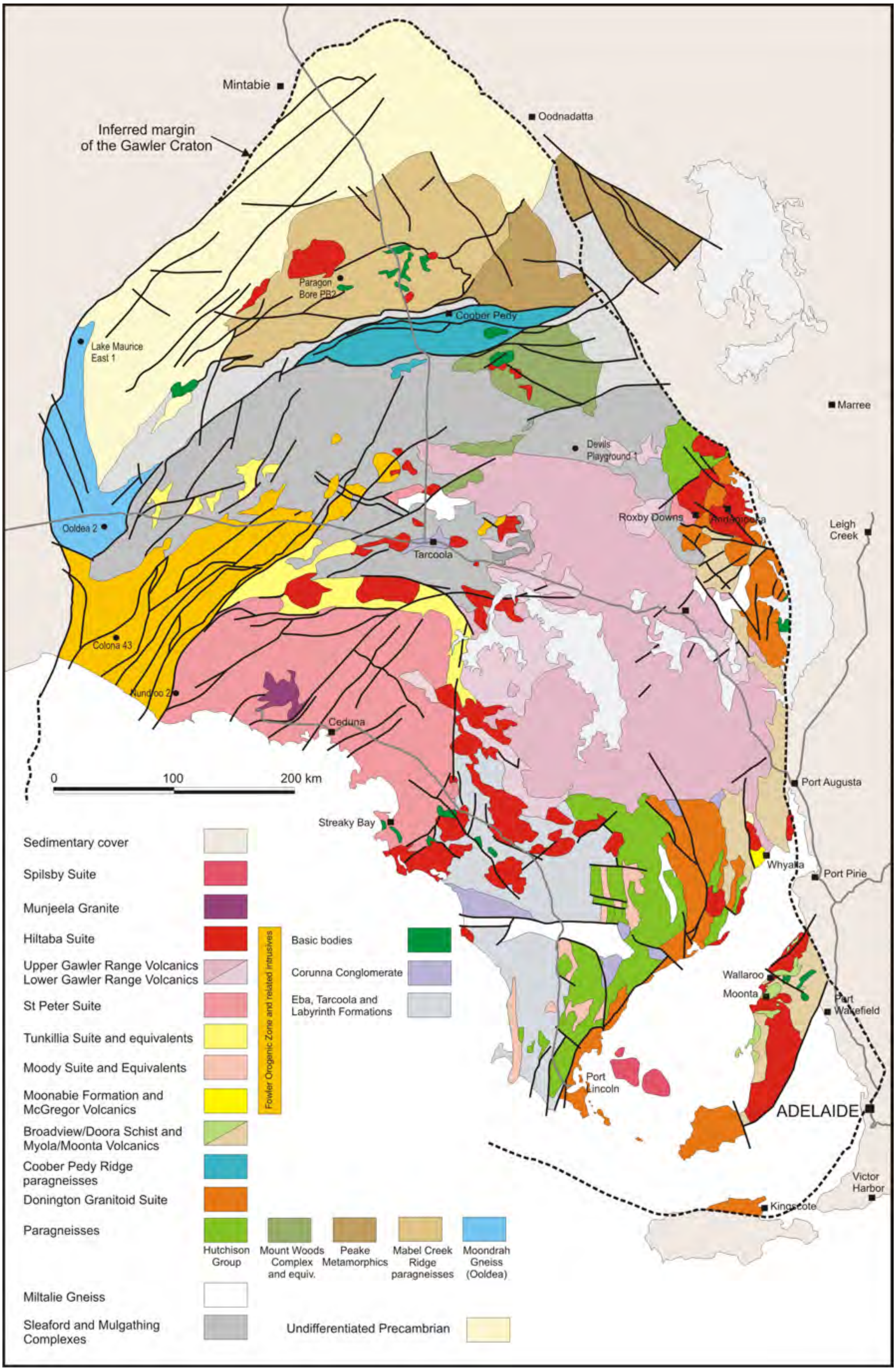

Figure 3. Gawler Craton solid geology interpretation (modified after Ferris et al., 2002).

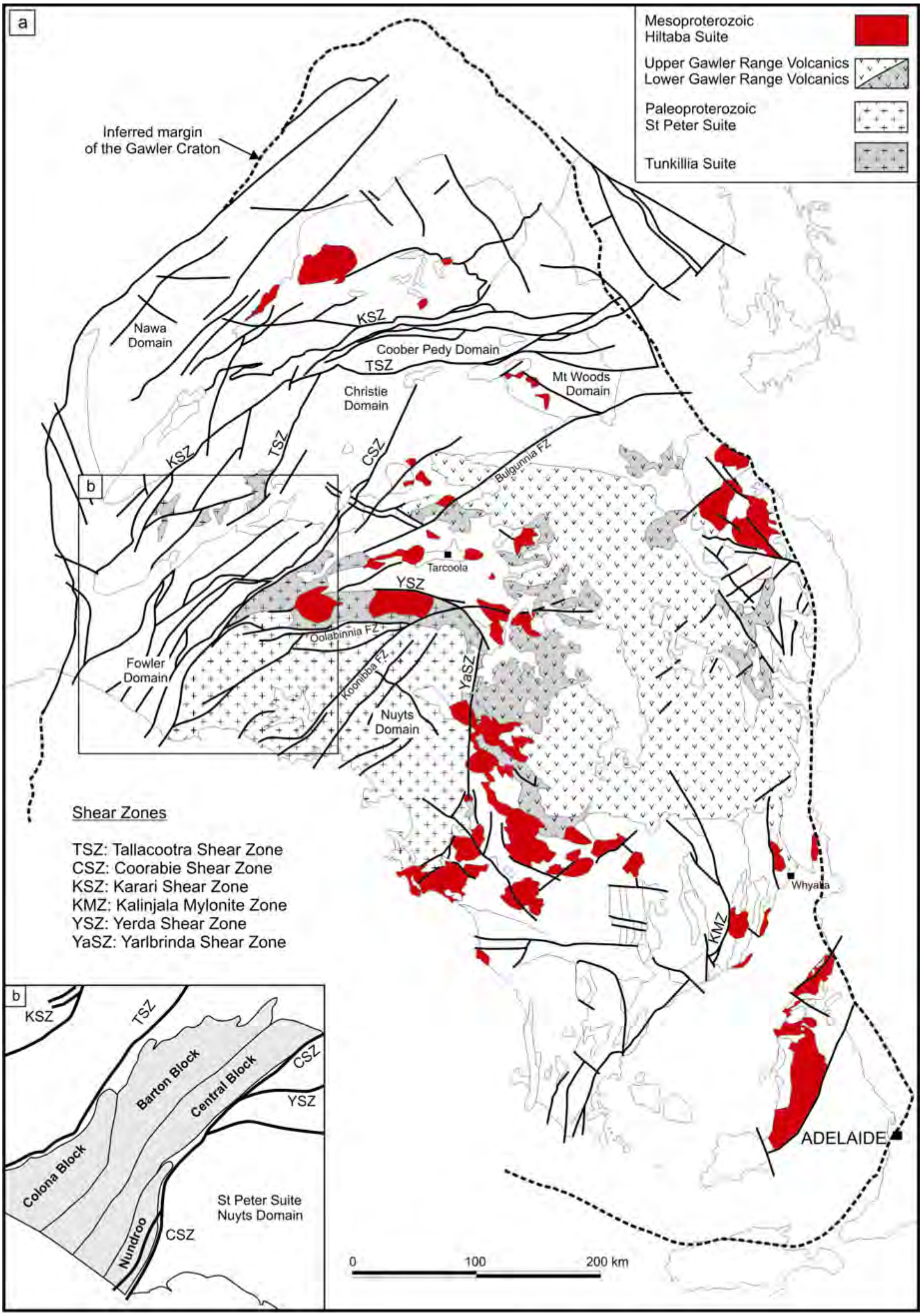

Figure 4. (a) Interpreted subsurface geology of the Gawler Craton (modified from Swain et al., 2005b), highlighting crustal-scale shear zones and major lithologies. (b) Shear zone relationships and the four blocks of the Fowler Domain (modified from Teasdale, 1997).

The Wilgena Domain occurs to the east of the Fowler and Christie Domains, across the Coorabie Shear Zone. The domain is bounded to the north and west by the Coorabie Shear Zone, to the south by the Yerda Shear Zone and Harris Greenstone Domain, and to the east by the Olympic Domain (Figure 4).

The Harris Greenstone Domain is a relatively small, fault-bounded region in the central Gawler Craton, composed of ultramafic-mafic extrusive rocks and interbedded metasedimentary rocks. The domain is bounded to the south by the Yerda Shear Zone, whereas its northern boundary is a lithological zone boundary with the Wilgena Domain.

The Nuyts Domain, situated in the southwest of the Gawler Craton, is one of the largest domains in the craton, but is also one of the least well understood, due to the lack of outcrop. Basement geology in the Nuyts Domain is interpreted to be dominated by late Paleoproterozoic to Mesoproterozoic magmatic intrusive rocks (Fairclough et al., 2003). The domain is bounded to the north by the Yerda Shear Zone and to the west by the Coorabie Shear Zone; its eastern boundary may be controlled by the Yarlbrinda Shear Zone (Ferris et al., 2002).

The Coulta Domain, in the southern Gawler Craton, is interpreted to consist primarily of late Archean metasedimentary and volcanic rocks of the Sleaford Complex (Ferris et al., 2002), and is intruded by granite plutons of the ~1590 Ma Hiltaba Suite. The boundary with the Cleve Domain to the east is interpreted as a major shear zone identified as a linear boundary in magnetic images. The western boundary of the Coulta Domain is also interpreted as a fault boundary, separating the Archean rocks of the Coulta Domain from interpreted younger intrusive rocks of the Nuyts Domain to the west (Fraser and Neumann, 2010).

The Cleve Domain (Ferris et al., 2002) lies to the west of the Spencer Domain, across the Kalinjala Mylonite Zone, and is characterised by a uniformly low magnetic signature. Geology of the Cleve Domain is dominated by Paleoproterozoic metasedimentary rocks of the Hutchison Group, as well as felsic gneisses. The western part of the Cleve Domain is intruded by granites of the ~1590 Ma Hiltaba Suite.

The Spencer Domain (Ferris et al., 2002), located on the easternmost side of Eyre Peninsula, is characterised by Donington Suite granites in the south. The northwestern part of the Spencer Domain contains Mesoarchean granites and gneisses, with Paleoproterozoic metasedimentary rocks in the northeast. The Spencer Domain is separated from the Domain to the west by the Kalinjala Mylonite Zone, and from the Olympic Domain to the east by the Roopena Fault.

The Gawler Range Volcanics Domain essentially comprises the extensive felsic, and lesser mafic, lavas of the Gawler Range Volcanics, its boundaries coincident with the elliptical magnetic signature of the Gawler Range Volcanics. Older components of the Gawler Craton below the volcanics are largely unknown.

The Olympic Domain (incorporating the Moonta Domain of Parker, 1993) forms the eastern margin of the Gawler Craton, bounded to the east by sedimentary rocks of the Adelaide Rift System. In its southern part, the Olympic Domain is separated from the Spencer Domain by the Roopena Fault. In the north, the western margin of the Olympic Domain abuts the Wilgena, Harris Greenstone and Gawler Range Volcanics domains. The Olympic Domain contains Donington Suite granites in the north, and Paleoproterozoic metasedimentary rocks of the Wallaroo Group.

The Peake and Denison Domain is situated on the northeastern margin of the Gawler Craton. The boundaries of the Peake and Denison Domain are defined by unconformable contacts with overlying Adelaidean sedimentary rocks (Ambrose et al., 1980; Ferris et al., 2002). The eastern boundary is obscured by thick Phanerozoic sediments, apparently fault-bounded because of the aeromagnetic signature.

In this report we use these domains of the Gawler Craton to facilitate discussion of its geological history and tectonic development, while acknowledging that some of these domains may require refinement. Such refinement is beyond the scope of this report; new data and geophysics should aid the revision of domain boundaries and manage current discrepancies, e.g., between figures 2, 3 and 4 (notably faults).

The Curnamona Province (Figures 1, 5, 6) is a large area of Paleoproterozoic to Mesoproterozoic rocks in the northeast of South Australia and western New South Wales (Robertson et al., 2006; Figure 5). The province lies east of the Archean to Mesoproterozoic Gawler Craton, from which it is separated at surface by the Neoproterozoic to Cambrian Adelaide Rift System (Preiss, 2000). Known rocks in the Curnamona Province fall into two main intervals: the older interval consists of a late Paleoproterozoic metasedimentary and partly metavolcanic succession (Willyama Supergroup) and intrusives, which were intensely deformed and metamorphosed by the Late Paleoproterozoic–Early Mesoproterozoic Olarian Orogeny (~1620-1580 Ma). The younger interval consists of early Mesoproterozoic volcanic and sedimentary rocks, and granitoids (Robertson et al., 2006).

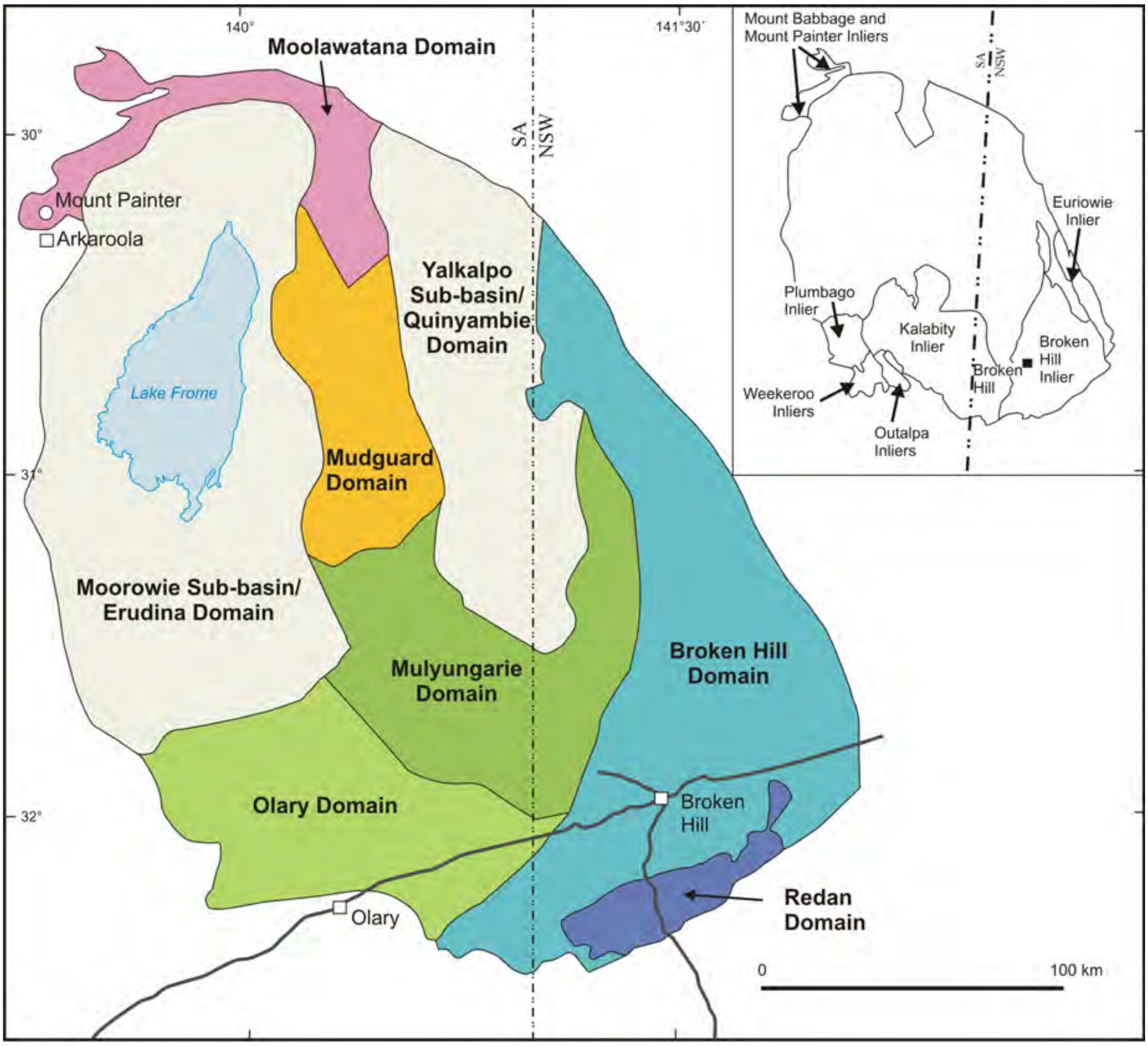

Figure 5. Subdivision of the Curnamona Province into geological domains defined by particular lithological, stratigraphic, geophysical or tectonic characteristics (after Conor and Preiss, 2008).

The Curnamona Province traditionally has been divided into the Broken Hill Domain and Olary Domain (Laing, 1996b), modified from the original Willyama Block of Thomson (1969). In this report, we follow Conor and Preiss (2008), who subdivide the Curnamona Province into the following geological domains, based on key differences in age, sedimentary facies and thickness, magmatism and metamorphism (Figure 5):

The Broken Hill Domain consists of relatively thick, essentially complete stratigraphy of the Willyama Supergroup intruded by Mesoproterozoic granite, exposed in the Broken Hill and Euriowie Inliers, and the southeastern portion of the Kalabity Inlier (Conor and Preiss, 2008; Figures 5 and 6).

The Redan Domain, located southeast of Broken Hill (Figure 5), contains rocks older than those in the Broken Hill Domain. It is characterised by calcalbitic Redan Gneiss, granulite-facies metamorphism, and a high Total Magnetic Intensity signature.

The Olary Domain is characterised by a thinner, less complete stratigraphy than in the Broken Hill Domain to the west (Figure 5, 6). The Olary Domain is recognised by restricted development of Broken Hill Group, and by the presence of the oldest known part of the Willyama Supergroup, the Curnamona Group (Conor and Preiss, 2008).

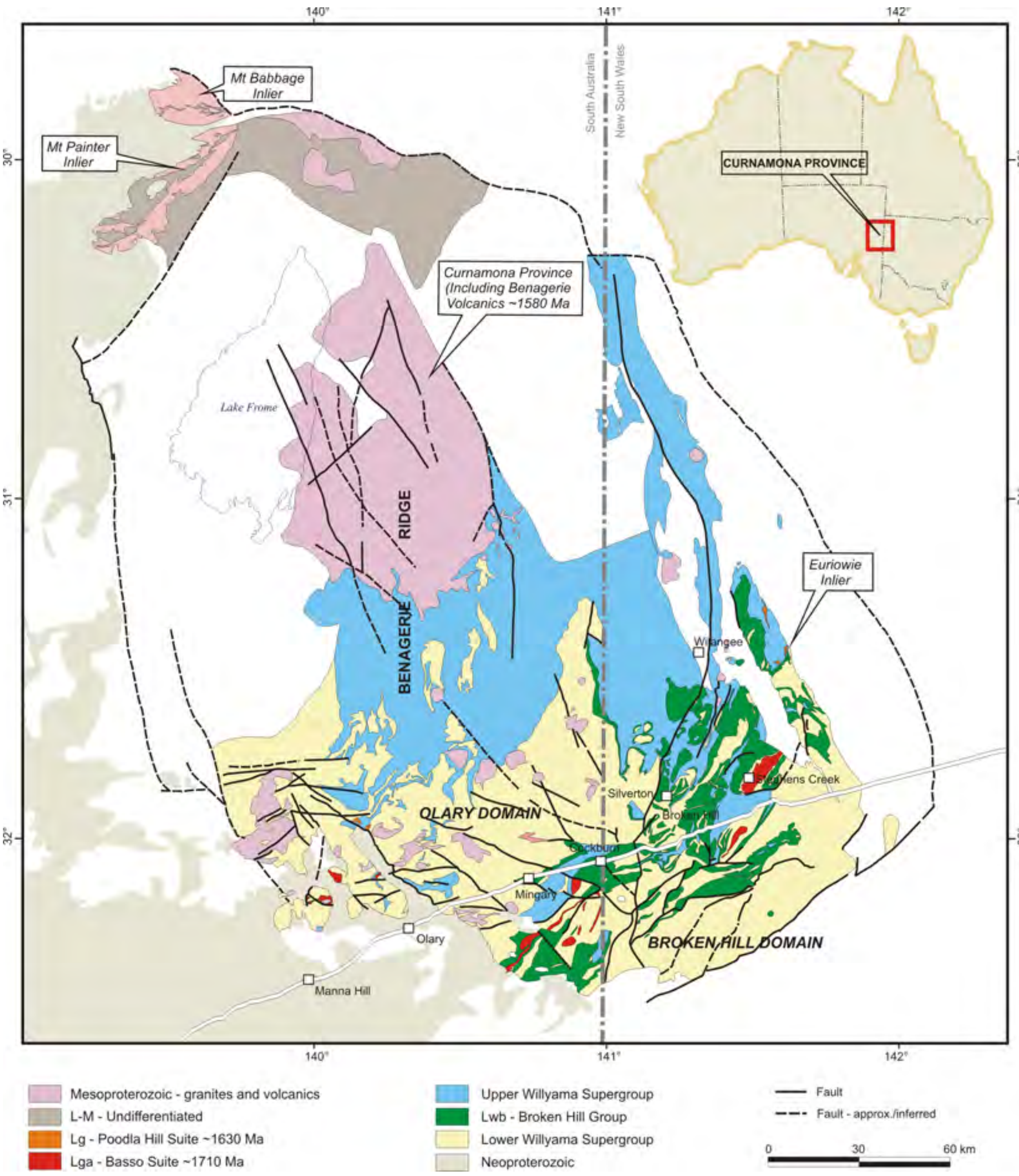

Figure 6. Outcrop and interpreted subsurface geology of the Curnamona Province (modified after Conor, 2006).

The Mulyungarie Domain occurs to the west of the Broken Hill Domain and to the north of the Olary Domain (Figure 5), and likely represents facies of the Willyama Supergroup. The domain is almost entirely blanketed by younger sediments, and has poorly defined boundaries.

The Moolawatana Domain, consists of ?Paleoproterozoic and early Mesoproterozoic rocks of the Mount Painter and Mount Babbage Inliers in the northwest, and their buried easterly extensions (Preiss, 2006).

The Mudguard Domain is characterised by the flat-lying Mesoproterozoic Benagerie Volcanics, inferred by geophysics to unconformably overlie deformed, granite-intruded Willyama Supergroup on the Benagerie Ridge (Preiss, 2006).

The Erudina and Quinyambie Domains consist of deeply buried and poorly known basement beneath the Cambrian Moorowie and Yalkapo Sub-basins of the Arrowie Basin, respectively (Conor et al., 2006; Conor and Preiss, 2008; Preiss, 2006). Their stratigraphy is not considered in this report.

These domain boundaries are subject to further refinement based on new mapping, drilling and seismic data. For example, the boundary between the Olary Domain and the Broken Hill Domain is currently being reassessed using solid geology, stratigraphic and geophysical interpretation on the Mingary 1:100 000 map sheet (Petrie et al., 2009; Crooks and Fricke, 2010).

In the following section, geological synthesis of the Gawler Craton and Curnamona Province is divided into Archean to earliest Paleoproterozoic (~3150-2400 Ma), Paleoproterozoic (~2020-1600 Ma) and Mesoproterozoic (~1600-~1450 Ma) time slices. Within these time slices, the geology is discussed by domain. Time-space plots for the Gawler Craton and Curnamona Province are presented, spanning the interval 3200 Ma to 1450 Ma (Figures 7, 8) and 1740 Ma to 1500 Ma (Figure 10), respectively.

Acknowledgements. The report was greatly assisted by detailed discussions on Gawler and Curnamona geology and geodynamics with R. Korsch, D. Champion, P. Henson, R. Blewett, G. Fraser, R. Skirrow and N. Neumann, (Geoscience Australia). We sincerely thank Wayne Cowley, Wolfgang Preiss, Anthony Reid, Tim Baker and Martin Fairclough (PIRSA) for their thorough reviews of the manuscript. We also thank Russell Korsch, Narelle Neumann, Geoff Fraser and David Huston who provided detailed internal reviews at Geoscience Australia.

SECTION 1. GEOLOGICAL SUMMARIES AND TIME-SPACE-EVENT PLOTS FOR THE GAWLER CRATON AND CURNAMONA PROVINCE

N. Kositcin

1.1. Archean to Earliest Paleoproterozoic – Gawler Craton

Geological and tectonic summary

Until recently, the oldest known rocks in the Gawler Craton were the Neoarchean to earliest Paleoproterozoic Mulgathing and Sleaford Complexes (Daly and Fanning, 1993; Figures 3, 7, 8), found in the Coulta, Cleve, Christie, Wilgena and Harris Greenstone domains. Together, the Sleaford and Mulgathing Complexes form an arcuate core to the Gawler Craton, with younger rocks both to the northeast and southwest. Recent SHRIMP U-Pb geochronology, however, has identified older (Mesoarchean) basement in the Spencer Domain (Fraser et al., 2008; 2010a), as discussed below.

The Coulta, Cleve, Christie, Wilgena and Harris Greenstone domains are composed predominantly of latest Archean or earliest Paleoproterozoic rocks, representing the original protolith on which subsequent tectonic units were superimposed (Parker et al., 1993). The Sleaford Complex occurs in the southeast of the Gawler Craton, on the Eyre Peninsula (Figure 3; Coulta and Cleve Domains), and consists of the Carnot Gneisses, Wangary Gneiss, the felsic Hall Bay Volcanics and high level intrusives of the Dutton Suite (Daly and Fanning, 1993).

In the northern Gawler Craton, within the Christie, Wilgena and Harris Greenstone domains, the Mulgathing Complex includes Archean metasedimentary rocks of the Christie Gneiss and meta-igneous Kenella Gneiss, calcalkaline andesite to rhyodacite volcanics of the Devils Playground Volcanics (Cowley and Fanning, 1991), and the pre- to syntectonic Glenloth Granite, as well as other mafic-ultramafic and orthogneiss bodies (Daly and Fanning, 1993; Figure 7, 8). The Harris Greenstone Domain contains supracrustal Archean komatiitic ultramafic and mafic volcanic rocks and Archean aluminous metasedimentary rocks, felsic extrusives and intrusives (Ferris et al., 2002; Hoatson et al., 2005), which are contemporaneous with, and form part of, the Mulgathing Complex.

All of these rocks experienced metamorphism during the ~2460 Ma to 2400 Ma Sleafordian Orogeny (Table 1; Daly and Fanning, 1993; Swain et al., 2005a). Very little is known of Sleafordian kinematics, as many original structures have been reworked by later events (Ferris et al., 2002). Sleafordian deformation is typically high-strain, characterised by isoclinal folding of the gneissic fabric and subsequent open, upright folding (Daly and Fanning, 1993; Tomkins and Mavrogenes, 2002). The most pervasive, peak granulite facies metamorphism occurred between 2440 Ma and 2420 Ma (Schwarz et al., 2006). The Devils Playground Volcanics and Harris Greenstone Domain were only metamorphosed to low metamorphic grades, although granulite facies metamorphic conditions (860°C at ~900 MPa, Daly et al., 1998) occurred in the Christie Gneiss during the Sleafordian Orogeny at 2447 ± 9 Ma (Fanning, 2002) to 2430 ± 6 Ma (Fanning, 1997). In the south of the craton, in the Sleaford Complex, Dutch et al. (2010) used EMPA monazite and garnet Sm-Nd geochronology to constrain the earliest stages of the Sleafordian event (D1) to ca. 2455 Ma, with metamorphic conditions of 5-7.5 kbar and 770-810˚C. Following the cessation of the Sleafordian Orogeny at ~2400 Ma, there was a period of relative tectonic quiescence until ~2000 Ma (Ferris et al., 2002).

Tectonic development of the late Archean Gawler Craton is considered by Swain et al. (2005a) to reflect interaction between a convergent margin and mantle plume, with deposition of sedimentary rocks occurring in a backarc or arc-rift environment. In their model, protoliths to the Christie Gneiss, Kenella Gneiss, Carnot Gneisses and Wangary Gneiss are envisaged to have been deposited between ca. 2535 Ma and ca. 2500 Ma in an extensional backarc setting, based on U-Pb zircon, geochemical and Sm-Nd isotopic data (Swain et al., 2005a). Extrusion of komatiites of the Harris Greenstone Domain occurred in the thinned backarc lithosphere, which was characterised by high heat flow. Trace element signatures of the Devils Playground Volcanics and Glenloth Granite (Mulgathing Complex), and the Hall Bay Volcanics, and Coulta Granodiorite of the Dutton Suite (Sleaford Complex) are characterised by distinctive Nb and Ti depletion anomalies which are attributed by Swain et al. (2005a) to indicate that they are part of a magmatic arc, associated with rollback of a steep subduction zone between ca. 2560 Ma and ca. 2500 Ma, behind which the protoliths to the Christie Gneiss and Carnot Gneisses were deposited. Possible shallowing of the subduction angle, or the incorporation of buoyant material led to a collisional regime, and allowed sediment deposition in the backarc to continue up until the beginning of orogenesis associated with the Sleafordian Orogeny between ca. 2460 Ma and 2400 Ma (Swain et al., 2005a). Thus, the Sleafordian Orogeny is interpreted to represent a change from extension to contraction in the overriding plate, with Swain et al. (2005a) suggesting that collision occurred due to closure of an ocean and arrival of a continent or continental fragment.

Geological history and explanation of Time-Space plot

Christie and Wilgena Domains

- The Christie Domain is characterised by predominantly latest Archean metasedimentary rocks of the Mulgathing Complex, which were metamorphosed to granulite facies in the earliest Proterozoic at 2480-2420 Ma (Sleafordian Orogeny; McFarlane, 2006). These metasedimentary rocks are collectively termed the Christie Gneiss, and were derived, at least in part, from an existing continental basement (Daly et al., 1998). The Mulgathing Complex also includes Archean metasedimentary rocks of the Kenella Gneiss, komatiite and tholeiitic basalt flows of the Harris Greenstone Domain (see below), the Devils Playground Volcanics and the pre- to syntectonic Glenloth Granite (Daly and Fanning, 1993; see below). The Christie Gneiss has been correlated with the Sleaford Complex of similar age in the Coulta Domain (Daly and Fanning, 1993; Daly et al., 1998; Swain et al., 2005a).

- Metasedimentary rocks of the Christie Gneiss include banded iron formation, chert, carbonate, calcsilicate, quartzite and aluminous sedimentary rocks (Daly and Fanning, 1993; Daly et al., 1998). Jagodzinski et al. (2009) report SHRIMP U-Pb data for three samples of Christie Gneiss, with maximum depositional ages of 2514 ± 4 Ma (at Golf Bore prospect), 2485 ± 4 Ma (at Mount Christie), and 2484 ± 7 Ma (at Tallaringa). These data indicate that at least some, and possibly all, of the metasedimentary successions in the Mulgathing Complex are significantly younger than previously interpreted (i.e., maximum depositional age of ca. 2515 Ma recorded by McFarlane et al., (2007) at Challenger Mine, and Swain et al., (2005b) for the Kenella Gneiss, southeast of Tarcoola; see below). Quartz norite from the Christie Domain gave a $^{207}Pb/^{206}Pb$ age of 2489 ± 4 Ma, considered to be the magmatic crystallisation age of the quartz norite protolith (Jagodzinski et al., 2009). Peak granulite facies metamorphism occurred in the Christie Gneiss during the Sleafordian Orogeny at 2447 ± 9 Ma (Fanning, 2002) to 2437 ± 11 Ma and 2430 ± 6 Ma (Fanning, 1997). A monazite age of 2416 ± 12 Ma was reported by Zang (2002) for the Christie Gneiss, and McFarlane (2006) provided a spread of monazite ages from 2454 Ma to 2428 Ma, with one slightly younger age at 2414 Ma, interpreted as the timing of high grade metamorphism and partial melting. Peak pressure and temperature conditions of the Christie Gneiss are estimated at ~7.5 ± 1.5 kbar and 800-850˚C (Tomkins et al., 2004).

- The Kenella Gneiss (Reid et al., 2007), consists of massive to layered K-feldspar-plagioclase-quartz gneiss, interpreted by Fanning et al. (2007) to be largely volcanic or volcaniclastic in origin. Previous SHRIMP geochronology on two samples of the Kenella Gneiss from DDH

Kenella 1A, yielded ages ranging between ca. 3540-2510 Ma, with an interpreted maximum depositional age of ca. 2535 Ma (Swain et al., 2005a). Analyses of metamorphic zircon overgrowths showed a spread of ages down concordia, with ages in the range ca. 2480 to 2420 Ma (Swain et al., 2005a). The minimum age of deposition is constrained by intrusion of the ca. 2500 Ma Glenloth Granite (see below; Daly and Fanning, 1993; Swain et al., 2005a). More recent SHRIMP U-Pb zircon data from a sample of Kenella Gneiss at Tarcoola provided inconclusive results, with ages of ca. 2495 Ma, 2507 ± 8 Ma or 2522 ± 5 Ma, possibly recording a magmatic age or the maximum depositional age of immature sediment (Jagodzinski et al., 2009). SHRIMP U-Pb zircon ages for samples of felsic (Kenella Gneiss?) and mafic metaigneous gneisses from the Aristarchus prospect (a Ni-Cu sulfide prospect hosted by an interpreted layered mafic-ultramafic intrusive complex) were reported by Jagodzinski et al. (2009). These data indicate that the magmatic protolith to the felsic gneiss crystallised at 2479 ± 3 Ma, and underwent high grade metamorphism during the Sleafordian Orogeny, possibly over an extended period of time, beginning at 2460 ± 4 Ma, and culminating at 2414 ± 7 Ma. Metagabbro from the Aristarchus prospect crystallised at 2494 ± 9 Ma and underwent high grade metamorphism during the Sleafordian Orogeny (Jagodzinski et al., 2009).

- The sedimentary facies of the Christie Gneiss and Kenella Gneiss suggest a depositional setting of shallow to moderate marine water depth, and Reid et al. (2007) suggest that deposition of these packages occurred within a restricted, actively subsiding, continental shelf setting.

- The Devils Playground Volcanics (Reid et al., 2009) are a bimodal suite of rhyodacite and andesite volcanic rocks (Cowley and Fanning, 1991), which have a SHRIMP U-Pb zircon age of 2553 ± 9 Ma (Fanning, 1997). Geochemically, these volcanics are calcalkaline and isotopically moderately juvenile, with $\varepsilon Nd_{(2558\ Ma)}$ values that range from +2.3 to +3.1 (Cowley and Fanning, 1991; Swain et al., 2005a; Reid et al., 2009). Swain et al. (2005a) considered that the Devils Playground Volcanics (and the Hall Bay Volcanics) formed part of a magmatic arc associated with rollback of a steep subduction zone, behind which protoliths to the Christie Gneiss and Carnot Gneiss were deposited as clastic sediments in an extensional backarc setting.

- Metanorite from the Blackfellow Hill Pyroxenite gives a zircon crystallisation age of 2461 ± 5 Ma and metamorphic ages of 2435 ± 11 Ma and 2425 ± 7 Ma, indicating it was emplaced and metamorphosed during the Sleafordian Orogeny (Fanning et al., 2007).

- The Glenloth Granite, composed of medium- to coarse-grained, leucocratic gneiss or gneissic granitoid, ranging in composition from granite to monzogranite and granodiorite (Blissett, 1985), is dated at 2499 ± 11 Ma by Fanning (1997). The Glenloth Granite also contains younger phases, at 2458 ± 10 Ma and 2435 ± 17 Ma (Fanning, 1997). The Glenloth Granite is interpreted to have evolved in the same magmatic system as the Devils Playground and Hall Bay Volcanics and Coulta Granodiorite, by Swain et al. (2005a). The Mobella Tonalite records SHRIMP U-Pb zircon magmatic crystallisation ages of 2496 ± 4 Ma and 2490 ± 4 Ma, and high grade metamorphism at 2466 ± 5 Ma (Jagodzinski et al., 2009).

Harris Greenstone Domain

- Komatiite and tholeiitic basalt flows (Hoatson et al., 2005), and pyroxenite and peridotite sills, are inferred to be contemporaneous with sedimentation of the Mulgathing Complex. The steeply dipping greenstone succession at Lake Harris include ultramafic-mafic extrusive rocks (Lake Harris Komatiite, Hopeful Hill Basalt, South Lake Gabbro), cumulate komatiite, high to low Mg komatiite, komatiitic and tholeiitic basalt, pyroxene cumulates, felsic volcanics, minor pelitic metasedimentary rocks and banded iron formation (Schwarz, 2004; Hoatson et al., 2005). Sedimentary structures suggest that the metasedimentary rocks associated with the mafic-ultramafic extrusive rocks in this domain might have been deposited in fluvial-estuarine to deltaic-shelf environments (Zang, 2007).

- Rhyodacite from the Harris Greenstone Domain provides a SHRIMP U-Pb age of 2509 ± 3 Ma, which constrains the time of igneous zircon crystallisation, and an age for volcanic activity (Fanning et al., 2007).

- Zircon from a volcaniclastic rock interbedded with komatiite in the Harris Greenstone Domain has a U-Pb SHRIMP zircon age of 2522 ± 8 Ma (Swain et al., 2005a). The older age limit of the succession is constrained by detrital zircons in a metasedimentary rock with an age of 2530 ± 8 Ma (Fanning, 2002).

- A rhyodacite dyke cutting komatiite in the Harris Greenstone Domain has an age of 2509 ± 3 Ma (Fanning, 2002), and is either a younger component of the magmatism associated with the komatiites or related to the Glenloth Granite, a quartz diorite vein from which has an age of 2499 ± 11 Ma (Fanning, 1997). The South Lake Gabbro is interpreted to be intrusive into the volcano-sedimentary pile prior to deformation, and is likely to represent a basaltic sill or dyke (Reid et al., 2007). ID-TIMS zircon dating by Fanning (1997) yielded a metamorphic age of 2440 ± 9 Ma for the South Lake Gabbro.

- These metamorphosed mafic and ultramafic rocks are inferred to represent regional attenuation of the Archean crust and may indicate the presence of oceanic crust during sedimentation (Budd, 2006). Daly et al. (1998) considered that the komatiites and basalts of the Harris Greenstone Domain formed during a period of extreme extension of Archean continental crust, which Hoatson et al. (2005) considered to have occurred in an intraplate setting. Geochemical and juvenile isotopic signatures in the komatiites ($\varepsilon Nd_{(2500\ Ma)}$ +2 to +4; Fanning, 2002) are indicative of a mantle plume source, emplaced between ca. 2520 Ma and 2510 Ma. Alternatively, Hand et al. (2002) and Swain et al. (2005a) considered that sufficient thinning of the lithosphere to allow the extrusion of komatiites and basalts occurred in a continental backarc setting, possibly linked to a mantle plume, at about 2522 Ma. In summary, sedimentation and ultramafic-mafic volcanism within the greenstone belt occurred at ~2520 Ma, followed by metamorphism and deformation at ca. 2440-2450 Ma during the Sleafordian Orogeny (Fraser and Reid, 2007).

Coulta and Cleve Domains

- The Sleaford Complex consists of late Archean medium- to high- metamorphic grade metasedimentary rock successions of the Carnot Gneisses, Wangary Gneiss, Hall Bay Volcanics and high-level intrusives of the Dutton Suite (Daly and Fanning, 1993; Swain et al., 2005a).

- The Carnot Gneisses consist of garnet quartz-feldspar paragneiss, augen gneiss and garnet gneiss (Daly and Fanning, 1993), with boudins of mafic granulite interpreted as tholeiitic basalt flows or gabbro sills emplaced prior to metamorphism (Daly et al., 1998; Tong et al., 2004). The paragneisses yield detrital U-Pb zircon ages between ca. 2850 Ma and ca. 2300 Ma, with a few individual analyses at ca. 3150-2950 Ma (Fanning, 1997). The Carnot Gneisses experienced high T- low P (770-800˚C at 5.5-7.5 kbar) granulite facies metamorphism during the Sleafordian Orogeny, and were subsequently reworked at high T- medium P (800-850˚C at 8-10 kbar) to high T- low P (800-850˚C at <6 kbar) granulite facies conditions during the Paleoproterozoic Kimban Orogeny (Dutch et al., 2010).

- The Wangary Gneiss consists of amphibolite facies, K-feldspar-plagioclase gneisses which are possibly lower grade equivalents of the Carnot Gneisses (Daly and Fanning, 1993). Protolith ages for rocks in the Sleaford Complex are poorly constrained, an exception being a SHRIMP age of 2679 ± 9 Ma for the Wangary Gneiss, interpreted by Fanning (1997) as being the age of crystallisation of the igneous protolith. Swain et al. (2005a) provides a maximum age of ca. 2510 Ma, and a minimum age for protolith deposition of ca. 2480 Ma, which is constrained by metamorphic zircon growth during the Sleafordian Orogeny (Swain et al., 2005a). Gneisses from Refuge Rocks are also as part of the Sleaford Complex with ages of 2411 ± 5 Ma (Fanning et al., 2007), and 2418 ± 7 Ma, interpreted as the crystallisation age of the igneous protolith to the gneiss (Fraser and Neumann, 2010). Recent age dating of the Minbrie Gneiss from near Mt Ghearthy and Elbow Hill, have maximum depositional or igneous protolith ages of ~2520 Ma and ~2455 Ma, respectively (plus older inheritance), indicating that this formation is also part of the Sleaford Complex. Both samples also have metamorphic populations consistent with high-grade metamorphism during the Kimban Orogeny (1762 ± 43 Ma and 1733 ± 7 Ma, Fraser and Neumann, 2010).

- SHRIMP U-Pb analysis of a migmatitic garnet-biotite gneiss at Waddikee Rocks (Coulta Domain) produced a range of zircon ages, interpreted to reflect source components in a metasedimentary rock. The youngest age grouping of 2597 ± 8 Ma is interpreted as a maximum age constraint on the timing of deposition of the sedimentary rocks (Fraser and Neumann, 2010).

- The Hall Bay Volcanics consist of low metamorphic grade aluminous metapelites interlayered with felsic volcanic and volcaniclastic units (Teale et al., 2000). The Hall Bay Volcanics have U-Pb zircon SHRIMP ages of 2529 ± 4 Ma (Swain et al., 2005a), 2524 ± 5 Ma and 2517 ± 7 Ma, with significant inheritance of ca. 2720 Ma (Fanning et al., 2007), and they have fractionated trace element patterns coupled with significant Nb and Ti depletion anomalies (Swain et al., 2005a).

- The pre- to syntectonic Dutton Suite was intruded into latest Archean to earliest Proterozoic sedimentary rocks (Fanning et al., 2007). Coeval with the Hall Bay Volcanics are older granites of the Dutton Suite, such as the Coulta Granodiorite, which has a SHRIMP U-Pb zircon age of 2519 ± 8 Ma (Fanning et al., 2007). The Dutton Suite also contains younger members, such as, the Kiana and Whidbey Granites (Daly et al., 1998; Daly and Fanning, 1993). The Kiana Granite has intrusive ages of 2466 ± 11 Ma and 2462 ± 15 Ma (Fanning et al., 2007).

- Fraser and Neumann (2010) reported U-Pb SHRIMP zircon ages of 2823 ± 37 Ma and 2427 ± 5 Ma for a sample of the Coolanie Gneiss (at Pinerow, Cleve Domain), interpreted as recording the timing of igneous crystallisation and high grade metamorphism, respectively. A sample of basement gneiss from Menninnie Dam (drillhole MD093) produced a dominant zircon age population of 2511 ± 4 Ma, interpreted to record the crystallisation age of the igneous protolith to the gneiss (Fraser and Neumann, 2010).

- Previous attempts to date the Carpa Granite (Cleve Domain, considered to be part of the Moody Suite, intruded during the Kimban Orogeny, ~1740-1690 Ma; Parker et al., 1993), have been unsuccessful (i.e., 1677 ± 125 Ma Rb-Sr, Webb et al., 1986; no coherent age population, Fanning et al., 2007). Fraser and Neumann (2010) report an igneous crystallisation age for the Carpa Granite of 2518 ± 8 Ma (although only defined by 4 analyses on poor quality zircon), with minor ~3400 Ma inheritance. A second sample gave an interpreted igneous protolith age of 2480 ± 6 Ma, with possible Sleafordian metamorphism recorded by groupings at 2439 ± 8 Ma and 2418 ± 9 Ma. New zircon growth occurred at 1724 ± 11 Ma, and is interpreted to be related to partial melting or melt injection from regional granites during the Kimban Orogeny (Fraser and Neumann, 2010).

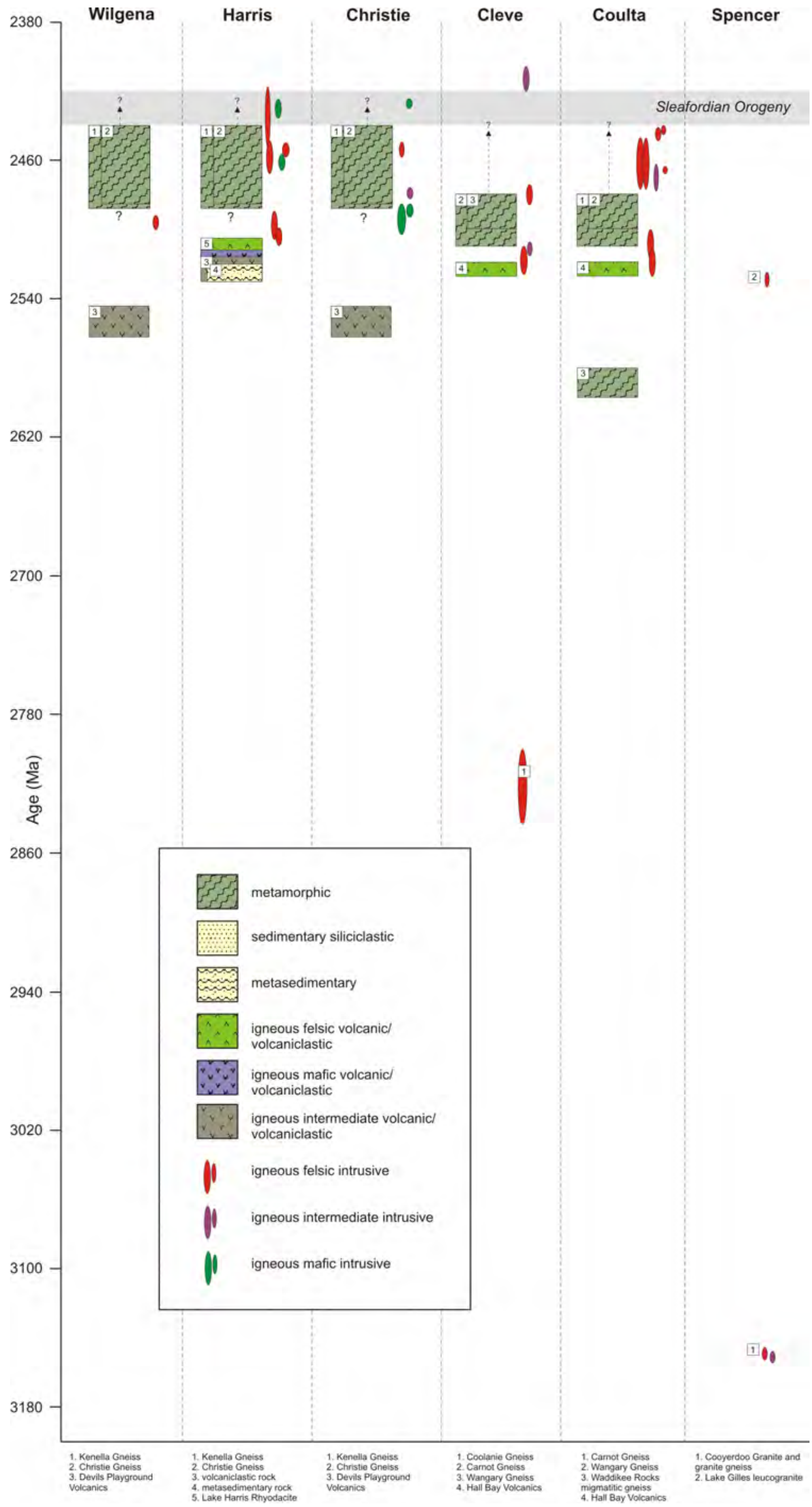

Figure 7. Time-space-event plot for the Archean to earliest Paleoproterozoic of the Gawler Craton.

- Granulite facies metamorphism occurred during the Sleafordian Orogeny, and associated extensive syntectonic granites, tonalites and norites have been dated at ~2450 Ma (Fanning, 1997).

Spencer Domain

- Intrusion of Mesoarchean gneissic granite occurred at ~3150 Ma (four samples with U-Pb SHRIMP zircon ages of 3149 ± 3 Ma, 3152 ± 19 Ma, 3155 ± 4 Ma, 3158 ± 2 Ma; Fraser et al., 2010a; Fraser and Neumann, 2010), forming basement to the Middleback Ranges. The ~3150 Ma age for the granite (Fraser et al., 2008; 2010a), formally known as the Cooyerdoo Granite (Reid et al., 2008a), represents the first direct evidence of basement of this age in the Gawler Craton, previously inferred by the presence of inherited zircons and extremely evolved Nd isotopic ratios in Proterozoic granites (Creaser and Fanning, 1993; Daly and Fanning, 1993; Belousova et al., 2006). Further rocks of Archean age are likely to be discovered, and ongoing work is in progress to define the regional extent of rocks of this age (Fraser et al., 2008; Fraser et al., 2010a). The presence of inherited zircons as old as ~3300 Ma within these ~3150 Ma granitic rocks implies the presence of Paleoarchean source regions, possibly still present at depth (Fraser et al., 2010a).

- Fraser and Neumann (2010) report U-Pb SHRIMP zircon results for an unnamed gneissic granite (quartz-Kfeldspar-plagioclase-biotite granite) located ~8 km west-northwest of Iron Knob. An age of 3151 ± 3 Ma is interpreted as the timing of igneous crystallisation, and an age of 2510 ± 24 Ma from the same sample is interpreted as recording the timing of high-grade metamorphic reworking and partial melting (Fraser and Neumann, 2010).

- A foliated leucogranite, from the northern shore of Lake Gilles, gave an interpreted U-Pb SHRIMP zircon crystallisation age of 2529 ± 4 Ma (Fraser and Neumann, 2010).

Table 1. Summary of deformational-metamorphic events in the Gawler Craton (after Hand et al., 2007).

Event	Distribution	Age range (Ma)	Metamorphic style	P-T conditions	Reference	P-T evolution	Magmatism
Sleafordian Orogeny	Entire proto-craton	~2480-2420 Ma	High heat flow, low to moderate P	Greenschist to 850˚C, 6 kbars	Tomkins and Mavrogenes (2004)	Unknown	Felsic
Miltalie Event	Eastern (?) craton	~2000 Ma	Unknown	Unknown	Fanning et al. (1998)	Unknown	Felsic
Cornian Orogeny	Eastern craton	~1860-1850 Ma	High heat flow, moderate P	750˚C, 6 kbars	Reid et al. (2008b)	Clockwise	Felsic, minor mafic
Kimban Orogeny	Entire craton	~1740-1690 Ma	High to low heat flow	Eyre Peninsula: greenschist to 800˚C-850˚C, 7-9 kbars; Northern and western Gawler Craton: low to medium-P amphibolite to granulite	Parker et al. (1993) Tong et al. (2004) Teasdale (1997) Payne et al. (2006b)	Anti-clockwise?	Felsic and mafic
Ooldean Event	Western craton	~1660-1630 Ma	High heat flow, low P	900˚C, 10 kbars	Teasdale (1997)	Isobaric cooling	Felsic?
Deformation associated with St Peter Suite	Southwestern craton	~1620-1610 Ma	High heat flow	Amphibolite	Hand et al. (2007)	Unknown	Felsic
Deformation associated with Hiltaba Suite	Central craton	~1595-1575 Ma	High heat flow	Greenschist to medium P (9 kbar) granulite	Holm (2004) Conor (1995) Ferris and Schwarz (2003)	Clockwise	Felsic and mafic
Kararan Orogeny	North and western craton	~1570-1540 Ma	High heat flow	Coober Pedy Ridge + Mabel Creek Ridge: medium-P amphibolite to granulite; Fowler Domain: 800˚C, 10 kbars	Daly et al. (1998) Teasdale (1997)	Clockwise	Unknown
Coorabie Orogeny	Northern and western craton	~1470-1450 Ma	High to moderate heat flow	500-800˚C, <4-6 kbar	Teasdale (1997)	Clockwise	Felsic

1.2. Paleoproterozoic – Gawler Craton

Geological and tectonic summary

After 400 million years of tectonic quiescence following the Sleafordian Orogeny, the next event recognised in the Gawler Craton occurred at approximately 2020-2000 Ma, at which time protoliths of the Miltalie Gneiss were intruded (Daly et al., 1998; Howard et al., 2006; Fanning et al., 2007) in the southern Gawler Craton (Cleve Domain). A period of sediment deposition in the Eastern Gawler Craton followed, with sedimentary rocks of the Hutchison Group, forming the largest basin succession in the eastern Gawler Craton, being interpreted to have been deposited on a passive margin in the time interval 2000-1860 Ma (Parker et al., 1993; Schwarz et al., 2002; Figure 8). The first major deformational and metamorphic event after the Sleafordian Orogeny was the Cornian Orogeny (Reid et al., 2008b; Table 1), a distinct event from ~1860-1850 Ma which produced east-southeast striking structural fabrics overprinted by east-west striking folds and late south-side down extensional ductile shearing (Reid et al., 2008b). Tectonism at ~1860-1850 Ma was previously termed the Neill Event (Ferris et al., 2002), or Lincoln Orogeny (Vassallo and Wilson, 2002). Metamorphism associated with the Cornian Orogeny is represented by a clockwise P-T path with peak granulite metamorphic conditions of c. 750˚C and c. 6 kbar (Reid et al., 2008b). The extent of the Cornian Orogeny is so far restricted to a small region east of the Kalinjala Mylonite Zone in the Yorke Peninsula region (Reid et al., 2008b; Payne et al., 2009).

Synchronous with the Cornian Orogeny was the intrusion of the voluminous Donington Granitoid suite at ca. 1850 Ma - an intrusive magmatic complex which occupies a north-south non-continuous trending belt within the Spencer and Olympic domains along the eastern margin of the Gawler Craton (Figure 3, 8; Schwarz, 2003). Following the Cornian Orogeny, the interval 1800-1740 Ma marks a time of extensive sediment deposition across much of the Gawler Craton. In the northern and western Gawler Craton, this sedimentation and volcanism is preserved in metasedimentary successions in the Fowler, Nawa, Peake and Denison and Mount Woods Domains (Payne et al., 2008b). Widespread deposition across the eastern Gawler Craton continued over the interval ca. 1770 to 1740 Ma (e.g., Wallaroo Group in the southeastern Olympic Domain of the Gawler Craton), with magmatic suites associated with these depositional basins being bimodal in nature (Hand et al., 2007). Sedimentation concluded before the onset of magmatism, deformation and metamorphism associated with Kimban Orogeny at ca. 1730-1690 Ma (Daly et al., 1998; Vassallo and Wilson, 2001, 2002; Ferris et al., 2002; Swain et al., 2005b; Payne et al., 2008a).

The Kimban Orogeny was defined originally as a relatively long-lived event (e.g., 1850-1540 Ma, Glen et al., 1977; ~1820-1580 Ma, Parker and Lemon, 1982; 1845-~1700 Ma, Daly et al., 1998), and was interpreted as the most pervasive orogenic event in the Gawler Craton (Daly et al., 1998; Fanning et al., 2007; Hand et al., 2007). On the eastern margin of the Gawler Craton, the Kimban Orogeny records bulk non-coaxial deformation associated with dextral transpression (Hand et al., 1995; Vassallo and Wilson, 2002), and was largely responsible for shaping the current crustal architecture (Swain et al., 2005b). More recent geochronology has confirmed it to be a cratonwide greenschist to granulite grade event with metamorphic and syntectonic magmatic ages confined to the range 1740-1690 Ma (Table 1; Fanning et al., 2007). High grade metamorphism, constrained by zircon and monazite geochronology, is reported from the Fowler, Nawa, Peake and Denison, Mount Woods, Cleve and Coulta Domains (Payne et al., 2008b). Specifically, the orogeny is represented by intrusion of the Middlecamp Granite at ~1730 Ma (Fanning et al., 2007; Fraser and Neumann, 2010), granulite facies metamorphism within the Kalinjala Mylonite Zone at ~1730 Ma (Hand et al., 1995) and intrusion of the late- to post-tectonic Moody Suite at ~1700 Ma (Fanning, 1997; Fanning et al., 2007).

There were two tectonic events associated with the Kimban Orogeny, KiD1 and KiD2 (Hoek and Schaefer, 1998; Vassallo and Wilson, 2001, 2002). The first deformation event, KiD1, was a high-grade metamorphic event producing tight isoclinal to recumbent folding with strong axial planar schistosity

(Parker and Lemon, 1982; Daly et al., 1998; Betts and Giles, 2006). This event was also associated with sheath folds and dextral transpression preserved in metasedimentary rocks of the Hutchison Group (Vassallo and Wilson, 2001, 2002). The second phase (KiD2) is defined by low-grade metamorphism, and was less intense than KiD1. Open folding, about northeast-southwest axes, produced upright, open to tight regional folds with variable plunges and major mylonite zones (Parker and Lemon, 1982; Daly et al., 1998; Betts and Giles, 2006). A collisional tectonic model was proposed by Tong et al. (2004) for KiD1 and KiD2 of the Kimban Orogeny. Swain et al. (2005b) states that the Kalinjala Mylonite Zone was the focus of a narrow, high strain orogenic belt that reworked older crust, while the similarity in timing of dextral transpression along both Kalinjala and Tallacootra Shear Zones implies the existence of a Gawler Craton-wide, dextral transpressional regime between ca. 1690 Ma and 1680 Ma (based on EMPA monazite ages; Swain et al., 2005b).

Metamorphic conditions during the Kimban Orogeny are poorly constrained, but are represented by 625-650˚C/5.5-6.5 kbar to 700-750˚C/8-9 kbar in the western Gawler Craton (Teasdale, 1997) and 600-675˚C/5-7 kbar to 800-850˚C/7-9 kbar in the southern Gawler Craton (Parker et al., 1993; Tong et al., 2004; Dutch et al., 2010). The Kimban Orogeny was interpreted by Betts et al. (2002, 2003a), Giles et al. (2004) and Betts and Giles (2006) to record the collision of the Archean nucleus of the Gawler Craton and its passive margin (Hutchison Group?), with the North Australian Craton, during north-northwest-dipping subduction between ca. 1740 Ma and 1690 Ma.

The period following the Kimban Orogeny and associated magmatism (late to post-tectonic magmatism of the 1690-1670 Ma Tunkillia Suite; Ferris, 2001; Ferris et al., 2002) is represented by a period of sedimentation in the central Gawler Craton (e.g., ca. 1660 Ma Tarcoola Formation, and possibly the Corunna Conglomerate; Daly et al., 1998; Figure 8). A high-pressure metamorphic event in the western Gawler Craton, poorly constrained in age at around 1660-1630 Ma, has been named the Ooldean Event by Hand et al. (2007). This event is marked by ultra high-T metamorphism (>950˚C and >9.5 kbar) in the Nawa Domain at around 1659 ± 6 Ma (Teasdale, 1997; Daly et al., 1998; Fanning et al., 2007; Table 1). Daly et al. (1998) originally termed this event the Kararan Orogeny but suggested it extended from 1650 Ma to 1540 Ma. This age range now incorporates a number of discrete events and the Kararan is now restricted to the orogenic event at the end of this time period (1570-1540 Ma, Hand et al., 2007; see next section). The Ooldean Event was followed shortly thereafter by the development of arc-like tonalitic to granodioritic magmatism of the St Peter Suite between 1630 Ma and 1615 Ma, which dominates the southwestern part of the craton (Swain et al., 2008).

Geological history and explanation of the Time-Space plot

Nawa Domain

- Geological observations and geochronological constraints for the Nawa Domain are limited to samples from several basement-intersecting drillholes. The southern Nawa Domain is characterised by a north-west trending, highly magnetised belt of rocks known as the Moondrah Gneiss (Daly et al., 1998), which consists of variably mylonitised, iron-rich, magnetite-bearing, migmatitic gneisses that preserve UHT metamorphic conditions (~950°C, 9.5 kbar). A sample of the Moondrah Gneiss from diamond drillhole Ooldea 2 yielded a U-Pb zircon age of 1659 ± 6 Ma, interpreted as a metamorphic zircon age (Daly et al., 1998; Fanning et al., 2007; the ~1660 Ma Ooldean Event of Hand et al., 2007); the sample contained inherited zircons at ~1700 Ma (Fanning et al., 2007). Monazite from drillhole Ooldea 1 yields an EMPA age of 1640 ± 12 Ma (Swain et al., 2005b). It is unknown whether inherited zircons at ~1700 Ma represent detrital zircons, providing a maximum deposition age for sedimentary protoliths to the Moondrah Gneiss or whether they represent an earlier phase of metamorphism (Fraser and Reid, 2007). Payne et al. (2008a) reported monazite ages of ~1690 Ma from Ooldea 2, and Fraser and Reid (2008) reported a zircon age of ~1690 Ma from Ooldea 2. Cutts et al. (2010)

constrain a younger high-strain UHT event at ca. 1660-1650 Ma using LA-ICPMS and EMPA dating of monazite.

- LA-ICP-MS analyses of zircon from drillhole AMPB3 yielded a range of ages, with the majority between ~1900 Ma and ~1700 Ma, and a few analyses with ages ~2500-2400 Ma (Payne et al., 2006b). Several samples dated by Payne et al. (2008a) record metamorphic monazite growth in the period 1730 Ma to 1690 Ma, coincident with the Kimban Orogeny.

- Quartz-feldspar-biotite-hornblende paragneiss sampled in drillhole GOMA DH4, gave a U-Pb SHRIMP zircon age pattern consistent with derivation from Mulgathing-aged sources (i.e., peaks at ~2460 Ma, ~2500 Ma, ~2518 Ma and ~2540 Ma), and an interpreted metamorphic age of 1523 ± 10 Ma from zircon rims (Jagodzinski et al., 2010).

- Based on limited data, Payne et al. (2006, 2008a) suggested that a minimum age for deposition of sedimentary rocks in the Nawa Domain is provided by metamorphic zircon with ages of ~1720 Ma, while U-Pb detrital zircon analysis was used to constrain deposition of sedimentary protoliths in the Nawa Domain to between ca. 1750 Ma and 1720 Ma (Payne et al., 2008a). Sedimentary rocks are shown to be enriched in REE, and sourced from evolved crust (εNd -7.0 to -4.3) (Payne et al., 2006a). On the basis of geochemical, Nd-isotope and U-Pb detrital zircon age data, Payne et al. (2006a) suggested that the Paleoproterozoic metasedimentary succession in the Nawa Domain may have been derived from central Australian terranes (i.e., Arunta Region).

Coober Pedy Domain

- A quartz-feldspar-biotite-magnetite orthogneiss, sampled in drillhole GOMA DH1 (~307 m), gave an interpreted U-Pb SHRIMP intrusive age of 1919 ± 9 Ma, with metamorphic rims at 1690 ± 7 Ma (Jagodzinski et al., 2010).

- Samples from drillhole CR9125 are interlayered ironstones and aluminous granulites, with peak P-T conditions of ~9 kbar and 950°C (Daly et al., 1998). A metagabbro from drillhole DD89LR22 yielded a SHRIMP U-Pb magmatic emplacement age of 1725 ± 7 Ma (Hand et al., 2007). A granodiorite gneiss from drillhole CR9119 yielded a magmatic crystallisation age of ~1590 Ma (Fanning et al., 2007).

Mount Woods Domain

- Supracrustal successions of interlayered banded iron formation, leucocratic gneiss, psammitic and pelitic schists and calcsilicates of the Mount Woods Complex were deposited in the Paleoproterozoic (Betts et al., 2003a). The bulk of the metasedimentary units in the Mount Woods Complex have been placed in a new group termed the Skylark Metasediments (Chalmers, 2007). The Mount Woods Complex incorporates the Skylark Metasediments, the Coodnambana Metaconglomerate and various intrusive units (e.g., Engenina Adamellite; Chalmers, 2007).

- The depositional age of the Mount Woods Complex is poorly known, with one sample of quartzofeldspathic gneiss yielding an upper intercept concordia age of 1742 ± 27 Ma, interpreted to be the timing of magmatic crystallisation (Fanning et al., 1988). SHRIMP U-Pb dating of pelitic gneiss and psammite samples from the Skylark Metasediments give maximum depositional ages of 1752 ± 6 Ma and 1749 ± 6 Ma (Jagodzinski et al., 2007; Chalmers, 2007). A maximum depositional age of 1749 ± 6 Ma is derived from the Coodnambana Metaconglomerate, which unconformably overlies units of the Skylark Metasediments, which also gave a metamorphic age of 1595 ± 10 Ma from metamorphic zircon rims (Jagodzinski et al., 2007; Chalmers, 2007).

- A second sedimentary package, of gneisses, conglomerates and sandstones, is inferred to unconformably overlie the Mount Woods Complex in this domain (Betts et al., 2003a). Samples from drillholes (Engenina 20 and 38) have detrital zircon age clusters from ~1860 Ma to ~1750 Ma and down to ~1670 Ma (O. Holm, OZCHRON). Several other samples (drillholes Warriner Creek 1 and Engenina 61) have detrital zircons with ages of ~1720 Ma and ~1650 Ma (O. Holm, OZCHRON), and indicate that sedimentary deposition of the protoliths of the gneisses in these drilholes postdate ~1650 Ma (Skirrow et al., 2006). Betts et al. (2003a) tentatively correlated this second sedimentary package with the Tarcoola Formation (see Wilgena Domain).

- The initial Kimban deformational phase (KiD1) produced isoclinal folding along with a bedding-parallel and axial-planar gneissic foliation in the Mount Woods Complex (Betts et al., 2003a). Deformation was accompanied by high-temperature, upper amphibolite to granulite metamorphism dated at 1736 ± 14 Ma by Fanning (1997).

- Granite sampled from a drillhole in the Mount Woods Domain has a SHRIMP zircon age of 1708 ± 3 Ma (Holm, 2004). The Engenina Adamellite is a synorogenic (KiD2, Betts et al., 2003a) foliated granite, which intruded the supracrustal metamorphic package in the Mount Woods Domain, and has a U-Pb zircon age of 1691 ± 25 Ma (Finlay, 1993; Fanning, 1997).

Christie Domain

- No Paleo- or Mesoproterozoic sedimentary rocks are known from the Christie Domain. The history of the Christie Domain after ~2400 Ma is one of variable metamorphic reworking, intrusion by sparsely distributed granites, and cooling, presumably related to exhumation.

- Granites equivalent in age to the Tunkillia Suite occur in this domain. For example, a granite from Wynbring gives a U-Pb zircon age of 1677 ± 9 Ma (Fanning et al., 2007). A second sample at the western end of the informally named Wynbring pluton gives a poorly constrained U-Pb zircon age of 1702 ± 42 Ma, and a third gives an age of 1681 ± 15 Ma (Fanning et al., 2007). Granite of the Tunkillia Suite, dated by Teasdale (1997) gives a U-Pb zircon age of 1672 ± 17 Ma.

- Fanning (2002) has reported U-Pb monazite ages from the Challenger Mine at ~1735 Ma, ~1722 Ma and ~1710 Ma. These results suggest that there was an episode of relatively high-grade reworking of the Christie Gneiss, coincident with the ca. 1740-1690 Ma Kimban Orogeny (Vassallo and Wilson, 2002; Swain et al., 2005a; Hand et al., 2007). McFarlane (2006) reported monazite growth ages between 2470 Ma and 2410 Ma from the Christie Gneiss at the Challenger mine. Tomkins et al. (2004) also report $^{40}Ar/^{39}Ar$ cooling ages ranging between ~2000-1800 Ma and sericite ages of ~1620 Ma, from the Challenger Mine. Elsewhere in the Christie Domain, the cooling history of the Christie Gneiss is somewhat constrained by a $^{40}Ar/^{39}Ar$ biotite age of ~1656 ± 12 Ma from Mount Christie, west of the Coorabie Shear Zone (Fraser and Lyons, 2006).

Fowler Domain

- The age and provenance of the stratigraphy in the Fowler Domain is poorly constrained, with only a few zircon U-Pb ages available from metasedimentary rocks from the domain. One metasedimentary sample yielded zircon cores with ages of ~1750 Ma and ~1685 Ma, and metamorphic rims with ages of ~1566 Ma (Fanning et al., 2007). More recently, LA-ICPMS U-

Pb zircon dating of metapelites by Howard et al. (2008) gives a maximum depositional age of 1770 ± 12 Ma for a metapelite in the Colona Block and maximum depositional ages of 1753 ± 19 Ma, 1738 ± 15 Ma and 1730 ± 8 Ma from metapelite samples in the Barton Block (Howard et al., 2008). In the bulk of metasedimentary samples, $\varepsilon Nd_{(1700\ Ma)}$ range from -4.3 to -3.8, suggesting comparatively evolved source regions (Howard et al., 2010c).

- Paleoproterozoic felsic magmatism within the Fowler Domain consists of intrusives with ages of 1726 ± 11 Ma (Kimban age) through to ~1675 Ma (Tunkillia age) (Fanning et al., 2007). LA-ICPMS U-Pb zircon magmatic crystallisation ages of 1707 ± 8 Ma and 1706 ± 8 Ma for orthogneiss from the Barton Block are interpreted to be syntectonic with the Kimban Orogeny (Howard et al., 2008).

- Mafic magmatism in the Fowler Domain is evidenced by metagabbro intersected by the Colona drillholes (Daly et al., 1994), a sample of which was emplaced at 1727 ± 8 Ma (Fanning et al., 2007). A metagabbro from the Barton Block gives a LA-ICPMS U-Pb zircon crystallisation age of 1690 ± 9 Ma (Howard et al., 2008), suggesting that it forms part of the 1690-1670 Ma Tunkillia Suite. The Tunkillia Suite in the Fowler Domain consists of variably deformed, medium-grained, equigranular leucocratic granite and adamellite. Intrusion of Tunkillia Suite equivalents near Pidinga Tank at Lake Ifould provides a magmatic crystallisation age of 1681 ± 15 Ma, and granite at White Gin Rockhole gives an age of 1670 ± 9 Ma (Fanning et al., 2007).

- Geochronological constraints on the metamorphism within the Fowler Domain indicate that there have been multiple events. The oldest age is a monazite SHRIMP age 1712 ± 10 Ma (Barton Block; Teasdale, 1997) which indicates Kimban-aged metamorphism. Metapelite samples analysed by Howard et al. (2008) produced metamorphic monazites, with LA-ICPMS U-Pb ages of 1690 ± 9 Ma (Colona Block), 1685 ± 25 Ma, 1681 ± 5 Ma and 1665 ± 7 Ma (Barton Block), possibly reflecting emplacement of, or thermal reworking associated with, Tunkillia Suite magmatism. EMPA dating by Thomas et al. (2008) of monazite from two samples of amphibolite grade metamorphics in the westernmost Fowler Domain produced ages of 1648 ± 15 Ma and 1638 ± 15 Ma, coincident with the timing of the Ooldean Orogeny, although the accuracy of these EMPA ages is questionable (Hand, Dutch, pers comm.).

Wilgena Domain

- The Wilgena Hill Jaspilite, a distinctive unit of finely-banded, red-black jaspilitic banded iron formation (Daly et al., 1998), overlies ortho- and paragneisses of the Mulgathing Complex in this domain. Age constraints are poor, permitting this unit to have been deposited any time between deformation and metamorphism of the underlying Mulgathing Complex (~2450 Ma) and deposition of the overlying Tarcoola Formation (~1650 Ma). If the folding in the Wilgena Hill Jaspilite is, indeed, of Kimban age, then deposition must have predated the Kimban Orogeny (i.e., >1730 Ma) (Cowley and Martin, 1991a, b).

- The Eba Formation, which contains pebbles of Wilgena Hill Jaspilite, forms a succession of thick-bedded, white quartzite, siltstone, shale, sandstone, deposited in a high to moderate energy fluvial or shallow marine environment, and interbedded amygdaloidal basalt (Daly et al., 1998). U-Pb LA-ICPMS zircon geochronology yield exclusively Archean ages (~2530-3300 Ma, consistent with evolved Nd and zircon Hf isotopic data (Howard et al., 2010a).

- The Labyrinth Formation unconformably overlies the Eba Formation, and consists of basal finely- laminated chert (after carbonate) with well preserved stromatolites, overlain by pebbly sandstone to boulder conglomerate (Cowley and Martin, 1991a, b; Daly et al., 1998). The succession is interpreted to have been deposited at about the same time as the Kimban Orogeny in shallow marine/lacustrine and fluvial environments, with sediments derived from local sources (Daly et al., 1998), as a thin rhyolite within the Labyrinth Formation has a SHRIMP U-Pb zircon age of 1715 ± 9 Ma (Fanning et al., 2007). Detrital zircon ages in the Labyrinth Formation range from Neoarchean to Paleoproterozoic, with isotopic zircon Hf data and Nd data suggesting a source with a mixed crustal evolution (εNd -4.5 to -6; Howard et al., 2010a).

- Granites of the Tunkillia Suite intrude the Wilgena Domain (e.g., Symons Granite, 1686 ± 10 Ma; Teasdale, 1997; Fanning et al., 2007). Three SHRIMP U-Pb zircon dates from different phases of the Paxton Granite at Tarcoola gave ages of 1722 ± 4 Ma to 1715 ± 3 Ma (Budd, 2006).

- Intrusion of granites of the Tunkillia Suite was followed by deposition of the Tarcoola Formation, an ~800 m thick, fining-upwards succession, ranging from the basal Peela Conglomerate through quartzites and sandstone (Fabian Quartzite Member) upwards into the upper Sullivan Shale Member, interpreted to have been deposited in fluvial to marginal-marine conditions (Daly et al., 1993). Water-lain tuffs of the Sullivan Shale Member give ages of 1657 ± 11 Ma and 1654 ± 3 Ma (Fanning, 1997). The basal Peela Conglomerate contains detrital zircons with age peaks of 1733 ± 5 Ma and 1715 ± 8 Ma (Holm, 2004), consistent with derivation from the nearby Paxton Granite. Zircon Hf isotopic and whole rock Nd data indicate the Tarcoola Formation was sourced from predominantly juvenile rocks ($\varepsilon Nd_{(1655\ Ma)}$ +1 to -2.5; Howard et al., 2010a).

- The age of the Muckannippie Anorthosite suite, consisting of diorite, gabbro, anorthosite and troctolite (Budd, 2006), is unknown, but is speculated to be either part of the 1690-1660 Ma Tunkillia Suite (Teasdale, 1997; Direen et al., 2005), the ~1630 Ma St Peter Suite, or the Gawler Range Volcanics (Stewart and Foden, 2003).

Harris Greenstone Domain

- The fluvial to marginal-marine clastic succession of the Tarcoola Formation (see Wilgena Domain) is also present in the Harris Greenstone Domain, with interbedded andesitic to dacitic tuffs and basaltic sills, deposited in this interval.

- Evidence for the Kimban Orogeny comes from a monazite age of 1721 ± 17 Ma from the Christie Gneiss in the Harris Greenstone Domain (Holm and Lyons, in Zang, 2002).

Nuyts Domain

- The earliest recorded event in the Nuyts Domain is the intrusion of the Investigator Granite Gneiss (U-Pb zircon age of 1762 ± 11 Ma; Cooper and Belousova, 2004).

- A sample of quartzofeldspathic granite gneiss from Poverty Corner gave a U-Pb SHRIMP zircon crystallisation age of 1717 ± 4 Ma (Fraser and Neumann, 2010). Banded biotite gneiss from Little Pinbong Rockhole gave a weighted mean $^{207}Pb/^{206}Pb$ crystallisation age of 1714 ± 4 Ma (Fraser and Neumann, 2010). A leucocratic segregation from within the banded gneiss from Little Pinbong Rockhole gave a zircon crystallisation age of 1712 ± 3 Ma, within uncertainty of the age for the host sample (Fraser and Neumann, 2010).

- A cluster of three U-Pb zircon ages at ~1680 Ma (1686 ± 5 Ma, 1680 ± 9 Ma, 1669 ± 13 Ma; Fanning et al., 2007) was determined on granites of the Tunkillia Suite in this domain. The Tunkillia Suite represents a suite of coeval late Paleoproterozoic, I-type intrusive plutons and rhyolite and mafic dykes, which crop out within the Yarlbrinda Shear Zone (YSZ) of the Nuyts Domain (Ferris and Schwarz, 2004). These intrusives are dominantly felsic, with compositions including diorite, granodiorite, syenite and granite (Ferris and Schwarz, 2004). The Suite includes other granites from outside the YSZ, including Barton South, Lake Ifould, Mulgathing Rocks (Symons Granite), Lake Tallacootra and Wynbring Rocks (Ferris, 2001).

- The Tunkillia Suite and age equivalents represent a major addition of crustal material to the western margin of the Gawler Craton at ~1690-1670 Ma (Ferris and Schwarz, 2004). Teasdale (1997) considered that the Tunkillia Suite was associated with an active plate margin. Ferris et al. (2002) supported this interpretation, and suggested that the voluminous, mafic to intermediate calcalkaline magmatic rocks of the Tunkillia Suite had a geochemical signature consistent with a magmatic arc setting. Alternatively, Ferris and Schwarz (2004) considered that the Tunkillia Suite was emplaced in a backarc environment behind an active magmatic arc, whereas Payne et al. (2006b) have suggested a possible post-orogenic (?post-collisional) setting for the Suite. According to Payne (2008), and Payne et al. (2010), geochemical evidence suggests that a suprasubduction setting for the suite cannot be substantiated and, thus, a late to post-tectonic setting is favoured.

- I-type calcalkaline tonalite to granodiorite magmatic rocks of the ca. 1631-1608 Ma St Peter Suite are inferred to underlie the majority of the Nuyts Domain (Fairclough et al., 2003). Ages of 1620 ± 10 Ma, 1619 ± 15 Ma, 1616 ± 6 Ma and 1615 ± 8 Ma are assigned to granites of the St Peter Suite (Fanning et al., 2007) in this domain. Three slightly younger ages of 1613 ± 4 Ma, 1611 ± 7 Ma and 1608 ± 6 Ma have been reported by Ferris (2001).

- The felsic Nuyts Volcanics (and associated granitoids, e.g., St Francis Granite) on islands in the Nuyts Archipelago define a northeast-trending zone of felsic magmatism. The volcanic rocks consist predominantly of dark grey to pink porphyritic rhyodacite to rhyolite (Daly et al., 1998). TIMS zircon ages for the Nuyts Volcanics of 1631 ± 3 Ma (Cooper et al., 1985) and 1627 ± 2 Ma (Flint and Rankin, 1991) are within uncertainty of ages for the St Peter Suite, and they are interpreted to be part of the same event.

- Granites of the St Peter Suite have compositions similar in chemistry to Archean tonalite-trondhjemite-granodiorite suites (Ferris and Schwarz, 2004). Geochemically, the St Peter Suite displays negative Nb and Ti anomalies, LREE and LILE enrichment and relatively juvenile $\varepsilon Nd_{(1620\ Ma)}$ isotope values (-0.8 to +3.7; Swain et al., 2008). Ferris and Schwarz (2004) interpreted the suite to represent emplacement within a backarc environment behind an active magmatic arc setting. Ferris et al. (2002) considered that both the Tunkillia and St Peter Suites were emplaced at the southwest margin of the Archean Gawler nuclei during subduction that dipped to the northeast. An alternative proposal by Swain et al. (2008) is that the St Peter Suite arc magmatism developed outboard of the current Gawler Craton due to a subduction zone that dipped to the south under lithosphere that is now preserved in Antarctica. They proposed that subduction ceased at about 1608 Ma as a result of collision of the volcanic arc of the St Peter Suite with the Archean Gawler nuclei.

Coulta Domain

- The Hutchison Group overlies basement gneisses and granites of the Sleaford Complex in the Coulta (and Cleve) Domains. Metasedimentary rocks of the Hutchison Group are composed of a mixed succession of chemical and clastic sedimentary and extrusive mafic and felsic volcanic rocks including the Warrow Quartzite, Katunga Dolomite, Lower Middleback Jaspilite, Burrawing Amphibolite, Cook Gap Schist, Upper Middleback Jaspilite, Yadnarie Schist and Bosanquet Formation (Rankin et al., 1988). The basal Warrow Quartzite contains detrital zircons as young as ~2000 Ma, providing a maximum depositional age for the Group (Fanning et al., 2007). Fraser and Neumann (2010) report a U-Pb zircon maximum depositional age of 2507 ± 5 Ma for the Warrow Quartzite at Thurlga Station. The Cook Gap Schist, part of the Middleback Subgroup thought to overlie the Warrow Quartzite, gave a U-Pb detrital zircon maximum depositional age of >2650 Ma (Szpunar et al., 2007a). Volcanic rocks from the Bosanquet Formation give a zircon crystallisation age of 1866 ± 10 Ma (Fanning et al., 2007). Recently, the upper part of the Group has been shown to contain ca. 1850 Ma and 1780 Ma detrital zircons, thus indicating the presence of several distinct sedimentary packages within the Hutchison Group (see Cleve Domain; Szpunar et al., 2007a; Barovich et al., 2008). Indeed, Szpunar et al. (in press) report maximum depositional ages of ~1780 Ma for the Cook Gap Schist in the Cleve and Tumby Bay areas, constraining the upper Hutchison Group to ~1780-1730 Ma. As a result of these new age constraints, the Hutchison Group is now divided into three groups: 1. the oldest (which contains banded iron formations of the Middleback Ranges), has a maximum depositional age of ~2570 Ma, is isotopically evolved and is geochemically similar to typical Archean rocks, 2. a younger succession with a maximum depositional age of ~1865 Ma, derived from isotopically evolved Gawler Craton sources and, 3. the youngest succession, deposited within the age range 1790-1730 Ma, and derived from a mix of isotopically evolved Gawler Craton sources and isotopically juvenile (non-Gawler?) sources (Szpunar and Fraser, 2010).

- The Price Metasediments have U-Pb SHRIMP zircon ages which constrain maximum depositional ages to 1766 ± 14 Ma and 1767 ± 16 Ma (Oliver and Fanning, 1997). These sedimentary rocks may have been derived from the Investigator Granite Gneiss of the Nuyts Domain, or magmatic rocks of the same age. The low grade phyllitic schists of the Price Metasediments may be a correlative of the Wallaroo Group of Yorke Peninsula (Cowley et al., 2003).

- Fraser and Neumann (2010) record a U-Pb SHRIMP zircon crystallisation age of 1724 ± 4 Ma for granite (at Peter Pan Platforms), with Sleaford-aged inheritance (~2425 Ma and ~2500 Ma). A sample of grey granite gneiss from the same location gave a U-Pb SHRIMP zircon crystallisation age of 1724 ± 3 Ma, with inheritance at ~2700 Ma, ~2550 Ma and 2500 Ma (Fraser and Neumann, 2010). Megacrystic gneiss from Peter Pan Platforms produced several ages: a 2470 ± 8 Ma zircon age is interpreted as the igneous protolith age of the gneiss. A distinct generation of high-grade metamorphic zircon is present at ~2430 Ma, consistent with Sleaford-aged metamorphism. Zircon of age 1723 ± 13 Ma is coeval with the age of adjacent granite and gneiss and is interpreted to result from melt injection of adjacent ~1725 Ma granite, or possibly from partial melting (Fraser and Neumann, 2010).

Cleve Domain

- The migmatitic, quartzofeldspathic Miltalie Gneiss, a group of granitic orthogneisses, was emplaced between 2002 ± 15 Ma and 1999 ± 9 Ma (Fanning et al., 2007) and, together with the Sleafordian Complex, forms basement to the Hutchison Group. A Miltalie Gneiss equivalent (known informally as the Glengarra Granite, located on Glengarra Station on central-eastern Eyre Peninsula west of Tumby Bay) gives a U-Pb age of 2001 ± 8 Ma (Fanning et al., 2007). Little is known about the tectonic setting of these gneisses (Reid et al., 2008b); Daly et al. (1998) suggested that the Miltalie Gneiss was intruded during a period of extensional tectonism. Fanning et al. (1995), Fanning (1997) and Daly et al. (1998) have proposed that, during the emplacement of the Miltalie Gneiss, the lower-crustal high-grade Carnot Gneiss was brought into juxtaposition with higher-crustal level, lower-grade intrusions of the Dutton Suite. Megacrystic granite gneiss at Bascombe Rocks gives a U-Pb zircon mean age of 2005 ± 3 Ma, interpreted as the igneous age of the protolith to the gneiss (Fraser and Neumann, 2010); an age equivalent to the Miltalie Gneiss.

- The Hutchison Group, a succession of highly-deformed, clastic and chemical metasediments, was deposited on a shallow continental shelf. The group consists of a basal quartzite (Warrow Quartzite), and overlying semipelitic metasediments, carbonates and banded iron formations of the Middleback Subgroup (Parker and Lemon, 1982; Parker et al., 1993). The succession is interpreted to have been deposited onto a composite Archean-Proterozoic continent (Sleaford Complex and Miltalie Gneiss) within a passive margin, continental shelf environment, which deepened to the east (Daly et al., 1998). Concordant amphibolites within the Hutchison Group are interpreted most likely to have been mafic volcanic rocks with a quartz tholeiite chemistry (Parker et al., 1993), formed during the extensional event. The Warrow Quartzite (Parker and Lemon, 1982), has a detrital zircon population with peaks at 2720 Ma, 2520 Ma, 2440 Ma and 2000 Ma (Fanning et al., 2007), consistent with derivation from local Sleafordian and Miltalie sources (Schwarz et al., 2002). Fraser and Neumann (2010) report U-Pb SHRIMP zircon maximum depositional ages for the Warrow Quartzite of 2019 ± 3 Ma and 2003 ± 4 Ma (from the western shore of Lake Gilles and Moseley Nobs, respectively). The Warrow Quartzite also has evolved $\varepsilon Nd_{(1850\ Ma)}$ values (-10), consistent with derivation largely from the Late Archean to earliest Paleoproterozoic Gawler Craton (Szpunar et al., 2007a). A tighter constraint on the age of deposition comes from a U-Pb zircon age of 1866 ± 10 Ma from the Bosanquet Formation, interpreted as a volcanic unit within the upper Hutchison Group (Fanning et al., 2007). There is considerable uncertainty, however, regarding the geological relationship between the Bosanquet Formation and the Hutchison Group, as the original nature of the contact is obscured by deformation. Consequently, alternative interpretations have been proposed in which the Hutchison Group is suggested to be younger than the Donington Suite (see Vassallo and Wilson, 2001). Nevertheless, the presence of metasedimentary enclaves within the Donington Suite, with detrital zircon populations at ~2500 Ma and ~2000 Ma, confirms that sedimentation did occur in the region prior to emplacement of the Donington Suite (Howard et al., 2006; Fanning et al., 2007), although there is no evidence to say that such sediments are Hutchison Group. Recently, samples from the overlying Yadnarie Schist have been shown to contain ~1850 Ma and ~1780 Ma detrital zircons (Szpunar et al., 2007a; Barovich et al., 2008; Szpunar et al., in prep). These units have significantly more juvenile initial $\varepsilon Nd_{(1850\ Ma)}$ values (-2) and geochemical affinities with a more mafic source than that for the Warrow Quartzite (Szpunar et al., 2007a). The Hutchison Group is currently divided into three groups (see Coulta Domain above).

- No examples of Donington Suite granites (~1850 Ma) have been identified in the Cleve Domain (see Spencer Domain).

- Magmatism associated with the Kimban Orogeny is represented by the Middlecamp, Moody and Tunkillia Suites. The Middlecamp Granite is a pre- to early-Kimban Orogeny granite suite that has been U-Pb dated at 1737 ± 7 Ma and 1726 ± 7 Ma (Fanning et al., 2007), and 1736 ± 6 Ma (with Sleaford-aged inheritance; Fraser and Neumann, 2010). It is a grey, medium-grained, foliated granite of monzogranite to granodiorite composition (Parker et al., 1993), and is intruded by a number of pegmatite veins. A possible Middlecamp equivalent at Refuge Rocks gives a U-Pb zircon age of 1740 ± 3 Ma (Fanning et al., 2007). These granites are foliated, interpreted as a result of the Kimban Orogeny, and their ages provide a maximum age constraint for the timing of at least some Kimban deformation. Grey granite gneiss from Refuge Rocks gives a magmatic age of 1738 ± 3 Ma with ~2450 Ma inheritance (Fraser and Neumann, 2010). A garnet leucogranite, west of Lake Gilles, gave a SHRIMP zircon crystallisation age of 1738 ± 4 Ma (Fraser and Neumann, 2010).

- The Moody Suite in the Spencer Domain on Eyre Peninsula was defined by Schwarz (1999) as a series of weakly-deformed to massive plutons, which were emplaced during the waning stages of the Kimban Orogeny at ~1700 Ma. A variety of lithologies occur within the Moody Suite, including mostly medium- to coarse-grained, pink monzogranite, monzonite and leucogranite (Parker et al., 1993). Intrusion of hornblende-bearing granites to muscovite-bearing leucogranites of the Moody Suite (Schwarz, 1999) occurred later in the Kimban Orogeny, with SHRIMP U-Pb ages in the range 1720 ± 9 Ma to 1701 ± 12 Ma (Fanning et al., 2007). A mean U-Pb zircon age of 1722 ± 4 Ma is interpreted as the crystallisation age of granite at Bascombe Rocks (Fraser and Neumann, 2010). Granite at Moody Tank has an Rb-Sr age of 1709 ± 14 Ma (Mortimer et al., 1988). Several granites of the Moody Suite dated by Fanning (1997), have ages between 1708 ± 7 Ma and 1692 ± 10 Ma. The Moreenia Adamellite and Chinmina Monzonite yielded U-Pb zircon crystallisation ages of 1720 ± 9 Ma and 1701 ± 12 Ma, respectively (Fanning et al., 2007). The Carappee Granite was dated at 1689 ± 59 Ma by Flint et al. (1988), recently redated at 1720 ± 3 Ma (U-Pb SHRIMP zircon; Fraser and Neumann, 2010).

Spencer Domain

- The Corny Point Paragneiss and metasedimentary rocks possibly equivalent to the Hutchison Group were deposited in the Spencer Domain in the Paleoproterozoic. The Corny Point Paragneiss is a layered migmatite, with a protolith depositional age of which is constrained by ~2400 Ma and ~1920 Ma detrital zircons probably sourced from the western to northwestern Gawler Craton, and deposited on a continental shelf (Zang and Fanning, 2001). Reid and Howard (2006) and Howard et al. (2006) reported younger detrital zircon ages with peaks at 1946 ± 18 Ma, 1889 ± 11 Ma, 1881 ± 18 Ma, and 1852 ± 18 Ma. Howard et al. (2009) constrain sedimentation of the Corny Point Paragneiss to the interval ca. 1870-1850 Ma, with detrital zircon age peaks at 2510 Ma, 2450 Ma and 2000 Ma. The rocks were subjected to high grade metamorphism and deformation during the Cornian Orogeny (1850-1840 Ma; Zang, 2006). Detrital zircon geochronology, in conjunction with Nd and Hf isotopic data, have been used to suggest that the Gawler Craton was not a viable source region for the Corny Point Paragneiss (Howard et al., 2009). Instead, the Pine Creek Orogen has been suggested as a plausible source region (Howard et al., 2009).

- Donington Suite granites dominate the Spencer Domain. The suite is an intrusive magmatic complex that occupies a non-continuous north-south trending belt around 600 km long and up to 80 km wide along the eastern part of the Gawler Craton (Schwarz, 2003). Compositions range from mafic gabbro-norite to evolved leucogranites, and were derived from the fractionation of a mafic magma, which incorporated a significant component of pre-existing crustal material (Schaefer, 1998; Reid et al., 2008b). Intrusion of the Donington Suite was coincident with the ~1850 Ma Cornian Orogeny, which is characterised by the emplacement of the Donington Suite into a contractional tectonic environment within which contractional strain and granulite-facies peak metamorphism were synchronous (Reid et al., 2008b). Reid et al. (2008b) demonstrated that the Cornian Orogeny produced a clockwise P-T path with peak metamorphic conditions of ~750˚C and 6 kbar. Fanning et al. (2007) reported SHRIMP isotopic ages for granites of the Donington Suite at 1850 ± 5 Ma (Memory Cove Charnockite) and 1853 ± 5 Ma (Colbert Granite). Essentially identical ages of 1850 ± 2 Ma to 1846 ± 3 Ma (six ages) have been reported by Jagodzinski et al. (2006). SHRIMP geochronology indicates that the Gleesons Landing Granite was emplaced at 1850 ± 5 Ma, 1850 ± 3 Ma and 1850 ± 2 Ma (Jagodzinski et al., 2006), while two samples of the Royston Granite yielded U-Pb zircon ages of 1850 ± 12 Ma and 1849 ± 11 Ma (Jagodzinski et al., 2006; Zang, 2006).

- Geochemically, granitoids of the Donington Suite show LREE enrichment, negative Nb, Sr, P and Ti anomalies and have εNd (at 1850 Ma) values between -2 and -4 (Schaefer, 1998). The Donington Suite is thought to be derived from a mixture of moderately juvenile mafic parent and pre-existing crustal material (Mortimer et al., 1988; Schaefer, 1998). The coincidence of deformation during the Cornian Orogeny and emplacement of the Donington Suite suggests a convergent tectonic setting at the time of emplacement (Ferris et al., 2002). Potentially, the Donington Suite either directly marks the existence of a magmatic arc, which developed along the margin of the Archean-Paleoproterozoic continent, or may have been produced in a backarc environment and then accreted into the eastern margin of the Gawler Craton during the Kimban

Orogeny (Ferris et al., 2002). Schwarz et al. (2002) suggested that the Donington Suite was either magmatic arc or syncollisional in character, related to a subduction zone located to the west of the present day position of the suite. Alternatively, Hand (2005) proposed that the Donington Suite formed during collision along the eastern margin of the Gawler Craton. Reid et al. (2008b) have proposed that the Donington Suite was generated in a (far-field?) continental backarc setting. In this scenario, lithospheric extension and consequent decompressional melting of the mantle in the continental backarc may have resulted in mantle-derived melts undergoing both crustal assimilation and fractional crystallisation as they ascended into the crust, thus forming the bulk of the Donington Suite. Reid et al. (2008b) suggest that the deformation and metamorphism of the Cornian Orogeny may have been focussed in this backarc, due to thermal softening of the thinned lithosphere, possibly the result of a change in subduction dynamics (e.g., the arrival of a buoyant collider at a far-field subduction zone).

- Dykes of the Tournefort Metadolerite (1812 ± 5 Ma; Schaefer, 1998, in Vassallo and Wilson, 2002; and Jussieu Metadolerite, Hoek and Schafer, 1998; Zang, 2002), are continental tholeiites which crosscut the Donington Suite. The dykes have a predominant north-south to north-northeast-south-southwest trend, indicating emplacement during an east-west extensional event (Hoek and Schaefer, 1998; Vassallo and Wilson, 2002). Reid et al. (2008b) suggest that the emplacement of the Tournefort Metadolerite at ca. 1815 Ma (Schaefer, 1998), and the subsequent development of localised basins (see below), indicates extension is likely to have continued periodically across the region for up to 80 million years (~1810-1730 Ma) following relatively short-lived convergence during the Cornian Orogeny (~1850 Ma).

- Subsequent to intrusion of the Tournefort dykes, a number of ca. 1790-1740 Ma volcano-sedimentary successions were deposited across the southeastern Gawler Craton (northeastern Eyre Peninsula). This extension-dominated system, may have developed as a series of slab rollback-initiated, extensional basins which formed inboard of a far-field subduction front, that is, possibly in a continental backarc (Reid et al., 2008b). The Myola Volcanics consist of interbanded porphyritic rhyolite and rhyodacite, fine-grained felsic and hornblende-bearing gneisses, and amphibolite (Parker et al., 1993; Curtis, 2007). A rhyolite within the succession gives a U-Pb zircon crystallisation age of 1791 ± 4 Ma (Fanning et al., 1988), with another sample reported by Fraser and Neumann (2010) giving an age of 1792 ± 5 Ma. Geochemically, trace element compositions are typical of within-plate volcanism, and the volcanics are geochemically similar to the Tidnamurkuna Volcanics in the Peake and Denison Domain (Parker et al., 1993; initial $\varepsilon Nd_{(1790\ Ma)}$ -4.4; Szpunar and Fraser, 2010). The Wertigo Granite intrudes the Myola Volcanics and two samples have produced U-Pb SHRIMP zircon crystallisation ages of 1792 ± 6 Ma and 1768 ± 4 Ma (the latter with significant inherited populations at 1860 Ma, 2500 Ma, 2800 Ma and 3400 Ma; Fraser and Neumann, 2010). The Myola Volcanics were presumably overlain (relationship not exposed) by sediments which were metamorphosed to fine-grained slaty to phyllitic quartz-muscovite schist and fine-grained quartzite and amphibolite of the Broadview Schist, the latter sourced predominantly from granitoids of the Donington Suite (Daly et al., 1998; Jagodzinski, 2005), and with a maximum depositional age of 1795 ± 5 Ma (Fraser and Neumann, 2010; initial $\varepsilon Nd_{(1790\ Ma)}$ -3.5 and -4.8; Szpunar and Fraser, 2010). This was followed by extrusion of rhyolite to dacite, with minor andesite-basalt, of the McGregor Volcanics (~1740 Ma, Fanning et al., 1988; Parker et al., 1993; 1754 ± 5 Ma, Fraser and Neumann, 2010; initial $\varepsilon Nd_{(1750\ Ma)}$ -2.1 and -3.3; Szpunar and Fraser, 2010). Bimodal felsic (porphyritic rhyolite, rhyodacite and dacite) and lesser basaltic volcanic rocks of the McGregor Volcanics have typical within-plate A-type characteristics, but also contain components of possible mantle- or subduction-derivation (Parker et al., 1993). U-Pb SHRIMP dating of zircons in a volcaniclastic sedimentary rock of the Moonabie Formation, a massive volcaniclastic grit (Parker et al., 1993) interlayered with the McGregor Volcanics, gave a syn-depositional age of 1756 ± 8 Ma (Jagodzinski, 2005). A sample of volcaniclastic gritstone from the Moonabie Formation gave a U-Pb SHRIMP zircon maximum depositional age of 1750 ± 5 Ma (Fraser and Neumann, 2010). A second sample of Moonabie Formation, a quartz sandstone, gave a zircon maximum depositional age of 1755 ± 5 Ma (Fraser and Neumann, 2010). Zircons analysed from a felsic volcanic rock within the Moonabie Formation gave a U-Pb SHRIMP crystallisation age of 1744 ± 6 Ma, and provides a minimum age for the host (Fraser and Neumann, 2010). Recent age and geochemical data are used to link these northeastern Eyre Peninsula successions with the Wallaroo Group (i.e., Wandearah Metasiltstone; Szpunar and Fraser, 2010) in the Olympic Domain (see below).

- Fraser and Neumann (2010) report U-Pb SHRIMP zircon crystallisation upper intercept ages of 1742 ± 42 Ma and 1755 ± 19 Ma for samples of Burkitt Granite (placed into the Moody Suite by Parker et al., 1993). The imprecise ages are attributed to the highly radiogenic nature of this granite (that is, it yields a heat production value of ~17 μWm^{-3}, and a whole-rock uranium abundance of 37 ppm; Neumann, 2001), and Pb loss from metamict zircon. Geochemical data were used originally to argue for formation of the suite in a magmatic arc to backarc environment (Teasdale, 1997; Ferris et al., 2002; Ferris and Schwarz, 2004), although Payne (2008) also used geochemical data to argue for a late to post-tectonic setting.

- Metamorphic monazite ages of ~1690 Ma (from garnet-biotite-plagioclase-quartz-Kfeldspar mylonites from the western central part of the Kalinjala Shear Zone; Swain et al., 2005b) are attributed to the late stages of the Kimban Orogeny, and probable movement along the Kalinjala Mylonite Zone that separates the Spencer Domain from the Cleve Domain.

Olympic Domain

- Granites of the Donington Suite intrude into possible equivalents of the Hutchison Group in the Olympic Domain. Jagodzinski (2005) reported SHRIMP ages for five granites ranging from 1860 ± 4 Ma to 1850 ± 4 Ma, apparently some 10 Ma older than granites of the Donington Suite in the Spencer Domain (Jagodzinski, 2005).

- Metasedimentary rocks of the Wallaroo Group, presumably sourced from granitoids or volcanic equivalents of the Donington Suite (Jagodzinski, 2005), were deposited in fluvial-aeolian to shallow-water and shelfal settings (Zang, 2002; Zang et al., 2002). The Wallaroo Group contains metasedimentary rocks and mafic and felsic metavolcanic rocks, and is divided into three main formations: the Wandearah Formation (metasedimentary rocks), the Weetulta Formation (A-type felsic volcanic rocks) and the Matta Formation (mafic volcanic rocks, including tholeiites) (Zang, 2002). Felsic porphyries of the Moonta Porphyry Member of the Weetulta Formation, intercalated with the Doora Member, the latter consisting of pelitic to psammitic mica schist, hornfels, amphibolite and iron formation (Parker et al., 1993), produced U-Pb zircon ages of 1753 ± 8 Ma and 1748 ± 15 Ma (Fanning et al., 2007). The Mona Volcanics Member of the Weetulta Formation has yielded zircon crystallisation ages of 1735 ± 10 Ma and 1740 ± 16 Ma (Fanning et al., 2007). The Wardang Volcanics Member in the Weetulta Formation has an extrusive age of 1772 ± 14 Ma (Fanning et al., 2007), while felsic porphyries interlayered in the Wandearah Formation, a succession of siliceous and haematitic siltstone, dolomite and calcsilicate (Parker et al., 1993) and Doora Schist, have crystallisation ages of 1763 ± 14 Ma, 1750 ± 7 Ma and 1741 ± 9 Ma (Fanning, 1997). Metasedimentary rocks of the Wallaroo Group contain detrital zircons of ~1850 Ma age, and a younger zircon population at ~1760-1750 Ma (Jagodzinski, 2005). This younger population corresponds broadly to the Wardang Volcanics at ~1770 Ma and volcanic rocks within metasedimentary rocks of the Wandearah Metasiltstone at ~1740 Ma. Szpunar and Fraser (2010) provide initial $\varepsilon Nd_{(1750\ Ma)}$ values of 0 to -3 for volcanic rocks of the Wallaroo Group and -1 to -5 for metasedimentary rocks from the Wallaroo Group. Conor (1995), Zang (2002) and Zang et al. (2002) have suggested that the Wallaroo Group on Yorke Peninsula was deposited on a rifted continental margin during this time period. Alternatively, Reid et al. (2008b) suggested that the episodic extension occurred in a backarc setting associated with slab roll-back, although there is no evidence of a magmatic arc at this stage. Deformation and low- to medium-grade metamorphism of the Wallaroo Group is assumed to have occurred during the Kimban Orogeny (Conor, 1995; Zang et al., 2002), although there is no direct geochronological evidence for this view. The important point here is that the most easterly occurrences of the Wandearah Formation show only extremely low metamorphism and very little deformation (of any age), showing that there is a distinct eastern limit to the Kimban Orogeny (Preiss, pers. comm.)

Peake and Denison Domain

- Deposition of the Peake Metamorphics (Ambrose et al., 1981), a mildly deformed, dominantly quartzitic succession (Baltucoodna Quartzite), with interbedded amygdaloidal basalts and rhyolite (Tidnamurkuna Volcanics), and quartz-chlorite-muscovite schists, occurred in a shallow marine setting (Ambrose et al., 1981; Parker et al., 1993). The Tidnamurkuna Volcanics consist of metamorphosed amygdaloidal basalt and porphyritic rhyolites with minor phyllite and marble (Parker et al., 1993). Extrusion of the lower Tidnamurkana Volcanics occurred at 1781 ± 8 Ma, and the upper Tidnamurkuna Volcanics at 1789 ± 10 Ma (Fanning et al., 2007).

- Zircon geochronology of greenschist to amphibolite quartz-rich sedimentary facies from the Peake, Eastern and Kingston Inliers, gave maximum depositional ages of ~1767 Ma, ~1836 Ma and ~1844 Ma, respectively (P. Betts, pers. comm.).

- Hopper (2001) reports TIMS U-Pb zircon ages of ~1800 Ma from both a pegmatite and mafic sill, as well as an age of ~1733 ± 13 Ma from a metatonalite in the Peake and Denison Domain.

- The Wirriecurrie Granite has tectonised contacts with the surrounding Peake Metamorphics. Augen gneiss of the Wirriecurrie Granite gives a U-Pb zircon age of 1787 ± 8 Ma (Fanning et al., 2007), within uncertainty of the age of the Tidnamurkuna Volcanics. The Wirriecurrie Granite is interpreted to have formed in an intracontinental setting that sampled a c. 2500 Ma subduction-modified mantle source (Hopper, 2001, cited in Payne et al., 2009).

- The felsic volcanics at Spring Hill are distinct from the Tidnamurkuna Volcanics, with an extrusive age of 1740 ± 6 Ma, and may be an equivalent of the McGregor Volcanics in the Spencer Domain (Fanning et al., 2007).

- Peak metamorphic conditions of about 5 kbars and 650˚C occurred at 1718 ± 31 Ma (Kimban Orogeny; Hopper, 2001).

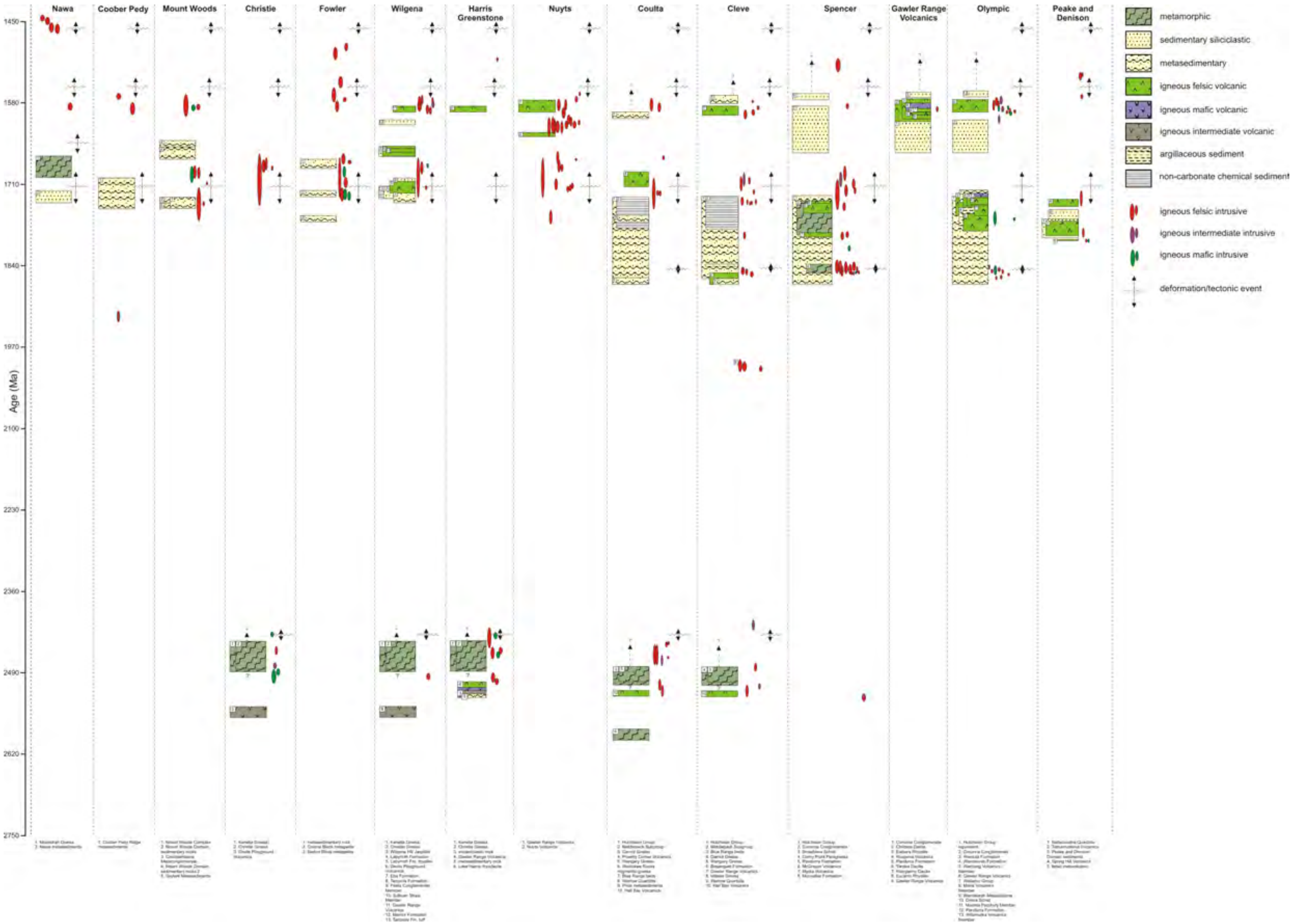

Figure 8a. Time-space-event plot for domains of the Gawler Province for the interval 2750 Ma to 1450 Ma. Note that the Cooyerdoo Granite (~3150 Ma) in the Spencer Domain is not plotted on this figure.

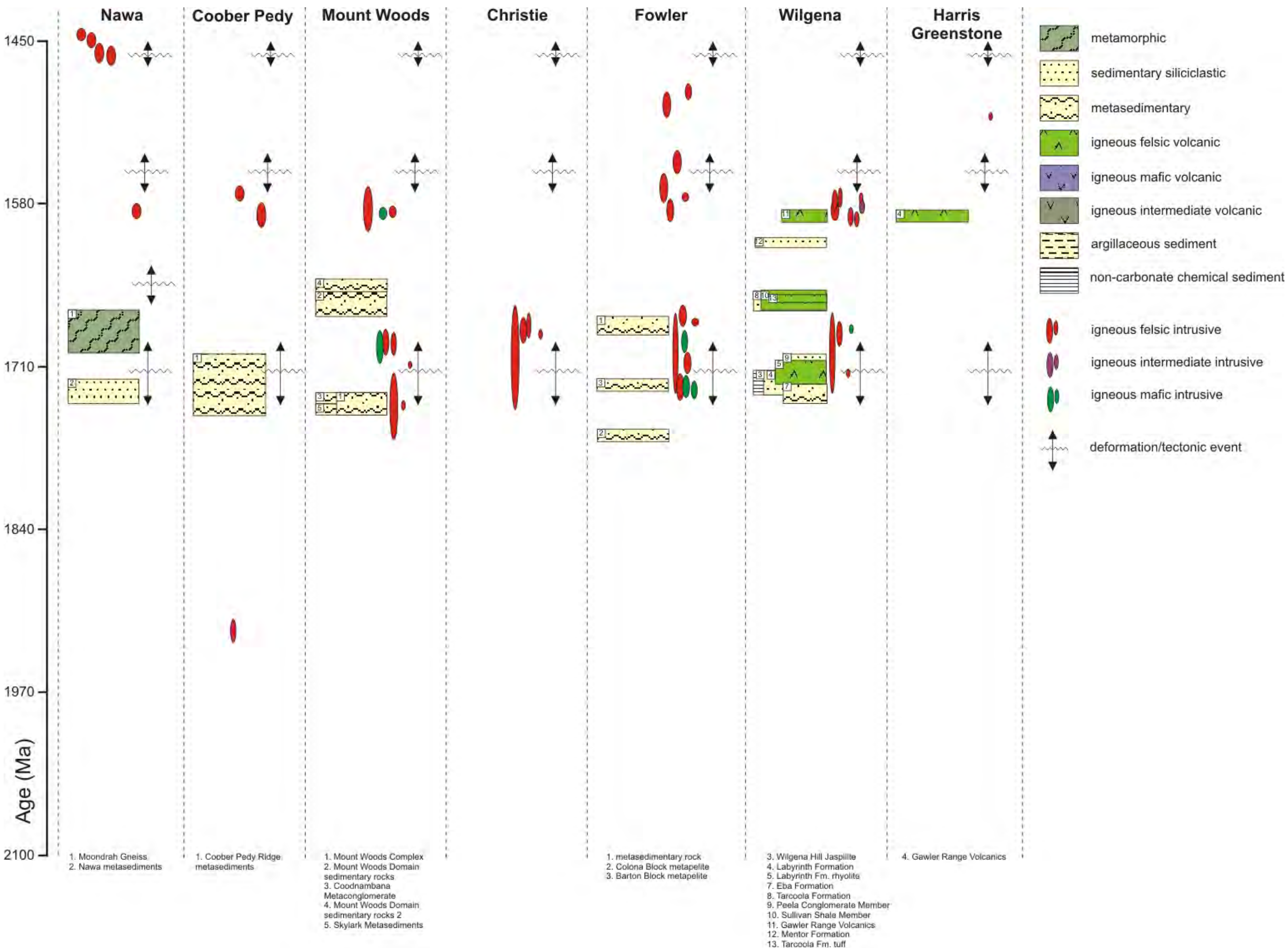

Figure 8b. Time-space-event plot of domains of the Gawler Craton in the interval ~2100-1450 Ma.

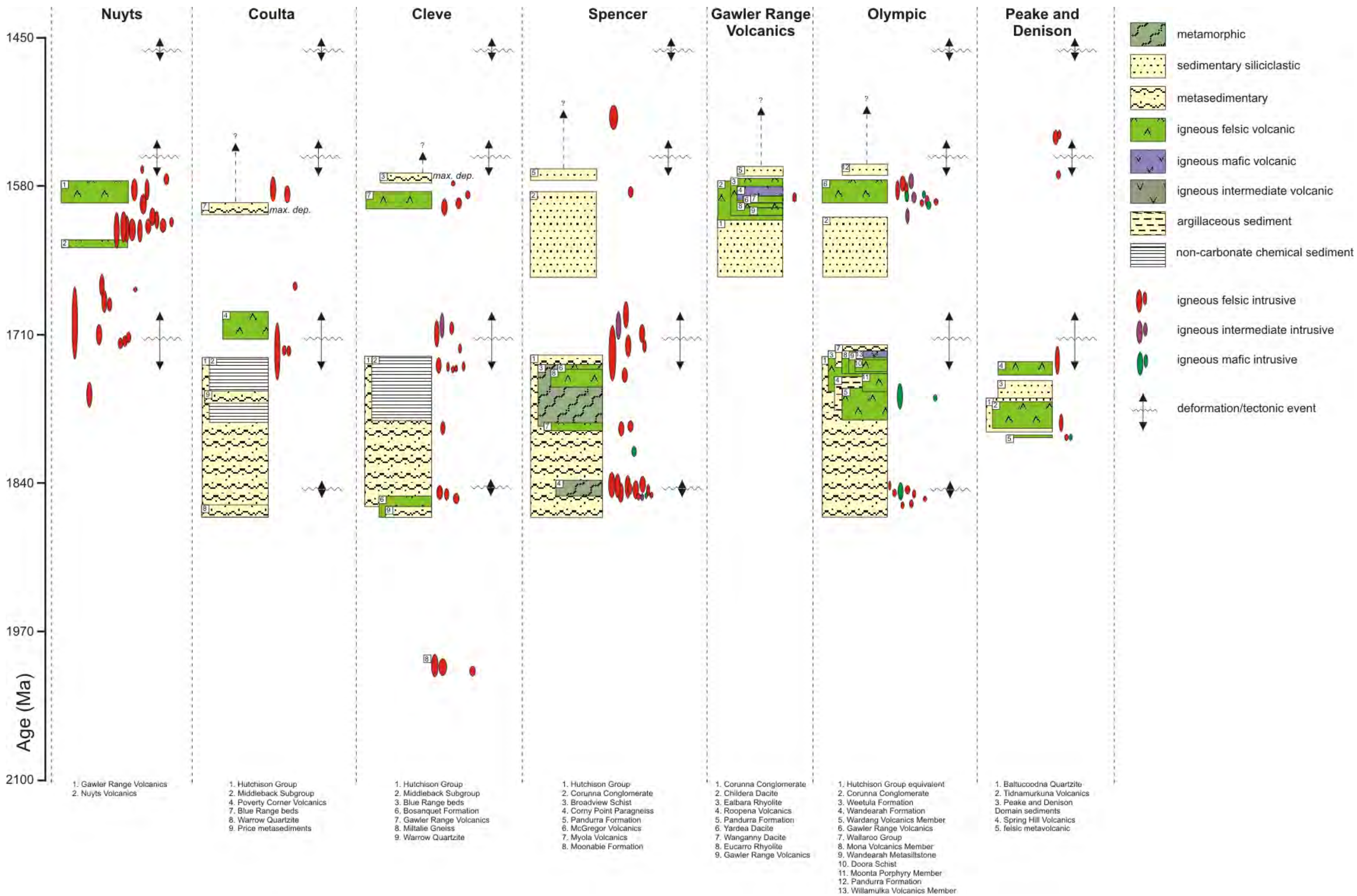

Figure 8c. Time-space-event plot of domains of the Gawler Craton in the interval ~2100-1450 Ma continued.

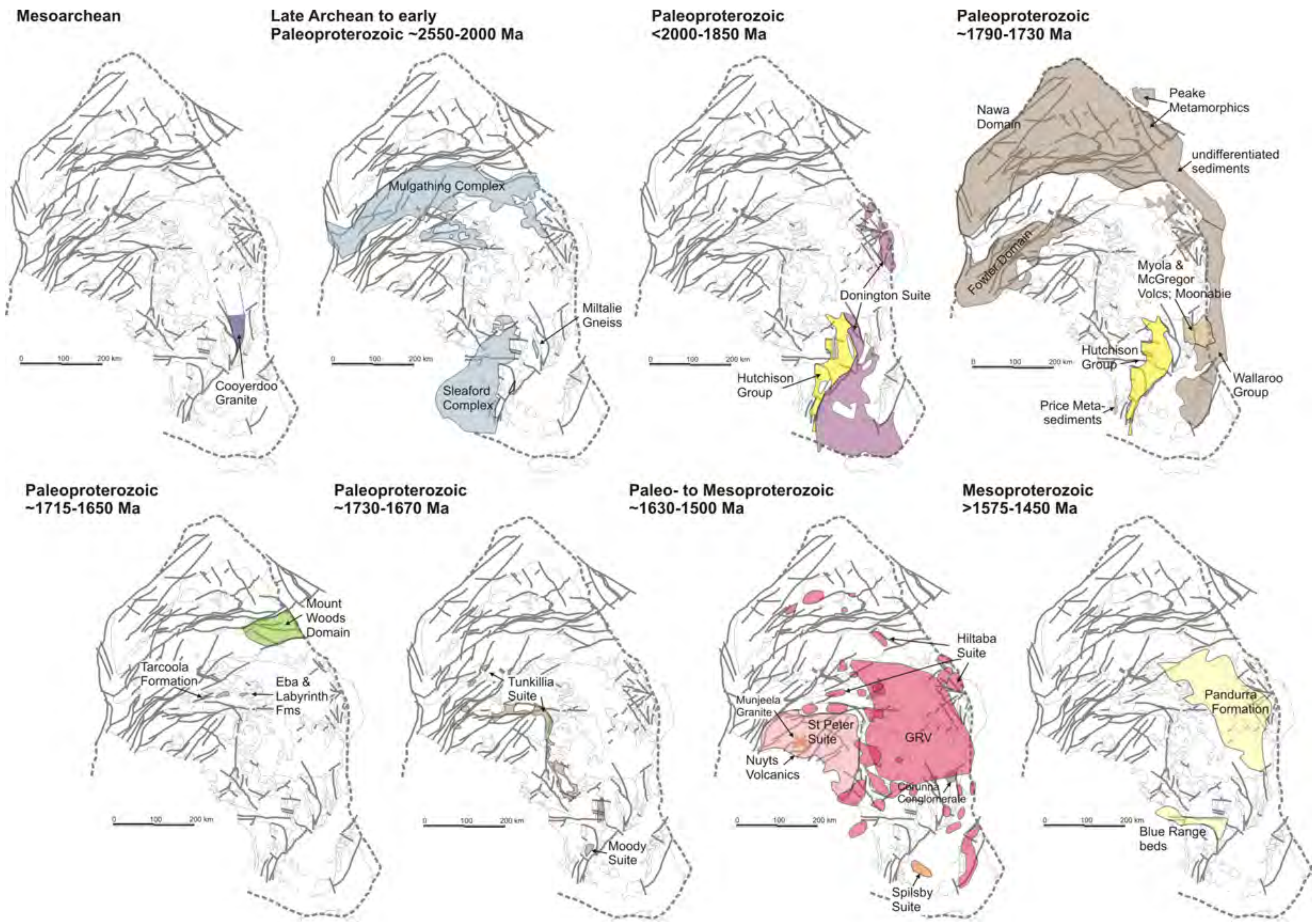

Figure 8d. Archean to Mesoproterozoic development of the Gawler Craton, showing the location of major rock units formed in each interval (modified after Hand et al., 2007).

1.3. Paleoproterozoic – Curnamona Province

Geological and tectonic summary

The ~1720-1640 Ma Willyama Supergroup (Page et al., 1998, 2000, 2003; 2005a, b; Stevens et al., 2008; Conor and Preiss, 2008; Figure 9) occupies the southern portion of the Curnamona Province. Basin fill consists of Paleoproterozoic metasedimentary and minor metavolcanic rocks, extensively intruded by syndepositional dolerite or gabbro (now amphibolite or mafic gneiss) dykes and sills and granite (now granite gneiss) sills, overlain by Mesoproterozoic volcanics and sedimentary rocks with granitoid intrusives (Stevens et al., 2008). The estimated thickness of the Willyama Supergroup ranges between 7 and 13 km (Willis et al., 1983; Stevens et al., 1988), although neither the base nor the top of the succession has been observed. Although basement rocks to the Willyama Supergroup are not exposed, inheritance indicates the presence of Archean and older Paleoproterozoic crust at depth (Cutten et al., 2006).

Rifting controlled the variable distribution and thickness of the Willyama Supergroup throughout the southern Curnamona Province (Conor and Preiss, 2008). There is a general increase in stratigraphic thickness from the western Olary Domain to the eastern Broken Hill Domain, with the greatest thickness of pelitic sediments in the Broken Hill Domain (Conor and Page, 2003). Equivalents in the Mulyungarie Domain are less complete or, more commonly, missing (Conor and Preiss, 2008; Figure 9). It has been proposed that the Broken Hill Domain was proximal to the depocentres of the Willyama basin, with the Olary Domain occupying a distal off-axis position (Conor and Page, 2003). Significant differences exist in the stratigraphic successions and synsedimentary intrusive histories between the Olary and Broken Hill domains (see Page et al., 2005b) and, whereas the Willyama Supergroup in the Olary Domain is thinner than that in the Broken Hill Domain, and subdivided into different units, correlation between the two domains has been achieved through systematic geochronology (e.g., Page et al., 2005a, b).

In the Olary Domain, the paleoflank exposes the ~1720-1715 Ma Curnamona Group, but overlying sediment packages are somewhat poorly developed. To the east, and closer to the rift axis, sedimentation in the Broken Hill Domain was more complete, consisting of ~1710-1705 Ma migmatitic-metasedimentary rocks, plus the overlying ~1705-1695 Ma Thackaringa Group, ~1695-1685 Ma Broken Hill Group and ~1685-1670 Ma Sundown Group (Conor and Preiss, 2008; Figure 8). Page et al. (2000), Payne et al. (2006a) and Barovich and Hand (2008) all proposed a northern Australian provenance for the source of sedimentary rocks from the lower Willyama Supergroup (e.g., Arunta Region). The ~1670-1640 Ma Paragon and Strathearn Groups, of the upper Willyama Supergroup of the Broken Hill and Olary Domains respectively, represent sag-phase deposition post-dating the climax of rifting, and coincided with a change in both basin configuration and provenance (Barovich et al., 2002; Barovich and Hand, 2008; Conor and Preiss, 2008). Earliest syndepositional magmatism is recorded in the Olary Domain by ~1720-1710 Ma A-type felsic volcanic and intrusive rocks and basaltic lavas (Conor and Preiss, 2008). The locus of magmatism subsequently shifted to the Broken Hill Domain, with partial melting of either the lower parts of the Willyama Supergroup or underlying rocks of similar composition (Barovich and Hand, 2004), to produce the S-type granitic sills of the ~1705-1685 Ma Silver City Suite (Stevens et al., 2008). Final magmatism occurred in the upper Broken Hill Group, with emplacement of Fe-rich tholeiitic sills and the ~1685 Ma felsic Hores Gneiss (Page et al., 2005a; Conor and Preiss, 2008).

Historically, the most favoured tectonic setting for the deposition of the Willyama Supergroup is a volcanically-active, intracontinental rift, with initial intercalated terrestrial, lacustrine to sabkha and marine successions being succeeded by deeper marine to lacustrine successions (Scheibner, 1974; Willis et al., 1983; Phillips et al., 1985; Laing, 1996b; Conor and Preiss, 2008). Lithostratigraphic evidence, including the generally shallow-water nature of the sediments, and geochemical characteristics of syndepositional bimodal magmatism, which are indicative of a continental rift environment (Ashley et al., 1996; Rutherford et al., 2006), has led workers to suggest that deposition of the Willyama Supergroup occurred in an intracontinental rifted-basin setting, with restricted ocean access (e.g., Willis et al., 1983; Raetz et al., 2002). A rift environment is also consistent with the presence of felsic volcanic rocks and high-Fe tholeiites (Rutherford et al., 2006).

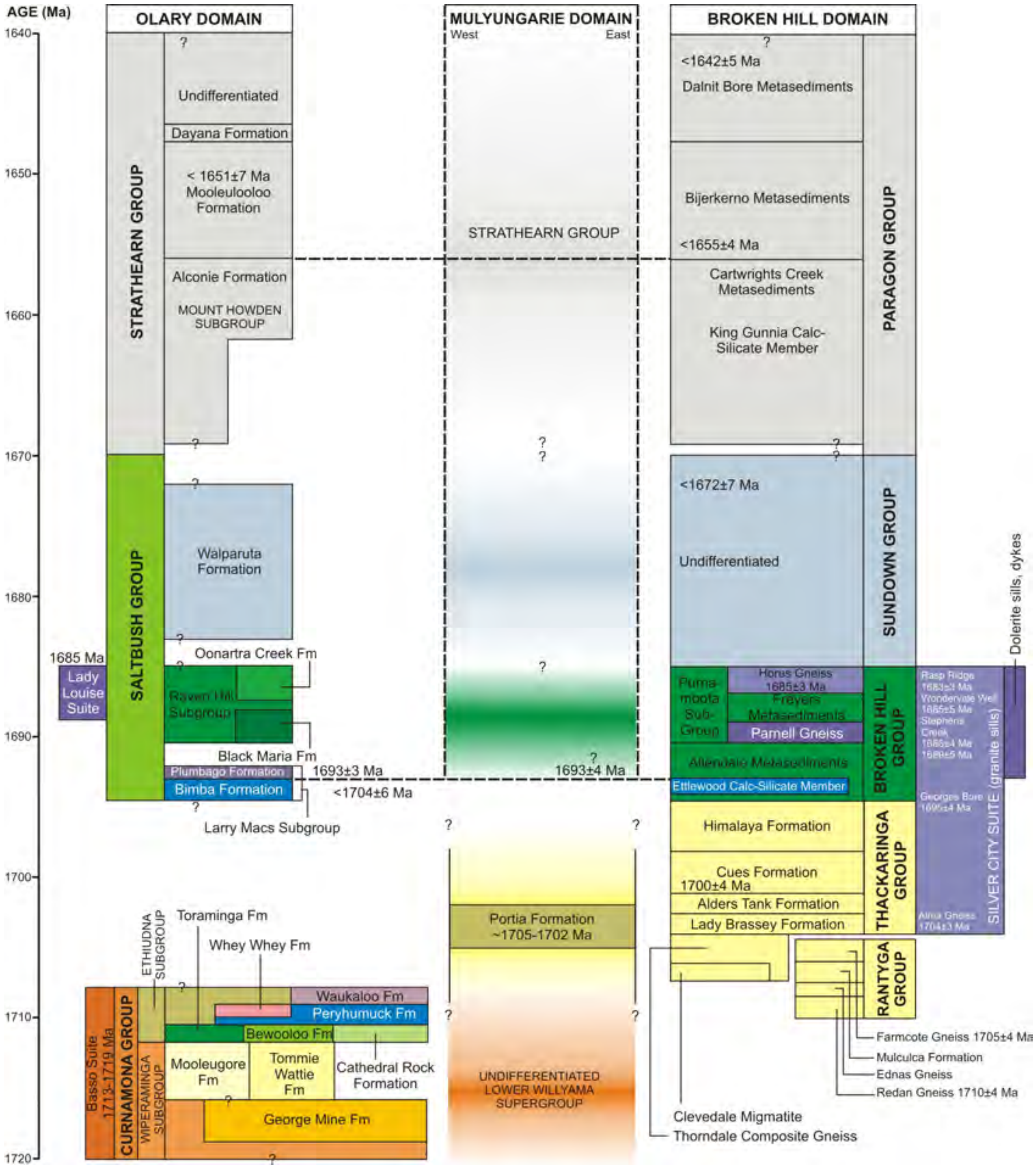

Figure 9. Lithostratigraphy of the Willyama Supergroup in the Olary, Mulyungarie and Broken Hill domains of the Curnamona Province (modified after Conor, 2006 and Conor and Preiss, 2008).

Giles et al. (2002, 2004), Fraser et al. (2007) and Gibson et al. (2008) have suggested that extension occurred in a backarc environment, which was part of a larger series of basins that include the Mount Isa Province and Georgetown Region (Calvert and Isan Superbasins) and that the Isan Orogeny correlates with the Olarian Orogeny. The tectonic environment of a backarc basin as the setting for the deposition of the Willyama Supergroup is based primarily on tectonic reconstructions which place the Curnamona Province adjacent to the Mount Isa Province and the southeastern margin of the North Australian Craton (NAC; Giles et al., 2002). In this model, the southern margin of the NAC is interpreted to be a long-lived, southward-migrating, convergent margin, with accretionary processes occurring sporadically between 1800 Ma and 1600 Ma (Giles et al., 2002).

The ca. 1600 Ma Olarian Orogeny provides a minimum depositional age for the Willyama Supergroup, although aspects of the deformational history, particularly in relation to timing, are still poorly understood and highly debated (Robertson et al., 2006).

Geological history and explanation of the Time-Space plot

Moolawatana Domain

- Previous work has suggested that basal metasedimentary units in the Mount Painter Inlier are equivalents of the Willyama Supergroup (e.g., Teale, 1993a). However, more recent age dating provides no current evidence for any time correlatives to the Willyama Supergroup in this domain (Fanning et al., 2003; Fraser and Neumann, 2010; Figure 10).

Olary Domain

- The Curnamona Group incorporates the oldest known metasedimentary rocks of the Willyama Supergroup, with no known equivalent in the Broken Hill Group (Conor, 2000; 2006; Figures 9, 10). The Curnamona Group is dated at ~1719 Ma, based on detrital zircons and the age of the intruding Basso Suite (SHRIMP U-Pb age of 1710 ± 9 Ma, Conor et al., 2006b; Page et al., in prep.).

- Thick, upward-fining successions, from psammite to andalusite-bearing pelite (George Mine and Tommie Wattie Formations), overlain by quartz-biotite metasedimentary rocks of the Mooleugore Formation, form the basal Wiperaminga Subgroup of the Curnamona Group (Figure 10). These units have an approximate minimum age of ~1718 Ma, and were deposited coevally with the extrusion and deposition of volcanic and epiclastic rocks of the Abminga Subsuite of the Basso Suite, dated at ~1719-1715 Ma (Page et al., 1998, 2000). Granitic sills belonging to the Basso Suite (Ameroo Subsuite) are slightly younger, to 1713 - ~1711 Ma (Page et al., 1998, 2000). The Basso Suite is largely A-type. A-type magmatism (1719-1711 Ma Basso Suite) and subsequent mafic volcanism (Montstephen Metabasalt Member) indicate that crustal extension and lithospheric thinning occurred during deposition of the Curnamona Group (Conor and Preiss, 2008).

- The Wiperaminga Subgroup is overlain by the Ethiudna Subgroup. The basal unit of the Ethiudna Subgroup is the Cathedral Rock Formation, consisting of quartz-rich metasedimentary and locally volcaniclastic rocks, which is overlain by the Beewooloo Member, a psammitic metasediment containing the Montstephen Metabasalt Member, the latter likely coeval with the Basso Suite. The remainder of the Ethiudna Subgroup, although lithologically diverse, is characterised by psammitic, psammopelitic and calcalbitite lithologies (Peryhumuck, Whey Whey and Waukaloo Formations) (Conor, 2006).

- The Ethiudna Subgroup is unconformably overlain by the Saltbush Group, a greatly-thinned equivalent of the Broken Hill and Sundown Groups in the Broken Hill Domain (Conor, 2006; Figure 10). The basal Larry Macs Subgroup of the Saltbush Group consists of the basal Bimba Formation, a unit of mixed sandstone, calcsilicate and carbonate, and the overlying tuffaceous Plumbago Formation (Conor, 2006), both of which have great regional continuity. The Bimba Formation is the stratigraphic equivalent to the Ettlewood Calc-Silicate Member of the Broken Hill Group (Clarke et al., 1986; Conor, 2006). This correlation indicates that there was a depositional hiatus in the Olary Domain while the Thackaringa Group was being deposited in the Broken Hill Domain, from ~1705 Ma to 1693 Ma (Conor, 2006). Page et al. (1998, 2000, 2005a) obtained a SHRIMP U-Pb zircon age of 1693 ± 3 Ma from tuffaceous metasandstone of the Plumbago Formation, and 1693 ± 4 Ma from a metasiltstone above the Ettlewood Calc-Silicate Member in the Broken Hill Domain. Above the Larry Macs Subgroup, above a possible unconformity, the Saltbush Group is dominated by siliciclastic metasedimentary rocks, mainly sandstone with thick mudstone intervals (Raven Hill Subgroup). The Raven Hill Subgroup is locally graphitic, contains calcsilicate ellipsoids and well-laminated, banded iron formations (Conor and Preiss, 2008). Another possible hiatus preceded deposition of the Walparuta Formation, considered the time equivalent of the Sundown Group in the Broken Hill Domain (Conor, 2006; Conor and Preiss, 2008). It is composed of metamorphosed mudstone and siltstone with minor sandstone layers and rare calcsilicate nodules.

- Felsic intrusives of the Lady Louise Suite have ages of ~1685 Ma (Conor and Fanning, 2001; Page et al., 2003). The Lady Louise Suite includes the mafic Woman-in-White Amphibolite dated at 1685 ± 4 Ma (Conor and Fanning, 2001), indicating that coeval felsic and mafic magmatism occurred at this time across the southern Curnamona Province. This age is similar to that of the S-type Rasp Ridge granite gneiss of the Silver City Suite of Stevens (2006), and is the temporal equivalent of the Hores Gneiss and rocks of the Parnell Formation in the Broken Hill Domain (Conor, 2006).

- The ~1667-1640 Ma Strathearn Group (Mount Howden Subgroup) is considered to be the direct equivalent of the Paragon Group in the Broken Hill Domain (Conor and Preiss, 2008; Figure 9). Metasedimentary rocks of the Strathearn Group are locally tuffaceous, and include aluminous mudstone, graphitic siltstone, albitic sandstone and brown mica schist (Conor, 2006). The basal Alconie Formation is a thickly layered succession of graphitic metasiltstone, sandstone and andalusite mudstone. The overlying Mooleulooloo Formation consists of thinly interlayered albitic sandstone and mica schist. Locally tuffaceous, the Mooleulooloo Formation has been dated at 1651 ± 7 Ma (Page, in preparation), an age similar to that obtained from the Bijerkerno Metasediments in the Broken Hill Domain (1655 ± 4 Ma; Page et al., 2000; Conor, 2006). The Dayana Formation, consisting of fine-grained sillimanite-andalusite-staurolite schist, forms the top of the Mount Howden Subgroup. Younger sedimentary rocks of the Willyama Supergroup were almost certainly deposited above the Strathearn and Paragon Groups to provide the cover necessary for metamorphism, and may be imaged on the north-south seismic transect (Korsch et al., 2010a). No syndepositional magmatism is recorded in this time interval (~1670-1640 Ma), and it is proposed that the Paragon and Strathearn Groups represent a thermal subsidence phase of sedimentation postdating rifting (Conor and Preiss, 2008).

Mulyungarie Domain

- Deposition of undifferentiated equivalents of the lower Willyama Supergroup, and possible equivalents of the Curnamona Group, occurred in the Mulyungarie Domain. Interpretation of geophysics suggests that thick successions of uppermost Willyama Supergroup are likely to be preserved under cover in this domain (Preiss, 2009). Interpreted equivalents of the Broken Hill Group are best developed in the eastern part of this domain, but are locally absent towards the west (Conor and Preiss, 2008).

- The Portia Formation (Conor, 2006) is a package of siltstone, calcareous siltstone and minor carbonate (Zang and Conor, 2006; Conor, 2006). The age of tuffs from the Portia Formation has been determined by U-Pb SHRIMP zircon dates of 1702 ± 6 Ma (Teale, 2000) and 1705 ± 4 Ma (Jagodzinski et al., 2006a). This places the Portia Formation within a sedimentary hiatus in the Olary Domain, but suggests it is temporally equivalent to the Thackaringa Group (Jagodzinski et al., 2006) (Figure 10).

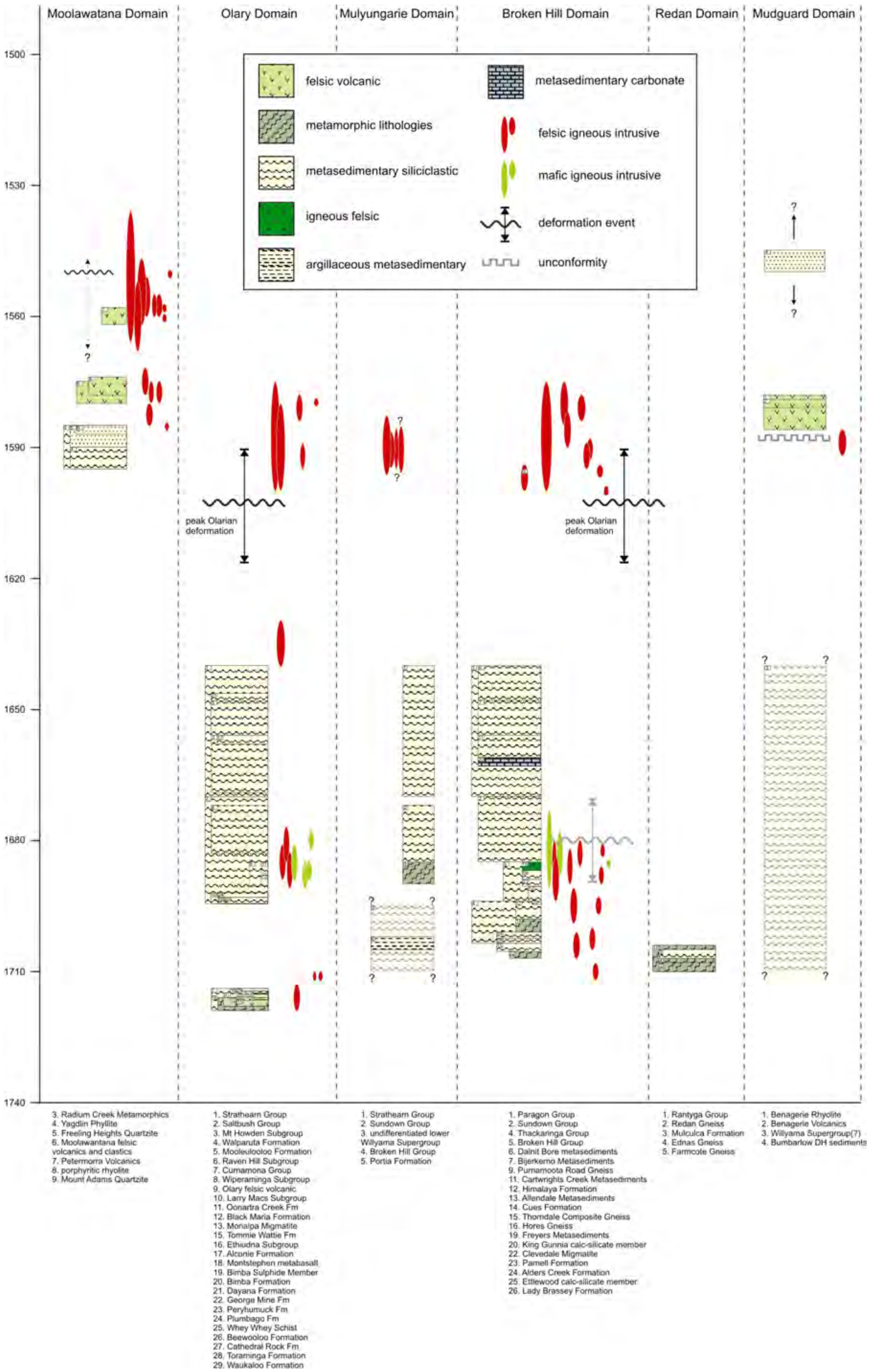

Figure 10. Time-space-event plot for domains of the Curnamona Province.

Broken Hill Domain

- The lower Willyama Supergroup in the Broken Hill Domain contains largely shallow water successions and includes migmatites, and metasedimentary rocks, psammitic to psammopelitic gneisses, often albitised and rich in magnetite, passing up into quartzofeldspathic gneisses, interpreted as metamorphosed rhyodacitic lava flows or tuffs (Page et al., 2005a). Intrusions in the Lower Willyama succession are represented by granitic gneisses and amphibolite sills. The basal units, the Clevedale Migmatite and Thorndale Composite Gneiss, have maximum U-Pb zircon depositional ages of ~1710 Ma. The S-type Alma Gneiss, dated at 1704 ± 3 Ma (Page et al., 2005a), intruded the Lady Brassey Formation and Cues Formation of the Thackaringa Group. The Cues Formation is dated by Stevens et al. (2008) at 1700 ± 4 Ma, from a leucogneiss interpreted to be a metavolcanic unit. The age of the Himalaya Formation (uppermost Thackaringa Group) is between that of the Cues Formation and the age of metasiltstone immediately overlying the Ettlewood Calc-Silicate Member in the overlying Broken Hill Group (see below).

- The upper Willyama Supergroup also contains largely shallow water successions and includes the Broken Hill Group, Sundown Group and Paragon Group (Conor and Preiss, 2008; Figures 9, 10). The Allendale Metasediments of the basal Broken Hill Group contain thin-bedded mudstone to siltstone, with minor mafic gneiss and calcsilicate nodules. This unit represents a transition in sedimentation from feldspar-rich psammitic metasedimentary rocks, found in the Thackaringa Group, to the more quartz-rich sandstone and mudstone metasedimentary rocks typical of the Broken Hill Group. The Allendale Metasediments contain the Ettlewood Calc-Silicate Member, a metamorphosed carbonate and an adjacent tuffaceous metasiltstone dated at 1693 ± 4 Ma (Page et al., 2005a). These are overlain by the Parnell Formation, characterised by the association of amphibolite-mafic granulite and garnetiferous quartzofeldspathic gneiss, interpreted as the product of bimodal rhyodacitic to tholeiitic basalt volcanism, with associated Zn- and/or Mn-rich rocks interpreted as exhalites. The overlying Freyers Metasediments are well-bedded mudstone to siltstone-sandstone with rare mafic gneisses (Page et al., 2005a).

- The upper unit of the Broken Hill Group is the Hores Gneiss, consisting of quartz-feldspar-garnet-biotite gneiss and quartzofeldspathic gneiss, with intercalated metasedimentary rocks. The protoliths are interpreted to have been mass-flow felsic volcanic rocks, and have a SHRIMP U-Pb zircon age of 1685 ± 3 Ma (Page et al., 2005a). The overlying Sundown Group consists of thin-bedded, commonly graded mudstone, siltstone and sandstone, with calcsilicate nodules. SHRIMP dating of detrital zircons give maximum depositional ages of 1688 ± 6 Ma and 1672 ± 7 Ma (Page et al., 2005a).

- The uppermost unit in the Willyama Supergroup is the Paragon Group, which has been divided into three formations (Figure 10). The Cartwrights Creek Metasediments consist of thin-bedded to laminated graphitic mudstone and siltstone and include the King Gunnia Calc-Silicate Member. These are overlain by the Bijerkerno Metasediments, consisting of thin-bedded to laminated, cross-bedded, fine-grained feldspathic graphitic sandstone, with a maximum U-Pb zircon depositional age of 1655 ± 4 Ma (Page et al., 2005a, in preparation), and by the Dalnit Bore Metasediments, consisting of laminated graphitic phyllite and rare graphitic psammite. A sample from a highly feldspathic siltstone bed, interpreted to represent an airfall tephra, produced a SHRIMP U-Pb age of 1642 ± 5 Ma (Page et al., 2005a). No other syndepositional magmatism is recorded at this time, and it is proposed that the Paragon Group represents a thermal subsidence phase of sedimentation postdating the climax of rifting (Conor and Preiss, 2008). Barovich et al. (2002) and Barovich and Hand (2008) provide isotopic evidence for a shift in the source of sediment supply during deposition of the upper Willyama Supergroup.

- Two preorogenic magmatic events have been recorded in the Broken Hill Domain, represented by the Alma Gneiss (1704 ± 3 Ma; Page et al., 2005a), which intruded as high as the Cues Formation, and the Rasp Ridge Gneiss (1683 ± 3 Ma; Page et al., 2005a), which intruded the Broken Hill Group. Both magmatic events include S-type granites, and the second event is bimodal, with mafic sills (now amphibolites). Mafic intrusives, now amphibolites, cutting the Thackaringa and Broken Hill Groups, have crystallisation ages of 1683 ± 5 Ma (Nutman and Ehlers, 1998) and 1682 ± 9 Ma (Page et al., 2005a). Recently, Stevens (2006) defined the Silver City Suite to include these two rock units plus three newly-identified magmatic units: the Oakdale Granite Gneiss (Georges Bore Granite) dated at 1695 ± 4 Ma, the Stephens Creek Granite Gneiss dated at 1689 ± 5 Ma in the lower part, and 1686 ± 4 Ma in the upper part, and the Wondervale Well Granite Gneiss, dated at 1685 ± 5 Ma. These units all intrude the Thackaringa Group and/or the Broken Hill Group (Figure 10). The geochronological data indicate that all of these bodies were emplaced at a very high level under little cover (Conor and Preiss, 2008). According to Barovich and Hand (2004), the S-type composition and isotopic characteristics of these magmas indicates their derivation by partial melting of either more deeply buried parts of the Willyama Supergroup, or underlying crust of similar composition.

Redan Domain

- The Redan Gneiss forms the basal unit in the Redan Domain (Figure 10). Metavolcanic gneiss within this unit has a SHRIMP zircon age of 1710 ± 4 Ma (Stevens et al., 2008). The Redan Gneiss is overlain by the Ednas Formation, the Mulculca Formation and the Farmcote Gneiss, together forming the Rantyga Group (Stevens et al., 2008). A-type felsic orthogneiss in the Farmcote Gneiss gives ages of 1703 ± 3 Ma and 1705 ± 3 Ma (Page et al., 2005a), while a metasedimentary rock in the Farmcote Gneiss contains detrital zircons dated at 1705 ± 5 Ma (Stevens et al., 2008), constraining the minimum age of the Redan succession at ~1705 Ma. Orthogneisses in the Farmcote Gneiss correlate with the S-type Alma Gneiss (1704 ± 3 Ma, Page et al., 2005a) in the Broken Hill Domain.

Mudguard Domain

- Undifferentiated Willyama Supergroup equivalents are interpreted to have been deposited in this domain (Figure 10). The only evidence comes from the base of drillhole Bumbarlow 1, where weakly cleaved siltstone is considered to be part of the Willyama Supergroup (Conor and Preiss, 2008; A. Schofield, pers. comm.).

1.4. Mesoproterozoic – Gawler Craton

Geological and tectonic summary

With the exception of the Nawa and Christie Domains in the west, felsic magmatism of ~1600-1570 Ma age is recorded throughout the Gawler Craton. Felsic and mafic lavas and ignimbrites of the Gawler Range Volcanics are coeval with, and genetically related to, A-type felsic plutons of the Hiltaba Suite (Giles, 1988; Blissett et al., 1993; Creaser, 1996; Figures 8, 11) and together they form the bimodal Gawler Range Volcanics-Hiltaba volcano-plutonic event (Johnson and Cross, 1995), which created a large igneous province covering a huge area in the central eastern Gawler Craton (Figure 11).

The Gawler Range Volcanics (GRV) have a maximum preserved thickness of ~1.5 km and cover >25 000 km^2 across the central Gawler Craton (Blissett et al., 1993; Allen et al., 2003). Volcanism occurred over a relatively short period of time between 1592 ± 3 Ma and 1591 ± 3 Ma (Fanning et al., 1988;

Creaser, 1995; Johnson and Cross, 1995). The Gawler Range Volcanics are divided into two broad groups. The lower group contains dacite-rhyodacite-rhyolite ignimbrites and flows, with thick, interlayered successions of basaltic lavas, whereas the upper group contains thick porphyritic dacite sheets predominantly ignimbritic in origin (Allen et al., 2003), emplaced in a subaerial, intracontinental setting (Daly et al., 1998). The predominance of rhyolites and dacites may, according to Allen et al. (2008), be related to an intracontinental setting in which crustal melting and stalling of mafic magmas favoured the production of felsic compositions. The Gawler Range Volcanics also extend beneath younger Proterozoic and Phanerozoic sedimentary successions of the Stuart Shelf in South Australia (Allen et al., 2008), and have been correlated with ~1600 Ma felsic volcanic clasts in glacial sedimentary units of the Terre Adélie Craton, Antarctica (Blissett et al., 1993; Peucat et al., 2002). The flat-lying, felsic Benagerie Volcanics were erupted at ca. 1583-1582 Ma (Fanning et al., 1998) in the Curnamona Province.

The extensive Hiltaba Suite is comagmatic with the Gawler Range Volcanics, with ages spanning the interval 1600 ± 16 Ma to 1575 ± 6 Ma (Flint et al., 1993; Creaser and Cooper, 1993; Creaser and Fanning, 1993; Stewart and Foden, 2003; Budd, 2006; Zang et al., 2007). The bulk of the Hiltaba Suite is highly fractionated granite to granodiorite, ranging in composition from quartz-monzonite through to two-mica granite; mafic rocks make up a minor but spatially widespread component (Hand et al., 2007). Outcrop is most abundant in the central Gawler Craton, particularly on the western and southwestern margins of the Gawler Range Volcanics. Plutonic rocks of the same age as the Hiltaba Suite also occur in the Curnamona Province and the Mount Painter Inlier (Figure 11). The Hiltaba Suite and Gawler Range Volcanics are regarded as comagmatic, with both having compositions similar to A-type magmas (Giles, 1988; Creaser and White, 1991; Creaser, 1996). Recent work, however, has shown that the Hiltaba Suite and Gawler Range Volcanics include both I- and A-type compositions, with a predominance of A-type granites and volcanic rocks in the eastern and central Gawler craton, whereas I-types are largely restricted to the central and western parts of the craton (Budd, 2006; Skirrow et al., 2006).

Traditionally, the Gawler Range Volcanics-Hiltaba Suite magmatic event has been attributed to development of a mantle plume within an anorogenic, intercontinental environment (Creaser, 1989; Flint et al., 1993). An anorogenic interpretation was applied because the Hiltaba Suite was thought to be undeformed and contain relatively homogeneous chemistry across the Craton. There is, however, evidence for deformation synchronous with emplacement of the Hiltaba Suite, as SHRIMP U-Pb data from metamorphic zircon gives ages between 1595-1583 Ma (Skirrow et al., 2006; Jagodzinski et al., 2007) in the Mount Woods Domain and Nawa Domain (e.g., Mabel Creek Ridge; Payne et al., 2008a). In the adjacent Coober Pedy Domain, SHRIMP data on metamorphic zircon (Fanning et al., 2007; Jagodzinski et al., 2007) and LA-ICPMS monazite ages (Payne et al., 2008a) suggest high temperature conditions may have prevailed as early as 1590 Ma, extending to 1565 Ma. These data have led to the suggestion of a syntectonic setting for the GRV-Hiltaba event. Deformation and metamorphism associated with the Gawler Range Volcanics-Hiltaba event occurred shortly before, and within uncertainty of, the ca. 1570-1540 Ma Kararan Orogeny (see below; Hand et al., 2007; Table 1). Hand et al. (2007) have suggested that syn-Hiltaba deformation and the Kararan Orogeny may actually be part of a continuum of magmatic emplacement, deformation and metamorphism from ca. 1595 Ma to ca. 1540 Ma.

Recent tectonic models for the setting of the Gawler Range Volcanics-Hiltaba Suite rocks have proposed a backarc setting behind an active subduction zone. Ferris et al. (2002) inferred a backarc continental setting behind a northeast-dipping subduction zone located at the southwest margin of the Gawler Craton. In this scenario, the Hiltaba magmatic event may be related to a northeast-dipping subduction zone located southwest of the Nuyts Domain, which produced the arc-related magmatism of the ~1620 Ma St Peter Suite. A similar setting was proposed by Betts et al. (2003b) and Betts and Giles (2006), who envisaged that the Hiltaba Suite was emplaced syn- to post-deformation in a continental backarc setting in the overriding plate during north or northeast-dipping flat slab subduction.

Alternatively, Swain et al. (2004, 2008) suggested that the Hiltaba Suite was related to south-dipping subduction under the northern margin of the Gawler Craton, resulting in the emplacement of the 1575-1595 Ma Hiltaba magmatic suite into a near margin continental backarc setting associated with the Musgrave magmatic arc. Betts et al. (2007) have suggested a model for A-type magmatism that evolved between 1600 Ma and 1500 Ma, in which the Hiltaba Granite Suite and the Gawler Range Volcanics define part of a hotspot track.

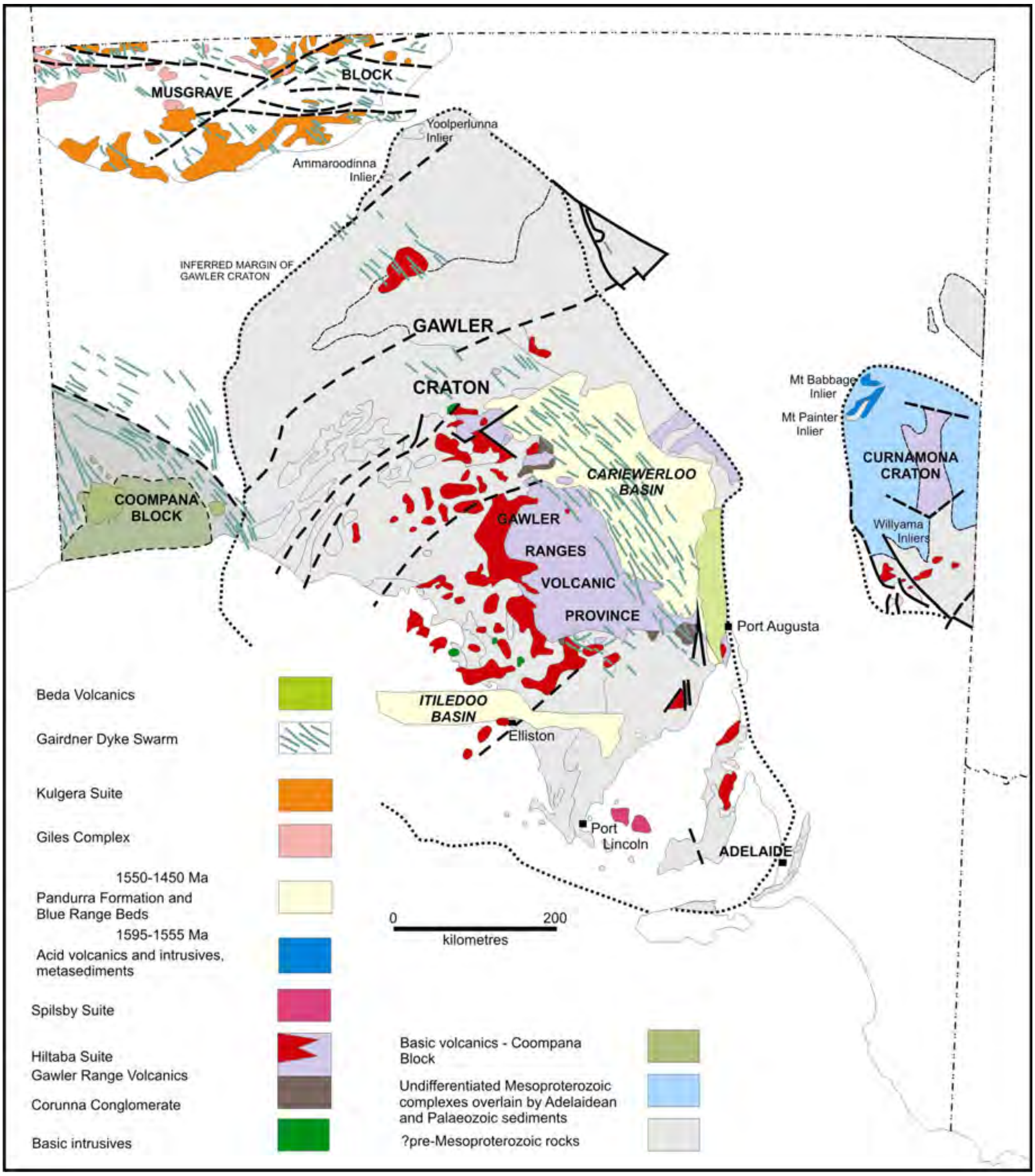

Figure 11. Distribution of Mesoproterozoic rocks in South Australia. Modified from Flint et al. (1993).

Only localised, ?undeformed magmatic rocks are recorded after 1575 Ma in the Gawler Craton (e.g., ~1490 Ma Spilsby Suite, Fanning et al., 2007; ca. 1450 Ma granites in the Nawa Domain; Howard et al., 2010b; ~1490 Ma granites in the Fowler Domain, Fanning, 1997; and ~1555-1530 Ma magmatism in the Peake and Denison Domain, Fanning et al., 2007). The Pandurra Formation (Cariewerloo Basin) and

Blue Range beds (Itiledoo Basin) may have been deposited in this interval (Fraser and Reid, 2007; see below; Figures 8, 11).

Geochronological evidence indicates the western Gawler Craton was reworked over the interval 1570-1450 Ma, accompanied by the reactivation of major northeast-trending faults and shear zones (e.g., Tallacootra Shear Zone, Coorabie Shear Zone, Karari Shear Zone; Fraser et al., 2002; Tomkins et al., 2004; Swain et al., 2005b; Fraser and Lyons, 2006). A small number of isotopic ages constrain the timing of the Kararan Orogeny to between 1570 Ma and 1540 Ma (Daly et al., 1998; Fanning et al., 2007; Hand et al., 2007; Table 1). Evidence for the Kararan Orogeny is largely restricted to the northern and western Gawler Craton, with peak metamorphic conditions of 800˚C and 10 kbar recorded in the Fowler Domain (Teasdale, 1997), and granulite-grade metamorphism in the Coober Pedy and Mabel Creek Ridge areas (Fanning et al., 2007; Payne et al., 2008a). Direen et al. (2005) introduced the term Coorabie Orogeny to encompass deformation in the Gawler Craton between ca. 1560 Ma and 1450 Ma, characterised by major reactivation of craton-scale, northeast-trending faults and shear zones, including the timing of movement on the Coorabie Shear Zone (Swain et al., 2005a; Fraser and Lyons, 2006; Thomas et al., 2008; Table 1). Currently, there is no evidence to support a single continuous orogenic event of this duration (Payne et al., 2008a). Therefore, both Hand et al. (2007) and Thomas et al. (2008) have suggested that the Coorabie Orogeny be restricted to the reactivation of shear zones and associated terrane-scale cooling (Teasdale, 1997; Swain et al., 2005b; Fraser and Lyons, 2006) in the time interval ca. 1470 Ma to 1450 Ma.

Geological history and explanation of the Time-Space plot

Nawa Domain

- Evidence for a high-grade, granulite facies metamorphic event within the Nawa Domain is recorded at ~1570-~1550 Ma (Payne et al., 2008a). This evidence comes from samples from drillholes within the Mabel Creek Ridge, in which metamorphic zircon and monazite ages of ~1570-1555 Ma are reported from paragneisses (Payne et al., 2006, 2008a).

- Recent monazite U-Pb age data has provided evidence for a group of granites intruded at ca. 1450 Ma in the Nawa Domain (Howard et al., 2010b), north of the Mabel Creek Ridge, accompanied by granulite facies metamorphism.

Coober Pedy Domain

- Granodioritic gneiss from the Coober Pedy Ridge is dated at 1589 ± 11 Ma (Fanning et al., 2007).

- Metamorphic zircon rims in cordierite-garnet-sillimanite gneiss yield a U-Pb SHRIMP age for metamorphism at ~1590 Ma (Fanning et al., 2007), synchronous with the emplacement of the Hiltaba Suite and Gawler Range Volcanics.

- Fine-grained biotite tonalite from drillhole GOMA DH1 (~308 m) produced a SHRIMP zircon igneous crystallisation age of 1571 ± 5 Ma (Jagodzinski et al., 2010).

- In situ LA-ICPMS dating of monazites from metapelitic granulites intersected in drillhole on the Coober Pedy Ridge, indicates that UHT metamorphism occurred at ~1590-1580 Ma (Cutts et al., 2010).

- Metamorphic rims on ca. 1750 Ma zircon cores from metapelite yield a SHRIMP U-Pb age of ~1590 Ma (Kararan Orogeny; Fanning et al., 2007).

Mount Woods Domain

- Granites of the Hiltaba Suite include the Balta Granite (intruding the Mount Woods Complex and the Engenina Adamellite), which has yielded a U-Pb zircon age of 1584 ± 18 Ma (Finlay, 1993). Other Hiltaba-aged intrusives have been dated at 1586 ± 3 Ma for leucogranite and 1587 ± 4 Ma for leucogabbro (Jagodzinski, 2005; Budd, 2006).

- In the Mount Woods Domain, a relatively tight cluster of ages occurs between 1594 ± 5 Ma and 1575 ± 3 Ma, including metamorphic zircon ages (Jagodzinski, 2005; Skirrow et al., 2006) from quartzites and felsic gneisses. Zircon U-Pb data from upper amphibolite to granulite-facies rocks gave metamorphic ages that range between 1587 ± 2 Ma and 1583 ± 18 Ma (Holm, 2004) to 1576 ± 5 Ma (Skirrow et al., 2006), indicating that Hiltaba Suite magmatism was coeval with high-temperature metamorphism in this domain.

Christie Domain

- Tomkins et al. (2004) dated K-feldspar from the Challenger gold mine by the $^{40}Ar/^{39}Ar$ method, and found consistent plateau ages, from the young component of the age spectrum, of 1535 ± 2 Ma to 1529 ± 2 Ma (three ages), which they considered to be related to a low-temperature thermal pulse associated with regional faulting.

- Fraser and Lyons (2006) determined a $^{40}Ar/^{39}Ar$ muscovite age of 1478 ± 11 Ma for possible shearing in the Christie Domain.

Fowler Domain

- A granitoid suite in the Central Block gave an interpreted magmatic crystallisation age of 1568 ± 11 Ma, with inherited components at ca. 1620 Ma and 1680 Ma (Fanning et al., 2007).

- High grade metamorphism is recorded by zircons from drillhole Nundroo 2, which have a U-Pb SHRIMP age of 1547 ± 9 Ma (Fanning et al., 2007). In the eastern Fowler Domain, an EMPA monazite age of 1557 ± 15 Ma for regional, lower-crustal metamorphism is recorded by Thomas et al. (2008), whereas Swain et al. (2005b) reported two EMPA ages of ~1510 Ma from monazite inclusions within garnet in rocks deformed by the Coorabie Shear Zone.

- A late, discordant pegmatite (Nundroo Block, DDH2) has an intrusive U-Pb SHRIMP zircon age of 1489 ± 4 Ma (Fanning et al., 2007; Teasdale, 1997), which may indicate it is coeval with the Spilsby Suite.

- Thomas et al. (2008) dated monazite from pelite in the Nundroo Block by EMPA at 1471 ± 14 Ma, which they interpreted to be related to transpressional reactivation.

Wilgena Domain

- The Mentor Formation, consisting of chloritic and sericitic mudstone, altered and tuffaceous breccia and tuffaceous rhyolite, overlies the Labyrinth Formation in this domain. This unit may be contemporaneous with extrusion of the Gawler Range Volcanics, although there are no age constraints (Flint et al., 1993). The Ealbara Rhyolite of the Gawler Range Volcanics produced a U-Pb zircon extrusive age of 1589 ± 16 Ma (Fanning, 1997).

- Archean and Paleoproterozoic sedimentary rocks in the Wilgena Domain were intruded by voluminous granitoids of the Hiltaba Suite at ~1590-1575 Ma (Budd, 2006), and mafic dykes (1582 ± 5 Ma; Budd and Fraser, 2004). Budd (2006) provided SHRIMP ages for plutons from the Hiltaba Suite ranging between 1592 ± 6 Ma and 1575 ± 4 Ma (five ages). The informal Ambrosia, Big Tank, Kychering, Partridge, and Pegler Granites and the Lady Jane Diorite have all been dated at between ~1590 Ma and ~1575 Ma (Budd, 2006).

Harris Greenstone Domain

- Younger Proterozoic components of the Harris Greenstone Domain consist of felsic intrusions (Hiltaba Suite granites), and felsic and mafic extrusive rocks of the Gawler Range Volcanics (Hoatson et al., 2005).

Nuyts Domain

- Granites of the Hiltaba Suite in this domain have ages from 1596 ± 9 Ma to 1562 ± 16 Ma. The Tyringa pluton has an age of 1583 ± 11 Ma (Fanning, 1997). The unfractionated, restite-rich, muscovite-biotite±garnet-bearing S-type Munjeela Granite gave an EMPA monazite emplacement age of 1562 ± 15 Ma (Ferris, 2001), and a U-Pb monazite age of ca. 1585 Ma (Payne et al., 2007). Stewart and Foden (2003) provide U-Pb Kober zircon ages of 1578 ± 12 Ma and 1575 ± 13 Ma, for the Arcoordaby and Frogs Eyes plutons, respectively.

- Evidence for tectonic activity post-dating the Hiltaba granites is limited to an EMPA monazite age of 1503 ± 16 Ma, reported by Swain et al. (2005b) from the Yerda Shear Zone along the northern margin of the Nuyts Domain.

Coulta Domain

- Granites of the Hiltaba Suite include the Calca Granite, with a SHRIMP U-Pb zircon age of 1583 ± 10 Ma (Fanning et al., 2007), and the Buckleboo Granite (1586 ± 6 Ma) (Fraser and Neumann, 2010).

- $^{40}Ar/^{39}Ar$ ages from muscovite and hornblende from the western side of Eyre Peninsula indicate muscovite closure to argon diffusion at ~1650 Ma in the Sleaford Complex, and local resetting of hornblende at ~1590 Ma, coincident with intrusion of the Hiltaba Suite (Foster and Ehlers, 1998).

- Fraser and Neumann (2010) record a U-Pb SHRIMP zircon maximum depositional age of 1605 ± 9 Ma for sandstone from the Blue Range beds (Itiledoo Basin, Figure 11) at Mount Wedge.

Cleve Domain

- Granites of the Hiltaba Suite and Gawler Range Volcanics, with ages ~1600-1590 Ma, occur particularly around the northern margin of the domain (e.g., Menninnie Dam; Fanning et al., 2007), and include the Cunyarie (Thomson, 1969) Granite (SHRIMP U-Pb zircon crystallisation age of 1598 ± 7 Ma; Fraser and Neumann, 2010).

- Zircons derived from red sandstone belonging to the Blue Range beds (near Ningana Station; Itiledoo Basin, Figure 11), give a SHRIMP U-Pb maximum depositional age of 1750 ± 14 Ma (Fraser and Neumann, 2010).

Spencer Domain

- Deposition of the Corunna Conglomerate took place somewhere in the interval between ~1660 Ma and 1585 Ma. It is a succession of unmetamorphosed fluvial to shallow marine sediments, comprising conglomerates, sandstones and siltstones deposited in active fault-bounded grabens (McAvaney and Reid, 2008), unconformably overlying the Moonabie Formation and

McGregor Volcanics (~1740 Ma) (Parker et al., 1988). U-Pb zircon maximum depositional ages of 1765 ± 15 Ma (Red Conglomerate Member, Tassie Creek Reservoir, Corunna Station) and 1746 ± 9 Ma (Red Conglomerate Member, Wartaka Station, Corunna Range) for the Corunna Conglomerate are recorded by McAvaney and Reid (2008). Three samples of Corunna Conglomerate from Baxter Hills gave SHRIMP U-Pb zircon maximum depositional ages of 1752 ± 5 Ma, 1718 ± 12 Ma and 1680 ± 7 Ma, while another two samples from Uno Range gave maximum depositional ages of 1690 ± 7 Ma and 1659 ± 7 Ma (Fraser and Neumann, 2010). A minimum age constraint is provided by the intrusion of porphyritic dykes of the Gawler Range Volcanics and local intrusions of Hiltaba granite.

- The Charleston Granite, a pluton consisting of megacrystic granite, intrudes the McGregor Volcanics and Corunna Conglomerate (Creaser and Fanning, 1993), and is dated at 1585 ± 5 Ma (Creaser and Fanning, 1993).

- Granite plutons of the Spilsby Suite and associated felsic dykes intrude Donington Suite granitoids in the Sir Joseph Banks Group of Islands in the southern Spencer Gulf (Flint et al., 1993). Lithologies within the suite include feldspar-bearing granite, granite to aplite and granodiorite (Fanning et al., 2007). A U-Pb SHRIMP age for zircon crystallisation of 1510 ± 12 Ma (Fanning, 1997) has been revised to 1497 ± 13 Ma by Fanning et al. (2007), who dated a second sample at 1497 ± 38 Ma. A post-tectonic dyke intruding the Corny Point Paragneiss and Donington Suite on southwestern Yorke Peninsula, dated at 1509 ± 9 Ma (Jagodzinski et al., 2006), may also belong to this suite.

Gawler Range Volcanics Domain

- Extensive mafic and felsic lavas of the Gawler Range Volcanics dominate this domain. U-Pb isotopic dating of the Yardea and Wanganny Dacites gives ages of 1592 ± 3 Ma and 1591 ± 3 Ma, respectively (Fanning et al., 1998; Creaser, 1993).

Olympic Domain

- The Olympic Domain experienced extensive intrusion of granites of the Hiltaba Suite and extrusion of felsic lavas and ignimbrites, with minor mafic lavas, of the Gawler Range Volcanics (Creaser and Cooper, 1993; Jagodzinski, 2005). Hiltaba Suite granites in this domain include the Tickera (1577 ± 7 Ma), Arthurton (1582 ± 7 Ma; Creaser and Cooper, 1993) and Roxby Downs granites, intruding the Wallaroo Group. Microdiorite at Snake Gully was dated at 1591 ± 3 Ma (Jagodzinski, 2005), and the Curramulka Gabbronorite (of continental tholeiite composition; Zang et al., 2007) gave a crystallisation age of 1589 ± 5 Ma (Zang et al., 2002; Zang et al., 2007). Other Hiltaba granites produced ages of 1600 ± 16 Ma, 1598 ± 7 Ma, 1586 ± 5 Ma and 1575 ± 6 Ma (Zang et al., 2007). U-Pb zircon dating of the bimodal Wirrda Suite, consisting of quartz monzodiorite, monzonite, syenite and aplite, has produced crystallisation ages of 1598 ± 2 Ma, 1590 ± 5 Ma, 1593 ± 4 Ma and 1593 ± 2 Ma (Creaser and Cooper, 1993).

- Deposition of the Pandurra Formation, a thick succession of undeformed mainly red beds, post-dated extrusion of the Gawler Range Volcanics (Cariewerloo Basin, Figure 11; Cowley, 1991). It is typically a medium- to coarse-grained quartz-rich and lithic-rich sandstone, but also includes pebble conglomerate and shale (Flint et al., 1993). Fanning et al. (1983) reports that the Pandurra Formation has a poorly-constrained minimum depositional age of 1424 ± 51 Ma, based on Rb-Sr geochronology. Recent U-Pb SHRIMP zircon geochronology gave a maximum depositional age of 1756 ± 5 Ma for the Pandurra Formation at Mount Laura, and 1575 ± 5 Ma for a sample from Red Rock Hill (Fraser and Neumann, 2010).

Peake and Denison Domain

- Fanning et al. (2007) record a magmatic event between ~1555 Ma and ~1530 Ma in this domain (granites and aplites at Lagoon Hill at 1555 ± 14 Ma and ~1530 Ma). Ferris et al. (2002) also report a SHRIMP U-Pb zircon age from an unnamed granite intruding the Peake Metamorphics at 1536 ± 6 Ma.

1.5. Mesoproterozoic – Curnamona Province

Geological and tectonic summary

During the early Mesoproterozoic (ca. 1610-1550 Ma), a continental-scale thermal event, characterised by regional, high-temperature metamorphism and bimodal magmatism was recorded in the Proterozoic of eastern Australia. In the Gawler Craton, this event included the extrusion of the bimodal Gawler Range Volcanics (ca. 1592 Ma) and the intrusion of the Hiltaba granite suite (ca. 1595-1575 Ma). Contemporaneous magmatism also occurred in the Curnamona Province, where vast quantities of A-, I- and S-type magma were generated. A-type magmas of the bimodal Benagerie Volcanics were extruded at ca. 1585-1581 Ma (Fanning et al., 1998), and emplacement of the I- and S-type granites of the Ninnerie Supersuite occurred at ca. 1590-1570 Ma throughout the Curnamona Province (Ludwig and Cooper, 1984; Fricke, 2009). In the Moolawatana Domain, A-type granites and felsic volcanic successions were emplaced at ca. 1575 Ma and ca. 1555 Ma (e.g., Teale, 1993a; Fanning et al., 1998; McLaren et al., 2006; Neumann et al., 2009; Fraser and Neumann, 2010). Widespread S-type granites of the Bimbowrie Suite were intruded at ~1590 Ma in the late stages of the Olarian Orogeny in the Olary Domain and the Mundi Mundi type granites in the Broken Hill Domain are comparable (~1600-1580 Ma; Page et al., 2005a).

The Olarian Orogeny, an episode of major, midcrustal basin inversion involving polyphase deformation and high-T, low-P metamorphism, occurred around the Paleoproterozoic-Mesoproterozoic boundary. The Willyama Supergroup was multiply deformed, metamorphosed and locally intensely metasomatically altered, particularly in the lower part of the succession in the Olary Domain (Page et al., 2005b). Early stages of the Olarian Orogeny involved development of large-scale recumbent to shallowly inclined, sheath-like folds during thin-skinned deformation (Majoribanks et al., 1980; Hobbs et al., 1984; Forbes et al., 2004; Conor and Preiss, 2008). These were overprinted by approximately northeast-trending upright folds, developed during thick-skinned deformation (Majoribanks et al., 1980; Hobbs et al., 1984; Forbes et al., 2004). The rocks of the southern Curnamona Province were metamorphosed to upper amphibolite to granulite facies conditions (Phillips and Wall, 1981; Willis et al., 1983; Page and Laing, 1992). Peak amphibolite to granulite facies conditions were reached at ca. 1600 Ma, attaining temperatures and pressures of at least 740˚C at ~5 kbar in the southern Broken Hill Block (Forbes et al., 2005). Spatially, there is an increase in metamorphic grade, from andalusite-bearing amphibolite facies to two-pyroxene granulite facies, from the northwest towards the southeast (Phillips and Wall, 1981; Willis et al., 1983; Hobbs et al., 1984; Clarke et al., 1987; Page and Laing, 1992).

Possibly the most debated issue pertaining to the Curnamona Province is the timing and duration of Olarian metamorphism and deformation (Nutman and Ehlers, 1998; Page et al., 1998, 2000, 2003, 2005a; Gibson and Nutman, 2004; Conor et al., 2005), with two contrasting models being proposed: 1. tectonism at ca. 1690 Ma and at 1600 Ma (Gibson and Nutman, 2004); and 2. tectonism only at ca. 1600 Ma (Page and Laing, 1992; Page et al., 1998, 2000, 2003, 2005a).

Based on SHRIMP dating, some workers (e.g., Nutman and Ehlers, 1998; Gibson and Nutman, 2004; Gibson et al., 2004) consider that high T metamorphism of the lower Willyama Supergroup occurred between ca. 1690 Ma and 1670 Ma, in addition to the later Olarian Orogeny. In this model, the development of migmatites and a layer-parallel foliation is attributed to crustal extension during lithospheric thinning and bimodal magmatism at ca. 1690-1670 Ma. The lower Willyama Supergroup was exhumed and juxtaposed against the upper Willyama Supergroup along a low angle detachment surface as a metamorphic core complex (Gibson and Nutman, 2004; Gibson et al., 2004). Evins et al. (2008) report matrix monazite EMPA age peaks from pelites in the Sundown Group of ~1675 Ma and ~1635 Ma, with a second sample yielding monazite ages ranging from >1700 Ma to ~1550 Ma, supporting an earlier metamorphic event. Similarly, White et al. (2006) report EMPA monazite ages from metasedimentary rocks throughout the stratigraphy, recording a broad peak from ca. 1720 Ma to 1600 Ma, also supporting an earlier metamorphic event. McFarlane and Frost (2009) record pulses of monazite growth at ca. 1657 Ma, ca. 1630 Ma and ca. 1602 Ma, which correspond to monazite growth during lower amphibolite facies, upper amphibolite facies and granulite facies metamorphism, respectively; but find no evidence of monazite growth at 1690-1670 Ma (c.f. Gibson and Nutman, 2004).

The model of Gibson and Nutman (2004) and Gibson et al. (2004) has been refuted by Conor et al. (2005), Page et al. (2005a), Rutherford et al. (2007) and Stevens et al. (2008), who find no evidence of a ca. 1690-1670 Ma event, and indicate that most of the metamorphic zircon and rutile ages, as well as the ages of syn- and late-tectonic granites indicate peak Olarian metamorphism was constrained to 1610-1590 Ma. Page et al. (2005a) bracketed the D1, D2 and D3 events as inseparable within the interval 1597 ± 3 Ma to 1591 ± 5 Ma on the basis of syn- and post-tectonic granites. Forbes et al. (2005, 2008) considered that the M1 metamorphism was associated with mid-crustal extension, and that in situ SHRIMP dating of monazite of ca. 1620 Ma provided an age for the timing of this prograde amphibolite metamorphic event. They considered that this extensional event was terminated by crustal shortening, basin inversion and M2 metamorphism during the Olarian Orogeny which they dated at ca. 1600 Ma by in situ SHRIMP analyses of monazite. Rutherford et al. (2007) record an additional episode of high grade metamorphism in the Broken Hill region, and retrograde metamorphism preserved in the Olary Domain and western Broken Hill Domain between ca. 1570 Ma and 1550 Ma using EMPA analyses of monazite. Rutherford et al. (2007) correlate this event to ~1550 Ma tectonism in the Mount Isa Province and Georgetown Region.

Geological history and explanation of the Time-Space plot

Moolawatana Domain

- Within the Mount Painter Inlier, Fanning et al. (2003) analysed detrital zircon samples from the Hot Springs Creek, Mawson Plateau, Gordon Springs Creek and Corundum Creek areas, concluding that deposition occurred later than 1630 Ma. More recent U-Pb zircon data indicate that the deposition of the Radium Creek Metamorphics (U-Pb SHRIMP zircon maximum depositional ages of 1591 ± 6 Ma and 1600 ± 8 Ma; Fraser and Neumann, 2010), Yagdlin Phyllite, Mount Adams Quartzite and Freeling Heights Quartzite occurred at or after ~1595-1590 Ma (maximum depositional ages of Fanning et al., 2003; Ogilvie, 2006; Shafton, 2006; Neumann et al., 2009 and Fraser and Neumann, 2010; Figure 10).

- Intrusion of the Mount Neill Granite (1569 ± 14 Ma, Fanning et al., 1998; 1585 ± 3 Ma, Fraser and Neumann, 2010), Pepegoona Porphyry, Box Bore Granite (~1560 Ma, Stewart and Foden, 2003; 1584 ± 2 Ma, Fraser and Neumann, 2010) and Hot Springs gneisses occurred at ~1585-1575 Ma (McLaren et al., 2006; Neumann et al., 2009). Felsic volcanics and quartz-rich sedimentary rocks (1580-1575 Ma) were deposited synchronously with the intrusives; porphyritic rhyolites have ages of 1576 ± 2 Ma (Fanning et al., 1998) and 1575 ± 14 Ma (Fanning et al., 1986).

- A second magmatic event, the Moolawatana Suite, includes the Terrapinna (1557 ± 6 Ma, Sheard et al., 1992; 1556 ± 4 Ma, Cooper et al., 1982; 1560 ± 3 Ma, Fraser and Neumann, 2010), Wattleowie (1563 ± 3 Ma, Fraser and Neumann, 2010), Old Camp and Yerila Granites (1556 ± 10 Ma, Cooper et al., 1982; 1558 ± 3 Ma, Fraser and Neumann, 2010), the Hodgkinson Granodiorite (1552 ± 4 Ma, Fraser and Neumann, 2010), as well as other smaller granite bodies, occurred at ~1565-1555 Ma (McLaren et al., 2006; Neumann, 2001; Neumann et al., 2009). Fanning et al. (1998) obtained SHRIMP U-Pb zircon ages for a granite from Radium Creek (1560 ± 8 Ma), the Old Camp Granite (1555 ± 8 Ma) and the Camel Pad Granite (1551 ± 15 Ma). Teale (1993a, 1993b) indicated that, geochemically, the granites had an A-type geochemistry, and that they plotted in the within-plate granite field on tectonic discriminant diagrams. Extrusion of the Petermorra Volcanics also occurred at this time. Porphyritic rhyodacite to rhyolite of the Petermorra Volcanics from the northern Mount Babbage Inlier has an interpreted crystallisation age of 1560 ± 3 Ma (Sheard et al., 1992).

- ~1565-1555 Ma magmatism was followed by high grade metamorphism (1553 ± 4 Ma), as recorded by the Hot Springs gneisses (magmatic crystallisation age of 1582 ± 6 Ma, Neumann et al., 2009, Fraser and Neumann, 2010).

- A highly altered calcsilicate sample collected from drillhole SPH1 (between 414 m and 418.9 m depth) was dated through U-Pb zircon SHRIMP geochronology. The majority of analyses provided a weighted mean age of 1558 ± 2 Ma, interpreted to record a magmatic crystallisation age (Fraser and Neumann, 2010), possibly indicating that this rock is a calc-silicate altered granite.

Olary Domain

- Intrusion and extrusion of Early Mesoproterozoic granites and related rocks of the Ninnerie Supersuite took place syn-late to post-Olarian Orogeny (Fricke, 2006, 2008, 2009), including the Bimbowrie and Crocker Well Suites and the near-flat-lying ~1580 Ma Benagerie Volcanics (Conor et al., 2006; Figure 10).

- In the Olary Domain, the Ninnerie Supersuite includes the dominantly potassic, S-type, two-mica granites of the Bimbowrie Suite (Barovich and Foden, 2002; Stewart and Foden, 2003), and the dominantly biotite-rich and sodic varieties of the Crockers Well Suite, which includes trondhjemite, alaskite, and more mafic I-type granodiorite and diorite (Barovich and Ashley, 2002). Dates recorded for the Bimbowrie Suite range from ~1592 Ma to 1580 Ma (Cook et al., 1994; Page et al., 2005a). Several U-Pb-Th isotope ages reported for the Crocker Well Suite range from ~1580 Ma to 1568 Ma (Ludwig and Cooper, 1984).

- According to Fricke (2006), the Bimbowrie and Crocker Well suites probably represent separate magmatic events or sources, illustrated through differences in spatial and temporal distribution and geochemistry. The majority of the Ninnerie Supersuite are of S-type composition, and have been interpreted to have been sourced from anatexis of metasedimentary rocks of the Willyama Supergroup or from underlying crust of similar composition (Barovich and Foden, 2002). The Bimbowrie Suite is coeval with the Mundi Mundi granites in the Broken Hill Domain, undeformed felsic volcanics found in the northern Benagerie Ridge (1580 Ma), and with the I-A type Hiltaba Suite granites, and Gawler Range Volcanics of the Gawler Craton (Robertson et al., 2006).

- The Billeroo alkaline magmatic complex (Knaak, 2002) in the northwestern Olary Domain is considered to be of similar age to the Ninnerie Supersuite, containing syenite and ijolite veined by lamprophyre (Rutherford et al., 2002, 2007).

Mulyungarie Domain

- At least four varieties of intrusive rocks have been identified in the Mulyungarie Domain, including: buried magnetic gabbro to granodiorite and non-magnetic granodiorite to monzogranite; buried non-magnetic granite (Honeymoon Type), and magnetic, anatectic granite (Mundaerno Type) (Fricke, 2008, 2009). Little is known about the geochemistry, petrology and age of these intrusive bodies, due to extensive cover and limited drilling. The granites in this region have A-type affinities, and an inferred age of ~1590 Ma (Teale, 2000; Fricke, 2008, 2009).

- Only four concordant zircon analyses were collected from an unnamed metasedimentary rock, from the ETMA5 1 drillhole in this domain (between 444.8 m and 449.2 m depth). The youngest age is 1623 ± 42 Ma, and three other analyses combine to provide a weighted mean age of 1705 ± 15 Ma (Fraser and Neumann, 2010). Although the paucity of data limits geological interpretation, the ~1625 Ma and ~1705 Ma ages are different from dominant age populations from samples from the Bumbarlow 1 and SPH 1 drillholes (Moolawatana and Mudguard domains).

Broken Hill Domain

- In the Broken Hill Domain, at least four types of Mesoproterozoic granites have been identified (i. Purnamoota Road leucocratic gneiss, ii. Mundi Mundi type, iii. Champion type, and iv. Umberumberka type; Stevens and Willis, 1983; Brown et al., 1983). The Purnamoota Road leucocratic gneiss is a coarse leucocratic granitoid which gives a crystallisation age of 1597 ± 3 Ma (Page et al., 2005a). Mundi Mundi type granites are massive post-D3 granites, one of which gives a crystallisation age of 1591 ± 5 Ma (Page et al., 2005a). This type also includes the Cusin Creek (1596 ± 3 Ma, Page et al., 2005a) and Brewery Creek (1580 ± 5 Ma, Page, 2006) plutons. The Mundi Mundi type granites are of similar age and composition to the Bimbowrie Suite (Conor and Preiss, 2008). Other granite types in the Broken Hill Domain include the Champion type (1581 ± 3 Ma, Page, 2006) and the Umberumberka type (1586 ± 4 Ma, Page, 2006).

- The Bimbowrie Suite granites and the Mundi Mundi type granites share similar mineralogy and geochemical compositions and, according to Fricke (2008), may loosely be classified as part of the same suite. No detailed geochemical investigations of these granite types in the Broken Hill Domain have been conducted.

- Intrusion of most of these granites occurred after peak metamorphism, although the 1596 ± 3 Ma Cusin Creek pluton is interpreted to have intruded prior to the development of the upright fabric (Page et al., 2005a) and, thus, D3 deformation is bracketed between 1591 ± 5 Ma (minimum defined by Mundi Mundi intrusion) and 1596 ± 3 Ma (maximum age defined by Cusin Creek pluton) (Page et al., 2005a).

Mudguard Domain

- Drillhole Bumbarlow 1 in the Mudguard Domain contains interbedded clastic sedimentary rocks and basalts (Schofield and Neumann, pers. comm.). A unit of black shale, metamorphosed to greenschist facies, probably representing weakly deformed Willyama Supergroup, unconformably underlies the sandstone and basalt unit (interpreted as Benagerie Volcanics) at the base of Bumbarlow 1 (Conor and Preiss, 2008). Detrital zircon geochronology of sandstone samples at 686-691 m (unnamed white sandstone below the basalt) and 469-474 m (unnamed red sandstone above the basalt) in Bumbarlow 1, gave U-Pb maximum depositional ages of 1591 ± 6 Ma and 1550 ± 6 Ma, respectively (Fraser and Neumann, 2010).

- Medium-grained porphyritic granite, collected from drillhole BRD 012 (between the 482 m to 488 m depth intervals), gave a U-Pb SHRIMP zircon magmatic crystallisation age of 1589 ± 5 Ma (Fraser and Neumann, 2010).

- The Mudguard Domain is host to the Benagerie Volcanics, which consists of A-type porphyritic rhyolite, rhyodacite, trachyte, andesite and rare basalt (Teale and Flint, 1993). Due to extensive cover, the Benagerie Volcanics are only known from drillholes (Fricke, 2008). The A-type felsic porphyritic volcanic rocks form a thick, flat-lying sheet which unconformably overlies metasedimentary rocks of the Willyama Supergroup. There is approximately a 10 Ma time period between the cessation of the D3 Olarian deformation and extrusion of the Benagerie Volcanics (Korsch et al., 2006).

- Reported ages for the Benagerie Volcanics range between 1585 Ma and 1581 Ma. For example, a porphyritic rhyolite from drillhole Mudguard 1 has a SHRIMP age of 1581 ± 4 Ma (Fanning et al., 1998) and two unpublished ages are 1583 Ma and 1582 Ma (M. Wingate, pers. comm.). The A-type Benagerie Volcanics and A- and I-type granites to the north (Moolawatana Suite and Mount Neill Granite) have geochemistry closely resembling that of the slightly older Gawler Range Volcanics and A-type Hiltaba Suite granites in the neighbouring Gawler Craton (Barovich et al., 2006).

- Associated with the Benagerie Volcanics are basaltic and andesitic volcanic breccias intersected in drillhole Quinyambie 1 (Fricke, 2008), and amygdaloidal basalts and felsic volcanics encountered in several drillholes (Fricke, 2008, 2009). The intermediate intrusives and volcanics of the Mudguard Domain show evidence of a mantle component (Conor et al., 2006).

1.6. Shear zones in the Gawler Craton: nature and movement history

As outlined in the Introduction, the Gawler Craton has been subdivided into several domains based on varying magnetic, gravity, structural, geochronological, isotopic and geochemical character (Teasdale, 1997; Ferris et al., 2002), the boundaries of which commonly correspond to major shear zones (Figures 2, 4, 12). The most prominent of these structures is the northeast-trending Karari Shear Zone, an ~800 km long structure that divides the Nawa Domain to the northwest from the Christie Domain and Coober Pedy Domain to the southeast (Swain et al., 2005b; Fraser and Lyons, 2006). The southern part of the Christie Domain is separated from the Fowler Domain by the Tallacootra Shear Zone, one of a set of anastomosing, northeast-striking structures that characterises both the Fowler and Christie Domains (Teasdale, 1997). East of the Tallacootra Shear Zone is the Coorabie Shear Zone, separating the Fowler and Christie Domains from the Wilgena and Nuyts Domains. The Wilgena and Nuyts Domains are separated by the east-west trending Yerda Shear Zone, which is truncated in the west by the Coorabie Shear Zone (Swain et al., 2005b; Fraser and Lyons, 2006), and the Harris Greenstone Domain. The Yarlbrinda Shear Zone has been interpreted as a crustal boundary between the Nuyts Domain in the west and the Gawler Range Volcanics Domain to the east (Chalmers et al., 2007). The Kalinjala Mylonite Zone, a >250 km long north- to northeast-trending high-grade dextral shear zone, forms the major crustal-scale structure in the eastern Gawler Craton (Hand, 2005; Vassallo and Wilson, 2002; Howard et al., 2006). Few direct constraints on the timing of shear zone development and their subsequent evolution exist (e.g., Teasdale, 1997; Ferris et al., 2002). Limited data indicate that the deformational history of these shear zones does not reflect a single period of terrane assembly or reworking, but a complex pattern of strain partitioning during tectonic events in the late Paleoproterozoic to early Mesoproterozoic (Swain et al., 2005b). An overview of these constraints is presented below.

Karari Shear Zone

The Karari Shear Zone forms a significant, northeast-trending, linear magnetic high which separates the Nawa Domain to the northwest from the Christie Domain to the southeast (Rankin et al., 1989; Figure 12). It bifurcates around the Coober Pedy Ridge in the northeast where it anastomoses with the Tallacootra Shear Zone (Teasdale, 1997). The southern parts of the Karari Shear Zone are interpreted to have undergone northwest-over-southeast, vertical displacement of at least 18-22 km (possibly up to 35 km), consistent with the contrasting metamorphic grades between the southern Nawa and Christie Domains (Teasdale, 1997). The trace of the Karari Shear Zone is identified from magnetic images, as no exposure has been found. Information about the composition, structural evolution and metamorphic grade of lithologies are largely derived from the Ooldea DDH3 drillhole (Teasdale, 1997), which intersects the Karari Shear Zone near its southern end.

Where intersected by DDH Ooldea 3, the Karari Shear Zone forms a 280 m thick zone of mylonitic gneisses (Rankin et al., 1989; Teasdale, 1997). The major rock type is quartz-magnetite- feldspar-biotite±garnet±sillimanite gneiss, with a variably developed mylonitic fabric. Sillimanite is aligned parallel to the mylonitic fabric suggesting mid- to upper amphibolite facies conditions during at least some of the deformation (Fraser and Reid, 2007).

Age constraints from the Karari Shear Zone come from two studies which used different methods and produced different results. Swain et al. (2005b) reported an EMPA monazite age of 1631 ± 12 Ma (monazite in mylonitic matrix of gneiss) from the Karari Shear Zone, and interpreted this age as the timing of peak sinistral strike-slip deformation along the shear zone. Fraser and Lyons (2006) reported muscovite and biotite $^{40}Ar/^{39}Ar$ ages from the mylonitic fabric of ~1440 Ma and interpreted this age as a period of deformation and mylonitic fabric development at low- to medium-metamorphic grade. Fraser and Lyons (2006) noted that the ~1630 Ma monazite age of Swain et al. (2005b) is within uncertainty of metamorphic zircon ages from the Moondrah Gneiss to the west of the Karari Shear Zone, and thus suggested that this monazite age reflects high-grade metamorphism prior to movement on the Karari Shear Zone. The alternative interpretation of Swain et al. (2005b), that the monazite reflects deformation along the Karari Shear Zone closely following peak metamorphism of the Moondrah Gneiss, is also possible and implies that there could have been multiple periods of movement along the Karari Shear Zone.

Fraser and Lyons (2006) recognised that the earliest event common to domains on either side of the Karari Shear Zone is the high grade metamorphism recorded in the Mabel Creek Ridge and Coober Pedy Ridge of the Nawa Domain at ~1550 Ma, which was synchronous with high grade metamorphism in parts of the Fowler Domain and relatively low temperature isotopic resetting in the Christie Domain. This may have been the time at which relatively deep crustal rocks of the Nawa Domain were exhumed along the Karari Shear Zone to shallower crustal levels against the Christie Domain (Fraser and Lyons, 2006). This is consistent with the model of Direen et al. (2005), who envisage collision of the Nawa Domain with the Gawler Craton between ~1600 Ma and 1550 Ma.

Tallacootra Shear Zone

The Tallacootra Shear Zone forms one of the most significant geophysical boundaries in the western Gawler Craton (Figure 12). It can be traced for over 700 km as a set of major, ~northeast-striking, anastomosing, ductile structures extending from the Peake and Denison Domain in the far northeastern Gawler Craton southwestward beyond the current coastline where it is truncated by the Karari Shear Zone (Teasdale, 1997; Stewart et al., 2009). The geometry of the Tallacootra Shear Zone has been determined from its interpreted geophysical expression (Teasdale, 1997) and through structural analysis of sparse outcrops at Lake Tallacootra (Teasdale, 1997; Swain et al., 2005b). Cross-cutting relationships, which are apparent in magnetic images, suggest the Tallacootra Shear Zone (and the Coorabie Shear Zone) predates final movement on the Karari Shear Zone to the west (Teasdale, 1997).

A polyphase evolution has been interpreted for the Tallacootra Shear Zone based on geochronological data (Teasdale, 1997; Swain et al., 2005b; Fraser and Lyons, 2006). Age constraints from within the Tallacootra Shear Zone come from both EMPA analyses of monazite (Swain et al., 2005b) and $^{40}Ar/^{39}Ar$ analyses of mica and K-feldspar (Fraser and Lyons, 2006). These two methods have yielded contrasting results. Swain et al. (2005b) report essentially identical results from a sample from within the main trace of the Tallacootra Shear Zone at Lake Tallacootra, and from a migmatite sample to the west, at Lake Ifould. Chemical dating of zoned monazites from Archean mafic rocks within the Tallacootra Shear Zone yielded bimodal age distributions of ca. 2340 Ma and ca. 1680 Ma (Swain et al., 2005b). The ca. 2340 Ma core ages are interpreted to represent metamorphism during the Sleafordian Orogeny, whereas the ca. 1680 Ma rim ages are interpreted to record deformation along the Tallacootra Shear Zone late in the Kimban Orogeny at relatively high-grade metamorphic conditions (Swain et al., 2005b).

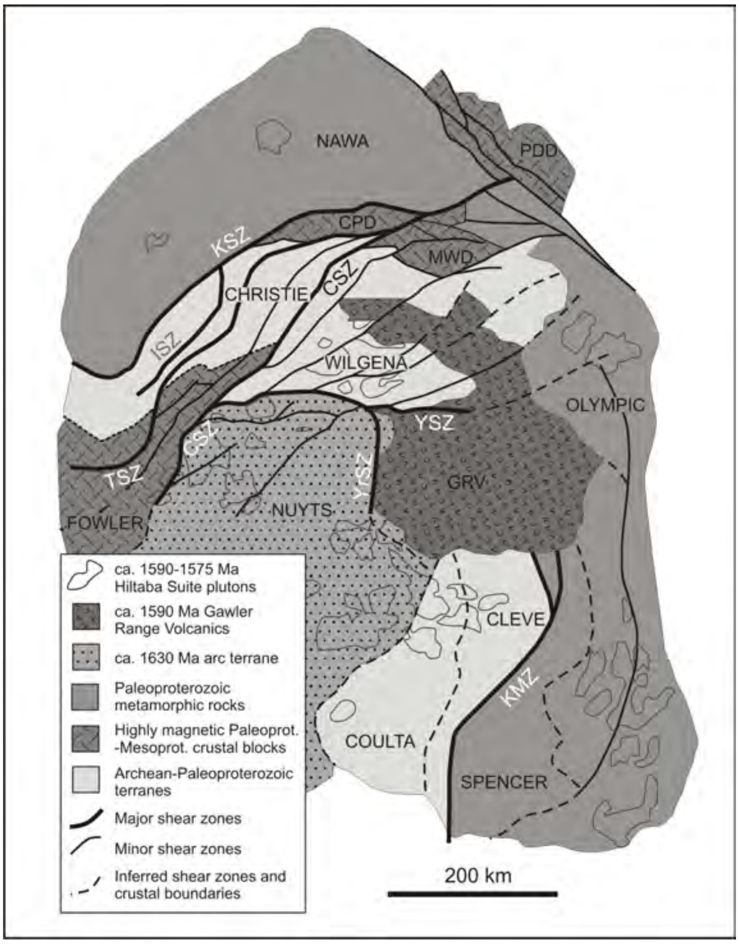

Figure 12. Interpreted map of Gawler crustal domains (modified after Daly et al., 1998; Ferris et al., 2002; Stewart et al., 2009). Main structures labelled from west to east:
KSZ – Karari Shear Zone;
ISZ – Ifould Shear Zone;
TSZ – Tallacootra Shear Zone;
CSZ – Coorabie Shear Zone;
YrSZ – Yarlbrinda Shear Zone;
YSZ – Yerda Shear Zone;
KMZ – Kalinjala Mylonite Zone;
GRV – Gawler Range Volcanics;
CPD – Coober Pedy Domain;
PDD – Peake and Denison Domain;
MWD – Mount Woods Domain.

Fraser and Lyons (2006) report $^{40}Ar/^{39}Ar$ mica ages from the mylonitic fabric at Lake Tallacootra of ~1440 Ma and suggest that these ages represent a period of deformation and hydrous fabric development. The differing age results from monazite and mica from the Tallacootra Shear Zone raise the question of whether this structure experienced multiple episodes of movement and, if so, which of these episodes were more important geologically in juxtaposing crustal blocks into their current relative crustal levels and along-strike positions (Fraser and Reid, 2007). Fraser and Lyons (2006) note that the ca. 1680 Ma monazite ages are within uncertainty of ~1680 Ma granites intruding the Christie Domain (Tunkillia Suite) and ~1720 Ma metamorphic monazites reported by Fanning (2002) from the Christie Gneiss at the Challenger Mine. On this basis, Fraser and Lyons (2006) raise the possibility that the ~1680 Ma monazite ages reported by Swain et al. (2005b) may represent metamorphic monazite that predates the development of the Tallacootra Shear Zone. Development of the Tallacootra Shear Zone after 1680 Ma is consistent with the evidence of high-grade metamorphism at ~1550 Ma within the

Fowler Domain, which is dissected by the Tallacootra Shear Zone. It is apparent that the Tallacootra Shear Zone must have been active post- ~1550 Ma, probably as late as ~1450 Ma, but periods of movement as old as ca. 1680 Ma are also possible (Fraser and Reid, 2007). Stewart et al. (2009) suggest that reactivation of the Tallacootra Shear Zone by dextral transpression during the ca. 1470-1450 Ma Coorabie Orogeny, exhumed the lower crust of the northwestern Fowler Domain.

Coorabie Shear Zone

The Coorabie Shear Zone is one of the longest and most prominent of a series of anastomosing northeast-striking structures visible in magnetic images of the western Gawler Craton. It forms the easternmost of a system of north-east-trending shear zones in the western Gawler Craton, and defines the boundary between the Fowler-Nuyts Domains and the Christie-Wilgena Domains, and can be traced for >300 km into the northeast Gawler Craton (Teasdale, 1997; Figure 12). To the south, the shear zone can be traced into the current continental shelf, suggesting it continues on the northern coast of Terra Adélie Land in Antarctica (Peucat et al., 1999, 2002; Teasdale et al., 2003). In the southwest part of the craton, the shear zone is defined by a set of anastomosing terranes with a total width of ~10 km (Swain et al., 2005b). Movement on the Coorabie Shear Zone exhumed lower-crustal (8-10 kbar) granulites of the Nundroo Block (Teasdale, 1997), and juxtaposed them with mid-crustal rocks belonging to the Nuyts Domain (Daly et al., 1998), implying at least 10-15 km of differential uplift (Teasdale, 1997). Cross-cutting relationships apparent in magnetic images suggest that the Coorabie Shear Zone (and Tallacootra Shear Zone) predate final movement on the Karari Shear Zone to the west (Teasdale, 1997).

Age constraints from within the Coorabie Shear Zone come from both EMPA analyses of monazite (Swain et al., 2005b) and $^{40}Ar/^{39}Ar$ analyses of mica and K-feldspar (Fraser and Lyons, 2006). Rocks from the Coorabie Shear Zone yield a range of radiometric ages from ~1530 Ma to ~1460 Ma. Samples from near the southern end of the Coorabie Shear Zone, from DDH Nundroo 5, and from outcrop at Cape Adieu, each yield two distinct monazite EMPA ages. Monazite inclusions in garnet yield ages of 1516 ± 18 Ma and 1508 ± 14 Ma, whereas monazite from within the mylonitic fabric yield ages of 1468 ± 12 Ma and 1457 ± 22 Ma (Swain et al., 2005b). In contrast, mica from an outcrop of protomylonitic granite within the northern part of the Coorabie Shear Zone yielded $^{40}Ar/^{39}Ar$ ages of ~1540 and ~1520 Ma (Fraser and Lyons, 2006). The ~1460 Ma monazite ages were interpreted by Swain et al. (2005b) to be the time of deformation, consistent with the ~1490 Ma U-Pb zircon age from a pegmatite from drillhole Nundroo 2, which itself is cut by narrow brittle-ductile shear zones (Teasdale, 1997; Fanning et al., 2007). Given the evidence for ~1460 Ma deformation in the southern part of the Coorabie Shear Zone, the significance of the older (~1530 Ma) $^{40}Ar/^{39}Ar$ mica ages from the northern part of the Coorabie Shear Zone is uncertain. The deformed granite sampled by Fraser and Lyons (2006) may not be representative of the main trace of the Coorabie Shear Zone, and the $^{40}Ar/^{39}Ar$ ages may simply reflect cooling ages (Fraser and Reid, 2007). This would be consistent with similar cooling ages of ~1530 Ma from K-feldspar from the northern Christie Domain at the Challenger mine, which have been tentatively linked to high-grade metamorphic events at this time in the Coober Pedy, Mabel Creek and southern Fowler Domains. Alternatively, the $^{40}Ar/^{39}Ar$ ages from the northern Coorabie Shear Zone may be minimum ages for deformation at this locality, implying multiple periods of movement along the Coorabie Shear Zone (Fraser and Reid, 2007).

Yerda Shear Zone

The Yerda Shear Zone separates rocks of the late Archean to early Paleoproterozoic Mulgathing Complex of the Wilgena Domain from the ca. 1690-1680 Ma Tunkillia Suite and ca. 1630-1610 Ma St Peter Suite of the Nuyts Domain in the central Gawler Craton, and forms the northern margin of a major (~250 km long) east-west trending corridor of deformation interpreted as a major zone of synmagmatic strike-slip movement which facilitated ascent of Hiltaba-age granites (Swain et al., 2005b). The shear zone is characterised by foliated granodiorite with a U-Pb zircon age of 1592 ± 11 Ma (Hiltaba Suite; Ferris, 2001), indicating that development of foliation postdated intrusion of at least some Hiltaba Suite granites. Interpretation of regional magnetics suggests that movement along the Yerda Shear Zone postdates deformation in the Yarlbrinda Shear Zone (Fraser et al., 2007a).

Fraser and Lyons (2006) interpret a biotite plateau-age of 1586 ± 11 Ma, from a sample from the Yerda Shear Zone, to reflect initial argon isotopic closure closely following granite intrusion, consistent with previous interpretations that movement on the Yerda Shear Zone provided an emplacement mechanism for some Hiltaba Suite granites (McLean and Betts, 2003). Swain et al. (2005b), however, interpret a monazite age of 1503 ± 16 Ma as the time of shearing along the Yerda Shear Zone, an age considerably younger than the $^{40}Ar/^{39}Ar$ biotite age of 1586 ± 11 Ma from the same structure (Fraser and Lyons, 2006).

Yarlbrinda Shear Zone

In the central Gawler Craton, the Yarlbrinda Shear Zone is located at the boundary between the Nuyts and Gawler Ranges Volcanics Domains. This structure strikes north-south, and is approximately 150 km long and 12 km wide (Ferris, 2001; Chalmers et al., 2007; Stewart and Betts, 2010; Figure 12). The shear zone both truncates, and is stitched by, granites belonging to the Hiltaba Suite (Ferris, 2001). Regional magnetics indicate that, toward its southern end, the Yarlbrinda Shear Zone is cut by the Kondoolka batholith, which yielded a U-Pb SHRIMP zircon age of 1580 ± 7 Ma (Ferris, 2001), indicating that pervasive deformation within the Yarlbrinda Shear Zone had ceased by this time. At its northern end, the Yarlbrinda Shear Zone swings to the west and joins the east-west-trending Yerda Shear Zone (Ferris, 2001; McLean and Betts, 2003; Fraser et al., 2007b). Granites of the ca. 1690-1680 Ma Tunkillia Suite are deformed by the Yarlbrinda Shear Zone; both dextral strike-slip and west-up dip-slip kinematics have been observed within outcrops of Tunkillia Suite within the Yarlbrinda Shear Zone (Ferris, 2001; Chalmers et al., 2007). Ferris (2001) interpreted west-up movements to overprint a phase of dextral strike-slip which, in turn, are overprinted by late-stage northeast-striking brittle structures observed at the regional scale. Stewart and Betts (2010) suggest dextral strike slip along the Yarlbrinda Shear Zone affected both the Tunkillia Suite and early phases of the St Peter Suite adjacent to the Paleoproterozoic southwestern margin (in present coordinates) of the Gawler Craton. In addition, the absence of dextral strike-slip within late St Peter Suite plutons (e.g., ca. 1611 Ma Yarlbrinda Hill pluton), constrains the upper limit timing of this dextral shearing event.

Ferris and Schwarz (2003) used Sm-Nd data compiled for granites of the Hiltaba Suite in the Gawler Craton by Stewart et al. (1999) to suggest that there is a significant change in composition from west to east, which broadly coincides with the Yarlbrinda Shear Zone. These data indicate that the Yarlbrinda Shear Zone marks the eastern boundary of a domain dominated by more primitive εNd values, most likely related to the presence of significant volumes of juvenile St Peter Suite I-type granites. Older granites of the Tunkillia Suite occur along parts of this boundary zone, although their distribution is not restricted to this setting. More negative εNd values to the east of the Yarlbrinda Shear Zone suggest a greater influence of older basement as a source for Hiltaba Suite magmatism (Fraser et al., 2007a).

Kalinjala Mylonite Zone

The Kalinjala Mylonite Zone is a >250 km long north- to northeast-striking high-grade, dextral transpressional shear zone that formed during the ca. 1740-1690 Ma Kimban Orogeny (Hand et al., 1995; Vassallo and Wilson, 2002; Figure 12). The southern extension of the shear zone can be traced geophysically into the current continental shelf, suggesting continuation on the northern coast of Terra Adélie Land in Antarctica (e.g., Peucat et al., 1999, 2002; Teasdale et al., 2003). At its northern end, it is unconformably overlain by ca. 1590 Ma Gawler Range Volcanics.

The mylonite zone is a steeply-dipping structure, up to 4 km wide, which separates two distinct geological regions. To the west of the shear zone, the oldest rocks consist of the late Archean metasedimentary and volcanic successions of the Sleaford Complex, the Dutton Suite, the Miltalie

Gneiss and metasedimentary rocks of the Hutchison Group (Daly et al., 1988; Swain et al., 2005a). To the east of the Kalinjala Mylonite Zone, the geology is dominated by the 1850 Ma Donington Suite (Hoek and Schaefer, 1998). The Kalinjala Mylonite Zone is inferred to have accommodated significant displacement during dextral transpression (Vassallo and Wilson, 2002). The core of the mylonite zone contains lower-crustal granulites, which form a narrow zone of deeply exhumed rocks, flanked by mid- to upper-crustal rocks (Hand et al., 1995), which contain comparatively well-preserved pre-Kimban Orogeny records.

Monazite inclusions in garnet porphyroclasts within the shear fabric have been dated by EMPA at 1693 ± 16 Ma, giving the age of the metamorphic mineral assemblage, and providing a maximum age constraint on shearing (Swain et al., 2005b). Monazite within upper amphibolite microshears between two plagioclase porphyroclasts is dated at 1682 ± 10 Ma, interpreted to record the age of shearing (Swain et al., 2005b). Swain et al. (2005b) suggests that the Kalinjala Mylonite Zone was the focus of a comparatively narrow, high-strain orogenic belt which reworked older crust. Dutch and Hand (2009) present EMPA monazite ages of ~1690 Ma and ~1710 Ma from Sleaford Complex units along the western flank of the Kalinjala Mylonite Zone, which they interpreted to be the age of fabric formation during the development of the shear zone in the Kimban Orogeny.

Dextral transpression recorded by the Kalinjala Mylonite Zone at ca. 1690-1680 Ma, occurred during the latter stages of the Kimban Orogeny, synchronous with the emplacement age of granites belonging to the Tunkillia Suite. In reconstruction models, Betts and Giles (2006) have suggested that the Kalinjala Mylonite Zone represents a suture along which two different continental blocks were joined at ~1740-1690 Ma (refer to Section 2).

On northern Eyre Peninsula, the Kalinjala Mylonite Zone is interpreted to be east-dipping, based on its magnetic signature and reflection-seismic interpretation (Fraser et al., 2010b). In this region, the Kalinjala Mylonite Zone separates Kimban-aged gneisses to the west from relatively low-metamorphic grade, pre-Kimban sedimentary rocks (Broadview Schist, Moonabie Formation), to the east (Fraser et al., 2010b).

SECTION 2. REGIONAL OVERVIEW OF THE TECTONIC DEVELOPMENT OF THE GAWLER CRATON AND CURNAMONA PROVINCE

N. Kositcin

2.1. Introduction

The Gawler Craton and Curnamona Province are part of a complex collage of Archean to Mesoproterozoic metasedimentary and metaigneous terranes in southern Australia forming the South Australian Craton (SAC). The South Australian Craton forms part of the Mawson Continent, which includes the correlative coastal outcrops of Terre Adélie and George V Land in Antarctica, and various other terrains of East Antarctica (Peucat et al., 1999; Fitzsimons, 2003; Payne et al., 2009). Limited outcrop, due to the widespread presence of younger cover which obscures much of the craton, has hampered the development of tectonic models which describe the time-space evolution of the South Australian Craton and, as such, difficulties exist in placing the South Australian Craton into the context of Australian and global Proterozoic continental evolution. As highlighted by Hand et al. (2007), significant ambiguities remain in terms of the timing and spatial distribution of the tectonic events within the craton, the tectonic settings of the major magmatic systems and the crust-mantle evolution of the craton through time.

2.2. Intraplate versus convergent margin plate tectonics

Various interpretations exist for the evolution of the Proterozoic Australian continent, which embrace plate tectonic processes to varying degrees. Etheridge et al. (1987) developed an intraplate tectonic model, which emphasised the importance of vertical accretion (widespread mafic underplating) and reworking (partial melting of the lower crust) in an intracontinental setting (Etheridge et al., 1987; Loosveld and Etheridge, 1990; Oliver et al., 1991). This intraplate model highlights the fact that many Proterozoic Australian terranes lack blueschist or other high-pressure metamorphic rocks, accretionary wedges or mélange zones, or belts of arc-type magmatic rocks and ophiolite complexes, which are typical of Phanerozoic-style convergent margin tectonics. Driving mechanisms of the intraplate model include: interaction between the continental lithosphere and mantle plumes, delamination of the subcontinental lithospheric mantle, and vigorous, small-scale mantle convection beneath a large, insulating continental mass (Etheridge et al., 1987; Loosveld and Etheridge, 1990; Oliver et al., 1991).

The model of Etheridge et al. (1987) differs significantly from convergent, plate-margin tectonic models for Proterozoic Australia. For example, Myers et al. (1996) interpreted the same geological record as a consequence of modern-style plate tectonics, in which various continental fragments were amalgamated via oceanic subduction followed by continental collision. A feature of the model of Myers et al. (1996) is the subdivision of Precambrian Australia into three large cratonic elements: the West Australian Craton (WAC), the North Australian Craton (NAC) and the South Australian Craton (SAC) (Figure 13). Myers et al. (1996) envisaged that these three cratonic elements were assembled separately by ~1830 Ma, and then amalgamated into their present relative positions during the interval 1300-1000 Ma. Following Myers et al. (1996), the past decade has seen the development of numerous tectonic reconstructions for Proterozoic Australia involving accretionary and subduction tectonics, with different sequences of events used to explain final amalgamation of the NAC, WAC and SAC, and differences in the proposed location, timing and polarity of convergent margins (e.g., Scott et al., 2000; Sheppard et al., 2001; Giles et al., 2002; Betts et al., 2002; Giles et al., 2004; Betts and Giles, 2006; Wade et al., 2006; Betts et al., 2007, 2008, 2009; Cawood and Korsch, 2008; Conor and Preiss, 2008; Gibson et al., 2008; Hand et al., 2008; Payne et al., 2009). Gibson et al. (2008), based on studies in Mount Isa, have proposed a Basin-and-Range-style model, in which crustal evolution was dominated by extension. In this interpretation, the relative positions of the North Australian and South Australian Cratons have not changed significantly through the Paleo- to Mesoproterozoic.

The relationship between the South Australian Craton (SAC) and the North Australian Craton (NAC) is poorly understood, but is crucial in understanding the assembly of Proterozoic Australia (Figure 13). As such, reconstruction models for Proterozoic Australia commonly focus on the interaction of the Gawler Craton and the Curnamona Province with the North Australian Craton (as defined by Myers et al., 1996) and the Musgrave Province (e.g., Myers et al., 1996; Karlstrom et al., 2001; Betts et al., 2002; Giles et al., 2002, 2004; Betts and Giles, 2006, Wade et al., 2006).

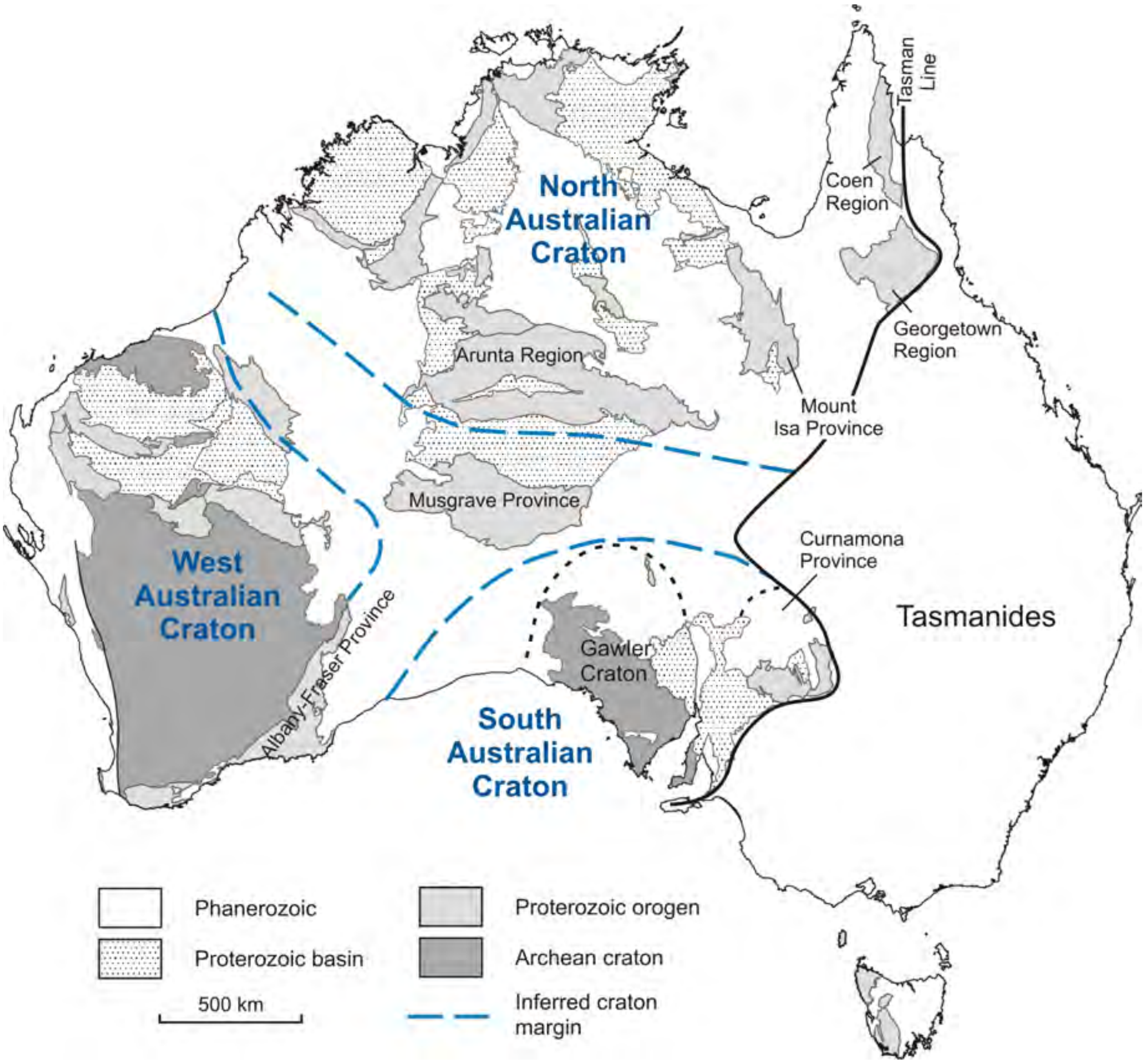

Figure 13. Map of Australia displaying the locations of Archean cratons and Paleoproterozoic-Mesoproterozoic terrains (modified after Cawood and Korsch, 2008).

2.3. Summary of tectonic models

A significant number of tectonic reconstruction models, which address aspects of the evolution of the Gawler Craton and Curnamona Province have been proposed in recent years (e.g., Daly et al., 1998; Dawson et al., 2002; Giles et al., 2002, 2004; Swain et al., 2005a, 2008; Betts and Giles, 2006; Wade et al., 2006; Betts et al., 2007, 2008; Cawood and Korsch, 2008; Conor and Preiss, 2008; Gibson et al., 2008; Hand et al., 2008; Betts et al., 2009; Payne et al., 2009). Most of these models focus on the Paleo- to Mesoproterozoic time period, with only a few models (e.g., Swain et al., 2005a; Payne et al., 2009) addressing possible tectonic scenarios in the Archean to earliest Paleoproterozoic interval.

The tectonic models for the latest Paleoproterozoic to earliest Mesoproterozoic evolution of eastern Australia are diverse, and either emphasise plume activity or plate margin activity. For the 1620-1500 Ma time period, Betts et al. (2009) divide the range of proposed tectonic models into two end-members: 1. plume-driven models, developed to account for widespread voluminous felsic A-type magmatism in the Gawler Craton, Curnamona Province and Mount Isa Province (e.g., Giles, 1988; Stewart and Foden, 2003; Betts et al., 2007), and, 2. plate margin models, in which the tectonic evolution of eastern Australia was driven by processes operating at one or more convergent plate margins (e.g., Betts et al., 2002; Giles et al., 2004; Betts and Giles, 2006; Wade et al., 2006; Cawood and Korsch, 2008; Swain et al., 2008). A brief overview of the current tectonic models, which deal with aspects of the Late Archean to Early Mesoproterozoic evolution of the Gawler Craton and Curnamona Province, is presented below.

2.3.1. Gawler Craton

Cooyerdoo Granite (ca. 3150 Ma)

The first recognised event in the Gawler Craton is intrusion of the ~3150 Ma Cooyerdoo Granite into unknown older crust, which is inferred to form basement to the Middleback Ranges iron-rich sedimentary rocks (Fraser et al., 2008, 2010a). Fraser et al. (2010a) suggest that the geochemistry of the Cooyerdoo Granite, particularly its elevated LILE and LREE content and low Na/K, is similar to the post-tectonic, potassic granites in the Yilgarn and Pilbara Cratons. The Pilbara and Yilgarn granites have been interpreted to be the products of melting of pre-existing Tonalite-Trondhjemite-Granodiorite (TTG) crust (Champion and Smithies, 2007; Van Kranendonk et al., 2007), which Fraser et al. (2010a) also infer for the origin of the Cooyerdoo Granite. The presence of inherited zircons as old as ~3300 Ma implies the presence of Paleoarchean source regions, possibly still present at depth (Fraser et al., 2010a). Granite chemistry, zircon ages and Nd-depleted mantle model ages (~3300-3150 Ma) are interpreted to indicate the existence of ~3400 to 3200 Ma TTG-like crust within at least parts of the eastern Gawler Craton (Fraser et al., 2010a).

Fraser et al. (2010a) note that this ancient crust adjacent to the eastern margin of the Gawler Craton may predict the presence of Mesoarchean crust beneath the NAC, following the model of Betts and Giles (2006), which envisages the eastern Gawler Craton as part of the NAC in the Paleoproterozoic (see below). Fraser et al. (2010a) also note that close temporal correlations can be made between the ~3150 Ma rocks in the eastern Gawler Craton and Archean crustal fragments in Antarctica (e.g., Miller Range, Prince Charles Mountains), based on the work of Goodge and Fanning (1999), Goodge et al. (2001), Boger et al. (2001) and Boger et al. (2006), and suggest that early Mesoarchean crust may occur more widely within the East Antarctic and Gawler Cratons (Mawson Continent).

Sleaford and Mulgathing Complexes and Sleafordian Orogeny (ca. <2560-2400 Ma)

The second recognised tectonic cycle on the Gawler Craton is the Late Archean magmatism and sedimentation of the Dutton Suite, Hall Bay Volcanics and Devils Playground Volcanics (ca. 2560 - 2520 Ma), and the Sleafordian Orogeny at ca. 2460-2430 Ma (Swain et al., 2005a). Swain et al. (2005a) suggest that the Late Archean lithologies of the Gawler Craton are consistent with a cycle of continental crustal growth between ca. 2560 Ma and 2500 Ma, interpreted to reflect an evolving arc environment analogous to modern plate tectonic-like geodynamic processes. These authors envisage a tectonothermal regime for development of the Late Archean Gawler Craton which reflects interaction between a convergent plate margin and mantle plume, in which sedimentary deposition occurred during active basin development in a backarc or arc-rift setting (Figure 14).

In this model, Swain et al. (2005a) use U-Pb zircon data, geochemical and Sm-Nd isotopic data from the Christie Gneiss and Kenella Gneiss of the Mulgathing Complex and the Carnot Gneisses and Wangary Gneiss of the Sleaford Complex to suggest that these rocks probably all formed in the same depositional environment. Arc-style volcanism at ca. 2650-2520 Ma, associated with the Devils Playground Volcanics and Hall Bay Volcanics, was probably contemporaneous with deposition of the protoliths of the Christie and Carnot gneisses, and the Kenella and Wangary gneisses in an extensional backarc setting. Arc-style magmatism continued with ca. 2520-2500 Ma magmatism of the Coulta Granodiorite and Glenloth Granite. Extension and sedimentation in the backarc continued, facilitating the extrusion of the ca. 2520-2500 Ma komatiites from an ascending mantle plume. Basin formation and associated magmatism was terminated by regional, granulite-grade metamorphism and crustal thickening during the Sleafordian Orogeny. The shift from extension to contraction in the backarc may have occurred through shallowing of the subduction angle and/or the incorporation of buoyant material in the subduction zone, or may have occurred due to closure of the ocean and the arrival of a continent or continental fragment. As some 400 Ma of tectonic quiescence followed the Sleafordian Orogeny, cratonisation of the Gawler Craton at this time may have resulted from the incorporation of a colliding continental terrane (Swain et al., 2005a; Figure 14).

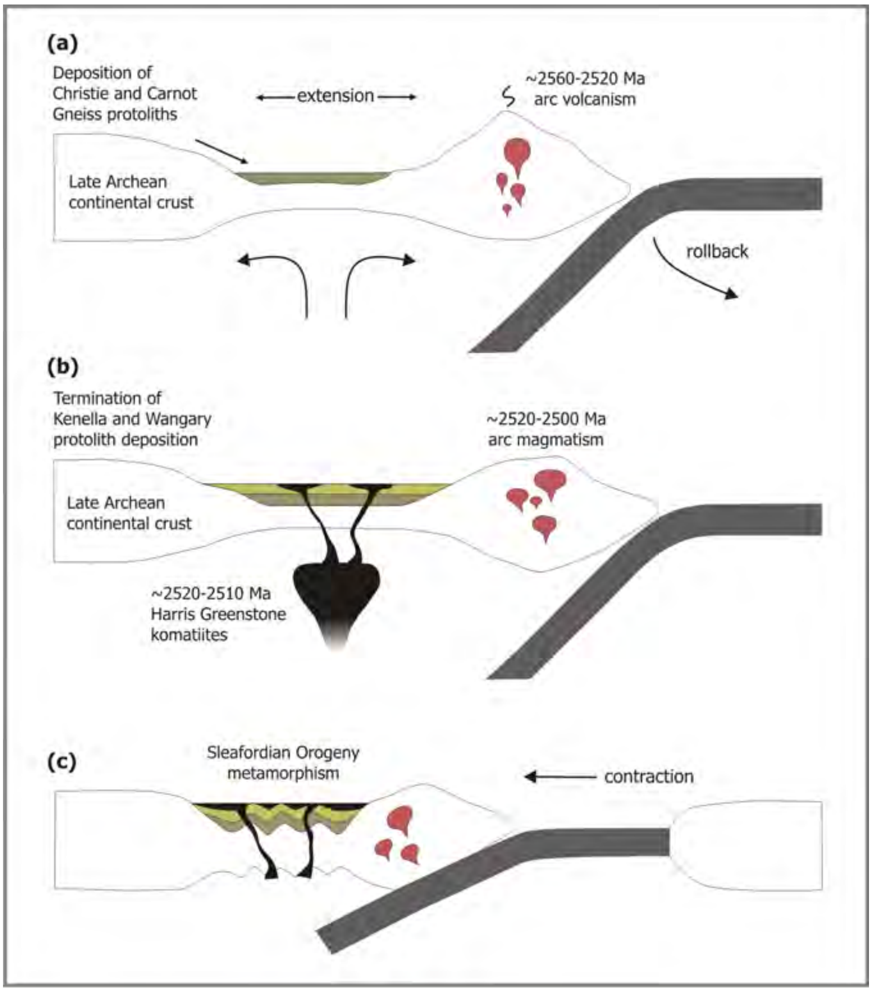

Figure 14. Tectonic cartoon illustrating the proposed crustal development of the late Archean Gawler Craton, modified after Swain et al. (2005a). (a) ca. 2560-2520 Ma arc volcanism associated with the Devils Playground Volcanics and the Hall Bay Volcanics contemporaneous with deposition of protoliths to the Christie and Carnot Gneisses, and the Kenella and Wangary Gneisses to the west in a backarc setting. (b) ca. 2520-2500 Ma arc magmatism of the Coulta Granodiorite and Glenloth Granite, and extrusion of Harris Greenstone Domain komatiites through mantle plume activity. (c) Cratonisation during the Sleafordian Orogeny.

Miltalie Gneiss (ca. 2000 Ma)

Intrusion of protoliths to the Miltalie Gneiss represents the first recognised event in the Gawler Province after 400 Ma of inactivity. There are currently no constraints on the tectonic setting of the Miltalie Gneiss and its protoliths. Daly et al. (1998) suggested the Miltalie Gneiss was emplaced during a period of extensional tectonism, and that this period of extension gave rise to a shallow continental shelf (passive margin) along the eastern margin of the Archean nucleus (Sleaford and Mulgathing complexes). Geochemically, the Miltalie Gneiss falls into the within plate field on Pearce et al. (1984) tectonic discrimination diagrams, consistent with a within plate extensional environment.

At ~2000 Ma, along what is now its eastern margin, the Gawler Craton probably underwent substantial extension to form a major elongate basin into which a mixed shallow-water clastic and chemical metasedimentary succession was deposited (Hutchison Group; Parker et al., 1993; Daly et al., 1998). The Hutchison Group has been interpreted to have accumulated as a passive margin succession on a shallow shelf that deepened to the east (Parker and Lemon, 1982; Parker et al., 1993).

Donington Suite and Cornian Orogeny (ca. 1860-1850 Ma)

Reid et al. (2008b) consider the early Paleoproterozoic tectonic evolution of the Gawler Craton, focussing specifically on the interval that includes the intrusion of Donington Suite magmas and the Cornian Orogeny at ~1850 Ma, and tectonic activity including volcano-sedimentary basin formation in the interval ca. 1810-1740 Ma, prior to the onset of the Kimban Orogeny.

The ca. 1860 Ma evolution of the southeastern Gawler Craton was dominated by the emplacement of granites of the Donington Suite, synchronous with high-grade contractional deformation (Cornian Orogeny of short duration, that is, 10-20 million years, which interrupted rifting in the eastern Gawler Craton), which was terminated by high temperature decompression and extensional deformation (Reid et al., 2008b). On the basis of geochemical characteristics, and the lack of intermediate calcalkaline magmatism, Mortimer et al. (1988) concluded that the Donington Suite was not related to convergent margin processes, and Parker et al. (1993) advocated a model of intracontinental magmatism for the tectonic evolution of the region at ca. 1860 Ma. Hoek and Schaefer (1998) noted the physical similarities that the Donington Suite shared with plate-margin magmas, namely the co-existence of mafic and felsic lavas, along with the geochemistry of crosscutting mafic dykes, which led to suggestions that the Donington Suite may have formed inboard of, or adjacent to, a subduction zone (Hoek and Schaefer, 1998). Schwarz et al. (2002) suggested that the Donington Suite was a collision or subduction-derived suite associated with northeast-directed subduction of Late Archean-Paleoproterozoic lithosphere beneath a younger continental region to the east and northeast. Ferris et al. (2002) suggested the Donington Suite either directly marks the existence of a magmatic arc, which developed along the margin of the Archean-Paleoproterozoic continent, or that it may have been produced in a backarc environment. Hand (2005) proposed that the suite formed during a collisional event along the eastern margin of the craton.

Reid et al. (2008b) suggest the Donington Suite was generated in a (far-field) continental backarc setting. In this scenario, mantle-derived melts undergoing both crustal assimilation and fractional crystallisation ascended into the crust through lithospheric extension and consequent decompression melting of the mantle in the continental backarc. In addition, the contractional deformation characterising the Cornian Orogeny was focussed into this proposed backarc due to thermal softening and lithospheric thinning. In their model, this contractional deformation (and the clockwise P-T path) may have been initiated as a result of a change in subduction dynamics, either a change in subduction angle or through the arrival of a buoyant collider. Reid et al. (2008b) do not, however, explicitly state the envisaged location and polarity of any ~1850 Ma subduction. Reid and Howard (2006) consider that, by 1843 ± 5 Ma, the contractional phase had ceased and extension had started. Convergence associated with the Cornian Orogeny was followed by a long-lived extensional and transtensional regime which produced regional scale mafic dyke swarms and localised subbasins (Schwarz et al., 2002). The emplacement of the Tournefort Metadolerite at ca. 1815 Ma (Schaefer, 1998), and the subsequent development of localised basins, indicates that this extension is likely to have continued periodically across the region for up to 80 million years over the interval ~1810-1730 Ma (Reid et al., 2008b). Indeed, several phases of sedimentation, together with pulses of bimodal volcanism, occurred along the eastern margin of the Gawler Craton in the interval ~1790-1740 Ma (e.g., 1791 ± 4 Ma Myola Volcanics, Fanning et al., 1988; ~1780 Ma Peake Metamorphics, Fanning et al., 1988; 1767 ± 16 Ma Price Metasediments, Oliver and Fanning, 1997; 1760-1740 Ma Wallaroo Group, Cowley et al., 2003; 1756 ± 8 Ma Moonabie Formation, Jagodzinski, 2005; 1754 ± 5 Ma, McGregor Volcanics, Fraser and Neumann, 2010). Zang (2002) suggested a rifted continental margin setting for sedimentation and volcanism in this interval. Reid et al. (2008b) proposed that the extension-dominated system, accounting for the ca. 1790-1740 Ma bimodal volcano-sedimentary successions in the Gawler Craton, may have developed as a series of slab rollback-initiated, extensional basins formed inboard of a far-field subduction front (continental backarc setting). The models of Giles et al. (2002) and Betts and Giles (2006) propose that the South Australian Craton was rotated and translated anticlockwise northeastwards such that the Wallaroo Group formed in a backarc setting relative to an easterly extension of a subduction zone along the southern margin of the North Australian Craton (see below). Extensional tectonics and major sedimentation shifted eastwards from the Gawler Craton to the Curnamona Province during the Kimban Orogeny, although minor localised syn-Kimban sedimentation continued in the Gawler Craton (e.g., ca. 1715 Ma Labyrinth Formation in the central Gawler Craton; Fanning et al., 2007).

Convergent margin tectonism (ca. 1800-1500 Ma), backarc basin sedimentation (ca. 1800-1600 Ma) and the Kimban - Strangways Orogeny (ca. 1740-1690 Ma)

Betts and Giles (2006) presented a model for the 1800-1000 Ma tectonic evolution of Proterozoic Australia, building upon concepts first published in Betts et al. (2002) and subsequently revised in Giles et al. (2002, 2004). A primary characteristic of each of these models is the presence of a long-lived accretionary margin on the southern margin of the North Australian Craton (Figure 15).

A diverse range of geological phenomena correlate between the component terranes of the North Australian Craton and South Australian Craton and are used by Giles et al. (2004) to suggest a shared tectonic evolution between approximately 1800 Ma and 1500 Ma. These correlations include:

- Synchronous deposition of the Willyama Supergroup (Curnamona Province) and Eastern Succession (Mount Isa Province) between ca. 1710-1600 Ma (Nutman and Gibson, 1998; Page and Sun, 1998; Page et al., 2000), including bimodal magmatism in both terranes at ca. 1710-1670 Ma (Nutman and Ehlers, 1998; Page et al., 2000).
- Low-pressure metamorphism to granulite facies in the Curnamona Province (Phillips and Wall, 1981), and upper amphibolite facies in the eastern Mount Isa Province (Rubenach and Barker, 1998) of the ca. 1600-1580 Ma Olarian and Isan orogenies, respectively.
- Similar chemistry of Tidnamurkuna Volcanics and Myola Volcanics of the Gawler Craton and volcanics in the Bottletree Formation, Argylla Formation and Eastern Creek Volcanics in the Mount Isa Province, all in the interval ca. 1790-1780 Ma (Wyborn et al., 1987).
- The Kararan Orogeny in the Gawler Craton (ca. 1580-1540 Ma) showing similar style and timing to phases of the Isan Orogeny in Mount Isa (Daly et al., 1998; MacCready et al., 1998).
- Similar timing of the Late Strangways Orogeny in the Arunta Province (ca. 1740-1690 Ma; Collins and Shaw, 1995; Claoue-Long et al., 2008) and the Kimban Orogeny in the Gawler Craton (ca. 1740-1690 Ma).
- Bimodal magmatism of the Gawler Range-Hiltaba Event in the Gawler Craton, and magmatism in the Curnamona Province between 1600 and 1580 Ma (Page et al., 2000), which was coincident with high-temperature, low-pressure metamorphism at mid-crustal levels in the

Arunta Province (Rubatto et al., 2001), the Curnamona Province (Page et al., 2000), and eastern Mount Isa Province (Page and Sun, 1998).

To account for these observations, in reconstruction models proposed by Betts et al. (2002) and Giles et al. (2002, 2004), the Kimban Orogeny is proposed to result from the collision between the Gawler Craton (as part of the Mawson Continent) and the North Australian Craton, between ca. 1740 Ma and 1690 Ma, during north-northwest-dipping subduction (Figure 15, 16). This was revised by Betts and Giles (2006) such that crust east of the Kalinjala Mylonite Zone in the Gawler Craton was originally part of the NAC and the proto-Gawler Craton was the colliding terrain during the Kimban-Strangways Orogeny. In these models, the Kimban Orogeny is interpreted to connect with the Strangways Orogeny in the Arunta Province to form an approximately east-west trending collisional belt. This interpretation requires the SAC to be rotated approximately ~52˚ counterclockwise about a pole located at ~136˚E and ~25˚S (present-day coordinates), relative to its current position. In this scenario, the Curnamona Province was directly adjacent with the current southern subsurface extent of the Mount Isa Province and Georgetown Region (Giles et al., 2004), linking the Eastern Succession and Willyama Supergroup as part of the same stratigraphic package, aligning the structural grains of the Mount Isa and Curnamona Provinces, and aligning ~1800-1690 Ma accretionary terranes of the Arunta Province and western Gawler Craton into a continuous orogenic belt along the southern margin of the Proterozoic continent. This belt was the product of protracted accretion and arc magmatism along a convergent margin that evolved from at least 1800 to 1600 Ma (Figure 15, 16; Betts et al., 2002; Giles et al., 2002).

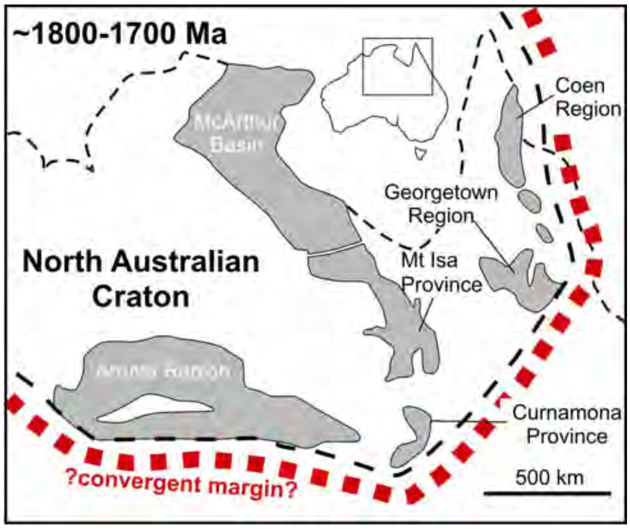

Figure 15. Proterozoic tectonic reconstruction highlighting the proximity of the Curnamona Province to the Mount Isa Province in northern Australia prior to the rotation of the South Australian Craton proposed by Giles et al. (2002). Both provinces are situated in the backarc to convergent margins at the southern and eastern margins of the NAC (after Giles et al., 2002).

Betts and Giles (2006) suggested the subduction zone dipped to the northeast and was located between the western Archean Gawler Craton (and Hutchison Group), as a continental fragment on the downgoing slab, and the Wallaroo Group in the eastern Gawler Province and the Curnamona Province on the upper plate to the east (Figure 16). By suggesting an active margin for the southern NAC, Giles et al. (2002, 2004) place the McArthur and Mount Isa basins into a farfield extensional backarc setting (Figure 15, 16). The 1800 to 1600 Ma volcano-sedimentary basins of the Mount Isa Province, Curnamona Province and northern Gawler Craton evolved in the continental interior of the overriding plate of this convergent margin (Giles et al., 2002), until they were deformed, metamorphosed and intruded by voluminous magmas during 1600-1500 Ma orogenesis.

Orogenesis at 1565-1500 Ma in the NAC and SAC is interpreted to be related to collision with Laurentia along the eastern margin of the NAC (Betts and Giles, 2006). The Gawler Craton is then interpreted to have subsequently (post 1500 Ma) rifted away from the NAC before again being accreted to the Australian continent in its present configuration during the Albany-Fraser and Musgravian orogenies (ca. 1330-1100 Ma, Betts et al., 2002; Giles et al., 2002) or, alternatively, rotated and translated during the same time period (Giles et al., 2004).

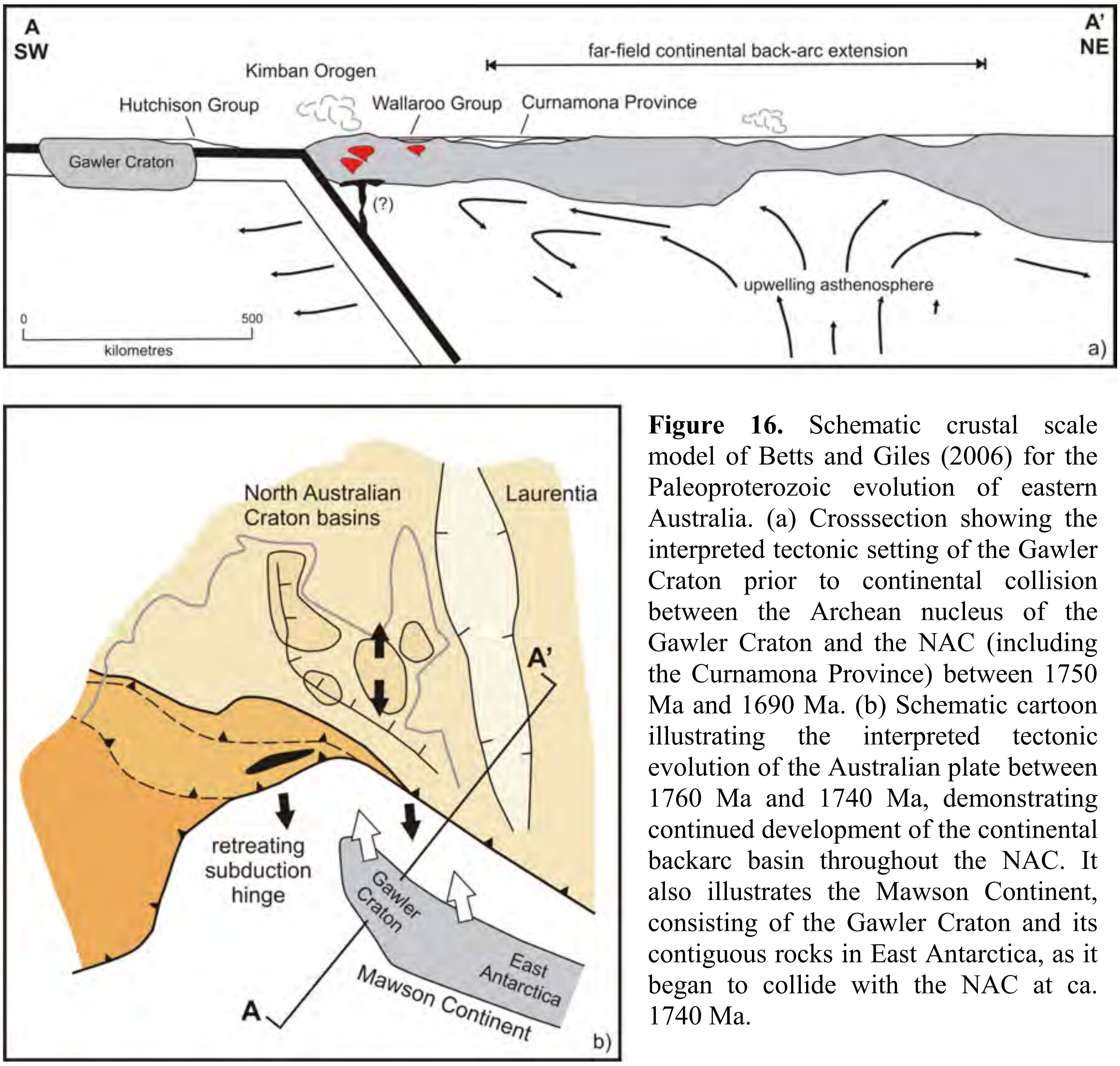

Figure 16. Schematic crustal scale model of Betts and Giles (2006) for the Paleoproterozoic evolution of eastern Australia. (a) Crosssection showing the interpreted tectonic setting of the Gawler Craton prior to continental collision between the Archean nucleus of the Gawler Craton and the NAC (including the Curnamona Province) between 1750 Ma and 1690 Ma. (b) Schematic cartoon illustrating the interpreted tectonic evolution of the Australian plate between 1760 Ma and 1740 Ma, demonstrating continued development of the continental backarc basin throughout the NAC. It also illustrates the Mawson Continent, consisting of the Gawler Craton and its contiguous rocks in East Antarctica, as it began to collide with the NAC at ca. 1740 Ma.

Cawood and Korsch (2008) suggest that the model of subduction of Betts and Giles (2006) does not easily account for Kimban deformation in the western domains of the Gawler Province (the rocks on the downgoing plate). Alternatively, Cawood and Korsch (2008) suggest that a subduction zone was located well to the southwest of the Archean rocks, with the Sleaford and Mulgathing complexes lying on the upper plate, and with orogenesis occurring in an accretionary, rather than collisional, tectonic setting.

The Kimban Orogeny was accompanied by a major pulse of igneous activity between ca. 1730 and 1670 Ma (~1737-1726 Ma Middlecamp Granite, 1720-1700 Ma Moody Suite, and 1690-1670 Ma Tunkillia Suite; Parker et al., 1993; Schwarz, 1999; Fanning et al., 2007). In the model of Betts and Giles (2006), following the Kimban Orogeny, the plate margin shifted to the southwestern edge of the Gawler Craton and is recorded by the Tunkillia Suite of supposed arc-like affinities. Ferris et al. (2002) suggested that the voluminous mafic to intermediate calcalkaline magmatic rocks of the Tunkillia Suite had a geochemical signature consistent with a magmatic arc setting. Alternatively, Ferris and Schwarz (2004) considered that the Tunkillia Suite was emplaced in a backarc environment behind an active magmatic arc. New age and geochemical data (Payne et al., 2010) do not support a subduction-related setting but rather a late- to post-tectonic setting, indicating that the Tunkillia Suite represents a continuation of the Kimban Orogeny.

Yilgarn-Gawler collision (ca. 1750 Ma)

Dawson et al. (2002) suggest that a sedimentary package in the southwest of Western Australia, the 1696 ± 7 Ma Mount Barren Group, is recycled from a foreland basin to the proposed Yilgarn-Gawler collision of Krapez (1999). The provenance of detrital zircon grains is used to negate a Yilgarn source, but also to infer a Gawler source for the detritus and, therefore, to suggest that the Yilgarn and Gawler Cratons were involved in orthogonal collision during the Late Paleoproterozoic assembly of proto-Australia. This collision is inferred to have occurred at ~1750 Ma and involved the Nawa Domain as the leading edge of the Yilgarn indenter (Dawson et al., 2002). Krapez (1999) and Krapez and Martin (1999) proposed pre-indentation (pre-1750 Ma) contiguity of the Pilbara and Gawler Cratons because of their similar rock records and tectonic histories for the ~2450-1680 Ma period.

Dawson et al. (2002) proposed that the onset of the Kimban Orogeny was synchronous with the Pilbara-Yilgarn collision and inferred Yilgarn-Gawler collision, thereby providing detritus for a pre-Mount Barren Group foreland basin on the Yilgarn Craton. There is no record of any orogenic or metamorphic event in the Gawler Craton (Daly et al., 1998; Ferris et al., 2002) at around 1750 Ma. The model of Dawson et al. (2002) would be more plausible if Yilgarn-Gawler collision occurred at ~1730-1700 Ma (Kimban Orogeny), rather than at ~1750 Ma.

Evolution of the Mawson Continent (ca. 2500-1500 Ma)

The model of Payne et al. (2009) highlights the presence of comparable sedimentary, magmatic, metamorphic and deformational event histories in the NAC and the Mawson Continent from at least the end of the Archean to the early Mesoproterozoic. Payne et al. (2009) suggest that the Gawler-Adélie Craton and the NAC formed a contiguous continental terrain during the entire Paleoproterozoic, with both the Gawler Craton and Curnamona Province sharing correlatable event histories with the NAC at ca. 2500-2430 Ma, 2000 Ma, 1865-1850 Ma, 1730-1690 Ma, and 1600-1550 Ma. As with the models of Betts et al. (2002), Giles et al. (2004) and Betts and Giles (2006), their model highlights the complexity and longevity of interaction between the NAC and SAC.

Some proposed correlations include:

- Temporal equivalents to ca. 2520 Ma Gawler magmatism in the Pine Creek and Tanami regions (e.g., see Lally, 2002; Cross et al., 2005; Crispe et al., 2007), and within basement inliers in the Mount Isa region (e.g., ~2500-2450 Ma Black Angel Gneiss of supposed subduction-related geochemical signature; McDonald et al., 1997), plus 2500-2450 Ma detrital ages of metasedimentary lithologies in the Arunta Region (Wade et al., 2008).
- Both the NAC and Gawler-Adélie Craton share a period of inactivity in the interval ca. 2440-2050 Ma, prior to the onset of sedimentation and magmatism (Daly et al., 1998; Fanning et al., 2007), possibly representing continental rifting and breakup.
- The NAC records metamorphism and deformation similar in age to the Cornian Orogeny in the Gawler Craton (e.g., post-1850 Ma accretion of the Kimberley Craton to the NAC and collision of the Kimberley Craton plus Western Halls Creek Orogen with the Tickalara arc at ca. 1850-1845 Ma; Sheppard et al., 1999).
- Events in the Tennant Creek-Davenport and Pine Creek regions correlate temporally with events in the Gawler-Adélie Craton (basin development and volcanics in Pine Creek at ~1850 Ma; Carson et al., 2008, contemporaneous with deposition of the upper Halls Creek Group on the western margin of the NAC).
- The Mount Isa Province records a similar sequence of sedimentary and igneous events to the Pine Creek and Tennant regions, and to the Gawler Craton, with volcanics at ca. 1865 Ma (Page and Williams, 1988).
- Basin formation in the Curnamona Province (~1720-1640 Ma) is approximately synchronous with a period of short-lived sedimentation in the Gawler Craton (i.e., ~1715 Ma Labyrinth

Formation; Fanning et al., 2007), and the Western Fold Belt in the Mount Isa Province (Calvert and Isa Superbasins).

Payne et al. (2009) refute certain aspects of the Betts and Giles models (e.g., Betts and Giles, 2000; Giles et al., 2004; Betts and Giles 2006), in that they suggest the long-lived southern accretionary margin in the model of Betts and Giles (2006) is inconsistent with the lack of evidence for subduction and collisional orogenesis in the 1760-1690 Ma time interval. The major difference in the model of Payne et al. (2009) compared to that of Betts and Giles (2006) is that Payne et al. (2009) interpret the Kalinjala Mylonite Zone as an intracratonic shear zone, as opposed to a ca. 1730-1690 Ma continental suture, and therefore they retain the Gawler-Adélie and North Australian Cratons as a single entity at this time. The inference is that the Kimban Orogeny does not preserve evidence for subduction-related magmatism, in that it has a craton-wide distribution, and the (current) aggregate geometry of Kimban and Strangways deformation is not readily reconcilable with an east-west trending, linear plate margin setting (Payne et al., 2009).

Alternatively, Payne et al. (2009) use the similarity in the timing of the Kimban Orogeny with metamorphism of the Nimrod Group (part of the Miller Range within the Trans-Antarctic Mountains; Nimrod Orogeny 1730-1723 Ma; Goodge et al., 2001) to infer a related and possibly contiguous event involving both regions, and use these data to suggest that the 1730-1690 Ma Kimban-Nimrod Orogeny records the accretion of the Miller Range terrain to the Gawler-Adélie Craton.

In addition, Payne et al. (2009), based on work by Payne (2008b), argue against a second proposed accretionary event in the interval 1690-1670 Ma. Although previous suggestions place the Tunkillia Suite in a subduction-related setting, new age and geochemical data (Payne, 2008b; Payne et al., 2010) do not support a subduction-related setting, but rather a late- to post-tectonic setting, perhaps indicating that the Tunkillia Suite represents a continuation of the Kimban Orogeny.

Convergent margin models (ca. 1630-1550 Ma)

Advocates for plate margin models for the Mesoproterozoic evolution of eastern and central Australia have derived their interpretations, at least in part, from the spatial and temporal patterns of orogenesis and associated metamorphism (Betts and Giles, 2006; Hand et al., 2007; Cawood and Korsch, 2008), together with geochemical data suggesting the presence of magmatic arc(s) proximal to the plate margin (e.g., Wade et al., 2006; Swain et al., 2008). The two related plate margin models of Wade et al. (2006) and Swain et al. (2008) suggest plate margin architectures based on the interpreted position of arc magmatic provinces on either side of the Gawler Craton.

The I-type calcalkaline tonalite to granodiorite magmatic rocks of the St Peter Suite intruded into the Nuyts Domain in the interval ca. 1631-1608 Ma (Ferris and Schwarz, 2004). The St Peter Suite magmas comprise granite, tonalite, granodiorite, diorite and gabbro, and are considered to have formed by fractionation of a chemically-enriched mantle, metasomatised by slab-derived fluids or silica-rich melts (Swain et al., 2008). The rocks have calcalkaline affinities, are LREE-enriched, HREE and Y-depleted, with relatively juvenile $\varepsilon Nd_{(1620\ Ma)}$ values varying between -0.8 and +3.7 (Swain et al., 2008). Based on geochemical data, Ferris and Schwarz (2004) interpreted the suite to have been emplaced within a backarc environment of an active magmatic arc setting. Ferris et al. (2002) suggested that the St Peter Suite was emplaced at the southwest margin of the Archean Gawler nuclei during subduction that dipped to the northeast.

Swain et al. (2008) use the geochemical characteristics of the St Peter Suite to suggest the existence of a ca. 1631-1608 Ma volcanic arc in southern Australia. These authors proposed that arc magmatism of the St Peter Suite developed outboard of the current Gawler Craton, due to a subduction zone that dipped to the south under lithosphere now preserved in Antarctica (Figure 17). In this model, subduction ceased at ca. 1608 Ma, as a result of collision of the Archean Gawler nuclei with the St Peter Suite volcanic arc, and conceivably jumped northward, consistent with the development of the south-dipping Musgrave arc terrane (Wade et al., 2006, see below). In this model, the voluminous 1595-1575 Ma Hiltaba magmatic suite was emplaced into a near margin continental backarc setting associated with the Musgrave magmatic arc (Figure 17).

This model potentially explains the distribution of the St Peter Suite arc and provides a tectonic driver for north-south crustal shortening associated with ca. 1610-1600 Ma orogenesis in the SAC. This model, however, ignores the geochronological evidence which indicates that Terra Adélie and King George V Land in East Antarctica and the Gawler Province were contiguous (forming the Mawson Continent) from at least 2440 Ma until at least the end of the Kimban Orogeny (Fanning et al., 1996; Peucat et al., 1999, 2002). Hence, Cawood and Korsch (2008) suggest it is unlikely that there was any suturing between the two continental masses at ~1608 Ma. These authors instead follow the model of Ferris and Schwarz (2004), invoking a subduction zone that dipped to the northeast, implying the St Peter Suite was a continental margin magmatic arc.

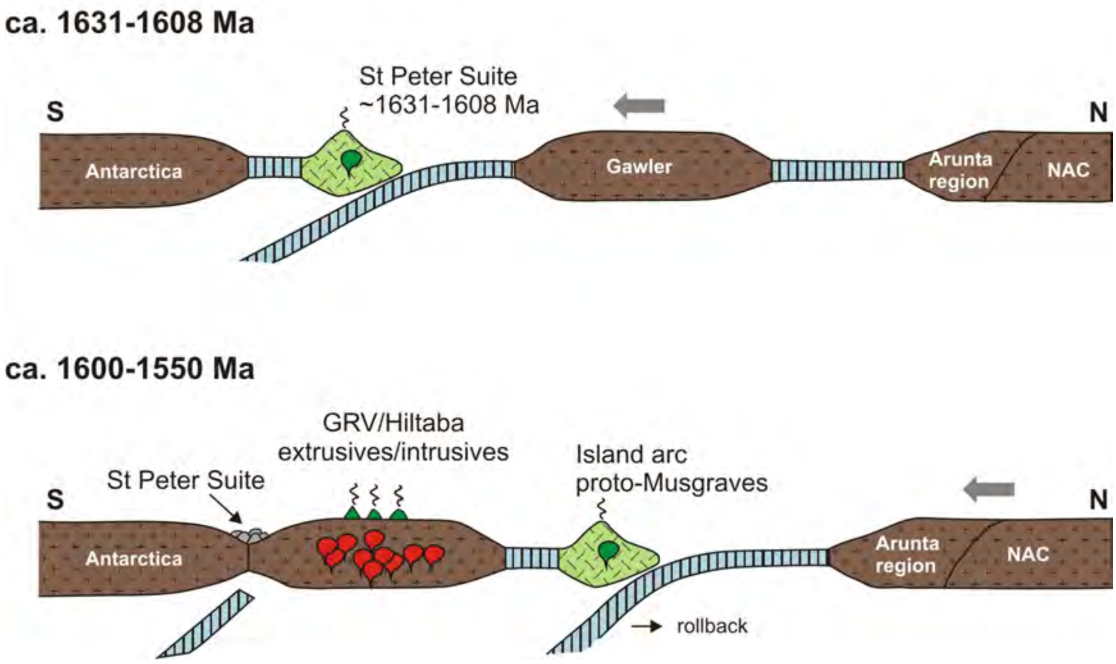

Figure 17. (top) St Peter Suite arc developed outboard of the Archean Gawler Craton via south-dipping subduction under lithosphere now preserved in Antarctica. (bottom) cessation of St Peter Suite magmatism and jump of subduction northwards, resulting in the development of the south-dipping Musgrave arc terrane (adapted from Wade et al., 2006 and Swain et al., 2008).

A suite of granulite to amphibolite facies felsic orthogneisses in the Musgrave Province are interpreted to be the reworked remnants of island arc rocks emplaced between ca. 1590 and 1550 Ma (Wade et al., 2006). These samples are REE-enriched, are characterised by negative anomalies in Nb, Ti and Y, and have $\varepsilon Nd_{(1550\ Ma)}$ values between -1.2 and +0.9. Wade et al. (2006) used this geochemical and isotopic evidence to suggest ca. 1590-1550 Ma arc-like magmatism in the Musgrave Province (Figure 17). This model suggests crustal addition in an intraoceanic island-arc setting via south-dipping subduction north of the Gawler Craton. In this model, the Musgrave Province represents the remnants of an arc terrane associated with suturing of the NAC and the SAC between ca. 1590-1500 Ma. The St Peter Suite and ca. 1590-1550 Ma Musgrave subduction-related arc magmatism are considered to represent the fundamental primer and heat engine to drive emplacement of the ca. 1595-1575 Ma GRV-Hiltaba Suite, which, in this scenario, is placed in a backarc setting (Figure 17). Collision between the Gawler Craton (Mawson continent) and the NAC, along a suture thought to be located at the northern margin of the Musgrave Province, occurred after the cessation of arc magmatism at 1550 Ma (Wade et al., 2006). This model explains the distribution of arc-related rocks in the Musgrave Province and the preservation of ca.

1580-1540 Ma deformation and metamorphism in the northern Gawler Craton (Kararan Orogeny). The model, however, provides no mechanism to drive ca. 1580-1540 Ma crustal shortening throughout the NAC, because this orogenesis predates the proposed Gawler-NAC collision (ca. 1550 Ma), nor is there any evidence for deformation in the Musgrave Province at ca. 1600-1550 Ma.

2.3.2. Overview of the Gawler Range Volcanics-Hiltaba Event

Granite magmatism of the St Peter Suite was followed by the voluminous ca. 1590 Ma Gawler Range Volcanics and 1600 ± 16 Ma to 1575 ± 6 Ma Hiltaba Suite across the Gawler Craton (Flint et al., 1993; Creaser and Cooper, 1993; Creaser and Fanning, 1993; Stewart and Foden, 2003; Budd, 2006; Zang et al., 2007). Volcanic and plutonic rocks of the same age as the Gawler Range Volcanics and Hiltaba Suite occur in the Mount Painter Inlier (Moolawatana Domain), and are also found on the formerly adjoining component of the Mawson Continent in Antarctica (Peucat et al., 2002), and in the Curnamona Province (Benagerie Volcanics and Ninnerie Supersuite). The Gawler Range Volcanics and Hiltaba Suite were interpreted as an anorogenic magmatic event linked to a mantle plume by Creaser (1995). Recent evidence points to contemporaneous metamorphism and deformation in the Gawler Craton at this time (Skirrow et al., 2006; Fanning et al., 2007), and to the suggestion of a syntectonic setting for the Gawler Range Volcanics-Hiltaba event. Deformation and metamorphism associated with the Gawler Range Volcanics-Hiltaba event occurred shortly before, and within uncertainty of, the ca. 1570-1540 Ma Kararan Orogeny (Hand et al., 2007). The timing of earliest Hiltaba Suite magmatism coincides with the Olarian Orogeny in the Curnamona Province, which may suggest that magmatism occurred in an overall compressional regime (Hand et al., 2007, 2008).

Ferris et al. (2002) inferred a backarc continental setting behind a northeast-dipping subduction zone located at the southwest margin of the Gawler Craton for granites of the Hiltaba Suite. In this scenario, granites of the Hiltaba Suite may be related to a northeast-dipping subduction zone located southwest of the Nuyts Domain, which produced the arc-related magmatism of the ~1631-1608 Ma St Peter Suite. A similar setting was proposed by Betts et al. (2003b), and Betts and Giles (2006), who envisaged that the suite was emplaced, syn- to post-deformation, in a continental backarc setting in the overriding plate, during north or northeast-dipping flat slab subduction.

Alternatively, Swain et al. (2004, 2008) suggested that the Hiltaba Suite was related to south-dipping subduction under the northern margin of the Gawler Craton, resulting in the emplacement of the ~1595-1575 Ma Hiltaba Suite into a near margin, continental backarc setting associated with the Musgrave magmatic arc. Wade et al. (2006) also place the Hiltaba Suite in a backarc setting, related to south-dipping subduction between the NAC and SAC to account for 1590-1550 Ma arc-like magmatism in the Musgrave Province. Collision between the Gawler Craton (Mawson Continent) and the NAC occurred presumably after the cessation of arc magmatism (ca. 1550 Ma). This model implies that the NAC and the Gawler Craton were disconnected until they were amalgamated at ca. 1550 Ma. Payne et al. (2008) used the temporal overlap between ca. 1595-1575 Ma A-type magmatism throughout the Gawler Craton, the ca. 1600-1590 Ma orogenesis in the Curnamona Province, and the ca. 1585-1540 Ma orogenesis in the northern Gawler Craton (Kararan Orogeny), to propose that magmatism occurred in a backarc setting that was undergoing contraction.

Betts et al. (2009) favour a plume-modified orogenic setting, where a north-dipping subducting slab along the southern Australian plate margin interacted with an impinging plume. This was the major tectonic driver for orogenesis, and was ultimately responsible for plume-related magmatism and the development of a north-trending hotspot track, defined by the distribution of A-type magmas from the Gawler through the Curnamona and into the Mount Isa Province. In this model, slab delamination and thermal assimilation of the plume and subducting slab caused a switch to crustal extension in the overriding plate, resulting in extensive, mantle-derived and crustal melting (Betts et al., 2007, 2009, see below).

Plume hotspot model (ca. 1595-1500 Ma)

Proponents for a plume origin for the Hiltaba Suite and Gawler Range Volcanics have developed models based primarily on geochemical data (i.e., the geochemistry of the Hiltaba Suite reflects interaction between continental lithosphere and a mantle plume or hotspot; Giles, 1988; Creaser and White, 1991; Flint, 1993; Daly et al., 1998), the short time frame of emplacement, the very large melt flux, isotopic trends towards juvenile compositions (Stewart and Foden, 2003), the size and circular extent of the system or, in the case of Betts et al. (2007), on the spatial and temporal distribution of A-type magmatic systems.

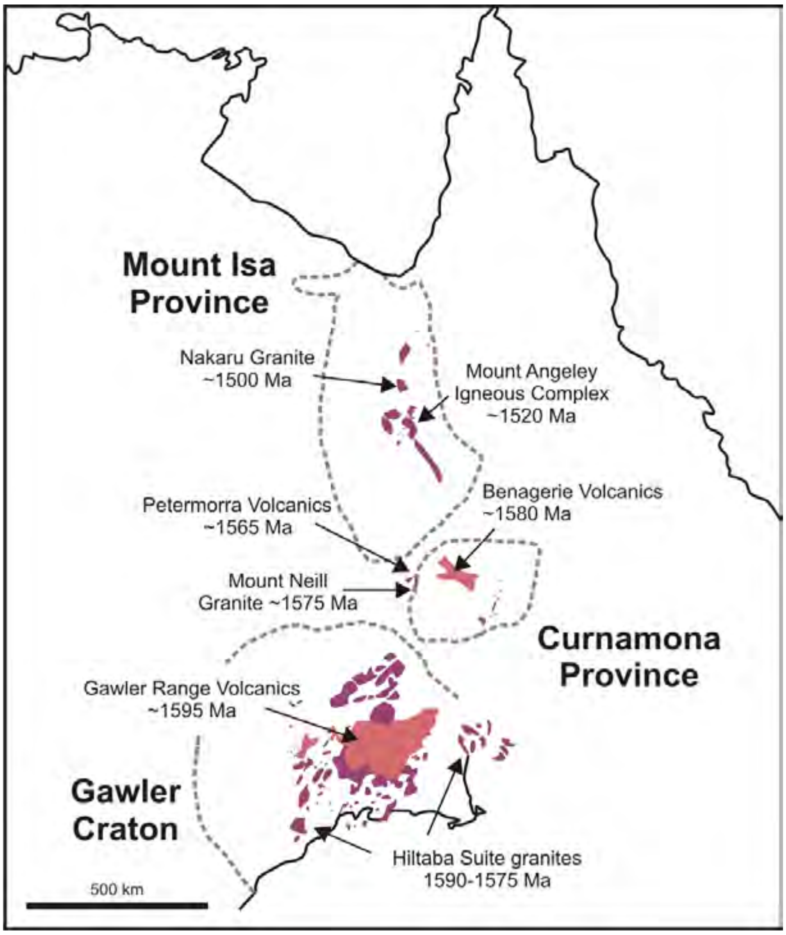

Figure 18. Schematic geological map showing the distribution of 1590-1500 Ma A-type magmatism throughout the Gawler Craton, Curnamona Province and Mount Isa Province (modified after Betts et al., 2007). The distribution of magmatism is based on the model of Giles et al. (2004) for the reconstruction of the South Australian Craton.

Betts et al. (2007) used the plate reconstruction of Giles et al. (2004) to show that ca. 1595-1500 Ma A-type magmatic rocks formed a curvilinear belt that extended from the Gawler Craton (Hiltaba Suite and Gawler Range Volcanics), through the Curnamona Province (Benagerie Volcanics and Mount Painter magmatism), and into the Mount Isa Province (Williams and Naraku Batholiths; Figure 18). Betts et al. (2007) propose that the A-type magmas along the inferred eastern margin of the Australian continent were generated in response to the interaction of a mantle plume and continental lithosphere. The distribution was inferred to record a ~1500 km portion of a hotspot track generated between ca. 1600 Ma and 1500 Ma as eastern Mesoproterozoic Australia migrated southward over a stationary plume. This interpretation is supported by the proposed geometry of the belt and the northward age progression of magmatism. Geochemically, the plume model is supported by the elevated heat flux throughout the magmatic belt, with a corresponding mantle component. The plume head is interpreted to have interacted with continental lithosphere of the Gawler Craton, where the hotspot is thickest, whereas the narrower magmatic belt in the Mount Isa Province is more consistent with a plume tail interacting with continental lithosphere. Betts et al. (2007) supported their interpretation with paleomagnetic data, by demonstrating that the A-type magmatic belt was positioned on the same trajectory as that defined by the NAC apparent polar wander path (ca. 1640-1590 Ma).

Plume-modified orogenesis (ca. 1620-1500 Ma)

A plume model in isolation does not provide a mechanism for the crustal shortening prevalent throughout eastern Australia, nor does it consider the timing and distribution of arc magmatism in the 1620-1500 Ma interval. As such, Betts et al. (2009) have attempted to reconcile the two end member models (that is, plume-driven magmatism model for the Hiltaba-GRV versus plate margin models) to account for all of the observations relating to each, and they have proposed a hybrid tectonic model of plume-modified orogenesis to explain the temporal and spatial evolution of magmatic and orogenic events in the interval 1620-1500 Ma for central and eastern Australia.

The components used to constrain the plume modified orogenic model include:

- the 1631-1608 Ma arc magmatism of the St Peter Suite and its distribution,
- the cessation of St Peter Suite magmatism and the onset of intraplate magmatism throughout the Gawler Craton at ca. 1595-1575 Ma,
- the onset of far-field orogenesis at ca. 1610-1590 Ma up to 2000 km into the continental interior,
- magmatism of the Hiltaba Suite following a major episode of north-south directed shortening in the Curnamona Province,
- the geochemistry of the Hiltaba Suite reflecting interaction between continental lithosphere and a mantle plume or hotspot,
- the spatial and temporal distribution of A-type magmas that are interpreted to indicate a hot spot track from the Gawler Craton to the Mount Isa Province (Betts et al., 2007), and
- the ca. 1590-1550 Ma interpreted arc magmatism in the Musgrave Province (Wade et al., 2006).

Betts et al. (2009) suggest that these observations can be explained where a north- to northeast- dipping subduction zone either migrated over an oceanic plume (during rollback), or a mantle plume arrived beneath a subducting slab at ca. 1620-1610 Ma along the margin (or just outboard) of the Gawler Craton (see Figure 11 of Betts et al., 2009).

In the model of Betts et al. (2009), arc magmas of the St Peter Suite formed along the southern margin of the Gawler Craton during north-dipping subduction. The cessation of arc magmatism at ca. 1610 Ma was followed by a 10-15 Ma hiatus in the Gawler Craton, interpreted to record continued flattening of the slab beneath the Gawler Craton and greater coupling between the oceanic lithosphere with the overriding plate. Flat subduction caused transient orogenesis in the overriding plate. A temporary insulating lid refrigerated the crust and removed the asthenospheric wedge, inhibiting further decompressive melting, and eventually switching off arc magmatism. Incubation of the plume beneath the subducting oceanic lithosphere and the Gawler Craton continental lithosphere in the interval 1610-1595 Ma led to thermal uplift and doming of the Gawler lithosphere, possibly explaining the depositional hiatus in the Gawler Craton at this time. Thermal assimilation or erosion of the subducting slab resulted in the plume interacting with the Gawler Craton continental lithosphere. Assimilation of the slab with the mantle plume resulted in slab delamination and the onset of renewed extension. The mantle plume, interacting with the continental lithosphere, resulted in mafic underplating, crustal melting and voluminous ca. 1595-1575 Ma volcanism and emplacement of A-type granite throughout the Gawler Craton, and over large areas of the central and northern Curnamona Province (ca. 1575-1555 Ma) and eastern Mount Isa Province (ca. 1550-1500 Ma).

Foreland basin tectonics (ca. 1600-1560 Ma)

Hand et al. (2008) have suggested the Gawler Range Volcanics and the Benagerie Volcanics constitute volcanic fill within a foreland basin, which formed within a broadly northwest-southeast-directed contractional regime. Some of the evidence for this model includes:

- crustal-scale early Mesoproterozoic deformation in the Gawler Craton,

- contractional fault reactivation, metamorphism and locally pervasive deformation associated with emplacement of the Hiltaba Suite (e.g., deformation in the Mount Woods and Coober Pedy regions, and Moonta-Wallaroo region),
- syndepositional folding, and unconformable accumulation of, the Corunna Conglomerate, a style typical of proximal foreland basins,
- high-grade metamorphism and deformation between 1590-1580 Ma in the Barossa Complex (Szpunar et al., 2007b), located between the Gawler Craton and Curnamona Province, and
- the northwestward decrease in early Mesoproterozoic metamorphic grade from the Barossa Complex through the eastern Gawler Craton.

Hand et al. (2008) suggest that this deformation forms part of a broader system that accompanied the Olarian Orogeny, and that the Gawler Range Volcanics-Benagerie Volcanics is a synorogenic system located in a foreland position to the flanking, hinterland regions that underwent broadly northwest-southeast driven crustal thickening.

2.3.3. Curnamona Province

The Willyama Supergroup is interpreted to have been deposited predominantly during active extension and thinning of the crust, with some models suggesting an intracontinental tectonic setting (e.g., Willis et al., 1983; Stevens et al., 1988; Raveggi et al., 2006; Conor and Preiss, 2008). Giles et al. (2002, 2004), Fraser et al. (2007) and Gibson et al. (2008) have suggested that this extension occurred in a backarc environment, and was part of a larger series of basins that include the Mount Isa Province and Georgetown Region (Calvert and Isan Superbasins). The backarc basin model is based primarily on the reconstructions that place the Willyama basin adjacent to the Mount Isa Province, at the southeastern margin of the North Australian Craton (Myers et al., 1996), and because the Willyama Supergroup has temporal similarities with the Mount Isa Superbasin with regard to its sedimentary, magmatic and metamorphic history (Page and Laing, 1992; Laing, 1996a, b; Barovich et al., 2002; Page et al., 2005b). Indeed, apart from the Leichhardt Superbasin, for which there is no obvious correlative in the Willyama Supergroup, the rocks of Broken Hill and Mount Isa share many similarities, including a common record of synextensional magmatism, deformation and high-T-low-P metamorphism linked to basin formation at upper crustal levels (O'Dea et al., 1997; Betts et al., 2006; Giles et al., 2006). This shared record is particularly evident for the period 1710-1670 Ma, when the bulk of synrift sedimentation occurred in the Willyama Supergroup and Calvert Superbasin (Gibson et al., 2008). Giles et al. (2004) argued that the Isan Orogeny correlates with the Olarian Orogeny, and that the Olary and Isa orogens were once continuous, observing, in particular, that deformational styles in both terranes changed from thin- to thick-skinned during the course of orogenesis. Gibson et al. (2008) suggest that there is no need for these two terranes to have been contiguous during the Paleoproterozoic. Rather, the two terranes may have occupied different parts of a single continental-scale rift system. In this interpretation, the Curnamona and Mount Isa Provinces evolved from a narrow intracontinental rift, represented by the 1790-1750 Ma Leichhardt Superbasin in northern Australia, to a fully fledged backarc basin at 1670 Ma, in both the Curnamona and Mount Isa Provinces (Gibson et al., 2008). The critical observation of Gibson et al. (2008), which differs significantly from the subduction-related models of Giles et al. (2002) and Betts and Giles (2006) for the Curnamona and Mount Isa Provinces, is that Gibson et al. (2008) invoke extensional, and not contractional, processes to have dominated the geological record, that is, the period ~1800 to ~1600 Ma was dominated by crustal extension and broadly east-west directed rifting.

Detailed geochemical and Nd isotopic analyses from the Paleoproterozoic Willyama Supergroup by Barovich and Hand (2008), suggest an intracratonic source for the lower parts of the succession, with introduction of a significantly more juvenile source region for the ca. ≤1650 Ma uppermost part, marking a significant change in tectonic regime. These authors infer a dominant intracratonic central and northern Australian provenance for sedimentary rocks of the lower Willyama Supergroup (that is, Arunta region). Geochemical and isotopic data are used to suggest that the provenance for sedimentary rocks of the upper Willyama Supergroup is from a previously undetected source not known in intracratonic Australia, although possibilities for such sources exist in southwestern Laurentia (based on pre-Rodinian plate configurations; e.g., SWEAT, Moores, 1991; AUSWUS, Karlstrom et al., 2001; AUSMEX, Wingate et al., 2002). According to Barovich and Hand (2008), a continental backarc basin setting for the Willyama basin, as suggested by Giles et al. (2002, 2004), is not supported by geochemical and isotopic data. There is no evidence for two sources (that is, a volcanic arc sediment source and a cratonic source) for sediments in the lower Willyama succession, as would be expected in a backarc setting, while the succession displays no geochemical characteristics consistent with the input of arc material (Barovich and Hand, 2008). The Nd isotopic and geochemical signature of the ca. 1680 Ma basaltic magmatism in the Curnamona Province is interpreted to suggest that the Broken Hill area was in an axial position in the forming rift (Raveggi et al., 2006; Rutherford et al., 2006, 2007).

The proposal that metamorphism in the southern Curnamona Province occurred at ca. 1690-1660 Ma (e.g., Nutman and Ehlers, 1998; Gibson and Nutman, 2004; Gibson et al., 2004; McFarlane et al., 2008) is used as supporting evidence in plate-tectonic reconstruction models featuring an Australia-southwest US connection at ca. 1690 Ma (AUSWUS) (Karlstrom et al., 1999; Burrett and Berry, 2000, 2002). This connection, however, is based on geochronological data which are widely disputed (Page et al., 2000, 2003, 2005a; Rutherford et al., 2007; Stevens et al., 2008). However, while a pre-Olarian metamorphic event has not been recorded using SHRIMP zircon geochronology, several studies using EMPA monazite geochronology have recorded earlier events, perhaps indicating that previous studies may have been limited by analytical spot size. For instance, McFarlane and Frost (2009) record pulses of monazite growth at ca. 1657 Ma, ca. 1630 Ma and ca. 1602 Ma, which correspond to monazite growth during lower amphibolite facies, upper amphibolite facies and granulite facies metamorphism, respectively; but find no evidence of monazite growth at 1690-1670 Ma (c.f. Gibson and Nutman, 2004). Evins et al. (2008) report matrix monazite EMPA age peaks from pelites in the Sundown Group of ~1675 Ma, ~1635 Ma, with a second sample yielding monazite ages ranging from >1700 Ma to ~1550 Ma. The ~1675 Ma age is interpreted to record peak metamorphism, in support of a pre-Olarian metamorphic event. Similarly, White et al. (2006) report EMPA monazite ages from metasedimentary rocks throughout the stratigraphy, recording a broad peak from ca. 1720 Ma to 1600 Ma, again, supporting a pre-Olarian metamorphic event.

Rutherford et al. (2007) considered that the Olarian Orogeny at ca. 1600-1585 Ma was due to convergence along the eastern margin of the North Australian Craton, and their model requires the North Australian Craton and South Australian Craton to have been a single entity at this time. Giles et al. (2004), Betts and Giles (2006), and Forbes et al. (2008) also considered that the North Australian Craton and South Australian Craton were a single entity at this time; Giles et al. (2004) and Forbes et al. (2008) have a subduction zone located to the southwest of the Gawler Craton, whereas Betts and Giles (2006) have subduction zones both to the southwest of the Gawler Craton and east of the North Australian Craton.

Forbes et al. (2008) place the Curnamona Province in a plate tectonic setting where multiple transient episodes of extension, followed by a fast switch to intense crustal shortening, can be accounted for. These authors suggest the tectonic environment may be in a backarc setting, located in the overriding plate of a subduction zone (Figure 19), which is consistent with previous suggestions for the setting of similar aged terranes (e.g., Mount Isa) during the Proterozoic (e.g., Giles et al., 2002).

In the model of Forbes et al. (2008), the Proterozoic evolution of the Broken Hill Domain is envisaged to have involved two extensional events. The first event involved rifting and opening of the basin, into which units of the Willyama Supergroup, up to the top of the Sundown Group, were deposited at ca. 1710-1670 Ma. Following early rifting, sag phase sediments of the Paragon Group were deposited at 1670-1640 Ma (Page et al., 2000). Forbes et al. (2008) suggest a transient mid-crustal extensional event (D1) occurred at 1620 Ma during prograde amphibolite facies metamorphism (M1) at temperatures and pressures of at least 600˚C and 2.8-4.2 kbar (Figure 19; Forbes et al., 2005). At ca. 1600 Ma, peak granulite facies metamorphism (M2) occurred contemporaneously with activity along lithology-parallel, high-temperature (D2) shear zones preserved throughout the Broken Hill Domain (Forbes et al., 2005, 2007; ca. 1600-1590 Ma Olarian Orogeny; Page et al., 2000).

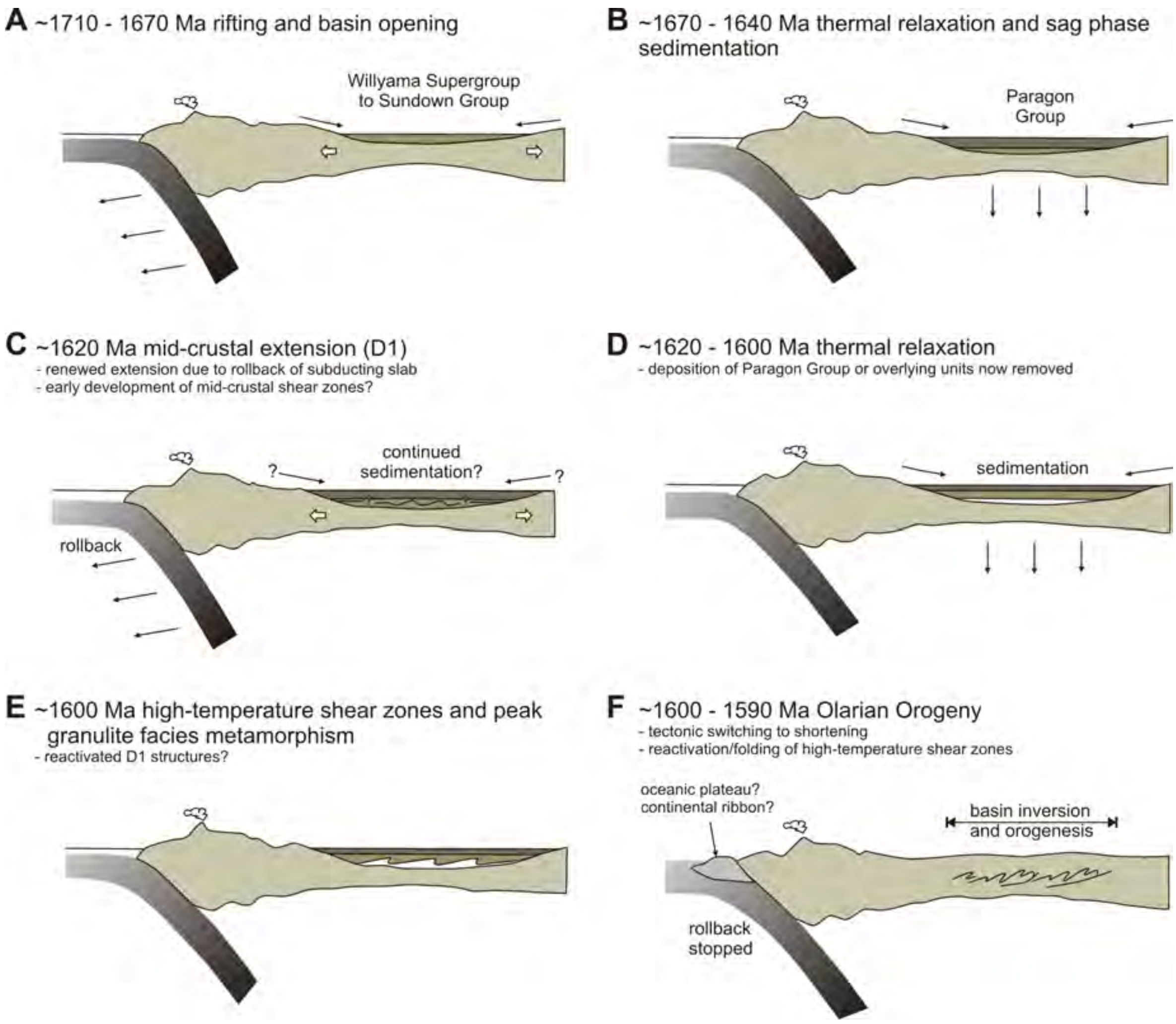

Figure 19. Schematic representation of the Proterozoic evolution of the Broken Hill Domain, placing it within a continental backarc setting (modified after Forbes et al., 2008). Note subduction polarity not given.

2.4. Links between the Gawler Craton and Curnamona Province

The Gawler Craton and Curnamona Province are separated by a zone of thick cover (Adelaide Rift System), with a minimum width of 170 km. This cover, ranging in age from Neoproterozoic to Holocene, obscures the critical relationship between the basement provinces, and makes it difficult to explore tectonic linkages between them (Szpunar et al., 2006). The Gawler Craton and Curnamona Province experienced broadly similar thermal, igneous and hydrothermal events between 1660 Ma and 1595 Ma (Figure 20), but when were these two provinces contiguous? Several lines of evidence point to the Gawler Craton and Curnamona Province being part of a coherent, and likely contiguous, crustal system by the Late Paleoproterozoic to early Mesoproterozoic, and perhaps earlier. These are:

- Likely correlation of the Gawler Range Volcanics and Benagerie Volcanics (Figure 20);
- Similarities in the timing of early Mesoproterozoic granitic magmatism and similar A-type chemistry (Figure 20);

- An overall younging of Paleoproterozoic sedimentation from the Gawler Craton to the Curnamona Province (Figure 21), over the interval ~1790 Ma to ~1640 Ma;
- Correlations in the timing of high grade metamorphism and deformation at ~1600-1580 Ma, and
- Recent seismic results and interpretations (section 2.4.1).

Similarities between the Benagerie Volcanics and the Gawler Range Volcanics have long been recognised (Giles and Teale, 1979). As outlined in Section 1 of this report, extrusion of the bimodal Gawler Range Volcanics was synchronous with intrusion of the 1590-1575 Ma Hiltaba Suite in the Gawler Craton (Creaser et al., 1991; Creaser and Cooper, 1993; Allen et al., 2003). Contemporaneous magmatism occurred in the Curnamona Province, where a vast quantity of A-, I- and S-type magmatism was generated: A-type magmas of the bimodal Benagerie Volcanics were extruded at ca. 1585-1581 Ma (Fanning et al., 1998) and emplacement of I- and S-type granites of the Ninnerie Supersuite occurred at ca. 1590-1570 Ma (Figure 20; Ludwig and Cooper, 1984; Cook et al., 1994).

On the eastern Gawler Craton, the ~1760 Ma Wallaroo Group metasedimentary and metavolcanic rocks overlap in age with the older rift packages at Mount Isa, but no package of similar age has been identified in the Curnamona Province (Conor, 1995; Figure 20). Metamorphic grade and intensity of deformation in the Wallaroo Group decreases to the east, so it is possible that equivalents could extend eastward, with little evidence of pre-Olarian deformation, beneath the Willyama Supergroup (Preiss, 2006; 2009). Such successions are potentially much thicker than the exposed Willyama Supergroup, and Fricke et al. (2010) suggest that this may account for much of the layered mid-crust imaged on the Curnamona seismic transects (section 2.4.1). If this is indeed the case, the Gawler Craton and Curnamona Province may have been contiguous from ~1760 Ma.

Szpunar et al. (2006, 2009) provide isotopic and geochemical data for the series of Paleoproterozoic basin successions (Hutchison, Wallaroo, Willyama), which geographically link the eastern margin of the Gawler Craton to the Curnamona Province. Geochronological data from metasedimentary units in basement inliers in the Barossa Complex, which represent Proterozoic outliers between the Gawler Craton and Curnamona Province, yield depositional ages between 1740-1715 Ma (with a younger succession at ~<1680 Ma; Szpunar et al., 2006) which are younger than the eastern Gawler Craton but similar to depositional ages in the lower Willyama Supergroup. Magmatic and metamorphic data from the basement inliers constrain magmatic events at ca. 1715 Ma and ca. 1580 Ma (Szpunar et al., 2005, 2006), and metamorphic events at ~1630 Ma (early, temporally constrained) and 1590-1580 Ma (later, granulite facies event; Szpunar and Fraser, 2010). Szpunar et al. (2009) used these data to indicate that the basin system progressively youngs from the eastern side of the Gawler Craton at ~1790 Ma through to the Curnamona Province at ~1710 Ma, thereby implying spatial and temporal linkages between the eastern Gawler Craton and Curnamona Province from as early as ~1710 Ma (Figure 21). In addition, age information, Nd isotopes, geochemistry, provenance, sedimentation and magmatism all point to the Barossa Complex being a likely equivalent of the Willyama Supergroup (Szpunar and Fraser, 2010). According to Conor (2009), the easterly younging sedimentation (from ~1790 Ma Myola, and Broadview, and 'younger' Hutchison, to ~1760 Ma Wallaroo Group to ~1720-1640 Ma Willyama Supergroup) implies basin migration to the east though time (Figure 21). Rollback of an oceanic slab farther to the east is suggested as the cause of crustal attenuation, and hence backarc basin formation (Conor, 2009).

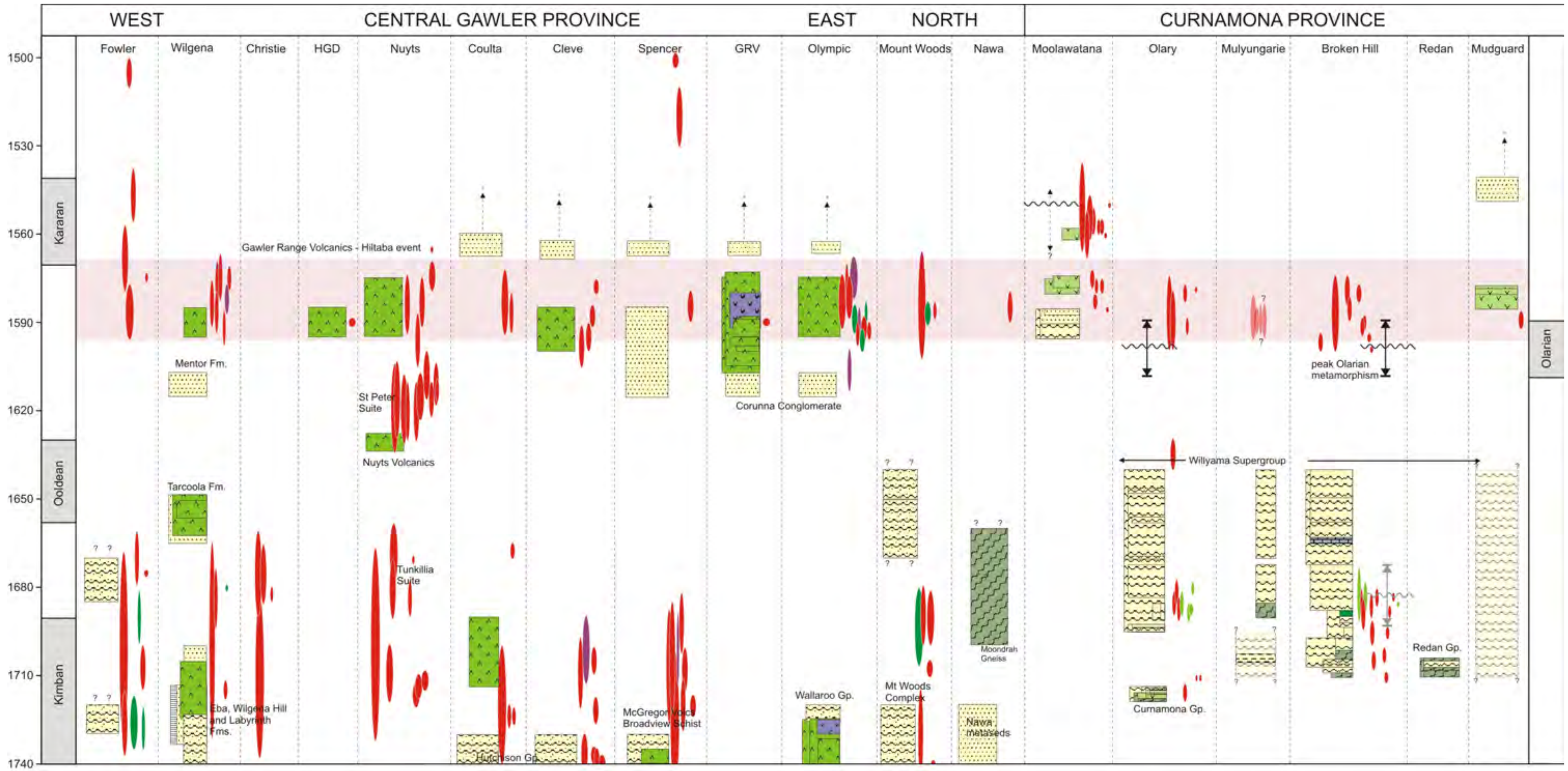

Figure 20. Time-space-event plot for the Gawler Craton and Curnamona Province, from 1740-1500 Ma, showing possible links between the two.

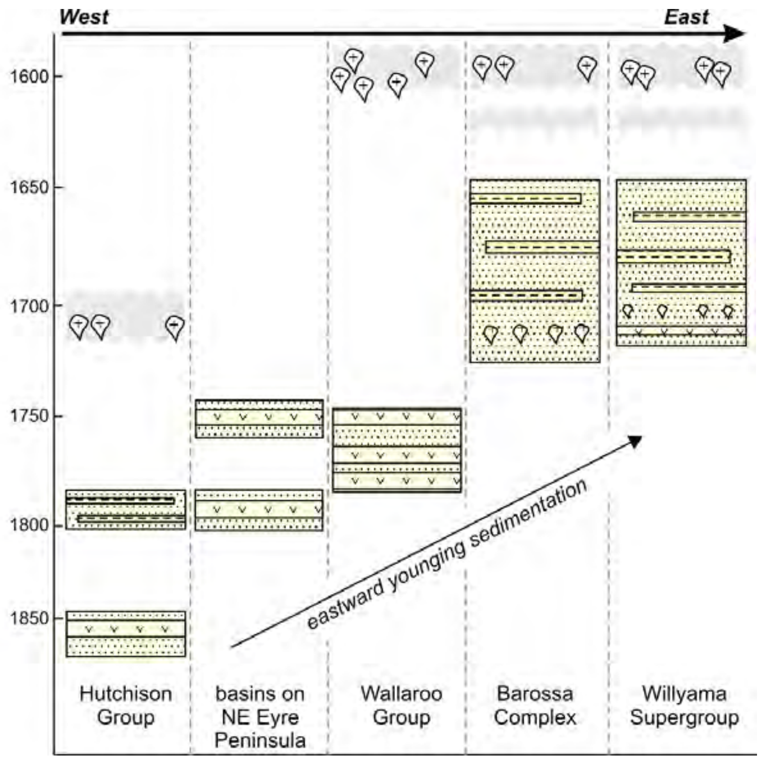

Figure 21. Schematic time-space-event plot of sedimentary basin systems from the Gawler Craton to the Curnamona Province, showing a gradual eastward younging in sedimentation.

On a broad scale, the timing of earliest Hiltaba Suite magmatism (1600 ± 16 Ma to 1575 ± 6 Ma), in part coincides with the northwest-southeast contractional 1600-1585 Ma Olarian Orogeny in the Curnamona Province. Evidence for syn-Hiltaba-aged deformation in the northern Gawler Craton appears to point to a contractional regime also operating in this area at this time. For example, zircon U-Pb data from upper amphibolite- to granulite-facies rocks in the Mount Woods Domain give metamorphic ages which range between 1587 ± 2 Ma and 1583 ± 18 Ma (Holm, 2004), indicating that Hiltaba magmatism was coeval with high-temperature metamorphism in that region. In the adjacent Coober Pedy Domain, a metapelitic granulite has a SHRIMP metamorphic age of ca. 1590 Ma (Fanning et al., 2007), coinciding with Hiltaba-aged magmatism and Olarian deformation elsewhere in the region.

Recent U-Pb LA-ICPMS dating of zircon and monazite in granulite facies metapelitic rocks in the Barossa Complex indicates high-grade metamorphism at ~1590-1580 Ma (Szpunar et al., 2007b, 2009). This age is temporally equivalent to high grade metamorphism in the Broken Hill region, magmatism in the Gawler Craton, and high grade metamorphism and deformation in the Mount Woods and Coober Pedy domains (Szpunar et al., 2007b). According to Szpunar et al. (2007b), these data may represent part of a large tectonothermal event expressed as collisional-style tectonics in the southern Curnamona Province and southeastern Gawler Craton.

The stratigraphy of the Mount Painter Inlier (Moolawatana Domain) has long been correlated with the Willyama Supergroup (Coats and Blissett, 1971; Willis et al., 1983; Teale, 1993a, b; Teale and Flint, 1993). Recent geochronology (e.g., Fanning et al., 2003; Ogilvie, 2006; Shafton, 2006; Neumann et al., 2009; Fraser and Neumann, 2010), however, has revealed that the stratigraphy at Mount Painter is younger than the Willyama Supergroup in the Broken Hill and Olary Domains, and indicates sedimentation occurred between the Gawler Craton and Curnamona Province during the early Mesoproterozoic.

Fanning et al. (2003) concluded that all the metasedimentary rocks in the Mount Painter area are Mesoproterozoic and that they did not experience Olarian deformation. Recent SHRIMP results indicate that the oldest identified stratigraphic units in the Mount Painter area were deposited at or after ~1595-1590 Ma (i.e., Radium Creek Metamorphics, Mount Adams Quartzite, Freeling Heights Quartzite;

Neumann et al., 2009; Fraser and Neumann, 2010). Detrital zircon geochronology (Sheard et al., 1992; Ogilvie, 2006; Shafton, 2006) suggests a likely Gawler Craton provenance for the succession at Mount Painter, with detrital zircon ages peaks of ~1592 Ma, 1684 Ma, 1712 Ma, 1860 Ma and 2480 Ma, matching events in the Gawler Craton (Shafton, 2006).

Magmatism in the Mount Painter area occurred between ~1585-1575 Ma (e.g., Mount Neill Granite, Box Bore Granite, Pepegoona Porphyry), with a second magmatic event recorded at ~1565-1555 Ma (i.e., Terrapinna, Wattleowie and Yerila Granites; Neumann et al., 2009; Fraser and Neumann, 2010). Magmatism at ~1565-1555 Ma was followed by high-grade metamorphism recorded by the Hot Springs gneisses (Neumann et al., 2009; Fraser and Neumann, 2010).

Geochronology and field relations suggest that deformation within the Mount Painter Inlier occurred after the ~1600 Ma Olarian Orogeny (Ogilvie, 2006; Shafton, 2006; Neumann et al., 2009). According to Ogilvie (2006), the Mount Painter Inlier shares a similar deformational evolution to both the Mount Isa Province and also the Curnamona Province, characterised by a D1 layer parallel fabric, D2 tight, recumbent folding that formed during ~north-south shortening and D3 upright, open folding which formed during ~east-west shortening. Ogilvie (2006) and Shafton (2006) constrain Mesoproterozoic deformation (the 'Painter Orogeny') to ~1590-1575 Ma, based on ~1595-1590 Ma maximum depositional ages for the sedimentary succession and the age of the Mount Neill Granite, which intruded into deformed metasedimentary rocks. They suggest also that the style and timing of deformation to be similar to the Isan Orogeny in the Mount Isa Province (~1600-1500 Ma, with peak metamorphism within the Eastern Succession at 1585 Ma; Page and Sun, 1998; Giles and Nutman, 2002; Giles and Nutman, 2003). In addition, Fraser and Neumann (2010) record high grade metamorphism of gneisses at Paralana Hot Springs at ~1550 Ma, while Rutherford et al. (2007) record an additional episode of high grade metamorphism, now preserved in the Broken Hill region, and retrograde metamorphism preserved in the Olary Domain and western Broken Hill Domain between ca. 1570 Ma and 1550 Ma (EMPA monazite). Rutherford et al. (2007) noted that this event had not been previously identified in the southern Curnamona Province but has correlatives in the Mount Isa Province and Georgetown Region.

2.4.1. Seismic data and implications

In mid-2008 and early 2009, Geoscience Australia, in conjunction with PIRSA, acquired ~720 km of deep seismic reflection data in the southern Gawler Craton and Curnamona Province (Figure 22, 23), which have shed some light on possible relationships both within, and between, the Gawler Craton and Curnamona Province. These data consisted of four separate profiles, namely, a 253 km long east-west profile in the southern Gawler Craton (08GA-G1, Fraser et al., 2010b), a 60 km long profile across the western Arrowie Basin (08GA-A1, Carr et al., 2010), a 262 km long, north-south oriented profile in the Curnamona Province in South Australia (08GA-C1, Korsch et al., 2009, 2010a) and a 144 km long profile which links the Gawler Craton to the Curnamona Province (09GA-CG1, Preiss et al., 2010). These data, together with those of previous seismic studies (lines 03GA-OD1, 03GA-OD2, 03GA-CU1, 96AGS-BH1A, 96AGS-BH1B; Gibson et al., 1998; Goleby et al., 2006; Korsch et al., 2006b), have been used to help characterise the crustal architecture and inferred geodynamics of this region, and have provided fundamental constraints on the geometries of the Gawler Craton and Curnamona Province, and previously unknown seismic provinces (see description below). The combined seismic profiles, from the western Eyre Peninsula to the Darling Basin, covering a distance of about 870 km, provide a near complete cross-section of the crust across the Gawler Craton, Adelaide Rift System, Curnamona Province, Koonenberry Belt and western Darling Basin (Figure 22, 23). Interpretation of seismic sections forming this transect are shown in Figure 23.

Korsch et al. (2010b) use the term seismic province to refer to a discrete volume of middle to lower crust, which cannot be traced to the surface (and hence there are no geological constraints on its age or composition), and whose crustal reflectivity is different to that of an adjoining seismic province. At least four discrete middle to lower crustal seismic provinces have been recognised in the Gawler-Curnamona region, which are, from west to east, the Yeltana, Warrakimbo, Yarramba and Tandou Seismic Provinces. These seismic provinces are described in Korsch et al. (2010b), Carr et al. (2010), Preiss et al. (2010) and Fraser et al. (2010b), and shown in Figure 23.

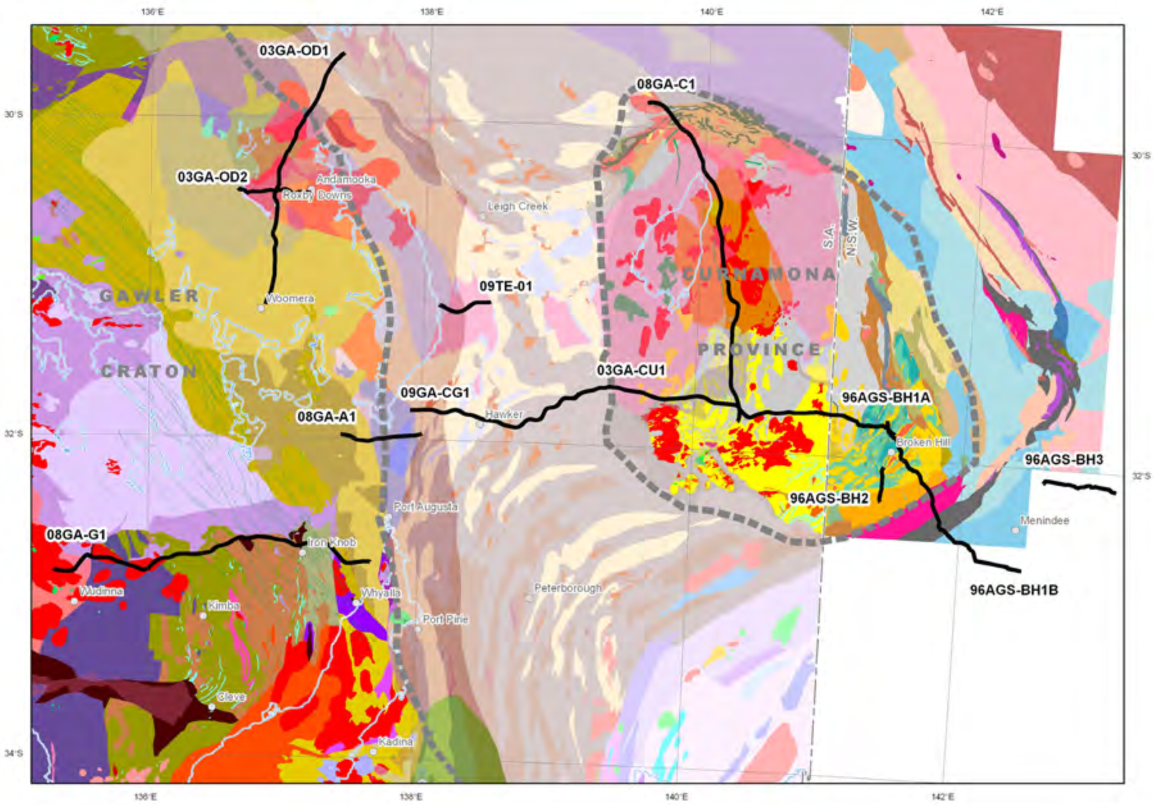

Figure 22. Map showing the solid geology of the area covered by the east-west seismic transect (after Cowley, 2006, which also contains the legend), the inferred boundaries, at the surface of the Gawler Craton and Curnamona Province (dashed lines), and the locations of the deep seismic lines in South Australia and New South Wales (from Korsch et al., 2010b).

All four seismic provinces depicted in Figure 23 are bounded by east-dipping, crustal-penetrating fault zones which extend to the Moho. Each seismic province has a markedly different seismic reflectivity to the adjacent seismic province, suggesting that each is a unique piece of middle to lower crust, consisting of differing lithologies with differing geological histories and architectures. Age control of the seismic provinces is poor, as they cannot be traced to the surface. The timing of their amalgamation with each other is also poorly constrained (Korsch et al., 2010b).

Preliminary results from the Curnamona-Gawler Link Line (09GA-CG1; Figure 22, 23) deep reflection seismic survey, an east-west transect 144.44 km long across the Flinders Ranges, are reported by Preiss et al. (2010). The primary aim of this seismic survey was to establish the relationship between the Paleo- to Mesoproterozoic Gawler Craton and the Curnamona Province. As mentioned earlier, the relationship between these basement provinces is obscured by a zone of thick sedimentary cover forming the Neoproterozoic to Cambrian Adelaide Rift System (Preiss, 2010). Seismic line 09GA-CG1 has provided new insights into the stratigraphy and tectonic architecture of a part of the Adelaide Rift System, in addition to its primary aim of deciphering the relationship, at depth, between the Gawler Craton and the Curnamona Province (Preiss et al., 2010).

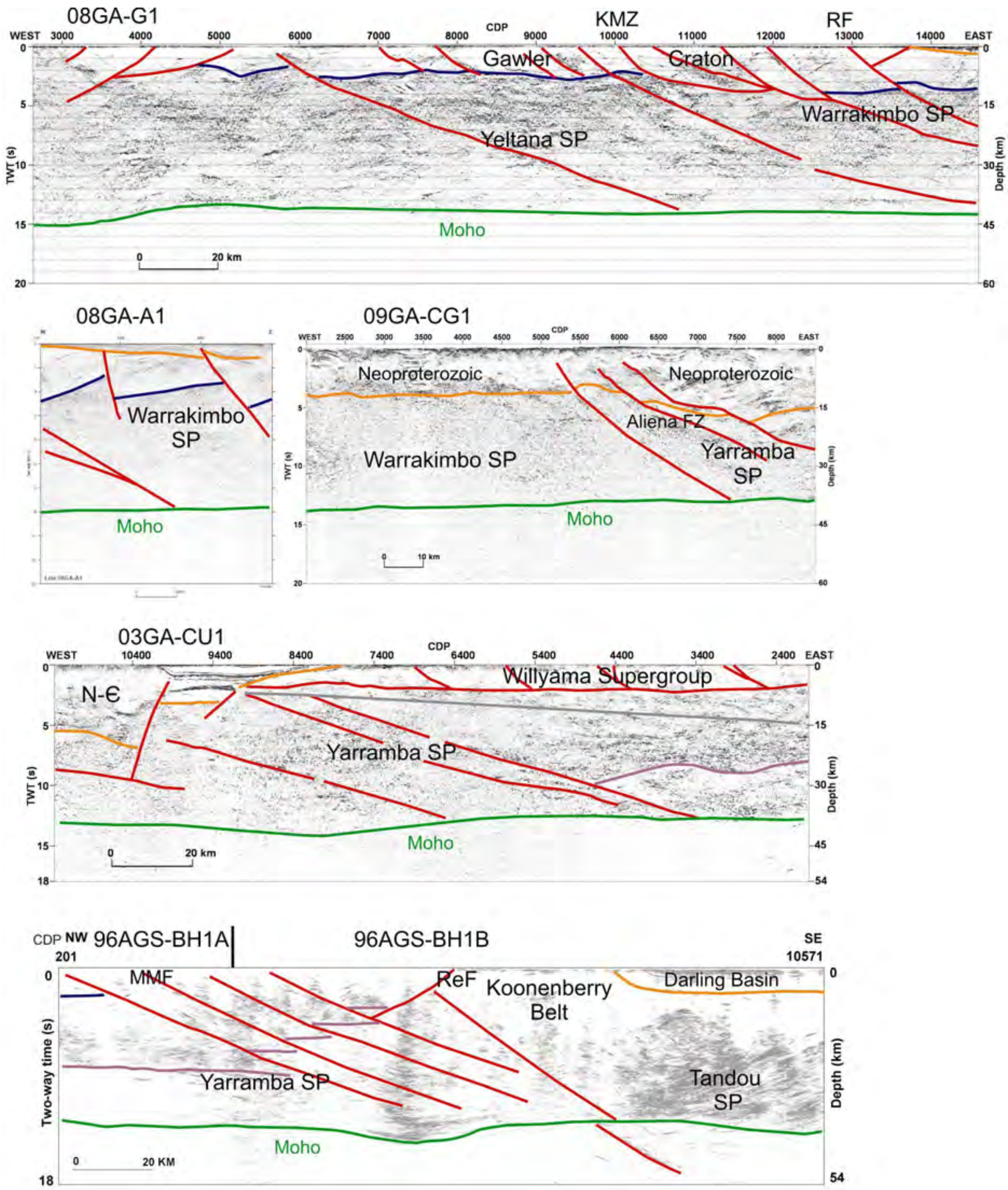

Figure 23. Interpretations of seismic sections forming a transect from the western Eyre Peninsula to the Darling Basin. Display shows the vertical scale equal to the horizontal scale, at a crustal velocity of 6000 km $^{s-1}$. Horizontal scale is based on 1 CDP equal 20 m. KMZ = Kalinjala Mylonite Zone, RF = Roopena Fault, ReF = Redan Fault, MMF = Mundi Mundi Fault. Red = faults, orange = base of Neoproterozoic succession (and base of Darling Basin on lowest section), blue = top of reflective middle crust, purple = base of reflective middle crust, grey = base of Willyama Supergroup (from Korsch et al., 2010b).

Seismic section 09GA-CG1 imaged a major, crustal-penetrating fault zone, the Aliena Fault Zone, which separates lower crust of very different seismic character (Figure 23; Preiss et al., 2010; Korsch et al., 2010a, b). The crust to the west of the fault zone (the Warrakimbo Seismic Province of Korsch et al., 2010b), below the Gawler Craton, is essentially nonreflective, whereas the lower crust to the east of the fault zone (the Yarramba Seismic Province of Korsch et al., 2010), below the Curnamona Province, is much more reflective. Korsch et al. (2010b) consider that the Yarramba Seismic Province can be tracked at depth beneath the entire exposed part of the Curnamona Province. Importantly, the Aliena Fault Zone represents a major crustal boundary, with significantly different lower to middle crust on either side, thus indicating that it may represent a zone across which there was an amalgamation of terranes, prior to the deposition of the Willyama Supergroup (Korsch et al., 2010b).

Preliminary results from deep seismic reflection line 08GA-C1 (a single traverse in the Curnamona Province; Figure 22, 23), oriented approximately north-south, are reported by Korsch et al. (2010a). The northern limit of the Curnamona Province is defined by a possible boundary between two different types of lower crust. In the northernmost part of the line, where rocks of the Mount Painter and Mount Babbage Inliers (part of the Moolawatana Domain) are exposed, the crust has a markedly lower reflectivity than that of the remainder of the line. The observed contrast in crustal reflectivity suggests that the crust beneath the Moolawatana Domain is different to that beneath the Willyama Supergroup of the Curnamona Province to the south, and a southeast-dipping boundary between the Moolawatana Domain and the Curnamona Province is inferred (Korsch et al., 2010a). This observation is consistent with the geochronological evidence, which points to the Moolawatana Domain having a different, and younger, succession to that observed in the Broken Hill and Olary Domains.

The location and orientation of the 08GA-1 seismic survey were selected in order to investigate the crustal architecture of the eastern part of the Gawler Craton (Fraser et al., 2010b), with the seismic line crossing the Moonta, Spencer, Cleve, Coulta and Nuyts Domains of Ferris et al. (2002). Preliminary interpretations of this line are reported in Fraser et al. (2010b), and infer that a crustal-scale, east-dipping fault zone separates crust of different seismic character beneath the central versus the eastern Eyre Peninsula (Figure 23). The boundary zone broadly corresponds to the Kalinjala Mylonite Zone, forming the boundary between the Cleve and Spencer Domains at the surface and the Yeltana and Warrakimbo Seismic Provinces in the middle to lower crust (Fraser et al., 2010b; Figure 23). This observation complements other geological datasets, which suggest a change in crustal age, density, geological history, radiogenic heat production, and mineralisation-style across the same structural zone, although the timing and kinematics of movement within the structural zone remain open to interpretation (Fraser et al., 2010b). The geodynamic implications of the preliminary seismic interpretations (Figure 23) are investigated in Section 3 below.

SECTION 3. GEODYNAMIC MODEL FOR THE GAWLER CRATON AND CURNAMONA PROVINCE

N. Kositcin, R.J. Korsch, D.C. Champion and P.A. Henson.

3.1. Introduction

The discussion in Section 2 detailed the various geodynamic models advocated for the evolution of the Gawler Craton and Curnamona Province, and highlighted apparent ambiguities in the geological evidence used to infer some of these tectonic settings. Given these uncertainties, and the differences between geodynamic models, we present our preferred model for the Archean to early Mesoproterozoic geodynamic evolution of the Gawler Craton and Curnamona Province.

A number of major orogenies have occurred within the Gawler Craton and Curnamona Province (Table 1). It is proposed here to describe the evolution of the Gawler Craton and Curnamona Province in terms of tectonic cycles, to encompass depositional, magmatic, and deformational histories of rock packages. Tectonic cycles are used in the sense of the Wilson cycle, to record the cycle starting after an orogeny, typically involving renewed extension, and ending by major orogeny (as adopted by Glen, 2005 for the Tasmanides of eastern Australia). A complete idealised cycle begins with a rift phase, which may pass into drift (passive margin), then into convergence (if subduction can be recognised), followed by deformation and/or collision which terminates the cycle, although a post-collisional phase may also be present (Glen, 2005). The cycle approach has been used to simplify, and attempt to unify, the Gawler Craton and Curnamona Province into a geodynamic framework.

We present geodynamic scenarios for the following time intervals:

- 3400-3150 Ma
- 2550-2400 Ma
- ~2000 Ma
- 1900-1830 Ma
- 1800-1550 Ma

For each of the above intervals, our model is presented in figures; the information for this is derived, in part, from the available geological evidence and time constraints, as summarised in Parts 1 and 2, and from the deep seismic reflection data and interpretations presented in Korsch and Kositcin (2010). It must be noted that this is only one possible interpretation for the evolution of the Gawler Craton and Curnamona Province, and alternative interpretations may also honour the available data. The models outlined therefore represent a simplification of the geological evidence, and we propose them to aid consideration of the geodynamic evolution of the Gawler Craton and Curnamona Province which, presently, is poorly understood.

3.2. Constraining the crustal components using Sm-Nd isotopes

Available Sm-Nd isotopic data for felsic magmatic samples (granites and orthogneisses predominantly, and volcanic rocks) were compiled for the Gawler Craton region, South Australia, to help characterise the geological basement and identify discrete crustal blocks (if present). Sm and Nd are both members of the rare earth series of elements (REE) and, as such, have very similar behaviour in felsic systems, such as granites. Accordingly, Sm-Nd isotopic data can be used to effectively 'see' through crustal processes, that is, to provide an 'average' (model-dependant) crustal age for the source of the rocks in question. For volumetrically abundant rocks such as granites, which commonly have a dominant crustal component (± varying mantle input), this provides a potentially powerful proxy to constrain the bulk age of the crust within which the granites occur. Sm-Nd datasets for felsic magmatic rocks within a region can, therefore, be used to provide constraints on the relative ages of the middle to lower crust, and to discriminate different crustal blocks (with some caveats; see Champion and Cassidy, 2008).

Two-stage Sm-Nd depleted-mantle model ages (T_{2DM}) were calculated for available Archean-Mesoproterozoic felsic intrusive and extrusive samples from the Gawler Craton, which were used to produce a map of relative crustal ages (following Champion et al., 2009b; Figure 24). The calculated model ages range from 3.45 Ga to 1.87 Ga. In ArcMap, the point locations and model ages were used to generate the Inverse Distance Weighted (IDW) grid, with bin values selected on the histogram. Owing to the method of gridding, the resulting distribution of ages is subject, in places, to computational artefacts, particularly in areas of low data density. Consequently, only broad trends are considered in the following discussion.

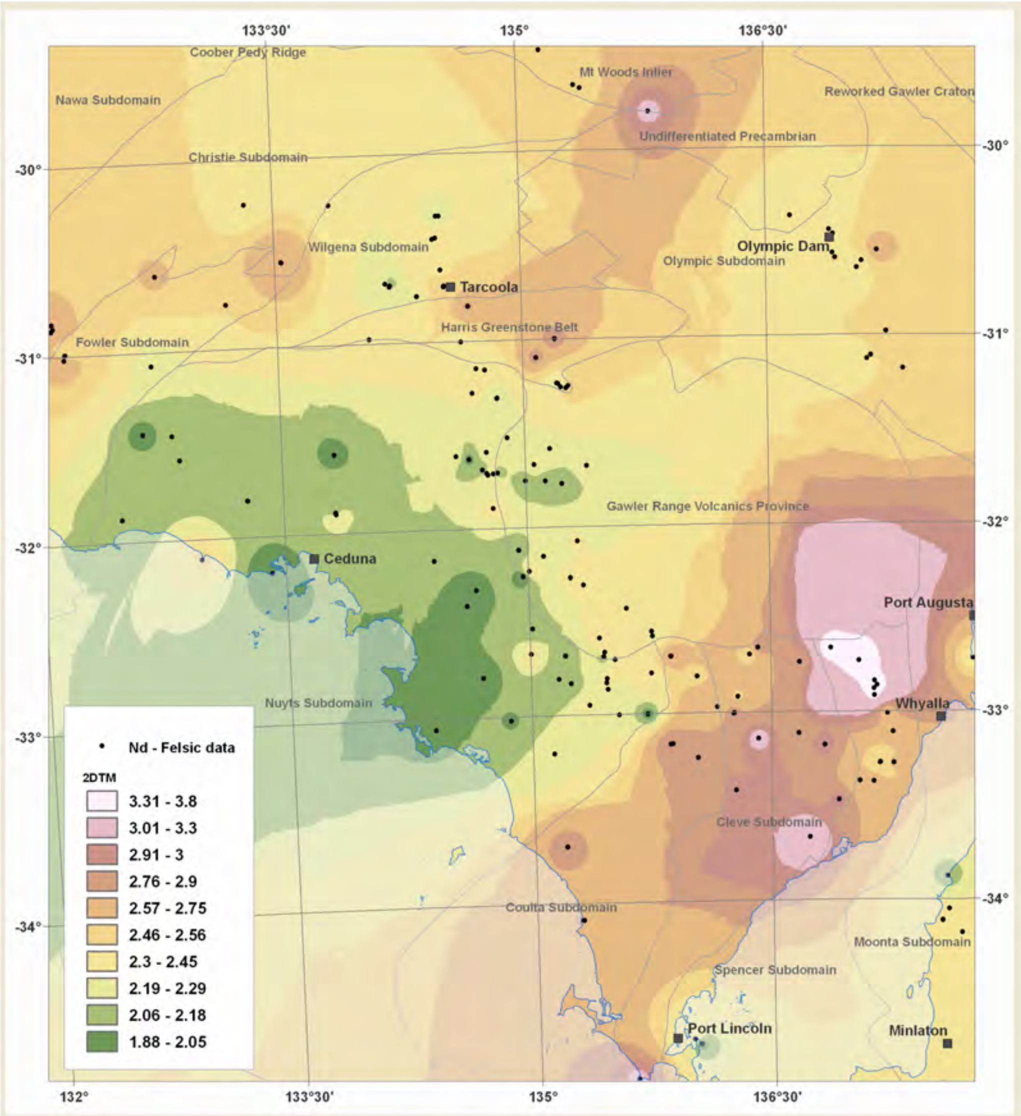

Figure 24. Gridded image of Sm-Nd isotope data, showing the 2-stage depleted mantle model crustal ages for felsic magmatic units of the Gawler Craton, with domain boundaries from Ferris et al. (2002). Data sources: Budd (2006), Creaser (1989), Fanning (2002), G. Fraser (unpublished data), Johnson (1993), Johnson and McCulloch (1995), Payne et al. (2010), Stewart and Foden (2003), Stewart (1994), Swain et al. (2005a), Turner et al. (1993), Wurst (1994). Black points represent location of Nd data used for Nd model ages and gridding.

The spatial distribution of model ages in the study area clearly suggest there are some significant trends in the data (Figure 24). These variations identify boundaries characterised by abrupt changes in T_{2DM} (Neumann, 2001; Stewart and Foden, 2003; Hand et al., 2007). Of note are the northeast orientation of gradients defined in two areas: 1) the boundary between the significantly younger Nuyts Domain in the central Gawler Craton and the oldest (>3 Ga) model ages in the Cleve and Spencer Domains, south of the Gawler Range Volcanics, and 2) the one >3 Ga model age recorded in the Mt Woods Domain (Figure 24). On the assumption that the model ages are generated from the middle-lower crust, the recorded model ages display significant heterogeneity in the maturity of the source region. This variation within the Gawler Craton can be caused by a variety of processes, including the amalgamation of different crustal blocks through collision tectonics, or the temporal evolution of the crust through plate margin dynamics. There are few definitive ways to evaluate these varied hypotheses, although analysis of the magmatic ages of granites compared to their model ages provides some clues.

A significant proportion of the Nd data used in this compilation is from the ~1590 Ma Hiltaba Suite. The Nd model ages for this magmatic suite show a distinct spatial change across the Gawler Craton, from younger model ages (<2.3 Ga) in the north and west, to older model ages (>2.3 Ga) in the south and east. The change in model ages in the Hiltaba Suite is coincident with the Nuyts-Coulta Domain boundary. Assuming that the Hiltaba-GRV magmatic event has a mantle component (e.g., Stewart, 1994), the variation in model ages observed in the Hiltaba Suite reflects either: 1) variation in the isotopic signature (~age) of the crustal component, or 2) variations in the amounts of crust and mantle end members within these granites (Stewart and Foden, 2003). In their study, Stewart and Foden (2003) concluded that the Nd variations observed in the Hiltaba Suite resulted from variations in the isotopic signature (~age) of the crust across the Gawler Craton.

3.3. Geodynamic interpretation by time slice

The development of the Gawler Craton and the Curnamona Province was influenced by geodynamic processes operating along their margins. In the following sections, we discuss geodynamic processes that we interpret to have operated at the eastern, southeastern and northern margins of the Gawler Craton and Curnamona Province.

3.3.1. Geodynamics driven by convergent processes to the east

~3400-3150 Ma

Identification of ~3150 Ma Paleoarchean-Mesoarchean crust (with inheritance as old as 3300 Ma; Fraser et al., 2010a) near the eastern margin of the Gawler Craton requires consideration of how it might link with other Australian cratonic elements. No direct age equivalents of these ~3150 Ma rocks have been identified elsewhere in Australia, apart from the possibility of the Mount Billroth Supersuite (~3200-3165 Ma) in the East Pilbara, which includes the Flat Rocks Tonalite with U-Pb zircon crystallisation ages of 3166 ± 5 Ma and 3164 ± 4 Ma (Van Kranendonk et al., 2006). Mesoarchean detrital zircons identified in metasedimentary rocks in the Mulgathing Complex in the Gawler Craton (Jagodzinski et al., 2009), metasedimentary rocks within the Pine Creek Orogen (Hollis et al., 2009), and metasedimentary rocks in the Mount Isa Province and Georgetown Region (Neumann et al., 2008, 2009b), suggest a Mesoarchean source area considerably larger than that currently identified in the eastern Gawler Craton (Fraser et al., 2010a).

Fraser et al. (2010a) suggest that the geochemistry of the Cooyerdoo Granite (~3150 Ma), particularly its elevated LILE and LREE content and low Na/K, is most similar to post-tectonic, potassic granites in the Yilgarn and Pilbara Cratons – certainly they are not similar to Archean TTGs. Archean potassic granites, such as those in the Pilbara and Yilgarn Cratons, are interpreted to be the products of melting of pre-existing TTG crust (Champion and Sheraton, 1997; Champion and Smithies, 2001, 2007) – an interpretation also favoured by Fraser et al. (2010a) for the Cooyerdoo Granite. This suggests that TTG precursors exist in parts of the Gawler Craton. Based on inherited zircon and Sm-Nd model ages for the Cooyerdoo Granite, these TTG precursors are inferred to have been ca. 3400 Ma to 3200 Ma in age. Archean TTGs are thought to be largely derived by partial melting of basaltic protoliths, mostly at high pressures, by melting either within subducting slabs or thickened continental crust (Figure 25; Champion and Smithies, 2007). Either option is possible for the source of the Cooyerdoo Granite, but, with the current data and limited outcrop extent, this cannot be constrained further.

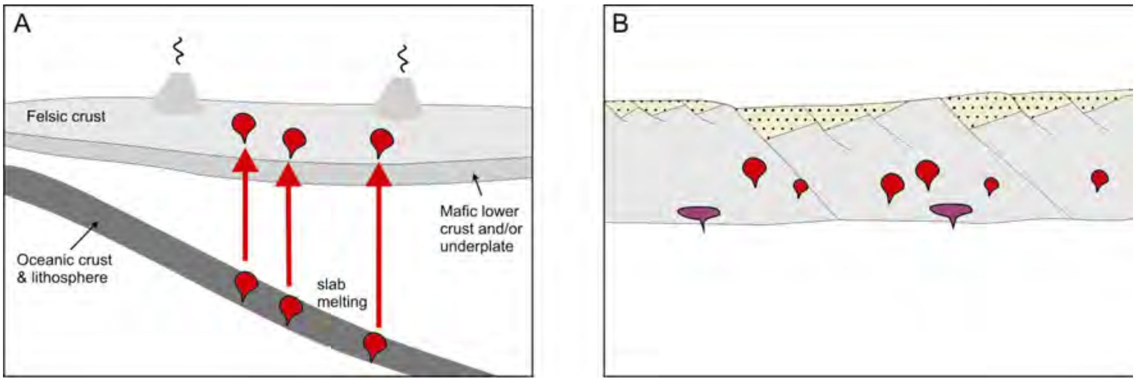

Figure 25. Generation of TTG magmatism through either (a). slab melting, or (b). thickened crust (modified from Champion and Smithies, 2007).

~2550-2400 Ma

For the time slice 2550-2400 Ma, we tentatively follow a model similar to that of Swain et al. (2005a), although we acknowledge that direct evidence for plate margin-driven tectonism in this time interval is limited. Swain et al. (2005a) interpreted the Devils Playground (~2553 Ma) and Hall Bay Volcanics (~2525 Ma) to have been generated within an arc environment. Interestingly, the Devils Playground Volcanics contain members characterised by high MgO, Mg#, Ni and Cr, not unlike Archean high-Mg diorites (or sanukitoids) (e.g., Smithies and Champion, 2000). These latter rocks are interpreted to have formed either in subduction environments or reworking of mantle previously metasomatised by slab melts – though in modern environments they are mostly found associated with active subduction (see Martin et al., 2005). Unfortunately, the chemistry of the Devils Playground Volcanics is not definitive – in particular the very low LILE, especially Sr, is atypical of such high-Mg rocks – though this may simply reflect subsequent loss of the LILEs.

An interpreted arc environment for the Hall Bay Volcanics is more problematical as these rocks appear to be dominated by dacitic and rhyolitic volcanics, and this limited range in geochemistry makes tectonic interpretations difficult. Accordingly, there is little evidence that the arc environment inferred for the Devils Playground Volcanics was still operative 30 million years later when the Hall Bay Volcanics were generated. In fact, the contemporaneity of the Hall Bay Volcanics with the mafic and ultramafic basalts and komatiites of the Harris Greenstone Belt (ca. 2520 Ma), strongly suggests a plume-related extensional environment – probably backarc at best (on crust of the Yeltana Seismic Province; Figure 26A). Evidence for such an environment is also provided by the Coulta Granodiorite (Swain et al., 2005a), which very much resembles the HFSE- and Fe-enriched felsic granites found on other Archean terranes, and is interpreted to result from high-temperature melting in a bimodal, possibly backarc, environment (e.g., Champion and Cassidy, 2008). We favour a backarc setting for the generation of these rocks, as well as the Sleaford and Mulgathing Complexes and Harris Greenstone Domain rocks in this time interval (Figure 26A). Further evidence for a possible convergent setting around this time is the chemistry of the slightly younger ca. 2500 Ma Glenloth Granite (Swain et al., 2005a), which is geochemically similar to Archean TTGs and, hence, may have been derived by melting of a subducting slab (Figure 26B1), or a change in subduction dynamics may have triggered lower crustal melting of thickened crust (Figure 26B2). Basin formation and associated magmatism was terminated by granulite-grade metamorphism and crustal thickening during the Sleafordian Orogeny, which may have been the result of collision following ocean closure, and the arrival of a continent or continental fragment from the east (Warrakimbo Seismic Province, Figure 26C).

Collision and obduction of the Warrakimbo Seismic Province is interpreted here to have created the proto-Kalinjala Mylonite Zone during the Sleafordian Orogeny (~2460-2400 Ma) (Figure 26C). Seismic line GA08-G1 images a crustal-scale, east-dipping fault zone, which separates crust of different seismic character beneath the central and the eastern Eyre Peninsula - the Yeltana and Warrakimbo Seismic Provinces (Korsch and Kositcin, 2010). The boundary zone broadly corresponds, at the surface, to the Kalinjala Mylonite Zone, which separates the Cleve and Spencer Domains (Fraser et al., 2010b) (Figure 23).

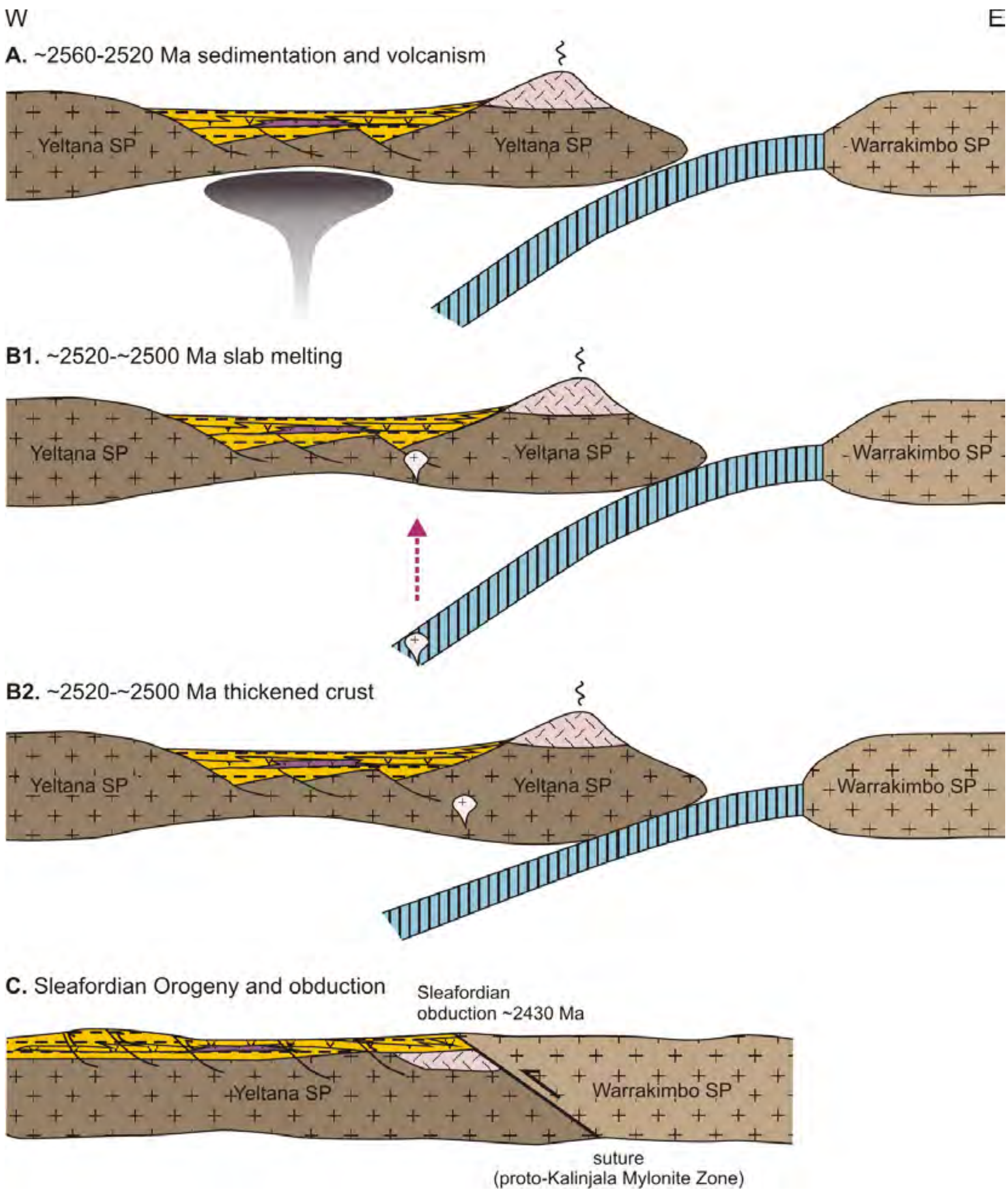

Figure 26. Schematic cross sections showing the proposed tectonic development of the Gawler Craton and Curnamona Province from ~2550 Ma to ~2430 Ma.

~2000 Ma

The Miltalie Gneiss is interpreted to have been emplaced during a period of extensional tectonism (Daly et al., 1998; Figure 26D). The Miltalie Gneiss is overlain by the Hutchison Group (<2000 Ma - 1730 Ma, Szpunar and Fraser, 2010; Szpunar et al., in press), a succession of clastic and carbonate units, which were deposited on a shallow continental shelf which deepened to the east (Daly et al., 1998). This suggests that, during the deposition of the Hutchison Group, the extensional event continued through to breakup with the formation of a passive continental margin at the eastern edge of the Archean nucleus (Figure 26E).

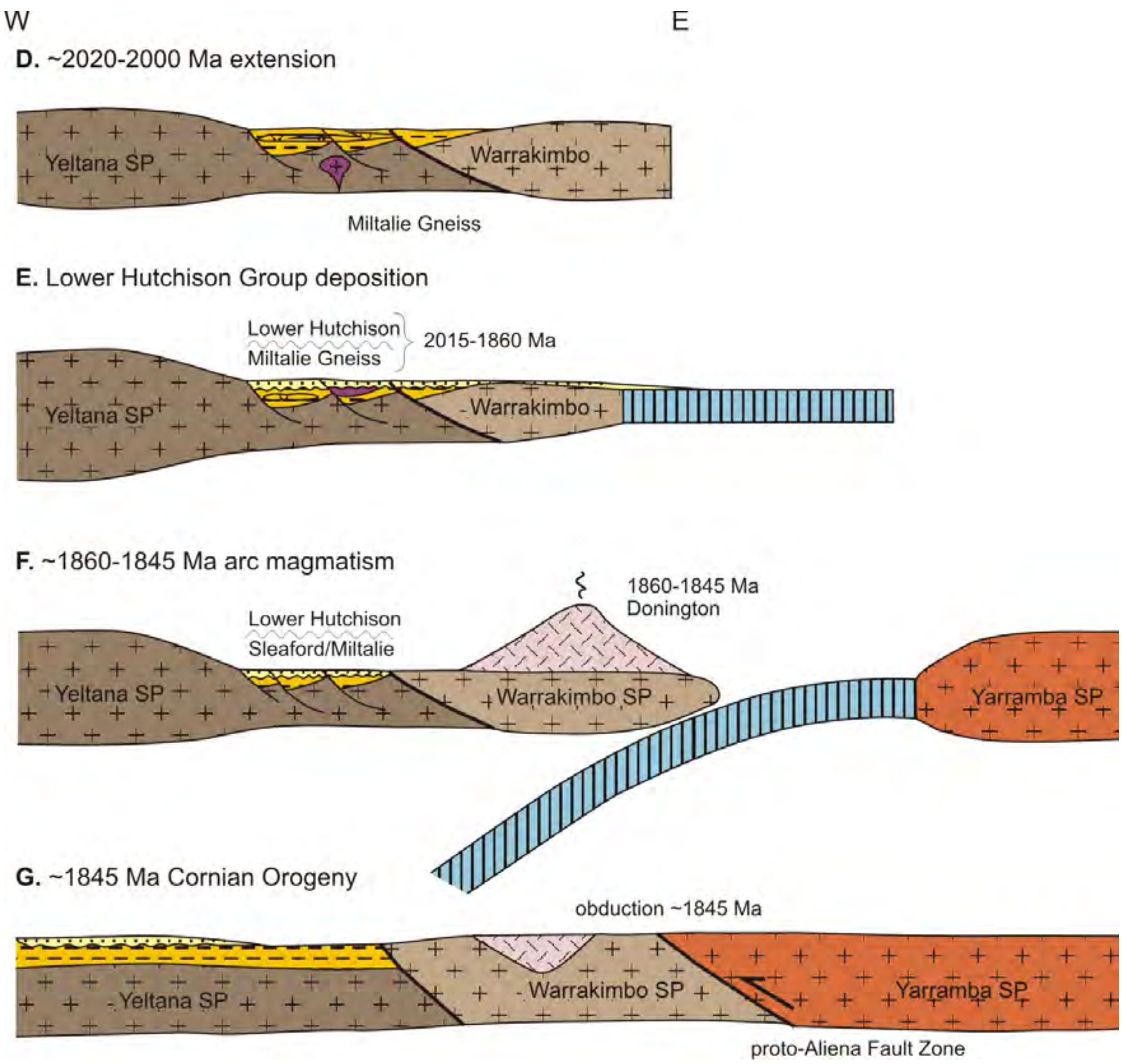

Figure 26 continued. Schematic cross sections showing the proposed tectonic development of the Gawler Craton and Curnamona Province from ~2020 Ma to ~1845 Ma.

~1900-1830 Ma

The lower Hutchison Group (<2000 Ma) was deposited on a shallow continental shelf, forming a passive margin setting in this interval (Figure 26E). We suggest that a west-dipping convergent margin commenced at the eastern margin of the Warrakimbo Province at >1850 Ma. We support the magmatic arc to collisional interpretation for the Donington Suite (e.g., Ferris et al., 2002), but prefer a subduction zone that was located to the east (Figure 26F). Thus, a continental margin arc, or backarc (given the chemistry), for the Donington Suite is envisaged, with west-dipping oceanic crust (attached to the

Yarramba Province of Korsch et al., 2010b; Figure 26F) subducting beneath the Gawler Craton (Warrakimbo Province of Korsch et al., 2010b). Subduction was followed by collision at ~1845 Ma, recorded by the Cornian Orogeny (Reid et al., 2008b). Collision and obduction of the Yarramba Seismic Province formed the proto-Aliena Fault Zone (Figure 26G). The Aliena Fault Zone represents a major crustal boundary, with significantly different seismic character of the lower to middle crust on either side, and Korsch et al. (2010b) indicate that it could represent a zone across which there was an amalgamation of terranes, prior to deposition of the Willyama Supergroup, to form the deep crust beneath this part of South Australia.

Volumetrically significant magmatism at ~1850 Ma occurs in broadly linear north-south belts along the eastern margin of the Gawler Craton (Donington Suite) and the inferred eastern margin of the NAC (Kalkadoon Supersuite, in the Mount Isa Province). We speculate (c.f. Fraser et al., 2007b) that these rocks form part of a possibly previously-continuous magmatic belt along the eastern margin of Proterozoic Australia, or they could represent arcs on two separate continents. This hypothesis requires testing via more detailed comparison of the geochemistry, Nd isotopic signature and inherited zircon populations of the Donington Suite compared with the Kalkadoon Supersuite, although Mortimer et al. (1998) suggested they were chemically similar, but did not favour an arc environment. Recent work by Bierlein et al. (2010) indicate that granites of the Kalkadoon Supersuite are characterised by late Archean to Paleoproterozoic crustal residence ages (TDM ca. 2.3 to 2.6 Ga), with the geochemical data suggesting strong within-plate affinities, implying that the Kalkadoon Supersuite intrusions underwent a greater degree of crustal assimilation and/or were emplaced further inboard of the active subduction margin located further to the east.

~1800-1550 Ma

Following cessation of the convergent Cornian Orogeny, several phases of sedimentation occurred in several domains in the Gawler Craton (Parker, 1993; Daly et al., 1998; Ferris et al., 2002). The Tournefort dykes (1812 ± 5 Ma; Schaefer, 1998), most likely heralded the start of a long-lived series of extensional events, from 1812 Ma to ~1685 Ma, recorded by sedimentation and contemporaneous bimodal magmatism, with sedimentation stepping (younging) eastwards (Broadview Schist, Wallaroo Group, Barossa Complex and Willyama Supergroup; Figure 21). This extension-dominated system may have developed as a series of slab rollback-initiated, extensional basins, which formed well inboard of a subduction zone, possibly in a backarc setting, as suggested by Reid et al. (2008b). This scenario is consistent with that of Conor (2009), who suggested that the easterly younging sedimentation implies basin migration to the east, and rollback of an oceanic slab further to the east is suggested as the cause of crustal attenuation, and hence, backarc basin formation.

Recent geodynamic models have suggested that the depositional environment for the Willyama Supergroup could have been an extensional backarc basin (e.g., Giles et al, 2002; 2004; Reid et al., 2008a; Conor, 2009). We place the Yeltana, Warrakimbo and Yarramba Seismic Provinces (and overlying basins, including the Wallaroo Group, Barossa Complex, and Willyama Supergroup) into a backarc regime, with a west-dipping subduction zone located to the east (Figure 26H). Giles et al. (2002, 2004), Fraser et al. (2007), and Gibson et al. (2008), all have suggested that the extension occurred in a backarc environment, and was part of a larger series of basins that include the Mount Isa Province and Georgetown Region in north Australia.

In the Curnamona Province, there is a record of magmatic activity (bimodal: mafic Fe-rich tholeiites and felsic magmatism), associated with sedimentation, from about ~1720 Ma to 1685 Ma. Based on geochemistry of the mafic units, Rutherford et al. (2006) and Raveggi et al. (2006) have suggested that the amount of lithospheric thinning was considerably greater under the Broken Hill Domain than under the Olary Domain. Raveggi et al. (2006) considered that maximum extension, that is, thinnest lithosphere, occurred at about 1685 Ma; again coincident with the cessation of mechanical extension. We prefer a backarc setting for deposition of the Willyama Supergroup and associated bimodal magmatism (Figure 26H). The cessation of magmatism at about 1685 Ma (Raetz et al., 2002; Page et al., 2005a; Rutherford et al., 2006) was probably associated with the cessation of mechanical extension, and subsidence was then driven by thermal relaxation until at least 1640 Ma (Raveggi et al., 2006). The introduction of a significantly more juvenile source region for the ca. ≤1650 Ma uppermost part of the Willyama Supergroup (εNd values between -3 to 0), is interpreted by Barovich and Hand (2008) to mark a significant change in tectonic regime (from the lower Willyama Supergroup: εNd between -7 and -4; Barovich and Hand, 2008). This change may also reflect a switch in provenance (e.g., from west– continental to the east– arc dominated. The data suggest proximity of a relatively juvenile and as yet unidentified source, possibly an arc, to the present-day east of the basin (Barovich and Hand, 2008).

The Curnamona Province was under extension in the period 1720-1685 Ma, which overlaps with the timing of the Kimban Orogeny (~1740-1690 Ma) in the Gawler Craton. In our scenario, we have the Gawler Craton and Curnamona Province on the same (upper plate), thus placing the Kimban Orogeny in a backarc regime, and inferring that at least part of the Kimban Orogeny (at least from ~1720-1690 Ma) was extensional. Thus, if contraction occurred, it must have been in the earliest stages, but overall, it was an extension-dominated regime. The Tunkillia Suite was initiated late in the Kimban Orogeny, with intrusive activity continuing to postorogenic time, that is, late to post-tectonic setting (Payne et al., 2010).

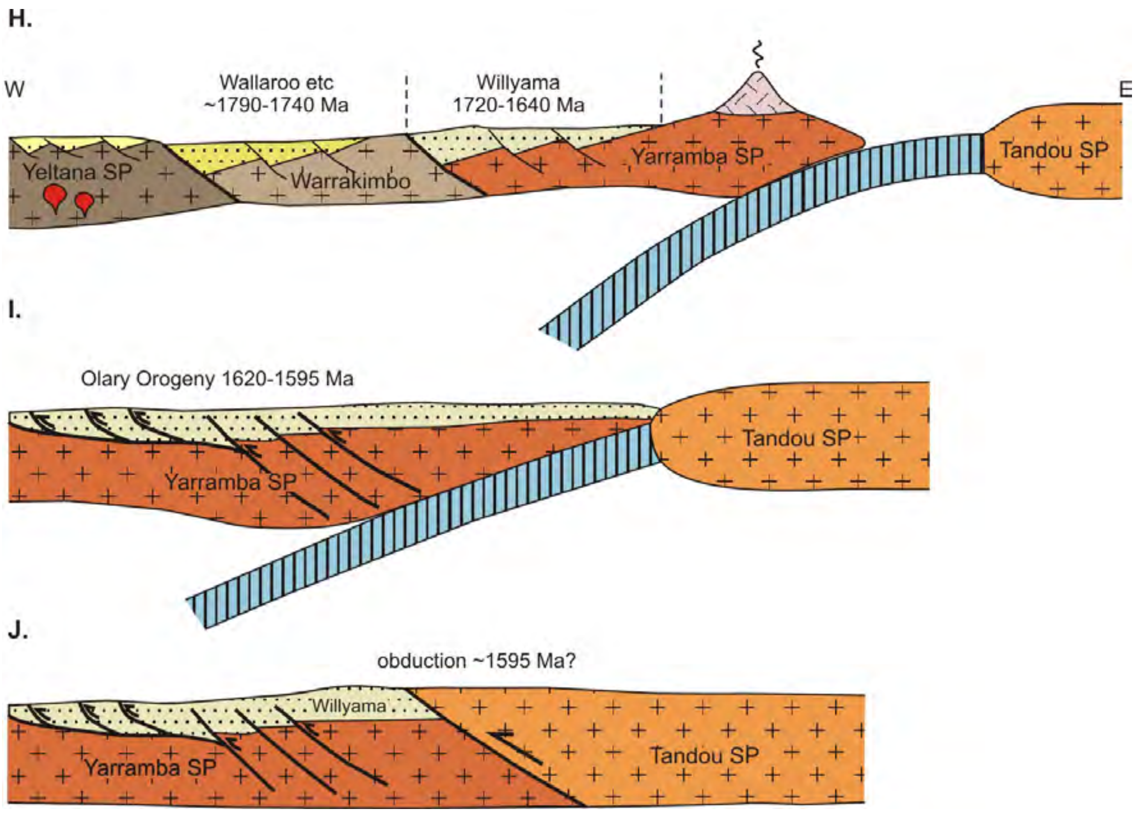

Figure 26 continued. Schematic cross sections showing the proposed tectonic development of the Gawler Craton and Curnamona Province from ~1790 Ma to ~1595 Ma.

At some time between ~1640 Ma and 1620 Ma, there was a change from extension to contraction in the upper plate. We correlate the Olarian Orogeny (1620-1595 Ma) in the Curnamona Province with the eventual collision, and obduction, of the Tandou Seismic Province with the Yarramba Seismic Province (Figure 26I, J). The boundary between the Yarramba and Tandou Seismic Provinces coincides approximately with strong, east-dipping reflections which penetrate the uppermost mantle. They possibly represent a crustal suture, with the Tandou Seismic Province being thrust over the Yarramba Seismic Province (Figure 26J; Korsch et al., 2010b).

3.3.2. Geodynamics associated with convergent processes to the southwest

The St Peter Suite (Ferris and Schwarz, 2004) consists of ca. 1631-1608 Ma magmatic rocks which were intruded into the Nuyts Domain of the southwest Gawler Craton forming a suite of I-type calcalkaline tonalite to granodiorite, similar in chemistry to Archean tonalite-trondhjemite-granodiorite suites. Knight (1997) and Swain et al. (2008) inferred that they formed as an island arc on oceanic crust, outboard of the current Gawler Craton with a subduction zone that dipped to the south under lithosphere that is now preserved in Antarctica. In this model, subduction ceased at 1608 Ma when the arc collided with the craton (Figure 27A). Ferris and Schwarz (2004) favoured a continental margin magmatic arc setting for the St Peter Suite, and Ferris et al. (2002) and Betts et al. (2009) considered that it was emplaced at the southwest margin of the Archean Gawler nuclei during subduction that dipped to the northeast (Figure 27B). The geochemical and isotopic signature of the St Peter Suite, however, appear more consistent with the possibility that it was a magmatic arc which formed at the edge of a continental margin (that is, the St Peter Suite compares more closely with modern day volcanics from continental margin arcs; Korsch et al., in review).

There is geochronological evidence to indicate that Terra Adélie and King George V Land in East Antarctica and the Gawler Craton were contiguous, forming the Mawson Continent of Fanning et al. (1996), from at least 2440 Ma until at least the end of the Kimban Orogeny at ca. 1680 Ma (Oliver and Fanning, 1997; Peucat et al., 1999; Di Vincenzo et al., 2007). Hence, it is likely that seafloor spreading occurred during the period 1680 Ma to 1630 Ma, forming a wide ocean to separate the two continents, prior to the onset of convergence and initiation of subduction at ca. 1630 Ma (Figure 27). We suggest that the subduction zone dipped to the northeast (c.f. Ferris and Schwarz, 2004; Cawood and Korsch, 2008; Figure 27).

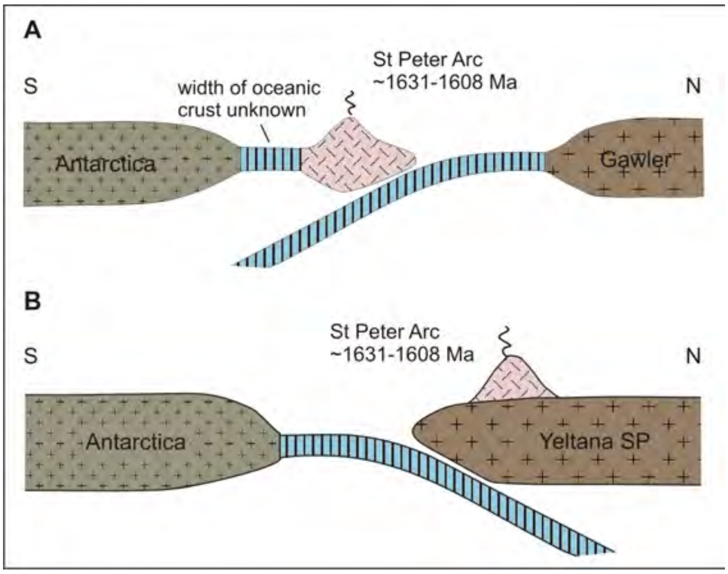

Figure 27. (a) Cartoon cross section showing tectonic scenario at ca. 1631-1608 Ma, during formation of the St Peter island arc (after Swain et al., 2008). (b) Cartoon cross section showing alternative tectonic scenario (Ferris and Schwarz, 2004), with the St Peter arc in a continental margin setting at the southwest margin of the Gawler Craton.

3.3.3. Geodynamics related to convergent processes to the north

The northern Gawler Craton occupies an important place in the architecture of Proterozoic Australia, in that this area links the Mesoarchean core of the central Gawler Craton to the comparatively juvenile Mesoproterozoic Musgrave Province. Due to an almost complete lack of outcrop and thick Neoproterozoic to Holocene cover, however, little is known about the Paleoproterozoic basement rocks from the northern Gawler Craton (Howard et al., 2010b).

In the Musgrave Province of central Australia, age and geochemical constraints of early Mesoproterozoic rocks are poor due to later overprinting tectonic events, but felsic orthogneisses (ca. 1607-1565 Ma) possibly represent juvenile felsic crust which was emplaced through subduction-related processes into an oceanic island arc (Wade et al., 2006). Values for ε_{Nd} (at 1550 Ma) determined by Wade et al. (2006) of +0.9 to -1.2 attest to the relatively juvenile character of these rocks. The mafic protolith gneisses compare well with the field for modern island arcs, whereas the felsic orthogneisses have a signature more comparable with modern continental margin arcs, and a continental environment is probably more consistent with the observed range in HREE contents (Korsch et al., in review). Wade et al. (2006) considered that the island arc was associated with a south-dipping subduction zone to the north of the Gawler Craton, and they proposed that the closure of the ocean basin eventually led to suturing of the North Australian Craton and South Australian Craton at 1590-1550 Ma. An alternative interpretation by Direen et al. (2005) has the Nawa Domain and Mabel Creek Ridge (both now part of the Gawler Craton) being located outboard of the main Gawler Craton at ca. 1600 Ma, and being amalgamated between ca. 1600 and 1550 Ma. However, the presence of Hiltaba-aged granites in the Nawa Domain suggests that these domains were continuous at ~1600 Ma and, therefore, appears to negate this suggestion.

We tentatively follow Wade et al. (2006), in part, with a south-dipping subduction zone operating from at least ca. 1607 Ma to the north of the Musgravian island arc or continental sliver (Figure 28A). An oceanic backarc (marginal sea) was present to the south of the island arc/continental sliver. Consumption of the oceanic crust eventually led to the suturing of the island arc to the North Australian Craton. We consider that the initiation of the Chewings Event in the southern Arunta Region is a reflection of this collision, with strong north-south directed deformation and zircon metamorphic ages from 1594 ± 6 Ma (Williams et al., 1996) to 1562 ± 4 Ma (Rubatto et al., 2001). The collision was probably initiated at about 1594 Ma and, following the initial contact, polarity of the subduction zone is interpreted to have flipped, with the oceanic crust of the marginal sea being consumed in a north-dipping subduction zone that dipped beneath the Musgravian arc (now proto-continental in setting), which was now being sutured to the North Australian Craton (Figure 28B). In the Gawler Craton, this resulted in the Kararan Orogeny of Hand et al. (2007). Deformation and associated high T-low P metamorphism of the Chewings Event continued until about 1562 Ma in a backarc setting. Arc magmatism continued in the Musgrave Province until the marginal sea was consumed, and the South Australian Craton collided with the expanded North Australian Craton. Younger granitic protolith phases in the Musgravian Gneiss (1550-1500 Ma) are tentatively related to collision and post-collision activity.

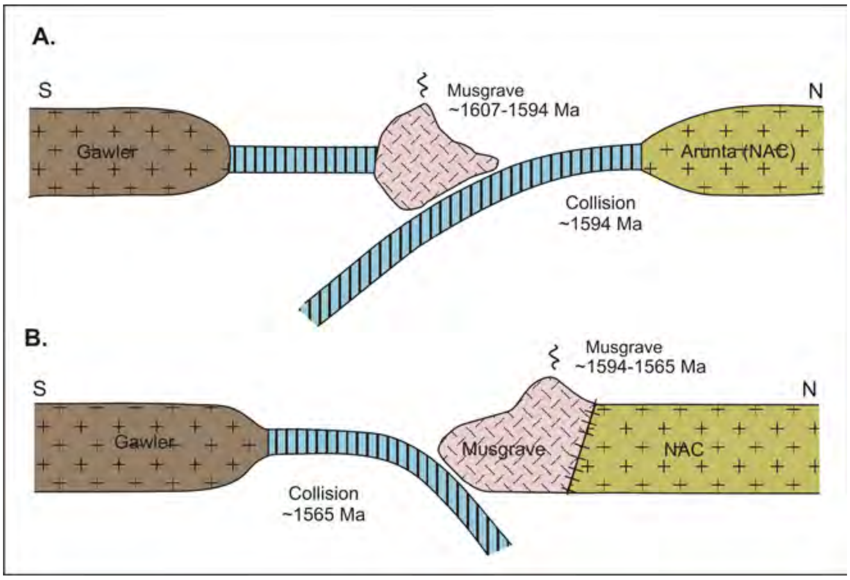

Figure 28. (a) Cartoon cross section showing tectonic scenario at ca. 1607-1594 Ma, during formation of the Musgrave island arc or continental sliver. (b) Cartoon cross section showing tectonic scenario following collision and amalgamation of the Musgrave Province with the North Australia Craton (modified after Wade et al., 2006).

A number of tectonic models for the setting of the Gawler Range Volcanics-Hiltaba Suite have proposed a backarc setting behind an active subduction zone. Ferris et al. (2002) inferred a backarc continental setting behind a northeast dipping subduction zone located to at the southwest margin of the Gawler Province. Giles et al. (2004), Betts and Giles (2006) and Rutherford et al. (2007) proposed that the Olarian Orogeny and Gawler igneous activity occurred in a backarc environment with subduction to the east or northeast beneath the combined South Australian-North Australian cratonic segment of protoAustralia. Alternatively, Swain et al. (2004) suggested that the Hiltaba Suite was related to south-dipping subduction under the northern margin of the Gawler Craton, resulting in the formation of the Hiltaba Suite in an extensional setting in a near margin continental backarc setting associated with the Musgrave magmatic arc. In our model, we consider that the Hiltaba event occurred in a backarc setting to the west of the thrust belt which formed during the Olarian Orogeny in the Curnamona Province to the east. Olarian deformation ceased by ~1595 Ma, which corresponds to the time of the start of the Gawler Range Volcanics-Hiltaba event in the Gawler Craton. We therefore envisage a switch from contraction to extension in the upper plate to accommodate the Gawler Range-Hiltaba event in an extensional regime.

The metamorphism and deformation observed in the Nawa Domain (including Mabel Creek Ridge, and Coober Pedy Ridge) is possibly related to events at the northern margin of the Gawler Craton (as mentioned above). We consider that granulite-grade metamorphism in the northern Gawler Craton reflects the collision between the South Australian Craton and the North Australian Craton; metamorphic ages for granulite grade rocks of 1565 ± 8 Ma in the Coober Pedy Ridge and ca. 1550 Ma in the Mabel Creek Ridge have been reported by Daly et al. (1998). Protoliths of the Musgravian gneiss aged between 1565 Ma and 1528 Ma are tentatively interpreted as collisional granites (Korsch et al., in review).

Sedimentation at Mount Painter, in the northern Curnamona Province, occurred at or after ~1595-1590 Ma (Fanning et al., 2003; Ogilvie, 2006; Shafton, 2006; Neumann et al., 2009; Fraser and Neumann, 2010). A-type granites and felsic volcanic successions were emplaced at ca. 1575 Ma and ~1565-1555 Ma (Teale, 1993a; Fanning et al., 1998; McLaren et al., 2006; Neumann et al., 2009; Fraser and Neumann, 2010), with metamorphism at ~1555 Ma (Fraser and Neumann, 2010). Thus, Mount Painter is considered to have affinities with the northern Gawler Craton, and was also affected by collision of the North and South Australian Cratons.

After 1550 Ma, the Gawler Craton was in an intraplate setting; post-1500 Ma deformational and igneous activity in the Gawler Craton (e.g., Spilsby Suite) is related to intraplate activity (Cawood and Korsch, 2008). The reactivation of shear zones and associated terrane-scale cooling occurred between ca. 1470 Ma and 1450 Ma in the western Gawler Craton (Teasdale, 1997; Fraser et al., 2002; Tomkins et al., 2004; Swain et al., 2005b; Fraser and Lyons, 2006), and the reactivation of the shear zone system was associated with regional denudation of much of the Gawler Craton. Magmatism at ca. 1450 Ma in the Nawa Domain was accompanied by granulite facies metamorphism (Howard et al., 2010b), and represents the youngest known Mesoproterozoic event in the craton.

3.4. Concluding remarks

A number of workers (e.g., Hand et al., 2007; Betts and Giles, 2006; Betts et al., 2009; Payne et al., 2009) have highlighted the significant ambiguities that exist in understanding the geologic framework of the Gawler Craton and Curnamona Province. Principally, these ambiguities revolve around the timing and spatial distribution of many of the tectonic events, the tectonic settings of the major magmatic systems, and the crust-mantle evolution through time. This ambiguity in part reflects the limited basement outcrop, and has consequently hampered the development of tectonic models which describe the evolution of the Gawler Craton and Curnamona Province in a space-time framework. Large tracts of crust are buried or have been reworked during younger tectonic events; evidence for plate and terrane margins are cryptic and poorly resolved, and the preserved geological record is complex.

We have presented one possible scenario for the evolution of the central and eastern margins of the Gawler Craton and Curnamona Province, and acknowledge there are numerous possible interpretations which could fit the available data. The northern and southwest margins of the Gawler Craton have been interpreted by previous authors. Clearly, the complexity of the tectonic evolution of the Gawler Craton and Curnamona Province in Australian continental reconstructions requires further studies involving specific tests designed to discriminate between geodynamic scenarios, before any consensus regarding tectonic evolution is possible. It is hoped that future studies will provide some insight into poorly understood events within southern Australia, and will stimulate further investigations on continental reconstruction models for the Archean to Mesoproterozoic of the South Australian Craton.

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