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Correlations and reconstruction models for the 2500–1500 Ma evolution of the Mawson Continent

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Abstract: Continental lithosphere formed and reworked during the Palaeoproterozoic era is a major component of pre-1070 Ma Australia and the East Antarctic Shield. Within this lithosphere, the Mawson Continent encompasses the Gawler–Adélie Craton in southern Australia and Antarctica, and crust of the Miller Range, Transantarctic Mountains, which are interpreted to have assembled during *c*. 1730–1690 Ma tectonism of the Kimban–Nimrod–Strangways orogenies. Recent geochronology has strengthened correlations between the Mawson Continent and Shackleton Range (Antarctica), but the potential for Meso- to Neoproterozoic rifting and/or accretion events prevent any confident extension of the Mawson Continent to include the Shackleton Range. Proposed later addition (*c*. 1600–1550 Ma) of the Coompana Block and its Antarctic extension provides the final component of the Mawson Continent. A new model proposed for the late Archaean to early Mesoproterozoic evolution of the Mawson Continent highlights important timelines in the tectonic evolution of the Australian lithosphere. The Gawler–Adélie Craton and adjacent Curnamona Province are interpreted to share correlatable timelines with the North Australian Craton at *c*. 2500–2430 Ma, *c*. 2000 Ma, 1865–1850 Ma, 1730–1690 Ma and 1600–1550 Ma. These common timelines are used to suggest the Gawler–Adélie Craton and North Australian Craton formed a contiguous continental terrain during the entirety of the Palaeoproterozoic. Revised palaeomagnetic constraints for global correlation of proto-Australia highlight an apparently static relationship with northwestern Laurentia during the *c*. 1730–1590 Ma time period. These data have important implications for many previously proposed reconstruction models and are used as a primary constraint in the configuration of the reconstruction model proposed herein. This palaeomagnetic link strengthens previous correlations between the Wernecke region of northwestern Laurentia and terrains in the eastern margin of proto-Australia.

This chapter outlines the Palaeoproterozoic to early Mesoproterozoic tectono-thermal evolution of the Mawson Continent of Australia and Antarctica (Fig. 1). The Mawson Continent comprises the Gawler Craton, South Australia, and the correlative coastal outcrops (e.g. Cape Hunter and Cape Denison) of Terre Adélie and George V Land in Antarctica and various other terrains of East Antarctica (Fig. 1, Oliver & Fanning 1997; Goodge *et al*. 2001; Fitzsimons 2003). Perhaps the most notable feature of the Mawson Continent is its lack of exposure. Excluding the flat-lying *c*. 1590 Ma Gawler Range Volcanics, the Gawler Craton portion is estimated to contain <5% basement exposure in an area approximately 530 800 km^2 (slightly smaller than France). The Antarctic component of the Mawson Continent contains even less exposure. Despite the impediment of limited basement exposure, numerous tectonic reconstruction models have been proposed to account for the evolution of the Mawson Continent and its interaction with other Proterozoic terrains, particularly other portions of the current Australian continent (Borg & DePaolo 1994; Daly *et al*. 1998; Betts *et al*. 2002; Dawson *et al*. 2002; Fitzsimons 2003; Giles *et al*. 2004; Betts & Giles 2006; Wade *et al*. 2006).

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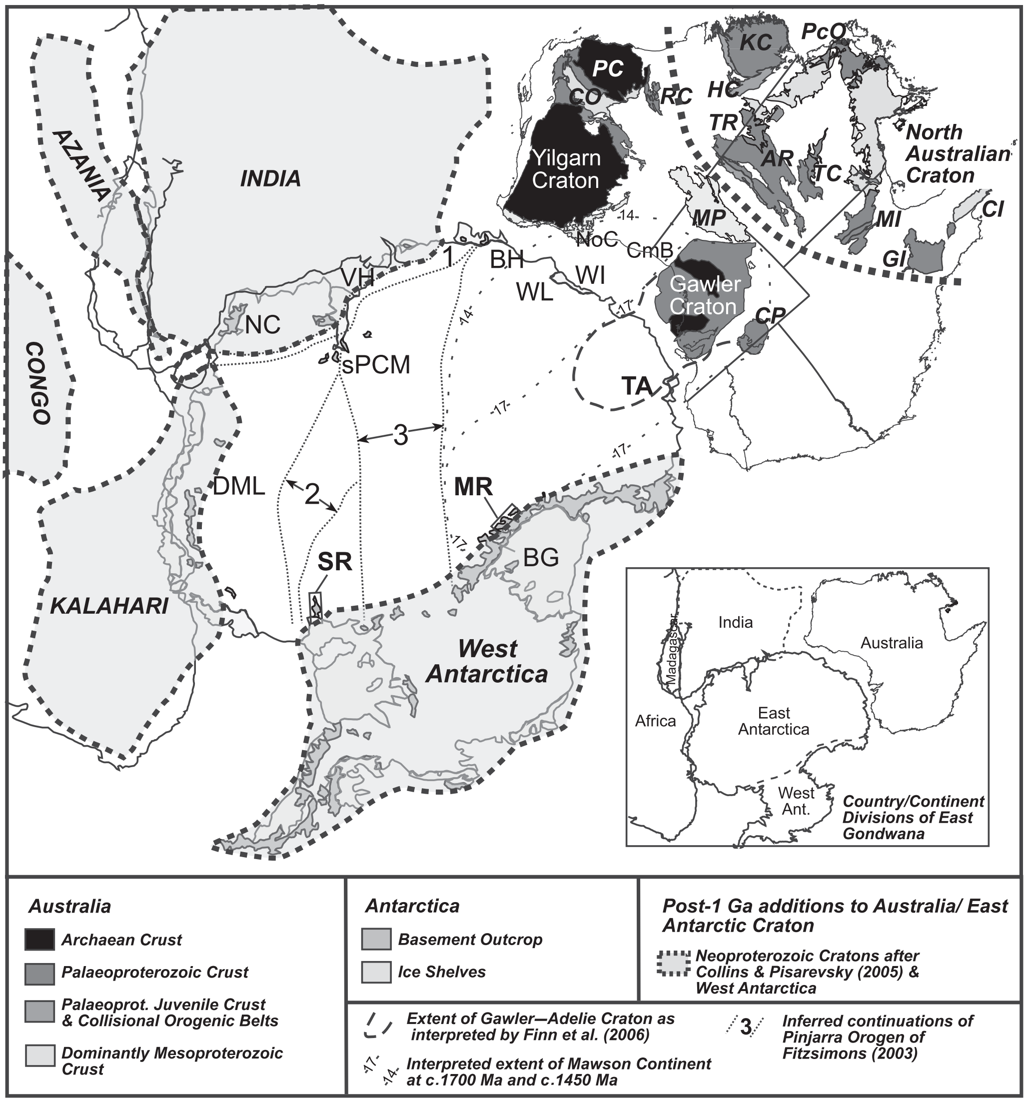

Fig. 1. Map of East Gondwana (modified from Collins & Pisarevsky 2005) displaying pre-Gondwana terrain locations in Antarctica (after Boger *et al*. 2006) and pre-1 Ga crustal provinces of Australia (after Betts *et al*. 2002; Payne *et al*. 2008). East Antarctica terrains are: BG, Beardmore Glacier; BH, Bunger Hills, DML, Dronning Maud Land; NC, Napier Complex, sPCM, southern Prince Charles Mountains; VH, Vestfold Hills, WI, Windmill Islands; WL, Wilkes Land. Bold abbreviations are MR, Miller Range; SR, Shackleton Range; and TA, Terre Adélie Craton, which have all experienced *c*. 1700 Ma tectonism (see text). Australian pre-1070 Ma terrains are: AR, Arunta Region, CI, Coen Inlier; CmB, Coompana Block; CO, Capricorn Orogen; CP, Curnamona Province; GI, Georgetown Inlier; HC, Halls Creek Orogen; KC, Kimberley Craton; MI, Mt Isa Inlier; MP, Musgrave Province; NoC, Nornalup Complex; PC, Pilbara Craton; PcO, Pine Creek Orogen; RC, Rudall Complex; TC, Tennant Creek Region; and TR, Tanami Region.

In attempting to reconstruct the evolution of the Mawson Continent, a particularly intriguing aspect of the geology of the Mawson Continent, and the Australian Proterozoic in general, is the comparative lack of evidence for subduction-related magmatism. The few Australia-wide examples of documented late Palaeoproterozoic and early Mesoproterozoic subduction-related magmatism can be summarized as follows: *c*. 1850 Ma magmatism associated with the accretion of the Kimberley Craton (Sheppard *et al.* 1999); volumetrically minor granites of the 1770–1750 Ma

Calcalkaline–Trondhjemite (CAT) Suite in the Arunta Region (Foden *et al*. 1988; Zhao & McCulloch 1995); the bimodal 1620–1600 Ma St Peter Suite of the Gawler Craton (Swain *et al*. 2008) and 1600–1550 Ma magmatism in the Musgrave Province (Wade *et al*. 2006). Reconnaissance geochemical data indicates *c*. 1690 Ma magmatism in the Warumpi Province on the southern margin of the Arunta Region may also represent subduction-related magmatism (Scrimgeour *et al*. 2005). In contrast, Palaeoproterozoic orogenic belts preserved in Laurentia and Baltica are commonly associated with identifiable subduction-related magmatism (e.g. Gandhi *et al*. 2001; Theriault *et al*. 2001; Ketchum *et al*. 2002; Mueller *et al*. 2002; Ansdell 2005; Whitmeyer & Karlstrom 2007; Åhall & Connelly 2008 and references therein). In addition, these orogenic belts are commonly quasi-linear belts often with associated inverted back-arc basins, accreted micro-continents and island arcs (e.g. Ketchum *et al*. 2002; St-Onge *et al*. 2006; Åhall & Connelly 2008 and references therein). This is in stark contrast to many orogenic events within Palaeo- to Mesoproterozoic Australia (e.g. summary of Betts & Giles 2006), which lack these elements and are commonly diffuse craton-wide events.

An example of the complex tectonic systems preserved in the Mawson Continent and Australia is the 1730–1690 Ma Strangways and Kimban orogenies in the Arunta Region and Gawler Craton, respectively. The Strangways Orogeny is preceded by the interpreted subduction-related CAT Suite magmatism, and forms a cornerstone of the argument for a long-lived accretionary system on the southern margin of the North Australian Craton (e.g. Betts & Giles 2006). However, the Strangways Orogeny has a very limited east–west extent and, as discussed later, is not easily reconcilable with an east–west trending accretionary margin. The temporally equivalent Kimban Orogeny does not preserve evidence for subduction-related magmatism (Hand *et al*. 2007) and 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.

The style of many tectonic events within the Palaeo- to Mesoproterozoic of Australia (McLaren *et al*. 2005) suggest that long-lived, pseudo-linear continental margins such as Phanerozoic Andean or Caledonian systems are not readily reconcilable with the geological record of the Australian Proterozoic. This appears to be a fundamental difference with continents such as Laurentia. While these differences are obviously generalized, reconstruction models for the Mawson Continent and Australia must take into account the nature of tectonic events within Australia, and be driven by available geological constraints rather than *a priori* postulated plate-tectonic models. McLaren *et al*. (2005) goes some way towards explaining many of the phenomena of the Australian Proterozoic by attributing high geothermal gradients and the predominance of high temperature–medium- to low-pressure metamorphism to the high heat producing nature of the North Australian Palaeoproterozoic crust. Although it seems likely that high heat production played a role in shaping the character of the Mawson Continent and Australia, it does not resolve many of the issues surrounding existing reconstruction models for the evolution of the Mawson Continent and Australia. This review provides a revised model for the 2500–1500 Ma evolution of the Mawson Continent for the purpose of outlining event correlations in associated terranes, thus enabling the revision of continent reconstruction models for the 2500–1500 Ma period.

The Mawson Continent

The name 'Mawson Continent' was first used to describe the Archaean–Mesoproterozoic southern Australian Gawler Craton and correlated terrains in Antarctica (Fanning *et al*. 1996). Alternative nomenclature has since included 'Mawson Block' (Oliver & Fanning 1996; Wingate *et al*. 2002*a*; Finn *et al*. 2006; Mikhalsky *et al*. 2006) and 'Mawson Craton' (Condie & Myers 1999; Fitzsimons 2003; Bodorkos & Clark 2004*b*). As 'continent' has first precedence and is a non-genetic descriptor, we favour the use of 'Mawson Continent' over alternative names.

The extent of the Mawson Continent is uncertain due to the extensive Neoproterozoic to Phanerozoic cover in Australia, and ice and snow cover in Antarctica. The Gawler Craton and the directly correlated coastal outcrops of Terre Adélie and George V Land in Antarctica (Oliver & Fanning 1997), form the nucleus of the Mawson Continent. In addition to these regions, the unexposed Coompana Block in South and Western Australia (Fig. 1) is often considered part of the Mawson Continent (Condie & Myers 1999; Bodorkos & Clark 2004*a*). In Antarctica, the Mawson Continent is commonly extended to include Palaeoproterozoic crust in the Miller and Shackleton ranges of the Trans-Antarctic Mountains (Fanning *et al*. 1999; Goodge *et al*. 2001). A recent compilation of airborne and satellite magnetic geophysical data (Finn *et al*. 2006) has suggested that fundamental differences in crustal petrophysical properties exist between the Gawler and Adélie cratons on the one hand, and the Miller Range and remainder of the East Antarctic Shield on the other. This is supported by differing geological evolutions of the various terrains with the presence of *c*. 1700 Ma tectonism considered as evidence for a single continent in the late Palaeoproterozoic period (Fanning *et al*. 1999; Goodge *et al*. 2001). In this review we adopt the terminology 'Mawson Continent' for the region encompassing the Gawler Craton, Terre Adélie Craton, Miller Range and Coompana Block. The former three of these domains are presumed to have acted as a coherent crustal fragment during the Proterozoic and early to mid-Phanerozoic after initial amalgamation at *c*. 1700 Ma, with proposed later addition of the Coompana Block at *c*. 1600–1550 Ma. The Mawson Continent was subsequently divided during the breakup of Gondwana Land.

The following section presents the tectonic histories for the proposed components of the Mawson Continent, which forms the basis for ensuing discussion on its *c*. 2500–1500 Ma amalgamation and evolution.

The Gawler Craton

The Gawler Craton (Fig. 2) is composed of late Archaean–early Palaeoproterozoic supracrustal and magmatic lithologies which are surrounded, overlain and intruded by Palaeoproterozoic (2000–1610 Ma) and Mesoproterozoic (1590–1490 Ma) units (Daly *et al*. 1998; Ferris *et al*. 2002; Swain *et al*. 2005*b*; Fanning *et al*. 2007; Hand *et al*. 2007). Tectonic domains have been delineated for the Gawler Craton based upon the interpretation of Total Magnetic Intensity (TMI) and gravity datasets combined with available geological evidence (Fig. 3, Ferris *et al*. 2002). These domains largely represent variations in structural trends and extent of crustal re-working as opposed to fundamental terrane boundaries (Hand *et al*. 2007).

Late Archaean–Early Palaeoproterozoic. The late Archaean stratigraphy in the central Gawler Craton consists of metasedimentary, volcanic and granite-greenstone lithologies (*c*. 2560–2500 Ma, Daly & Fanning 1993; Swain *et al*. 2005*b*) that were deformed during the Sleafordian Orogeny (2460–2430 Ma, Daly *et al*. 1998; McFarlane 2006). The *c*. 2560–2500 Ma Devil's Playground Volcanics and Dutton Suite are interpreted to have formed in a magmatic arc setting, which terminated shortly before or during the Sleafordian Orogeny (Swain *et al*. 2005*b*). Sleafordian Orogeny magmatism includes the Kiana Granite suite (*c*. 2460 Ma, Fanning *et al*. 2007) and leucogranites of the Whidbey Granite (*c*. 2445 Ma, Jagodzinski *et al*. 2006). Metamorphic grade of the Sleafordian Orogeny ranges from sub-greenschist to granulite facies (Daly & Fanning 1993). Peak metamorphism is recorded by P–T estimates of 800–850 °C and *c*. 7.5 kbar (Tomkins & Mavrogenes 2002) and 750–800 °C and 4.5–5.5 kbar (Teasdale 1997) for localities within the Mulgathing Complex (Fig. 2). Sleafordian Orogeny-aged structures in the central Gawler Craton (Mulgathing Complex) consist of shallowly NNE–NE plunging folds which have been subjected to some degree of block rotation by later shear zone movement (Teasdale 1997; Direen *et al*. 2005).

Circa 2000 Ma Miltalie Event. The Miltalie Event represents the first recognized tectonic activity after approximately 400 Ma of tectonic quiescence following the Sleafordian Orogeny (Webb *et al*. 1986; Daly *et al*. 1998). The Miltalie Gneiss has protolith magmatic ages of 2002 ± 15 Ma and 1999 ± 13 Ma (Fanning *et al*. 2007). Our field observations indicate that the Miltalie Gneiss map unit (Parker 1983) incorporates apparently metasedimentary lithologies; however, the age of this sequence is yet to be determined, and the tectonic setting of the Miltalie Gneiss and its protoliths has not yet been constrained.

2000–1860 Ma sediment deposition and the c. 1850 Ma Cornian Orogeny. The Miltalie Gneiss is overlain by sequences of the Hutchison Group which are interpreted to have been deposited on a passive margin in the time interval 2000–1860 Ma (Parker 1993; Schwarz *et al*. 2002). Final sedimentation prior to the onset of the Cornian Orogeny is relatively tightly constrained by the Bosanquet Formation volcanics at 1866 ± 10 Ma (Fanning *et al*. 2007). Apparently time equivalent sediment deposition is also evident in the Corny Point region of the Yorke Peninsula (Howard *et al*. 2007). The Cornian Orogeny (1850–1840 Ma, Reid *et al*. 2008) has associated voluminous felsic magmatism of the Donington Suite (Hoek & Schaefer 1998; Reid *et al*. 2008). This orogenic event produced ESE striking structural fabrics overprinted by east–west striking folds and late south-side down extensional ductile shearing. Metamorphism associated with the Cornian Orogeny is represented by a clockwise P–T path with peak metamorphic conditions of *c*. 750 °C and *c*. 6 kbar (Reid *et al*. 2008). The Donington Suite intrusions and Cornian Orogeny appear to be restricted to east of the Kalinjala Shear Zone in the southern Gawler Craton (Fig. 2). Temporally equivalent magmatic lithologies also exist as basement in the Olympic Dam region in the central eastern Gawler Craton (Fig. 2, Jagodzinski 2005), indicating the Cornian Orogeny system affected much of what is now the eastern Gawler Craton.

1800–1740 Ma magmatism and sedimentation. The 60 Myr period from approximately 1800–1740 Ma marks an interval of extensive sediment deposition

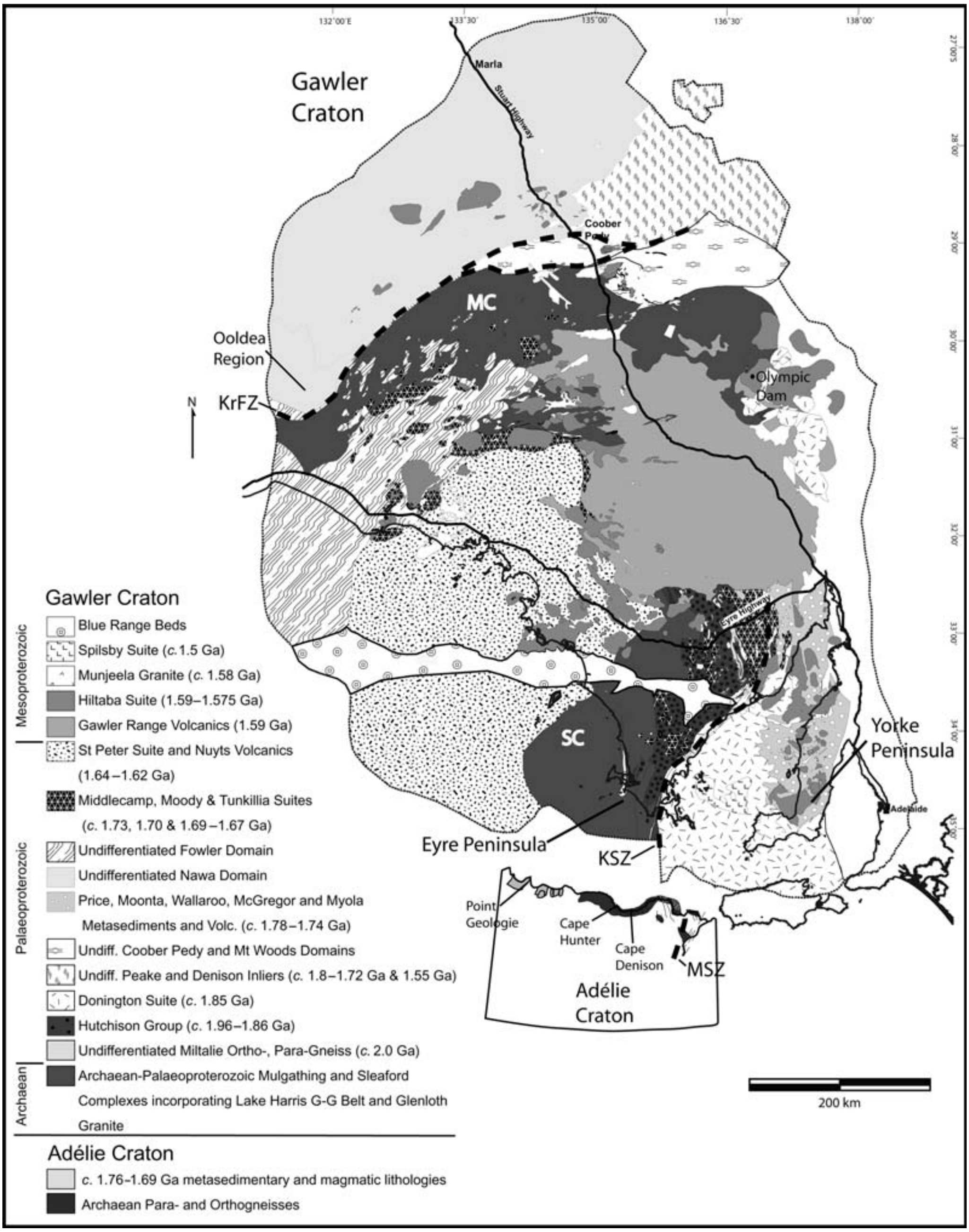

Fig. 2. Simplified geological map of the Gawler Craton and Adélie Craton in restored configuration. Relative positions of the two cratons after Oliver & Fanning (1997). Geology of the Gawler Craton after Fairclough *et al*. (2003). Geology of the Adélie Craton after Pelletier *et al*. (2002) and Ménot *et al*. (2005). Abbreviations are: KSZ, Kalinjala Shear Zone; KrFZ, Karari Fault Zone; MC, Mulgathing Complex; and SC, Sleafordian Complex.

across much of the Gawler Craton. In the southern Gawler Craton this includes the Myola Volcanics (1791 ± 4 Ma, Fanning *et al.* 1988), McGregor Volcanics and Moonabie Formation (*c.* 1740 Ma, Fanning *et al.* 1988), Wallaroo Group (Parker 1993; Cowley *et al.* 2003), Price Metasediments (*c.* 1770 Ma, Oliver & Fanning 1997) and successions previously incorporated into the Hutchison Group (Szpunar *et al.* 2006). Volcanic sequences in the Wallaroo Group constrain the age of deposition to 1772 ± 14 Ma (Wardang Volcanics, Fanning *et al.* 2007), 1753 ± 8 Ma (Moonta Porphyry rhyolite, Fanning *et al.* 2007) and 1740 ± 6 Ma (Mona Volcanics, Fanning *et al.* 2007). In the northern Gawler Craton, metasedimentary lithologies intersected in drillholes in the Nawa Domain are interpreted to have been deposited in the interval *c.* 1750–1730 Ma (Payne *et al.* 2006) and may correlate with the Peake Metamorphics within the Peake and Denison Inliers in the northeastern Gawler Craton (Hopper 2001). The depositional ages for the protoliths of the Peake Metamorphics are constrained by ages of 1789 ± 10 and 1740 ± 6 Ma for the Tidnamurkuna and Spring Hill volcanics, respectively (Fanning *et al.* 2007). The Wirriecurrie Granite in the Peake and Denison Inlier is constrained to 1787 ± 8 Ma (Parker 1993; Fanning *et al.* 2007) and is interpreted to have formed in an intracontinental setting that sampled a *c.* 2500 Ma subduction-modified mantle source (Hopper 2001).

1730–1690 Ma Kimban Orogeny. The Kimban Orogeny is interpreted as the most pervasive orogenic event in the Gawler Craton (Daly *et al.* 1998; Fanning *et al.* 2007; Hand *et al.* 2007; Payne *et al.* 2008). Geochronology has confirmed it to be a craton-wide event with metamorphic and syntectonic magmatic ages in the range 1730–1690 Ma reported from the southern (Vassallo 2001; Fanning *et al.* 2007), western (Fowler Domain, Teasdale 1997) and northern Gawler Craton (Hopper 2001; Betts *et al.* 2003; Payne *et al.* 2008).

Kimban Orogeny structures in the southern Gawler Craton formed during dextral transpression (Parker 1993; Vassallo & Wilson 2001, 2002) with NE trending-structures curving to north trends further north. In the northern Gawler Craton, within the Peake and Denison Inlier, early north–south trending structures are correlated with the Kimban Orogeny (Hopper 2001). Payne *et al.* (2008) suggest the Kimban Orogeny is expressed by the prominent NE trends in the regional aeromagnetic (TMI) data for the northwestern Gawler Craton.

Craton-wide metamorphic conditions of 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 1993; Tong *et al.* 2004). However, considerable variation in metamorphic grade is observed within the Kimban Orogeny as evidenced by the regions in the southern Gawler Craton that preserve greenschist-facies metamorphism (Price Metasediments, Oliver & Fanning 1997) adjacent to granulite regions.

Magmatism associated with the Kimban Orogeny is represented by the Middlecamp, Moody and Tunkillia suites. The Middlecamp Suite in the eastern Gawler Craton is a pre- to early Kimban Orogeny granite suite with ages in the range 1737 ± 7–1726 ± 7 Ma (Fanning *et al.* 2007). The Moody Suite is later in the Kimban Orogeny (1720 ± 9–1701 ± 12 Ma) and contains intrusives ranging from hornblende-bearing granitoids to muscovite-bearing leucogranites (Fanning *et al.* 2007). The Tunkillia Suite is constrained to 1690–1670 Ma (Ferris & Schwarz 2004) and interpreted to be a post-tectonic magmatic suite based upon Nd-isotope and trace element geochemistry (Payne 2008).

Localized syn-Kimban Orogeny basin formation is also preserved in the central Gawler Craton. Here the *c.* 1715 Ma Labyrinth Formation is typified by upward coarsening sequences (Cowley & Martin 1991). The depositional environment is interpreted to be within a fault-bounded basin with sediments derived from local sources (Daly *et al.* 1998). A rhyolite within the Labyrinth Formation constrains the timing of deposition to *c.* 1715 ± 9 Ma (Fanning *et al.* 2007).

1660 Ma Ooldean Event. The Ooldean Event as defined by Hand *et al.* (2007) is currently constrained to 1659 ± 6 Ma (Fanning *et al.* 2007) and is represented by UHT metamorphic conditions of *c.* 950 °C and 10 kbar (Teasdale 1997). The mineral assemblage associated with these P–T conditions defines a fine-grained mylonitic fabric that overprints a high-grade metamorphic assemblage (Teasdale 1997). The earlier assemblage is interpreted to represent Kimban Orogeny metamorphism (*c.* 1690 Ma, Teasdale 1997; Payne *et al.* 2008). The tectonic setting of the Ooldean Event is unconstrained and evidence for the event is currently confined to the Ooldea region (Fig. 2). Elsewhere, at *c.* 1660 Ma, the Gawler Craton appears to have undergone extension as evidenced by the deposition of the Tarcoola Formation (Cowley & Martin 1991; Daly *et al.* 1998), and also potentially sedimentary packages in the Mt Woods region (Fig. 2, Betts *et al.* 2003; Skirrow *et al.* 2006). Further petrographically constrained geochronology is required to determine if the *c.* 1660 Ma age obtained by

Fanning *et al.* (2007) does represent the age of UHT metamorphism, as this age has not been found in subsequent geochronology studies of the Ooldea 2 lithologies (Payne *et al.* 2008).

Palaeo-Mesoproterozoic transition events. The Gawler Craton preserves evidence of a complex sequence of events in the period from 1630–1540 Ma. The co-magmatic mafic and felsic intrusions of the St Peter Suite (1620–1608 Ma, Fig. 2, Flint *et al.* 1990) have been interpreted to have a subduction-related petrogenesis (Swain *et al.* 2008). St Peter Suite magmatism was followed by the voluminous and metallogenically significant Gawler Range Volcanics (GRV, *c.* 1592 Ma) and 1595–1575 Ma Hiltaba Suite intrusives (Flint 1993; Daly *et al.* 1998; Budd 2006). The GRV and Hiltaba Suite have previously been interpreted as an anorogenic magmatic event and linked to a plume (Flint 1993; Creaser 1995). However, new evidence of contemporaneous high-grade metamorphism and deformation in the Mt Woods and Coober Pedy Ridge domains (Skirrow *et al.* 2006; Fanning *et al.* 2007) has led to a suggestion of a syntectonic setting for the GRV–Hiltaba event. However, this syntectonic setting does not negate the potential role of a mantle plume in magma generation or necessarily require a direct causal link between the tectonism and magmatism (Betts *et al.* 2007). Deformation and low-grade metamorphism at this time is also reported (Direen & Lyons 2007; Hand *et al.* 2007) from the Eyre Peninsula region (Foster & Ehlers 1998) and in the Wallaroo Group (Conor 1995). Deformation and metamorphism associated with the GRV–Hiltaba event occurred shortly before, and within uncertainty of, the *c.* 1565–1540 Ma Kararan Orogeny (Hand *et al.* 2007).

The Kararan Orogeny, as defined by Hand *et al.* (2007), represents the final episode of high-grade metamorphism and deformation within the Gawler Craton (Teasdale 1997; Fraser & Lyons 2006; Fanning *et al.* 2007; Hand *et al.* 2007; Payne *et al.* 2008) before a final period of shear-zone activity and subsequent cratonization at *c.* 1450 Ma (Webb *et al.* 1986; Fraser & Lyons 2006). 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 regions (Fanning *et al.* 2007; Payne *et al.* 2008). East–west to NE-trending structures evident in regional aeromagnetic data of the northern Gawler Craton and in outcrop in the Peake and Denison Inliers, are interpreted to have formed during the Kararan Orogeny (Hopper 2001; Payne *et al.* 2008). The extent of shear zone activity at *c.* 1450 Ma (Fraser & Lyons 2006) and its influence of Gawler Craton geometry is yet to be constrained.

Archaean–Palaeoproterozoic Antarctica

Three regions of outcropping Palaeoproterozoic geology are commonly assigned to the Mawson Continent in Antarctica: Terre Adélie Craton, the Miller Range and the Shackleton Range (Fig. 1, Fitzsimons 2003; Finn *et al.* 2006). The Terre Adélie Craton (represented by outcrop in George V Land and Terre Adélie Land) represents the conjugate rifted margin of the Gawler Craton (Stagg *et al.* 2005). This correlation is supported by reconstruction of the rifted margin and correlation of Archaean to Palaeoproterozoic lithologies of the Eyre Peninsula, Gawler Craton, and George V and Terre Adélie Land (Oliver & Fanning 1997).

Terre Adélie Craton. The Terre Adélie Craton is known from *c.* 400 km of discontinuous outcrops along the coast of George V and Terre Adélie Land (Fig. 2, Peucat *et al.* 2002). The outcrop consists of late Archaean–Early Palaeoproterozoic gneisses with overlying Palaeoproterozoic metasedimentary lithologies that directly correlate to lithologies and tectono-thermal events of the Gawler Craton (Oliver & Fanning 1997). Granodiorites at Cape Denison yield a crystallization age of *c.* 2520 Ma (correlating with the Dutton Suite of the Gawler Craton) and, along with metasedimentary lithologies, were deformed during 2440–2420 Ma orogenesis (Monnier 1995). Garnet-cordierite granites, outcropping from Point Martin to Cape Denison, were generated during *c.* 2440 Ma orogenesis (Monnier 1995) and are similar to garnet-cordierite granites of the Whidbey Granite association (*c.* 2443 Ma, Daly & Fanning 1993; Jagodzinski *et al.* 2006). Deformation associated with *c.* 2440 Ma orogenesis displays a NW–SE trend associated with NE–SW shortening (Ménot *et al.* 2005).

Migmatitic Palaeoproterozoic metasedimentary lithologies at Pointe Geologie are intruded syntectonically by dolerites and gabbros, which show mingling relationships with anatectic melts (Peucat *et al.* 1999). The Cape Hunter Phyllite is suggested to be a stratigraphic equivalent which has only experienced greenschist facies metamorphism (Oliver & Fanning 1997). Structures within the Cape Hunter Phyllite are sub-vertical with an approximate north–south orientation (Oliver & Fanning 1997). The Point Geologie HT–LP event is constrained by monazite ages of 1694 ± 2 and 1693 ± 2 Ma (U–Pb TIMS multi-grain dissolution and evaporation, respectively, Peucat *et al.* 1999). U–Pb SHRIMP zircon ages constrain metamorphism to 1696 ± 11 Ma with a reported

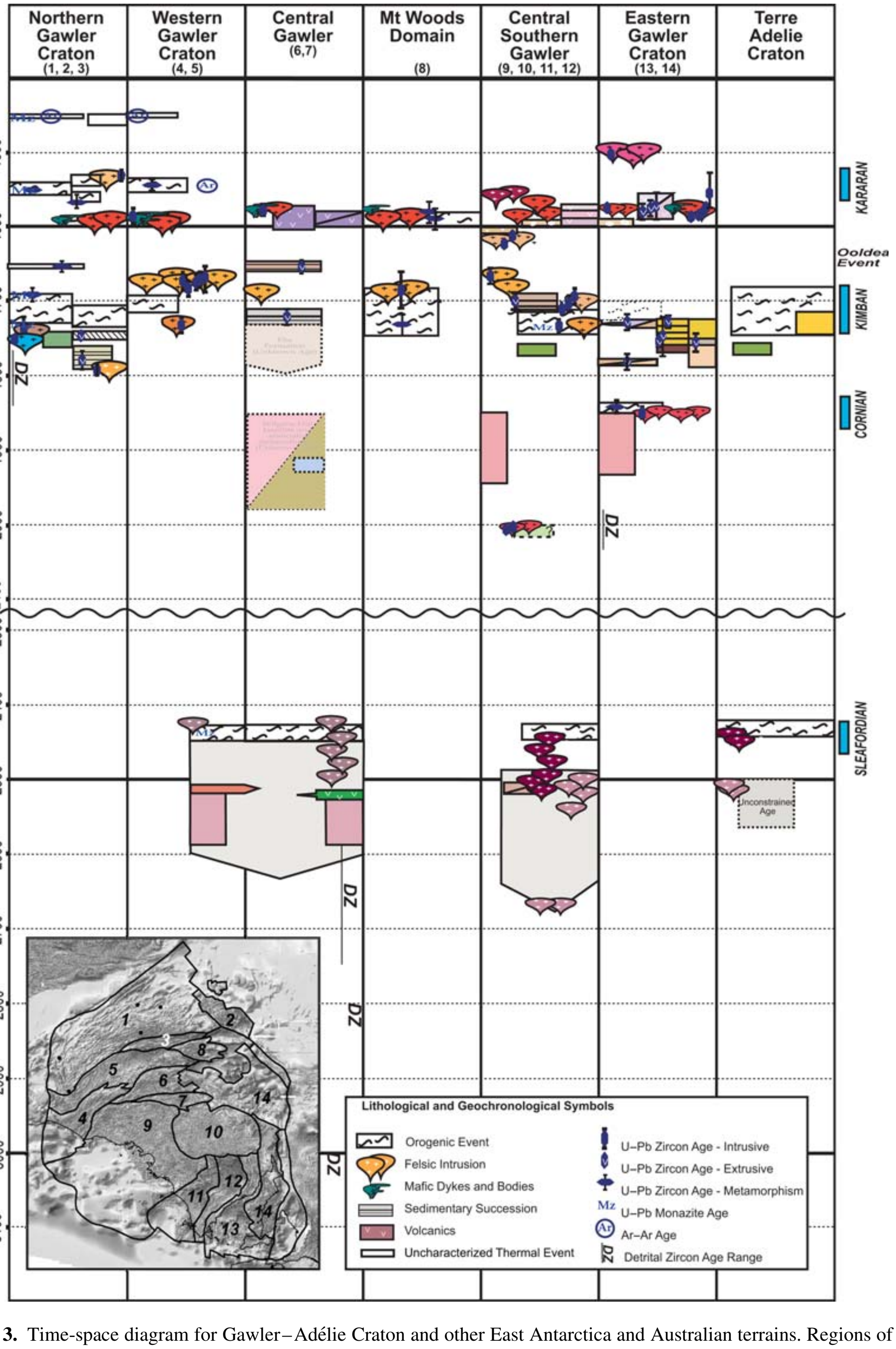

Fig. 3. Time-space diagram for Gawler–Adélie Craton and other East Antarctica and Australian terrains. Regions of the Gawler Craton correspond to domains as numbered and represented in inset map. Domains are: 1, Nawa Domain; 2, Peake and Denison Inliers; 3, Coober Pedy Ridge Domain; 4, Fowler Domain; 5, Christie Domain; 6, Wilgena Domain; 7, Lake Harris Greenstone Domain: 8, Mount Woods Domain; 9, Nuyts Domain; 10, Gawler Range Domain; 11, Coulta Domain; 12, Cleve Domain; 13, Spencer Domain; and 14, Olympic Domain. T-S plot for Gawler Craton

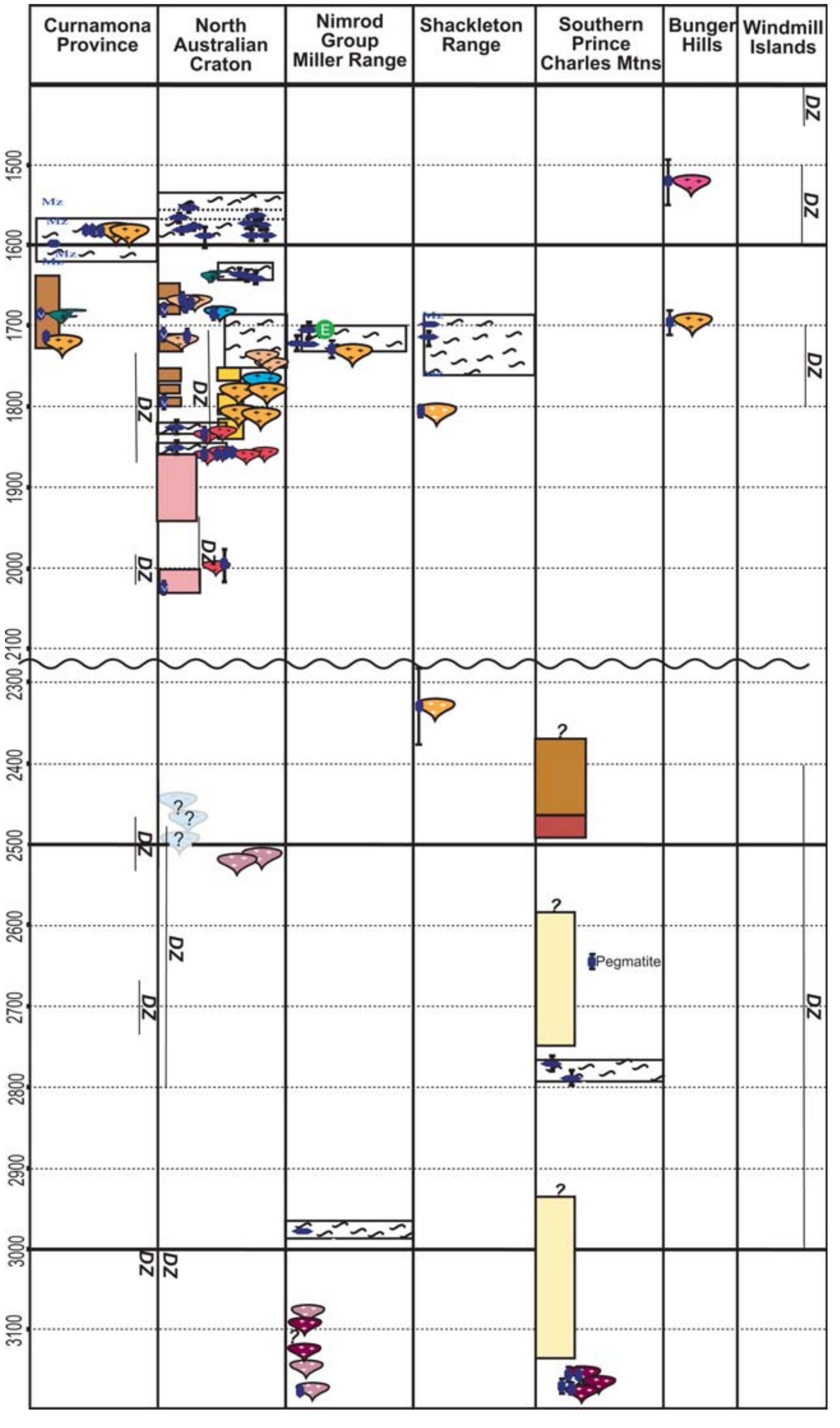

Fig. 3. (*Continued*) modified after Ferris *et al*. (2002) with additional data from Peucat *et al*. (1999, 2002), Holm (2004), Jagodzinski *et al*. (2006), Fraser & Lyons (2006), Payne *et al*. (2006), Howard *et al*. (2007) and Hand *et al*. (2007). Data sources for remainder of T-S plots are as discussed in text. North Australian Craton timeline is a simplified representation with greater detail provided in text and Figures 8 and 9. E, eclogite facies metamorphism.

maximum deposition age of 1740–1720 Ma from zircon cores (Peucat *et al.* 1999). This suggests the metasediments of the Point Geologie region were deposited in a basin setting associated with the development of the early Kimban Orogeny (after *c.* 1720 Ma, potentially equivalent to the Labyrinth Formation) and were metamorphosed under HT–LP conditions during the late Kimban Orogeny (*c.* 1690 Ma).

Miller Range. The Nimrod Group is a relatively localized group of Archaean–Palaeoproterozoic exposures in the Miller Range within the Trans-Antarctic Mountains (Fig. 1). The Nimrod Group includes quartzofeldspathic, mafic and calc-silicate gneiss, pelitic schist, amphibolite, orthogneiss, along with relict eclogite and ultramafic pods (Goodge *et al.* 2001).

Protolith U–Pb zircon ages from the gneisses are in the age range 3290–3060 Ma with metamorphism at *c.* 2975 Ma (Bennett & Fanning 1993; Goodge & Fanning 1999). These U–Pb data agree with Nd-isotope evidence (Borg & DePaolo 1994) for initial crustal growth in the mid- to late-Archaean (Goodge *et al.* 2001). Late Palaeoproterozoic metamorphism is constrained by U–Pb SHRIMP zircon ages of 1723 ± 14 and 1720 Ma for biotite-hornblende gneisses and 1723 ± 29 Ma for relict eclogites (Goodge *et al.* 2001). Syntectonic orthogneisses yield a U–Pb SHRIMP zircon age of 1730 ± 10 Ma. This orogenic event was termed the Nimrod Orogeny by Goodge *et al.* (2001). The presence of eclogitic material indicates a likely collisional-setting with the potential for a proximal terrane boundary (Kurz & Froitzheim 2002 and referencers therein). Structural information for the Nimrod Orogeny is largely overprinted and/or re-oriented by reworking during the Phanerozoic Ross Orogeny.

Shackleton Range. The lithologies of the Shackleton Range (Fig. 1) preserve a complex history of early Palaeoproterozoic to Cambrian tectonism. Links with the Nimrod Group and Gawler Craton are based upon the correlation of age-equivalent Palaeoproterozoic tectonism (Fanning *et al.* 1999; Goodge *et al.* 2001; Zeh *et al.* 2004). The southern Shackleton Range consists of upper amphibolite-granulite ortho- and paragneisses that were metamorphosed at 1763 ± 32 Ma as constrained by a Rb–Sr whole-rock isochron age (Schubert & Will 1994; Talarico & Kroner 1999). The northern Shackleton Range records extensive overprinting and tectonic juxtaposition by the *c.* 500 Ma (Pan-African) Ross Orogeny. Orthogneisses in the Haskard Highland and La Grange Nunataks preserve U–Pb zircon protolith ages of 2328 ± 47 and 1810 ± 2 Ma, respectively (Brommer *et al.* 1999; Zeh *et al.* 1999). Upper amphibolite to granulite facies metamorphism of these orthogneisses and associated paragneisses is constrained by a 1715 ± 6 Ma U–Pb zircon age from a syntectonic leucosome (Brommer *et al.* 1999) and a U–Pb monazite age of 1737 ± 3 Ma (Zeh *et al.* 1999). *P–T* data from these locations and the central-northern Shackleton Ranges indicate Barrovian-style metamorphism with peak conditions of 630–750 °C and 7–11 kbar (Schubert & Will 1994; Brommer *et al.* 1999; Zeh *et al.* 1999). Monazite and zircon from Meade Nunatak in the NE Shackleton Range record metamorphic ages of *c.* 1700 and 1686 ± 2 Ma, respectively (Zeh *et al.* 2004), suggesting it too was part of a regional *c.* 1730–1690 Ma tectono-thermal event. Interpreted cooling ages of 1650–1550 Ma, provided by Sm–Nd, Rb–Sr and K–Ar data, are reported from both the northern and southern Shackleton Ranges (Zeh *et al.* 2004).

Assembling the Mawson Continent

The regions considered as key components of the Mawson Continent (Gawler and Terre Adélie cratons, Miller Range and Shackleton Range) share few similar tectono-thermal events, with the basis for comparison being solely *c.* 1700 Ma tectonism (Fig. 3). The exceptions to this generality are the Gawler and Terre Adélie cratons which have coincident late Archaean to early Proterozoic histories. The numerous tectono-thermal and relatively precise geographical correlations between these two (Oliver & Fanning 1997) means they can be considered to have formed a continuous crustal block from the Archaean until Cretaceous rifting. Hence we propose the name of Gawler–Adélie Craton to refer to this terrain. In this section we discuss the validity of assigning Antarctic terrains to the Mawson Continent and the various proposals for the Palaeo- to early Mesoproterozoic extent of the Mawson Continent.

The Gawler–Adélie–Miller Range–Shackleton Range 1700 Ma connection

The near identical timing of Nimrod Group metamorphism at 1730–1723 Ma with early Kimban Orogeny metamorphism (1730–1720 Ma) is suggestive of a related and possibly contiguous tectono-thermal event involving both regions. Given the presence of relict eclogite (*c.* 1730 Ma, Goodge *et al.* 2001) and the lack of evidence for later terrane accretion, we suggest that the 1730–1690 Ma Kimban–Nimrod Orogeny records the accretion of the Miller Range terrain to the Gawler–Adélie Craton. The suture zone that accommodated this amalgamation is potentially at or near the location of the Nimrod Group, as suggested by the presence of eclogite-facies metamorphic lithologies within this sequence.

The Shackleton Range is located approximately 3500 km from outcrops of the Gawler–Adélie Craton and approximately 1800 km from the Nimrod Group in the Miller Range (Fig. 1). Despite these distances, the geochronology of Zeh *et al.* (2004) highlights a temporal correlation between *c.* 1730–1690 Ma tectonism in all three terrains. Nd-isotope data of Borg & DePaolo (1994) indicate a Palaeoproterozoic model age ($T_{DM} = 2.2–1.6$ Ga) for granite source regions along the Transantarctic Mountains from Victoria Land to the Beardmore Glacier region, in contrast to Mesoproterozoic model ages beyond the Beardmore Glacier (Fig. 1). This change in crustal evolution is utilized by Fitzsimons (2003) to suggest that one of three possible paths for the *c.* 550–500 Ma Pinjarra Orogen (Path 3, Fig. 1) bisects the Transantarctic Mountains between the Miller and Shackleton ranges, with the latter considered a Neoproterozoic or Cambrian addition to the proto-East Antarctic Shield. The non-unique nature of bulk-rock Nd-isotope data implies that Mesoproterozoic model ages may result from a variety of processes, and the Shackleton Range may still have formed part of the Mawson Continent. However, given the high degree of uncertainty, for the purposes of this review the Shackleton Range is excluded from the Mawson Continent.

The western extent of the Mawson Continent

The western extent of the Mawson Continent (eastern extent in Antarctica) is unclear. The Coompana Block and Nornalup Complex in Australia (Fig. 1, Fitzsimons 2003; Bodorkos & Clark 2004*a*) and Bunger Hills and Windmill Islands in Antarctica (Fig. 1, Fitzsimons 2000, 2003) have typically been assigned to the Mawson Continent. The vast majority of this region, marked on Figure 1, is unexposed, with the current level of geophysical characterization insufficient to adequately constrain potential crustal-scale terrane boundaries.

Of the above regions, the Bunger Hills is the only location to preserve pre-1500 Ma crust, with two magmatic protolith conventional U–Pb zircon ages of 1699 ± 15 and 1521 ± 29 Ma (Fig. 3, Sheraton *et al.* 1992). The Windmill Islands preserve metasedimentary units similar in age and provenance to lithologies of the Nornalup Complex (*c.* 1400–1340 Ma, see discussion of Fitzsimons 2003 and references therein). As summarized by Fitzsimons (2003), the Nornalup Complex of the Albany–Fraser Belt is predominantly composed of syn-orogenic granites (1330–1290 Ma) with preserved *c.* 1440 Ma zircon xenocrysts and paragneisses with an interpreted depositional range of *c.* 1550–1400 Ma (Nelson *et al.* 1995). The non-outcropping Coompana Block has a single chronological constraint of 1505 ± 7 Ma for a juvenile, anorogenic orthogneiss intersected by drillhole (Wade *et al.* 2007).

The vast majority of the Coompana–Albany–Fraser–Wilkes region appears to be composed of distinctly different crust to the Gawler–Adélie Craton. The aeromagnetic signature of the Coompana Block appears different to that of the Gawler–Adélie Craton, with numerous large, approximately circular magnetic lows interpreted to represent undeformed plutons (Cowley 2006) that are not present on the Gawler–Adélie Craton. The apparently younger magmatism and basement (*c.* 1500–1400 Ma) and slightly more juvenile nature of the magmatism is also not consistent with a continuous Archaean-floored Palaeoproterozoic continent. However, metasedimentary lithologies from this region preserve old Nd-isotope model ages (e.g. 3.2–2.4 Ga from Windmill Islands, Post 2001) and pre-1500 Ma detrital zircon ages that appear to be consistent with derivation from a Gawler–Adélie Craton source (Fig. 3, Post 2001). This may indicate some genetic link with the Gawler–Adélie Craton at or after *c.* 1500 Ma. The model adopted herein suggests accretion of the Coompana Block and Antarctic equivalents to the Mawson Continent at *c.* 1600–1550 Ma (Betts & Giles 2006).

The extent of the Mawson Continent south of the Wilkes Province in East Antarctica (Fig. 1) has little constraint. As noted by Boger *et al.* (2006) the Archaean lithologies of the southern Prince Charles Mountains share no common timelines with the Archaean crust of the Mawson Continent. It was previously thought that the southern Prince Charles Mountains represented a distinct terrane to the adjoining Napier Complex and Vestfold Hills until Palaeozoic amalgamation (Fitzsimons 2000; Boger *et al.* 2001). However, recent detrital zircon geochronology suggests these terranes may have been amalgamated as early as the late Archaean–early Palaeoproterozoic (Phillips *et al.* 2006). This further highlights the differing evolution of these terranes compared to the Gawler–Adélie Craton and Mawson Continent. It would appear that the southern Prince Charles Mountains and associated terranes were not part of the Palaeoproterozoic Mawson Continent, with probable amalgamation with the East Antarctic Shield occurring during a later event such as the Pinjarra Orogen associated with East Gondwana-Land assembly (Fitzsimons 2003).

Palaeomagnetic constraints

As a general principle, palaeomagnetic data are well-suited to disproving proposed cratonic connections, but they cannot definitively substantiate any particular reconstruction. Nonetheless, long-lived connections between two or more cratons can be supported by a well-populated palaeomagnetic database as follows. One craton, along with its palaeomagnetic apparent polar wander (APW) path, can be rotated by the same Euler parameters into the reference frame of another craton. If this results in both the direct spatial juxtaposition of those cratons, and also superposition of the two APW paths with precise age matches, then a long-lived, direct connection is allowable for the given interval of time represented by the palaeomagnetic poles. The more poles that rotate into alignment, the more powerful the connection is supported; and if there are tectono-stratigraphic similarities that are brought together in the reconstruction, even more compelling does that model become. The palaeomagnetic technique requires assumptions of a constant-radius Earth and geocentric-axial-dipole (GAD) hypothesis for the Earth's magnetic field, the latter verified to first order by Evans (2006).

In the case of the Palaeo-Mesoproterozoic interval, a sparse palaeomagnetic dataset exists from each of the three blocks considered here: Mawson Continent, North Australian Craton and Laurentia. Data from Australia are summarized by Idnurm (2000) and Wingate & Evans (2003) and those from Laurentia are reviewed by Irving *et al.* (1972, 2004) and supplemented by a new isotopic age of *c.* 1590 Ma (Hamilton & Buchan 2007) for the Western Channel Diabase (Irving *et al.* 1972). The Mawson Continent data, namely the pole from the Gawler Range Volcanics, are first restored to North Australia by an Euler pole with parameters (18°S, 134°E, 51°CCW), as described above with minor modifications from Giles *et al.* (2004). Thereafter, this pole and the 1725–1640 Ma APW path from the McArthur Basin and Lawn Hill Platform (Idnurm 2000) are rotated to the Laurentian reference frame by an Euler pole with parameters (31.5°N, 98°E, 102.5°CCW). As shown in Figure 4, these rotations bring the Australian cratons in direct juxtaposition with northwestern Laurentia, in a reconstruction that is reminiscent of the SWEAT model for Rodinia (Moores 1991). Although the Rodinian SWEAT hypothesis has been shown to be palaeomagnetically untenable for the ages 755 Ma (Wingate & Giddings 2000), 1070 Ma (Wingate *et al.* 2002*b*) and *c.* 1200 Ma (Pisarevsky *et al.* 2003), our figured proto-SWEAT reconstruction appears attractive for the pre-Rodinian interval of 1740–1590 Ma.

A long-lived connection between reconstructed Australian cratons and Laurentia through to the end of the Palaeoproterozoic begs the question of when it could have initially formed, and when it ultimately fragmented. In principle, successively older palaeomagnetic poles can be compared in the same rotated reference frame until discordance of a precisely coeval pole pair is identified, and this provides a maximum age estimate for the formation of the cratonic juxtaposition. Laurentia assembled *c.* 1810 Ma (St-Onge *et al.* 2006), thus prior to that age we must compare data only from its more proximal components to the proto-SWEAT juxtaposition, that is, the Slave Craton and conjoined portions of the Churchill Province. A preliminary APW path for those regions is developing (Buchan *et al.* 2007; Evans & Raub 2007), but comparative data from ages immediately older than 1800 Ma in the North Australian Craton are lacking. It thus remains unclear when our proposed connections between Australian cratons and NW Laurentia initiated (Fig. 5). Presumably, there were collisions associated with the Barramundi Orogeny, as discussed above, but the current palaeomagnetic database is inadequate to test whether small-scale Wilson cycles of separation and reunification occurred between Australian and Laurentian blocks (as queried in Fig. 6b, Option 2).

A long-lived connection as proposed here also begs the question of when it fragmented. Successively younger palaeomagnetic poles, of precisely the same ages across all of the reconstructed cratons, will ultimately result in a discordance, which then provides the minimum age of breakup. For our proposed reconstruction, the Laurentian database following 1590 Ma is well constrained by high-quality poles (Evans & Pisarevsky 2008), but the Australian database lacks high-reliability results until *c.* 1200 Ma from the Albany–Fraser belt and southern Yilgarn Craton (Pisarevsky *et al.* 2003). Those poles are broadly compatible with our reconstruction, which permits the intriguing possibility that the North (combined with West) Australian craton remained fixed to NW Laurentia until the late Mesoproterozoic, while the Mawson Continent rifted from Laurentia and rotated into Albany–Fraser–Musgrave orogenesis at *c.* 1300 Ma. Such kinematics would be similar to the Neogene rotation of Arabia away from Africa and towards collision with Eurasia, those larger continents being relatively stationary. If so, then mid-ocean ridge propagation into the originally unified Laurentia–Mawson plate could be manifested by mafic magmatism of either 1470 Ma (Sears *et al.* 1998) or 1370 Ma (Doughty & Chamberlain 1996) in western North America. High-quality palaeomagnetic data from the Gawler Craton in the interval following 1590 Ma are needed to test this hypothesis.

Existing tectonic reconstruction models

In recent years a significant number of tectonic reconstruction models that address the evolution of

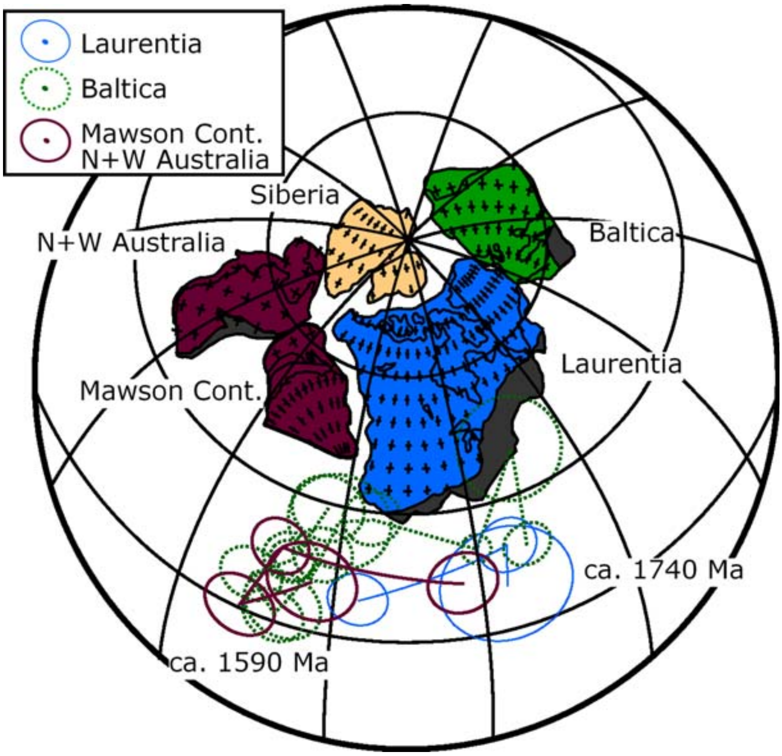

Fig. 4. Palaeomagnetic reconstruction, in the present North American reference frame, of the 'proto-SWEAT' connection between Australia and Laurentia, along with possibly adjacent cratons Baltica and Siberia, for the interval *c*. 1740–1590 Ma, and likely immediately earlier and later times. For geographic reference, late Mesoproterozoic ('Grenvillian') orogenic belts are shown in dark grey. As discussed in the text, the Mawson Continent (Gawler, Terre Adélie and proposed adjacent regions of Antarctica) is reconstructed to the united North and West Australian cratons by the Euler parameters (18°S, 134°E, +51°CCW), largely following Giles *et al*. (2004) but with minor modifications. Thereafter, the reconstructed Australian craton in North Australian reference frame is rotated to Laurentia (31.5°N, 098°E, +102.5°). Following Smethurst *et al*. (1998), again with minor modifications, the northwestern portion of Siberia is restored to the Aldan Shield by closing the Devonian Vilyuy rift (60°N, 115°E, −25°). Thereafter, the reconstructed Siberian Craton in Aldan reference frame is rotated to Laurentia (77.1°N, 113.2°E, +138.7) with minor modifications from the Rodinia models of Rainbird *et al*. (1998) and Li *et al*. (2008). Baltica is restored to Laurentia in the NENA connection of Gower *et al*. (1990) as quantified (47.5°N, 001.5°E, +49°) by Evans & Pisarevsky (2008). Palaeomagnetic poles are rotated by the same parameters as their host cratons, sharing the same colour codes. Australian paleomagnetic poles are selected from Idnurm (2000), inclusion here requiring satisfaction of a field-stability test on the age of magnetization. Laurentian and Baltic poles are illustrated in Evans & Pisarevsky (2008).

the Mawson Continent, or components thereof, have been proposed (Daly *et al*. 1998; Karlstrom *et al*. 2001; Betts *et al*. 2002; Dawson *et al*. 2002; Giles *et al*. 2002, 2004; Fitzsimons 2003; Direen *et al*. 2005; Betts & Giles 2006; Wade *et al*. 2006). Discussions regarding many of these models can be found in Betts & Giles (2006), Payne *et al*. (2006, 2008) and Hand *et al*. (2007). For the purposes of this review we focus on the most recent tectonic reconstruction model for the Mawson Continent, Betts & Giles (2006), and earlier versions of this model.

Betts & Giles (2006) present a model for the 1800–1000 Ma tectonic evolution of Proterozoic Australia that builds 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 (Fig. 5). In each of the Betts & Giles models (listed above), the Kimban Orogeny is interpreted to align/connect with the Strangways Orogeny in the Arunta Region to form a roughly east–west trending

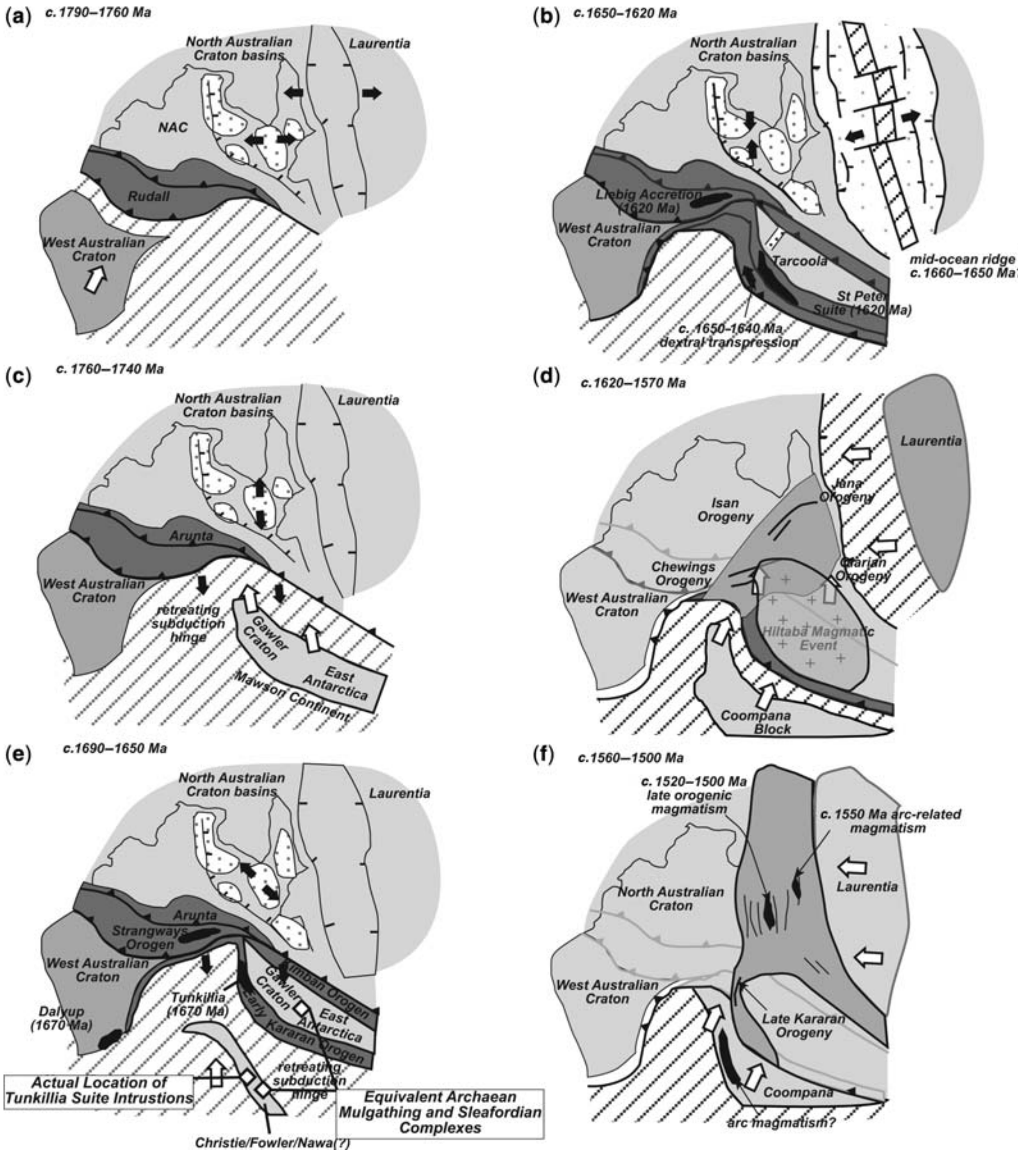

Fig. 5. Reconstruction model of Betts & Giles (2006) showing hypothesized multiple accretion events in the Gawler Craton. Specific tectono-thermal events highlighted are those referred to within the text.

collisional belt (Fig. 5). This rotation is formalized by utilizing palaeomagnetic data at *c*. 1590 Ma to infer a counter-clockwise rotation of 52° about an Euler pole of 136°E and 25°S (modern-day coordinates, Giles *et al*. 2004) with minor modification here to accommodate the broader 1740–1590 Ma dataset (Gawler to North Australia: −18°, 134°, 51° CCW). By assigning an active margin to the southern North Australian Craton, Giles *et al*. (2002, 2004) place the intracratonic McArthur and Mt Isa basins in northern Australia into a far-field extensional back-arc setting.

In the models of Betts *et al*. (2002) and Giles *et al*. (2002, 2004), the Gawler Craton was thought to have accreted to the North Australian Craton during the Kimban–Strangways Orogeny. This was revised in Betts & Giles (2006) such that crust east of the Kalinjala Shear Zone in the Gawler Craton was originally part of the North Australian Craton and the proto-Gawler Craton (namely *c*. 2560–2420 Ma

Sleafordian Complex and 200–1780 Ma cover sequences) was the colliding terrain during the Kimban–Strangways Orogeny (Fig. 5b). The remainder of the Gawler Craton is then interpreted to have accreted at *c*. 1690–1650 Ma (Fig. 5c, d). Orogenesis at 1565–1500 Ma in the North and South Australian Cratons is interpreted to be related to collision with Laurentia along the eastern margin of the North Australian Craton.

The model of Betts & Giles (2006) honours many geological constraints and consequently a number of aspects of this model are adopted in the reconstruction model proposed herein. However, there are a number of apparent inconsistencies between the model of Betts & Giles (2006) and known geological constraints from the North Australian Craton and Mawson Continent. The long-lived southern accretionary margin of Betts & Giles (2006) is not readily reconcilable with the apparent lack of evidence for subduction and collisional orogenesis in the 1760–1690 Ma time period outside of the easternmost Arunta Region (Claoue-Long *et al*. 2008). The lack of evidence for subduction and/or collision along the strike of the proposed southern margin cannot be attributed to preservation because the remainder of the Arunta Region is one of the best preserved and exposed regions of the Australian Proterozoic. Second, within the Gawler Craton the proposed division into three terranes at *c*. 1690 Ma is largely unsupported and in some cases irreconcilable with geological constraints. The distinction of the eastern Gawler Craton (east of the Kalinjala Shear Zone) from the proto-Gawler Craton Archaean is largely based on the lack of 1850 Ma Donington Suite granitoids and associated Cornian Orogeny to the west of the Kalinjala Shear Zone (Fig. 2). Recently collected detrital zircon Hf-isotope data potentially supports the distinction as it suggests pre-1850 Ma sedimentary rocks deposited east of the Kalinjala Shear Zone were not sourced from the currently outcropping Gawler Craton (Howard *et al*. 2007). In contrast, the presence of temporally equivalent sediment deposition on either side of the proposed suture at both pre-1850 Ma (Fanning *et al*. 2007; Howard *et al*. 2007) and 1780–1740 Ma time periods (Daly *et al*. 1998; Cowley *et al*. 2003; Fanning *et al*. 2007) suggests some common tectonic context for the two regions prior to the Kimban Orogeny. Metasedimentary rocks west of the Kalinjala Shear Zone, deposited after 1850 Ma, have a high proportion of *c*. 1860–1850 Ma detrital zircons (Jagodzinski 2005). This suggests that the proto-Gawler Craton was already associated with a significant volume of 1850 Ma magmatic lithologies prior to the 1730–1690 Ma Kimban Orogeny. In addition, there is little or no evidence for deformation of the proposed overriding plate during the Kimban Orogeny, even immediately adjacent to the proposed suture: major deformation in the Yorke Peninsula region is synchronous with *c*. 1590 Ma Hiltaba Suite intrusion (Cowley *et al*. 2003).

The second proposed accretionary event of Betts & Giles (2006, Fig. 5c) is difficult to reconcile with current geological constraints. The Tunkillia Suite (1690–1670 Ma) in the central and western Gawler Craton has previously been suggested to represent subduction-related magmatism based upon trace-element tectonic discrimination diagrams (Teasdale 1997; Betts & Giles 2006). This classification has since been demonstrated to be questionable (post-tectonic petrogenesis, Payne 2008) and hence there is no evidence for subduction beneath the Gawler Craton at this time. Regardless of the tectonic setting of the Tunkillia Suite, the bulk of its magmatism occurs on the proposed underthrust Nawa–Christie–Fowler plate of Betts & Giles (2006), meaning the proposed model is internally inconsistent. The proposed allocthonous Nawa–Christie–Fowler collider also separates the late-Archaean Mulgathing (on Christie Plate, Fig. 5c) and Sleafordian Complexes, despite their identical Archaean–early Palaeoproterozoic tectonic history (Swain *et al*. 2005*b*). Furthermore, the metamorphic and magmatic expressions of the Kimban Orogeny effectively stitch the Gawler Craton together at 1730–1690 Ma (Payne *et al*. 2008), which argues against any younger accretionary events.

Towards a unified model

Internal architecture of the Gawler Craton

Due to the large degrees of freedom in reconstruction models for the Palaeoproterozoic, an exact geometry of the continental blocks is commonly not required and typically not possible. However, the potential for large intra-cratonic architectural rearrangements must be assessed to ascertain the validity of utilizing palaeomagnetic and structural geology constraints. At the craton-scale, the Gawler–Adélie Craton has an architecture in which the Archaean to Palaeoproterozoic lithologies appear to have been wrapped around a younger Palaeo- to Mesoproterozoic core (namely St Peter Suite magmatic lithologies) in the regional interpreted geology (Fig. 2). This architecture has led to suggestions incorporating a hypothesized ‘bending’ of a previously more linear Gawler–Adélie Craton through oroclinal folding (Swain *et al*. 2005*a*), or large degrees of lateral movement through late left-lateral shear zone movement (Direen *et al*. 2005). Proponents of a bending or transposition of the Gawler Craton cite the north–south to NNE–SSW structural trends in the

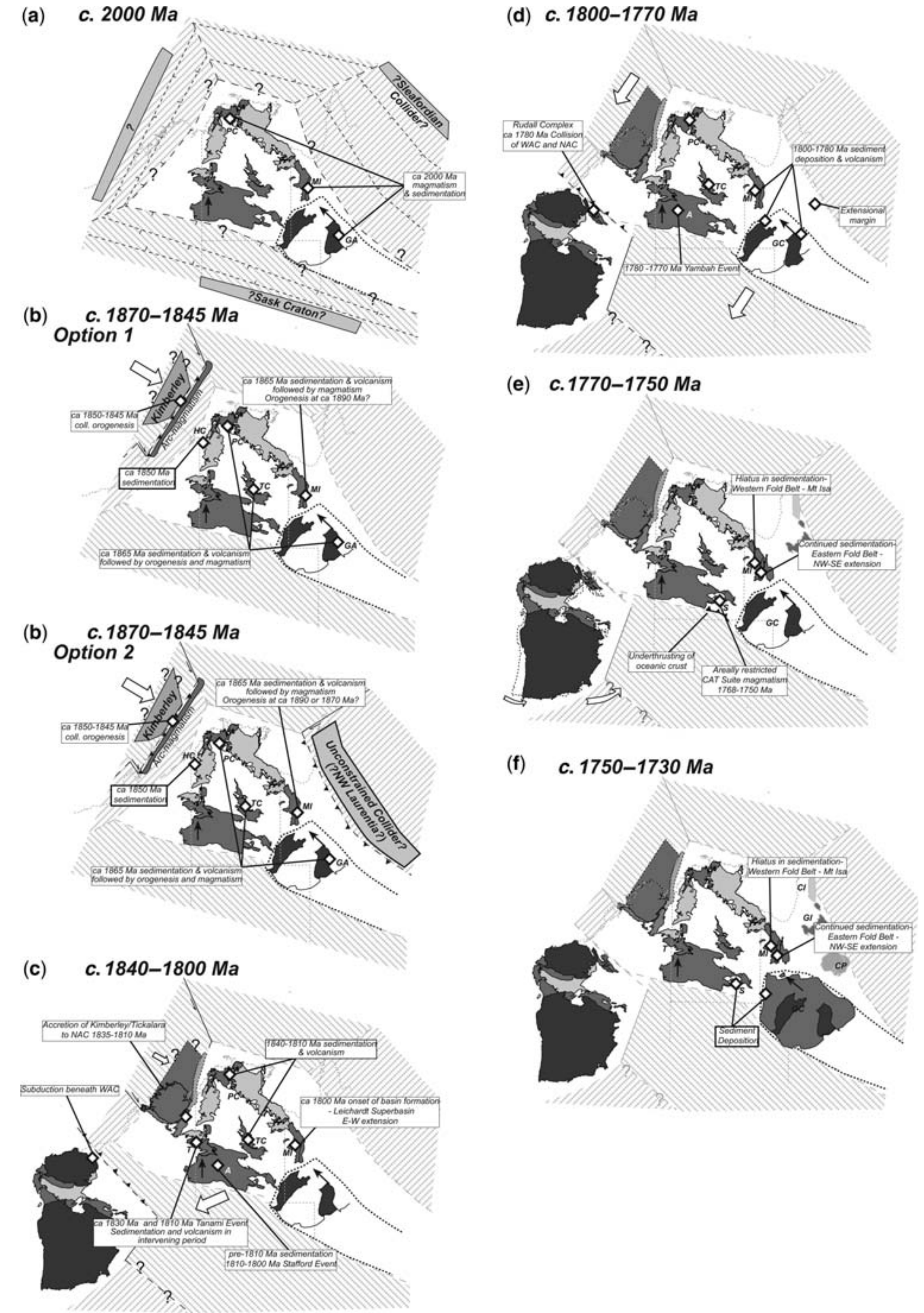

Fig. 6. Proposed reconstruction model for the development and tectonic evolution of the Mawson Continent. Small black arrow in Gawler–Adélie Craton and Arunta Region represents current north. Large arrows represent potential plate movement directions. Figure 6b provides two alternative scenarios for the *c*. 1850 Ma timeline. Going forward from this timeline, Option 1 is adopted but the geometry can be readily exchanged such that rifting in the

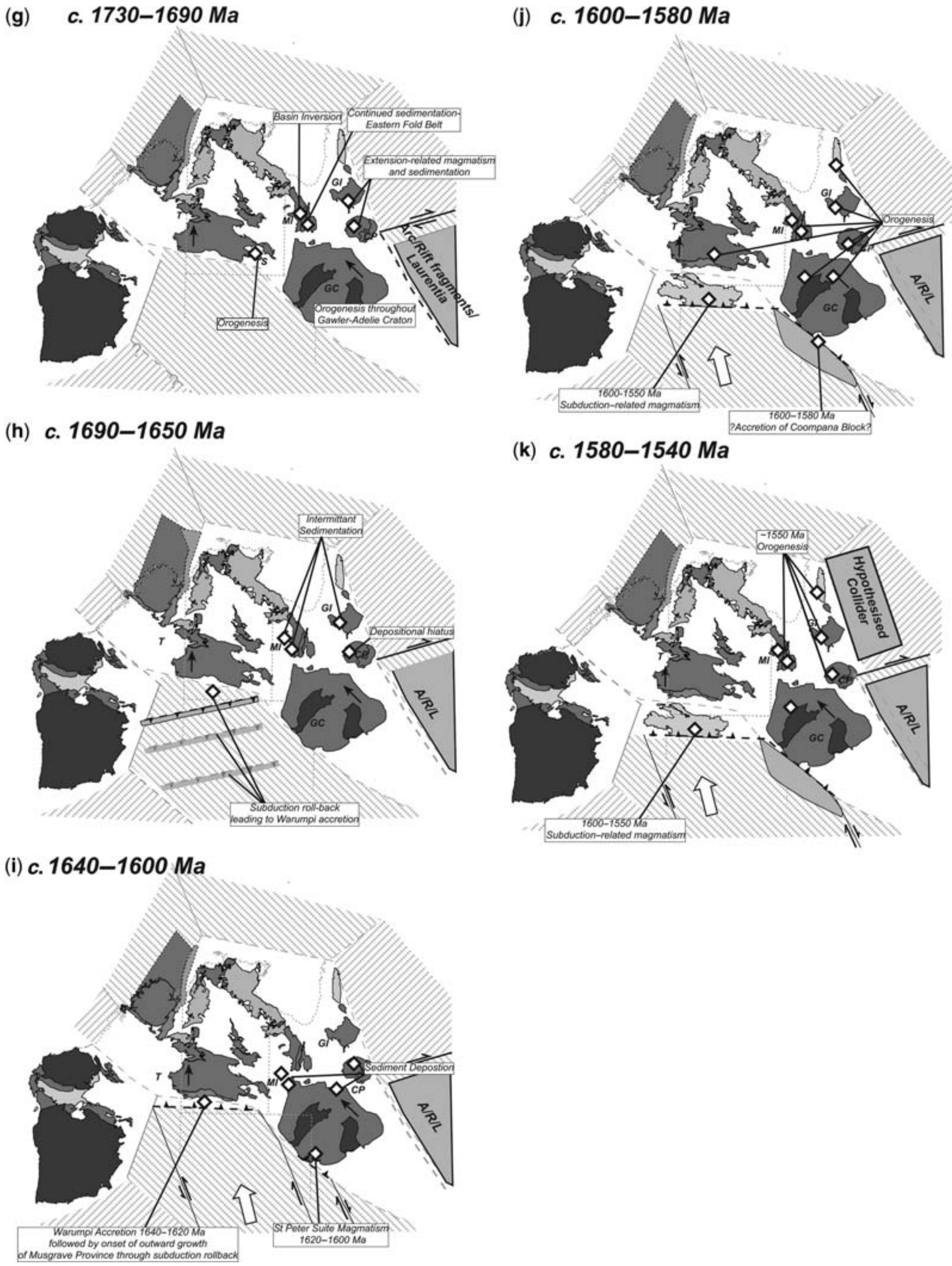

Fig. 6. (*Continued*) northeastern North Australian Craton relates to the rifting away of the collided terrain. Terrain abbreviations are: A, Arunta Region; CI, Coen Inlier; CP, Curnamona Province; GC, Gawler-Adelie Craton; GI, Georgetown Inlier; HC, Halls Creek; MI, Mount Isa Inlier; PI, Pine Creek; R, Rudall Complex; S, Strangways Complex; T, Tanami Region; TC, Tennant Creek/Davenport Region; and W, Warumpi Province. Striped grey regions represent oceanic lithosphere, dashed black lines represent active plate boundaries, grey dashed lines represent inactive boundaries, dotted line represents extent of the Gawler–Adelie crust. Greyscale shading of Australian terrains as per shading in Figure 1.

southeastern Gawler Craton and Adélie Craton and NE–SW to east–west structural trends in the northern and northwestern Gawler Craton as supporting evidence. If this hypothesis is correct, it implies that oroclinal bending or transposition of the northern parts of the craton occurred between *c.* 1608 and 1592 Ma, as the upper Gawler Range Volcanics (in the core of the apparently-arcuate shaped cratonic domains) are relatively flat-lying and undeformed (Daly *et al.* 1998). Because convincing evidence demonstrating the bending of the Gawler Craton has not been recorded, we adopt the simplest model for the Palaeoproterozoic geometry of the Gawler–Adélie Craton, restoring only the Mesozoic rifting between Australia and Antarctica (Figs 1 & 2).

Evolution of the Gawler–Adélie Craton and Mawson Continent in a global setting

This section outlines a new model for the evolution of the Gawler–Adélie Craton and Mawson Continent (Fig. 6). The model focuses on identifying correlatable timelines within other Archaean–Mesoproterozoic terrains and summarizing all correlations in a geological constraint-driven model.

Late Archaean–early Palaeoproterozoic. The first correlatable tectonic cycle on the Gawler–Adélie Craton is late Archaean magmatism and sedimentation, including subduction-related magmatism of the Dutton suite and Devil's Playground Volcanics (*c.* 2560–2520 Ma) and the Sleafordian Orogeny at *c.* 2460–2430 Ma (Swain *et al.* 2005*b*). Convergent tectonic settings of this age are rare among the world's cratons (Fig. 7) and the Sleafordian Orogeny in particular represents a relatively uncommon timeline. Three terrains record evidence for similar-aged metamorphism/orogenesis with peak metamorphism in the earliest Palaeoproterozoic: the North Australian Craton, the Sask Craton and the very poorly-known North Korean peninsula.

The *c.* 2520 Ma magmatism in the Gawler Craton has temporal equivalents in the North Australian Craton in the Pine Creek Inlier and Tanami Region (Lally 2002; Cross *et al.* 2005; Crispe *et al.* 2007). The nature and timing of metamorphism of the North Australian Archaean lithologies is yet to be reliably constrained. Recent reconnaissance geochronology has reported an age of 2473 $\pm$ 12 Ma from three analyses of zircon rims in a *c.* 2633 Ma orthogneiss (Worden *et al.* 2006). Within basement inliers in the Mt Isa region, McDonald *et al.* (1997) report 2500–2420 Ma magmatic ages for the Black Angel Gneiss and identified subduction-related arc-geochemical signatures. The classification of the Black Angel Gneiss as a magmatic Archaean-age lithology is disputed, with alternative *c.* 1850 Ma interpreted ages suggested (Page & Sun 1998). Furthermore, detailed evaluation of geochemical data used to identify a subduction-related petrogenesis is not provided in McDonald *et al.* (1997). In addition to Archaean basement lithologies of the North Australian Craton, ages of 2500–2450 Ma are obtained from detrital zircon geochronology of metasedimentary lithologies within the eastern Arunta Region (Wade *et al.* 2008). These provenance data provide further evidence for similar-aged Archaean to early Palaeoproterozoic lithologies within the North Australian and Gawler–Adelie cratons. The similarity in age, and potentially tectonic setting, of tectono-thermal events in the late Archaean and early Palaeoproterozoic lithologies of the North Australian and Gawler–Adelie Cratons appears to continue throughout the Palaeoproterozoic (see below). Based upon the outlined temporal correlations and apparent longevity of interaction between the two cratons, we suggest they formed a single continental domain in the late Archaean–early Palaeoproterozoic.

The crustally evolved Nd-isotope composition of Archaean lithologies within the Gawler Craton (Nd-depleted mantle model ages – *c.* 3.4–2.8 Ga, Swain *et al.* 2005*b*) and presence of detrital zircon ages (2720–2600 Ma with minor inheritance at *c.* 3000–2800 Ma) that are not consistent with derivation from the currently exposed Gawler–Adélie Craton (Swain *et al.* 2005*b*), suggest that the Gawler–Adélie Craton was built upon pre-existing Mesoarchaean crust. The Sask Craton, Trans-Hudson Orogen, Laurentia, yields evidence of Mesoarchaean crust that is consistent with potential Gawler–Adelie Craton protoliths (Chiarenzelli *et al.* 1998) and records orogenesis similar in age to the Sleafordian Orogeny (Chiarenzelli *et al.* 1998; Rayner *et al.* 2005). Magmatic lithologies and inherited zircons within the Sask Craton (Chiarenzelli *et al.* 1998; Ashton *et al.* 1999; Rayner *et al.* 2005) correspond to all major detrital zircon age populations in the late Archaean metasedimentary lithologies of the Gawler–Adélie Craton (Fig. 7, Swain *et al.* 2005*b*). Magmatism at *c.* 2520–2450 Ma with metamorphic reworking of the crust at *c.* 2450 Ma (Rayner *et al.* 2005) correlates well with the late Archaean–early Palaeoproterozoic lithologies of the Gawler–Adélie Craton. The combined evidence suggests the Sask Craton could have been contiguous with the Gawler–Adélie and North Australian cratons, and also provides potential equivalent lithologies for the unexposed basement to the Gawler–Adélie Craton lithologies.

In addition to the North Australian Craton and Sask Craton, Sleafordian-age metamorphism is also recorded in the North Korean Peninsula (Zhao *et al.* 2006*a*). The North Korean Peninsula records

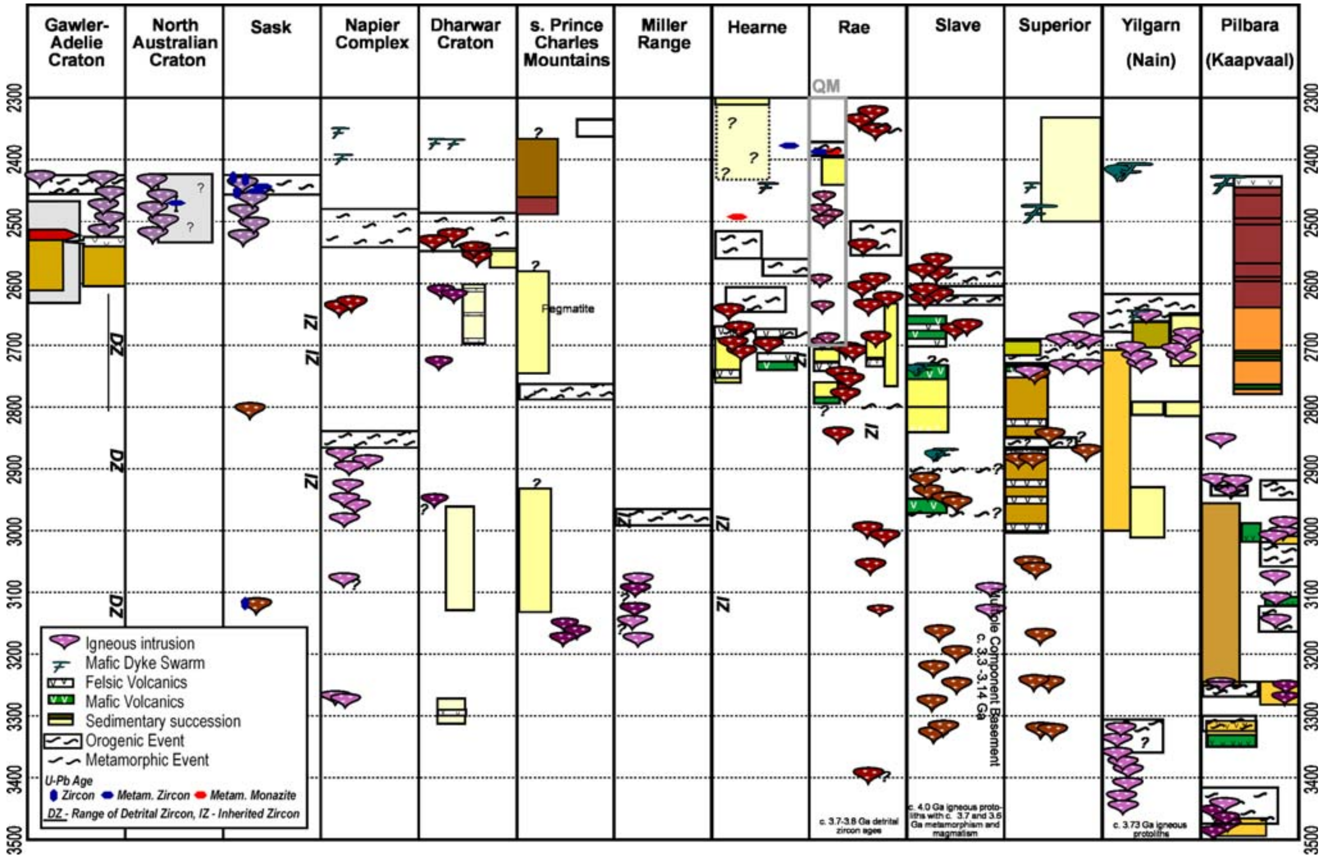

Fig. 7. Geological event plot displaying tectono-thermal evolutions of the Archaean components of the Mawson Continent and various other Archaean cratons/terranes. For simplicity, cratons which have previously been highlighted as similar to those portrayed in the figure (and generally not suitable correlatives of the Gawler–Adélie Craton) are noted in brackets in the column heading. Gray rectangle within Rae Craton column represents the Queen Maud Block (QM). Age-constrained metamorphic mineral growth that is not demonstrated to represent orogenesis is represented as a 'Metamorphic Event'. Data sources not mentioned in text are: *Yilgarn*: Cassidy *et al*. (2006) and references therein; *Pilbara*: Hickman (2004), Pawley *et al*. (2004), Trendall *et al*. (2004) and references therein; *Napier Complex*: Kelly & Harley (2005) and references therein; *Sask*: Chiarenzelli *et al*. (1998), Ashton *et al*. (1999), Rayner *et al*. (2005); *Superior*: Stott (1997), Bleeker (2003), Tomlinson *et al*. (2004), Lin *et al*. (2006), Melynk *et al*. (2006), Parks *et al*. (2006), Percival *et al*. (2006); *Slave*: Bleeker *et al*. (1999), Bowring & Williams (1999), Davis & Bleeker (1999), Bleeker (2003), Ketchum *et al*. (2004), Goodwin *et al*. (2006); *Rae*: Bethune & Scammell (2003), Carson *et al*. (2004), Berman *et al*. (2005), Hartlaub *et al*. (2004, 2005, 2007) and references therein; *Hearne*: Davis *et al*. (2004, 2006), Hanmer *et al*. (2004), MacLachlan *et al*. (2005), van Breemen *et al*. (2007) and references therein; *Queen Maud*: Schultz *et al*. (2007); *Dharwar*: Nutman *et al*. (1992, 1996), Peucat *et al*. (1993, 1995), Chadwick *et al*. (2001), Vasudev *et al*. (2000), Jayananda *et al*. (2006), Halls *et al*. (2007).

magmatism at *c.* 2640 and 2540 Ma with *c.* 2460–2430 Ma metamorphism (Zhao *et al.* 2006*a*). Further work is required to better characterize this terrain and its relationship with the eastern block of the North China Craton (Zhao *et al.* 2006*a*), but these preliminary data suggests possible links with the Gawler–Adélie and North Australian cratons.

Immediately prior to the time of the Sleafordian Orogeny, the western block of the North China Craton (Zhao *et al.* 2006*b* and references therein), Dharwar Craton (India, Friend & Nutman 1991), Napier Complex and Vestfold Hills (Antarctica, Kelly & Harley 2005; Zulbati & Harley 2007) and Rae and Hearne cratons (Laurentia, see Fig. 7 caption for references) underwent orogenesis over the interval 2550–2470 Ma with peak metamorphism prior to 2500 Ma. The Rae Craton and Queen Maud Block (Fig. 7) also record a later episode(s) of orogenesis in the 2390–2320 Ma period (Arrowsmith Orogeny, Schultz *et al.* 2007). The relationship of the Sleafordian Orogeny with deformation in these terrains is unclear; however, given the close temporal relationship of orogenesis in relation to the apparent extensional or cratonized state of most other Archaean cratons (Bleeker 2003), the potential for palaeogeographic proximity is worthy of further consideration.

Circa 2000–1850 Ma rifting and sedimentation. Both the North Australian Craton and Gawler–Adélie Craton share a period of inactivity (*c.* 2440–2050 Ma) prior to the onset of sedimentation and felsic magmatism (Daly *et al.* 1998; Fanning *et al.* 2007; Worden *et al.* 2008), possibly representing continental rifting and breakup. In order to honour the late Archaean link proposed with the Sask Craton, the Sask Craton must have rifted away from the Gawler–Adélie and North Australian cratons sometime prior to *c.* 1830 Ma collision and incorporation into the Trans-Hudson Orogen (Ansdell 2005; Rayner *et al.* 2005). The *c.* 2000 Ma timeframe appears to be the most suitable time for the required rifting of the Sask Craton away from the proposed Gawler–Adelie and North Australian craton lithosphere. That sedimentation and magmatism developed within previously stabilized continental domains and is followed by an extended period of sediment deposition is also circumstantial evidence for a rift or extensional setting at *c.* 2000 Ma (as proposed by Daly *et al.* 1998). The central Pine Creek Inlier metasedimentary and volcanic lithologies suggest deposition in east-deepening tilted-block basins with coarsening of fluvial fan material indicating topographic relief to the west and approximately east–west directed extension at the time of deposition (*c.* 2020–2000 Ma, Worden *et al.* 2008 and references therein). McDonald *et al.* (1997) outline a subduction-related magmatic event in the Mt Isa inlier at *c.* 2000 Ma. As the age of this magmatism is disputed (Page & Sun 1998), we have not included it in the model proposed here.

Globally, the 2000 Ma timeline is late in the time period typically assigned to the final breakup of late Archaean supercontinent/s before the onset of extensive continent amalgamation starting at *c.* 1900 Ma (Trans-Hudson and North China Orogenies, Condie 2002, 2004; St-Onge *et al.* 2006; Zhao *et al.* 2006*b* and references therein).

Circa 1890–1810 Ma orogenesis and magma generation. The period of the Palaeoproterozoic from *c.* 1950–1800 Ma is commonly cited as representing the final amalgamation of the proposed supercontinent Nuna or Columbia (Zhao *et al.* 2002; Rogers & Santosh 2003; Zhao *et al.* 2004; Kusky *et al.* 2007). In the case of continents such as Laurentia and Baltica, this period is relatively well-characterized and has been demonstrated to represent the amalgamation of multiple cratons and continent stabilization prior to a period of terrain accretion. By contrast, the events of this period in the Gawler–Adélie and North Australian Cratons are yet to be fully understood.

The *c.* 1850 Ma Cornian Orogeny in the eastern Gawler Craton is effectively delineated by the areal extent of the syntectonic Donington Suite (Daly *et al.* 1998; Reid *et al.* 2008). This areal distribution has led to the aforementioned proposal of a collisional suture along the Kalinjala Shear Zone during the Kimban Orogeny (Betts & Giles 2006). The model proposed herein differs from the Betts & Giles (2006) model, instead interpreting the Kalinjala Shear Zone as an intra-cratonic shear zone. This interpretation is consistent with the information outlined earlier in this review, which does not support a *c.* 1730–1690 Ma continental suture along the Kalinjala Shear Zone (Betts & Giles 2006) and retains the Gawler–Adélie and North Australian Cratons as a single entity at this time.

The North Australian Craton records metamorphism and deformation similar in age to the Cornian Orogeny in a number of terrains. The stepwise accretion of the Kimberley Craton to the western margin of the North Australian Craton is the best understood of these events. The Kimberley Craton is interpreted to have collided with the Tickalara arc at *c.* 1850–1845 Ma (Sheppard *et al.* 1999, 2001; Griffin *et al.* 2000), before both terrains were accreted to the North Australian Craton at *c.* 1835–1810 Ma (Fig. 6b, c, Sheppard *et al.* 2001). Within the interior of the North Australian Craton, events in the Tennant Creek–Davenport and Pine Creek regions of the North Australian Craton (Fig. 8) correlate temporally with events in the Gawler–Adélie Craton. Within the Pine Creek

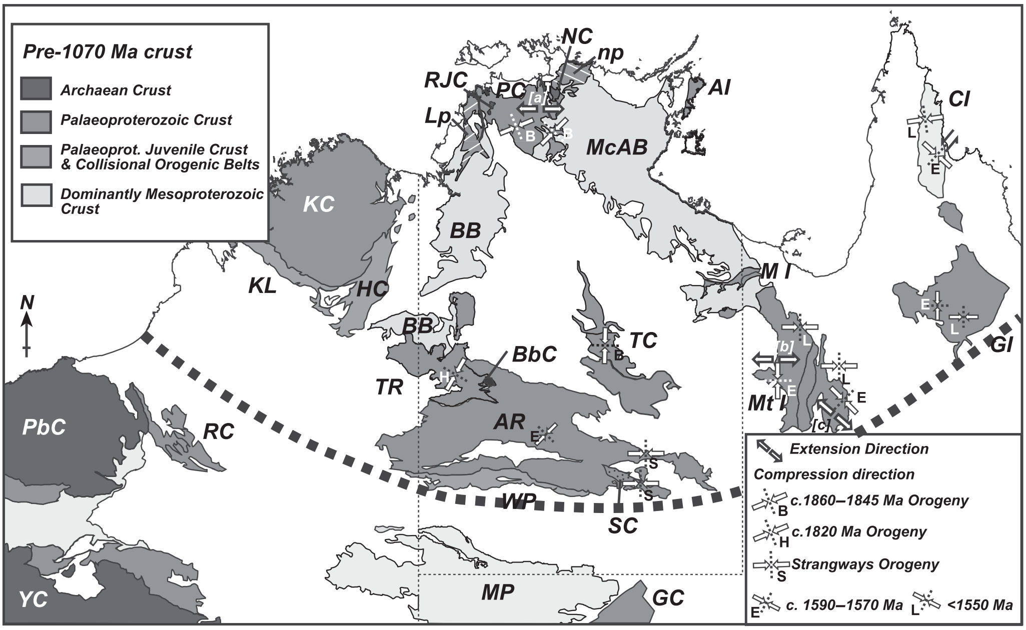

Fig. 8. Geology of the North Australian Craton with orientation of principle stress for tectonic events in the Palaeo- to early Mesoproterozoic. Thick dashed line represents widely used southern extent of the North Australian Craton. Region abbreviations are: AI, Arnhem Inlier; AR, Arunta Region; BB, Victoria-Birindudu Basin; BbC, Billabong Complex (Archaean); CI, Coen Inlier; GC, Gawler Craton; GI, Georgetown Inlier; HC, Halls Creek Orogen; KC, Kimberley Craton; KL, King Leopold Orogen; McAB, McArthur Basin; MI, Murphy Inlier; Mt I, Mt Isa Inlier; MP, Musgrave Province; NC, Nanumbu Complex (Archaean); PC, Pine Creek Orogen with white hatched regions representing; Lp, Litchfield Province and np, Nimbuwah Province; PbC, Pilbara Craton; RC, Rudall Complex; RJC, Rum Jungle Complex (Archaean); SC, Strangways Complex (stippled region); TC, Tennant Region; TR, Tanami Reigon; WP, Warumpi Province; and YC, Yilgarn Craton. Shortening directions for 1860–1845 Ma (Pine Creek, Tennant Region), *c*. 1820 Ma (Halls Creek Orogeny, Tanami Event) tectonic events, Strangways Orogeny, *c*. 1600–1570 Ma (Early Isan and equivalent and Chewings Orogeny, Blewett & Black 1998; Boger & Hansen 2004) and 1550–1520 Ma (Late Isan and equivalent, Black *et al*. 1998; Boger & Hansen 2004) tectonic events. Dominant extension directions at time of basin formation represent: (**a**) *c*. 2000 Ma in Pine Creek Orogen; (**b**) *c*. 1800 Ma in Mt Isa Inlier; (**c**) 1780–1750 Ma in Mt Isa Inlier. Refer to text for references.

region basin development, including volcanics at 1864 ± 3–1861 ± 4 Ma, is terminated by metamorphism and deformation at 1853 ± 4 Ma with deformation complete by 1847 ± 1 Ma (Carson *et al*. 2008; Worden *et al*. 2008 and references therein). This deformation and metamorphism is contemporaneous with the deposition of the upper Halls Creek Group on the western margin of the North Australian Craton (Olympio Formation, Blake *et al*. 1998). The setting of the Halls Creek Group has been equated to passive margin sedimentary sequences (Sheppard *et al*. 1999), and importantly, these do not show evidence for *c*. 1850 Ma orogenesis. South of the Halls Creek Orogen, recent research has identified a *c*. 1865 Ma volcanic sequence in the western Tanami Region (Bagas *et al*. 2008). This sequence is the first of this age identified in the region and is suggested to have undergone tectonism as early as 1850 Ma (Bagas *et al*. 2008). If this proves to be the case, it further highlights the extent of *c*. 1850 Ma tectonism in the North Australian Craton.

In the eastern North Australian Craton, the Mt Isa Inlier records a similar sequence of sedimentary and igneous events to the Pine Creek and Tennant regions and Gawler Craton, with volcanics at *c*. 1865 Ma (Page & Williams 1988), and voluminous felsic magmatism of the Kalkadoon-Ewen Batholith. However, metamorphism and deformation (and magmatism) within the Mt Isa Inlier is considered to have occurred significantly prior to that of the Pine Creek region and Gawler Craton. Metamorphism (the basis for the Barramundi Orogeny nomenclature) is constrained by a single SHRIMP I U–Pb zircon age of 1890 ± 8 Ma (Page & Williams 1988). However, Bierlein *et al*. (2008) suggest this age is older than the age of metamorphism and assign an orogenic

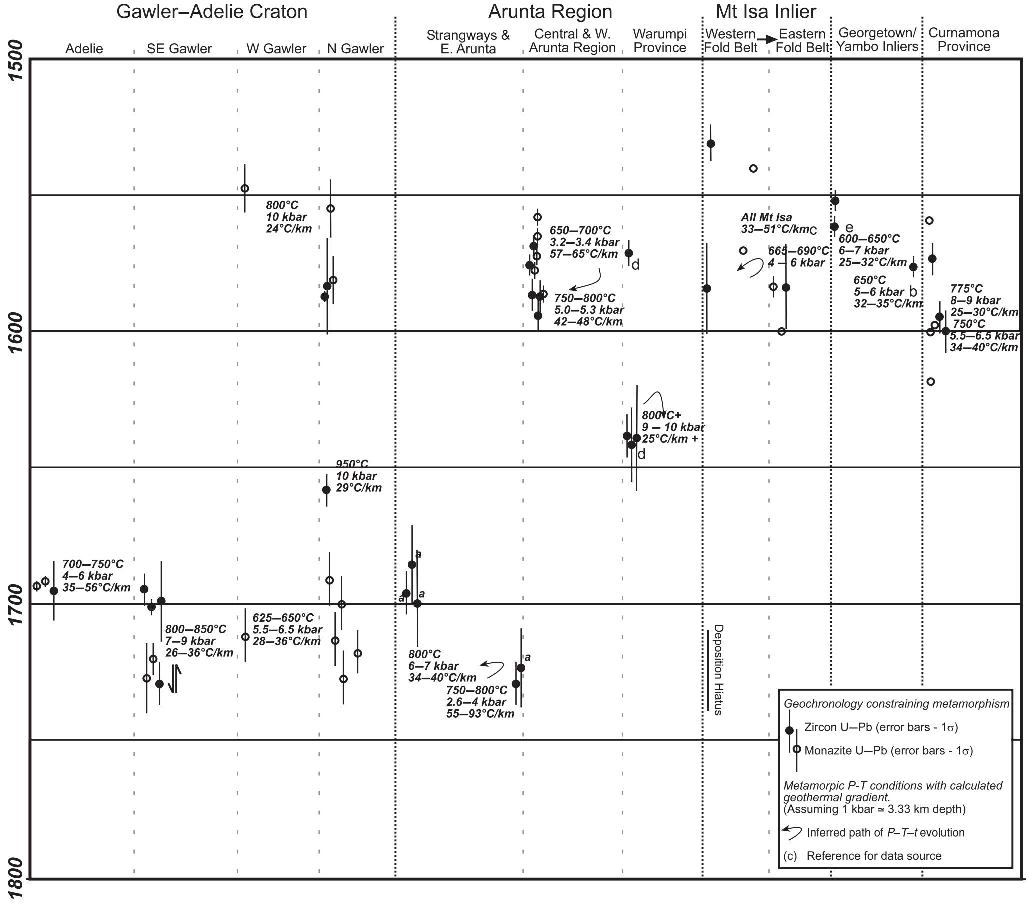

Fig. 9. Detailed Time-Space diagram of *P–T* conditions, *P–T* paths and geochronology of late Palaeoproterozoic to early Mesoproterozoic tectono-thermal events in northern and southern Australia. Source data is referred to in text. Information not referred to in text is: (**a**) Claoue-Long & Hoatson (2005); (**b**) Blewett & Black (1998), Blewett *et al.* (1998) and references therein; (**c**) Mark *et al.* (1998) with regional geothermal gradients from Foster & Rubenach (2006) and references therein; (**d**) Scrimgeour *et al.* (2005); (**e**) Black *et al.* (1998), Boger & Hanson (2004).

age of *c.* 1870 Ma based upon U–Pb zircon ages of granitic intrusions. Given the poorly constrained nature of metamorphism, further work is required to resolve the relationship of Mt Isa tectonism with *c.* 1850 Ma orogenesis elsewhere in North Australian and Gawler–Adélie Cratons.

Two alternative models are envisaged to accommodate the widely distributed metamorphism and tectonism within the 1870–1845 Ma time period. The first (Fig. 6b, Option 1) considers the *c.* 1850 Ma tectonism to be a far-field effect of the Kimberley accretion. Close timing and event duration correlations between accretion of the Kimberley Craton to the Tickalara Arc and events throughout the North Australian Craton supports such a model. However, the lack of deformation within the geographically closer, passive margin Olympio Formation, seemingly contradicts this model. The second scenario (Fig. 6b, Option 2) considers there to have been multiple active margins during this period, potentially with collisional orogenesis occurring on both the eastern and western margins of the North Australian Craton (similar to suggestion of Betts *et al.* 2002). In such a model, tectonism in the Gawler–Adélie, Pine Creek, Mt Isa and Tennant Creek-Davenport regions would be related to a separate collisional event on the eastern or northeastern margin of Australia. Support for such a scenario is provided by the suggested arc-affinities of the geochemical signature for basement granitoids within the Mt Isa Inlier (McDonald *et al.* 1997; Bierlein & Betts 2004). The apparent earlier timing of tectonism within the Mt Isa Inlier, compared to the remainder of the North Australian Craton, may be consistent with a convergent margin in this region.

At *c*. 1830 and 1810 Ma, the Tanami Region underwent metamorphism and deformation associated with the Tanami Event (Crispe *et al*. 2007). The latter part of this event is temporally equivalent to the Stafford Event in the Arunta Region (Scrimgeour 2003). These events are magmatically dominated systems with metamorphism driven by magmatic heat advection and, particularly in the case of the Arunta Region, show limited deformation (Scrimgeour 2003; White *et al*. 2003). The Tanami and Stafford Events (and Murchison Event in Tennant Region) are coeval with the final accretion of the combined Kimberley/Tickalara terrains to the North Australian Craton. As no other plate margin activity is evident within the North Australian Craton at this time, we agree with recent suggestions (Crispe *et al*. 2007; Worden *et al*. 2008) that *c*. 1830–1810 Ma magmatism and deformation within the Tanami and Arunta regions is related to the accretion of the Kimberley/Tickalara crust to the North Australian Craton (Fig. 6c).

Circa 1800–1770 Ma. The Rudall Complex, Western Australia, is interpreted to record the collision of the North Australian and West Australian cratons during the period *c*. 1795–1765 Ma (Fig. 6d). This timing is constrained by the age of granitoid intrusions which are interpreted to pre- and post-date high-pressure metamorphism (*c*. 800 °C, 12 kbar) interpreted to record the collision event (Smithies & Bagas 1997; Bagas 2004). The geometry of subduction leading to this interpreted collision is effectively unconstrained. Smithies & Bagas (1997) propose a NE-dipping subduction geometry that is also adopted by Betts & Giles (2006). We also adopt a NE-dipping subduction geometry but recognize that subduction is equally as likely to have occurred with a SW-dipping geometry and further research is required to resolve this issue.

Apparently synchronous with collision in the Rudall Complex is the 1780–1770 Ma Yambah Event (previously termed Early Strangways) in the Arunta Region (Hand & Buick 2001; Scrimgeour 2003). The Yambah Event is a dominantly magmatic event that does not appear to represent major crustal thickening (Scrimgeour 2003). In the central Arunta Region, Hand & Buick (2001) suggest the Yambah Event was associated with NE–SW directed shortening. A similar-style event occurred in the Tanami Region at *c*. 1800–1790 Ma (Crispe *et al*. 2007), dominated by magmatism with WSW–ENE to east–west shortening represented by thrust faulting (Wygralak *et al*. 2005; Crispe *et al*. 2007). In our proposed model the compressional deformation associated with these two events is linked to the ongoing collision with the West Australian Craton to the SW and west of the Tanami and Arunta Regions. Within the Mt Isa Inlier, basin formation and sedimentation was initiated at *c*. 1800 Ma and continued until basin inversion at *c*. 1740 Ma, as represented by the Leichhardt Superbasin. Initial extension was east–west-directed (O'Dea *et al*. 1997) up to and including the deposition of the Eastern Creek Volcanics. Neumann *et al*. (2006) revise the Mt Isa stratigraphy such that the Eastern Creek Volcanics were emplaced at the initiation of the Myally Supersequence (*c*. 1780–1765 Ma, Neumann *et al*. 2006). The Myally Supersequence was previously interpreted to have been deposited in a north–south extensional regime (O'Dea *et al*. 1997) but this interpretation may no longer be accurate due to the recently revised stratigraphy (see Foster & Austin 2008 for discussion). Initial *c*. 1800 Ma basin extension is interpreted to represent the onset of intracontinental rifting, either within a trailing edge of the North Australian Craton, or between the North Australian Craton and another protocontinent, potentially Laurentia, as hypothesized in Betts & Giles (2006).

Circa 1770–1740 Ma. Within the Arunta Region, *c*. 1770–1750 Ma CAT suite magmatism has been identified as having a subduction-related petrogenesis (Foden *et al*. 1988; Zhao & McCulloch 1995) and has led to the proposal of long-lived north-dipping subduction under the southern margin of the North Australian Craton (Scott *et al*. 2000; Giles *et al*. 2002; Betts & Giles 2006). However, the CAT suite is localized in the easternmost Arunta Region and has a much smaller volume than the predominant 1780–1770 Ma granites, which do not have a subduction-related petrogenesis and are found throughout the southern part of the North Australian Craton (Zhao & McCulloch 1995). The model proposed herein relates the generation of the CAT Suite to rotation of the West Australian Craton comparative to the North Australian Craton following the initial *c*. 1780 Ma interpreted collision between the two cratons. This is interpreted to result in the rupture and forced under-thrusting of the intervening oceanic crust under the eastern Arunta region (Fig. 6e), generating the CAT Suite magmatism. In the western Mt Isa Inlier, the *c*. 1765–1750 Ma period represents a break in sedimentation between the Myally and Quilalar Supersequences (Neumann *et al*. 2006). Conversely, within the Eastern Fold Belt of the Mt Isa Inlier, this period correlated to NW–SE directed extension during the deposition of the Malbon Group (Potma & Betts 2006).

CAT Suite magmatism in the Arunta Region ceased approximately 20–30 Ma prior to the onset of the Strangways Orogeny. The intervening period appears to have been dominated by short-lived basin formation, *c.* 1750–1740 Ma (Fig. 6e), in the eastern Arunta Region and northern Gawler–Adélie Craton (Payne *et al.* 2006; Wade *et al.* 2008). We hypothesize that this may have represented the increased influence of subduction/oceanic crust consumption to the east of the continent, prior to the proposed accretion with Laurentia, resulting in an extensional regime in the Australian plate. Sedimentation also resumed within the Western Fold Belt of the Mt Isa at this time in a sag basin setting (Neumann *et al.* 2006 and references therein).

Circa 1730–1690 Ma orogeny and sediment deposition. The *c.* 1730–1690 Ma Kimban Orogeny timeline has been nominated as a primary correlation event for Proterozoic continental reconstruction models (Goodge *et al.* 2001; Karlstrom *et al.* 2001; Giles *et al.* 2004; Betts & Giles 2006). Equivalent timelines exist in Antarctica (outlined Miller and Shackleton range events) and the Arunta region of the North Australian Craton (Strangways Orogeny). The Yavapai Orogeny in southern Laurentia also records similar timing (Duebendorfer *et al.* 2001; Jessup *et al.* 2006 and references therein).

The Strangways Orogeny in the eastern Arunta Region is currently constrained to *c.* 1730–1690 Ma (Möller *et al.* 2003; Claoue-Long & Hoatson 2005; Maidment *et al.* 2005; Clarke *et al.* 2007), consistent with the timing of the Kimban Orogeny. Within the Strangways Complex (Figs 6, 8 & 9), peak P–T conditions of 800 °C and up to 8 kbar are recorded (Ballevre *et al.* 1997; Möller *et al.* 2003). High-grade gneissosity is deformed by upright folds with a near vertical, north–south trending foliation defined by sillimanite and biotite in garnet-cordierite-quartz metapelites (Hand *et al.* 1999). In the eastern part of the Strangways Metamorphic Complex, Strangways-age metamorphism is consistent with near isobaric heating-cooling paths that reach peak metamorphic conditions of 2.6–4.0 kbar and 750–800 °C (Deep Bore Metamorphics) with metamorphic zircon rims recording a SHRIMP U–Pb age of 1730 ± 7 Ma (Scrimgeour *et al.* 2001; Scrimgeour & Raith 2002). A north–south trending upright sillimanite-bearing fabric (Scrimgeour *et al.* 2001) post-dates melt crystallization, and either represents a second event after minor isobaric cooling or a pressure increase after melt crystallization in an anticlockwise P–T path (Scrimgeour *et al.* 2001). Metamorphic grade decreases to greenschist facies to the NW (Shaw *et al.* 1975; Warren & Hensen 1989; Scrimgeour *et al.* 2001; Scrimgeour & Raith 2002).

Effects of the Strangways Orogeny extend with decreasing intensity into the Tanami in the form of a NW-trending belt of magmatism (Scrimgeour 2003). Within the Tennant Creek region there is Strangways-age magmatism and low-grade metamorphism (Compston & McDougall 1994; Compston 1995). The above summary of the evidence for the Strangways Orogeny highlights the approximate north–south trend of the compressive phase of the orogeny, supported by north–south trending structures and east–west decrease in metamorphic grade, and the limited east–west extent of *c.* 1730–1690 Ma deformation in the southern North Australian Craton (Scrimgeour 2003). Hence the Strangways Orogeny does not appear to provide strong evidence for a previously proposed active southern margin of the North Australian Craton at this time (Giles *et al.* 2002; Betts & Giles 2006).

Given the metamorphic and structural characteristics of the Kimban and Strangways orogenies we consider the reconstruction model in Figure 6g the most appropriate. This interpretation considers the current-day eastern margin of Proterozoic Australia to have undergone active rifting initiating at *c.* 1800 Ma. The conjugate rifted fragment may be Laurentian, if earlier collision at *c.* 1850 Ma (Fig. 6b, Option 2) brought the North Australian Craton and Laurentia together. If so, this rifting is similar to that proposed by Betts *et al.* (2002). Following the collision of the West and North Australian Craton, and possibly related to this event, active rifting ceased and the consumption of oceanic crust commenced in a subduction zone to the west of current-day Laurentia. This resulted in the accretion of proto-Australia to the western margin of the Laurentian plate at *c.* 1730–1720 Ma in the configuration allowed by palaeomagnetic constraints (Fig. 4). We suggest this collision was centred around the margin of the Gawler–Adélie Craton crust, with margin geometry resulting in an initial dextral margin. In the proposed scenario we interpret the Miller Range to represent an orphaned fragment of Laurentia, possibly a fragment of the Slave Craton. Although the Archaean history of the Miller Range is extremely poorly-known it is not inconsistent with that of the Slave Craton, as both terrains contain evidence for pre-3100 Ma magmatic lithologies and *c.* 2980 Ma metamorphism (Fig. 7). An interpreted period of intra-orogenic crustal relaxation and/or extension is marked by sedimentation and volcanism within the Gawler–Adélie Craton (*c.* 1715–1710 Ma, Labyrinth Formation) prior to a second period of high-grade metamorphism at *c.* 1700–1690 Ma (Fanning *et al.* 2007; Payne *et al.* 2008). This is interpreted to represent a second episode of compressional deformation, either within a single collisional event or perhaps representing the consumption of a back-arc basin and accretion of an arc and proto-Australia to the main Laurentian continent.

Sedimentary basins in the Mt Isa Inlier in northern Australia record differing histories during the 1730–1690 Ma period. Within the western Mt Isa region a basin inversion event at *c*. 1740–1710 Ma represents east–west shortening (Betts 1999) and is followed by a short period of sedimentation (including basal conglomerates) and volcanism (Bigie Formation and Fiery Creek Volcanics, *c*. 1710 Ma). Subsequent deposition of the Prize Supersequence does not commence until *c*. 1688 Ma (Neumann *et al*. 2006). These event timings correlate well with periods of compression and the intervening sedimentation/volcanism recorded in the eastern Arunta Region and Gawler–Adélie Craton. The Eastern Fold Belt of the Mt Isa Inlier does not record evidence of basin inversion in the 1740–1710 Ma period but, in the revised stratigraphy of Foster and Austin (2008), also did not undergo rifting and sedimentation in the period 1730–1680 Ma. The mechanisms for this difference in tectonic history are uncertain. We provide a speculative hypothesis that the Kalkadoon/Leichardt belt basement inlier may have provided some form of structural continuity with deforming crust to the south, resulting in weak compression of crust to the west of the inlier, i.e. the Western Fold Belt.

Evidence to support the adopted model linking Australia and Laurentia is found in the Yukon region of northwestern Laurentia. There the Wernecke Supergroup is deposited some time during the *c*. 1840–1710 Ma period, intruded by the *c*. 1710 Ma Bonnet Plume River intrusions, deformed during the Racklan Orogeny and unconformably overlain by the Slab Volcanics (Thorkelson *et al*. 2001, 2005; Laughton *et al*. 2005). The Racklan Orogeny produced tight north-trending, east-verging folds that are related to the main schistosity development with peak metamorphic temperatures of 450–550°C (Laughton *et al*. 2005; Thorkelson *et al*. 2005 and references therein). These structures were overprinted by open to tight, south-verging folds. The timing of the Racklan Orogeny is uncertain but is pre-1600 Ma as constrained by the cross-cutting Wernecke Breccia (Laughton *et al*. 2005). Thorkelson *et al*. (2005) note that 'few if any of the (Bonnet Plume) intrusions' are foliated, suggesting the possibility of a post-kinematic petrogenesis; however, given the relatively low grade of deformation and metamorphism the apparent lack of foliation does not preclude a pre-kinematic petrogenesis. The model proposed here suggests that the initial east–west compression relates to the accretion of proto-Australia to the Laurentian plate. The Racklan Orogeny may be correlated to the intracratonic Forward Orogeny further to the east (Cook & MacLean 1995; Thorkelson *et al*. 2005). The Forward Orogeny is represented by NW–SE directed compression with south- and north-directed vergent folds and thrust faults followed by later wrench faulting along more northerly directed faults (Cook & MacLean 1995). The Forward Orogeny is constrained by syntectonic sedimentation and volcanism at 1663 ± 8 Ma (Bowring & Ross 1985). The apparent difference in direction of compression between the two orogenic systems highlights the uncertainty regarding the correlation between the Racklan and Forward orogenies, and the need for direct geochronological constraints on Racklan Orogeny metamorphism and deformation.

Palaeomagnetic constraints for proto-Australia with respect to Laurentia result in an open passive margin east of the Georgetown/Curnamona region in the rotated model (Fig. 4). Although potentially a fortuitous coincidence, this palaeogeometry allows for the extension and basin development in the Georgetown/Curnamona region as recorded by the Etheridge and Willyama sequences and associated magmatism (Black *et al*. 2005; Stevens *et al*. 2008 and references therein). Recorded extension commenced at *c*. 1700 Ma in the Georgetown region and *c*. 1720–1715 Ma in the Curnamona Province. Basin formation within the Curnamona Province is approximately synchronous with the period of short-lived volcanism and sedimentation recorded in the Gawler–Adélie Craton (Labyrinth Formation and Point Geologie migmatite protolith) and Western Fold Belt of the Mt Isa Inlier. Unlike the western Mt Isa Inlier, sediment deposition continued in the Curnamona Province until *c*. 1690–1680 Ma before a hiatus in deposition until *c*. 1650 Ma (Conor 2004; Page *et al*. 2005; Stevens *et al*. 2008 and references therein). The Georgetown Inlier appears to have recorded continuous deposition until Mesoproterozoic orogenesis. The continuation of rifting and sedimentation after *c*. 1720 Ma is interpreted to represent the continued extension of the northeastern North Australian Craton crust due to the continued consumption of oceanic crust within a subduction zone further to the east or northeast of proto-Australia.

A potential problem with the model proposed here, is the possibility for Siberia to occupy a position to the NW of Laurentia during the Mesoproterozoic (Fig. 4), as would be the case for the hypothesized extension of Rodinia reconstruction models backwards in time to the early Mesoproterozoic (Frost *et al*. 1998; Rainbird *et al*. 1998; Pisarevsky & Natapov 2003; Pisarevsky *et al*. 2008). This would remove the possibility of an ocean-facing margin to the east of northeastern Australia and hence require alternative mechanisms to explain the continued extension (1700–1600 Ma) and subsequent *c.* 1550 Ma orogenesis in the Mt Isa, Georgetown and Coen Inlier regions.

Circa 1690–1620 Ma accretion, UHT metamorphism and sedimentation. The interpreted accretion of the Warumpi Province to the southern margin of the Arunta Region at *c.* 1640–1620 Ma (Scrimgeour *et al.* 2005) provides the only direct evidence for plate margin processes during the *c.* 1690–1620 Ma time period. The proposed model considers this to have occurred via south-dipping subduction beneath the Warumpi Province initiating shortly after the Kimban–Strangways Orogeny (Close *et al.* 2005; Scrimgeour *et al.* 2005). This model provides a petrogenetic framework for *c.* 1690 Ma interpreted subduction-related magmatism in the Warumpi Province (Close *et al.* 2005; Scrimgeour *et al.* 2005). It also satisfies geophysical constraints that image a south-dipping boundary in the lithospheric mantle extending beneath the Warumpi Province (Selway 2007). Given the potential antiquity of hypothesized oceanic lithosphere between the Gawler–Adélie Craton and West Australian Craton subduction initiation may have occurred through oceanic plate foundering with subsequent subduction-zone rollback resulting in the accretion of the Warumpi Province to the Arunta Region. Alternatively, in the proposed model the advancement of the Warumpi plate is interpreted to relate to the docking of the proto-Australian plate with Laurentia, re-establishing the previous regime of differing relative movement of proto-Australia and oceanic lithosphere to its south. The transform margin interpreted along the southwestern margin of the Gawler–Adélie Craton provides the mechanism for advancement of the Warumpi Province and is also interpreted to relate to the formation and exhumation of the Ooldea ultra-high temperature granulite lithologies at *c.* 1660 Ma.

Convergent tectonism at *c.* 1650–1620 Ma is largely absent from the remainder of the Mawson Continent and West and North Australian cratons. Minor basin formation is evidenced at *c.* 1650 Ma within the Gawler–Adélie Craton, leading to the deposition of the Tarcoola Formation. The McArthur Basin, Mt Isa Inlier, Curnamona Province and Georgetown Inlier all underwent active sedimentation during the *c.* 1660–1600 Ma period, and do not appear to record basin inversion associated with the accretionary Leibig Orogeny (Black *et al.* 2005; Betts *et al.* 2006; Neumann *et al.* 2006; Foster & Austin 2008; Stevens *et al.* 2008 and references therein). The continued basin development is interpreted to reflect the continued extensional nature of the northeastern margin of proto-Australia. Palaeomagnetic constraints allow static relative positions of proto-Australia and Laurentia until at least *c.* 1590 Ma (Fig. 4), suggesting the two continents did not separate prior to then. We interpret this to indicate the continued extension within northeastern proto-Australia may have been related to continued consumption of oceanic crust north of the Laurentia plate until final *c.* 1550 Ma collisional orogenesis.

Circa 1620–1500 Ma arc-magmatism and orogenesis. Subduction-related magmatism was initiated in the southern Gawler Craton at *c.* 1620 Ma as represented by the St Peter Suite (Flint *et al.* 1990; Fanning *et al.* 2007; Swain *et al.* 2008). This is interpreted to represent the onset of subduction outboard of the Warumpi accretion, progressing to northward-dipping subduction under the trailing edge of the Warumpi Province at *c.* 1600 Ma. This results in the formation of the interpreted arc-related magmatism in the Musgrave Province from 1.6–1.55 Ga (Wade *et al.* 2006). A north-dipping subduction zone may also provide mechanisms for *c.* 1590–1550 Ma magmatism in the Rudall Complex (Maidment & Kositcin 2007). Convergence along the southern margin of proto-Australia is synchronous with the early stages of the intracratonic Olarian, Isan and Chewings orogenies (Hand 2006). Shortening directions for these events are represented in Figure 8. For orogenesis in the North Australian Craton, the initial phase of orogenesis represents north–south or NW–SE compression constrained to *c.* 1590–1570 Ma (Hand & Buick 2001; Betts *et al.* 2006). Early Olarian Orogeny structures are typically north–south to NE–SW trending and are associated with peak metamorphism (Wilson & Powell 2001; Conor 2004 and references therein). Rotation of the Gawler–Adélie Craton and Curnamona Province into the alignment utilized in the proposed model results in the early Olarian Orogeny structures implying an approximately north–south to NE–SW compression direction. Although a simplistic assessment of the palaeo-tectonic stress orientations, this broadly correlates to the transport direction recorded in the inliers of the North Australian Craton (Hand & Buick 2001). *Circa* 1600–1580 Ma orogenesis within Australia is consistent with compression caused by the proposed north-dipping subduction beneath the Musgrave Province. Alternatively or additionally, as hypothesized by Betts & Giles (2006), intracratonic orogenesis at *c.* 1600–1580 Ma may be linked to a collision on the palaeo-southern margin of the Gawler–Adélie Craton in which crust now forming the Coompana Block was accreted. This is potentially consistent with the cessation of St Peter Suite magmatism at *c.* 1600 Ma. Recent evidence for crustal thickening and anatexis reported from the southern

Gawler–Adélie Craton (Payne 2008) may also support such a scenario. As discussed by Betts & Giles (2006), the petrogenesis of the voluminous GRV and Hiltaba Suite can be considered to represent a mantle plume coincident with tectonothermal activities occurring in the over-riding plate and may not be genetically linked to these events.

The late Isan Orogeny and related events record a distinct second phase of deformation. In the Mt Isa, Georgetown and Coen inliers, the principal direction of compression is approximately east–west at 1560–1550 Ma (Blewett & Black 1998; Boger & Hansen 2004). The westerly directed transport direction is consistent with a collision along the northeastern margin of proto-Australia (Betts & Giles 2006), who suggest this event represents collision of Australia with Laurentia. This proposed scenario is not supported by our interpretation of the existing palaeomagnetic constraints as Australia and Laurentia would already have been amalgamated at this stage. Currently the colliding terrain is unidentified. As outlined in Betts & Giles (2006), the potential exists for subduction to have occurred beneath the Georgetown Inlier leading up to 1550 Ma orogenesis based upon granite geochemistry (Champion 1991).

Subsequent to *c*. 1550 Ma tectonism, proto-Australia does not appear to have undergone orogenesis until the *c*. 1300–1100 Ma Musgrave and Albany-Fraser (Camacho & Fanning 1995; Condie & Myers 1999; Bodorkos & Clark 2004*b*). As summarized in Betts & Giles (2006), the proto-Australian continent underwent a period of extension, basin formation and magmatism in the *c*. 1500–1400 Ma period. In the Mawson Continent this includes *c*. 1500 Ma anorogenic magmatism in the Coompana Block (Wade *et al*. 2007) and the formation of the Cariewerloo Basin in the Gawler–Adélie Craton (Cowley 1993). Palaeomagnetic constraints for *c*. 1070 Ma (Wingate *et al*. 2002*b*) indicate proto-Australia and Laurentia did not stay amalgamated in their *c*. 1590 Ma configuration for the duration of the Mesoproterozoic, suggesting the two separated at some point in the 1590–1070 Ma period. As suggested by Betts & Giles (2006), this may have occurred during the 1500–1400 Ma time period. However, as outlined above, palaeomagnetic data permit the continued association of northern Australia and Laurentia until as late as *c*. 1200 Ma.

Conclusions

In constructing the proposed model we have attempted to draw together geological and palaeomagnetic constraints to provide an internally consistent reconstruction of the evolution of the Mawson Continent and associated proto-Australia terrains. The model highlights the presence of comparable timelines in the North Australian Craton and Gawler–Adélie Craton for the duration of the Palaeoproterozoic. This is interpreted such that the two 'cratons' formed a single entity in the late Archaean to middle Proterozoic. As with other recently proposed models (Betts *et al*. 2002; Giles *et al*. 2004; Betts & Giles 2006), this model highlights the complexity and longevity of interaction between the so-called North and South Australian cratons.

The identification of reliable palaeomagnetic data for the *c*. 1750–1730 Ma and *c*. 1595 Ma timelines provides new constraints on the palaeogeometry of proto-Australia and Laurentica. These data indicate Australia and Laurentia may have been contiguous from *c*. 1730–1595 Ma. The proposed geometry lends additional weight to previous suggestions (Thorkelson *et al*. 2005 and references therein) for correlations between basement geology of the Wernecke Mountains and the eastern Australian Proterozoic.

Although the proposed model will probably not provide the ultimate answer to the evolution of the Mawson Continent and Australia, it is hoped that it will provide some insight into poorly understood events within Australia, and will stimulate further investigations on continental reconstruction models for the Palaeoproterozoic.

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