Tectonic evolution of Proterozoic Australia

John S. Myers, Russell D. Shaw, and Ian M. Tyler

Abstract. Proterozoic Australia has long been interpreted as a single intact continent in which all tectonic and magmatic activity was intracratonic. This paper proposes an alternative hypothesis in which numerous fragments of continental crust were assembled by plate tectonic processes. The assembly was completed between 1300 and 1100 Ma when the crustal fragments were combined as an early component of the Rodinian supercontinent. Rifting and fragmentation of Archaean continents began in the late Archaean and continued into the Proterozoic. Passive margin deposits, such as those of the Hamersley Basin, accumulated on isolated fragments of Archaean crust. These numerous fragments were subsequently assembled into three cratons by ~1830 Ma. A West Australian Craton was established by collision of the Archaean Pilbara and Yilgarn cratons, which were joined along the Capricorn Orogen. Similarly, a South Australian Craton developed by amalgamation of the proto-Gawler and proto-Curnamona cratons along the Kimban Orogen. A North Australian Craton appears to have formed by accretion of numerous crustal fragments, including the Kimberley, Pine Creek, Lucas, and Altjawaarra cratons, with sutures marked by the King Leopold, Halls Creek, Tennant Creek and proto-Isan orogens. The southern margin of the North Australian Craton was the site of repeated terrane accretion and orogenic activity between ~1880 Ma and 1400 Ma. This included an orogenic event at ~1880 - 1850 Ma; the Strangways (1780 - 1730 Ma), Argilke (1680 - 1650 Ma), and Chewings (1620 - 1580 Ma) orogenies; and the intracratonic Anmatjira uplift (1500 - 1400 Ma). Intracratonic rifting at ~1750 to 1710 Ma and ~1640 to 1600 Ma produced the McArthur Basin and related minor basins, parts of which were deformed by the Isan Orogeny at ~1600 and ~1550 Ma. Rifting along the line of the Capricorn Orogen led to deposition in the overlying intracratonic Bangemall Basin between 1630 and 1300 Ma. Along the eastern margin of the South Australian Craton, the 1670 to 1600 Ma Olarian Orogeny marks interaction with now obscured continental crust to the east. Tectonic activity between 1300 and 1100 Ma led to the assembly of Proterozoic Australia as an early component of the supercontinent of Rodinia. This first involved the amalgamation of the West Australian and North Australian cratons, followed by collision with the South Australian Craton. The Centralian Superbasin developed over the junction of the North, South, and West Australian cratons between ~830 and 750 Ma. Rifting to the east formed the "Adelaide Geosyncline" at ~830 Ma. This was followed by the breakup of Rodinia, with the rifting apart of Laurentia and Gondwanaland along the eastern margin of Proterozoic Australia at ~750 Ma, and the subsequent formation of the Palaeo-Pacific Ocean. After the breakup of Rodinia, a series of northeast-southwest compressional events followed by periods of relaxation, reflect the assembly of a new supercontinent. Old lines of weakness were reactivated, culminating in the intracratonic King Leopold, Paterson, Petermann Ranges, and Pinjarra orogenies between 620 and 540 Ma. Subsequent reactivation continued into the Phanerozoic, with the widespread eruption of continental flood basalts and the formation of intracratonic basins (540 - 530 Ma).

Introduction

The Proterozoic geology of Australia has long been thought to reflect the evolution of a single intact raft of crust. This interpretation was largely based on palaeomagnetic evidence [e.g., Veevers and McElhinny, 1976; Idnurm and Giddings, 1988], and a supposed lack of evidence for the subduction of oceanic lithosphere. Geophysical, geochemical, structural, and metamorphic evidence was also combined to establish models in which all tectonic and magmatic activity was intracratonic, dominated by vertical and extensional tectonic processes [Rutland, 1973; Gee, 1979; Etheridge et al., 1987].

This essentially "stabilist" view of the Proterozoic evolution of Australia contrasts with the more dynamic interpretations of the Proterozoic of North America, Europe, and Africa. Models such as those of Hoffman [1988, 1989, 1991] and Windley [1993] envisage the aggregation of early and middle Proterozoic continents by plate tectonic processes similar to those operating at the present. These culminated in the formation of a middle and late Proterozoic supercontinent called Rodinia [McMenamin and Schulte McMenamin, 1990; Dalziel, 1992a; Hoffman, 1991].

Dynamic plate tectonic interpretations of substantial Proterozoic parts of Australia were proposed by Myers [1990a, 1993] who suggested that an early Proterozoic, West Australian Craton, was assembled from diverse Archaean crustal fragments by plate tectonic processes. Heterogeneity was also recognized in the pre-Phanerozoic crystalline basement of northern and central Australia by Shaw et al. [1995], on the basis of gravity and magnetic data.

Powell and Li [1994] cautioned that the currently available palaeomagnetic data for the early and middle Proterozoic do not provide unequivocal evidence as to whether or not all the Australian Proterozoic crust was a single entity during that period. The single apparent polar wander path constructed by Idnurm and Giddings [1988, 1995] connecting all the palaeopoles is only one of a number of possible interpretations of these data. Some of the shortcomings of the Proterozoic palaeomagnetic data from Australia are discussed by Idnurm and Giddings [1995] and Idnurm et al. [1995]. Their recent detailed palaeomagnetic study of the McArthur Basin suggests that this part of Australia was close to northwestern America at ~1700 - 1600 Ma, but does not demonstrate any relation with other Proterozoic parts of Australia.

We propose that the Proterozoic tectonic evolution of Australia can be interpreted as a consequence of plate tectonic processes by which various crustal fragments were combined, first into...
three separate continents by \( \sim 1830 \) Ma, and then into a single continent at \( \sim 1300 \) Ma. This model is consistent with the plate reconstructions of Dalziel [1991, 1992a], Hoffman [1991], and Moores [1991], and the recognition that, on a global scale, numerous continental fragments were amalgamated as a supercontinent between \( \sim 1300 \) and \( 1100 \) Ma. Thereafter basin formation and deformation were intracratonic and were controlled by reactivation of old lines of weakness.

This interpretation is based on the cumulative work of a large number of people. It incorporates recent geological and geophysical mapping by the Australian state and federal geological surveys, as well as many new, largely unpublished, observations and interpretations of the field evidence, and recent geochronology, geochemistry, and metamorphic and structural studies.

**Main Tectonic Features**

The main outcrops of Precambrian rocks in Australia are shown in Figure 1. Between these outcrops, the structure of the Precambrian basement below Phanerozoic cover has been interpreted from integrated aeromagnetic and gravity maps [Fraser et al., 1977; Shaw et al., 1995; Tucker et al., 1979; Wellman, 1988, 1992a, 1992b]. Proterozoic Australia comprises three main tectonic units, here called the North, West, and South Australian cratons (Figure 2). These units were independently accreted from older crustal fragments by \( \sim 1830 \) Ma. The present boundaries of these units are faults that truncate the structures within the units. Therefore, before \( \sim 1830 \) Ma, these units probably formed parts of larger continental masses.

The central Australian terranes (Figure 2) comprise a collage of strip-like crustal units that formed between \( \sim 1900 \) and \( 1300 \) Ma. They are analogous to the outboard terranes that accreted to the western margin of the North American continent during the Phanerozoic [Coney et al., 1980], and may indicate that the southern margin of the North Australian Craton faced an open ocean at that time. The ocean was closed during collision of the North and West Australian cratons at \( \sim 1300 \) Ma. The combined West and North Australian cratons were then joined, also at \( \sim 1300 \) Ma, to the South Australian Craton along the Albany-Fraser Orogen [Myers, 1993]. The West Australian Craton may also have been joined to greater India along the Pinjarra Orogen by \( \sim 1080 \) Ma [Bruguier et al., 1994].
Figure 2. The main Precambrian tectonic units of Australia: ~ 1830 Ma cratons; ~ 1900 to 1300 Ma central Australian terranes; and the ~ 1300 Ma Albany-Fraser orogen. The central Australian terranes were accreted onto the North Australian Craton prior to its ~ 1300 Ma suturing with the West and South Australian cratons.

The sequential tectonic development of Proterozoic Australia is illustrated a table (Figure 3), and by a series of maps (Figures 4 - 9 and 11 - 12), which are based on an integrated interpretation of the Precambrian geology seen in outcrop with that inferred below the intervening cover of Phanerozoic rocks. For each time period, the maps show: (1) older continental crust; (2) the main structures that were active; (3) the main areas of sedimentary deposition; and (4) areas of substantial volcanism. Gaps indicate areas, of unknown extent, inferred to have been occupied by either oceanic crust or lost fragments of continental crust, that separated the fragments of continental crust at the times indicated. The common legend for these maps is given in Figure 4.

**2500-1950 Ma Rifted Archaean Crustal Fragments**

At this time the Archaean continents were fragmented and passive margin deposits accumulated on Archaean crust (Figure 4). This is best illustrated from the West Australian Craton, where the Archaean Pilbara and Yilgarn cratons are well exposed. The two cratons have completely different Archaean histories, and the margins along which they were joined contain different sequences of sedimentary and volcanic rocks [Thorne and Seymour, 1991; Tyler, 1991; Myers, 1993].

The Pilbara Craton comprises a composite basement of 3500 - 2950 Ma granite-greenstone terranes unconformably overlain by 2765 - 2687 Ma flood basalt (Fortescue Group) and ~ 2500 - 2400 Ma banded iron formations (Hamersley Group) [Hickman, 1983; Arndt et al., 1991; Trendall, 1983]. In contrast, the Yilgarn Craton was not assembled until ~ 2650 Ma from diverse crustal fragments [Myers, 1995a], and is not overlain by banded iron formations of Hamersley Basin type.

Early Proterozoic rifting along the southern margin of the Pilbara Craton resulted in the deposition of shallow marine and deltaic deposits and the eruption of basalt, (lower part of the Wyloo Group) (Figures 4) [Thorne and Seymour, 1991].

The northern margin of the Yilgarn Craton is overlain by a different sequence of early Proterozoic clastic shelf deposits, including carbonates and evaporites, succeeded by volcanic-derived wackes interbedded with basalt erupted in shallow water (Yerrida Group) (Figure 4) [Pirajno et al., 1996].

In northern Australia, the Archaean continental fragments and associated sedimentary rocks (Figure 4) are partly inferred from airborne magnetic and gravity data [Shaw et al., 1995]. The only
exposed Archaean rocks are granite and granitic gneiss in the Rum Jungle and Nanambu complexes within the Pine Creek fragment (Figure 4) [Needham et al., 1988]. The Rum Jungle basement complex is overlain by the arkosic Namoona Group, which may represent platform cover deposited before 1950 Ma. In addition, U-Pb zircon geochronology and Nd isotopic evidence points to the presence of Archaean crust in the Granites - Tanami region within the Lucas Craton (Figure 4) [Page and Sun, 1994; Page et al., 1995]. These Archaean fragments are partly covered by arkosic and conglomeratic clastic rocks, similar to those of the Namoona Group. In the rest of northern Australia, the inferred Archaean ages of the basement fragments and their inferred Proterozoic sedimentary cover (Figure 4) are partly supported by Sm-Nd and Pb isotope data [Page et al., 1984; Windrum and McCulloch, 1986; Sun et al., 1995].

In South Australia, the oldest known parts of the proto-Gawler Craton (Sleaford and Mulgathing complexes) comprise late Archaean orthogneiss, paragneiss, and minor greenstones, intruded by granite, granodiorite, and tonalite [Daly and Fanning, 1993; Daly et al., 1996]. These rocks were involved in deformation and granulite facies metamorphism during the Sleafordian Orogeny between ~2640 and 2300 Ma (Figure 3) [Daly and Fanning, 1993].

1950-1700 Ma Assembly of the West, North and South Australian Cratons

During this period, diverse older crustal fragments were assembled into three cratons (Figure 2). The best exposed record of this episode of crustal assembly is seen in the West Australian Craton where the Archaean Pilbara and Yilgarn cratons were joined along the Capricorn Orogen (Figures 3 and 5) [Myers, 1990a; Tyler and Thorne, 1990].

On the southern margin of the Pilbara Craton, the shelf deposits of the lower Wyloo Group (Figure 4) are deformed in a north-verging fold-and-thrust belt that resulted from oblique collision of this margin with continental crust to the south [Tyler and Thorne, 1990]. Related uplift and erosion led to the deposition of conglomerate, sandstone and mudstone followed by carbonate on a subsiding shelf (lower part of upper Wyloo Group) [Tyler and Thorne, 1990; Thorne and Seymour, 1991].

Intensification of the collision led to recumbent folding and northward-directed thrusting of the foreland basin deposits.
To the south, the center of the orogen is dominated by strongly deformed, fault-bounded belts of early Proterozoic granitic gneiss, high-grade metasedimentary rocks and granitic plutons (collectively called the Gascoyne Complex) [Williams, 1986]. Granite is especially abundant in a 75-km-wide belt that may represent a cordilleran batholith (Minnie Creek Batholith) above a southward-dipping subduction zone [Myers, 1990b, 1993]. Further south, rocks of the Gascoyne Complex were transported southward by thrusts onto Archaean gneisses and granites on the edge of the Yilgarn Craton [Myers, 1989, 1993].

Clastic shelf deposits and rift-related basalt (Yerrida Group) (Figure 4), deposited on Archaean basement along the northern
margin of the Yilgarn Craton, were overridden by southeast-directed thrusts (Figure 5). The thrusts carried slices of mafic and ultramafic plutonic and volcanic rocks (Trillbar Complex and Narracoota Volcanics) (including boninites Hynes and Gee, [1986]) that may represent an obducted part of a volcanic arc [Myers, 1989]. These thrust slices were overridden by a thrust carrying clastic sedimentary rocks (Padbury Group) [Martin, 1994; Occhipinti et al., 1996] forming a piggyback basin. These rocks and structures are unconformably overlain by marine shelf deposits of the Earaheedy Group (Figure 5), which accumulated on the southern margin of an intracratonic rift developed on the site of the Capricorn Orogen.

Between 1800 and 1700 Ma, deformation continued within the Capricorn Orogen (Figure 6) with dextral strike-slip faulting along the southern margin of the Pilbara Craton, and sinistral strike-slip faulting along the northern margin of the Yilgarn Craton, accompanying the westward extrusion of material caught between the oblique margins of the two cratons [Tyler and Thorne, 1990]. Sandstone and conglomerate (Capricorn Formation and Mount Minnie Group) (Figure 6) [Thorne and Seymour, 1991] were deposited in pull-apart basins. Southeasterly-directed extension at the southeastern corner of the Pilbara Craton resulted in the deposition of arkosic sandstone and conglomerate (Bresnahan Group) (Figure 6) [Hunter, 1990] in fault-bounded basins.

A small raft of sialic crust (Rudall Complex, Figure 6), comprising early Proterozoic metasedimentary and metavolcanic rocks, was intruded by granite at ~ 1790 Ma [Nelson 1995]. The rocks were strongly deformed and recrystallized at high-grade and then intruded by more granite between 1790 and 1760 Ma [Nelson, 1995]. Further deformation led to west-verging thrusts, recumbent folds and high-grade, high-pressure metamorphism [Smithies and Bagas, 1996]. The Rudall Complex is now in fault contact with the eastern edge of the Archaean Pilbara Craton, but its relative position at ~ 1790 - 1760 Ma is unknown.

In northern Australia, the oldest rocks exposed along the southern and western margins of the Kimberley Craton (Figure 4) are turbiditic sandstone, siltstone, and mudstone (Marboo Formation) that were deposited at ~ 1870 Ma [Griffin and Tyler, in press]. To the southeast, mafic volcanic, siliciclastic and carbonate rocks (Tickalara Metamorphics) formed before ~ 1865 Ma when they were deformed in the Hooper Orogeny (Figure 5) [Tyler et al., 1995]. Between ~ 1865 and 1850 Ma the protoliths of the Tickalara Metamorphics were intruded by numerous acid and basic to intermediate sheets with geochemical signatures indicative of subduction [Ogasawara, 1988; Sheppard et al., 1995; Tyler and Page, 1996], together with layered mafic-ultramafic bodies [Hooton and Tyler, 1993; Hooton, 1993; Page et al., 1995]. The Tickalara Metamorphics and associated intrusions may have formed as a volcanic arc or fore-arc along the eastern...
1830–1700 Ma

Figure 6. 1830–1700 Ma intracratonic deformation and accretion of central Australian terranes.

margin of the Kimberley Craton. To the northwest, within the edge of the craton, voluminous granitic and gabbroic rocks (Bow River batholith) and comagmatic, dacitic and rhyodacitic volcanic rocks (Whitewater Volcanics) were emplaced into and over the Marboo Formation between ~ 1865 and 1850 Ma [Griffin and Tyler, in press; Griffin et al., 1994]. This may represent an Andean-type magmatic arc, developed in an extensional setting on the margin of the continent, above a northwesterly-dipping subduction zone.

The Tickalara Metamorphics were deformed and metamorphosed during the Hooper Orogeny (Figure 5) [Tyler and Page, 1996], probably during thrusting of the volcanic arc or fore-arc onto the margin of the Kimberley Craton. They are overlain by felsic and mafic volcanics (Koongie Park Formation) that were erupted during an episode of rifting at ~ 1840 Ma [Griffin and Tyler, 1992; Page et al., 1994].

Along the eastern passive margin of the Lucas Craton (Figure 4), extension led to the eruption of mafic and felsic volcanic rocks (Ding Dong Downs Volcanics) and intrusion of high-level granitic magma at ~ 1910 Ma [Tyler et al., 1995]. Unconformably overlying conglomerate and sandstone (Saunders Creek Formation) form the basal unit of the ~ 1880 Ma lower part of the Halls Creek Group, comprising mafic and minor felsic volcanics and associated clastic and carbonate-rich metasedimentary rocks [Page and Sun, 1994; Tyler et al., 1995]. In the same region, trachyte and trachyandesite produced by submarine alkaline volcanism between 1870 and 1855 Ma may reflect further extension along this margin [Tyler et al., 1995; Page and Sun, 1994]. These volcanics are overlain by mudstone, siltstone, and sandstone (upper part of the Halls Creek Group) that formed a submarine fan on the margin of the Lucas Craton.

The Kimberley and Lucas cratons (Figure 4) were joined during the Halls Creek Orogeny (Figure 3 and 6) to form part of the North Australian Craton at ~ 1830 Ma [Tyler and Page, 1996]. During and immediately following this continental collision, acid and basic to intermediate plutons were intruded [Griffin et al., 1994; Sheppard et al., 1995], stitching the suture. Large layered mafic-ultramafic bodies were intruded into the same belt at the same time [Matheson and Hamlyn, 1987; Hoatson and Tyler, 1993; Hoatson, 1993; Page et al., 1995]. Folding and southeasternly directed thrusting accompanied metamorphism in the central part of the orogen as the Tickalara Metamorphics were overridden by the Kimberley Craton.

The Kimberley Basin (Figure 6) overlies inferred Archaean crust forming the Kimberley Craton (Figure 4). The lower part of the basin (Speewah Group) consists of conglomerate, sandstone, siltstone, mudstone, and minor ~ 1835 Ma felsic igneous rocks [Page and Sun, 1994], derived from the elevation and erosion of the central part of the Halls Creek Orogen during late stages of the collision between the Kimberley and Lucas cratons. By 1800 Ma northerly-derived conglomerate, sandstone, siltstone, mudstone, stromatolitic carbonate, and mafic volcanic
rocks (Kimberley Group) were shed from the Kimberley Craton, across the orogen, onto the margin of the Lucas Craton [Plumb and Gemuts, 1976]. A huge volume of mafic magma (Hart Dolerite) was intruded throughout the Kimberley Basin at ~ 1800 Ma [Page and Sun, 1994]. This magma may have been derived from a mantle plume associated with disruption of the Kimberley Craton from continental crust to the north and northwest [Griffin and Tyler, in press].

Extension of the northeastern part of the North Australian Craton between ~ 1835 and 1780 Ma led to the deposition of clastic sedimentary rocks together with basaltic volcanism in the Davenport and Mount Isa rifts (Figure 6) (Hatches Creek and Haslingden groups respectively). In the western part of the North Australian Craton, felsic magmatic activity occurred between ~ 1835 and 1815 Ma in the Pine Creek and Granites-Tanami regions (Figure 6), and again at 1790 Ma in the Granite-Tanami region and in the southern part of the Halls Creek Orogen [Stuart-Smith et al., 1993; Page and Sun, 1994].

A belt of mafic to intermediate granulite (Narweetooma metamorphic complex) occurs along the southern margin of the North Australian Craton. These granulites are marked as mafic rocks in the Arunta Complex on Figure 5. They may be derived from a volcanic arc that formed during or immediately before the emplacement of ~ 1880 - 1860 Ma calcalkaline granite (Atnarpna Complex), reflecting subduction and consumption of oceanic crust along the southern margin of the North Australian Craton [Zhao and Cooper, 1992; Zhao and Bennett, 1995].

There was another episode of magmatic activity along the southern margin of the North Australian Craton between ~ 1820 and 1790 Ma. This included bimodal volcanics, gabbro and anorthosite in the Arunta Complex (Figure 6) [Shaw et al., 1984]. Studies of the petrology, geochemistry and Sm-Nd isotopes of these intrusions interpret the environment as a magmatic arc [Foden et al., 1988; Shull, 1988]. Nd isotopic data suggest that a substantial amount of these intrusions are derived from newly generated crust [Sivell and McCulloch, 1991]. These events may reflect the development of volcanic and magmatic arcs along, or to the south of, the southern margin of the North Australian Craton.

This magmatic activity was immediately followed between 1780 and 1730 Ma by the Strangways Orogeny (Figure 6) [Collins and Shaw, 1995]. Granitic plutons were emplaced during the orogeny, especially between 1770 and 1750 Ma [Zhao and Bennett, 1995; Zhao and McCulloch, 1995; Sun et al., 1995]. They were followed between 1750 and 1730 Ma by deformation and high-grade metamorphism [Collins and Shaw, 1995]. The central part of the Arunta Complex is dominated by northwest- and west-directed thrusts that formed in a dextral transpressive regime [Collins and Shaw, 1995; Goscombe, 1991]. This deformation may be related to oblique convergence and accretion of magmatic arcs along this margin of the North Australian Craton.

In the northern part of the North Australian Craton, extension and deposition continued until ~ 1710 Ma in both the McArthur and Mount Isa rifts (Figure 6). Granite was intruded in the region of the Davenport rift at 1720 - 1710 Ma.

Along the eastern margin of the proto-Gawler Craton (Figure 4) in South Australia, clastic shallow marine sedimentary rocks, volcanics, carbonates and banded iron formations (Hutchison Group) (Figure 5) were deposited between ~ 1950 and 1840 Ma [Parker, 1993]. These rocks were deformed, metamorphosed at high-grade, and intruded by large volumes of granite (Lincoln Complex) during the Kimban Orogeny between ~ 1845 and 1700 Ma [Parker, 1993], associated with collision and amalgamation of the proto-Gawler and proto-Curnamona cratons to form the South Australian Craton (Figures 4 - 6). Associated volcanic and sedimentary rocks were deposited at ~ 1790 Ma (ryholite and rhyodacite, Myola Volcanics) [Fanning et al., 1988], ~ 1780 Ma (Peake Metamorphics derived from quartzite and volcanic rocks), 1765 - 1735 Ma (Wallaroo Group of shallow marine sedimentary and volcanic rocks) and at ~ 1723 Ma (shallow marine to fluvial Labyrinth Formation) [Daly et al., 1996].

### 1700-1600 Ma Accretion and Intracratonic Deformation

Deformation continued along the southern margin of the North Australian Craton (Figure 7). Major compression in the Argilke event between ~ 1680 and 1650 Ma may reflect accretion of a strip of continental crust onto this margin of the craton. This strip of continental crust is inferred by geophysical data [Shaw et al., 1995]. The compression was followed by extension between ~ 1650 and 1620 Ma, and then further compression in the intracratonic Chewings Orogeny from ~ 1620 to 1580 Ma (Figure 7) [Warren and Shaw, 1995; Collins and Shaw, 1995].

Older lines of weakness continued to be reactivated within the North Australian Craton (Figure 7). Granites were emplaced in the Mount Isa belt in an extensional regime between ~ 1680 and 1650 Ma [Connors and Page, 1995], followed by the deposition of clastic sedimentary rocks and the eruption of rhyolite at ~ 1630 Ma during further extension. Localized, renewed rifting occurred in the McArthur Basin at ~ 1640 Ma (Batten and Walker troughs) [Plumb et al., 1990; Etheridge and Wall, 1994; Page et al., 1994]. Deformation and renewed granite emplacement occurred in the Davenport belt [Blake and Page, 1988].

Shallow marine, immature clastic sedimentary rocks, and bimodal volcanics (Willyama Supergroup) (Figure 7) were deposited on the eastern margin of the South Australian Craton at ~ 1690 Ma [Parker, 1993; Page and Laing, 1992]. They were intensely deformed and metamorphosed at high-grade during the Olorian Orogeny between ~ 1670 and 1600 Ma [Page and Laing, 1992]. At about the same time, granites were emplaced along the former Kimban Orogen between ~ 1700 and 1600 Ma [Parker, 1993]. Shallow water sandstone, shale and dolomite, together with volcanic rocks ranging from basalt to rhyolite, were deposited as alluvial fans in grabens at ~ 1655 Ma (Tarcoola Formation) (Figure 7) [Daly et al., 1996].

The western margin of the South Australian Craton (Figure 7) is marked by the Karari Fault, considered by Daly et al. [1996] to represent the leading edge of another raft of sialic crust that was accreted onto the edge of the South Australian Craton at ~ 1660 Ma during the Kararin Orogeny. This raft of older crust (Archaean to early Proterozoic) (Nawa subdomain and possibly including the Coompana Block) is not exposed and is only known from drilling and geophysics [Parker, 1993; Daly et al., 1996].

Extensional faulting within the West Australian Craton after ~ 1630 Ma [Collins and McDonald, 1994], led to the deposition of clastic sedimentary rocks in the Bangermain Basin (Edmund Subgroup) over the suture between the former Pilbara and Yilgarn cratons.
1700 - 1600 Ma

In what follows, we describe the tectonic evolution of Proterozoic Australia from 1700 to 1600 Ma. This period is characterized by the accretion of South and North Australian cratons and intracratic deformation.

1600 - 1300 Ma Intracratic Deformation

Extensional faulting and associated clastic sedimentary deposition continued within the Bangemall Basin (1630 - 1450 Ma Edmund Subgroup) on the West Australian Craton (Figures 3 and 8). Sandstone, siltstone, shale and carbonates accumulated in shallow water and were intruded by numerous dolerite sills. This was followed by compressional deformation, and then by further deposition of similar shallow water sedimentary rocks (1450 - 1300 Ma Manganese and Collier subgroups) [Muhling and Brakel, 1985; Williams, 1990a].

Intracratic deformation continued within the North Australian Craton (Figure 8), and east-west compression led to intense and widespread deformation (D2) and associated metamorphism in the Isan Orogen culminating at ~ 1530 Ma [Connors and Page, 1995]. Faulting, and associated deposition of clastic sedimentary rocks, occurred in the Birrindudu Basin at or before ~ 1560 Ma [Sweet, 1977; Blake et al., 1979]. This was followed by deposition in the MacArthur Basin (Roper Group of sandstone, black shale and minor carbonate) at about 1430 Ma [Plumb et al., 1990]. At this time there was major north-south compression with sinistral shear along the older north-south faults within the Roper Group (Figure 8). Deformation along the southern margin of the craton led to the Anmatjira uplift along the Redbank Thrust (Figure 8) [Shaw and Black, 1991; Shaw et al., 1992a].

Deformation and metamorphism in the Georgetown inlier (Figure 8) at ~ 1550 Ma [Black and Withnall, 1993] may have been associated with its amalgamation with the North Australian Craton at this time.

On the South Australian Craton (Figure 8), there was widespread intracontinental volcanism with the eruption of rhyolite, dacite, andesite and basalt (Gawler Range Volcanics) and intrusion of associated granites (Hiltaba Suite) at ~ 1595 - 1585 Ma, perhaps related to a mantle plume [Campbell and Hill, 1991; Flint, 1993]. Felsic volcanism and granite intrusion also occurred in the eastern part of the craton (Mount Babbage, Mount Painter, and Willyama inliers) between 1600 and 1560 Ma [Flint, 1993]. Granulites within the Birksgate Complex (Figure 8), derived from 1600 - 1300 Ma felsic volcanics and sedimentary rocks [Major and Conor, 1993], may have originated along the northern margin of the South Australian Craton.

1300-1000 Ma Assembly of Australian Cratons as part of Rodinia

The North, South and West Australian cratons were combined at ~ 1300 (Figures 3 and 9) as an early component of the Rodinian supercontinent (Figure 10) [McMenamin and Schulte McMenamin, 1990; Dalziel, 1991, 1992a; Hoffman, 1991; Moores, 1991].

Figure 7. 1700 - 1600 Ma accretion to South and North Australian cratons and intracratic deformation.
The North Australian Craton was first joined to the northeastern margin of the West Australian Craton. During this assembly, the intervening Rudall Complex (Figure 9) was intruded by granite at ~ 1300 Ma and overlain by shallow marine alternations of sandstone, siltstone, shale, and dolomite (Yeneena Supergroup) between ~ 1300 and 1100 Ma [Williams, 1990b; Nelson, 1995]. These rocks were stacked by southwest-directed thrusts and metamorphosed at low-grade [Clarke, 1991; Smithies and Bagas, 1996]. The adjacent, older sedimentary rocks in the Bangemall Basin (Figure 8) also developed a related cleavage, and the intensity of this deformation and metamorphism decreased toward the southwest [Williams, 1990a].

The combined West and North Australian cratons were joined to the South Australian Craton along the Albany-Fraser Orogen (Figure 9). The orogen contains gabbro and granite that was emplaced, intensely deformed, and recrystallized in granulite facies between ~ 1300 and 1280 Ma [Myers, 1990a, 1993]. The Musgravian Orogen may represent a former continuation of the Albany-Fraser Orogen along the northern margin of the South Australian Craton that was later displaced westward during younger intracratonic deformation.

In the Albany-Fraser Orogen the northwestern margin of the South Australian Craton, comprising 2630 Ma, 1700-1600 Ma and 1300 Ma granites [Nelson et al., 1995], was shortened and stacked as thrust slices onto the margin of the West Australian Craton [Myers, 1993, 1995b,c]. Substantial crustal thickening led to regional granulite grade metamorphism at ~ 1300 - 1280 Ma. The repeated asymmetry of minor structures indicates that the collision was oblique, with the West Australian Craton moving northeastward and the South Australian Craton moving westward. A large volume of granite was emplaced as sheets and plutons into the southern and eastern parts of the Albany-Fraser Orogen between 1180 - 1140 Ma, and was deformed by continuing dextral movements under amphibolite facies conditions.

In the Musgravian Orogen, there was substantial crustal thickening and high-grade metamorphism at ~ 1200 - 1150 Ma [Maboko et al., 1991; Comacho and Fanning, 1995] of 1600 - 1300 Ma gneiss and 1250 - 1150 Ma granite during the Musgravian Orogeny (Figure 9) [Major and Conor, 1993]. This was followed by extensive granite intrusion at ~ 1190 - 1150 Ma. Thereafter, uplift and erosion was followed at ~ 1080 Ma by the deposition of post-tectonic basalt and rhyolite (Tollu Volcanics), the intrusion of associated high-level plutons of granite, and the emplacement of large layered intrusions of gabbro and ultramafic rocks (Giles Complex) (Figure 3) [Maboko et al., 1991; Ellis and Maboko, 1992; Sun and Sheraton, 1992], perhaps during exten-
Proto-Pinjarra Orogeny - WEST AUSTRALIAN CRATON 1300-1000 Ma

Yampi Orogeny

Yeneena Supergroup

Victoria River Basin 1200

Proto-Pinjarra Orogeny

WEST AUSTRALIAN CRATON

Rudall Complex

Northampton Complex 1080

Gnowangerup dykes

Northampton Complex 1080

Gnowangerup dykes

SOUTH AUSTRALIAN CRATON

Torrens Hinge Zone

Figure 9. 1300 - 1000 Ma assembly of Australian cratons as part of Rodinia.

Psammitic gneiss (Northampton Complex) within this orogen was deformed and metamorphosed to granulite facies at ~ 1080 Ma [Bruguière et al., 1994].
1000 - 750 Ma Intracratonic Basin

An extensive intracratonic basin (Centralian Superbasin) (Figure 11) developed over the junction between the North, South, and West Australian cratons at ~830 Ma [Walter et al., 1995]. It comprises shallow marine and fluvialite sandstone overlain successively by carbonates, evaporites, and glaciogene deposits. These deposits extend to the southeast into an epicontinental rift basin known as the Adelaide Geosyncline [Preiss, 1993; Powell et al., 1994a]. A major dyke swarm (Gairdner dykes) (Figure 11) intruded at ~800 Ma is related by Flint [1993] and Zhao et al. [1994] to a mantle plume centered beneath the Adelaide Geosyncline.

750 - 540 Ma Breakup of Rodinia

Jostling of the largely intracratonic plates during the breakup of the Rodinian supercontinent led to movements along older lines of weakness and related sedimentary deposition (Figure 12).

During the early stages of breakup at ~750 Ma, northeast-southwest compression was dominant [Shaw et al., 1991]. The Centralian Superbasin (Figure 11) was broken into a number of smaller fault-bounded basins (Amadeus, Ngalia, and Georgina basins) (Figure 12). These basins formed as a consequence of tilting and flexuring of fault blocks controlled by reactivation of structures, including the proto-Woodroffe and Redbank thrusts, during the Areyonga and South Range tectonic movements [Shaw et al., 1991].

Structures within the Capricorn Orogen were also reactivated in the Blake fold-and-thrust belt at about the same time [Williams, 1992], and major orthogonal dolerite dyke swarms were emplaced within the West Australian Craton (Figure 12) [Myers, 1993].

In the "Adelaide Geosyncline" (Figures 11 and 12), northeast-southwest extension accompanied the deposition of the Sturtian glacial deposits at ~700 to 725 Ma [Preiss, 1993]. Extension to the southeast along the southeastern margin of the Australian plate (the "Tasman Line") (Figure 12) at about 600 Ma may have been related to initiation of the Palaeo-Pacific Ocean [Powell et al., 1994a], with the northwesterly oriented orogenic belts being reactivated as major transfer fault zones. Deposition in the "Adelaide Geosyncline", including the Marinoan glacial deposits, took place on a passive continental margin [Powell et al., 1994a].

Major folding and thrusting, followed by substantial strike-slip movements, occurred between ~620 and 540 Ma during the Paterson and Petermann Ranges orogenies (Figures 3 and 12). These resulted from reactivation of the sutures between the North Australian Craton, and the South and West Australian cratons in a northeast-southwest compressive regime. This deformation may have been related to large-scale movements of plates and events at plate margins beyond the present extent of Australia (Figure 10) [Hoffman, 1991; Dalziel, 1992a and b]. Powell et al., [1994a and b] have previously related the reactivation to an anti-clockwise (dextral) rotation of northern Australia relative to southern Australia, involving dextral strike-slip or transpressive movements. However, this is inconsistent with the southwest- and northeast-directed thrusting along the northwest-trending
750–540 Ma intracratonic deformation during breakup of Rodinia.

King Leopold and Paterson orogens (Figure 12) [Tyler and Griffin, 1990; Myers, 1990a, 1993; Williams, 1992].

Thick sequences of sedimentary rocks were deposited in foreland basins associated with the Paterson and proto-Petermann Ranges orogenies [Myers, 1990a], and granite (Crofton Granite) was intruded into these sedimentary rocks at ~ 620 Ma [Nelson, 1995].

Major uplift along the Woodroffe Thrust at ~ 550 Ma [Maboko et al., 1991] did not result in a foreland basin of appropriate width in the Amadeus Basin (Figure 12) [Shaw et al., 1991]. This may reflect large sinistral strike-slip movements along the Woodroffe - Mann fault complex, truncating the Albany-Fraser Orogen and displacing the Musgravian Orogen to the west. Shallow thrusting may form part of a strike-slip related, crustal-scale, positive flower structure. Reactivation also occurred in northern Australia with development of the King Leopold Orogeny at ~ 560 Ma (Figures 3 and 12) [Tyler and Griffin, 1990; Shaw et al., 1992b].

In the southwest corner of Western Australia, granites were emplaced into the Leeuwin Complex at ~ 780 and 695 Ma (Figure 12) [Myers, 1994; Nelson, 1996]. They were intensely deformed and recrystallized to form granulite facies gneisses at ~ 615 Ma [Nelson, 1996] during the Pinjarra Orogeny, perhaps reflecting accretion and continental collision associated with closure of the Mozambique Ocean [Dalziel, 1992]. These gneisses were intruded by peralkaline granite at ~ 535 Ma [Myers, 1994; Nelson, 1996; Wilde and Murphy, 1990], during extensional collapse in a high thermal regime.

The Proterozoic rocks of southeastern Australia and Tasmania, east of the "Tasman Line" (Figure 12), appear to be unrelated to those of the North, West, and South Australian cratons. They may represent exotic terranes that formed elsewhere, and were accreted to the Australian plate during the Palaeozoic [e.g., Powell et al., 1994b]. The interpretation of these terranes has major consequences for supercontinent reconstructions, but the Proterozoic geology of these terranes is complicated because they occur as isolated remnants that were deformed and metamorphosed during the Paleozoic. Therefore they are not discussed in this paper.

In the latest Proterozoic to earliest Cambrian, at ~ 550 to 520 Ma, the southeastern margin of Australia may have undergone southeasterly directed extension in a back-arc setting inboard of the Palaeo-Pacific Ocean. The eruption of flood basalts (Antrim Plateau, Kulyong, and Table Hill volcanics), and the formation of the early Cambrian Ord, Bonaparte, Wiso, and Warburton basins in central and northern Australia may be related to this episode of extension [cf. Shaw et al., 1991], with the northwesterly oriented Proterozoic orogenic belts being reactivated as major transfer fault zones.

Conclusions

Proterozoic Australia comprises a number of rafts of continental crust that show a dynamic history of repeated aggregation and dispersal. The boundaries of these crustal fragments, once established, remained important zones of crustal weakness that were intermittently reactivated as intracontinental mobile belts, or lines along which continental crust was assembled and then pulled apart.
The main crustal components of Proterozoic Australia, the North, South, and West Australian cratons, were first assembled as parts of three separate continents by ~1830 Ma. The North Australian Craton formed part of a much larger continent, which may have included North America. Part of the former continental margin is preserved along the southern part of the craton. At different times this was both an active and passive margin. As an active margin it was the site of crustal accretion, including volcanic and magmatic arcs, and microcontinental slices. It was also the site of passive margin deposition, rifted continental margin volcanism, and major transcurrent faulting.

The South Australian Craton was part of a larger continent including East Antarctica, referred to by Daly et al. [1996] as the Mawson Continent. The West Australian Craton is also a fault-bounded fragment of a much larger continent, that may have included greater India.

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