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Evolution of the Australian lithosphere

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The evolution of the Australian plate can be interpreted in a plate-tectonic paradigm in which lithospheric growth occurred via vertical and horizontal accretion. The lithospheric roots of Archaean lithosphere developed contemporaneously with the overlying crust. Vertical accretion of the Archaean lithosphere is probably related to the arrival of large plumes, although horizontal lithospheric accretion was also important to crustal growth. The Proterozoic was an era of major crustal growth in which the components of the North Australian, West Australian and South Australian cratons were formed and amalgamated during a series of accretionary events and continent–continent collisions, interspersed with periods of lithospheric extension. During Phanerozoic accretionary tectonism, approximately 30% of the Australian crust was added to the eastern margin of the continent in a predominantly supra-subduction environment. Widespread plume-driven rifting during the breakup of Gondwana may have contributed to the destruction of Archaean lithospheric roots (as a result of lithospheric stretching). However, lithospheric growth occurred at the same time due to mafic underplating along the eastern margin of the plate. Northward drift of Australia during the Tertiary led to the development of a complex accretionary margin at the leading edge of the plate (Papua New Guinea).

KEY WORDS: Archaean, Australia, Gondwana, lithosphere, Palaeozoic, Proterozoic, tectonics.

INTRODUCTION

The lithosphere has been defined as the rigid outer part of the Earth where heat transfer occurs mainly by conduction (Lliboutry 1974; Turcotte 1987; Muirhead & Drummond 1991; O'Reilly *et al.* 2001). However, the physical nature of the lithosphere and the underlying asthenosphere remain controversial. This can be attributed to the different criteria used to define the lithosphere within the earth science community. Mechanical definitions of the lithosphere suggest that it is rigid and able to support loads such as sedimentary basins and mountain belts, whereas the underlying asthenosphere flows during flexing of the lithosphere (Muirhead & Drummond 1991; O'Reilly *et al.* 2001). The base of the lithosphere corresponds to the depth at which mantle rocks begin to deform in a viscous or plastic manner. This is related to temperature (commonly inferred to be ~1300°C: Turcotte & Schubert 1982) and to composition (O'Reilly & Griffin 1985; Green 1991; Rudnick & Taylor 1991; O'Reilly *et al.* 2001). The lithosphere consists of a crustal component and a mantle component. For continental lithosphere, the crustal component can vary in thickness from 10 km to over 70 km. The mantle component varies in thickness up to 400 km (Zhang & Tanimoto 1993). This mantle component is considered by geochemists to be a chemically depleted reservoir that is the residue of partial melting of the Earth's asthenosphere (O'Reilly *et al.* 2001).

Conventional geological techniques lend themselves to understanding the uppermost parts of the crust. Studies of the deep lithosphere can only be undertaken remotely using geophysical techniques and by the sampling of mantle xenoliths in igneous rocks (O'Reilly *et al.* 2001). Geophysical studies of the Australian lithosphere include seismic reflection and refraction at the crustal-scale (Goleby *et al.* 1989; Drummond *et al.* 1998; MacCready *et al.* 1998), as well as continental tomographic studies (SKIPPY: Simons *et al.* 1999; Zielhuis & van der Hilst 1996; Clitheroe *et al.* 2000; Debayle & Kennett 2000a, b). Mantle xenolith studies reveal that lithosphere is heterogeneous (O'Reilly & Griffin 1991; McDonough *et al.* 1991; O'Reilly *et al.* 2001) and is distinct from asthenospheric sources of mid-ocean ridge and ocean-island basalts.

Dating of the lithosphere beneath Archaean cratons suggests that it may be as old as the crust that it underlies (Pearson 1999). The relative buoyancy of sub-continental lithosphere renders it more difficult to destroy and recycle into the mantle than oceanic lithosphere during subduction (O'Reilly *et al.* 2001). Consequently, the continental lithosphere is often older and more complex than oceanic lithosphere, and is subject to reworking during repeated large-scale tectonic processes such as basin formation and the building and destruction of mountain belts.

Isotopic and geochemical studies suggest that the formation of the mantle lithosphere and the crust is a paired process (Pearson 1999). Pearson (1999) recognised the following mechanisms for the formation of the sub-crustal lithospheric mantle: (i) relatively rapid vertical accretion such that the age of the overlying crust is the same as the mantle lithosphere; (ii) gradual vertical accretion or cooling such that the lithosphere thickens and becomes younger with depth; (iii) accretion due to subduction that results in a complex vertical and horizontal structure; and (iv) continent–continent collisions (Figure 1). These mechanisms relate to tectonic processes such as plume activity, rifting and magmatic underplating, subduction, crustal accretion and continental collision, which leave a signature in the geological record. These same tectonic

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processes may also modify or destroy existing lithosphere. Our challenge is to infer the evolution of the Australian lithosphere based on our understanding of the crustal geological record and the limited data available from geophysical and geochemical studies.

In this paper, we use published tomographic data (Figure 2) (Simons *et al.* 1999; Chevrot & van der Hilst 2000; Debayle & Kennett 2000a, b), reflection and refraction data (Goleby *et al.* 1989; Drummond *et al.* 1998, 2000; MacCready *et al.* 1998) and isotopic data (O'Reilly & Griffin 1985; Handler *et al.* 1997; Graham *et al.* 1999) to illustrate the morphology, and to some extent the age, of the Australian continental lithosphere. The spatial resolution in these datasets is coarse. The tomographic models, in particular, vary depending on the type of data used, the size of the dataset and the processing techniques undertaken. The large-scale (500–1000 km) observations in many of the models are similar, although there are significant variations at the craton-scale (Simons *et al.* 1999; Debayle & Kennett 2000a). We will attempt to address the extent to which the evolution of the crust can be related to the present-day lithospheric architecture.

MORPHOLOGY OF THE AUSTRALIAN LITHOSPHERE

Tomographic studies have shown that the Australian lithosphere is complicated with highly variable thickness and seismic wave speeds (Figure 2a) (Simons *et al.* 1999; Chevrot & van der Hilst 2000; Clitheroe *et al.* 2000; Debayle & Kennett 2000a, b). There is no clear relationship between the thickness and composition of the lithosphere and the age of the overlying crust, as has been observed in global tomographic studies. The lithosphere beneath Archaean

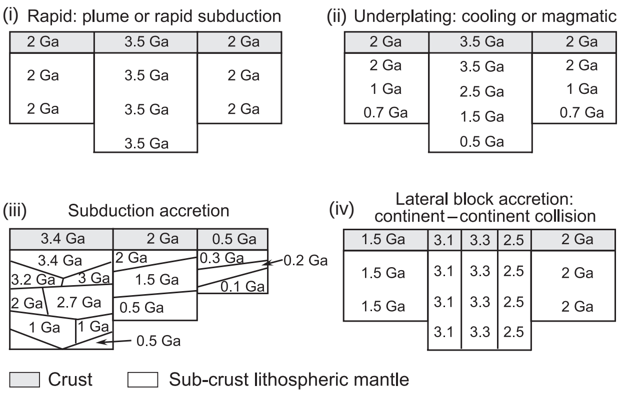


Figure 1 Possible mechanisms of continental lithospheric growth (after Pearson 1999). (i) Plume or rapid accretion where the crust and the lithosphere have the same age. (ii) Underplating: the lithosphere becomes progressively younger with depth. (iii) Subduction-related accretion in which there is a decrease in crustal and lithospheric age with depth and towards the edge of the continent. The decrease in the age of the lithosphere towards the continental margin results from accretion of new material onto the continental margin. The decrease in the age of the lithosphere with depth reflects vertical accretion of new material with time. (iv) Continent–continent collision or lateral block accretion. Lithospheric growth occurs by amalgamation of blocks of different ages.

and Proterozoic Australia shows heterogeneity at scales smaller than the component cratons (Simons *et al.* 1999; Debayle & Kennett 2000a, b).

Data for the lithosphere beneath the Archaean Yilgarn, Pilbara and Gawler Cratons is less reliable than the Proterozoic terranes of North Australia and the eastern Phanerozoic terranes (Simons *et al.* 1999; Debayle & Kennett 2000a, b). Tomographic models based on early SKIPPY data (Zielhuis & van der Hilst 1996; Simons *et al.* 1999) revealed heterogeneous wave speeds and significant thickness variations in the lithosphere beneath Archaean cratons. However, the most recent models have shown consistently high wave speeds to depths of ~150 km beneath the Yilgarn, Pilbara and Gawler Cratons (Debayle & Kennett 2000a) and a relatively high wave-speed 'keel' beneath the Yilgarn Craton that persists to depths of >350 km (Figure 2b, c). These may be the remnants of an Archaean lithospheric root beneath the Australian continent.

The crustal thickness of the Yilgarn and Pilbara Cratons (see Figure 3 for locations) is between 30 and 35 km thick (Figure 2a). The Yilgarn Craton is characterised by north-trending granite–greenstone belts bounded by major strike-slip faults and shear zones, the southwestern gneiss terranes and the gneissic Narryer terrane in the northwestern part of the craton (Figure 4a). Seismic-reflection data over the Eastern Goldfields Province (Figure 4b) suggests a three-layered crust (Drummond *et al.* 2000). Greenstone belts in the central and eastern parts of the section form the uppermost layer. The mid-crustal section (10–20 km) consists of an east-dipping, west-vergent duplex system, whereas the lower crust is characterised by shallowly dipping reflectors that are interpreted to represent ductile deformation (Drummond *et al.* 2000) (Figure 4b). Most of the structures developed in the greenstone belts sole into a detachment at ~10 km depth that is essentially undeformed (Archibald 1998). The high-grade metamorphic rocks in the southwestern gneiss terranes (Figure 4a) (Middleton *et al.* 1995) may represent the exposed remnants of the mid-crustal duplex system, or terranes accreted to the craton margin (Myers 1993).

The Pilbara Craton comprises a central granite–greenstone belt characterised by an outcrop pattern dominated by 50–100 km diameter domal granitoid complexes, separated by synformal greenstone belts (Hickman 1983; Oliver & Cawood 2001) (Figure 5a). The western parts of the Pilbara Craton are composed of polydeformed volcano-sedimentary successions, granite and gneiss (Barley *et al.* 1998). Palaeoproterozoic sedimentary successions of the Hamersley Basin (Martin *et al.* 1998; Powell *et al.* 1999) are exposed extensively throughout the southern Pilbara Craton (Figure 5a).

Regional potential-field interpretation (Wellman 2000) shows that the Archaean rocks of the Pilbara Craton are oval-shaped, extending beneath the North West Shelf to the north and beneath the Hamersley Basin to the south (Figure 5b). Drummond (1983) showed that the Pilbara Craton crust is relatively thin (~30 km), with a mid-crustal increase in the velocity of seismic waves at approximately 15 km, suggesting that the lower crust is more mafic than the upper crust (Figure 5c). Based on regional potential-field forward modelling, Wellman (2000) suggested that the base of the granitoid bodies and large greenstone belts also extends to this depth (Figure 5c). A ~4 km-thick transition seismic velocity layer between the crust and mantle is characterised by a sharp increase in seismic velocity at ~28 km (Drummond 1983) (Figure 5c).

Beneath the Proterozoic Australian cratons, relatively high seismic velocities exist to depths greater than 200 km, implying a relatively thick lithosphere (Figure 2b, c). The seismic velocities are generally slower than those beneath the Archaean crust, but extend to slightly greater depths and there is greater heterogeneity at shallower depths (<150 km) (Figure 2b). The Kimberley Craton (Figure 3) is a notable exception (Simons *et al.* 1999) (Figure 2b). Tomographic datasets do not appear to delineate interpreted tectonic boundaries recognised from the surface geology, although, long-wavelength gravity features (Archibald *et al.* 2000) often delineate such boundaries.

Proterozoic crust is inferred to have an average thickness of 45 km (Chevrot & van der Hilst 2000). The greatest

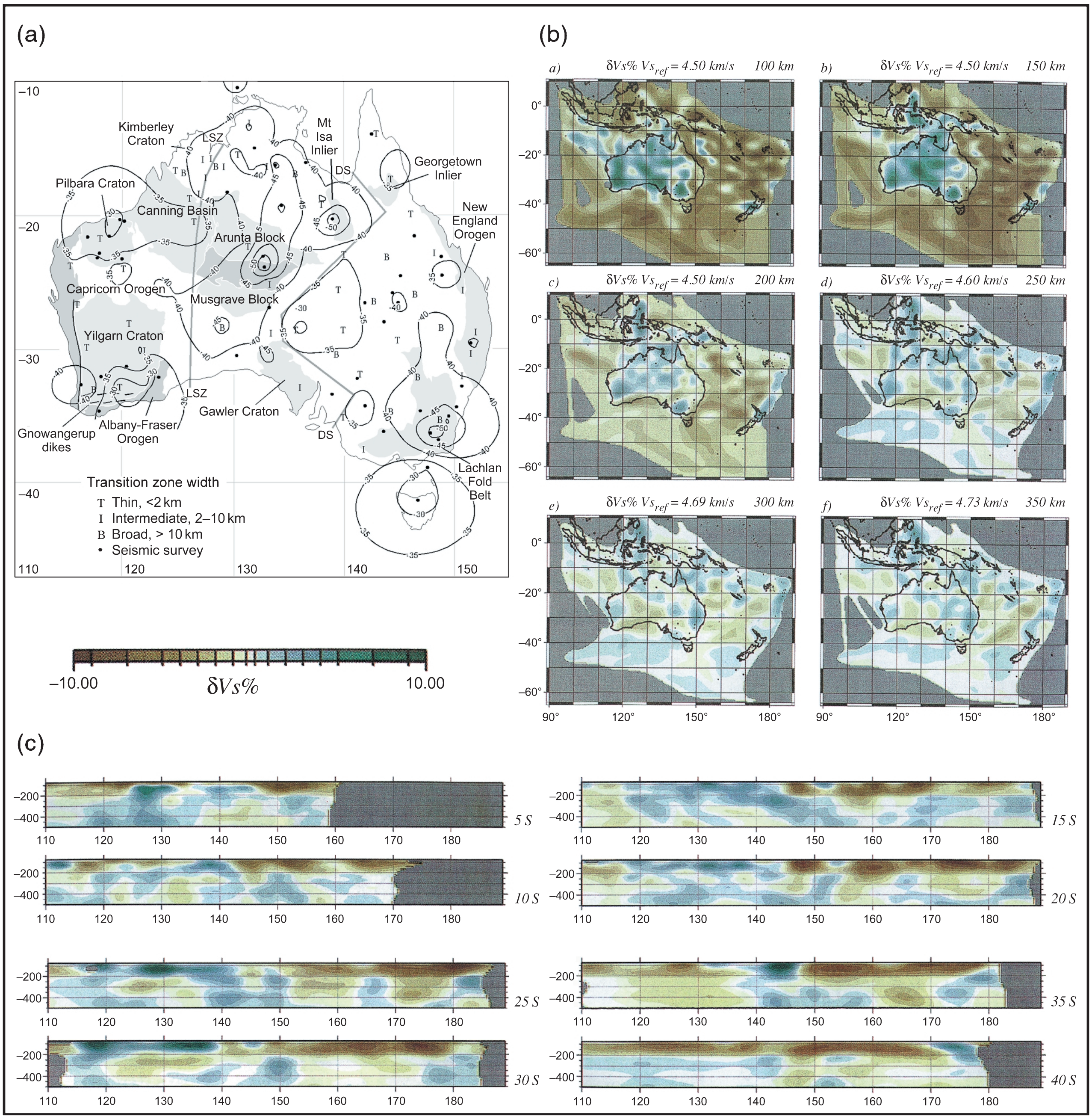


Figure 2 (a) Contour map showing the crustal thickness of the Australian continent (after Clitheroe *et al.* 2000). The light and dark shades of grey indicate the location of major crustal blocks in Australia. (b) Tomographic maps across Australia showing SV wave-speed models at different depths. The colours represent perturbations (in percent) from a reference velocity. Grey tones define areas of poor resolution (images adapted after Debayle & Kennett 2000a). (c) Tomographic sections across Australia showing SV wave-speed models at different latitudes. The colours represent perturbations (in percent) from a reference velocity. Grey tones define areas of poor resolution (images adapted after Debayle & Kennett 2000a).

depths to the Moho occur beneath the Mt Isa Inlier, McArthur Basin, Tennant Creek and the Arunta Inlier (see locations on Figure 3), where crust thickness is up to 55 km. Short-wavelength variations in crustal thickness, for example, that offset the Moho by ~10 km along the Redbank Zone (Goleby *et al.* 1989) are not imaged in the regional tomographic data. The thick Proterozoic crust of the North Australian Craton is often characterised by a zone of transitional seismic wave speeds that has been interpreted as mafic and ultramafic underplate (Drummond & Collins 1986). At Mt Isa, seismic refraction data indicate shallowly west-dipping, high-velocity layers at mid-crustal levels, which may represent underplated mafic rocks thrust into the mid-crust during crustal shortening (Goncharov *et al.* 1996; MacCready *et al.* 1998) (Figure 6).

In general, seismic velocities of the Phanerozoic lithosphere are slower than those beneath the Archaean and Proterozoic crust (Figure 2). The high-velocity lithosphere beneath Phanerozoic Australian terranes is heterogeneous and relatively thin, extending to a depth of approximately 80 km (Simons *et al.* 1999). This is underlain by a pronounced low-velocity zone that extends to approximately 200–220 km depth (Simons *et al.* 1999), but is centred at a depth of 150 km (Zielhuis & van der Hilst 1996).

Lower than average wave speeds, in regions such as central Queensland, are likely to reflect thermal processes related to the opening of the Tasman Sea and accompanying volcanism (Storey 1995), volatile infusion during pre-extensional subduction and elevated heat flow (Cull & Denham 1979) during recent volcanism (Johnson *et al.* 1989; Simons *et al.* 1999). Higher than average wave speeds occur to a depth of approximately 200 km beneath the Curnamona Craton and the Murray Basin (Zielhuis & van der Hilst 1996), to at least 400 km beneath the Delamerian Orogen, and in a discrete layer between 200 and 400 km beneath central and northern Queensland (Simons *et al.* 1999) (Figure 2b). The last of these may represent the underthrust remnant of a west- or south-dipping lithospheric slab related to either Palaeozoic or Mesozoic subduction. The high-velocity zones beneath the Curnamona Craton, Murray Basin and the Delamerian Orogen may indicate the presence of an inherited, seismically fast Proterozoic lithospheric root beneath the Delamerian Orogen and the western Lachlan Orogen (Simons *et al.* 1999). This is supported by Re–Os studies of peridotite xenoliths in Victoria, which show that the lithospheric mantle becomes older to the west, with mantle depletion ages ranging from less than 500 Ma to 1960 Ma (Handler *et al.* 1997).

The crustal depth across the eastern Australian Phanerozoic terranes is highly variable. For the most part the depth to the Moho is between 32 and 45 km (Finlayson 1993; Gray *et al.* 1998; Debayle & Kennett 2000a). However, beneath Bass Strait and Tasmania it may be as little as 26 km (Figure 2b) (Roach *et al.* 1993; Clitheroe *et al.* 2000). Anomalously thick crust (~50 km) occurs in the eastern Lachlan Orogen and is centred beneath the Eastern

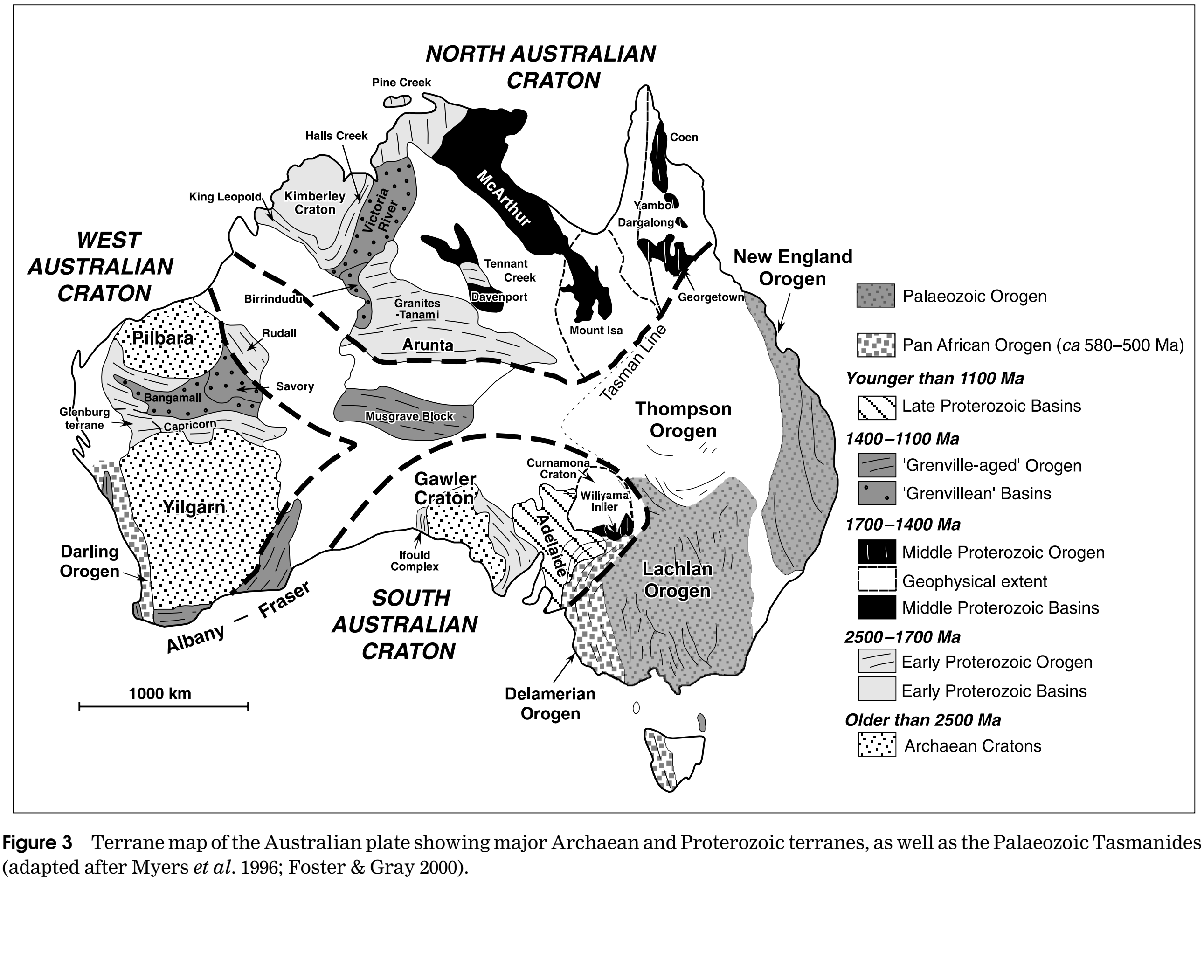


Figure 3 Terrane map of the Australian plate showing major Archaean and Proterozoic terranes, as well as the Palaeozoic Tasmanides (adapted after Myers *et al.* 1996; Foster & Gray 2000).

Australian Highlands (Figure 2a). In eastern Victoria, the boundary between the crust and the mantle is transitional between 45 and 52 km (Finlayson *et al.* 1980; Gray *et al.* 1998), whereas in the Otway Basin and in the northern Gippsland Basin a 2 km transition between the lower crust and the mantle occurs at a depth of 30 km (Gray *et al.* 1998). In the interior of Victoria the upper crust extends to approximately 15 km depth (Gray *et al.* 1998), whereas towards Bass Strait the upper crust may be as thin as 10 km (Collins *et al.* 1992). Limited seismic-reflection profiling in the southern Lachlan Orogen indicates a structurally complex middle to lower crust characterised by detachment faults or shear zones that intersect or link with the crust–mantle boundary (Gray *et al.* 1991). Seismic data from the northeastern Lachlan Orogen (Korsch *et al.* 2001) suggest that the crust is also complexly deformed and is characterised by east-directed thrusting, with west-dipping back thrusts.

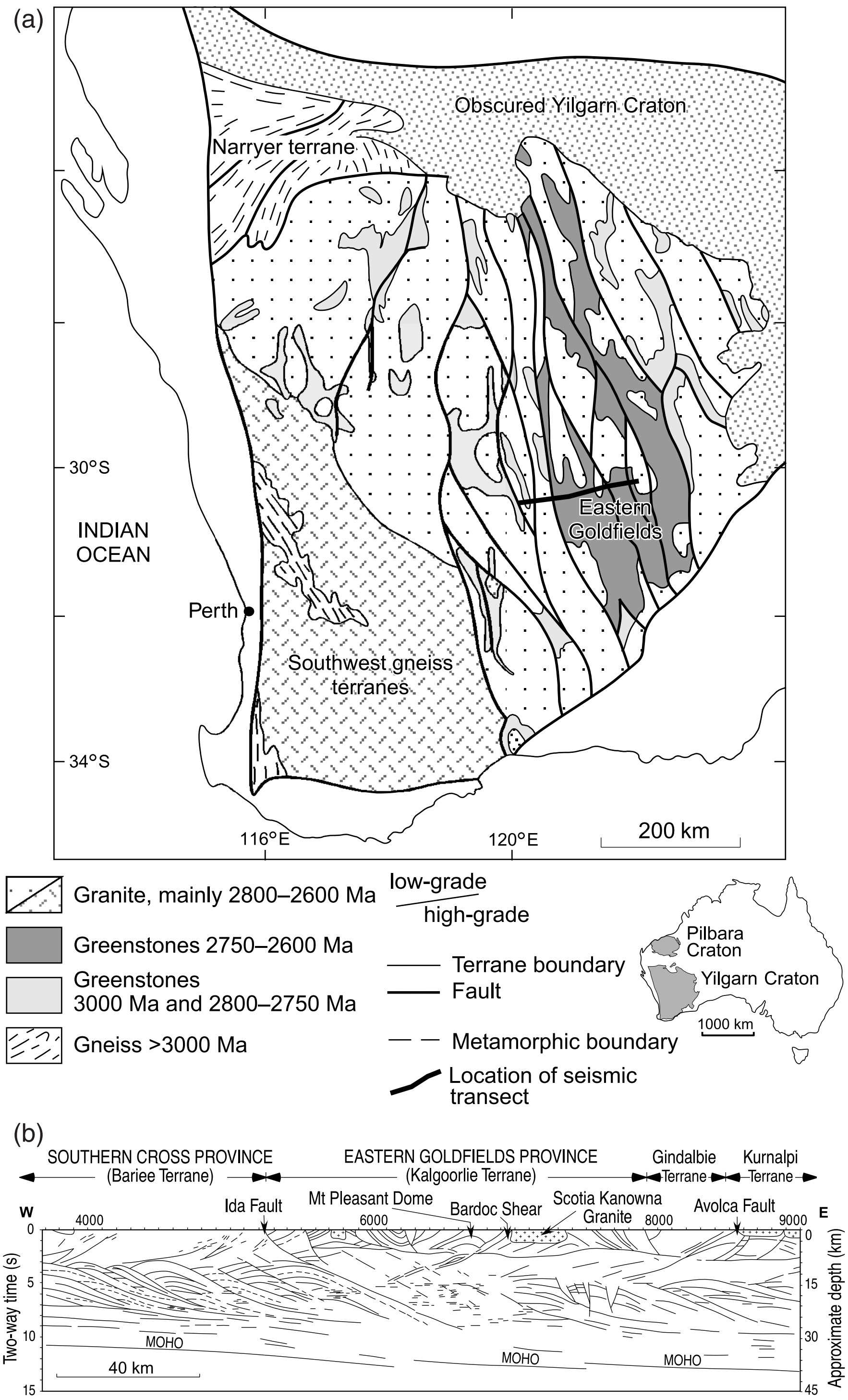


Figure 4 (a) Simplified geological terrane map of the Yilgarn Craton (after Myers 1993). (b) Seismic section interpretation of the crustal structure of the Eastern Goldfields in the central Yilgarn Craton (adapted after Drummond *et al.* 2000).

ARCHAEAN TECTONICS

There has been much speculation as to the specific nature of Archaean tectonic processes (Davies 1992, 1995, 1997; Condie 1997; Collins *et al.* 1998; de Wit 1998). What is the relative importance of processes similar to modern plate tectonics, and what is the importance of long-term secular changes, the most important of which is the decreasing radiogenic production of the Earth, which may have been four–sixfold that of today (McKenzie & Weiss 1975; Hargraves 1981)? This heat production must have been balanced by greater heat loss at the Earth's surface via more vigourous mantle convection, a greater length of mid-ocean ridges, faster spreading rates, greater magmatism at plate margins, greater heat transfer through continental crust or plume activity. The consequence is that Archaean tectonism is likely to have been more dynamic than the present Earth.

If plate tectonics, or some variant of plate tectonics, operated in the Archaean there may have been significant differences from the present day. Plates and the continental masses within plates may have been smaller as a consequence of the greater length of oceanic ridges (Davies 1997; Taylor & McLennan 1997). Assuming that the earth did not expand, longer ridges and faster spreading rates must have been balanced by a greater cumulative length of subduction/obduction zones or perhaps by more rapid subduction. Arc magmatism along these zones would have contributed to heat loss and to continental growth. Enhanced melting at mid-ocean ridges due to higher mantle temperatures may have resulted in thicker, more buoyant oceanic crust (Davies 1992, 1997). This thick crust

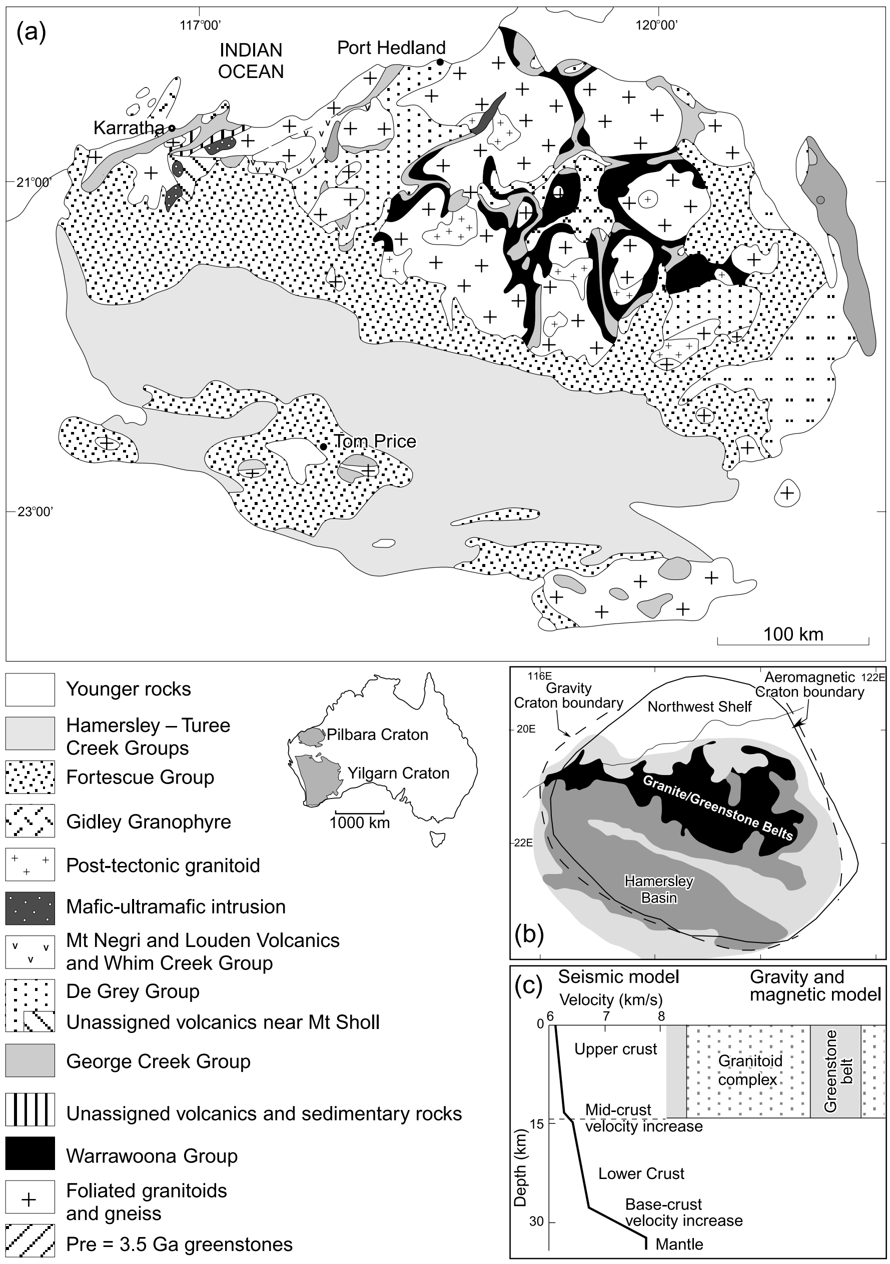


Figure 5 (a) Simplified geological terrane map of the Pilbara Craton (after Witt *et al.* 1998). (b) Inferred potential-field boundaries of the Pilbara Craton and surrounding area (adapted after Wellman 2000). (c) Average seismic velocity structure of the Pilbara Craton (after Wellman 2000; original data from Drummond 1983).

may have been resistant to subduction, and as a result of lateral accretion, obduction or underthrusting of oceanic crust may have been an important process of formation of continental lithosphere (Davies 1992).

Global tomographic data and studies of the age and thermobarometry of diamonds from beneath Archaean cratons suggest that thick lithospheric roots exist beneath Archaean continental crust (Dziewonski & Ekstrom 1998; Pearson *et al.* 1999; O'Reilly *et al.* 2001). These roots are long lived because of their buoyant and refractory nature (O'Reilly *et al.* 2001). Gravitation forces are insufficient to delaminate Archaean lithosphere, and the lithosphere is too depleted to be dispersed by melting (O'Reilly *et al.* 2001).

Evolution of the Pilbara Craton

The Pilbara Craton (Figure 5a) preserves a tectonic evolution between *ca* 3.65 and 2.0 Ga (Barley *et al.* 1998). Crustal growth occurred throughout the eastern part of the craton between 3.65 and 3.15 Ga. In the western parts of the craton, crustal growth is interpreted to have occurred in tectonic environments comparable to modern magmatic arc and backarc tectonic settings (Barley *et al.* 1998) and accretion of outboard island arcs and collisions of continental fragments (Smith *et al.* 1998). Volcano-sedimentary (greenstone) packages and associated igneous rocks were deposited or emplaced on a basement of sialic crust (Collins *et al.* 1998; Wellman 2000). Metamorphosed komatiite, basalt and dacite at the base of the succession have been dated at *ca* 3.51 Ga (Buick *et al.* 1995). These are overlain by tholeiitic basalt (up to 10 km thick), komatiites and calc-alkaline volcanics (*ca* 3.47–3.45 Ga) intercalated with chert horizons (Warrawoona Group) (Barley *et al.* 1998). These units were deposited in shallow- to deep-marine rift basins in a volcanic arc or near-arc tectonic setting (Barley *et al.* 1998). Extensive calc-alkaline batholiths were also emplaced in the eastern Pilbara Craton between 3.50 and 3.43 Ga, contemporaneous with greenstones (Williams & Collins 1990). This package of rocks is overlain by tholeiitic basalt, komatiite, calc-alkaline volcanics and clastic and chemical sediments (3.32–3.27 Ga) (Barley *et al.* 1998). These are in turn overlain by terrigenous and volcaniclastic sediments deposited in a continental backarc environment between *ca* 3.30 and 3.24 Ga (Barley *et al.* 1998).

Two competing models for the development of domal granite–greenstone belts in central and eastern Pilbara Craton have been proposed. Collins *et al.* (1998) and Collins and Van Kranendonk (1999) suggested that the development of 100 km-scale granitoid domes resulted from 'convective overturn' where a gravitationally unstable two-layered crust with thick and dense greenstone belts was underlain by hot, low-viscosity, sialic crust. Episodes of wholesale partial melting and thermal softening of this lower crust occurred and are interpreted to have resulted from unusually high levels of radiogenic heat production or by the arrival of a mantle plume at the base of the crust (Collins & Van Kranendonk 1999). Large domal structures began to develop as the dense greenstone belts sank into the thermally weakened sialic substrate and granite domes 'rose' into the upper crust (Figures 5a, 7). This process may have repeated several times with distinct granite emplacement events recorded at *ca* 3.45, 3.31, 3.24, 2.93 and 2.76 Ga (Collins *et al.* 1998). The result was an outcrop pattern dominated by 50–100 km-diameter domal granitoid complexes, separated by synformal greenstone belts (Hickman 1983; Oliver & Cawood 2001). 'Convective overturn' (Figure 7) appears to have been important for crustal differentiation, cratonisation and heat loss. However, its importance to lithospheric growth remains uncertain.

Zegers *et al.* (1996) proposed an alternative model to 'convective overturn' for the formation of large domal culminations of the Pilbara Craton. In their model, they recognised large shear zones and interpreted these to have evolved in an extensional environment (*ca* 3.45 Ga) analogous to a metamorphic core complex.

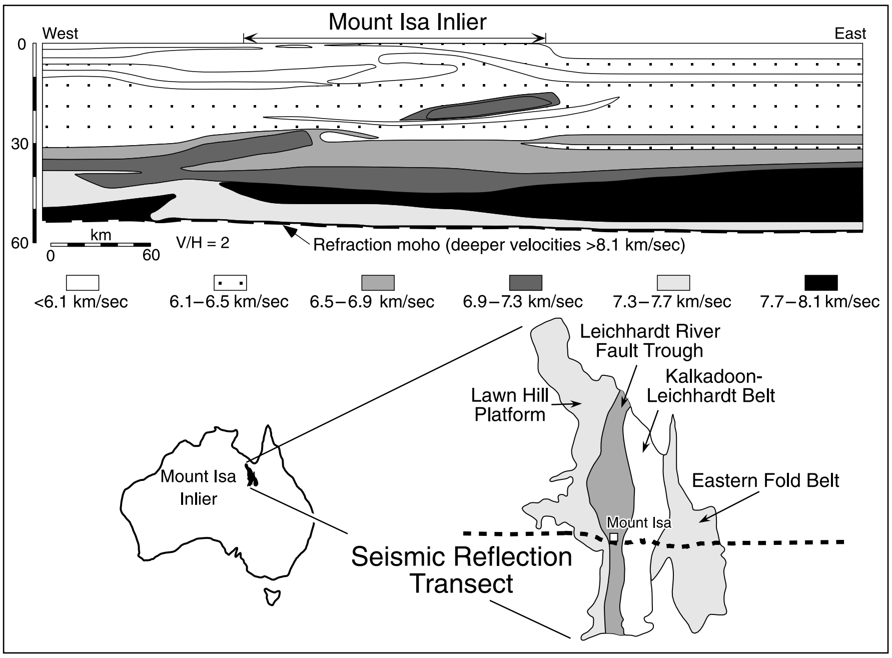


Figure 6 Seismic-refraction data across the Mt Isa Inlier collected along the seismic-reflection transect showing several high seismic velocity layers at mid-crustal levels. These layers have been interpreted as mafic underplate material that has been thrust into mid-crustal levels during the *ca* 1.6–1.5 Ga Isan Orogeny (adapted after MacCready *et al.* 1998; original data after Goncharov *et al.* 1996).

Contemporaneous deposition of the lower parts of the Warrawoona Group (Figure 5a) were interpreted to have been controlled by brittle extensional faults structurally above the shear zone (Zegers *et al.* 1996). This model is similar to modern terranes (Lister & Davis 1989) in which extensional tectonism involving regional detachments results in the uplift of gneissic rocks.

There are numerous geological features that are similar for both the 'convective overturn' and 'metamorphic core complex' models with regard to the formation of large diapiric domes in the eastern Pilbara (Collins *et al.* 1998). For example both models will contain: (i) synkinematic granitoids; (ii) marginal sheeted granitoids; (iii) discontinuous high-strain migmatitic carapaces; (iv) detachments; (v) normal arrangement of stratigraphy in the upper plate; and (vi) normal metamorphic zonation (Collins *et al.* 1998). Collins *et al.* (1998) argued that there are fundamental differences between the Pilbara granitoid domes and metamorphic core complexes. The domes are an order of magnitude larger than modern metamorphic core complexes; the steep-sided geometry of the domal granitoidal culminations (Wellman 2000) is inconsistent with low-attitude metamorphic core complexes, suggesting that detachments have been rotated into vertical, and in some cases their attitudes are overturned (Collin *et al.* 1998). Radial stretching lineations have been used to argue for the convective overturn (Van Kranendonk *et al.* 2001), although Van Haaften and White (2001) suggested that mineral lineations near the domes are unidirectional, which suggests horizontal tectonics.

It is possible that the domes began as metamorphic core complexes characterised by relatively shallow-dipping detachment faults and brittle normal faults in the upper plate. However, as the domes evolved detachment faults have been rotated into steep dips (Collins *et al.* 1998). It is likely that evidence of the early stages of 'convective overturn' may be indistinguishable from evidence for a metamorphic core complex.

The western parts of the northern Pilbara Craton (Figure 5a) contain elongate granitoids and greenstone belts (Figure 5a) that formed between *ca* 3.32 and 2.80 Ga (Barley *et al.* 1998). The evolution of these rocks has been interpreted in the context of an active plate margin with the episodic development of arc and backarc basins (Krapez & Eisenlohr 1998) and tectonic accretion at the western margin of the continent between *ca* 3.15 and 2.78 Ga (Smith *et al.* 1998). Komatiite, tholeiitic basalt, calc-alkaline volcanic rocks and clastic sediments were deposited and coeval granites were intruded (Barley *et al.* 1998; Krapez & Eisenlohr 1998; Smith *et al.* 1998) and subsequently deformed as the craton grew westward. During this time, anastomosing strike-slip faults developed between the domal culminations in the central Pilbara Craton (Van Kranendonk & Collins 1998). Deformation and craton growth is interpreted to have ceased by *ca* 2.86–2.83 Ga, when post-tectonic granites were emplaced (Bickle *et al.* 1989; Collins & Gray 1990).

HAMERSLEY BASIN

During the Late Archaean and Palaeoproterozoic the Hamersley Basin developed along the southern margin of the Pilbara Craton (Figure 5a) (Martin 1999; Oliver & Cawood 2001). Initial sedimentation occurred in a divergent margin setting into which the Fortescue Group (pre-2.6 Ga) was deposited (Martin *et al.* 1998). Banded iron-formation of the Hamersley Group was deposited between 2.60 and 2.45 Ga on a continental-margin platform (Barley *et al.* 1997; Martin *et al.* 1998; Powell *et al.* 1999). These are overlain by sedimentary rocks of the McGrath Trough, which marks a transition into a foreland basin environment associated with the northward advancing Ophthalmia Fold Belt at *ca* 2.2 Ga (Powell *et al.* 1999).

Evolution of the Yilgarn Craton

The Yilgarn Craton (Figure 4a) is composed mainly of low metamorphic grade granite–greenstone rocks that formed between *ca* 3.73 and 2.55 Ga. The oldest exposed rocks in the craton occur in the Narryer Terrane (*ca* 3.73 Ga: Myers 1993) (Figure 4a) in the northwestern Yilgarn Craton. Detrital zircons from the region yield an age of *ca* 4.3 Ga (Mojzsis *et al.* 2001), suggesting that continental crust of unknown size and character existed at this stage. The majority of the granite–greenstone belts comprise rocks that were deposited, deformed and metamorphosed between *ca* 2.8 and 2.65 Ga.

In the Eastern Goldfields (Figure 4), the lower part of the greenstone stratigraphy is dominated by high-MgO basalt and komatiites intercalated with felsic volcanic, volcaniclastic and clastic successions (*ca* 2.71 and 2.70 Ga: Nelson 1997). These are overlain by crustally contaminated komatiitic lavas (Paringa Basalt) and turbidite sequences (Kapai Slate) 2.692–2.687 Ga (Archibald 1998; Bateman *et al.* 2001). Other greenstone successions in the central Yilgarn have yielded a minimum age of 3.023 Ga (Nelson 1999; Chen

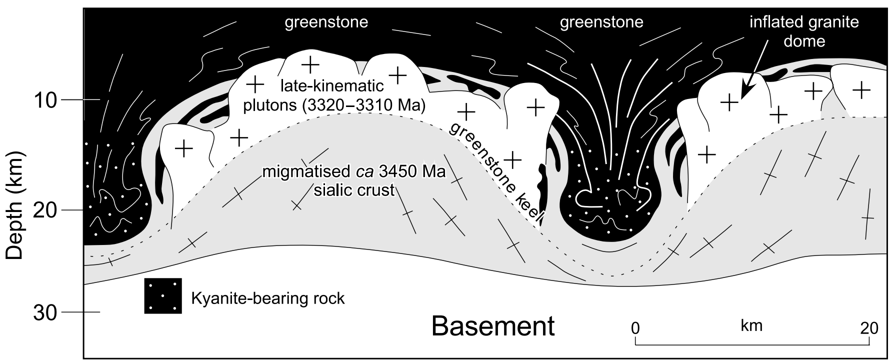


Figure 7 Schematic cross-section showing geometry produced by convective overturn of the crust as interpreted for the Pilbara Craton (adapted after Collins & Van Kranendonk 1999).

et al. 2001), implying that some greenstone belts could be considerably older. Between 2.680 and 2.675 Ga, sediments were deposited and tholeiitic intrusives were emplaced in an extensional environment. These are in turn overlain by flysch and molasse sediments (Archibald 1998).

The gneissic terranes in the western craton consist of deformed granites, gneiss, metasedimentary rocks, greenstone, and layered basic intrusives (Gee *et al.* 1986) that have been metamorphosed to upper amphibolite–granulite facies (Dentith *et al.* 2000). These rocks have been interpreted as a series of suspect terranes accreted to the Yilgarn Craton between *ca* 2.68 and 2.55 Ga (Myers 1993), and as mid crustal equivalents of granite–greenstone belts farther to the east.

Shortening in the Yilgarn Craton began with low-angle thrusts, isoclinal and sheath folds formed during north–south shortening (Chen *et al.* 2001). This evolved to a predominantly east–west shortening regime involving early thrusting, upright folding (Chen *et al.* 2001), regional greenschist to amphibolite facies metamorphism in the upper crust (Dalstra *et al.* 1999), and the formation of west-vergent duplexes in the middle crust (Drummond *et al.* 2000) (Figure 4b). Continued east–west shortening was accommodated by upright folding and strike-slip faulting (Chen *et al.* 2001) (*ca* 2.665–2.655 Ga).

Multiple generations of granitoids (2.71–2.60 Ga: Nelson 1997) have been identified with the most volumetrically significant being emplaced at *ca* 2.6 Ga (Archibald 1998; Bateman *et al.* 2001). These granites form elongate domes (Figure 4a) separated by synformal greenstone belts. Like the Pilbara Craton, the emplacement of these granites may be related to episodic 'convective overturn' of the crust. However, this process may have been relatively minor with thrust stacking and significant lateral translations, typical of modern tectonism, being the major influence on the crustal architecture (Drummond *et al.* 2000) (Figure 4b).

Evolution of the Gawler Craton

A large proportion of the central nuclei of the Gawler Craton (Figure 3) is considered to consist of Archaean gneiss with exposures located along the southern margin and northwestern parts of the craton (Daly *et al.* 1998). The tectonic setting and processes responsible for the formation of these rocks are poorly constrained. U–Pb zircon SHRIMP analysis of a metasedimentary gneiss in the southern part of the craton yielded a maximum depositional age of *ca* 2.64 Ga (Daly *et al.* 1998) and included inherited zircon, suggesting that the rocks were derived from continental crust as old as *ca* 3.15 Ga (Daly *et al.* 1998). The gneissic rocks were deformed and metamorphosed during the Sleafordian Orogeny (*ca* 2.55–2.40 Ga) and intruded by syntectonic granites aged between *ca* 2.55 and 2.46 Ga (Fanning 1997).

Growth of the Archaean lithosphere

Archaean crustal evolution is often viewed as a contest between vertical (plume tectonics) and horizontal (subduction, crustal accretion) processes. There is evidence that supports both tectonic situations in the Archaean geological record (Myers 1993; Archibald 1998; Barley *et al.* 1998; Collins *et al.* 1998; Collins & Van Kranendonk 1999; Bateman *et al.* 2001). Extensional tectonism has often been overlooked in understanding Archaean lithospheric growth. Extensional detachment faults have been described in the Yilgarn Craton (Hammond & Nesbit 1992; Williams & Whitaker 1993). In this context, extensional detachments may be responsible for the juxtaposition of 'greenstone lids' against underlying granite gneiss terranes. However, in a tectonic era of elevated mantle temperatures (Vlaar *et al.* 1994; de Smet *et al.* 1999) small amounts of lithospheric stretching will induce mantle melting, which may lead to voluminous crustal melting.

It has been speculated that the first continental lithospheric fragments formed by 'thrust-stacking' of buoyant, hydrated, thick oceanic lithosphere followed by internal differentiation (de Wit 1998; also see Trendall 1996). Trendall (1996) suggested that the earliest continental fragments formed when mantle convection slowed such that buoyant silicic crust could not be recycled into the mantle. Supracrustal magmatism and mafic underplating associated with arc and backarc environments may also have contributed to lithospheric growth. The arrival of mantle plumes at the base of the lithosphere has been attributed to the relatively high percentage of komatiite in the Archaean cratons (Hill *et al.* 1992; Bateman *et al.* 2001). Increased plume activity may have had an influence on lithospheric growth and the style of Archaean tectonics. Plume or mantle material that is depleted from melt extraction becomes refractory and buoyant (O'Reilly *et al.* 2001). Vertical accretion and the development of lithospheric keels (Pearson 1999) (Figure 1) results from mantle differentiation and cooling of refractory mantle. Plume-driven tectonism has been considered in the context of an intraplate environment (Bateman *et al.* 2001) and convective overturn of the crust (Collins *et al.* 1998), but little consideration has been given to the interaction between supra-subduction and a mantle plume [although see Abbott (1996) for one of many possible interactions between Archaean hot spots and subduction of oceanic material]. A recent analogy of this situation is the overriding of the North American plate above the Yellowstone hot spot during the Tertiary (Murphy *et al.* 1998). The Yellowstone hot spot interacted with the subducting Fallon and/or Kula slab, resulting in slab flattening and a magmatic hiatus (Murphy *et al.* 1998). Evidence of increased mantle-plume activity, as well as plate-tectonic processes, does not preclude one or the other. It is likely that the two processes were contemporaneous, with plume activity being an efficient way of cooling the interior of the Earth and plate tectonics being an efficient mechanism of heat loss at the surface.

PROTEROZOIC TECTONICS

Many Proterozoic Australian terranes lack blueschist or other high-pressure metamorphic rocks, accretionary complexes or mélange zones, belts of arc-type magmatic rocks and ophiolite complexes that are typical of Phanerozoic style subduction tectonics. Consequently, many Australian Proterozoic tectonic models emphasised the importance of vertical accretion (widespread mafic underplating) and reworking (partial melting of the lower crust) in an intracontinental setting (Etheridge *et al.* 1987; Loosveld & Etheridge 1990; Oliver *et al.* 1991). The driving mechanisms of these intraplate tectonic models include: (i) interaction between the continental lithosphere and mantle plumes (Oliver *et al.* 1991); (ii) delamination of the sub-continental lithospheric mantle (Loosveld & Etheridge 1990); (iii) vigourous, small-scale (<1000 km) mantle convection beneath a large, insulating continental mass (Etheridge *et al.* 1987); and (iv) deformation enhancement in regions of radiogenic heat production (McLaren *et al.* 1999). These mechanisms and processes are largely independent of plate-boundary forces or conditions.

Recently, plate-tectonic models of the Australian Proterozoic have become more popular (Zhao & McCulloch 1995; Myers *et al.* 1996; Giles & Betts 2000a, b; Scott *et al.* 2000) as increasing evidence provides support for subduction in the Proterozoic. Interpretations of convergent margin tectonism during the Proterozoic include the Kimberley Craton (Sheppard *et al.* 1999; Griffin *et al.* 2000; Tyler 2001), the Rudall Complex (Smithies & Bagas 1997), Capricorn Orogen (Occhipinti *et al.* 1999), Arunta Inlier (Zhao & McCulloch 1995; Giles *et al.* 2001), Kimban Orogen, Mt Isa Inlier, Curnamona Craton and northern Gawler Craton (Betts & Giles 2000), and the Musgrave and Albany–Fraser Orogens (Myers *et al.* 1996) (see locations on Figure 3). However, the increasing acceptance that plate-tectonic processes operated during the Proterozoic does not necessitate that these processes were exactly the same as on the modern Earth (McKenzie & Weiss 1975). For example, high-heat-producing granites in the continent interior may have had significant influence on the deformation and metamorphic history (McLaren *et al.* 1999). McLaren and Sandiford (2001) have suggested that the vertical migration of heat-producing elements in highly radiogenic granites may be integral to crustal strengthening and cratonisaton in the Proterozoic.

Continent–continent collisions and accretionary belts

The Palaeoproterozoic evolution of the Australian continent was dynamic and was dominated by the amalgamation of various Precambrian cratonic elements that comprise the continent. However, the initial amalgamation between the cratons began during the Palaeoproterozoic Glenburg and Capricorn Orogenies (Occhipinti *et al.* 1999; Sheppard *et al.* 2001). Amalgamation of the Yilgarn Craton and Glenburg terrane (Figure 3) (southern Gascoyne Province) occurred at *ca* 2.0–1.96 Ga, followed by docking of the Pilbara Craton and renewed deformation of the foreland basin succession of the Hamersley Province during the Capricorn Orogeny (*ca* 1.84–1.79 Ga: Tyler & Thorne 1990; Occhipinti *et al.* 1998).

ACCRETION OF THE KIMBERLEY CRATON

In the Late Archaean to Palaeoproterozoic Kimberley Craton (Figures 3, 8a) episodic basin development, arc accretion and ensialic volcanism occurred during west-dipping subduction (*ca* 1.91–1.85 Ga: Sheppard *et al.* 1999; Tyler *et al.* 1999; Griffin *et al.* 2000; Tyler 2001) (Figure 8b). Thereafter, turbidite sequences deposited along the southern and eastern continental margin of the Kimberley Craton were accreted onto the North Australian Craton followed by continent–continent collision of the Kimberley and North Australian Cratons during the *ca* 1.82 Ga Halls Creek Orogeny (Tyler & Griffin 1990; Tyler 2001) (Figure 8b).

BARRAMUNDI OROGENY

The Barramundi Orogeny (Etheridge *et al.* 1987) was an orogenic event that was preceded by an episode of rift and sag-phase sedimentation, followed shortly thereafter by crustal shortening, voluminous igneous activity (Wyborn *et al.* 1988) and high-temperature, low-pressure metamorphism (Etheridge *et al.* 1987) between *ca* 1.87 and

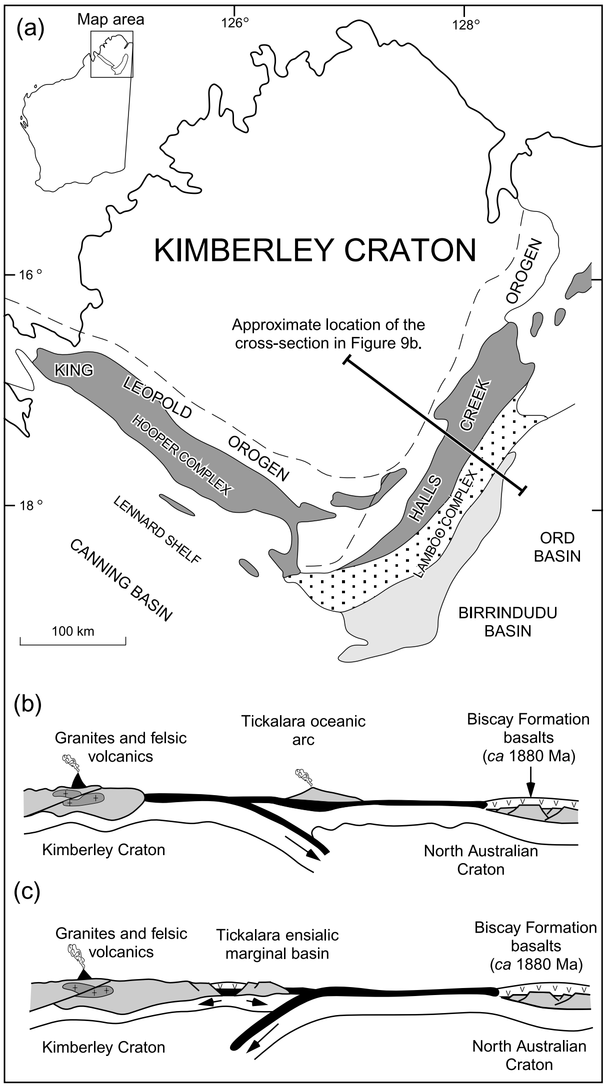


Figure 8 (a) Simplified geological map of the Kimberley Craton. (b) Lithospheric-scale cross-section showing the formation of the Tickalara Arc above a west-dipping subduction zone at *ca* 1.88 Ga (after Sheppard *et al.* 1999). (c) Accretion of the Tickalara Arc and the development of a west-dipping subduction that led to the amalgamation of the Kimberley Craton and the North Australian Craton during the Halls Creek Orogen (after Sheppard *et al.* 1999).

1.84 Ga. Evidence of the Barramundi Orogeny exists throughout many of the Proterozoic inliers of the North Australian Craton, such as the Halls Creek Orogen (Plumb 1990), Pine Creek Inlier (Needham & De Ross 1990), Tennant Creek Inlier (Le Messurier *et al.* 1990) and the Mt Isa Inlier (Etheridge *et al.* 1987) (Figure 3). Etheridge *et al.* (1987) interpreted the evolution of the Barramundi Orogeny in an intraplate context within a continent that had amalgamated by *ca* 2.3 Ga. However, McDonald *et al.* 1997) interpreted the *ca* 1.87–1.84 Ga Kalkadoon–Leichhardt Belt in the Mt Isa Inlier as the remnants of a magmatic arc. It is possible that the Barramundi Orogeny represents a continent-wide accretionary event in which several Archaean cratons amalgamated to form the North Australian Craton in a manner that is analogous to the Palaeoproterozoic amalgamation of Laurentia along the Trans-Hudson Orogen (Hoffman & Bowring 1984; Hoffman 1988).

ACCRETION ON THE SOUTHERN MARGIN OF THE CONTINENT

The tectonic evolution of Proterozoic Australia between 1.80 and 1.60 Ga was influenced by a long-lived accretionary margin and a postulated north-dipping subduction system along the southern margin of the North Australian Craton (Scott *et al.* 2000; Giles *et al.* in press). This margin

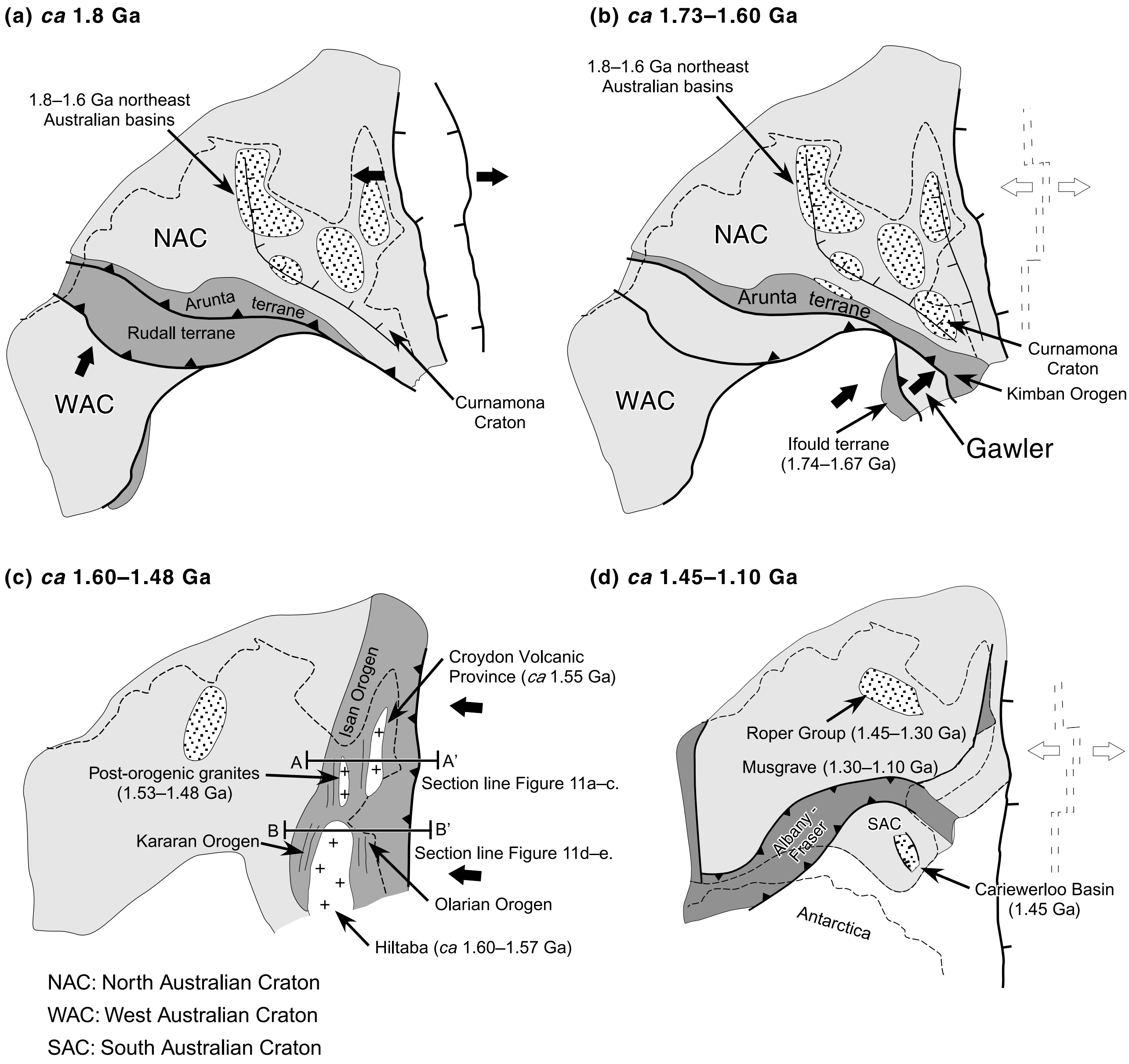


Figure 9 (a) Schematic diagram illustrating the amalgamation of the North Australian Craton and the West Australian Craton at *ca* 1.80 Ga. (b) Amalgamation of the Archaean nuclei of the Gawler Craton and the North Australian Craton at *ca* 1.73–1.60 Ga during the Kimban Orogeny. (c) Orogenic belt development along the eastern margin of the Australian continent between *ca* 1.60 and 1.50 Ga, and emplacement of late to post-orogenic granites *ca* 1.53–1.48 Ga. (d) Reconfiguration of the South Australian Craton during continental breakup (*ca* 1.45 Ga) and re-amalgamation with the North Australian Craton during the Grenville-aged orogenic events (*ca* 1.30–1.10 Ga).

preserves evidence of arc-related magmatism (Zhao & McCulloch 1995), moderate to high-pressure metamorphism (Norman & Clarke 1990; Smithies & Bagas 1997) and continent–continent collisional events (Li 2000). The discontinuous remnants of this accretionary margin are preserved in the Rudall Complex (Smithies & Bagas 1998), the Arunta Inlier (Collins & Shaw 1995) and along the Kimban Orogen on the eastern margin of the Gawler Craton (Parker & Lemon 1982; Vassallo & Wilson 2001). The Rudall Complex (Figure 3) records the collision between the West Australian and North Australian Cratons at *ca* 1.80–1.77 Ga (Smithies & Bagas 1997; Li 2000) (Figure 9a). In the northern and central Arunta Inlier (Figure 3), deformation and metamorphism associated with the Early Strangways (*ca* 1.78–1.77 Ga) and Late Strangways (*ca* 1.74–1.73 Ga) Orogenies (Collins & Shaw 1995) were interspersed by a period of Cordilleran arc-type magmatism (*ca* 1.77–1.75 Ga: Zhao & McCulloch 1995) (Figure 9).

At the same time as continued accretionary orogenesis was occurring during the Late Strangways Orogeny in central Australia, the Archaean nuclei of the Gawler Craton collided with the southeastern margin of the North Australian Craton along the Kimban Orogen (*ca* 1.74–1.73 Ga) farther to the southeast (Parker & Lemon 1982; Vassallo & Wilson 2001) (Figure 9b). This period of mountain building along the Kimban Orogen is coincident with a significant depositional hiatus and basin inversion (*ca* 1.73 Ga) in the Mt Isa Inlier and McArthur Basin (Bull & Rogers 1996; Betts 1999). This suggests that crustal shortening may have propagated up to 1500 km into the hinterland of the main orogenic belt (Giles *et al.* 2001; in press). Arc-related magmas were intruded along the western margin of the Gawler Craton (Ifould Complex) (Figure 3) between *ca* 1.74 and 1.67 Ga (Teasdale 1997; Daly *et al.* 1998), suggesting that a new subduction zone developed outboard of the Gawler Craton following its amalgamation with the North Australian Craton (Betts & Giles 2000). Renewed crustal shortening occurred in central Australia during the *ca* 1.60 Ga Chewings Orogeny (Collins & Shaw 1995; Collins *et al.* 1995). During this event polyphase deformation dominated the north-directed thrusts (Collins & Shaw 1995) and syntectonic granites were emplaced (Collins *et al.* 1995).

Continental backarc basins

Contemporaneous with subduction and continent collision (*ca* 1.8 and 1.6 Ga) along the southern margin of the North Australian Craton was a protracted period of extensional basin development in the continent interior (O'Dea *et al.* 1997a, b; Betts *et al.* 1998; Jackson *et al.* 2000; Page *et al.* 2000a; Southgate *et al.* 2000; Giles *et al.* 2001). The remnants of these Palaeoproterozoic basins are preserved in the McArthur Basin, Mt Isa Inlier, Georgetown Inlier and the Curnamona Craton (Figure 3). A characteristic of these intraplate basins is their high levels of heat flow expressed as magmatism (Connors & Page 1995; Pearson *et al.* 1992) and low-pressure, high-temperature metamorphism (Oliver *et al.* 1991) in an environment of intermittent crustal extension. The evolution of these basins can also be linked temporally with accretionary processes along the continent margin, suggesting a paired evolution of the terranes (Giles *et al.* 2001; in press). Page *et al.* (2000b) correlated Palaeoproterozoic sedimentary succession in the Curnamona Craton with the contemporaneous sedimentary rocks in the eastern and western fold belts of the Mt Isa Inlier, suggesting that they were once linked. Giles *et al.* (2001; in press) propose that the extensional stress regime in the overriding plate is strongly influenced by 'rollback' of the subducting slab (Figure 10).

Hiltaba event

At *ca* 1.60–1.58 Ga, the eastern margin of the Australian continent underwent a widespread tectono-thermal event characterised by voluminous outpouring of bimodal volcanic rocks (Gawler Range Volcanics) and emplacement of the Hiltaba Granites over the Gawler Craton and southern North Australian Craton (*ca* 1.58–1.56 Ga) (Creaser & White 1991; Creaser & Cooper 1993). During the same interval, high-temperature, low-pressure metamorphism culminated in the Curnamona Craton (Page *et al.* 2000b), the eastern Mt Isa Inlier (Page & Sun 1998; Giles 2000) and the Reynolds Range in the Arunta Inlier (Rubatto *et al.* 2001). This widespread thermal event has been attributed to the arrival of a mantle plume in an

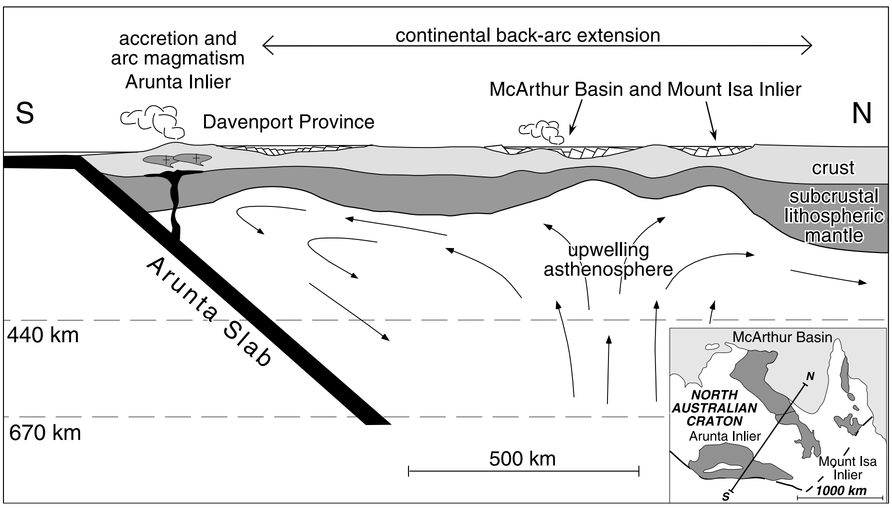


Figure 10 Lithospheric-scale cross-section showing the relationship between north-dipping subduction along the southern margin of the Australian craton and the development of extensional basins in a continental backarc setting (adapted after Giles *et al.* in press).

intra-continental setting (Flint *et al.* 1993), although magmatism and high-temperature metamorphism was also contemporaneous with orogeny along the eastern margin of the North Australian Craton.

1.60–1.50 Ga eastern Australian orogenic belts

Mesoproterozoic (*ca* 1.60–1.50 Ga) orogenesis along the eastern margin of the Australian Proterozoic continent interrupted basin development across the North Australian Craton (Figure 9c). Evidence of this *ca* 1.60–1.50 Ga orogenic event is preserved in the Mt Isa Inlier (Isan Orogeny), Georgetown Inlier, Curnamona Craton (Olarian Orogeny) and the northern Gawler Craton (Late Kararan Orogeny) (see location on Figure 3). These orogens are interpreted to have formed a once-continuous mountain belt as a consequence of west-dipping subduction and ultimately continental collision (Figure 9c). A common

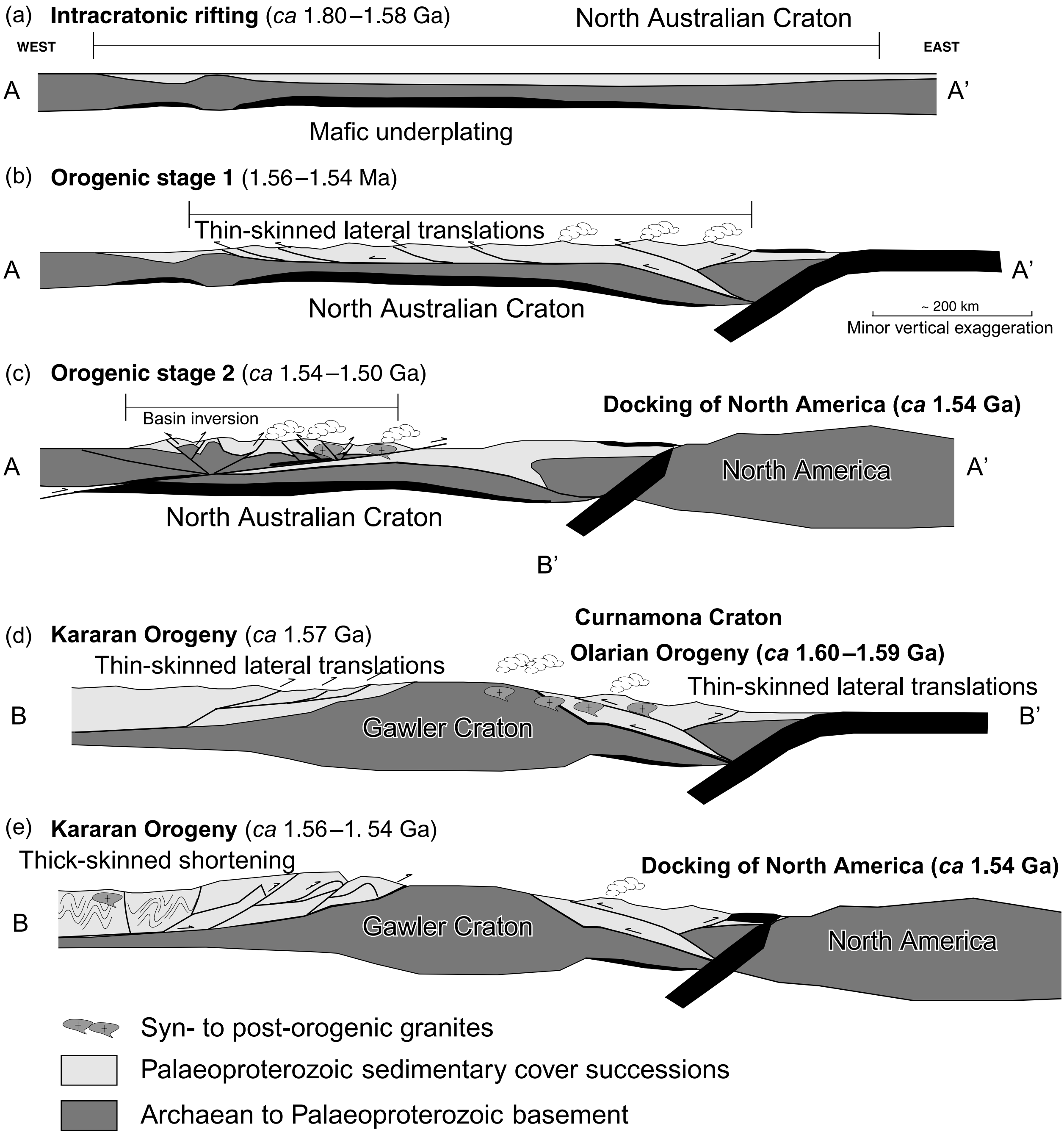


Figure 11 Crustal shortening associated with west-dipping subduction along the eastern margin of the Australian continent. (a) Pre-orogenic basin formation. (b) Thin-skinned deformation and lateral translations of the crust in the Georgetown and Mt Isa Inliers. (c) Thick-skinned crustal shortening and basin inversion in the Western Fold Belt of the Mt Isa Inlier. Thick-skinned tectonics associated with the collision between North America and Australia. (d) Thin-skinned deformation and lateral translations of the crust in the Broken Hill terrane, Gawler Craton and Kararan Orogen. (e) Thick-skinned crustal shortening in the Broken Hill terrane, Gawler Craton and Kararan Orogen. See Figure 9 for the location of cross-sections.

feature of the orogens is that they appear to have undergone a two-stage evolution (Figure 11) (Betts & Giles 2001). The first stage (*ca* 1.60–1.55 Ga) is characterised by high-strain, thin-skinned lateral translation (White *et al.* 1997; MacCready *et al.* 1998; Betts *et al.* 2000) (Figure 11b, d). The second stage (*ca* 1.55–1.50 Ga) involved shortening of deeper crustal levels and is often characterised by upright to inclined folding and brittle faulting (Figure 11c, e) (Hobbs *et al.* 1984; MacCready *et al.* 1998; Betts *et al.* 2000). Contemporaneous with crustal shortening in the Georgetown Inlier (Figure 3) were the extrusion of the predominantly rhyolitic Croydon Volcanic Group and the emplacement of the Forsyth Batholith and the Forest Home Supersuite, the latter of which yields arc-type magmatic signatures (Champion 1991; Zhao & McCulloch 1995) (Figure 9c).

Proterozoic plate reconfiguration

After the 1.60–1.50 Ga orogeny, the tectonic evolution of northern and eastern Australia was dominated by extensional basin development and the relatively slow exhumation of mid-crustal rocks in the 1.60–1.50 Ga orogenic belts (Foster *et al.* 1996). Intracontinental extensional basin development occurred in the Gawler Craton (Cariewerloo Basin at *ca* 1.45 Ga: Cowley 1993) and in the North Australian Craton (Roper Group of the McArthur Basin at *ca* 1.45 Ga: Jackson *et al.* 1987; Plumb 1993) (Figure 9d). The abundance of dolerite–gabbro sills in the stratigraphic pile of the Roper River Group possibly reflects injection of mantle melts during rifting.

Between *ca* 1.30 and 1.10 Ga, the South Australian Craton was reattached to the North Australian Craton and the West Australian Craton (Figure 9d) along a continuous Grenville-aged orogenic belt that is preserved in the Musgrave Block (Clarke *et al.* 1995) and the Albany–Fraser Belt (Myers 1993; Myers *et al.* 1996) (Figures 3, 9d). Granite and gabbro in the Albany–Fraser Orogen (Figure 1a) have been emplaced, intensely deformed and metamorphosed in the interval *ca* 1.30–1.28 Ga as Archaean and Palaeoproterozoic terranes of the South Australian Craton were shortened and stacked as thrust slices along the eastern margin of the West Australian Craton.

In the Musgrave Block, this event was associated with crustal thickening, high-grade metamorphism [500 MPa, >750°C (Clarke *et al.* 1995); *ca* 1.20–1.15 Ga (Maboko *et al.* 1991; Camacho & Fanning 1995)] and granite intrusion (*ca* 1.16–1.15 Ga: Clarke *et al.* 1995). Thereafter, mafic and ultramafic magmatism resulted in the emplacement of large layered intrusions of the Giles Complex (*ca* 1.08 Ga), comagmatic microgranite dykes, the Kulgera and Stuart Dyke swarms (*ca* 1.09–1.08 Ga: Camacho *et al.* 1991) and the

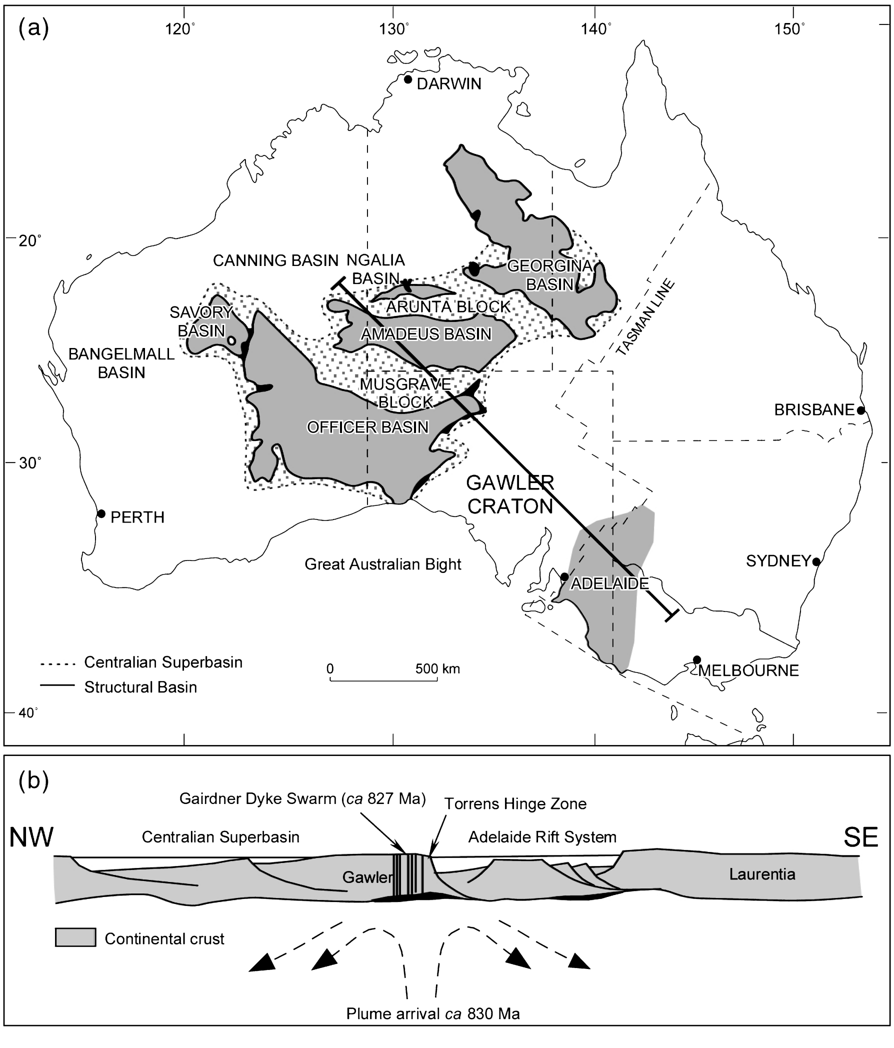


Figure 12 (a) Map showing the distribution of the basin elements of the Centralian Superbasin in the interior of the continent during the Neoproterozoic (after Walter *et al.* 1995). The Tasman Line is the inferred boundary between the Precambrian terrane to the west and Phanerozoic orogens to the east. (b) Cross-section showing the development of continental rifts along the eastern margin of the Australian continent during the Neoproterozoic. The approximate location of the cross-section is shown in (a).

bimodal Tollu Volcanics (Myers *et al.* 1996). The tectonic setting during this igneous event is conjectural, but it has been suggested that it occurred during a post-orogenic extensional event (Myers *et al.* 1996).

In the northern parts of the Gawler Craton this orogenic event is represented by the development of numerous strike-slip faults and shear zones, including the Karari Fault Zone. Metamorphic zircon within the Karari Shear Zone has yielded *ca* 1.17 Ga ages (Teasdale 1997). Geophysical data suggest that the Karari Fault zone has sinistral strike-slip offset, consistent with the oblique collision along the Albany–Fraser Orogen in which the South Australian Craton moved southwest with respect to the West Australian Craton (Myers *et al.* 1996).

Continental breakup

After the Grenville-aged orogenic event, a period of tectonic quiescence (*ca* 1.10–0.83 Ga) was followed by a protracted era of basin development associated with continental rifting (*ca* 830–600 Ma). Rift and sag basins developed along the eastern margin of the continent (e.g. Adelaide Rift Complex: Powell *et al.* 1994) (Figure 12a, b) and in the continent interior (Centralian Superbasin: Korsch & Lindsay 1989; Lindsay & Korsch 1991; Walter *et al.* 1995) (Figure 12a). Mafic dyke swarms were emplaced in the Gawler Craton (Gairdner Dyke Swarm at *ca* 827 Ma: Zhao *et al.* 1994; Wingate *et al.* 1998), the Musgrave Block (Amata Dyke Swarm: Walter & Veevers 1997) and the Broken Hill terrane (Wingate *et al.* 1998) (Figure 13). At the same time, basalts with similar geochemistry to these dykes were extruded in the Amadeus Basin (R. J. Korsch pers. comm. 2001). It has been inferred that these dykes and basalts record the arrival of a plume centred near Adelaide (Zhao & McCulloch 1993; Zhao *et al.* 1994; Walter & Veevers 1997). Anomalously deep basin depocentres occur proximal to the location of mafic dykes as well as recognised regions of growth faulting (e.g. southern Georgina Basin: Figure 12a) (Hand & Sandiford 1999).

Rifting along the eastern margin of the continent resulted in the development of a conjugate set of northeast-striking transfer faults and northwest-striking rift-bounding normal faults, which largely define the eastern margin of Proterozoic Australia—the Tasman Line (Powell *et al.* 1994) (Figure 12a). However, in the interior of the continent Lindsay and Korsch (1991) recognised north-northwest-trending transfer faults and inferred approximate east–west extension at this time. This rifting eventually led to the formation of a palaeo-Pacific ocean (Young 1995).

There are several interpretations for the exact timing of continental breakup between Australia and Laurentia. Subsidence curves from Laurentia, interpretation of the breakup unconformity in the Adelaidean sedimentary successions and palaeogeographical interpretations were used by Veevers *et al.* (1997) to infer a breakup age of *ca* 560 Ma. This interpretation conflicts with comparisons of palaeomagnetic data between Australia and Laurentia (Powell *et al.* 1993, 1994; Wingate & Giddings 2000). Powell *et al.* (1993, 1994) suggested that breakup between Laurentia and Australia occurred before *ca* 720–700 Ma, whereas palaeomagnetic data presented in Wingate and Gidding (2000) suggests that breakup occurred before *ca* 755 Ma.

Late Neoproterozoic crustal shortening

Transpressional deformation along reverse and strike-slip faults and partial melting of the crust and high-temperature medium-pressure metamorphism

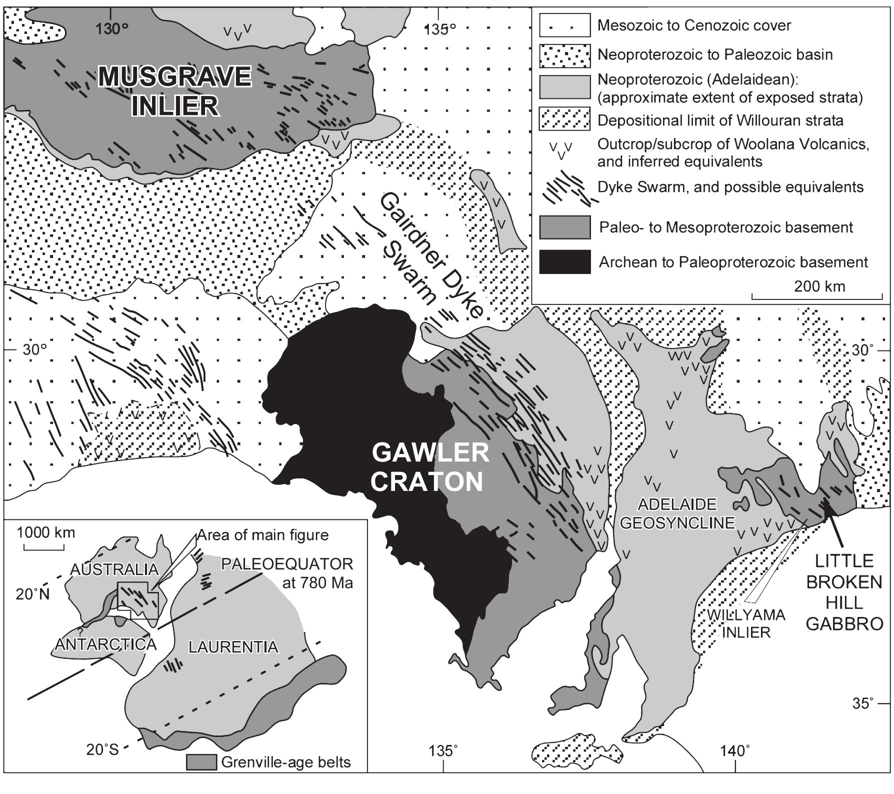


Figure 13 Map showing the distribution of *ca* 827 Ma mafic dyke swarms of the Gawler Craton, Musgrave Block and the Broken Hill terrane. Inset shows the global setting of Australia at the time of emplacement (adapted after Wingate *et al.* 1998 and Zhao *et al.* 1994).

(120–130 MPa and 700–750°C), resulted in localised inversion of the Centralian Superbasin during the Petermann Ranges Orogeny in the northern Musgrave Block (*ca* 580–540 Ma: Close *et al.* 1998; Hand & Sandiford 1999; Scrimgeour & Close 1999). Strain localisation and metamorphism during this event has been attributed to the presence of buried Mesoproterozoic high-heat-producing granites in the middle crust (Hand & Sandiford 1999; Scrimgeour & Close 1999). Contemporaneous with the Petermann Ranges Orogeny was an episode of sequential, transtensional magmatism and crustal shortening along the Darling Orogen to the west of the Yilgarn Craton (Figure 3) (Wilde & Murphy 1990; Harris 1994; Libby & de Laeter 1998).

GROWTH OF THE PROTEROZOIC LITHOSPHERE

The crustal evolution of the Australian continent during the Proterozoic involved several episodes of continent–continent and accretionary collision along several subduction fronts, interspersed by periods of extensional tectonism at the plate margins and in the continental interior. There is a strong spatial relationship between the location of major intracontinental basins and the location of thick crust. Deposition of thick sedimentary and volcanic piles (up to 15 km) (O'Dea *et al.* 1997a, b; Betts *et al.* 1998; Rawlings 1999; Jackson *et al.* 2000; Southgate *et al.* 2000) occurred during several episodes of lithospheric extension and subsequent thermal subsidence (Betts & Lister 2001). Crustal thickening may also have been achieved by vertical accretion associated with mafic underplating. This is supported by thick seismic wave-speed transition zones beneath several Proterozoic terranes (e.g. Mt Isa: Etheridge *et al.* 1987). The only part of the North Australian Craton that did not undergo significant crustal extension and basin development between *ca* 1.80 and 1.60 Ga was the Kimberley Craton. This part of the North Australian Craton is characterised by a relatively thin crust and lithosphere (Figure 2a), despite isotopic signatures from *ca* 1.2 Ga diamondiferous kimberlites of the Kimberley Craton that suggest derivation from

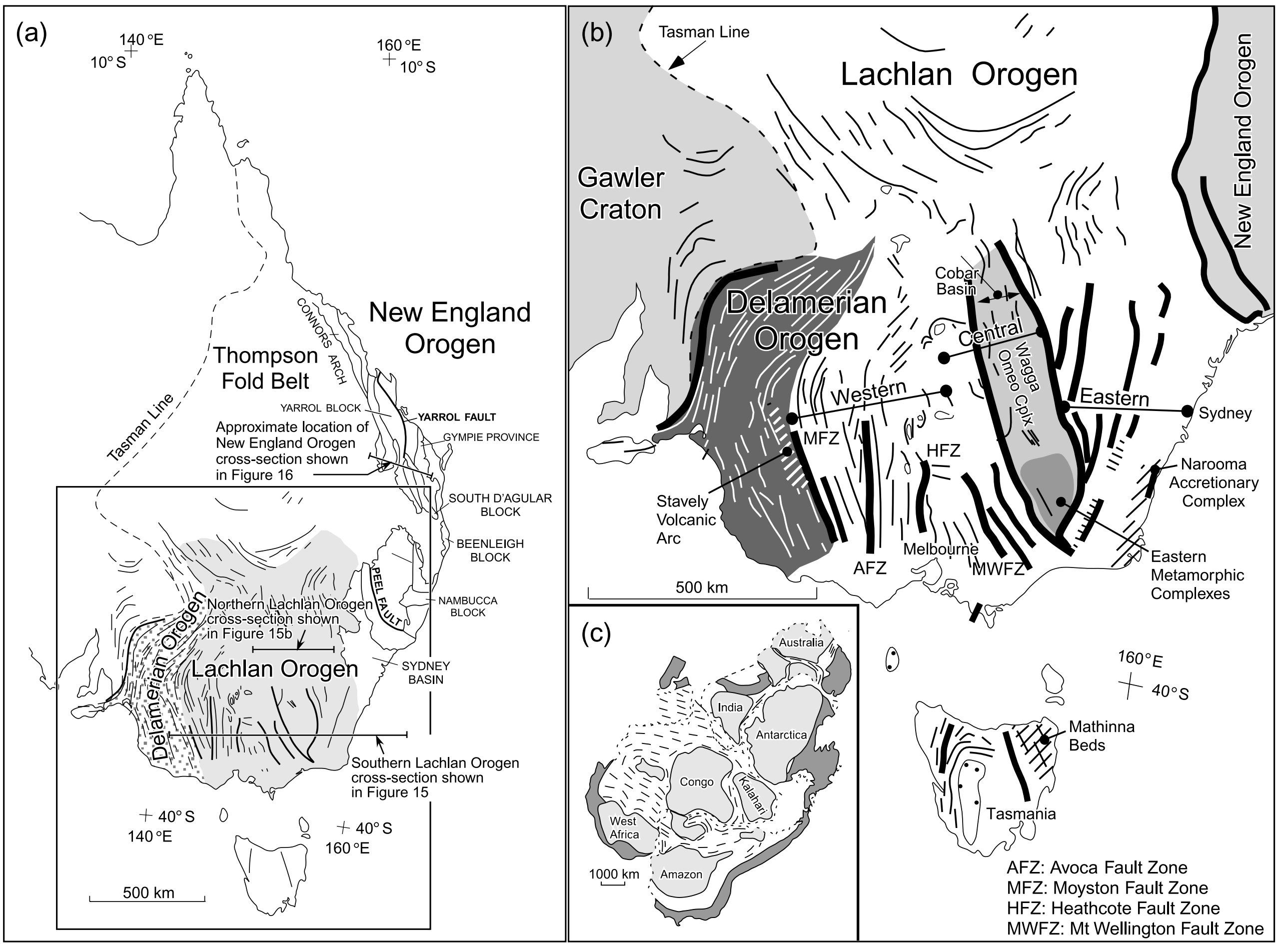


Figure 14 (a) Map of the Phanerozoic terranes of eastern Australia showing major elements of the crustal architecture. The major terranes are the Delamerian Orogen, the Lachlan Orogen and the New England Orogen. (b) Map of the Lachlan Orogen showing the major tectonic elements and fault zones. This map also shows the location of western, central and eastern subprovinces, which have been delineated by their differing deformation evolutions, structural styles and sedimentation histories. Major faults of the Lachlan Orogen are shown. These faults delineate the boundaries of major crustal blocks in the orogen (adapted after Foster & Gray 2000). (c) Simplified geological map of Gondwana showing the development of an accretionary margin that extended from Australia to South America (after Foster & Gray 2000).

an Archaean continental lithosphere that was metasomatised during the Palaeoproterozoic–Mesoproterozoic (Graham *et al.* 1999). If there was an Archaean lithospheric root it must have since been removed.

McCulloch (1987), Wyborn *et al.* (1988) and Wyborn (1998) interpreted granite ϵNd model ages of 2.30–2.10 Ga throughout much of northern Australia to reflect a widespread mafic underplate of this age. However, the Nd data can also be reconciled by a mixing of Archaean crustal material and juvenile mantle melts that would be consistent with the formation of mafic underplate during lithospheric extension at *ca* 1.80–1.60 Ga (Giles 2000). If this is correct, the crust and the lithosphere (?) on which Proterozoic basins and orogens developed may have been Archaean, and it is possible that the thick mantle lithosphere beneath much of Proterozoic Australia may have been inherited from the Archaean building blocks of the continent. Evidence of an Archaean lithospheric root beneath much of Proterozoic Australia is unequivocal and speculative. Re–Os isotopic data suggest that Archaean lithosphere once existed beneath the Kimberley Craton (Graham *et al.* 1999). U–Pb analysis on zircon from basement successions (Farmcote Gneiss) in the Curnamona Craton yield ages inherited between *ca* 2.67 and 2.55 Ga (Nutman & Ehlers 1998), suggesting a possible Archaean provenance. Similarly, McDonald *et al.* (1996) suggested the presence of Late Archaean crust in the Mt Isa Inlier from Nd isotopic studies and conventional U–Pb geochronology. Page *et al.* (1997) refuted this interpretation based on SHRIMP U–Pb analysis, which indicates that this crust is likely to be Early Palaeoproterozoic in age with a substantial component of Archaean crust.

However, protracted and episodic extension during the Palaeoproterozoic should have resulted in thinning of any inherited lithospheric root. Variation in seismic wave speeds in the North Australian Craton indicated by tomographic studies (Simons *et al.* 1999) may be related to heterogeneity in lithospheric architecture, which may reflect heterogeneities in lithospheric thinning during basin formation. For example, Betts *et al.* (1998) suggested that the development of the Calvert and Isa Superbasins involved asymmetric extension of the lithosphere in which crustal extension and lithospheric extension were offset. Renewed extension at *ca* 1.45 and 830 Ma might be expected to have resulted in local thinning of the lithosphere. The relatively thick lithosphere beneath the South and North Australian Cratons appears to have been little affected by post-1.50 Ga extension.

The thick lithosphere must have been re-established during subsequent thermal subsidence and/or orogeny. Thermal subsidence appears to have been important between *ca* 1.65–1.58 Ga in the north Australian basins (Betts *et al.* 1998; Betts & Lister 2001). Numerical modelling suggests that thermal subsidence will act to re-establish a lithospheric thickness similar to that prior to extension (Lister *et al.* 1991). This may be accomplished by a combination of vertical thermal accretion of new lithosphere (Pearson 1999) and lateral flow of reworked lithosphere.

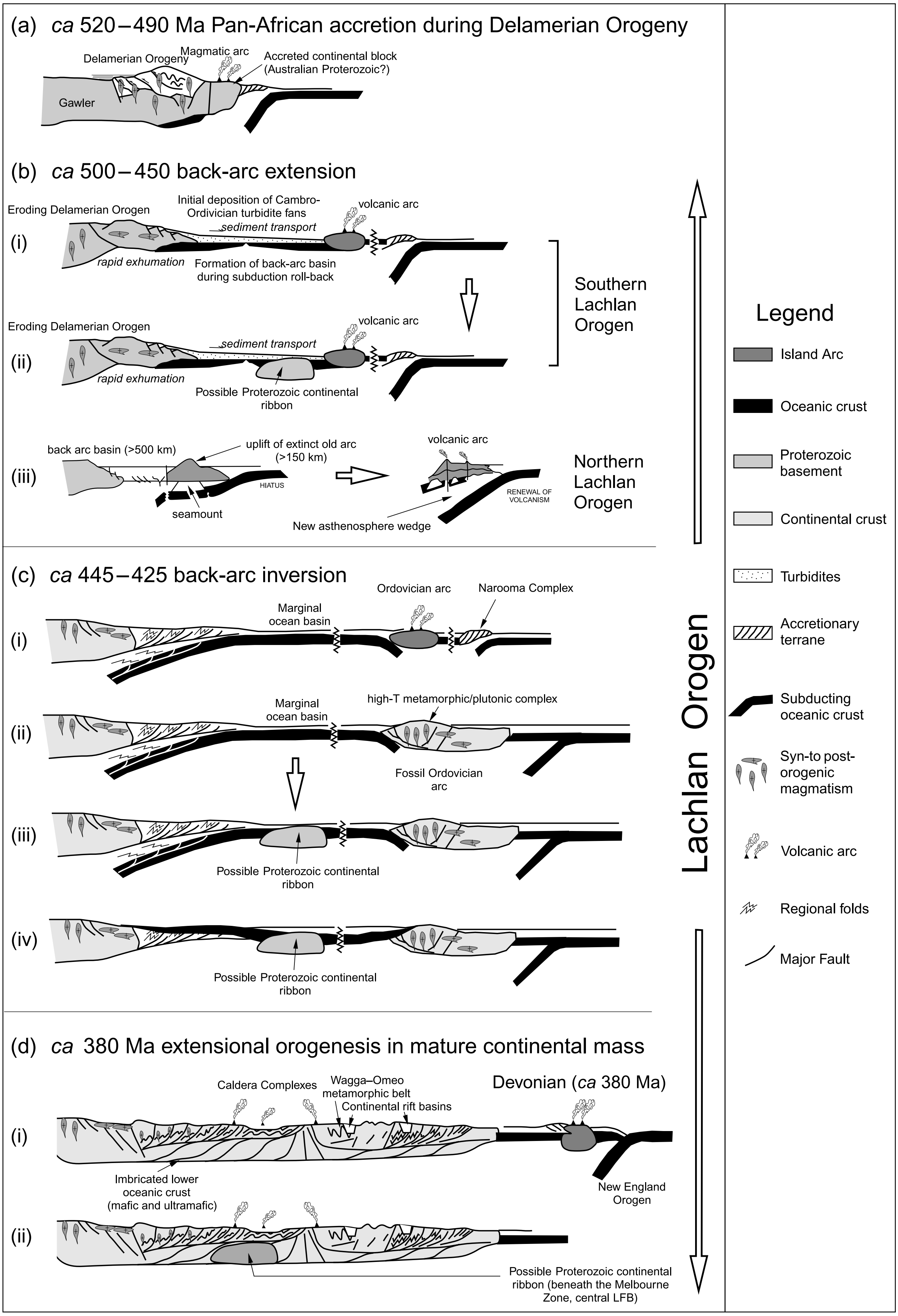
Another mechanism for lithospheric thickening may be post-extensional orogenesis. Along the Arunta–Kimban orogenic belts, lithospheric growth occurred by a combination of vertical accretion associated with partial melting of the asthenosphere in the mantle wedge and horizontal accretion of pre-existing lithospheric blocks (e.g. the Gawler Craton). In the orogenic belt that developed along the eastern margin of Proterozoic Australia (*ca* 1.60–1.50 Ga) crustal thicknesses may have been in excess of 60 km (Figure 2b) (Clitheroe *et al.* 2000). The Isan Orogen is characterised by several mid-crustal high seismic-wave velocity zones (Figure 6) (Goncharov *et al.* 1996; MacCready *et al.* 1998). These zones are interpreted as remnants of mafic underplate that were structurally emplaced during crustal shortening (MacCready *et al.* 1998). Crustal thickening must have been accompanied by significant lithospheric thickening (*ca* 1.65–1.60 Ga). However, such a mechanism may not have been particularly effective in the continent interior where the effects of orogenesis were relatively mild (Giles *et al.* 2001), although lateral flow of the lithosphere may have propagated the thickening toward the continent interior. Similar processes may have led to lithospheric thickening during the development of the Grenville-aged orogens (*ca* 1.30–1.10 Ga).

The chemistry of the voluminous late- to post-orogenic ultramafic and mafic sills of the Giles Complex (*ca* 1.2 Ga) in the Musgrave Block suggests that they were derived from a fractionating, continuously replenished, olivine-normative parent melt (Ballhaus & Berry 1991) consistent with a partially molten asthenosphere. This possibly implies synorogenic lithospheric thinning, which may indicate the arrival of a plume. If this is the case, the thinned lithosphere has not been preserved.

The Tasman Line (Figure 12a) has been interpreted to define the boundary between Precambrian cratons to the west and Phanerozoic terranes to the east (Powell *et al.* 1994). The dip of the Tasman Line is poorly constrained. This is critical for understanding how far Precambrian rocks extend beneath the Phanerozoic terranes of eastern Australia. For example, a shallow eastward dip may allow Precambrian rocks to extend a considerable distance beneath the western Lachlan Orogen (Figure 14). Tomographic studies (Zielhuis & van der Hilst 1996; Simons *et al.* 1999; Debayle & Kennett 2000a) (Figure 2b) and isotope studies of mantle xenoliths (McBride *et al.* 1996; Handler *et al.* 1997; Zhang & O'Reilly 1997) suggest that the Precambrian lithosphere extends well to the east of this line. The presence of Proterozoic lithosphere to the east of the Tasman Line may simply indicate that one or several Proterozoic continental ribbons were rifted from Australia during the Neoproterozoic and that these ribbons were subsequently reattached during Phanerozoic accretionary tectonism along eastern Australia.

PALAEOZOIC EVOLUTION OF THE CONTINENT

There has been lack of consensus regarding the evolution and plate tectonic setting of Phanerozoic terranes of eastern Australia (Figure 14) (Coney 1992). This has revolved around four key questions: (i) what is the nature of the lower crust and lithosphere (Proterozoic continental or Palaeozoic oceanic) that underlies the widespread Palaeozoic supracrustal successions; (ii) what is the relationship between granite chemistry, granite


(a) ca 520–490 Ma Pan-African accretion during Delamerian Orogeny
Delamerian Orogeny
Magmatic arc
Accreted continental block (Australian Proterozoic?)
Gawler
(b) ca 500–450 back-arc extension
Eroding Delamerian Orogen
Initial deposition of Cambro-Ordivician turbidite fans
sediment transport
volcanic arc
(i)
rapid exhumation
Formation of back-arc basin during subduction roll-back
Southern Lachlan Orogen
Eroding Delamerian Orogen
volcanic arc
sediment transport
(ii)
rapid exhumation
Possible Proterozoic continental ribbon
back arc basin (>500 km)
uplift of extinct old arc (>150 km)
volcanic arc
(iii)
HIATUS
RENEWAL OF VOLCANISM
seamount
New asthenosphere wedge
Northern Lachlan Orogen
(c) ca 445–425 back-arc inversion
Ordovician arc
Narooma Complex
Marginal ocean basin
(i)
Marginal ocean basin
high-T metamorphic/plutonic complex
(ii)
Fossil Ordovician arc
(iii)
Possible Proterozoic continental ribbon
(iv)
Possible Proterozoic continental ribbon
(d) ca 380 Ma extensional orogenesis in mature continental mass
Caldera Complexes
Wagga–Omeo metamorphic belt
Continental rift basins
Devonian (ca 380 Ma)
(i)
Imbricated lower oceanic crust (mafic and ultramafic)
New England Orogen
(ii)
Possible Proterozoic continental ribbon (beneath the Melbourne Zone, central LFB)
Lachlan Orogen
Legend
Island Arc
Oceanic crust
Proterozoic basement
Continental crust
Turbidites
Accretionary terrane
Subducting oceanic crust
Syn-to post-orogenic magmatism
Volcanic arc
Regional folds
Major Fault

source and tectonic setting; (iii) was there progressive deformation throughout the Palaeozoic (Gray & Foster 1997), or were there a series of discrete tectonic events (VandenBerg 1999); and (iv) what was the tectonic setting of these terranes?

Cambrian accretion

Along the eastern margin of the continent, the passive margin that formed during the 800–600 Ma continental rifting persisted until the Middle Cambrian when convergence along the proto-Pacific margin led to development of the Delamerian Orogen (*ca* 520–490 Ma: Figure 15a). During this event a west-vergent foreland-type fold and thrust belt overprinted the Neoproterozoic Adelaidean successions (Marshak & Flöttmann 1996) and underlying Palaeoproterozoic to Mesoproterozoic basement units of the Curnamona and Gawler Cratons. Amphibolite facies high-temperature, low-pressure metamorphism accompanied polydeformation and intrusion of syn- and post-tectonic granites (Sandiford *et al.* 1992). At *ca* 500 Ma emplacement of post-orogenic granites and gabbro occurred, and parts of the Delamerian Orogen were rapidly exhumed and uplifted (Turner *et al.* 1996) providing the source of clastic sediments of this age in the Lachlan Orogen (Cas 1983; Turner *et al.* 1996) (Figure 15b).

Figure 15 (a) Pan-African accretion and inversion of Neoproterozoic Adelaidean Rift Complex successions along the eastern edge of Gondwana during the Delamerian Orogeny (after Foster & Gray 2000). Uplift during this inversion provided the source for turbidite successions throughout the Lachlan Orogen. (b) (i) Model cross-section showing the development of a Cambrian–Ordovician oceanic backarc and the deposition of Ordovician turbidite successions in the western and central Lachlan Orogen (adapted after Foster & Gray 2000). Development of this oceanic backarc may have resulted from the rollback of a west-dipping subduction zone outboard of the Australian continent. (ii) Model cross-section showing the same tectonic setting as (i) except for the presence of a Proterozoic continental ribbon with the oceanic substrate. (iii) Ordovician tectonic environment in the northern Lachlan Orogen showing the deposition of turbidites in an oceanic backarc setting and the accretion of a buoyant seamount during subduction (adapted after Glen *et al.* 1998). (c) (i) Model cross-section showing backarc inversion during subduction via duplexing of oceanic crust and overthrusting of overlying turbidite successions in the western and central Lachlan Orogen. In the eastern Lachlan Orogen an Ordovician arc developed and the Narooma Accretion Complex formed in response to west-dipping subduction (adapted after Foster & Gray 2000). (ii) Model cross-section showing the evolution of the Lachlan Orogen with development of a metamorphic and plutonic complex on the Ordovician Arc (adapted after Foster & Gray 2000) during continued backarc inversion. (iii) Same model as (ii) except for the presence of Proterozoic continental ribbon in the oceanic substrate. (iv) Alternative model where oceanic material is obducted during backarc inversion. (d) (i) Model cross-sections showing the development of the Lachlan Orogen into a mature continental extensional orogen in the overriding plate of an outboard west-dipping subduction zone. Thickening of the continental crust was achieved by duplexing of continental and oceanic crust and chevron folding. Backarc extension has resulted in the development of extensional basins and these are intermittently inverted during later shortening events (e.g. Tabberabberan Orogeny: *ca* 380 Ma). (ii) Same model as (i) except for the presence of a continental ribbon incorporated into the Palaeozoic crust during backarc inversion. The approximate locations of the cross-sections are shown in Figure 14a. LFB, Lachlan Fold Belt.

Lachlan Orogen

During the Late Cambrian and Ordovician, thick accumulations of turbiditic successions were deposited in an oceanic backarc setting to the east of the eroding Delamerian Orogen (Gray & Foster 1997; Foster & Gray 2000; Collins in press). This sedimentation marks the beginning of the Lachlan and New England Orogens (Figures 14–16) (Coney 1992). The evolution of the Lachlan and New England Orogens (Figure 15) between the Cambrian and the Late Permian involved episodic backarc development, accretionary collisional events and backarc inversion along several subduction zones (Scheibner 1976; Cawood 1982; Cas 1983; Murray *et al.* 1987; Coney *et al.* 1990; Collins & Vernon 1992; Coney 1992; Little *et al.* 1992; Gray & Foster 1997; Collins 1998; Foster & Gray 2000; Bryan *et al.* 2001). The Lachlan and New England Orogens belong to an orogenic system that extended approximately 20 000 km along the eastern margin of Gondwana between the northern Andes and eastern Australia (Gray *et al.* 1998) (Figure 14c).

CAMBRIAN–ORDOVICIAN BACKARC BASINS

Remnants of Cambrian–Ordovician oceanic crust are preserved throughout the central and western Lachlan Orogen in fault slices of greenstones consisting of

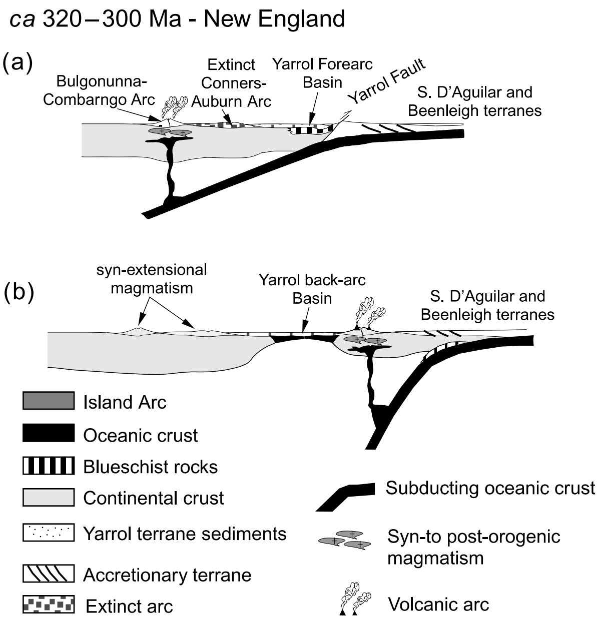


Figure 16 (a) Cross-section of arc magmatism and forearc basin development during the evolution of the New England Orogen (simplified after Little *et al.* 1992). (b) Alternative evolution of the New England Orogen where the Yarrol terrane formed as a backarc basin instead of a forearc basin (Bryan *et al.* 2001). The approximate locations of the cross-sections are shown in Figure 14a.

calc-alkaline andesite, boninite, subalkaline tholeiitic basalt, dolerite and gabbro with oceanic chemical affinities (Coney 1992; Fergusson 1998). Thrust-bounded Cambrian greenstone units of the western and central Lachlan Orogen are interpreted to represent remnant oceanic crust (Gray & Foster 1997; Foster & Gray 2000). These units formed the basement onto which a thick (>5 km) pile of Ordovician turbidites were deposited (Cas 1983; Crawford *et al.* 1984; Crawford *et al.* 1988; Foster & Gray 2000) (Figure 15b). The tectonic setting of this oceanic crust is ambiguous. The formation of this oceanic basin may have been in a backarc environment, which developed in response to rollback of a west-dipping subduction zone located outboard of the present exposures of the Lachlan Orogen (Collins & Vernon 1992; Collins 1998; Fergusson 1998; Foster & Gray 2000). The presence of boninite in the greenstones (Crawford *et al.* 1988) may indicate that the parts of the oceanic crust occurred in forearc or protoarc environment (Stern & Bloomer 1992) during their evolution. Emplacement of boninite into the greenstone pile may relate to later backarc inversion in the western Lachlan Orogen, and suggests the presence of multiple subduction zones across the region. Contemporaneous arc-related volcanic and sedimentary rocks (*ca* 500–480 Ma: Glen *et al.* 1998) structurally interfinger Ordovician turbidites in the eastern parts of the belt.

The presence of arc and oceanic basement in Ordovician turbidites is at odds with interpretations of crustal architecture based on the restite model for granite genesis (Chappell *et al.* 1987). According to this model the granites of the Lachlan Orogen were derived from partial melting of homogeneous source regions of either sedimentary (S-type), igneous (I-type) or refractory igneous (A-type) rocks in the basement (Chappell *et al.* 1987). This single source approach implies that there were crustal-scale tectonic boundaries (for example, the I–S line of the eastern Lachlan Orogen) between areas of differing granite chemistry. The corollary of the restite model is that the major element, trace-element and isotopic signature of the granites could be extrapolated to determine the nature and age of the source region—the much hypothesised basement to the Lachlan Orogen supracrustal successions. These data, combined with Proterozoic Nd model ages for the granites were used by Chappell *et al.* (1988) to argue for a Proterozoic source region beneath the Lachlan Orogen.

More recently Keay *et al.* (1997) and Collins (1998) have shown that a Proterozoic source region is not required to explain the ϵNd–Sr signature and U–Pb zircon populations of the Lachlan Orogen granites. Moreover, Sr isotope data indicate that the granites of the Lachlan Orogen are not sufficiently radiogenic to be derived from Precambrian crust exposed in the interior of the continent (Gray *et al.* 1998). Collins (1998) proposed a three-component mixing model involving Cambrian greenstones, Ordovician turbidites and an isotopically primitive end member that is a crust–mantle mix of tonalitic chemistry. Proterozoic Nd model ages and inherited zircon populations are consistent with provenance of the Ordovician sediments from an eroding orogen containing Proterozoic rocks (Collins 1998). This is not to say that there is no Proterozoic crust beneath the Lachlan Orogen, only that there is no clear evidence for it from granite geochemistry.

The presence of a relatively fast lithosphere (Simons *et al.* 1999; Debayle & Kennett 2000a) and Proterozoic Re–Os model ages from mantle xenoliths to the east of the Tasman line (McBride *et al.* 1996; Handler *et al.* 1997), suggest that Proterozoic lithosphere exists beneath the western Lachlan Orogen. One possibility is that there are a number of ribbons of continental crust, for example underlying the Melbourne Zone (Figure 15b(ii), d) of the central Lachlan Orogen, within a basement dominated by oceanic crust (Foster & Gray 2000).

BACKARC INVERSION

Inversion of the backarc basins of the western and central Lachlan Orogen may have been coincident with the formation of several small and relatively short-lived, opposite verging, subduction zones during the Benambran Orogeny (*ca* 445–440 Ma: Collins & Vernon 1992; Gray & Foster 1997; Foster & Gray 2000) (Figure 15c). Melange zones (Miller & Gray 1996, 1997; Spaggiari *et al.* 1998) and high-pressure metamorphic assemblages (Spaggiari *et al.* 1998) are interpreted to define the approximate location of the accretionary wedges formed during subduction and backarc-basin inversion (Foster & Gray 2000).

The timing of shortening and backarc inversion in the Lachlan Orogen has been constrained by argon thermochronology (Foster *et al.* 1998, 1999; Gray & Foster 1998). $^{40}Ar/^{39}Ar$ ages from white mica in fault zones vary between the Late Ordovician (*ca* 450 Ma) and Early Carboniferous (*ca* 340 Ma) across the belt (Gray & Foster 1998 figure 17). This pattern of ages was interpreted by Foster *et al.* (1998, 1999) and Gray and Foster (1998) to indicate progressive shortening, initiating in the western and central Lachlan Orogen and Narooma Accretionary Complex during the Ordovician. They inferred that deformation migrated to the east in the western Lachlan Orogen, to the southwest in the central Lachlan Orogen, and to the east in the eastern Lachlan Orogen (excluding the Narooma Accretionary Complex) during the Silurian and Devonian. In contrast, VandenBerg (1999) used the same data to argue for episodic deformation at *ca* 455, 440 and 425 Ma separated by periods of relative tectonic quiescence for the western Lachlan Orogen. At this stage, the evidence for progressive or episodic crustal shortening is equivocal: nevertheless, each model has important implications regarding the tectonic setting and processes that operated during the development of the orogen. Gray and Foster (1998) interpreted continuous sedimentation on a progressively shortening crust for the western and central parts of the Lachlan Orogen. However, for the eastern parts of the orogen, Collins (in press) interprets an 'extensional accretionary orogen' in which episodic shortening occurred in a regime dominated by lithospheric extension. It is worth noting that the tectonic setting proposed by Gray and Foster (1998), Foster and Gray (2000) and Collins (in press) in the backarc of a west-dipping subduction zone, is essentially the same in all models. Differences of opinion are only in the processes inferred to have operated in this tectonic setting. Modern tectonic settings demonstrate that the backarc environments can be exceedingly complex, with switches in tectonic mode routinely associated with individual accretion events for example (Rawling & Lister

1999). The difficulties are that a lack of outcrop and correlation make the task of deciphering the ancient record far more difficult.

In the eastern parts of the orogen, deformation focused in the Narooma Accretionary Complex (Figure 15c) and west-vergent thrust systems developed along with high-temperature plutonic complexes (Foster & Gray 2000), which may represent the remnants of arc and backarc magmatism (Collins in press) (Figure 15d). An estimated 50–70% shortening and thickening of the crust to ~30 km (Collins 1998) was accomplished by duplexing of the crustal pile and the development of chevron-style folds (Gray & Foster 1997) (Figure 15d). The lack of crust in excess of 30 km at the time of the orogen is consistent with the absence of high metamorphic grades across the belt (Collins 1998).

CONTINENTAL EXTENSION

It has been proposed that by the Early–Middle Silurian to Late Devonian – Early Carboniferous the Lachlan Orogen evolved into a mature continental orogen (Coney 1992) (Figure 15e) coincident with the evolution of a thin oceanic lithosphere (~30 km) to thicker 'continental' lithosphere (~80 km) (Collins in press). At this time, bimodal volcanism was concentrated in the eastern part of the orogen and localised marine basins (Cas 1983; Coney 1992) (e.g. the *ca* 390 Ma Tumut and Cowra Troughs) formed in a continental backarc setting (Collins 1998; in press) (Figure 15d). Crustal extension during this time was driven by either collapse of Early Silurian thickened crust (Foster & Gray 2000) or renewed rollback of a west-dipping subduction zone that was located outboard of the present exposure of the orogen (Collins in press). The continental lithosphere must have been relatively 'hot', with granites emplaced at this time comprising approximately one-third of the orogen. Early Devonian basins developed during this time were subsequently inverted in the Middle Devonian Tabberabberan Orogeny (*ca* 380 Ma: Foster & Gray 2000). During the Late Devonian – Early Carboniferous a series of northwest-trending basins (Mansfield, McAlister and Avon Basins) were developed above the Cambrian–Devonian successions of the Lachlan Orogen. These basins have been interpreted as regional dextral transtensional pullapart basins (Powell 1984) and as foreland basins (O'Halloran & Cas 1995).

New England Orogen

The New England Orogen (Figure 14a) (Cawood 1982; Murray *et al.* 1987; Coney *et al.* 1990; Little *et al.* 1992; Holcombe *et al.* 1997a, b) is characterised by numerous north-northwest-trending belts representing volcanic arc, forearc/backarc basins and subduction complexes (Figure 14b). Throughout the Palaeozoic it was a convergent plate margin at the edge of the Australian continent (Leitch & Cawood 1987; Murray *et al.* 1987), related to a west-dipping subduction zone (Figure 16a). It has been speculated that the New England Orogen may have begun to evolve as early as the Late Neoproterozoic (*ca* 570 Ma) during which time eclogite formed along an arc–trench system (Watanabe *et al.* 1996). The presence of Cambrian volcaniclastic sequences suggests that the orogen evolved contemporaneously with the Lachlan Orogen (Cawood 1976; Leitch & Cawood 1987). The presence of blueschist facies rocks in serpentinite melange along the Peel Fault, which yield K–Ar ages of *ca* 480–465 Ma, provide evidence of Early Palaeozoic subduction in the New England Orogen. Subsequent deposition of Silurian – Middle Devonian sedimentary and volcanic successions may be related to an island arc located to the east of the Australian continent (Murray *et al.* 1987). This island arc may have been separated from the Australian continent by a marginal sea (Marsden 1972) located in the approximate position of the Bowen Basin, or by a much wider oceanic basin that was subsequently consumed (Veevers *et al.* 1982). During the Early to Middle Devonian a forearc basin developed to the west of the Peel Fault (Murray *et al.* 1987) (Figure 14a). This forearc basin was located between a magmatic arc, which is now buried beneath the Sydney and Gunnedah Basins (see Korsch *et al.* 1997 for discussion), and an uplifted accretionary prism to the east of the Peel Fault (Cawood 1982). Regional folding during the Middle Devonian extended as far west as the Anakie Inlier and is possibly related to marginal sea closure at this time (Murray *et al.* 1987).

During the Late Devonian – Middle Carboniferous (*ca* 380–325 Ma) an Andean-type continental margin was established inland of the continental margin (Murray *et al.* 1987; Little *et al.* 1992), resulting in calc-alkaline magmatism dominated by andesitic flows and, to a lesser extent, bimodal volcanism (Connors–Auburn arc). The large distance between the inferred continental margin and the developed arc suggests relatively shallow subduction of the Pacific Plate (Little *et al.* 1992). Crustal thickening was accompanied by blueschist facies metamorphism (Little *et al.* 1992). Sedimentary basins developed unconformably over pre-existing arc terranes and have been interpreted as forearc volcaniclastic successions derived from contemporaneous arcs to the west (Yarrol Basin) (Murray *et al.* 1987) (Figure 16a). Holcombe *et al.* (1997a) proposed that Middle Carboniferous magmatic rocks are not arc-related, but instead developed in an extensional tectonic environment perhaps associated with the development of a backarc basin. This interpretation is supported by a recent study by Bryan *et al.* (2001), which suggests that the Yarrol Basin formed in a backarc environment during Upper Devonian to Lower Carboniferous extension (Figure 16b).

Murray *et al.* (1987) considered the New England Orogen to be a dextral transform margin during the Middle Carboniferous (*ca* 325–308 Ma), before reverting to a convergent margin at the Carboniferous–Permian boundary. Alternatively, Holcombe *et al.* (1994, 1997a) suggested that during the Late Carboniferous, syntectonic magmatism was accompanied by extensional tectonism that exhumed deeply subducted rocks, perhaps within metamorphic core complexes, juxtaposing them against higher levels of the accretionary complex. This extensional event marks the onset of the evolution of the Bowen Basin in a backarc setting (Korsch *et al.* 1992; Holcombe *et al.* 1994, 1997a; Fielding *et al.* 1997).

During the Early Permian, a new island arc began to evolve (Gympie arc) to the east of the Devonian–

Carboniferous arc (Sivell & McCulloch 1997). A backarc basin may have evolved between this arc and continental eastern Australia (Sivell & McCulloch 1997) as a consequence of rollback of an eastward-retreating hinge (Sivell & McCulloch 2001). Marine sedimentation ensued within fault-bounded extensional basins superimposed on the underlying accretionary complex (Holcombe *et al.* 1994, 1997a). Greenschist-facies metamorphism and intrusion of S-type granites took place at this time (*ca* 306–300 Ma: Little *et al.* 1992). This metamorphic event is interpreted to reflect an upwelling mantle in response to detached oceanic lithosphere or slab rollback (Little *et al.* 1992).

Arc-related magmatism in the southern Sydney Basin is interpreted to have been related to west-dipping Andean-style subduction that operated between 270 and 225 Ma (Cawood 1984; Fergusson & Leitch 1993). Deformation during the Late Permian Hunter–Bowen Orogeny (*ca* 265–225 Ma) is characterised by the sequential development of folding, southwest-directed thrusting, back thrusting and reactivation of the ancestral Peel Fault (Collins 1991; Holcombe *et al.* 1997b). The Hunter–Bowen Orogeny is responsible for the dominant north-northwest structural grain of the New England Orogen (Holcombe *et al.* 1997b). In southern Queensland, there appears to have been two episodes of crustal shortening with an intervening episode of calk-alkaline magmatism (Holcombe *et al.* 1997b). During this event continental detritus flooded into the Sydney Basin, which occupied a foreland setting (Collins 1991), while the Bowen Basin evolved into a foreland basin (Korsch *et al.* 1992; Holcombe *et al.* 1994, 1997b). This was followed by extensional tectonism and bimodal volcanism in the Late Triassic (Holcombe *et al.* 1994). The final stages of the evolution of the New England Orogen are recorded in the Surat and Bowen Basins, where a major basin-inversion event at *ca* 245–215 Ma resulted in the uplift and erosion of up to 4 km of section (Korsch *et al.* 1998b). It has been speculated that this uplift resulted from the docking of the Gympie terrane to the New England Orogen (Harrington & Korsch 1985), suggesting that subduction continued outboard of the New England Orogen until the mid-Cretaceous (Korsch *et al.* 1998b). However, Holcombe *et al.* (1997b) speculated that the Gympie terrane represents a northerly terrane of the northern New England Orogen that has been displaced southwards by dextral strike-slip movement during the waning stages of the Hunter–Bowen Orogeny.

The Alice Springs connection

Contemporaneous with the development of the Lachlan and New England Orogens was a long-lived period of deformation in central Australia. A metamorphic event has been recognised in the Arunta Inlier, central Australia (*ca* 475 Ma: Hand *et al.* 1999), during which time peak metamorphic conditions reached 800°C and 1005 MPa (Mawby *et al.* 1999). These rocks were partly exhumed during north–south to northeast–southwest directed extension in the Ordovician (Mawby *et al.* 1999). Sedimentary rocks of the overlying Amadeus and Georgina Basins formed part of a larger extensional system that linked with the Canning Basin to the northwest (Romine *et al.* 1994). The heat source for metamorphism is interpreted to be crustal and related to highly radiogenic plutons in the Palaeoproterozoic basement (Sandiford & Hand 1998). However, given the extensional environment, elevated geothermal gradients caused by attenuation of the lithosphere should also be considered as a potential cause of metamorphism.

The Devonian–Carboniferous (*ca* 450–300 Ma) Alice Springs Orogeny (Collins & Teyssier 1989) was a significant crustal-shortening event that exhumed the Arunta Inlier (Ballévre *et al.* 2000) during south-directed thrusting (Figure 17a, b). Inversion of the Amadeus Basin (Flöttmann & Hand 1999) led to foreland basin development along the northern Amadeus and Ngalia Basins (Sandiford & Hand 1998) (Figure 17a). Crustal shortening may have begun as early as *ca* 450 Ma (Mawby *et al.* 1999), although peak metamorphic conditions (600°C and 500 MPa) were not attained until *ca* 380 Ma (Ballévre *et al.* 2000). Basement-controlled thick-skinned deformation resulted in up to 20 km exhumation of the Arunta Inlier along deep crustal-penetrating shears zones (Sandiford & Hand 1998) (Figure 17), some of which offset the Moho by ~10 km (e.g. Redbank Thrust Zone: Goleby *et al.* 1989). Deformation in the surrounding Amadeus Basin was thin-skinned, producing foreland fold-thrust structures (Shaw *et al.* 1991; Korsch *et al.* 1998a; Sandiford & Hand 1998).

Attempts to reconcile the Palaeozoic evolution of central Australia with the evolution of the Lachlan and New England Orogens have been hampered by two factors. First, the inferred shortening directions are orthogonal: south-directed thrusting during the Alice Springs Orogeny

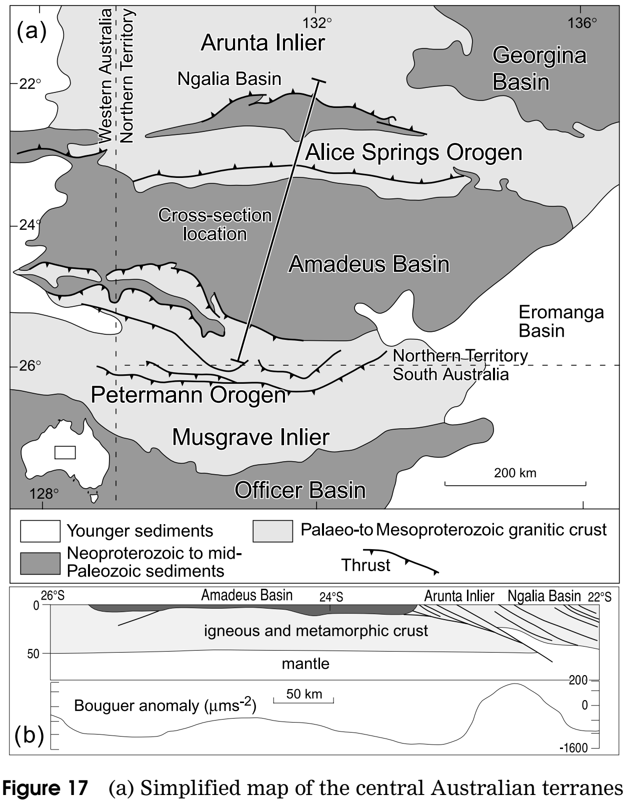


Figure 17 (a) Simplified map of the central Australian terranes and basins (adapted after Sandiford & Hand 1998). (b) Cross-section of the Alice Springs Orogen (adapted after Korsch *et al.* 1998a).

compared with east–west shortening in the Lachlan and New England Orogens. East–west shortening along the Lachlan and New England Orogens at this time was largely influenced by plate-margin processes comparable to the present southwest Pacific-type setting (Collins & Vernon 1992). The Palaeozoic evolution of the northern margin of the Australia is not well understood. It has been interpreted as a convergent margin, linking with subduction zones along the eastern Gondwana margin. The northern margin of the continent interacted with numerous microplates including north and south China (Pan 1994). This setting is remarkably similar to the present-day setting of the Australian plate in which active tectonism is occurring along the northern and eastern margins of the plate. Therefore, it is hardly surprising that present Australian stress field (Hillis *et al.* 1999) resembles that inferred for the Palaeozoic.

Second, the geochronological resolution allows for only broad temporal correlations between the regions. Nevertheless, it is tempting to make correlations between several tectonic events. For example, Ordovician extension in the Lachlan Orogen (*ca* 475–450 Ma) and blueschist-facies metamorphism in the New England Orogen (Fukui *et al.* 1995) correlate with extensional tectonism in central Australia (Mawby *et al.* 1999). Subsequent crustal shortening at *ca* 450–300 Ma in central Australia (Sandiford & Hand 1998; Mawby *et al.* 1999) occurred at a time when the Lachlan Orogen underwent several shortening events at *ca* 455, 440, 425, 380 and 340 Ma, and the timing of peak metamorphism during the Alice Springs Orogeny (*ca* 380 Ma: Ballévre *et al.* 2000) correlates with the Tabberabberan Orogeny. It may be that the evolution of the Alice Springs Orogen involved a number of discrete events (see Bradshaw & Evans 1988) that have yet to be dated. If the correlations between the Lachlan Orogen and central Australia are valid it may be that the tectonic stresses applied to the plate margin were propagated through the crust and affected large areas of the continent interior. Deformation may have been focused in regions where highly radiogenic granites in the basement rocks thermally weakened the crust (Sandiford & Hand 1998; Hand & Sandiford 1999).

GROWTH OF THE PALAEOZOIC LITHOSPHERE

The Lachlan and New England Orogens are probably not a unique geodynamic setting, but may be unusual in that they have evolved over ~500 million years without continent–continent collision. It appears that the last stage of the 'Wilson Cycle' had not occurred in eastern Australia during the Palaeozoic. During this period the Australian continent grew by 30% as a consequence of eastward accretion above a long-lived west-dipping subduction zone (Gray *et al.* 1998). As pointed out by Collins (in press), this tectonic setting will result in long-lived extensional tectonics with transient episodes of crustal shortening. We are not surprised, therefore, that the Lachlan and New England Orogens are underlain by relatively slow and thin lithospheric mantle (Simons *et al.* 1999; Debayle & Kennett 2000a) typical of extended lithosphere. In this context, the relatively fast lithosphere beneath the western Lachlan Orogen and the Delamerian Orogen may represent the vestiges of extended Proterozoic lithosphere preserved since continental breakup occurred during the Neoproterozoic. The relatively fast lithosphere

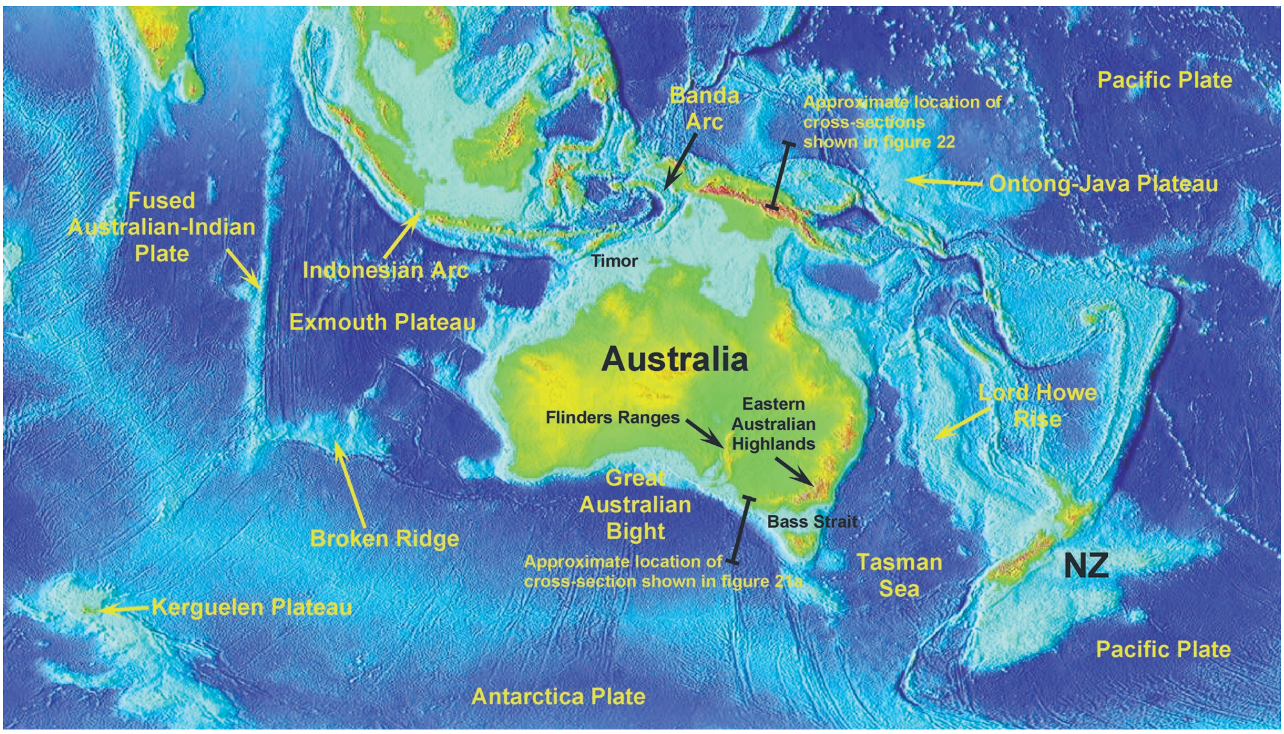


Figure 18 Topographic and bathymetric map of the Australian Plate showing the location of the major geological features discussed in the text. The approximate locations of cross-sections in Figures 21a and 22 are shown.

at depths between 200 and 400 km beneath central and northern Queensland may be an underthrust relic of the west-dipping subduction zone above which the New England Orogen developed during the Palaeozoic.

Seismic-refraction profiles of the eastern Lachlan Orogen show increasing seismic velocity in the lower crust, suggesting that it becomes more mafic with depth (Finlayson *et al.* 1979). It has been suggested that the mafic lower crust formed by imbricate stacking of oceanic crust in the basement of the western and central Lachlan Orogen (Gray & Foster 1998). It has also been argued that the mafic lower crust formed during late orogenic mafic underplating in the eastern Lachlan Orogen (*ca* 390 Ma) (Collins in press). This underplating is consistent with a lithosphere that evolved in an 'extensional accretionary orogen' dominated by backarc extension (Collins in press) and provides a heat source for the voluminous Lachlan Orogen granites.

MESOZOIC EVOLUTION OF AUSTRALIA

Extensional tectonism and the breakup of Gondwana

The existence of the Gondwanan supercontinent modified planetary dynamics in two significant ways that may have eventually contributed to its destruction (Veevers 1990). First, the supercontinent may have acted as an 'insulating blanket', reducing the efficiency of heat loss and leading to a thermal anomaly beneath the supercontinent. Second, the resultant large oceans that coexisted with the supercontinent must have become increasingly old and dense. Eventually the age of the subducting lithosphere may have become such that its increased density drove accelerated seaward retreat of the subduction zone. Rollback of the hinge of the subducting lithosphere would have acted as a powerful driving force leading to extension in the backarc and to magmatism and volcanism at the margins of the continent. Breakup was also coincident with the arrival of numerous plumes (e.g. Kerguelen, Karoo–Ferrar). The evolution of much of the Australian Plate since the Triassic has been dominated by the dispersion of Gondwana fragments that resulted in the development of passive margins preserved on the east, south and west coasts of Australia today (Figure 18).

In Western Australia, Permian to Triassic extensional tectonism led to the formation of the Perth Basin, which extended inland over the Yilgarn Craton (Hocking & Preston 1998). Carboniferous to Triassic dolerite dykes were intruded during northeast–southwest extension associated with sinistral transtension and rifting between the West Australian Craton and Greater India (Harris & Li 1995). Continued breakup resulted in the formation of Permian to Triassic intracontinental rift basins. Lithospheric extension continued into the Jurassic (e.g. Vulcan Sub-basin: Baxter *et al.* 1998). Eventual breakup (*ca* 135 Ma) led to formation of the North West Shelf passive margin and the initiation of oceanic spreading during the Late Jurassic (Etheridge & O'Brien 1994; Baxter *et al.* 1998).

The earliest evidence for the initial continental extension of Gondwana in eastern Australia is recorded in the voluminous Jurassic (*ca* 180–175 Ma) dolerites that exist over a large area of central and eastern Tasmania

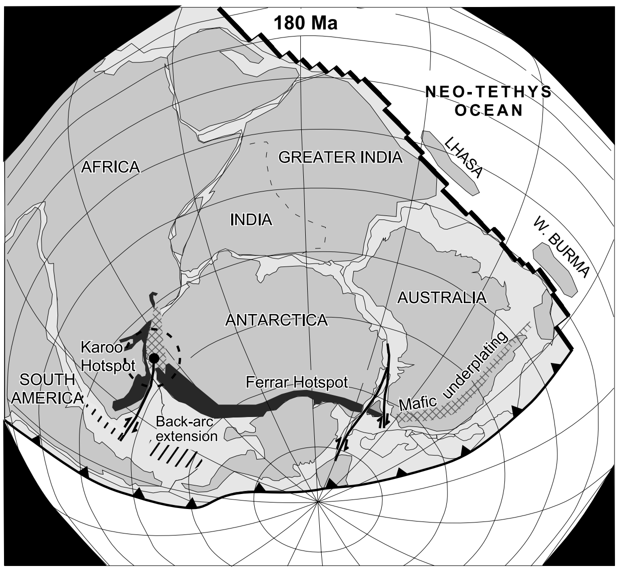


Figure 19 Reconstruction of East Gondwana supercontinent showing the position of a major Large Igneous Province (LIP) that extended from the Karoo hot spot through the Transantarctic Mountains and into Tasmania. Uplift along eastern Australia during the formation of this LIP possibly indicates mafic underplating and a northern extension of the LIP. Reconstruction developed using the PlatyPlus software.

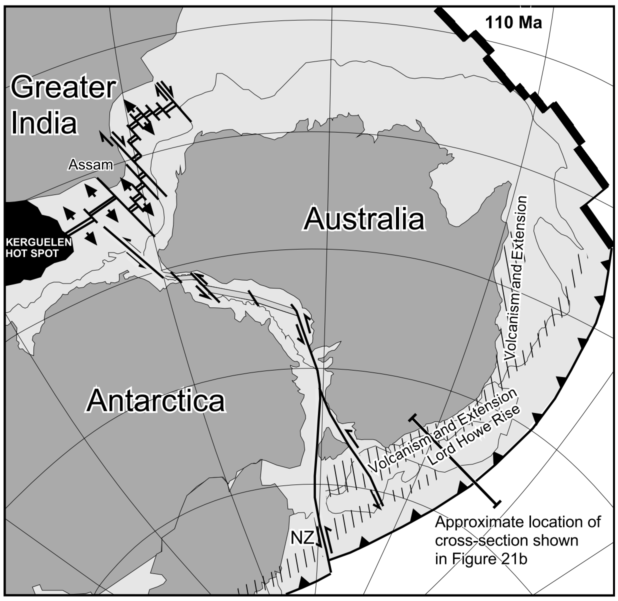


Figure 20 Reconstruction of Australia and Antarctica at *ca* 110 Ma showing the distribution of continental volcanism that may be attributed to the rollback of the Pacific subduction zone during the onset of rifting along the eastern margin of Australia. The location of the Kerguelen Plume is shown and the architecture of the southern Australian basins is dominated by east-northeast dextral strike-slip faults. Reconstruction developed using the PlatyPlus software.

(Hergt *et al.* 1989; Hergt & Brauns 2001). These dolerites form part of a Large Igneous Province that extended 3500 km through the Transantarctic Mountains (Ferrar) (Heimann *et al.* 1994) and into the Karoo Province of South Africa (Karoo) (Encarnacion *et al.* 1996) between *ca* 194 and 175 Ma (Figure 19). This Large Igneous Province is widely regarded as the result of a mantle plume (White & McKenzie 1989), although it was emplacement in a zone of extension and subduction along the proto-Gondwana margin (Storey 1995) and its magma geochemistry that indicate a lithospheric source and subduction-related signature (Hergt *et al.* 1989; Hergt & Brauns 2001). It has been proposed that dolerites were emplaced in a continental backarc environment during rollback of the subducting Pacific Plate (Vaughan & Storey 2000). Dalziel *et al.* (2000) alternatively suggest that the Karoo–Ferrar Large Igneous Province impinged on the subducting slab causing slab flattening. Thermal erosion of the slab in front of a newly formed subduction zone resulted in Large Igneous Province development in a continental backarc environment. The implication of this model is that the Karoo–Ferrar Large Igneous Province was a direct agent of Gondwana breakup (Dalziel *et al.* 2000), not a consequence of breakup.

The surface expression of the *ca* 184–175 Ma Large Igneous Province terminates at Tasmania (Hergt *et al.* 1989; Hergt & Braun 2001) and may have affected a large area of eastern Australia in the form of mafic underplates (Figure 19). Nott and Horton (2000) recorded the establishment of drainage divides in Cape York at 180 Ma and these data may provide the first clue of asthenospheric upwelling and underplate-related uplift during this *ca* 180 Ma magmatic event.

The basin signature of Gondwana breakup is not recorded in Australia until the Upper Jurassic with the development of rift basins preserved in the Great Australian Bight, the Perth Basin (*ca* 155 Ma) and in Bass Strait (*ca* 145 Ma) (Norvick & Smith 2001). The early stage

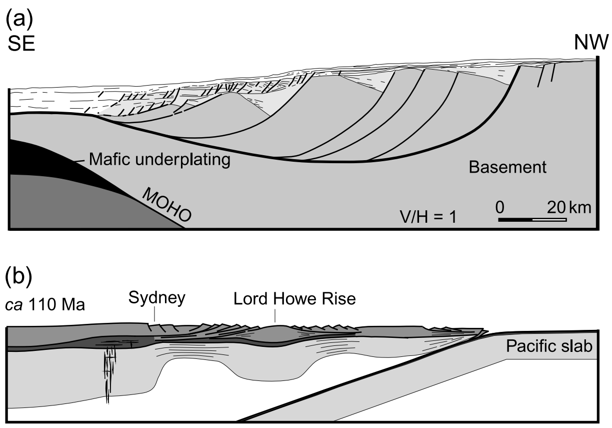


Figure 21 (a) Crustal cross-section of the southern Australian passive margin. The upper part of the section is constrained by seismic-reflection data. See Figure 18 for location. (b) Inferred crustal cross-section of the eastern Australian passive margin at *ca* 110 Ma showing detachment style faulting and asymmetric extension before the opening of the Tasman Sea. Lithospheric extension is interpreted to have occurred during the hinge retreat of the subducting Pacific plate. See Figure 20 for location.

of rifting between Australia and Antarctica was dominated by left-lateral strike-slip faulting from 125 to 110 Ma (Figure 20). The extensional architecture is dominated by west-northwest transform faults that may have provided the underlying structural template for continuing extension in the Gippsland and Otway Basins (Lavin 1997). In the context of this architecture, it is possible that Tasmania may have been translated to the southeast during movement along these strike-slip faults (Harrington 1979).

Basaltic volcanism in the Perth Basin at *ca* 132 Ma may have been associated with the arrival of the Kerguelen Plume (Frey *et al.* 1996) (Figure 20). These basalts have been interpreted to reflect incubation of the Kerguelen Plume beneath the West Australian Craton before the formation of the tholeiitic Large Igneous Province (*ca* 110 Ma: Frey *et al.* 2000) preserved in Broken Ridge and the Kerguelen Plateau (Figure 18) (Frey *et al.* 1996).

Episodic rifting (Williamson *et al.* 1990) continued along the southern marginal basins during the Cretaceous and appears to have formed rift valleys, which propagated eastward along the entire southern margin of Australia (Willcox 1990). After the initial stages of rift basin formation between Tasmania and Victoria, faulting began to wane and a sag-phase basin started to evolve (*ca* 110–100 Ma: Norvick & Smith 2001). Renewed rifting began at *ca* 95 Ma (Williamson *et al.* 1990), coincident with the final breakup of Antarctica and Australia (Cande & Mutter 1982). This breakup appears to have been little influenced by the geometry or structural grain of pre-existing cratons. The Gippsland, Bass and Otway Basins failed at this time, but were immediately inverted such that Tasmania remained attached to the remainder of the Australian continent (Hill *et al.* 1995a, b). In contrast, the western Otway Basin eventually became a lower plate passive margin (Lister *et al.* 1991) (Figure 21a).

Between 132 and 95 Ma, voluminous extension-related pyroclastic volcanism extended along a belt between the Whitsunday Province and the Otway/Gippsland Basin (Bryan *et al.* 1997, 2000). Magmatic episodes were spread over the interval, but were concentrated between 120 and 105 Ma, with peak magmatic activity at *ca* 110 Ma (Figure 20). Bryan *et al.* (1997) suggested that these volcanic events are related to 'passive margin formation', but may equally be related to a major period of continental extension in long swaths parallel to a retreating subduction zone (Figure 21b) or to dynamic platform tilting associated with west-dipping subduction (Waschbusch *et al.* 1997). The *ca* 110 Ma age correlates extensional denudation ages in New Zealand (Forster 2001) and Australia (Foster & Gleadow 1993). Major crustal and lithospheric-scale fault zones in Victoria were also reactivated with kilometre-scale vertical displacements (Foster & Gleadow 1993).

During the Cretaceous, the eastern margin of Australia has been interpreted as an upper plate margin (Lister *et al.* 1991) that was rapidly denuded at *ca* 95 Ma (O'Sullivan *et al.* 1996, 1999) because of uplift and erosion of tilt-blocks (Foster & Gleadow 1993), thermal heating of the lithosphere caused by underplating of basaltic sills (Etheridge *et al.* 1989; Lister *et al.* 1991) and basin inversion (Hill *et al.* 1995a). Uplift along the eastern and southeastern margin of

Australia continued into the Tertiary (Kohn *et al.* 1999; O'Sullivan *et al.* 1999; van der Beek *et al.* 2001). This episode of uplift appears to occur along vertical dip-slip basement faults (van der Beek *et al.* 2001) and coincides with the onset of onshore volcanism, possibly indicating continued mafic underplating.

At 95 Ma, the Lord Howe Rise had not yet separated from the Australian continent (Lister *et al.* 1991) and the Tasman Sea was a zone of extended continental crust (Figure 21b). Cretaceous continental extension was highly asymmetric and resulted in the development of detachment style faulting during continental breakup (Figure 21b) (Etheridge *et al.* 1989; Lister *et al.* 1991). This led to the formation of conjugate, highly extended lower plate (Lord Howe Rise) and mildly extended, underplated upper plate (southeastern Australia) margins. Recent studies of the evolution of the Tasman Sea suggest a complicated interaction of northward-migrating failed rifts and strike-slip faults (Gaina *et al.* 1998). Sea-floor spreading began at *ca* 80 Ma and several internal rift basins, including the Lord Howe Rise (Figure 18), were stranded on the lower plate margin (Etheridge *et al.* 1989). It has been speculated that sea-floor spreading was caused by the arrival of the Balleny Plume (Lanyon *et al.* 1993). The sea-floor spreading ridge terminated in the Coral Sea, although continental extension may have continued in this region.

Northward migration and Tertiary plate reorganisations

Since final breakup at *ca* 80 Ma, Australia has travelled north and is colliding with various microplates of southeast Asia. During this period there have been several plate reconfigurations and adjustments (Hall 1996; Veevers 2000). Sea-floor spreading terminated in the Tasman Sea at *ca* 54 Ma (Norvick & Smith 2001), but continued in the Southern Ocean. By *ca* 44–43 Ma, fast spreading had begun in the Southern Ocean, leading to rapid marine transgressions and subsidence on the margins of the Australian continent (Norvick & Smith 2001). This time corresponds to an increase in spreading rate along the southeastern Indian ocean ridge and a fundamental shift in the movement vector of the Pacific Plate (Veevers 2000) (Figure 21). The Pacific Plate changed direction (Veevers 2000) coincident with collision between fragments of Gondwana and Tethyan arcs and continent–continent collision between India and Asia (Searle *et al.* 1987; Le Pichon *et al.* 1992). The Indian Plate fused with the Australian Plate at *ca* 43 Ma (Liu *et al.* 1983; Veevers 2000) to compensate the radically different torque balances caused by the initiation of the Himalayan collision (Coblenz *et al.* 1995). Once this plate reorganisation had taken place the relative movement across many convergent zones in the southwest Pacific changed resulting in major episodes of magmatism and extension (Rawling & Lister 1999). This episode of plate reorganisation is manifested in the Australian continent by the formation of arches, domes and broad monoclines in the interior of the continent, fault reactivation and graben formation along the eastern Queensland coast (Veevers 2000).

Collision along the northern margin

The New Guinean Orogeny began at *ca* 30–25 Ma when several arc complexes, oceanic plateaux and microcontinents collided with the northern margin of the Australian Plate (Pigram & Davies 1987; Hall 1996) (Figure 22). North-dipping subduction is interpreted to have occurred beneath an island arc located to the north of the continental margin (Crowhurst *et al.* 1996), although south-dipping subduction beneath continental crust, separated from Papua New Guinea by a marginal ocean, has also been inferred (Hill *et al.* 1993) (Figure 22a). Arc accretion had occurred by the Late Oligocene (Figure 22b). During the Early Miocene, a retreating subduction hinge extended the overriding plate to produce metamorphic core complexes and starved sedimentary basins (Crowhurst *et al.* 1996) (Figure 22c). Voluminous Middle Miocene arc-related volcanism formed the Maramuni Arc (Crowhurst *et al.* 1996) (Figure 22d). The arrival of the buoyant Caroline Plate resulted in an episode of Late

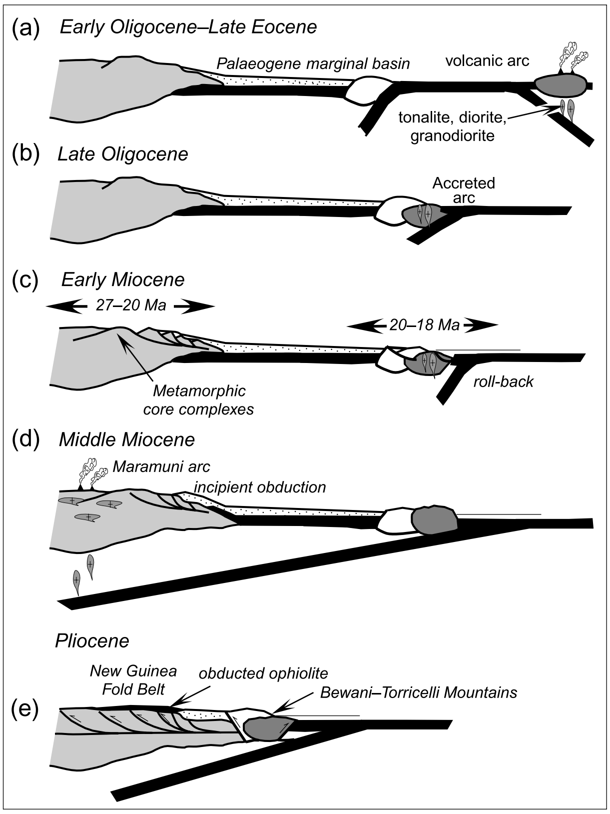


Figure 22 Schematic cross-section illustrating the evolution of the northern Australian margin since the Oligocene (adapted after Crowhurst *et al.* 1996). (a) Island-arc formation outboard of the Australian plate. There is conjecture as to whether the slab dips to the north or south. (b) Arc accretion and the development of new subduction zone during the Late Oligocene. (c) Early Miocene extension resulting in the development of metamorphic core complexes. Extension may be related to rollback of the subducting slab. (d) Middle Miocene obduction of marginal basin and arc magmatism. (e) Development of New Guinea Fold Belt and accretion of the island arc during the Pliocene. See Figure 18 for location.

Oligocene – Pliocene contractional orogenesis and obduction of oceanic crust from the marginal basin (Crowhurst *et al.* 1996) (Figure 22d). Fold and thrust belt development during the Pliocene (8–5 Ma) (Figure 22e) was followed by transpression along the northern margin of the Australian Plate (northern Papua New Guinea) (Hall 1996).

The most recent event to significantly affect Australia was the collision between the buoyant *ca* 130–110 Ma Ontong–Java Plateau (Figure 18) along the northern margin of the Australian Plate at *ca* 6 Ma. This oceanic plateau appears to have jammed the subduction system, changing torque balance, and thus forcing a reorganisation of global plate kinematics. Retreat of the Banda Arc since 5 Ma has resulted in overthrusting of Timor onto the Australian margin (Figure 18). Extensional tectonism along the northern margin of the Australian Plate since 3 Ma has resulted in the development of metamorphic core complexes of the Solomon Sea (Hill *et al.* 1995c) and detachment tectonics in the Woodlark Basin that have partly dismembered the Oligocene–Pliocene collisional orogen.

Neotectonics and southeastern Australian hot spot

In southeastern Australia, regional tilting and the development of disconformities within the Bass Strait basins, fault reactivation (up to 1 km) and localised uplift in the Strzelecki Ranges, Mt Lofty Ranges and Flinders Ranges (Sandiford & Gleadow 2001) (Figure 18), as well as the 'Kosciuszko Uplift', continues to the present day. This activity is interpreted by Sandiford and Gleadow (2001) to relate to changes in the intraplate stress regime associated with far-field plate boundaries. In addition, there has been significant intraplate tholeiitic and alkalic basaltic volcanism of the Newer Volcanic Province (*ca* 5–0 Ma) in central and western Victoria. This province has been interpreted by Wellman and McDougall (1974) to have occurred above a fixed mantle hot spot, with individual volcanoes becoming younger to the south.

Growth and destruction of the Australian lithosphere during the Mesozoic

Tomographic data show the presence of anomalously high-velocity mantle at depths of 300 km beneath Queensland. This is interpreted by Müller *et al.* (2000) to have originated from Late Eocene – Late Oligocene subduction from Australia's northern margin.

Tomographic data show that the southwestern parts of the Yilgarn lithospheric root may have been destroyed (Simons *et al.* 1999), the most likely causes of which are crustal extension during the breakup of the West Australian Craton and Greater India and/or thermal erosion of the root during the arrival of the Kerguelen Plume (Frey *et al.* 1996).

Along the eastern margin of Australia, continental growth additions as a result of mafic underplating may have occurred at *ca* 180 Ma and *ca* 95 Ma (Etheridge *et al.* 1989; Lister *et al.* 1991). Some of the thickest crust in eastern Australia is located beneath the Eastern Australian Highlands (Figure 2a), where mafic underplating occurred during asymmetric extension of the lithosphere during the opening of the Tasman Sea (Etheridge *et al.* 1989; Lister *et al.* 1991) (Figure 21b).

Plate tectonics and the Australian lithosphere

A reoccurring theme of Australian geology is the importance of the concepts and processes of modern plate-tectonic theory to ancient terranes. There has been continuing debate about the relative importance of lateral accretion (arc magmatism, Cordilleran-style tectonism and continental collision) versus vertical accretion (mafic underplating related to lithospheric extension or the arrival of mantle plumes) and there are numerous examples from the Archaean to the Mesozoic in which 'intraplate tectonism' has been interpreted outside the context of plate tectonics. This topic is likely to remain controversial because evidence of plate-margin processes will be poorly preserved and subject to interpretation in the ancient terranes. In addition, the nature of the plate-tectonic process may have changed throughout time.

At issue is: (i) the recognition of the cryptic evidence for plate-margin processes; and (ii) the relative importance of, stress propagation from the plate margins, the influence of mantle plumes and the architecture and composition of the continent interior, to the control of intraplate tectonism. With increased geochronological resolution, we are beginning to determine the temporal links between the evolution of the plate margins and the continent interiors that may reconcile the long-running 'fixist' versus 'mobilist' arguments.

In the past decade, our knowledge of the architecture and chemistry of the Australian lithosphere has improved significantly. There has been significant progress in our understanding of the tectonic evolution of the continent through all geological eras. This is not to say that we fully understand how the Australian continent has evolved. In this paper we have attempted to relate our outlook on the tectonic evolution of the Australian plate with the present-day architecture of the Australian lithosphere as defined by others. We have also attempted to highlight some of the more contentious and controversial issues concerning Australian tectonics. As our knowledge of both the lithospheric architecture and tectonic evolution of Australia increases, perhaps we will better understand and constrain how the tectonic evolution of the Australian Plate has influenced the Australian lithosphere as we see it today.

CONCLUSION

The Australian lithosphere has been shaped during a complicated evolution over the past 4.6 billion years. The Archaean Yilgarn, Pilbara and Gawler Cratons (*ca* 4.6–2.5 Ga) are underlain by thick lithosphere. This lithosphere is refractory and has survived subsequent Proterozoic and Phanerozoic tectonism. The tectonic evolution of these Archaean cratons has been influenced by a combination of lateral lithospheric accretion associated with processes analogous to plate tectonics, and vertical lithospheric accretion associated with the arrival of mantle plumes. Convective crustal reworking appears to have been a significant process for crustal differentiation during the middle Archaean, especially in the Pilbara Craton.

The Pilbara and Yilgarn Cratons collided during the Early Palaeoproterozoic to form the West Australian Craton. Thereafter, there was a protracted period of continental growth (*ca* 2.0–1.50 Ga) during which time the north, south and west Australian cratons were amalgamated. A series of extensional basins formed in the cratonic interior coincident with episodic accretion and magmatic-arc development at the southern and eastern continent margins (*ca* 1.80–1.50 Ga). During this time lithospheric thinning was offset by mafic underplating and sedimentation in extensional basins.

Between 1.45 and 1.10 Ga, the South Australian Craton broke away from the North Australian Craton and was reattached in its present orientation during Grenville-aged orogenic events. Subsequent Neoproterozoic continental breakup resulted in the development of the Centralian Superbasin in the continental interior and ribbons of Proterozoic lithosphere rifted from the eastern margin of the continent. These ribbons form part of the basement to the orogenic belts that formed along the eastern margin of the continent during the Palaeozoic evolution of eastern Australia.

The Palaeozoic belts developed mainly in a backarc environment with transient episodes of accretionary tectonism related to west-dipping subduction. During this time the Australian crust grew by ~30% without any evidence for continent–continent collision. The lithosphere beneath this part of the continent is relatively thin, reflecting its protracted extensional evolution along a predominantly accretionary orogen. Contemporaneous deformation and metamorphism in the continent interior may be related to accretionary tectonism along the northern margin of the continent.

Passive margins preserved on the east, west and south coasts of the Australian continent formed as a result of Gondwana breakup. Breakup was strongly influenced by several plumes (e.g. Kerguelen and Karoo–Ferrar), although slab retreat may have influenced breakup throughout eastern Gondwana. The Archaean lithosphere beneath the West Australian Craton may have been thermally eroded, whereas mafic underplating beneath the uplifted Eastern Australian Highlands contributed to crustal thickening. The northward migration of Australia during the Tertiary resulted in complex collisional tectonism along the northern margin of the plate.

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