Crustal architecture of the Capricorn Orogen, Western Australia and associated metallogeny


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A 581 km vibroseis-source, deep seismic reflection survey was acquired through the Capricorn Orogen of Western Australia and, for the first time, provides an unprecedented view of the deep crustal architecture of the West Australian Craton. The survey has imaged three principal suture zones, as well as several other lithospheric-scale faults. The suture zones separate four seismically distinct tectonic blocks, which include the Pilbara Craton, the Bandee Seismic Province (a previously unrecognised tectonic block), the Glenburgh Terrane of the Gascoyne Province and the Narryer Terrane of the Yilgarn Craton. In the upper crust, the survey imaged numerous Proterozoic granite batholiths as well as the architecture of the Mesoproterozoic Edmund and Collier basins. These features were formed during the punctuated reworking of the craton by the reactivation of the major crustal structures. The location and setting of gold, base metal and rare earth element deposits across the orogen are closely linked to the major lithospheric-scale structures, highlighting their importance to fluid flow within mineral systems by the transport of fluid and energy direct from the mantle into the upper crust.

KEYWORDS: Ashburton Basin, Capricorn Orogen, Collier Basin, Edmund Basin, Fortescue Basin, Gascoyne Province, Glenburgh Terrane, Narryer Terrane, Ophthalmian Orogeny, Pilbara Craton, seismic reflection study, suture zone, West Australian Craton, Yilgarn Craton.

INTRODUCTION

Like most continental regions on Earth, the Australian continent is particularly well endowed with several large, high-quality (or giant) orebodies including iron ore in the Hamersley Ranges of the Pilbara, lead–zinc at Broken Hill, uranium–copper–gold at Olympic Dam and lead–zinc–copper–gold at Mount Isa, as well as numerous high-quality gold and nickel deposits. The large range of commodities and abundance of deposits has been the main driver for Australia’s recent economic success and stability; however, these resources are being depleted at a rate greater than they are being discovered. The decline in exploration success is due mostly to the inaccessibility of the bedrock, up to 80% of which is covered by a thick layer of regolith, or is locally deeply weathered (Australian Academy of Science 2012). It also suffers from a lack of detailed geological knowledge in some underexplored (greenfields) parts of the continent. A variety of geological evidence has shown that the formation of giant orebodies or deposit clusters in the upper crust, are directly related to specific tectonic settings (e.g. Groves et al. 2005; Kerrich et al. 2005; Begg et al. 2010; Hronsky et al. 2012), and that they commonly occur above deep crustal structures that mark the edges of discrete continental blocks, such as fossil suture zones or the extended margins of rifted continents.

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Regional-scale geophysical surveys, including aeromagnetic, gravity and a range of passive and active seismic techniques, provide a relatively rapid method for greatly increasing our geological understanding of deeply buried or weathered geological terranes, as well as providing critical first-order constraints on crustal architecture (e.g. Golbey et al. 2006; Aitken & Betts 2009; Cayley et al. 2011). When these geophysical methods are combined with some degree of geological knowledge, regions of preferential mineral endowment can be identified, providing a focus for mineral exploration and targeting.

The Capricorn Orogen of Western Australia is a ~1000 km long, 500 km wide region of variably deformed rocks located between the Pilbara and Yilgarn Cratons (Figure 1) and records the Paleoproterozoic assembly of these cratons to form the West Australian Craton, as well as over one billion years of subsequent intracratonic reworking (Cawood & Tyler 2004; Sheppard et al. 2010a, b; Johnson et al. 2011a). Apart from the extensive high-grade iron-ore deposits along the southern Pilbara margin, the region has few working mines and has had little history of recent world-class orebody discovery. The western part of the orogen is particularly well exposed, and as a result the surface geology, geological history and plate tectonic setting of the orogen are mostly well understood (e.g. Sheppard et al. 2010b; Johnson et al. 2011a). However, owing to extensive crustal reworking, the location of major crustal structures and broad architecture of the orogen are poorly constrained. To improve the exploration potential of the region, a better understanding of the crustal architecture across the orogen is critical, especially identifying the location and orientation of major crustal structures and craton edges, as well as any island arc, or exotic accreted crustal material.

In April and May 2010, 581 km of vibroseis-source, deep seismic reflection data were acquired along three traverses (10GA-CP1, 10GA-CP2, and 10GA-CP3) through the well-exposed western part of the orogen. These data were interpreted in conjunction with 400 m-line-spaced aeromagnetic and 2.5 km-spaced regional gravity data, as well as detailed gravity and magnetotelluric data that were collected along the length of the survey. The traverses started in the southern part of the Pilbara Craton, crossed the Gascoyne Province, and ended in the Narrryer Terrane of the Yilgarn Craton (Figures 2, 3). The project was a collaboration between the Geological Survey of Western Australia (GSWA), AuScope (a component of NCRIS, the National Collaborative Research Infrastructure Strategy), and Geoscience Australia. The seismic data were processed by the Seismic Acquisition and Processing team of the Onshore Energy and Minerals Division at Geoscience Australia and interpreted by the authors of this paper during several interpretation sessions. The seismic processing steps, 2.5D geophysical forward modelling and geological interpretation of the seismic data are outlined in detail in Johnson et al. (2011b) and in the Supplementary Paper. This paper presents the results of that final interpretation in order to construct a rigid architectural and geological framework for the orogen, and to better understand the nature and geometry of events associated with both the assembly and punctuated reactivation of the West Australian Craton. This improved and integrated geological dataset also provides critical

![Figure 1](image1.png)

**Figure 1** Elements of the Capricorn Orogen and surrounding cratons and basins, showing the location of the Capricorn Orogen seismic transect. Modified from Martin & Thorne (2004). The thick grey lines delimit the northern and southern boundaries of the Capricorn Orogen. Inset shows location of the Capricorn Orogen and Paleoproterozoic crustal elements (KC, Kimberley Craton; NAC, North Australian Craton; SAC, South Australian Craton; WAC, West Australian Craton) and Archean cratons (YC, Yilgarn Craton; PC, Pilbara Craton; GC, Gawler Craton). Other abbreviations: MI, Marymia Inlier; YGC, Yarlarweelo Gneiss Complex.
Figure 2 Geological map of the western Capricorn Orogen, showing the location of the Capricorn Orogen seismic lines 10GA–CP1, 10GA–CP2, and 10GA–CP3.
geodynamic information that may be used to identify regions of the upper crust that may contain a preferential mineral endowment or be more prospective for particular commodities.

GEOLOGICAL SETTING

Early interpretations of the Capricorn Orogen (e.g. Horwitz & Smith 1978; Gee 1979) favoured an intracratonic setting. These were followed by plate tectonic models in which the Proterozoic orogeny was driven by collision between the hitherto unrelated Pilbara and Yilgarn Cratons (Muhling 1988; Tyler & Thorne 1990; Blake & Barley 1992; Powell & Horwitz 1994; Evans et al. 2003).

More recent work, however, has highlighted the complexity of the Capricorn Orogen, which has now been shown to record at least seven major orogenic events (Figure 4; e.g. Sheppard et al. 2010b), the oldest two of which, the 2215–2145 Ma Ophthalmian Orogeny and the 2005–1950 Ma Glenburgh Orogeny, relate to the punctuated assembly of the West Australian Craton (Blake & Barley 1992; Powell & Horwitz 1994; Occhipinti et al. 2004; Sheppard et al. 2004; Rasmussen et al. 2005; Johnson et al. 2011a), while the younger five, record over one billion years of intracratonic reworking (Cawood & Tyler 2004; Sheppard et al. 2010b; Johnson et al. 2011a).

The Pilbara Craton and Archean to Proterozoic sedimentation in the northern Capricorn Orogen

During the late Archean and Paleoproterozoic, prior to the collision of the Pilbara Craton with the Glenburgh Terrane of the Gascoyne Province, the southern margin of the craton evolved from an intracontinental rift to a passive margin, before being converted to an active margin, and subsequently into a series of foreland basins (Blake & Barley 1992; Krapez 1999; Martin et al. 2000; Thorne & Trendall 2001; Trendall et al. 2004; Martin & Morris 2010). This history is marked by the deposition of the Fortescue, Hamersley, Turee Creek, and lower Wyloo Groups. Following the assembly of the West Australian Craton, Paleoproterozoic intracontinental reactivation resulted in the deposition of the upper Wyloo and Capricorn Groups (Figures 4, 5) in a possible foreland basin setting. Based on the known stratigraphy, stacking of all the Archean and Paleoproterozoic supracrustal units present in the northern Capricorn Orogen gives a maximum cumulative thickness of ~26 km (Figure 5).
Figure 3 Regional geophysical data for the western Capricorn Orogen with 2.5 km Bouguer anomaly gravity data (colour) draped over 300–500 m line-spaced TMI aeromagnetic data (black and white). The location of the structural-metamorphic zone boundaries of the Gascoyne Province, major fault and mineral deposits is also shown.
PILBARA CRATON BASEMENT

Within the northern Capricorn Orogen, granite–greenstone basement rocks of the Pilbara Craton are exposed in the Wyloo, Rocklea and Milli Milli Domes, and in the Sylvania Inlier (Figures 1, 2). Greenstones are dominated by low-grade metamorphosed mafic volcanic and siliciclastic sedimentary rocks. The minimum age of the granite–greenstones is fixed by the ca 2775 Ma age of the overlying basal Fortescue Group (Trendall et al. 2004). Their maximum age is unknown, although comparison with similar granite–greenstone assemblages in the northern Pilbara Craton (Van Kranendonk et al. 2007; Hickman & Van Kranendonk 2008) suggests they probably formed sometime between ca 3800 and ca 2830 Ma.

FORTESCUE GROUP: RIFTING OF THE PILBARA CRATON

Granite–greenstone rocks are unconformably overlain by mixed volcano-sedimentary rocks of the Fortescue Group, formed during protracted rifting of the Pilbara Craton between ca 2775 and ca 2630 Ma (Blake 1993, 2001; Thorne & Trendall 2001; Blake et al. 2004; Trendall et al. 2004). In the northwestern and northeastern Pilbara, Fortescue Group sedimentation and volcanism was controlled by a northeast-striking extensional fault system (Blake 2001), whereas in the southern Pilbara, Thorne & Trendall (2001) argued that the fault system was principally oriented east-southeast.

In the South Pilbara Sub-basin (Thorne & Trendall 2001), the Fortescue Group is up to 6.5 km thick
Figure 5 Generalised stratigraphy and deformation history of the northern Capricorn Orogen. The tectonic interpretation is modified after Martin & Morris (2010).

(Figure 5), and is subdivided, into seven formations, which are grouped into four major tectono-stratigraphic units. From the base upwards, Unit 1 consists of the Bellary Formation and Mount Roe Basalt; Unit 2 is the Hardey Formation; Unit 3 consists of the Boongal, Pyracle, and Bunjinah formations; and Unit 4 is the Jeerinah Formation. Thick mafic to ultramafic sills intrude much of the succession above the Mount Roe Basalt. The origin of the four major tectono-stratigraphic units of the Fortescue Group are interpreted to result from deposition in an extensional tectonic setting (Thorne & Trendall 2001).

HAMERSLEY GROUP: PASSIVE- TO ACTIVE-MARGIN SEDIMENTATION AND IGNEOUS ACTIVITY

The 2.5–3.0 km thick Hamersley Group conformably overlies the Fortescue Group (Figure 5), and was deposited between ca 2630 and ca 2450 Ma, when the southern Pilbara evolved from a rift setting to a passive continental margin and finally to an active margin (Morris & Horwitz 1983; Blake & Barley 1992; Krapež & McNaughton 1999; Thorne & Trendall 2001; Trendall et al. 2004; Martin & Morris 2010). Rocks of the Hamersley Group reflect this largely deeper-marine distal setting, and are dominated by banded iron-formation, shale, chert and fine-grained carbonate.

Seven major lithostratigraphic units are recognised within the Hamersley Group. In ascending order, these are the Marra Mamba Iron Formation, Wittenoom Formation, Mount Sylvia Formation, Mount McRae Shale, Brockman Iron Formation, Weeli Wolli Formation, and Boolgeeda Iron Formation (Trendall & Blockley 1970; Simonson et al. 1993; Trendall et al. 2004).

TUREE CREEK GROUP AND LOWER WYLOO GROUP: FORELAND BASIN SEDIMENTATION AND VOLCANISM DURING THE OPHTHALMIAN OROGENY

Deposition of the Turee Creek and lower Wyloo Groups, took place during the early foreland-basin stage of the 2215–2145 Ma Ophthalmian Orogeny, when the Pilbara Craton collided with the Glenburgh Terrane to the south (Blake & Barley 1992; Martin et al. 2000; Occhipinti et al. 2004; Sheppard et al. 2004, 2010b; Rasmussen et al. 2005; Johnson et al. 2010, 2011a; Martin & Morris 2010). An alternative view by Krapež & McNaughton (1999), considered the post-Turee Creek Group history in terms of two mega-sequences that record the opening and closure of an Atlantic-type ocean.

Along the southern Pilbara margin, the Turee Creek and lower Wyloo groups have maximum thicknesses of about 4 and 3 km, respectively, although the Turee Creek Group was not crossed during the survey. The upper part of the lower Wyloo Group is dominated by the ~ 2 km thick, ca 2210 Ma Cheela Springs Basalt, and both the Turee Creek and lower Wyloo Groups are intruded by similar-aged dolerite sills and dykes (Martin
et al. 1998; Müller et al. 2005; Martin & Morris 2010). The two groups are separated by a significant angular unconformity (Figure 5), and, together with another occurring within the upper Turee Creek Group, are interpreted to reflect the northward propagation of the Ophthalmia fold-and-thrust belt into the foreland basin (Martin & Morris 2010).

UPPER WYLOO AND CAPRICORN GROUPS: FORELAND-BASIN SEDIMENTATION AND VOLCANISM DURING THE CAPRICORN OROGENY

The upper Wyloo Group (Figure 5) has an estimated thickness of about 7.5 km, and, in ascending order, consists of the Mount McGrath Formation, Duck Creek Dolomite, and Ashburton Formation (Thorne & Seymour 1991). A ~120 m-thick, mafic and felsic volcanic unit, the June Hill Volcanics, overlies the Duck Creek Dolomite north of the Wylooo Dome, although this unit was not crossed during the survey. The Ashburton Formation is unconformably overlain by siliciclastic, carbonate and felsic volcanic rocks of the Capricorn Group, deposited following the early deformation stage of the 1820–1770 Ma Capricorn Orogeny.

The age of the upper Wyloo Group is still poorly constrained. Evans et al. (2003) obtained an age of ca 1800 Ma for the June Hill Volcanics, whereas a tuffaceous unit in the overlying Ashburton Formation has been dated at ca 1804 Ma (Sircroome 2003). The apparent contradiction in these ages, coupled with the ca 1804 Ma age of felsic volcanic rocks in the Capricorn Group (Hall et al. 2001), has led Evans et al. (2003) to suggest that the deposition of the upper part of the upper Wyloo Group and of the Capricorn Group were strongly diachronous owing to oblique basin closure during the Capricorn Orogeny.

The tectonic setting of the upper Wyloo Group is considered to be a retro-foreland basin associated with the onset of the 1820–1770 Ma Capricorn Orogeny (Tyler & Thorne 1990; Thorne & Seymour 1991; Evans et al. 2003), although Blake & Barley (1992) and Martin & Morris (2010) suggest that the setting may have been extensional in its early stages.

The Narryer Terrane of the Yilgarn Craton

The Yilgarn Craton is an extensive region of Archean continental crust dominated by granite–greenstone terranes. On the basis of recent geological mapping and a re-evaluation of geological data at all scales (Cassidy et al. 2006), this craton has been divided into several distinct tectonic entities—the Narryer, Youanmi and South West Terranes, and the Eastern Goldfields Superterran e. The Narryer and South West Terranes are dominated by granite and granitic gneiss with only minor supracrustal greenstone inliers, whereas the Youanmi Terrane and the Eastern Goldfields Superterran e contain substantial greenstone belts separated by granite and granitic gneiss. The Narryer Terrane contains the oldest rocks in Australia—the ca 3730 Ma Manfred Complex—as well as the oldest detrital zircons on Earth, at ca 4400 Ma (Froude et al. 1985; Compston & Pidgeon 1986; Wilde et al. 2001). Part of the Narryer Terrane was subject to extensive metamorphism and reworking during the 1820–1770 Ma Capricorn Orogeny and these strongly deformed parts have been termed the Yarlarweelor Gneiss Complex (Figures 1, 2; Sheppard et al. 2003). The Murchison Domain consists of granite–greenstones and contains rocks as old as ca 3000 Ma. All of the northern Yilgarn Craton terranes were intruded by granitic rocks between ca 2800 and ca 2600 Ma. All of the northern Yilgarn Craton (super)terranes were intruded by high-Ca TTG granitic rocks between ca 2690 and ca 2660 Ma, and low-Ca granites between ca 2660 and ca 2620 Ma (Cassidy et al. 2006; Champion & Cassidy 2007).

The Glenburgh Terrane and assembly of the West Australian Craton

The Glenburgh Terrane of the Gascoyne Province represents a continental fragment, exotic to both the Pilbara and Yilgarn Cratons. The oldest component of the Glenburgh Terrane, the 2555–2430 Ma Halfway Gneiss (Figures 2, 4), consists of heterogeneous granitic gneisses that contain abundant older inherited zircons, some of which are as old as ca 3447 Ma (Johnson et al. 2011c). Although no older crust (>2555 Ma) is exposed, the Lu–Hf compositions and crustal model ages of both magmatic and inherited zircons indicate a long crustal history, ranging back to ca 3700 Ma. These isotopic data also demonstrate that large parts of the unexposed terrane formed via juvenile crustal growth processes between ca 2730 and ca 2660 Ma. Formation of the 2555–2430 Ma gneisses occurred mainly by the in situ reworking of these older crustal components (Johnson et al. 2011c). A comparison of the U–Pb zircon ages and zircon–Hf isotopic compositions of the Halfway Gneiss with those of the bounding Pilbara and Yilgarn cratons, indicates that the Halfway Gneiss (and thus the Glenburgh Terrane) is exotic to, and evolved independently from, these cratons (Johnson et al. 2011c). The Glenburgh Terrane is interpreted to have collided with, and accreted to, the Pilbara Craton during the 2215–2145 Ma Ophthalmian Orogeny (Blake & Barley 1992; Powell & Horwitz 1994; Occhipinti et al. 2004; Johnson et al. 2010, 2011a, c).

The Moogie Metamorphics (Figures 2, 4) are dominated by psammitic schists, the protoliths to which were deposited across the Glenburgh Terrane in a time frame coincident with the 2215–2145 Ma Ophthalmian Orogeny. These rocks contain detrital zircon sourced from the southern Pilbara region, and are interpreted to have been deposited in a foreland basin that formed in response to uplift of the southern margin of the Pilbara Craton during the collision of the Glenburgh Terrane with the Pilbara Craton (Occhipinti et al. 2004; Johnson et al. 2010, 2011a; Martin & Morris 2010). Deposition of the Moogie Metamorphics occurred in a similar time frame to the Beasley River Quartzite of the lower Wyloo Group, which was also deposited in a foreland basin setting during the Ophthalmian Orogeny (Martin & Morris 2010).

Sometime after the collision between the Glenburgh Terrane and the Pilbara Craton, continental-margin arc-magmatic activity was initiated along the southern margin of this newly amalgamated block (Sheppard et al. 2004; Johnson et al. 2010, 2011a). The 2005–1975 Ma
Dalgaringa Supersuite is exposed in the southern part of the province (Figures 2, 4), and consists of granitic gneisses, which have major-, trace-, and rare earth element concentrations consistent with formation in a supra-subduction zone setting (Sheppard et al. 2004). Their whole-rock Sm–Nd, and magmatic zircon Lu–Hf, isotopic signatures indicate the incorporation of Neoarchean granitic gneisses with isotopic compositions similar to those of the Halfway Gneiss (Sheppard et al. 2004; Johnson et al. 2011a), suggesting that magmatism occurred in a continental-margin arc, the Dalgaringa Arc, which formed along the southern margin of the Glenburgh Terrane. This magmatic event records the progressive closure and northward subduction of oceanic crust under the combined Pilbara Craton–Glenburgh Terrane.

Terminal ocean closure, the collision between the amalgamated Pilbara Craton–Glenburgh Terrane and the Yilgarn Craton, and the formation of the West Australian Craton, all took place during the 1865–1950 Ma collisional phase of the Glenburgh Orogeny (Occhipinti et al. 2004; Johnson et al. 2010, 2011a). The collision resulted in the imbrication of the northern margin of the Yilgarn Craton with slices of the Glenburgh Terrane along the Errabiddy Shear Zone (Figure 2), and the high-grade tectonometamorphism of metasedimentary and meta-igneous rocks along the southern margin of the Glenburgh Terrane. Substantial uplift and exhumation of the Narryer Terrane also occurred during this event (Muhleng et al. 2008). The collisional event was accompanied by the intrusion of granitic stocks and dykes of the 1865–1945 Ma Bertibubba Supersuite, which are the first common magmatic element of the northern margin of the Yilgarn Craton, the Yarlarweelor Gneiss Complex, the Errabiddy Shear Zone, and the Paradise Zone of the Glenburgh Terrane (Figures 2, 4).

Intracratonic reworking, magmatism and sedimentation

Subsequent to the assembly of the West Australian Craton during the Glenburgh Orogeny, the history of the Capricorn Orogen is dominated by more than one billion years of episodic intracontinental reworking and reactivation, including sedimentation, magmatism, metamorphism and deformation (Figure 4).

THE 1820–1775 MA CAPRICORN OROGENY

Although the Capricorn Orogeny had been widely interpreted to be the result of oblique collision between the Yilgarn and Pilbara Cratons (Myers 1990; Tyler & Thorne 1990; Krapež & McNaughton 1999; Occhipinti & Myers 1999; Sheppard & Swager 1999; Pirajno et al. 2000; Sheppard et al. 2003, 2010b; Martin et al. 2005; Muhring et al. 2012). The orogeny is characterised by extensive compressional and possibly strike-slip deformation at low- to medium-metamorphic grades, and was accompanied by the intrusion of voluminous, felsic magmatic stocks and plutons of the Moorarie Supersuite, including the Minnie Creek and Landor batholiths (Figures 2, 4). Intense structural and magmatic reworking of the northern margin of the Narryer Terrane formed the Yarlarweelor Gneiss Complex (Figures 1, 2; Sheppard et al. 2003). Sedimentation is recorded across the orogen with the deposition of the upper Wyloo and Capricorn groups in the Ashburton Basin, and the Leake Spring Metamorphics in the Gascoyne Province (Figure 4).

THE 1680–1620 MA MANGAROON OROGENY

The Mangaroon Orogeny encompasses complex deformation, metamorphism, sedimentation and granite magmatism. Extensional-related faults and fold structures and metamorphic assemblages related to the orogeny appear to be restricted entirely to the Mangaroon Zone (Figures 2–4) in the northern part of the Gascoyne Province, although granite magmatism (the Durlacher Supersuite) and sedimentation (the Pooranoo Metamorphics) took place across the entire province (Figures 2, 4). In the southern part of the Gascoyne Province, the Pooranoo Metamorphics consist of a lower fluvial succession of sandstones, which are overlain by shallow-marine sandstones. Farther north, in the Mangaroon Zone, deep-water turbiditic sandstones dominate the succession. Deformation and medium-grade metamorphism is restricted to the Mangaroon Zone of the Gascoyne Province, which records high-temperature-low-pressure metamorphism, presumably under extensional crustal regimes (Sheppard et al. 2005). Metamorphism was accompanied by the intrusion of voluminous granitic stocks and plutons of the 1860–1620 Ma Durlacher Supersuite. Following peak metamorphism in the Mangaroon Zone, magmatism stepped across into the other parts of the Gascoyne Province, especially the Matherbukiin Zone (Figures 3, 4), where large volumes of megacrystic K-feldspar-phyric monzogranite (the Davey Well batholith; Figures 2, 4) were emplaced between ca 1670 and ca 1650 Ma. Although the driver of orogenesis is currently unknown, the high-temperature-low-pressure metamorphic conditions, and short duration of metamorphism, magmatism and sedimentation imply extension-dominated orogeny (Sheppard et al. 2005).

MESOAMPHROZOIC SEDIMENTATION IN THE EDMUND AND COLLIER BASINS

Following the Mangaroon Orogeny, mostly fine-grained siliciclastic and carbonate sediments were deposited in
the Edmund and Colliers basins (Figures 2, 4). The Edmund and Collier groups consist of 4–10 km of siliciclastic and carbonate metasedimentary rocks that were deposited under fluviatile to deep marine conditions. The groups have been divided into six depositional packages, each being separated by an unconformity or basal marine-flooding surface (Martin & Thorne 2004).

The Edmund Basin extends from the Pingandy Shelf in the north, and continues south across the Talga Fault as an extensional basin over 180 km in width (NE–SW) and 400 km in length (NE–SE) (Figures 1, 2). The sedimentary rocks unconformably overlie granites of the 1680–1620 Ma Durlacher Supersuite of the Gascoyne Province (Sheppard et al. 2010b), and in the uppermost part of the basin, locally includes volcanioclastic rocks dated at 1463 ± 8 Ma (Figure 4; Wingate et al. 2010). These constraints indicate deposition sometime between ca 1620 and ca 1465 Ma. Deposition within the Edmund Basin was controlled mainly by extensional movements on the Talga Fault (Martin & Thorne 2004), with the basin fill dramatically thickening to the southwest from the Pingandy Shelf. U–Pb SHRIMP dating of detrital zircons from the Edmund Group, and associated paleocurrent directions indicate that detritus was sourced predominantly from the northern part of the orogen, including the Fortescue and Hamersley groups of the southern Pilbara region (Martin et al. 2008). Subsequent basin inversion took place during the 1385–1200 Ma Mutherbukin Tectonic Event and the 1030–955 Ma Edmundian Orogeny, with minor fault reactivation taking place during the ca 570 Ma Mulka Tectonic Event.

Sedimentary rocks within the Collier Basin were deposited across both basement of the Gascoyne Province and locally deformed sedimentary rocks of the Edmund Basin. In the western part of the Capricorn Orogen, in the region of the seismic traverse, the basin is relatively restricted; although it widens significantly to the east, to a maximum of ~200 km (Figures 1, 2). The depositional age of the Collier Group is poorly constrained. Faulting associated with the 1385–1200 Ma Mutherbukin Tectonic Event (Johnson et al. 2011d) does not appear to affect rocks of the Collier Group, thus providing an upper age constraint for deposition. The lower age constraint is provided by voluminous dolerite sills of the ca 1070 Ma Kulkatharra Dolerite (Figure 4)—which form part of the Warakurna Large Igneous Province (Wingate et al. 2002)—and intrusion rocks of both the Edmund and Collier basins, thus deposition took place sometime between ca 1200 and ca 1070 Ma. Deposition also appears to have been less influenced by significant fault movements, which characterised the deposition within the Edmund Basin (Martin & Thorne 2004). U–Pb SHRIMP dating of detrital zircons from the Collier Group, and associated paleocurrent directions imply that detritus was sourced predominantly by the erosion of the underlying Edmund Basin (Martin et al. 2008). Subsequent basin inversion took place during the 1030–955 Ma Edmundian Orogeny.

THE 1385–1200 MA MUTHERBUKIN TECTONIC EVENT

The Mutherbukin Tectonic Event is a poorly defined tectono-thermal event, known primarily from the Mutherbukin Zone in the central part of the Gascoyne Province (Johnson et al. 2011d), although hydrothermal alteration and transpressional strike-slip faulting also affected rocks of the Edmund Basin. This event does not appear to have been associated with any significant periods of magmatism or sedimentation.

In the Gascoyne Province, the event is characterised by amphibolite facies metamorphism and deformation, which is restricted to a narrow 50 km wide corridor, bounded by the Ti Tree and Chalba shear zones (Figures 2–4). Dating of metamorphic monazite, mainly from garnet–staurolite schists, from widely spaced localities provides a range of ages between ca 1280 and ca 1200 Ma, interpreted as the age of deformation and metamorphism (Johnson et al. 2011d). In the overlying Edmund Group, evidence for Mutherbukin-age deformation is more cryptic, owing to its very low metamorphic grade and restriction to narrow shear zones and faults, which have subsequently been reactivated. Nevertheless, Mutherbukin-aged hydrothermal monazite and xenotime within these sedimentary rocks indicates that they were subject to low-grade metamorphism and hydrothermal alteration at this time (Rasmussen et al. 2010; Johnson et al. 2011d).

Although the driver of tectonism is not precisely known, it is noted that the timing and duration of deformation and hydrothermal fluid flow were synchronous with Stages I (1345–1260 Ma) and II (1225–1140 Ma) of the Albany-Fraser Orogeny (Spaggiari et al. 2011), as well as the Mount West (1345–1293 Ma) and Musgrave Orogeny (1220–1150 Ma) in the west Musgrave Province (Howard et al. 2011). The synchroneity of tectonism across the West Australian Craton, implies a period of continent-wide activity, possibly associated with a major period of continent-scale plate reorganisation, i.e. during progressive break-up of the Nuna Supercontinent (e.g. Evans & Mitchell 2011; Johnson 2013).

THE 1030–955 MA EDMUNDIAN OROGENY

The latest Mesoproterozoic to earliest Neoproterozoic Edmundian Orogeny is best known for widespread folding and low-grade metamorphism in the Edmund and Collier basins (Martin & Thorne 2004), although the orogeny was also responsible for reworking a southeast-striking corridor between the Chalba and Ti Tree shear zones in the Gascoyne Province (Figures 2–4; Sheppard et al. 2007). Within this corridor, deformation and metamorphism were accompanied, and post-dated, by the intrusion of leucocratic granite stocks and sheets, and rare earth element-bearing pegmatites, of the 1030–925 Ma Thirty Three Supersuite.

In the Edmund and Collier basins, the Edmundian Orogeny was responsible for low- to very-low-grade metamorphism, thrust faulting and transpressional-related upright and open folding (Martin & Thorne 2004; Martin et al. 2005). The fold and fault structures strike west-east to northwest–southeast, and are concordant with both the general basin architecture and the regional-scale structures in the underlying Gascoyne Province basement (Figure 2).

The interval between ca 1050 and ca 1000 Ma is commonly thought to mark the assembly of the Rodinia...
supercontinent (e.g. Li et al. 2008 and references therein), of which the Australian continent may have been an integral part. Collision between the eastern margin of Australia and a partly assembled Rodinia is estimated at ca 1000 Ma (Li et al. 2008), the timing of which coincides with the growth of peak metamorphic phases during the Edmundian Orogeny. In all reconstructions of Rodinia (e.g. Pisarevsky et al. 2003; Li et al. 2008), however, the western margin of the West Australian Craton is shown to face an open ocean. If so, then the deformation and metamorphism associated with the Edmundian Orogeny may have been a response to plate reorganisation and collisions elsewhere in Rodinia, since no other impinging crustal block was inferred to be present to the west.

THE CA 570 MA MULKA TECTONIC EVENT

The Mulka Tectonic Event is responsible for a series of anastomosing shear zones or faults that cut rocks of the Gascoyne Province and Edmund and Collier groups across the southwestern part of the Capricorn Orogen (Figure 4). This tectonic event is characterised by fault reactivation, rather than reworking. Mulka-aged faults are generally concentrated within discrete corridors such as the Chalba and T Tree shear zones (Figures 2–4) and show small dextral offsets in the order of 10–100 m, although cumulative offsets across the Chalba Shear Zone are up to 35 km (Sheppard et al. 2010b). The Mulka Tectonic Event is coeval with the Petermann, Paterson and King Leopold Orogenies, and reflects an episode of intracontinental reactivation during the assembly of the Gondwana Supercontinent (Johnson 2013).

GEOLOGICAL INTERPRETATION OF THE SEISMIC DATA

The seismic reflection profiling provided very clear images, identifying many structures that extend right through the crust (Figures 6–9). The depth to the base of the crust varies significantly, changing from a shallow but variable Mohorovičić (‘Moho’) character visible beneath the Pilbara Craton (Figures 6, 9), to a deep, indistinct crust–mantle boundary beneath the southern part of the Capricorn Orogen (Figures 7–9), and passing into a shallower Moho once the northern edge of the Yilgarn Craton is reached at the southern end of line 10GA–CP3 (Figures 8, 9). The upper crust can be sub-divided into several provinces, basins and zones, based principally on surface geological mapping and potential-field data (Johnson et al. 2011b). By comparison, the lower crust appears to consist of at least three discrete seismic provinces. Following Korsch et al. (2010), we use the term ‘seismic province’ to refer to a discrete...
volume of middle to lower crust, which cannot be traced to the surface, and whose crustal reflectivity is different to that of laterally or vertically adjoining provinces.

The Pilbara Craton (seismic line 10GA–CP1)

The Moho is weakly defined and has a gently undulating character, at a depth of 11.5–12.3 s two-way travel time (TWT) (about 34–37 km using an average crustal velocity of 6000 m/s) (Figures 6, 9). The Baring Downs Fault is a moderate to steeply north-dipping fault that extends through the crust to the Moho. The seismic character of the mid and lower crust on either side of this fault is markedly different. To the north of the Baring Downs Fault, beneath the Archean to Paleoproterozoic supracrustal rocks, the crust shows a broad twofold subdivision into a generally weakly reflective upper crust, corresponding to the exposed granite–greenstones of the Pilbara Craton, and a moderately reflective lower crust referred to as the Carlathunda Seismic Province, as its composition is not known, since it has not been tracked to the surface. The boundary between these crustal divisions is undulating and offset by the major faults, but generally occurs at a depth of 4–7 s TWT (12–21 km). To the south of the Baring Downs Fault, the mid- to lower-crustal levels can be subdivided into a generally strongly reflective middle crust, and a weakly reflective lower crust, the boundary undulating at depths of 8.3–9.5 s TWT (25–27.5 km). The crust to the south of the Baring Downs Fault is interpreted as a separate seismic entity, the Bandee Seismic Province, which also is not exposed at the surface.

In addition to the Baring Downs Fault, other major crustal structures—the Nanjilgardy, Soda, Moona, Beasley and Blair faults—are imaged in the seismic data (Figures 6, 9). Both the Soda and Nanjilgardy faults are interpreted to extend through the crust to the Moho. The Nanjilgardy Fault is a single, steep, north- to northeast-dipping structure in middle- to lower-crustal levels, but splays upwards into a complex, transpressive flower structure defined by steep to flat-lying, northeast- or southwest-dipping, minor faults (Figure 6, 9). Both the Moona and Soda faults are steep to subvertical close to the surface, but flatten out to form irregular, northeast- or southwest-dipping listric structures in the middle to lower crust. The Blair and Beasley faults are moderately to steeply southward-dipping in the upper parts of the profile, although become flatter at depth. Although all the major faults across the orogen are subparallel (Figure 2), the Baring Downs Fault marks a major axis across which the dip direction of the major faults change, with those to the north, all dipping moderate to steeply north-eastward, and those to the south, dipping moderately south-westward (Figure 6).

At the northeastern end of 10GA–CP1 (Figure 6), granite–greenstones of the Pilbara Craton imaged in the
Rocklea Dome are generally weakly reflective, although a number of prominent southwest-dipping to flat-lying reflections are recorded at depth. These reflections are interpreted as discontinuous greenstone sheets within the granitic crust.

The Fortescue Group is most prominent in the northeastern part of the section (Figures 6, 9) where it dips gently to the southwest, and is imaged as a series of weak to strong, layered reflections. Based on their seismic reflectivity, three units can be recognised within the Fortescue Group: a lower layer of weak reflections 0.25 s TWT (~0.8 km) thick that possibly corresponds to sedimentary rocks of the Hardey Formation; a middle layer of strong reflections 0.7 s TWT (~2.2 km) thick that equates to the mostly basaltic rocks of the Boongal, Pyradie and lower Bunjinah formations; and an upper layer of weak reflections 0.5 s TWT (~1.4 km) thick, which includes the upper Bunjinah Formation and Jeerinah Formation. Although the seismic reflections lose some of their definition beneath the Turner Syncline, they suggest a southwesterly thickening of the Fortescue Group from about 1.5 s TWT (~4.5 km) to about 2 s TWT (~6 km) approaching the Moona Fault. The seismic data also suggest that a section of the Hamersley Group about 1 s TWT (~3 km) thick is preserved in the Turner Syncline.

A thick, almost complete, succession of the Fortescue Group crops out on the southwestern limb of the Rocklea

Figure 8 Migrated seismic section for line 10GA-CP3, showing both uninterpreted and interpreted versions. Display is to 60 km depth, and shows vertical scale equal to the horizontal scale, assuming a crustal velocity of 6000 m/s. Common Depth Point (CDP) locations are shown in Figures 2 and 3 (100 CDP = 2 km).
Dome, and is cut by the Karra Well and Soda faults (Figures 6, 9). Farther west, between the Soda and Nanjilgardy faults, a succession of weakly to strongly reflective lower and middle Fortescue Group rocks 1.7 s TWT (~5 km) thick is interpreted to overlie the weakly reflective granite–greenstone rocks.

The stratigraphic succession between the Nanjilgardy Fault and southwestern end of 10GA–CP1 (Figures 6, 9) has been difficult to interpret from the seismic reflection data alone. The surface geology consists of polyphase-deformed fold- and fault-repeated succession of rocks of the upper Wyloo Group, consisting of the Mount McGrath Formation, Duck Creek Dolomite, and Ashburton Formation, although this structural complexity is not imaged in the subsurface, probably because of a lack of continuous reflective markers within the stratigraphy. In the surface geology, the presence of both Fortescue and Hamersley group rocks to the south of the Nanjilgardy Fault along strike of the seismic traverse, implies that both successions should be present beneath the Ashburton Basin. The 2.5D forward modelling of gravity and aeromagnetic data (Johnson et al. 2011b)

Figure 9 Cross-section of the transect across the Capricorn Orogen, combining seismic lines 10GA–CP1, 10GA–CP2, and 10GA–CP3, and showing key faults, terranes, zones, basins and seismic provinces.
suggest that banded iron formation of the Hamersley Group must be present to the south of the Baring Downs Fault and so these units, although somewhat thinned, have been extend southward onto the northern part of line 10GA–CP2 (Figures 7, 9). Although these layers do not have the same reflective signature as the exposed Hamersley and Fortescue groups observed at the northern end of 10GA–CP1 (Figure 6), this may be due to an increased structural complexity and, in particular, the presence of numerous low-angle faults, beneath the Ashburton Basin. On the northern part of 10GA–CP2 (Figures 7, 9), these layers are dissected by a series of north-verging thrust faults which terminate within the lower Wyloo Group.

The Gascoyne Province and Yilgarn Craton (seismic lines 10GA–CP2 and 10GA–CP3)

The location and character of the Moho varies considerably across 10GA–CP2 and 10GA–CP3 (Figures 7–9). Under the Gascoyne Province, in 10GA–CP2 (Figure 7), the Moho is deep and undulating, varying between 12 and 15.3 s TWT (36–46 km), and continues this trend into the northern part of 10GA–CP3 (Figure 8), where the Moho is at 15 s TWT (~45 km) depth. In the central part of 10GA–CP3, the Moho is interpreted to be at about 16 s TWT (~48 km) depth, but at the southernmost end of the seismic line, the Moho is difficult to interpret. Line 10GA–CP3 ties to another, more recently acquired, line across the southern Carnarvon Basin, 11GA–SC1 (Korsch et al. 2013), and ends 40 km to the west of the start of the Youanmi seismic line, 10GA–YU1 (Zibra et al. 2013). Based on these three seismic lines, the Moho at the southern end of 10GA–CP3 is interpreted to have been faulted and duplicated by crustal-scale thrusting (Figure 8) marking the subsurface boundary between the Yilgarn Craton and the Glenburgh Terrane. Therefore, the Glenburgh Terrane may be present below the Narryer Terrane on seismic line 10GA–YU1 at around 15–16 s TWT (45–48 km).

CRUSTAL TERRANES, SEISMIC PROVINCES AND SUTURE ZONES

On the basis of seismic character and surface geology, two terranes—the Glenburgh and Narryer Terranes—and three seismic crustal provinces—the Yarraquin, MacAdam and Bandee Seismic Provinces—have been identified within lines 10GA–CP2 and 10GA–CP3 (Figures 7–9). Most of these are separated by major crustal structures, which coincide with mapped surface faults or shear zones (Figures 2, 3).

Within line 10GA–CP2, the Glenburgh Terrane and MacAdam Seismic Province are separated from the Bandee Seismic Province by the moderately south-dipping Lyons River Fault (Figures 7, 9), which in the middle and upper crust spays into the Ti Tree Shear Zone. Since rocks of the Bandee Seismic Province are not exposed at the surface, it is not possible to determine if this province is similar to the Glenburgh Terrane in terms of its age, lithological makeup and composition. However, the contact between the two, along the Lyons River Fault, is marked by a zone of strong seismic reflections that parallel the fault (areas A and B in Figure 10); there is a significant step in the Moho where the Lyons River Fault intersects the upper mantle (area C in Figure 10); and although the seismic characters of the middle crust of both the Glenburgh Terrane and Bandee Seismic Province are relatively similar, the Glenburgh Terrane is characterised by a non-reflective lower crust—the MacAdam Seismic Province—that is not evident in the Bandee Seismic Province immediately north of the Lyons River Fault (Figures 7, 10). These observations suggest that the Lyons River Fault represents a major suture zone that separates two, distinct pieces of continental crust.

Based on the interpretation of deep seismic reflection data in 11GA–SC1 (Korsch et al. 2013), which crosses the

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**Figure 10** Interpreted migrated seismic section of part of line 10GA–CP2, showing the location of the suture zone between the Glenburgh Terrane and Bandee Seismic Province at the Lyons River Fault. Display shows vertical scale equal to the horizontal scale, assuming a crustal velocity of 6000 m/s.
southern part of 10GA–CP3, and 10GA–YU1 (Zibra et al. 2013) to the east, the Yilgarn Craton at the southern end of 10GA–CP3 has been subdivided into the Narryer Terrane and the Yarrraquin Seismic Province, the two separated by the Yalgar Fault. The Yarrraquin Seismic Province (Korsch et al. 2013) forms a highly reflective lower crustal package that underlies the entire Youanmi Terrane (Zibra et al. 2013). The position of the Yalgar Fault in 10GA–CP3 has been tied from 11GA–SC1 (Korsch et al. 2013) where the fault is imaged as an east-verging, west-dipping thrust. The flat geometry of the Yalgar Fault in this section is due to the thrust fault being imaged normal to its movement direction, with the Narryer Terrane moving out of the plane of section. Movement on this fault and the juxtaposition of the Narryer Terrane with the Yarrraquin Seismic Province is interpreted to predate the final collision of the Yilgarn Craton and the Glenburgh Terrane since the Yalgar Fault is cut by Glenburgh Orogeny-aged (ca 1950 Ma) structures (Korsch et al. 2013).

The Narryer Terrane and Yarrraquin Seismic Province are separated from the Glenburgh Terrane by a series of anastomosing, north-dipping faults known as the Errabiddy Shear Zone, and a single, moderately south-dipping fault named the Cardilya Fault, which is interpreted to be the main suture (Figures 8, 9). These two fault systems also segment the Glenburgh Terrane into two crustal elements:

- The Dalgaringa Supersuite, including the Nardoo Granite, which at the surface, represents the exhumed mid-crustal portions of a continental-margin arc known as the Dalgaringa Arc (Sheppard et al. 2004; Johnson et al. 2010, 2011a), and which occurs structurally above (the hangingwall) both the Errabiddy Shear Zone and the Cardilya Fault.

- The remainder of the Glenburgh Terrane, including the basement gneisses into which the continental-margin arc magmas were intruded, occur in the footwall of the Cardilya Fault, lying structurally below the Narryer Terrane.

The Cardilya Fault, which separates the Yilgarn Craton (the Narryer Terrane–Yarrraquin Seismic Province) from the Glenburgh Terrane and MacAdam Seismic Province, transects the entire crustal profile and can be imaged down to 15.5 s TWT (~46.5 km). It is interpreted to offset and duplicate the Moho, which under the Narryer Terrane, is much shallower at 12.5 s TWT (~37.5 km) depth (Figures 8, 9). These results are comparable to those obtained by passive-seismic methods (Reading et al. 2012) and gravity modelling (Hackney 2004). At the northern end of line 10GA–CP3, the footwall of the Cardilya Fault is marked by a thick package, up to 3 s TWT (~9 km) thick, of strong seismic reflections (area A in Figure 8) that are parallel to the fault. Similar packages are also observed adjacent to the boundary with the MacAdam Seismic Province (area B in Figure 8), and these areas are interpreted to represent strongly deformed parts of the Glenburgh Terrane. Lithological variations within the Errabiddy Shear Zone, such as imbricate slices of Narryer Terrane, or intrusions of the Bertibubba Supersuite, cannot be differentiated seismically.

### THE TALGA AND GODFREY FAULTS

Although the Talga and Godfrey faults do not appear to separate crust of differing seismic character, they are important structures because they transect the entire crustal profile, joining together before intersecting the Moho (Figures 7, 9, 11). Both faults are listric in the upper 10 s TWT (~30 km) of crust offsetting rocks as young as the 1620–1465 Ma Edmund Group, indicating they are Paleoproterozoic to Mesoproterozoic extensional structures which accommodated the deposition of sediments into the Edmund Basin. However, from 3 to 7 s TWT (9–21 km), the Talga Fault is parallel to an imbricate set of faults (area A in Figure 11) that are defined by series of parallel reflectors (area B in Figure 11) indicating that it is a zone of intense deformation. The faults offset units of both the Fortescue and Hamersley groups (area C in Figure 11) and terminate within the lower Wyloo Group, suggesting the faulting may be related to thrusting during the Ophthalmian Orogeny. The northward-directed thrust sense of movement is parallel to the transport direction of exposed northward-verging thrusts in the Ophthalmia Fold and Thrust Belt (Tyler 1991). Therefore, the Talga Fault appears to be a major crustal shear zone that forms part of a north-verging fold and thrust system that was subsequently reactivated as an extensional fault, accommodating the deposition of the Edmund Group.

### BATHOLITHS AND GRANITE INTRUSIONS

Within the Glenburgh Terrane and upper part of the Bandee Seismic Province, the upper 5 s TWT (~15 km) of the crust is defined by numerous, irregular and seismically non-reflective bodies (Figures 7, 9, 10), which are indicated by the mapped surface geology (Figure 2) to be granite plutons of the 1820–1775 Ma Moorarie Supersuite and 1680–1620 Ma Durlacher Supersuite. In the Mutherbukin Zone (Figures 2, 3, 9, 10), a single, large intrusion—the Davey Well batholith—belonging to the 1680–1620 Ma Durlacher Supersuite is imaged between the Chalba and Ti Tree shear zones (CDPs 11075–13325). The intrusion has a concave basal contact with the underlying Glenburgh Terrane (Figure 10) and ranges in thickness from 0.6 s TWT (~1.8 km) in the centre of the Mutherbukin Zone, to 3.8 s TWT (~11.5 km) at its northerly contact where it is truncated by the Ti Tree Shear Zone. The pluton thins rapidly to about 2 s TWT (~6 km) on the southern side of the Chalba Shear Zone, where it has been downthrown across this fault to the south. Weak seismic reflections within and beneath the intrusion are parallel to the concave basal contact, suggesting that the shape and orientation of the pluton is due to folding during the Mutherbukin Tectonic Event. Prior to folding, the pluton would have been a flat-lying or gently dipping, tabular body greater than, or equal to, 11.5 km at its thickest.

In the Limejuice Zone, the Minnie Creek batholith (Figures 2, 3, 7, 9, 10) is imaged north of the Ti Tree Shear Zone at CDP 11075, and is interpreted to continue northward, underlining rocks of the Edmund Basin in the Cobra Syncline to CDP 8975 (Figure 10). North of the Edmund Fault at CDP 9450, the batholith is cut by...
numerous listric normal faults, which have produced a series of rotated (southwest-side down) half grabens. The Minnie Creek batholith has not been imaged north of these half grabens (between CDP 8975 and the Lyons River Fault at CDP 8725), where the area is dominated by rocks of high seismic reflectivity (Figures 7, 10). The base of the Minnie Creek batholith is mostly interpreted as a fault contact, but between CDP 9900 and 10400, the batholith has a relatively sharp, flat, possibly intrusive contact with the underlying Glenburgh Terrane. In this section, the batholith has a thickness of 2.25 s TWT (C24 6.75 km), but its maximum thickness, immediately north of the Ti Tree Shear Zone, is 4.5 s TWT (C24 13.5 km); nevertheless, as the basal contact in this region is tectonic in origin, it is possible that the Minnie Creek batholith has been thickened during folding or faulting.

At several localities within the batholith (e.g. CDP 10300, 10500, and 11000), regions of highly planar seismically reflective material dip at moderate angle toward the batholith centre (Figure 10). At the surface, these bodies have been shown to be kilometre-scale rafts of low-metamorphic grade pelitic and semipelitic schist belonging to the 1840–1810 Ma Leake Spring Metamorphics (Figure 2). These packages, some of which are up to 8 km in length, most likely represent vestiges of a formerly coherent sedimentary succession, into which the granites were intruded. The preservation of these tabular metasedimentary packages suggest that the batholith may have formed by a series of sheet-like plutons, an interpretation supported in part by field evidence, which indicates that many of the granites in the Moorarie Supersuite, in the Limejuice Zone, have sheet-like geometries (Sheppard et al. 2010a, b).

Weakly reflective crust is imaged in the Mangaroon Zone (Figures 7, 11, 12), which is bounded by the Lyons River Fault in the south (CDP 8720), and the Godfrey Fault to the north (CDP 6575). The area north of CDP 8300 is covered by sedimentary rocks of the Edmund Basin, and there are no surface exposures of basement rocks in this region (Figure 2). The crust in the Mangaroon Zone is interpreted as mostly granitic rocks of the 1680–1620 Ma Durlacher Supersuite, and metasedimentary rocks of the 1760–1680 Ma Pooranoo Metamorphics. In the centre of the zone (CDP 8300–8550), a package of moderately north-dipping, seismically reflective material (Figure 12) is interpreted to be a 6 km long raft of metasedimentary material of the Pooranoo Metamorphics. The reflective package also parallels numerous other weak seismic reflections that occur throughout the zone, and which are interpreted to reflect relict sedimentary or lithological layering within the Pooranoo Metamorphics. The presence of parallel seismic reflections down to ~2.0 s TWT indicates that the Pooranoo basin was at least ~6 km thick in the Mangaroon Zone.

Farther north, in the area underlying the Edmund Basin (south of the Godfrey Fault; Figure 11), it is difficult to determine if the non-reflective packages are granites of the Moorarie or Durlacher Supersuites, and for simplicity they are tentatively shown as a continuation of the interlayered granitic and metasedimentary material of Durlacher Supersuite and Pooranoo Metamorphics.
ARCHITECTURE OF THE EDMUND AND COLLIER BASINS

Although the Collier Basin succession in the Wanna Syncline is relatively thin, both Edmund and Collier basins are well imaged in line GA10–CP2 (Figures 7, 11). The maximum thickness of the Edmund Basin is 2.25 s TWT (~6.75 km) thick on the southern side of the Godfrey Fault, and in the core of the Wanna Syncline, the Collier Basin is ~0.25 s TWT (~750 m) thick. A relatively large package, up to 1 s TWT (~3 km) thick, of highly seismically reflective material occurs between the base of Package 2 and the top of Package 4 (Figure 11); these reflections are interpreted to be abundant dolerite sills of the ca 1465 Ma Narimbunna Dolerite and ca 1070 Ma Kulkatharra Dolerite, which intrude the upper parts of the Edmund Basin.

Three principal southwest-dipping structures, the Talga, Godfrey and Lyons River faults define three half grabens (Figures 7, 9, 11), into which sediments of the Edmund Basin were deposited (Martin & Thorne 2004), and across which significant variations in sediment thickness are evident. On the Pingandy Shelf, to the north of the Talga Fault (Figure 11), the maximum thickness of Packages 1 and 2 are ~0.5 s TWT (~1.5 km), whereas to the south they increase to ~1.25 s TWT (~3.75 km). Across the Godfrey Fault, Packages 1 and 2 increase from 2.0 s TWT (~6 km) to ~2.75 s TWT (~8.25 km) thick. Packages 1 and 2 are consistently thicker in the hangingwall of the basin-bounding extensional faults, suggesting that extensional downthrow on these major faults was toward the southwest. Packages 1 and 2 are not present south of CDP 8300, presumably having been incised and eroded away prior to the deposition of Packages 3 and 4 (Martin et al. 2008). In this region, Package 3 is at least 1 s TWT (~3 km) thick.

Between the Godfrey Fault (CDP 6575) and the Edmund Fault (CDP 9465), the Edmund Basin (Package 3) is intensely folded, with numerous imbricate thrust faults truncating the limbs of the tight, upright folds (Figure 11).

DISCUSSION

Crustal architecture, assembly and reactivation of the West Australian Craton

The new deep seismic reflection imaging, extending from the southern Pilbara Craton to the northern Yilgarn Craton, provides, for the first time, a holistic view of the crustal architecture of the Capricorn Orogen (Figure 9). These data are important because they not only provide critical information on the timing, assembly and reworking history of the West Australian Craton, but also provide a rare insight into the anatomy of continental collision zones in general.

THE PILBARA CRATON–BANDEE SEISMIC PROVINCE SUTURE

The Pilbara Craton granite–greenstones and underlying Carathaunda Seismic Province are separated from the Bandee Seismic Province by the steeply north-dipping Baring Downs Fault. The timing of amalgamation is uncertain, but must pre-date the deposition of the 2775–2630 Ma Fortescue Group, since these rocks were deposited across the suture. The pre-ca 2775 Ma age for the collisional episode suggests that it may be contemporary with other suturing events during the 3070–3060 Ma Prinsep Orogeny, which amalgamated the East Pilbara Terrane and the numerous terranes of the West Pilbara Superterrane (Van Kranendonk et al. 2007; Hickman & Van Kranendonk 2008; Hickman et al. 2010).

Deposition of over 20 km of sediments into the Fortescue, Hamersley, Turee Creek, and Ashburton basins has covered the original suture, but the seismic reflection data indicate that reactivation of this structure has extended the fault to the surface, where it is mapped as the present-day Baring Downs Fault. The seismic data also show evidence of post-1800 Ma (the minimum age of the upper Wyloo Group within the Ashburton Basin) extensional displacements on this structure with significant downthrow on the northeastern side (Figures 6, 9). The exact age or setting for reactivation is not known, nor is it clear whether any fault reactivation accompanied deposition of the upper Wyloo Group in the Ashburton Basin.

Although the Nanjilgardy Fault, which lies to the north of the Baring Downs Fault (Figures 6, 9), does not represent a suture zone, the fault transects the entire crustal profile, rooting in the Moho. At the surface, the fault forms the major lithological boundary between the 1840–1800 Ma upper Wyloo Group in the Ashburton Basin and older ca 2200 Ma metasedimentary and metavolcanic rocks of the lower Wyloo, Turee Creek, Hamersley and Fortescue groups. The age of initiation, or significance of the Nanjilgardy Fault, is not known, but the present-day architecture as a positive flower structure must have been imparted during or after the 1820–1770 Ma Capricorn Orogeny as rocks of the upper Wyloo Group have been displaced.

THE BANDEE SEISMIC PROVINCE–GLENBURGH TERRANE SUTURE

The Glenburgh Terrane–MacAdam Seismic Province and Pilbara Craton–Bandeel Seismic Province are sutured along the moderately south-dipping Lyons River Fault (Figures 2, 7, 9). From a variety of surface geological information, the collision is interpreted to have occurred during the 2215–2145 Ma Ophthalmian Orogeny (Occipinti et al. 2004; Johnson et al. 2010, 2011a, c; Martin & Morris 2010). However, intense crustal reworking of the entire northern Gascoyne Province and the intrusion of voluminous plutonic granitic rocks during episodic reworking (including the 1820–1770 Ma Capricorn and 1680–1620 Ma Mangarooorogenies) has obscured any evidence of this collision (Sheppard et al. 2005, 2010a, b). The suture zone is now represented at the surface by faults that have propagated through the younger magmatic and metamorphic rocks. To the north of the Lyons River Fault, metasedimentary and metavolcanic rocks of the Fortescue and Hamersley groups are interpreted to underlie much of the Ashburton Basin, occurring as far south as the Talga Fault in the Gascoyne Province (Figures 6, 7, 9). Both groups appear to be faulted and dissected by a series of north-directed thrusts that terminate within,
and are progressively truncated by the lower parts of the 2450–2210 Ma lower Wyloo Group, suggesting that thrusting was related to collision during the Ophthalmian Orogeny. Uplift and erosion of the Fortescue and Hamersley groups along this axis during thrusting, may have provided much of the detritus for the foreland basin deposits of the lower Wyloo Group (Martin & Morris 2010) and the Moogie Metamorphics in the Gascoyne Province (Johnson et al. 2011a).

During intracontinental reactivation, the Lyons River and Geegin faults (Figure 12) were reactivated as extensional structures allowing the deposition of up to 6 km of turbiditic sedimentary rocks (the 1740–1680 Ma the Pooranoo Metamorphics) into the Mangaroon Zone, immediately prior to the 1680–1620 Ma Mangaroon Orogeny. Outside the Mangaroon Zone, the Pooranoo Metamorphics consist of less than 700 m of medium- to coarse-grained fluvial to shallow-marine sedimentary rocks.

Following the Mangaroon Orogeny, the Lyons River, Talga and Godfrey faults, appear to have controlled much of the depositional history of the 1820–1465 Ma Edmund Basin (Johnson et al. 2011a). These three Ophthalmian-aged structures were reactivated in an extensional setting to form a series of half grabens (Figures 7, 9, 11) into which sediments of the Edmund Group were deposited. Package thickness variations across the major faults, especially from the Pingandy Shelf southward across the Talga Fault (Figure 11), imply intermittent and variable amounts of extensional movements on all three faults. The progressive truncation of older packages by units higher in the succession indicates syn-depositional uplift and erosion during periods of short-lived and punctuated reverse movements. Regional-scale post-depositional basin inversion is evident as a series of thrusts and hangingwall anticlines on fault splays between the Lyons River and Godfrey faults (Figure 11). This corridor of intense deformation has been mapped in the surface geology (Figures 2, 3), and does not appear to have significantly affected rocks of the Collier Basin, suggesting either that the deformation is related to the 1385–1200 Ma Mutherbukin Tectonic Event, or that the Godfrey Fault acted as a backstop to the northward-propagating thrust system during the 1030–955 Ma Edmundian Orogeny.

THE GLENBURGH TERRANE–YILGARN CRATON SUTURE

The final assembly of the West Australian Craton occurred when the northern margin of the Yilgarn Craton was sutured to the composite Pilbara–Bandee–Glenburgh craton at about 1965 Ma, during the Glenburgh Orogeny, along the Cardilya Fault. Owing to the intrusion of voluminous post-collisional granitic rocks of the 1820–1775 Ma Moorarie Supersuite across the Cardilya Fault during the Capricorn Orogeny (Sheppard et al. 2010a), the significance of this structure as the main suture zone was not recognised until the interpretation of the seismic data. Prior to the seismic investigation, the moderately well-exposed north-dipping Errabiddy Shear Zone, which contains an imbricate assemblage of lithologies from both the Glenburgh and Narryer Terranes, was thought to be the main suture zone (e.g. Sheppard et al. 2010b).

Although there is little evidence of post-collisional reactivation on the Cardilya Fault and Errabiddy Shear
Zone in the seismic data, the Cardilya Fault truncates and displaces plutonic magmatic rocks, specifically the Landor batholith, of the 1820–1775 Ma Moorangie Supersuite, indicating reactivation during, or after, the 1820–1770 Ma Capricorn Orogeny. Uplift and reworking across the Errabiddy Shear Zone are also recorded by several \(^{40}\text{Ar}/^{39}\text{Ar}\) mica dates between ca 960 and ca 820 Ma (Occhipinti 2007), indicating a long-lived post-collisional history for this fault system.

**Implications for Continental Collision and Subsequent Crustal Reworking**

The West Australian Craton was built progressively over one billion years from the punctuated collision of continental blocks (the Bandee Seismic Province, the Glenburgh Terrane and the Yilgarn Craton) to the southern Pilbara Craton margin. Owing to significant post-collisional tectonomagmatic reworking and vertical extension of the major fault structures into younger rock packages, the principal crustal sutures only have a weak expression at the surface, and ophiolites, high-pressure rocks or melanges, which are usually considered key components of continental suture zones, are either not preserved or not exposed. However, the suture zones, as well as other major crustal-scale structures related to the assembly of the West Australian Craton, are well preserved and imaged to great depth in the seismic reflection data (Figures 6–9). The preservation of ancient reworked suture and crustal-scale structures within the mid- to lower-crust appears to be a common feature of many continental suture zones, where examples have been documented in seismic reflection data from the Trans-Hudson Orogen (e.g. White et al. 2005), Slave Province (e.g. Clowes et al. 2005; Cook & Erdmer 2005) and numerous Australian orogenic belts (e.g. Korsch et al. 2010, 2012; Cayley et al. 2011; Wyche et al. 2013).

The preservation of ancient suture zones in the deep crust is therefore critical for understanding the crustal architecture and evolution of the collision zone, but does it reliably preserve pre-collisional geometric relationships such as the subduction polarity? Furthermore, in complex collision zones, what stage during the collisional/orogenic cycle does the present-day architecture reflect? In the Capricorn survey, both the 2215–2145 Ma Ophthalmanian and 2005–1950 Ma Glenburgh Orogeny suture zones (the Lyons River and Cardilya faults, respectively) dip moderately southward (Figure 9). The simplest interpretation implies south-directed subduction of oceanic crust during both events. However, the wealth of whole-rock geochemical, isotopic and \textit{in situ} zircon Lu–Hf isotopic data (Sheppard et al. 2004; Johnson et al. 2011a) demonstrates that during the 2005–1950 Ma Glenburgh Orogeny, the subduction of oceanic crust was north-directed with the formation of the Dalgarri Arc within the southern margin of the Glenburgh Terrane, opposite to the present-day dip of the suture zone. The current crustal architecture implies significant syn- to post-collisional modification of the crustal profile either by a style of ‘crocodile tectonics’ (e.g. Meissner 1989; Sadowial et al. 1991; Xu et al. 2002) or by the interplay of complex syn- to post-collisional events such as crustal obduction, backthrusting and crustal delamination. This implies that the crustal architecture was preserved relatively late in the orogenic cycle and that pre-collisional features may not be inherently preserved in the orogenetic architecture. Additional information, such as whole-rock geochemical and isotopic data (e.g. Sheppard et al. 2004), or where the subduction-related magmatic rocks are not accessible, \textit{in situ} Hf isotopic composition of appropriately aged inherited zircon grains, are thus required to constrain pre-collisional relationships. However, such information is not available to constrain the subduction polarity leading up to the pre-ca 2775 Ma collision between the Pilbara Craton and Bandee Seismic Province or the 2215–2145 Ma Ophthalmanian-aged collision between the Bandee Seismic Province and the Glenburgh Terrane.

Following the final assembly of the West Australian Craton, the orogen was subject to over one billion years of punctuated intracontinental reactivation and reworking (e.g. Sheppard et al. 2010b), including present-day seismic movements (Revets et al. 2009; Keep et al. 2012). The three principal sutures associated with the assembly of the West Australian Craton, as well as most of the other crustal structures, show evidence for multiple reactivations (Figure 9), in extensional, compressional and oblique-slip settings. Reactivation of these major crustal structures has led to their vertical extension into younger sedimentary or plutonic rocks (a process common to ancient collision zones; e.g. O’Driscoll 1986; Hronsky et al. 2012), and so the suture zones, and other associated crustal-scale faults, are commonly only weakly expressed at the surface by much younger faults or fault systems (Figures 2, 3, 9). The accommodation of regional-scale strain (extensional, compressional and oblique-slip) during the numerous reworking events appears to have been accommodated mostly by the reactivation of the pre-existing faults and shear zones, rather than by the generation of new crustal-scale structures. Furthermore, in the Gascoyne Province, the orientation of most tectonometamorphic fabrics and syn-tectonic magmatic intrusions such as the Minnie Creek, Landor and Davey Well batholiths (Figure 2) are also coaxial to the main crustal-scale structures. The depositional architecture of the 1620–1465 Ma Edmund and 1200–1070 Ma Collier basins, and subsequent style of basin inversion were controlled by extensional and transpressional movements on these structures and, more importantly, their orientation (Johnson et al. 2011a). Therefore, the crustal architecture, imparted during the assembly of the West Australian Craton ca 2200 to ca 1950 million years ago, has strongly influenced the partitioning of deformation, metamorphism and magmatism during the numerous intracratonic reworking events.

**Mineral prospectivity**

The Capricorn seismic survey has identified a series of major crustal structures, many of which cut through the crust to the mantle and show evidence for multiple reactivation (Figure 9). Numerous mineral deposit types have been recognised throughout the orogen (Figure 3) and include the world-class hematite iron-ore deposits of the Hamersley Basin; volcanic-hosted massive sulfide
(VHMS) copper–gold deposits in the Bryah Basin on the Yilgarn Craton margin; orogenic lode-gold mineralisation, such as that at Peak Hill in the southern Capricorn Orogen margin, Glenburgh and the Star of Mangaroon in the Gascoyne Province, and Paulsens and Mount Olympus on the northern Capricorn Orogen margin; various intrusion- and shear zone-related base metal, tungsten, rare earth element, uranium and rare-metal deposits in the Gascoyne Province; and lead–copper–zinc sediment-hosted mineralisation at Abra within the Edmund Basin (Figure 3). However, the tectonic setting of these deposits or their relation to the crustal architecture has yet to be adequately assessed.

Central to the orogen’s prospectivity is the growing understanding that the formation of ore deposits is an expression of much larger earth-system processes, which operate on a variety of scales to focus mass and energy flux. Many factors including the geodynamic setting, architecture, metal source, fluid-flow drivers and pathways, and depositional mechanisms (Wyborn et al. 1994; Knox-Robinson & Wyborn 1997), influence the type, style and location of mineralisation. One of the aims of the seismic survey was to identify large-scale crustal structures, including those that cut through the crust to the mantle, which may form pathways for fluid flow to mineral systems. A recent mineral prospectivity study of the Gascoyne Province, using the mineral systems approach, has also highlighted the significance of these major crustal structures to sites of potential mineralisation (Aitken et al. 2013). Although factors such as rock type, rock composition and metal source are important for mineralisation, the faults and, in particular, their constant reactivation during subsequent and punctuated orogenesis is key to focussing the fluids and energy flux of the mineral system. Especially important are the structures that transect the entire crustal profile (Figure 3). The Soda and Karra faults in the northern Capricorn Orogen are associated with the premium martite-microplaty hematite ores. The Nanjilgardy Fault hosts significant gold deposits in both the Fortescue Group of the Wyloo Dome at Paulsens and Proterozoic rocks of the Ashburton Basin such as Mount Olympus. The Baring Downs Fault hosts several gold and base metal deposits in the Ashburton Basin such as Star of the West, Glen Florrie and Soldiers Secret. In the Gascoyne Province, the composite Lyons River–Minnie Creek–Minga Bar Fault system hosts rare earth elements at Gifford Creek, and gold at the Star of Mangaroon. An along-strike extension of the Lyons River Fault, the Quartzite Well Fault, hosts the polymetallic Abra base metal deposit. The Ti Tree and Chalba shear zones host numerous tungsten, molybdenum and copper–gold base metal deposits, and the Cardilya–Deadman Fault hosts gold at Glenburgh. Apart from the Mt Olympus gold deposit, which has been dated at ca 1740 Ma (Sener et al. 2005), the age of the various mineral deposits across the orogen are not precisely known, and so it is difficult to determine whether they represent primary deposits related directly to tectonic setting (i.e. craton edge or island arc), or whether they are younger, secondary hydrothermal deposits related to fluid flow during intracontinental reactivation.

In the case of the Mt Olympus deposit, which lies along the Nanjilgardy Fault, the gold is interpreted to be of hydrothermal origin (Sener et al. 2005). The seismic data show the Nanjilgardy Fault to have been (re)activated during the 1820–1770 Ma Capricorn Orogeny consistent with the age dating and demonstrate that hydrothermal fluid flow within this major-crustal structure is an important mechanism for gold mineralisation.

The polymetallic Pb–Cu–Zn deposit at Abra in the lower part of the Edmund Group (which lies on an extension of the Lyons River Fault) shows complex history of monazite and xenotime growth between ca 1385 and ca 1235 Ma (Rasmussen et al. 2010). The deposit has been interpreted to be SEDEX in origin (Vogt & Stumpf 1987; Collins & McDonald 1994), but the polyphase growth of phosphate minerals implies significant secondary upgrading associated with hydrothermal fluid flow (Pirajno et al. 2008; Rasmussen et al. 2010) during fault reactivation.

The Lyons River Fault also hosts significant rare earth elements within the Gifford Creek Carbonatite Suite (formerly the Gifford Creek Complex; Pearson 1996; Pearson & Taylor 1996; Pearson et al. 1996).
(Figures 3, 13). The suite consists of a series of feroan carbonate sills that are distributed within a 30 km long northwest-trending belt that parallels the Lyons River Fault (Pearson 1996; Pearson & Taylor 1996; Pearson et al. 1996). The sills were emplaced into metasedimentary and granitic rocks of the Pooranoo Metamorphics and Durlacher Supersuite, respectively and so are younger than ca 1620 Ma. During subsequent tectonometamorphism, hydrothermal alteration of the carbonatites led to extensive upgrading of rare earth elements into a series of mineralised ironstone veins and dykes. Although the ages of the carbonate sills and subsequent rare earth element-rich ironstones are not precisely known (< ca 1620 Ma), the emplacement of the alkaline rocks implies the interaction of mantle-derived melts with thinned, metasomatitised, subcontinental lithospheric mantle (Foley 2008). Emplacement of the sills in the upper crust was controlled by ?extensional movements on the Lyons River Fault, and subsequent rare earth element-upgrading took place during hydrothermal activity.

These three examples highlight the importance and interplay between hydrothermal activity, the reactivation of lithospheric-scale faults, and gold, base metal and rare earth element mineralisation. Although the ages or structural settings of other deposits in the orogen are not precisely known, it is noted that the main lithospheric-scale structures (at the surface) host most of the known deposits. This association highlights the importance of major crustal-scale structures as conduits of fluid flow to mineral systems by the transport of fluid and energy direct from the mantle into the upper crust. However, these structures also coincide with former craton edges and possible island arcs (areas of preferentially endowed crust) and so, with regard to the source of metals, there may be a deeper, underlying ‘tectonic’ control (e.g. Groves et al. 2005; Kerrich et al. 2005; Begg et al. 2010; Hronsky et al. 2012) on the location of particular commodities within the orogen. The clustering of gold and base metal deposits around the three main suture zones (Figure 13) may reflect this underlying tectonic control, but some deposits, such as the rare earth element deposits at Gifford Creek, show a direct mantle-derived component for the metal source (the carbonatite sills). In reality, mineralisation is likely to reflect both end-member components and various combinations of the two. However, since most of the former island arcs and craton edges are now deeply buried or reworked, the most important factor for mineral prospectivity is the presence and location of repeatedly reactivated lithospheric-scale faults (Figure 13).

CONCLUSIONS

The Capricorn Orogen deep seismic reflection survey has imaged the crust and upper mantle structure, providing, for the first time, a holistic view of the crustal architecture of the Western Australian Craton. The survey demonstrates that the craton was built from four main crustal components: the Pilbara and Yilgarn Cratons, the Glenburgh Terrane and the Bandee Seismic Province, by the progressive and punctuated collision of continental blocks to the southern Pilbara Craton margin. This collisional architecture has fundamentally controlled the style and orientation of over one billion years of intracontinental reactivation, including deformation, metamorphism, magmatism and sedimentation.

The three main crustal suture zones, as well as other major crustal structures, are well preserved in the mid and lower crust, but are only weakly expressed at the surface by vertical extensions of these structures through younger (meta)sedimentary and (meta)igneous rock packages during intracontinental reactivation. The present-day architecture is the culmination of numerous collision-related processes that may have been preserved relatively late in the orogenic cycle. Therefore, the imaged architecture may not be used to reconstruct the pre-collisional tectonic settings such as the subduction polarity.

The majority of known mineral deposits in the Capricorn Orogen are spatially associated with the surface expression of the major crustal suture zones and other lithospheric-scale faults. These faults show evidence of having been multiply reactivated in an intracontinental setting, and highlight the importance of major crustal-scale structures and fluid flow to mineral systems by the transport of fluid and energy direct from the mantle into the upper crust.

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SUPPLEMENTARY PAPER

Seismic Reflection Data Processing


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