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THE PALEOPROTEROZOIC CAPRICORN OROGENY: INTRACONTINENTAL REWORKING NOT CONTINENT – CONTINENT COLLISION

by S Sheppard, S Bodorkos, SP Johnson, MTD Wingate, and CL Kirkland

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Cover photograph:
Creek pavements showing intimate intermingling of several phases belonging to the 1820–1775 Ma Moorarie Supersuite. Biotite-rich metatonalite forms abundant inclusions within a heterogeneous metagranodiorite to metamonzogranite, which is in turn veined by a leucocratic biotite metasyenogranite to metapegmatite

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The Paleoproterozoic Capricorn Orogeny: intracontinental reworking not continent–continent collision

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Abstract

The virtual obliteration of primary architecture and tectonostratigraphic relationships in most Precambrian orogens hinders the reconstruction of tectonic events, even at the first-order level. The Capricorn Orogen in Western Australia is a prime example of these difficulties. Although widely regarded as recording the collision of the Archean Yilgarn and Pilbara Cratons during the Capricorn Orogeny at 1820–1770 Ma, a recent program of regional mapping, ion microprobe (SHRIMP) U–Pb zircon dating, and whole-rock geochemistry in the Gascoyne Province, at the western end of the orogen, has identified no evidence of subduction-related magmatism, or of a suture that could be related to the Capricorn Orogeny. Instead, the spatial and temporal patterns of sedimentation and magmatism, along with the chemical and isotopic compositions of the granites in the province, are best explained by intracontinental reworking during the Capricorn Orogeny. The Capricorn Orogen formed by accretion of the Glenburgh Terrane (which forms the basement to the Gascoyne Province) to the Pilbara Craton during the Ophthalmian Orogeny at 2215–2145 Ma, followed by collision of this combined entity with the Yilgarn Craton during the Glenburgh Orogeny at c. 1950 Ma. Subseqent tectonism during the Capricorn Orogeny was probably driven by distal plate-margin processes.

KEYWORDS: Paleoproterozoic, orogenic events, ion microprobe U–Pb geochronology, granites, intracontinental reworking, Capricorn Orogen, Gascoyne Province, Capricorn Orogeny

Introduction

Unlike many Phanerozoic orogens where primary architecture and tectonostratigraphy can be either directly observed or confidently inferred (e.g. Goldfarb, 1997; Busby et al., 2006), in older orogens most of these primary relationships have been obliterated, and workers are forced to rely on less direct constraints (e.g. lithostratigraphy, spatial distributions, and relative chronologies of events; Holdsworth et al., 2001). As a result of these restrictions, the attendant interpretations are more sensitive to inaccurate input parameters. It is becoming increasingly clear that many Precambrian orogens that contain former sutures have undergone repeated reworking and reactivation, some for 1.5 Ga or more (e.g. Hand and Buick, 2001; Scrimgeour, 2003; Sheppard et al., 2008b). In this scenario, it is easy to produce '…*apparent* histories that point toward tectonic processes that may not have occurred' (Hand and Buick, 2001, p. 237). The Capricorn Orogen in Western Australia (Fig. 1) is a primary example of the problem of identifying even a first-order sequence of tectonic events in a multiply deformed orogen.

The Capricorn Orogen comprises low-grade metasedimentary and minor mafic metavolcanic rocks in the Bryah, Yerrida, Padbury, Earaheedy, Ashburton, Edmund, and Collier Basins, and granitic and medium- to high-grade metasedimentary rocks of the Gascoyne Province* (Fig. 1; Cawood and Tyler, 2004). Early intracratonic interpretations of the region invoking vertical tectonics (Daniels, 1975; Gee, 1979; Williams, 1986) have been superseded by the work of Tyler and Thorne (1990), who concluded that the orogen represented the juxtaposition of two previously unrelated cratons. Since then, it has become almost axiomatic that the Archean Yilgarn and Pilbara Cratons collided during the Capricorn Orogeny, following a period of oblique convergence. However, the work of Tyler and Thorne (1990) pre-dated the routine use of robust U–Pb geochronology, and what these workers considered to be a roughly 400 million year long orogenic event (2000–1600 Ma) is now known to comprise three distinct events: the 2005–1950 Ma Glenburgh Orogeny, the 1820–1770 Ma Capricorn Orogeny, and the 1680–1620 Ma Mangaroon Orogeny

* Use of the term 'Gascoyne Complex' is discontinued, because 'complex' is a lithostratigraphic term; therefore, the term 'Gascoyne Province' is reinstated. The following definition of a 'province' is modified from GSWA Memoir 2 (Trendall, 1975, p. 30): a volume of the earth's crust in which the rocks have some combination of geological characters in common, including either age, metamorphic grade, structural style or type of mineralization.

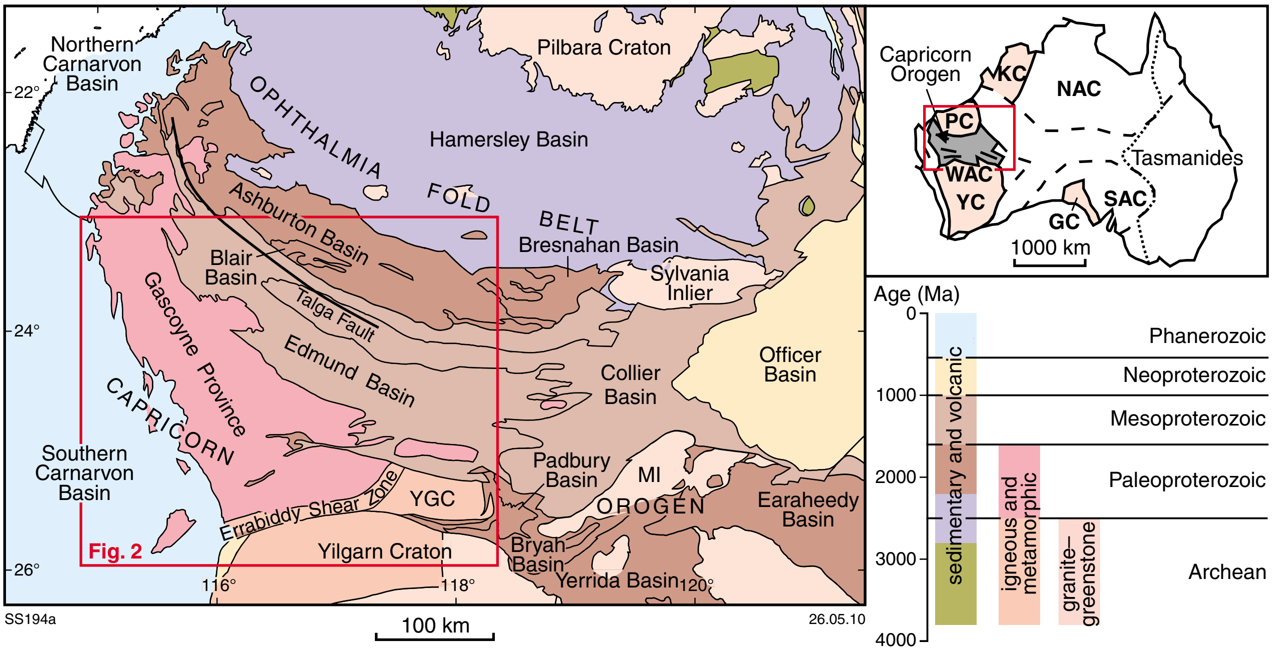

Figure 1. **Elements of the Capricorn Orogen and surrounding cratons and basins; modified from Martin and Thorne (2004, fig. 1). Inset shows location of the Capricorn Orogen, Paleoproterozoic crustal elements (NAC — North Australian Craton; SAC — South Australian Craton; WAC — West Australian Craton; KC — Kimberley Craton), and Archean cratons (YC — Yilgarn Craton; PC — Pilbara Craton; GC — Gawler Craton; YGC — Yarlarweelor Gneiss Complex; MI — Marymia Inlier); modified from Myers et al. (1996, fig. 2)**

(Occhipinti et al., 2001; Sheppard et al., 2005). Therefore, many of the structures, igneous rocks, and metamorphic assemblages attributed to the Capricorn Orogeny likely formed during either older or younger orogenic events (Occhipinti et al., 2004; Sheppard et al., 2005). Recent work has shown that some of the regional metamorphic mineral assemblages and structures in the centre of the Gascoyne Province widely attributed to the Capricorn Orogeny, are actually as young as Neoproterozoic (Sheppard et al., 2007; Johnson et al., 2009).

Existing models for the 1820–1770 Ma Capricorn Orogeny are either based largely on interpretations of poorly dated metasedimentary successions in the northern part of the orogen, or else take no account of the ages, spatial distribution, and composition of granitic magmatism in the Gascoyne Province. Most models invoke oblique convergence between the Yilgarn and Pilbara Cratons leading to collision at about 1800 Ma (Myers, 1990; Tyler and Thorne, 1990; Powell and Horwitz, 1994; Evans et al., 2003). The majority regard the Minnie Creek batholith (Fig. 2) in the central part of the Gascoyne Province as having stitched this suture, although Myers (1990) suggested that the batholith instead reflects subduction of oceanic crust leading up to the collision. In contrast, Krapež (1999) interpreted the Capricorn Orogeny as a strike-slip megashear caused by the convergence and collision of the Yilgarn Craton and Gawler Craton at c. 1740 Ma. Despite their differences, all of these models propose the existence of a long-lived subduction-related arc in the Gascoyne Province, albeit with varying geometries.

However, U–Pb geochronological data, obtained by secondary ion mass spectrometry (SIMS) using the sensitive high-resolution ion microprobe (SHRIMP), plus whole-rock Nd isotopic data for granites in the Gascoyne Province, are not consistent with subduction leading up to the Capricorn Orogeny, instead suggesting the intracontinental reworking of a collision zone that formed about 150 m.y. earlier. This reworking was probably a reaction to plate collisions elsewhere in the supercontinent of which the West Australian Craton was part, or was developed well inboard of an active convergent margin.

Previous work on the Capricorn Orogeny

Vertical tectonics

Daniels (1975) interpreted the Gascoyne Province as an ensialic trough, largely based on the apparent continuity of the Wyloo Group into the province from the northern margin of the Capricorn Orogen. Indicative of the poor state of knowledge of the Gascoyne Province at the time, this interpretation was based on '…photogeological interpretation of parts of the Glenburgh and Mount Phillips Sheet areas, [and] several critical traverses across the area...' (Daniels, 1975, p. 107). Horwitz and Smith (1978, p. 300) could find '…no reason why the Pilbara and Yilgarn Blocks could not have been continuous and part of a single unit and a tectonic plate', and consequently preferred an intracratonic interpretation for the Capricorn Orogen. A number of important differences between the Archean Yilgarn and Pilbara Cratons were subsequently recognized (Gee, 1979; Williams, 1986; Tyler and Thorne,

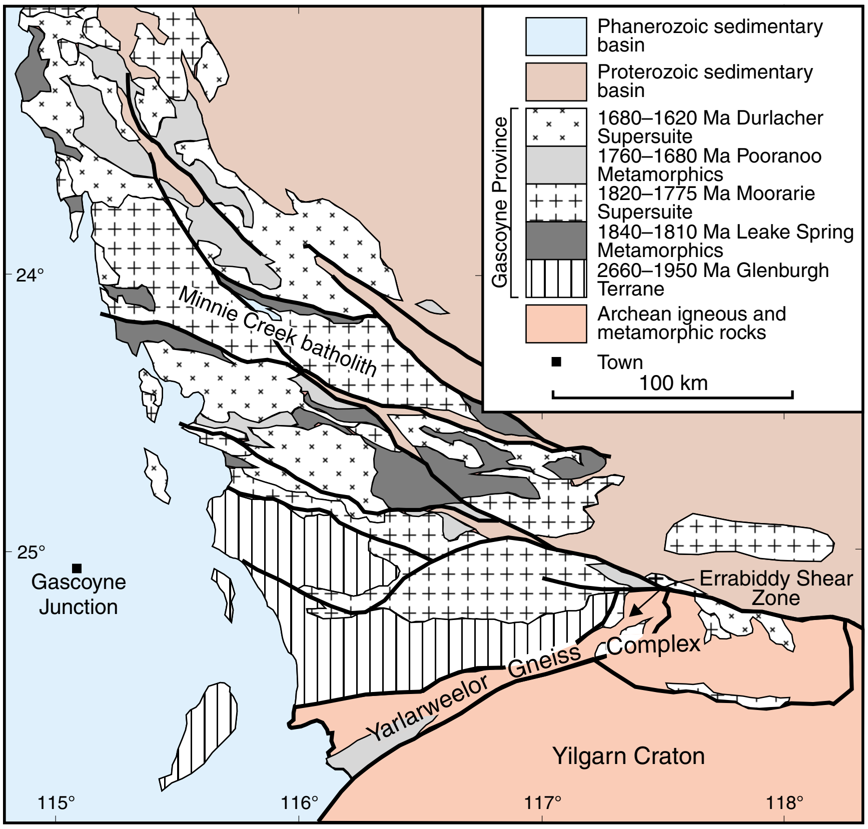

Figure 2. Simplified solid geology map of the Gascoyne Province.

1990); however, Gee (1979) and Williams (1986) both retained interpretations of the Capricorn Orogen as an ensialic mobile belt with limited horizontal movements of crustal elements. Gee (1979) was strongly influenced by lithological correlations between greywackes across the orogen in the Ashburton Fold Belt, Gascoyne Province and 'Glengarry Sub-Basin' (Bryah and Padbury Basins), and by the presence of Archean gneissic basement beneath the Gascoyne Province. Little other supporting evidence was presented for an ensialic model: Gee (1979, p. 364) stated that 'Primarily because of its ensialic nature no evidence for Phanerozoic-type plate tectonics in the Capricorn Orogen is recognised.'

In his synthesis of the first edition mapping of the Gascoyne Province by the Geological Survey of Western Australia (GSWA), Williams (1986) interpreted the Capricorn Orogen as an ensialic mobile belt on the basis of '…the continentally-derived nature of the geosynclinal sediments; the lack of any large ultramafic or mafic sequences; the voluminous, crustal-derived, or crustal-reworked granitoids; and the structural-tectonic style' (p. 77). He also pointed to the lack of andesitic volcanic rocks and volcaniclastic rocks in the 'Morrissey Metamorphic Suite', and to paleomagnetic data (McElhinny and Embleton, 1976) that suggested that the Yilgarn and Pilbara Cratons have not moved substantially relative to each other since about 2500 Ma. Williams (1986) favoured a model for the Capricorn Orogeny in which vertical tectonics dominated, driven by granitic diapirs and rising gneiss domes. Curiously though, he regarded the Yilgarn and Pilbara Cratons sufficiently different '…to make it unlikely that the northern Yilgarn Block ever adjoined the southern Pilbara Craton' (Williams, 1986, p. 71). How this was to be reconciled with his suggestion of the Gascoyne Province as an ensialic mobile belt was not explained.

Second edition mapping of the Gascoyne Province by the GSWA, in conjunction with SHRIMP U–Pb zircon and monazite geochronology, has shown that some 'basement gneiss domes', such as the 'Yinnetharra Gneiss Dome' of Williams (1986), consist of c. 1665 Ma granites that post-date the Capricorn Orogeny and are overprinted by Mesoproterozoic to Neoproterozoic deformation and regional metamorphism. Furthermore, application of modern sense-of-shear criteria (e.g. Hanmer and Passchier, 1991) to shear zones bounding the Minnie Creek batholith has failed to substantiate the diapiric emplacement of the batholith advocated by Williams (1986), and, therefore, removes much of the evidence for vertical tectonics.

Fletcher et al. (1983) presented a transect of Sm–Nd model ages from the northwest Yilgarn Craton (c. 3.5 Ga), across the Errabiddy Shear Zone, and into the central Gascoyne Province (c. 2.0 Ga). They interpreted these data as '…lending support to a model of crustal accretion against an older stabilised craton' (Fletcher et al., 1983, p. 167), although they did not specify a tectonic setting.

Horizontal tectonics

Muhling (1986; 1988) studied the metamorphic and structural evolution of Archean gneisses along the northern edge of the Yilgarn Craton. She identified evidence for Proterozoic crustal shortening followed by significant uplift and granite intrusion, interpreted this in terms of regional north–south compression, and suggested that the Capricorn Orogen marked a zone of collision between the Yilgarn and Pilbara Cratons. This explicit recognition of the role of horizontal tectonics was a clear break from earlier intepretations, which emphasised the importance of vertical tectonics.

Tyler and Thorne (1990) emphasised substantial differences between the Yilgarn and Pilbara Cratons, and pointed out that 'No unequivocal evidence is currently available that *requires* the Pilbara and Yilgarn Cratons to have formed part of a single, continuous craton prior to the development of the Capricorn Orogen. There is also, therefore, no *requirement* for intracratonic, i.e. A-subduction-style, orogeny.' (Tyler and Thorne, 1990, p. 698). These authors also suggested a continent–continent collisional model for the Capricorn Orogeny (Fig. 3) based on the fact that some of the expected features of subduction and collisional tectonics are present in the orogen. Owing to a lack of robust geochronology, Tyler and Thorne (1990) had only loose constraints on the age of the orogeny, of between 2000 Ma and 1600 Ma. The very long orogenic interval was also in part a function of the authors regarding structures in the Ophthalmia Fold Belt and younger Ashburton Fold Belt as products of progressive deformation and sedimentation. They interpreted the Ophthalmia Fold Belt (Fig. 3a), which deformed the lower Wyloo Group (Beasley River Quartzite and overlying Cheela Springs Basalt, the latter subsequently dated at c. 2210 Ma by Martin et al., 1998), as reflecting the initial collision between the two cratons. Continued convergence and subsequent suturing was thought to be marked by structures in the unconformably overlying upper Wyloo Group (Mount McGrath Formation, Duck Creek Dolomite, June Hill Volcanics, and Ashburton Formation in ascending order) in the Ashburton Fold Belt, and by granites of the Gascoyne Province. In other words, the Wyloo Group was thought to represent a single package deposited at c. 2000 Ma, with no significant hiatus separating the formation of structures in the two fold belts (Thorne and Seymour, 1991; Tyler, 1991).

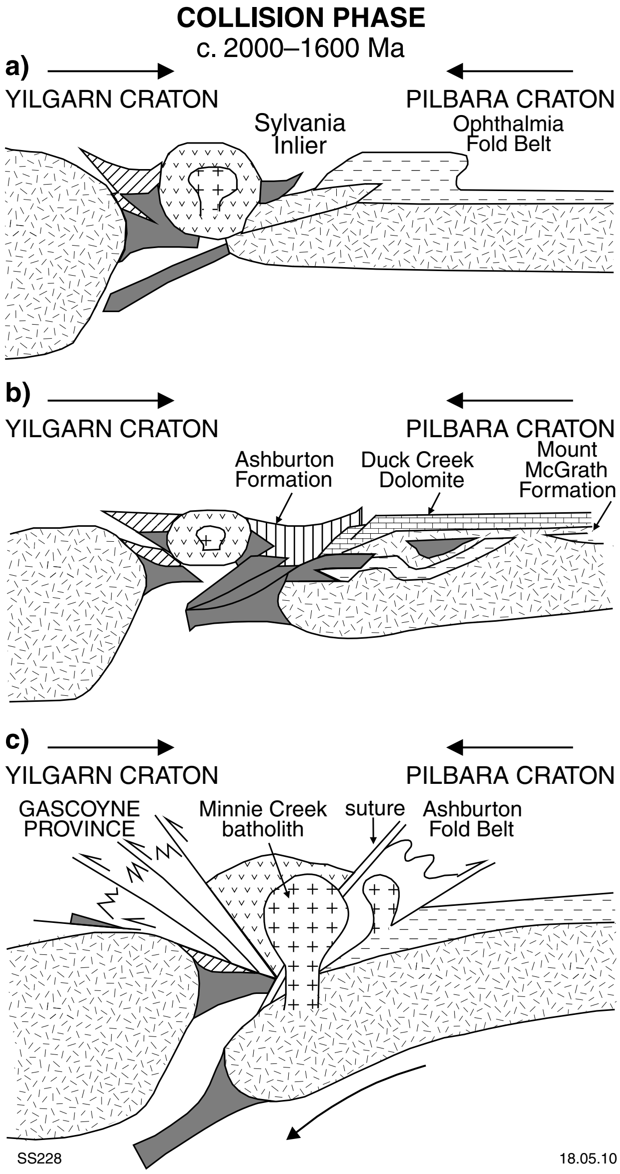

Figure 3. Cartoons illustrating the interpretation of the Capricorn Orogeny by Tyler and Thorne (1990): (a) formation of the Ophthalmia Fold Belt; (b) deposition of the Duck Creek Dolomite and Ashburton Formation; (c) suturing and emplacement of the Minnie Creek batholith. Modified from Tyler and Thorne (1990, fig. 8)

Myers (1989, 1990) inferred widespread thrust-stacking in the Capricorn Orogen, and interpreted this in terms of a plate collisional model. He regarded the Gascoyne Province as being underlain by allochthonous Pilbara Craton in the north, and by allochthonous Yilgarn Craton in the south, with a suture in the northern Gascoyne Province (Fig. 4). The placement of this inferred south-dipping suture was based on the presence of granite batholiths to the south (in particular the Minnie Creek batholith), which were regarded as a product of subduction leading up to collision. However, this model was not without its problems. Firstly, the ages of these granites was never established in relation to the suture. Secondly, in the Gascoyne Province the Ashburton Formation (along the southern edge of the Pilbara Craton) and Leake Spring Metamorphics (formerly the 'Morrissey Metamorphic Suite'), which were regarded as equivalents by Williams (1986), appear to cross the proposed suture.

In contrast with both Thorne and Seymour (1991) and Tyler and Thorne (1990), who regarded both the upper and lower parts of the Wyloo Group as a single foreland basin, Powell and Horwitz (1994), Martin et al. (1998), and Martin et al. (2000) each proposed the presence of two stacked basins. These authors regarded the lower Wyloo Group as having been deposited in a fold-and-thrust belt (McGrath Trough) prior to the deposition of the unconformably overlying upper Wyloo Group in a separate

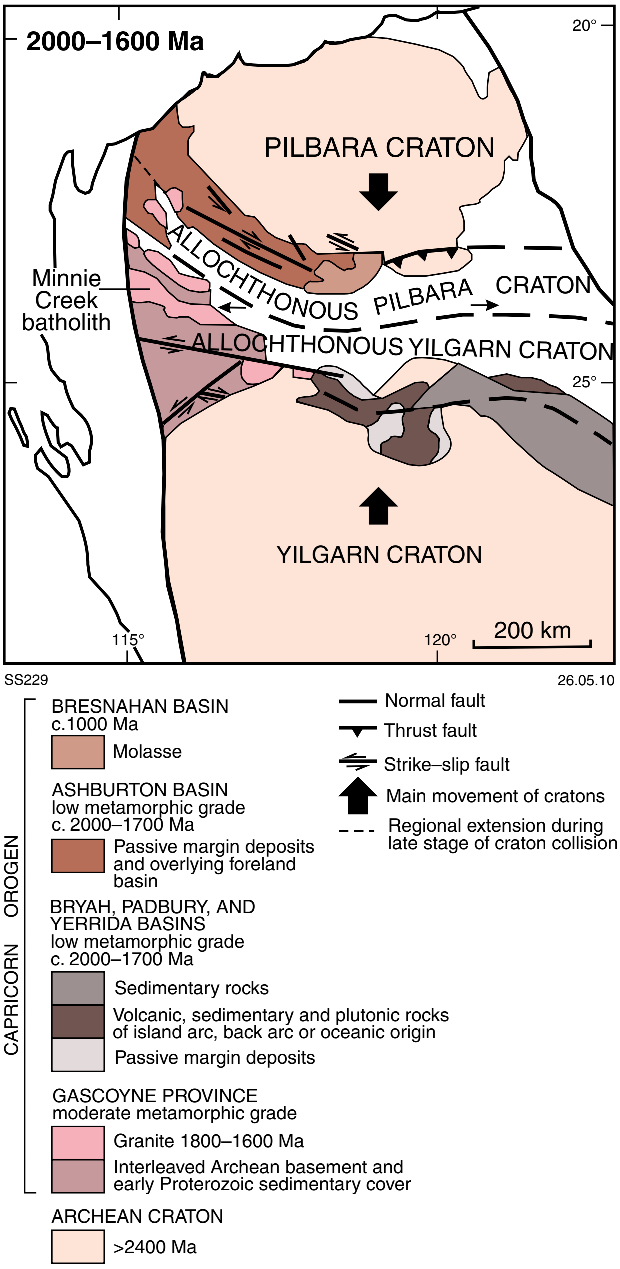

Figure 4. Main features of the Capricorn Orogen interpreted as resulting from collision of the Yilgarn and Pilbara Cratons during the Capricorn Orogeny according to Myers (1990). Note the position of the inferred suture north of the Minnie Creek batholith. Modified from Myers (1990, fig. 2)

basin, which evolved from a rift to a passive margin to a foreland setting. In addition, whereas Tyler and Thorne (1990) considered that structures in the Ophthalmia Fold Belt and Ashburton Fold Belt were both manifestations of the Capricorn Orogeny, Blake and Barley (1992), Powell and Horwitz (1994), Martin et al. (1998), and Martin et al. (2000) considered the structures in the Ophthalmia Fold Belt to belong to an older (pre-2200 Ma) orogeny (Ophthalmian Orogeny) unrelated to collision of the Pilbara and Yilgarn Cratons. This latter interpretation has subsequently been confirmed by SHRIMP U–Pb geochronology, although there is some disagreement as how to interpret the data. Rasmussen et al. (2005) obtained ages of c. 2215 to c. 2145 Ma for monazite and xenotime related to low-grade metamorphism during the Ophthalmian Orogeny, whereas Müller et al. (2005) inferred an age of between c. 2210 and c. 2030 Ma for the Ophthalmian Orogeny by dating baddeleyite from mafic dykes and sills in the lower Wyloo Group, and zircon from a tuffaceous sedimentary rock from the upper Wyloo Group. Regardless of which interpretation is correct, structures formed in the Ophthalmia Fold Belt clearly predate the Capricorn Orogeny.

Krapež (1999) and Dawson et al. (2002) viewed the Capricorn Orogen as a sinistral megashear driven by the collision from the west-northwest of the Yilgarn Craton with the Gawler Craton, which they interpreted as being contiguous with the Pilbara Craton (Fig. 5). These authors proposed that the Gascoyne Province formed a west-northwestly facing Andean-type margin or back-arc setting between c. 1865 Ma, and c. 1815 Ma or c. 1775 Ma. Much of this interpretation was based on the application of sequence stratigraphy to poorly dated metasedimentary successions forced into a theoretical eustatic curve for the Paleoproterozoic, and on rather contentious similarities between the Pilbara and Gawler Cratons.

Evans et al. (2003) presented a revised collisional model for the Capricorn Orogeny (Fig. 6) based on new SHRIMP U–Pb zircon ages for the June Hill Volcanics in the upper Wyloo Group. In this model, they showed accretion of the Glenburgh Terrane to the Yilgarn Craton occurred

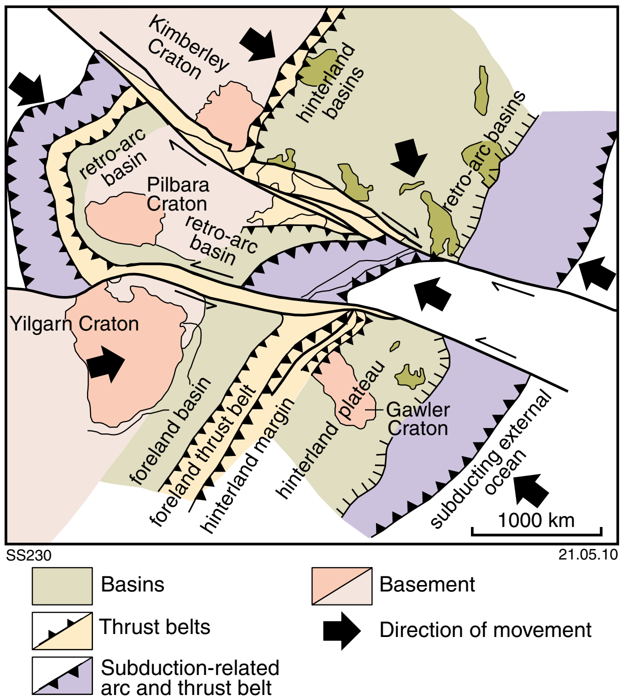

Figure 5. Summary of the Krapež (1999) model for the Capricorn Orogeny. At c. 1800 Ma, the Gascoyne Province is interpreted as a back-arc that is cut by a sinistral transcurrent megashear. Taken from Krapež (1999, fig. 2)

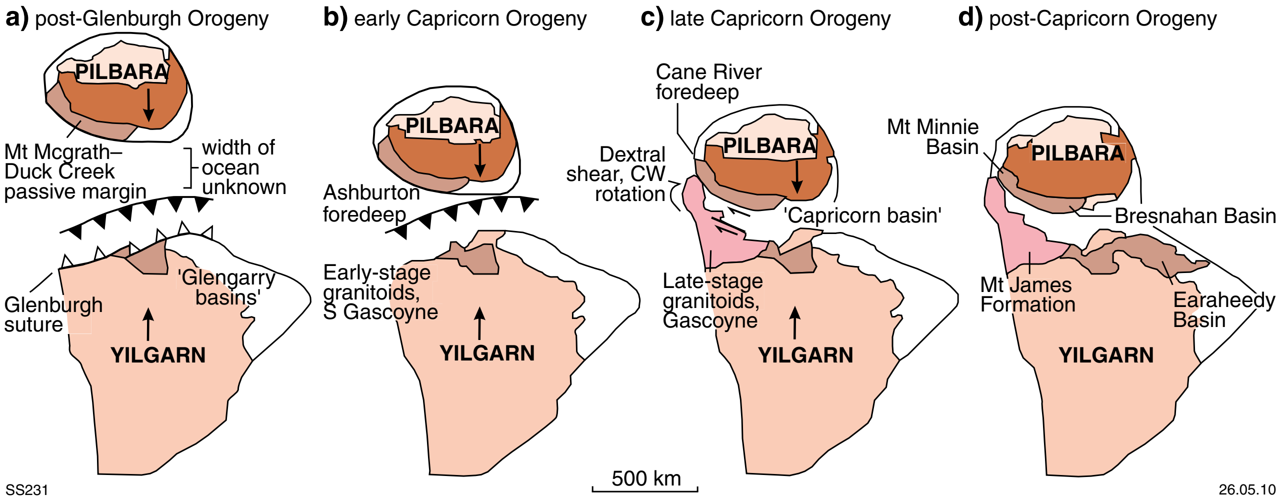

Figure 6. Plate tectonic model of Evans et al. (2003) for the Capricorn Orogeny. Taken from Evans et al. (2003, fig. 5).

at c. 1950 Ma, following Occhipinti et al. (1999), and further suggested that collision was diachronous over about 20–30 m.y., with a westwards younging. Their model incorporated the foredeep interpretation of the upper half of the upper Wyloo Group (June Hill Volcanics and Ashburton Formation) postulated by Thorne and Seymour (1991), but concurred with Powell and Horwitz (1994) and Martin et al. (2000) that the lower part of the upper Wyloo Group (Mount McGrath Formation and overlying Duck Creek Dolomite) represented a foreland basin setting related to the Ophthalmia Fold Belt. Evans et al. (2003) also obtained a date of 1795 ± 8 Ma for a porphyritic monzogranite in the Minnie Creek batholith, which they inferred to stitch the suture between the Pilbara and Yilgarn Cratons.

Despite the widespread application of continent–continent collision models for the Capricorn Orogeny, some objections to a collisional model, as documented by Williams (1986), have yet to be adequately addressed. Additionally, most of these models have been based on poorly dated sedimentary successions, with little consideration for the Gascoyne Province, which is essential to any interpretation of the tectonic setting for the Capricorn Orogeny. These issues regarding interpretation of the Gascoyne Province are discussed in greater detail in the following section on regional geology.

Regional geology

This section briefly discusses the lithotectonic elements and orogenic events that constitute the Capricorn Orogen. The Gascoyne Province is treated in more detail than the craton margins and the various basins, as recent ideas regarding the evolution of the province are as yet unpublished. Orogenic events younger than the Capricorn Orogeny are also discussed, as many of these younger granites, metamorphic assemblages, and structures were previously considered part of the Capricorn Orogeny.

Lithotectonic elements

Craton margins

The Capricorn Orogen encompasses the deformed margins of the Yilgarn and Pilbara Cratons, with the most obvious expression of this being the extensive Paleoproterozoic reworking within the Yarlarweelor Gneiss Complex of the Yilgarn Craton (Fig. 1; Occhipinti et al., 1998; Occhipinti and Myers, 1999; Sheppard and Swager, 1999; Sheppard et al., 2003). Additionally, the Yilgarn Craton was affected by fault and shear zone formation or reactivation during the Paleoproterozoic, possibly during the Capricorn Orogeny, for at least 100 km south of the Errabiddy Shear Zone (Fig. 1; Spaggiari et al., 2008). Farther to the east, the Marymia Inlier (Fig. 1) is a segment of Yilgarn Craton overprinted by a tectonothermal event at c. 1720 Ma (Vielreicher et al., 2002). The southern margin of the Pilbara Craton, mainly the Hamersley Basin and Sylvania Inlier, was deformed during the Ophthalmian and Capricorn Orogenies (Tyler and Thorne, 1990; Tyler, 1991). The Sylvania Inlier was also cut by a network of shear zones, one of which has been dated using $^{40}Ar/^{39}Ar$ on muscovite at c. 1650 Ma (Sheppard et al., 2006).

Paleoproterozoic and Mesoproterozoic basins

The Capricorn Orogen includes numerous siliciclastic sedimentary basins, some of which also contain substantial volumes of volcanic rock. These basins include the Paleoproterozoic Ashburton, Blair, and Bresnahan Basins (Hunter, 1990a; Thorne and Seymour, 1991; Hall et al., 2001; Evans et al., 2003; Sircombe, 2003; Martin et al., 2005) along the northern edge of the orogen; the Mesoproterozoic Edmund and Collier Basins in the centre of the orogen (Martin and Thorne, 2004; Martin et al., 2008); and the Paleoproterozoic Padbury, Bryah, Yerrida, and Earaheedy Basins (Pirajno and Adamides, 2000; Pirajno et al., 2000, 2004; Halilovic et al., 2004) along the southern margin of the orogen (Fig. 1). All of these basins are metamorphosed only at low grade, but have been deformed to varying degrees. The timing of metamorphism and deformation in these basins is very poorly known owing to the absence of bracketing relationships with intrusive felsic igneous rocks.

Gascoyne Province

The Gascoyne Province is divided into several tectonometamorphic zones, largely bounded by east-southeasterly trending faults or shear zones (Fig. 7). Although each zone is characterized by a distinctive and episodic history of deformation, metamorphism, and granitic magmatism (Fig. 8), exotic terranes are juxtaposed only along the Errabiddy Shear Zone.

Glenburgh Terrane

The oldest crust in the Gascoyne Province is the Glenburgh Terrane, which is exposed in the Paradise, Mooloo, Mutherbukin, and Limejuice Zones in the southern part of the province (Fig. 9a). The Glenburgh Terrane comprises granitic rocks with igneous crystallization ages of 2660–2430 Ma (Halfway Gneiss); psammitic and pelitic rocks of the Moogie Metamorphics, deposited between c. 2240 Ma and c. 2125 Ma; and a 2005–1970 Ma Andean-type batholith (Dalgaringa Supersuite; Occhipinti and Sheppard, 2001; Johnson et al., 2010). The bulk of the Halfway Gneiss is younger than c. 2610 Ma, and has no counterparts in either the Yilgarn or Pilbara Cratons. The Halfway Gneiss is unconformably overlain by the Moogie Metamorphics, which include relict pelitic migmatites, now pervasively retrogressed to chloritoid-bearing schists (Johnson et al., 2010). The Halfway Gneiss and Moogie Metamorphics were intruded by igneous rocks of the Dalgaringa Supersuite, which range from tonalite to syenogranite. This supersuite is not present in the adjacent Yilgarn Craton.

The 2005–1970 Ma Dalgaringa Supersuite is interpreted to have formed above a northwestward-dipping subduction zone along the southern margin of the Glenburgh Terrane (Sheppard et al., 2004). Collision with the passive margin of the Yilgarn Craton was marked by medium- to high-grade metamorphism during the later stages of the Glenburgh Orogeny between 1965 and 1950 Ma, and was accompanied by intrusion of granites of the Bertibubba Supersuite across the suture (Occhipinti, 2004; Occhipinti et al., 2004; Johnson et al., 2010).

The results of a magnetotelluric survey across the western Capricorn Orogen (Selway, 2008; Selway et al., 2009), suggest that, although not exposed at the surface in the northern part of the province (the Mangaroon and Boora Boora Zones; Figs 7, 9a), the Glenburgh Terrane forms basement to the whole Gascoyne Province. The survey also shows that the electrical character of the Glenburgh Terrane is unlike that of either bounding Archean craton, with the margins of the terrane marked by the Errabiddy Shear Zone to the south and the Talga Fault to the north (Fig. 9). These results are consistent with the Glenburgh Terrane being exotic to both bounding cratons, and, therefore, with the presence of two sutures in the Capricorn Orogen.

Camel Hills Metamorphics

The Camel Hills Metamorphics, which is confined to the Errabiddy Shear Zone (Figs 8, 9), comprises pelitic gneiss and schist (Quartpot Pelite); calc-silicate gneiss and schist with quartzite (Petter Calc-silicate); and undivided amphibolite and metamorphosed banded iron-formation. The Quartpot Pelite was deposited between 2000 Ma and 1955 Ma as a fore-arc deposit to the Dalgaringa continental margin arc, whereas the Petter Calc-silicate was deposited sometime between 2610 Ma and 1965 Ma on the passive margin of the Yilgarn Craton (Johnson et al., 2010).

Leake Spring Metamorphics

The Leake Spring Metamorphics comprises metasedimentary and minor meta-igneous rocks that were deposited after the Glenburgh Orogeny and prior to the Capricorn Orogeny. This package was formerly known as the ‘Morrissey Metamorphics’ (e.g. Martin et al., 2007, 2008; Sheppard et al., 2008a) or the ‘Morrissey Metamorphic Suite’ (Williams et al., 1978, 1983; Williams, 1986). Based on field mapping and geochronology, most of what was thought to be Morrissey Metamorphics, including the type areas for the original ‘Morrissey Metamorphic Suite’ (Williams et al., 1978), has been reassigned to either the older Moogie Metamorphics and Camel Hills Metamorphics, or to the younger Pooranoo Metamorphics (e.g. Occhipinti and Sheppard, 2001; Martin et al., 2005, 2007). Consequently, the names ‘Morrissey Metamorphic Suite’ and ‘Morrissey Metamorphics’ are no longer valid.

The Leake Spring Metamorphics are dominated by pelitic and psammitic schists, with minor amounts of quartzite and calc-silicate rock. Amphibolite lenses and layers up to a few metres thick are a minor component of the succession, although amphibolite becomes abundant toward the southeastern end of the Limejuice Zone. The amphibolites are mainly foliated, fine grained, and homogeneous, but are locally finely interlayered with quartzite at their margins, with some layers consisting of mixtures of quartzite and mafic rock (Fig. 10). These textures imply that the mafic rocks were deposited at the same time as the protoliths to the quartzite, and were originally volcanic or volcaniclastic.

The precursor siliciclastic sediments (and minor mafic volcanic rock) of the Leake Spring Metamorphics were deposited across the northern two-thirds of the Gascoyne Province between c. 1840 Ma and c. 1810 Ma (Fig. 9b; Sheppard et al., 2010). The maximum depositional age of the Leake Spring Metamorphics is provided by SHRIMP U–Pb dating of detrital zircons in only three samples; two of these samples are from a single locality in the Nardoo Hills area (Mutherbukin Zone), whereas the third is from the southern edge of the Limejuice Zone. Varvell (2001) sampled a 100 m thick band of semipelitic schist within a distinctive unit of pelitic schist about 2 km wide and characterized by abundant large staurolite and garnet porphyroblasts. Thirty-seven of 63 analyzed zircons in this sample (CV065) yielded a weighted mean $^{207}Pb^{*}/^{206}Pb^{*}$ date of 1840 ± 4 Ma, interpreted as the maximum depositional age of the precursor sediment

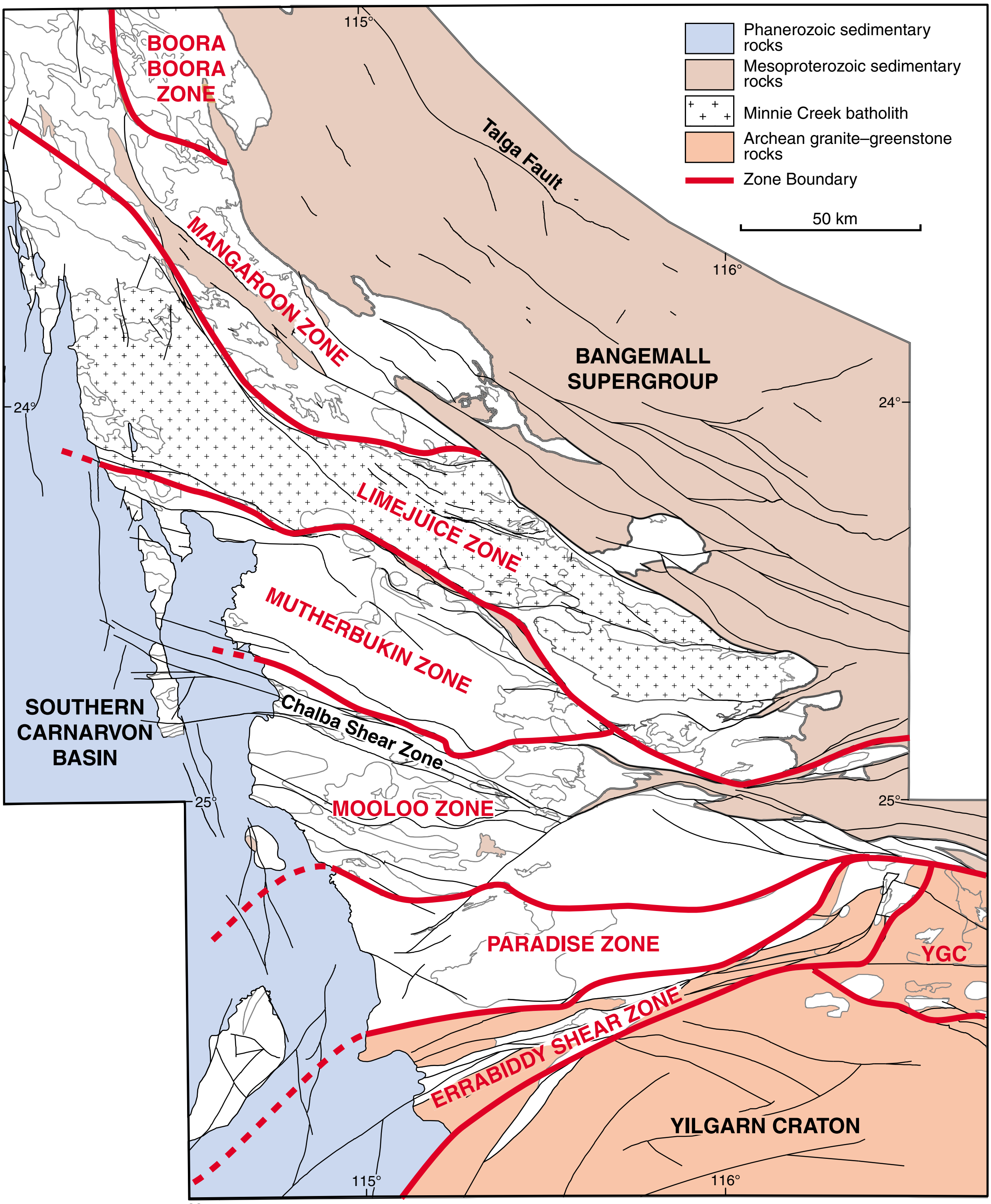

Figure 7. Structural–metamorphic zones of the Gascoyne Province. YGC — Yarlarweelor Gneiss Complex.

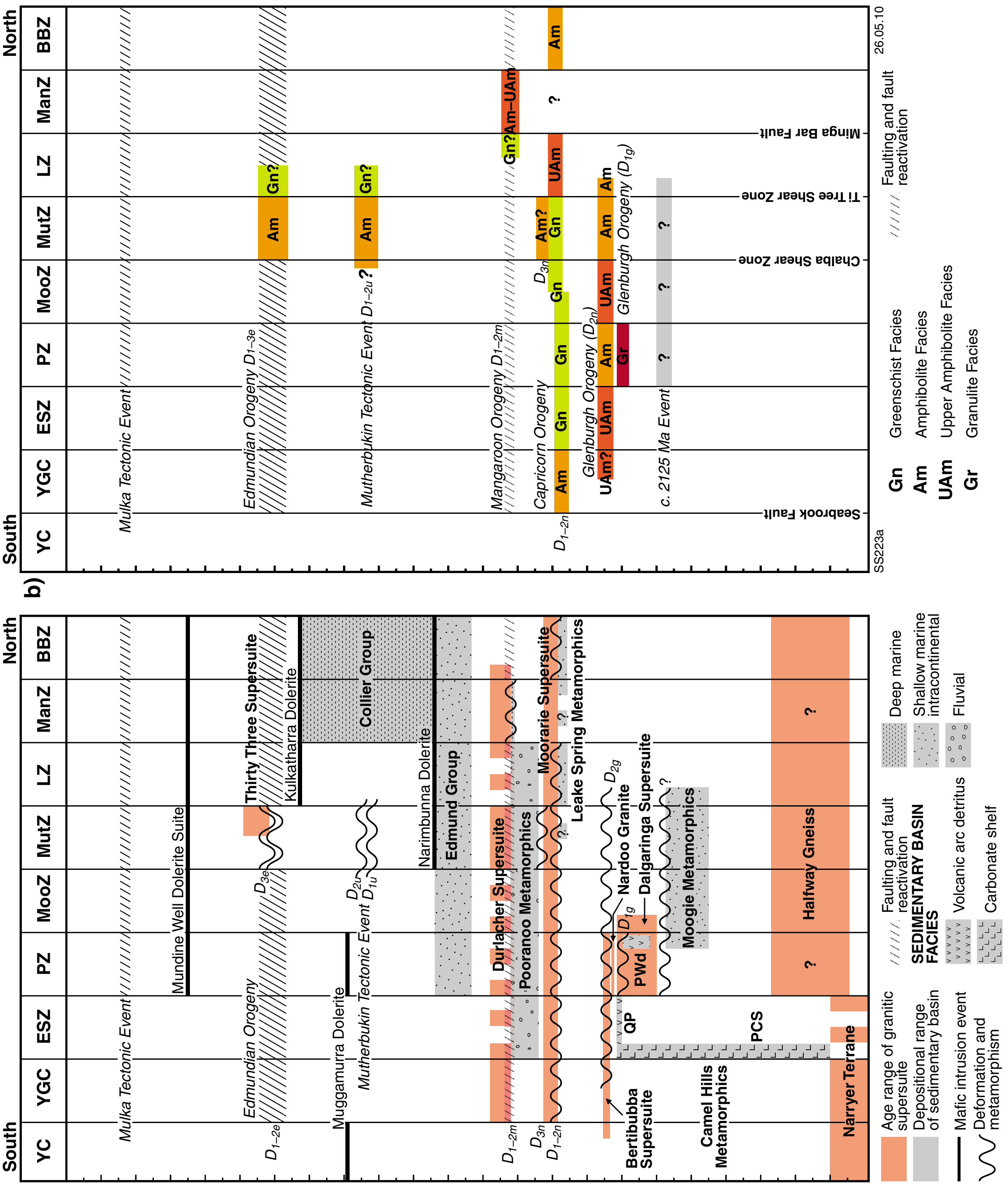

Figure 8. **Simplified time–space plot for the Gascoyne Province, showing lithostratigraphic units (a), and peak metamorphism (b). Key to zone divisions: YC — Yilgarn Craton; YGC — Yarlarweelor Gneiss Complex; ESZ — Errabiddy Shear Zone; PZ — Paradise Zone; MooZ — Mooloo Zone; MutZ — Mutherbukin Zone; LZ — Limejuice Zone; ManZ — Mangaroon Zone; BBZ — Boora Boora Zone. Other abbreviations: QP — Quartpot Pelite; PWd — Paradise Well diatexite; PCS —Petter Calc-silicate.**

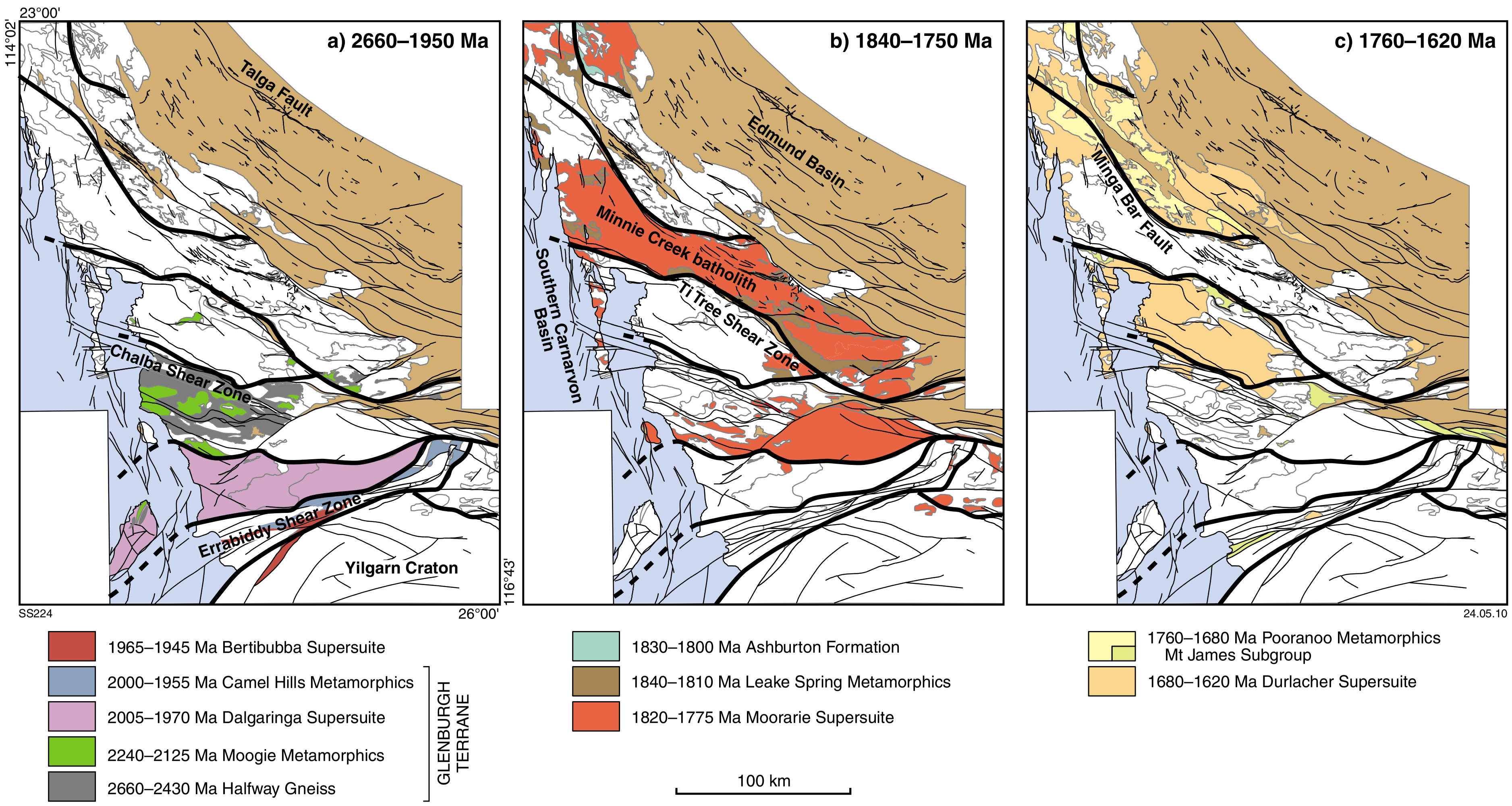

Figure 9. Distribution of the main supersuites and metasedimentary rock packages in the Gascoyne Province: a) 2660–1950 Ma; b) 1840–1750 Ma; c) 1760–1620 Ma.

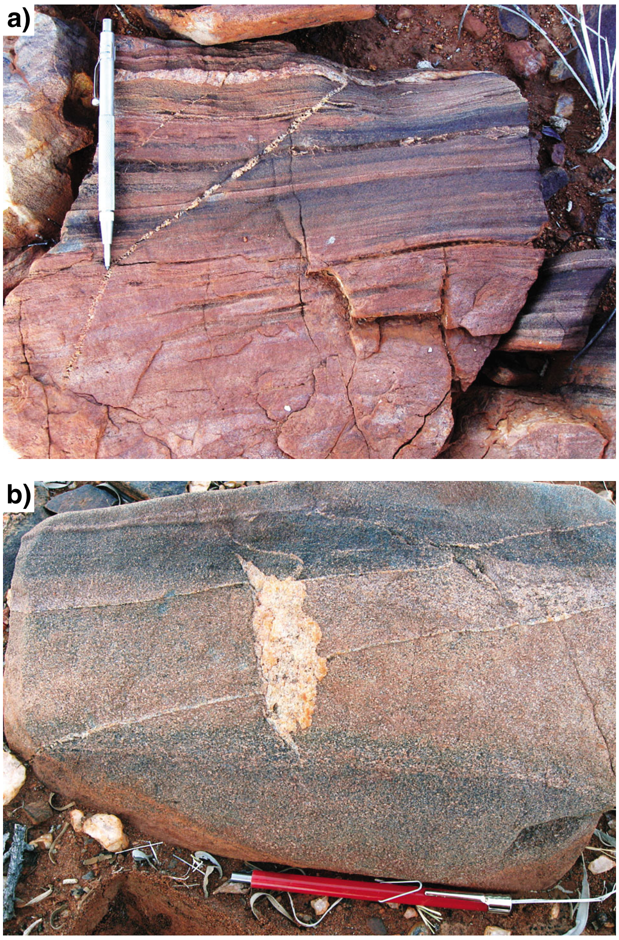

Figure 10. Metamorphosed mafic volcanic or volcaniclastic rocks, and quartzite on northwestern Pink Hills: (a) finely interlayered amphibolite and quartzite, with some layers of quartz amphibolite or quartz–amphibole–plagioclase rock, consistent with synsedimentary reworking of mafic volcanic rock; (b) layer of quartz–amphibole–plagioclase rock with amphibolite above and quartz amphibolite below. Note the gradational contact between the quartz–amphibole–plagioclase rock and the quartz amphibolite.

(Varvell, 2001). More material was recovered from this site during remapping, in order to better define the maximum depositional age due to a lack of zircon crystals from the surrounding pelitic units. In this repeat sample (GSWA 190668), 36 of 46 concordant analyses yielded a weighted mean $^{207}Pb^{*}/^{206}Pb^{*}$ date of 1842 ± 5 Ma, interpreted as the maximum depositional age (Wingate et al., in prep.) The only other constraint on the maximum depositional age of the Leake Spring Metamorphics is provided by five analyses of five detrital zircons from a sample of psammitic schist (GSWA 180935) intruded by the Minnie Creek batholith at Leake Spring (Mount Augustus), which yielded a weighted mean $^{207}Pb^{*}/^{206}Pb^{*}$ date of 1961 ± 7 Ma (Kirkland et al., 2009).

At the northern end of the Gascoyne Province in the Boora Boora Zone (Fig. 9b), metasedimentary rocks of the Leake Spring Metamorphics appear to grade northwards and northeastwards into lower grade metasedimentary, and mafic and felsic metavolcanic rocks of the upper Wyloo Group, particularly the Ashburton Formation (Daniels, 1968; Williams, 1986; Myers, 1990). Although minor felsic volcanic and volcaniclastic rocks are present in the Ashburton Formation (Thorne and Seymour, 1991; Sircombe, 2003; Martin et al., 2005), these felsic volcanic rocks are yet to be identified in the Leake Spring Metamorphics. The Ashburton Formation comprises two distinct parts — a lower package of deep marine turbidites derived from a source to the southeast of the basin, overlain by turbidites deposited in prograding fans derived largely from an uplifted Gascoyne Province to the southwest (Thorne and Seymour, 1991) — and was deposited between c. 1830 Ma and c. 1800 Ma (Hall et al., 2001; Sircombe, 2002, 2003; Evans et al., 2003; Nelson, 2004).

Moorarie Supersuite

The Moorarie Supersuite is dominated by granodiorite, monzogranite, and their metamorphosed equivalents (Fig. 9b). The Minnie Creek batholith, which is part of this supersuite, is notable for the presence of tonalite, quartz diorite, and some gabbro. Forty-five samples from the supersuite, collected across the western Capricorn Orogen, have been dated at between c. 1820 Ma and c. 1775 Ma.

In the Boora Boora Zone, granites of the supersuite consist of foliated and gneissic metagranodiorite, with subordinate metatonalite and metamonzogranite. Low-strain areas show the rocks contain primary biotite, or biotite and minor muscovite; there is no evidence for igneous hornblende. In the Mangaroon Zone, exposures of the Moorare Supersuite are restricted to small inliers of biotite metamonzogranite and metagranodiorite gneiss (Gooche Gneiss; Martin et al., 2005).

In the Limejuice Zone, the Minnie Creek batholith comprises medium- to coarse-grained, biotite-bearing, monzogranite, granodiorite and tonalite plutons, with minor quartz diorite and syenogranite. The granodiorites and tonalites are typically equigranular to sparsely porphyritic, whereas many of the monzogranites contain up to 30% tabular or round microcline or microperthite phenocrysts up to 6 cm in length. Igneous muscovite is a very minor component of some granites. Hornblende-bearing granites are rare, being mainly restricted to the quartz diorites, where hornblende is intergrown with biotite. Some tonalites contain intergrowths of recrystallized biotite and epidote, with the shapes of some intergrowths suggestive of original amphibole. Many of the granites are massive or weakly foliated, with well-preserved igneous structures and textures. However, even these apparently undeformed rocks have undergone at least low-grade static metamorphism. Along the margins of the Minnie Creek batholith, and at the southeastern end of the batholith, many of the granites are foliated or gneissic.

Rocks of basic and intermediate composition, including fragments of layered mafic–ultramafic intrusions, are scattered as rafts throughout the batholith. Some of these basic rocks have associated net-vein structures indicative of mingling between basic and acid magmas. Round to elliptical inclusions with basic to intermediate chemistry are widespread in the granites, although not abundant.

In the Mutherbukin Zone, Moorarie Supersuite rocks comprise foliated to gneissic biotite(–muscovite) metamonzogranite, metagranodiorite, and metatonalite. Farther south in the Mooloo Zone, the Moorarie Supersuite forms several large, sheet-like plutons and associated dykes striking in an easterly direction. These are typically massive and contain medium-grained, equigranular biotite(–tourmaline–muscovite) monzogranite, biotite(–muscovite) monzogranite, and porphyritic fine- to medium-grained biotite metagranodiorite (Occhipinti and Sheppard, 2001).

In the Yarlarweelor Gneiss Complex, Moorarie Supersuite rocks comprise sheets, dykes, and veins of metamorphosed coarse grained to very coarse grained leucocratic biotite granodiorite, monzogranite, and syenogranite; biotite (–muscovite) pegmatite; and metamorphosed medium-grained, equigranular to moderately porphyritic biotite granodiorite, monzogranite, and leucocratic tonalite (Sheppard et al., 2003). Mafic rocks are absent.

Pooranoo Metamorphics

Siliciclastic sedimentary protoliths to the Pooranoo Metamorphics were deposited across the Gascoyne Province following the Capricorn Orogeny (Fig. 9c). At the base of the package, metamorphosed fluvial conglomerates and sandstones, and shallow-marine sandstones are grouped together as the Mount James Subgroup (Sheppard et al., 2008a, 2010). Formerly the Mount James Formation (Hunter, 1990b), this unit was elevated to subgroup status with the recognition of a coherent stratigraphy, and the identification of two internal formations (Spring Camp Formation and Biddenew Formation; Sheppard et al., 2008a, 2010). In the northern part of the Mutherbukin Zone and in the Mangaroon Zone, these rocks grade upwards into metamorphosed turbiditic sandstones, siltstones, and shales that appear to mark a deepening of the basin to the north. Rare amphibolites and amphibole- and plagioclase-rich calc-silicate rocks represent metamorphosed mafic–ultramafic extrusive or high-level intrusive rocks.

Field relationships, combined with SHRIMP U–Pb zircon dating of detrital zircons from metasedimentary rocks, indicate that the protoliths to the Mount James Subgroup were deposited after c. 1760 Ma. A sample of pelitic gneiss from the Pooranoo Metamorphics in the Mangaroon Zone yields a youngest population, at 1680 ± 13 Ma, of four small euhedral grains or grain fragments (GSWA 169094; Sheppard et al., 2005). In cathodoluminenscence (CL) images, these zircons display complex growth-zoning typical of granitic rocks (see Corfu et al., 2003), and they do not have the low Th/U ratios common in many metamorphic zircons (Rubatto and Gebauer, 2000), suggesting that they are detrital in origin. The euhedral nature of the zircons, combined with their lack of pitted surfaces, may indicate that they were sourced from proximal syndepositional volcanic material. Therefore, at least part of the succession in the Pooranoo Metamorphics may have been deposited shortly before, or synchronous with, the onset of the 1680–1620 Ma Mangaroon Orogeny.

Durlacher Supersuite

The 1680–1620 Ma Durlacher Supersuite was emplaced across the Gascoyne Province during the Mangaroon Orogeny (Fig. 9c). In the Mangaroon Zone, the supersuite comprises peraluminous biotite–muscovite monzogranite, granodiorite, and syenogranite, plus muscovite–tourmaline(–biotite) monzogranite all dated at between c. 1675 Ma and c. 1660 Ma (Martin et al., 2005). These rocks comprise undeformed or weakly deformed large plutons and plugs, with associated dykes and veins. Many of the plutons are elongate in an east-southeasterly direction, parallel to the prevailing structures in the Mangaroon Zone. Large plutons of peraluminous leucocratic muscovite–tourmaline(–biotite) monzogranite as young as c. 1620 Ma also intrude the Boora Boora Zone to the north (Martin et al., 2005).

In the Mutherbukin Zone, the Durlacher Supersuite comprises biotite metamonzogranite, and biotite–muscovite(–tourmaline) metamonzogranite and metasyenogranite. These rocks are typically moderately to strongly foliated, owing to overprinting by the Mutherbukin Tectonic Event and Edmundian Orogeny (Fig. 8), and have igneous crystallization ages of c. 1665 Ma to c. 1650 Ma. In the Yarlarweelor Gneiss Complex, the Durlacher Supersuite comprises two large plutons of the Discretion Granite, the bulk of which was emplaced at c. 1620 Ma (Sheppard and Swager, 1999).

Edmund and Collier Groups

The Mesoproterozoic Edmund and Collier Basins are the youngest elements of the Capricorn Orogen. The Edmund Basin rests unconformably on the Gascoyne Province — including rocks of the Durlacher Supersuite — as well as on various other tectonic units (Martin and Thorne, 2004). The Edmund Basin is in turn unconformably overlain by the Collier Basin. The fine-grained siliciclastic sediments and carbonates of both basins were deposited in response to intracratonic extensional reactivation of the Capricorn Orogen (Martin et al., 2008). The Edmund Basin was intruded by c. 1465 Ma dolerite sills (Wingate, 2002), whereas the Collier Basin (and to a lesser extent, the Edmund Basin) was intruded by c. 1070 Ma dolerite sills and dykes of the Warakurna large igneous province (Wingate et al., 2004; Morris and Pirajno, 2005).

Orogenic events

Ophthalmian Orogeny

Although the Ophthalmia Fold Belt is widely regarded as a northerly verging fold-and-thrust belt, there is little agreement about the timing of deformation and sedimentation, particularly in regard to the age of the Turee Creek Group and overlying lower Wyloo Group. In any case, the age of the Ophthalmian Orogeny is provided by SHRIMP U–Pb dates of c. 2215 Ma to c. 2145 Ma, derived from monazite and xenotime crystals that grew during low-grade metamorphism (Rasmussen et al., 2005), and by SHRIMP U–Pb baddeleyite and zircon dates of c. 2210 and c. 2030 Ma for the mafic sills and volcaniclastic rocks that bracket the orogeny (Müller et al., 2005). The style of deformation is consistent with collision between the Pilbara Craton and a continent to the south (Blake and Barley, 1992), the most likely candidate being the older elements of the Glenburgh Terrane (Johnson et al., 2010).

Glenburgh Orogeny

The Glenburgh Orogeny comprises two discrete tectonothermal episodes (D_{1g} and D_{2g}), both of which have been dated by the growth of metamorphic zircon and monazite. The D_{1g} event, which is known only from the Glenburgh Terrane, is dated at c. 2000 Ma, and is coeval with rocks of the Dalgaringa Supersuite. Metamorphic assemblages related to D_{1g} indicate temperatures of 800–1000°C and 7–10 kb, and are interpreted to reflect construction of a continental arc in the middle to deep crust (Johnson et al., 2010). The D_{2g} event is the earliest common structural element to the Yarlarweelor Gneiss Complex, Errabiddy Shear Zone, and Glenburgh Terrane (Fig. 8), suggesting that this event marks suturing of the combined Glenburgh Terrane – Pilbara Craton with the Yilgarn Craton. The D_{2g} event yielded ages between 1965 Ma and 1950 Ma (Johnson et al., 2010) and coincides with intrusion of the Bertibubba Supersuite. The ubiquitous presence of migmatitic textures and now-retrogressed garnet and sillimanite porphyroblasts within pelitic to semipelitic rocks indicate that peak D_{2g} metamorphism occurred under high temperature and moderate pressure conditions (Johnson et al., 2010).

Capricorn Orogeny

Deformation and metamorphism related to the Capricorn Orogeny is recorded in the Gascoyne Province, in the Ashburton and Hamersley Basins, and Sylvania Inlier along the northern margin of the orogen; and, along the southern margin of the orogen, in the Bryah, Padbury and Yerrida Basins, and the northern edge of the Yilgarn Craton. During the Capricorn Orogeny, sedimentary and volcanic rocks of the Leake Spring Metamorphics in the Gascoyne Province, and upper Wyloo Group rocks in the Ashburton Basin, were deformed, metamorphosed, and intruded by voluminous biotite granites of the 1820–1775 Ma Moorarie Supersuite (Occhipinti et al., 1998, 2001; Krapež and McNaughton, 1999; Evans et al., 2003; Martin et al., 2005). Deformation, metamorphism, and granite intrusion during the Capricorn Orogeny also affected the Errabiddy Shear Zone (Sheppard and Occhipinti, 2000; Occhipinti et al., 2001) and Yarlarweelor Gneiss Complex to the southeast (Occhipinti et al., 1998; Sheppard et al., 2003). Furthermore, the orogeny was probably responsible for cryptic, early high-grade metamorphism in the Mutherbukin Zone, indicated by c. 1770 Ma rims on detrital zircons from a quartzite sample on northwestern YINNETHARRA (GSWA 187403; Wingate et al., 2010). The effects of the Capricorn Orogeny on the region are more fully described in later sections of this report.

Mangaroon Orogeny

During the 1680–1620 Ma Mangaroon Orogeny, the Pooranoo Metamorphics underwent recumbent and upright folding accompanied by low-pressure and high-temperature metamorphism in the Mangaroon Zone (Sheppard et al., 2005). High-grade metamorphism of the Pooranoo Metamorphics produced metatexite and diatexite migmatites, dated at 1677 ± 5 Ma using SHRIMP U–Pb zircon (Sheppard et al., 2005). This was followed by the intrusion of peraluminous biotite(–muscovite) monzogranite, syenogranite, and granodiorite, with crystallization ages for these rocks predominantly between 1675 Ma and 1660 Ma. The Mutherbukin Zone was also extensively intruded by voluminous peraluminous two-mica granites of the Durlacher Supersuite at c. 1665 Ma. Tectonism probably continued as late as c. 1620 Ma, when granites of the Durlacher Supersuite were emplaced into the Yarlarweelor Gneiss Complex.

Mesoproterozoic and Neoproterozoic events

Much of the Mutherbukin Zone has been overprinted by pervasive, amphibolite facies, low-pressure and high-temperature regional metamorphism and deformation. Preliminary SHRIMP U–Pb dating of monazite, xenotime, and titanite in the northernwestern part of the zone suggests that deformation and regional metamorphism during the Mutherbukin Tectonic Event took place between c. 1280 Ma and c. 1250 Ma (Johnson et al., 2009). The dated fabrics can be traced across much of the Mutherbukin Zone, suggesting that the majority of the zone was overprinted during the Mesoproterozoic.

Published SHRIMP U–Pb monazite and xenotime dating in the Nardoo Hills along the northern edge of the Mutherbukin Zone (Martin et al., 2007; Sheppard et al., 2007), combined with preliminary SHRIMP U–Pb zircon dating from the same area (GSWA, unpublished data), indicate that greenschist- to amphibolite-facies regional metamorphism and deformation is related to the 1030–955 Ma Edmundian Orogeny, and not to the Capricorn Orogeny as widely supposed. This tectonism was accompanied by intrusion of leucocratic granite plutons and rare element pegmatites between c. 995 Ma and c. 900 Ma (Johnson et al., 2009). Outside of this zone, the Edmundian Orogeny was responsible for the widespread folding and low-grade metamorphism of the Edmund and Collier Groups (Martin and Thorne, 2004).

At c. 755 Ma, the Gascoyne Province and adjacent tectonic units were intruded by dykes of the Mundine Well Dolerite Suite (Wingate and Giddings, 2000). Within the Gascoyne Province, this dyke suite is cut by a series of prominent east-southeasterly trending shear zones with dextral strike-slip kinematics, formed during the Mulka Tectonic Event. In situ $^{40}Ar/^{39}Ar$ dating of fine-grained white mica on the S-planes of an S–C fabric within the Chalba Shear Zone yielded a likely crystallization age of 570 ± 10 Ma (95% confidence; Bodorkos and Wingate, 2007), which may date the fabric formation and deformation in the shear zone.

The dextral shear zones probably represent an episode of intracontinental reactivation related to other 'pan-African' or 'pan-Gondwana' events (Veevers, 2003).

Problems with existing collisional models for the Capricorn Orogeny

Extensive work has recently been carried out on the Gascoyne Province, including regional mapping and SHRIMP U–Pb zircon geochronology carried out by GSWA; SHRIMP U–Pb monazite and xenotime geochronology by the University of Western Australia and Curtin University of Technology; and a study on the use of magnetotelluric data to examine the deep structure of the western Capricorn Orogen conducted by GSWA and The University of Adelaide. This work has highlighted problems with existing collisional models for the Capricorn Orogeny. These problems, discussed in the following sections, may be summarized as follows:

- There are two Paleoproterozoic sutures in the western Capricorn Orogen not one, following from the recognition of three, rather than two, disparate crustal elements.
- The age and distribution of the Leake Spring Metamorphics appears at odds with suggestions that the Capricorn Orogeny marks a continent–continent collision.
- It is clear that many of the tectonic fabrics and metamorphic mineral assemblages in the Gascoyne Province are unrelated to the Capricorn Orogeny.
- The age patterns of granites in the Gascoyne Province are not necessarily consistent with the Capricorn Orogeny marking a continent–continent collision.
- It has never been established whether the geochemical and isotopic signatures of the Moorarie Supersuite are best explained by a convergent continental margin setting.

Location of sutures and identification of main crustal elements

One of the problems that has plagued interpretations of the Capricorn Orogen to date is uncertainty about the number, location, and age of possible sutures. As a consequence, the number of crustal fragments involved in assembly of the orogen is presently unclear. Collisional models for the orogen have generally assumed the presence of one suture within the Gascoyne Province, along which the Yilgarn and Pilbara Cratons were juxtaposed. However, recent studies using a range of datasets indicate that there are two sutures, and three main crustal fragments, in the orogen: the Glenburgh Terrane, the Yilgarn Craton, and the Pilbara Craton. The Glenburgh Terrane and Pilbara Craton probably amalgamated during the 2215–2145 Ma Ophthalmian Orogeny, and this combined entity collided with the Yilgarn Craton during the 2005–1950 Ma Glenburgh Orogeny (Johnson et al., 2010).

Up until 1994, the Paradise and Mooloo zones of the Glenburgh Terrane were thought to comprise reworked Archean rocks of the Yilgarn Craton (Williams, 1986; Myers, 1990), although subsequent field mapping and geochronology have indicated that before c. 1950 Ma the southern Gascoyne Province had a geological history distinct from that of the Errabiddy Shear Zone and Yilgarn Craton (Nutman and Kinny, 1994; Kinny et al., 2004; Occhipinti et al., 2004; Fig. 8). Metagranitic rocks of the Halfway Gneiss, which forms the basement to the Glenburgh Terrane in these zones, have igneous crystallization ages between c. 2660 Ma and c. 2430 Ma (Occhipinti et al., 2001; Kinny et al., 2004; GSWA, unpublished data), whereas the foliated to gneissic granites of the Dalgaringa Supersuite (which has suprasubduction geochemical affinities) in the Paradise Zone, have igneous crystallization ages between c. 2005 Ma and c. 1970 Ma (Occhipinti et al., 2004; Sheppard et al., 2004). Rocks with crystallization ages younger than c. 2610 Ma have no counterparts in the northwestern Yilgarn Craton, indicating that the Errabiddy Shear Zone is a Paleoproterozoic suture (Occhipinti et al., 2004; Sheppard et al., 2004; Johnson et al., 2010). This conclusion is supported by the magnetotelluric data, which show a clear contrast between the Glenburgh Terrane and Yilgarn Craton across the Errabiddy Shear Zone (Selway, 2008; Selway et al., 2009). At 1965–1950 Ma, the southern edge of the Glenburgh Terrane, the Errabiddy Shear Zone, and the northern edge of the Yilgarn Craton were all deformed, metamorphosed, and intruded by biotite–muscovite granites of the Bertibubba Supersuite. This reflects collision and suturing of the Pilbara Craton – Glenburgh Terrane with the Yilgarn Craton, following the northwestwardly directed subduction of oceanic crust beneath the Glenburgh Terrane (Johnson et al., 2010).

Magnetotelluric data also show a clear distinction between the Glenburgh Terrane and Pilbara Craton across a vertical or steeply dipping structure roughly coincident with the Talga Fault (Selway et al., 2009). This structure appears to be blanketed by the Ashburton Formation and Leake Spring Metamorphics, indicating that the suture is older than c. 1820 Ma. There is no evidence of the Glenburgh Orogeny in the northern part of the Capricorn Orogen, but there is a northerly verging fold-and-thrust belt related to the 2215–2145 Ma Ophthalmian Orogeny (Blake and Barley, 1992; Powell and Horwitz, 1994; Martin et al., 1998, 2000; Rasmussen et al., 2005). Blake and Barley (1992) noted that the style of deformation in this belt was consistent with collision between the Pilbara Craton and a craton to the south. The presence of two sutures indicates that the evolution of the Capricorn Orogen did not involve a single collision between the Yilgarn and Pilbara Cratons, but two collisions between three pieces of crust.

The exotic nature of the Glenburgh Terrane relative to the Pilbara and Yilgarn Cratons is also supported by a P-wave tomography study in northwestern Australia (Abdulah, 2007). Maps of P-wave seismic perturbation at 35 km depth show a wedge-shaped region of slow P-wave velocities corresponding to the Glenburgh Terrane, with sharp boundaries to regions of higher P-wave velocities corresponding to the Pilbara and Yilgarn Cratons (Abdulah, 2007). The boundaries between the Glenburgh Terrane and its bounding cratons agree with the locations identified from surface studies and the magnetotelluric survey.

Age, distribution, and composition of the Leake Spring Metamorphics

Age and distribution

Protoliths of the Leake Spring Metamorphics were deposited after c. 1840 Ma, the age of the youngest detrital zircon population in the two samples recorded from the unit (Varvell, 2001; Wingate et al., in prep.), and prior to c. 1810 Ma, the age of the oldest granites in the cross-cutting Moorarie Supersuite. The unit outcrops within the Mutherbukin and Limejuice zones, and farther to the north in the Boora Boora Zone (Fig. 9b). The Leake Spring Metamorphics is not recognized in the intervening Mangaroon Zone, primarily due to deposition of the Pooranoo Metamorphics at c. 1680 Ma, and intense reworking at upper amphibolite facies grade during the 1680–1620 Ma Mangaroon Orogeny. Therefore this unit, which should pre-date collision in the models of Tyler and Thorne (1990), Myers (1990), and Evans et al. (2003), appears to continue across the proposed suture located in the northern Gascoyne Province (along or just north of the Minnie Creek batholith).

At the northern end of the Boora Boora Zone, the Leake Spring Metamorphics have been interpreted to grade into low-grade metasedimentary rocks of the Ashburton Formation (Williams, 1986). This observation is supported by the fact that both units are intruded by granites of the 1820–1775 Ma Moorarie Supersuite, and both share a common deformation history (Martin et al., 2005).

The stratigraphy and depositional age of the Ashburton Formation is presently unclear, in part owing to structural complexity and a lack of marker horizons (Martin et al., 2005). The lower Ashburton Formation was deposited in a deep marine basin with detritus derived from the southeast, and is overlain by a series of northward prograding fans with sediments sourced from the Gascoyne Province (Thorne and Seymour, 1991). The lower part of the Ashburton Formation was probably deposited at c. 1830 Ma (Sircombe, 2003), approximately coeval with deposition of precursors to the Leake Spring Metamorphics (c. 1840 to c. 1810 Ma). The upper part of the Ashburton Formation was deposited after c. 1806 Ma (Martin et al., 2005) and before 1786 ± 5 Ma, the igneous crystallization age of the cross-cutting Boolaloo Granodiorite (Krapež and McNaughton, 1999); therefore, the upper Ashburton Formation appears to be younger than the Leake Spring Metamorphics. Overall, the age dating does not prove, but is consistent with, the lower Ashburton Formation being equivalent to the Leake Spring Metamorphics. In the absence of evidence to the contrary, these units are correlated across the boundary between the Glenburgh Terrane and Pilbara Craton. Therefore, it is considered that protoliths to the Leake Spring Metamorphics were deposited across the suture between the Glenburgh Terrane and Pilbara Craton, indicating that the two crustal elements were juxtaposed prior to the 1820–1770 Ma Capricorn Orogeny.

Composition of the metavolcanic rocks

Volcanic rocks are present in both the Ashburton Formation and June Hill Volcanics (Evans et al., 2003; Sircombe, 2003), with minor volumes of mafic metavolcanic rocks identified in the Leake Spring Metamorphics in the southeastern part of the Limejuice Zone. However, as all these units are younger than either of the identified sutures, there is no evidence for a magmatic arc in the Gascoyne Province leading up to the Capricorn Orogeny. The only identified arc in the province is the 2005–1970 Ma Dalgaringa Supersuite and unexposed older arc components in the Camel Hills Metamorphics (Fig. 8; Johnson et al., 2010) preceding the Glenburgh Orogeny. Nevertheless, it is worth briefly considering the composition of the metamorphosed mafic volcanic rocks in the Leake Spring Metamorphics for evidence of a subduction-related origin, as mantle-derived rocks are more likely to be a sensitive recorder of subduction process than crustally derived rocks (i.e. granites).

Whole-rock analyses have been obtained from 12 metamorphosed mafic volcanic or subvolcanic rocks in the Leake Spring Metamorphics. All the rocks are now amphibolites, and range from weakly to strongly foliated. As the rocks have been metamorphosed and deformed, some comment is warranted on the possibility of major and trace element mobility in the samples. There is no correlation between loss on ignition (LOI) and any of the element ratios (Fig. 11a–c), and the analyses all form tight coherent groupings on normalized multi-element diagrams ('spider diagrams'; Fig. 11d–f), with the exception of the mobile large ion lithophile elements (LILE) Cs, Rb, K, Ba, Pb, Sr (not shown), and to some extent, Th. For this reason, the LILE have not been used.

Analysed samples from the Leake Spring Metamorphics are basaltic (<52 wt% SiO_2 on an anhydrous basis) with slightly to moderately evolved compositions, as indicated by Mg#s ($100 \times Mg^{2+})/(Mg^{2+}+Fe^{T}$) of 43–52. All the samples have flat rare earth element (REE) patterns when normalized to chondrite (Fig. 11d). The amphibolites have Ce/Yb values of 3.3–6.9, which are at the lower end of the range for mafic magmas from modern-day destructive plate margins (Ce/Yb mostly ~4–40+; Hawkesworth et al., 1991, p. 579), and more comparable with back-arc basin basalts (e.g. from the Central Lau Spreading Centre; Pearce et al., 1995). None of the samples from the Leake Spring Metamorphics has a significant Eu anomaly (Eu/Eu* = 0.84 – 1.05), and the middle and heavy REE abundances are similar to mid-ocean ridge basalt (MORB). The amphibolites have La/Nb values of 1.2 – 2.4, which lie between those for arc basalts (~3.6; Hawkesworth et al., 1991) and for normal (N-) MORB and enriched (E-) MORB (~1.07 and ~0.80; Sun and McDonough, 1989), but are most similar to back-arc basin basalts (~1.5 – 3.0; Pearce et al., 1995). The amphibolites from the Leake Spring Metamorphics have very weak or no Nb, Ta, and TiO_2 depletions relative to N-MORB, unlike arc-related

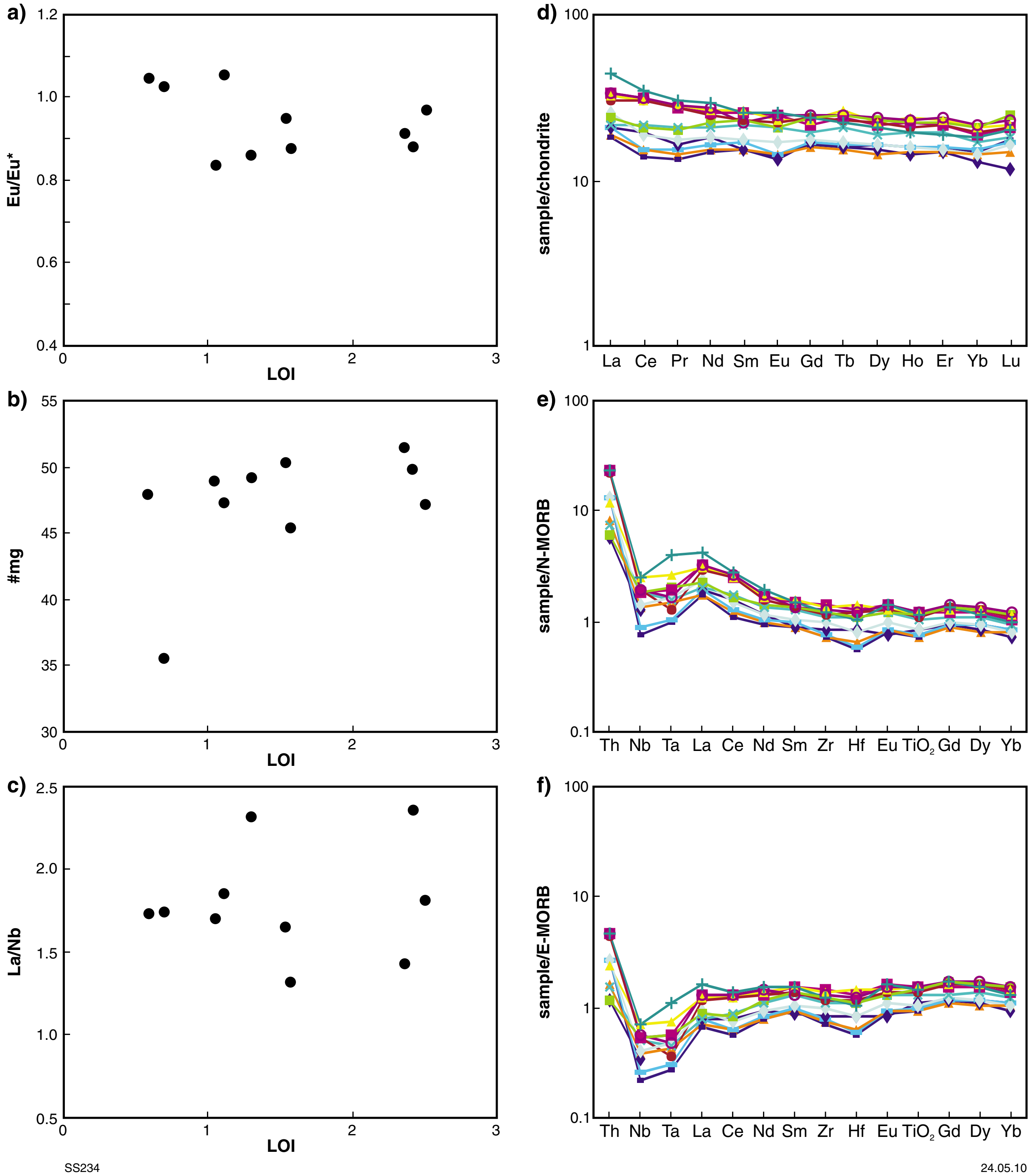

Figure 11. Whole-rock chemistry of amphibolites from the Leake Spring Metamorphics: plots of loss on ignition (LOI) versus Eu/Eu* (a), Mg# (b), and La/Nb (c); (d) chondrite-normalized REE plot; multi-element plots normalized to N-MORB (e) and E-MORB (f). Normalizing values for (d), (e) and (f) from Sun and McDonough (1989).

basalts (e.g. Wilson, 1989) and, to a lesser extent, back-arc basin basalts (Gamble et al., 1995; Pearce et al., 1995). In fact, Nb and Ta are weakly enriched relative to N-MORB in amphibolites from the Leake Spring Metamorphics.

Overall, the composition of the amphibolites in the Leake Spring Metamorphics lies between that of N-MORB and E-MORB (Fig. 11e,f). There are very few similarities with island arc basalts and, hence, little evidence of a subduction component in the amphibolites. Nevertheless, there are similarities with back-arc basin basalts, although the Th content in the amphibolites (and high Th/Nb of 0.1 – 0.9) is much higher than in back-arc basin basalts (≤ 0.1; Pearce et al., 1995). Therefore, there is nothing in the whole-rock chemistry that requires subuction-related processes to explain the composition of the amphibolites, although there is nothing to preclude them either. If the basaltic precursors to the Leake Spring Metamorphics amphibolites were erupted in a back-arc setting, it is not clear where the arc was.

Deformation and metamorphism

Unravelling the orogenic events

Most of the deformation and metamorphism in the Gascoyne Province (and Capricorn Orogen) was originally thought to be the product of one orogenic event, the Capricorn Orogeny (Williams, 1986), the age of which was loosely defined to between c. 2200 Ma and c. 1600 Ma by Rb–Sr and Sm–Nd geochronology (Tyler and Thorne, 1990). However, extensive SHRIMP U–Pb studies have now shown that the history of the orogen is far more complicated, and that many of the structures and prograde metamorphic assemblages are actually related to orogenic events that are either older or younger than the 1820–1770 Ma Capricorn Orogeny.

As previously mentioned, recent dating indicates that there were at least four discrete Paleoproterozoic tectonothermal events, as well as two Mesoproterozoic to Early Neoproterozoic events (Fig. 8). The Paleoproterozoic events are the 2215–2145 Ma Ophthalmian Orogeny (Rasmussen et al., 2005), the 2005–1950 Ma Glenburgh Orogeny (Occhipinti et al., 2004; Sheppard et al., 2004; Johnson et al., 2010), the 1820–1770 Ma Capricorn Orogeny (Occhipinti et al., 1998; Sheppard and Occhipinti, 2000; Evans et al., 2003; Martin et al., 2005), and the 1680–1620 Ma Mangaroon Orogeny (Sheppard et al., 2005). During the Mesoproterozoic, the western part of the Capricorn Orogen was affected by a tectonothermal event at 1280–1250 Ma (the Mutherbukin Tectonic Event; Johnson et al., 2009), and the centre of the Gascoyne Province was extensively reworked or reactivated during the Edmundian Orogeny at 1030–955 Ma (Sheppard et al., 2007). In addition, the orogen underwent intracontinental reactivation during the Mulka Tectonic Event at c. 570 Ma, an event marked by anastomosing, narrow shear zones. These tectonothermal events do not have a uniform distribution throughout the Gascoyne Province, but are commonly concentrated in one or two zones.

In the Mooloo and Paradise Zones, and Errabiddy Shear Zone, medium- to high-grade metamorphism, including the migmatization and ductile deformation of sedimentary rocks, is linked to the 2005–1950 Ma Glenburgh Orogeny (Fig. 8; Occhipinti et al., 2001; Occhipinti and Sheppard, 2001; Occhipinti et al., 2004; Johnson et al., 2010).

In the Mutherbukin Zone, SHRIMP U–Pb monazite and xenotime geochronology indicates that widespread medium-grade metamorphism and deformation is related to both the Mutherbukin Tectonic Event between c. 1280 and c. 1250 Ma (Johnson et al., 2009), and the 1030–955 Ma Edmundian Orogeny (Sheppard et al., 2007; Johnson et al., 2009). SHRIMP monazite and xenotime ages of 1030–990 Ma, obtained in situ from garnet–staurolite grade pelitic schists, have estimated pressure–temperature conditions of 3–5 kbar and 500–550°C (Sheppard et al., 2007). Pegmatite melt pockets within a c. 1650 Ma gneissic metamonzogranite in the centre of the Mutherbukin Zone contain zircon grains comprising rims on igneous cores. These rims yield a precise U–Pb SHRIMP age of 1000 ± 8 Ma, interpreted to date regional metamorphism (GSWA 185945; Wingate et al., in prep.). All of these metamorphic assemblages and structures in the Mutherbukin Zone were previously considered to have formed during the Capricorn Orogeny (Williams, 1986; Culver, 2001; Varvell, 2001; Varvell et al., 2003; Fitzsimons et al., 2004).

In the Mangaroon Zone, deformation and medium- to high-grade metamorphism, which produced widespread migmatization of Pooranoo Metamorphics metasedimentary rocks, and which was accompanied by the intrusion of peraluminous two-mica granites, is related to the 1680–1620 Ma Mangaroon Orogeny (Fig. 8; Martin et al., 2005; Sheppard et al., 2005). Protoliths to the Pooranoo Metamorphics were therefore deposited after the Capricorn Orogeny (Sheppard et al., 2005).

Along the northern edge of the Yilgarn Craton, just south of the Errabiddy Shear Zone, Muhling (1988) documented crustal shortening and uplift, which she attributed to the Capricorn Orogeny; however, no robust dating was available at the time, and more recent work (Muhling et al., 2008) suggests that uplift is instead related to the Glenburgh Orogeny.

Deformation and metamorphism during the Capricorn Orogeny

Although many of the structures and peak metamorphic assemblages in the western part of the Capricorn Orogen did not form during the 1820–1770 Ma Capricorn Orogeny, there was still substantial deformation and metamorphism at this time.

In the Yarlarweelor Gneiss Complex, gneissic fabrics, upright folds, and metamorphic assemblages up to amphibolite facies (Fig. 8) formed during the Capricorn Orogeny. Metamorphism and deformation were accompanied by voluminous granite and pegmatite intrusions (Occhipinti et al., 1998; Sheppard et al., 2003).

In the Errabiddy Shear Zone, structures and metamorphic assemblages formed during the 2005–1950 Ma

Glenburgh Orogeny were overprinted by greenschist-facies metamorphic assemblages, with the formation of decametre-scale upright folds, and anastamosing shear zones (Occhipinti et al., 2001, 2003; Johnson et al., 2010). These structures and assemblages appear to be a continuation of those formed during the Capricorn Orogeny in the Yarlarweelor Gneiss Complex. Rocks in the Paradise Zone also show some low-grade recrystallization of Glenburgh Orogeny assemblages; the age of this retrogression is unclear, but it may also be related to the Capricorn Orogeny.

In the Mooloo Zone, greenschist-facies metamorphism and deformation is bracketed by granite intrusions at c. 1810 Ma and c. 1800 Ma (Occhipinti and Sheppard, 2001). This event was responsible for the retrogression of garnet–sillimanite pelitic and semipelitic schists of the Moogie Metamorphics (Johnson et al., 2010), with garnet porphyroblasts pseudomorphed by chloritoid, and sillimanite replaced by sericite. The widespread presence of chloritoid as a stable phase in both pelitic and psammitic rocks throughout the Mooloo Zone indicates that Capricorn metamorphism peaked at 425–500°C (Spear and Cheney, 1989) and at relatively low pressures (<4 kbar).

In the Mutherbukin Zone, evidence for the Capricorn Orogeny is limited to c. 1770 Ma low-U rims found on detrital zircon grains from a sample of quartzite (GSWA 187403; Wingate et al., 2010; Johnson et al., 2010). However, the grade of metamorphism during this period of new zircon rim growth is presently unknown, and it is also unclear as to what structural fabrics formed at this time.

In the Limejuice Zone, the sedimentary and mafic volcanic protoliths of the Leake Spring Metamorphics were folded and metamorphosed at medium- to high-grade during the Capricorn Orogeny, before being intruded by granites of the Minnie Creek batholith. The mafic metavolcanic or volcaniclastic rocks are now amphibolites, whereas pelitic gneisses, which are almost entirely retrogressed, contain relict peak metamorphic assemblages of andalusite–cordierite–biotite–muscovite–plagioclase, implying high-temperature and low-pressure metamorphism. The same gneissic fabric is also present in some c. 1790 Ma granites in the southeastern part of the Limejuice Zone. These granites are intruded by weakly to moderately foliated granites with igneous crystallization ages indistinguishable from the gneissic granites, indicating that medium- to high-grade metamorphism and deformation in the Limejuice Zone was broadly contemporaneous with the intrusion of parts of the Minnie Creek batholith.

Deformation and metamorphism in the Boora Boora Zone, and in the Wyloo and Capricorn Groups of the Ashburton Fold Belt to the north (Tyler and Thorne, 1990; Thorne and Seymour, 1991; Krapež and McNaughton, 1999; Martin et al., 2005), is also related to the Capricorn Orogeny*. Of the two major deformation events recognized, D_{1a} and D_{2a}, the first of these is constrained to between c. 1805 Ma and c. 1790 Ma (Martin et al., 2005, p. 14). Evidence for D_{1a} consists of a cleavage subparallel to bedding, which increases in metamorphic grade and develops into a schistosity towards the Gascoyne Province, scattered F_{1a} folds, and an unconformity between the Ashburton Formation and overlying Capricorn Group. The maximum metamorphic grade during M_{1a} is recorded by biotite–cordierite–andalusite–garnet schist. The second event, D_{2a}, which is loosely dated at between c. 1790 Ma and c. 1620 Ma, could conceivably be a product of the 1680–1620 Ma Mangaroon Orogeny rather than the Capricorn Orogeny. This event consists of upright, non-cylindrical folds with an axial planar cleavage (S_{2a}) that dips steeply to the southwest or northeast, and numerous dextral strike-slip faults that are either parallel to F_{2a} folds, or cut them at a low angle (Martin et al., 2005). Coeval metamorphism (M_{2a}) was at chlorite–muscovite grade.

Interpretation of the metamorphic evolution during the Capricorn Orogeny is hampered by an absence of quantitative or semiquantitative constraints on pressure and temperature. For example, regional metamorphism in the Yarlarweelor Gneiss Complex reached upper amphibolite facies conditions, but there are no assemblages that constrain the pressure (Sheppard and Swager, 1999). Nevertheless, some estimates can be made based on the metamorphic facies associated with the orogeny.

In the Paradise Zone of the southern Gascoyne Province, metamorphism is restricted to zones of very low grade recrystallization and retrogression of former higher grade assemblages (Fig. 8). In the Mooloo Zone, the abundance of chloritoid indicates metamorphism at moderate temperature and low pressure (Fig. 8). Further north in the Limejuice Zone, an inferred prograde assemblage of andalusite–cordierite–biotite–muscovite–plagioclase in pelitic rocks implies high-temperature and low-pressure metamorphism (e.g. Spear, 1993). To the south of the Minnie Creek batholith, the grade of metamorphism appears to have been in the low to mid greenschist facies, with peak temperatures increasing from the Paradise Zone toward the Minnie Creek batholith (Fig. 8). Peak pressures were consistently low (<4 kbar) and deformation confined to discrete shear zones and corridors of open and upright folding. To the north of the Minnie Creek batholith in the Boora Boora Zone, pelitic rocks contain the prograde assemblage of muscovite–biotite–plagioclase–garnet–?andalusite, which is almost wholly overprinted by chlorite, muscovite, and epidote (Martin et al., 2005). The prograde assemblage is consistent with lower amphibolite facies conditions at low pressures (Fig. 8). Locally, mats of fibrolite are present in muscovite, implying upper amphibolite facies conditions (Spear, 1993).

Although metamorphism associated with the Capricorn Orogeny peaked at different temperatures throughout the Gascoyne Province (from localized lower greenschist facies to amphibolite facies), the associated pressures are always low and there is no evidence for formation of the substantially thickened crust that would be expected during continental collision. Instead, metamorphism

* It is not clear as to how fabrics and metamorphic assemblages in the Boora Boora Zone relate to D_{1n} and D_{2n} fabrics farther south in the Gascoyne Province. However, geochronology indicates that D_1 and D_2 structures in the Boora Boora Zone are comparable in age with D_{1n} and D_{2n} in the adjacent Ashburton Fold Belt (Thorne and Seymour, 1991), with which they are correlated.

may be associated with emplacement of the Moorarie Supersuite, and specifically the construction of the Minnie Creek batholith, around which some of the highest grade metamorphism and most intense deformation is recorded.

Ages of granite magmatism

Overall patterns

Granitic rocks make up a large proportion of the Gascoyne Province and they were intruded in three main pulses during the Glenburgh, Capricorn, and Mangaroon orogenies (Fig. 12). About half of the granites in the province thought to have been emplaced during the Capricorn Orogeny were actually emplaced during the 2005–1950 Ma Glenburgh Orogeny or the 1680–1620 Ma Mangaroon Orogeny (Fig. 9). All of the granites in the Mangaroon Zone mapped thus far (the 'Edmund Batholith' of Williams, 1986) were intruded during the Mangaroon Orogeny, as were the majority of granites in the Mutherbukin Zone, including the 'Yinnetharra Gneiss Dome' augen gneiss of Williams (1986). Similarly, the 'Mount Marquis Batholith' (Williams, 1986) emplaced into the Yarlarweelor Gneiss Complex, is also related to the Mangaroon Orogeny.

Despite much of the granite in the Gascoyne Province being older or younger than the Capricorn Orogeny, the Moorarie Supersuite still represents a substantial volume of felsic melt (Fig. 9b). Current collisional models for the Capricorn Orogeny invoke the presence of a long-lived subduction zone somewhere in the central to northern part of the Gascoyne Province (Tyler and Thorne, 1990; Myers, 1990; Evans et al., 2003). Considering this, there should be a record of extended suprasubduction-related magmatism somewhere in the Gascoyne Province, particularly near the Minnie Creek batholith. However, what is immediately apparent is the episodic nature of the magmatism (Fig. 12), and the absence of any magmatic activity in the period 1950–1820 Ma; that is, between the end of the Glenburgh Orogeny and the start of the Capricorn Orogeny. These

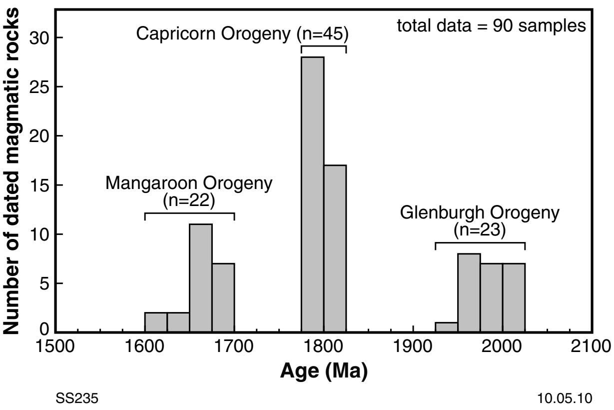

Figure 12. Histogram of all igneous crystallization ages by sample for the Gascoyne Province (bin width 25 Ma). Xenocrysts have not been plotted. Data from Geological Survey of Western Australia (2009).

age patterns are not consistent with protracted subduction, either between the Yilgarn and Pilbara Cratons (e.g. Tyler and Thorne, 1990; Myers, 1990; Powell and Horwitz, 1994; Evans et al., 2003), or along the west-facing arc proposed by Krapež (1999) in which the Capricorn Orogeny was the result of subsequent collision along the axis of the Minnie Creek batholith.

During the Capricorn Orogeny

The oldest (c. 1820–1800 Ma) Moorarie Supersuite granites were emplaced across the Gascoyne Province during the Capricorn Orogeny. Although the oldest granites in each region are within error of each other, granitic magmatism of this age is more voluminous in the southern part of the province than in either the north or within the Minnie Creek batholith (Fig. 13). Regardless of which statistical method of comparison is used (mean, median, or Tukey's Biweight Mean) magmatism in the southern part of the Gascoyne Province is distinguishably older than in the Minnie Creek batholith and in the northern part of the province. This younging to the north was suggested, from a much more limited dataset, by Evans et al. (2003, p. 862), who speculated that it may reflect '…advancement of a north-verging thrust stack.' This interpretation is not supported by the high-temperature / low-pressure nature of Capricorn metamorphism, although the reason for the northward younging is presently unclear.

The oldest Moorarie Supersuite rocks in the southern part of the province form a suite of volumetrically minor, peraluminous, two-mica granite dykes and plugs, which have been dated at 1827 ± 14 Ma and 1824 ± 9 Ma (GSWA 142924 and 142931; Nelson, 1998a,b; Occhipinti and Sheppard, 2001). However, these samples were dated without the benefit of CL imaging, and the interpreted igneous crystallization ages may have been affected by minor inherited material in the analytical sites. New CL imaging and preliminary dating indicates that the zircons of sample GSWA 142931 do not contain obvious rims, and 11 analyses of 11 zircons, most with euhedral growth zoning, yield a weighted mean $^{207}Pb^{*}/^{206}Pb^{*}$ date of 1817 ± 11 Ma. The zircons in sample GSWA 142924 contain rims enriched in U and common Pb; one rim was successfully dated. This rim, together with six analyses of six other zircons, most with euhedral growth zoning, yield a weighted mean $^{207}Pb^{*}/^{206}Pb^{*}$ date of 1812 ± 9 Ma. The igneous crystallization ages of these two granites is similar to many other granites of the Moorarie Supersuite in the southern part of the Gascoyne Province.

The oldest granite in the Minnie Creek batholith is a schistose metatonalite dated at 1808 ± 6 Ma. The metatonalite forms screens and rafts scattered through much of the batholith, suggesting that older phases may have largely been displaced and disrupted during intrusion of younger granites. This conclusion is consistent with the pattern of inherited zircons in the batholith — inherited zircons are rare (a total of 14 analyzed grains), with the most abundant population at c. 1815 Ma (Fig. 14). As this population is younger than the youngest detrital population identified in the Leake Spring Metamorphics, it was probably derived from older granitic components of the batholith. Therefore, the Minnie Creek batholith

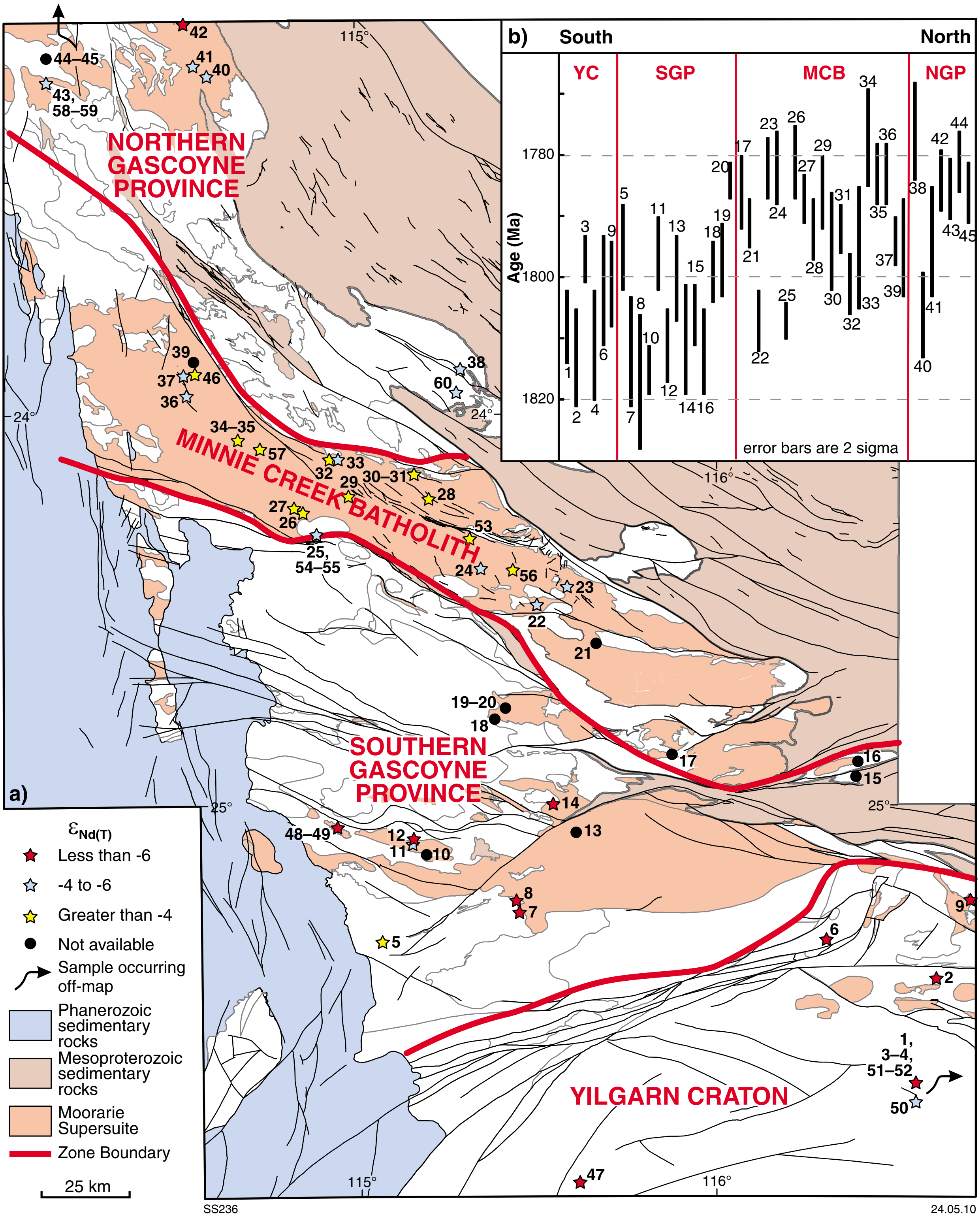

Figure 13. Summary of whole-rock Nd isotope compositions (a), and SHRIMP U–Pb zircon ages (b) for granites of the Moorarie Supersuite from south to north. Age data from Geological Survey of Western Australia (2009), Perring (1996), Krapež and McNaughton (1999), and Evans et al. (2003). Numbers on figure refer to sample numbers in Table 1. Key to abbreviations: YC — reworked Yilgarn Craton; SGP — southern Gascoyne Province; MCB — Minnie Creek batholith; NGP — northern Gascoyne Province.

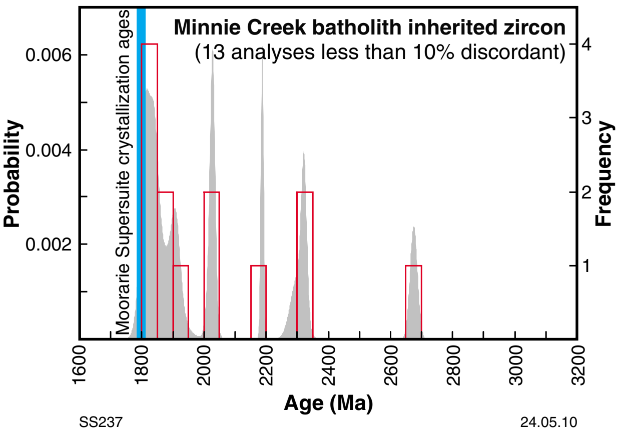

Figure 14. Combined histogram (bin width 50 Ma), and probability density plot, showing individual xenocryst analyses from the Minnie Creek batholith. Data from Geological Survey of Western Australia (2009)

probably contains elements as old as granites to the north and south; this age pattern does not match the predictions arising from the interpretation that the Minnie Creek batholith stitched a suture (Tyler and Thorne, 1990; Evans et al., 2003).

The distribution of the Moorarie Supersuite, relative to a previously proposed suture somewhere near the Minnie Creek batholith, contrasts sharply with that of the Dalgaringa Supersuite and its successor, the Bertibubba Supersuite (Figs 8, 9). Whereas the Dalgaringa Supersuite is confined to one side of a suture — the Errabiddy Shear Zone — with the successor Bertibubba Supersuite intruding both sides of the suture, the Moorarie Supersuite was intruded across the Gascoyne Province throughout its history.

Composition of the Moorarie Supersuite

Whole-rock chemistry

It is not the intention of this report to examine the petrogenesis of the Moorarie Supersuite granites in any detail, but rather to broadly classify the compositions of the granites, and discuss what bearing, if any, those chemical features have on interpretations of the tectonic setting of the supersuite.

There are myriad ways of classifying granites (see reviews by Barbarin, 1990, 1999), of which several have achieved wide currency, most notably the I–S scheme of Chappell and White (1974). All such schemes have deficiencies. Some problems relate to assumptions that granitic compositions primarily reflect a specific source rock type (Chappell and White, 1974) or a particular tectonic setting (Pearce et al., 1984), whereas other schemes are unable to deal with the wide range of granite types (e.g. Ishihara, 1977; De la Roche et al., 1980). For these reasons, we have chosen to classify the granites of the Moorarie Supersuite using the non-genetic scheme of Frost et al. (2001). This scheme uses three tiers to classify granites: the FeO/(FeO+MgO) ratio of the rock, followed by the modified alkali–lime index (Na_2O+K_2O-CaO), and finally the aluminium saturation index (ASI: molecular Al/(Ca-1.67P+Na+K). Unfortunately, most of the elements used in this classification scheme are amongst the more mobile during deformation and metamorphism (e.g. Rollinson, 1993, p. 72); however, for the Moorarie Supersuite there is no correlation between any of the indexes used in the scheme and loss on ignition, H_2O^+, or CO_2 (GSWA, unpublished data). Furthermore, samples from individual units (such as the Dumbie Granodiorite and Scrubber Granite; Occhipinti and Sheppard, 2001) define tight groupings on normalized multi-element diagrams ('spider diagrams'), even for the LILE other than Cs and Ba. These observations suggest that although some element mobility is possible, it is unlikely to be on a scale sufficient to change the classification of the granites.

A total of 106 whole-rock analyses were obtained from various units of the Moorarie Supersuite, including the Minnie Creek batholith. According to the scheme of Frost et al. (2001), the bulk of the granites are magnesian (Fig. 15a), although some silicic rocks are ferroan granites. In terms of the modified alkali–lime index, most of the granites are calc-alkalic, with a small proportion of calcic granites (Fig. 15b) in the Minnie Creek batholith and southern Gascoyne Province. There are also some alkali-calcic compositions in the Dumbie Granodiorite (Occhipinti and Sheppard, 2001) of the Mooloo Zone. Most of the rocks in the supersuite are weakly peraluminous, with a small number of metaluminous rocks, almost all of which are calcic granites of the Minnie Creek batholith (Fig. 15c). Therefore, the vast majority of the granites in the Moorarie Supersuite are magnesian, calc-alkalic, and (weakly) peraluminous. Granites of this type are common in the inboard portion of Cordilleran batholiths (Frost et al., 2001), such as the Idaho batholith (Hyndman, 1984). These batholiths may be located 600 km or more away from the convergent plate boundary in an intracontinental collisional setting (Driver et al., 2000). The Minnie Creek batholith also includes some magnesian, calcic, and metaluminous compositions, which are common on the outboard parts of Cordilleran batholiths and in island arcs (Frost et al., 2001).

Granites of the Moorarie Supersuite mostly range from about 61 wt% SiO_2 to 77 wt% SiO_2 (Fig. 16a), although of the 23 analyses with <68 wt% SiO_2, 19 are from the Minnie Creek batholith and include a quartz diorite inclusion (~55 wt% SiO_2) and a gabbro (~50 wt% SiO_2). The less silicic nature of the batholith compared with the remainder of the Moorarie Supersuite, coincides with the presence of mafic–ultramafic intrusions and mafic inclusions in the batholith. The range of Moorarie Supersuite compositions is much narrower within individual zones; for example, granites emplaced into the reworked northern margin of the Yilgarn Craton are exclusively silicic (>71 wt% SiO_2; Sheppard et al., 2003).

The Moorarie Supersuite is considerably more silicic than Phanerozoic granites from convergent continental margins. Both the Trans-Himalaya batholith (Fig. 16b)

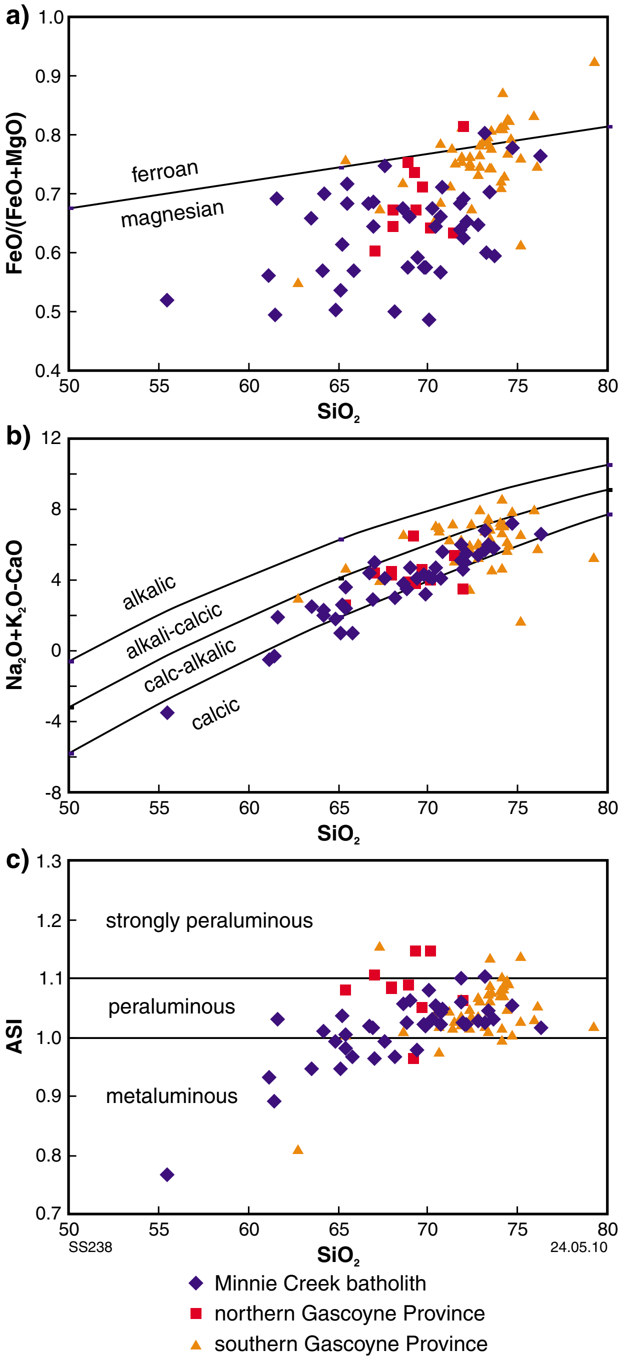

Figure 15. Plots of compositional patterns of the Moorarie Supersuite as a function of SiO_2 (wt%): (a) FeO/[FeO+MgO]; (b) Na_2O+K_2O-CaO; (c) ASI [molecular Al/(Ca-1.67P+Na+K)]. Fields from Frost et al. (2001).

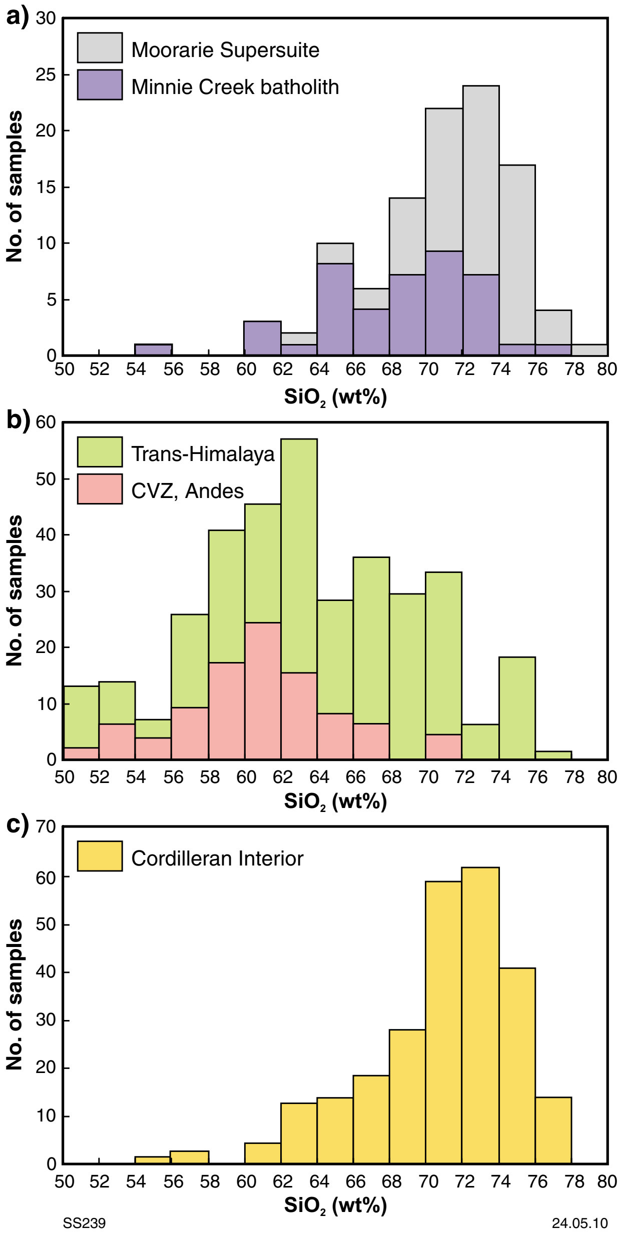

Figure 16. Histogram of SiO_2 contents for the Moorarie Supersuite (a), compared with Phanerozoic granites from convergent continental margins (b), and Cordilleran Interior batholiths (c). Parts (b) and (c) taken from De et al. (2000, fig. 12).

and the North American Peninsular Ranges batholith have an average SiO_2 content of about 64 wt% (Silver and Chappell, 1988). The eastern part of the Peninsular Ranges batholith is slightly more silicic (66 wt% SiO_2), a result consistent with the paucity of gabbros seen in comparison with the western part (Silver and Chappell, 1988), but it is still much less silicic than the Moorarie Supersuite. The distribution of SiO_2 content in the Moorarie Supersuite is similar to that of the Cordilleran Interior batholiths, such as the southeast British Columbia batholiths, and the Idaho and Cassiar batholiths (Fig. 16c). These batholiths are also dominated by calc-alkalic, weakly peraluminous (ASI 1.0 to <1.1), biotite(–muscovite) monzogranite, granodiorite, and syenogranite (Driver et al., 2000), and are emplaced well inboard of active continental margins. Therefore, based on the whole-rock compositions of the Moorarie Supersuite granites alone, a tectonic setting analogous to the Cordilleran Interior batholiths is more appropriate than that of the outboard parts of Cordilleran batholiths.

Sm–Nd isotopes

Granites emplaced during the Capricorn Orogeny show a wide range in initial ε_{Nd} values from -1.7 to -14.3 (Table 1, Fig. 17), although most of the granites have initial ε_{Nd} values between -1.7 and -8.1. Values between -11.6 and -14.3 belong to very silicic, leucocratic, pegmatitic granites in the Yarlarweelor Gneiss Complex that were derived from melting of Archean crust (Sheppard et al., 2003). Granites from individual regions show a more restricted range of values (Figs 13, 17). Minnie Creek batholith granites have initial ε_{Nd} values of -1.7 to -5.5, which are less negative than coeval granites to the north and south (-4.0 to -8.1), and on the whole have younger T_{DM} model ages (15 of 20 samples between 2570 and 2250 Ma) than contemporaneous granites elsewhere in the province (22 of 27 samples with T_{DM} model ages older than 2570 Ma). These data imply a more juvenile source for the Minnie Creek batholith, on average. The presence of gabbros and widespread mafic inclusions in the granites of the batholith suggests that this juvenile component was probably derived directly from the mantle. Nevertheless, all of the granites, including those of the Minnie Creek batholith, have much lower ε_{Nd} values than depleted mantle at this time (~+6, using the depleted mantle model of Goldstein et al., 1984), implying that the granites were largely derived by melting of older crust, which must have one component older than c. 2570 Ma.

The exclusively negative ε_{Nd} values of the Moorarie Supersuite are in contrast to Phanerozoic batholiths at convergent continental margins, such as the Trans-Himalaya batholith, Coast Mountains batholith, and the Sierra Nevada batholith, which mostly contain granites with initial ε_{Nd} values greater than CHUR (e.g. Debon et al., 1986; Crawford and Searle, 1992; Ducea and Barton, 2007; DeCelles et al., 2009). Although these batholiths also contain magmatic 'flare-up' events marked by negative initial ε_{Nd} excursions, unlike the Moorarie Supersuite they are set against a background of ongoing juvenile magmatism (Haschke et al., 2006; Ducea and Barton, 2007; DeCelles et al., 2009). Instead, the range of negative initial ε_{Nd} values in the Moorarie Supersuite resembles the distributions in Cordilleran Interior batholiths (Driver et al., 2000). Granites from the European Caledonides and Hercynides are also dominated by negative $\varepsilon_{Nd(T)}$ values (e.g. Jahn et al., 2000). However, these collisional belts, unlike the Gascoyne Province, contain abundant evidence of thrusting and early low-temperature and high-pressure regional metamorphism.

Discussion

Locations and ages of sutures in the western Capricorn Orogen

The model of Tyler and Thorne (1990) for collision of the Yilgarn and Pilbara Cratons during the Capricorn Orogeny proposed a single suture located on the northern side of the Minnie Creek batholith. However, recognition that the Errabiddy Shear Zone is a Paleoproterozoic suture (Nutman and Kinny, 1994) juxtaposing the Glenburgh Terrane and Yilgarn Craton during the latter stages of the Glenburgh Orogeny at 1965–1950 Ma (Kinny et al., 2004; Occhipinti et al., 2004; Sheppard et al., 2004; Selway et al., 2009; Johnson et al., 2010), not during the 1820–1770 Ma Capricorn Orogeny, suggests that the orogen has a more complicated history than previously appreciated. The possibility of a younger suture on the northern side of the Minnie Creek batholith is considered unlikely for two reasons. First, magnetotelluric data, which support the conclusion from field mapping and geochronology that the Errabiddy Shear Zone is a suture, do not show any contrast across the Minnie Creek batholith, suggesting that the Glenburgh Terrane forms basement to the entire Gascoyne Province (Selway et al., 2009). Second, the 1840–1820 Ma Leake Spring Metamorphics continue from the Limejuice Zone northwards across the proposed suture into the Boora Boora Zone (Sheppard et al., 2005).

The aforementioned magnetotelluric survey (Selway et al., 2009) shows a subvertical boundary between the Pilbara Craton and Glenburgh Terrane roughly coincident with the Talga Fault, which is interpreted as a suture, although the age of this suture is not well constrained at present. If the lower Ashburton Formation and Leake Spring Metamorphics are correlatives, then the suture must be older than c. 1810 Ma, the minimum depositional age of the upper Ashburton Formation and Leake Spring Metamorphics (Sircombe, 2003; Kirkland et al., 2009; Wingate et al., in prep.), and therefore must have formed before the Capricorn Orogeny. Given the lack of evidence for any structures and metamorphic assemblages attributable to the Glenburgh Orogeny along the northern margin of the Capricorn Orogen, the suture most likely formed during the Ophthalmian Orogeny. This orogeny produced a northerly verging, fold-and-thrust belt along the southern margin of the Pilbara Craton (Powell and Horwitz, 1994). Monazite and xenotime geochronology indicates that deformation and metamorphism also youngs from south (c. 2215 Ma) to north (c. 2145 Ma) during this event (Rasmussen et al., 2005). Blake and Barley (1992) proposed that the fold-and-thrust belt formed during a collision between the Pilbara Craton and a continent to the south; we suggest that this continent was the Glenburgh

Table 1. Summary of igneous crystallization ages and Nd isotope compositions of samples from the Moorarie Supersuite

Sample no.	*Tectonic region*	*Sample ID*	*Easting[(a)]*	*Northing*	*Rock type*	*Magmatic age*	*Uncertainty (95% conf.)*	*εNd(T)*
1	Yilgarn Craton	142851	576200	7174200	granite	1808	6	-6.9
2	Yilgarn Craton	142849	560760	7186340	granite	1813	8	-14.3
3	Yilgarn Craton	142852	591429	7187811	granite	1797	4	-8.0
4	Yilgarn Craton	142854	588400	7191700	granite	1811	9	-7.5
5	southern Gascoyne Province	191995	405118	7196088	granite	1795	7	-3.6
6	Yilgarn Craton	142900	529856	7197186	granite	1802	9	-6.9
7	southern Gascoyne Province	142924	443590	7204850	granite	1812	9	-6.3
8	southern Gascoyne Province	142931	442540	7208100	granite	1817	11	-6.7
9	Yilgarn Craton	142856	570300	7208200	granite	1801	7	-8.0
10	southern Gascoyne Province	159724	417420	7220900	granite	1815	4	–
11	southern Gascoyne Province	159996	413420	7224140	granite	1796	6	-5.5
12	southern Gascoyne Province	159995	413700	7225400	granite	1811	6	-6.3
13	southern Gascoyne Province	168939	459480	7227250	granite	1800	7	–
14	southern Gascoyne Province	159987	453100	7235160	granite	1810	9	-8.1
15	southern Gascoyne Province	169050	538150	7242920	granite	1806	5	–
16	southern Gascoyne Province	169052	538770	7247050	granite	1812	7	–
17	Minnie Creek batholith	188974	486418	7249142	granite	1786	6	–
18	southern Gascoyne Province	185952	436639	7259064	granite	1799	5	–
19	southern Gascoyne Province	185950	439741	7262124	granite	1797	6	–
20	southern Gascoyne Province	185951	439741	7262124	granite	1784	3	–
21	Minnie Creek batholith	190660	465066	7280382	granite	1791	4	–
22	Minnie Creek batholith	180947	448344	7291435	granite	1807	5	-5.0
23	Minnie Creek batholith	180938	456845	7296488	granite	1782	5	-4.8
24	Minnie Creek batholith	180933	432693	7301516	granite	1782	6	-5.4
25	southern Gascoyne Province	183205	386327	7310931	granite	1807	3	-4.7
26	Minnie Creek batholith	88419	382401	7317220	granite	1781	6	-3.4
27	Minnie Creek batholith	88420	380093	7318519	granite	1787	4	-3.2
28	Minnie Creek batholith	88405	417899	7321184	granite	1792	5	-4.0
29	Minnie Creek batholith	183269	395353	7321676	granite	1786	6	-3.7
30	Minnie Creek batholith	88407	413885	7328156	granite	1794	8	-3.5
31	Minnie Creek batholith	88408	413783	7328204	granite	1792	4	-3.0
32	Minnie Creek batholith	88412	389940	7332230	granite	1801	5	-3.2
33	Minnie Creek batholith	88411	392360	7332260	granite	1795	10	-4.2
34	Minnie Creek batholith	88415	364313	7337732	granite	1777	8	-3.2
35	Minnie Creek batholith	88414	364335	7337800	granite	1783	5	-3.7
36	Minnie Creek batholith	178024	349960	7350180	granite	1783	5	-3.0
37	Minnie Creek batholith	178025	352200	7356340	granite	1794	4	-1.7
38	northern Gascoyne Province	169058	426780	7357910	granite	1776	8	-5.7
39	Minnie Creek batholith	MD02MC[(b)]	351900	7359550	granite	1795	8	–
40	northern Gascoyne Province	169088	355440	7440020	granite	1806	7	-5.1
41	northern Gascoyne Province	169087	351550	7442960	granite	1794	9	-4.1
42	northern Gascoyne Province	169086	348780	7454740	granite	1784	5	-6.1
43	northern Gascoyne Province	88444	371564	7464594	granite	1786	5	-5.7
44	northern Gascoyne Province	W431	303120	7479240	granite	1781	5	–
45	northern Gascoyne Province	Krapez & McN[(c)]	397104	7517011	granite	1786	5	–
46	Minnie Creek batholith	178755	349140	7355800	granite	**1785**[(d)]	–	-2.7
47	southern Gascoyne Province	144814	460514	7128830	granite	**1800**	–	-6.7
48	southern Gascoyne Province	159788	392300	7228600	granite	**1800**	–	-7.1
49	southern Gascoyne Province	159789	392300	7228600	granite	**1800**	–	-7.6
50	Yilgarn Craton	164302	600154	7186324	granite	**1812**	–	-5.9
51	Yilgarn Craton	164305	589660	7191785	granite	**1812**	–	-11.5
52	Yilgarn Craton	164306	582124	7188607	granite	**1812**	–	-13.4
53	Minnie Creek batholith	183203	429537	7310046	metagabbro	**1810**	–	-3.0
54	Minnie Creek batholith	183205	386327	7310931	granite	**1808**	–	-4.7
55	Minnie Creek batholith	183206	386327	7310931	granite	**1810**	–	-5.1
56	Minnie Creek batholith	183268	441573	7301070	metagabbro	**1800**	–	-2.4
57	Minnie Creek batholith	174503	370484	7335236	metagabbro	**1800**	–	-2.1
58	northern Gascoyne Province	88466	377458	7516482	granite	**1786**	–	-5.5
59	northern Gascoyne Province	88452	397323	7517353	granite	**1786**	–	-5.2
60	northern Gascoyne Province	168780	425540	7351240	granite	**1776**	–	-4.4

NOTES:
(a) Eastings and northings refer to GDA94, Zone 50
(b) Sample from Evans et al. (2003)
(c) Sample from Krapez and McNaughton (1999)
(d) Magmatic ages in bold are constrained by field relationships or correlation with dated samples

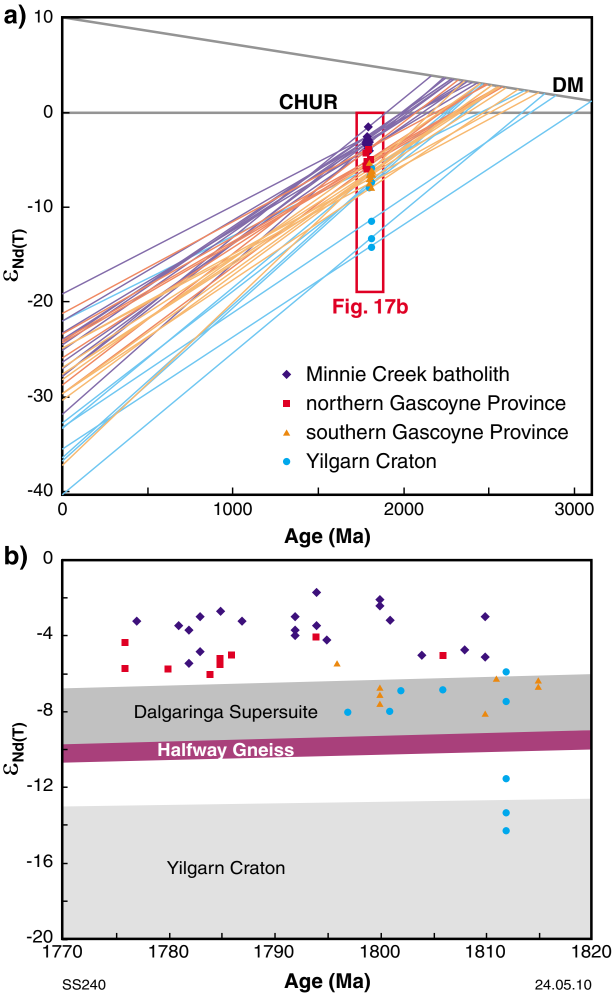

Figure 17. Plot of initial ε_{Nd} values (determined using method outlined in Waight et al., 2000) versus igneous crystallization age for granites of the Moorarie Supersuite. Data for granites and amphibolites from the Dalgaringa Supersuite from Sheppard et al. (2004); granitic rocks of the northwestern Yilgarn Craton from Nutman et al. (1993); c. 1800 Ma granites intruding the Yilgarn Craton from Sheppard et al. (2003).

Terrane, which was accreted to the Pilbara Craton along the suture identified by the magnetotelluric survey (Selway et al., 2009).

Vale the Capricorn collision

If conclusions about the ages of the sutures between the Gascoyne Province and its bounding cratons are correct, then the Capricorn Orogeny does not reflect continent–continent collision as these three pieces of crust would have already been juxtaposed by c. 1950 Ma (Johnson et al., 2010). This conclusion is consistent with the regional patterns of metamorphism, deformation, and magmatism outlined earlier. The apparent presence of high-temperature/low-pressure metamorphism across the Gascoyne Province during the Capricorn Orogeny is not consistent with substantial shortening and burial and, as a corollary, models of continent–continent collision and crustal thickening. Moreover, regional metamorphism is better attributed to a protracted period of thermal metamorphism associated with the intrusion of voluminous granitic magmas, especially construction of the Minnie Creek batholith.

Although current models for the Capricorn Orogeny invoke convergent margin magmatism leading up to a collision at c. 1800 Ma, there are no exposed rocks in the area with depositional or crystallization ages between c. 1950 Ma (the end of the Glenburgh Orogeny) and c. 1840 Ma (the maximum depositional age of the Leake Spring Metamorphics). The Ashburton Formation contains some detrital zircons with ages of c. 1860 Ma, which were sourced from the present-day southeast, but the source rocks themselves are not exposed (Sircombe, 2002). Therefore, the magmatic history of the Gascoyne region is not consistent with protracted subduction leading up to the Capricorn Orogeny, either between the Yilgarn and Pilbara Cratons (e.g. Myers, 1990; Tyler and Thorne, 1990; Powell and Horwitz, 1994; Evans et al., 2003), or along the west-facing arc proposed by Krapež (1999). It follows that the Ashburton Formation along the northern margin of the orogen is unlikely to represent a peripheral foreland basin as proposed by Thorne and Seymour (1991).

Tectonic implications of the igneous rock compositions

The calc-alkalic compositions that dominate the Moorarie Supersuite are common in Cordilleran batholiths, although it was noted by Frost et al. (2001, p. 2035) that ‘Trace element compositions of granitoids are a function of the sources and crystallization history of the melt; the tectonic environment is secondary’. They also point out that ‘… there is no *a piori* reason that would restrict any particular granitoid type to any particular tectonic setting’ (Frost et al., 2001, p. 2044) — that is, there is no one-to-one relationship between granite composition and tectonic setting. The whole-rock chemistry must be interpreted within the framework of rock associations, the spatial and temporal distribution of sedimentation and magmatism, and the history of deformation and metamorphism.

Nevertheless, the overwhelmingly silicic composition of the Capricorn Orogeny granites (monzogranite, granodiorite, and subordinate tonalite) distinguish them from batholiths formed at Andean-type margins and at island arcs. Although granites from the Minnie Creek batholith have less negative initial ε_{Nd} values and a greater proportion of intermediate rocks than other granites emplaced during the Capricorn Orogeny to the north and south, they do not represent an arc. The somewhat more radiogenic Nd isotope composition of the Minnie Creek batholith relative to the remainder of the Moorarie Supersuite is best explained as a zone of greater lithospheric thinning in the centre of the province with channelling of mantle melts into this zone. This is consistent with the suggestion that all granites, with the exception of the peraluminous or ‘Himalayan-type’ leucogranites, contain at least some direct mantle component (e.g. Patiño-Douce, 1999; Kemp

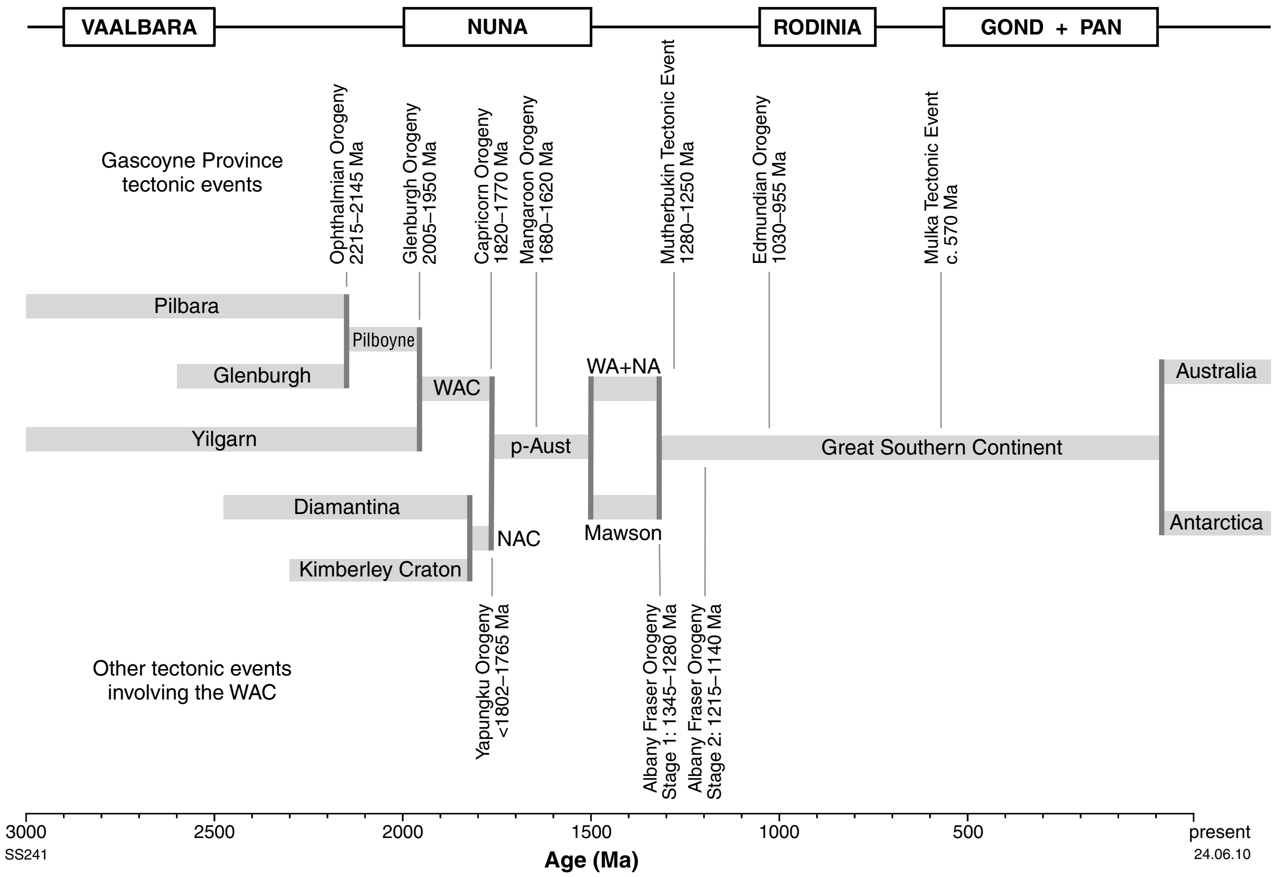

Figure 18. Tectonic events in the Gascoyne Province since 3000 Ma, and their relationship to the assembly and breakup of cratonic blocks that constitute Australia. Also shown are the supercontinent cycles (top of figure). Modified from Cawood and Korsch (2008). Key to abrreviations: WAC — West Australian Craton; NAC — North Australian Craton; p-Aust — Paleoproterozoic Australia; WA+NA — combined West and North Australian Cratons; Gond — Gondwana; Pan —Pangea.

and Hawkesworth, 2005). Underplating and intraplating of mantle-derived mafic melts appears to be necessary to achieve the high temperatures required for large-scale melting in the crust (e.g. Hildreth and Moorbath, 1988; Huppert and Sparks, 1988; Annen et al., 2006).

Isotopic data indicate that all the granites in the Moorarie Supersuite reflect reworking of older crust rather than substantial addition of new crust, which is in marked contrast to Phanerozoic convergent margins. In combination with the absence of a suture in the Gascoyne Province younger than c. 1950 Ma, it is very unlikely that any of the granites intruded during the Capricorn Orogeny were directly related to the subduction of oceanic crust. Nevertheless, there is still the possibility that the 1820–1775 Ma Moorarie Supersuite occupied a position analogous to Codilleran Interior batholiths, which may be located 600 km or more away from the convergent plate boundary. Driver et al. (2000) suggested that these batholiths formed following thickening and associated heating of a crust that included a radiogenic layer in the middle crust. They discounted the role of mafic magmas in generating the granites. Although thickening probably preceded intrusion of the batholiths, it is unlikely that this alone was responsible for generating the granites; melts formed by solely by thickening are leucocratic and strongly peraluminous, and have major and trace element compositions dissimilar to the vast majority of granites, which require a direct mantle input, including those from the Cordilleran Interior (Patiño-Douce, 1999; Kemp and Hawkesworth, 2005). Based on this, we argue that crustal thickening is not a precondition for the generation of Cordilleran Interior batholiths, but basaltic underplating or intraplating probably is.

Tectonic setting of the Capricorn Orogeny

The Gascoyne Province is unique amongst the lithotectonic elements of the Capricorn Orogen in having a prolonged and episodic history of crustal reworking and granite intrusion. There are no granites coeval with the Moorarie Supersuite intruding any other tectonic region of the Capricorn Orogen, including the Bryah, Padbury and Yerrida Basins, or the Archean Sylvania and Marymia Inliers. Furthermore, pervasive reworking during the Capricorn Orogeny is restricted to the Gascoyne Province (and Yarlarweelor Gneiss Complex); at the same time and farther east, the Bryah and Padbury Basins, and the northern edge of the Yerrida Basin, underwent folding and low-grade metamorphism (Pirajno and Adamides, 2000; Pirajno et al., 2000). The same pattern of magmatism and deformation is evident during the subsequent Mangaroon and Edmundian Orogenies (Sheppard et al., 2007; Pirajno et al., 2009).

The spatial distribution of reworking and granite intrusions in the Gascoyne Province coincides with slow velocities at about 35 km depth in the P-wave tomography of Abdulah (2007, fig. A.6). In contrast, farther to the east, the Capricorn Orogen is characterized by fast velocities, similar to those underneath the Yilgarn and Pilbara Cratons. Fishwick and Reading (2008) demonstrate that at about 75 km depth beneath the eastern Capricorn Orogen the mantle has fast shear wavespeeds, which appear to be typical of fast cratonic lithosphere, which they suggest can be reasonably explained by low temperatures beneath the Moho. They also show that the western end of the Capricorn Orogen is characterized by comparatively slow shear wavespeeds, consistent with the findings of Abdulah (2007). Therefore, the nature of the lithospheric mantle beneath different parts of the Capricorn Orogen may have exerted a strong control on the style of deformation and amount of magmatism, an effect observed in orogens and cratons worldwide (Begg et al., 2009). Given the two-stage collisional history of the Glenburgh Terrane, it is possible that the subcontinental lithospheric mantle beneath the terrane was substantially modified and refertilized by early subduction events.

If the Capricorn Orogeny does not reflect collision of the Yilgarn Craton with the Pilbara Craton, then another mechanism is required to drive the tectonism. The most conceivable mechanisms are far-field stresses related either to a plate collision, or collisions elsewhere in the continent to which the West Australian Craton belonged, or to a distal convergent continental margin. Tectonism and magmatism are recorded in many parts of the world between 1900 Ma and 1800 Ma (see Zhao et al., 2002), including across the North Australian Craton (e.g. Cawood and Korsch, 2008). In Paleoproterozoic Australia, plate collisions have been proposed between the Kimberley Craton and North Australia Craton at c. 1820 Ma (Halls Creek Orogeny; Tyler et al., 1995; Sheppard et al., 1999, 2001), and the North Australian Craton and West Australian Craton between c. 1800 Ma and c. 1765 Ma (Yapungku Orogeny; Bagas, 2004; Smithies and Bagas, 1997; Figs 1, 18). Additionally, a long-lived subduction zone may have been established along the southern margin of the North Australian Craton following the collision of the Kimberley and North Australian Cratons (Zhao and McCulloch, 1995; Scrimgeour, 2006), although the presence of this proposed subduction zone has recently been disputed (Payne et al., 2009).

At present, it is impossible to state which of these possibilities is the most plausible, as there is no evidence to indicate whether or not the West Australian Craton was joined to the North Australian Craton at this time. If the Halls Creek Orogeny or the subduction zone along the southern margin of the North Australian Craton was the mechanism, then the West Australian Craton must have already been joined to the North Australian Craton. There was no prior connection if far-field stresses from the collision between the North and West Australian Cratons during the Yapungku Orogeny was the origin of the tectonism. Without further data linking the two cratons, little more can be said on which of these two mechanisms is more likely.

The Yapungku Orogeny involved thrusting, isoclinal folding, medium- to high-pressure (up to 12 kbar) granulite facies metamorphism, and granitic magmatism. Smithies and Bagas (1997), and Bagas (2004) interpreted the metamorphism as following a clockwise P–T–t path that recorded crustal thickening related to continent–continent collision. The main episode of folding, metamorphism, and granite intrusion (D_2/M_2) started at c. 1790 Ma and continued until c. 1765 Ma. However, the age of the preceding D_1/M_1 event, which may have involved thrusting and recumbent folding, is not well known beyond being considered older than 1802 ± 14 Ma and younger than 1972 ± 4 Ma (Bagas, 2004). If the Yapungku Orogeny commenced before c. 1820 Ma — the maximum age limit for the Capricorn Orogeny — then far-field stresses from the collision between the North and West Australian Cratons could have triggered the Capricorn Orogeny.

Alternatively, if the Yapungku Orogeny is too young to be the cause of the Capricorn Orogeny, then it may well be that a collision (or collisions) to the west of the orogen was the driving force for the Capricorn Orogeny. At present, it is unclear what lay to the west of the Capricorn Orogen at c. 1800 Ma.

The Gascoyne Province was the locus of episodic intracontinental reworking for much of the Proterozoic. The susceptibility of the crust to reworking events is increased profoundly by refertilization and hydration of the underlying lithosphere, which profoundly reduces its rigidity (Griffin et al., 2008; Begg et al., 2009). Therefore, it is likely that intracontinental reworking was largely concentrated in the Gascoyne Province due to the extensively refertilized subcontinental lithosphere produced under the province during the long history of plate collisions before 1950 Ma. Regardless of which of the tectonic mechanisms outlined above is correct, the Moorarie Supersuite probably occupied a tectonic setting similar to that of the Cordilleran interior batholiths of North America.

Conclusions

Our results do not support the interpretation of the Capricorn Orogeny as a collision between the Yilgarn and Pilbara Cratons, proposed by Tyler and Thorne (1990) and others. In addition, we find nothing to sustain an interpretation of the orogeny as a sinistral megashear with a northwest-facing arc leading up to collision between the Yilgarn and Gawler Cratons, as advocated by Krapež (1999). Therefore, there is no evidence to link the combined North and West Australian Cratons with the South Australian Craton before c. 1800 Ma (cf. Krapež, 1999). Collision of the combined Glenburgh Terrane – Pilbara Craton with the Yilgarn Craton — and, therefore, the assembly of the West Australian Craton — was complete by c. 1950 Ma (Johnson et al., 2010) and the absence of any younger sutures implies that the history of the Capricorn Orogen after 1950 Ma was one of intracontinental reworking and reactivation. The 1820–1770 Ma Capricorn Orogeny is a major episode of intracontinental reworking, and is probably a far-field response to plate collisions elsewhere in the supercontinent to which the West Australian Craton belonged.

The western end of the Capricorn Orogen records episodic magmatism and thermotectonic activity over more than 1500 million years. In addition to a two-stage collisional history before c. 1950 Ma, the Gascoyne Province was the site of repeated intracontinental reworking for the remainder of the Proterozoic: namely, the 1820–1770 Ma Capricorn Orogeny, 1680–1620 Ma Mangaroon Orogeny, 1280–1250 Ma Mutherbukin Tectonic Event, 1030–955 Ma Edmundian Orogeny, and c. 570 Ma Mulka Tectonic Event. This prolonged history of reactivation and reworking has obliterated kinematic indicators and many metamorphic assemblages from the Paleoproterozoic orogenic events, and probably profoundly altered the Paleoproterozoic architecture of the Gascoyne Province. Many of the structures and prograde metamorphic assemblages formed during the different events are similar in style and grade, and were previously attributed to a single, albeit prolonged, collisional orogenic ‘event’: the Capricorn Orogeny. SHRIMP U–Pb zircon, monazite, and xenotime dating, allied with systematic regional mapping, has allowed us to identify the first-order sequence of events in the Gascoyne Province and in so doing, provides an indication of the tectonic processes responsible for the shaping the province.

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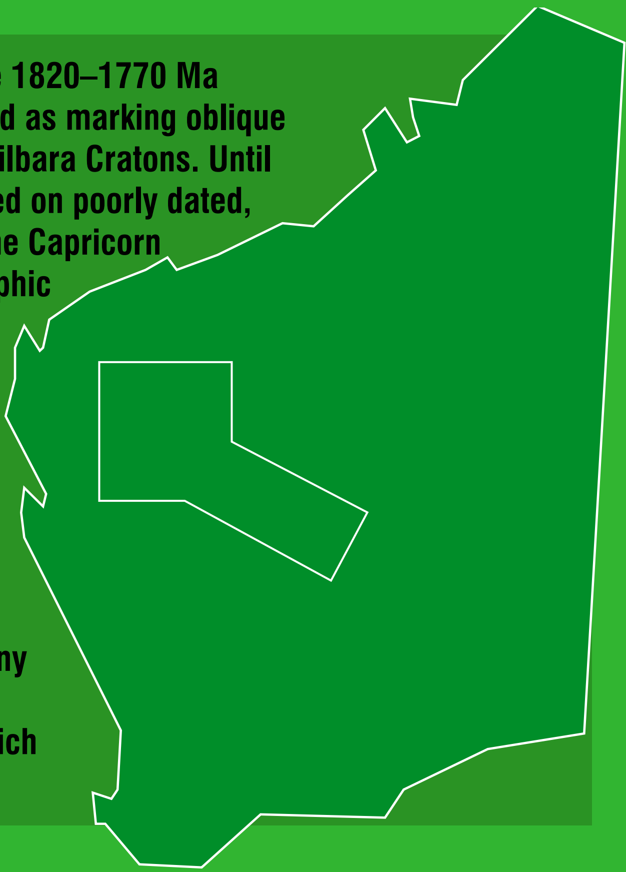
This report provides a reinterpretation of the 1820–1770 Ma Capricorn Orogeny, which is usually regarded as marking oblique collision between the Archean Yilgarn and Pilbara Cratons. Until now, models for the orogeny have been based on poorly dated, low-grade metasedimentary basins within the Capricorn Orogen, with no explanation of the metamorphic and granitic rocks in the Gascoyne Province at the western edge of the orogen. A program of regional mapping, ion microprobe (SHRIMP) dating, and whole-rock geochemical and isotopic studies in the Gascoyne Province does not support models that invoke subduction-related magmatism and continent–continent collision. The effects of the Capricorn Orogeny are better explained by intracontinental reworking of the West Australian Craton, which had assembled by c. 1950 Ma.



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