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JIMPERDING AND CHITTERING METAMORPHIC BELTS, SOUTHWESTERN YILGARN CRATON, WESTERN AUSTRALIA — A FIELD GUIDE

by S. A. Wilde

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Geological Survey of Western Australia

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

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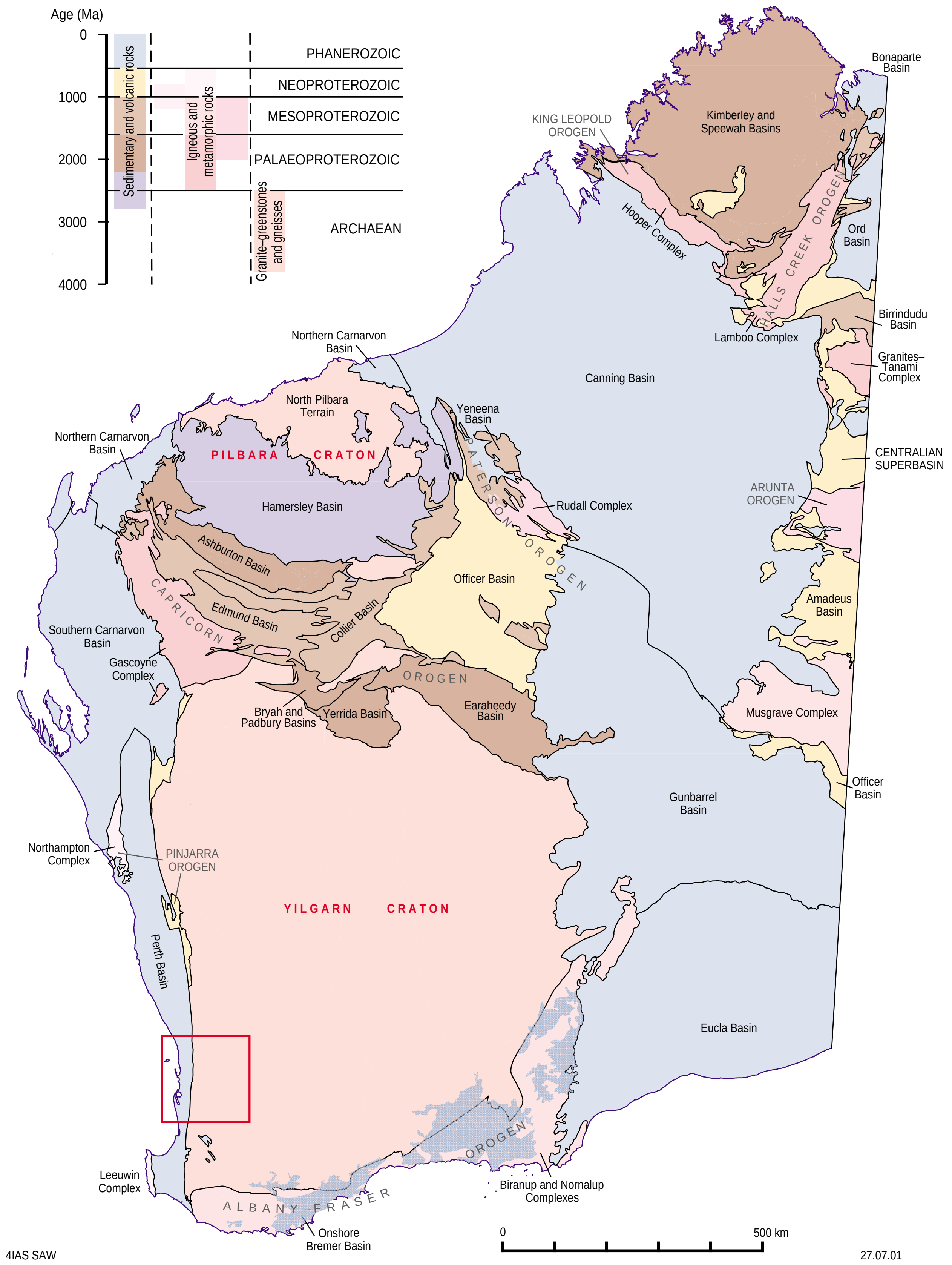

Jimperding and Chittering Metamorphic Belts, southwestern Yilgarn Craton, Western Australia — a field guide

by

S. A. Wilde[1]

Introduction

The aim of this excursion is to examine some of the diagnostic features of the Chittering and Jimperding Metamorphic Belts, which are exposed close to the western margin of the Yilgarn Craton in Western Australia (Fig. 1). In order that these outcrops may be appreciated within their regional context, a brief overview of the geological evolution of southwestern Australia is also provided. This is based largely on the review by Wilde et al. (1996).

Geological setting

Crustal evolution of the southwestern Yilgarn Craton

This excursion guide provides an update on our geological knowledge of the area, based on a review of the previous literature and incorporating major advances in the period since the 3rd International Archaean Symposium (Wilde, 1990). The modern view of the regional geology for this area commenced with mapping by the author between 1972 and 1976 for the Geological Survey of Western Australia (Wilde, 1980, 1981, 1990, 1994). Two major studies have been undertaken since the 3rd International Archaean Symposium field excursion in 1990: a geochronological investigation of the Darling Range Batholith (Nemchin, 1996; Nemchin and Pidgeon, 1997) and the acquisition of a deep seismic traverse that crossed all major crustal components (Middleton et al., 1995), thus leading to the establishment of a terrane-accretion model (Wilde et al, 1996).

The Yilgarn Craton

The Yilgarn Craton was originally subdivided by Gee et al. (1981) into four distinct regions: an older, polymetamorphic, high-grade domain along the western margin (the Western Gneiss Terrain) and three younger, low-grade granite–greenstone terranes to the east (the Murchison, Southern Cross, and Eastern Goldfields Provinces). The latter were considered to have developed, at least in part, on an older gneissic basement and to have been deposited within ensialic basins. The Western Gneiss Terrain was

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considered to extend beneath the granite–greenstone provinces (Groves and Batt, 1984; Hallberg, 1986; Blake and Groves, 1987; Campbell and Hill, 1988). The model of Gee et al. (1981) implied that the Yilgarn Craton had essentially been in existence from at least 3.0 Ga, finally becoming cratonized by the emplacement of voluminous granitoids at c. 2660 Ma.

The identification of older inherited zircons within the greenstone sequences of the eastern part of the Yilgarn Craton (Compston et al., 1985; Campbell and Hill, 1988) apparently substantiated the presence of an older granitic basement. However, detailed mapping of the Eastern Goldfields region (Swager et al., 1990), and the application elsewhere of Phanerozoic-style plate-tectonic models to the Archaean (Hoffman, 1988), led to a re-evaluation of the available evidence and to the development of an alternative model. Myers (1993) presented evidence for terrane accretion in all the Precambrian areas of Western Australia, including the Yilgarn Craton. A number of discrete terranes were identified within the northern and eastern portions of the craton and this approach has been extended into the southwest by Wilde et al. (1996), based on new U–Pb zircon data, evidence from the deep-crustal geophysical survey, and a re-evaluation of the regional geology.

Southwestern Yilgarn Craton

Major geological units

The area lies within the Western Gneiss Terrane of Gee et al. (1981), which is terminated to the west by the Darling Fault, a major crustal feature extending 1000 km along the western margin of Australia. The area is characterized by discrete, linear metamorphic belts enveloped by diffuse areas of migmatite, and all intruded by later granitoids (Wilde, 1994). The present distribution of gneiss and migmatite is largely controlled by the emplacement of the late Archaean granitoids. The granitoids in the west form several batholiths, the largest of which is the Darling Range Batholith (Wilde and Low, 1978), which post-dates metamorphism and regional tectonism. Granitoids in the east lie within the ‘Wheat Belt’ region and include charnockites. The high-grade gneisses and supracrustal rocks have been grouped into three units: the Jimperding, Chittering, and Balingup Metamorphic Belts (Wilde, 1980, 1990). There are two low-grade greenstone sequences in the southwestern Yilgarn Craton (the Saddleback and Morangup Greenstone Belts), as well as a number of high-grade rocks in the east of the area (Fig. 1) that are interpreted to be remnants of former greenstones (Wilde, 1994; Wilde et al., 1996). Migmatite is locally developed at the margins of all these metamorphic belts and also forms more extensive areas in the eastern part of the region. All sequences are intruded by a variety of granitoids, which include charnockites in the east.

Terrane model

A deep seismic-reflection survey was carried out approximately 110 km north of Perth, near New Norcia (Middleton et al., 1993). The seismic data revealed a multitude of shallow east-dipping reflectors that, on further enhancement, led to the identification of seven major zones (Fig. 2), based on reflection character and supplemented by modelling of ground magnetics. Within the Yilgarn Craton, one of the most significant features is the three-layered nature of the crust, substantiating the results of an earlier survey by Mathur (1974). The transition to the upper mantle occurs between 30 and 40 km depth and shows a marked easterly dip (Fig. 2). This passes upward into the lower crustal component, which has previously been considered to be intermediate granulite in character (Mathur, 1974; Gee et al., 1981). This component also contains some easterly dipping reflectors, which may represent ductile shear zones.

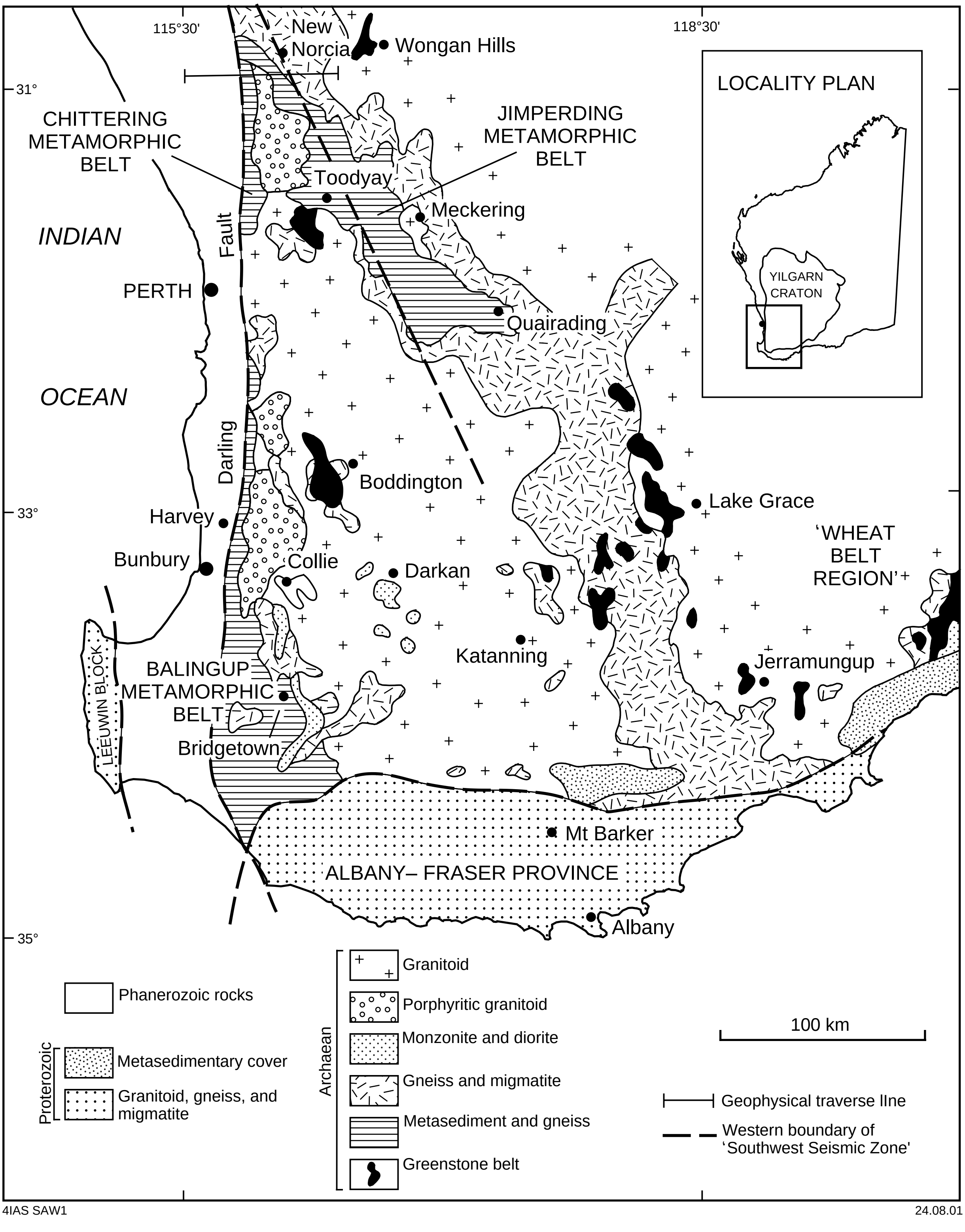

Figure 1. Regional geology of the southwestern Yilgarn Craton, showing location of the geophysical traverse line (after Middleton et al., 1995)

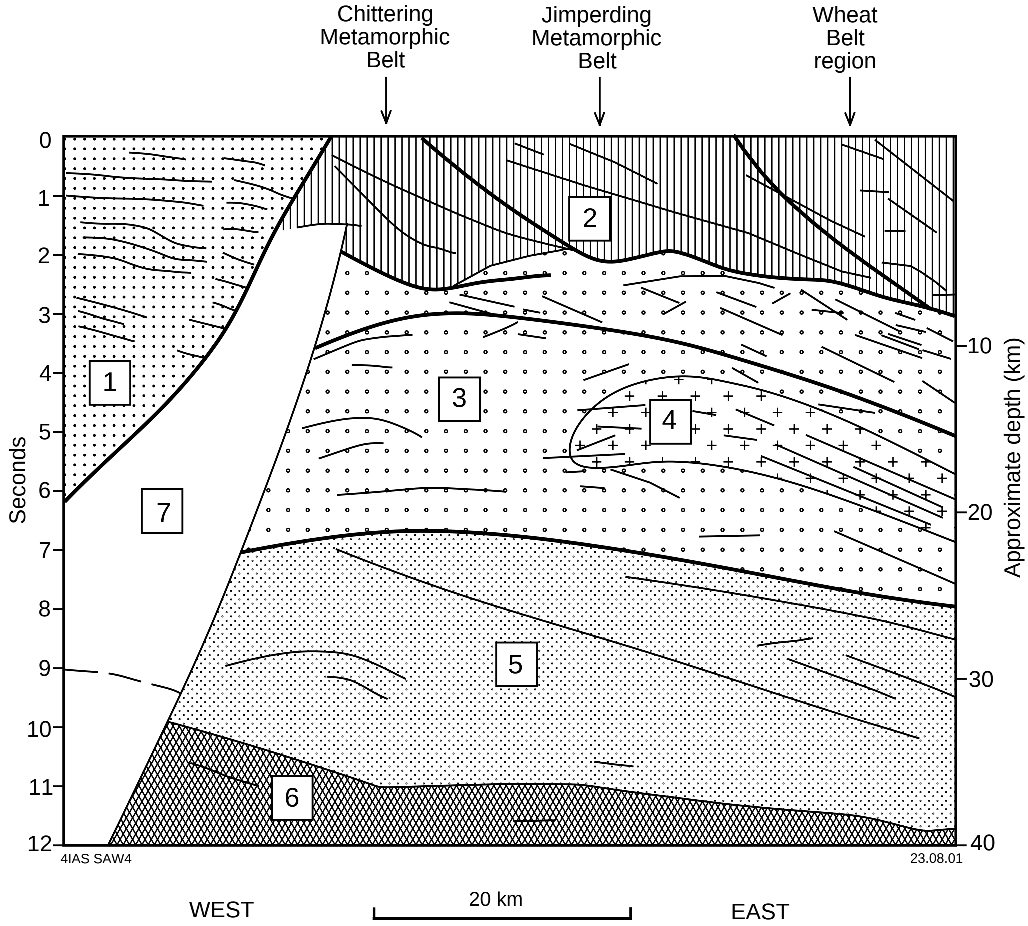

Figure 2. Line diagram of major reflectors from the deep seismic traverse, simplified to show the preferred interpretation of the deep-seismic data. There are seven zones of distinct seismic reflection character: 1) Perth Basin; 2) Western Gneiss Terrane (Gee et al., 1981), Yilgarn Craton; 3) an intermediate crustal zone within the Yilgarn Craton; 4) a zone of strong reflections interpreted as a layered mafic intrusion; 5) a deep crustal zone that contains several easterly dipping reflection events; 6) the Moho Zone, which is interpreted to be the reflection package that forms the transition from crust into upper mantle; 7) the 'proto-Darling Fault', which is a zone of non-reflection (modified from Middleton et al., 1995)

The overlying intermediate crustal layer is believed to consist chiefly of granitic rocks and is separated from the upper and lower crustal layers by major décollements. There is also a lens-like unit within this zone that is characterized by strong reflectors that are coincident in the east with two distinct magnetic bodies identified by the ground magnetics, and these have been interpreted as part of a large, layered mafic intrusion (Middleton et al., 1993).

Close to the Darling Fault, there is a distinctive segment that has virtually no seismic response. It may lie in the 'shadow zone' of the Phanerozoic Darling Fault, where no seismic energy can penetrate (Middleton et al., 1993), or it may be the result of extreme deformation associated with the Archaean–Proterozoic proto-Darling Fault Zone (Blight et al., 1981; Wilde, 1990) that outcrops in this area and is characterized by subvertical ductile structures.

Detailed geological mapping along the seismic traverse line (Fig. 1) by Morawa (1992) has shown that a number of reflections, which are interpreted as thrust faults, have surface expression (Fig. 2), including major faults that appear to define the junction between the Chittering and Jimperding Metamorphic Belts and between the Jimperding belt and the Wheat Belt region. These faults dip steeply at the surface and then 'bottom-out' onto a basal detachment in the vicinity of 2 to 3 seconds (about 7 to 9 km deep). The continuity of this detachment surface suggests that this zone was formed by 'thin-skinned' compressional tectonics, with the major crustal components (the Chittering, Jimperding and Wheat Belt regions) transported to their present positions as thrust slices (Fig. 2). Significantly, the boundary between the Jimperding and Wheat Belt regions also corresponds with a marked fall in gravity values (Fraser and Pettifer, 1980). In addition, the décollement surface appears to have been locally displaced by normal faulting (Fig. 2), suggesting reactivation under an extensional regime. This may have accompanied mafic dyke emplacement during the Proterozoic or else be associated with Phanerozoic tectonism that resulted in the development of the Perth Basin and/or its subsequent dismemberment during the Cretaceous episode of continental dispersal.

The recognition that shear zones at the surface can be correlated with major, shallow east-dipping seismic reflectors in the subsurface is of major significance. The reflectors are present throughout the total thickness of continental crust in the southwestern Yilgarn Craton and imply a uniformity in crustal construction in this region.

The coincidence of shear zones and seismic reflectors with the boundary between the Chittering and Jimperding Metamorphic Belts, and between the latter and the Wheat Belt region, implies that these areas are separate crustal segments that have been brought into their present position by transport along major east-dipping thrusts. The Jimperding Metamorphic Belt – Wheat Belt boundary is also marked by a fall in the gravity values (Middleton et al., 1993), the position of which also corresponds with a regional change in the Bouguer anomalies recognized by Fraser and Pettifer (1980). The Avon Gravity High in the west gives way sharply to a region of intermediate gravity referred to as the Narambeen Gravity Shelf (Fraser and Pettifer, 1980). Similarly, Everingham (1968) postulated that this change in gravity is centred along the currently active Southwest Seismic Zone and is related to a major change in the deep crustal structure.

When the events are considered in the light of the recently acquired geochronological data, it is possible to apply an alternative evolutionary model to that of Gee et al. (1981) to the southwestern Yilgarn Craton.

If evidence obtained along the geophysical traverse line is extended southward, it is possible to delimit the main crustal components in the southwestern Yilgarn Craton. This leads to the recognition of at least three separate terranes that, from west to east, are referred to as the Balingup, Boddington, and Lake Grace Terranes (Wilde et al., 1996; Fig. 3). In addition, a small portion of the Murchison Terrane (Myers, 1993) is present in the north around Wongan Hills. The characteristics of the newly defined terranes are summarized in Table 1 and discussed below, based on information presented in Wilde et al. (1996).

Before considering the geology in more detail, it is pertinent to review the geochronological evidence, since this also appears to substantiate the terrane model.

Geochronology of the southwestern Yilgarn Craton

Overview

The antiquity of the Western Gneiss Terrain was first established by Arriens (1971), who obtained a number of ages in excess of 3.0 Ga from gneisses using Rb–Sr techniques. A Rb–Sr study by de Laeter et al. (1981) identified gneissic rocks at least 3.3 billion years old in the northwestern part of the Western Gneiss Terrain near Mount Narryer. This led to extensive investigations within the Narryer Gneiss Complex (Myers,

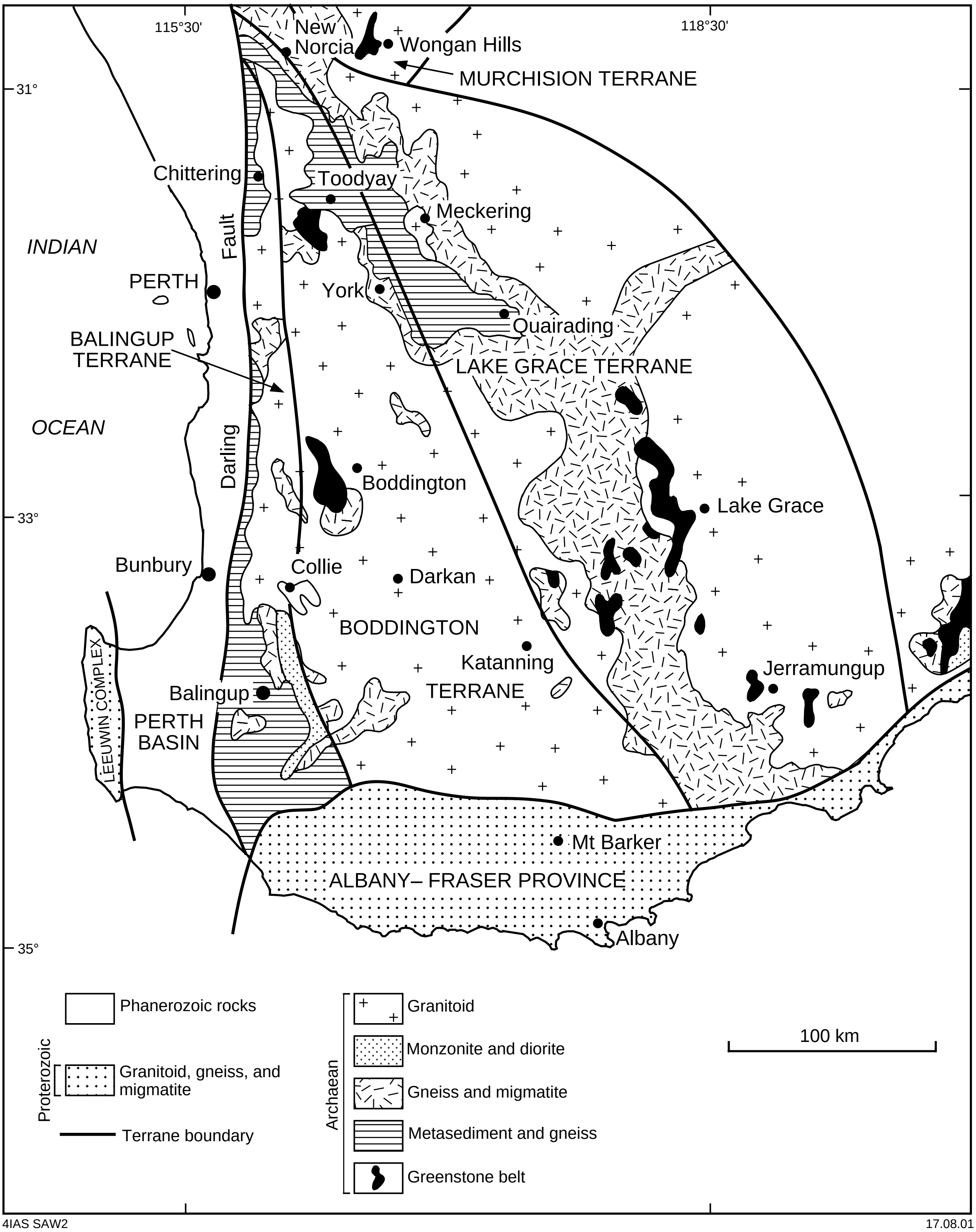

Figure 3. Terranes of the southwestern Yilgarn Craton (after Wilde et al., 1996).

Table 1. Characteristic features of the Balingup, Boddington, and Lake Grace Terranes

Balingup	*Boddington*	*Lake Grace*
Trough sediments	Shelf sediments	Shelf sediments
Medium-pressure metamorphism	Low-pressure metamorphism	Low-pressure metamorphism
No greenstone belts	Low-grade greenstones	High-grade greenstones
Ductile deformation extensive	Nappes and brittle faults	Upright folds; steep plunges
2612–2527 Ma granitoids	2677–2640 Ma granitoids	2640 and 2587 Ma granitoids
Granitoids; greenschist facies	Granitoids; greenschist facies	Granitoids (2640 Ma); granulite facies

SOURCE: Wilde et al. (1996)

1988) using Rb–Sr, Sm–Nd, Pb–Pb (de Laeter et al., 1985; Fletcher et al., 1988) and U–Pb zircon (Kinny et al., 1988, 1990; Pidgeon and Wilde, 1998) techniques. The oldest rocks so far identified in the area are gneisses that are 3730 m.y. old (Kinny et al., 1988), based on sensitive high-resolution ion microprobe (SHRIMP) U–Pb–Th data. Detrital zircons as old as 4.18 Ga (Froude et al., 1983) have been identified at Mount Narryer, and as old as 4.27 Ga (Compston and Pidgeon, 1986) and 4.4 Ga (Wilde et al., 2001) from Jack Hills. The mature clastic sedimentary rocks that contain the ancient zircons also contain considerably younger detrital zircons and, based on an evaluation of these, the age of deposition at both Jack Hills and Mount Narryer is considered to be c. 3.1 Ga (Compston and Pidgeon, 1986; Kinny et al., 1990).

Nieuwland and Compston (1981) obtained c. 3.3 Ga detrital multigrain zircon ages from mature orthoquartzites in the Toodyay area of the southwestern Yilgarn Craton, which were believed to be intruded by granite approximately 3.25 billion years ago. These rocks of the Jimperding Metamorphic Belt (Fig. 1) near Toodyay are the oldest known crustal components in the southwestern Yilgarn. Indeed, P. D. Kinny (quoted in Wilde, 1990) has identified a range of zircon ages up to 3735 Ma from the Windmill Hill quartzite.

The earliest granitoids in the region were the protoliths of the orthogneisses that intrude quartzites of the Jimperding Metamorphic Belt at Toodyay. These were emplaced approximately 3250 m.y. ago (Nieuwland and Compston, 1981) and appear to have originally been porphyritic granites (Wilde and Low, 1978). No other granitoids of this age are known from the region. However, in the southwestern Yilgarn, granitoid activity with an age of approximately 2.8 Ga is recorded for the Gibralter Quartz Monzonite near Bridgetown (Fig. 1), where Fletcher et al. (1983) obtained a Sm–Nd T_{CHUR} model age of 2.74 ± 0.09 Ga. This age, although rather imprecise, is nonetheless in agreement with the Rb–Sr model age of Rosman et al. (1980).

The bulk of the granitoids across the Yilgarn Craton belong to the major c. 2660 Ma event, which was first identified by the early Rb–Sr study of Wilson et al. (1960) and later substantiated by Arriens (1971). Geochronological results up to January 1980 were summarized and evaluated by de Laeter et al. (1981). A number of more recent studies using U–Pb zircon dating reveal a spectrum of granitoid ages from 2670 to 2587 Ma. Granitoids that intrude into the Jimperding Metamorphic Belt at Toodyay were dated by conventional multigrain U–Pb methods by Nieuwland and Compston (1981). The Mortigup granodiorite has an isochron age of 2677 +50/–43 Ma and the Monday Hill adamellite an age of 2642 +315/–166 Ma. Regression of both data sets gives an age of 2670 ± 45 Ma for granitoid crystallization in the Toodyay district. McCulloch et al. (1983) undertook a Sm–Nd study of both these rock types and calculated T_{CHUR} model ages of 2777 and 2680 Ma for the Mortigup and Monday Hill intrusives respectively (both with analytical errors of approximately ± 40 Ma).

Porphyritic granite that intrudes into the approximately three billion-year-old Wongan Hills Greenstone Belt gave an isochron age of 2651 ± 4 Ma, using combined ion microprobe and conventional single-grain U–Pb results on zircon (Pidgeon et al., 1990). Further south near Boddington (Fig. 1), monzogranite that intrudes the younger Saddleback Greenstone Belt gave a conventional multigrain isochron age of 2640 + 26/–24 Ma (Wilde and Pidgeon, 1986).

Slightly younger ages have been reported from the porphyritic adamellites and granites of charnockitic affinity from the Wheat Belt region (Wilde and Pidgeon, 1987). Using conventional multigrain zircon methods, Wilde and Pidgeon (1987) obtained U–Pb ages of 2652 ± 24 Ma on a sample from near Lake Grace (Fig. 1) and 2639 ± 19 and 2634 ± 18 Ma on granitoids from Badgebup, near Katanning. These results, when combined with a more intensely deformed variant from Badgebup, gave a pooled isochron age of 2627 ± 12 Ma.

The porphyritic Logue Brook Granite (Blight et al., 1981) near Harvey (Fig. 1) gave a Sm–Nd T_{CHUR} model age of 2905 ± 35 Ma (Fletcher et al., 1985). Although this was not necessarily considered to be the age of emplacement, it was significantly older than most other Sm–Nd model ages from Yilgarn granitoids. However, an ion microprobe U–Pb zircon study by Compston et al. (1986) established that the crystallization age was considerably younger at 2612 ± 5 Ma.

The youngest granitoids identified in the southwestern Yilgarn Craton are present in the Wheat Belt region and along the western margin of the craton in the Balingup Metamorphic Belt. Late, medium- to coarse-grained granodiorite and adamellite intrusions post-date granulite-facies metamorphism and have been emplaced into both the migmatite and granite terranes in the Wheat Belt region. Two samples from near Dumbleyung and Jerramungup gave conventional multigrain U–Pb zircon ages of 2576 ± 54 and 2595 ± 65 Ma, respectively (Wilde and Pidgeon, 1987). These give a pooled isochron age of 2587 ± 25 Ma that, although it overlaps with younger members of the general 2660 Ma granitoid suite, is nonetheless considered significant since the rocks are post-metamorphic and clearly later than the charnockitic granitoids of the Wheat Belt.

In the Balingup Metamorphic Belt, a SHRIMP zircon study of the Logue Brook Granite, located east of Bunbury (Fig. 1), by Compston et al. (1986) established a crystallization age at 2612 ± 5 Ma. In the Greenbushes area, north of Bridgetown (Fig. 1), ages of approximately 2577 Ma were recorded for the Cowan Brook Dam and Millstream Dam granitoids, whereas the Greenbushes Pegmatite has an age of 2527 Ma (Partington et al., 1986). All these rocks have undergone ductile deformation, with portions of the Logue Brook Granite infolded with paragneisses of the Balingup belt (Wilde and Walker, 1982). The Greenbushes Pegmatite is a late-stage granitoid that was emplaced synkinematically into a major sinistral shear zone under medium-pressure, amphibolite-facies conditions (Partington et al., 1986).

Recent geochronological data

A comprehensive review of the available geological and geochronological data for the southwestern Yilgarn has previously been undertaken by Wilde (1990, 1994). The main differences since the last review are discussed below.

The main changes from those presented by Wilde (1994) relate to further geochronological work undertaken on the Jimperding Metamorphic Belt in the Toodyay and Wheat Belt areas, and in the Darling Range Batholith. A SHRIMP study of the Windmill Hill quartzite (P. D. Kinny quoted in Wilde, 1990) gave a spectrum of ages from 3735 to 3177 Ma. Unpublished work on granitic sheets (orthogneisses) intruding into the quartzites indicates that the age of intrusion was c. 2.6 Ga and that the 3250 Ma age obtained by Nieuwland and Compston (1981) was most likely the result of incorporation of significant amounts of detrital zircon from digestion of older quartzite xenoliths (Kinny, P. D., 1998, pers. comm.). Whereas the depositional age of the quartzites still appears to be greater than 3.0 Ga, there is no evidence for an early pre-3.0 Ga granite-forming event in the region.

Work on the Katrine Syenite near Northam by Pidgeon et al. (1996) defined a crystallization age of 2654 ± 5 Ma. The rock did, however, contain older zircon grains and cores with an age of c. 3250 Ma, similar to the age of granitic gneisses obtained a few kilometres to the north near Toodyay by Nieuwland and Compston (1981) using multigrain zircon techniques.

New evidence on the age and metamorphism of the metasedimentary sequence comes from a conventional, single-grain U–Pb zircon study of mafic granulites (high-grade greenstones) in the Wheat Belt region southeast of Toodyay (Nemchin et al., 1994). Granulite-facies metamorphism took place between 2649 and 2640 m.y. ago, which is significantly younger than the age of 2671 to 2654 Ma recorded from the greenschist-facies Saddleback Greenstone Belt (Wilde and Pidgeon, 1986) to the west near Boddington (Fig. 1). As suggested by Wilde (1990), this implies a separate evolution for these adjoining regions.

A detailed study of the Darling Range Batholith by Nemchin (1996) identified zircons cores with ages of 2690–2650 Ma, which were interpreted as dating either the source rocks or various stages of isotopic resetting due to recrystallization and melting of the protolith (Nemchin and Pidgeon, 1997). The cores are surrounded by oscillatory zoned rims with ages of 2648–2626 Ma, interpreted as resulting from extended crystallization of the granite magma. The youngest zircon ages of 2628–2616 Ma were obtained from unzoned to weakly zoned outer rims that transgress the earlier zircon structures and appear to be the result of corrosion and recrystallization during slow cooling of the magma. These ages are similar to that obtained from the Logue Brook Granite (Compston et al., 1986).

Terrane geology

Balingup Terrane

The Balingup Terrane (Fig. 3) is a narrow, north-trending belt delimited to the west by the Darling Fault and less precisely defined in the east by later intrusive granitoids (Wilde et al., 1996). The terrane consists of the Balingup and Chittering Metamorphic Belts, which are characterized by turbiditic sediments and intensely deformed orthogneisses. The age of sedimentation is not precisely constrained, with Sm–Nd model ages (Fletcher et al., 1985) suggesting that the Balingup belt was deposited between 3.07 and 2.83 Ga and the Chittering belt after 2.89 Ga (Wilde, 1990).

The rocks were metamorphosed under medium-pressure, amphibolite-facies conditions, with the development of kyanite, sillimanite, and staurolite in metapelites being diagnostic. Metamorphism in the Balingup Metamorphic Belt has possibly been dated by the Rb–Sr age of c. 2838 Ma obtained near Bridgetown by D. A. Nieuwland (quoted in Wilde, 1980) from a quartz–feldspar–biotite gneiss; there are no comparable data for the Chittering Metamorphic Belt.

Strong ductile deformation is prominent along the western boundary of this terrane and is associated with periodic movement along the Darling Fault Zone, commencing about 2577 m.y. ago (Blight et al., 1981). The eastern boundary is defined in the extreme southeast by the Gibralter Quartz Monzonite (Wilde and Walker, 1982, 1984). This is a metamorphic tectonite believed to have been emplaced during the peak of regional metamorphism. The Sm–Nd model age of 2.74 ± 0.09 Ga (Fletcher et al., 1983) and Rb–Sr model age of c. 2828 Ma (Rosman et al., 1980), although not precise, are within error of D. A. Nieuwland's Rb–Sr isochron age.

Granitoids associated with the Balingup Terrane are younger than in most other parts of the Yilgarn Craton. An ion microprobe U–Pb–Th zircon study of the Logue Brook Granite, located east of Bunbury (Fig. 1), by Compston et al. (1986) established a crystallization age of 2612 ± 5 Ma. In the Greenbushes area, north of Bridgetown (Fig. 1), ages of c. 2577 Ma were recorded for the Cowan Brook Dam and Millstream Dam granitoids, whereas the Greenbushes Pegmatite has an age of 2527 Ma (Partington et al., 1986). All these rocks have undergone ductile deformation, with portions of the Logue Brook Granite infolded with paragneisses of the Balingup belt (Wilde and Walker, 1982). The Greenbushes Pegmatite is a late-stage granitoid that was emplaced synkinematically into a major sinistral shear zone under medium-pressure, amphibolite-facies conditions (Partington et al., 1986).

The Chittering Metamorphic Belt, which will be visited on the excursion, is composed of quartz–feldspar–biotite gneiss with interleaved metamorphosed pelite, semi-pelite, and greywacke. Banded iron-formation and quartzite are notably absent and this association was interpreted to be the result of rapid, trough-style sedimentation along a continental margin (Gee et al., 1981; Wilde, 1990). Both the Chittering and Balingup Metamorphic Belts are chiefly at amphibolite facies and have undergone moderate-pressure Barrovian-type metamorphism (Wilde, 1990). This appears to be controlled in part by their location, with the higher pressure assemblages present along the western margin of the Yilgarn Craton, and associated with ductile shear zones related to early movement along the Darling Fault Zone (Blight et al., 1981; Bretan, 1985). This zone has been reactivated at several later periods, resulting in local retrogression to greenschist-facies assemblages.

Boddington Terrane

The Boddington Terrane (Fig. 3) is separated in the north from the Balingup Terrane by a 2 km-wide shear zone characterized by phyllonite and quartz boudins, marking the junction of the Chittering and Jimperding Metamorphic Belts (Middleton et al., 1995). Further south, the original line of contact appears to have been obliterated by later granite intrusion (stitching plutons), which includes three porphyritic granite plutons (Wilde, 1990), the most southerly being the 2612 Ma Logue Brook Granite, which implies that the two terranes were in contact at this time (Wilde et al., 1996). The Boddington Terrane is only about 10 km wide in the north, but widens to around 120 km in the south (Fig. 3).

The Boddington Terrane includes the northern part of the Jimperding Metamorphic Belt. Further south, the major gravity change previously discussed occurs within this belt, and the Boddington Terrane is defined to include only the western, flat-lying portion near Toodyay and York (Fig. 3). The dominant rock type is quartz–feldspar–biotite gneiss interleaved with aluminous schist, orthoquartzite, banded iron-formation, and rare calc-silicate rocks. Some units are paragneiss and show gradations to arkosic quartzite and quartz–mica schist. This association is a characteristic feature of the Jimperding Metamorphic Belt east and southeast of Toodyay and has been interpreted as indicating stable shelf sedimentation on a pre-existing sialic basement (Gee et al., 1981; Wilde, 1990). The metapelites contain andalusite, sillimanite, and cordierite, indicating low-pressure, high-temperature metamorphic conditions. The Jimperding metasediments in the Boddington Terrane form part of a major nappe that has been strongly refolded (Wilde and Low, 1978), with metamorphic grade increasing eastward from lower amphibolite to amphibolite–granulite transition facies, with andalusite and iron-rich cordierite indicating low-pressure conditions. It appears from the detrital zircon suite that the orthoquartzites were deposited between 3.1 and 3.0 Ga (Wilde, 1990). However, the oldest zircons are c. 3735 Ma and were interpreted by P. D. Kinny (Wilde, 1990) as being derived from a granite–gneiss terrane. Although there is no other evidence of crust of this age in the southwestern Yilgarn Craton, it has been identified in the Narryer Gneiss Complex, some 800 km to the north (Kinny et al., 1990; Pidgeon and Wilde, 1998).

The Saddleback and Morangup Greenstone Belts form part of the Boddington Terrane and consist of mafic and felsic lavas and pyroclastic rocks, with minor amounts of sediment metamorphosed to low-pressure greenschist facies and largely downfaulted against adjacent gneiss and granitoid. The Saddleback belt formed between 2671 and 2654 Ma (Wilde and Pidgeon, 1986) and the Morangup sequence near Toodyay (Fig. 1) is considered to be of similar age (Wilde and Pidgeon, 1990). The geochemical nature of both sequences is consistent with either a mature island arc or a continental Andean-type magmatic arc setting (Wilde, 1990, 1994).

Granitoids in the Boddington Terrane appear to have developed within a short time span between 2677 and 2640 Ma (Wilde, 1990; Wilde et al., 1996). Conventional multigrain U–Pb dating of zircons gave ages from 2677 to 2642 Ma for granitoids intruding the Jimperding Metamorphic Belt at Toodyay (Nieuwland and Compston, 1981). Further south near Boddington (Fig. 1), monzogranite intruding the Saddleback Greenstone Belt gave a conventional multigrain isochron age of 2640 +26/–24 Ma (Wilde and Pidgeon, 1986). This age pattern is supported by the work of Nemchin and Pidgeon (1997), although some recrystallization of zircon continued until approximately 2616 Ma, which is similar to the crystallization age of the Logue Brook Granite in the Balingup Terrane to the west. The granitoids of the Darling Range Batholith are quite diverse and show considerable textural variation. They range in composition from granodiorite to granite; the compositional variations being commonly independent of textural changes. Where cross-cutting relations can be identified, granodiorite is generally the earliest phase. Most granitoids are undeformed, although plutons of porphyritic granite that are present close to the eastern boundaries of the Chittering and Balingup Metamorphic Belts show evidence of intense ductile shearing.

There are also a number of small bodies of quartz-poor granitoids of dioritic, monzonitic, and syenitic affinity within the granite batholiths of the Boddington Terrane. More extensive areas of quartz monzonite are present south of Darkan (Fig. 1) and these are rich in amphibolite xenoliths.

Migmatites form significant areas in the Boddington Terrane (Fig. 3). They are composed of earlier gneissic rocks invaded by younger, less deformed, unmetamorphosed granitoid components. They are developed at the margins of the metamorphic belts, with much of the gneissic palaeosome derived from reworking of the metamorphic rocks. Some granitic neosomes appear to have been formed by in situ partial melting of the gneisses (Wilde, 1990).

The eastern boundary of the Boddington Terrane is clearly defined by gravity data and coincides with the Yandanooka – Cape Riche Lineament of Everingham (1968), and the boundary between the Avon Regional Gravity High in the west and the Narembeen Regional Gravity Shelf to the east, as defined by Fraser and Pettifer (1980). Significantly, this zone is also the focus of present-day seismic activity and has been referred to as the ‘Southwest Seismic Zone’ (Doyle, 1971).

Lake Grace Terrane

The Lake Grace Terrane (Fig. 3) extends east from the major change in gravity discussed above for almost 200 km and corresponds closely with the area originally referred to as the ‘Wheat Belt’ region by Wilson (1958). It narrows markedly in the north, where it comes into contact with the Murchison Terrane (Fig. 3). The nature of this boundary is uncertain and may need to be revised in the light of future work. The eastern boundary of the Lake Grace Terrane is not precisely defined, owing to a lack of detailed geological mapping. The terrane is characterized by sediments, greenstones, granitoids, and migmatite, mostly at granulite facies.

Included within the terrane is the eastern portion of the Jimperding Metamorphic Belt and isolated outcrops of metamorphic rocks incorporated in the migmatite zone that extends south from near Meckering to the southern limit of the Yilgarn Craton (Fig. 3); this is the most extensive zone of gneiss and migmatite in the southwestern Yilgarn. The metamorphic rocks consist of thin units of orthoquartzite, arkosic paragneiss, and banded iron-formation, interleaved with a variety of garnetiferous orthogneiss and ultramafic units, and all enclosed within granitoids. All rocks are metamorphosed to granulite facies, with the local development of cordierite in metapelites indicating low-pressure conditions. The boundary within the Jimperding Metamorphic Belt is marked by the easterly change from amphibolite–granulite transition facies to granulite-facies metamorphism and the structural change from flat-lying nappes in the west to north-plunging upright folds in the east (Wilde and Low, 1978).

A number of greenstone remnants are present within the Lake Grace Terrane (Fig. 3), with ages up to c. 2790 Ma (Wilde and Pidgeon, 1987; Nemchin et al., 1994). These were interpreted as ‘keels’ of original greenstone belts by Wilson (1969). The mineralogical features indicate that this whole area underwent low- to moderate-pressure granulite-facies metamorphism and the enclosing granitoids commonly include hypersthene-bearing charnockites (Wilde, 1990). A two-pyroxene mafic granulite from Badgebup, 33 km northeast of Katanning (Fig. 3), has been interpreted as an original basalt (Wilde and Pidgeon, 1987) and has a concordant, conventional multigrain zircon U–Pb isochron age of 2798 ± 16 Ma. Felsic gneisses associated with the mafic granulite give a comparable, though less precise, age of 2750 ± 130 Ma. A felsic granulite from west of Lake Grace (Fig. 3) has a poorly constrained, discordant isochron age of c. 2780 Ma (Wilde and Pidgeon, 1987) and a T_{CHUR} Sm–Nd model age of 2.77 ± 0.04 Ga (Fletcher, I. R., 1987, written comm.). The consistency of these ages across the Lake Grace Terrane and their similarity to the 2795 ± 38 Ma age for the lower succession Kathleen Valley greenstones of the Eastern Goldfields Province (Cooper and Dong, 1983) supports the view of Wilson (1969) that these granulites represent uplifted portions of the lower ‘root’ zones of greenstone belts. Nemchin et al. (1994) also identified zircon components in the mafic granulites with ages of 2649 to 2640 Ma. These are ascribed to later zircon growth during the regional granulite-facies metamorphism.

The earliest granitoids within the Lake Grace Terrane are hypersthene-bearing charnockites (Wilde and Pidgeon, 1987). Reaction textures indicate that hypersthene and subsequent mafic minerals followed a magmatic crystallization sequence and that these charnockites are of igneous origin. Conventional multigrain U–Pb zircon ages of 2652 ± 24 Ma were obtained near Lake Grace and ages of 2639 ± 19 and 2634 ± 18 Ma were recorded from Badgebup, 33 km northeast of Katanning (Fig. 3). These results, when combined with a more intensely deformed variant, give a pooled isochron age of 2627 ± 12 Ma. More precise single-zircon data from some of the same samples gave a consistent age of 2640 Ma (Nemchin, A. A., 1998, pers. comm.), considered to reflect zircon growth during peak regional metamorphism which, for these rocks, is believed to be coincident with their emplacement (Wilde and Pidgeon, 1987). These high-grade charnockitic rocks formed at the same time as 'normal' monzogranites were intruding the Saddleback Greenstone Belt in the Boddington Terrane to the west, again indicating the fundamental nature of the boundary between these two terranes. Younger granitoids are also present in the Lake Grace Terrane, with medium- to coarse-grained granodiorite and monzogranite cutting the charnockites and post-dating granulite-facies metamorphism. These give a pooled multigrain zircon U–Pb isochron age of 2587 ± 25 Ma (Wilde and Pidgeon, 1987).

Granulite facies metamorphism in the Lake Grace Terrane was of regional extent and took place at relatively low pressure, with the local development of sapphirine and cordierite suggesting pressures below 6 kb (Wilde and Pidgeon, 1987). Estimates of peak metamorphic temperatures also indicate a uniformity across the area. Wilde and Pidgeon (1987) calculated a temperature of approximately 700°C at Lake Grace, based on limited two-pyroxene geothermometry. Lindsley (1983) calculated mean temperatures of 740 ± 45°C (clinopyroxene) and 693 ± 28°C (orthopyroxene) for rocks at Quairading (Fig. 3), using the pyroxene analyses of Davidson (1968). Similarly, temperature estimates by Wilson and Green (1971), based on the fractionation of O^{18}/O^{16} in coexisting mineral pairs from South Quairading, gave a maximum temperature of 695 ± 15°C. All these data support the idea that a portion of intermediate crust was transported to the surface along the thrust zone identified in a previous section of this report, and marked by the major change in gravity signature. The timing of this event is unknown but, since there is no evidence of the c. 2640 Ma granulite-facies metamorphic event in the Boddington Terrane to the west, it must have occurred after 2640 Ma.

Murchison Terrane

A small portion of the Murchison Terrane (Myers, 1993) lies within the southwestern Yilgarn Craton and is represented by the 3008 Ma Wongan Hills Greenstone Belt (Pidgeon et al., 1990). The age of many greenstone sequences in the Murchison Terrane is 3.0 Ga and they are distinctly older than similar lithologies to the south and east (Pidgeon and Wilde, 1990). Closely associated with the greenstone volcanics are granitoids with an age of 2800 ± 9 Ma, based on zircon xenocrysts in a migmatite south of Wongan Hills (Pidgeon et al., 1990; Wilde, 1990).

Terrane assembly

The above data have been integrated with the terrane model of Myers (1993), and the results have been presented in Wilde et al. (1996). The Lake Grace Terrane is unique, being the only region characterized by charnockitic granitoids. Its relationship to the Murchison Terrane is unclear, since it appears to cut across this terrane at a high angle close to the Wongan Hills Greenstone Belt (Fig. 3). However, it may be significant that both these terranes show evidence of migmatization, regional metamorphism, and possibly granitic intrusion at c. 2.8 Ga (Wilde, 1990). The major difference is that the greenstone sequences in the Lake Grace Terrane formed in the interval 2.8 to 2.7 Ga, whereas those in the Murchison chiefly formed at 3.0 Ga (Pidgeon and Wilde, 1990) and are amongst the oldest in the craton.

The Lake Grace Terrane has close similarities with the Yellowdine and Barlee Terranes to the east, in what was originally referred to as the Southern Cross Province by Gee et al. (1981). The youngest greenstones in the Yilgarn Craton (2.7 to 2.65 Ga) are present both in the Boddington Terrane in the southwest and further east within the Kalgoorlie, Kurnalpi, and Gindalbie Terranes (the Norseman–Wiluna Belt of Gee et al., 1981). There is thus a repetition of greenstone ages across the craton, contradicting the earlier view of Gee et al. (1981) that implied eastward younging of events away from the western margin.

There are insufficient precise dates available on granitoids throughout the Yilgarn Craton to make meaningful comparisons regarding the nature and timing of granite magmatism in the different terranes. However, it is pertinent to note that the youngest Archaean granitoids in the Yilgarn Craton crystallized at about 2580 Ma and have so far been identified in the Balingup Terrane (Partington, 1990) and in the Lake Grace Terrane (Wilde and Pidgeon, 1987). In the Balingup Terrane, there are also large plutons of porphyritic granite, including the Logue Brook Granite with a U–Pb zircon age of 2612 ± 5 Ma (Compston et al., 1986), which is considerably younger than dates so far obtained from the bulk of the granitoids elsewhere in the craton. The view was originally put forward by Gee et al. (1981) that there was a widespread granite emplacement event at about 2660 Ma that led to cratonization of the Yilgarn Craton. This must now be questioned, since U–Pb zircon data from Nemchin and Pidgeon (1997) showed a range of crystallization ages from 2690 to 2616 Ma. Furthermore, there is a marked contrast between granites of the same age, with 2640 Ma high-grade charnockitic granitoids in the Lake Grace Terrane and c. 2640 Ma ‘low-grade’ to unmetamorphosed granites in the Boddington Terrane.

The existence of major terrane boundaries in the southwestern Yilgarn Craton can also be used to explain the apparent paradox regarding the timing of high-grade metamorphism in the region. Upper amphibolite- to granulite-facies metamorphism took place at 3180 Ma in the Jimperding Metamorphic Belt near Toodyay (Nieuwland and Compston, 1981), at approximately 2800 Ma at Wongan Hills (Pidgeon et al., 1990), and at 2640 Ma near Lake Grace (Wilde and Pidgeon, 1987; Nemchin et al., 1994). It has long been an enigma as to why regional metamorphism at such high grades was apparently localized and had no apparent effect on adjacent areas. Evidence discussed here indicates that these events took place within separate terranes (the Boddington, Murchison and Lake Grace Terranes, respectively) that had independent histories prior to collision and accretion during the Late Archaean.

The exact timing of terrane accretion is not precisely constrained at present. The simplest interpretation of the geophysical data presented in Figure 2 is that the terranes were successively thrust from east to west over an earlier granitic basement (Middleton et al., 1995). The marked contrast between the Lake Grace and Boddington Terranes indicates they did not attain their present relative configuration until after the 2640 Ma granulite-facies event. The boundary between the Boddington and Balingup Terranes is more difficult to define, since the relationship is obscured by Tertiary laterite. However, it is defined in the extreme south by a linear belt of quartz monzonite (Fig. 3), which is over 100 km long and less than 10 km in width (Wilde, 1990). It appears that the three large bodies of porphyritic granite lie close to this boundary and the age of 2612 ± 5 Ma for the Logue Brook Granite (Compston et al., 1986) may provide the minimum age for accretion. The boundary shown in Figure 3 corresponds closely with a line separating Rb–Sr biotite ages of 500 to 430 Ma in the west from 2600 to 2300 Ma ages to the east, as determined by Libby and de Laeter (1979) and de Laeter and Libby (1993). The mineral ages are not reflected in the Rb–Sr whole-rock ages, which are consistently greater than 2300 Ma, even at the western edge of the craton. De Laeter and Libby (1993) interpreted this change in mineral age as being the result of cooling during uplift in the Early Palaeozoic. However, they did recognize that it could have resulted from the presence of a major crustal boundary.

Table 2. Timing of major Archaean events in the southwestern Yilgarn Craton within the Balingup, Boddington, Lake Grace, and Murchison Terranes

Age (Ma)	*Event*
BALINGUP TERRANE	
2527	Emplacement of Sn–Ta–Li Greenbushes Pegmatite during amphibolite-facies metamorphism and sinistral shearing along Darling Fault Zone
2577	Amphibolite-facies metamorphism and shearing (dextral ?) along Darling Fault Zone, plus local granitoid emplacement
2612	Emplacement of Logue Brook Granite near Darling Fault
2740	Emplacement of Gibralter Quartz Monzonite along eastern margin of the Balingup Terrane
2838	Medium-pressure amphibolite-facies metamorphism
3070–2830	Development of Chittering and Balingup Metamorphic Belts. Trough sedimentation along evolving continental margin
BODDINGTON TERRANE	
2560–2530	Static upper amphibolite-facies metamorphism in the Jimperding Metamorphic Belt at Toodyay
2677–2640	Emplacement of main ‘Darling Range’ granitoids
2671–2654	Formation of Saddleback Group greenstones
c. 2800	Low-pressure metamorphism
3177–3100	Deposition of shelf sediments in Jimperding Metamorphic Belt
3735–3340	Early Archaean craton. Variety of granitoid source rocks contributing zircons to Jimperding quartzites near Toodyay
LAKE GRACE TERRANE	
2587	Post-tectonic granitoid emplacement
2640	Emplacement of charnockitic granitoids
2649–2640	Granulite-facies metamorphism in the Lake Grace Terrane
2790	Earliest development of greenstones
MURCHISON TERRANE	
2800	Granitic activity near Wongan Hills, possibly associated with migmatization and upper amphibolite-facies metamorphism
3008	Development of greenstones at Wongan Hills, possibly accompanied by granitoid emplacement

SOURCE: Modified from Wilde et al. (1996)

Crustal assembly of the eastern Yilgarn Craton was completed prior to 2411 Ma, since this is the age of the major east–west Widgemooltha Dyke Suite (Fletcher et al., 1987; Nemchin and Pidgeon, 1998). Mafic dykes in this suite traverse the Gindalbie, Kalgoorlie, Barlee, and Yellowdine Terranes (Myers, 1993). One of these dykes, the Binneringee Dyke (Wilde and Walker, 1982), extends through into the Lake Grace and Boddington Terranes, but appears to terminate abruptly to the southeast of Collie. It might be terminated by later Proterozoic strike-slip faulting or, alternatively, it could terminate at the boundary between the Boddington and Balingup Terranes. A summary of the timing of major events in the various terranes, based on Wilde et al. (1996), is presented in Table 2.

Excursion localities

The excursion will examine aspects of the Balingup and Boddington Terranes, including the Chittering and Jimperding Metamorphic Belts, the Morangup Greenstone Belt, and the post-tectonic granitoids. The excursion route and localities are shown on Figure 4, which is a simplified geological sketch map of the PERTH* 1:250 000 geological map sheet (Wilde and Low, 1978).

Balingup Terrane

Chittering Metamorphic Belt

Locality 1: Kyanite schist, South Chittering

One of the characteristic features of the Balingup Terrane is the development of medium-pressure Barrovian metamorphic assemblages in the pelitic units. Kyanite, staurolite, and sillimanite are widely distributed, particularly in the southern part of the Chittering Metamorphic Belt where pelitic schists are best developed. Much of the belt is composed of schist and granofels tectonically interleaved with orthogneiss. The sequence has been interpreted as resulting from trough-style sedimentation, possibly along a developing continental margin (Wilde, 1990).

It appears likely that medium-pressure metamorphism was associated with later tectonism related to the evolving western margin of the Yilgarn Craton. Blight et al. (1981) and Bretan (1985) have presented evidence for Archaean transcurrent movement along the ‘proto-Darling Fault’, an ancient deformation zone lying subparallel to the present Darling Fault. Evidence for this is present along the western margin of the Chittering Metamorphic Belt where metamorphic minerals, including garnet and amphibole, overprint the mylonitic fabric (Wilde and Low, 1978).

The unit of kyanite schist at this locality in South Chittering (AMG 414600E 6501700N) is lensoid and cannot be traced for more than a few kilometres to the north and south. The schistosity has a general trend of 170°, with a steep lineation plunging approximately 76° to the south. The rock is fine grained and composed mainly of quartz (60%), with biotite and kyanite making up most of the remainder. The proportion of biotite varies across the unit and shows an inverse relationship with kyanite, which forms ragged porphyroblasts. The biotite is a pale bleached variety with some alteration to chlorite and muscovite. Rounded detrital zircons are a common accessory mineral and there are also thin fibres of sillimanite (fibrolite) developed adjacent to the biotite flakes.

Locality 2: Granofels, Chittering Lake

Whereas schistose units are best developed in the southern part of the Chittering Metamorphic Belt, more massive granofels units are prominent in the central area, where they are locally interleaved with quartz–feldspar–biotite gneiss and thin layers of biotite-rich schist. The rock has been interpreted as an original greywacke metamorphosed to amphibolite facies (Wilde, 1990).

This disused quarry overlooking Chittering Lake (AMG 412900E 6522000N) demonstrates the main features of the granofelses in the area. It is a massive, bluish quartz–feldspar–biotite granofels, with a weak foliation trending 009/33°W. However, the foliation is indistinct and subordinate to a strong lineation that trends 001° and plunges 5°S. The rock is composed of an equigranular mosaic of microcline, andesine–

* Capitalized names refer to standard 1:250 000 map sheets, unless otherwise indicated.

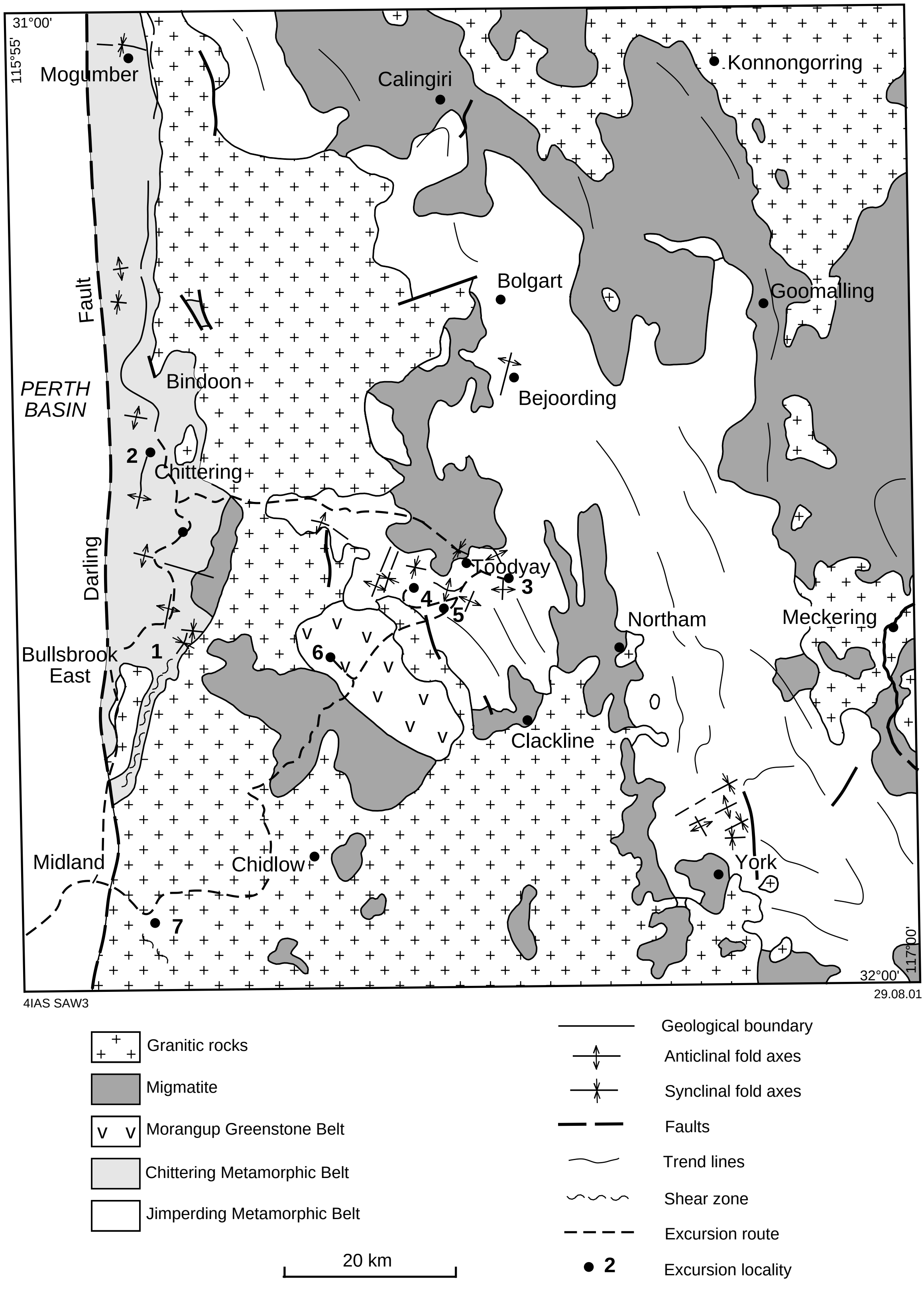

Figure 4. Simplified geological map of the PERTH 1:250 000 sheet, showing the excursion localities (adapted from Wilde and Low, 1978, and Wilde, 1990)

oligoclase, and quartz, with small interstitial flakes of dark-brown biotite that are aligned in the lineation. There are also some larger flakes of biotite and scattered hornblende crystals. Locally, there are bands richer in biotite and these help to define the weak foliation. There are also variations in the relative proportions of the two feldspars, with much of the plagioclase being untwinned. Pyrite is also present.

Boddington Terrane

Jimperding Metamorphic Belt

Locality 3: Windmill Hill railway cutting, Toodyay

A strongly weathered sequence of schist, gneiss and amphibolite is present at the eastern end of the cutting (AMG 454100E 6507500N). This passes westward into quartzite with a foliation trending 170/45°E. The quartzite consists of alternating flaggy and more massive units, with foliation surfaces sparsely coated with green chrome muscovite (fuchsite). There are some thicker layers up to 1 cm wide composed largely of fuchsite with minor sillimanite. Continuing westward, a 35 m-thick unit of well-foliated amphibolite is concordant within the quartzites. This rock is composed of a xenoblastic aggregate of calcic plagioclase and hornblende, with some quartz and biotite (showing chlorite alteration).

Further west, the quartzite is more flaggy and has a prominent lineation. There are some transgressive quartz and pegmatite veins, many of which are subhorizontal. Some excellent examples of herringbone cross-bedding are preserved in the quartzite close to where the cutting is widest and these indicate that the sequence is the correct way up. **Note this area is classified as a Geological Monument Site under the Heritage Act and the site must not be damaged or samples removed.**

An extensively weathered ultramafic intrusion, approximately 15 m wide, is present to the west. This rock is particularly unstable and collapsed during construction of the cutting. A subvertical dolerite dyke cuts the quartzites to the west and is unmetamorphosed. At the extreme western end, granitic gneiss is faulted against the quartzite.

Fifty individual zircon grains were analysed using the SHRIMP method from a sample of fuchsitic quartzite collected from the northern side of this railway cutting (P. D. Kinny quoted in Wilde, 1990). Zircons from this unit had previously been studied by Nieuwland and Compston (1981) using multigrain techniques. From analyses of four grain-size fractions, Nieuwland and Compston (1981) obtained a discordia line with an upper intercept age of 3341 +133/–73 Ma. They interpreted the good alignment of the analyses on the concordia diagram as indicating that the range in ages of rocks in the source area was rather small. However, it is clear from the ion microprobe analyses that the provenance ages span at least 550 m.y. and a heterogeneous provenance for the quartzite is indicated. Two-thirds of the detrital zircon populations have $^{207}Pb/^{206}Pb$ ages in the range 3350 to 3200 Ma (median 3270 Ma), whereas the remaining one third are older, having $^{207}Pb/^{206}Pb$ ages up to 3500 Ma. Additionally, one grain yielded an age of 3735 ± 10 Ma (2σ). Using these results, the positions of the four multigrain analyses of Nieuwland and Compston (1981) can be explained readily in terms of the admixture of all the above components identified by the ion probe, allowing for higher relative proportions of radiogenic Pb in the multigrain analyses to have been derived from discordant high-U zircon (P. D. Kinny quoted in Wilde, 1990).

The youngest concordant analysis of low-U zircon provides an upper limit to the time of deposition of the quartzite at 3177 ± 15 Ma (2σ). This is consistent with the c. 3100 Ma upper limit for the deposition of the Mount Narryer quartzite, Jack Hills metaconglomerate, and related metasedimentary rocks from the northern part of the Western Gneiss Terrain (Compston and Pidgeon, 1986; Kinny et al., 1990; Wilde et al., 2001), and suggests that the quartzites of the Jimperding Metamorphic Belt may be contemporaneous.

Locality 4: Poison Creek, Toodyay

The area was first mapped and described by Prider (1934). Later work (Prider, 1944) extended the mapping to include most of the Toodyay region, including the previous locality of Windmill Hill. Rocks from the Poison Creek section (AMG 439900E 6505600N) were also sampled by Nieuwland (1980) and are critical to Nieuwland and Compston's (1981) interpretation of their U–Pb zircon data.

Poison Creek is partially controlled by a fault line. The exposed rocks consist of a sequence of quartzites and orthogneisses. Within the quartzites are thin schistose units, some of which contain abundant sillimanite. There are also thin layers of banded amphibolite in which amphibole is far in excess of plagioclase.

The quartzites are fine to medium grained, with a xenoblastic granular texture. Accessory amounts of muscovite (locally fuchsite) and feldspar (both microcline and oligoclase) are present. The orthogneisses are medium to coarse grained, with variations in the intensity of the gneissic foliation. Most have an augen texture, with porphyroclasts of microcline strongly aligned in the fabric. The original rock was a porphyritic granite, as microcline is in excess of plagioclase. Biotite is the mafic mineral phase and is commonly altered to chlorite.

One of the most important features of the Poison Creek section is the exposed contact between orthogneiss and quartzite. The orthogneiss is intrusive, being discordant to the bedding and extending as veins into the quartzite. In addition, xenoliths of quartzite are present within the orthogneiss close to the contact. Later deformation has lessened the discordance and the contact is now subparallel to bedding in the quartzite. The granite appears to have been intruded as sheets (Prider, 1944).

Using conventional multigrain zircon U–Pb isotopic data, Nieuwland and Compston (1981) defined an age of c. 3250 Ma for the igneous precursor of the orthogneiss This would constrain the minimum depositional age of the supracrustal succession. However, unpublished data of P. D. Kinny (1998, pers. comm.) indicates that this age is most likely the result of contamination by older quartzite xenoliths and the true age of intrusion is c. 2.6 Ga.

Locality 5: Andalusite schist, Lovers Lane

This locality (AMG 439900E 6503700N) provides an excellent example of andalusite schist within the Jimperding Metamorphic Belt. In this region it is associated with units of quartzite and banded iron-formation. The rock is composed of large, bluish porphyroblasts of andalusite up to 2 cm in length, set in a reddish matrix rich in mica, with quartz and minor feldspar and sillimanite. The rock has a strong schistosity that trends 092/10°S. The presence of andalusite plus sillimanite in the Jimperding Metamorphic Belt contrasts with the kyanite–sillimanite assemblage that characterizes the aluminous pelites in the Chittering Metamorphic Belt. The low-pressure metamorphic conditions in the Jimperding Metamorphic Belt have resulted in the local development of cordierite at several localities near Toodyay.

Locality 6: Morangup Greenstone Belt, Morangup Hill

The Morangup Greenstone Belt was discovered in 1989 by S. A. Wilde and is a further extension of Late Archaean mafic and felsic volcanism in the western part of the Yilgarn

Craton (Wilde and Pidgeon, 1990). It closely resembles the Saddleback Group at Boddington, with both sequences being metamorphosed to greenschist facies.

The belt is extremely poorly exposed, with this outcrop (AMG 433600E 6498100N) being the largest known. It is composed of fine-grained, blue-grey metabasalt with a brown weathering rind. The rock consists of tremolite–actinolite, albite, and epidote, with accessory opaque oxides and titanite. The texture is variable, ranging from static (relict igneous) to more dynamic with strongly aligned blades of amphibole. Most amphibole laths are ragged and show acicular growth into neighbouring epidote. The opaque oxides and titanite are present as small granules preferentially associated with tremolite–actinolite. Where relict igneous textures are preserved, granular epidote pseudomorphs the original plagioclase. Small quartz–epidote veins are abundant in the metabasalt and are accompanied by brecciation in small outcrops to the east of Morangup Hill.

The mineralogy of the basalts is consistent with greenschist-facies metamorphism and the rocks are thus at a lower grade than those in the Jimperding Metamorphic Belt a few kilometres to the north and east. The exact nature of the contact between the two belts is unknown but, based on an analogy with the Saddleback Group at Boddington and photo-interpretation, it is likely that the Morangup Greenstone Belt has been downfaulted into the older Archaean basement.

Locality 7: Mountain Quarry, Boya

Quarries in the Boya–Darlington area were amongst the earliest to be opened up in the Darling Range to supply aggregate for the Perth market. All have been abandoned for many years and the exposed surfaces are variously discoloured.

The rocks are typical of the large Darling Range Batholith (Wilde and Low, 1978; Nemchin, 1996) located near the western margin of the Yilgarn Craton. A detailed examination reveals that a number of textural phases are present. The earliest is a fine-grained, grey, mesocratic granodiorite composed of plagioclase, quartz, and microcline, with ragged interstitial biotite. The plagioclase is andesine to oligoclase in composition and has been sericitized. The quartz has undulose strain extinction and has also undergone marginal granulation. Microcline forms large anhedral crystals that enclose both the quartz and plagioclase.

The grey granodiorite is cut by a medium- to coarse-grained, leucocratic granitoid that ranges from granodiorite to monzogranite in composition. This tends to be slightly finer grained near the contacts and also encloses xenoliths of the earlier granodiorite. The younger granitoids are richer in microcline, which is present as more subhedral, poikilitic crystals enclosing plagioclase and quartz. The quartz has a strong undulose extinction, but is generally devoid of marginal granulation. The amount of biotite varies and allows two subtypes to be identified: a leucocratic variety with only minor chloritized biotite and a more mesocratic type containing flakes and aggregates of greenish biotite. These subtypes are locally in sharp contact and may form subhorizontal bands and schlieren. A few thin, subhorizontal pegmatite veins are also present and appear to have been emplaced along joints.

Wilson et al. (1960) obtained a Rb–Sr whole-rock age of 2700 Ma from the Mountain Quarry (AMG 411800E 6468400N), whereas biotite separates gave a range of ages from 548 to 452 Ma (Libby and de Laeter, 1979). The younger biotite ages indicate resetting, possibly related to dolerite dyke emplacement. The age of dolerite intrusion is not precisely known, although palaeomagnetic evidence presented by Giddings (1976) indicates at least three periods of intrusion in the Boya area. The earliest set of dykes was emplaced more than 1700 m.y. ago and these have been partially remagnetized by a later set that were emplaced approximately 1500 m.y. ago. In these quarries, the youngest suite was emplaced between 750 and 700 m. y. ago.

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