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Geology and Tectonic Evolution of the Archean North Pilbara Terrain, Pilbara Craton, Western Australia

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Abstract

Results from a multidisciplinary geoscience program since 1994 are summarized for the North Pilbara terrain of the Pilbara Craton. Major findings include the recognition of three separate terranes with unique stratigraphy, geochronological, and structural histories; the ca. 3.72 to 2.85 Ga East Pilbara granite-greenstone terrane, the ca. 3.27 to 2.92 Ga West Pilbara granite-greenstone terrane, and the ≤3.29 Ga Kuranna terrane in the southeast. These are separated by two late, dominantly clastic sedimentary basins deposited within tectonically active zones; the ca. 3.01 to 2.93 Ga Mallina basin in the west and the undated Mosquito Creek basin in the east.

The oldest supracrustal rocks are the ca. 3.51 to 3.50 Ga Coonterunah and ca. 3.49 to 3.31 Ga Warrawoona Groups in the East Pilbara granite-greenstone terrane, deposited on fragments of older sialic crust to 3.72 Ga. The Warrawoona Group is subdivided into three main (ultra)mafic-felsic volcanic cycles including from base to top, the Talga Talga (3.49–3.46 Ga), Salgash (3.46–3.43 Ga), and newly defined Kelly (3.43–3.31 Ga) Subgroups. These dominantly basaltic rocks include chert beds containing Earth's oldest stromatolites and are interbedded with significant felsic volcanics erupted intermittently from 3.49 to 3.43 Ga during emplacement of sheeted sodic granitoid sills. Estimates of autochthonous stratigraphic thickness range from 9 to 18 km. Deformation involved extensional growth faulting, local folding, and tilting of greenstones away from synvolcanic granitoid domes. Rapid partial convective overturn of upper and middle crust occurred at 3.32 Ga during voluminous potassic felsic magmatism, followed by deposition of the Budjan Creek Formation at 3.31 Ga.

Granitoid plutonism at ca. 3.29 Ga in the Kuranna terrane preceded deposition of ultramafic through felsic volcanics and chert in the West Pilbara granite-greenstone terrane (3.27–3.25 Ga Roebourne Group) and western margin of the East Pilbara granite-greenstone terrane (3.26–3.24 Ga Sulphur Springs Group). Geochemical and isotopic data suggest that volcanism resulted from plume-related rifting of the East Pilbara granite-greenstone terrane, which was accompanied by granitoid plutonism and deformation. Following this was ca. 100 m.y. of relative quiescence during which locally economic concentrations of banded iron-formation and siliciclastics of the Gorge Creek Group were deposited in the East Pilbara granite-greenstone terrane.

Thereafter, geologic events are more consistent with microplate tectonics, commencing with deformation at 3.15 Ga followed by deposition of 3.13 to 3.11 Ga bimodal volcanics in the West Pilbara granite-greenstone terrane (Whundo Group), which have juvenile Nd isotope signatures and thus may represent either a rift or island-arc succession. Basaltic rocks and minor felsic tuff were deposited in the East Pilbara granite-greenstone terrane at 3.06 Ga and possibly in the West Pilbara granite-greenstone terrane (Regal Formation). At 3.02 Ga, the Whundo and Roebourne Groups share a common history of deposition of banded iron-formation and granitoid plutonism across the Sholl shear zone, suggesting accretion at, or immediately preceding, this time. This was followed by deposition in the Mallina basin of the volcanic Whim Creek Group at 3.01 Ga, possibly as an arc, and then the 2.97 to 2.93 Ga volcanic Bookingarra (west) and clastic De Grey (east) Groups during periods of intracontinental rifting interspersed with compression and granitoid intrusion. The geochemistry of 2.95 Ga high Mg diorites (sanukitoids) indicates a previous episode of subduction during either the Whundo or Whim Creek Groups or both. Final events include emplacement of ultramafic-mafic layered intrusions (2.925 Ga in the West Pilbara granite-greenstone terrane), local shearing and lode Au mineralization (2.92 Ga in the West Pilbara granite-greenstone terrane, 2.90 Ga in the Mosquito Creek basin, 2.89 Ga in the East Pilbara granite-greenstone terrane), and intrusion of fractionated, Sn-Ta-Li-bearing granites to 2.85 Ga (East Pilbara granite-greenstone terrane).

Introduction

THE OVOID Pilbara Craton of Western Australia is composed of two principal components: a Paleo- to NeoArchean (3.72–2.85 Ga) basement of granites and greenstones, which is exposed in several inliers including a large area in the north of the craton known as the North Pilbara terrain,[1] and unconformably overlying volcano-sedimentary rocks of the NeoArchean (2.77–2.40 Ga) Mount Bruce Supergroup deposited in the Hamersley basin (Griffin, 1990; Fig. 1). In this paper, we provide a geologic framework of the North Pilbara terrain for studies relating to specific mineralization styles described in this issue.

This overview is based on the results of 1:100 000 scale mapping, geochronology, geophysics, and geochemistry undertaken since 1994 as part of a National Geoscience Mapping Accord project between the Geological Survey of Western

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[1] In this paper, terrain refers to a geographical area, whereas terrane is used to indicate a distinct geologic element.

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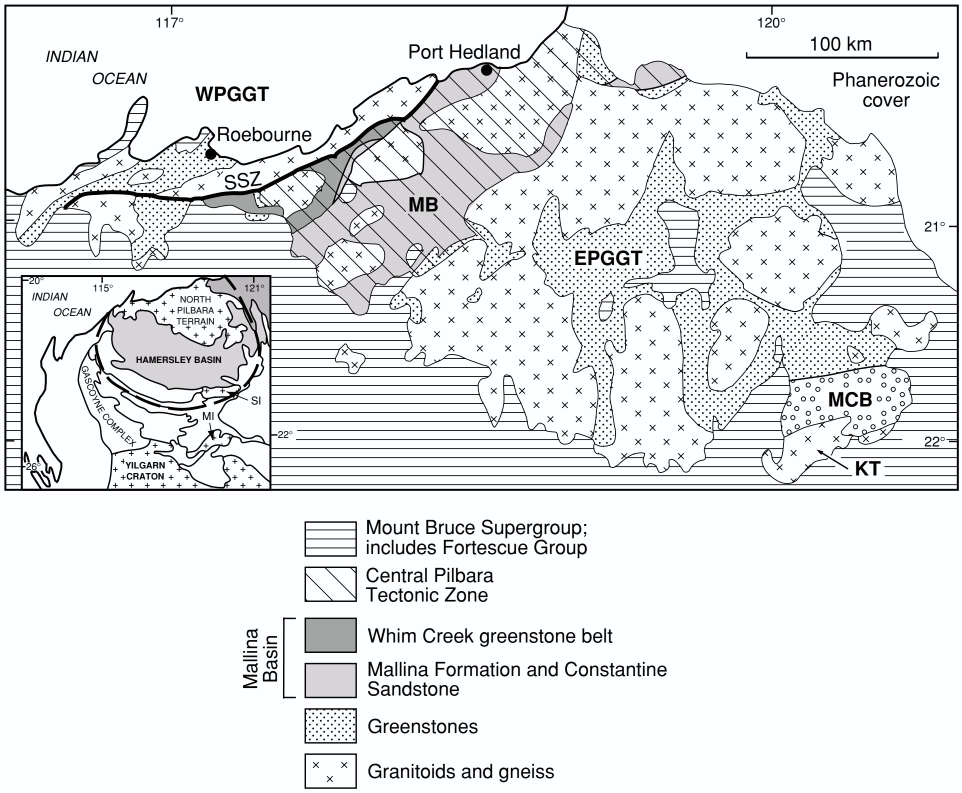

FIG. 1. Simplified geologic map of the North Pilbara terrain, showing component features discussed in the text. EPGGT = East Pilbara granite-greenstone terrane, KT = Kuranna terrane, MB = Mallina basin, MCB = Mosquito Creek basin, SSZ = Sholl shear zone, WPGGT = West Pilbara granite-greenstone terrane. Inset shows the position of the Pilbara Craton in Western Australia (heavy dashed line) and the North Pilbara terrain: MI = Marymia inlier of the Yilgarn Craton, SI = Sylvania inlier of the Pilbara Craton.

Australia and Geoscience Australia, in cooperation with Monash University and the University of Newcastle. The project has involved over 36 geoscientist years of work since 1995.

New data indicate that the North Pilbara terrain is composed of three separate granite-greenstone terranes with distinct stratigraphy and unique geochronological and structural histories, separated by sedimentary basins formed in active tectonic zones. Based on this data, the North Pilbara terrain is subdivided into five lithotectonic elements (Fig. 1): (1) the East Pilbara granite-greenstone terrane, which represents the ancient crustal nucleus of the Pilbara Craton that formed between 3.72 to 2.85 Ga; (2) the ca. 3.27 to 2.92 Ga West Pilbara granite-greenstone terrane; (3) the intervening, ca. 3.01 to 2.94 Ga Mallina basin of volcanic and sedimentary rocks with associated granitoid rocks and deformation at ca. 2.95 to 2.93 Ga; (4) the little-known Kuranna terrane in the southeast, which is composed largely of granitoid rocks dated at ca. 3.3 to 3.2 Ga and separated from the East Pilbara granite-greenstone terrane by; (5) the Mosquito Creek basin of undated turbiditic clastic rocks deformed into a fold and thrust belt at ca. 2.9 Ga.

Overviews of the stratigraphy, magmatic history, and structure of each of these elements are presented, followed by a discussion of the tectonic evolution. The following authors contributed detailed information for individual areas, based on their mapping; A.H.H. and R.H.S. for the West Pilbara granite-greenstone terrane, R.H.S. and G.P. for the Mallina basin, M.V.K. and A.H.H. for the East Pilbara granite-greenstone terrane. D.R.N. provided age dates for samples collected by the mapping geologists. A lithostratigraphic approach is used in this paper, as results from the National Geoscience Mapping Accord project do not support previous sequence stratigraphic schemes. All sets of structures (D_1, etc.) described below refer only to the local hierarchy within the five identified lithotectonic elements, even though some structure sets transgress lithotectonic element boundaries. This paper represents a summary of work in progress as our knowledge of some areas—particularly the Mosquito Creek basin and the Kuranna terrane—is as yet incomplete and will be tested through further studies founded on mapping conducted through to 2002. The evolution of the Mount Bruce Supergroup is outside the scope of this paper, but readers interested in the Fortescue Group should refer to detailed studies by Blake (1993) and Thorne and Trendall (2001).

Previous Interpretations of the North Pilbara Terrain

Hickman (1983, 1984) provided a regional lithostratigraphic interpretation of the Pilbara block, based on reconnaissance geologic mapping during the 1970s, and presented a tectonic model whereby the major tectonic structures resulted from essentially solid-state granitoid diapirism. Without precise geochronology, stratigraphic correlations between greenstone belts led to the conclusions that a relatively uniform lithostratigraphic succession was present in the east Pilbara, and the lower, dominantly volcanic part of that succession (Warrawoona Group) was also preserved in the west. Subsequent geochronology has shown that the second conclusion was incorrect (Horwitz and Pidgeon, 1993; Hickman, 1997a; Nelson, 1997, 1998; Smith et al., 1998), and also that greenstones of the east Pilbara are more laterally variable than previously envisaged (e.g., Van Kranendonk and Morant, 1998).

During the 1980s, many individual research projects were undertaken in the East Pilbara granite-greenstone terrane, yielding a variety of interpretations of stratigraphy and structural geology. Objections to the regional lithostratigraphic interpretation arose from interpretations of structural duplication due to horizontal tectonics (Bickle et al., 1980, 1985; Boulter et al., 1987). Since these early studies, the operation of horizontal tectonics in various parts of the North Pilbara terrain has been advocated by several workers (Ohta et al., 1996; Zegers et al., 1996; Kiyokawa and Taira, 1998; van Haaften and White, 1998). Nevertheless, other studies have continued to provide strong supporting evidence of diapirism in the East Pilbara granite-greenstone terrane (Collins, 1989; Collins et al., 1998; Collins and Van Kranendonk, 1999; Van Kranendonk and Collins, 2001, in press) that was interrupted by a period of sinistral transpression at ca. 2940 Ma (Van Kranendonk and Collins, 1998; Zegers et al., 1998).

Horwitz and Krapez (1991) and Krapez (1993) applied sequence stratigraphy to the North Pilbara terrain, the latter suggesting that much of the stratigraphic succession has been tectonically repeated and intercalated, that some units might be allochthonous, and that several tectonostratigraphic cycles might be present. The terrain was divided into five tectonostratigraphic domains separated by northeast trending lineaments, following Krapez and Barley (1987). These lineaments were stated to form the dominant structural fabric of the Pilbara Craton, and have a long history of development and reactivation. Krapez (1993) concluded that the Pilbara stratigraphy formed in response to a steady state, global tectonic regime related to supercontinent cycles.

Barley (1997) summarized the evolution of the Pilbara Craton in terms of westerly continental growth through accretion, based on modern plate tectonic analogues. Smith et al. (1998) presented geochronological data in support of two lithotectonic domains within the West Pilbara granite-greenstone terrane, separated by the sinistral Sholl shear zone. Smith et al. (1998) and Krapez and Eisenlohr (1998) used this data and new geochronology from the west Pilbara to develop a model of NeoArchean continental growth involving the tectonic accretion of an outboard island arc (Roebourne Group and related granitoid rocks) and a back-arc basin (Whundo Group) onto the East Pilbara granite-greenstone terrane.

Krapez and Eisenlohr (1998) further developed the sequence stratigraphic model, interpreting two Megacycle Sets spanning 3500 to 2775 Ma, which they divided into four Megacycles of 190 to 175 m.y. duration each. Each Megacycle was inferred to contain a Megasequence that could be divided into supersequences or basins. These authors reiterated that the domain boundaries have a long history of development and reactivation, but admitted that evidence for any pre-3000 Ma history was only "…cryptic, and related to interpreted stratigraphic patterns" (Krapez and Eisenlohr, 1998, p. 177). In contrast, Van Kranendonk and Collins (1998) interpreted that the stratigraphy was continuous across one of the proposed domain boundaries in the East Pilbara granite-greenstone terrane (i.e., the Lalla Rookh-Western Shaw structural corridor) and that it represented a relatively late zone of intracontinental transpression. Barley and Pickard (1999) suggested that a period of voluminous felsic magmatism at ca. 3315 Ma in the East Pilbara granite-greenstone terrane was the result of extension analogous to the Basin and Range of the North American Cordillera.

East Pilbara Granite-Greenstone Terrane

The East Pilbara granite-greenstone terrane represents the ancient cratonic nucleus of the North Pilbara terrain, with a geologic history spanning 870 m.y., from 3.72 to 2.85 Ga. It is composed of some of the best-preserved, oldest rocks on Earth, contains evidence for the oldest life on Earth, is host to some of the oldest mineral deposits on Earth, and represents the type example of a dome and basin map pattern that is unique to Archean terrains. The East Pilbara granite-greenstone terrane is also well known because of the ongoing controversy regarding the tectonic evolution that gave rise to this map pattern.

The East Pilbara granite-greenstone terrane is characterized by large (35–120-km-diam in map view), ovoid, domical granitoid complexes (Griffin, 1990) and flanking, curviplanar, generally synclinal tracts of generally steeply dipping volcano-sedimentary rocks collectively referred to as greenstones (Fig. 2). Greenstone belts are defined as relatively well preserved tracts of generally coherent greenstone stratigraphy bounded by faults, intrusive and/or sheared intrusive contacts with granitoid complexes, unconformities with granitoid rocks, and unconformably overlying supracrustal rocks of the Fortescue Group and younger cover. Greenstone complexes are areas of high structural complexity in which a coherent stratigraphy is generally lacking and faulting and/or shearing is common. Geochronological, structural, and gravity data indicate that granitoid complexes are nearly vertical cylinders extending to depths of 14 km and represent structural domes of a continuous mid-crustal granitoid layer that is linked beneath greenstone synclines (Hickman, 1984; Collins et al., 1998; Wellman, 2000).

Lithostratigraphy

The Pilbara Supergroup of the East Pilbara granite-greenstone terrane is formally divided into five volcano-sedimentary groups and two formations that include, from base to top (Fig. 3): the ca. 3.51 to 3.50 Ga Coonterunah Group; the ca. 3.49 to 3.31 Ma Warrawoona Group; the ca. 3.26 to 3.24 Ga Sulphur Springs Group; the undated Gorge Creek Group deposited between 3.24 Ma and 2.94 Ga; the ca. 2.94 Ga De Grey Group. The ca. 3.31 Ga Budjan Creek Formation lies unconformably on the Wyman Formation in the Kelly greenstone belt and is unconformably overlain by the Gorge Creek Group. The undated Golden Cockatoo Formation in the Yule Granitoid Complex is bounded by faults and intrusive contacts with granitoid rocks but is probably related to deposition of the adjacent Sulphur Springs Group.

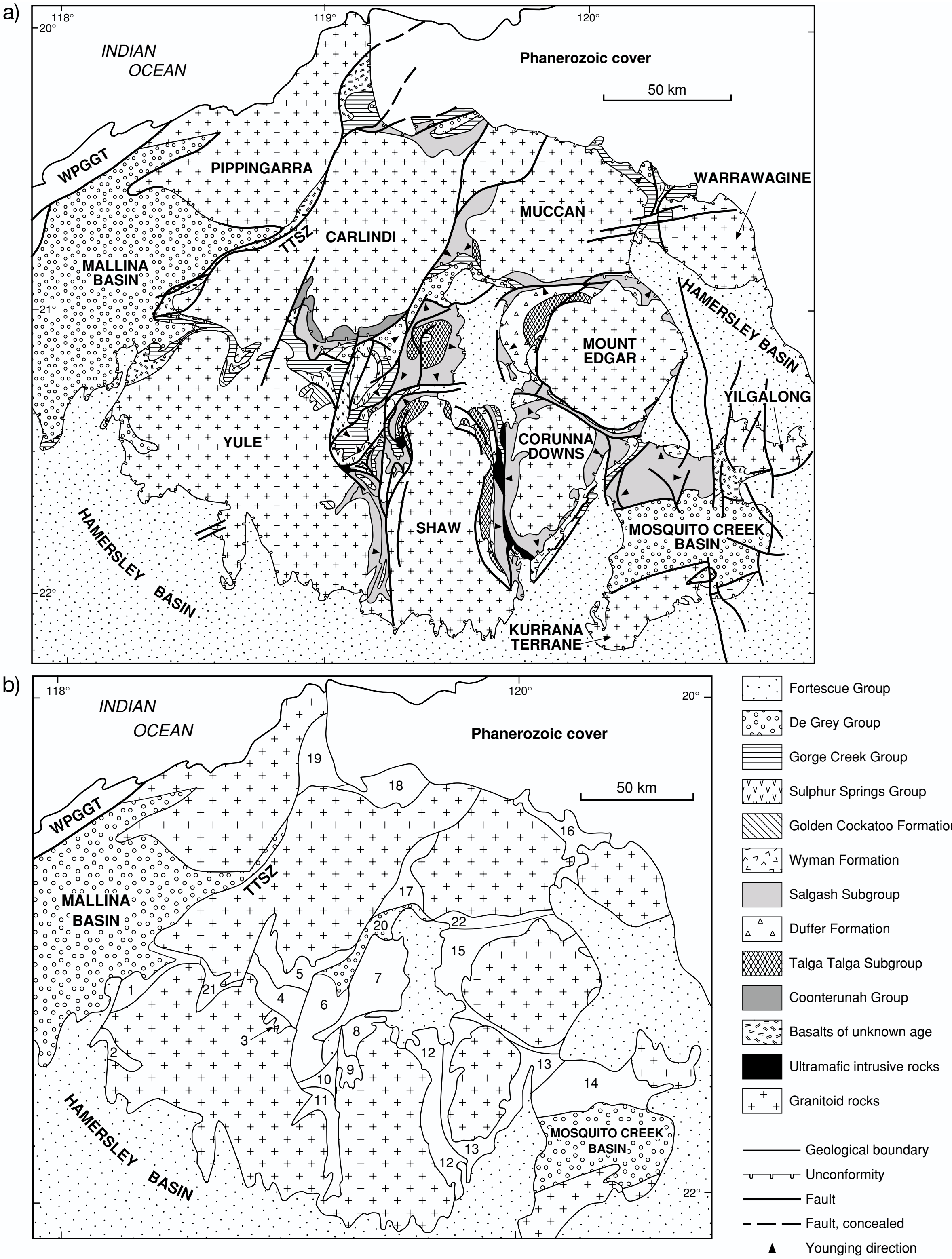

FIG. 2. a. Generalized geologic map of the East Pilbara granite-greenstone terrane. Bold names are of granitoid complexes. WPGGT = West Pilbara granite-greenstone terrane. b. Greenstone belts and complexes: 1 = Pilbara Well greenstone belt, 2 = Cheearra greenstone belt, 3 = Abydos greenstone belt, 4 = Pincunah greenstone belt, 5 = East Strelley greenstone belt, 6 = Soanesville greenstone belt, 7 = Panorama greenstone belt, 8 = North Shaw greenstone belt, 9 = Emerald Mine greenstone complex, 10 = Tambina greenstone complex, 11 = Western Shaw greenstone belt, 12 = Coongan greenstone belt, 13 = Kelly greenstone belt, 14 = Yilgalong greenstone belt, 15 = Marble Bar greenstone belt, 16 = Shay Gap greenstone belt, 17 = Warralong greenstone belt, 18 = Goldsworthy greenstone belt, 19 = Ord Range greenstone belt, 20 = Lalla Rookh Synclinorium, 21 = Wodgina greenstone belt, 22 = Dooleena Gap greenstone belt.

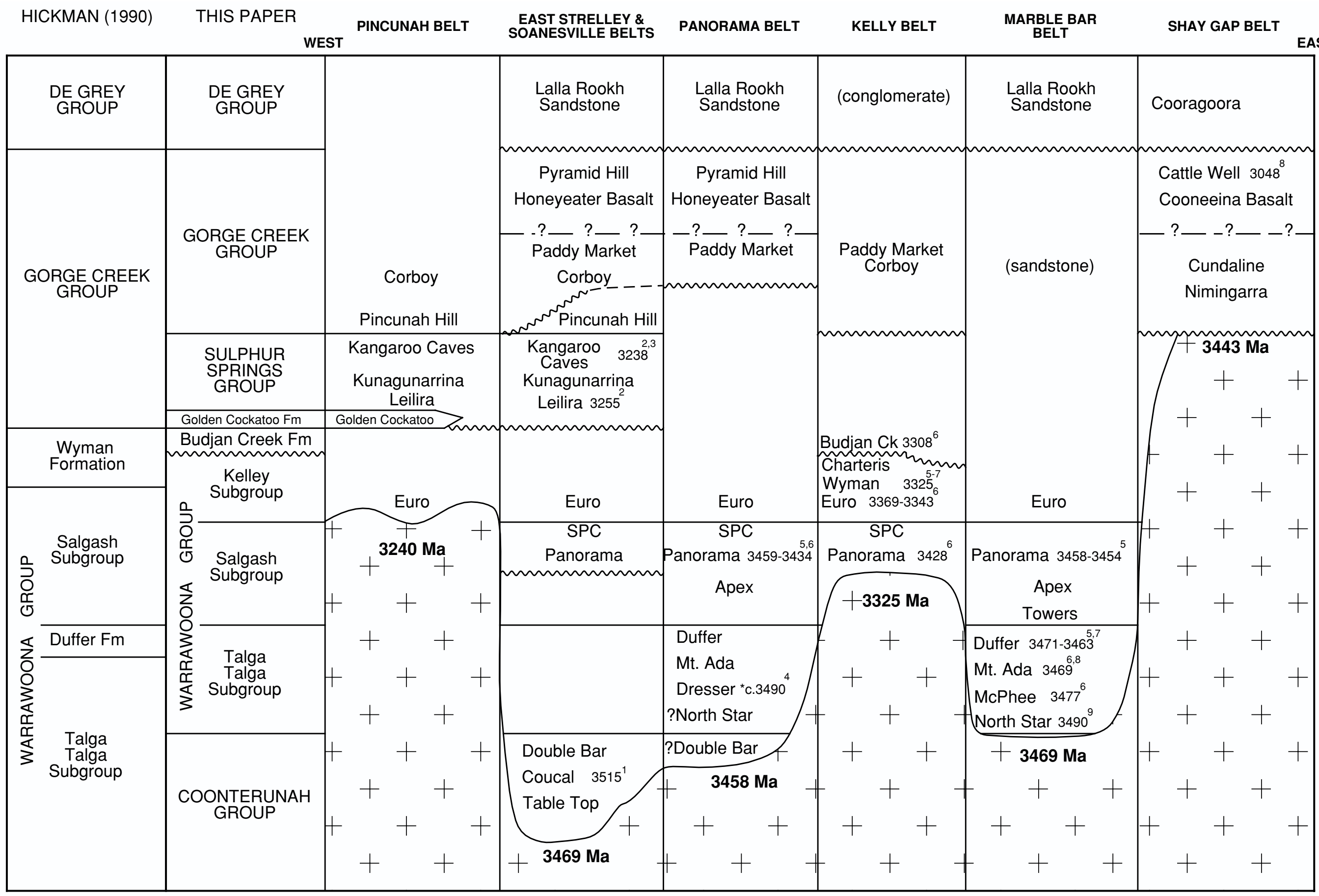

FIG. 3. Schematic stratigraphic profiles of greenstone belts across the East Pilbara granite-greenstone terrane. Age data are U-Pb zircon ages except * = Pb-Pb isochron age on galena. Age date references: 1 = Buick et al. (1995), 2 = Brauhart (1999), Buick et al (2002), 3 = R. Thorpe, Geological Survey of Canada (writ. commun., 1999), 4 = Thorpe et al. (1992b), 5 = Thorpe et al. (1992a), 6 = Nelson (1998), 7 = Nelson (2001, 2002), 8 = Nelson (1999), 9 = van Koolwijk et al. (2001).

Coonterunah Group: The Coonterunah Group (Buick et al., 1995; Van Kranendonk and Morant, 1998; Van Kranendonk, 2000) is a ≤5.9-km-thick succession of dominantly tholeiitic basalt and subordinate intermediate to felsic volcanic rocks distributed around the southern flank of, and intruded by, the Carlindi Granitoid Complex (Fig. 2). This group is defined on the basis of a local high-angle unconformity with the upper part of the Warrawoona Group and U-Pb zircon ages of 3515 ± 3 and 3498 ± 2 Ma (Buick et al., 1995; Nelson, 2002). Two Ar-Ar dates of ca. 3500 Ma on hornblende from amphibolite-facies metabasalts (Davids et al., 1997; Zegers et al., 1999) suggest that the Coonterunah Group may be more widespread than presently recognized.

Warrawoona Group: The Warrawoona Group consists predominantly of basaltic rocks that are interlayered, and locally interfinger, with intermediate to felsic volcanic rocks in a series of repeated (ultra)mafic-felsic volcanic cycles (Barley, 1981; Hickman, 1983; Van Kranendonk et al., 2001b). Glikson and Hickman (1981) and Glikson et al. (1987) provided geochemical analyses of the volcanic rocks of the group, which also contains numerous and lithologically varied chert beds (Buick and Barnes, 1984). Barley et al. (1979) studied the volcano-sedimentary facies of the group and showed that at least some of it was deposited under shallow-water conditions.

Hickman (1977, 1983, 1990) subdivided the lithostratigraphy of the Warrawoona Group into two major basaltic divisions; the lower Talga Talga Subgroup and the overlying Salgash Subgroup (Fig. 3). These were separated by the ca. 3470 Ma felsic volcanic Duffer Formation and unconformably overlain by the felsic, ca. 3325 Ma Wyman Formation. Various units of intercalated thick chert and basalt in different greenstone belts were correlated with the Towers Formation, at the base of the Salgash Subgroup in the type area near Marble Bar, and used in craton-wide stratigraphic correlations.

Since then, additional geochronology and detailed mapping have shown the need for the six following major changes to the previous lithostratigraphic subdivisions (Fig. 3):

1. Chert assemblages, previously correlated as the Towers Formation, instead represent three separate, compositionally distinct units (Van Kranendonk, 2000). The oldest of these include the stromatolitic chert-barite units of the ca. 3490 Ma Dresser Formation in the Panorama greenstone belt (Van Kranendonk and Morant, 1998). Slightly younger is the ca. 3460 Marble Bar Chert Member of the Towers Formation in

the Marble Bar greenstone belt (Hickman, 1983). The youngest is the Strelley Pool Chert, a stromatolitic unit now recognized across the East Pilbara granite-greenstone terrane that lies conformably on rocks of the 3458 to 3426 Ma Panorama Formation (Figs. 3, 4; Van Kranendonk, 2000).

2. The ca. 3474 to 3463 Ma Duffer Formation is locally interbedded with, and the same age as, the upper part of the Mount Ada Basalt (Nelson, 2000) and is thus included in the Talga Talga Subgroup.

3. The Towers Formation in the Marble Bar greenstone belt contains interbedded components of the Talga Talga Subgroup (Chinaman Creek and Marble Bar Chert Members representing the last products of the Duffer Formation) and the Salgash Subgroup (unnamed komatiitic basalts representing the onset of the Apex Basalt) and is thus regarded as a transitional formation between, and separate from, the two subgroups.

4. The Wyman Formation has a generally conformable, gradational contact on the Euro Basalt (Hickman, 1983; L. Bagas, Geological Survey of Westerdan Australia, unpub. data) and is thus included in the Warrawoona Group, as originally proposed (Hickman, 1983).

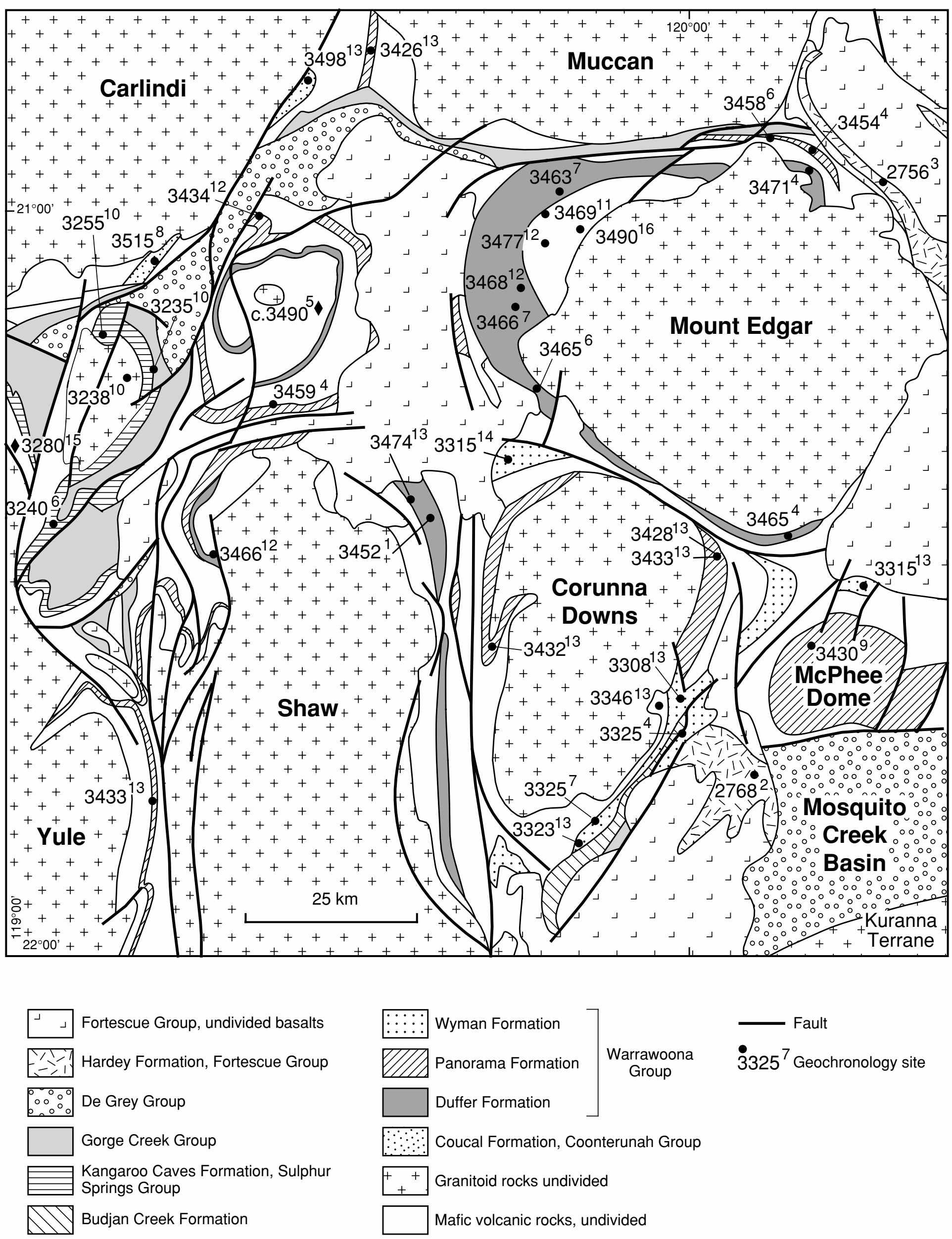

FIG. 4. U-Pb zircon and Pb-Pb age data on felsic volcanic units across the East Pilbara granite-greenstone terrane. References: 1 = Pidgeon (1978), 2 = Pidgeon (1984), 3 = Arndt et al. (1991), 4 = Thorpe et al. (1992a), 5 = Thorpe et al. (1992b), 6 = R. Thorpe, Geological Survey of Canada (writ. commun., 1999), 7 = McNaughton et al. (1993), 8 = Buick et al. (1995), 9 = Barley et al. (1998), 10 = Brauhart (1999), Buick et al. (2002), 11 = Nelson (1999), 12 = Nelson (2000), 13 = Nelson (2001, 2002), 14 = Age cited in Barley and Pickard (1999), 15 = P. Morant, Sipa Resources NL (pers. commun., 1999).

5. The Euro Basalt and Wyman Formation have been placed together in the newly defined Kelly Subgroup at the top of the Warrawoona Group in order to better reflect recurrent volcanic processes in the Warrawoona Group (L. Bagas and M. Van Kranendonk, Geological Survey of Westerdan Australia, unpub. data).

6. Mapping in the Kelly greenstone belt has shown that the Charteris Basalt lies conformably on the Wyman Formation (L. Bagas and I.R. Williams, Geological Survey of Westerdan Australia, pers. commun., 2001) and thus it is reassigned from the Gorge Creek Group (Hickman, 1983) to the top of the Warrawoona Group.

The type area of the Talga Talga Subgroup is in the Marble Bar greenstone belt (Hickman, 1983), where geochemical data (Glikson and Hickman, 1981; Hickman, 1983), several new age dates (Figs. 3, 4, and references therein), and structural mapping (A.H.H. *in* Van Kranendonk et al., 2001b) has confirmed that the mildly deformed, low-grade rocks represent a coherent, right way-up succession (Van Kranendonk et al., 2001a, b), rather than a downward-younging tectonic assemblage, as suggested by van Haaften and White (1998, 2001). The age data includes an Ar-Ar date of ca. 3490 Ma from hornblende in basalt from the stratigraphically lowest North Star Basalt (van Koolwijk et al., 2001), a 3477 ± 2 Ma age from a thin felsic tuff horizon in the conformably overlying McPhee Formation, six U-Pb zircon dates of between 3471 to 3463 Ma from andesite to rhyolitic rocks of the Duffer Formation, and two dates of 3458 to 3454 Ma on rhyolitic volcaniclastic rocks in the Panorama Formation (see Figs. 3, 4, for references).

Granitoid rocks intrude the basal contact of this subgroup, and thus the 8 km of stratigraphic thickness in the Marble Bar greenstone belt is a minimum. The Duffer Formation reaches a maximum of 5 km thick, ranging from andesite interbedded with basalt and intrusive dacite sills at the base, through dacite lava and pyroclastic units, to rhyolite, rhyolitic tuff, and reworked tuff and sandstone at the top. Geochemical data suggest the Duffer Formation derived from fractionation of a basaltic parent magma (Cullers et al., 1993), and the formation is host to syngenetic volcanic-hosted massive sulfide (VHMS) mineralization (Hickman, 1983; Thorpe et al., 1992a, b).

In the Panorama greenstone belt (North Pole dome), the lower part of the Talga Talga Subgroup includes the ca. 3490 Ma Dresser Formation, which contains stromatolitic, carbonate- and sulfate-bearing chert units interbedded with pillow basalt (age from Thorpe et al., 1992b; Nijman et al., 1998a; Van Kranendonk, 2000). Whereas it was previously thought that the cherts represent silicified carbonates that were deposited together with evaporative gypsum in a supratidal, quiet marine environment and then replaced by silica and barite, respectively (e.g., Groves et al., 1981; Buick and Dunlop, 1990), more recent studies have shown that most of the chert and barite were precipitated directly from hydrothermal fluids that erupted from swarms of chert-barite veins emplaced into, and above, active growth faults during the formation of a caldera over a synvolcanic laccolith (Nijman et al., 1998a; Ueno et al., 2001; Van Kranendonk, in press).

The Salgash Subgroup conformably overlies the Talga Talga Subgroup. The lowermost Apex Basalt reaches a maximum thickness of 3 to 4 km in the Panorama and Western Shaw greenstone belts. Elsewhere the Apex Basalt is either very thin or absent, and the Panorama Formation lies directly on the Duffer Formation (DiMarco and Lowe, 1989; M. Van Kranendonk, unpub. data). The rhyolitic succession of volcaniclastic rocks and lavas known as the Panorama Formation has now been identified in greenstones across the East Pilbara granite-greenstone terrane and dated at 3459 to 3426 Ma (Figs. 2-4). The formation is up to 1.5 km thick in its type area in the Panorama greenstone belt (DiMarco and Lowe, 1989; Van Kranendonk, 2000) but varies along strike to a thickness of only a few meters in some belts and is locally absent. Stratigraphic variations suggest it was erupted from several volcanic sources, including a vent exposed in cross section in the northern part of the Panorama greenstone belt (North Pole dome; Van Kranendonk, 2000). The North Pole Monzogranite in the core of the dome is the same age as extrusive rhyolite from the lower part of the Panorama Formation in the southern margin of the belt (Thorpe et al., 1992a), and their emplacement caused synvolcanic doming (DiMarco and Lowe, 1989). Related small porphyry stocks intruding the Apex Basalt in the North Pole dome are between 3449 to 3434 Ma and contain epithermal Au, Cu, Ag, and Bi mineralization (Marston, 1979; Thorpe et al., 1992a; Amelin et al., 2000; R. Thorpe, Geological Survey of Canada, writ. commun., 1992). Cullers et al. (1993) presented geochemical evidence that the Panorama Formation was derived from melting of eclogite.

The Strelley Pool Chert (Lowe, 1983) conformably overlies the Panorama Formation and the two are considered to be genetically related (Van Kranendonk, 2000, in press). The chert contains distinctive stratigraphic elements, including stromatolitic carbonate laminites (Hoffman et al., 1999), which can be correlated over a distance of 130 km across most of the East Pilbara granite-greenstone terrane (Van Kranendonk, 2000, in press).

The Euro Basalt conformably overlies the Strelley Pool Chert or Panorama Formation across the East Pilbara granite-greenstone terrane. It is up to 9.4 km thick across the southwestern flank of the Panorama greenstone belt (Van Kranendonk, 2000) and composed of dominantly basaltic volcanic rocks interbedded with numerous thin cherts. Typically, the Euro Basalt contains a basal unit, <3 km thick, of high Mg basalt to komatiite, which is overlain by tholeiitic basalt and then several units of alternating high Mg and tholeiitic basalts (e.g., Glikson et al., 1986; Van Kranendonk, 2000). A silicified felsic tuff ~1 km below the top of the formation in the Kelly greenstone belt contains zircon populations at ca. 3346 and 3363 Ma (Nelson, 2001), the former interpreted as the eruption age of the tuff and the latter interpreted to be a xenocrystic population (L. Bagas and M. Van Kranendonk, Geological Survey of Western Australia, unpub. data). These data, combined with ages of detrital zircons from unconformably overlying clastic rocks (Nelson, 1998, 1999) and stratigraphic data, suggest essentially continuous deposition of the Euro Basalt from at least ca. 3395 to 3325 Ma.

The 3325 to 3312 Ma, conformably overlying Wyman Formation is composed of 500 to 1,100 m of massive and columnar-jointed rhyolite, volcaniclastic tuff and sandstone, and shale (Hickman, 1983). It is preserved around the Corunna

Downs Granitoid Complex, with which it is partly coeval, and in the northern part of the McPhee dome (Figs. 2, 4; Pidgeon, 1984; McNaughton et al., 1993; Barley and Pickard, 1999; Nelson, 2000, 2002; I.R. Williams, Geological Survey of Western Australia, pers. commun., 2001). Either a fault or an unconformity with younger rocks generally defines the top of the formation, but in the McPhee dome it is conformably overlain by stromatolitic chert fed by hydrothermal black chert dikes (I.R. Williams, Geological Survey of Western Australia, pers. commun., 2000). In the northern part of the Kelly greenstone belt and in the Warrawoona area, komatiite and basalt of the Charteris Basalt overlie the formation.

Budjan Creek Formation: The ca. 3308 Ma Budjan Creek Formation (Noldart and Wyatt, 1962; L. Bagas, Geological Survey of Western Australia, unpub. data) is a 150- to 1,200-m-thick succession of conglomerate, sandstone, siltstone, wacke, and felsic volcanic rocks that unconformably overlies folded rocks of the Wyman Formation in the southeastern part of the Kelly greenstone belt. The formation dips shallowly to the southeast, away from the Corunna Downs Granitoid Complex, and is unconformably overlain by the Gorge Creek Group (Fig. 4).

Golden Cockatoo Formation: The Golden Cockatoo Formation is a highly folded sequence of amphibolite-facies quartzite, pelite, rhyolite, and banded iron-formation that is found only within the Abydos greenstone belt in the northeastern part of the Yule Granitoid Complex (Figs. 2 and 3; Van Kranendonk and Morant, 1998; Van Kranendonk, 2000). Map data show that these rocks were deposited unconformably on ca. 3450 Ma orthogneiss and deformed into a small-scale dome and basin pattern during intrusion of synkinematic granitoid rocks dated at ca. 3240 Ma, which cut the formation (M. Van Kranendonk, unpub. U-Pb SHRIMP zircon data). Although the precise depositional age of these rocks is unknown, they must be older than the crosscutting 3240 Ma granitoid rocks and younger than the Warrawoona Group (ca. 3312 Ma), which lacks significant siliciclastic sedimentary rocks.

Sulphur Springs Group: The Sulphur Springs Group is a dominantly volcanic succession that unconformably overlies the Euro Basalt in the western part of the East Pilbara granite-greenstone terrane (Soanesville, Pincunah, East Strelley, and Warralong greenstone belts) and may also underlie much of the Pilbara Well greenstone belt (Fig. 2; Van Kranendonk and Morant, 1998; Van Kranendonk, 2000). Whereas previously the group was divided into four formations, new map and geochemical data (Geological Survey of Western Australia, unpub. data) suggest that the lowermost Six Mile Creek Formation should be reassigned to the underlying Euro Basalt.

The maximum age for deposition of the group is ca. 3255 Ma, based on a SHRIMP age of detrital zircons in lithic wacke of the basal Leilira Formation (Buick et al., 2002). However, model Pb-Pb ages of ca. 3280 Ma on sulfide mineralization from small felsic intrusions near the base of the group suggest a possibly older maximum age of deposition (P. Morant, Sipa Resources Ltd., writ. commun., 1998). Conformably overlying rocks of the ≤2.4-km-thick Kunagunarrina Formation include komatiite, high Mg basalt, and chert. These rocks have distinctive, strongly light rare earth element (LREE) depleted profiles, with ε_{Nd} values of between +3.2 to –2.3, becoming more negative up stratigraphic section, indicating progressive crustal contamination (S.-S. Sun, Geoscience Australia, unpub. data). Conformably overlying this formation is up to 1.5 km of basalt-andesite, dacite, rhyolite, and chert of the Kangaroo Caves Formation (Vearncombe et al., 1998; Brauhart, 1999; Van Kranendonk, 2000), which has Nd $T_{(2\text{-stage})}$ model ages (Sun et al., 1995) of between 3507 to 3467 Ma (Brauhart, 1999). Felsic volcanism was coeval with emplacement of the Strelley Granite laccolith, both of which have been dated at ca. 3238 Ma (Buick et al., 2002). Intrusion of the laccolith drove hydrothermal circulation that caused widespread alteration of overlying rocks, silicification of epiclastic sediments at the top of the formation, and deposition of syngenetic Zn-Cu massive sulfide mineralization at intervals of 5 to 8 km around the granite (Morant, 1998; Vearncombe et al., 1995, 1998; Brauhart et al., 1998; Brauhart, 1999; Van Kranendonk, 2000). Vearncombe et al. (1995, 1998) interpreted sulfide mineralization at Sulphur Springs as Earth's oldest black smoker. Late-stage caldera collapse caused local deposition of a coarse olistostrome breccia at the top of the formation. This was accompanied by late, volumetrically minor felsic volcanism at ca. 3235 Ma (Brauhart, 1999).

Gorge Creek Group: The Gorge Creek Group is an extensive, but as yet undated, component of many greenstone belts across the East Pilbara granite-greenstone terrane (Fig. 2) and reaches a maximum thickness of 3,500 m in the Soanesville greenstone belt (Van Kranendonk, 2000). The group lies conformably to unconformably on a variety of older supracrustal rocks and granitoid rocks (Dawes et al., 1995; Williams, 1999; Van Kranendonk, 2000). It is characterized by a lower succession of clastic rocks (Soanesville Subgroup), including thick ferruginous shale and/or locally economic concentrations of banded iron-formation at the base (Yarrie, Goldsworthy, and Ord Range deposits), overlain by sandstone and shale. Deposition of this succession occurred in a basin undergoing horst and graben faulting in the Pincunah greenstone belt (Wilhelmij and Dunlop, 1984).

In the Soanesville greenstone belt, basal units of the group disconformably overlie relict topography in the Sulphur Springs Group and have been affected by alteration related to Sulphur Springs mineralization (Vearncombe et al., 1998; Van Kranendonk, 2000). Slightly younger model Pb-Pb ages of ca. 3220 Ma obtained from some of the galenas dated from the Sulphur Springs deposit (see Huston et al., 2002) may reflect the depositional age of these lower sedimentary rocks. Elsewhere, the age of the Gorge Creek Group is constrained only by the age of underlying units, such as the 3308 Ma Budjan Creek Formation, and by the youngest age of contained detrital zircon populations, which is 3362 ± 12 Ma (Williams, 1999).

Higher up in the Gorge Creek Group, across what may represent a significant time gap, are the Honeyeater and Cooneeina Basalts. Whereas the Honeyeater Basalt is conformable on the Paddy Market Formation in the Soanesville greenstone belt, the Cooneeina Basalt lies unconformably on correlative rocks in the northeastern part of the Marble Bar greenstone belt (Williams, 1999). Basaltic volcanism was accompanied by the intrusion of thick, locally differentiated, ultramafic and mafic sills of the Dalton Suite that hosts PGE minerals and Ni sulfides (Van Kranendonk, 2000). Pb-Pb dates of ca. 3070 Ma on some galena samples from the ca. 3240 Ma Sulphur Springs deposit (see Huston et al., 2002) may date the age of this mafic magmatism, as the only possible source of heat and fluids that could have reset, or introduced new, galena into this synvolcanic VMS deposit are thick, differentiated ultramafic sills of the Dalton Suite. The Honeyeater and Cooneeina Basalts are disconformably overlain by felsic tuffs, tuffaceous sandstones, and ferruginous shales of the undated Pyramid Hill Formation of the Gorge Creek Group, and ca. 3048 Ma felsic tuff of the Cattle Well Formation of the De Grey Group, respectively (Nelson, 1999; Williams, 1999; Van Kranendonk, 2000).

De Grey Group: Scattered, isolated synclines and synclinoria of coarse clastic rocks belonging to the De Grey Group lie unconformably on older rocks across the East Pilbara granite-greenstone terrane (Fig. 2; Wilhelmij and Dunlop, 1984; Hickman, 1990; Williams, 1999; Van Kranendonk, 2000). The largest of these is the Lalla Rookh Synclinorium, located along strike northeast of the Soanesville greenstone belt and deposited in a fault-bound basin (Krapez, 1984; Krapez and Barley, 1987) during intracontinental sinistral transpression at ca. 2940 Ma (Van Kranendonk and Collins, 1998). Regional mapping has shown that at least some of the scattered synclinoria formed part of a larger basin during the onset of late, regional deformation (Van Kranendonk and Collins, 1998; Van Kranendonk, 2000).

Granitoid complexes

Nine ovoid, domical granitoid complexes occur in the East Pilbara granite-greenstone terrane (Fig. 2). These form structural domes with generally outward dipping margins, although some dip steeply inward. The contacts of granitoid complexes vary from being locally intrusive (e.g., all of Corunna Downs, northern Shaw, southern Carlindi, northeastern Mount Edgar), or the locus of shearing (e.g., southwestern Mount Edgar, western and eastern Shaw), to an unconformity with younger supracrustal rocks (e.g., Shay Gap greenstone belt; Dawes et al., 1995). Whereas granitoid complexes have simple outlines, they may have a complex and even chaotic internal geometry (e.g., Shaw Granitoid Complex; Van Kranendonk and Collins, in press). In many cases, older granitoid components are preserved along the margins of the complexes, whereas successively younger phases occupy the core (e.g., Van Kranendonk and Collins, 1998; Champion and Smithies, 2000). In the Yule and Mount Edgar Granitoid Complexes, roof pendants of amphibolite and ca. 3450 Ma orthogneiss form trains of enclaves separating lobes of younger material (Hickman, 1984; Collins, 1989; Champion and Smithies, 2000).

All mapped complexes contain gneissic to foliated remnants of ca. 3600 to 3420 Ma tonalite-trondhjemite-granodiorite in addition to younger, more potassic suites dated at 3321 to 3303, 3252 to 3242, ca. 2935, and ca. 2850 Ma (Hickman, 1983; Williams and Collins, 1990; Collins, 1993; Barley and Pickard, 1999; Nelson, 1999, 2000, 2001, 2002; Champion and Smithies, 2000). The complexes contain greatly varying proportions of the different age suites (Fig. 5). Of particular significance is the presence of ancient gneisses in Warrawagine (Williams, 2000), the large volume of ca. 3490 to 3410 tonalite-trondhjemite-granodiorite in the Shaw, the large volume of ca. 3315 Ma granite in Corunna Downs, and the dominance of ca. 2935 Ma monzogranites in the Yule, Carlindi, and Pippingarra Granitoid Complexes. It is significant, too, that plutonic phases coincide with episodes of felsic volcanism (Hickman, 1983; Williams and Collins, 1990; Thorpe et al., 1992a; Barley and Pickard, 1999; Brauhart, 1999), as are the large gaps between felsic plutonic/volcanic events, especially the 300 m.y. gap between ca.3240 and 2940 Ma.

The 3490 to 3420 Ma tonalite-trondhjemite-granodiorite suite is silicic and sodic and derived through high-pressure melting of a mafic source (Smithies, 2000). Bickle et al. (1983, 1993) compared the tonalite-trondhjemite-granodiorite suite favourably to modern calc-alkaline, subduction-related arc suites, but Smithies (2000) showed that the compositions are incompatible with an origin from slab melting in a modern-style steep subduction setting and suggested instead that tonalite-trondhjemite-granodiorites were more likely generated by melting of lower mafic crust. Bickle et al. (1993) presented geochemical evidence that at least some of the suite was derived from melting of crustal sources >3600 Ma old. Bickle et al. (1989) and Collins (1993) showed that the younger, more potassic suites were derived through progressive episodes of melting of the >3420 Ma tonalite-trondhjemite-granodiorite suite. Porphyry Cu-Mo mineralization is associated with the ca. 3315 Ma Coppin Gap pluton in the Mount Edgar Granitoid Complex (Hickman, 1983). The ca. 2850 Ma suite of highly fractionated monzogranites and pegmatites emplaced into the cores of some complexes contains Sn-Ta-Li mineralization that was mined extensively from placer deposits since the 1890s (Hickman, 1983; Collins and Gray, 1990; Nelson, 1998; Kinny, 2000).

Distribution of ancient crust

Indications of ancient sialic crust in the East Pilbara granite-greenstone terrane were first detected from Sm-Nd studies of Warrawoona Group volcanic rocks, which returned model isochron ages of 3560 ± 32 Ma (Hamilton et al., 1981) and 3712 ± 98 Ma (Gruau et al., 1987). Inherited and detrital zircon studies from a variety of different host rocks support the presence of ancient sialic crust during sedimentation and magmatism across the East Pilbara granite-greenstone terrane (Fig. 6). The oldest date of >3724 Ma is from a xenocrystic zircon from the Panorama Formation in the Panorama greenstone belt (1 in Fig. 6; Thorpe et al., 1992a). The oldest dated rock is an inclusion of tightly folded, gneissic tonalite from the Warrawagine Granitoid Complex (Williams, 2001), which yielded zircon populations at 3655 ± 6, 3637 ± 12, 3595 ± 4, and 3576 ± 6 Ma (the probable igneous age), in addition to populations with typical Warrawoona ages and an Nd model age of ca. 3.65 Ga (2 in Fig. 6; Nelson, 1999). The rest of the zircon data indicate peak crust formation times at ca. 3650, 3575, and 3530 to 3500 Ma, including gabbroic anorthosite at 3578 ± 4 Ma (McNaughton et al., 1988). Nd-depleted mantle (T_{DM}) model ages from ca. 3.45 Ga tonalite-trondhjemite-granodiorite from the Carlindi, Shaw, Warrawagine, and Yule Granitoid Complexes fall in the range ca. 3700 to 3564 Ma (Bickle et al., 1993; Geological Survey of Western Australia, unpub. data), indicating contamination by, and in part derivation from, older crustal

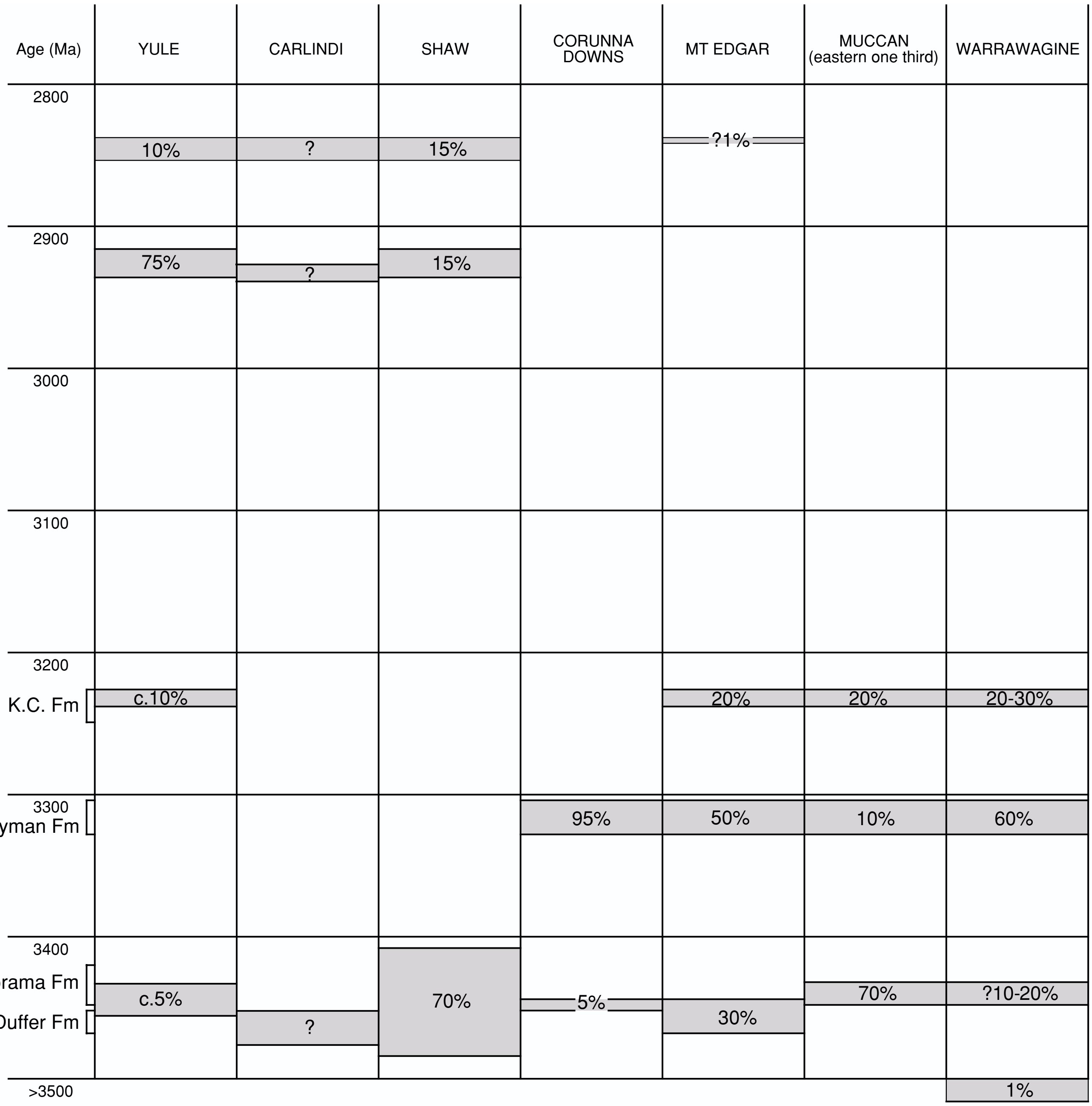

FIG. 5. Estimated volume and age of components in granitoid complexes of the East Pilbara granite-greenstone terrane. Age range of felsic volcanic components shown for reference in age column. K.C. = Kangaroo Caves Formation of the Sulphur Springs Group.

sources. The occurrence of ca. 3650 and 3580 Ma detrital zircons in a quartzite from the Apex Basalt (W.J. Collins, University of Newcastle, unpub. data) indicate that parts of the ancient sialic crust were exposed by ca. 3460 Ma.

Structural geology

Two types of regional structures occur in the East Pilbara granite-greenstone terrane; a dome and basin pattern comprising granitoid-cored domes (Hickman, 1984) and a 5- to 15-km-wide, north-south–striking, curvilinear zone of complex folds and faults referred to as the Lalla Rookh-Western Shaw structural corridor, which transects the East Pilbara granite-greenstone terrane (Fig. 7; Van Kranendonk and Collins, 1998; Zegers et al., 1998).

Structural domes include a core of granitoid rocks and a rind of greenstones that are attached to the granitoid rocks by intrusive or sheared intrusive contacts. Ring faults in the axes of intervening greenstone synclines separate the structural domes (Fig. 7). Granitoid complexes typically contain areas of migmatitic gneiss with evidence of multiphase deformation (e.g., Bettenay et al., 1981), but their margins commonly have a single foliation that dips steeply and is parallel to, and passes

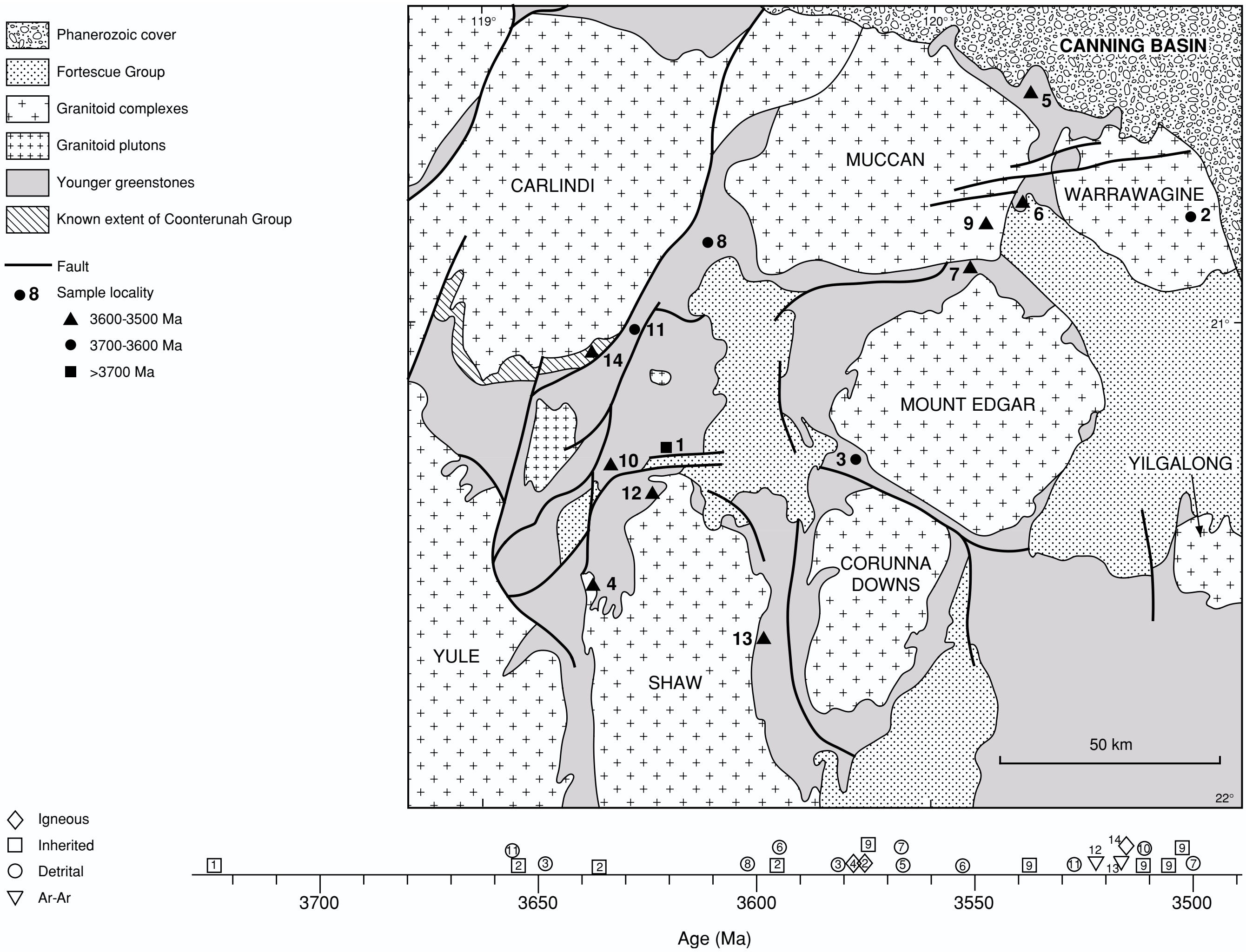

FIG. 6. Location and time line of samples with U-Pb zircon evidence of ancient crust in the East Pilbara granite-greenstone terrane. References: 1 = Thorpe et al. (1992a), 2 = Nelson (1999: sample 142870), 3 = W.J. Collins (University of Newcastle, written comm., 1996), 4 = McNaughton et al. (1988), 5 = Nelson (1999; sample 142867), 6 = Nelson (1998; sample 143995), 7 = Nelson (1998; sample 143994), 8 = Nelson (1998; sample 142836), 9 = Nelson (1998; sample 142828), 10 = Buick et al. (1995; sample 94001), 11 = Nelson (2000; sample 142951), 12 = Zegers et al. (1999), 13 = Davids et al. (1997), 14 = Buick et al. (1995; sample 70601).

across, contacts with greenstones (Hickman, 1983, 1984; Collins, 1989). Adjacent greenstones have shallow (e.g., Panorama and northwestern Marble Bar greenstone belts) to steep (e.g., East Strelley and Coongan greenstone belts) dips and generally display a single foliation. Farther away from granitoid contacts, greenstones commonly have subvertical dips and display a single foliation, although bedding-parallel high-strain zones may contain multiple, overprinting foliations (e.g., Blewett, 2002; Zegers et al., 2002).

A precise age is often difficult to ascribe to structural fabric elements in greenstones since the metamorphic minerals that define these elements have been largely reset by successive generations of intrusive granitoid phases or other regional events (e.g., Davids et al., 1997; Zegers et al., 1999). Attempts at dating synkinematic pegmatites from granitoid complexes have also proved frustratingly difficult, as they contain small, high U zircons unsuitable for dating by SHRIMP. As a result, some structures in the following section have been grouped together in broad events. Blewett (2002) presented an alternative interpretation of regional structural development.

D1 (ca. 3490–3410 Ma): D_1 structures include local tilting of the Coonterunah Group and the Talga Talga Subgroup away from the cores of granitoid complexes prior to deposition of the Panorama Formation and Apex Basalt, respectively (Hickman, 1983; Van Kranendonk, 2000). D_1 folds in the Coonterunah Group verge away from the Carlindi Granitoid Complex and predate deposition of the unconformably overlying Strelley Pool Chert (Van Kranendonk and Collins, 1998).

D_1 structures also include synvolcanic, listric, normal growth faults in the Dresser Formation (Nijman et al., 1998a; Van Kranendonk et al., 2001b), the Duffer Formation (Zegers et al., 1996), and the Panorama Formation (Nijman et al., 2001). Growth faults in the Dresser Formation are filled by hydrothermal chert-barite veins and have been interpreted either to have formed as a result of craton-wide extension

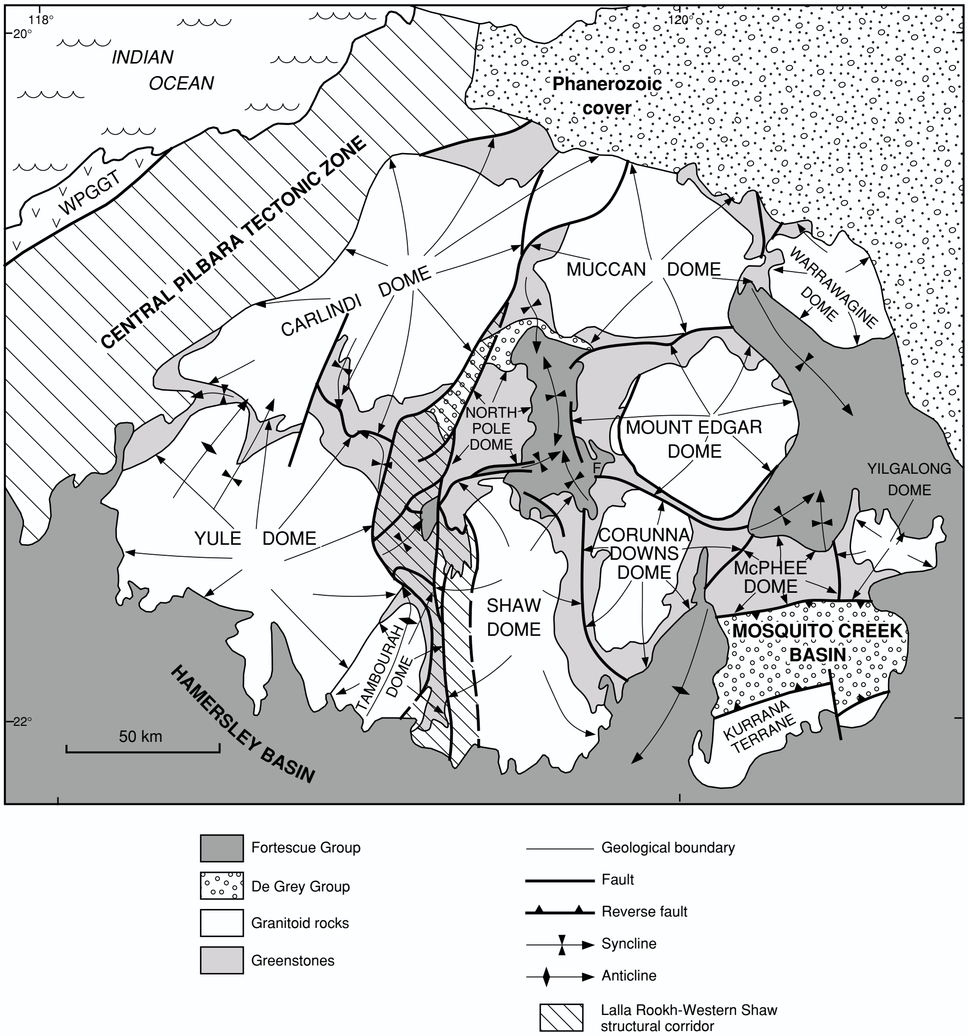

FIG. 7. Schematic map of large-scale structures of the East Pilbara granite-greenstone terrane. Bold names refer to granitoid-cored domes that formed throughout the development of the terrane. The sinistral transpressional Lalla Rookh-Western Shaw structural corridor formed at ca. 2940 Ma as a result of compression across the Central Pilbara tectonic zone and was coeval with deposition of the Lalla Rookh Sandstone.

(Nijman et al., 2001) or caldera collapse during local intrusion of a synvolcanic laccolith (Van Kranendonk, in press).

Complex folds, overturned bedding, and anomalously high-pressure metamorphic mineral assemblages (kyanite-grade, P = 6 kbars) occur in the northwest Shaw area, which were interpreted to represent D_1 structures resulting from Alpine-style thrusting that preceded doming of the Shaw Granitoid Complex (Bickle et al., 1980, 1985). Zegers et al. (1996) suggested that this thrusting occurred prior to 3467 Ma and that associated crustal thickening led to the formation of a core complex (Shaw Granitoid Complex) during orogenic collapse. This model was based on the interpretation of syndepositional extensional faults in the 3467 Ma Duffer Formation in the Coongan greenstone belt and of reverse kinematics in the Split Rock shear zone along the eastern margin of the Shaw Granitoid Complex. The shear zone was interpreted to have formed at 3467 Ma, synchronous with tonalite-trondhjemite-granodiorite plutonism, and to represent the detachment fault between the developing metamorphic core zone and the cover sequence (Zegers et al., 1996). The zone was shown to extend across the northern margin of the complex and to have sinistral kinematic indicators along this segment (but see "Discussion").

Gneissic fabrics and rootless isoclinal folds of leucosome veins in >3.45 Ga protoliths of the Shaw Granitoid Complex are also interpreted as D_1 structures, as they are cut by weakly foliated granitoid rocks dated at ca. 3.43 Ga (Van Kranendonk et al., 2001b). Partial melting in the Shaw Granitoid Complex continued to ca. 3400 Ma as determined from zircon ages on melt veins and diatexite (Van Kranendonk, 2000). A similar age of 3410 ± 7 Ma was obtained from a population of zircons from the >3600 Ma gneiss in the Warrawagine Granitoid Complex, which is interpreted to indicate the age of leucosome generation (Nelson, 1999).

D_2 (ca. 3315 Ma): These structures were generated between deposition of the Wyman and Budjan Creek Formations during intrusion of granitoid rocks at ca. 3315 Ma in the Corunna Downs and Mount Edgar Granitoid Complexes.

Collins (1989) and Collins et al. (1998) documented a set of structures within and around the Mount Edgar Granitoid Complex, which they related to diapirism (Fig. 8). These structures include a 1- to 3-km-wide, horseshoe-shaped shear zone within the southwestern margin of the complex that has granite side-up displacement indicators across which uplift of the granitoid complex was achieved (Collins et al., 1998). Dating of deformed granitoid rocks within the shear zone and of undeformed rocks that cut the shear zone indicate uplift and emplacement of the granitoid dome at ca. 3315 Ma (Williams and Collins, 1990; Collins et al., 1998). Collins and Gray (1990) presented geochronological evidence that

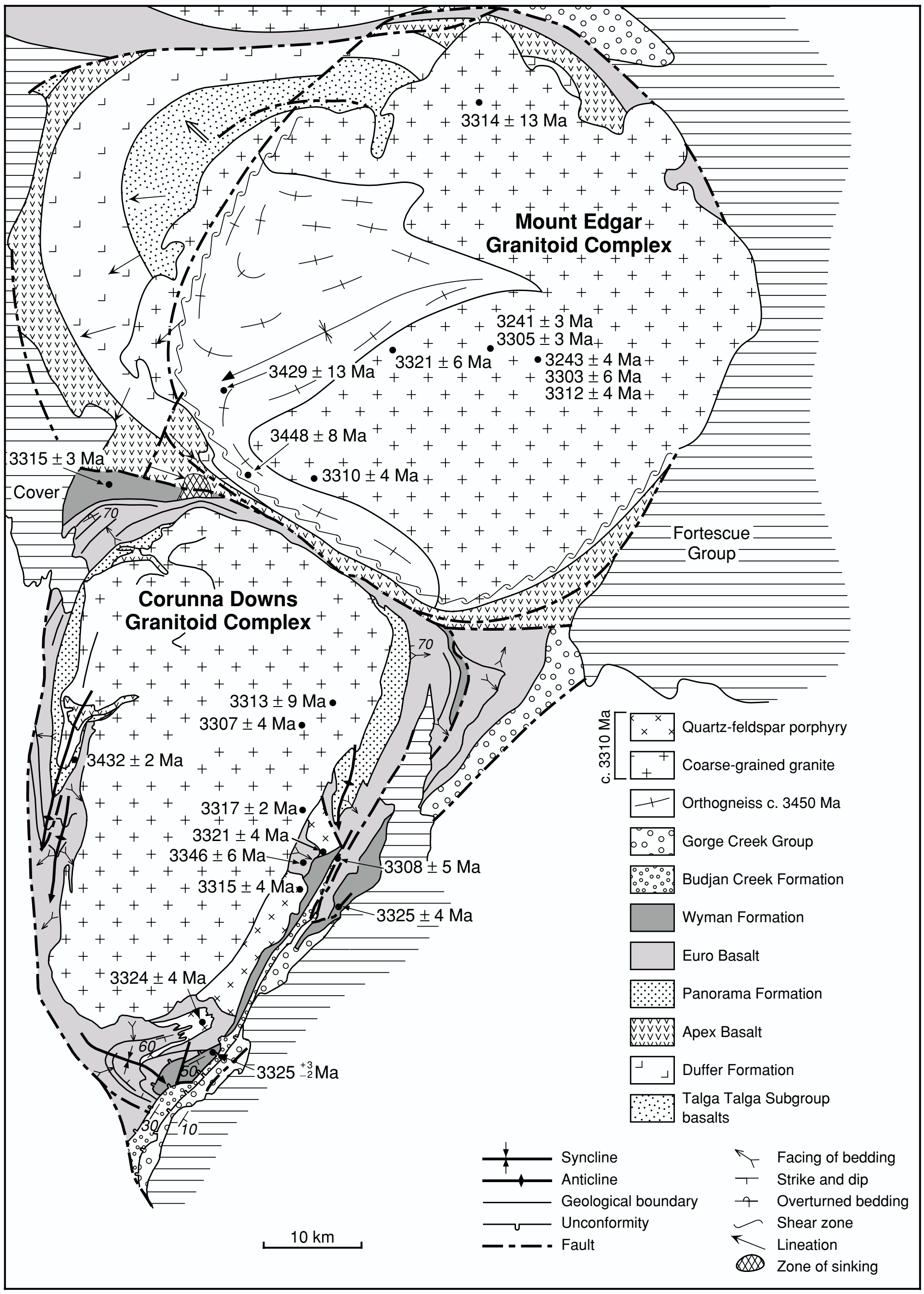

FIG. 8. Geologic sketch map of the Mount Edgar and Corunna Downs Granitoid Complexes, showing the main structures generated by granitoid diapirism (see text for discussion). Open arrow denotes area of northwest-verging recumbent folds in the McPhee Reward area, described by Collins (1989). Zone of sinking between the granitoid complexes denotes area of vertical L tectonites, described by Collins et al. (1998). Note the three hook-shaped folds within greenstones inside of the bounding ring fault of the Corunna Downs Granitoid Complex. Lineation trends from Collins (1989) and van Haaften and White (1998). Age data from Williams and Collins (1990), Thorpe et al. (1992a), McNaughton et al. (1993), Barley and Pickard (1999), and Nelson (2001).

doming of the Mount Edgar Granitoid Complex was largely completed by ca. 3200 Ma.

A set of tight, asymmetrical D_2 folds in the McPhee Formation of the Marble Bar greenstone belt verge away from the Mount Edgar Granitoid Complex to the northwest (Fig. 8). These folds have been interpreted as either having formed as a result of gravitational sliding of greenstones off the rising dome at ca. 3315 Ma (Collins, 1989) or by horizontal compression (van Haaften and White, 1998).

Greenstones in the Warrawoona syncline between the coeval Mount Edgar and Corunna Downs Granitoid Complexes were deformed into a D_2 zone of sinking characterized by a core of intense vertical stretching (Fig. 8; Collins et al., 1998), although Blewett (2002) interpreted these structures as the result of intersecting cleavages (see "Discussion"). Ring faults parallel to bedding in the flanking greenstones are the sites of epigenetic gold mineralization (see Huston et al., 2002) and are interpreted as accommodation structures to the uplifted granitoid complexes (Collins and Van Kranendonk, 1999).

Three large-scale, hook-shaped folds are present in greenstones around the Corunna Downs Granitoid Complex, bound by, but not affecting, a ring fault in the greenstones (Fig. 8). The synplutonic, D_2 age of the southern fold is well constrained as it affects the 3325 Ma Wyman Formation but is unconformably overlain by the Budjan Creek Formation, dated at 3308 Ma (Fig. 8). The saddle reef position of the fold in the northwest is occupied by weakly deformed, syntectonic granite with an axial planar foliation and a minimum age of ca. 3270 Ma (Rb-Sr isochron; Cooper et al., 1982), although mapping indicates this body is continuous with the voluminous ca. 3315 Ma suite dated by zircon and forming the bulk of the complex. The fold in the northeast is developed in rocks of the 3430 to 3325 Ma Euro Basalt but does not affect adjacent 3315 Ma porphyries or the 3308 Ma Budjan Creek Formation. Thus all three folds are considered coeval and to have formed between 3325 to 3308 Ma during emplacement of the Corunna Downs Granitoid Complex. After fold formation, episodes of younger tilting occurred after deposition of the Gorge Creek, De Grey, and Fortescue Groups, as indicated by progressively more shallow dips upsection (Fig. 8); this is also a feature common to greenstones flanking other domes.

Van Haaften and White (1998, 2001) described a set of bedding-parallel shears between steeply dipping formations of the Talga Talga Subgroup in the Marble Bar greenstone belt, which they interpreted to have formed prior to intrusion of granitoid rocks at ca. 3315 Ma and thus must be D_2 in age. The significance of these structures is reviewed in the "Discussion."

D_3 (ca. 3240 Ma): Intrusion of the Strelley Granite resulted in a suite of synvolcanic growth faults in the Kangaroo Caves Formation during hydrothermal circulation and massive sulfide deposition at 3240 Ma (Vearncombe et al., 1998). Late caldera collapse resulted in renewed faulting and deposition of an olistostrome breccia with synsedimentary slump folds (Van Kranendonk, 2000).

Sulphur Springs Group volcanism was accompanied by intrusion of granitoid rocks in the northeastern part of the Yule Granitoid Complex, including the Tambourah dome (Van Kranendonk, 1997; Van Kranendonk and Collins, 1998). Dome and basin folding of the Golden Cockatoo Formation accompanied granite intrusion, generating the same hook-shaped folds and ring faults as adjacent to the Corunna Downs Granitoid Complex (Van Kranendonk and Collins, in press). Doming was accompanied by amphibolite-facies contact metamorphism of the Western Shaw greenstone belt (Wijbrans and McDougall, 1987).

Deposition of the Gorge Creek Group occurred during the late stages of D_3 doming in the Pincunah greenstone belt, where the basin was affected by syndepositional horst and graben faults (Wilhelmij and Dunlop, 1984). Nijman et al. (1998b) described extensional and compressional structures in the Gorge Creek Group from the Marble Bar greenstone belt (see "Discussion").

D_4 (ca. 2940 Ma): Deformation affected much of the North Pilbara terrain at ca. 2940 Ma and resulted in the formation of the northerly striking Lalla Rookh-Western Shaw structural corridor in the East Pilbara granite-greenstone terrane and deposition of the De Grey Group (Fig. 7; Van Kranendonk and Collins, 1998). The Lalla Rookh-Western Shaw structural corridor is characterized by coeval sets of large, northeast-southwest–trending folds, north- to northwest–striking sinistral faults and shear zones, and northeast-striking dextral faults. Collectively, these structures indicate a northwest-southeast direction of maximum compression at this time (Van Kranendonk and Collins, 1998). Stratigraphic links across the Lalla Rookh-Western Shaw structural corridor show that it is an entirely intracontinental structure and not the site of terrane accretion as previously suggested by Krapez (1993) and Eriksson et al. (1994) (Van Kranendonk and Collins, 1998; Van Kranendonk, 2000). The shear zones of the Lalla Rookh-Western Shaw structural corridor are host to epigenetic Au deposits in the East Strelley and Western Shaw greenstone belts and emeralds in the Emerald mine greenstone complex (Van Kranendonk, 2000, 2002).

Two strands of sinistral shear formed along the western margin of the Shaw Granitoid Complex at this time and are known collectively as the Mulgandinnah shear zone (Van Kranendonk and Collins, 1998; Zegers et al., 1998). Linear fabric elements within the shear strands are colinear with lineations and fold axes in the intervening area of complex folds in the northwest Shaw area, and these structures have all been ascribed to D_4 deformation for the reasons discussed below (see "Discussion").

D_4 deformation also caused amplification of the Tambourah dome, low-grade metamorphism of adjacent greenstones (Wijbrans and McDougall, 1987), and development of a north-northeast–striking foliation defined by oblate quartz grains in granitoid rocks of the Yule and Shaw Granitoid Complexes (Van Kranendonk, 2002). Related monzogranite plutonism was widespread across the western part of the East Pilbara granite-greenstone terrane (Nelson, 1998, 2000) but is unknown east of the Shaw Granitoid Complex.

D_5 (ca. 2890 Ma): Sinistral shearing on north-northeast–striking zones affected the northwestern part of the East Pilbara granite-greenstone terrane at ca. 2890 Ma and was accompanied by epigenetic gold mineralization in the Mount York and Lynas districts (Neumayr et al., 1993). It is not known to what tectonic event this shearing relates, but it is the last deformation to affect the basement rocks prior to

cratonization marked by the emplacement of post-tectonic granitoids at 2850 Ma.

D_6 (ca. 2760 Ma): The unconformably overlying, predominantly basaltic volcanic rocks of the Fortescue Group record a late component of granitoid doming (Hickman, 1984). This is evidenced by the fact that outliers of the Fortescue Group in the East Pilbara granite-greenstone terrane are preserved only in synclines that are developed over older synclines in greenstones between domical granitoid rocks. Further evidence comes from paleocurrent data in the Hardey Formation in the Marble Bar sub-basin, which indicate transport of sand detritus radially inward to the basin center (Blake, 1993). Evidence for deformation during deposition of the group is preserved in the tight syncline between the Shaw and North Pole domes, where an angular unconformity is developed between tightly folded rocks of the basal Mount Roe Basalt and overlying, gently dipping rocks of the Hardey Formation (Van Kranendonk, 2000).

Metamorphism

Metamorphic temperatures in greenstones decrease away from granitoid complexes from lower amphibolite facies, to greenschist and prehnite-pumpellyite facies, to very low temperature metamorphism of the De Grey and Fortescue Groups. Two-pyroxene granulite remnants have been recorded in amphibolite in the Shaw and Yule Granitoid Complexes. The highest pressure estimates of ca. 6 kbars have been obtained from kyanite-bearing felsic schists in the northwestern Shaw area (Bickle et al., 1985) and southern Marble Bar greenstone belt (Delor et al., 1991). Whereas Bickle et al. (1985) suggested that assemblages in the northwestern Shaw area developed in response to thrusting, a thermal anomaly associated with the granitoid rocks was unexplained. Collins and Van Kranendonk (1999) showed that this problem can be resolved if the metamorphic rocks and related structures result not from thrusting but from partial convective overturn of the upper and middle crust and provided supporting data from the Warrawoona syncline.

Amphibolite-facies metamorphism of the Western Shaw greenstone belt was dated at ca. 3234 Ma and greenschist-facies metamorphism at ca. 2950 Ma (Wijbrans and McDougall, 1987)—the two dates ascribed to D_3 doming and D_4 transpression, respectively (Van Kranendonk and Collins, 1998). Results of extensive Ar-Ar dating of metamorphic amphiboles in greenstones show a spread of ages throughout the tectonic evolution of the East Pilbara granite-greenstone terrane, including end-member dates similar to those of the Coonterunah and Fortescue Groups (Davids et al., 1997; Zegers et al., 1999). Much of the rest of the Ar-Ar data show that greenstones adjacent to granitoid complexes cooled over the protracted development of the terrane.

West Pilbara Granite-Greenstone Terrane

The West Pilbara granite-greenstone terrane extends from the base of the Whim Creek Group (Mallina basin) in the east, to the unconformity with the Fortescue Group in the south and west (Fig. 1). It is comprised of a completely different stratigraphic succession than the East Pilbara granite-greenstone terrane and has a distinct tectonic style (Fig. 9). The oldest preserved rocks of the West Pilbara granite-greenstone terrane are undated basaltic rocks beneath 3270 Ma felsic volcanic rocks of the Roebourne Group and coeval granitoid rocks. The West Pilbara granite-greenstone terrane is also distinct because it contains the 3130 to 3115 Ma Whundo Group, the ca. 3020 Ma Cleaverville Formation, the undated Regal Formation, and 3130 to 3060 Ma granitoid rocks, for which there are no known equivalents in the East Pilbara granite-greenstone terrane.

The most distinctive tectonic feature of the West Pilbara granite-greenstone terrane is a northeasterly structural grain, defined by the trend of granitoid complexes and greenstone belts, and by numerous closely spaced easterly and northeasterly striking faults. These include the kilometer-wide Sholl shear zone that bisects the West Pilbara granite-greenstone terrane and has a long history of reactivation (Fig. 9). Granitoid rocks are structurally relatively simple and discordant to the greenstone stratigraphy, in contrast to the East Pilbara granite-greenstone terrane where granitoid rocks crop out in broadly circular domical complexes and where greenstone belts have no preferred structural trend.

Lithotectonic elements

Previously, Krapez and Eisenlohr (1998) and Smith et al. (1998) recognized two tectonostratigraphic domains separated by the Sholl shear zone: the 3270-3250 Ma Roebourne Group in the north, and the 3130-3115 Ma Whundo Group in the south. Krapez and Eisenlohr (1998) subdivided each domain into various tectonostratigraphic units based on previous mapping (Smith et al., 1998). However, recent mapping (Hickman, 2001, 2002; Hickman et al., 2000) and geochronological data (Nelson, 1996, 1997, 1998) indicate that this nomenclature requires revision.

The West Pilbara granite-greenstone terrane is herein divided into four granitoid complexes (Dampier, Cherratta, Harding, and Caines Well) and three tectonostratigraphic domains (Karratha, Cleaverville, and Sholl), the latter underlain by rocks assigned to three groups (Fig. 10). The Karratha domain to the north of the Sholl shear zone is composed predominantly of ultramafic-felsic volcanic rocks of the ca. 3270 to 3250 Ma Roebourne Group and by the coeval Karratha Granodiorite. The top contact of this domain is along an early (D_1) layer-parallel shear zone, the Regal thrust, that separates it from the Cleaverville domain over an area measuring 70 km east-northeast–west-southwest and 25 km north-northwest–south-southeast (Fig. 10; Hickman, 1997b, 2002; Hickman et al., 2000, 2001).

To the south of the Sholl shear zone are the Cherratta Granitoid Complex and the Sholl domain that are separated by the Maitland shear zone (Hickman, 1997a). The Sholl domain comprises a succession of interbedded mafic and felsic volcanic rocks mainly composed of the 3130 to 3115 Ma Whundo Group but which includes the Cleaverville Formation in the east. The Maitland shear zone is older than the crosscutting 2925 Ma suite of layered mafic-ultramafic intrusions of the West Pilbara granite-greenstone terrane and was truncated by the last movement on the Sholl shear zone.

The Cleaverville domain is underlain by the Regal thrust and bounded to the south by the Sholl shear zone. Stratigraphic components are the Regal Formation and the ca.

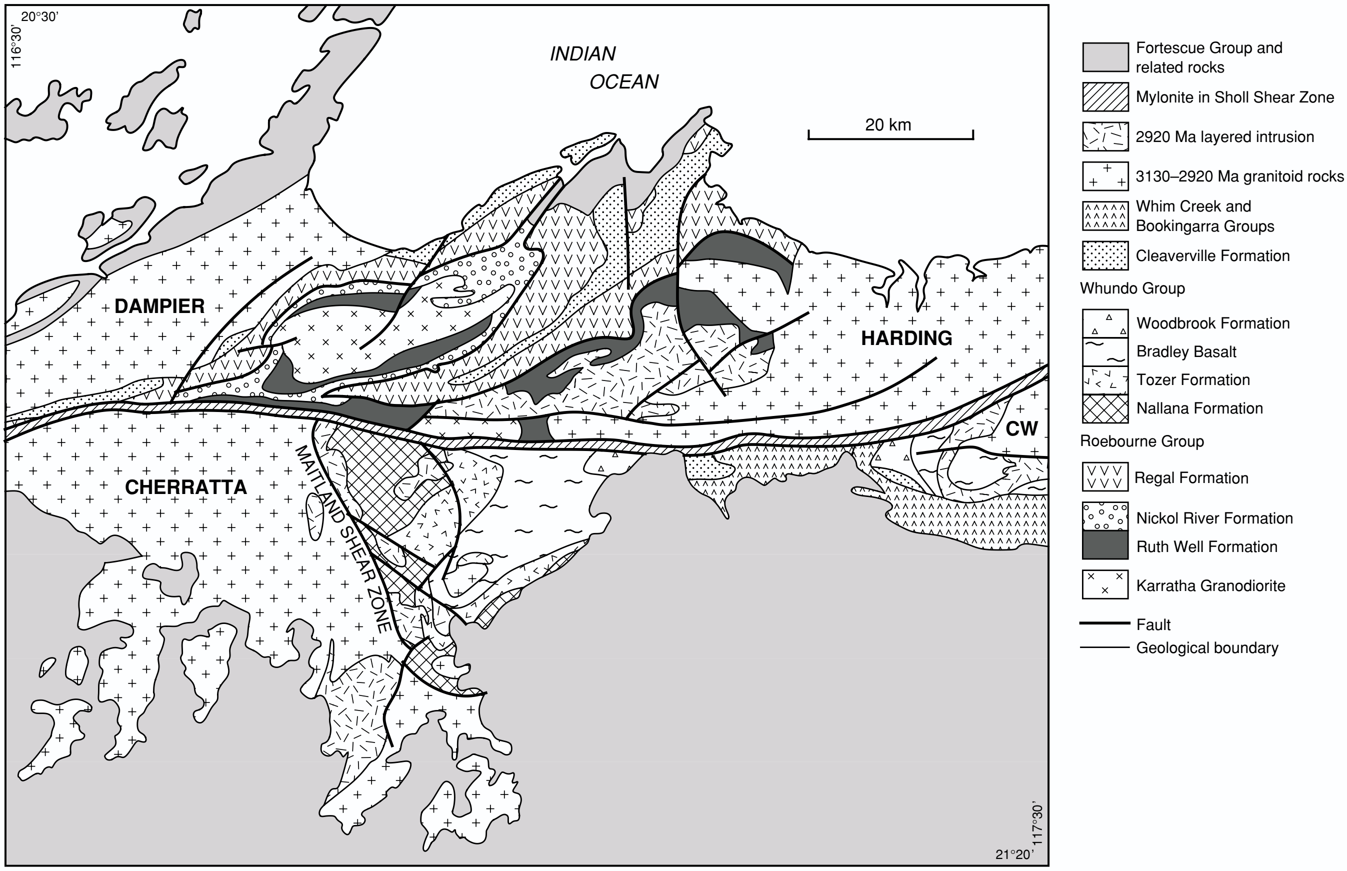

FIG. 9. Lithostratigraphy of the West Pilbara granite-greenstone terrane. CW = Caines Well Granitoid Complex.

3020 to 3015 Ma Cleaverville Formation. Very poor exposure obscures the relationship between the Dampier Granitoid Complex and the Regal Formation, but granite veins in the Regal Formation suggest an intrusive relationship. The age of the Regal Formation has not been directly established, but a general similarity of lithology with the Ruth Well Formation of the Roebourne Group is used to infer a similar age. Sandstone or finer grained clastic sedimentary rocks at the base of the overlying Cleaverville Formation lie disconformably above the Regal Formation.

Lithostratigraphy

Greenstones of the West Pilbara granite-greenstone terrane were deposited over 250 m.y. in three unconformity-, fault-, and/or intrusion-bounded successions (Fig. 9). The volcanic Roebourne Group was deposited north of the Sholl shear zone at, or before, ca. 3270 to 3250 Ma, in part synchronously with the Karratha Granodiorite. The volcanic Whundo Group was deposited south of the Sholl shear zone at 3130 to 3115 Ma. The ca. 3020 Ma Cleaverville Formation conformably to disconformably overlies both of these older groups and is deformed within the Sholl shear zone. For now, the Cleaverville Formation is assigned to the Gorge Creek Group, based on lithostratigraphic correlation with undated banded iron-formation and chert units on the western and northwestern margins of the East Pilbara granite-greenstone terrane (Pilbara Well and Ord Ranges greenstone belts; Fig. 2). This correlation may have to be reviewed should dating of the correlative rocks in the East Pilbara granite-greenstone terrane show them to be distinct from the Cleaverville Formation. A suite of layered mafic-ultramafic intrusions with significant Ni-Cu-PGE-V mineralization (Hoatson et al., 1992) was emplaced within the West Pilbara granite-greenstone terrane at ca. 2924 Ma (Nelson, 1998).

Roebourne Group: Field relationships have established that serpentinized peridotitic komatiite, derived talc-chlorite schist, metabasalt, and chert of the Ruth Well Formation are the oldest preserved components of the Roebourne Group (Table 1). These rocks are undated but must be older than ca. 3270 to 3250 Ma, the age of conformably overlying rocks of the Nickol River Formation and coeval, intrusive granitoid rocks of the Karratha Granodiorite (Table 1; Nelson, 1998; Smith et al., 1998; Hickman, 1999).

Peridotitic komatiite, metabasalt, and chert of the Regal Formation overlie the Nickol River Formation in the central and western parts of the West Pilbara granite-greenstone terrane across the layer-parallel Regal thrust. This tectonic contact raises questions as to the assignment of the Regal Formation to the Roebourne Group (Hickman, 1997a), but stratigraphic redefinition has been deferred pending precise geochronological data on the Regal Formation. A minimum depositional age for the Regal Formation is provided by a

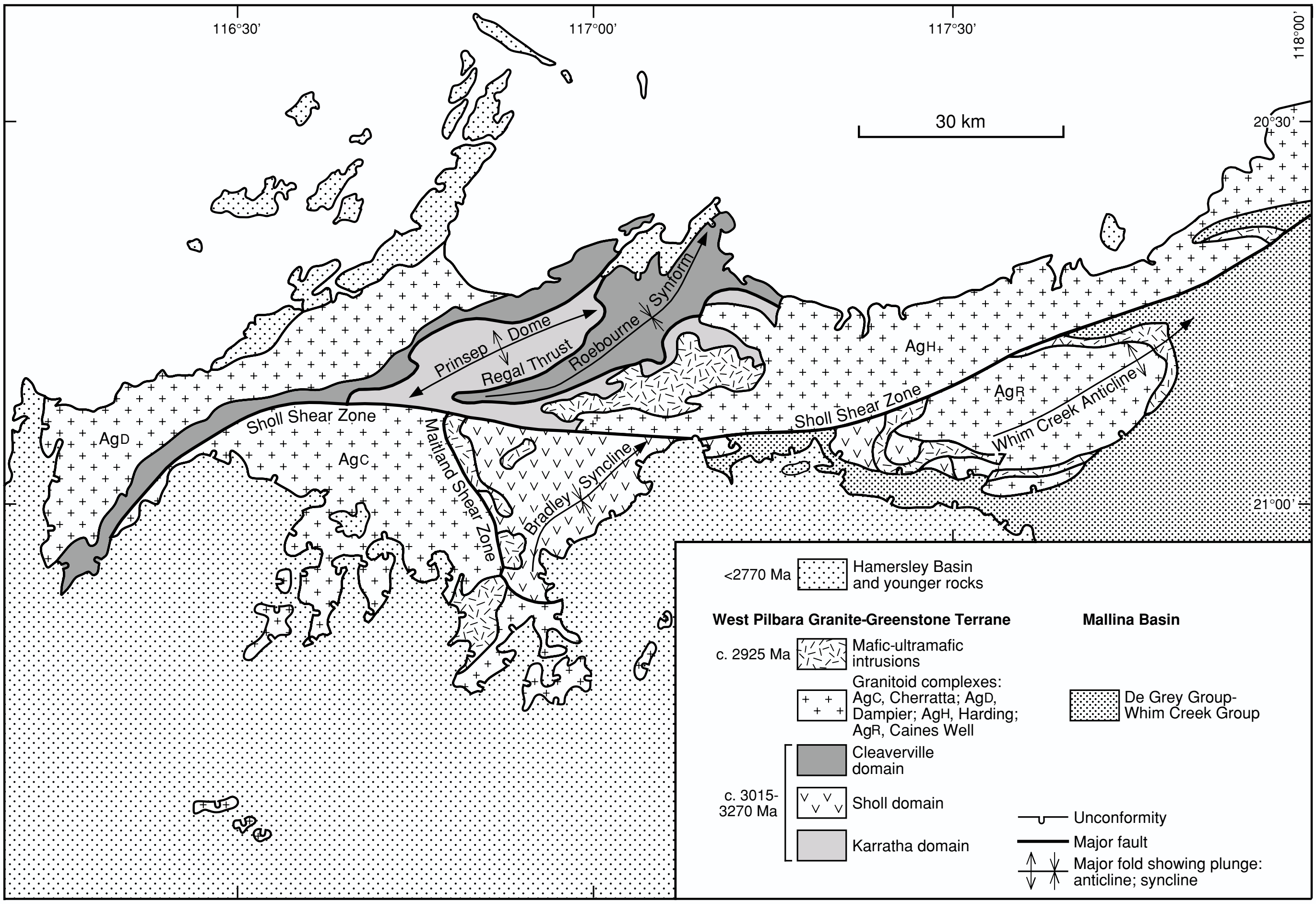

FIG. 10. Lithotectonic domains and granitoid complexes of the West Pilbara granite-greenstone terrane. The Regal thrust marks the boundary between the Karratha and Cleaverville domains.

TABLE 1. Stratigraphy of the Roebourne Group (after Hickman, 1997)

Formation	Thickness (m)	Lithology and relationships
Regal	2,000	Basal peridotitic komatiite overlain by pillow basalt and local chert units; intruded by microgranite and ca. 3015 Ma felsic porphyry
	Tectonized contact along Regal Thrust	
Nickol River	100–500	Banded chert, iron-formation, ferruginous clastic sedimentary rocks, quartzite, felsic volcanic, carbonate rocks, and volcanogenic sedimentary rocks and local conglomerate; felsic schist dated at 3269 ± 2 Ma and rhyolite dated at 3251 ± 6 Ma
Ruth Well	1,000–2,000	Basalt and extrusive peridotite with thin chert units; intruded by granodiorite dated at 3270 ± 2 Ma

3018 ± 2 Ma granite sill and a slightly older age of ca. 3022 Ma for the basal Gorge Creek Group (Nelson, 1998).

Whundo Group: The Whundo Group (Table 2; Hickman, 1997a) is an ~10-km-thick succession of mafic and felsic metavolcanic rocks that crops out south of the Sholl shear zone between the Cherratta and Caines Well Granitoid Complexes (Fig. 9). The basal Nallana Formation is 2000 m thick and comprises metabasalt, ultramafic rocks, intermediate-felsic pyroclastic rocks dated at 3125 ± 4 Ma (Nelson, 1996), and sills of dolerite. Whereas the maximum age of this formation is unknown, it must be older than the intrusive granodiorite of the ca. 3130 Ma Cherratta Granitoid Complex (Nelson, 1998). The conformably overlying Tozer Formation is ca. 3120 Ma and consists of ~2,500 m of alternating metamorphosed basalt, andesite, dacite, rhyolite, and thin metasedimentary units, including chert and banded iron-formation (age data from Horwitz and Pidgeon, 1993; Hickman, 1997a; Nelson, 1997, 1998). The next, conformably overlying unit is the homogeneous Bradley Basalt, which exceeds 4,000 m in

TABLE 2. Stratigraphy of the Whundo Group (after Hickman, 1997)

Formation	Thickness (m)	Lithology and relationships
Woodbrook 3117 ± 3 Ma	1,000	Rhyolite tuff and agglomerate; minor basalt and thin banded iron-formation
Bradley Basalt 3115 ± 5 Ma	>4,000	Pillow basalt, massive basalt, minor units of felsic tuff and chert
Tozer ca. 3120 Ma	2,500	Calc-alkaline volcanics, including felsic pyroclastic units; minor chert and thin banded iron-formation
Nallana 3125 ± 4 Ma	2,000	Dominantly basalt but includes minor ultramafic and felsic units; base of formation intruded by 3130 Ma granitoids and truncated by the Maitland shear zone

thickness. Massive and pillowed basalt dominate the formation but spinifex-textured high Mg basalt is present particularly near its base, and felsic tuffs occur in the upper part of the formation, one of which was dated at 3115 ± 5 Ma (Nelson, 1996). Units of rhyolite tuff and agglomerate in the uppermost 1,000 m of the Whundo Group have been assigned to the Woodbrook Formation (Hickman, 1997a), which is 3117 ± 3 Ma (Nelson, 1998). The Fortescue Group conceals the stratigraphic top of the Whundo Group.

Gorge Creek Group: The Cleaverville Formation (Ryan and Kriewaldt, 1964) is composed of banded iron-formation, ferruginous chert, gray-white and black chert, shale, siltstone, and minor volcanogenic sedimentary rocks. Dolerite sills intrude the formation in the Karratha domain. Clastic sedimentary rocks at the base of the Cleaverville Formation suggest an erosional, disconformable contact in pillow basalt of the underlying Regal Formation.

The Cleaverville Formation locally contains detrital zircons with near-concordant ages of ca. 3287 to 3236 Ma and one 3461 ± 8 Ma zircon grain (Nelson, 1998). Whereas the former population was probably derived by erosion of the Karratha Granodiorite, the ca. 3461 Ma age and suggests that the East Pilbara granite-greenstone terrane-aged crust may have been exposed in the Roebourne area during deposition of the Cleaverville Formation at ca. 3020 Ma. South of the Sholl shear zone, the depositional age of the Cleaverville Formation is closely constrained by clastic zircons dated at 3018 ± 3 Ma (Nelson, 1998) and by intrusive granophyre dated at 3014 ± 6 Ma (Nelson, 1997).

Granitoid rocks

Granitoid rocks of the West Pilbara granite-greenstone terrane crop out as parts of four granitoid complexes that include three age components: the ca. 3270 to 3260 Ma Karratha Granodiorite, granitoid rocks in the range 3160 to 3060 Ma, and others between ca. 3015 and 2940 Ma.

The 3270 to 3260 Ma Karratha Granodiorite ranges from allotriomorphic granular tonalite to granodiorite (Nelson, 1998; Smith et al., 1998). Much older Nd T_{DM} model ages of 3480 to 3430 Ma from this body (Sun and Hickman, 1998) indicate that magma generation involved older crust or enriched lithospheric mantle. One of the samples (JS17) dated by Smith et al. (1998) contained near-concordant zircon cores with $^{207}Pb/^{206}Pb$ ages up to ca. 3311 Ma, supporting the recycling of older crust.

The Sholl shear zone bounds the Cherratta Granitoid Complex to the north, whereas to the southwest the Fortescue Group unconformably overlies it. To the east, the Maitland shear zone separates the complex from the Whundo Group, although small stocks of monzogranite and syenogranite assigned to the complex locally intrude the greenstones. Small inliers of the complex in the Fortescue Group are ca. 3236 Ma (Nelson, 1999). Elsewhere, the oldest components of the complex are 3150 to 3060 Ma that include banded gray granite-tonalite gneiss with xenoliths of amphibolite-facies mafic gneiss and sheets of leucocratic gneiss. These rocks are intruded by porphyritic granite and hornblende-biotite monzogranite. Other components range from granite to granodiorite, including hornblende-rich and porphyritic granitoids. Gneiss dated at 2995 ± 11 Ma (Nelson, 1997) returned a Nd T_{DM} model age of ca. 3246 Ma, whereas other gneiss samples returned Nd T_{DM} model ages of ca. 3139 and 3149 Ma (S-S. Sun, Geoscience Australia, writ. commun., 1997, 2000), similar in age to those obtained from the Whundo Group (Sun and Hickman, 1998).

The Dampier Granitoid Complex intrudes the Regal Formation in the southwest and is intruded by the ca. 2725 Ma Gidley Granophyre. Most of the complex consists of porphyritic granite to granodiorite and a mixed assemblage of even-grained granite to granodiorite containing banded gneissic granitoids, syenogranite, and pegmatite. The complex is thus compositionally similar to the Cherratta Granitoid Complex, and limited geochronology (two samples dated) indicates a similar age of ca. 2990 Ma (Nelson, 1998, 1999). Reconstruction of post-3020 Ma horizontal displacement across the intervening Sholl shear zone indicates no more than 30 to 40 km of dextral offset (see below), and this combined with the similar age and compositional data shows that the two granitoid complexes represent parts of an originally contiguous unit. One sample of the Dampier Granitoid Complex, dated at 2997 ± 3 Ma (Nelson, 1998), included zircon cores with $^{207}Pb/^{206}Pb$ ages ranging between ca. 3106 and 3255 Ma. Nd T_{DM} model ages range from 3387 to 3247 Ma (Geological Survey of Western Australia, unpub. data), suggesting that the Karratha Granodiorite and/or the lower part of the Roebourne Group contributed material to the complex.

The Harding Granitoid Complex is almost entirely concealed by Cenozoic alluvial and marine deposits, but aeromagnetic data indicate that it extends for ~130 km east-northeast and Bouguer anomalies (Blewett et al., 2000) suggest it is 40 to 80 km wide, including its offshore extent. At its southern margin, a 1-km-wide belt of mylonitized granitoid rocks and tectonic lenses of sheared greenstones within the Sholl shear zone separate the complex from the Whundo and Whim Creek Groups. Rare outcrops of the complex are composed of weakly foliated to compositionally banded monzogranite and granodiorite. A sample of little deformed monzogranite is 2970 ± 5 Ma and contains a population of xenocrystic zircons at 3018 ± 19 Ma (Nelson, 1999). The same rock provided an Nd T_{DM} model age of ca. 3309 Ma (Geological Survey of Western Australia, unpub. data). Immediately north of the Sholl shear zone, a sample of granodiorite gneiss was dated at 3014 ± 3 Ma (Nelson, 1997) and gave an Nd T_{DM} model age of ca. 3276 Ma (Sun and Hickman, 1998). A granitoid sample (JS43) assigned to this complex by Smith et al. (1998) is reinterpreted herein to be part of the Karratha Granodiorite.

The Caines Well Granitoid Complex forms a doubly plunging anticline in the far northeast of the West Pilbara granite-greenstone terrane and is in contact with rocks of the Whim Creek greenstone belt to the east, south, and west across an unconformity or faults (Fig. 10). To the north, geophysical data and limited outcrop identify the northern portion of the granitoid complex and mantling rocks of the Whim Creek greenstone belt approximately 40 km in a dextral sense across the Sholl shear zone (Smithies, 1998a). Strongly foliated trondhjemite in the east of the complex has an intrusive age of 3093 ± 4 Ma (Nelson, 1997) that is only marginally younger than its Nd T_{DM} model age of ca. 3118 Ma (Sun and Hickman, 1998) and may indicate derivation via partial melting of the Whundo Group. The main phase of the complex is porphyritic monzogranite, dated at 2990 ± 5 Ma (Nelson, 2000). Minor amounts of leucocratic monzogranite have been dated at 2925 ± 4 Ma (Nelson, 1997).

Sholl shear zone

The Sholl shear zone bisects the West Pilbara granite-greenstone terrane (Fig. 9) and has a long history of displacement and reactivation. It is a subvertical zone of mylonite and schist 1 to 2 km wide over most of its 250- to 350-km length. Smith et al. (1998) placed the timing of the movement in the shear zone at 2991 to 2925 Ma, whereas Krapez and Eisenlohr (1998) placed it between ca. 3000 and 2955 Ma. Both these interpretations advocated major sinistral movement, up to 150 to 200 km (Hickman et al., 2000; Hickman, 2002). However, displacement after 3020 Ma was predominantly dextral, as seen by the 30- to 40-km displacement of the Whim Creek Group and Caines Well Granitoid Complex and of ca. 2925 Ma layered intrusions (Fig. 10: Smithies, 1998a; Hickman, 2002).

Crustal variation north and south of the Sholl shear zone is indicated by differences in lithostratigraphy and ages of greenstones (Smith et al., 1998). Sm-Nd data (Sun and Hickman, 1998; Hickman et al., 2000) supports reworking of 3500 to 3300 Ma (East Pilbara granite-greenstone terrane?) crust north of the shear zone but indicates that such an old component is lacking to the south, where crust is <3300 Ma.

Structural geology

Structures in the West Pilbara granite-greenstone terrane range in age from 3160 to <2920 Ma and may be divided into seven local events. Some of these events have also been recognized in the Mallina basin but have been placed in a separate local hierarchy for that area.

D_1 (3160–3130 Ma): The earliest recognizable tectonic structures in the West Pilbara granite-greenstone terrane are bedding-parallel, shallow-dipping high-strain zones and recumbent folds, produced by southerly directed thrusting (Hickman, 2001). The largest fault is the Regal thrust at the top of the Nickol River Formation, but low-angle faults also occur within the Regal Formation. Intrafolial isoclinal folds are relatively common within the Nickol River Formation of the Mount Regal area (Hickman et al., 2000) and are interpreted as minor structures related to the recumbent folds. A bedding-parallel tectonic foliation (S_1), which is preserved in metasedimentary rocks of the Nickol River Formation (Hickman et al., 2000) and in metabasalt of the Regal Formation, is interpreted to have initially formed parallel to the D_1 thrusts but was reactivated by parallel shearing during later tectonic events.

A thermotectonic event at 3160 to 3150 Ma is indicated by zircon data (Smith et al., 1998) and K-Ar geochronology (Kiyokawa and Taira, 1998) from north of the Sholl shear zone, which is interpreted to coincide with D_1. Granitoid rocks of this age from the Cherratta Granitoid Complex (Nelson, 1999) and from close to the southeastern margin of the Mallina basin (Nelson, 2001) confirm a contemporaneous magmatic event at this time. This age of D_1 deformation is further supported by the fact that such structures have not been recognized in the Whundo Group, which is younger than ca. 3130 Ma.

D_2 (ca. 3050-3020 Ma): Rotated feldspar porphyroclasts within fine-scale mylonitic laminations of the Sholl shear zone consistently indicate an episode of sinistral shear prior to dextral reactivation of the zone at <2925 Ma (D_6; Hickman, 2001). This early deformation must have predated deposition of the Cleaverville Formation at ca. 3020 Ma, as these rocks have only been affected by the younger episode of dextral displacement across the zone.

D_3 (3015-3010 Ma): Tight to isoclinal east-northeasterly trending folds in the Cleaverville Formation north of the Sholl shear zone are attributed to D_3 and predate intrusion of a ca. 3000 Ma rhyolite sill on the northern limb of the Prinsep dome. Easterly trending, upright, tight to isoclinal D_3 folds affect a ca. 3014 Ma granophyre sill in the Cleaverville Formation (Nelson, 1997) and are unconformably overlain by the ca. 3010 Ma Warambie Basalt of the Whim Creek Group (see "The Mallina Basin").

D_4 (ca. 2945-2930 Ma): The D_4 event formed major northeasterly trending, tight to open folds such as the Prinsep dome, Roebourne synform, and Bradley syncline (Fig. 10). D_4 folds are oblique to the Sholl shear zone and related strike-slip faults of the West Pilbara granite-greenstone terrane, and are interpreted to have formed in response to northwest-southeast compression prior to, or during, the onset of major faulting across the Sholl shear zone. D_4 structures in the West Pilbara granite-greenstone terrane are chronologically correlated with the major D_3 folds in the Mallina basin (Smithies, 1998a) and interpreted to be equivalent to phase 4 structures of Krapez and Eisenlohr (1998). Geochronology established the age of these structures in the Mallina basin at younger than 2938 ± 4 Ma (Smithies, 1998b), and in the West Pilbara granite-greenstone terrane they are cut by ca. 2930 Ma D_5 structures, thus they are not 2906 to 2863 Ma as suggested by Krapez and Eisenlohr (1998). Shear zones also occur within the Cherratta Granitoid Complex where gneiss, which developed as a result of the shearing, contains synkinematic zircon populations dated at 2944 ± 5 Ma (Nelson, 1997).

D_5 (ca. 2940 Ma): The north-northwesterly striking Maitland shear zone truncates major, northeasterly trending D_4 folds of the Mount Sholl area. A parallel tectonic foliation (S_5)

is developed in the adjacent greenstones of the Whundo Group and in the Cherratta Granitoid Complex. The Sholl shear zone locally truncates all these structures.

D_6 (ca. 2920 Ma): The latest movement in the Sholl shear zone was dextral and accounted for 30 to 40 km of displacement of the Whim Creek greenstone belt and ca. 2925 Ma ultramafic layered intrusions. A subsidiary dextral strike-slip fault, the Black Hill shear zone, displaces the Andover intrusion by 10 km (Fig. 10; Hickman, 2002). As noted by Krapez and Eisenlohr (1998), zircon geochronology in several rock units close to the Sholl shear zone has revealed a metamorphic disturbance event at about 2920 Ma, which is interpreted to date D_6. Minor D_6 structures in the Sholl shear zone include dextral drag folds and isoclinal folds of S_2 mylonite lamination and associated small-scale faults and brecciation.

D_7 (<2920 Ma): The Sholl shear zone and earlier structures are deformed by a conjugate system of north-northeasterly striking sinistral faults and west-northwesterly striking dextral faults (Hickman, 2001, 2002). The precise age of D_7 faults is unknown, but well-developed conjugate structures have not been observed in the Fortescue Group.

Metamorphism

Granitoid complexes contain greenstone enclaves that have been metamorphosed to amphibolite grade, and the granitoid rocks themselves show evidence of retrograde metamorphism to greenschist facies from amphibolite facies. Amphibolite facies assemblages are also developed in greenstones of the Roebourne Group, whereas the Whundo Group contains rocks at lower greenschist grade. An exception to the above generalization occurs in the Cleaverville domain between Karratha and Cleaverville, where downward movement in steep faults on the northwestern side of the Prinsep dome has preserved greenschist facies sections of the Regal Formation. The metamorphic change across the Sholl shear zone is attributed to upward movement on its northern side, although the timing of this is uncertain.

The Mallina Basin

The ca. 3016 to 2940 Ma Mallina basin is >200 km long, up to 90 km wide, and located across the boundary between the East and West Pilbara granite-greenstone terranes (Fig. 1). Supracrustal rocks of the basin include a volcanic rift margin succession of the Whim Creek greenstone belt in the northwestern part and siliciclastic turbidites of the De Grey Group, which dominate the southern and eastern two-thirds of the basin.

Previous investigations have interpreted the faulted contact between the Whim Creek greenstone belt and the De Grey Group to be a long-lived tectono-stratigraphic domain boundary (Krapez, 1993; Krapez and Eisenlohr, 1998). However, recent geological and geochronological evidence suggests that the upper stratigraphic part of the Whim Creek greenstone belt (the Bookingarra Group) forms an intrinsic part of the Mallina basin (Fig. 11; Smithies et al., 1999). Thus, the western margin of the Mallina basin is now taken as the locally faulted, unconformable contact between the little deformed volcanic rocks of the Whim Creek greenstone belt and the more strongly deformed and folded rocks of the underlying Cleaverville Formation, Whundo Group, and the oldest (ca. 3100 Ma) phase of the Caines Well Granitoid Complex (Barley, 1987; Smithies, 1998a). The southeastern margin of the basin is a locally faulted unconformity or disconformity, where a chert-cobble conglomerate at the base of the Constantine Sandstone of the De Grey Group overlies folded chert and banded iron-formation and older greenstones of the East Pilbara granite-greenstone terrane (Fig. 12).

A maximum depositional age of 3016 ± 13 Ma was obtained on detrital zircons from a volcanolithic sandstone that conformably underlies chert immediately below the Constantine Sandstone in the southeastern part of the basin (Nelson, 1998; Smithies et al., 1999). This age is consistent with correlation of the chert with the Cleaverville Formation exposed farther to the west. Similar cherts are exposed in the cores of major anticlines in the southeastern part of the zone and are regarded as basement to the basin. However, Nd T_{DM} model ages of 3.1 to 3.2 Ga from late granites within the basin suggest that much of the concealed basement is of similar age to the Whundo Group but has not been identified in outcrop. The basin was intruded by widespread granitoid rocks and deformed late in the history of deposition to form the Central Pilbara tectonic zone.

Whim Creek greenstone belt

Evidence presented by Pike and Cas (in press) and Pike et al. (2002) suggests that the dominantly volcanic rocks of the Whim Creek greenstone belt formed in two distinct, upward-deepening successions separated by up to 40 m.y. (Fig. 11). As a result, the previous stratigraphic model of Hickman (1983) has been upgraded such that the volcanic stratigraphy of the belt is now subdivided into the older, stratigraphically lower Whim Creek Group and the younger Bookingarra Group.

The Whim Creek Group contains a complex association of approximately coeval felsic to mafic volcanic, intrusive and volcaniclastic rocks (Pike and Cas, in press) that have been dated at 3016 ± 3 to 3009 ± 4 Ma (Nelson, 1998, 2002). Turbidity currents and debris flows deposited abundant juvenile felsic volcaniclastic material, with limited paleocurrent data suggesting a source to the south. These rocks are interleaved with subaqueous basaltic lavas and breccia, and a voluminous, synsedimentary, dacitic-rhyodacitic sill intrudes the entire sequence with apophyses into the overlying volcaniclastic rocks (Pike and Cas, in press).

The Cistern Formation forms the base of the Bookingarra Group (Fig. 11) and contains abundant volcaniclastic material derived from the underlying Whim Creek Group. Rocks of the Cistern Formation show an upward increase in siliciclastic material and parallel decrease in grain size, culminating with the deposition of the Rushall Slate. The rocks are overlain by locally spinifex-textured and variolitic high Mg basalts of the Louden and Mount Negri Volcanics (Hickman, 1977). Contacts between the basalts and the volcaniclastic and siliciclastic rocks are generally conformable (Smithies, 1998a) or a low-angle unconformity (Hickman, 1997a). However, Pike and Cas (in press) recognized peperite-like contacts where basalt locally intruded unlithified sediment, indicating that basaltic volcanism and clastic deposition in the upper part of

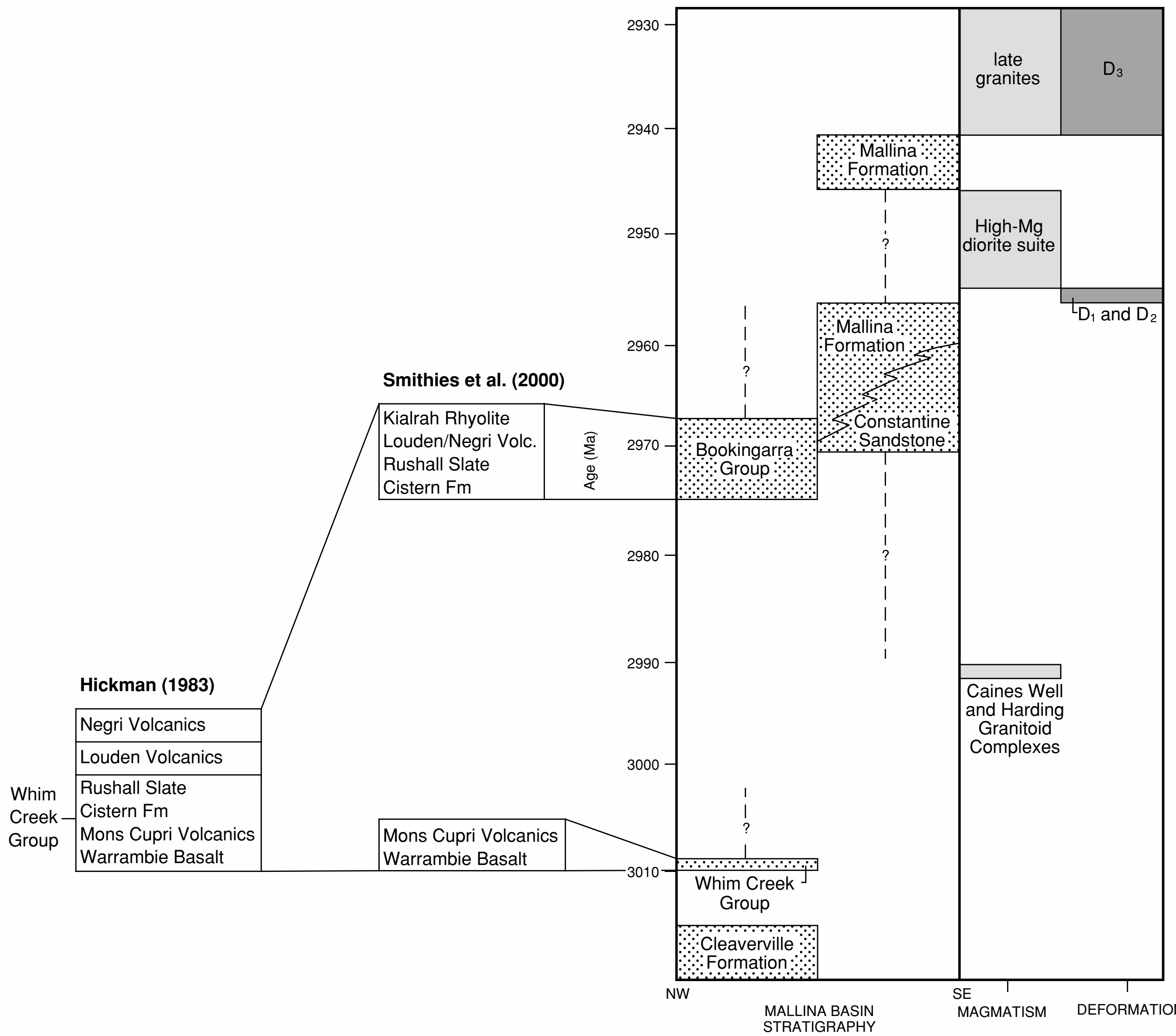

FIG. 11. Revised stratigraphy and evolutionary history of the Mallina basin. Left column is the stratigraphy of mafic volcanic rocks from the Whim Creek area, proposed by Hickman (1983). Second column is the revised stratigraphy of the Whim Creek greenstone belt by Smithies et al. (2000), showing the division of greenstones into the Whim Creek and Bookingarra Groups. Main column shows the schematic relationship between basin components, magmatism, and deformation.

the group overlapped in time. Limited paleocurrent data indicate derivation of the upper part of the group from a northerly source.

In the southwestern part of the Whim Creek greenstone belt, the Kialrah Rhyolite overlies the Louden Volcanics and has a maximum depositional age of 2975 ± 4 Ma (Nelson, 1998; Smithies et al., 2001). Detrital zircons from a sample of the Cistern Formation also provide a maximum depositional age of ca. 2975 Ma (Nelson, 2000; Smithies et al., 2001). Sulfide mineralization within the Bookingarra Group has a Pb isotope model age of ca. 2948 to 2942 Ma, which provides a minimum depositional age for this group irrespective of whether the mineralization is interpreted as syngenetic or epigenetic (Huston et al., 2002). This depositional age range is identical to that for the Constantine Sandstone and the Mallina Formation, in the southern part of the Mallina basin (see below).

De Grey Group

The siliciclastic rocks of the De Grey Group, to the south and east of the Whim Creek greenstone belt, have been divided into two units, the Constantine Sandstone and the Mallina Formation (Figs. 11, 12). The Mallina Formation, comprising interbedded, well-graded, fine- to medium-grained wacke and shale, overlies the Constantine Sandstone, which comprises medium- to coarse-grained, poorly sorted subarkose to wacke, locally with thick conglomerate layers (e.g., Fitton et al. 1975; Hickman, 1990). Both units were deposited in a submarine fan (e.g., Hickman, 1977; Eriksson, 1982) but represent contrasting depositional environments. The

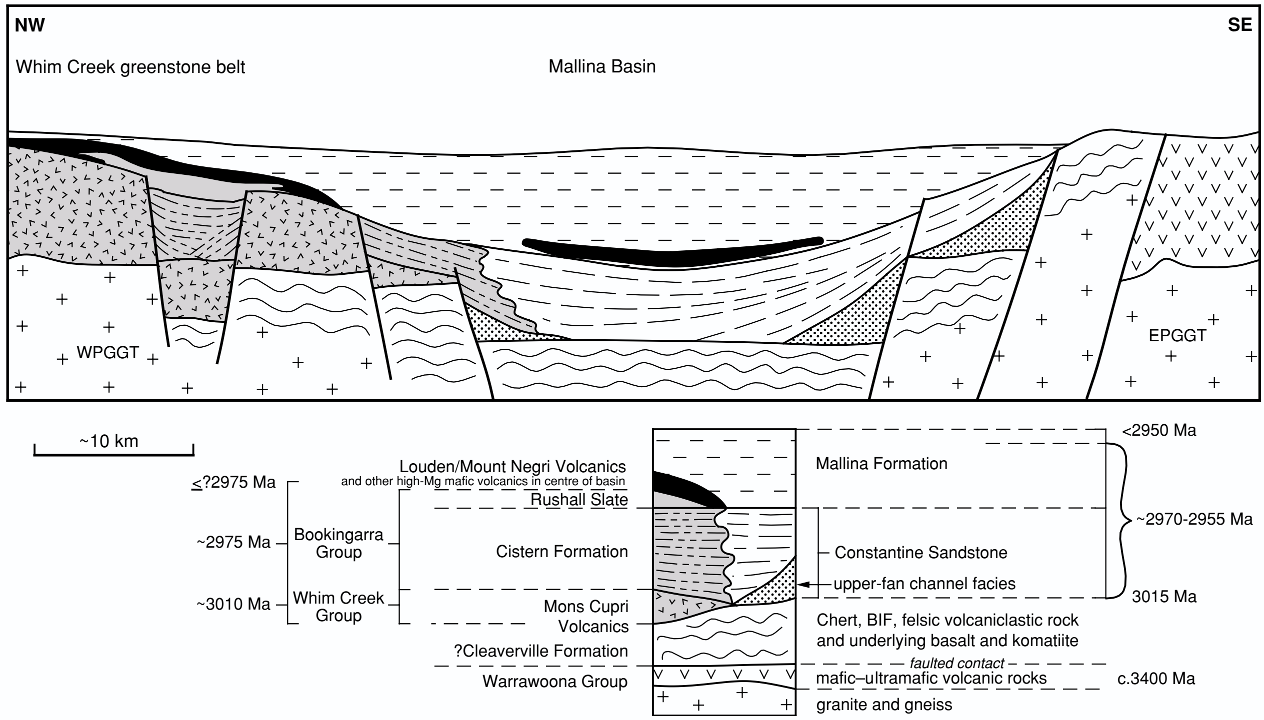

FIG. 12. Schematic northwest-southeast cross section of the Mallina basin, showing the interpreted interbedding between the rift margin Whim Creek greenstone belt and the clastic sedimentary rocks of the De Grey Group. Adapted from Smithies et al. (1999).

Mallina Formation is typically a fine-grained facies that probably reflects generally low rates of sediment supply, whereas the coarse-grained Constantine Sandstone typically reflects generally higher rates of sediment supply. Thin, hyaloclastite-bearing mafic lavas in the lower parts of the basin are compositionally transitional between basalt and high Mg basalt, with close compositional similarity to the Mount Negri and Loudens Volcanics of the Whim Creek greenstone belt (Geological Survey of Western Australia unpub. data).

Detrital zircons within the Constantine Sandstone indicate a maximum depositional age of ca. 2990 Ma, while a study of xenocrystic zircons from granites that intrude the basin suggest that the maximum depositional age for the basin may be as young as ca. 2970 Ma. (Smithies et al., 2001). The youngest population of detrital zircons from a wacke, presumed to be of the Mallina Formation, within the southeastern part of the Mallina basin, gave a maximum depositional age of 2941 ± 9 Ma (Smithies et al., 1999). Folding of the Mallina Formation predates intrusion of granites (see below), which began at ca. 2955 Ma, and so the sedimentary host to these young detrital zircons was deposited after the onset of folding. This suggests the presence of a cryptic unconformity within these rocks and a second, post-ca. 2945 Ma, depositional cycle within the Mallina basin (Smithies and Farrell, 2000; Smithies et al., 2001).

Intrusive rocks

Three suites of layered mafic-ultramafic rocks have intruded the Mallina basin. The Sherlock intrusion, comprising leucogabbro and pyroxenite, has intruded the contact between the Caines Well Granitoid Complex and the Whim Creek greenstone belt and hosts significant Ti-V mineralization. The Opaline Well intrusion, comprising layered gabbro and olivine gabbro, intrudes the Whim Creek greenstone belt. Smithies (1998a) interpreted this intrusion to be the subvolcanic equivalent to the Louden and/or Mount Negri Volcanics. Similar gabbro sills intrude the siliciclastic turbidites in the southern part of the basin and have compositions that are virtually indistiguishable from the Louden Volcanics. The Millindinna intrusion is a regionally extensive but generally thin (<200 m) sill that intrudes the siliciclastic turbidites in the southern part of the Mallina basin. This sill varies from lherzolite at the base to melanogabbro at the top. The intrusive age of the Millindinna intrusion and the gabbros is constrained between ca. 2970 Ma (max. depositional age of the basin) and 2950 Ma (the age of later intrusions) and thus may be similar in age to the Louden Volcanics.

Granitic intrusions into the Mallina basin can be divided broadly into three compositional groups; alkaline granite, high Mg diorite (sanukitoid), and high K monzogranite. The 2946 ± 6 Ma (Nelson, 1999) Portree Granitoid Complex is a large nested plutonic complex of alkaline granite, which is locally hypersolvus and contains sodic pyroxene.

Intrusion of the high Mg diorite, or sanukitoid, suite was coeval with intrusion of the alkaline granite, with ages ranging between 2954 ± 4 and 2945 ± 6 Ma (Nelson, 2000). Rocks of the sanukitoid suite range from diorite to granodiorite and have compositions that can best be explained by derivation from a mantle source strongly metasomatized by a subduction-derived melt component rich in La, Ce, Sr, and Ba (e.g., Shirey and Hanson, 1984; Smithies and Champion, 2000). The suite comprises a belt of intrusions that parallels the axis of the Mallina basin, indicating a strong structural control. Distinct aeromagnetic anomalies suggest the presence of a number of these intrusions under Cenozoic cover within and adjacent to the major east-northeast–trending Mallina shear zone. Low-pressure metamorphic aureoles and geobarometric calculations constrain the depths of emplacement to less than 7 km (Smithies and Champion, 2000).

A series of late, generally porphyritic monzogranites has intruded the siliciclastic rocks of the Mallina basin between 2941 ± 4 and 2931 ± 5 Ma (Nelson, 1998, 2000). The largest intrusion is the Satirist Granite in the southern part of the Mallina basin, which has associated, locally extensive, quartz-tourmaline veining. The K feldspar porphyrytic Opaline Well Granite lies adjacent to the intersection of the Mallina shear zone and the Loudens fault and locally contains quartz-tourmaline orbicules up to 10 cm in diameter. However, with an age of 2765 ± 5 Ma (Nelson, 1997), this granite is significantly younger than the other monzogranites and is coeval with basaltic magmatism that forms the lower part of the Fortescue Group.

Structural geology

Rocks of the Mallina basin have been folded during at least three contractional deformation events. The first (D_1), resulted in open east-trending folds that are poorly preserved around the Satirist Granite. Although undated, these structures must be younger than ca. 2970 Ma, the age of the basin fill. Thrusts in the ca. 3010 Ma Warambie Basalt of the Whim Creek Group (Hickman, 2002) predate dextral movement in the Sholl shear zone at ca. 2920 Ma and may therefore also represent D_1 structures.

D_2 produced large north- to northeast-trending folds that have locally exposed the Constantine Sandstone and underlying Cleaverville Formation in the cores of anticlines around the Satirist Granite. Granites dated at ca. 2950 Ma truncate these folds.

D_3 resulted in the dominant east-northeast–trending structural fabric of the basin, forming the Central Pilbara tectonic zone, and is correlated with large-scale folds with similar orientations in the West Pilbara granite-greenstone terrane and across the western part of the East Pilbara granite-greenstone terrane. Fabrics related to this third event overprint the marginal phase of the Satirist Granite, dated at 2938 ± 4 Ma (Nelson, 1998).

The D_2 and D_3 events are separated by a period of extension, during which sediment that comprises the upper part of the Mallina basin was deposited and granitoid rocks of the sanukitoid suite were emplaced (Smithies and Farrell, 2000; Smithies et al., 2001). Numerous east-northeast–trending shear zones are thought to reflect reactivation of early basin-developing faults within the basement during this event (Smithies, 1999). The largest of these is the Mallina shear zone, which locally juxtaposes slivers of Constantine Sandstone against the Mallina Formation. Examination of the western extent of this shear zone in the eastern part of the Mallina basin, where it is known as the Tabba Tabba shear zone, indicates a dominantly normal, north-side down displacement with a minor component of sinistral displacement that occurred during, or slightly after, granite emplacement at ca. 2950 Ma (Beintema et al., 2001).

Kuranna Terrane

Granitoid rocks are the dominant lithology of the Kuranna terrane (Fig. 1) and can be divided into three main types: lineated biotite monzogranite and granodiorite gneiss with xenoliths of previously metamorphosed sedimentary, mafic and ultramafic rocks; foliated to gneissic biotite monzogranite and syenogranite; post-tectonic, muscovite-bearing rocks of the Bonnie Downs Granite (Hickman, 1983; Williams, 1989). The former two rock types have magmatic crystallization ages of ca. 3.29 to 3.28 Ga, while Ar-Ar cooling ages of hornblende fall in the range 3.1 to 3.2 Ga, and those of muscovite to 3.0 Ga (J. Wijbrans, Vrije Universiteit, The Netherlands, writ. commun., 2000). The granitoid rocks are in contact with fine-grained sedimentary rocks of the Mosquito Creek basin to the north across a high-strain shear zone and are unconformably overlain to the south by rocks of the Mount Bruce Supergroup.

Nd T_{DM} model ages of ca. 3.26 to 3.28 Ga with negative ε_{Nd} values of –2.4 were obtained on rocks of the Kuranna Granitoid Complex, using the depleting mantle model of De Paolo (1981) (Tyler et al., 1992). Since these results were significantly different from those from the East Pilbara granite-greenstone terrane, it was proposed that the Kuranna terrane represented a distinct terrane within the North Pilbara terrain (Tyler et al., 1992). However, when the model ages are recalculated using a two-stage mantle evolution model, the results indicate model ages of ca. 3.45 Ga, similar to crystallization ages of felsic rocks in the East Pilbara granite-greenstone terrane (Geological Survey of Western Australia, unpub. data).

Mosquito Creek Basin

The Mosquito Creek basin to the north of the Kuranna terrane contains low-grade sandstone, shale, and conglomerate of the Mosquito Creek Formation (Hickman, 1983). Thin units of metamorphosed mafic and ultramafic schist and chert occur at the base of the formation in both the northern and southern parts of the basin. The northern margin locally preserves a shallow-dipping, angular unconformity on previously deformed volcanic rocks of the Warrawoona Group, but most of the contact is a north-verging thrust fault. Aeromagnetic data suggest the basin fill thickens dramatically to the south away from the low-angle unconformity along the northern contact (T. Farrell, Geological Survey of Western Australia, pers. commun., 2000). In the northern part of the basin, facing directions are predominantly to the south, although there are local reversals due to tight folding. The rocks become tightly folded and more highly metamorphosed in the southern part of the basin and lie in tectonic contact with rocks of the Kuranna terrane across a wide zone of mylonite (Williams, 1989).

The basin is preserved in an east-west–striking, linear fold-thrust belt that has a fan-shaped cross section. Generally it has been affected by three main sets of structures, although Blewett (2002) recognized up to six. D_1 structures include a slaty cleavage, moderate to steeply plunging S_0-S_1 intersection lineations, and tight folds. D_2 structures include tight to isoclinal east-west–trending folds that are upright to steeply inclined and the main regional schistosity. North-verging D_3 thrusting was concentrated in shear zones in pelitic rocks that host the Au mineralization (Hickman, 1983). Two sets of late, minor structures include kink bands and late-stage, north-south–striking faults (D_4) that continue south into the Kuranna terrane. Northwest-southeast–striking dextral faults (D_5) with local drag folds also affect the Fortescue Group.

The precise age of the basin fill is unknown. However, a maximum age for deposition of the Mosquito Creek Formation is 3312 Ma (Wyman Formation), the youngest age of unconformably underlying rocks in the McPhee dome to the north, or possibly 3240 Ma, the minimum age for the onset of dome formation in the East Pilbara granite-greenstone terrane. White et al. (2001) state that thrusting and folding occurred at 3250 to 3200 Ma, although no supporting evidence was presented. A minimum age of deposition is given by ca. 2900 Ma Pb-Pb model ages on galena in epigenetic Au deposits from this area, which may reflect the age of the D_3 deformation (see Huston et al., 2002).

Discussion

The data presented above show that the North Pilbara terrain evolved through a long, complex history and that individual lithotectonic elements may have partly distinct tectonic histories. The recent data also provide critical new constraints on previous models and show that many of these are seriously flawed.

Horizontal vs. vertical tectonics in the East Pilbara granite-greenstone terrane

Horizontal tectonics?: Previous models of vertical tectonics for the East Pilbara granite-greenstone terrane have been challenged on the grounds that the deformation history of the terrane is more complex than previously thought, involving periods of horizontal compression and extension that have been interpreted to explain the dome and basin map pattern (see references below). In particular, three additional and/or alternate deformation mechanisms to diapirism have been proposed: (1) early extension, (2) early thrusting followed by extension and/or doming, and (3) later thrusting during deposition of the Gorge Creek Group. These are examined below.

1. Early rifting. Extensional growth faults have been recognized in three different formations deposited over 60 m.y. in the lower Warrawoona Group and cited as evidence for early rifting (Zegers et al., 1996; Nijman et al., 1998a, 2001). However, the faults only occur in local areas and only accommodate tens of meters of offset at most, so that the regional significance of these structures is dubious. It is equally possible that some of the growth faults developed during local collapse in the flanks of synvolcanic domes formed during the emplacement of subvolcanic sills or in the margins of calderas (e.g., DiMarco and Lowe, 1989; Van Kranendonk et al., 2001b; Van Kranendonk, in press).

2. Early thrusting followed by extension. Four studies have suggested that thrusting preceded doming in the East Pilbara granite-greenstone terrane. Bickle et al. (1980, 1985) recognized overturned bedding and relatively high grade metamorphic mineral assemblages in an area of complex folding in the northwestern Shaw area, which they concluded arose from a period of early Alpine-style thrusting. In this model, the folds and thrusts were tilted onto their side during later doming of the Shaw Granitoid Complex. Zegers et al. (1996) suggested that this doming occurred at ca. 3467 Ma as a metamorphic core complex during orogenic collapse following the Alpine-style thrusting event identified by Bickle et al. (1985). The core complex model was based on the observation of growth faults in the ca. 3467 Ma Duffer Formation, the observation of reverse, greenstone side-up kinematic indicators within the Split Rock shear zone on the eastern margin of the Shaw Granitoid Complex, a 3467 Ma age from a granitoid sheet cutting early shear fabrics in the shear zone(s) and sinistral kinematic indicators in a westerly continuation of the shear zone that cuts across the top of the Shaw Granitoid Complex. Van Haaften and White (1998) suggested that the greenstone stratigraphy of the Marble Bar greenstone belt was thickened and tectonically inverted by horizontal thrusting following extension during deposition of the Duffer Formation.

Detailed mapping and geochronology has revealed inconsistencies with each of the proposed models for thrusting. The Alpine-style thrusting model for the northwestern Shaw area (Bickle et al., 1985) warrants reconsideration because (1) the tight folds and interpreted thrusts only occur in this local area, (2) the stratigraphy can be followed around the folds and faces away from the granitoids throughout the area, (3) a thermal anomaly associated with the Shaw Granitoid Complex was unexplained by thrusting as pointed out by the authors of the model, and (4) there is only a single foliation that contains metamorphic mineral elongation lineations that are colinear with the axes of the large-scale folds and pass gradationally into D_4 linear fabric elements in the ca. 2940 Ma Mulgandinnah shear zone and thus are more consistent with having formed as D_4 structures (Van Kranendonk, 2000, 2002). Indeed, the geology of the northwestern Shaw area can be intepreted as the result of localized constriction, having formed as a flower structure during 2940 Ma deformation within a restraining bend between stepped splays of the Mulgandinnah shear zone (Van Kranendonk, 1997, 2002; Van Kranendonk et al., 2001b). In this model, the thermal anomaly associated with the Shaw Granitoid Complex was formed during an earlier period of partial convective overturn (Collins and Van Kranendonk, 1999).

The core complex model of Zegers et al. (1996) for the Shaw Granitoid Complex is inconsistent with the following geologica data: (1) the proposed westerly continuation of the extensional detachment of the core complex across the top of the Shaw Granitoid Complex does not exist; rather this well-exposed area is underlain by a set of tight, north-plunging folds defined by several different rock types (including ca. 3445–3410 Ma leucogranite) and foliations (Van Kranendonk, 2000); (2) the metamorphic and structural features of the Shaw Granitoid Complex and surrounding greenstones are grossly dissimilar to typical core complexes in that there is no sharp metamorphic discontinuity at the granite-greenstone interface, univergent extension across the dome is unproven (see point 1 above), and a synexhumation, unconformably overlying cover sequence is not present (cf. Coney, 1980); (3) recent mapping has shown that the proposed extensional faults in the Duffer Formation do not cause original stratigraphic thickness variations in the formation as described, but are late structures that crosscut foliations (Geological Survey of Western Australia, unpub. map data); (4) the proposal that the core complex formed as a result of earlier thrusting in the northwestern Shaw area is precluded by the fact that folds and foliations related to this deformation affect the ca. 3430 Ma South Daltons pluton (age from McNaughton et al., 1993) as well as the <3235 Ma Gorge Creek Group (Van Kranendonk, 2000, 2002).

The model of van Haaften and White (1998) was contested on geometrical and geochronological grounds (Van Kranendonk et al., 2001a), as well as because it ignored previously available geochemical evidence of unique compositional trends across stratigraphy (Hickman, 1980; Van Kranendonk et al., 2001b). More importantly, recent geochronology has confirmed that the greenstones are a continuously upward-younging succession (see Figs. 3, 4). Thus the shear zones along unit contacts identified by van Haaften and White (1998) are viewed as relatively insignificant structures and are more consistent with having formed in response to bedding-parallel slip associated with tilting of greenstones (Van Kranendonk et al., 2001b).

Blewett (2002) reinterpreted some of the linear fabrics in the zone of sinking as the product of intersecting oblique cleavages, but these comprise only a small proportion of the linear fabric elements in the zone, which is also defined by metamorphic mineral elongation lineations (e.g., kyanite, sillimanite, muscovite, chlorite, fuchsite), stretched agglomerate clasts in dacite and/or rhyolite, and fold axes. Intersection of oblique cleavages cannot account for the observed change in plunge of all of these linear fabric elements from shallow to vertical in all directions toward the zone of sinking nor the change in shape of the finite strain ellipsoid as documented by Collins et al. (1998).

3. Later thrusting. Boulter et al. (1987) described evidence for tilting and local folding of lower parts of the Gorge Creek Group in the northwestern Shaw area, and Nijman et al. (1998b, p. 83) described extensional and compressional structures in correlative rocks north of Marble Bar. In the latter, depocenter shifts were interpreted to result from differential doming of adjacent granitoid complexes, whereas subsequent reverse faults were ascribed to "…regional thrusting and interbatholith folding." Although this is one possible interpretation, localized compressional structures are also consistent with diapirism whereby they form from constriction of bed length in greenstone synclines between pairs of rising granitoid diapirs (cf. Dixon and Summers, 1983). Indeed, deposition of the Gorge Creek Group across a transition from extensional to compressional structures is consistent with progressive diapirism (cf. Chardon et al., 1996), although this does not preclude thrusting at this time. Thus, further studies in other parts of the Gorge Creek Group are required in order to resolve the possibility of regional thrusting in the East Pilbara granite-greenstone terrane.

Partial convective overturn of the East Pilbara granite-greenstone terrane: In addition to the specific problems that need to be resolved concerning the proposed models of thrusting for the East Pilbara granite-greenstone terrane, any model involving Alpine-style thrusting is generally at odds with the evidence presented above for an essentially autochthonous stratigraphy and regional contact style of metamorphism in the East Pilbara granite-greenstone terrane. Hickman (1983) summarized the total stratigraphic thickness of individual belts and, although the names of some formations have changed, the thickness estimates of between 9 and 15 km remain valid. This must be considered a minimum, as granitoid rocks intrude the base of greenstone belts, their tops are commonly unconformities with younger groups, and the greenstones have experienced some degree of tectonic flattening. Indeed, an estimate for the stratigraphic thickness of the best preserved part of the stratigraphy across the southwestern flank of the Panorama greenstone belt includes 18 km of the Warrawoona Group, and this is unconformably overlain laterally by 1 to 3 km of the Sulphur Springs Group and 3.5 km of the Gorge Creek Group (Van Kranendonk, 2000). How such a vast thickness of autochthonous volcano-sedimentary stratigraphy (up to 23 km!) can be deposited while the base of the pile remains at relatively low metamorphic grade is a major challenge to any tectonic model for the terrane.

Hickman's (1984) model of long-lived, solid-state diapirism explained how such a thickness of autochthonous stratigraphy could accumulate under low-grade metamorphic conditions, as well as provided an explanation for many other regional features of the terrane. This regional model was supported by more detailed studies around the Mount Edgar Granitoid Complex (Collins, 1989, 1993; Williams and Collins, 1990; Collins et al., 1998). A significant refinement to these models suggested that magmatic diapirism was a major feature of granitoid dome emplacement and accompanied solid-state remobilization of older granitoids and greenstones during partial convective overturn of the upper and middle crust (Collins et al., 1998; Collins and Van Kranendonk, 1999; Van Kranendonk and Collins, 2001, in press). In this model, domical granitoid complexes were interpreted as having developed partly through ductile remobilization of the early, synvolcanic tonalite-trondhjemite-granodiorite suite, and partly through the emplacement of successive generations of plutonic suites (magmatic diapirs) into the cores of progressively evolving domes. The younger granitoid suites were interpreted to derive from partial melting of the older granitoids (Bickle et al., 1989; Collins, 1993), and doming was interpreted to have formed in response to sinking of dense, upper-crustal greenstones into a thermally softened, more buoyant granitoid middle crust during partial convective overturn (Collins et al., 1998).

A two-stage process of early crust formation in the East Pilbara granite-greenstone terrane was identified as responsible for the creation of the density inversion in the crustal section (Collins et al., 1998; Van Kranendonk et al., 2001b). In the first stage (3.49–3.43 Ma), sheeted tonalite-trondhjemite-granodiorite sills were emplaced at a gravitationally stable level within the lower part of the greenstone succession (Coonterunah and lower parts of the Warrawoona Group) and fed felsic volcanics (Duffer and Panorama Formations) from point sources that developed over synvolcanic laccoliths. This uneven emplacement of granitoids and eruption of felsic volcanics instigated lateral variations in lithology and thickness in both the mid-crustal sheeted sill complex and the supracrustal succession, and synvolcanic doming commenced at this time (fig. 15 in Van Kranendonk et al., 2001b).

During stage 2 (3420–3325 Ma), eruption of the thick (5–8 km) Euro Basalt generated a gravitational instability that subsequently drove partial convective overturn between a dense greenstone upper crust and more buoyant granitoid middle crust at ca. 3315 Ma. Areas of more voluminous tonalite-trondhjemite-granodiorite intrusion under felsic volcanoes were the sites of later granitoid upwelling, and flanking areas of thicker greenstone accumulations were the sites of greenstone sinking. The hook-shaped folds, ring faults, and vertical zone of sinking in greenstones flanking the Mount Edgar and Corunna Downs Granitoid Complexes are characteristic of modeled and natural diapirs (e.g., Ramberg, 1967; Dixon and Summers, 1983; Jackson et al., 1990; Chardon et al., 1996) and are a key piece of evidence supporting the diapir hypothesis (Van Kranendonk and Collins, in press).

Overturn of the inverted density profile was accomplished by thermal softening of the middle crustal tonalite-trondhjemite-granodiorite layer. The heat required to melt the tonalite-trondhjemite-granodiorite was derived either from external thermotectonic events (e.g., active extension; cf. Barley and Pickard, 1999), or high heat-producing granitoids buried under greenstones (Rey et al., 2001; Weinburg and Sandiford, 2001). The resultant granitic melts (Bickle et al., 1989; Collins, 1993) were driven out from under sites of greenstone sinking and migrated up the preexisting perturbations in the tonalite-trondhjemite-granodiorite sill complex. These melts accumulated upward and were emplaced as little deformed plutons into the cores of rising granitoid complexes. Once established, subsequent thermotectonic events amplified the dome and basin geometry.

Episodic uplift of the domes during deposition of supracrustal rocks solves the conundrum posed by the presence of a thick autochthonous stratigraphy at low metamorphic grade. This can only be achieved if lower units are uplifted by doming of granitoid complexes while younger units are being deposited, such that basal units never undergo burial by the entire stratigraphic pile (Fig. 13; Hickman, 1984; Van Kranendonk et al., 2001b). Contact metamorphism of greenstones occurred along the margins of progressively more steeply inclined granitoid domes during the intrusion of successive phases of granitoids. The domes thus acted as conduits for the escape of heat from the mantle and lower crust to the surface at punctuated intervals through the history of the region.

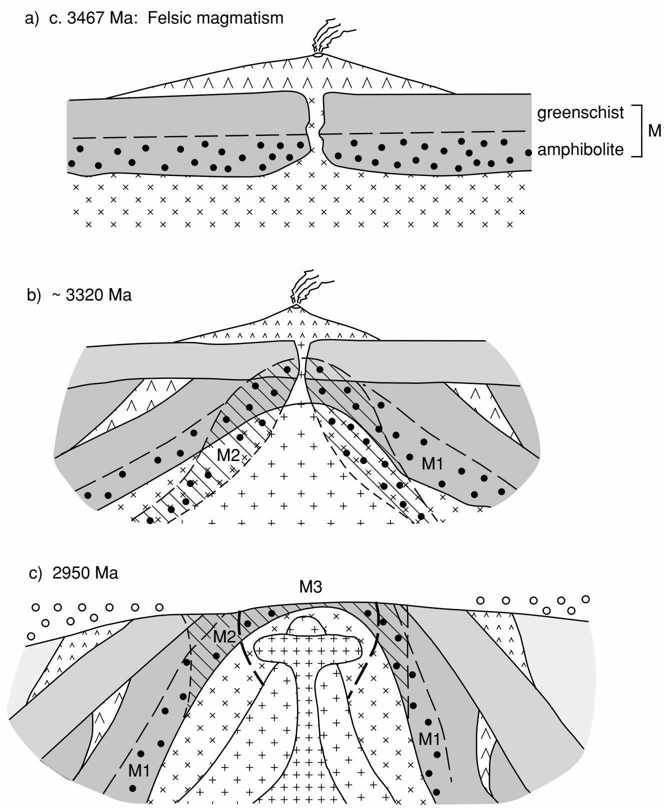

FIG. 13. Model for the metamorphic development of the East Pilbara granite-greenstone terrane. a. Contact metamorphism (M1) of lowermost greenstones (shaded fill) during emplacement of a sheeted sill complex of sodic granitoids (crosses) and eruption of the Duffer Formation (hashed pattern) at ca. 3467 Ma. b. and c. Intervals in the punctuated evolution of the dome and syncline geometry in which thermal aureoles (M2 and M3) from successive generations of granitoid plutons are superimposed on parts of older metamorphic assemblages in the lower greenstones and cause initial metamorphism of younger greenstones. Synchronous doming of granitoid complexes with granite emplacement causes an overlap in the areas of amphibolite-facies contact metamorphic aureoles through time. Doming results in the deposition of greenstone wedges that accumulate laterally in progressively developing basins while basal components of the succession are uplifted such that they never experience high-grade burial metamorphism.

Terrane accretion in the West Pilbara granite-greenstone terrane

Smith et al. (1998) suggested that terrane accretion occurred in the West Pilbara granite-greenstone terrane between ca. 2991 and 2925 Ma. However, although the new data supports the general concept of terrane accretion, new geochronological data indicate that it must have occurred prior to ca. 3020 Ma, the age of deposition of the Cleaverville Formation across the West Pilbara granite-greenstone terrane, Mallina basin, and the western part of the East Pilbara granite-greenstone terrane. Detrital zircon populations in Cleaverville samples support the concept that the domains that contain the Roebourne and Whundo Groups, together with the East Pilbara granite-greenstone terrane, evolved as a coherent unit after ca. 3020 Ma (Smithies et al., 2001).

Krapez and Barley (1987) suggested that the Whim Creek greenstone belt was deposited in a pull-apart basin developed on older sialic crust during sinistral strike-slip faulting. Although some aspects of this model may still apply, new geochronological data indicate that the greenstones were deposited in two distinct stages (ca. 3010 Ma Whim Creek Group and ca. 2970 Ma Bookingarra Group), the younger of which passes eastward into the De Grey Group. Further, Smithies (1998a) showed that the proposed bounding fault along the eastern edge of the proposed pull-apart basin was a post-Fortescue Group structure. Pike et al. (2002) present stratigraphic and geochemical data and discuss the tectonic setting of the Whim Creek and Bookingarra Groups as a possible arc and intracontinental rift assemblage, respectively.

Current Tectonic Model

The data presented herein indicate that the North Pilbara terrain evolved through a long and complex geologic history, which spans 870 m.y. from the oldest xenocrystic zircon at 3.72 Ga to the age of post-tectonic granites at 2.85 Ga. Our current, testable working hypothesis is that the North Pilbara terrain evolved through four principal stages (Fig. 14).

Stage 1 (3.72–3.53 Ga) represents a period of early crust formation in the East Pilbara granite-greenstone terrane with the development of a largely cryptic, partly sialic, basement as

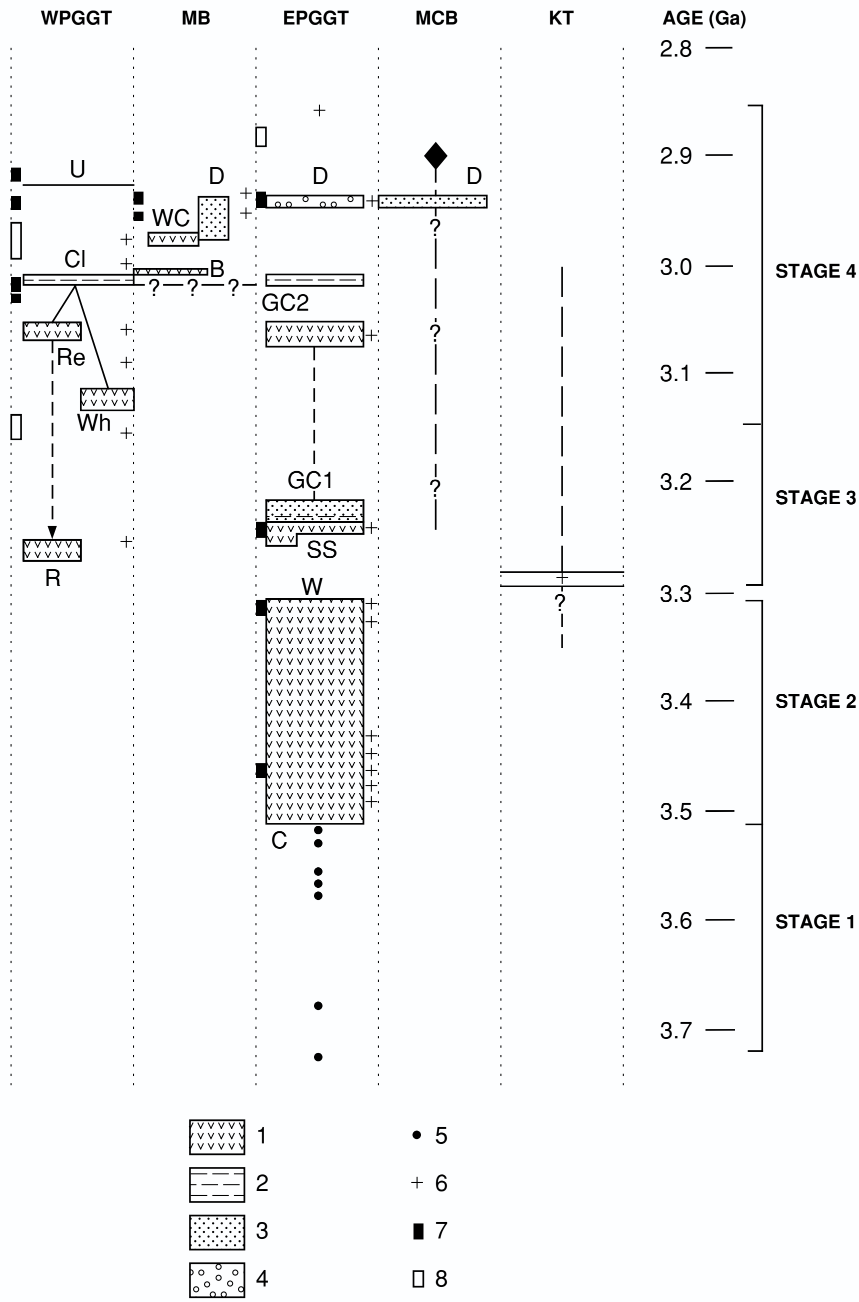

FIG. 14. Schematic time line showing the geologic evolution of the North Pilbara terrain. Columns: WPGGT = West Pilbara granite-greenstone terrane, MB = Mallina basin, EPGGT= Pilbara granite-greenstone terrane, MCB = Mosquito Creek basin, KT = Kuranna terrane. Solid diamond represents Pb-Pb ages of galena in epigenetic Au deposits (from Huston et al., 2002). 1 = volcanic rocks, 2 = shale and banded iron-formation, 3 = sandstone, 4 = conglomerate, 5 = inherited and detrital zircons, 6 = granites, 7 = deformation event—age known, 8 = deformation event—posible age range. C = Coonterunah Group, Cl = Cleaverville Formation, D = De Grey Group, GC1 = Gorge Creek Group, lower (includes Soanesville Subgroup), GC2 = Gorge Creek Group, upper (see text), R = Roebourne Group, Re = Regal Formation, SS = Sulphur Springs Group, U = Ultramafic layered intrusions, W = Warrawoona Group, WC = Whim Creek Group, Wh = Whundo Group.

determined from xenocrystic and detrital zircon populations as well as from xenoliths of gabbroic anorthosite and migmatitic tonalite orthogneiss.

Stage 2 (3.53–3.31 Ga) was a period of effusive, nearly continuous volcanism in the East Pilbara granite-greenstone terrane over 220 m.y., forming an autochthonous stratigraphic succession at least 9 to 18 km thick. The succession is composed of thick (ultra)mafic-felsic volcanic cycles capped by hydrothermal chert precipitates containing evidence of Earth's oldest life. Episodes of felsic volcanism from 3.49 to 3.43 Ga were accompanied by the emplacement of sodic granitoids as a sheeted sill complex at a gravitationally stable level in the mid- to upper crust. Earth's oldest barite, VHMS, and epigenetic Au deposits formed at this time. The subsequent eruption of the 5- to 9-km-thick Euro Basalt created an inverted density profile of the mid- to upper crust in the East Pilbara granite-greenstone terrane. Heating of mid-crustal granitoids caused partial melting of these rocks, resulting in felsic volcanism and monzogranite plutonism that accompanied the onset of partial convective overturn of the mid- to upper crust at 3.32 to 3.31 Ga, with locally associated porphyry Cu-Mo mineralization.

The bimodal composition of the Coonterunah and Warrawoona Groups, the absence of evidence for thrusts between formations, and the geochemical evidence for contamination of even the oldest basalts by felsic crust (e.g., Green et al., 2000) precludes the formation of these groups in a mid-ocean ridge setting. Rather, the evidence for crustal contamination in 3.51 to 3.47 Ga basalts (Green et al., 2000) and in ca. 3.47 Ga granitoid rocks (Bickle et al., 1993), combined with evidence of ancient sialic crust at ca. 3.47 to 3.43 Ga, suggest that the greenstones were erupted onto the ancient basement formed during stage 1.

Although Barley (1981) suggested that the Warrawoona Group formed from independently fractionating calc-alkaline and tholeiitic magmas, and others suggested that the felsic volcanic Duffer Formation and coeval tonalite-trondhjemite-granodiorite sill complex were generated in volcanic- or near-arc settings (Barley, 1993; Bickle et al., 1993; Barley et al., 1998), these modern analogues do not explain repeated mafic-felsic cycles throughout the 220 m.y. history of stage 2 volcanism. Rather, Smithies (2000) showed that Archean tonalite-trondhjemite-granodiorite with characteristic high La/Yb ratios are most likely to be melts of hydrated lower mafic crust generated at depths ≥55 km, which can occur either during flat subduction, when no mantle wedge is generated, or through melting at the base of a crust overthickened by voluminous magmatism. That the latter process may have been important during stage 2 is supported by the great thickness (up to18 km) of generally autochthonous stratigraphy in the group and by total crustal thickness estimates of 55 km or more (Bickle et al., 1985). Some of us (M.V.K. and A.H.H.) consider that the continuous, cyclical nature of the volcanism reflects magmatism derived from nearly continuous melting of the mantle (recurrent mantle plumes?) to generate the large volumes of volcanic rocks, the compositional variation dependent on depth of melting in the mantle (cf. Campbell et al., 1989), the amount of crustal contamination from stage 1 basement, and fractional crystallization. Felsic volcanic horizons higher up in the succession (e.g., the Panorama Formation) may have resulted from melting of the lower, mafic crust in what had already become an anomalously thick volcanic plateau by ca. 3430 Ma.

Stage 3 (3.3–3.07 Ga) commenced with diachronous volcanism and granitoid plutonism across the North Pilbara terrain at ca. 3.29 Ga in the Kuranna terrane, at 3.27 to 3.25 Ga in the West Pilbara granite-greenstone terrane (Roebourne Group and Karratha Granodiorite), and at 3.26 to 3.24 Ga in the East Pilbara granite-greenstone terrane (Sulphur Springs Group and Strelley Granite). The presence of Nd T_{DM} model ages that overlap with the Warrawoona Group of the East Pilbara granite-greenstone terrane in the Roebourne Group and granitoid rocks of the Kuranna terrane suggests that these terranes may be underlain by East Pilbara granite-greenstone terrane crust and/or mantle and may thus represent rifted fragments of the East Pilbara granite-greenstone terrane. Rifting may have occurred during deposition of the Sulphur Springs Group, as a back-arc setting was proposed for the felsic volcanic rocks and associated VHMS mineralization of the group (Vearncombe et al., 1998; Brauhart, 1999; Vearncombe and Kerrich, 1999). Alternatively, the presence of voluminous komatiites in the lower part of the Sulphur Springs Group, combined with the evidence for crustal contamination of felsic magmas, suggest that the group may represent the products of a mantle plume erupted beneath, and contaminated by, continental crust (cf. Dostal and Mueller, 1997) and that rifting followed on from this event.

This period of magmatism was followed by nearly 100 m.y. of relative quiescence across the North Pilbara terrain during which the lower part of the Gorge Creek Group was deposited in the East Pilbara granite-greenstone terrane (Soanesville Subgroup), including locally economic concentrations of banded iron-formation.

Stage 4 (3.15–2.93 Ga) represents a period of rapid, diverse geologic events and widespread deformation, probably as a result of the onset of microplate tectonics (i.e., Smith et al., 1998). Events of this stage commenced with a restricted period of deformation and granitoid plutonism in the West Pilbara granite-greenstone terrane followed by bimodal volcanism in the Whundo Group. The tectonic setting of the Whundo Group is undetermined, pending analysis of geochemical data, but juvenile ε_{Nd} values suggest it is newly formed, uncontaminated crust and thus must have formed either in a rift, or as an island arc, between the Roebourne Group and the East Pilbara granite-greenstone terrane. The Roebourne and Whundo Groups had accreted across the Sholl shear zone by ca. 3.02 Ga, the age of common granitoid rocks and banded iron-formation of the Cleaverville Formation. This was followed by deposition of volcanic and sedimentary rocks in the Mallina basin during possible arc volcanism (Whim Creek Group?) at 3.01 Ga and several episodes of intracontinental rifting interspersed with contractional deformation and plutonism from ca. 2.97 to 2.93 Ga.

Geochemical data from the ca. 2.95 Ga sanukitoid suite in the Mallina basin indicate that the underlying mantle had been affected by subduction prior to this time, which we infer may have occurred at either 3.13 to 3.11 Ga (Whundo Group), 3.07 to 3.05 Ga (during deposition of the upper part of the Gorge Creek Group and lower De Grey Group in the East Pilbara granite-greenstone terrane; i.e., back-arc rift succession), and/or 3.01 Ga (Whim Creek Group). Widespread deformation and granitoid plutonism affected the whole of the West Pilbara granite-greenstone terrane and extended across the western half of the East Pilbara granite-greenstone terrane at 2.935 Ga. This was followed by the emplacement of ultramafic-mafic layered intrusions (2.925 Ga) and dextral shearing (2.92 Ga) in the West Pilbara granite-greenstone terrane and by deformation in the Mosquito Creek basin to 2.90 Ga. Epigenetic Au mineralization accompanied local shear reactivation of the East Pilbara granite-greenstone terrane at 2.89 Ga. Final cratonization was achieved through the emplacement of post-tectonic, highly fractionated granites in the East Pilbara granite-greenstone terrane at 2.85 Ga with Sn-Ta(-Li) mineralization.

Further detailed mapping, geochronology, and geochemistry are required to further establish the history of events in the North Pilbara terrain, particularly in the little-known Mosquito Creek basin and the Kuranna terrane.

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APPENDIX 1

Summary Geochronological Data for the North Pilbara Terrain

Group/ Complex	Formation; lithology; greenstone belt	Sample	Latitude	Longitude	Age (Ma)	Method	Interpretation	Source
Eastern Pilbara granite-greenstone terrane								
Greenstones								
Coonterunah	Coucal; hyaloclastic rhyolite; East Strelley	70601	–21.0709	119.2118	3515 ± 3	U-Pb SHRIMP	Crystallization	Buick et al. (1995)
Coonterunah	Double Bar; chert pebble sandstone fed by hydrothermal veins; Warralong	168993	–21.0705	119.2127	3508 ± 3	U-Pb SHRIMP	Depositional age	Nelson (2002)
Warrawoona	North Star Basalt; pyroxenite lens; Marble Bar	N/A			3490 ± 15	Ar-Ar hornblende	Igneous cooling	van Koolwijk et al. (2001)
Warrawoona	North Star Basalt; pillowed metabasalt; Marble Bar	UWA98078	–21.0192	119.8008	3326 ± 10	U-Pb SHRIMP	Hydrothermal	McNaughton et al. (1993)
Warrawoona	Dresser; barite; Panorama	Pb415			ca. 3490	Pb-Pb galena	Barite crystallization	Thorpe et al. (1992b)
Warrawoona	McPhee; silicified felsic tuff; Marble Bar	148498	–21.0531	119.7858	3477 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	McPhee; dolerite sill; Marble Bar	UWA98077	–20.9947	119.8136	3487 ± 8	U-Pb SHRIMP	Xenocrystic or igneous	McNaughton et al. (1993)
					3432 ± 6		Igneous or hydrothermal	
					3308 ± 4		Probable hydrothermal	
					3238 ± 6		Probable hydrothermal	
Warrawoona	Mount Ada Basalt; lapilli tuff; Marble Bar	148500	–20.9931	119.8219	3469 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1999)
Warrawoona	Duffer; crystal lithic tuff; Coongan	142975	–21.3669	119.5786	3474 ± 7	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	Duffer; felsic volcanic; Marble Bar	100512	–20.9372	120.1694	3471 ± 5	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
Warrawoona	Duffer; volcaniclastic sandstone; Coongan	168918	–21.3665	119.5788	3470 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Duffer; subvolcanic porphyritic granophyre intrusion in Mount Ada Basalt; Coongan	142976	–21.3886	119.5719	3469 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	Duffer; porphyritic tuffaceous andesite; Marble Bar	148509	–21.0989	119.7750	3468 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	Duffer; felsic agglomerate; Coongan	168921	–21.3777	119.6047	3468 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Duffer; columnar-jointed porphyritic dacite; Coongan	168920	–21.3652	119.6018	3467 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Duffer; rheomorphic ignimbrite; North Shaw	142964	–21.4475	119.3036	3466 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	Duffer; intermediate pyroclastic; Marble Bar	UWA98076	–21.1075	119.7578	3466 ± 4	U-Pb SHRIMP	Crystallization	McNaughton et al. (1993)
Warrawoona	Duffer; felsic schist; SE Marble Bar	100515	–21.4375	120.1347	3465 ± 3	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
Warrawoona	Duffer; porphyritic dacite; Marble Bar	ANU89-332	–20.9533	119.8497	3463 ± 2	U-Pb SHRIMP	Crystallization	McNaughton et al. (1993)
Warrawoona	Panorama; felsic lava; Panorama	100507	–21.2653	119.3919	3458 ± 2	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
					3724 ± 1	U-Pb conventional	Xenocryst	
Warrawoona	Panorama; felsic schist; NE Marble Bar	100511	–20.8965	120.1134	3458 ± 2	U-Pb conventional	Crystallization	Williams (1999)
Warrawoona	Panorama; quartz porphyritic felsic volcanic; NE Marble Bar	94770	–20.9211	120.1836	3454 ± 1	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
					3470 ± 2		Xenocryst	
Warrawoona	Panorama; felsic sill in Mount Ada Basalt; Marble Bar	94750	–21.1161	119.8031	3449 ± 3	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
Warrawoona	Panorama; quartz porphyry intrusion in Apex Basalt; Panorama	103283	–21.1506	119.7383	3449 ± 2	U-Pb conventional	Crystallization	Thorpe et al. (1992a)

APPENDIX 1 *(Cont.)*

Group/ Complex	Formation; lithology; greenstone belt	Sample	Latitude	Longitude	Age (Ma)	Method	Interpretation	Source
Warrawoona	Panorama; felsic volcanic; Marble Bar	N/A	–20.8987	120.0807	3446 ± 2	U-Pb SHRIMP	Crystallization	J. Wijbrans and W. Nijman, unpublished data, in Van Kranendonk et al. (2001b)
Warrawoona	Panorama; quartz-porphyritic felsic tuff; Panorama	142952	–21.0042	119.3783	3434 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	Panorama; porphyritic tuffaceous rhyolite; Coongan	148502	–21.4706	120.0547	3433 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2000)
Warrawoona	Panorama; volcaniclastic; Western Shaw	142972	–21.8528	119.2222	3433 ± 6	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Panorama; quartz phenocrystic rhyolite; east Coongan	168915	–21.5683	119.7334	3432 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Panorama; rhyolitic tuff; west Coongan	160221	–21.4687	119.6486	3430 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2002)
Warrawoona	Panorama; dacite; Kelly	MP1	–21.6206	120.1755	3430 ± 3	U-Pb SHRIMP	Crystallization	Barley et al. (1998)
Warrawoona	Panorama; dacitic agglomerate; Kelly	168914	–21.4609	120.0456	3428 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Panorama; plagioclase-porphyritic dacite; Kelly	168913	–21.4616	120.0444	3427 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2001)
Warrawoona	Panorama; rhyodacite flow; Kelly	K1	–21.5753	120.0114	3417 ± 9 ca. 3304	U-Pb SHRIMP	Crystallization Metamorphic	Barley et al. (1998)
Warrawoona	Euro Basalt; silicified felsic tuff; Kelly	168909	–21.6548	119.9633	3363 ± 6 3346 ± 6 3311 ± 6	U-Pb SHRIMP	Xenocryst Crystallization Metamorphic	data from Nelson (2001), interpretation by L. Bagas and M. Van Kranendonk, GSWA
Warrawoona	Euro Basalt; silicified felsic tuff; Panorama	94001	–21.3066	119.2818	3463 ± 7 ca. 3515 3547 ± 6	U-Pb SHRIMP	Detrital Detrital Detrital	Buick et al. (1995)
Warrawoona	Wyman; columnar-jointed rhyolite flow; Kelly	UWA98074	–21.8411	119.8625	3325 ± 4	U-Pb SHRIMP	Crystallization	McNaughton et al. (1993)
Warrawoona	Wyman; quartz-porphyritic felsic lava; NW Coongan	94754	–21.7042	119.9883	3325 ± 3	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
Warrawoona	Wyman; Kelly subvolcanic porphyry; SE Kelly	UWA98075	–21.8056	119.8536	3324 ± 4	U-Pb SHRIMP	Crystallization	McNaughton et al. (1993)
Warrawoona	Wyman; felsic volcaniclastic; Kelly	168910	–21.8691	119.8204	3323 ± 3	U-Pb SHRIMP	Maximum deposition	Nelson (2001)
Warrawoona	Wyman; columnar-jointed, quartz-feldspar rhyolitic porphyry; NW Kelly	94003	–21.3187	119.7697	3315 ± 3	U-Pb SHRIMP	Crystallization	Buick et al. (2001)
Warrawoona	Wyman; volcaniclastic sandstone; southern Coongan	169000	–21.8720	119.7452	3312 ± 4	U-Pb SHRIMP	Maximum deposition	Nelson (2002)
Warrawoona	Budjan Creek; crystal-lithic tuff; Kelly	168908	–21.6689	119.9826	3308 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2001)
Sulphur Springs	Leilira; felsic volcaniclastic arenite; Soanesville	60925	–21.1626	119.1557	3255 ± 4 ~3450 ~3510	U-Pb SHRIMP	Crystallization Inherited Inherited	Buick et al. (2002)
Sulphur Springs	Kangaroo Caves; silicified rhyolite breccia; Soanesville	72263	–21.2301	119.2782	~3240	U-Pb SHRIMP	Crystallization	Buick et al. (2002)
Sulphur Springs	Kangaroo Caves; dacite; Soanesville	111868	–21.4294	119.0734	3240 ± 2	U-Pb conventional	Crystallization	R. Thorpe, written comm., 1996
Sulphur Springs	Strelley Granite synvolcanic laccolith, outer phase; Soanesville	RM1	–21.1706	119.1507	3239 ± 5	U-Pb SHRIMP	Crystallization	Buick et al. (2002)
Sulphur Springs	Strelley Granite synvolcanic laccolith, outer phase; Soanesville	203366	–21.2784	119.1044	3239 ± 2	U-Pb SHRIMP	Crystallization	Buick et al. (2002)
Sulphur Springs	Strelley Granite synvolcanic laccolith, inner phase; Soanesville	203368	–21.2133	119.1929	3238 ± 3	U-Pb SHRIMP	Crystallization	Buick et al. (2002)
Sulphur Springs	Kangaroo Caves; rhyolite; Soanesville	72081	–21.1903	119.2292	3238 ± 3	U-Pb SHRIMP	Crystallization	Buick et al. (2002)
Sulphur Springs	Kangaroo Caves; spherulitic amygdaloidal andesite; Soanesville	70727	–21.2776	119.0772	3235 ± 4	U-Pb SHRIMP	Crystallization	Buick et al. (2002)

APPENDIX 1 *(Cont.)*

Group/ Complex	Formation; lithology; greenstone belt	Sample	Latitude	Longitude	Age (Ma)	Method	Interpretation	Source
Sulphur Springs	Kangaroo Caves; rhyodacite breccia; Soanesville	94002	–21.2336	119.2371	3235 ± 3	U-Pb SHRIMP	Crystallization	Buick et al. (2002)
Gorge Creek	Unassigned; quartzite; Warralong	142836	–20.8531	119.4850	3426 ± 10	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Gorge Creek	Nimingarra Iron Formation; basal quartzite; Shay Gap	143995	–20.7547	120.1992	3403 ± 10	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Gorge Creek	Nimingarra Iron Formation; quartzite; NE Marble Bar	143994	–20.8831	120.0881	3362 ± 13	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Gorge Creek	Cundaline; quartzite; Shay Gap	143996	–20.5356	120.1789	3398 ± 15	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Gorge Creek	Cleaverville; felsic sediments; Pilbara Well	142842	–20.9603	117.9150	3016 ± 5	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
De Grey	Lalla Rookh Sandstone; pebbly conglomerate; Soanesville	142951	–21.0053	119.3550	3304 ± 6	U-Pb SHRIMP	Maximum deposition	Nelson (2000)
De Grey	Cattle Well; volcaniclastic sandstone; Shay Gap	142867	–20.5167	120.2067	3048 ± 19	U-Pb SHRIMP	Maximum deposition	Nelson (1999)
Granitoid complexes								
Carlindi	Biotite granodiorite	153188	–21.0464	119.0835	3484 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1999)
Carlindi	Leucogranite	153190	–21.0460	119.0416	3469 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1999)
Carlindi	Porphyritic microgranite	95058	–21.1331	119.0028	3468 ± 4	U-Pb SHRIMP	Crystallization	Buick et al. (1995)
Carlindi	Foliated biotite granodiorite in Tabba Tabba shear zone	160745	–20.5410	119.0122	2940 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2001)
Corunna Downs	Monzogranite	160211	–21.4824	119.8143	3427 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2002)
Corunna Downs	Mondana suite, biotite monzogranite	Mond	–21.6303	119.9444	3317 ± 2	U-Pb SHRIMP	Crystallization	Barley and Pickard (1999)
Corunna Downs	Carbana Pool suite, porphyritic biotite monzogranite	Carb	–21.5364	119.9425	3313 ± 9	U-Pb SHRIMP	Crystallization	Barley and Pickard (1999)
Corunna Downs	Monzogranite	160212	–21.4644	119.8298	3313 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2002)
Corunna Downs	Carbana Pool porphyritic monzogranite	142978	–21.5522	119.9156	3307 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Corunna Downs	Porphyritic biotite tonalite	168922	–21.4266	119.8415	3307 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2001)
Mt Edgar	Fl-bearing, hypersolvus alkali granite, subvolcanic to the Duffer Formation	142865	–21.1694	119.7609	3466 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Mt Edgar	Mafic trondhjemitic gneiss	LTU6419			3448 ± 8	U-Pb SHRIMP	Crystallization	Williams and Collins (1990)
Mt Edgar	Felsic trondjemitic gneiss	LTU6417			3443 ± 10	U-Pb SHRIMP	Crystallization	Williams and Collins (1990)
Mt Edgar	Granitoid	169031	–21.2148	120.0130	3430 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2002)
Mt Edgar	Banded tonalitic gneiss	LTU5416			3429 ± 13	U-Pb SHRIMP	Crystallization	Williams and Collins (1990)
Mt Edgar	Wilina pluton	N/A			3324 ± 6	U-Pb SHRIMP	Crystallization	Collins et al. (1998)
Mt Edgar	Orthogneiss	142984	–21.1850	120.0306	3321 ± 6	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Rhyolitic porphyry	103279	–20.8901	120.1078	3317 ± 1	U-Pb conventional on titanite	Crystallization	R. Thorpe, written comm. (1992)
Mt Edgar	Pegmatite-banded orthogneiss	142985	–21.1881	119.9631	3315 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Granitoid	169038	–21.2675	120.2205	3315 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2002)
Mt Edgar	Coppin Gap Granodiorite	LTU5577	–20.9763	120.0831	3314 ± 13	U-Pb SHRIMP	Crystallization	Williams and Collins (1990)
Mt Edgar	Foliated biotite-hornblende tonalite	142981	–21.2011	120.1547	3312 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Hornblende granodiorite	142974	–21.1369	119.8103	3310 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Foliated biotite granodiorite	142982	–21.1947	120.1014	3305 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Boodallana suite, biotite tonalite	LTU5472			3304 ± 10	U-Pb SHRIMP	Crystallization	Williams and Collins (1990)
Mt Edgar	Magnetite monzogranite	142980	–21.2011	120.1547	3243 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Foliated biotite monzogranite	142983	–21.1950	120.0844	3241 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2000)
Mt Edgar	Moolyella Monzogranite	142977	–21.1464	119.9039	3240 ± 11	U-Pb SHRIMP	Inherited	Nelson (2000)
Mt Edgar	Small granodiorite pluton	142825	–20.9672	120.1398	2757 ± 7	U-Pb SHRIMP	Crystallization	Nelson (1998)
Muccan	Heterogeneous banded gneiss	142828	–20.8225	120.1089	3470 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1998)
Muccan	Sunrise Hill monzogranite	124755	–20.4481	120.0383	3443 ± 6	U-Pb SHRIMP	Crystallization	Nelson (1996)

APPENDIX 1 *(Cont.)*

Group/ Complex	Formation; lithology; greenstone belt	Sample	Latitude	Longitude	Age (Ma)	Method	Interpretation	Source
Muccan	Kennedy Gap tonalite	143807	–20.5757	120.2829	3438 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1998)
Muccan	Monzogranite	143803	–20.5433	120.0828	3313 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Muccan	Monzogranite	143806	–20.7286	120.0098	3303 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Muccan	Wolline Monzogranite	143805	–20.6058	120.0403	3252 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Muccan	Wolline Monzogranite	143810	–20.7952	120.0645	3244 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Pippingarra	Sanukitoid	142935	–20.7458	118.8156	2954 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Pippingarra	Chillerina Granodiorite	160730	–20.6532	118.7054	2946 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2001)
Pippingarra	Chillerina Granodiorite	160744	–20.6637	118.7923	2945 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2001)
Pippingarra	Biotite leucogranite	160728	–20.6637	118.7923	2940 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2001)
Pippingarra	Biotite leucogranite	160727	–20.6637	118.7923	2928 ± 6	U-Pb SHRIMP	Crystallization	Nelson (2001)
Shaw	Gabbroic anorthosite xenolith, South Daltons pluton	N/A			3578 ± 4	U-Pb SHRIMP	Crystallization	McNaughton et al. (1988)
Shaw	North Shaw suite	N/A			3499 ± 22	Pb-Pb whole-rock isochron	Crystallization	Bickle et al. (1983)
Shaw	North Shaw suite	91084			3493 ± 4	U-Pb SHRIMP	Crystallization	McNaughton et al. (1988)
Shaw	Gray tonalitic gneiss	N/A	–21.7572	119.2381	3485 ± 30	U-Pb SHRIMP	Crystallization	Williams et al. (1983)
Shaw	Gray tonalitic gneiss	N/A			3470 ± 25	U-Pb SHRIMP	Crystallization	Williams et al. (1983)
Shaw	Coolyia Creek Granodiorite	142962	–21.3869	119.3531	3469 ± 2	U-Pb SHRIMP	Crystallization	Nelson (2000)
Shaw	Mylonitic granite	T94/227	–21.4918	119.5827	3469 ± 3	U-Pb SHRIMP	Crystallization	Zegers et al. (2001)
Shaw	Biotite granodiorite	T94/193	–21.8311	119.5975	3468 ± 3	U-Pb SHRIMP	Crystallization	Zegers et al. (2001)
Shaw	Coolyia Creek Granodiorite	91085			3467 ± 6	U-Pb SHRIMP	Crystallization	McNaughton et al. (1988)
Shaw	Quartz diorite	T94/222	–21.8440	119.3985	3463 ± 2	U-Pb SHRIMP	Crystallization	Zegers et al. (2001)
Shaw	Hornblende quartz diorite	168923	–21.8216	119.3460	3462 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2001)
Shaw	Biotite-granodiorite gneiss	T94/221	–21.8440	119.3985	3451 ± 1	U-Pb SHRIMP	Crystallization	Zegers et al. (2001)
Shaw	Leucogranite	142878	–21.4475	119.5325	3445 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Shaw	South Daltons pluton, granodiorite	UWA98053	–21.5253	119.2417	3431 ± 4	U-Pb SHRIMP	Crystallization	McNaughton et al. (1993)
Shaw	Schlieric granodiorite	142966	–21.6972	119.3311	3430 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Shaw	Homogeneous granodiorite	M95-78	–21.6086	119.2836	~3430	U-Pb SHRIMP	Crystallization	M. Van Kranen-donk, unpublished data
Shaw	Blue-gray tonalite	142968	–21.7572	119.2381	3425 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Shaw	Leucogranite diatexite	142881	–21.4737	119.3451	~3410	U-Pb SHRIMP	diatexis	D. Nelson, written comm., 1998
Shaw	Syn-kinematic biotite granite dike	T94/31	–21.5424	119.3366	2934 ± 2	U-Pb SHRIMP	Crystallization	Zegers et al. (2001)
Shaw	Monzogranite dike	142965	–21.6972	119.3311	2929 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Shaw	Biotite monzogranite	142967	–21.6972	119.3311	2928 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2000)
Shaw	Mulgandinnah Monzogranite	142882	–21.4855	119.3405	2928 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Shaw	Porphyritic monzogranite	142883	–21.5070	119.3833	2919 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
	dike in dextral shear zone	142879	–21.5207	119.4014	2851 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Shaw	Spear Hill Monzogranite							
Warrawagine	Folded tonalite gneiss inclusion	142870	–20.8128	120.5675	3655 ± 6	U-Pb SHRIMP	Igneous or xenocryst	Nelson (1999)
					3637 ± 12		Igneous or xenocryst	
					3595 ± 4		Igneous or xenocryst	
					3576 ± 6		Probable igneous	
					3410 ± 7		Metamorphic	
Warrawagine	Biotite monzogranite	143809	–20.7853	120.4183	3313 ± 6	U-Pb SHRIMP	Crystallization	Nelson (1998)
Warrawagine	Porphyritic biotite granodiorite	142871	–20.8958	120.6039	3303 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1999)
Warrawagine	Porphyritic biotite granodiorite	142869	–20.8144	120.5675	3244 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1999)
Warrawagine	Hornblende granodiorite	142874	–20.8272	120.5247	3242 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1999)
Warrawagine	Quartz-feldspar porphyry dike	142875	–20.6925	120.5203	2758 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1999)
Yule	Gneissic tonalite	169019	–21.9701	119.0080	3428 ± 6	U-Pb SHRIMP	Crystallization	Nelson (2002)
Yule	Foliated biotite monzogranite	142170	–21.0914	118.5125	3421 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1999)
Yule	Tonalite	142948	–21.6783	117.8739	3166 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2000)
Yule	Foliated tonalite	142946	–21.6619	117.8717	3164 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Yule	Biotite hornblende tonalite	142938	–21.3647	118.2047	2945 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2000)

APPENDIX 1 *(Cont.)*

Group/ Complex	Formation; lithology; greenstone belt	Sample	Latitude	Longitude	Age (Ma)	Method	Interpretation	Source
Yule	Foliated, porphyritic biotite monzogranite	142176	–21.2711	118.4836	2938 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1999)
Yule	Leucocratic monzogranite	142936	–21.2353	118.4042	2937 ± 7	U-Pb SHRIMP	Crystallization	Nelson (2000)
Yule	Leucocratic monzogranite	142937	–21.3244	118.3933	2935 ± 3	U-Pb SHRIMP	Crystallization	Nelson (2000)
Yule	Foliated biotite monzogranite	160442	–21.2353	118.4042	2933 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
Yule	Schlieric biotite syenogranite emplaced into fold nose of Tambourah dome	142884	–21.6503	119.1406	2933 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Yule	Syntectonic biotite monzogranite	142885	–21.6617	119.0946	2927 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
	North Pole Monzogranite; Panorama	100510	–21.1148	119.3751	3459 ± 18	U-Pb conventional	Crystallization	Thorpe et al. (1992a)
	Copper Hills porphyry; Kelly	86473	–21.6604	119.9778	3321 ± 4	U-Pb SHRIMP	Crystallization	Barley and Pickard (1999)
	Boobina porphyry; Kelly	86483	–21.6778	119.9588	3315 ± 4	U-Pb SHRIMP	Crystallization	Barley and Pickard (1999)
	Gobbos stock; granodiorite; McPhee	86358	–21.5531	120.2936	3313 ± 4	U-Pb SHRIMP	Crystallization	Barley and Pickard (1999)
	Unnamed porphyry; Pilbara Well	142941	–21.2075	118.3467	2946 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
	Keep It Dark Monzogranite; North Shaw	M95-327	–21.4410	119.2906	2936 ± 5	U-Pb SHRIMP	Crystallization	M. Van Kranendonk, unpublished data
West Pilbara granite-greenstone terrane								
Greenstones								
Roebourne	Nickol River; volcanigenic metasediment	136819	–20.7508	117.0056	3269 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Roebourne	Nickol River; porphyritic dacite	118975	–20.8186	116.7588	3251 ± 6	U-Pb SHRIMP	Crystallization	Nelson (1997)
Whundo	Nallana; schistose metadacite	114350	–20.9364	116.9189	3125 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1996)
Whundo	Tozer; porphyritic rhyolite	114358	–20.9303	116.9222	3122 ± 7	U-Pb SHRIMP	Crystallization	Nelson (1997)
Whundo	Tozer; rhyolite	114356	–20.9525	116.9275	3118 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1996)
Whundo	Woodbrook; rhyolite tuff	144256	–20.9067	117.3369	3118 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Whundo	Woodbrook; welded felsic tuff	127378	–20.9194	117.0842	3117 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Whundo	Bradley Basalt; metamorphosed rhyodacite	144210	–20.9258	117.4650	3116 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Whundo	Bradley Basalt; tuffaceous rhyolite	114305	–20.9250	117.0139	3115 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1996)
Whundo	Tozer; cleaved coarse felsic tuff	W197	–20.9333	116.9833	3112 ± 6	U-Pb conventional	Crystallization	Horwitz and Pidgeon (1993)
Gorge Creek	Cleaverville; volcaniclastic sediment	127330	–20.6594	117.0222	3022 ± 12	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Gorge Creek	Cleaverville; volcaniclastic sediment	142830	–20.9206	117.1620	3018 ± 3	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Gorge Creek	Cleaverville; volcaniclastic sediment	136899	–20.6881	117.0853	3015 ± 5	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
Granitoid complexes								
Caines Well	Biotite monzogranite gneiss	118965	–20.8686	117.5986	3093 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1997)
Caines Well	Porphyritic biotite monzogranite	142950	–20.8233	117.7619	2990 ± 5	U-Pb SHRIMP	Crystallization	Nelson (2000)
Caines Well	Foliated granite	118964	–20.8475	117.7825	2925 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1997)
Cherratta	Foliated hornblende-biotite tonalite	142535	–21.3894	116.5031	3236 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Cherratta	Tonalitic gneiss	142835	–21.1117	116.9233	3130 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1999)
Cherratta	Foliated migmatitic tonalite	JS33	–21.1100	116.8988	3114 ± 5	U-Pb SHRIMP	Crystallization	Smith et al. (1998)
Cherratta	Foliated biotite tonalite	142661	–21.1454	116.8696	3068 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1998)
Cherratta	Post-tectonic granitoid	JS35	–21.0954	116.9223	3013 ± 4	U-Pb SHRIMP	Crystallization	Smith et al. (1998)
Cherratta	Porphyritic granodiorite	168932	–21.1161	117.1848	3006 ± 12	U-Pb SHRIMP	Crystallization	Nelson (2001)
Cherratta	Biotite-hornblende tonalite gneiss	136826	–20.9769	116.7664	2995 ± 11	U-Pb SHRIMP	Crystallization	Nelson (1997)
Cherratta	Foliated porphyritic hornblende granodiorite	118974	–20.8871	116.6673	2994 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1997)
Cherratta	Granodiorite	142657	–21.1200	116.7973	2990 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1999)
Cherratta	Granodiorite	142438	–21.0867	116.8672	2988 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1999)
Cherratta	Biotite monzogranite	168934	–21.2115	117.0723	2988 ± 7	U-Pb SHRIMP	Crystallization	Nelson (2001)

APPENDIX 1 *(Cont.)*

Group/ Complex	Formation; lithology; greenstone belt	Sample	Latitude	Longitude	Age (Ma)	Method	Interpretation	Source
Dampier	Granite	136844	–20.7564	116.7214	2997 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1998)
Dampier	Schlieric, pegmatite-veined monzogranite	142893	–21.0461	116.2422	2982 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1999)
Harding	Monzodiorite	168936	–20.8601	117.1116	3016 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2001)
Harding	Forrestier Bay gneiss	118966	–20.7461	117.6956	3014 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1997)
Harding	Monzogranite	142430	–20.8778	117.0750	2970 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1999)
Felsic intrusions								
Roebourne	Karratha Granodiorite; Ruth Well	142433	–20.8328	116.7803	3270 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Roebourne	Karratha Granodiorite; Ruth Well	JS43	–20.8792	116.9729	3265 ± 4	U-Pb SHRIMP	Crystallization	Smith et al. (1998)
Roebourne	Karratha Granodiorite; Ruth Well	JS17	–20.8295	116.8512	3261 ± 4	U-Pb SHRIMP	Crystallization	Smith et al. (1998)
Whundo	Foliated monzogranite	JS20	–20.8946	116.8406	3114 ± 5	U-Pb SHRIMP	Crystallization	Smith et al. (1998)
	Mylonitic tonalite in Sholl shear zone	JS25	–20.8964	116.9501	3024 ± 10	U-Pb SHRIMP	Crystallization	Smith et al (1998)
Roebourne	Dacite sill in Nickol River Fm.	118976	–20.8867	116.9578	3023 ± 9	U-Pb SHRIMP	Crystallization	Nelson (1997)
Roebourne	Porphyry intrusion; Ruth Well	144224	–20.7186	117.1897	3021 ± 3	U-Pb SHRIMP	Crystallization	Nelson (1999)
Roebourne	Dacite porphyry sill; Regal	127327	–20.6997	117.0142	3018 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1998)
Gorge Creek	Quartz granophyre sill; Cleaverville	127320	–20.9167	117.1625	3014 ± 6	U-Pb SHRIMP	Crystallization	Nelson (1997)
Roebourne	Deformed quartz-feldspar porphyry; Ruth Well	118979	–20.8625	116.9706	3014 ± 2	U-Pb SHRIMP	Crystallization	Nelson (1997)
Ultramafic intrusions								
Munni Munni	Magnetite ferrogabbro pegmatite	103227	–21.1166	116.8500	2925 ± 16	U-Pb SHRIMP	Crystallization	Arndt et al. (1991)
Munni Munni	Granite dike	142436	–21.1197	116.8617	2924 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1998)
<u>Mallina basin</u>								
Supracrustal rocks								
Whim Creek	Mons Cupri; volcanolithic sandstone	168924	–20.8576	117.8419	3016 ± 3	U-Pb SHRIMP	Maximum deposition	Nelson (2001)
Whim Creek	Mons Cupri; welded vitric tuff	141936	–20.9620	117.5224	3009 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1998)
Bookingarra	Cistern; metasandstone	142949	–20.8761	117.8331	2978 ± 5	U-Pb SHRIMP	Maximum deposition	Nelson (2000)
Bookingarra	Kialrah Rhyolite; plagioclase-porphyritic rhyolite	144261	–20.9639	117.3522	2975 ± 4	U-Pb SHRIMP	Maximum deposition	Nelson (1998)
De Grey	Constantine Sandstone	142943	–21.1214	117.8070	3201 ± 17	U-Pb SHRIMP	Maximum deposition	Nelson (2000)
De Grey	Constantine Sandstone	142942	–22.9308	117.8564	2994 ± 4	U-Pb SHRIMP	Maximum deposition	Nelson (2000)
De Grey	Mallina; fine-grained graywacke	118969	–20.9533	117.8431	2997 ± 20	U-Pb SHRIMP	Maximum deposition	Nelson (1997)
De Grey	Mallina; subarkose	142188	–21.0909	118.2464	2941 ± 9	U-Pb SHRIMP	Maximum deposition	Nelson (1999)
Granitoid rocks								
	Opaline Well Monzogranite	118972	–20.9869	117.7211	2765 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1997)
	Peawah Granodiorite (sanukitoid)	118967	–20.9886	117.9914	2948 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1997)
	Plagioclase porphyry high Mg diorite (sanukitoid)	142945	–21.4653	117.9347	2946 ± 20	U-Pb SHRIMP	Crystallization	Nelson (2000)
	Portree (alkali-granite)	142889	–20.8308	118.0678	2946 ± 6	U-Pb SHRIMP	Crystallization	Nelson (1999)
	High Mg diorite	160498	–20.9879	117.7746	2945 ± 6	U-Pb SHRIMP	Crystallization	Nelson (2000)
	Porphyry	142892	–21.0678	118.1386	2941 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1999)
	Biotite monzogranite	142934	–20.6836	118.6389	2941 ± 4	U-Pb SHRIMP	Crystallization	Nelson (2000)
	Satirist Granite	141973	–21.0919	117.9811	2938 ± 4	U-Pb SHRIMP	Crystallization	Nelson (1998)
	Satirist Granite	141977	–21.2028	117.9811	2931 ± 5	U-Pb SHRIMP	Crystallization	Nelson (1998)