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ARCHEAN CRUSTAL EVOLUTION AND MINERALIZATION OF THE NORTHERN PILBARA CRATON — A FIELD GUIDE

by AH Hickman and MJ Van Kranendonk

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by
AH Hickman and MJ Van Kranendonk

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Archean crustal evolution and mineralization of the northern Pilbara Craton — a field guide

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Preface

This guide was prepared to support a six-day excursion across the northern Pilbara Craton (Figs 1 and 2). It describes and compares the 3525–3165 Ma East Pilbara Terrane and the 3275–3060 Ma West Pilbara Superterrane. Once thought to be geologically similar (Hickman, 1983), these two parts of the northern Pilbara Craton are now interpreted to have evolved very differently: formation of the East Pilbara Terrane involved several mantle plumes with both vertical and horizontal tectonics, and the West Pilbara Superterrane is a product of plate-tectonic processes. The excursion also includes a brief examination of the 3020–2930 Ma De Grey Superbasin, which unconformably overlies the older terranes, and covers the different styles of Archean mineralization in the Pilbara Craton.

There are two parts to this guide. **Part 1** describes the geology and mineralization of the northern Pilbara Craton and **Part 2** provides descriptions of excursion localities and directions for travel between them.

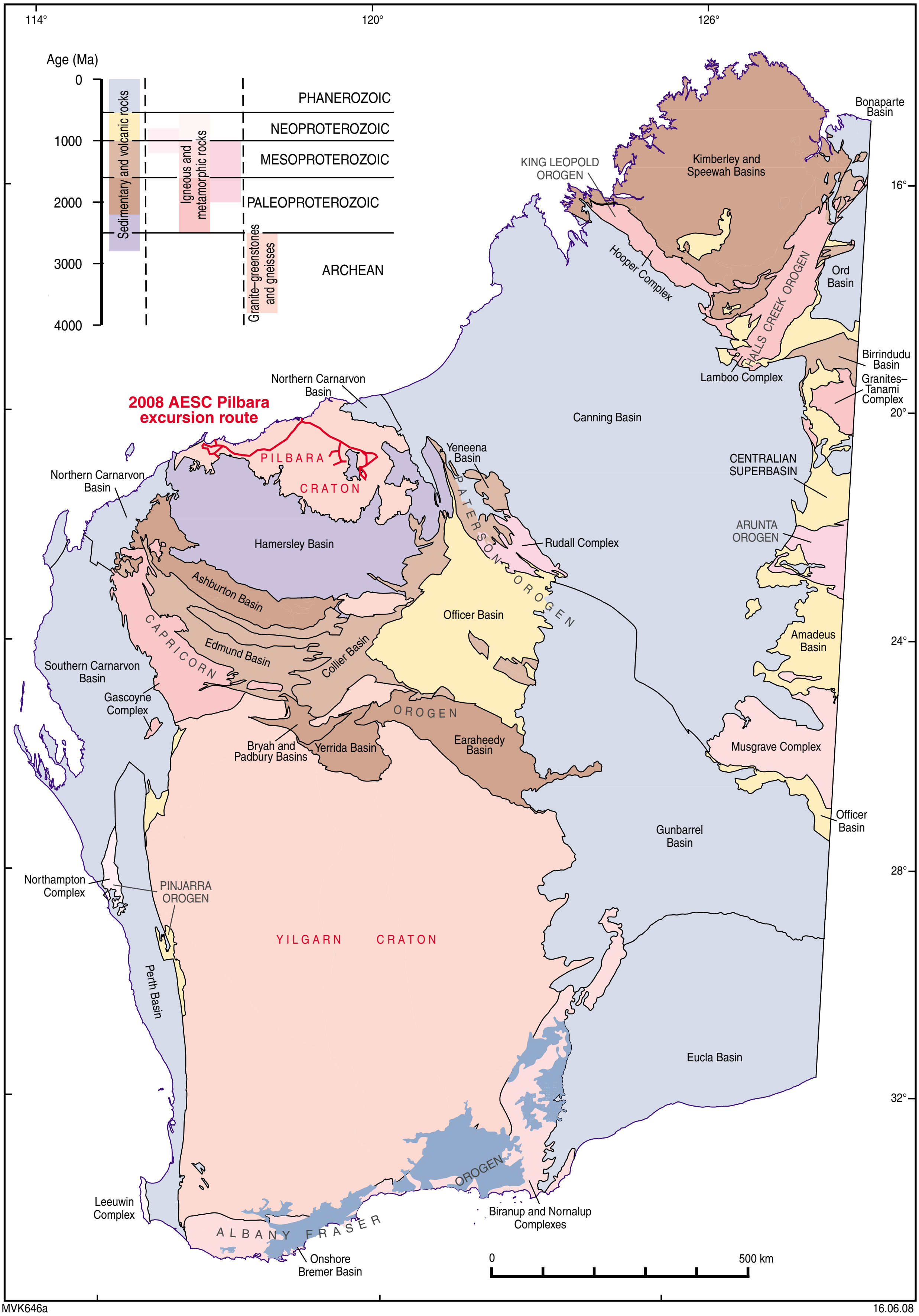

Figure 1. Map showing tectonic units of Western Australia and the excursion route covered by this field guide

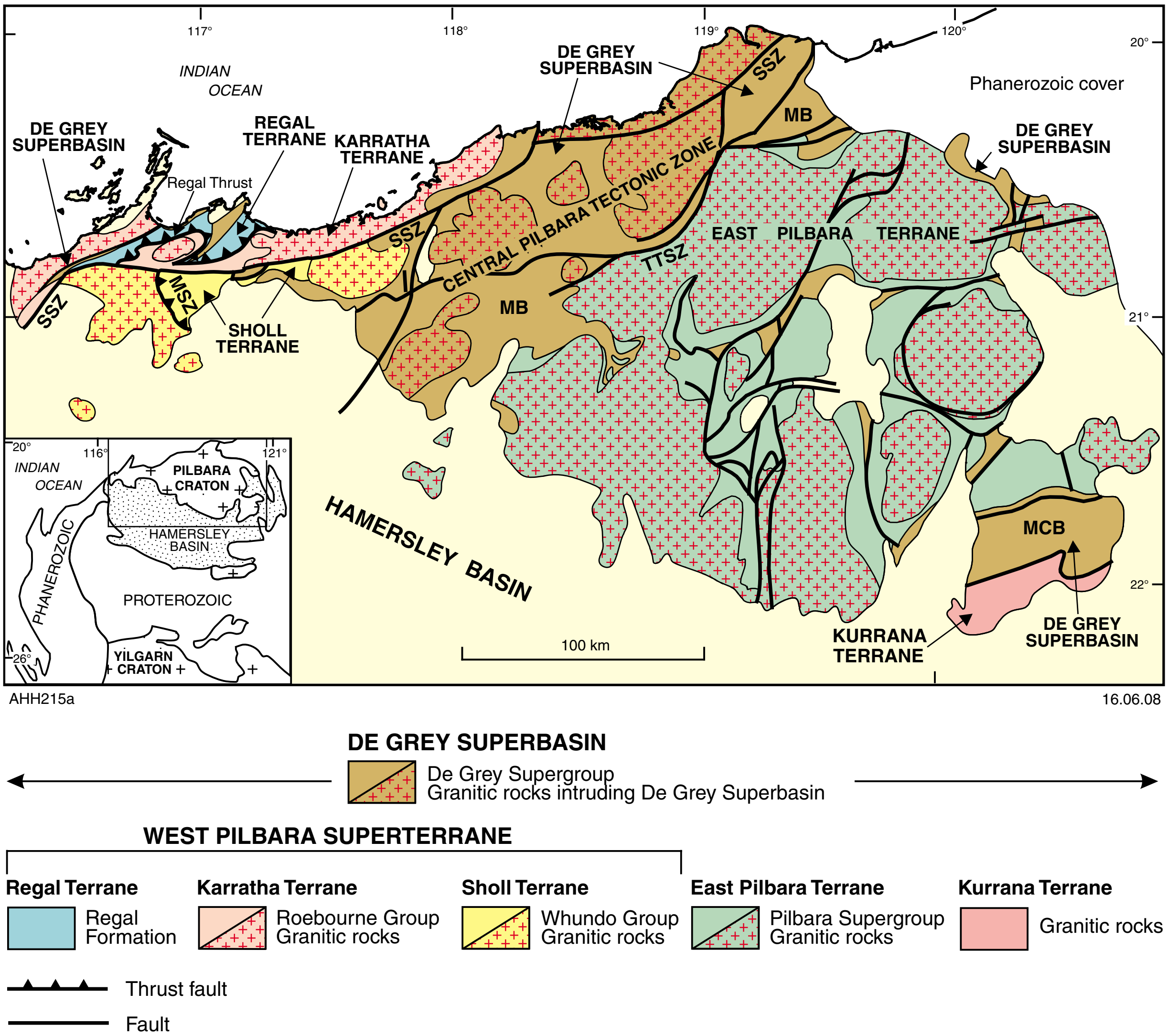

Figure 2. Simplified geological map of the northwestern Pilbara Craton showing terranes and the De Grey Superbasin. MB = Mallina Basin; MCB = Mosquito Creek Basin; SSZ = Sholl Shear Zone; MSZ = Maitland Shear Zone; TTSZ = Tabba Tabba Shear Zone

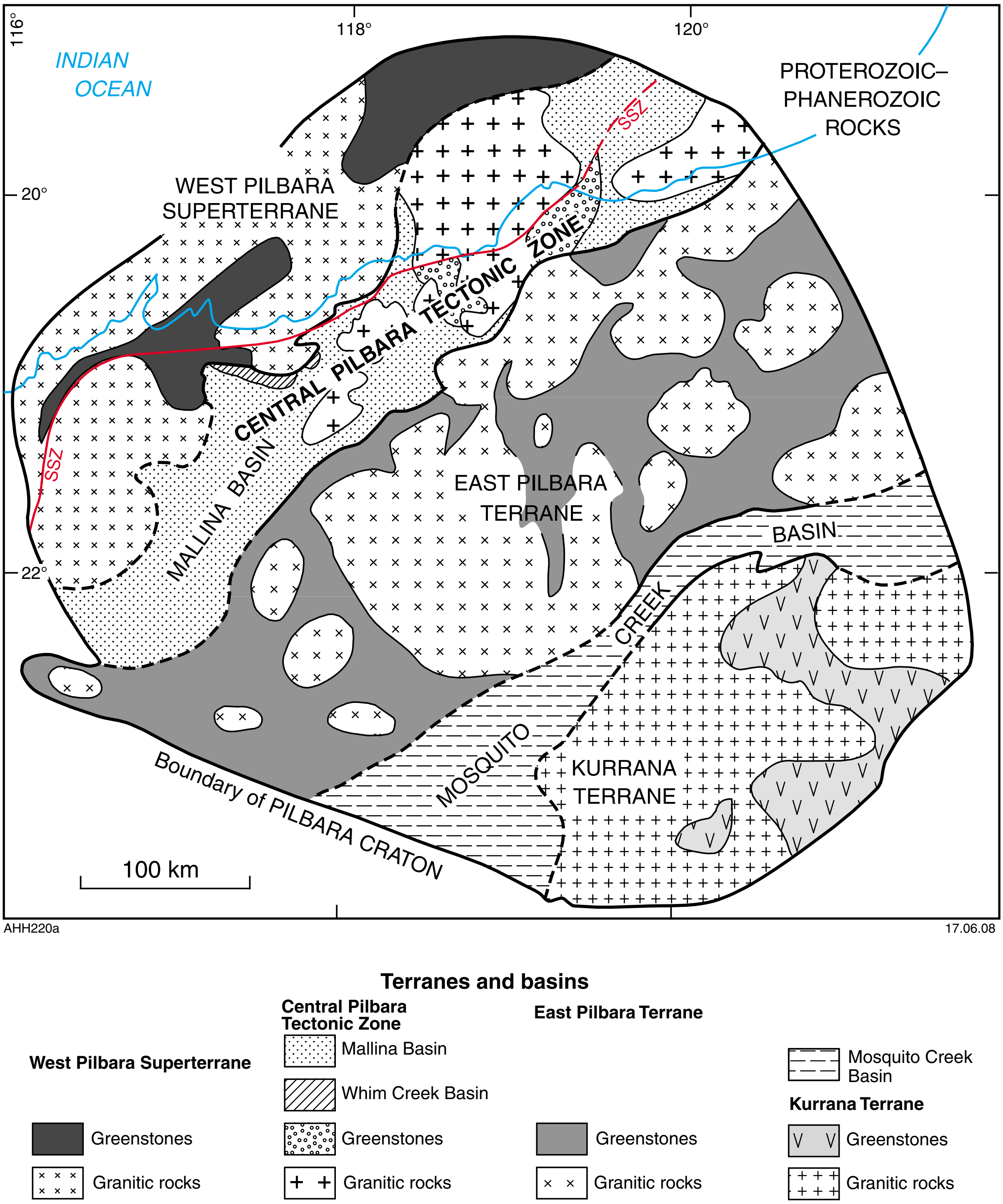

Figure 3. Interpretation of the granite–greenstone geology of the Pilbara Craton beneath the Fortescue and Hamersley Basins (modified from Hickman, 2004a; based on regional gravity and magnetic data from Blewett et al., 2000)

Part 1

Geology and mineralization of the northern Pilbara Craton

Geology

The entire Pilbara Craton is a 250 000 km^2 ovoid segment of 3800–2830 Ma Archean crust underlying northwestern Western Australia. The southern 70% of the craton is concealed by unconformably overlying 2770–2450 Ma rocks of the Fortescue and Hamersley Basins, but in the north the craton is exposed within a 60 000 km^2 inlier. Geological knowledge of the Pilbara Craton comes almost entirely from this northern inlier, where it comprises six main components, in order of decreasing age:

1. Early crust, 3800–3530 Ma (remnants are only very rarely exposed, and its existence is inferred mainly from geochronological data);
2. East Pilbara Terrane, 3525–3165 Ma (a granite–greenstone terrane consisting of the predominantly volcanic supracrustal succession of the Pilbara Supergroup and five contemporaneous granitic supersuites);
3. West Pilbara Superterrane, 3270–3060 Ma (three granite–greenstone terranes);
4. Kurrana Terrane, 3200–2895 Ma (two granitic supersuites, with minor greenstone units of uncertain age);
5. De Grey Superbasin, 3020–2930 Ma (five predominantly sedimentary basins, and three contemporaneous, mainly granitic, supersuites);
6. Split Rock Supersuite, 2890–2830 Ma (late- to post-tectonic granitic intrusions).

The Pilbara Craton was recently redefined (Van Kranendonk et al., 2006a) to exclude the overlying Fortescue and Hamersley Basins. Trendall (1990) included the Hamersley Basin (which in 1990 included the Fortescue Basin) on the grounds that both it and the underlying granite–greenstones did not attain stability until 2.4 Ga. However, recent mapping and geochronology have revealed that stability of the terranes and basins of the Pilbara Craton was essentially attained at c. 2895 Ma. This followed the final stage of terrane accretion, with a major event of orogenic deformation and metamorphism; in essence this was the ‘cratonization event’. Rifting of the Pilbara Craton at c. 2775 Ma, and the first outpouring of continental flood basalts of the Fortescue Group (Thorne and Trendall, 2001), marked the beginning of a distinctly separate stage in the evolution of the Pilbara crust. Some of the major structures (especially synforms) in the granite–greenstones of the East Pilbara Terrane were reactivated during deposition of the lower part of the 2775–2630 Ma Fortescue Group, but this deformation was comparatively minor, and occurred at least 140 m.y. after the 2.9 Ga cratonization event.

The interpreted bedrock geology of the Pilbara Craton is shown in Figure 3, the regional lithostratigraphy of the craton is summarized in Figure 4, and its tectonic evolution is diagrammatically illustrated in Figure 5.

Previous investigations

In 1956, the Geological Survey of Western Australia (GSWA) embarked on a program of regional geological mapping in the Pilbara. By 1964, fieldwork had been completed across the northern part of the craton. Resulting maps (1:250 000 scale) and reports greatly improved geological understanding of the geology and mineral potential of the Hamersley Basin, although the geology of the underlying granite–greenstone terranes of the Pilbara Craton remained poorly understood. Between 1972 and 1976, GSWA undertook a second program of 1:250 000-scale mapping in the east Pilbara region, which resulted in the first comprehensive account of the regional geology and mineralization of the northern Pilbara Craton (Hickman, 1983).

Between 1976 and 1993, the Pilbara Craton was geologically investigated by a large number of Australian and overseas researchers. Using new technology (e.g. remote sensing and precise geochronology) and an expanding range of models for crustal evolution, these investigators reached a wide range of new geological interpretations. This is probably partly explained by the fact that most of these workers attempted to draw regional conclusions from data collected over very small areas that were unrepresentative of the craton as a whole. Nevertheless, the previously widely accepted regional interpretation based on GSWA mapping in the 1970s (Hickman, 1983) was seriously challenged on several fronts. For example, new geochronological data indicated that some of the regional stratigraphic correlations between the east and west Pilbara (Hickman, 1983) were incorrect, leading to suggestions of totally different terranes and a crustal evolution involving terrane accretion (Barley, 1987). It became obvious that the northern Pilbara Craton required remapping, in conjunction with additional geochronology and geochemistry, to provide a geological interpretation constrained by data at craton-wide scale. In 1995, GSWA and the Australian Geological Survey Organization (now Geoscience Australia) started the Northern Pilbara Project, which included systematic mapping at 1:25 000 scale (for publication of 1:100 000 maps), high-quality regional airborne magnetic and radiometric surveys, regional geochronology and geochemistry, and local mineralization and structural

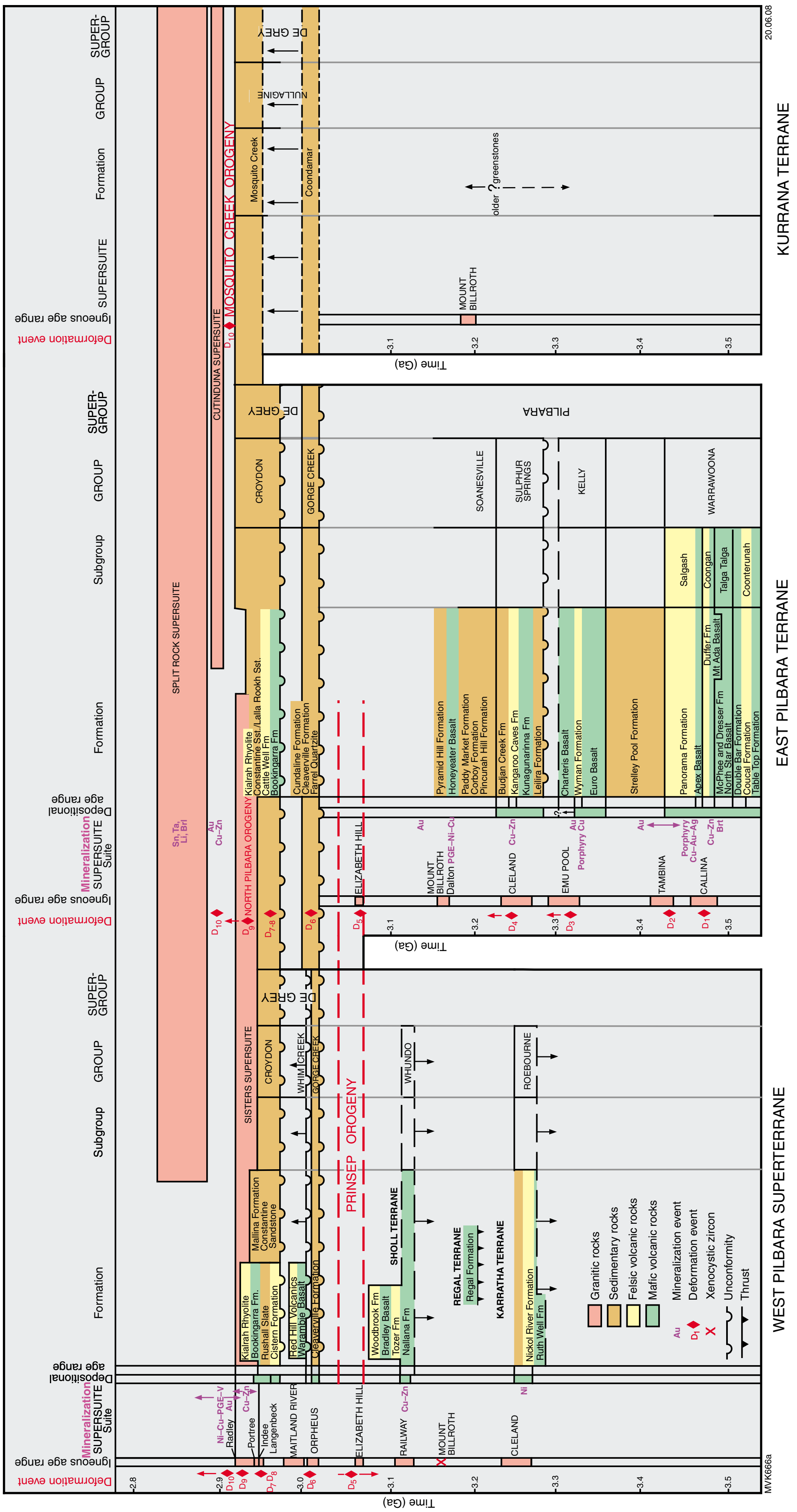

Figure 4. Time–space diagram showing the distinct evolution of terranes of the northern Pilbara Craton (from Van Kranendonk et al., 2006a)

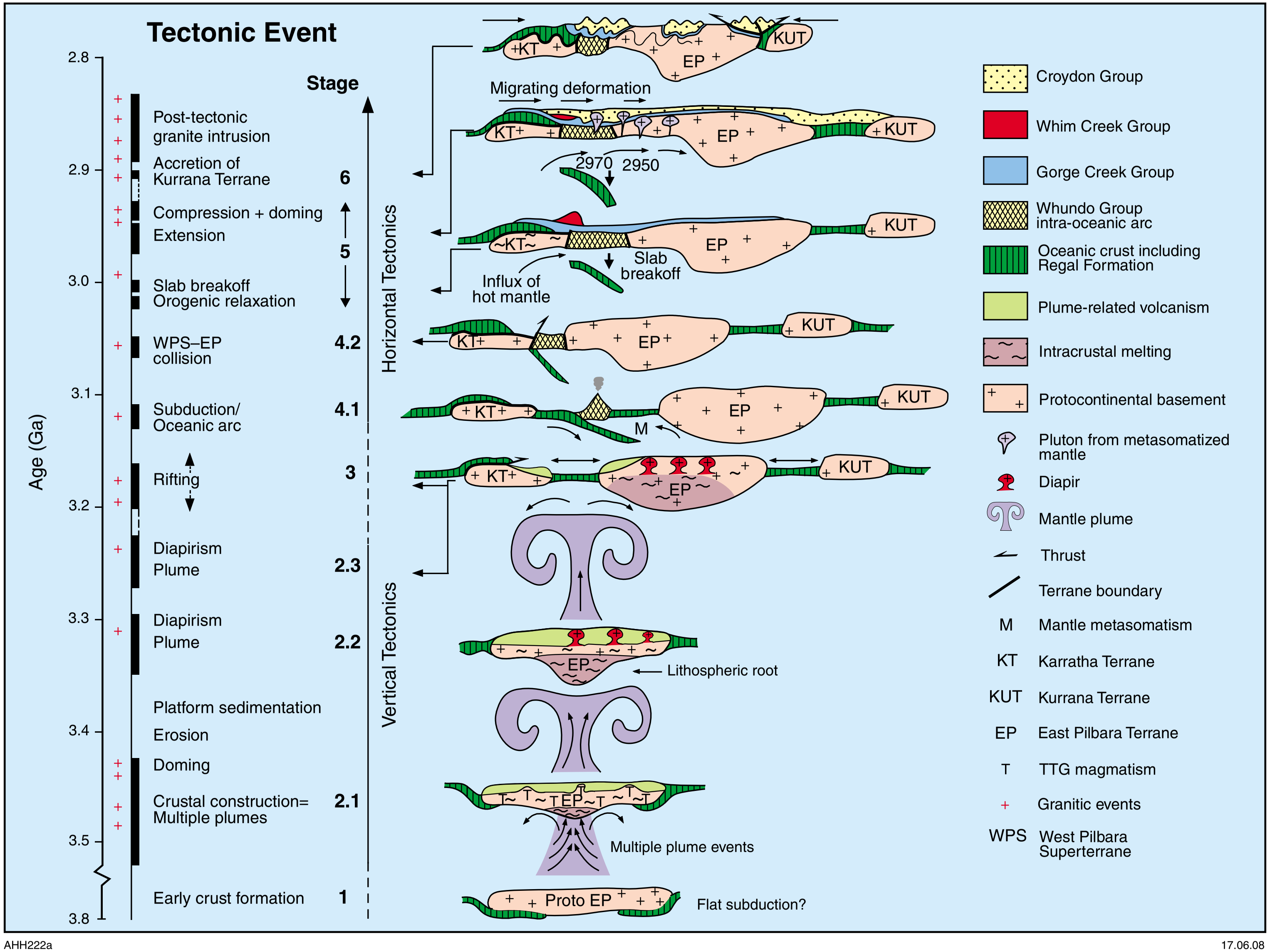

Figure 5. Diagrammatic summary of events in the formation of the Pilbara Craton (modified from Van Kranendonk et al., 2006a). The change from vertical tectonics in the East Pilbara Terrane to horizontal tectonics in the West Pilbara Superterrane occurred at c. 3.2 Ga following rifting at East Pilbara Terrane margins. Post-orogenic relaxation at 3.02 Ga resulted in crustal extension and deposition of the De Grey Superbasin until 2.94 Ga (Stages 1–6 are described by Van Kranendonk et al., 2006a,b)

studies. Hickman (1999) reported that an important early conclusion of the project was that the northern Pilbara Craton is composed of western and eastern terranes characterized by different tectonic styles and different lithostratigraphies. Further important conclusions were reported in GSWA publications, at various conferences, and in externally published papers to 2004. The completion of field mapping in 2003 established a large new geoscientific database that is now being released progressively to the public (Geological Survey of Western Australia, 2008).

The new geological mapping has led to a vastly improved understanding of the region's Archean geology and crustal evolution. Major findings have been that the west Pilbara region is composed largely of Archean crust that is 250 m.y. younger than the oldest crust in the east Pilbara, and that between 3.17 and 3.07 Ga the two areas evolved separately and in different tectonic environments (Van Kranendonk et al., 2002, 2006a, 2007; Hickman, 2004a).

East Pilbara Terrane

The East Pilbara Terrane provides the world's most complete record of Paleo- to Mesoarchean crustal evolution. The terrane represents the nucleus of the craton, formed through a succession of mantle plumes (3525–3235 Ma) that produced a dominantly basaltic volcanic succession, known as the Pilbara Supergroup, on an older sialic basement. Volcanism was accompanied by subvolcanic granitic magmatism (intrusion of the 3.47 Ga Callina, 3.43 Ga Tambina, 3.31 Ga Emu Pool, and 3.25 Ga Cleland Supersuites), which over time became increasingly restricted to a number of progressively inflating granitic domes (Hickman and Van Kranendonk, 2004). The East Pilbara Terrane provides an outstanding example of gravity-driven diapiric deformation of the upper and middle crust (Hickman, 1984; Collins et al., 1998; Sandiford et al., 2004; Van Kranendonk et al., 2004). The multiple melting events related to the plumes led to formation of a thick, depleted lithospheric keel. At c. 3.20 Ga, this was followed by rifting of the East Pilbara Terrane margins and related intrusion of granitic rocks (Mount Billroth Supersuite).

The East Pilbara Terrane is host to a variety of mineral deposit types and styles that have been exploited since the late nineteenth century. Gold and tin–tantalite–lithium have been the most important commodities, and have been mined since the start of the twentieth century. At present, the Wodgina tantalite mine is the only operational mine in East Pilbara Terrane rocks, although a world-class copper–molybdenum mining operation is being developed at Spinifex Ridge, ~50 km northeast of Marble Bar, and a copper–zinc mine development is well advanced at Sulphur Springs, ~50 km west of Marble Bar.

Lithostructural elements

The East Pilbara Terrane is characterized by large, ovoid (35–120 km diameter) granitic complexes (formerly 'batholiths' of Hickman, 1983) that are flanked by curviplanar belts of commonly steeply dipping volcano-sedimentary rocks collectively referred to as greenstones (Fig. 6). Hickman (1983, 1984) showed that the granitic complexes commonly form domes, and the greenstones form synclinal belts. Van Kranendonk (1998) noted that some greenstone belts are monoclines (e.g. Western Shaw greenstone belt, 11 on Fig. 7), and confirmed the existence of greenstone domes (e.g. Panorama greenstone belt of the North Pole Dome, 7 on Fig. 7) previously recognized by Hickman (1975, 1983). Van Kranendonk (1998) proposed that structural domes include both granitic rocks and attached greenstones and showed that adjacent domes are separated by faults within greenstones.

Greenstone belts are defined as relatively well-preserved tracts of commonly coherent greenstone stratigraphy bounded by faults, or by intrusive or sheared-intrusive contacts (or both) with granitic complexes, or unconformities with older supracrustal rocks and, locally, with granitic rocks (Buick et al., 1995; Dawes et al., 1995). Greenstone complexes are areas of high structural complexity in which a coherent stratigraphy is typically lacking and faulting or shearing (or both) is common (Van Kranendonk, 1998).

The domical granitic complexes are multicomponent bodies that evolved contemporaneously with deposition of felsic volcanic components of the greenstones (Hickman, 1984, 1990; Williams and Collins, 1990; Barley and Pickard, 1999; Van Kranendonk et al., 2002, 2006a). Similar-aged suites of plutonic rocks are present in many individual complexes. This, combined with structural and gravity data, suggests that the complexes are linked at depth beneath intervening synclinal greenstone keels and thus represent structural domes of a continuous, mid-crustal granitic layer (Hickman, 1984; Wellman, 2000; Van Kranendonk et al., 2004). Early, more sodic components (c. 3500–3420 Ma) were emplaced as a sheeted sill complex into the Warrawoona Group and older crustal remnants.

Granitic complexes commonly have outward dipping margins, although some outer contacts with greenstones are locally overturned (e.g. the southern margins of the Mount Edgar and Muccan Granitic Complexes). The contacts of granitic complexes vary from being locally intrusive (e.g. all contacts of the Corunna Downs, the northern part of the Shaw, the southern part of the Carlindi, and the northern part of Mount Edgar Granitic Complexes), to the locus of shearing (e.g. southwestern Mount Edgar, and western and eastern Shaw Granitic Complexes), to an unconformity with younger supracrustal rocks (e.g. Shay Gap greenstone belt; Dawes et al., 1995). Although granitic complexes have simple outlines, they are characterized by complex and even chaotic internal geometries (Van Kranendonk et al., 2004). Older components are preserved principally around the outer margins of granitic complexes, pushed up and out by successive generations of increasingly potassic components that were emplaced into the cores of progressively amplifying domes at intervals punctuated throughout the history of the terrane, as exemplified by the Yule Granitic Complex (Champion and Smithies, 2000; Hickman and Van Kranendonk, 2004). In the Yule and Mount Edgar Granitic Complexes, roof pendants of amphibolite and orthogneiss form trains of enclaves separating domical lobes of younger granitic rocks. Along the margins of other granitic complexes, particularly

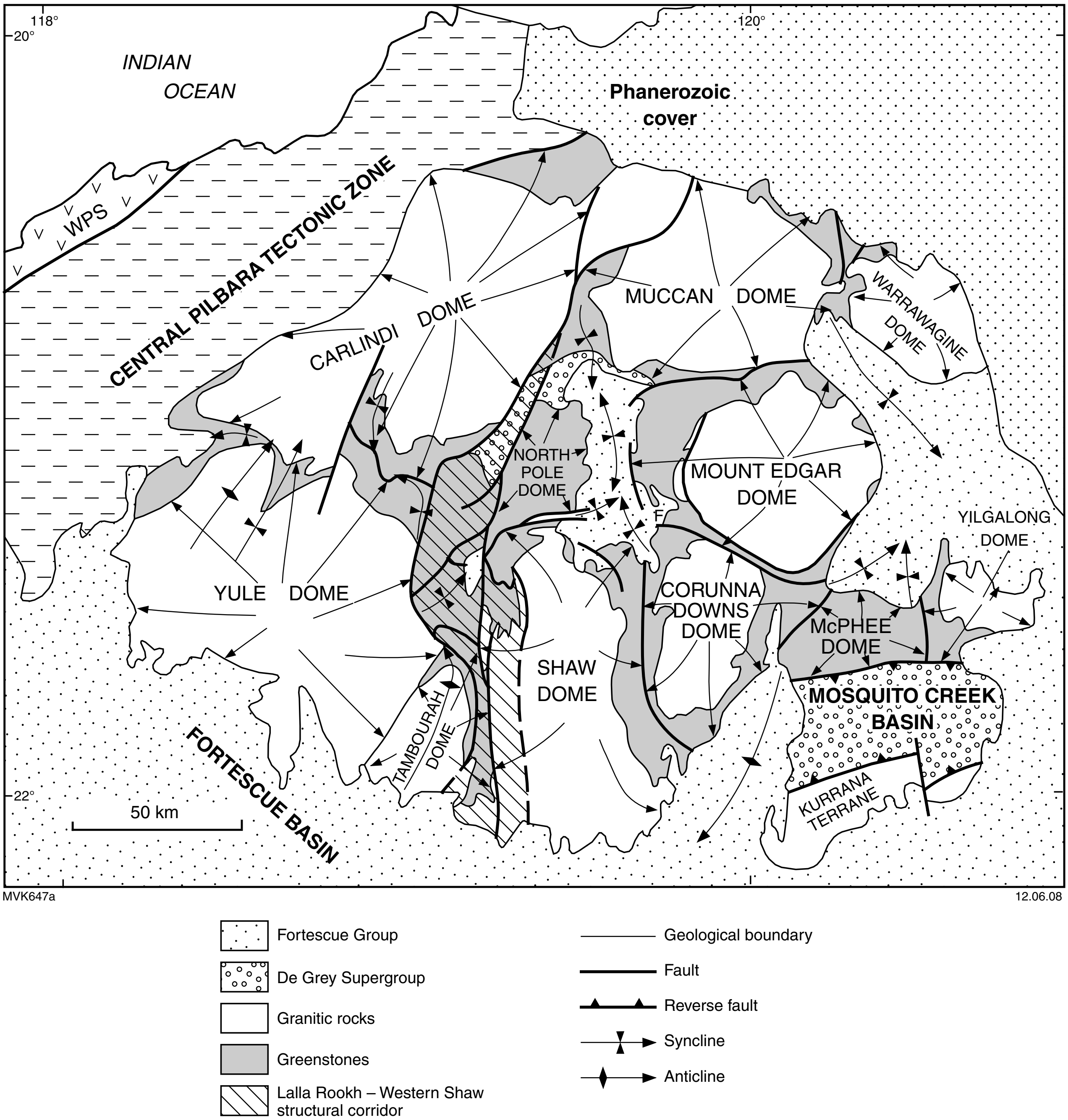

Figure 6. Principal structural elements of the East Pilbara Terrane showing structural domes with granitic cores and flanking greenstones, axial faults in greenstone synclines, and the Lalla Rookh–Western Shaw structural corridor of concentrated c. 2940 Ma deformation. The Central Pilbara Tectonic Zone is another zone of concentrated c. 2940 Ma deformation in the Mallina Basin (modified from Van Kranendonk et al., 2002)

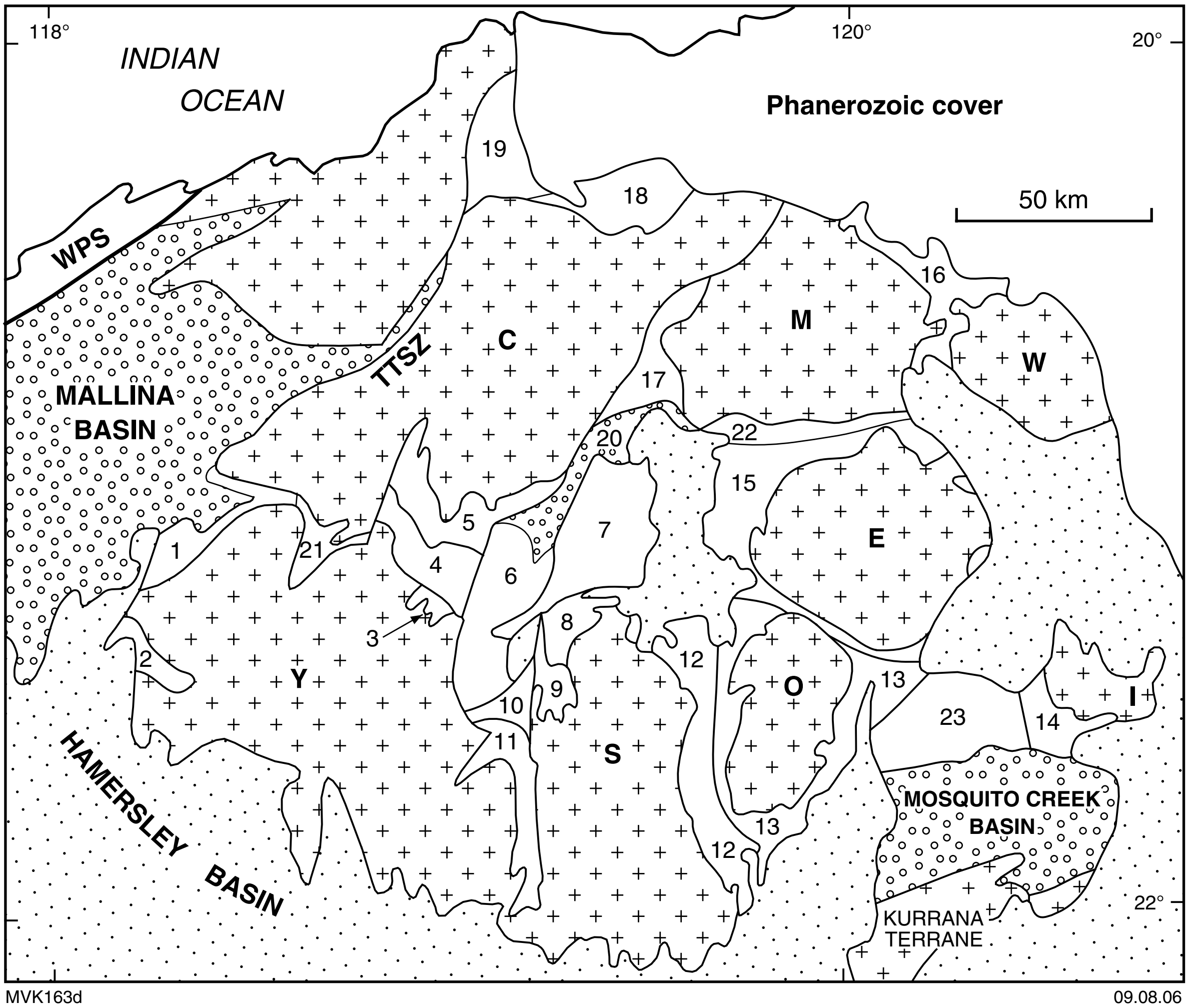

Figure 7. Lithostructural division of the East Pilbara Terrane into greenstone belts, greenstone complexes, and granitic complexes. 1 = Pilbara Well greenstone belt; 2 = Cheearra greenstone belt; 3 = Abydos greenstone belt; 4 = Pincunah greenstone belt; 5 = East Strelley greenstone belt; 6 = Soanesville 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 = Mount Elsie 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; 23 = McPhee greenstone belt. Granitic complexes: C = Carlindi; E = Mount Edgar; I = Yilgalong; M = Muccan; O = Corunna Downs; S = Shaw; W = Warrawagine; Y = Yule. TTSZ = Tabba Tabba shear zone; WPS = West Pilbara Superterrane (after Van Kranendonk et al., 2006a)

the Corunna Downs and Shaw Granitic Complexes, greenstones have been deformed into complex folds analogous to those found around salt diapirs (Van Kranendonk et al., 2004).

Lithostratigraphy

The Pilbara Supergroup of the East Pilbara Terrane is divided into four demonstrably autochthonous volcano-sedimentary groups and one formation (Van Kranendonk et al., 2006; revised by Hickman, 2008). These units were deposited across erosional unconformities, disconformities, or paraconformities. The succession, in order of decreasing age, comprises:

- 3525–3427 Ma Warrawoona Group;
- 3426–3350 Ma Strelley Pool Formation;
- 3350–3315 Ma Kelly Group;
- 3270–3230 Ma Sulphur Springs Group;
- 3230–3165 Ma Soanesville Group.

Warrawoona Group

The Warrawoona Group is a thick succession of dominantly basaltic volcanic rocks with lesser felsic volcanic and minor sedimentary rocks and was deposited between 3525 and 3426 Ma above continental crust, but mainly in marine environments. Geochronological data (U–Pb ages on detrital and xenocrystic zircon grains and

Nd model ages on granitic rocks) indicate that the age of the underlying crust was between 3.8 and 3.55 Ga, and rare remnants of this ancient crust have been identified in some of the granitic complexes. The Warrawoona Group is composed of four subgroups that include, from base to top, the Coonterunah, Talga Talga, Coongan, and Salgash Subgroups (Fig. 8). However, as noted below, the Coonterunah Subgroup, which is confined to the greenstone belts around the Carlindi Dome, is nowhere in contact with the Talga Talga Subgroup, and the two subgroups are of similar age. The type area of the Warrawoona Group is in the Marble Bar greenstone belt, where all units except the Coonterunah Subgroup are well exposed and form a right-way-up, upward-younging succession of 12 km thickness. Deposition of the Warrawoona Group was as a series of (ultra-)mafic through felsic volcanic cycles, of c. 15 m.y. duration, at rates of accumulation consistent with a mantle plume origin (Van Kranendonk et al., 2002, 2004, 2006a).

The Coonterunah Subgroup (Van Kranendonk et al., 2006a) is a 5.9 km-thick succession of mainly tholeiitic basalt and subordinate intermediate to felsic volcanic rocks distributed around the southern and eastern flanks of, and intruded by, the c. 3470 Ma Carlindi Granitic Complex. This subgroup was first identified as a 'succession' (Buick et al., 1995) on the basis of a local high-angle unconformity with the Strelley Pool Formation and the oldest age (3515 ± 3 Ma) ever recorded from the Warrawoona Group. Van Kranendonk and Morant (1998) separated the Coonterunah succession from the Warrawoona Group, and formally named it the Coonterunah Group. However, more recent geochronology has indicated that the Coonterunah succession is at least partly contemporaneous with the Talga Talga Subgroup of the Warrawoona Group, as originally interpreted by Hickman (1980, 1981, 1983). Moreover, geochemical studies have shown no difference in composition between basaltic rocks of the Coonterunah Subgroup and other Warrawoona Group basalts (Green et al., 2000; Smithies et al., 2005b); thus, Van Kranendonk et al. (2006a) changed its status to a subgroup of the Warrawoona Group. Geochemical data indicate significant contamination of Coonterunah Subgroup basalts by older sialic crust (Green et al., 2000; Smithies et al., 2005b).

The type area of the Talga Talga Subgroup is in the Marble Bar greenstone belt, where it consists of the North Star Basalt and the McPhee Formation. The subgroup is also mapped in the Coongan, Panorama, and North Shaw greenstone belts. Two Ar–Ar dates of c. 3500 Ma from hornblende in amphibolite-facies metabasalts of the North Star Basalt in the Panorama and Coongan greenstone belts (Davids et al., 1997; Zegers et al., 1999) suggest that the North Star Basalt and Coonterunah Subgroups are of similar age, a conclusion supported by a c. 3490 Ma Ar–Ar date (van Koolwijk et al., 2001) from the North Star Basalt in its type area north of Marble Bar. The McPhee Formation, comprising laminated grey-and-white chert (possibly a silicified fine-grained volcaniclastic or carbonate rock), iron formation, shale, and intercalated mafic and ultramafic rocks, was dated at 3477 ± 2 Ma (Nelson, 2000a) in the Marble Bar greenstone belt. The basal contact of this subgroup is everywhere intrusive with synvolcanic or younger granitic rocks, and it is conformably overlain by the Coongan Subgroup.

The 3474–3463 Ma Coongan Subgroup consists of the Mount Ada Basalt and the felsic volcanic Duffer Formation, including the Marble Bar Chert Member at the top. These rocks are best exposed in the Marble Bar greenstone belt where they are conformably overlain by the c. 3460–3426 Ma Salgash Subgroup, which consists of the Apex Basalt and overlying felsic volcanic Panorama Formation. The Apex Basalt is present only in the Marble Bar, Warralong, and Western Shaw greenstone belts; elsewhere, the Panorama Formation lies directly on the Duffer Formation (cf. DiMarco and Lowe, 1989; Van Kranendonk et al., 2004). The Panorama Formation has been identified across the East Pilbara Terrane and consists of a dominantly rhyolitic succession of massive volcaniclastic rocks 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 silicified volcaniclastic sandstone a few metres thick in some belts. Stratigraphic variations and geochemical data suggest that it was erupted from several volcanic sources, including the vent of the Panorama volcano, which is exposed in cross section in the northern part of the Panorama greenstone belt (Van Kranendonk, 2000; Smithies et al., 2007).

Strelley Pool Formation

The Strelley Pool Formation was first named informally as the 'Strelley Pool Chert' by Lowe (1980, 1983); this name was eventually formalized by Van Kranendonk and Morant (1998). Detailed sedimentological studies by Allwood et al. (2007a) and Wacey et al. (in prep.), combined with mapping in most greenstone belts of the East Pilbara Terrane, have revealed that, regionally, the formation is not predominantly composed of chert, and in those areas where chert is an important constituent it is mainly a secondary alteration effect, the main protoliths having been carbonate rocks and fine-grained siliciclastic rocks. Accordingly, Hickman (2008) formally renamed the unit as the Strelley Pool Formation. The formation has been identified in eleven of the twenty greenstone belts in the East Pilbara Terrane of the Pilbara Craton, establishing that its depositional area was at least 30 000 km^2. Regional geochronological data indicate that the interval of time during which the Strelley Pool Formation was deposited was from 3426 to 3350 Ma, and it is probable that deposition occurred throughout this interval. The sedimentary facies of the formation indicate that it is predominantly a shallow-water marine sequence that includes beach, estuarine, sabkha, stromatolite reef, fluviatile, and lacustrine environments. Hickman (2008) reviewed the stratigraphic assignment of the formation and concluded that the lower and central parts of the succession, previously assigned to the 'Strelley Pool Chert', should be separated from both the Kelly and Warrawoona Groups.

The Strelley Pool Formation has acquired international recognition for the important evidence it provides on early Archean depositional environments and early life. It contains the world's best-preserved and most diverse and abundant early Archean fossil stromatolites, as well

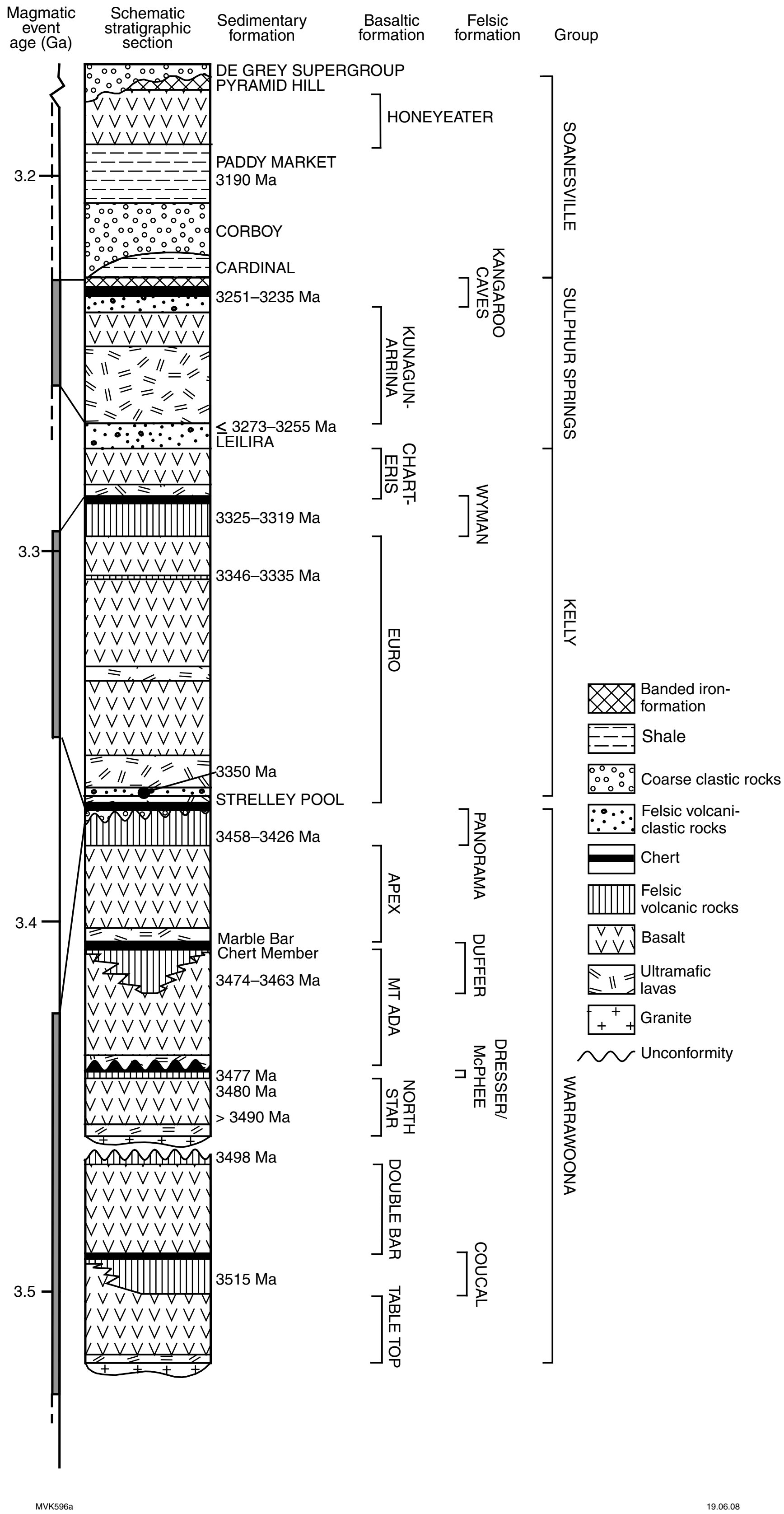

Figure 8. Generalized lithostratigraphic column of the Pilbara Supergroup in the East Pilbara Terrane showing available age constraints. The unconformably overlying, topmost coarse clastic unit represents the De Grey Supergroup in the Lalla Rookh Basin. The timeline on left shows the time span of depositional events (from Van Kranendonk et al., 2006a)

as well-preserved evidence of carbonaceous microbial fossils and possible biogenic microtubular structures. The formation is also important because its shallow-water deposition, commonly above an erosional unconformity, confirms that the Pilbara Supergroup, one of the world's oldest and best preserved greenstone successions, was deposited on continental crust.

Kelly Group

The Kelly Group consists of three formations: the 3350–3325 Ma Euro Basalt, the 3325–3315 Ma Wyman Formation, and the c. 3315 Ma (not yet dated) Charteris Basalt. Unlike all formations in the Warrawoona Group, the Euro Basalt is present in almost all greenstone belts of the East Pilbara Terrane, with the possible exception of poorly dated greenstone belts on the western and northern flanks of the Yule Dome (Fig. 6). The stratigraphic thickness of the formation varies between 9.4 km and 1 km, and it comprises a succession of pillowed tholeiite interlayered with thin cherts (upper part of formation), komatiitic basalt, and peridotitic komatiite (lower part of formation). Despite its thickness, composition, and widespread distribution, the Euro Basalt cannot be an ophiolite or deep-water oceanic succession, as interpreted by some workers (Isozaki et al., 1997; Kitajima et al., 2001; Furnes et al., 2007). The basal part of the Euro Basalt was erupted at or slightly below sea level (evidence from the immediately underlying Strelley Pool Formation). The underlying stratigraphy of the Warrawoona Group and the presence of several intrusive >3.4 Ga granitic supersuites prove that it was deposited above thick (>20 km) continental crust. The Euro Basalt's thick succession of predominantly subaqueous lavas proves major subsidence of its depositional basin. However, regional variations in thickness indicate that some parts of the basin sank by up to 10 km during the 25 m.y. between 3350 Ma and 3325 Ma, whereas other parts either remained stable or subsided then rose. The thickest sections of the Euro Basalt are in areas where the granitic cores of the domes are most widely separated (Fig. 6). This indicates that subsidence was a complementary process to doming, as suggested by the diapiric models of crustal evolution applied to the East Pilbara Terrane (Hickman, 1975, 1983, 1984; Collins, 1989; Van Kranendonk et al., 2002, 2004; Hickman and Van Kranendonk, 2004).

The 3325–3315 Ma Wyman Formation is a dacitic to rhyolitic succession of lavas and volcaniclastic rocks with local intercalations of siliciclastic sedimentary rocks, chert, and basalt (komatiitic and tholeiitic). In some areas, the top of the formation includes stromatolitic carbonate rocks, black shale, and beds of barite up to 1 m thick (Williams and Bagas, 2007). The formation varies in thickness up to 2 km, but is absent from most greenstone belts in the western half of the terrane. This distribution is very similar to 3325–3290 Ma monzogranite intrusions of the Emu Pool Supersuite (Van Kranendonk et al., 2006a; Fig. 9), and the formation is interpreted to be the volcanic part of this supersuite. A striking feature of the Wyman Formation in several greenstone belts is the presence of thick flows of columnar rhyolite, one of which is visited on this excursion (Locality 13).

The Charteris Basalt is undated, but is interpreted to be of similar age to the Wyman Formation because the two are conformable; rare basalt intercalations within the Wyman Formation may belong to the Charteris Basalt. The formation is composed predominantly of komatiitic basalt and tholeiitic basalt, with sills of dolerite. Many of the lavas are pillowed. The formation is up to 2 km thick, but it is apparently restricted to the Kelly greenstone belt.

Sulphur Springs Group

The Sulphur Springs Group (Fig. 8; Van Kranendonk and Morant, 1998; Van Kranendonk, 2000; Van Kranendonk et al., 2006a) is a succession of dominantly volcanic rocks in the western part of the East Pilbara Terrane and may also underlie the Wodgina and Pilbara Well greenstone belts (Fig. 7). The group is divided into four formations, three of which are well developed in the type area of the Pincunah and Soanesville greenstone belts (Leilira, Kunagunarrina, and Kangaroo Caves Formations), and a fourth (Budjan Creek Formation) that is present only in the southeastern part of the terrane.

The Leilira Formation at the base of the Sulphur Springs Group is composed of conglomerate, lithic wacke, felsic volcaniclastic sandstone, tuff, and dacite. It unconformably overlies the Euro Basalt in the southwestern extension of the Panorama greenstone belt and in the East Strelley greenstone belt (Fig. 7; Van Kranendonk, 2000). A maximum age of deposition for this formation is c. 3255 Ma, based on SHRIMP dating of detrital zircons (Buick et al., 2002). However, Pb–Pb model ages of c. 3280 Ma (using the model of Cumming and Richards, 1975) of sulfide mineralization from small felsic intrusions near the base of the group suggest a possibly older maximum age of deposition (Van Kranendonk et al., 2006b), as do the presence of c. 3270 Ma detrital zircons in nearby quartzite (Geological Survey of Western Australia, 2006, sample 178045).

The conformably overlying Kunagunarrina Formation includes komatiite, komatiitic basalt, chert, and local felsic volcaniclastic rocks. The type section of this formation is 2.4 km thick in the Pincunah greenstone belt. These rocks have distinctive strongly depleted light-rare-earth element (LREE) profiles, with εNd values of between +3.2 and –2.3 (on the depleted-mantle evolution curve), becoming more negative up the stratigraphic section (Van Kranendonk et al., 2007), which indicates progressive crustal contamination. A felsic volcaniclastic unit within this formation from the East Strelley greenstone belt formed at 3251 ± 4 Ma (Geological Survey of Western Australia, 2006, sample 160957).

Conformably overlying the Kunagunarrina Formation is the Kangaroo Caves Formation, comprising up to 1.5 km of basalt–andesite to rhyolitic volcanic rocks, banded iron-formation, and chert (Brauhart, 1999; Van Kranendonk et al., 2006a). Felsic volcanism was coeval with emplacement of the Strelley Monzogranite, a synvolcanic laccolith; both of these have been dated at c. 3238 Ma (Buick et al., 2002). The intermediate to felsic volcanic rocks of the Kangaroo Caves Formation have more-fractionated LREE profiles, and εNd values are consistently between –1 and –0.3.

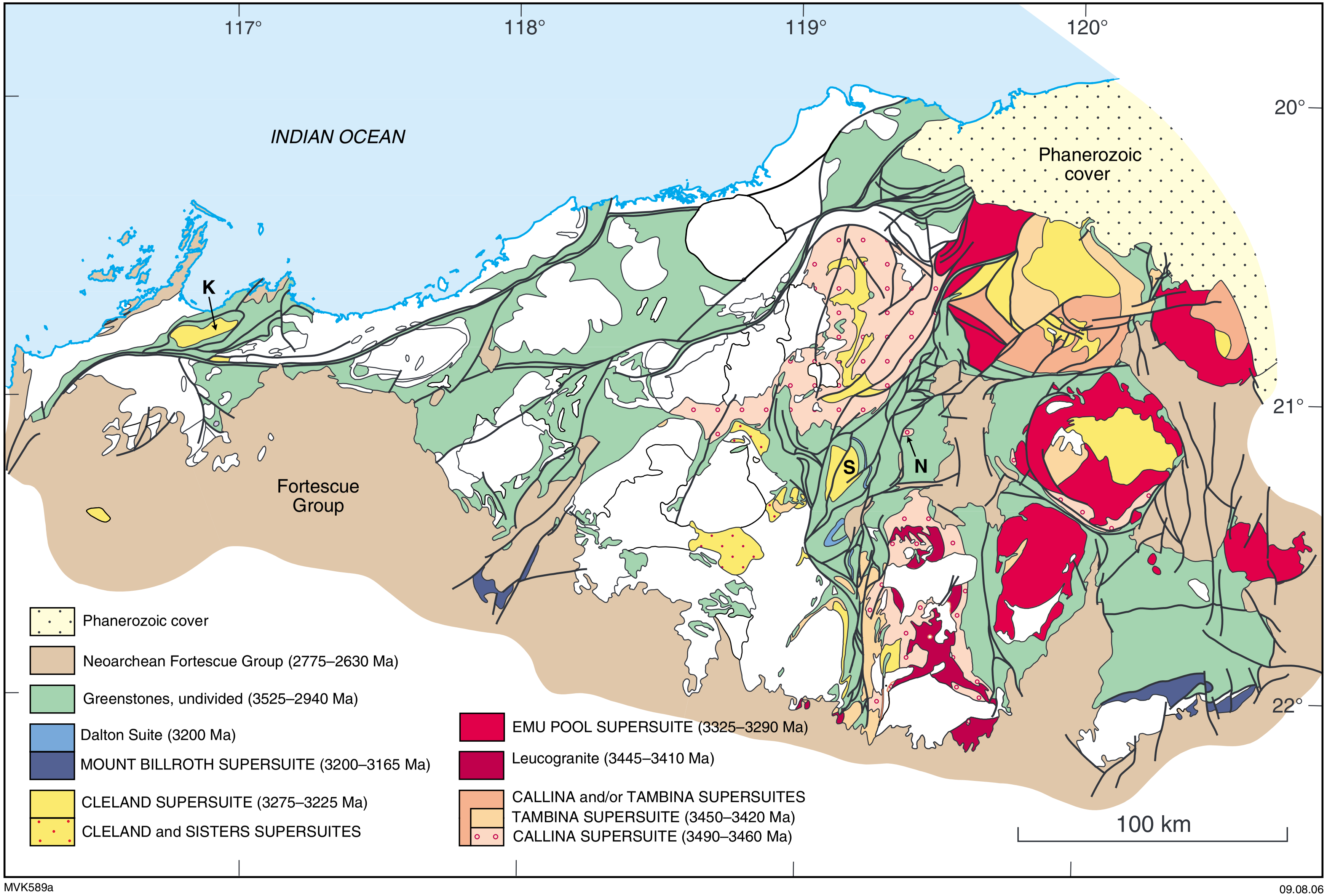

Figure 9. Map of the northern Pilbara Craton showing that the Callina, Tambina, and Emu Pool Supersuites are restricted to the East Pilbara Terrane, whereas the Cleland Supersuite is more widespread. K = Karratha Granodiorite; N = North Pole Monzogranite; S = Strelley Monzogranite (from Van Kranendonk et al., 2006a)

Intrusion of the synvolcanic Strelley Monzogranite laccolith caused hydrothermal circulation that silicified a unit of fine-grained epiclastic sedimentary rocks at the top of the Sulphur Springs Group and developed several zinc–copper volcanic-hosted massive sulfide (VHMS) deposits spaced at regular intervals around the monzogranite (Morant, 1998; Vearncombe et al., 1995, 1998; Brauhart et al., 1998; Brauhart, 1999; Van Kranendonk, 2000; Pirajno and Van Kranendonk, 2005).

The Budjan Creek Formation is a succession of coarse siliciclastic rocks, felsic volcaniclastic sandstone, shale, local chert, and minor amounts of dacitic tuff in the Kelly greenstone belt. This formation lies unconformably on the Wyman Formation (Kelly Group) and is unconformably overlain by rocks of the Gorge Creek Group, placing it within the stratigraphic framework of the Sulphur Springs Group. The depositional age of the formation is very loosely constrained by U–Pb zircon data (Nelson, 2001, sample 168908) interpreted to indicate deposition at c. 3228 Ma. However, this interpretation relies on analysis of a single zircon grain; the main zircon population of the rock showed an age of c. 3308 Ma.

Soanesville Group

The 3200–3180 Ma Soanesville Group (Fig. 8; Van Kranendonk et al., 2006a) outcrops in the western part of the East Pilbara Terrane, and typically overlies the Sulphur Springs Group. The group consists of clastic sedimentary rocks in the lower part, and up to 1.5 km of basalt and minor amounts of banded iron-formation in the upper part. The lower, clastic sedimentary formations of the group disconformably overlie relict topography of the Sulphur Springs Group. Rasmussen et al. (2007) dated monazite within these clastic rocks at c. 3.19 Ga and interpreted that to be the approximate age of deposition.

Deposition of the Honeyeater Basalt in the upper part of the Soanesville Group was accompanied by the intrusion of thick, locally differentiated ultramafic and mafic sills of the Dalton Suite (Van Kranendonk and Morant, 1998; Van Kranendonk, 2000); the sills host platinum group element (PGE) minerals and nickel sulfides. This mafic magmatism is interpreted to have occurred at c. 3180 Ma on the basis of U–Pb zircon dating of one of the intrusions at 3182 ± 2 Ma (Wingate, in prep. a, sample 178185). Intrusion of the dykes and sills of the c. 3.18 Ga Dalton Suite is interpreted to have been a stage in rifting of the East Pilbara Terrane (Hickman, 2004a; Van Kranendonk et al., 2006a).

Granitic supersuites

Granitic rocks of the East Pilbara Terrane are distributed in eight ovoid, domical granitic complexes, two synvolcanic laccolith intrusions (the c. 3458 Ma North Pole Monzogranite and 3240 Ma Strelley Monzogranite), and the syntectonic Keep It Dark Monzogranite in the Emerald Mine greenstone complex (Fig. 7). These rocks were previously mapped and classified on the basis of their distribution within granitic complexes. Subsequent geochronological and geochemical data, combined with new mapping, has allowed division of the granitic rocks of the Pilbara Craton into several suites and supersuites (Figs 9–11; Van Kranendonk et al., 2006a). There are three supersuites of granitic rocks that are unique to the East Pilbara Terrane and reflect the older part of its history, and five other supersuites that are also present in one or several other terranes of the craton.

All mapped granitic complexes, except the Corunna Downs and Yilgalong Granitic Complexes, contain surface exposures of gneissic to foliated remnants of the oldest rocks of the Callina Supersuite of metamorphosed tonalite–trondhjemite–granodiorite (TTG) and derived gneisses, dated between c. 3500 and 3460 Ma and coeval with the lower parts of the Warrawoona Group. Many complexes also contain remnants of the slightly younger Tambina Supersuite (3450–3420 Ma) of tonalite–monzogranite that was coeval with eruption of the Panorama Formation. The predominantly monzogranitic rocks of the 3325–3290 Ma Emu Pool Supersuite are restricted to the southeastern part of the East Pilbara Terrane and constitute nearly the whole of the Corunna Downs and Yilgalong Granitic Complexes. The 3275–3225 Ma Cleland Supersuite is also predominantly of monzogranitic composition and is present in granitic complexes across the northern part of the East Pilbara Terrane, as well as in the Karratha Terrane of the West Pilbara Superterrane (Figs 2, 4).

Two younger granitic supersuites outcrop over areas in the western part of the Yule Granitic Complex along the western margin of the East Pilbara Terrane. These include the c. 3165 Ma Flat Rocks Tonalite of the Mount Billroth Supersuite, thought to be associated with rifting of the East Pilbara Terrane margins at this time, and the c. 3068 Ma Cockeraga Leucogranite of the Elizabeth Hill Supersuite, thought to be associated with collision between the East Pilbara Terrane and West Pilbara Superterrane during the Prinsep Orogeny (Fig. 4; Van Kranendonk et al., 2006a).

Dominantly monzogranitic rocks of the Sisters Supersuite were emplaced in a significant portion of the western half of the East Pilbara Terrane between 2945 and 2920 Ma, contemporaneous with regional transpressional deformation (Fig. 10; Van Kranendonk and Collins, 1998; Zegers et al., 2001).

Synkinematic monzogranitic rocks of the Bamboo Springs Monzogranite of the Cutinduna Supersuite were emplaced into the southeastern part of the East Pilbara Terrane at c. 2908 Ma (Fig. 11), contemporaneous with deformation of the Mosquito Creek Basin. This pluton is characterized by a pervasive synmagmatic foliation defined by aligned K-feldspar phenocrysts that are subparallel to the main tectonic fabric in the Mosquito Creek Basin.

The youngest granitic rocks intruding the East Pilbara Terrane are those of the post-tectonic Split Rock Supersuite, which occupy the cores of many of the domical granitic complexes in a linear northwest-trending belt across the East Pilbara Terrane (Fig. 11).

The early TTG suites are silicic and sodic and are derived from high-pressure melting of a mafic source. Bickle et al. (1983, 1993) compared the TTG suite favourably with modern calc-alkaline, subduction-related arc suites,

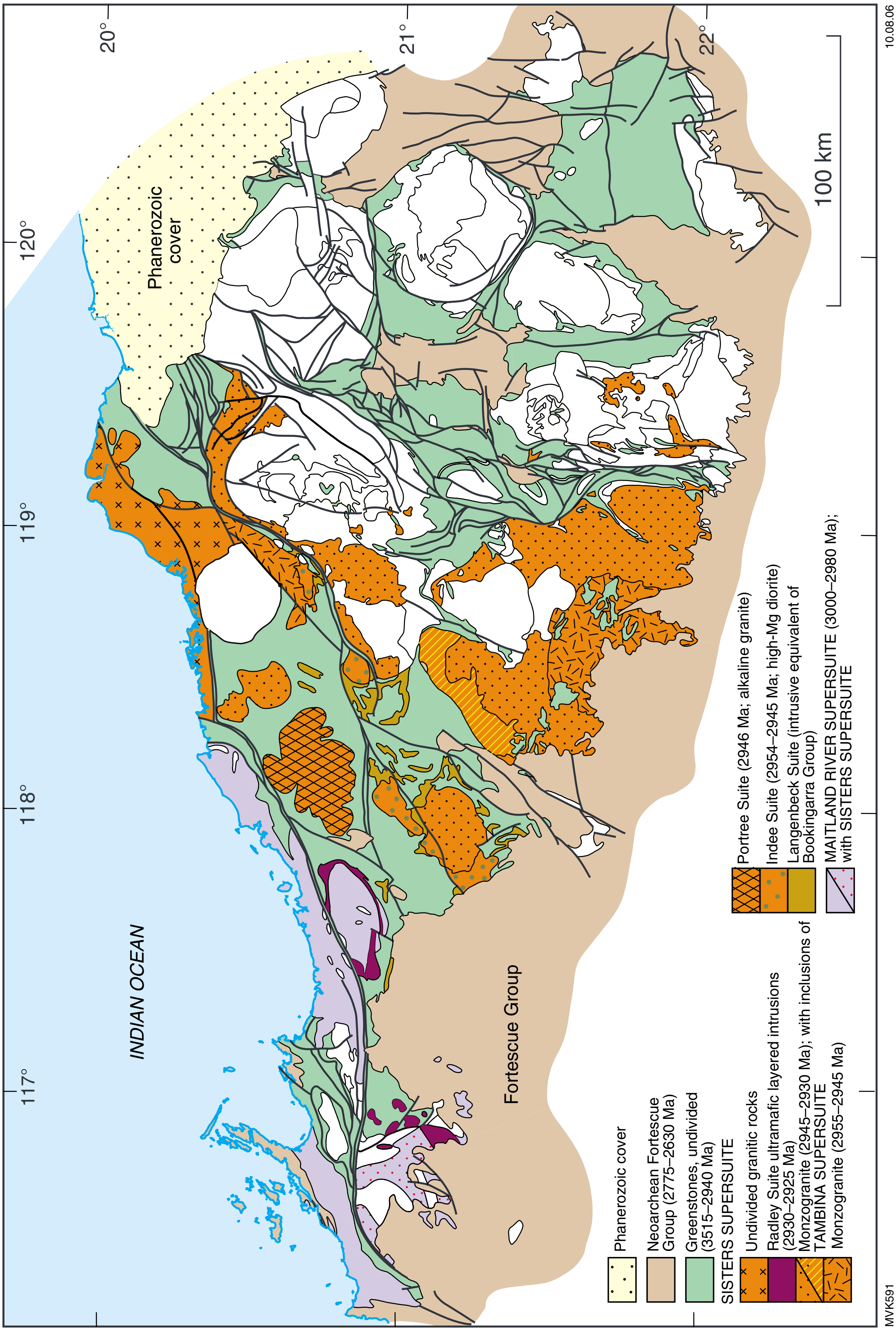

Figure 10. Map of the northern Pilbara Craton showing the distribution of the Maitland River and Sisters Supersuites: note the decreasing ages from west to east across the craton over this time period (from Van Kranendonk et al., 2006a)

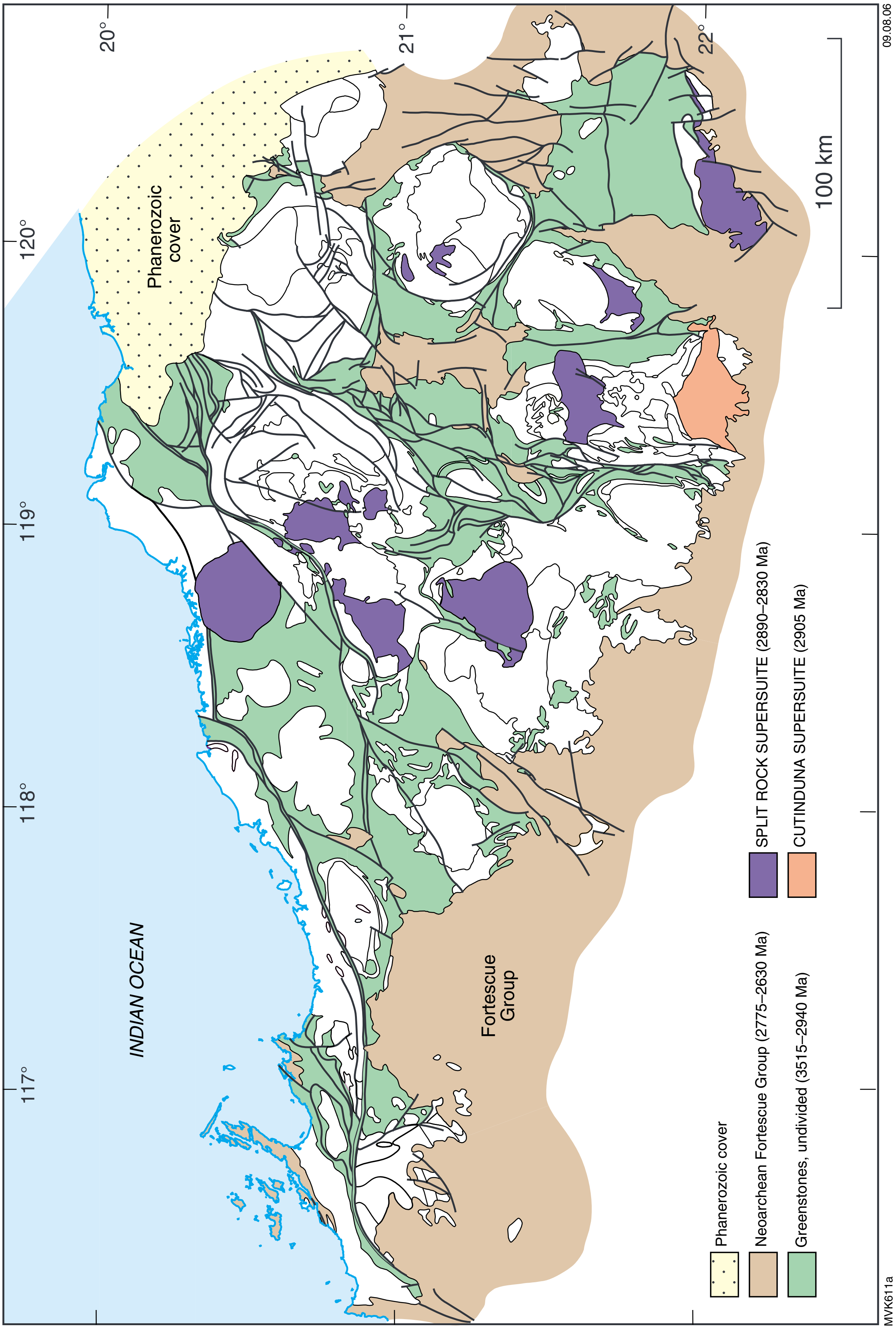

Figure 11. Map of the northern Pilbara Craton showing the distribution of the Cutinduna and Split Rock Supersuites (from Van Kranendonk et al., 2006a)

whereas Smithies (2000) showed that the compositions of the TTG suites are incompatible with an origin from slab melting in a modern-style, steep subduction setting. Rather, he suggested that the TTGs were more likely to be generated by melting of lower mafic crust — an interpretation in part supported by geochemical evidence presented by Bickle et al. (1993) who showed that at least some of the North Shaw Suite was derived from melting of crustal sources older than c. 3600 Ma. Isotopic and geochemical data show that the younger granitic supersuites were formed by progressive episodes of melting and recycling of felsic crust, including the TTG rocks of the Callina Supersuite as well as older felsic crust (Bickle et al., 1989; Collins, 1993; Smithies et al., 2003).

West Pilbara Superterrane

The West Pilbara Superterrane forms the northwestern ~20% of the Pilbara Craton (Fig. 3) and outcrops between Port Hedland and Cape Preston. The superterrane (Figs 2, 3, 12) is a collage of the 3.27 – 3.25 Ga Karratha Terrane (Roebourne Group and Cleland Supersuite, interpreted to be a rifted fragment of the East Pilbara Terrane), the c. 3.2 Ga Regal Terrane (2 km-thick basalt succession of the Regal Formation), and the 3.13 – 3.11 Ga Sholl Terrane (Whundo Group volcanics and Railway Supersuite). These terranes accreted together during collision with the East Pilbara Terrane at 3.07 – 3.05 Ga. This collision produced the Prinsep Orogeny and was accompanied by granite magmatism of the Elizabeth Hill Supersuite.

The most distinctive tectonic feature of the West Pilbara Superterrane is a northeasterly structural grain defined by elongation of the granitic complexes and the trend of greenstone belts, and by numerous closely spaced east- and northeast-striking strike-slip faults. The Karratha and Sholl Terranes are separated by the 1–2 km-wide Sholl Shear Zone, which bisects the West Pilbara Superterrane over an exposed length of 250 km (Figs 2, 12). The Sholl Shear Zone is steeply inclined to the northwest and is composed of mylonite and schist derived from granitic, volcanic, and intrusive rocks. The Karratha Terrane, and the overlying Regal Terrane, outcrop north of the Sholl Shear Zone and are everywhere separated by the Regal Thrust (Figs 2, 12).

The structural geology of the West Pilbara Superterrane, and the geochemistry of its volcanic and granitic rocks, establish that this superterrane is a product of Phanerozoic-style plate-tectonic processes (Krapez, 1993; Smith et al., 1998; Krapez and Eisenlohr, 1998; Van Kranendonk et al., 2002, 2006a; Hickman, 2004a; Smithies et al., 2005a, 2007). Smithies et al. (2005a) presented geochemical evidence that the Whundo Group (Sholl Terrane) formed in an intra-oceanic environment.

Mineralization in the West Pilbara Superterrane includes komatiite-hosted nickel–copper in the Roebourne Group, VHMS copper–zinc in the Whundo Group, and epigenetic gold and copper in sheared units immediately underlying the Regal Thrust. Nickel–copper, PGE, and vanadium–titanium mineralization is present in mafic–ultramafic intrusions of the Sisters Supersuite.

Karratha Terrane

The Karratha Terrane comprises the 3.27 – 3.25 Ga Roebourne Group (Table 1) and the 3.27 Ga Karratha Granodiorite.

The Roebourne Group contains two formations, the 1.5 km-thick Ruth Well Formation, consisting of metamorphosed ultramafic and mafic volcanic rocks, and the overlying 1.5 km-thick, 3.27 – 3.25 Ga Nickol River Formation, which is a metamorphosed package of ultramafic–felsic volcanic rocks overlain by metamorphosed siliciclastic and chemical sedimentary rocks. In the area south of Cleaverville, a conglomerate unit separates the main part of the Nickol River Formation from the Regal Formation. Stratigraphic relations are complicated by faulting, but available mapping evidence suggests that the conglomerate is part of the Nickol River Formation. If, however, this conglomerate is actually a basal member of the Regal Formation, the latter could not be oceanic. More mapping of this area is therefore needed.

The 3.27 Ga Karratha Granodiorite is a subvolcanic tonalite–granodiorite intrusion emplaced beneath the Roebourne Group and correlated with the Cleland Supersuite of the East Pilbara Terrane. The Karratha Granodiorite ranges from allotriomorphic granular tonalite to granodiorite (Nelson, 1998; Smith et al., 1998). Much older Nd T_{DM} model ages of 3.48 – 3.43 Ga from this intrusion (Sun and Hickman, 1998) indicate that magma generation involved older crust or enriched lithospheric mantle.

The Karratha Terrane is intruded by granitic intrusions of the 3.0 – 2.98 Ga Maitland River and 3.02 – 3.01 Ga Orpheus Supersuites within the Dampier and Harding Granitic Complexes and is separated from the Regal Terrane by the Regal Thrust in all parts of the West Pilbara Superterrane. In most parts of the West Pilbara Superterrane, the Karratha Terrane is separated from the Sholl Terrane by the Sholl Shear Zone, but the 3.24 Ga Tarlwa Pool Tonalite of the Cleland Supersuite in the Cherratta Granitic Complex (south of the Sholl Shear Zone), and isotopic evidence of c. 3.27 Ga crust underlying the Cherratta Granitic Complex south of the Whundo Group on COOYA POOYA* (Hickman, 2004b), together reveal that parts of the Karratha Terrane also underlie the area south of the Sholl Shear Zone.

The Sholl Shear Zone is a 1–2 km-wide belt of mylonite and schist that bisects the West Pilbara Superterrane over an exposed length of 250 km. An abrupt change in metamorphic grade across the Sholl Shear Zone, from lower greenschist facies in the Whundo Group (Sholl Terrane) to amphibolite facies in the Roebourne Group (Karratha Terrane), indicates either uplift of the Karratha Terrane relative to the Sholl Terrane along the Sholl Shear Zone, or metamorphism of the Karratha Terrane prior to its c. 3.07 Ga collision with the Sholl Terrane. The Regal Thrust is discussed below under the heading **Regal Terrane**.

* Capitalized names refer to standard 1:100 000 map sheets, unless otherwise indicated.

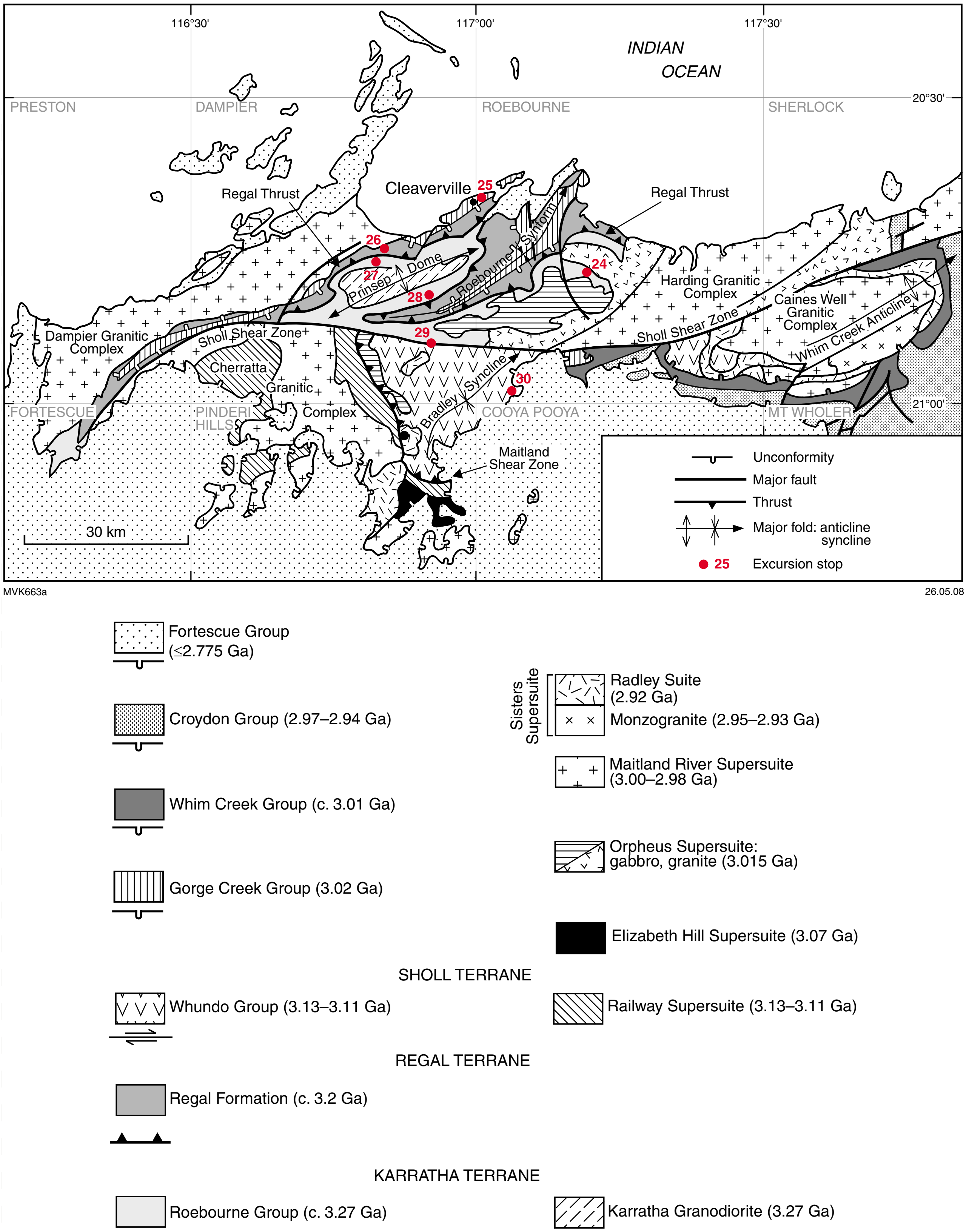

Figure 12. Simplified geological map of the northwestern Pilbara Craton showing lithostratigraphy, tectonic units, and excursion Localities 24–30

Sholl Terrane

The Sholl Terrane comprises the 3.13 – 3.11 Ga Whundo Group and the 3.13 – 3.11 Ga Railway Supersuite. Most Railway Supersuite outcrops are within the Cherratta Granitic Complex, but mafic and felsic intrusions of the supersuite also intrude the lower stratigraphic levels of the Whundo Group. The Sholl Terrane is intruded by granitic rocks of the c. 3.07 Ga Elizabeth Hill and 3.0 – 2.98 Ga Maitland River Supersuites in the Cherratta Granitic Complex, and by mafic–ultramafic intrusions of the Radley Suite (Sisters Supersuite).

The Whundo Group is divided into four lithostratigraphic formations (Table 1; Hickman, 1997). The lowermost Nallana Formation is >2.0 km thick and comprises metabasalt, ultramafic rocks, intermediate pyroclastic rocks, and dolerite sills. The conformably overlying Tozer Formation is 2.5 km thick and contains basalt, andesite, dacite, rhyolite, and thin metasedimentary units. The conformably overlying Bradley Basalt is more than 4.0 km thick and consists of metamorphosed, massive and pillowed basalt, with komatiitic basalt near its base. The Bradley Basalt is conformably overlain by felsic volcanic rocks of the Woodbrook Formation, which comprises metamorphosed rhyolite tuff and agglomerate, and only minor amounts of metabasalt.

A detailed chemostratigraphic analysis of the Whundo Group (Smithies et al., 2005a) revealed that it developed from a much more complex set of igneous processes than can be identified from lithological mapping. A lower volcanic package (lower and middle Nallana Formation) contains calc-alkaline and boninite-like rocks, a middle package (upper Nallana Formation, Tozer Formation, and lower Bradley Basalt) consists of tholeiitic rocks with minor units of boninite-like rocks and rhyolites, and an upper package (upper Bradley Basalt and Woodbrook Formation) contains calc-alkaline rocks, including adakites, Mg-rich basalts, and rhyolites (Fig. 13). Smithies et al. (2005a) concluded that the geochemistry of the group was generally consistent with deposition in an intra-oceanic arc environment. However, there is also published zircon and Nd T_{DM} model age data to suggest that significantly older crust underlies the Whundo Group.

Hickman and Kojan (2003) cited previously published Nd T_{DM} 3.34 – 3.25 Ga model ages from the Cleland, Railway, and Maitland River Supersuites in the Cherratta Granitic Complex. These ages are significantly older than the c. 3.2 Ga rifting event interpreted to have formed oceanic crust between the East Pilbara Terrane and Karratha Terrane (Fig. 5). Subsequently, Hickman (2004b) noted that c. 3.27 Ga xenocrystic zircons in 3.00 – 2.98 Ga rocks of the Cherratta Granitic Complex on PINDERI HILLS (southeast of the Sholl Terrane) suggest partial inheritance from crust of Karratha Terrane age. Unless the Whundo Group is highly allocthonous upon the underlying Cherratta Granitic Complex, the evidence outlined above indicates that the crust on which it was deposited must have included >3.25 Ga crust, or clastic sedimentary rocks derived from such crust.

Table 1. Archean lithostratigraphy of the Roebourne area

Group	*Formation*	*Thickness (m)*	*Lithology and relationships*
Gorge Creek	Cleaverville Formation	1500	Banded iron-formation, chert, fine-grained clastic sedimentary rocks. Age c. 3.020 Ma
~ ~ ~ *low-angle unconformity* ~ ~ ~			
Whundo Group	Woodbrook Formation	1000	Rhyolite tuff and agglomerate; minor basalt and thin banded iron-formation. Age 3117 ± 3 Ma
	Bradley Basalt	>4000	Pillow basalt, massive basalt, minor units of felsic tuff and chert. Age 3115 ± 5 Ma
	Tozer Formation	2500	Calc-alkaline volcanics, including felsic pyroclastic units. Minor chert and thin banded iron-formation. Age c. 3120 Ma
	Nallana Formation	2000	Dominantly basalt, but includes minor ultramafic and felsic units. Felsic tuff dated at 3125 ± 4 Ma. Base of formation intruded by 3310 Ma granitoids and truncated by Maitland shear zone
- - - *Sholl Shear Zone* - - -			
	Regal Formation	2000	Basal peridotitic komatiite overlain by pillow basalt and local chert units. Intruded by microgranite and c. 3015 Ma felsic porphyry
- - - *Regal Thrust* - - -			
Roebourne Group	Nickol River Formation	100–500	Banded chert, iron-formation, ferruginous clastic sedimentary rocks, quartzite, felsic volcanic, carbonate rocks, and volcanogenic sedimentary rocks and local conglomerate. Schist protolith younger than 3269 ± 2 Ma, and rhyolite dated at 3251 ± 6 Ma
	Ruth Well Formation	1000–2000	Basalt and extrusive peridotite with thin chert units. Intruded by granodiorite and tonalite dated at 3270 ± 2 Ma

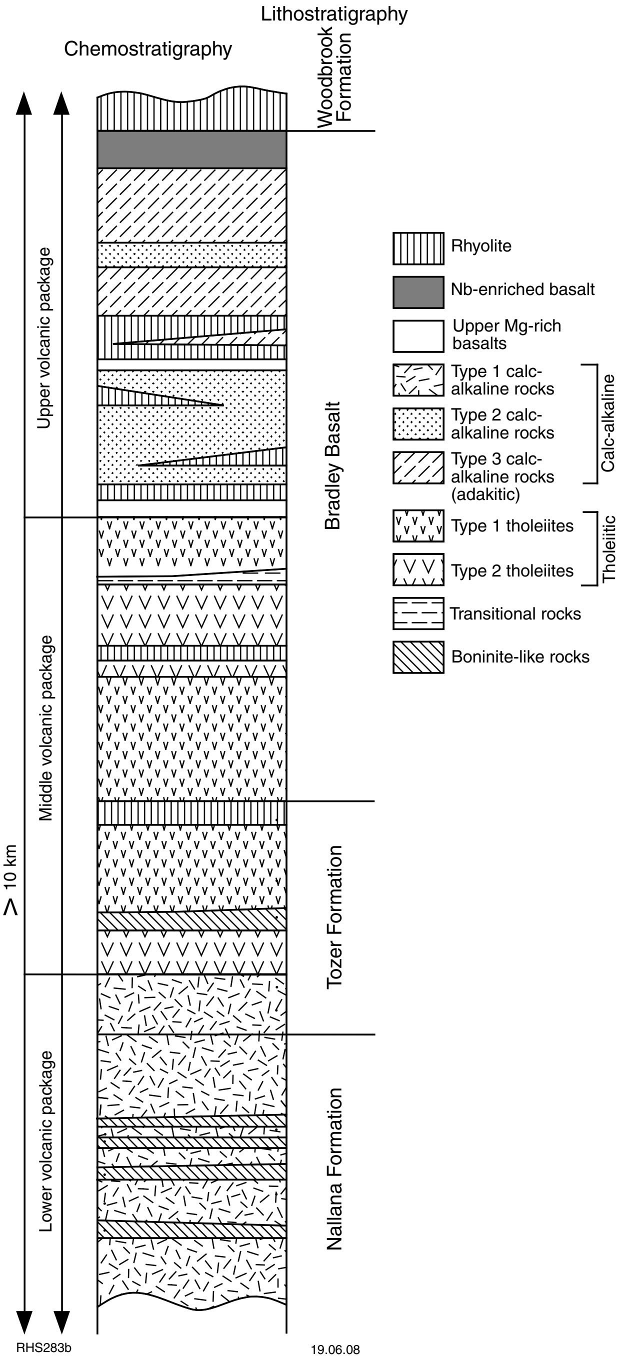

Figure 13. Chemostratigraphic division of the Whundo Group into three volcanic packages, with lithostratigraphic formation boundaries shown for comparison (after Van Kranendonk et al., 2006a)

Recent U–Pb zircon dating of the Bullock Hide Intrusion (layered mafic–ultramafic intrusion), which was previously assigned to the 3025–3010 Ma Orpheus Supersuite (Van Kranendonk et al., 2006a), indicates a 3117 ± 4 Ma intrusive age (Wingate, in prep.b, sample 178164). This suggests that the Railway Supersuite includes a suite of major mafic–ultramafic intrusions, which probably also includes the Andover Intrusion (interpreted to be a faulted section of the Bullock Hide Intrusion, Hickman, 2002).

Regal Terrane and the Regal Thrust

The Regal Terrane is entirely composed of the Regal Formation, a 2 km-thick volcanic pile of metamorphosed basalt and basal komatiitic peridotite, with rare chert. Locally, there are numerous dolerite dykes (Locality 25) of uncertain age in the formation. Sills of dolerite and gabbro also locally intrude the formation. In all areas except Cleaverville (Fig. 12), the formation has been metamorphosed to upper greenschist and amphibolite facies. Lower grade metamorphism in the outcrop at Cleaverville is attributed to separation from the main outcrop area by faults. The Regal Terrane is separated from the underlying Karratha Terrane by the Regal Thrust, is intruded by the 3.025 – 3.01 Ga Orpheus Supersuite and the 3.00 – 2.98 Ga Maitland River Supersuite, and is disconformably overlain by the 3.02 Ga Cleaverville Formation (Gorge Creek Group). The Regal Formation has not been dated, but Nd isotopic data (Smithies et al., 2007) is consistent with a depositional age between 3.2 and 3.0 Ga.

The origin of the Regal Terrane remains problematic. The lithological composition and geochemistry of the only component unit, the Regal Formation, is consistent with oceanic basalt, but the formation is sandwiched between underlying and overlying continental crust (Karratha Terrane below, and Cleaverville Formation above) over an area of 3000 km^2. Before the regional geochemistry of the Regal Formation became known (Ohta et al., 1996; Sun and Hickman, 1998; Smithies et al., 2007), the formation was considered to stratigraphically overlie the Nickol River Formation, although it was recognized that the contact was regionally tectonized (Hickman, 1997). However, geochemical evidence combined with the existence of a bedding-parallel mylonitic zone between the Regal Formation and the Roebourne Group led to interpretation of the Regal Terrane as a slab of oceanic crust obducted across the Karratha Terrane (Sun and Hickman, 1998; Hickman, 2004a). This interpretation is dependent on geochemical data from the Regal Formation (a flat REE pattern), although, if dating of the dyke swarm at Karratha (Locality 25) shows it to be older than 3.02 Ga, this would provide some support for derivation of the formation from oceanic crust.

Although the depositional environment of the Regal Formation is uncertain, we here reservedly adopt the interpretation that the Regal Thrust is a regional sole thrust beneath the Regal Formation (Hickman, 2004a; Van Kranendonk et al., 2006a). The mylonite in the thrust contains refolded isoclinal folds, the orientation of which led Hickman (2001b, 2004a) to suggest that obduction of the Regal Terrane was from the north or northwest. A tectonothermal event at c. 3.16 Ga (K–Ar data, Kiyokawa and Tairo, 1998; zircon geochronology, Smith et al., 1998) in the Karratha Granodiorite beneath the thrust was interpreted by Hickman (2004a) to coincide with the age of the Regal Thrust. However, Van Kranendonk et al. (2006a,b) suggested that the Regal Formation could have been derived from within the juvenile mafic crust in the zone of rifting between the East Pilbara Terrane and the Karratha Terrane, requiring thrusting from the south or southeast. The same authors also suggested that the thrusting occurred during collisions among the Karratha Terrane, Sholl Terrane, and East Pilbara Terrane, interpreted to be at 3.07 Ga (Fig. 5).

Elizabeth Hill Supersuite

The Elizabeth Hill Supersuite intrudes the West Pilbara Superterrane and the western part of the East Pilbara Terrane. In the west Pilbara, the Elizabeth Hill Supersuite is represented by the Cliff Pool Tonalite, which outcrops over 70 km^2 as part of the Cherratta Granitic Complex east of the Munni Munni Intrusion, and possibly also by gneiss in the Caines Well Granitic Complex. A sample of Cliff Pool Tonalite collected approximately 6 km south-southeast of the Elizabeth Hill mine was dated by Nelson (1998, sample 142661) at 3068 ± 4 Ma. The Cliff Pool Tonalite is a foliated biotite tonalite, which includes xenocrystic zircons of Railway Supersuite age (Nelson, 1998, sample 142661), supporting the field observation of an intrusive relationship to the c. 3.13 Ga Pinnacle Hill Gneiss. Its northern contact with the Pinnacle Hill Gneiss is diffuse and interlayered with sheets of the tonalite intruding rafts of the gneiss. The southern contact of the tonalite is intruded by granitic rocks of the 3.0 – 2.98 Ga Maitland River Supersuite. In the Caines Well Granitic Complex (dominantly composed of the 3.0 – 2.98 Ga Maitland River Supersuite), an enclave of biotite monzogranite gneiss was dated by Nelson (1997, sample 118965) at 3093 ± 4 Ma. The gneiss contained a xenocrystic zircon population dated at 3111 ± 13 Ma, which coincides with the age of the Railway Supersuite. The two youngest near-concordant zircons used to calculate the 3093 Ma age have a model age of 3084 Ma, which is similar to the age of many zircons in the c. 3070 Ma Elizabeth Hill Supersuite. Therefore, the Caines Well Granitic Complex, which is poorly exposed, may include intrusions of both the Railway and Elizabeth Hills Supersuites.

De Grey Superbasin

In the west Pilbara, the De Grey Superbasin (Fig. 2) unconformably overlies the West Pilbara Superterrane, and contains sedimentary and volcanic rocks of the 3.02 – 2.93 Ga De Grey Supergroup. It is intruded by contemporaneous rocks of the Orpheus, Maitland River, and Sisters Supersuites. Over the entire northern Pilbara Craton, the De Grey Supergroup is composed of four groups, deposited in five basins: the 3.02 Ga Gorge Creek Group, deposited in the Gorge Creek Basin across the West Pilbara Superterrane and East Pilbara Terrane; the 3.01 Ga Whim Creek Group, deposited in the Whim Creek Basin (restricted to the Whim Creek greenstone belt) on the southeastern margin of the West Pilbara Superterrane; the 2.97 – 2.94 Ga Croydon Group, deposited in the Mallina Basin between the West Pilbara Superterrane and East Pilbara Terrane, and in the Lalla Rookh Basin within the East Pilbara Terrane; and the >2.926 Ga Nullagine Group, deposited in the Mosquito Creek Basin between the East Pilbara Terrane and Kurrana Terrane.

The 3.02 Ga Gorge Creek Basin, which in the west Pilbara is filled by the Cleaverville Formation (banded iron-formation and siliciclastic sedimentary rocks) of the Gorge Creek Group, was deposited soon after accretion of the West Pilbara Superterrane and collision with the East Pilbara Terrane (3.07 Ga, Prinsep Orogeny). This, combined with evidence from growth faults in the lower part of the Gorge Creek Group, indicates deposition during extensional relaxation following collisional orogenesis. It was deformed prior to deposition of the unconformably overlying Whim Creek and Croydon Groups.

The Whim Creek Basin contains the c. 3.01 Ga Whim Creek Group, comprising the Warambie Basalt and Red Hill Volcanics. Previously interpreted to represent a continental arc, the Whim Creek Basin is now re-interpreted as a pull-apart basin that formed as a result of extension caused by break-off of the subducted slab associated with formation of the c. 3.13 – 3.11 Ga Whundo Group intra-oceanic arc (Van Kranendonk et al., 2006a).

In the Mallina Basin (Figs 2, 3), the 2.97 – 2.94 Ga Croydon Group consists of a variety of clastic sedimentary and volcanic rocks. On the western margin of the basin are the c. 2964 Ma Cistern Formation of quartz-rich clastic sedimentary and felsic volcaniclastic rocks, the Rushall Slate, the Bookingarra Formation of silicic high-Mg basalts, and the Kialrah Rhyolite. The main part of the basin fill consists of the Constantine Sandstone, Mallina Formation (fine sandstone and shale), Bookingarra Formation (mafic volcanic rocks), and Kialrah Rhyolite.

Granitic supersuites younger than 3.02 Ga

Orpheus Supersuite

Felsic intrusions of the Orpheus Supersuite have been extensively dated. They include the Mount Gregory Monzodiorite, the Forrestier Bay Granodiorite, the Snell Well Granodiorite, the Stone Yard Granophyre, and many individual dacite stocks, dykes, and sills assigned to either the Rea Dacite or the Mount Wangee Dacite. The Stone Yard Granophyre, dated at 3014 ± 6 Ma, intrudes the 3.02 Ga Cleaverville Formation, which is the oldest stratigraphic formation preserved on both sides of the Sholl Shear Zone. The stratigraphic mismatch of the Karratha Terrane and Sholl Terrane across the Sholl Shear Zone suggests that there was a major strike-slip movement of at least 200 km (Sun and Hickman, 1998) before deposition of the Cleaverville Formation, and that intrusion of the Orpheus Supersuite was therefore relatively late in the history of the shear zone.

Maitland River Supersuite

The 3.00 – 2.98 Ga Maitland River Supersuite is the most voluminous granitic supersuite of the west Pilbara (Fig. 10), and is represented by large granitic intrusions in four granitic complexes (the Dampier, Cherratta, Harding, and Caines Well Granitic Complexes) that invade and separate the three >3.11 Ga terranes of the West Pilbara Superterrane. In the Cherratta Granitic Complex, the

Maitland River Supersuite includes the Jean Granodiorite, Whyjabby Granodiorite, Brill Monzogranite, Toorare Tonalite, Waloo Waloo Granodiorite, and Mardeburra Granodiorite. In the Dampier Granitic Complex, it includes the Miaree Granite, Hearson Cove Monzogranite, and the Eramurra Monzogranite, and in the Caines Well Granitic Complex it is represented by the Caines Well Granite. The Harding Granitic Complex is very poorly exposed, but is interpreted to include parts of the Toorare Tonalite in addition to large areas of concealed, unnamed granitic intrusions.

The various intrusions of the Maitland River Supersuite collectively form a 300–400 km-long northeast-trending belt on both sides of the Sholl Shear Zone (Figs 2, 10), indicating intrusion during a major regional event. Crystallization ages (3.00 – 2.98 Ga) are only slightly younger than the Red Hill Volcanics of the Whim Creek Group (c. 3009 Ma), but the distribution of this group is apparently restricted to a relatively small continental pull-apart basin (Barley, 1987; Van Kranendonk et al., 2006a). Van Kranendonk et al. (2006a) suggested that the Maitland River Supersuite marks the early stages of regional extension and crustal melting following the c. 3.05 Ga collision between the West Pilbara Superterrane and the East Pilbara Terrane. Nd T_{DM} model ages from intrusions of the supersuite differ systematically according to location. Granitic intrusions in the Dampier Granitic Complex have Nd T_{DM} model ages of 3.387 – 3.247 Ga (Hickman and Smithies, in prep.), indicating that the Karratha Terrane (and older underlying crust) contributed material to this part of the Maitland River Supersuite. By contrast, Nd T_{DM} model ages from the supersuite in the Cherratta Granitic Complex fall within the range 3.25 – 3.14 Ga (Sun and Hickman, 1998), and xenocrystic zircons are all younger than 3.28 Ga, indicating that any underlying crust is younger than c. 3.28 Ga.

Sisters Supersuite (including the Indee, Langenbeck, and Portree Suites)

The c. 2.95 – 2.92 Ga Sisters Supersuite consists of a wide variety of igneous rocks distributed across the western central part of the northern Pilbara Craton, including a major component that intruded the Mallina Basin and western part of the East Pilbara Terrane in the Central Pilbara Tectonic Zone (Fig. 2). The oldest components of the Sisters Supersuite were emplaced into clastic sedimentary rocks of the Mallina Basin between 2.95 and 2.94 Ga. These include high-Mg diorite (sanukitoid) rocks (Smithies and Champion, 2000) of the Indee Suite, alkaline granite of the Portree Suite, and ultramafic–mafic intrusions of the Langenbeck Suite. The Indee and Langenbeck Suites host orthomagmatic PGE–gold mineralization. This magmatism was followed by widespread and voluminous granitic magmatism (2.94 – 2.92 Ga) that was dominant in the western half of the East Pilbara Terrane, but was continuous across the Mallina Basin and into the West Pilbara Superterrane. The youngest component of the Sisters Supersuite is the Radley Suite of layered mafic–ultramafic intrusions in the West Pilbara Superterrane, some of which host PGE and nickel–copper mineralization.

The Langenbeck, Indee, and Portree Suites resulted from melting of mantle sources but they also show clear evidence of contamination by continental crust (i.e. high Th, Zr, and light-rare-earth element (LREE) concentrations; high Th/high-field-strength element (HFSE) and LREE/HFSE ratios) (Smithies et al., 2004). Although the geochemistry of individual members of the Langenbeck Suite differs significantly, they share surprisingly similar La/Sm and La/Zr ratios that cannot be accounted for through assimilation by any known outcropping Pilbara crustal component. Likewise, sanukitoids of the Indee Suite show enrichment in large-ion-lithophile elements and in Cr, Ni, and Mg number, which cannot be tied to crustal assimilation. Rather, the Langenbeck and Indee Suite magmatism is interpreted to be derived from a mantle source that was modified (enriched) during an earlier subduction event, probably that which formed the Whundo intra-oceanic arc. The later remobilization of the enriched (presumably lithospheric) mantle may be linked to break-off of the subducted slab. The widespread granitic rocks of the Sisters Supersuite represent melts derived from pre-existing granitic crust, and were derived from conductive heating related to remobilization of enriched mantle material during the end phase of the slab break-off event.

Split Rock Supersuite

The 2.89 – 2.83 Ga Split Rock Supersuite consists of post-tectonic, sheet-like granitic plutons distributed in a northwest-trending belt across the northern Pilbara Craton. These highly fractionated intrusions are typically fringed by pegmatite veins containing tin–tantalum–lithium mineralization. The Mallina Basin is intruded by one of these granites, the Myanna Leucogranite (Van Kranendonk et al., 2006a), and tin and tantalum mineralization at Pippingarra and Bore Creek are interpreted to be related to this intrusion.

Deformation events

East Pilbara Terrane

There are two types of regional-scale structures in the East Pilbara Terrane: a dome-and-basin pattern of granite-cored domes and intervening synclinal tracts of greenstones that evolved over the entire history of the craton (Fig. 6; Hickman, 1984; Van Kranendonk, 2003a; Van Kranendonk et al., 2004); and structures within the Lalla Rookh – Western Shaw structural corridor that formed during a discrete event of sinistral transpression at c. 2940 Ma (Van Kranendonk and Collins, 1998; Zegers et al., 1998).

A general characteristic of the dome-and-basin pattern is that the margins of granitic complexes commonly have steeply dipping foliations that are parallel to, and pass across, contacts with greenstones (Hickman, 1984; Van Kranendonk et al., 2004). Adjacent greenstones are metamorphosed to amphibolite to greenschist facies in contact metamorphic aureoles, at pressures of up to 6 kbar (Delor et al., 1991). Farther away from granitic contacts, greenstones are still commonly steeply dipping, but unaffected by penetrative foliations, and are metamorphosed at only greenschist to prehnite–pumpellyite facies. This reflects either a decrease in temperature and strain away from the granitic heat source, or a younger age of deposition of the stratigraphically higher rocks, or both. Although the foliations must be younger than their host rocks, it is difficult to constrain the precise age of these structures. Many of the host rocks have yet to be dated, or the metamorphic minerals defining the structures have been largely reset by successive generations of intrusive granite, or other regional events, or both. In fact, attempts at Ar–Ar dating of metamorphic minerals in foliations around the Shaw Granitic Complex (Davids et al., 1997; Zegers et al., 1999) indicate a long history of heating and cooling that can be interpreted as progressive and commonly coaxial development of the dome-and-basin structures. As a result, it is generally difficult to ascribe a particular set of structures to a precise time interval, thus some structures have been grouped together in broad categories. The geological history presented below is based on craton-wide mapping by GSWA as presented by Van Kranendonk et al. (2007). Blewett (2000) presented a different interpretation of the structural evolution of the northern Pilbara Craton.

D_1–D_2 deformation at 3490–3410 Ma

The earliest recognized structures span a period from c. 3490 to 3410 Ma and can locally be divided into two separate deformation events. These include the local development of synvolcanic, listric, normal growth faults in the Dresser Formation (Nijman et al., 1998a, 1999a; Van Kranendonk and Hickman, 2000), Duffer Formation (Zegers et al., 1996), and Panorama Formation (Nijman et al., 1998b, 1999b). Faults in the Dresser Formation are filled by hydrothermal chert–barite veins and were interpreted to have formed as a result of caldera collapse during intrusion of a subvolcanic laccolith (Nijman et al., 1998b, 1999b; Nijman and de Vries, 2004; Van Kranendonk, 2006). Zegers et al. (1996) presented an extensional core complex model for the origin of extensional faults in the Duffer Formation (but see **Evidence for thrusting** below).

D_1 deformation also caused tilting of the Coonterunah Group and Talga Talga Subgroup away from the cores of synvolcanic granitic rocks, prior to deposition of the Salgash Subgroup and intrusion of the Tambina Supersuite of granitic rocks (Hickman, 1983; Van Kranendonk, 2000; Van Kranendonk et al., 2006a). DiMarco and Lowe (1989) argued that these structures probably formed as a result of granitic intrusion and doming during episodes of felsic volcanism. Local D_1 folds were also developed in the Coonterunah Subgroup in the East Strelley greenstone belt, where their geometry is consistent with formation during gravitational sliding off the rising Carlindi Granitic Complex (Van Kranendonk and Collins, 1998; Van Kranendonk, 2006), as proposed for folds of similar geometry but different age adjacent to the Mount Edgar Granitic Complex (Collins, 1989).

Gneissic fabrics were developed in >3450 Ma protoliths of the Shaw Granitic Complex prior to the intrusion of homogeneous, weakly foliated granitic rocks of the c. 3430 Ma Tambina Supersuite; this is indicated by the presence in these rocks of xenoliths of migmatitic orthogneiss (see fig. 13a of Van Kranendonk, 2003b). Gneissic layering was the result of partial melting of TTG protoliths and lit-par-lit injection of the resultant leucogranite during synmagmatic doming (Van Kranendonk and Hickman, 2000). Partial melting in the Shaw Granitic Complex continued to c. 3400 Ma as determined from zircon dates for melt veins and diatexite (Van Kranendonk, 2000). A similar age of 3410 ± 7 Ma was obtained for partial melting of >3600 Ma gneiss from the Warrawagine Granitic Complex (Nelson, 1999b).

D_3 deformation at 3325–3290 Ma

D_3 structures accompanied deposition of the felsic volcanic Wyman Formation and intrusion of related granitic rocks at c. 3325–3308 Ma over the eastern part of the East Pilbara Terrane, but most notably in the Corunna Downs and Mount Edgar Granitic Complexes (Fig. 14). Collins (1989) and Collins et al. (1998) documented a set of structures within and around the Mount Edgar Granitic Complex that are related to diapirism. These structures include a 1–3 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 granitic complex was achieved. This shear zone gradually dies out to the northwest and northeast in undeformed synkinematic granites emplaced into greenstones at a high level and locally associated with porphyry copper–molybdenum mineralization (see **Part 2, Locality 16**). Associated ring faults in the flanking greenstones accommodated sinking of the greenstones as well as later exhumation of marginal components adjacent to granitic complexes (Collins and Van Kranendonk, 1999; Van Kranendonk et al., 2004). A synformal roof pendant of greenstones and c. 3450 Ma gneisses within the Mount Edgar Granitic Complex formed during diapiric emplacement of two domical lobes of synkinematic c. 3300 Ma granite. In the McPhee Formation of the adjacent Marble Bar greenstone belt, a set of tight asymmetric folds that verge away from the Mount Edgar Granitic Complex formed as a result of gravitational sliding of greenstones off the rising dome (Collins, 1989). Greenstones in the Warrawoona Syncline were deformed into a zone of sinking between the coeval Mount Edgar and Corunna Downs Granitic Complexes, which is characterized by a core of intense vertical stretching (see **Part 2, Locality 10**). This deformation event is clearly identified as separate from subsequent events by a high-angle unconformity between the Kelly and Sulphur Springs Groups (Van Kranendonk, 2004, fig. 20; Van Kranendonk et al., 2004, fig. 12).

D_4 deformation at c. 3240 Ma

D_4 structures were formed in the western part of the East Pilbara Terrane during eruption of the Sulphur Springs Group and widespread granite intrusion at c. 3240 Ma. In the Soanesville greenstone belt, intrusion of the Strelley Granite resulted in a suite of synvolcanic growth faults in the Kangaroo Caves Formation that were sites of

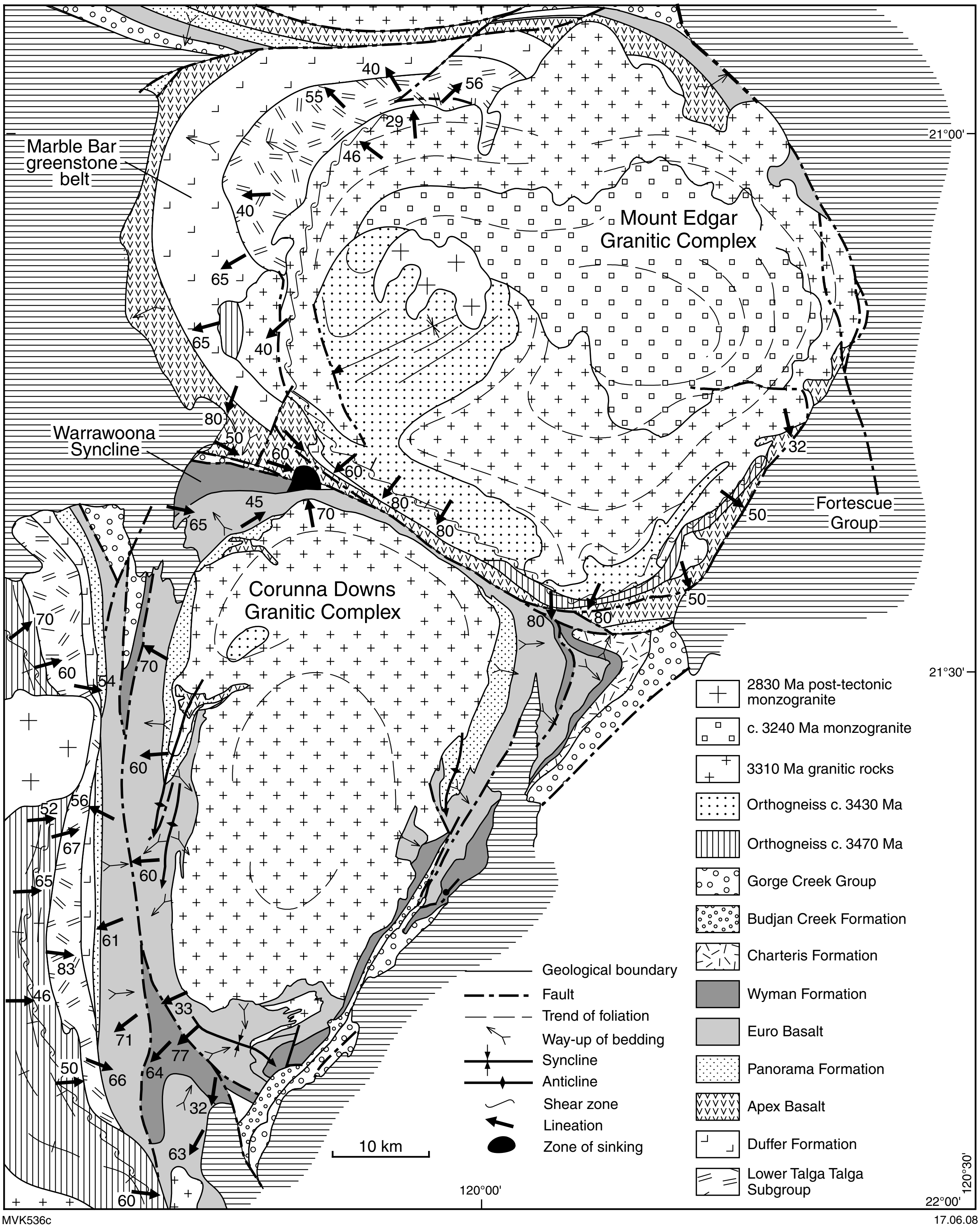

Figure 14. Geological map of part of the East Pilbara Terrane showing the radial distribution of metamorphic mineral elongation lineations around the Mount Edgar and Corunna Downs Granitic Complexes that support a model of partial convective overturn (from Van Kranendonk et al., 2004)

hydrothermal venting and massive sulfide deposition (Vearncombe et al., 1998). Late caldera collapse resulted in renewed faulting and deposition of an olistostrome breccia with synsedimentary folds (Van Kranendonk, 2000; Pirajno and Van Kranendonk, 2005). Intrusion of the Strelley Granite and deposition of the Kangaroo Caves Formation were accompanied by dome-and-basin folding and granite intrusion in the northeastern part of the Yule Granitic Complex, including the Tambourah Dome. Van Kranendonk (1997) showed that this dome-and-basin pattern has similar hook-shaped folds and ring faults to those characteristic of diapiric emplacement of the Corunna Downs Granitic Complex (Fig. 14; Van Kranendonk et al., 2004). Doming was accompanied by amphibolite-facies contact metamorphism of the Western Shaw greenstone belt (Wijbrans and McDougall, 1987).

Rifting event at 3220–3165 Ma

Rifting of the East Pilbara Terrane (Fig. 5) followed deposition of the Sulphur Springs Group (Hickman, 2004a) and, by c. 3165 Ma, had evolved to produce northwest–southeast separation of the Karratha Terrane (Van Kranendonk et al., 2007). This established a northeast–southwest trending rift zone, which was intruded by c. 3.18 Ga mafic intrusions of the Dalton Suite (see below) and c. 3.17 Ga granitic intrusions of the Mount Billroth Supersuite. The rift is interpreted to have included formation of oceanic crust, because widespread Nd–isotope data from post-3.13 Ga rocks within it indicate underlying juvenile (<3.2 Ga) crust (Van Kranendonk et al., 2004, 2006a). The approximate margins of the rift are now interpreted to be the northwestern margin of the East Pilbara Terrane in the southeast and the Sholl Shear Zone in the northwest. Previous suggestions that juvenile crust also extends into the western part of the East Pilbara Terrane (Van Kranendonk et al., 2004) are now contradicted by U–Pb zircon data. For example, the Carlindi Granitic Complex, along the Tabba Tabba Shear Zone, includes c. 3255 Ma granitic rocks that contain 3629–3464 Ma xenocrysts (Beintema, 2003) and 3252 ± 4 Ma rhyolite of the Sulphur Springs Group (Bodorkos, in prep., sample 160258). Likewise, the northwestern part of the Yule Granitic Complex includes the 3421 ± 2 Ma Yallingarrintha Tonalite (Nelson, 1999a, sample 142170). If any juvenile crust underlies the western part of the East Pilbara Terrane, as suggested by Nd data from some granitic rocks (Van Kranendonk et al., 2004, 2007), it must be mixed with segments of older crust typical of the remainder of the East Pilbara Terrane.

Deposition of the c. 3.2 Ga Soanesville Group was influenced by horst and graben faults (Wilhelmij and Dunlop, 1984), suggesting that clastic sedimentation accompanied early rifting. In the western part of the East Pilbara Terrane, the 3.18 Ga mafic and ultramafic intrusions of the Dalton Suite were intruded into the Soanesville Group as sills and dykes, whereas farther east, similar intrusions (not yet dated) were emplaced along many dome contacts, indicating crustal extension across the entire East Pilbara Terrane. In the West Pilbara Superterrane, the upper clastic part of the Nickol River Formation may be of similar age to the Soanesville Group, and large mafic and ultramafic sills also intrude this formation. Along the southeastern margin of the East Pilbara Terrane, an undated clastic sedimentary succession is intruded by a suite of mafic and ultramafic sills in the Lionel area, north of Nullagine, and there are similar intrusions along the northern and southern margins of the Mosquito Creek Basin (a 3.29–2.93 Ga rift basin).

West Pilbara Superterrane

The geological record of the west Pilbara starts at 3.27 Ga with deposition of the Roebourne Group and intrusion of the Karratha Granodiorite. Isotopic evidence indicates that the age of older underlying crust is between 3.5 and 3.3 Ga (Sun and Hickman, 1998); this is similar to the East Pilbara Terrane where the Sulphur Springs Group overlies 3.5 – 3.3 Ga crust. This similarity suggests that the Karratha Terrane is a rifted fragment of the East Pilbara Terrane.

West Pilbara Superterrane D_1 deformation at c. 3160 Ma

D_1 of the West Pilbara Superterrane is restricted to the area northwest of the Sholl Shear Zone. The age of D_1 is established by a 3.16 – 3.15 Ga thermotectonic event in the c. 3.27 Ga Karratha Granodiorite. This event is recorded by zircon geochronology from a granodiorite sample from 20 km southeast of Karratha (Smith et al., 1998) and K–Ar geochronology on hornblende from a sample of the Karratha Granodiorite from about 5 km south of Karratha (Kiyokawa, 1993). Both samples, though collected from widely separated locations, were from positions approximately 2 km beneath the Regal Thrust.

Hickman (2004a) proposed that D_1 of the West Pilbara Superterrane included obduction of oceanic crust (Regal Formation) from a plate to the north or northwest of the Karratha Terrane. At this time, the Karratha Terrane was rapidly separating from the East Pilbara Terrane (see **Rifting event at 3220–3165 Ma** above), and this northerly or northwesterly drift of Karratha Terrane is likely to have impinged on any plate to the north. If this plate was oceanic, collision could have resulted in its partial obduction onto the Karratha Terrane. Under this interpretation, the 3.16 Ga thermotectonic event marks plate collision and metamorphism of the Karratha Terrane as the northern oceanic plate was thrust across it. The Regal Thrust and several parallel thrusts may therefore be 3.16 Ga structures. Large-scale isoclinal folds in the Nickol River Formation (Karratha Terrane) southeast of Mount Regal were recumbent and southwesterly facing prior to tilting by the D_9 Prinsep Dome (see **D_9 deformation** below). Intrafolial isoclines are relatively common within the Nickol River Formation of the Mount Regal area (Hickman et al., 2000) and are probably minor structures related to the recumbent folds, and to the Regal Thrust. A bedding-parallel tectonic foliation (S_1) 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. An alternative interpretation (Van Kranendonk et al., 2006a,b) is that the Regal Formation is c. 3.2 Ga oceanic crust tectonically emplaced from the south at 3.07 Ga at the time of collision of the West Pilbara Superterrane and East Pilbara Terrane. Dating and structural analysis of the Regal Thrust is complicated by several phases of deformation between its initial development and 2.92 Ga. In particular, the D_9 (2.95 – 2.93 Ga) folding responsible for the Prinsep Dome should be represented by structures and metamorphism superimposed on the mylonite within the Regal Thrust.

Deformation common to the east and west Pilbara regions (3070 Ma onwards)

Following collision of the West Pilbara Superterrane and East Pilbara Terrane at c. 3.07 Ga, the east and west Pilbara regions shared major deformation events and some magmatic events, but each also shows evidence of more localized tectonomagmatic events. In the following description, the deformation histories of both regions are combined and numbered sequentially, with notes on distribution.

D_5 deformation at c. 3070 Ma (Prinsep Orogeny)

The D_5 deformation is most evident in the Sholl Terrane and along the Sholl Shear Zone and was accompanied by intrusion of the Elizabeth Hill Supersuite at 3.07 Ga. This event is recognized both in the Sholl Terrane and in the northwestern part of the East Pilbara Terrane, and was interpreted by Van Kranendonk et al. (2006a, 2007) to coincide with north–south convergence and collision between the West Pilbara Superterrane and East Pilbara Terrane, which caused the Prinsep Orogeny. Deformation included southwest thrusting of the Whundo Group onto the Cherratta Granitic Complex across the Maitland Shear Zone. An alternative interpretation (Van Kranendonk et al., 2006a) of the Regal Thrust is that this could be a 3.07 Ga rather than a 3.16 Ga structure.

D_5 also included major sinistral strike-slip movement on the Sholl Shear Zone. Measurable strike-slip movement on the Sholl Shear Zone is dextral, later than c. 2.95 Ga (Bookingarra Formation), and extends approximately 30–40 km (measured displacement of the Whim Creek and Croydon Groups). However, the dextral movement was preceded by greater sinistral movement prior to 3.02 Ga deposition of the Cleaverille Formation (Hickman, 2001b). Direct evidence for sinistral movement is provided by porphyroclasts within fine-scale mylonitic lamination. These shear-sense indicators indicate sinistral movement along the foliation planes of the mylonite. Evidence for major early movement is also provided by the stratigraphic mismatch across the Sholl Shear Zone.

D_5 structures formed in the western part of the East Pilbara Terrane include synmagmatic shear zones in granitic rocks and broad folds defined by semicoherent trains of xenoliths, which are intruded by the granitic rocks of the Elizabeth Hills Supersuite (Smithies, 2003).

D_6 deformation at c. 3020 Ma

D_6 deformation is represented by extensional growth faults in rocks of the c. 3020 Ma Gorge Creek Group, which was deposited in a post-collision extensional basin.

D_7 deformation at 3015–3010 Ma

D_7 structures are most strongly developed along major northeast-trending faults in the Gorge Creek Basin in the west Pilbara. Tight-to-isoclinal east-northeasterly trending folds in the Cleaverville Formation between Karratha and Cleaverville are attributed to the D_7 event. The same type of folding is observed in the Cleaverville Formation between Miaree Pool and Mount Regal. In the Regal Formation, S_7 is locally synchronous with intrusive sheets of 3015 Ma quartz–feldspar porphyry and porphyritic microgranite of the Orpheus Supersuite. At Mount Ada (15 km south of Roebourne), easterly trending, upright tight-to-isoclinal D_7 folds in the Cleaverville Formation contain a sill of 3014 Ma granophyre (Nelson, 1997), and these fold structures are unconformably overlain by the c. 3010 Ma Warambie Basalt of the Whim Creek Group.

D_8 deformation at 3000–2960 Ma

D_8 structures are restricted mainly to the Whim Creek and Croydon Basins. Thrusts in the Warambie Basalt east of Mount Ada (Hickman, 2002) pre-date dextral movement on the Sholl Shear Zone, and may be equivalent to 'Phase 3' that Krapez and Eisenlohr (1998) recognized in the Whim Creek area. Originally east–west trending 'D_1 folds' on MOUNT WOHLER (Smithies, 1998) and SATIRIST (Smithies and Farrell, 2000) may belong to the same event. These correlations suggest that D_8 occurred between 2.99 and 2.96 Ga.

D_9 deformation at 2950–2930 Ma

In the west Pilbara, D_9 deformation was most strongly developed in the Mallina Basin (major D_3 folds of Smithies, 1998) and are equivalent to 'Phase 4' structures of Krapez and Eisenlohr (1998), but this event also formed major northeasterly trending tight-to-open folds such as the Prinsep Dome, Roebourne Synform, and Bradley Syncline (Fig. 12). Geochronology in the Mallina Basin (Smithies, 1998) establishes that the age of these structures must be 2.95 – 2.93 Ga, not 2.906 – 2.863 Ga as suggested by Krapez and Eisenlohr (1998). D_9 folds are oblique to the Sholl Shear Zone and to other strike-slip faults of the West Pilbara Superterrane and Central Pilbara Tectonic Zone. They are probably transpressional folds within a post-2.95 Ga, east-northeasterly belt of dextral strike-slip movement. Minor D_9 structures include a steeply dipping, east-northeasterly striking axial-plane foliation (S_9) in the Prinsep Dome and in an anticline east of Mount Sholl. Minor D_9 folds deform West Pilbara Superterrane S_1

southeast from Mount Regal and 1 km southwest from Nickol Well. The Mount Regal folds plunge southwest and the Nickol Well folds plunge northeast.

In the east Pilbara (Fig. 15), D_9 deformation produced the north to northeasterly trending Lalla Rookh – Western Shaw structural corridor, which pre-dated, or was accompanied by, deposition of the coarse clastic Lalla Rookh Sandstone in a trough developed in advance of the north-moving Strelley Monzogranite and associated greenstones (Van Kranendonk and Collins, 1998; Van Kranendonk, 2008). The Lalla Rookh – Western Shaw Structural Corridor is characterized by large northeast-trending folds, north- to northwest-striking sinistral faults and shear zones, and northeast-striking dextral faults that collectively indicate a northwest–southeast direction of maximum compression (Fig. 15). The two strands of the Mulgandinnah Shear Zone formed along the western margin of the Shaw Granitic Complex at this time (Van Kranendonk and Collins, 1998; Zegers et al., 1998) and were responsible for tight folding of the northwestern margin of the Shaw Granitic Complex (Van Kranendonk et al., 2004), the structures of which were thought to be the product of c. 3470 Ma Alpine-style recumbent folding and thrusting (Bickle et al., 1980, 1985; Zegers et al., 2001). The Lalla Rookh – Western Shaw structural corridor represents a significant zone of translation and was responsible for exhumation and tilting of the Strelley Monzogranite and flanking greenstones. However, 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 proposed by Krapez (1993) (Van Kranendonk and Collins, 1998; Van Kranendonk et al., 2004).

The D_9 deformational event also caused amplification of most of the domes, including the Tambourah Dome where it was accompanied by low-grade metamorphism of adjacent greenstones (Wijbrans and McDougall, 1987) and development of a north-northeasterly striking quartz foliation across the Yule and Shaw Granitic Complexes. Synkinematic monzogranite plutonism was widespread across the western part of the East Pilbara Terrane at this time (Van Kranendonk et al., 2002, 2006a), but is unknown east of the Shaw Granitic Complex.

D_{10} deformation at c. 2930 Ma

In the west Pilbara, there is localized deformation along the north-northwesterly striking section of the Maitland Shear Zone where it truncates major northeasterly trending D_9 folds of the Mount Sholl area. A parallel tectonic foliation (S_{10}) is developed in the adjacent greenstones of the Whundo Group and in the rocks of the Cherratta Granitic Complex. Shear zones are also present within the Cherratta Granitic Complex where gneiss, interpreted to include deformed syntectonic granitic veins, contains late zircon populations dated at 2944 ± 5 Ma and 2925 ± 2 Ma (Nelson, 1997).

D_{11} deformation at c. 2920 Ma

In the west Pilbara, the latest movement on the Sholl Shear Zone, and probably on most other major faults in the West Pilbara Superterrane and Mallina Basin, was dextral. A subsidiary dextral strike-slip fault, the Black Hill Shear Zone, displaces the Andover Intrusion by 10 km south of Roebourne. As noted by Krapez and Eisenlohr (1998), zircon geochronology on several rock units close to the Sholl Shear Zone has revealed a metamorphic disturbance event at about 2.92 Ga; this event could have coincided with D_{11}. Minor D_{11} structures in the Sholl Shear Zone include dextral drag folding and isoclinal folding of mylonite lamination, and associated small-scale faulting and brecciation.

D_{12} deformation at 2905–2890 Ma

A major orogenic event occurred in the Mosquito Creek Basin between 2905 and 2890 Ma. Sinistral shearing on north-northeasterly striking zones affected the northwestern part of the East Pilbara Terrane at c. 2890 Ma and was accompanied by gold mineralization in the Mount York and Lynas districts (Neumayr et al., 1993, 1998). Deformation of the East Pilbara Terrane was largely completed by c. 2850 Ma, the time of emplacement of large, undeformed tin–tantalum-bearing granites of the Split Rock Supersuite into the cores of granitic complexes across the terrane (Kinny, 2000; Van Kranendonk et al., 2006a).

Mineralization

In the following discussion of mineralization of the northern Pilbara Craton the primary division is on the basis of two geographically defined regions, the east Pilbara and west Pilbara, rather than the geologically defined East Pilbara Terrane and West Pilbara Superterrane.

East Pilbara region

The east Pilbara region is host to a variety of mineralization formed in a range of tectonic environments. Gold mineralization in the De Grey Supergroup is restricted to shear-zone hosted c. 2905 Ma gold deposits in the Mosquito Creek Basin and to banded iron-formation in the c. 3020 Ma Gorge Creek Basin. Recent reviews of mineralization in the east Pilbara (Ferguson and Ruddock, 2001; Huston et al., 2001, 2002) summarize the styles and ages of mineralization, and provide references to previous investigations focused on particular mineral commodities. A more detailed regional review is beyond the scope of this field guide, but it is important to relate the mineralization of the east Pilbara to the current interpretation of its geological history. Figure 4 identifies the main types of mineralization formed during the evolution of the northern Pilbara Craton. The following summary expands on this with reference to tectonic units, suites and supersuites, deformation events, and inferred controls of mineralization in the East Pilbara Terrane and De Grey Supergroup. Figure 5 shows the six stages of evolution of the Pilbara Craton to which mineralization is related in the following discussion.

Mineralization of the East Pilbara Terrane commenced with the onset of voluminous volcanic activity during

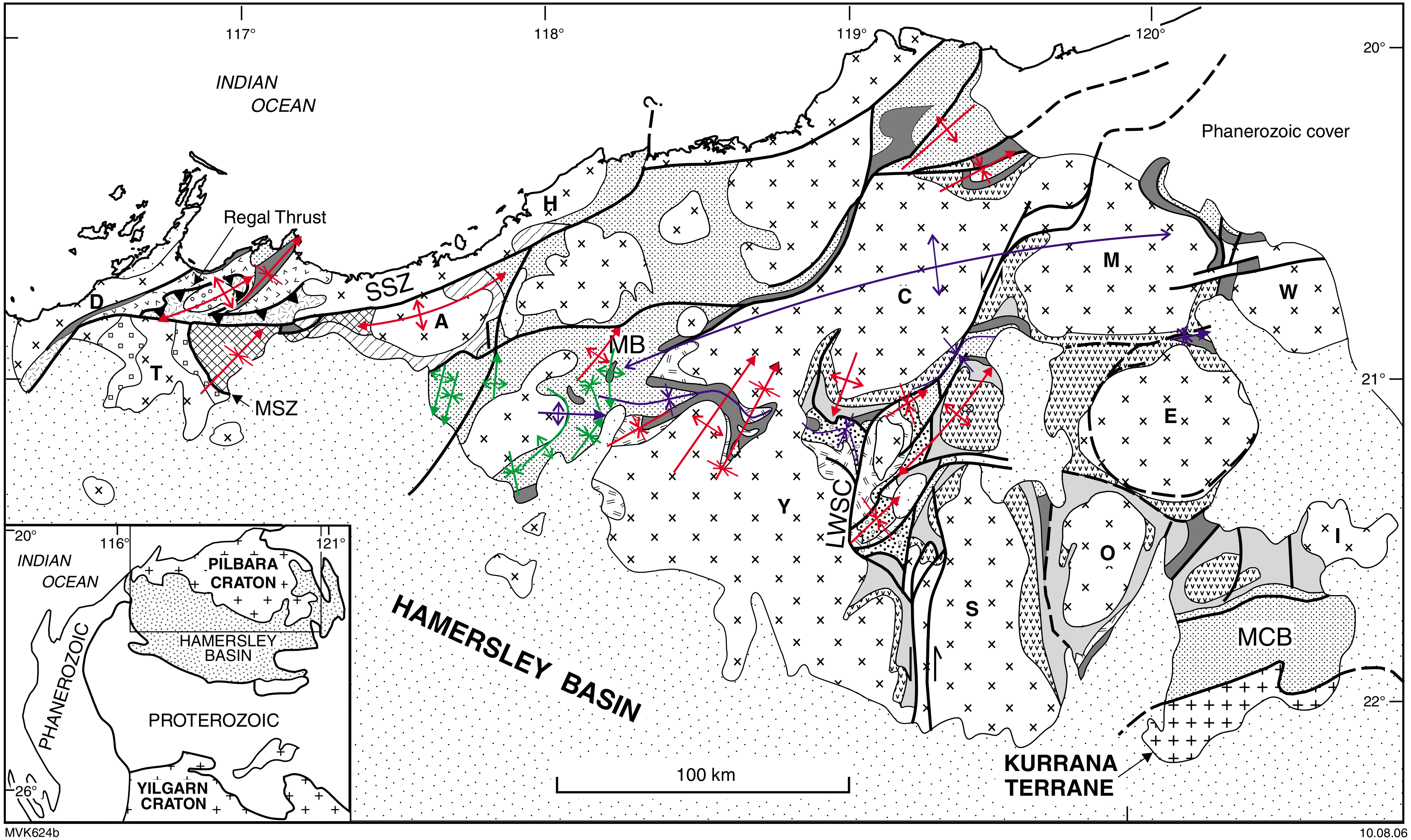

Figure 15. Main structures associated with late regional compression (2970–2930 Ma) across the northern Pilbara Craton. MB = Mallina Basin; MCB = Mosquito Creek Basin; LWSC = Lalla Rookh–Western Shaw structural corridor; MSZ = Maitland Shear Zone; SSZ = Sholl Shear Zone. Single bold letters refer to granitic complexes: A = Caines Well; D = Dampier; E = Mount Edgar; H = Harding; I = Yilgalong; M = Muccan; O = Corunna Downs; S = Shaw; T = Cheeratta; W = Warrawagine; Y = Yule (from Van Kranendonk et al., 2007)

Stage 2 of the tectonic evolution of the Pilbara Craton (Fig. 5; Van Kranendonk et al. 2006a,b); Stage 1 pre-dates the preserved geological record.

Rocks associated with Stage 3 (rifting event) are not known to contain economically significant mineralization, but future exploration of the c. 3.18 Ga Dalton Suite may reveal nickel, copper, and PGE mineralization. Stage 4 mineralization is restricted to the West Pilbara Superterrane, as discussed by Hickman et al. (2006), but significant mineralization is associated with Stages 5 and 6 events in the East Pilbara Terrane and the Mosquito Creek Basin. Mineralization also accompanied depositional and tectonic events associated with the Fortescue Group.

Stage 2: Warrawoona Group, and Callina and Tambina Supersuites

Rocks of the Warrawoona Group are host to a variety of synvolcanic mineralization styles including:

- sediment-hosted, hydrothermal massive sulfates in the Dresser Formation;
- volcanic-hosted massive sulfides (VHMS) in the Duffer Formation;
- vein and hydrothermal base metals in the Panorama Formation;
- precious metals.

Sediment-hosted, hydrothermal massive sulfates in the Dresser Formation

The oldest mineralization of the Pilbara Craton includes several million tonnes of barite, with minor amounts of sphalerite and galena, in rocks of the Dresser Formation in the Barite Range on the eastern flank of the North Pole Dome (Hickman, 1973, 1983). Between 1970 and 1990, 129 505 t of barite was mined at the Dresser mining centre (MGA Zone 50, 752790E 7659155N), which included leases with total published reserves of 500 000 t (Sargeant and Sampson, 1980). A review of the deposits and their origin was presented by Abeysinghe and Fetherston (1997).

The main barite deposits outcrop in an 8 km-long zone within and immediately underlying the Dresser Formation. Mineralization takes the form of either stratabound layers and mounds or veins of coarsely crystalline barite within hydrothermal chert–barite veins. The veins intrude highly altered metabasalt immediately underlying chert of the Dresser Formation. Barite is also present as pebbles in basal diamictite layers, as finely laminated rock containing stromatoloid laminae (Buick et al., 1981), and as coarsely crystalline mounds adjacent to synsedimentary growth faults (Nijman et al., 1998a). The main deposits are where the Dresser Formation is thickest and they consist of three stratigraphic horizons of chert–barite interlayered with metabasalt. The three mineralized beds are bound by an array of listric normal growth faults, the main set of which passes just south of and through the Dresser mining centre. The geometry of the faults and their close relation to barite mineralization (Nijman et al., 1998a) indicates that the barite was deposited during normal faulting and development of a volcanic caldera associated with felsic volcanism (Nijman and de Vries, 2004; Van Kranendonk, 2006). Galena and sphalerite are finely disseminated through the system, suggesting the possibility of massive sulfides at depth. Sulfide mineralization was explored for in the underlying rocks, but to no avail (Sargeant, 1979).

Hickman (1973) ascribed a sedimentary origin to the North Pole barite deposits and presented several points against a hydrothermal origin. The presence of barite in chert–barite veins was interpreted to be the result of tectonic reactivation of sedimentary chert–barite during structural doming. Dunlop et al. (1978) and Sargeant (1979) suggested that the barite deposits formed by diagenetic replacement of an original gypsum–carbonate sequence; gypsum pseudomorphs in barite were reported by Buick and Dunlop (1990). Sargeant (1979) inferred that the deposits formed by exhalative chemical precipitation. In this model, the chert–barite veins were interpreted to represent the footwall vein stockwork system of bedded chert–barite rocks of the Dresser Formation, a conclusion supported by subsequent studies (Nijman et al., 1998a; Van Kranendonk, 2006) that documented variations in both lithofacies type and thicknesses of barite and chert adjacent to growth faults and showed contemporaneous development. The most recent interpretation of the environment of deposition was a shallow and probably restricted marine setting, with precipitation of barite and diagenetic replacement of carbonate rocks by barite during low-temperature hydrothermal volcanic emissions (Van Kranendonk, 2006).

Volcanic-hosted massive sulfides in the Duffer Formation

VHMS copper–zinc mineralization is associated with rocks of the 3470 Ma Duffer Formation at Marble Bar – Big Stubby, and at Lennons Find, both of which are around the southern flank of the Mount Edgar Granitic Complex. At Marble Bar – Big Stubby, mineralization is interpreted to be associated with intrusion of the subvolcanic 3466 Ma Homeward Bound Granite (Locality 4).

Vein and hydrothermal base metal occurrences in the Panorama Formation

Lead–zinc mineralization at Quartz Circle in the McPhee greenstone belt (McPhee Dome) and at the Breens and Miralga Creek localities in the Panorama greenstone belt (North Pole Dome) is associated with felsic volcanic rocks and related porphyry intrusions of the Panorama Formation. At Quartz Circle, both massive and vein-type base metal mineralization appear to be present, possibly with epithermal affiliations (Ferguson and Ruddock, 2001).

At Miralga Creek, gold, zinc, lead, and copper mineralization is associated with a c. 3450 Ma (Thorpe et al., 1992a) felsic porphyry stock emplaced into rocks of the Warrawoona Group in the eastern part of the North Pole Dome (Goellnicht et al., 1988). There is disseminated, stringer, vein, and hydrothermal breccia-hosted mineralization around the marginal intrusive breccia of the stock and in association with porphyry dykes. Van Kranendonk (1999, 2000) related the mineralization to emplacement of the c. 3458 Ma North Pole Monzogranite in the core of the North Pole Dome. Galena at Miralga Creek gave a range of Pb-model ages between 3447 and 3451 Ma (Groves, 1987).

The Breens Copper deposit in the central western part of the North Pole Dome is a complex northeasterly striking belt of stratabound massive sulfide mineralization (pyrite, chalcopyrite, chalcocite, covellite, and neodiginite) and native copper, 500 m wide, and consisting of stockworks and silicified breccia zones (Ferguson and Ruddock, 2001) in mafic to felsic volcanic rocks of the Dresser Formation (Van Kranendonk, 1999). Mineralization is associated with a small dyke of felsic porphyry, interpreted to be related to the North Pole Monzogranite (Van Kranendonk, 1999).

Precious metals

A few gold and silver deposits in the East Pilbara Terrane contain galena that yield Pb–Pb model ages of mineralization at 3430–3400 Ma (Thorpe et al., 1992b; Huston et al., 2002; Zegers et al., 2002). However, some of these deposits lie within younger host rocks of the c. 3350 Ma Euro Basalt (e.g. Bamboo mining centre; Williams, 1998), suggesting that the dated Pb, in at least some of the deposits, is derived from old crustal sources remobilized during later tectonothermal events.

Stage 2: Kelly Group and Emu Pool Supersuite

Voluminous intrusion of Emu Pool Supersuite granitic rocks at the end of eruption of the Kelly Group caused widespread metamorphism and hydrothermal alteration of greenstones and the development of several types of mineralization. Precipitation of gold in shear zones was the most significant mineralization during this period, with significant deposits distributed around the Mount Edgar Granitic Complex, in particular within the Warrawoona Syncline, and at McPhee Reward (Ferguson and Ruddock, 2001; Huston et al., 2002). Lead and silver were also precipitated in shear zones along the southern flank of the Muccan Granitic Complex at Doolena Gap. A porphyry copper–molybdenum system is developed at Spinifex Ridge, at the northern tip of the Mount Edgar Granitic Complex, associated with intrusion of a granodiorite (Locality 16). Base metals were deposited at the Copper Hills and Kellys mining centres on the eastern margin of the Corunna Downs Granitic Complex, on either side of the 3315 ± 4 Ma (Barley and Pickard, 1999) Boobina Porphyry, with which they are genetically related.

Stage 2: Sulphur Springs Group and Cleland Supersuite

Widespread intrusion of the Cleland Supersuite at the end of eruption of the Sulphur Springs Group produced significant mineralization. This included syngenetic VHMS (zinc–copper) mineralization at Sulphur Springs, and other prospects along strike, associated with emplacement of the Strelley Monzogranite (Brauhart et al., 1998; Brauhart, 1999; Vearncombe et al., 1995, 1998; Van Kranendonk, 2000, 2006). Minor molybdenum, tin, and gold mineralization is also associated with emplacement of this intrusion. Significant volumes of barite (c. 64 000 t) are present in veins near Cooke Bluff Hill, hosted by rocks of the Sulphur Springs Group and underlying Kelly Group (Hickman, 1977b; Van Kranendonk, 2004).

Stage 3

Minor copper, nickel, and PGE mineralization is associated with the Dalton Suite of layered ultramafic and mafic intrusions that were emplaced into the Soanesville Group in the Soanesville greenstone belt. One of these sills, which are interpreted to be associated with rifting of the East Pilbara Terrane between c. 3200 and 3160 Ma, was dated by the U–Pb zircon method at 3182 ± 2 Ma (Wingate, in prep.a, sample 178185). Chromite pods, also possibly associated with serpentinized metadunite of the Dalton Suite, are present along the northeastern contact between the Yule Granitic Complex and the Pincunah greenstone belt. Farther west, the Flat Rocks lead mine is hosted by rocks of the Mount Billroth Supersuite and is thus considered to be related to the rifting event. Gold mineralization at Lalla Rookh is associated with galena that returned a Pb–Pb model age of 3188 Ma (Thorpe et al., 1992b).

Stage 5

Intracontinental extension and basin development during deposition of the c. 3020 Ma Gorge Creek Group was accompanied by deposition of widespread banded iron-formation (Cleaverville Formation) across much of the east Pilbara region. This continues to be mined at Yarrie, on the northeastern flank of the Muccan Granitic Complex (Williams, 1999; Van Kranendonk et al., 2001b). Iron ore is mined both from hard rock (c. 65% Fe) and from overlying bedded hematite conglomerate of the Neoproterozoic Eel Creek Formation (c. 61% Fe; Williams, 1999). Iron ore of the Cleaverville Formation was also mined at Goldsworthy, and has been extensively prospected for in the Ord Ranges, where both supergene enrichment and magnetite deposits are currently being evaluated as possible economic sources of iron ore (Atlas Iron Limited, 2008a,b). Prior to recent iron ore exploration in the Ord Ranges, the area was intermittently worked for tiger-eye iron formation (silicified reibeckite, used as building stone and gemstone). There are shear-zone hosted gold deposits within, and immediately adjacent to the Lalla Rookh – Western Shaw structural corridor. Alluvial gold has been mined from the rocks of the Lalla Rookh Sandstone at the Keep It Dark mine, and emeralds have been mined from pegmatite emplaced within a shear zone on the northwestern margin of the Shaw Granitic Complex.

Stage 6

Copper–lead–zinc–(silver–gold) mineralization is present at Coondamar Creek in the basal parts of the Nullagine Group (De Grey Supergroup) in the Mosquito Creek Basin. Primary mineralization consists of short intervals of pyrite, chalcopyrite, sphalerite, and galena in schistose metavolcanic host rocks (Ferguson and Ruddock, 2001). Shear-zone hosted epigenetic gold is widespread throughout the Mosquito Creek Basin (Bagas et al., 2004) and has been dated at 2905 Ma (Huston et al., 2002). There is shear-zone hosted gold of similar age (2888 ± 6 Ma; Neumayer et al., 1998) in the Mount York area in the northwestern part of the East Pilbara Terrane. Mining from 1994 to 1998 by Lynas Gold produced a total of 3822.9 kg of gold from a series of small deposits in this area (Ferguson and Ruddock, 2001).

Tin–tantalum–lithium has been mined from pegmatite veins along the margins of post-tectonic monzogranites of the c. 2850 Ma Split Rock Supersuite. Three types of tin–tantalum–lithium–beryllium pegmatites have been defined (Hickman, 1983): (1) simple pegmatite with cassiterite and tantalite–columbite minerals and rarer lithium and beryllium compounds; (2) layered albite pegmatites with a wide variety of mineral species and with cassiterite concentrated at finer albitic margins; and (3) complex rare earth pegmatites, which mainly intrude greenstones (Ferguson and Ruddock, 2001). Mining was particularly widespread in the Shaw Granitic Complex and recovered tin and tantalum from predominantly very shallow (<2.5 m) alluvial workings in recent drainage basins (Van Kranendonk, 2002). Mining went underground in the Moolyella Monzogranite in the core of the Mount Edgar Granitic Complex, where there is cassiterite at the margins of layered albite pegmatite veins emplaced along the western side of the intrusion. Pegmatite-hosted tantalite is currently being mined from the Wodgina area by Sons of Gwalia, from opencut workings in shallow-dipping, thick pegmatite sheets (Ferguson and Ruddock, 2001; Sweetapple et al., 2001).

Younger mineralization

There are several styles of mineralization hosted in the Fortescue Group or younger rocks. These include shear-zone hosted gold mineralization (e.g. Blue Bar mine in the Coongan greenstone belt; Comet mine just south of Marble Bar) and alluvial gold workings in conglomerates in the lower part of the Fortescue Group. Alluvial diamonds have been reported from around the town of Nullagine in Cenozoic alluvium and in conglomerate of the Hardey Formation of the Fortescue Group. Diamonds have been found in situ in samples of kimberlite from the Brockman Kimberlite Dyke that cuts across the Corunna Downs and Mount Edgar Granitic Complexes and the intervening Warrawoona Syncline.

West Pilbara region

All the terranes and basins of the west Pilbara region are mineralized but, because these units formed in a range of tectonic environments, the styles of mineralization and the processes responsible for them differ. Recent reviews of mineralization in the west Pilbara (Ruddock, 1999; Huston et al., 2001, 2002) summarize the styles and ages of mineralization and provide references to previous investigations focused on particular mineral commodities. A more detailed regional review is beyond the scope of this field guide, but it is important to relate the mineralization of the west Pilbara to the current interpretation of its geological history. Figure 4 identifies the main types of mineralization during the evolution of the terranes and basins. The following summary expands on this with reference to tectonic units, suites and supersuites, deformation events, and inferred controls of mineralization. Only those units and events known to be associated with significant mineralization are discussed.

Karratha Terrane

The Ruth Well nickel–copper deposits (Tomich, 1974; Marston, 1984) are hosted by serpentinized komatiite in the >3.27 Ga Ruth Well Formation. In Western Australia, komatiite-associated nickel sulfide deposits from the Eastern Goldfields Superterrane of the Yilgarn Craton are the best documented examples, but these deposits are approximately 550 m.y. younger than the Ruth Well mineralization. Ruddock (1999) questioned the classification of the Ruth Well deposits as volcanic on the basis of textural evidence that the host unit might be intrusive and belong to the >3016 Ma Andover Intrusion (Orpheus Supersuite), which outcrops extensively to the east of Ruth Well. However, no economically significant nickel–copper mineralization has been found in the main section of the Andover Intrusion south of Roebourne.

The Nickol River Formation contains copper–zinc and gold mineralization at several localities around the Karratha Granodiorite, and in the Roebourne – Carlow Castle area. The copper–zinc deposits are hosted mainly by sheared felsic volcaniclastic units, whereas gold mineralization is fault controlled, and interpreted to be related to the Regal Thrust (see **Regal Terrane** below).

Sholl Terrane

Mineralization in the Sholl Terrane includes c. 3.12 Ga VHMS copper–zinc deposits in the Tozer and Nallana Formations of the Whundo Group, and epigenetic gold and copper in shears adjacent to the Maitland Shear Zone south of Whundo. Copper–zinc mineralization is also present in the Whundo Group at the Orpheus prospect (MGA Zone 50, 499800E 7687500N), approximately 20 km north-northeast of Whundo. Although the Orpheus mineralization is focused along northeast-striking D_6 faults, and also contains significant amounts of lead, silver, and gold (Hickman, 2001a) suggesting partly epigenetic mineralization, the base metal content suggests derivation from older synvolcanic mineralization in the Whundo Group. The Whundo Group outcrops over approximately 500 km^2 of the Sholl Terrane and has not been intensively explored. Additionally, the group is concealed beneath shallow Hamersley Basin cover over an area of approximately 300 km^2 on Cooya Pooya, Roebourne, and Pinderi Hills (Hickman, 2004b).

The depositional environment of the Whundo Group is interpreted to be a subduction-related intra-oceanic magmatic arc (Smithies et al., 2005a).

Regal Terrane

Known mineralization associated with the Regal Terrane is limited to shear-hosted gold immediately below, within, and immediately above, the Regal Thrust. Although the basal section of the Regal Formation contains locally thick komatiite units, no nickel–copper mineralization similar to that in the Ruth Well Formation (see above) has yet been reported.

Gold mineralization in the West Pilbara Superterrane correlates strongly with the position of the Regal Thrust (e.g. at mining areas such as Lower Nickol, Weerianna, Carlow Castle, and Sing Well) (Hickman, 2002). The thrust includes tectonic slivers of ultramafic and mafic rock derived mainly from the overlying Regal Formation, as well as sheared and brecciated sedimentary rocks from the underlying Nickol River Formation. In many areas the zone of thrusting is veined by hydrothermal quartz or intruded by veins and sills of granite. The timing of gold mineralization spatially associated with the thrust has not been established by direct dating, but is expected to range between 3.16 and 2.94 Ga, reflecting multiple episodes of deformation along the thrust zone (e.g. folding of the Regal Thrust).

Orpheus Supersuite

Felsic intrusive rocks of the Orpheus Supersuite locally contain minor copper mineralization, whereas gabbro of the Andover Intrusion includes deposits of magnetite, and vanadium–titanium, although none of these deposits have been mined for these commodities. Exploration of the Andover Intrusion for nickel–copper and PGE mineralization has failed to reveal any economic deposits.

Gorge Creek Basin

The Cleaverville Formation in the Gorge Creek Basin contains thick units of banded iron-formation that are locally enriched in iron. Exploration for economically mineable deposits has been centred on the area northwest of Roebourne.

Central Pilbara Tectonic Zone

The Central Pilbara Tectonic Zone is the most richly mineralized part of the Pilbara Craton. Deposits within this zone include VHMS deposits (e.g. Whim Creek, Salt Creek, and Turner River), epigenetic base metal deposits (e.g. Mons Cupri and Comstock), layered mafic intrusion-hosted vanadium–titanium deposits (e.g. Andover and Balla Balla), layered mafic–ultramafic intrusion-hosted nickel–copper–PGE deposits (e.g. Radio Hill, Munni Munni, Mount Sholl, and Three Kings), and numerous lode-gold and gold–antimony deposits related to hydrothermal alteration in shear zones (e.g. deposits along the Mallina and Tabba Tabba Shear Zones). Most of these deposits are hosted by stratigraphic units of either the Croydon Group (Mallina Basin) or the Sisters Supersuite.

Mallina Basin

The Whim Creek greenstone belt is one of the most mineralized areas of the west Pilbara. Although the Whim Creek Basin (Warambie Basalt and Red Hill Volcanics of the Whim Creek Group) forms about 30% of the outcrop in this greenstone belt, it hosts none of the significant mineral deposits. Most mineralization is concentrated in the lower part of the Mallina Basin, and includes lead–zinc and copper mineralization in sedimentary rocks of the Cistern Formation at Mons Cupri and Salt Creek, and stratabound copper–zinc mineralization in the shale facies of the Rushall Slate at Whim Creek. Mafic volcanic rocks of the Bookingarra Formation contain small hydrothermal vein deposits of copper, lead, zinc, and antimony. The Mallina Formation hosts stratabound copper–zinc at Egina, vein- and shear-hosted gold and gold–antimony mineralization in the Mallina district, and high-level epithermal gold mineralization.

Sisters Supersuite (Radley, Langenbeck, and Indee Suites)

One of the most significant mineralizing events in the west Pilbara was the intrusion of large layered mafic–ultramafic intrusions of the Radley Suite at c. 2.93 Ga. These intrusions host significant nickel–copper–PGE deposits at Radio Hill, Munni Munni, and Mount Sholl, and vanadium–titanium deposits at Balla Balla. In addition, nickel–copper mineralization is hosted in banded quartz–magnetite–amphibole schist adjacent to the Sherlock Intrusion. The mafic–ultramafic sills of the Langenbeck Suite host nickel–copper–PGE and orthomagmatic gold deposits, commonly in close proximity to high-Mg diorite (sanukitoid) rocks of the Indee Suite.

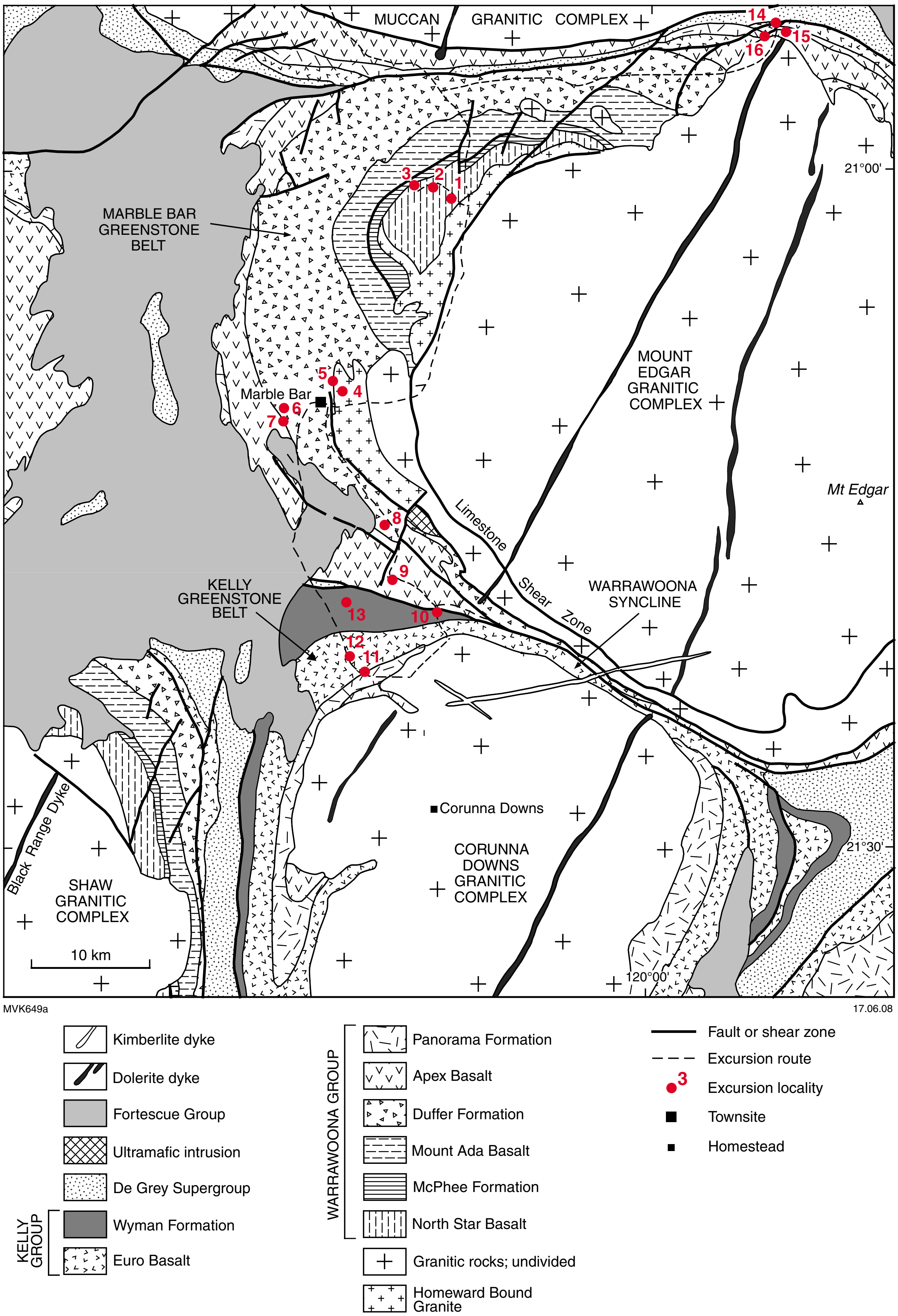

Figure 16. Geological map of the Marble Bar area showing excursion Localities 1–16 (modified from Van Kranendonk et al., 2001b)

Part 2

Excursion localities

Geological sites included in this six-day excursion were selected to provide participants with a brief overview of the crustal evolution of the northern Pilbara Craton. The localities visited are described in sufficient detail to permit later use of the guide by anyone wishing to visit particular sites independently; however, it should be noted that access to many of the sites requires a 4WD vehicle, and prior approval from landowners may be necessary. The excursion starts at Port Hedland and ends six days later at Karratha. Excursion localities are shown on Plates 1 and 2 (Localities 1–23), and Figure 12 (Localities 24–30).

Day 1: Investigation of part of the 3525–3426 Ma Warrawoona Group, including a granite–porphyry felsic volcanic system in the Marble Bar greenstone belt

The Marble Bar greenstone belt surrounds the Mount Edgar Granitic Complex. In the well-exposed, low-strain western part of the belt near Marble Bar, the succession contains up to 12 km of mafic and felsic volcanic rocks, chert, and subordinate metasedimentary rock of the Warrawoona Group, including the type sections for the Talga Talga, Coongan, and Salgash Subgroups (Fig. 16). All volcanic and sedimentary strata of the belt face consistently away from, and young up section from, the granitic complex. Felsic volcanic components were fed by subvolcanic intrusions emplaced within the flanking Mount Edgar Granitic Complex. At the base of the succession, closest to the centre of the dome, strata dip about 20° to the northwest. However, farther northwest towards the centre of the greenstone syncline, dips progressively increase until near the top of the Duffer Formation, at which point strata pass through the vertical and become overturned (dipping steeply southeast, but still younging to the west).

Previous models suggest that the Marble Bar greenstone belt contains a coherent stratigraphic section tilted on edge by diapiric doming of the Mount Edgar Granitic Complex (Hickman, 1983, 1984; Collins, 1989). However, the total thickness of the Warrawoona Group, combined with the presence of local bedding-parallel shears along major lithological contacts in this area, led van Haaften and White (1998) to propose that regional northeasterly directed thrusting produced a tectonostratigraphic assemblage of inverted greenstones deformed in a thrust-stack culmination into which granitic rocks of the Mount Edgar Granitic Complex were later emplaced. Van Kranendonk et al. (2001a) rejected the structural arguments behind the thrust culmination model, although these were rebutted by van Haaften and White (2001). More recently, Van Kranendonk et al. (2002, 2004) identified radial mineral elongation lineations around the Mount Edgar and adjacent granitic complexes (Fig. 14) that are consistent with partial convective overturn models, and summarized new geochronological data that shows that the greenstone succession youngs consistently up section (Fig. 17), thereby supporting the view that the Marble Bar greenstone belt is a fitting type area for much of the Warrawoona Group.

Locality 1*: Volcanic cycles in the c. 3490 Ma North Star Basalt (Marble Bar, MGA Zone 50, 795650E 7673000N)

From Port Hedland, drive east on Highway 1 for approximately 50 km to the Marble Bar turnoff. Follow the Marble Bar road (Highway 138) for approximately 127 km to the McPhee Reward track (MGA Zone 50, 797100E 7673000N), turn right, and continue westwards for 1.6 km to a point 100 m west of a section where the track follows a rocky creek bed. Locality 1 is 200 m south of the track, and can be accessed by walking or 4WD vehicle (MGA Zone 50, 795650E 7673000N).

The track from Highway 138 to the abandoned McPhee Reward gold mine traverses the North Star Basalt, the oldest formation in the Talga Talga Subgroup. Recent mapping (Hickman and Van Kranendonk, 2008) identified three ultramafic–mafic cycles (members) in the formation and older mafic rocks intruded by granitic rocks east of the highway that are remnants of an underlying basaltic section of the formation. West of the highway, the upper two members are capped by chert. A metamorphic date (Ar–Ar) of c. 3490 Ma was obtained by van Koolwijk et al. (2001) close to this locality; c. 3500 Ma metamorphic dates were obtained on the North Star Basalt in other greenstone belts (see **Warrawoona Group** in **Part 1**). The age of the formation is therefore at least c. 3500 Ma, although it may span >20 m.y. A silicified fine-grained felsic volcaniclastic rock in the overlying McPhee Formation was dated at 3477 ± 2 Ma (Nelson, 2000a).

* Please note that access to many localities visited during this excursion involves crossing pastoral leases and mining tenements, and the use of private tracks. Accordingly, anyone intending to visit these localities at a future date should first obtain approval from the landowners and tenement holders involved.

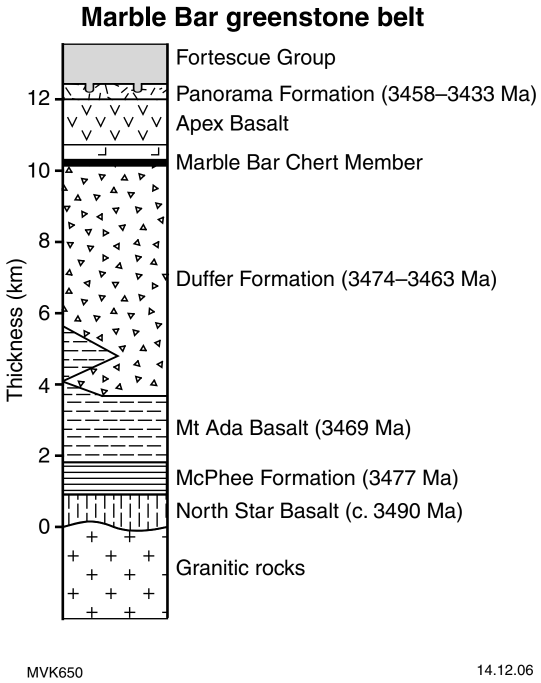

Figure 17. Simplified stratigraphic section of the Marble Bar greenstone belt showing geochronological data and unit thicknesses (modified from Van Kranendonk et al., 2001b)

Geochemical traverses undertaken by Glikson and Hickman (1981; Figs 18, 19) revealed much of the North Star Basalt to be tholeiitic, but sampling was concentrated on least-altered volcanic rocks and therefore biased towards tholeiites rather than the komatiite and komatiitic basalt in the lower parts of each cycles. The basaltic rocks of this area were reported by Glikson and Hickman (1981) to have relatively high contents of Zr, Nb, Y, La, and Ce, suggesting crustal contamination. This interpretation agrees with a conclusion reached 20 years later by Green et al. (2000) that some basalts in the lower part of the Warrawoona Group have compositions indicating up to 25% contamination by felsic crust. Smithies et al. (2007) noted that samples they analysed from the North Star Basalt were high-titanium basalts indistinguishable from those of tholeiitic basalt in the c. 3515–3498 Ma Coonterunah Subgroup. The lower part of the formation is locally intruded by c. 3465 Ma granitic rocks of the Callina Supersuite, and a dyke of porphyritic rhyolite intrudes the North Star Basalt farther west along the track. The metamorphic grade of the North Star Basalt in this area is lower amphibolite to upper greenschist facies.

The view to the north of Locality 1 provides good examples of dip slopes, with basaltic flows inclined at about 20° to the northwest. Westwards, towards the margin of the Mount Edgar Dome, dips progressively increase until they become vertical or slightly overturned. Carbonated serpentinite at Locality 1 is interpreted to be part of a layered sill because it is immediately overlain by metapyroxenite and metadolerite. The composition of the serpentinite includes 40.30% SiO_2, 33.58% MgO, 2.31% H_2O, 3.75% CO_2, 4280 ppm Cr, and 2214 ppm Ni. Some parts of the rock preserve relict olivine-cumulate texture.

Locality 2: Top of a volcanic cycle in the North Star Basalt (Marble Bar, MGA Zone 50, 794200E 7674100N)

The oldest identified chert unit in the North Star Basalt of the Marble Bar greenstone belt outcrops at Locality 2. This marks the top of a volcanic cycle, 500 m above the stratigraphic level of Locality 1. A unit of grey-and-white layered chert and pyritic iron formation is overlain by peridotite and komatiitic basalt at the base of the next cycle, and underlain by variably silicified and carbonated basaltic rocks. The overlying cycle, which is about 400 m thick and predominantly composed of komatiitic basalt, is intruded by sills of gabbro and dolerite and overlain by silicified felsic tuff, grey-and-white layered chert, and carbonated volcaniclastic rocks assigned to the lower part of the c. 3477 Ma McPhee Formation. Along strike to the southwest, the chert at Locality 2 is immediately underlain by extensively silicified basaltic rocks. Alteration was probably hydrothermal during a break in volcanism when the underlying volcanic pile cooled.

Locality 3: TalgaTalga Subgroup and shear-hosted gold mineralization (3420–3310 Ma) at McPhee Reward gold mine (Marble Bar, MGA Zone 50, 793100E 7674150N)

Continue westward on the McPhee Reward mine track for 1.5 km as far as the entrance to a gorge.

The type section of the McPhee Formation (Hickman, 1977a) is exposed in the gorge at Locality 3 (Fig. 20). Entering the gorge from the southeast, the basal member of the formation is a 25 m-thick grey-and-white banded chert. The chert stratigraphically overlies silicified komatiitic basalt at the top of the North Star Basalt. The komatiitic basalt shows relict clinopyroxene spinifex texture, and is intruded by veins of chert immediately beneath the banded chert. Approximately 6 km to the southwest, a silicified fine-grained tuff that locally overlies the komatiitic basalt is assigned to the base of the McPhee Formation, and was dated at 3477 ± 2 Ma (Nelson, 2000a). The McPhee Formation is stratigraphically overlain by the Mount Ada Basalt and, about 2 km north of the gorge, the lower part of this formation contains a thin felsic tuff dated at 3469 ± 3 Ma (Nelson, 1999b).

The grey-and-white banded chert at this locality is stratigraphically overlain by a 50 m-thick succession of carbonate–chlorite–quartz schist, ferruginous chert, and altered metabasalt. High nickel (about 1000 ppm) and chromium (about 2000 ppm) contents in the carbonate–chlorite–quartz schist indicate that it is a deformed ultramafic rock, and interlayering with thin beds of ferruginous chert suggests that it originated as lava or tuff. This is supported by the regional extent of the carbonate–chlorite–quartz schist at this stratigraphic

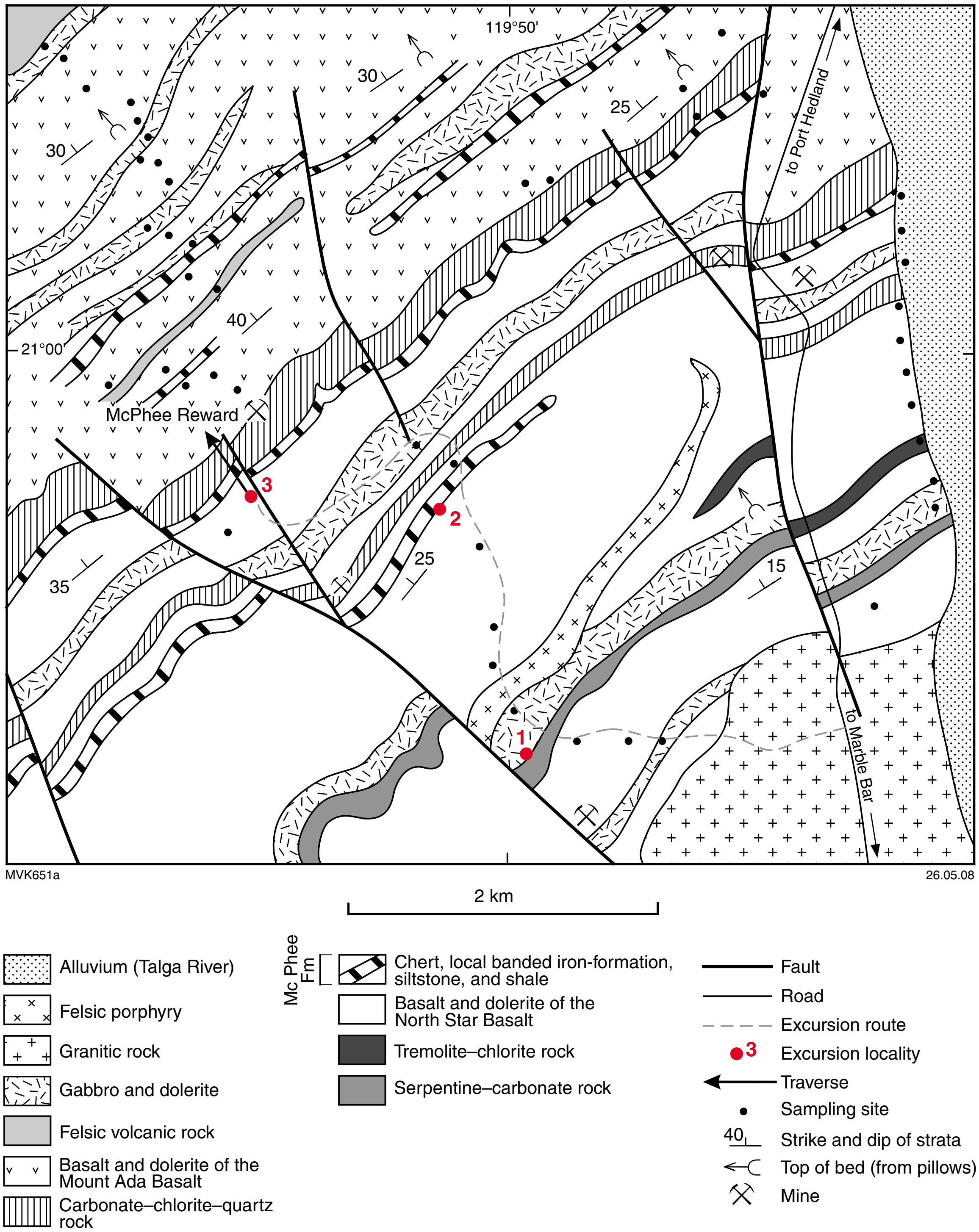

Figure 18. Geological map of the Talga Talga Subgroup in the McPhee Reward area showing excursion Localities 1–3 and geochemical sampling sites for data shown in Figure 19 (after Hickman, 1980)

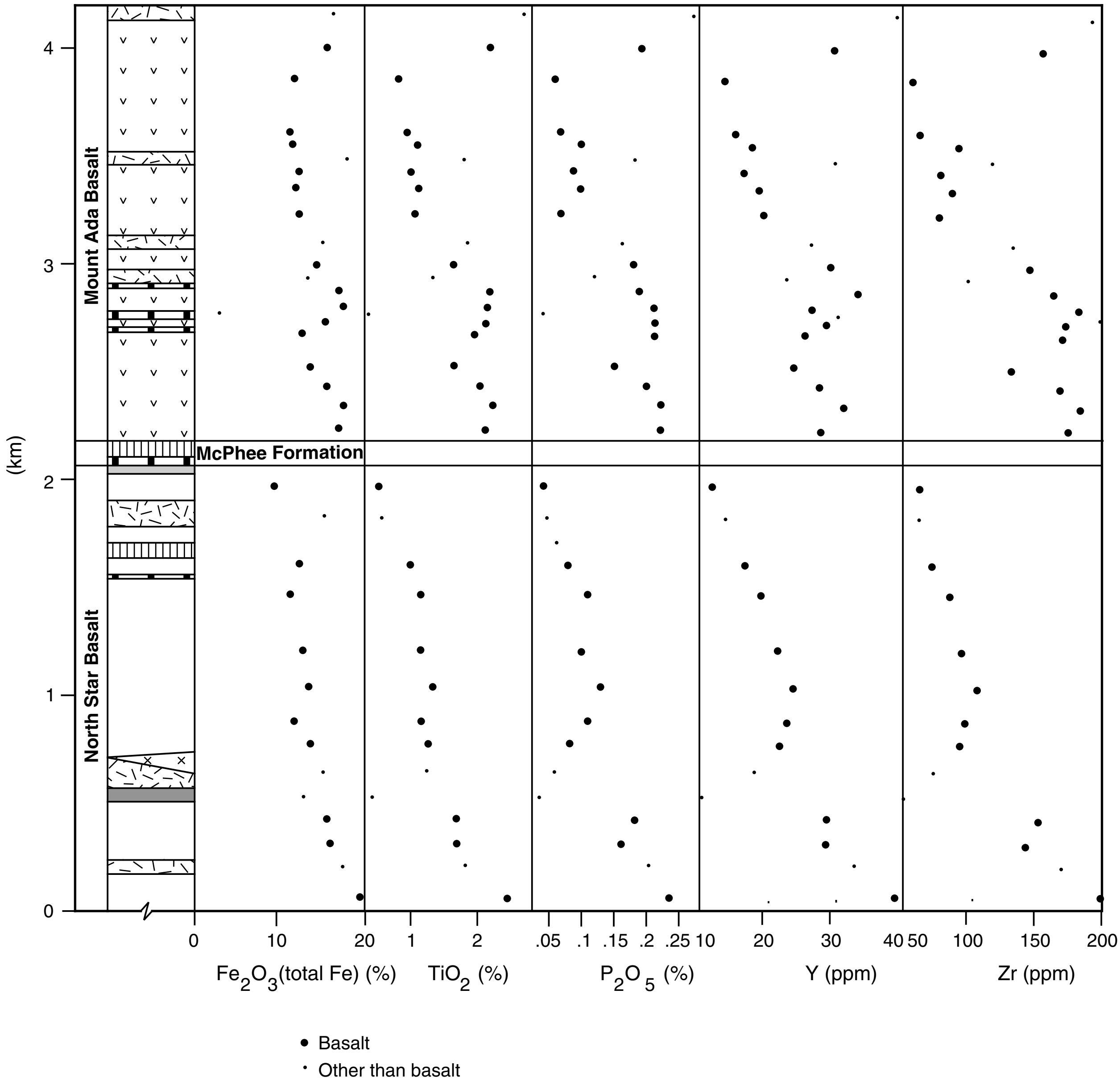

Figure 19. Geochemical trends in the lower Warrawoona Group in the McPhee Reward section (after Hickman, 1980)

position, not only within the Marble Bar greenstone belt, but also 50 km to the southwest in the Coongan greenstone belt. This stratigraphic correlation to the succession of the Shark Gully area of the Coongan greenstone belt has been supported by SHRIMP U–Pb zircon geochronology. Nelson (2000b) dated a granophyre sill in the Mount Ada Basalt of the Coongan belt at 3469 ± 3 Ma, and the overlying Duffer Formation at 3474 ± 7 Ma. Three other samples of the Duffer Formation from the Shark Gully succession have been dated at between 3470 and 3467 Ma. Exposures in the floor of the gorge reveal a northwesterly striking fault, which explains why the detailed stratigraphy does not match on both sides of the gorge.

In a disputed structural interpretation, van Haaften and White (1998) claimed that the McPhee Formation is not a stratigraphic unit, but represents a mylonitic shear zone between the Mount Ada and North Star Basalts. They interpreted this shear zone to mark a major thrust that involved tectonic stacking of material from the west. The thrust was interpreted to have reactivated an earlier syndepositional normal fault. Most prominent lineations in the sheared rocks of the McPhee Formation plunge west-northwesterly in the direction of dip and, in combination with other kinematic indicators, were reported to indicate west-side-up movement (van Haaften and White, 1998). Thus, van Haaften and White (1998) suggested that

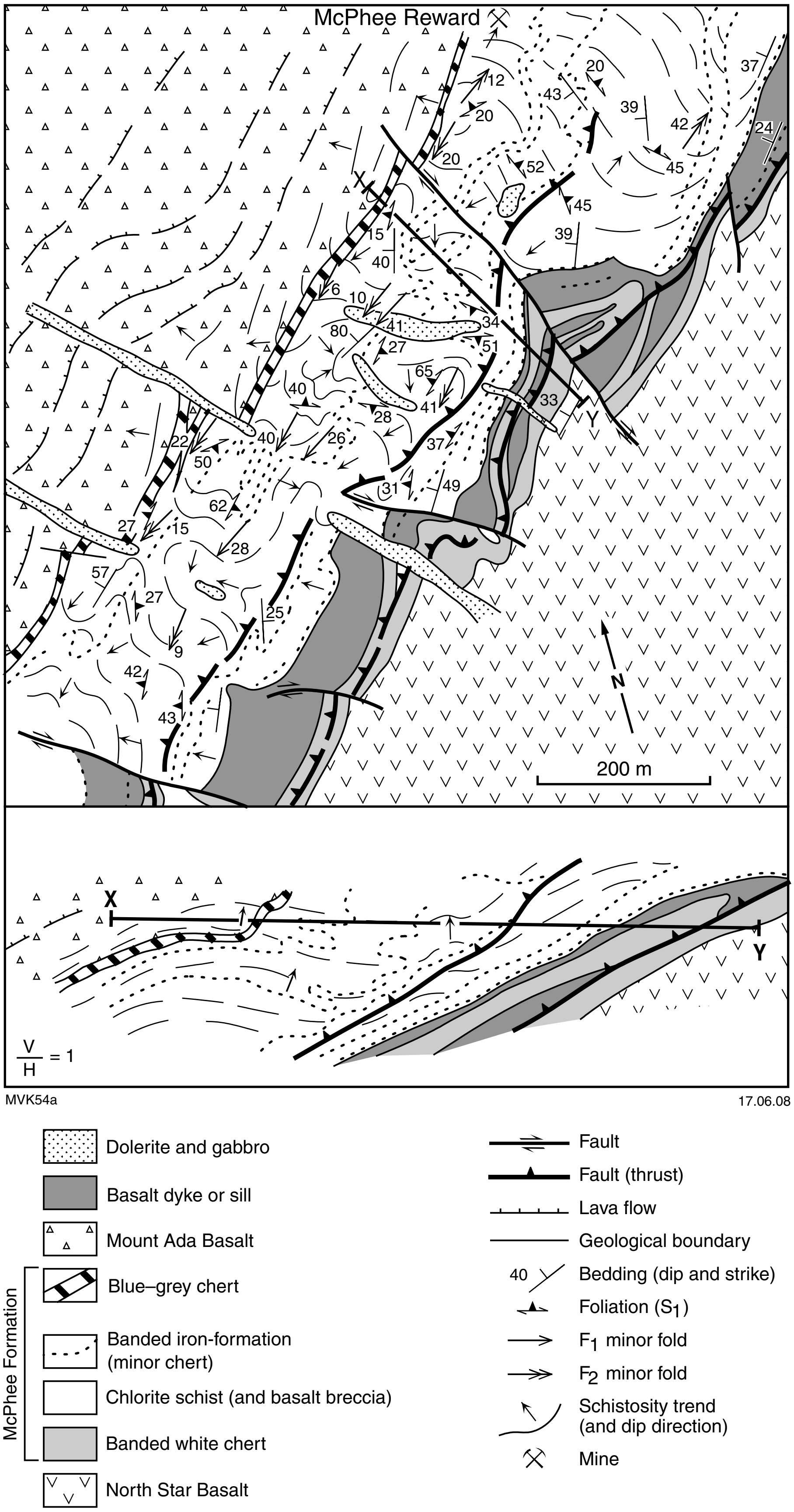

Figure 20. Structural geology of the McPhee Formation at Locality 3. Note the vergence of folds away from the Mount Edgar Granitic Complex and the later reverse faults that displace greenstones up and to the east towards the granitic complex (after Collins, 1989)

the Mount Ada Basalt (west of the shear zone) is older than the North Star Basalt to the east. However, recent geochronology (see above) has established that this is incorrect.

The McPhee Formation is a well-defined stratigraphic unit of regional extent, and is of closely similar age to the c. 3480 Ma Dresser Formation in the North Pole area. At North Pole, the top of the Dresser Formation marks a significant break in mafic volcanism between the North Star Basalt and the overlying Mount Ada Basalt, and includes chert, carbonate rocks, and felsic volcaniclastic rocks. The shear zone along the upper contact of the McPhee Formation at McPhee Reward is relatively minor and has not changed the primary stratigraphic succession (Van Kranendonk et al., 2006b). The apparent thrust movement described by van Haaften and White (1998) is consistent with layer-parallel shear related to bed-length shortening of greenstones within diapiric synclines that formed during doming of the Mount Edgar Granitic Complex (Collins et al., 1998; Hickman, 2001a; Van Kranendonk et al., 2001a, 2004). The earlier, normal sense of movement recognized by van Haaften and White (1998) is consistent with gravity sliding during the early stages of doming (Collins, 1989). Accordingly, this area does not provide evidence of the syndepositional extension proposed by van Haaften and White (1998).

Above the McPhee Formation, the basal part of the Mount Ada Basalt displays well-preserved pillow structures confirming that the northwesterly dipping succession is right way up.

Locality 4: 3466 Ma Homeward Bound Granite intruded by 3460 Ma dolerite dykes of the Salgash dyke swarm (Marble Bar, MGA Zone 50, 786900E 7656800N)

From McPhee Reward, return to Highway 138 and drive southwards almost to Marble Bar. Two kilometres east of the town, where the road rises to cross a granite ridge (MGA Zone 50, 786900E 7656800N), turn right on a minor track. Locality 4 is approximately 500 m north of the road, on the east side of a high granite tor (MGA Zone 50, 786900E 7656800N).

Locality 4 is a geochronology site in the Homeward Bound Granite (Fig. 16; Hickman et al., in prep.). The SHRIMP date obtained on the alkali granite at this locality (3466 ± 2 Ma; Nelson, 1998, sample 142865) supports earlier interpretations (Reynolds et al., 1975; Hickman and Lipple, 1978a,b) that the granite represents a subvolcanic intrusion beneath the 3474–3463 Ma Duffer Formation. The Homeward Bound Granite is a massive to weakly foliated, medium- to coarse-grained, quartz–perthite–biotite–alkali-feldspar granite carrying minor amounts of fluorite, apatite, sericite, and fine muscovite. Recent mapping indicates that the Homeward Bound Granite was probably intruded as two large laccoliths, each up to 3 km thick, over a north–south strike length of at least 30 km. The northern laccolith penetrates the North Star and Mount Ada Basalts in the area of Locality 4, whereas the southern intrusion, east of Marble Bar, invades the lower part of the Duffer Formation. Dykes and sills of porphyritic rhyolite intrude the Duffer Formation and the underlying basalt formation, but none of them reach the top of the Duffer Formation.

About 6 km south of Marble Bar, Reynolds et al. (1975) mapped seven rhyolite domes in the Duffer Formation; these are now interpreted to be high-level intrusions of the Homeward Bound Granite. The domes are up to 1 km long and 120 m thick, and are capped by massive pyrite and finely laminated red, black, and white chert. Black chert locally contains tourmaline, and zones of massive barite on the flanks of the domes contain zinc, lead, and silver mineralization.

A second interesting feature of the Homeward Bound Granite is that it is intruded by the Salgash dyke swarm. These dykes trend approximately east–west at Marble Bar, but on a regional scale they are arranged radially around the western and northern parts of the Mount Edgar Dome. Where the dykes intrude the Duffer Formation they locally show emplacement as subhorizontal sheets or sills. Sills are thickest and most numerous close to the top of the Duffer Formation, presumably due to the presence of the overlying Marble Bar and Chinaman Pool Chert Members. Some dykes cut through the cherts and pass into the Apex Basalt, where they are difficult to map further. The dykes are metamorphosed and, where exposed in the Duffer Formation, contain a continuation of the bedding-parallel foliation that is present in the Duffer Formation. This foliation formed at c. 3310 Ma during uplift of the Mount Edgar Granitic Complex. Based on these observations, the dykes are interpreted to be feeders of the Apex Basalt, and therefore c. 3460 Ma in age. The radial orientation of the dykes around the flanks of the Mount Edgar Dome indicates that crustal extension was related to an early stage of domal uplift, and not to a more regional northeast–southwest crustal extension event as previously interpreted by Kloppenburg (2003).

Locality 5: Homeward Bound Granite (3466 Ma) and related porphyritic rhyolite intruding the Duffer Formation near the Ironclad gold mine (Marble Bar, MGA Zone 50, 787000E 7656000N)

From the Marble Bar Road 500 m south of Locality 4, return 1.5 km westwards to the crossroads 500 m east of Marble Bar (MGA Zone 50, 785800E 7656000N). Turn right, and follow the track northwards for 2 km, passing the Ironclad mine turnoff (MGA Zone 50, 786100E 7657700N). Continue 400 m northwards, and turn right on a minor track (MGA Zone 50, 786300E 7657750N). Continue east for 800 m to Locality 5 (MGA Zone 50, 787000E 7656000N).

The track between Locality 5 and the main track 800 m to the west crosses numerous exposures of porphyritic rhyolite close to the western margin of the Homeward Bound Granite. At Locality 5, the contact between the granite and the porphyritic rhyolite is partly faulted (carbonated shear zone striking north-northeast), but the Homeward Bound Granite outcrops extensively to the southeast. Some localities along the contact show granite veins in the porphyry, suggestive of a similar age, but more than one phase of granite intrusion. The small shear zone at this locality is approximately parallel to a larger mineralized shear zone 500 m to the west and, based on several test pits, may also be mineralized. This suggests that the porphyritic rhyolite, which south of Marble Bar locally contains up to 10% pyrite (Reynolds et al., 1975), may be mineralized over a large area, and deserves further exploration. Sulfide mineralization, similar to the type associated with the porphyritic rhyolite intrusions south of Marble Bar (see **Locality 4**), may have been reworked during c. 3300 Ma shearing and hydrothermal activity in the major shear zone at the Marble Bar mining centre, where pyrite, chalcopyrite, and minor amounts of galena accompany gold mineralization.

Day 2: Stratigraphy and structure of the Warrawoona and Kelly Groups in the Marble Bar area and Warrawoona Syncline

A remarkable feature of the Marble Bar greenstone belt is the transition along strike from essentially undeformed rocks in most of the western part of the belt, to penetratively deformed and highly transposed schists to the south and southeast in the Warrawoona Syncline. This transition is accompanied by a dramatic structural thinning of the sequence, from at least 12 km thick where little deformed, to less than 1.5 km thick in the core of the syncline (Fig. 14).

As the rocks become more intensely sheared from west to east towards the core of the syncline, metamorphic foliations are accompanied by the progressive development of mineral elongation lineations that gradually change orientation from moderately east plunging to vertical in the core of the syncline and moderately west plunging farther to the east (Collins, 1989; Teyssier and Collins, 1990). Teyssier and Collins (1990) showed that in combination with the steepening of the lineations, fabric shape elements changed from S>L, to L>>S, to pure L-tectonites in a zone of sinking (Fig. 21a,b) of identical geometry to that predicted by centrifuge models of diapirs (Fig. 21c,d). The data are consistent with the diapiric model first proposed by Hickman (1975, 1983, 1984). Subsequently, Collins et al. (1998) used more detailed data to support a model involving partial overturn of the crust. Hickman (1984) used Rb–Sr metamorphic age data to suggest that this vertical deformation occurred between 3270 and 2950 Ma, whereas Collins et al. (1998) used geochronological data published by Williams and Collins (1990) to interpret

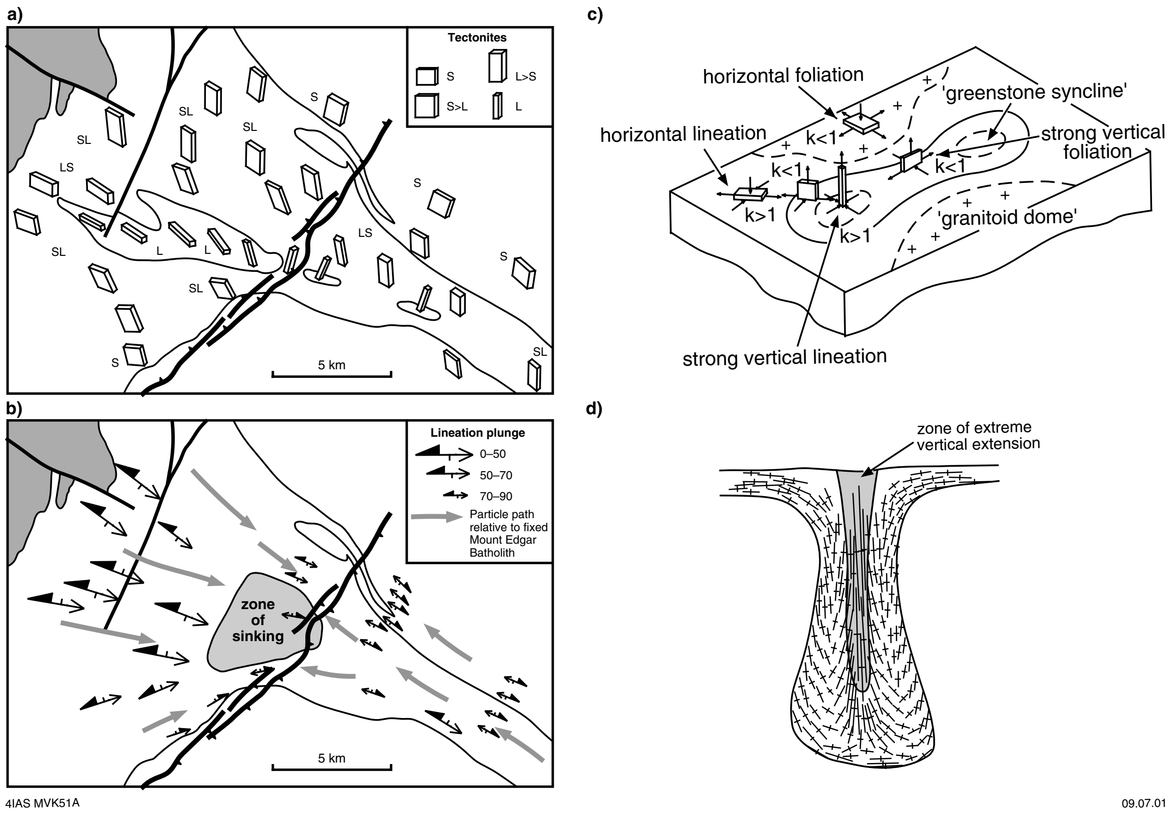

Figure 21. Strain patterns across the margin of the Mount Edgar Granitic Complex and Warrawoona Syncline: a) strain pattern variation; b) lineation pattern showing convergence toward a central point along the synclinal axis. Combined with kinematic data, this indicates that the greenstones sank through a zone of constrictional strain in the axis of the Warrawoona Syncline; c) variation in strain across a modelled diapiric syncline. Note the similarity of the modelled strain relative to that in the Warrawoona Syncline shown in a); d) cross-sectional profile showing a zone of extreme vertical extension in the core of a modelled diapiric syncline. a) and b) after Teyssier and Collins (1990); c) and d) modified from Dixon and Summers (1983)

an age of c. 3300 Ma. More recent work (Hickman and Van Kranendonk, 2004) has supported Hickman's (1984) interpretation that doming was a progressive process spanning at least 500 m.y. between 3450 and 2950 Ma. Major doming at c. 3.0 Ga is indicated by regional deformation of the post-3.02 Ga Gorge Creek Group in the east Pilbara (see Locality 14).

The Klondyke Shear Zone, just north of the axis of the Warrawoona Syncline, is a ring fault close to the southwestern margin of the Mount Edgar Dome (Fig. 6), but recent mapping (Hickman et al., in prep.) has shown it to terminate along the Salgash Fault, which is a scissor fault with east-side-up displacement (Fig. 16). The shear zone is host to numerous small but high-grade gold deposits mined in the early 1900s; similar mineralized shear zones are present in other greenstone belts of the east Pilbara region. There is a wedge of kyanite-bearing schist between the Klondyke and Limestone Shear Zones along the margin of the Mount Edgar Granitic Complex. Pressure–temperature estimates of 5.5 – 6 kbar and 500–600°C have been obtained from the schists and associated metabasites (Delor et al., 1991). These indicate initial burial to depths of about 25 km. Collins and Van Kranendonk (1999) showed that these deeply buried rocks were exhumed within the kyanite stability field, and that this is consistent with heating along the side of a rising diapir as indicated by numerical modelling (Mareschal and West, 1980).

Locality 6: Transect across the upper contact of the Duffer Formation, commencing at Chinaman Pool (Marble Bar, MGA Zone 50, 781600E 7655700N)

From Marble Bar, take the road leading to Comet gold mine and Hillside. Approximately 1 km from Marble Bar turn right on the sealed road to Chinaman and Marble Bar Pools. Proceed 2 km and turn right to Chinaman Pool. Follow the track to the northern end of the picnic area (MGA Zone 50, 781600E 7655700N).

There are excellent outcrops of the 3474–3463 Ma Duffer Formation at the northern end of the Chinaman Pool picnic area (Fig. 22). Here, felsic agglomerate and volcaniclastic conglomerate in the upper section of the Duffer Formation are exposed in rock pavements and other rocky outcrops. The Duffer Formation is approximately 8 km thick in this area and outcrops continuously for 10 km to the northeast. Mapping and geochemical data from systematic sampling traverses across strike (Glikson and Hickman, 1981) show no evidence of structural repetition; this locality exposes rocks at the top of an immensely thick volcanic pile. Clasts in the volcaniclastic rock at Chinaman Pool are angular to subrounded, and composed mainly of felsic lava, felsic volcaniclastic rock, grey-and-white chert, and red-and-white layered chert; all of these lithologies are present in the underlying Duffer Formation. At the southern end of the picnic area, across a small creek and pool, the felsic volcanic rocks are conformably overlain by a discontinuous 10–20 m-thick unit of centimetre-thick-layered blue, white, and red chert. This is the Chinaman Pool Chert Member of the Duffer Formation. The chert dips 80° to the east-northeast and, as seen from pillow structures in overlying basalt to the south, is overturned.

Southwards from the Chinaman Pool Chert Member, a rough walking trail parallels the eastern bank of the Coongan River. The first outcrops include a thin unit of komatiitic basalt that passes southwards into variolitic pillow basalts with well-preserved textures, including small vesicles around pillow margins, and crackle rinds, as well as interpillow hyaloclastite and way-up structures that have survived minor to moderate flattening. The pillows are slightly overturned to the east-northeast, but face west. Continuing south, the pillows become increasingly altered, with bleached pillow rinds, and gas cavities in pillows filled by carbonate and quartz. At the southern end of the rock platform is a thick almost undeformed dolerite sill. This basaltic succession between the Chinaman Pool Chert Member and the Marble Bar Chert Member has a strike length of 8 km and is up to 300 m thick. Its thickness changes along strike, mainly due to its transgression across the irregular surface of the underlying felsic volcanics, but also as a result of local dolerite sills and faulting. The Chinaman Pool Chert Member is far less extensive, having a strike length of less than 1 km. The Chinaman Pool Chert Member and the overlying basaltic succession were deposited in a 200–300 m-deep depression at the top of the felsic volcanic pile, possibly in a collapse structure or in some form of erosional feature; detailed local mapping would probably establish the depositional setting.

Locality 7: Marble Bar Chert Member at Marble Bar Pool (Marble Bar, MGA Zone 50, 781500E 7654800N)

West of the car park at Marble Bar Pool are the well-known water-polished outcrops of blue, white, and red layered chert that forms a rocky 'bar' (the Marble Bar) across the bed of the Coongan River. This area is an important reserve; hammering and rock sampling are not allowed.

At the basal contact of the 100 m-thick Marble Bar Chert Member, altered massive volcanic rocks are cut by an anastomosing network of weakly folded, massive, blue–black chert veins (Fig. 23). Near the basal contact of the chert, the veins contain numerous fragments of country rock, many with jigsaw fit consistent with phreatic brecciation. Some of these veins extend up into the layered chert as breccia dykes and become quite large, reaching up to 12.5 m wide.

The Marble Bar Chert Member is composed of three distinct colour varieties that are interlayered at a centimetre to decimetre scale: thinly bedded red jasper, milky-white chert, and blue–black chert (Fig. 24). At Marble Bar Pool, jasper is present only in the top third of the member. The lower parts of the chert member are composed of more massive white and blue–black chert, and all of the chert is transected by veins of white, grey, and black chert. Jasper is bedded at millimetre scale, defined by slight changes in colour and the degree of faint granularity (Fig. 25a). The fine layering of the jasper indicates deposition under quiet-water conditions and the presence of pillowed

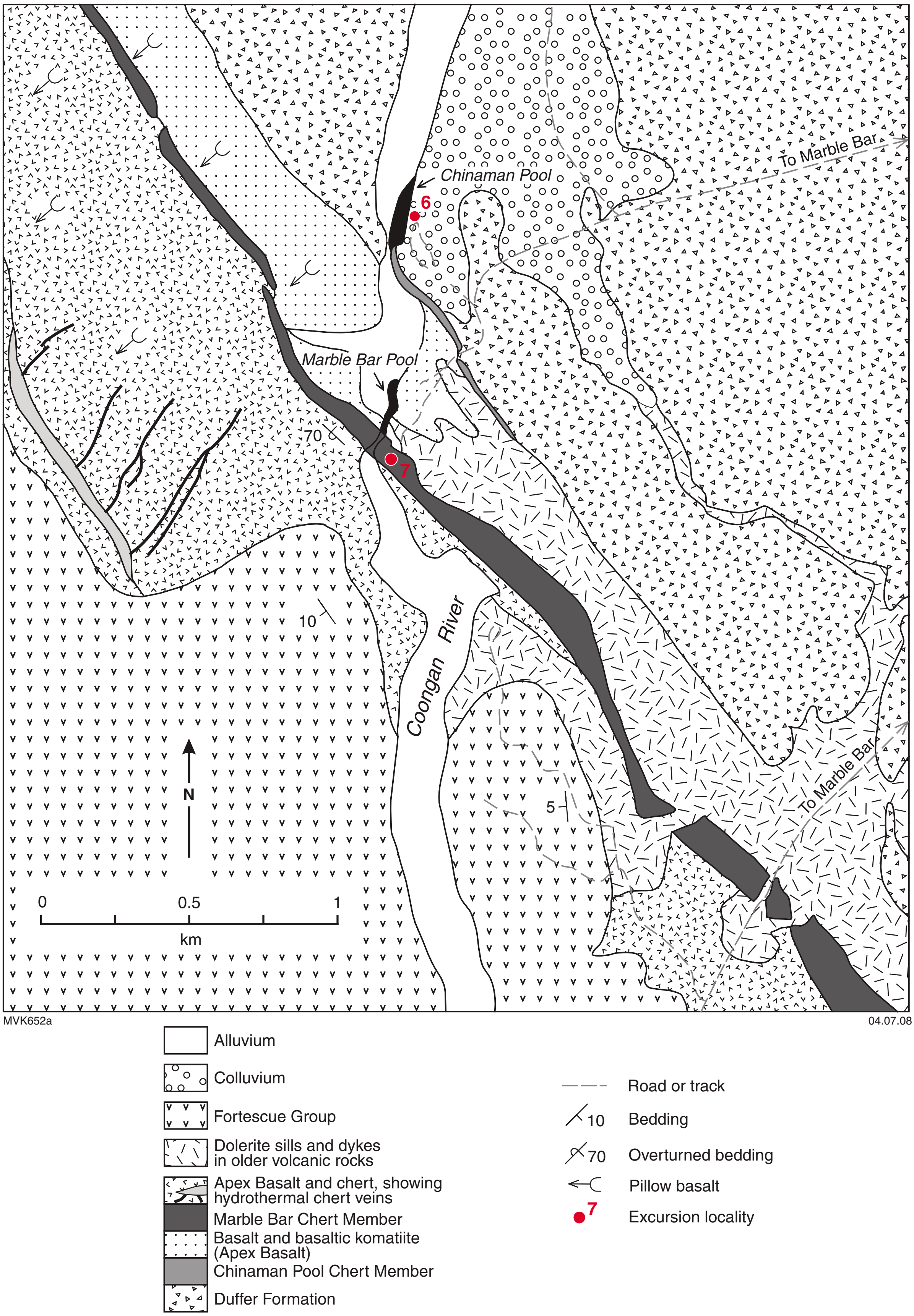

Figure 22. Geological map of the Marble Bar and Chinaman Pool area showing excursion Localities 6 and 7 (modified from Van Kranendonk et al., 2001b)

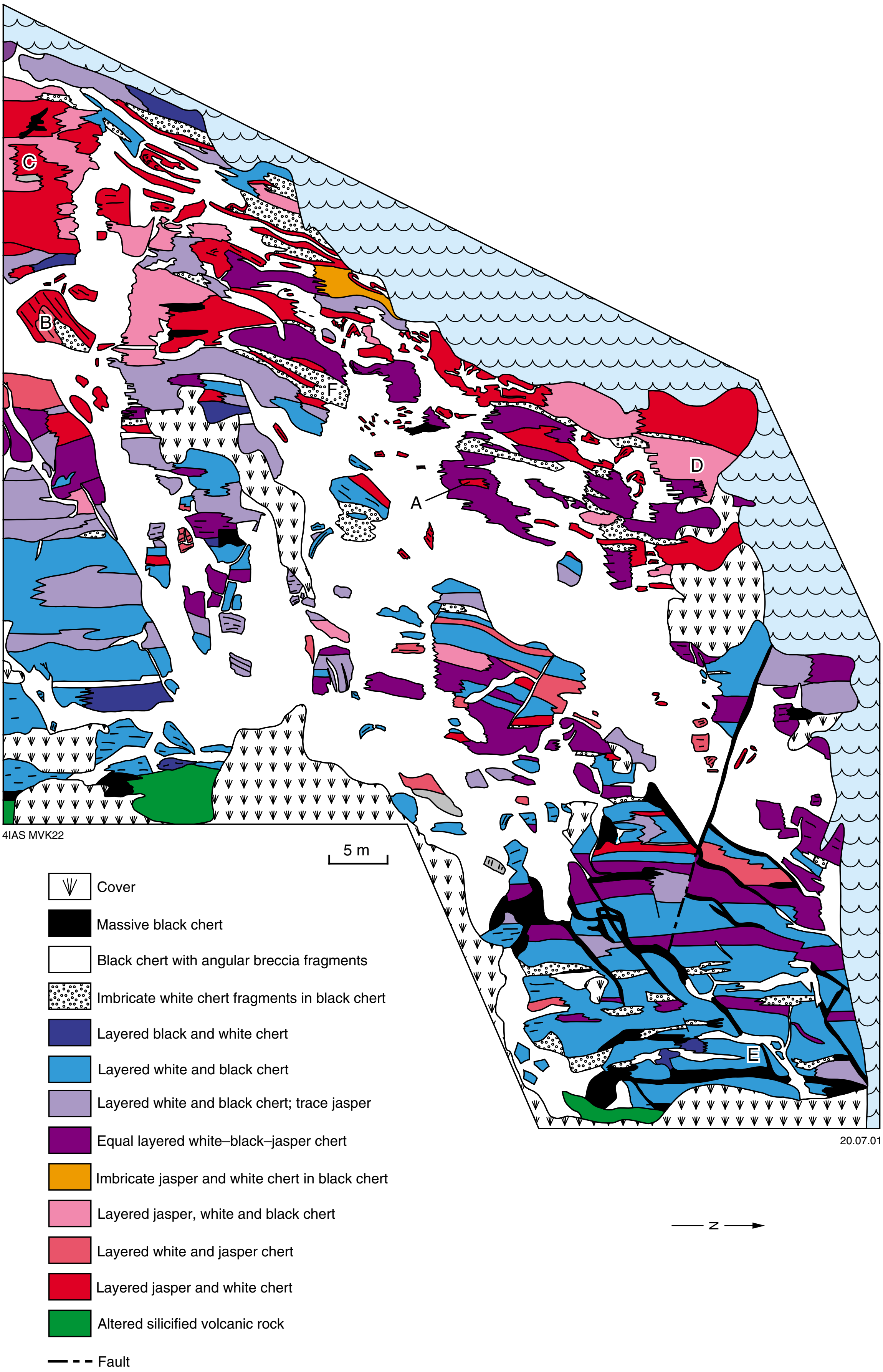

Figure 23. Geological map of the Marble Bar Chert Member at Marble Bar Pool (from Van Kranendonk et al., 2001b)

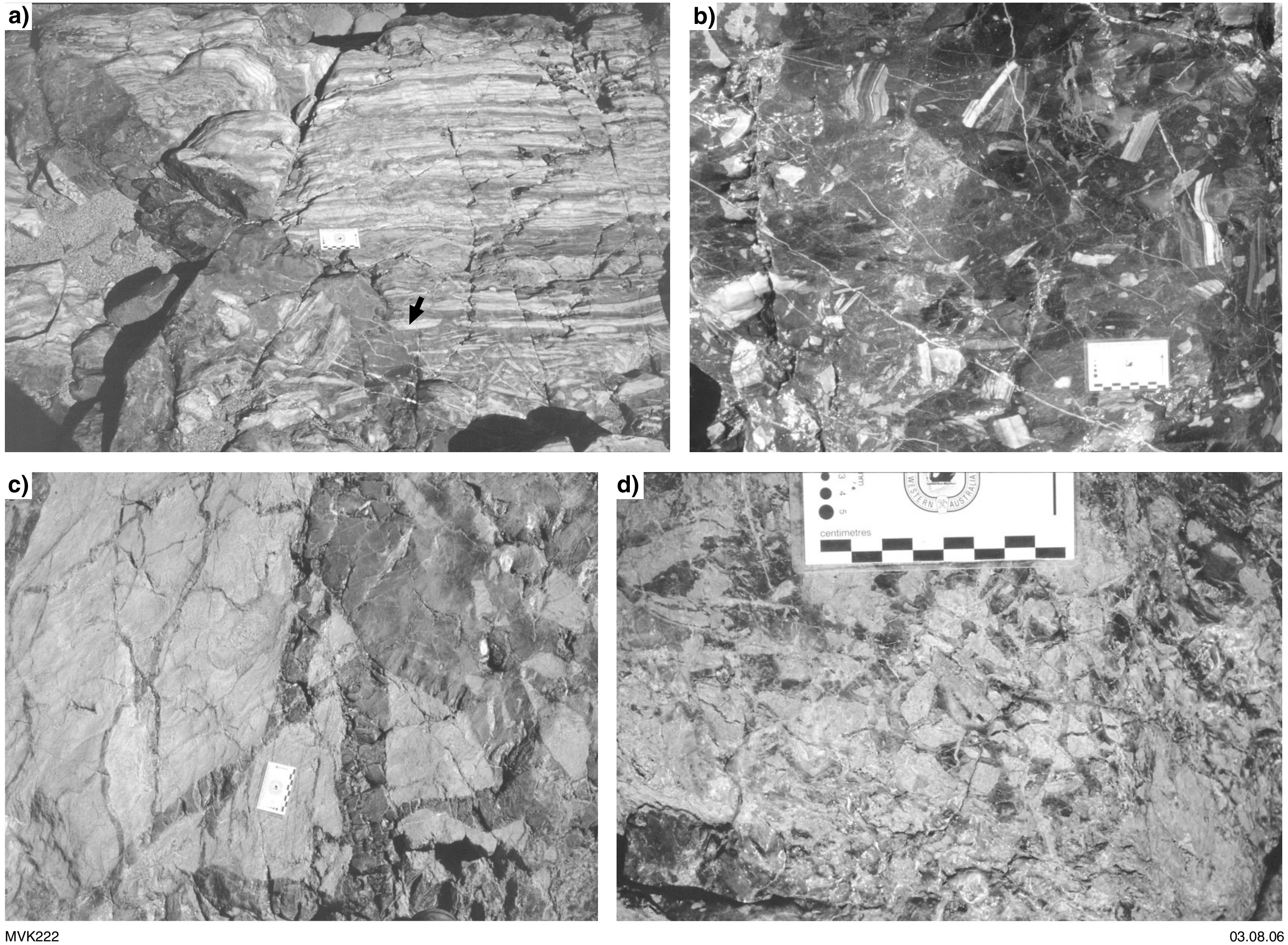

Figure 24. Outcrop photographs of phreatomagmatic silica breccia veins at and beneath the Marble Bar Chert Member at Marble Bar Pool: a) a discordant blue–black chert vein cutting layered blue–black and white chert. Note how the vein margin loses clarity below and below right of the marked arrow and penetrates laterally into a bedding-parallel breccia horizon; b) textures in the core of a large blue–black chert breccia vein showing juxtaposition of angular and well-rounded bedded chert fragments; c) and d) a net of massive blue–black chert veins cutting silicified, altered metabasalt immediately under the base of the Marble Bar Chert Member. Note the jigsaw-puzzle fit of angular fragments in c) and the more rounded nature of fragments in d), where vein material is more voluminous (from Van Kranendonk, 2006)

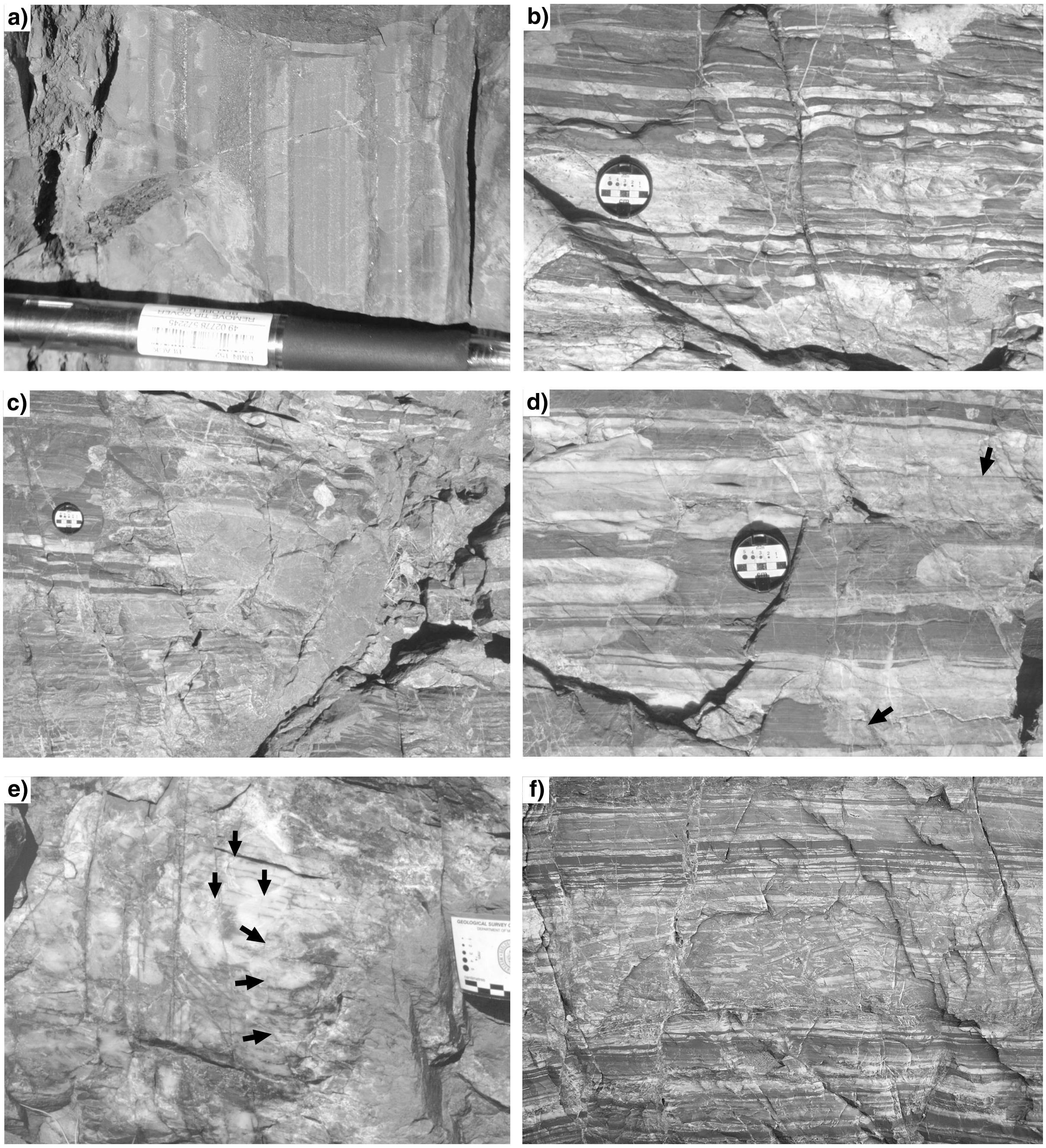

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Figure 25. Photographs showing outcrop features of the Marble Bar Chert Member at Marble Bar Pool: a) bedded jasper chert defined by millimetre-scale variations in granularity and colour; b) characteristic features of the layering between primary bedded jasper and milky white chert, showing local areas of regular white chert layering, pinch-and-swell structures, and large blobs of white chert that cut across bedding (e.g. under lens cap); c) discordant grey chert vein in jasper feeding bedding-parallel layers of grey to white chert; d) Detail of white chert layering in jasper, showing smooth reaction fronts (contacts on either side of lens cap) and jagged reaction fronts (lower arrow) of white chert replacing bedded jasper. Note the faint traces of bedding in the white chert, inherited from bedding in the jasper (upper arrow); e) example of complete replacement of jasper by white chert, which retains faint traces of inherited bedding (vertical arrows) and has pushed hematite granules into an irregular curved line at a high angle to the bedding along a reaction front (horizontal arrows); f) a typical breccia layer of white chert fragments in remobilized jasper (dark grey). Note the low angle of discordance of the breccia layer to bedding (width of view is 1.5 m (from Van Kranendonk, 2006)

basalts above and below suggests relatively deep water. However, current-bedded sandstone and conglomerate in the upper part of the Duffer Formation along strike to the north and south of Marble Bar Pool indicate shallow-water sedimentation prior to deposition of the member. Lenticular sandstone units in the lower part of the overlying Apex Basalt suggest moderate water depths.

Along strike, the Marble Bar Chert Member locally exhibits abrupt changes in thickness where its lower section is intruded by dolerite; in some areas the chert is entirely truncated by cross-cutting dolerite dykes of the Salgash Dyke Swarm. Lateral depositional thinning of the chert is also evident, especially in sections where the chert is not underlain by the basaltic succession but directly overlies felsic volcanic or clastic sedimentary rocks of the Duffer Formation. At Locality 8, 12 km southeast of Marble Bar Pool, only a 2 m-thick chert separates the felsic volcanics of the Duffer Formation from the overlying Apex Basalt.

The white chert forms centimetre-thick layers spaced at irregular intervals through the jasper, but is not bedded. Characteristically, the white chert forms lenses in the jasper, often with pinch-and-swell structures, and in many places the bedded jasper strikes directly into, and is replaced by, white chert (Fig. 25b), much of which appear to have emanated sideways out of white to grey chert veins (Fig. 25c). In some places, the white chert is clearly intrusive into the red jasper chert (Fig. 25c, lower part of Fig. 25d), but elsewhere clearly represents the product of chemical replacement in situ of the jasper chert (middle part of Fig. 25d,e; Van Kranendonk, 2006). Bedding-subparallel breccia horizons in the layered white and jasper chert are composed of elongate angular fragments of white chert within a homogeneous matrix of remobilized, homogeneous jasper or blue–grey chert (Fig. 25f) and, through mapping, can be seen to have emanated from discordant chert veins, thus representing the products of silica-saturated fluids emplaced into the chert under high fluid pressures.

In stark contrast to the thinly bedded jasper, wide, branching veins of blue–black chert and chert breccia cut through the underlying silicified and altered basalts in areas of high vein density, and across most of the bedded chert (Fig. 23). Blue–black chert also forms layers parallel to bedding throughout the chert, but predominantly in the lower third of the outcrop. Breccia veins contain numerous fragments of layered jasper and white chert, the shapes of which vary from angular to round with decreasing size, indicating mechanical milling or thermal erosion processes (or both). Black chert veins are almost entirely absent from the top 20% of the chert as well as from the conformably overlying pillow basalts, indicating that they were emplaced during chert formation and not as a result of later events. Blue–black chert veins feed sills injected laterally into the pre-existing layering. Locally, the force of emplacement of these sills formed layers of pseudoconglomerate comprising a matrix of blue–black chert and imbricated fragments of dismembered white chert layers. The greater volume of black chert towards the bottom of the member, combined with its intrusive origin, indicates that the Marble Bar Chert Member thickened downward as a result of intraplating and underplating of chert.

Although the origin of the jasper is unequivocally depositional and that of the blue–black chert is clearly intrusive, the origin of the white chert is more complex. A depositional origin for the white chert is superficially suggested by the observations that it is interlayered with jasper at such a regular spacing, lies in contact with jasper across contacts deformed by pinch-and-swell structures, and forms shingled dismembered fragments within jasper in some bed-parallel breccia layers. However, several other features suggest that the white chert was intrusive. First, the white chert displays no bedding, except where inherited through replacement of bedded jasper (see below). Second, veins of white chert commonly truncate bedded jasper and link with bedding-parallel sills of white chert. Where intrusion of white chert is extensive, it replaces bedded jasper at grain scale, replacing the jasper but retaining the bedding characteristics of the host and pushing the finely disseminated hematite of the precursor into diffuse, colloform bands along replacement fronts whose margins may be 90° to bedding. The presence of pinch-and-swell and load structures between the jasper and white chert sills indicates that the white chert was probably intruded before the jasper was lithified.

Thus, the sequence of events that formed the Marble Bar Chert Member include deposition of finely bedded jasper during hydrothermal emanations (Sigutani, 1992), subsurface intrusion of white chert as sills and veins, and final intrusion of blue–black chert veins. As with other chert horizons visited during this excursion, the original material deposited was an oxidized species, namely hematitic jasper, although in this case deposition was in quiet water.

The top contact of the chert is exposed to within 1 or 2 cm of overlying beautifully preserved pillow basalt of the Apex Basalt. Pillows commonly contain one or multiple pillow shelves. No chert veins extend up through the chert into the pillow basalt, indicating that deposition of the basalt occurred after chert veining. The age of the Apex Basalt is between 3463 Ma (youngest date obtained for the Duffer Formation, McNaughton et al., 1993) and 3458 Ma (date from a felsic sill, Williams, 1999).

Locality 8: Stretched clasts in the upper Duffer Formation (Marble Bar, MGA Zone 50, 789934 7645932N)

Return to Marble Bar and, 600 m east of the town, take the gravel road leading south towards Corunna Downs. Follow this road for approximately 13 km and turn off to the west (no track, 4WD) at MGA Zone 50, 790600E 7646400N. Drive west-southwest for 600 m to the base of low hills of felsic schist.. Walk up the slopes westwards to Locality 8 (MGA Zone 50, 789934E 7645932N) on the crest of a northeast-trending ridge.

Locality 8 is the first of three excursion stops (Localities 8–10) selected to show structural evidence, visible at outcrop scale, demonstrating that vertical tectonic processes (diapiric deformation) were primarily responsible for the regional dome-and-basin pattern seen in the East Pilbara Terrane (Hickman and Van Kranendonk, 2004). Alternative models have attempted to explain the structural evolution of the terrane using Phanerozoic-style horizontal plate-tectonic processes (Bickle et al., 1980, 1985, 1993; Boulter et al., 1987; Barley, 1993, 1997; Zegers et al., 1996, 2001; van Haaften and White, 1998; Kloppenburg et al., 2001; Blewett, 2002). These models differ in detail, but most include suggestions of Alpine-style thrusting at 3450 and 3300 Ma and subsequent periods of major regional extension during which the granitic domes formed as metamorphic core complexes.

Locality 8 shows small-scale structures within rhyolitic schist near the top of the Duffer Formation. At Marble Bar, the Duffer Formation is not strongly sheared, and its succession is 8 km thick. In the area of Locality 8, on the northern limb of the Warrawoona Syncline, the formation is both tectonically attenuated and intruded by the 3466 Ma Homeward Bound Granite, and as a result is only 1.8 km thick. Participants will immediately observe that the felsic volcanic rocks of this area are strongly schistose. The schistosity is part of the steeply inclined to vertical tectonic foliation that rims the 50 km-diameter granitic core of the Mount Edgar Dome. The sense of shear movement, demonstrated by changes in metamorphic grade and shear fabrics, is granite side up. At Locality 8, on the southwestern flank of the dome, the dominant foliation dips 80° towards 225°, although the structural history of the area included development of more than one foliation, and cross-cutting structures such as strain-slip cleavage are commonly observed. The main foliation in the schists locally appears to bear an axial-planar relationship to tight-to-isoclinal folding defined by layers of altered mafic rocks in the otherwise felsic succession. Regional evidence indicates that the first major deformation in this area (D_3) occurred at c. 3.31 Ga; this is the interpreted age of the main tectonic foliation (S_3) at Locality 8.

The schist is a metamorphosed rhyolitic volcaniclastic rock containing clasts of vein quartz, felsic volcanic rock, and chert. These clasts are stretched, with the long axis plunging 60° towards 115° (Fig. 26). The same elongation direction is repeated at other localities on the northern limb of the syncline, as also recorded by Kloppenburg (2003). Locality 9 shows stretching lineations with a similar southeasterly plunge. Eastwards from Locality 9, the stretching fabrics steepen to almost vertical in the more highly deformed and deeper axial region of the syncline at Locality 10. The orientation of stretching is normal to the general plunge of the Warrawoona Syncline and consistent with sinking of the greenstone succession relative to the adjacent granitic rocks in the Mount Edgar Dome. In detail, however, various minor folds within the northern limb of the Warrawoona Syncline plunge southeast as well as northwest, indicating primary noncylindricality or superposed folding. Shear intensity in the greenstones varies with lithology and proximity to ring faults, but there is a general trend of increasing shear towards the centre of the syncline and into deeper structural levels of the syncline (east-southeastwards). The east-southeast plunging elongation of the clasts at Locality 8 is consistent with the diapiric model for development of the Warrawoona Syncline (Hickman, 1975, 1983, 1984; Collins, 1989; Collins et al., 1998), but is far less easily reconciled with the regional unidirectional southwest–northeast extension and stretching that is essential to the metamorphic core complex model (Kloppenburg et al., 2001; Kloppenburg, 2003).

AHH325 28.04.08

Figure 26. Stretched clasts in the Duffer Formation at Locality 8

AHH326 28.04.08

Figure 27. View of the upper Duffer Formation at Locality 8. A thin chert unit at the top of the ridge is correlated with the Marble Bar Chert Member

Other significant features at Locality 8 include the presence of secondary chert units along silicified shear zones in the felsic schists, and the observation that on the top of the ridge, a few hundred metres to the west of Locality 8 (Fig. 27), the Duffer Formation is separated from the Apex Basalt by only a few metres of ferruginous and grey-and-white layered chert (MGA Zone 50, 789709E 7645873N). This chert is at the same stratigraphic level as the Marble Bar Chert Member at Locality 7, but it is very much thinner and different in appearance. It is impossible to map lateral facies changes in the Marble Bar Chert Member over most of the 12 km between Localities 7 and 8 because the member is concealed by basalt of the Fortescue Group. However, it is clear that the Marble Bar Chert Member is not laterally continuous, but instead forms lenticular units at this stratigraphic level; the distribution of the lenses probably depends on local hydrothermal activity.

Locality 9: Shallow–moderate east-plunging lineations pointing towards the zone of sinking in the Warrawoona Syncline (Marble Bar, MGA Zone 50, 790500E 7641800N)

Proceed for 4.5 km along the Marble Bar – Corunna Downs road, stopping about 500 m past the sharp left bend in the road (MGA Zone 50, 790500E 7641800N).

The steep hill on the south side of the track contains a thin chert unit in metamorphosed schistose metabasalt. The rocks show well-developed elongation lineations plunging moderately to the east-southeast, parallel in trend to the axis of the Warrawoona Syncline and plunging towards the zone of vertical L-tectonites in the centre of the zone of sinking farther along strike to the east (see **Locality 10**; Collins, 1989; Collins et al., 1998). Note that at the top of the hill there is a set of en echelon quartz veins in tension gashes that are oriented at 90° to the lineation direction; these are interpreted as having formed during a late component of the stretching during sinking of the greenstones.

Locality 10: Zone of sinking — vertical L-tectonites in the core of the Warrawoona Syncline (Marble Bar, MGA Zone 50, 794480E 7639020N)

Proceed for 6 km towards Corunna Downs, stopping 100 m past a creek crossing and small gorge through a chert ridge (silicified sheared Wyman Formation). At this point the road bends sharply left. Space for safe parking is available about 50 m east of the bend. Walk to the top of the hill in the angle of the bend in the road (MGA Zone 50, 794480E 7639020N).

This locality is in the core of the Warrawoona Syncline at the centre of the zone of sinking of the greenstones (Fig. 21) and is less than 200 m south of the axial shear zone that separates the Mount Edgar and Corunna Downs Domes. This axial shear, which also marks the boundary between the Marble Bar and Kelly greenstone belts, excises at least 7 km of the sequence between the Apex Basalt and the top of the Wyman Formation. West of Locality 10 it is intruded by peridotite of the 3.18 Ga Dalton Suite. The rocks at Locality 10 are characterized by vertical L-tectonite fabrics in silicified felsic tuffaceous rocks of the 3.32 Ga Wyman Formation (Fig. 28). Although cleavages are visible cutting relict, highly transposed (millimetre-scale) bedding in the chert, the cleavages are in all directions and reflect equal compression from all directions during vertical stretching of the rock in the zone of sinking of the greenstones. Once established, this post-3.32 Ga zone in the core of the syncline was reactivated during regional extension at 3.18 Ga (Dalton Suite), and very probably again after 3.02 Ga during renewed doming. Sandstone and pelite of either the 3.2 Ga Soanesville Group or the 3.02 Ga Gorge Creek Group were displaced by later movement along the shear zone 10 km west of Locality 10. Throughout the East Pilbara Terrane, the Gorge Creek Group was strongly deformed by D_9 (2950–2930 Ma).

Locality 11: Panorama Formation, Strelley Pool Formation, and Euro Basalt at Camel Creek on the southern limb of the Warrawoona Syncline (Marble Bar, MGA Zone 50, 790500E 7635100N)

Continue on the road towards Corunna Downs. After 2.3 km the road crosses a low ridge of komatiitic basalt and serpentinized peridotite (lower part of the Euro Basalt) and goes through a gate onto granitic rocks of the Corunna Downs Granitic Complex. Keep to the main track, which swings southwest, and after 4.3 km, cross Camel Creek. Continue on the track for 300 m past the creek and take the right fork signed to Old Corunna Downs Airfield. After 800 m take a minor track to the right, leaving the Old Corunna Downs Airfield track. Follow this old track northwards for 1.5 km to an old gate. Here, a large ridge of chert is cut by the gorge of Camel Creek (MGA Zone 50, 790500E 7635100N).

This locality is in the northwestern part of the Kelly greenstone belt (Fig. 16) and provides a section from the Warrawoona Group through the Strelley Pool Formation into the Kelly Group. In contrast to the southwestern margin of the Mount Edgar Granitic Complex north of Warrawoona, the northwestern margin of the Corunna Downs Granitic Complex is not strongly sheared and, except where excised by granitic intrusion (mainly at 3.31 Ga), original stratigraphic relations are well preserved. The Panorama Formation in this area consists of variably silicified felsic volcanic breccia and volcaniclastic rocks that have been intruded by veins of black or grey chert. The chert veins extend into the base of the layered chert of the Strelley Pool Formation, but do not pass through the formation. Accordingly, they have been interpreted by some workers as hydrothermal cherts representing feeders to the bedded chert. Another possibility is that the veins filled extensional fissures extending deep into the surface on which the Strelley Pool Formation was deposited, and that the grey-to-black chert includes not only hydrothermal deposits but also silicified sedimentary rock. In other parts of the East Pilbara Terrane, the upper sections of similar dykes below the Strelley Pool Formation contain debris from the formation, and locally include sandstone (Wacey et al., in prep.).

The Strelley Pool Formation extends over a strike length of 30 km around the northwestern side of the Corunna Downs Dome, and would extend continuously over more than 70 km into the eastern side of the dome if it had not been intruded by granitic rocks of the Emu Pool Supersuite. Its thickness varies from 3 to 30 m, and across the 70 km it includes layered grey-and-white chert, sandstone, conglomerate, wavy laminated carbonate rocks that locally

Figure 28. Outcrop view of the zone of sinking in the core of the Warrawoona Syncline, showing L-tectonites derived from silicified felsic volcaniclastic rocks of the Wyman Formation

contain stromatolites, and local evaporites. On the eastern side of Camel Creek at Locality 11, the Strelley Pool Formation is intruded by ultramafic sills, probably related to the overlying Euro Basalt. The lower chert section is approximately 20 m thick and is predominantly composed of layered grey-and-white chert with minor brown chert, but no observed jasper. At Camel Creek, this lower chert overlies silicified felsic volcanic breccia. The upper chert section is 10 m thick and contains more grey chert than the lower part of the formation. Approximately 20 m above the upper chert section, and separated from it by ultramafic rock, there is a slightly sulfidic, finely layered, dark-grey to black chert (probably a silicified carbonaceous shale). Two kilometres to the east of Locality 11, this uppermost chert unit is a finely granular grey rock, and in that area is probably a silicified siltstone. In the same eastern area, there is an additional 2 m-thick grey chert separated from the lower chert by a 10 m-thick felsic volcanic unit and a thin ultramafic sill, but this lowermost chert has not been identified near Locality 11. Approximately 4 km east of Locality 11, ultramafic intrusion has separated the formation into four chert units.

In the northwestern part of the Kelly greenstone belt, the Strelley Pool Formation is overlain by komatiitic basalt and peridotite of the Euro Basalt. Spinifex textures in parts of the peridotite suggest lava flows, but sills may also be present in the volcanic pile.

Locality 12: Ultramafic flows in the Euro Basalt at Camel Creek (Marble Bar, MGA Zone 50, 790000E 7635700N)

Continue on the track northwards for 650 m to a crossing over a small creek (MGA Zone 50, 790200E 7635750N). Walk 250 m west-southwest along the creek.

Locality 12 is immediately above the upper contact of a 400 m-thick peridotite unit in the basal section of the Euro Basalt, and provides an exposure of overlying komatiitic basalt. Layers of pyroxene spinifex texture in these ultramafic rocks indicate lava flows. Geochemical data through this area (Smithies et al., 2007) have shown that the peridotite immediately south of Locality 12 contains

44.33% SiO_2, 0.26% TiO_2, 5.37% Al_2O_3, 28.12% MgO, 5.62% CaO_2, 2474 ppm Cr, and 1354 ppm Ni (GSWA sample 179816), whereas the komatiitic basalt close to Locality 12 contains 49.48% SiO_2, 0.50% TiO_2, 11.99% Al_2O_3, 11.02% MgO, 13.40% CaO_2, 1785 ppm Cr, and 417 ppm Ni (GSWA 179817). Smithies et al. (2007) reported that the komatiitic rocks at Camel Creek are not depleted in aluminium, and have low incompatible trace-element concentrations and flat normalized trace-element patterns.

Locality 13: Columnar rhyolite in the Wyman Formation at Camel Creek (Marble Bar, MGA Zone 50, 787400E 7640200N)

Continue on the track northwards for 9 km to a major creek crossing (Camel Creek). Approximately 100 m north of the creek crossing (MGA Zone 50, 785000E 7640800N) turn right on a track that leads 2.5 km eastwards to a crossing with a tributary creek (MGA Zone 50, 787400E, 7640100N).

Approximately 100 m up the tributary creek there is a spectacular cliff exposure of columnar rhyolite in the Wyman Formation of the Kelly Group (Hickman, 1983, fig. 11). Similar columnar rhyolite is exposed in several other areas of the East Pilbara Terrane, for example, at Budjan Creek (MGA Zone 50, 796000E 7582300N), in the northern part of the McPhee greenstone belt (MGA Zone 51, 222200E 7619200N), and at Wallabirdee Ridge (MGA Zone 51, 219000E 7621000N). The rock at Locality 13 consists of euhedral phenocrysts of beta-quartz and K-feldspar set in a groundmass of quartz, twinned K-feldspar, and sericite, with variable epidote and chlorite alteration. Three samples from this area, analysed by Jahn et al. (1981), contained between 73.36% and 79.34% SiO_2, and between 6.11% and 11.05% K_2O.

Geochronological data are consistent with a genetic link between the 3325–3315 Ma rhyolite units of the Wyman Formation and the 3325–3290 Ma granitic intrusions of the Emu Pool Supersuite. The Emu Pool Supersuite intrudes the lower part of the Euro Basalt along the northern margin of the Corunna Downs Granitic Complex, and dykes and sills of c. 3303 Ma porphyritic rhyolite (Nelson, 2002, GSWA 160218) intrude the upper Euro Basalt immediately southwest of Camel Creek. The elevated SiO_2 , K_2O, and Rb (up to 480 ppm Rb) levels of the rhyolite probably reflect strong fractionation combined with pneumatolytic alteration by Si- and K-fluids (Glikson et al., 1987). It is uncertain whether the columnar rhyolite at Locality 13 is part of a thick flow or represents a local high-level intrusion. The columns plunge moderately to the southeast, which is approximately normal to the local north-northwesterly dip of the succession.

Return to the main Camel Creek track (MGA Zone 50, 785000E 7640800N), turn right and drive 7 km to the Marble Bar – Hillside road (MGA Zone 50, 782200E 7645800N). Turn right and drive approximately 14 km to Marble Bar.

Day 3: The world's oldest copper–molybdenum deposit and an introduction to the De Grey Supergroup

Locality 14: Cleaverville Formation of the Gorge Creek Group at Coppin Gap

From Marble Bar, drive towards Port Hedland for 33 km and turn right on the road signed to Bamboo Creek. Follow this road for 15 km, then turn right at the Bamboo Creek sign. Continue eastwards for 13 km, then turn left on the road signed to Coppin Gap. Follow this road northwards for 7 km to its termination at a parking area 100 m south of a large gorge (Coppin Gap). Follow the footpath into the gorge.

The deep gorge of Coppin Gap (Fig. 29) provides superb exposures of red-and-black jaspilite (Fig. 30), banded chert, and banded iron-formation of the Cleaverville Formation, which is a c. 3020 Ma formation of the Gorge Creek Group. The Gorge Creek Group is the oldest group of the De Grey Supergroup and unconformably overlies the Pilbara Supergroup, which at Locality 14 is represented by the c. 3350 Ma Euro Basalt. Coppin Gap is in the tight, east-plunging Coppin Gap Syncline, which separates the Muccan and Mount Edgar Granitic Complexes. The northern limb of the syncline is locally overturned and sheared out along the South Muccan Shear Zone. Remnants of the Farrel Quartzite, the basal formation of the Gorge Creek Group, are locally preserved on both sides of the syncline, but are not exposed at Coppin Gap. South of the gap, the Cleaverville Formation is in sheared contact with ultramafic rocks of the Euro Basalt.

Significant features of the Cleaverville Formation at Coppin Gap are its steep dip, and the complexity of folding it displays. This degree of deformation is typical of the Gorge Creek Group throughout the East Pilbara Terrane, where the group forms the outer flanks of domes, conforming to the same dome-and-basin pattern shown

Figure 29. Folded jaspilitic banded iron-formation in the Cleaverville Formation at Coppin Gap

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Figure 30. View of Coppin Gap from the south. Cliffs on both sides of the gap are composed of the c. 3020 Ma Cleaverville Formation in the core of a faulted syncline between the Mount Edgar and Muccan Domes

by the Pilbara Supergroup. This deformation, which was post-3020 Ma, is attributed to D_9, and provides direct evidence of the degree to which the domes of the East Pilbara Terrane were enhanced at c. 2950 Ma. Hickman (1983, 1984) interpreted this stage of doming to be the result of solid-state diapirism, and it is significant that extensive geochronology on the granitic rocks of the terrane has revealed no intrusions of the 2955–2920 Ma Sisters Supersuite in the eastern part of the East Pilbara Terrane.

Locality 15: Olivine spinifex texture in peridotite flows of the Euro Basalt

Exposures of the Euro Basalt immediately east of the parking area at Coppin Gap show good examples of carbonate-altered olivine spinifex texture in komatiite flows.

Locality 16: Spinifex Ridge molybdenum–copper mine development (Moly Mines Ltd)

From Coppin Gap, return 1.5 km on the road towards Marble Bar, and turn right to the mine office of Moly Mines Limited.

The following description of the Spinifex Ridge mine was provided by Brendan Cummins (Moly Mines Limited).

The Spinifex Ridge molybdenum deposit is situated at the apex of the Coppin Gap Granodiorite, a 3314 ± 3 Ma (Williams, 1999) intrusion into the Warrawoona Group on the northern margin of the Mount Edgar Granitic Complex (Fig. 16). The granodiorite has generated an extensive alteration halo dominated by quartz stockwork and potassic and phyllic alteration. Molybdenum and copper mineralization occurs as vein-hosted molybdenite and chalcopyrite.

The steep northerly inclination of the Warrawoona Group at Coppin Gap is the result of doming of the area during and after intrusion of the Coppin Gap Granodiorite. Part of this tilting occurred after 3020 Ma, as demonstrated by the almost vertical inclination of the Cleaverville Formation at Locality 14, and must be partly a result of D_9. Therefore, the present attitude of the ore system is not the original attitude, but includes a northerly tilt of approximately 70°. The intrusion of the Coppin Gap Granodiorite created the fault system into which the younger mineralized granodiorite was intruded and deposited the molybdenum and copper mineralization. The mineralized granodiorite has sharp subvertical and commonly brecciated contacts, which are interpreted to represent these fault zones. However, the distribution of molybdenum and copper overprints these structures (Figs 31–33). Some molybdenite was deposited in secondary fractures and faults resulting from local overpressure.

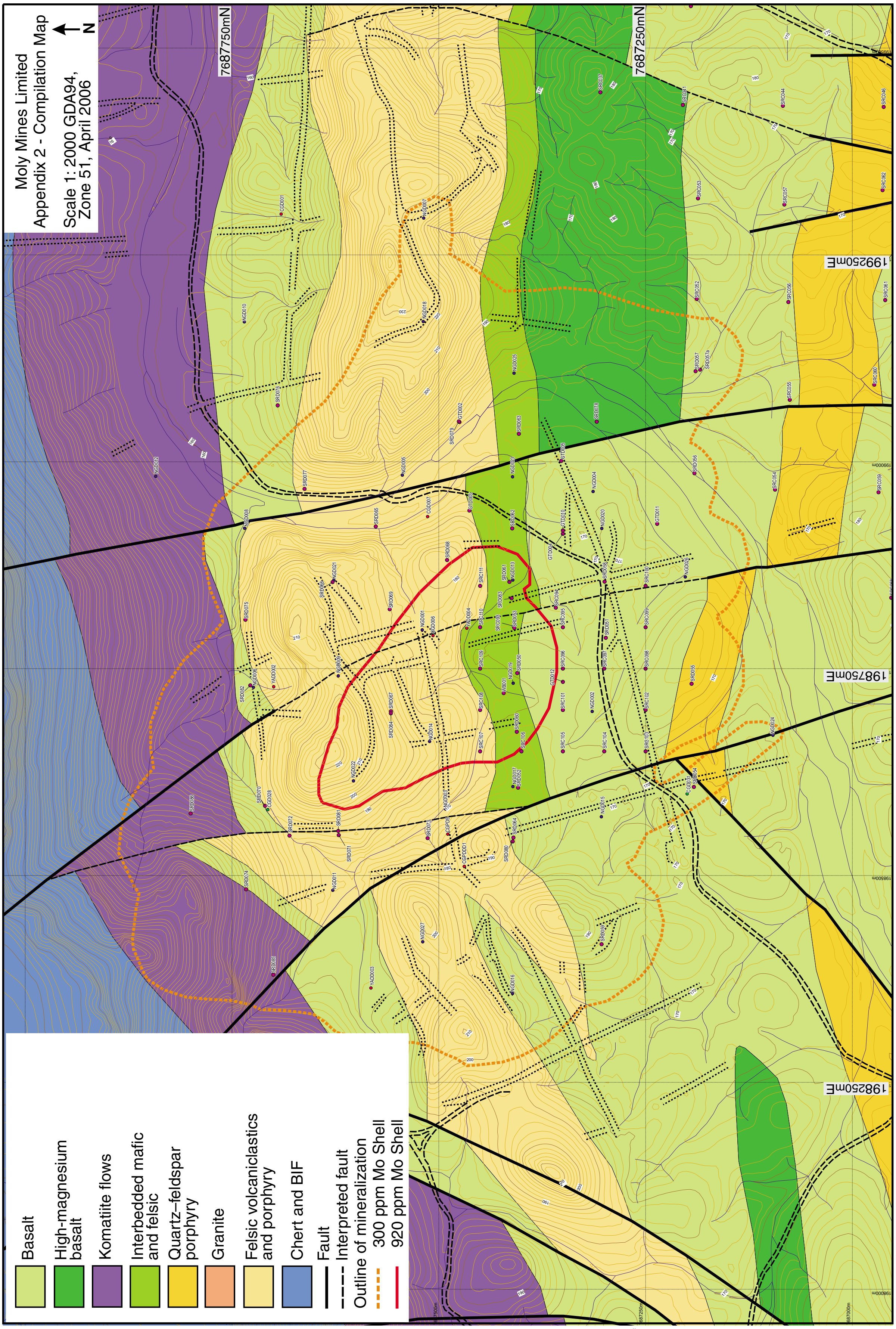
Since 1969, there has been intermittent exploration by several companies at Spinifex Ridge. Most companies completed detailed mapping, extensive geochemical programs, minor geophysical investigations, and variable amounts of percussion and diamond drilling. The geology and grade data presented in Table 2 are derived from historical drill and assay records in conjunction with more recent drilling by Moly Mines (Moly Mines Limited 2007, 2008).

Spinifex Ridge can be classified as an Archean, low-F porphyry molybdenum deposit (Sinclair, 1995). These deposits are characterized by stockworks of molybdenite-bearing quartz veinlets and fractures hosted by intermediate to felsic intrusive and associated country rocks. Chalcopyrite, scheelite, and galena may be present but are generally subordinate. The deposits vary in shape from an inverted cup, to roughly cylindrical, to highly irregular. They are typically hundreds of metres across and range from tens to hundreds of metres in vertical extent. They are typically low-grade, large-tonnage deposits that are amenable to bulk mining methods.

These deposits are thought to originate from large volumes of magmatic, highly saline aqueous fluids under pressure. Multiple stages of brecciation related to explosive fluid-pressure release from the upper parts of small intrusions resulted in deposition of ore and gangue minerals in cross-cutting fractures, veinlets, and breccias in the outer carapace of the intrusions and in associated country rocks. Incursion of meteoric water during the waning stages of the magmatic–hydrothermal system may have resulted in late alteration of the host rocks, but did not play a significant role in the ore-forming process (Sinclair, 1995).

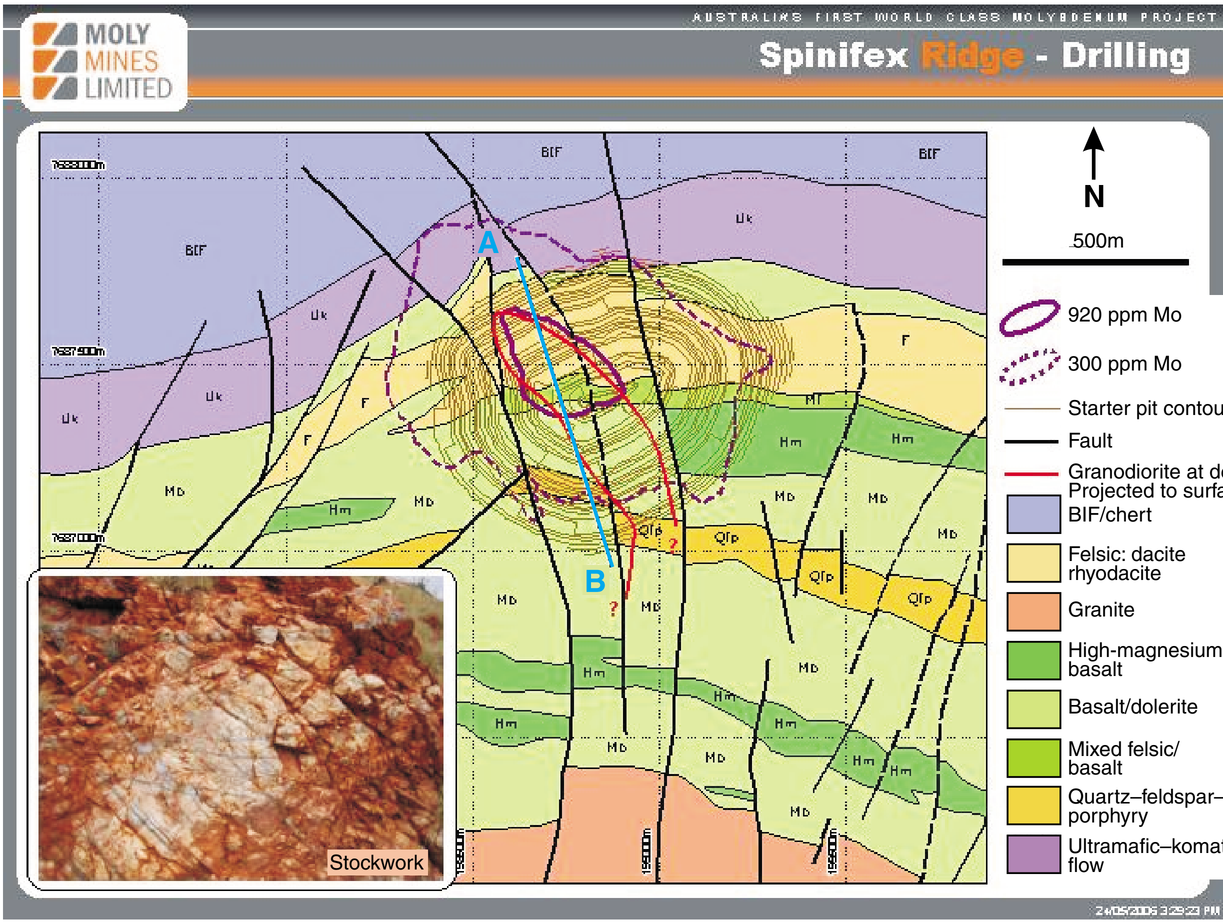
As is typical of most porphyry deposits, the bulk of the metal content at Spinifex Ridge is in or adjacent to a granodiorite intrusion that provided the metals, mineralizing fluids, and heat source. The intrusion is now subhorizontal, but was almost vertical when it was emplaced.

The mineralized granodiorite is 50 to 80 m wide, up to 200 m thick, and does not outcrop; it is about 120 m



MVK712 27.06.08

Figure 31. Surface geology of the Spinifex Ridge prospect showing outline of mineralization (after Moly Mines Limited)



MVK711 09.08.06

Figure 32. Geology of the Spinifex Ridge prospect showing proposed pit contours and granodiorite at depth projected to the surface. Line of Section A–B (Fig. 33) indicated (after Moly Mines Limited)

below the current topographic surface. The mineralized granodiorite strikes to the northwest and plunges to the southeast at about 35°, is about 500 m long, and is open to the northwest and southeast. The country rocks dip to the north at about 70°. The sharpness of the vertical contacts is attributed to bounding subvertical structures, which have been confirmed from inclined, oriented drillholes. There is a series of thin (<5 m thick) dykes and sills radiating from the mineralized granodiorite, which generally appear to be quite flat, and some are well mineralized and quartz veined. Molybdenite extends in a radial pattern away from the mineralized granodiorite for up to 400 m. High-grade molybdenum mineralization is restricted to within 100 m of the mineralized granodiorite.

The molybdenum–copper deposit is hosted within four lithologies:

- mafic rocks – basalt, pillowed basalt, and dolerite;
- felsic rocks – quartz- or feldspar-dominated porphyry, rhyolites, dacites, and rhyodacites;
- granodiorite;
- ultramafic rocks – spinifex textured komatiites and high-Mg basaltic flows.

The mineralization consists of a complex series of multiphase stockwork veins, which contain coarse molybdenum and copper sulfide grains. The stockworking is intense and molybdenum grades tend to increase with increasing vein density. As well as being vein related, disseminated copper sulfides are also present in the mineralized granodiorite wallrock. The mineralized veins vary in width from millimetre to centimetre scale and have a preferred orientation to the northwest and northeast at variable dips forming a conjugate set. Surface outcrop shows the veins to be tens of metres in length.

The veins are typically quartz dominated with accessory K-feldspar. Sulfides are found in vein selvages or as large irregular blebs within veins. In the more distal part of the mineralization, some molybdenite veins are virtually devoid of quartz. Alteration at the vein selvages is weak, apart from localized sericite bleaching. Veins hosted in the mafic and felsic rocks tend to have sharp to slightly irregular contacts, but veins in the mineralized granodiorite may have either sharp or diffuse contacts. The diffuse contacts result from the resorption of the vein and are commonly only discernable by a lack of feldspar or biotite.

Molybdenum and copper grades are broadly correlated, although controlled individually to some extent by host lithology. In some parts of the system, high copper grade is not accompanied by high molybdenum grade. For example, the mineralized granodiorite contains significantly more

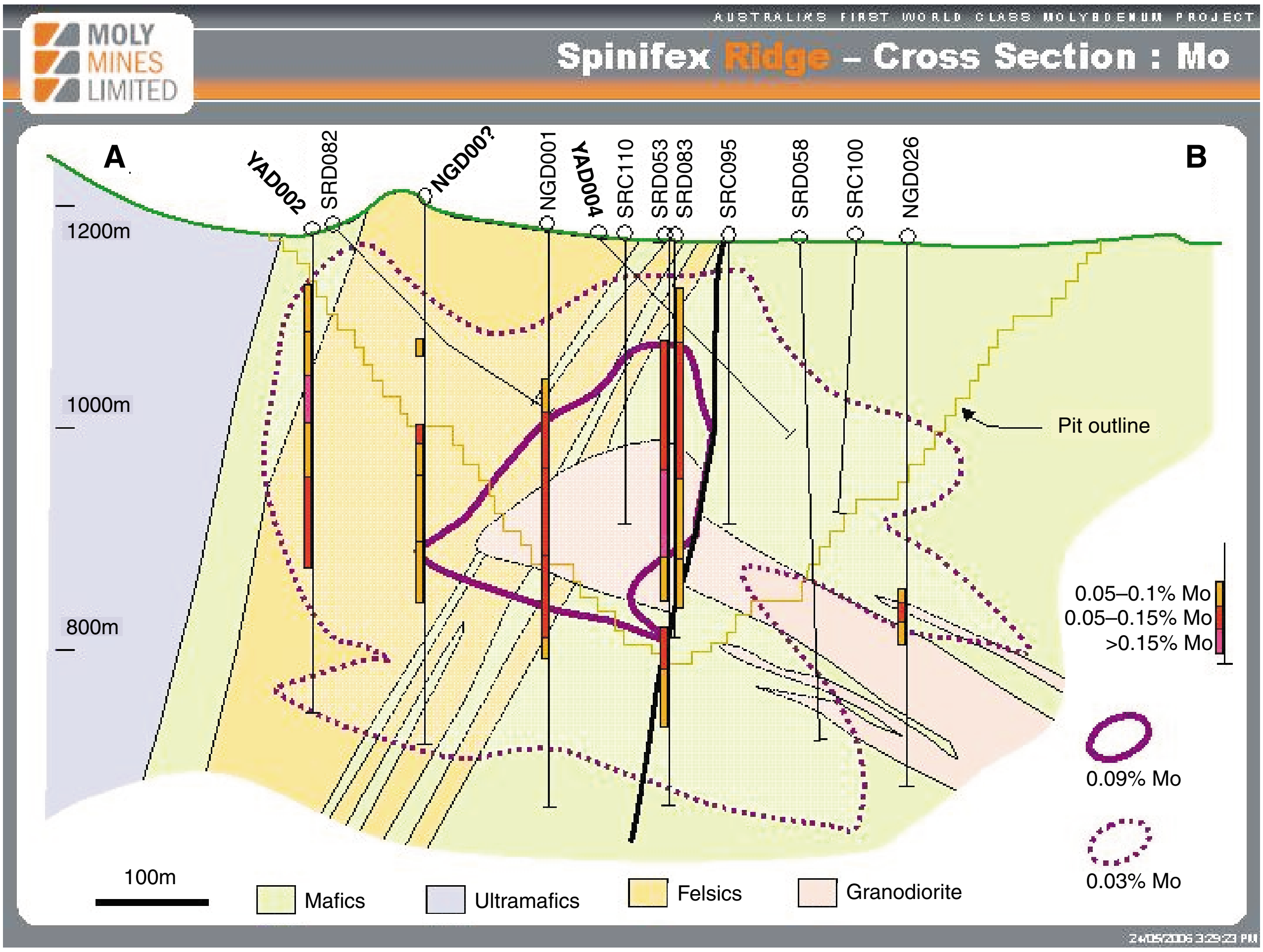


Figure 33. Cross section A–B through the Spinifex Ridge prospect; see Figure 32 for location of cross section (after Moly Mines Limited)

Table 2. Spinifex Ridge summary resource estimate (0.02% Mo cut-off as at 31 March 2008)

Classification	*Tonnes*	*Mo (%)*	*Cu (%)*	*Ag (g/t)*
Measured	207 127 000	0.060	0.099	1.53
Indicated	451 288 000	0.039	0.071	1.15
Measured plus indicated	658 415 000	0.045	0.080	1.27
Inferred	390 494 000	0.039	0.069	1.22

SOURCE: Moly Mines Limited (2007, 2008)

chalcopyrite than molybdenite, especially to the southeast, where it is high in copper content and virtually barren of molybdenum. Further to the northwest, both the copper and molybdenum grades increase within the mineralized granodiorite. At the northwest end of the intrusion, both molybdenum and copper form a high-grade core within and asymmetrically surrounding the mineralized granodiorite. Lower grade molybdenum mineralization is evenly distributed and forms a broader halo around the mineralized granodiorite.

This distribution of mineralization is common in porphyry systems where the higher grades are associated with a strongly silicified core that grades outwards into a potassic zone. At Spinifex Ridge, potassic alteration is preserved as K-feldspar veins, the replacement of plagioclase by K-feldspar, and pervasive biotite alteration at the periphery of the high-grade core. Biotite alteration is especially well developed and preserved in the mafic rocks. Phyllic alteration at Spinifex Ridge is the dominant alteration assemblage and manifests as intense to very intense sericite alteration that surrounds and, in many cases, totally overprints a large portion of the potassic core. In the mineralized granodiorite, strong sericite alteration has in part replaced virtually all other mineral species, excluding quartz. The main sulfide associated with phyllic alteration is pyrrhotite. Any propylitic alteration within the deposit is likely to be observed as chlorite and epidote. Chlorite is commonly seen to replace biotite in the mineralized granodiorite; this may represent a retrograde metamorphic reaction.

The dominant ore-carrying sulfide species are molybdenite (MoS_2) and chalcopyrite ($CuFeS_2$). Silver is closely related to copper, possibly substituting into the matrices of tetrahedrite ($Cu_{12}Sb_4S_{13}$). Non-ore sulfides include pyrrhotite and subordinate pyrite–marcasite (FeS_2), which are commonly in the interstitial regions of pillow lavas. Rare sulfides of interest include sphalerite ((Zn,Fe) S), galena (PbS), and stibnite (Sb_2S_3). One occurrence of native copper has been observed. Tungsten is present as scheelite ($CaWO_4$) and is mainly orange in colour. Disseminated magnetite (Fe_3O_4) is the most significant oxide but also manifests as rare blebs in the mafic minerals. Magnetite is not common in the felsic units.

Locality 17: Lalla Rookh Sandstone, De Grey Supergroup, Western Shaw – Lalla Rookh structural corridor (North Shaw, MGA Zone 50, 744930E 7675610N)

From Moly Mines Ltd, return to the Marble Bar–Port Hedland road, a distance of about 30 km. Turn right and proceed towards Port Hedland for approximately 60 km before turning left on the North Pole road (MGA Zone 50, 745100E 7706900N) 1.5 km east of the Shaw River crossing. Continue down this track for 25 km to Locality 17. Locality 17 is reached by leaving the road at MGA Zone 50, 745100E 7675460N, and driving or walking 200 m northwest.

As you pass from the flat terrain of the Carlindi Granitic Complex into the greenstone hills, you cross the Lalla Rookh – Western Shaw fault, a 150 km-long, curviplanar fault that marks the western boundary of the Lalla Rookh – Western Shaw structural corridor (Van Kranendonk and Collins, 1998; Van Kranendonk et al., 2002; Van Kranendonk, 2008). This corridor represents a 1–30 km-wide zone of dominantly sinistral transpressional deformation that formed during the c. 2.93 Ga North Pilbara Orogeny, one of the last events to affect the craton.

To the east of the fault are coarse sandstones and conglomerates of the Lalla Rookh Sandstone of the Croydon Group (De Grey Supergroup), which were deposited from c. 2.97 to 2.94 Ga (Smithies et al., 1999; Van Kranendonk et al., 2006a, 2007). These rocks represent one of the youngest units deposited on the Pilbara Craton. They consist of coarse clastic sediments deposited in a fluviatile to lacustrine environment, including large alluvial fans (Krapez, 1984; Erikkson et al., 1994). Clasts are predominantly of white quartz and grey to black chert, but include silicified siltstone and, locally, silicified felsic volcanic rocks. Clasts of altered mafic volcanic rocks have been observed in the lowermost parts of the formation.

The Lalla Rookh Sandstone was previously interpreted to have been deposited in a pull-apart basin (Krapez, 1984; Eriksson et al., 2004), but more recent mapping shows that the bounding faults of the Lalla Rookh Synclinorium are a sinistral and dextral pair (Van Kranendonk and Collins, 1998; Van Kranendonk, 2000) and that the formation is part of a much more widespread succession that was deposited during regional extension (3.02 – 2.94 Ga), pre-dating the North Pilbara Orogeny (Van Kranendonk et al., 2002, 2004, 2007).

A low cliff face exposes cross-bedded Lalla Rookh Sandstone dipping northwestwards at 30°.

Day 4: North Pole Dome — c. 3500 Ma depositional environments, and early life

Continue south on the North Pole access road for 20 km to the North Pole mining area.

The North Pole Dome (Plate 2) is a structural dome of dominantly mafic volcanic rocks of the Warrawoona and Kelly Groups intruded by the 3459 ± 18 Ma North Pole Monzogranite (Thorpe et al., 1992a). The stratigraphic succession of the dome forms the Panorama greenstone belt, and ranges from the c. 3500 Ma North Star Basalt in the core of the dome to the 3350 Ma Euro Basalt on its outer flanks. Younging directions, from pillow structures in lavas and graded bedding in volcaniclastic rocks, establish that all the rocks young consistently away from the core in a radial pattern. Despite the >15 km thickness of the greenstone succession, there is only local evidence of tectonic thickening; the accumulated thickness can be explained by subsidence of a local depositional basin between 3480 and 3350 Ma as a result of the vertical deformation processes responsible for of the dome-and-basin pattern of the East Pilbara Terrane (Hickman and Van Kranendonk, 2004). The dip of strata increases from 30° in the core of the dome to 80° towards the dome margins. The rocks have been metamorphosed to prehnite–pumpellyite to greenschist facies, although a lower amphibolite facies contact metamorphic aureole is developed around the North Pole Monzogranite.

The mafic volcanic rocks range between dominantly tholeiitic basalt with N-MORB REE profiles, to komatiitic basalts that, in some places, immediately overly chert horizons in the succession. Geochemical data, and the association of pillow basalts with thin cherts, has led some workers to interpret the succession as a deep-water oceanic assemblage (Isozaki et al., 1997; Kitajima et al., 2001), despite previous compelling evidence that both the Dresser and Panorama Formations are shallow-water deposits (Dunlop et al., 1978; Barley et al., 1979; Lowe, 1980, 1983; Buick and Barnes, 1984; DiMarco and Lowe, 1989; Buick and Dunlop, 1990). The Strelley Pool Formation overlies a regional erosional unconformity (Buick et al., 1995), which establishes that this formation and the overlying Euro Basalt were deposited on continental crust. Because the geochemistry of the Euro Basalt is very similar to that of the Warrawoona Group basalts that underlie the shallow-water Strelley Pool Formation (Smithies et al., 2007), and because the basalt–chert association is also a feature of the Euro Basalt, the interpretation of a deep-water oceanic environment cannot be sustained. Regional geological evidence shows that the Warrawoona Group was deposited on continental crust (Van Kranendonk et al., 2002, 2004, 2006a).

Recent SHRIMP U–Pb zircon dating of a felsic volcaniclastic sandstone from the top of the lowermost chert unit of the Dresser Formation has yielded a maximum depositional age of 3480 Ma (Geological Survey of Western Australia, in prep.), which is interpreted to represent the age of deposition of the unit. The stratiform chert–barite units were fed by a set of chert–barite dykes that were emplaced within and immediately above listric normal growth faults that were active during deposition of the cherts. Above the Dresser Formation, the Mount Ada Basalt includes a thin unit of felsic sandstone and layered blue, grey, and red chert known as the Antarctic Creek Member (Van Kranendonk, 1999, 2000), which contains impact-generated spherules. This unit has yielded a conventional U–Pb zircon age of 3470 ± 2 Ma (Byerly et al., 2002), the same age as the Mount Ada Basalt in the Coongan and Marble Bar greenstone belts.

Overlying the Mount Ada Basalt there is a succession of up to 1.3 km of felsic volcanic and volcaniclastic rocks of the Duffer and Panorama Formations, including up to 150 m of the Marble Bar Chert Member in the far southeastern corner of the dome. A sample of massive rhyolite from the lower part of the felsic volcanic rocks in the southern part of the belt has been dated at 3458 ± 2 Ma, the same age as the North Pole Monzogranite (Thorpe et al., 1992a). A sample of bedded tuff with local pumice layers from the upper part of the volcaniclastic apron of the Panorama volcano in the northwestern part of the dome returned a SHRIMP U–Pb zircon date of 3434 ± 5 Ma (Nelson, 2000b).

Disconformably overlying the Panorama Formation is the Strelley Pool Formation, a shallow-water sedimentary succession containing locally abundant c. 3430 Ma conical stromatolites (Lowe, 1983; Hofmann et al., 1999; Van Kranendonk et al., 2003; Allwood et al., 2006, 2007a,b). In the northeastern and southwestern parts of the dome, the Strelley Pool Formation lies on the underlying rocks across an angular unconformity and contains a basal unit of cobble conglomerate and quartz sandstone that is up to 1 km thick. The Euro Basalt overlies the Strelley Pool Formation, as at Locality 11, and in the Panorama greenstone belt it comprises a succession of interbedded high-Mg and tholeiitic basalts reaching a maximum thickness of 9.4 km.

Locality 18: North Pole Monzogranite at North Pole Dome (North Shaw, MGA Zone 50, 746600E 7663350N)

Continue southeast for ~12 km. Take the right hand fork in the road and continue south for ~2.5 km and turn right onto a road leading to the North Pole mining area. On arrival, report to the mine office (this is a mining area where approval to use haulage roads is required). From the mine office, follow a track leading westwards for ~2.5 km to Locality 19.

Low outcrops of coarse-grained undeformed granite of the North Pole Monzogranite are present at this locality and have been dated at 3459 ± 18 Ma (Thorpe et al., 1992a). A more recent attempt to obtain a date with a smaller error failed because of Pb loss in zircons, despite using a fresh sample obtained by blasting. In neither attempt were older xenocrysts identified. The monzogranite has a coarse-grained to weakly porphyritic texture, and miarolitic cavities have been observed locally, indicating crystallization under relatively shallow crustal conditions.

The North Pole Monzogranite is a 2–3 km-thick subvolcanic laccolith that is coeval with felsic volcanic and volcaniclastic rocks of the overlying Panorama Formation (3458 ± 2 Ma; Thorpe et al., 1992a). The Panorama Formation was fed by a series of porphyritic granite dykes that radiate out from the North Pole Monzogranite across the Mount Ada Basalt (Van Kranendonk, 1999, 2000).

The North Pole battery and cyanide tanks at the hilltop were used in the processing of 400.52 oz of gold from the North Pole mining centre between 1899 and 1931. The Breens Reward mine (one of the mines of the North Pole mining centre) exploited a polymetallic epithermal deposit with gold, copper, silver, and bismuth mineralization. The deposit is associated with a small intrusive felsic porphyry dated at 3431 ± 7 Ma (Thorpe, R, 1992, written communication), the same age as the upper part of the Panorama Formation.

Continue 5 km southwest to a fork, and turn left. Follow this track southwards for 11 km to Antarctic Creek (MGA Zone 50, 740500E 7653900N). Overnight camping.

Locality 19: 'Trendall locality' — stromatolites in the 3430–3350 Ma Strelley Pool Formation (North Shaw, MGA Zone 50, 739600E 7652500N) (protected locality)

From Antarctic Creek, follow the track southwards along the eastern side of the Shaw River for 2 km to the 'Trendall locality' (MGA Zone 50, 739600E 7652500N), where the river has eroded through a ridge of chert and carbonate rocks. Note that this locality is a protected Reserve and collecting from the outcrops is prohibited.

Structures resembling Archean fossil stromatolites were discovered by Dr Alec Trendall of the GSWA in 1984 at this now world-famous locality on the east bank of the Shaw River. The site, which is now informally known as the 'Trendall locality' (Hofmann et al., 1999; Van Kranendonk et al., 2003), was revisited in 1990 by Drs Trendall, Thorpe, and Hickman, who discovered better-preserved examples of stromatolites over a wider area. In 1997, Arthur Hickman showed Prof. Hans Hofmann and Dr Kathleen Grey outcrops that had been photographed during the 1990 trip, and a more detailed investigation resulted in the discovery of additional coniform stromatolites. Removal of chert from above one stromatolitic bedding plane exposed a 1 m-long surface of extremely well-preserved conical stromatolites (Hofmann et al., 1999, fig. 2D), since referred to as 'egg-carton' stromatolites. Because publication of the results of this investigation required revealing the location of such rare fossils, making them vulnerable to unauthorized collecting, the authors arranged for the bedding plane of egg-carton stromatolites to be permanently moved to the Western Australian Museum in July 1999.

Subsequent investigations of the stromatolites at the 'Trendall locality' were undertaken during geological mapping of the area (Van Kranendonk, 2000) and in more detailed studies by Van Kranendonk et al. (2001b, 2003), Allwood et al. (2004, 2006, 2007a,b), and Van Kranendonk (2006, 2007). Van Kranendonk (2000, 2006) suggested correlations between members identified at this outcrop with members recognized by Lowe (1983) in the East Strelley greenstone belt, but also commented on significant differences of stromatolite morphology in the two areas. Van Kranendonk et al. (2001b) presented a detailed map of the 'Trendall locality' and showed the presence of onlapping stromatolite biostromes in laminated carbonates, and of kerogenous stromatolites in overlying sandstones (Marshall, 2007).

Van Kranendonk et al. (2003) presented trace-element geochemical data that provided evidence of stromatolitic carbonate deposition from seawater, rather than from hydrothermal solutions as had previously been proposed (Van Kranendonk and Nijman, 2001; Lindsay et al., 2003). Van Kranendonk et al. (2003) also documented stromatolite growth under moderate energy conditions that supported the shallow-water to episodically exposed setting interpreted by Lowe (1983), which countered a hydrothermal precipitate origin for the stromatolites as suggested by Lindsay et al. (2003, 2005).

Allwood et al. (2004, 2006, 2007a,b) reported on the stratigraphy, facies, and stromatolite morphology along a 15 km-long, relatively continuous strike length of the Strelley Pool Formation northwest and southeast of the Shaw River, including the 'Trendall locality'. Detailed measurements were made on 27 vertical stratigraphic sections to determine spatial and temporal variability among stromatolites, sedimentary facies, and depositional environments. The lateral and vertical facies associations documented in their study record a southeastward-deepening peritidal carbonate platform developed over a drowned landmass. Van Kranendonk (2007) provided further morphological evidence for biogenicity of the stromatolites from this and nearby outcrops.

The excursion includes visits to exposures of the Strelley Pool Formation on both sides of the Shaw River and examination of the basal regional unconformity best exposed on the west bank. A sample collected immediately below this unconformity indicated a SHRIMP U–Pb zircon maximum depositional age of c. 3445 Ma, suggesting possible correlation to the Panorama Formation.

The Strelley Pool Formation at Locality 19 consists of four members (Fig. 34):

1. basal conglomerate, consisting of boulders of black-and-red layered chert;
2. millimetre-scale laminated carbonates, commonly stromatolitic and with widespread development of weakly upward-radiating crystal splays;
3. centimetre-scale bedded black chert, with local siltstone, beds of radiating crystal splays replaced by white quartz, and beds of finely laminated material with centimetre-high domical stromatolites;
4. an unconformably overlying series of very coarse to very fine-grained clastic rocks and mafic ash tuff, deposited as a series of five fining-upwards successions, interpreted to represent a series of receding submarine fan deposits, and containing black beds of kerogenous material with wavy laminated texture interpreted to be stromatolites.

Locality 20: Earth's oldest stromatolites in the c. 3480 Ma Dresser Formation at North Pole Dome (protected locality)

Return to the North Pole mining area and proceed back to the main access road to North Pole. Turn left. Continue north for ~800 m and turn right onto a smaller track. Continue eastwards for 4.5 km to the ruins of Panorama Homestead. As the next three stops are of major geological significance, their precise location is not given.

Note: This is a locality of major geological significance and lies within a proposed Geological Monument area. Hammering or sample collection are not permitted.

This is the type locality of the oldest known fossils in the world, which are within the main lower chert–barite

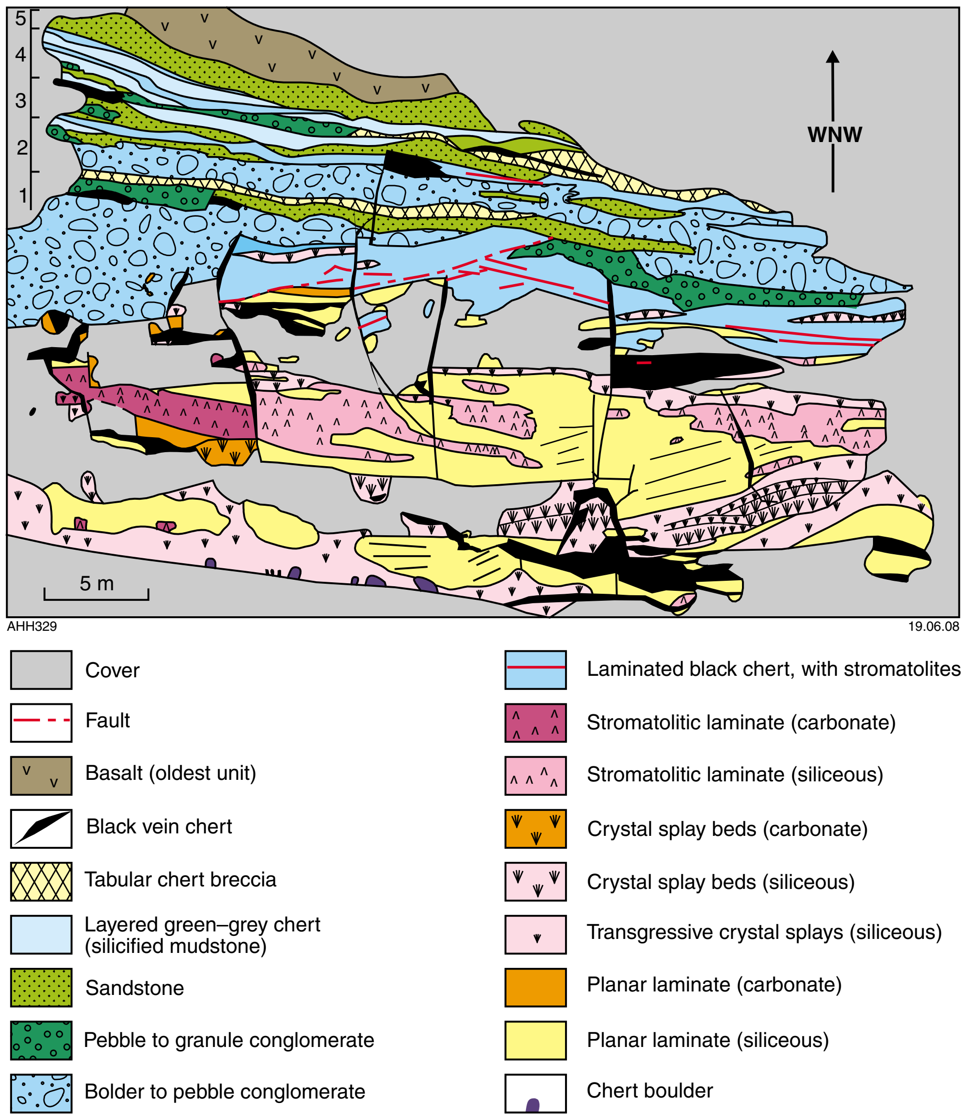


Figure 34. Outcrop map of the 'Trendall locality'. The basal section of the Strelley Pool Formation (composed of sandstone and silicified carbonate rocks) overlies sandstone, mudstone, and tabular chert breccia correlated with the Panorama Formation. The oldest unit exposed is altered basalt of the Mount Ada Basalt. The bar at top left indicates the position of five fining-upward sequences, possibly representing deposits from a series of receding alluvial fans

unit of the 3.49 – 3.48 Ga Dresser Formation in the lower part of the Warrawoona Group (Walter et al., 1980). Previously interpreted to have been deposited in a shallow, quiet-water marine basin (Groves et al., 1981; Buick and Dunlop, 1990), more recent work has shown that deposition of this unit across the North Pole Dome was in a much more dynamic environment that included block faulting on growth faults and hydrothermal fluid circulation (Nijman et al., 1998a; Van Kranendonk and Pirajno, 2004; Van Kranendonk, 2006; Van Kranendonk et al., 2006b).

At this locality, however, the effects of block faulting are not readily apparent and bedding is well preserved, dipping 34° to the east and only broadly warped on east-northeasterly plunging axes (Van Kranendonk, 2006). Otherwise, these rocks are unaffected by penetrative deformation.

The stromatolites here form black-weathered wrinkled mats, smooth broad domes, elliptical low-amplitude domes, small individual columns, coniform types, and mounds of wrinkle-laminated sulfides and barite (Fig. 35). The immediately underlying clastic rocks display well-preserved cross-bedding, indicative of shallow-water conditions, and there are some ripple crests. Bladed crystals of diagenetic barite can be seen to have pushed up what was still wet sedimentary bedding in now silicified carbonates. More massive veins of coarsely crystalline barite are also present in the outcrop.

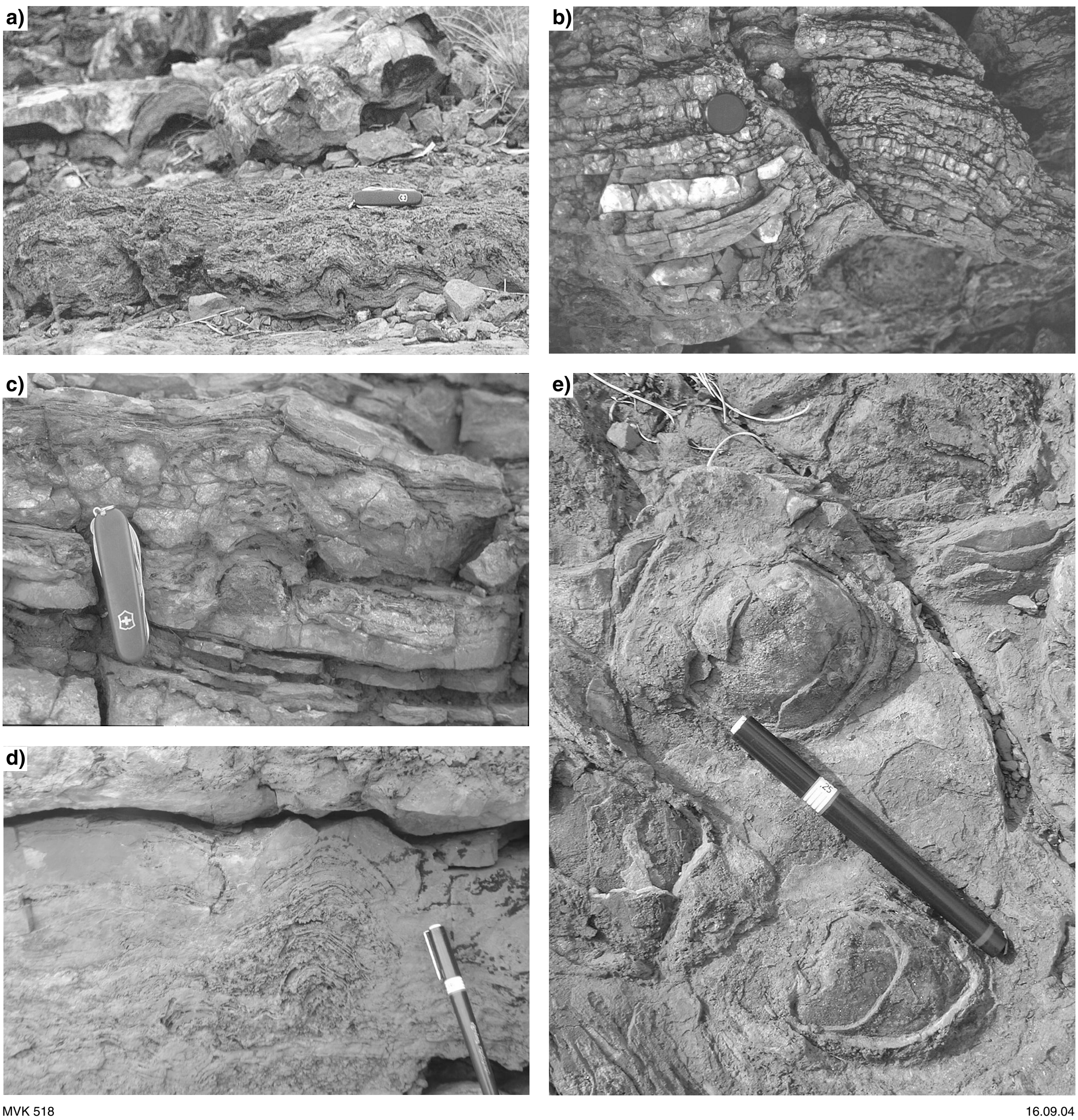


Figure 35. Probable stromatolites from the Dresser Formation, North Pole Dome: a) wrinkly laminated mat and broad smooth domes; b) broad dome of wrinkly laminates with sediment wedges draped on flank; c) small columnar form; d) cross-sectional view of conical form with wrinkly laminations; e) plan view of conical forms. Knife in a) and c) is 15 cm long; diameter of lens cap in b) is ~5 cm; pen cap in d) and e) is 3 cm long (from Van Kranendonk, 2006)

Locality 21: Dresser Formation, primary carbonate and stromatolites overlain by pillow basalt (protected locality)

At this locality, the lower chert unit of the Dresser Formation is only 5 m thick and is exposed in a cliff beside a creek. The creek bed below the cliff exposes altered basalt of the footwall, cut by black chert hydrothermal feeder veins. At the top of the cliff section, primary bedded carbonate is preserved that has a trace-element geochemical signature indicating deposition in seawater of composition similar to that of today (Van Kranendonk et al., 2003). The carbonates have been partly replaced by spherical radiating crystal fans (now filled by aragonite), and by light-green silica. Although the green silica layers may represent original chemical precipitates, thin-section petrography suggests that they more likely represent preferential hydrothermal replacement along porous beds of the carbonate.

Conformably overlying the carbonates there are wrinkly black laminates similar to, and stratigraphically correlated with, the stromatolites viewed at Locality 20. These are directly overlain by basalt with large undeformed pillows.

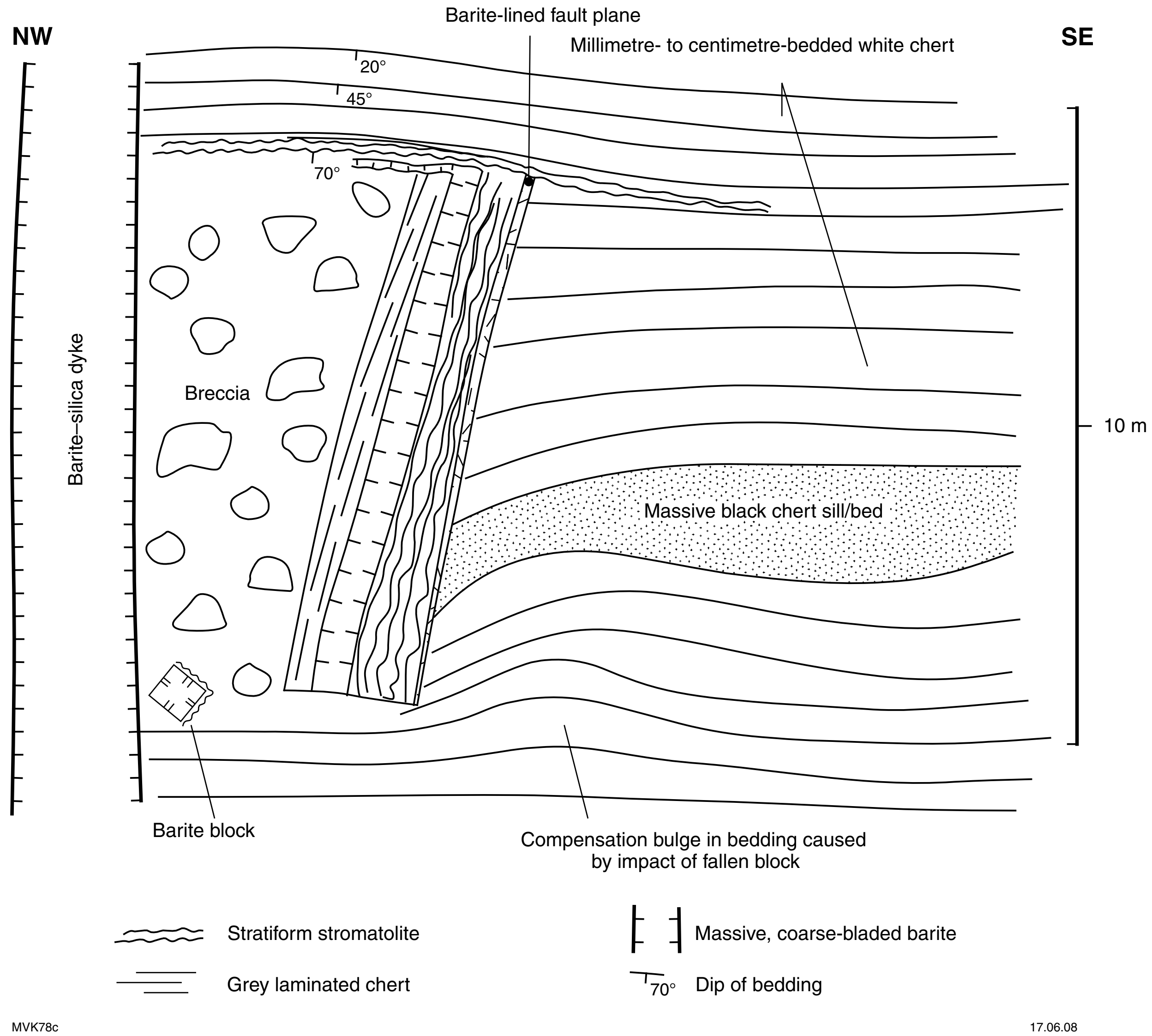


Figure 36. Sketch of a vertical cliff face through bedded chert of the Dresser Formation showing a syndepositional breccia containing coarse barite blocks, which is overlain by stromatolitic laminite and further layers of bedded chert. Note that the stratiform stromatolites are concentrated around areas of vein barite, suggesting colonization of sites near hydrothermal fluid flow and the later vein of coarsely crystalline barite (from Van Kranendonk, 2006)

The lack of evidence of strain in the rocks underlying and overlying the contact between the laminates and the basalts suggests that this is a primary depositional contact, whereas the relative thinness of the chert unit here, compared with Localities 20 and 22, suggests that this was a horst block during deposition of the chert unit.

Locality 22: Dresser Formation, barite mound and diamictite (protected locality)

Nijman et al. (1998a) showed that the Dresser Formation was deposited during growth faulting accompanied by exhalation of sulfate- and silica-rich hydrothermal fluids. This model was further developed by Van Kranendonk (2006) who showed that deposition was in a much more dynamic environment than previously supposed, probably within a felsic caldera setting. Part of the evidence for this change of view from the previous model of deposition in quiet, shallow-water lagoonal conditions (Groves et al., 1981; Buick and Dunlop, 1990) is evident at this locality.

Locality 22 provides a cross section through the chert–barite unit of the Dresser Formation (Fig. 36). Most of the exposed unit is composed of gently dipping grey-and-white cherts. Recent drilling to beneath the surficial weathering profile, and rare surface outcrops, show that most of the cherts were originally carbonate sedimentary rocks. However, immediately along strike from the bedded cherts there is a massive block of coarsely crystalline barite and wrinkly laminate, which is oriented at a high angle to bedding, and there are numerous large, subrounded to angular blocks in a coarse breccia. More layered grey-and-white chert overlies the breccia, and bedding indicates recovery up section from steep dips back to shallow dips. Higher in the section, this unit is overlain by a coarse diamictite with clasts of barite and wrinkly laminated rocks. All of these rocks are cut by a 2 m-wide vertical vein of coarsely crystalline barite, indicating the repetitive nature of barite fluid infiltration into the Dresser Formation system.

This outcrop shows the dynamic nature of the depositional environment (Fig. 36), which varied from periods of quiet-water carbonate deposition to very high-energy events associated with block faulting, growth-fault development, and hydrothermal fluid circulation. The breccia unit contains a matrix with devitrified felsic glass shards, indicating formation during periods of felsic eruption. A sample of one such tuffaceous rock, obtained from drillcore through the formation to the south of Locality 22, contained a large zircon population with a SHRIMP U–Pb age of 3525 ± 2 Ma (Van Kranendonk et al., 2007). However, the sample also contained a zircon grain dated at c. 3480 Ma, which is therefore interpreted to be the depositional age of the Dresser Formation.

Locality 23: Evidence of a 3470 Ma asteroid impact event, Antarctic Creek Member, Mount Ada Basalt (North Shaw, MGA Zone 50, 759500E 7660600N)

Return to the Panorama Homestead road and turn right. Drive eastwards across Miralga Creek, and continue for ~9 km south to Locality 22.

The Antarctic Creek Member of the c. 3.47 Ga Mount Ada Basalt is a 1–15 m-thick unit of laminated grey, white, and red chert and less common sandstone, conglomerate, breccia, and felsic volcaniclastic rocks.

Lowe and Byerly (1986) reported millimetre- to sub-millimetre-scale silicified spherules exhibiting quench-crystallization and glass-devitrification textures in sandstone within the Antarctic Creek Member. Byerly et al. (2002) reported a $^{207}Pb/^{206}Pb$ age of 3470.1 ± 1.9 Ma for euhedral zircons from the spherule-bearing unit, which is within the error margin for the age of the Mount Ada Basalt and coeval Duffer Formation (3474–3463 Ma, Van Kranendonk et al., 2002). Of 30 zircons of 50–100 μm size that were extracted from 2 kg of rock, two grains yielded ages of c. 3510 Ma, suggesting derivation of some of the clastic material from the older components of the Warrawoona Group (Coonterunah Subgroup, Buick et al., 1995; Van Kranendonk et al., 2002, 2006a). Byerly et al. (2002) correlated the spherule-bearing unit with a similar unit in the South African Hoogenoeg Formation (Barberton greenstone belt, Kaapvaal Craton), which yielded a $^{207}Pb/^{206}Pb$ zircon age of 3470.4 ± 2.3 Ma.

Lowe and Byerly (1986) identified spherule-bearing chert and arenite over a distance of 1 km along strike in the Antarctic Creek Member; Glikson et al. (2004) extended this to about 15 km along strike in the northeastern part of the North Pole Dome (Plate 2). Spherules are present within at least two beds of 15–50 cm-thick current-deposited arenite and form <10% of the arenite, but are locally in higher proportions. Spherules are mostly 0.10 – 0.75 mm in diameter. Irregular particles, compound spherules, dumbbell-shaped particles, and broken spherules are present. Spherules within chert show high to very high sphericity. In the arenites, the silicified, partly broken, corroded and matrix-resorbed state of scattered spherules renders their identification more difficult.

Lowe and Byerly's (1986) spherule-bearing chert–arenite unit at the type locality of the Antarctic Creek Member (MGA Zone 50, 759685E 7663583N) consists of ~14 m-thick massive to finely laminated chert, which contains up to three units of interbanded chert, recrystallized feldspathic arenite with millimetre-scale chert fragments, and gritstone to intraclast conglomerate consisting of recrystallized arenite grains and millimetre- to several centimetre-scale, mostly layer-oriented chert fragments in a siliceous matrix. Sub-millimetre-scale, poorly to moderately well-preserved spherules are scattered within the matrix of the arenite, gritstone, and intraclast conglomerate and are also present in stratiform bands and lenses within the chert. There are some microkrystite spherules both within chert fragments and in the matrix of the same arenite–microconglomerate units (Fig. 37), attesting to multiple stages of spherule deposition.

Lowe and Byerly (1986) and Byerly et al. (2002) documented quench-like spinifex-textures, cross-cutting crystallites, and fan-shaped radiating devitrification features replaced by quartz, mica, and iron oxide in the spherule-bearing arenite. Other components of the spherule-bearing arenites include altered lithic fragments (many of which include microlite and lath-like pseudomorphs) in matrices of chert and sericite with minor amounts of chlorite. Lowe and Byerly (1986) identified little or no volcanic detritus in the arenite–chert units. In arenite samples, Glikson et al. (2004) observed coexisting lithic microlitic to microporphyritic particles of possible volcanic origin (Fig. 37a,b) and distinct highly spherical spherules with quench and devitrification textures (Fig. 37c,d).

Day 5: From North Pole to the West Pilbara Superterrane and the De Grey Superbasin in the west Pilbara

Return to the North Pole access road and drive north 45 km to the Marble Bar – Port Hedland road. Turn left and drive approximately 100 km to Port Hedland. From Port Hedland, drive 185 km westwards along Highway 1 to Mount Hall, 6 km east of Roebourne. Locality 24 provides roadside exposures of komatiite flows in the c. 3270 Ma Ruth Well Formation of the Roebourne Group.

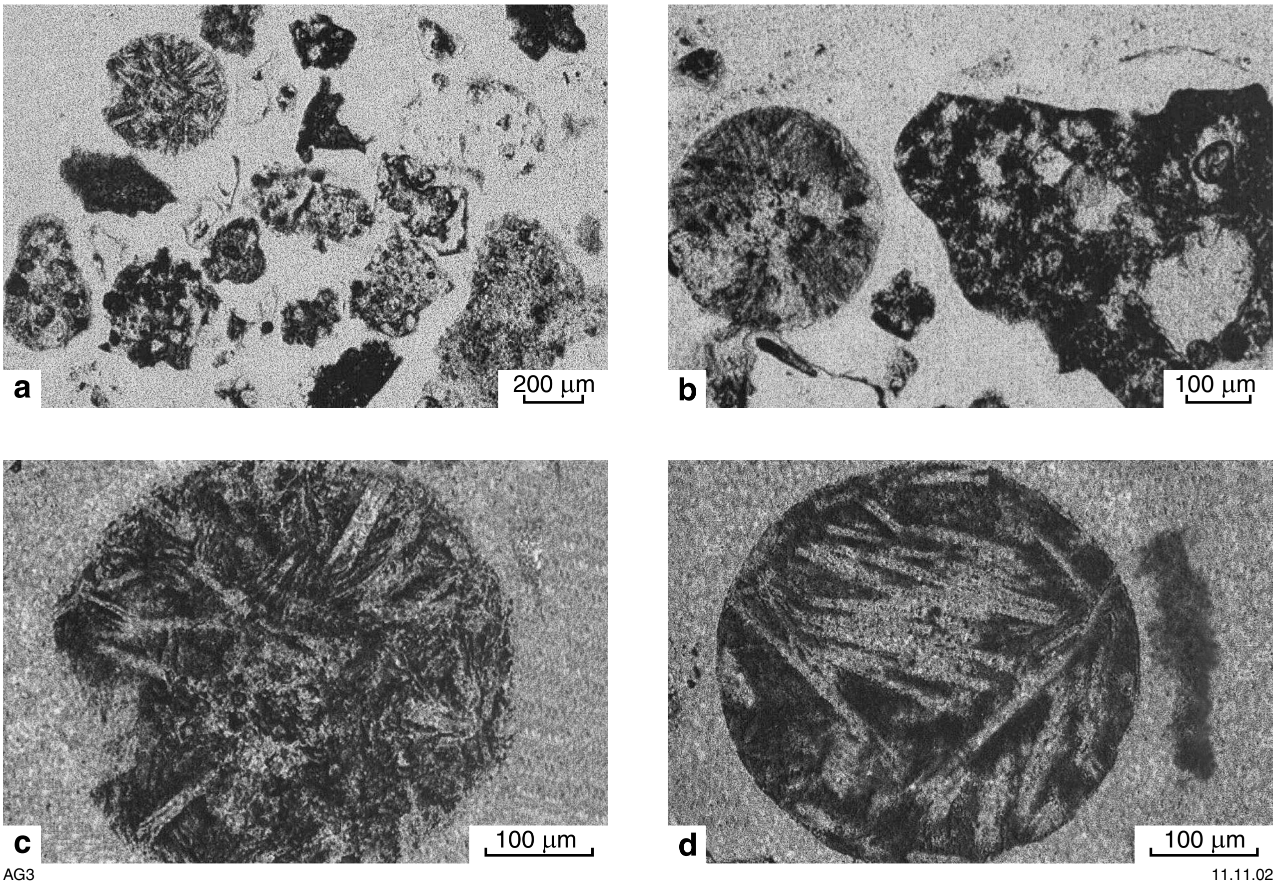


Figure 37. Photomicrographs of spherules from the Antarctic Creek Member (from Glikson and Vickers, in prep.): a) arenite consisting of clouded and altered irregularly shaped fragments, many of probable volcanic origin, and including a spherule showing quench and devitrification textures, set in chert matrix; b) spherule showing quench and devitrification textures, next to a microporphyritic fragment of probable volcanic derivation; c) enlarged image of the spherule shown in (a). Note the inward-radiating quench texture at lower right; d) spherule showing needle-like pseudomorphs of quartz, probably after feldspar. The dominance of parallel orientation suggests this may have been a microtektite rather than microkrystite

Locality 24: Komatiite flows in the c. 3270 Ma Ruth Well Formation, Roebourne Group (Roebourne, MGA Zone 50, 520700E 7701400N)

Locality 24 is on a bend of Highway 1, 6 km east of Roebourne.

Roadside exposures at Mount Hall provide one of the few accessible exposures of well-preserved sheaf and random spinifex-textured ultramafic komatiite flows in the oldest formation of the Roebourne Group, the Ruth Well Formation. This formation is approximately the same age as the Kunagunarrina Formation of the Sulphur Springs Group in the East Pilbara Terrane, and may have been rifted into its present position in the west Pilbara during the 3220–3165 Ma rifting event (see **Deformation events**). Olivine plates (pseudomorphed by serpentine and tremolite) are up to 30 cm long. Microscopic examination of the sheaf spinifex texture reveals a highly magnesian rock of mainly tremolite, serpentine, chlorite, and a pale-yellow birefringent phyllosilicate (probably vermiculite). Serpentine and tremolite or vermiculite have replaced olivine blades. Tremolite and serpentine also form very fine intergrowths, and chlorite forms flakes of random orientation. Opaques are disseminated through the rock as anhedral grains and granular aggregates, and as discontinuous linings along narrow fractures or veinlets. Some of the opaques have a translucent, reddish brown character around their outer margins, suggesting they are iron oxides such as hematite. Layers of different spinifex texture indicate that the ultramafic flows are about 2 m thick.

The komatiitic flows are overlain by a thin unit of quartzite and ferruginous chert, the bedding of which shows that the succession dips 30° to 50° southwards. Farther south, this is overlain by amphibolite (metabasalt), with gabbro, pyroxenite, and serpentinized peridotite of the Andover Intrusion. At Mount Wangee, 7 km to the north, similar spinifex-textured ultramafic flows are overlain by massive and pillowed basalts of the Regal Formation, but the succession in that area dips northwestward. There are no Archean exposures between these localities, but aeromagnetic data indicate that the Harding Granitic Complex underlies Cenozoic deposits of the coastal plain. The opposing inclinations of the

Ruth Well Formation at Mount Hall and Mount Wangee indicate that the Harding Granitic Complex occupies the core of an anticline in the Ruth Well Formation.

Locality 25: Cleaverville Formation at Cleaverville Beach, and recent geoscientific drilling (ROEBOURNE, MGA Zone 50, 502300E 7716300N)

From Roebourne, drive 12 km west on Highway 1 to the turnoff signed to Cleaverville Beach (MGA Zone 50, 503600E 7705350N). Follow this road for approximately 11 km and then turn right to the beach. Follow the graded gravel road 2 km to its end at the most easterly part of the beach (MGA Zone 50, 502300E 7716300).

Significance of the Cleaverville area

Cleaverville is the type area of the Cleaverville Formation, which Hickman (1983) correlated with the Gorge Creek Group of the east Pilbara. He interpreted the Cleaverville Formation to be a major stratigraphic marker that extends across most of the northern Pilbara Craton; that interpretation is still maintained. SHRIMP U–Pb zircon geochronology (Nelson, 1998) indicates that all outcrops correlated with the Cleaverville Formation in the west Pilbara, and outcrops of similar banded iron-formation in the extreme northwestern part of the east Pilbara, are the same age (c. 3020–3015 Ma). Several attempts to precisely date the major iron formations of the Gorge Creek Group in the east Pilbara have been unsuccessful. Extension of the Cleaverville Formation across the northern Pilbara Craton establishes that terrane accretion was completed before 3020 Ma. As noted in **Part 1**, the time of accretion is currently interpreted to be 3070 Ma, coinciding with the Prinsep Orogeny.

In the west Pilbara, the Cleaverville Formation is up to 1500 m thick, and is composed of banded iron-formation, ferruginous chert, grey–white and black chert, shale, siltstone, and minor amounts of volcanogenic sedimentary rocks. Because the Cleaverville Formation contains magnetite-bearing iron formation it is readily identifiable on regional aeromagnetic images, and the formation includes iron ore deposits southeast of Cleaverville. In the Roebourne–Dampier area, aeromagnetic images confirm previous conclusions from geological mapping (Hickman, 1980) that the Cleaverville Formation is folded around the Prinsep Dome (Fig. 12). Outcrops at Cleaverville lie on the northwestern limb of this fold, whereas extensive outcrops of the formation in the Roebourne–Wickham area are on the southeastern limb. Southwest of Roebourne, almost continuous outcrops of the Cleaverville Formation extend 30 km along the axial region of the Roebourne Synform. Southwest from Cleaverville, the formation can be traced for a distance of 70 km through Karratha to Mount Regal, Maitland River, and farther west to Devil Creek (Hickman and Strong, 2003; Hickman and Smithies, in prep.). The only other major outcrop of the Cleaverville Formation in the west Pilbara is south of the Sholl Shear Zone at Mount Ada, 20 km south of Roebourne (Hickman, 2002). Correlation of the Mount Ada iron formations with the Cleaverville Formation is based on similarities in lithology and thickness, and on precise SHRIMP U–Pb zircon geochronology.

Previous tectonic interpretations of the Cleaverville area

Some workers have interpreted the Cleaverville area as a terrane that is separate from the rest of the west Pilbara. Ohta et al. (1996) interpreted the geology of the Cleaverville area to indicate accretion of a MORB–trench succession onto the northwestern margin of the Pilbara Craton. Kiyokawa and Taira (1998) also interpreted the area as an accreted unit, but showed evidence supporting their conclusion that the Cleaverville succession formed in an oceanic island-arc environment. There is a narrow belt of clastic sedimentary rocks along the southeastern margin of the Cleaverville area, which Kiyokawa and Taira (1998) interpreted as a syn- to post-orogenic basin succession in a suture zone. Krapez and Eisenlohr (1998) used information from Kiyokawa (1993) and Kiyokawa and Taira (1998) to propose that the Cleaverville succession formed on a platform within an intra-arc basin, but emphasized that relationships to successions farther southeast indicate that it was not part of an exotic terrane.

Ohta et al. (1996) based their interpretation that the Cleaverville area contains a tectonically repeated (seven tectonic slices) MORB–trench succession on lithological associations. They recognized a single succession of pillow basalt overlain by chert and banded iron-formation, with an upper assemblage of volcanic and clastic rocks. They interpreted this sequence to be indicative of an oceanic plate moving from a mid-ocean ridge (basalt) setting, through deep-sea pelagic (chert and banded iron-formation) to hemipelagic to a trench setting. However, the interpretation given by Kiyokawa and Taira (1998) did not include the complex tectonic duplication proposed by Ohta et al. (1996), and they interpreted the Cleaverville succession to represent three volcano-sedimentary cycles. Each cycle was said to commence with basaltic volcanism, followed by rhyolite volcanism, and to be terminated by deposition of chemical sediments and volcano-sedimentary rocks.

Ohta et al. (1996) and Kiyokawa and Taira (1998) assumed that the chert and banded iron-formation units of the Cleaverville Formation were deposited in deep-water oceanic environments remote from any influx of continental material. However, Sugitani et al. (1998) used evidence provided by sedimentary structures and mineralogy to interpret the Cleaverville Formation of the adjacent Roebourne area as a shallow-water deposit. It is also significant that the Cleaverville Formation contains clastic zircons dated at 3461 ± 8 and 3287 ± 17 Ma (Nelson, 1998). These ages coincide with those of the Warrawoona Group in the East Pilbara Terrane, and the Roebourne Group and Karratha Granodiorite of the Karratha Terrane, respectively, suggesting deposition above, or adjacent to, much older continental crust.

Present tectonic interpretation of the Cleaverville area

The regional geology of the northern Pilbara Craton precludes interpretation of Gorge Creek Group depositional environments in terms of oceanic and convergent margin settings. The chert, banded iron-formation, and clastic sediment sequences in the Gorge Creek Group unconformably overlie pre-3020 Ma crust that is at least 20 km thick (Hickman, 2004a; Van Kranendonk et al., 2006a). In the northeast Pilbara, the Gorge Creek Group rests directly on major erosional unconformities (Dawes et al., 1995; Williams, 1999). The Cleaverville Formation is not confined to the Cleaverville area, as would be required by the accretionary models. The pillow basalt unit that underlies the Cleaverville Formation in the west Pilbara is assigned to the c. 3200 Ma Regal Formation (Hickman, 1997; Van Kranendonk et al., 2006a).

Beach exposures, and geoscientific drilling

Wave-cut platforms at Cleaverville Beach provide excellent, but weathered, exposures of steeply inclined banded iron-formation and interbedded shale and mudstone. Deformation of the succession includes tight-to-isoclinal folding and minor shear zones. In August 2007, Japanese researchers, mainly from Kyushu University, in a collaborative project with GSWA, drilled three diamond drillholes through the Cleaverville Formation at Cleaverville Beach. Two of these holes, CL-1 and CL-2, were close to Locality 24, and the third, DX-1, was drilled 1.5 km to the west. As the succession is almost vertical, all holes were inclined at the lowest practical drilling angle of 52° from horizontal. The two CL holes were inclined to the southeast and the DX hole to the northwest. The aim was to investigate two units previously mapped by Kiyokawa (1993) and Kiyokawa and Taira (1998) as the 'Snapper Beach Formation' (CL holes) and the 'Dixon Island Formation' (DX hole) (both informal stratigraphic names). Surface mapping (Kiyokawa and Taira, 1998) indicated that the drilled lower section of the 'Snapper Beach Formation' is composed of laminated black-and-white chert, and ferruginous chert, whereas the 'Dixon Island Formation' is exposed at the surface as black-, white-, and red- laminated chert, black-and-green chert, and red shale.

The main objectives of the drilling were to establish the detailed stratigraphy of the sections drilled and to search for Archean microfossils and other evidence of Archean biological activity. Near-surface alteration at Cleaverville extends to vertical depths of between 30 and 50 m, making the upper sections of each drillhole unsuitable for detailed study. Below the alteration zone, diamond drillcore was collected for geochemical investigation (major, trace, and rare earth element analysis; and organic carbon, nitrogen, and sulfur isotope studies), petrological examination, and paleomagnetic measurements.

Hole CL-1, drilled stratigraphically above CL-2, commenced in altered black and grey shale, and below 40 m passed stratigraphically upwards into a thick succession of finely bedded black shale with thin layers of fine sandstone, with several altered intervals of reddish black shale. Fragmentation between the black and the reddish black shale units suggests oxidation of pyrite along faults. At 105.5 m, the hole was terminated in black shale containing pyrite nodules and beds of graded and cross-laminated sandstone. Hole CL-2 commenced in altered basalt, but below 47 m passed into fragmented black and grey shale that continued to a depth of 92 m. Hole DX-1 was drilled in altered pillow basalt from 48 to 69 m. From 69 to 148 m the hole intersected a black shale succession that the research team subdivided into well-laminated black shale, laminated black shale with thin pyrite layers, and massive black shale. Zones of fragmentation extend to a downhole depth of approximately 110 m. In summary, the drilling revealed that the various types of chert mapped by Kiyokawa and Taira (1998) at the surface are, at depth, black shales and silstones with varying pyrite content.

Results of research on the drillcore will be presented by Shoichi Kiyokawa at the Australian Earth Sciences Convention in July 2008.

From Cleaverville, return to Highway 1 and drive 35 km west to Dampier.

Day 6: West Pilbara Superterrane

Locality 26: Metamorphosed basalt of the Regal Formation at Karratha (DAMPIER, MGA Zone 50, 484200E 7706350N)

Locality 25 is on the ridge overlooking Karratha, and close to the water tanks northwest of the Karratha Tourist Bureau (DAMPIER MGA Zone 50, 484200E 7706350N).

The ridge south of Karratha provides excellent exposures of metabasalt of the Regal Formation. The Regal Formation is a 2 km-thick slab of c. 3.2 Ga oceanic crust that was thrust onto the 3.27–3.25 Ga Karratha Terrane. The thrust, known as the Regal Thrust (Hickman, 1997), has been mapped over a wide area of the west Pilbara (see **Part 1 — Regal Terrane**). Locally preserved pillow structures show flattening parallel to the east–west-trending tectonic foliation in the basalt, but some better-preserved pillows are sufficiently well preserved to indicate northerly younging. East-northeast from Karratha, the Regal Formation is overlain by the Cleaverville Formation, as at Cleaverville (Locality 24). A swarm of northerly trending dolerite dykes is exposed on the ridge, but appears to terminate at the Regal Thrust, approximately 2 km to the south. The age of the dykes has not been determined and it is uncertain if they pre-date the thrust. Rare northerly trending dolerite dykes are also exposed in the Karratha Terrane, but it is unknown if these are the same age as the Regal Formation dykes.

Locality 27: Komatiite flows in a tectonic slice along the Regal Thrust (DAMPIER, MGA Zone 50, 482500E 7704900N)

From the Karratha Tourist Bureau, drive 2 km south to the outdoor cinema. Immediately past the cinema, turn right along a well-used track leading westwards. After 2 km (MGA Zone 50, 482500E 7704650N), leave the track and drive 250 m northwards to Locality 27 on the ridge.

At Locality 27 there are outcrops of well-preserved, spinifex-textured flows of ultramafic lava. These flows are preserved within a tectonized lens of serpentinized peridotite, which is probably a metamorphosed ultramafic intrusion along the zone of thrusting. The komatiite flows probably belong to the basal part of the Regal Formation. The best exposures are on the southern side of the ridge and west of the end of a valley in the ultramafic succession.

Individual ultramafic lava flows are 1–2 m thick and dip northwestwards at 10–30°. The most visually striking feature of the flows is the excellent development of sheaf-spinifex texture in which blades of serpentinized olivine are up to 0.5 m long. The sheaf-spinifex zones of each flow are overlain by zones of random spinifex texture, and these underlie fractured, aphanitic, and vesicular flow tops. The basal parts of the flows are poorly exposed due to rubble from overlying flow components, but consist of massive serpentinite with locally visible olivine-cumulate texture. Thin sections from similar rocks at nearby localities suggest that the olivine plates of the sheaf-spinifex zones are likely to have been completely replaced by serpentine and tremolite.

Locality 28: Mylonite in the Regal Thrust southeast of the Karratha Granodiorite (DAMPIER, MGA Zone 50, 492000E 7696800N)

Return to the Karratha cinema, turn right, and drive to Highway 1. Turn right, and after 50 m turn sharp left on a road that follows the water pipeline southeastwards. Follow this road for 7 km (MGA Zone 50, 491200E 7696800N). Leave the track about 100 m southeast of the creek, and drive east for about 600 m to Locality 28.

Locality 28 provides excellent exposures of the Regal Thrust on the southern limb of the Prinsep Dome. Close examination of the chert-like rock forming the low cliffs reveals a finely laminated silicic mylonite that has been isoclinally folded. Figure 38 shows a particularly good outcrop of the mylonite, and Figures 39 and 40 provide close-ups of the isoclinal folding. Parts of the outcrop (Fig. 40) show that the isoclinal folds have been refolded by tight-to-isoclinal folds. Plunge of the early isoclines is generally shallow (up to 30°) east or west, and the prevailing dip of the mylonite is 60–80° south.

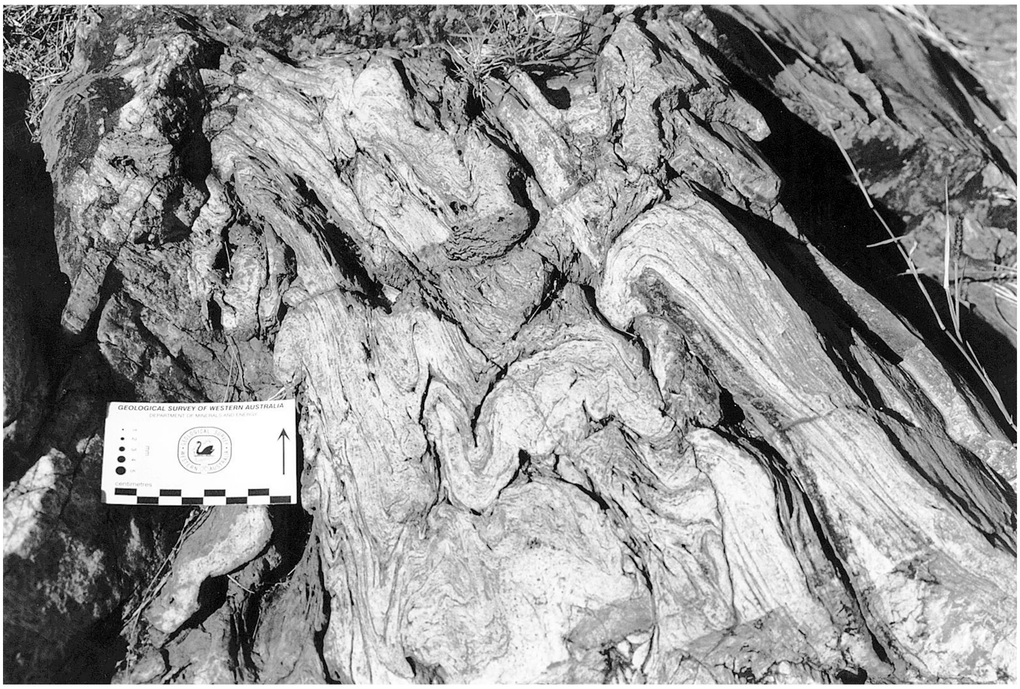
Above the mylonite to the south is the 2 km-thick metabasalt succession of the Regal Formation. Underlying the mylonite to the north is a mixed lithological assemblage of ultramafic schist, felsic schist, volcanogenic metasedimentary rocks, quartzite, and chert of the Nickol River Formation. Outcrops of the mylonite and schist between these formations have been mapped over a distance of 25 km on the southeastern limb of the Prinsep Dome. Between Karratha and the Mount Regal area, the Regal Thrust has been traced for 30 km. Although this continuity of outcrop is broken by a combination of structural complexity and limited exposure at the northeastern and southwestern ends of the dome, it is evident that the Regal Thrust is folded around the dome. Detailed measurement of small-scale structures have not been made but would be expected to reveal the primary orientation of isoclinal folds and lineations in the Regal Thrust. Preliminary observations are that lineations, which in some outcrops appear to be stretching lineations, generally plunge northeast or southwest.

4IAS AHH FIG27 10.08.01

Figure 38. Intensely folded silicic mylonite at Locality 28 (after Hickman et al., 2001)

4IAS AHH FIG28 10.08.01

Figure 39. Close-up view of isoclinal folds in the mylonite shown in Figure 38 (after Hickman et al., 2001)



4IAS AHH FIG29 10.08.01

Figure 40. Close-up view of a refolded isoclinal fold in the mylonite shown in Figure 38 (after Hickman et al., 2001)

Locality 29: Sholl Shear Zone between the Karratha and Sholl Terranes at Nickol River (Dampier, MGA Zone 50, 494800E 7689600N)

Follow the water pipeline southeastwards for 11 km to the crossroad intersection with the Roebourne–Cherratta road (MGA Zone 50, 499700E 7689200N). Turn right, and follow the Roebourne–Cherratta road west for about 4 km to a 4WD track about 200 m before the crossing over Nickol River. Take the track to the right, drive north for about 1 km, then turn left on an indistinct track for 200 m to rock pavements in the bed of Nickol River.

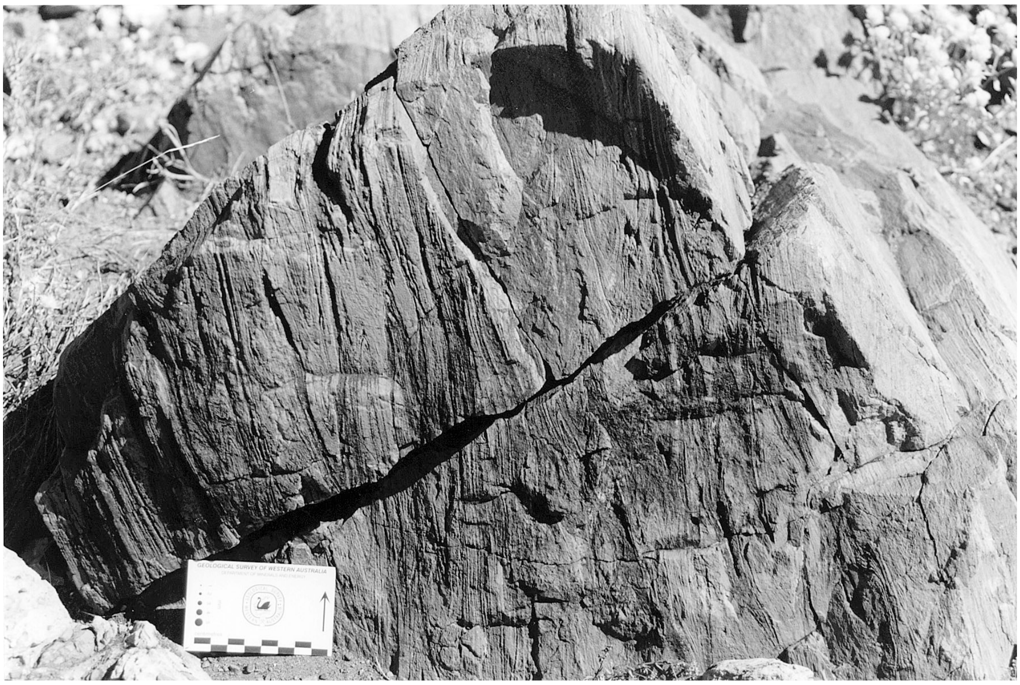
Rock pavements in the Nickol River provide excellent exposures (Fig. 41) of the northern section of the Sholl Shear Zone. The mylonite is dominantly silicic and represents extremely sheared granitic rocks, but there are also layers of amphibolite. The mylonitic lamination is folded by tight west-plunging Z-folds, which may be related to late dextral movement, and isoclinal folds (Fig. 42). All of these structures are displaced by late brittle fractures (Fig. 43) that are locally filled by pseudotachylite. These are probably related to a post-lower Fortescue Group (post-2725 Ma) north–south compressional event that produced a conjugate fault system in the Dampier–Roebourne area (Hickman, 2001b).

Smith et al. (1998) reported a zircon age of 3024 ± 4 Ma from the mylonite, which is interpreted to be the age of the local granitic precursor. Shear-sense indicators elsewhere (Hickman, 2001b) indicate that the dominant early shearing was sinistral and may have been partly contemporaneous with the Orpheus Supersuite, which intrudes the Sholl Shear Zone and outcrops sporadically for several kilometres to the north. Late movement (post-3.01 Ga Whim Creek Group) was clearly dextral because the Whim Creek Group and the Caines Well Granitic Complex are displaced by about 30–40 km across the shear zone. Metamorphic grades immediately north of the shear zone range from upper greenschist to amphibolite facies, whereas grades to the south are greenschist facies. However, both the timing of the implied north-side-up movement and the amount of early sinistral movement are unclear. This is largely because, in the Dampier–Roebourne area, the Maitland River Supersuite intruded large areas of greenstones north and south of the shear zone. These granitic intrusions appear to have totally obscured pre-3.01 Ga displacements of the stratigraphy.

Evidence that early movement on the Sholl Shear Zone was much greater than the 30–40 km late dextral movement is provided by Sm–Nd data (Sun and Hickman, 1998) from the Sholl Terrane and the Cherratta Granitic Complex to the south, and from the Karratha Terrane to the north. Sm–Nd data south of the shear zone show no evidence of underlying c. 3.48 Ga crust, whereas Sm–Nd data north of the shear zone do. This difference in age of underlying crust on either side of the Sholl Shear Zone supports evidence from the stratigraphy, and from the width of the mylonite zone, that the early sinistral movement was probably at least 200 km.

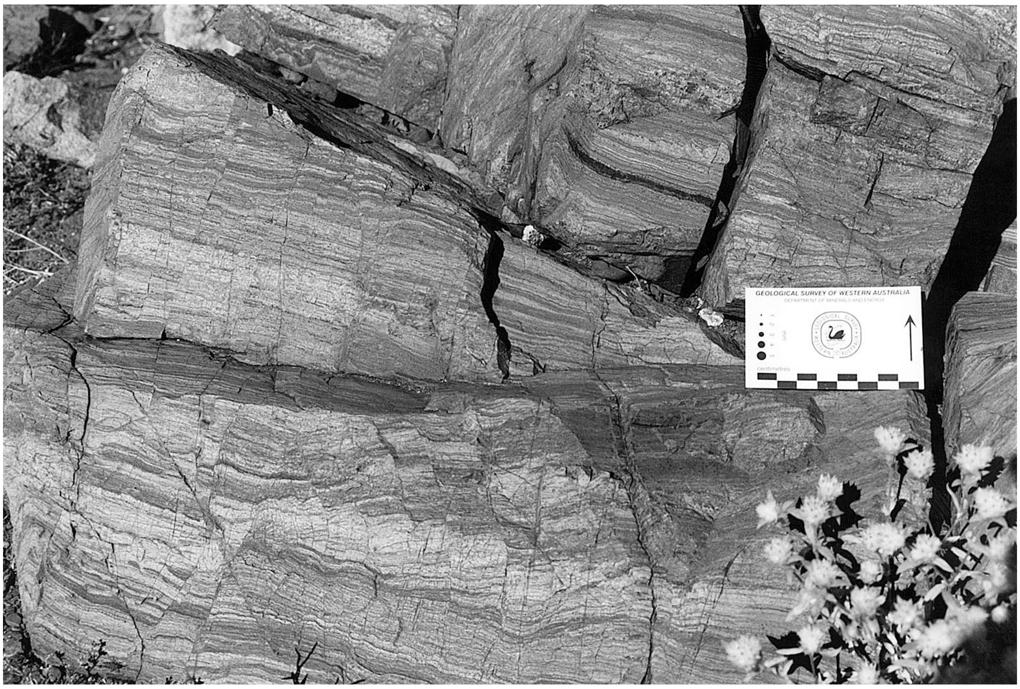
4IAS AHH FIG22 10.08.01

Figure 41. Mylonite of the Sholl Shear Zone in the Nickol River at Locality 29 (after Hickman et al., 2001)



4IAS AHH FIG25 10.08.01

Figure 42. Isoclinal folding of mylonitic foliation at Locality 29 (after Hickman et al., 2001)



4IAS AHH FIG26 10.08.01

Figure 43. Close-up view of compositional banding in the mylonite at Locality 29. Late brittle structures that displace the mylonitic fabrics are probably post-2725 Ma (after Hickman et al., 2001)

Locality 30: Pillow basalt in the Bradley Basalt (Roebourne, MGA Zone 50, 508900E 7682000N)

Return to the water-pipeline road, turn right, and continue southeast for 14.5 km where the road bends sharply to the left. A few hundred metres past the bend, a disused track leads into a gravel pit (MGA Zone 50, 508900E 7681850N). Locality 29 is on the western side of the small hill 150 m north of the gravel pit.

This locality provides excellent three-dimensional exposures of pillow basalt (Fig. 44) in the Bradley Basalt of the Whundo Group. Pillow structures show that the lava flows, which dip 55° towards the northeast, are right-way-up. The basalt at this locality is at a stratigraphic level 1000 m above the base of the formation and, from geochronological data along strike and higher in the succession, is dated at 3.12 – 3.11 Ga. Nd T_{DM} model ages (3250–3150 Ma) from the Whundo Group (Sun and Hickman, 1998) suggest that it was not generated by melting of source rocks significantly older than its depositional age. This contrasts with the volcanic rocks of the c. 3.27 – 3.25 Ga Roebourne Group, which have Nd T_{DM} model ages of 3.48 – 3.43 Ga (Sun and Hickman, 1998).

The top of the hill at Locality 30 provides views of formations of the Fortescue Group to the north, east, and south. To the north, the basal Mount Roe Basalt (c. 2.77 Ga) unconformably overlies the Whundo Group, and dips southeastwards at about 20°. To the east, the Mount Roe Basalt is overlain by the c. 2.76 Ga Hardey Formation, which locally comprises sandstone, tuff, ultramafic lava, and dolerite.

The water-pipeline track leads onto a well-maintained gravel road near Harding Dam. Follow this road northwards to Roebourne and Highway 1. This is the end of the excursion. Participants can drive west to Karratha or east to Port Hedland.

4IAS AHH FIG19 10.08.01

Figure 44. Pillow lava in the Bradley Basalt, 2.5 km northwest of Harding Dam, Locality 30 (after Hickman et al., 2006)

References

Abeysinghe, PB, and Fetherston, JM, 1997, Barite and Fluorite in Western Australia: Geological Survey of Western Australia, Mineral Resources Bulletin 17, 97p.

Allwood, AC, Burch, IW, and Walter, MR, 2007a, Stratigraphy and facies of the 3.43 Ga Strelley Pool Chert in the southwestern part of the North Pole Dome, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 2007/11, 22p.

Allwood, AC, Walter, MR, Burch, IW, and Kamber, BS, 2007b, 3.43 billion-year-old stromatolite reef from the Pilbara Craton of Western Australia: Ecosystem-scale insights to early life on Earth: Precambrian Research, v. 158, p. 198–227.

Allwood, AC, Walter, MR, Kamber, BS, Marshall, CP, and Burch, IW, 2006, Stromatolite reef from the Early Archaean era of Australia: Nature, v. 441, p. 714–718.

Allwood, AC, Walter, MR, Marshall, CP, and Van Kranendonk, MJ, 2004, Life at 3.4 Ga; paleobiology and paleoenvironment of the stromatolitic Strelley Pool Chert, Pilbara Craton, Western Australia: Geological Society of America, Annual Meeting 2004, Denver, USA, Abstracts with Programs, p. 196.

Amelin, Y, Lee, D-C, and Halliday, AN, 2000, Early-middle Archaean crustal evolution deduced from Lu-Hf and U-Pb isotopic studies of single zircon grains: Geochimica et Cosmochimica Acta, v. 64, p. 4205–4225

Atlas Iron Limited, 2008a, EPA recommends Pardoo DSO Project for Ministerial approval: Report to Australian Stock Exchange, 16 June 2008, 1p.

Atlas Iron Limited, 2008b, Ridley Magnetite pre-feasibility study (PFS) technical update: Report to Australian Stock Exchange, 17 June 2008, 5p.

Bagas, L, Beukenhorst, O, and Hos, K, 2004, Nullagine, WA Sheet 2954: Geological Survey of Western Australia, 1:100 000 Geological Series.

Barley, ME, 1987, The Archaean Whim Creek Belt, an ensialic fault-bounded basin in the Pilbara Block, Australia: Precambrian Research, v. 37, p. 199–215.

Barley, ME, 1993, Volcanic, sedimentary and tectonostratigraphic environments of the ~3.46 Ga Warrawoona Megasequence: a review: Precambrian Research, v. 60, p. 47–67.

Barley, ME, 1997, The Pilbara Craton, *in* Greenstone belts *edited by* MJ de Wit and L Ashwal: Clarendon Press, Oxford University, Monographs on Geology and Geophysics, no. 35, p. 657–664.

Barley, ME, Dunlop, JSR, Glover, JE, and Groves, DI, 1979, Sedimentary evidence for Archaean shallow-water volcanic-sedimentary facies, eastern Pilbara Block, Western Australia: Earth and Planetary Science Letters, v. 43, p. 74–84.

Barley, ME, and Pickard, AL, 1999, An extensive, crustally-derived, 3325 to 3310 Ma silicic volcanoplutonic suite in the eastern Pilbara Craton: evidence from the Kelley Belt, McPhee Dome and Corunna Downs Batholith: Precambrian Research, v. 96, p. 41–62.

Barley, ME, Sylvester, GC, and Groves, DI, 1984, Archaean calc-alkaline volcanism in the Pilbara Block, Western Australia: Precambrian Research, v. 24, p. 285–319.

Beintema, KA, 2003, Geodynamic evolution of the west and central Pilbara Craton: a mid-Archaean active continental margin: University of Utrecht, Geologica Ultraiectina, No. 232, 248p.

Bickle, MJ, Bettenay, LF, Barley, ME, Chapman, HJ, Groves, DI, Campbell, IH, and de Laeter, JR, 1983, A 3500 Ma plutonic and volcanic calc-alkaline province in the Archaean East Pilbara Block: Contributions to Mineralogy and Petrology, v. 84, p. 25–35.

Bickle, MJ, Bettenay, LF, Chapman, HJ, Groves, DI, McNaughton, NJ, Campbell, IH, and de Laeter, JR, 1989, The age and origin of younger granitic plutons of the Shaw batholith in the Archaean Pilbara Block, Western Australia: Contributions to Mineralogy and Petrology, v. 101, p. 361–376.

Bickle, MJ, Bettenay, LF, Boulter, CA, Groves, DI, and Morant, P, 1980, Horizontal tectonic intercalation of an Archaean gneiss belt and greenstones, Pilbara Block, Western Australia: Geology, v. 8, p. 525–529.

Bickle, MJ, Bettenay, LF, Chapman, HJ, Groves, DI, McNaughton, NJ, Campbell, IH, and de Laeter, JR, 1993, Origin of the 3500–3300 Ma calc-alkaline rocks in the Pilbara Archaean: isotopic and geochemical constraints from the Shaw Batholith: Precambrian Research, v. 60, p. 117–149.

Bickle, MJ, Morant, P, Bettenay, LF, Boulter, CA, Blake, TS, and Groves, DI, 1985, Archaean tectonics of the Shaw Batholith, Pilbara Block, Western Australia: structural and metamorphic tests of the batholith concept, *in* Evolution of Archean Supracrustal Sequences *edited by* LD Ayers, PC Thurston, KD Card, and W Weber: Geological Association of Canada, Special Paper 28, p. 325–341.

Blewett, RS, 2000, North Pilbara 'Virtual' structural field trip: Australian Geological Survey Organisation, Record 2000/45.

Blewett, RS, 2002, Archaean tectonic processes: a case for horizontal shortening in the North Pilbara granite–greenstone terrane, Western Australia: Precambrian Research, v. 113, p. 87–120.

Blewett, RS, Shevchenko, S, and Bell, B, 2004, the North Pole Dome: a non-diapiric dome in the Archaean Pilbara Craton, Western Australia: Precambrian Research, v. 133, p. 105–120.

Blewett, RS, Wellman, M, Ratajkoski, M, and Huston, DL, 2000, Atlas of North Pilbara Geology and Geophysics, 1:1.5 Million Scale: Australian Geological Survey Organisation, Record 2000/04, 36p.

Bodorkos, S, in prep., GSWA 160258, Geochronology dataset 399, *in* Compilation of geochronology data, June 2008 update: Geological Survey of Western Australia.

Boulter, CA, Bickle, MJ, Gibson, B, and Wright, RK, 1987, Horizontal tectonics pre-dating upper Gorge Creek Group sedimentation, Pilbara Block, Western Australia: Precambrian Research, v. 36, p. 241–258.

Brauhart, C, 1999, Regional alteration systems associated with Archean volcanogenic massive sulphide deposits at Panorama, Pilbara, Western Australia: University of Western Australia, PhD thesis (unpublished).

Brauhart, C, Groves, DI, and Morant, P, 1998, Regional alteration systems associated with volcanogenic massive sulphide mineralization at Panorama, Pilbara, Western Australia: Economic Geology, v. 93, p. 292–302.

Buick, R, and Barnes, KR, 1984, Cherts in the Warrawoona Group: Early Archaean silicified sediments deposited in shallow-water environments, *in* Archaean and Proterozoic Basins of the Pilbara, Western Australia: Evolution and Mineralization Potential *edited by* JR Muhling, DI Groves, and TS Blake: University of Western Australia, Geology Department and University Extension, Publication 9, p. 37–53.

Buick, R, Brauhart, CW, Morant, P, Thornett, JR, Maniw, JG, Archibald, NJ, Doepel, MG, Fletcher, IR, Pickard, AL, Smith, JB, Barley, ME, McNaughton, NJ, and Groves, DI, 2002, Geochronology and stratigraphic relationships of the Sulphur Springs Group and Strelley Granite: a temporally distinct igneous province in the Archaean Pilbara Craton, Australia: Precambrian Research, v. 114, p. 87–120.

Buick, R, Dunlop, JSR, and Groves, DI, 1981, Stromatolite recognition in ancient rocks: an appraisal of irregular laminated structures in an early Archaean chert–barite unit from North Pole, Western Australia: Alcheringa, v. 5, p. 161–181.

Buick, R, and Dunlop, J, 1990, Evaporitic sediments of early Archaean age from the Warrawoona Group, North Pole, Western Australia: Sedimentology, v. 37, p. 247–277.

Buick, R, Thornett, JR, McNaughton, NJ, Smith, JB, Barley, ME, and Savage, M, 1995, Record of emergent continental crust ~3.5 billion years ago in the Pilbara Craton of Australia: Nature, v. 375, p. 574–577.

Byerly, GR, Lowe, DR, Wooden, JL, and Xie, X, 2002, An Archean impact layer from the Pilbara and Kaapvaal cratons: Science, v. 297, p. 1325–1327.

Champion, DC, and Smithies, RH, 2000, The geochemistry of the Yule Granitoid Complex, East Pilbara Granite–Greenstone Terrane; evidence for early felsic crust: Geological Survey of Western Australia, Annual Review 1999–2000, p. 42–48.

Collins, WJ, 1989, Polydiapirism of the Archaean Mt Edgar batholith, Pilbara Block, Western Australia: Precambrian Research, v. 43, p. 41–62.

Collins, WJ, 1993, Melting of Archaean sialic crust under high a_{H2O} conditions: genesis of 3300 Ma Na-rich granitoids in the Mount Edgar Batholith, Pilbara Block, Western Australia: Precambrian Research, v. 60, p. 151–174.

Collins, WJ, Van Kranendonk, MJ, and Teyssier, C, 1998, Partial convective overturn of Archaean crust in the east Pilbara Craton, Western Australia: driving mechanisms and tectonic implications: Journal of Structural Geology, v. 20, p. 1405–1424.

Collins, WJ, and Van Kranendonk, MJ, 1999, Model for the development of kyanite during partial convective overturn of Archaean granite–greenstone terranes: the Pilbara Craton, Australia: Journal of Metamorphic Geology, v. 17, no. 2, p. 145–156.

Cumming, GL, and Richards, JR, 1975, Ore lead isotope ratios in a continuously changing Earth: Earth and Planetary Science Letters, v. 28, p. 155–171.

Davids, C, Wijbrans, JR, and White, SH, 1997, $^{40}Ar/^{39}Ar$ laserprobe ages of metamorphic hornblendes from the Coongan Belt, Pilbara, Western Australia: Precambrian Research, v. 83, p. 221–242.

Dawes, PR, Smithies, RH, Centofanti, J, and Podmore, DC, 1995, Sunrise Hill unconformity: a newly discovered regional hiatus between Archaean granites and greenstones in the northeastern Pilbara Craton: Australian Journal of Earth Sciences, v. 42, p. 635–639.

Delor, C, Burg, J-P, and Clarke, G, 1991, Relations diapirisme–metamorphisme dans la Province du Pilbara (Australie-occidentale): implications pour les régimes thermiques et tectoniques à la Archéen: Comptes Rendus de l'Academie des Sciences Paris, v. 312, p. 257–263.

DiMarco, MJ, and Lowe, DR, 1989, Stratigraphy and sedimentology of an Early Archaean felsic volcaniclastic sequence, eastern Pilbara Block, Western Australia, with special reference to the Duffer Formation and implications for crustal evolution: Precambrian Research, v. 44, p. 147–169.

Dunlop, JSR, Muir, MD, Milne, VA, and Groves, DI, 1978, A new microfossil assemblage from the Archaean of Western Australia: Nature, v. 274, p. 676–678.

Eriksson, KA, Krapez, B, and Fralick, P, 1994, Archaean sedimentation: Earth-Science Reviews, v. 37, p. 1–88.

Ferguson, KM, and Ruddock, I, 2001, Mineral occurrences and exploration potential of the east Pilbara: Geological Survey of Western Australia, Report 81, 114p.

Furnes, H, Banerjee, NR, Staudigel, H, Muehlenbachs, K, McLoughlin, N, de Wit, M, and Van Kranendonk, MJ, 2007, Comparing petrographic signatures of bioalteration in recent to Mesoarchean pillow lavas: Tracing subsurface life in oceanic igneous rocks: Precambrian Research, v. 158, p. 156–176.

Geological Survey of Western Australia, 2006, Compilation of geochronology data, June 2006 update, Geological Survey of Western Australia.

Geological Survey of Western Australia, 2008, Pilbara: Geological Survey of Western Australia, 1:100 000 Geological Information Series.

Geological Survey of Western Australia, in prep., Compilation of geochronology data, June 2008 update, Geological Survey of Western Australia.

Glikson, AY, Allen, C, and Vickers, J, 2004, Multiple 3.47-Ga-old asteroid impact fallout units, Pilbara Craton, Western Australia. Earth and Planetary Science Letters, v. 221(1 4), p. 383–396.

Glikson, AY, Davy, R, Hickman, AH, Pride, C, and Jahn, B, 1987, Trace element geochemistry and petrogenesis of Archaean felsic igneous rocks, Pilbara Block, Western Australia: Australian Bureau of Mineral Resources, Geology and Geophysics, Record 1987/30, 63p.

Glikson, AY, and Hickman, AH, 1981, Geochemistry of Archaean volcanic successions, eastern Pilbara Block, Western Australia: Australia BMR, Record 1981/36, 83p.

Glikson, AY, and Vickers, J, in prep, Archean to early Proterozoic asteroid impact fallout units and related tsunami deposits, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Report 102.

Goellnicht, NM, Groves, IM, Groves, DI, Ho, SE, and McNaughton, NJ, 1988, A comparison between mesothermal gold deposits of the Yilgarn Block and gold mineralization at Telfer and Miralga Creek, Western Australia — indirect evidence for a non-magmatic origin for greenstone-hosted gold deposits, *in* Advances in understanding Precambrian gold deposits, volume II *edited by* SE Ho and DI Groves: University of Western Australia, Department of Geology and Extension Service, Publication no. 12, p. 23–40.

Green, MG, Sylvester, PJ, and Buick, R, 2000, Growth and recycling of early Archaean continental crust: geochemical evidence from the Coonterunah and Warrawoona Groups, Pilbara Craton, Australia: Tectonophysics, v. 322, p. 69–88.

Groves, IM, 1987, Epithermal/porphyry style base- and precious-metal mineralization in the Miralga Creek area, eastern Pilbara Block: University of Western Australia, BSc Hon thesis (unpublished).

Groves, DI, Dunlop, JSR, and Buick, R, 1981, An early habitat of life: Scientific American, v. 245, p. 64–73.

Hickman, AH, 1973, The North Pole barite deposits, Pilbara Goldfield: Geological Survey of Western Australia, Annual Report 1972, p. 57–60.

Hickman, AH, 1975, Precambrian Structural Geology of Part of the Pilbara region: Geological Survey of Western Australia, Annual Report 1974, p. 68–73.

Hickman, AH, 1977a, New and revised definitions of the Warrawoona Group, Pilbara Block: Geological Survey of Western Australia, Annual Report 1976, p. 53.

Hickman, AH, 1977b, Barite deposits near Cooke Bluff Hill, Port Hedland 1:250 000 sheet: Geological Survey of Western Australia, Record 1976/22, 22p.

Hickman, AH, 1980, Archaean geology of the Pilbara Block Excursion Guide, 2nd International Archean Symposium: Geological Society of Australia, WA Division, Perth, 55p.

Hickman, AH, 1981, Crustal evolution of the Pilbara Block, *in* Archaean Geology *edited by* JE Glover and DI Groves: 2nd International Archaean Symposium, Perth, 1980: Geological Society of Australia, Special Publication no. 7, p. 57–69.

Hickman, AH, 1983, Geology of the Pilbara Block and its environs: Geological Survey of Western Australia, Bulletin 127, 268p.

Hickman, AH, 1984, Archaean diapirism in the Pilbara Block, Western Australia, *in* Precambrian Tectonics Illustrated *edited by* A Kröner and R Greiling: E Schweizerbarts'che Verlagsbuchhandlung, Stuttgart, p. 113–127.

Hickman, AH, 1990, Granite–greenstone terrain, *in* 3rd International Archaean Symposium Excursion Guidebook *edited by* SE Ho, JS Myers, and JR Muhling: University of Western Australia, Geology Department and University Extension, Publication no. 21, p. 213.

Hickman, AH, 1997, A revision of the stratigraphy of Archaean greenstone successions in the Roebourne–Whundo area, west Pilbara: Geological Survey of Western Australia, Annual Review 1996–97, p. 76–82.

Hickman, AH, 1999, New tectono-stratigraphic interpretations of the Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 1999/6, p. 4–6.

Hickman, AH, 2001a, East Pilbara diapirism: new evidence from mapping, *in* GSWA extended abstracts: new data for WA explorers: Geological Survey of Western Australia, Record 2001/5, p. 23–25.

Hickman, AH, 2001b, Geology of the Dampier 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 39p.

Hickman, AH, 2002, Geology of the Roebourne 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 35p.

Hickman, AH, 2004a, Two contrasting granite–greenstones terranes in the Pilbara Craton, Australia: evidence for vertical *and* horizontal tectonic regimes prior to 2900 Ma: Precambrian Research, v. 131, p. 153–172.

Hickman, AH, 2004b, Geology of the Cooya Pooya 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 25p.

Hickman, AH, 2008, Regional review of the 3426–3350 Ma Strelley Pool Formation, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 2008/15, 27p.

Hickman, AH, Huston, DL, Van Kranendonk, MJ, and Smithies, RH, 2006, Geology and mineralization of the west Pilbara — a field guide: Geological Survey of Western Australia, Record 2006/17, 50p.

Hickman, AH, and Kojan, CJ, 2003, Geology of the Pinderi Hills 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 36p.

Hickman, AH, and Lipple, SH, 1978a, Marble Bar, W.A. Sheet SF 50-8: Geological Survey of Western Australia, 1:250 000 Geological Series.

Hickman, AH, and Lipple SL, 1978b, Marble Bar, W.A. (2nd edition): Geological Survey of Western Australia, 1:250 000 Geological Series Explanatory Notes, 23p.

Hickman, AH, and Smithies, RH, in prep., The West Pilbara Superterrane, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Report series.

Hickman, AH, Smithies, RH, and Huston, DL, 2000, Archaean geology of the West Pilbara Granite–Greenstone Terrane and Mallina Basin, Western Australia — a field guide: Geological Survey of Western Australia, Record 2000/9, p. 61.

Hickman, AH, and Strong, CA, 2003, Dampier – Barrow Island, W.A., (2nd Edition): Geological Survey of Western Australia, 1:250 000 Geological Series Explanatory Notes, 75p.

Hickman, AH, and Van Kranendonk, MJ, 2004, Diapiric processes in the formation of Archaean continental crust, East Pilbara Granite–Greenstone Terrane, Australia, *in* The Precambrian Earth: Tempos and Events *edited by* PG Eriksson, W Altermann, DR Nelson, WU Mueller, and O Catuneau: Elsevier, p. 54–75.

Hickman, AH, and Van Kranendonk, M J, 2008, Marble Bar, W.A. Sheet 2855: Geological Survey of Western Australia, 1:100 000 Geological Series.

Hickman, AH, Van Kranendonk, MJ, and Glikson, AY, in prep., Geology of the Mount Edgar Dome, Pilbara Craton, Geological Survey of Western Australia, Report series

Hofmann, HJ, Grey, K, Hickman, AH, and Thorpe, R, 1999, Origin of 3.45 Ga coniform stromatolites in Warrawoona Group, Western Australia: Geological Society of America, Bulletin, v. 111, p. 1256–1262.

Huston, DL, Blewett, RS, Sweetapple, M, Brauhart, C, Cornelius, H, and Collins, PLF, 2001, Metallogenesis of the north Pilbara granite–greenstones, Western Australia — a field guide: Geological Survey of Western Australia, Record 2001/11, p. 87.

Huston, DL, Sun, S-S, Blewett, R, Hickman, AH, Van Kranendonk, MJ, Phillips, D, Baker, D, and Brauhart, C, 2002, The timing of mineralisation in the Archaean Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 733–755.

Isozaki, Y, Kabashima, T, Ueno, Y, Kitajima, K, Maruyama, S, Kato, Y, and Terabayashi, M, 1997, Early Archean mid-ocean ridge rocks and early life in the Pilbara Craton, W. Australia: Eos, v. 78, p. 399.

Jahn, B, Glikson, AY, Peucat, JJ, and Hickman, AH, 1981, REE geochemistry and geochronology of Archaean silicic volcanics and granitoids from the Pilbara Block, Western Australia: Geochimica et Cosmochimica Acta, v. 45, p. 1633–1652.

Kinny, PD, 2000, U–Pb dating of rare-metal (Sn–Ta–Li) mineralized pegmatites in Western Australia by SIMS analysis of tin and tantalum-bearing ore minerals, *in* Beyond 2000, New Frontiers in Isotope Geoscience, Lorne, Victoria, 2000, Abstracts and Proceedings, p. 113–116.

Kitajima, K, Maruyama, S, Utsunomiya, S, and Liou, JG, 2001, Seafloor hydrothermal alteration at an Archean mid-ocean ridge: Journal of Metamorphic Geology, v. 19, p. 583–599.

Kiyokawa, S, 1993, Stratigraphy and structural evolution of a Middle Archaean greenstone belt, northwestern Pilbara Craton: University of Tokyo, PhD thesis (unpublished).

Kiyokawa, S, and Taira, A, 1998, The Cleaverville Group in the west Pilbara coastal granite–greenstone terrane of Western Australia: an example of a mid-Archaean immature oceanic island-arc succession: Precambrian Research, v. 88, p. 102–142.

Kloppenburg, A, 2003, Structural evolution of the Marble Bar Domain, Pilbara granite greenstone terrain, Australia: the role of Archaean mid-crustal detachments: University of Utrecht, Geologica Ultraiectina No. 237, 256p.

Kloppenburg, A, White, SH, and Zegers, TE, 2001, Structural evolution of the Warrawoona Greenstone Belt and adjoining granitoid complexes, Pilbara Craton, Australia: implications for Archaean tectonic processes: Precambrian Research, v. 112, p. 107–147.

Krapez, B, 1984, Sedimentation in a small, fault-bounded basin: the Lalla Rookh sandstone, east Pilbara Block, *in* Archaean and Proterozoic basins of the Pilbara, Western Australia: Evolution and Mineralization potential *edited by* JR Muhling, DI Groves, and TS Blake : University of Western Australia, Geology Department and University Extension, Publication 9, p. 89–110.

Krapez, B, 1993, Sequence stratigraphy of the Archaean supracrustal belts of the Pilbara Block, Western Australia: Precambrian Research, v. 60, p. 1–45.

Krapez, B, and Eisenlohr, B, 1998, Tectonic settings of Archaean (3325–2775 Ma) crustal–supracrustal belts in the West Pilbara Block: Precambrian Research, v. 88, p. 173–205.

Lindsay, JF, Brasier, MD, McLoughlin, N, Green, OR, Fogel, M, McNamara, KM, Steele, A, and Mertzman, SA, 2003, Abiotic Earth – establishing a baseline for earliest Life, data from the Archean of Western Australia: Lunar and Planetary Science Conference, Proceedings, LPI, Houston, Texas, p. 1137

Lindsay, JF, Brasier, MD, McLoughlin, N, Green, OR, Fogel, M, Steele, A, and Mertzman, SA, 2005, The problem of deep carbon — an Archaean paradox: Precambrian Research, v. 143, p. 1–22.

Lowe, DR, 1980, Stromatolites 3,400-Myr old from the Archean of Western Australia: Nature v. 284, p. 441–443.

Lowe, DR, 1983, Restricted shallow-water sedimentation of early Archaean stromatolitic and evaporitic strata of the Strelley Pool chert, Pilbara block, Western Australia: Precambrian Research, v. 19, p. 239–283.

Lowe, DR, and Byerly, GR., 1986, Early Archean silicate spherules of probable impact origin, South Africa and Western Australia: Geology, v. 14, p. 83–86.

Mareschal, J-C, and West, GF, 1980, A model for Archean tectonism. Part 2. Numerical models of vertical tectonism in greenstone belts: Canadian Journal of Earth Sciences, v. 17, p. 60–71.

Marshall, CP, 2007, Organic geochemistry of Archaean carbonaceous cherts from the Pilbara Craton, Western Australia *in* Earth's Oldest Rocks, *edited by* MJ Van Kranendonk, RH Smithies, and VC Bennett: Developments in Precambrian Geology, v. 15, Elsevier, Amsterdam, p. 897–921.

Marston, RJ, 1984, Nickel mineralization in Western Australia: Geological Survey of Western Australia, Mineral Resources Bulletin 14, 217p.

McNaughton, NJ, Compston, W, and Barley, ME, 1993, Constraints on the age of the Warrawoona Group, eastern Pilbara Block, Western Australia: Precambrian Research, v. 60. p. 69–98.

Miller, LJ, and Gair, HS, 1975, Mons Cupri copper–lead–zinc–silver deposit: Australasian Institute of Mining and Metallurgy Monograph 5, p. 195–202.

Moly Mines Limited, 2007, Spinifex Ridge Molybdenum–Copper Deposit, Western Australia, Technical Report NI 43-101 *compiled by* SRK Consulting: viewed 26 June 2008 <http://www1.molymines.com/documents/4/8/SRMo_NI43_101_TSX_301007_Final1.pdf>.

Moly Mines Limited, 2008, Doubling of Spinifex Ridge Resource: Report to Australian Stock Exchange, 31 March 2008, 3p.

Morant, P, 1998, Panorama zinc–copper deposits: Australasian Institute of Mining and Metallurgy, Monograph 22, p. 287–292.

Nelson, DR, 1997, Compilation of SHRIMP U–Pb zircon geochronology data, 1996: Geological Survey of Western Australia, Record 1997/2, 189p.

Nelson, DR, 1998, Compilation of SHRIMP U–Pb zircon geochronology data, 1997: Geological Survey of Western Australia, Record 1998/2, 242p.

Nelson, DR, 1999a, 142170: foliated biotite monzogranite, Kangan Homestead; Geochronology dataset 399, *in* Compilation of geochronology data, June 2007 update: Geological Survey of Western Australia.

Nelson, DR, 1999b, Compilation of SHRIMP U–Pb zircon geochronology data, 1998: Geological Survey of Western Australia, Record 1999/2, 222p.

Nelson, DR, 2000a, 148498: tuffaceous chert, Eight Mile Bore; Geochronology dataset 245, *in* Compilation of geochronology data, June 2007 update: Geological Survey of Western Australia.

Nelson, DR, 2000b, Compilation of geochronology data, 1999: Geological Survey of Western Australia, Record 2000/2, 251p.

Nelson, DR, 2001, 168908: crystal-lithic tuff, Copper Hills; Geochronology dataset 222, *in* Compilation of geochronology data, June 2007 update: Geological Survey of Western Australia.

Neumayr, P, Groves, DI, Ridley, JR, and Koning, CD, 1993, Syn-amphibolite facies Archaean lode gold mineralization in the Mt York District, Pilbara Block, Western Australia: Mineralium Deposita, v. 28, p. 457–468.

Neumayr, P, Ridley, JR, McNaughton, NJ, Kinny, PD, Barley, ME, and Groves, DI, 1998, Timing of gold mineralization in the Mount York district, Pilgangoora greenstone belt and implications for the tectonic and metamorphic evolution of an area linking the western and eastern Pilbara Craton: Precambrian Research, v. 88, p. 249–265.

Nijman, W, de Bruijne, CH, and Valkering, ME, 1998a, Growth fault control of early Archaean cherts and barite mounds and chert-barite veins, North Pole Dome, Eastern Pilbara, Western Australia: Precambrian Research, v. 88, p. 25–52.

Nijman, W, De Bruijne, CH, and Valkering, ME, 1999a, Growth fault control of Early Archaean cherts, barite mounds and chert–barite veins, North Pole Dome, Eastern Pilbara, Western Australia (Erratum to article in Precambrian Research, v. 88, p. 25–52): Precambrian Research, v. 95, p. 247–274.

Nijman, W, and de Vries, ST, 2004, Early Archaean crustal collapse structures and sedimentary basin dynamics, *in* The Precambrian Earth: Tempos and Events *edited by* PG Eriksson, W Altermann, DR Nelson, WU Mueller, and O Catuneau: Amsterdam, Elsevier, , p. 139–154.

Nijman, W, Willigers, BA, and Krikke, A, 1998b, Tensile and compressive growth structures: relationships between sedimentation, deformation and granite intrusion in the Archaean Coppin Gap greenstone belt, Eastern Pilbara, Western Australia: Precambrian Research, v. 88, p. 83–107.

Nijman, W, Willigers, BA, and Krikke, A, 1999b, Tensile and compressive growth structures: relationships between sedimentation, deformation and granite intrusion in the Archaean Coppin Gap greenstone belt, Eastern Pilbara, Western Australia (Erratum to article in Precambrian Research, v. 88, p. 83–107): Precambrian Research, v. 95, p. 277–302.

Ohta, H, Maruyama, S, Takahashi, E, Watanabe, Y, and Kato, Y, 1996, Field occurrence, geochemistry and petrogenesis of the Archaean Mid-Oceanic Ridge Basalts (AMORBs) of the Cleaverville area, Pilbara Craton, Western Australia: Lithos, v. 37, p. 199–221.

Pirajno, F, and Van Kranendonk, MJ, 2005, A review of hydrothermal processes and systems on earth and implications for Martian analogues: Australian Journal of Earth Sciences, v. 52, p. 329–351.

Rasmussen, B, Fletcher, IR, and Muhling, J, 2007, In-situ U–Pb dating and element mapping in three generations of monazite: Unravelling cryptic tectonothermal events in low-grade terranes: Geochimica et Cosmochimica Acta, v. 71. p. 670–690.

Reynolds, DG, Brook, WA, Marshall, AE, and Allchurch, PD, 1975, Volcanogenic copper zinc deposits in the Pilbara and Yilgarn Archaean Blocks, *in* Economic Geology of Australia and Papua New Guinea — Metals *edited by* CL Knight: Australasian Institute of Mining and Metallurgy Monograph 5, p. 185–195.

Ruddock, I, 1999, Mineral occurrences and exploration potential of the west Pilbara: Geological Survey of Western Australia, Report 70, 63p.

Sandiford, M, Van Kranendonk, MJ, and Bodorkos, S, 2004, Conductive incubation and the origin of dome-and-keel structure in Archean granite-greenstone terrains: a model based on the eastern Pilbara Craton, Western Australia: Tectonics 23, TC1009, DOI: 10.1029/2002TC001452.

Sargeant, DW, 1979, Annual report on mineral claim numbers 45/8674 to 45/8684 inclusive, North Pole area, Pilbara W.A.; Newmont Pty Ltd: Geological Survey of Western Australia, Statutory mineral exploration report, A9368 (unpublished).

Sargeant, DW, and Sampson, PG, 1980, Final report on Temporary reserve 6712H, North Pole area, Pilbara Goldfield; Newmont Proprietary Ltd: Geological Survey of Western Australia, Statutory mineral exploration report, A9093 (unpublished).

Sigutani, K, 1992, Geochemical characteristics of Archean cherts and other sedimentary rocks in the Pilbara Block, Western Australia: evidence for Archean seawater enriched in hydrothermally derived iron and silica: Precambrian Research, v. 57, p. 21–47.

Sinclair, WD, 1995, Porphyry molybdenum (Low-F-type), *in* Selected British Columbia Mineral Deposit profiles, Volume 1 — Metallics and Coal *edited by* DV Lefebure and GE Ray: British Columbia Ministry of Energy of Employment and Investment, Open File 1995-20, p. 93–96.

Smith, JB, Barley, ME, Groves, DI, Krapez, B, McNaughton, NJ, Bickle, MJ, and Chapman, HJ, 1998, The Sholl Shear Zone, West Pilbara; evidence for a domain boundary structure from integrated tectonostratigraphic analysis, SHRIMP U–Pb dating and isotopic and geochemical data of granitoids: Precambrian Research, v. 88, p. 143–172.

Smithies, RH, 1998, Geology of the Mount Wohler 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 19p.

Smithies, RH, 2000, The Archaean tonalite-trondhjemite-granodiorite (TTG) series is not an analogue of Cenozoic adakite: Earth and Planetary Science Letters, v. 182, p. 115–125.

Smithies, RH, 2003, Geology of the White Springs 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 16p.

Smithies, RH, and Champion, DC, 2000, The Archaean high-Mg diorite suite: links to tonalite–trondhjemite–granodiorite magmatism and implications for Early Archaean crustal growth: Journal of Petrology, v. 41, p. 1653–1671.

Smithies, RH, Champion, DC, and Cassidy, KF, 2003, Formation of Earth's early Archaean continental crust: Precambrian Research, v. 127, p. 89–101.

Smithies, RH, Champion, DC, and Sun, S-S, 2004, Evidence for early LREE-enriched mantle source regions: Diverse magmas from the ca. 3.0 Ga Mallina Basin, Pilbara Craton, NW Australia: Journal of Petrology, v. 45, p. 1515–1537.

Smithies, RH, Champion, DC, Van Kranendonk, MJ, and Hickman, AH, 2007, Geochemistry of volcanic units of the northern Pilbara Craton: Geological Survey of Western Australia, Report 104, 48p.

Smithies, RH, Champion, DC, Van Kranendonk, MJ, Howard, HM, and Hickman, AH, 2005a, Modern-style subduction processes in the Mesoarchaean: geochemical evidence from the 3.12 Ga Whundo intraoceanic arc: Earth and Planetary Science Letters, v. 231, p. 221–237.

Smithies, RH, and Farrell, T, 2000, Geology of the Satirist 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 42p.

Smithies, RH, Hickman, AH, and Nelson, DR, 1999, New constraints on the evolution of the Mallina Basin, and their bearing on relationships between the contrasting eastern and western granite–greenstone terrains of the Archaean Pilbara Craton, Western Australia: Precambrian Research, v. 94, p. 11–28.

Smithies, RH, Van Kranendonk, MJ, and Champion, DC, 2005b, It started with a plume — early Archaean basaltic proto-continental crust: Earth and Planetary Science Letters, v. 238 (3–4), p. 284–297.

Sugitani, K, Yamamoto, K, Adachi, M, Kawabe, I, and Sugisaki, R, 1998, Archaean cherts derived from chemical, biogenic, and clastic sedimentation in a shallow restricted basin: examples from the Gorge Creek Group in the Pilbara Block: Sedimentology, v. 45, p. 1045–1062.

Sun, S-S, and Hickman, AH, 1998, New Nd-isotopic and geochemical data from the west Pilbara — implications for Archaean crustal accretion and shear zone development: Australian Geological Survey Organisation, Research Newsletter, no. 28, p. 25–29.

Sweetapple, MT, Cornelius, H, and Collins, PLF, 2001, Part four: The Wodgina tantalum–tin pegmatite district, *in* Metallogenesis of the North Pilbara granite–greenstones, Western Australia — a field guide *edited by* DL Huston, RS Blewett, M Sweetapple, C Brauhart, H Cornelius, and PLF Collins: Geological Survey of Western Australia, Record 2001/11, p. 41–58.

Teyssier, C, and Collins, WJ, 1990, Strain and kinematics during the emplacement of the Mount Edgar Batholith and Warrawoona Syncline, Pilbara Block, Western Australia, *in* 3rd International Archaean Symposium, Extended Abstracts Volume *compiled by* JE Glover and SE Ho: Perth, Western Australia, Geoconferences (W.A.) Inc, p. 481–483.

Thorne, AM, and Trendall, AF, 2001, The geology of the Fortescue Group, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Bulletin 144, 249p.

Thorpe, RI, Hickman, AH, Davis, DW, Mortensen, JK, and Trendall, AF, 1992a, U–Pb zircon geochronology of Archaean felsic units in the Marble Bar region, Pilbara Craton, Western Australia: Precambrian Research, v. 56, p. 169–189.

Thorpe, RI, Hickman, AH, Davis, DW, Mortensen, JK, and Trendall, AF, 1992b, Constraints to models for Archaean lead evolution from precise U–Pb geochronology from the Marble Bar region, Pilbara Craton, Western Australia, *in* The Archaean: Terrains, processes and metallogeny *edited by* JE Glover and SE Ho: University of Western Australia, Geology Department and University Extension, Publication no. 22, p. 395–408.

Tomich, BNV, 1974, The geology and nickel mineralisation of the Ruth Well area, Western Australia: University of Western Australia, BSc (Hons) thesis (unpublished).

Trendall, AF, 1990, Pilbara Craton, Introduction, *in* Geology and mineral resources of Western Australia: Geological Survey of Western Australia, Memoir 3, p. 128.

van Haaften, WM, and White, SH, 1998, Evidence for multiphase deformation in the Archean basal Warrawoona Group in the Marble Bar area, East Pilbara, Western Australia: Precambrian Research v. 88, p. 53–66.

van Haaften, WM, and White, SH, 2001, Reply to comment on 'Evidence for multiphase deformation in the Archaean basal Warrawoona Group in the Marble Bar area, East Pilbara, Western Australia' by MJ Van Kranendonk, AH Hickman, and WJ Collins: Precambrian Research, v. 105, p. 79–84.

van Koolwijk, ME, Beintema, KA, White, SH, and Wijbrans, JR, 2001, Petrogenesis and structures of the basal Warrawoona Group, Marble Bar Belt, Pilbara Craton, W.A, *in* 4th International Archaean Symposium, Extended Abstracts *edited by* KF Cassidy, JM Dunphy, and MJ Van Kranendonk: Geoscience Australia, Record 2001/37, p. 102–103.

Van Kranendonk, MJ, 1997, Results of field mapping, 1994–1996, in the North Shaw and Tambourah 1:100 000 sheet areas, eastern Pilbara Craton, northwestern Australia: Australian Geological Survey Organisation, Record 1997/23, 44p.

Van Kranendonk, MJ, 1998, Litho-tectonic and structural map components of the North Shaw 1:100 000 sheet, Archaean Pilbara Craton: Geological Survey of Western Australia, Annual Review 1997–98, p. 63–70.

Van Kranendonk, MJ, 2000, Geology of the North Shaw 1:100 000 sheet: Geological Survey of Western Australia, 1:100á000 Geological Series Explanatory Notes, 86p.

Van Kranendonk, MJ, 2001, Volcanic degassing, hydrothermal circulation and the flourishing of early life on new evidence from the c. 3.46 Ga Warrawoona Group, Pilbara Craton, Western Australia: West Australian Geologist, no. 402, p. 3.

Van Kranendonk, MJ, 2002, Tambourah, WA Sheet 2754: Geological Survey of Western Australia, 1:100 000 Geological Series.

Van Kranendonk, MJ, 2003a, Stratigraphic and tectonic significance of eight local unconformities in the Fortescue Group, Pear Creek Centrocline, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Annual Review 2001–2002, p. 70–79.

Van Kranendonk, MJ, 2003b. Geology of the Tambourah 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 59p.

Van Kranendonk, MJ, 2004, Geology of the Carlindie 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 45p.

Van Kranendonk, MJ, 2006, Volcanic degassing, hydrothermal circulation and the flourishing of life on Earth: New evidence from the c. 3.46 Ga Warrawoona Group, Pilbara Craton, Western Australia: Earth-Science Reviews, v. 74(3–4), p. 197–240.

Van Kranendonk, MJ, 2007, A review of the evidence for putative Paleoarchean life in the Pilbara Craton, Western Australia, *in* Earth's Oldest Rocks *edited by* MJ Van Kranendonk, RH Smithies, and VC Bennett: Developments in Precambrian Geology, v. 15, Elsevier, Amsterdam, p. 855–877.

Van Kranendonk, MJ, 2008, Structural geology of the central part of the Lalla Rookh – Western Shaw structural corridor, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Report 103, 29p.

Van Kranendonk, MJ, and Collins, WJ, 1998, Timing and tectonic significance of Late Archaean, sinistral strike-slip deformation in the Central Pilbara Structural Corridor, Pilbara Craton, Western Australia: Precambrian Research, v. 88, p. 207–232.

Van Kranendonk, MJ, Collins, WJ, Hickman, AH, and Pawley, MJ, 2004, Critical tests of vertical vs horizontal tectonic models for the Archaean East Pilbara Granite–Greenstone Terrane, Pilbara Craton, Western Australia: Precambrian Research, v. 131, p. 173–211.

Van Kranendonk, MJ, and Hickman, AH, 2000, Archaean geology of the North Shaw region, East Pilbara Granite–Greenstone Terrain, Western Australia — a field guide: Geological Survey of Western Australia, Record 2000/5, 64p.

Van Kranendonk, MJ, Hickman, AH, and Collins, WJ, 2001a, Comment on 'Evidence for multiphase deformation in the Archaean basal Warrawoona Group in the Marble Bar area, East Pilbara, Western Australia': Precambrian Research, v. 105, p. 73–78.

Van Kranendonk, MJ, Hickman, AH, Huston, DL, and Cummins, B, 2006b, Geology and mineralization of the east Pilbara — a field guide: Geological Survey of Western Australia, Record 2006/16.

Van Kranendonk, MJ, Hickman, AH Smithies, RH, and Nelson, D, 2002, Geology and tectonic evolution of the Archean North Pilbara terrain, Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 695–732.

Van Kranendonk, MJ, Hickman, AH, Smithies, RH, Williams, IR, Bagas, L, and Farrell, TR, 2006a, Revised lithostratigraphy of Archean supracrustal and intrusive rocks in the northern Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 2006/15, 57p.

Van Kranendonk, MJ, Hickman, AH, Williams, IR, Nijman, W, 2001b, Archaean geology of the East Pilbara Granite–Greenstone Terrane, Western Australia — a field guide: Geological Survey of Western Australia, Record 2001/9, 134p.

Van Kranendonk, MJ, and Morant, P, 1998, Revised Archaean stratigraphy of the North Shaw 1:100 000 sheet, Pilbara Craton: Geological Survey of Western Australia, Annual Review 1997–98, p. 55–62.

Van Kranendonk, MJ, and Nijman, W, 2001, Geology of the North Pole Dome, *in* Van Kranendonk, MJ, Hickman, AH, Williams IR, and Nijman, W, 2001, Archaean geology of the East Pilbara Granite–Greenstone Terrane, Western Australia a field guide: Geological Survey of Western Australia, Record 2001/9, p. 53–72.

Van Kranendonk, MJ, and Pirajno, F, 2004, Geological setting and geochemistry of metabasalts and alteration zones associated with hydrothermal chert ± barite deposits in the ca. 3.45 Ga Warrawoona Group, Pilbara Craton, Australia: Geochemistry: Exploration, Environment, Analysis, v. 4, p. 253–278.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH, Bagas, L, Williams, IR, and Farrell, TR, 2004, Event stratigraphy applied to 700 million years of Archaean crustal evolution, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Annual Review 2003–04, p. 49–61.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH, and Champion, DC, 2007, Secular tectonic evolution of Archaean continental crust: interplay between horizontal and vertical processes: Terra Nova, v. 19(1), p. 1–38.

Van Kranendonk, MJ, Webb, GE, and Kamber, BS, 2003, Geological and trace element evidence for a marine sedimentary environment of deposition and biogenicity of 3.45 Ga stromatolitic carbonates in the Pilbara Craton, and support for a reducing Archaean ocean: Geobiology v. 1, p. 91–108.

Vearncombe, S, Barley, ME, Groves, DI, McNaughton, NJ, Mikucki, EJ, and Vearncombe, JR, 1995, 3.26 Ga black smoker-type mineralization in the Strelley Belt, Pilbara Craton, Western Australia: London, Journal of the Geological Society, v. 152, p. 587–590.

Vearncombe, S, Vearncombe, JR, and Barley, ME, 1998, Fault and stratigraphic controls on volcanogenic massive sulphide deposits in the Strelley Belt, Pilbara Craton, Western Australia: Precambrian Research, v. 88, p. 67–82.

Wacey, D, McLoughlin, N, Stoakes, A, Kilburn, MR, Green, OR, Brasier, MD, and Hickman, AH, in prep., The ~3.4 Ga Strelley Pool Formation in the East Strelley greenstone belt — a field and petrographic guide: Geological Survey of Western Australia, Record Series.

Walter, MR, Buick, R, and Dunlop, JSR, 1980, Stromatolites 3,400–3,500 Myr old from the North Pole area, Western Australia: Nature, v. 284, p. 443–445.

Wellman, P, 2000, Upper crust of the Pilbara Craton, Australia: 3D geometry of a granite-greenstone terrain: Precambrian Research, v. 104, p. 187–206.

Wijbrans, JR, and McDougall, I, 1987, On the metamorphic history of an Archaean granitoid greenstone terrane, East Pilbara, Western Australia, using the $^{40}Ar/^{39}Ar$ age spectrum technique: Earth and Planetary Science Letters, v. 84, p. 226–242.

Wilhelmij, HR, and Dunlop, JSR, 1984, A genetic stratigraphic investigation of the Gorge Creek Group in the Pilgangoora syncline, *in* Archaean and Proterozoic Basins of the Pilbara, Western Australia: evolution and mineralization potential *edited by* JR Muhling, DI Groves, and TS Blake: University of Western Australia, Geology Department and University Extension, Publication no. 9, p. 68–88.

Williams, IR, 1998, Muccan, WA sheet 2956: Geological Survey of Western Australia, 1:100 000 Geological Series.

Williams, IR, 1999, Geology of the Muccan 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 39p.

Williams, IR, and Bagas, L, 2007, Geology of the Mount Edgar 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 62p.

Williams, IS, and Collins, WJ, 1990, Granite-greenstone terranes in the Pilbara block, Australia, as coeval volcano-plutonic complexes: Evidence from U–Pb zircon dating of the Mount Edgar batholith: Earth and Planetary Science Letters, v. 97, p. 41–53.

Wingate, MTD, in prep.a, 178185: Dalton Sill; Geochronology dataset, *in* Compilation of geochronology data, June 2008 update: Geological Survey of Western Australia.

Wingate, MTD, in prep.b, 178164: Bullock Hide; Geochronology dataset, *in* Compilation of geochronology data, June 2008 update: Geological Survey of Western Australia.

Zegers, TE, Barley, ME, Groves, DI, McNaughton, NJ, and White, SH, 2002, Oldest gold: deformation and hydrothermal alteration in the early Archean shear zone-hosted Bamboo Creek Deposit, Pilbara, Western Australia: Economic Geology, v. 97, p. 757–773.

Zegers, TE, De Keijzer, M, Passchier, CW, and White, SH, 1998, The Mulgandinnah shear zone; an Archean crustal-scale strike-slip zone, eastern Pilbara, Western Australia: Precambrian Research, v. 88, p. 233–248.

Zegers, TE, Nelson, DR, Wijbrans, JR, and White, SH, 2001, SHRIMP U–Pb zircon dating of Archean core complex formation and pancratonic strike-slip deformation in the East Pilbara Granite–Greenstone Terrain: Tectonics, v. 20 (6), p. 883–908.

Zegers, TE, White, SH, de Keijzer, M, and Dirks, P, 1996, Extensional structures during deposition of the 3460 Ma Warrawoona Group in the eastern Pilbara Craton, Western Australia: Precambrian Research, v. 80, p. 89–105.

Zegers, TE, Wijbrans, J, and White, SH, 1999, $^{40}Ar/^{39}Ar$ age constraints on tectonothermal events in the Shaw area of the eastern Pilbara granite–greenstone terrain (Western Australia): 700 Ma of Archaean tectonic evolution: Tectonophysics, v. 311, p. 45–81.

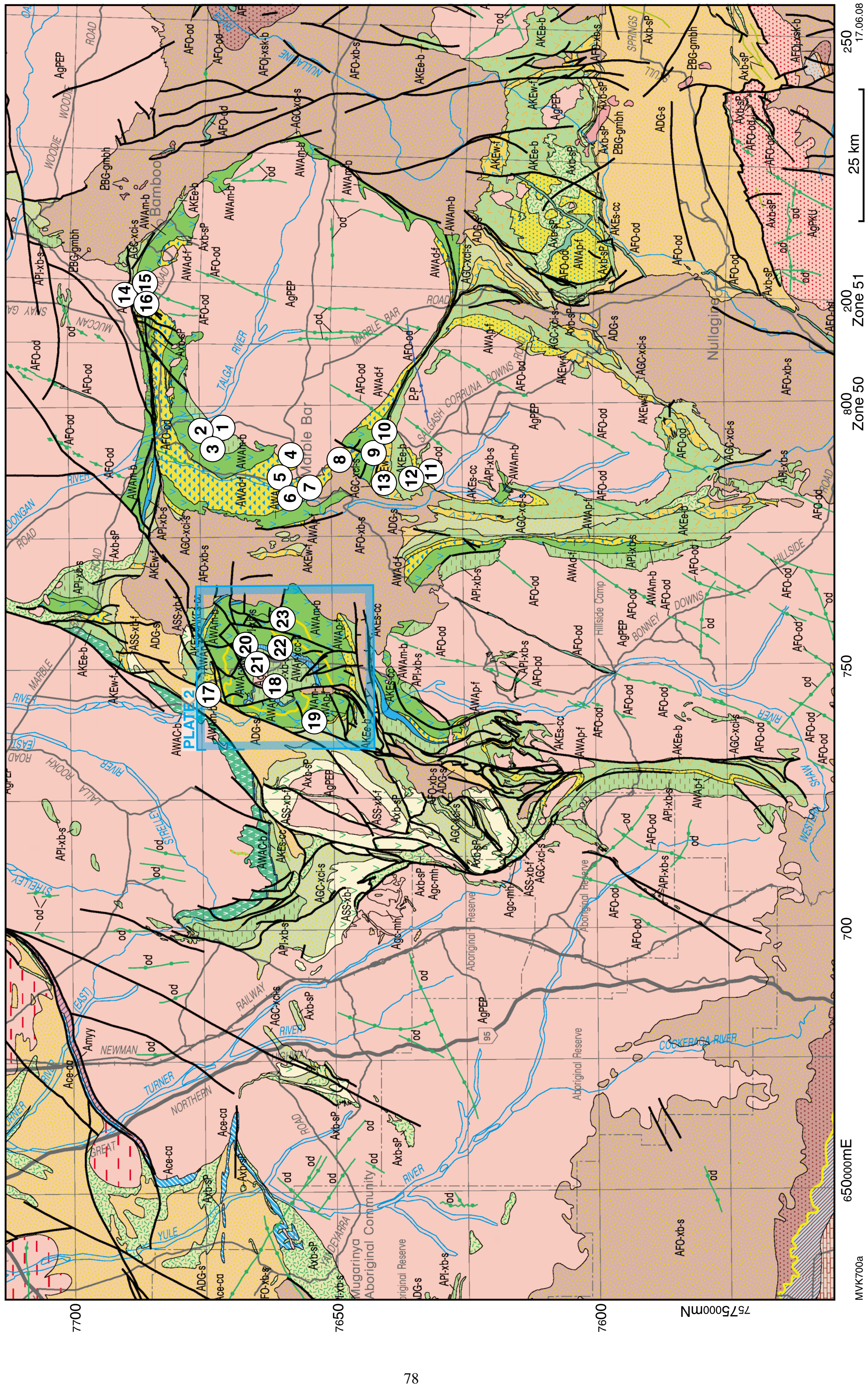
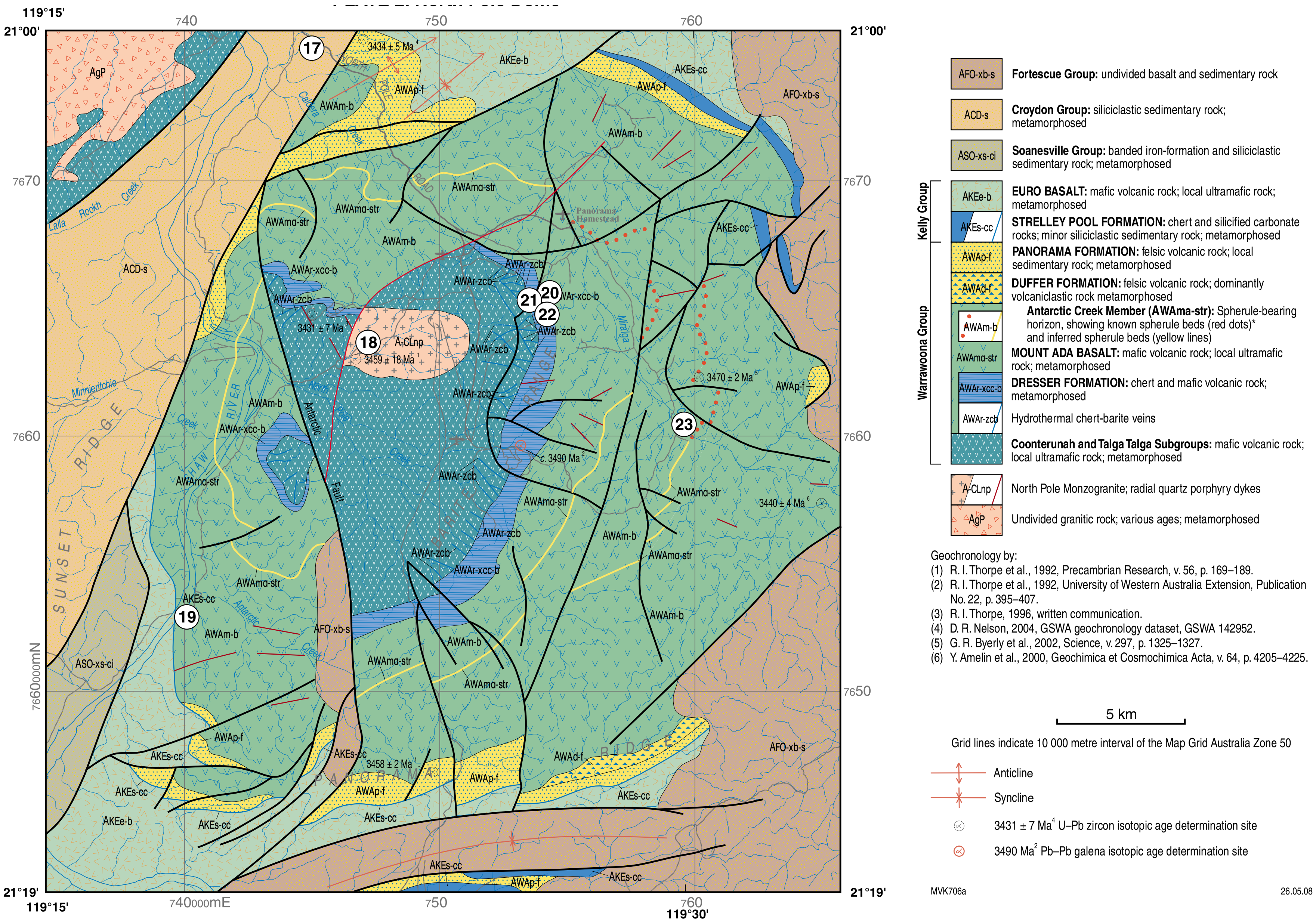


Plate 1. Geological map of the east Pilbara, showing excursion Localities 1–23

Plate 2. Interpreted bedrock geology map of the North Pole Dome showing excursion Localities



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