

Geology of the northern Capricorn Orogen

by

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Introduction

Seismic transects 10GA–CP1, 10GA–CP2, and 10GA–CP3 are located in the western part of the Capricorn Orogen, a major zone of Proterozoic sedimentation, deformation, metamorphism, and magmatism lying between the Archean Yilgarn and Pilbara Cratons (Gee, 1979; Cawood and Tyler, 2004; Plate 1; Frontispieces 1–3). The orogen includes metamorphic and igneous rocks of the Gascoyne Province, and a number of low-grade metasedimentary and metavolcanic basins, including the late Paleoproterozoic to Mesoproterozoic Edmund and Collier Basins. It also includes the deformed margins of the Yilgarn and Pilbara Cratons. The northern transect, 10GA–CP1, crosses granite–greenstones of the Pilbara Craton and an overlying supracrustal succession that begins within the Archean Fortescue Basin and ends within the Paleoproterozoic Blair Basin (Plate 1; Frontispiece 1). Transect 10GA–CP2 starts in the Edmund and Collier Basins, and extends southwards across Paleoproterozoic to Neoproterozoic rocks of the Mangaroon, Limejuice, Mutherbukin, Mooloo, and Paradise Zones of the Gascoyne Province. The southern transect, 10GA–CP3, begins in the Paradise Zone of the Gascoyne Province, crosses the Errabiddy Shear Zone, and ends in the Archean Narryer Terrane of the Yilgarn Craton.

The seismic transects across the Capricorn Orogen provide an important opportunity to gain insight into the deep crustal structure and geological evolution of a Proterozoic orogenic belt and its associated Archean cratons (Kennett et al., 2011). Early interpretations of the Capricorn Orogen — e.g. Horwitz and Smith (1978), and Gee (1979) — favoured an intracratonic setting. These were followed by plate tectonic models in which Proterozoic orogeny was driven by collision between the hitherto unrelated Pilbara and Yilgarn Cratons (Muhling, 1988; Tyler and Thorne, 1990; Evans et al., 2003). More recent work has highlighted the complexity of the Capricorn Orogen, which has now been shown to record seven major orogenic events: the 2215–2145 Ma Ophthalmian Orogeny, the 2005–1950 Ma Glenburgh Orogeny, the 1820–1770 Ma Capricorn Orogeny, the 1680–1620 Ma Mangaroon Orogeny, the 1385–1200 Ma Mutherbukin Tectonic Event, the 1030–950 Ma Edmundian Orogeny

(Occhipinti et al., 2001, 2004; Rasmussen et al., 2005; Sheppard et al., 2005, 2007, 2010a,b; Johnson et al., 2009, 2010, 2011a,b), and the c. 570 Ma Mulka Tectonic Event (Sheppard et al., 2010a; Johnson et al., 2011a). The oldest of these, the 2215–2145 Ma Ophthalmian Orogeny, is now believed to be the result of a collision between the Pilbara Craton and the Glenburgh Terrane of the Gascoyne Province (Occhipinti et al., 2004; Sheppard et al., 2004; Johnson et al., 2010, 2011b). It was followed by collision between the combined Pilbara Craton – Glenburgh Terrane and the Yilgarn Craton, an event that resulted in the 2005–1950 Ma Glenburgh Orogeny (Occhipinti et al., 2004; Sheppard et al., 2004; Johnson et al., 2010, 2011b). Subsequent orogenic events, including the 1820–1770 Ma Capricorn Orogeny, which was formerly thought to record the main Pilbara–Yilgarn collision (Tyler and Thorne, 1990; Evans et al., 2003), have since been shown to result from intracratonic reworking within the orogen (Sheppard et al., 2005, 2007, 2010a,b; Johnson et al., 2009).

Geological setting of seismic line 10GA–CP1

Seismic line 10GA–CP1 extends across the northern margin of the Capricorn Orogen, from the northern limb of the Turner Syncline, through the Rocklea Dome and southern Hardey Syncline, before crossing the Nanjilgardy Fault system into the Ashburton Basin to the south (Plate 1; Frontispiece 1). In doing so, it crosses greater than one billion years of geological history, ranging from the Archean Pilbara Craton granite–greenstone basement, to the overlying Archean to Paleoproterozoic supracrustal succession consisting of, in ascending order, the Fortescue, Hamersley, Turee Creek, Wyloo, and Capricorn Groups (Fig. 1). The Wyloo Group is generally discussed in terms of lower and upper divisions, based on the presence of a significant unconformity in the middle of the succession, representing a hiatus of up to 200 million years.

During the late Archean and Paleoproterozoic, the northern part of the orogen evolved from a rifted margin to a passive margin, before being converted to an active margin, and subsequently into a series of foreland

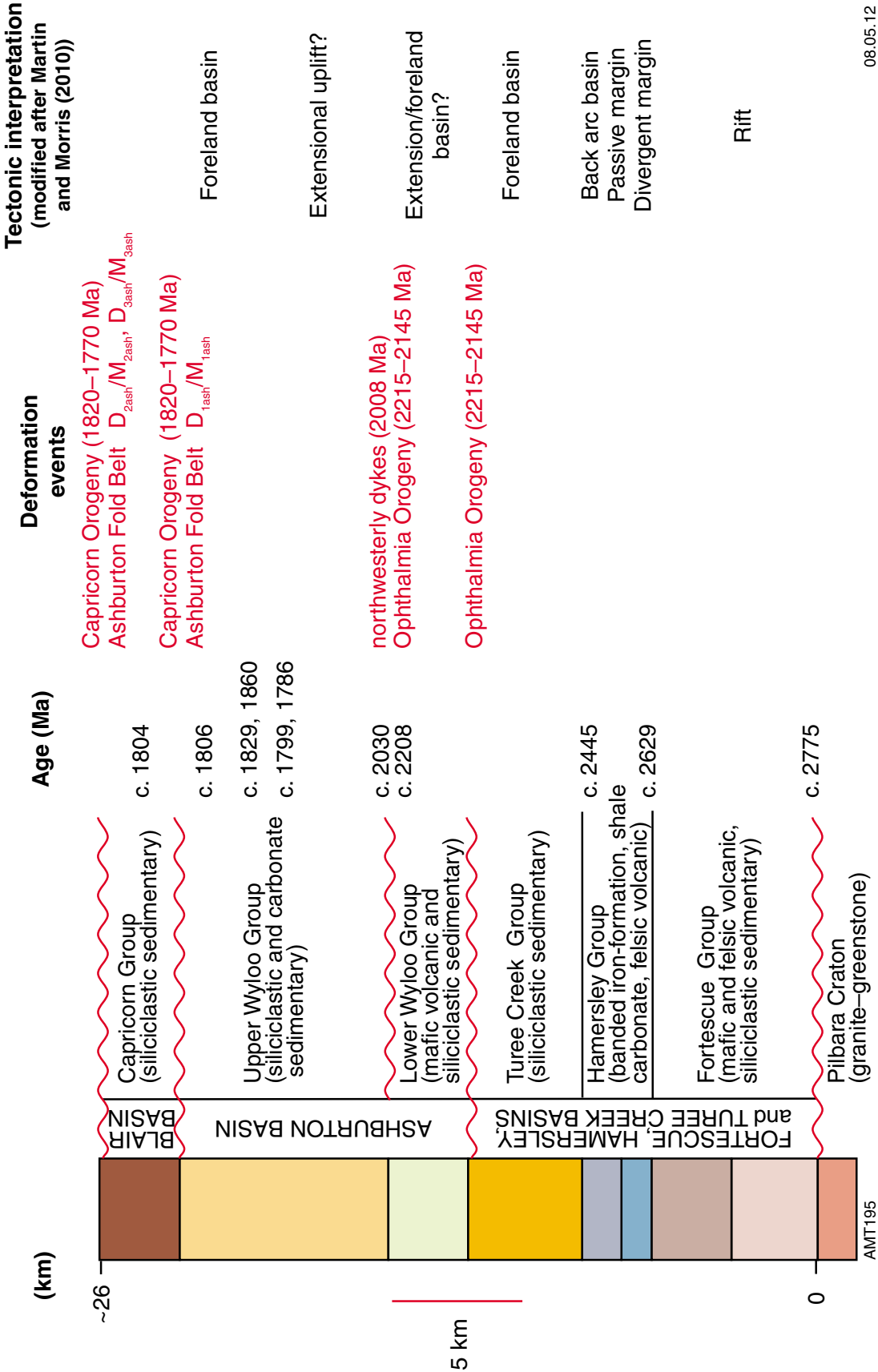


Figure 1. Generalized stratigraphy and deformation history of the northern Capricorn Orogen. The tectonic interpretation is modified after Martin and Morris (2010). Geochronology sources are cited in the text.

basins (Blake and Barley, 1992; Martin et al., 2000; Trendall et al., 2004; Martin and Morris, 2010). As part of this evolution, basement and supracrustal rocks were subject to two major orogenic events, the 2215–2145 Ma Ophthalmian Orogeny (Rasmussen et al., 2005) and the 1820–1770 Ma Capricorn Orogeny (Sheppard et al., 2010a,b).

Pilbara Craton basement

Within the northern Capricorn Orogen, Pilbara Craton granite–greenstone basement rocks are exposed in the Wyloo, Rocklea, and Milli Milli Domes, and in the Sylvania Inlier (Frontispiece 1; Plate 1). Greenstones are dominated by low-grade metamorphosed mafic volcanic and siliclastic sedimentary rocks. Granitic rocks are mostly biotite monzogranite and minor pegmatite, and are locally intruded by pre- Fortescue Group metadolerite dykes. The minimum age of the granite–greenstones is fixed by the c. 2775 Ma age of the overlying basal Fortescue Group (Trendall et al., 2004). Their maximum age is unknown, although comparison with similar granite–greenstone assemblages in the northern Pilbara Craton (Hickman and Van Kranendonk, 2008) suggests they probably formed between 3800–2830 Ma.

Fortescue Group: Pilbara Craton rifting

Granite–greenstone rocks are unconformably overlain by mixed volcano-sedimentary rocks of the Fortescue Group, formed during protracted rifting of the Pilbara Craton between c. 2775 and c. 2630 Ma (Blake, 1993, 2001; Thorne and Trendall, 2001; Blake et al., 2004; Trendall et al., 2004). In the northwestern and northeastern Pilbara, Fortescue Group sedimentation and volcanism was controlled by a northeast-trending extensional fault system (Blake, 2001), whereas in the southern Pilbara, Thorne and Trendall (2001) argued that the fault system was principally oriented east-southeast. It is also likely that during much of the early to middle rifting stages, these areas were separated from each other by topographic highs, such as the Yule–Sylvania high. These architectural controls created separate sub-basins (Blake, 1984; Thorne and Trendall, 2001), in which the local Fortescue Group stratigraphies have a broadly similar framework, but are each quite different in detail.

In the South Pilbara Sub-basin (Thorne and Trendall, 2001), the Fortescue Group is up to 6.5 km thick, and is subdivided, in ascending order, into seven formations, which are grouped into four major tectono-stratigraphic units (Fig. 2). From the base upwards, Unit 1 comprises the Bellary Formation and Mount Roe Basalt; Unit 2 comprises the Hardey Formation; Unit 3 comprises the Boongal, Pyradie, and Bunjinah Formations; and Unit 4 is the Jeerinah Formation. Thick mafic to ultramafic sills intrude much of the succession above the Mount Roe Basalt.

The Bellary Formation is a localized, 0.4 km thick, lacustrine or shallow-marine unit at the base of the Fortescue Group. It is conformably overlain by the 0 – 1.4 km thick Mount Roe Basalt that consists largely

of subaerial basaltic lavas and subaqueous volcanoclastic rocks. These are unconformably overlain by the Hardey Formation, which is up to 1.8 km thick, and consists of a wide range of sedimentary and mafic and felsic volcanic rocks, laid down in a continental to shallow-marine setting. The middle to upper parts of the Fortescue Group conformably overlie the Hardey Formation, and form a 3 km thick succession of subaqueous basaltic to komatiitic lavas and volcanoclastic rocks (the Boongal, Pyradie, and Bunjinah Formations). The Jeerinah Formation, the uppermost unit within the Fortescue Group, is up to 1.8 km thick, and consists of argillaceous sedimentary rocks interbedded with significant amounts of basaltic lava and volcanoclastic deposits.

The origin of the four major tectono-stratigraphic units of the Fortescue Group are interpreted as the result of deposition in an extensional tectonic setting. Units 1 and 2 were deposited in isolated fault-bounded sub-basins between c. 2775 and c. 2745 Ma (Trendall et al., 2004), under largely subaerial conditions. The separate sub-basins coalesced during the time of Unit 3 (2745–2715 Ma), when regional subsidence, accentuated by further normal faulting and tilting in the south, resulted in a change to coastal and deeper-shelf volcanism and sedimentation. Further subsidence during the time of Unit 4 (2715–2630 Ma) resulted in a major marine transgression and the establishment of deeper marine-shelf conditions over the entire Fortescue Basin. During this evolution, growth faulting appears to have had a major effect on Fortescue Group stratigraphy, with a marked increase in thickness and paleo- water depth being recorded when traced from north to south across the main faults that cut the sub-basin. Similarly, the thickness and frequency of intrusive sills in the Fortescue Group increases in a southward direction across the sub-basin fault system.

Hamersley Group: passive- to active-margin sedimentation and igneous activity

The 2.5 – 3.0 km thick Hamersley Group conformably overlies the Fortescue Group, and was deposited between c. 2630 and c. 2450 Ma, when the southern Pilbara evolved from a rifted to a passive continental margin and finally to an active margin (Morris and Horwitz, 1983; Blake and Barley, 1992; Thorne and Trendall, 2001; Trendall et al., 2004; Martin and Morris, 2010). Hamersley Group rocks reflect this largely deeper-marine distal setting, and are dominated by banded iron-formation (BIF), shale, chert, and fine-grained carbonate. Thin spherule layers, interpreted as resedimented impact debris, have also been recorded (Simonson and Hassler, 1997). Average (compacted) depositional rates for BIF, carbonate, and shale were 180 m per million years, 12 m per million years, and 5 m per million years, respectively (Trendall et al., 2004).

Seven major lithostratigraphic units are recognized within the Hamersley Group (Fig. 3). In ascending order, these are the Marra Mamba Iron Formation, Wittenoom Formation, Mount Sylvia Formation, Mount McRae Shale,

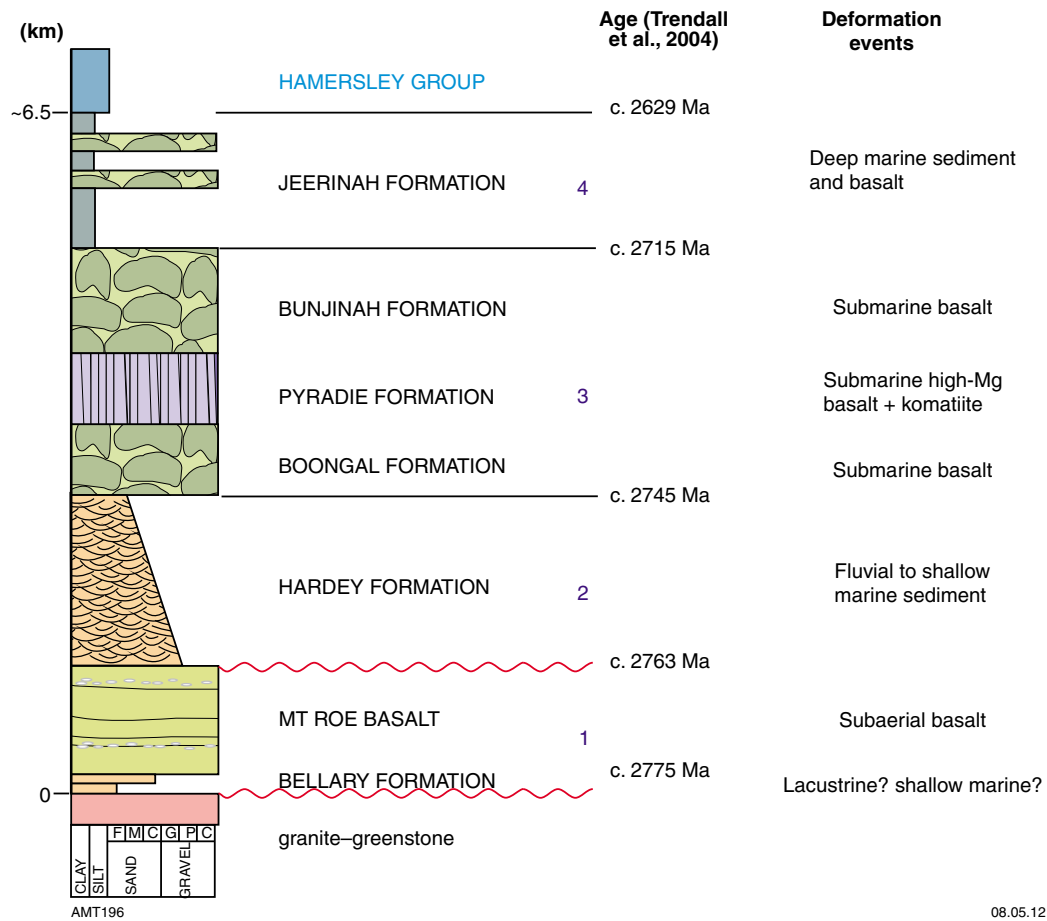


Figure 2. Stratigraphy of the Fortescue Group, showing tectono-stratigraphic units 1–4 and associated depositional environments. Major erosion surfaces are shown in red.

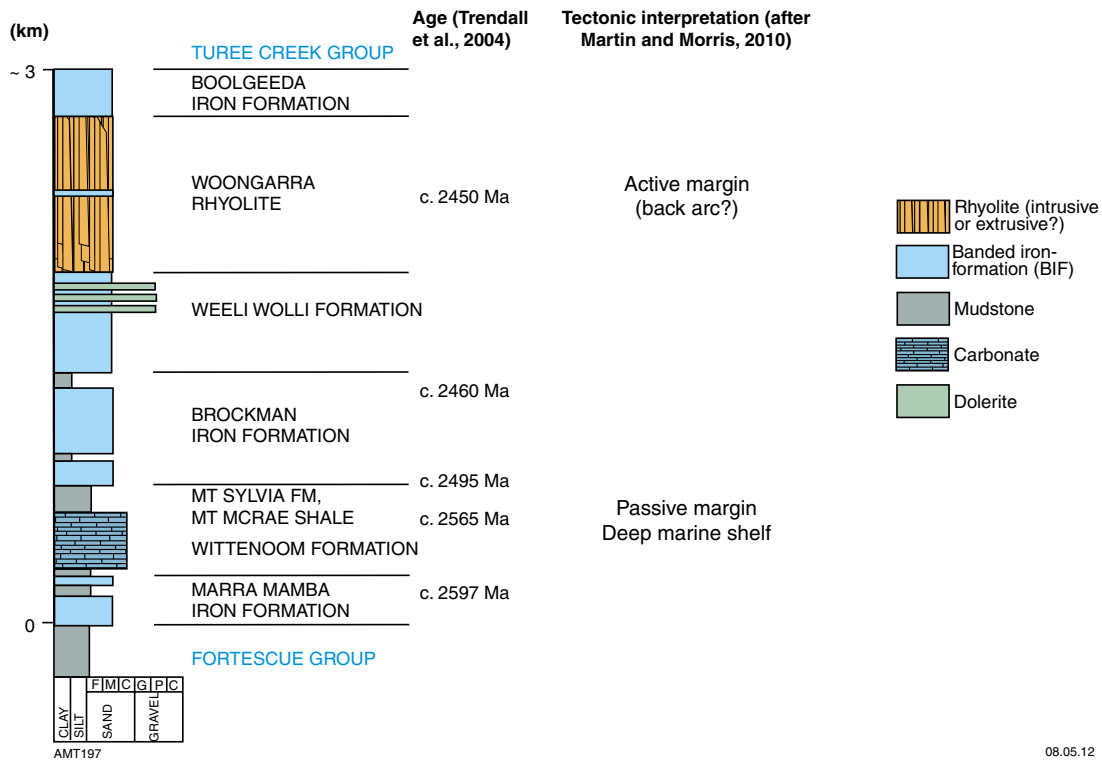


Figure 3. Stratigraphy of the Hamersley Group, showing the distribution of the principal rock types and interpreted tectonic setting.

Brockman Iron Formation, Weeli Wolli Formation, and Boolgeeda Iron Formation (Trendall and Blockley, 1970; Trendall, 1990; Simonson et al., 1993; Trendall et al., 2004). The status of the ~400 m thick Woongarra Rhyolite, which occurs below the Boolgeeda Iron Formation, is controversial. Trendall (1995) favoured an intrusive origin for this unit, whereas Doyle et al. (2001) suggested that it has both intrusive and extrusive components. Significant thicknesses of mafic intrusive rocks are interlayered within the Weeli Wolli Formation; however, it is not known if these are coeval with the Woongarra Rhyolite.

Turee Creek Group and lower Wyloo Group: foreland basin sedimentation and volcanism during the Ophthalmian Orogeny

Deposition of the uppermost Hamersley Group (Boolgeeda Iron Formation), plus the overlying Turee Creek and lower Wyloo Groups, took place during the early foreland-basin stage of the 2215–2145 Ma Ophthalmian Orogeny, when the Pilbara Craton collided with a continent to the south (Blake and Barley, 1992; Martin et al., 2000; Rasmussen et al., 2005). Recent evidence suggests that this southern continent was the Glenburgh Terrane of the Gascoyne Province (Occhipinti et al., 2004; Sheppard et al., 2004, 2010a; Martin and Morris, 2010; Johnson et al., 2010, 2011b). An alternative view was put forward by Krapež and McNaughton (1999), who considered the post-Turee Creek Group history in terms of two megasequences that record the opening and closure of an Atlantic-type ocean.

Along the southern Pilbara margin, the Turee Creek and lower Wyloo Groups have a maximum thickness of about 4 km and 3 km, respectively, and are separated by a significant angular unconformity (Fig. 4). This unconformity, together with another occurring within the upper Turee Creek Group, are interpreted to reflect the northward propagation of the Ophthalmia fold-and-thrust belt into the foreland basin (Martin and Morris, 2010). The maximum age of the Turee Creek Group – lower Wyloo Group stratigraphy is c. 2450 Ma, the age of the Woongarra Rhyolite, clasts of which are present in the Boolgeeda Iron Formation (Martin, 1999). The minimum age is c. 2210 Ma, the age of the Cheela Springs Basalt and the dolerite sills that intrude the upper Turee Creek Group (Martin et al., 1998; Müller et al., 2005; Martin and Morris, 2010). A c. 2030 Ma age for the Woolly Dolomite (Müller et al., 2005), usually considered to be the uppermost unit of the lower Wyloo Group (Thorne and Seymour, 1991), instead suggests that this unit was deposited in the interval between lower and upper Wyloo Group deposition.

The Turee Creek Group is subdivided, in ascending order, into the Kungarra, Koolbye, and Kazput Formations (Thorne and Tyler, 1996). Much of the Kungarra Formation is dominated by deep-marine mudstones and turbidite sandstones, interbedded with minor dolostone. A prominent glaciogenic diamictite, the Meteorite Bore Member (Trendall, 1976), outcrops in the upper part of the formation. Fluvial to shallow-marine sandstones of the Koolbye Formation overlie the Kungarra Formation,

and are themselves overlain by deltaic and shallow-marine sandstones, siltstones, and dolostones belonging to the Kazput Formation.

Fluvial to shallow-marine sandstones and conglomerates belonging to the ~300 m thick Beasley River Quartzite (lower Wyloo Group) unconformably overlie the Kazput Formation, and are conformably overlain by the ~2.5 km thick Cheela Springs Basalt. Martin and Morris (2010) argue convincingly that these continental tholeiite lavas are coeval with c. 2210 Ma dolerite sills that intrude the underlying Turee Creek Group. Throughout most of the southern Pilbara, the Cheela Springs Basalt is unconformably overlain by the upper Wyloo Group. An exception to this occurs on the Wyloo Dome, where the basalts are conformably overlain by the Woolly Dolomite, a 300 m thick succession of high- and low-energy shelf carbonates.

Upper Wyloo and Capricorn Groups: foreland-basin sedimentation and volcanism during the Capricorn Orogeny

Turee Creek and lower Wyloo Group rocks were deformed during the 2215–2145 Ma Ophthalmian Orogeny (Rasmussen et al., 2005), overlain locally by the c. 2030 Ma Woolly Dolomite, and then intruded by c. 2010 Ma northwest-trending dolerite dykes (Müller et al., 2005). Both the Ophthalmian folds and the c. 2010 Ma dykes were then truncated by the unconformity at the base of the upper Wyloo Group.

The upper Wyloo Group (Fig. 5) has an estimated thickness of about 7.5 km, and comprises, in ascending order, the Mount McGrath Formation, Duck Creek Dolomite, and Asburton Formation (Thorne and Seymour, 1991). A ~120 m thick, mixed mafic and felsic volcanic unit, the June Hill Volcanics, overlies the Duck Creek Dolomite north of the Wyloo Dome. The Ashburton Formation is unconformably overlain by siliciclastic, carbonate, and felsic volcanic rocks of the Capricorn Group, deposited following the early deformation stage (D_{1ash} — Ashburton Fold Belt) of the 1820–1770 Ma Capricorn Orogeny.

The age of the upper Wyloo Group is still poorly constrained. Evans et al. (2003) obtained an age of c. 1800 Ma for the June Hill Volcanics, whereas a tuffaceous unit in the overlying Ashburton Formation has been dated at c. 1829 Ma (Sircombe, 2003). The apparent contradiction in these ages, coupled with the c. 1804 Ma age of felsic volcanic rocks in the Capricorn Group (Hall et al., 2001) has led Evans et al. (2003) to suggest that deposition of the upper part of the upper Wyloo Group and of the Capricorn Group were strongly diachronous due to oblique basin closure during the Capricorn Orogeny.

The ~1.2 km thick Mount McGrath Formation comprises ferruginous conglomerate and sandstone, quartz sandstone, mudstone, and locally developed carbonate. Coarse-grained siliciclastic detritus was laid down on gravelly, braided deltas that fringed the northeastern margin of

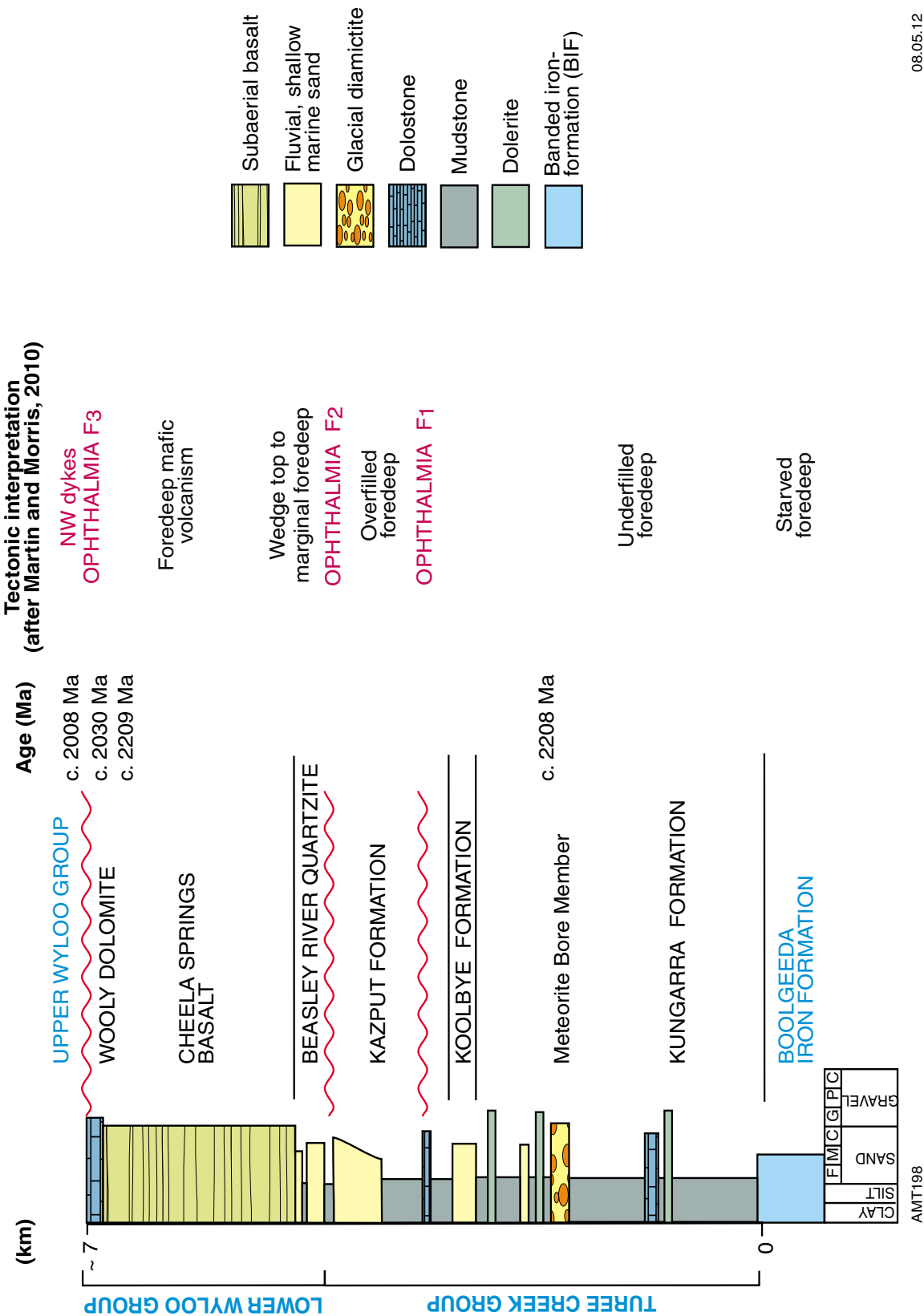
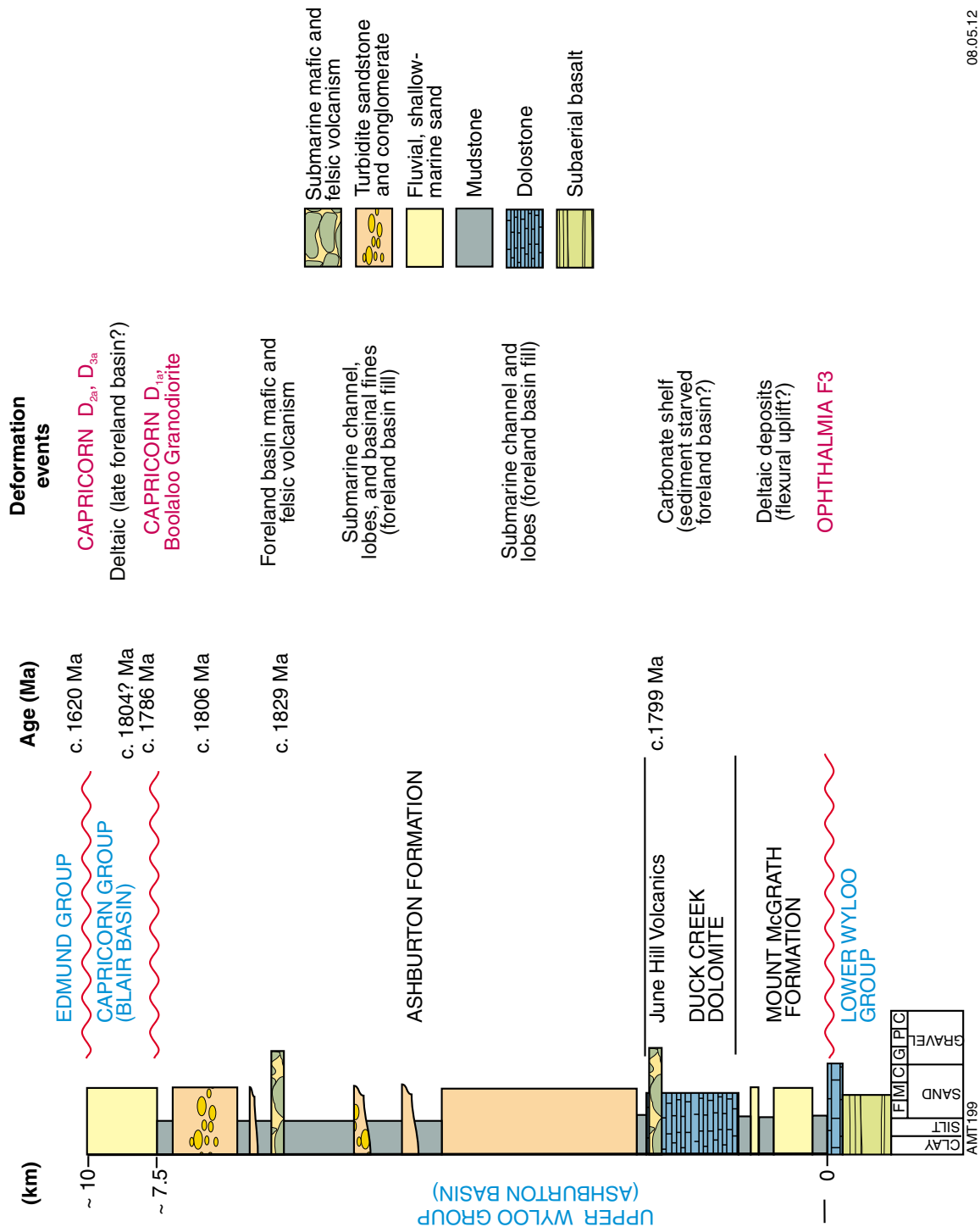


Figure 4. Stratigraphy of the Turee Creek Group and lower Wyloo Group, showing the distribution of the principal rock types, major deformation events, and interpreted tectonic settings. The sources for the geochronology are discussed in the text.



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Figure 5. Stratigraphy of the upper Wyloo Group, showing the distribution of the principal rock types, major deformation events, and interpreted tectonic settings. The sources for the geochronology are discussed in the text.

the Ashburton Basin. Mud, silt, and carbonate were deposited in open shelf waters further to the south and west. Maximum development of the deltaic complex occurred during deposition of the lower Mount McGrath Formation. The middle and upper parts of the formation record a gradual waning of siliciclastic supply; this was accompanied by a deepening of shelf waters throughout the basin (Thorne and Seymour, 1991).

Carbonate rocks of the ~1.0 km thick Duck Creek Dolomite conformably overlie the Mount McGrath Formation, and are dominated by stromatolitic inner-shelf facies and thin-bedded or conglomeratic outer-shelf and slope facies. Deposition of the lower Duck Creek Dolomite took place in a relatively deep-water, shelf, slope, and basin environment that existed along the northeastern margin of the Ashburton Basin. To the north and east, a belt of shallow, marine, inner-shelf carbonate deposition fringed the low-relief Pilbara Craton. The carbonate shelf underwent one major episode of progradation during the deposition of the middle Duck Creek Dolomite. The upper parts of the formation were laid down in a gradually deepening basin, which received increasing amounts of fine-grained siliciclastic detritus (Thorne and Seymour, 1991).

The Ashburton Formation has an estimated thickness of about 5 km, and conformably overlies the Duck Creek Dolomite, except in the northwestern part of the Ashburton Basin, where it is disconformable upon the June Hill Volcanics. Principal rock types are: mudstone; thin- to thick-bedded sandstone; conglomerate; BIF and chert; and local mafic and felsic volcanic rock. These deposits were laid down in a deep marine basin, with most sediment being derived from a Paleoproterozoic granitic source to the southeast, south of the exposed southern Pilbara margin (Thorne and Seymour, 1991; Sircombe, 2002).

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

Ophthalmia and Ashburton Fold Belts

The southern Pilbara margin was deformed during two Paleoproterozoic orogenic events related to the assembly and subsequent tectonic reworking of the West Australian Craton. The earliest of these, the 2215–2145 Ma Ophthalmian Orogeny (Rasmussen et al., 2005), formed the Ophthalmia Fold Belt and reflects collision of the southern Pilbara Craton with the Glenburgh Terrane of the Gascoyne Province (Occhipinti et al., 2004; Sheppard et al., 2004, 2010a; Martin and Morris, 2010; Johnson et al., 2010, 2011b). Subsequent crustal reworking during the 1820–1770 Ma Capricorn Orogeny created the Ashburton Fold Belt (Thorne and Seymour, 1991; Sheppard et al., 2010a,b).

Ophthalmia Fold Belt

Ophthalmia Fold Belt structures affect the Pilbara Craton basement and very low grade supracrustal rocks, up to, and including, the lower Wyloo Group (Tyler, 1991; Thorne and Seymour, 1991; Martin and Morris, 2010). These structures include regional scale, en-echelon, open to tight, upright folds; e.g. the Hardey Syncline, Turner Syncline, and Rocklea Dome (Frontispiece 1; Plate 1). These structures are cut by 2010 Ma, northwest-trending dolerite dykes (Müller et al., 2005), and are overprinted by younger Ashburton Fold Belt structures. Ophthalmia Fold Belt structures show decreasing strain when traced from southwest to northeast.

Three pre- and one post- lower Wyloo Group Ophthalmian events are recognized. The earliest structures, F_{1oph} , are represented by localized, though widely distributed, tight to isoclinal, bedding-parallel folds that are refolded by the later regional-scale Ophthalmian folds (Tyler, 1991). Martin and Morris (2010) record a minimum of three phases of roughly east–west trending Ophthalmian folds in the Hardey syncline. The earliest of these, F_{2oph} (F_{1oph} of Martin and Morris, 2010), affected all strata up to, and including, the middle Kazput Formation; however, these folds are truncated by the unconformity at the base of the upper Kazput Formation. Subsequent F_{3oph} folds (F_{2oph} of Martin and Morris, 2010) are open and upright, and affected all strata up to, and including, the upper Kazput Formation, but are truncated by the unconformity at the base of the lower Wyloo Group. Later F_{4oph} folding (F_{3oph} of Martin and Morris, 2010) post-dates the c. 2210 Ma dolerite sills that intrude the Beasley River Quartzite, but are cut by c. 2010 Ma northwest-trending dolerite dykes (Müller et al., 2005).

Martin and Morris (2010) also record the presence of a younger folding event that refolds the Hardey Syncline. Although it could be Ophthalmian in age, the absolute timing of this event is unclear, and it may be related to younger deformation; e.g. the Panhandle event (Taylor et al., 2001) or Capricorn Orogeny.

Ashburton Fold Belt

Ashburton Fold Belt structures affect the Pilbara Craton basement and very low grade supracrustal rocks, up to, and including, the lower Capricorn Group. These structures overprint those belonging to the earlier Ophthalmia Fold Belt (Tyler and Thorne, 1990; Thorne and Seymour, 1991; Martin and Morris, 2010). One pre- Capricorn Group event (D_{1ash}/M_{1ash}) and two post- Capricorn Group – pre- Edmund Group events (D_{2ash}/M_{2ash} and D_{3ash}) are recognized within the Ashburton Fold Belt. The timing of D_{1ash}/M_{1ash} was between c. 1806 and c. 1786 Ma, the latter being the age of the post- D_{1ash} Boolaloo Granodiorite (Krapež and McNaughton, 1999; Martin et al., 2005). D_{2ash}/M_{2ash} occurred between c. 1786 and c. 1738 Ma, the younger age limit being set by the age of gold mineralization at the Mount Olympus deposit (Sener et al., 2005). The youngest event, D_{3ash} , occurred between c. 1738 and c. 1620 Ma, based on the minimum age of

D_{2ash}/M_{2ash} and the maximum age of the overlying Edmund Group (Martin et al., 2005).

Thorne and Seymour (1991) recognized three structural zones (A, B, and C) within the Ashburton Fold Belt, based on the geometry of the D_{2ash} structures and the preservation of D_{1ash} structures. Zone A is dominated by dextral strike-slip faulting and refolding of Ophthalmia Fold Belt folds along the southwestern Pilbara margin. Zone B is developed in the Ashburton Formation, northeast of the Baring Downs Fault. It represents a relatively high-strain zone formed during D_{2ash} . As a result of this, the recognition of D_{1ash} structures within Zone B is generally difficult, because strong overprinting by D_{2ash} has resulted in the early cleavage (S_{1ash}) being coaxial, and often coplanar, with the later fabric (S_{2ash}). Zone C occupies the remainder of the Ashburton Fold Belt, between the southwestern boundary of Zone B and the Edmund Group unconformity. It is distinguished from Zone B by its generally lower level of D_{2ash} strain leading to the better preservation of D_{1ash} structures, and also by the presence of large-scale F_{2ash} folds and dextral strike-slip faults. Zone C also preserves evidence of D_{3ash} structures in the western Capricorn Range.

Ashburton Fold Belt D_{1ash}/M_{1ash}

Most of the evidence for the D_{1ash} deformation is based on the widespread S_{1ash} foliation — the marked angular unconformity between the Ashburton Formation and the Capricorn Group — and rare F_{1ash} folds (Thorne and Seymour, 1991; Martin et al., 2005). In many outcrops, evidence for D_{1ash} is present in the form of an early S_{1ash} cleavage developed subparallel to bedding. This cleavage is often crenulated by S_{2ash} , and with the increase in metamorphic grade in southwestern parts of the fold belt, it develops into a metamorphic schistosity.

In the Capricorn Range, the marked angular unconformity at the base of the Capricorn Group provides clear evidence that the Ashburton Formation was folded prior to deposition of the Capricorn Group. In addition, the Ashburton Formation shows a well-developed S_{1ash} penetrative cleavage that is not developed in the overlying Capricorn Group (Thorne and Seymour, 1991). At the eastern end of the Capricorn Range, the basal unconformity of the Capricorn Group dips north at 50° as a result of D_{2ash} folding. The S_{1ash} cleavage in the underlying Ashburton Formation dips $15\text{--}45^\circ$ south, whereas bedding dips (and youngs) southward at $0\text{--}30^\circ$. The rotation of bedding and S_{1ash} to their pre- D_{2a} orientation indicates that the tight F_{1ash} fold was characterized by a steeply southward-dipping axial surface.

The metamorphic grade is low throughout most of the Ashburton Basin; however, there is a general increase in grade and intensity of schistosity towards the west and southwest. Thorne and Seymour (1991) note that much of the Ashburton Basin is characterized by the quartz–chlorite–muscovite(–sericite) mineral assemblage in pelitic and psammitic rocks. To the southwest of the Capricorn

Range, medium-grade metamorphosed equivalents of the Ashburton Formation are represented by quartz–muscovite–biotite–cordierite–andalusite–garnet schists. Textural evidence suggests that porphyroblastic minerals (biotite, andalusite, cordierite, and garnet) grew both during and after the D_{1ash} deformation event, overgrowing a quartz, muscovite, and chlorite groundmass. The metamorphic schistosity (S_{1ash}) is typically deformed by F_{2ash} folds and crenulation cleavage.

Ashburton Fold Belt D_{2ash}/M_{2ash}

Most obvious folding and faulting in the Ashburton Fold Belt results from the second deformation event, D_{2ash} (Thorne and Seymour, 1991; Martin et al., 2005). Within Zone B, D_{2ash} deformation has resulted in tight to isoclinal, non-cylindrical F_{2ash} folds, with wavelengths of 5–200 m. Folds trend west to northwest, and are associated with a pronounced axial-plane cleavage (S_{2ash}) that generally dips $60\text{--}90^\circ$ to the southwest or northeast. Within the main body of Ashburton Formation rocks, faults are generally associated with subparallel quartz veins, and are frequently observed to cut out all, or part, of the northern limb of the F_{2ash} folds. This fact, coupled with lack of marker horizons in the Ashburton Formation and the tight to isoclinal folding, creates a false impression of a simple stratigraphy with southwesterly dipping beds throughout much of Zone B. In such cases, evidence for F_{2ash} folding comes from local reversals in younging direction, and the presence of isolated F_{2ash} fold closures within the otherwise uniformly dipping Ashburton Formation. F_{2ash} folds in Zone C are large (100–6000 m wavelength), non-cylindrical, and trend west to northwest. Most plunge $10\text{--}40^\circ$ (up to 80° locally) to the southeast or northwest. Axial planes dip steeply to the southwest or northeast. Close to the northern margin of Zone C, folds are open to tight (or locally isoclinal), but are generally more open in central and southwestern parts of the fold belt. S_{2ash} is a penetrative slaty cleavage in the more easterly outcrops, but is present as a crenulation cleavage further west.

Numerous west-northwesterly to north-northwesterly trending strike-slip faults are either parallel to the F_{2ash} fold axes, or crosscut them at a shallow angle. The most prominent of these faults are the Nanjilgardy, Baring Downs, and Blair Faults. Locally, a dextral displacement of up to 3 km can be measured. However, in general, the lack of marker horizons within the Ashburton Formation makes it difficult to accurately assess the amount of relative movement. The northern margin of the Capricorn Range is locally marked by a steep southward-dipping fault. Many faults are marked by a line of en echelon quartz veins, dipping $30\text{--}90^\circ$ northeast or southwest. Locally, they are associated with a second suite of steeply dipping veins that trend between north-northwest and north-northeast. Most veins consist of equant to prismatic, subhedral to anhedral quartz, with irregular goethite vugs (after sulfide). Locally, quartz crystals are kinked as a result of progressive fault movement. Most of the copper, gold, lead, and silver mineralization discovered to date in the Ashburton Fold Belt is associated with D_{2ash} faults and quartz veins.

Low-grade metamorphism accompanied D_{2ash} , causing the retrogression of biotite to chlorite and andalusite to sericite, and the growth of porphyroblastic muscovite.

Ashburton Fold Belt D_{3a}

In the western Capricorn Range, the traces of F_{2ash} folds in the Capricorn Group are folded such that the regional west-northwest trend swings firstly southwest, and then west-northwest near Irregully Creek. Locally, the fold limbs are cut by a steep, southwesterly dipping fracture cleavage. The geometry of the F_{3ash} fold structure suggests it may have formed in response to a localized late-stage sinistral movement, on a pair of strike-slip faults that occur along the northern margin of the Capricorn Range and beneath the northern edge of Edmund Group outcrop.

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