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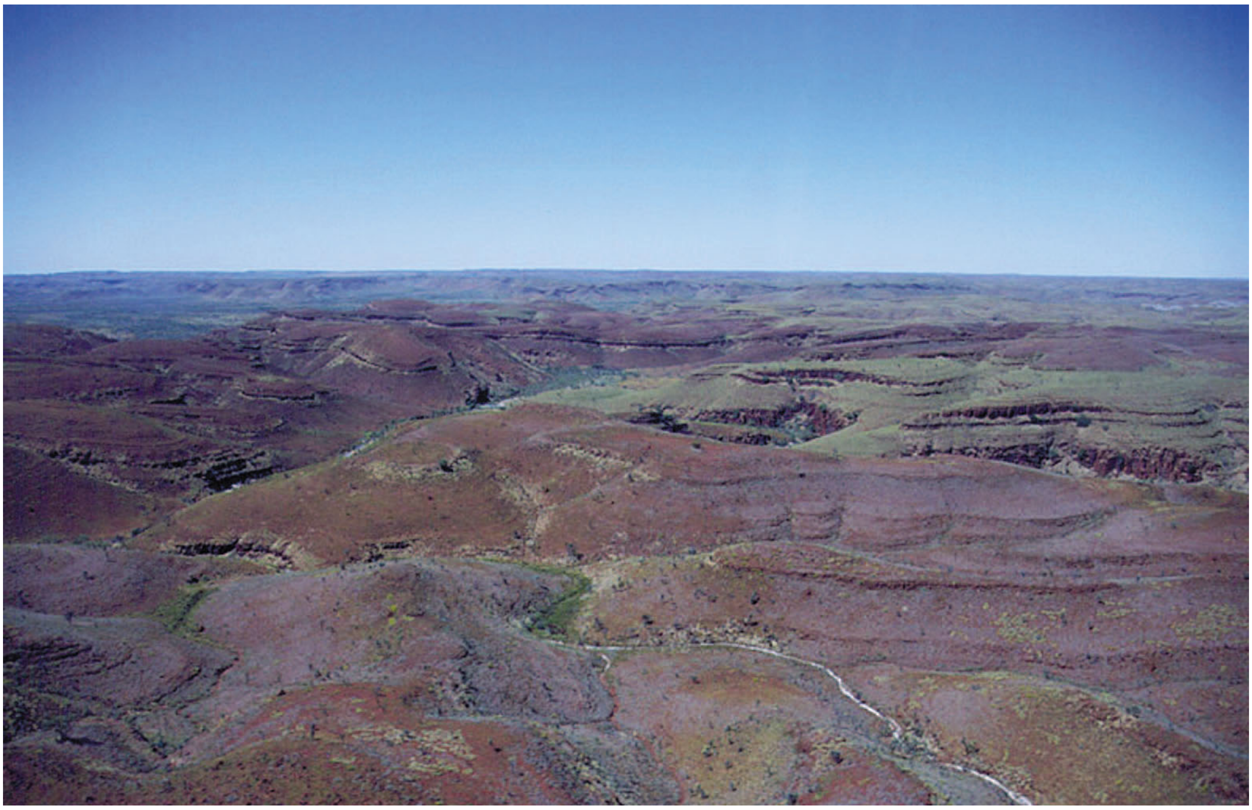
GOVERNMENT OF
WESTERN AUSTRALIA

GEOLOGY OF THE FORTESCUE GROUP PILBARA CRATON WESTERN AUSTRALIA

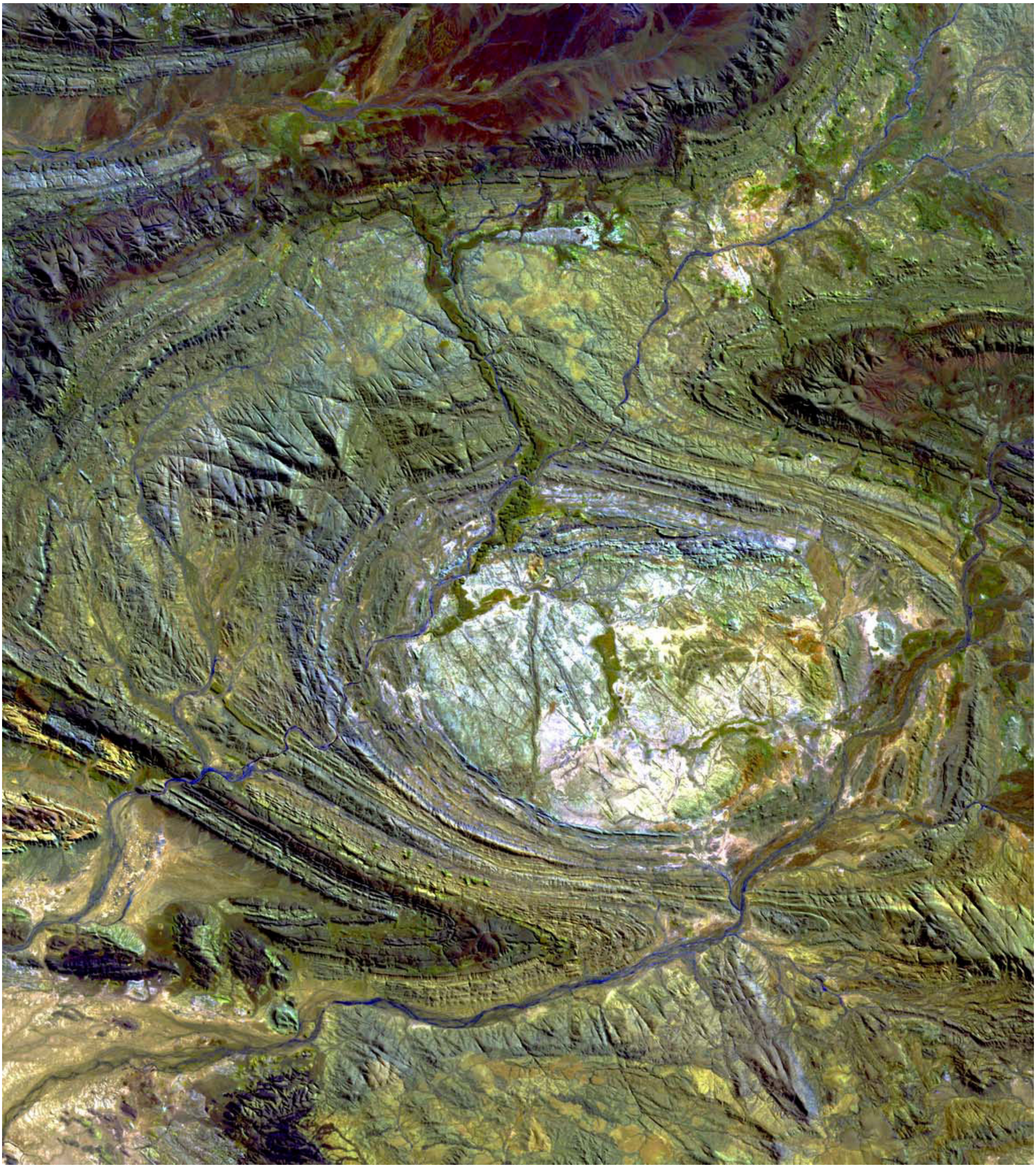
by A. M. Thorne and A. F. Trendall



GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

DEPARTMENT OF MINERALS AND ENERGY

GEOLOGY OF THE FORTESCUE GROUP, PILBARA CRATON, WESTERN AUSTRALIA



FRONTISPIECE

Landsat image (bands 4, 5 7) of the Fortescue Group unconformably overlying granite–greenstones of the Rocklea dome (right centre) in the south central Pilbara. Here, the Fortescue Group consists of the Mount Roe Basalt, preserved in the southeastern closure of the dome, overlain successively by the Hardey, Boongal, Pyradie, Bunjinah, and Jeerinah Formations. Hamersley Group rocks overlie the Fortescue Group and are preserved in the Boolgeeda, Hardey, and Turner synclines, located in the top left, bottom left, and right centre of the image respectively.

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Cover photograph:
Oblique aerial view of the Fortescue Group in central Chichester Range on the southwestern part of the MARBLE BAR 1:250 000 map sheet. Gently dipping mafic lava of the Kylena Formation is overlain by sedimentary and volcanosedimentary rocks of the Tumbiana Formation. Photograph by R. H. Smithies.

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Geology of the Fortescue Group, Pilbara Craton, Western Australia

by

A. M. Thorne and A. F. Trendall

Abstract

The Fortescue Group is a thick sequence of Archaean mafic and felsic volcanic and associated sedimentary rocks, which unconformably overlies Pilbara Craton granite–greenstones in the northwest of Western Australia. The succession has a maximum thickness of about 6.5 km and is divided into seven formations, which are grouped into four major tectono-stratigraphic units. From the base upwards, Unit 1 incorporates the Mount Roe Basalt and pre-Mount Roe Basalt sedimentary units, including the Bellary Formation; Unit 2 comprises the Hardey Formation; Unit 3 consists of the Kylena, Tumbiana, and Maddina Formations, and their lateral equivalents the Boongal, Pyradie, and Bunjinah Formations; and Unit 4 is the Jeerinah Formation. Uranium–lead zircon geochronology indicates that the Fortescue Group was deposited between about 2775 and 2630 Ma.

The Mount Roe Basalt is up to 2.5 km thick and consists largely of subaerial basaltic lavas, subaqueous basaltic (pillow) lavas, and water-lain volcaniclastic rocks. These are overlain by the Hardey Formation, which is up to 3 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. Middle to upper parts of the Fortescue Group are dominated by subaerial basaltic flows (Kylena and Maddina Formations) and coastal and nearshore-shelf sedimentary and volcaniclastic rocks (Tumbiana Formation) in the north Pilbara, and by subaqueous basaltic to komatiitic lavas and volcaniclastic rocks (Boongal, Pyradie, and Bunjinah Formations) in the south. The Jeerinah Formation, the uppermost unit within the Fortescue Group, consists largely of argillaceous rocks in the north, whereas basaltic lava and volcaniclastic rocks are abundant in the south. With the exception of the Mount Roe Basalt, a stratigraphy typical of the northern Pilbara is also recognized in the Gregory Range inlier.

Minor intrusions within the Fortescue Group consist mainly of mafic sills and dykes, the former being most abundant in the south Pilbara and northwest Pilbara sub-basins. The north-northeasterly trending Black Range Suite of mafic dykes is best developed in the northeast Pilbara; their intrusion was coeval with deposition of the Mount Roe Basalt.

Chemical analyses of 448 Fortescue Group rocks shows that virtually all ‘basalts’ are, chemically, basaltic andesites. Excluding felsic rocks, the only samples of clearly different composition are local high-Mg komatiitic lavas, mainly within the Pyradie Formation, and strongly mafic sills within the Jeerinah Formation. Sedimentary and volcanic rocks of the Fortescue Group have been associated with the small-scale production of gold, copper, lead, silver, zinc and fluorite, and an exploration interest in uranium.

The origin of the four major stratigraphic sequences of the Fortescue Group are interpreted as the result of deposition in an extensional tectonic setting. Sequences 1 and 2 (pre-Mount Roe Basalt to Hardey Formation) were deposited under largely subaerial conditions in isolated fault-bounded sub-basins. The separate sub-basins coalesced during Sequence 3 times 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 at the beginning of Sequence 4 times resulted in a major marine transgression and the establishment of deeper marine-shelf conditions over the entire Hamersley Basin. The major structural controls on Fortescue Group sedimentation and volcanism are discussed and comparisons made with Phanerozoic rift settings. Although the Fortescue Group has been described as a continental flood basalt province (CFBP), it differs from ‘typical’ CFBPs in at least three important respects: the group is not solely ‘continental’, its emplacement took considerably more time, and the main magma component was basaltic andesite rather than basalt.

KEYWORDS: Fortescue Group, Archaean, Hamersley Basin, Pilbara Craton, Bellary Formation, Mount Roe Basalt, Hardey Formation, Kylena Formation, Tumbiana Formation, Maddina Formation, Boongal Formation, Pyradie Formation, Bunjinah Formation, Jeerinah Formation, rifting, sedimentation, volcanism, continental flood basalt province.

Chapter summaries

Chapter 1: Introduction

The Fortescue Group is a thick sequence of mafic and felsic volcanic and associated sedimentary rocks covering an area of about 40 000 km^2 of the Pilbara region, in the northwest of Western Australia. Its main outcrop is an irregular strip extending from the lower Fortescue River, in the west, for about 550 km to the Gregory Range, in the east; important inliers lie farther south, within the Hamersley Range area, and a few scattered outliers also occur to the north. The Fortescue Group is part of the Pilbara Craton, which has an older granite–greenstone component formed largely between 3.5 and 2.9 Ga, and an unconformably overlying supracrustal succession called the Mount Bruce Supergroup. The three major and regionally concordant subdivisions of the Mount Bruce Supergroup, the Fortescue Group, Hamersley Group and Turee Creek Group, formed between about 2.8 and 2.4 Ga. The boundaries of the Pilbara Craton are either tectonic, against younger Precambrian rocks of the Capricorn Orogen, or are concealed by unconformably overlying Precambrian and Phanerozoic rocks.

The warm to hot, semi-arid climate of the Pilbara region is associated with generally thin soil and sparse vegetation cover, and the topography of the area reflects closely the underlying geology.

The earliest European inland exploration took place in 1861, and was followed quickly by subdivision into pastoral stations. Neither the pastoral industry, nor early mining, prospered. However, since the establishment of globally significant iron ore mining in the 1960s, and the later discovery of major gasfields offshore, the Pilbara has experienced massive advancement in communications, construction, and population growth.

Although some studies of the Fortescue Group (as part of the 'Nullagine' succession) were made in the first decade of the century, the first systematic mapping, by the Geological Survey of Western Australia (GSWA), took place in the 1960s. Important contributions were later made by universities, Australian (federal) Government organizations, and exploration companies.

This Bulletin is designed to provide a systematic descriptive account of the geology of the Fortescue Group. It focuses on the depositional environment of the group rather than post-depositional events, such as deformation and metamorphism, which are dealt with only briefly. Both deformation and metamorphism of the Fortescue Group tend to increase towards the margins of the Pilbara Craton, but even in the most tectonized and metamorphosed areas the original identity of the rocks remains confidently recognizable.

Chapter 2: Stratigraphic framework and geochronology

The first formal lithostratigraphic definition of the Fortescue Group, in 1963, superseded older reference to these rocks as the Nullagine Beds, Series, or Formation. Since then, 47 formal stratigraphic names have been defined within the group, and a further 30 informal names have been used in the literature. An independent set of unconformity-bounded units, based on sequence-stratigraphic concepts, has also been proposed.

This Bulletin applies an integrated approach to stratigraphic subdivision of the Fortescue Group. It uses a broad system of lithostratigraphy as the basis for description, as well as a range of additional stratigraphic tools. The Fortescue Group of the northern Hamersley Basin is divided into six formations, which are directly correlated with six formations of the southern Hamersley Basin, although the Bellary Formation, the lowest unit in the south, has no formally named equivalent in the north. The formations are grouped into four major tectono-stratigraphic units based on the presently known distribution of unconformities, nature of the volcanic and sedimentary facies, and geochronological and geochemical data. Unit 1 incorporates the Mount Roe Basalt and pre-Mount Roe Basalt sedimentary units; unit 2 comprises the Hardey Formation; unit 3 consists of the Kylena, Tumbiana, and Maddina Formations and their equivalents; and unit 4 is the Jeerinah Formation. This four-fold subdivision forms the basis for interpreting the tectonic setting of the Fortescue Group (**Chapter 14**).

Four Fortescue Group sub-basins have been recognized within the Hamersley Basin; these are named the northwest Pilbara sub-basin, Marble Bar sub-basin, northeast Pilbara sub-basin, and the south Pilbara sub-basin. The sub-basin margins are not clearly defined, and their recognition is largely for descriptive convenience.

Sensitive high-resolution ion microprobe (SHRIMP) analyses of zircons, especially of those from felsic volcanic units, has shown that earlier geochronological results from both the Rb–Sr and Pb–Pb methods reflect the ages of secondary (mainly metamorphic, and in the range 2.4–2.0 Ga) events rather than primary ages, and that Sm–Nd results from basalts give ages of mantle extraction rather than extrusive ages. SHRIMP ages from samples of Mount Roe Basalt suggest that deposition took place between 2775 and 2663 Ma. Data from the Bamboo Creek and Koongaling Volcanic Members of the Hardey Formation, and from a sill in the upper part of the Hardey Formation in the Gregory Range area all lie within the range 2768–2756 Ma.

A mixed zircon population was obtained from a tuffaceous sandstone in the Tumbiana Formation (Pillingini Tuff) near Mount Herbert, northwest Pilbara sub-basin. Results indicate an extrusion age of 2715 ± 6 Ma with detrital and/or xenocrystic grains giving ages of 2936 ± 6 and 3200–3350 Ma. A U–Pb zircon date of 2717 ± 2 Ma from a rhyolite in the Maddina Formation shows that this formation is of very similar age to the underlying Tumbiana Formation.

The younger age limit for the Fortescue Group is constrained by SHRIMP U–Pb zircon dates of 2690 ± 16 and 2684 ± 6 Ma gathered from the upper part of the Jeerinah Formation. In addition, a SHRIMP age of 2629 ± 5 Ma has been obtained from the very top of the Jeerinah Formation on NEWMAN*. Comparison between this date and those from the lower part of the stratigraphy implies that the Fortescue Group was deposited over an interval of about 140 m.y.

The youngest SHRIMP zircon ages from rocks of the North Pilbara Terrain unconformably underlying the Fortescue Group are slightly older than 2.85 Ga, implying that a depositional hiatus of about 80 m.y. preceded the abrupt onset of Fortescue Group deposition.

Chapter 3: Sedimentary and volcanic units below the Mount Roe Basalt

Various sedimentary rock units overlie the irregular surface of the basal unconformity of the Fortescue Group, and are conformably overlain by the Mount Roe Basalt. Substantial thicknesses occur in only two areas: in the Marble Bar sub-basin, and in part of the south Pilbara sub-basin, where they make up the Bellary Formation. Pre-Mount Roe Basalt Fortescue Group rocks consist mainly of conglomerate, sandstone, and argillite; basalt flows, hyaloclastite, and tuffaceous deposits also occur locally. Three non-volcanic sedimentary facies associations are recognized: fan-delta, braided fluvial, and topographic hollow fill.

A variety of volcanic rock types, including massive basalt and basaltic breccia, vesicular basalt, and mafic tuff occur in lower and upper levels of the Bellary Formation. They are interbedded with both subaerial and subaqueous fan-delta deposits and form about 25–30% of the exposed stratigraphy.

Pre-Mount Roe Basalt Fortescue Group rocks have been recorded from the four principal sub-basins of the Hamersley Basin. Attempts to construct a depositional model are hampered by sparse exposure, difficulties of correlation, and uncertainty as to whether presently existing sections represent original thicknesses. Notwithstanding these difficulties, basal Fortescue Group sedimentary units apparently were deposited on a rugged landscape of low to moderate relief, in which local topographic highs were controlled by the more resistant of the greenstone lithologies; areas underlain by granitoid were generally of more subdued relief. Thin, laterally impersistent bodies of alluvial/colluvial conglomerate and sandstone were deposited in small valleys and topographic hollows. Locally, finer grained sandstone and argillite accumulated in shallow, ephemeral lakes. Major sandy, braided fluvial systems drained into the Marble Bar area from granitoid uplands to the north and southwest. Near Bellary Dome, in the southwestern Hamersley Basin, a small fan-delta system built out toward the west-northwest in a small, possibly fault-bounded lake or marine embayment. Fan-delta sedimentation was accompanied locally by the extrusion of basaltic lava.

Chapter 4: Mount Roe Basalt

The Mount Roe Basalt, a widely distributed unit up to 2.5 km thick, consists mainly of basaltic flows and local pillow lava interbedded with minor tuff, hyaloclastite, and epiclastic rocks; subordinate non-volcanogenic sedimentary rocks include clast- and matrix-supported conglomerate, feldspathic quartz sandstone, and argillite. The formation is absent from the Sylvania Inlier. It either overlies granite–greenstone basement unconformably, or rests conformably upon localized, mostly thin, basal sedimentary units (**Chapter 3**). The Mount Roe Basalt is unconformably overlain by the Hardey Formation in all sub-basins except the south Pilbara sub-basin, where this contact is either conformable or unconformable.

Volcanic rocks within the Mount Roe Basalt can be subdivided into subaerial basaltic lavas, subaqueous basaltic (pillow) lavas, and water-lain volcaniclastic rocks. Subaerial basaltic flows are the dominant volcanic facies throughout the Hamersley Basin. Measured flow thicknesses range from less than a metre to 82 m; average thicknesses vary between 10 m in the south Pilbara sub-basin and nearly 25 m in the northwest Pilbara sub-basin. Most of the lavas originated as low- to moderate-viscosity flows that were close to their liquidus temperatures at eruption, with a low volatile content.

Basaltic pillow lava is estimated to form less than 2% of the Mount Roe Basalt. Most occurrences are recorded at or near the base of the succession in the northwest Pilbara, Marble Bar, and northeast Pilbara sub-basins. In the majority of outcrops the lavas are intimately associated with hyaloclastite and together they form laterally discontinuous units which are up to 800 m thick. These pillow lavas probably formed in lakes, themselves formed either by volcanic damming in areas of rugged topography, or by local faulting.

Volcaniclastic rocks are estimated to form less than 5% of the Mount Roe Basalt. They generally consist of hyaloclastite, and mafic and felsic lapilli tuff, although mafic agglomerate has been reported locally. Most occurrences of volcaniclastic rocks occur at, or near the base of, the formation. Thin and laterally impersistent accumulations of non-volcanogenic sedimentary rocks, apparently derived from local basement, are interbedded with the basalt flows in the northwest and south Pilbara sub-basins.

* Capitalized names refer to standard map sheets. Where 1:250 000 and 1:100 000 sheets bear identical names, the 1:250 000-scale sheet is implied unless otherwise specified.

Penecontemporaneous weathering profiles (palaeosols) are developed over flows of the Mount Roe Basalt. They are the oldest well-studied palaesols in the stratigraphic record, and provide evidence that the partial pressure of atmospheric oxygen in Mount Roe Basalt times was less than 8% of present day levels.

The regional thickness pattern of the Mount Roe Basalt suggests that it formed in separate northern and southern basalt provinces, divided by a west-northwesterly trending area devoid of basalt.

Chapter 5: Hardey Formation

The Hardey Formation includes a wide range of sedimentary and volcanic rocks, is up to 3 km thick, and is absent only from the western portion of the northeast Pilbara sub-basin. The sedimentary rocks include clast- and matrix-supported conglomerate, feldspathic quartz sandstone and pebbly sandstone, and argillite; the volcanic rocks include felsic and mafic volcaniclastic deposits, basaltic flows and local pillow lava. Dolerite and layered mafic sills locally form a significant part of the stratigraphy. In the north, the formation unconformably overlies both the Mount Roe Basalt and the granite–greenstone basement; in the south, it either unconformably overlies granite–greenstone rocks or rests conformably or disconformably on the Mount Roe Basalt.

The sedimentary rocks were deposited in five major palaeoenvironments corresponding to the following facies: alluvial fan and coarse-grained braided alluvial; sandy braided fluvial; lacustrine; deltaic; and shoreline. Alluvial fan and coarse-grained braided fluvial deposits consist of localized occurences of thick- to very thick bedded cobble and boulder conglomerate, interlayered with pebbly sandstone and coarse-grained sandstone. The sandy braided fluvial facies comprises thin to very thick beds of medium- to very coarse grained sandstone, granule and pebbly sandstone, and pebble conglomerate. Lacustrine facies rocks form lenticular units that generally range in thickness from a few metres to about 30 m. Principal rock types within this facies are mudstone and siltstone interbedded with variable amounts of fine- to very coarse grained sandstone and conglomerate. The deltaic facies is characterized by a mixed assemblage of argillite, sandstone, and conglomerate, which displays a wide range of sedimentary structures. These rocks typically consist of a succession of upward-coarsening and thickening units that range in thickness from about 20 m to over 300 m. The shoreline facies consists of medium- to very coarse grained quartz-rich sandstone and subordinate conglomerate; these rocks typically form units 1–10 m thick.

Volcanic rocks include felsic and mafic volcaniclastic deposits, basaltic flows and local pillow lava. In addition, massive quartz–feldspar porphyry has also been recorded. A varied suite of felsic volcaniclastic rocks, ranging from tuff to lapilli tuff and lapilli tuff breccia, occurs interbedded with non-volcanic facies or else forms part of the large felsic volcanic complexes represented by the Bamboo Creek and Lyre Creek Members. Massive felsic porphyry makes up most of the 800 m thick Bamboo Creek Member of the northeast Pilbara sub-basin. The porphyry exists in association with lapilli tuff, and lapilli tuff–breccia units of probable pyroclastic flow origin. This member is part of a felsic volcanic complex, but the mode of emplacement of many of the rocks is unclear. Mafic lava is interbedded with non-volcanic and volcaniclastic rocks in the Hardey Formation.

Chapter 6: Kylena and Boongal Formations

The names Kylena Formation and Boongal Formation are applied, within the northern sub-basins and the south Pilbara sub-basin respectively, to a widely distributed unit of mafic lava and volcaniclastic rock with subordinate felsic volcanic and sedimentary carbonate rocks which overlies the Hardey Formation. These rocks are absent from the Marble Bar sub-basin, a small area of the central Chichester Range, and from much of the Sylvania Inlier. The Kylena Formation is generally less than 600 m thick, but reaches a thickness of 2000 m in the northeast Pilbara sub-basin. The formation rests either conformably (especially in the northwest) or unconformably (especially in the northeast) upon the Hardey Formation, or unconformably upon granite–greenstone terrane. The Boongal Formation is 400–1000 m thick, and conformably overlies the Hardey Formation.

Volcanic rocks can be divided into subaerial and subaqueous basaltic lavas, and volcaniclastic rocks. Subaerial basaltic flows, up to 70 m thick, are the dominant volcanic facies within the Kylena Formation. Flows are divisible into thin amygdaloidal and massive to amygdaloidal types; both are bounded by well-defined flow surfaces, including scoriaceous flow-tops. Locally, thick massive to amygdaloidal flows have a lower cumulate zone. Subaqueous basaltic lavas are the dominant volcanic facies within the Boongal Formation. They comprise pillow lava and massive basalt flows and are intimately associated with hyaloclastite. Massive basalt flows range in thickness from 5 m to more than 100 m, and have undulatory to planar bounding surfaces; many flows can be traced laterally and vertically into pillow lava and hyaloclastite. The association of pillow lava, massive lava, and hyaloclastite indicates that much of the Boongal Formation was formed in a subaqueous environment. Volcanic rocks of the Kylena and Boongal Formations show a marked change from subaerial to submarine facies when traced from north to south across the Pilbara Craton. In addition, the succession generally thickens toward the south.

A variety of non-basaltic volcanic rocks have been reported from the Kylena Formation in the west and east Pilbara. These include andesite, high-Mg andesite, dacite and rhyolite.

Non-volcanic sedimentary rocks form less than 1% of the Kylena Formation, and have not been recorded from the Boongal Formation. They are most abundant in the Meentheena Centrocline and consist of thin (<3 m) units of partially silicified stromatolitic carbonate and volcaniclastic debris overlying subaerial basalt flow tops.

Chapter 7: Tumbiana and Pyradie Formations

The Tumbiana Formation outcrops in a narrow strip extending from Cape Preston in the northwest Pilbara to Warrawagine Homestead in the northeast. The lateral equivalent of the Tumbiana Formation, the Pyradie Formation, extends throughout the the south Pilbara sub-basin and is equivalent to ultramafic rocks within a previously recognized upper mafic volcanic unit in the southern part of the Sylvania Inlier. The Tumbiana Formation of the northern Pilbara Craton is commonly less than 200 m thick and is either conformable upon the Kylena Formation or is unconformable upon granite–greenstone terrane. In the former situation, the widespread occurrence of flow tops on the uppermost Kylena Formation flows suggests that there was little or no erosion immediately prior to deposition of the Tumbiana Formation. In the south Pilbara sub-basin, the Pyradie Formation is up to 1200 m thick and the contact between it and the underlying Boongal Formation appears to be conformable.

The Tumbiana Formation consists mainly of coastal and nearshore shelf facies comprising stromatolitic and clastic carbonate, argillite, sandstone, primary and reworked tuff, and minor conglomerate; there is a minor mafic lava component. In contrast the Pyradie Formation consists mainly of volcanic and volcaniclastic rocks laid down in a deeper shelf setting. These include pyroxene spinifex-textured basalt flows and pillow lava, hyaloclastite, komatiite, and minor chert and tuffaceous argillite.

The palaeogeographic reconstruction for the Tumbiana and Pyradie Formations is similar to that for the Kylena and Boongal Formations: volcanic rocks and non-volcanic sedimentary rocks both show a change from coastal and shallow-marine shelf facies in the north to deeper, outer-marine shelf facies in the south. In addition, there is a general thickening of the succession towards the south. The distribution of pyroclastic air-fall tuff suggests there were at least two major hydrovolcanic eruptive centres in the north Pilbara during Tumbiana Formation deposition. The first of these, in the northeast Pilbara, was active mostly during deposition of the lower Tumbiana Formation; the second centre affected the northwest Pilbara during deposition of the upper part.

Chapter 8: Maddina and Bunjinah Formations

The laterally equivalent Maddina Formation and Bunjinah Formation are present over most of the central and southern Pilbara Craton, and conformably overlie the Tumbiana Formation and Pyradie Formation respectively; the formation is absent from the Marble Bar sub-basin.

A U–Pb zircon date of 2717 ± 2 Ma from a rhyolite from the Maddina Formation indicates that this formation is similar in age to the underlying Tumbiana Formation.

The Maddina Formation and Bunjinah Formation consist mainly of basalt flows, pillow lava, fine- to coarse-grained and mafic volcaniclastic rocks with subordinate dacite and rhyolite. Non-volcanic sedimentary rocks, including stromatolitic carbonate and quartz sandstone, are also recorded. The Maddina Formation of the northern Pilbara Craton is commonly between 500 and 1000 m thick; in the south Pilbara sub-basin the Bunjinah Formation attains a maximum thickness of 2.4 km, but is mostly between 800 and 1200 m thick.

Volcanic facies are similar to those in the Kylena and Boongal Formations and can be subdivided into subaerial basaltic lavas, subaqueous basaltic lavas, and volcaniclastic rocks. Subaerial basaltic flows, up to 60 m thick, are the dominant volcanic facies within the Maddina Formation. Flows are divisible into thin amygdaloidal and massive to amygdaloidal types; both are bounded by well-defined flow surfaces. Most flows appear to be laterally discontinuous over distances under 2 km. Flow tops generally exhibit irregular to broadly symmetrical undulations and are rarely scoriaceous. Massive to amygdaloidal flows are the most abundant flow type in the Maddina Formation. Exceptionally large amygdales, up to 1 m across, are a feature of the subaerial flows in the upper Maddina Formation of the northwest Pilbara sub-basin.

The basaltic composition and aphyric texture of most of the lavas suggest that they originated as low- to moderate-viscosity flows that were close to their liquidus temperatures at eruption. In addition, the general lack of evidence for contemporaneous explosive fragmentation suggests either that the initial content of exsolved volatiles was low, or that levels were high, but were reduced by non-explosive degassing or by volatile loss during a previous phase of explosive activity.

Subaqueous basaltic lavas are the dominant volcanic facies within the Bunjinah Formation. They comprise pillow lava, massive basalt flows, and minor vesicular basalt; all facies are intimately associated with hyaloclastite. Pillowed units range in thickness from a few metres to many tens of metres. Massive basalt flows range in thickness from about 5 m to over 100 m and have undulatory to planar bounding surfaces; many can be traced laterally and vertically into pillow lava and hyaloclastite. The association of pillow lava, massive lava, and hyaloclastite indicates that most, if not all, of the Bunjinah Formation was formed in a subaqueous environment, probably submarine. Strongly vesicular flows, despite their similarity to subaerial lavas in the Maddina Formation, are also thought to have been deposited in a submarine setting.

Volcaniclastic rocks comprise hyaloclastite, mafic tuff, and lapilli tuff. Mafic tuff and lapilli tuff, and associated non-volcanic sedimentary rocks, are an important facies locally within the Maddina Formation; hyaloclastite may be the dominant rock type within parts of the Bunjinah Formation. The basaltic breccias are interpreted as hyaloclastites, produced by the rapid cooling and fracturing of basalt during contact with water.

A suite of non-basaltic volcanic rocks is present in the Maddina Formation in the west Pilbara. These include andesite, high-Mg andesite, dacite, and rhyolite.

Non-volcanic sedimentary rocks comprise quartz sandstone and stromatolitic carbonate. They have been recorded only from the Maddina Formation where they are commonly found in association with volcaniclastic facies. The compositional and textural maturity of the sandstones, and the varied assemblage of sedimentary structures, indicate that deposition took place in an environment characterized by fluctuating low- to high-energy conditions. A coastal setting is also considered probable for the stromatolitic carbonate units, although they display few features which would allow identification of the specific depositional environment.

The palaeogeographic reconstruction for the Maddina and Bunjinah Formations is similar to that for the Kylena and Boongal Formations. Volcanic rocks show a change from subaerial to submarine facies from north to south across the Pilbara and, in addition, there is a general thickening of the succession towards the south.

Chapter 9: Jeerinah Formation

The Jeerinah Formation is present over most of the central and southern Pilbara Craton but absent from the Marble Bar sub-basin. No evidence of angular discordance was observed between the Jeerinah Formation and the underlying Maddina or Bunjinah Formations; it is conformably overlain by the Marra Mamba Iron Formation of the Hamersley Group. The formation is commonly between 100 and 500 m thick in central parts of the craton. In the south it is up to 1250 m thick, although up to half of this thickness may consist of coarse-grained mafic sills.

The main lithologies of the Jeerinah Formation are argillite, sandstone, dolomite, and chert, and a variety of volcanic rocks including basalt flows, pillow lava, fine- to coarse-grained mafic volcaniclastic rocks, and felsic volcaniclastic rocks. Argillaceous sedimentary deposits are the dominant lithology in northern areas, whereas basaltic lava and volcaniclastic rocks are abundant in parts of the southern Pilbara. The only formal subdivisions recognized are the Woodiana (Sandstone) Member and the Nallanaring (Volcanic) Member, both locally restricted.

Non-volcanic sedimentary rocks can be grouped into two major facies: nearshore shelf and offshore shelf. The former characterizes the lower part of the Jeerinah Formation, whereas the latter constitutes most of the middle to upper levels. Nearshore facies comprise quartz sandstone and subordinate argillite, laminated and stromatolitic chert, and chert breccia. These rocks are the main component of the Woodiana Member, the 15–60 m-thick basal unit of the Jeerinah Formation on northern parts of the Pilbara Craton. Offshore facies comprise mudstone and siltstone, with subordinate chert, carbonate, and sandstone. On northern parts of the Pilbara Craton these rocks are the main component of the Jeerinah Formation above the Woodiana Member. In the south Pilbara they occur throughout the stratigraphy and are interbedded with volcanic facies.

Volcanic rocks within the Jeerinah Formation are dominated by subaqueous facies comprising basaltic pillow lava and massive lava, basaltic volcaniclastic rocks, and felsic volcaniclastic rocks. Subaqueous basaltic lavas are the dominant volcanic facies; they comprise pillow lava and massive basalt flows and are intimately associated with hyaloclastite. Volcaniclastic rocks comprise basaltic hyaloclastite, and felsic tuff and lapilli tuff.

The palaeogeographic reconstruction for the lower Jeerinah Formation is similar to that proposed for the Maddina and Bunjinah Formations. Although subaerial facies are not recorded, nearshore marine-shelf facies dominate this part of the stratigraphy in northern and northeastern parts of the craton. South of a line from Western Creek in the Sylvania Inlier to the northern margin of the Jeerinah Anticline, equivalent levels of the Jeerinah Formation comprise deeper marine argillite and chert, and submarine mafic volcanic rocks. The upper part of the Jeerinah Formation records a craton-wide change to deeper marine conditions. Argillaceous sediments with minor chert, carbonate, storm-deposited sandstones, and felsic tuff were deposited in the north and northeast. Submarine mafic and felsic volcanic material accumulated with the offshore sediments in more southern areas.

Chapter 10: Gregory Range inlier

The Fortescue Group rocks of the Gregory Range inlier form a 120 km-long strip along the eastern edge of the Pilbara Craton. Phanerozoic rocks of the Wallal Embayment, along the Oakover Valley, separate the inlier from the main northeast Pilbara sub-basin to the west. To the east they abut the Gregory Granitic Complex, an extensive belt of heterogeneous granitoid rocks not dealt with in this Bulletin. The Gregory Range inlier is dealt with in a separate chapter because of the structural complexity and unusual stratigraphic features of this outcrop area of the Fortescue Group.

The inlier is cut by abundant faults subparallel to its length. Selected larger ones, usually vertical or steeply east-dipping and commonly marked on the ground by conspicuous quartz filling, are identified and named as main faults that divide the rocks into tectonically discrete elongate belts, or 'slices'. These five main faults are, from west to east, the Southwest, Antiform, Baramine, Y, and Camel Hump faults. The five slices which they bound, again from west to east, are called the Warri Warri, Newdegate, Tanguin, Isabella, and Lochinvar slices. Penetrative deformation, associated with a well-developed steep cleavage subparallel to the length of the inlier, is heterogeneously distributed, both geographically and stratigraphically, throughout the inlier.

The thickness of the Fortescue Group varies markedly, and sharply, between these five slices; thus the complete thickness to the top of the Jeerinah Formation, is about 3 km in the Warri Warri and Isabella slices but over 7 km in the adjacent Newdegate slice. However, these thicknesses can be only roughly estimated, since the basal unconformity of the Fortescue Group is not present in the inlier. Instead, the lowest stratigraphic unit (the Koongaling Volcanic Member of the Hardey Formation) is gradational downwards through granophyre into the granitoids of the Gregory Granitic Complex, over which it formed a coeval volcanic and hypabyssal carapace. Differences in both thickness and lithology between the five major tectonic slices are confined largely to the lower part of the Fortescue Group succession. Thus, the Jeerinah Formation, Maddina Formation, and Tumbiana Formation all closely resemble in both respects their equivalents in the northwest Pilbara sub-basin, to the west; however, the Kylena Formation, and most particularly the Hardey Formation, are substantially different. The Mount Roe Basalt is not recognized at all.

The Koongaling Volcanic Member is present in all five slices. It consists of a sequence of porphyritic rhyolites and cogenetic tuffs at least a kilometre thick; unusual large-scale curved banding (ogives) that is locally present has probably been caused by shallow sub-surface shearing during extrusion rather than being a reflection of discrete lava flows. Overlying this member, a variable thickness of both epiclastic and tuffaceous shallow-water sedimentary rocks with locally intercalated felsic and mafic volcanics has been designated the Warri Warri Member in the Warri Warri and Newdegate slices and the Tanguin Member in the Tanguin slice. The application of the name Hardey Formation in the Gregory Range inlier is based on the similarity between the Bamboo Creek Member of the northeast Pilbara sub-basin and the demonstrably coeval rhyolitic rocks of the Koongaling Volcanic Member. The upper limit of the Hardey Formation is arbitrarily designated in each slice at the base of thick stacked basaltic lava flows similar to those of the Kylena Formation, although these are intercalated in the Newdegate slice with felsic and tuffaceous sedimentary rocks.

Above the Kylena Formation the upper units of the Fortescue Group of the Gregory Range inlier (Tumbiana, Maddina, and Jeerinah Formations) are generally similar in both facies and thickness to those units in the northeast Pilbara sub-basin, apart from a greater abundance, thickness, and variety of volcanogenic rocks in the Jeerinah Formation.

Chapter 11: Intrusive igneous rocks

Minor intrusions (mainly mafic sills and dykes) are associated with all formations of the Fortescue Group, but vary in thickness and abundance between different stratigraphic levels, and also between different outcrop areas of individual formations.

Sills are most abundant in the south Pilbara and northwest Pilbara sub-basins. Only two, both in the northwest Pilbara sub-basin, have been formally named: the Cooya Pooya Dolerite intrudes the Lyre Creek Member of the Hardey Formation over a strike length of 85 km, and consists of a lower sill about 120 m thick and an upper sill about 40 m thick; the Gidley Granophyre, a sheet of granophyre and associated quartz gabbro about 900 m thick, was intruded along the basal unconformity of the Fortescue Group on the Burrup Peninsula and in the Dampier Archipelago. Other important sill occurrences include the mafic sills of the Jeerinah Formation, which locally form about 60% of the formation thickness, and layered (ultramafic to felsic) sills of the Hardey and Pyradie Formations of the south Pilbara sub-basin.

The Fortescue Group is cut by west-southwesterly to west-northwesterly trending, and north-northeasterly trending mafic dykes, particularly along the southern and western margin of its outcrop area. The north-northeasterly trending Black Range Suite is well developed in the northeast Pilbara sub-basin and has a U–Pb baddeleyite age of 2772 ± 2 Ma. This set is coeval with the Mount Roe Basalt and is overlain by the upper part of the Hardey Formation. These dykes probably fed early mafic flows of the Fortescue Group, although the relationship is hard to demonstrate unequivocally. Coeval felsic dykes in the Meentheena area have similar stratigraphic relationships.

Chapter 12: Geochemistry

Most (376) of 448 of Fortescue Group samples analysed are of mafic to ultramafic rock from stratiform units to which the name ‘basalt’ has been generally applied. The remainder include six samples from mafic dykes related to the Fortescue Group, ten samples of felsic rocks, and 56 analyses of various pyroclastic and sedimentary rocks. All analyses are included in a PC-based file called FORTGRP.GDA. Of these data, 248 analyses are from published sources, 103 are new analyses, and 97 originate from unpublished theses.

To detect any systematic correlation of chemical composition of the mafic and ultramafic rocks with either stratigraphic position or geographic location within the total outcrop area, these 376 mafic and ultramafic analyses are classified into 42 groups based on these two criteria. These 42 groups are then further split into 11 ‘clusters’, lettered A to K. Assessment using these groupings shows that virtually all ‘basalts’ of the Fortescue Group are, chemically, basaltic andesites with a silica content about 57.5% and total alkalies about 4.6%. The only rocks of clearly different composition are the sills of the Jeerinah Formation, which are markedly more mafic (mean SiO_2 about 50.6%), and local high-Mg komatiitic rocks, mainly within the Pyradie Formation. The effects of weathering and/or metamorphism are regarded as having minimal impact and therefore the compositions, as analysed, reflect those of the erupted magmas. The mafic lavas of the Fortescue Group do not appear to have a well developed ‘chemical stratigraphy’ of the type that has been proposed for other major mafic lava provinces.

Chapter 13: Economic geology

No major mineral deposits have been located within rocks of the Fortescue Group, and there is no current mineral production from the group. However, there has been small-scale production of gold, copper, lead, silver, zinc, and fluorite, and exploration interest in uranium.

Small quantities of gold have been won from conglomerates of the Hardey Formation, and in the basal part of the Mount Roe Basalt in the vicinity of Nullagine, and in the Marble Bar sub-basin; these are probably palaeoplacer deposits. Gold has also been won from Cainozoic alluvium and colluvium close to these occurrences. Minor quantities of gold have also been reported from Capricorn Orogen veins and shears in Fortescue Group rocks in the South Pilbara sub-basin, where principal occurrences are the Paulsen and Belvedere sites, and a small prospect near Paraburdoo.

The only recorded production of copper is from Capricorn Orogen quartz veins cutting the lower to middle Fortescue Group in the Wyloo Dome and from the Jeerinah Formation in the Wonmunna area, but other minor occurrences have been recorded from the Jeerinah Formation of the south Pilbara sub-basin and the Braeside area of the Gregory Range inlier. Small quantities of lead, silver, and zinc have also come from the Braeside area. There are significant quantities of fluorite in veins cutting the Fortescue Group of the Meentheena area, and although there was some production, this has now ceased.

Most exploration for uranium has been directed towards possible palaeoplacers within Hardey Formation conglomerates in the northeast and northwest Pilbara sub-basins, in the same 'Witwatersrand' environment as the small gold discoveries; no deposits of economic grade have been discovered.

Chapter 14: Synthesis and discussion

The lithostratigraphic units of the Fortescue Group can be arranged into four informal 'depositional sequences', numbered upwards as follows: 1. Mount Roe Basalt, Bellary Formation, and other pre-Mount Roe Basalt units; 2. Hardey Formation; 3. Kylena and Boongal Formation, Tumbiana and Pyradie Formations, Maddina and Bunjinah Formations; 4. Jeerinah Formation and Hamersley Group. The rocks of the group were deposited on granite–greenstone terrain of the Pilbara Craton; prior to Fortescue Group deposition, this probably extended outside the present craton margins.

Sequence 1 is up to 4.5 km thick and is present in the southwest, northwest, and northeast of the Hamersley Basin; it is absent from the Gregory Range inlier and much of the central and southeast Pilbara. The rocks of Sequence 1 are of two main types: shallow-water epiclastic (alluvial, fluvial, and colluvial) sediments, and generally overlying stacked basaltic lava flows. The latter were extruded in separate northern and southern areas from a north-northeasterly trending dyke suite. Sequence 1 rocks were deposited on a rugged, subaerial landscape with local relief up to more than 500 m. Local topographic highs were controlled by more resistant supracrustal lithologies; areas underlain by granitoids were generally of more subdued relief.

The transition from Sequence 1 to Sequence 2 is marked by an expansion of the depositional area, accompanied by a change of the volcanism associated with the epiclastic sedimentation from stacked basaltic flows to large-scale but localized felsic activity. In the northwest and northeast Pilbara there was local faulting, tilting, and erosion of Sequence 1 rocks before deposition of Sequence 2, but elsewhere the contact between the two is conformable or disconformable. The time interval between them too short for measurement by the best current geochronological methods.

Sequence 2 is up to 3 km thick; deposition was dominantly continental, including braided alluvial, alluvial fan, and lacustrine sediments. Local depositional environments were largely controlled by growth faults. The thickness and nature of the felsic volcanism associated with Sequence 2 deposition show marked local variation. The strongest volcanic development is in the Gregory Range inlier, where Sequence 2 is the lowest Fortescue Group unit recognized. The rhyolitic volcanic material formed the roof of an underlying coeval and consanguineous plutonic complex, the Gregory Granitic Complex.

Sequence 3 is generally conformable over Sequence 2, but there is local angular discordance; in the southern part of the basin the contact is a major marine-flooding surface. Total sequence thickness varies from about 3 km in the south to 1.5 km in the north, although the locally anomalous development in the Newdegate slice of the Gregory Range inlier reaches 3.5 km. Sequence 3 consists largely of basaltic volcanic rocks: stacked subaerial flows in the north and submarine lavas in the south. However, there was an important median interlude of felsic volcanism and associated sedimentation.

The conformable transition from Sequence 3 to Sequence 4 is again related to a major marine transgression, which extended across the Pilbara from the south and east. On northern parts of the craton, transgressive shoreface sandstone (Woodiana Member) is overlain by 'deeper' shelf argillite, whereas in the south, submarine sedimentation and volcanism prevailed throughout deposition of that part of Sequence 4 included in the Fortescue Group — the Jeerinah Formation. This formation is always less than a kilometre thick, and tends to have a higher volcanic content close to the margins of the Pilbara Craton.

These four numbered depositional sequences reflect the evolving tectonic setting of the basin. Initial continental sedimentation and volcanism (Sequences 1 and 2) took place in isolated, faulted sub-basins, and were accompanied by intermittent regional-scale uplift in the north and central Pilbara. The separate sub-basins coalesced during the time of Sequence 3 when regional subsidence, accentuated by further normal faulting and/or tilting in the south, resulted in a change to coastal and deeper shelf volcanism and sedimentation. Further subsidence at the begining of Sequence 4 times resulted in a major marine transgression and the establishment of 'deeper' marine shelf sedimentation and volcanism over the entire Hamersley Basin. The major structural controls on Fortescue Group sedimentation and volcanism in the southern Pilbara are discussed and comparisons made with Phanerozoic rift settings.

The Fortescue Group has been described as a continental flood basalt province (CFBP), but the application of such a label is unhelpful unless it can be demonstrated that the name is a valid one, in the sense that it denotes a unique class of volcanic province whose genesis is well understood. This is not currently the case for CFBPs, which tend to be seen as a subclass of large igneous provinces (LIPs) with diverse origins, sizes, compositions, and tectonic settings. Although the Fortescue Group certainly has the status of a LIP it differs from 'typical' CFBPs, such as the Deccan and Columbia River in at least three important respects: it is not solely 'continental', its emplacement lasted for a much longer time, and the main magma component was basaltic andesite rather than basalt.

Chapter 1

Introduction

Definition and outcrop

The name Fortescue Group was first used by MacLeod et al. (1963a, reprinted identically as 1963b) for a thick sequence of mafic lava flows and associated clastic and volcaniclastic sedimentary rocks which MacLeod and his co-authors had systematically mapped in the southern Hamersley Range area of the Pilbara region during the preceding (1962) field season. They noted that the succession was a component of the 'Proterozoic' rocks of the Pilbara region which had long been referred to as the 'Nullagine' Beds or Series (e.g. Maitland, 1909).

Although the internal stratigraphic divisions originally defined by MacLeod et al. (1963a) have been substantially modified and extended, the Fortescue Group essentially retains its identity. As established by these authors, it is the lowest of three groups which were successively laid down in a major Precambrian basin — the Hamersley Basin — covering much of what is now the Pilbara region (Fig. 1.1).

The principal outcrop area of the Fortescue Group (Fig. 1.2) forms an irregular, elongate, east–west strip lying mainly to the north of the Fortescue River, with apophyses extending farther northwards at each end: in the west into the Dampier Archipelago, and in the east along the western side of the Oakover River catchment. To the north, a number of outliers rest on the north Pilbara granite–greenstone terrain, including the large and important outlier west of Marble Bar. In addition to these, four smaller areas of outcrop are physically separated from the main strip by the overlying Hamersley Group. These are: the Gregory Range area, which has special characteristics and is dealt with under a separate heading below; the area adjacent to the northern and western edges of the Sylvania Inlier; an inlier in the core of the Wyloo Dome, in the southwest; and a major irregular inlier in the southwestern part of the Hamersley Ranges.

From Figure 1.2 it can be seen that the greatest longitudinal dimension of Fortescue Group outcrop is about 550 km, from its exposures on the lower Fortescue River in the west to the Gregory Range area in the east. Its greatest latitudinal extent is some 400 km, from the Dampier Archipelago in the north to the exposures on the southern side of the Sylvania Inlier in the south. The total area of Fortescue Group outcrop is about 40 000 km^2.

Regional geological setting

The following brief account of the geology of the area covered in Figure 1.2 is based on a summary of the geology of Western Australia (GSWA, 1990), which contains systematic descriptions of each of the tectonic units, as well as extensive references. The principal, and oldest, tectonic unit of the area is the Pilbara Craton, whose extent is shown on Figure 1.1, together with the positions of adjacent tectonic units. The Pilbara Craton has two components: an older granite–greenstone terrane, and the overlying Mount Bruce Supergroup, which includes the Fortescue Group.

The main outcrop area of the granite–greenstone terrane of the Pilbara Craton lies to the north of the basal unconformity of the Fortescue Group, from which it extends northwards to the coast. The quality of exposure decreases northwards and the older rocks are mainly concealed by an overlying Phanerozoic veneer immediately inland from the flat coastline. In the south Pilbara, granite–greenstone rocks are also exposed in a number of inliers within the outcrop of the Mount Bruce Supergroup. These lie both within the Hamersley Range area, where gentle folding has exposed granite–greenstone 'basement' in the Rocklea Dome and Milli Milli Dome, and farther east in the major Sylvania Inlier and a cluster of smaller inliers north-northeast of it, the largest of which is the Cooninia Inlier.

The geology of the main outcrop of the granite–greenstone terrane was comprehensively described by Hickman (1983), who recognized a stratigraphic sequence with a thickness up to 30 km. The lower part of the sequence (Warrawoona Group) consists principally of mafic, ultramafic, and felsic volcanic rocks with a subordinate sedimentary component, and the upper part (mainly the Gorge Creek Group) is made up mainly of mixed volcanic and epiclastic sedimentary rocks. The entire sequence has undergone low- to medium-grade metamorphism, and is folded into rectilinear to sinusoidal synformal greenstone belts that enclose large complex granitoid batholiths with shapes ranging from nearly circular (Mount Edgar Batholith), through ovoid (Corunna Downs Batholith), to irregular (Yule Batholith). These granitoid rocks have discordant, intrusive margins against the lower units of the greenstone sequence, but isotopic ages show that both greenstones and granitoid rocks developed mainly between c. 3.5 and c. 2.9 Ga, so that

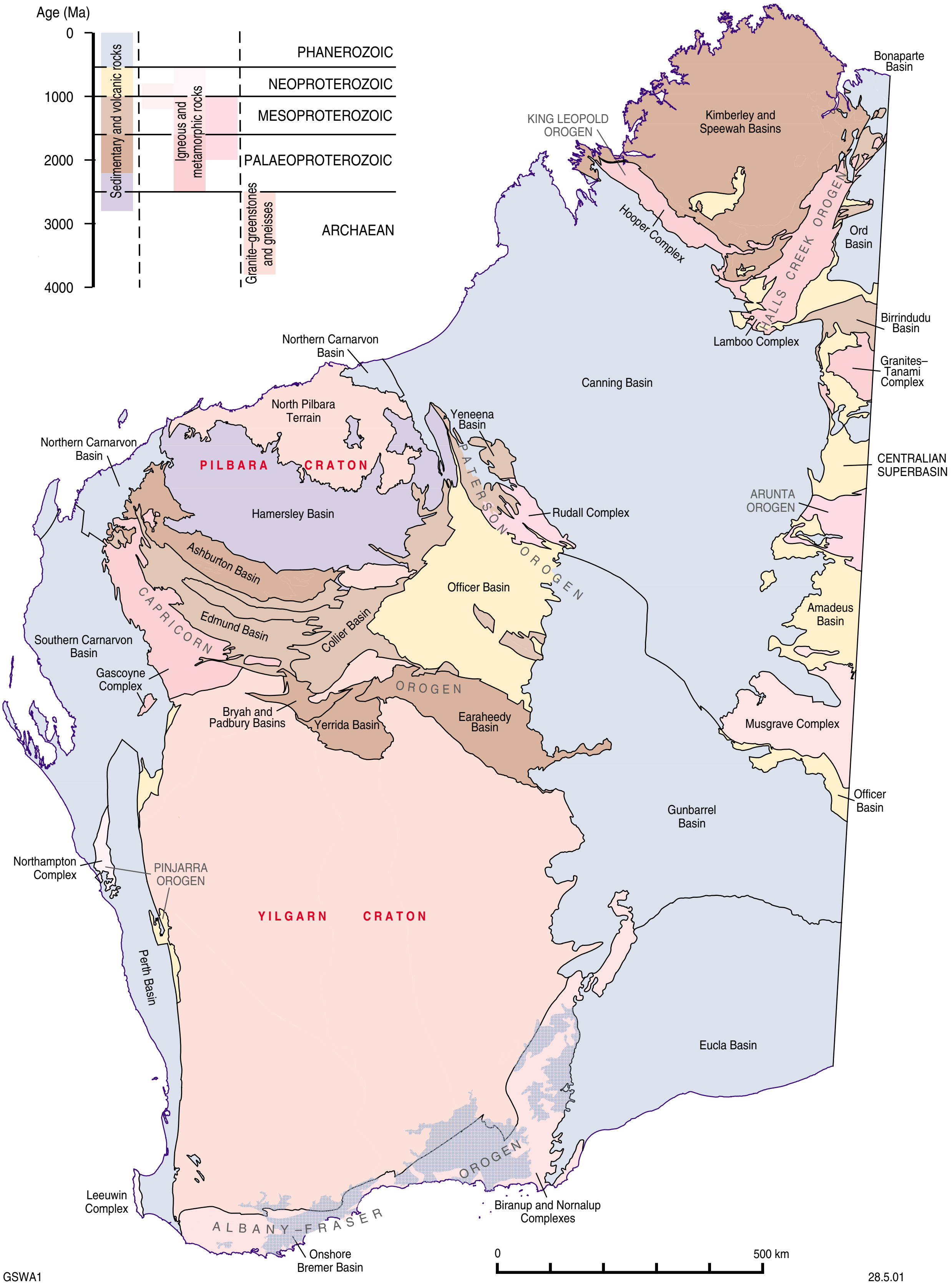

Figure 1.1. Main tectonic units of Western Australia

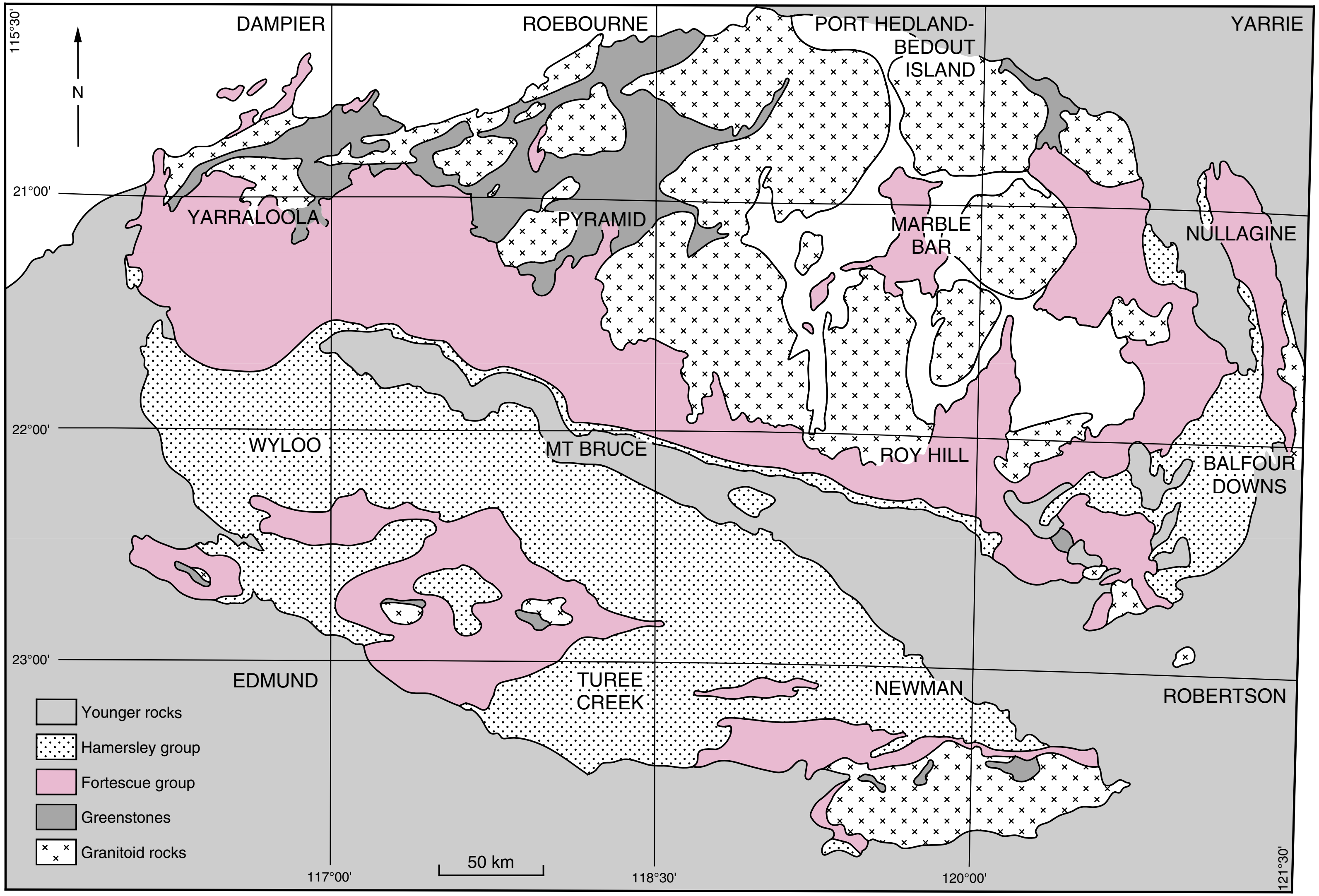

Figure 1.2. Geological map of the Pilbara showing the Fortescue Group outcrop and location of 1:250 000 map sheets mentioned in the text. Dampier and Barrow Island (Kriewaldt, 1964a,b); Roebourne (Ryan, 1966); Port Hedland – Bedout Island; (Low, 1965; Hickman and Gibson, 1982); Yarrie (Wells, 1959); Yarraloola (Williams, 1968); Pyramid (Kriewaldt and Ryan, 1967); Marble Bar (Noldart and Wyatt, 1962; Hickman and Lipple, 1978); Nullagine (Noldart and Wyatt, 1962; Hickman, 1978); Wyloo (Daniels, 1970; Seymour et al., 1988); Mount Bruce (de la Hunty, 1965; Thorne and Tyler, 1997a); Roy Hill (MacLeod and de la Hunty, 1966; Thorne and Tyler, 1997b); Balfour Downs (de la Hunty, 1964; Williams, 1989); Turee Creek (Daniels, 1968; Thorne et al., 1991); Newman (Daniels and MacLeod, 1965; Tyler et al., 1991); Robertson (de la Hunty, 1969; Williams and Tyler, 1989)

their deposition and evolution were contemporaneous, and intimately interlinked.

The younger of the two components of the Pilbara Craton is the Mount Bruce Supergroup of the Hamersley Basin. This was laid down unconformably over an eroded surface of the granite–greenstone terrane. The extent of the Fortescue Group, the lowest group within the Mount Bruce Supergroup, has already been described. This group is succeeded conformably by the Hamersley Group. The Hamersley Group is about 2.5 km thick, and is characterized by abundant banded iron-formations (BIFs), which have acted as the host rocks for large bodies of high-grade iron ore. The BIF units are separated by thinner 'shales', which are also iron-rich; many have a volcanogenic component. The Wittenoom Formation, in the lower part, consists largely of dolomitic rocks. A thick rhyolitic unit in the upper part of the Hamersley Group, the Woongarra Volcanics of MacLeod (1966), was originally thought to represent an extrusive volcanic sequence, but has been reinterpreted as intrusive, and redefined as the Woongarra Rhyolite (Trendall, 1995). The uppermost formation of the Hamersley Group, the Boolgeeda Iron Formation, is conformably overlain by the lowest (Kungarra) formation of the Turee Creek Group. This group consists largely of epiclastic sedimentary rocks, and its outcrop is confined to a few separate localities within the southern outcrop area of the Mount Bruce Supergroup. The age of the base of the Mount Bruce Supergroup is well established as c. 2.77 Ga (Arndt et al., 1991) and that of the the Woongarra Rhyolite is c. 2.43 Ga (Pidgeon and Horwitz, 1991), so that deposition in the Hamersley Basin spanned some 340 m.y. Although there are local stratigraphic discordances within the Mount Bruce Supergroup which represent significant periods of non-deposition, the stratigraphic integrity of the entire unit is emphasized by the fact that the main regional structures, and the pervasive regional burial metamorphism, appear to have affected all of its components equally and penecontemporaneously.

The Mount Bruce Supergroup is locally overlain unconformably by basal epiclastic sedimentary rocks of the younger Wyloo Group of the Ashburton Basin, but most boundaries of the Mount Bruce Supergroup against this group in the western and southern parts of its outcrop areas are faulted. These tectonic boundaries are associated with the Capricorn Orogeny, during which the thick, mainly turbiditic, sedimentary rocks that form the major part of the Wyloo Group, were tightly folded and cleaved. The fold axial traces wrap around the southern and western margins of the Mount Bruce Supergroup. The Wyloo Group (Thorne and Seymour, 1991) was probably laid down between c. 2.0 Ga and c. 1.8 Ga, whereas the Capricorn Orogeny reached its zenith at about 1.8 Ga. The area of Figure 1.1 also includes the c. 1.78 Ga Boolaloo Granodiorite (Krapez and McNaughton, 1999) whose emplacement was associated with the orogeny.

Farther south, the folded rocks of the Ashburton Basin are unconformably overlain by relatively gently dipping Precambrian rocks of the Bangemall Group. The Mount Bruce Supergroup is also overlain by various relatively undeformed late Precambrian sedimentary sequences along the eastern and southern limits of its outcrop area. These include the Manganese Subgroup of the Bangemall Group, the Yeneena Group, and the Bresnahan Group.

Undeformed Phanerozoic sedimentary rocks of the Phanerozoic Carnarvon Basin and Canning Basin unconformably overlie the Pilbara Craton and younger Precambrian tectonic units on the west and east respectively, and delimit the outcrop of Precambrian rocks.

Physiography, climate, and vegetation of the Pilbara region

The outcrop of the Fortescue Group lies entirely within the Pilbara region, a name loosely applied to the northwestern area of Western Australia extending eastward from the Indian Ocean coastline to the Great Sandy Desert, between latitudes 20° and 24° south.

The Pilbara region may be divided from north to south into five main physiographic units:

1) the coastal plain and remnant ranges of a dissected plateau, which is traversed by a number of major rivers flowing generally from south to north;
2) the north-facing scarp of the Chichester Range, which runs roughly east–west and defines the southern edge of the coastal plain;
3) the flat valley of the Fortescue River, which forms a plateau capping the Chichester Range;
4) the main north-facing scarp of the Hamersley Range, which defines the southern edge of the Fortescue River valley; and
5) the high, dissected country of the Hamersley and Ophthalmia Ranges, to the south of the main north-facing scarp of the Hamersley Range.

These divisions become blurred in the east, as the Chichester, Hamersley, and Ophthalmia Ranges become ill-defined, and the broken country of the Gregory Range merges eastward into the seif dunes of the Great Sandy Desert.

The distribution and structural disposition of the Fortescue Group has played an important role in the evolution and establishment of these broad divisions. Thus, the Chichester Range is formed by the gently south-dipping rocks of the group, whose stratigraphic thickness effectively controls the scarp for most of its length; prominent dissected range areas rising above the coastal plain are due to Fortescue Group outliers. These include the outliers west of Marble Bar, and east of Whim Creek. In a similar way, the major apophyses curving northward from each end of the main east–west outcrop are both associated with hilly country breaking the continuity of the coastal plain. To the south of the main Hamersley Range, the Fortescue Group inliers within the outcrop area of the Hamersley Group also form conspicuous dissected strike ridges.

Most constituent formations of the Fortescue Group generally form rough terrain of high relief throughout their outcrop area. This is especially so for the basaltic rocks, which characteristically form steep, irregular, boulder-strewn hillslopes. Exceptionally, the shale component of the Jeerinah Formation locally provides smooth intervening flat-bottomed valleys. Access to the Fortescue Group away from the main road system is therefore difficult, since off-track travel by 4WD vehicle, where possible, is always slow.

The Pilbara region has, in effect, two six-month climatic seasons: a hot summer lasting from November to March, and a pleasantly warm winter from April to October. Temperatures vary according to latitude, distance from the coast, and topographic height, but average daily maxima and minima* range from about 40° and 25°C in the hotter months to about 26° and 12°C in the cooler months.

The average annual rainfall of the Pilbara region is about 300–350 mm, but decreases to below 300 mm in the east and south. The rainfall is erratically distributed: 80% of the annual average falls during the hot summer months, and the remainder during the winter. During the summer months the rainfall is largely associated with the immediate passage of the intense cyclones which periodically follow a curving southerly course across the area; rainfall immediately connected with such storms commonly exceeds the annual average rainfall, so that the rainfall recorded at any one centre for a given year is mainly dependent on the number of such events (rarely more than two and often none) during that year. The erratic distribution of rainfall also means that there are no permanently flowing surface rivers in the region; the major rivers flood briefly, and spectacularly, immediately after rainfall, and then quickly dry again.

Average annual evaporation of the Pilbara region is about 3600 mm, over ten times the annual rainfall. Coupled with the high temperatures experienced, this means that, away from the few major rivers with permanent surface pools or shallow groundwater, vegetation is everywhere relatively sparse, and consists of an almost ubiquitous covering of spinifex grass. This is relieved by small scattered trees, mainly *Eucalyptus* spp and *Acacia* spp. Larger tree growth, including *Eucalyptus* and *Melaleuca*, are confined to the immediate vicinity of drainage lines.

Social and economic development

The Pilbara was first occupied by the ancestors of the present Aboriginal population, who migrated southwards to the Australian continent via land bridges in the Indonesian Archipelago during the last glacial maximum. By the time of European settlement the small Aboriginal population of related groups of semi-nomadic hunter–gatherers, lived in equilibrium with the natural productivity of the region.

Although there were early European landings on the Pilbara coast, notably by Dampier in 1688 and 1699, and by Darwin in the Beagle in 1836, no permanent European settlement took place until after the extensive exploration journey of F. T. Gregory in 1861. As a result of Gregory's enthusiastic advocacy of the suitability of the Pilbara region for grazing, much of the region was divided into large pastoral leases. The consequent erection of fencing, failure to recognize prior Aboriginal occupation of the land, and the obvious material advantages of the European lifestyle, were all factors which led to a rapid change of Aboriginal social economy, from hunting and gathering to a loose association with the pastoralists, in which Aboriginal communities became increasingly settled on stations, working in various capacities for pastoral leaseholders.

Pastoral settlement was accompanied by prospecting. The discovery of gold near Roebourne in 1877 (Maitland, 1909), and richer discoveries such as those at Pilbara and Mallina in 1888, led to an early gold rush, which declined rapidly after the 1892 and 1893 discoveries at Coolgardie and Kalgoorlie. Although the mining and pastoral industries continued to develop during the first half of the present century, neither fulfilled the hopes of the early pioneers. A major industrial dispute between pastoralists and their Aboriginal workforce in 1950 reinforced social divisions, and the Aboriginal population moved to designated Reserves and onto pastoral leases transferred to their occupation with Government funding. By 1960, prices for most minerals, as well as those for wool and beef, were not conducive for either industry to expand. The Pilbara region had no railways, and no sealed roads outside the scattered towns, and other public services and facilities were at correspondingly elementary levels.

This situation began to alter very rapidly in late 1960, when the Australian Government lifted an embargo on the export of iron ore, which had been imposed in 1938 (Blockley et al., 1990); the export embargo also blocked exploration. Massive new discoveries of high-grade hematitic ore were the immediate result of the recommencement of iron ore exploration. Negotiation of export contracts, mainly with Japanese mills, soon followed, and large mines were established, served by railways connecting them with coastal port facilities, and by the construction of new towns to house the resultant workforce (Lord and Trendall, 1976). Dampier, Karratha, Tom Price, Paraburdoo, Pannawonica, Wickham, Newman and South Hedland, with a total population of about 30 000, were all established after 1960. The subsequent construction of natural-gas processing and shipping facilities on the Burrup Peninsula to serve the offshore gasfields has reinforced development of facilities in the area. Sealed roads now connect all the main towns, and there are frequent air services from Perth to and from Paraburdoo, Karratha, Port Hedland, and Newman.

Summary of previous investigations

In reviewing published work on the geology of the Fortescue Group, studies can be divided into those which preceded, and those which followed, its formal definition by MacLeod et al. (1963a,b). The latter uses the current name, whereas the former generally refers to rocks of the Fortescue Group as components of the 'Nullagine' Beds or Series.

* Climate data from Commonwealth Bureau of Meteorology website, 2000

The explorer F. T. Gregory (Gregory and Gregory, 1884) included comments on some Fortescue Group rocks in his account of the geological results of his 1861 journeys, and Woodward (1890, 1891) later provided a general account of the geology of the Pilbara region. However, the first systematic field descriptions of Fortescue Group rocks, especially of the eastern part of the main outcrop area, were given by Maitland (1904, 1905, 1906, 1908, 1909). Talbot (1919, 1920) augmented the field observations of Maitland. For many years, following the work of Maitland and Talbot, the only new geological information published on the Fortescue Group consisted of local investigations of specific mining localities during the course of the Aerial Geological and Geophysical Survey of Northern Australia (AGGSNA) in the 1930s (Finucane, 1935, 1936, 1938a,b, 1939).

The foundations of present knowledge of the Fortescue Group were laid by a program of systematic regional geological mapping of the Pilbara region at a scale of 1:250 000 carried out by GSWA. Although this was initiated in 1956 (Noldart and Wyatt, 1958; Noldart, 1960a,b), its momentum increased in 1961 due to strong interest in the regional geology of the Hamersley Group caused by the lifting of the iron ore export embargo. Fortescue Group rocks occur on fifteen contiguous 1:250 000 map sheets covering the Pilbara region, nine of which were subsequently remapped between 1972 and 1993; these are all accompanied by Explanatory Notes, in many of which stratigraphic names are established. Hickman (1983) included a review of the Fortescue Group in the northern part of the Pilbara Craton based on the investigations prior to 1978.

Because GSWA mapping started separately in a number of non-contiguous sheet areas (Trendall, 1975b, p. 121) there were considerable local differences in stratigraphic nomenclature. Trendall (1975b, 1990), tried to limit these problems by grouping the formally established units into six informal divisions. Subsequent remapping of the northeastern Pilbara during the late 1970s enabled Hickman (1983, plate 1) to apply a single formal stratigraphic subdivision of the Fortescue Group across the northern Pilbara. Thorne and Blake (1990) later modified this scheme and applied it to northern and southwestern parts of the Pilbara.

Blake (1984a) was the first to recognize the role of intra-Fortescue Group unconformities and separate depositional basins in the northern Pilbara Craton. He used the major depositional breaks as a basis for a four-fold numerical stratigraphy that was applied to most of the northern part of the craton. This approach was later expanded into a complete, sequence-based stratigraphy for the Mount Bruce Supergroup (Blake and Barley, 1992; Blake, 1993).

As already noted, the name Fortescue Group was established near the start of this mapping program, the early results of which were synthesized by MacLeod (1966). Supplementary contributions to Fortescue Group geology published by GSWA in association with the program include that of Kriewaldt (1964b) on the Fortescue Group of the Roebourne area, and those of Trendall (1963, 1965), which include petrographic descriptions of Fortescue Group rocks.

As a result of the upsurge in mineral exploration activity that flowed initially from iron ore development, the Commonwealth Scientific and Industrial Research Organisation (CSIRO) established a laboratory of its Applied Mineralogy Division (now Division of Exploration Geoscience) in Perth, and staff of both the Perth and Sydney branches contributed to specialized aspects of Fortescue Group geology. These included the work of R. E. Smith and others (Smith, 1975, 1979; Smith and Smith, 1976; Smith et al., 1982), R. C. Horwitz (1976, 1978, 1987; Horwitz and Smith,1978), and B. J. Embleton and others (Embleton, 1978; Schmidt and Embleton, 1985). Smith's contributions were geochemical and mineralogical, and included the establishment of a regional scheme of metamorphic zones for mafic rocks of the Fortescue Group and a local study of heterogeneous alteration of the Maddina Formation and its relationship to post-consolidation fluid movements. Horwitz studied stratigraphic thickness variations within the Fortescue Group and, with Smith, related these to the regional tectonic control of its deposition. Schmidt and Embleton (1985) carried out substantial palaeomagnetic studies on both mafic volcanic rocks of the Fortescue Group and possible feeder dykes.

Prior to the 1970s, virtually no analyses of Fortescue Group rocks other than assays directed towards mineral exploration had been carried out, but at the time of writing between 700 and 800 whole-rock analyses exist, including major elements and a range of trace elements. In 1975, a geochemical study of the volcanic rocks of the Pilbara Craton was initiated jointly by the Commonwealth Bureau of Mineral Resources (BMR) and GSWA; although the project had no formal title, it is convenient to refer to it here as the Pilbara Volcanic Study. The main, and initially the only, concern of the project was the geochemistry of the mafic and ultramafic volcanic rocks of the granite–greenstone terrane of the Pilbara Craton, but sample collection was extended stratigraphically upwards between 1981 and 1983 to include the Fortescue Group. All the analytical data (964 analyses) resulting from the Pilbara Volcanic Study have been published (Glikson et al., 1986b). A full evaluation of the results has yet to be made by the participants, although contributions relevant to the Fortescue Group have appeared (Glikson et al., 1986a). Work by Cowley (1979) also made a significant contribution to chemical information available, as did later studies by Sieber (1989, pers. comm.), and by Nelson, D. R., et al. (1992) in association with a geochronological project, described further below. Meakins (1990) gave a brief appraisal of a body of 366 chemical analyses of Fortescue Group rocks made for CRA Exploration Pty Ltd.

Additional previous work on specialized aspects of Fortescue Group geology is also mentioned in later chapters.

Scope and objectives of this Bulletin

The primary objective of this Bulletin is to provide a systematic descriptive geology of the Fortescue Group, integrating the results of our own original work with the published (and some unpublished) results of others so as to provide an overview of current knowledge. We hope that non-specialist readers approaching the Fortescue Group for the first time will find all that they need within its covers, and that those who seek more detail on specific topics will be able to find out whether the information is available, and if so where. Although a descriptive geology has been our primary objective we also provide, in the final chapter, an interpretative summary of the development of the Hamersley Basin during Fortescue Group time, as well as a comparative assessment of the status of the group as an ancient continental flood basalt province.

The extent of the Fortescue Group is such that we could not address every aspect of its geology in detail. We therefore elected to concentrate on those aspects relevant to an understanding of its depositional environment, such as sedimentology and volcanology, as distinct from topics largely relevant to its post-depositional history, such as structure and metamorphism, existing knowledge of which is only briefly summarized below.

Structure

In the central and southwestern part of the Fortescue Group outcrop there is a general increase in the intensity of folding from north to south MacLeod (1966, p. 63). In the Chichester Range the dip is generally less than 5° southwards. On the southern side of the main Hamersley Range Synclinorium the northward dips are of the same order. Folding gradually increases in intensity to the south, where fold limbs commonly dip over 45°. The intensity reaches a maximum close to the southernmost limit of Fortescue Group outcrop.

This north-to-south increase in deformational intensity noted by MacLeod (1966) in the Hamersley Range area also applies in a general way over the total Fortescue Group outcrop: thus the central part, in the central Chichester Range, can be seen as a virtually undeformed core away from which the intensity of deformation increases in all directions. As the strongly deformed southern boundary, already noted, is followed westwards, this zone of more intense deformation curves northwards along the western extremity of the Fortescue Group, and this swing continues as the boundary is followed farther north, so that there is significant faulting and folding of the Fortescue Group along its western margin. Similarly, at its eastern extremity in the Gregory Range, there is a marked increase in deformation. In the southeast segment of the outcrop area, the Fortescue Group is overlapped by gently dipping younger Precambrian sedimentary rocks (Manganese Subgroup of the Bangemall Group) so that it is not known whether there is an equivalent increase in deformation in that direction. In the north, where the Fortescue Group occurs only in outliers resting on the older granite–greenstone terrane, no northward increase in deformation is apparent, except for localized structural complexity in the southern part of the inlier west of Marble Bar.

Trendall (1983) used this perceived radial outward increase in structural complexity of the Mount Bruce Supergroup to support his argument for the existence of a ‘Pilbara egg’, an imaginary ellipse of uncertain significance bounding the outcrop of the supergroup. Miyano (1976) had previously suggested that the craton showed concentric elliptical zones with different structural characteristics. Whatever the significance, if any, of the ‘Pilbara egg’ may be, it is true that the peripheral zones of the Fortescue Group are zones of structural complexity (Plates 1a,1b). More recent interpretations of the tectonic history of the craton and its adjacent units are given below and are also summarized in GSWA (1990). These accounts relate the strong deformation in the south to tectonism associated with the Palaeoproterozoic Capricorn Orogeny, and the intense deformation on the east to Palaeoproterozoic to Neoproterozoic tectonism in the Paterson Orogen.

Capricorn Orogen structures

The southern Pilbara forms the northern margin of the Capricorn Orogen, a major zone of deformed, low- to high-grade metamorphic rocks and granitoid intrusions formed during continental crustal collision between the Pilbara and Yilgarn Cratons about 2200–1600 Ma (Gee, 1979; Myers, 1990). Detailed accounts of deformational events in the southern Pilbara are provided by Tyler and Thorne (1990), Tyler (1991), and Thorne and Seymour (1991). These workers separate the major structures into two fold belts: the Ophthalmia Fold Belt, and the younger Ashburton Fold Belt.

In the southwestern part of the Hamersley Basin the Ophthalmia Fold Belt is characterized by broad-scale, open folds having a mainly northwesterly trend. These correspond to the central structural zone of MacLeod et al. (1963a,b). Tyler (1991) identified a regional-scale foreland fold-and-thrust belt in the southeastern Hamersley Basin characterized by easterly trending close to tight folds with short wavelengths, corresponding to the southern structural zone of MacLeod et al. (1963a,b). Structural and stratigraphic relationships along the margin between the Ophthalmia and Ashburton Fold Belts suggest that these two groups of folds represent different events. Deformation to produce the open folds was attributed by Tyler and Thorne (1990) to dextral transpression along the southern Pilbara margin during the early stages of the Capricorn Orogeny of Gee (1979). Horwitz and Powell (1992) and Blake and Barley (1992), however, have suggested that this deformation was related to the development of the McGrath Trough (Horwitz, 1982), which was initiated either as a foreland basin or a backarc compressive cratonic basin during a collision between the Pilbara Craton and an unknown southern continent sometime after c. 2440 Ma but before 1840 Ma.

Tyler and Thorne (1990) interpreted the foreland fold-and-thrust belt in the southeastern Hamersley Basin as the result of an oblique collision between the Pilbara and Yilgarn Cratons at c. 1840 Ma, which commenced in the east and migrated westwards. Uplift of the Sylvania Inlier supplied granitic sediment to the Ashburton Formation in the Ashburton Basin. Initial deformation of the Ashburton Basin to form the Ashburton Fold Belt was attributed to thrusting. Associated uplift provided sediment to the

Mount Minnie Group and the Capricorn Formation. Later deformation was related to a dextral wrench-fault system produced by the extrusion of material westwards from between the two approaching craton margins.

Paterson Orogen structures

The effects of orogenies of the Paterson Orogen in the Gregory Range are discussed by Williams and Trendall (1998a,b,c). Deformation of the Fortescue Group and exposure of the Gregory Granitic Complex is attributed to uplift along a series of steep reverse faults with a sinistral strike-slip component. These faults have successively thrusted deeper crustal components upwards and westwards onto the overlying volcanic edifice. The major transpressional regime represented by this faulting is the product of collisional tectonics (probably continent–continent) that took place to the east of the present Pilbara Craton. These events commenced in the Late Palaeoproterozoic (Yapungki Orogeny), and were followed by successive compressive reactivation of pre-existing fracture lines in the Late Mesoproterozoic – Early Neoproterozoic (Miles Orogeny) and Late Neoproterozoic (Paterson Orogeny). All of these events are recorded in the Paterson Orogen, which lies to the east and southeast of the Pilbara Craton (Myers, 1993; Hickman et al., 1994; Williams and Bagas, in prep.).

Metamorphism

Although MacLeod (1966) did not mention any metamorphism of Fortescue Group rocks in his initial description, and Trendall and Blockley (1970, p. 294) specifically asserted that the overlying ‘Hamersley Group has nowhere undergone regional metamorphism’, it is now clear that all rocks of the Fortescue have been modified by some degree of secondary alteration.

The first systematic study of Fortescue Group metamorphism was that of Smith et al. (1982), who recognized four zones, from lowest to highest grade:

- Zone 1 prehnite–pumpellyite zone
- Zone II prehnite–pumpellyite–epidote zone
- Zone III prehnite–pumpellyite–epidote–actinolite zone
- Zone IV actinolite zone.

The surface distribution of these four zones is shown on Figure 1.3. The work of Smith et al. (1982) involved thin-section observations at more than 1000 sample sites, concentrated in the eastern part of the main northern outcrop, and in the Hamersley Range area. Their work concentrated on the mafic lava units, and they carried out extensive microprobe analysis of epidotes and pumpellyites from these rocks, demonstrating within each mineral a broad range of compositions with superimposed trends related to their zones.

Smith et al. (1982) argued from the relationship of their sample set to folds, especially in the Hamersley Range area, that depth of burial was the dominant control of increases in metamorphic grade. They argued from published experimental work that a relatively high temperature gradient of 80 to 100°C/km was likely for the shallow part of the sequence, but that the deeper part of the sequence, below about 2.5 km, had a gradient of 40°C/km. They suggested that the prehnite–pumpellyite facies corresponded to a fluid pressure of 0.5 to 1 kilobar (kb) and a temperature range between 100 and 300°C, and that the prehnite-bearing pumpellyite–actinolite facies developed at about 1.5 kb over a temperature range of 300 to 360°C.

Tyler (1990a, 1991) provided a metamorphic map of the Sylvania Inlier and the adjacent part of the Hamersley Basin (Tyler, 1990a, fig. 2–163) and revised the Hamersley Basin geotherm of Smith et al. (1982). Tyler estimated that Mount Bruce Supergroup overburden pressures were between 240 and 350 MPa (2.4–3.5 kb).

The study of Smith et al. (1982) did not extend farther east than longitude 120°40'E, and excluded the Gregory Range, where the Fortescue Group shows heterogeneous but generally substantial metamorphism, which is discussed later.

Terminology

Although the metamorphism of the Fortescue Group is pervasive and locally significant, it rarely impedes lithostratigraphic mapping, or field observation of the primary rock types and textural features required for sedimentological and volcanological interpretation. In the following text the prefix ‘meta-’ is omitted from all rock names for ease of description.

Grain-size terms for clastic sedimentary rocks are based on the Wentworth scheme (Folk, 1974). An explanation of symbols used in the graphic logs is given in Figure 1.4.

The following terms are used to describe bed and cross-stratified set thickness:

Descriptive term	*Thickness range (m)*
Bedding	
Very thick	>1.5
Thick	0.5–1.5
Medium	0.2–0.5
Thin	<0.20
Cross-stratification	
Large-scale	>0.5
Medium-scale	0.05–0.5
Small-scale	<0.05

Authorship responsibility

A. F. Trendall is responsible for **Chapters 1, 10,** and **12**. A. M. Thorne wrote the other chapters, except for **14**, which is the work of both authors.

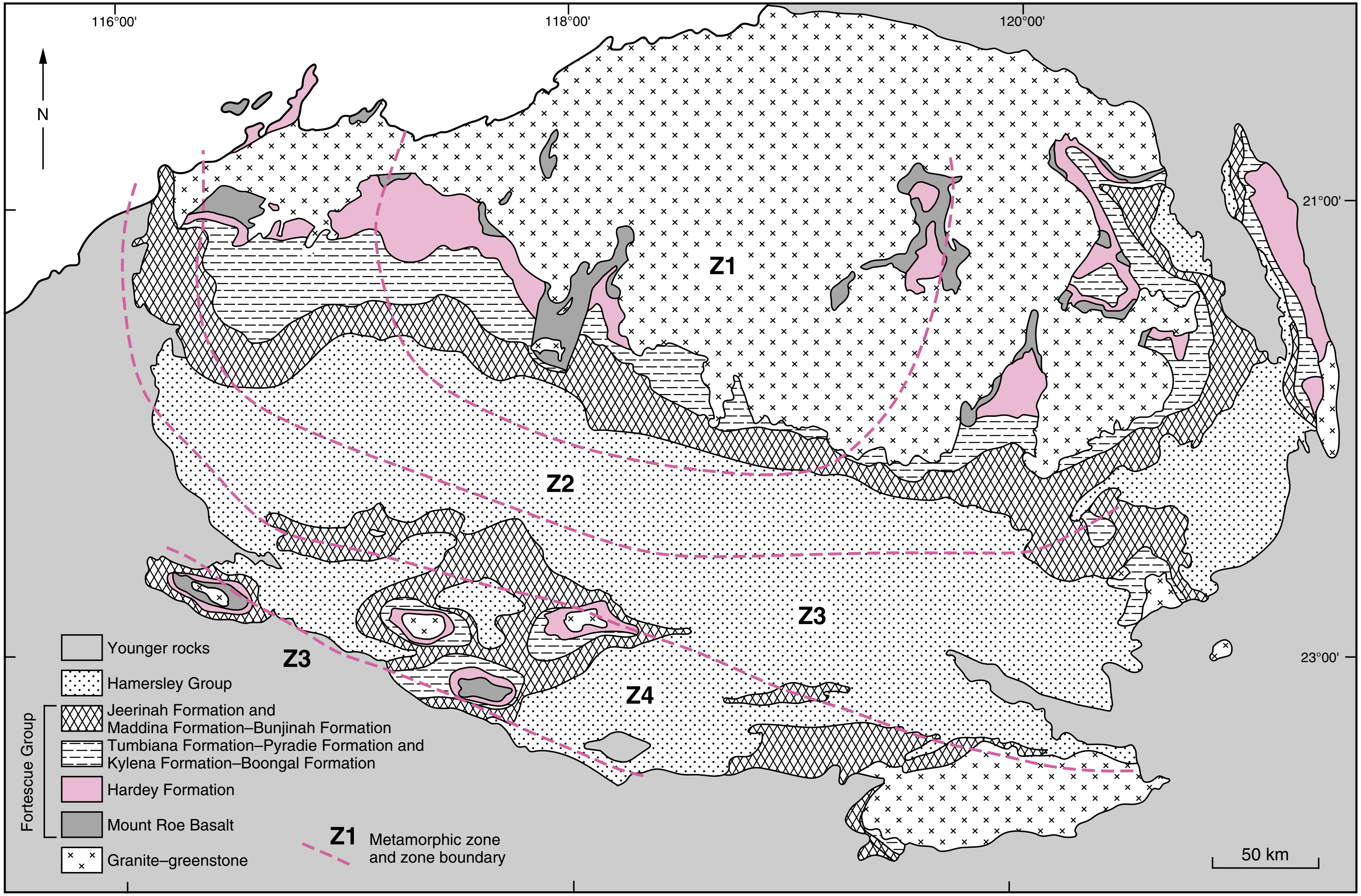

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Figure 1.3. Geological map of the Hamersley Basin showing the burial metamorphic zones of Smith et al. (1982). Zone 1 — prehnite–pumpellyite zone; Zone 2 — prehnite–pumpellyite–epidote zone; Zone 3 — prehnite–pumpellyite–epidote–actinolite zone; Zone 4 — actinolite zone

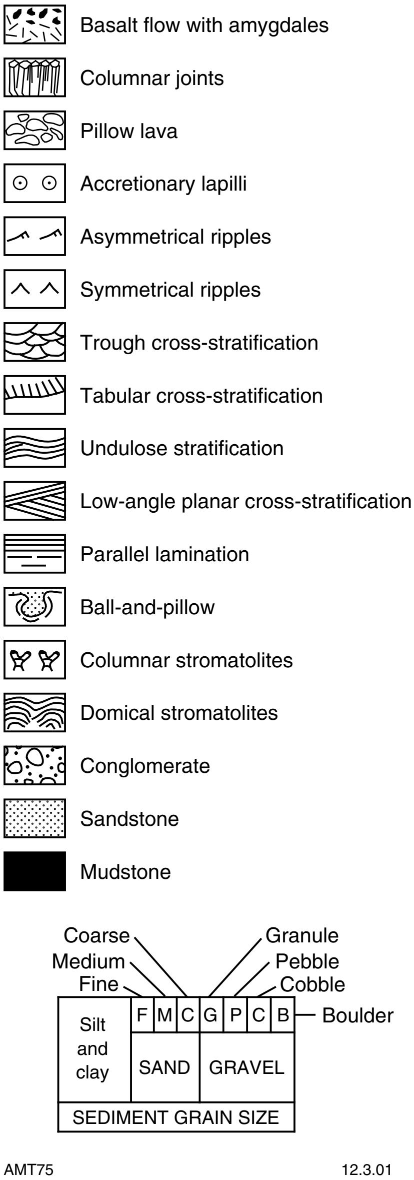

Figure 1.4. Explanation of graphic log symbols used in this Bulletin

Acknowledgements

We gratefully acknowledge the kindness of Timothy Blake, who shared his extensive knowledge of the Fortescue Group with us and also provided us with pre-publication copies of his 1:100 000-scale maps covering much of the Fortescue Group outcrop. We also found discussion with the staff of Rio Tinto Exploration Pty Ltd particularly helpful, and appreciate the generosity of that company in allowing us access to diamond drillcore.

Chapter 2

Stratigraphic framework and geochronology

Previous lithostratigraphy

Prior to the formal definition of the Fortescue Group by MacLeod et al. (1963a,b), most published descriptions followed the stratigraphic nomenclature of Maitland (1904, 1905, 1906, 1908, 1919), in which he referred to these rocks variously as part of the Nullagine Beds, Nullagine Series, or Nullagine Formation. Maitland (1919, p. 23) subdivided the Nullagine Formation into a lower Nullagine Volcanic Series, a middle Carawine Dolomite Series, and an upper, unnamed assemblage of sedimentary rocks. This subdivision formed the basis for description until Noldart and Wyatt (1962) published the first named stratigraphy for what is now the Fortescue Group in the northeastern Pilbara.

Noldart and Wyatt's (1962) work on MARBLE BAR and NULLAGINE formed part of a regional 1:250 000-scale mapping program carried out by GSWA between 1954 and 1965. The results of this study (MacLeod et al., 1963a,b; authors listed in Fig. 2.1) form the basis for most subsequent descriptions of the Fortescue Group. However, because mapping commenced independently in a number of non-contiguous sheet areas (Trendall, 1975b, p. 121), there are considerable local differences in stratigraphic subdivision of the Fortescue Group, with a total of 33 formal names documented during the 1960s. Studies carried out during the period 1970 to 1998 resulted in a further 17 formal and 30 informal stratigraphic names being introduced to the literature. A list of formal, and published informal, names that have been applied to the Fortescue Group since 1962 is given in Appendix 1; the list of formal names is also summarized in Figure 2.1.

Trendall (1975b), in a review of Fortescue Group stratigraphy, attempted to minimize nomenclature problems by limiting the use of formal stratigraphic names, and by grouping the formally established units into six informal divisions. Subsequent remapping of the MARBLE BAR, NULLAGINE, PORT HEDLAND – BEDOUT ISLAND, and YARRIE map sheets during the 1970s resulted in better correlation between the northeast Pilbara succession and that occurring farther west (Hickman and Lipple, 1978; Hickman, 1978; Hickman and Gibson, 1982; Hickman et al., 1983). This remapping enabled Hickman (1983, plate 1) to publish the first map in which a single formal stratigraphic subdivision of the Fortescue Group was applied to the entire northern Hamersley Basin. The scheme of Hickman (1983) is also given in Figure 2.1.

Sequence stratigraphy

The concept of the depositional sequence was introduced by Sloss et al. (1949), and later modified by Sloss (1963) and Krumbein and Sloss (1963). It was used to describe the geological history of part of North America in terms of a succession of stratigraphic packages that are internally varied but separated from each other by continent-wide unconformities. During the past 20 years, sequence concepts have been used, and modified, by the oil industry during the exploration of Phanerozoic passive-margin basins (e.g. Vail et al., 1977). Today, the most widely accepted definition of a sequence is that of Mitchum et al. (1977), who described it as 'a relatively conformable succession of genetically related strata bounded at its top and base by unconformities or their correlative conformities.'

Blake (1984a) was the first to recognize the importance of intra-Fortescue Group unconformities in the lower part of the succession in the northern Pilbara Craton. He used these hiatuses as a basis for a four-fold numerical classification in which the various lithological units are separated by angular unconformities or disconformities. This framework was later expanded into a complete, sequence-based stratigraphy for the Mount Bruce Supergroup (Blake and Barley, 1992).

Stratigraphy used in this Bulletin

Previous stratigraphic subdivisions of the Fortescue Group based entirely on lithostratigraphic mapping were hampered by discontinuous outcrop area, absence of marker horizons or biostratigraphic zonation, marked lateral and vertical facies variations, and limited geochronological control. On the other hand a stratigraphy founded on sequence analysis is essentially interpretative, particularly when based on outcrop information alone. For example, in cases such as the Fortescue Group, where several unconformities are commonly present, their correlation across areas of no exposure incorporates a significant element of subjectivity. Similarly, the judgement as to where to place the correlative conformity of an unconformity can be arbitrary unless supported by other data.

Because of the drawbacks associated with both lithostratigraphy and sequence stratigraphy, we have

NORTHERN HAMERSLEY BASIN (Hickman, 1983)	*ROEBOURNE (Ryan, 1966)*	*YARRALOOLA (Williams, 1968)*	*PYRAMID (Kriewaldt and Ryan, 1967)*	*MOUNT BRUCE (de la Hunty, 1965)*	*ROY HILL (MacLeod and de la Hunty, 1965)*	*BALFOUR DOWNS (de la Hunty, 1964)*	*NULLAGINE AND MARBLE BAR (Noldart and Wyatt, 1962)*
JEERINAH FORMATION		JEERINAH FM. Roy Hill Shale Mbr Nallanaring Volc. Mbr Warrie Member Woodiana Sst. Mbr	JEERINAH FM. Roy Hill Shale Mbr Warrie Member Woodiana Sst. Mbr	JEERINAH FM.	JEERINAH FM. Roy Hill Shale Mbr Warrie Member Woodiana Sst. Mbr	LEWIN SHALE (lower part)	mapped as part of CARAWINE DOLOMITE and TUMBIANA PISOLITE
MADDINA BASALT		MADDINA VOLCANICS	MADDINA BASALT	MOUNT JOPE VOLCANICS Bunjinah Pillow Lava Member	MOUNT JOPE VOLCANICS Maddina Basalt Member	LITTLE DE GREY LAVA (in part)	'UPPER' LITTLE DE GREY LAVA
KURUNA SILTSTONE NYMERINA BASALT TUMBIANA FORMATION		PILLINGINI TUFF	PILLINGINI TUFF	Pyradie Pyroclastic Member	Kuruna Siltstone Mbr Nymerina Basalt Mbr Tumbiana Pisolite Mbr	TUMBIANA PISOLITE	TUMBIANA PISOLITE
KYLENA BASALT		KYLENA VOLCANICS	KYLENA BASALT	Boongal Pillow Lava Member	Kylena Basalt Mbr	LITTLE DE GREY LAVA (in part)	'LOWER' LITTLE DE GREY LAVA and 'UPPER' COONGAN VOLCANICS
HARDEY SANDSTONE	Sandstone south of Mount Ada	CLIFF SPRINGS FORMATION Lyre Creek Agglomerate Mbr	CLIFF SPRINGS FORMATION Lyre Creek Agglomerate Mbr	HARDEY SANDSTONE		BEATONS CREEK CONGLOMERATE	GREEN HOLE CONGL. 'LWR' COONGAN VOLC. GLEN HERRING SHALE BEATONS CREEK CONGLOMERATE
MOUNT ROE BASALT	MOUNT ROE BASALT	MOUNT ROE BASALT	MOUNT ROE BASALT				

AMT21 NOTE: Mbr=Member; Sst=Sandstone; Congl.=Conglomerate; Volc.=Volcanics 12.3.01

Figure 2.1. A summary of previous formal stratigraphic subdivisions of the Fortescue Group from 1:250 000-scale geological mapping and Hickman (1983)

adopted in this Bulletin an integrated approach to subdivide the Fortescue Group. We use a broad system of lithostratigraphy as the basis for description, but at the same time recognize the importance of regional unconformities and geochronological, sedimentological, geochemical and volcanological data as additional stratigraphic tools.

The stratigraphic framework is similar to that of Thorne and Blake (1990) and subdivides the Fortescue Group of the northern Hamersley Basin into six formations; these are directly correlated with the six major formations of the southern Hamersley Basin (Fig. 2.2). The lowermost unit in the southern Pilbara, the Bellary Formation (Thorne et al., 1991), is confined to the northern part of TUREE CREEK and has no formally named equivalent in the northern Hamersley Basin. In the northern Pilbara, eight members are identified within the Fortescue Group; their names and stratigraphic position are also given in Figure 2.2.

The formations shown in Figure 2.2 can be grouped into four major tectono-stratigraphic sequences based on the nature of the intra-Fortescue Group bounding surfaces, volcanic and sedimentary facies character, and geochronological and geochemical data. Sequence 1 incorporates the Mount Roe Basalt and pre-Mount Roe Basalt sedimentary units; Sequence 2 comprises the Hardey Formation; Sequence 3 consists of the Kylena Formation, Tumbiana Formation, and Maddina Formation, and their equivalents; and Sequence 4 is the Jeerinah Formation. This four-fold subdivision forms the basis for interpreting the tectonic setting of the Fortescue Group (**Chapter 14**).

Sequences 1, 2, and 3 correspond to the Mount Roe Sequence, Hardey Sequence Package, and Mount Jope Supersequence respectively of Blake (1993), whereas Sequence 4 is equivalent to the lower part of his Marra Mamba Supersequence Package. Notwithstanding this correspondence we prefer not to use Blake's (1993) sequence stratigraphic units because our subdivision is not entirely unconformity-based. For example, in the case of the boundary between units 3 and 4 there is no documented proof that this is an unconformity. No angular discordance has been recorded, and there appears to have been little, or no, erosion between unit 3 and unit 4. In addition, using current geochronological data, it is not known if this surface represents a significant time break. At present, the main indication that the base of the Jeerinah Formation represents a major stratigraphic boundary lies with the fact that, in the north Pilbara, this level marks the change from coastal and subaerial volcanism to deeper shelf sedimentation. In the south Pilbara, the entire middle to upper Fortescue Group was dominated by subaqueous mafic volcanism. Here the boundary between units 3 and 4 appears to be marked by an increased amount of argillite.

Status of the Fortescue Group

The International Stratigraphic Guide (Salvador, 1994) recommends that 'the union of adjacent strata separated

NORTHERN HAMERSLEY BASIN			***SOUTHERN HAMERSLEY BASIN***	***TECTONO-STRATIGRAPHIC SEQUENCES***
MARBLE BAR SUB-BASIN	*NORTHEAST PILBARA SUB-BASIN (Excluding Gregory Range)*	*NORTHWEST PILBARA SUB-BASIN*	*SOUTH PILBARA SUB-BASIN*	
	JEERINAH FORMATION Roy Hill Member Warrie Member Woodiana Member	JEERINAH FORMATION Roy Hill Member Nallanaring Member Warrie Member Woodiana Member	JEERINAH FORMATION	SEQUENCE 4
	MADDINA FORMATION Kuruna Member	MADDINA FORMATION	BUNJINAH FORMATION	SEQUENCE 3
	TUMBIANA FORMATION Meentheena Member Mingah Member	TUMBIANA FORMATION	PYRADIE FORMATION	
	KYLENA FORMATION	KYLENA FORMATION Gidley Granophyre	BOONGAL FORMATION	
HARDEY FORMATION	HARDEY FORMATION Bamboo Creek Member	HARDEY FORMATION Cooya Pooya Dolerite Lyre Creek Member	HARDEY FORMATION	SEQUENCE 2
MOUNT ROE BASALT	MOUNT ROE BASALT	MOUNT ROE BASALT	MOUNT ROE BASALT	SEQUENCE 1
GRANITE–GREENSTONE BASEMENT			BELLARY FORMATION ?	

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Unconformity

Figure 2.2. Stratigraphic subdivision of the Fortescue Group used in this Bulletin. The locations of the various sub-basins are shown on Figure 2.3

by regional unconformities or major hiatuses into a single lithostratigraphic unit should preferably be avoided even if no more than minor lithologic differences can be found to justify the separation'. Although we accept the presence of unconformities or disconformities within the Fortescue Group, particularly on the northern Pilbara Craton (Blake, 1984a, 1993), we choose to retain the term 'Group' when referring to these rocks. The reasons for this are twofold. Firstly, because a large part of the Fortescue Group outcrop on the northern Pilbara Craton was not mapped by us during the present study, our interpretation of the stratigraphy in this area relies heavily on the work of others such as Blake (1984a, 1993), and those GSWA authors listed in Figure 1.2. Although we recognize the valuable contribution of these workers, we feel it would be premature for us to undertake a re-definition of the Fortescue Group without having examined all the evidence at first hand.

Secondly, the basin-wide extent of various unconformities and disconformities present on the northern Pilbara Craton has not been clearly demonstrated. Of the unconformable sequence boundaries recognized on the northern Pilbara Craton by Blake and Barley (1992) and Blake (1993), all but one are traced into correlative conformities in the southern Pilbara. The one possible exception is the base of the Hardey Formation, which is disconformable on the Mount Roe Basalt in the Rocklea

Dome but appears to be conformable in the neighbouring Bellary and Wyloo Domes.

Despite its shortcomings, therefore, we retain the term ‘Fortescue Group’ as a convenient way of referring to the diverse succession of volcano-sedimentary rocks outcropping above the Pilbara Craton granite–greenstones and below the Hamersley Group.

Hamersley Basin and principal Fortescue Group sub-basins

The term Hamersley Basin refers to the depositional basin of the Mount Bruce Supergroup and is used in the sense of Trendall (1990a) as ‘an area underlain by a substantial thickness of sedimentary rocks which possess unifying characteristics of stratigraphy and structure, due to their deposition during a regionally restricted episode of crustal depression or a related sequence of such episodes. As well as being applied to the deposited rocks, or to their outcrop area after erosion, the term is used also for the actual crustal depression in which they accumulated’. Under this scheme the Mount Bruce Supergroup represents a broad, unified stratigraphy that was deposited in a series of related tectonic settings, which evolved from rifted continent, to a passive margin, to a convergent margin over an interval of about 340 m.y. (Thorne, 1990; Barley et al., 1992; Blake, 1993).

Based on the work of Blake (1984a) and Thorne and Blake (1990), four Fortescue Group sub-basins, each showing a distinctive variation on the overall stratigraphic framework, are recognized within the Hamersley Basin. These are named the northwest Pilbara sub-basin, Marble Bar sub-basin, northeast Pilbara sub-basin, and the south Pilbara sub-basin (Fig. 2.3). Our sub-basins are closely equivalent to the ‘basins’ of Blake (1984a), but we prefer use of the term sub-basins (of the Hamersley Basin) to emphasize the depositional integrity of the Mount Bruce Supergroup. Like Blake’s (1984a) basins, the former extent of each sub-basin is uncertain, and their boundaries, as shown on Figure 2.3, are imprecise. This is due in part to the limited amount of outcrop and drillhole data from large parts of the southern Pilbara, but it also results from the fact that the location of the sub-basin boundaries changed over time. The south Pilbara and northwest Pilbara sub-basins and the Marble Bar and northeast Pilbara sub-basins were probably unconnected during deposition of the Mount Roe Basalt. By the time the Hardey Formation was deposited, the south Pilbara and northwest Pilbara sub-basins were separated from the northeast Pilbara and Marble Bar sub-basins by an upland area, termed the Yule–Sylvania High by Blake (1993). Whether the boundaries between the southern part of the northwest Pilbara sub-basin and the northwestern part of the south Pilbara sub-basin had merged by this time is uncertain due to lack of outcrop or drillhole data from the area immediately north of the Jeerinah Anticline. By the time the middle to upper formations (Kylena Formation/Boongal Formation to Jeerinah Formation) were deposited, boundaries between the northeast, northwest, and south Pilbara sub-basins appear to have merged completely. The results of this evolution are that the lower Fortescue Group (Mount Roe Basalt and Hardey Formation) shows considerable lithological variation within the various sub-basins, and middle to upper formations (Kylena Formation or Boongal Formation to Jeerinah Formation) show greater lateral continuity and exhibit less regional variation than is observed in the lower part of the Fortescue Group. For these units the principal regional differences occur between rocks of the south Pilbara sub-basin and the combined northwest and northeast Pilbara sub-basins.

Descriptions of Fortescue Group stratigraphy in each sub-basin are presented formation by formation in **Chapters 3–9**. The only exception relates to Fortescue Group rocks exposed in the Gregory Range, in the extreme eastern part of the Pilbara (Fig. 2.3). Here, structural complexity and stratigraphic complication make it appropriate to discuss the Fortescue Group geology of this area in a separate chapter (**Chapter 10**).

Geochronology

A substantial amount of geochronological work has been carried out on the Fortescue Group during the past 25 years (Table 2.1), and these results have been reviewed by Trendall (1983, 1990), Blake and McNaughton (1984), Trendall et al. (1990), Arndt et al. (1991), and Nelson, D. R. et al. (1992). It is apparent from these reviews that the Rb–Sr and Pb–Pb methods give anomalously young emplacement or depositional ages, owing to isotopic or chemical modification associated with subsequent events such as uplift and cooling, low-grade metamorphism, and hydrothermal alteration. In contrast, model Sm–Nd ages are unrealistically old, and may reflect either derivation from an older, fractionated mantle source, or crustal contamination (Nelson, D. R. et al., 1992). Zircon U–Pb analyses have the greatest potential to reliably measure the age of the Fortescue Group, and the following summary will be restricted to results gathered using this method.

Summary of zircon U–Pb results

The most complete study of Fortescue Group U–Pb zircon geochronology is that presented by Arndt et al. (1991). These workers report the results from nine widely separated felsic lava or tuff units occurring near the base, middle, and top of the group.

Lower Fortescue Group

Arndt et al. (1991) collected two samples from the Mount Roe Basalt. The first, which gave an age of 2763 ± 13 Ma, was obtained from a thin accretionary lapilli tuff located immediately above the basal unconformity in an area southwest of Whim Creek. The second sample came from a 60 m-thick lenticular unit of felsic volcanic rock some 700 m above the base of the formation on the northern Wyloo Dome; it yielded an age of 2775 ± 10 Ma. These dates are within error of the date of 2765 ± 2 Ma reported from a small rhyolite dyke which intrudes the Mount Roe Basalt and lower Hardey Formation near Meentheena Homestead (Thorpe, R., 1992, pers. comm.), and the age of 2772 ± 2 Ma reported from the Black Range Suite (Wingate, 1994, 1997, 1999). On this basis the best estimate for the age range of the Mount Roe Basalt is 2775–2663 Ma.

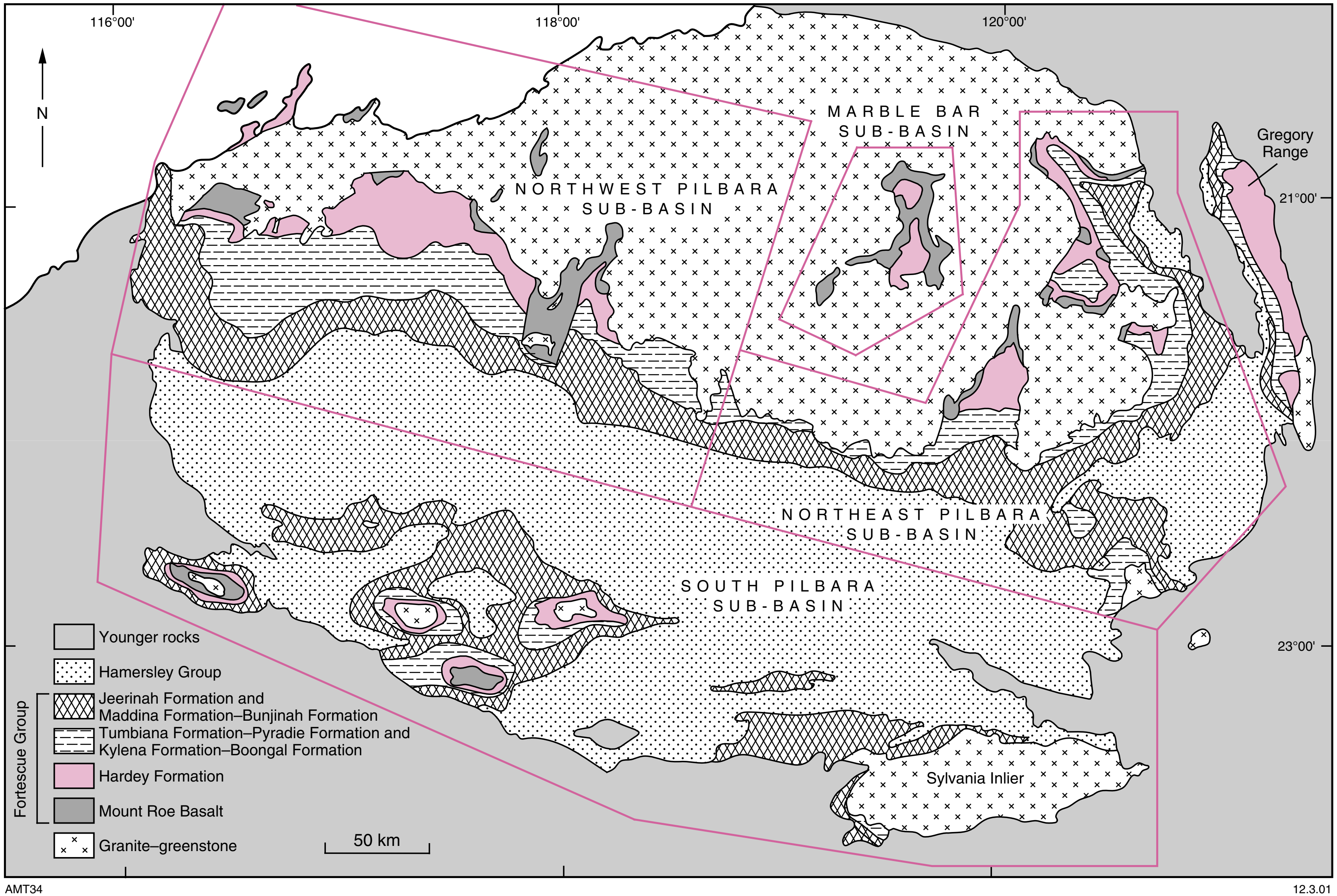

Figure 2.3. Geological map of the Pilbara Craton showing the location of the northwest Pilbara, Marble Bar, northeast Pilbara and south Pilbara sub-basins, based on the work of Blake (1984a) and Thorne and Blake (1990). As noted by Blake (1984a), the basin boundaries are for descriptive purposes only and are not meant to accurately reflect the extent of original depositional basins

Table 2.1. Compilation of available isotopic ages for the Fortescue Group

Method	*Age (Ma)*	*Unit and comments*	*Reference*
1	2629 ± 5	Jeerinah Formation	Nelson et al. (1999)
1	2690 ± 16	Jeerinah Formation	Arndt et al. (1991)
1	2684 ± 6	Jeerinah Formation	Arndt et al. (1991)
5	1487 ± 305	Jeerinah Formation	Blockley et al. (1980)
6	2144 ± 100[a]	Nallanarring Volcanic Member (Jeerinah Formation)	Compston and Arriens (1968)
1	2717 ± 2	Maddina Formation	Nelson (1998); Kojan and Hickman (1998)
5	2070 ± 114	Bunjinah Formation	Nelson et al. (1992)
3	2063 ± 245	Bunjinah Formation	Nelson et al. (1992)
4	2760 ± 20	Kylena Basalt (Formation) vein	Richards (1977), Richards et al. (1981)
1	2715 ± 6	Pillingini Tuff (Tumbiana Formation)	Arndt et al. (1991)
5	2150 ± 26[a]	Gidley Granophyre (two possible ages from	de Laeter and Trendall (1971)
	2557 ± 56[a]	(same data)	
7	2725 ± 1.3	Gidley Granophyre	Wingate (1997)
1	2760 ± 10	Sill intrudes contact between Kylena Basalt (Formation) and an unnamed sedimentary unit	Arndt et al. (1991)
4	2760 ± 30	Kylena Basalt (Formation) amygdales	Richards and Blockley (1984)
5	2100–2300	Various units	de Laeter and Trendall *quoted in* Trendall (1983)
2	3078 ± 107	Various West Pilbara units	Nelson et al. (1992)
4	2720 ± 20	Vein in intrusive dolerite	Richards (1977), Richards et al. (1981)
6	2878[a]	Cliff Springs Sandstone (Hardey Formation)	Compston and Arriens (1968)
1	2900–3250	Lyre Creek Agglomerate (Member)	Arndt et al. (1991)
5	2707 ± 151	Hardey Sandstone (Hardey Formation)	Hickman and de Laeter (1977)
5	2820 ± 516	Bamboo Creek Porphyry (Bamboo Creek Member)	Trendall (1975a)
5	2079 ± 195[a]	Spinaway Porphyry (Bamboo Creek Member)	Trendall (1975a)
1	2768 ± 16	Spinaway Porphyry (Bamboo Creek Member)	Pidgeon (1984)
4	2610 ± 20	Spinaway Porphyry (Bamboo Creek Member)	Unpublished data *in* Blake and McNaughton (1984)
1	2764 ± 8	Koongaling Volcanics (Member)	Arndt et al. (1991)
1	2756 ± 8	Bamboo Creek Porphyry (Bamboo Creek Member, Hardey Formation)	Arndt et al. (1991)
5	2028 ± 168	Mount Roe Basalt	de Laeter et al. *quoted in* Trendall (1983)
3	2545 ± 292	Mount Roe Basalt	Unpublished data *quoted in* Blake and McNaughton (1984)
3	2375 ± 200	Mount Roe Basalt	Unpublished data *quoted in* Blake and McNaughton (1984)
3	2466 ± 65	Mount Roe Basalt	Nelson et al. (1992)
5	2115 ± 447	Mount Roe Basalt	Nelson et al. (1992)
4	2770 ± 10	Basal Fortescue Group lavas	Richards (1981)
1	2763 ± 13	Mount Roe Basalt	Arndt et al. (1991)
1	2775 ± 10	Mount Roe Basalt	Arndt et al. (1991)
4	2720 ± 20	Vein in intrusive dolerite	Richards (1977), Richards et al. (1981)
5	2280 ± 89[a]	Black Range Suite, dolerite	Lewis et al. (1975)
7	2772 ± 2	Black Range Suite, dolerite	Wingate (1994, 1997)

NOTE: Key to method numbers:

1 U–Pb (zircon)
2 Sm–Nd (isochron)
3 Pb–Pb (whole rock/mineral isochron)
4 Pb (model)
5 Rb–Sr (whole rock/mineral isochron)
6 Rb–Sr (model)
7 U–Pb baddeleyite

(a) Ages have been reduced by a factor of 1.39/1.42 from the values reported by the authors who used λ^{87} Rb = 1.39×10^{-11} yr^{-1} in their original calculations. Rubidium–strontium ages not marked by an asterisk are from publications using λ^{87} Rb = 1.42×10^{-11} yr^{-1}

An age of 2756 ± 8 Ma was obtained from a rhyolite porphyry within the Bamboo Creek Member of the Hardey Formation, northeast Pilbara sub-basin. This member is thought to correlate broadly with the Lyre Creek Member of the northwest Pilbara sub-basin and the Koongaling Volcanic Member of the Gregory Range. Data from the Lyre Creek Member are equivocal and appear to reflect derivation from a 2900–3250 Ma source; zircons from a macroscopically banded rhyolite in the Koongaling Volcanic Member gave an age of 2764 ± 8 Ma; (Arndt et al., 1991). Another sample from the Gregory Range came from a rhyolite porphyry sill intruded along the contact between the base of the Kylena Formation and an underlying unnamed sedimentary unit. Its age of 2760 ± 10 Ma (Arndt et al., 1991) is within error of other dates from the lower Fortescue Group and the age of 2768 ± 16 Ma obtained by Pidgeon (1984) from the Spinaway Porphyry (Bamboo Creek Member) near Nullagine.

An important feature of lower Fortescue Group geochronological data is that they do not resolve the Mount Roe Basalt and Hardey Formation into separate volcano-sedimentary events. This is despite the considerable thickness of strata involved (about 2.5–5 km) and the fact that these formations are separated by an unconformity or disconformity locally.

Middle Fortescue Group

A mixed zircon population was obtained from a tuffaceous sandstone in the Tumbiana Formation (Pillingini Tuff) near Mount Herbert, northwest Pilbara sub-basin. Results indicate an extrusion age of 2715 ± 6 Ma with detrital and/or xenocrystic grains giving ages of 2936 ± 6 Ma and 3200–3350 Ma.

A U–Pb zircon date of 2717 ± 2 Ma from a rhyolite in the Maddina Formation (Nelson, 1998; Kojan and Hickman, 1998) shows that this formation is of very similar age to that of the underlying Tumbiana Formation, but is probably younger than the Gidley Granophyre, which has been dated at 2725 ± 1.3 Ma using U–Pb baddeleyite geochronology (Wingate, 1997)

Upper Fortescue Group

The younger age limit for the Fortescue Group is constrained by the following dates: 2690 ± 16 Ma, obtained from a tuffaceous sandstone which outcrops close to the top of the Jeerinah Formation near Paraburdoo, south Pilbara sub-basin; and 2684 ± 6 Ma, gathered from an andesitic ignimbrite in the Nallanaring Member, northwest Pilbara sub-basin (Arndt et al., 1991). In addition, a SHRIMP age of 2629 ± 5 Ma has been obtained by Nelson et al. (1999) from the very top of the Jeerinah Formation on NEWMAN (Nelson et al., 1999). Comparison between this date and those from the lower part of the stratigraphy implies the Fortescue Group was deposited over an interval of about 140 m.y.

Age of the Fortescue Group compared with that of the Black Range Suite

In the northeast Pilbara, a number of dolerite dykes belonging to the Black Range Suite are intimately associated with lower Fortescue Group rocks. Wingate (1994, 1997, 1999) obtained an age of 2772 ± 2 Ma from this suite using U–Pb baddeleyite geochronology. This age supports the views of earlier workers who suggested that these dykes may have been part of a feeder system for the Mount Roe Basalt (Lewis et al., 1975; Hickman and Lipple, 1975; Hickman, 1983, Blake, 1993).

Age of the Fortescue Group compared with that of the granite–greenstone basement and the Hamersley Group

The youngest reported U–Pb zircon ages from the granite–greenstone basement are 2851 ± 2 Ma from the Cooglegong Monzogranite (Nelson, 1998) and 2925 ± 16 Ma from the Munni Munni Complex (Arndt et al., 1991). These data, taken together with the c. 2770 Ma age of the lower Fortescue Group, suggest that there may have been a hiatus of about 80 m.y. between the emplacement of the youngest basement rocks and deposition of the lower Fortescue Group (cf. Blake, 1984a; Arndt et al., 1991).

Trendall et al. (1998) reported an U–Pb zircon age of 2595 ± 5 Ma from the Marra Mamba Iron Formation and an age of 2561 ± 8 Ma from the Wittenoom Formation, both in the lower part of the Hamersley Group. The age of the middle to upper part of the Hamersley Group is fixed by U–Pb zircon dates of 2490 ± 20 Ma from the Dales Gorge Member of the Brockman Iron Formation, and 2470 ± 30 Ma from the Woongarra Volcanics (Compston et al., 1981). When compared with the 2687 Ma age for the top of the Jeerinah Formation, these data imply either very slow rates of deposition for the lower Hamersley Group (<10 m/m.y.), or the presence of subtle hiatuses within the stratigraphy, or a combination of both (cf. Trendall et al., 1990; Blake,1993; Blake and Barley, 1992).

Chapter 3

Sedimentary and volcanic units below the Mount Roe Basalt

Although the Mount Roe Basalt is the lowest formally named stratigraphic division of the Fortescue Group, it is underlain locally by a variety of unnamed sedimentary rock units. These deposits overlie the irregular surface of the basal unconformity and are conformably overlain by the Mount Roe Basalt. The sedimentary units are mostly thin, ranging from a few decimetres to 30 m thick. More substantial accumulations of sedimentary rock are confined to two areas: firstly, in the Marble Bar sub-basin where Blake (1984a, 1993) documented that up to 1.8 km of sandstone and conglomerate outcrop as an unnamed unit above the basal unconformity, and secondly on TUREE CREEK in the south Pilbara sub-basin, where a unit of sedimentary and volcanic rocks at least 400 m thick, the Bellary Formation, conformably underlies the Mount Roe Basalt (Thorne et al., 1991).

Pre-Mount Roe Basalt Fortescue Group rocks consist mainly of clast- and matrix-supported conglomerate, lithic and feldspathic sandstone, and argillite; basalt flows, hyaloclastite, and tuffaceous deposits also occur locally. Three non-volcanic sedimentary facies associations are recognized: fan-delta, braided fluvial, and topographic hollow fill.

Fan-delta facies association

The fan-delta facies association has been recorded only from the Bellary Dome on TUREE CREEK, where it can be subdivided into subaqueous fan-delta and subaerial fan-delta facies. Principal rock types are argillite, fine- to very coarse grained sandstone, and conglomerate. Using the nomenclature of Folk (1974), non-volcanic sandstones from subaerial and subaqueous facies are classified as arkose, lithic arkose, feldspathic sedarenite, or sedarenite.

Subaqueous fan-delta facies

Description: This facies consists of grey-green mudstone and siltstone, interlayered with single sandstone beds or amalgamated sandstone units (Fig. 3.1a,b). The proportion of argillite to sandstone varies from 8:2 to 1:9.

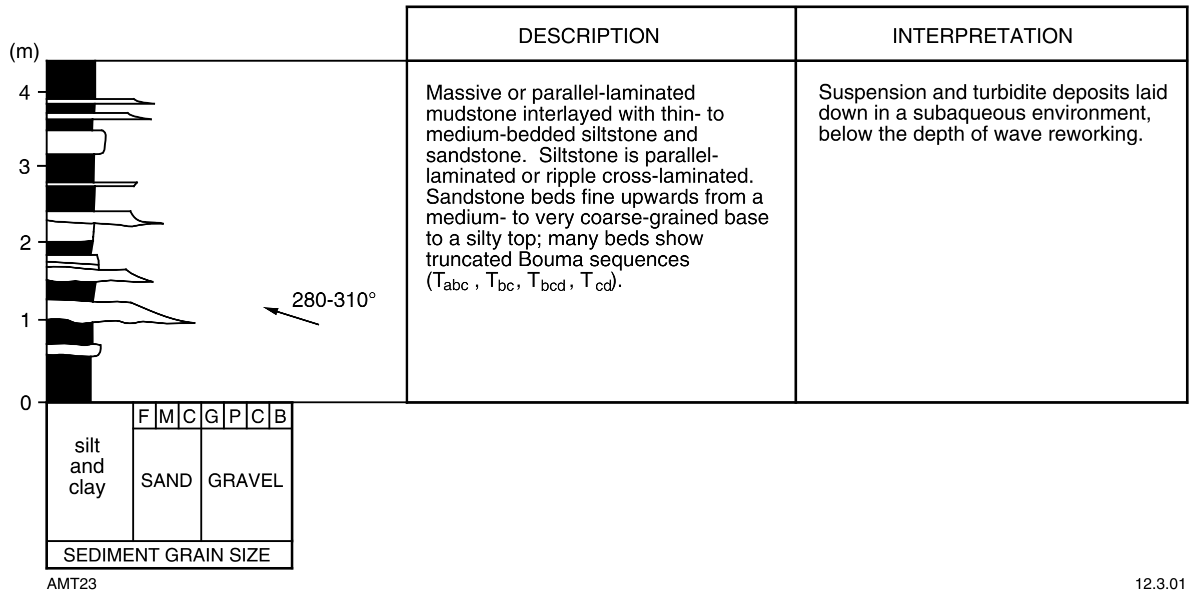

Figure 3.1. a) A representative section of argillaceous subaqueous fan-delta facies showing variation in bedding thickness, grain size, and internal structure, Bellary Formation, Bellary Dome, south Pilbara sub-basin. An explanation of graphic log symbols is given in Figure 1.4

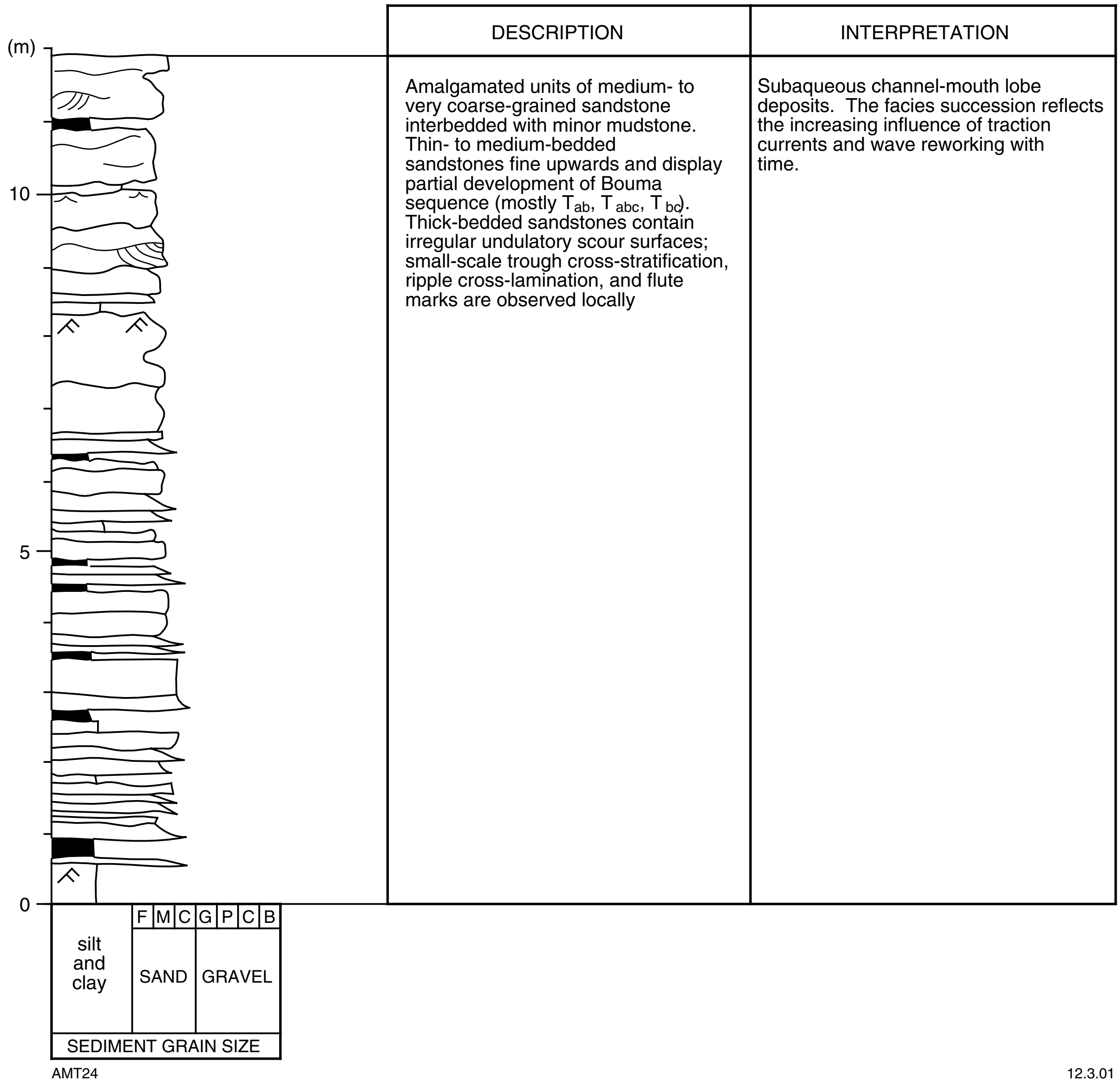

Figure 3.1. b) A representative section of sandstone-dominated subaqueous fan-delta facies showing variation in bedding thickness, grain size, and internal structure. Bellary Formation, Bellary Dome, south Pilbara sub-basin. An explanation of graphic log symbols is given in Figure 1.4

Mudstone and siltstone beds range in thickness from a few centimetres to about 1.5 m. They are generally massive or exhibit weak parallel lamination; climbing ripple cross-lamination is present locally in some coarse-grained siltstones.

Individual beds of fine- to very coarse grained sandstone are laterally persistent or sub-lenticular and generally range in thickness up to 0.3 m. Lower contacts are sharp and often scoured, and flute markings are preserved locally (Fig. 3.2). Top contacts are planar and generally grade upwards into siltstone or mudstone. The thickest sandstone beds are massive generally. Most thin- to medium-bedded sandstones fine upwards and many show partial development of the Bouma sequence of sedimentary structures (Allen, 1984); divisions Tabc or Tbc are preserved in thicker beds, whereas divisions Tbcd or Tcd characterize thinner layers.

Very thin to thick beds (0.01–0.35 m) of medium- to coarse-grained sandstone form lenticular or prismatic amalgamated units up to 20 m thick (Fig. 3.1b). Individual sandstone beds within these units show considerable lateral variation in thickness, resulting in irregular interbedding. Sandstone layers in lower parts of the unit are thin or thick bedded and fine- to very coarse grained, and the beds are often separated by siltstone and mudstone partings. These deposits are transitional with underlying argillite and sandstone units. Middle and upper levels of the amalgamated units are coarse grained and thick bedded, and are overlain by conglomeratic channel facies. The internal structure of the thick sandstone beds is

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Figure 3.2. Load-casted flute markings from subaqueous fan-delta sandstone. Palaeoflow was toward the west-northwest (top to bottom), Bellary Formation, Bellary Dome, south Pilbara sub-basin

dominated by irregular, discontinuous scour surfaces; small- to medium-scale trough cross-stratification is also observed locally. Parallel planar to undulatory lamination is present in many of the thinner, finer grained beds and symmetrical and asymmetrical ripple profiles are present on some bedding contacts. Flutes are observed locally on lower bedding surfaces.

Interpretation: The fine grain size, parallel lamination, and lack of scouring suggest that the mudstones were laid down from suspension in a low-energy environment. Parallel-laminated siltstones are also interpreted as suspension deposits for similar reasons. Siltstones that display climbing ripple cross-lamination were formed under lower flow regime conditions, where there was moderate to high sediment fallout from suspension (Harms et al., 1982).

Thin- to medium-bedded sandstones interlayered with the argillaceous deposits are interpreted as turbidites (Harms and Fahnestock, 1965; Walker, 1967). The various examples of truncated Bouma sequence are thought to represent deposition from turbidites of differing flow strengths and sediment grain sizes. Thick beds with a well-developed Ta division were deposited by powerful flows laden with coarse-grained sediment; thinner beds, preserving the Tbcd or Tcd divisions, were laid down from weaker currents that carried fine-grained detritus.

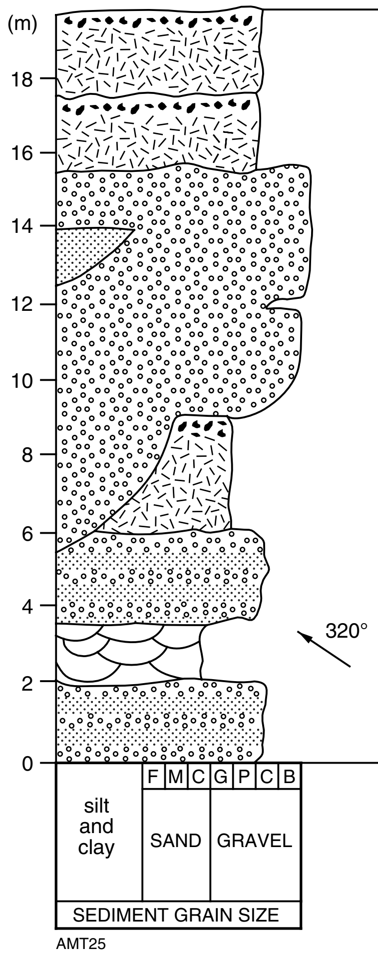
Amalgamated sandstone units are interpreted as channel-mouth lobes on the following basis:

- Their lenticular or prismatic geometry.
- The stratigraphic position of these units; they are transitional with turbidite and suspension deposits below, and are succeeded by conglomeratic channel facies.
- Their complex internal organization, which reflects a variety of depositional and sediment reworking processes. Thin- to thick-bedded sandstones in the lower parts of the units were probably emplaced as sediment gravity flows. Medium- to thick-bedded sandstones in middle and upper parts of the units record deposition from alternating sediment gravity flows and traction currents; the former are responsible for deposition of massive beds, whereas the latter produced localized trough cross-stratification. The occurrence of symmetrical ripples and undulatory lamination in some beds suggests that there was limited reworking of these deposits by wave processes.

Subaerial fan-delta facies

Description: This facies consists of massive and parallel-stratified conglomerate, pebbly sandstone, and coarse-grained sandstone. These rock types are interlayered with thin basalt flows and reworked tuff.

Massive, cobble to boulder conglomerate outcrops in erosively-based, lenticular units up to 9 m thick (Figs 3.3, 3.4). Clasts are generally well-rounded, maximum diameter is 0.6 m, but most lie in the range 0.05–0.35 m.



DESCRIPTION	INTERPRETATION
Parallel-stratified pebble and cobble conglomerate interbedded with channelized cobble to boulder conglomerate, trough cross-stratified sandstone, and vesicular basalt flows. Tabular bedded parallel-stratified conglomerate occurs interlayered with horizontally stratified sandstone; both contain sub-horizontal scour surfaces. Channelized polymictic conglomerate shows mixed clast and sand matrix support; clasts are rounded and up to 0.6 m diameter.	Subaerial fan-delta deposits. Parallel-stratified sandstone and conglomerate, and trough cross-stratified sandstone were laid down on outer parts of the subaerial fan surface. Erosively based bodies of cobble and boulder conglomerate are interpreted as fan channel deposits.

12.3.01

Figure 3.3. A representative section of subaerial fan-delta facies showing details of internal organization, Bellary Formation, Bellary Dome, south Pilbara sub-basin. An explanation of graphic log symbols is given in Figure 1.4

The conglomerate is predominatly clast-supported with a sandy matrix.

Parallel-stratified conglomerate occurs in tabular to lenticular units up to 12 m thick. These units exist in isolation or else form composite bodies up to 20 m thick. Clast-size is generally in the range 4–100 mm; most fragments are well rounded with low to high sphericity. Internal structure is dominated by planar parallel-stratification, although low-angle scours and cross-stratified sandstone lenses occur locally.

Tabular or lenticular bodies of coarse-grained sandstone and pebbly sandstone are intimately associated with the parallel-stratified conglomerate units. Most bodies are 0.5–5 m thick and medium to thick bedded. Sedimentary structures include horizontal stratification, with or without primary current lineation, low-angle scour surfaces, and medium-scale trough cross-stratification.

Interpretation: Erosively based bodies of cobble and boulder conglomerate are interpreted as alluvial fan-channel deposits. The absence of lag conglomerate, lack of sorting, and absence of stratification suggests they result from major erosive floods rather than prolonged channel occupation (cf. Heward, 1978a,b).

Parallel-stratified conglomerates are deposited under upper flow regime conditions in which coarse detritus is transported as bed load (Harms et al., 1982). The lack of channelling associated with many of the deposits described above implies that they were laid down as unrestricted sheetfloods, probably on distal parts of the subaerial fan.

Tabular beds of horizontally stratified sandstones are also interpreted as outer-fan sheetflood deposits. The presence of low-angle scours and current lineations, in association with the horizontal lamination, suggests that this facies was deposited in upper regime flows (Harms et al., 1982). Trough cross-stratified sandstones formed as a result of the downcurrent migration of small dunes under lower flow regime conditions (Harms and Fahnestock, 1965).

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Figure 3.4. Very thick bedded cobble to boulder conglomerate consisting of rounded fragments of monzogranite, metabasalt, metafelsic volcanic rock, and metasandstone, set in a sandy quartzofeldspathic matrix. Subaerial fan-delta facies, Bellary Formation, Bellary Dome, south Pilbara sub-basin. Jacobs staff is 1.5 m long

Braided fluvial facies association

The braided fluvial facies association occurs locally throughout the Hamersley Basin, but is best developed in the Marble Bar sub-basin. A summary of this facies association is given by Blake (1984a, 1993). Blake (1993) notes that in the Marble Bar sub-basin, up to 700 m, and in one area possibly more than 1.8 km, of dominantly very coarse grained to granular quartzofeldspathic sandstone and pebble to boulder conglomerate outcrop discontinuously beneath the Mount Roe Basalt. Palaeocurrent data for this unit suggest a convergence toward the present North Pole Dome and Lalla Rookh Syncline, with the major fluvial systems entering the sub-basin from the north and particularly from the southwest. The sandstones are dominantly arkosic and in both cases were derived from the surrounding granite batholiths Blake (1984a).

Topographic hollow-fill facies association

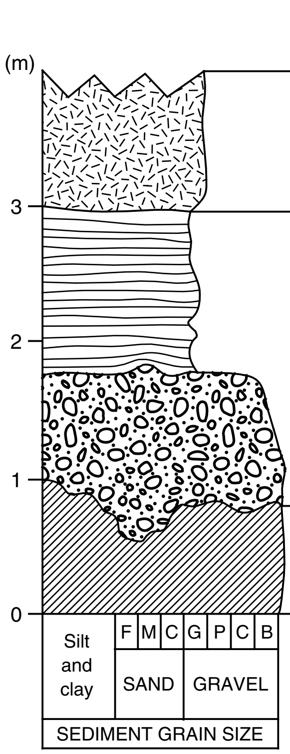
Topographic hollow-fill deposits are the most widespread of the pre-Mount Roe Basalt sedimentary facies associations and are recorded in each of the Fortescue Group sub-basins. They range in thickness from a few decimetres to about 35 m and consist of an impersistent, laterally variable assemblage of conglomerate, sandstone, siltstone, and mudstone. Two principal facies are recorded within this association: alluvial and colluvial, and lacustrine.

Alluvial and colluvial facies

Description: Alluvial and colluvial facies consist of matrix- and clast-supported conglomerate, pebbly sandstone, and coarse-grained sandstone.

Matrix- and clast-supported conglomerate outcrops in laterally impersistent layers up to 10 m thick. These deposits contain angular fragments of locally derived bedrock, up to 0.7 m in diameter. Some lower contacts are sharp, others display a rubbly, gradational contact with the underlying basement. Matrix-supported cobble and boulder conglomerates are usually polymodal, unstratified, and lack fabric; the matrix consists of sand- and granule-sized rock fragments (Fig. 3.5). Beds of cobble and pebble conglomerate may show both clast and matrix support and crude layering, parallel to the basal unconformity, exists locally.

Lenses of coarse-grained sandstone and pebbly sandstone are intimately associated with the conglomeratic units. Many display isolated or stacked sets of small- to medium-scale trough cross-stratification; parallel stratification, commonly accompanied by low-angle scouring, is also present locally.



DESCRIPTION	INTERPRETATION
Vesicular basalt flow	Subaerial Mount Roe Basalt
Laterally variable assemblage of massive, very poorly sorted, matrix- and clast-supported conglomerate and sandstone. Conglomerate clasts are angular and derived locally. Sandstone beds are parallel-stratified with local grading.	Palaeocolluvium overlain by sheetflood or lacustrine sandstone
Cleaved felsic volcanic rock	Pre-Fortescue Group greenstone basement

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Figure 3.5. A representative section of pre-Mount Roe Basalt topographic hollow-fill deposits from the Marble Bar sub-basin, showing variation in internal structure. An explanation of graphic log symbols is given in Figure 1.4

Interpretation: Matrix-supported cobble and boulder conglomerates show little evidence of sediment transport and are interpreted as colluvium and scree material which has accumulated close to the bedrock source. Structureless beds of mixed clast- and matrix-supported pebble and cobble conglomerate were probably deposited as a sliding, cohesionless sediment gravity flow (Postma, 1986). Clast composition and internal structure suggest short travel distances, and steep slopes would have been required to overcome internal friction. Conglomerate beds that display diffuse, crude layering may have originated as fully sheared cohesionless debris flows (Postma, 1986).

Solitary and stacked sets of cross-stratified sandstone and pebbly sandstone record deposition from lower flow regime currents (Harms et al., 1982). Parallel-stratified, scoured sandstones were probably deposited from localized upper flow regime sheet floods (McGowen, 1970).

Lacustrine facies

Description: This facies consists of grey-green and purple mudstone and siltstone interlayered with thin beds of fine- to coarse-grained sandstone, and minor conglomerate. Lacustrine units generally range in thickness from a few decimetres to 6 m and either overlie the basal unconformity directly, or else occur above alluvial and/or colluvial facies.

Mudstones are usually parallel-laminated; siltstones and fine-grained sandstones display either parallel lamination or ripple cross-lamination. Normal grading is observed in some sandstone and conglomerate layers.

Interpretation: These rocks were laid down in a subaqueous environment where deposition of mud and silt from suspension was interrupted by periods of higher energy sand and gravel sedimentation. The limited lateral extent of these deposits and their association with alluvial and colluvial facies indicates a lacustrine rather than marine origin.

Volcanic facies

A variety of volcanic rock types, including massive basalt and basaltic breccia, vesicular basalt, and mafic tuff occur in lower and upper levels of the Bellary Formation. They are interbedded with both subaerial and subaqueous fan-delta deposits and form about 25–30% of the exposed stratigraphy.

Massive basalt and basaltic breccia

Description: This facies association consists of thin- to thick-bedded massive, basalt flows interlayered with beds of basaltic breccia. These rocks form the lowermost 100 m of the Bellary Formation and are conformably overlain by subaqueous fan-delta sedimentary rocks.

Basaltic lava flows are 1–15 m thick and have flat or undulatory to lobate bounding surfaces. Flows are either massive or contain up to 15% irregular amygdales;

bedding-parallel lenticular cavities, ranging in size from 20–100 mm, are also developed locally. Amygdales and larger cavities are infilled by quartz and carbonate.

The basalts are fine grained (mostly 50–200 µm) and aphyric. Extensive post-depositional alteration has converted the original mineralogy to carbonate (40–60%), quartz, sericite, and chlorite. Major and trace element geochemical data from one flow are given in Table 12.2

Tabular to lenticular beds of poorly sorted, matrix-supported basaltic breccia are interlayered with the basalt flows. The beds are generally 0.1–3.0 m thick and lack any internal stratification. Contacts with the underlying basalt are very irregular and are either sharp or gradational; boundaries with overlying flows are generally sharp. Clasts are entirely basaltic in appearance and most range in size between 2 and 60 mm. They are non-vesicular and very angular with planar and curved grain margins.

Interpretation: Although their flow morphology provides little direct evidence as to the environment of deposition, this can be inferred from the nature of the interbedded breccias and the overlying sedimentary succession.

The matrix-supported breccias are interpreted as hyaloclastites, produced by the quench-fracturing of basaltic flows, on the following basis:

- Their homogeneous, non-vesicular composition.
- The predominance of poorly sorted, angular clasts.
- Local gradational (mosaic) contacts with massive to sparsely vesicular basalt flows, with which they are intimately associated.

Hyaloclastite breccias are produced when there is sudden contact between hot, coherent magma and cold water or water-saturated sediment (Cas and Wright, 1987). They form in a variety of situations: where magma is erupted subaqueously or subglacially; where lava flows into water or over water-saturated sediments; or where magma is intruded into water-saturated sediment or country rock. In the case of the breccias described here, geological setting rules out an intrusion-related, or subglacial origin. Likewise, the sharp lower boundaries and transitional upper contacts of the flows argue against either the flow-into-water or flow-over-wet-sediment hypotheses.

The only depositional setting that satisfies all the available evidence is that in which the breccias originated from the quench-shattering of subaqueously erupted basaltic magma. This interpretation is also consistent with data from the conformably overlying sedimentary unit, which is characterized by a thick (~100 m) assemblage of subaqueous fan-delta deposits.

The lack of stratification in the hyalocalastite breccias, combined with their matrix support, poor sorting, angular clasts, and presence of transitional boundaries with the basalt flows suggests that most of them accumulated close to source, either as a flow-top mantle, or marginal scree and sediment gravity-flow deposits.

Amygdaloidal basalt flows

Description: The upper part of the Bellary Formation locally contains up to seven vesicular basalt flows, ranging in thickness from 1–7.8 m. They are generally interbedded with subaerial fan-delta facies, but are also associated locally with subaqueous deposits.

Amygdaloidal basalt flows range in thickness from 1 to 8 m. They tend to maintain uniform thickness within each outcrop, although variation in exposure quality precludes the tracing of a single flow laterally for more than a few hundred metres. Lower parts of the lava flows are massive to sparsely amygdaloidal; middle and upper parts contain 10–40% spherical to amoeboid amygdales, either distributed randomly or concentrated in deformed, flow-banded patches. Flow tops are sharp and irregular and are commonly highly amygdaloidal (>50% amygdales); small pockets of breccia occur locally.

The basalts are mostly fine-grained (50–1200 µm) and aphyric, although plagioclase-phyric varieties have also been recorded. Elongate crystals define a random, or weak, pilotaxitic texture. Equant to prismatic, albitized plagioclase forms the largest grains and phenocrysts. These make up 30–45% of the rock, and contain chloritized inclusions and embayments locally. Intergranular and groundmass material consists of carbonate, chlorite, quartz, sericite, and granular sphene.

The average of eight analyses from three flows of this facies appears as group BEPD (Paraburdoo) in Table 12.2; the individual analyses are plotted on Figure 12.2 (see **Chapter 12**). The average is close to the basaltic andesitic composition (Le Maitre, 1989) typical of the mafic lavas higher in the Fortescue Group succession, although the silica content (52%) is marginally lower. The trace element pattern also has the enrichment of incompatible elements relative to MORB (Fig. 12.9) associated with stratigraphically higher flows.

Interpretation: The morphology, thickness, and uniform character of these flows, together with their lack of pahoehoe or aa surfaces and lava tubes, indicate that they probably accumulated as rapidly emplaced, ponded lava flows.

Mafic tuff

Description: Two laterally persistent beds of ferruginous mafic tuff are associated with subaerial and subaqueous fan-delta deposits in the upper part of the Bellary Formation. The tuff beds range in thickness from 1.5 to 10 m and display sharp, non-erosive lower contacts and sharp, or transitional upper contacts. Internal structure is heterogeneous; some beds are massive, apart from randomly distributed, small, irregular vesicles, but other laterally equivalent levels show horizontal stratification or isolated sets of trough cross-stratification.

The tuff layers show a considerable range in composition. Massive portions contain about 30% relic crystal fragments (now replaced by carbonate and quartz) set in a groundmass of chlorite and albitized feldspar. Locally, however, the tuff is more acidic in composition, consisting of fractured beta-quartz phenocrysts in a matrix of fine-grained quartz and feldspar. Stratified portions of the tuff contain sparse to abundant blocky to highly cuspate vitric fragments (replaced by chlorite) mixed with various proportions of non-volcanic fragments, including polycrystalline quartz, sandstone, basalt, and chert. The matrix generally consists of very fine grained chlorite of indeterminate origin.

Interpretation: The precise mode of origin of these tuff beds is unclear. Their fine grain size, uniform thickness range in subaerial and subaqueous settings, non-erosive lower contacts, generally basaltic composition, and presence of blocky and cuspate vitric fragments suggest they accumulated as a series of air-fall deposits from a vesiculated phreatomagmatic source. In this context, the paucity of horizontal stratification and grading is puzzling, and the local occurrence of small-scale trough cross-stratification has to be explained as the result of post-depositional reworking. Irregular vesicles, evident in some massive beds, imply that trapped air or steam could not escape from what must have been a water-saturated ash at the time of deposition (Cas and Wright, 1987).

Distribution of facies associations

Pre-Mount Roe Basalt Fortescue Group rocks have been recorded from the four principal sub-basins of the Hamersley Basin. The major occurrences are summarized below.

Northwest Pilbara sub-basin

Thin arkosic sandstone and conglomerate units outcrop below the basal lavas of the Mount Roe Basalt in a number of localities on DAMPIER AND BARROW ISLAND, PYRAMID, and ROEBOURNE. Although these units have not been examined during the course of the present study, and published descriptions of them are sketchy, their lenticularity, thickness, and composition suggest they are mostly topographic hollow-fill deposits.

Kriewaldt (1964a), and Kriewaldt et al. (1964) recorded the presence of arkose lenses at the base of the Mount Roe Basalt 9.0 km north-northeast, and 7.5 km southeast of Mount Wilkie. These authors noted that similar rocks are also present along the contact between the Gidley Granophyre and the Mount Roe Basalt on East Lewis Island in the Dampier Archipelago. Because of the intrusive nature of the Gidley Granophyre, it is not known with certainty whether these East Lewis Island arkoses are basal Fortescue Group units, or sedimentary deposits occurring higher in the Mount Roe Basalt.

During their mapping on ROEBOURNE, Ryan (1966) and Ryan et al. (1965) recorded the presence of sandstone and argillite at the base of the Mount Roe Basalt between Mount Ada and Mount Oscar. More recent work indicates that these deposits pre-date the Fortescue Group and belong to sedimentary units intercalated within the Warrambie Basalt (Hickman, A. H., 1999, pers. comm.).

On PYRAMID, thin basal arkoses and tuffaceous sandstones outcrop 2.5 km north of Bluff Hill, 1 km south of Croydon Top Camp, 7.5 km south of Nunyerry, and 10 km west-southwest of Teichman Well (Kriewaldt and Ryan, 1963; Blake, 1990). No details of these occurrences have yet been published.

Marble Bar sub-basin

Up to 500 m of sedimentary rocks are recorded locally at the base of the Fortescue Group on MARBLE BAR and PORT HEDLAND – BEDOUT ISLAND. Maitland (1908), Finucane (1938a), and Noldart and Wyatt (1962) provided early descriptions of auriferous basal conglomerates at the Just-in-Time and Tassy Queen mines, and later summaries are given by Hickman and Lipple (1978), Hickman (1983), and Hickman and Harrison (1986). These, and other pre-Mount Roe Basalt sedimentary units within the Marble Bar sub-basin, have been mapped by Hickman and Lipple (1978), Hickman and Gibson (1982), and Blake (1990).

The discontinuous basal sedimentary unit generally consists of thin topographic hollow-fill deposits, overlain either by braided fluvial sandstone and conglomerate or by the Mount Roe Basalt. In areas where the topographic hollow-fill deposits are absent, braided-fluvial facies overlie the basal unconformity directly.

Topographic hollow-fill deposits consist of lenticular beds of poorly sorted, cobble to boulder conglomerate (colluvial and/or alluvial facies), overlain locally by argillite and sandstone (lacustrine facies). Earlier published descriptions (Maitland, 1908; Finucane, 1938a), and mapping carried out in the Big Stubby area during the course of this study, have shown that there is often a strong correlation between conglomerate clast type and the underlying bedrock lithology. This evidence, combined with the textural characteristics of the conglomerate, indicates that most of these deposits were derived locally.

Blake (1984a, 1993) records up to 700 m of braided fluvial sandstone and conglomerate in the Marble Bar sub-basin. These deposits outcrop in the Gorge Range and Halleys Comet gold mine area, and in an arcuate belt in the southwestern part of the Marble Bar sub-basin (between NORTH SHAW AMG 455446 and TAMBOURAH AMG 260116) (Blake, 1990). Palaeocurrent data from these rocks are shown in Figure 3.6 (modified from Blake, 1984a, fig. 4) and suggest a convergence toward the present day North Pole Dome, and Lalla Rookh Syncline. Major fluvial systems entered the Marble Bar sub-basin from the north, and particularly from the southwest. The sediments deposited were dominantly arkosic and derived from the granitoid batholiths (Blake, 1984a).

Northeast Pilbara sub-basin

Only minor occurrences of discontinuous basal conglomerate, sandstone, and siltstone (topographic hollow-fill deposits) exist in the northeast Pilbara sub-basin. The principal outcrops lie on NULLAGINE between Yandicoogina Well and Police Creek and south of Mount Elsie (Thom et al., 1979; Blake, 1990).

South Pilbara sub-basin

Pre-Mount Roe Basalt sedimentary units have been recorded from the Wyloo Dome (Horwitz, 1978; Seymour

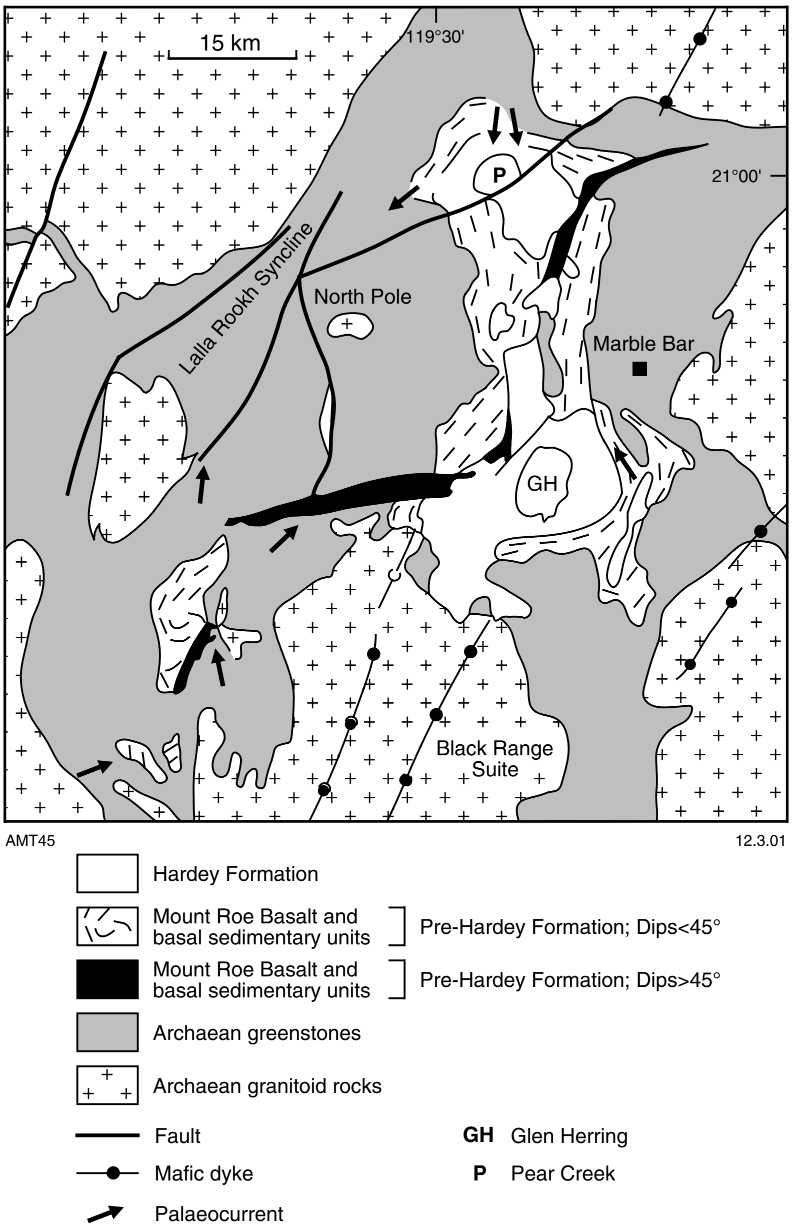

Figure 3.6. Geological sketch map for the Marble Bar sub-basin showing fluvial palaeocurrents (trough cross-strata) for the pre-Mount Roe Basalt sedimentary unit. The solid black area is the post-Mount Roe Basalt – pre-Hardey Formation fold belt. P = Pear Creek; GH = Glen Herring; NP = North Pole; LRS = Lalla Rookh Syncline; BR = Black Range Suite; MB = Marble Bar. Modified from Blake (1984a, fig. 4)

et al., 1988), Bellary Dome (Thorne et al., 1991; Thorne and Tyler, 1994), southeastern margin of the Rocklea Dome (Thorne and Tyler, 1996), and in the SGS–1 diamond drillhole (CRA Exploration Pty Ltd, 1988) in north-central MOUNT BRUCE (see Appendix 2).

On the northern flank of the Wyloo Dome (WYLOO AMG 366978) the Mount Roe Basalt is underlain by several metres of topographic hollow-fill deposits comprising trough cross-stratified arkose (Fig. 3.7) and pebbly arkose, overlain locally by up to 0.5 m of parallel-laminated siltstone. Coarser grained facies occur adjacent to the Metawandy Granite (WYLOO AMG 443914). Here, there is an upward transition from monzogranite bedrock, through angular cobble and boulder conglomerate with sandstone matrix, into more rounded pebble and cobble conglomerate and pebbly sandstone (Seymour et al., 1988).

A 400 m-thick succession of sedimentary and volcanic rock, named the Bellary Formation, outcrops in the core of the Bellary Dome on TUREE CREEK (Thorne et al., 1991; Thorne and Tyler, 1994). The lower contact of the Bellary Formation is not exposed but is thought to be unconformable upon granite–greenstone basement; its upper contact is with conformably overlying Mount Roe Basalt.

Figure 3.7. Photomicrograph of pre-Mount Roe Basalt topographic hollow-fill sandstone overlying the Metawandy Granite, Wyloo Dome, south Pilbara sub-basin. Scale bar is 0.2 mm long. Photograph by D. F. Blight

Bellary Formation rocks have previously been assigned to the Hardey Formation (Daniels et al., 1967) and, subsequently, to a basement greenstone succession (Blight, 1985). The first interpretation is rejected on the grounds that the Bellary Formation is clearly overlain by the Mount Roe Basalt, whereas the second view is inconsistent with a gradational, conformable contact between the Bellary Formation and the Mount Roe Basalt (Thorne and Tyler, 1994).

The lower part of the Bellary Formation comprises over 100 m of subaqueous basalt flows and hyaloclastite overlain by 200 m of subaqueous fan-delta argillite and sandstone (Fig. 3.8). In the eastern Bellary Dome, subaqueous facies are overlain by the lowest flows of the Mount Roe Basalt; farther west, they are capped by 100 m of subaerial fan-delta sandstone and conglomerate, interbedded with vesicular basalt flows and basaltic tuff. The limited amount of palaeocurrent data, obtained from both subaerial and subaqueous facies, indicates that sediment transport was towards the west-northwest.

On the southern margin of the Rocklea Dome (ROCKLEA AMG 406702), a thin unit of medium- to coarse-grained feldspathic sandstone, pebbly sandstone, and conglomerate outcrops between basement ultramafic flows and the overlying Mount Roe Basalt. Although field relationships are not clear, these deposits may represent basal Fortescue Group sedimentary facies belonging to either the topographic hollow-fill or braided-fluvial facies associations. The only other reported occurrence of basal Fortescue Group sedimentary rocks on MOUNT BRUCE is from the SGS–1 diamond drillhole (CRA Exploration Pty Ltd, 1988). Here, 22 m of coarse-grained feldspathic sandstone (?braided-fluvial facies association) overlie basement granitoid rocks with a sharp contact.

Depositional model and palaeogeographic reconstruction

Attempts to construct a depositional model for pre-Mount Roe Basalt Fortescue Group sedimentation are hampered by three major difficulties:

- Basal sedimentary rocks are exposed only in a relatively small area of the Hamersley Basin. In addition, the original extent of the Hamersley Basin is currently unknown.
- There is a lack of those data which would allow effective correlation of the various basal sedimentary units across the Pilbara. Theoretically, these sedimentary rocks could have been deposited at any time between the cratonization of Pilbara basement and deposition of the Mount Roe Basalt — a time span of about 80 m.y. (**Chapter 2**). The gradational upper contact of the Bellary Formation suggests, however, that this unit was probably deposited a relatively short time before the Mount Roe Basalt was extruded. Similarly, Blake (1984a, 1993) argues that the basal sedimentary units of the Marble Bar area were also deposited in response to the same extensional event as that which produced the Mount Roe Basalt.
- Intra-Fortescue Group erosion may have removed many of the basal deposits from large areas of the Pilbara.

Notwithstanding the above-mentioned difficulties, the uniform character of basal sedimentary facies over a large area of the Hamersley Basin, combined with variations observed in the Marble Bar outlier and Bellary Dome, suggests that the following depositional model may be

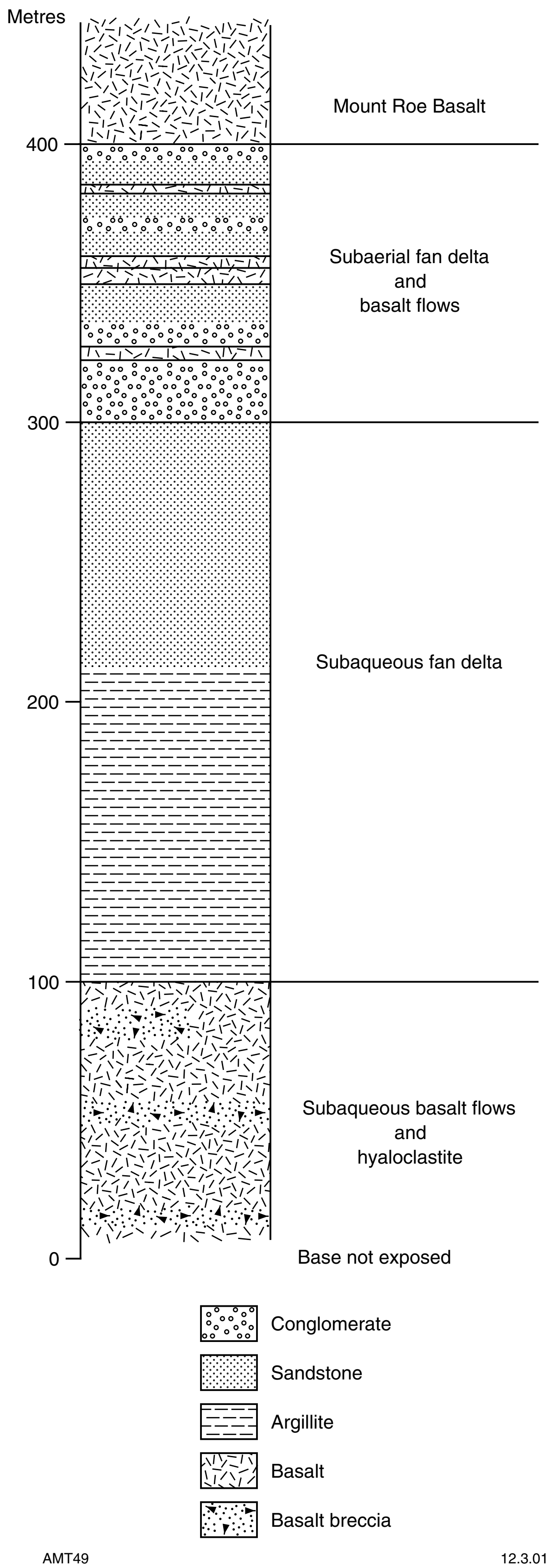

Figure 3.8. Generalized stratigraphy of pre-Mount Roe Basalt sedimentary units (Bellary Formation) in the Bellary Dome, south Pilbara sub-basin

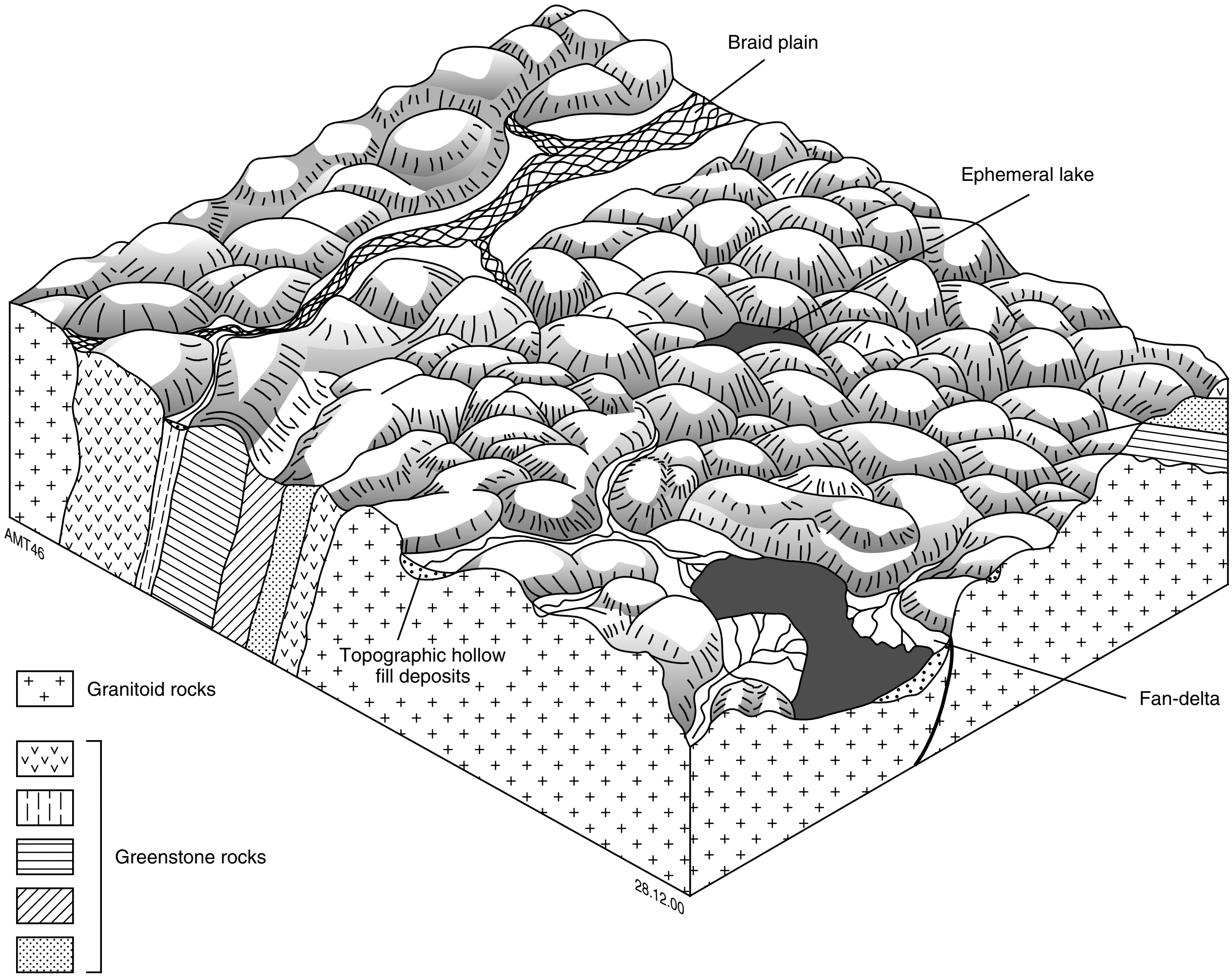

Figure 3.9. Depositional model for pre-Mount Roe Basalt sedimentary units of the Fortescue Group

applicable. It is based upon the nature of the facies, and their vertical and spatial distribution, and summarizes the main depositional environments of the Fortescue Group immediately before extrusion of the Mount Roe Basalt. The principal features of the model are shown in Figure 3.9.

Basal Fortescue Group sedimentary units were deposited on a rugged landscape of low to moderate relief. Local topographic highs were controlled by more-resistant greenstone lithologies such as cherts, acid volcanics, and silicified sandstones; areas underlain by granitoid were generally of more subdued relief. Thin, laterally impersistent bodies of alluvial/colluvial conglomerate and sandstone were deposited in small valleys and topographic hollows. Locally, finer grained sandstone and argillite accumulated in shallow, ephemeral lakes.

Major sandy, braided-fluvial systems drained into the Marble Bar area from granitoid uplands to the north and southwest. Near Bellary Dome, in the southwestern Hamersley Basin, a small fan-delta system built out toward the west-northwest in a small, possibly fault-bounded lake or marine embayment. Fan-delta sedimentation was accompanied locally by the extrusion of basaltic lava.

Chapter 4

Mount Roe Basalt

The Mount Roe Basalt (Kriewaldt, 1964b) occurs in discontinuous outcrops throughout most of the Hamersley Basin, but is absent from the Sylvania Inlier in the extreme southeast, and from the Gregory Range (Fig. 4.1). The formation is up to 2.5 km thick (Blake, 1993) and either overlies granite–greenstone basement unconformably, or rests conformably upon localized, mostly thin, basal sedimentary units (**Chapter 3**). The Mount Roe Basalt is unconformably overlain by the Hardey Formation in the northwest Pilbara, Marble Bar, and northeast Pilbara sub-basins (Blake, 1984a); in the south Pilbara sub-basin this contact is either disconformable or conformable.

The age of the Mount Roe Basalt (**Chapter 2**) is constrained by U–Pb zircon dates of 2763 ± 13 Ma and 2775 ± 10 Ma obtained for the base and lower part of the formation respectively on ROEBOURNE and WYLOO (Arndt et al., 1991). Uranium–lead zircon data from the overlying Hardey Formation and the Koongaling Volcanic Member of the Gregory Range lie within the range 2768 ± 16 Ma to 2756 ± 8 Ma (Pidgeon, 1984; Arndt et al., 1991). Wingate (1994,1997, 1999) obtained an age of 2772 ± 2 Ma from dolerite of the Black Range Suite, using U–Pb baddeleyite geochronology. This age supports the views of earlier workers who suggested that these dykes may have been part of a feeder system for the the Mount Roe Basalt (Lewis et al., 1975; Hickman and Lipple, 1975; Hickman, 1983; Blake, 1993)

Previous descriptions of the Mount Roe Basalt are given by Daniels (1970), Hickman (1978, 1983), Hickman and Gibson (1982), Hickman and Lipple (1978), Hickman et al. (1983), Kriewaldt (1964a,b), Kriewaldt and Ryan (1967), Ryan (1966), Seymour et al. (1988), Thorne and Tyler (1996, 1997a), and Williams (1989). The formation has also been described under various other formal and informal names (Blake, 1984a, 1993; Noldart and Wyatt, 1962; Trendall, 1975b, 1990b). This chapter describes and interprets the principal volcanic and sedimentary facies associations present in the Mount Roe Basalt. Most of the data have been obtained from seven measured sections and 1:40 000-scale mapping of selected outcrop areas in the south and northwest Pilbara sub-basins. Information from the northeast Pilbara and Marble Bar sub-basins has been gathered from reconnaissance mapping, supplemented by descriptions from other workers cited in the text.

The Mount Roe Basalt consists mainly of basalt and basaltic andesite flows and local pillow lava, interbedded with minor tuff, hyaloclastite, and epiclastic rocks. Non-volcanogenic sedimentary rocks are also recorded; these include clast- and matrix-supported conglomerate, feldspathic quartz sandstone, and argillite. A palaeosol has also been described from the Mount Roe Basalt on ROEBOURNE (Macfarlane and Holland, 1990).

Volcanic facies

Volcanic rocks within the Mount Roe Basalt can be subdivided into subaerial basaltic lavas, subaqueous basaltic lavas, and water-lain volcaniclastic rocks.

Subaerial basaltic flows

Subaerial basaltic flows are the dominant volcanic facies within the Mount Roe Basalt throughout the Hamersley Basin. Flows range from less than a metre to 82 m thick. The average thickness of 55 flows measured at Paraburdoo in the south Pilbara sub-basin was 10 m, with a median thickness of 7 m (Fig. 4.2). In contrast, data from 70 flows in the Mount Roe, Mount Anketel, and Mount Leopold areas of the northwest Pilbara sub-basin show average and median thickness of 25 and 17 m respectively.

As noted by Blake (1993), two broad categories of basaltic flow can be recognized on the basis of thickness and flow structure. These are here referred to as thick, massive to amygdaloidal flows, and thin pahoehoe-type flows.

Massive to amygdaloidal flows

Most massive to amygdaloidal flows vary in thickness from about 2 to 78 m with most values lying in the range 6–15 m. They are the dominant flow type in the south Pilbara, northwest Pilbara and Marble Bar sub-basins.

Apart from their generally greater thickness, most massive to amygdaloidal flows are characterized by a high proportion (25–60% of the rock) of large spherical to irregularly rounded or streaked amygdales in upper parts of the flow. These upper levels contrast with lower and middle parts of the flow, which commonly contain only a few percent of amygdales. Most amygdales range in size up to about 30 mm and are either scattered randomly throughout the rock or concentrated in folded, flow-aligned clusters or bedding-parallel sheets. The majority

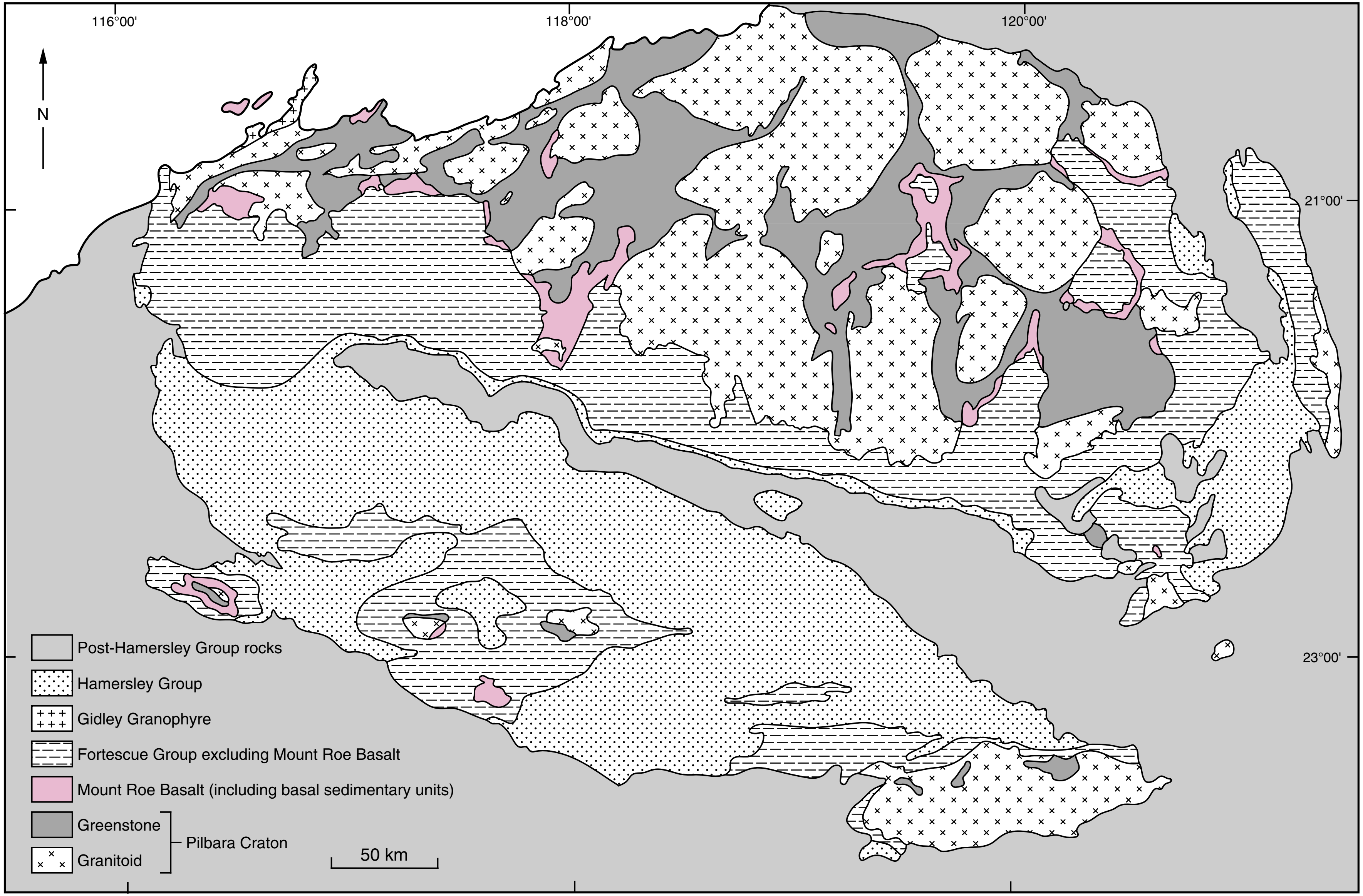

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Figure 4.1. Principal outcrop areas of the Mount Roe Basalt

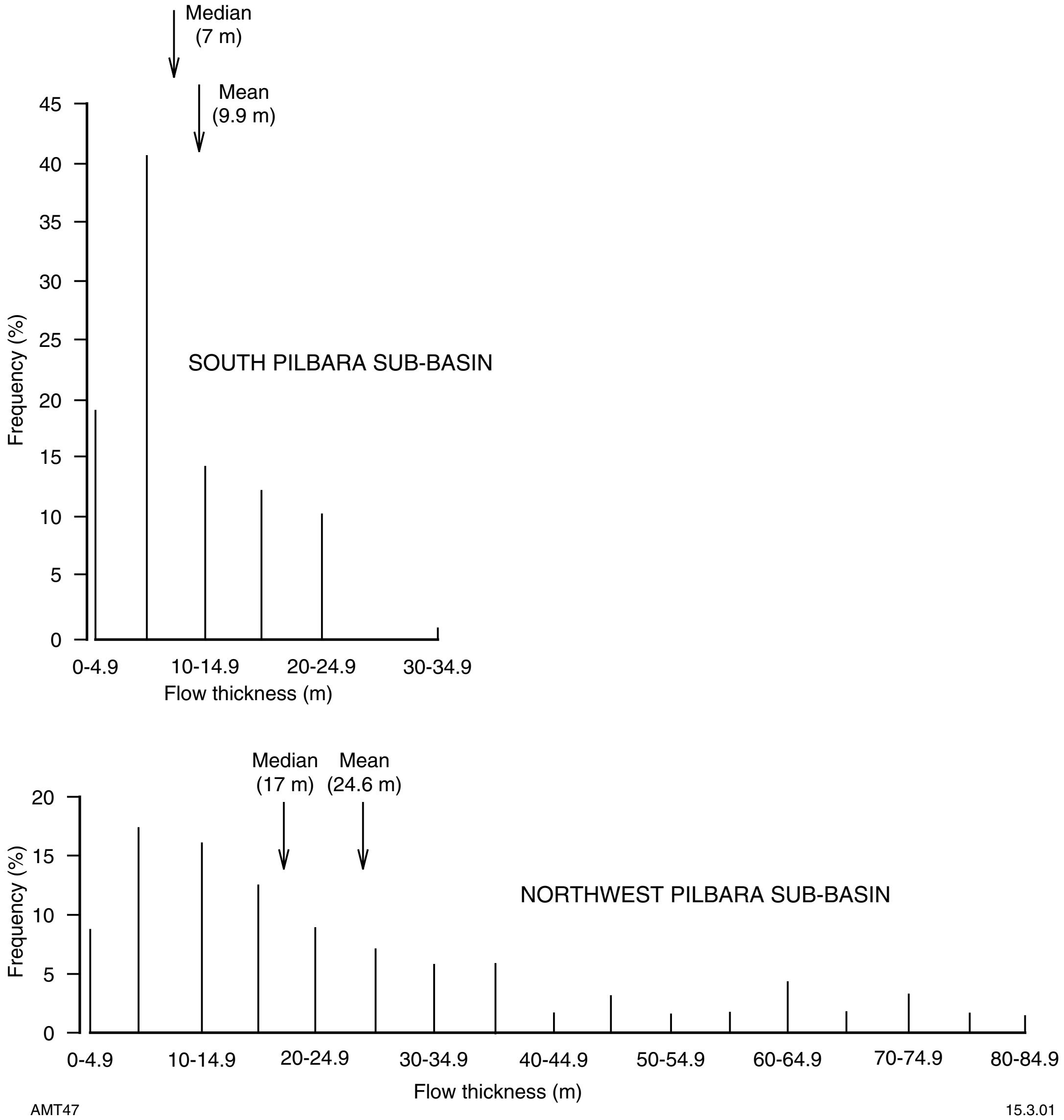

Figure 4.2. Histograms summarizing Mount Roe Basalt flow thickness in the northwest and south Pilbara sub-basins. Flow numbers for the northwest and south Pilbara sub-basins are 70 and 55 respectively

of amygdales are infilled by various combinations of quartz, carbonate, and chlorite.

Pipe amygdales and amygdale cylinders are recorded locally in massive to amygdaloidal flows. Planar-convex cavities are rarely observed; however, irregular voids, 0.05–0.15 m in diameter, are present near the top of several of the thicker flows.

Massive to amydaloidal flows are bounded by tabular flow surfaces which exhibit gentle, irregular undulations at outcrop scale. Thick flows (>20 m) are laterally persistent over distances of 1–3 km, but poor outcrop continuity and general absence of diagnostic features make it difficult to trace them further. One possible exception is a 78 m-thick columnar-jointed flow 140 m above the base of the formation at Mount Roe, which is closely similar to a 72 m-thick flow at the same stratigraphic level at Mount Anketel, 25 km farther north. The lateral extent of thinner (<20 m) flows is unknown.

Massive to amygdaloidal flow tops generally exhibit irregular to broadly symmetrical undulations. Wavelength and amplitude vary with flow thickness but are generally in the range 0.5–10 m and 0.1–1.2 m respectively. Flow tops are almost everywhere scoriaceous and may merge upwards into an aa-like rubbly breccia which may be up to 6 m thick locally (e.g. Mount Leopold). Flow bases are generally smooth and mimic irregularities in the underlying surface. More rarely the lower parts of the flow comprise a hyaloclastite breccia, which is transitional upwards into massive or amygdaloidal basalt.

Most massive and amygdaloidal basalts show little systematic jointing, although columnar joints and irregular joint networks have been observed locally. Columnar joints have been recorded in only eight flows, all over 20 m thick, and all in the northwest Pilbara sub-basin. Individual columns are mostly 0.15–0.3 m in diameter and are present in lower and middle to upper parts of the flows; column axes are generally oriented 45–90° to the flow surfaces. No examples of the classic collonade–entablature structure (Long and Wood, 1986) were observed.

In the Paraburdoo area the lower levels of thick, massive flows are cross-cut locally by an irregular network of curved and planar (tortoiseshell) joint surfaces (Fig. 4.3). These joints are similar in plan and cross section and give the rock an appearance that ranges from irregular net-like veining to a pseudobreccia. Joint surfaces generally display a plumose fracture pattern and there is no evidence of significant relative movement between adjacent 'fragments'.

Pahoehoe-type flows

Blake (1993) recorded thin (0.1–3.0 m) basalt flows from the northern Pilbara that are similar to dense, hummocky pahoehoe lava. These flows have not been reported from the southern part of the craton.

Summary of petrography: Massive to amygdaloidal flows are generally fine to coarse grained (40–1500 µm) and aphyric, although plagioclase-phyric varieties are present locally (Fig. 4.4). In the latter, plagioclase (replaced by albite, with or without epidote and chlorite) phenocrysts are up to 3 mm wide by 15 mm long and occur singly, or in distinctive elongate glomoporphyritic aggregates (Hickman, 1983).

Equant to prismatic, albitized plagioclase is abundant within the groundmass and lath-like to prismatic varieties generally define a random, or less commonly a pilotaxitic, texture. Equant to tabular, subhedral to euhedral pyroxene rarely forms more than 5–10% of the rock and is commonly replaced by tremolite, chlorite, and serpentine. Interstitial material is altered to a variety of minerals including tremolite, chlorite, carbonate, pumpellyite, quartz, serpentine, and epidote. Despite this pervasive alteration, it is often possible to distinguish original intersertal and intergranular textures.

Summary of geochemistry: A total of 109 analyses are available of subaerial basaltic flows of the Mount Roe Basalt. The analysed samples come from a wide area of the basin. In terms of the scheme used in **Chapter 12** to subdivide the analytical data available for mafic rocks of the Fortescue Group they include, firstly, the MRPD (Paraburdoo), MRRD (Rocklea Dome), MRBM (Corehole BMW2) and MRSG (Corehole SGS1) groups; the 30 analyses in these four groups all come from the south Pilbara sub-basin and form, with BEPD, cluster A (Tables 12.1, 12.2). Secondly, a further 26 analyses are available from the northwest Pilbara sub-basin, in groups MRGS (Gudrun Sieber) and MRWC (Whim Creek). And finally, 53 analyses have been published from the northeast Pilbara sub-basin, in groups MRMB (Marble Bar), MRGH (Glen Herring), MRNG (Nullagine) and MRYC (Yandicoogina). These latter 79 analyses form cluster B.

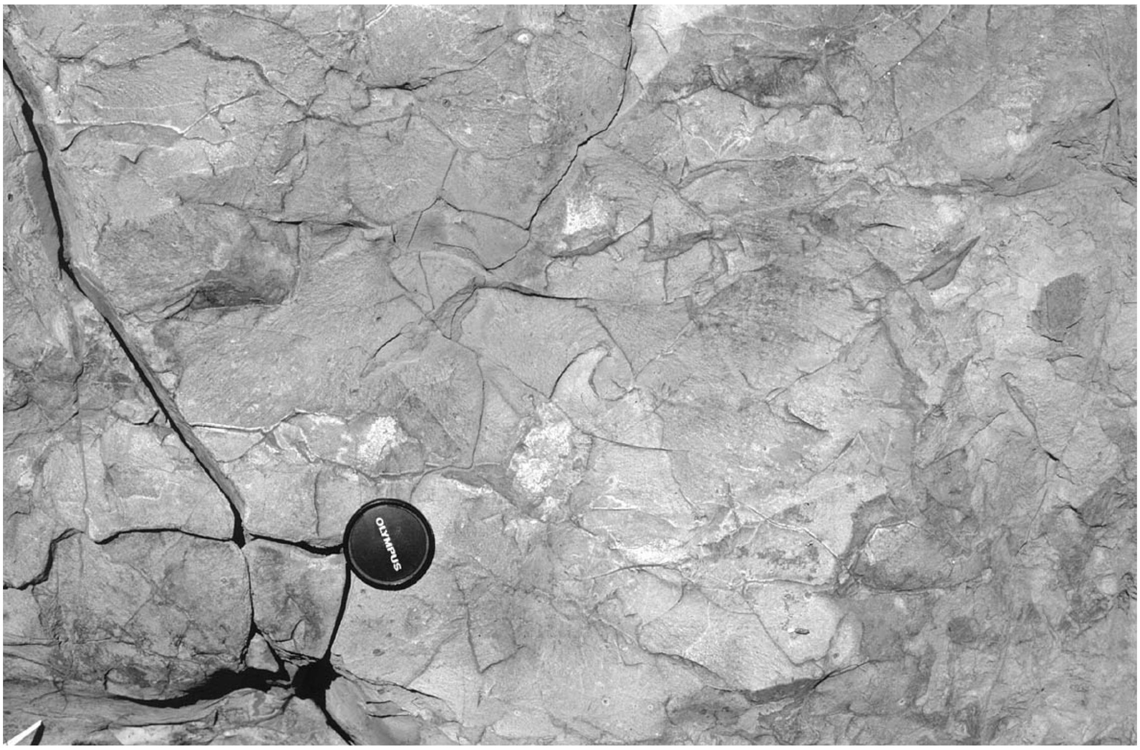

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Figure 4.3. Network of planar to curved 'tortoiseshell' joint surfaces in the lower part of a thick massive to amygdaloidal Mount Roe Basalt flow, Bellary Dome, south Pilbara sub-basin. The 3D intersection of these joints gives the rock an appearance which ranges from irregular net-veining to a pseudobreccia. Although the example shown comes from a subaerial flow, this style of jointing is most abundant in massive subaqueous flows of the Boongal and Bunjinah Formations

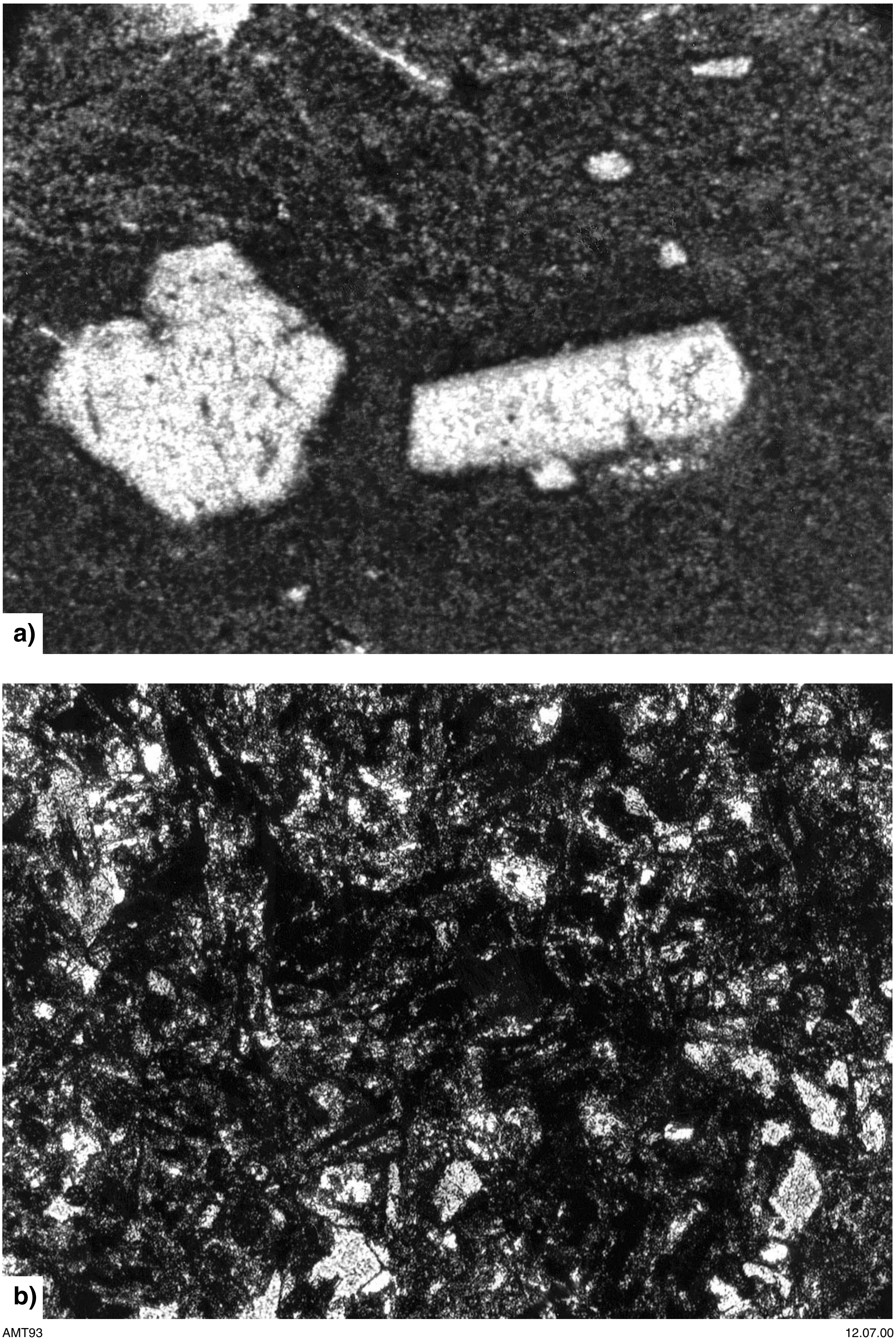

Figure 4.4. a) Photomicrograph of plagioclase-phyric basalt, from the Mount Roe Basalt. Plane polarized light, field of view = 2.5 mm; b) Photomicrograph of aphyric basalt from the Mount Roe Basalt. Cross polarized light, field of view = 2.5 mm

Although there is substantial compositional spread both between individual analyses of these ten groups (Fig. 12.2) and between the group means (Table 12.2), the centroid of the individual analyses approximates to the basaltic andesitic composition (in terms of the silica/total alkalies scheme of Le Maitre, 1989) which characterizes the bulk of the stacked mafic lavas of the Fortescue Group (Fig. 12.6). The mean silica content of all Mount Roe Basalt groups is close to 57%, significantly higher than typical Phanerozoic flood basalts; individual analyses extend across the calc-alkaline and tholeiitic fields of the Jensen Cation Plot (Jensen, 1976; Fig. 12.7). Trace elements have the enrichment relative to MORB that is characteristic of the later Fortescue Group mafic lavas (Fig. 12.9), and which is a common feature of within-plate continental flood basalts. In terms of the Ti–Zr–Y plot of Pearce and Cann (1973; see Fig. 12.11), however, they fall mainly in the calc-alkali field (C) rather than the within-plate field (D). There is no clearly discernible regional variation in composition of lavas of the Mount Roe Basalt.

Interpretation: The basaltic composition and aphyric texture of most of the lavas suggest that they originated as low- to moderate-viscosity flows that were close to their liquidus temperatures at eruption (cf. Green, 1989). In addition, the general lack of evidence for contemporaneous explosive fragmentation suggests either that the initial content of exsolved volatiles was low, or that levels were high but were reduced by non-explosive degassing or by volatile loss during a previous phase of explosive activity (Cas and Wright, 1987).

Despite their narrow compositional range, the flows show considerable variation in thickness and shape, and in their surface and internal structure. These differences reflect local and regional variations in the environmental setting and post-extrusion history of the flows.

Very thick (>40 m), columnar jointed, massive to amygdaloidal flows are similar in thickness, internal structure, surface features, and petrographic character to many flows in the Columbia River Basalt (Waters, 1961; Long and Wood, 1986). Their irregular columnar structure, lacking entablature, most closely resembles the Type 1 flows of Long and Wood (1986) although their thickness (up to 72 m) is greater, and column diameter (<0.3 m) is smaller than the Columbia River examples. By analogy with these North American lavas, the columnar jointed massive to amydaloidal flows of the Mount Roe Basalt are interpreted as rapidly emplaced subaerial, ponded lava flows. Other massive to amygdaloidal flows in the Mount Roe Basalt are commonly very thick (up to 82 m) but do not show evidence of columnar jointing. It is likely that these very thick basalts also formed ponded flows but did not develop the necessary thermal stress pattern for joint formation (Spry, 1962; Degraff and Aydin, 1987)

It is not known to what extent many of the thinner (2–20 m) massive to amygdaloidal lavas in the Mount Roe Basalt represent ponded flows. On the one hand, these basalts are thinner and have a more varied morphology than typical ponded flows, but on the other, they lack many of the features associated with flowing lava, such as lava tubes and surface channels.

Although the precise origin of pipe amygdales and amygdale cylinders is uncertain, they are both believed to result from the exsolution of magmatic gases and are abundant in modern pahoehoe lava (Walker, 1987; Philpotts and Lewis, 1987). Walker (1987) has observed that pipe vesicles in Hawaiian pahoehoe form only when the flow comes to rest on slopes of less than 4°. In both modern and ancient lavas, the tilt of inclined pipe amygdales and amygdale cylinders is thought to record the final direction of lava movement. Data from the Mount Roe Basalt are too few to permit a reliable interpretation of palaeoflow directions.

Basaltic pillow lavas

Description: Basaltic pillow lava (Fig. 4.5a,b) is estimated to form less than 2% of the Mount Roe Basalt. Most occurrences are reported from, or near, the base of the succession in the northwest Pilbara, Marble Bar and northeast Pilbara sub-basins. In the majority of outcrops the lavas are intimately associated with hyaloclastite (described below) and together they form laterally discontinuous units which locally may exceed 50 m in thickness.

Individual pillows are up to 1.8 m wide and have a sac-shaped or elongate, finger-like morphology (Fig. 4.5a,b). Many contain spheroidal, arcuate, or planar convex cavities up to 0.3 m wide and 0.1 m high; in addition, radial pipe-amygdales are observed locally. Pillows have an altered glassy selvage up to 10 mm thick and interpillow cavities are filled by variable amounts of quartz, carbonate, and hyaloclastite.

No chemical analyses are available of lavas of this facies.

Interpretation: The formation of pillow lava and associated hyaloclastite clearly indicates that a small part of the Mount Roe Basalt was either erupted subaqueously or flowed into water. Of the three principal settings for pillow lava formation, marine, sub-glacial, and lacustrine, the first two can be ruled out for the following reasons:

- The Mount Roe Basalt is underlain by subaerial or lacustrine sedimentary facies and there is no evidence of an intervening marine transgression.
- The overwhelming mass of Mount Roe Basalt comprises subaerial basaltic lavas; pillow lava occurs only in laterally impersistent units at, or near the base of, the formation.
- The top of the basalt pile was eroded subaerially and overlain by continental sedimentary and volcanic facies.
- No other glacial facies rocks have been reported from the Fortescue Group.

The limited lateral extent, thickness, and distribution of the pillow lava and hyaloclastite, coupled with their association with subaerial flows, suggest that they were deposited in small lakes. Very thick accumulations of hyaloclastite and pillow lava, which contain little other interbedded sedimentary rock, may reflect the presence of deep lakes of limited lateral extent. Alternatively, they

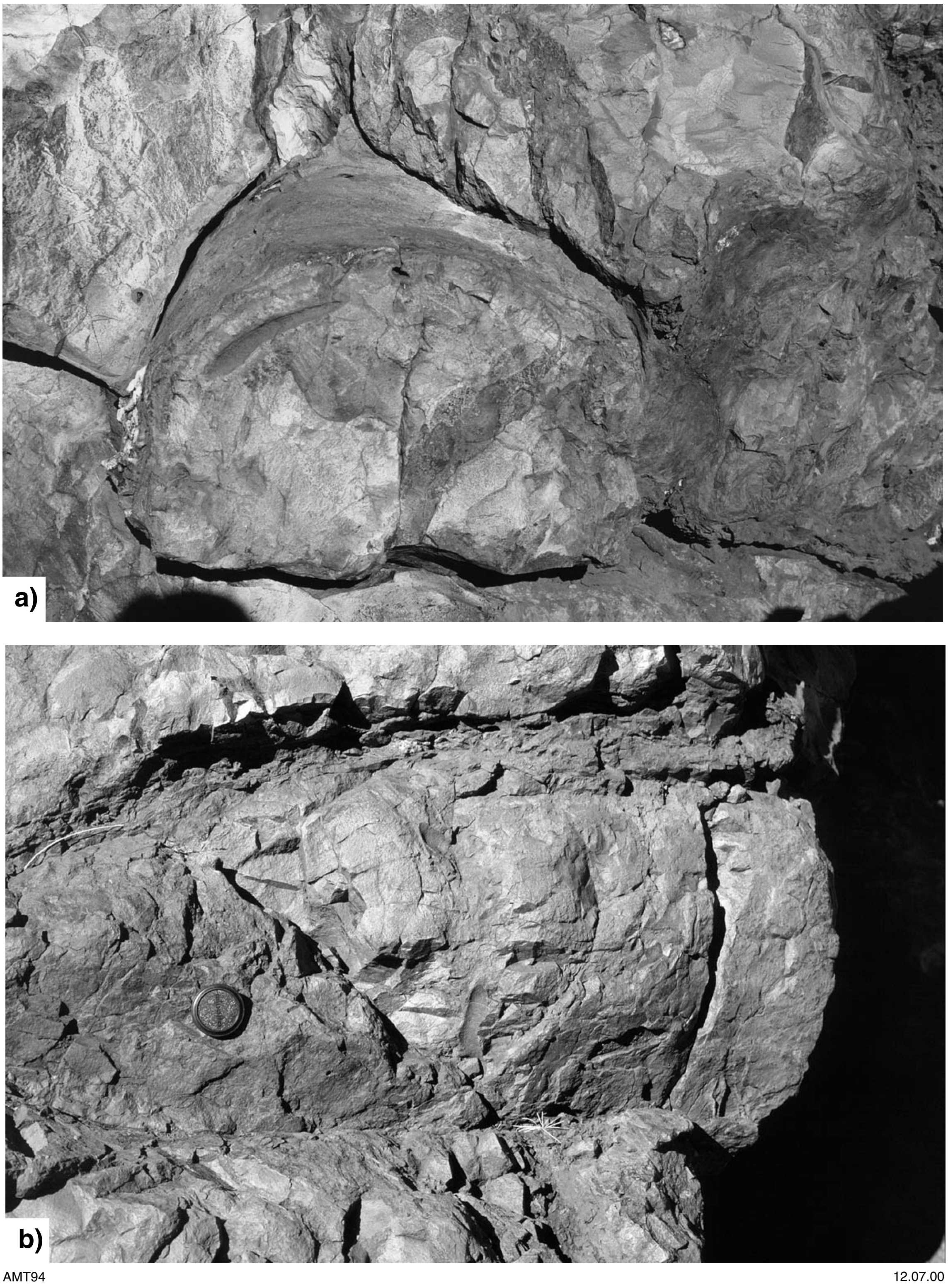

Figure 4.5. a) Cross section of tubular pillow lava, Mount Roe Basalt, Whim Creek, northwest Pilbara sub-basin. Pillow width is 0.65 m; b) Longitudinal section of tubular pillow lava from same outcrop as Figure 4.6a. Although the pillow long axes show a consistent orientation at this locality, the direction of budding, and hence lava flow, is unknown

may have originated in areas which had a history of flooding and rapid, possibly fault-controlled, subsidence. By analogy with similar rocks in the Columbia River Basalt, some of the lakes may have formed as a result of lava flows damming and disrupting the local drainage system (Swanson and Wright, 1978).

Volcaniclastic facies

Description: Volcaniclastic rocks are estimated to form less than 5% of the Mount Roe Basalt. They generally consist of hyaloclastite, and mafic and felsic lapilli tuff, although mafic agglomerate has been reported locally (Blight, 1985). Most volcaniclastic rocks lie at, or near, the base of the formation.

Hyaloclastite is commonly associated with pillow lava and occurs in laterally discontinuous units, generally ranging from a few metres to some 300 m thick. Most hyaloclastites are massive, very poorly sorted breccias composed of angular fragments of sparsely vesicular basalt up to 0.5 m across. Weak cross-stratification is sometimes observed in thin beds of fine-grained hyaloclastite. Clasts of all sizes are cut by a network of polygonal fractures, which vary in width from hair-like cracks to wider veins filled by quartz, chlorite, and carbonate. Even when separated by a few centimetres, many clasts exhibit a jigsaw fit with neighbouring fragments.

Thin sheets and lenses of mafic, intermediate, and felsic lapilli tuff occur locally at the base of the Mount Roe Basalt and higher up the succession between subaerial flows. Tuff units range in thickness from less than one metre to over 60 m (mostly 1–10 m), and whereas many are laterally impersistent, the thicker units may be traced over a distance of up to 4.5 km. Individual beds may be normally graded (Fig. 4.6a) or reverse graded, and the tuff unit as a whole commonly displays an upward decrease in grain size. Internal structure consists of parallel-stratification or trough cross-stratification (Fig. 4.6b), and ripple cross-lamination is observed in finer grained beds.

Most of the mafic varieties fall into the category of lithic vitric tuff and lapilli tuff and consist of non-welded vitric fragments and subordinate basalt, tuff peloids, quartz, plagioclase, and argillite set in a chloritic groundmass. Vitric fragments, now replaced by chlorite, are irregularly elongate to equant and often display scalloped margins. Many contain sparse to moderately abundant spherical vesicles and rare coalesced vesicles.

Felsic and intermediate lapilli tuff contain anhedral to subhedral ß-quartz, albitized plagioclase, K-feldspar, felsic lithic fragments, and devitrified shards set in a matrix of fine-grained quartz, feldspar, sericite and chloritized biotite. Many rocks contain whole or broken accretionary lapilli up to 12 mm in diameter. These have an outer margin of very fine (>40 µm) tuff, a middle zone of concentric fine (40–100 µm) tuff bands, and a core with the same texture and composition as the interlapilli matrix.

Interpretation: The basaltic breccias are interpreted as hyaloclastites, produced by the rapid cooling and fracturing of basalt during contact with water. This interpretation is based upon the following evidence:

- Their homogeneous, non-vesicular clast composition.
- The predominance of poorly sorted, angular clasts.
- Their intimate association with pillow lava.

As with the pillow lavas, the geological setting of the hyaloclastite provides evidence that they were formed when subaerially erupted basalt flows entered lakes or rivers.

Mafic vitric and lithic vitric tuff are interpreted as the products of hydroclastic basaltic eruptions. The low vesicle content of many vitric grains suggests that while the magma was partly vesiculated, expansion of magmatic gases alone was probably insufficient to produce explosive fragmentation. Instead, the broken, scalloped margins of many vitric grains most closely matches grain textures caused by steam-explosion fracturing resulting from the interaction of basaltic magma and external water (Wohletz, 1983). The occurrence of small- to medium-scale trough cross-stratification, ripple cross-lamination, and abundant non-volcanogenic detritus in many mafic tuffs suggests that these have undergone some reworking and are not primary eruptive deposits. Although there is strong evidence for resedimentation, the majority of fragments show no evidence of derivation from pre-existing volcanic rocks; only a few basalt clasts and tuff peloids are likely to be epiclastic fragments.

Felsic tuffs are interpreted as pyroclastic airfall deposits on the basis of the widespread occurrence of concentrically zoned accretionary lapilli, presence of normal and reverse grading, lack of erosional structures, and (less reliably) absence of welding. Cross-stratified and ripple cross-laminated tuff containing a mixed population of whole and broken accretionary lapilli are probably reworked airfall deposits. The most likely alternative explanation, that they are primary pyroclastic surge deposits, is discounted due to the absence of upper flow regime structures, climbing dune forms, and topographic mantling (cf. Fischer and Schmincke, 1984).

Non-volcanogenic sedimentary rocks

Localized accumulations of non-volcanogenic sedimentary rocks are interbedded with the basalt flows in the northwest and south Pilbara sub-basins. In most areas these deposits are thin and laterally impersistent, and are estimated to form considerably less than 1% of the Mount Roe Basalt stratigraphy. Only in the southern Wyloo Dome, in the south Pilbara sub-basin, are significant thicknesses recorded (Seymour et al., 1988).

One inter-flow unit, a 2 m-thick pebbly quartz sandstone exposed at Mount Roe, was examined during the course of this study. The sandstone forms an upward-coarsening sheet-like body which is laterally persistent over 500 m. Internal structure is dominated by stacked sets of small- to medium-scale trough cross-strata and sub-planar parallel-stratification. Pebbles consist of rounded to angular fragments of felsic volcanic rock. The sandstone unit is interpreted as a braided sheetwash deposit.

Published descriptions of other interflow sedimentary rocks are sketchy but suggest that most consist of basement-derived conglomerate, pebbly sandstone, and sandstone together with argillite and dolomite (Ryan, 1966; Hickman et al., 1983; Seymour et al., 1988).

The thickest recorded occurrences of interflow sedimentary rocks are on WYLOO (Blight, 1985; plate 1), where a number of lenticular conglomerate and sandstone bodies outcrop in the south and west central part of the Wyloo Dome. The most extensive of these, a 400 m-thick conglomeratic unit, contains rounded cobbles and boulders of basalt (50%), granitoid (30%), chert (10%), and felsic volcanic rock, vein quartz, and phyllite (10%) in a foliated sandy matrix. The conglomerate appears to pass laterally into arkosic sandstone and argillite, although outcrop is incomplete and the intense foliation makes it difficult to determine original lithology.

Palaeosols

A weathering profile has developed over flows of the Mount Roe Basalt southeast of Whim Creek, on ROEBOURNE (Macfarlane and Holland, 1990). Detailed work on this and a second profile nearby was later published by Macfarlane and Holland (1991) and by Macfarlane et al. (1994a,b), who designated them the Mount Roe #1 and Mount Roe #2 palaeosols. The following account is summarized mainly from those papers.

Field relationships

The Mount Roe #1 palaeosol is exposed for more than 2 km along strike, but is best exposed to the east of Whim Creek (SHERLOCK AMG 933904), where the succession dips west-southwest at about 30°. The uppermost part consists of about 1–10 m of very fine grained, grey to cream or white, massive, sericitic, conchoidally fractured rock. This top layer grades down into a dark green, chloritic, Fe-rich zone, generally 2–4 m thick which, in turn, grades down into unweathered basalt. Relict vesicles and veinlets throughout the upper part of the sericite zone are clearly inherited from textures of the basalt flow-top protolith. The contact between the sericite and chlorite zones is irregular, and pods of chlorite-rich material several metres across persist within the lower part of the sericite zone.

The palaeosol at this locality is sharply overlain by 2–6 m of thin-bedded to laminated sandstone and pebbly sandstone, grading upward into siltstone; local pockets of palaeosol material bulging upwards into the overlying sediments are thought to be associated with burial and diagenesis. Mud-cracks and ripple marks within the sediments clearly indicate shallow-water deposition. Small rip-up clasts of palaeosol material in the lowermost coarse sands are consistent with this. Thin, discontinuous, calcareous horizons within the sediments may reflect evaporation within a shallow, volcanically dammed lake.

The Mount Roe #2 palaeosol, which lies 4 km northeast of the bettter-exposed Mount Roe #1 locality and an unknown thickness stratigraphically below it, is closely similar. One significant difference is that there are no sediments intervening between the Mount Roe #2 palaeosol and the overlying basalt, which appears to have produced some baking and metasomatism in the upper few decimetres. The sericite zone locally exceeds 15 m in thickness, and its gradational contact with the underlying thin chlorite zone is more irregular.

Petrography and geochemistry

In thin section the texture of the lower part of the palaeosol is gradational to the upper part, and appears to reflect progressively greater diagenetic compaction of the weathered material from the bottom to the top. Many samples from the sericite-rich upper zone show well-preserved textures mimetic (although distorted by compaction) of those in the unaltered basalts. In addition to the predominant fine-grained muscovite and chlorite of the sericite and chlorite zones respectively, minor quartz is present throughout. Rutile, apatite, Al, Ca–Al, and REE phosphates have been identified in the sericite zone (S. Sutton *in* Macfarlane et al., 1994a); some kaolinite is also present in the lower part of the zone.

From extensive major oxide and trace element analyses of samples collected systematically through a number of profiles of both palaeosols, Macfarlane et al. (1994a) divided the elements into three groups, characterized by differing chemical behaviour during weathering: those which remained immobile (Group 1), those which were partly or almost completely leached out (Group 2), and those whose present distribution appears to be controlled by processes other than weathering (Group 3). Group 1 elements (Zr, Ti, Al, Th, Nb) increase upwards until about the middle of the sericite zone, and then remain constant. Pairs of these elements plot on straight lines, consistent with their essential immobility during weathering, and enrichment by removal of other elements. The fact that unaltered basalt samples fall on the same lines confirms the field evidence for derivation of the palaeosols by chemical weathering of the basalts over which they lie. Chromium, V and U follow the same trends, although somewhat less perfectly.

Group 2 elements almost entirely removed during weathering include Fe, Mn, Mg, and Zn. Lead, Na, and Ni were also depleted, but to a lesser extent. Calcium behaves somewhat erratically, possibly due to late metamorphism. The most striking Group 3 element is K, whose concentration in the sericite zone (K_2O>10%) is too great to be explained as the result of residual enrichment. Rubidium, Na, and Sr also appear to have been added by some process other than weathering. On the basis of a Rb–Sr isochron of c. 2168 ± 9.7 Ma, Macfarlane and Holland (1990) suggested that enrichment in alkalis took place during regional metamorphism (long after the palaeoweathering) whose age is constrained between c. 2765 and c. 2715 Ma (Arndt et al., 1991).

Macfarlane et al. (1994a) suggested a maximum value of 0.016 for P_{O_2}/P_{CO_2} in the atmosphere during the

a)

Figure 4.6. a) Thin beds of graded mafic volcanic sandstone, Mount Roe Basalt, Wyloo Dome, south Pilbara sub-basin; b) Cross-stratified volcanic sandstone containing whole and broken fragments of accretionary lapilli, Mount Roe Basalt, Bellary Dome, south Pilbara sub-basin. Photographs by D. F. Blight

b)

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formation of the Mount Roe palaeosols, using arguments based on the relative mobility of Fe, Ca, Mg, Na, K, and Mn during weathering (Pinto and Holland, 1988; Holland, 1992). They then combined this maximum value with Kasting's (1992) estimate of P_{CO_2} in the early atmosphere to arrive at a value of 1% present atmospheric level for the upper limit of P_{O_2} at the time of palaeosol formation.

Macfarlane et al. (1994b) reported extensive REE and Sm–Nd isotopic analyses from the same suite of palaeosol and basalt samples. They found that heavy REEs were relatively immobile during weathering, and behaved essentially as Group 1 elements. Light REEs, by contrast, showed substantial relative mobilization, and enrichment in the central parts of both palaeosols. Macfarlane et al. (1994b) concluded, from a Sm–Nd reference isochron of 2151 ± 360 Ma, that substantial mobilization of REEs took place during late metamorphism and metasomatism as well as during weathering. They also concluded, from the absence of Ce anomalies, that there was insufficient oxygen in the soil profile to convert a significant amount of Ce^{3+} to Ce^{4+}, qualitatively confirming the 'very low' P_{O_2} argued from the major-element chemistry.

Their Sm–Nd isotopic analyses of unweathered basalt samples revealed ξNd values of between -4.0 and -7.4, consistent with the findings of Nelson, D. R. et al. (1992).

Other possible palaeosol occurrences

Three other discontinuous horizons of white weathered rock interbedded with flows of the Mount Roe Basalt occur in the outlier southeast of Whim Creek. A brief examination of the lowermost horizon, in the northeast part of the outlier (SHERLOCK AMG 962946), suggests that it is similar to the Mount Roe #1 and #2 palaeosols. The rock is white, pseudostratified, and displays a prominent conchoidal fracture. No overlying sedimentary unit was observed.

Distribution of facies

The Mount Roe Basalt exists in discontinuous outcrops throughout the Pilbara, but is absent from the Sylvania Inlier. Principal outcrop areas are shown in Plates 1a and 1b; this information is summarized in Figure 4.1.

Northwest Pilbara sub-basin

On DAMPIER AND BARROW ISLAND, major outcrops correlated with the Mount Roe Basalt are confined to the area between Mount Wilkie and Mount Leopold, East Lewis, West Lewis, and Enderby Islands in the Dampier Archipelago, and to a small linear outcrop extending southwest from Mount Sholl.

Five kilometres west-southwest of Mount Leopold, the Mount Roe Basalt has a minimum thickness of 400 m and comprises thin to very thick, massive to amygdaloidal subaerial flows, which locally display both columnar jointing and brecciated flow tops. The total thickness of the basalt stratigraphy in this area is probably less than 1000 m, although reliable estimates are hampered by the relief of the basement topography, variable dip, and post-Mount Roe Basalt faulting. Kriewaldt (1964a) estimated that the maximum thickness of the basalt pile at nearby Mount Wilkie was about 2400 m.

The basalt succession on the Dampier Archipelago was not examined during this study. Kriewaldt (1964a) records about 1200 m of basaltic lava interbedded with some agglomerate and acid igneous rock.

On ROEBOURNE, the major occurrences of Mount Roe Basalt are confined to three areas: Mount Anketel to Cape Lambert, Mount Roe to Warambie Homestead, and a chain of outcrops extending from southeast of Whim Creek to Peewah Hill.

At Mount Roe and Mount Anketel, the Mount Roe Basalt consists of 700 and 800 m respectively of thick, or very thick, massive to amygdaloidal lava flows. Although these localities are 25 km apart they have similar stratigraphies, particularly in the lowest 300 m of the succession, where several unusually thick flows are recorded. On this evidence, there is a possibility that some of the lavas may be correlatives. One thin, sheet-like body of interflow sandstone and conglomerate is present at Mount Roe but is not observed at Mount Anketel. Trough cross-strata preserved within this unit record a southerly directed palaeoflow.

The measured thickness of some 700 m for the Mount Roe Basalt at Mount Roe contrasts with the figure of 2.4 km quoted for this area by Kriewaldt (1964b). One consequence of this revised figure is that the thinning of the succession between Mount Roe and southwest of Pinanular Pool, where approximately 220 m of basalt is recorded, is less marked than previously believed. Such variation could be the result of localized penecontemporaneous subsidence as originally suggested by Kriewaldt (1964b), but it could also reflect a combination of variable basement topography and post-Mount Roe Basalt – pre-Hardey Formation erosion. The latter explanation probably accounts for the presence of no more than 200 m of Mount Roe Basalt in the area south of Mount Ada; here, the local palaeotopography is estimated to be at least 100–150 m and basalt clasts are abundant in the basal conglomerate of the Hardey Formation.

Southeast of Whim Creek (SHERLOCK AMG 953877), the basal unit of the Mount Roe Basalt is exposed on the plain immediately east of the main outcrop. This basal unit comprises subaqueous pillow lava and minor hyaloclastite; pillow long axes are consistently aligned toward 030–080°. The main body of the inlier is made up of an estimated 1 km of subaerial basalt flows interbedded with sandstone and argillite. These rocks are especially abundant in the upper part of the succession, exposed at the southwestern corner of the inlier (Ryan et al., 1965). Macfarlane and Holland (1990) have documented the occurrence of a palaeosol from the middle part of the Mount Roe Basalt near Whim Creek (SHERLOCK AMG 933904). This feature has been described earlier (see **Palaeosols**).

On PYRAMID, the Mount Roe Basalt is dominated by subaerial basalt flows. These are exposed in discontinuous outcrops between Bluff Hill and Nunyerry, and in a large fault-bounded outcrop between Nunyerry and the area northeast of Teichman Well.

Marble Bar sub-basin

Here, the Mount Roe Basalt has a maximum thickness of about 2500 m (Blake, 1993) and consists largely of subaerial flows, underlain locally by subaqueous pillow lava and hyaloclastite. Hickman and Lipple (1978) also recorded columnar lava in this succession.

Northeast Pilbara sub-basin

The thickest succession of Mount Roe Basalt in the northeast Pilbara sub-basin outcrops in the Meentheena area, where Blake (1984a) recorded up to 1600 m of predominantly subaerial lavas above the granite–greenstone basement. There are other outcrops in the Nullagine and Bamboo Creek districts, where about 500 m of lava is recorded (Blake, 1984a). Williams (1999) reported up to 700 m of Mount Roe Basalt on MUCCAN but noted that the unit is absent from the keel area of the Oakover Syncline. Coarse-grained basaltic agglomerate, tuff, and basalt–dacite breccia are present at the base of the formation on the western side of the syncline (Williams, 1999).

On BALFOUR DOWNS, Williams (1989) recorded less than 350 m of subaerial lavas along the eastern margin of the Rooney Inlier.

South Pilbara sub-basin

The Mount Roe Basalt has been recorded from the Wyloo Dome (Horwitz, 1978; Seymour et al., 1988), the Bellary Dome and the southeastern flank of the Rocklea Dome (Blight, 1985, Thorne and Tyler, 1996, 1997a), and the SGS–1, BMW–1, and BMW–2 diamond drillholes, located in north-central and south-central MOUNT BRUCE and northern TUREE CREEK respectively (CRA Exploration Pty Ltd, 1988, 1989).

Up to 1500 m of predominantly subaerial basalt flows and subordinate felsic tuff and sedimentary rock outcrop on the northeastern margin of Wyloo Dome. On the southern limb the succession is apparently thinner, although reliable thickness estimates are hampered by the stronger deformation in this area. The thickest recorded occurrence (>400 m) of interflow sedimentary rocks within the Mount Roe Basalt outcrops on this southern limb, in the area southeast of Big Metawandy Well (Blight, 1985, plate 1).

Tuffaceous interflow sedimentary facies constitute a minor part of the Mount Roe Basalt stratigraphy in the Bellary Dome. Here, the succession ranges in thickness from about 700 to 1000 m and consists largely of subaerial flows with hyaloclastite developed locally near the base. A much thinner succession (~100 m), consisting entirely of subaerial flows, is preserved on the southeastern flank of Rocklea Dome. Geochemical comparisons suggest that these lavas correlate with similar flows in the upper part of the Mount Roe Basalt at Bellary Dome.

The only other major occurrence of Mount Roe Basalt in the south Pilbara sub-basin is reported from the SGS–1 diamond drillhole (CRA Exploration Pty Ltd, 1988). Here the succession is 1050 m thick and comprises massive to amygdaloidal (?subaerial) lava and minor tuff. The total thickness of Mount Roe Basalt below the southern Turner Syncline is unknown because drilling in the BMW–1 diamond drillhole was abandoned after passing through only 20 m of the succession (CRA Exploration Pty Ltd, 1989).

Results from the BMW–2 core are difficult to interpret. The hole was collared below the top of the Mount Roe Basalt and does not appear to have penetrated the complete Fortescue Group succession. In addition, the presence of significant interflow sedimentary units at the base and middle of the cored interval makes it difficult to locate the boundary between the Mount Roe Basalt and the underlying Bellary Formation.

Palaeogeographic reconstruction

Figure 4.7 summarizes the present-day thickness values for the Mount Roe Basalt; it also shows the inferred zero isopach for the Mount Roe Basalt, as indicated by the present outcrop pattern and CRAE drillhole data. As noted by Blake (1993), these data point to the presence of northern and southern basalt provinces, separated by a west-northwesterly trending area that is devoid of basalt. Hickman (1975a) and Hickman and Lipple (1975) provided a similar interpretation for the area north of the Fortescue Valley. The meaning of this distribution is uncertain and two end-member interpretations are possible. Firstly, the present regional pattern is primary and the central area was never covered by Mount Roe Basalt, and secondly, basalts covered the entire craton and were partially removed during regional uplift and erosion prior to, and during, deposition of the Hardey Formation (see **Chapter 5**).

The evidence currently available appears to favour the first explanation for the following reasons:

- Palaeogeographic reconstructions for the Hardey Formation, which are based on a considerable amount of thickness, palaeocurrent, and petrographic data, indicate that a major topographic high existed in the same area after deposition of the Mount Roe Basalt (Blake, 1993; see also **Chapter 5**).
- In the south Pilbara sub-basin there is a marked northward thinning of the basalt pile between the Bellary and Rocklea Domes. On the basis of their geochemistry, the Rocklea Dome basalts can only be correlated with the uppermost lavas of the Bellary Dome, implying a northward onlap onto a basement high.
- Although there is clear evidence for post-Mount Roe Basalt – pre-Hardey Formation erosion in many parts of the Pilbara, there is a marked contrast between the thickness of the preserved Mount Roe Basalt sections and the amount of basalt-derived sedimentary rocks in the Hardey Formation. If the basalt pile originally covered the

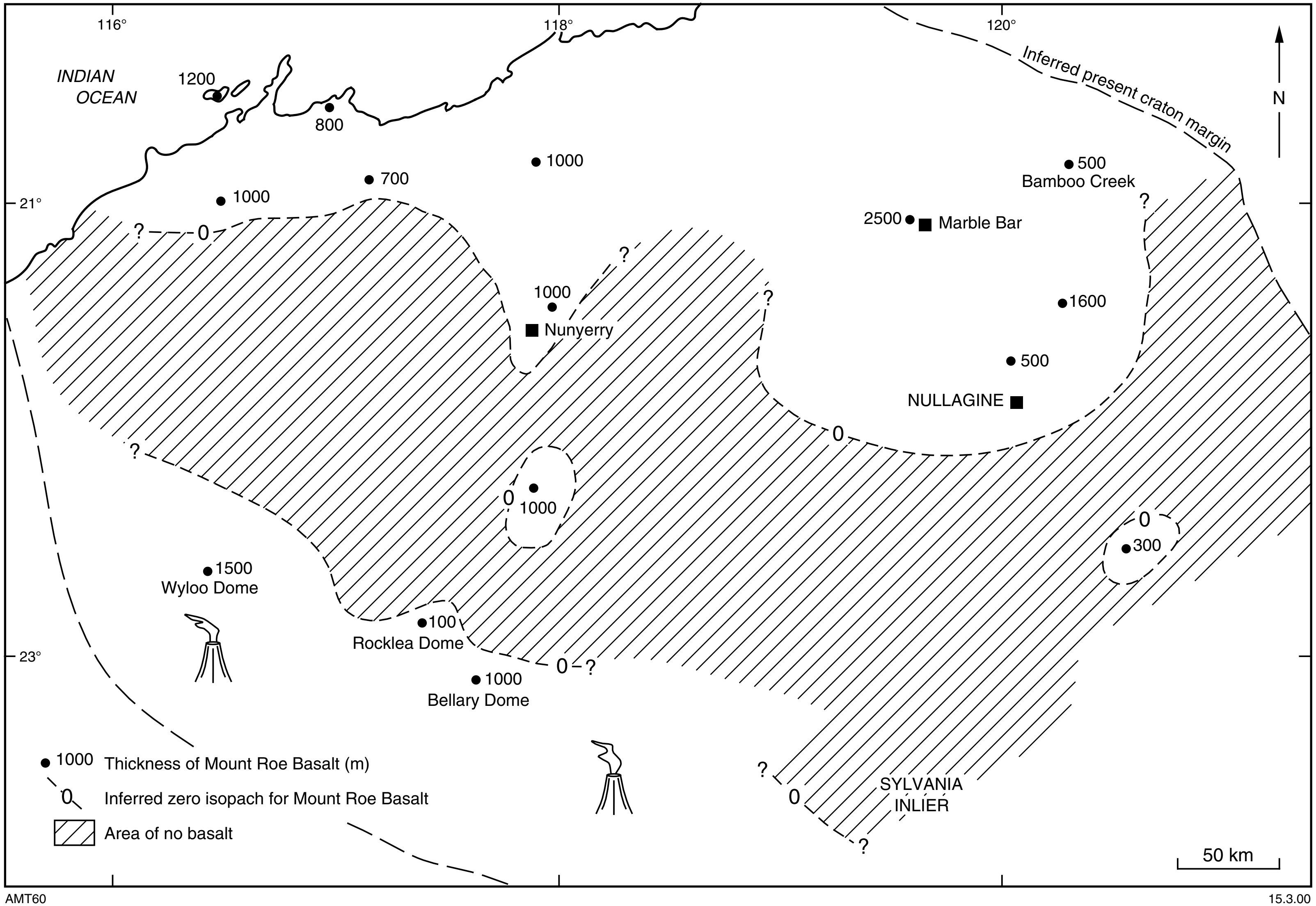


Figure 4.7 Palaeogeographic reconstruction for the Mount Roe Basalt showing present-day thickness variation. Thickness data from Nunyerry, Marble Bar, and the Nullagine to Bamboo Creek area are from Blake (1984a, 1993)

Pilbara and was eroded prior to deposition of the Hardey Formation, an enormous quantity (100 000 km^3) of basalt would have had to have been removed without trace.

To the north and south of the palaeohigh, Mount Roe Basalt volcanism was dominated by low-viscosity basalt floods with high effusion rates. The succession was probably built up of many hundreds of localized flows which originated mainly from fissures. To date, however, no feeder system has been positively identified, although geochronological evidence (Wingate 1994, 1997, 1999) suggests that intrusion of part of the north-northeasterly trending Black Range Suite was coeval with deposition of the lower Fortescue Group. Similarly, no eruptive vents for the localized tuff units within the Mount Roe Basalt have been identified. The greater abundance of these deposits in the Wyloo and Bellary Domes suggests the source volcano(es) was closest to these areas.

The volcanic style of the Mount Roe Basalt appears to fall between plains volcanism and Columbia River Basalt flood volcanism (cf. Waters, 1961; Greeley 1982; Long and Wood, 1986). The Mount Roe Basalt shares the following characteristics with plains basalts:

- the presence of pahoehoe flows,
- relatively thin (1–5 m) flows, and
- eruptions were probably fissure-fed.

The following differences are apparent, however. Firstly, no basaltic shield volcanoes or lava tubes have been identified in the Mount Roe Basalt. Secondly, most lava flows in the Mount Roe Basalt are thicker than 5 m and display limited relief on flow surfaces. In addition, cinder and tuff cones, and collapse structures have not been recorded from the Fortescue Group. On the other hand, the volcanic style of the Mount Roe Basalt differs from that of the Columbia River Basalt in having a greater proportion of thin flows of apparently limited lateral extent.

Chapter 5

Hardey Formation

The Hardey Formation (Thorne et al., 1991) is up to 3 km thick and is present throughout most of the Hamersley Basin, except for the western portion of the northeast Pilbara sub-basin (Fig. 5.1). Correlation of the Hardey Formation with the Fortescue Group sequence of the Gregory Range area is discussed in **Chapter 10**. In the Sylvania Inlier, the Hardey Formation is probably equivalent to the combined lower metasedimentary unit, lower mafic volcanic unit, and felsic pyroclastic unit of Tyler (1986).

In the northern Pilbara, the Hardey Formation unconformably overlies both the Mount Roe Basalt and the granite–greenstone basement. In the southern Pilbara, the formation unconformably overlies granite–greenstone rocks in the Milli Milli and western Rocklea Domes and Sylvania Inlier, and disconformably overlies the Mount Roe Basalt in the southeastern Rocklea Dome and in the SGS–1 borehole in northern MOUNT BRUCE. In the Bellary and Wyloo Domes, the contact with the Mount Roe Basalt appears to be conformable.

The nature of the boundary between the Hardey Formation and the overlying Kylena or Boongal Formations varies from place to place. In the northwest Pilbara sub-basin, it is apparently conformable, whereas in the northeast Pilbara sub-basin, the contact varies from an angular unconformity in the Nullagine area to a disconformity or non-depositional unconformity elsewhere (Blake, 1993). The upper contact of the Hardey Formation is apparently conformable in the south Pilbara sub-basin.

Uranium–lead zircon dates (**Chapter 2**) from the Hardey Formation (Bamboo Creek Member) of the northeast Pilbara sub-basin and the Gregory Range area (Koongaling Volcanic Member) lie within the range 2768 ± 16 to 2756 ± 8 Ma (Pidgeon, 1984; Arndt et al., 1991). Using the same method Arndt et al. (1991) obtained dates of 2763 ± 13 and 2775 ± 10 Ma respectively for the base and lower part of the underlying Mount Roe Basalt. No zircon data are available from the overlying Kylena or Boongal Formations.

Previous descriptions of the Hardey Formation are given by: Blight (1985), Daniels (1970), Hickman (1978, 1983), Hickman and Gibson (1982), Hickman and Lipple (1978), Hickman et al. (1983), Kriewaldt (1964a,b), Kriewaldt and Ryan (1967), Ryan (1966), Seymour et al. (1988), Thorne and Blake (1990), Thorne et al. (1991), Thorne and Tyler (1994, 1996, 1997a,b) and Williams (1989). The formation has also been described under various other formal and informal names (Blake, 1984a, 1993; Noldart and Wyatt, 1962; Trendall, 1975b, 1990b).

Post-Mount Roe Basalt – pre-Hardey Formation unconformity

In the northwest and northeast Pilbara sub-basins, the Mount Roe Basalt and basal sedimentary units of the Fortescue Group were faulted, tilted, and eroded before and during deposition of the Hardey Formation (Blake, 1993). This resulted in negligible or slight angular discordances between the Mount Roe Basalt and the Hardey Formation in these areas. In the Marble Bar sub-basin, a narrow, sinuous belt of gentle to tight folds with steeply dipping axial surfaces, and mostly gently plunging fold axes, formed after basaltic volcanism. This was in turn followed by faulting and erosion prior to deposition of the Hardey Formation (Blake, 1984a,b, 1993). This has resulted in up to 90° of discordance between Hardey Formation and Mount Roe Basalt bedding in this area.

The post-Mount Roe Basalt – pre-Hardey Formation history of the south Pilbara sub-basin is poorly known. Horwitz (1990) refers to the boundary between these units as an angular unconformity; however, the contact appears to be disconformable in the southeastern Rocklea Dome and no evidence of erosion or angular discordance was observed in the Bellary and Wyloo Domes.

Facies analysis

This section describes and interprets the principal sedimentary and volcanic facies associations present in the Hardey Formation. Most of the data have been obtained from eight measured sections and 1:40 000-scale mapping of selected outcrop areas. These data have been supplemented by descriptions from other workers, most notably Blake (1993).

The Hardey Formation comprises a diverse suite of non-volcanic sedimentary rocks and volcanic rocks. The former include clast- and matrix-supported conglomerate, feldspathic quartz sandstone and pebbly sandstone, and argillite. Volcanic rocks include felsic and mafic

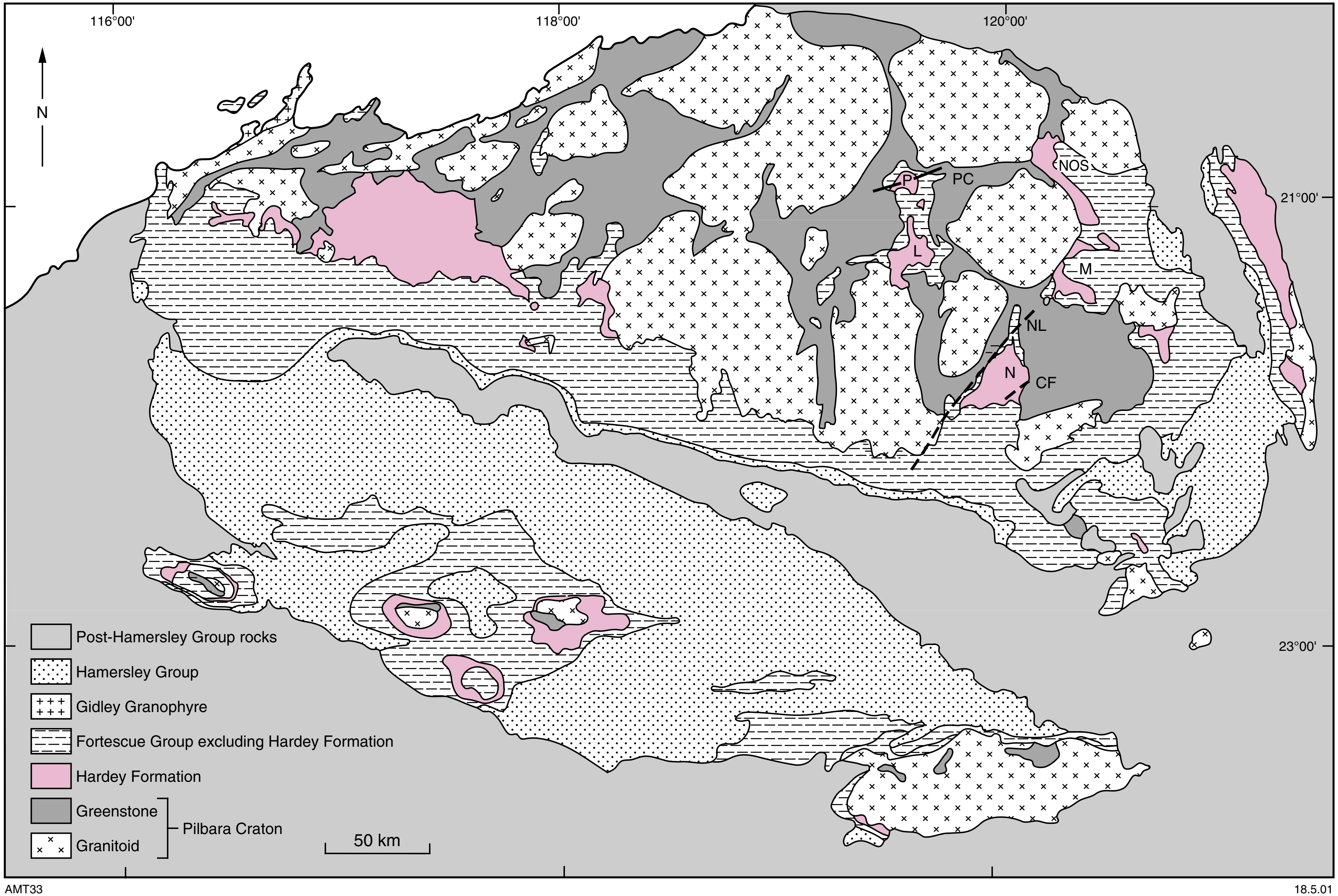


Figure 5.1. Principal outcrop areas of the Hardey Formation. P = Pear Creek Centrocline, PC = Pear Creek Fault Complex, L = Langwell Creek Centrocline, N = Nullagine Synclinorium, NL = Nullagine Lineament, CF = Conglomerate Creek Fault, M = Meentheena Centrocline, NOS = Northwest Oakover Syncline. Structural nomenclature after Blake (1993)

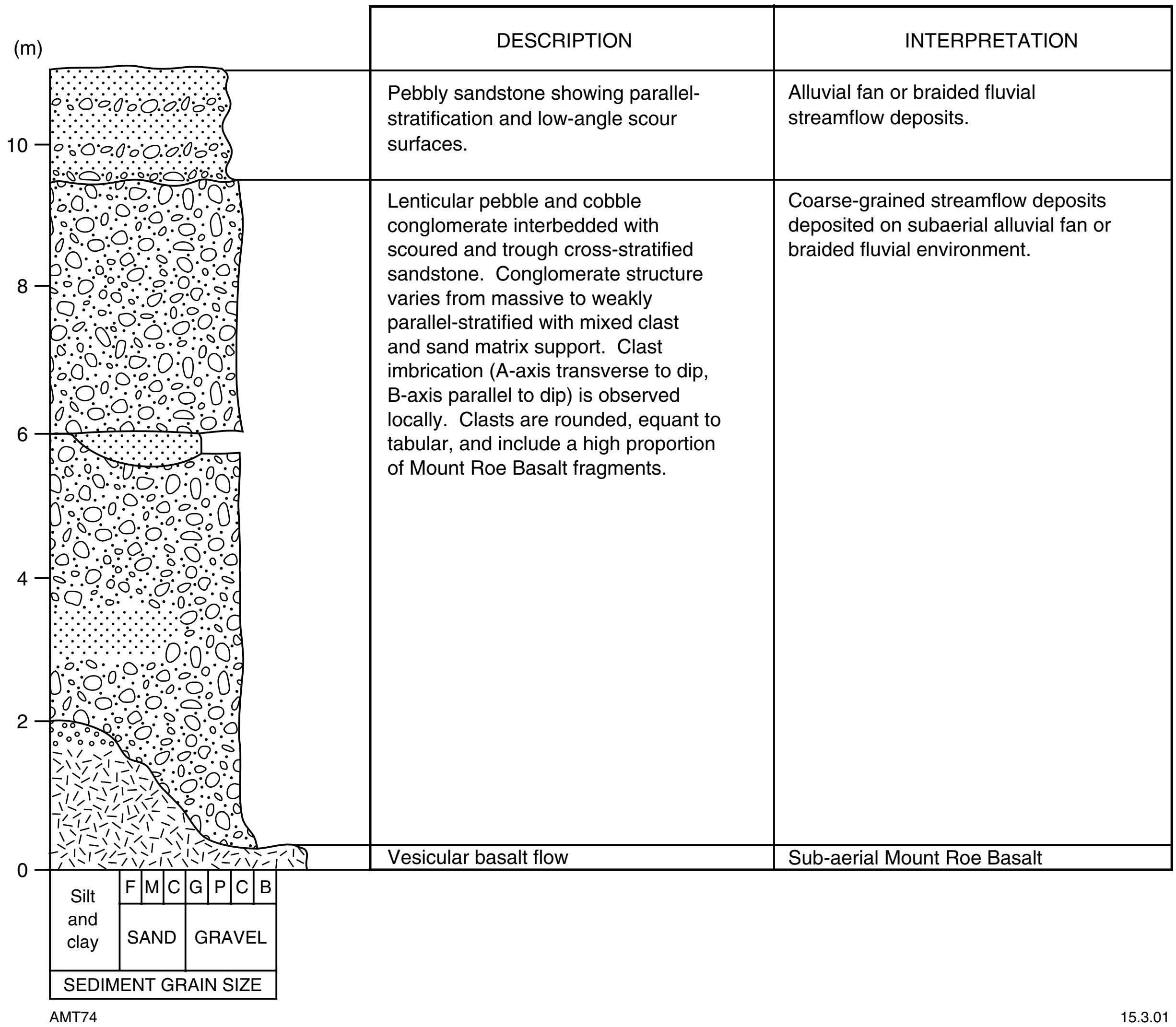


Figure 5.2. Representative vertical profile through alluvial fan and coarse-grained braided fluvial facies, showing variation in grain size, bed thickness, and internal structure, northwest Pilbara Sub-basin. An explanation of graphic log symbols is given in Figure 1.4

volcaniclastic deposits, basaltic flows, and local pillow lava. Dolerite and layered mafic sills form a significant part of the stratigraphy in the northwest Pilbara and south Pilbara sub-basins (**Chapter 11**).

Non-volcanic sedimentary rocks

The Hardey Formation incorporates a diverse range of palaeoenvironments and five major sedimentary facies are recognized: alluvial fan and coarse-grained braided alluvial; sandy braided fluvial; lacustrine; deltaic; and shoreline.

Alluvial fan and coarse-grained braided fluvial facies

Description: Alluvial fan and coarse-grained braided fluvial deposits consist of localized occurences of thick- to very thick bedded cobble and boulder conglomerate, interlayered with pebbly sandstone and coarse-grained sandstone (Fig. 5.2). These rock types are grouped into vertical successions ranging from 20 to several hundred metres thick.

Cobble and boulder conglomerates (Fig. 5.3) generally show either clast support or a combination of clast and matrix (sand and granule) support. Clast diameters are up to 1.5 m, but most lie in the range 0.05–0.5 m. In general, conglomeratic units as a whole show an upward-fining grain-size trend, whereas individual beds may be normally graded or non-graded.

Many conglomerates are either massive, or exhibit a weak bedding-parallel orientation of the tabular clasts. In others, there is a pronounced horizontal or inclined stratification produced by alternating layers of different grain size. Clasts vary in shape from platy to spherical and are angular to rounded; consistent clast imbrication (A-axis transverse to dip, B-axis parallel to dip) is observed

Figure 5.3. Thick-bedded cobble to boulder conglomerate containing abundant rounded fragments of Mount Roe Basalt. Hardey Formation alluvial fan and coarse-grained braided fluvial facies, northwest Pilbara sub-basin. Height of section is 5 m

locally (Fig. 5.4). Clast composition varies with local bedrock geology, although many conglomerates contain a high proportion of fragments derived from the Mount Roe Basalt.

Sandstone and pebbly sandstone forms lenses and laterally discontinuous sheets within the conglomerate units. Internal structure varies from massive to parallel-stratified or trough cross-stratified.

Interpretation: This facies is interpreted as alluvial fan and coarse-grained braided fluvial deposits on the basis of its geometry and localized occurrence, nature of internal structure compositional and textural immaturity, and character of associated facies. Alluvial fans were constructed by debris-flow and streamflow events; coarse-grained braided fluvial deposits were entirely water-lain.

Massive, very thick bedded conglomerates that show clast and/or matrix support, very poor sorting, and little evidence of clast abrasion are interpreted as debris-flow deposits. Lowe (1982) placed this type of deposit into a broad category of cohesive debris-flow sediments, but he noted that they were deposited in a flow in which the clasts were lubricated, but not fully supported or suspended, by the sandy matrix.

Clast-supported conglomerates that display parallel or inclined stratification, moderate sorting, normal grading, and clast imbrication are interpreted as streamflow deposits, similar to those reported from modern proximal braided rivers (reviewed by Miall, 1977, 1978; Collinson, 1978a,b; Cant, 1982). In this environment, the major sites of sediment accumulation are diffuse gravel sheets and complex braid bars within channels; both settings are characterized by gravel showing horizontal or gently inclined stratification. Where clast imbrication is developed, it is almost exclusively of the transverse A-axis – inclined B-axis variety, and is the result of clasts rolling along the river bed (Harms et al., 1982).

Thin sheets of structureless, coarse-grained and pebbly sandstone that are interbedded with streamflow conglomerates are regarded as suspension deposits, laid down during the later stages of flooding events. Trough cross-stratified sandstone and pebbly sandstone are the result of the downstream migration of small dunes under lower flow regime conditions; those rocks that exhibit parallel stratification were probably deposited from transitional to upper flow regime currents (Harms et al., 1982). In both cases these finer grained sediments were probably deposited during the waning stage of successive streamflow events.

Sandy braided fluvial facies

Description: The sandy braided fluvial facies forms units that range in thickness from a few tens of metres to some 400 m. These units are most commonly associated with a diverse range of facies including alluvial fan, coarse-grained and sandy braided fluvial, lacustrine, and deltaic. The facies comprises thin to very thick beds of medium- to very coarse grained sandstone, granule and pebbly sandstone, and pebble conglomerate. Sandstones are classified as arkose, lithic arkose, feldspathic volcanic arenite, or volcanic arenite, using the nomenclature of Folk (1974).

The internal structure of this facies is complex. In many outcrops it consists of stacked sets of trough cross-strata (or, less commonly, tabular cross-strata), 0.05–1.5 m high and up to 15 m wide, which together form irregular tabular to lenticular cosets 0.2–5 m thick (Fig. 5.5a,b). Some cosets show an upward decrease in grain size and scale of cross-stratification. Pebbly sandstone occurs as lags, or else is concentrated on individual foresets; thicker layers of pebble conglomerate commonly mark coset boundaries. Troughed cosets may also be separated by discontinous sheets of horizontal to gently inclined or undulatory stratified sandstone; intervening sets of small-scale planar tabular cross-stratification are also recorded. Primary current lineation is observed on many horizontally stratified sandstone bedding surfaces.

Palaeocurrent data from medium- to large-scale trough cross-strata are unidirectional at any stratigraphic level for a given locality and are generally characterized by a high

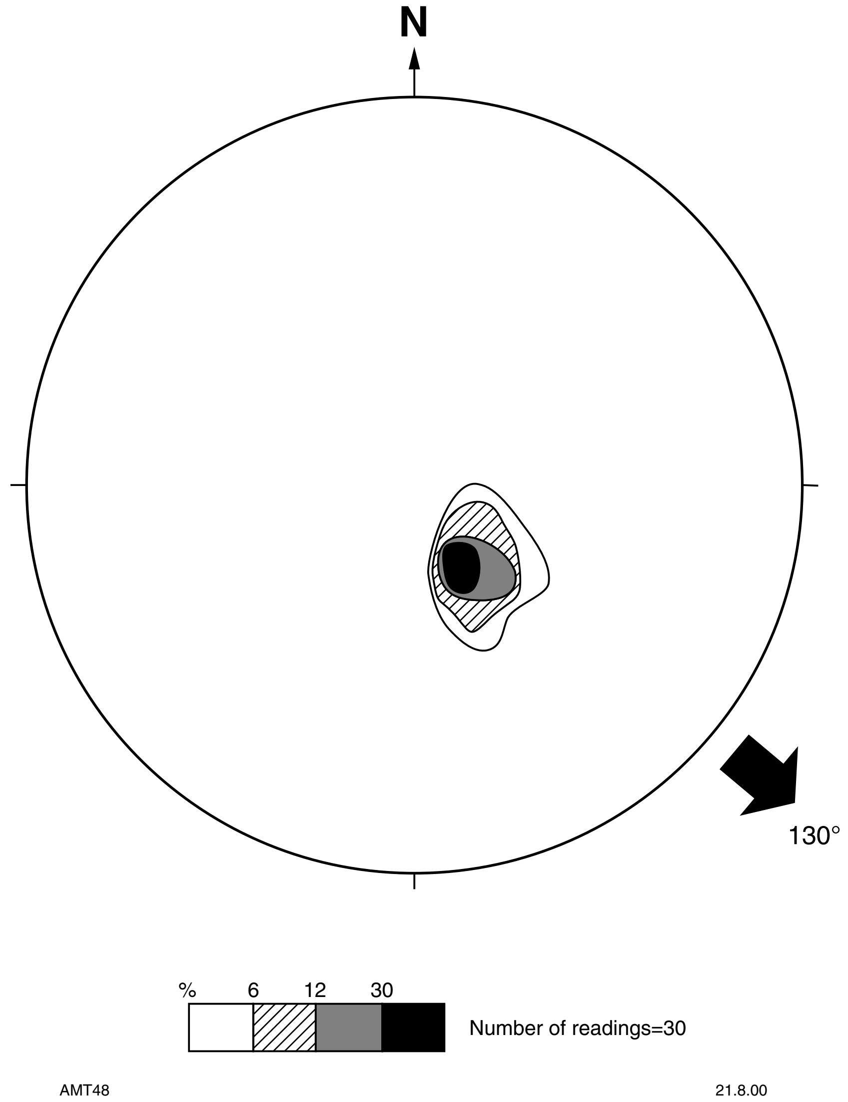


Figure 5.4. Contoured plot of cobble imbrication in Hardey Formation alluvial fan and coarse-grained braided fluvial facies, Pinanular Pool, northwest Pilbara sub-basin. Imbrication is defined by long (A) axis transverse to dip, intermediate (B) axis parallel to dip. Data are plotted as poles to the A–B plane. The inferred palaeocurrent trend of 130° is consistent with an easterly directed palaeoflow recorded from trough cross-stratified sandstone above the conglomerate. Number of readings = 30

(>0.8) vector mean magnitude (Fig. 5.6a–m). Palaeocurrent azimuths from small-scale troughs may differ from those of the larger scale structures by 50–90°.

Interpretation: This facies is interpreted as a braided fluvial deposit on the basis of its lithology, sheet- to ribbon-like geometry, diverse assemblage of traction current structures, and strongly unidirectional palaeoflow data.

Trough cross-stratification is formed by the downstream migration of lunate to sinuous dunes during moderate to high streamflow velocities (Harms et al., 1982). Tabular cross-stratification and the horizontal to undulatory lamination formed when flow velocities were lower and higher respectively (Harms et al., 1982).

Stacked cosets of structurally uniform medium- to very large scale trough cross-strata are interpreted as the deposits of major braided channels. Coset geometry and the unidirectional palaeocurrents with high vector mean magnitudes suggest channels were of low sinuosity, whereas the thickness of the troughed sets (<1.5 m) gives a minimum value for the channel depth.

Sandstone units that are characterized by a diverse range of sedimentary structures probably represent mixed channel and braid-bar deposits. By analogy with similar assemblages in the Devonian Battery Point Sandstone in Canada (Cant and Walker, 1976), the basal conglomerate marks the channel scour surface. The overlying upward-fining sandstone assemblage, with its upward-decreasing scale of trough cross-strata and interbedded planar–tabular

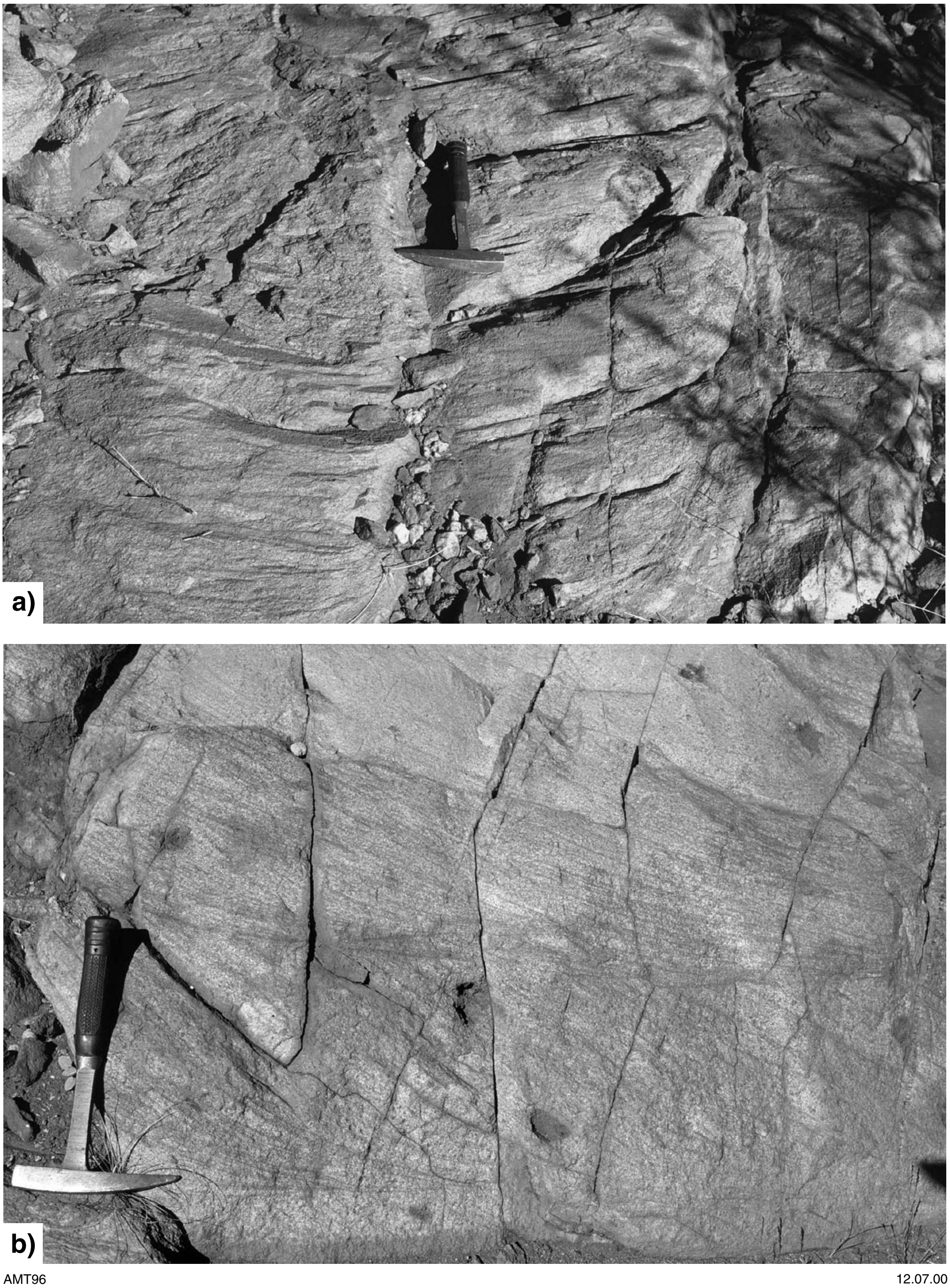


Figure 5.5. a) Oblique plan view of trough cross-stratification in Hardey Formation, sandy braided-fluvial facies, Bellary Dome, south Pilbara sub-basin. Palaeoflow direction is from bottom to top of the photograph; b) Vertical section trough cross-stratification in Hardey Formation, sandy braided-fluvial facies, Milli Milli Dome, south Pilbara sub-basin

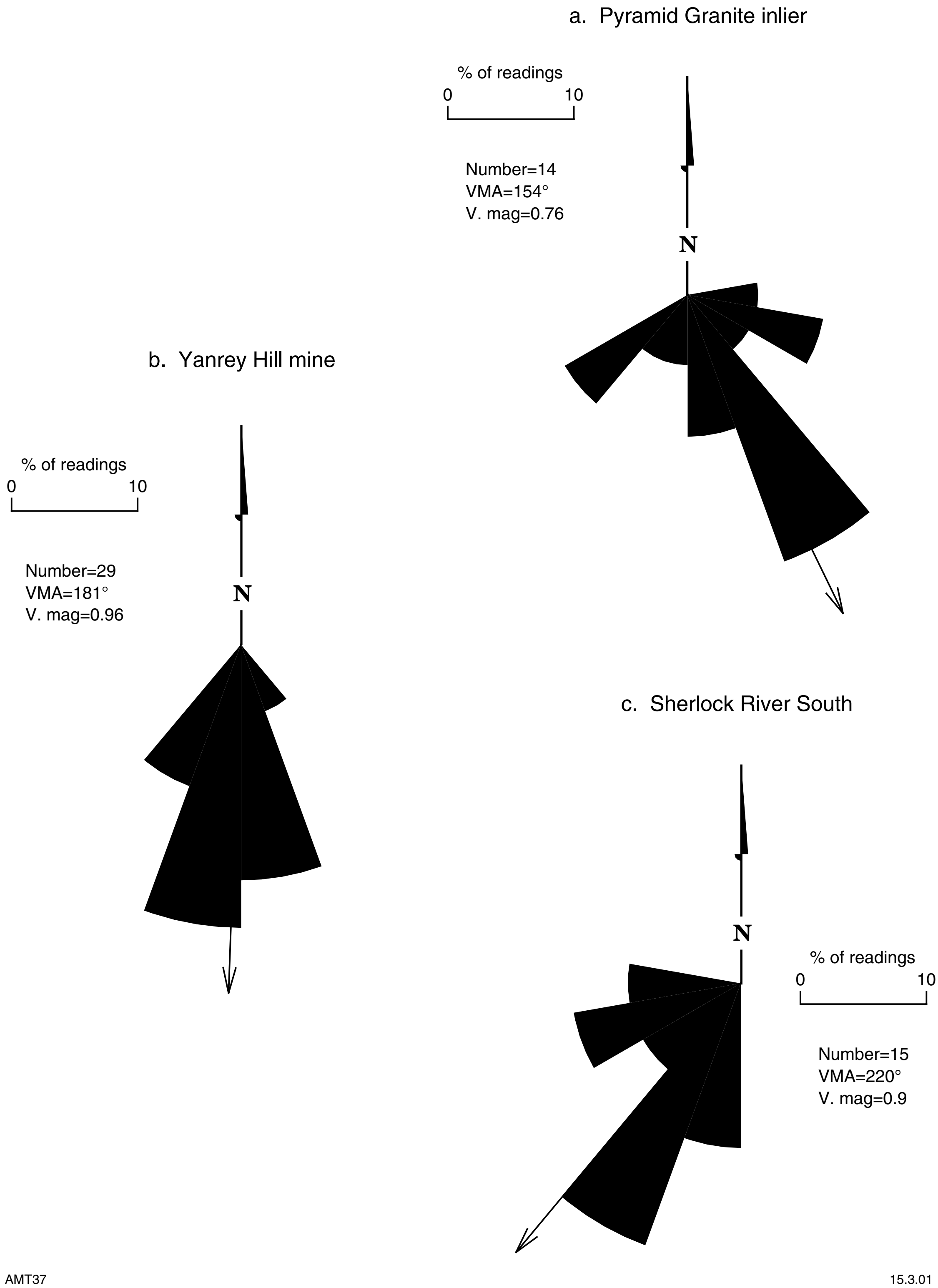


Figure 5.6. a) Palaeocurrent rose diagrams for Hardey Formation sandy braided-fluvial facies from the northwest (a–c), northeast (d) and south Pilbara (e–m) sub-basins. Data are recorded from axes of trough cross-strata. VMA = vector mean angle, V. mag = vector magnitude

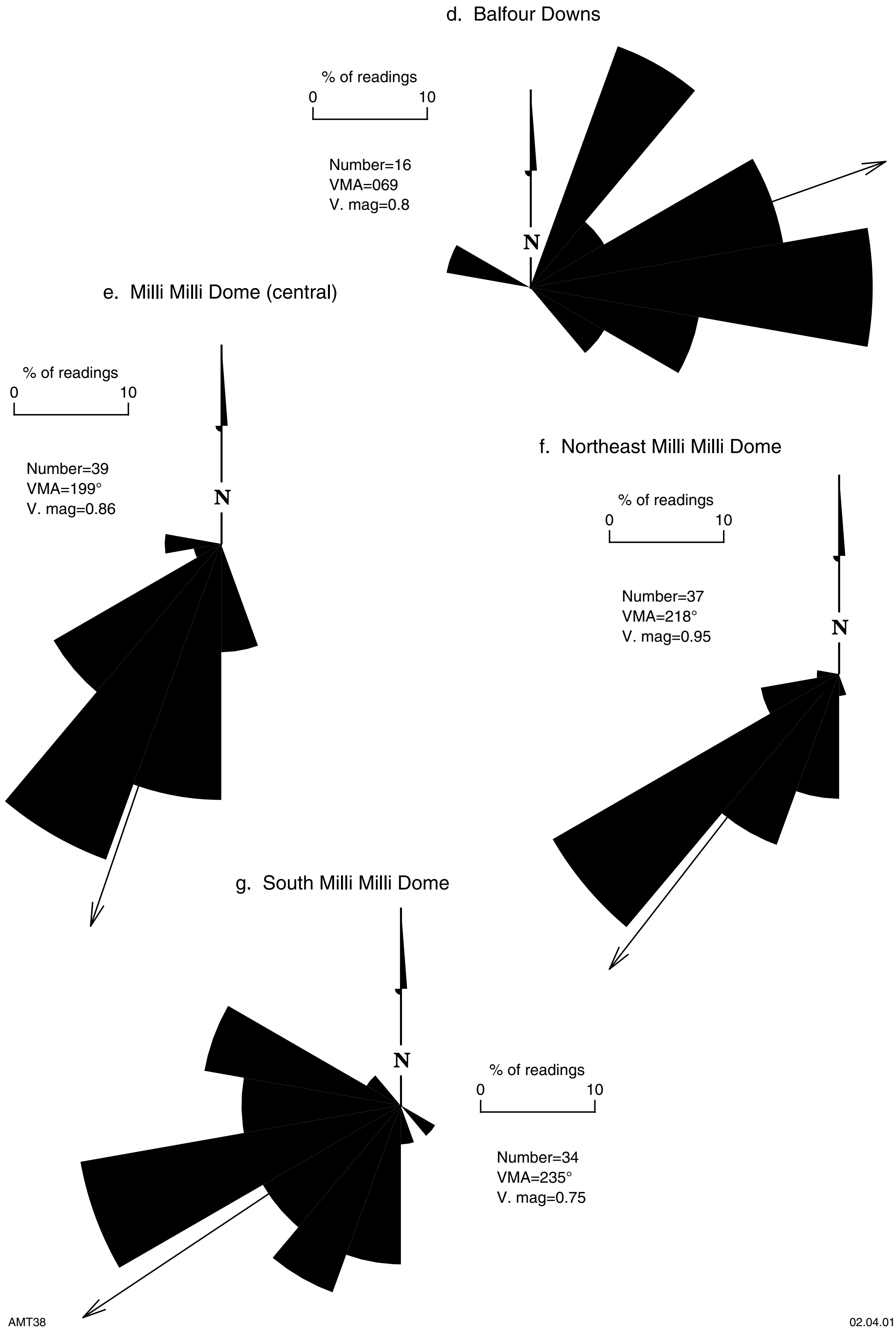
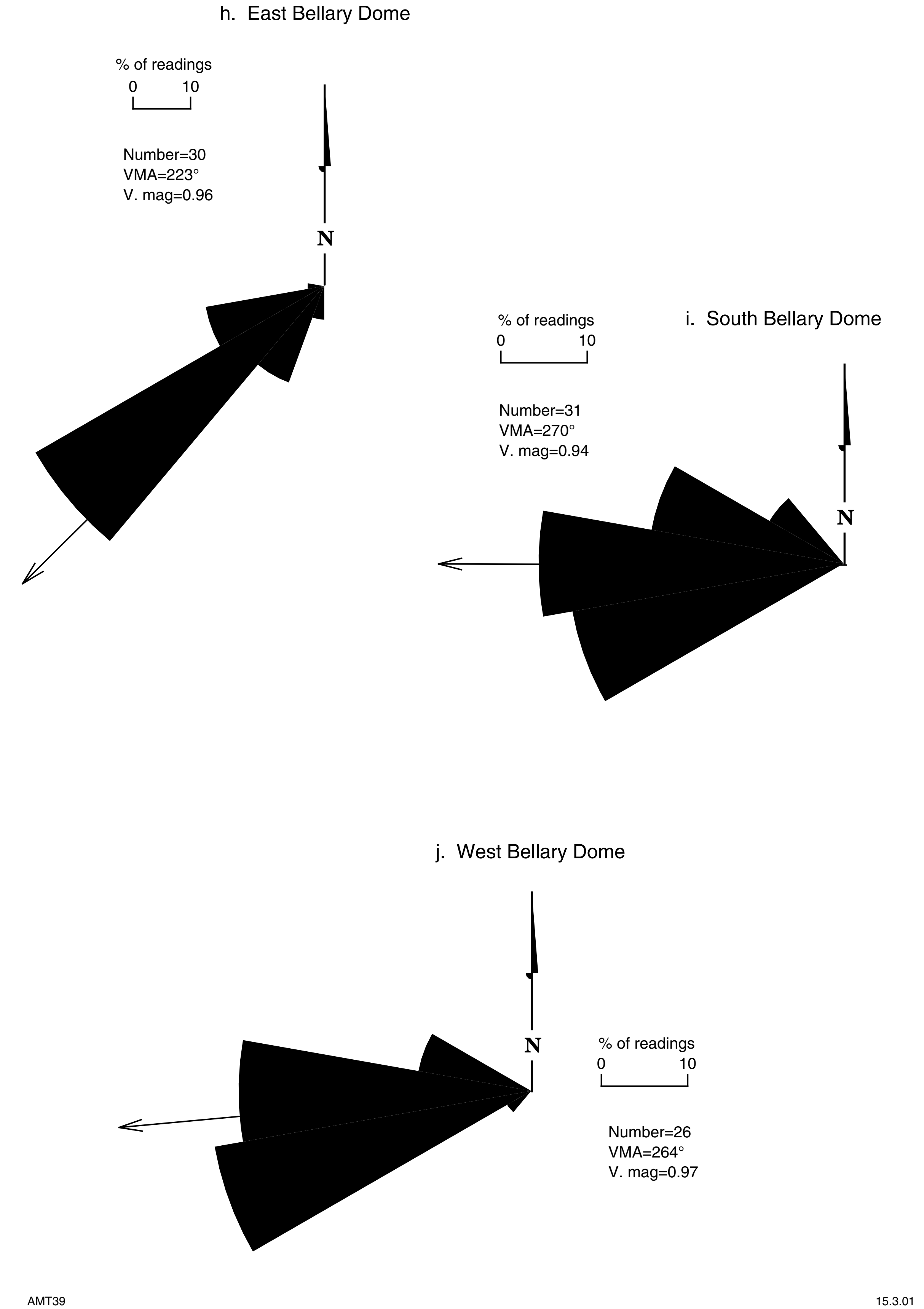


Figure 5.6. (continued)



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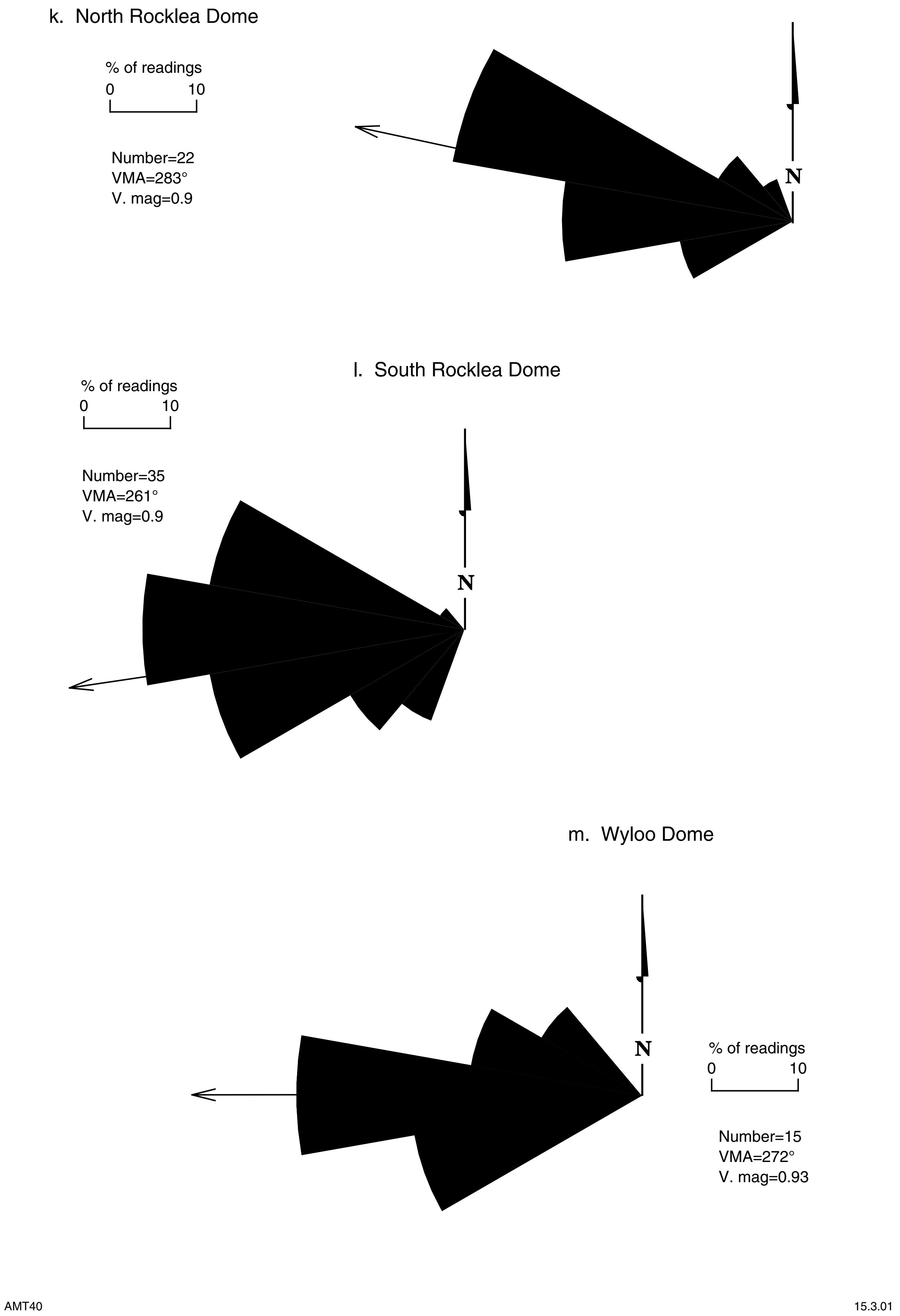


Figure 5.6. (continued)

cross-stratification and horizontal to undulatory lamination, represent in-channel and bar-top deposits. Variable palaeocurrent trends from smaller scale structures in the upper part of the bar were formed during lowering water levels, when flow directions were influenced by sandbar topography.

Lacustrine facies

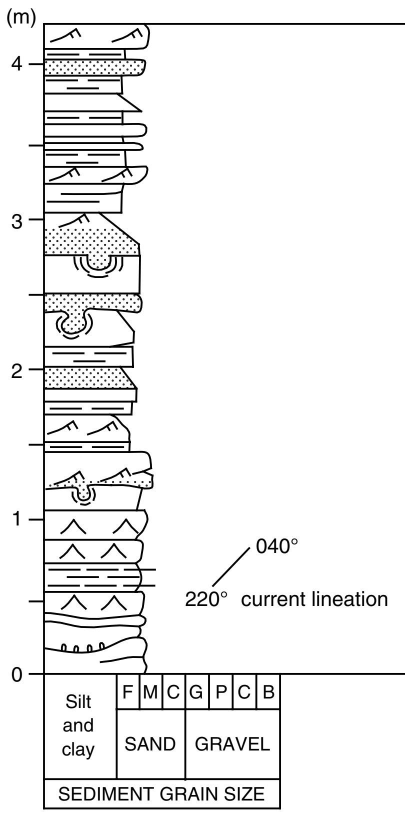
Description: Lacustrine facies rocks (Fig. 5.7) form lenticular units that generally range in thickness from a few metres to about 30 m. In most outcrops these rocks are associated with alluvial fan, coarse-grained braided fluvial, sandy braided fluvial, and deltaic facies. Principal rock-types within this facies are mudstone and siltstone interbedded with variable amounts of fine- to very coarse grained sandstone and conglomerate.

Parallel planar to undulatory lamination is the dominant sedimentary structure in many mudstone, siltstone, and sandstone layers. Siltstones and fine- to medium-grained sandstones also exhibit ripple cross-lamination and occur locally in symmetrical and asymmetrical ripple linsen. Straight crested or lunate bedforms are present locally on bedding surfaces (Fig. 5.8); desiccation cracks are also recorded.

Beds of erosively based, medium- to very coarse grained sandstone and granule to pebble conglomerate are interlayered with the argillite and finer sandstone layers. These beds, which may or may not show normal grading, are either massive or show horizontal stratification; diffuse trough cross-stratification is also present locally. Evidence of soft-sediment deformation, usually in the form of flame or ball-and-pillow structures, is observed locally at the contact between fine- and coarse-grained rock types.

Interpretation: These rocks were laid down in an environment where mud, silt, and fine sand were deposited from suspension and weak currents, interrupted by periods of higher energy sand and gravel sedimentation. The thickness and lenticular geometry of this facies, and its close association with braided fluvial and alluvial-fan facies, is indicative of a lacustrine rather than submarine setting.

Horizontally laminated mudstone and siltstone layers represent low-energy lake-floor sediments. Siltstone and sandstone that display straight-crested symmetrical and asymmetrical ripples and lunate ripples reflect the influence of small waves and lower flow regime currents respectively. Erosively based sandstone and conglomerate beds are interpreted as higher energy sediment gravity



DESCRIPTION	INTERPRETATION
An assemblage of thin- to thick-bedded, fine- to medium-grained micaceous or chloritic sandstone and chloritic siltstone. Sandstones are tabular to sublenticular bedded and display a variety of structures including: bifurcating, straight-crested symmetrical ripples; asymmetrical ripples, horizontal lamination with rare primary current lineation; and sediment load structures. Some beds fine upwards to a silty top, others are massive. Siltstones are massive or parallel laminated.	Lacustrine facies rocks comprising suspension (siltstone) and turbidity current deposits (graded sandstone), interbedded with wave- and current-modified shoreline deposits (rippled and lineated sandstone).

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Figure 5.7. Representative section through Hardey Formation lacustrine facies showing bedding thickness and grain-size variations, and details of internal structure, Meentheena, northeast Pilbara sub-basin. An explanation of graphic log symbols is given in Figure 1.4

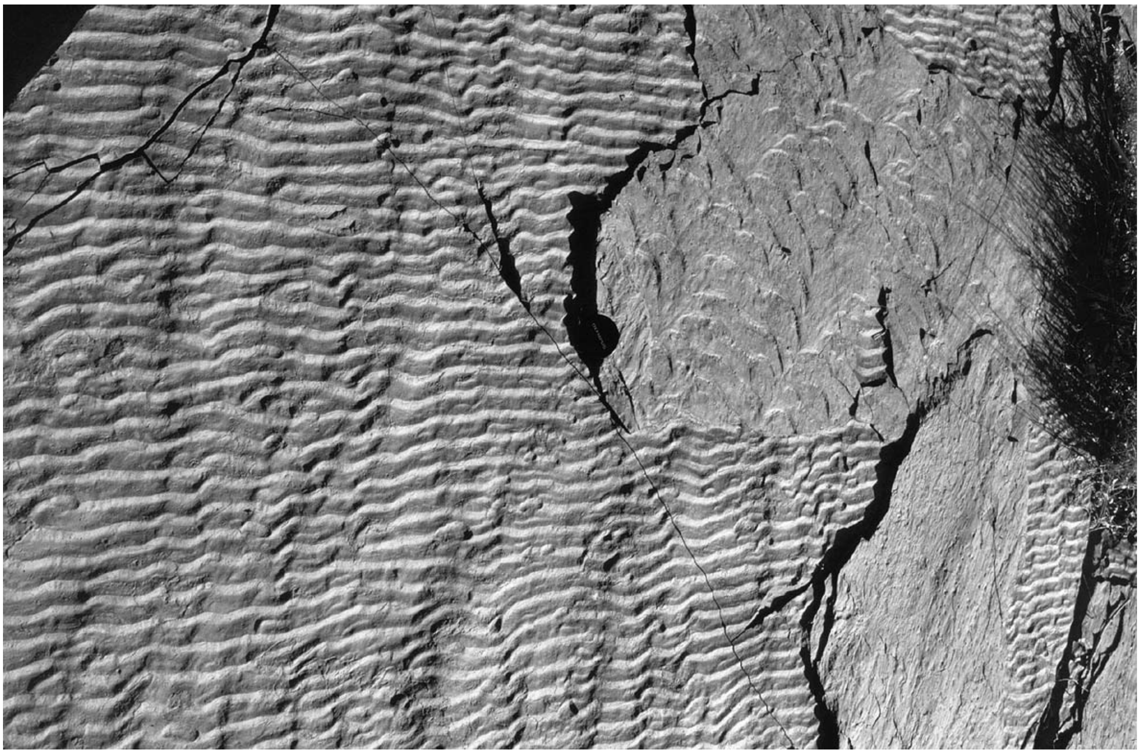

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Figure 5.8. Plan view of straight-crested, symmetrical and asymmetrical ripples with bifurcations. Lens cap for scale in centre of photograph. Hardey Formation lacustrine facies, Hales Grave Well, northeast Pilbara sub-basin

flows and traction-current deposits, and were probably introduced from adjacent fluvial or alluvial fan systems.

Deltaic facies

Description: The deltaic facies is characterized by a mixed assemblage of argillite, sandstone, and conglomerate and displays a wide range of sedimentary structures. These rocks mostly form units which vary in thickness from about 20 m to over 700 m. Deltaic deposits are commonly associated with sandy braided fluvial, coarse-grained braided fluvial, and lacustrine facies.

Deltaic facies typically consist of a single or, more commonly, repeated succession of upward-coarsening and thickening rock units (Fig. 5.9). These rock units range in thickness from about 20 m to over 300 m and consist of a lower argillaceous assemblage overlain by successively coarser grained sandstone beds, which in turn may be capped by pebbly sandstone or conglomerate.

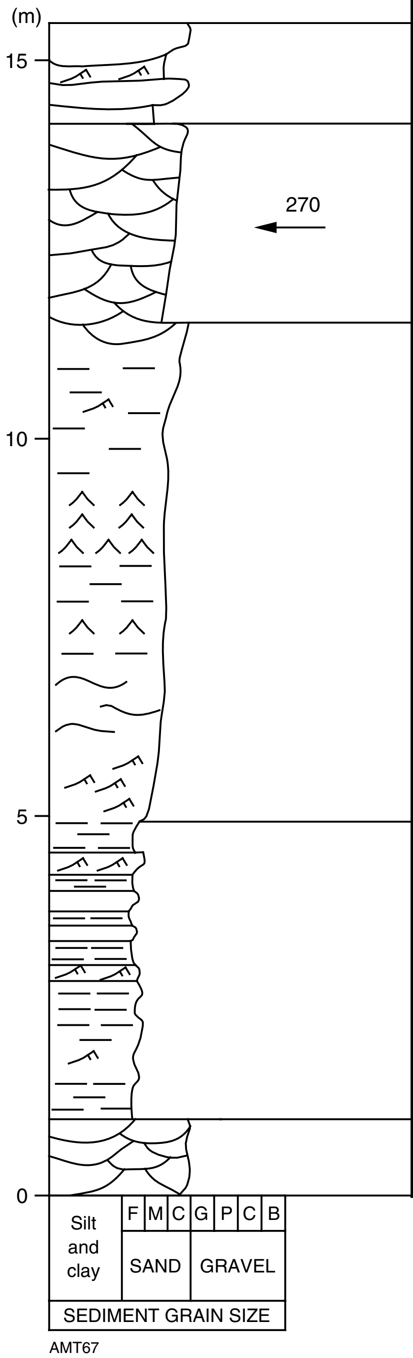
The lower argillaceous assemblage constitutes 20–95% of the thickness of most deltaic units. It consists of mudstone and siltstone interbedded with minor, thin beds of fine- to coarse-grained sandstone. Mudstone is structureless or parallel-laminated; siltstone and fine-grained sandstone are parallel-laminated or ripple cross-laminated. Thicker sand layers are erosively based and may fine upwards.

The lower part of the sandstone assemblage consists of thin- to medium-bedded fine- to medium-grained sandstone interlayered with minor argillite. Sandstone beds exhibit a variety of sedimentary structures including parallel- and undulatory-lamination, complex ripple cross-lamination, and small-scale trough cross-stratification. Symmetric and asymmetric ripple profiles are also observed. Upper parts of the sandstone assemblage are medium to thick bedded and consist of medium- to very coarse grained sandstone with minor granule to pebble conglomerate. Internal structure is dominated by small- to large-scale trough cross-stratification; these record mostly unimodal palaeoflow directions, although bipolar trends are present locally.

Interpretation: The upward-coarsening and thickening sections are analogous to the prograding deltaic successions described by Coleman et al. (1974) and Elliott (1986). The varied internal structure of these bodies reflects deposition in an environment characterized by both wave and current activity.

Symmetrical and asymmetrical ripples with complex internal structure are interpreted as wave ripples, formed under conditions of low to moderate sediment supply. (de Raaf et al., 1977; Allen, 1984). Undulose lamination probably also formed during periods of rapid sedimentation. Trough cross-stratification results from the down-current migration of dunes under lower flow regime conditions, at current velocities greater than those required to form current ripples (Harms et al., 1982).

Sands and silts that display a variety of wave- and current-formed structures are the principal deposits of



DESCRIPTION	INTERPRETATION
Scoured, quartzo-feldspathic, medium-grained sandstone, interbedded with ripple-laminated fine-grained sandstone and siltstone	Interdistributary-bay or delta-plain ?deposit
Medium- to coarse-grained quartzo-feldspathic sandstone. Stacked sets of medium-scale trough cross-stratification preserve unidirectional palaeocurrents	Distributary-channel outlet
Fine- to medium-grained quartz sandstone with symmetrical and asymmetrical ripples, and planar to undulatory parallel lamination	Bar-crest and distal bar
Fine-grained sandstone and siltstone showing parallel lamination or ripple cross-lamination	Transgressive offshore deposits
Trough cross-stratified, medium- to coarse-grained sandstone	Channel-mouth deposits

15.03.01

Figure 5.9. Representative section through upward-coarsening deltaic facies of the Hardey Formation summarizing grain-size variation, bed thickness and internal organization, Bellary Dome, south Pilbara sub-basin. An explanation of graphic log symbols is given in Figure 1.4

modern distributary mouth bars, formed at the seaward termination of major deltaic distributary channels. In this setting, sediment deposition takes place close to the mouth of the distributary channel and constructs a series of discrete or coalesced bars. Bar crests aggrade rapidly during times of flood; however, the new sediment is subsequently reworked and transported to the seaward side of the bar during falling water levels. By this mechanism, distributary mouth-bars and their associated distributary channels prograde seawards over the prodelta (offshore) deposits (Coleman et al., 1974). This results in a thick, upward-coarsening succession in which the offshore muds and silts are overlain by distal bar sands that have been reworked by waves and currents. These, in turn, are succeeded by coarser grained sand, displaying current-formed structures and which record the offshore migration of the topmost bar and distributary channel facies.

Shoreline facies

Description: The shoreline consists of medium- to very coarse grained quartz-rich sandstone and subordinate conglomerate. These rocks commonly form units which vary in thickness from about 1–10 m. Shoreline facies are associated with deltaic and lacustrine deposits.

Internal structure is dominated by parallel-stratification and very low angle planar cross-stratification. Primary current lineation, shallow scours, and straight-crested symmetrical and asymmetrical ripples occur on some bedding surfaces.

Interpretation: The compositional and textural maturity of the sandstone, coupled with widespread parallel-stratification, low-angle planar cross-stratification with associated scours, and primary current lineation indicates that deposition took place in high-energy environment. Harms et al., 1982) suggested that low-angle planar cross-stratification is the product of non-uniform, unsteady flow, and pointed out that it is the dominant structure formed in the swash zone of modern sandy shorelines.

Volcanic facies

Volcanic rocks include felsic and mafic volcaniclastic deposits, basaltic flows, and local pillow lava. In addition, massive quartz–feldspar porphyries of uncertain origin are also recorded.

Felsic volcaniclastic facies

Description: A varied suite of felsic volcaniclastic rocks, ranging from tuff to lapilli-tuff and lapilli tuff breccia, is present within the Hardey Formation. These rocks are interbedded with non-volcanic facies, or else form part of the large felsic volcanic complexes represented by the Bamboo Creek and Lyre Creek Members.

Tuff and lapilli tuff beds range in thickness from 0.05–3 m, and bed geometry varies from sheet-like to tabular or lenticular. Thin-bedded, fine-grained tuff is generally parallel planar to undulatory laminated or ripple-cross laminated, and straight-crested symmetrical ripples are observed locally on bedding surfaces. Locally, the beds show normal grading and are interlayered with thick argillite units. The composition of these rocks shows a complete spectrum from vitric and lithic tuff to volcanic arenites that contain only small amounts of pumice and shard material. No evidence of welding was observed in any of these thin tuff beds.

Thin (20–150 mm), laterally persistent beds of packed accretionary lapilli are interlayered with other volcaniclastic facies. They have sharp, non-erosive bases and drape over irregularities in the underlying surface. Most beds consist of one, or more, non-graded to normally graded layers in which the size of the accretionary lapilli generally varies between 2 and 12 mm.

Thick-bedded lapilli tuff and breccia forms irregular lenticular or tabular beds, which generally range in thickness from 1 to 10 m. Irregular, lenticular-bedded lapilli tuffs and breccias form the major component of the Lyre Creek Member in the northwest Pilbara sub-basin. They are characterized by strongly erosional lower bedding contacts (Fig. 5.10) which may cut as much as 10 m into underlying beds. Locally, these erosional surfaces display vertical and overhanging walls; load structures may develop where the underlying bed is of much finer grain size. Most lapilli tuff and breccia units are massive, others contain horizontal layers of accretionary lapilli, or parallel planar- to undulatory stratified tuff. Normal grading is present in the upper parts of some beds. In thin section, the lapilli tuffs are very poorly sorted and show a great diversity in shape and composition. They include monocrystalline and polycrystalline quartz, plagioclase and K-feldspar, vitric fragments, and volcanic lithic fragments (rhyolite, dacite, and basalt). Whole and broken accretionary lapilli are abundant in some samples. No evidence of welding was observed.

Tabular-bedded dacitic breccia and lapilli tuff is recognized in the Bamboo Creek Member of the northeast Pilbara sub-basin (Fig. 5.11). Beds generally vary from 2 to 5 m thick, are erosively based, and are separated by up to 0.3 m of tuff. Lapilli tuff and breccia beds are massive apart from a weak to moderate, bedding-parallel foliation. Clasts, which vary up to 1 m, consist entirely of quartz–feldspar porphyry. The intervening tuff beds have sharply gradational bases and are generally massive or parallel stratified. Some exhibit normal grading and accretionary lapilli are present in some beds.

In thin section the thick-bedded lapilli tuff and tuff breccia consists of dacite porphyry fragments set in a matrix of vitric lapilli tuff. Within the matrix anhedral, often embayed, monocrystalline quartz and subhedral plagioclase phenocrysts are set in a devitrified, quartz–feldspar groundmass. Most quartz phenocrysts have undergone slight to extensive fracturing in situ. In some beds the feldspar phenocrysts are similarly affected, whereas perlitic cracks may be abundant in the groundmass. Notwithstanding the devitrified nature of much of the matrix, angular to rounded and cuspate shards and fragments of collapsed pumice are common (Fig. 5.13a).

Interpretation: The field and petrographic characteristics of this facies suggest that it is made up of primary pyroclastic deposits and reworked volcanic debris.

Thin beds of densely packed accretionary lapilli that occur interbedded with fine-grained tuff and non-volcanic facies are interpreted as pyroclastic air-fall deposits on the following basis: their thickness and lateral continuity, good sorting, non-erosional lower contacts, presence of single or multiple graded layers, and lack of associated pyroclastic flow or surge deposits. The presence of accretionary lapilli alone is not considered diagnostic of an air-fall deposit, since these clasts have been reported from a variety of pyroclastic flow and surge deposits (Cas and Wright, 1987). In addition, beds which contain whole or broken accretionary lapilli mixed with a variety of other volcanic and non-volcanic detritus probably represent reworked pyroclastic material.

Thick, tabular-bedded lapilli tuff, lapilli tuff breccia and their associated thin tuff layers have many similarities

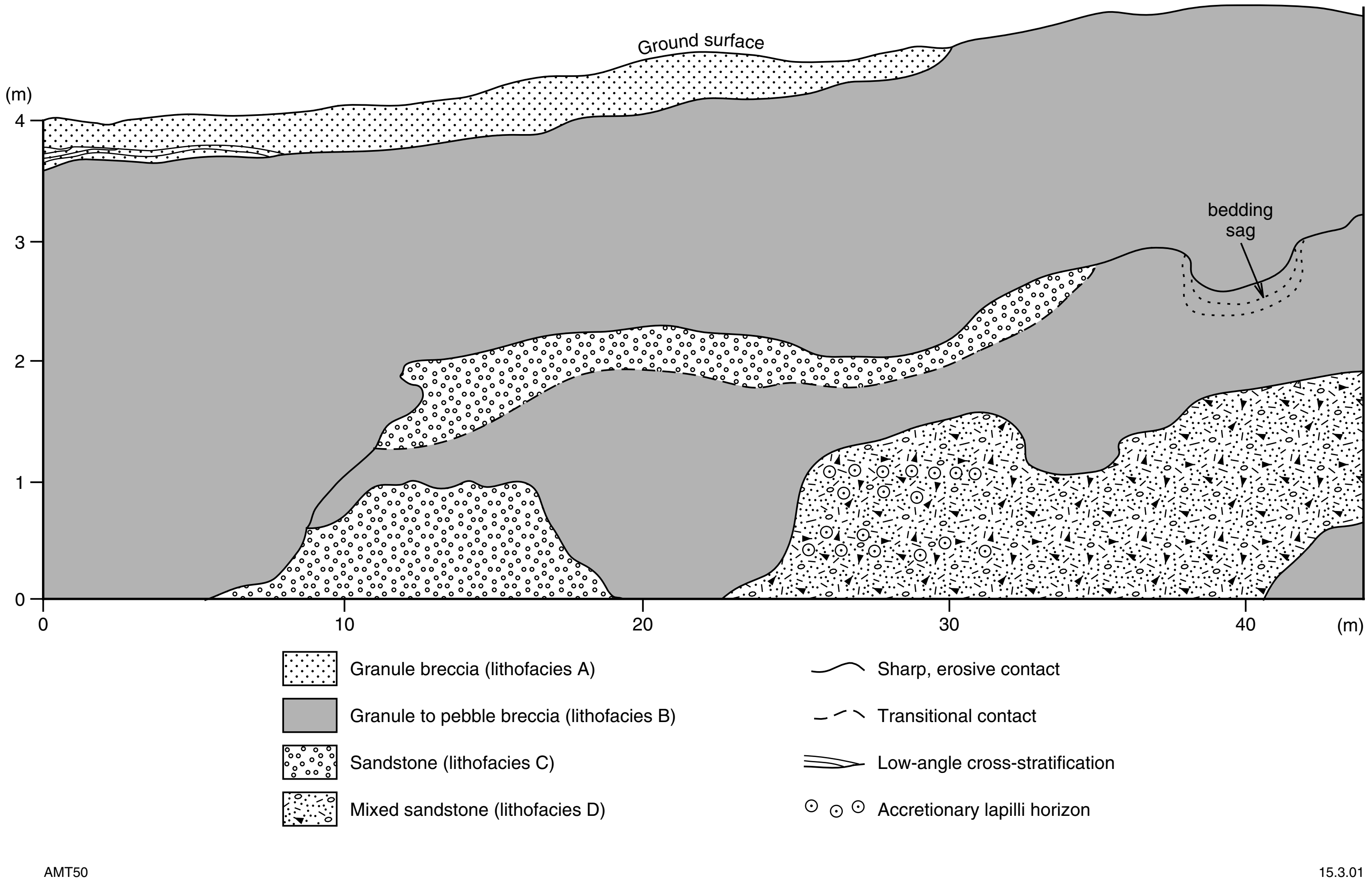


Figure 5.10. Contact relationships and lithofacies in felsic to intermediate volcaniclastic rocks, Lyre Creek Member, northwest Pilbara sub-basin. Lithofacies (A) = granule breccia showing low-angle cross-stratification locally. (B) = massive, matrix-supported granule to pebble breccia. (C) = sandstone and granule to pebble sandstone. (D) = mixed unit comprising both massive and parallel-stratified granule and pebble sandstone; discontinuous horizons of accretionary lapilli, up to 20 mm in diameter occur locally. Paraburdoo–Dampier railway cutting at AMG 507300E, 7648700N

Figure 5.11. Interbedded dacitic breccia and volcanic sandstone, Bamboo Creek Member, Helen Well, northeast Pilbara sub-basin. Boulder-sized clasts of dacitic porphyry occur in many breccia beds and accretionary lapilli are present in some of the volcanic sandstones

with modern and Phanerozoic pyroclastic flow deposits (Fisher and Schmincke, 1984; Cas and Wright, 1987; Cole and DeCelles, 1991). A primary volcanic origin is indicated by the high proportion of juvenile and accessory lithic grains, pumice, vitric groundmass, and embayed ß-quartz. The absence of cannibalized volcaniclastic lithic clasts, and the virtual absence of non-volcanic lithic clasts, indicate that the various fragments were not derived from pre-existing consolidated deposits. In addition, the presence of collapsed pumice within the matrix is widely considered to reflect high temperature, and therefore pyroclastic, emplacement (Ross and Smith, 1961). Widespread perlitic cracks and fractured and shattered phenocrysts provide further evidence for elevated temperatures (and rapid quenching) at the time of deposition (Cole and DeCelles, 1991).

In the case of pyroclastic flows from the Miocene Tecuya Formation, Cole and DeCelles (1991) proposed that the association of dacite tuff breccia and interbedded laminated tuff is indicative of subaqueous deposition. A similar lithological association is present in the Hardey Formation, but here there is no direct evidence from which to infer a subaqueous setting. In this case it seems more likely that the tuff beds, particularly those containing accretionary lapilli, represent air-fall ash deposits laid down during the final stages of the pyroclastic flow.

Irregular, lenticular-bedded lapilli tuff and breccia are interpreted as the reworked deposits of a subaerial hydroclastic volcanic complex. The lack of welding or other evidence for high-temperature emplacement, occurrence of non-pyroclastic debris, presence of beds of broken accretionary lapilli, and strongly erosional bedding contacts argue against these rocks being primary pyroclastic deposits. Instead, their morphology, internal structure, and composition is more analogous to the hyperconcentrated streamflow and debris-flow deposits of modern lahars (Rodolfo, 1989). The deposits described here are characterized by a complex history of channel formation, infill, and later erosion, most of which took place while the volcanic debris was in an unconsolidated or weakly consolidated state.

Thin-bedded, fine-grained tuff and volcanic sandstone that display symmetrical and asymmetrical ripple bedforms and ripple cross-lamination are interpreted as pyroclastic debris that has been reworked by wave and current action in a shallow marine or lacustrine setting. Similarly, normally graded tuffaceous rocks that contain a significant component of non-volcanic detritus, and occur interbedded with thick argillaceous units, are more likely to be resedimented volcaniclastic turbidites than primary pyroclastic material.

Massive quartz–feldspar porphyry facies

Description: The massive quartz–feldspar porphyry facies makes up most of the 800 m-thick Bamboo Creek Member of the northeast Pilbara sub-basin. In many outcrops the internal structure is completely massive and the unit is cross cut by large, irregularly undulating bedding-parallel joints. Other, smaller scale, apparently randomly oriented joints also combine to give the rock a brecciated appearance in many localities. Poor columnar jointing with variable long-axis orientation has been observed locally. The most commonly observed variation within the massive porphyry facies is a weak to strong bedding-parallel fabric which, in some outcrops, gives the rock a foliated appearance. In other cases the layering is defined by variations in colour and grain size within the fragmental porphyry. The porphyry contains layers of septarium-like lithophysae (Fig. 5.12); lithophysae fragments are also recorded.

In thin section, there is a complete gradation from non-brecciated to fragmental quartz–feldspar porphyry (Fig. 5.13a,b,c). The former consist of quartz, plagioclase, and minor mafic phenocrysts set in a microcrystalline granular quartz–feldspar matrix. Quartz phenocrysts are anhedral to subhedral and show varying degrees of embayment and corrosion. Pink plagioclase is commonly anhedral to euhedral oligoclase or andesine, and microcline is observed locally (Trendall, 1975b). Anhedral to subhedral mafic phenocrysts are much less abundant than quartz and feldspar and are represented by aggregates of biotite, chlorite, epidote, and altered ilmenite.

Fragmental quartz–feldsar porphyry is similar petrographically to the thick-bedded lapilli tuff and tuff breccia facies. This consist of quartz and plagioclase phenocrysts, and dacite porphyry fragments set in a devitrified, quartz–feldspar groundmass. Most quartz phenocrysts have undergone slight to extensive fracturing in situ. In some beds the feldspar phenocrysts are similarly affected, whereas perlitic cracks may be abundant in the groundmass. Despite extensive devitrification of much of the matrix, angular to rounded and cuspate shards, and fragments of collapsed pumice are observed locally.

Interpretation: Massive quartz–feldspar porphyry occurs in association with lapilli tuff, and lapilli tuff breccia units of probable pyroclastic flow origin. It is likely, therefore, that the massive porphyries also form part of a felsic volcanic complex, although in many cases, their precise mode of origin is unclear.

Massive, foliated, and stratified units of fragmental quartz–feldspar porphyry are similar texturally to the thick-bedded lapilli tuff, and lapilli tuff breccia units, though they lack the thin tuff interbeds. For this reason they are also interpreted as pyroclastic flow deposits (see *Interpretation* of **Felsic volcaniclastic facies**).

Porphyry units which have a less obvious fragmental appearance, but which contain horizons of abundant lithophysae, may also have originated as pyroclastic flows. Lithophysae are believed to form as the result of laminar flow modifying the shape of an expanding vesicle (Cas and Wright, 1987). Bonnichsen and Kauffman (1987) recorded that swarms of lithophysae are present in the basal portions of Phanerozoic ash-flow sheets in the Cougar Point Tuff, Idaho, USA, but are rare in rhyolite flows from the same area.

Units that consist only of massive to weakly layered, non-fragmental quartz–feldspar porphyry are difficult to

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Figure 5.12. Lithophysae in massive quartz–feldspar porphyry facies, Bamboo Creek Member, Meentheena Centrocline, northeast Pilbara sub-basin

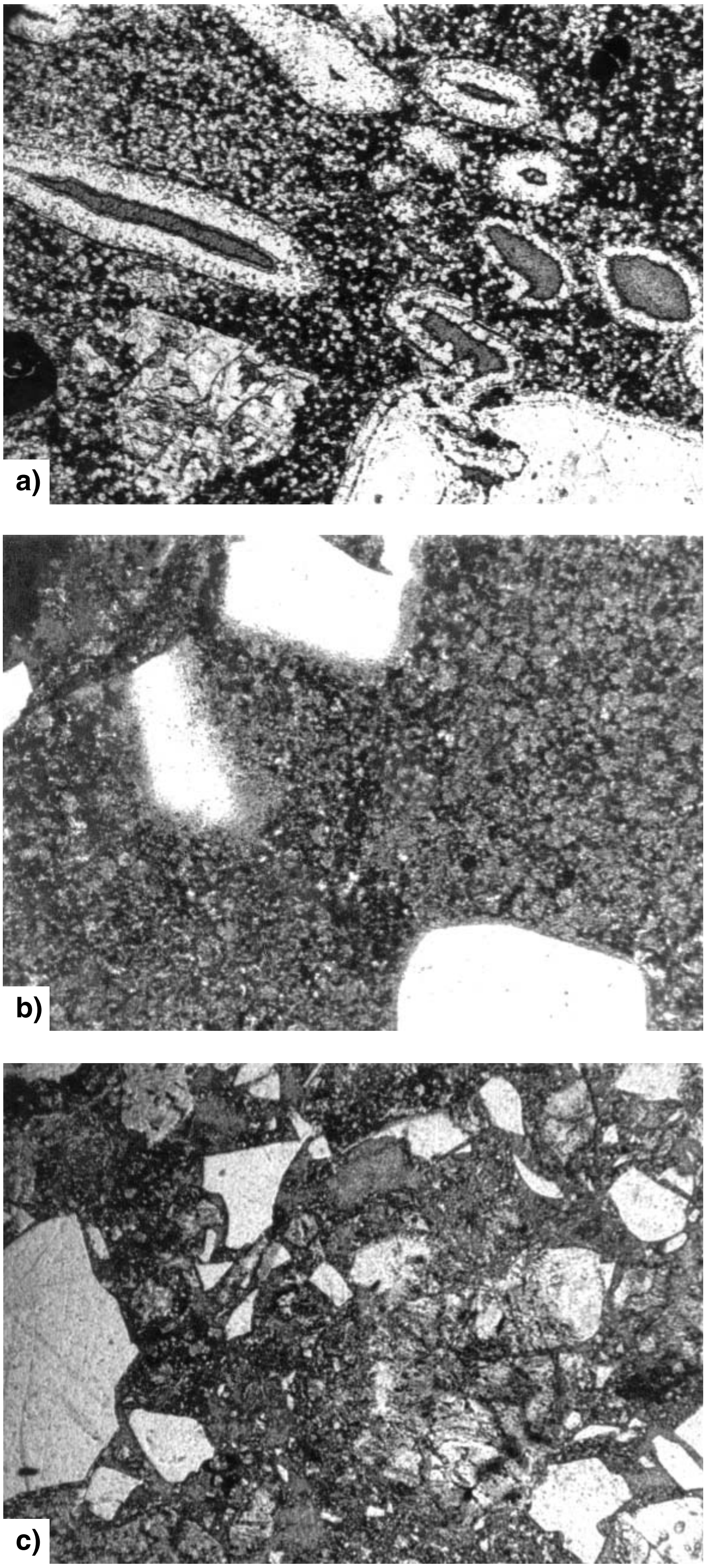

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Figure 5.13. a) Photomicrograph of dacitic volcaniclastic breccia, Bamboo Creek Member, northeast Pilbara sub-basin. Plane polarized light, field of view = 3 mm; b,c) Photomicrographs showing variation in degree of fragmentation within massive quartz–feldspar porphyry facies; b) massive, non-brecciated porphyry showing reaction rims around quartz phenocrysts; c) massive, volcaniclastic rock consisting of angular quartz and plagioclase phenocrysts, and dacite porphyry fragments, set in a devitrified, quartz–feldspar groundmass, Bamboo Creek Member, northeast Pilbara sub-basin. Plane polarized light, field of view = 3 mm

interpret. Their physical character suggests that they could represent dacitic lava flows or subvolcanic intrusive rocks. Another possibility is that they are rheoignimbrites, formed when welding in an ignimbrite has continued to the point where the identity of the individual glass s-hards has become lost and the whole mass has been reconstituted as secondary lava, capable of viscous flowage (Ekren et al., 1984).

Mafic volcaniclastic facies

Description: Tuff and lapilli tuff beds range from 0.05 to 2.0 m and exhibit a variety of internal structures ranging from massive to parallel-stratified or cross-laminated. Thin beds tend to be normally graded. Lithic and vitric fragments, and the matrix, are basaltic to andesitic in composition and are now composed of fine-grained chlorite, actinolite, plagioclase, quartz, stilpnomelane, and secondary calcite (Blight, 1985). Other fragments include rounded quartz and K-feldspar, sandstone, basalt, tuff, and accretionary lapilli.

Interpretation: Tuff and lapilli tuff beds probably originated as volcaniclastic deposits associated with subaerial hydroclastic eruptions. The varied clast composition includes a significant component of non-pyroclastic debris, however, and suggests that many of these deposits have been redeposited.

Basaltic flows

Description: Basaltic flows and subordinate pillow lava and hyaloclastite occur interbedded with non-volcanic and volcaniclastic rocks (Blake, 1984a, 1993; Blight, 1985; Thorne and Tyler, 1994). Commonly, the basalt units are thin (<100 m); however, in the Marble Bar sub-basin they attain a thickness of 1100 m (Blake, 1993). Flow morphologies are similar to those described for the Mount Roe Basalt (Blake, 1993) although no examples of very thick (>20 m) massive to amygdaloidal lava flows have been recorded. No chemical analyses are available for these lavas.

Interpretation: Massive to amygdaloidal basalts originated from voluminous subaerial flows that retained a high volatile content during emplacement. The geological setting of the pillow lava and hyaloclastite suggests that these formed when subaerially erupted basalt flows entered lakes or rivers.

Distribution of facies

The Hardey Formation exists in all the major sub-basins; principal outcrop areas are shown in Plates 1a and 1b, and this information is summarized in Figs 5.1 and 5.14.

Northwest Pilbara sub-basin

Throughout most of the northwest Pilbara sub-basin, excluding the Nunyerry area, the Hardey Formation can be subdivided into two main units (Blake, 1993): a lower clastic sedimentary unit with abundant tuff horizons and rare basalt flows, and an upper unit, here referred to as the Lyre Creek Member, consisting of felsic pyroclastic and reworked pyroclastic deposits. These units are intruded by an extensive dolerite sill complex — the Cooya Pooya Dolerite (**Chapter 11**). The Hardey Formation is absent from the Cape Preston area and is either thin or absent from parts of PINDERI HILLS (Kojan and Hickman, 2000).

The lower unit typically varies in thickness from about 150 to 500 m, although the thickest single section is in excess of 1000 m (Blake, 1993). Principal facies recognized are, in order of decreasing abundance: sand-dominated braided fluvial, shallow lacustrine, alluvial fan and coarse-grained braided alluvial, deeper lacustrine, deltaic, and thin talus deposits above Mount Roe Basalt and granite–greenstone basement. Fluvial palaeocurrents are overwhelmingly toward the south (Fig. 5.6a–c; Blake, 1993); data from alluvial fan and coarse-grained braided alluvial facies are more variable. The Lyre Creek Member ranges in thickness up to 1000 m but in most areas is less than 200 m. The unit is dominated by felsic volcaniclastic rocks, consisting mostly of primary and redeposited, unwelded air-fall and ashflow tuff. Thickness variations and sizes of cognate blocks suggest the presence of three major eruptive centres, forming the George River Volcanic Complex, in the central part of the sub-basin (Blake, 1993).

On YARRALOOLA, the Hardey Formation outcrops in the upper reaches of the Yanyare River and also, semi-continuously, between Moondle Creek and Yannery Hills mine. The formation is also recorded in the MF–1 diamond drillhole near the confluence of the Portland and Fortescue Rivers (CRA Exploration Pty Ltd, 1987a).

In the vicinity of Yanyare River, the Hardey Formation either overlies the Mount Roe Basalt disconformably, or rests unconformably upon granitoid basement. Southwest of Mount Leopold (PINDERI HILLS AMG 561701), the Hardey Formation overlies foliated granitoid and is only 80 m thick. Most of the succession consists of medium- to very coarse grained quartz sandstone, feldspathic sandstone, and lithic sandstone (sandy braided fluvial facies) interbedded with ripple laminated fine- to medium-grained tuffaceous sandstone and argillite (lacustrine facies). Parallel-stratified tuff and lapilli tuff (reworked pyroclastic air-fall deposits) dominate the upper 15 m of the succession and are overlain by the lowermost flows of the Kylena Formation.

Near Yannery Hills mine, the Hardey Formation unconformably overlies granite–greenstone felsic schist and consists of about 60 m of cross-stratified micaceous quartz sandstone and feldspathic quartz sandstone (sandy braided fluvial facies). Very large scale cross-stratification, which presumably formed in the deeper channels, is confined to hollows in the basement topography; palaeocurrent data from small- to very large scale trough axes are directed toward the south and southwest. Approximately 60 m of the overlying succession is exposed southeast of Yannery Hills mine (PINDERI HILLS AMG 965690) and consists of fine- to very coarse grained, locally pebbly, lithic sandstone and micaceous sandstone (?deltaic facies) with palaeocurrents directed toward the southwest and northeast.

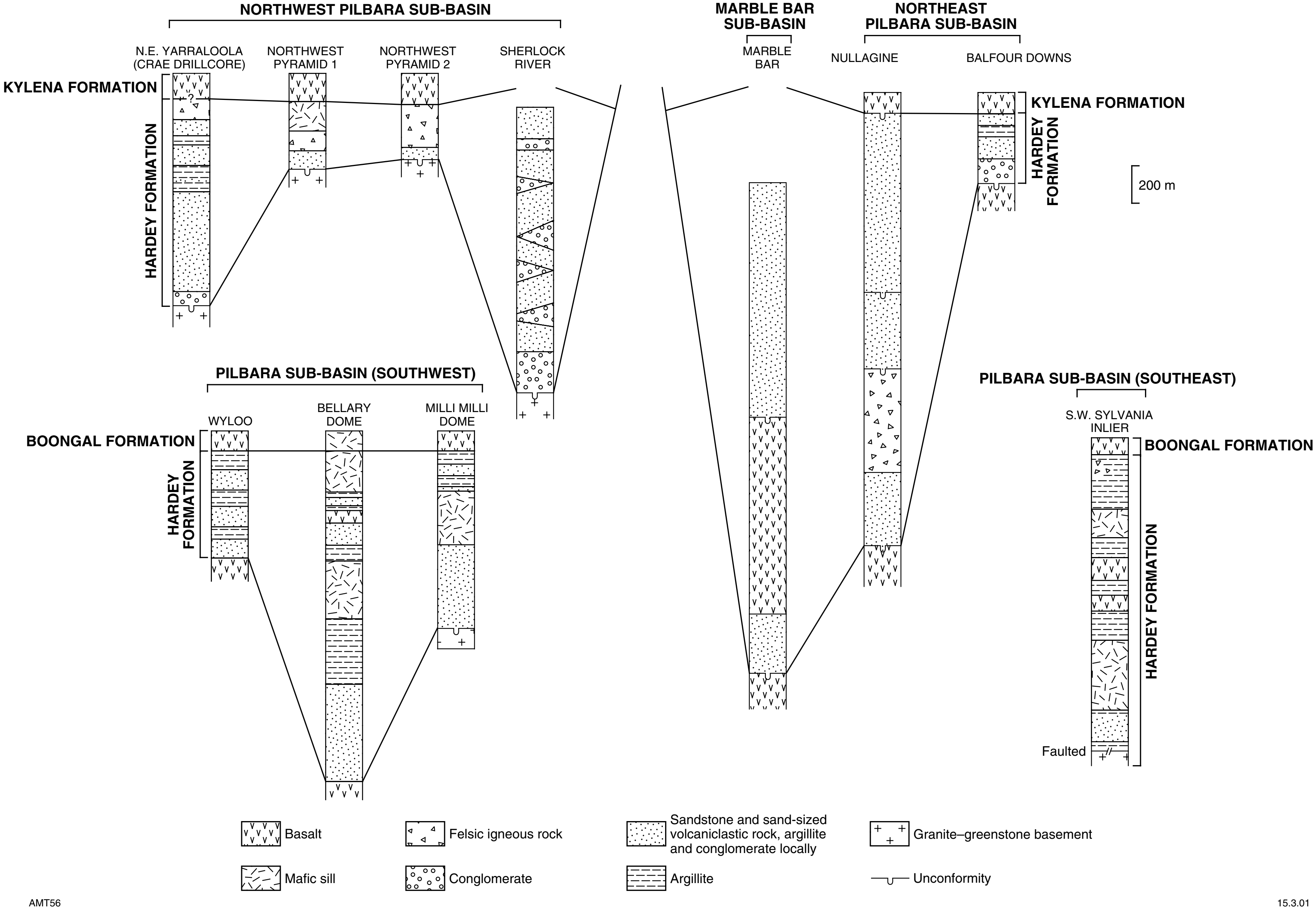

Figure 5.14. Generalized stratigraphic profiles showing variation in thickness and rock type through the Hardey Formation in the northwest, northeast, and south Pilbara sub-basins. Data from the Sherlock River, Marble Bar, and Nullagine areas are from Blake (1984a, 1993)

One thousand and fifty metres of Hardey Formation rocks are recorded from the MF–1 diamond drillhole on east central YARRALOOLA (CRA Exploration Pty Ltd, 1987a), immediately south of the east-southeasterly trending Portland Fault (Kojan and Hickman, 2000). The formation, which unconformably overlies altered basalt considered to part of the greenstone basement, consists of a 50 m-thick lower unit of basalt-derived breccia (?alluvial fan facies or talus deposits) overlain successively by: 575 m of fine- to very coarse grained quartz sandstone, lithic sandstone, and conglomerate (?braided fluvial facies); 320 m of fine- to coarse-grained lithic sandstone, argillite, and tuff (?deltaic facies); and 100 m of tuff and lapilli tuff (?pyroclastic air-fall deposits). The upper contact with the overlying Kylena Formation is sharp.

On DAMPIER, the lower part of the Hardey Formation outcrops on Rosemary Island. Kriewaldt (1964a) reported that this succession consists of, '. . . bedded and cross-bedded, pale-green, fine-grained tuffaceous rocks with white quartz sandstone and purple and green shales.'

On ROEBOURNE, occurrences of Hardey Formation are confined to a small area immediately northwest of Mount Anketel, and to a strip of outcrop extending from south of Mount Roe to south of Warambie Homestead. At Mount Anketel, the Hardey Formation disconformably overlies the Mount Roe Basalt and consists of 60 m of granite–greenstone and basalt-derived cobble to boulder conglomerate (alluvial fan and coarse-grained braided alluvial facies). Sparse palaeocurrent data from cobble imbrication suggest a westerly palaeoflow. Hickman (1999) considered this unit to be a sedimentary interval within the Mount Roe Basalt. Laterally discontinuous units of alluvial fan and coarse-grained braided alluvial facies disconformably overlie the Mount Roe Basalt in the area southeast of Mount Roe. At Pinanular Pool these deposits are about 60 m thick and consist largely of basalt-derived detritus. Cobble imbrication and trough cross-stratification record an easterly directed palaeoflow. The overlying unit is poorly exposed and consists of ripple cross-laminated tuff and fine-grained sandstone (lacustrine facies).

On PYRAMID, the Hardey Formation is exposed over a large area between Western Creek and Mumbillina Bluff. In the vicinity of Western Creek (COOYA POOYA AMG 000649), the Hardey Formation unconformably overlies granitoid basement and comprises about 80 m of cross-stratified quartz and lithic sandstone (sandy braided fluvial facies) and interbedded tuffaceous argillite and sandstone (lacustrine facies). Palaeocurrent trends from the top of this unit (COOYA POOYA AMG 011609) are directed toward the south. Here, these beds, and the overlying 70 m thick felsic lapilli tuff unit (air-fall and reworked pyroclastic deposits), are folded into open folds with strongly arcuate axial traces. This has resulted in a localized angular unconformity with the overlying horizontally dipping Cooya Pooya Dolerite.

The unconformity between the Hardey Formation and the underlying granitoid basement is also exposed to the southeast of Western Creek (COOYA POOYA AMG 066547). Here, the lower sandstone and tuffaceous argillite unit (braided fluvial and lacustrine facies) is 50 m thick and is overlain by 30 m of massive, felsic lapilli tuff (?pyroclastic ash-flow deposit). Palaeocurrents from trough axes in the braided fluvial facies are toward the south and southeast; data from current ripples are toward the south and southwest. The remainder of the Hardey Formation is estimated to be 250–300 m thick and comprises thin- to very thick bedded felsic lapilli tuff, accretionary lapilli tuff, and tuff breccia (primary air-fall and reworked pyroclastic flow and air-fall deposits). Most lithic fragments are sand to pebble sized.

Approximately 130 m of massive felsic lapilli tuff and lapilli tuff breccia (non-welded pyroclastic flow or reworked pyroclastic flow deposits) form the upper part of the Hardey Formation at Pyramid Hill. Here, lithic blocks within the breccia are up to 0.4 m in diameter.

Hardey Formation rocks outcrop over a strike length of about 60 km between the Nunyerry Creek and Sherlock River Fault Systems (Blake, 1993; Plate 1a). This succession, referred to as the Pilbara Creek Sequence by Blake (1993), is up to 1480 m thick and is dominated by terrigenous and epiclastic volcaniclastic sedimentary rock, and reworked tuff with minor tholeiitic basalt (Blake, 1993). The sedimentary rocks are dominated by two complex interfingering associations. Firstly, an association dominated by cobble to boulder conglomerate that outcrops mostly close to the western side of the Sherlock River Fault System and is interpreted by Blake (1993) as a fault-controlled alluvial fan system. Conglomerate clasts are dominated by Mount Roe Basalt. The second succession is dominated by terrigenous and volcanic arenites that were deposited in a fluvial system that flowed to the south-southwest, parallel to the Sherlock River Fault System (Blake, 1993).

Near the southwestern end of the Sherlock River Fault System (MOUNT BILROTH AMG 896001) the Hardey Formation is only 110 m thick and comprises a 25 m-thick basal unit of cobble to pebble conglomerate (alluvial fan and braided alluvial facies) overlain by 50 m of fine- to very coarse grained quartz sandstone, micaceous sandstone, pebbly sandstone, and tuffaceous argillite (sandy braided fluvial and lacustrine facies). Palaeocurrent data from the sandy braided fluvial facies are directed toward the south-southwest.

Approximately 470 m of Hardey Formation is recorded from the SV–1 diamond drillhole in the Sherlock Valley (CRA Exploration Pty Ltd, 1985a). The succession, which unconformably overlies gneissic granitoid, consists of tuffaceous and arkosic sandstone, lithic conglomerate, and black siltstone.

Marble Bar sub-basin

The Hardey Formation of the Marble Bar sub-basin was not examined in detail during this study and the following account is based upon that of Blake (1984a; 1993).

The Hardey Formation of the Marble Bar sub-basin is preserved mostly in the Marble Bar Outlier (Fig. 5.1), which comprises two linked centroclines: the Pear Creek Centrocline in the north, and the Limestone Well Centrocline in the south. In both areas, the stratigraphy of the Hardey Formation consists of a lower unit dominated by subaerial basalt, with basal interfingering terrigenous sedimentary rocks (Unit 2 of Blake, 1984a; Glen Herring Creek Sequence of Blake, 1993), and an upper unit of terrigenous and epiclastic volcanic sedimentary rocks (Unit 3 of Blake, 1984a; the Pear Creek Sequence of Blake, 1993).

The base of the lower succession comprises a succession of mudrock and very coarse grained, commonly granular sandstone, with subordinate tuff and locally developed cobble to boulder conglomerate. Depositional environments include lacustrine and fluvio-deltaic systems, and local alluvial fans and talus slopes mantling topographic highs. The stratigraphy of this succession is complex, but in essence comprises an early suite of rocks derived from the north, and a younger suite derived from the southeast (Blake, 1984a). Basal sedimentary rocks are overlain by up to 1100 m of dominantly subaerial basaltic lava, with subordinate hyaloclastite and pillow lava.

Most of the upper unit is restricted to the cores of the Pear Creek and Limestone Well Centroclines (Blake, 1993). In the Limestone Well Centrocline, about 400 m of coarse-grained to granular sandstone rests unconformably on the underlying basalts, and locally, two 200 m-deep palaeovalleys filled with sandstone and conglomerate are preserved. Sandstones were deposited in braided alluvial settings and transport directions were toward the north-northeast. The Pear Creek Centrocline is bisected by a major northeasterly trending fault — the Pear Creek Fault. Northwest of the fault, about 1600 m of sedimentary rocks rests with apparent conformity upon the underlying basalt, whereas southeast of the fault, only 250 m of coarse-grained and granular sandstone rests unconformably on the basalt. The northwestern succession is subdivided by Blake (1993) as follows:

- A lower fluvio-lacustrine assemblage (up to 830 m) dominated by reworked mafic tuff with subordinate epiclastic basaltic and basement-derived arenites, and felsic tuff.
- A middle boulder to cobble conglomerate assemblage (up to 440 m) that was deposited on alluvial fans or in gravel-dominated braided alluvial environments and consists mostly of basalt clasts derived from the Mount Roe Basalt and lower part of the Hardey Formation.
- An upper assemblage of about 300 m of coarse-grained to granular sandstone that is probably equivalent to the sandstone to the southeast of the fault. Both sandstone bodies are quartzo-feldspathic and were deposited in braided alluvial environments.

Stratigraphic and sedimentological studies (Blake, 1993) suggest that the Pear Creek Fault was active, with a northwest-block-down normal movement, probably during deposition of the lower volcarenite assemblage, and certainly during deposition of the middle conglomerate assemblage, when local alluvial fans built up along the fault scarp. Palaeocurrent analysis suggests, however, that most of the conglomerate was derived from the northwest. The upper arkosic arenite was also derived from the northwest and appears to have been transported across the Pear Creek Fault without deflection or significant facies change, suggesting that the fault was not active at this time (Blake, 1993).

In summary, Blake (1993) proposed that, initially, Hardey Formation sedimentation was centred on the Marble Bar outlier. During this time, sedimentation was fault controlled and currents flowed generally from the north, although in the Limestone Well Centrocline a younger pre-basalt succession was derived from the southeast. During deposition of the upper Hardey Formation, fluvial transport was toward the north-northeast in the Limestone Well Centrocline and was associated with major north-northeasterly trending basal channels. In the Pear Creek Centrocline, flow was toward the southeast, normal to the northeasterly trending, west-block-down, Pear Creek growth fault (Blake, 1993).

Northeast Pilbara sub-basin

The Hardey Formation of the northeast Pilbara sub-basin outcrops in four main areas on NULLAGINE, YARRIE, and BALFOUR DOWNS. The three largest areas are the Northwest Oakover Syncline, the Meentheena Centrocline, and the Nullagine Synclinorium (Fig. 5.1), using the terminology of Blake (1993). The fourth, and smallest outcrop area, is on the eastern side of the Rooney Inlier on BALFOUR DOWNS.

Blake (1993) has subdivided the Hardey Formation of the Northwest Oakover Syncline, Meentheena Centrocline, and Nullagine Synclinorium into lower and upper parts, which are separated by an intra-basinal unconformity. The lower part corresponds to Unit 2 of Blake (1984b) and was later named the Hales Grave Well Sequence by Blake (1993). The upper part equates to Unit 3 of Blake (1984a) and has been referred to as the Taylor Creek Sequence (Blake, 1993).

The lower part of the Hardey Formation, below the intra-basinal unconformity, comprises a basal conglomerate assemblage, overlain by sandstone and mudrock, in turn overlain by a felsic porphyry complex (Blake, 1990; 1993) which is here referred to as the Bamboo Creek Member.

The basal conglomerate assemblage comprises a laterally discontinuous, dominantly cobble to boulder conglomerate derived largely from the Mount Roe Basalt. The conglomerates, which are up to 370 m thick, were deposited in alluvial fans or gravel-dominated braided alluvial environments and lie adjacent to a major north-northeast trending lineament, the Nullagine Lineament (Fig. 5.1; Blake, 1984a, 1993). In the southern part of the sub-basin, this lineament approximates the western margin of the Hardey Formation outcrop.

The conglomerates are overlain by up to 600 m of dominantly terrigenous mudrock, medium- to very coarse grained sandstone, and felsic tuff that were deposited in braided alluvial and shallow, ephemeral lacustrine systems (Blake, 1993).

The overlying Bamboo Creek Member has previously been described as an intrusive body (Noldart and Wyatt, 1962; Hickman, 1983). However, Blake (1984a, 1990, 1993) recorded that the porphyry complex contains extrusive felsic rocks, including pyroclastic breccias, lapilli tuff containing blocks of porphyry, accretionary lapilli tuff, amygdaloidal porphyry, and probably substantial volumes of rheoignimbrite (see *Description* of **Felsic volcaniclastic facies** and **Massive quartz–feldspar porphyry facies**). In addition, the upper contact of the Bamboo Creek Member, which has been cited as evidence for intrusion (Hickman,1983), is now known to be an unconformity (Blake, 1984a, 1993).

The upper part of the Hardey Formation, corresponding to the Taylor Creek Sequence of Blake (1993), comprises a relatively thin basal succession dominated by terrigenous mudrock, overlain by very coarse grained quartzofeldspathic sandstone with subordinate pebble conglomerate and felsic tuff (Blake, 1993). Sandstone and conglomerate were deposited mostly in a braided fluvial environment and the basal mudrock and some turbidite sandstone were laid down in a lacustrine setting. On the western edge of the Nullagine Synclinorium, the basin margin is marked by a major north-northeasterly trending, west-block-down growth-fault system named the Conglomerate Creek Fault, which is associated with about 800 m of alluvial-fan and fan-delta conglomerate (Blake, 1993). Deep drilling has also located similar deposits in the lower Hardey Formation in this area, which suggests that the growth fault was active before emplacement of the felsic porphyry complex (Blake, 1993).

In the eastern Rooney Inlier on BALFOUR DOWNS, the Hardey Formation is up to 350 m thick. Here, this unit comprises a lower pebble to boulder conglomerate (alluvial fan and coarse-grained braided alluvial facies) overlain by very coarse grained sandstone with minor mudstone (sandy braided fluvial and lacustrine facies) and subordinate accretionary lapilli tuff (pyroclastic air-fall deposits). The larger clasts consist of felsic volcanic rock, basalt, granitoid, chert, sandstone, and vein quartz. Trough axes in braided fluvial sandstones record an easterly directed palaeoflow.

South Pilbara sub-basin

The Hardey Formation of the south Pilbara sub-basin outcrops in five areas: Wyloo Dome, Rocklea Dome, Bellary Dome, Milli Milli Dome, and the Sylvania Inlier. The formation has also been recorded from the FVG–1, WRL–1, and SGS–1, and BMW–1 diamond drillholes, located respectively on central ROY HILL, northeast MOUNT BRUCE, north-central MOUNT BRUCE, and southern MOUNT BRUCE (CRA Exploration Pty Ltd, 1985b, 1987b, 1988, 1989). In most of these areas the succession can be subdivided into two parts, a lower unit of very coarse grained quartzofeldspathic sandstone and minor conglomerate, and an upper unit of argillite, sandstone, basalt, and felsic and mafic tuff. In many parts of the south Pilbara sub-basin the upper unit is intruded by several thick, layered mafic sills.

Around Wyloo Dome, the formation is up to 550 m thick and is apparently conformable with the Mount Roe Basalt. Where it was possible to examine the contact directly, these rocks were found to rest upon a highly vesicular basalt flow top. In addition, no basalt-derived material has been reported from the Hardey Formation in this area (Seymour et al., 1988). The lower unit is about 100 m thick and comprises medium- to very coarse grained and pebbly quartzofeldspathic sandstone, and minor conglomerate and tuff (sandy braided fluvial facies). Palaeocurrent data from these rocks are directed toward the west-southwest (Fig. 5.6m). The upper unit is about 450 m thick and consists of poorly outcropping argillite and fine- to coarse-grained sandstone (deltaic facies).

In the Rocklea Dome, the Hardey Formation is up to 1800 m thick, rests disconformably upon the Mount Roe Basalt near Chillemarringa Well on the southeastern limb, but elsewhere is unconformable upon granite–greenstone basement. Near Chillemarringa Well, the lower Hardey Formation comprises 7–20 m of cobble to boulder conglomerate, which contains abundant clasts of vesicular basalt and granitoid, vein quartz, and chert. The conglomerate is overlain by 80 m of argillite and minor sandstone (lacustrine facies) interbedded with two, 10 m-thick, parallel-stratified conglomerate layers (coarse-grained braided alluvial facies). Conglomerate clast size and the proportion of basalt fragments decrease up-section.

Elsewhere on the Rocklea Dome, the lower unit of the Hardey Formation comprises about 100 m of coarse-grained to pebbly quartzofeldspathic sandstone (sandy braided fluvial facies and minor shoreline facies) with westerly directed palaeocurrents (Fig. 5.6l–m). The much thicker upper unit is interpreted as a deltaic assemblage and comprises argillite and fine-to coarse-grained sandstone interbedded with lenticular bodies of cross-stratified sandstone, tuff and lapilli tuff, and minor basalt. Palaeocurrent data from the sandstone bodies are toward the south-southwest and southwest. The upper unit of the Hardey Formation is intruded by up to four layered ultramafic to mafic sills and several dolerite sills; these are described in **Chapter 11**.

The Hardey Formation of the Bellary Dome has a maximum thickness of about 1400 m and rests with apparent conformity upon the Mount Roe Basalt. Where it was possible to examine the contact directly, the succession was found to rest upon vesicular basalt flow-top material. As is the case with the Wyloo Dome occurrences, no significant basalt-derived material was recorded in the formation. The lower trough cross-stratified sandstone unit (sandy braided fluvial facies) is generally 250–400 m thick and trough axes record a palaeoflow toward the southwest quadrant (Fig. 5.6h–j). The upper part of the succession is similar to the stratigraphically equivalent unit on Rocklea Dome. This part consists of up to 1100 m of argillite and sandstone (deltaic facies) interbedded with felsic and mafic tuff and lapilli tuff (primary and reworked pyroclastic fall and flow deposits) and basalt flows. Layered mafic sills and dolerite sills are a characteristic feature of the upper Hardey Formation in this area.

Throughout the Milli Milli Dome, the Hardey Formation rests unconformably on granite–greenstone basement. In the northeast, it is about 800 m thick and can be subdivided into a 400 m-thick lower unit of cross-stratified feldspathic sandstone and minor conglomerate (sandy braided fluvial facies), a middle 100 m-thick dolerite sill, and an upper mixed sandstone–argillite unit (sandy braided fluvial and deltaic facies). Palaeocurrent data from the braided fluvial facies are toward the south and west (Fig. 5.6e–g).

On NEWMAN, the Hardey Formation is equivalent to the combined lower metasedimentary unit, lower mafic volcanic unit, and felsic pyroclastic unit of Tyler (1986). The formation outcrops along the southwestern margin of the Sylvania Inlier and has a maximum thickness of at least 1500 m. The thickest section is in the Sandy Creek Well area and is dominated by deltaic facies and comprises a lower argillite and sandstone unit containing layered mafic sills; a middle unit of mafic tuff, tuffaceous sandstone, basalt, and mafic sills; and an upper unit of redeposited felsic tuff and argillite.

Data from the FVG–1, WRL–1, and SGS–1 diamond drillholes (CRA Exploration Pty Ltd, 1985a, 1987b, 1988) indicate that the Hardey Formation of the northern part of the south Pilbara sub-basin is much thinner than in the south. In SGS–1 the succession is 60 m thick and consists of a lower, basalt-derived conglomerate (?alluvial fan and coarse-grained braided alluvial facies) overlain by a composite unit of sandstone, argillite, and tuff (?sandy braided fluvial and pyroclastic air-fall deposits). The succession is 100 m thick in WRL–1 and unconformably overlies granitoid basement. Here, it comprises a lower conglomerate and sandstone unit containing abundant granitoid detritus, and an upper tuffaceous sandstone and argillite unit. In FVG–1, the formation is 75 m thick and consists of alternations of very fine to coarse-grained sandstone, conglomerate, argillite, and pisolitic (?accretionary lapilli) tuff.

Palaeogeography and depositional history

The principal characteristics of Hardey Formation sedimentation and volcanism on the northern Pilbara Craton have been summarized by Blake (1993) as follows:

- Depositional environments were dominantly continental, including braided alluvial, alluvial fan, and lacustrine.
- Sedimentation and possibly felsic volcanic centres were associated with dominantly north-northeasterly trending extensional structures, with consistent west-block-down normal movement.
- In the northeast and northwest Pilbara sub-basins, fluvial transport was predominantly parallel to associated north-northeasterly trending tectonic structures, except in the Northwest Oakover Syncline, where flow was toward the southeast. In the Marble Bar sub-basin, sedimentation trends varied with stratigraphic position but were either parallel to, or normal to, the main tectonic grain.
- Most of the non-volcanogenic arenites are texturally and compositionally immature and are derived from basement granitoid rocks.
- In the northwest Pilbara sub-basin, syn-Hardey Formation northeasterly or north-northeasterly trending structures are fairly evenly spaced over a distance of about 200 km with an averaged spacing of about 20 km.
- Relics of a major growth-fault system have been located in the north Pilbara with the following estimated syndepositional throws: Conglomerate Creek Fault in excess of 1000 m, Pear Creek Fault Complex 450–1300 m, Sherlock River Fault Complex in excess of 1500 m. Each major growth fault is subparallel to, and offset a similar distance southeast from, a major basement structure. The Conglomerate Creek Fault is about 20 km south-east of the Nullagine Lineament; the Pear Creek Fault is about 20 km southeast of the western main boundary strike-slip fault of the Lalla Rookh Basin (Krapez, 1984); and the Sherlock River Fault Complex is up to 17 km southeast of the Nunyerry Creek Fault System (Fig. 5.1). In the northeast and northwest Pilbara sub-basins, there is evidence that the Nullagine Lineament and the Nunyerry Creek Fault Complex were active during deposition of the Hardey Formation, but their influence on sedimentation patterns (where adjacent sedimentary rocks are preserved) is much less than nearby growth faults to the southeast.

In addition, the east-southeasterly trending Portland Fault (Kojan and Hickman, 2000) in the northwest Pilbara sub-basin also appears to have acted as a major Hardey Formation growth fault, with a syndepositional throw of about 1000 m.

In the south Pilbara sub-basin, Hardey Formation depositional environments evolved from braided fluvial to deltaic. As is the case in the sub-basins to the north, braided fluvial sandstones are compositionally and texturally immature and were probably derived from basement granitoids. Sediment transport directions in fluvial and deltaic distributary channels were principally toward the southwest and west and indicate the presence of a basement high to the north and northeast of the Milli Milli Dome. The greatly reduced thickness (<100 m) of the Hardey Formation in the FVG–1, WRL–1, and SGS–1 diamond drillholes (CRA Exploration Pty Ltd, 1985b, 1987b, 1988) and the absence of Hardey Formation from northern ROY HILL, further supports this view. In the Newman area, there is a marked northward thinning of the Hardey Formation across the Sylvania Inlier. This evidence, combined with the absence of thick subaqueous deltaic facies along the northern margin, suggests that the basement high to the north of the Milli Milli Dome may have extended close to the northern margin of the Sylvania Inlier.

With the possible exception of the Fortescue Valley Fault on NEWMAN, no syn-sedimentary growth faults have yet been identified in the south Pilbara sub-basin. Furthermore, attempts to locate such structures using thickness and palaeocurrent trends, and sedimentary facies distributions, are hampered by the limited outcrop of the Hardey Formation in this area. As noted earlier, the available stratigraphic and sedimentological data suggest the presence of a west-northwesterly trending palaeohigh that extended from the Sylvania Inlier to the headwaters of the Yule River on PYRAMID — the Yule–Sylvania High of Blake (1993). The orientation of this palaeohigh is parallel to the main pre- and post-Fortescue Group tectonic grain in the southern Pilbara, and the possibility exists that some of these structures were active during deposition of the Hardey Formation and overlying parts of the Fortescue Group (Thorne, 1990; see also **Chapters 6, 7, 8, 9**).

A major drawback with such an inferred west-northwesterly trending active fault system in the southern Pilbara during deposition of the Hardey Formation is that it is difficult to reconcile with the well defined north-northeasterly trending growth-fault system of the northern Pilbara Craton. Assuming growth faults are an integral part of Hardey Formation development in the southern Pilbara, Blake (1993) suggests the following interpretations are possible:

1. The development of the Hardey Formation in the southern Pilbara was associated with west-northwesterly striking south-block-down normal faults, but is significantly younger than the Hardey Formation of the northern craton. If correct, this would suggest that the Hardey Formation of the southern Pilbara was more closely linked tectonically to the overlying Mount Jope (Boongal to Bunjinah Formations) succession.
2. The development of the southern Hardey Formation was associated with north-northeasterly trending normal faults and is essentially coeval with the Hardey Formation in the north. The southward, down-palaeocurrent thickening trend in the south Pilbara sub-basin reflects along-basin thickening away from the Yule–Sylvania High (Fig. 5.15).

Although there is currently no geochronological evidence to support or refute argument 1, it is considered suspect in view of the similar depositional histories and common source areas for the lower Hardey Formation in the northern and southern Pilbara. On the other hand, a major objection to argument 2 is the limited evidence for

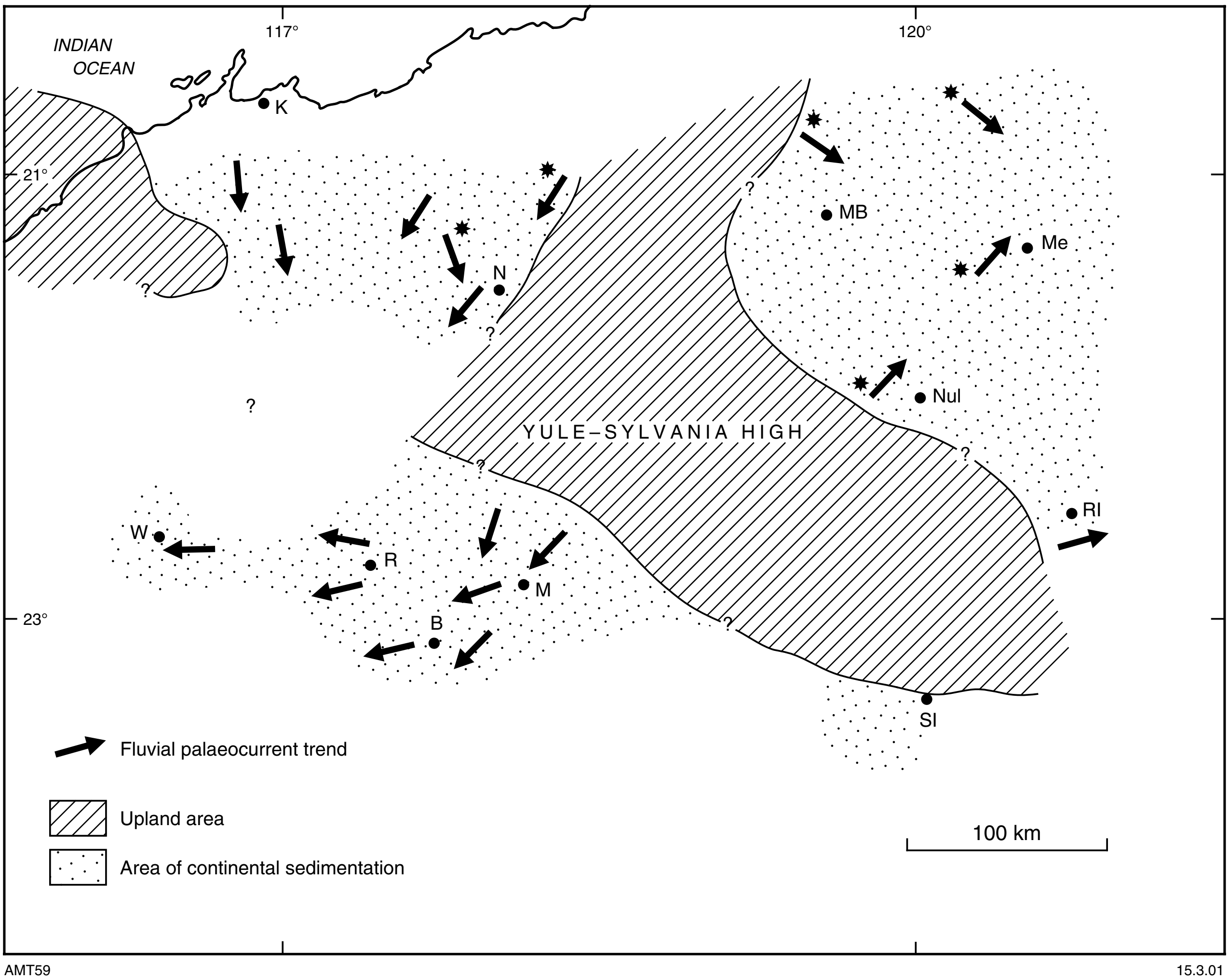

Figure 5.15. Palaeogeography of the Pilbara during deposition of the lower Hardey Formation. Localities shown for reference are: B = Bellary Dome, K = Karratha, MB = Marble Bar, Me = Meentheena, M = Milli Milli Dome, Nul = Nullagine, N = Nunyerry, R = Rocklea, RI = Rooney Inlier, W = Wyloo Dome. Palaeocurrent data marked with * are from Blake (1993)

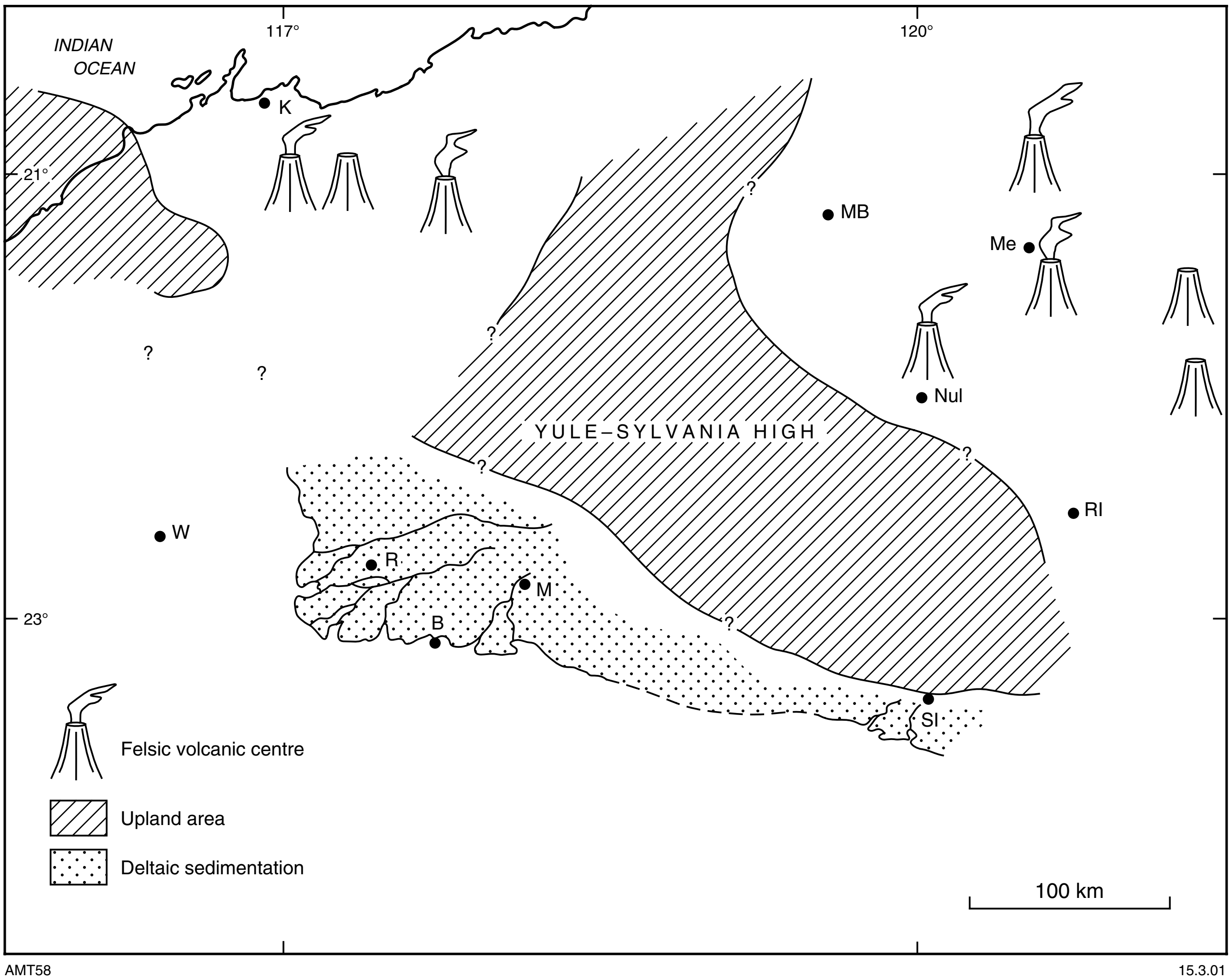

Figure 5.16. Palaeogeography of the Pilbara during deposition of the middle to upper Hardey Formation. Localities are as shown for Figure 5.15

north-northeasterly trending growth faults in the South Pilbara sub-basin. Although the Fortescue Valley Fault in the Sylvania Inlier appears to have acted as a west-block-down normal fault throughout Fortescue Group deposition, no other fault of this type has been identified in the southern Pilbara. This may be because the relatively small area of lower Fortescue Group and basement outcrop in the southern Pilbara limits the chances of identifying such structures. An alternative explanation is, however, that the north-northeasterly trending fault system of the northern Pilbara did not extend southwards across the Yule–Sylvania High. Some support for this view comes from Hardey Formation fluvial palaeocurrent trends in the southern Pilbara which are toward the southwest and west, and are therefore not entirely consistent with an active, north-northeasterly trending growth-fault system.

To summarize, while there is clear evidence for a north-northeasterly trending growth-fault system in the northern Pilbara, these structures are generally not recognized in the southern part of the craton. Similarly, although it seems likely that a west-northwesterly to northwesterly trending basement high existed in the central Pilbara during Hardey Formation deposition, most of the evidence for an active growth-fault system parallel to its southern margin is based on indirect evidence and is discussed further in **Chapter 14**.

Figure 5.16 is a generalized palaeogeographic reconstruction of the preserved portion of the Pilbara Craton during deposition of the lower Hardey Formation. Thickness, facies, and palaeocurrent data point to the presence of a major, south to southwesterly flowing sandy braided fluvial system in the northwestern part of the craton. Here, alluvial fan and gravelly braided rivers were associated with small palaeohighs and north-northeasterly trending fault scarps in the braid-plain. The zero isopach for the Hardey Formation suggests that one of the larger topographic highs extended southeast from Cape Preston.

The southeastern margin of this northwest Pilbara sub-basin was marked by the Nunyerry Creek – Sherlock River Fault System, and to the southeast, a major palaeoridge extended toward the present day Sylvania Inlier. In the Marble Bar district to the north of the ridge, sandy braided rivers flowed mostly from the north and southeast. Farther to the east and southeast in the

Meentheena and Nullagine districts, sandy braided streams flowed toward the north-northeast. Southwest of the palaeohigh, in the Wyloo–Milli Milli area, braided streams flowed toward the west and southwest.

Deposition of the middle and upper Hardey Formation was characterized by increased levels of felsic volcanism in the northwest and northeast Pilbara (Fig. 5.16). In the central northwest Pilbara sub-basin, three volcanic centres were developed adjacent to, and to the west of, major north-northeasterly trending normal faults and lineaments (Blake, 1993). Here, the volcanic centres were active during the remainder of Hardey Formation deposition. In the northeast Pilbara sub-basin, the main period of felsic volcanism was followed by significant uplift and erosion and, finally, by a return to continental siliciclastic sedimentation. Levels of felsic volcanism were significantly less in the southern Pilbara. Here, the lower Hardey Formation braid-plain evolved into a fluvial-dominated deltaic complex, which persisted until the close of Hardey Formation times.

Chapter 6

Kylena and Boongal Formations

The Kylena Formation or its lateral equivalent, the Boongal Formation, is present over much of the southern and central Pilbara Craton and in the Gregory Range, but is absent from the Marble Bar sub-basin, the Sylvania Inlier, and a small area of the central Chichester Range on ROY HILL (Fig. 6.1). In the southern Sylvania Inlier, these rocks are equivalent to the lower part of the upper mafic volcanic unit of Tyler (1986).

In the northern part of the Pilbara Craton, the Kylena Formation is typically less than 600 m thick. It rests either conformably or unconformably upon the Hardey Formation, or unconformably upon granite–greenstone terrane. In the northwest Pilbara sub-basin, no angular discordance was observed between the Kylena Formation and the Hardey Formation although there is some evidence that this contact could be a nondepositional unconformity (Blake, 1993). In the Nullagine Synclinorium of the northeast Pilbara sub-basin, the boundary with the Hardey Formation is an angular unconformity (Blake, 1993). Elsewhere, however, according to Blake (1993), this contact is either a disconformity or a nondepositional unconformity. Blake's (1993) reconstruction of Hardey Formation depositional systems around the anticline between the Meentheena Centrocline and the Northwest Oakover Syncline suggested that hundreds of metres of Hardey Formation were eroded prior to deposition of the Kylena Formation. The northeast Pilbara Basin, therefore, appears to have been gently folded and the cores of the major anticline eroded before deposition of the Kylena Formation. In the south Pilbara sub-basin, the Boongal Formation is up to 1200 m thick, and the contact between this unit and the underlying Hardey Formation appears to be conformable.

The older age limit of the Kylena Formation (**Chapter 2**) is constrained by U–Pb zircon dates from the underlying Hardey Formation which lie within the range 2768 ± 16 to 2756 ± 8 Ma (Pidgeon, 1984; Arndt et al., 1991). The younger age limit for the Kylena Formation is fixed by a U–Pb age of 2715 ± 6 Ma from the youngest zircon population in a sample from the Tumbiana Formation (Arndt et al., 1991) and also by a U–Pb zircon date of 2717 ± 2 Ma from a rhyolite in the Maddina Formation (Nelson, 1998; Kojan and Hickman, 1998). Wingate (1999) suggested that 2747 ± 4 Ma north-northeasterly trending mafic dykes in the Sylvania Inlier may have acted as feeders to the Kylena Formation; however, these dykes are not seen in contact with any part of the Fortescue Group.

Previous descriptions of the Kylena and Boongal Formations are given by: Daniels (1970), Hickman (1978, 1983), Hickman et al. (1983), Horwitz (1976), Kojan and Hickman (1998), Kriewaldt (1964a,b), Kriewaldt and Ryan (1967), Seymour et al. (1988), Thorne et al. (1991) Thorne and Tyler (1994, 1996, 1997a,b), and Williams (1989). The formation has also been described under various other formal and informal names (Blake, 1984a, 1993; Noldart and Wyatt, 1962; Trendall, 1975b, 1990b). This chapter describes and interprets the principal volcanic and sedimentary facies associations present in the Kylena and Boongal Formations. Most of the data have been obtained from five measured sections in the northwest and northeast Pilbara sub-basins and 1:40 000-scale mapping of selected outcrop areas in the south Pilbara sub-basin.

The Kylena and Boongal Formations consist mainly of basalt flows, pillow lava, fine- to coarse-grained mafic volcaniclastic rocks, and sedimentary carbonate rocks.

Volcanic facies

Volcanic facies can be subdivided into subaerial basaltic lavas, subaqueous basaltic lavas, and volcaniclastic rocks.

Subaerial basaltic lavas

Physical volcanology: Subaerial basaltic flows are the dominant volcanic facies within the Kylena Formation. Flows range in thickness from less than a metre to 60 m. Average flow thickness and number of flows in each measured section are as follows: Cape Preston (7.76 m, 30), Harding River (10 m, 25), Mount Herbert area (32 m, 5), Sherlock River (18.8 m, 12), Newman–Port Hedland Road (35 m, 1), and Meentheena (14.7 m, 41). Two broad categories of basaltic flow can be recognized on the basis of thickness and flow structure: thin amygdaloidal flows, and massive to amygdaloidal flows.

Thin (commonly <5 m) amygdaloidal flows have been recorded from most localities in the east and west Pilbara but are less abundant in the central Chichester Range. Most amygdaloidal flows contain 10–20% of spherical amygdales in lower parts of the flow, whereas upper levels are characterized by a much greater proportion (30–65% of the rock) of large, spherical to irregularly rounded or streaked amygdales. Amygdales typically range in size up to about 30 mm and are either scattered randomly

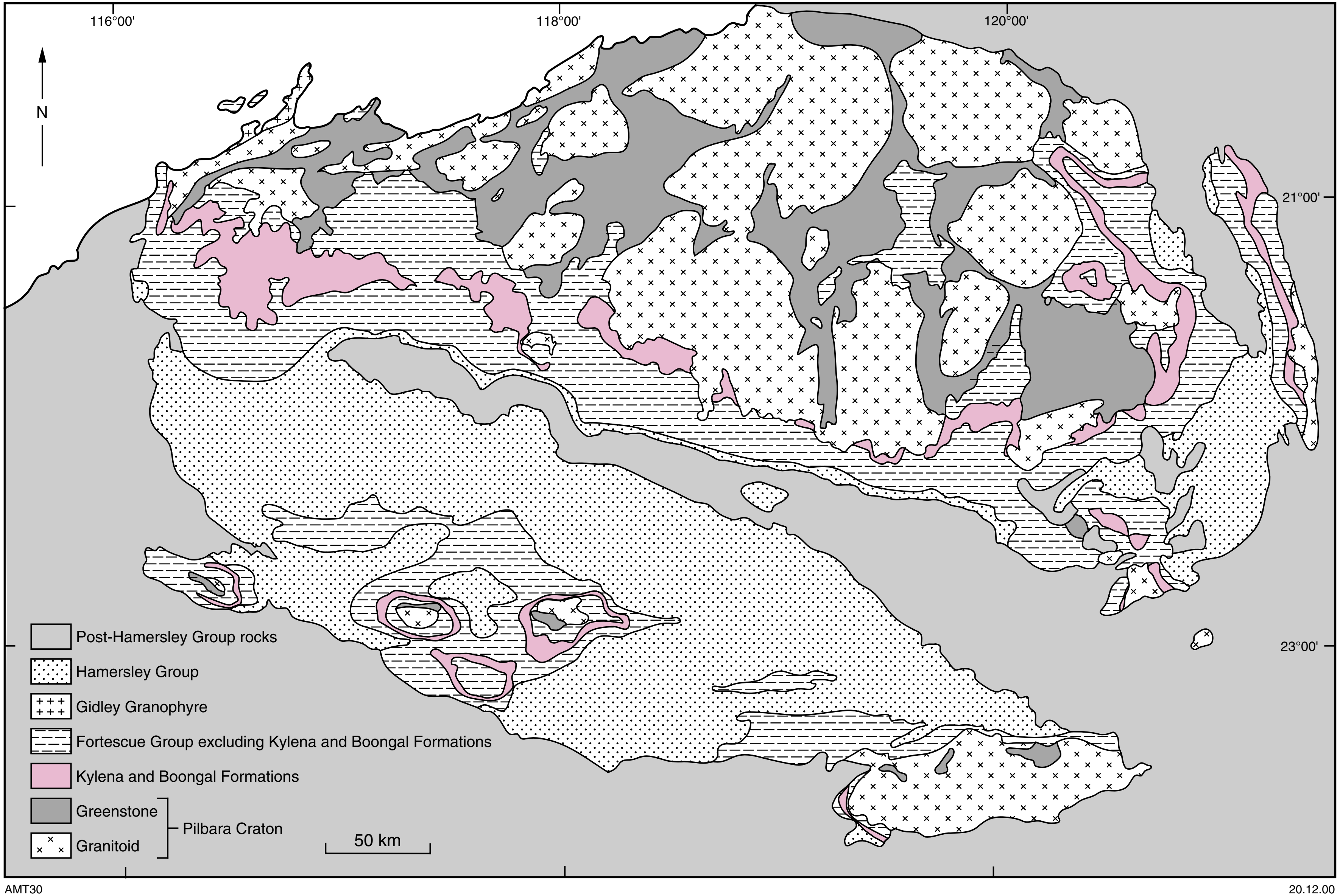

Figure 6.1. Principal outcrop areas of the Kylena Formation and Boongal Formation. In Wyloo Dome, the Boongal Formation has not been separated from the rest of the post-Hardey Formation Fortescue Group stratigraphy (cf. Seymour et al., 1988)

throughout the rock or concentrated in flow-aligned clusters. Most amygdales are filled by quartz, carbonate, and chlorite.

Thin amygdaloidal flows are bounded by parallel or sub-parallel flow surfaces, which exhibit gentle, irregular undulations. Most flows appear to be laterally discontinuous over distances exceeding 2 km. Flow tops typically exhibit irregular to broadly symmetrical undulations. Wavelength and amplitude vary with flow thickness but most lie in the range 0.5–5 m and 0.1–1 m respectively. Flow tops are rarely scoriaceous. Flow bases are generally smooth and follow irregularities in the underlying surface. No systematic jointing was observed in these flows.

Massive to amygdaloidal flows range in thickness from about 5 to 70 m. They have been recorded from all measured sections and are the dominant flow type in the central Chichester Range. Most massive to amygdaloidal basalts are characterized by a high proportion (20–65% of the rock) of large spherical to irregularly rounded or streaked amygdales in upper parts of the flow (Fig. 6.2). These upper levels contrast with lower and middle parts of the flow, which commonly contain only a few percent of amygdales. Most amygdales range in size up to about 30 mm and occur either scattered randomly throughout the rock or concentrated in folded, flow-aligned clusters or bedding-parallel sheets. The majority of amygdales are filled by various combinations of quartz, carbonate, and chlorite.

Massive to amygdaloidal flows are bounded by tabular flow surfaces which exhibit gentle, irregular undulations at outcrop scale. The thickest flows (>20 m) are laterally persistent over distances of 1 to at least 5 km; the lateral extent of thinner (<20 m) flows is unknown. Flow tops generally exhibit irregular to broadly symmetrical undulations. Wavelength and amplitude vary with flow thickness but lie typically in the range 0.5–15 m and 0.2–2 m respectively. Locally, the upper flow surface displays a network of interconnected ridges up to 0.2 m high. Flow tops are commonly scoriaceous and evidence of autobrecciation is also present locally. Flow bases tend to be smooth and mimic irregularities in the underlying surface. More rarely, the lower parts of the flow comprise a hyaloclastite breccia which is transitional upwards into massive or amygdaloidal basalt.

Most massive and amygdaloidal basalts show little systematic jointing, but columnar joints and irregular joint networks have been observed locally. Columnar joints have been recorded only in flows thicker than 15 m. Individual columns are mostly 0.15–0.3 m in diameter and are present in lower and middle to upper parts of the flows; column axes are generally oriented 45–90° to the flow surfaces. No examples of the collonade–entablature structure (Long and Wood, 1986) were observed. Locally, the thick massive to amygdaloidal flows are cross-cut by an irregular network of curved and planar joint surfaces. These are observed in plan and cross section and give the rock an irregular, net-veined appearance. Joint surfaces commonly display a plumose fracture pattern and there

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Figure 6.2. Amygdaloidal flow top, massive to amygdaloidal basalt flow, Kylena Formation, Harding River, northwest Pilbara sub-basin

is no evidence of significant relative movement between adjacent 'fragments'. Very thick, acicular-textured basalt flows may display discontinuous partings aligned parallel to flow boundaries.

Petrography: Massive to amygdaloidal flows are typically fine to coarse grained (50–2000 μm) and aphyric with intersertal and intergranular textures (Fig. 6.3a). Equant to prismatic, albitized plagioclase is abundant within the groundmass, and lath-like to prismatic varieties normally define a random or, less commonly, a pilotaxitic texture. Equant to tabular, anhedral to subhedral pyroxene occurs in all samples and is most commonly replaced by tremolite, chlorite, and serpentine. Interstitial material is altered to a variety of minerals including tremolite, chlorite, carbonate, pumpellyite, quartz, serpentine, and epidote.

Thick, massive to amygdaloidal flows on PYRAMID are very coarse grained and may display a lower cumulate zone comprising a framework of equant to platy clinopyroxene crystals, up to 10 mm long, enclosing altered intersertal pyroxene, plagioclase, quartz, and iron oxides. Other coarse-grained flows are characterized by a framework of equant to bladed plagioclase and clinopyroxene, separated by graphic intergrowths of plagioclase and quartz, with local pyroxene (Fig. 6.3b).

Summary of geochemistry: Most (70) of the 84 available analyses of rocks of this facies come from the northeast Pilbara sub-basin, and form the KYHC (Hays Creek), KYMT (Meentheena), KYPP (Pelican Pool), and KYMB (Marble Bar) groups (all cluster F) in terms of the scheme used in **Chapter 12**. These analyses were carried out as part of the Pilbara Volcanic Study (see **Chapter 1**). The remaining 14 analyses come from the west Pilbara sub-basin and form the KYGS (Gudrun Sieber) group. The average compositions of all five of these groups appear in Table 12.2.

The analyses from the northeast Pilbara sub-basin have little spread (Fig. 12.2) and lie close to the basaltic andesitic composition (Le Maitre, 1989) that characterizes all of the stacked mafic lavas of the Fortescue Group. Their average silica content, close to 57% (Fig. 12.6), is significantly higher than that of typical Phanerozoic flood basalts, and on the Jensen Cation Plot (Jensen, 1976; Fig. 12.7), the analyses are clustered across the division between the calc-alkaline and tholeiitic fields. They are also gradational into Jensen's high-iron tholeiitic basalt. The analyses from the west Pilbara sub-basin have a significantly higher average content of silica, and lower total alkalies, but nevertheless also fall within the basaltic andesite field in the nomenclature of Le Maitre (1989). Whether the different chemical composition of the west Pilbara samples represents a systematic regional compositional trend remains an open question. As in all other mafic lava facies whose major-element content approximates to the basaltic andesite composition typical of the stacked mafic lavas of the Fortescue Group, there is a strong enrichment of incompatible trace elements relative to MORB (Fig. 12.9). This is a common feature of within-plate continental flood basalts. However, in terms of the Ti–Zr–Y plot of Pearce and Cann (1973; see Fig. 12.11), these lavas fall in their calc-alkali field (C) rather than the within-plate field (D). Trace element enrichment relative to MORB is particularly strongly present in all four groups of the F cluster; inspection of Figure 12.2 shows this with particular clarity for La, Y, Ce, and Zr, underlining the possibility of significant regional (east–west) differences in the compositions of erupted magmas.

Interpretation: The basaltic composition and aphyric texture of most of the lavas suggest that they originated as low- to moderate-viscosity flows that were close to their liquidus temperatures at eruption (cf. Green, 1989). In addition, the general lack of evidence for contemporaneous explosive fragmentation suggests either that the initial content of exsolved volatiles was low, or that levels were high, but were reduced by non-explosive degassing or by volatile loss during a previous phase of explosive activity (Cas and Wright, 1987).

Very thick (>40 m), columnar-jointed, massive to amygdaloidal flows are similar in thickness, internal structure, surface features, and petrographic character to many flows in the Columbia River Basalt (Waters, 1961; Long and Wood, 1986). Their irregular columnar structure, lacking entablature, most closely resembles the Type 1 flows of Long and Wood (1986) although their thickness (up to 70 m) is greater, and column diameter (<0.3 m) is smaller than the Columbia River examples. By analogy with these North American lavas, the columnar-jointed massive to amydaloidal flows of the Kylena Formation are interpreted as rapidly emplaced, subaerial, ponded lava flows. Other massive to amygdaloidal flows in the Kylena Formation are often very thick but do not show evidence of columnar jointing. It is likely that these very thick basalts also formed ponded flows but did not develop the necessary thermal stress pattern for joint formation (Spry, 1962; Degraff and Aydin, 1987).

It is not known to what extent many of the thinner (5–20 m) massive to amygdaloidal lavas in the Kylena Formation represent ponded flows. On the one hand, these basalts are thinner and have a more varied morphology than typical ponded flows, but on the other, they lack many of the features associated with flowing lava, such as lava tubes and surface channels.

Subaqueous basaltic lavas

Physical volcanology: Subaqueous basaltic lavas are the dominant volcanic facies within the Boongal Formation. They comprise pillow lava and massive basalt flows and are intimately associated with hyaloclastite. Individual pillows are up to 2.5 m wide and have a sac-shaped or elongate, finger-like morphology (Fig. 6.4a). Many contain spheroidal, arcuate, or planar convex cavities up to 0.2 m wide, and radial pipe amygdales are observed locally. Pillows have an altered glassy selvage up to 20 mm thick and interpillow cavities are filled by variable amounts of quartz, carbonate, and hyaloclastite.

Massive basalt flows range in thickness from 5 m to over 100 m and have undulatory to planar bounding surfaces. Many flows can be traced laterally and vertically into pillow lava and hyaloclastite. Most flows are either non-vesicular or sparsely vesicular; in the latter situation,

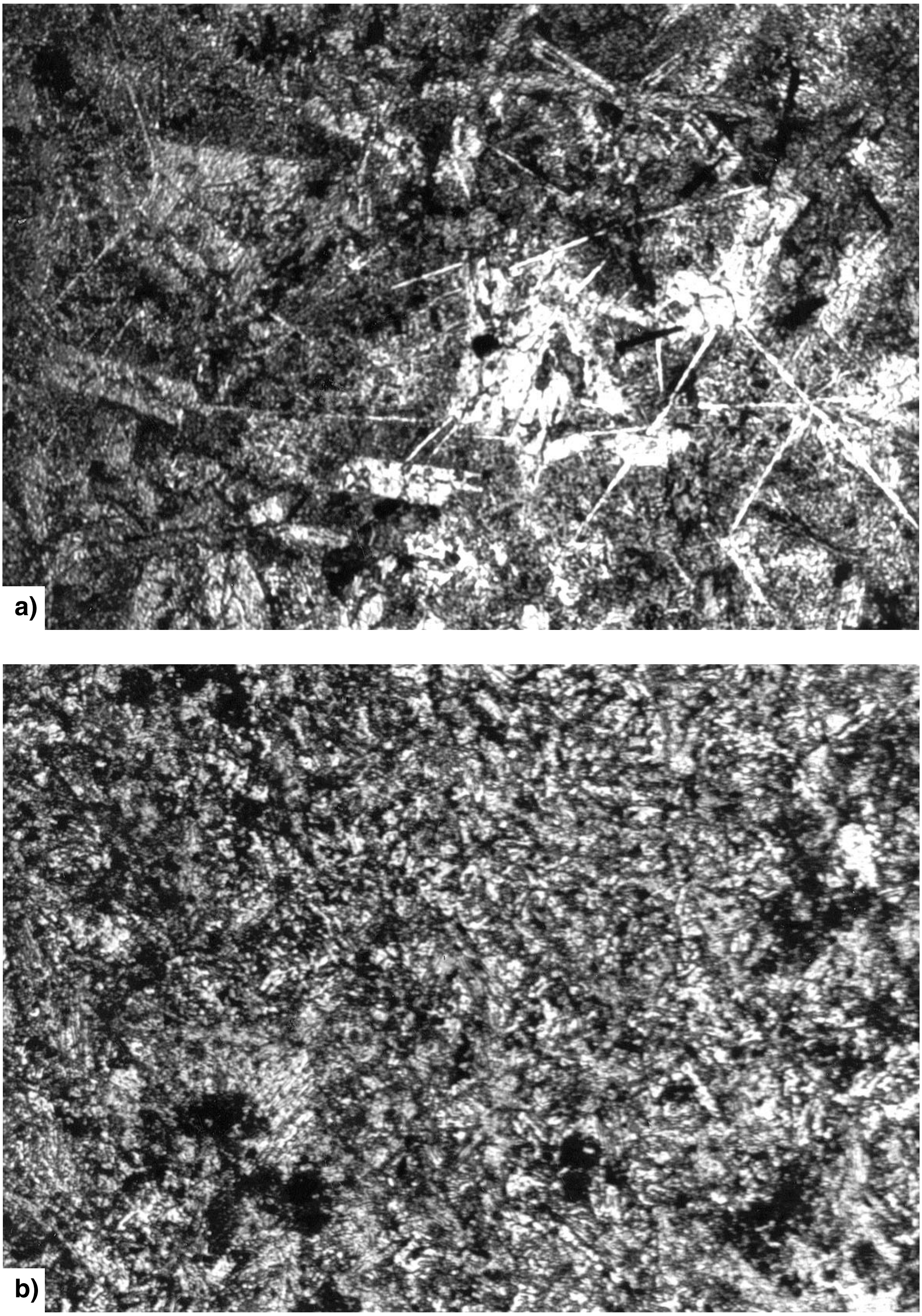

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Figure 6.3. a) Photomicrograph of massive to amygdaloidal basalt showing randomly oriented prismatic and acicular, albitized plagioclase, Kylena Formation, northwest Pilbara sub-basin. Plane polarized light, field of view = 3.5 mm; b) Photomicrograph of aphyric massive basalt, Bunjinah Formation, south Pilbara sub-basin. Plane polarized light, field of view = 3.5 mm

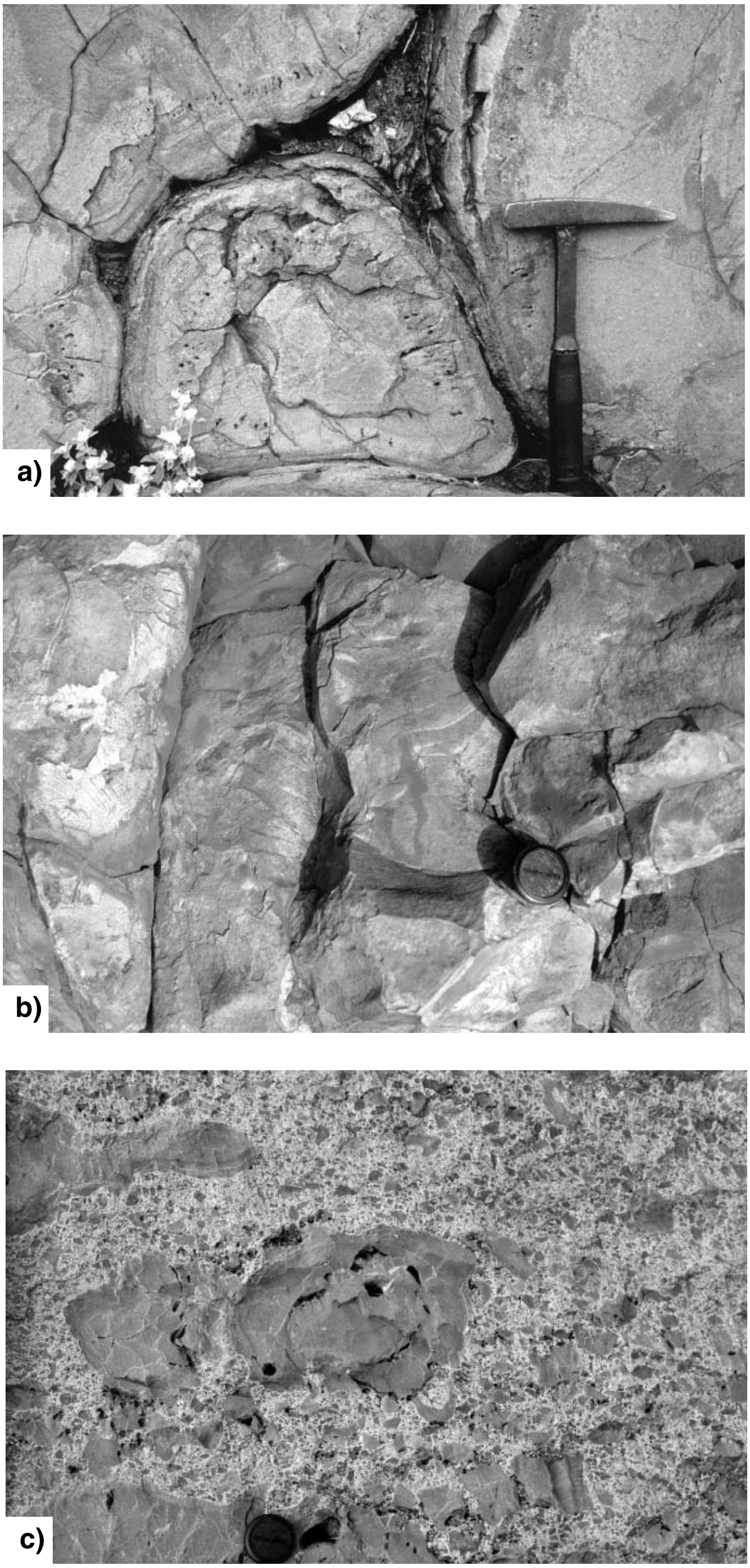

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Figure 6.4. a) Basaltic pillow lava showing radial amygdales, outer selvage, and inter-pillow material, Boongal Formation, eastern Wyloo Dome, south Pilbara sub-basin. Photograph by D. B. Seymour; b) Fracture surfaces in irregularly jointed massive basalt, Boongal Formation, Milli Milli Dome, south Pilbara sub-basin; c) Pillow fragment in massive basaltic breccia, Boongal Formation, Rocklea Dome, south Pilbara sub-basin

amygdales are filled by combinations of quartz, carbonate, and chlorite. Thick massive flows are cut by a system of irregular joint planes aligned parallel to, or normal to, flow-bounding surfaces. Locally, flows are cross-cut by an irregular network of curved and planar joint surfaces (Fig. 6.4b) similar to those observed in some thick subaerial lavas. These joints give the rock an irregular, net-veined appearance on bedding sufaces and in cross section.

Petrography: Most samples of subaqueous facies basalt are aphyric and fine to medium grained (50–500 µm). They consist of a framework of ragged, acicular to bladed actinolite–tremolite separated by anhedral to subhedral epidote–clinozoisite, quartz, albite, chlorite, calcite, and iron oxide. The effects of low-grade metamorphism make it difficult to distinguish original intersertal and intergranular textures.

Geochemistry: Only six analyses of this facies from the Boongal Formation of the south Pilbara sub-basin are available, compared with the 84 of subaerial basaltic lava from the northern outcrop area of the Kylena Formation. They form the BOBC (Bellary Creek) and BOCA (Cowley area) groups (part of cluster D) in terms of the scheme used in **Chapter 12**. Although the average silica contents (Table 12.2) of both groups are slightly higher than those of the subaerial basaltic facies (except the KYGS group), they also lie close to the basaltic andesitic composition (Le Maitre, 1989) that characterizes all of the stacked mafic lavas of the Fortescue Group. There is no evidence from the available analyses that the magma whose subaqueous eruption formed the lavas of the Boongal Formation differed in any significant respect from that erupted to form the subaerial flows farther north.

Interpretation: The association of pillow lava, massive lava, and hyaloclastite indicates that much of the Boongal Formation was formed in a subaqueous environment. By analogy with modern subaqueous flows (Moore et al., 1973), most pillow lavas in the Boongal Formation formed by the budding of subaqueous lava tubes, a process that is comparable to the digital advance of subaerial pahoehoe. Discrete pillows are also present in the Boongal Formation and probably formed on steeper slopes where individual lava toes became detatched from the rest of the tube. Ballard et al. (1979) reported the occurrence of massive sheet flows of basaltic lava from the Galapagos Rift Valley. These flows are associated with pillow lava and the transition from pillowed to massive lithology probably reflects an increase in basalt discharge rate.

Pillow lavas have been reported from three principal environments: marine, sub-glacial, and lacustrine. In the case of the Boongal Formation, deposition in a sub-glacial setting can be ruled out on the basis of the thickness and lateral extent of the lava units and the absence of other evidence for a glacial event. A lacustrine setting for the pillow lava is also thought to be unlikely as there are no associated subaerial facies in the Boongal Formation or overlying parts of the Fortescue Group. In addition, the lateral extent of subaqueous facies would neccessitate the former presence of a lake covering an area of at least 50 000 km^2. In view of these constraints, it is thought most likely that the Boongal Formation was deposited in a submarine environment.

Volcaniclastic rocks

Description: Volcaniclastic rocks comprise hyaloclastite and lapilli tuff. In most outcrops these rocks are subordinate to basalt flows and pillow lava, although hyaloclastite may be the dominant rock type locally within the Boongal Formation.

Basaltic hyaloclastite is typically associated with pillow lava and massive lava in the Boongal Formation but is also recorded beneath subaerial flows in the Kylena Formation. In the former situation, the hyaloclastite occurs in laterally discontinuous units, ranging from a few metres to over 100 m thick. Most hyaloclastites are massive, very poorly sorted breccias composed of angular fragments (≤0.5 m across) of sparsely vesicular to vesicular basalt. Others are finer grained (sand to granule sized) and may exhibit weak stratification. Clasts of all sizes are cut by networks of polygonal fractures that vary in width from hair-like cracks to wider veins filled by quartz, chlorite, and carbonate. Even when separated by a few centimetres, many clasts exhibit a jigsaw fit with neighbouring fragments. Detatched pillows or pillow fragments are observed locally in hyaloclastites within the Kylena and Boongal Formations (Fig. 6.4c).

Beds of basaltic lapilli tuff are present locally in association with subaerial and subaqueous facies. Descriptions by Hickman (1983) and Blight (1985) suggest that some beds contain accretionary lapilli.

Interpretation: The basaltic breccias are interpreted as hyaloclastites, produced by the rapid cooling and fracturing of basalt during contact with water. This interpretation is based upon the following evidence:

- Their homogeneous, generally non-vesicular clast composition.
- The predominance of poorly sorted, angular clasts and evidence for fracturing in situ.
- Their intimate association with pillow lava.

Hyaloclastite breccias are produced when there is sudden contact between hot, coherent magma and cold water or water-saturated sediment (Cas and Wright, 1987). They form in a variety of situations: where magma is erupted subaqueously or subglacially; where lava flows into water or over water-saturated sediments; or where magma is intruded into water-saturated sediment or country rock. In the case of the breccias described here, geological setting rules out an intrusion-related or subglacial origin. Localized accumulations of hyaloclastite at the base of some subaerial Kylena Formation flows were probably formed by lava either flowing into shallow water or over wet sediment. In most cases, however, hyaloclastite occurs in association with pillow lava and massive lava in the Boongal Formation. For these units the only interpretation that satisfies all the available evidence is the one in which the breccias formed in a marine environment. The lack of stratification in most hyalocalastites, combined with their matrix support, poor sorting, angular clasts, and presence of transitional boundaries with the basalt flows suggests that they accumulated close to source, either as a flow-top mantle, or marginal scree and sediment gravity-flow deposits.

It is not known whether the mafic tuff and lapilli tuff beds are primary or reworked pyroclastic deposits. Those that contain accretionary lapilli probably originated from subaerial hydroclastic eruptions.

Non-basaltic volcanic rocks

A variety of non-basaltic volcanic rocks has been reported from the Kylena Formaion in the west and east Pilbara.

Williams (1968, 1999) recorded local occurrences of felsic lava on YARRALOOLA, and andesite and possible andesitic dacite on MUCCAN.

During their mapping on PINDERI HILLS, Kojan and Hickman (1998) noted that a high proportion of the Kylena Formation consisted of high-Mg andesite with andesite, dacite, and rhyolite. The andesite and high-Mg andesites contain plagioclase and clinopyroxene with interstitial quartz and K-feldspar, whereas the dacite and rhyolite consist of plagioclase intergrown with K-feldspar and granular quartz. Dacite is typically granophyric, whereas rhyolite commonly exhibits spherulitic or snowflake textures. All rocks show varying degrees of propylitic alteration with development of chlorite, carbonate and leucoxene, and more rarely quartz and prehnite. Sulfides, including chalcopyrite and sphalerite, were noted in several samples of dacite and rhyolite from the upper part of the Kylena Formation.

Geochemical analysis of these rocks (Kojan and Hickman, 1998) supports the petrographic data and indicates a compositional range from basalt to rhyolite, with many samples plotting very close to the shoshonite and trachyandesite fields. The Kylena Formation lavas are medium- to high-K, and show mixed calc-alkaline and tholeiitic affinities. In addition, absolute abundances of Ba, Ce, La, Rb, and Th are much greater than for most Archaean basaltic rocks. The andesites contain 45.4–49.9 ppm Ni, 36.5–87.0 ppm Cr, 208.3–237.4 ppm V, and 3.96–4.65% MgO. High-Mg andesite contains higher concentrations of MgO, Ni and Cr, and lower V.

Non-volcanic sedimentary facies

Description: Non-volcanic sedimentary rocks form less than 1% of the Kylena Formation and have not been recorded from the Boongal Formation. They are most abundant in the Meentheena Centrocline and consist of thin (<3 m) units of partially silicified stromatolitic carbonate and volocaniclastic debris. These units overlie subaerial basalt flow tops and may be subdivided locally (Fig. 6.5) into: a 0.5 m-thick lower unit of irregular, horizontally laminated carbonate; a middle unit of small (700 mm high by 1000 mm wide), recrystallized domical stromatolites; and an upper unit of 0.1 – 0.4 m bedded, chert carbonate breccia and recrystallized micro-columnar stromatolites. Thin layers of peloidal grainstone with abundant siliciclastic detritus are also present in this upper division.

Interpretation: Carbonate units which display the previously described three-fold subdivision of rock types are broadly similar to the upward-shallowing carbonate cycles reported from many Precambrian sedimentary basins (White, 1981; James, 1984; Thorne, 1985; Grotzinger, 1986). In such cases the vertical variation in rock type results from initial transgressive, and subsequent regressive, shoreline sedimentation. Using this model as a basis, the lower planar laminated unit was probably laid down in a subaqueous environment, during and immediately after the initial transgression. The overlying units formed during the following regressive phase when,

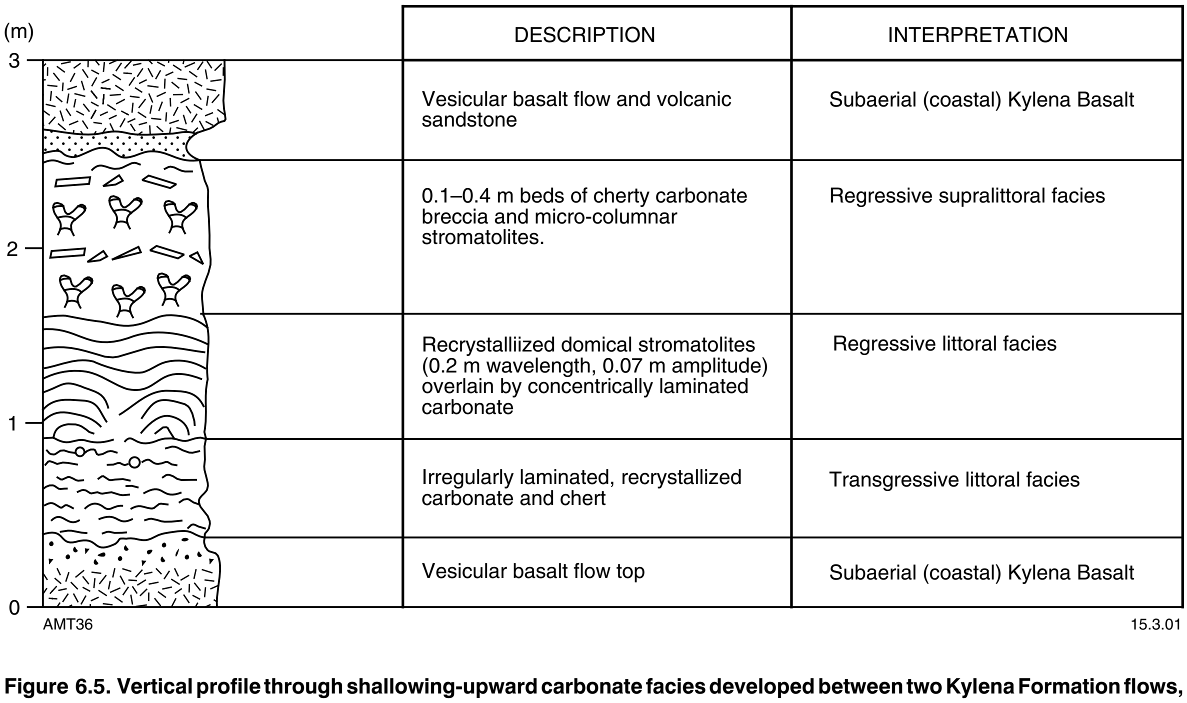


Figure 6.5. Vertical profile through shallowing-upward carbonate facies developed between two Kylena Formation flows, Meentheena Centrocline, northeast Pilbara sub-basin

firstly, littoral facies (domical stromatolites), and then supralittoral facies (carbonate breccia and microcolumnar stromatolites), were deposited.

Distribution of facies

Northwest Pilbara sub-basin

The Kylena Formation forms a semicontinuous outcrop from Cape Preston to the eastern margin of the northwest Pilbara sub-basin. It is also recorded from the MF–1 and SV–1 diamond drillholes on eastern PYRAMID and east-central YARRALOOLA (CRA Exploration Pty Ltd, 1985a, 1987a).

Subaerial basalt flows, interbedded locally with thin tuff, are the dominant facies in all outcrops studied. The thickest sections, some 200–300 m thick and comprising 20–30 flows, lie in the western part of the sub-basin. In eastern outcrops, the Kylena Formation is not only thinner but made up of fewer (2–12) flows. An exception to this occurs in the Sherlock River area, where 230 m of Kylena Formation have been recorded from the SV–1 diamond drillhole (CRA Exploration Pty Ltd, 1985a). Many flows are over 25 m thick and very coarse grained (e.g. at Python Pool) and some display a basal division of coarse-grained pyroxene cumulate, such as those near the Sherlock River Fault System (MOUNT BILROTH AMG 887010).

On PINDERI HILLS the Kylena Formation has a maximum thickness of 520 m (Kojan and Hickman, 1998). The formation comprises a lower unit, with low radiometric values, which is composed of basalt and high-Mg andesite with minor andesite and dacite, and an upper unit, with high radiometric values, which consists of dacite and rhyolite. Basalt, basaltic andesite, and andesite are encountered beneath the Tumbiana Formation, south of the Fortescue River, in the southwest corner of PINDERI HILLS (Kojan and Hickman, 1998). Three hundred and fifty metres of Kylena Formation have been recorded in the MF–1 drillhole (CRA Exploration Pty Ltd, 1987a).

Marble Bar sub-basin

Hickman and Lipple (1978) and Hickman and Gibson (1982) assigned the thick succession of subaerial basalt, minor hyaloclastite, and pillow lava, which outcrops in the Limestone Well and Pear Creek Centroclines, to the Kylena Basalt (Formation). Blake (1993) argued that the basalt pile is up to 1100 m thick, however, and is also atypical of the Kylena Formation in that it is overlain by thick, coarse-grained siliciclastic facies. Stratigraphic and sedimentological studies of the overlying unit indicate that it is probably part of the upper Hardey Formation, a relationship which would place the basalt succession in the lower to middle Hardey Formation and not in the Kylena Formation (Blake, 1984a, 1990, 1993).

A small, faulted outlier of Kylena Formation unconformably overlies basement greenstone rocks immediately north of Soansville. The basaltic succesion is conformably overlain by a thin development of volcaniclastic rocks belonging to the Tumbiana Formation.

Northeast Pilbara sub-basin

The Kylena Formation of the northeast Pilbara sub-basin ranges in thickness from 0 to 100 m on northern ROY HILL and northwest BALFOUR DOWNS to about 600 m in the Meentheena Centrocline, Northwest Oakover Sycline, and eastern Rooney Inlier. Subaerial basalt flows, up to 40 m thick, dominate the stratigraphy in all areas but are interbedded with significant amounts of hyaloclastite, pillow lava, mafic tuff, and stromatolitic carbonate locally, for example, in the Meentheena Centrocline (MOUNT EDGAR AMG 2704544).

South Pilbara sub-basin

The Boongal Formation is the lateral equivalent of the Kylena Formation in the South Pilbara sub-basin. This unit outcrops around the Bellary, Milli Milli, Rocklea, and Wyloo Domes, and along the southwestern margin of the Sylvania Inlier; it has also been recorded in the FVG–1, WRL–1, SGS–1, and BMW–1 diamond drillholes (CRA Exploration Pty Ltd, 1985b, 1987b, 1988, 1989). Submarine volcanic facies, comprising pillow lava, massive lava, and fine- to coarse-grained volcaniclastic rocks, are the dominant rock type in most outcrops; subaerial facies are recorded only in the FVG–1, SGS–1, and WRL–1 boreholes and the northeastern Sylvania Inlier.

The Boongal Formation has an estimated thickness of 1000 m on the northeastern flank of the Wyloo Dome. No figure is available from the more deformed southwestern part of the Wyloo Dome because the Boongal Formation cannot be separated from the lithologically similar Pyradie and Bunjinah Formations. The Boongal Formation has a thickness range of 400–1000 m in the Bellary–Milli Milli–Rocklea area, whereas about 700 m of strata are present along the southwestern margin of the Sylvania Inlier. Neither the Boongal Formation and/or Kylena Formation, nor underlying parts of the Fortescue Group, appear to be present on the northern margin of the Sylvania Inlier. Basal sedimentary and mafic volcanic rocks beneath the Jeerinah Formation in the Spearhole Yard, Shovelanna Hill, and Murramunda – Jigalong Road areas are correlated with the upper part of the Upper mafic volcanic unit (Tyler, 1986) and are probably equivalent to the Bunjinah Formation (see Tyler, 1991, p. 31).

Three hundred and seventy metres of massive and amygdaloidal basaltic lava, pillow lava, and breccia, which lie above the Hardey Formation in the SGS–1 borehole on north-central MOUNT BRUCE (CRA Exploration Pty Ltd, 1988), are closely comparable with submarine volcanic facies in the Boongal Formation. Equivalent units in the FVG–1 and WRL–1 boreholes (CRA Exploration Pty Ltd, 1985b, 1987b) contain mixed subaerial and submarine facies and appear to mark the transition from a Boongal Formation style of volcanism to one more typical of the Kylena Formation.

Palaeogeography

The forgoing description and the accounts of previous authors (e.g. Horwitz, 1980, 1987; Morris and Horwitz,

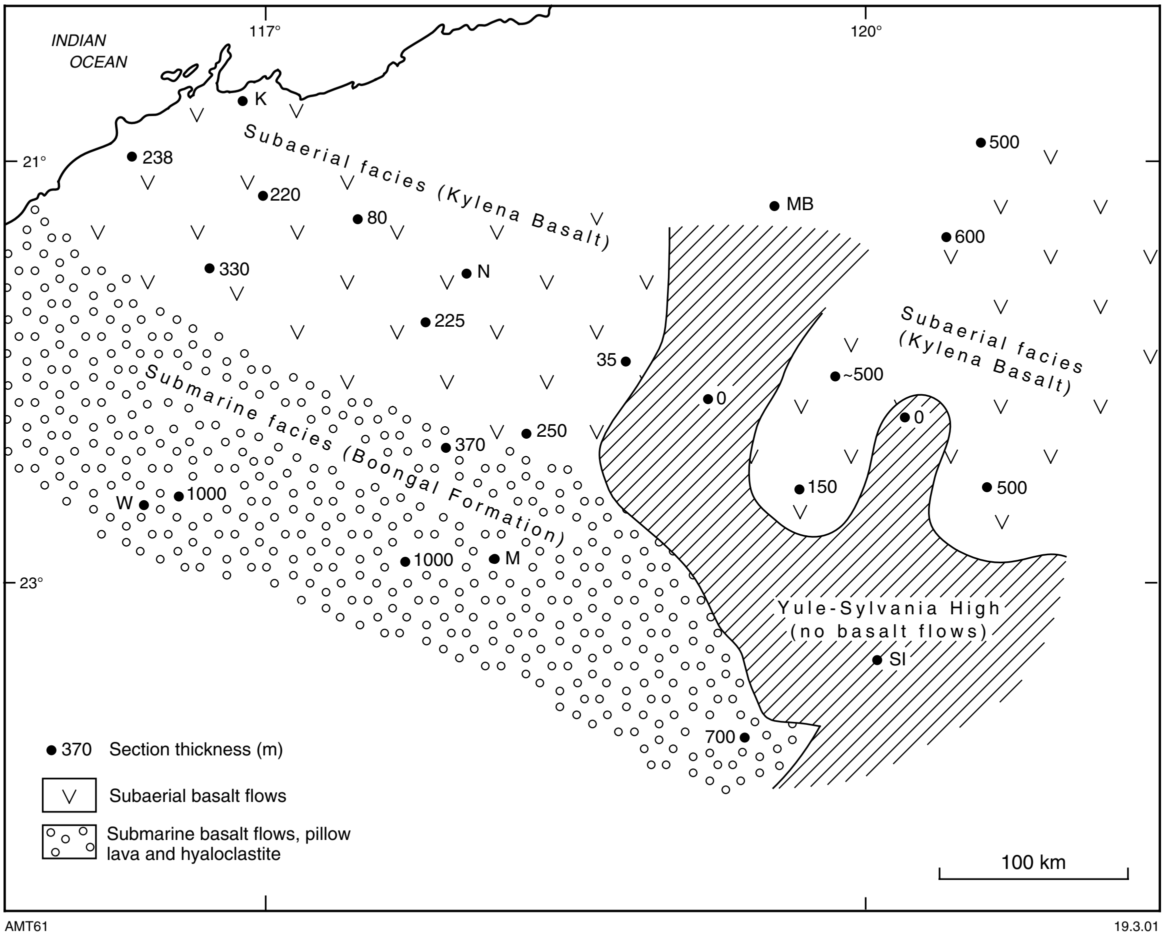


Figure 6.6. Palaeogeography of the Pilbara during deposition of the Kylena Formation and Boongal Formation. Locality abbreviations as in Figure 5.15

1983) affirm that volcanic rocks within the Kylena and Boongal Formations show a marked change from subaerial to submarine facies when traced from north to south across the Pilbara Craton (Fig. 6.6). In addition, the succession thickens toward the south (Horwitz and Smith, 1978; Morris and Horwitz, 1983).

Thickness data suggest that the Yule–Sylvania High, which was a dominant feature of Lower Fortescue Group palaeogeography, was considerably reduced by the time the Kylena and Boongal Formations were deposited. On the northern Pilbara Craton there is a tendency for the thickest sections of Kylena Formation to be preserved above the main north-northeasterly trending rift basins which were active during deposition of the lower Fortescue Group. Blake (1993) concluded from such data that the Nunyerry Creek Fault Complex and the southern extension of the Conglomerate Creek Fault were reactivated with a west-block-down movement (the same sense as in Hardey Formation times), and that the Sherlock River Fault Complex was reactivated with an east-block-down movement (the opposite sense to syn-Hardey Formation movement).

Indirect evidence of syn-Boongal Formation faulting in the southern Pilbara Craton has been found around the Sylvania Inlier. Here, units below the Bunjinah Formation, which are up to 4 km thick in the Deadman Hill area, are not recorded north of the easterly trending Western Creek Fault (Tyler, 1991). The south-block-down movement on this fault during Fortescue Group deposition was later reversed by compression associated with the Proterozoic Capricorn Orogeny (Tyler, 1991). There is marked variation in upper Fortescue Group (Bunjinah Formation to Jeerinah Formation) thickness and rock type across the north-northeasterly trending Fortescue River Fault. Although this evidence suggests the structure may have acted as a west-block-down growth fault during this time, its influence on pre-Bunjinah Formation volcanism and sedimentation is unknown.

Syn-Boongal Formation structures have not been recognized elsewhere in the South Pilbara sub-basin. Nevertheless, the sharp east-southeasterly trending boundary between subaerial and submarine volcanic facies, and the marked increase in the formation's thickness immediately southwest of the Yule–Sylvania High, prompted Thorne (1990) to suggest that an easterly to east-southeasterly trending, south-block-down growth-fault system may have been active in the southern Pilbara at this time. The inferred system is parallel to the main tectonic grain in the southern Pilbara and it is likely that many of the faults were reactivated during the subsequent Capricorn Orogeny.

Chapter 7

Tumbiana and Pyradie Formations

The Tumbiana Formation outcrops in a narrow strip extending from Cape Preston in the northwest Pilbara to Warrawagine Homestead in the northeast and is also found in the Gregory Range (Fig. 7.1). The lateral equivalent of the Tumbiana Formation, the Pyradie Formation, is recorded from all Fortescue Group inliers in the south Pilbara sub-basin and is equivalent to ultramafic rocks within Tyler's (1986) upper mafic volcanic unit on NEWMAN.

The Tumbiana Formation of the northern Pilbara Craton is commonly less than 200 m thick and is either conformable upon the Kylena Formation or unconformable upon granite–greenstone terrain. In the former situation, the widespread occurrence of flow tops on the uppermost Kylena Formation flows suggests that there was little or no erosion immediately prior to deposition of the Tumbiana Formation. This observation does not, however, preclude the possibility that the contact represents a significant time break.

In the south Pilbara sub-basin, the Pyradie Formation is up to 1.2 km thick and the contact between it and the underlying Boongal Formation appears to be conformable.

The age of the Tumbiana Formation is indicated by a U–Pb date of 2715 ± 6 Ma from the youngest zircon population in a tuffaceous sandstone from the middle of the formation in the northwest Pilbara sub-basin (Arndt et al., 1991). Uranium–lead zircon dates from the underlying Hardey Formation lie within the range 2768 ± 16 to 2756 ± 8 Ma (Pidgeon, 1984; Arndt et al., 1991) while a rhyolite in the overlying Maddina Formation has a U–Pb zircon date of 2717 ± 2 Ma (Nelson, 1998; Kojan and Hickman, 1998).

Previous descriptions of the Tumbiana and Pyradie Formations are given by: Daniels (1970), de la Hunty (1965), Hickman (1978, 1983), Hickman et al. (1983), Kojan and Hickman (1998), Kriewaldt (1964a,b), Kriewaldt and Ryan (1967), Lipple (1975), MacLeod and de la Hunty (1966), Packer (1990), Seymour et al. (1988), Thorne et al. (1991), Thorne and Tyler (1994, 1996, 1997a,b), and Williams (1968, 1989). The formations have also been described under various other formal and informal names (Blake, 1984a, 1993; Noldart and Wyatt, 1962; Trendall, 1975b, 1990b; Tyler, 1986, 1991).

This chapter describes and interprets the principal volcanic and sedimentary facies associations present in the Tumbiana and Pyradie Formations. Most of the data have been obtained from four measured sections in the northwest and northeast Pilbara sub-basins and 1:40 000-scale mapping of selected outcrop areas in the south Pilbara sub-basin.

The Tumbiana Formation consists of stromatolitic and clastic carbonate, argillite, sandstone, primary and reworked tuff, and minor conglomerate. In contrast, the Pyradie Formation comprises pyroxene spinifex-textured basalt flows and pillow lava, hyaloclastite, komatiite, and minor chert and tuffaceous argillite.

Non-volcanic sedimentary facies

Non-volcanic sedimentary rocks can be grouped into two major facies: coastal and nearshore shelf, and offshore shelf.

Coastal and nearshore-shelf facies

Coastal and nearshore-shelf facies comprise large-scale cross-stratified sandstone, complexly cross-bedded calcareous sandstone, stromatolitic and fenestrate carbonate, and micaceous quartz sandstone. These deposits are often interlayered with volcaniclastic rocks, especially primary and reworked air-fall tuff.

Large-scale cross-stratified sandstone

Description: Large-scale cross-stratified sandstone forms laterally impersistent units close to the base of the Tumbiana Formtion in the northwest Pilbara sub-basin. Units are up to 30 m thick and are transitional along strike and upwards into complexly cross-bedded calcareous sandstone. Internal structure is dominated by cosets of stacked trough cross-strata in which individual sets range from about 0.3 to 1.3 m thick and 2 to 12 m wide (Fig. 7.2). Palaeocurrent data are unimodal and directed toward the south (Fig. 7.3). Using the scheme of Folk (1974), most sandstones are classified as chlorite-cemented, medium- to coarse-grained litharenites. The well-rounded grains are composed of monocrystalline and polycrystalline quartz, plagioclase and K-feldspar, and epiclastic mafic and felsic volcanic rock.

Interpretation: Trough cross-stratification is formed by the downstream migration of lunate to sinuous dunes during

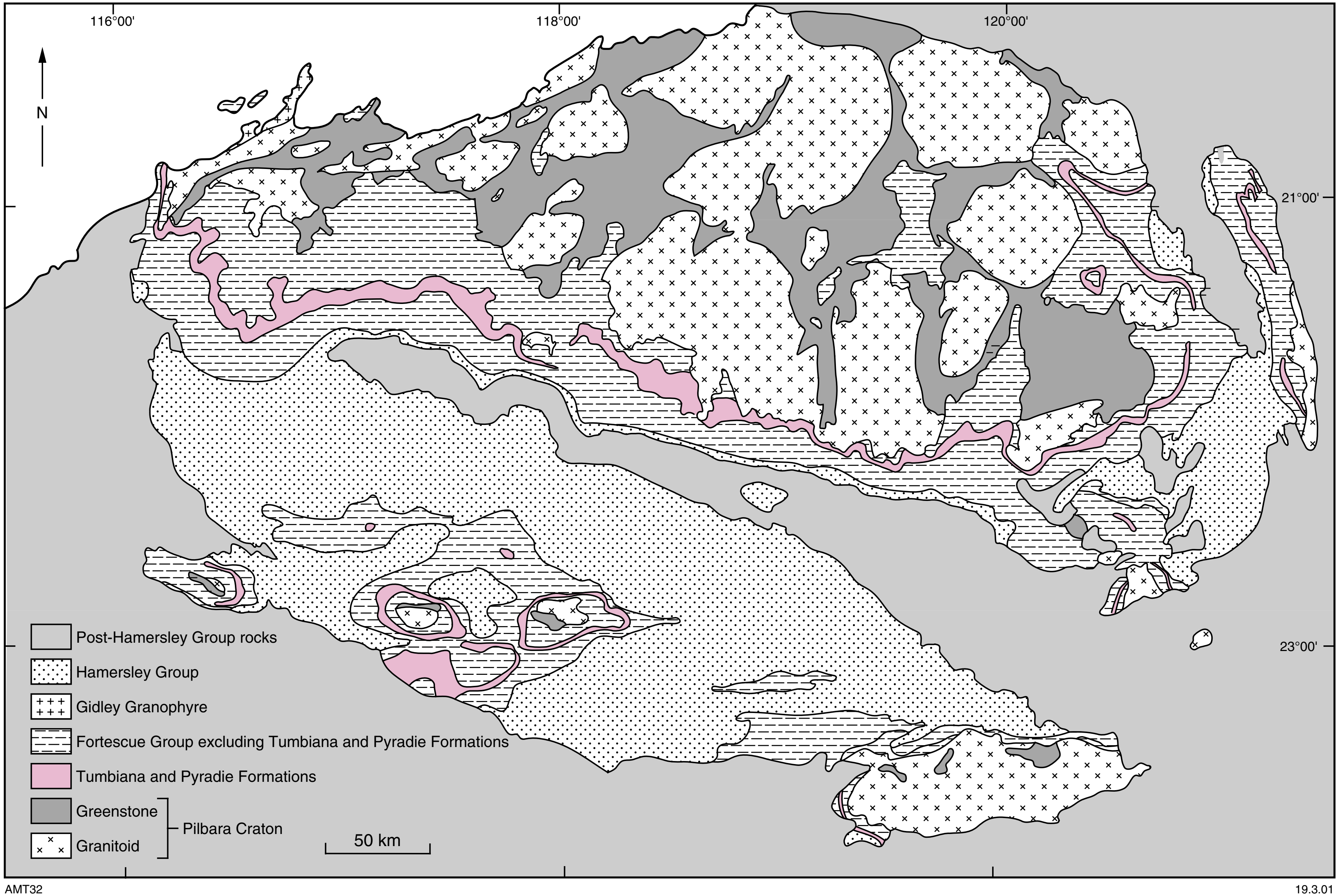
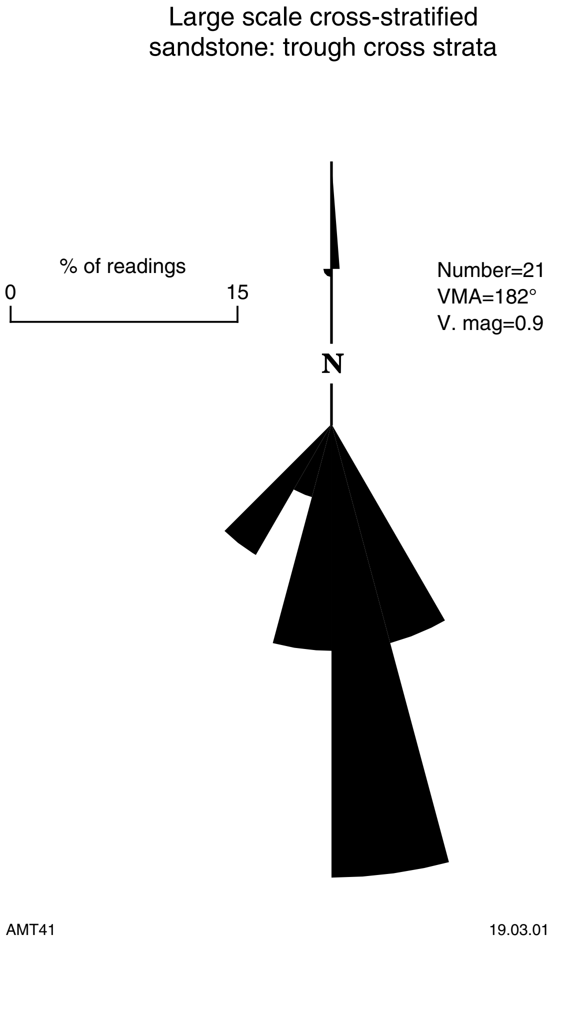


Figure 7.1. Principal outcrop areas of the Tumbiana and Pyradie Formations. In Wyloo Dome, the Pyradie Formation has not been separated from the rest of the post-Hardey Formation Fortescue Group stratigraphy (cf. Seymour et al., 1988)

Figure 7.2. Large-scale cross-stratified sandstone, coastal and nearshore shelf facies of the Tumbiana Formation, Horseshoe Gorge, northwest Pilbara sub-basin. Jacobs staff is 1.5 m long



moderate to high flow velocities (Harms et al., 1982). This, combined with the unidirectional palaeocurrents, limited lateral extent of the facies normal to the palaeoflow, scale of cross-stratification, and association with shallow marine facies suggests that the sandstones are channel sediments deposited near the mouth of a major, southward-flowing river.

Complexly cross-bedded calcareous sandstone

Description: Complexly cross-bedded calcareous sandstones occur in 2–30 m thick units in the lower part of the Tumbiana Formation in the northwest Pilbara sub-basin. Here, the units appear to persist along strike for tens of kilometres, but may pass laterally into large-scale cross-stratified sandstone. Units are medium to thick bedded and display a variety of cross-stratification styles; small stromatolitic buildups are also present locally.

Beds consist of sets or cosets of small- to medium-scale trough cross-strata (Fig. 7.4), or tabular cross-strata interlayered with sets of parallel or gently inclined stratification. Palaeocurrent data from cross-stratified sets are bimodal with vector means toward 146° and 276° (Fig. 7.5a). Upper boundaries of cross-stratified sets

Figure 7.3. Palaeocurrent rose diagram for large-scale cross-stratified sandstone, coastal and nearshore-shelf facies, Tumbiana Formation, Horshoe Gorge area. Data are recorded from axes of trough cross-strata. VMA = vector mean angle, V. mag. = vector magnitude

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Figure 7.4. Small-scale cross-stratification in complexly bedded calcareous sandstone, coastal and nearshore-shelf facies, Tumbiana Formation, Horseshoe Gorge, northwest Pilbara sub-basin

and cosets are modified locally into straight-crested symmetrical and asymmetrical rippled bedforms. Rippled linsen with similar morphology are also abundant in sets showing parallel or gently inclined stratification. Bedding plane exposures indicate most ripple crests strike easterly (Fig. 7.5b).

Irregular, undulatory organo-sedimentary lamination is evident in many sets showing parallel or gently inclined stratification. Locally, the laminae are organized into nodular to columnar stromatolites whose synoptic relief ranges from 0.03 to 0.15 m. Reworked stromatolites of this type are also recorded in some cross-stratified sets.

Complexly bedded sandstones commonly consist of chlorite and carbonate cemented lithic arenites. Most comprise 40–75% terrigenous fragments (monocrystalline and polycrystalline quartz, microcline, plagioclase, and felsic and mafic volcanic rock) with the remaining clasts consisting of rounded carbonate peloids.

Interpretation: Complexly cross-bedded calcareous sandstones were deposited in a shallow subaqueous environment characterized by intermittent moderate to strong currents and considerable reworking by wave processes. Trough and tabular cross-stratification formed in response to the downcurrent migration of lunate- to straight-crested dunes under lower flow regime conditions (Harms et al., 1982). Straight-crested symmetrical and asymmetrical ripples are interpreted as wave ripples formed under conditions of low to moderate sediment fallout (de Raaf et al., 1977; Allen, 1984). The easterly orientation of wave-ripple crests is normal to the palaeoflow in underlying channel facies and probably reflects the trend of the ancient shoreline. If this interpretation is correct, the bimodal palaeocurrent trend evident in cross-stratified units probably refects longshore currents (toward 275°) and offshore-directed rip currents (toward 145°).

Stromatolitic and fenestrate carbonate and tuffaceous carbonate

Description: Stromatolitic (Fig. 7.6) and fenestrate carbonates occur in 0.5–20 m-thick units and are generally interbedded with primary and reworked air-fall tuff and complexly cross-bedded sandstones. These thin- to thick-bedded units may persist along strike for several kilometres. In addition to containing stromatolites and/or displaying a fenestrate texture, many units include a variety of other primary and diagenetic sedimentary structures including ripple cross-lamination, tepee structures, pebble rosettes, desiccation cracks, and evaporite pseudomorphs (Packer, 1990). Many beds have also undergone varying amounts of silicification.

Packer (1990) described four stromatolite types from the northeast Pilbara sub-basin. Her type A stromatolite is *Alcheringa narrina* Walter (1972), which forms individual columns and domed to tabular biostromes. The shape of the individuals is cumulate to columnar (bulbous) and column shape is stubby with erect to inclined column attitude. Columns bifurcate infrequently and the branching

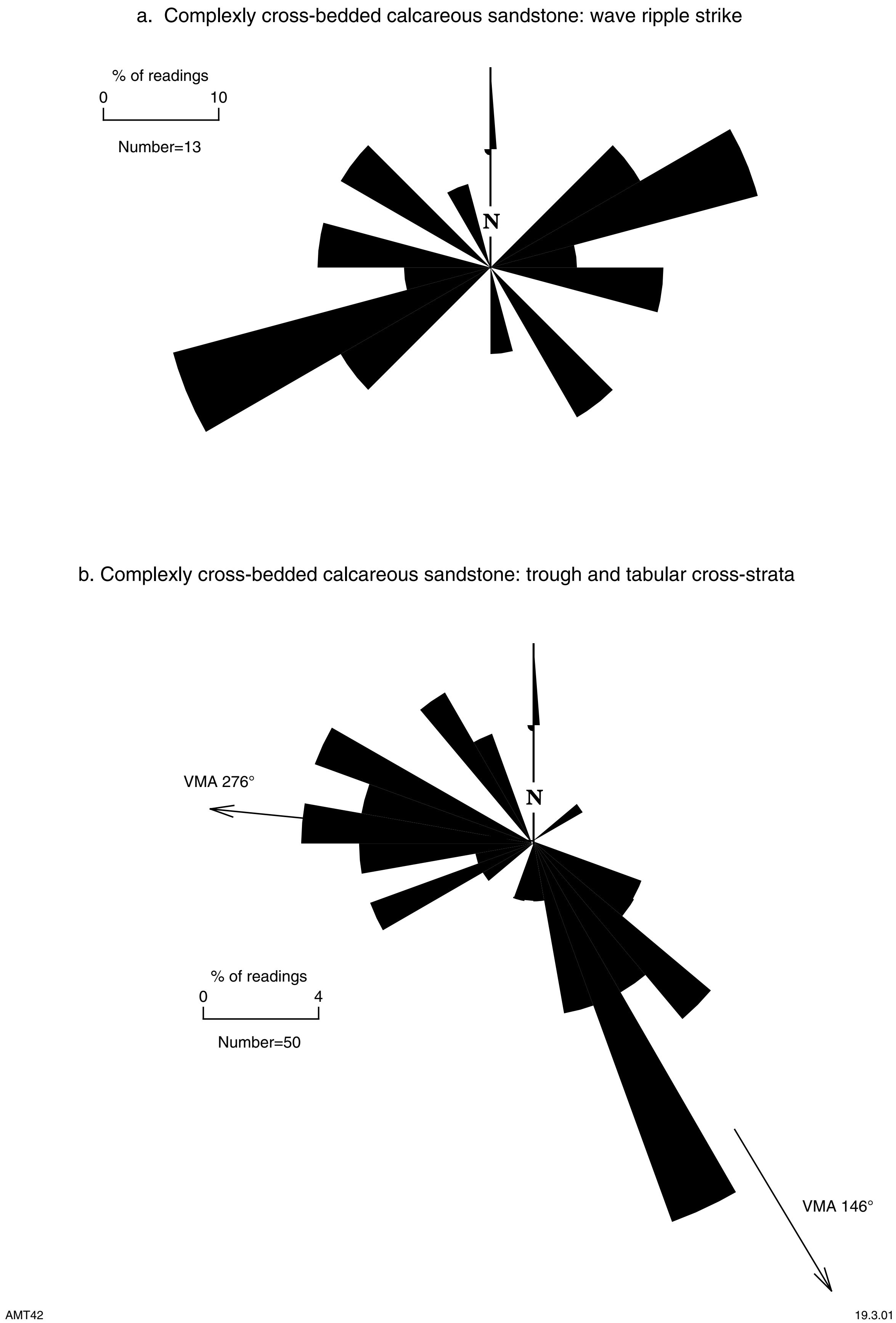

Figure 7.5. a) Palaeocurrent rose diagram for trough and tabular cross-strata in complexly cross-bedded calcareous sandstone, Tumbiana Formation, Horshoe Gorge, northwest Pilbara sub-basin. VMA = vector mean angle; b) Rose diagram showing wave-ripple strike, coastal and nearshore shelf facies

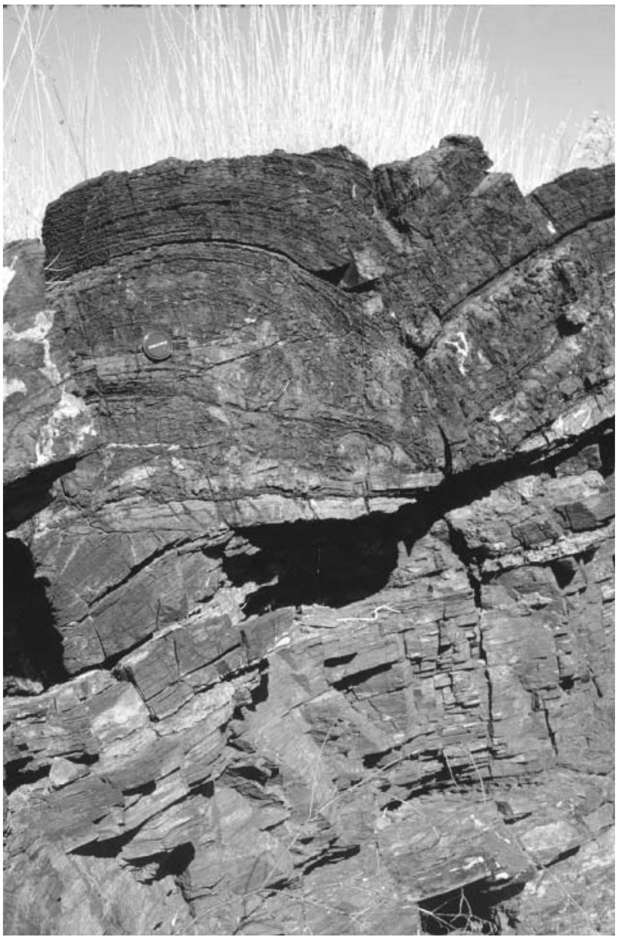

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Figure 7.6. Large domical stromatolite buildup, coastal and nearshore-shelf facies, Tumbiana Formation, Meentheena Centrocline, northeast Pilbara sub-basin. Width of photograph is 1.2 m. Photograph by K. Grey

angle is moderately divergent; the branching method is beta (cf. Grey, 1989). In profile, the smooth laminae are parabolic to slightly conical and have a low to moderate synoptic relief with a high degree of inheritance.

The type B stromatolite of Packer (1990) forms biostromes and bioherms with relief locally in excess of 0.5 m. Individuals are columnar in shape and column attitude is erect to slightly inclined. Stromatolites are linked and display no branching. In profile, the laminae are gently to steeply convex, smooth, and show a low synoptic relief with a moderate to high degree of inheritance.

Type C, or tufted stromatolites (Packer, 1990), form domed to tabular biostromes. The shape of individuals is linked conical to cumulate; in thin section they resemble the outline of a pine tree. They have a narrow top with one or more apices and widen toward the base with apices along the margins. Apices at the top point upward; those along the margin point upward or horizontally. In profile, the smooth laminae are cuspate, show moderate synoptic relief, and a high degree of inheritance.

Type D stromatolites form domed mushroom-shaped bioherms. The shape of individuals is columnar layered to pseudocolumnar; column shape is nodular and erect to inclined in attitude. Stromatolites branch repeatedly and branches are markedly divergent. Laminar profile is gently to steeply convex and laminae are very wrinkled and wavy with a moderate to low degree of inheritance (Packer, 1990).

Fenestrae occur in stromatolitic and non-stromatolitic carbonates. They are irregular spherical to laminoid in shape and are infilled by various combinations of sparry calcite, chert, and megaquartz crystals. Packer (1990) noted that some fenestrae contain chert pseudomorphs after gypsum.

Straight- and sinuous-crested symmetrical and asymmetrical ripples are abundant locally (Fig. 7.7). In profile, the rippled sets display both subcritical and supercritical climbing ripple cross-lamination. Many ripple surfaces are draped by organo-sedimentary lamination, which may pass upwards into small stromatolitic biostromes or build-ups. Pebble rosettes, termed beach rosettes by Packer (1990), are clusters of vertically aligned, tabular clasts of micritic carbonate. Clasts are generally 10–20 mm thick and up to 0.2 m long and may be colonized by stromatolitic lamination. Desiccation cracks are observed locally in micritic and fine-grained tuffaceous carbonates; these are up to 50 mm deep and are filled by calcarenite or coarser grained volcaniclastic material, including accretionary lapilli.

Interpretation: The association of sedimentary and diagenetic structures present in stromatolitic and fenestrate carbonates and tuffaceous carbonates suggests that they were deposited in a low-energy coastal setting.

Straight- and sinuous-crested symmetrical and asymmetrical ripples are interpreted as wave ripples formed under conditions of low to high sediment fallout. Subcritical climbing ripple cross-lamination forms during ripple migration when low to moderate rates of sediment fallout cause the angle of climb to be less than the slope of the ripple stoss. Supercritical cross-lamination forms when the sediment fallout, coupled with the slow migration of the ripple, allows the ripple to climb at an angle greater than that of the ripple stoss (Allen, 1984).

Clusters of vertically stacked pebbles, similar to the pebble rosettes in the Tumbiana Formation, have been described from modern marine and lacustrine beaches (Sanderson and Donovan, 1974). In these modern examples, this edgewise alignment reflects optimum clast packing under the influence of moderate wave agitation. Areas of edgewise packing on modern beaches are generally very stable, a feature which enables them to support a large sessile and semi-sessile biota. The stability of the stacked pebbles in the Tumbiana Formation is attested to by the fact that the tops of the rosettes are often colonized by stromatolites (Packer, 1990).

Fenestrate cavities could have formed by a number of different processes including repeated and prolonged desiccation, removal of organic matter, gas evolution due to the decay of organic material, and growth expansion

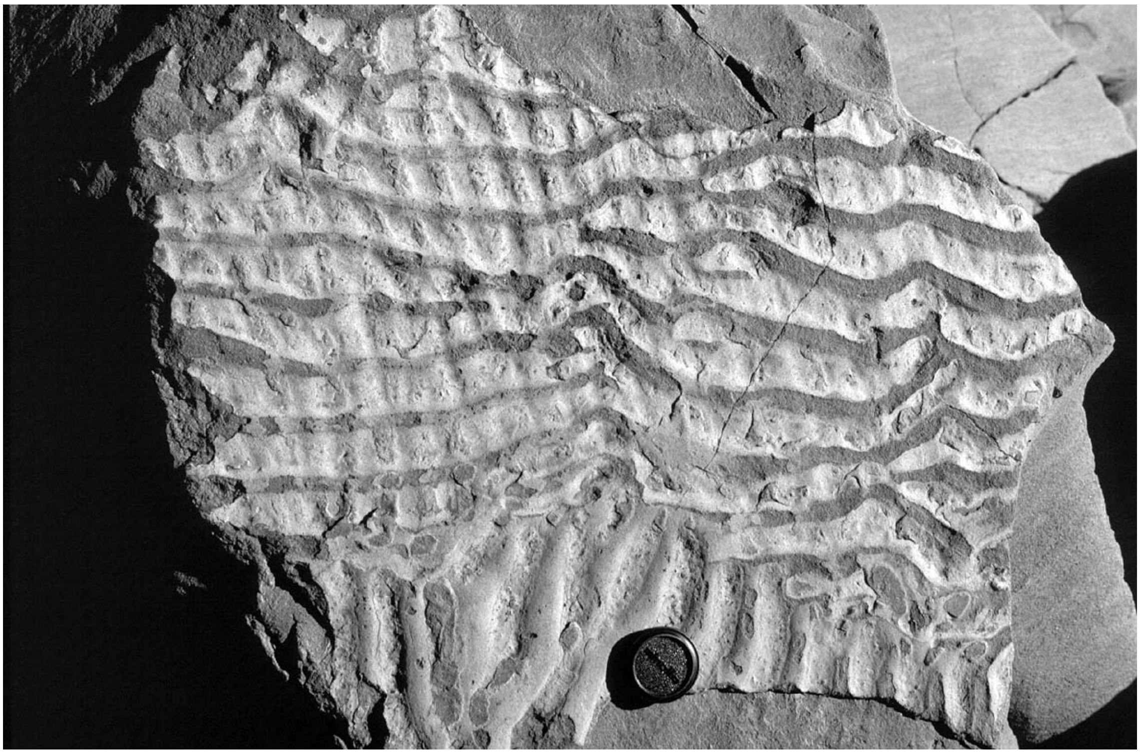

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Figure 7.7. Superimposed sets of sinuous-crested bifurcating ripples, stromatolitic and fenestrate carbonate facies, Tumbiana Formation, Horseshoe Gorge area, northwest Pilbara sub-basin

of algal mats (Logan, 1974; Shinn, 1968, 1983). They are recorded from shallow sublittoral to supralittoral environments but are not diagnostic of any particular setting.

Desiccation cracks in some carbonates and tuffaceous carbonates record intermittent periods of subaerial exposure. The tepee structures reported by Packer (1990) represent the buckled margins of large supralittoral polygonal structures formed by the periodic desiccation, wetting, cementation, and mechanical fracturing of sediment layers. Their formation often reflects seasonal changes in the height of the watertable; formation is also influenced by the location of groundwater springs (cf. Ferguson et al., 1982; Warren, 1983).

Packer's (1990) stromatolite types A, B, and C are associated mainly with fine-grained facies, interpreted as having been deposited in a strandline to supralittoral setting. The fourth stromatolite, type D, is associated with coarse-grained sediment and appears to have developed in a higher energy coastal channel environment.

The morphological similarity between Tumbiana Formation microfossils and modern oxygen-producing cyanobacteria has been used to suggest that oxygen-producing photoautotrophs were present in the Pilbara at about 2.7 Ga (Schopf and Walter, 1983). These workers described two microfossils from the Knossos locality that were morphologically similar to existing cyanobacteria. Packer (1990) suggests, however, that such a comparison is probably unwarranted because not only does morphology often correlate imperfectly with phylogeny, but convergent evolution of morphological characteristics is a common theme among procaryote microorganisms.

Micaceous quartz sandstone

Description: Micaceous quartz sandstone forms 2–25 m-thick laterally restricted units near the top of the Tumbiana Formation in the northwest Pilbara sub-basin. Units are thin to very thick bedded and fine to coarse grained. Medium to thick beds display unidirectional trough cross-strata; thin beds preserve straight-crested symmetrical and asymmetrical ripples. Current lineations are preserved on some bedding planes in thick- and thin-bedded units.

Interpretation: Micaceous quartz sandstones were deposited in a subaqueous environment characterized by fluctuating current energies and periods of wave reworking. Trough cross-stratification formed by the downcurrent migration of medium-sized dunes under lower flow regime conditions, whereas current lineations record periods of upper flow regime sediment-ation (Harms et al., 1982). Straight-crested symmetrical and asymmetrical ripples are interpreted as wave ripples.

Offshore-shelf facies

Offshore-shelf facies comprise thin-to thick-bedded sandstone and argillite, and minor thin-bedded chert and carbonate. These deposits are most abundant in the south Pilbara sub-basin, where they are often interlayered with volcanic facies, especially subaqueous basaltic lavas, hyaloclastite, and primary and reworked air-fall tuff.

Thin- to thick-bedded sandstone and argillite

Description: Thin- to thick-bedded sandstone and argillite units range in thickness up to about 50 m. Taken overall, most units are dominated by argillite although the proportion of sandstone may be high locally (Fig. 7.8). Argillaceous portions are commonly pyritic and carbonaceous and most are parallel-laminated, although cross-lamination is observed in some coarse-grained siltstones. Thin- and medium-bedded sandstones are tabular bedded and either ungraded or normally graded. Lower parts of beds may be parallel-laminated; upper parts display parallel lamination or cross-lamination. Thick-bedded sandstones are massive. Sandstone composition varies from quartzitic calcarenite and volcanic arenite to feldspathic arenite.

Interpretation: The fine grain size, parallel lamination, and lack of scouring suggest that the argillites were laid down from suspension in a low-energy subaqueous environment. A marine rather than lacustrine setting is favoured because of the large area over which these rocks are exposed. Siltstones that display cross-lamination formed as a result of ripple migration under lower flow regime conditions (Harms et al., 1982). Thin- to medium-bedded graded sandstones are interpreted as waning turbidity-current deposits (cf. Harms and Fahnestock, 1965; Walker, 1967). The precise method of deposition of the massive thick-bedded sandstones is not clearly understood. Walker (1970) suggested beds of this type are products of deposition from turbidity currents; others, reviewed by Rupke (1978), have stressed the importance during

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Figure 7.8. Parallel-laminated argillite and fine-grained sandstone, offshore-shelf facies, Pyradie Formation, Bellary Dome, south Pilbara sub-basin

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Figure 7.9. Evenly laminated, light-coloured chert nodules set in a darker chert matrix, offshore-shelf facies, Pyradie Formtion, Milli Milli Dome, south Pilbara sub-basin

deposition of alternative processes such as grain–grain interaction and movement of pore fluids.

Thin-bedded chert and carbonate

Description: Thin-bedded chert units range in thickness up to about 8 m and consist of 0.03–0.15 m-thick beds of evenly laminated white to black or red chert, interlayered with thin argillite. Many chert layers are nodular (Fig. 7.9). Carbonate beds are tabular to nodular and deformed locally by soft-sediment deformation.

Interpretation: Finely laminated cherts are interpreted as chemical precipitates that were laid down during periods of low clastic supply (cf. McConchie, 1984). The origin of the carbonate beds is unclear; they could have originated as chemical precipitates or as redeposited clastic material.

Volcanic facies

Volcanic facies within the Tumbiana and Pyradie Formations can be subdivided into subaerial basaltic lavas, subaqueous basaltic lavas, subaqueous komatiite lavas, and volcaniclastic rocks.

Subaerial basaltic lavas

Description: Subaerial basaltic flows are a minor component of the Tumbiana Formation and occur interbedded with volcaniclastic facies. Flows are thin, range from less than a metre to 3 m thick, and contain 5–25% of spherical amygdales in the lower parts of the flow. The upper levels are characterized by a much greater proportion (30–65% of the rock) of large, spherical to irregularly rounded or streaked amygdales. Most flows are aphyric. Amygdales usually range in size up to about 30 mm and occur either scattered randomly throughout the rock or concentrated in flow-aligned clusters. Most amygdales are filled by quartz, carbonate, and chlorite.

Thin amygdaloidal flows are bounded by parallel or sub-parallel flow surfaces which exhibit gentle, irregular undulations; most flows appear to be discontinuous laterally. Flow tops generally exhibit irregular to broadly symmetrical undulations and are rarely scoriaceous. Flow bases are generally smooth and follow irregularities in the underlying surface. No systematic jointing was observed in these flows.

No chemical analyses are available for lavas of this facies.

Interpretation: The basaltic composition and aphyric texture of most of the lavas suggest that they originated as low- to moderate-viscosity flows that were close to their liquidus temperatures at eruption (cf. Green, 1989). In addition, the general lack of evidence for contemporaneous explosive fragmentation suggests either that the initial content of exsolved volatiles was low, or that levels were high, but were reduced by non-explosive degassing, or by volatile loss during a previous phase of explosive activity (Cas and Wright, 1987).

It is not known to what extent many of the thin amygdaloidal lavas in the Tumbiana Formation represent ponded flows. On the one hand, these basalts are thinner and have a more varied morphology than typical ponded flows, but on the other, they commonly lack many of the features associated with flowing lava, such as lava tubes and surface channels.

Subaqueous basaltic lavas

Description: Subaqueous basaltic lavas are the dominant volcanic facies within the Pyradie Formation. They comprise pyroxene spinifex-textured basalt flows and pillow lava and are intimately associated with hyaloclastite, minor komatiite, and offshore-shelf sedimentary facies.

Pyroxene spinifex-textured basalt flows (Figs 7.10a–d, 7.11) are tabular bodies which range in thickness from a few metres up to 50 m. The thicker flows may persist laterally over several kilometres. Flow bases are sharp, planar to gently undulating, and fine to medium grained. Flow tops are irregular and fine grained, and are gradational into a hyaloclastite breccia, which may contain flame-like projections (Fig. 7.10a) and small pillows of the underlying flow material.

Internal structure of the flows is diverse, although most consist of a mesh of randomly oriented needles and blades of relic pyroxene, less than 30 mm long, set in a matrix of smaller crystals. Rosette-like aggregates of former pyroxene are observed locally within the matrix of some flows, whereas others are characterized by lenses and discontinuous seams, up to 0.1 m thick, containing needles and blades up to 50 mm long (Fig. 7.10b,c). The upper few metres of some flows contain quartz- and carbonate-filled amygdales, where they form about 5–20% of the rock (Fig. 7.10c). These are typically confined to a thin layer 0.1–0.5 m below the flow top. Most flows show no systematic fracture pattern; however, columnar joints, with some suggestion of a collonade–entablature structure, are well developed locally (Fig. 7.10a).

Pillowed pyroxene spinifex-textured basalts are subordinate to the tabular flow bodies. They form lenticular units, mostly 10–50 m thick, within the tabular flow pile and comprise sac or finger-like pillows less than 1 m in diameter. Most pillows are non-vesicular or sparsely vesicular and are fine to medium grained; most display a 10–50 mm-thick fine-grained selvage. Inter-pillow voids are filled with various combinations of quartz, carbonate, and hyaloclastite.

Flow mineralogy has been modified by low-grade burial metamorphism (cf. Smith et al., 1981), although relic textures are commonly well preserved (Fig. 7.12). Needles and blades of augite are partially or completely replaced by actinolite, chlorite, and minor epidote–clinozoisite. Interstitial material comprises various combinations of anhedral to subhedral actinolite, carbonate, epidote–clinozoisite, leucoxene, saussuritized and albitized plagioclase, quartz, sericite, sphene, and talc.

Summary of geochemistry: Figure 12.2 provides a total stratigraphic overview, constituent by constituent, for the 367 available analyses of mafic and ultramafic rocks of the Fortescue Group. The MgO and mg number plots of that figure show clearly that high-MgO rocks are concentrated in a stratigraphic section just below the (vertical) centre of the plots. The vertical scale on those plots has only a general relationship with the real stratigraphy; their high-MgO ‘spikes’ are caused by analyses of lavas and sills from the Pyradie Formation, and to a lesser extent by sills of the Cooya Pooya Dolerite and others in the Hardey Formation of the south Pilbara sub-basin.

In terms of the scheme used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group, the total of 24 analyses available for lavas of the Pyradie Formation includes 10 analyses of the PYBC (Bellary Creek) 4–letter group, 11 analyses of the PYHR (Hardey River) group, and three analyses of the PYCA (Cowley area) groups; all three groups fall within cluster E.

If, following the statistical treatment of Grunsky et al. (1992), a minimum of 10.20% MgO is taken arbitrarily as the lower cutoff for the application of the name komatiite (whether ultramafic or basaltic), then both the PYBC and PYHR groups contain komatiitic and non-komatiitic basalts, so that their mean group compositions given in Table 12.2 are not as significant as the averages of more homogeneous groups discussed in other chapters. Ten of these 24 analyses are of komatiitic rocks, whose chemical characteristics are summarized below. The remaining analyses differ compositionally from the major stacked flow sequences of the Fortescue Group in that they straddle the boundary between basalt and basaltic andesite in terms of the silica/total alkalies scheme of Le Maitre (1989). Although the scatter is wide, the silica content of most is below 55% (Fig. 12.6). More significantly, these rocks lie mainly within the high-magnesian field of Jensen (1976), whereas their associated komatiitic rocks of the same group extend into the komatiitic field of Jensen's plot (Fig. 12.7).

In summary, the limited data available suggest that the subaqueous basaltic lavas of the Pyradie Formation are chemically distinct from the major stacked flow sequences of the Fortescue Group, and have affinities with the associated, and both chemically and physically distinctive, komatiitic lavas.

Interpretation: Pyroxene spinifex-textured basalts are interpreted as subaqueous flows on the basis of their intimate association with pillowed units, hyaloclastite, and offshore-shelf sedimentary facies. Their sharp, planar lower contacts and irregular gradational upper boundaries, and presence locally of amygdales and columnar joints, clearly distinguish these subaqueous flows from superficially similar layered intrusive bodies in the middle part of the Fortescue Group (**Chapter 11**). Their relatively high Mg content, low aspect ratio (thickness:area; *sensu* Cas and Wright, 1987), and paucity of pillow development indicates most units originated as high-temperature, low viscosity, sheet- or ribbon-like flows.

Subaqueous komatiite lavas

Description: Subaqueous komatiite lavas are a minor component of the Pyradie Formation. They are present locally in central and southeastern parts of the south Pilbara sub-basin, where they make up 0–15% of the stratigraphy. Komatiites are interbedded with pyroxene spinifex-textured basalt flows, hyaloclastite, and offshore-shelf sedimentary facies.

Most of the information on komatiite flow morphology has been obtained from the area between Rocklea Dome and western Bellary Dome. Here, a single komatiite layer, ranging in thickness from 50–100 m, lies at or close to the base of the Pyradie Formation. Although poor outcrop continuity does not allow direct correlation across the region, the fact that the komatiite occupies roughly the same stratigraphic level, and maintains the same general thickness and internal structure throughout, suggests that it forms part of a single flow. If this reasoning is correct, the present-day distribution and thickness of the komatiite implies that it was a ribbon-like to tabular flow, which extended over an area 80–100 km long by 15–40 km wide. Another, geographically separate flow near the southwestern Milli Milli Dome, appears to have been less extensive and was probably confined to an area of about 20 × 5 km.

Komatiite flow bases are sharp and planar to gently undulating; flow tops are irregular and are gradational into hyaloclastite breccia. The internal structure of the komatiite flows shows little regional variation and can be subdivided, in ascending order, into:

- A basal fine- to medium-grained chilled margin, 0.1–1.5 m thick;
- Massive, pyroxene (augite) cumulate, 5–10 m thick. (Fig. 7.13a)
- Porphyritic (tending to glomoporphyritic locally) olivine orthocumulate comprising subhedral to euhedral olivine phenocrysts 5–15 mm long, in a matrix of 0.5–1.5 mm subhedral to euhedral olivine and intercumulus material. Phenocrysts become less abundant upwards. Unit is 4.5–10 m thick. (Fig. 7.13b)
- Massive or, more rarely, weakly parallel stratified olivine orthocumulate. Comprises subhedral to euhedral or embayed olivine, 5–10% anhedral to euhedral opaques, and 10–20% intercumulus material. Unit is 30–60 m thick and grades sharply into the overlying lithology (Fig. 7.13c).
- Pyroxene spinifex-textured unit in which clino-pyroxene occurs as random, feathery needles and blades, and as macroscopic rosettes and downward diverging sheafs. Unit is 4–15 m thick. (Fig. 7.13d)
- Fine-grained flow top. It is sparsely vesicular locally and grades up into flow-top breccia and hyaloclastite.

Komatiite mineralogy has been modified by low-grade burial metamorphism (cf. Smith et al., 1981), although relic textures are generally well preserved. Olivine is replaced by antigorite, chlorite, and magnetite; pyroxene has altered to tremolite. Interstitial material in pyroxenite units consists largely of various combinations of anhedral to subhedral actinolite, leucoxene, saussuritized and albitized plagioclase, and quartz.

Summary of geochemistry: It has been noted above that lavas and sills of the Pyradie Formation contribute to a distinctive stratigraphically restricted 'spike' in MgO content and mg number within the Fortescue Group (Fig. 12.2).

Ten of the analyses in the PYBC (Bellary Creek) and PYHR (Hardey River) 4-letter groups used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group have an MgO content of 10.20% or above. From the viewpoint of magma chemistry, this may be taken as a convenient arbitrary lower cutoff figure for description of the resultant lavas as 'komatiitic'. These lavas typically have silica contents between 45 and 52%. Except for one sample with 34.05% MgO, the magnesia content of these samples lies between 10 and 20%, so that they are basaltic komatiites. Their identity is clearly displayed on Figure 12.7, where most lie within the komatiitic field of Jensen's (1976) cation plot.

In summary, the limited data available show that the subaqueous komatiitic lavas of the Pyradie Formation are chemically distinct from the major stacked flow sequences of the Fortescue Group. The extent to which the subaqueous basaltic lavas represent a distinct and unrelated magma type, or whether there may be a range of flows intermediate between ultramafic komatiite and high-magnesian basalts, is at present unknown.

Interpretation: Komatiites in the Pyradie Formation are interpreted as subaqueous flows for similar reasons that the pyroxene spinifex-textured lavas are, namely their close association with pillowed units, hyaloclastite, and offshore-shelf sedimentary facies. They are unlikely to be intrusive bodies because of their dissimilar upper and lower contacts; the former are sharp and planar, whereas the latter are irregular and transitional into hyaloclastite. The presence of amygdales in upper parts of the komatiite is also more in keeping with an extrusive origin.

Their high Mg content, low aspect ratio (thickness:area; *sensu* Cas and Wright, 1987), and paucity of pillow development indicate that most komatiites originated as turbulent, very high temperature, low-viscosity, sheet- or ribbon-like flows (cf. Huppert et al., 1984). The komatiite flow structure described above is unusual in that it is characterized by a basal pyroxene cumulate unit, overlain by olivine orthocumulate, pyroxene spinifex, and flow-top material. This mineralogical layering is thought to reflect a basin-wide change to a more MgO-rich liquid. Pyroxene crystallized early, whereas throughout most of the later history of the flow, firstly olivine, then pyroxene were the main liquidus phases (Gole, M., 1990, pers. comm.). The predominance of subhedral to euhedral and embayed olivine in orthocumulates in the middle parts of the komatiite point to a moderate degree of super-cooling, followed by slow cooling of the flow interior (cf. Hill et al., 1987).

Figure 7.10. a) Projection of flow-top material into overlying hyaloclastite breccia, subaqueous, spinifex-textured basalt, Pyradie Formation, Rocklea Dome, south Pilbara sub-basin; b) Seam of vertically aligned relic pyroxene needles in a subaqueous pyroxene spinifex-textured basalt flow, Pyradie Formation, Rocklea Formation, south Pilbara sub-basin; c) Amygdaloidal flow top to submarine pyroxene spinifex-textured basalt. Amygdales are concentrated in a narrow zone about 0.1 m below the top of the flow (coinciding with the top of this sample), Pyradie Formation, Milli Milli Dome, south Pilbara sub-basin; d) Columnar jointing in submarine pyroxene spinifex-textured basalt, Pyradie Formation, Rocklea Dome, south Pilbara sub-basin

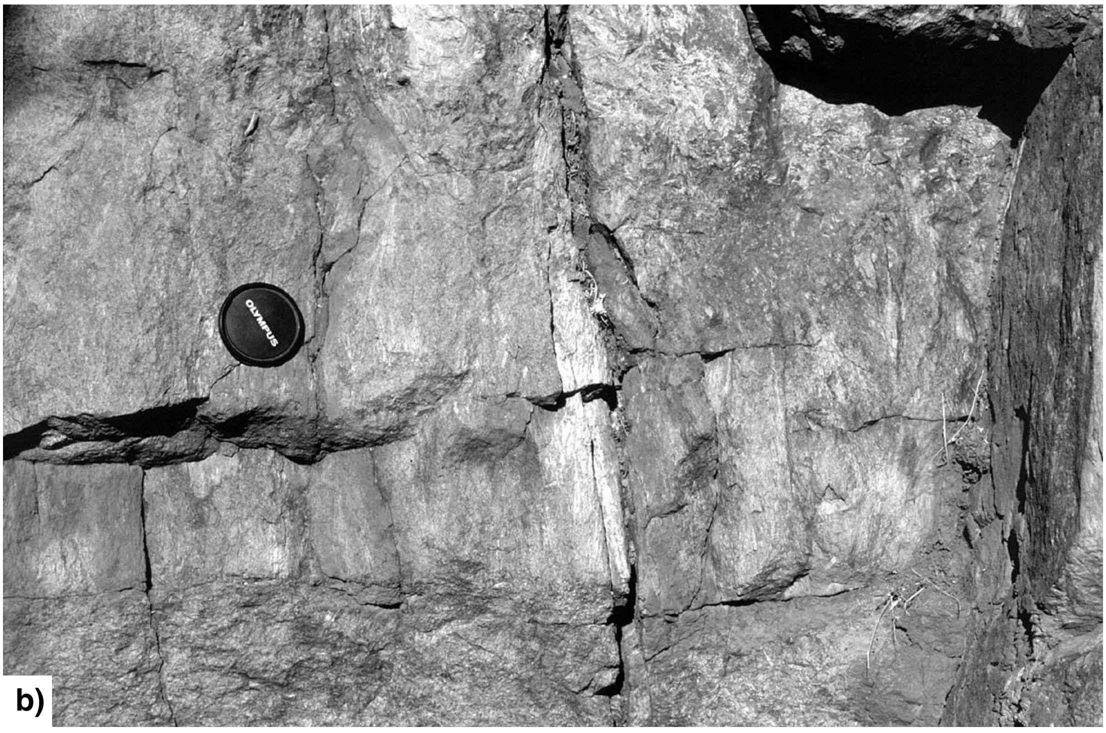

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Figure 7.10 (continued)

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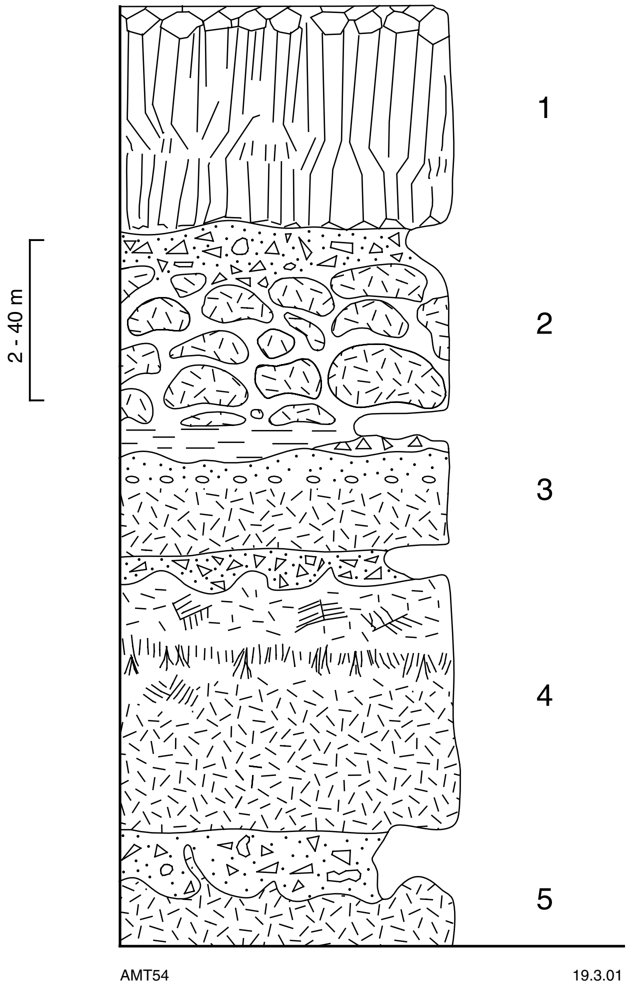

Figure 7.11. Summary of pyroxene spinifex-textured basalt flow morphology. 1 = columnar-jointed flow; 2 = pillow lava; 3 = amygdaloidal flow; 4 = massive flow comprising random crystals of relic pyroxene and horizontal seams of vertically oriented pyroxene; 5 = interflow breccia

Figure 7.12. Photomicrograph of pyroxene spinifex-textured basalt. Plane polarized light, field of view = 4 mm

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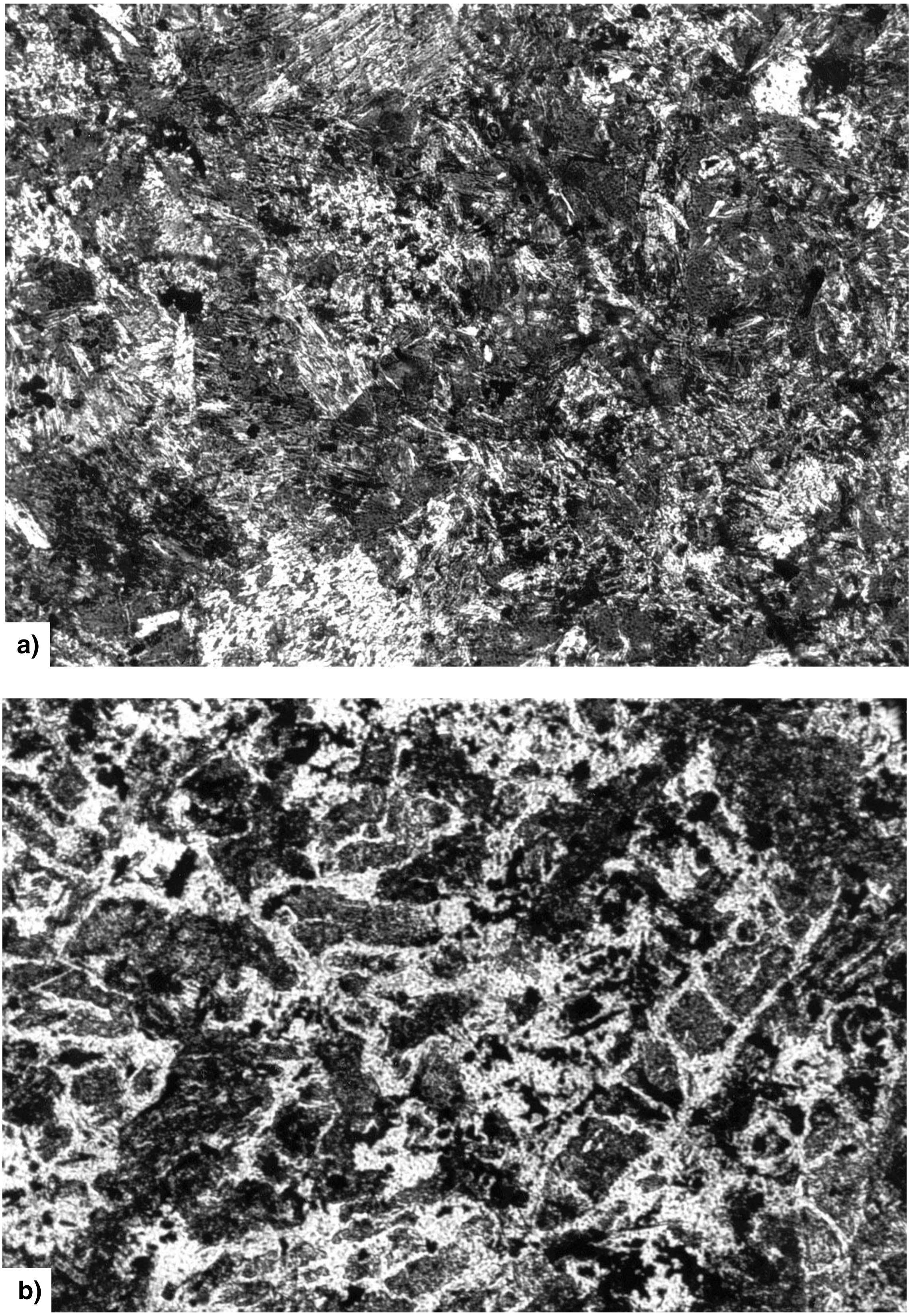

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Figure 7.13. a–d) Series of photomicrographs showing textural and mineralogical variation within an individual submarine komatiite flow, south Pilbara sub-basin. Plane polarized light, field of view = 4 mm

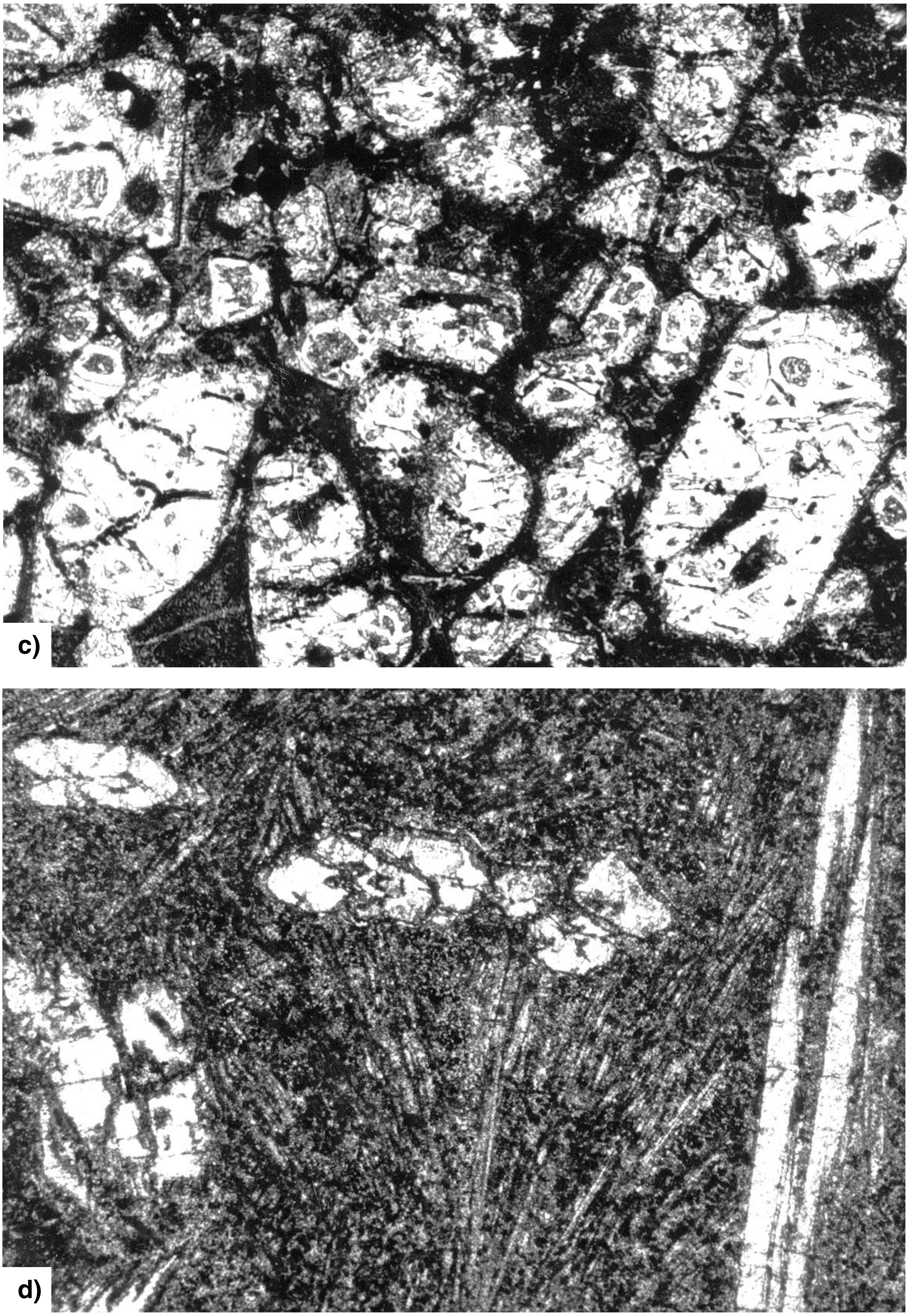

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Figure 7.13 (continued)

Volcaniclastic rocks

Description: Volcaniclastic rocks comprise hyaloclastite, tuff and lapilli tuff. Hyaloclastite is generally associated with pyroxene spinifex-textured basalt and pillow lava in the Pyradie Formation; primary and reworked tuff and lapilli tuff are abundant in the Tumbiana Formation.

Hyaloclastite occurs in laterally discontinuous units, ranging in thickness from around 0.2 to 10 m. Most hyaloclastites are massive, very poorly sorted breccias composed of angular fragments of sparsely vesicular basalt or komatiite that are up to 0.2 m across in a sand- to granule-sized matrix. Clasts are cut by networks of polygonal fractures that vary in width from hair-like cracks to wider veins filled by quartz, chlorite, and carbonate. Many clasts exhibit a jigsaw fit with neighbouring fragments. Detatched pillows or pillow fragments are observed locally in some hyaloclastites.

Silt and sand-sized tuff and lapilli tuff are present in friable, grey-green weathering units up to 30 m thick. Units are tabular bedded and internal structure consists primarily of horizontal lamination and ripple cross-lamination; there are small stromatolitic buildups and isolated accretionary lapilli within some beds. Thin (0.02–0.25 m) beds of closely packed accretionary lapilli tuff exist within most tuff units. Most layers are laterally persistent, others are discontinuous and fill hollows and cracks in the underlying surface. Beds have sharp, non-erosive bases and sharp or gradational tops; some are ungraded, others show normal or reverse grading (Fig. 7.14). Accretionary lapilli range in size from around 2 to 10 mm and most are characterized by a concentric internal structure.

Interpretation: The basaltic breccias are interpreted as hyaloclastites, produced by the rapid cooling and fracturing of basalt during contact with water. This interpretation is based upon the following evidence:

- Their homogeneous, generally non-vesicular clast composition.
- The predominance of poorly sorted, angular clasts and evidence for fracturing in situ.
- Their occurrence locally in interpillow voids.

Hyaloclastite breccias are produced when there is sudden contact between hot, coherent magma and cold water or water-saturated sediment (Cas and Wright, 1987). They form in a variety of situations: where magma is erupted subaqueously or subglacially; where lava flows into water or over water-saturated sediments; or where magma is intruded into water-saturated sediment or country rock. In the case of the breccias described here, geological setting rules out an intrusion-related or subglacial origin. The only interpretation that satisfies all the available evidence is the one in which the breccias formed in a marine environment. The lack of stratification in most hyalocalastites, combined with their matrix support, poor sorting, angular clasts, and presence of transitional boundaries with the basalt and komatiite flows suggests that they accumulated close to source, either as a flow-top mantle, or marginal scree and sediment gravity-flow deposits.

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Figure 7.14. Graded and non-graded layers of packed accretionary lapilli, Tumbiana Formation, Cape Preston, northwest Pilbara sub-basin

Beds of closely packed accretionary lapilli are interpreted as primary pyroclastic fall deposits, either from subaerial hydroclastic eruptions, or from eruptions where the moisture was supplied by rain falling through the eruption cloud (cf. Fisher and Schmincke, 1984). Mafic tuff and lapilli tuff units that display parallel lamination or ripple cross-lamination and contain localized stromatolitic buildups are regarded as reworked pyroclastic deposits which accumulated in a shallow marine or lacustrine setting.

Distribution of facies

The Tumbiana Formation or Pyradie Formation is recorded from most of the Pilbara, except the Marble Bar sub-basin. A summary of the stratigraphy of each sub-basin is shown in Figure 7.15.

Northwest Pilbara sub-basin

The Tumbiana Formation forms a continuous outcrop from Cape Preston to the eastern margin of the northwest Pilbara sub-basin. At Cape Preston, the formation appears to have a conformable, transitional contact with the underlying Kylena Formation. The 70 m-thick succession comprises alternating tuffaceous sandstone and minor stromatolitic carbonate (coastal and nearshore shelf facies), subaerial basalt flows, and accretionary lapilli tuff (primary and reworked pyroclastic air-fall material).

A thickness of 220 m of Tumbiana Formation was recorded in the Horseshoe Gorge area on northwestern PYRAMID. Here, contact with the Kylena Formation appears conformable, with little evidence of post-Kylena Formation – pre-Tumbiana Formation erosion. Coastal and nearshore-shelf facies dominate the entire thickness of the Tumbiana Formation. Lower levels consist largely of medium- to large-scale cross-stratified calcareous sandstone; middle to upper levels comprise reworked tuff and tuffaceous sandstone, stromatolitic carbonate, and micaceous sandstone. Layers of primary air-fall tuff are also more abundant in upper parts of the stratigraphy. One hundred and ninety metres of Tumbiana Formation were recorded from CRAE's MF–1 diamond drillhole on east central YARRALOOLA (CRA Exploration Pty Ltd, 1987a). The succession consists of tuffaceous sandstone and stromatolitic carbonate (coastal and nearshore-shelf facies), alternating with black argillite (offshore-shelf facies) and lapilli tuff (primary and reworked pyroclastic air-fall deposits).

Northwest of Nunyerry mine (MOUNT WOHLER AMG 862038), the Tumbiana Formation is about 130 m thick. The formation consists of coastal and nearshore-shelf facies and comprises cross-bedded and ripple-laminated calcareous sandstone and tuffaceous sandstone, stromatolitic carbonate, and tuffaceous carbonate, interbedded with primary and reworked air-fall tuff. This succession thickens to about 270 m in the SV–1 diamond drillhole (CRA Exploration Pty Ltd, 1985a).

Northeast Pilbara sub-basin

The Tumbiana Formation of the northeast Pilbara sub-basin ranges in thickness from 90 m on west-central ROY HILL, 80–100 m on BALFOUR DOWNS (Williams, 1989), less than 200 m on NULLAGINE (Hickman, 1978), to 120–150 m on YARRIE (Hickman et al., 1983). In these areas, the formation can be divided into lower and upper units, named the Mingah Tuff Member and Meentheena Carbonate Member respectively (Lipple, 1975). Both the Mingah Tuff and Meentheena Carbonate Members are dominated by coastal and nearshore-shelf facies. The former consists largely of reworked, fine- to coarse-grained volcaniclastic material and minor subaerial basalt and carbonate; the latter is dominated by stromatolitic and fenestrate carbonate, tuffaceous carbonate, and cross-bedded calcareous sandstone.

South Pilbara sub-basin

The Pyradie Formation is the lateral equivalent of the Tumbiana Formation in the south Pilbara sub-basin, where it outcrops around the Bellary, Milli Milli, Rocklea, and Wyloo Domes, in the core of the Jeerinah Anticline, and along the southwestern margin of the Sylvania Inlier. The Pyradie Formation has also been recorded in the FVG–1, WRL–1, SGS–1, and BMW–1 diamond drillholes (CRA Exploration Pty Ltd, 1985b, 1987b, 1988, 1989). Submarine volcanic facies and offshore-shelf facies are dominant in most southern outcrops. In the FVG–1, WRL–1, and SGS–1 boreholes in the northern part of the sub-basin, coastal and nearshore-shelf facies, and offshore-shelf facies are recorded in association with ?subaerial basalt and volcaniclastic rocks.

The Pyradie Formation has an estimated thickness of 800 m on the northeastern flank of the Wyloo Dome and consists of pyroxene spinifex-textured basalt and pillow lava, hyaloclastite, layered sills (**Chapter 10**), argillite, and minor chert. No thickness figure is available for the more deformed southwestern part of the Wyloo Dome because this formation is indistinguishable from the Boongal and Bunjinah Formations. The Pyradie Formation has a thickness range of around 200–800 m in the Bellary–Milli Milli–Rocklea area, whereas only the uppermost few tens of metres of strata are present in the Jeerinah Anticline. Nine hundred metres of Pyradie Formation are reported from the BMW–1 borehole (CRA Exploration Pty Ltd, 1989). Pyroxene spinifex-textured basalt and pillow lava, hyaloclastite, and argillite are the dominant lithologies; komatiite is found locally near the base of the formation between Paraburdoo and Rocklea Dome, and near the southwestern flank of Milli Milli Dome.

Data from the FVG–1, WRL–1, and SGS–1 diamond drillholes (CRA Exploration Pty Ltd, 1985b, 1987b, 1988) indicate that the Pyradie Formation of the northern part of the south Pilbara sub-basin is generally thinner than in the south. The FVG–1 stratigraphy consists of 360 m of tuffaceous argillite and minor thin-bedded tuffaceous sandstone (?offshore-shelf facies); and stromatolitic carbonate and vesicular basalt (?coastal and nearshore-shelf facies), interbedded with abundant pisolitic tuff (?pyroclastic air-fall material). In SGS–1 and WRL–1, the succession is 365 m and 290 m thick respectively, and consists of tuffaceous argillite and sandstone (?offshore-shelf facies), interbedded with lapilli tuff and minor hyaloclastite. Small amounts of pillow lava and massive

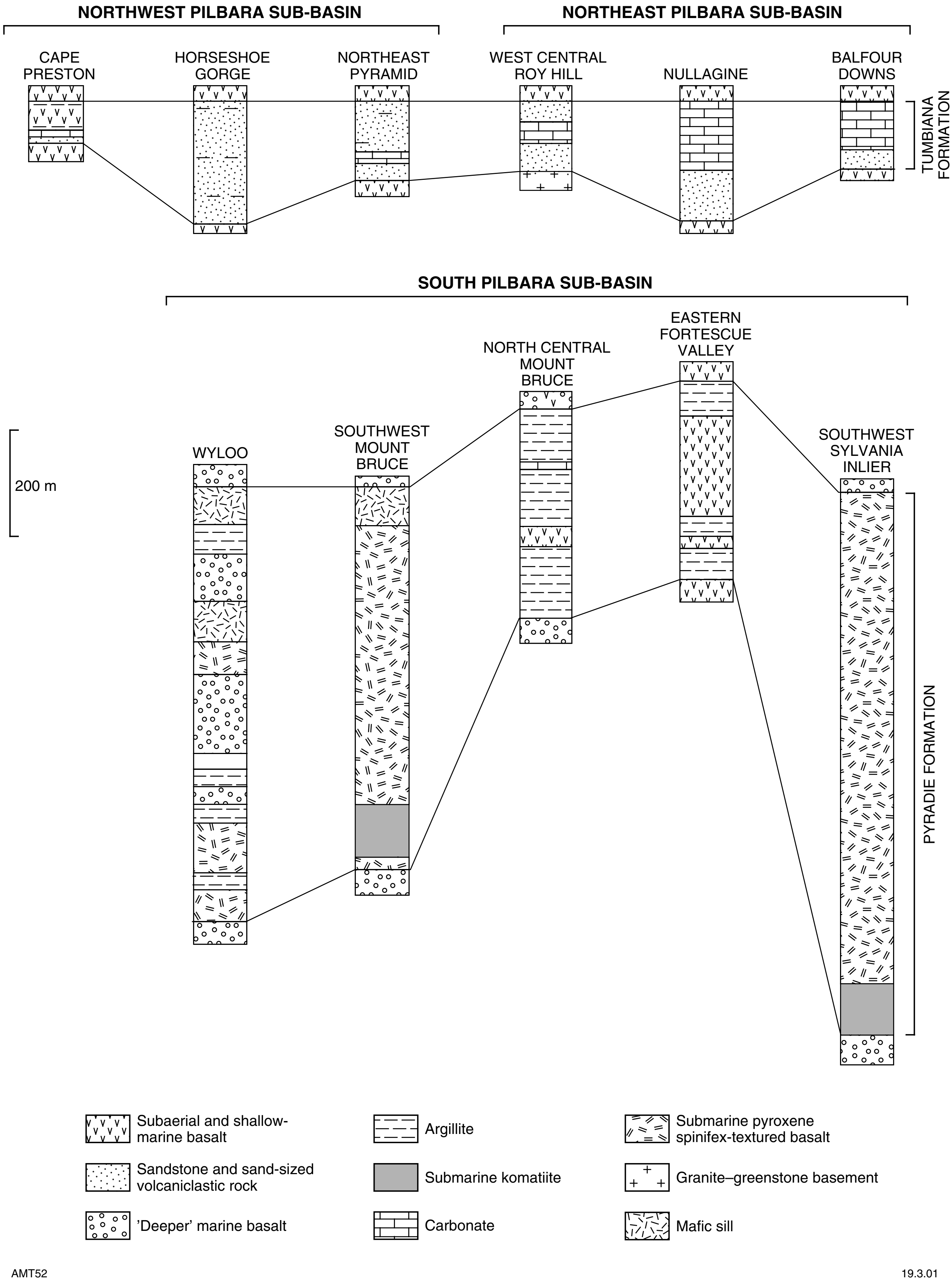

Figure 7.15. Generalized stratigraphy for the Tumbiana and Pyradie Formations in the northwest, northeast, and south Pilbara sub-basins

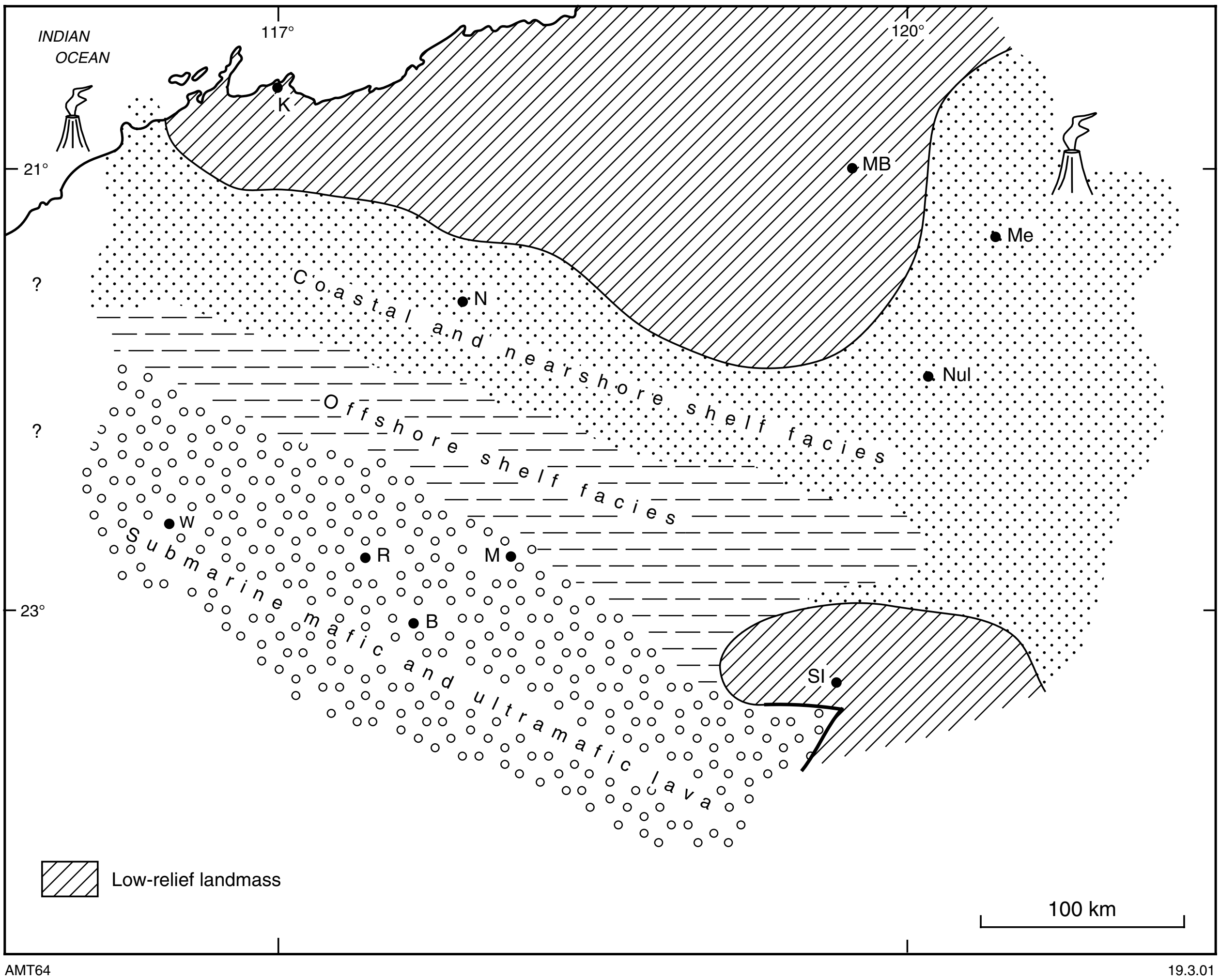

Figure 7.16. Palaeogeography of the Pilbara during deposition of the Tumbiana and Pyradie Formations. Locality abbreviations as in Figure 5.15

lava are recorded from SGS–1, whereas stromatolitic carbonates are present in lower and upper levels of WRL–1 (CRA Exploration Pty Ltd, 1987b, 1988).

In the Sylvania Inlier, the Pyradie Formation is confined to the Deadman Hill area where it is equivalent to the middle part of the upper mafic volcanic unit of Tyler (1991). The formation is about 1 km thick and consists of a lower submarine komatiite flow, about 150 m thick, overlain by pyroxene spinifex-textured basalt flows, hyaloclastite, and argillite.

Palaeogeography

The palaeogeographic reconstruction for the Tumbiana Formation and Pyradie Formation is similar to that for the Kylena and Boongal Formations (Fig. 7.16). Volcanic and non-volcanic sedimentary rocks show a change from coastal and shallow-marine shelf facies in the north Pilbara to deeper, outer-marine shelf facies in the south. In addition, there is an overall thickening of the succession towards the south (cf. Horwitz, 1980, 1987; Horwitz and Smith, 1978; Morris and Horwitz, 1983). The Yule–Sylvania High, which was a major feature of lower Fortescue Group palaeogeography, appears to have exercised little influence on Tumbiana and Pyradie Formation deposition.

Only small amounts of non-volcanic terrigenous sediment were supplied to the Hamersley Basin, mostly during early and late Tumbiana Formation deposition. In the northwest Pilbara sub-basin, sands derived from basement granite–greenstones were introduced to the shallow-marine shelf by a southward-flowing river system that entered the region in the Horseshoe Creek area. On eastern Roy Hill, basaltic conglomerate and sandstone occur locally at the base of the Tumbiana Formation suggesting that the underlying Kylena Formation acted as a temporary sediment source in this area.

The spatial and temporal distributions of pyroclastic air-fall tuff suggest that there were at least two major hydrovolcanic eruptive centres in the north Pilbara during Tumbiana Formation deposition. The first of these was in the northeast Pilbara, and was active mostly during deposition of the lower Tumbiana Formation; the second centre affected the northwest Pilbara during deposition of the upper part of the stratigraphy. Neither of these volcanic sources has been located during this study, although the virtual absence of proximal volcanic facies in the Tumbiana Formation suggests that the centres may have been outside the present Fortescue Group outcrop.

Direct evidence for syn-Pyradie Formation faulting is confined mostly to the Sylvania Inlier. Here, the Pyradie Formation, which is over 1 km thick in the Deadman Hill area, disappears when traced northward across the easterly trending Western Creek Fault and eastward across the north-northeasterly trending Fortescue River Fault (Tyler, 1991). Syn-Pyradie Formation structures have not been recognized elsewhere in the south Pilbara sub-basin. As with the Kylena and Boongal Formations, however, a sharp west-northwesterly trending boundary exists between coastal and nearshore-shelf facies in the north and the contrasting deeper marine volcanic and sedimentary facies in the south. This, and the marked change in the thickness of the Pyradie Formation south of the Fortescue Valley prompted Thorne (1990) to suggest that an easterly to east-southeasterly trending, south-block-down growth-fault system may have been active in the southern Pilbara at this time.

Chapter 8

Maddina and Bunjinah Formations

The Maddina Formation, or its lateral equivalent, the Bunjinah Formation, is present over most of the central and southern Pilbara Craton and the Gregory Range (Fig. 8.1) but is above the stratigraphic exposure level in the Marble Bar sub-basin. In the Sylvania Inlier, these rocks are equivalent to the upper part of the Upper mafic volcanic unit of Tyler (1986, 1991).

The Maddina Formation of the northern Pilbara Craton is typically between 500 and 1000 m thick. No evidence of angular discordance or erosion was observed between the Maddina Formation and the underlying Tumbiana Formation. A similar relationship exists in the southern Pilbara where, except for outcrops on the northern margin of the Sylvania Inlier, the Bunjinah Formation is apparently conformable with the underlying Pyradie Formation.

A U–Pb zircon date of 2717 ± 2 Ma from a rhyolite in the Maddina Formation (Nelson, 1998; Kojan and Hickman, 1998) shows that this formation is of very similar age to the underlying Tumbiana Formation, but is probably younger than the Gidley Granophyre. Gabbro at the base of this granophyre has been dated at 2725 ± 1.3 Ma using U–Pb baddeleyite geochronology (Wingate, 1997).

Previous descriptions of the Maddina and Bunjinah Formations are given by: Daniels (1970), Hickman (1978, 1983), Hickman et al. (1983), Horwitz (1976, 1978), Kojan and Hickman (1998), Kriewaldt (1964a,b), Kriewaldt and Ryan (1967), Seymour et al. (1988), Smith (1975, 1976, 1979), Smith et al. (1982), Thorne and Tyler (1994, 1996, 1997a,b), and Williams (1989). The formation has also been described under various other formal and informal names (Blake, 1984a, 1993; Noldart and Wyatt, 1962; Trendall, 1975b, 1990b).

This chapter describes and interprets the principal volcanic and sedimentary facies associations present in the Maddina and Bunjinah Formations. Most of the data have been obtained from four measured sections and reconnaissance studies in the northwest and northeast Pilbara sub-basins, and 1:40 000-scale mapping of selected outcrop areas in the South Pilbara sub-basin.

The Maddina and Bunjinah Formations consist mainly of basalt flows, pillow lava, fine- to coarse-grained and mafic volcaniclastic rocks. Non-volcanic sedimentary rocks, including stromatolitic carbonate, quartz sandstone, conglomerate, and argillite are also recorded locally.

Volcanic facies

Volcanic facies are similar to those in the Kylena and Boongal Formations and can be subdivided into subaerial basaltic lavas, subaqueous basaltic lavas, and volcaniclastic rocks. Non-basaltic volcanic rocks form part of the formation in the west Pilbara.

Subaerial basaltic lavas

Physical volcanology: Subaerial basaltic flows are the dominant volcanic facies within the Maddina Formation. Flows range from less than a metre to over 60 m thick. Average flow thickness and number of flows recorded in each incomplete measured section is as follows: Cape Preston (20.1 m, 44), Pannawonica railway (10.3 m, 20), Newman–Port Hedland road (~15 m, 8), Newman–Port Hedland railway (10 m, 13), and Warrie Homestead (16 m, 7) (Fig. 2). Two broad categories of basaltic flow can be recognized on the basis of thickness and flow structure: thin amygdaloidal flows, and massive to amygdaloidal flows.

Thin (generally <5 m thick) amygdaloidal flows are recorded from most localities in the northeast and northwest Pilbara but are generally less abundant than massive to amygdaloidal flows. Most thin amygdaloidal flows co ntain 10–20% of spherical amygdales in lower parts of the flow, whereas upper levels are characterized by a much greater proportion (30–65% of the rock) of large, spherical to irregularly rounded or streaked amygdales. Amygdales usually range in size up to about 30 mm and are either scattered randomly throughout the rock or concentrated in flow-aligned clusters. Most amygdales are filled by quartz, carbonate, and chlorite.

Thin amygdaloidal flows are bounded by parallel or subparallel flow surfaces which exhibit gentle, irregular undulations. Most flows appear to be laterally discontinuous over distances exceeding 2 km. Flow tops typically exhibit irregular to broadly symmetrical undulations and are rarely scoriaceous. Flow bases are generally smooth and follow irregularities in the underlying surface. No systematic jointing was observed in these flows.

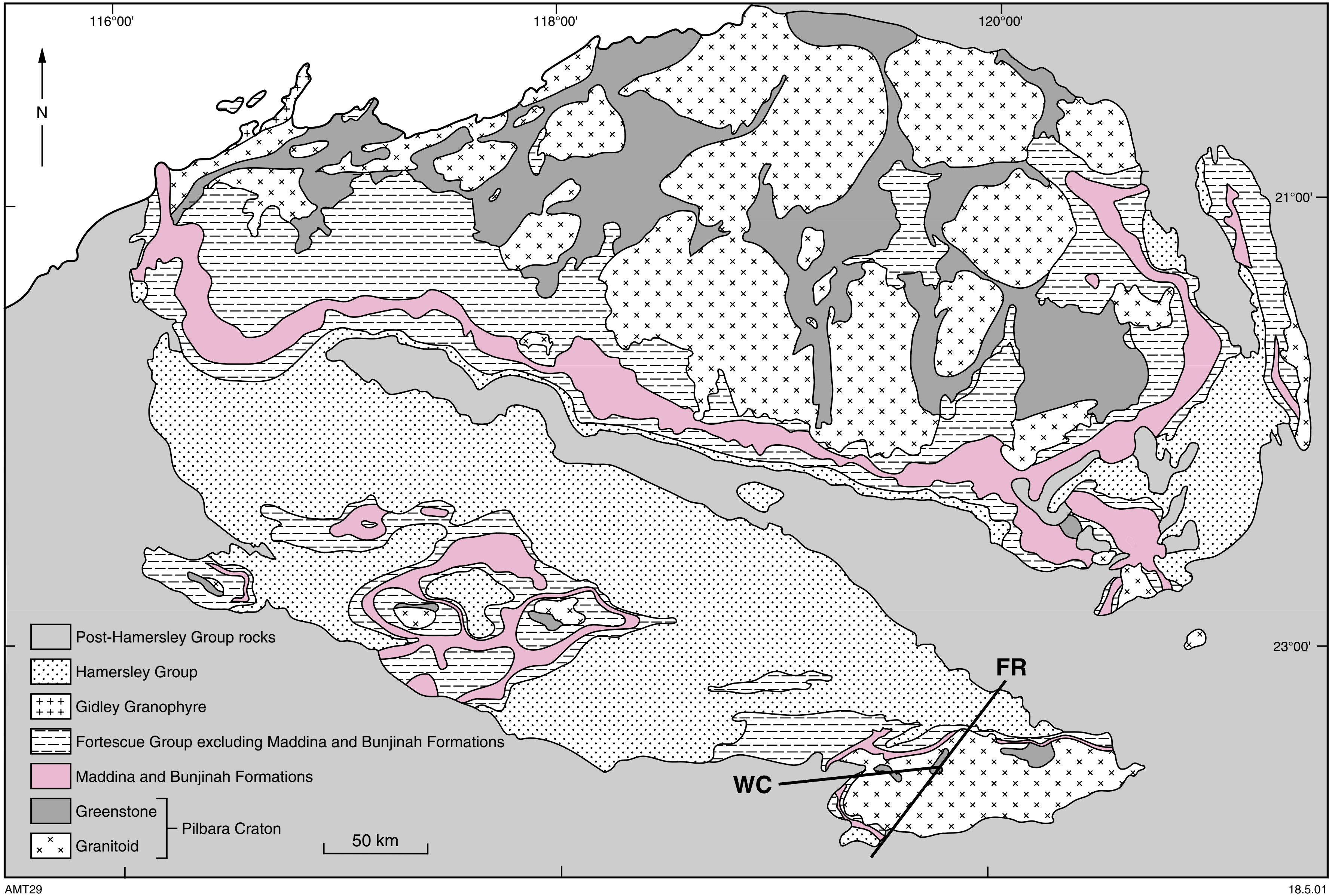

Figure 8.1. Principal outcrop areas of the Maddina Formation and Bunjinah Formation. The Bunjinah Formation has not been separated from the rest of the post-Hardey Formation – pre-Jeerinah Formation stratigraphy in the western Wyloo Dome. WC = Western Creek Fault, FR = Fortescue River Fault

Massive to amygdaloidal flows, equivalent to the altered flood lavas and altered flows of Smith (1975, 1976, 1979), are the most abundant flow type in the Maddina Formation. Flows range in thickness from about 5 to over 60 m and are characterized by a high proportion (20–65% of the rock) of large spherical to irregularly rounded or streaked amygdales in upper parts of the flow. These upper levels contrast with lower and middle parts of the flow, which commonly contain only a few percent of amygdales. Most amygdales range in size up to about 30 mm and are either scattered randomly throughout the rock or concentrated in folded, flow-aligned clusters and plumes (Fig. 8.2a), or bedding-parallel sheets. Exceptionally large amygdales, up to 1 m across, are a feature of the subaerial flows in the upper Maddina Formation of the northwest Pilbara sub-basin. Amygdale cylinders, (Fig. 8.2b) up to 0.15 m diameter and over 1.0 m long are recorded in the lower parts of some flows, whereas upper levels may contain bedding-parallel, blister-like cavities with abundant gas-escape structures on the cavity floor. The majority of amygdales are filled by various combinations of quartz, carbonate, and chlorite.

Massive to amygdaloidal flows are bounded by tabular flow surfaces which exhibit gentle, irregular undulations at outcrop scale. The thickest flows (>20 m) are laterally persistent over distances of 1 to at least 5 km; the lateral extent of thinner (<20 m) flows is unknown. Flow tops generally exhibit irregular to broadly symmetrical undulations (Fig. 8.2c). Wavelength and amplitude vary with flow thickness but are generally in the range 0.5–15 m and 0.2–2 m respectively. Strongly undululatory flow tops characterize the upper part of the Maddina Formation in the northwest Pilbara sub-basin; here wavelength and amplitude are in the order of 40–100 m and 2–8 m respectively (Fig. 9.2). Flow tops may be scoriaceous and evidence of autobrecciation is also present locally. Flow bases are typically smooth and mimic irregularities in the underlying surface. More rarely the lower parts of the flow comprise a hyaloclastite breccia, which is transitional upwards into massive or amygdaloidal basalt.

Most massive to amygdaloidal basalts show little systematic jointing; however, columnar joints and bedding-parallel partings have been observed locally. Columnar joints have been recorded only in flows thicker than 15 m. Individual columns are mostly 0.15–0.3 m in diameter and are present in lower and middle to upper parts of the flows; column axes are generally oriented 45–90° to the flow surfaces. No examples of the collonade–entablature structure (Long and Wood, 1986) were observed. Locally, the thick massive to amygdaloidal flows are cross-cut by an irregular network of small fractures and bedding-parallel joints. Their combined effect is to give the flow a bedded, pseudobrecciated appearance similar to that of weathered mafic tuff.

Petrography: Massive to amygdaloidal flows are fine to very coarse grained (50–3000 μm) and aphyric (Fig. 8.3a,b), although Smith (1975b) noted the presence of plagioclase-phyric varieties locally. Intersertal and intergranular textures dominate most basalts, whereas ophitic clinopyroxene is present in some thick flows.

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Figure 8.2. a) Plumose vesicle alignment at the base of a thick subaerial basalt flow. The darker unit at the bottom of the photograph is the underlying flow-top breccia, Maddina Formation, Gregory Gorge, northwest Pilbara sub-basin; b) Vertical amygdale cylinder in the lower part of a thick subaerial basalt flow, Maddina Formation, Bonnie Downs area, northeast Pilbara sub-basin; c) Plan view of domal flow-top structure in subaerial basalt. Note presence of accretionary lapilli in interdomal crevices, Maddina Formation, Bonnie Downs area, northeast Pilbara sub-basin

The most complete study of Maddina Formation petrography to date is that of Smith (1975b) and Smith et al. (1982), who recorded the effects of low-grade metamorphism on basalt mineralogy. Samples from the Maddina Formation fall in the prehnite–pumpellyite facies and two distinct associations may be recognized (Smith, 1975b). The first association bears pumpellyite in various combinations of albite–epidote–(prehnite)–pumpellyite–chlorite, in addition to quartz, sphene, and traces of carbonate. Of the original mineralogy, relic pyroxene is present in some samples. In most, however, pyroxene has been extensively altered to chlorite, carbonate, or quartz. Calcic plagioclase has been extensively replaced by albite, which may be clear, dusty, or contain pumpellyite or epidote–clinozoisite grains. Authigenic quartz forms anhedral patches in the groundmass, fills vesicles, and may

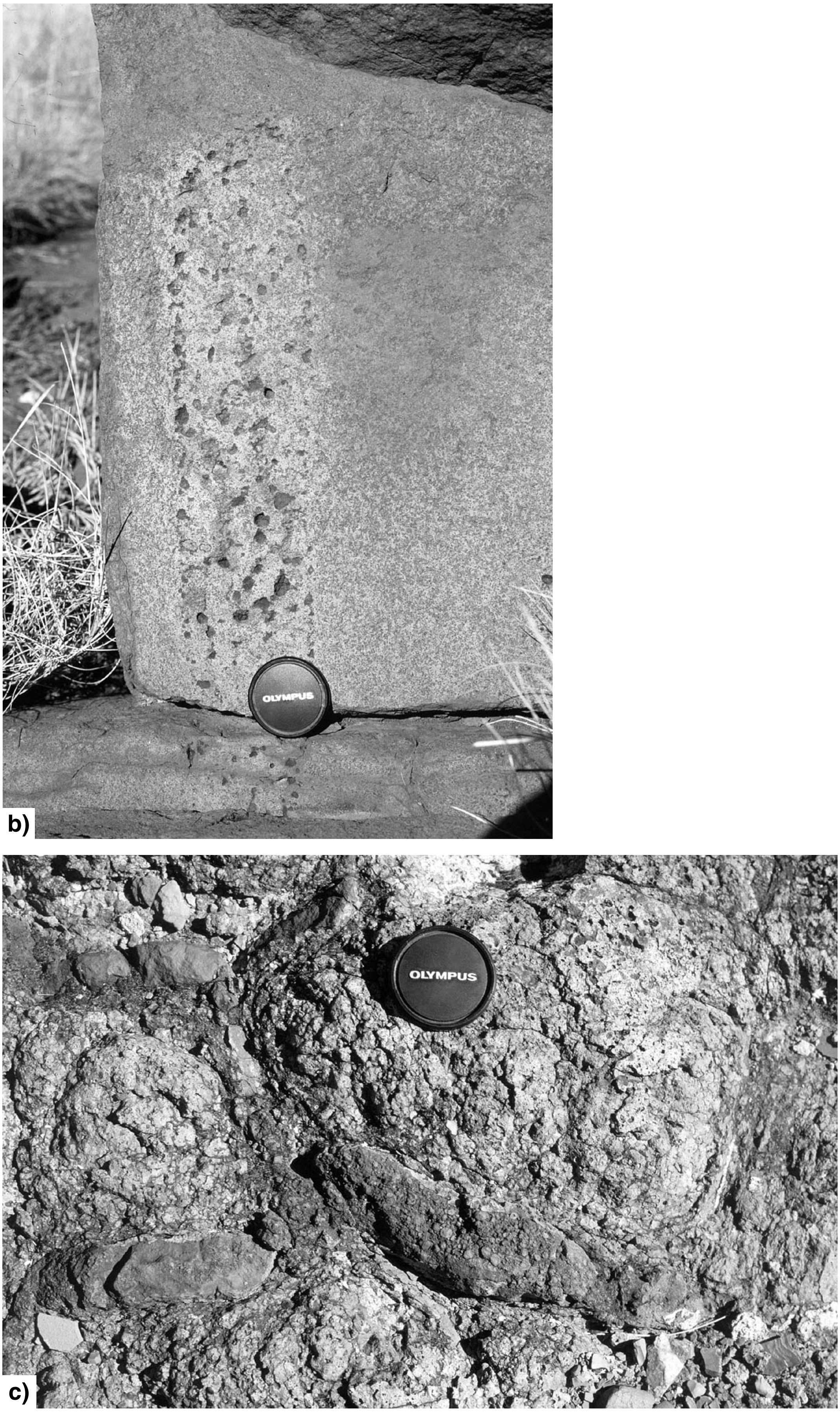

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Figure 8.2. (continued)

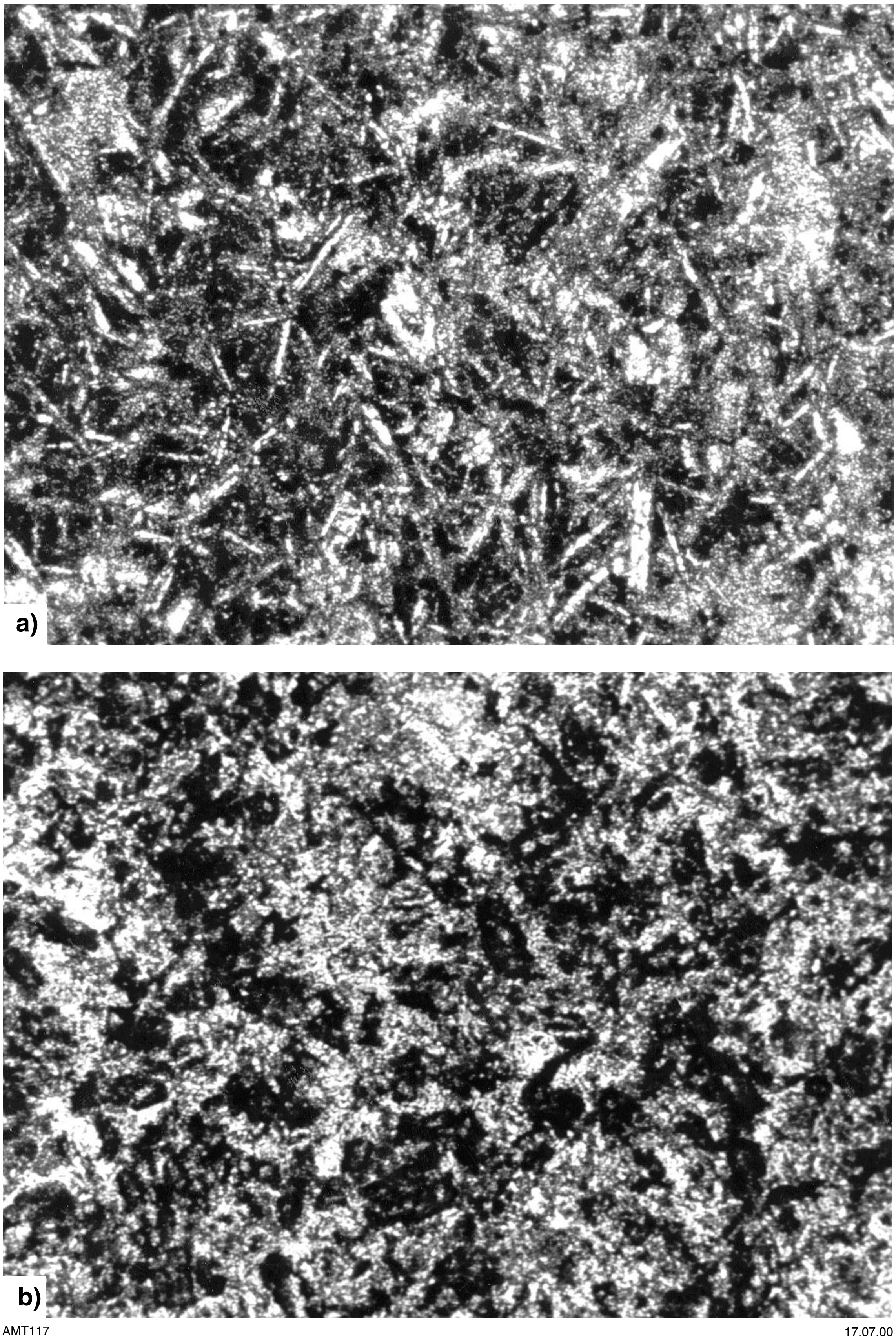

Figure 8.3. a) Photomicrograph of Maddina Formation basalt showing intersertal texture; b) Photomicrograph of aphyric basalt from the Bunjinah Formation. Plane polarized light, field of view = 3.5 mm

exist on feldspar sites. Epidote–clinozoisite exists sporadically in the groundmass, in vesicles, or on original feldspar. Prehnite is similarly distributed but not as abundant, whereas pumpellyite occurs locally within albite. The degree of alteration varies throughout the flow; characteristically, the most extensive replacement by pumpellyite and quartz occurs near the flow top.

The second association recognized by Smith (1975b) lacks pumpellyite but is placed in the prehnite–pumpellyite facies because of the close proximity of the two associations. In general, some combination of the following authigenic minerals is seen: (albite)–(epidote)–chlorite–sericite–carbonate–quartz–sphene. In contrast to the first association, albitization is very slight and calcic plagioclase relics are characteristic, although some show minor alteration to sericite. Pyroxene remains as relics or shows some degree of alteration to chlorite carbonate.

Geochemistry: Eighty-one analyses are available of subaerial basaltic flows from the Maddina Formation, all from the northern outcrop area. They include, in terms of the scheme used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group, the 17 analyses included in the 4-letter groups MARS (Smith et al., 1982), MAGS (Gudrun Sieber) and MAPS (Pindana Spring). All these are from the northwest Pilbara sub-basin, and form cluster H in the figures of **Chapter 12**. From the northwest Pilbara sub-basin and the Gregory Range inlier a further 63 analyses are available. These comprise the 34 analyses of cluster I, which includes the MAHC (Hays Creek), MAMT (Meentheena) and MAGR (Gregory Range) 4-letter groups; and the 29 analyses of cluster G, which includes the NYHC (Hays Creek) and NYMT (Meentheena) groups. The bulk of all these analyses (from Meentheena and Hays Creek) are derived from the Pilbara Volcanic Study (see **Chapter 1**).

Although there is substantially greater compositional scatter within the 81 analyses of subaerial mafic flows of this major eruptive unit than within those of the last major mafic unit below (the Kylena and Boongal Formations of **Chapter 6**) the flows of the Maddina and Bunjinah Formations maintain the basaltic andesitic composition (in terms of the silica/total alkalies scheme of Le Maitre, 1989) that characterizes the bulk of the stacked mafic lavas of the Fortescue Group (Fig. 12.6). The silica content is mainly close to 57%, significantly higher than that of Phanerozoic flood basalts. There are suggestions of locally higher percentages in groups MAGS and MAPS, in the northwest, tending towards an andesitic composition, and in the northeast the NYMT and NYHC groups have a somewhat lower silica, basaltic composition. Data are still inadequate to define a consistent regional trend. On the Jensen Cation Plot (Jensen, 1976; Fig. 12.7) most analyses of clusters G and I plot close to the division between the calc-alkaline and tholeiitic fields, which also divides andesite from basalt. Trace elements again have the enrichment relative to MORB characteristic of later stacked mafic lavas of the Fortescue Group (Fig. 12.9), a common feature of within-plate continental flood basalts. But in terms of the Ti–Zr–Y plot of Pearce and Cann (1973; see Fig. 12.11), these lavas fall mainly in their calc-alkali field (C) rather than the within-plate field (D).

Interpretation: The basaltic composition and aphyric texture of most of the lavas suggest that they originated as low- to moderate-viscosity flows that were close to their liquidus temperatures at eruption (cf. Green, 1989). In addition, the general lack of evidence for contemporaneous explosive fragmentation suggests either that the initial content of exsolved volatiles was low, or that levels were high, but were reduced by non-explosive degassing or by volatile loss during a previous phase of explosive activity (Cas and Wright, 1987).

Very thick (>40 m), columnar jointed, massive to amygdaloidal flows are similar in thickness, internal structure, surface features, and petrographic character to many flows in the Columbia River Basalt (Waters, 1961; Long and Wood, 1986). Their irregular columnar structure, lacking entablature, most closely resembles the Type 1 flows of Long and Wood (1986), although their thickness (up to 70 m) is greater and column diameter (<0.3 m) is smaller than the Columbia River examples. By analogy with these North American lavas, the columnar-jointed massive to amydaloidal flows of the Maddina Formation are interpreted as rapidly emplaced, subaerial, ponded lava flows. Other massive to amygdaloidal flows in the Maddina Formation are often very thick but do not show evidence of columnar jointing. It is likely that these very thick basalts also formed ponded flows but did not develop the necessary thermal stress pattern for joint formation (Spry, 1962; De Graff and Aydin, 1987)

It is not known to what extent many of the thinner (5–20 m) massive-to- amygdaloidal lavas and amygdaloidal lavas in the Maddina Formation represent ponded flows. On the one hand, these basalts are thinner and have a more varied morphology than typical ponded flows, but on the other, they lack many of the features associated with flowing lava, such as lava tubes and surface channels.

Subaqueous basaltic lavas

Physical volcanology: Subaqueous basaltic lavas are the dominant volcanic facies within the Bunjinah Formation. They comprise pillow lava, massive basalt flows, and minor vesicular basalt; all facies are intimately associated with hyaloclastite.

Pillowed units range in thickness from a few metres to many tens of metres. Most are massive, although large-scale, crude lenticular bedding has been observed in some pillowed units. Individual pillows (Fig. 8.4) are up to 2.0 m wide and have a sac-shaped or elongate, finger-like morphology. Elongate forms may show a preferred long-axis orientation locally. Many pillows contain spheroidal, arcuate, or planar convex cavities and radial pipe-amygdales are observed locally. Pillows have an altered glassy selvage up to 25 mm thick, and interpillow cavities are filled by variable amounts of quartz, carbonate, and hyaloclastite.

Massive basalt flows range in thickness from about 5 to over 100 m and have undulatory to planar bounding surfaces. Many flows can be traced laterally and vertically into pillow lava and hyaloclastite. Most flows are either non-vesicular or sparsely vesicular; in the latter situation,

Figure 8.4. Pillow lava exposed in a road cutting 25 km northeast of Paraburdoo, Bunjinah Formation, south Pilbara sub-basin

amygdales are filled by combinations of quartz, carbonate, and chlorite. Thick massive flows are cut by a system of irregular joint planes aligned parallel to, or normal to, flow bounding surfaces. Locally, flows are cross-cut by an irregular network of curved and planar joint surfaces similar to those observed in some thick subaerial lavas. These joints give the rock an irregular, net-veined appearance on bedding surfaces and in cross section.

Vesicular basalt flows are recorded from near the top of the Bunjinah Formation. Flows range in thickness from 0.3 to 15 m and are often characterized by rusty-coloured weathered surfaces. Thin flows are 20–50% vesicular throughout; thicker lavas are only strongly vesicular (up to 60%) near their upper contacts.

Petrography: Most samples of subaqueous facies basalt (Fig. 8.3b) are aphyric and fine to coarse grained (50–2000 μm). They consist of a random framework of ragged, acicular to bladed actinolite–tremolite after original pyroxene, separated by anhedral to subhedral epidote–clinozoisite, quartz, albite, chlorite, calcite, and iron-oxide. The effects of lower greenschist facies metamorphism (Smith, 1975b; Smith et al., 1982) make it difficult to distinguish original intersertal and intergranular textures.

Geochemistry: The 14 available analyses of this facies all come from the Bunjinah Formation of the south Pilbara sub-basin. They include, in terms of the scheme used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group, the two analyses of the 4-letter group BUBC (Bellary Creek), the two analyses of the group BUSG (Corehole SGS1), and the ten analyses of the group BUCA (Cowley's (1979) area). Together they form cluster J (Tables 12.1 and 12.2). The limited data represented by these analyses suggest that the magma of the subaqueous flows of the south Pilbara sub-basin was essentially identical to that whose extrusion in the north formed the stacked subaerial flows of the correlative Maddina Formation; it had the major oxide composition of a basaltic andesite, with trace elements strongly enriched relative to MORB.

Interpretation: The association of pillow lava, massive lava, and hyaloclastite indicates that most, if not all, of the Bunjinah Formation was formed in a subaqueous environment. By analogy with modern subaqueous flows (Moore et al., 1973), most pillow lavas in the Bunjinah Formation formed by the budding of subaqueous lava tubes. This process is comparable to the digital advance of subaerial pahoehoe. Discrete pillows are also found in the Bunjinah Formation and probably formed on steeper slopes where individual lava toes became detatched from the rest of the tube. Ballard et al. (1979) reported the occurrence of massive sheet flows of basaltic lava from the Galapagos Rift Valley. Those flows are associated with pillow lava, and the transition from pillowed to massive lithology probably reflects an increase in basalt discharge rate.

Pillow lavas have been reported from three principal environments: marine, sub-glacial, and lacustrine. In the case of the Bunjinah Formation, deposition in a subglacial setting can be ruled out on the basis of the thickness and lateral extent of the lava units and the absence of other evidence for a glacial event. A lacustrine setting for the pillow lava is also thought to be unlikely as there are no associated subaerial facies in the Bunjinah Formation or overlying parts of the Fortescue Group. In addition, the lateral extent of subaqueous facies would necessitate the former presence of a lake covering an area of at least 500 × 100 km^2. In view of these constraints, it is thought most likely that the Bunjinah Formation was deposited in a submarine environment.

Strongly vesicular flows, despite their similarity to subaerial lavas in the Maddina Formation, are also thought to have been deposited in a submarine setting. Principal evidence for this comes from the nature of the associated rock types, which consist of hyaloclastite, pillow lava, and argillite. The increased vesicularity of these basalts may have resulted from the magma having a higher than usual volatile content. It could also reflect a basin-wide drop in relative sea level, the consequent fall in hydrostatic pressure allowing greater than normal vesiculation.

Volcaniclastic rocks

Description: Volcaniclastic rocks comprise hyaloclastite, mafic tuff, and lapilli tuff. Mafic tuff and lapilli tuff, and associated non-volcanic sedimentary rocks, is an important facies locally within the Maddina Formation; hyaloclastite may be the dominant rock type within parts of the Bunjinah Formation.

Basaltic hyaloclastite is typically associated with pillow lava and massive lava in the Bunjinah Formation but it is also recorded beneath subaerial flows in the Maddina Formation. In the former situation, the hyaloclastite occurs in laterally discontinuous units, ranging from a few metres to over 200 m thick. Most hyaloclastites are massive, very poorly sorted breccias composed of angular fragments of sparsely vesicular to vesicular basalt up to 1.5 m across. Others are finer grained (sand to granule sized) and may exhibit parallel stratification and scour-and-fill structures. Clasts of all sizes are cut by networks of polygonal fractures that vary in width from hair-like cracks to wider veins filled by quartz, chlorite, and carbonate. Even when separated by a few centimetres, many clasts exhibit a jigsaw fit with neighbouring fragments. Detatched pillows or pillow fragments are observed locally in hyaloclastites within the Bunjinah Formation.

Silt- and sand-sized mafic tuff and lapilli tuff occur in friable, grey-green weathering units up to 30 m thick. Units are tabular bedded and internal structure typically consists of horizontal lamination and ripple cross-lamination; small stromatolitic buildups and isolated accretionary lapilli exist within some beds. Thin (0.02–0.25 m) beds of closely packed accretionary lapilli tuff are contained within most mafic tuff units. Most layers are laterally persistent, others are discontinuous and fill hollows and cracks in the underlying surface. Beds have sharp, non-erosive bases and sharp or gradational tops; some are ungraded, others show normal or reverse grading. Accretionary lapilli normally range in size from 2 to 12 mm and most are characterized by a concentric internal structure.

Interpretation: The basaltic breccias are interpreted as hyaloclastites, produced by the rapid cooling and fracturing of basalt during contact with water. This interpretation is based upon the following evidence:

- Their homogeneous, generally non-vesicular clast composition.
- The predominance of poorly sorted, angular clasts and evidence for fracturing in situ.
- Their intimate association with pillow lava.

Hyaloclastite breccias are produced when there is sudden contact between hot, coherent magma and cold water or water-saturated sediment (Cas and Wright, 1987). They form in a variety of situations: where magma is erupted subaqueously or subglacially; where lava flows into water or over water-saturated sediments; or where magma is intruded into water-saturated sediment or country rock. In the case of the breccias described here, geological setting rules out an intrusion-related or subglacial origin. Localized accumulations of hyaloclastite at the base of some subaerial Maddina Formation flows were probably formed by lava either flowing into shallow water or over wet sediment. In most cases, however, hyaloclastite is found in association with pillow lava and massive lava in the Bunjinah Formation. For these units the only interpretation that satisfies all the available evidence is the one in which the breccias formed in a marine environment. The lack of stratification in most hyalocalastites, combined with their matrix support, poor sorting, angular clasts, and presence of transitional boundaries with the basalt flows, suggests that they accumulated close to source, either as a flow-top mantle, or marginal scree and sediment gravity-flow deposits.

Beds of closely packed accretionary lapilli are interpreted as primary pyroclastic fall deposits, either from subaerial hydroclastic eruptions, or from eruptions where the moisture was supplied by rain falling through the eruption cloud (cf. Fisher and Schmincke, 1984). Mafic tuff and lapilli tuff units, which display parallel lamination or ripple cross-lamination and contain localized stromatolitic buildups, are regarded as reworked pyroclastic deposits that accumulated in a shallow marine or lacustrine setting.

Non-basaltic volcanic rocks

A variety of non-basaltic volcanic rocks has been reported from the Maddina Formation in the west Pilbara. Williams (1968) recorded local occurrences of felsic lava on YARRALOOLA, and Smith (1979) described a variety of rhyolitic domes, plugs, flows and breccias on PYRAMID and YARRALOOLA (see **Distribution of facies**).

Kojan and Hickman (1998) noted that the lower part of the Maddina Formation comprises a lower unit of high-Mg basalt, basalt and basaltic andesite, and an upper unit of dacite and rhyolite. The high-Mg basalt of the lower unit is 50–100 m thick, and is characterized by relatively high Cr and MgO values. The upper Maddina Formation consists of a lower unit, with highly variable radiometric values, which is composed of basalt, basaltic andesite and andesite; and an upper unit, with generally very high radiometric values, which is composed of interbedded dacite, rhyolite and basaltic andesite.

Non-volcanic sedimentary facies

Description: Minor occurrences of non-volcanic sedimentary rocks, consisting mostly of quartz sandstone and stromatolitic carbonate, are recorded from the Maddina Formation where they typically occur in association with volcaniclastic facies. In addition, localized occurrences of conglomerate, pebbly sandstone and argillite have been recorded at the base of the Bunjinah Formation on NEWMAN and ROBERTSON.

Fine- to coarse-grained quartz sandstone forms in tabular units up to 7 m thick. Lower contacts, where observed, are erosional, upper contacts are sharp and commonly overlain by a subaerial basalt flow. Internal structure is dominated by planar to undulatory parallel-lamination, small- to medium-scale trough cross-stratification, or ripple cross-lamination. Current lineation is present on some bedding surfaces.

Stromatolitic carbonate, commonly partially silicified, forms 0.05–0.4 m-thick beds or small lenticular buildups within mafic tuff units. Beds are often characterized by very irregular, horizontal to inclined lamination, which merges laterally into small domical, bulbous, or nodular stromatolites. Larger, bulbous forms are similar in gross morphology to *Alcheringa narrina* Walter (1972).

A varied assemblage of clast- and matrix-supported polymict conglomerate, pebbly sandstone, sandstone and argillite is recorded from the base of the Bunjinah Formation along the northern margin of the Sylvania Inlier (Tyler, 1991). The thickness of these deposits varies from about 10 to 500 m and most of the coarser grained rock types appear to have been derived locally.

Interpretation: The compositional and textural maturity of the sandstones and varied assemblage of sedimentary structures indicate that deposition took place in an environment characterized by fluctuating low- to high-energy conditions. Parallel-lamination with current lineation is indicative of transitional to upper regime currents, whereas trough cross-stratification and ripple cross-lamination record dune and ripple migration under lower regime flows (Harms et al., 1982). Although this assemblage of structures has been recorded in sandy fluvial deposits (see review by Collinson, 1986), the tabular form of the units and the nature of the associated facies suggests a shoreline setting is more likely.

A coastal setting is also considered probable for the stromatolitic carbonate units, although they display few features that would allow identification of the specific depositional environment.

The mixed assemblage of conglomerate, sandstone, and argillite recorded at the base of the Bunjinah Formation on NEWMAN and ROBERTSON are interpreted as fan-delta or valley-fill deposits.

Distribution of facies

A summary of the stratigraphy of the Maddina and Bunjinah Formations in the northwest, northeast, and south Pilbara sub-basins is shown in Figure 8.5.

Northwest Pilbara sub-basin

The Maddina Formation forms a continuous outcrop from Cape Preston to the eastern margin of the northwest Pilbara sub-basin. The maximum measured thickness for the formation is 900 m for an incomplete section at Cape Preston; the total thickness in this area is estimated at approximately 1100 m. Five hundred and forty metres of Maddina Formation are recorded in the MF–1 borehole (CRA Exploration Pty Ltd, 1987a). This section, which is also incomplete, is close to the estimated thickness of 600 m for the formation on PYRAMID and YARRALOOLA (Kriewaldt and Ryan, 1967; Williams, 1968).

The Maddina Formation of the northwest Pilbara sub-basin is dominated by subaerial basalt flows with minor intercalated hyaloclastite and quartz sandstone. The formation also contains a significant proportion of volcanic rocks ranging in compositon from rhyolite to high-Mg basalt (Kojan and Hickman, 1998). Many of the dacitic to rhyolitic plugs, domes, and flows reported by Smith (1979) are here regarded as extensively silicified basalt flows and flow-top material for two reasons. Firstly, their flow thickness, morphology, and internal structure are typical of other subaerial basalt flows in the Maddina Formation but are atypical of most acid lavas (cf. Bonnichsen and Kauffman, 1987). Secondly, in many flows the highly siliceous upper levels are gradational downwards into less-altered rock with a relic basaltic mineralogy. Similarly, the siliceous volcanic sinter horizons and fumorolic vents reported by Smith (1979) from the Maddina Formation–Jeerinah Formation contact have been reinterpreted as silicified sedimentary carbonate and sandstone (Packer, 1990; see also **Chapter 9**).

Kojan and Hickman (1998) reported that on PINDERI HILLS, the Maddina Formation has a total estimated vertical thickness of 740 m and can subdivided into lower and upper divisions. The lower Maddina Formation comprises a lower unit of high-Mg basalt, basalt and basaltic andesite, and an upper unit of dacite and rhyolite. The high-Mg basalt of the lower unit is 50–100 m thick, and has a distinctive brown radiometric signature.

The upper Maddina Formation consists of a lower unit, with highly variable radiometric values, which is composed of basalt, basaltic andesite and andesite; and an upper unit, with typically very high radiometric values, which is composed of interbedded dacite, rhyolite and basaltic andesite. This upper unit, which forms a low range of hills on the southern boundary of PINDERI HILLS, has a thickness of 130–200 m (Kojan and Hickman, 1998).

Northeast Pilbara sub-basin

The Maddina Formation of the northeast Pilbara sub-basin, which is equivalent to the combined Nymerina Basalt, Kuruna Siltstone, and Maddina Formation of Hickman (1983), has maximum thicknesses of about

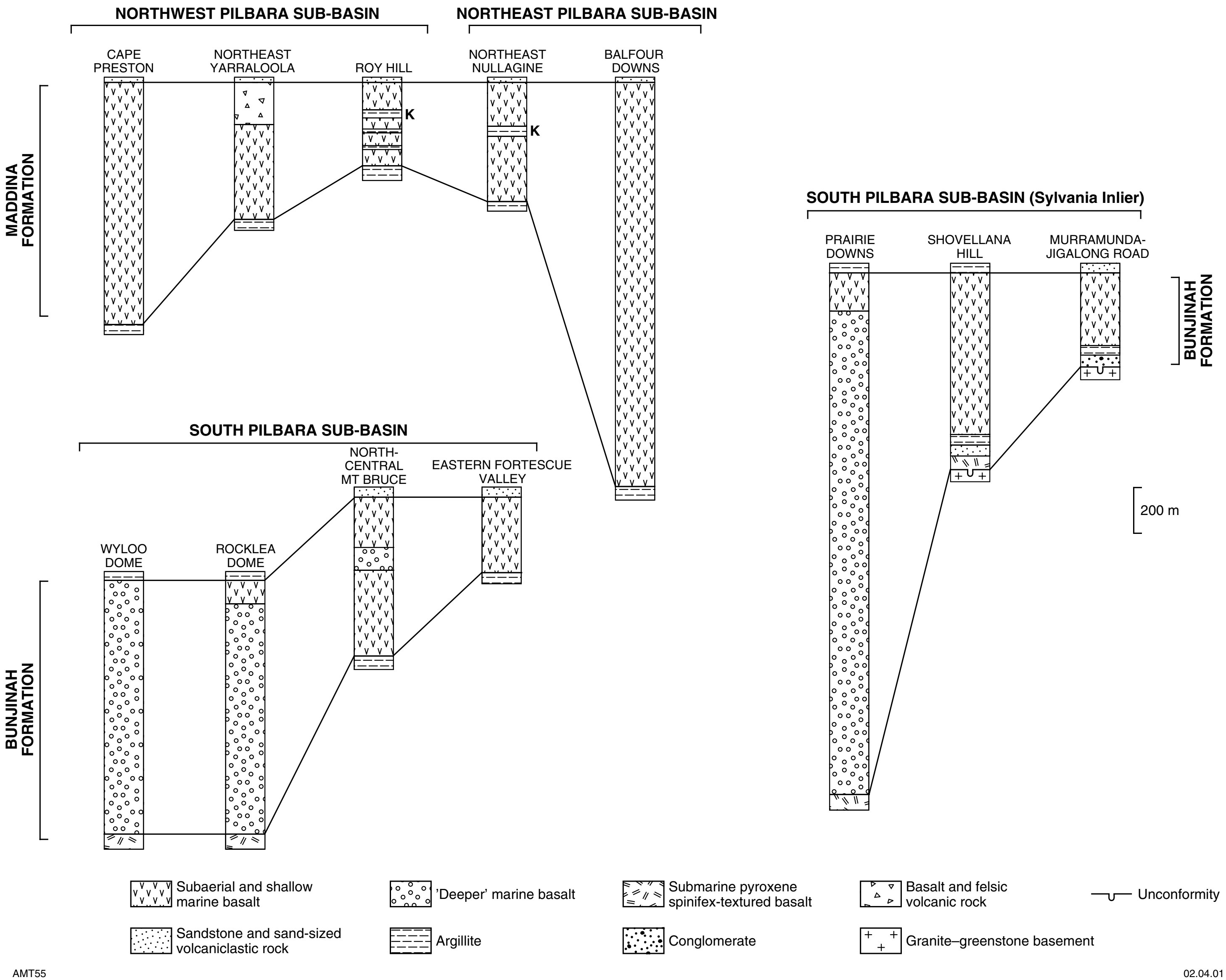

Figure 8.5. Generalized stratigraphy for the Maddina Formation and Bunjinah Formation in the northwest, northeast and south Pilbara sub-basins. K = Kuruna Member

2000 m on BALFOUR DOWNS (Williams, 1989), 500 m on NULLAGINE, and 340 m on ROY HILL.

The formation comprises subaerial basaltic lava flows interbedded with sedimentary deposits consisting of reworked mafic tuff, primary air-fall tuff, stromatolitic carbonate, and quartz sandstone. Three distinctive sedimentary units, ranging in thickness up to 30 m, are recognized on northeastern ROY HILL. The upper and more persistent of the three units is here referred to as the Kuruna Member and is equivalent to the Kuruna Siltstone of Hickman (1983). In most outcrops, the Kuruna Member shows a threefold subdivision into: a lower silt- to sand-sized mafic tuff division; a middle division consisting of thin beds of packed accretionary lapilli, alternating with layers of partly silicified stromatolitic carbonate; and an upper division of silt- to sand-sized mafic tuff, capped locally by quartz sandstone.

The Kuruna Member and underlying sedimentary units are not present on southwestern NULLAGINE or on BALFOUR DOWNS. The Kuruna Member is, however, recorded on northeastern NULLAGINE, where it is 50 m thick and shows the same threefold lithological subdivision as on ROY HILL. Elsewhere, on eastern NULLAGINE, the Kuruna Member is represented by a thin, silicified pisolitic tuff or chert (Hickman, 1978).

South Pilbara sub-basin

In the south Pilbara sub-basin, the Bunjinah Formation is the lateral equivalent of the Maddina Formation. It outcrops around the Bellary, Milli Milli, Rocklea, and Wyloo Domes, the Jeerinah Anticline, and along the northern and southwestern margins of the Sylvania Inlier. The Bunjinah Formation has also been recorded in the FVG–1, WRL–1, SGS–1, and BMW–1 diamond drillholes (CRA Exploration Pty Ltd, 1985b, 1987b, 1988, 1989). Submarine volcanic facies, comprising pillow lava, massive and vesicular lava, and fine- to coarse-grained volcaniclastic rocks are the dominant rock type in most southern outcrops, whereas subaerial and subaqueous facies are recorded in the Jeerinah Anticline and SGS–1 and WRL–1 boreholes.

The Bunjinah Formation has an estimated thickness of 1200 m on the northeastern flank of the Wyloo Dome. No figure is available for the more deformed southwestern part of the Wyloo Dome because this formation cannot be separated from the lithologically similar Pyradie and Boongal Formations. The Bunjinah Formation has a thickness range of 800–1200 m in the Bellary–Milli Milli–Rocklea area, and about 1100 m of strata are present in the Jeerinah Anticline. Pillow lava long axes on MOUNT BRUCE and TUREE CREEK trend mostly west-northwesterly (Fig. 8.6); the direction of budding is unknown.

In the Sylvania Inlier, the Bunjinah Formation is equivalent to the upper part of the upper mafic volcanic unit of Tyler (1991). Around the western part of the inlier the formation is about 2.4 km thick in the Prairie Downs–Deadman Hill area, but thins to around 1700 m north of Spearhole Yard. At Shovelanna Hill, and on the Murramunda–Jigalong road on the northeastern

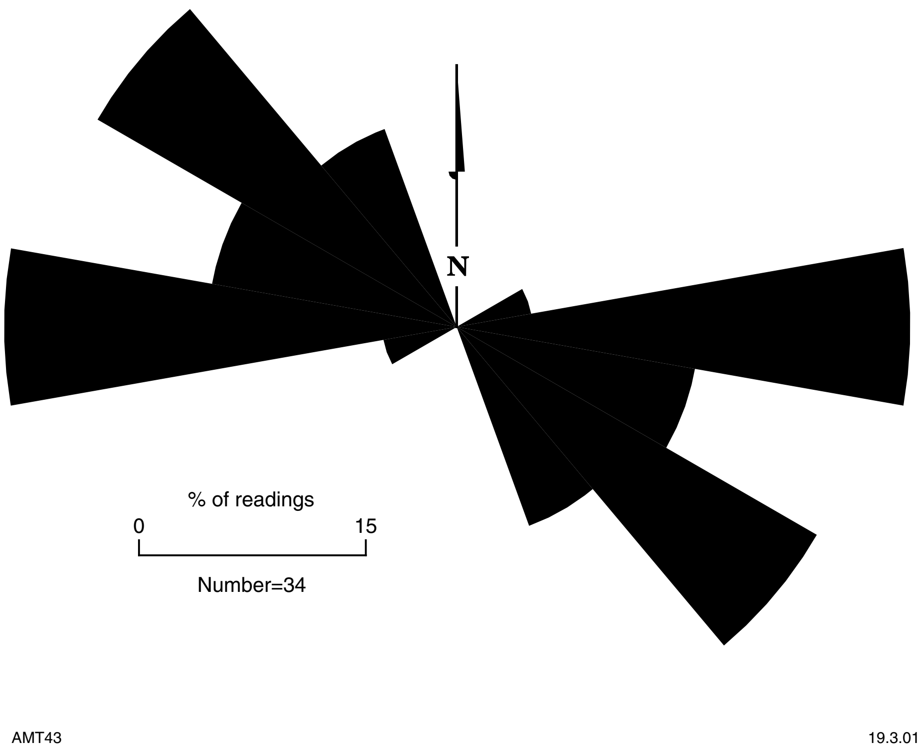

Figure 8.6. Rose diagram for pillow lava long axes in the Bunjinah Formation, south Pilbara sub-basin. In all cases, the direction of pillow lava budding is unknown

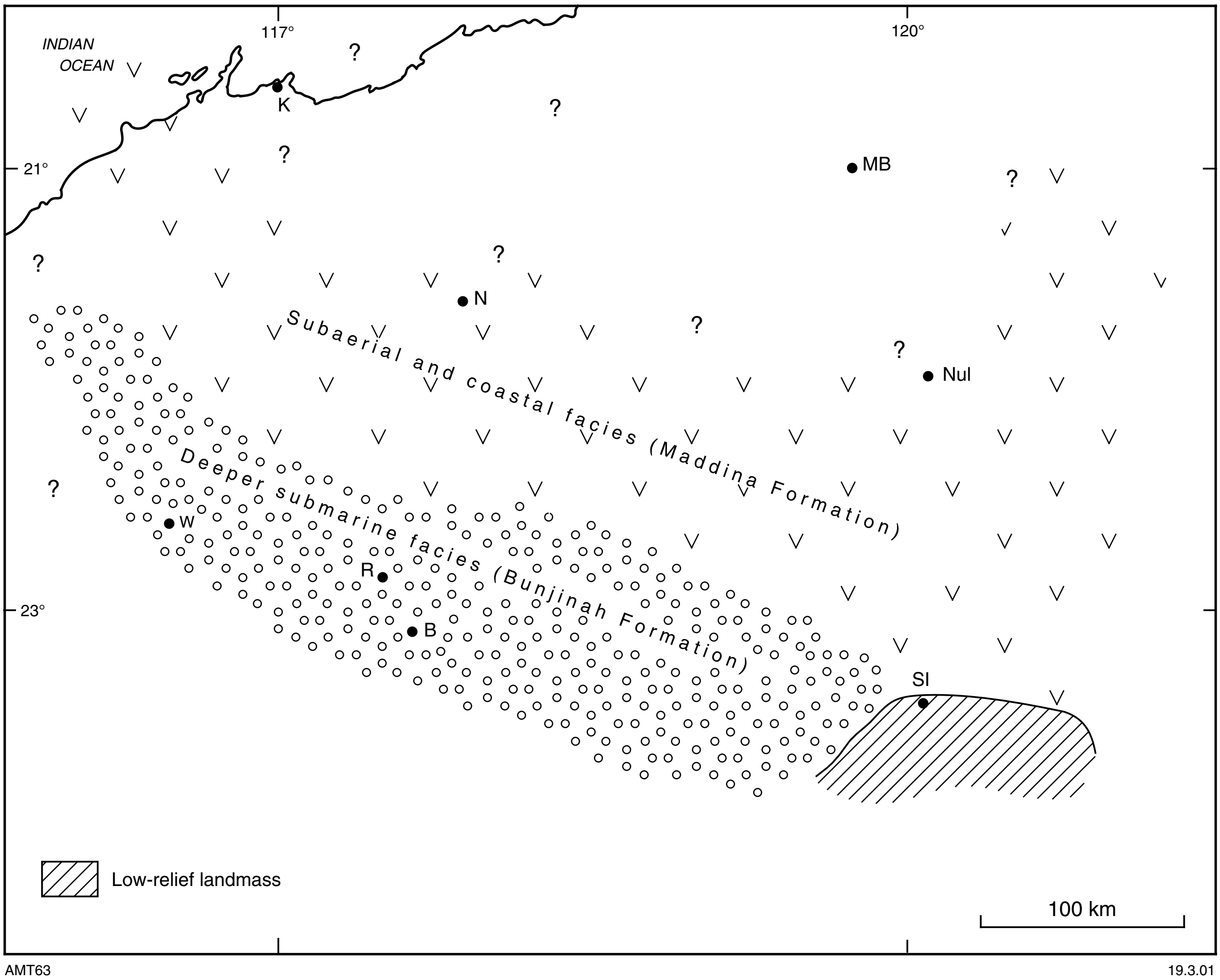

Figure 8.7. Palaeogeography of the Pilbara during deposition of the Maddina Formation and Bunjinah Formation. Locality abbreviations as in Figure 5.15

margin of the inlier, the formation is 900 and 400 m thick respectively. The Spearhole Yard to Murramunda–Jigalong Road occurrences are atypical of the Bunjinah Formation in that they incorporate a lower sedimentary unit, up to 500 m thick, consisting of clast- and matrix-supported polymict conglomerate, pebbly sandstone, and argillite (Tyler, 1991). Clasts are varied; they include locally derived metasedimentary rocks, banded iron-formation, mafic volcanics, and vein quartz. Tyler (1991) noted that the type and nature of the clasts is apparently controlled by local sources within the underlying granite–greenstone. The nature and thickness of the basal sedimentary unit varies along the northern margin of the inlier, particularly east of the Fortescue River. South of Shovelanna Hill, a 10–20 m-thick conglomeratic arkose lies unconformably upon an irregular granitoid surface (Tyler, 1991). Tyler (1991) reported that, 3 km to the east, the basal unit consists of at least 100 m of iron-stained argillite.

Palaeogeography

The palaeogeographic reconstruction for the Maddina Formation and Bunjinah Formation is similar to that for the Kylena and Boongal Formations (Fig. 8.7). Volcanic rocks show a change from subaerial to submarine facies from north to south across the Pilbara and, in addition, there is a general thickening of the succession towards the south (cf. Horwitz, 1980, 1987; Horwitz and Smith, 1978; Morris and Horwitz, 1983). The Yule–Sylvania High, which was a major feature of lower Fortescue Group palaeogeography, appears to have exercised little influence on Maddina Formation and Boongal Formation deposition.

On the northern Pilbara Craton, a relatively thick section of Maddina Formation outcrops between the southern extensions of the Conglomerate Creek Fault and the Nullagine Lineament (Fig. 5.1), suggesting that this north-northeasterly trending fault system may have been active at this time (cf. Blake, 1993). This view is supported by the truncations of the Kuruna Member across the line of the Conglomerate Creek Fault although whether this disappearance is the result of non-deposition or erosion, or a combination of both, is unknown.

Direct evidence for syn-Bunjinah Formation faulting is mostly confined to the Sylvania Inlier. Here, the succession thins markedly when followed northward across the easterly trending Western Creek Fault and eastward across the north-northeasterly trending Fortescue River Fault (Fig. 8.1; Tyler, 1991). The conglomerate–sandstone–argillite assemblage at the base of the Bunjinah Formation along the northern margin of the Sylvania Inlier may represent fan-delta deposits shed from a local basement high. Incomplete outcrop, coupled with a lack of palaeocurrent data, makes it difficult to establish the exact position of the palaeohigh. The fact that Marra Mamba Iron Formation rocks rest unconformably upon granitoid basement at AMG 796525E, 7415365N on NEWMAN suggests that the palaeohigh may have been located south and southeast of Shovelanna Hill.

There is little direct evidence for syn-Bunjinah Formation faulting elsewhere in the South Pilbara sub-basin. Immediately west of Mount Turner (ROCKLEA AMG 359914), vesicular basalt flows in the upper part of the Bunjinah Formation pass sharply into a 350 m-thick pillow-breccia unit when traced across across a small west-northwesterly trending fault. Although the present outcrop pattern indicates a north-block-down displacement, the sense of movement during Bunjinah Formation times is unknown. Syn-Bunjinah Formation structures have not been recognized elsewhere in the south Pilbara sub-basin. As with the Kylena and Boongal Formations, there is a sharp west-northwesterly trending boundary between subaerial and subaqueous facies along the line of the Fortescue Valley. This facies change is accompanied by a marked southward thickening and suggests that an easterly to east-southeasterly trending, south-block-down growth-fault system may have been active in the southern Pilbara at this time Thorne (1990).

Chapter 9

Jeerinah Formation

The Jeerinah Formation is present over most of the central and southern Pilbara Craton, and Gregory Range, but is absent from the Marble Bar sub-basin (Fig. 9.1). In the east central Pilbara, it is equivalent to the greater part of the Lewin Shale (Hickman, 1978). The formation is commonly between 100 and 500 m thick in the central parts of the craton; in the south it is up to 1250 m thick, although up to 50% of this value may consist of coarse-grained mafic sills. The Jeerinah Formation is conformably overlain by the Marra Mamba Iron Formation of the Hamersley Group.

The three-fold subdivision of the Jeerinah Formation into Woodiana Sandstone, Warrie Shale, and Roy Hill Shale Members (MacLeod and de la Hunty, 1966) is not used in this Bulletin. There are two main reasons for this. Firstly, it was found that the subdivision cannot be applied to the South Pilbara sub-basin because of the high proportion of basaltic volcanic rocks in this area. Secondly, in northern parts of the craton it was often very difficult to separate the Warrie and Roy Hill Shale Members, as the thin dolomite that separates these units was often not recognized in the field. The Woodiana Sandstone Member was, however, found to be an easily identifiable, widespread stratigraphic unit in this area. For the purposes of this Bulletin, therefore, the Woodiana Member (equivalent to the Woodiana Sandstone Member) is recognized as the basal unit of the Jeerinah Formation on the northern Pilbara Craton. The other formal division recognized within the Jeerinah Formation is the Nallanaring Member, equivalent to the Nallanaring Volcanic Member of Williams (1968). This name is applied to the predominantly mafic volcanic unit in the upper part of the formation on YARRALOOLA and PYRAMID, although it appears that this subdivision has only local significance.

Three U–Pb zircon ages are available from the upper levels of the Jeerinah Formation (**Chapter 2**). A sample of andesitic ignimbrite from the Nallanaring Member on PYRAMID yielded an age of 2684 ± 6 Ma, whereas a tuffaceous sandstone from TUREE CREEK gave a date of 2690 ± 16 Ma (Arndt et al., 1991). In addition, a SHRIMP age of 2629 ± 5 Ma has been obtained by Nelson et al. (1999) from the very top of the Jeerinah Formation on NEWMAN. No zircon dates are available from the lower part of the Jeerinah Formation. The older age limit for this part of the formation is constrained by a U–Pb zircon date of 2717 ± 2 Ma from a rhyolite in the underlying Maddina Formation (Nelson, 1998; Kojan and Hickman, 1998).

Previous descriptions of the Jeerinah Formation are given by: Daniels (1968, 1970), Daniels and MacLeod (1965), Davy and Hickman (1988), de la Hunty (1965, 1969), Hickman (1978, 1983), Horwitz (1976, 1978), Kriewaldt (1964a,b), Kriewaldt and Ryan (1963, 1967), MacLeod and de la Hunty (1966), Packer (1990), Seymour et al. (1988), Smith (1979), Thorne et al. (1991), Thorne and Tyler (1994, 1996, 1997a,b), Tyler et al. (1991), Williams (1968, 1989), and Williams and Tyler (1989). The formation has also been described under various other formal and informal names (Blake, 1993, Noldart and Wyatt, 1962; Trendall, 1975b, 1990b).

Basal contact

No evidence of angular discordance was observed between the Jeerinah Formation and the underlying Maddina or Bunjinah Formations. Kriewaldt (1964b) noted significant differences between rocks above and below this boundary on the northern craton, an observation reiterated by several later workers (e.g. Hickman, 1978; Trendall, 1983; Williams, 1989). Kriewaldt and Ryan (1967) and Williams (1968) first suggested that parts of this boundary are an unconformity (disconformity), and Horwitz and Smith (1978) described the entire boundary on the northern craton as an unconformity (disconformity). However, as noted by Blake (1993), there is no documented proof of significant erosion of the rocks underlying this contact.

Evidence collected during this study suggests that there was little post-Maddina Formation – pre-Jeerinah Formation erosion on the central Pilbara Craton. The Maddina Formation – Jeerinah Formation contact was examined at nine localities in the northwest and northeast Pilbara sub-basins. At eight of these sites, the Jeerinah Formation was observed to rest upon strongly vesicular basalt flow tops. In the remaining locality the underlying flow, although it is not highly vesicular, contains very large amygdales, found typically in the upper parts of flows in this area.

Although the uppermost Maddina Formation appears to have undergone little erosion prior to deposition of the Jeerinah Formation, it locally displays the following unusual features:

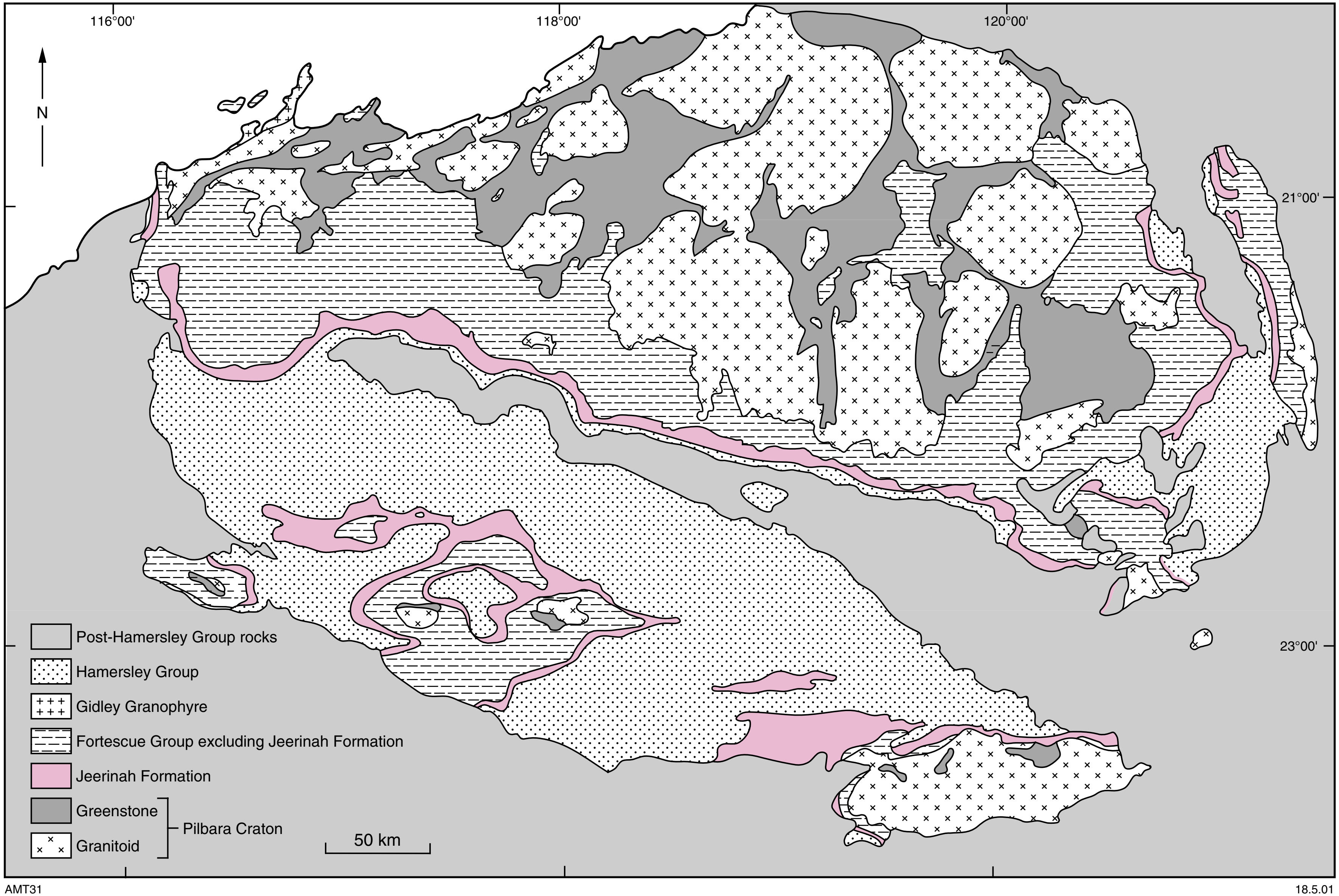

Figure 9.1. Principal outcrop areas of the Jeerinah Formation in the northwest, northeast and south Pilbara sub-basins

- In the northwest Pilbara sub-basin, the upper contact of the Maddina Formation is characterized by large-amplitude (4–12 m), long-wavelength (~80 m) hummocky undulations (Fig. 9.2). Similar undulations are not observed in the underlying basalt stratigraphy. Nor are they present in the Jeerinah Formation above the basal Woodiana Member, for although the lower beds partially drape the underlying topography, mimicking the dome-and-basin outcrop pattern, the overlying units become more planar bedded as they onlap successively higher parts of the flow surface. These relationships suggest that the hummocky upper contact of the Maddina Formation is unlikely to have formed as a result of post-Fortescue Group deformation. Post-Maddina Formation – pre-Jeerinah Formation erosion is also considered an unlikely explanation because vesicular flow-top lithologies are present along the entire upper contact of the basalt. Instead, the most probable explanation is that the hummocky upper contact is a primary surface feature of the uppermost basalt flow and became preserved beneath the Jeerinah Formation.
- The basalt flow immediately below the Jeerinah Formation contact is commonly characterized by a ferruginous, arenaceous texture. This sedimentary appearance is commonly accentuated by the local presence of closely spaced, bedding-parallel joints, which give the rock a laminated appearance. In thin section, the relic basalt mineralogy, although silicified and albitized, can still be clearly recognized. The rusty appearance is caused by the presence of aggregates of subhedral to anhedral iron oxide. Either these are scattered throughout the rock as an apparent replacement of primary ferromagnesian minerals, or they are concentrated along, and overprint, the bedding-parallel partings. Although it is mostly silicified, the lowest Jeerinah Formation does not show any of these features.
- The pseudo-bedding appears to be a primary feature of the basalt flows. Macdonald (1967) noted that some basalts develop prominent joints aligned parallel to the flow planes, and that these may be so closely spaced that they give the rock a sheeted or shaly appearance. They are interpreted as the result of late-stage basalt movement; slight shearing has occurred along flow planes after the basalt became partly solidified. The exact timing of the ferruginous alteration is uncertain. It post-dates the silicification and development of the bedding-parallel fabric in the basalt, but it is unclear whether the alteration occurred prior to, or following, deposition of the Jeerinah Formation. The fact that the lowermost Jeerinah Formation was affected by major silicification suggests, however, that the ferruginous alteration also took place after the Maddina Formation was buried.

In summary, the evidence presented above suggests that there was little post- Maddina Formation–pre-Jeerinah Formation erosion on the central Pilbara Craton.

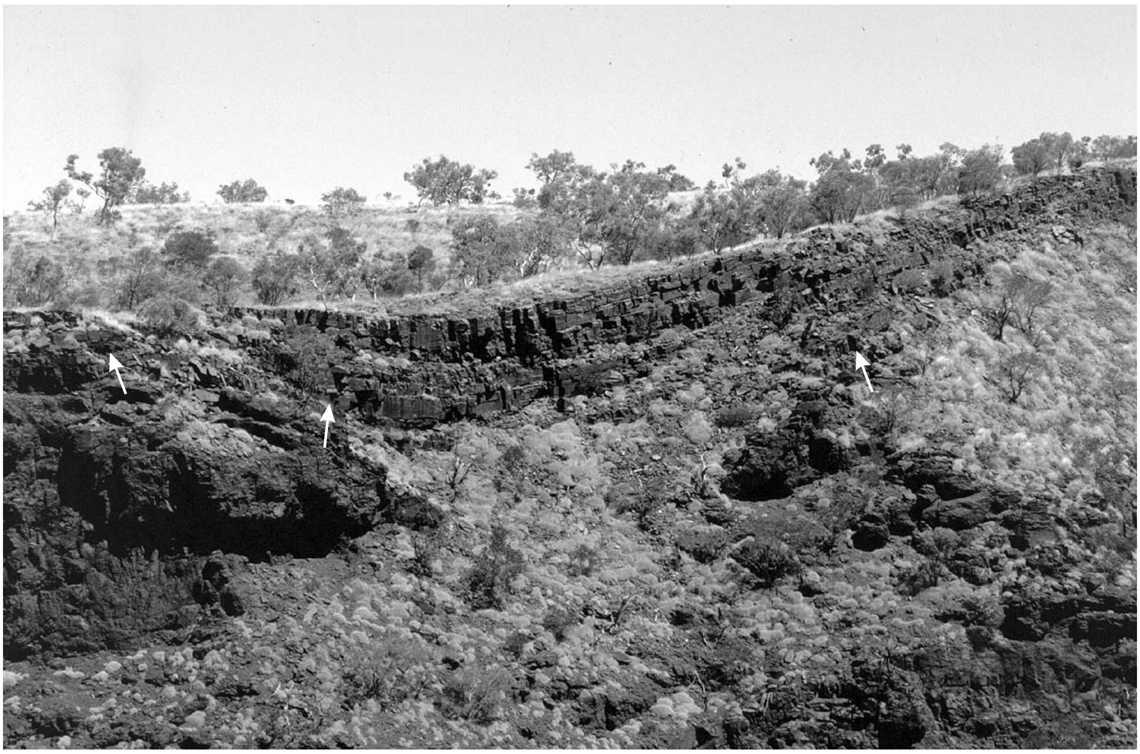

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Figure 9.2. Basal Jeerinah Formation (Woodiana Member) overlying large-scale hummocky flow top of Maddina Formation, Gregory Gorge, northwest Pilbara sub-basin. Contact is marked by arrows

Moreover, certain unusual features of this contact, such as the large hummocks, bedding-parallel fabric, and ferruginous alteration appear to be either primary features of the underlying basalt flows or the result of post-Jeerinah Formation modification. It is still possible, however, that this interval represents a significant time-break, although this cannot be proved using the currently available geochronological data. Even if no large time-break is involved, the Maddina Formation–Jeerinah Formation contact clearly represents an important boundary within the Fortescue Group, as it marks the craton-wide change from subaerial to submarine conditions. As such, this contact represents a major marine-flooding surface (Van Wagoner et al., 1988) and is correlated with the marine Bunjinah Formation – Jeerinah Formation contact of the southern Pilbara Craton (Fig. 9.3).

Facies analysis

This section describes and interprets the principal volcanic and sedimentary facies associations present in the Jeerinah Formation. Most of the data have been obtained from three measured sections and reconnaissance studies in the northwest and northeast Pilbara sub-basins, and 1:40 000-scale mapping of selected outcrop areas in the South Pilbara sub-basin.

The Jeerinah Formation consists of argillite, sandstone, dolomite, chert, and a variety of volcanic rocks including basalt flows, pillow lava, fine- to coarse-grained mafic volcaniclastic, and felsic volcaniclastic rocks. Argillaceous sedimentary deposits are the dominant lithology in northern areas, whereas basaltic lava and volcaniclastic rocks are abundant in parts of the southern Pilbara.

Non-volcanic sedimentary facies

Non-volcanic sedimentary rocks can be grouped into two major facies: nearshore, and offshore. The former characterizes the lower part of the Jeerinah Formation, whereas the latter constitutes most of the middle to upper levels.

Nearshore facies

Nearshore facies comprise quartz sandstone and subordinate argillite, laminated and stromatolitic chert, and chert breccia. These rocks are the main component of the Woodiana Member, the 15–60 m-thick basal unit of the Jeerinah Formation on northern parts of the Pilbara Craton.

Sandstone lithofacies: The sandstone lithofacies forms a 15–60 m-thick, probably sheet-like body at the base of the Jeerinah Formation in the northwest and northeast Pilbara sub-basins. Lower contacts with the Maddina Formation are sharp, with evidence of minor erosion; upper contacts are usually transitional, with thin-bedded sandstone and argillite. Sandstones are fine to very coarse grained and, using the scheme of Folk (1974), are classified as subarkose and sublitharenite. Most have undergone some degree of silicification.

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Figure 9.3. Basal Jeerinah Formation argillite overlying massive to amygdaloidal Bunjinah Formation basalt flows, Milli Milli Dome, south Pilbara sub-basin

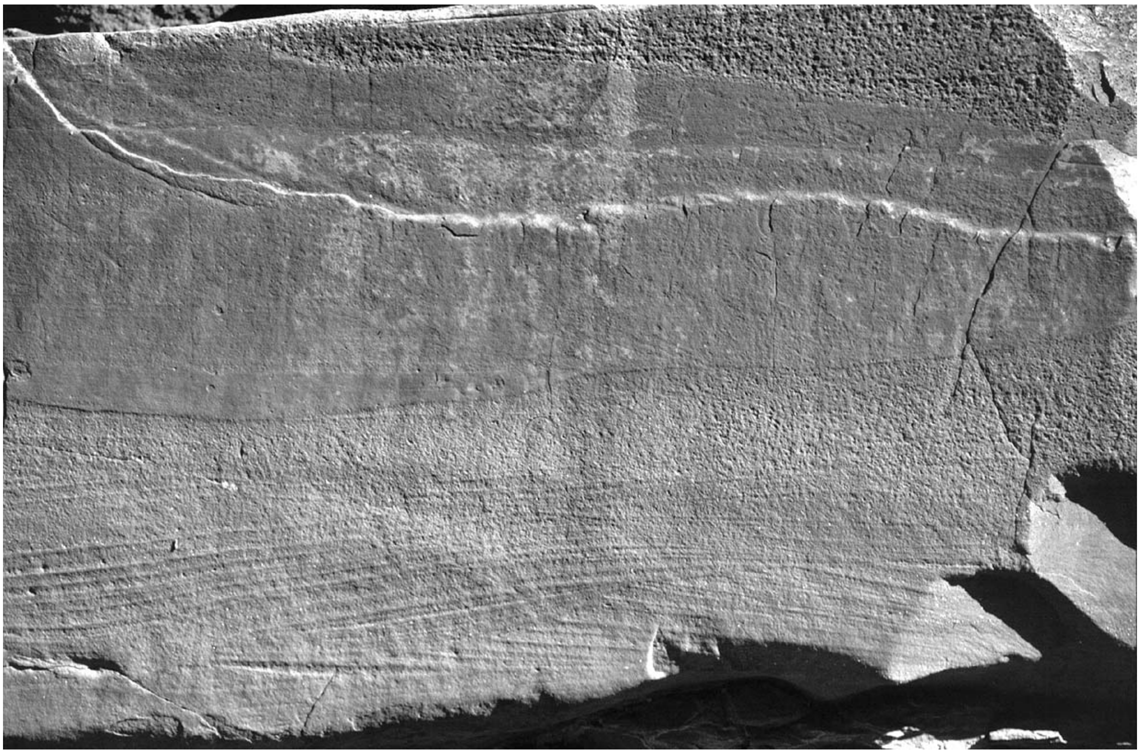
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Figure 9.4. Undulatory (hummocky) cross-stratification in Woodiana Member, nearshore facies. Gregory Gorge, northwest Pilbara sub-basin. Width of view is 1.2 m

Sandstone beds are mostly 0.05–1.5 m thick. The internal structure of the sandstone units includes a diverse range of sedimentary structures. Thick-bedded sandstones contain stacked sets of undulatory parallel-stratification and hummocky cross-stratification in up to 2 m-thick cosets (Fig. 9.4). Some lower set and coset boundaries are marked by chert granule or pebble lags; others display symmetrical and asymmetrical straight-crested ripples. Isolated sets of planar tabular cross-stratification up to 0.8 m thick and medium-scale trough cross-stratification occur locally and generally record complex flow patterns.

Thin-bedded sandstones are interlayered with argillite and the proportion of the latter generally increases up section. Sandstone beds are tabular to sublenticular with sharp erosive bases and gradational tops. Many are normally graded. The thickest beds have undulatory lamination, others are planar parallel-laminated and low-angle ripple cross-lamination is observed locally.

Hummocky cross-stratification is a form of medium- to large-scale cross-stratification, in which the undulating and gently dipping nature of the laminae preserve a three-dimensional bedform comprising shifting hummocks and depressions (Johnson and Baldwin, 1986). Although this structure has not been observed either in the natural environment or experimentally, most workers conclude that it forms in response to high-energy currents with a strong oscillatory component (Harms, 1975; Dott and Bourgeois, 1982; Hunter and Clifton, 1982). To date, hummocky cross-stratification has been described most frequently from shallow-marine environments, where deposition occurred above the storm-wave base (Elliott, 1986; Johnson and Baldwin, 1986).

In the idealized hummocky sequence of Dott and Bourgeois (1982), the basal erosion surface is overlain successively by (a) the main hummocky cross-stratified unit, (b) wave ripple-laminated sand, and finally by (c) an argillite capping. This vertical assemblage is thought to reflect the transition from the main period of storm erosion and deposition (erosion surface and unit a), through a waning storm stage (unit b), to a post-storm or fair-weather stage (unit c). This idealized sequence is relatively uncommon in thick-bedded parts of the Woodiana Member; instead, the internal organization of these units is similar to the cut-out and lag-type amalgamated sequences of Dott and Bourgeois (1982). These formed when the upper (b) and (c) divisions were removed during subsequent storm events.

Thin-bedded sandstone and argillite units are interpreted as storm and fair-weather deposits respectively. The thickness and internal structure of the sandstones indicates that they were deposited in an environment characterized by relatively little sand and weak to moderate storm-generated currents. By analogy with the storm assemblages recognized by Dott and Bourgeois (1982), this setting was probably deeper and farther offshore than that in which the thick-bedded sandstones formed.

Chert lithofacies: Packer and Walter (1986) and Packer (1990) described the chert succession that exists within the lower Jeerinah Formation at Tambrey on PYRAMID. The chert deposits occur in large, mostly 50–400 m-diameter, saucer-shaped erosional remnants on the undulating upper surface of the Maddina Formation. Five major chert lithofacies were recognized:

1. Chert breccia with flat clasts ranging from a few millimetres to 0.75 m in length and commonly 10–20 mm in width. Clasts are white to grey and have internal lamination; they are cemented by coarsely crystalline quartz. Rhombohedral carbonate pseudomorphs occur on some bedding planes. Chert breccia may grade laterally and vertically into chert with a contorted, undulatory lamination; together, they form composite units up to 12 m thick.

2. Black, irregularly laminated chert, with or without stromatolitic relief. It is highly discontinuous and commonly grades into silicified quartzofeldspathic sandstone. Two forms of stromatolite are recognized within this lithofacies: (a) cumulate to bulbous, slightly inclined columns that branch frequently and are anastomosed (Fig. 9.5a,b); and (b) tabular biostromes of stubby to bulbous, closely spaced columns that do not show branching.

3. Red weathering, volcanic and quartzofeldspathic detritus with chert matrix. Internal structure consists of low-angle cross-stratification; some units are brecciated.

4. Coarsely crystalline laminated chert with a botryoidal texture and abundant rhombic and cubic-crystal casts. The latter are aligned perpendicular or parallel to bedding.

5. White to grey weathering, volcanic and quartzofeldspathic detritus with chert matrix. This lithofacies is similar to (3) but is less iron rich. Wave ripples are recognized on bedding surfaces.

All lithofacies are impersistent, and their stratigraphic order varies from site to site. For example, lithofacies 1 commonly lies at the base of the chert succession and is overlain by various combinations of the remaining rock types. Locally, however, lithofacies 2, 3 and 5 form the basal unit and lithofacies 1 is found higher in the succession.

Kerogen ($\delta^{13}C$) values for the Tambrey cherts average -45.59, whereas sulfur ($\delta^{34}S$) values average -2.51 (Packer, 1990)

The Tambrey chert facies have previously been interpreted as hot-spring deposits (Smith, 1979; Packer and Walter, 1986). This interpretation has since been questioned by Packer (1990), who notes that their morphology, distribution, sedimentology, biota, and isotope geochemistry are unlike modern hot-spring deposits but instead are probably silicified shallow-water carbonate and siliciclastic sediments.

Two lines of evidence suggest a secondary origin for the chert (Packer, 1990). Firstly, the high proportion of quartzofeldspathic detrital material (10–90%) in the chert suggests much of it is altered terrigenous sand. This is supported by the presence of cross-stratification and wave ripples, which indicate it was deposited as non-cohesive particulate sediment. Secondly, petrographic examination indicates that some of the chert was originally carbonate.

Much of the lateral variation in lithology appears to be due to the fact that sediment deposition took place in shallow water on an undulating basalt topography. Lithofacies distributions suggest that a discontinuous quartzofeldspathic sand sheet mantled the underlying basalt surface. Local stromatolitic carbonate build-ups developed within the sand sheet and probably also upon the basalt floor. These initial differences in rock type were subsequently exaggerated by the selective nature of silica replacement. Silicification probably also contributed to some of the brecciation (Packer, 1990).

Offshore facies

Offshore facies comprise mudstone and siltstone, with subordinate chert, carbonate, and sandstone (Fig. 9.6). On northern parts of the Pilbara Craton, these rocks are the main component of the Jeerinah Formation above the Woodiana Member. In the south Pilbara, they are present throughout the stratigraphy and are interbedded with volcanic facies.

Description: Argillaceous units tend to be parallel-laminated or, more rarely, ripple cross-laminated, and weather to a white, pink, or red colour. Fresh material is black and finely laminated with thin layers of chert, sandstone, carbonate, and massive sulfide.

A summary of the mineralogy and geochemistry of argillite core, taken from the upper part of the Jeerinah Formation on Mount Bruce, is given by Davy (1986) and Davy and Hickman (1988). The silt- to fine sand-sized component consists of quartz, chert, carbonate, volcanic lithic fragments, and K-feldspar. The finer component, identified by X-ray diffraction, is largely made up of chlorite, dolomite, K-feldspar, magnetite, minnesotaite, mica, pyrite, quartz, siderite, and stilpnomelane. In terms of chemical composition, the rocks are unusually rich in K_2O (up to 9.9%) and have a high K_2O:Na_2O ratio. Free carbon is present in levels ranging from 1 to 14.5%; subeconomic amounts of Zn and anomalous Cu have also been recorded (Davy and Hickman, 1988).

Thin beds and pods of chert are interlayered with the argillaceous units. Most are finely laminated and white to grey or black.

Thin chert-carbonate layers within the argillite units consist of alternating laminae of green (chlorite-rich), white (chert- and carbonate-rich) and black (argillite) layers. The carbonate component is principally dolomite and siderite (Davy and Hickman, 1988). Thicker dolomite beds are tabular to nodular with wavy stylolitic lamination. Although most are recrystallized, some appear to contain relic domical stromatolites, ripple cross-lamination, and structureless dolorudite.

Sandstone (felsic volcanic arenite) occurs in isolated thin (5–20 mm) layers within the argillaceous units or, more rarely, forms thin- to thick-bedded lenticular to tabular units up to 10 m thick. Thin sandstone layers are

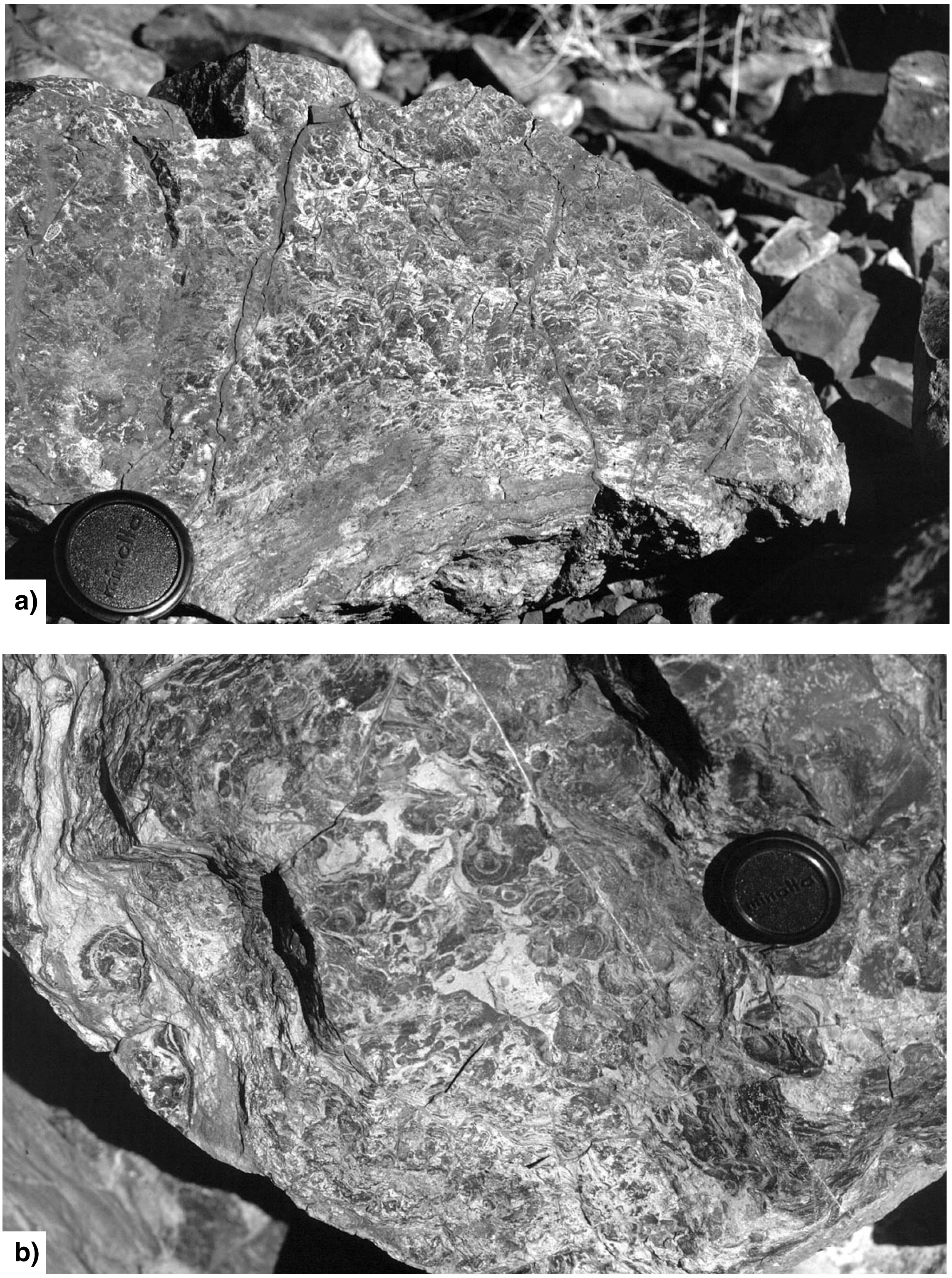

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Figure 9.5. a) Cross section of cumulate microcolumnar stromatolites, nearshore facies (chert lithofacies), Tambrey, northwest Pilbara sub-basin; b) Transverse section of cumulate microcolumnar stromatolites, nearshore facies (chert lithofacies), Tambrey, northwest Pilbara sub-basin

normally graded or non-graded and may display faint parallel-lamination or ripple cross-lamination. Medium- to thick-bedded, fine- to very coarse grained sandstones have erosive bases and are commonly normally graded. Some show Bouma T_{ab} or T_{abc} divisions (Fig. 9.7), although small-scale hummocky cross-stratification is developed locally instead of the T_b division. Some medium- to thick-bedded units can be traced laterally into

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Figure 9.6. Parallel-laminated argillite and cherty argillite, offshore facies, Milli Milli Dome, south Pilbara sub-basin. Jacobs staff is 1.5 m long

amalgamated, very thick bedded (0.6–1.0 m), massive to parallel-stratified sandstone.

Interpretation: The fine grain size, parallel lamination, and lack of scouring suggest that the mudstones were laid down from suspension in a low-energy environment. Parallel-laminated siltstones are also interpreted as suspension deposits for similar reasons. Siltstones that display ripple cross-lamination were formed under lower flow regime conditions, when there was low to moderate sediment fall-out from suspension. The chemistry and mineralogy of the argillite suggest that deposition of clastic material, including quartz sand, felsic to intermediate tuff, and organic detritus, was accompanied by carbonate precipitation (Davy and Hickman, 1988).

Finely laminated cherts are interpreted as chemical precipitates that were laid down during periods of low clastic supply (cf. McConchie, 1984). Many of the thinner dolomite beds could also have originated as chemical precipitates; others that contain ripple cross-lamination or dolorudite layers represent former clastic material.

Sandstones which display incomplete Bouma sequences are interpreted as waning turbidity current deposits (cf. Harms and Fahnestock, 1965; Walker, 1967). The graded T_a division forms in response to very rapid settling of sediment at the base of the flow. The overlying T_b or plane-parallel laminated and ripple cross-laminated divisions are the result of sediment traction under upper and lower flow regime conditions respectively. The presence of hummocky cross-stratification in place of the T_b division in some beds points to a strong oscillatory (storm wave) component in these flows (Dott and Bourgeois, 1982). The precise method of deposition of the massive sandstones is not clearly understood. Walker (1970) suggested that beds of this type are products of deposition from turbidity currents; others, reviewed by Rupke (1978), have stressed the importance during deposition of alternative processes such as grain–grain interaction and movement of pore fluids.

The various rock types discussed above were deposited in an environment where deposition of mud and silt from suspension and weak currents was interrupted by infrequent periods of higher energy sand sedimentation. A marine rather than lacustrine setting is favoured beca use of the large area (90 000 km^2) over which these rocks are exposed. The presence of hummocky cross-stratification in some sandstones suggests that water depths were probably close to the limit of about 200 m

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Figure 9.7. Normally graded turbidite sandstone with well-developed ripple-laminated division, offshore facies, Paraburdoo, south Pilbara sub-basin

for storm-wave reworking on modern marine shelves (Johnson and Baldwin, 1986).

Volcanic facies

Volcanic rocks within the Jeerinah Formation are dominated by subaqueous facies comprising basaltic pillow lava and massive lava, basaltic volcaniclastic rocks, and felsic volcaniclastic rocks.

Subaqueous basaltic lavas

Description: Subaqueous basaltic lavas are the principal volcanic facies within the Jeerinah Formation. They comprise pillow lava and massive basalt flows, and are intimately associated with hyaloclastite.

Pillowed units range in thickness from a few metres to many tens of metres. Most are massive, although large-scale, crude lenticular bedding has been observed in some pillowed units. Individual pillows are up to 1.2 m wide and have a sac-shaped or elongate, finger-like morphology. Elongate forms may show a preferred long-axis orientation locally. Many pillows contain spheroidal, arcuate, or planar convex cavities, and radial pipe-amygdales are observed locally. Pillows have an altered glassy selvage up to 20 mm thick and interpillow cavities are filled by variable amounts of quartz, carbonate, and hyaloclastite.

Massive basalt flows range in thickness from about 5 to over 50 m and have undulatory to planar bounding surfaces. Many flows can be traced laterally and vertically into pillow lava and hyaloclastite. Most flows are either non-vesicular or sparsely vesicular; in the latter situation, amygdales are filled by combinations of quartz, carbonate, and chlorite. Thick massive flows are cut by a set of irregular joint planes aligned either parallel or normal to flow-bounding surfaces. Locally, flows are cross-cut by an irregular network of curved and planar joint surfaces; these joints give the rock an irregular, net-veined appearance on bedding surfaces and in cross section.

As summarized in **Chapter 11**, the chemical composition of the mafic sills of the Jeerinah Formation is consistently and markedly different from that characteristic of the stacked mafic flows of all the underlying units of the Fortescue Group except the Pyradie Formation. Unfortunately, no chemical analyses are available of Jeerinah Formation lavas; knowledge of their consanguinity, or otherwise, with the sills would be important for fixing the timing of this important change in available magma chemistry in the development of the basin.

Interpretation: The association of pillow lava, massive lava, and hyaloclastite indicates that all the basaltic rocks of the Jeerinah Formation accumulated in a subaqueous environment. By analogy with modern subaqueous flows (Moore et al., 1973), most pillow lavas in the Jeerinah Formation formed by the budding of subaqueous lava tubes, a process that is comparable to the digital advance of subaerial pahoehoe. Ballard et al. (1979) reported the occurrence of massive sheet flows of basaltic lava from the Galapagos Rift Valley. These flows are associated with pillow lava and the transition from pillowed to massive lithology probably reflects an increase in basalt discharge rate.

Pillow lavas have been reported from three principal environments: marine, subglacial, and lacustrine. In the case of the Jeerinah Formation, deposition in a sub-glacial setting can be ruled out on the basis of the thickness and lateral extent of the lava units and the absence of other evidence for a glacial event. A lacustrine setting for the pillow lava is also thought to be unlikely as there are no associated subaerial facies in the Jeerinah Formation or overlying parts of the Fortescue Group. Furthermore, the lateral extent of subaqueous facies would necessitate the former presence of a lake covering an area of at least 90 000 km^2. In view of these constraints, it is thought most likely that the Jeerinah Formation was deposited in a submarine environment.

Volcaniclastic rocks

Description: Volcaniclastic rocks comprise basaltic hyaloclastite, and felsic tuff and lapilli tuff.

Basaltic hyaloclastite is mostly associated with pillow lava and massive lava. It occurs in laterally discontinuous units, ranging from a few metres to several tens of metres in thickness. Most hyaloclastites are massive, very poorly sorted breccias composed of angular fragments of sparsely vesicular to vesicular basalt, up to 0.5 m across. Others are finer grained (sand to granule sized) and may exhibit parallel stratification and scour-and-fill structures. Clasts of all sizes are cut by networks of polygonal fractures, which vary in width from hair-like cracks to wider veins filled by quartz, chlorite, and carbonate. Even when separated by a few centimetres, many clasts exhibit a jigsaw fit with neighbouring fragments. Detatched pillows or pillow fragments are observed locally in hyaloclastites within the Jeerinah Formation.

Silt- and sand-sized felsic tuff and lapilli tuff occur in white to grey weathering, thin- to thick-bedded units ranging from a few tens of centimetres up to 80 m thick. Units are tabular bedded, and internal structure mostly varies from massive to horizontally laminated or ripple cross-laminated; small-scale hummocky cross-stratification is also present locally. Arndt et al. (1991) described a 7 m-thick massive felsic unit interbedded with chert and argillite. The unit is not internally stratified, but from about 1 m above the base to within about 0.5 m of the top there is a slight fissility, parallel to the plane of the sheet, related to the parallel orientation of thin, darker discoidal fragments up to a few centimetres across. In thin section, the rock has a microcrystalline quartz–feldspar–chlorite matrix in which lies a variety of clasts, up to 3 mm, but mostly smaller than 1 mm. These include angular quartz, quartz that retains its volcanic ß-quartz form, and indeterminate fragments. The macroscopically visible discoid fragments appear in thin section as subparallel, grossly compacted, vesiculated pumice flakes up to 8 mm long and about 0.3 mm wide.

Thin beds of felsic accretionary lapilli tuff are observed locally within the tuff units. At Millstream, in the northwest Pilbara sub-basin, a 10–50 mm-thick coarse tuff grades sharply into a 0.2 m-thick graded accretionary lapilli tuff in which most accretionary lapilli are in the range 3–7 mm. The top of the bed is ripple cross-laminated and overlain by an upward-fining tuff.

Interpretation: The basaltic breccias are interpreted as hyaloclastites, produced by the rapid cooling and fracturing of basalt during contact with water. This interpretation is based upon the following evidence:

- Their homogeneous, typically non-vesicular clast composition.
- The predominance of poorly sorted, angular clasts and evidence for fracturing in situ.
- Their intimate association with pillow lava.

Hyaloclastite breccias are produced when there is sudden contact between hot, coherent magma and cold water or water-saturated sediment (Cas and Wright, 1987). They form in a variety of situations: where magma is erupted subaqueously or subglacially, where lava flows into water or over water-saturated sediments, or where magma is intruded into water-saturated sediment or country rock. In the case of the breccias described here, geological setting rules out an intrusion-related or subglacial origin. Localized accumulations of hyaloclastite at the base of some subaerial Jeerinah flows were probably formed by lava either flowing into shallow water or over wet sediment. In most cases, however, hyaloclastite occurs in association with pillow lava and massive lava in the Jeerinah Formation. For these units, the only interpretation that satisfies all the available evidence is the one in which the breccias formed in a marine environment. The lack of stratification in most hyaloclastites, combined with their matrix support, poor sorting, angular clasts, and presence of transitional boundaries with the basalt flows, suggests that they accumulated close to source, either as a flow-top mantle, or marginal scree and sediment gravity-flow deposits.

Graded beds of accretionary lapilli tuff that occur interlayered with argillaceous marine sediments are interpreted as redeposited air-fall tuff. Their geological setting suggests that the accretionary lapilli formed either from subaerial hydroclastic eruptions, or from eruptions where the moisture was supplied by rain falling through the eruption cloud (cf. Fisher and Schmincke, 1984). The presence of ripple cross-lamination in upper parts of the beds indicates, however, that they are not primary air-fall tuff but have been remobilized by storms or turbidity currents. A similar origin is likely for other tuff beds which display hummocky cross-stratification in their upper levels.

Arndt et al. (1991) interpreted the massive felsic sheet as a stongly welded ignimbrite, whose present matrix represents the glassy welded phase of the original rock.

Facies distribution

Northwest Pilbara sub-basin

The Jeerinah Formation outcrops in a semi-continuous belt from Cape Preston to the eastern margin of the northwest Pilbara sub-basin. The formation ranges in thickness from 150 to 200 m and shows relatively little lateral variation in stratigraphy across the region (Fig. 9.8).

The lowest 15–25 m of the Jeerinah Formation, the Woodiana Member, comprises silicified quartz lithic sandstone with local stromatolitic chert and chert breccia (nearshore facies). Chert facies are most abundant in the area between Millstream and Mount Florence Stations. On PYRAMID and eastern YARRALOOLA, the remainder of the succession consists of black argillite interbedded with minor chert, carbonate, sandstone, and felsic tuff (offshore facies). On western YARRALOOLA, the offshore facies include a 0–60 m-thick unit of massive basaltic lava, pillow lava, and felsic tuff. This unit, named the Nallanaring Member, can be traced from 8 km northeast of the junction of Kumina Creek and Robe River to a point about 5 km southwest of Bilanoo Hill. It reaches its maximum thickness in the area south of Mount Enid (Williams, 1968).

Northeast Pilbara sub-basin

The Jeerinah Formation of the northeast Pilbara sub-basin has a stratigraphy similar to that of the northwest Pilbara sub-basin but is generally thicker, particularly in the east. On ROY HILL, the Jeerinah Formation is about 150 m thick and comprises a lower Woodiana Member (60 m), consisting of nearshore-facies sandstone, and an upper division of offshore-facies argillite interbedded with minor chert, sandstone, and dolomite. The same two-fold subdivision is present on NULLAGINE, although the succession here is up to 500 m thick. Hickman (1978) notes that the formation includes vesicular basalt flows and dolerite sills at Coonabunna Creek. The Jeerinah Formation on BALFOUR DOWNS is up to 800 m thick (Williams, 1989) and can be subdivided into: a lower, 60 m-thick, Woodiana Member, consisting of quartz and lithic sandstone (nearshore facies); a middle basaltic pillow lava and tuff unit, up to 300 m thick; and a 400 m-thick upper division, comprising argillite, chert, and dolomite (offshore facies).

South Pilbara sub-basin

The two-fold subdivision of the Jeerinah Formation into a lower sandstone and an upper, predominantly argillaceous unit is recognized only in the far north of the south Pilbara sub-basin. In most outcrops, the formation consists of a laterally variable assemblage of argillite, massive basaltic lava, pillow lava, hyaloclastite, and minor volcanic arenite, chert, and dolomite. The succession, which is believed to be entirely marine, is intruded by numerous dolerite sills; these constitute up to 60% of the formation thickness locally.

The Jeerinah Formation on WYLOO is 940 m thick and typically has a transitional, conformable contact with the overlying Marra Mamba Iron Formation of the Hamersley Group. Locally, however, these units are reported to be separated by an unconformity (Horwitz, 1978).

On southern MOUNT BRUCE and TUREE CREEK the Jeerinah Formation is mostly between 1000 and 1200 m thick and up to 50% of the non-intrusive stratigraphy

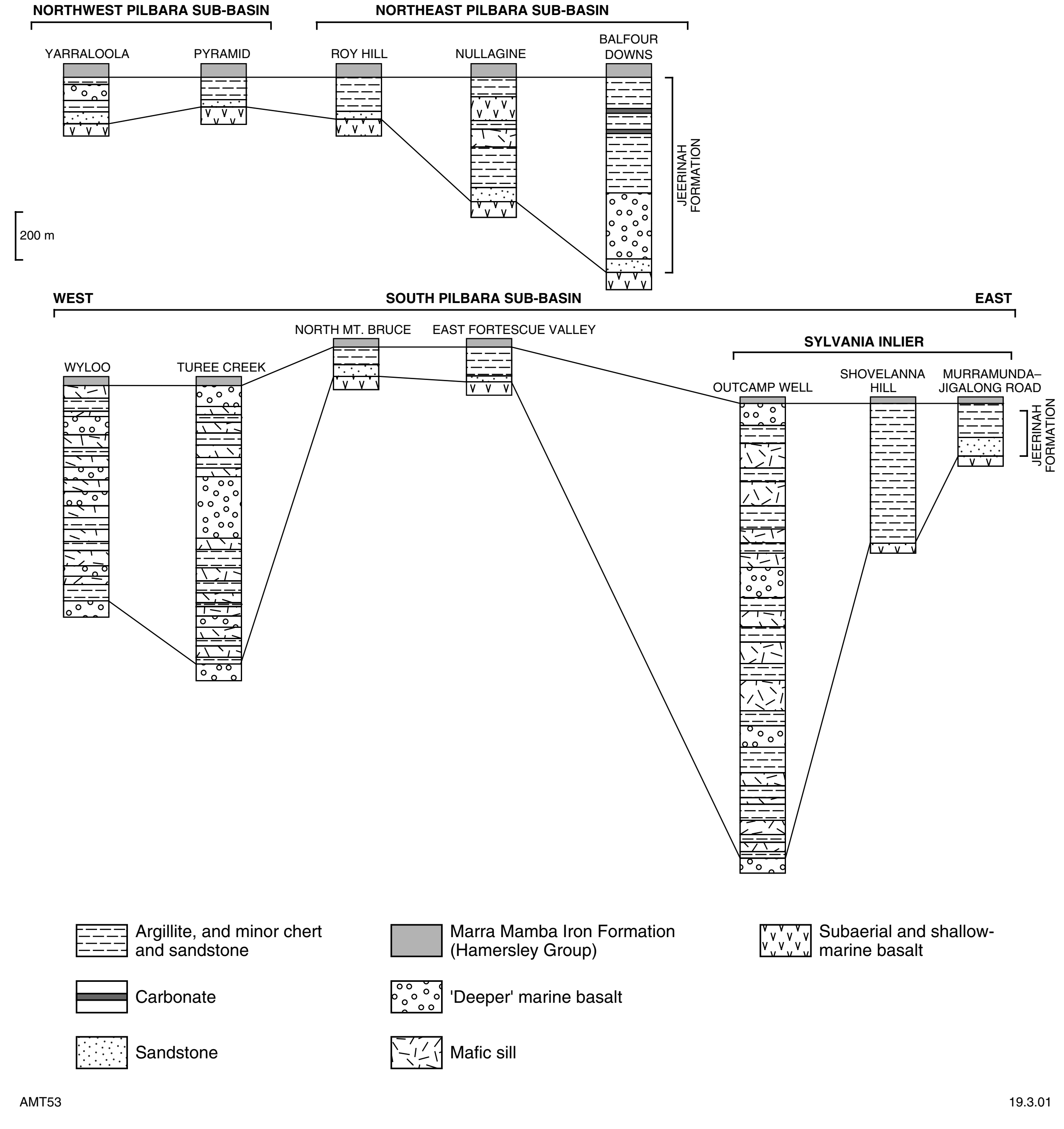

Figure 9.8. Generalized stratigraphy of the Jeerinah Formation in the northwest, northeast and south Pilbara sub-basins

consists of mafic lava, pillow lava, and hyaloclastite. Although most of the succession is variable laterally, mafic volcanic rocks form a persistent 100–200 m-thick unit in the middle of the formation. On parts of northeastern TUREE CREEK, pillow lava directly underlies the Marra Mamba Iron Formation. The thickness of the Jeerinah Formation and proportion of mafic volcanic rock and sills decrease northward on MOUNT BRUCE. In the Jeerinah Anticline, the formation is about 750 m thick and, excluding sills, consists of a lower unit of black argillite, minor chert and sandstone, and an upper unit of felsic tuff and massive to pillowed basalt. In contrast, the equivalent stratigraphy in the FVG–1, WRL–1, and SGS–1 boreholes is only 100–150 m thick and shows the typical northern craton subdivision into a lower sandstone and upper argillaceous assemblage (CRA Exploration Pty Ltd, 1985b, 1987b, 1988).

The Jeerinah Formation on NEWMAN and ROBERTSON has been described by Tyler (1991). In the Prairie Downs – Deadman Hill area, the Jeerinah Formation is dominated by phyllite interbedded with 2–3 m-thick silicified mudstone and chert units. Including mafic sills, the formation is at least 900 m thick.

Tyler (1991) records similar lithologies from the Outcamp Well area, where the formation is 2 km thick. Of this thickness, 850 m is made up of mafic sills. At the

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Figure 9.9. Jeerinah Formation argillite unconformably overlying granitoid basement, Sylvania Inlier, south Pilbara sub-basin

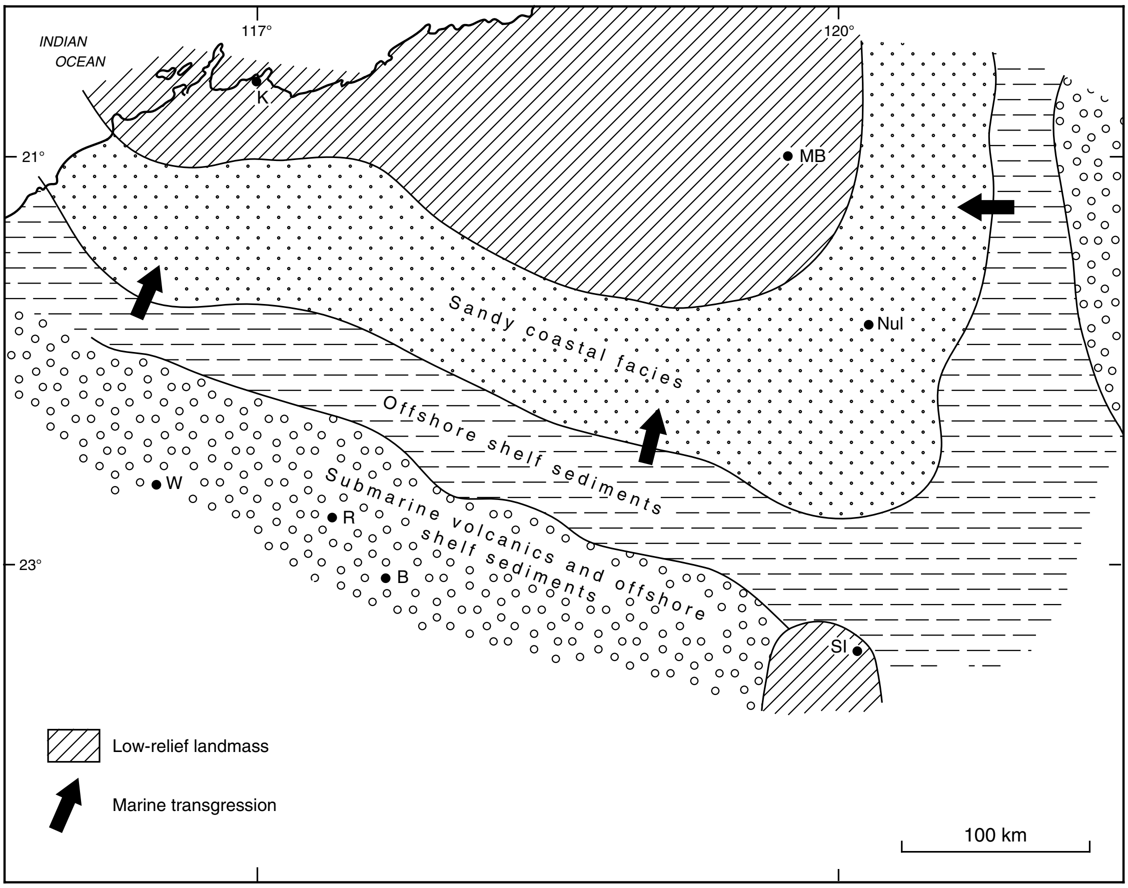


Figure 9.10. Palaeogeography of the Pilbara during deposition of the Jeerinah Formation. Locality abbreviations as in Figure 5.15

base of the formation, Horwitz (1976) recorded 100 m of ‘chertified current-bedded sediment’ overlain by ‘125 m of sandy and gritty, ferruginised quartzitic sandstone and some shales.’ He later correlated these rocks with the Woodiana Sandstone Member of the northern Pilbara Craton (Horwitz, 1987). East of Western Creek the lower part of the Jeerinah Formation comprises interbedded mudstone and laminated siltstones, and no cross-laminated sandstone is recognized. Felsic tuff and thin dolomites occur in the upper part of the unit and are well exposed in the Wonmunna Anticline. Near Mount Whaleback, a basalt that has well-developed pillows is exposed beneath the Marra Mamba Iron Formation (Blockley et al., 1980). This unit is developed in both the Wonmunna and the Alligator Anticlines (Tyler, 1991).

East of the Fortescue River, the Jeerinah Formation is locally unconformable on granite–greenstone basement (Fig. 9.9). Here, mafic sills are absent from the succession and the principal lithologies are phyllite and chert, with minor tuff (Tyler, 1991). Two chert horizons, each up to 5 m thick, are recognized: one towards the base, and the other very near the top. The actual top is marked by a transition from phyllite into cherts distinctive of the lower part of the Marra Mamba Iron Formation.

Farther east, north of Junction Pool, a cross-laminated, medium-grained sandstone appears at the base of the formation. This probably correlates with the Woodiana Member (Tyler, 1991). Above the sandstone is a relatively thin succession of interlayered phyllite and chert.

Palaeogeography

The palaeogeographic reconstruction for the lower Jeerinah Formation (Fig. 9.10) has some similarity with that proposed for the Maddina Formation and Bunjinah Formation. Although subaerial facies are not recorded, nearshore marine-shelf facies dominate this part of the stratigraphy in northern and northeastern parts of the craton. South of a line from Western Creek in the Sylvania Inlier to the northern margin of the Jeerinah Anticline, equivalent levels of the Jeerinah Formation comprise deeper marine argillite and chert, and submarine mafic volcanic rocks.

The upper part of the Jeerinah Formation records a craton-wide change to deeper marine conditions. Argillaceous sediments with minor chert, carbonate, storm-deposited sandstones, and felsic tuff were deposited in the north and northeast. In areas south of the Jeerinah–Western Creek line and on parts of BALFOUR DOWNS, NULLAGINE, and YARRALOOLA, submarine mafic and felsic volcanic material accumulated with the offshore sediments.

No direct evidence of syn-Jeerinah Formation faulting has been recognized. As noted for the immediately underlying units (**Chapters 6–8**), an apparently sharp, northwest-trending boundary separates deeper water argillite and volcanic facies from nearshore deposits in the lower Jeerinah Formation. This boundary, which also marks the northern limit of Jeerinah Formation sills, is parallel to the main tectonic grain in the south Pilbara sub-basin and may, therefore, be fault controlled.

Chapter 10

Gregory Range inlier

The easternmost outcrop area of the Fortescue Group (Plate 1b) forms a belt of dissected rocky terrain, about 120 km long and less than 10 km in average width, along the eastern edge of NULLAGINE, and extending northwards onto YARRIE; this belt is referred to as the Gregory Range inlier (Williams and Trendall, 1998a,b,c). We also use the name Gregory Range area to include, with the inlier, the rocks of the Gregory Granitic Complex immediately to the east.

The eastern limit of Fortescue Group outcrop in the inlier is here taken to be the eastern outcrop limit of the Koongaling Volcanic Member of the Hardey Formation; this member is the lowest stratigraphic unit of the Fortescue Group recognized in the inlier. The western margin of the Gregory Granitic Complex is either gradational into rhyolite of the basal Koongaling Volcanic Member of the Hardey Formation or faulted against higher stratigraphic units of the Fortescue Group. The Gregory Granitic Complex also encloses remnants of metamorphosed sedimentary and volcanic rocks (Warroo Hill Member) that are probably correlative with parts of the Fortescue Group (Williams and Trendall, 1998a).

The western limit of Fortescue Group outcrop in the Gregory Range inlier is defined either by the base of the conformably overlying Carawine Dolomite, correlated with the Wittenoom Formation of the Hamersley Group (Trendall, 1983), or by a tectonic contact with the same formation. All the rocks of the Gregory Range area are unconformably overlain by younger sedimentary rocks, mostly by the late Precambrian Yeneena Group to the east and at the southern end, and by Permian rocks of the Canning Basin to the west. Thus the Fortescue Group of the Gregory Range inlier is cut off from the eastern end of the main northern outcrop area by a finger of Phanerozoic rocks, the Wallal Embayment of the Canning Basin, extending southwards along the valley of the Oakover River.

Within the Gregory Range inlier, the intensity of folding and faulting is such that confident stratigraphic identification is often uncertain. In addition, the stratigraphic succession appears to be closely related to structure, and in a number of respects differs radically from that of any other outcrop area of the Fortescue Group; the area has much the same status as the separately named sub-basins of the main outcrop area. It is for this reason that the inlier is here described in a separate chapter.

Previous work on the Gregory Range area was recorded by Trendall (1991) in a progress report on the work whose final evaluation, in respect to the Fortescue Group, is given here. In the course of our work in support of the preparation of this Bulletin, the Fortescue Group rocks of the five sheets that include parts of the Gregory Range inlier (PEARANA, ISABELLA, BRAESIDE, WARRAWAGINE, and YILGALONG) have been remapped at a scale of 1:100 000, and the relevant part of Plate 1b has been simplified from those maps, the three first named being separately published (Williams and Trendall, 1998a,b,c).

Structure

The Fortescue Group of the Gregory Range inlier is cut by abundant faults subparallel to its length. Trendall (1991) identified a relatively small number of these as main faults, dividing the rocks into tectonically discrete elongate belts, here called ‘slices’. Although continuing mapping since 1990 (Williams and Trendall, 1998a,b,c.) has shown that the number of subparallel faults is greater than originally thought, and that many of the minor ones approach the importance of the five named by Trendall (1991), it is convenient here to retain Trendall’s (1991) major fault and slice names. Thus the five main faults, from west to east, are shown on Figure 10.1 as the Southwest, Antiform, Baramine, Y, and Camel Hump faults. Trendall (1991) described the course of these individually; in many places each is marked by a thick (5–10 m) quartz ‘dyke’ dipping steeply eastwards (Trendall, 1991, figs 4,5), immediately adjacent to which is intense shearing and brecciation. The five slices which they bound, from west to east, are called the Warri Warri, Newdegate, Tanguin, Isabella, and Lochinvar slices.

Apart from the heterogeneity of strain implicit in the existence of fault-bounded slices within which a rational stratigraphy is recognizable, a notable structural feature of the Gregory Range inlier is the marked heterogeneity of strain distribution within the rocks of each slice. This is expressed in two ways, respectively related and unrelated to the stratigraphy.

Throughout the inlier there is a stratigraphically controlled strain heterogeneity related to the physical

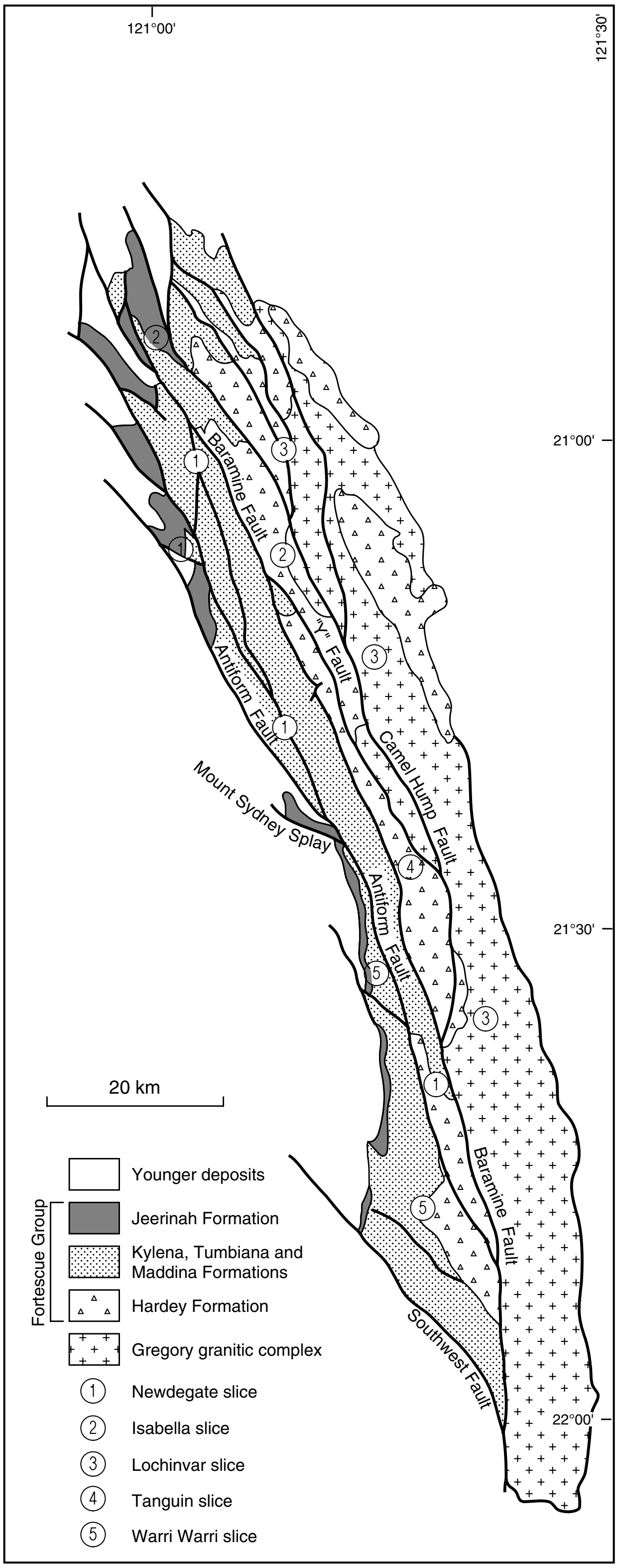


Figure 10.1. Simplified solid geological map of the Gregory Range showing the major fault slices discussed in the text (modified after Trendall, 1991)

properties of the various formations. Thus thick sequences of stacked basaltic flows, such as those of the Kylena and Maddina Formations, rarely show any detectable effects of tectonism. On the other hand, the fine-grained tuffaceous rocks of the Tumbiana Formation characteristically possess a strongly developed near-vertical cleavage with a strike parallel to the length of the inlier, to the extent that bedding is often hard to distinguish. But superimposed on this stratigraphic or, more correctly, lithological relationship is a within-formation strain heterogeneity whose cause is not always clear. This is particularly well shown within the rhyolitic rocks of the Koongaling Volcanic Member, whose strain varies markedly, and erratically, within quite small exposures.

Correlation with the main outcrop area

The stratigraphic nomenclature previously used for the Fortescue Group of the Gregory Range inlier has, because of correlational doubts arising from its isolation, differed in some respects from that applied to the main northern outcrop area of the group. The more significant differences include Hickman's (1978) preference, on the grounds of historical priority, for the name Lewin Shale for the uppermost unit of the group, which we call the Jeerinah Formation. Hickman (1975a, 1978) also applied the name Pearana Basalt to the uppermost unit of stacked basaltic flows, immediately under the Jeerinah Formation, which we call the Maddina Formation; in addition, Hickman (1975a, 1978) introduced the name Koongaling Volcanics for the lowest unit of the group, and suggested its correlation with the Mount Roe Basalt (Hickman, 1983). Except for the choice between the names Jeerinah Formation and Lewin Shale, Trendall (1991) followed Hickman's preferences, but opted specifically not to apply many names from the main outcrop area to avoid the implication of correlation in what was a progress report on the work whose final results are summarized here.

We establish and explain here a preferred stratigraphic nomenclature for the Fortescue Group of the inlier, based not only on its completed 1:100 000 mapping (Williams and Trendall, 1998a,b,c) but also on geochronological results not available to Hickman (Arndt et al., 1991; Nelson, 1996). A primary objective of the scheme adopted is to opt for a simple set of names, with the possible sacrifice of some correlational precision. The alternative course, that of erecting new names wherever correlation is uncertain, tends to result in a nomenclatural scheme which emphasizes strict lithostratigraphic propriety at the expense of understanding.

Our preferred correlation between the inlier and the eastern parts of the main northern outcrop area of the group proceeds from the top down. Thus we accept, as did Hickman (1983, p. 123), the lithostratigraphic equivalence of the Lewin Shale and Jeerinah Formation, and use the latter term. We have accepted the name Baramine Volcanic Member of Hickman et al. (1983) for a volcanic unit within this formation confined to the inlier, but have modified its definition to accomodate the additional Isabella Member (Williams and Trendall, 1998b).

To the sequence of stacked basaltic flows immediately underlying the Jeerinah Formation we apply the name Maddina Formation rather than Pearana Basalt. We do so because the Kuruna Siltstone (MacLeod and de la Hunty, 1966) is clearly a locally restricted unit whose sporadic presence does not justify (Hickman 1975a, 1978) a separate stratigraphic name for lithologically and chemically indistinguishable basaltic flows below (Nymerina) and above (Maddina).

The underlying tuffaceous and carbonate unit is also the obvious lithostratigraphic equivalent of the Tumbiana Formation of the western part of NULLAGINE. The correlation is reinforced by its divisibility in the central and northern areas of the inlier into lower tuff and upper carbonate members, the clear equivalents of the Mingah Tuff Member and Meentheena Carbonate Member of the formation on NULLAGINE (Hickman, 1978).

It is in the stratigraphy of the Fortescue Group below the base of the Tumbiana Formation that serious difficulties arise for confident lithostratigraphic correlation between the inlier and the main outcrop area. In some parts of the inlier, and most particularly in the thick sequence of the Newdegate slice, the unit of stacked basaltic flows below the Tumbiana Formation contains felsic intercalations of much greater extent than those within the Kylena Formation of the main outcrop area. Moreover, between any arbitrarily defined Kylena Formation equivalent in the inlier and the basal Koongaling Volcanics (of Hickman, 1975a, 1978), there are extensive sequences of rocks with no clear lithostratigraphic equivalents in the main outcrop area.

We have resolved these difficulties by noting the ages of 2764 ± 8 and 2760 ± 10 Ma found in the Gregory Range inlier respectively for the top of Hickman's (1978) Koongaling Volcanics and an intrusive sill near the top of the overlying unit determined by Arndt et al. (1991). We also note the age of 2756 ± 8 Ma found by Arndt et al. (1991) for the Bamboo Creek Porphyry (Member) of the Hardey Formation of the main outcrop area. The essential synchroneity of all these felsic magmatic events leads us to apply the name Hardey Formation both to the Koongaling Volcanics (as Koongaling Volcanic Member) and to all sedimentary, igneous, and tuffaceous units overlying the Koongaling Volcanic Member but below a Kylena Formation equivalent.

The main problem raised by our wish to simplify stratigraphic nomenclature in the Gregory Range inlier, by carrying into it the lithostratigraphic names Kylena Formation and Hardey Formation from the main outcrop area, is that of defining the boundary between them. In practice, this was done during 1:100 000 mapping by accepting the base of the Kylena Formation as the base of the first continuous thick sequence of stacked basaltic flows above rocks assigned to the Hardey Formation, as described above. In some cases, noted below, this acceptance has involved arbitrary stratigraphic divisions which may not reflect precise lithostratigraphic correlations with the main outcrop area. But given that such correlation is simply not feasible, we believe that the virtues of the system we have adopted outweigh the difficulties that would inevitably be caused to future workers by the establishment of a plethora of local names.

Our preferred stratigraphic scheme is shown in Figure 10.2, which is amplified by unit-by-unit descriptions later in this chapter, after a slice-by-slice summary of stratigraphic variations.

Relationship between the Fortescue Group and the Gregory Granitic Complex

The Gregory Granitic Complex is an extensive and complex unit of plutonic and hypabyssal granitoid rocks (Williams and Trendall, 1998a,b,c) whose outcrop forms the eastern limit of Fortescue Group outcrop. The complex has been variously interpreted as an intrusive post-Fortescue Group 'Gregory Range Granite' (Noldart and Wyatt, 1962) and as a component of the 'Archaean' pre-Fortescue Group granite–greenstone terrane of the Pilbara Craton (Hickman, 1978), as distinct from the 'Proterozoic' and unconformably overlying Fortescue Group. In an attempt to reconcile the evidence which had led to these two contrasting views, Trendall (1990b, p. 178) suggested that the Gregory Granitic Complex was the plutonic and hypabyssal core of a massive silicic volcanic complex whose extrusive component was the rhyolite of the Koongaling Volcanics (here, the Koongaling Volcanic Member of the Hardey Formation). This intepretation, in which the extrusion of the rhyolite of the Koongaling Volcanic Member is seen as essentially contemporaneous with the emplacement of the underlying hypabyssal and plutonic rocks of the Gregory Granitic Complex, is now strongly supported by SHRIMP zircon U–Pb geochronological results (Nelson, 1996) and the available geochemical data (Williams and Trendall, 1998a,b).

Two plutonic granitoids from the complex, an alkali granite and a gneissic rapakivi granite, gave respective ages of 2762 ± 4 and 2757 ± 5 Ma, and a hyabyssal granophyre gave an age of 2763 ± 8 Ma. The relative ages of the two plutonic rocks are consistent with field evidence that rapakivi granite is the youngest of the plutonic rocks of the complex. The age of the granophyre, which is basal and transitional to the rhyolite of the Koongaling Volcanic Member, is virtually identical to the 2764 ± 8 Ma previously reported by Arndt et al. (1991) for the rhyolite. Both these ages are slightly older than either of the plutonic rocks, thus consistent with the concept that the earliest manifestation of Fortescue Group volcanism in the Gregory Range area was the formation of a thick blanket of silicic lavas whose extrusion was closely followed by intrusion and cooling of cogenetic granitic plutons.

Limited geochemical data show that rocks of the Gregory Granitic Complex and felsic lavas of the Koongaling Volcanic Member of the Hardey Formation are alkaline, both petrologically and chemically (Williams and Trendall, 1998a,b). The lavas of the Koongaling Volcanic Member and granitoid rocks of the Gregory Granitic Complex are silicic (SiO_2 range 71–77 wt%) and potassic (K_2O range 4.9–5.9 wt%). Increasing concentration of SiO_2 is accompanied by a decrease in Al_2O_3, Ba, Sr, and Zr. In terms of tectonic discrimination diagrams, Williams and Trendall (1998a) placed these igneous rocks in the alkaline, A–type granite

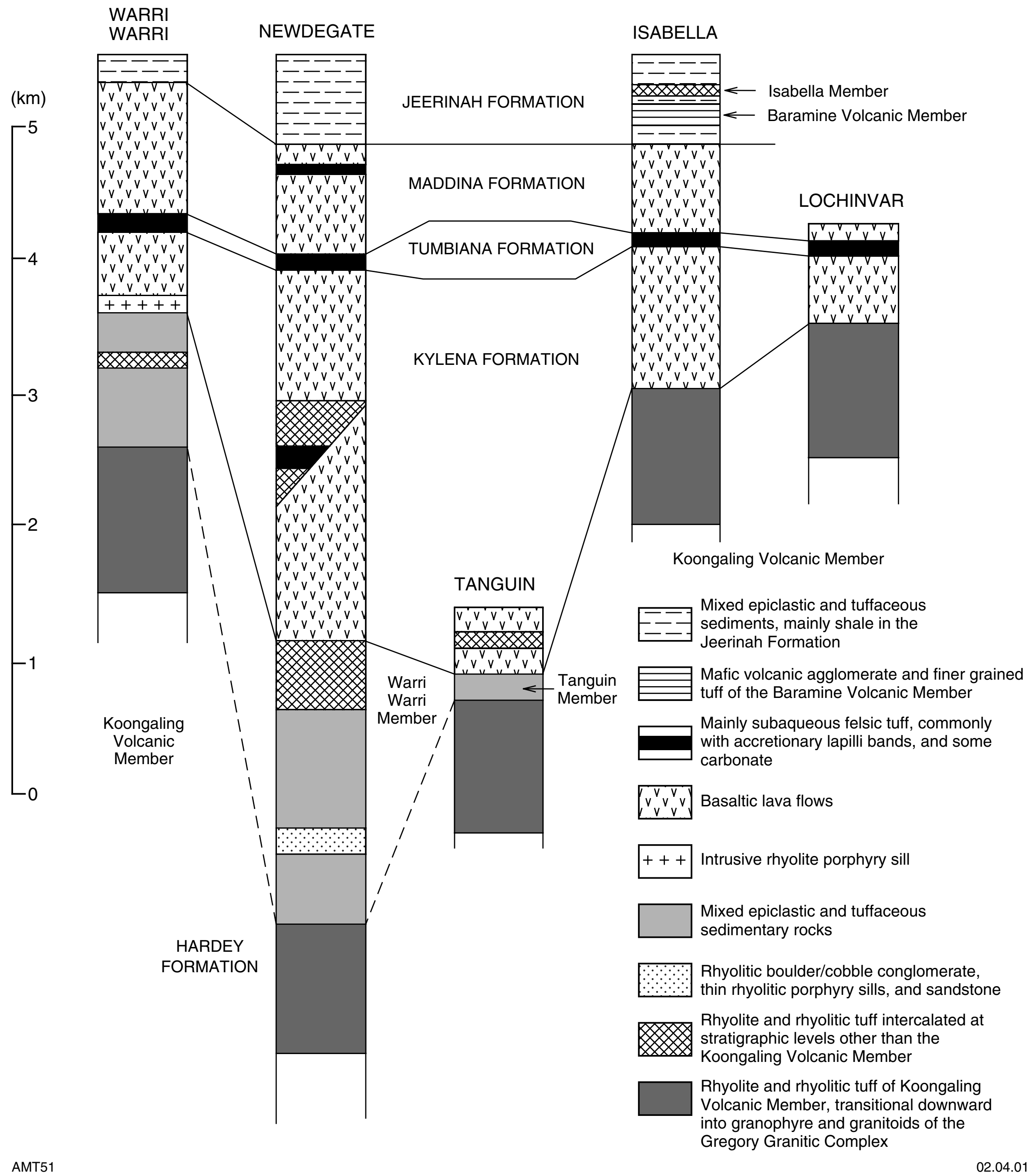


Figure 10.2. Composite lithological and lithostatigraphic summary of the Gregory Range inlier. The five columns show the individual stratigraphic successions in each of the five tectonic slices. Correlative formation names are shown by solid lines between the columns; member boundaries by pecked lines

category (Zr plotted against Ga:Al). Plotting Rb:Nb ratios against Y:Nb ratios shows that the rocks belong to the A_2 group, indicating that the igneous suite has an eclectic origin and is derived from a mixed crustal source rather than a mantle source. Also supporting the A_2 group characteristic of the Gregory Granitic Complex is the occurrence of rapakivi-textured, porphyritic syenogranite plutons. The mixed crustal source for the origin of the felsic volcanic rocks of the Koongaling Volcanic Member is supported by negative ξNd values derived for these rocks (Nelson, D. R. et al., 1992). Hence, a possible source for the Gregory Granitic Complex may be the older granite–greenstone rocks of the Pilbara Craton. It is possible that some of the large xenoliths and rafts of metasedimentary and mafic rocks in the Gregory Granitic Complex may be relicts of this older crustal material, although some of the rafts may also be Fortescue Group rocks. Finally, the tectonic discrimination diagrams of Rb plotted against Y + Nb, and Nb plotted against Y, show that rocks of the Koongaling Volcanic Member and the Gregory Granitic Complex fall in the 'within-plate granite' domain (Williams and Trendall, 1998a,b).

Comparative summaries of the slices

Warri Warri slice

The Warri Warri slice, bounded by the Southwest, Antiform, and Baramine faults, is named after Warri Warri Creek, an eastern tributary of the Oakover River; the slice extends for about 70 km from the southern tip of the Gregory Range inlier to near Mount Sydney. The prevalent gentle westerly dip of the Fortescue Group throughout the southern part of the slice defines the western limb of the gently north-plunging antiform from which the Antiform fault is named, and exposes a well-defined Fortescue Group sequence about 3 km thick. A generalized stratigraphy for the slice is given in Figure 10.2.

The principal stratigraphic features of the succession within the Warri Warri slice are the clear equivalence of the upper part (Jeerinah Formation, Maddina Formation, Tumbiana Formation, and Kylena Formation) with the corresponding formations of the main Fortescue Group outcrop area west of the Wallal Embayment, contrasting with a marked dissimilarity of the lower part, exposed in the southeastern part of the slice. This lower part of the Fortescue Group consists of a thick (~1 km) succession of turbiditic sedimentary rocks which are defined as the Warri Warri Member of the Hardey Formation; the member forms the lowest stratigraphic unit of the Fortescue Group in the Warri Warri slice.

Newdegate slice

The Newdegate slice, named after Mount Newdegate, is bounded by the Antiform and Baramine faults, and extends northwards for some 120 km from the confluence of those faults near Pearana Rock Hole (PEARANA AMG 150850). The stratigraphic sequence within the Newdegate slice, whose thickness of at least 7.5 km exceeds that of any of the other slices of the inlier, is shown in Figure 10.2. As in the contiguous Warri Warri slice to the west, the lowest part of the succession is present at the southern end, and there is a general upward progression northwards, related to a gentle northerly plunge. In the southern part, a consistent, low north-easterly dip represents the eastern limb of the antiform from which the Antiform fault is named. In the central part of the slice the dip is undulating, and often near-horizontal. In the north, the internal structure of the slice becomes increasingly complex, and internal faults parallel with the margins effectively split the Newdegate slice into a package of subslices.

Also, as in the contiguous Warri Warri slice, the formations of the upper part of the Fortescue Group (Jeerinah Formation, Maddina Formation, Tumbiana Formation, and the upper part of the Kylena Formation) are clearly equivalent to the corresponding units of the main Fortescue Group outcrop area west of the Wallal Embayment. However, the rocks of the lower 5 km, extending in this slice from the lowermost Koongaling Volcanic Member of the Hardey Formation upwards to the central part of the Kylena Formation, show substantial lithological differences from either the main outcrop area or the adjacent Warri Warri slice, and arguments for their correlation with the succession of that slice appear below.

Tanguin slice

The Tanguin slice is named from Tanguin Hill. It is the smallest of the five slices named by Trendall (1991), having an average width of only about 2 km between the subparallel Baramine and Y faults, and a length of about 40 km. Its status as a separately identified slice is justified by the presence, between the base of the Kylena Formation and the top of the basal Koongaling Volcanic Member of the Hardey Formation, of a sediment–basalt–rhyolite sequence unlike that of any other slice; this sequence, which has a low northerly dip, is identified as the Tanguin Member of the Hardey Formation. The succession is summarized in Figure 10.2. Fortescue Group units above the Kylena Formation are not present in the Tanguin slice.

Isabella slice

The Isabella slice is named after the Isabella Range, which is formed by the rugged topography of the Koongaling Volcanic Member of the Hardey Formation. The southern part of the slice (south of BRAESIDE AMG 950770) is entirely occupied by the rhyolites of that member, which pass southward (and stratigraphically downward) gradationally into granophyre and granitoid of the Gregory Granitic Complex, within which the identity of the slice becomes lost. Northwards from that latitude the Fortescue Group succession of the slice (Fig. 10.2) is relatively simple, both structurally and stratigraphically. The Kylena Formation directly and conformably overlies the Koongaling Volcanic Member, and is succeeded by the characteristic sequence of the main Fortescue Group outcrop area west of the Wallal Embayment (Tumbiana, Maddina, and Jeerinah Formations). All these formations dip gently northeastwards, and strike obliquely to the length of the slice, in a manner structurally similar to the succession in the southern part of the Newdegate slice. The total thickness of the Fortescue Group in the Isabella slice is about 3.5 km.

Lochinvar slice

The Lochinvar slice exists only at the northern end of the Gregory Range inlier; it is bounded to the west by the Camel Hump fault, and its eastern margin is covered by the unconformably overlying rocks of the Yeneena Group, which define the eastern boundary of the inlier. When Trendall (1991) first used the name he noted that mapping to that date suggested that the stratigraphic sequence within it, apart from the lowermost Koongaling Volcanic Member, could not be matched with those of other slices. Subsequent mapping (Williams and Trendall, 1998a,b) has revised this view, and shown that, as in the contiguous Isabella slice to the west, the succession overlying the Koongaling Volcanic Member is clearly correlative with that of the main Fortescue Group outcrop area west of the Wallal Embayment. However, only the Kylena and Tumbiana Formations are normally present, since the latter unit has acted as an incompetent layer controlling the locations of successive thrusts, subparallel to the length of the slice, which cause repetition of the lower part of the Fortescue Group.

The total thickness of the Fortescue Group of the Lochinvar slice is uncertain, as neither the base of the lowermost unit, the Koongaling Volcanic Member of the Hardey Formation, nor the top of the uppermost exposed unit, the Tumbiana Formation, is exposed. However, the Kylena Formation between these units has an average thickness of about 0.5 km, about half of that in the adjacent Isabella slice to the west.

Formation descriptions and interpretation

Hardey Formation

The application of the name Hardey Formation to the lower units of the Fortescue Group in the Gregory Range area has already been discussed. It is a corollary of that discussion that application of the name implies no certain correlation with any subdivisions of the Hardey Formation of the type area, or elsewhere in the main outcrop area of the group. As summarized in Figure 10.2, the formation is subdivided into four members. The Koongaling Volcanic Member is the lowest of these, and is overlain by the Warri Warri Member. The Tanguin Member is present only in the Tanguin slice, and its correlation with the other two members is uncertain. The Warroo Hill Member is a newly recognized unit of the Hardey Formation (Williams and Trendall, 1998a), which lies east of the Gregory Granitic Complex on BRAESIDE.

Koongaling Volcanic Member

Description: The Koongaling Volcanic Member of the Hardey Formation, the lowest stratigraphic unit of the Fortescue Group present in the Gregory Range inlier (Fig. 10.2), has an extensive and continuous outcrop area along the eastern side of the inlier, within the Lochinvar, Isabella, and Tanguin slices. A small area, tectonically isolated by the Baramine fault, also lies close to the southern end of the Newdegate slice (Plate 1b); Arndt et al. (1991) reported a U–Pb zircon age of 2764 ± 8 Ma from this area of rhyolite near Warri Warri Creek (PEARANA AMG 276897) and gave a petrographic description of the sample.

The unit typically consists of thick, uniform, porphyritic rhyolitic rocks, whose mode of emplacement is largely obscured by a pervasive near-vertical foliation along the length of the inlier. The intensity of development of this foliation is characteristically heterogeneous on a range of scales, and may vary markedly within small, single exposures. Although easily visible in weathered exposures, where a near-vertical lineation is commonly also present, the foliation may be hard to detect in completely fresh rock, which appears flinty and structureless. Primary textures are normally obscured in thin section by extensive recrystallization of the fine-grained matrix, related to low-grade metamorphism. The predominant rock type of the member is a tough, flinty, dark-grey or brown rhyolite in which no original flows or flow structure are normally discernible, although an important exception to this generalization is described below. The rocks are relatively resistant to weathering, and characteristically form broken, poorly accessible terrain with excellent exposure. The rhyolite is normally porphyritic, with K-feldspar phenocrysts up to about 5 mm across.

The foliation is relatively weakly developed within the rhyolite of the narrow north-northwesterly projecting fingers of the Isabella slice, and in some places spectacular and undeformed spherulitic structure is present, for example in the Isabella Range (ISABELLA AMG 950940). The normally pervasive foliation is also very weakly developed in the southern exposures of the member, close to the southern end of the Newdegate slice. In these exposures, undulating terraces 3–5 m high in the rugged hills of the outcrop area have the appearance of flows, but more probably reflect a coarsely sheeted structure acquired by the thick viscous magma body during extrusion. This structure may be related to unusual ovoid concentric structures best developed in this unit on BRAESIDE (Williams and Trendall, 1998a).

These structures, which have a 2:1 north-northwesterly elongation and are of the order of a kilometre in diameter, are particularly well developed within the rhyolite north of Koongaling Hill, in the Isabella slice (BRAESIDE AMG 023692, 028696, and 0207688). Others are present farther north along the slice, and tectonically distorted remnants are scattered elsewhere within the outcrop area of the member. They are well defined on aerial photographs as fingerprint-like structures defined by concentric dark lines, between 15 and 50 m apart, with the average separation about 25 m. On the ground they are more difficult to see, but the paler ground between the dark lines can be distinguished as poorly defined but evenly curving topographic ridges whose orientation is independent of the superimposed drainage pattern and the interesecting joint systems to which that pattern is related. Although there are subtle differences in colour, texture, and weathering susceptibility between hand specimens of rhyolite from the central and marginal parts of each ridge, no clear petrographic distinction is apparent in thin section.

The base of the Koongaling Volcanic Member is nowhere clearly demarcated. The rocks are taken to be gradational downwards into granophyre of the Gregory Granitic Complex, as this rock type is found consistently at the lowest stratigraphic levels within the complex; for example, the areas east of the Woodie Woodie Mining Centre (PEARANA AMG 270060, 160430, and 080775).

The Warri Warri Member of the Hardey Formation appears to overlie the Koongaling Volcanic Member conformably where the latter unit is exposed in the southern part of the Newdegate slice. In this area, the lowest component of the Warri Warri Member, a turbiditic tuffaceous sandstone, dips consistently away from the northern and western rhyolite hills of the Koongaling Volcanic Member. Farther north, the stratigraphic top of the member is clearly identifiable in the Tanguin slice, where it is conformably overlain by tuffaceous sedimentary rocks of the Tanguin Member. In both the Isabella and Lochinvar slices, rhyolite of the Koongaling Volcanic Member is immediately overlain by basal lava flows of the Kylena Formation.

The top of the member is not normally well exposed, but in a small north-draining gully in the northern part of the Isabella slice (BRAESIDE AMG 990787), crudely stratified coarse, mixed agglomerate dips northward at about 20°, and appears to mark the northerly plunging, uppermost part of the unit.

From the extent of its outcrop along almost the whole length of the Gregory Range inlier, coupled with the shallow northerly plunge, the thickness of the Koongaling Volcanic Member is probably at least 2 km, but the structural complexity within and about its outcrop area precludes reliable measurement.

Interpretation: The pervasive foliation and faulting that affect the Koongaling Volcanic Member, as well as the poor exposure of its commonly tectonic boundaries, make interpretation of this unit difficult. The member consists essentially of a stratiform body of 'lavalike' rhyolite that shows little variation in lithology, either along its exposed strike length of about 120 km or through its probable stratigraphic thickness of about 2 km. Trendall (1995) has discussed the various criteria which may be used to determine whether such bodies (Giant Lavalike Felsic Sheets) are extrusive or intrusive, using the Woongarra Rhyolite of the Hamersley Group as an example. The term lavalike (Ekren et al., 1984) is used for silicic volcanic rocks which lack any of the features characteristic of ashflow tuffs, and which are either massive or, if porphyritic, have evenly distributed phenocrysts. It is clear from Trendall's (1995) discussion that the emplacement mechanism of such rocks cannot be reliably determined without exceptional quality of exposure and preservation, especially at the upper contact. This is not the case for the Koongaling Volcanic Member, and in the absence of definitive emplacement evidence it must be accepted that the rhyolite may result from extrusion as lava flows, from the secondary melting of thick ash flows, or from the injection of shallow sills.

The unusual ovoid concentric structures described above from BRAESIDE are believed to relate to spaced shears formed during forcible non-explosive extrusion of highly viscous magma. Such repetitive shearing, or ramping, has been described from recent lavas, and is associated with the formation of ovoid or circular ridges, known as ogives, on the surfaces of rhyolite flows close to the sites of their extrusion (Cas and Wright, 1987; Clough, 1981; Richardson, 1978). Ogives have so far been reported only from modern or recent volcanic rocks, and it is noteworthy that the excellently preserved examples of ogives, or more correctly of their roots, in the Koongaling Volcanic Member, should be in rocks as old as about 2760 Ma.

The presence of these ogives strongly suggests that their constituent rhyolites were close to the surface at the time of formation, and it seems most likely that the whole member represents a complex of flows and interdigitating sills associated with a single pulse of silicic magma effusion. Interpretation of the relationship of the Koongaling Volcanic Member to the Gregory Granitic Complex is discussed in **Chapter 14**.

Warri Warri Member

Description: The Warri Warri Member (Williams and Trendall, 1998c) of the Hardey Formation is present only in the Warri Warri and Newdegate slices. It appears to overlie the Koongaling Volcanic Member conformably where the latter unit is exposed in the southern part of the Newdegate slice. In this area the lowest component of the Warri Warri Member, a turbiditic tuffaceous sandstone, dips consistently away from the northern and western rhyolite hills of the Koongaling Volcanic Member. The anticlinal structure so defined extends westward across the Antiform fault into the Warri Warri slice, and the consistent lithology and structure of this basal sandstone unit of the Warri Warri Member provides compelling evidence for its lithostratigraphic equivalence in both the Warri Warri and Newdegate slices. This basal tuffaceous sandstone is well exposed over a wide area of the headwaters of Warri Warri Creek (PEARANA AMG 280860). It is typically a massive green sandstone with subordinate siltstone in which bedding may not be conspicuous; but close examination of apparently massive exposures usually reveals graded Bouma cycles between 10 and 50 cm thick. Thin sections show a fine-grained matrix recrystallized to biotite and chlorite, and a uniformly fine-grained mixture of both quartz and feldspar, and lithic grains; many of the latter show a trachytic flow texture confirming their volcanic derivation, although such rocks are not known locally in situ. Associated pale-green siltstones are commonly finely laminated.

Intercalated within this tuffaceous sandstone unit in the Warri Warri slice is a stratiform body of massive felsic volcanic rock, about 100 m thick, whose outcrop defines a double fold pattern in the anticline core; no definitive evidence for intrusion or extrusion was found. Towards the top of the sandstone, a rhyolite porphyry sill forms a prominent east-facing scarp along its outcrop. The existence of isolated patches of the turbiditic sandstone unit above the sill confirms its intrusive nature, which is suggested both by its lithological uniformity and by a complete absence of any flow-banding or other textural feature suggestive of extrusion. However, a U–Pb zircon age of 2763 ± 13 Ma reported by Arndt et al. (1991) for a sample from this sill collected in the vicinity of Warri Warri Creek (PEARANA AMG 250930) is within error of the age they determined for the Koongaling Volcanic Member, stratigraphically below, so that rhyolite injection must have closely followed deposition. Although the rhyolite of the sill differs from that of the Koongaling Volcanic Member in having abundant quartz phenocrysts, both rocks evidently belong to the same virtually paradepositional volcanic episode, and the sill is therefore described here rather than as an unrelated intrusive rock.

The tuffaceous sandstone, together with its central massive felsic unit and the porphyry sill close to the top, has a total thickness of about 1 km in the Warri Warri slice. It is conformably overlain either directly by Kylena Formation or locally by hybrid mafic/felsic lavas, included in the Warri Warri Member; such lava is restricted in this slice to the small area centred near the headwaters of Warri Warri Creek (PEARANA AMG 250920). Although this lava resembles that overlying the Kylena Formation in being fine grained, greenish, well jointed and massive, it differs in having a much paler colour, in lacking clearly stacked flows, and in possessing, in thin section, a significant fine-grained quartz content and K-feldspar phenocrysts with textural similarities to those of the Koongaling Volcanic Member and other rhyolite units of the Fortescue Group. Immediately below this lava, a band of tuff at the top of the massive felsic unit has abundant accretionary lapilli, and is locally associated with black manganiferous shale; it is well exposed around Warri Warri Creek (PEARANA AMG 258937).

The part of the Warri Warri Member above the basal turbidite sandstone unit in the Newdegate slice differs radically from that of the Warri Warri slice, both in lithology and thickness. The tuffaceous sandstone and shale pass upwards into a thick-bedded porphyry conglomerate unit. In the predominant oligomictic conglomerate of this unit, well-rounded boulders of rhyolite porphyry are closely packed, and the interstices between them are filled with finer debris of the same material (Trendall, 1991, fig. 6b). The boulders in the coarser conglomerate beds are up to a metre across, but the larger boulders in most conglomerate beds have diameters about 30 cm. This conglomerate unit has a minor sandstone content, and rare tuffs with accretionary lapilli, and is intruded by a number of thin discontinuous sills of porphyry closely similar to porphyry clasts in the conglomerate. The unit is gradational upwards, over a stratigraphic thicknesss of about 170 m, into coarse cross-bedded quartz sandstone; the transition is marked both by a gradual decrease in size and abundance of porphyry clasts and a steady increase in quartz content.

Northwards in the Newdegate slice from the conglomerate unit, whose outcrop is centred east of Warri Warri Creek (PEARANA AMG 280925), there appears to be upward stratigraphic continuity in the exposed rocks, also assigned here to the Warri Warri Member; these have a consistent north-northwesterly strike and east-northeasterly dips of about 10–20°. The rocks include epiclastic sandstone, massive green tuff, hybrid lava like that of the adjacent Warri Warri slice, well-stratified green tuffaceous sandstone with accretionary lapilli, and flow-banded to massive rhyolite which locally is spectacularly spheroidal, with spheroids up to 10 cm in diameter. The rhyolite is capped by agglomerate with mixed angular volcanic clasts. Dolerite sills are concordant within this sequence.

Although exposure is not continuous in the central part of the Newdegate slice, between the oligomictic conglomerate unit and the base of the Kylena Formation, the consistent dip and strike make it reasonable to assume that the sequence is continuous. If this is so, the Warri Warri Member has a total stratigraphic thickness of at least 4 km in the Newdegate slice, some four times thicker than that of the Warri Warri slice. Apart from the argument for the correlation of these two disparate sequences provided by their common position below the Kylena Formation, outcrops of tuff with accretionary lapilli south of the Woodie Woodie Mining Centre (PEARANA AMG 265005) show very close resemblance to those in the Warri Warri slice area (PEARANA AMG 250893).

Most units of the Warri Warri Member have a steeply dipping penetrative cleavage axial to the antiform jointly defined by the Warri Warri and Newdegate slices. The intensity of the cleavage is highly variable, and appears to depend jointly on the competence of the rock and its proximity either to the crest of the antiform or to one or other of the major faults bounding the slices. Thus the oligomictic conglomerate of the Newdegate slice locally shows spectacular flattening of the boulders where it abuts the east side of the Antiform fault (PEARANA AMG 274924). However, in exposures in the hills 2 km east-southeast of this point the cleavage is barely detectable.

Interpretation: The lowermost massive sandstone facies of the Warri Warri Member are interpreted as high-density deposits laid down in a lacustrine or marine environment (Lowe, 1982). The porphyry conglomerate and associated porphyry sheets, which overlie the tuffaceous turbiditic rocks of the Warri Warri Member in the Newdegate slice, present problems for the interpretation of their depositional environment. The high degree of rounding of the large porphyry boulders and cobbles, if ascribed to normal processes of fluviatile transport, would require a substantial distance of transport from their source. But if these clasts are far removed from their source, it is anomalous that there should have been no mixture with other rock types during travel, unless the parent rock type were to cover a very large area, which would be unusual for any hypabyssal rock.

A depositional mechanism is required that will produce a strictly monomictic boulder conglomerate, implying proximity to source, with a high degree of rounding, implying distance. Two explanations of this anomalous rounding seem possible: firstly, the boulder may have been ground during repeated back-and-forward transport, for example in an immediate offshore environment with varying current directions, or secondly, the boulders may have been rounded by grinding against each other during mass flow (autogenous grinding), rather than by normal subaerial erosional mechanisms. Using either of these mechanisms, a palaeogeographic setting can be postulated in which the rhyolite magma responsible for the rhyolite porphyry sill at a similar stratigraphic level in the Warri Warri slice breached the surface along the line of a paradepositional fault to form a line of small rhyolitic volcanoes. Either these were then contemporaneously eroded in a shallow marine or lacustrine environment, or they periodically collapsed down the face of a submarine escarpment along the line of the Antiform fault to form mass-flow boulder conglomerates. The intercalation of the porphyry boulder conglomerates with thin porphyry sills, and the rare occurrence of accretionary lapilli intercalated tuffs, would both be consistent with this model. The coarse, cross-bedded sandstone overlying the unit with abundant porphyry conglomerate probably corresponds to a shallowing of the local depository as volcanism filled it.

An interpretation of the rhyolite in the upper part of the member in the Newdegate slice as a flow is preferred on the basis that coarse spheroidal and flow-banded rhyolite is typically extrusive (Cas and Wright, 1987).

Tanguin Member

Description: The name Tanguin Member is applied here to a locally restricted sequence of sedimentary and volcanic rocks in the Tanguin slice; the outcrop of the member is restricted to an area of about 15 km^2 in the vicinity of Tanguin Hill, isolated on the west and east sides by the Baramine and Y faults respectively. These rocks appear to overlie the Koongaling Volcanic Member conformably. The lowest part of the member consists of about 200 m of dark-green, poorly stratified volcanic sandstone, interbedded with coarser quartz sandstone.

These sedimentary rocks are overlain by about 200 m of basalt, in stacked flows, which are in turn overlain by about 100 m of rhyolite. Above the rhyolite again lie basalt flows lithologically indistinguishable from those of the Kylena Formation, from whose contiguous outcrop on the west they are separated by the Baramine fault. These uppermost flows in this part of the Tanguin slice are accepted as the lowermost part of the Kylena Formation, so that the approximately 500 m thickness of the Tanguin Member lies, like the Warri Warri Member of PEARANA (Williams and Trendall, 1998c), between the top of the Koongaling Volcanic Member and the Kylena Formation. Any primary physical connection between the Warri Warri Member and Tanguin Member is speculative, and the two units have therefore been distinguished by different names. However, it is worth noting that lithic grains in some thin sections of tuffaceous beds of the Tanguin Member resemble similar rocks of the Warri Warri Member in possessing a trachytic flow texture confirming their volcanic derivation, although such rocks are not known locally in situ.

Interpretation: The basal volcanic sandstone of the Tanguin Member is interpreted as a high-density turbidite, deposited in a lacustrine or marine setting. The restriction of the member to a narrow fault-bounded outcrop suggests that its deposition may again have been related to volcanism close to a trough defined by syndepositional faults. There is insufficient outcrop area of the rhyolitic and basaltic lavas of the Tanguin Member to do other than accept the same interpretations as applied (in both cases as flows) to such rhyolites as those in the upper part of the Warri Warri Member and lower Kylena Formation, and such stacked basaltic lavas as those of the Kylena and Maddina Formations.

Warroo Hill Member

Description: The Warroo Hill Member (Williams and Trendall, 1998a) lies east of the Gregory Granitic Complex on BRAESIDE. Earlier mapping assigned the poorly exposed metasedimentary and metavolcanic rocks of this region to the Bangemall or Yeneena Groups (Hickman, 1978, 1983).

The member is at least 1 km thick and appears to conformably overlie the metamorphosed rocks in the Koongaling Volcanic Member. It is, in turn, unconformably overlain by the Neoproterozoic Tarcunyah Group. The Warroo Hill Member consists of interbedded quartzite, micaceous quartzite, metamorphosed quartz-pebble conglomerate, and siltstone. This is overlain to the east by metamorphosed basalt, amygdaloidal basalt, and dolerite interbedded with, and overlain by, pelitic schist. Some chlorite–tremolite–actinolite/talc schists in this area may be metamorphosed high-Mg basalt. There is a small area of metadolomite 2 km southeast of Warroo Hill.

In thin section the psammitic rocks are strongly recrystallized, whereas the metamorphic foliation in the pelitic schists is consistently cut by a later, spaced crenulation cleavage. The metasedimentary rocks contain a simple quartz–muscovite–biotite(–feldspar–garnet) assemblage indicative of upper greenschist – lower amphibolite facies.

Interpretation: The Warroo Member is regarded as a possible correlative of the Tanguin and Warri Warri Members.

Kylena Formation

Description: The Kylena Formation consists principally of a thick sequence of stacked mafic flows of unequivocal basaltic appearance; it is present, with generally similar characteristics, in all five named tectonic slices of the Gregory Range inlier. In the Newdegate slice, the formation is not only very much thicker than in any other slice but has an unusual intercalated wedge of felsic volcanics. In the Lochinvar slice, the Kylena Formation occurs in a number of fault-repeated sub-slices, sandwiched conformably, and directly, between the Koongaling Volcanic Member below and the Tumbiana Formation above.

The penetrative cleavage that is variably expressed within the underlying Warri Warri Member is rarely detectable within the highly competent stacked flows of the Kylena Formation.

The basalt flows form uniformly rough and dissected terrain, across which the individual flows are clearly defined on aerial photographs by thin subparallel lines marking the flow-tops. The flows, which are subaerial, are up to about 20 m thick, with an average thickness of about 15 m; they may be traced continuously for many kilometres along strike. It is uncertain whether slight local angular discordances between packages of stacked flows result from penecontemporaneous erosion or from minor strike faulting.

On the ground, the body of each successive flow consists of tough, massive, dark-green, well-jointed aphanitic lava; amygdales, generally filled by quartz, chlorite or carbonate, may be distributed throughout the flows, but are mostly concentrated in the upper parts. The uppermost parts of each flow, which range from about one to a few metres in thickness are, as well as being more strongly amygdaloidal, commonly strongly bleached and silicified, criss-crossed by quartz veins, and iregularly brecciated. In thin section, the original pyroxene and labradorite of the primary basalt are almost entirely altered to a fine-grained felted mass of amphibole, set in a matrix of albite and quartz and, in most cases, epidote.

An unusual feature of the Kylena Formation in the Newdegate slice is the presence of a wedge of felsic volcanic (including pyroclastic) rocks. These have a strike length of some 12 km, from the area east of Two Sisters (PEARANA AMG 206230) in the south to its northern termination near Mount Sydney (PEARANA AMG 146344); the sequence within the wedge dips gently (about 10°) east-northeast. The rocks include a dark aphanitic, flinty rhyolite at the base, succeeded upwards by a sequence of well-stratified green tuffaceous siltstone and sandstone, locally with abundant accretionary lapilli bands. These pyroclastic rocks are in turn overlain by rhyolite similar to that at the base, but locally associated with coarse angular felsic agglomerate. A very coarse unstratified band of mafic agglomerate caps the succession within this wedge.

This lithologically distinct wedge is represented as Tumbiana Formation on the NULLAGINE 1:250 000 sheet (Hickman, 1978), with its northern end shown, by implication, as a south-plunging syncline. The area was closely examined during remapping carried out for this Bulletin, with this stratigraphic correlation in mind. There is no doubt that the pyroclastic rocks within this wedge are closely similar lithologically to the tuffaceous rocks in the lower part of the Tumbiana Formation elsewhere on NULLAGINE. However, there is no doubt either that the sequence described within the wedge is stratigraphically coherent, as described above, and although complicated by oblique faults, represents a concordant stratigraphic wedge within a pile of stacked basaltic flows. Therefore, in spite of the lithological similarity of the pyroclastic unit to tuffs of the Tumbiana Formation, both the total lithology of the wedge, and its greatest thickness, of about 800 m, argue against such a correlation. This conclusion is in accord with two further points. Firstly, if the wedge were Tumbiana Formation, it would be impossible to reconcile the consequent identification of the overlying basaltic flows as Maddina Formation with their evident continuity with those of the Kylena Formation as they are followed northwards. Secondly, tuffs in both the Maddina Formation and Jeerinah Formation elsewhere on NULLAGINE also closely resemble those of the Tumbiana Formation, so that little weight can be placed on simple lithological similarity of such rocks.

In the Warri Warri and Newdegate slices, the base of the Kylena Formation is arbitrarily taken to be marked by the first appearance of a thick sequence of stacked mafic flows of unequivocal basaltic appearance. In the Warri Warri slice such basalts lie, with apparent conformity, either on tuffaceous sandstone or hybrid lava of the Warri Warri Member; however, this boundary is mostly concealed or complicated by a rhyolite porphyry sill emplaced close to the top of the underlying tuffaceous sandstone unit. The sill was mapped by Hickman (1978) as a felsic lava and described as a component of the Koongaling Volcanic Member of the Hardey Formation, but Hickman (1978, p. 12) also noted that 'this rock may be a high-level sill'. Although no evidence has been found within this stratiform porphyry body for intrusive or extrusive origin, both the uniformity of its lithology and its lateral extent in relation to thickness argue for its intrusive origin. The most convincing evidence for intrusion is the existence of slight stratigraphic discordance of both the upper and lower margins. The common presence of columnar jointing perpendicular to the plane of the intrusion (Trendall, 1991, fig. 6a) probably suggests a fairly shallow depth. In the Newdegate slice, the lowest stacked basalt flows overlie either rhyolite or agglomerate of the Warri Warri Member, in both cases with apparent conformity, although slight discordance is suggested by the lithological variation beneath the contact along strike.

The base of the Kylena Formation in the Tanguin slice outcrops near Tanguin Hill (PEARANA AMG 065574 to 084564). The identity of a few square kilometres of basaltic flows in the Tanguin slice north of this as Kylena Formation is accepted on the basis that they are lithologically indistinguishable from the flows of the main Kylena Formation outcrop in the Newdegate slice to the west, from which they are separated by the Baramine fault. The Kylena Formation here rests conformably over the Tanguin Member. The base is also present in the Isabella slice, in the southern Isabella Range (PEARANA AMG 014758 to 038752), where it appears to overlie the top of the Koongaling Volcanic Member of the Hardey Formation conformably. However, the actual contact is not exposed along this line, and a topographically smooth strip along the contact may conceal sedimentary rocks equivalent to the Tanguin Member.

The topmost flow of the Kylena Formation is concordantly overlain by basal Tumbiana Formation rocks throughout its outcrop within the Gregory Range inlier. There is no sign of erosional removal of any part of this flow before Tumbiana Formation deposition, nor is there any unusually deep weathering of it.

In the Warri Warri slice, between 30 and 40 flows are present in the approximately 500 m total thickness of the Kylena Formation. The total thickness of the Kylena Formation of the Newdegate slice is uncertain, because of the complexity of the fold and fault pattern of its main outcrop area within the central and northern parts of the slice. However, it is unlikely that less than 1 km of stacked basaltic flows (60–80 flows) exist above the wedge of felsic volcanic rocks near the southern edge of BRAESIDE; a further 1.8 km of stacked basaltic flows (about 120 flows) lie below the felsic volcanic wedge. If the maximum 800 m thickness of the felsic rocks is added to the 2.8 km of stacked flows, the total thickness of the Kylena Formation in the Newdegate slice is over 3.5 km. From outcrop width and average dip, the Kylena Formation of the Isabella slice is about 1 km thick, thinning to about half of this in the Lochinvar slice. Thus, some 50 stacked flows are probably present in the Isabella slice, reducing to about 25 in the Lochinvar slice.

Richards and Blockley (1984, p. 264) proposed a model Pb age for galena associated with the stacked basaltic flows of this unit of 2.76 ± 0.03 Ga. This age is based on analyses of galena from the interior of amygdales in a flow in the northern Gregory Range (PEARANA AMG 973747). They describe the sample as part of a 'pillow-like' fragment within a flow top; the flows in this locality are undoubtedly subaerial, and the rounded forms described may be related to the presence of a local shallow lake on the surface of the cooling flow. Their age is stated to support the earlier model ages of Richards (1977) obtained from galena in a quartz vein within dolerite associated with the unit east of Yownama Creek (PEARANA AMG 098453), as well as from galena in a fault cutting basalt of this unit nearby (PEARANA AMG 026655). Although these data suggest that the Kylena Formation of the Gregory Range inlier is only slightly younger than the Koongaling Volcanic Member of the Hardey Formation, the inherent interpretational uncertainties of galena model Pb ages are such that Richards and Blockley's (1984) result does not provide definitive evidence for the precise relative ages of the sampled level of this unit in the inlier, and the Kylena Formation of the main northern outcrop area of the Fortescue Group. A younger age limit for the latter is suggested by the 2715 ± 6 Ma age of detrital zircons in a sample (94775) of Tumbiana Formation (Arndt et al., 1991).

Interpretation: As in the Kylena Formation flows of the main outcrop area of the Fortescue Group (**Chapter 6**), the magmas of the stacked basaltic flows of this unit in the Gregory Range inlier appear to have been of low to moderate viscosity and to have been close to their liquidus temperature at the time of eruption. There must also have been a rather low volatile content, as there is little evidence within the flow sequences of associated explosive fragmentation, such as interstratified pyroclastic deposits. On the contrary, the repetitive similarity in thickness, lithology, and such features as vesicularity and flow-top structure in successive flows rival such classical plateau successions as those of the Deccan or Columbia River. This repetition must presumably be interpreted as the reflection of some regular periodicity in the magma generation and extrusion mechanism, but the nature of this is unknown. The absence of any lateral variation of flow thicknesses detectable on air photos indicates that the flows spread over an even subaerial plain of low gradient and negligible relief.

An entirely different interpretation must be placed on the intercalated wedge of felsic volcanic rocks in the southern part of BRAESIDE; the base of this wedge marks an important local interruption in the regular extrusion of extensive low-viscosity basaltic magma flows. The lowest unit of this wedge is a lavalike rhyolite sheet, whose interpretation as a flow is preferred purely on the basis that the units overlying it in the wedge seem to represent a local cycle of volcanic events. Thus the basal rhyolite is overlain by subaqueous tuffs whose abundant accretionary lapilli suggest that the local felsic event, which commenced with lava extrusion, continued on a decreasing scale with explosive activity. Finally, the coarse mafic agglomerate, which forms the uppermost unit of the wedge sequence, marks the beginning of a return to a period of regular mafic magma supply during the further extrusion of the stacked basaltic flows that form the upper part of the Kylena Formation.

Tumbiana Formation

Description: It has already been noted that the lithology of the Tumbiana Formation, unlike that of all units underlying it in the Fortescue Group of the Gregory Range inlier, differs so little from the main northern outcrop area of the group that its lithostratigraphic correlation across the Oakover valley (Wallal Embayment) to the west is beyond reasonable doubt. The Tumbiana Formation outcrops in four of the five tectonic slices, being absent only from the thin and restricted Tanguin slice. Notwithstanding the confidence of its identification, and the equal certainty of its correlation between the four slices in which it occurs, there are some differences between the Tumbiana Formation in each of these slices which make it convenient to describe its stratigraphy slice-by-slice; this is done below from west to east.

At the southern end of the inlier, the outcrop of the Tumbiana Formation in the Warri Warri slice stands out clearly on aerial photographs as a pale band conformably sandwiched between the top of the underlying Kylena Formation and the base of the overlying Maddina Formation. The Tumbiana Formation forms a gently north-plunging anticline, whose western limb is truncated by the Antiform fault. The unit is made up of greyish green tuffaceous siltstone and minor fine-grained sandstone; thin bands of breccia occur rarely, and are restricted to the southern section of the outcrop. Beds up to 10 cm thick of close-packed accretionary lapilli 2–4 mm across are abundant throughout the thickness of the formation. A near-vertical cleavage parallel to the axial plane of the anticline is strongly developed throughout the Tumbiana Formation outcrop in the Warri Warri slice. Although exposure over much of the narrow outcrop area is generally poor, in many places the unit forms scenically spectacular outcrops, such as around Warri Warri Creek (PEARANA AMG 205915), where the tuffaceous sediments have weathered out along the steep north–south cleavage into stacks of thin vertical plates, like giant packs of playing cards. Stratigraphic thickness in the Warri Warri slice is estimated from dip and outcrop width to be between 100 and 150 m.

From the northern end of Tumbiana Formation outcrop in the Warri Warri slice there is an exposure gap of some 35 km north-northwestwards to the southern end of the formation's outcrop in the Newdegate slice. These southernmost exposures form a gently northward-dipping cap on a conspicuous hill just north of the Port Hedland–Telfer road (PEARANA AMG 075420). Northwards from this hill its outcrop becomes increasingly discontinuous. However, the consistent stratigraphy within all these remnants indicates that their present scattered distribution is due entirely to tectonic dismemberment and displacement, and does not reflect local depositional discontinuity. The structural style of the remnants varies from systematically displaced gently dipping segments, in the area west of Moxom Well (PEARANA AMG 092460 to 025567), to tightly folded strips with local complex faulting. The dismemberment of the formation is clearly due to its relative incompetence between two thick competent mafic volcanic sequences during tectonism; both the carbonates and tuffs have acted as a lubricating layer facilitating and localizing deformation.

In contrast to the Warri Warri slice, the Tumbiana Formation of the Newdegate slice is divisible into two components: a lower tuffaceous unit and an upper carbonate unit, clearly equivalent to the Mingah Tuff Member and Meentheena Carbonate Member respectively of the formation on NULLAGINE (Hickman, 1978). As in the Warri Warri slice, both the upper and lower margins appear perfectly conformable.

The lithology of the lower tuffaceous unit is almost identical to that of the formation in the Warri Warri slice. It is made up mainly of greyish green tuffaceous siltstone and minor fine-grained tuffaceous sandstone; thin bands of agglomerate occur more rarely. Beds up to about 10 cm thick of close-packed accretionary lapilli 2–4 mm across are abundant throughout the thickness of the unit. In a locality on BRAESIDE, the lapilli of one such bed were seen to fill a network of spaced cracks, wedging out downwards from a width of a few centimetres over a depth of 10–15 cm, in the underlying siltstone. These crack-fillings represent a multiple example of the single lapilli dyke that Boulter (1985) has described from the Tumbiana Formation of ROY HILL. The upper carbonate unit consists of tough, dark-brown-weathering dolomite of varying lithology. This unit includes finely to coarsely stratified grey dolomite, in which the bedding is defined by these siliceous laminae which may be either continuous or discontinuous, and either slightly rippled or flat. Such siliceous laminae, together with slight colour variations of the dolomite, also locally define small branching stromatolites, as well as domal stromatolites up to about 50 cm in diameter and 20 cm in height. Tepee structures and breccia bands are also present, but many thicker (30–50 cm) beds within the dolomite appear massive and structureless. In many exposures, a terraced effect is present which reflects the occurrence, at intervals of a few metres, of minor bands of greenish shaly dolomite. Ripple marks are common on well-exposed bedding planes, although no consistent orientation of these was observed, and in some exposures the ripple direction varies substantially over small stratigraphic intervals.

The formation is about 90 m thick at the southernmost end of its outcrop, in the hilltop section close to the Port Hedland – Telfer road; of this thickness, the lower unit is about 60 m thick and the upper unit about 30 m. Although complete measurable sections become steadily rarer northwards, it appears to be the case that the lower tuffaceous unit steadily thins in that direction. The upper carbonate unit thickens proportionately to occupy most of the total thickness, which remains at about 90 m, towards the northern margin of the sheet.

In the Isabella slice, the Tumbiana Formation is well exposed in a single northwest-trending belt about 7 km long, in which the formation dips gently to the northeast, and lies with perfect conformity between the underlying Kylena Formation and overlying Maddina Formation. A well-exposed section was measured near Lightning Ridge (PEARANA AMG 950610), in which the carbonate facies is predominant. The total thickness of 103 m consists of a number of steps formed by bands of well-stratified pale to dark-grey, buff-weathered carbonate alternating with thinner and darker shaly carbonates that form flatter terraces between the steps. The shaly carbonate is thinly and evenly bedded, but the main carbonate type is stratified on a scale of 10–20 cm. Such beds may lack internal stratification, but in others thin siliceous laminae commonly define small domal stromatolites, tepee structures, and brecciated bands. Local massive chert bands are also present. The terraced effect at this section, associated with the alternation of two carbonate types, appears consistently throughout the strike length of the formation in the Isabella slice.

In the Lochinvar slice, the formation outcrops in a number of separate, poorly exposed strips parallel to, and partly bounded by, the north-northwest faults which repeat the lower part of the Fortescue Group sequence within thin sub-slices. The Tumbiana Formation of the Lochinvar slice is less well exposed, and its tectonic dismemberment precludes any attempt to determine precise thickness. In some places in the northern Gregory Range (e.g. ISABELLA AMG 950870) exposure is sufficiently good to indicate that its lithology is identical to that of the Isabella slice, but in many of the outcrop strips farther east only general recognition of the rocks as carbonate is possible.

Interpretation: The well-stratified tuffaceous component of the Tumbiana Formation is interpreted as a primary pyroclastic fall deposit accumulated in a shallow-water environment, either lacustrine or marine. Local cross-bedding and redistribution of broken accretionary lapilli are evidence of some reworking, but the thicker beds of close-packed lapilli are almost certainly direct air-fall deposits. Moreover, the incidence of observed infilled cracks in the underlying beds suggests that the water body in which they accumulated was sufficiently shallow to permit local or periodic desiccation.

The depositional environment of the upper carbonate component is interpreted, like the stromatolitic carbonate of the main outcrop area (**Chapter 7**), as a low-energy shallow coastal setting.

Maddina Formation

Description: The Maddina Formation, like the Kylena Formation, consists predominantly of a sequence of stacked mafic lava flows, and like the underlying Tumbiana Formation is present in four of the five tectonic slices; it is absent only from the thin and restricted Tanguin slice. The penetrative axial cleavage that is locally conspicuous in the underlying Tumbiana Formation of the Gregory Range inlier is rarely detectable within the highly competent massive flows of the Maddina Formation. In areas of structural complexity, the Maddina Formation and Kylena Formation can be confidently identified only if their stratigraphic relationships with the Tumbiana Formation can be established — respectively conformably above or conformably below.

The stacked flows of the Maddina Formation have an unequivocal basaltic appearance. They form uniformly rough and dissected terrain across which the individual flows are clearly defined on aerial photographs by thin subparallel lines marking the flow tops. The flows, which are subaerial, are up to about 50 m thick, with an average thickness of about 20 m. They may be traced continuously for many kilometres along strike, except in the southern part of the Gregory Range inlier, where the stratigraphy is compressed and complicated by minor strike faulting. The body of each successive flow consists of tough, massive, dark-green, well-jointed aphanitic lava; amygdales, mainly filled by quartz, chlorite or carbonate, may be distributed throughout the flows, but are mostly concentrated in the upper parts. The uppermost parts of each flow, from about a metre to a few metres in thickness, as well as being more strongly amygdaloidal, are often strongly bleached and silicified, quartz veined, and irregularly brecciated. In thin section, the original pyroxene and labradorite of the primary basalt are almost entirely altered to a fine-grained felted mass of amphibole, set in a matrix of albite and quartz, and commonly epidote.

Apart from their generally greater average thickness, the flows of the Maddina Formation are not distinguishable from those of the Kylena Formation in the field, nor, so far as is known at present, can they be distinguished by other means. An exception to this is a distinctive porphyritic lava commonly present in the upper part of the formation in the southern part of its exposure, in the Warri Warri slice. This basaltic rock has pale-green euhedral feldspar phenocrysts up to 15 mm long, commonly in porphyritic clusters. It is unclear whether the restricted distribution of this feldspar-phyric lava type is due to its removal from a wider area by erosion prior to the deposition of the overlying Jeerinah Formation, or whether its primary extrusion was restricted to the present area of occurrence.

In the northern part of the Newdegate slice, a unit of stratified tuffaceous sediments 60–70 m thick occurs in the upper part of the formation, and stands out on airphotos as a pale band within the dark basalt flows. It is well exposed near Parsons Creek (PEARANA AMG 925760 to 940762) sandwiched between conformable lava flows below and above. The rocks consist of greyish green, fine-grained, commonly flinty-looking, structureless or finely stratified felsic tuffaceous siltstone and rarer sandstone. Accretionary lapilli between 2 and 4 mm in diameter occur both as scattered lapilli and in concentrated bands up to 5 cm thick. This stratified tuffaceous unit is also present within the Maddina Formation of the Isabella slice, and may be correlated either with the Kuruna Siltstone (now Kuruna Member) of NULLAGINE (Hickman, 1978) or with one of the two other lithologically similar units that are present within the Maddina Formation of ROY HILL, in addition to the Kuruna Member equivalent there (Thorne and Tyler, 1997b).

The Maddina Formation of the Warri Warri slice is estimated from average dip and outcrop width to be 1 km, and is therefore made up of about 50 stacked flows. Although pervasive faulting within the Newdegate slice renders thickness estimation more difficult, an estimate from average dip and outcrop width of about 800 m is reasonable, so that here the formation may be made up of between 30 and 40 stacked flows. The total thickness in the Isabella slice can be estimated from average dip and outcrop width as close to 700 m, so that the lava component below and above the tuff band probably contains some 30 stacked flows. No thickness estimate is possible in the Lochinvar slice. The thickness of the Maddina Formation thus varies little between tectonic slices, but there appears to be a slight thinning from south to north within the inlier.

Interpretation: Interpretation of the basaltic flows of the Maddina Formation, like those of the Kylena Formation, follows that already made for the same unit in the main outcrop area of the Fortescue Group (**Chapter 8**). The magmas of the stacked basaltic flows of this unit must have been of low to moderate viscosity and close to their liquidus temperature at the time of eruption. There must also have been a rather low volatile content, as there is little evidence within the flow sequences of associated explosive fragmentation, such as interstratified pyroclastic deposits. Again, in this unit the repetitive similarity in thickness, lithology, and such features as vesicularity and flow-top structure in successive flows rival such classical plateau successions as those of the Deccan or Columbia River; this must also be interpreted as the reflection of some regular periodicity of unknown origin in the magma generation and extrusion mechanism. The absence of any lateral variation of flow thicknesses detectable on airphotos indicates that the flows spread over an even subaerial plain of low gradient and negligible relief.

The intercalated tuffaceous shale unit within the Maddina Formation represents a recurrence of the depositional environment of the tuffaceous component of the Tumbiana Formation.

Jeerinah Formation

Description: The Jeerinah Formation outcrops over the entire north–south length of the Gregory Range inlier in the Warri Warri, Newdegate, and Isabella slices. Although overall there is more evidence of a tuffaceous content for the sedimentary rocks of the formation than in the main outcrop area of the Fortescue Group to the west, there is also a strongly marked tendency for this volcanogenic content to increase from south to north, to the extent that two volcanic members are now recognized within the formation at the northern end of the inlier. Both of these, the mainly mafic volcaniclastic Baramine Volcanic Member (as redefined by Williams and Trendall, 1998b), and the stratigraphically higher and silicic Isabella Member, are components of the originally defined Baramine Volcanic Member of Hickman et al. (1983).

The undifferentiated sedimentary rocks of the Jeerinah Formation include white, buff, or pale-purple thinly bedded shales and siltstones, and medium-grained, more coarsely stratified sandstones. Exposures of the finer grained facies, which occurs throughout the length of the inlier, tend to be of poor quality. A characteristic outcrop feature is a scattering of chert cobbles derived from interbedded bands of coarsely laminated pale chert up to about 10 cm thick within the epiclastic sediments; such bands tend to be irregularly spaced rather than concentrated. Bands of accretionary lapilli tuff are common in the lower part of the formation, and beds up to 30 cm thick with close-packed lapilli were noted locally.

The more coarsely epiclastic rocks are restricted to the northern part of the inlier, and include sandstone with individual beds up to 30 cm thick, although mostly much less.

Dolerite sills commonly intrude the sedimentary rocks of the Jeerinah Formation, and vary widely in both local abundance and thickness.

One of the few well exposed complete sections of the Jeerinah Formation is found in the northern part of the inlier (ISABELLA AMG 922880 to 935880), an area which also includes the type section of the Baramine Volcanic Member and the Isabella Member. Throughout the inlier, the lowest sedimentary rocks of the Jeerinah Formation appear to overlie the uppermost lavas of the Maddina Formation conformably, with the possible exception, already noted, of the feldspar-phyric facies of the Maddina Formation in the southernmost part of the inlier.

There is a steady increase in thickness of the Jeerinah Formation from about 200–250 m in the Warri Warri slice, at the southern end, to a thickness of at least 1 km in the Newdegate slice of the central part of the inlier. The thickness of 700 m in the Isabella slice (Fig. 10.2) indicates that the greatest thickness is in the central area. Although much of this thickness increase is directly attributable to the greater volcanic content, there is also a thicker sedimentary component which is not demonstrably volcanogenic.

Interpretation: Interpretation of the non-volcanic sedimentary rocks of the Jeerinah Formation follows that already described for the main outcrop area (**Chapter 9**). The fine grain size, parallel lamination, and lack of scouring indicate that the shales and siltstones were laid down from suspension in a low-energy environment. A marine environment deeper than 200 m is indicated by the very large area over which such rocks were deposited without evident wave disturbance. A variable component of fine-grained tuffaceous material is supported by the local intercalation of accretionary lapilli beds. Inter-calated laminated cherts are interpreted as chemical precipitates that accumulated during periods of low sediment supply. In the northern part of the inlier an increasing component of coarser grained and locally cross-stratified sandstone, especially in the upper part of the formation, indicates deposition in shallower water.

Baramine Volcanic Member

Description: Mapping at 1:100 000 scale by Williams and Trendall (1998b) enabled Hickman et al.'s (1983) Baramine Volcanic Member to be subdivided at the northern end of the Gregory Range inlier. To avoid proliferation of new stratigraphic names, that member was redefined by Williams and Trendall to apply only to a discrete mafic pyroclastic unit within the originally defined member, with a distinctive dark field appearance and a characteristic pyroclastic lithology.

Although neither the lower nor the upper contact of the Baramine Volcanic Member is clearly exposed on ISABELLA, the unit appears to be conformable within the enclosing undifferentiated sedimentary rocks of the Jeerinah Formation. In the type section the redefined member has a thickness of about 175 m. This is consistent with Hickman et al.'s (1983) specification of the the thickness of the Baramine Volcanic Member as 250 m, since Hickman et al.'s unit included the Isabella Member, as defined here, as well as sedimentary rocks between the Isabella and Baramine Volcanic Members.

The rocks of the member characteristically form almost karst-like exposures, due to the significant carbonate content and preferential dissection along multiple joint sets. Their usual dark-brown to black colour makes them conspicuous both on the ground and on airphotos. The unit consists of mafic volcaniclastic rocks with a distinctive style of stratification. Individual beds with internally consistent lithology are mostly between one and five metres thick, although units up to twenty metres thick are locally present. All of the thicker, and many of the thinner beds are made up of a chaotic, unstratified, aggregation of mafic lava clasts, both angular and rounded, mainly between 5 and 20 cm across, although large (50–100 cm) boulders are not uncommon. The material of the clasts varies from 'spongy' and vesiculated to massive fine-grained basaltic rock. The rocks are everywhere heavily carbonated, and in many exposures there is a peripheral concentration of carbonate around each clast, so that each has a recessed rim on weathered faces. The matrix of the clasts consists of progressively finer grained mafic debris. Although there is some tendency for thicker beds to contain larger clasts, there are many exceptions to this relationship. The size distribution within individual beds varies wildly, so that thin beds with mainly quite fine material may have much larger angular blocks widely distributed within them. Rare clasts of bedded dolomite, pale sandstone, and bedded chert are present, mostly comparable with rocks present in the sedimentary rocks of the underlying part of the Jeerinah Formation. Some thinner beds made up of finer (<1 cm) mafic debris possess an internal stratification, which tends to be on a scale of a few centimetres, and to show little consistent pattern of grain-size variation from bed to bed. Both normal and reverse grading of grain size are locally present but do not show the well-defined patterns of epiclastic water-lain turbidites; cross-bedding is also present in some exposures.

The distinctive dark-weathering carbonate-rich pyroclastic rocks present in the type area and elsewhere on ISABELLA (Williams and Trendall, 1998b) extend for about 20 km southwest on BRAESIDE and YILGALONG, and thin steadily in that direction. Williams and Trendall (1998b) also noted an anomalous occurrence of mixed volcanic rocks in the Isabella Range (ISABELLA AMG 922898 to 994532), which have a different aspect to the Baramine Volcanic Member of the type section. The rocks include thick basaltic flows, associated dolerite sills, mafic pyroclastic bands, and a variety of thin felsic flows and associated tuffs, including thick bands with close-packed accretionary lapilli. Although Williams and Trendall (1998b) mapped these rocks as a component of the Baramine Volcanic Member they stressed that their mapping was insufficiently detailed to establish whether the correlation was valid. The total thickness of these anomalous rocks was estimated as about 1 km.

Interpretation: Although insufficient detailed work has been done on the Baramine Volcanic Member, as typified by the pyroclastic rocks of the type section, for a definitive palaeovolcanological interpretation it is possible to make some first-order assessments of their depositional environment. The presence of units up to 20 m thick with clasts up to 50 cm in diameter in the northern parts of the outcrop, on Isabella, suggests that these beds may be direct fallout deposits no more than a few kilometres from the eruptive source (Fisher and Schmincke, 1984). However, the virtual absence of internal stratification within them makes it more likely that they are volcaniclastic gravity-flow deposits, but whether they were synchronous with eruption, or formed later as secondary lahars associated with collapse of more proximal volcanic products, is uncertain. Their intercalation with finer grained well-stratified tuffs also favours their identification as transported mass flows. The stratification of the finer tuffs closely resembles, particularly in grain-size distribution, many of the base-surge deposits illustrated by Cas and Wright (1987), but no accretionary lapilli have been seen in this facies. It is therefore likely that all the volcaniclastic rocks of the member are subaqueous, though from a subaerial eruption, as subaqueously erupted mafic magma might be expected to have produced some intercalated pillowed flows. A shallow subaqueous environment close to an explosive subaerial, possibly island, source would be consistent with the concordant occurrence of the Baramine Volcanic Member within subaqueous sedimentary rocks of the Jeerinah Formation.

The consistent and distinctive association of pyroclastic lithologies of the Baramine Volcanic Member suggests that it probably formed rapidly during a relatively brief episode of semi-continuous eruption, and that the rocks represent direct products of the volcanism. This is in contrast, for example, to the reworked turbiditic tuffaceous sandstones of the Warri Warri Member of the Hardey Formation, which result from secondary resedimentation of volcanic material, and in whose stratification no direct record of volcanic processes is preserved.

Isabella Member

Description: The Isabella Member of the Jeerinah Formation (Williams and Trendall, 1998b) forms a prominent southwest-facing scarp, whose highest peak (Isabella AMG 948882) lies in the Isabella slice, about 5 km southeast of the abandoned Baramine Homestead. The member is also well exposed in the hilly country extending directly south of Baramine, where it is complexly folded adjacent to the Baramine fault, as well as in the Newdegate slice on the eastern side of the northwesterly valley centred about 8 km south-southwest of the site of the abandoned homestead. In the latter location it also forms a conspicuous southwest-facing scarp.

The Isabella Member consists of about 50 m of even-grained massive rhyolite, which is grey when fresh and reddish-brown on weathered faces; no flow-banding or other structure suggestive of intrusion is present. Small (2–4 mm) spherical carbonate-filled amygdales are commonly abundant in the rock.

The rhyolite of the Isabella Member is both overlain and underlain by conformable fine-grained epiclastic rocks. The fact that the member wedges out within these rocks a few kilometres north of the type section suggests that it may be a sill, but no definitive evidence for or against intrusion could be found at either margin.

Interpretation: The Isabella Member is a lavalike felsic sheet which may be either a sill or a flow; there is insufficient evidence at present to prefer either alternative.

Chapter 11

Intrusive igneous rocks

Sills

Sills are common throughout the Hamersley Basin but are most prevalent in the south Pilbara and northwest Pilbara sub-basins (Plates 1a, 1b). Most are mafic, although layered ultramafic–mafic and granophyric bodies are also present locally. Only two sills, the Gidley Granophyre (de Laeter and Trendall (1971) and the Cooya Pooya Dolerite (Kriewaldt and Ryan, 1967) have been formally named, and both lie in the northwest Pilbara sub-basin.

Gidley Granophyre

The Gidley Granophyre is a thick sheet of granophyre and associated quartz gabbro intruded along the basal unconformity of the Fortescue Group in the Dampier Archipelago. It was first mapped by Kriewaldt (1964a), who referred to it as 'granophyre in the Dampier Archipelago'. Further description, using the informal name 'Dampier Granophyre', was given by Trendall (1963) before the unit was formally named by de Laeter and Trendall (1971). Subsequent mapping of the Gidley Granophyre was carried out by Biggs (1979, 1980), and Wallace and Hoatson (1990) summarized aspects of its petrology and geochemistry.

Wingate (1997) obtained a U–Pb baddeleyite age of 2725 ± 1.3 Ma from the basal gabbro of the Gidley Granophyre. Previous efforts to obtain a U–Pb zircon age from the granophyre had been unsuccessful owing to the absence of suitable zircon crystals in the samples collected. The only other geochronological data came from 14 samples of granophyre, which yielded two Rb–Sr isochrons: a younger one of 2196 ± 26 Ma and an older one of 2612 ± 56 Ma (de Laeter and Trendall, 1971). Comparison with the U–Pb baddeleyite age of Wingate (1997) and other Rb–Sr data from the Fortescue Group suggests that these dates may reflect post-emplacement modification of the isotope systematics rather than the age of intrusion (cf. Nelson, D. R. et al., 1992).

The Gidley Granophyre has an estimated thickness of about 870 m (Trendall, 1990b) and its field relationships were summarized by Kriewaldt (1964a, p. 9) as being 'intruded at the unconformity between Archaean granitic rocks' (the Dampier Granitoid Complex) 'and the overlying gently dipping basaltic lavas of the Proterozoic' (Mount Roe Basalt). 'At the top of the granophyre, metasomatized by it, are small lenses of altered arkose, basal to the lava. Small dykes, apparently offshoots of the granophyre, intrude the overlying basalt on Enderby Island'. Locally, felsite dykes also intrude the granophyre (Wallace and Hoatson, 1990). The assimilation of country rock close to the lower and upper contacts of the intrusion has produced a variety of hybrid rocks, commonly ranging from gabbro to tonalite.

In hand specimen the granophyre is normally a massive, homogeneous to mottled green, purple, blue, or black rock. Its petrography has been described by Trendall (1963), de Laeter and Trendall (1971), and Wallace and Hoatson (1990). Most granophyres are medium-grained rocks consisting predominantly of complexly intergrown strained quartz and K-feldspar. Common boundaries between these two minerals are diffuse owing to the development of fine, radiating micrographic intergrowths, which give the rock its distinctive granophyric texture. Scattered accessory clinopyroxene is commonly zoned, ophitic with respect to K-feldspar, and subophitically encloses rare sericitized albite (Wallace and Hoatson, 1990).

The basal gabbro ranges from microcrystalline to very coarsely crystalline; Biggs (1980) recorded crystals of hornblende up to 100 mm long. De Laeter and Trendall (1971) described a gabbro sample as comprising groups of subhedral, partially altered clinopyroxene, set in a matrix consisting mainly of an aggregate of saussuritized, subhedral andesine–labradorite. Some orthopyroxene is also present. The interstices between the plagioclase laths are occupied by pyroxene, quartz, or a micrographic intergrowth of quartz and albite. Irregular areas of chlorite are probably derived from alteration of pyroxene, and a few flakes of biotite are also present. De Laeter and Trendall (1971) noted that other gabbros have a noritic mineralogy of equigranular hypersthene, augite, and labradorite, although these also have interstitial pockets of quartz or micrographic quartz–albite intergrowths.

Wallace and Hoatson (1990) also reported analyses of three gabbro samples from this intrusion. In terms of the scheme used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group, these form, the GGKA (Karratha) 4-letter group (Tables 12.1, 12.2). These samples show no clear chemical affinity with either the characteristic basaltic andesite composition of the stacked mafic flows throughout the Fortescue Group, the high-Mg komatiitic rocks of the Cooya Pooya Dolerite and Pyradie Formation, or the basaltic sills of the Jeerinah Formation. More work is needed to establish, physically and chemically, their relationship to the main felsic mass of the overlying granophyre.

Cooya Pooya Dolerite

The Cooya Pooya Dolerite (Kriewaldt and Ryan, 1967) is a large sill complex which outcrops over a strike length of 85 km in the northwest Pilbara sub-basin. The dolerite comprises two main bodies (Blake, 1993): a lower sill, which is about 120 m thick and intrudes the middle part of the Hardey Formation, and an upper sill, which is about 40 m thick and intrudes close to and along the Hardey Formation – Kylena Formation boundary. The Cooya Pooya Dolerite is therefore restricted to the Lyre Creek Member of the Hardey Formation.

Swindells (1986) noted that the basal contact of the lower sill is transitional with the Lyre Creek Member, the dolerite becoming increasingly fine grained until it is difficult to distinguish it from hornfelsed country rock. Small-scale feeder dykes have been recognized in this part of the succession at Lockyer Gap (Swindells, 1986). Descriptions of the upper contact of the dolerite suggest that it too has metamorphosed the adjacent country rock. However, there is some doubt as to whether this material overlies the sill, or is a rafted body as suggested by Swindells (1986).

Wallace and Hoatson (1990) described samples taken from the lower sill of the Cooya Pooya complex at Lockyer Gap. They noted that lower levels of this sill comprise quench-textured clinopyroxene rocks. Clinopyroxene is present as uralitized skeletal rods and unaltered platy phenocrysts poikilitically enclosing numerous small talc–chlorite pseudomorphs after orthopyroxene and olivine.

Wallace and Hoatson (1990) reported analyses of 11 samples of Cooya Pooya Dolerite. In terms of the scheme used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group, these form the CPLG (Lockyer Gap) 4-letter group (Tables 12.1, 12.2). These samples contribute to the stratigraphically median high-MgO 'spike' apparent in Figure 12.2 and already noted in **Chapter 7**. The samples plot in a field stretching across the komatiite field of Grunsky et al. (1992) on the Jensen (1976) diagram into the high-Mg tholeiitic basalt field (Fig. 12.7). The highest MgO content of the reported analyses is 20.52%, and five others of the group are above the 10.2% MgO limit for classification as komatiite (Grunsky et al., 1992). This compositional range precisely mirrors, within the limits of the available data, that of the subaqueous basaltic and komatiitic flows of the Pyradie Formation, and suggests the possibility that this sill, or sills, is part of a feeder complex for those stratigraphically overlying extrusive rocks.

Other mafic sills

Mafic sills are interlayered with most Fortescue Group units, but are most abundant within the Jeerinah Formation, where they can make up over 60% of the exposed section. There are marked local variations in their abundance. Sills are particularly abundant in the south Pilbara sub-basin, where most lie south of a line extending from the northern Jeerinah Anticline to the northwestern Sylvania Inlier. In this latter area Tyler (1991) noted that mafic sills in the Jeerinah Formation are confined to the area west of the Fortescue River Fault. Sills are also locally numerous in the Gregory Range inlier, where their abundance increases from south to north; they are rare or absent on PEARANA, but abundant and locally thick on BRAESIDE.

Mafic sills range in thickness from a few metres to about 200 m. Most are impersistent laterally, although the thicker bodies may be traced along strike for about 60 km. Many sills are transgressive, whereas others contain lenticular bodies of country rock that range in size from a few metres to several hundred metres across.

In common with other volcanic rocks in the Fortescue Group, sill mineralogy has been modified by low-grade burial metamorphism (cf. Smith et al., 1982). Relic textures are commonly well preserved, however, particularly in the coarse-grained rocks. Most sills consist of fine- to coarse-grained dolerite or gabbro that preserve ophitic or subophitic igneous textures (Tyler, 1991; Thorne et al., 1991). Clinopyroxene may be preserved as plates and blades up to 12 mm across, which show alteration to chlorite and/or actinolite. Plagioclase laths are invariably albitized and locally replaced by intergrowths of sericite and epidote/clinozoisite. Iron oxides are leucoxenized and pumpellyite is present in some samples.

Twenty-six analyses are available of dolerite sills from the Jeerinah Formation, from the south Pilbara sub-basin and the Gregory Range inlier. In terms of the scheme used in **Chapter 12** to subdivide the large volume of analytical data available for mafic rocks of the Fortescue Group, they include one analysis from the JECA (Cowley area) 4-letter group, four analyses of the JEBD (Bellary Dome) group, five analyses of the JETP (Tom Price) group, nine analyses of the JENE (Newman) group, and seven analyses of the JEGR (Gregory Range) group; all 26 analyses from the five geographically separate groups form the K cluster (Figure 12.2; Tables 12.1, 12.2).

The mean of these relatively tightly clustered analyses differs in a number of ways from the generally basaltic andesite composition typical of all the underlying stacked mafic lavas of the Fortescue Group except those of the Pyradie Formation. Importantly, the average silica content of 50.7% for cluster K is substantially lower than the 57.5% of the underlying stacked flows (Fig. 12.2). The total alkali content is also lower, so that in terms of the classification scheme of Le Maitre (1989), the magma supplying these sills was truly basaltic. Complementary to this, their Al_2O_3, MgO and CaO contents are above those of the stratigraphically lower basaltic andesites. On a Jensen cation plot, analyses of the K cluster form a tight group on the boundary between the high-Fe and high-Mg tholeiitic basalt fields. And finally, their incompatible trace element level is substantially lower than that of the basaltic andesites.

Layered sills

Layered mafic or ultramafic–mafic sills occur interbedded with the Fortescue Group, particularly the Hardey and Pyradie Formations, in the south Pilbara sub-basin. They are most abundant around the Bellary, Rocklea, and Milli Milli Domes and southwestern Sylvania Inlier (Plates 1a,1b) and are not recorded north of a line extending from the northern Jeerinah Anticline to the Western Creek Fault in the west-central Sylvania Inlier.

Most sills range in thickness from 30 to 350 m and the thicker bodies may be traced along strike for about 20 km. Some are markedly transgressive and many contain rafts and screens of altered country rock. Cowley (1979) records one sill in the Hardey Formation of the northern Rocklea Dome that merges laterally into a dyke that cuts the overlying Boongal Formation.

The majority of sills appear to consist of a single differentiated unit ranging in composition from pyroxenite at the base to mesocratic or leucocratic dolerite or gabbro at the top. In addition, several intrusions contain a basal serpentinite layer, commonly less than 25 m thick, below the pyroxenite division. In many sills the boundary between the various major compositional layers, although transitional, is reasonably sharp. Locally, the internal structure of the sill appears to reflect two periods of magma injection, with a lower pyroxenite–dolerite division succeeded by an upper unit with a similar differentiated structure.

Sill mineralogy has been modified by low-grade burial metamorphism (Smith et al., 1982), although relic textures are commonly well preserved. The lower pyroxenite division consists of a medium- to very coarse grained augite and/or orthopyroxene cumulate (typically an orthocumulate or mesocumulate). Subhedral to euhedral augite and orthopyroxene is partially or completely replaced by pseudomorphic and epitaxial actinolite, with or without chlorite, and carbonate. Interstitial material comprises various combinations of anhedral to subhedral actinolite, chlorite, epidote–clinozoisite, leucoxene, saussuritized or albitized plagioclase, serpentine, quartz, sericite, sphene, and talc. The upper dolerite to gabbro division comprises ophitic to subophitic relic pyroxene, partially or completely replaced by actinolite and/or chlorite and carbonate; and former calcic plagioclase, mostly altered to albite, epidote–clinozoisite, and sericite. Interstitial material comprises various combinations of anhedral to subhedral actinolite, biotite, carbonate, chlorite, epidote–clinozoisite, leucoxene, saussuritized or albitized plagioclase, quartz, sericite, and sphene.

The similar geographic distribution and mineralogy of layered sills in the Hardey and Pyradie Formations, and pyroxene spinifex-textured basalt and komatiitic flows in the Pyradie Formation, suggest that the sills may represent subvolcanic equivalents of the extrusive flow units (Blight, 1985; Meakins, 1990). Two analyses of the ultramafic parts of such sills within the Hardey Formation in the area of the south Pilbara sub-basin mapped by Cowley (1979; see Figure 12.1a) are highly magnesian, with MgO 29.71% (SiO_2 39.25%) and 31.34% (SiO_2 39.12%). However, the lowermost pyroxenite component of a layered sill within the Pyradie Formation of the same area had MgO 13.32% (SiO_2 51.00%). Unfortunately, no systematic vertically spaced analyses through any layered sill are available, so that the bulk composition of the original magma cannot be shown to be that of the Pyradie Formation flows.

Layered pyroxenite–dolerite sills are also present locally in the Boongal Formation; for example, in the southwestern Milli Milli Dome. It is uncertain whether these represent an earlier pulse of high-magnesian magmatism, or whether they are intrusions linked to the main Pyradie Formation activity.

Dykes

Mafic dykes

Based on the classification of Tyler (1990b) three groups of dykes are recognized on the Pilbara Craton:

- An older group of foliated dykes that pre-dates deformation of the greenstone belts.
- A predominantly north-northeasterly trending suite, present on the northern Pilbara Craton, which post-dates deformation of the greenstone belts but pre-dates most of the Fortescue Group. This is the Black Range Suite of Hickman and Lipple (1978) and has been correlated with a similar group of northerly to north-northeasterly trending dykes in the Sylvania Inlier (Tyler, 1990b, 1991). However, these dykes have been shown to be about 25 m.y. younger than the Black Range Suite (Wingate, 1999).
- A younger group of west-southwesterly to west-northwesterly trending, and north-northeasterly trending dykes that developed along the southern and western margin of the craton during and after the Capricorn Orogeny.

In the northeast Pilbara, a number of dykes belonging to the Black Range Suite are intimately associated with lower Fortescue Group rocks. Wingate (1994, 1997, 1999) obtained an age of 2772 ± 2 Ma from this suite using SHRIMP U–Pb baddeleyite geochronology. This age supports the views of earlier workers who suggested that these dykes may have been part of a feeder system for the the Mount Roe Basalt (Lewis et al., 1975; Hickman and Lipple, 1975; Hickman, 1983; Blake, 1993).

Dykes of the Black Range Suite outcrop over a distance of about 150 km on MARBLE BAR and NULLAGINE. The longest single intrusion is the Black Range dyke which is over 100 km long and up to 150 m wide, although the shorter Cajuput dyke is almost 1 km wide in places (Lewis et al., 1975; Hickman, 1983).

Field relationships between Fortescue Group rocks and dykes belonging to the Black Range Suite are generally ambiguous because they are only rarely seen in contact with one another. Blake (1993) noted that in the Marble Bar sub-basin, dykes intrude the lower sedimentary unit, Mount Roe Basalt and lower part of the Hardey Formation, and that conglomerates in the upper Hardey

Formation contain dolerite clasts that are petrographically similar to the Black Range Suite. In the east Pilbara sub-basin the relationship of dykes to the Mount Roe Basalt is unclear. Hickman (1983) comments that the Cajuput dyke penetrates the Mount Roe Basalt 38 km north-northeast of Nullagine but is uncertain whether this is an intrusive relationship or whether the lavas have draped around the dyke. Ten kilometres south of Nullagine, the Cajuput dyke is overlain by the upper part of the Hardey Formation. About 28 km north-northwest of Nullagine, another similarly oriented dyke is overlain by the Mount Roe Basalt (Blake, 1993). On ROY HILL, the north-northeasterly trending dyke swarm is overlain by the Kylena Formation and Tumbiana Formation. In the Sylvania Inlier, the Mount Roe Basalt is absent and the relationship of the dykes to the locally preserved Hardey Formation is unknown (Tyler, 1991).

In summary, some of the Black Range Suite apparently pre-dates parts of the Mount Roe Basalt, although in other cases the field relationships are unclear. In the Marble Bar sub-basin, dykes of this trend cross-cut Fortescue Group units as high as the lower Hardey Formation. The younger age limit for these intrusives has not been clearly demonstrated, however, since dolerite clasts in upper Hardey Formation conglomerates could have been derived from part of the Black Range Suite that pre-dates the Fortescue Group. Elsewhere in the northern Pilbara, all north-northeasterly trending dykes pre-date the Kylena and Tumbiana Formations.

North-northeasterly trending dolerite dykes which intrude granite–greenstone rocks of the Sylvania Inlier are not seen in contact with the Fortescue Group, but have a probable age of 2747 ± 4 Ma, about 25 m.y. younger than the Black Range Suite (Wingate, 1999). These dykes may have acted as feeders for basaltic rocks in the Kylena and Boongal Formations (Wingate, 1999). An alternative interpretation is that they were associated with basaltic volcanism in the upper part of the Hardey Formation.

Felsic dykes

In the northeast Pilbara sub-basin, Fortescue Group rocks are intruded by two sets of felsic dykes. The older set comprises a northwesterly trending suite of rhyolitic dykes that cut the Mount Roe Basalt and lower part of the Hardey Formation in the Meentheena area (Hickman, 1974). Uranium–lead zircon data from a sample of dyke material gave an age of 2765 ± 2 Ma (Thorpe, R., 1992, pers. comm.). This date is within error of U–Pb zircon results obtained from the Mount Roe Basalt and Bamboo Creek Member of the Hardey Formation (Pidgeon, 1984; Arndt et al., 1991). The fact that the dykes cut lower Hardey Formation rocks and are of rhyolitic composition suggests they are more likely to be associated with the Bamboo Creek Member than with the Mount Roe Basalt.

Williams (1999) described a small, composite pluton of medium-grained hornblende granodiorite and hornblende monzogranite on the southwestern part of the MUCCAN 1:100 000 map sheet in the northeast Pilbara sub-basin. The hornblende granodiorite has yielded a SHRIMP U–Pb zircon age of 2757 ± 7 Ma, which is within error of the age of felsic igneous rocks in the Bamboo Member of the Hardey Formation.

Chapter 12

Geochemistry

In this chapter we evaluate the significance of the 448 chemical analyses of Fortescue Group rocks available to us. This total is made up of 248 analyses already published, 103 analyses of rocks selected by us during the course of this study, and 97 analyses in unpublished theses. As noted in **Chapter 1**, some 350 additional whole-rock analyses, briefly discussed by Meakins (1990), remain unavailable, and 99 analyses from the Kylena and Maddina Formations of the northwest Pilbara sub-basin were reported (Kojan and Hickman, 1998) after this Bulletin was compiled. All the Fortescue Group whole-rock analyses, excluding isotopic data, used during this study will be made available from GSWA in delimited ASCII file format.

The great bulk (376) of these 448 whole-rock analyses are of mafic to ultramafic rocks, to which either the name 'basalt', if extrusive, or 'dolerite', if intrusive, has been historically applied. These 376 analyses are mainly of mafic rocks, but they also include a small number for which, by virtue of various conventional criteria such as low silica or high MgO content, the label ultramafic would be more appropriate, and we have therefore used the term 'mafic and ultramafic' throughout. Although, as we show later, the bulk of the lavas analysed would more correctly be described as basaltic andesites, they have the field aspect of basaltic rocks, and we have not used the chemically derived name either in stratigraphic names or in descriptive sections of earlier chapters. The 72 analyses of other rocks include six samples from mafic dykes either associated with or possibly related to the Fortescue Group, 12 samples of felsic rocks, and 54 analyses of various pyroclastic and sedimentary rocks; the bulk of these last 54 analyses are the 37 analyses reported by Davy (1985).

The database

Geochemical data from the Fortescue Group were analysed using a PC-based geochemical data-processing software package developed within the Australian Geological Survey Organisation (AGSO) in Canberra, and called GDA (Geochemical Data Analysis). The system, which is described by Sheraton and Simons (1988), stores collections of analyses as MS–DOS files, and has facilities for interrogating files, printing tables of analyses, generating a range of plots, and calculating norms, as well as various statistical functions. The contents of the GDA database for the Fortescue Group are derived from several sources, which are listed below in order of number of analyses; the stratigraphic positions and locations of the rocks included in each source are described thereafter.

- One hundred and seventy-seven analyses of Fortescue Group mafic volcanic rocks carried out as part of the Pilbara Volcanic Study (PVS: see **Chapter 1**). These analyses include the 11 standard major and minor oxides (including Fe_2O_3 and FeO, but see below), and also H_2O+, H_2O-, and CO_2. All have a suite of 15 trace elements (Ba, Ce, Cu, Ga, La, Nb, Ni, Rb, S, Sc, Sr, V, Y, Zn, Zr) and many include Co, Cr, Pb, Th, and U. A total of 18 samples have either seven (Ce, Nd, Sm, Eu, Tb, Yb, Lu), six (Lu not determined), or five (Lu and Nd not determined) REE (Glikson et al., 1986a). Analytical methods are specified in Glikson et al. (1986b); note that the separate reporting of FeO and Fe_2O_3 is based partly on titration for Feo, and partly on an assumption that $FeO{:}Fe_2O_3{=}9$, after XRF determination of total Fe as Fe_2O_3
- One hundred and three analyses carried out at Chemistry Centre (W.A.) on rocks selected by us for analysis during the course of this project. These analyses include 11 major and minor oxides (including Fe_2O_3 and FeO), H_2O+, H_2O-, CO_2, and 22 trace elements (Ba, Ce, Cr, Cu, Ga, La, Li, Nb, Ni, Pb, Rb, S, Sc, Sn, Sr, Th, U, V, W, Y, Zn, Zr). None of these analyses has been previously published. Analytical methods are essentially as reported by Nelson, D. R. et al. (1992).
- Fifty-eight analyses of Fortescue Group rocks carried out by G. Sieber at the Max-Planck Institut for Geochemie, Mainz, in the course of a PhD study concerned with the geochemical evolution of mafic and ultramafic rocks during the stabilization of the Pilbara Craton. These analyses have ten major and minor oxides (Fe as FeO), LOI (loss on ignition), and normally ten trace elements (Ba, Co, Cr, Nb, Ni, Sr, Rb, V, Y, Zr), but some also include Cu or Sc. Twenty-five have a suite of 14 REE. These analyses are identified by numbers terminating in 'G' in the database; forty-five were made on samples collected by Sieber, and the remaining 13 on samples collected by A. M. Thorne. Separate pieces of five of the latter were also analysed by Chemistry Centre (W.A.), and these are counted as separate analyses in the preceding item. None of these analyses has been published elsewhere at the time of writing.

- Thirty-seven analyses reported by Cowley (1979). These analyses have 11 major and minor oxides (including Fe_2O_3 and FeO), and LOI, as well as up to 14 trace elements (Ba, Ce, Cr, Cu, Li, Nb, Nd, Ni, Pb, Rb, Sr, Y, Zn, Zr). Cowley (1979) also reported analyses of two dykes, which are not included in our database. Analytical methods are fully described by Cowley (1979).
- Thirty-seven analyses reported by Davy (1985) from successive continuous lengths of diamond drillcore representing 65.55 m of the uppermost part of the Roy Hill Shale Member of the Jeerinah Formation. These analyses have 11 major and minor oxides (including Fe_2O_3 and FeO), H_2O+, H_2O-, CO_2, and 28 trace elements (Ag, As, B, Ba, Be, Cd, Ce, Cl, Co, Cr, Cu, F, Ga, La, Li, Mo, Nb, Ni, Pb, Rb, S, Sr, Th, U, V, Y, Zn, Zr); however, detection levels for many of these elements are too high to provide useful results. Elemental C is also reported.
- Twenty-two analyses carried out at Chemistry Centre (W.A.) in support of MERIWA Project 82 (see **Chapter 1**). These analyses have 11 major and minor oxides (including Fe_2O_3 and FeO), H_2O+, H_2O- and CO_2. Fifteen of these analyses report 14 trace elements (Co, Cr, Cu, Rb, Nb, Ni, Sc, Sr, Th, U, V, Y, Zn, Zr); S is reported variably. Eighteen analyses include a suite of 11 REE. All these analyses, together with isotopic analyses of many of the same suite of samples, have been published elsewhere (Nelson, D. R. et al., 1992).
- Five analyses of unaltered Maddina Basalt (Maddina Formation) reported by Smith et al. (1982, table 1–3 analyses) and Smith (1975a, table 1–2 analyses). All five analyses include major oxides only.
- Two analyses of Kylena Formation from the southwestern corner of the MARBLE BAR 1:250 000 sheet reported by Hickman and Lipple (1975, Table 13). These two analyses were carried out at Chemistry Centre (W.A.), and have major oxides only.
- Three gabbro analyses from the basal part of the Gidley Granophyre (**Chapter 11**), reported by Wallace and Hoatson (1990, table 2). These analyses include major oxides and a suite of 21 trace elements.

Making sense of the data

With a total of 448 analyses (>9000 element determinations) of various rock types, distributed over the whole outcrop area of the Fortescue Group and collected from most stratigraphic units, some basis needs to be established to seek order within the database. In view of the fact that over 80% of available analyses are of mafic and ultramafic rocks, the bulk of this chapter must be concerned with these. The first natural division of the database for evaluation purposes is therefore between the 376 analyses of mafic and ultramafic rocks and the 70 analyses of other rock types.

Mafic and ultramafic rocks

Sample groups and clusters

The analyses of mafic and ultramafic rocks available to us have been made at various times, for various purposes, and in different laboratories. To provide an intelligible summary of their significance the analyses need to be grouped for comparative discussion and display. The system of subdivision used in most figures illustrating this chapter must first be explained.

The 376 analyses are first divided into 42 groups, each identified by a 4-letter mnemonic name, the **first two letters** of which indicate **stratigraphic position** and the **last two letters** of which indicate **geographic area**.

Letter pairs used for the first two letters of group mnemonic names are, in alphabetical order:

BE	Bellary Formation
BO	Boongal Formation
BU	Bunjinah Formation
CP	Cooya Pooya Dolerite
GG	Gidley Granophyre
HY	Hardey Formation
JE	Jeerinah Formation
KY	Kylena Formation
MA	Maddina Formation
MR	Mount Roe Basalt
NY	'Nymerina Basalt'
PY	Pyradie Formation

Letter pairs used for the last two letters are, again, in alphabetical order:

BC	Bellary Creek
BD	Bellary Dome
BM	CRAE Corehole BMW2
CA	Area studied by Cowley (1979)
CS	Same area as CA: sills
GH	Glen Herring
GR	Gregory Range
HR	Hardey River
KA	Karratha
GS	Collection area of Gudrun Sieber
HC	Hays Creek
LG	Lockyer Gap
MB	Marble Bar
MT	Meentheena
NE	NEWMAN 1:250 000 sheet
NG	Nullagine area
PD	Paraburdoo
PP	Pelican Pool
PS	Pindana Spring

2RD	Rocklea Dome
RS	Ray Smith (Smith et al., 1982)
SG	CRAE Corehole SGS1
TP	Tom Price
WC	Whim Creek area
YC	Yandicoogina

Thus, for example, the group identified by the 4-letter mnemonic name KYHC contains Kylena Formation samples from the Hays Creek area, and the mnemonic BUPD indentifies Bunjinah Formation samples from the vicinity of Paraburdoo. All the localities (listed above) identified by the last two letters of such names are shown in Figure 12.1a.

Greater clarification is also achieved, especially in plots with large numbers of data points, by plotting stratigraphically or regionally related groups together on a single diagram. The 42 groups with 4-letter mnemonic names are therefore further reduced to 11 clusters, each identified by a single capital letter. These single-letter clusters (A–K) run roughly in ascending stratigraphic sequence. The 4-letter mnemonic names of all of the 42 groups are listed in Table 12.1, which also shows their relationships to the 11 single-letter clusters.

Most of the 42 mnemonic-named groups include analyses from only one of the sources listed above, but some contain analyses from more than one. Figure 12.1b graphically summarizes both the stratigraphic and regional relationship of all 42 mnemonic groups, and of the 11 single-lettered clusters. That figure emphasizes the upward stratigraphic sequence of the clusters and the generalized framework these provide to look for systematic geographic and stratigraphic differences in chemical composition. The logic of the clusters, as shown in Figure 12.1b, may be summarized as follows:

- Cluster A — Mount Roe Basalt (MRPD, MRRD, MRBM, MRSG) and Bellary Formation basalt (BEPD) samples from the south Pilbara sub-basin;
- Cluster B — Mount Roe Basalt (MRGS, MRWC, MRMB, MRNG, MRYC, MRGH) samples from the main Fortescue Group outcrop area north of the Fortescue River;
- Cluster C — Kylena Formation (KYGS) samples from the western part of the main outcrop area, combined with the Cooya Pooya Dolerite (CPLG) and gabbro of the Gidley Granophyre (GGKA);
- Cluster D — Boongal Formation (BOBC, BOCA) and and Hardey Formation basalt (HYBC, HYCS) samples from the south Pilbara sub-basin;
- Cluster E — Pyradie Formation (PYBC, PYHR, PYCA, PYCS) samples from the south Pilbara sub-basin;
- Cluster F — Kylena Formation (KYHC, KYMT, KYPP, KYMB) samples from the eastern part of the main outcrop area;
- Cluster G — Nominal 'Nymerina Basalt' (Maddina Formation) (NYHC, NYMT) samples from the eastern part of the main outcrop area;
- Cluster H — Maddina Formation (MARS, MAGS, MAPS) samples from the western part of the main outcrop area;
- Cluster I — Maddina Formation samples from the eastern part of the main outcrop area (MAHC, MAMT) and the Gregory Range (MAGR);
- Cluster J — Bunjinah Formation (BUBC, BUSG, BUCA) samples from the south Pilbara sub-basin;
- Cluster K — Dolerite sills from the Jeerinah Formation (JEBD, JETP, JECA, JENE, JEGR).

Within most of the 42 groups the relative stratigraphic positions of the constituent samples are not defined, so that all the samples together represent a 'grab' sampling of the stratigraphic level and geographic area which the groups represents. However, some of the groups are made up of samples systematically collected within the unit they represent. Such groups are:

- MRRD — 9 samples equally spaced across the outcrop of the Mount Roe Basalt of the Rocklea Dome;
- MRPD — 17 of the 19 samples in this group are spaced across the outcrop of the Mount Roe Basalt near Paraburdoo;
- BEPD — 6 samples come from a single flow near the top of the lava sequence, arranged in two lots of three samples, each spaced upwards through the 8–9 m thick flow, and separated laterally by about 70 m. Two other samples of the BEPD group come from other flows near the base and the centre of the whole lava unit.
- MRGH — 23 samples spaced across the outcrop of the Mount Roe Basalt near Glen Herring;
- MRYC — 20 samples spaced across the outcrop of the Mount Roe Basalt at Yandicoogina;
- KYPP — 24 samples spaced across the outcrop of the Kylena Formation near Pelican Pool;
- KYMT — 13 samples spaced across the outcrop of the Kylena Formation near Meentheena;
- KYHC — 31 samples spaced across the outcrop of the Kylena Formation at Hays Creek;
- NYMT — 10 samples spaced across the outcrop of the lower part of the Maddina Formation (formerly Nymerina Basalt) near Meentheena;
- NYHC — 19 samples spaced across the outcrop of the lower part of the Maddina Formation (formerly Nymerina Basalt) at Hays Creek;
- MAMT — 10 samples spaced across the outcrop of the upper part of the Maddina Formation near Meentheena;
- MAHC — 7 samples spaced across the outcrop of the upper part of the Maddina Formation at Hays Creek;
- MAGR — 14 of the 17 samples spaced across the outcrop of the Maddina Formation of the Gregory Range.

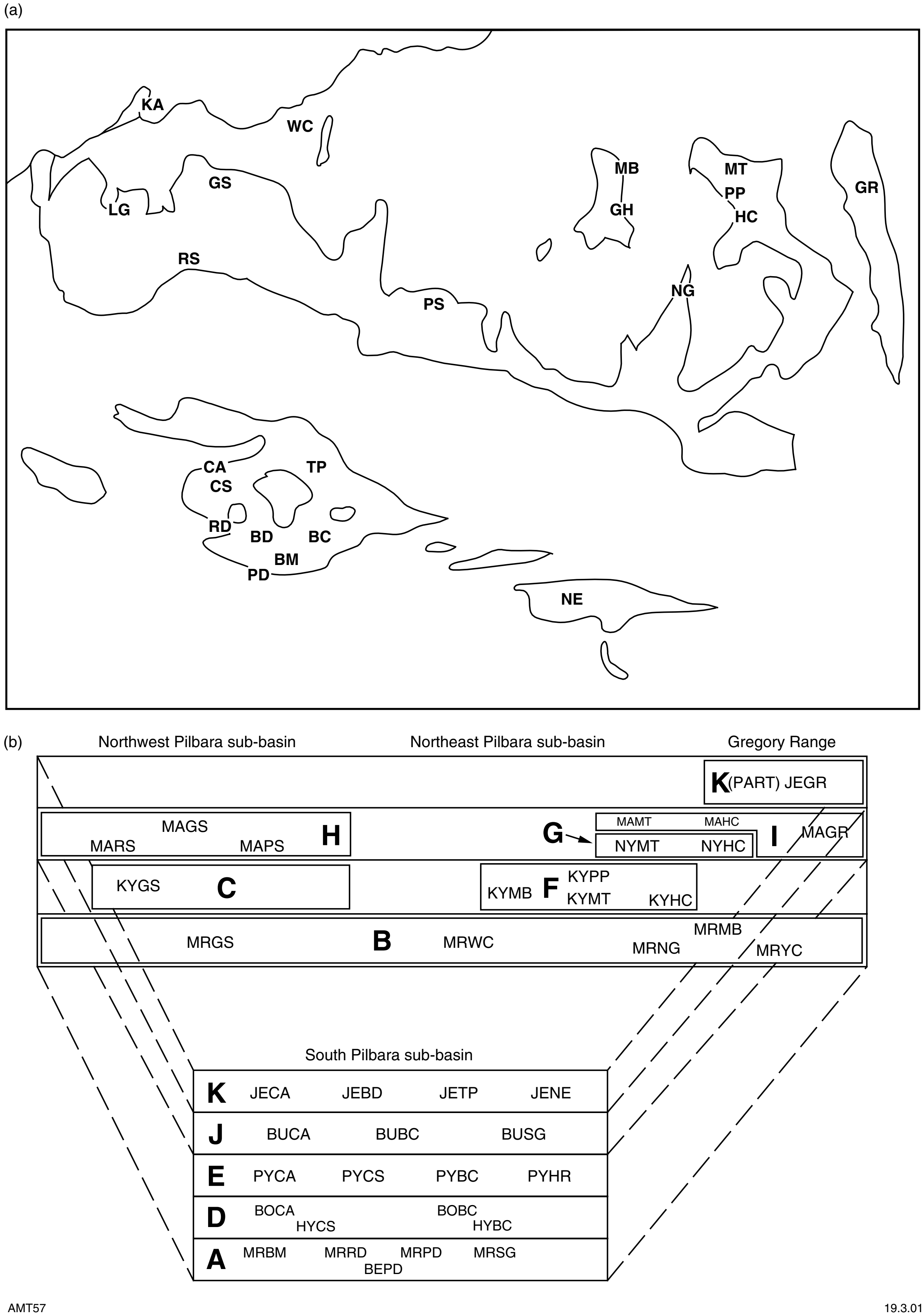


Figure 12.1. **Simplified geological map (a) of the Fortescue Group outcrop area, showing the general locations of 376 analysed mafic and ultramafic rocks, and (b) the stratigraphic and regional relationships between the forty-two 4-letter mnemonic groups and the eleven single-lettered (A–K) clusters into which small numbers (between 2 and 6) of the groups are combined. See Table 12.1 for a complete tabulation of the groups and clusters**

Table 12.1. Four-letter mnemonic groups and single-letter clusters used for ordering analyses of mafic and ultramafic rocks

Group mnemonic	*No.*[(a)]	*Number of samples*	*Single-letter clusters in group*
JECA	1	1	
JEBD	2	4	
JETP	3	5	**K**
JENE	4	9	
JEGR	5	7	
BUBC	6	2	
BUSG	7	2	**J**
BUCA	8	10	
MAHC	9	7	
MAMT	10	10	**I**
MAGR	11	17	
MARS	12	5	
MAGS	13	8	**H**
MAPS	14	4	
NYHC	15	19	**G**
NYMT	16	10	
KYHC	17	31	
KYMT	18	13	**F**
KYPP	19	24	
KYMB	20	2	
PYBC	21	10	
PYHR	22	11	**E**
PYCA	23	3	
PYCS	24	6	
HYCS	25	4	
HYBC	26	1	
BOBC	27	2	**D**
BOCA	28	4	
KYGS	29	14	
CPLG	30	11	**C**
GGKA	31	3	
MRGS	32	21	
MRWC	33	5	
MRMB	34	7	**B**
MRNG	35	3	
MRYC	36	20	
MRGH	37	23	
MRPD	38	19	
BEPD	39	8	
MRRD	40	9	**A**
MRBM	41	1	
MRSG	42	1	

NOTE: (a) The group numbers in the second column are of interest only to users of the GDA system

General features of present composition

Average compositions of all 42 groups are set out in Table 12.2. Figure 12.2 displays major and minor oxide, and selected trace element plots of the 376 available analyses of mafic and ultramafic rocks. The format of the plots is explained in the caption: it provides, for each chemical constituent, a graphic display of the broad variation within the entire dataset, as well as some indication of the coherence of the data within each cluster and within the component groups of the cluster. It is clear that a large percentage of the samples are altered. This is indicated by high values for LOI, H_2O+, or CO_2 in Table 12.2, and by the degree of compositional scatter for many components on Figure 12.2. Because the approach adopted here is to include all available data, rather than only those data from least-altered samples, only very general observations are made on the present compositions of the rocks. In particular, secondary addition of CO_2 and H_2O has effectively lowered the percentage contents of the major oxides, and values shown for these are therefore not directly comparable to values typically present in less altered rocks.

Major oxides

Figure 12.2 shows clearly that the coherence of the entire dataset varies widely in different major constituents. Thus total Fe, MgO, Al_2O_3, and SiO_2 vary less than CaO, K_2O, and Na_2O, which all show extreme scatter.

The K cluster, containing all sills from the Jeerinah Formation, stands out compositionally from all other clusters, having the lowest SiO_2, Na_2O, K_2O, and P_2O_5, as well as the highest Al_2O_3 and CaO. It can be argued that individual groups within other clusters (e.g. HYCS within D, and GGKA within C) are similar in major oxide compositions; this point is pursued further below, in discussion of cluster homogeneity.

Apart from the low-SiO_2 K cluster, the average SiO_2 compositions of clusters do not vary greatly from the 376-sample average of 56.5% (all SiO_2 averages rounded to nearest 0.1%), as listed below:

Cluster	Average SiO_2 (Wt %)
K	50.3
J	58.7
I	58.2
H	61.1
G	54.9
F	56.9
E	54.5
D	55.0
C	55.7
B	57.9
A	58.6

The only other exceptional cluster in these terms is H, whose average exceeds that of the whole dataset significantly; two of its component groups (MAGS and MAPS) have averages of 62% and 63.5% respectively. However, two component groups within cluster B, MRGS and MRMB, have SiO_2 averages of 57% and 58% respectively. These exceptions apart, there is a rather uniform SiO_2 content at all stratigraphic levels, with no consistent variation with height (=time). Although the mean SiO_2 content of the 117 Mount Roe Basalt and

Table 12.2. Average chemical compositions of forty 4-letter sample groups

Cluster			*A*					*B*			
Group	*MRPD*	*BEPD*	*MRRD*	*MRBM*	*MRSG*	*MRGS*	*MRWC*	*MRMB*	*MRNG*	*MRYC*	*MRGH*
No. of samples	*19*	*8*	*9*	*1*	*1*	*21*	*5*	*7*	*3*	*20*	*23*
					Percentage						
SiO_2	55.87	57.39	58.18	62.92	58.74	56.92	59.43	60.64	57.21	54.77	57.93
TiO_2	1.10	.94	1.29	1.61	.91	0.83	1.52	1.16	1.07	1.17	0.86
Al_2O_3	16.34	14.03	15.18	15.53	14.33	15.15	14.20	14.90	15.29	14.65	15.17
Fe_2O_3	1.44	1.83	1.46	1.32	2.36	1.16	1.51	14.93	2.28	0.99	1.51
FeO	6.84	9.43	6.74	5.32	7.71	9.82	8.22	–	7.34	11.83	9.44
MnO	.18	.20	.20	.09	.17	0.16	0.17	.13	.17	0.15	0.14
MgO	3.46	4.21	5.88	3.02	4.03	5.20	4.86	5.08	4.91	4.50	3.91
CaO	8.87	8.30	5.94	2.47	7.23	7.19	6.28	.85	6.80	8.47	7.14
Na_2O	4.28	3.17	3.12	4.50	3.89	2.78	1.97	1.09	4.26	2.50	2.58
K_2O	1.38	.39	1.69	2.79	.50	0.58	1.25	1.13	0.49	0.77	1.18
P_2O_5	.24	.11	.32	.43	.13	0.22	0.59	0.09	0.18	0.20	0.14
Total	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**
Volatiles	8.89	10.31	7.93	2.88	2.41	5.1	5.17	4.25	4.75	19.03	7.45
					Trace elements (parts per million)						
As	–	–	–	–	–	–	–	–	–	–	–
Ba	467	153	938	–	–	809	791	151	–	316	564
Ce	43	32	72	86	52	41	151	14	33	35	44
Co	–	–	–	42	71	46	40	50	45	–	48
Cr	135	22	269	310	83	183	178	367	120	124	27
Cu	48	221	34	16	88	40	21	62	52	87	74
Ga	18	16	17	–	–	–	12	16	–	14	15
Hf	–	–	–	–	–	–	–	–	–	3	2
La	21	15	32	43	28	21	75	7	16	15	22
Li	34	47	42	–	–	–	–	30	–	–	–
Nb	8	10	11	12	8	7	13	5	4	4	5
Nd	–	–	–	42	24	18	79	–	18	15	20
Ni	110	134	109	92	96	138	86	105	90	184	96
Pb	7	5	13	–	–	–	11	12	–	5	7
Rb	49	15	45	66	18	12	40	40	17	20	39
Pr	–	–	–	–	–	4	–	–	–	–	–
S	.–	–	–	–	.01	–	80	1 104	–	284	189
Sc	19	14	17	26	26	24	19	–	25	25	24
Sn	3	3	2	–	–	–	–	–	–	–	–
Sr	191	190	133	177	332	313	909	39	280	327	303
Th	4	4	6	4	7	–	7	6	3	3	7
U	1	1	1	1	2	–	1	1	1	1	1
V	203	178	196	220	199	180	200	275	208	204	201
W	2	2	2	–	–	–	–	–	–	–	–
Y	25	17	36	30	27	24	39	16	22	23	19
Zn	80	91	118	91	95	119	120	188	101	101	97
Zr	198	121	280	266	160	121	328	123	152	153	138

Bellary Formation samples of clusters A and B is marginally lower than that of the 48 Maddina Formation samples of I and J, the variation does not suggest a significant difference between the initial and final lavas of the Fortescue Group succession.

In spite of the somewhat consistent average SiO_2 contents throughout the dataset, it appears from Figure 12.2 that there is enough heterogeneity within clusters C, D, and E for their averages to be misleading. From Table 12.2 it is clear that the allocation of CPLG and GGKA to C on the basis of their stratigraphic association with the Kylena Formation of the western part of the main northern outcrop area (KYGS) is not reflected by any uniformity in major oxide composition of these three groups. While KYGS shares the high-SiO_2 attribute of H, groups CPLG and GGKA, with SiO_2 of about 54% and 52% respectively, show some similarity to cluster K. Similarly, within D, the four samples of HYCS have an average of about 49% SiO_2, whereas within E, the three samples of PYCA have an average SiO_2 content of about 52%.

From Figure 12.2 and Table 12.2, it is plain that these within-cluster heterogeneities are expressed in other major oxides. Thus, within cluster C, CPLG has low TiO_2 and Al_2O_3 combined with high MgO, whereas both CPLG and GGKA have very low K_2O. Within the D cluster, HYCS has low TiO_2, Al_2O_3, and K_2O, but very high MgO. And within the E cluster, there is even greater spread of the

Table 12.2. (continued)

Cluster		*C*				*D*				*E*	
Group	*KYGS*	*CPLG*	*GGKA*	*HYCS*	*HYBC*	*BOBC*	*BOCA*	*PYBC*	*PYHR*	*PYCA*	*PYCS*
No. of samples	*14*	*11*	*3*	*4*	*1*	*2*	*4*	*10*	*11*	*3*	*6*
						Percentage					
SiO_2	61.01	54.03	51.89	48.68	56.30	57.32	57.49	55.07	54.78	52.03	55.6
TiO_2	1.05	.55	1.31	.43	1.52	.75	.69	1.25	.91	1.55	1.15
Al_2O_3	12.97	12.49	14.54	10.25	13.36	15.07	15.25	11.95	11.63	13.98	13.29
Fe_2O_3	1.65	1.58	3.11	3.51	2.96	2.40	2.29	1.82	1.35	3.07	2.13
FeO	9.41	8.04	8.71	7.34	8.69	6.26	6.39	9.68	9.62	9.02	7.90
MnO	.14	.14	.16	.17	.17	.14	.16	.21	.14	.18	.15
MgO	3.81	13.12	6.82	21.76	5.34	6.41	5.86	10.30	11.67	6.87	8.17
CaO	5.16	7.67	10.62	6.16	8.09	8.45	7.78	6.48	7.49	10.11	8.40
Na_2O	2.94	1.60	2.05	1.09	3.24	2.44	2.88	2.47	1.91	2.65	2.33
K_2O	1.72	.70	.64	.54	.08	.66	1.10	.59	.36	.29	.67
P_2O_5	.14	.08	.15	.07	.25	.10	.11	.18	.14	.25	.21
Total	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**
Volatiles	3.89	–	3.22	6.13	3.46	3.09	2.17	5.57	5.00	2.24	2.66
						Trace elements (parts per million)					
As	–	–	3	–	–	–	–	–	–	–	–
Ba	546	302	394	299	88	237	448	117	121	153	155
Ce	34	14	20	15	27	31	36	26	17	33	35
Co	42	–	–	–	–	–	–	76	66	–	–
Cr	163	1 376	158	–	46	199	201	734	990	292	–
Cu	–	50	63	64	12	66	56	65	45	59	39
Ga	–	10	16	–	17	15	–	13	14	–	–
Hf	–	–	–	–	–	–	–	–	–	–	–
La	16	8	10	–	9	15	–	10	13	–	–
Li	–	–	10	8	15	16	25	28	12	5	24
Nb	8	2	6	2	8	6	39	6	5	13	27
Nd	15	–	15	3	–	–	10	12	11	32	14
Ni	58	377	106	554	53	118	71	261	246	55	81
Pb	–	5	4	12	6	8	–	6	7	50	10
Rb	51	29	33	14	1	28	34	14	8	9	25
Pr	4	–	–	–	–	–	–	2	2	–	–
S	–	–	–	–	–	–	–	–	–	–	–
Sc	30	25	33	–	32	26	–	28	27	–	–
Sn	–	–	–	–	2	3	–	3	9	–	–
Sr	221	209	186	94	142	162	213	45	79	198	95
Th	–	2	2	–	4	6	–	3	1	–	–
U	–	1		–	2	1	–	1	1	–	–
V	173	145	240	–	271	181	–	228	177	–	–
W	–	–	–	–	2	2	–	2	2	–	–
Y	26	11	20	9	44	22	18	30	25	44	31
Zn	–	59	90	69	60	81	71	93	77	89	66
Zr	134	58	89	37	198	125	93	146	113	183	145

data in oxides other than SiO_2; for example, MgO is high in both PYBC and PYHR, and both also have low Al_2O_3. In spite of their internal heterogeneity, the general affinity of the C, D and E clusters is underlined by the mg diagram, in which they are the only clusters to include analyses with values greater than 70, although some groups within them (e.g. KYGS) are in accord with the average of the whole dataset. All groups within E are notably low in K_2O, as are all in C and D except KYGS and BOCA.

Apart from C, D, and E the lower two clusters, A and B, are of interest not only in that they generally show more scatter than most higher lava clusters (e.g. F and I) but also because some oxides show some indication of consistent differences between component groups. This is well shown for B by TiO_2, where the six component groups appear to fall into distinct low-TiO_2 (MRGS, MRGH) and high-TiO_2 types (MRWC, MRMB, MRNG, MRYC). In cluster A, the Al_2O_3 values seem to show a comparable pattern, with MRPD and MRBM high and the remaining groups (BEPD, MRRD, MRSG) low.

Trace elements

In terms of trace element data, cluster K again appears as compositionally distinctive. It has by far the lowest average values for Ba, REE (as judged by La and Ce), Rb, Th, Y, and Zr, for many of these elements by a factor of ×0.5 compared with any other cluster; its

Table 12.2. (continued)

Cluster	F				G		H		
Group	*KYHC*	*KYMT*	*KYPP*	*KYMB*	*NYHC*	*NYMT*	*MARS*	*MAGS*	*MAPS*
No. of samples	*31*	*13*	*24*	*2*	*19*	*10*	*5*	*9*	*4*
					Percentage				
SiO_2	56.54	56.68	57.63	56.87	54.27	55.31	57.30	62.39	63.60
TiO_2	1.21	1.07	1.11	.73	1.39	1.17	.96	1.19	1.32
Al_2O_3	14.13	14.52	14.09	14.90	13.06	13.19	14.97	12.51	12.17
Fe_2O_3	1.70	1.40	1.52	1.00	2.23	2.49	1.39	1.41	2.57
FeO	8.85	8.46	8.75	8.27	9.79	7.94	8.49	12.71	9.07
MnO	.16	.13	.14	.12	.15	.15	.16	.17	.22
MgO	4.89	4.66	4.07	6.10	6.11	4.66	5.93	2.68	2.15
CaO	6.91	8.13	7.55	8.06	7.74	10.89	6.40	3.51	3.09
Na_2O	3.25	2.67	2.42	2.19	3.92	2.84	3.10	1.36	3.14
K_2O	2.07	2.02	2.48	1.07	1.11	1.21	1.62	1.91	2.50
P_2O_5	.29	.26	.24	.69	.23	.15	.13	.16	.17
Total	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**
Volatiles	3.99	4.03	4.56	3.50	6.00	6.06	4.3	5.27	3.84
					Trace elements (parts per million)				
As	–	–	–	–	–	–	–	–	–
Ba	560	591	569	–	520	493	–	356	–
Ce	86	76	90	–	53	50	–	53	76
Co	–	40	38	–	34	37	–	47	51
Cr	26	52	14	–	375	73	–	241	21
Cu	46	45	50	–	114	116	–	54	50
Ga	14	14	14	–	13	14	–	–	–
Hf	–	6	–	–	–	4	–	–	–
La	44	38	46	–	24	24	–	24	38
Li	–	–	–	–	–	–	–	–	–
Nb	19	17	19	–	8	8	–	11	13
Nd	–	34	–	–	–	27	–	22	34
Ni	45	52	35	–	131	82	–	95	21
Pb	7	8	10	–	5	8	–	–	–
Rb	78	75	95	–	20	34	–	45	62
Pr	–	–	–	–	–	–	–	5	–
S	300	261	382	–	223	284	–	–	–
Sc	25	39	26	–	21	20	–	25	34
Sn	–	–	–	–	–	–	–	–	–
Sr	149	171	236	–	348	380	–	73	76
Th	11	8	12	–	4	6	–	–	13
U	2	1	2	–	1	1	–	–	4
V	211	197	202	–	198	155	–	213	257
W	–	–	–	–	–	–	–	–	–
Y	46	38	44	–	24	25	–	34	42
Zn	80	78	81	–	105	82	–	104	108
Zr	361	304	355	–	170	174	–	182	234

average Sc content is the highest, and Ni the second highest.

Cluster F stands out in a way that could not be anticipated from its unexceptional major oxide composition. Although this is closely similar to most other lava clusters, its average incompatible element values — notably Ba, REE, Nb, Rb, Y, and Zr — are conspicuously high; it also has the lowest Ni content among those clusters.

On the other hand, the internal heterogeneity of clusters C, D, and E indicated by major oxide compositions is also reflected by trace elements, in a way that indicates that some of their constituent groups have compositional affinitiy with cluster K. Thus, in REE content, CPLG (within C), HYCS (within D), and PYBC (within E) are all very low. Both Ba and Rb are also low in HYBC (D), and Rb is also low in PYCA and PYHR (both E), while Zr is low in CPLG (C) and HYCS (D).

Summary of main features

An assessment based jointly on major oxides and trace elements suggests that the following general conclusions can be drawn concerning the 376 rocks for which analyses are available:

1. Cluster K, consisting of Jeerinah Formation sills, is compositionally distinctive in terms of both major

Table 12.2. (continued)

Cluster		*I*			*J*				*K*		
Group	*MAHC*	*MAMT*	*MAGR*	*BUBC*	*BUSG*	*BUCA*	*JECA*	*JEBD*	*JETP*	*JENE*	*JEGR*
No. of samples	*7*	*10*	*17*	*2*	*2*	*10*	*1*	*4*	*5*	*9*	*7*
					Percentage						
SiO_2	59.32	57.34	57.92	55.85	62.63	57.41	48.96	50.12	49.91	50.19	51.12
TiO_2	1.14	.88	1.20	.94	1.23	.93	.64	1.25	.91	0.94	2.17
Al_2O_3	13.83	14.93	14.32	16.91	13.13	15.11	16.84	14.98	15.59	15.56	14.33
Fe_2O_3	2.26	.83	3.56	2.34	1.08	2.13	1.86	2.90	2.38	2.07	1.98
FeO	7.86	8.83	6.86	6.48	8.50	7.42	8.21	10.09	8.61	9.04	11.58
MnO	.15	.14	.14	.12	.16	.16	.17	.25	.22	.22	.20
MgO	4.16	5.38	4.24	7.02	2.93	5.59	10.63	7.51	8.74	8.10	5.93
CaO	5.26	7.37	4.95	5.24	6.39	7.60	11.19	10.57	10.15	10.87	7.95
Na_2O	4.21	2.95	4.61	3.94	1.10	2.47	1.05	2.03	3.15	2.52	3.72
K_2O	1.58	1.23	1.95	1.03	2.70	1.04	.39	.19	.28	.44	.73
P_2O_5	.23	.12	.25	.13	.15	.14	.06	.11	.06	.05	.29
Total	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**100.00**	**101.00**	**100.00**
Volatiles	7.50	4.33	4.87	3.98	7.81	2.59	3.61	4.70	4.28	–	–
					Trace elements (parts per million)						
As	–	–	–	–	–	–	–	–	–	4	6
As	–	–	–	–	–	–	–	–	–	5	<4
Ba	477	573	669	285	–	488	101	43	60	106	457
Ce	75	48	85	33	67	54	20	8	4	4	45
Co	32	43	–	–	49	–	–	–	–	–	–
Cr	49	125	86	165	50	141	–	233	286	383	158
Cu	139	64	158	62	57	63	81	97	108	103	145
Ga	14	13	18	16	–	–	–	16	16	14	19
Hf	–	6	–	–	–	–	–	–	–	–	–
La	37	22	40	12	34	–	–	<5	3	2	19
Li	–	–	13	25	–	9	21	18	22	–	–
Mo	–	–	–	–	–	–	–	–	–	6	7
Nb	12	8	14	9	13	14	13	3	3	3	12
Nd	–	25	–	–	30	22	2	–	–	–	–
Ni	44	94	82	117	40	89	230	131	159	120	85
Pb	6	8	8	5	–	12	11	2	2	2	4
Pr	–	–	–	–	–	–	–	–	–	–	–
Rb	28	33	37	33	129	33	14	4	8	9	15
Pr	–	–	–	–	–	–	–	–	–	–	
S	173	212		–	–	–	–	–	–	2	2
Sc	13	31	12	31	30	–	–	37	35	–	–
Sn	–	–	3	3	–	–	–	<4	3	2	2
Sr	211	197	301	171	76	131	189	145	136	112	249
Th	5	6	5	8	11	–	–	<2	<2	1	4
U	1	1	1	2	3	–	–	<2	<2	<2	<2
V	145	213	146	219	301	–	–	292	247	241	274
W	–	–	2	3	–	–	–	<4	<4	–	–
Y	22	30	27	27	39	34	9	23	17	15	37
Zn	92	79	128	92	103	89	80	113	119	91	115
Zr	207	166	238	166	209	161	28	78	52	65	243

oxide and trace element composition. These dolerites have substantially lower SiO_2 and lower incompatible trace elements than those of any other cluster, together with higher Al_2O_3, MgO, and CaO.

2. Most ‘basalt’ units of the Fortescue Group have an average composition (recalculated to 100% after removal of volatile components) approximating to:

SiO_2	57.50%
TiO_2	1.05%
Al_2O_3	14.40%
Fe_2O_3	2.21%
FeO	8.12%
MnO	0.15%
MgO	4.83%
CaO	6.93%
Na_2O	3.20%
K_2O	1.40%
P_2O_5	0.21%

These values represent the (unweighted) arithmetic mean of the average values for the six clusters A, B, F, G, I, and J. They thus include all (264) analyses of Mount Roe Basalt, Kylena and Maddina Formation samples except the KYGS group and the H cluster. As in Table 12.2, these values sum to 100% and are on a

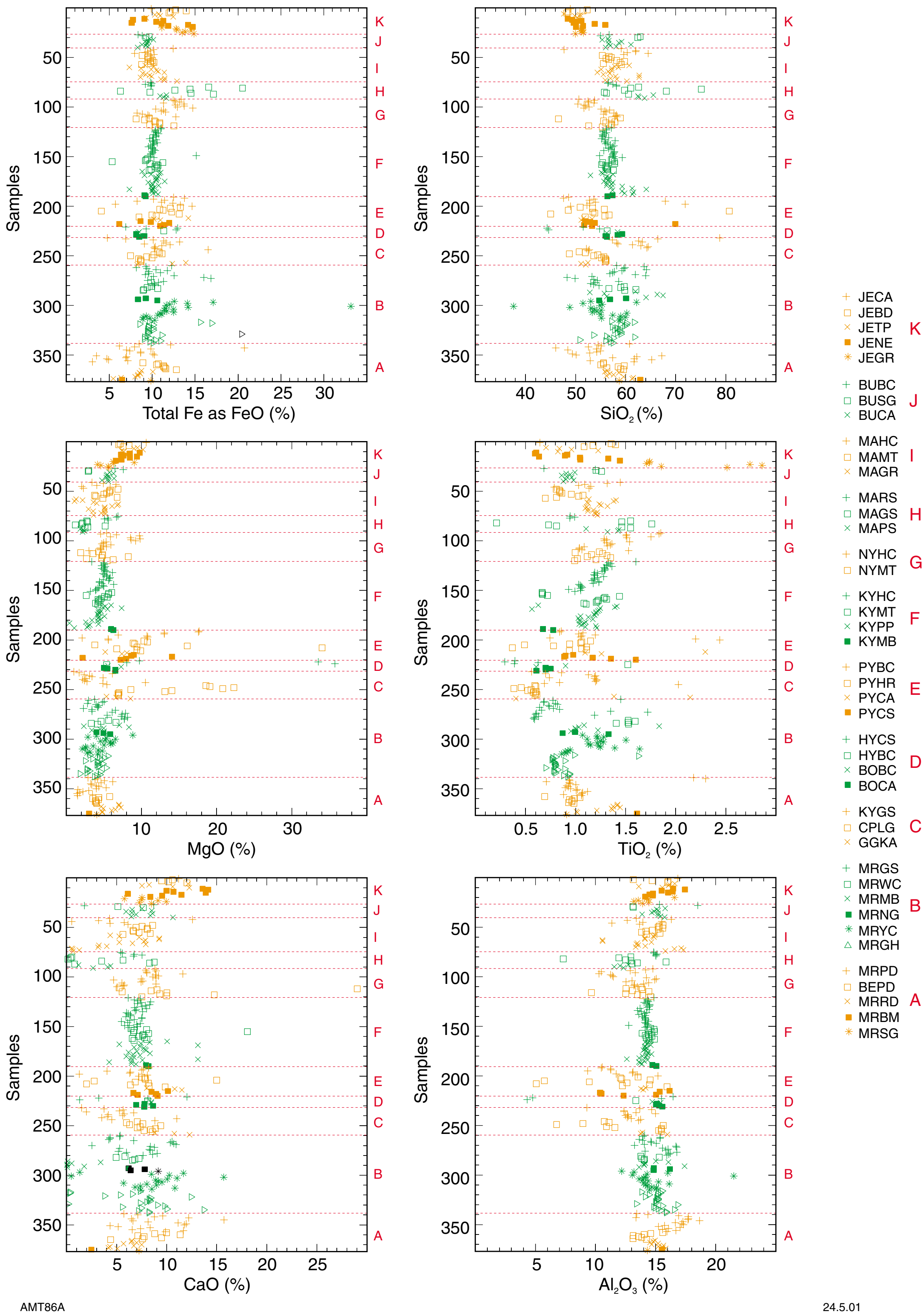
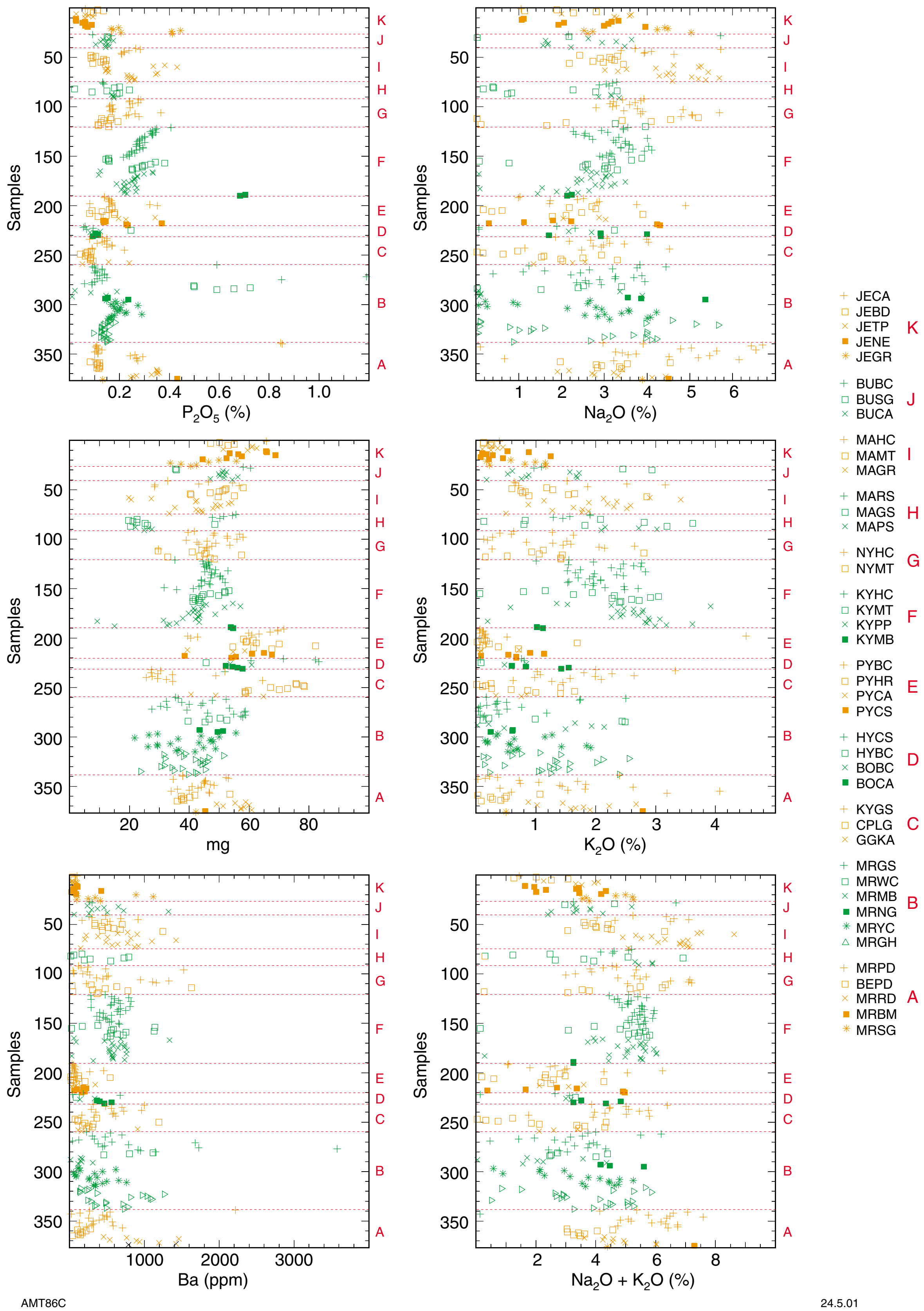


Figure 12.2. Plots displaying major and selected minor element compositions of 376 mafic and ultramafic rocks of the Fortescue Group. Each rectangular plot displays a single major oxide or trace element. Weight percent or ppm values are marked on the X coordinates. Analyses are plotted upwards at equal intervals on the Y coordinates, following the sequence of single-letter clusters. Within each cluster the groups are individually identified, from top to bottom, by the following sequence of symbols: upright cross (+), open square (□), diagonal cross (×), solid square (■), star (✱), open triangle (△). The symbols are applied in the sequence of Table 12.1, and restart for each cluster.



Thus JECA and BUBC both appear as +, and B is the only cluster which uses all six symbols. Clusters are demarcated by horizontal lines linking all diagrams. The effect is, for each constituent represented, to provide a graphic display, in roughly stratigraphic sequence, of its occurrence within the mafic and ultramafic rocks of the Fortescue Group, as well as of its variability. Thus the high-magnesian rocks within clusters C and E stand out clearly on the MgO diagram, with all other groups showing little variation. The mobility of such elements as alkalies shows up clearly on the $Na_2O + K_2O$ diagram

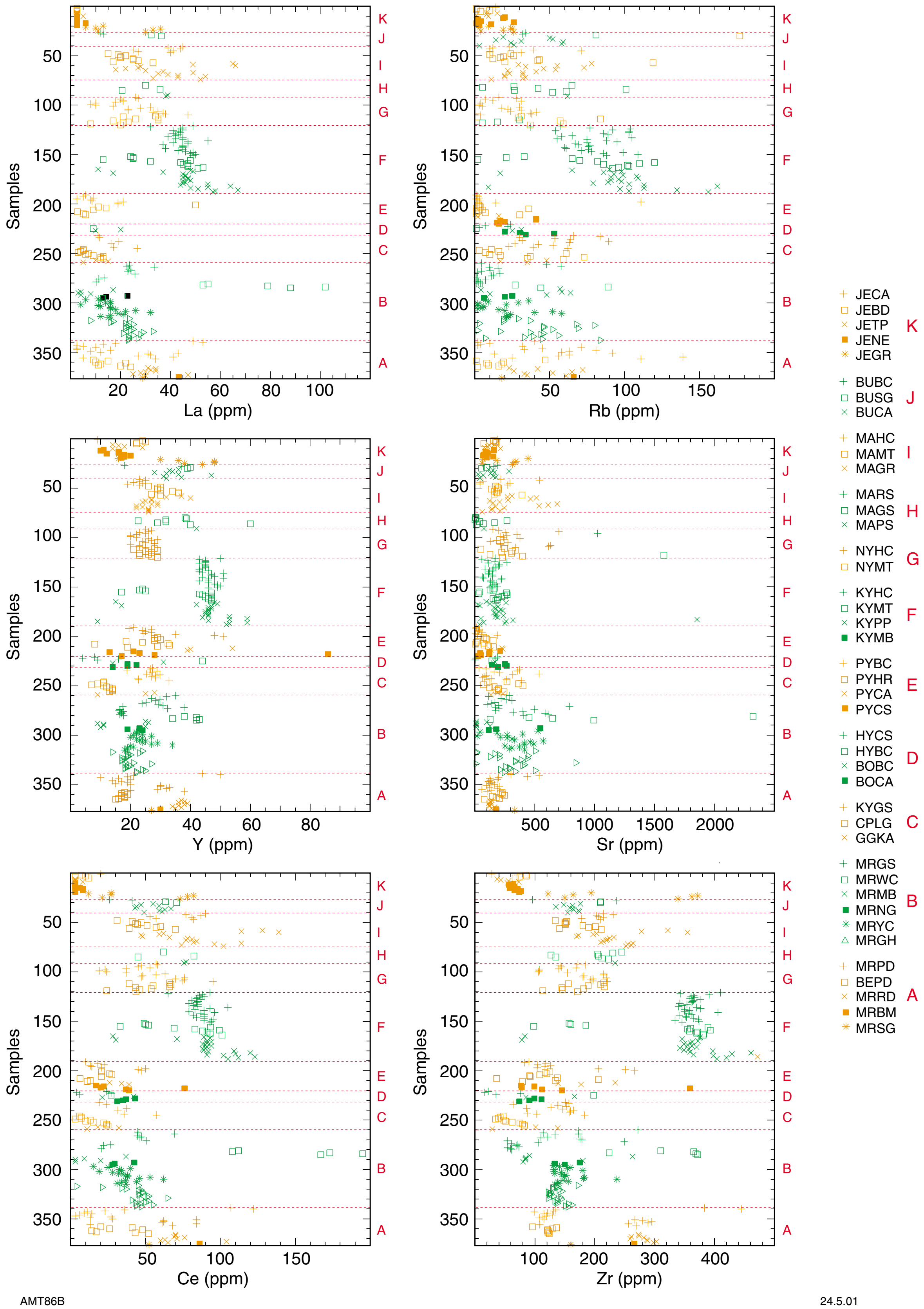


Figure 12.2. (continued)

volatile-free basis. Three important exceptions to the general homogeneity are emphasized as points 3–5 below.

3. Some groups from the Pyradie and Hardey Formations, particularly PYBC and PYHR (within E) and HYCS (within D) differ from the common composition set out in point 2, as do CPLG and GGKA, both intrusive rocks from within cluster C. All these rocks show some affinity with the unique cluster K, but differ in having conspicuously high MgO, particularly relative to both total Fe and CaO.

4. The KYGS group of the C cluster, as well as the MRGS and MRMB groups of the B cluster, have exceptionally high SiO_2, but differ in no other consistent way from the normal 'basaltic' composition listed above.

5. The bulk of samples of the the F cluster are characterized by very high incompatible elements, but are otherwise unexceptional in composition.

Alteration

The analytical data have so far been presented solely in terms of the actual compositions of the samples, as collected and analysed. To what extent these compositions correspond to those of erupted or intruded magmas is a vital question for the application of the data for nomenclatural purposes, for petrogenetic interpretation, or even for empirical comparison with volcanic provinces, either modern or ancient. Following the eruption of any lava flow there are three main times at which chemical modification may be expected: during or immediately after initial cooling; during any subsequent metamorphism that may have occurred; and during weathering, before the sample is collected for analysis. Evidence that the data under discussion have or have not been modified by one or all of these processes is now examined.

Effects of weathering: There are three main reasons to suppose that the compositions of the 376 analysed samples collected as mafic or ultramafic rocks have not been greatly modified by modern weathering:

- All authors reporting analyses compiled into our database describe in their collection procedure appropriate awareness of the need to remove all obvious evidence of weathering from the samples collected. The 103 samples for which we report analyses here for the first time were all collected from good exposures from which 'fresh', hard, dark-green or grey rock could be broken with a sledge hammer, and were so selected.
- Samples from drillcore have a similar appearance to such fresh surface rock, and give chemical results which show similar features. While only seven of the 376 analyses available to us were made on drillcore (all four samples from the BUSG, MRBM and MRSG groups, two from the MAGS groups, and one from MRGS), results reported (but not published) from drillcore by Meakins (1990) support the view that surface weathering has not had a significant influence on the 353 samples making up our database that were collected from surface exposures.
- Although many analyses in our database were made by XRF, with FeO and Fe_2O_3 partitioned on the assumption that $FeO:Fe_2O_3=9$, the majority include titration for FeO; one of the most obvious chemical effects of modern weathering is the oxidation of FeO, and it is evident from Table 12.2 that $FeO:Fe_2O_3$ ratios are generally of primary igneous aspect.

It is unfortunately not possible to obtain from thin-section examination any definitive evidence for or against the bulk chemical alteration of Fortescue Group mafic and ultramafic rocks. The mineral composition that may be assumed to have existed following initial cooling of the lavas has clearly undergone major modification in the later history of the rocks. Most importantly, primary pyroxene and labradorite are rarely present, and are represented mainly by a fine-grained mesh of secondary amphiboles and albite–oligoclase. It is not possible on petrographic grounds to differentiate between isochemical post-crystallization modification, and systematic large-scale change of initial chemical composition.

Effects of metamorphism: Detailed studies of chemical modification of lavas during low-grade metamorphism (Smith, 1968; Smith and Smith, 1976) have demonstrated significant mobility of some elements, but have not indicated systematic modification of the original bulk chemical compositions of the rocks. Smith (1968) introduced the concept of three-dimensional alteration domains with varied and complex relationships to the metamorphosed flows, and showed significant and different chemical alteration trends in domains of different types. However, Smith (1968, p. 217) apparently accepted that there had been no (or little) change in the bulk composition of the lavas that were the subject of that study.

Smith et al. (1982), in their examination of Fortescue Group lavas in the light of Smith's (1968) earlier work, identified such alteration domains within flows of the Maddina Basalt (Maddina Formation). They proposed (Smith et al., 1982, p. 77) that 'The massive central layer of each flow has a uniform composition *that is taken to be essentially the original composition of the flow . . .*' (our italics), although at the same time they stated that 'Metasomatic alteration of the pile is very significant' due to alteration in the upper parts of the flows.

It is methodologically difficult to obtain definitive evidence that determines whether lavas that have undergone low-grade metamorphism do or do not retain their primary bulk composition: that is, to demonstrate whether metamorphism has or has not been isochemical on a regional scale. However, there is empirical evidence within the body of analyses examined here to suggest that original lava compositions may not have undergone major regional modifications. Within cluster F, for example, three of its component groups, KYHC, KYMT, and KYPP, were collected across the outcrop of the Kylena Formation in three separate traverses within the same general area (Fig. 12.1). Figure 12.3 shows that, with aberrant samples

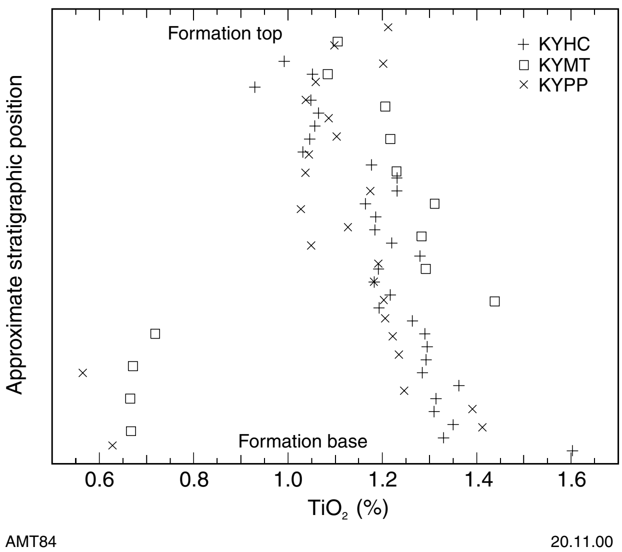


Figure 12.3. TiO_2 plotted against stratigraphic height for groups KYHC (vertical crosses, 31 analyses), KYMT (open squares, 13 analyses), and KYPP (diagonal crosses, 24 analyses). The samples are distributed at equal intervals across the full stratigraphic thickness of the Kylena Basalt; this thickness is not quantified, since it varies slightly between the three sample profiles, and is represented by the vertical height of the plot. Although the stratigraphic equivalence is thus diagrammatic, a consistent decrease in TiO_2 content with height is evident in each group

in the lower parts of KYMT and KYPP, there is a steady upward decrease in TiO_2 through the unit, an effect only likely to result from metamorphic modification if that process were stratigraphically selective. In passing, it may be noted that this argument also provides a fourth argument against any important influence of weathering, which would be equally unlikely to take place in a stratigraphically selective way.

Titanium is generally accepted as an element resistant to secondary redistribution (Smith and Smith, 1976) and it may still be argued that elements regarded as mobile may well have been radically affected. Figure 12.4a indicates the close coherence of K_2O and Rb for these same three groups, and Figure 12.4b the general concordance of these two components for all clusters. Although such a correlation would be expected in an unaltered suite of lavas, we cannot exclude the coupled mobility of K and Rb as a means of retaining the relationship through substantial modification. Most tests of this kind show substantial scatter, and it is obvious from Figure 12.2 that the spread of analysed values for alkalies, for example, is larger than that present in most modern, or Phanerozoic volcanic suites. However, in Figure 12.5, selected elements, including some normally accepted as 'mobile' under metamorphic conditions, are plotted against mg number. The spread of data seems consistent with the view that while metamorphism has indeed resulted in significant movement of such elements, the rearrangement has been intraformational, so that a sufficient number of analyses will yield a bulk composition reflecting that of the erupted magma. This view is contrary to that of Meakins (1990, p. 37), for example, who did not accept present SiO_2 values as primary, and concluded instea d that 'extensive secondary silicification has altered the original lava geochemistry . . .'

Early post-eruptive alteration: As described in earlier chapters, flow tops in 'basalt' units of the Fortescue group are easily recognized, and commonly show extreme alteration, which is obvious from field observation. The uppermost parts of flows are commonly brecciated, highly vesicular or amygdaloidal, and bleached to a very pale green or grey colour in freshly broken rock; often the material is clearly siliceous, and chert-like in general appearance.

Smith et al. (1982, fig. 1, table 1) have shown that such material is chemically enriched in SiO_2 and CaO, and grossly depleted in most other major constituents,

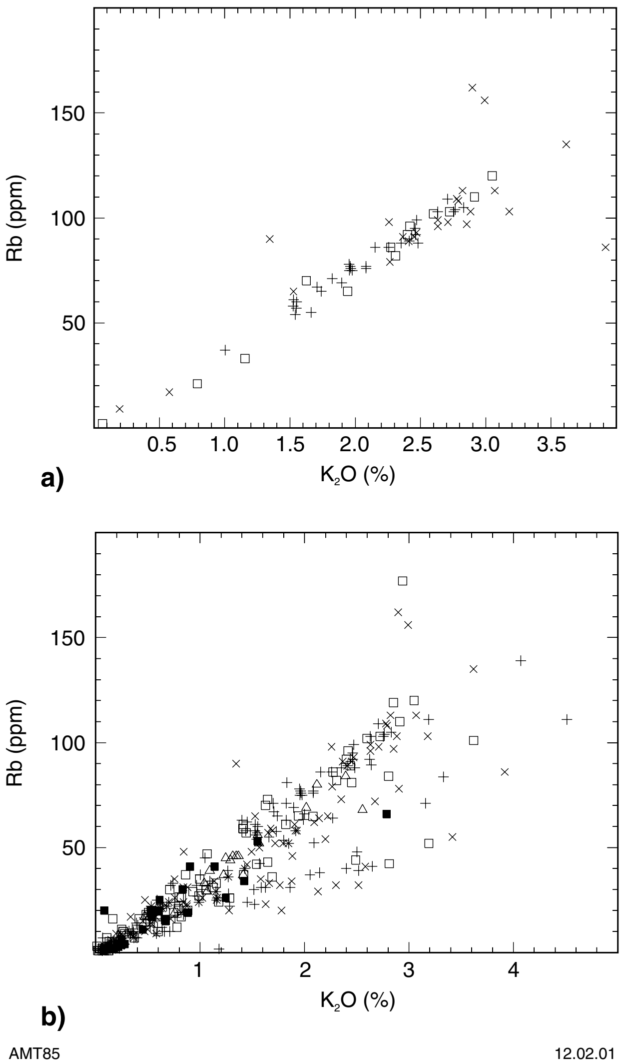


Figure 12.4. a) K_2O plotted against Rb for all samples of KYHC (vertical crosses, 31 analyses), KYMT (open squares, 13 analyses), and KYPP (diagonal crosses, 24 analyses); b) K_2O plotted against Rb for all 4-letter groups (376 analyses)

specifically Fe, MgO, and alkalies. It seems likely that such extreme alteration in a lithologically identifiable flow top is more likely to be the result of early post-eruptive processes, when volatile loss associated with the cooling and crystallization of the flow took place, and the exposed upper part of the flow was exposed to the atmosphere. None of the 376 analyses within our database includes such material.

Evidence from Ventersdorp comparison: We noted in **Chapter 1** the close resemblance of the Fortescue Group to the Ventersdorp Supergroup of the Kaapvaal Craton; the similarities include many general stratigraphic and lithological features, age, and status in the broad evolution of the Pilbara Craton and Kaapvaal Craton respectively. Myers et al. (1990) have recently published a geochemical study of the uppermost (Klipriviersberg) lava group of the Ventersdorp Supergroup. Their table 1 shows average chemical compositions of the chemical stratigraphic units which they define. A comparison of the rounded mean of columns 5 and 6 of their table 1 (recalculated by us on a volatile-free basis) representing an average composition of 206 rocks, is compared below with the average of the 264 nominal Fortescue Group 'basalts' already set out:

	Fortescue Group 'basalts' (%)	*Klipriviersberg Group lavas (%)*
SiO_2	57.50	57.19
TiO_2	1.05	1.18
Al_2O_3	14.40	15.39
Total Fe	7.40	8.55
MnO	0.15	0.15
MgO	4.83	5.02
CaO	6.93	7.48
Na_2O	3.20	3.42
K_2O	1.40	1.50
P_2O_5	0.21	0.11

It is not feasible to evaluate the significance of this similarity numerically, but, given the large numbers of analyses involved, it is difficult to accept that the close resemblance is not significant. We find it hard to believe that similar metamorphic modifications and/or modern weathering effects have coincidentally produced such similar compositions in the rocks of two widely separated volcanic suites of similar age on different continents. We conclude that, while some secondary alteration has undoubtedly taken place since both suites of lavas were erupted, this comparative evidence reinforces the conclusion from internal coherence, already noted, that the bulk compositions of Fortescue Group lavas reflect their primary compositions, in non-volatile oxides as well as trace elements.

Compositions and nomenclature of the erupted magmas

Major components: If it be accepted that present bulk compositions of the Fortescue Group rocks reflect primary composition, then based on major oxide composition, it follows that three magma types are represented: a rather uniform magma making up the bulk of the erupted lava, a more mafic magma confined to the sills of the Jeerinah Formation, and a high-magnesian type typically present in the Pyradie Formation.

The International Union of Geological Sciences Subcommission on the Systematics of Igneous Rocks (Le Maitre, 1989) currently recommends the silica/total alkalies diagram for the nomenclature of volcanics rocks from chemical composition. All the 376 analyses examined here are plotted on this diagram in Figure 12.6, separately for each of the 11 single-letter clusters. The same groupings that were discussed earlier can now be examined in terms of nomenclature. Analyses of the K cluster, dolerite sills of the Jeerinah Formation, uniquely fall almost entirely within the basalt field. Clusters D, F, G, H, I and J show variable scatter, but tend to be concentrated in the upper part of the basaltic andesite field close to, and spreading into, or beyond, the adjacent andesite field. This is especially clear for F, in which the individual analyses show least spread. Clusters C and E again show heterogeneity between their constituents. Within cluster C, CPLG and GGKA rocks are spread between the basalt and basaltic andesite fields, while the high-SiO_2 KYGS group scatters widely between the basalt and rhyolite fields. In cluster E, PYBC samples are spread between basalt and rhyolite, although the rocks of PYHR are more closely grouped, and centred between the basalt and basaltic andesite fields. Consistent with earlier discussion, both of the Mount Roe Basalt clusters (A and B) show much greater scatter than the upper 'basaltic' units; also, there is some indication that the A cluster (especially MRPD) is more alkaline than cluster B.

On Figure 12.7, all 11 clusters are plotted on a Jensen cation diagram (Jensen, 1976), which provides an alternative system for volcanic rock nomenclature. Similar nomenclatural groups are indicated, but with some modifications from the names suggested by the silica/total alkalies diagram (Fig. 12.6). Once again, the K cluster is both coherent and spatially distinct from all other clusters: the data fall roughly on the dividing line between high-iron tholeiitic basalt and high-magnesium basalt, and the rock name consequently suggested by the diagram is simply tholeiitic basalt. The uniform lava composition represented by clusters F (particularly), G, H, I, and J group close to the part of the central solid line separating the basalt field from that of high-iron tholeiitic basalt, but the scatter in many groups (for example MAGR of the F cluster) extends into the andesite field, and in places into dacite. As in the silica/total alkalies diagram, clusters A and B are centred about the same general area, but are scattered more widely than the stratigraphically higher clusters; there appears to be a significant difference in the preferred spread of these two clusters, with the data for A tending to lie below the tholeiite/calc-alkaline line and the B data above it. Clusters C, D, and E all show up clearly as chemically heterogeneous in this diagram, with CPLG (in C), HYCS (in B), and both PYBC and PYHR (in E) extending into the komatiitic field.

The terms calc-alkaline and tholeiitic do not appear to be helpful in describing the chemistry of these rocks, and particularly of the main lava composition, since in an AFM diagram most clusters lie along the line commonly used to distinguish these two chemical types (Fig. 12.8). In terms of major element composition, the bulk of the

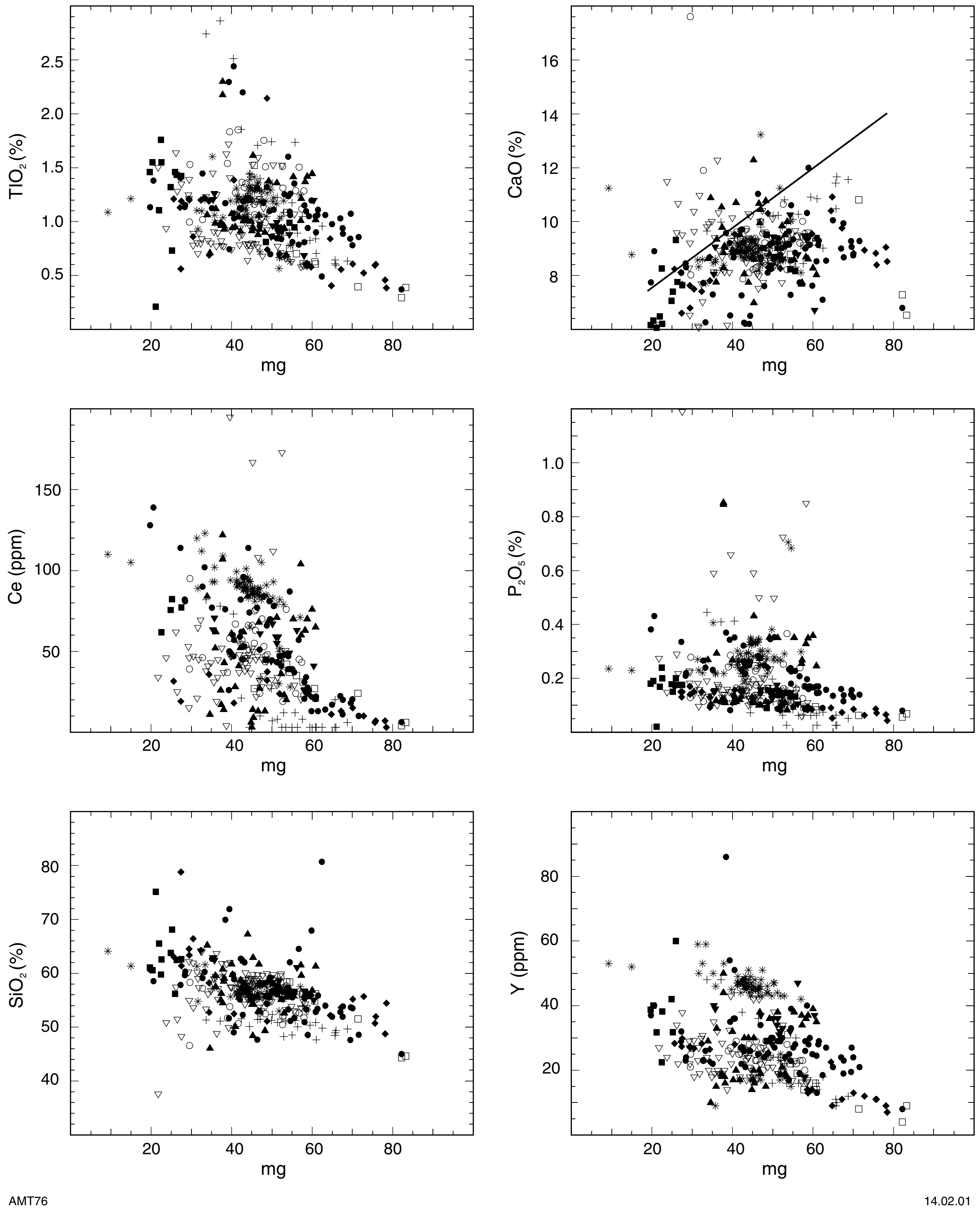


Figure 12.5. Magnesium number (mg) plotted against various components. On the CaO and Al_2O_3 (all groups) diagrams, lines have been marked showing the typical MORB trends suggested by Wilkinson (1981)

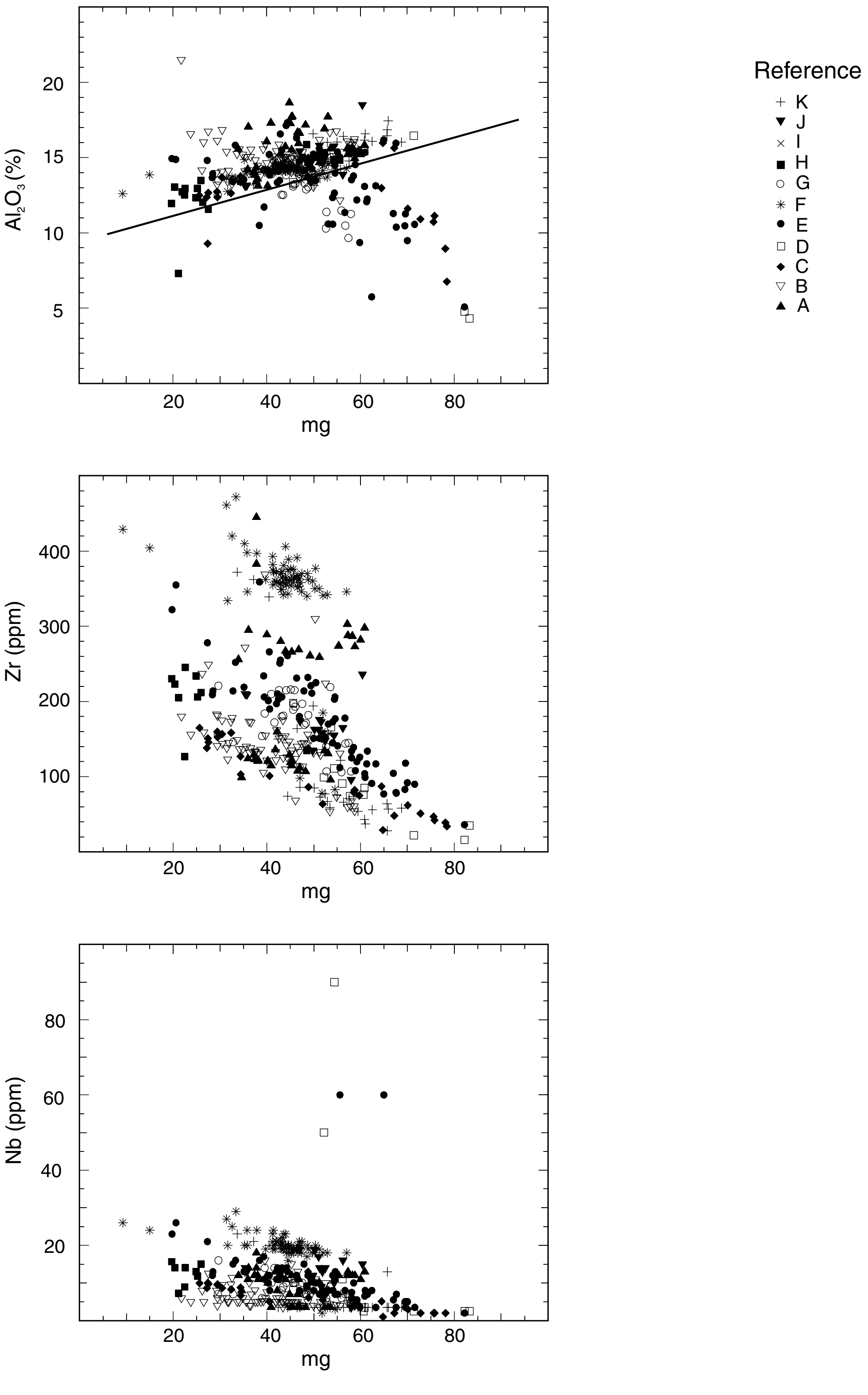


Figure 12.5. (continued)

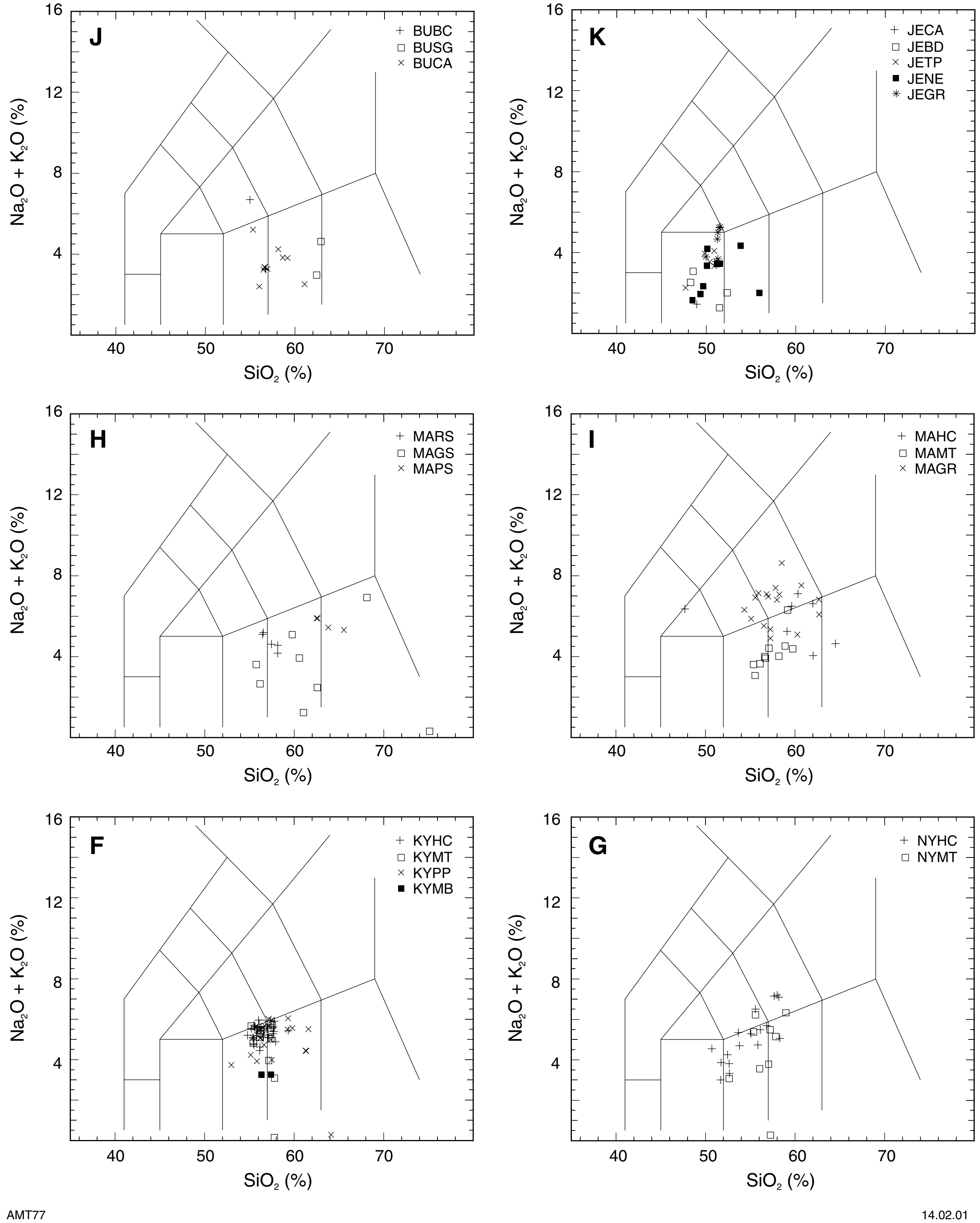


Figure 12.6. Silica/total alkalies plots for the eleven clusters of mafic and ultramafic rocks. The field divisions on each diagram are those of Le Maitre (1989) recommended by the International Union of Geological Sciences Subcommission on the Systematics of Volcanic Rocks. Lower case letters in relevant fields of the reference diagram indicate the following recommended names: a: andesite; b: basalt; ba: basaltic andesite; bta: basaltic trachyandesite; d: dacite; r: rhyolite; t: trachyte; ta: trachyandesite; tb: trachybasalt; td: trachydacite

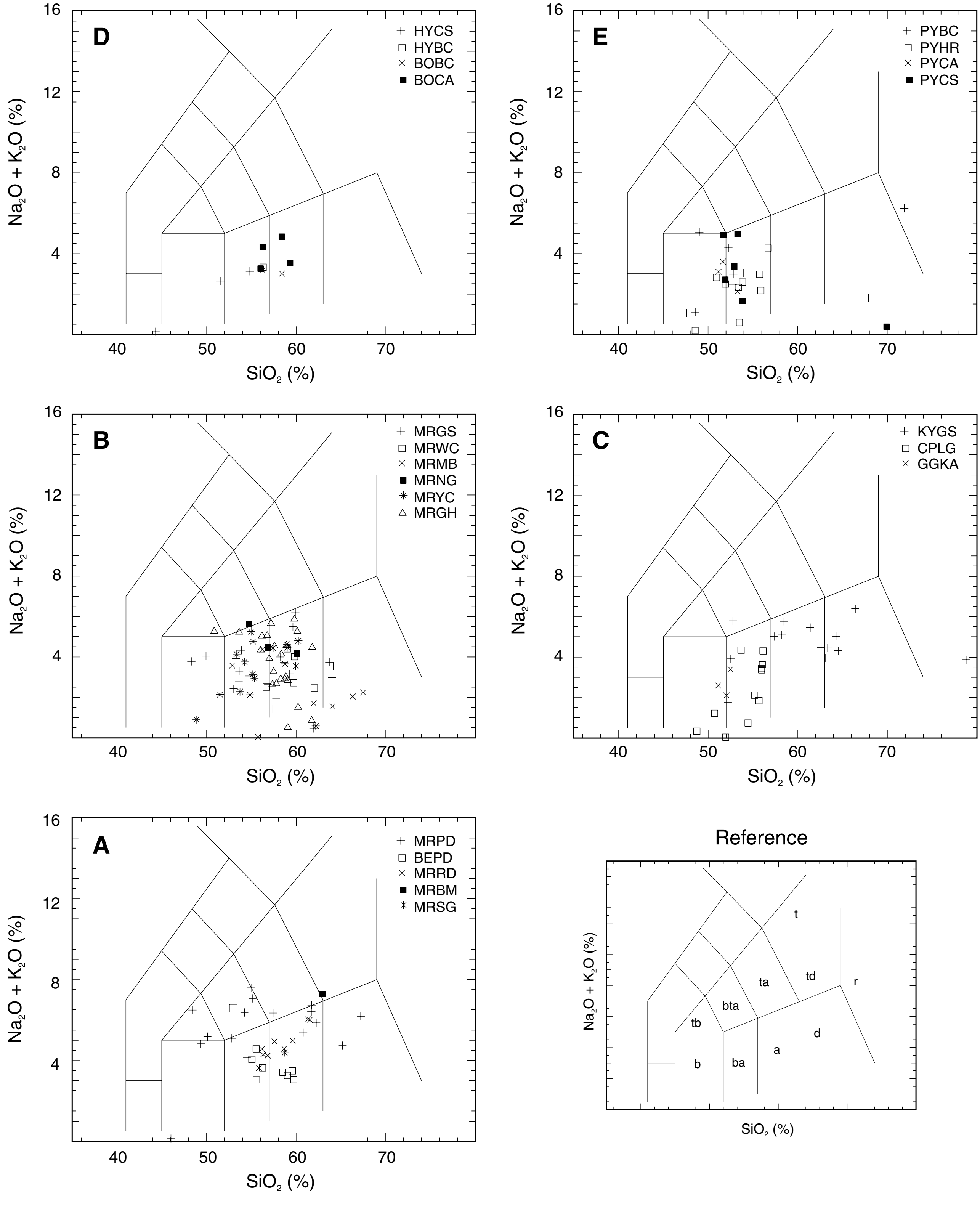


Figure 12.6. (continued)

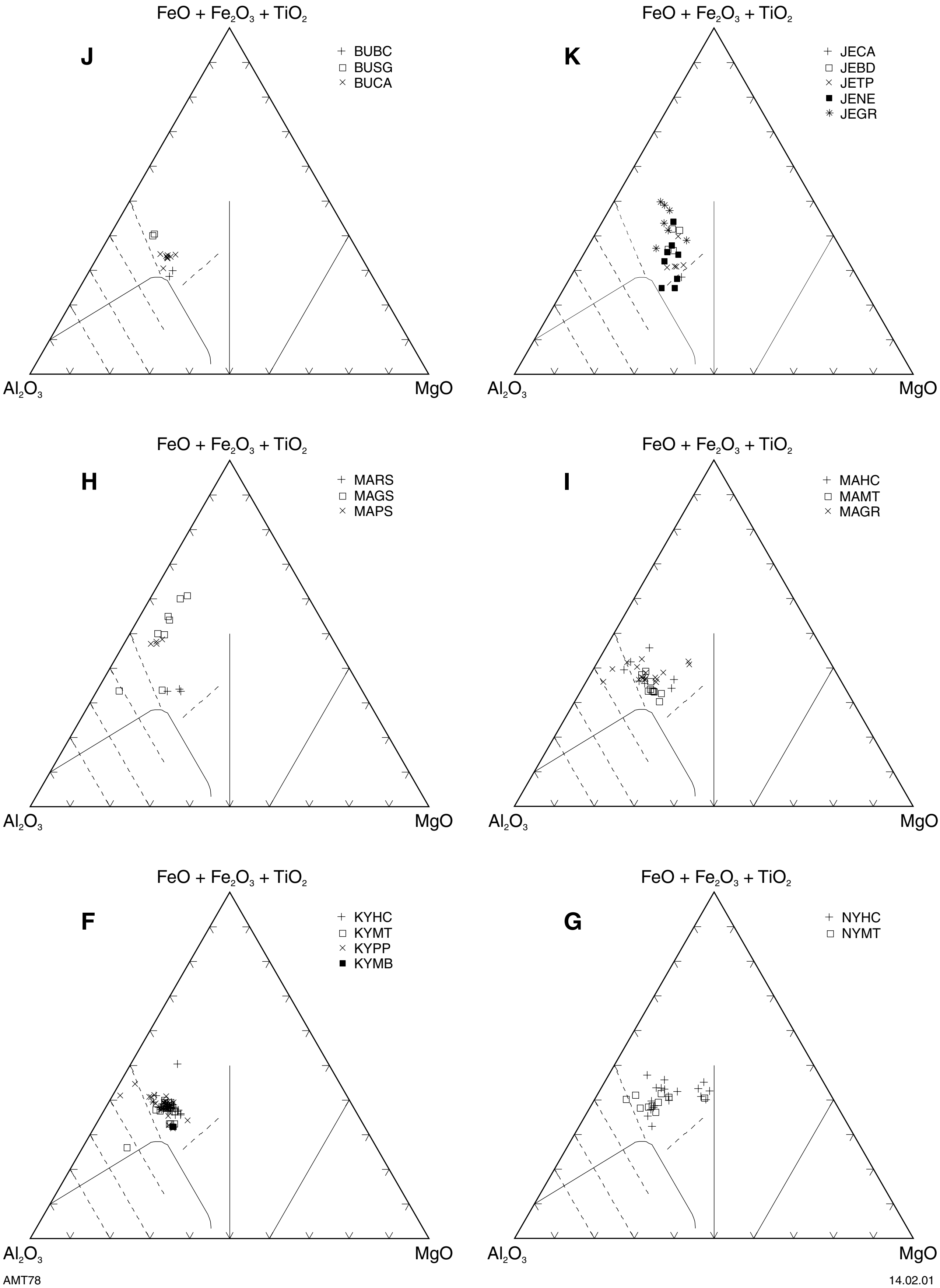


Figure 12.7. Jensen cation diagrams (Jensen, 1976) for the eleven clusters of mafic and ultramafic rocks

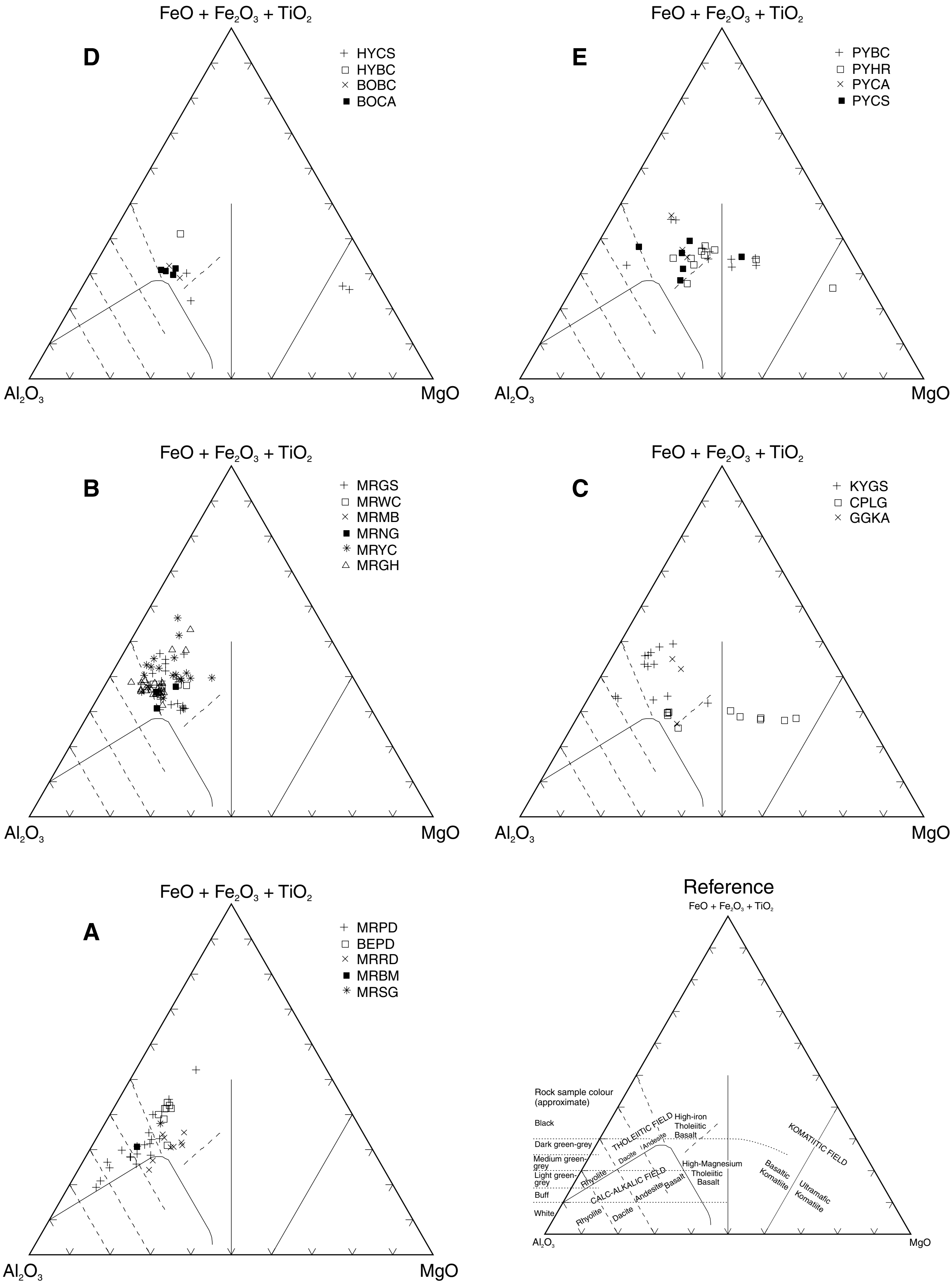


Figure 12.7. (continued)

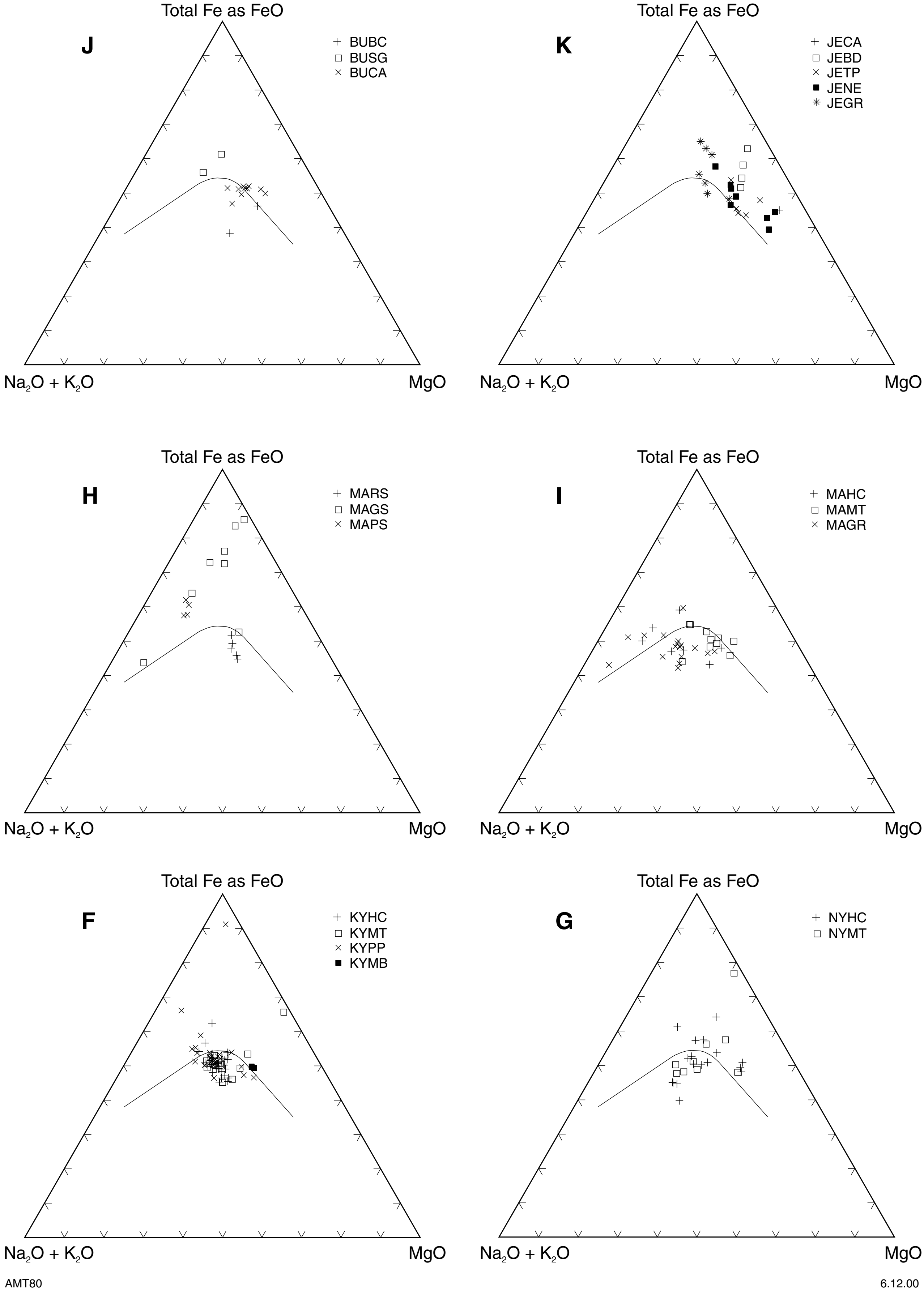


Figure 12.8. AFM diagrams for the eleven clusters of mafic and ultramafic rocks. The curved line within each diagram (Irvine and Baragar, 1971) is commonly taken as a division between tholeiitic rocks (above) and calc-alkaline rocks (below)

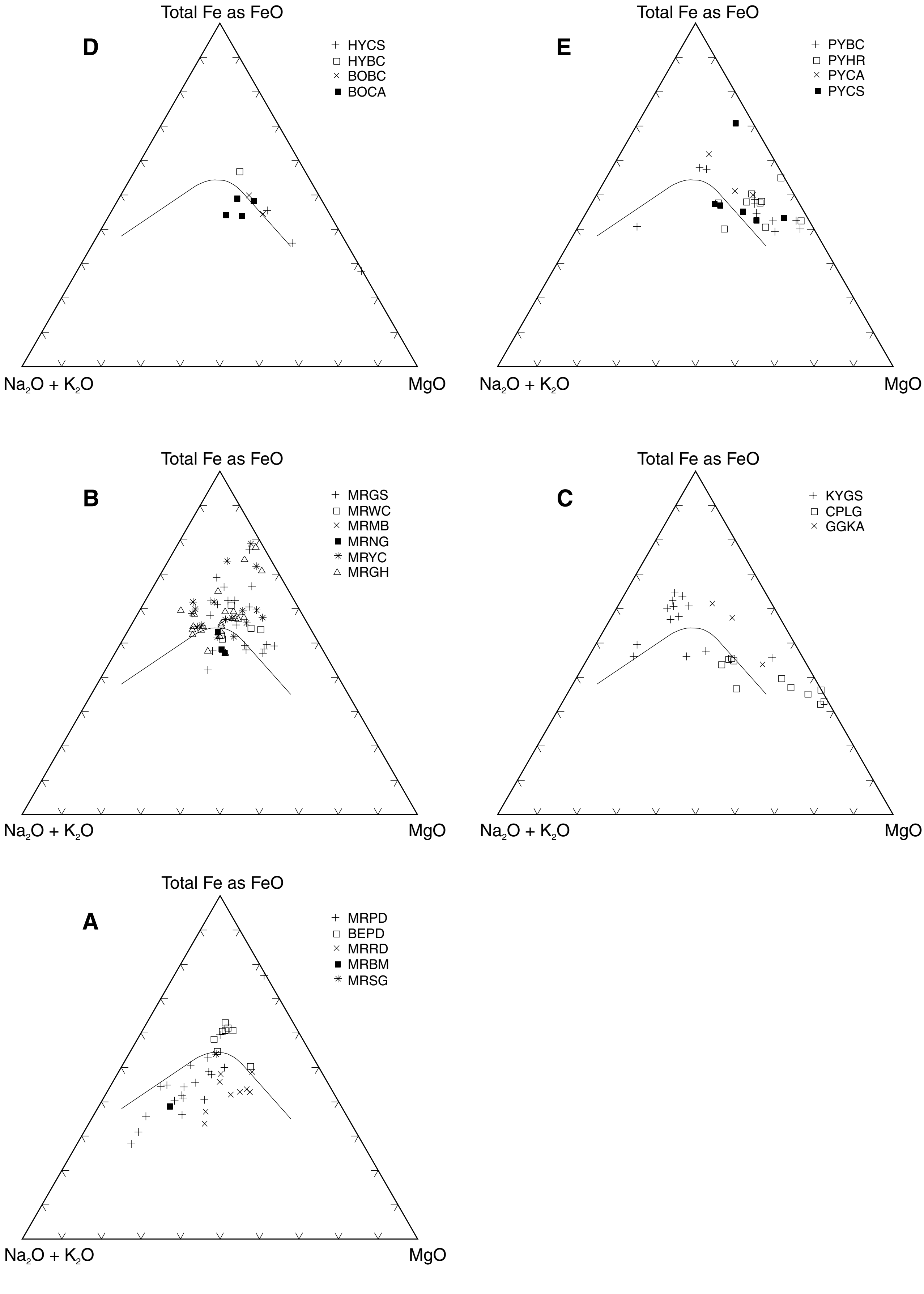


Figure 12.8. (continued)

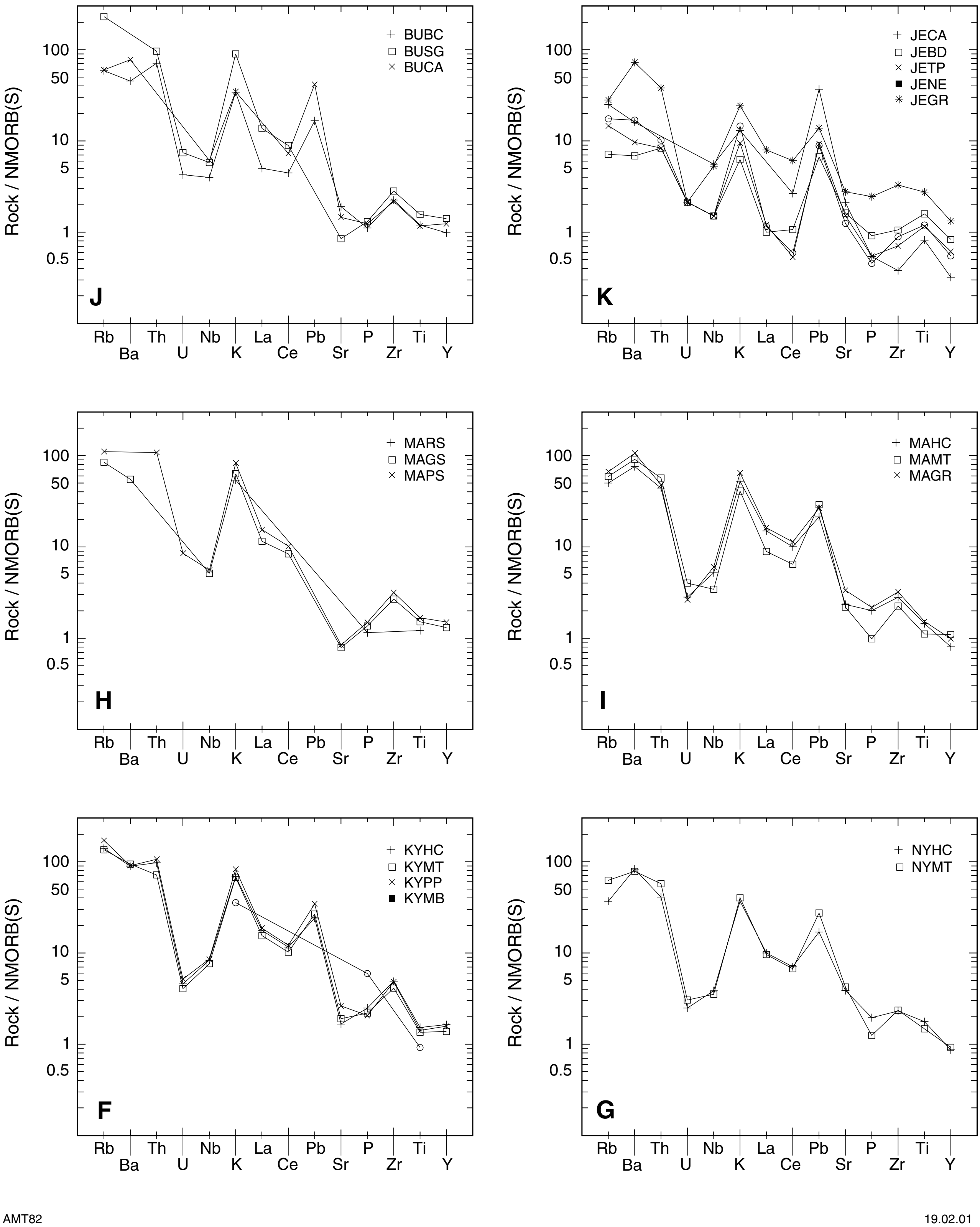


Figure 12.9. Trace and minor element spidergrams for the eleven clusters of mafic and ultramafic rocks, normalized to the MORB values of Sun and McDonough (1989); the logarithmic scale of the y-axis shows enrichment or depletion relative to those values

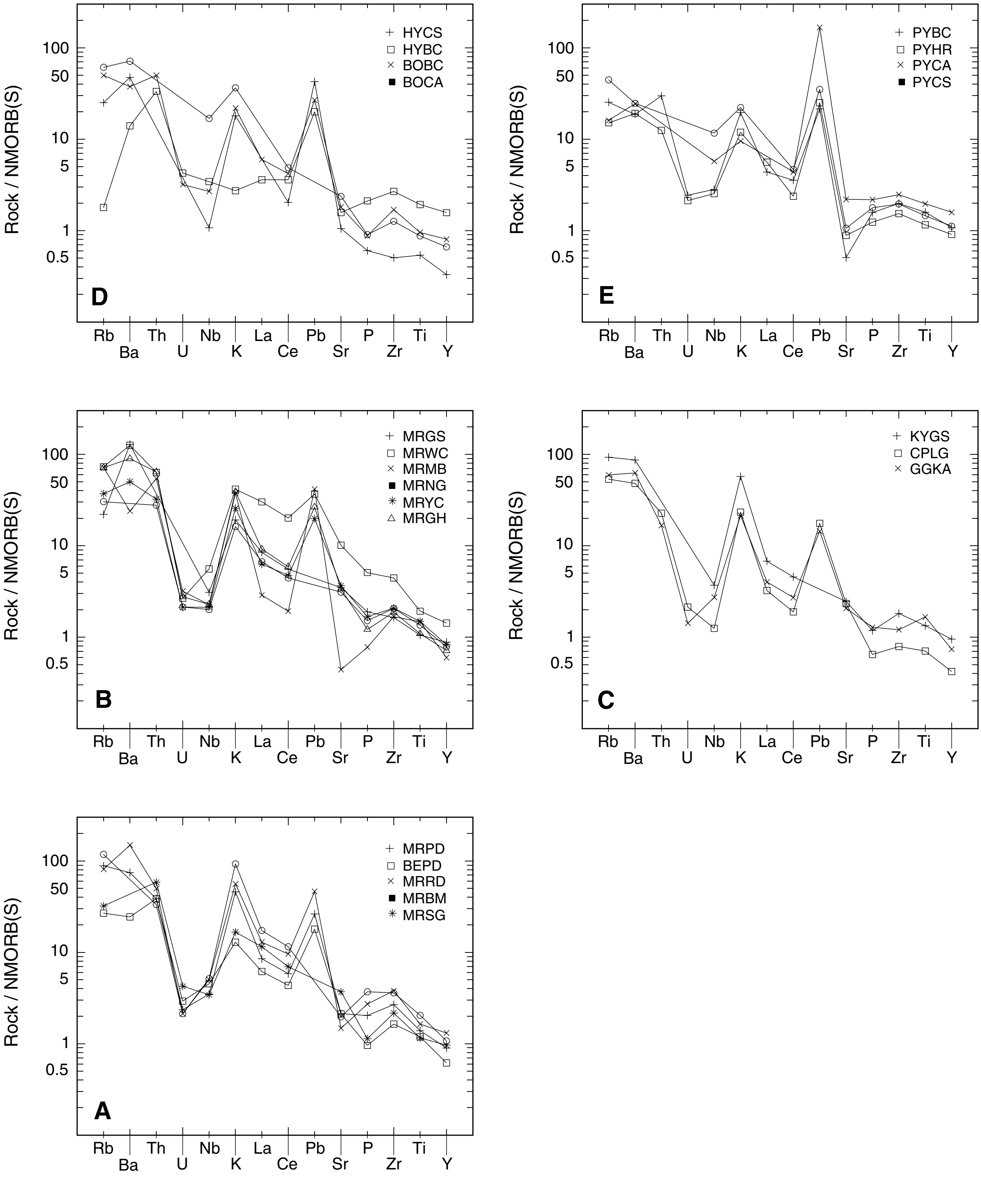


Figure 12.9. (continued)

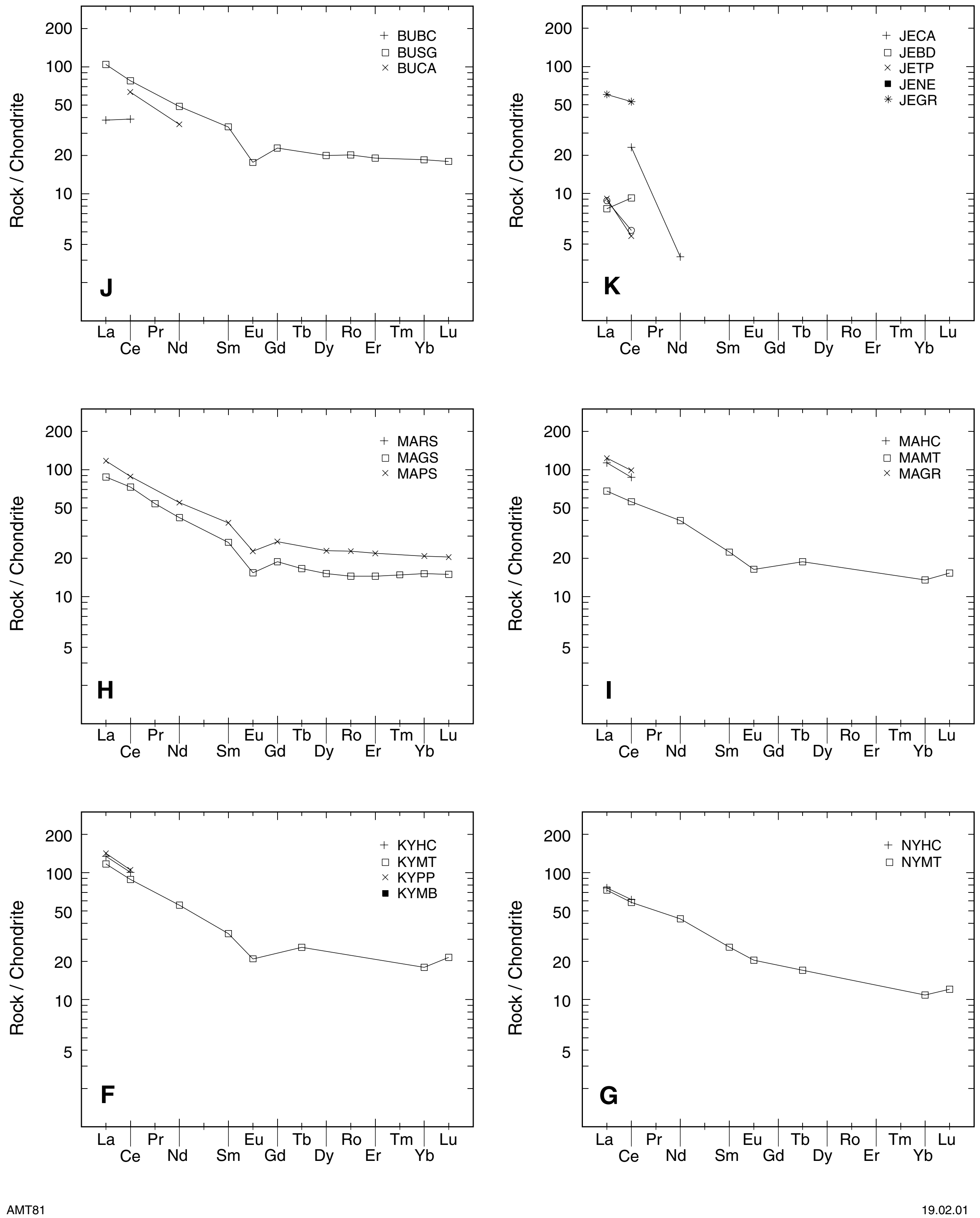


Figure 12.10. Chondrite-normalized REE distribution patterns for the eleven clusters of mafic and ultramafic rocks

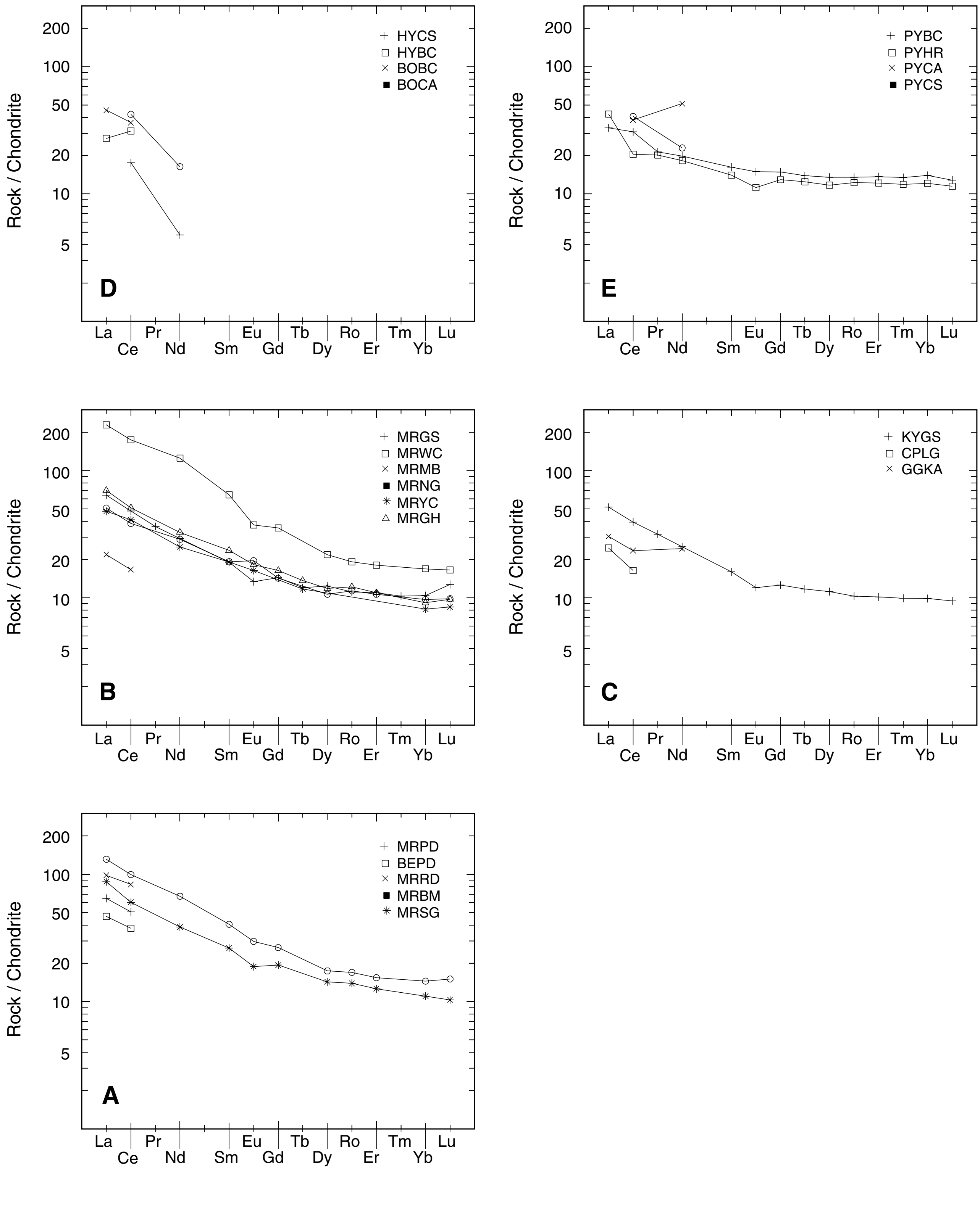


Figure 12.10. (continued)

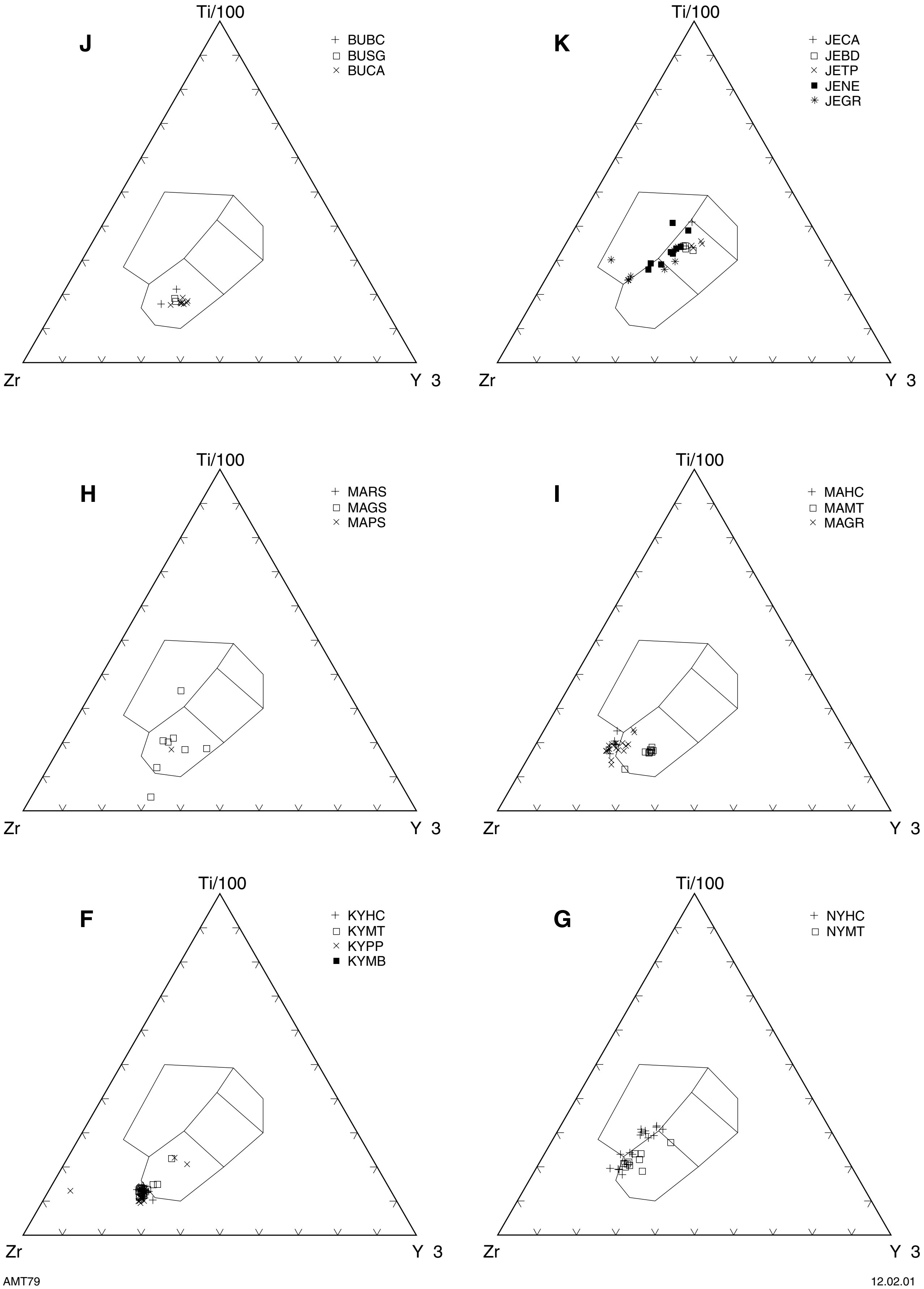
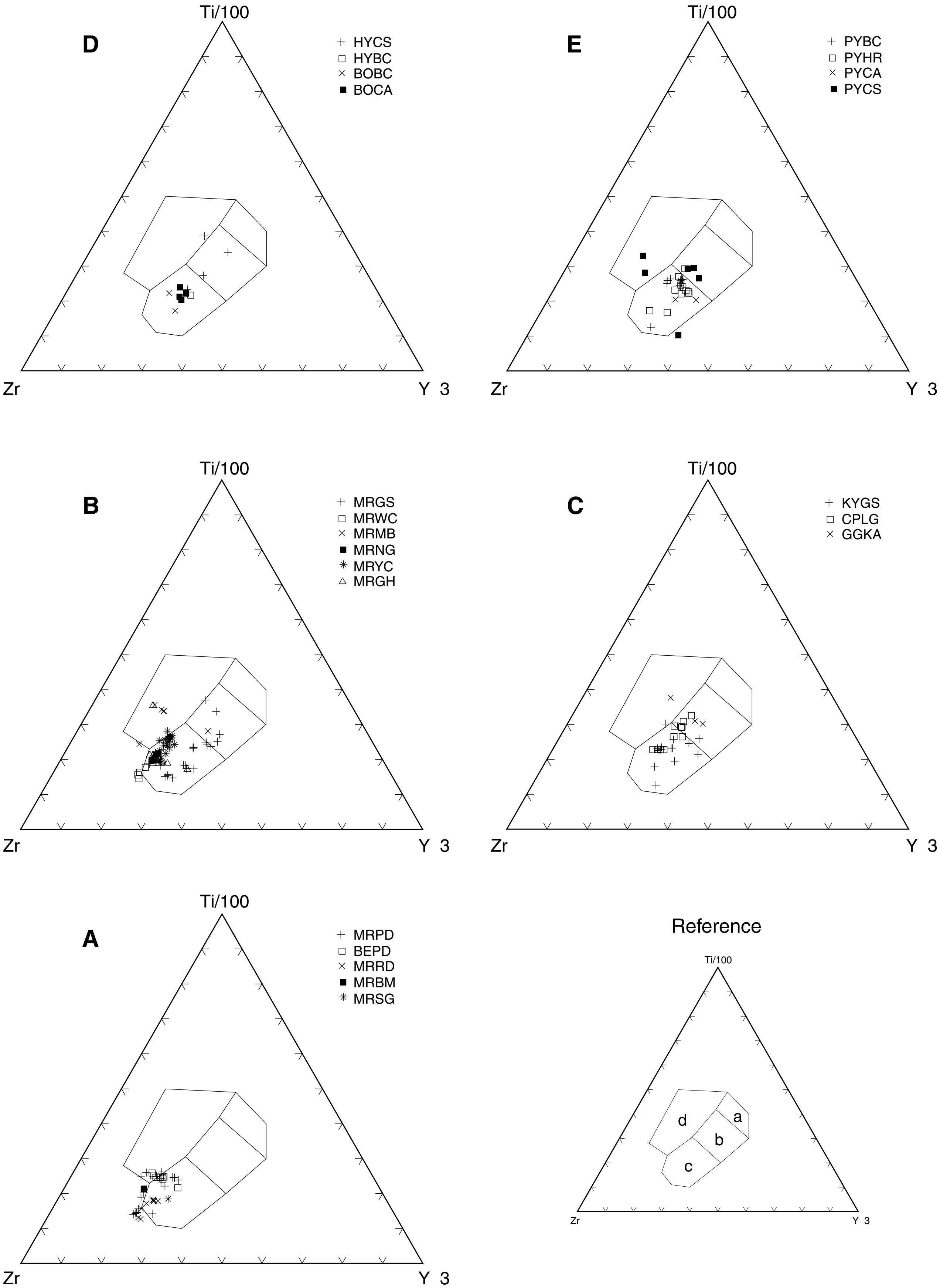


Figure 12.11. Pearce and Cann diagrams for the eleven clusters of mafic and ultramafic rocks. Lower case letters in relevant fields of the reference diagram correspond to the following tectonic interpretations of Pearce and Cann (1973):



a, b: low-K (island arc) tholeiites; b: ocean-floor basalts; b, c: calc-alkali basalts (island arc); d: within-plate basalts (ocean island and continental basalts)

lavas erupted during Fortescue Group volcanism, and historically called basalts, should clearly be called basaltic andesites in terms of current nomenclatural convention. The Jeerinah Formation has sills of basaltic composition, and basaltic komatiites are locally present at restricted stratigraphic levels.

Trace elements: We believe also that the individual trace element data displayed in Figure 12.2, although exhibiting scatter due to variable alteration of individual samples, do serve to indicate the average trace element contents of the primary magmas. We recognize, however, that the more mobile elements, such as the alkalies, have undergone substantial redistribution, so that analyses of individual samples cannot be taken to reflect primary composition. Indeed, an additional argument for this belief lies in the remarkable uniformity of trace element patterns in virtually all clusters. These are plotted individually as spidergrams in Figure 12.9, where the group averages of Table 12.2 are normalized to the MORB values of Sun and McDonough (1989).

As noted later (**Chapter 14**), the general trace element pattern corresponds with that regarded as typical of within-plate continental flood basalt; incompatible elements are substantially enriched relative to MORB. It should be noted that the trace element patterns show generally similar shapes throughout, even in those clusters which are distinctive in terms of major oxide composition. Thus the basalts of cluster K show the same pattern as the basaltic andesites of clusters from the major lava units; however, that cluster is still distinguishable from virtually all others in that the concentrations of all trace elements are uniformly, and proportionally, much lower — above MORB by a factor of between 10 and 30, rather than between 30 and 100. It is interesting to note that the logarithmic ordinates of these spidergrams substantially reduce the distinctive enrichment of cluster F in elements such as Ti and Zr, which marks this cluster out on many other diagrams (e.g. Fig. 12.5 — mg/Zr). The consistently exceptional and heterogeneous clusters C and D are again unusual in showing the low trace element concentrations of cluster K, but the same general pattern is present.

The relative uniformity of trace element signature applies also to REE patterns, which consistently have the typical LREE enrichment and slight or absent Eu anomalies of continental flood basalts (Fig. 12.10). This 'tectonic' characterization of the clusters is not, however, supported by their positions on the triangular Ti–Zr–Y diagram of Pearce and Cann (1973). In terms of these plots, which appear in Figure 12.11, cluster K occupies a different area from all other clusters, but the plot does not appear to provide a useful guide to the tectonic provenance of the Fortescue Group lavas as a whole.

'Chemical stratigraphy': regional and stratigraphic variations of composition

Chemical stratigraphies have been claimed for at least six regionally extensive volcanic provinces with extensive superimposed mafic flows: the Ventersdorp Supergroup (Bowen et al., 1986; Myers et al., 1990), the Parana province (Peate, 1990; Peate et al., 1990), the Karroo province (Marsh and Eales, 1984), the Columbia River Basalt Group (Reidel and Hooper, 1989, and further references therein), the Deccan Traps (Mahoney, 1988, and further references therein), and the Siberian Traps (Lightfoot et al., 1990). Comparative discussion of the precise implications of the expression is outside present scope. In broad terms, to suggest that a volcanic province has chemical stratigraphy implies that individual flows, or sequences of successive flows, have sufficiently distinctive chemical 'signatures' for these to be used for stratigraphic identification throughout the province, independently of other criteria. The term may also be used to denote a secular change in chemical 'signature' with stratigraphic height. None of the references cited above uses the term chemical stratigraphy to refer to regional variations in the chemical composition of lava sequences with broad stratigraphic equivalence, although this phenomenon has been recorded, for example, from the Karroo province (Erlank, 1984, and papers therein). The 376 analyses of mafic and ultramafic rocks of the Fortescue Group available to us are clearly an inadequate sample for the establishment of a chemical stratigraphy of any kind. However, there are some indications of both lateral and vertical chemical variations, at various scales, and these are discussed below.

Lateral (regional) variations: From data already presented and discussed it is clear that within all clusters consisting of nominal basalts there is a generally uniform composition. Only four exceptions to this deserve note:

- Komatiitic lavas are concentrated in the south Pilbara sub-basin. They are concentrated in the Pyradie Formation. To the extent that the only regionally equivalent mafic rocks in the northern areas are in the Kylena Formation there is a strong regional bias in the occurrence of these rocks.
- There also appears to be a slight north–south contrast at the level of the Mount Roe Basalt. This is expressed in such ways as a lower alkali content in the northern rocks (Fig. 12.6) as well as a slightly lower alumina content relative to iron and magnesia (Fig. 12.7).
- Within the northern Mount Roe Basalt (cluster B) the MRGS group is quite distinctive, in particular with its high SiO_2. This is also shown by KYGS and MAGS, and laboratory bias might be a tempting explanation for this regionally high SiO_2 if it were not clearly indicated also by the independently analysed MAPS group.
- The F cluster, including northeastern areas of the Kylena Formation, has been shown to be chemically unique in a number of respects, all basically reflecting a high concentration of incompatible elements. Meakins (1990) has also noted this distinctive regional composition.

Vertical (stratigraphic) variations: In terms of vertical chemical variation, the clusters consisting of nominal basalts, as in the case of lateral variation, do not show any clear compositional changes with stratigraphic height within the Fortescue Group. The erupted lavas consisted largely of basaltic andesite of rather consistent composition throughout, and the only trend that might be suggested from the available data is that the mafic lavas

Table 12.3. Chemical compositions of five felsic volcanic rocks from the Hardey Formation

Sample no.	*94759*	*38018*	*38023*	*94760*	*94761*
			Percentage		
SiO_2	74.80	70.10	70.70	78.80	73.70
TiO_2	0.41	0.41	0.44	0.25	0.47
Al_2O_3	11.00	12.10	12.40	11.20	13.10
Fe_2O_3	1.75	2.80	2.30	1.50	0.97
FeO	2.30	2.42	2.23	0.91	1.11
MnO	0.08	0.08	0.10	0.04	0.05
MgO	0.67	0.40	0.30	0.15	0.51
CaO	0.13	1.53	0.93	0.06	0.64
Na_2O	0.08	3.40	3.45	1.41	2.67
K_2O	7.04	5.10	5.90	4.40	4.82
P_2O_5	0.11	0.12	0.11	0.02	0.12
H_2O+	–	0.49	0.35	–	–
$H_2O–$	–	0.17	0.15	–	–
CO_2	–	1.13	0.58	–	–
LOI	1.57	–	–	1.55	2.14
Rest	0.20	–	–	0.15	0.14
Total	**100.14**	**100.25**	**100.94**	**100.44**	**100.44**
			Trace elements (parts per million)		
Ba	–	1 090	1 410	–	–
Ce	210	–	–	210	159
Co	59	–	–	55	50
Cr	<4	–	–	<4	<4
Cu	<4	–	–	<4	<4
F	–	140	350	–	–
La	110	–	–	109	87
Li	–	10	10	–	–
Rb	267	125	180	161	167
Nb	30	–	–	23	13
Nd	90	–	–	84	67
Ni	12	–	–	17	15
Sc	5	–	–	3	5
Sn	–	5	5	–	–
Sr	15	40	50	41	68
Th	23	–	–	21	19
U	5	1	1	5	4
V	9	–	–	3	28
Y	80	–	–	55	47
Zn	53	–	–	67	27
Zr	632	720	760	345	331

NOTE: Sample 94759 is a rhyolite lava from the Koongaling Volcanic Member (PEARANA AMG 250930). Sample 94760 is a rhyolite porphyry from a sill close to the top of the Hardey Formation (PEARANA AMG 167802). Sample 94761 is a rhyolite porphyry from the Bamboo Creek Member at Green Hole (YARRIE AMG 182820). Samples 38018 and 39023 are from the Koongaling Volcanic Member (MUCCAN AMG 054699 and 025699)

of the opening phase (Mount Roe Basalt) had slightly greater compositional diversity.

The only indication of systematic vertical variation within the total lava sequence has been already noted, since it is also expressed as a lateral difference; this is the characteristic occurrence of magnesian rocks in the central part of the succession in the south Pilbara sub-basin, between the 'standard' basaltic andesites of the Mount Roe Basalt and the Bunjinah Formation.

Near the close of Fortescue Group time, the sills of the Jeerinah Formation demonstrate availability of a distinctive and different magma type, but analyses at this stratigraphic level are locally restricted, and until more are available it is not known whether this represents a 'chemical stratigraphic' change.

Figure 12.3 has already demonstrated a systematic chemical variation with stratigraphic height in the Kylena Formation on a local scale. Meakins (1990, p. 39) has referred to chemical trends within the Maddina Formation in drillcores, which may be on a similar scale, but the data on which these are based, and their nature, have not been published.

Felsic rocks

At the time of writing, only five analyses of felsic rocks from the Fortescue Group are available, three rhyolitic lavas from the Koongaling Volcanic Member, a rhyolite porphyry from a sill close to the top of the Hardey Formation, and a rhyolite porphyry from the Bamboo Creek Member. All five analyses have been published previously (Hickman et al., 1983; Nelson, D. R. et al., 1992). All these rocks (Table 12.3) have typical compositions of rhyolites, and show trace element (including REE where available) patterns comparable with those of the mafic lava clusters (Fig. 12.12).

Isotope systematics and petrogenesis of the igneous rocks

Current general models for the origin of mafic and ultramafic volcanic rocks have a consensus view that the primary magmas originate by partial melting of the mantle (or lower crust), and that, possibly after storage in a magma chamber, they reach the surface through fissures or vents. The chemical composition of the magma finally erupted will depend mainly on:

- The composition of the source region;
- The degree of partial melting;
- The length of time which the primary magma spends in an intermediate magma chamber, during which its composition may be affected by:

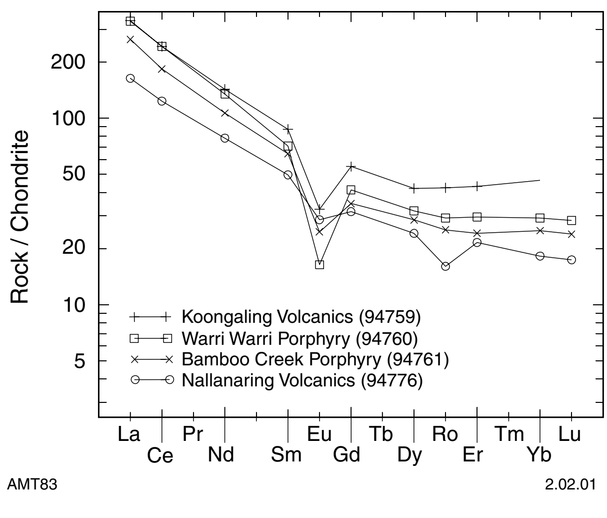


Figure 12.12. Chondrite-normalized REE distribution patterns and MORB-normalized spidergrams for four felsic rocks

  - # differentiation by settling of early-formed crystals,
  - # contamination by assimilation of wall material,
  - # magma mixing;
- The possible interactions of this complex set of processes is such that for most extensive volcanic provinces of the Earth the relative influence of each remains controversial. In seeking to determine this it is necessary to consider the total evidence, and petrogenetic models for Fortescue Groups volcanism need to take the isotope systematics into account, equally with the geochemistry. The only published model to do this is that of Nelson, D. R. et al. (1992) and the following summary is essentially based on that paper.

Nelson, D. R. et al. (1992) reviewed new and published Rb–Sr, Pb–Pb and Sm–Nd isotopic data. The first two systems indicate substantial disturbance at ages between about 2.0 and 2.4 Ga, while the Nd isotopic data for Fortescue Group lavas show ξNd values between -1.5 and -4.4. Nelson, D. R. et al. (1992) concluded that these values probably do not result from crustal assimilation within crustal-level magma chambers; they believe instead that their results are consistent with either extensive crustal contamination of primitive (komatiitic or picritic) Fortescue Group magmas prior to their differentiation, or derivation from negative ξNd (or enriched) mantle sources.

Clastic and pyroclastic rocks

Analyses of seventeen sedimentary and pyroclastic rocks of the Fortescue Group are included within our geochemical database. They include an interflow sandstone from the Mount Roe Basalt, coarse tuffs from the Lyre Creek Member of the Hardey Formation, and a variety of clastic rocks from the Hardey, Boongal, Pyradie, and Bunjinah Formations, principally from the analyses of Cowley (1979) listed under an earlier heading of this chapter.

None of these analyses of clastic and pyroclastic rocks yields data of sufficient signifiance in Fortescue Group interpretation to warrant further discussion here.

Chapter 13

Economic geology

For the purposes of administration and statistical records the Department of Minerals and Energy divides the outcrop area of the Fortescue Group between four mineral fields: Ashburton, Marble Bar, West Pilbara, and Peak Hill. The location of the principal mineral occurrences described in the text are shown on Plates 1a and 1b. Only those occurrences that are associated with the Fortescue Group are discussed in this chapter.

Gold

Gold was first recorded from the Fortescue Group in 1890 (Woodward, 1891). The deposits are classified into three principal types: auriferous pyritic conglomerates, Palaeoproterozoic (post-Wyloo Group) quartz veins, and Cainozoic alluvial–colluvial deposits which are derived from the Fortescue Group. With few exceptions, the deposits are small and production data are unreliable. The larger workings are at Beatons Creek, near Nullagine, and at Just-in-Time, near Marble Bar. In both areas, gold production historically has been grouped with output from other types of deposit, but available records and the size of the workings suggest that production was less than 100 kg from individual deposits associated with the Fortescue Group.

Auriferous pyritic conglomerates

Gold has been mined from palaeoplacer deposits in the lower Fortescue Group near Nullagine and in the Marble Bar sub-basin.

Nullagine district

About 2 km northwest of Nullagine, the Mosquito Creek Formation is unconformably overlain by the conglomeratic lower part of the Hardey Formation. Within the lower 30 m of this conglomerate are several 0.5–4 m-thick pyritiferous and auriferous units (Maitland, 1905; Finucane, 1935; Noldart and Wyatt, 1962; Hickman, 1983; Hickman and Harrison, 1986). Gold is typically fine grained and forms small flakes and rounded particles in the matrix. Lithic fragments suggest derivation from the Mosquito Creek Formation and possibly from the Kurrana Batholith (Hickman, 1983).

Descriptions of the workings by Hickman (1983) indicate that mining has been concentrated in areas where hematite, limonite, and jarosite predominate over pyrite. The Hardey Formation dips gently west at Grants Hill, but farther west its attitude is reversed across the axis of a southwesterly plunging syncline. This syncline was recognized by Finucane (1935), who noted a continuation of gold mineralization into the western limb.

Results from recent company exploration programs in the Nullagine area are summarized by Carter and Gee (1988). Highest reported values were 0.5 ppm Au over 0.1 m.

Marble Bar sub-basin

Beds of auriferous conglomerate up to 2 m thick at the base of the Mount Roe Basalt have been worked at the Just-in-Time and Tassy Queen mines, 9 km south southwest of Marble Bar (Finucane, 1938a; Noldart and Wyatt, 1962).

Hickman and Gibson (1982) describe an auriferous conglomerate at the base of the Mount Roe Basalt 10 km east of Cooke Bluff Hill. This conglomerate consists of well-rounded pebbles and cobbles of chert, quartzite, and gossan, set in a coarse-grained ferruginous matrix. Two specimens of the matrix assayed 2.9 ppm and 9 ppm gold, a specimen of conglomerate assayed 2.6 ppm gold, and a gossan pebble 3.7 ppm gold. Two gossan pebbles and a second sample of conglomerate assayed 0.2 ppm or less.

Noldart and Wyatt (1962, p. 136) refer to minor gold production from auriferous conglomerate in the Apex–Salgash area, but the exact location of these workings has not been recorded.

Carter and Gee (1988) summarize the results of exploration in the Hardey Formation of the Marble Bar sub-basin. A conglomerate near Shady Camp Well, in the centre of the sub-basin (MARBLE BAR AMG 665572) yielded 3.85 ppm gold. However, follow-up close-spaced rock-chip sampling at 25 m intervals recorded few gold values greater than 1 ppm and, in a majority of samples, gold was either very low or not detected. A diamond drillhole in this bed recorded a maximum gold value of only 0.6 ppm.

Ten kilometres to the northeast of the Shady Camp Well area, two coarsening-upward sequences of shale, sandstone, and conglomerate have been explored. An exploratory shaft (6.1 m deep) exposed a metre of flat-lying conglomerate immediately below the central basalt of the Hardey Formation. A best value of 0.07 ppm gold was returned from analyses of shaft samples (Carter and Gee, 1988).

Post-Fortescue Group quartz veins

Minor quantities of gold have been reported from Capricorn Orogen veins and shears in Fortescue Group rocks in the south Pilbara sub-basin. Principal occurrences are the Paulsen and Belvedere Groups on WYLOO, and a small prospect near Paraburdoo on TUREE CREEK.

Paulsen (Melrose Group)

The Paulsen mine is located 24 km southeast of Mount Stuart Homestead; descriptions of the workings have been given by Forman (1938), Finucane (1939), Marston (1979), and Blight (1985).

Auriferous quartz veins strike approximately 310°, and dip northeast at 20–30°. They cut across basalt and sandstone, close to the contact between the Hardey Formation and the Boongal Formation, and can be traced on the surface for approximately 60 m. Finucane (1939) noted that the deposit shows evidence of secondary enrichment; there are lower average grades and significant amounts of pyrite below a depth of about 17 m. Small amounts of gold were reported from other northwesterly trending quartz veins, north, east, and south of the main prospect (Finucane, 1939).

Belvedere Group

The Belvedere Group, located 6.5 km southeast of the Paulsen mine, has been described by Forman (1938), Finucane (1939), Blockley (1971), Marston (1979), and Blight (1985). Gold production to the end of 1986 amounted to 14.4 kg.

Gold, copper, lead, and silver mineralization occurs in two sets of quartz veins. One, the No. 1 reef of Finucane (1939), strikes north and dips west at approximately 45°; the other, the No. 2 reef, strikes 215° and dips northwest at 75–85°. The No. 2 reef parallels a northeast-trending fault that displaces the No. 1 reef 3–4 m to the northeast. In addition, there are a number of barren veins striking at 310°. Country rocks include mafic and felsic volcanic rock, and sandstone of the Mount Roe Basalt. Lead-isotope ratios reported by Richards et al. (1981) indicate that mineralization may be late Proterozoic, but the data are poor and this conclusion is equivocal.

Paraburdoo

A small gold prospect within the Bellary Formation lies approximately 4 km north of Paraburdoo (PARABURDOO AMG 668375). Gold values generally ranging from 0.2–0.7 g/t (22 g/t maximum) have been reported from a sheared conglomerate in the upper part of the formation (Zeelanberg, 1976).

Cainozoic alluvial–colluvial deposits

Small amounts of gold have been obtained from Cainozoic alluvium and colluvium close to many of the previously described gold occurrences, and also near deposits hosted by the granite–greenstone rocks. Some of the latter were described by Hickman (1983). One of the most important areas is near Nullagine, where creeks draining auriferous Hardey Formation conglomerate have probably yielded over 150 kg of gold (Finucane, 1935).

Uranium

Although no uranium deposits of economic grade have been discovered in the Fortescue Group, these rocks are known to be uraniferous at several localities (Robertson, 1974; Hickman, 1983; Carter and Gee, 1988). Most of the occurrences are associated with pyritic conglomerate and sandstone in the lower part of the Fortescue Group. These have been reviewed by Carter and Gee (1988) and the following account is summarized from that work.

Exploration history

It was not until 1955 that the possibility of uranium occurring in Fortescue Group conglomerates was considered. In that year, Enterprise Exploration Pty Ltd conducted a brief survey based on the similarity of these conglomerates to the uranium-bearing conglomerates of Blind River, Canada. Ground scintillometer traverses were made but no anomalous radioactivity was discovered.

In the same year, the Australian Bureau of Mineral Resources flew an airborne reconnaissance scintillograph survey of part of the east Pilbara region.

No further exploration activity is known until 1968, when large exploration operations began. The general approach was to employ airborne radiometric surveys to locate radioactive outcrops and to follow up with ground radiometric surveys, geological mapping, and drilling. By 1982, when operations had largely ceased, no economic occurrence of uranium had been identified despite a number of good drillhole intersections.

Mineral explorers obtained very high uranium and gold values from analyses of surface rock-chip samples. Values of uranium in excess of 4000 ppm U_3O_8 are mentioned in reports, but analyses of drillcore from corresponding horizons reveal uranium values generally below 500 ppm U_3O_8. Surface enrichment of uranium is readily explained. The large pyrite content of Fortescue Group conglomerates and sandstone, when oxidized, produces acid weathering conditions in which limonite is ultimately formed. Uranium is mobilized during hydrolytic reactions accompanying the formation of limonite, and the metal is concentrated by coprecipitation with, and absorption by, hydrous iron oxides. Most gold values above 1 ppm probably result from near-surface enrichment.

A weakness in some exploration work is the failure to determine the mode of occurrence of uranium responsible for high uranium values. A number of occurrences are known to have been derived in part from heavy mineral resistates such as monazite, xenotime, and zircon minerals commonly found in Fortescue Group sedimentary rocks.

Exploration areas

Most exploration for palaeoplacer deposits of uranium has been conducted in the Nullagine Synclinorium of the northeast Pilbara sub-basin, and in the Marble Bar and Northwest Pilbara sub-basins. Here, the cutoff value employed in the selection of uranium intersections (with one or two exceptions) is 250 ppm U_3O_8.

Northeast Pilbara sub-basin (Nullagine Synclinorium)

Exploration: Radioactive conglomerate in the Hardey Formation was explored for uranium in the southern part of the synclinorium. Here, facies variations appear to reflect the presence of a structural high, which subdivides the basin along its long axis. From west to east across the structural high, pyritic, oligomictic conglomerate passes into polymictic conglomerate with diminishing pyrite content; argillaceous sediments become more abundant, with greater thicknesses of basal conglomerate. Drilling near Nullagine has proved more than 700 m of conglomerate.

Most conglomerates contain fragments of vein quartz, chert, basalt, andesite, granitoid, quartz sandstone, and siltstone. The matrix consists of poorly sorted quartz and feldspar, and a high proportion of sericite; in some cases up to 60% carbonaceous material is recorded. Pyrite and thucholite occur as disseminations and as concentrations around quartz grains. Buckshot pyrite, of detrital origin, ranges in size from 0.2–2 mm in diameter and typically has abraded and pitted surfaces. Thucholite pellets are less abundant and usually smaller than pyrite grains.

Feldspathic sandstones are the predominant lithology in the southwest of the synclinorium. Most contain pyrite and thucholite in thin, discontinuous layers on bedding surfaces and foresets, whereas carbonaceous material occurs as thin stringers and specks in the matrix. Many sandstones yield enhanced uranium values and an intersection of 305 ppm U_3O_8 over 0.6 m has been reported.

Siltstone and mudstone occurs as thin partings and as thick beds; uranium values of up to 80 ppm U_3O_8 have been recorded from some layers

Mode of occurrence of uranium: In most analyses from drillcore, thorium values are low, indicating that resistate minerals such as monazite do not contribute significantly to the uranium content. Although high uranium values such as 1085 ppm U_3O_8 were obtained by drilling, no detrital uraninite grains have been identified. Cominco Exploration Pty Ltd (quoted in Carter and Gee, 1988) recorded that uranium occurs as fine-grained uraninite inclusions in carbonaceous pellets within the conglomerate matrix. Petrographical and mineralogical work reported by Marathon Petroleum Australia Ltd (quoted in Carter and Gee, 1988) records thucholite pellets that contain line inclusions, mainly of galena and thorite ($ThSiO_4$). The thorite contains minor uranium, almost certainly as partial substitution of Th rather than as a specific phase. Persistent detrital accessory grains of cellular anatase carry minute grains of a uraniferous phase, almost certainly brannerite.

Anatase, also a common detrital accessory, has an average size of 0.2 mm and an average abundance of 1–2% (Carter and Gee, 1988). Anatase grains contain small (1–3 µm) inclusions of a uraniferous phase, probably brannerite but possibly uraninite. Marathon also reported that there is a direct correlation between the abundance of U_3O_8 and Pb in carbon (thucholite) pellets. Also, these pellets appear to be more prevalent in samples that also contain more abundant pyrite and quartzose pebbles. With regard to grains of detrital zircon and cellular anatase, Marathon reported that these also appear to be more abundant in samples containing most pyrite and thucholite pellets.

Marble Bar sub-basin

Exploration in the Marble Bar sub-basin has focused upon siliciclastic rocks in the lower Mount Roe Basalt and Hardey Formation. Results from the former are low; those from the latter have encouraged some investigation, with most attention having been focused on arenaceous units below and above a middle basalt division. This basalt, which was previously assigned to the Kylena Basalt (Hickman, 1983; Carter and Gee, 1988), is here considered to be part of the Hardey Formation (see Blake, 1984a; **Chapter 5** this volume)

Hardey Formation: A diamond drillhole in a quartz-pebble conglomerate near Shady Camp Well (MARBLE BAR AMG 665572) intersected 356 ppm U_3O_8 over 0.3 m. The mode of occurrence of uranium in this drillcore is not known but surface samples of conglomerate returning high uranium analyses are accompanied by high thorium and rare earth element values, indicating that some uranium is bound up in resistate minerals such as monazite and xenotime.

Ten kilometres to the northeast of the Shady Camp Well area, two coarsening-upward sequences of shale, sandstone, and conglomerate have been explored. An exploratory shaft (6.1 m deep) exposed a metre of flat-lying conglomerate immediately below the middle basalt unit, and bottomed in sandstone. A best value of 33 ppm U_3O_8 was returned from analyses of shaft samples.

Carter and Gee (1988) recorded that green arkosic sandstone from the Shady Camp Well locality showed finely disseminated grains (2–3 µm) of an unidentified uranium mineral associated with phosphorus and occurring both in lead-bearing graphite grains and along narrow fractures containing pyrite. These graphite grains are probably metamorphosed thucholite particles and thus, as in the Nullagine Synclinorium, uranium is likely to occur at least in part as fine uraninite inclusions in carbonaceous pellets.

Arenaceous and conglomeratic units in the upper Hardey Formation near Limestone Well were tested for palaeoplacers during three exploration programs, two of which included diamond drilling (Carter and Gee, 1988). High uranium intersections were made initially by Alcoa of Australia but were not reproduced during subsequent analyses when values as low as 10–50% of the originals were obtained. International Nickel Australia Ltd, however, obtained 613 ppm U_3O_8 over 0.25 m in a drillhole sited close to the Alcoa hole that had yielded a value of 1300 ppm U_3O_8 over 2.75 m. It is possible therefore that Alcoa's high values may have substance, but these clearly need to be treated with caution.

Summarizing other occurrences, Carter and Gee (1988) noted that detrital pyrite is present in many sandstones, although thucholite has not been identified. The best (but qualified) uranium intersection of 1510 ppm U_3O_8 over 0.3 m was made in pyritic pebbly sandstone. Conglomerate, which makes up about 5% of the section, is uraniferous near the base of the upper Hardey Formation

Mode of occurrence of uranium: International Nickel Australia Ltd obtained 613 ppm U_3O_8, and 831 ppm Th, over 0.25 m from core taken from the lower part of the upper Hardey Formation (Carter and Gee, 1988). Bands of heavy minerals within the core consist of detrital pyrite and small amounts of molybdenite, galena, chalcopyrite, zircon, monazite, xenotime, tourmaline, and composite radioactive grains. The monazite contains significant amounts of thorium, and the xenotime carries thorium and a small amount of uranium. The major sources of radioactivity are in small (up to 100 µm) composite grains of two types. One type consists of intergrowths of illite, uraninite, and galena; the second type consists of an intergrowth of illite, coffinite, uraninite, and galena. Titanium is abundant in both types and may be present as brannerite. Uraninite is 'earthy' but shows remnant cleavage traces indicating an original crystalline nature. Other radioactive sources, not positively identified, include uranium- and thorium-rich biotite as inclusions in quartz grains, and uranium-rich coatings on zircon (Carter and Gee, 1988).

This mineralogical composition points to a primary source of either an uranium-enriched granite, or a pegmatite that contained intergrowths of uraninite and feldspar. Coffinite may either have formed by alteration of uraninite, or be a primary mineral.

Northwest Pilbara sub-basin

Exploration for uranium and gold has been undertaken at several localities on PYRAMID. Exposures of Hardey Formation near Coorbeelie River were the principal scene of exploration, which included three separate drilling programs (Carter and Gee, 1988). The formation outcrops over an area of about 10 km^2, but only 120 m of conglomeratic sandstone is present. Coarse feldspathic sandstone and conglomerate are largely similar in composition and pyrite content to those in the Nullagine Synclinorium; they also consist of fluvial sedimentary facies. There are three conglomeratic intervals; within these, individual beds range up to 3 m in thickness. Abundant detrital pyrite, locally 10% of the total matrix, is found in thin layers. Pale-green sericite, after feldspar, constitutes 5% of some rocks. Carbonaceous grains presumed to be thucholite have been noted in the middle one of the three conglomerate intervals.

The principal uranium intersections were made in conglomerate; the best was 684 ppm U_3O_8 over 0.1 m. Drilling has demonstrated that down-dip extensions of the uraniferous conglomerates are extensively disrupted by intrusions of dolerite (Carter and Gee, 1988).

No known mineralogical investigation of the mode of occurrence of uranium at Coorbeelie River has been made. Marathon Petroleum Australia Ltd reported what was considered to be thucholite in the middle conglomerate interval, with disseminated pyrite and clay galls — an association commonly seen in Nullagine conglomerates (Carter and Gee, 1988). It is most likely, therefore, that part of the uranium occurs as fine inclusions of uraninite in thucholite, or as partial substitutions of thorium in thorite.

South Pilbara sub-basin

Blight (1985) investigated slightly radioactive, fluvial arkosic sandstone in the Paraburdoo area. The best assay is 25 ppm U_3O_8 and 370 ppm Th and he concluded that the anomalous radioactivity was due to thorium and not uranium.

Discussion

Carter and Gee (1988) commented that concentrations of uranium within the Hardey Formation are high when considered beside the largely unpromising source rocks now exposed in the Pilbara Craton which, apart from the 'tin granites' are rather uranium poor. Evidently, some granitic sources were exposed because uraninite travelled within granitic fragments to deposit in palaeoplacers, along with pyrite and other heavy minerals. No uraninite has been identified in the 'tin granites', and illite–uraninite grains in the lower Fortescue Group may have eroded from uranium-enriched upper levels of intrusions such as those of Moolyella and Spear Hill.

Most uranium in the Hardey Formation is found in detrital thucholite pellets as fine uraninite inclusions within thorite grains carrying minor uranium, and as detrital anatase containing brannerite (Carter and Gee, 1988). This suite suggests mineralization of a modified placer, with uranium moving as very fine particles, and possibly also in solution. Microbial colonies existed locally during the deposition of the lower Fortescue Group and could have concentrated uranium by both mechanical filtration of particles and biochemical precipitation of solutes. Subsequent erosional destruction of the colonies would have produced thucholite pellets. Brannerite in anatase could form as a result of replacement of titanium by uranium from solution (Carter and Gee, 1988).

Fluorite

Significant quantities of fluorite were discovered within the Fortescue Group of the Meentheena area (MOUNT EDGAR AMG 390480) in 1971. About 8 kt of fluorite was extracted, although no production has been officially recorded (Hickman, 1983). Total reserves for the larger veins were calculated at 13 000 t per vertical metre of 50% CaF_2 ore; total reserves for the entire Meentheena area are thought to exceed 30 000 t per vertical metre, but most of this material is of low grade (Hickman, 1974).

The most detailed description of the Meentheena fluorite occurrences is that given by Hickman (1974) and the following summary is based largely upon that work.

Geological setting

The fluorite deposits occur as fissure veins within folded and faulted Mount Roe Basalt and lower Hardey Formation rocks. No large igneous intrusions are exposed close to the prospects; however, quartz veining is extensive and several narrow rhyolitic dykes intrude the area. West of the Nullagine River, host rocks mostly belong to the basal Hardey Formation and comprise massive to parallel-stratified conglomerate and pebbly sandstone (the agglomerate, lapilli tuff, and gritty tuff of Hickman, 1974). Clasts consist of well-rounded to angular fragments of basalt, schist, granite, vein quartz, quartzite, and chert; most are up to 0.1 m in diameter. East of the Nullagine River, host rocks comprise vesicular subaerial flows of the Mount Roe Basalt. The lowest flows are generally porphyritic.

The fluorite deposits are situated on the faulted culmination of a broadly domal structure. This structural high is truncated sharply to the west of the prospects by the Meentheena Fault, a north-trending dextral fracture, with a horizontal displacement of about 4 km. Vertical movement (west-block-down) is probably in the order of a few hundred metres. There is a conjugate system of minor fractures between the larger faults, and these fractures are most abundant in the area of the prospects, where they are mineralized by fluorite and quartz. Slickensides, zones of crush-breccia, and sheared quartz veins are also associated with the fissures.

Deposits

The distribution of the main fluorite-bearing veins is described by Hickman (1974), who recorded two sets of veins striking 060° and 135°. A typical vein consists of an outer zone of quartz and a wide central zone of interlayered quartz and massive, coarse-grained, anhedral fluorite. Locally, the fluorite displays banding parallel to the vein walls.

Fluorite bodies within individual veins are rarely tabular but outcrop as pinch-and-swell structures. The walls of the fissure veins are generally composed of silicified and brecciated country rock. This breccia contains angular fragments of quartz, chert, and bleached country rock in a matrix of white fluorite. Purple fluorite appears to have been introduced at a late stage. Fragmentation of the early quartz and fluorite reflects some disturbance and collapse during mineralization. More than one generation of quartz is present; some of the veins contain sheared (platy) quartz, intruded by undeformed quartz and fluorite (Hickman, 1974).

Thick veins of calcite are present on the periphery of the mineralized area, but no barite, a mineral commonly associated with fluorite, was observed (Hickman, 1974). Small occurrences of galena, malachite, and goethite have been noted east of the Nullagine River; atacamite and brochantite are also present. Veins of rhyolite breccia containing fluorite have been noted within the prospect area although most contain no fluorite.

Origin

Hickman (1974) interprets the Meentheena fluorite occurrences as an epigenetic fracture-fill deposit and suggests two possible sources:

- A deep-seated alkalic magma beneath the Meentheena area.
- A fluorite-rich basement granitoid.

The mineralogy of the deposits, particularly the absence of tin, tungsten, and molybdenum, suggests that they are epithermal (Hickman, 1974). In addition, Hickman (1974) noted that there is little evidence to suggest the area is underlain by a large late Archean to Proterozoic alkalic intrusion. On the other hand, a fluorite-rich basement granitoid occurs 33 km south of Meentheena, at Cookes Creek. This rock contains about 2% accessory fluorite and is extensively veined by fluorite, barite, and quartz. A similar granitoid beneath the Meentheena area may have acted as the source for the prospect fluorite (Hickman, 1974).

Copper

Cupriferous mineralization within the Fortescue Group is sparse and most occurrences are reported from the southern Hamersley Basin. The only recorded production is from Capricorn Orogen quartz veins cutting the lower to middle Fortescue Group in the Wyloo Dome, and from the Jeerinah Formation in the Wonmunna area. Other minor occurrences have also been recorded from the Wyloo Dome (Marston, 1979; Blight, 1985), and Marston (1979) described several prospects in the central Hamersley Basin and Gregory Range.

Wyloo Dome

Copper mineralization accompanies gold-bearing quartz veins at the Paulsen and Belvedere mines in the west-central Wyloo Dome. Small quantities of ore were produced in 1916, 1939, and 1949; total production was reported to be 1.6 t of copper ore (Marston, 1979). Descriptions of the Paulsen and Belvedere mines are given above in the section describing gold mineralization.

A number of small copper prospects have also been reported from Fortescue Group rocks elsewhere in the Wyloo Dome. No production has been reported from these occurrences to the end of 1991.

Blacks (Belfrey, Metawandy) copper prospect is located on WYLOO (AMG 289987). Descriptions of the occurrence were given by Maitland (1909), Forman (1938), Finucane (1939), Marston, (1979), and Blight (1985). Marston (1979) recorded that the prospect is located in an apparently south-dipping assemblage of chert and mafic to felsic volcanic rock (Mount Roe Basalt). At the prospect, the chert forms a conspicuous ridge and is structurally underlain by a sheared, fine-grained, siliceous, metavolcanic rock. This lower unit is pyritic in part and green stained over the basal 10 m of outcrop. The mineralization is weak and apparently of small extent and is probably iron sulfide at depth (Marston, 1979).

Blight (1985) recorded green secondary copper staining in cleaved Mount Roe Basalt in the Wyloo Dome (WYLOO AMG 426877). He also reported an anomalous copper value (235 ppm) from a gossan-like cap overlying undifferentiated Boongal–Pyradie–Bunjinah Formation basalt in an area 8 km southeast of Mount de Courcy. Daniels (1970) recorded a copper occurrence in the same unit near Mindel Well (WYLOO AMG 083060), but this has not been described.

Hamersley Basin, excluding Wyloo Dome and the Gregory Range

Most exploration in the Hamersley Basin has focused attention on carbonaceous and sulfidic shales in the Jeerinah Formation with few anomalies being reported from other parts of the Fortescue Group.

Jeerinah Formation

Marston (1979) summarized the results of a Pilbara-wide exploration program of the Jeerinah Formation, conducted by Western Mining Corporation between 1967 and 1972. Although much of the upper part of the Jeerinah Formation was found to be sulfidic at depth and contained copper values up to 3% locally, no deposits of economic significance were found. The most prospective areas were centred on old workings in the Wonmunna district, 60 km west-northwest of Newman, the northeastern flank of the Milli Milli Dome, and the northern limb of the Jeerinah Anticline.

Wonmunna district: The Bull prospect is located 3.8 km west of the old Wonmunna workings (Marston, 1979), the mineralization being exposed in a small creek. Zones of malachite and chrysocolla veinlets from a few centimetres to 0.3 m thick are visible within a 25 m strike width. The host rocks are pale grey slates, with carbonate-rich beds, dipping 65–75° south. The slaty cleavage is subparallel to bedding and some mineralization is located within the cleavage rather than the bedding. There seems to be little continuity of the mineralization along the strike length of 0.7 km. The best results obtained from drilling were 0.76 m of 1.79% Cu from oxidized slate in diamond drillcore, and 10.67 m drilled width of 3.43% Cu from one rotary drillhole.

The Mount Robinson prospect is 15.7 km west of Wonmunna, just north of a dolerite sill. Five north–south costeans along a 140 m strike length exposed a 25 m-wide zone of calcareous shale with localized stainings of malachite, cuprite, and chrysocolla. Drilling by Western Mining Corporation indicated the presence of an oxidized mineralized zone at least 270 m long, 4.5 m wide and 18 m deep, assaying 0.2–2.3% copper (Marston, 1979). The Bend prospect is 8 km west of the Mount Robinson prospect and consists of similar but weaker mineralization. At the Ironstone prospect six costeans expose about 250 m of bedrock beneath a lateritic cap. White-weathering shales strike 290° and dip moderately to steeply northwards and display an axial planar cleavage associated with west-plunging folds. Two zones of ochrous limonitic shale less than 2 m thick, and 25 m apart, contain malachite and chrysocolla staining (Marston, 1979). The Central (Sleepy Hollow) prospect, investigated by Amax in 1964, consists of five costeans over about 250 m strike length and a small opencut in shales containing fine, pisolitic carbonate horizons. Thin stratabound zones up to 0.4 m thick contain laminae and veins of limonite plus rare malachite and chrysocolla.

The Wonmunna (Wanna–Munna) prospect has a recorded production of 13.53 t of copper ore (25.6% Cu) and 5.96 t of cupreous ore (23.75% Cu) during 1953 (Low, 1963). Today it consists of 12 costeans, cut by Amax, into south-dipping pisolitic carbonate and siliceous shale (Marston, 1979). Mineralization is similar in style to the other prospects in the area but is weak, thin, and has little persistence along strike. Fine-grained disseminated malachite occurs in some carbonate-rich shale. Subsequent drilling by Western Mining Corporation encountered little mineralization, the best intersection being 3 m assaying 2.9% Cu.

Other prospects: The Minthicoondunna (Mindi) prospect is on the northeastern flank of the Milli Milli Dome close to the contact with the Marra Mamba Iron Formation (Marston, 1979). A 110 m vertical drillhole sunk by Western Mining Corporation encountered pyrite- and pyrrhotite-bearing graphitic shale assaying a maximum of 2080 ppm Cu. Most assays were, however, considerably less than 1000 ppm.

The Brockman and Edneys prospects are adjacent to each other in shales with interbedded cherts on the north-dipping limb of the Jeerinah Anticline (Marston, 1979). Five vertical drillholes bored by Western Mining Corporation encountered a maximum of 1.62 m of 0.4% Cu. Low-grade oxidized copper mineralization over an average width of 8 m (3 Mt averaging 0.4% Cu) was indicated by 67 percussion drillholes, of 30 m average depth, put down at the Brockman prospect in 1968. The best intersection was 1.6 m assaying 0.7% Cu. Further shallow drilling in 1971 yielded a peak value of 4.6 m of 0.61% Cu.

Surface copper anomalies associated with Jeerinah Formation shale in the Gregory Gorge and eastern Wyloo Dome were investigated respectively by Westfield in 1971–73 and Asarco in 1971. No encouraging results were reported (Marston, 1979).

Other Fortescue Group units

Minor anomalous copper values have been reported from the Hardey Formation in the western Rocklea Dome (Marston, 1979). The anomalies were recorded at the Beasley River prospect in carbonate-bearing, weakly pyritic feldspathic sandstone, and graphitic shale. Surface mineralization occurs in a north trending 60 × 30 m area. Two diamond-drill holes (total length 280 m) intersected 1.6 m assaying 0.13% Cu and 2.4 m assaying 0.19% Cu.

Hickman (1975a) and Hickman and Lipple (1975) recorded minor cupriferous mineralization in Mount Roe Basalt and Hardey Formation, near Meentheena and Marble Bar respectively.

Gregory Range area

Minor copper mineralization has been reported from the Braeside Lead Field (see following section) where it is generally associated with north-northwesterly striking faults cutting Kylena Formation basalt. Copper minerals have been reported from the Ragged Hills and Lightning Ridge workings, and the north Koongalin prospect.

The Baramine prospect is 2.5 km south of the abandoned Baramine Homestead in a small cupriferous quartz vein filling a fault between Pearana Basalt (Maddina Formation) and Carawine Dolomite (Hamersley Group) (Marston, 1979). Blatchford (1925) described the vein as being 0.3 – 0.4 m thick and opened up by a 3 m deep underlay shaft from which 2–3 t of malachite, cuprite, and chalcocite ore had been extracted. At the nearby Baramine South prospect, a silicified fault zone striking north-northwest contains cuprite and malachite locally. Blatchford (1925) reported taking a sample across a width of 1.4 m, which assayed 25.32% Cu, 279 ppm Ag, and a trace of lead.

Camel Hump South prospect is 11 km southwest of Baramine Homestead and consists of a cupriferous quartz vein in Maddina Formation (Marston, 1979). About 1.5 km to the north, at the small hill known as Camel Hump, cleaved felsic volcanic rock belonging to the Baramine Volcanic Member of the Jeerinah Formation contains veinlets and small pods of malachite.

Lead, silver, and zinc

Between 1925 and 1959, 32 362 t of lead, 25 t of zinc, and 920 kg of silver were mined from the Fortescue Group. Most of this production came from the Braeside field in the Gregory Range, with minor contributions from the Meentheena area and Wyloo Dome (Table 13.1).

Braeside

The Braeside Lead Field is situated in the northern part of the Gregory Range between latitudes 21°00' to 21°20'S and longitudes 121°00' to 121°15'E. Descriptions of the various mines within the field are given by Blatchford (1925), Finucane (1938b), Low (1963), Blockley (1971), and more recently by Ferguson (1999). A summary was given by Hickman (1983). By far the greatest production (2937 t of lead, 870 kg of silver, 25 t of zinc) has come from the Ragged Hills mine (Blockley, 1971).

Argentiferous galena occurs in quartz veins associated with north-northwesterly trending, steeply dipping, silicified faults within the Kylena and Tumbiana Formations (Blockley, 1971; Hickman, 1983). The faults can often be traced for several kilometres, but lead minerals are restricted to shoots of 150 m in length. Where present in the larger faults, lead minerals are confined to marginal veins which form only part of the whole fault zone and appear to post-date the main period of quartz deposition. In the smaller, narrower, reefs lead may be present throughout the width of the veins.

Blockley (1971) noted that the lead-silver ore is found either as massive lenses 0.15 – 1.5 m in width, or as disseminations through quartz and quartz-filled breccia. In most mines the zone of weathering is shallow, ranging from 0–15 m. The main sulfide mineral in the primary ore is galena, which occurs with small amounts of sphalerite, chalcopyrite, bornite, and pyrite. The oxidized ore minerals are chiefly cerussite and anglesite, with minor amounts of pyromorphite. Some deposits contain significant amounts of malachite and cuprite, and the vanadium minerals mottramite, descloite, and vanadinite have been recorded from the field (Blockley, 1971).

Table 13.1. Production of lead, zinc, and silver from centres hosted by Fortescue Group rocks

Locality	*Period*	*Lead (t)*	*Zinc (t)*	*Silver (kg)*
Braeside	1925–59	3 229	25.3	917
Meentheena	1949	5.6	nd	nd
Wyloo	1949	1.8	nd	2.5
Total		**3 236.4**	**25.3**	**919.5**

NOTE: nd: no data

Meentheena area

Four tonnes of lead was reported to have been extracted from the Meentheena prospect about, 32 km southeast of Meentheena Homestead (Blockley, 1971). No further details of the prospect are known.

Wyloo Dome

In 1949, 1.8 t of lead and 2.5 kg of silver were extracted from the Belvedere mine (see description above of gold deposits). Principal lead minerals were galena, cerussite, beudantite, and plumbojarosite (Blockley, 1971).

Blight (1985) observed galena in quartz veins within the Mount Roe Basalt, 4.5 km southeast of Horse Well.

Other occurrences

The Barramine deposit is reputed to have produced a small quantity of lead ore (Blatchford, 1925; Finucane, 1938b). The mine is on the northern extension of the Braeside Lead Field on a quartz vein striking 320° and dipping steeply east. An inspection by Blockley (1971) uncovered only copper carbonate and chalcocite at the prospect; no lead minerals were found.

Marston (1979) notes that silver-rich veinlets and zinc values up to 0.4% are associated with copper mineralization at the Wonmunna prospect, northwest of Newman (see description above of copper deposits).

Blight (1985) recorded that lead and mercury mineralization is associated with quartz veins on the southern side of the Milli Milli Dome, immediately west of Coppin Pool. Galena and cerussite, and traces of cinnabar exist in an east-trending quartz vein that is totally contained within a dolerite sill in the Hardey Formation.

Blight (1985) quoted J. R. Richards as having dated the lead at 2.4 Ga.

Other minerals

Fibrous chrysotile asbestos has been observed within the Hardey and Pyradie Formations in the Rocklea Dome. The Hardey Formation occurrences are in veins cutting layered sills; those in the Pyradie Formation form veins in pyroxene spinifex-textured basalt and komatiite. Blight (1985) observed fibres up to 0.2 m long and interpreted them as having formed in dilation seams by hydrothermal alteration of the enclosing ultramafic material.

Vanadium minerals, including mottramite, descloizite, and vanadinite have been recorded from the Braeside Lead Field (see description above of lead deposits).

Anomalous nickel (0.2% NiO) and chromite (0.7% Cr_2O_3) values have been recorded in a komatiite flow unit within the Pyradie Formation on MOUNT BRUCE (**Chapter 12**).

Residual manganese and ferromanganese deposits are associated with the Marra Mamba Iron Formation, immediately overlying Jeerinah Formation shales at Mount Nicholas and Mount Fraser on BALFOUR DOWNS. The Mount Nicholas deposit, from which a total of 3642 t at an average grade of 45% Mn was extracted (Fetherston, 1990), was last mined in 1966.

Chapter 14

Synthesis and discussion

Preceding chapters of this Bulletin systematically describe and interpret the lithostratigraphy of the Fortescue Group. In this chapter we first synthesize these interpretations into a single broad account of the initiation and development of the Hamersley Basin up to the deposition of the Hamersley Group. We then discuss various aspects of the tectonic evolution of the basin.

Depositional history of the basin

Nature and former extent of the basement

Fortescue Group rocks were deposited on granite–greenstone terrane of the Pilbara Craton. A comprehensive descriptive account of the geology of this basement was given by Hickman (1983); Griffin (1990) and Hickman (1990) provided later summaries, while Trendall (1995) has reviewed subsequent work to that date. Hickman (1983) recognized a stratigraphic succession whose lower part (Warrawoona Group) consists principally of mafic, ultramafic, and felsic volcanic rocks with a subordinate sedimentary component, and whose upper part (mainly the Gorge Creek Group) is made up mainly of mixed volcanic and epiclastic sedimentary rocks. The entire sequence has undergone low- to medium-grade metamorphism, and is folded into rectilinear to sinusoidal, synformal greenstone belts which enclose large complex granitoid batholiths with shapes ranging from nearly circular (Mount Edgar Batholith), through ovoid (Corunna Downs Batholith), to irregular (Yule Batholith). These granitoid batholiths, for which Griffin (1990) preferred the term granitoid complexes, have discordant, intrusive margins against the lower units of the greenstone sequence, but isotope geochronology indicates that both greenstones and granitoids developed synchronously, during the approximate period 3.5 – 2.9 Ga, so that their deposition and evolution were contemporaneous, and presumably interlinked. An exception is provided by the uppermost units of the stratigraphic sequence, the ~3.0 Ga Whim Creek Group and the 2.95 Ga De Grey Group.

The tectonic evolution of the granite–greenstone basement is controversial. Hickman (1983) considered that the granitoid domes were essentially solid-state diapirs, and that the later greenstones accumulated in synclinal areas between the rising granitoid complexes. Within the greenstones, volcanism, sedimentation and deformation were controlled largely by the diapiric growth of the granitoids. This view has been challenged by Bickle et al. (1980, 1985) and Bettenay et al. (1981), who recorded evidence of major northeasterly thrusting and recumbent folding of the older greenstones and granitoids in the Shaw Granitoid Complex. According to this view, granitoid diapirism occurred later and was a consequence, rather than the cause, of the deformation and metamorphism. Other tectonic models have been reviewed by Krapez and Eisenlohr (1998). The youngest deformation event in the basement evolution resulted from north–south compression and produced northeast-trending sinistral wrench faults and associated strike-slip basins such as the Lalla Rookh Basin (Krapez, 1984) and Whim Creek Belt (Barley, 1987).

The extent of the granite–greenstone terrane of the Pilbara Craton prior to Fortescue Group deposition is unknown, but there is evidence to suggest that it formed part of a substantially larger area of continental crust, which extended outside the present margins of the craton.

Firstly, palaeocurrent and provenance data from the Hardey and Tumbiana Formations point to the presence of a source area to the north of the present-day northern craton margin. Secondly, there are no major sedimentological or volcanological facies changes when the Fortescue Group is traced from east to west across the craton; and if the granite–greenstone terrane of the Pilbara Craton was restricted to its present extent during Fortescue Group deposition, some evidence of systematic facies variations inwards from the margins might be expected. A possible exception in the Gregory Range area is noted below.

Finally, the geological evolution of the Pilbara points to two major periods of continental rifting during Fortescue Group times (Blake, 1984a; Blake and Groves, 1987). In this regard, the dominantly west-facing, syn-Hardey Formation extensional structures on the northern craton are consistent with the view that this area once formed part of a continent that extended beyond the present-day western Pilbara margin (Blake, 1993). Similarly, there is strong evidence that the middle to upper Fortescue Group of the southern Pilbara formed during a later rifting event produced by north–south extension; this implies that the granite–greenstone basement also extended farther south than the existing outcrop would suggest.

If the present Pilbara Craton is indeed part of a much larger continent, the present positions of any missing parts are unknown. It is unlikely that the northern Yilgarn Craton represents the other half of the southern rift system, because most evidence suggests the Pilbara and Yilgarn were brought together by continental collision about 1800 Ma (Myers, 1990; Tyler and Thorne, 1990; Thorne and Seymour, 1991).

The only area in which a possible margin of the early Pilbara Craton may have been located is immediately east of the Gregory Range inlier, where the stratigraphy of the lower part of the Fortescue Group (**Chapter 10**), especially in the presence of the thick and extensive Koongaling Volcanic Member of the Hardey Formation, is markedly different from outcrop areas farther west. This would be consistent with de Laeter et al.'s (1977) argument, based on Rb–Sr analyses, that the older sialic crust of the Pilbara Craton did not extend east of the Gregory Range area. However that may be, we still see the question of the pre-Fortescue Group extent of the Pilbara granite–greenstone terrane as a problem awaiting resolution; and the fact that most of the present craton boundaries are concealed beneath younger sedimentary rocks adds to the difficulty of finding definitive evidence.

The basal unconformity

The basal unconformity of the Fortescue Group over the granite–greenstone terrane is now a single, well-defined stratigraphic entity. However, its present physical continuity conceals wide local differences in 'age'. The quotation marks here emphasize that the attribution of a single age even to a single point on a surface of unconformity is conceptually unsound. The ages of the rocks immediately above and below are two potentially determinable age parameters of an unconformity, the difference between which defines its maximum time interval. But ages intermediate between these limits may be more significant for the geological interpretation of an unconformity, and may be harder to determine: these include, for example, the time at which the lower surface achieved erosional maturity, and in the case of a major basal unconformity, the time at which the tectonic unit to which the underlying basement belongs achieved tectonic stabilization.

The age difference between rocks immediately above and below the unconformity between the Pilbara granite–greenstone terrane and the Fortescue Group varies greatly. This is partly because the Fortescue Group rests on Pilbara Craton rocks of different ages, but it is also a consequence of the different ages of the basal Fortescue Group rocks at various points above the unconformity. As a result, although it is often convenient to use the term 'basal unconformity of the Fortescue Group' when referring to this lower contact, it should be noted that the age significance of this surface varies widely across the Pilbara.

The major depositional sequences

The ten lithostratigraphic formations of the Fortescue Group reflect a wide range of volcanic and sedimentary environments that prevailed within various parts of its total outcrop area between the initiation of the depository at ~2775 Ma and the completion of Jeerinah Formation deposition at ~2630 Ma. We find it convenient here to group these ten formations into four 'depositional sequences' which, for reasons set out below, appear to correspond to discrete stages in the development of the basin. The four depositional sequences, each of which consists of assemblages of genetically related strata, are listed below and numbered in ascending stratigraphic order:

Depositional sequence	*Stratigraphic unit*
4	Jeerinah Formation and Hamersley Group
3	Maddina, Bunjinah, Tumbiana, Pyradie, Kylena, and Boongal Formations
2	Hardey Formation
1	Mount Roe Basalt, Bellary Formation, and other pre-Mount Roe Basalt units.

The term 'depositional sequence', normally abbreviated to 'sequence', was introduced between quotation marks above to emphasize that we do not apply it in a way exactly corresponding either to its usage in the set of sequence stratigraphic units of Vail et al. (1977) or its usage in the sense implicit in the term unconformity-bounded sequence (UBS) (ISSC, 1987). Both of these have previously been applied to the Fortescue Group, and both have connotations which we prefer to avoid.

Cheney et al. (1988) initially proposed the subdivision of the Fortescue Group into four UBS, closely similar to those we use, on the basis of a comparison of the stratigraphy of the supracrustal successions of the Kaapvaal Craton and Pilbara Craton; Cheney (1996) appears to have partly revised these subdivisions. We prefer not to identify our sequences specifically as unconformity-bounded because the presence of such unconformities is not a mandatory feature of them. On a basin-wide scale, their general concordance is as noteworthy a feature as the unconformities or disconformities which certainly separate them locally.

We also prefer not to follow the subdivision of the Fortescue Group into Supersequences, Supersequence Packages, and Sequences and Sequence Packages, used by Blake (1993), Blake and Barley (1992), and Barley et al. (1992). All those authors based their nomenclature on the stratigraphic concepts of Vail et al. (1977), in which the different sequence terms relate to cyclic changes of global sea level, and imply an order of duration for each unit. We do not consider that the absolute depositional chronologies of the Fortescue Group, or of the Mount Bruce Supergroup, are sufficiently well established for these terms to be appropriately applied. However, because all such alternative stratigraphic subdivions are based on the same set of formal lithostratigraphic units, our Sequences 1 and 2 correspond to the Mount Roe Sequence and Hardey Sequence Package respectively of Blake (1993), Blake and Barley (1992), and Barley et al. (1992). Our Sequence 3 equates to their Mount Jope Supersequence; and our Sequence 4 equates broadly to their combined Marra Mamba and Brockman Supersequence Packages, Woongarra Supersequence, and lower Turee Creek Supersequence.

Our numbered depositional sequences described here are an order of magnitude thicker than those described from Phanerozoic passive margin settings (e.g. Vail et al.,1977; Posamentier et al., 1988). In terms of the time interval they represent, Sequences 1 and 2 are broadly comparable to the third-order cycles of Vail et al. (1977), whereas Sequences 3 and 4 appear more analogous to the second-order cycles of these workers.

Sequence 1: Mount Roe Basalt; Bellary Formation and pre-Mount Roe Basalt Fortescue Group units

Sequence 1 is up to 4.5 km thick and comprises the Mount Roe Basalt and various pre-Mount Roe Basalt Fortescue Group sedimentary units, such as the Bellary Formation (**Chapters 3** and **4**). Sequence 1 rocks are recorded from the northwest and southwest Pilbara, and from the northeast Pilbara, including the Marble Bar area. They are not present in the Gregory Range inlier, and are also absent from much of the central and southeast Pilbara.

No geochronological data are available from the pre-Mount Roe Basalt sedimentary units. Uranium–lead zircon data indicate that the lower part of the Mount Roe Basalt was extruded between 2775 ± 10 and 2763 ± 13 Ma (Arndt et al., 1991), about 125–175 m.y. after the youngest Pilbara greenstone was deposited (Trendall et al., 1990), and possibly 30–116 m.y. after the youngest basement granitoid was intruded (Blake, 1993).

The duration of Sequence 1 is unknown, as ages from Sequence 2 are within error of those cited above. A period of 1–10 m.y. would be consistent both with the available geochronological data and with the best current information on the rapidity with which stacked basalts may be erupted (Fig. 14.4).

Sequence 1 rocks were deposited on a rugged, subaerial landscape with local relief up to more than 500 m (Blake, 1993). Local topographic highs were controlled by more resistant lithologies such as chert, felsic volcanics, and silicified sandstone; areas underlain by granitoid were commonly of more subdued relief.

Over most of the Pilbara, the earliest Sequence 1 deposits comprise thin, laterally impersistent, alluvial and colluvial conglomerate and sandstone, associated locally with lacustrine argillite and sandstone. There are notable exceptions to this sedimentation pattern in two areas: firstly near Marble Bar, where major sandy braided fluvial systems drained into the area from granitoid uplands located to the north and southwest; and secondly near Bellary Dome, where a small fan-delta system built out toward the west-northwest in a lake or marine embayment.

In the northwest, northeast, and southwest Pilbara, basal sedimentary units, and much of the pre-Fortescue Group topography, were buried beneath the Mount Roe Basalt, a 1 km-thick pile of mostly subaerial basaltic flows and minor lacustrine pillow lava and hyaloclastite.

Sequence 1 is not represented in the Gregory Range inlier.

The northern and southern basaltic provinces of Sequence 1 are today separated by a west-northwesterly trending area which is largely devoid of basalt. The meaning of this distribution is somewhat uncertain and two interpretations are possible. Firstly, that the regional pattern is primary, and the central area was never covered by Mount Roe Basalt; and secondly, basalts covered the entire craton but were subsequently removed by erosion before and during Sequence 2 times. The evidence available favours the first interpretation, largely because of the great thickness of the preserved Mount Roe Basalt sections when compared with the amount of basalt-derived material in overlying sedimentary rocks.

The volcanic style of the Mount Roe Basalt included thick (20–80 m) ponded lavas; medium thickness (5–20 m) flows of uncertain origin; and thin, non-ponded pahoehoe flows. This association appears to fall in between plains volcanism and Columbia River Basalt flood volcanism (cf. Waters, 1961; Greeley, 1982; Long and Wood, 1986). Features in the Mount Roe Basalt that are characteristic of plains basalts include:

- the presence of pahoehoe flows,
- many relatively thin (1–5 m) flows, and
- eruptions were probably fissure-fed.

The following differences with plains volcanism are apparent, however. Firstly, no basaltic shield volcanoes or lava tubes have been identified in the Mount Roe Basalt. Secondly, most lava flows in the Mount Roe Basalt are over 5 m thick and display limited relief on flow surfaces. In addition, cinder and tuff cones, and collapse structures have not been recorded from the Fortescue Group. On the other hand, the volcanic style of the Mount Roe Basalt differs from that of the Columbia River Basalt in having more pahoehoe flows and a greater proportion of other thin flows of apparently limited lateral extent.

Earlier suggestions that the north-northeasterly trending dykes of the Black Range Suite may represent a feeder system for the Mount Roe Basalt (e.g. Lewis et al., 1975; Hickman, 1983; Blake, 1993) were to some extent speculative, since field relationships between Fortescue Group rocks and these dykes are unclear. However, recent geochronological data (Wingate, 1994, 1997, 1999) show that the ages of two major dykes of the Black Range Suite are statistically indistinguishable from those of the Mount Roe Basalt, and thus confirm this association.

In the northwest and northeast Pilbara, Sequence 1 rocks were faulted and tilted locally, then eroded before deposition of Sequence 2 (Blake, 1993). This resulted in either a disconformity or slight angular discordance between Sequence 1 and Sequence 2 rocks in these areas. In the Marble Bar sub-basin, a narrow, sinuous belt of gentle to tight folds with steeply dipping axial surfaces and mostly gently plunging fold axes formed after Sequence 1 deposition. This period of folding was followed in turn by an interval of faulting and erosion prior to deposition of Sequence 2 (Blake, 1993). The post-Sequence 1 – pre-Sequence 2 history of the south Pilbara sub-basin is poorly known. Horwitz (1990) refers to the boundary between these units as an angular unconformity, but the contact appears to be disconformable in the southeastern Rocklea Dome and no evidence of erosion or angular discordance was observed by us in the Bellary and Wyloo Domes.

We take the local discordance between Sequences 1 and 2, together with a sharp increase in depositional area, as well as a change from eruption of stacked basaltic lavas to epiclastic sedimentation and felsic volcanism, as joint justification for acceptance of the Sequence 1 – Sequence 2 boundary as that between separate stages in the devlopment of the depository.

A summary of the palaeogeography and its tectonic controls during Sequence 1 is given in Figure 14.1a.

Sequence 2: Hardey Formation

Sequence 2 is up to 3000 m thick and consists of one major stratigraphic unit, the Hardey Formation, which extends throughout most of the main outcrop area of the Hamersley Basin and the Gregory Range. A SHRIMP U–Pb zircon date of 2756 ± 8 Ma (Arndt et al., 1991) from the Bamboo Creek Porphyry, confirming the conventional multi-grain age of 2768 ± 16 Ma reported by Pidgeon (1984), is within error of ages already noted for the lower part of Sequence 1. The depositional hiatus between the two must therefore have been relatively short. Arndt et al. (1991) also reported a SHRIMP U–Pb zircon date of 2764 ± 8 Ma from the Koongaling Volcanic Member of the Hardey Formation of the Gregory Range inlier, supporting the lithostratigraphic equivalence of the extensive felsic volcanics of the inlier, together with their related overlying epiclastic rocks, as components of Sequence 2.

The principal characteristics of Hardey Formation sedimentation and volcanism on the northern Pilbara Craton have been well summarized by Blake (1993) as follows:

- Depositional environments were dominantly continental, including braided alluvial, alluvial fan, and lacustrine.
- Sedimentation and possibly felsic volcanic centres were associated with dominantly north-northeasterly trending extensional structures, with consistent west-block-down normal movement.
- In the northeast and northwest Pilbara sub-basins, fluvial transport was predominantly parallel to associated north-northeasterly trending tectonic structures, except in the Northwest Oakover Syncline, where flow was toward the southeast. In the Marble Bar sub-basin, sedimentation trends varied with stratigraphic position but were either parallel to or normal to the main tectonic grain.

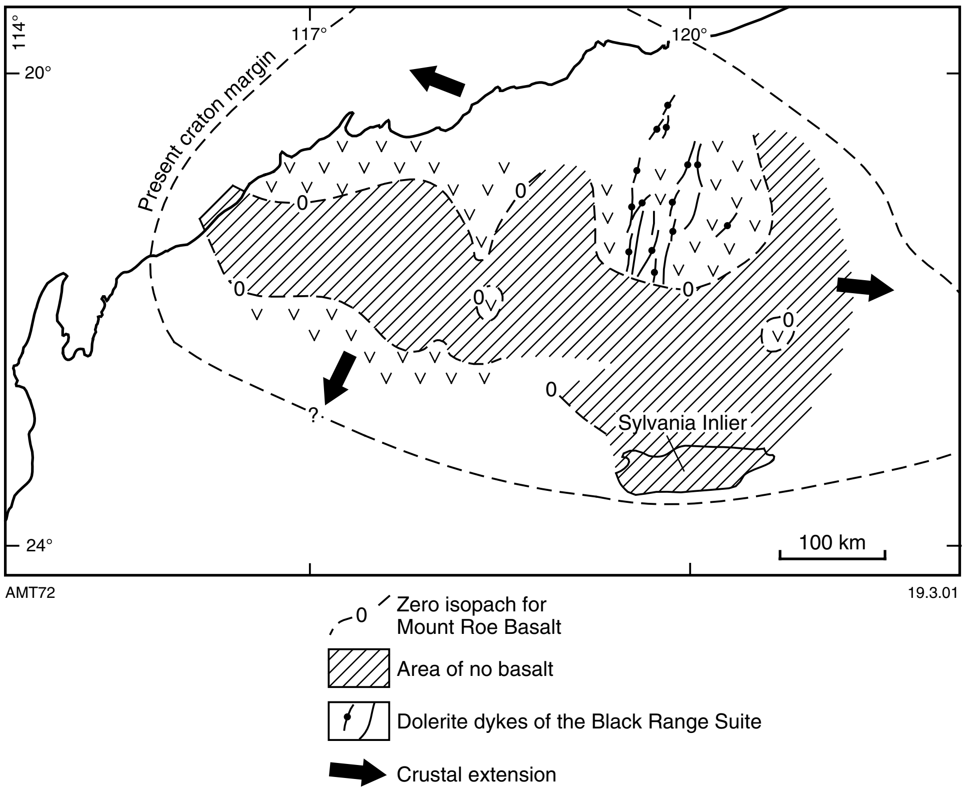
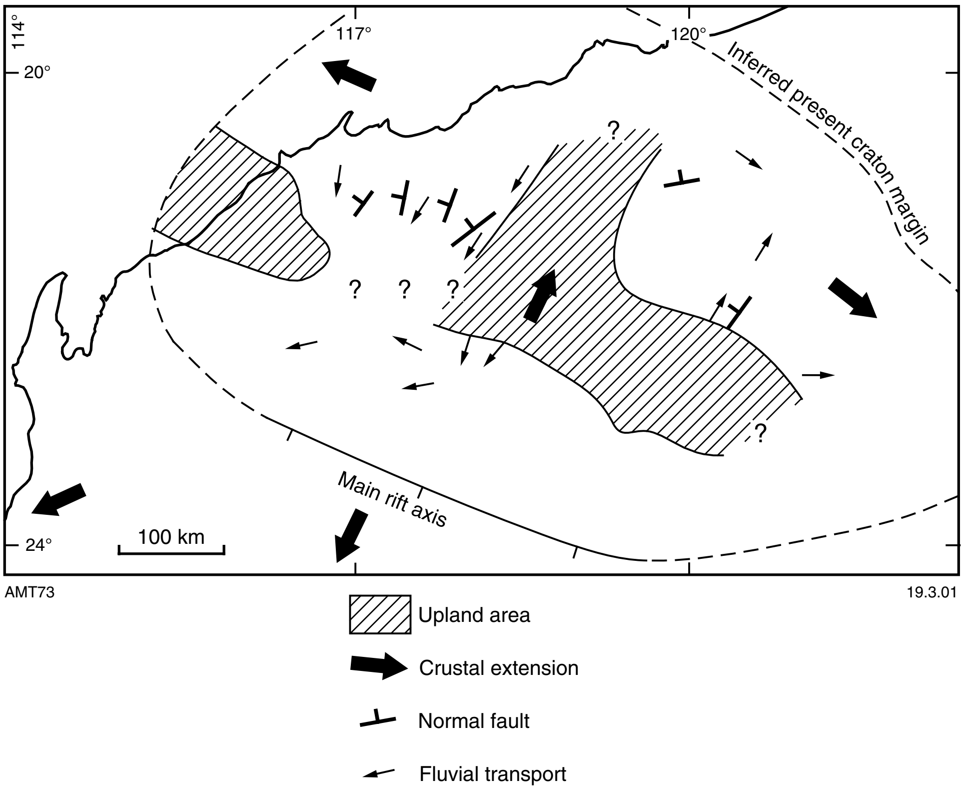


Figure 14.1. Tectonic and palaeogeographic reconstructions for the Pilbara during deposition of Sequence 1 (14.1a — shown above), Sequence 2 (14.1b) and Sequence 3 (14.1c). Reconstructions are based on data presented in Chapters 3 to 10 and from Blake (1984a, 1993)



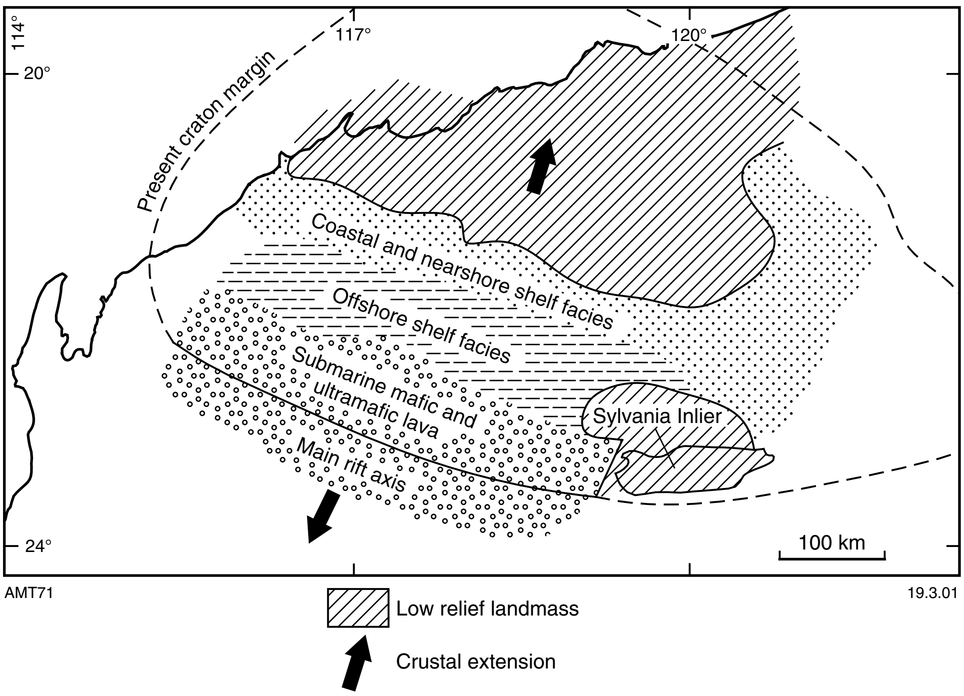


Figure 14.1. (continued); b (top); c (bottom)

- Most of the non-volcanogenic sandstones are texturally and compositionally immature and are derived from basement granitoids.
- In the northwest Pilbara sub-basin, syn-Hardey Formation northeasterly or north-northeasterly trending structures are fairly evenly spaced over a distance of about 200 km with an average spacing of about 20 km.
- Relics of a major growth-fault system have been located in each sub-basin with the following estimated syndepositional throws: Conglomerate Creek Fault over 1000 m, Pear Creek Fault 450–1300 m, Sherlock River Fault Complex over 1500 m. Each major growth fault is subparallel to, and offset a similar distance southeast from, a major basement structure. The Conglomerate Creek Fault is about 20 km southeast of the Nullagine Lineament; the Pear Creek Fault is about 20 km southeast of the western main boundary strike-slip fault of the Lalla Rookh Basin (Krapez, 1984); and the Sherlock Fault Complex is up to 17 km southeast of the Nunyerry Creek Fault System (although the two fault systems converge north-northeastwards along strike). In the northeast and northwest Pilbara sub-basins, there is evidence that the Nullagine lineament and the Nunyerry Creek Fault System were active during deposition of the Hardey Formation, but their influence on sedimentation patterns (where adjacent sedimentary rocks are preserved) is much less than that of nearby growth faults to the southeast.

In the Gregory Range inlier, the rocks of Sequence 2 have a closer and more continuous volcanogenic association than in other areas of Fortescue Group deposition. The lowest unit of the Hardey Formation here, which is also the lowest unit of the Fortescue Group, is the extensive rhyolite of the Koongaling Volcanic Member. Early components of this thick-sheeted volcanic accumulation of ash-flows, lavas flows, and sills may have been extruded over a granite–greenstone basement, but no trace of this is now preserved, because the cogenetic Gregory Granitic Complex was emplaced beneath the volcanic material penecontemporaneously with its formation. Wherever the basal contact of the Koongaling Volcanic Member is not tectonic there is a downward gradation through high-level granophyre into lower units of the complex, which represents the plutonic and hypabyssal core of a massive silicic volcanic complex whose extrusive component was the rhyolite of the Koongaling Volcanic Member. All units from which zircon U–Pb ages are available, including the granophyre, have ages within error of the overlying volcanic rocks (Nelson, 1996).

Rocks of the Gregory Granitic Complex are anhydrous, alkaline A-type granites, derived from a mixed crustal source. They imply an anorogenic setting and are probably the result of crustal melting of a source that has been through a previous subduction or collisional history (R. H. Smithies, quoted in Williams and Trendall, 1998a). Their presence provides further support for a comagmatic relationship between the Gregory Granitic Complex and the overlying Koongaling Volcanic Member.

The sedimentary rocks of the overlying Warri Warri Member and Tanguin Member are closer in depositional facies to the Sequence 2 rocks of the main outcrop area. However, the turbiditic sandstones of the Warri Warri Member suggest a deeper water environment than the fluvial to alluvial-fan and deltaic sandstones of the main outcrop area. They differ also in having a consistent volcaniclastic component.

In the south Pilbara sub-basin, depositional environments of Sequence 2 evolved from braided fluvial to deltaic. As is the case in the north Pilbara, braided fluvial sandstones are compositionally and texturally immature and were probably derived from basement granitoid rocks. Sediment transport directions in fluvial and deltaic distributary channels were principally toward the southwest and west and indicate the presence of a basement high to the north and northeast of the Milli Milli Dome. The greatly reduced thickness (<100 m) of the Hardey Formation in the SGS–1, WRL–1, and FVG–1 diamond drillholes, and the absence of Hardey Formation from the central north Pilbara, further supports this view. In the Newman area, there is a marked northward thinning of the Hardey Formation across the Sylvania Inlier. This evidence, combined with the absence of thick subaqueous deltaic facies along the northern margin, suggests that the basement high to the north of the Milli Milli Dome may have extended close to the northern margin of the Sylvania Inlier.

With the exception of the Western Creek and Fortescue River Faults on NEWMAN, no syn-sedimentary growth faults have yet been identified in the south Pilbara sub-basin. As noted earlier, the available stratigraphic and sedimentological data suggest the presence of a west-northwesterly trending palaeohigh that extended from the Sylvania Inlier to the headwaters of the Yule River on PYRAMID — the Yule–Sylvania High of Blake (1993). The orientation of this palaeohigh is parallel to the main pre- and post-Fortescue Group tectonic grain in the southern Pilbara and the possibility exists that some of these structures were active during deposition of the Hardey Formation and overlying parts of the Fortescue Group (Thorne, 1990; see also **Chapters 6–9**).

To summarize, while there is clear evidence for a north-northeasterly trending growth-fault system in the northern Pilbara, these structures are generally not recognized in the southern part of the craton. Although it seems likely that a west-northwesterly to northwesterly trending basement high existed in the central Pilbara during Hardey Formation deposition, there is only indirect evidence for an active growth-fault system parallel to its southern margin (see **Structural controls on sedimentation and volcanism in the southern Pilbara** below)

Figure 14.1b shows the palaeogeography and tectonic controls for the Pilbara Craton during deposition of the lower Hardey Formation. Thickness, facies, and palaeocurrent data point to the presence of a major, southerly flowing, sandy braided fluvial system in the northwestern part of the craton. Here, alluvial fan and gravelly braided rivers were associated with small palaeohighs and north-northeasterly trending fault scarps in the braid plain. The zero isopach for the Hardey Formation suggests that one of the larger topographic highs extended southeast from Cape Preston.

The southeastern margin of this northwest Pilbara sub-basin was marked by the Nunyerry–Sherlock fault system, and to the southeast, a major palaeoridge extended toward the present-day Sylvania Inlier. In the Marble Bar district to the north of the ridge, sandy braided rivers flowed from the north and southeast. Farther to the east and southeast in the Meentheena and Nullagine districts, sandy braided streams flowed toward the north-northeast fault-bounded troughs. Southwest of the palaeohigh, in the Wyloo–Milli Milli area, braided streams flowed toward the southwest and west.

Deposition of the upper Hardey Formation was characterized by increased levels of felsic volcanism in the northwest and northeast Pilbara. In the central northwest Pilbara sub-basin, three volcanic centres were developed adjacent to, and to the west of, major north-northeasterly trending normal faults and lineaments (Blake, 1993). Here, the volcanic centres were active during the remainder of Hardey Formation deposition. In the northeast Pilbara sub-basin, the main period of felsic volcanism was followed by significant uplift and erosion, and finally by a return to continental siliciclastic sedimentation. Levels of felsic volcanism were significantly lower in the southern Pilbara. Here, the lower Hardey Formation braid plain evolved into a fluvial-dominated deltaic complex, which persisted until the close of Hardey Formation times.

On the northern Pilbara craton, the lower boundary of Sequence 3 is either conformable or unconformable upon the Hardey Formation, or else is unconformable upon granite–greenstone basement. In the northwest Pilbara sub-basin, evidence of an angular discordance was observed between the Kylena and Hardey Formations, but is confined to a small area near the western boundary of PYRAMID. In the Nullagine Synclinorium of the northeast Pilbara sub-basin, the boundary with the Hardey Formation is an angular unconformity (Blake, 1993). Elsewhere, however, this contact is either a disconformity or a nondepositional unconformity (Blake, 1993). Blake's (1993) reconstruction of Hardey Formation depositional systems around the inlier between the Meentheena Centrocline and the Northwest Oakover Syncline suggests that hundreds of metres of Hardey Formation were eroded prior to deposition of the Kylena Formation. The northeast Pilbara sub-basin therefore appears to have been gently folded, and the cores of the major anticline eroded, before deposition of the Kylena Formation.

In the south Pilbara sub-basin, no angular discordance, or evidence of significant erosion has been observed at the Sequence 2 – Sequence 3 boundary. Instead, this contact represents a major marine-flooding surface which separates deltaic sedimentary and volcanic rocks of the Hardey Formation from the submarine volcanic rocks of the Boongal, Pyradie, and Bunjinah Formations.

The Sequence 2 – Sequence 3 boundary in the Gregory Range inlier is more difficult to define as a precise boundary than in other areas. Williams and Trendall (1998a,b) accepted the nominal base of the Kylena Formation as the base of the first thick, continuous-stacked basaltic flows above the Hardey Formation. However, in both the Warri Warri and Newdegate slices, the presence of both felsic volcanic rocks and mafic lavas of unusual type in the upper part of the Warri Warri Member complicates the transition. Nowhere in the inlier is there evidence of major discordance.

Although the boundary between Sequence 2 and Sequence 3 is regionally concordant, the presence of local discordance, combined with the abrupt cessation of the association of largely felsic volcanism and epiclastic sedimentation, and the recommencement of large-scale eruption of stacked basaltic flows, all justify acceptance of the boundary between these sequences as that of a further stage in the evolution of the depository.

Sequence 3: Kylena and Boongal Formations, Tumbiana and Pyradie Formations, and Maddina and Bunjinah Formations

Sequence 3 includes that part of the Fortescue Group between the top of the Hardey Formation and the base of the Jeerinah Formation. On the northern part of the craton, the sequence is about 1.5 km thick and comprises the Kylena, Tumbiana, and Maddina Formations; in the south Pilbara the equivalent units are respectively the Boongal, Pyradie, and Bunjinah Formations, which have a combined thickness of about 3 km.

No reliable age data are available from the Kylena or Boongal Formations. The youngest well-defined age population from Sequence 3 (2715 ± 6 Ma) comes from zircons in a tuffaceous sandstone from the middle of the Tumbiana Formation (Arndt et al., 1991). These workers interpret this as the likely age of penecontemporaneous volcanism; the more abundant older zircons were all older than the Mount Roe Basalt, and may either have been entrained during ascent of the magma or derived by erosion of exposed basement. A U–Pb zircon date of 2717 ± 2 Ma from a rhyolite in the Maddina Formation (Nelson, 1998; Kojan and Hickman, 1998) shows that this formation was deposited very shortly after the Tumbiana Formation was laid down.

The major tectonic and palaeogeographic controls on Sequence 3 volcanism and sedimentation are summarized in Figure 14.1c and the principal characteristics of this sequence may be summarized as follows:

- From north to south across the Pilbara Craton, there is a general change from subaerial and shallow-marine continental volcanism and sedimentation to 'deeper' submarine volcanism. This change is accompanied by a marked thickening of the sequence toward the south (Horwitz, 1980; Morris and Horwitz, 1983).
- The Yule–Sylvania High, which was a dominant feature of Lower Fortescue Group palaeogeography, was gradually onlapped and buried by the end of Tumbiana Formation deposition (Blake, 1993).
- In the north and south Pilbara, lower and upper parts of Sequence 2 (Kylena or Boongal

Formations and Maddina or Bunjinah Formations) are dominated by 'tholeiitic' basalts, although high-Mg varieties, together with rhyolite and dacite, occur in the northwest Pilbara sub-basin. Thick (5–70 m) massive to amygdaloial flows, and thinner (<5 m) amydaloidal flows characterize subaerial and shallow-marine palaeoenvironments on northern parts of the craton, whereas massive flows, associated with pillow lava and hyaloclastite breccia, are abundant in the 'deeper' marine-shelf setting in the south. Strongly amygdaloidal flows, evident in upper parts of the Bunjinah Formation, are also interpreted as submarine facies. Although reasons for the increased vesicularity are unclear, it may be related to regional shallowing, and a consequent decrease in hydrostatic pressure caused by rapid aggradation of the volcanic pile.

- The contrast between north and south Pilbara successions is strongest when middle units (Tumbiana and Pyradie Formations) of Sequence 3 are compared. In the north Pilbara, the Tumbiana Formation comprises 100–200 m of sedimentary carbonate and siliciclastics, volcaniclastic rock and mafic lava, laid down in coastal and shallow-marine environments. Palaeocurrent and provenance data suggest that much of the continental detritus in the northwestern stratigraphy was introduced from a low-relief granite–greenstone landmass which existed to the north of the present-day Chichester Range. The spatial and temporal distribution of pyroclastic air-fall tuff suggests that there were at least two major hydrovolcanic eruptive centres in the north Pilbara during Tumbiana Formation deposition. The first of these was in the northeast Pilbara, and was active mostly during deposition of the lower Tumbiana Formation; the second centre affected the northwest Pilbara during deposition of the upper part of the stratigraphy. Coastal and shallow-marine facies of the Tumbiana Formation extend as far south as a poorly defined line extending from the northern Sylvania Inlier to the northern limb of the Jeerinah Anticline. South of this line, equivalent rocks of the Pyradie Formation are about 1 km thick and comprise 'deeper' marine pyroxene spinifex-textured basalt and hyaloclastite, komatiite, argillite, and chert.

The environment of Sequence 3 volcanism and sedimentation in the Gregory Range area was essentially similar to that of the northern Pilbara sub-basins, in which the shallow-water sediments of the Tumbiana Formation are sandwiched between the stacked subaerial basaltic flows of the Kylena Formation below and the Maddina Formation above.

On the northern Pilbara Craton the north-northeasterly trending normal fault system, which exercised a strong control on Sequence 1 and 2 sedimentation and volcanism, had relatively little influence on Sequence 3 facies distribution. There is however, a tendency for the thickest sections of Kylena Formation to be preserved above the main north-northeasterly trending rift basins. Blake (1993) concluded from such data that the Nunyerry Creek Fault Complex and the southern extension of the Conglomerate Creek Fault were reactivated with a west-block-down movement (the same sense as in Sequence 2 times), and that the Sherlock River Fault System was reactivated with an east-block-down movement (the opposite sense to Sequence 2 times).

It has been suggested that the Sequence 3 – Sequence 4 boundary represents a significant disconformity (Kriewaldt and Ryan, 1967; Williams, 1968; Horwitz and Smith, 1978). However, in almost all instances where this contact has been observed during this study, basal Sequence 4 rocks overlie strongly vesicular, basaltic flow-top material, which does not support either the idea of appreciable erosion between the deposition of Sequence 3 and Sequence 4 or, by consequent implication, significant tectonism. Similarly, certain unusual features of the basalt immediately below this contact, such as large hummocky undulations, bedding-parallel fabric, and ferruginous alteration, are not the result of erosion or weathering, but appear instead to be either primary features of the underlying basalt flows or the result of post-Jeerinah Formation modification.

Although there is no good evidence of significant post-Sequence 3 – pre-Sequence 4 erosion, it is still possible that this boundary represents a significant non-depositional interval. Even if no large time-break is involved, the Sequence 3 – Sequence 4 contact clearly represents an important boundary within the Fortescue Group since, on the north and central Pilbara, it marks a change from subaerial to subaqueous conditions. As such it represents a major flooding surface (van Wagoner et al., 1988), correlated with the marine Bunjinah Formation – Jeerinah Formation contact of the southern Pilbara Craton, and is accepted by us as the boundary of a further stage in basin development.

Sequence 4: Jeerinah Formation and Hamersley Group

Sequence 4 comprises the Jeerinah Formation and Hamersley Group, which have a combined thickness of about 1.5 to 2.5 km, excluding intrusive volcanic rocks.

Uranium–lead zircon geochronology has provided a good chronostratigraphic framework for the Jeerinah Formation and Hamersley Group. From a review of the available data, Trendall (1997) has suggested a total time span of 267 m.y. for Sequence 4, ranging from 2710 Ma for the base of the Jeerinah Formation to 2443 Ma for the top of the Boolgeeda Iron Formation. Ages from a substantial number of intermediate stratigraphic levels are consistent with a semi-continuous deposition throughout the sequence, and do not support the presence of major non-depositional intervals.

Sequence 4 consists mostly of argillite and fine-grained volcaniclastic rock, dolomite, chert, banded iron-formation (BIF), sandstone, and felsic and mafic igneous rocks of extrusive and intrusive origin. Lower parts of Sequence 4 (Jeerinah Formation) show the greatest regional variation in lithology and thickness; middle and upper levels (Hamersley Group) are characterized by a more uniform stratigraphy.

Sequence 4 stratigraphy is summarized in Figure 9.8c. Over much of the central and eastern Pilbara craton, the lowest lithological division (the Jeerinah Formation) is about 150 m thick and is subdivided into: a lower division of nearshore shelf deposits (quartz sandstone and subordinate argillite and chert), and an upper division of 'deeper' offshore shelf facies (carbonaceous argillite, carbonate, chert, volcanic sandstone, and BIF). In contrast, the Jeerinah Formation in the south Pilbara ranges from 1 to 2 km thick and consists of a laterally variable assemblage of 'deeper' shelf deposits comprising massive basaltic lava, pillow lava, hyaloclastite, argillite, chert, dolomite, and sandstone.

The Jeerinah Formation of the Gregory Range inlier is up to a kilometre thick, and like that of the south Pilbara includes, in addition to fine-grained epiclastic rocks, a wide range of mafic and felsic volcanic and volcaniclastic rocks.

The Hamersley Group forms the remainder of Sequence 4 and consists mostly of 'deeper' marine-shelf deposits (BIF, argillite, carbonate, and fine-grained volcaniclastic rock) intruded by dolerite and acid igneous rock. The major stratigraphic units show relatively little regional variation in thickness or lithology (Trendall and Blockley, 1970; Trendall, 1983) and, in essence, reflect an alternation of BIF and rocks other than BIF (Trendall, 1990). The notable exception to the dominance of 'deeper' shelf facies is seen in the northeast Pilbara where the Marra Mamba Iron Formation is overlain by a shallow-marine carbonate, the Carawine Dolomite. Although relationships are somewhat equivocal, this formation appears to be, in part at least, a correlative of the Wittenoom Dolomite (Formation) (MacLeod and de la Hunty, 1966; Simonson, 1992).

Facies distributions indicate that the Jeerinah Formation was deposited during a major marine transgression, which extended across the Pilbara from the south and east (Fig. 9.10). On northern parts of the craton, transgressive shoreface sandstone (Woodiana Member) is overlain by 'deeper' shelf argillite, whereas in the south, submarine sedimentation and volcanism prevailed throughout deposition of the Jeerinah Formation.

With the exception of the Carawine Dolomite and parts of the Wittenoom Formation (Kargel et al., 1996), shallow-water sedimentary structures are absent from the Hamersley Group. This observation, combined with the dominance of argillaceous and laminated chemical rocks, has prompted several workers to suggest that much of the Hamersley Group was deposited in a sediment-starved basin or shelf, or on a shelf platform (Trendall and Blockley, 1970; Morris and Horwitz, 1983; Simonson et al., 1993). Arguments put forward by Simonson et al. (1993) indicate that the water depth in the main body of the Hamersley Basin was probably in the order of 100 m during deposition of the Wittenoom Formation, and that the regional palaeoslope during this time was from northeast to the south and west.

Mafic sills form a significant part of both the Jeerinah Formation and the Weeli Wolli Formation of the Hamersley Group. Within the Jeerinah Formation, sills constitute up to 60% of the stratigraphy, and are most abundant south of a line extending from the northwestern Sylvania Inlier to the northern Jeerinah Anticline. In the former area, Tyler (1991) noted that the sills are confined to the area west of the Fortescue River Fault. Similarly, in the Milli Milli Dome – Jeerinah Anticline area, there is a relatively sharp boundary between those Jeerinah Formation outcrops that contain mafic sills and those that do not. This boundary coincides with the line of separation between shallow-shelf and deeper shelf facies in Sequence 4 and may also be fault-controlled.

Mafic sills form about 60% of the Weeli Wolli Formation; at present it is unclear whether their distribution is related to the occurence of basement faults.

Trendall (1995) has argued that the thick felsic Woongarra Rhyolite, in the upper part of the Hamersley Group, is an intrusive sill rather than a volcanic unit interstratified with the enclosing BIF units. The available geochronlogy has not so far been able either to confirm or negate this hypothesis.

Structural controls on sedimentation and volcanism in the southern Pilbara

Blake (1984a, 1993) recognized a number of north-northeasterly trending faults in the north Pilbara which have exercised a strong control on Fortescue Group sedimentation and volcanism. The most important of these structures have been discussed in **Chapters 4–8**. In the south Pilbara, the major structural controls have been harder to identify, largely because most of the Fortescue Group in this area is concealed beneath the overlying Hamersley Group. Here, evidence relating to the nature of the syndepositional fault system is based upon four entities: firstly, the geometry of probable syn-Fortescue Group growth faults in the southern Pilbara; secondly, the local and regional variations in Fortescue Group stratigraphy; thirdly, the nature of the younger Capricorn Orgogen fault system; and fourthly, the use of analogues from Phanerozoic extensional settings.

Geometry of syn-Fortescue Group growth faults in the southern Pilbara

Most of the evidence for syn-Fortescue Group faulting in the south Pilbara comes from the western Sylvania Inlier and western Turner Syncline. In the Sylvania Inlier, pre-Bunjinah Formation units, which are up to 4 km thick in the Deadman Hill area, are not recorded north of the easterly trending Western Creek Fault (Tyler, 1991). The inferred syn-depositional, south-block-down movement on this fault was later reversed by compression associated with the Proterozoic Capricorn Orogeny (Tyler, 1991). Marked variation in upper Sequence 3 (Bunjinah Formation) and Sequence 4 (Jeerinah Formation) thickness and rock type occur across the north-northeasterly trending Fortescue River Fault, suggesting that this structure may have acted as a west-block-down growth fault during this time.

In the central south Pilbara, localized accumulations of basaltic breccia and dolerite are found within the Bunjinah Formation, immediately northwest of Mount Turner (Thorne et al., 1995). The succession, which is up to 150 m thick, is bound by steep, west-northwesterly to north-northwesterly trending faults. As is the case with the Sylvania Inlier structures, the original fault movement, which caused the central block to be downthrown, was reversed by later tectonism.

Variations in Fortescue Group stratigraphy

The Yule–Sylvania High (Blake, 1993) was a narrow upland area that extended from the southeast to the central west Pilbara and exerted a strong control on Sequence 2 sedimentation (**Chapter 5**). Although the influence of this feature is thought to have waned during Sequence 3 times, it still appears to have acted as an east-southeasterly trending boundary between the predominantly subaerial succession in the north Pilbara and the generally deeper marine succession of the south. This boundary also coincides with a marked change in the thickness of Sequences 3 and 4 stratigraphy, with some of formations in the south being up to five or six times thicker than the equivalent units in the north. The nature of these contrasts prompted Thorne (1990) to suggest that an easterly to east-southeasterly trending, south-block-down growth-fault system may have been active in the southern Pilbara at this time. This fault system controlled the main tectonic grain in the southern Pilbara and was reactivated during subsequent orogeny.

The Capricorn Orogen fault system

Figure 14.2 shows the main Palaeoproterozoic Capricorn Orogen and the faults and folds in the southern Pilbara, based on the work of Seymour et al. (1988), Thorne et al. (1991), Tyler et al. (1991) and Thorne and Tyler (1997a,b). In general, the degree of deformation is greatest in the south and east Pilbara but decreases northwards before dying out in the area of the former Yule–Sylvania High. Within this deformation zone the principal faults consist of steep, southward-dipping thrusts in the east and similarly dipping dextral wrench faults in the west. Both sets of structures are associated with generally upright to northward vergent folds. Throughout much of this zone, the faults and folds do not link up to form a continuous line, but instead form an overlapping, en echelon arrangement of east–west trending structures, which are offset northward and westward from the adjacent structure to the south. The southern limit of the exposed Pilbara Craton is also defined by a set of en echelon, east–west trending structures, although here the offsets are linked by steep northwesterly or north-northwesterly trending faults.

Although many of these faults and associated folds are seen only in the post-breakup Hamersley Group succession, their orientation and systematic en echelon distribution appears to reflect a strong basement control. The fact that this deformation zone coincides with the main area of Fortescue Group subsidence and volcanism suggests that many of these faults represent reactivated extensional structures, formed during syn-Fortescue Group rifting.

Analogues from Phanerozoic rift settings

Studies of Phanerozoic rift systems, particularly those in central and eastern Africa, have highlighted the fact that the rift zones are rarely made up of a single continuous fault line but instead consist of a series of in-line, or more commonly, en echelon half-graben segments (Rosendahl, 1987; Nelson, R. A. et al., 1992). Within the rift, the fault segments are separated by transfer zones which accommodate the extension of the individual fault sections.These exercise a strong control on sedimentation as they can act as topographic highs, lows, or ramps, depending on such features as fault-dip direction and degree of fault overlap (Gawthorpe and Hurst, 1993). Similarly, the individual rift segments may be arranged in-line or else form an en echelon geometry, with each segment separated by unfaulted or faulted jumps, or fault-connected offsets and passes (Nelson, R. A. et al., 1992).

A striking feature of continental rift systems is that, on a regional scale, coeval extension zones are commonly developed at a high angle to one another (e.g. Bishop and Trendall, 1967; Rosendahl, 1987, fig. 1c). Examples of this include the Nyanza rift unit, adjacent to the Gregory Rift Zone, and the Upemba–Mweru–Sumbu Chisi (UMSC) fault system adjacent to the Tanganyika rift zone. The latter example is likely to be a close analogue for the early stages of syn-Fortescue rifting, not only in terms of the scale of the rift system and the relative orientation of the major faults, but also in regard to the regional topography resulting from the interplay of the various structural elements (Fig. 14.3).

The Fortescue Group as a continental flood basalt province (CFBP)

The Fortescue Group has been compared with continental flood basalt terranes by Blake and Groves (1987), and was specifically described as a continental flood basalt sequence by Nelson, D. R. et al. (1992). Arndt et al. (1991) also noted a ‘clear affinity‘ of flows of the Mount Roe Basalt and Kylena Formation with those of young CFBPs, and Meakins (1990) directly described the Fortescue Group as ‘the oldest documented flood basalt terrain in the world’. The application of such a label implies firstly that there does exist some clearly defineable class of volcanic field called a CFBP, secondly that the attributes of the Fortescue Group identify it as a member of that class, and thirdly that an understanding of its geodynamic significance follows from this identification. Here we address these points in turn, so that the following discussion deals with definition of a CFBP and whether the Fortescue Group fulfils the requirements of that definition.

There is no published consensus on the defining characters of a CFBP (see Swanson, 1989), and most authors who have used the term have done so without definition, relying on a list of examples to communicate their concept of its meaning. A trial definition, based on two examples universally accepted as CFBPs — the Deccan Traps and Columbia River Basalt Group — might run as follows: a CFBP is a region, typically over 10^5 km^2 in area, which was intermittently flooded, at short (?10^{3-5} years) and rather regular intervals, with individually voluminous, relatively thin, regionally extensive, basaltic lava flows erupted in rapid succession, so that within a

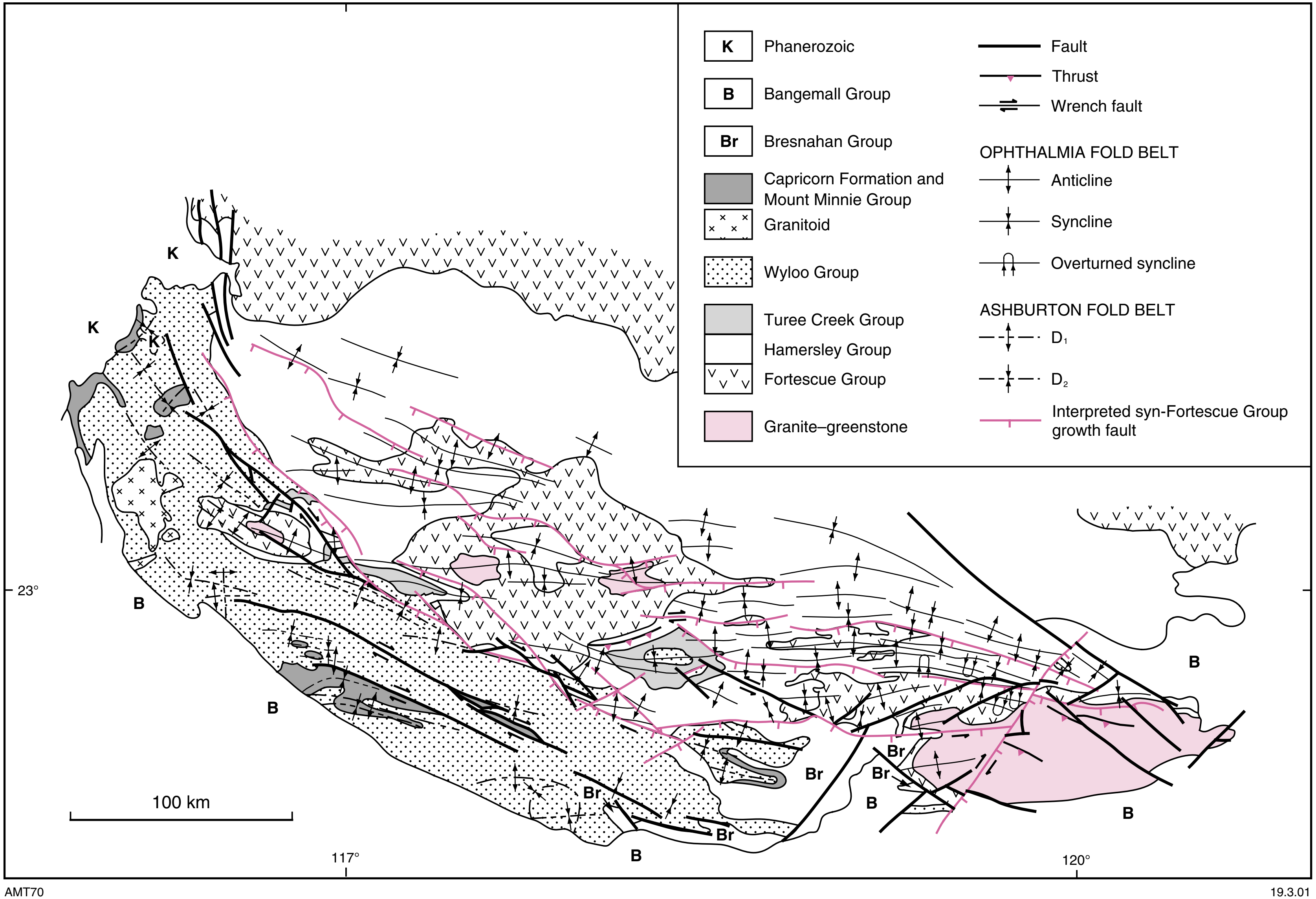
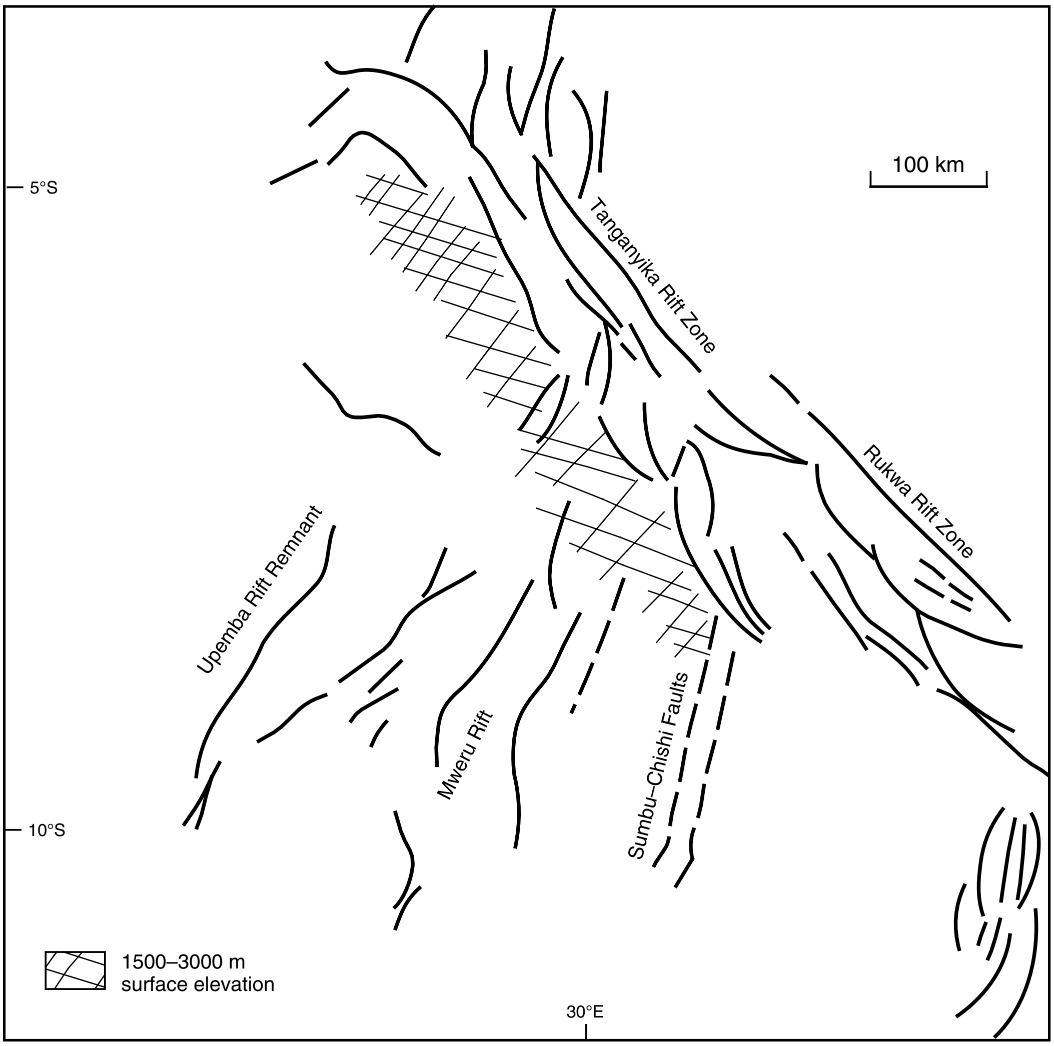
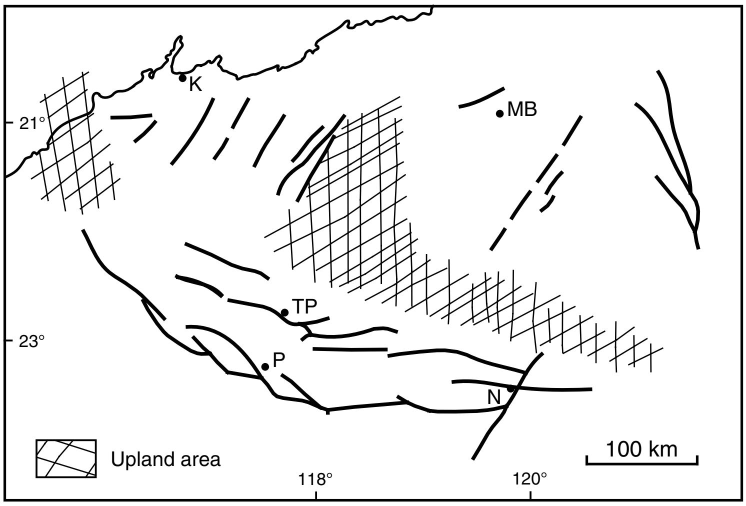


Figure 14.2. Simplified geological map of the southern Pilbara showing the main Capricorn Orgogen structural features and the probable location and orientation of the major syn-Fortescue Group faults. Modified after Tyler (1991)



a.



b.

Figure 14.3. (a) Structural architecture of the Tanganyika Rift Zone showing the orientation of principal faults and the distribution of upland areas (modified after Rosendahl, 1987). (b) Interpreted structural architecture of the Pilbara during syn-Fortescue (Sequence 2) rifting showing the location of major faults and upland areas. Structural data from the northern Pilbara based on Blake (1993). Towns of Karratha (K), Marble Bar (M), Tom Price (TP), Paraburdoo (P), and Newman (N) are shown for reference

geologically short period (typically $<10^7$ years) a pile of (typically $>10^2$) superimposed flows, with a relatively minor interflow sedimentary component, attained an aggregate thickness in excess of a kilometre.

Because they were chosen as 'type' examples (see BVSP, 1981, p. 30), the Deccan Traps and the Columbia River Basalt Group both fit such a definition well. In both these provinces, which are respectively end-Cretaceous and Tertiary in age, the flows are still, on a regional scale, only mildly deformed, and apart from uplift, erosional incision, and normal diagenesis, remain much as they were immediately following eruption. There can be no doubt that each results from related multiple effusions, which can together be considered as a single volcanic event, and that these two events have sufficient common features to justify a single identifying name.

It is when the number of examples of CFBPs is extended that differences appear between the various published lists; for example Wilkinson and Binns (1977), BVSP, (1981), Thompson et al. (1984), Blake and Groves (1987), Yoder (1988), Wilson (1989), White and McKenzie (1989), Hutchinson et al. (1990), and Blake (1993). There at least four reasons for this:

- a number of the papers cited are concerned with particular aspects of petrogenesis, rather than questions of definition;
- recognition of older CFBPs becomes more difficult as such parameters of the volcanism as age limits, areal extent, and magma volume become harder to establish;
- the possibility of identifying very old CFBPs has only been appreciated quite recently — Nelson, D. R. et al. (1992) have pointed out that the benchmark BVSP (1981) compilation recognized only one Precambrian CFBP just over a decade ago;
- the absence of an agreed definition means that differences between the concepts of various authors remain hidden, rather than being exposed and discussed.

The following list of possible CFBPs, in age order of initial volcanism, is a composite one taken from the compilations of the various authorities cited above:

CFBP	*Age (Ma)*
Snake River, USA	17
Columbia River, USA	17.5
Ethiopia, Africa	43
Deccan Traps, India	60
North Atlantic	65
Parana–Etendeka, Brazil/SW Africa	135
Karoo (South Africa)	206
Lebombo–Nuanetsi, Lesotho/S Africa	193
Ferrar–Kirkpatrick, Tasmania/Antarctic	179 ± 7
Siberia, Russia	248
Antrim Plateau, northern Australia	600
Keeweenawan, USA	1096
Coppermine River, Canada	1200
Ventersdorp, South Africa	2720
Fortescue Group, Australia	c. 2770

The identification of all those listed as CFBPs raises difficulties with the rigid application of the trial definition above; the difficulties include at least the five following points:

- the 'continental' requirement;
- the necessity of a 'short' time interval;
- the name 'basalt';
- the requirement for eruption;
- the associated sedimentation.

The 'continental' requirement: Although we discuss below the possible association of flood basalt provinces with mantle plumes, it is necessary to anticipate that association here in raising the significance of the term 'continental'. According to some authors (e.g. Morgan, 1971, 1981; Richards et al., 1989) flood basalt provinces are associated with the rise of plume heads to the base of the lithosphere. Richards et al. (1989) specifically make the point that plume heads may be located beneath either oceanic or continental crust, and that the CFBPs that result in the latter case are the genetic analogues of large oceanic basaltic plateaus, such as Ontong Java.

A 'short' time interval: Rapid eruption was implicit in early concepts of flood or plateau basalt volcanism, but the idea was qualitative rather than quantitative, and based on the impression from typical flow sequences that each basalt eruption closely followed the preceding one, with little intervening opportunity for weathering or erosion. Following relatively recent increases in geochronological precision it seems to be emerging that as more measurements are made, the total time taken for the emplacement of some of the larger provinces listed becomes less (see especially Baksi, 1989, for Columbia River; Renne and Basu, 1991, for Siberia; Duncan and Pyle, 1988, for Deccan). Few of the Precambrian examples are sufficiently tightly constrained for confidence that comparable rates of magma effusion were achieved, and a requirement in any definition for rapid emplacement automatically makes identification of Precambrian examples difficult.

'Basalt': As far as magma composition is concerned the name 'basalt' is not only a simplistic component of the term CFBP, but to some extent misleading if it is taken to exclude other magma types. Both the Ventersdorp Supergroup and Fortescue Group consist predominantly of basaltic andesite, and the validity of the name CFBP is arguable unless the B is taken to stand for basaltic andesite. Like Parana–Etendeka, and others, but unlike Columbia River and the Deccan, they also have a significant component of felsic rocks.

The requirement for eruption: The name CFBP was restricted in the trial definition above to areas of erupted basaltic rocks. But Thompson et al. (1984, table 2) accompany their list with the following very general definition: '(a) major, basalt-dominated, fissure-erupted, plateau-forming lava fields; and (b) dyke and sill swarms which appear, from their sizes and tectonic situations, to have been emplaced beneath ancient plateau basalt lava fields (now eroded).' Sill swarms constitute a major, and largely unmeasured, component of many CFBPs (Karroo, Ferrar, and Tasmanian dolerites). In Figure 14.4 the Hart Dolerite of Western Australia, which has apparently not been listed elsewhere as a CFPB, has been included to illustrate this point. White and McKenzie (1989) include not only shallow intrusions but also underplating in the lower crust in estimating total magma volumes of CFBPs.

Associated sedimentation: The name CFBP excludes a very significant component of many examples: the associated sedimentary rocks. If all the mafic rocks were removed from the two 'type' examples — Columbia River and the Deccan — little other material would remain. But if the same were done with the Lesotho Karroo, Parana and Siberia, significant sedimentary basins would remain. Any complete definition of a CFBP must take this into account.

Finally, a sixth problem of definition arises from the proposed association of CFPBs with mantle plumes, which has already been referred to. In commenting on the problem of CFBP definition, Thompson et al. (1984, p. 159) justify the generality of their definition, already given above, with the comment: 'Perhaps the best way to look at continental flood basalts is as a group of igneous rocks whose larger representatives are obvious, but whose smaller examples grade imperceptibly into other types of continental magmatism. This is a typical arbitrary scientific carve-up of a range of natural phenomena.' The variability of CFBPs is emphasized by the comment of Cox (1988, p. 239) that 'every continental flood basalt province has its own particular flavor'.

If mantle plumes and CFBPs have a close genetic relationship then it would be expected that CFBPs would either be very large or would not be formed. A CFBP–plume association would not lead to the expectation that there should be an imperceptible gradation into other (non-CFBP) volcanic phenomena, unless the plume concept includes a corresponding gradation in plume sizes. This question has not been seriously addressed in the relevant literature.

In summary, although the most typical CFBPs, such as Columbia River and the Deccan, are spectacular volcanic features which seem to form a uniquely identifiable class, there are a number of serious difficulties which preclude exact definition of a CFBP. Partly for this reason there is now a preference for regarding CFBPs as a subclass of large igneous provinces (LIPs), which embrace all 'voluminous emplacements of predominantly mafic extrusive and intrusive rock whose origins lie in processes other than "normal" seafloor spreading' (Coffin and Eldholm, 1992). Coffin and Eldholm (1991) have classified and catalogued LIPs into six types: continental flood basalts; volcanic passive margins; oceanic plateaus; submarine ridges; seamount groups; and ocean basin flood basalts.

Since the depository of the Fortescue Group embraced both continental and submarine environments, including a passive continental margin, it is preferable here to compare it with the broader category of LIPs, rather than the narrow one of CFBPs. In the context of such a comparison the key quantifiable parameters of any LIP are duration (in m.y.) and volume, which combine to give the derivative parameter of magma extrusion rate; the rate referred to here is the integrated rate over the total lifespan of the LIP, and note the short-term rate during extrusion of single large flows. While it can be argued that volume is derived from the separable parameters of area and thickness it is equally true that, given the probably constant rheology of mafic magma throughout Earth history, the correlation between area and volume of LIPs is sufficiently close for volume to be taken as a single-size index.

Figure 14.4 is a plot of duration against volume for selected LIPs, with a bias towards CFBPs. Although Coffin and Eldholm (1992) maintain that CFBPs are 'the best studied of all LIPs', they also emphasize that accurate basic information on LIP size and duration of emplacement are lacking for almost all their catalogued examples. A comparison of the Fortescue Group with typical CFBPs is thus difficult, but within the limitations of the data the following generalizations are possible:

- The total areal extent, thickness, and volume of mafic lavas of the Fortescue Group comfortably exceed all normally accepted minima for CFBPs. We estimate a total volume of mafic magma extrusion of about 2.5×10^5 km^3 for the entire group, which, for example, far exceeds the 1.8×10^5 km^3 of the Columbia River Basalt Group (Fig. 14.4).
- On the other hand, the Fortescue Group had a longer duration than many younger CFBPs. The total time from the base of the Mount Roe Basalt to the top of the Jeerinah Formation is about 145 m.y. (2775–2629 Ma). But the age of the top of the Maddina Formation/Bunjinah Formation is not well fixed, and could be as old as c. 2700 Ma, to give a duration for all of the three main pulses of mafic flow emplacement (Mount Roe Basalt, Kylena Formation/Boongal Formation, and Maddina Formation/Bunjinah Formation) of 75 m.y., or even less. Although the emplacement of these three units may have occupied less than 1 m.y. the total volume of each (respectively 72, 68, and 110×10^5 km^3) precludes each from being regarded as a separate LIP in its own right.
- The Fortescue Group has a significant sedimentary component (e.g. Hardey Formation), which differentiates it from, for example, Columbia River and the Deccan. However, the non-volcanic component of the Fortescue Group is substantially less than, for example, the Lesotho Karroo.
- The Fortescue Group is chemically unlike many younger, 'typical', examples of CFBPs. Most of its stacked mafic flow sequences consist of basaltic andesites rather than basalts.
- Although the depository of the Fortescue Group is floored throughout by continental crust, only the stacked flows of the northern sub-basins have the subaerial flows typical of CFBPs. The south Pilbara sub-basin is clearly closer to the passive continental margin sub-class of Coffin and Eldholm's (1991; 1992) LIPs.

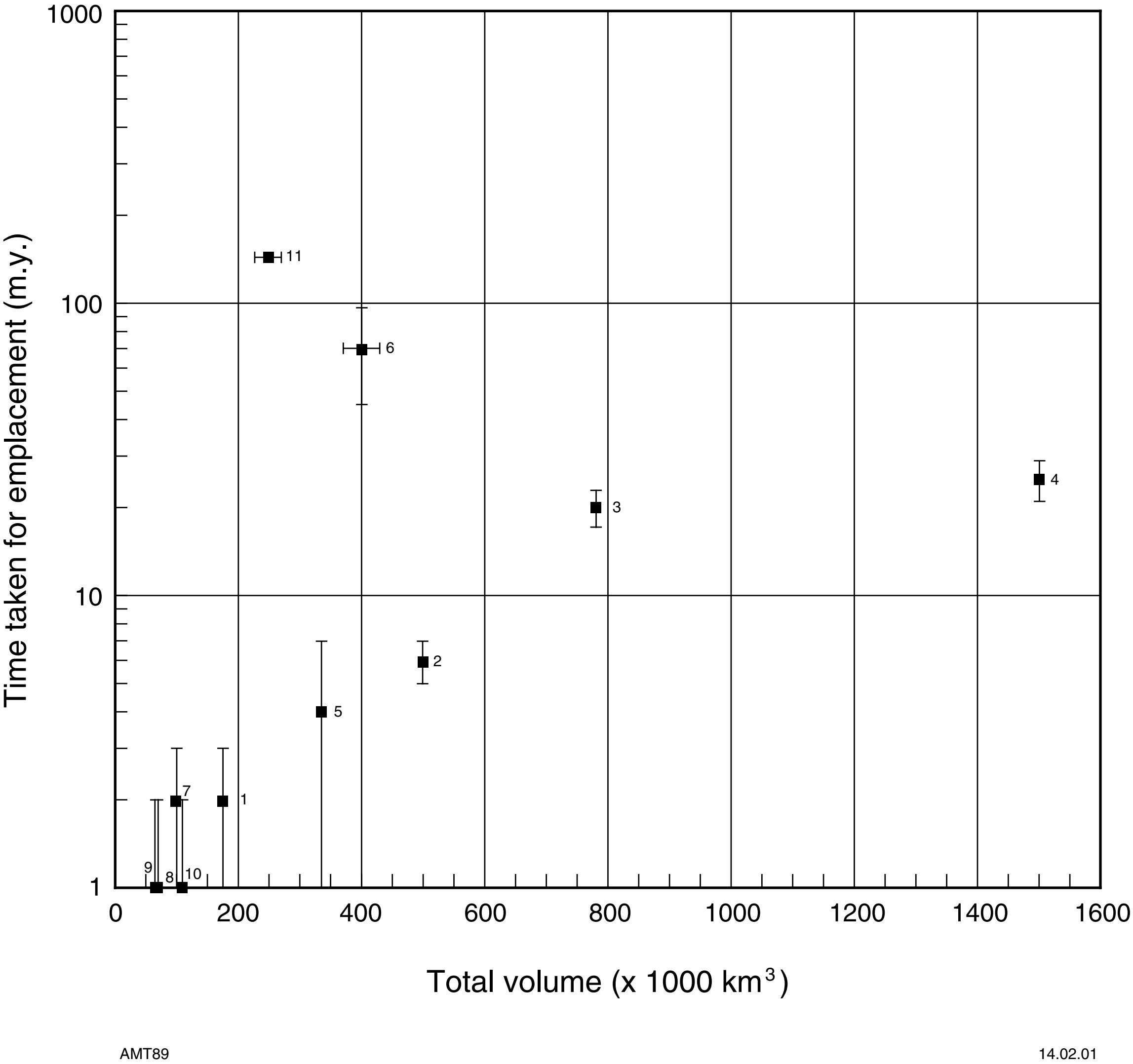


Figure 14.4. Plot of total volume of mafic lava (x, cubic kilometres) against time taken for emplacement (y, millions of years) for some major continental flood basalts and the Fortescue Group. Points are numbered as follows: 1 = Columbia River; 2 = Deccan Traps; 3 = Parana; 4 = Karroo; 5 = Siberia; 6 = Antrim Plateau; 7 = Keeweenawan; 8 = Fortescue Group, Mount Roe only; 9 = Fortescue Group, Kylena–Boongal only; 10 = Fortescue Group, Maddina–Bunjinah only; 11 = All Fortescue Group. Taking 1, 2, 3, 4 and 5 as 'classic' flood basalt areas, it is possible to speculate that there may be an upper regional magma-generation rate of about 100 km^3 per million years, controlled by the rate of radioactive heat generation within the Earth, which constrains the positions of these examples on this diagram. In that context, as elaborated in the text, the total volume of Fortescue Group mafic lavas falls within the range of flood basalt provinces, but the time taken for extrusion is unusually long

In summary, although several of these points suggest significant differences from the more typical younger examples, the Fortescue Group is sufficiently close to many areas accepted as CFBPs that we need to examine whether an understanding of its origin follows from acceptance of this identity.

The origin of CFBPs: the continental rift and plume models

The evolution of Phanerozoic CFBPs have been explained by two types of mantle plume model (see discussion in Blake, 1993). In the first of these, here referred to as the Richards et al. model (Richards et al., 1989, Campbell and Griffiths, 1990), lithospheric thinning, rifting, and associated flood basalt volcanism is caused by an ascending mantle plume head impinging on the base of the underlying lithosphere. In the second model, here labelled the White and McKenzie model (White and McKenzie, 1989), flood basalts and large volumes of intrusive igneous material form as a mantle plume rises passively beneath the stretched and thinned lithosphere of major rift zones. In both models, basalts are produced by adiabatic decompression melting of anomalously hot athenospheric mantle, whereas basaltic underplating of the lithosphere normally results in crustal uplift, rather than subsidence.

At present, neither the Richards et al. nor the White and McKenzie models are entirely satisfactory in explaining Fortescue Group evolution. The principal objections to these models are:

- As already noted, the Fortescue Group contains not one but three main basaltic units, separated by sedimentary deposits of the Hardey and Tumbiana Formations. This distribution is made more complex by the fact that in the southern Pilbara, the Pyradie Formation, which is the lateral equivalent of the Tumbiana Formation, is characterized by submarine high-Mg basalt flows and local komatiite.
- Flood basalt volcanism probably continued, in a punctuated fashion, over a period of at least 75 m.y., and possibly much longer.
- The history of uplift during Fortescue Group deposition is inconsistent with either plume model.

In order to explain the first two objections, the Richards et al. model would have to argue the unlikely case of three separate mantle plumes being developed beneath the Pilbara during Fortescue Group evolution. The White and McKenzie model faces a similar difficulty since it would require a stationary Pilbara Craton to undergo a puntuated history of rifting above a single long-lived (~70 m.y.) plume. In addition, both plume models predict that the main periods of mafic volcanism should be accompanied by significant regional uplift. Although this relationship appears valid for the Mount Roe Basalt, it does not apply to the other basalt units; their depositional history indicates that basalt extrusion either just kept pace with, or was rather less than, the rate of basin subsidence.

Many workers (e.g. Windley, 1984; Bickle, 1986, 1990) argue that because late Archaean mantle temperatures were high, substantial volumes of basaltic melt would have formed by decompression melting beneath stretched and thinned lithosphere in most tectonic settings. Anomalously hot mantle plumes such as those envisaged for the Richards et al. and White and McKenzie models are therefore not a prerequisite for large-volume basaltic volcanism (Blake, 1993). Discrete periods of lithospheric extension alone, related to continental breakup, could account for the history of flood basalt volcanism observed in the Fortescue Group.

Tectonic setting

The tectonic setting for the Fortescue Group and the overlying Hamersley Group is discussed in terms of the four depositional sequences described earlier in this chapter. Under this scheme, initial continental sedimentation and volcanism (Sequences 1 and 2) took place in isolated, faulted sub-basins, and was accompanied by intermittent regional-scale uplift in the north and central Pilbara. The separate sub-basins coalesced during Sequence 3 times when regional subsidence, accentuated by further normal faulting and/or tilting in the south, resulted in a change to coastal and deeper shelf volcanism and sedimentation. Further subsidence at the begining of Sequence 4 times resulted in a major marine transgression and the establishment of 'deeper' marine-shelf sedimentation and volcanism over the entire Hamersley Basin.

Studies of Phanerozoic sedimentary basins suggest that the Hamersley Basin formed as a result of one of the following mechanisms (Allen and Allen, 1990):

- Lithospheric stretching,
- Flexure on continental and oceanic lithosphere,
- Strike-slip or megashear-related deformation.

The style and history of Fortescue and Hamersley Group sedimentation and volcanism, the large size of the depository, and general absence of associated compressive deformation or major strike-slip faulting make it unlikely that the Hamersley Basin formed by either the second or third mechanisms. In contrast, Hamersley Basin development can be adequately explained using a lithospheric stretching model, in which early intracratonic rifting of a more extensive Pilbara Craton led to crustal separation and ocean-basin formation. Viewed in this light, the Hamersley Basin represents a composite suite of superimposed, interrelated basins formed as the rifted Pilbara crust evolved to a subsiding, passive continental margin.

Previous rift-to-passive margin models

Blake (1990, 1993), Blake and Groves (1987), and Blake and Barley (1992) have interpreted the Fortescue Group, up to the base of the Jeerinah Formation, as the rock record of a two-phase continental breakup involving two differently oriented extension events. Initial crustal extension was oriented west-northwest – east-southeast and resulted in the formation of large volumes of tholeiitic basalt (Mount Roe Basalt) followed by block faulting and the development of extensional intracratonic sedimentary basins and associated felsic and mafic volcanism (Hardey Formation). Basement uplift pre-dated and post-dated mafic volcanism and an ocean basin may have formed to the west of the present craton (Blake, 1993). The later eruption of large volumes of dominantly mafic volcanic rock (Kylena to Maddina Formation) resulted from the development of a younger east-southeasterly trending rift along the present southern craton margin, although reactivation of earlier northeasterly and north-northeasterly trending growth faults suggests that west-northwest – east-southeast extension continued to some extent (Blake and Barley, 1992). Continued east-southeast rifting resulted in the formation of an ocean basin south of the present-day craton margin. Blake and Barley (1992) and Blake (1993) interpret the Jeerinah Formation and lower part of the Hamersley Group (Marra Mamba Iron Formation to base of Mount Sylvia Formation) as having been deposited on a thermally subsiding continental shelf bordering this ocean (cf. Horwitz, 1980; Morris and Horwitz, 1983).

Tectonic model used in this Bulletin

We are in broad agreement with previous workers such as Blake and Groves (1987), Blake and Barley (1992), and Blake (1993) who interpret the Fortescue Group and lower

Hamersley Group in terms of a rift-to-passive margin tectonic model. In contrast to these workers, however, we propose that rifting of the Pilbara Craton was not a two-stage process involving two differently oriented extension regimes, but occurred instead as one protracted breakup event. Principal arguments against the 'two-stage, two rift' model may be summarized as follows:

- There is little evidence for syn-Fortescue Group west-northwest – east-southeast extension in the southern Pilbara. All north-northeasterly trending dykes in the southern Pilbara either pre-date or post-date the Fortescue Group; there are none which can be correlated with the syn-Fortescue (cf. Blake, 1993) Black Range Suite of the northeast Pilbara. Similarly, only one north-northeasterly trending syndepositional fault, the Fortescue River Fault in the Sylvania Inlier (Tyler, 1991), has been reported from the southern Pilbara. Here, most thickness and facies changes are apparently related to a west-northwesterly trending system of en echelon east–west growth faults; for example, the Western Creek Fault in the Sylvania Inlier (Tyler, 1991). Palaeocurrent data for the Hardey Formation (Fig. 5.16) are also inconsistent with a northeasterly trending growth-fault system in the southern Pilbara and appear instead to reflect more of an east–west tectonic control (cf. Blight, 1985).
- The distribution of north-northeasterly trending, syn-Fortescue Group dykes and growth faults in the northern Pilbara do not support the concept of an early rift axis to the west of the present craton margin. The north-northeasterly trending Black Range Suite, which forms a possible feeder system for basalts of Sequences 1 and 2, is confined to the northeast Pilbara. No syn-Fortescue Group feeder dykes have been recognized in the northwest Pilbara, an area which is closer to the proposed rift axis and which would be expected therefore to have undergone significant crustal extension. Similarly, the major north-northeasterly trending growth faults and faulted sub-basins are not concentrated in the far west Pilbara, but extend across the entire northern Pilbara Craton, a horizontal distance of over 500 km. This figure would require the north-northeasterly trending rift zone to have a total width in excess of 1000 km, compared with typical values of 100–400 km for Phanerozoic African rift zones (cf. Rosendahl, 1987).
- Thickness data for Sequences 1 and 2 sedimentary and volcanic rocks in the northwest Pilbara do not indicate proximity to a major rift axis. The combined thickness of Sequences 1 and 2 rocks in this area is generally in the order of 1–1.5 km, compared with the 4–5.6 km thickness recorded for equivalent units in the northeast Pilbara.

While the preceeding arguments question the evidence for a major rift axis west of the present Pilbara Craton margin, it is clear from the work of Blake (1984a, 1993) that lower Fortescue Group volcanism and sedimentation in the northern Pilbara was associated with a limited amount of west-northwesterly directed crustal extension. In the following tectonic model, we assert that this extension was not associated with an early, separate north-northeasterly trending rifting event, but rather, was coeval with the development of a major west-northwesterly trending rift zone that formed along the southern Pilbara margin during deposition of Sequences 1–3. In this model, Pilbara Craton rifting took place as a protracted event, during which the principal extension was directed along a north–south or north-northeast – south-southwest line. The north-northeasterly trending extensional fault and dyke system in the northern Pilbara formed at a high angle to the principal rift axis in a manner analagous to a failed rift arm. Possible analogues for this model are seen in the Phanerozoic African Rift System (e.g. Rosendahl, 1987, fig. 1c; Nelson, R. A. et al., 1992, fig. 6a); in particular, geometric relationships are similar to those described for the UMSC fault system and the adjacent Tanganyika rift zone (see **Structural controls on sedimentation and volcanism in the southern Pilbara** above).

Three principal stages are recognized in our tectonic model for the Fortescue Group and lower Hamersley Group: (a) early crustal extension, (b) breakup of the Pilbara Craton, (c) post-breakup shelf subsidence.

Early crustal extension (Sequences 1 and 2)

Sequences 1 and 2 were deposited on a granite–greenstone terrain, considerably larger in extent than the present day Pilbara Craton. Pre-Mount Roe Basalt sedimentary rocks are typically thin and reflect early subaerial degradation of the landscape, whereas the thickest deposits, such as those in the Marble Bar and Paraburdoo areas, probably accumulated in small, intermontane basins.

Initial syn-Fortescue Group crustal extension, at about 2770 Ma, resulted in the formation of a major west-northwesterly trending rift system along the present-day southern Pilbara margin. In the north Pilbara, north-northeasterly trending normal fault and dyke orientations reflect coeval west-northwesterly directed extension in areas north of main rift axis (cf. the Gregory–Kavirondo rift zone of the East African System). Crustal thinning, and the associated decompression melting of the underlying mantle, gave rise to the extrusion of about 1 km of mostly subaerial basalt (Mount Roe Basalt) in the northwest, northeast, and southwest Pilbara. The general lack of interbedded sedimentary rocks, available geochronological data, and comparisons with similar Phanerozoic flood basalt successions (Fig. 14.4) suggest that most of the lava pile was extruded in a relatively short period, perhaps 1–5 m.y.

The post-Sequence 1, pre-Sequence 2 interval was characterized by local block faulting, uplift, and erosion of the subaerial basalt pile. In the southern Pilbara, Sequence 2 (Hardey Formation) depositional environments evolved from braided fluvial to deltaic. During this time, the principal sediment transport direction was westward, away from a basement high along the rift shoulder, and parallel to the main fault elements in the rift. In northern parts of the Pilbara, north of the main rift axis, Hardey Formation sedimentation and volcanism was strongly influenced by the subordinate north-northeasterly trending growth-fault system, with its dominant west-block-down sense of movement. Here, principal depositional environments were braided fluvial, alluvial fan, and lacustrine, with fluvial transport directions being mostly parallel to the north-northeast structural grain.

Further crustal extension occurred during middle to late Hardey Formation times and resulted in the development of three major felsic volcanic provinces, in the northwest and northeast Pilbara, and in the Gregory Range area. Here, several of the principal volcanic centres appear to have developed in association with the main north-northeasterly trending faults. In the south Pilbara, the main rift zone underwent major subsidence at this time and submarine deltaic conditions were established over much of the region.

Breakup of the Pilbara Craton (Sequence 3)

Major crustal extension caused further rifting and breakup of the Pilbara Craton during Sequence 3 times. South to south-southwesterly directed crustal extension resulted in widespread normal faulting and accompanying submarine mafic volcanism in the southern Pilbara. Here, the thickest accumulation of submarine basalt occurred south of a west-northwesterly trending zone of en echelon east–west faults which extended from the northern Sylvania Inlier to north of the Jeerinah Anticline. Crustal separation was probably synchronous with Pyradie Formation deposition (c. 2715 Ma), the presence of komatiite in this unit pointing to the decompression melting of large volumes of the underlying mantle at this time.

On the north Pilbara, some folding and erosion of Sequence 2 rocks preceeded deposition of Sequence 3. The north-northeasterly trending fault system, which had been a major influence on Sequence 2 deposition, became largely inactive during Sequence 3 times. Instead, sedimentary and volcanic facies reflect the presence of a west-northwesterly trending coastline, adjoining a southward-deepening shallow-marine shelf. As is the case with Sequence 1, the principal mafic volcanic units of the north Pilbara (Kylena and Maddina Formations) comprise a monotonous succession of subaerial to shallow marine lavas that are closely similar to those that exist in Phanerozoic flood basalt provinces.

Post-breakup shelf subsidence (Sequence 4)

The Jeerinah Formation and overlying Hamersley Group were deposited on a largely sediment-starved, thermally subsiding passive continental margin. On the north Pilbara, initial lithospheric cooling and subsidence took place at around 2690 Ma and resulted in the deposition of a transgressive marine succession (Jeerinah Formation) that extended across the region toward the north and northwest. In the south, early subsidence was accompanied locally by submarine basalt extrusion, suggesting further crustal extension may have taken place locally, after the main separation event was completed. The overlying Hamersley Group was deposited on the continental shelf during a period of further lithospheric cooling and subsidence.

Concluding statement

While our preferred general model is consistent with current knowledge, we emphasize finally that systematic study of the Fortescue Group is still at an early stage. The vast areal extent and thickness of the group, which represents the preserved physical products of a complex succession of ancient tectono-magmatic events, means that the geological resources devoted to its study since its first systematic mapping in the 1960s have been grossly inadequate to provide a complete appreciation of its significance in terms of the overall magmatic history of the Earth. In comparison with a relatively recent modern large igneous province such as the North Atlantic Tertiary Province, the Fortescue Group has probably received two or three orders of magnitude less attention. But only if such early provinces as the Fortescue Group are as well understood as more recent examples can a properly informed appraisal be made of secular changes in the Earth's tectono-magmatic development. We hope that augmenting and centralizing knowledge in this Bulletin will both stimulate a wider interest in the group and provide a foundation for the more extensive specialized studies which are needed.

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Appendix 1

Formal and informal stratigraphic names that have been applied to Fortescue Group rocks

Bamboo Creek Porphyry

Proposed by: Noldart and Wyatt (1962, p. 89).

Lithology/thickness: Quartz feldspar porphyry or 'granite porphyry'/not stated.

Geographic feature: Bamboo Creek, southwest YARRIE.

Type section or area: Named from the largest development on Bamboo Creek. Original reference 310384.

Upper and lower limits: 'Intrusive into the lower sedimentary beds of the Little de Grey synclinal and Meentheena basinal structures'. An extrusive origin is favoured north of Nullagine.

Original correlations: By implication, with a similar unit (the Spinaway Porphyry of Lipple (1975)) outcropping 4.8 km north of Nullagine.

Status and comments: The unit is here referred to as the Bamboo Creek Member of the Hardey Formation. It outcrops on MARBLE BAR, NULLAGINE, and YARRIE, and includes the Spinaway Porphyry.

Baramine Volcanic Member

Proposed by: Hickman et al. (1983).

Lithology/thickness: 'A basal unit of banded chert and dolomite is overlain by carbonated tuff, breccia, and agglomerate of felsic and intermediate composition. Carbonated mafic and felsic lavas are dispersed throughout the member.' The unit also includes dolerite sills and dykes/250 m.

Geographic feature: Baramine Homestead.

Type section or area: 'type area 5 km southeast of Baramine Homestead'.

Upper and lower limits: 'conformably overlies the Pearana Basalt' Upper limit not defined, but presumed to be upper limit of volcanic content within Lewin Shale, of which BVM is a member.

Original correlations: Nallanaring Volcanic Member.

Status and comments: The unit is here referred to as the Baramine Member of the Jeerinah Formation. Use of this name is confined to descriptions of the Gregory Range.

Beatons Creek Conglomerate

Proposed by: Noldart and Wyatt (1962, p. 82–85).

Lithology/thickness: '. . . a very coarse conglomerate. . . ' /not stated but acknowledged to be lenticular.

Geographic feature: 'the Beatons Creek locality, northwesterly of Nullagine township (Ref. 299269)'.

Type section or area: 'their largest development' is as above.

Upper and lower limits: Rests unconformably on 'basement rocks'; upper boundary not clearly defined.

Original correlations: Forms the basal beds of the Nullagine succession 'in most localities'.

Status and comments: An obsolete name. It is equivalent to part of the lower Hardey Formation in the Nullagine area.

Bellary Formation

Proposed by: Thorne et al. (1991).

Lithology/thickness: Mudstone, siltstone, and sandstone interbedded with smaller amounts of conglomerate, basalt, basaltic breccia and tuff/400 m minimum.

Geographic feature: Bellary Creek.

Type section or area: Core of the Bellary Dome, immediately north of Paraburdoo.

Upper and lower limits: Base is not exposed; conformably overlain by the Mount Roe Basalt.

Original correlations: None.

Status and comments: Represents the lowermost Fortescue Group unit on TUREE CREEK.

Boongal Pillow Lava Member

Proposed by: de la Hunty (1965, p. 9–10).

Lithology/thickness: Pillow lava, ophitic basalt, and 'intercalations of blue-grey carbonate rocks and soft grey chloritic rocks with dark elongated pellets'/about 2500 feet (760 m).

Geographic feature: Boongal Well.

Type section or area: Not specified.

Upper and lower limits: Overlies Hardey Sandstone. Overlain by Pyradie Pyroclastic Member.

Original correlations: Kylena Basalt of main outcrop area.

Status and comments: Upgraded to formation status by Thorne et al. (1991). It is recognized throughout the southern Hamersley Basin.

Boongal Formation

Proposed by: Thorne et al. (1991).

Lithology/thickness: Massive mafic lava, pillow lava, hyaloclastite, and mafic tuff/about 1000 m.

Geographic feature: As for Boongal Pillow Lava Member.

Type section or area: Not specified.

Upper and lower limits: Overlies Hardey Formation; overlain by Pyradie Formation.

Original correlations: None.

Status and comments: Equivalent to the Boongal Pillow Lava Member of de la Hunty (1965). Unit was raised to formation status because it forms a major lithological unit that can be recognized throughout the southern Hamersley Basin. It is regarded as a correlative of the Kylena Formation of the main outcrop area.

Bunjinah Pillow Lava Member

Proposed by: de la Hunty (1965, p. 10).

Lithology/thickness: Pillow lava, vesicular basalt, and 'volcanic breccia'/about 2500 feet (760 m).

Geographic feature: Bunjinah Spring.

Type section or area: Not specified. Good exposures at Fish Pool, Bunjinah, and Jeerinah Anticline.

Upper and lower limits: Overlain by Jeerinah Formation. Overlies Pyradie Pyroclastic Member.

Original correlations: Maddina Basalt of Chichester Range.

Status and comments: Raised to formation status by Thorne et al. (1991).

Bunjinah Formation

Proposed by: Thorne et al. (1991).

Lithology/thickness: Massive mafic lava, pillow lava, vesicular lava flows, hyaloclastite, and tuff/900 m.

Geographic feature: Bunjinah Spring.

Type section or area: As for Bunjinah Pillow Lava Member.

Upper and lower limits: Conformably overlies Pyradie Formation; conformably overlain by Jeerinah Formation.

Original correlations: None.

Status and comments: Equivalent to the Bunjinah Pillow Lava Member of de la Hunty (1965). It was raised to formation status because it forms a major lithological unit that can be recognized throughout the southern Hamersley Basin. The Bunjinah Formation is regarded as a correlative of the Maddina Basalt of the northern Hamersley Basin.

Cliff Springs Formation

Proposed by: Williams (1968, p. 28).

Lithology/thickness: 'Predominantly tuffaceous, but with clastic rocks and subordinate calcareous rocks . . .'/up to 500 feet (152 m).

Geographic feature: Cliff Springs, on the Maitland River.

Type section or area: Type area on Western Creek, a tributary of Harding River

Upper and lower limits: Conformably overlain by Kylena Basalt. Overlies Mount Roe Basalt or 'Archaean'.

Original correlations: Hardey Sandstone, Green Hole Conglomerate Glen Herring Shale, Beaton [sic] Creek Conglomerate, Lower Coongan Volcanics.

Status and comments: The Cliff Springs Formation is equivalent to the Hardey Formation of this Bulletin.

Coongan Volcanics

Proposed by: Noldart and Wyatt (1962, p. 92–94).

Lithology/thickness: '. . . acid-intermediate lava flows and fine fragmental ash beds . . .'/not stated.

Geographic feature: 'the Coongan River in the vicinity of the Comet mining centre'.

Type section or area: No formal specification; 'to be seen on Beatons Creek westerly of Nullagine' and at geographic feature quoted above.

Upper and lower limits: Not clearly stated. Table VI shows Green Hole Conglomerate above and Glenn Herring Shale below.

Original correlations: '. . . with the bottom phase of the Little de Grey Lava formation.'

Status and comments: Name never adequately defined, and should be abandoned.

Cooya Pooya Dolerite

Proposed by: Kriewaldt and Ryan (1967, p. 38).

Lithology/thickness: 'Altered dolerite with very subordinate granophyres; locally with amygdales'/up to 300 feet (91 m).

Geographic feature: Cooya Pooya Homestead.

Type section or area: Lockyer Gap.

Upper and lower limits: 'Intrusive into the Cliff Springs Formation, commonly at the contact with the overlying Kylena Basalt'.

Original correlations: 'with gabbro–granophyre complex of the Dampier Archipelago'.

Status and comments: Intrusive into the upper Hardey Formation on PYRAMID and YARRALOOLA. Commonly occurs at the contact with the overlying Kylena Formation.

Duffers Creek Porphyry

Proposed by: Noldart and Wyatt (1962, p.88).

Lithology/thickness: Feldspar porphyry/not stated.

Geographic feature: Duffer Creek.

Type section or area: '. . . near the junction of Duffers Creek with the Coongan River (Ref. 259363)'. This locality is believed to be near Marble Bar (MARBLE BAR 862646).

Upper and lower limits: Occurs as 'a series of dykes intruding the metamorphics of the Archaean basement'.

Original correlations: Tentatively correlated with 'An extensive outcrop of similar rock type outcropping At Taylors Creek (Ref. 299278) on the Nullagine–Marble Bar road'.

Status and comments: The name Duffers Creek Porphyry has been abandoned. The location and description of the type locality indicates that this unit is part of the greenstone basement (the Duffer Formation of Lipple (1975)), not part of the Fortescue Group. Noldart and Wyatt's proposed correlation with a Fortescue Group unit, now referred to as the Bamboo Creek Porphyry, is therefore invalid.

Glenn Herring Shale

Proposed by: Noldart and Wyatt (1962, p. 82).

Lithology/thickness: 'Red and grey shales . . . often interbedded with sandstones and fine conglomerates'/Up to 80 feet (24 m).

Geographic feature: Glenn [sic] Herring Creek.

Type section or area: Not formally stated. '. . . the headwaters of Glenn Herring Creek . . . and in . . . outcrops extending northerly to Warralong Creek'.

Upper and lower limits: Not specifically described. Table VI shows Coongan Volcanics above and Beatons Creek Conglomerate below.

Original correlations: None.

Status and comments: The name has not been properly defined and could be abandoned.

Green Hole Conglomerate

Proposed by: Noldart and Wyatt (1962, p. 81).

Lithology/thickness: 'Fine conglomerate grading to coarse arkosic sandstones . . .'/up to 1100 feet (335 m).

Geographic feature: Green Hole, Little de Grey River.

Type section or area: Not properly defined. Authors refer to Traves et al.'s (1956) description of section on west side of river at Green Hole.

Upper and lower limits: Not accurately defined. Table VI shows above Coongan Volcanics and below Tumbinna Pisolite.

Original correlations: None directly stated, but similarities noted between Green Hole conglomerate and sedimentary rocks in the Just-in-Time area.

Status and comments: Not properly defined. The name could be abandoned.

Fortescue Group

Proposed by: MacLeod et al. (1963a,b).

Lithology/thickness: 'Lavas and clastic sediments . . .'/ 14 000 feet (4270 m).

Geographic feature: Fortescue River, by implication.

Type section or area: '. . . in the vicinity of Moonah Well on the southern limb of the Rocklea Anticline'.

Upper and lower limits: Unconformably overlies Archaean basement; conformably overlain by the Marra Mamba Iron Formation of the Hamersley Group.

Original correlations: None.

Status and comments: The lowermost Group in the Hamersley Basin succession. Together with the overlying Hamersley and Turee Creek Groups it forms the Mount Bruce Supergroup.

Gidley Granophyre

Proposed by: de Laeter and Trendall (1971).

Lithology/thickness: '. . . granophyre and associated quartz gabbro . . .'/about 10 000 feet (3000 m).

Geographic feature: Gidley Island, by implication.

Type section or area: Not strictly defined. Good exposures occur between the AMG Zone 50 limits 440000E and 488000E, and 7743000N and 7708000N.

Upper and lower limits: It 'is intruded at the unconformity between Archaean granitic rocks and the overlying gently dipping basaltic lavas of the Proterozoic'.

Original correlations: None proposed. The authors state that 'the relationship between the granophyre and the gabbros and dolerites of the same general area is not clear . . .'.

Status and comments: A useful name applied to a localized unit. The 'gently dipping basaltic lavas of the Proterozoic' referred to above are possibly equivalent to the Kylena Formation.

Hardey Sandstone

Proposed by: MacLeod et al. (1963a,b).

Lithology/thickness: '. . . a white to reddish brown and green quartz sandstone. It is commonly arkosic and in places rather calcareous'/about 4000 feet (1220 m).

Geographic feature: Hardey River.

Type section or area: 'type area in the core of the Rocklea Anticline . . .'.

Upper and lower limits: unconformably overlies 'Archaean', conformably overlain by Boongal Pillow Lava Member of Mount Jope Volcanics.

Original correlations: None.

Status and comments: Name changed to Hardey Formation (Thorne et al., 1991).

Hardey Formation

Proposed by: Thorne et al. (1991).

Lithology/thickness: Sandstone, siltstone, mudstone, tuff, basalt, and chert. Intruded by mafic sills/about 1500 m in the type area, but elsewhere it may exceed 3 km.

Geographic feature: Hardey River.

Type section or area: As for the Hardey Sandstone.

Upper and lower limits: Unit unconformably overlies granite–greenstone basement or else has unconformable or conformable contacts with the Mount Roe Basalt. In the type area The Hardey Formation is conformably overlain by Boongal Formation; in the northern Hamersley Basin it is succeeded conformably or unconformably by the Kylena Formation.

Original correlations: None.

Status and comments: Equivalent to the Hardey Sandstone of MacLeod et al. (1963a,b). The change in name was considered appropriate in view of the high proportion of non-arenaceous rocks contained in this part of the stratigraphy. The name currently refers to the mixed suite of sedimentary and volcanic rocks occurring above the Mount Roe Basalt and below the Kylena or Boongal Formations.

Isabella Member

Proposed by: Williams and Trendall (1998b).

Lithology/thickness: Fine-grained rhyolite with quartz or carbonate amygdales/ about 50 m.

Geographic feature: Isabella Range.

Type section or area: About 5 km southeast of abandoned Baramine Homestead.

Upper and lower limits: Unit appears to be intrusive into fine-grained epiclastic rocks of the Jeerinah Formation.

Original correlations: None.

Status and comments: Unit is confined to the northern part of the Gregory Range inlier.

Jeerinah Formation

Proposed by: MacLeod et al. (1963a,b, p. 47).

Lithology/thickness: Shale, chert, jaspillite, mudstone, quartzite and dolerite/3000 feet (914 m).

Geographic feature: Jeerinah Rockhole.

Type section or area: Type section about 1.5 km south of Mount Turner.

Upper and lower limits: Conformably overlies Mount Jope Volcanics. Overlain by Marra Mamba Iron Formation.

Original correlations: None.

Status and comments: The uppermost formation of the Fortescue Group.

Koongaling Volcanics

Proposed by: Hickman (1978, p. 12).

Lithology/thickness: Rhyolite and dacite lavas, and locally thick agglomerate and tuff/At least 1000 m.

Geographic feature: Koongaling Hill.

Type section or area: Type area 'East of Ragged Hills'.

Upper and lower limits: Conformably underlies Kylena Basalt; 'grades downwards into granophyre (*Pgy*)' according to Hickman et al. (1983).

Original correlations: 'Correlated with Mount Roe Basalt'. Forms part of the Gregory Range Granite of Noldart and Wyatt (1962).

Status and comments: In this Bulletin the Koongaling Volcanics is referred to as the Koongaling Volcanic Member and is correlated with the Bamboo Creek Member (Hardey Formation).

Kuruna Siltstone Member

Proposed by: MacLeod and de la Hunty (1966, p. 11).

Lithology/thickness: 'Siltstone, mudstone, oolite'/60 feet (18 m).

Geographic feature: Kuruna Bore.

Type section or area: Not formally defined; thickest section '4 miles southeast of Nymerina Spring'.

Upper and lower limits: Overlain by Maddina Basalt Member. Underlain by Nymerina Basalt Member.

Original correlations: None.

Status and comments: In this Bulletin the unit is referred to as the Kuruna Member of the Maddina Basalt. It is best developed on ROY HILL, but is not recognized on neighbouring sheets to east (PYRAMID) or west (BALFOUR DOWNS), though it does occur on northeast NULLAGINE. This unit had previously been upgraded to formation status by Hickman and Lipple (1978).

Kylena Basalt Member

Proposed by: MacLeod and de la Hunty (1966, p. 10).

Lithology/thickness: Dark green vesicular and amygdaloidal lava/100 feet (30 m).

Geographic feature: Kylena Well.

Type section or area: Type section not formally defined; 'best exposure extends northeast from Kylena Well, in a belt 2 miles wide'.

Upper and lower limits: 'overlies gneiss and outcrops of the Warrawoona Series'; overlain by Tumbiana Pisolite Member.

Original correlations: None.

Status and comments: Raised to formation status on PYRAMID by Kriewaldt and Ryan (1967). In this Bulletin the name refers to the major basalt unit above the Hardey Formation and below the Tumbiana Formation. It is correlated with the Boongal Formation of the southern Hamersley Basin.

Kylena Volcanics

Proposed by: Williams (1968, p. 9).

Lithology/thickness: 'Basic, intermediate, acid lavas, amygdaloidal and vesicular textures, thin intercalated tuffaceous pyroclastic rocks with sandstone and arkose.'/ 500 to 1500 feet (150 to 460 m).

Geographic feature: None.

Type section or area: None.

Upper and lower limits: Overlies the Cliff Springs Formation; overlain by the Pillingini Tuff.

Original correlations: Kylena Basalt, 'Lower' Little de Grey Lava, Upper Coongan Volanics, Bunjinah Pillow Lava Member.

Status and comments: Not properly defined and can probably be abandoned. It is equivalent to the Kylena Formation.

Kylena Formation

Proposed by: Kojan and Hickman (1998).

Lithology/thickness: Massive or amygdaloidal basaltic flows and breccia, andesite, dacite and rhyolite; minor pillow lava, local pyroxene spinifex-textured basalt and stromatolitic carbonate/not stated.

Geographic feature: None.

Type section or area: None.

Upper and lower limits: Conformably or disconformably overlies the Hardey Formation; conformably overlain by the Tumbiana Formation.

Original correlations: Boongal Formation.

Status and comments: This name replaces the Kylena Basalt. Use of the term 'formation' was preferred due to the locally high proportion of non-basaltic rocks in the northwest and northeast Pilbara.

Lewin Shale

Proposed by: de la Hunty (1964).

Lithology/thickness: 'Shales, banded chert and jaspillite, mudstone, thin dolomite; iron and chert concretions.'/ 800 feet (244 m).

Geographic feature: Mount Lewin.

Type section or area: Not stated.

Upper and lower limits: Overlies the Little de Grey Lava and is itself overlain by the Carawine Dolomite.

Original correlations: None.

Status and comments: The name Lewin Shale has been used on BALFOUR DOWNS (1st edition), NULLAGINE, and YARRIE to describe rocks that are broadly equivalent to the combined Jeerinah Formation and Marra Mamba Iron Formation (Hamersely Group). Subsequent mapping has shown that both the Jeerinah Formation and the Marra Mamba Iron Formation are recognized in the type area on BALFOUR DOWNS (Williams, 1989). The term Lewin Shale has therefore been abandoned.

Little de Grey Lava

Proposed by: Noldart and Wyatt (1962, p. 91).

Lithology/thickness: Intermediate to basic lavas, strongly amygdaloidal in part/not stated.

Geographic feature: Little de Grey River.

Type section or area: Poorly defined; in the Little de Grey Syncline.

Upper and lower limits: The name refers to 'two separate sequences of intermediate to basic lavas. The lower sequence rests unconformably upon the Green Hole Beds and is in turn overlain by beds of the Carawine Dolomite formation. The upper, and final phase, is interbedded with the Carawine Dolomite, occuring approximately in the centre of that formation'.

Original correlations: With 'highly sheared amydaloidal lavas of the Gregory Range'.

Status and comments: The lower sequence of Little de Grey Lava forms part of the Kylena Formation; the upper sequence equates with the Maddina Formation. The term Little de Grey Lava has limited usefulness and can probably be abandoned.

Lyre Creek Agglomerate Member

Proposed by: Williams (1968, p. 28).

Lithology/thickness: 'Green, bedded and massive tuff, crystal tuff, agglomerate, with bombs of amygdaloidal lava; basal quartzitic tuff; some calcareous beds; volcanic pisoliths in places.'/0–250 feet (0–76 m).

Geographic feature: Lyre Creek Well.

Type section or area: Western Creek.

Upper and lower limits: Occupies the upper part of the Cliff Springs Formation. Overlain conformably by the Kylena Basalt.

Original correlations: Hardey Sandstone, Green Hole Conglomerate, Glen Herring Shale, Beaton Creek Conglomerate wholly, or in part.

Status and comments: The name is changed to Lyre Creek Member for this Bulletin. It forms the upper part of the Hardey Formation in the northwestern Hamersley Basin.

Madang Tuff Member

Proposed by: Blight (1985, p. 13).

Lithology/thickness: Intermediate to mafic tuff which contains accretionary lapilli locally; amygdaloidal basalt/ up to 130 m.

Geographic feature: Madang Well.

Type section or area: 1 km east of Madang Well.

Upper and lower limits: not accurately defined. Occurs 'approximately mid-way through the (Hardey Sandstone) sequence . . .'.

Original correlations: part of Hardey Sandstone near Nullagine. *Efhw* unit on MARBLE BAR.

Status and comments: Forms part of the Hardey Formation. Its lateral impersistence, coupled with the presence of other tuff units within the Hardey Formation, limits its usefulness.

Maddina Basalt Member

Proposed by: MacLeod and de la Hunty (1966, p. 11).

Lithology/thickness: Vesicular and amygdaloidal basalt/ 200 feet (61 m).

Geographic feature: Maddina Spring.

Type section or area: Not accurately defined.

Upper and lower limits: A member within the Mount Jope Volcanics. Overlies the Kuruna Silstone Member, conformably overlain by the Jeerinah Formation.

Original correlations: None presented.

Status and comments: Elevated to formation status on PYRAMID by Kriewaldt and Ryan (1967, p. 14). In current usage it incorporates the Nymerina Basalt and the Kuruna Siltstone Member, and correlates with the Bunjinah Formation of the southern Hamersley Basin.

Maddina Volcanics

Proposed by: Williams (1968, p. 9).

Lithology/thickness: Basic, intermediate, acid lavas, amygdaloidal and vesicular textures, thin intercalated tuffaceous sedimentary rocks/2000 feet (610 m).

Geographic feature: None.

Type section or area: None.

Upper and lower limits: Overlies the Pillingini Tuff; overlain by the Jeerinah Formation.

Original correlations: Maddina Basalt Member, 'Upper' Little de Grey Lava, Boongal Pillow Lava Member.

Status and comments: The name has not been properly defined and can probably be discarded. It is equivalent to the Maddina Basalt of this Bulletin.

Maddina Formation

Proposed by: Kojan and Hickman (1998).

Lithology/thickness: Massive or amygdaloidal basaltic flows and breccia, andesite, dacite, and rhyolite; local thin beds of sandstone and dolomite/not stated.

Geographic feature: None.

Type section or area: None.

Upper and lower limits: Conformably overlies the Tumbiana Formation; conformably overlain by the Jeerinah Formation.

Original correlations: Bunjinah Formation.

Status and comments: This name replaces the Maddina Basalt. Use of the term 'formation' was preferred due to the locally high proportion of non-basaltic rocks in the northwest and northeast Pilbara.

Meentheena Carbonate Member

Proposed by: Lipple (1975, p. 63).

Lithology/thickness: 'Ripple-bedded dark grey siliceous carbonate rocks, containing algal stromatolites and syndepositional slump structures'/20 m.

Geographic feature: Meentheena Homestead.

Type section or area: Type area '3 km northeast of Pelican Pool' on Nullagine River.

Upper and lower limits: Underlies Nymerina Basalt, overlies Mingah Tuff Member.

Original correlations: See Tumbiana Formation.

Status and comments: In this Bulletin the name has been shortened to Meentheena Member of the Tumbiana Formation.

Mingah Tuff Member

Proposed by: Lipple (1975, p. 63).

Lithology/thickness: Basaltic to intermediate tuff; carbonate. Minor siltstone, mudstone, and basalt/150 m.

Geographic feature: Mingah Well.

Type section or area: Pelican Pool, on the Nullagine River.

Upper and lower limits: Underlies Meentheena Carbonate Member; conformably overlies Kylena Basalt.

Original correlations: See Tumbiana Formation.

Status and comments: The unit is here referred to as the Mingah Member of the Tumbiana Formation.

Mount Jope Basalt

Proposed by: MacLeod et al (1963a,b, p. 46).

Lithology/thickness: 'Pillow lavas and pyroclastics'/ 7000 feet (2130 m).

Geographic feature: Mount Jope.

Type section or area: 'type locality around Mount Jope'.

Upper and lower limits: 'conformably overlies the Hardey Sandstone' conformably overlain by Jeerinah Formation.

Original correlations: None.

Status and comments: de la Hunty (1965) revised the name to Mount Jope Volcanics. It is equivalent to the Boongal, Pyradie, and Bunjinah Formations of this Bulletin.

Mount Jope Volcanics

Proposed by: de la Hunty (1965, p. 9).

Lithology/thickness: 'Pillow lavas and pyroclastics'/ 7000 feet (2130 m).

Geographic feature: Mount Jope.

Type section or area: 'type locality around Mount Jope'.

Upper and lower limits: See Mount Jope Basalt.

Original correlations: None.

Status and comments: Renaming of Mount Jope Basalt of MacLeod et al. (1963a,b). It is equivalent to the combined Boongal, Pyradie, and Bunjinah Formations. With further modification (?Mount Jope Subgroup) the name could be retained for areas, such as the western Wyloo Dome, where these three formations cannot be easily separated.

Mount Roe Basalt

Proposed by: Kriewaldt (1964b, p. 32).

Lithology/thickness: Basalt and andesite; 'restricted occurences of agglomerate, local basal arkoses, and local intercalations of shales and ripple-marked quartz sandstones'/8000 feet (2440 m) maximum.

Geographic feature: Mount Roe.

Type section or area: 'The type area is from the unconformity north of Mount Roe to the overlying agglomerate bed at Pinanular Pool'.

Upper and lower limits: The unit is unconformable upon 'Archaean basement'. The top of the formation is taken at the contact between the basalt and the overlying sedimentary rocks.

Original correlations: Outliers of volcanic rock which unconformably overlie basement on ROEBOURNE and DAMPIER, and possibly on YARRALOOLA.

Status and comments: Recognized as the lowermost formation of the Fortescue Group over most of the Hamersley Basin. Exceptions to this occur on MARBLE BAR and TUREE CREEK, where sedimentary rock units underlie the Mount Roe Basalt.

Nallanaring Volcanic Member

Proposed by: Williams (1968, p. 28).

Lithology/thickness: 'Predominantly basalt, some minor pyroclastics, volcanic pisoliths and acid lavas towards the bottom'; local pillow lava/0–250 feet (0–76 m).

Geographic feature: Nallanaring Creek.

Type section or area: Nallanaring Creek.

Upper and lower limits: Lies conformably below the Roy Hill Shale and above the Warrie Member of the Jeerinah Formation.

Original correlations: None presented.

Status and comments: The unit is here referred to as the Nallanaring Member of the Jeerinah Formation. It forms a typically thin, though laterally persistent, unit on YARRALOOLA.

Nymerina Basalt Member

Proposed by: MacLeod and de la Hunty (1966, p. 11).

Lithology/thickness: '. . . dark-green vesicular and amygdaloidal basalt with some pillow lava bands, also interbedded pyroclastics and sediments'/up to 500 feet (150 m).

Geographic feature: Nymerina Spring.

Type section or area: Not defined.

Upper and lower limits: Forms the middle part of the Mount Jope Volcanics. It overlies the Tumbiana Pisolite Member and is in turn overlain by the Kuruna Siltstone Member.

Original correlations: By inference with part of the Mount Jope Volcanics on MOUNT BRUCE.

Status and comments: This unit was upgraded to fomation status by Hickman and Lipple (1978). For all practical purposes, the Nymerina Basalt can only be distinguished from the Maddina Formation where the latter is underlain by the Kuruna Member (Siltstone). In this Bulletin the Maddina Formation is equivalent to the combined Maddina Basalt, Kuruna Siltstone and Nymerina Basalt of Hickman and Lipple (1978).

Pearana Basalt

Proposed by: Hickman (1978, p. 12).

Lithology/thickness: 'Vesicular and porphyritic basalt. Felsic lava in upper part'/About 500 m.

Geographic feature: Not stated.

Type section or area: Not stated.

Upper and lower limits: Conformably overlies the Tumbiana Formation and conformably underlies the Lewin Shale.

Original correlations: 'It is directly correlated with the Nymerina and Maddina Basalts'.

Status and comments: The name given to the upper volcanic unit of the Fortescue Group in the Gregory Range. In this Bulletin the Pearana Basalt is correlated with the Maddina Formation of the main outcrop area.

Pillingini Tuff

Proposed by: Kriewaldt and Ryan (1967).

Lithology/thickness: Tuff, 'tuffaceous siltstone, shale and sandstone; beds of volcanic pisoliths, . . . calcareous beds with stromatolites', basic to intermediate lavas/ 150–500 feet (45–135 m).

Geographic feature: Pillingini Creek, a tributary of the George River.

Type section or area: Mount Herbert.

Upper and lower limits: Conformably overlies the Kylena Basalt; conformably overlain by the Maddina Basalt.

Original correlations: Kuruna Siltstone Member, Nymerina Basalt Member, Pyradie Pyroclastic Member, and, less certainly, the Tumbiana Pisolite.

Status and comments: The Pillingini Tuff is a direct lateral equivalent of the Tumbiana Formation of the eastern Hamersley Basin (Hickman, 1983). Although there is some nomenclatural confusion concerning the name Tumbiana Formation, it is given priority on the basis of its historical preference and more widespread usage. The term Pillingini Tuff is therefore discarded.

Pyradie Pyroclastic Member

Proposed by: de la Hunty (1965, p. 10).

Lithology/thickness: '. . . finely banded grey ash beds with agglomerate and volcanic bombs . . . Minor amounts of chert, shale and jaspilite have been noted, and there is some coarse-grained dolerite and pillow lava'/about 2000 feet (610 m).

Geographic feature: Pyradie Well.

Type section or area: Not stated.

Upper and lower limits: Underlain by the Boongal Pillow Lava Member, overlain by the Bunjinah Pillow Lava Member.

Original correlations: None.

Status and comments: The Pyradie Pyroclastic Member has been upgraded to formation status, and is recognized throughout the southern Hamersley Basin.

Pyradie Formation

Proposed by: Thorne et al. (1991).

Lithology/thickness: Pyroxene spinifex-textured basalt flows and pillow lava, tuff, minor komatiite and chert/up to 1000 m.

Geographic feature: As for Pyradie Pyroclastic Member.

Type section or area: Not stated.

Upper and lower limits: Conformably overlies the Boongal Formation, conformably underlies the Bunjinah Formation.

Original correlations: None.

Status and comments: This unit is equivalent to the Pyradie Pyroclastic Member (de la Hunty, 1965). It is recognized throughout the southern Hamersley Basin, but is characterized by the presence of spinifex-textured basalt and minor komatiite rather than by pyroclastic deposits as the original name suggests. The Pyradie Formation is correlated with the Tumbiana Formation of the northern Hamersley Basin.

Roy Hill Shale

Proposed by: MacLeod and de la Hunty (1966, p. 12).

Lithology/thickness: 'Leached pyritic shale'/100 feet (30 m).

Geographic feature: Roy Hill Homestead.

Type section or area: Poorly defined, but unit is well exposed on the Great Northern Highway, approximately 26 km north of Roy Hill Homestead.

Upper and lower limits: Conformably overlies the Warrie Member; conformably overlain by the Marra Mamba Iron Formation.

Original correlations: None.

Status and comments: The unit is here referred to as the Roy Hill Member of the Jeerinah Formation. It is not recognized in the southern Hamersley Basin. In northern outcrops it can be difficult to separate from the underlying Warrie Member, particularly in areas where the Warrie Member's upper dolomite unit is missing. On YARRALOOLA the Nallanaring Volcanic Member separates the Warrie and Roy Hill Members.

Spinaway Porphyry

Proposed by: Lipple (1975).

Lithology/thickness: '. . . coarse-grained plagioclase quartz dacite porphyry with abundant euhedral calcic oligoclase phenocrysts. . .'/Not stated.

Geographic feature: Spinaway Well.

Type section or area: '. . . 18 km south of Spinaway Well near the Great Northern Highway'.

Upper and lower limits: 'Intrusive sill into the Hardey Sandstone . . .'.

Original correlations: Correlated with the Bamboo Creek Porphyry of Noldart and Wyatt (1962).

Status and comments: The Spinaway Porphyry is here regarded as a direct equivalent of the Bamboo Creek Member. Since this latter name has historical preference the term Spinaway Porphyry is dispensed with.

Tumbiana Formation

Proposed by: Lipple (1975, p. 63).

Lithology/thickness: 'Upper carbonate member lower tuff member'/50–200 m.

Geographic feature: 'Tumbiana' Pool in the Nullagine River.

Type section or area: Type area 'from Pelican Pool . . . to 3 km northeast on the Nullagine River'.

Upper and lower limits: Conformably overlies Kylena Basalt. Conformably overlain by Nymerina Basalt.

Original correlations: Pillingini Tuff, Pyradie Pyroclastic Member of Mount Jope Volcanics.

Status and comments: There is confusion as to the original name of this unit. The term Tumbiana Formation was first used by Lipple (1975) who, along with de la Hunty (1963, p. 29–30) appears to have misquoted Noldart and Wyatt (1962) (see Tumbinna Pisolite entry). This incorrect usage was later adopted by other workers; for example, de la Hunty (1965), Hickman,(1983).

Despite this apparent misspelling of the original name, the term 'Tumbiana Formation' has gained wide acceptance in the geological literature. In this Bulletin it refers to the major sedimentary unit lying above the Kylena Formation and below the Maddina Formation. It is correlated with the Pyradie Formation of the southern Hamersley Basin.

Tumbinna Pisolite

Proposed by: Noldart and Wyatt (1962, p. 80).

Lithology/thickness: 'Dark green-greyish calcareous volcanic (?) and/or sedimentary rocks containing innumerable pea-sized oolites or pisolites'/unknown.

Geographic feature: Tumbinna Pool on Nullagine River.

Type section or area: By implication Tumbinna Pool, but no proper formal definition or description.

Upper and lower limits: Noldart and Wyatt (1962) show the unit as overlain by 'Little de Grey Lava', and underlain by Green Hole Conglomerate.

Original correlations: Along Yilgalong Creek and in the Gregory Range.

Status and comments: There is a problem with later changes in spelling. De la Hunty (1963, p. 29) states erroneously, 'Noldart named this formation from its outcrop at Tumbiana Pool'. A similar misquote appears in the explanatory notes for BALFOUR DOWNS (de la Hunty, 1965): 'Noldart and Wyatt (1962) named the Tumbiana Pisolite from its outcrop at Tumbiana Pool . . .'.

It is unclear why de la Hunty and later authors (e.g. Lipple, 1975) have misquoted or modified Noldart and Wyatt's original definition. The result of these changes, however, has been that the term Tumbiana Pisolite, and later, Tumbiana Formation, have been widely accepted whereas Tumbinna Pisolite has been discarded.

Warrie Member

Proposed by: MacLeod and de la Hunty (1966, p. 12).

Lithology/thickness: 'Shale, chert, mudstone, dolomite'/270 feet (82 m).

Geographic feature: Warrie Homestead, by implication.

Type section or area: Poorly defined, appears to be the Great Northern Highway roadside section on ROY HILL.

Upper and lower limits: Overlies the Woodiana Sandstone Member; is conformably overlain by the Roy Hill Shale Member.

Original correlations: None.

Status and comments: This unit is not recognized in the southern Hamersley Basin. In northern outcrops it can be difficult to separate from the overlying Roy Hill Shale Member, particularly in areas where its upper dolomite unit is missing. On YARRALOOLA, the Nallanaring Volcanic Member separates the Warrie and Roy Hill Members.

Warri Warri Member

Proposed by: Williams and Trendall (1998c).

Lithology/thickness: Rhyolite, tuffaceous sedimentary rock, felsic agglomerate, conglomerate and sandstone/ 0.5–3 km.

Geographic feature: Warri Warri Creek, by implication.

Type section or area: Not given.

Upper and lower limits: Overlies the Koongaling Volcanic Member (Hardey Formation). Overlain by the Kylena Basalt.

Original correlations: Tanguin Member (Hardey Formation).

Status and comments: This unit is not recognized outside the Gregory Range area on NULLAGINE.

Warroo Hill Member

Proposed by: Williams and Trendall (1998a).

Lithology/thickness: Quartzite, micaceous quartzite, metamorphosed quartz pebble conglomerate, siltstone, metamorphosed basalt, dolerite, pelitic schist, and metadolomite/ >1 km.

Geographic feature: Warroo Hill, by implication.

Type section or area: Not given

Upper and lower limits: Overlies the Koongaling Volcanic Member (Hardey Formation). Unconformably overlain by the Tarcunyah Group.

Original correlations: Tanguin Member (Hardey Formation).

Status and comments: This unit is not recognized outside the Gregory Range area on NULLAGINE.

Woodiana Sandstone

Proposed by: MacLeod and de la Hunty (1966, p. 11).

Lithology/thickness: Sandstone, mudstone/200 feet (60 m).

Geographic feature: Woodiana, an old camp 22.4 km north-northeast of Cowra Outcamp.

Type section or area: Not defined but presumably the Woodiana area.

Upper and lower limits: Conformably overlies the Mount Jope Volcanics; conformably overlain by the Warrie Member.

Original correlations: None.

Status and comments: The unit is here referred to as the Woodiana Member. It is not recognized in the southern Hamersley Basin.

Informal names listed

Units 1–9

Proposed by: Kriewaldt (1964b).

Area covered: Roebourne region.

Original correlations: Units 7, 8, 9 (Jeerinah Formation), Units 5, 6 (‘upper’ Little de Grey Lava, Bunjinah Member), Unit 4 (Tumbiana Pisolite, Bunjinah Member), Unit 3 (‘lower’ Little de Grey Lava and ‘upper’ Coongan Volcanics, Bunjinah Member.), Unit 2 (Beaton Creek Conglomerate, Green Hole Conglomerate, Budjan Creek Formation, Glen Herring Shale, ‘lower’ Coongan Volcanics, Pyradie and Boongal Members of the Mount Jope Basalt, and Hardey Sandstone), Unit 1 (None presented).

Status and comments: Most of the names listed above are no longer in current use. Kriewaldt’s (1964b) units equate to the following formal names: Jeerinah Formation (Units 7, 8, 9), Maddina, Bunjinah, Tumbiana, Pyradie, Kylena and Boongal Formations (Units 3, 4, 5, 6), Hardey Formation (Unit 2) and Mount Roe Basalt (Unit 1).

Upper Sediments, Upper Lavas, Middle Sediments, Middle Lavas, Lower Sediments, Lower Lavas

Proposed by: Trendall (1975b).

Area covered: Hamersley Basin.

Original correlations: Upper Sediments (Jeerinah Formation, lower part of Lewin Shale), Upper Lavas (Maddina Basalt, Maddina Volcanics, Bunjinah Pillow lava Member, Little de Grey Lava (in part), ‘upper’ Little de Grey Lava), Middle Sediments (Pillingini Tuff, Pyradie Pyroclastic Member, Kuruna Siltstone Member, Nymerina Basalt Member,Tumbiana Pisolite Member, Tumbiana Pisolite), Middle Lavas (Kylena Volcanics, Kylena Basalt, Kylena Basalt Member, Little de Grey Lava (in part), ‘lower’ Little de Grey Lava, ‘upper’ Coongan Volcanics), Lower Sediments (Unnamed sandstone on ROEBOURNE, Cliff Springs Formation, Hardey Sandstone, Beatons Creek Conglomerate, Green Hole Conglomerate, ‘lower’ Coongan Volcanics, Glen Herring Shale), Lower Lavas (Mount Roe Basalt).

Status and comments: This scheme formed the basis for a later informal subdivision of the Fortescue Group

(Trendall, 1990b). The formal names used in this Bulletin and the corresponding units of Trendall (1975b) are as follows: Jeerinah Formation (Upper Sediments) Maddina and Bunjinah Formations (Upper Lavas), Tumbiana and Pyradie Formation (Middle Sediments), Kylena and Boongal Formations (Middle Lavas), Hardey Formation (Lower Sediments), Mount Roe Basalt (Lower Lavas).

Upper sedimentary unit, Upper volcanic unit, Middle sedimentary unit, Middle volcanic unit, Lower sedimentary unit, Lower volcanic unit

Proposed by: Trendall (1990b).

Area covered: Hamersley Basin.

Original correlations: See informal subdivision of Trendall (1975b).

Status and comments: Based upon the subdivision of Trendall (1975b). Corresponding formal names as used in this Bulletin are as follows: Jeerinah Formation (Upper sedimentary unit), Maddina and Bunjinah Formations (Upper volcanic unit), Tumbiana and Pyradie Formation (Middle sedimentary unit), Kylena and Boongal Formations (Middle volcanic unit), Hardey Formation (Lower sedimentary unit), Mount Roe Basalt (Lower volcanic unit).

Unit 4, Unit 3, Unit 2, Unit 1

Proposed by: Blake (1984a).

Area covered: Northern Pilbara Craton.

Original correlations: With the Mount Roe Basalt, Hardey Sandstone, and Kylena Basalt.

Status and comments: This study recognizes the importance of intra-Fortescue Group unconformities in northern parts of the Hamersley Basin. Original correlations are adopted for this Bulletin although Hardey Sandstone and Kylena Basalt are here referred to as Hardey Formation and Kylena Formation respectively.

Upper metasedimentary unit, Upper mafic volcanic unit, Felsic pyroclastic unit, Lower mafic volcanic unit, Lower metasedimentary unit

Proposed by: Tyler (1986).

Area covered: Southeastern Hamersley Basin.

Original correlations: Upper metasedimentary unit (Jeerinah Formation), Upper mafic volcanic unit (Bunjinah Pillow Lava Member), Felsic pyroclastic unit (Pyradie Pyroclastic Member), Lower mafic volcanic unit (Boongal Pillow Lava Member), Lower metasedimentary unit (Hardey Sandstone).

Status and comments: The correlations used in this Bulletin are based upon those of Thorne and Blake (1990) and are as follows: Jeerinah Formation (Upper metasedimentary unit), Boongal, Pyradie, and Bunjinah Formations (Upper mafic volcanic unit), Hardey Formation (Felsic pyroclastic unit, Lower mafic volcanic unit and Lower metasedimentary units).

Mount Bruce Megasequence Set, Chichester Range Megasequence, Nullagine Supersequence, Mount Jope Supersequence, Mount Roe Sequence, Hardey Sequence Package, Kylena Sequence, Tumbiana Sequence, Maddina Sequence Package, Marra Mamba Supersequence Package

Proposed by: Blake and Barley (1992).

Area covered: Hamersley Basin.

Original correlations: Mount Bruce Megasequence Set (Mount Bruce Supergroup), Chichester Range Megasequence (Fortescue Group and lower Hamersley Group), Nullagine Supersequence (Mount Roe Basalt and Hardey Formation), Mount Jope Supersequence (Kylena, Boongal, Tumbiana, Pyradie, Maddina and Bunjinah Formations) Mount Roe Sequence (Mount Roe Basalt), Hardey Sequence Package (Hardey Formation), Kylena Sequence (Kylena and Boongal Formations), Tumbiana Sequence (Tumbiana and Pyradie Formations), Maddina Sequence Package (Maddina and Bunjinah Formations), Marra Mamba Supersequence Package (Jeerinah Formation and lower Hamersley Group).

Status and comments: This scheme formed the basis for Barley et al. (1992) and Blake's (1993) interpretation of the Fortescue Group and the overlying part of the Mount Bruce Supergroup. Our correlations are the same as given above.

Glen Herring Creek, Pear Creek, Hales Grave Well, Taylor Creek, Pilbara Creek, and Langwell Creek Sequences

Proposed by: Blake (1993).

Area covered: Northern Pilbara Craton.

Original correlations: Hardey Formation in the Marble Bar, East Pilbara, and West Pilbara Basins. The upper Hales Grave Well Sequence equates to the Bamboo Creek Porphyry (Hickman, 1983). The Langwell Creek Sequence includes the Lyre Creek Agglomerate Member (Williams, 1968).

Status and comments: These units have current status. They have not been used outside the northern Pilbara Craton.

Appendix 2

Gazetteer of localities referred to in text

Locality name	*1:250 000 map sheet*	*AMG reference*	*Locality name*	*1:250 000 map sheet*	*AMG reference*
Alligator Anticline	Newman	660000E, 7449000N	East Lewis Island	Dampier	465265E, 7722077N
Bamboo Creek	Yarrie	209400E, 7686000N	Enderby Island	Dampier	450000E, 7721000N
Baramine Homestead	Yarrie	287267E, 7787219N	Fortescue River Fault	Newman	780000E, 7400000N
Baramine prospect	Yarrie	287601E, 7690329N	Fullbrook Well	Roebourne	523692E, 7691661N
Baramine South prospect	Yarrie	288480E, 7689417N	FVG–1 diamond drillhole	Roy Hill	756231E, 7503224N
Barramine deposit	Yarrie	297218E, 7683989N	Gidley Island	Dampier	480000E, 7736000N
Beasley River prospect	Mount Bruce	522524E, 7480131N	Glen Herring Creek	Marble Bar	773046E, 7633129N
Beatons Creek	Nullagine	195000E, 7576400N	Gorge Range	Port Hedland – Bedout Island	766966E, 7684931N
Bellary Creek	Turee Creek	568221E, 7434174N	Grants Hill	Nullagine	200321E, 7577249N
Bellary Dome	Turee Creek	563977E, 7438806N	Green Hole	Yarrie	218200E, 7682000N
Belvedere Group	Wyloo	426917N, 7499038N	Gregory Gorge	Yarraloola	491200E, 7616500N
Big Metawandy Well	Wyloo	435600E, 7490400N	Halleys Comet Mine	Marble Bar	783132E, 7649578N
Big Stubby	Marble Bar	784318E, 7651404N	Hardey River	Mount Bruce	541500E, 7466200N
Bilanoo Hill	Yarraloola	411000E, 7643600N	Hays Creek	Nullagine	268000E, 7586000N
Blacks prospect	Wyloo	428916E, 7498740N	Horse Well	Wyloo	437481E, 7498777N
Bluff Hill	Pyramid	565805E, 7672026N	Horseshoe Gorge	Pyramid	507100E, 7640800N
BMW–1 diamond drillhole	Mount Bruce	563000E, 7645700N	Ironstone prospect	Newman	711623E, 7440519N
BMW–2 diamond drillhole	Turee Creek	568700E, 7442800N	Jeerinah Anticline	Mount Bruce	508000E, 7520000N
Bonnie Downs Station	Roy Hill	802300E, 7543800N	Jeerinah Rockhole (Donkey Pool)	Mount Bruce	511500E, 7524000N
Boongal Well	Mount Bruce	551800E, 7477200N	Junction Pool	Robertson	243400E, 7401400N
Brockman/Edneys prospects	Newman	530889E, 7528392N	Just-in-Time mine	Marble Bar	781946E, 7647751N
Bull prospect	Newman	715064E, 7442316N	Karratha	Dampier	483000E, 7707000N
Bunjinah Spring	Mount Bruce	586400E, 7500400N	Knossos locality	Roy Hill	707200E, 7562400N
Burrup Peninsula	Dampier	481000E, 7727000N	Koongaling Hill	Nullagine	304000E, 7668000N
Camel Hump South prospect	Yarrie	192009E, 7682685N	Kumina Creek	Yarraloola	470000E, 7593100N
Cape Lambert	Roebourne	518000E, 7722000N	Kuruna Bore (Corona Well)	Roy Hill	768100E, 7536500N
Cape Preston	Dampier	417000E, 7696000N	Kylena Well	Roy Hill	773900E, 7552000N
Central prospect	Newman	717029E, 7440441N	Lalla Rookh Syncline	Marble Bar	737219E, 7666920N
Chillemarringa Well	Mount Bruce	543982E, 7476300N	Limestone Well	Marble Bar	772000E, 7646000N
Cliff Springs	Yarraloola	493000E, 7660000N	Limestone Well Centrocline	Marble Bar	771562E, 7647927N
Cooke Bluff Hill	Port Hedland – Bedout Island	746600E, 7681500N	Little de Grey River (Miningarra Creek)	Yarrie	222400E, 7695800N
Cookes Creek	Nullagine	236081E, 7603736N	Lockyer Gap	Roebourne	499000E, 7678200N
Coonabunna Creek	Nullagine	251386E, 7666768N	Lyre Creek Well	Pyramid	523800E, 7659200N
Coorbeelie River	Pyramid	568000E, 7640000N	Madang (Yandie) Well	Mount Bruce	547300E, 7472500N
Cooya Pooya	Pyramid	514400E, 7674100N	Maddina Spring	Roy Hill	754700E, 7539300N
Coppin Pool	Mount Bruce	617105E, 7468932N	Marble Bar	Marble Bar	785568E, 7656922N
Cowra (Mulga Downs) Outcamp	Roy Hill	703800E, 7529800N	Meentheena Centrocline	Nullagine	226922E, 7634975N
Croydon Top Camp	Pyramid	570600E, 7660325N	Meentheena Homestead	Nullagine	235800E, 7646100N
Dampier	Dampier	470000E, 7715000N	Meentheena prospect	Nullagine	257909E, 7644696N
Deadman Hill	Newman	746100E, 7365500N	MF–1 diamond drillhole	Yarraloola	484600E, 7627200N
Duffer Creek	Marble Bar	786664E, 7661828N	Milli Milli Dome	Mount Bruce	619714E, 7474448N

Appendix 2 (continued)

Locality name	*1:250 000 map sheet*	*AMG reference*	*Locality name*	*1:250 000 map sheet*	*AMG reference*
Millstream	Pyramid	507700E, 7612000N	Pear Creek	Port Hedland – Bedout Island	766877E, 7679393N
Mingah Well	Nullagine	233100E, 7642100N	Pear Creek Centrocline	Marble Bar	772018E, 7675617N
Minthicoondunna (Mindi)	Newman	629490E, 7484209N	Pear Creek Fault	Marble Bar	767500E, 7679000N
Moonah Well	Mount Bruce	543900E, 7483100N	Pelican Pool	Nullagine	226500E, 7637200N
Moondle Creek	Yarraloola	465395E, 7659350N	Pewah Hill	Roebourne	597233E, 7717222N
Mount Ada	Roebourne	517330E, 7686135N	Pillingini Creek	Pyramid	540500E, 7642500N
Mount Anketel	Yarraloola	507000E, 7718500N	Pinanular Pool	Roebourne	510684E, 7682452N
Mount de Courcey	Wyloo	428500E, 7487000N	Pindana Spring	Pyramid	555960E, 7612020N
Mount Elsie	Nullagine	250524E, 7599966N	Police Creek	Nullagine	229746E, 7621173N
Mount Enid	Yarraloola	440000E, 7600000N	Prairie Downs	Newman	719200E, 7393600N
Mount Florence Station	Pyramid	589300E, 7590200N	Pyradie Well	Mount Bruce	551200E, 7481300N
Mount Fraser	Balfour Downs	245500E, 7482000N	Pyramid Hill	Pyramid	538930E, 7659352N
Mount Herbert	Pyramid	526100E, 7641600N	Python Pool	Pyramid	524482E, 7640626N
Mount Jope	Mount Bruce	549200E, 7462500N	Ragged Hills	Nullagine	308000E, 7648000N
Mount Leopold	Dampier	460000E, 7679000N	Robe River	Yarraloola	460000E, 7593500N
Mount Lewin	Balfour Downs	213879E, 7496231N	Rocklea Dome	Mount Bruce	534202E, 7474882N
Mount Newdegate	Nullagine	300800E, 7662800N	Rooney Inlier	Balfour Downs	258270E, 7508075N
Mount Nicholas	Balfour Downs	2462000E, 749150N	Rosemary Island	Dampier	456548E, 7734968N
Mount Oscar	Roebourne	532208E, 7687497N	Roy Hill Homestead	Roy Hill	803800E, 7495500N
Mount Roe	Roebourne	508700E, 7686000N	Sandy Creek Well area	Newman	737614E, 7375803N
Mount Sholl	Dampier	491335E, 7685220N	SGS–1 diamond drillhole	Mount Bruce	587277E, 7535250N
Mount Sydney	Nullagine	312700E, 7632600N	Sherlock River Fault System	Pyramid	603488E, 7607415N
Mount Turner	Mount Bruce	543500E, 7489100N	Shovelanna Hill	Robertson	194922E, 7416422N
Mount Whaleback	Newman	774500E, 7414000N	Soansville	Marble Bar	725500E, 7621000N
Mount Wilkie	Dampier	395390E, 7788198N	South Hedland	Port Hedland – Bedout Island	667000E, 7744000N
Moxom Well	Nullagine	300100E, 7662800N	Spearhole Yard	Newman	739952E, 7397310N
Mumbillina Bluff	Pyramid	634548E, 7608416N	Spinaway Well	Nullagine	195153E, 7607167N
Murramunda–Jigalong Road	Robertson	248898E, 7352798N	Sylvania Inlier	Newman	754984E, 7380434N
Nallanaring Creek	Yarraloola	413300E, 7633500N	Tambrey ruins	Pyramid	562600E, 7607100N
Newman	Newman	779000E, 7414000N	Tanguin Hill	Nullagine	310000E, 7654500N
Newman – Port Hedland railway	Roy Hill	704000E, 7559000N	Tassy Queen	Marble Bar	779751E, 7539137N
Newman – Port Hedland road	Roy Hill	685600E, 7561200N	Teichman Well	Pyramid	626511E, 7647227N
North Pole Dome	Marble Bar	747585E, 7664922N	Tom Price	Mount Bruce	581000E, 7491000N
Nullagine	Nullagine	201300E, 7576400N	Tumbinna Pool	Nullagine	239700E, 7650500N
Nullagine Synclinorium	Nullagine	810127E, 7582590N	Turner Syncline	Mount Bruce	560000E, 7487000N
Nunyerry	Pyramid	593800E, 7616000N	Warambie Homestead	Roebourne	538300E, 7683200N
Nunyerry Creek	Pyramid	596800E, 7620000N	Warri Warri Creek	Nullagine	320000E, 7595000N
Nymerina Spring	Roy Hill	760100E, 7543100N	Warrie Homestead	Roy Hill	778900E, 7535800N
Oakover Syncline	Nullagine	240115E, 7675811N	Waroo Hill	Nullagine	319800E, 7658400N
Outcamp Well	Newman	761000E, 7402800N	West Lewis Island	Dampier	461000E, 7723000N
Paddy Market Creek	Marble Bar	733017E, 7615290N	Western Creek	Pyramid	508652E, 7661240N
Pannawonica	Yarraloola	430000E, 7606000N	Western Creek	Newman	765000E, 7396500N
Pannawonica railway	Yarraloola	485000E, 7635000N	Whim Creek	Roebourne	585000E, 7694600N
Paraburdoo	Turee Creek	568500E, 7434000N	Wickham	Roebourne	514000E, 7714000N
Paulsen mine	Wyloo	421756E, 7503318N	Wonmunna Anticline	Newman	690000E, 7437000N

Appendix 2 (continued)

Locality name	*1:250 000 map sheet*	*AMG reference*	*Locality name*	*1:250 000 map sheet*	*AMG reference*
Wonmunna prospect	NEWMAN	719057E, 7442873N	Yandicoogina Well	NULLAGINE	206727E, 7627699N
Woodiana Well (abandoned)	ROY HILL	716800E, 7546600N	Yannery Hills mine	YARRALOOLA	484417E, 7670151N
WRL–1 diamond drillhole	MOUNT BRUCE	624600E, 7545300N	Yanyare River	YARRALOOLA	448063E, 7668540N
Wyloo Dome	WYLOO	430000E, 7495000N			

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The Fortescue Group, in the Pilbara region of Western Australia, is one of the best preserved examples of Archaean volcanic and sedimentary rocks in the world. These deposits were formed during a protracted period of crustal breakup that occurred between about 2775 and 2630 million years ago. The Fortescue Group unconformably overlies older granite–greenstone rocks of the Pilbara Craton and is itself overlain by banded iron-formations of the Hamersley Group. The succession has a maximum thickness of about 6.5 km and is divided into seven major formations. This Bulletin discusses the stratigraphy, volcanology, sedimentology, and geochemistry of each formation and also summarizes the likely tectonic controls on sedimentation and volcanism. Sedimentary and volcanic rocks of the Fortescue Group have been associated with the small-scale production of gold, copper, lead, silver, zinc, and fluorite, and an exploration interest in uranium.

Further details of geological publications and maps produced by the Geological Survey of Western Australia can be obtained by contacting:

Information Centre
Department of Minerals and Energy
100 Plain Street
East Perth WA 6004
Phone: (08) 9222 3459   Fax: (08) 9222 3444
www.dme.wa.gov.au

DEPARTMENT OF MINERALS AND ENERGY
L. C. RANFORD, DIRECTOR GENERAL

GOVERNMENT OF WESTERN AUSTRALIA
HON. NORMAN MOORE, M.L.C.
MINISTER FOR MINES

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
DAVID BLIGHT, DIRECTOR

SCALE 1:500 000

5000 Metres 0 5 10 15 20 25 30 35 40 45 50 Kilometres

UNIVERSAL TRANSVERSE MERCATOR PROJECTION
HORIZONTAL DATUM: GEOCENTRIC DATUM OF AUSTRALIA 1994
VERTICAL DATUM: AUSTRALIAN HEIGHT DATUM

Grid lines indicate 25 000 metre interval of the Map Grid Australia Zone 50

The Map Grid Australia (MGA) is based on the Geocentric Datum of Australia 1994 (GDA94)
GDA94 positions are compatible within one metre of the datum WGS84 positions

⊙ Reference points to align maps based on the previous datum, AGD84, have been placed near the map corners

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

BULLETIN 144 PLATE 1A

GEOLOGY OF THE FORTESCUE GROUP—WEST PILBARA CRATON

DEPARTMENT OF MINERALS AND ENERGY
L. C. RANFORD, DIRECTOR GENERAL

GOVERNMENT OF WESTERN AUSTRALIA
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GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
DAVID BLIGHT, DIRECTOR

SCALE 1:500 000

5000 0 5 10 15 20 25 30 35 40 45 50
Metres Kilometres

UNIVERSAL TRANSVERSE MERCATOR PROJECTION
HORIZONTAL DATUM: GEOCENTRIC DATUM OF AUSTRALIA 1994
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Grid lines indicate 25 000 metre interval of the Map Grid Australia Zone 50

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GDA94 positions are compatible within one metre of the datum WGS84 positions

Reference points to align maps based on the previous datum, AGD84, have been placed near the map corners

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

BULLETIN 144 PLATE 1B

GEOLOGY OF THE FORTESCUE GROUP—EAST PILBARA CRATON

REFERENCE

PHANEROZOIC — CAINOZOIC

Qa Qs Qm

Qa Alluvium—unconsolidated silt, sand, and gravel; in river channels
Qs Undivided Quaternary deposits; includes colluvium, reworked alluvium, eolian sand, and clay
Qm Clay, mud, silt, and sand; tidal and supratidal deposits, mangroves, lagoons, and coastal dunes

Czs Czk Czp

Czs Undivided Cainozoic deposits; includes partly consolidated colluvium and alluvium, and silcrete and laterite
Czk Calcrete—sheet carbonate; found along major drainage lines
Czp ROBE PISOLITE: pisolitic limonite deposits; developed along palaeodrainage lines

MESOZOIC — JURASSIC AND CRETACEOUS

JK Sandstone, conglomerate, siltstone, and mudstone; includes CALLAWA FORMATION, FREZIER SANDSTONE, JARLEMAI SILTSTONE, NANUTARRA FORMATION, PARDA FORMATION, WALLAL SANDSTONE, and YARRALOOLA CONGLOMERATE

PALAEOZOIC — PERMIAN

P Sandstone, conglomerate, siltstone, and mudstone; includes PATERSON FORMATION, GRANT GROUP, and POOLE SANDSTONE

NORTHERN CARNARVON AND CANNING BASINS

PROTEROZOIC

Savory Group — ES Quartz sandstone, siltstone, and mudstone; minor conglomerate

Tarcunyah Group — EU Fine- to coarse-grained sandstone, siltstone, mudstone, and stromatolitic and non-stromatolitic dolomite; minor conglomerate (c. 800 Ma)

Yeneena Supergroup — ET Throssell Group: quartzitic and micaceous sandstone, siltstone, mudstone, conglomerate, stromatolitic and non-stromatolitic dolomite, and dolomitic conglomerate; includes Lamil Group Malbina Hill (c. 900 Ma)

Bangemall Group — EM Sandstone, siltstone, mudstone, conglomerate, chert, stromatolitic and non-stromatolitic dolomite, and dolomitic mudstone, and conglomerate (c. 1600–1000 Ma)

Bresnahan Group — EB Pebble to boulder conglomerate, pebbly sandstone, sandstone, siltstone, and mudstone

Mount Minnie Group — Ei Quartz sandstone, conglomerate, siltstone, and mudstone

Ec CAPRICORN FORMATION: conglomerate, ferruginous and quartzitic sandstone, ferruginous siltstone and mudstone, dolomite, and felsic volcanic rock

Savory Sub-basin; OFFICER BASIN; Yeneena Basin; PATERSON OROGEN; BANGEMALL BASIN; BRESNAHAN BASIN; Mount Minnie Basin; Bryah Basin

Wyloo Group (c. 1840 Ma)

EWa ASHBURTON FORMATION: mudstone, siltstone, thin- and thick-bedded sandstone, minor banded iron-formation, conglomerate, mafic and felsic volcanic rocks, and dolomite; metamorphosed
EWj JUNE HILL VOLCANICS: mafic lava, pillow lava, and volcaniclastic rock; includes intermediate and felsic volcanic rock, sandstone, and mudstone; metamorphosed
EWd DUCK CREEK DOLOMITE: thin- to thick-bedded dolomite; local stromatolitic dolorudite; minor chert and argillite; metamorphosed
EWm MOUNT McGRATH FORMATION: ferruginous sandstone and conglomerate, mudstone, siltstone, and dolomite; metamorphosed
EWc CHEELA SPRINGS BASALT: amygdaloidal basalt and mafic volcaniclastic rock; metamorphosed
EWq BEASLEY RIVER QUARTZITE: fine- to coarse-grained quartz sandstone, conglomerate, mudstone, and siltstone; metamorphosed

Ecb PINJIAN CHERT BRECCIA: chert breccia and banded chert; overlies CARAWINE DOLOMITE

Ashburton Basin; CAPRICORN OROGEN

Turee Creek Group — ETU Mudstone, siltstone, sandstone, and conglomerate; local stromatolitic dolomite and basalt; metamorphosed

Hamersley Group

AEH Undivided Hamersley Group; metamorphosed
EHu BOOLGEEDA IRON FORMATION: fine-grained, finely laminated iron-formation; mudstone, siltstone, and chert; metamorphosed
EHw WOONGARRA RHYOLITE: rhyolite, rhyodacite, rhyolitic breccia, and banded iron-formation; metamorphosed (c. 2440 Ma)
EHj WEELI WOLLI FORMATION: banded iron-formation (often jaspilitic), mudstone, siltstone, and numerous dolerite sills; metamorphosed
EHb BROCKMAN IRON FORMATION: banded iron-formation, chert, mudstone, and siltstone; metamorphosed (c. 2490 Ma)
AHs MOUNT McRAE SHALE and MOUNT SYLVIA FORMATION: mudstone, siltstone, chert, banded iron-formation, and dolomite; metamorphosed (2500 Ma)
AHd WITTENOOM FORMATION: thin- to medium-bedded dolomite, dolomitic mudstone, chert, and felsic to mafic volcanic sandstone; metamorphosed (c. 2560 Ma)
AHc CARAWINE DOLOMITE: thin- to thick-bedded dolomite, stromatolitic dolomite, and chert; metamorphosed
AHm MARRA MAMBA IRON FORMATION: chert, banded iron-formation, mudstone, and siltstone; metamorphosed

ARCHAEAN — Mount Bruce Supergroup — Fortescue Group

SOUTHERN PILBARA

AFj JEERINAH FORMATION: mudstone, siltstone, sandstone, chert, massive basaltic flows, basaltic pillow lava, basaltic breccia, and minor felsic volcaniclastic rock; intruded by numerous dolerite sills; metamorphosed (c. 2687 Ma)
AFjb Pillowed and massive basalt flows, and basaltic breccia; metamorphosed
AFupo Undivided BOONGAL, PYRADIE and BUNJINAH FORMATIONS in the Wyloo Dome; metamorphosed
AFu BUNJINAH FORMATION: pillowed and massive basaltic flows, basaltic breccia, and basaltic volcaniclastic sandstone; minor chert; amygdaloidal basalt flows occur in upper parts of formation; metamorphosed (c. 2717 Ma)
AFus Quartzofeldspathic conglomerate and sandstone, mudstone, and siltstone; metamorphosed
AFp PYRADIE FORMATION: pyroxene spinifex-textured basaltic flows and pillow lava, mafic volcaniclastic sandstone; minor chert; local komatiite; metamorphosed
AFo BOONGAL FORMATION: pillowed and massive basaltic flows, basaltic breccia, basaltic volcaniclastic sandstone, and argillite; minor chert; metamorphosed
AFh HARDEY FORMATION: argillite, sandstone, pebbly sandstone, conglomerate, basaltic flows and breccia, and basaltic volcaniclastic sandstone; metamorphosed (c. 2725 Ma)
AFhb Basaltic flows and volcaniclastic rock; metamorphosed (c. 2757 Ma; c. 2757–2764 Ma)
AFhs Feldspathic sandstone, pebbly sandstone, and conglomerate; minor argillite; metamorphosed
AFr MOUNT ROE BASALT: amygdaloidal basaltic flows and breccia, and basaltic volcaniclastic sandstone; metamorphosed
AFrs Conglomerate, sandstone, and argillite; metamorphosed
AFrf Felsic volcanic rock; metamorphosed (c. 2772 Ma)
AFb BELLARY FORMATION: argillite, feldspathic sandstone, conglomerate, and basaltic flows and pillow lava; metamorphosed

NORTHERN PILBARA

AFj JEERINAH FORMATION: carbonaceous mudstone and siltstone, thin-bedded sandstone, chert, felsic volcaniclastic rock, basalt, and dolomite; metamorphosed
AFjn Nallanaring Member: pillowed and massive basaltic flows, and basaltic breccia; includes felsic volcaniclastic rock locally; metamorphosed
AFjl Pillowed and massive basaltic flows, basaltic breccia, and mafic to felsic volcaniclastic sandstone; minor felsic volcaniclastic rock; includes the Barramine Volcanic Member in the Gregory Range; metamorphosed
AFji Isabella Member: fine-grained massive rhyolite; metamorphosed
AFjo Woodiana Member: quartzitic sandstone, chert (locally stromatolitic), chert breccia, and argillite; metamorphosed
AFm MADDINA FORMATION: massive to amygdaloidal basaltic flows and breccia, andesite, dacite, and rhyolite; local thin beds of sandstone and dolomite; metamorphosed
AFmk Kuruna Member: mafic to intermediate volcaniclastic sandstone, argillite, chert, and dolomite; local accretionary lapilli and stromatolites; metamorphosed
AFms Argillite, mafic to intermediate volcaniclastic sandstone, chert, and stromatolitic dolomite; local abundant accretionary lapilli; metamorphosed
AFt TUMBIANA FORMATION: mafic to intermediate volcaniclastic sandstone, argillite, basaltic flows and breccia, chert, stromatolitic limestone and dolomite, and calcareous sandstone; mainly lapilli tuff in East Pilbara; metamorphosed
AFtc Meentheena Member: stromatolitic limestone and dolomite, mudstone, siltstone, and mafic and felsic volcaniclastic sandstone; metamorphosed
AFtl Basaltic flows and associated volcaniclastic rock; metamorphosed
AFa Unassigned Fortescue Group basalt and minor sedimentary rock; metamorphosed; in Dampier Archipelago; possibly equivalent of KYLENA FORMATION
AFk KYLENA FORMATION: thin to very thick, massive or amygdaloidal basaltic flows, andesite, dacite, and rhyolite; minor basaltic pillow lava and basaltic breccia; local pyroxene spinifex-textured basalt and thin stromatolitic carbonate beds; metamorphosed
AFka Rhyolitic flows and breccia, and mafic and felsic volcaniclastic sandstone; metamorphosed
AFh HARDEY FORMATION: sandstone, mudstone, siltstone, conglomerate, felsic and mafic volcaniclastic sandstone and conglomerate, and basaltic flows and breccia; metamorphosed
AFhy Lyre Creek Member: massive, felsic to intermediate volcaniclastic sandstone and conglomerate; local abundant accretionary lapilli; metamorphosed
AFhb Basaltic flows and associated volcaniclastic rock; metamorphosed
AFhf Massive quartz–feldspar porphyry, medium- to thick-bedded porphyritic felsic breccia, and silt- or sand-sized felsic volcaniclastic rock; metamorphosed; includes Bamboo Creek Member, Koongaling Volcanic Member, Warrie Member, and Spinaway Porphyry
AFr MOUNT ROE BASALT: amygdaloidal basaltic flows and breccia, and basaltic volcaniclastic sandstone; metamorphosed
AFrb Basaltic breccia; metamorphosed
AFrs Quartzofeldspathic sandstone, conglomerate, mudstone, and siltstone; metamorphosed

Hamersley Basin; PILBARA CRATON

SOUTHERN AND NORTHERN PILBARA

Pilbara Supergroup (c. 2800–3500 Ma)

Apf Feldspar porphyry; metamorphosed

Af Aa Ac Aci

Af Rhyolite and dacite flows, and associated volcaniclastic rocks; metamorphosed
As Sedimentary rock, undivided; includes oligomictic and polymictic conglomerate, pebbly sandstone, mudstone, siltstone, and chert; metamorphosed
Ac Chert; metamorphosed
Aci Banded iron-formation, chert, and jaspilite; metamorphosed

Ab Au

Ab Basaltic and andesite flows, and associated volcaniclastic rocks; metamorphosed
Au Ultramafic rock, undivided; includes metamorphosed peridotite, dunite, pyroxene peridotite, and serpentinite, talc and amphibole schist

INTRUSIVE ROCKS

d Dolerite dykes, sills, and small intrusions (various ages)

Eg Granodiorite, biotite-bearing and mainly coarse-grained; contains mafic xenoliths; intruded into Wyloo Group

Egn Hornblende monzogranite and porphyry (age uncertain, but postdates≥ 2560 Ma)

AFd Medium- to coarse-grained dolerite intruded into Fortescue Group; metamorphosed

AFi Layered sills intruded into Fortescue Group in southern Pilbara; sills generally consist of a coarse-grained pyroxenite base, overlain by leucocratic gabbro or dolerite; local basal serpentinite; metamorphosed

AFdc COOYA POOYA DOLERITE: fine- to medium-grained dolerite; metamorphosed

AyG AyGo

AyG GOLDEY GRANOPHYRE: fine- to medium-grained granophyre, commonly porphyritic; includes hybrid rocks; metamorphosed
AyGo Gabbro containing quench-textured, acicular pyroxene crystals; metamorphosed

Agn Hornblende monzogranite and porphyry; metamorphosed

Agr GREGORY GRANITIC COMPLEX: metamorphosed, even-grained to porphyritic granitoid rock, porphyritic to seriate syenogranite, rapakivi-textured syenogranite, and granophyre

AFdb AFdbr Black Range Dolerite suite: dolerite, medium- to coarse-grained gabbro; weakly metamorphosed
AFdbr BLACK RANGE DOLERITE: dolerite and gabbro; weakly metamorphosed

Ad Adm

Ad Dolerite sills and dykes; metamorphosed
Adm Rafts of metamorphosed dolerite and gabbro; restricted to GREGORY GRANITIC COMPLEX

Aoo Aou

Layered intrusion
Aoo Gabbro and dolerite; metamorphosed
Aou Ultramafic rock; metamorphosed

Ao Gabbro sills and dykes; metamorphosed

Granitoid rocks, various ages

Agm Granite to granodiorite, variably foliated

Agp Agx

Agp Porphyritic granite; foliated
Agx Alkali feldspar granite, metamorphosed

Agf Agt Agx

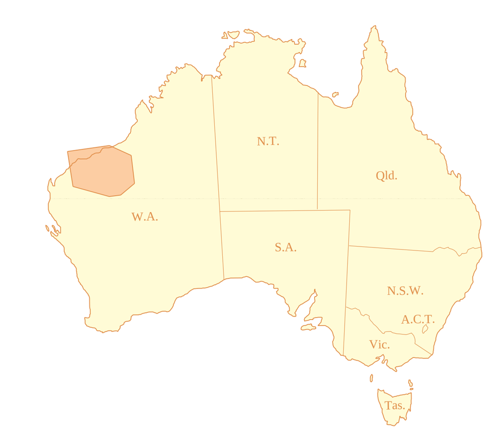
Agf Foliated, gneissic, or migmatitic biotite monzogranite, granodiorite, and tonalite; metamorphosed
Agt Tonalite and granodiorite, foliated
Agx Numerous greenstone xenoliths in foliated granitoid rock

SYMBOLS

Geological boundary
Fault
exposed
concealed
reverse
Major folds, showing axial trace and generalized plunge direction
anticline; exposed, concealed
syncline; exposed, concealed
overturned anticline; exposed
Bedding, showing strike and dip
horizontal
inclined
vertical
overturned
strike and dip estimated from aerial photography
0–15°
15–45°
45–90°
Igneous layering, showing strike and dip
inclined
Isotopic age determination site with identification number; Fortescue Group only
Highway with national route marker
Formed road
Track
Gas pipeline
Railway
Airport
Airfield
Landing ground
Townsite
population; less than 1000
1000–10000
greater than 10000
Point Samson
Dampier
KARRATHA
Millstream
Homestead
Building
National Park boundary
Reserve boundary
Horizontal control; major
Watercourse
Lighthouse
Abandoned
(abd)

Mineral field boundary
Mineral field district boundary
Mining group
WOODIE WOODIE
Mine or openpit from MINEDEX data (gold, unless otherwise indicated)
Bamboo Creek
Closed mine or undeveloped deposit from MINEDEX data
Closed mine or undeveloped deposit from published GSWA maps
Mining area or stockpiles
Mineral occurrence

Mineral	Symbol
Antimony	Sb
Asbestos, chrysotile	Ac
Asbestos, crocidolite	Ad
Barite	Ba
Beryl	Be
Building or dimension stone	Bs
Chromium	Cr
Chrysoprase	Chy
Copper	Cu
Diamond	Di
Fluorite	F
Gemstone	Gs
Gold	Au
Iron	Fe
Lead	Pb
Limestone	Ls
Lithium	Li
Manganese	Mn
Mica	Mi
Molybdenum	Mo
Nickel	Ni
Niobium	Nb
Platinum group element	PGE
Salt (solar evaporation)	Na
Silver	Ag
Tantalum	Ta
Tin	Sn
Titanium	Ti
Tungsten	W
Uranium	U
Vanadium	V
Zinc	Zn



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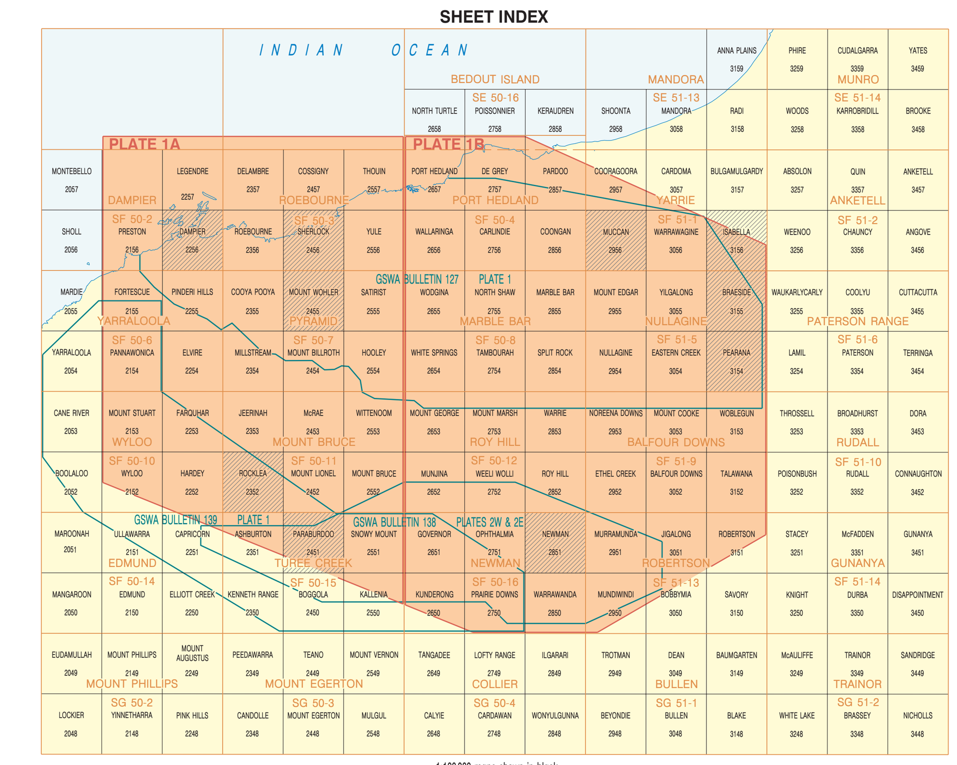

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DEPARTMENT OF MINERALS AND ENERGY
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GEOLOGICAL SURVEY OF WESTERN AUSTRALIA
DAVID BLIGHT, DIRECTOR

SCALE 1:500 000
5000 0 5 10 15 20 25 30 35 40 45 50
Metres Kilometres

UNIVERSAL TRANSVERSE MERCATOR PROJECTION
HORIZONTAL DATUM: GEOCENTRIC DATUM OF AUSTRALIA 1994
VERTICAL DATUM: AUSTRALIAN HEIGHT DATUM
Grid lines indicate 20 000 metre interval of the Map Grid Australia Zone 50

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

BULLETIN 144 PLATE 1C

GEOLOGY OF THE FORTESCUE GROUP — EAST PILBARA CRATON