GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

REPORT 19

PROFESSIONAL PAPERS FOR 1984

DEPARTMENT OF MINES
WESTERN AUSTRALIA

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D. R. KELLY

DIRECTOR OF THE GEOLOGICAL SURVEY
A. F. TRENDALL

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METAMORPHIC PATTERNS IN THE GREENSTONE BELTS OF THE SOUTHERN CROSS PROVINCE, WESTERN AUSTRALIA

by A. L. Ahmat

ABSTRACT

The greenstone belts of the Southern Cross Province have been variously metamorphosed between prehnite-pumpellyite and lower granulite facies, under low to intermediate pressures (200-500 MPa) and temperatures ranging up to 700°C. The observation that metamorphic grades are higher in the southern third of the province is consistent with the proposal, by Gee and others (1981), that there is a regional decrease of grades in a northeasterly direction across the Yilgarn Block. Within individual greenstone belts the distribution of metamorphic grades is largely concentric (and approximately U-shaped in cross-section), with higher grades around the margins and lower grades in the centres. The structure and sequence of metamorphic grades are best explained by a process of granitoid domes and diapirs warping and folding a pre-existing layer-cake supracrustal assemblage, within which a high geothermal gradient had already been established. The higher grade marginal zones are characterized by dynamic-type (high-strain) metamorphism and commonly show evidence of polyphase deformation and metamorphism. Field, petrographic and geochronological data suggest that this was related to continuous ("rolling") deformation and metamorphism, which occurred during a diapiric rise of the granitoid plutons in the comparatively short time interval of 2.7-2.6 Ga. There is little evidence of discrete, unrelated metamorphic events (except for the effects of local shear zones), contact metamorphism related to post-tectonic granitoids, or widespread retrograde metamorphism.

Evidence of an additional process—namely differential uplift of a lower crustal layer—that was involved in the metamorphism of the province is provided by: the juxtaposition of a granulite terrain to the western side of the Forrestania greenstone belt; the presence of granulite facies rocks; and an asymmetrical distribution of metamorphic patterns within the greenstone belt. It is believed that this additional process was penecontemporaneous with the granitoid doming and diapirism, occurring during the 2.7-2.6 Ga tectonothermal event, which was important in cratonizing much of the Yilgarn Block.

INTRODUCTION

The Archaean greenstone belts of the eastern Yilgarn Block of Western Australia have been variously metamorphosed between prehnite-pumpellyite and lower granulite facies (Binns and others, 1976). The metamorphism of greenstone belts has sometimes been viewed as an obstacle to investigations, as for example in rock identification and geochemical studies. But in recent years, metamorphic studies have been used to explain the complex tectonothermal history of the granitoid-greenstone belts, (Binns and others, 1976; Andersen and others, 1976; Glikson and Lambert, 1976; Archibald and others, 1976, 1978, 1981; Chapman and others, 1981; Gee and others, 1981; Hallberg and Glikson, 1981; Bickle and others, 1983). Although crustal evolution models for the Yilgarn Block differ (*cf.* Archibald and others, 1981, with Hallberg and Glikson, 1981) close temporal and spatial relationships between granitoid emplacement, deformation and metamorphism have now been demonstrated (Archibald, 1981; Bickle and others, 1983; McCulloch and others, 1983b). It is therefore evident that the documentation of metamorphic patterns, within the greenstone belts, will further the understanding of tectonic evolution in Archaean granitoid-greenstone terrains. In addition, metamorphism has been shown to be involved in some types of mineralization (Binns and others 1976; Barrett and others, 1976, 1977; Groves and others, 1979; Porter and McKay, 1981; Phillips and Groves, 1983).

This study deals with the distribution and origin of metamorphic patterns in the greenstone belts of the Southern Cross Province (Fig. 1A), and is part of a larger project aimed at producing a metamorphic map of the greenstone belts of the entire Yilgarn Block. With the recent completion of the 1:250 000 geological mapping of the area (Table 1), the Southern Cross Province was seen as an ideal starting point from which to extend, and partly complement, the first systematic study of regional metamorphism which was carried out by Binns and others (1976) for the eastern half of the Yilgarn Block (Fig. 2).

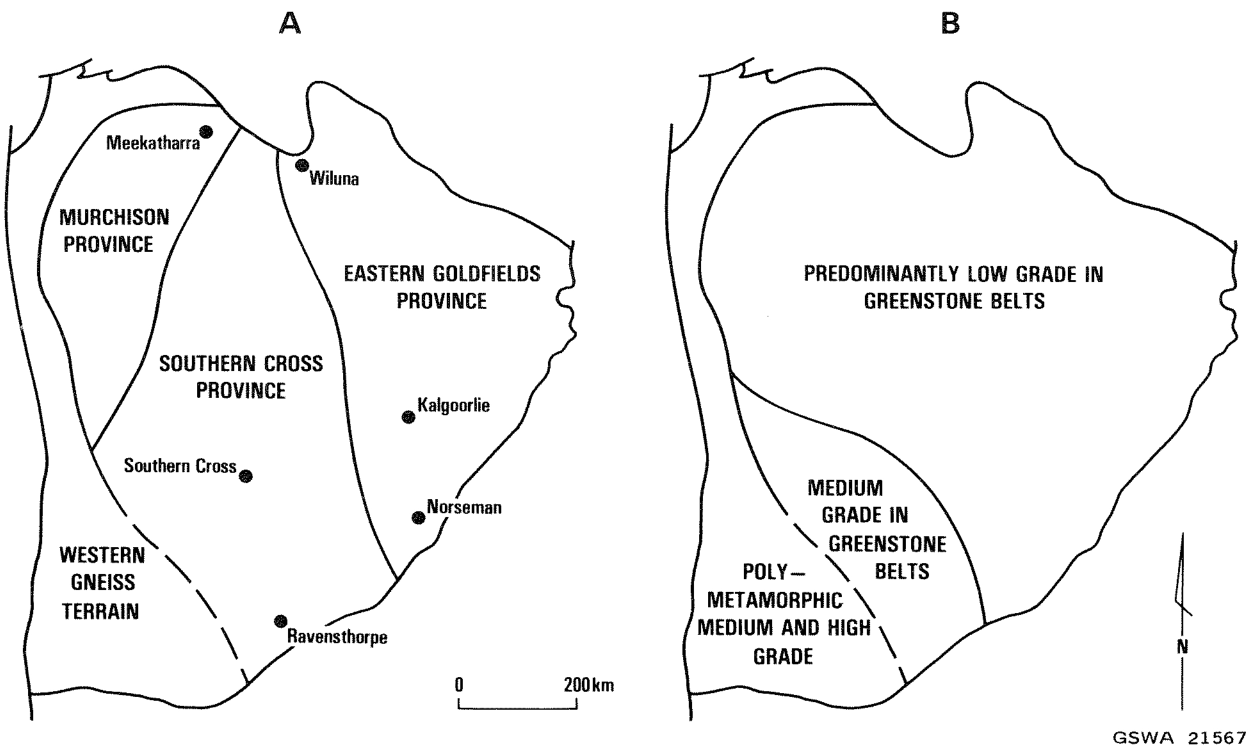


Figure 1. A—Regional subdivisions and B—gross metamorphic patterns of the Yilgarn Block (Gee and others, 1981).

TABLE 1. YEARS OF MAPPING AND PUBLICATION OF THE 1:250 000 GEOLOGICAL MAPS COVERING THE AREA STUDIED IN THE SOUTHERN CROSS PROVINCE.

Map Sheet	*Period of Mapping*	*Reference (Publication)*
Sandstone	1981	Tingey (1985)
Youanmi	1980	Stewart and others (1983)
Barlee	1979	Walker and Blight (1981)
Jackson	1979	Chin and Smith (1981)
Hyden	1978	Chin and others (1982)
Glengarry	1976-7	Elias and others (1982)
Southern Cross	1973-6	Gee (1982)
Lake Johnston	1971	Gower and Bunting (1976)
Menzies	1967	Kriewaldt (1970)
Kalgoorlie	1964-5	Kriewaldt (1969)
Boorabbin	1960	Sofoulis (1963)

While in general supporting the results of Binns and others (1976), the denser sampling used in the present study of the Southern Cross Province has allowed several refinements and modifications to be made to the previously published map of metamorphic patterns. These changes produce a regional picture which supports the hypothesis proposed by Gee and others (1981) that metamorphic grades broadly decrease in a northeasterly direction across the Yilgarn Block (Fig. 1B). However, it is clear that, eventually, more detailed studies may be carried out and these will improve upon the map produced in this report.

PREVIOUS METAMORPHIC WORK

Until the comprehensive regional study by Binns and co-workers in the early 1970s, metamorphic studies in the Southern Cross Province were very scanty, and mainly restricted to local areas (*e.g.* gold mines). Rocks in the greenstone belts were generally referred to rather simply as having been "deformed" and "metamorphosed". Except in the vicinity of a few gold mining leases, virtually no metamorphic data were available for the northern half of the province. However, in the southern third of the province, a number of occurrences of sillimanite, staurolite, andalusite and garnet had been recorded (Saint-Smith and Farquharson, 1913; Blatchford, 1915; Carroll, 1936, 1939; Ellis, 1939; see also Simpson, 1952 for a summary). Prider (1948; Presidential address delivered to the Royal Society of Western Australia in 1945) made one of the first statements on the grade of metamorphism in the Southern Cross Province, and used this information to suggest that "the greater part of the Yilgarn area (Southern Cross district) lies within the sillimanite zone". The "sillimanite zone", named after the

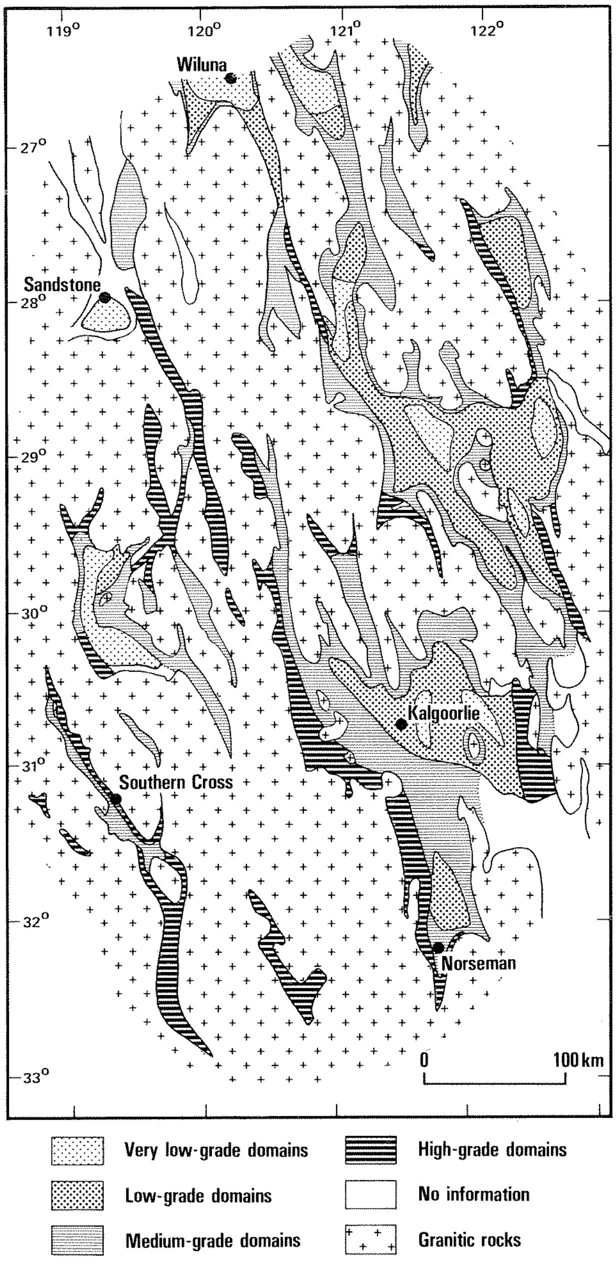


Figure 2. Metamorphic map of the eastern Yilgarn Block (Binns and others, 1976). Compare with Figure 5.

highest grade of Barrow's zones of metamorphism in the Scottish Highlands (Barrow, 1912), is equivalent to the upper amphibolite facies. Prider also made the broader statement that the metamorphic grade across the Yilgarn Block, between Perth and Kalgoorlie, decreased from west to east: a view that is now widely accepted (Gee and others, 1981).

At approximately the same time, Miles (1946) recognized that many banded iron-formations (BIF's) in the "Yilgarn Goldfields" were high-grade metamorphic rocks. The iron-rich amphibole, grunerite (in places cummingtonitic), was recorded by Miles from the Mount Palmer district, Koolyanobbing Range, Nevoria, Southern Cross and Corinthian. He also noted that grunerite was commonly well developed near intrusive granite contacts, and therefore suggested that the high-grade BIF's were produced principally by contact metamorphism. At Heaney's Find, where fayalite and hedenbergite are locally present, he concluded that the metamorphic grade was probably slightly higher than at most other places in the Southern Cross district, but was still related to contact metamorphism by the intruding granitoids.

Wilson (1953) determined that the metamorphic grade at Nevoria belonged to Turner and Verhoogen's (1951) "cordierite-anthophyllite subfacies" of the amphibolite facies. This subfacies was later called the hornblende-hornfels facies (Fyfe, Turner, and Verhoogen, 1958). Bekker (1957) established the same metamorphic grade for the Southern Cross area, and Williamson and Barr (1965) suggested that this was the grade throughout much of the Southern Cross district.

Following the nickel boom in the late 1960s, large numbers of new areas, outside existing gold mining districts, were explored by mining companies, and this eventually led to a marked increase in metamorphic studies (Porter 1971, 1982; Bettennay, 1972; Davis, 1972; Peters, 1972; Witt, 1972; Sullivan, 1973; Tyrer, 1974; King, 1974; Walker, 1974; Barrett and others, 1976, 1977; Binns and others, 1976; Purvis, 1978; Porter and McKay, 1981). Many of these studies were related to the regional metamorphic project instigated by the University of Western Australia in 1972, pertaining to the relationship between metamorphism and nickel mineralization.

Other recent metamorphic studies include P-T estimates from metamorphosed BIF at Heaney's and Meier's Find (Gole and Klein, 1981), and from metapelites at Westonia (Blight and Barley, 1981). Gore (1981) has documented, in detail, assemblages of metamorphosed BIF's throughout the Yilgarn Block, and brief accounts of metamorphism are found in most of the explanatory notes accompanying the 1:250 000 map sheets (listed in Table 1).

CURRENT METAMORPHIC VIEW

Binns and others (1976) presented important new data on Archaean metamorphism, which showed that metamorphic patterns in the eastern Goldfields were distinctive and could not be directly compared with Phanerozoic metamorphic-tectonic environments, and that metamorphic grades were governed largely by structural setting (and not stratigraphy) within broadly synformal greenstone belts. They recognized two contrasting

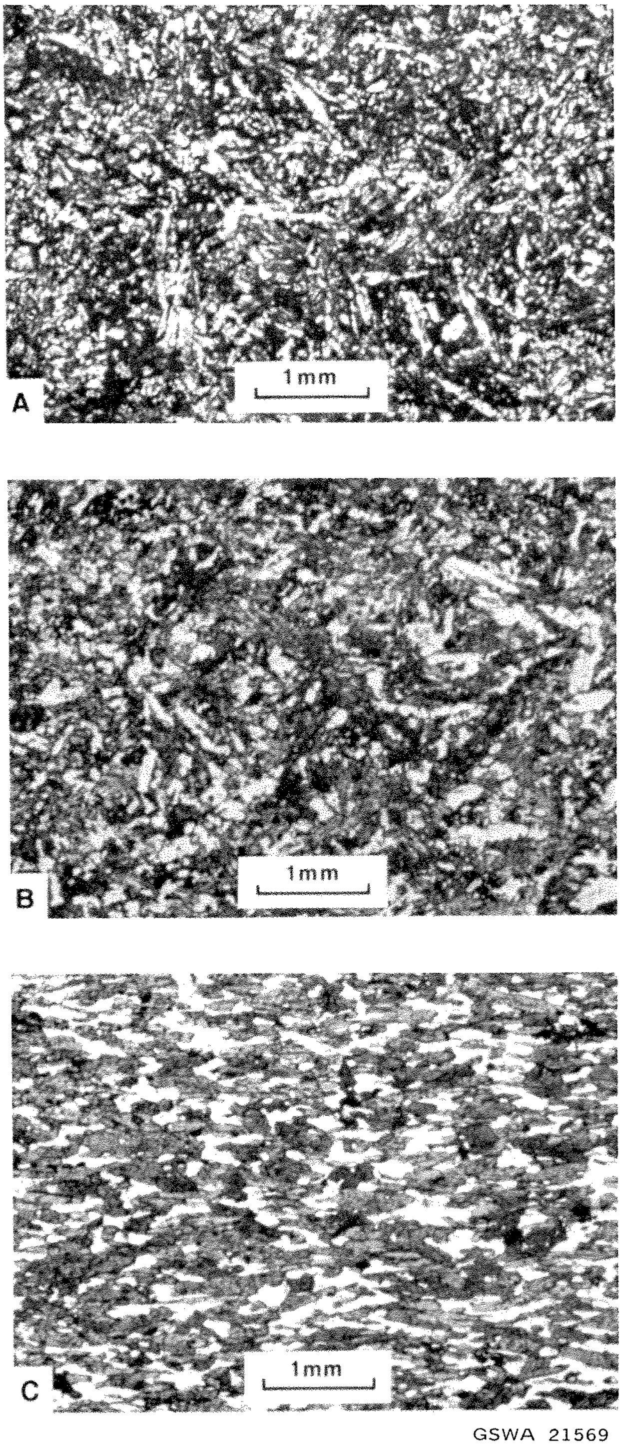


Figure 3. **Photomicrographs of tholeiitic metabasites to illustrate variation in metamorphic grade and style. A—Metabasalt with relic clinopyroxene and plagioclase from a very low-grade static-style domain. Sample 73726, 5.7 km southwest of Mount Jackson, Marda greenstone belt (plane-polarized light). B—Metabasalt with hornblende and partly recrystallized labradorite from a medium-grade domain. Although moderately recrystallized and weakly foliated, relic texture is preserved. Sample 73765A, Bell Chambers area, Sandstone greenstone belt (plane-polarized light). C—Thoroughly recrystallized metabasite from a high-grade, dynamic-style domain, showing aligned hornblende and granoblastic plagioclase. Sample 28777, 6 km south-southeast of Hatters Hill, Forrestania greenstone belt (plane-polarized light).**

styles of metamorphism: static-style metamorphism, characterized by weak penetrative fabrics and associated with prehnite-pumpellyite to mid-amphibolite facies (Fig. 3A); and dynamic-style metamorphism, characterized by penetrative fabrics and associated with mid-amphibolite to lower granulite facies (Fig. 3C).

Dynamic-style metamorphism is generally restricted to linear belts along the present margins of greenstone belts, which are typically characterized by high-strain and synkinematic granitoid diapirs. Binns and others suggested that, within the constraints imposed by regimes of low- to moderate-pressure, these higher grade, high-strain zones represented deep-seated rocks uplifted by diapirism relative to the central portion of the belts. Static-style metamorphism is generally restricted to central portions of the greenstone belts, and a close correlation with Bouguer anomalies in these areas led Binns and others to the conclusion that metamorphic grade and metamorphic style were related to structural height above a granitic substratum. They also concluded that the large-scale metamorphic patterns arose during a single, widespread event which was linked with the generation or remobilization of granitic bodies. This was based on the evidence of: the ubiquity of radiometric dates around 2.6-2.7 Ga; the apparent contemporaneity of static- and dynamic-style metamorphism; and the unusual distribution of relic igneous mineral phases in mafic and ultramafic rocks at various metamorphic grades.

The above hypothesis was elaborated upon by Archibald and others (1978, 1981), and has become the standard "sialic" model of granitoid-greenstone evolution and consequent metamorphism. However, the model has been rejected by other workers (Glikson, 1976; Glikson and Lambert, 1976; Hallberg and Glikson, 1981), who believe that: there have been two distinct periods of mafic-ultramafic development (the "lower" and "upper" greenstones); the granitoids have been derived by partial melting of the greenstones; and there have been distinct, separate phases of metamorphism. The debate is still unresolved (McCulloch and others, 1983b), but recent geochronological studies clearly demonstrate that the time span, between granitoid-greenstone generation and the period of tectonism-metamorphism, was comparatively short (Chapman and others, 1981; McCulloch and Compston, 1981; Bickle and others, 1983; McCulloch and others, 1983). Bickle and others (1983b), for example, suggest that the Archaean crust in the Diemals area formed between 2.80 and 2.65 Ga ago, and that the main phase of granitoid development (banded gneisses, syn- and post-kinematic granitoids), with its associated deformation and metamorphism, occurred between about 2.723 and 2.685 Ga ago.

The geochronological studies of de Laeter and others (1981), Nieuwland and Compston (1981), and McCulloch and others (1983a) provide evidence of high-grade metamorphism prior to about 3.2 Ga in the Western Gneiss Terrain. No evidence of this is found in the Southern Cross or Eastern Goldfields Provinces, where published dates of crustal formation generally do not exceed 2.8 Ga (McCulloch and others, 1983b). However, ages of about 3.0 Ga (Sm-Nd) are now available (Fletcher and others, 1984) for the Diemals-Marda area and Claoué-Long and others (1984) have recently dated the Kambalda greenstones at 3.26 Ga. This regional variation in ages, of both crustal development and metamorphism across the Yilgarn Block, does not support the model of Glikson and Lambert (1976), in which variations in metamorphic grade are interpreted in terms of coeval, vertical zonation.

METHODS

Sampling

The metamorphic study is based on the examination of about 1 030 thin sections from the collection of the Geological Survey of Western Australia, representing some 650 data points (Fig. 4). Most of the rocks were collected during systematic 1:250 000 geological mapping of the region (Table 1). Many of these were collected solely for rock-type identification, and they have little diagnostic metamorphic value. In addition, detailed structural relationships are not known for most of the rock samples, thus precluding petrofabric studies. Sample points have been transposed from aerial-photography data and plotted at a scale of 1:250 000 on bedrock maps, obtained from the explanatory notes of the respective map sheets. The error associated with plotting the sample is generally less than 200 m, which is insignificant at the scale of the map produced in this report. As will be discussed later, there is more inaccuracy associated with the position of the margins of the greenstone belts.

Metamorphic grades

The grades are based largely on the mineralogical criteria given by Binns and others (1976), who establish four metamorphic domains namely: very low grade; low grade; medium grade; and high grade (Table 2). These domains do not correspond

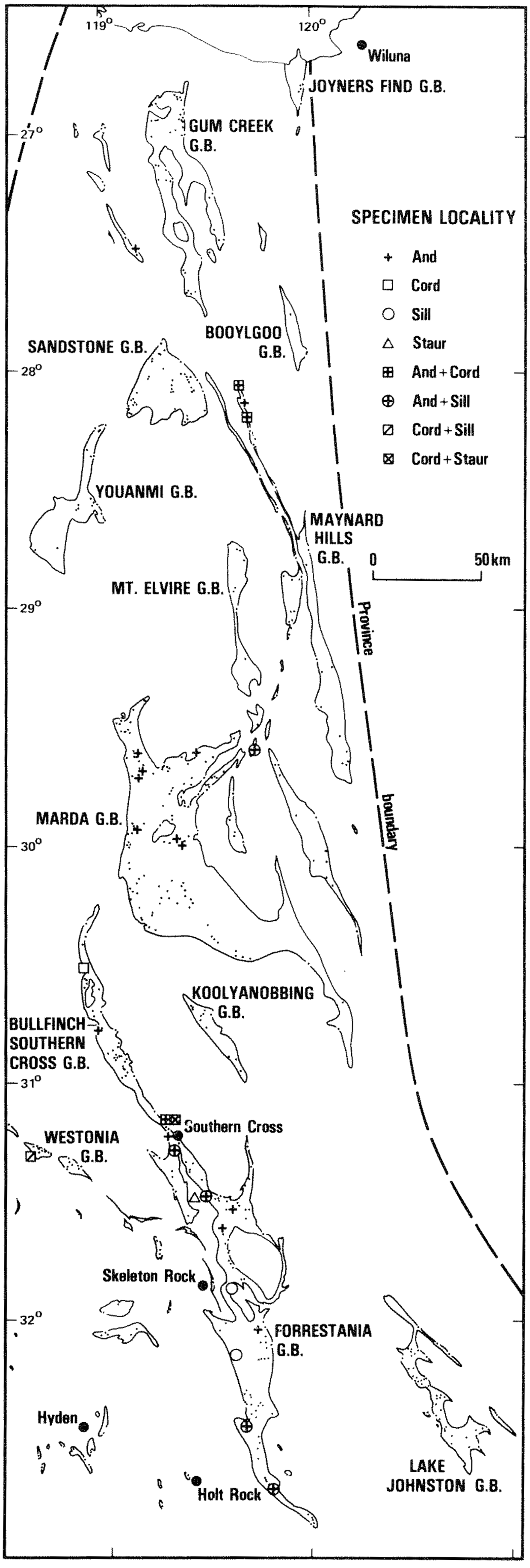
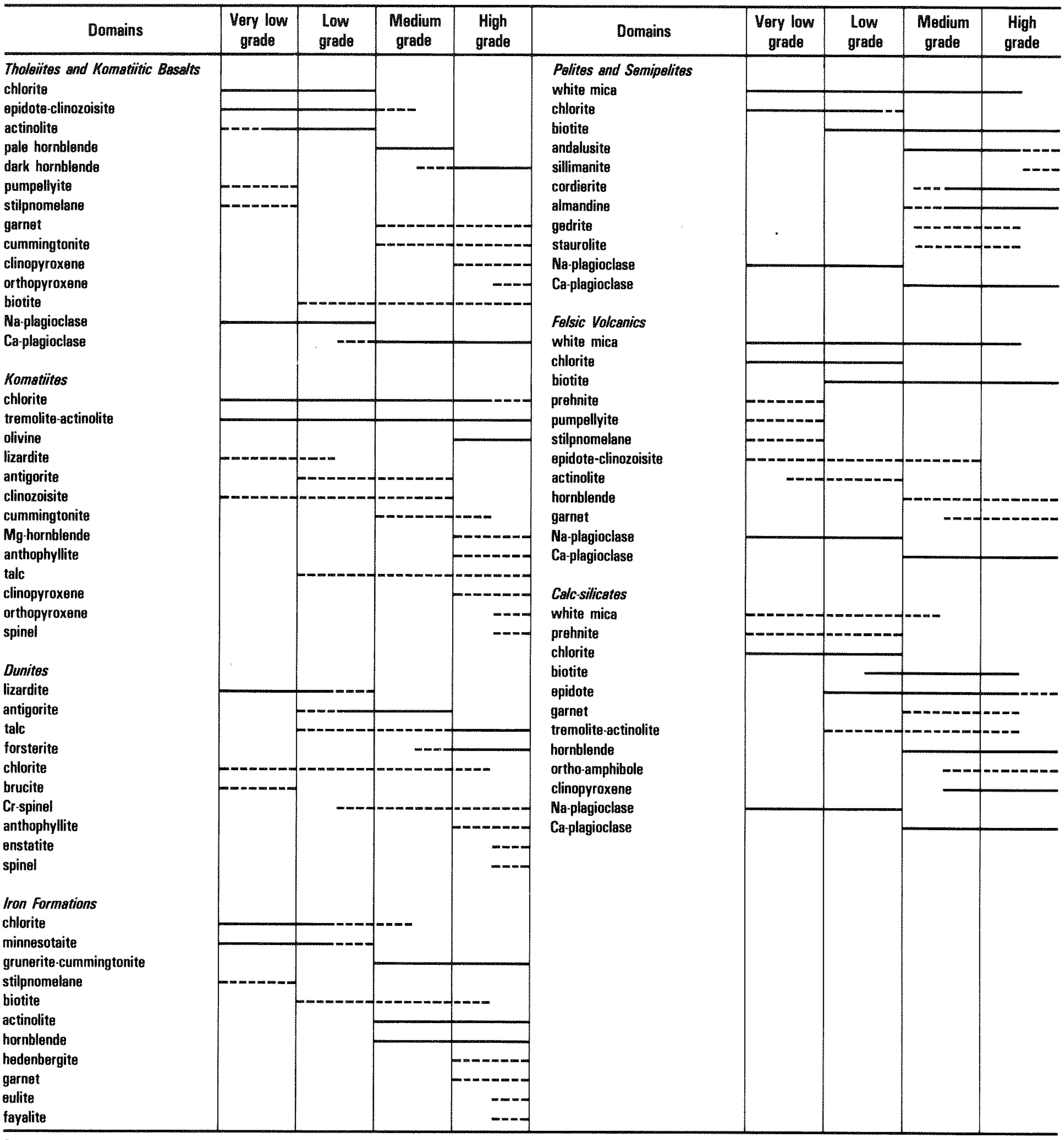


Figure 4. Sample location (dots) and distribution of andalusite, cordierite, sillimanite and staurolite.

TABLE 2. METAMORPHIC ASSEMBLAGES IN THE SOUTHERN CROSS PROVINCE.



Quartz occurs in all assemblages except dunites and komatiites. K-feldspar occurs in felsic volcanics, metasediments and calc-silicates. Carbonate and opaque assemblages, and titanite (sphene), not listed. Domains are as shown on Figures 2 and 5, but do not correspond to the nomenclature of Winkler (1979) (see Table 3).

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TABLE 3. SUBDIVISIONS OF METAMORPHIC GRADE.

Turner (1981)	Prehnite-pumpellyite facies	Greenschist facies	Amphibolite facies		Granulite facies
Winkler (1979)	Very low grade	Low grade	Medium grade		High grade
Binns *et al.* (1976); this study	Very low grade	Low grade	Medium grade	High grade	

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exactly to Winkler's (1979) classification (Table 3), which is based largely upon reactions in pelitic rocks, but the domains have been found suitable for the Yilgarn Block where pelitic rocks are scarce. Precise subdivision and correlation with the conventional classification must await the availability of microprobe data.

A number of additions and small changes have been made to the table of mineral parageneses given by Binns and others (1976).

Additions include the following: a section on calc-silicates; inclusion of komatiitic basalts in the general basaltic group; the minerals pumpellyite, garnet and cummingtonite are added to the parageneses of tholeiites and komatiitic basalts; the minerals cummingtonite, biotite, hornblende and garnet are added to the BIF paragenesis; and the mineral gedrite is added to the pelitic and semipelitic assemblages.

Minor changes include: the consideration of cordierite and garnet as common phases of pelitic rocks in medium-grade domains; and the adjustment of the cordierite range to a slightly higher grade, in order to conform with the conventional view that the first appearance of cordierite marks the beginning of the medium-grade (or amphibolite) facies as defined by Winkler (1979). Pyroaurite, stichtite, and chloritoid have not been observed in the Southern Cross Province and are therefore excluded from Table 2.

The metamorphic map produced from this study is shown in Figure 5, and is discussed in more detail in the subsequent section.

GEOLOGY

The geology of the province is shown on the respective map sheets (Table 1) and described in several of the references cited earlier (Gee and others, 1981; Williams 1975; Nesbitt and others, 1984). Briefly, the greenstone belts consist predominantly of metavolcanic and metasedimentary rocks, but also include mafic to ultramafic sills and dykes. The belts are arcuate in plan and synformal in cross section, and trend mainly north-northwest; several trend north-northeast (*e.g.* the Evanston-Diemal area). Up to four deformational events have been recognized throughout the province (Archibald and others, 1981), but in most places it is difficult to recognize more than two. The main deformational event (equivalent to the D_2-D_3 of Archibald and others, 1981, and the D_1 of Porter and McKay, 1981) was associated with the diapiric intrusion of granitoids, and resulted in the synformal folding of the greenstone sequences and widespread metamorphism. The next deformational event, that can be recognized, was associated with small-scale folding and crenulation of pelitic schists.

Metabasites, predominantly metamorphosed basaltic rocks, represent the bulk of the rocks examined. Other rock types included in the study are, in order of decreasing abundance: ultramafics;

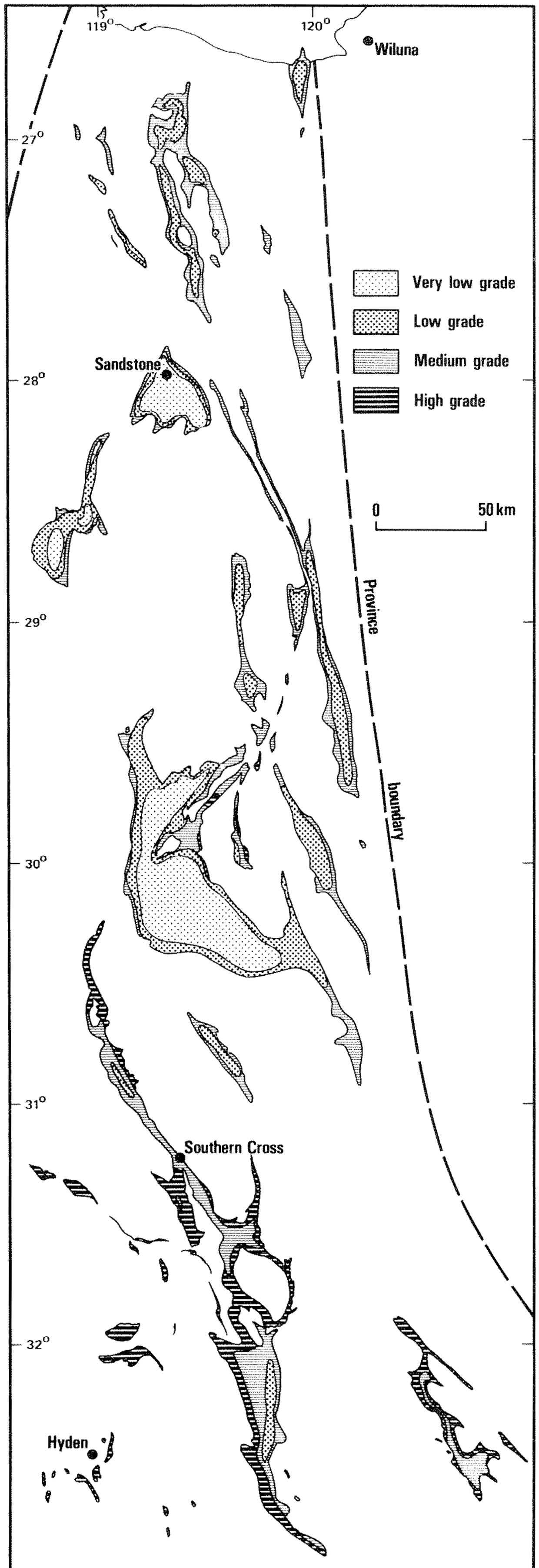

Figure 5. Metamorphic map of the greenstone belts of the Southern Cross Province. Compare with Figure 2.

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banded iron-formations; quartzofeldspathic rocks; semipelitic and pelitic rocks; and calc-silicates. Pelites and semipelites are relatively rare and, where exposed, are commonly deeply weathered. Fresh pelites crop out at Westonia, the Southern Cross area (*e.g.* Hopes Hill), and at several places in the northern portion of the Maynard Hills greenstone belt. Some fresh pelites have been encountered in diamond-drill core at Forrestania.

Retrograde alteration

The majority of rocks examined show evidence of retrograde alteration. It is most conspicuous in the higher grade rocks: for example, amphibolites contain granoblastic plagioclase grains which are zoned or commonly replaced by white mica (Fig. 6A), and hornblende crystals which have developed ragged terminations and/or have been overgrown by actinolite. Metabasites of lower grade generally contain plagioclase that has been saussuritized or replaced by coarsely crystalline clinozoisite, and amphibole that is chloritized. Higher grade ultramafics may show serpentinization and talc-carbonate alteration, and semipelitic to pelitic rocks may be sericitized, chloritized, and (where originally cordierite-bearing) pinitized.

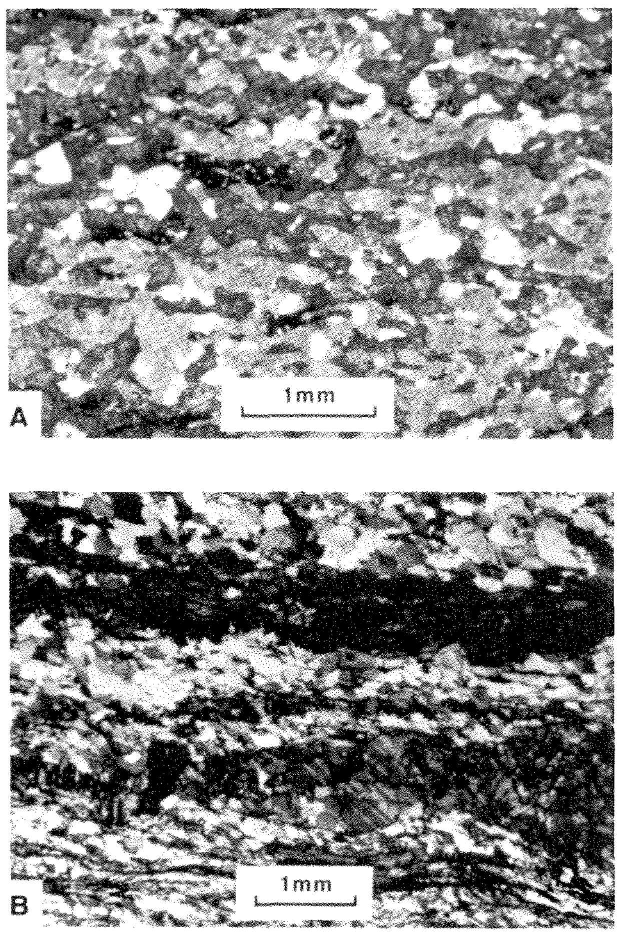

Figure 6. Photomicrographs to illustrate late-stage alteration. A—Foliated amphibolite containing granoblastic plagioclase clouded with sericitic alteration products. Clear areas represent unaltered plagioclase. Sample 74931A, high-grade, dynamic-style domain, Round Top Hill, Lake Johnston greenstone belt (plane-polarized light). B—Deformed grunerite-bearing meta-BIF containing highly strained and sutured quartz grains. Sample 74983, southern shore of Lake Deborah East, Koolyanobbing Range (crossed polars).

Much of the retrograde alteration can be related to late-stage deformation which has in many places produced cataclastic features and recrystallization of previous, well-crystallized (*e.g.* granoblastic) textures (Fig. 6B). Some of it, however, is probably related simply to retrograde metamorphism following peak P-T conditions (*e.g.* the white-mica replacement of polygonal plagioclase in amphibolites, which still retain relic granoblastic-polygonal fabrics).

Because of the widespread, retrograde alteration, evidence of peak metamorphic conditions may be locally obscured and some metamorphic grades may therefore be underestimated. Future studies will probably result in slight alterations to the isograd boundaries.

Timing of crystallization

Petrographic studies and field relationships suggest that the main phase of metamorphism occurred in both static and dynamic domains. The two domains are intergradational (Binns and others, 1976; Archibald and others, 1978). In static domains there is no evidence to suggest polymetamorphism as having been important (*e.g.* relic pre-metamorphic textures are well preserved). Yet in dynamic domains, there is evidence of both single and polyphase deformation and metamorphism. Single deformation or metamorphism in dynamic domains can be seen in many metabasites (*e.g.* metabasalts, metadolerites and metagabbros) where, despite penetrative fabrics, relic blasto-intergranular, -subophitic and -ophitic textures can still be distinguished. In rocks showing multiple deformation and metamorphism two foliations are present (Fig. 7), and in some of the pelitic rocks, helicitic garnets show evidence of an earlier foliation, which is not visible in the host rock (Fig. 8A, B). In dynamic domains, the common developments of granoblastic polygonal textures and randomly orientated porphyroblasts indicate that crystallization outlasted the main period of deformation of the greenstone belts (Fig. 8). Post- or late-syntectonic minerals include: porphyroblasts of amphibole (Fig. 8C); biotite and chlorite in mafic rocks; anthophyllite (Fig. 8D), tremolite (Fig. 8E), olivine (Fig. 8E), talc and chlorite in ultramafic rocks; and large, elongate, helicitic poikiloblasts of andalusite and cordierite

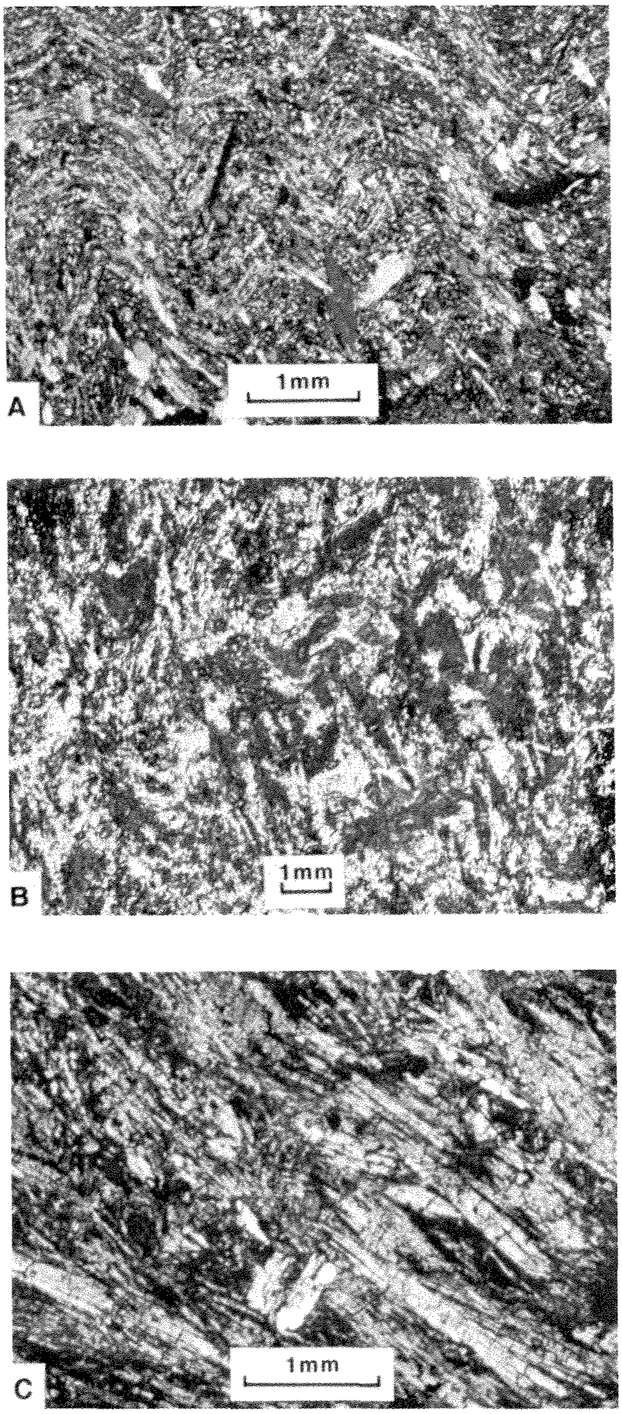

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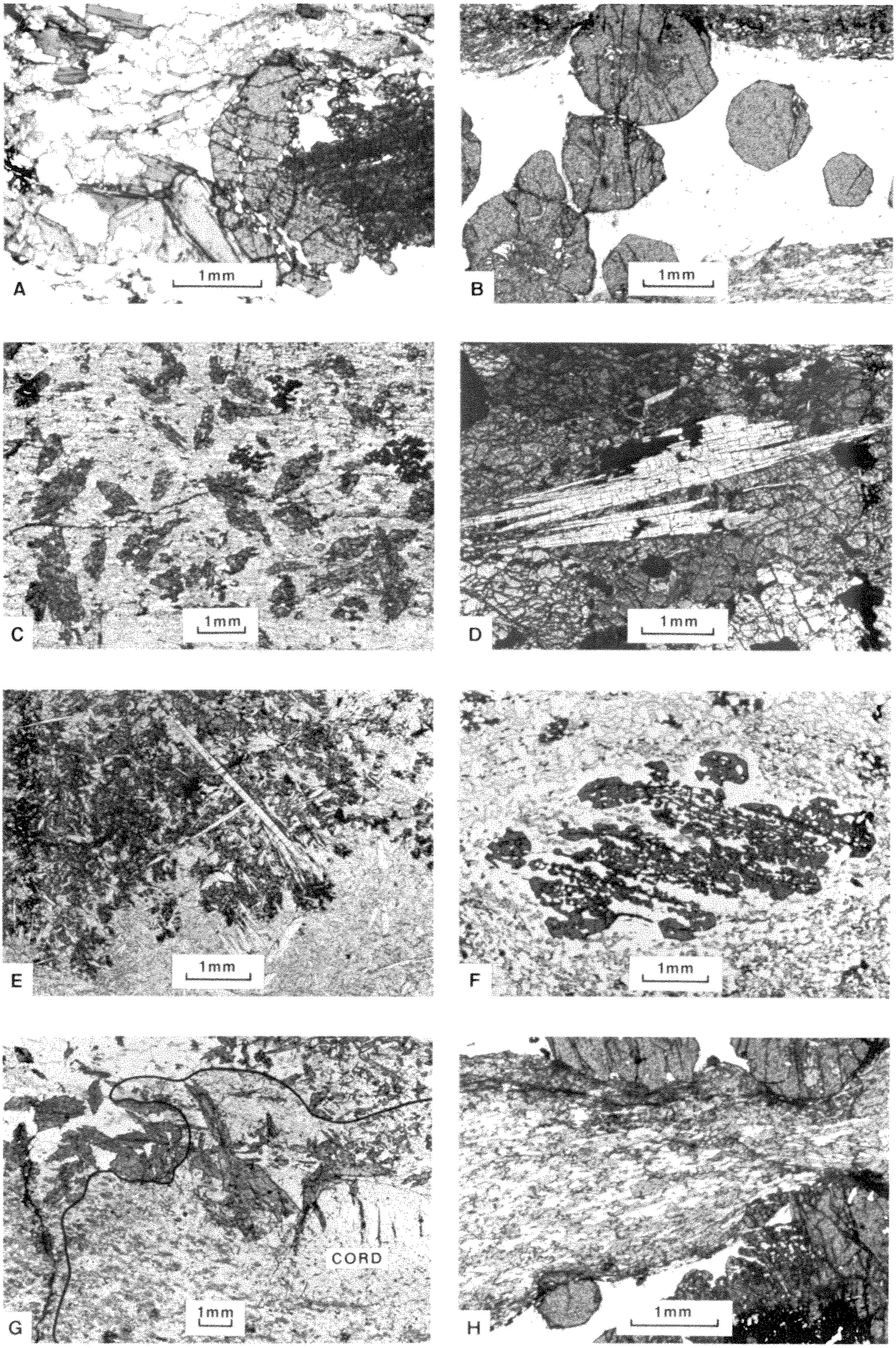
Figure 7. **Photomicrographs to show polyphase deformation/metamorphism. A—Crenulated amphibolite from a medium-grade domain. *Porphyroblasts of hornblende* have developed within both foliations, suggesting crystallization during penecontemporaneous deformation. Sample 73833A, 1.8 km southwest of Red Knob, northern Maynard Hills greenstone belt (crossed polars). B—Poly-deformed metabasite (metagabbro) with hornblende of similar composition in both foliations. Sample 74923, Hatters Hill, Forrestania greenstone belt (plane-polarized light). C—Anthophyllite-chlorite schist with porphyroblasts of anthophyllite aligned at a high angle to a foliation defined by the chlorite. Sample 59551B Evanston, northeast portion of the Marda greenstone belt (crossed polars).**

in pelites (Fig. 8G). Garnet crystallization extended over a wide time span (Fig. 8A, B, F) and the minerals may show post- to pre-tectonic features in the same rock: for example, cores with rotational textures surrounded by massive rims (Fig. 8A, H).

The above evidence of polyphase deformation and crystallization is viewed as a continuum, and not as a sequence of discrete events. This agrees with the interpretations made for the Norseman-Widgiemooltha belt (Archibald and others, 1978), for the Forrestania area, (Porter and McKay, 1981), and for much of the Yilgarn Block (Gee and others, 1981). It is also consistent with the geochronological data which suggest a comparatively short period of intense igneous and metamorphic activity between about 2.8 and 2.6 Ga ago. Consequently, there is little textural (or geochronological) evidence of regional-scale superimposition of separate, unrelated metamorphic events except for the effects of retrograde alteration (discussed above). Mantled grains reflecting progressive metamorphism have not yet been observed in the Southern Cross Province, and this, together with the unusual distribution of relic igneous phases in some mafic and ultramafic rocks (Binns and others, 1976) (Fig. 9), is consistent with the concept of a single phase of rapidly imposed metamorphism, whereby higher grade domains did not pass through an earlier phase of recrystallization equivalent to that present in lower grade domains. This conclusion is disputed by Hallberg and Glikson (1981), who maintain that the unusual distribution of relic phases relates to hydrous versus anhydrous metamorphic conditions (and consequently there was progressive metamorphism) and that distinct metamorphic events did occur.

Metasomatism

Metasomatism has been demonstrated at the local scale (Miles, 1942; Wilson, 1953; Porter and McKay, 1981, p. 1529), but as yet there is no evidence to suggest that it has occurred on a regional scale. The most common manifestation of local metasomatism is the development of porphyroblasts of diopsidic clinopyroxene in amphibolites (Fig. 10). The mineral usually occurs in distinct layers or discordant veins, and displays equilibrium textures consistent with those of the host, thus indicating that it developed during the main metamorphic event. In many of these amphibolites, the clinopyroxene is associated with quartz-epidote(-garnet)(-sphene), which is a typical calc-silicate assemblage. Most of these rocks occur within several hundreds of metres of granitoid bodies.



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Other examples of metasomatism are the development of porphyroblasts of carbonate and sulphide crystals, especially in retrograded schists associated with shear zones of gold-mineralized areas, and the widespread development of talc-carbonate alteration in metadunites, such as in the Forrestania greenstone belt (Porter and McKay, 1981).

METAMORPHIC PATTERNS

The main features shown by the metamorphic map (Fig. 5) are listed below:

(1) A complete range in grade from very low to high metamorphic grade is found. Very low-grade domains occur mainly in the central and northern portions of the province (*e.g.* Marda and Sandstone greenstone belts, respectively), whereas high-grade domains occur predominantly in the southern portion (*e.g.* Forrestania-Southern Cross greenstone belt). Mafic granulites (many with co-existing ortho- and clinopyroxene) are abundant in the western part of the Hyden sheet, and a few occur close to the western side of the Forrestania - Southern Cross greenstone belt (*e.g.* 4 km east-northeast of Skeleton Rock).

The general distribution of metamorphic domains agrees with the model proposed by Gee and others (1981), with high-grade domains restricted principally to the southern portion of the province (Fig. 1B).

(2) In most places there is a regular zonation of domains within the greenstone belts, with higher grades occurring around the margins, and lower grades in the centres (*cf.* Binns and others, 1976). The higher grade margins are characterized by dynamic-style metamorphism and the cores by static-type metamorphism.

Most metamorphic domains vary between 1 and 10 km in surface width, and many typically range between 2 and 5 km. Greenstone belts less than about 5 km wide (and ranging up to about 10 km) are therefore usually characterized by a specific metamorphic grade.

(3) In several places, however, asymmetrical patterns exist (*e.g.* in the Forrestania and Mount Manning greenstone belts).

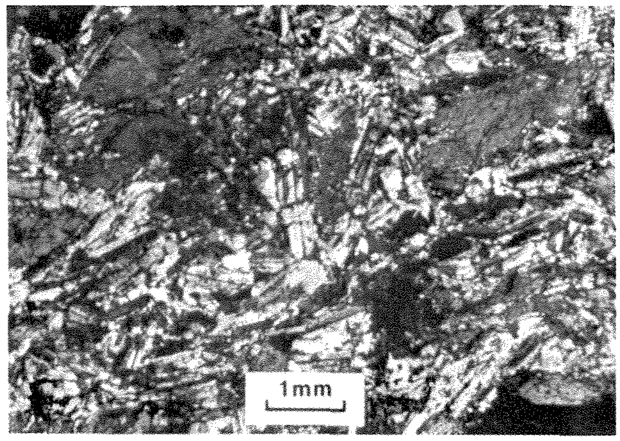

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Figure 9. Photomicrograph of a metagabbro from a medium-grade domain with relic labradorite laths. Although *all* the former pyroxenes have been replaced by hornblende, most of the plagioclase is relic. Locally it has recrystallized into granoblastic andesine. Sample 73771B, Black Range, Sandstone greenstone belt (crossed polars).

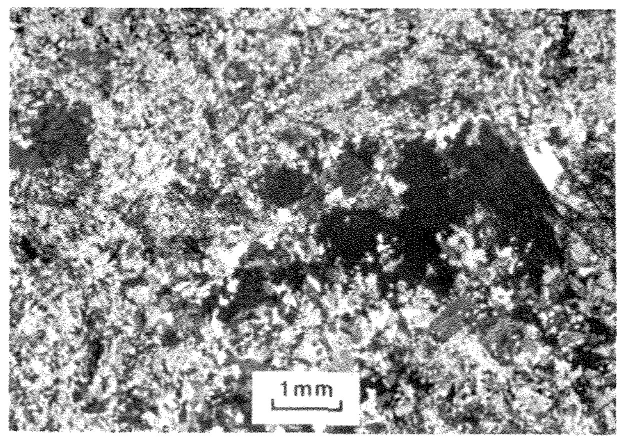

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Figure 10. Photomicrograph of a diopside vein in a high-grade foliated amphibolite. Sample 74970, 3.5 km northwest of Bodallin (crossed polars).

Figure 8. Photomicrographs illustrating textural and mineralogical evidence of polyphase metamorphism and deformation. A—Garnet contains inclusions which represent at least two growth stages. Sigmoidal pattern in the outer zone suggests rotation during syntectonic growth. Sample 74905, pelitic schist, 1.8 km southwest of Red Knob, Maynard Hills greenstone belt (plane-polarized light). B—Garnets with syntectonic rotational cores and late syntectonic to post-tectonic subhedral rims. Sample 73707, semipelitic schist, Hopes Hill area, 8 km northwest of Southern Cross (plane-polarized light). C—Post-tectonic, randomly-orientated porphyroblasts of gedrite in a semipelitic schist. Garnet is syn- to post-tectonic. Sample 74980A, Hopes Hill area, 8 km northwest of Southern Cross (plane-polarized light). D—Post-tectonic anthophyllite in a weakly foliated metadunite. Sample 60870, Liquid Acrobat, Forrestania greenstone belt (crossed polars). E—Post-tectonic olivine (aggregates of anhedral crystals) and tremolite in a tremolite-chlorite schist. Sample 37839, Agamemnon, 13 km south-southeast of Southern Cross (plane-polarized light). F—Pre- to syntectonic garnet in pelitic schist. Sample 73834D, 1.8 km southwest of Red Knob, Maynard Hills greenstone belt (plane-polarized light). G—Syntectonic cordierite (large porphyroblast—cord.), spongy andalusite (high relief, lower left) and post-tectonic gedrite. Sample 74979B, semipelitic schist, Hopes Hill area, 8 km northwest of Southern Cross. H—Small post-tectonic garnet enclosing thin layer of opaques. Larger garnet is pre- to syntectonic. Same sample (74905) as Fig. 8A.

(4) Metamorphic domains cut across the stratigraphy in a number of places. This feature and the asymmetrical metamorphic patterns support the idea that the control of metamorphism is structural rather than stratigraphic, as originally proposed by Binns and others (1976).

(5) Lower grade domains correlate with regional Bouguer anomalies, a feature which Binns and others (1976) interpreted as indicating structural control of metamorphism: structurally high sequences are less metamorphosed.

The rocks exposed within the lower grade domains show differences between the various localities. For example, the very low-grade domain of the Marda greenstone belt comprises a very large component of felsic volcanics (the Marda Complex) as well as tholeiitic and komatiitic basalts, whereas the very low-grade Sandstone greenstone belt comprises predominantly tholeiitic basalts, and the smaller very low-grade domains within the Youanmi area mainly occur over gabbroic rocks. The evidence, assuming the gravity anomalies are roughly proportional to the depth of the greenstone sequences, does not necessarily support a unified stratigraphy.

(6) Several of the small greenstone belts (excluding those within the granulite-facies domain in the southwest) do not show a higher metamorphic grade than adjacent larger ones. This suggests that they are not the "deep roots" of eroded greenstone belts (Glikson and Lambert, 1976).

Comparison with the metamorphic map of Binns and others (1976)

The distribution of the regional metamorphic patterns is basically the same as that of Binns and others (1976). However, there are a number of major changes involving the distribution of metamorphic grades.

Firstly, there is reasonable evidence to suggest that the Forrestania and Lake Johnston greenstone belts should each be subdivided into smaller domains; Binns and others (1976) have designated both areas as high-grade domains. Also, the lower grade domains recognized in the Forrestania greenstone belt correspond to a substantial, positive, Bouguer anomaly. The latter improves the overall correlation between structural depth and metamorphic grade, as first established by Binns and others (1976) for most other parts of the goldfields.

Secondly, there seems enough evidence to suggest that the area immediately south and southwest of Southern Cross (*i.e.* the Yilgarn Range) can be put into the high-grade category.

Thirdly, Binns and others (1976) designated the Mount Elvire greenstone belt as high grade, whereas this study suggests it is mainly composed of low- to medium-grade domains.

Fourthly, all of the Maynard Hills greenstone belt was designated as high grade, whereas this study suggests it is mainly low to medium grade, with only local areas of high-grade metamorphism.

In addition, there are a few smaller refinements, generally relating to local areas. For example, the Koolyanobbing greenstone belt contains a low-grade component (*cf.* Griffin, 1981), and the northwest and southwest portions of the Marda greenstone belt are probably better allocated to the medium-grade domain. Furthermore, this study shows that in the southwest portion of this same belt, the metamorphism was gradational, whereas the data of Binns and others (1976) show a high-grade zone juxtaposed to a very low-grade zone.

P-T CONDITIONS OF METAMORPHISM

P-T conditions (Table 4) have been determined by several workers at three localities in the province: Meier's Find-Heaney's Find (Gole and Klein, 1981); Westonia (Blight and Barley, 1981) and Forrestania (Porter and McKay, 1981). The P-T data for Westonia are probably too high, because the calibration used makes no allowance for the effect of water in stabilizing cordierite at higher pressures. Recalculation of the original data, in accordance with Martignole and Sisi's (1981) technique, suggests a temperature of 665 ± 30°C and a pressure of 300 ± 50 MPa. Lonker's (1981) calibration indicates slightly higher pressures, possibly 400 ± 100 MPa.

The three localities represent high-grade domains and the P-T data probably reflect temperatures and pressures which approach the highest attained within the greenstone sequence of the province. On a P-T diagram (Fig. 11) the values plot about the amphibolite-granulite transition facies, and indicate geothermal gradients of 45°C/km, or greater. This compares with a more typical range of about 25-30°C/km for continental crust (Bickle, 1978; Watson, 1978), and thus clearly indicates a low-pressure metamorphic facies-series ("Abukuma" type). The geobarometry data suggest that the greenstone belts have

TABLE 4. P-T CONDITIONS OF METAMORPHISM FOR THREE LOCALITIES IN THE SOUTHERN CROSS PROVINCE

Location	*Rock type*	*Geothermometer*	*Geobarometer*	T(°C)	*P(MPa)*	*Source*
Westonia	Pelite	Cord-garnet (Currie, 1971; Wells, 1979)	Cord-garnet (Currie, 1971; Wells, 1979)	657-760	(a) 460-520	Blight and Barley (1981)
Heaney's Find and Meier's Find	BIF	Opx-Cpx Ol-Opx Absence of pigeonite	Opx-O1-Qtz	670±50	300-500	Gole and Klein (1981)
Forrestania	Pelites	Biotite-garnet (Thompson, 1976)	Constraints imposed by various discontinuous reactions	(b) 655±30	300-500	Porter and McKay (1981)

(a) Maximum value because H_2O in cordierite not considered in calibrations (see text).

(b) Porter (1982) has redetermined this to be 668±30°C.

probably not been buried more than 15-18 km, which implies that the maximum thickness of the original greenstone sequences was equal to, or less than, this value.

Although the three P-T values represent only isolated points in the southern part of the province, the more widespread occurrence of andalusite, cordierite, and sillimanite (plus the coexistence of andalusite-cordierite, andalusite-sillimanite, and cordierite-sillimanite) is consistent with low pressures and relatively high temperatures throughout most of the province (Fig. 4). Further evidence of low-pressure conditions is the scarcity of garnet and the frequent presence of accessory amounts of cummingtonite in the amphibolites (Miyashiro, 1973). In addition, staurolite is rather rare in outcrops although widely present in soils around Southern Cross (Carroll, 1939); but in the few rocks where it has been observed, it coexists with cordierite, or cordierite-andalusite. These parageneses are thought to be low-pressure ones

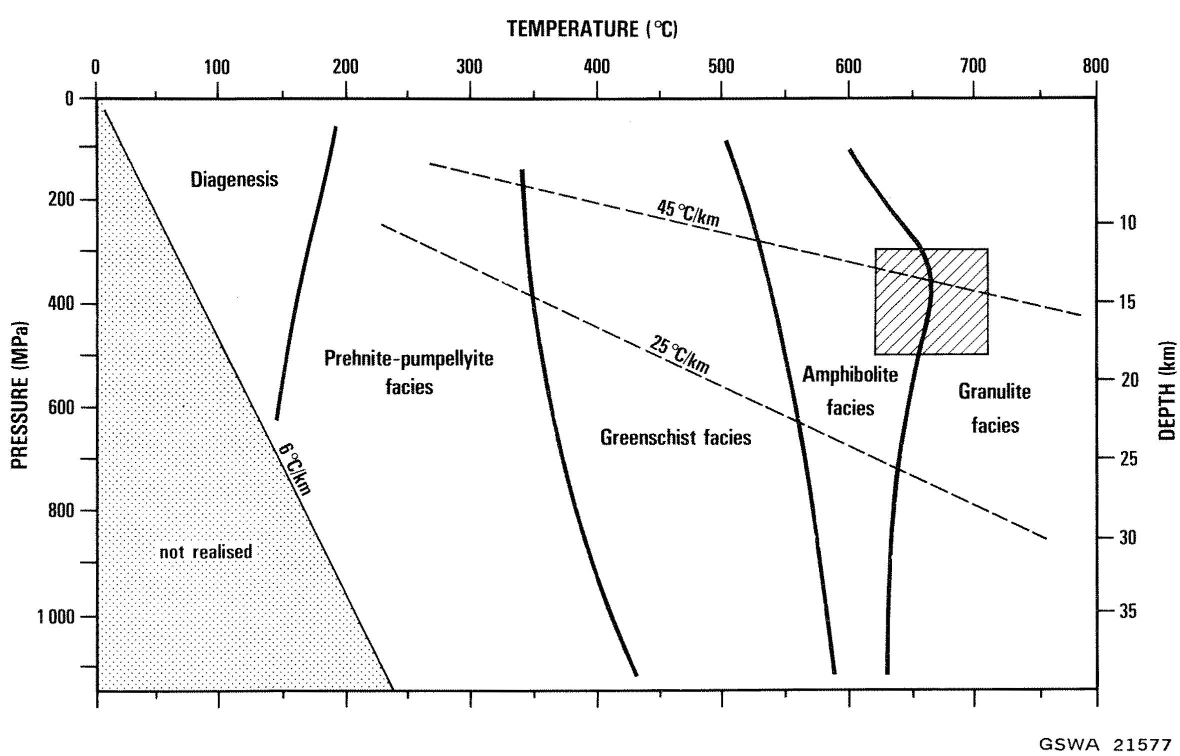


Figure 11. P-T diagram showing field (diagonal lines) of values listed in Table 4. This field straddles the amphibolite-granulite facies boundary and approximates maximum metamorphic conditions experienced in the greenstone belts of the Southern Cross Province. A geothermal gradient of ≃ 45°C/km is indicated. This compares with a continental crust average of ≃25° C/km.

(Richardson, 1968). Kyanite has been recorded in several soil samples from the Southern Cross district (Carroll, 1936; 1939), but it has not been found in any rock examined to date. Even if kyanite does exist in the province, its presence may not necessarily be indicative of high-pressure regional metamorphism. Binns and others (1976) suggested that the few known occurrences of kyanite in the Eastern Goldfields Province (Miles, 1943) are "restricted to probably anomalous structural settings." A similar explanation was proposed by Ayres (1978) for the rare occurrences of kyanite in the granite-greenstone terrain of the Superior Province of Canada, where kyanite occurs principally near the tectonic boundaries of sub-provinces.

Some indications of slightly higher pressures of metamorphism are found in rare garnet-bearing mafic rocks 4 km east-northeast of Skeleton Rock and near Holt Rock. The former locality is about 5 km west of the Southern Cross - Forrestania greenstone belt and the latter about 3 km west, which is well within the "Wheatbelt" granulite terrain. The assemblage of the Skeleton Rock sample is plagioclase-hornblende-orthopyroxene-garnet and belongs to Winkler's (1979) hypersthene-plagioclase granolite subzone of the regional hypersthene zone of high-grade metamorphism. The Holt Rock assemblage is clinopyroxene-plagioclase-garnet-sphene and probably belongs to Winkler's clinopyroxene-almandine-quartz granolite subzone of the regional hypersthene zone. Green and Ringwood (1967) show experimentally that, in rocks of basaltic composition, garnet is unstable below 500 MPa at 700°C. For a typical quartz tholeiite the pressure is even higher (about 700 MPa). Thus the mere presence of garnet in the Skeleton Rock and Holt Rock mafic rocks suggests pressures greater than those determined in the nearby Forrestania greenstone belt.

According to de Waard (1965), Green and Ringwood (1967), and Winkler (1979), the assemblage represented by the Holt Rock sample is indicative of higher pressures in granulite-facies metamorphism than the orthopyroxene-bearing assemblage, and suggests pressures in the 500-700 MPa range. Consequently, on these grounds, there is suggestive evidence of a gradual pressure gradient increasing from roughly 400 MPa in the Southern Cross - Forrestania greenstone belt to 600 MPa in the Holt Rock area.

Turner (1981) and Percival (1983) attach a different significance to the clinopyroxene-plagioclase-garnet assemblage and suggest that it is only an intermediate paragenesis between the amphibolite and granulite facies, and not the high-pressure equivalent of the orthopyroxene-bearing assemblage. This interpretation would tend to negate a gradual pressure gradient between the greenstone belt and the Holt Rock area, and instead suggest that there is an abrupt change, occurring over a distance of a few kilometres. If so, this has important implications, providing *prima facie* evidence of a major crustal discontinuity along the western margin of the Southern Cross - Forrestania greenstone belt.

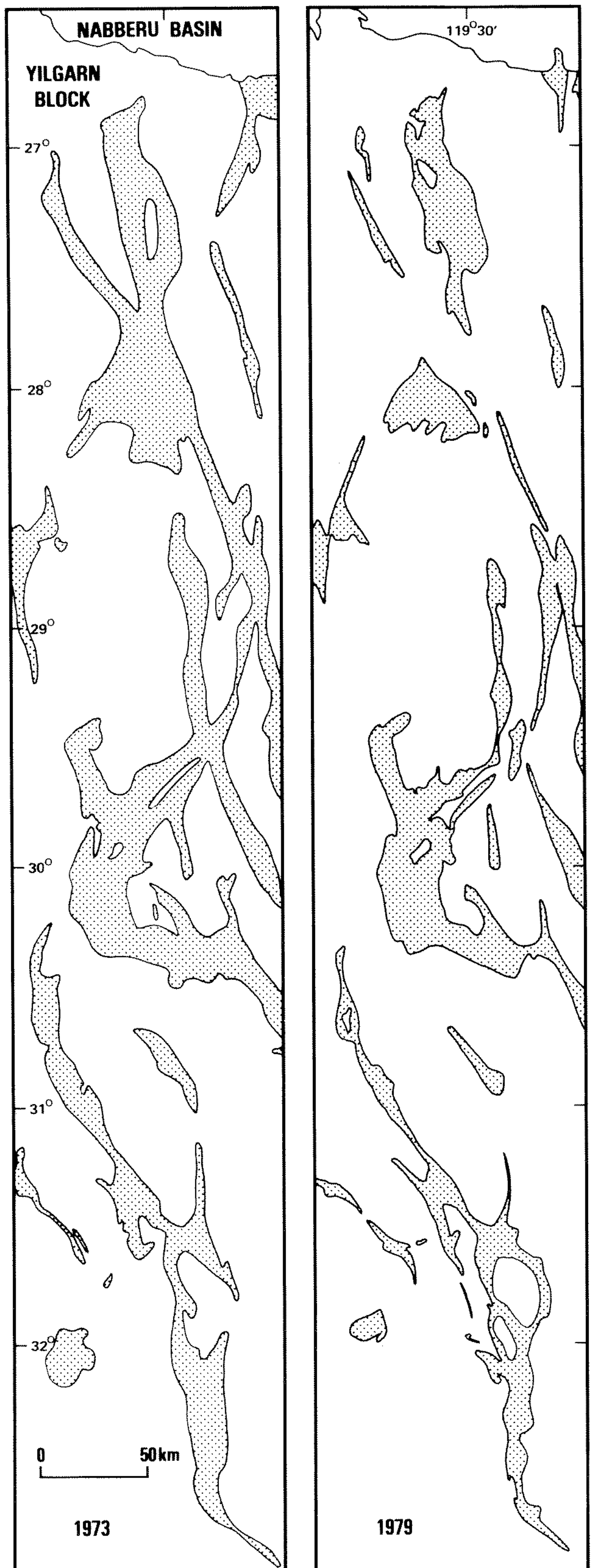
In summary, the general distribution of metamorphic grades and inferred P-T conditions is consistent with the proposed regional picture (Wilson, 1958; Gee and others, 1981).

DISCUSSION

Relationship to granitoids

Four problems complicate the relationship between the granitoids and the greenstone-belt metamorphism. The first is the poor definition of the greenstone-belt margins (Figs 12 and 13). The continuity of some greenstone belts is in question (*e.g.* north of Sandstone), and the boundaries have generally not been accurately located, so that lateral shifts over many kilometres are still possible. The second problem is the dispute over the types of granitoid rocks that are present in the Yilgarn Block: contrast Archibald and others (1981) and Gee and others (1981) with Hallberg and Glikson (1981). The third is the distribution of granitoid types within the province. For example, those who support the kinematic classification (Archibald and others, 1981; Gee and others, 1981) show wide variations in their interpretations of the data (Fig. 14). The fourth problem relates to the place of origin and the place of crystallization of the granitoid plutons. Bettenay (1977) and Archibald and others (1981) conclude that most of the intrusive granitoids had "restricted upward movement", and were thus essentially autochthonous. On the other hand, Gee and others (1981) suggest that the intrusive granitoids ranged from autochthonous to allochthonous.

The first problem (imprecise definition of greenstone-belt margins) clouds the effect and extent of contact metamorphism, especially in those places where the contact is not known to within several kilometres, whereas the three other problems (type, distribution, and locus of origin of bodies) preclude any distinction between the effects that the different granitoid types may have had on the degree and style of metamorphism. It is therefore apparent that more meaningful conclusions will be made when the above anomalies are resolved.



In spite of these limitations, the moderately regular concentric distribution of metamorphic grades within most of the greenstone belts, with higher grades around the margins and lower grades within the centres, appears to be directly related to the granitoids. The question of whether the relationship is either causal or coincidental remains problematical, but it is of paramount importance in understanding the metamorphism and development of metamorphic patterns in the greenstone belts. Any model construed to explain the phenomenon is constrained by the geochronological, structural and petrographic evidence, which suggests that the development and

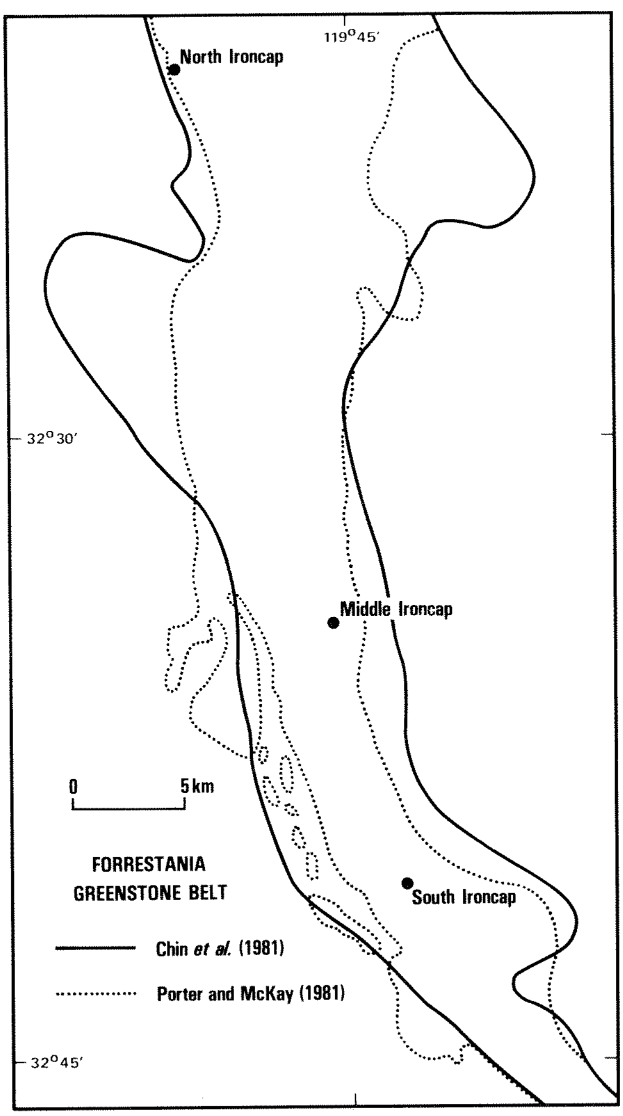


Figure 13. Part of the Forrestania greenstone belt as defined by Chin and others (1981) and Porter and McKay (1981). Such major differences preclude detailed interpretation, especially of the granitoid-greenstone relationships.

Figure 12. Outline of greenstone belts in the Southern Cross Province as determined by the Geological Survey for 1973 and 1979. Differences are gross and complicate structural and metamorphic interpretation. Binns and others (1976) metamorphic map is based on the 1973 version, whereas this study is based on the 1979 version, with updates from more recent 1:250 000 mapping results (see Table 1).

emplacement of the granitoids was synchronous with the deformation and metamorphism of the greenstone belts. Two alternatives that may be envisaged are: the granitoids were responsible for the metamorphism (*i.e.* contact metamorphism); or the granitoids were just part of a larger crustal process and were only involved by determining the final shape of the greenstone belts. The second model assumes that immediately prior to emplacement of the granitoids, the crust had a "layer-cake" stratigraphy and that high-temperature isotherms were already present. Mobilization of the granitoid domes or diapirs merely acted to deform the isotherms and the stratigraphy, resulting in synformal configurations within greenstone belts and antiform configurations within and above the domes or diapirs.

Several lines of evidence suggest that contact metamorphism was not the principal mechanism of producing the concentric pattern of isograds. Wherever contact metamorphism has been recognized it is limited to thin aureoles, generally less than 1 km wide (Binns and others, 1976), associated with post-tectonic granitoids. Such granitoids are comparatively rare (Fig. 14) and the influence is therefore largely local. More voluminous are the gneisses and synkinematic granitoids, and since these two types are essentially autochthonous, the concept of contact metamorphism becomes irrelevant because the adjacent greenstone-belt rocks would have been under similar P-T conditions to the granitoids. Further evidence, which reduces the direct role of granitoid-related metamorphism, is a consideration of the P-T conditions of granitoid formation and crystallization. There are data to show that the granitoids (including the gneisses) originated and crystallized at pressures of 400-500 MPa and temperatures between 650° and 680°C (Bettenay, 1977; also quoted in Archibald and others, 1981). Such temperatures could produce a narrow zone of high-grade rocks, but cannot adequately explain the wide tracts of high-grade rocks that occur in the southern third of the province. Also, the P-T conditions of metamorphism at three localities in this region (Table 4) are virtually identical to those inferred for the granitoids. The picture to emerge, then, is one where the P-T conditions between the granitoids and greenstone belts merge, and the most reasonable explanation is to consider that both structures were under a similar P-T environment. That is, elevated high-temperature isotherms were already present in the crust, and were not produced by the granitoids. This model is also advocated by Archibald and others (1981) for the eastern goldfields, and by K. P. Watkins (pers. comm., 1984) for the Murchison Province.

To obtain the observed arrangement of metamorphic domains, the pre-tectonic structure demands that the greenstone belts overlay the granitoids. Structurally high layers were relatively cool and, subsequent to folding and metamorphism,

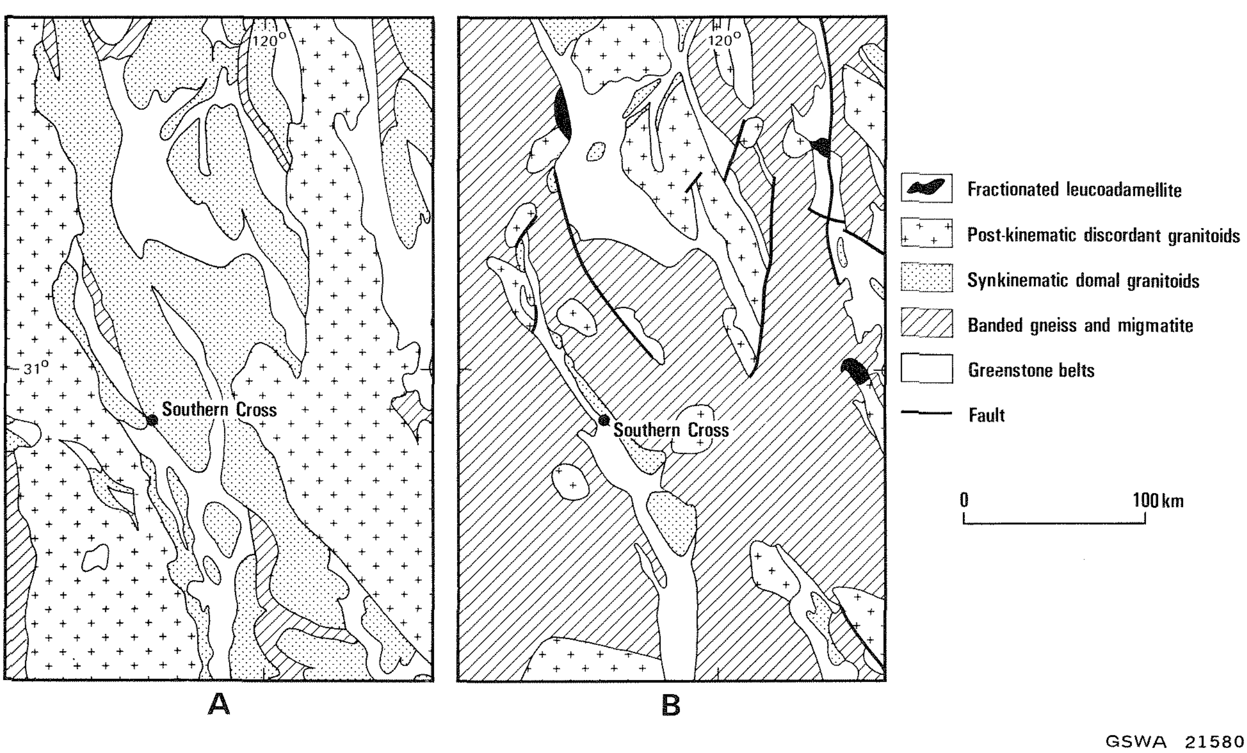


Figure 14. Interpretative geological maps of part of the Southern Cross Province by Gee and others (1981) and Archibald and others (1981).

are now represented as lower grade domains within the cores of the synformal greenstone belts (Fig. 15B). The hottest isotherms were present at greater depth, adjacent to the underlying granitoids, and are now reflected as high-grade domains around the margins of the synformal greenstone belts. Binns and others (1976) originally proposed that the metamorphic domains were related to structural height above a granitic substratum, but because they advocated that the metamorphism was imposed on previously deformed, yet broadly synformal greenstone belts, their configuration of isograds was different (Fig. 15A). Thus, with their model, the intersection of isograds with the present surface reflects subsequent crustal tilting and/or tectonism.

With the model advocated in this paper, the isograds should mimic the original layered sequence and parallel the granitoid-greenstone boundary. In several places, however, isograds are discordant (see above). The cause of this feature may have involved local rupture of the original granitoid-greenstone boundary by the intrusion or stoping mechanism of the diapiric granitoids, which then introduced perturbations in the original distribution of isotherms.

In summary, the relationship between the granitoids and the metamorphism of the greenstone belts appears to have been largely coincidental. Metamorphic grades were controlled predominantly by the distribution of isotherms in a sub-horizontally layered crust. The granitoids have only been important in deforming both the layered structure and the isotherms during the latter stages of a major tectonothermal event.

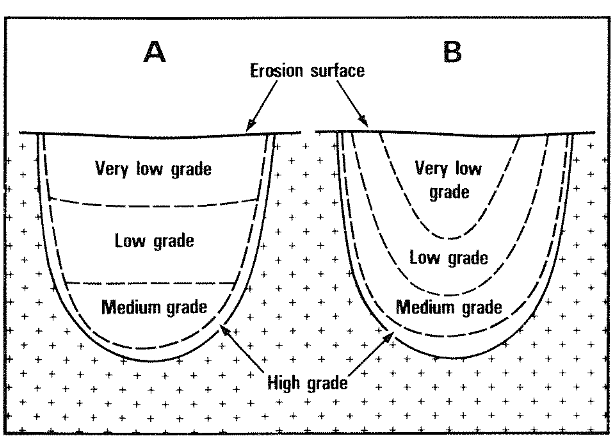


Figure 15. Diagrammatic cross-sections of greenstone belts showing distribution of metamorphic domains: A—as determined by Binns and others (1981), and B—as determined by this study. In B, domains are thicker at depth than at the sides because, following synformal folding of the greenstone belt, the isotherms would have risen slightly when a new thermal equilibrium had been established in the crust.

Regional northeast trend

Metamorphic grades have been shown above to be higher in the southern part of the Southern Cross Province than in the north. This difference is most noticeable across a northwest-trending line separating the Marda and Southern Cross greenstone belts, and is consistent with the scheme proposed by Gee and others (1981); see also Figure 1B. Pressure determinations (Table 4) and the wide distribution of low-pressure aluminosilicate phases (Fig. 4) in this southern portion constrain the metamorphism to relatively low pressures, and imply that the zone does not represent a deeper level of the crust. The higher grades and low pressures, therefore, indicate a higher geothermal gradient, and an explanation of the phenomenon must account for this. There is a possible similarity in South Africa. Saggerson and Turner (1976) explained a regional, low-pressure metamorphic zonation, symmetrically disposed across the Zimbabwe and Kaapvaal cratons, as having been produced by a large thermal anticline sited beneath the axis of the Limpopo Mobile Belt. While it is attractive to invoke a similar model for the regional zonation in the Southern Cross Province (and most of the Yilgarn Block), it faces one major obstacle: a symmetrical zonation pattern has not been observed.

In the extreme southwestern zone of the Southern Cross Province, the juxtaposition of a granulite terrain to the Forrestania greenstone belt (which is asymmetrically zoned from high grade in the west to low grade in the east) provides evidence for alternative mechanisms that could have produced, or contributed to, the regional pattern. Either the structure and distribution of metamorphic grades represent a single, extensive metamorphic event which decreased in intensity to the east, or the granulite terrain represents a tectonic-metamorphic episode in which an uplifted block of lower crustal material brought high-temperature isotherms closer to the surface, and produced a high, sub-horizontal thermal gradient across the margin of the greenstone belt (Archibald and others, 1981).

The first alternative would require an enormous, high-level heat source (lateral to the greenstone belt and centred in the granulite terrain), and a low-pressure style of granulite metamorphism. Whilst a low-pressure type of high-grade metamorphism has been documented, the likelihood of a high-level heat source having been present in the adjacent granulite terrain is low. For example, there is no evidence for the presence of large mafic intrusions, one of the few structures which could conceivably produce high-level granulite metamorphism. Other limitations of this model are: it does not explain the abrupt changes of metamorphic domains once into the greenstone belt; geochronological evidence further to the west suggests that granulite metamorphism may have occurred at about 3.2 Ga (Nieuwland and Compston, 1981); the event should have caused vast amounts of partial melting in the lower crust, for which no evidence exists; and it does not explain the high-strain domain along the western margin of the Forrestania greenstone belt.

The second alternative, that of differential uplift, provides a better explanation of the rather abrupt changes in metamorphic grade within the Forrestania greenstone belt and the high-strain zone; moreover it does not require a high-level heat source with associated melting at depth. Because of the inferred high geothermal gradients (*i.e.* 40°C/km), the necessary increase in temperature to produce the observed high-grade metamorphism would only require a relative uplift of 2-3 km, which is small compared with the overall crustal thickness of about 35 km (*cf.* Mathur and others, 1977). Archibald and others (1981) applied this differential-uplift model to the evolution of the eastern half of the Yilgarn Block, but were constrained by the lack of high-pressure assemblages which might be expected to occur, although a possible change of only 100 MPa, or so, may not have produced observable petrographic changes. However, the existence of several garnet-bearing mafic granulites—possibly representing slightly higher pressure assemblages—to the west of the Forrestania greenstone belt may represent an example of the model proposed by them. Whether the inferred differential uplift accounts for all the regional-scale, high-grade metamorphism in the southern portion of the province is less clear. The asymmetry of metamorphic domains in the Forrestania greenstone belt and the distribution of mafic granulites suggest that it is restricted only to the southwest portion of the province.

Comparisons with other Archaean terrains

Broad similarities in metamorphic patterns, grades and styles exist between the Southern Cross Province and other granitoid-greenstone terrains, such as the Canadian Shield (Ayres, 1978; Ermanovic and Froese, 1978; Weber and Scoates, 1978), the Zimbabwe craton (Saggerson and Turner, 1976) and the Kaapvaal craton (Viljoen and Viljoen, 1969; Saggerson and Turner, 1976). These similarities include the following: comparable "local" zonations within greenstone belts, with grades increasing towards the margins; "regional" zonations occurring across broad tracts of granitoid-greenstone terrain; metamorphic grades ranging from sub-greenschist to granulite facies; and low- to intermediate- pressure-series metamorphism, typically associated with high geothermal gradients. There is also a similar age of metamorphism, particularly 3.0-2.6 Ga ago, which is a time predicted by Lambert (1980) for maximum heat flow in the Earth's history.

Most of the above workers attribute "local" zonations in greenstone belts to thermal metamorphism by the surrounding intrusive granitoids, and "regional" zonations to the former presence of regional-scale thermal anticlines (Richardson, 1970) in the crust. The lower grade cores of some greenstone belts are thought to be products of burial metamorphism (Ayres, 1978; Jolly, 1978), having formed during crustal downwarping of the developing supracrustal sequence. Whilst elements of all these processes are envisaged in the metamorphism of the Southern Cross Province, other mechanisms are also considered to be important, including the warping of pre-existing, high-temperature isotherms in the supracrustal sequences, and possibly the uplift of hotter, lower crustal layers. Thermal metamorphism is not considered to be so effective, because the temperature estimated by Bettenay (1977) for the generation of granitoids (670°C) is relatively low to invoke widespread high-grade metamorphism. In contrast, Ayres (1978) estimates that the temperature of the granitoids in the Superior Province was 1 000°C upon emplacement, a temperature more likely to cause thermal metamorphism over considerable distances.

CONCLUSIONS

In addition to corroborating many of the ideas proposed by Binns and others (1976), this study of the metamorphic patterns in the greenstone belts of the Southern Cross Province shows several new significant features. These are listed below.

(1) A greater correspondence between the metamorphic patterns and the granitoid plutons. Metamorphic patterns are mostly concentric and are gradational from high grade, high strain around the margins of greenstone belts to low grade, low strain in the centres.

(2) A regional pattern with higher grades more prevalent in the southern portion of the province; a feature first recognized by Gee and others (1981).

(3) The high-grade western side of the Forrestania greenstone belt is juxtaposed to a granulite-facies terrain, providing some evidence of a "differential uplift" of a lower crustal layer; a model proposed by Archibald and others (1981).

These three points therefore suggest that the metamorphic patterns are the products of a combination of at least two processes: namely, the emplacement of the diapiric granitoid plutons; and the differential uplift of a slightly hotter, lower crustal layer. The textural and geochronological evidence together suggest that these processes were essentially penecontemporaneous, presumably related to the 2.8-2.6 Ga tectonothermal event, which led to the final cratonization of most of the Yilgarn Block (Gee, 1979).

Metamorphic grades, on the other hand, appear to be related to the pre-tectonic distribution of high-temperature isotherms in a layered crust. The granitoids are a consequence of the high temperatures—not a cause—and have been crucial only in deforming the isotherms through doming/diapiric mechanisms. Metamorphism by the granitoids is minimal, being restricted to narrow zones of contact metamorphism around both high-level (allochthonous) and post-kinematic plutons.

The present study has also highlighted several deficiencies in our geological knowledge of the area. Little is known about the granitoid-greenstone contact relationships and even less is known about the types and distribution of the granitoids. A better understanding of the metamorphism in the greenstone belts will depend strongly upon resolving these problems.

Furthermore, closer sampling and detailed mineralogical studies of the greenstone belts will significantly improve the contouring of the metamorphic map, and enable definitive isograds to be delineated. More analytical data are required to document the P-T conditions of metamorphism throughout the province.

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PROBLEMATIC MICROSTRUCTURES IN THE PROTEROZOIC DISCOVERY CHERT, BANGEMALL GROUP, WESTERN AUSTRALIA. AMBIENT GRAINS OR MICROFOSSILS?

by Kathleen Grey

ABSTRACT

Tubular microstructures with terminal bodies occur in the Discovery Chert in the Proterozoic of the Bangemall Basin, Western Australia. Attempts to demonstrate that the structures are either ambient grains or microfossils produced ambiguous results. A biogenic origin cannot be dismissed entirely, although an inorganic origin as trails of ambient grains seems more probable.

INTRODUCTION

Tubular microstructures with terminal bodies occur in large numbers in a sample from the Discovery Chert of the Bangemall Group, Western Australia. The microstructures were originally regarded as microfossils (Grey, 1977), but more recently M. R. Walter (pers. comm. 1978) pointed out their similarity to the trails formed by ambient pyrite grains described by Tyler and Barghoorn (1963) and Knoll and Barghoorn (1974). Although the microstructures resemble ambient-grain trails, they exhibit several puzzling features.

To investigate their origin, the microstructures were examined against a set of criteria used for demonstrating that a given assemblage of objects is of biogenic origin (Muir, 1978a). Petrological thin sections and slightly thicker sections—cut both parallel to, and normal to, the bedding—were examined. Size ranges were measured and the results analyzed using a series of statistical tests proposed by Schopf (1976). A portion of the sample was macerated with hydrofluoric acid but no biogenic macerate was recovered from the residue. Material is stored in the GSWA fossil collection.

GEOLOGICAL SETTING AND AGE

Sample F11143 is from an outcrop of the Discovery Chert on the Collier 1:250 000 area, near Coobarra Creek at Long. 118° 43′ 30″S, Lat. 24° 34′ 30″E (Fig.1). The sample was collected by P. C. Muhling during regional mapping of the Bangemall Basin. The Discovery Chert is an important marker horizon in the western facies of the Bangemall Group and is a distinctive, persistent, laminated black chert (Muhling and Brakel, 1985). Most of the chert is too recrystallized for microfossils to be preserved, but the dark colour can be attributed to the presence of abundant organic carbon. Sample F11143 is not typical of Discovery Chert lithology, but is from a fine-grained, banded, ferruginous siltstone with chert lenses. Similar lenses occur in the more typical black chert and in both cases the lenses are syndepositional. Based on the Rb/Sr age determinations, the Bangemall Group is approximately 1.1 Ga old (Gee, 1980).

DESCRIPTION

In thin section, sinuous tubes occur in one of the chert lenses either singly (Fig. 2C-G), or in large clusters (Fig. 2B). The tubes have random orientation, and are usually intertwined without cross-cutting each other. Most are at least 30 μm long, are not flattened, and are usually hollow (Fig. 2G). The majority are sinuous; although they may be arcuate, planispiral, or helical; one specimen has an almost perfect helical shape (Fig. 2D).

Tube diameter is more or less constant in any one tube, although borders show some irregularity. Tube diameter—measured approximately 1 μm from the end of the tube—ranges from 2.8 to 12.5 μm with a mean of 6.5 μm for 100 specimens. The frequency distribution of the tube diameters is shown in Figure 3. Some tubes are discontinuous and terminate abruptly against crystal boundaries in the matrix. They appear to be broken and displaced. This indicates that the tubes pre-dated recrystallization of the matrix.

Borders of the tubes are approximately 1 μm thick and are composed of a mosaic of small, angular, greenish-grey grains (Figs 2C and E). In the absence of microprobe examination it is difficult to determine the composition of the grains because of their small size. They are probably of carbonate composition, and are equigranular and spherical rather than elongated. In some tubes the borders are coated with iron oxide, especially in the region of flexures. The cavities of some tubes contain irregularly shaped grains of iron oxide (Fig. 2F).

Tubes usually terminate in a dark, spheroidal body—the terminal spheroid. Whole tubes are rare (most specimens pass outside the plane of the thin section), but there are some apparently complete tubes in which the spheroid is absent. Commonly the spheroid is present at only one end, but some tubes apparently have spheroids at both ends, although this is difficult to confirm where tubes are densely clustered. Few of the terminal spheroids have a distinct form; where shape can be determined, it is spherical or trefoil with rounded margins. Faceted surfaces were not observed. The spheroids are usually coated with, or composed of, iron oxide; they may be surrounded by patches of iron staining which results in a diffuse appearance, and obscures the grain boundaries (Fig. 2E). The tube immediately adjacent to the terminal spheroid may also be coated or stained with iron.

The diameter of the spheroid is larger than that of the accompanying tube. Spheroid diameters range from 4.6 to 20.3 μm with a mean of 9.0 μm for 100 spheroids measured. The frequency distribution of diameters is plotted in Figure 3. The diameters for paired tubes and terminal spheroids were measured and plotted as a scatter diagram (Fig. 4). Isolated spheroids of apparently similar composition to the terminal spheroids, but lacking accompanying tubular structures, are also present in the sample. They are diffuse, and rounded or slightly elongate; diameters range from 3.6 to 19.6 μm with a mean of 8.3 μm for 100 specimens. The relative frequency of diameters is plotted in Figure 3.

The terminal spheroids and the isolated spheroids were measured to test the possibility that they are from the same population. A chi-square test shows that the frequency distribution of the terminal spheroids is not normal. No explanation can be offered for the bimodality; and, because the frequency distribution is not normal, comparisons between the two samples were not made.

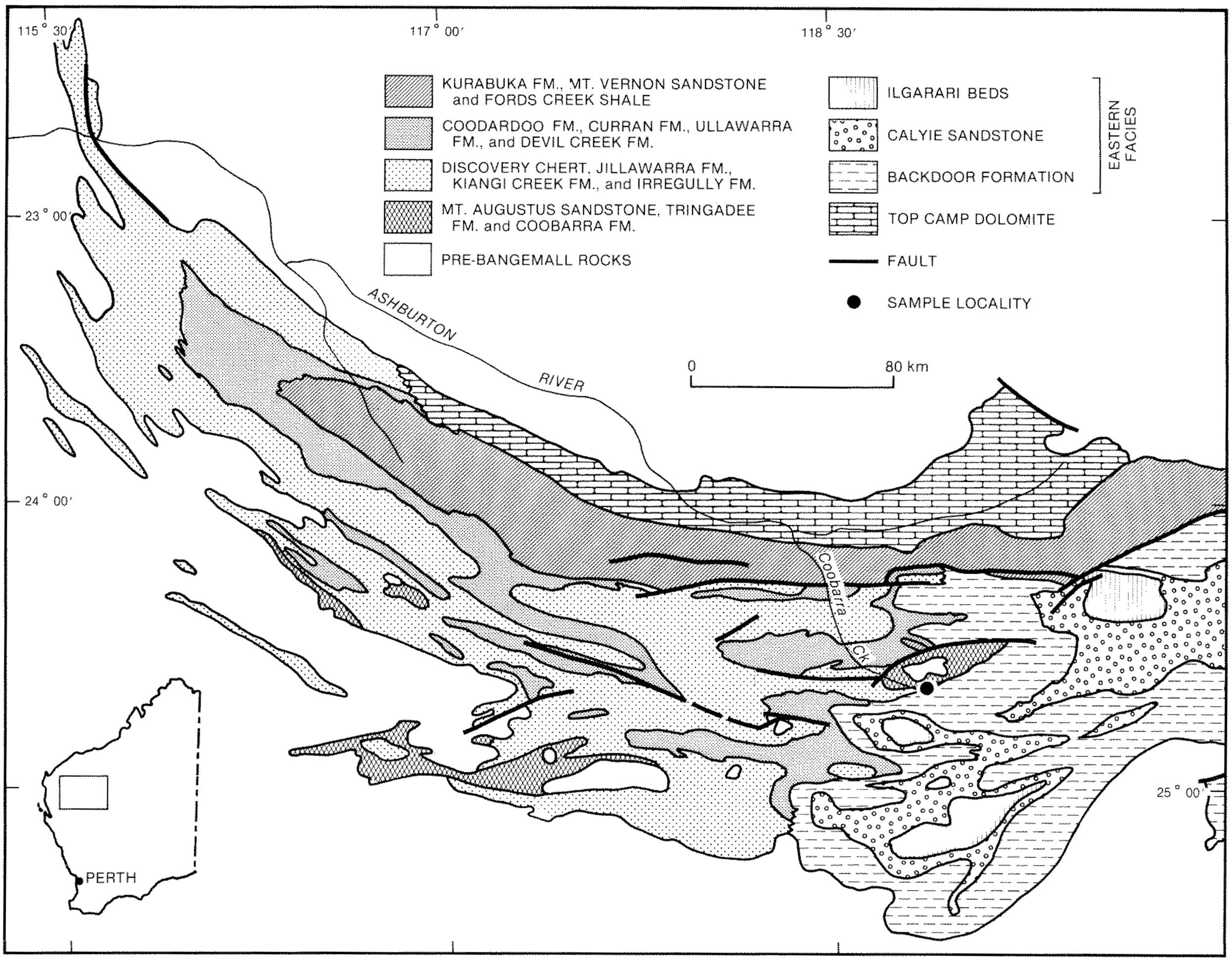
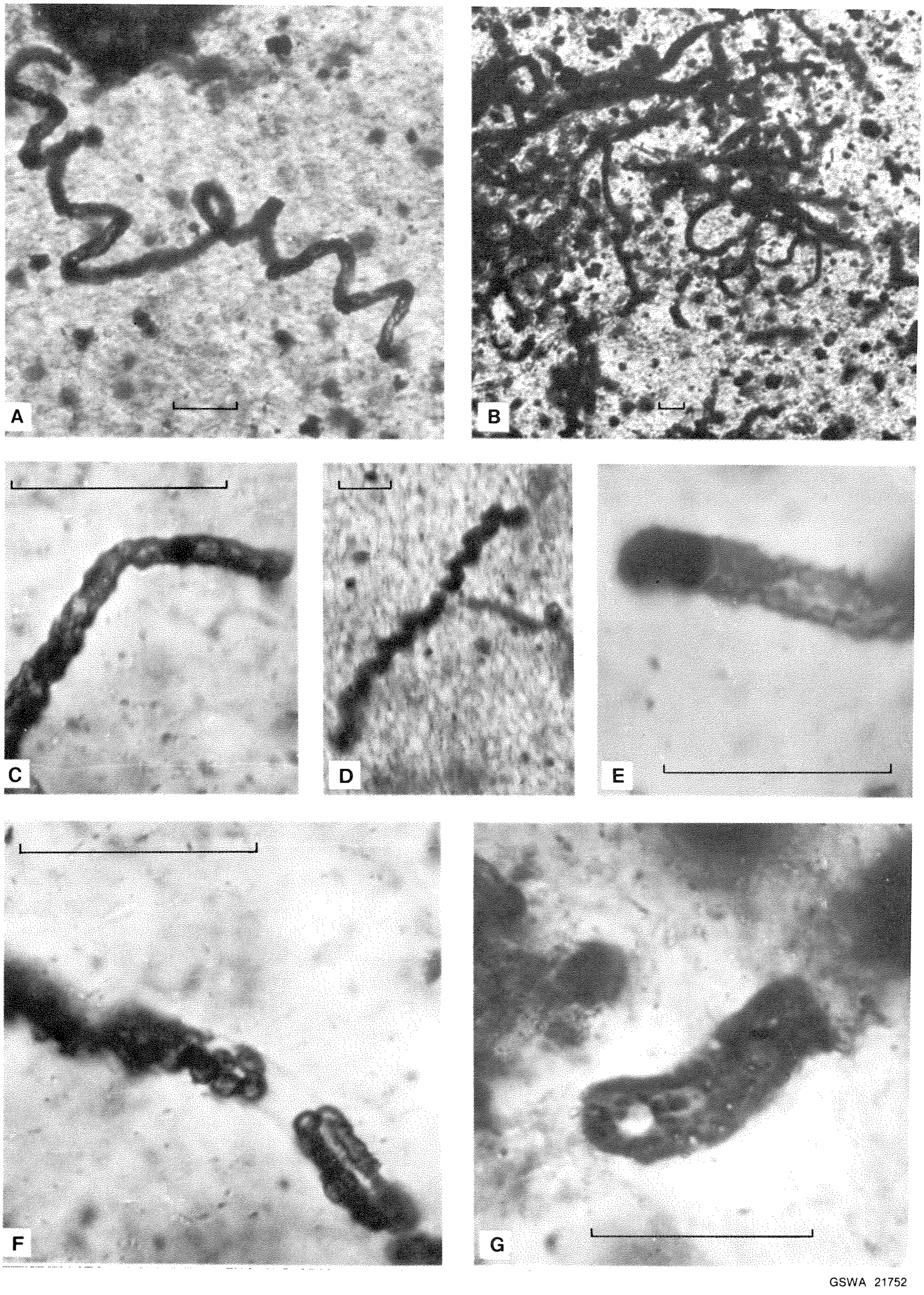


Figure 1. Western Bangemall Basin, Western Australia, (after Brakel and Muhling, 1976) showing location of sample GSWA F11143.



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ORIGIN OF THE STRUCTURES

Tyler and Barghoorn (1963) and Knoll and Barghoorn (1974) described microstructures, which they called "ambient grains", from the Gunflint and Biwabik Formations of the Lake Superior region. Similar structures have also been reported from the Warrawoona Group, northwestern Australia (Awramik and others, 1983). The inorganic nature of all these structures can be readily demonstrated. Their crystalline structure, striated tubular appearance, and the presence of faceted pyrite grains at the termination of the tubes, precludes a biogenic origin. In the Discovery Chert specimens these features cannot be observed, and a non-biogenic origin is less obvious. Although the tubular forms show some morphological similarities to trails of ambient grains, they also satisfy some of the criteria for the recognition of microfossils. Size distributions of both isolated and terminal spheroids coincide with probability plots of size ranges from known spheroidal microfossils and modern unicellular, spheroidal algae and bacteria. For this reason, the evidence for both an ambient origin and a possible biogenic origin is reviewed.

TABLE 1. COMPARISON OF DISCOVERY CHERT MICROSTRUCTURES WITH AMBIENT GRAINS PREVIOUSLY DESCRIBED BY BARGHOORN AND TYLER (1963) AND KNOLL AND BARGHOORN (1974)

Ambient grains	*Discovery Chert*
Microstructures consist of a mineral grain and appendage.	Microstructures consist of a spheroid and tube.
Appendages solid, formed by large, inequigranular, elongate crystals.	Tubes originally hollow, now filled with fine-grained chert. Occasionally contain iron-rich grains.
Appendages without distinct margin.	Tubes with margin of fine grained (approx. 1 μm) mosaic of equigranular crystals, possibly carbonate, usually iron-coated.
Appendages increase in smoothness nearer the grain.	Margins irregular throughout.
Some appendages cross-cut each other.	Some tubes intertwine, but do not cross-cut each other.
Clustered appendages form starbursts.	Clustered tubes do not form starbursts, but are interwoven in random orientation.
Length of appendages from less than 1 μm to more than 10 μm.	Tube length 30 μm or longer. No short tubes.
Appendage diameter decreases towards the grain.	Tube diameter constant.
Grains present at only one end of the appendage.	Some tubes apparently have spheroids at both ends.
All appendages have grains.	Some tubes have no spheroid.
Appendage diameter=grain diameter.	Tube diameter less than grain diameter (mean tube diameter=6.5 μm at 1 μm from the spheroid, mean spheroid diameter=9.0 μm, 100 specimens measured).

Figure 2. Tubular microstructures from Discovery Chert sample GSWA F11143. Bar scale represents 10 μm. A—Isolated sinuous tube. B—Cluster of tubes. C—Grains forming tube border. D—Helical tube. E—Iron-oxide spheroid at the tube termination. Note the granular border of the tube. F—Iron-oxide grains in the tube. G—Hollow tube.

Although the Discovery Chert forms have a basic morphology similar to previously described trails of ambient grains, several important differences can be observed, and these are summarized in Table 1.

From the differences described in Table 1, it is clear that several problems arise in interpreting the Discovery Chert microstructures as ambient-grain trails. The proposed mechanism of formation (Tyler and Barghoorn, 1963; and Knoll and Barghoorn, 1974) cannot account for the apparent presence of spheroids at both ends of some structures, nor for the lack of spheroids in others. Knoll and Barghoorn (1963) suggested that trails formed when grains were propelled through a gel by gases emitted by decomposing organic matter surrounding the grain. The grain, therefore, decreased slightly in diameter as it moved, and the resulting appendage tapered. Both Tyler and Barghoorn

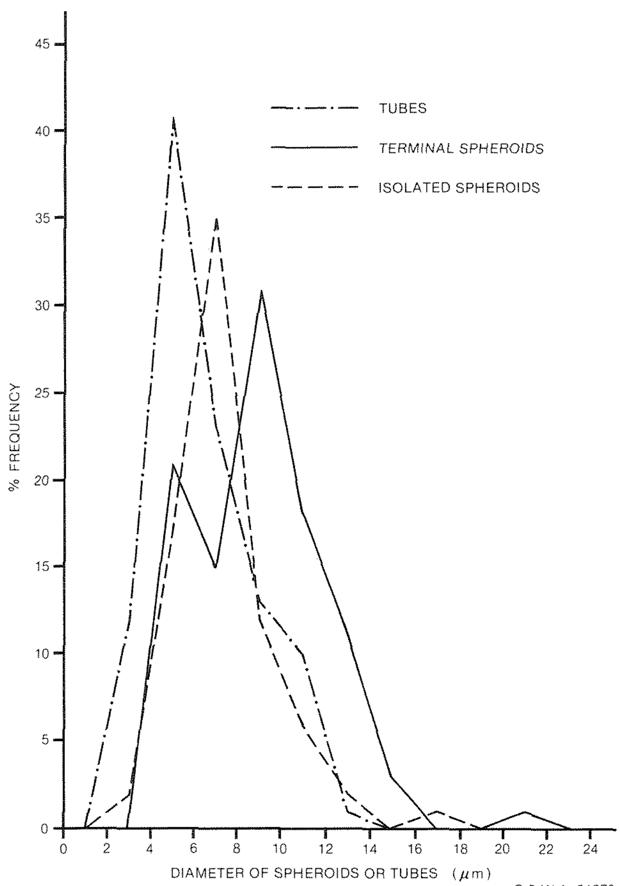

Figure 3. Frequency diagrams for populations of tubes, terminal spheroids and isolated spheroids from the Discovery Chert.

(1963) and Knoll and Barghoorn (1974) stress the fact that the appendage is either slightly larger than, or the same size as, the grain: in the Discovery Chert, the tubes do not taper and the tube diameter is smaller than the spheroid diameter. These authors also attributed variations in appendage length to variable quantities of organic matter which propelled the grains for variable distances. Such variations in length do not occur in the Discovery Chert specimens.

The relationships of the Lake Superior microstructures to each other, and to the matrix, suggest that appendages formed at different times, with some movements occurring as radiating bursts through a gel, and others postdating crystallization. In some examples, crystals apparently grew as the grain moved, giving rise to the large, elongate grains which now occupy the former trails. In the Discovery Chert, the tubes are infilled by the same fine-grained chert which forms the matrix.

The composition of the terminal spheroid cannot be determined with certainty. Iron oxide is present but may only be a coating surrounding some other mineral or organic matter. Iron oxide also coats parts of the tube borders and extends backwards along the tube border from the terminal spheroid. If this were residual material from a moving grain, there would be a decrease in tube width resulting from the passage of a smaller grain.

The above differences, although not disproving an inorganic origin for the structures, raise sufficient doubts for consideration of an alternative origin, possibly as microfossils.

The problems of demonstrating a biogenic origin for assemblages of microstructures are well known (Cloud and Licari, 1968; Cloud, 1976;

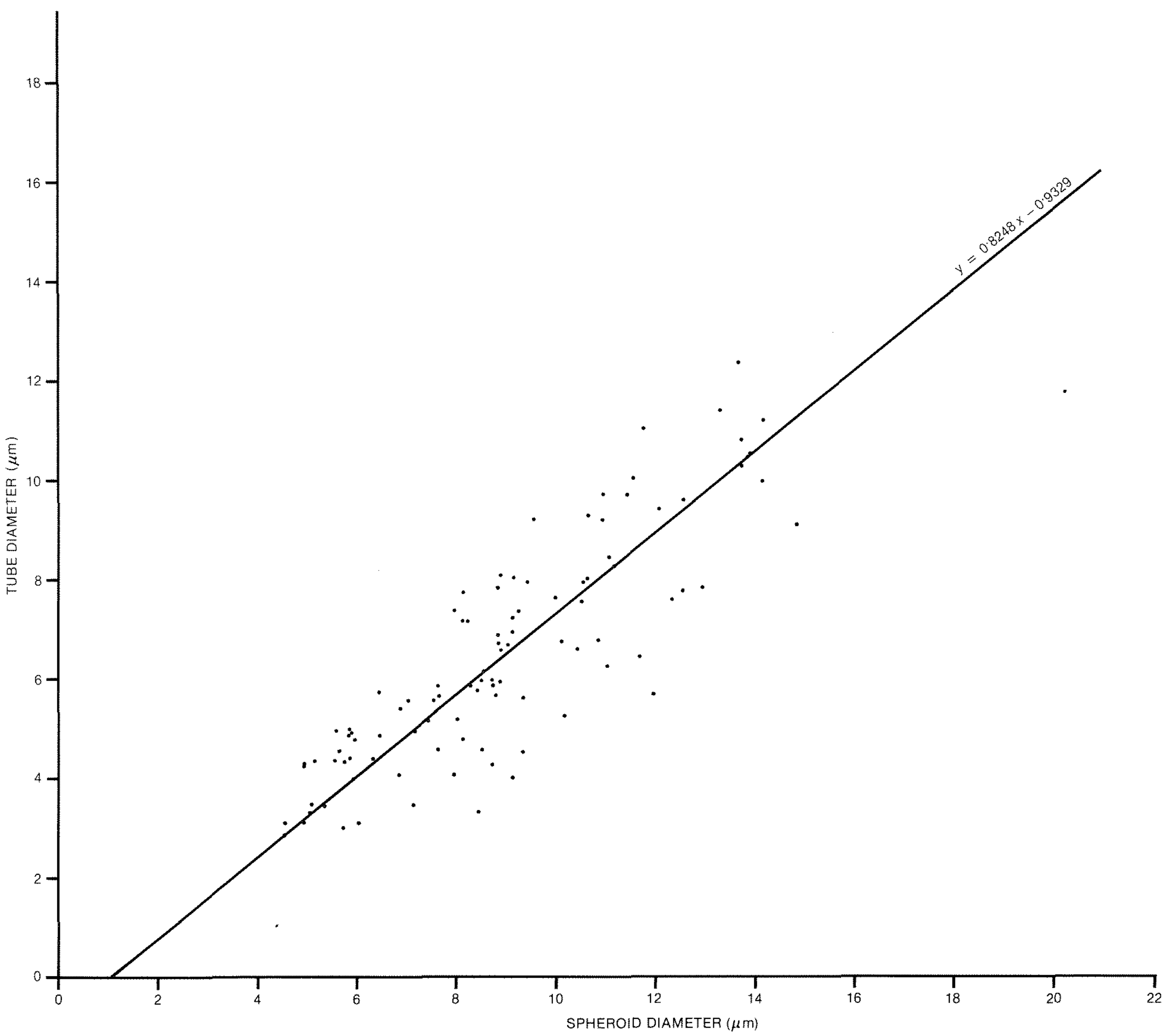

Figure 4. Scatter diagram and reduced major axis of tube diameter plotted against terminal-spheroid diameter for specimens from the Discovery Chert.

Cloud and Morrison, 1979; Schopf, 1976; Muir, 1978a). Muir (1978a) lists seven criteria for determining whether objects are biological in origin. The Discovery Chert microstructures satisfy five of them:

(a) they occur in rocks of demonstrably sedimentary origin;

(b) they are syndepositional with the sediment;

(c) their distribution and relationship to surrounding minerals in thin sections rules out the possibility of contamination;

(d) there has been very little mineralogical distortion; and

(e) all material of possible organic origin has undergone a similar diagenetic history.

Two of Muir's criteria require discussion. These are that:

(f) the objects must be demonstrably of organic matter; and

(g) they should be abundant and of a distinctive morphology which shows a reasonable resemblance to living organisms, such as those found in modern mat-building microbial communities. Their morphologies should also be comparable with other Precambrian microfossils.

The criterion of organic composition is probably the most critical factor. The nature of the tube borders cannot be readily determined because of the coating of carbonate or iron oxide. Several instances of the mineralization of organic matter, probably as a result of pre-lithification processes, have been reported (for example, Barghoorn and Tyler, 1965; Walter and others, 1976; Oehler, 1977; Muir, 1978b). Preservation is frequently in the form of replacement by pyrite, but replacement by iron oxides, iron hydroxides or manganese dioxide has also been recorded (Muir, 1978b).

Pacltova (1976) illustrates an assemblage of spheres and filaments apparently covered by colonies of filamentous prokaryotic organisms which produce a specific crystal-like mineral aggregation around each filament. Individual crystals are perforated by openings through which thecae protrude. At present it is not possible to demonstrate the presence of organic matter, and the origin of Pacltova's specimens remains doubtful. There are some similarities to both the spheroids and the tubes in the Discovery Chert. However, if Pacltova's evidence of mineralized microfossils is discounted, the crystalline structure of the Discovery Chert tube borders, though less distinct than that previously illustrated (Tyler and Barghoorn, 1963; Knoll and Barghoorn, 1974; Awramik and others 1983), suggests an abiotic origin. Because no organic material has been recognized, a biogenic origin cannot be proved.

Muir's final criterion is that the structures should be abundant, and of a distinctive morphology showing a reasonable resemblance to fossil and living organisms. There are two facets of comparative morphology which must be considered. One is the comparison of the morphological appearance of the supposed fossils with that of undoubted fossil filaments, and with modern algae and bacteria, while the other is the use of a series of statistical parameters (Schopf, 1976).

It is difficult to compare the Discovery Chert material directly with extant species of bacteria or algae, few of which have terminal structures or consist of long, spiral filaments. The high iron-oxide concentration suggests that analogous forms might occur among the iron-oxidizing bacteria. Harder (1919) illustrates several of these, including *Gallionella, Leptothrix* and *Spirophyllum. Gallionella* forms colonies of small, bean-shaped bacterial cells at the branched tips of excreted stalks, which are heavily impregnated with iron (Stanier and others, 1977). *Leptothrix* is thread-like, may be straight or curved, and its sheath is usually heavily coated with ferric hydroxide (Harder, 1919). Thickened ends have been described both for *Leptothrix*, and for a related species, *Chlamydothrix sideropu*, which has a well-developed holdfast (Harder, 1919). *Spirophyllum* resembles the helical specimen (Fig. 2D) in general morphology.

One of the principal arguments against a bacterial origin is the large size of the Discovery Chert tubes. Most bacteria are less than 2 μm in diameter, whereas the Discovery Chert microstructures have a mean diameter of 6.5 μm. Filamentous, cyanophytic bacteria and eukaryotic algae may have larger diameters, but rarely show mineralization and also differ in morphology. Francis and others (1978) have recently demonstrated a two-fold increase in diameter of *Spirochaeta stenostrepta* during silicification, and suggest that silica may be absorbed by the sheath in some bacteria during fossilization. If the Discovery Chert microstructures are of biogenic origin, this mechanism could explain their apparent large size. It is possible that the Discovery Chert structures could represent larger, but now extinct, varieties of bacteria, but there is little evidence of such organisms in the fossil record. A few large filaments and complexly twisted filaments have been previously recorded as fossils (Peat and others, 1978, Fig. 6; Cloud, 1976, pl. 5, Fig. 1; and Hofmann, 1976, Fig. 7E), but, in general, it does not seem possible to compare the Discovery Chert microstructures directly with the morphologies of either living or fossil microbial communities. This contrasts with many other convincing examples of Precambrian microfossils, including the putative microfilaments from the Warrawoona Group, which Awramik and others (1983) demonstrate as satisfying the morphological criteria for biogenic origin.

The use of statistical parameters (Schopf, 1976) to test populations for evidence of biogenic origin is inconclusive in the case of the Discovery Chert microstructures. Although Schopf confined his studies to spheroidal objects, there is no reason why his methods should not be extended to filamentous structures, because his objective was to discover characteristics unique to biological material. Tube-diameter data for the Discovery Chert are therefore included with data for terminal spheroids and isolated grains (Fig. 3 and Table 2).

Schopf (1976, p. 34) concludes that, "a simple biological population, composed of the vegetative cells of a single species can be expected to be unimodal, to closely approximate a Gaussian distribution, to exhibit a low to moderate standard deviation (a coefficient of variation in the order of 10 to 25%), and to exhibit a relatively low Divisional Dispersion Index (a DDI generally well less than 7 or 8 and commonly about 4)."

The Discovery Chert microstructures are unimodal for isolated spheroids and tubes, but bimodal for terminal spheroids, a feature for which no explanation can be offered. When probability plots for size data are constructed (Fig. 5), both groups show a similar distribution to that for biogenic material plotted by Schopf (1976). They have standard deviations and coefficients of correlation slightly higher than those quoted by Schopf for biogenic samples, but considerably outside the range of the populations considered to be abiogenic, with the exception of those from Swaziland.

Although the Divisional Disperson Index is of questionable value (see Appendix) it was calculated for the Discovery Chert material. The DDI's of 7 and 8 are relatively high for biogenic material, but are lower than those for abiogenic material.

Schopf recognizes that populations of unicell-like objects of known abiotic origin could exhibit similar patterns of distribution to those of biological origin. For the moment the Discovery Chert structures cannot be assigned either a biological or non-biological origin on the basis of Schopf's statistical parameters.

CONCLUSIONS

Comparisons between tubular structures present in the Discovery Chert and ambient grains reported by Tyler and Barghoorn (1963), Knoll and Barghoorn (1974), and Awramik and others (1983) indicate that, although there are some basic morphological similarities, there are also some unexplained differences, such as the relative sizes of the tube and the terminal spheroid, and the distribution of the tubes in interwoven clusters. They therefore differ from previously reported examples in several respects and their mode of origin is by no means certain.

TABLE 2. SUMMARY OF STATISTICAL PARAMETERS FOR MICROSTRUCTURES

		Diameters of objects					
	Number of specimens measured	*Range (μm)*	*Modal size groupings (μm)*	*Mean (μm)*	*Standard deviation*	*Coefficient of variation (%)*	*Divisional Dispersion Index (DDI)*
		A. Discovery Chert					
Tubes	100	2.8-12.5	2-4	6.5	2.4	36	7
Tubes-spheroids	100	4.6-20.3	bimodal 4-6 8-10	9.0	2.4	32	7
Isolated spheroids	100	3.6-19.6	6-8	8.3	2.7	33	8
		B. Selected data from Schopf (1976)					
Chlorella sp.	426	1.5-5.0	2-3	2.7	0.7	24	5
Gloeocapsa sp.	338	2.5-5.5	3-5	4.2	0.5	12	4
Tetraspora sp.	292	6-12	8-10	9.1	1.4	15	3
Bitter Springs Fm	519	5.5-15.5	8-10	9.3	1.9	21	5
Gunflint Fm	805	1.1-15.8	2-5	5.3	2.5	46	12
Orgueil Meteorite	268	2-36	7-11	10.9	5.1	46	13
Swaziland 1	54	10-39	16-20	20.3	6.8	33	6
Swaziland 2	456	7-193	polymodal?	36.0	31.6	88	15
Swaziland 3	192	6-74	polymodal?	27.2	14.0	51	11

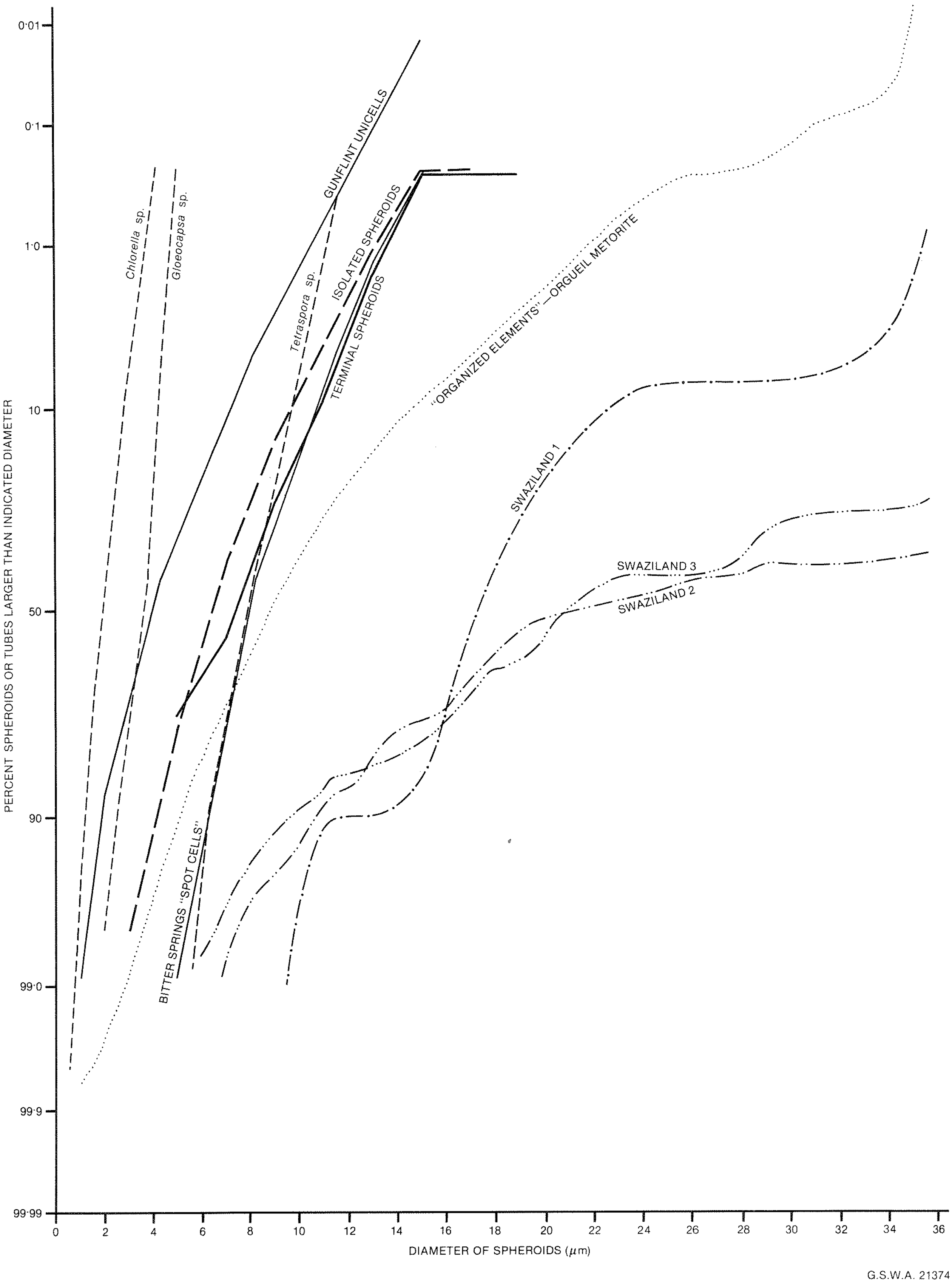

Figure 5. Probability plots of size data for populations of spheroids and terminal-spheroids from Discovery Chert sample, compared with populations illustrated by Schopf (1976).

However, an alternative theory—that the tubes were formerly bacteria with iron-encrusted sheaths which were preserved after decomposition of the original organic components—cannot be confirmed. If Muir's (1984a) criteria are rigorously applied, the microstructures must be rejected as microfossils. They cannot be compared directly with any extant taxa; in general they are much larger than most living micro-organisms, they show only a slight resemblance to some iron bacteria, and there is no evidence for an organic composition. When the microstructures are tested against statistical criteria for a biogenic origin, ambiguous results are obtained. The Discovery Chert material produces values which fall between the statistical parameters for biogenic and abiogenic microstructures outlined by Schopf (1976), although they are only slightly outside the values for biogenic material. Because of the lack of conclusive evidence, they must therefore be regarded as pseudofossils (Cloud and Morrison, 1979).

Rejection of objects as convincing microfossils does not necessarily rule out a biogenic origin, it merely emphasizes the lack of criteria for recognizing traces of biotic activity in the rock record. Schopf's statistical criteria were primarily devised for testing the biogenic origin of microfossil-like objects in the Archaean. They could, however, prove useful in evaluating the origin of problematic objects in younger rocks, where biologic activity may have played a significant role in sedimentation, and where diagenetic history could be interpreted from subsequent alterations in composition. Although results from the Discovery Chert are ambiguous, and a non-biogenic origin is more plausible using other criteria, Schopf's methods could be applied to test other cases where presumptive evidence of biogenic activity could have important consequences in palaeoenvironmental interpretation.

APPENDIX

by A. E. Cockbain

Schopf (1976, p. 30) introduced his Divisional Dispersion Index (DDI) in order to characterize the observed size range of a population and to provide a measure of relative dispersion. He defined DDI as "... the least number of sequential vegetative divisions required to mathematically 'reduce' the largest cell of a population to the smallest cell of that population ..." He further noted "... that the concept it represents could be expressed numerically in several other ways as well ..."

If D = diameter of largest member of the population and d = diameter of the smallest member of population, and $x = DDI$

then from

$$(D/d)^3 = 2^x$$

obtain, by taking logs on both sides,

$$3 \ln (D/d) = x \ln 2$$

Solve for x

$$x = 3 \ln (D/d) / \ln 2$$

and simplify to obtain

$$x = 4.33 \ln (D/d)$$

or

$$DDI = 4.33 \ln (D/d)$$

The DDI enables populations with different means to be compared and in this respect is similar to the coefficient of variation (V), with the important difference that V utilizes the standard deviation as a measure of 'spread', while DDI uses the observed range.

Simpson and others (1960) suggest that relative dispersion may be recorded as 100 (D/d). They state (p. 94) "of all measures of relative dispersion, this is undoubtedly the worst; ... For the more usual small samples, such a measure bears little rational relationship to the real variability". They conclude "Several other measures of relative dispersion are in occasional use but they have no advantage over V ..."

These two measures of relative dispersion are related:

$$100\,(D/d) = 100\,e^{(DDI/4.33)}$$

Simpson and others' (1960) comments apply equally to the DDI. If the observed range must be used as a measure of relative dispersion, then the computationally much simpler D/d is quite adequate. However, it would be preferable to compute the coefficient of variation rather than adopt a special index.

DDI	*D/d*	*DDI*	*D/d*
0	1.00	6	4.00
1	1.26	7	5.04
2	1.59	8	6.35
3	2.00	9	8.00
4	2.52	10	10.08
5	3.17	15	32.00

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STROMATOLITE EVIDENCE SUPPORTING A CORRELATION OF THE PROTEROZOIC UAROO AND BANGEMALL GROUPS, WESTERN AUSTRALIA

by Kathleen Grey

ABSTRACT

The occurrence of the stromatolite *Baicalia capricornia* Walter 1972 in the Mulya Dolomite at the northernmost extent of the Uaroo Group, northern Gascoyne Province, supports a correlation with the 1.1 Ga Bangemall Group, which outcrops to the east.

INTRODUCTION

Although van de Graaff and others (1980), Gee (1980), Hocking and others (1983), and Muhling and Brakel (1985) have discussed a possible correlation between the Uaroo and Bangemall Groups on broad stratigraphic grounds, there has never been any concrete evidence for direct correlation of formations within the two groups. The recent discovery, in the Uaroo Group, of the stromatolite *Baicalia capricornia* Walter 1972, previously known with certainty only from the Bangemall Group, supports a correlation of Mulya Dolomite in the Uaroo Group with part of the Irregully Formation in the Bangemall Group. This paper describes the occurrence of the stromatolites, and examines some of the regional implications.

GEOLOGICAL SETTING AND AGE

The Uaroo Group (van de Graaff and others, 1980) which derives its name from Uaroo Station (Lat. 22°46′S, Long. 115°22′E) forms outcrops only in the southeast corner of the Yanrey-Ningaloo sheet, and along the northern margin of the Winning Pool-Minilya 1:250 000 sheets (Fig. 1). It lies to the west of the Bangemall Basin at the northern tip of the Gascoyne Province, and is exposed in a north-northwesterly trending tectonic structure. It unconformably overlies a basement sequence of sedimentary, metamorphic, and igneous rocks of probable Proterozoic age, and various unassigned units. Its top is an erosion surface. The Uaroo Group consist of a series of arenite, shale, dolomite, and chert units, with a total thickness of about 8 km. Four formations have been recognized (van de Graaff and others, 1980), which, in ascending order, are the Rouse Creek Arenite, Mulya Dolomite, Mandorah Shale, and Tinkers Dolomite. The generalized distribution of the formations is shown in Figure 2, and the stratigraphy is described by van de Graaff and others (1980).

Previously, it was thought that the deposition of the Uaroo Group took place in a local basin during folding and uplift of older rocks (van de Graaff and others, 1980; Hocking and others, 1983). Direct correlation with other Proterozoic sediments of comparable age was considered unlikely because the rock types and relative thickness of units do not correlate satisfactorily. More recently, however, Muhling and Brakel (1985) and Williams (1986) favoured correlation with part of the Bangemall Group, despite disparities in thickness and lithology, but did not attempt to correlate individual formations.

A series of undifferentiated arenites and dolomites crops out to the west of the main Uaroo structure (Fig. 1). Their relationship to the Uaroo Group proper is not clear. A major tectonic lineament can be traced (Fig. 1) as a series of faults extending northwards from near Mount Tucker on the Winning Pool-Minilya sheet, along the western edge of the Nyang Range to near Mica Bore on the North West Coastal Highway. The lineament is not shown on the Yanrey-Ningaloo sheet, but is depicted in part on the State 1:2 500 000 geological map, and can be deduced from Landsat Scene 122-076, Winning Pool. Because of the problem of interpreting the undifferentiated units, the remainder of this paper is concerned only with the sediments to the east of the lineament (Fig. 2).

There is no direct evidence for the age of the Uaroo Group. The sequence post-dates early deformation and metamorphism in the Gascoyne Province; but, because of the similarity of folding, was presumably deformed at the same time as the Bangemall Group to the east. Radiometric dating of the Bangemall Group indicates an approximate age of 1.1 Ga (Gee, 1980).

STROMATOLITES IN THE UAROO GROUP

Stromatolites were reported from localities in both the Mulya and Tinkers Dolomites (Fig. 2, YNR 4 and 5) in the Uaroo Group (van de Graaff and others, 1980), but preservation is too poor for identification. Recently, samples were collected from two other localities (Fig. 2, YNR 6 and 7) near the northern end of the basin. Structural deformation has affected both localities, and at one, YNR 6, recrystallization is too extensive for taxonomic determination. Here stromatolites occur as a series of boulders along the strike of the Tinkers Dolomite, but there is no actual outcrop.

Air-photo interpretation shows that locality YNR 7 lies along strike from the Muyla Dolomite, in the nose of a tightly folded anticline. At this locality there are steeply dipping beds with numerous domed bioherms of branching-columnar stromatolites. The bioherms are strongly deformed (Fig. 3), reflecting their location in the axial zone of a chevron fold in the northern culmination of the basin. The columns, which are normally subcircular in cross-section, are compressed parallel to the strike of the bedding, and are elongate parallel to the dip. Nevertheless, tectonism was less severe here than at locality YNR 7. Details of the characteristic *Baicalia capricornia* lamination and microstructure are well preserved.

BIOSTRATIGRAPHIC SIGNIFICANCE

Baicalia capricornia has been previously (Walter, 1972) recorded with certainty only from the Bangemall Group, with the type locality in the Irregully Formation (Fig. 1) and additional, but poorly preserved, material from the Irregully and other formations of the western Bangemall Group (Grey, 1985). However, recently collected, better preserved material has prompted the review

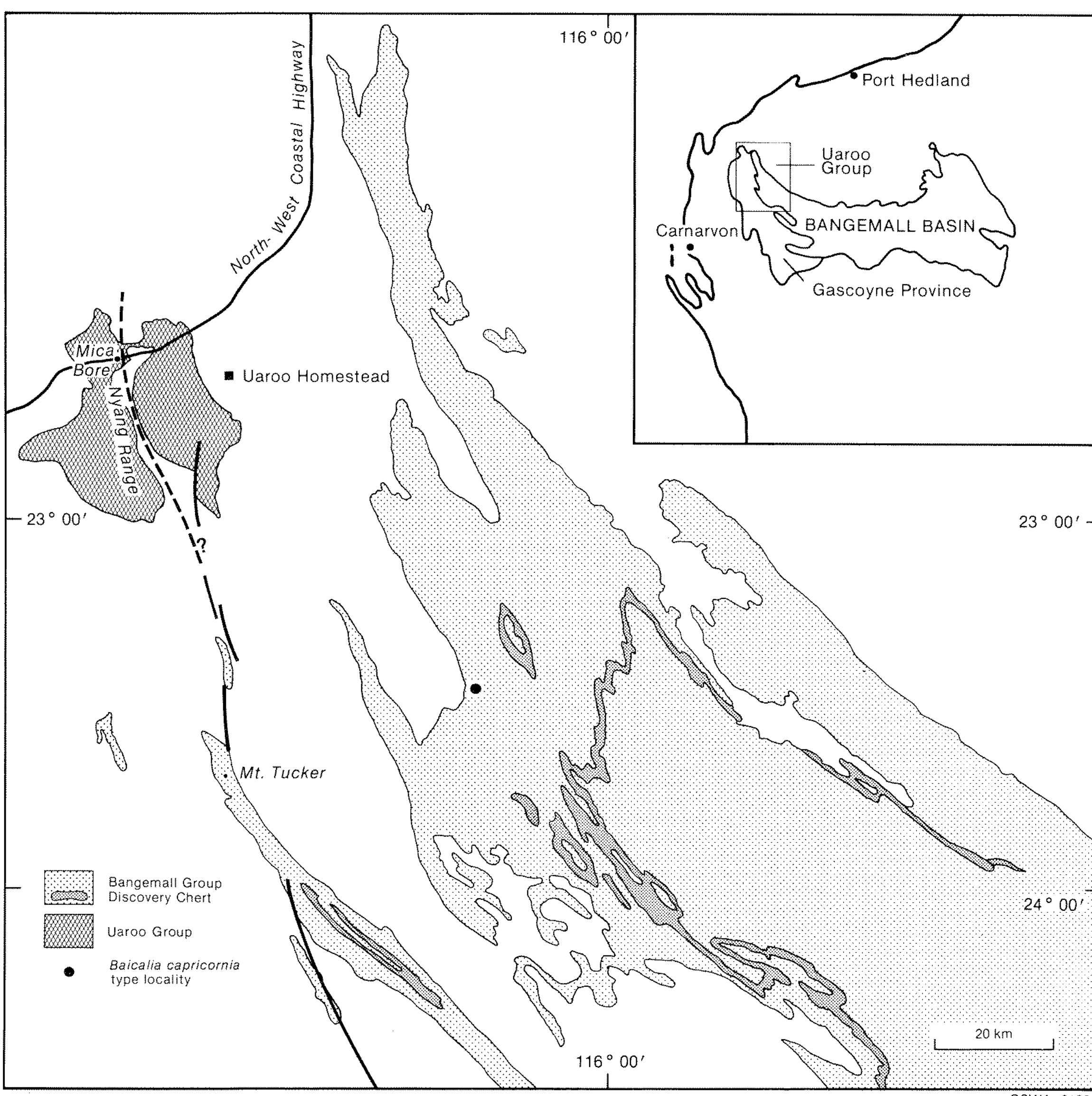

Figure 1. Outcrop distribution of the Uaroo Group and western part of the Bangemall Group.

of these identifications, together with specimens from the Moora Group tentatively identified as *B. capricornia* by Grey (1982).

The study of *B. capricornia*, still in progress, suggests that some of the specimens, in fact, belong to two new taxa, and that none of the specimens from the lower part of the Irregully Formation are *B. capricornia. B. capricornia* is apparently restricted to the upper part of the formation (above the Gooragoora Sandstone Member) and possibly to the Devil Creek Formation east of the area shown in Figure 1. The locality should be 20 miles (32 km) north-northwest of Coodardoo Gap and not "20 miles north-northeast" as reported by Walter (1972, p. 126-132).

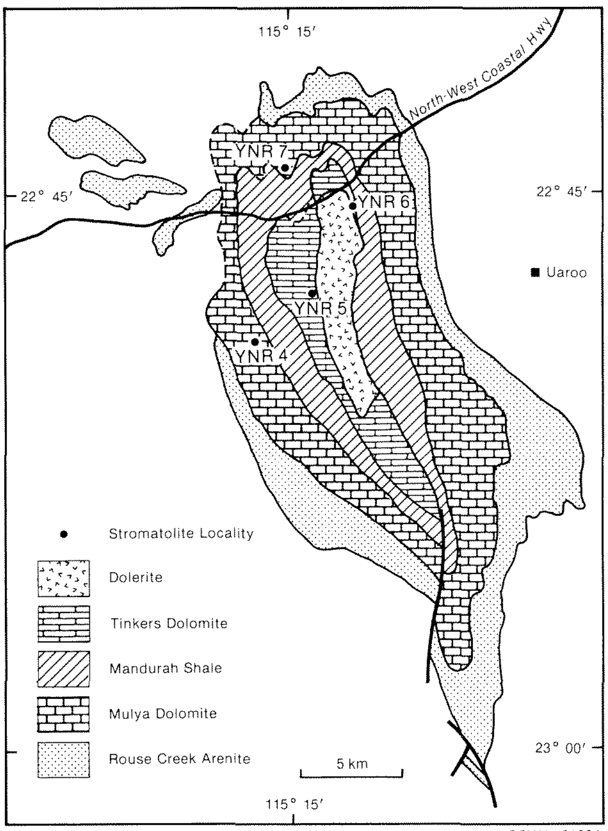

Figure 2. Stromatolite localities and generalized geology of the eastern part of the Uaroo Group.

Because the stromatolites in the Devil Creek Formation are "disoriented or roughly oriented, fragmented columns in a slump breccia" (Walter, 1972, p. 127), it is not clear whether they were derived from the Devil Creek Formation, or from the Irregully Formation; and the upper limit of the range of *B. capricornia* remains uncertain.

Consequently, although the stromatolites support a correlation with the Bangemall Group, they do not provide evidence for a precise correlation of formations. This is partly because the stratigraphy of the western part of the Bangemall Group is not clear, and is currently being revised (Chuck, 1984). Using existing terminology (Muhling and Brakel, 1985), the Mulya Dolomite is most probably equivalent to the upper part of the Irregully Formation, *i.e.* the Wannery and Warrada Members and their lateral equivalents (the "Unnamed carbonate sequence" of Chuck, 1984, Table 1). The Rouse Creek Arenite would then be equivalent to the Gooragoora Sandstone Member of the Irregully Formation. Correlations above the Mulya Dolomite are even less certain. However, both the Mandorah Shale and the Tinkers Dolomite may correspond to the Kiangi Creek Formation, which contains some stromatolites; or they could be equivalent to units higher in the Bangemall Group.

Some of these problems may be resolved once revisions of both the Bangemall stratigraphy and

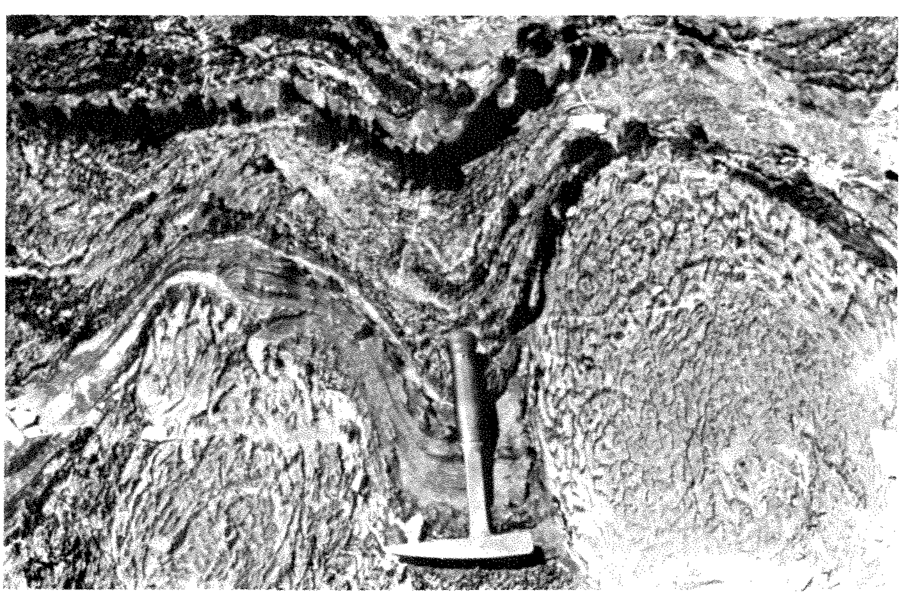

Figure 3. Two contiguous bioherms showing typical deformation of columns.

stromatolite taxonomy have been completed. Better preserved stromatolites from the Tinkers Dolomite would also assist in a more detailed correlation. The occurrence of *B. capricornia* in both the Uaroo and the Bangemall Groups is nevertheless significant because it provides the first convincing evidence supporting correlation of the two groups.

SYSTEMATIC DESCRIPTION

Baicalia capricornia Walter 1972
(figs. 3-7)

Algal stromatolites of *Cryptozoon*-type. Daniels 1966, p. 50, pl. 21A.

Baicalia capricornia Walter 1972, p. 126-132; pl. 5, fig. 1; pl. 17; pl. 18, figs. 1-2; text figs. 6, 34-36.

Material and localities

Holotype: University of Adelaide, Geology Museum Collection, sample S200.

New Material: GSWA F 46347-49 from Yanrey 1:250 000 sheet, GSWA fossil locality YNR 7, 3.5 km north-northeast of Tinkers Well, Lat. 22°43′09″S, Long. 115°14′05″E from the Mulya Dolomite, Uaroo Group.

Diagnosis

"*Baicalia* only rarely having peaks and cornices and predominantly with almost straight, subparallel columns; constrictions at the bases of branches are infrequent: the microstructure is banded" (Walter, 1972, p. 127).

Remarks

The stromatolites from the Mulya Dolomite correspond closely to the description given by Walter (1972), except for features which can be attributed to deformation.

Columns are oval to elongate in plan view with elongation in a general northeast direction. In vertical view, columns have been compressed, appear narrow, have steeply dipping laminae (Fig. 7) on faces at right angles to the direction of elongation, and are broad, with flattened laminae on faces parallel to the direction of elongation. The stromatolites occur in closely spaced, contiguous bioherms (Fig. 3), usually domed or hemispherical.

Branching columns develop from broad, low domes, approximately one metre in diameter, and show a radiating pattern around the dome (Fig. 4). Broad columns develop on top of the domed laminations and become narrower by successive alpha

GSWA 21636

Figure 4. Bioherm showing wrinkled lamination near base developing upwards into broad columns of *Baicalia capricornia* Walter 1972.

GSWA 21637

Figure 5. Columns of *Baicalia capricornia* showing typical branching pattern and detail of laminations.

GSWA 21638

Figure 6. ***Baicalia capricornia*** **F46348, polished slab showing details of lamination and branching pattern of slightly deformed columns.**

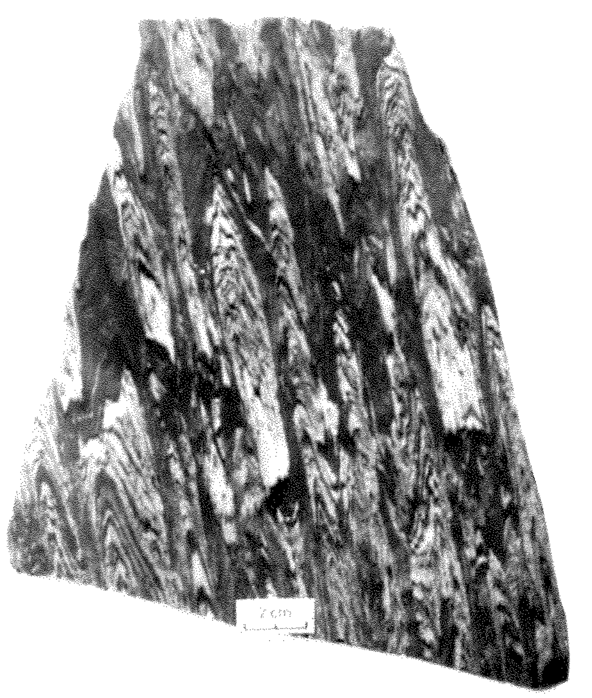

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Figure 7. ***Baicalia capricornia*** **F46349, polished slab showing deformed columns.**

parallel branching (Fig. 5) similar to the pattern described for the type locality by Walter (1972). In column shape (after allowing for tectonic distortion), margin structure, lamina shape and fabric,the stromatolites conform to the description of *Baicalia capricornia.*

Cleavage is strongly developed in the intercolumnar and interbiohermal spaces, but is less apparent in the columns, which show a more plastic deformation.

Distribution

Baicalia capricornia has only been recorded previously from the 1.1 Ga. old Bangemall Group, in the western part of the Bangemall Basin (Walter, 1972; Grey, 1985). The type locality is 180 km south of the Mulya Dolomite outcrop. Other forms of the group *Baicalia* occur in the Middle and Late Riphean of USSR.

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THE MINERAL POTENTIAL OF LAYERED IGNEOUS COMPLEXES WITHIN THE WESTERN GNEISS TERRAIN

by P. H. Harrison

ABSTRACT

Geological and geophysical evidence suggests that remnants of large basic/ultramafic layered complexes are present in several parts of the Western Gneiss Terrain. Various lines of evidence indicate that these rocks are some of the oldest preserved within the Western Gneiss Terrain. In spite of this, many are relatively undeformed and show only greenschist-facies assemblages. The rocks appear to be derived from a magma which had tholeiitic affinities; and they are considered to have potential for mineral deposits of the types normally associated with large tholeiitic layered intrusions. This includes potential for copper-nickel, chromite, titaniferous vanadiferous magnetite and platinum-group metals mineralization.

INTRODUCTION

In the course of studies directed towards a future bulletin on the potential for platinum mineralization within Western Australia, occurrences of basic* and ultramafic rocks and various mineral prospects in the Western Gneiss Terrain have been examined. These studies suggest that some, at least, of these occurrences are remnants of—formerly more extensive—large layered intrusions. These intrusions appear to be some of the oldest rocks preserved in the Western Gneiss Terrain. The individual remnant bodies appear to be much more extensive than has previously been recognized.

REGIONAL SETTING

The Western Gneiss Terrain (Gee, 1979) is that older, western part of the Yilgarn Block which has undergone polyphase medium- to high-grade metamorphism. This contrasts with generally lower grade metamorphism in the granitoid-greenstone terrain which forms most of the block.

Voluminous granitoids with whole-rock Rb-Sr ages of 2.6 to 2.7 Ga intrude both the granitoid-greenstone and gneiss terrains. In the case of the Western Gneiss Terrain, these granitoids intrude rocks of (dominantly) sedimentary and igneous composition which have reached metamorphic grades of amphibolite to granulite facies (Wilde, 1974).

The gneisses have been derived from rocks which span a period of geological history of close to 1 000 million years. In the north, protoliths of the gneisses at Mount Narryer have given Sm-Nd ages of 3.63 Ga and 3.51 Ga (de Laeter and others, 1981); in the central part of the Western Gneiss Terrain, the protoliths of the granofelsic paragneisses of the Chittering Metamorphic Belt have given Sm-Nd ages of 2.76 Ga and 2.89 Ga (Fletcher, and others, 1985). The absolute age range will probably be shown to be even greater than this, as ion probe U-Pb dating of zircon grains from quartzites at Mount Narryer has given clusters of ages around 3.75 Ga, 3.65 Ga, 3.4 Ga and 3.3 Ga (de laeter and others, 1985) and zircon ages of 4.1 Ga and 4.2 Ga have been obtained from orthoquartzites (Froude and others, 1983). This suggests that older protoliths may yet be found.

Several metamorphic belts, composed dominantly of schist and paragneiss of largely sedimentary origin, have been described within the Western Gneiss Terrain. The main belts are the Narryer (Williams and others, 1980), Jimperding and Chittering (Wilde, 1974), and Balingup (Wilde and Low, 1978) Metamorphic Belts (Fig. 1).

METAMORPHIC BELTS

In the Narryer, Jimperding, and Balingup Metamorphic Belts, the dominant rock is banded quartz-feldspar-biotite gneiss. While some of these gneisses may be orthogneiss derived from granitoid rocks, or pre-existing migmatites, the majority, from field evidence, appear to be derived

* The term 'basic' rather than 'mafic' is used in this paper, as anorthosites form a significant component of the sequences which are discussed.

from quartzo-feldspathic sediments. The paragneisses are associated with quartz-magnetite-amphibole rocks (metamorphosed banded iron-formation) and with orthoquartzites with well preserved cross-bedding. Gee (1979) suggested that these mature sediments originated from stable-shelf sedimentation.

In the Chittering Metamorphic Belt, on the other hand, a major rock type is quartz-feldspar-cordierite-biotite-garnet granofels. This is con-

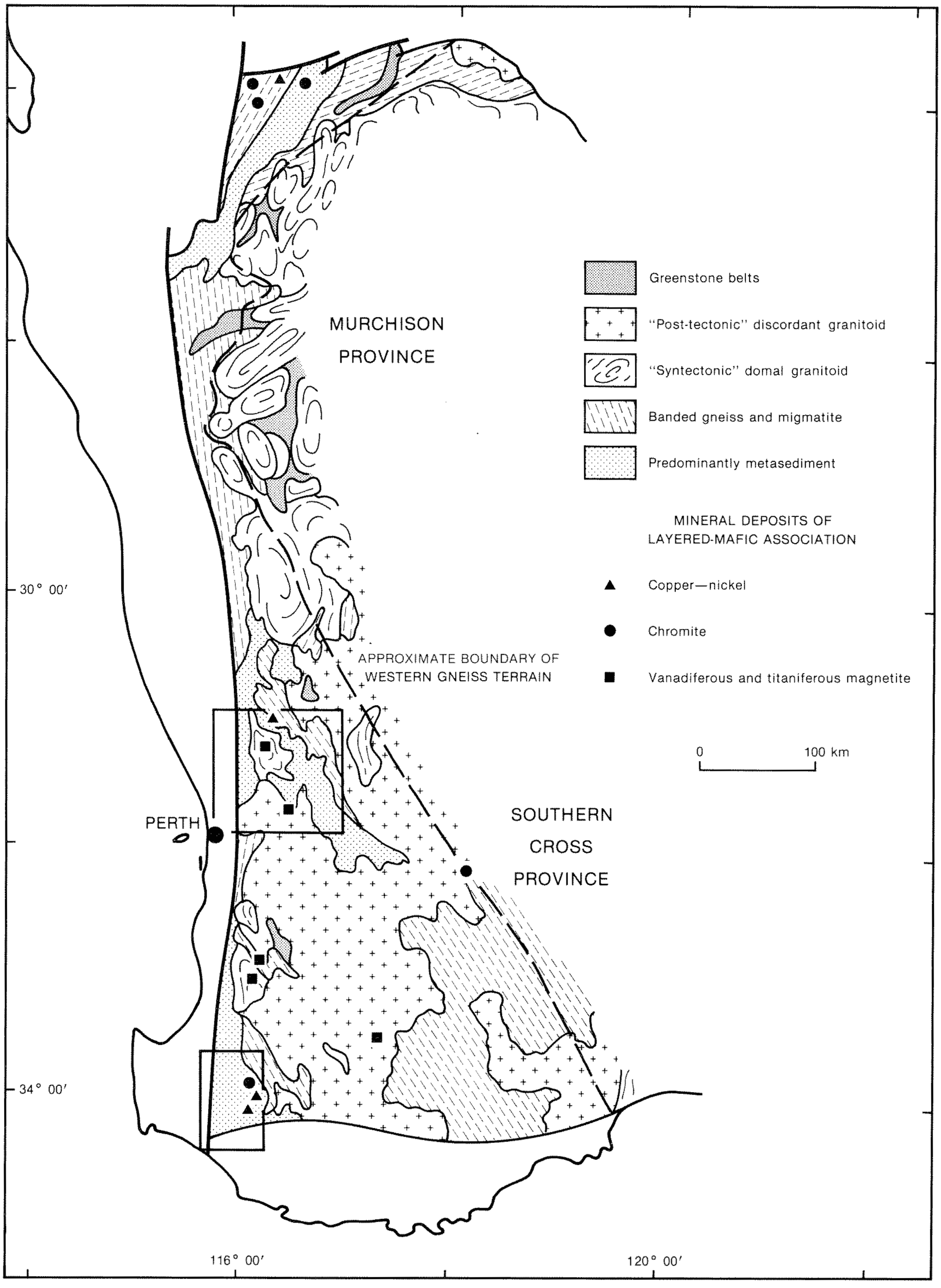


Figure 1. A simplified geological map of the Western Gneiss Terrain, showing the location of the main metamorphic belts, mineral occurrences of a possible layered-basic origin, and areas discussed in detail in this paper.

sidered to be metagreywacke and to indicate a greywacke-flysch type of sedimentation, less mature than in the other belts. Gneiss developed from this second facies also seems to be present, together with the shelf facies, in the Balingup Metamorphic Belt. Extensive remnants of a probable fifth metamorphic belt, of mainly Jimperding type, are present in the eastern part of the Western Gneiss Terrain on the Dumbleyung, Corrigin, and Kellerberin 1:250 000 sheets. This belt has not been formally named.

Amphibolites and/or mafic granulites, as small intrusive bodies, have been reported from all of the belts. Bodies of ultramafic rocks of various types have been identified in each of the metamorphic belts with the exception of the (younger) Chittering Metamorphic Belt.

A number of mineral occurrences, of a type normally associated with layered basic intrusive rocks, have been reported from within the Western Gneiss Terrain. The locations of these occurrences, taken from the map "Mineral Deposits of Western Australia 1981" together with additional information from Mines Department records, are shown on Figure 1, which also shows the locations of two areas examined in more detail below.

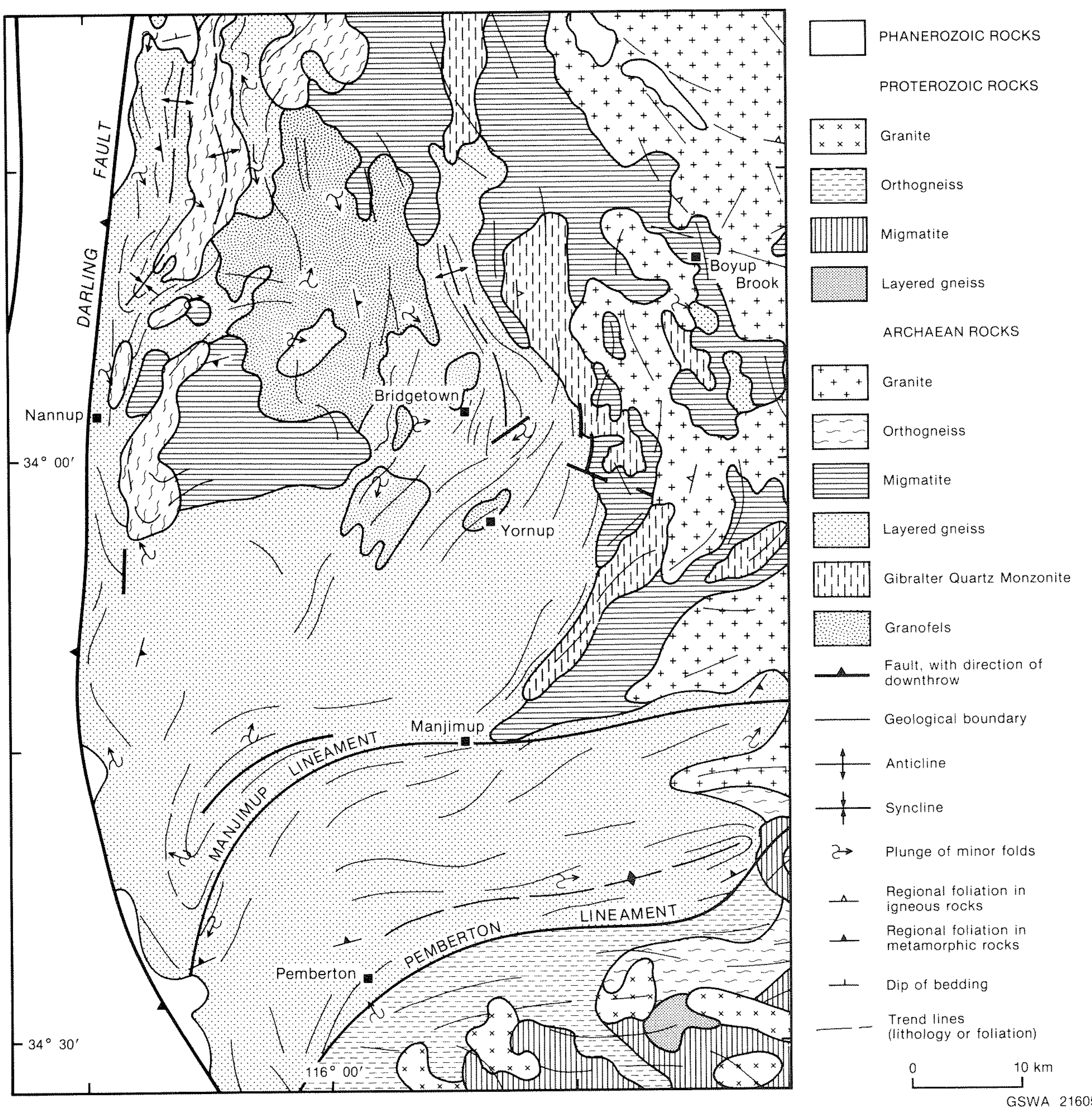


Figure 2. Generalized geology of the southern part of the Balingup Metamorphic Belt. Modified from Wilde and Walker (1979, 1981) and Fletcher and others (1983).

BALINGUP-BRIDGETOWN-DONNELLY RIVER AREA

The main elements of the geology of the southern part of the Balingup Metamorphic Belt, as interpreted from the very limited outcrop information, are shown in Figure 2. Here the paragneiss sequence appears to include both "shelf" and "trough" facies of the "Jimperding" and "Chittering" types; the relationships of which in this case, are not clear. A particular feature of the Balingup Metamorphic Belt is the high proportion of granitic orthogneiss (perhaps 30% of the total) which has intruded and subsequently been infolded with the paragneiss sequence.

Amphibolite and ultramafic rocks, described by Wilde and Walker (1979, 1981) as small lensoid occurrences, are present within the gneiss sequence. They apparently increase in proportion southwards and are most abundant in the area east of Bridgetown and near to Yornup (Fig. 3). This apparent increase may merely be a function of the better exposure in the more deeply incised area of the Blackwood River drainage system. A number of ultramafic units are also present in a belt from west of Balingup to north of Nannup.

South of a line between Yornup and Nannup, the belt is largely obscured by laterite and scattered areas of Tertiary sedimentary rock.

The Balingup Metamorphic Belt is structurally complex. It has undergone a number of periods of deformation and has been metamorphosed to amphibolite facies (with localized areas of granulite-

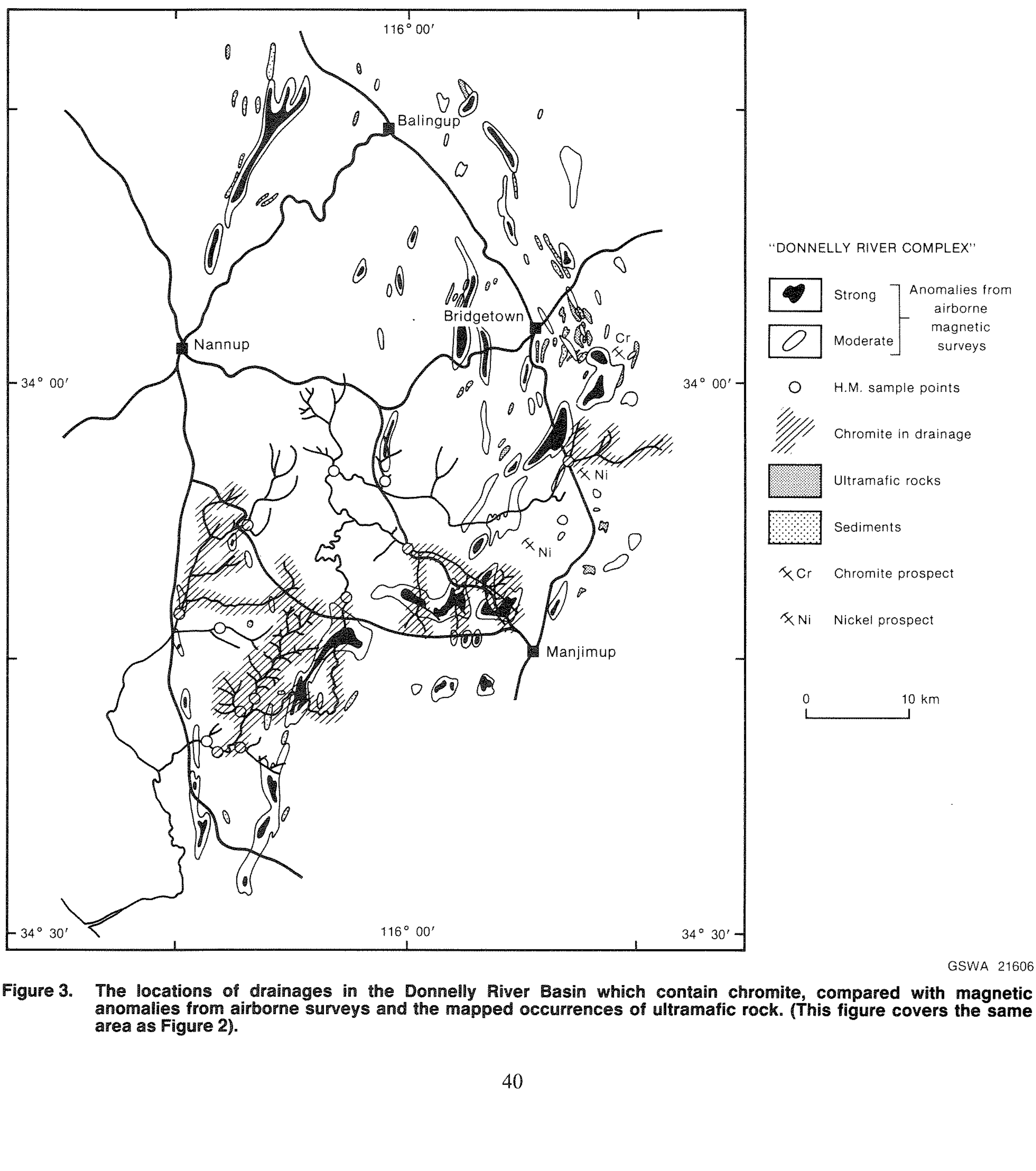


Figure 3. The locations of drainages in the Donnelly River Basin which contain chromite, compared with magnetic anomalies from airborne surveys and the mapped occurrences of ultramafic rock. (This figure covers the same area as Figure 2).

facies metamorphism). There is some suggestion of a regional doming around a large area of migmatite to the east of Nannup. The Manjimup Lineament corresponds closely with a change of lithological and structural trends: to the north of the lineament these are generally northerly; to the south, easterly and northeasterly trends are dominant. The Pemberton Lineament marks the southern limit of the Balingup Metamorphic Belt *sensu stricto*, as south of the lineament the gneiss appears to comprise orthogneiss of Proterozoic age (Fletcher and others, 1983).

A number of mineral prospects are known to be associated with the ultramafic rocks of the Balingup Metamorphic Belt.

At the Yornup prospect (34°11′50″S, 116°30′40″E) in the early 1970s, Planet Management and Research Pty Ltd completed 13 diamond-drill holes to test geophysical and geochemical anomalies associated with an area of lateritized and silicified ultramafic rocks. These holes encountered serpentinized peridotite and pyroxenite (and schistose talc-chlorite-anthophyllite derivatives) associated with metamorphosed norite and leuconorite. Disseminations and veinlets of pyrite, pyrrhotite, pentlandite, chalcopyrite, cubanite, bravoite, and rare millerite, were present; but the total amount of sulphide was very small and nickel assays rarely exceeded 0.3%. Planet also tested lateritic enrichments of nickel at Yornup and at the Palgarup prospect (34°13′20″S, 116°08′15″E). The best intersection, however, was only 0.83% nickel over 8.8 metres (Marston, 1984).

Outcrops of chromitite in ultramafic rocks occur close to the Blackwood River, approximately 6 km east of Bridgetown (Western Australian Government Chemical Laboratories, Annual Report for 1979).

Simpson (1914) reported that a prospector had submitted a sample of mineral concentrate from "24 miles (39 km) south of Nannup on the left-hand branch of the Donnelly River", in the general area of the present Vasse Highway. This sand sample, which was composed largely of cassiterite and monazite, carried highly anomalous quantities of platinum and assayed 8 910 g/t (291.05 oz/ton) platinum, 1681 g/t (54.90 oz/ton) osmiridium and 1 061 g/t (34.65 oz/ton) gold. At that time, ultramafic rocks were not known from within the Donnelly River Drainage basin. Saint-Smith (1912) noted that the sample closely resembled platinum-bearing beach sands which were then being mined on the northern coast of New South Wales.

"DONNELLY RIVER COMPLEX"

The general area of the platinum occurrence reported by Simpson has recently been examined by the writer. Heavy-mineral samples were panned from the Donnelly River and various tributaries. These were further concentrated by tetrabromoethane heavy-media separation, and classified using a Frantz magnetic separator before microscopic examination. The distribution of the sampling points, shown in Figure 3, was dictated by forest tracks and quarantine regulations. No platinum, platinum-group minerals, gold, or cassiterite were found in any of the samples; however, several did contain chromite (Ahmat, 1983a). The drainages from which chromite was reported have been indicated on Figure 3, together with the mapped outcrops of ultramafic rocks (and the orthoquartzites and metamorphosed banded iron-formations which are frequently associated with them). Superimposed on Figure 3 are the locations of magnetic anomalies, which have been empirically interpreted from several semi-detailed airborne magnetic surveys that are held in Mines Department "M series" records and from Bureau of Mineral Resources 1:250 000 scale magnetic maps.

The close association between the magnetic anomalies and ultramafic rocks in the areas of outcrop, and between the magnetic anomalies and drainages that contain chromite, is considered to indicate that extensive bodies of ultramafic and basic rocks, similar to those near Bridgetown and Yornup, are present below the laterites of the Donnelly River area. This belt of ultramafic and basic rocks is here informally termed the "Donnelly River Complex".

Exposed parts of the complex in the Bridgetown-Yornup area, and a number of samples of drill material from drill holes near Yornup, have been examined in an attempt to ascertain the nature of the rocks involved.

Most of the ultramafic rocks show greenschist-facies assemblages in areas of predominantly amphibolite-facies assemblages. For this reason they were regarded by Wilde and Walker (1981) as having been intruded after the amphibolite-facies metamorphism. East of Bridgetown, however, drilling carried out by Western Mining Corporation (W.M.C.) showed that the ultrabasic bodies are themselves intruded by granitic rocks (Mazzucchelli, 1981). It appears probable that the latter are either granitic phases of the Gibralter Quartz Monzonite or orthogneiss related to the Logue Brooke Granite; both of these rock units have themselves been affected by the amphibolite-facies metamorphism, which has been dated at

Figure 4. Chromite-rich metadunite from near Bridgetown. A relic cumulate texture after olivine — now replaced by serpentine and chlorite — is outlined by chromite grains. (Sample 78128. Field of view 20 mm. Plane-polarized light).

2.84 ± 0.20 Ga (D. A. Nieuwland, pers. comm., 1977; Fletcher and others, 1983). Cuttings taken from holes drilled by W.M.C. near Yornup (G.S.W.A. sample 78140) show retrograde greenschist-facies metamorphism of tremolite-olivine rocks previously at mid-amphibolite facies (Ahmat, 1983b). This demonstrates that the Yornup ultramafic bodies, at least, pre-date the amphibolite-facies metamorphism.

Nevertheless, many of the ultramafic bodies examined are relatively undeformed and show well-preserved primary igneous textures which suggest that they have never been above greenschist facies. It appears probable that the central parts of these relatively thick, competent, and anhydrous bodies, which are composed of high-temperature minerals, have been shielded from many of the thermal and dynamic effects of the amphibolite-facies metamorphism. Further evidence of this shielding effect was noted in the field. In the Bridgetown area, a number of the bodies are "rimmed" by foliated talc-chlorite-serpentinite rocks. In the Palgarup area, anthophyllite-talc schist appears to be a more deformed equivalent of nearby ultramafic rocks (Wilde and Walker, 1981).

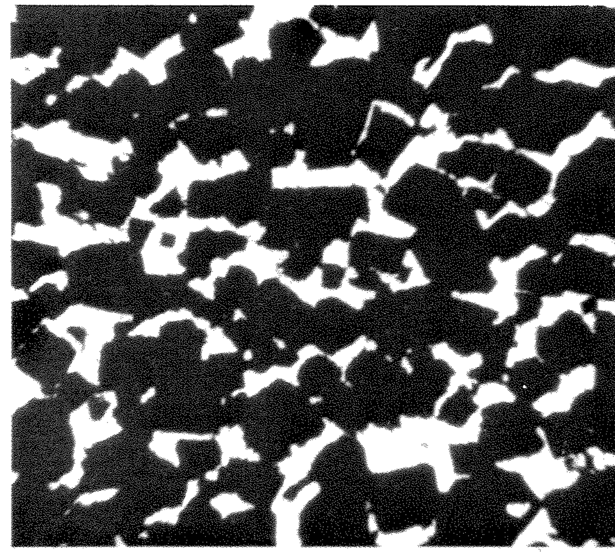


Figure 5. A chromitite layer from near Bridgetown. The chromite grains show a well-developed "chain-link" texture. (Sample 78128. Field of view 2.5mm. Plane-polarized light).

A number of the undeformed rocks sampled near Bridgetown show relic cumulate textures of elongate olivine grains—now replaced by chlorite and serpentine—outlined by chromite (Fig. 4). The chromitite horizons show well-developed "chain-link" textures indicative of primary igneous layering (Fig. 5).

Analyses of a number of chromitites and the host metadunites are presented in Table 1.

TABLE 1. ANALYSES OF CHROMITITES AND AN ULTRAMAFIC ROCK FROM NELSON LOCATION 575 (EAST OF BRIDGETOWN)

Sample	*78125*	*78127*	*78129*	*78131*
		Percentage		
Fe(total)	31.6	27.6	30.8	14.2
Cr_2O_3	22.0	33.6	26.8	6.58
FeO	19.5	20.5	19.2	9.34
SiO_2	11.6	6.83	9.03	31.6
Al_2O_3	5.24	8.24	6.77	3.29
MgO	11.4	8.72	9.73	28.5
		Parts per million		
Cu	44	28	44	39
Ni	2000	1400	1800	2000
Pt	<[0.04]	<[0.04]	<[0.04]	<[0.04]
Pd	<[0.005]	<[0.005]	<[0.005]	<[0.005]

Analyses for platinum-group metals by Kalgoorlie Metallurgical Laboratory using a specially developed fire-assay method. Other analyses by the Government Chemical Laboratories.

78125, 78127, 78219, are channel samples of chromitite. 78131 is a serpentinized metadunite.

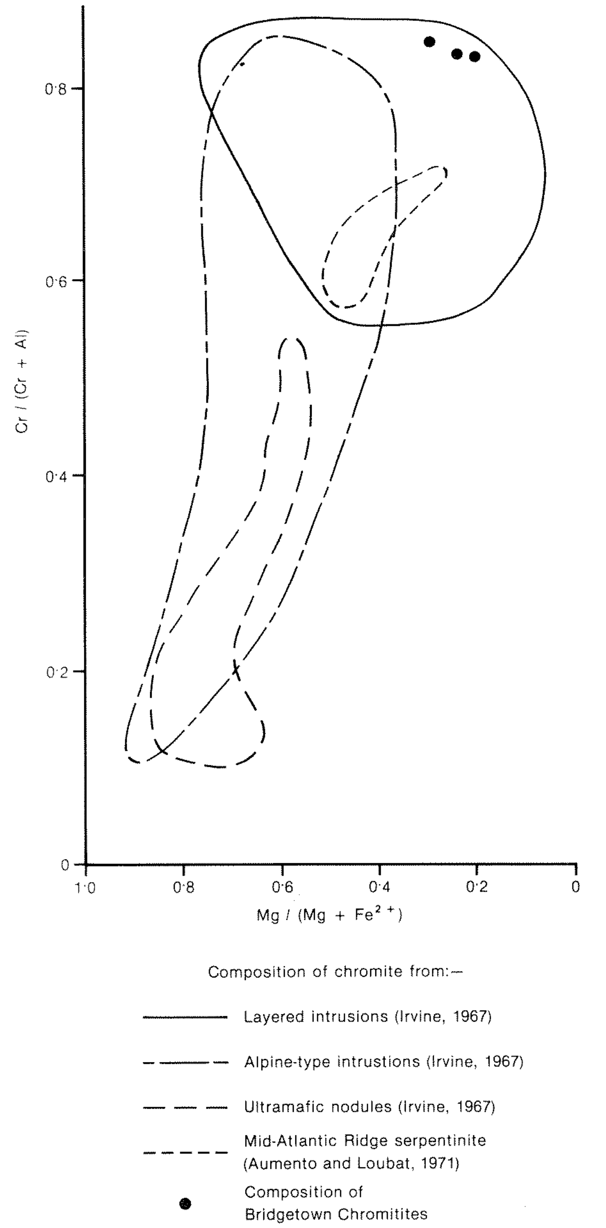


Figure 6. Composition of chromitites from Bridgetown compared with the compositional fields of chromite from layered-intrusive complexes, alpine-type ultramafic bodies and deep-ocean serpentinites.

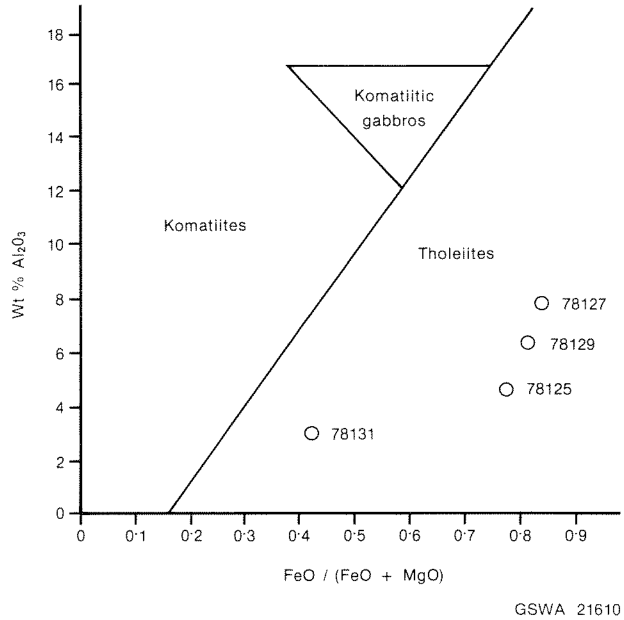


Figure 7. Bridgetown rocks plotted on chemical-variation diagrams for Al_2O_3 versus FeO*/(FeO* + MgO) constructed after Naldrett and Cabri (1976) and Naldrett and Goodwin (1977). FeO* = Total Fe as FeO.

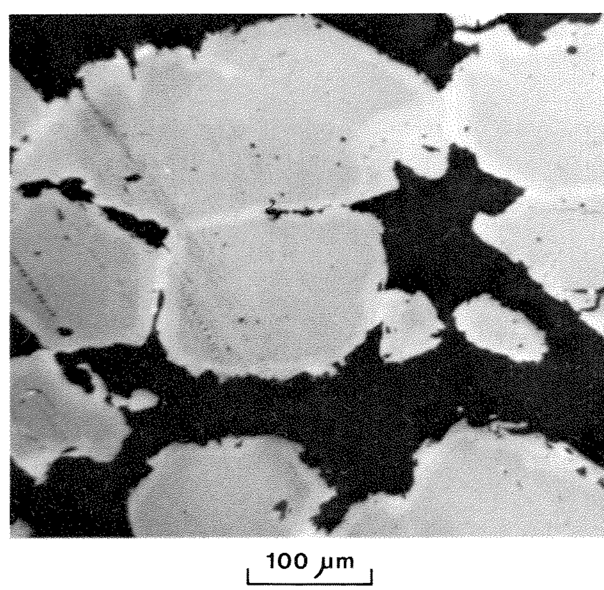


Figure 8. Chromitite from near Bridgetown showing ferritchromit alteration of chromite grains: alteration of the rims of grains to magnetite. (Sample 78128. Field of view 625 microns. Reflected light).

An examination of the Cr/(Cr+Al) versus Mg/(Mg+Fe) ratios of these chromites (Fig. 6) supports an origin in a layered intrusion. A plot of the Al_2O_3 versus FeO/(FeO+MgO) ratio (Fig. 7) suggests that the magma had a tholeiitic rather than a komatiitic affinity. However, these data have to be treated with some caution as ferritchromit alteration—the replacement of the rims of the chromite grains by magnetite, (Fig. 8)—suggests that during metamorphism, under high pO_2 conditions, there has been some migration of chromium and magnesium from the chromite into chromium-bearing magnesian chlorites which replace the former olivine grains.

Potential for mineralization

The combination of the evidence suggests that the "Donnelly River Complex" is a series of remnants of one, or more, layered tholeiitic intrusions. The recognition of chromitite horizons in the area east of Bridgetown, together with the extensive presence of detrital chromite in the drainage samples, suggest a considerable potential for minable deposits of chromite. The economically detrimental ferritchromit alteration, caused by the metamorphism (resulting in an increase in the $FeO:Cr_2O_3$ ratio), means that exploration should be restricted to the central, less altered, parts of the larger ultramafic remnants, where higher grade chromite may still be preserved.

The unusually low values in the analyzed samples suggest that the magma, at the time of formation of the Bridgetown chromitites, was deficient in the platinum-group metals. This downgrades the potential of the complex for platinum mineralization, although it is conceivable that the deficiency might result from the precipitation of platinum (with sulphides or chromite) lower in the sequence.

The copper and nickel sulphides intersected by the Planet drilling, near Yornup, demonstrate that there is some potential for nickel sulphide mineralization. The potential for platinum-group metals associated with this mineralization cannot be assessed, as analyses for these elements were not carried out. Platinum and palladium analyses of drill core from an (unsuccessful) exploration programme for chromite carried out by the Shell Company of Australia Ltd - Metals Division, in the area of the Yornup nickel prospect, were all less than the detection limit of the method used: <0.005 ppm Pd, <0.05 ppm Pt (Richards, 1981).

THE ARCHAEAN PORTION OF THE PERTH 1:250 000 SHEET

The Archaean geology of the Perth sheet (Figure 9) is dominated by paired metamorphic belts, which are composed mainly by gneisses of metasedimentary origin.

Most of the rocks of the Jimperding Metamorphic Belt appear to be derived from quartzo-feldspathic sediments formed under stable-shelf sedimentation. In the northern part especially, generally flat-lying metasedimentary sequences make the belt appear (superficially) structurally relatively simple. These flat-lying sequences may be the preserved remnants of large, early-generation recumbent folds formed by horizontal-folding tectonics (Gee, 1979).

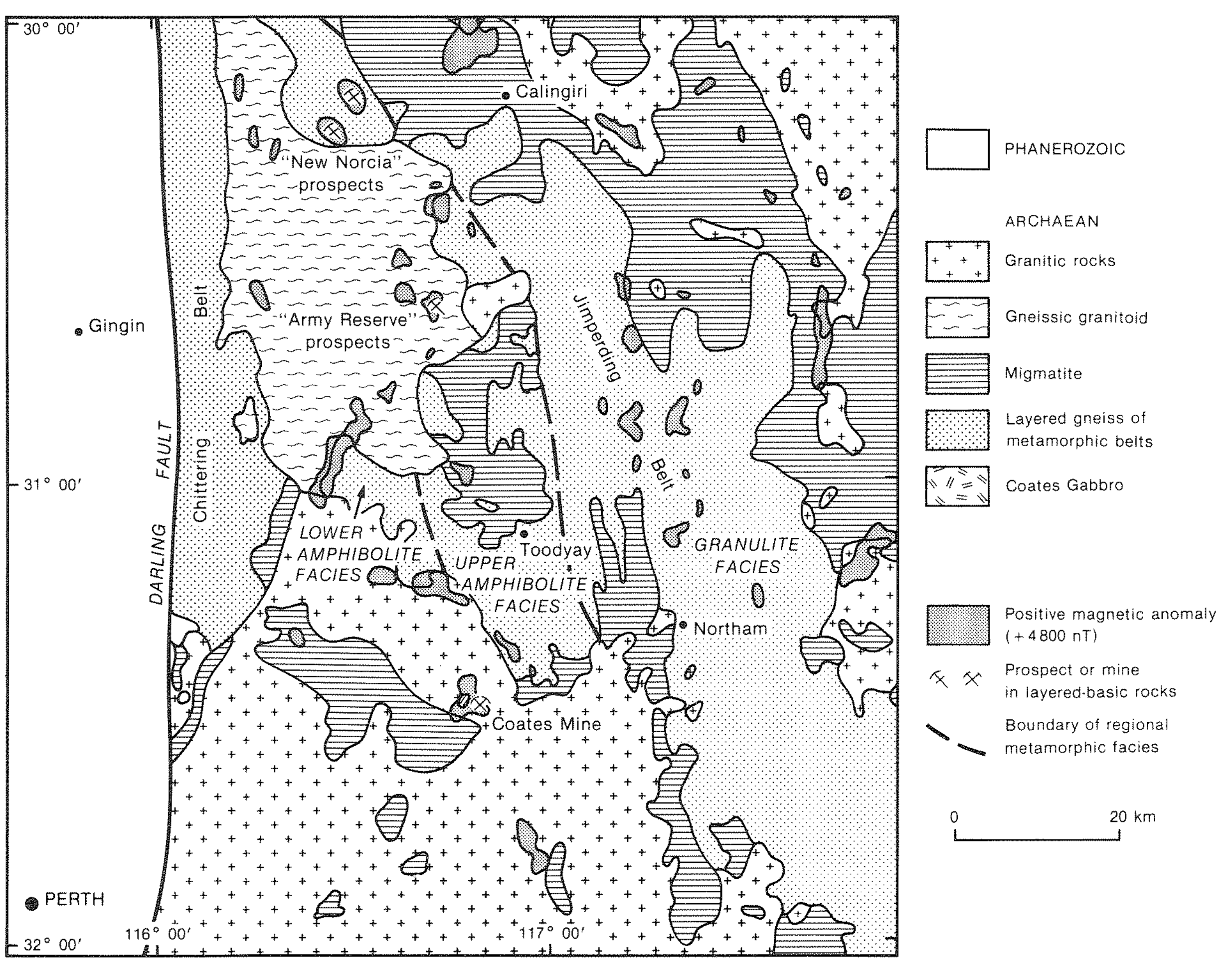


Figure 9. Simplified geology of the Archaean portion of the Perth 1:250 000 sheet, showing major magnetic anomalies and mineral prospects associated wth layered-mafic rocks. (Partly modified from Wilde and Low, 1978).

The Chittering Metamorphic Belt, on the other hand, is characterized by gneisses probably derived from greywacke-flysch sediments. It contains extensive tracts of steeply inclined gneiss with cataclastic and mylonitic zones. This combination was considered by Gee (1979) to be characteristic of vertical-shear tectonics.

Approximate metamorphic isograds, recognized from the regional mapping, are shown on Figure 9.

The Chittering Metamorphic Belt is characterized by (high-pressure series) amphibolite-facies metamorphism, while the Jimperding Metamorphic Belt shows (low-pressure series) metamorphism, grading from amphibolite facies in the west to granulite facies in the east.

Subordinate units of amphibolite are present in the Chittering Metamorphic Belt while both amphibolite and mafic granulite are known in the Jimperding Metamorphic Belt. Ultramafic rocks are apparently restricted to the Jimperding Metamorphic Belt. Wilde and Low (1978) regarded them as subconcordant weakly metamorphosed intrusions into the layered gneissic sequence, although they do pre-date the final deformation.

The metamorphic belts are surrounded by extensive areas of migmatite, and are intruded and disrupted by voluminous granitoids. The majority of the granitoids are post-tectonic, but one in the north of the sheet appears to be an older, syntectonic unit (Figure 9).

A body of gabbro at Coates Siding, recognized during the mapping, consists of several phases including leucogabbro, magnetite-gabbro, and gabbro. The body is enclosed by granite, but as no good exposures of the gabbro-granite contact are observed, the age relationships are unclear (Baxter 1978). Dolerite dykes intrude both the gabbro and the granitoid rocks.

A limited amount of isotopic dating has been completed within the area of the Perth Sheet. McCulloch and others (1983) showed that two groups of orthogneisses from the Toodyay and Northam areas have Sm-Nd models ages of 3.15 to 3.24 Ga and 2.95 to 3.05 Ga. A mafic granulite gave a model age of 3.05 Ga. U-Pb studies of zircons by Nieuwland and Compston (1981) showed that the protoliths of the banded paragneisses, in the same area, had dates of approximately 3.25 Ga. While detrital zircons with dates of 3.34 Ga, obtained from orthoquartzites from near Toodyay, suggest the presence of still older crustal material in the Jimperding Metamorphic Belt. A preliminary, unpublished, Sm-Nd model age of material from the Coates Gabbro suggests this unit may be older than any of the gneisses so far dated from the Jimperding Metamorphic Belt (Fletcher, pers. comm., 1984). Fletcher and others (1985) have shown, by Sm-Nd model dating, that the protoliths of the granofelsic paragneiss in the Chittering Metamorphic Belt are considerably younger than the rocks of the Jimperding Metamorphic Belt, giving ages of 2.76 and 2.98 Ga. Arriens (1971) showed that the post-tectonic granitoids have whole-rock Rb-Sr ages of around 2.6 Ga. No dates have been reported for the syntectonic granitoid, but this can be shown, from field evidence, to be younger than the gneisses of the Chittering Metamorphic Belt.

Magnetic anomalies of over 4 800 nT (taken from the B.M.R. regional magnetic survey) have been superimposed on Figure 9. In the eastern part of the sheet, a number of these anomalies occur over mapped outcrops of ultramafics or metamorphosed banded iron-formation rocks, but the relationship is far from precise. Many of the anomalies apparently relate to migmatites or gneisses: a large number occur over areas obscured by superficial deposits. In the west of the sheet, a north-trending line of anomalies occurs over rocks which are largely obscured by laterite, alluvial, and colluvial deposits. One anomaly includes the Coates Gabbro (although the anomaly is much more extensive than the mapped outcrop), and amphibolite and ultramafic rocks crop out on the flanks of a large anomaly which occurs over a laterite-covered area of the Julimar Forest (centred on Lat. 31°30'S, Long. 116°14'E).

"JULIMAR COMPLEX"

A number of mineral deposits and mineral-exploration prospects occur along this (western) north-trending line of magnetic anomalies (Fig. 9). Drilled material from these, together with the sparse outcrops, provides some evidence as to the cause of the anomalies.

The magnetite-gabbro phase of the Coates Gabbro contains 20 to 40% magnetite and ilmenite. The balance is made up of labradorite, hornblende, biotite, epidote, and augite. An estimated minimum of 39 Mt of primary unoxidized mineralization, containing an average of 0.51% V_2O_5 (using a 0.4% V_2O_5 cut-off) is present in this unit. This resource is overlain by an inferred 5.9-7.0 Mt of oxidized rock averaging 0.55% V_2O_5. Near surface this is, in turn, overlain by a laterite caprock which has indicated reserves of 1.2 to 1.5 Mt averaging 0.88% V_2O_5 (Baxter, 1978).

Thin section examination of material from a test shaft at Coates has shown evidence for a (basal?) ultramafic layer to the gabbro body. One sample consists of 95% serpentine minerals after olivine, in a metadunite with a well-preserved adcumulate texture (Figure 10). The outlines of former olivine grains are marked by chromite and a number of secondary veinlets of titaniferous, vanadiferous magnetite are present. Here again, a greenschist-facies assemblage is present in rocks which probably pre-date the amphibolite-facies gneisses and provides further evidence of the resistance of thick, layered-intrusive bodies to many of the more severe effects of metamorphism. In this case, "wavy-textured" serpentinization is the only evidence of any significant dynamic effect of metamorphism.

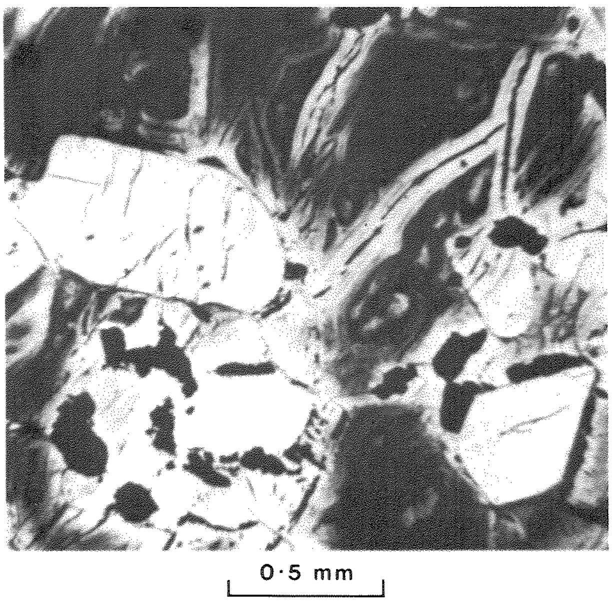


Figure 10. Serpentinized dunite from Coates Siding. A relic adcumulate texture after olivine grains is outlined by chromite. Small secondary veins are titaniferous magnetite. A "wavy" serpentinization is the only evidence of significant stress effects during metamorphism. (Sample 78180. Field of view 2.5mm. Plane-polarized light).

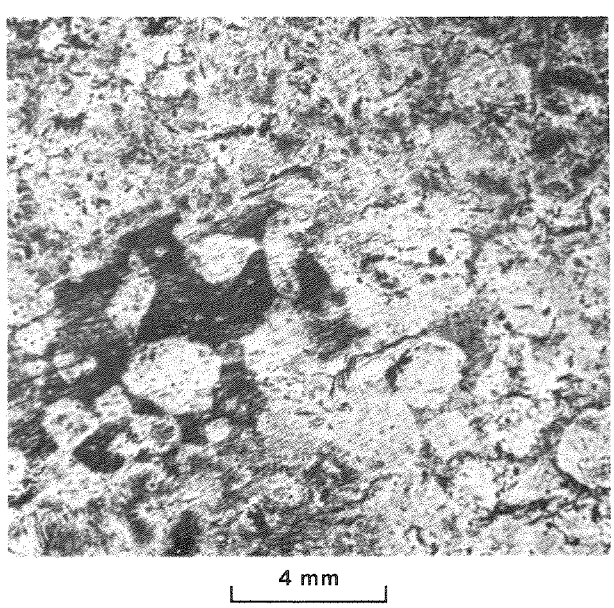


Figure 11. Metaperidotite from Julimar Creek. Cumulate grains after olivine are replaced by serpentine. The former olivine grains were enclosed by poikilitic orthopyroxene which has been replaced by tremolite and chlorite. The intercumulus material is "dusted" by fine-grained opaque minerals. (Sample 78169. Field of view 20 mm. Plane-polarized light).

Further north, in the Julimar Creek exposures, primary igneous textures of the type associated with layered-mafic intrusions are preserved. One sample shows serpentinized grains (after cumulate olivine) surrounded by a mixture of tremolite and chlorite which has replaced poikilitic orthopyroxene (Figure 11).

East of the Bindoon Army Training Area (centred on 31°20′S. 116°21′E), Hamersley Exploration Limited and Alcoa of Australia Limited carried out a drilling programme to test one of the north-trending magnetic anomalies. Existing regional mapping suggests that this area consists of laterite and scattered outcrops of deeply weathered (gneissic) syntectonic granitoid. The exploration programme of six percussion-drilled and three diamond-cored holes showed that the magnetic anomaly is underlain by an unexposed basic complex, consisting of leucogabbro, magnetite-gabbro, and anorthosite. A number of thin pyroxenitic layers are also present within the gabbro sequence, which has been intruded by (late stage?) granophyric differentiates and younger dolerite dyke rocks. The basic body appears to have been intruded into an older sequence of metasediments and (possible) acid volcanics (Pontifex, 1979). The whole sequence appears to now form a roof pendant to the older gneissic granite; a typical section is shown in Figure 12.

The drilling showed that vanadiferous, titaniferous magnetite-bearing layers (up to 25 metres thick) were present in lenses with strike lengths of at least 1.1 km. It appears, however, that the lateritic enrichment, which is a feature of the Coates mineralization is largely absent. The joint venturers estimated that the tonnage potential was 90 000 tonnes per vertical metre, at a grade of 0.3% V_2O_5 or 20 000 tonnes per vertical metre at 0.45% V_2O_5. An interesting feature is the presence of up to 10% apatite in some of the leucogabbros. This and the granophyric rocks suggest that the sequence includes the more alkaline upper part of a thick layered-intrusive body.

It is anticipated that this, and a number of other layered-basic bodies from the Western Gneiss Ter-

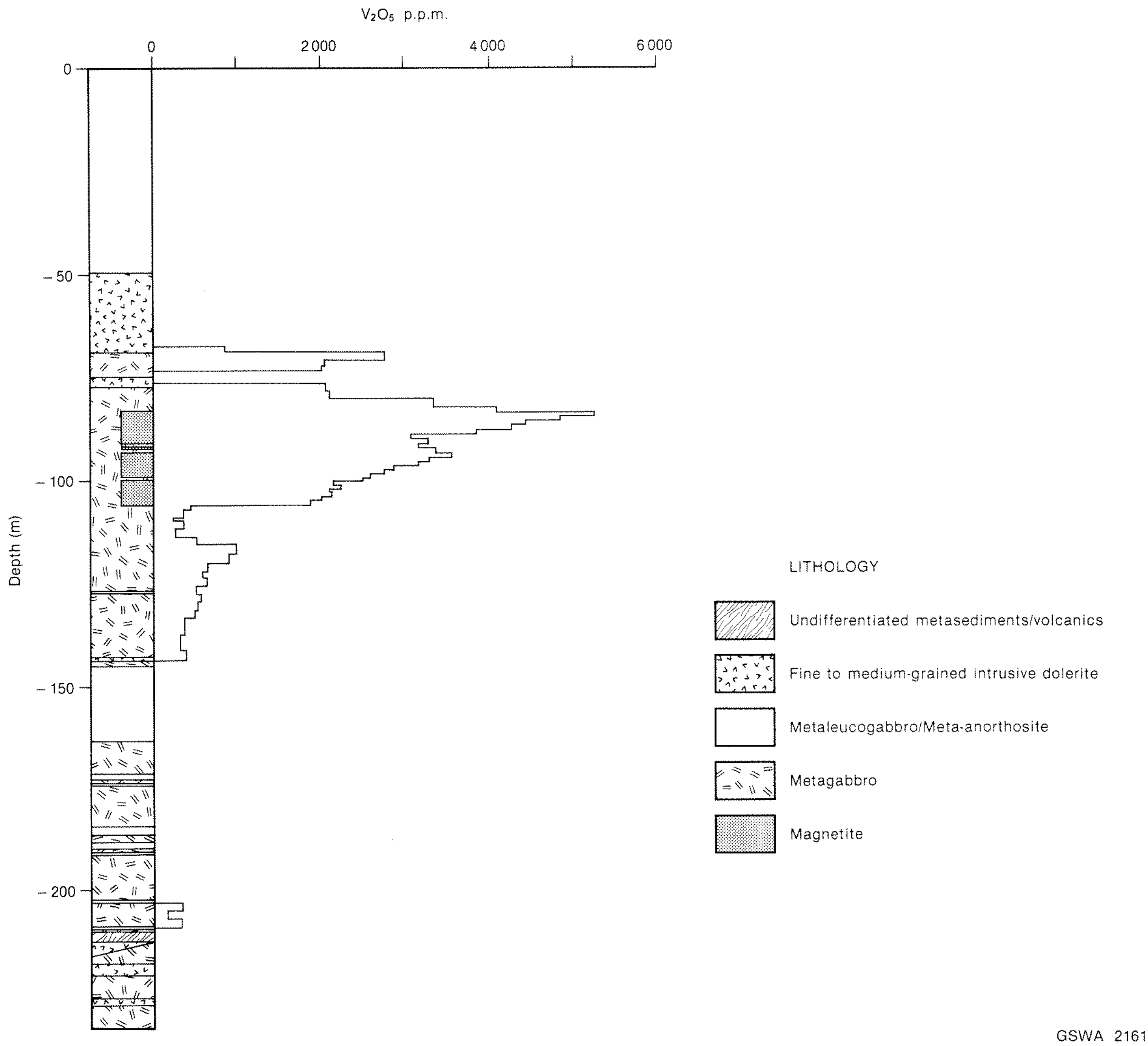
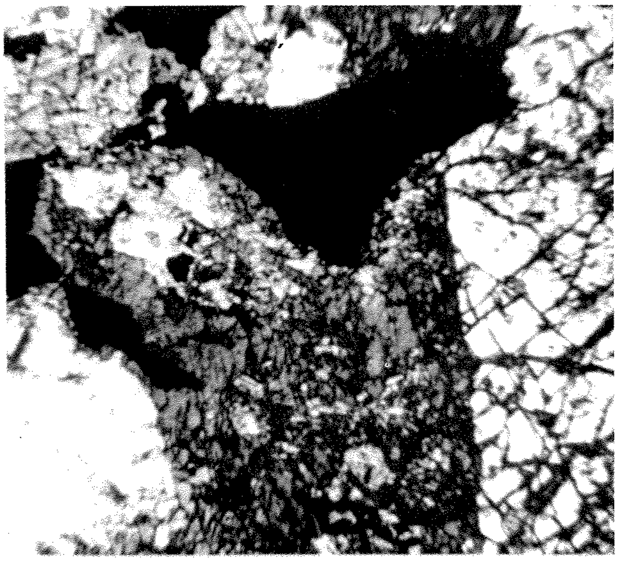


Figure 12. A typical drill-hole section, "Army Reserve" prospect, D.D.H.1., showing analysis for V_2O_5.

rain, will shortly be dated using the Sm-Nd method. To this end, a number of specimens of drill core have been examined in thin section. The samples examined include a metamorphosed leucogabbro with cumulus and intercumulus textures, in which actinolite replaces clinopyroxene; the plagioclase (of andesine composition) is partly saussuritized and, in places, replaced by grossular garnet (Figure 13). Ahmat (1984) regards the presence of the garnet as indicating calcic metasomatism. The intercumulus oxide mineralization in this sample (Figure 14) includes both titanomagnetite (with ilmenite exsolution lamellae) and primary ilmenite. Minor amounts of pyrite and chalcopyrite are present in some samples.

Another north-trending magnetic anomaly, the New Norcia Prospect (31°05′50″S, 116°15′50″E), has been examined in some detail by Otter Exploration N.L.—for a time in joint venture partnership with Shell Minerals Exploration (Australia) Pty Ltd. Their exploration programme, which included intensive rotary air-blast drilling, percussion drilling and eleven diamond-cored holes, has shown the presence of a layered sequence of basic and ultramafic rocks. These contain extensive zones of disseminated sulphides with rare thin massive zones. The higher grade intersections, assaying 1.5% nickel with similar values of copper, are quite narrow: 0.2 to 0.4 m (Marston, 1984).

A typical drill section is shown in Figure 15. Although the structural relationships from hole to hole are not clear, it appears that there is major phase layering involving metaperidotite (now serpentine-tremolite-chlorite rock), metapyroxenite (now largely tremolite-chlorite rock), and various metamorphosed melagabbros and leucogabbros. These rocks appear to be rich in tremolite rather than actinolite, indicating a probable original noritic composition. The distribution of copper and nickel may indicate the presence of some cyclicity in the layering (Figure 15).



GSWA 21616

Figure 13. **Meta-leucogabbro, D.D.H.1. "Army Reserve" prospect. Clinopyroxene has been replaced by blue-green actinolite and some minor chlorite. Plagioclase has been partly saussuritized and is partly replaced by grossular garnet. The intercumulus opaques are mainly titaniferous magnetite and ilmenite. (Sample 78157. Field of view 2.5 mm. Plane-polarized light).**

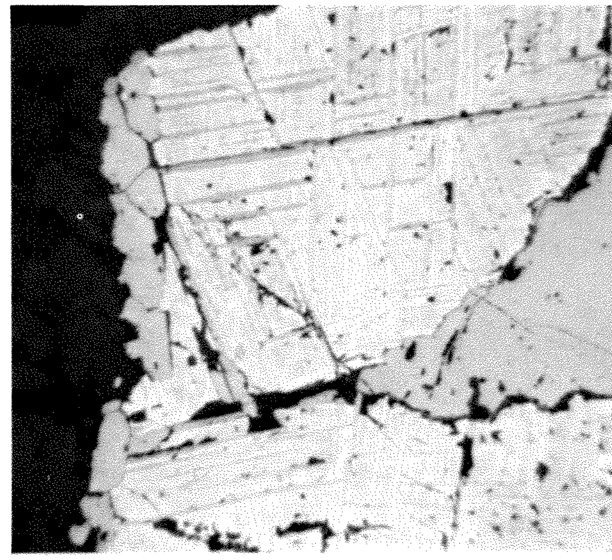


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Figure 14. **Oxide mineralization, D.D.H.1. "Army Reserve" prospect. Titanomagnetite shows well-developed exsolution lamellae of ilmenite. Coarse primary ilmenite is also present. (Sample 78157. Field of view 625 microns. Reflected light).**

TABLE 2. "JULIMAR COMPLEX": ANALYSES

G.S.W.A Sample	*Locality*	*Description*	*Cu (ppm)*	*Ni (ppm)*	*Cr_2O_3 (ppm)*	*Pt (ppm)*	*Pd (ppm)*	*Fe(a) (%)*	*TiO_2 (%)*	*V_2O_5 (%)*
78125	New Norcia prosp.	? Gossanous ultramafic	1830	2810	9000	0.06	0.24			
78153	New Norcia PDH	Gossanous serpentinite	3960	2850	3030	0.01	0.58			
78154	New Norcia PDH	Tremolitic metagabbro	260	120	130	<0.04	0.01			
78155	New Norcia PDH	Tremolite rock	940	390	2900	0.04	<0.006			
80781	DNN7—332 m	?Metapyroxenite	1310	470	150	<0.04	0.01			
80782	DNN7—309 m	Meta-leucogabbro	2600	980	230	<0.04	0.01			
80783	DNN7—292 m	Meta-melagabbro	1650	540	100	<0.04	0.01			
80784	DNN7—249 m	Metapyroxenite	4520	2910	740	0.06	0.02			
80785	DNN7—225 m	Metaperidotite cumulate	1450	2560	1360	0.11	0.37			
80787	DNN7—162 m	(Cumulate) metapyroxenite	4200	3720	1820	0.08	0.28			
80788	DNN7—152 m	Metapyroxenite	6980	4420	850	0.06	0.20			
80789	DNN7—130 m	Metapyroxenite (C.G.)	7680	2610	870	0.63	1.00			
78158	Army Reserve PDH	Anorthosite	160	40	25	<0.04	0.02	18.0	1.31	0.078
78159	Army Reserve PDH	Magnetite in leucogabbro	190	40	70	<0.04	<0.006	31.4	8.33	0.36
78161	Army Reserve PDH	Magnetite in leucogabbro	190	50	60	<0.04	<0.006	30.3	8.89	0.30
78162	Army Reserve PDH	Magnetite in leucogabbro	190	70	60	<0.04	<0.006	29.8	8.61	0.37
78163	Army Reserve PDH	Leucogabbro/mag gossanous?	300	120	80	<0.04	<0.006	37.3	2.68	0.37
78179	Coates Gabbro	Serpentinite	30	2390	1950	<0.04	0.01	8.44	<0.01	0.37
78180	Coates Gabbro	Serpentinized dunite	40	2390	8380	<0.04	<0.006	8.10	0.075	0.048

(a) Total iron

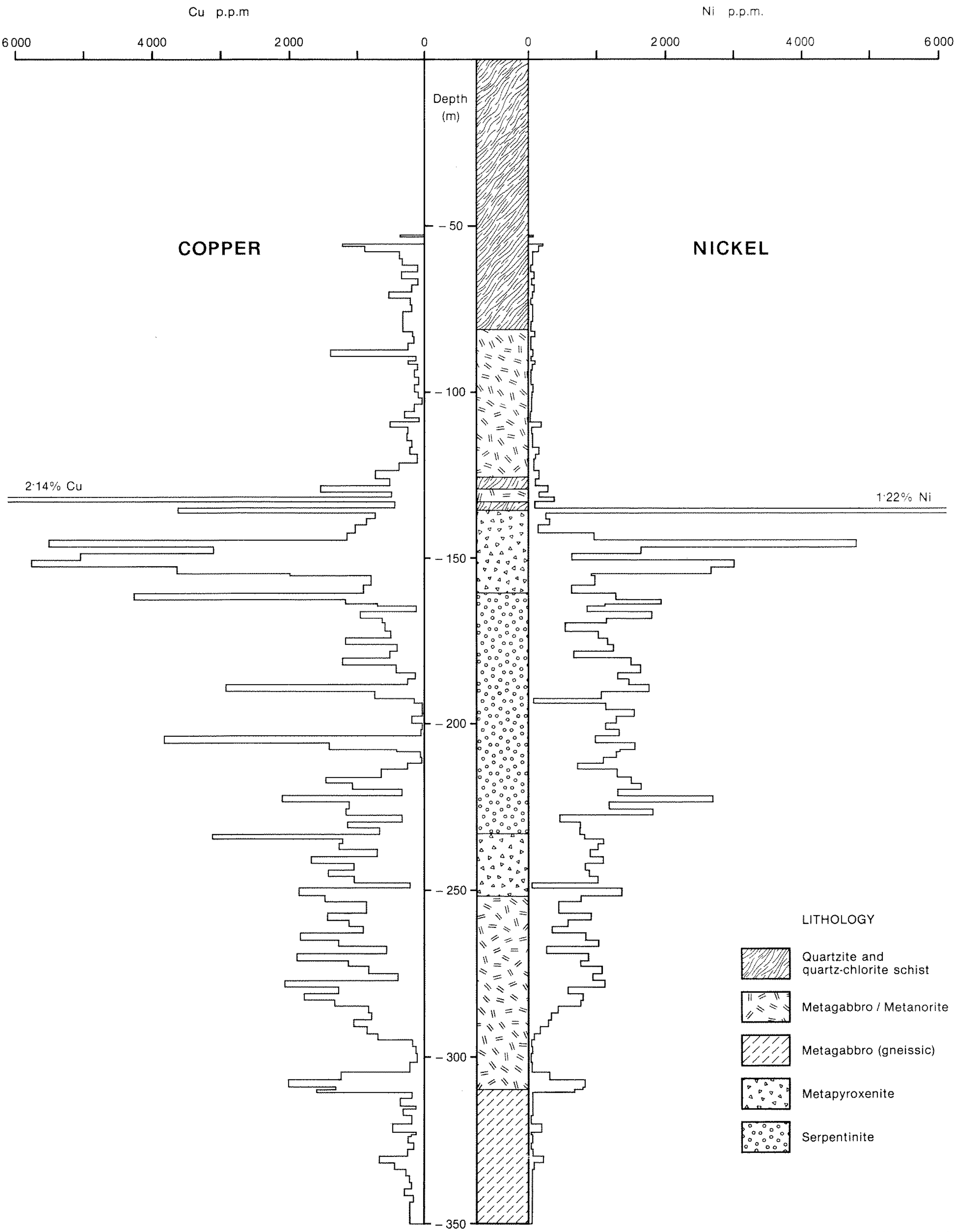
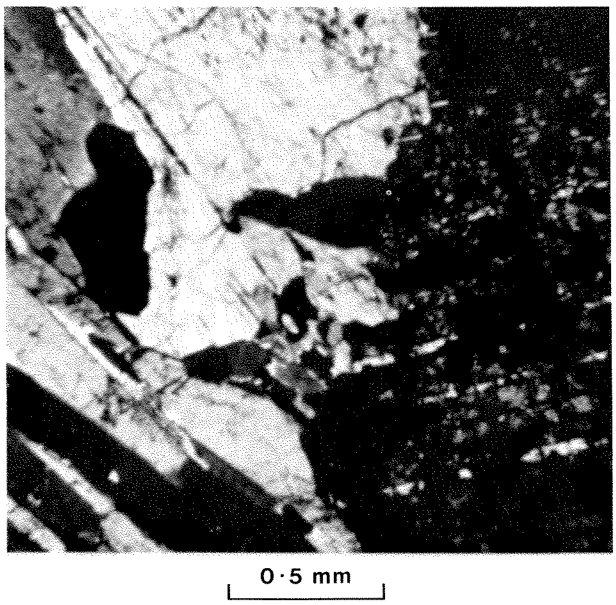


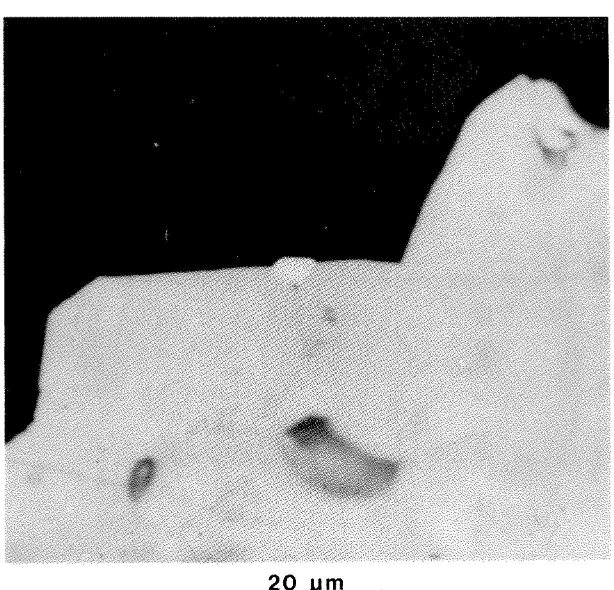
Figure 15. A typical drill section, New Norcia nickel prospect: D.D.H.7. Also showing analyses for copper and nickel.

Although this intrusion has been metamorphosed to amphibolite facies, many recognizable igneous textures are preserved in the samples examined. Figure 16, for example, shows coarse-grained metamorphosed norite. In this, a first-generation growth of tremolite replaces recognizable cumulate textures (after orthopyroxene), and a second-generation growth of granoblastic tremolite partly obscures these textures. Labradorite, although highly saussuritized, is still partly preserved. The sulphides (mainly pyrrhotite, subordinate chalcopyrite and pentlandite) show remnant intercumulus textures. There is minor remobilization of sulphides into fine hair-like cracks. Under high magnification (x 1 250) rare, small, (4 µm) very bright grains of sulphide within chalcopyrite were observed (Figure 17). These grains are considered to contain platinum-group minerals. This is borne out by the assays included in Table 2.



GSWA 21619

Figure 16. Metanorite from D.D.H.7. New Norcia nickel prospect. Plagioclase, of labradorite composition, has been extensively saussuritized. A "first-generation" growth of tremolite has replaced cumulate orthopyroxene, however, primary textures are partly obscured by a "second-generation" growth of nematoblastic metamorphic tremolite. (Sample 80733. Field of view 2.5 mm. Crossed polars).



GSWA 21620

Figure 17. Platinum-bearing sulphide mineralization, D.D.H.7. New Norcia nickel prospect. The small (4 micron diameter) bright grain on the margin of intercumlus textured chalcopyrite is a probable platinum-group mineral. (Sample 80789. Field of view 100 microns. Reflected light, oil immersion).

The evidence from the four areas discussed suggests that many of the north-trending magnetic anomalies in the western part of Figure 9 may be related to one, or several, layered basic/ultramafic intrusions. These are here informally named the "Julimar Complex".

Potential for mineralization

Work completed to date clearly demonstrates the potential of the "Julimar Complex" for vanadiferous, titaniferous magnetite and for nickel sulphides. The presence of chromite in some of the ultramafic samples indicates that chromitites may be developed in the sequence, while the apatite in the leucogabbros suggests some potential for phosphate deposits in the laterite profile and in residual deposits derived from these. The most interesting possibility, is for the development of economic deposits of platinum-group metals. The assays for platinum and palladium, from the New Norcia prospect (Table 2), are not in themselves economic. The presence of anomalous values over such a thick interval (at least 100 m), however, does indicate that the intrusion is derived from a magma which was highly anomalous in platinum-group metals. In the case of the New Norcia prospect, the widespread disseminated sulphides, which are present, result in platinum being disseminated through the intrusion. (This is perhaps a result of extensive sulphuration caused by contamination from the metasediment sequence which has been intruded.) If, however, some of the other bodies in the complex were derived from a similar magma, but have only a thin, concentrated zone of sulphide mineralization, such mineralization may contain economic grades of platinum-group metals.

Plots of Pt/(Pt+Pd) and Cu/(Cu+Ni) ratios of the New Norcia mineralization suggest that the parent magma had tholeiitic affinities and showed some similarities to flood-basalt related intrusions of the Norilsk type (Fig. 18). If the "Julimar Complex" intrusions formed in a similar tectonic setting to Norilsk (fundamental rifting close to the margins of a stable craton with magma derived from the immediately underlying mantle), this provides evidence of major fracturing in this very ancient Archaean crustal area.

Platinum and palladium analyses from the New Norcia prospect, recalculated to the value in 100% sulphides, and normalized with respect to their average abundance in chondritic meteorites (McBryde, 1972), are shown in Figure 18. Similarly normalized values are shown for the Merensky Reef and Norilsk deposits together with fields of values for the Sudbury nickel deposits and komatiitic nickel deposits, after Naldrett and others (1979). Values for certain ophiolitic chromitites, after Page and others (1982), are also plotted.

The four samples from New Norcia show a marked enrichment in platinum and palladium, relative to chondrite—and hence presumed mantle values—which is typical of deposits from layered intrusions. This is in marked contrast to komatiite-derived samples which normally show values close to chondrite values, and to samples from ophiolitic sequences which are characterized by a very strong depletion relative to chondrite values.

Sample 80789, which is the stratigraphically lowest sample from the New Norcia prospect and presumably contains the earliest precipitated sulphides, plots very close to values reported from the Norilsk deposit. Two samples from stratigraphically higher positions (80785 and 80788) show less enrichment relative to chondrite values and show some overlap with the field for the Sudbury nickel deposits. This "stratigraphic" variation suggests that the mineralization in the New Norcia prospect may have been derived by a series of successive segregations of sulphide. The more highly chalcophile elements, such as platinum and palladium, would partition strongly in the earliest segregated sulphides. This would result in the remaining magma being (relatively) depleted in these elements. The fact that this relative depletion is not uniform (sample 80787 in Figure 8 is more enriched than sample 80788, which is stratigraphically lower) may indicate that the chamber was receiving new pulses of magma during the formation of the sulphides.

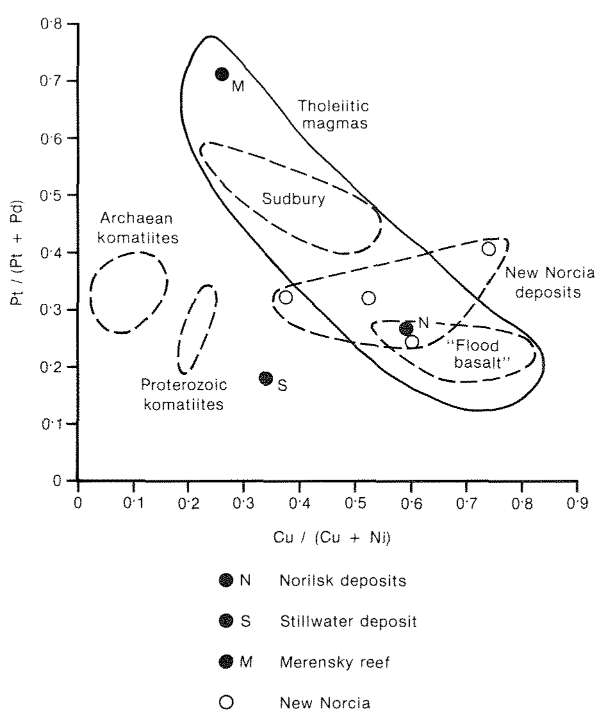


Figure 18. The relationship between the Pt/(Pt + Pd) and Cu/(Cu+Ni) ratios of mineralization from the New Norcia prospect compared with the fields for ores for various magma types, after Naldrett and Cabri (1976), and certain deposits with platinum-group metals, after Cabri (1981).

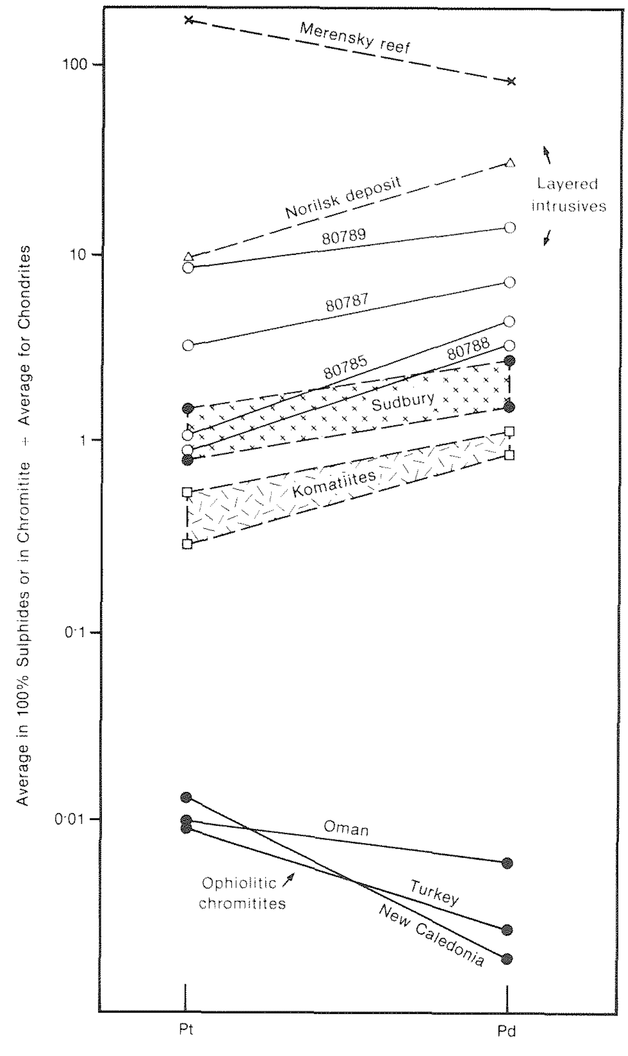


Figure 19. Chondrite-normalized platinum and palladium values, from the New Norcia prospect, compared with values for Norilsk deposit, the Merensky Reef, and certain ophiolitic chromitite deposits.

Each of the mineral prospects and many of the magnetic anomalies described from the Perth sheet are associated with areas, frequently hilly, of ferruginous laterite. The iron-rich composition of the basic intrusives may be the reason that particularly well-developed ferruginous layers have formed in the laterite profile. This, in turn, probably accounts for the relative lack of exposure of these rocks.

BYRO 1:250 000 SHEET

On the Byro 1:250 000 sheet, several areas of mafic granofels, basic granulite, and ultramafic rocks have been mapped. In spite of the medium- to high-grade metamorphism, primary igneous texture and gross igneous layering are preserved in several places.

These features, coupled with drill-core evidence from several localities, indicate that these rocks are derived from layered basic-ultramafic bodies, which included peridotite, lherzolite, norite, gabbro, anorthositic gabbro, and anorthosite (Williams and others, 1980). These remnants of early layered-basic complexes have been termed "Manfred Complex" by Williams and Myers (in prep).

were considered to be uneconomic because of the poor quality of the chromite and the limited tonnage (Williams and others, 1980). Drilling by Pacminex Pty Ltd near Taccabba Well, (26°05′27″S, 116°37′57″E), located a 1.5 km long magnetite-chromite band associated with meta-lherzolite in a concealed metamorphosed mafic-ultramafic body contained within granulite-facies terrain. A little pyrite and some chalcopyrite were associated with the chromite (Horsley, 1974). Exploration of ultramafic bodies on the sheet by various companies in the late 1960s and early 1970s located a number of possible nickel gossans. None of these proved to be of any significance.

Samples of chromitite collected on the Byro Sheet (held by the Government Chemical Laboratories in the "Simpson Collection" of minerals) have recently been analyzed for a variety of elements, including chromium, platinum, and palladium. These analyses (Table 3) returned low platinum and palladium values, all of which were below the detection limits of the method used. This suggests that the magma contained a low concentration of platinum-group elements. The layered intrusives of the Byro sheet, therefore, may have limited potential for platinum mineralization.

TABLE 3. ANALYSES OF CHROMITITES ON THE BYRO 1:250 000 GEOLOGICAL SHEET.

Sample number	Cr_2O_3 *(%)*	*FeO (%)*	*Cu (ppm)*	*Ni (ppm)*	*Pd (ppm)*	*Pt (ppm)*
MDC 838	30.5	12.1	82	850	<0.005	<0.04
MDC 1012	47.8	10.2	43	1000	<0.005	<0.04
MDC 1171	29.3	9.76	91	1400	—	—
MDC 2379	33.8	13.2	57	1300	<0.005	<0.04
MDC 2454	41.8	8.28	280	730	(a)<0.01	(a)<0.1
MDC 2607	33.8	18.1	41	1000	(a)<0.01	(a)<0.1
MDC 3712	46.3	9.14	147	680	(a)<0.01	(a)<0.1
MDC 4443	29.0	16.5	26	940	<0.005	<0.04

(a) Limits of detection reduced because of small sample size.

The metamorphosed basic bodies appear to pre-date orthogneisses in the Mount Narryer area which have given Sm-Nd model ages of 3.51 Ga (de Laeter and others, 1981).

A number of mineral occurrences within these layered-basic intrusive bodies have been examined by mining companies. Near Iniagi Well (26°11′53″S, 116°12′33″E), a chromite prospect has been explored by Electrolytic Zinc Company Australasia Limited and by Western Mining Corporation. Drilling by W.M.C. showed that chromitite lenses are associated with the ultramafic layers on the western side of the body and that iron-rich chromite layers are present in a metanorite higher in the sequence. The deposits

OTHER AREAS

There is evidence for the existence of remnants of layered basic-ultramafic intrusive bodies from several other parts of the Western Gneiss Terrain. Davidson (1968) interpreted mafic granulite from Quairading as having been derived from a layered-mafic intrusion. Morgan (1982) has described rhythmic layering in a metamorphosed layered harzburgite-lherzolite-anorthosite intrusion from West Bendering, near Kondinin. Up to 10% chromite has been recorded from the meta-lherzolite member of this body. Baxter and Harris (1980), in a report of drilling results from a copper prospect 25 km northeast of Mingenew, noted the presence of metagabbro rocks within the gneiss sequence adjacent to the Darling Fault. Also close to the Darling Fault, Drake (1976) showed unusual mineralogically and texturally zoned ultramafic bodies within banded gneiss, formed as a result of the boudinage of mafic-ultramafic igneous rocks during deformation. The deformation was accompanied by amphibolite-facies metamorphism followed by a static greenschist-facies metamorphism.

A metagabbro containing lenticular bands of vanadiferous titaniferous magnetite, largely obscured by laterite, occurs south of Tallanalla, 33°09'S 116°08'E (Baxter, 1978). Recent discoveries of titaniferous magnetite near Katanning and the discoveries of mafic granulites, which are believed to be derived from layered-basic intrusive rocks, during the mapping of the Dumbleyung 1:250 000 sheet (R. J. Chin, pers. comm.), suggest that other remnants of basic intrusions will be found.

CONCLUSION

Remnants of layered basic intrusions seem to be more common within the Western Gneiss Terrain than have previously been recognized. These rocks, once thought to be late-stage intrusions, are now considered to be amongst the oldest rocks of the Western Gneiss Terrain. Their thickness and composition have been responsible for protecting them from the more extreme effects of metamorphism; in many of the larger remnants, recognizable igneous textures are well preserved. The limited geochemical data suggest that the intrusives are derived from a magma of tholeiitic affinities, and they are considered to have potential for mineralization of the kind normally associated with layered intrusions of this type. Exploration which has been completed, thus far, has clearly demonstrated the potential for vanadiferous titaniferous magnetite, for chromite, and for nickel sulphides. Data presented in this paper suggest that some of the bodies may also have considerable potential for platinum-group metal mineralization.

ACKNOWLEDGEMENTS

The assistance of Otter Exploration N.L. and Alcoa of Australia Ltd in providing access to drill cores from their exploration projects on the Perth 1:250 000 sheet is gratefully acknowledged. Analyses for platinum-group metals were carried out by the Kalgoorlie Metallurgical Laboratory using a specially developed fire-assay method; other analyses were completed by the Government Chemical Laboratories.

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LIQUID-WASTE DISPOSAL IN PERTH
A HYDROGEOLOGICAL ASSESSMENT

by K-J. B. Hirschberg

ABSTRACT

Most of the Perth area is covered by a sedimentary sequence which must be considered unsuitable for liquid-waste disposal. The importance of hydrogeological investigations prior to commencement of disposal is stressed. Examples are given where the lack of such investigations had undesirable consequences, *e.g.* excessive hydraulic mounding, unknown direction of groundwater movement, and therefore misplacement of monitoring bores and contamination of production bores. A stricter application of the existing legal framework is considered necessary if a marked improvement is to be achieved.

INTRODUCTION

With the increasing use of groundwater in the Perth area over recent years, awareness has also been growing of the potential dangers of pollution of this resource.

Of all pollution sources, liquid wastes are potentially the most dangerous: the volume is often large; in many cases they are toxic; and their application in the form of shock loads often prevents sufficient adsorption by the soils, biodegradation, or dilution by the groundwater. The danger of groundwater pollution is particularly great in areas of sediments with low adsorption capacity, which is the case with large parts of the Perth region.

Liquid wastes can be broadly classified as follows:

domestic—septic systems
municipal—sewage, stormwater
industrial

The control and supervision of liquid-waste disposal, which has been in many ways less than satisfactory, is now slowly improving. The legal framework is presented by the *Health (Liquid Waste Disposal) Regulations* 1983, and by the *Rights in Water and Irrigation Act* 1914-1981.

For the Perth area, the administration of licensing of industrial effluents has been delegated to the Metropolitan Water Authority, which is assisted by multi-departmental panels. This paper will demonstrate the importance of a hydrogeological input to any groundwater-pollution investigation.

GEOLOGY

To understand the occurrence and behaviour of groundwater, a picture of the geological environment is required. In the context of groundwater pollution in the Perth region, only the superficial formations are of direct concern. They are shown on Figure 1, and on the geological cross-section on Figure 2, and are, from west to east:

Safety Bay Sand—bioclastic lime sand

Tamala Limestone—sandy, cavernous eolianite

Bassendean Sand—leached quartz sand

Guildford Formation—predominantly sandy clays

These formations are underlain by less permeable strata, *e.g.* the dark clays of the Osborne Formation, or the top clays of the Leederville Formation.

HYDROLOGY

The superficial formations contain unconfined groundwater. Recharge to this aquifer is mainly by direct infiltration of a proportion of the rainfall, and is greatest in clean and coarse sands. In the central regions of thick Bassendean Sand, two large groundwater mounds have formed, well known as the Gnangara and Jandakot Mounds, as shown in Figure 3. This figure also shows the positions of all past and present municipal liquid-waste disposal sites.

The water-table can be contoured by lines of equal elevation above datum. Groundwater flow is always perpendicular to the contours (from high elevations to lower ones) towards discharge zones, which can be either water courses or the coast. The rate of flow is dependent on the hydraulic conductivity of a particular formation, its porosity, and the hydraulic gradient. The former two are usually obtained from test pumping and other tests, whereas the gradient is calculated from a water-table contour plan as the drop of the water-table over a measured distance. Average flow rates in a medium sand are in the order of 10-20 m a^{-1}.

The above comments outline the importance of establishing detailed water-table contour plans to derive groundwater flow direction and flow rate, and therefore the expected extent of any pollution plume.

AMOUNTS OF LIQUID WASTE

The respective amounts of liquid waste have to be put into a proper perspective in order to assess their relative importance.

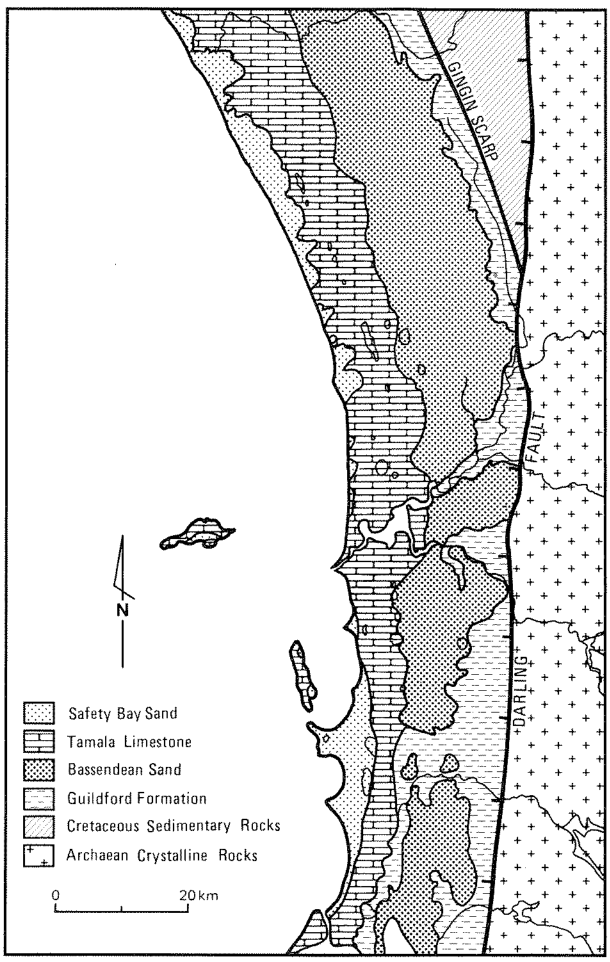

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Figure 1. Simplified geology, Perth area.

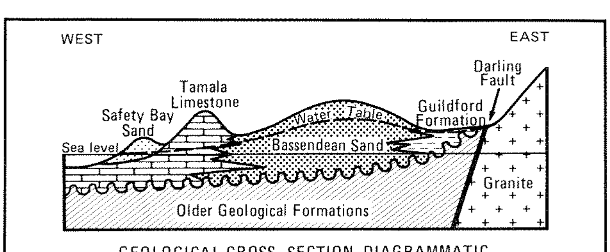

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Figure 2. Diagrammatic geological cross-section.

Figure 4 shows that of the 80 million m^3a^{-1} of total liquid waste for Perth, about half is discharged to sea, whereas the remainder is disposed into the ground. The majority of the latter is from septic systems and from industry, and only a minor amount, namely about 0.3%, is col-

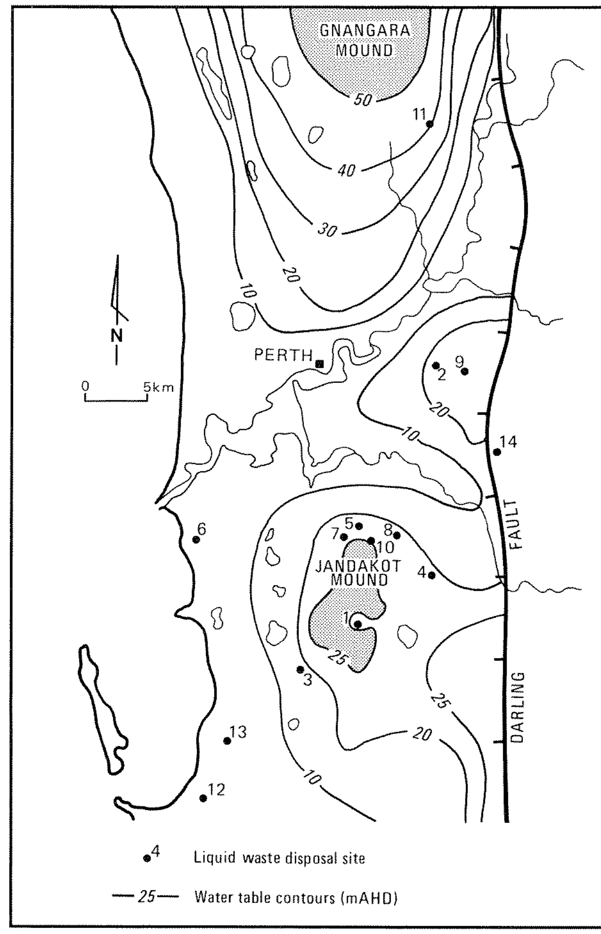

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Figure 3. Groundwater mounds and liquid-waste disposal sites.

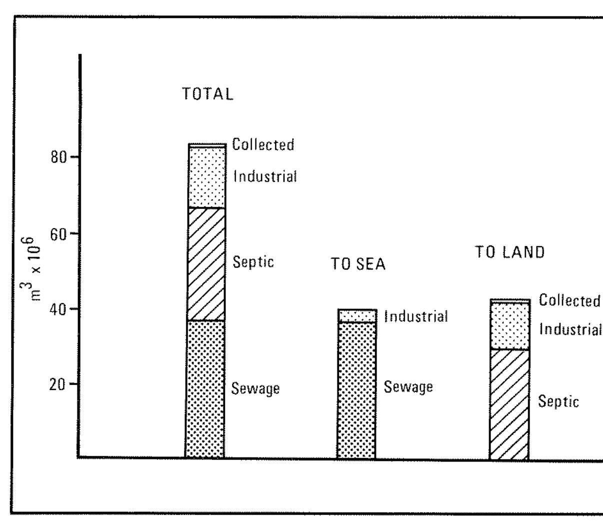

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Figure 4. Liquid waste, Perth area.

lected and disposed of on specific sites. Quite clearly, therefore, the major pollution threat comes from septic systems, and from industrial on-site disposal. The latter is further exacerbated by the often toxic nature of these industrial effluents.

SPECIFIC EXAMPLES

In the following examples of bad and good liquid-waste disposal practice, localities and authority or company names were, by necessity, largely omitted.

MUNICIPAL LIQUID-WASTE DISPOSAL SITES

Originally, such sites were put in areas convenient with respect to access and population, and scant regard was given to potential groundwater pollution. Figure 3 shows that a large number of these sites were located in the formerly undeveloped areas of leached Bassendean Sand, which is considered generally unsuitable for liquid-waste disposal. Futhermore, housing development has now encroached upon most of these areas, with a real danger of pollution of private bores or wells. Most of these sites have now been closed down. Three sites are at present used for biodegradable waste, and one site has been established for industrial liquid waste.

Disposal Site 4 (Fig. 5) was allowed to receive, over a large number of years, practically any amount of any type of waste. The local authority received a fixed annual payment by each cartage contractor. Control at the gate was minimal.

The water-table was originally about 1 m below the bottoms of the lagoons, which is, at best, considered insufficient for satisfactory adsorption and biodegradation. Problems were compounded by the disposal of substantial amounts of liquid waste, which resulted in a large groundwater mound, with the water-table actually on or near the surface, and a much increased flow of groundwater radially away from the site.

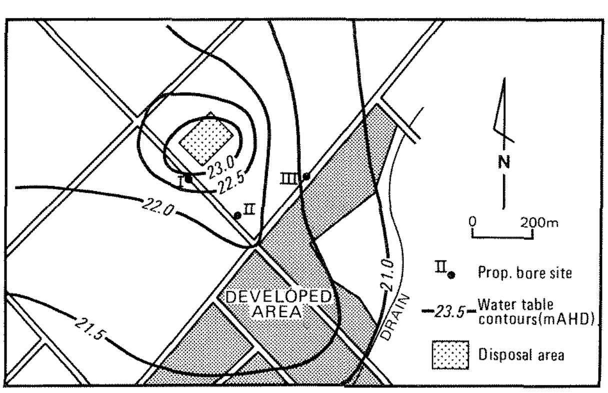

Figure 5. Disposal Site 4: Water-table contours and proposed monitor-bore sites.

Breakdown of pollutants is insufficient, and several private bores in a developing area to the southeast have been found to be contaminated. It is considered essential that monitoring bores be installed in this area, as the pollution plume will move with the groundwater for many years to come.

The Disposal Site 1 (Fig. 6) is on top of the Jandakot Mound, in a groundwater protection area. A production bore of the Metropolitan Water Authority is located about 200 m northeast of the site, and a system of monitoring bores has been installed between the disposal site and the production bore.

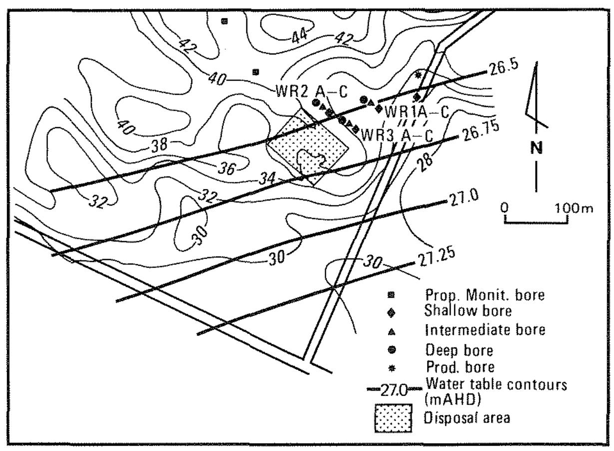

Figure 6. Disposal Site 1: Water-table contours and monitor-bore sites.

The hydrogeological investigation revealed that the undisturbed groundwater flow is to the northwest. However, an analysis of results indicates that pumping of the production bore creates a cone of depression, with induced groundwater flow from the disposal site to the bore. The monitoring bores show strong signs of pollution, and the production bore should in fact be rested until the pollution plume has dissipated. Additional monitoring to the northwest appears necessary.

The two new disposal sites installed during 1982 deserve mentioning as they show how a site assessment should, and can, be done. The Canning Vale site which is taking most of the carted septic waste since the closure of the majority of the old sites, was the subject of an intensive hydrogeological study including the drilling of a comprehensive network of monitoring bores. Figure 7 shows groundwater flow with a low gradient to the northwest, towards the sanitary landfill site and the old liquid-disposal site. The flow is therefore away from the groundwater protection area. Although it is generally towards a new housing development region, the danger of pollution of private water supplies is small as the rate of flow is slow, and adsorption and degradation of pollutants will be substantial due to the elevation of the lagoons well above the water-table.

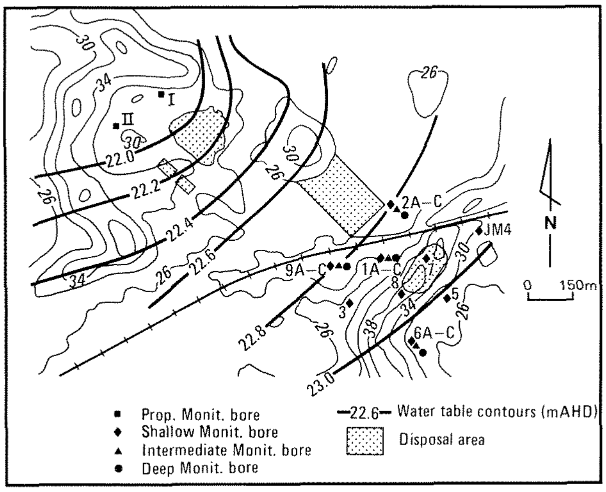

Figure 7. Disposal Site 10, Canning Vale: Water-table contours and monitor-bore sites.

It must be said that this site was established as an interim measure only, and does not meet all requirements for an ideal site. Any pollution originating here will eventually be superimposed on plumes which must be expected to exist already down-gradient of the landfill site and the old liquid-waste site. Monitoring bores are required between these two sites and the new housing area.

Figure 8 shows the new disposal site for industrial liquid waste, located within the Kelvin Road landfill site at Orange Grove. From a hydrogeological point of view, this site comes close to being ideal; it is situated just east of the Darling Fault, and lies on a wedge of clayey Leederville Formation, which is underlain by granitic bedrock at about 40 m depth. Downward leakage will be slow, with consequent good adsorption and degradation in the clayey soils. The water-table is well below the lagoons.

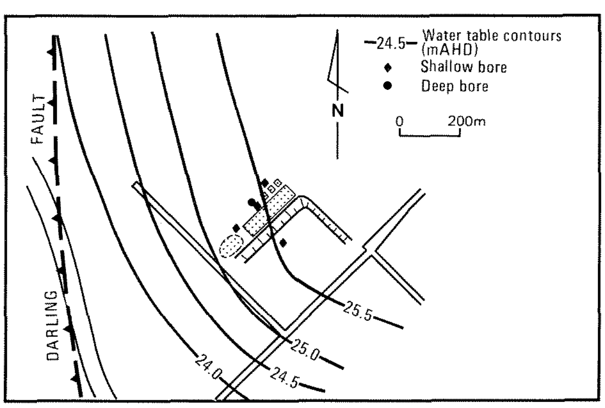

Figure 8. Disposal Site 14, Orange Grove: Water-table contours and monitor-bore sites.

The groundwater flow, as determined by a census of nearby bores, is to the southwest. The flow rate will be slow due to the expected low values for hydraulic conductivity. Rates of 1-2 m a^{-1} appear realistic. Groundwater quality is being monitored in several bores, and no pollution has been detected to date.

OTHER POTENTIAL SOURCES OF POLLUTION

Septic systems

Septic systems are still widely used in Perth, and the amount of liquid waste is substantial as was demonstrated earlier (Fig. 4). Practically all of this waste seeps into the ground and finally the groundwater. However, other studies have shown that the contamination generally appears to be confined to a small area because of strong adsorption and biodegradation in the unsaturated soil column.

Wastewater-treatment plants

Wastewater-treatment plants receive an enormous amount of liquid waste, however a substantial proportion of this is storm-water which results in a marked dilution of the effluent. Treatment reduces the contaminant strength further, and subsequently, 99% of the treated effluent is discharged to sea. The remaining small proportion is disposed of by spray-irrigation, and does not pose any significant threat to the groundwater.

Canning Vale recharge experiment

In the experiment, secondary effluent is disposed of via a system of soak lagoons, under controlled conditions (*i.e.* rotational application, wetting and drying cycles, and varying amounts of effluent). The experiment is being monitored extensively, and useful data are being obtained concerning the adsorption capacity of the soils and the fate of the various pollutants with time and distance.

The site is on Bassendean Sand, with generally poor adsorption capacity. The large amounts of effluent have resulted in a marked groundwater mound, and a pollution plume has been detected, migrating radially away from the site. The relative remoteness of the site prevents the pollution of production bores at present.

Large industries

The disposal of the substantial amounts of liquid waste, which originate from several large industries in the Perth area, is the subject of special agreements between the Government and the individual companies. The agreements generally stipulate certain disposal methods, site preparations, and monitoring requirements; companies usually adhere to the stipulated conditions. Leakages and spillages of some magnitude have occurred in the past despite stringent precautions. When monitoring has shown the presence of contamination, the companies concerned have gone to great lengths to rectify the situation.

The activities of these large industries are the subject of numerous individual reports and it is beyond the scope of this paper to go into any further detail.

Smaller industries

The disposal of liquid waste by numerous smaller industries constitutes a dark chapter in waste disposal in the Perth area. Businesses in this category are, to name just a few, plating shops, woolscourers, starch mills, chemical factories, and laboratories. Most of these do not have the means to install extensive treatment facilities or disposal systems, and disposal of effluent is usually into soak pits or bores within the premises. Although the amounts are generally only moderate, these effluents are a definate threat to the groundwater due to their often toxic nature, and the haphazard disposal practices. Furthermore, many of these companies are in well-developed areas.

Under the *Rights in Water and Irrigation Act*, companies which dispose of any effluent on their premises are now required to obtain a licence to do so. A licence usually sets maximum allowable levels for certain parameters, as well as methods of disposal, and monitoring requirements. Prior to the issue of a licence, a hydrogeological and hydrochemical investigation is done as a matter of course. The following examples show the importance of such investigations for the definition of a pollution plume and for the appropriate positioning of monitoring bores.

The first example is that of a corrosion-proofing business, called here company A, which had been dumping its effluent for several years into two soak pits within their plant site. The effluent contained high levels of cyanide, cadmium and zinc.

In order to show that groundwater leaving the premises was not polluted (requirement stipulated in the Act), the company drilled two monitoring bores. The first one was placed between the two disposal lagoons, the other near the boundary fence, downgradient of the expected groundwater flow. Analyses showed indeed that the first bore was heavily polluted, as expected, whereas the second bore showed no contamination.

When the licensing authority decided to expand the monitoring network, a detailed water-table contour plan was established (Fig. 9) which shows the surprising feature of a local flow direction to the southeast, in variance to the regional flow to the southwest. This means that the second monitoring bore could not show any pollution, as it was sited well outside the flow path of the pollution plume.

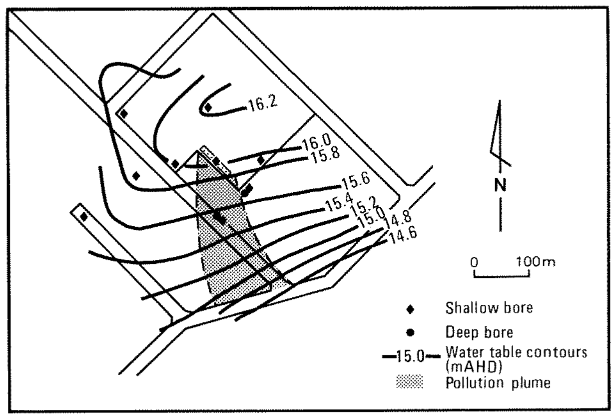

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Figure 9. Company A: Water-table contours and groundwater flow lines.

New bores were drilled subsequently directly downgradient of the lagoons, and they delineated a strong pollution plume moving in a southeasterly direction. In consequence, the company is now required to review and improve its disposal system to meet the conditions set out in the licence.

The second example is that of a large woolscouring business (company B) producing a considerable volume of effluent with high values for nitrogen, phosphorous, and BOD_5 (5-day biochemical oxygen demand). This disposal system consists of a number of lagoons, and is considerably larger than some of the municipal liquid-disposal sites.

The hydrogeological investigation resulted in the determination of a water-table contour plan which showed a groundwater mound, with increased flow rates towards some farmlets and towards a nearby lake. The investigation revealed also that there were already several monitoring bores nearby, drilled and sampled in 1973. Strong groundwater pollution was detected even then, but for some unknown reason, monitoring was discontinued.

Recommendations in this case are to resume monitoring of the existing bores, to improve the monitoring system with additional bores farther downgradient, and, if possible, to improve the quality of the effluent. The nearby lake is certainly in danger of eutrophication by the strong nutrient input, and some action is required in this respect.

The third example concerns a chemical factory (company C) which produces, among other things, herbicides. The effluent, which until recently was mixed with process cooling water, and then discharged into a soak pit, contains high amounts of phenols, 2,4-D, and 2,4,5-T.

The geological sequence in this area consists of Safety Bay Sand separated from the underlying sandy Tamala Limestone by a thin layer with high silt and clay content, which largely inhibits downward percolation.

The hydrogeological investigation revealed that the large amount of combined effluent and cooling water had produced a marked groundwater mound with elevations of up to 2 m above the regional water-table. This localized elevation of groundwater created greatly increased flow rates radially away from the site, until the regional gentle flow to the west prevails again, as shown on Figure 10. The existence of the mound requires a monitoring network which ideally covers the disposal area on all sides, and preferably at various distances.

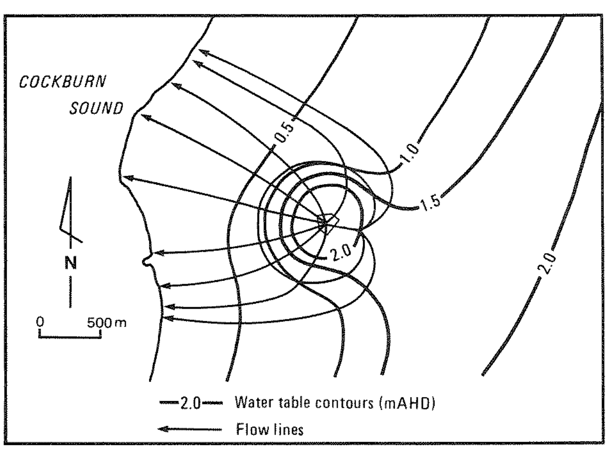

Figure 10. Company C: Water-table contours and groundwater flow lines.

A reasonably adequate bore network has now been completed, and the results of analysis are as expected—high to very high pollution levels in practically all bores. One bore, in fact, shows concentrations several times higher than the present neat effluent, indicating that the effluent was much stronger some years ago. Although the disposal system is no longer in use, monitoring will have to continue for years to come.

The results of the hydrogeological and hydrochemical investigations led to stringent measures being imposed on this company's effluent disposal.

GENERAL RESULTS OF INVESTIGATIONS

One of the aims of these hydrogeological and hydrochemical investigations has been to delineate the pollution plumes, and to predict their migration and decline in pollutant concentrations with time.

With a detailed water-table contour plan and the knowledge of the basic hydrogeological parameters, the direction and rate of flow can be ascertained with a fair degree of accuracy. This was done in the case of the last example, as it was thought important to know when the pollution plume would arrive at the discharge zone.

The calculations suggest a first arrival of the pollution plume at the discharge zone, along the shortest flow lines, in about 70 years, and progressively later along the longer flow paths. The expected maximum width of the pollution plume is also marked on Figure 10.

CONCLUSIONS

Although the few examples given above are by no means comprehensive, they help to show that liquid-waste disposal practice in the Perth area cannot be considered to have been very satisfactory. Disposal at municipal sites has greatly improved in the recent past; however, the disposal of industrial liquid waste still leaves a lot to be desired.

The Bassendean Sand and the Tamala Limestone must be considered unsuitable for safe disposal of liquid waste. This is because the retention times in the clean Bassendean Sand, and in the cavernous Tamala Limestone, are often too short for good biological degradation in the unsaturated zone above the water-table. In addition, the adsorptive capacities of both formations are generally too low for sufficient removal of chemical pollutants. Unfortunately, these formations cover a large part of the Perth region, and relocation of all industries concerned will not be possible. This means that improved treatment and disposal methods are the only means of reducing the pollution threat to our groundwater. First steps are being taken, as outlined above, by the requirement of licensing with more stringent licence conditions and monitoring requirements.

Hydrogeological and hydrochemical investigations are being done nowadays as a matter of course before a licence is issued, and their importance in this context has been demonstrated. Perth is lucky enough to have large usable groundwater resources, and it is certainly worth every effort to maintain this resource for the future. The legal framework exists already, however, stricter application and a more determined approval appear desirable.

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OCCURRENCE, DISTRIBUTION, AND ORIGIN OF SMITHSONITE IN THE No. 2 LEAD-ZINC DEPOSIT AT NARLARLA, WESTERN AUSTRALIA

by C. R. Ringrose

ABSTRACT

The No. 1 and No. 2 deposits at Narlarla are carbonate-hosted, epigenetic, lead-zinc deposits of post-Devonian age and are broadly analogous to ores of the Mississippi Valley-type.

Unmined remnants of the larger No. 2 deposit include gossanous secondary ore, and small pods and lenses (to 2 m x 1 m) and distinct beds (to 0.15 m wide) of high-grade, primary sulphide ore.

Smithsonite occurs in intimate association with unoxidized sulphide-bearing ore, and is abundant in the secondary, oxidized orebody overlying and adjoining this zone.

Smithsonite occurs as petal-shaped euhedra (to 2 mm) filling the interstices of sphalerite-galena dendrites, and as rounded and corroded ('sanded') grains (to 1 mm in diameter) with chamosite as a wall-rock alteration zone around the sulphide ore. Smithsonite's most characteristic form in the secondary orebody is as coarse, translucent, botryoidal masses up to 2 m across.

Primary versus secondary modes of origin for the smithsonite associated with unoxidized sulphide-bearing ore are presented. Primary smithsonite is rare and is atypical of Mississippi Valley-type deposits in general. The formation of smithsonite as a primary mineral requires very high concentrations of zinc in solution. A secondary mode of origin for smithsonite during supergene alteration of lead-zinc sulphide ore is favoured.

INTRODUCTION

Two carbonate-hosted lead-zinc deposits, the No. 1 and the No. 2, occur at Narlarla approximately 150 km east of Derby in the West Kimberley Mineral Field. Prior to mining, each deposit had a gossanous secondary orebody consisting mainly of cerussite, smithsonite, hydrozincite, and limonite, which had developed upon a sphalerite- and galena-rich primary orebody.

Only the larger, No. 2, deposit has been mined extensively (between 1947 and 1966), and a large, elongate water-filled pit now marks the site. In 1983, dewatering of the pit was carried out by the Geological Survey of Western Australia, and several unmined remnants of primary sulphide-bearing ore were exposed.

Detailed petrological and mineralogical examination of sulphide samples (including X-ray diffraction studies) revealed that smithsonite and a little cerussite are (intimately) mixed with these unoxidized sulphide occurrences.

The mineralogy, texture, and distribution of smithsonite from the No. 2 deposit is discussed, and two possible modes of its origin are presented.

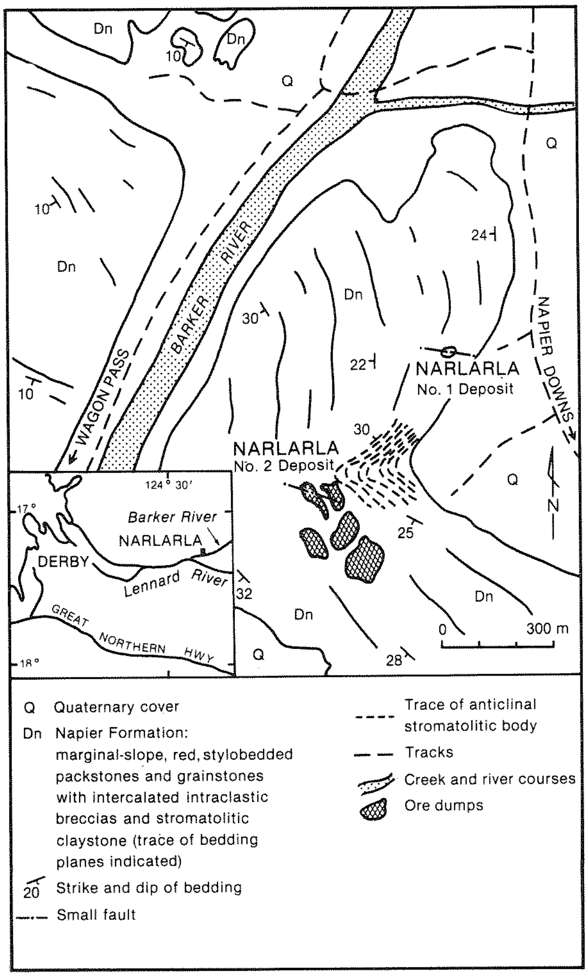
GEOLOGICAL SETTING

The deposits at Narlarla occur within Devonian reef complexes in the northwestern portion of the Lennard Shelf (Fig. 1). The host sediments, the Napier Formation, rest directly upon Precambrian basement and dip at 15° to 30° to the southwest. They consist mainly of partially dolomitized, red-brown, stylobedded packstones and grainstones, but include units of pink stromatolitic boundstone and intraclastic breccia. Collectively these sediments constitute a distal fore-reef facies.

Both deposits are just below the modern erosion surface which is approximately coincident with the unconformity at the base of the Late Carboniferous Grant Group. The No. 2 deposit (Fig. 2) lies at the nose of a large, depositional stromatolitic anticline which, together with a set of small east-trending faults, is believed to have localized ore fluids (Ringrose, 1984).

SECONDARY OREBODY

A large remnant of the secondary orebody of the No. 2 deposit outcrops on the northern corner of the pit (Fig. 2). This zone is intensely and pervasively iron-stained; it contains small patches (to 2 m x 1 m) of white or pink hydrozincite-rich ore and large calcite-lined vughs (to 0.3 m across)



towards its upper surface. Minerals in this remnant comprise smithsonite (to 40%), cerussite (to 15%), iron oxides (to 25%), and variable minor amounts of accessory hydrozincite (locally abundant), chamosite, quartz, calcite, azurite, and malachite. Chamosite, as used herein, is synonymous with "berthierine", a serpentine-like mineral of the septechlorite group.

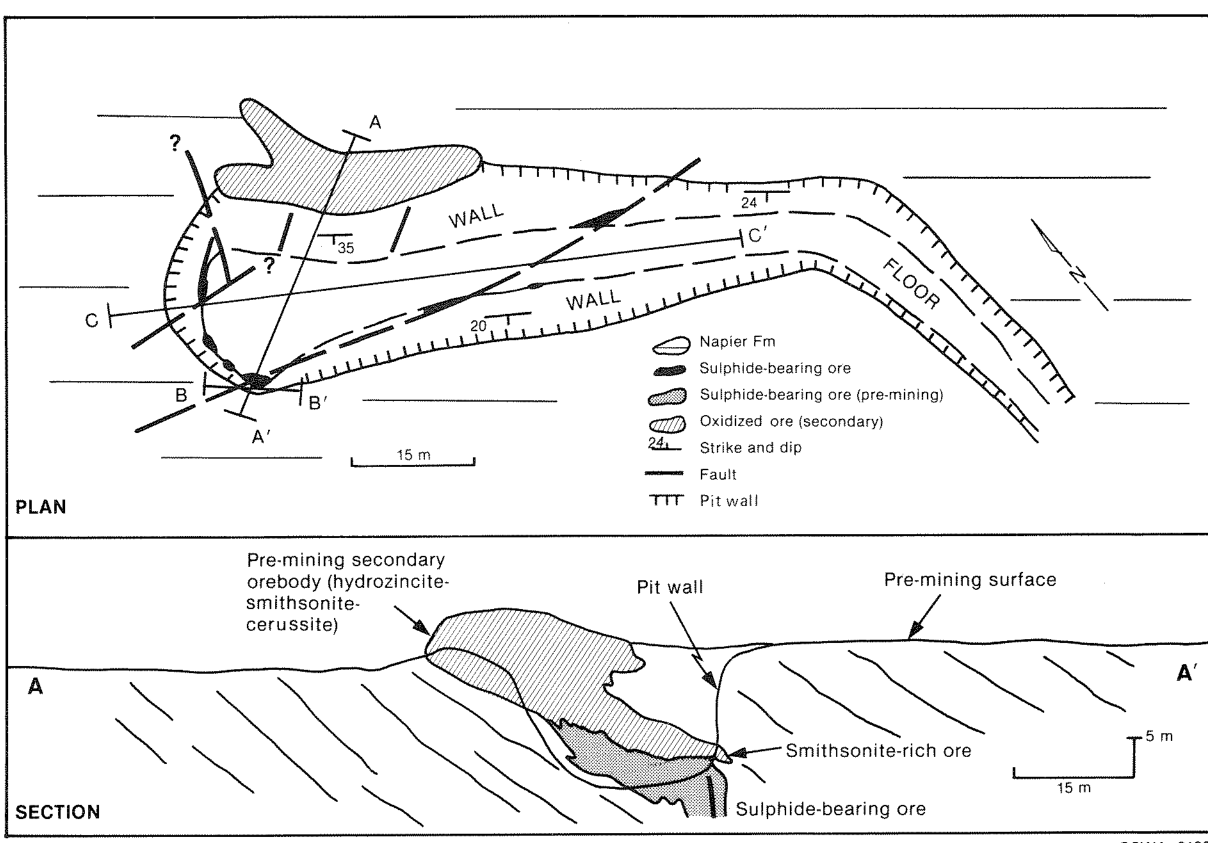
A second remnant, consisting almost entirely of iron oxides and botryoidal smithsonite, occurs down-dip from the larger remnant within the southwestern wall of the pit.

These remnants of the secondary orebody overlie apparently unoxidized sulphide along an irregular, approximately horizontal contact, which coincides with a change in the wall-rock discolouration from red-brown to green with increasing depth.

Secondary ore has resulted from the supergene alteration of sulphides. The observed zonation of the secondary ore is from cerussite-smithsonite-hydrozincite-iron oxides (up-dip) to smithsonite-iron oxides (down-dip) (Fig. 2). This is generally consistent with the description given by Sangameshwar and Barnes (1983) for the oxidation of similar carbonate-hosted lead-zinc-silver sulphide orebodies.

◄ **Figure 1. Locality map and general geology of Narlarla.**

▼ **Figure 2. Geology of the No. 2 deposit at Narlarla.**



SULPHIDE-BEARING ORE

Most unmined remnants of the primary, sulphide-bearing ore exposed in the dewatered pit occur towards the northwestern end (Fig. 2). These epigenetic pods, lenses, and beds are very high grade, and consist of 70-90% sulphides, minor amounts of interstitial gangue, and some primary porosity. The ores are composed of sphalerite (to 60%), galena (to 30%) and minor, but locally significant, amounts of iron sulphides (to 10%).

Skeletal-botryoidal textures are dominant in these occurrences, and are considered to have formed by the very rapid precipitation of ore minerals in regions of high porosity (Ringrose, 1984). These textures are characteristic of Mississippi Valley-type ore deposits in general (Anderson, 1975), and are indicative of a high degree of supersaturation of the ore fluid. The required high levels of supersaturation and ore precipitation probably resulted from a fluid-mixing process (Ringrose, 1984). The fluids involved in the ore formation are considered to have been basin-derived, metal-bearing brines that mixed with meteoric fluids (possibly of pre-Grant Group age).

SULPHIDE-SMITHSONITE ZONE

Adjacent to the remnants of sulphide-bearing ore, the partially dolomitized limestone wall rocks have been converted: an altered zone of grey-green, friable, and incompetent rock, which consists of "sanded" smithsonite grains (to 50%), siliciclastic detritus (to 30%), and chamosite (to 20%), occurs up to a distance of 0.5m.

This grey-green, smithsonite-chamosite alteration aureole passes outwards, gradationally, through zones of iron-bearing smithsonite rock (up to 2 m in width and containing about 40-50 wt% Zn0) and ankeritic (or iron-bearing) dolomite, into unaltered dolomitic limestone host rock within 5 to 10 m (Fig. 3).

In the inner smithsonite-chamosite alteration zone, smithsonite is mainly in the form of small, rounded to subequant, (hexagonal or elongate) dark-rimmed grains (to 1 mm across) in a friable aggregate (Fig. 4). The zone of smithsonite rock consists of interlocking to subeuhedral grains (to 2 mm) that include a proportion of siliciclastic detritus (Fig. 5). Euhedral crystals of hexagonal cross-section or elongate, petal-shaped form occur less commonly throughout these facies.

Bedded and skeletal-botryoidal samples of sulphide-bearing ore also include crystals of smithsonite in pore spaces. Smithsonite is translucent bluish-grey, and fine-grained (to 1 mm in length). It forms aggregates of pyramidal crystals that line vughs between dendrites and botryoidal masses of sphalerite and galena (Fig. 6). Hexagonal and petal-shaped euhedra are commonly observed in thin section.

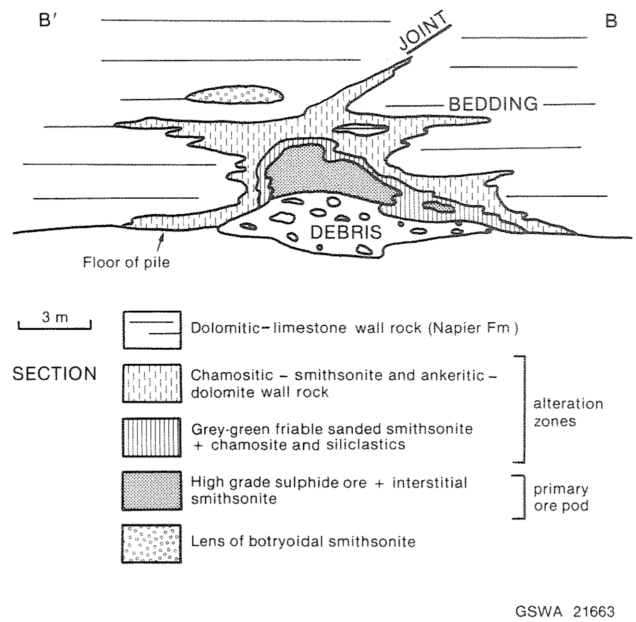


Figure 3. Sketch cross-section illustrating the distribution of alteration zones around a remnant of primary sulphide ore.

Smithsonite grains are colourless to cloudy, and possess weak to moderate cleavage, high relief (varying upon rotation), extreme birefringence, and symmetrical extinction. Most crystals are finely zoned.

PARAGENESIS OF SMITHSONITE

The paragenetic position of smithsonite, which surrounds and is interstitial to unoxidized sulphide, is problematic. Smithsonite in this zone may have formed locally by metasomatic replacement of dolomitic limestone, and elsewhere by precipitation at about the same time as the sulphides. Alternatively, the smithsonite surrounding unoxidized sulphide may be the lowermost portion of the secondary orebody.

A primary origin for smithsonite is suggested by the following features:

(a) the smithsonite-chamosite wall-rock alteration zone, which surrounds sulphide occurrences, decreases in intensity outwards;

(b) locally, smithsonite-chamosite pinches out upwards along faults and fractures;

(c) cross-cutting relationships between smithsonite-chamosite and sulphide are rare; and

(d) sulphide pods and beds, which are surrounded by friable and incompetent smithsonite-chamosite wall rock, show limited signs of fragmentation or disruption. This suggests that wall-rock alteration is contemporaneous with sulphide precipitation.

A secondary origin for the smithsonite, surrounding unoxidized sulphide, is suggested by its proximity to the oxidized secondary orebody which contains abundant smithsonite throughout. Further, the textural evidence indicates that smithsonite present between sulphide dendrites is the last-formed mineral of the paragenesis, and clearly post-dates sulphide precipitation.

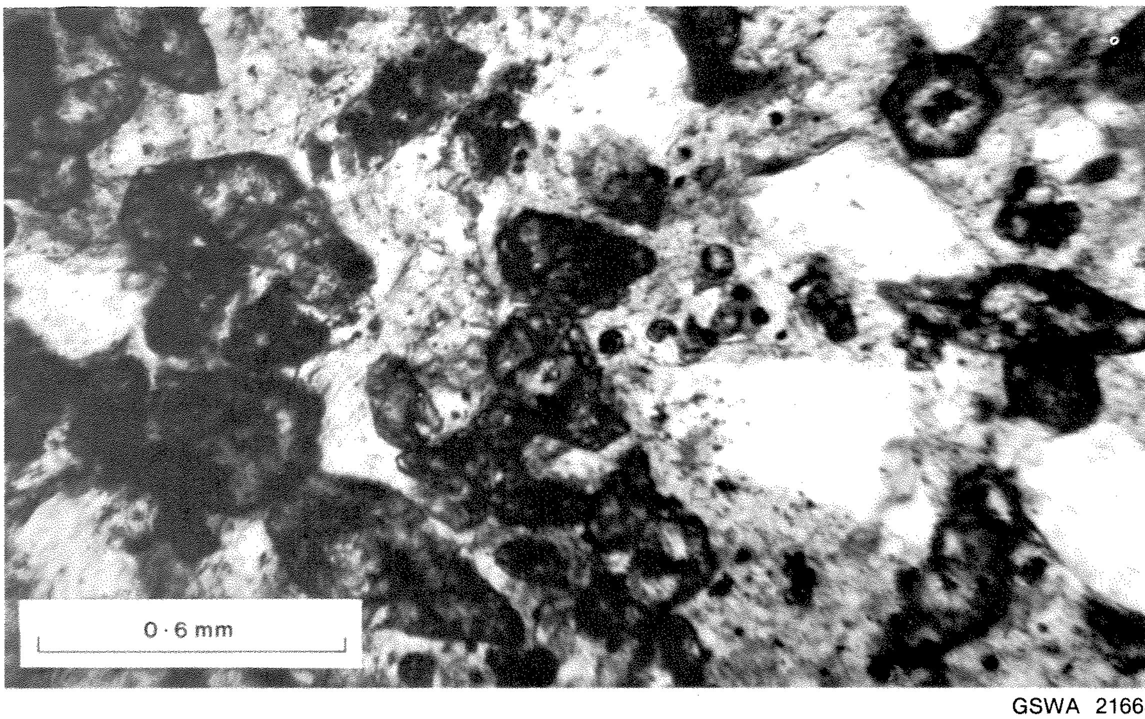


Figure 4. Rounded and corroded ('sanded'), hexagonal and elongate dark grains of smithsonite with groundmass of chamosite in a friable aggregate.

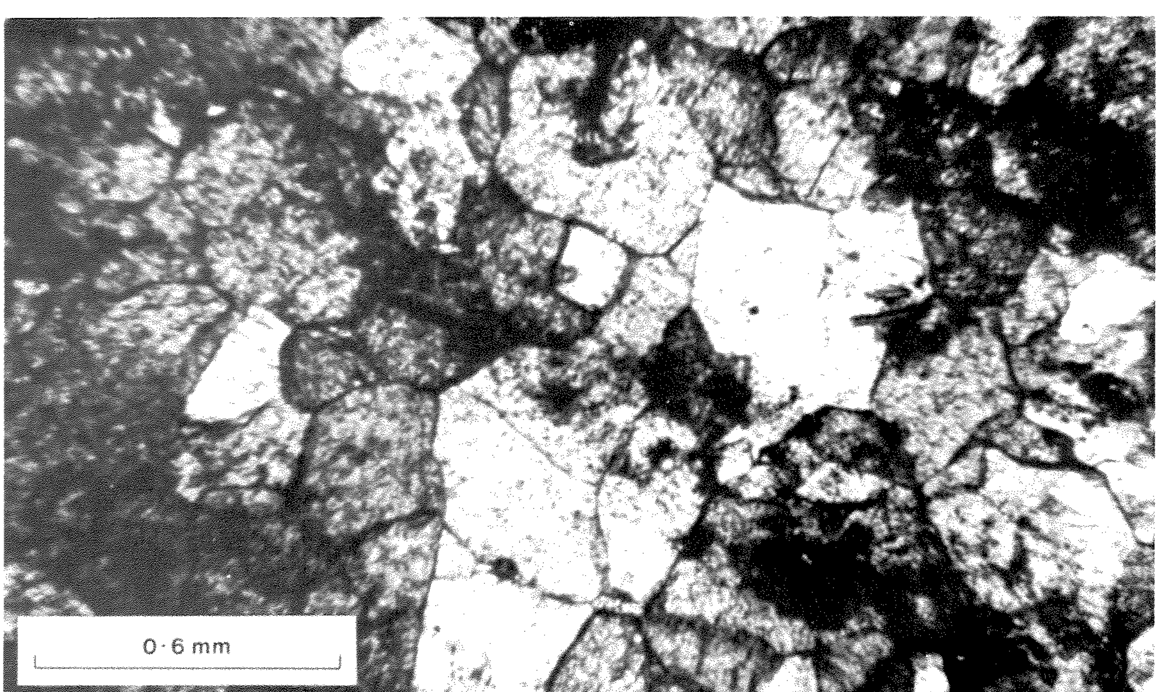


Figure 5. Interlocking anhedral grains of smithsonite which often include grains of biotite, feldspar, and quartz.

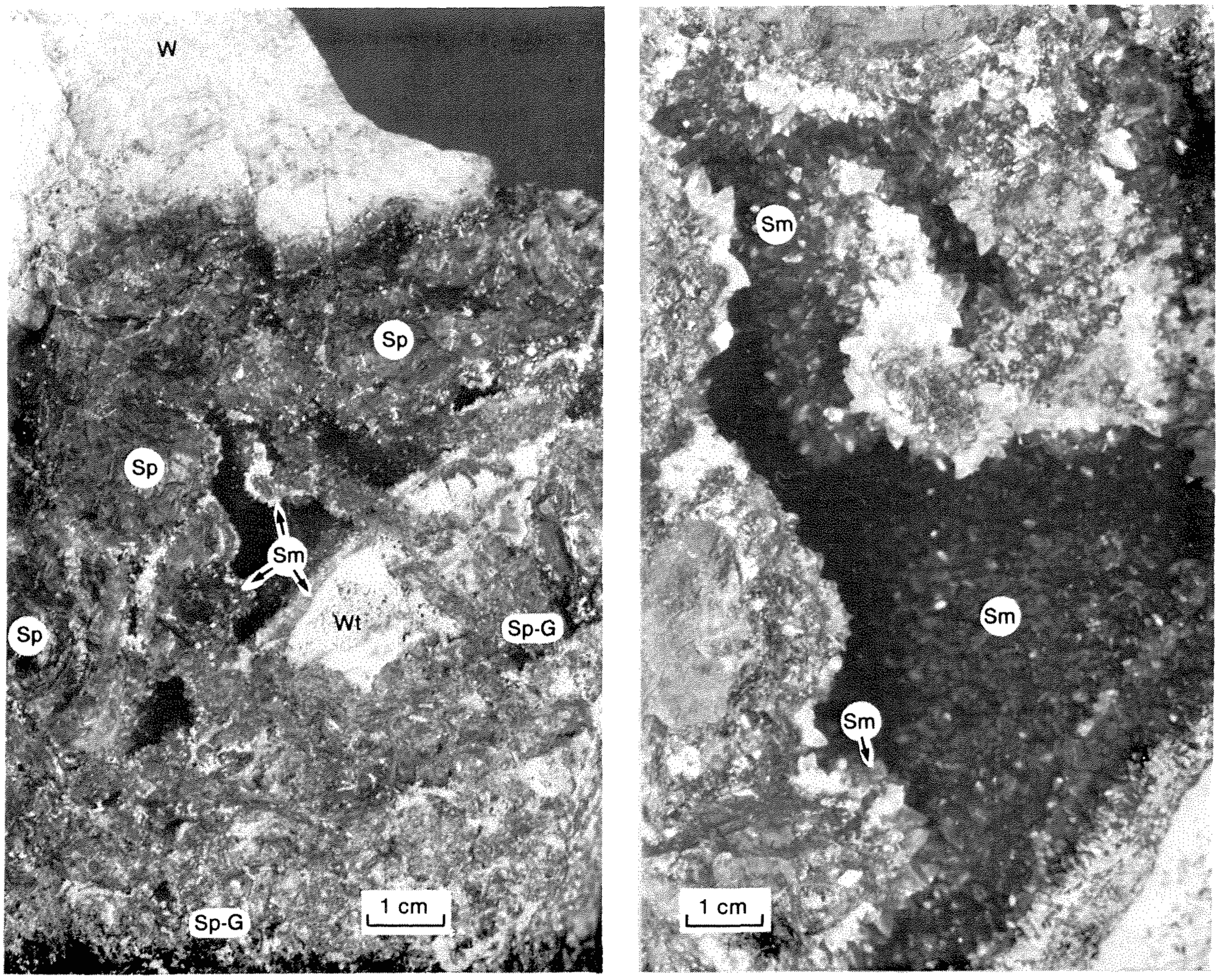


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Figure 6. Photographs showing details of a high-grade sulphide (sphalerite and galena) ore sample illustrating in A—spatial relationship of smithsonite (lining vughs) to unoxidized sulphide, and in B—smithsonite morphology. ('Sp' = sphalerite; 'Sp-G' = intergrown sphalerite and galena; 'W' = smithsonite-chamosite wall rock, Fig. 4; 'Wt' = wall rock fragment; and 'Sm' = smithsonite.)

FORMATION OF SMITHSONITE FROM ORE-FORMING SOLUTIONS

A prime candidate for the ore-forming solution at Narlarla, in view of the overall similarity of these orebodies to those of the Mississippi Valley-type (Ringrose, 1984), is a warm, saline, compaction brine (Jackson and Beales, 1967; Gustafson and Williams, 1981).

The formation of smithsonite from such a fluid within a carbonate host rock may be simulated by adding $ZnCl_2$ to a 3*M* NaCl solution saturated with smithsonite and calcite. This simulation can be used to calculate the total zinc concentration (as opposed to Zn^{2+}) required to replace calcite with smithsonite.

According to PATH, a computer program by Brown and Perkins (University of British Columbia) that simulated this reaction, over 23 000 ppm zinc is required to effect replacement of calcite or to precipitate primary smithsonite in the presence of calcite (G. Anderson, pers. comm) (assuming the following probable conditions of ore formation: pH 7, f CO_2 = 10^{-1}, and 100°C). Although the pH and temperature conditions at Narlarla may have been slightly different, a higher temperature or a lower pH would have only a limited effect on the total zinc content required to produce smithsonite by replacement of calcite.

Although such high concentrations of zinc in ore fluids (i.e. about 2.3% zinc) seem unlikely, it is notable that considerable controversy still exists concerning the metal content of ore-forming solutions. Very few reliable analyses of the metal content in fluid inclusions have been made (Roedder, 1972). Analyses of fluid inclusions from fluorite in the Cave-in-Rock deposits in Illinois (Czmanske and others, 1963) included values as high as 10 000 ppm zinc, clearly indicating that metal levels in the order of 20 000 ppm in ore fluids are possible.

Analyses of oil-field brines, which are analagous to Mississippi Valley-type ore-forming solutions (in terms of major-element composition, salinity, deuterium/hydrogen and $^{18}O/^{16}O$ ratios, and temperture; Sverjensky, 1984) indicate levels of heavy metals generally less than 40 ppm. For example, the Pleasant Bayou Field in Texas (Kharaka and others, 1980) and the Rayleigh Field in Central Mississippi (Carpenter and others, 1974) give lead levels of 1.1 ppm and 17 ppm and zinc levels of 1.5 ppm and 33 ppm respectively. These data clearly imply that the smithsonite-forming fluid at Narlarla, which may have been the ore-forming fluid, was not analagous to a typical oilfield brine.

FORMATION OF SMITHSONITE FROM GROUNDWATERS

In many occurrences, smithsonite has a secondary origin and is formed by the *in situ* oxidation of primary zinc sulphide and by precipitation from zinc-bearing groundwaters in the vicinity of lead-zinc orebodies during their weathering.

The formation and distribution of zinc and lead minerals in the oxidized zone of a lead-zinc sulphide orebody is dependent upon:

(a) the solubilities of their compounds, which set the limits for the relative amounts of zinc and lead transported out of the ore zone in solution;

(b) the relative amounts of these metals present in the sulphide ore;

(c) the physical and chemical nature of the host rock; and, in particular,

(d) the iron and copper content of the ore (Sangameshwar and Barnes, 1983).

At Narlarla, the host rock is considered to be fairly homogeneous in terms of physical and chemical properties. The sulphide ore has approximately equal amounts of lead and zinc; small amounts of iron and copper also occur. The principal control, therefore, upon the distribution of lead and zinc is the solubility of their compounds. Zinc sulphate is much more soluble than lead sulphate, and zinc carbonate is slightly more soluble than lead carbonate. This has resulted in a zonation of these compounds at Narlarla, with smithsonite forming a distal aureole to cerussite in the oxidized orebody.

The precipitation of zinc as smithsonite from groundwaters passing through and downwards within a sulphide orebody is favoured by: decreasing redox potential; an increase in the concentration of zinc; and a decrease in groundwater acidity.

A suitable decrease in redox potential may occur at or near the water-table. Exposures in the dewatered pit indicate that the zone of red-brown, oxidized secondary ore has a very irregular and approximately horizontal contact with the underlying, grey-green, sulphide-chlorite-smithsonite facies. This contact may be coincident with the top of the fluctuating water-table.

The required reduction in acidity may be brought about by reaction of acid, zinc-bearing groundwaters with calcium carbonate thus:

$$2H^{+}_{(aq)} + Zn^{2+}_{(aq)} + 2(Ca,Mg)CO_{3(s)} \longrightarrow ZnCO_{3(s)} + 2(Ca,Mg)^{2+}_{(aq)} + CO_{2(g)}$$

DISCUSSION

The formation of smithsonite in the primary, sulphide-forming environment would require exceedingly high levels of zinc to be present, in the absence of reduced sulphur, during ore-fluid activity. Specifically, if the ore fluid was similar to a basinal brine (in terms of acidity, salinity and temperature), over 23 000 ppm zinc would be required for smithsonite formation.

Such high levels of zinc in basinal brines are highly unlikely, given the known data from the analysis of oilfield brines. This would account for the absence of primary smithsonite in Mississippi Valley-type deposits. However, limited analyses (Czmanske and others, 1963) indicate that such high zinc levels in ore-forming solutions are possible. A primary origin for smithsonite at Narlarla implies, therefore, that the ore fluid was not analagous to a typical basinal brine.

It is conceivable that the Narlarla ore fluid was a basinal brine, containing a large proportion of fluid from a deep-seated ?magmatic source, and that it carried sufficient metals in solution to form primary smithsonite. Alternatively, fluid-mixing processes at the ore site, suggested by the evidence of ore textures (Ringrose, 1984), may have created local concentrations of zinc in solution by dissolution of early-formed sulphide. However, neither of these possibilities is considered likely. There is no isotopic or fluid-inclusion evidence (Etminan and others, 1984) to indicate the activity of an ore fluid with a magmatic-fluid component. Further, the remnant sulphide-bearing ore pods and lenses show no evidence of a large-scale dissolution during growth.

Although the textural evidence remains equivocal, it is considered more likely that the sulphide-smithsonite facies is secondary, and represents the boundary zone between truly unaltered primary ore below, and oxidized ore above. This boundary

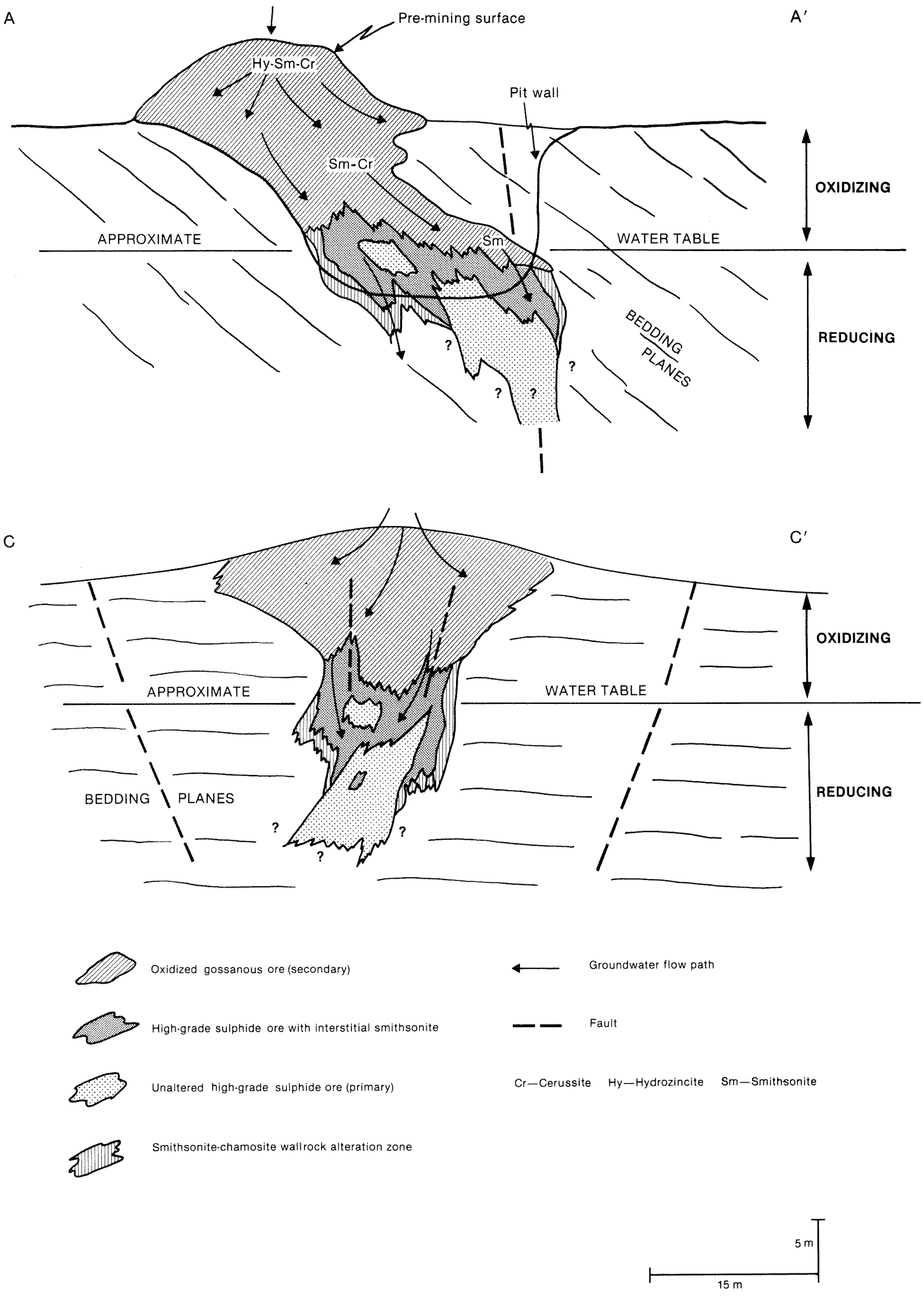

Figure 7. Cross sections (marked on Fig. 2) illustrating the distribution of primary- and secondary-ore, and alteration zones. Groundwater flow paths and the approximate level of the water table during oxidation of ore are indicated.

within the secondary ore body is coincident with the top of the recent, fluctuating water-table (Fig. 7).

Zinc-bearing groundwaters, percolating downwards through the sulphide orebody, precipitated smithsonite when they became sufficiently reduced or by reaction with carbonate wall rock (equations 1 and 2, and Fig. 7). "Sanded" smithsonite and smithsonite rock were formed by the replacement of carbonate wall rock; whereas euhedral smithsonite, between sulphide dendrites and botryoids close to the wall-rock contact, was precipitated from solution in open spaces. The redox potential of these groundwaters was also sufficiently low during smithsonite precipitation to preserve sulphides in an unoxidized state.

The manner in which smithsonite alteration surrounds sulphide pods may be explained by the percolation of groundwaters through high-grade sulphide and out into wall rock in a manner illustrated diagrammatically in Figure 7.

Primary wall-rock alteration most probably consisted of chamositic and ankeritic dolomite, upon which secondary smithsonite replacement is superimposed. Some physical resedimentation of this chamosite has taken place during groundwater activity, creating chamositic-rich infillings to skeletal-botryoidal sulphide porosity.

ACKNOWLEDGEMENTS

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THE SEDIMENTOLOGY OF A TIDE-INFLUENCED FAN-DELTA SYSTEM IN THE EARLY PROTEROZOIC WYLOO GROUP ON THE SOUTHERN MARGIN OF THE PILBARA CRATON, WESTERN AUSTRALIA

by A. M. Thorne and D. B. Seymour

ABSTRACT

The approximately 2.0 Ga old Beasley River Quartzite of the southwest Pilbara contains four major facies assemblages: mid fan, outer fan, tidal channel-tidal sand-bar, and offshore. These assemblages record the presence of an ancient tide-influenced fan-delta system which was developed in the vicinity of the eastern closure of the Wyloo Dome. The sedimentary history of the Beasley River Quartzite incorporates two phases of tectonically induced sedimentation. Each phase began with a period of uplift and fan progradation, followed by a more protracted period of tectonic quiescence characterized by tidal-shoreline sedimentation.

INTRODUCTION

Fan deltas have been defined as alluvial fans that prograde into a standing body of water from an adjacent highland (Wescott and Ethridge, 1980). The essential elements for the development of fan deltas are high relief, adjacent to the coastal zone, and high-gradient, bed-load streams that are braided to the coast, resulting in fan-shaped sedimentary deposits.

Rust (1979) has criticized the term "fan delta" on the basis that the fans are dominated by terrestrial processes, and do not show the distinct separation at sea level, between subaerial and subaqueous processes, which is a characteristic of deltas.

Rust's viewpoint is not supported by evidence from the Beasley River Quartzite which shows that subaqueous facies associations do vary from their subaerial counterparts, reflecting differences in the nature of the original depositional processes and sediment-reworking processes in these settings.

GEOLOGICAL SETTING AND PREVIOUS WORK

The Beasley River Quartzite forms part of the Wyloo Group, which unconformably overlies the Fortescue, Hamersley and Turee Creek Groups in the Western Pilbara (Fig. 1). The precise age of the Wyloo Group is uncertain but lies between 2.4 and 1.7 Ga (Grey, 1985). The Wyloo Group can be informally subdivided into: a lower sequence of terrestrial and shallow-marine shelf deposits; and an upper association of shelf-edge, submarine-fan and basin-plain sediments. Mafic volcanics occur in both the upper and lower subdivisions of the Wyloo Group.

The 370 m thick Beasley River Quartzite is the lowermost formation within the Wyloo Group. It is well exposed in a series of outcrops flanking the western and southern margins of the Hamersley Range. Brief descriptions of the formation were given by Daniels (1968, 1970, 1975) and Horwitz (1981, 1982). A more detailed account of the succession, outcropping between the eastern closure of the Wyloo Dome and the Hardey Syncline, 40 km to the east, is given by Trendall (1979). Seymour and Thorne (in press) include a summary of the major lithologies and provide a broad palaeoenvironmental interpretation of the Beasley River Quartzite on the Wyloo 1:250 000 map sheet.

This work describes and interprets the major facies associations present in the Beasley River Quartzite in the area immediately to the east and northeast of the Wyloo Dome. Most of the information was obtained from four measured sections, which are located on Figure 1, and 1:40 000 scale mapping of the adjacent outcrops.

FACIES ASSOCIATIONS

Four major facies associations can be identified within the Beasley River Quartzite:

mid-fan facies association

outer-fan facies association

tidal channel-tidal sand-bar facies association

offshore-marine facies association

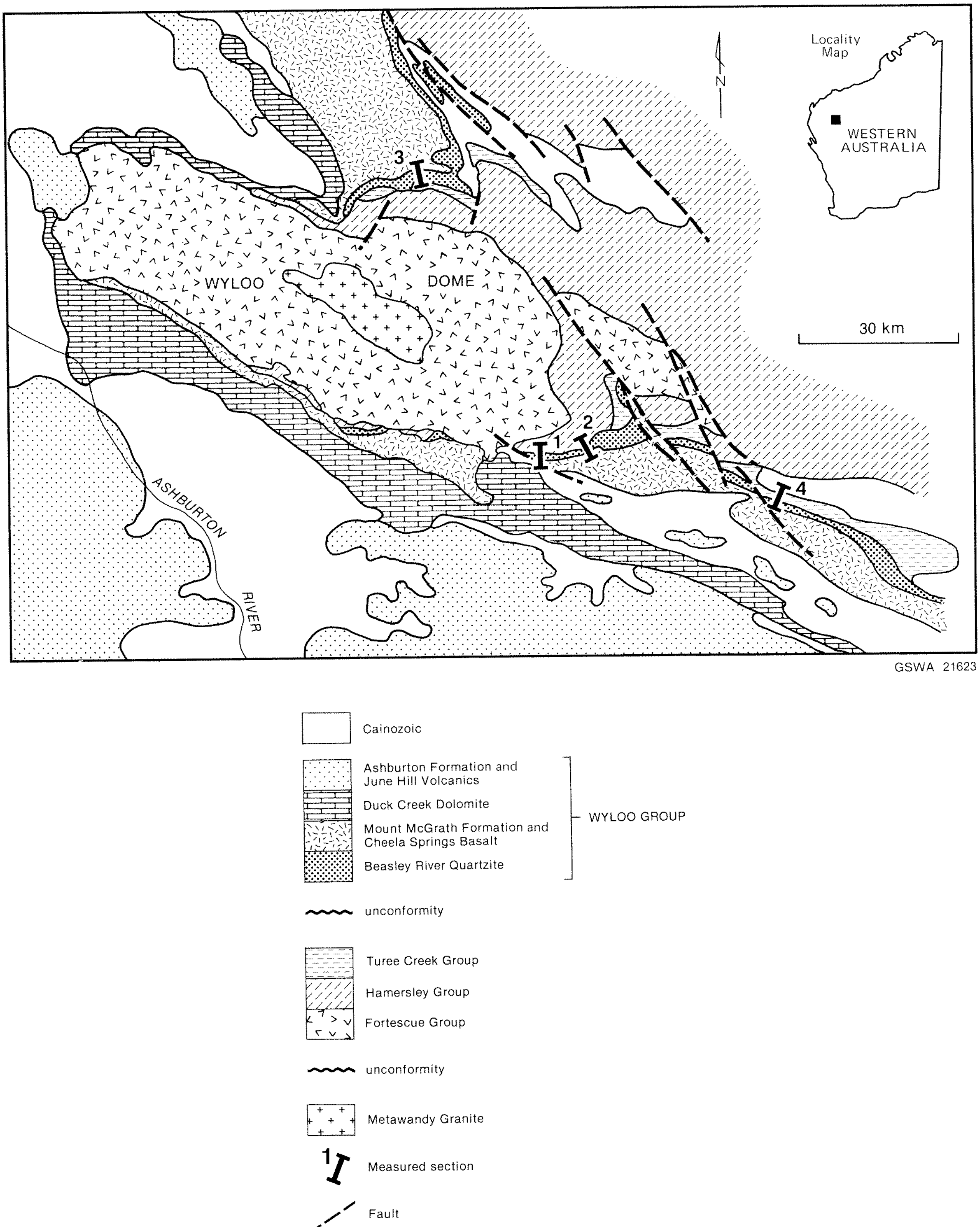


Figure 1. Generalized geological map of the Wyloo Dome showing locations of measured sections.

MID-FAN FACIES ASSOCIATION

The mid-fan association comprises mid-fan lobe and interlobe facies (Fig. 2). Much of this association is composed of conglomerates and coarse sandstones derived from the Hamersley Group. The clasts are predominantly angular fragments of banded iron-formation (BIF) and chert with lesser amounts of rounded rhyodacite and dolerite; the latter often exhibit a bleached outer rind.

MID-FAN LOBE FACIES

Description

This facies can be broadly divided into thick-bedded conglomerates and inversely graded conglomerates. Each lobe is usually composed of one of these lithotypes occurring either as a single unit or a stacked sequence. Thick-bedded and inversely graded conglomerates are sharp based, but show little evidence of erosion into the underlying beds.

Thick-bedded conglomerates occur in laterally continuous (>75 m) beds, 4-17 metres in thickness. Clast size varies up to 45 cm, but most lie in the range 2-15 cm. The conglomerates are predominantly clast supported and rarely exhibit any stratification other than a localized bedding-parallel orientation of the tabular clasts. Many beds display a mixture of clast and matrix support, and bedding-parallel or random fabric. In addition, the thicker conglomerates may also contain lenses of stratified coarse sand, locally slightly modified by soft-sediment deformation.

Inversely graded, thick (3-4 m) bedded conglomerates outcrop in laterally continuous beds with sharp, non-erosive bases. Grain size becomes coarser from a granule or, rarely, a sandy conglomerate to a cobble conglomerate with an average maximum clast size of 10 cm. Clast support is predominant in all but the sandy conglomerate horizons. Inversely graded conglomerates usually exhibit a random fabric.

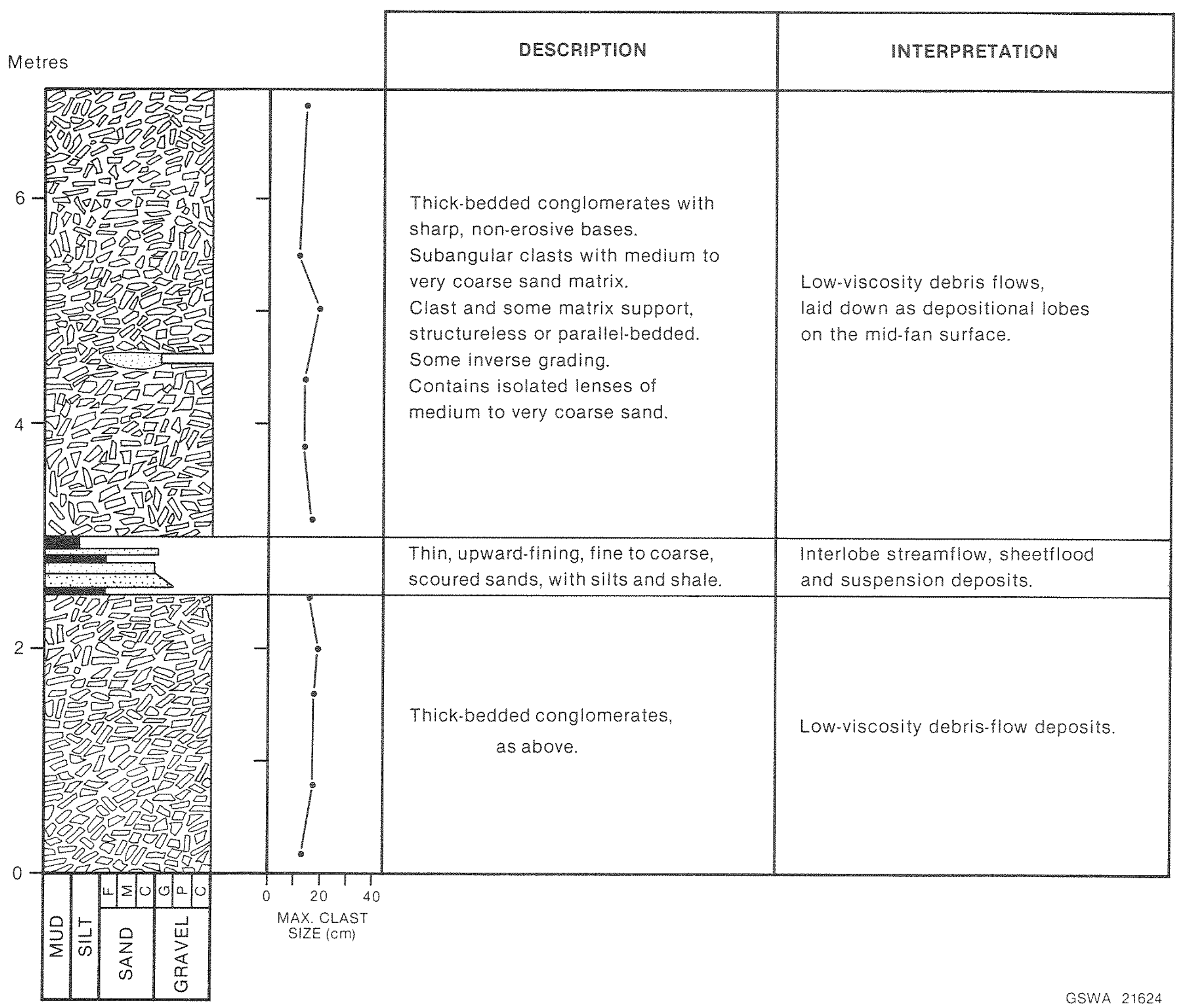


Figure 2. **The mid-fan association showing variations in thickness, grain size and internal structure between lobe and interlobe deposits.**

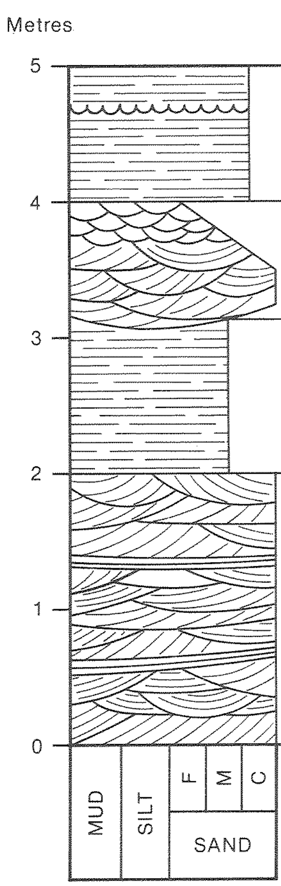
Interpretation

The thick-bedded conglomerates are interpreted as debris-flow deposits on the basis of the following: sharp, non-erosive bases; occurrence in thick well-defined units; lack of sorting or clast abrasion; and mixture of clast and matrix support. Lowe (1982) places deposits of this type into a broad category of cohesive debris-flow sediments, but noted that they were deposited by a flow in which the clasts were lubricated but not fully supported or suspended by the matrix. Lewis (1981) prefers the term non-cohesive or low-viscosity debris flows to describe these deposits, and suggests that intergranular support mechanisms (*e.g.* dispersive grain pressure) dominated.

The inversely graded conglomerates are also assigned to a debris-flow origin. This type of fabric is regarded as being a common feature of highly concentrated flows and possibly results from the dispersive pressure. The latter is caused by clast collision which forces smaller fragments downwards between the larger ones (Walker, 1975). Lowe (1982) notes that inverse grading is widespread in cobble-conglomerate beds greater than 0.4 metres in thickness, and refers to such deposits as density-modified grain flows. Here, the inverse grading reflects the high-dispersive pressure between large (cobble-sized) clasts.

The fabric of many of the thick-bedded conglomerates suggests that these units were often deposited from a single flow. However, the less homogeneous beds (especially those that incorporate lenses of well-stratified conglomerate and stratified sand) indicate that periods of streamflow sedimentation were intimately associated with the debris-flow events. This may indicate deposition from a series of pulses within a depositional event, or several periods of accumulation.

The thickness and coarse-grained nature of the debris-flow deposits described above suggest they occupy a comparatively proximal position on the fan surface. The planar boundaries and lateral extent of these bodies indicate that they were not confined to discrete channels, but instead accumulated as lobes on the mid-fan surface (Heward, 1978).



DESCRIPTION	INTERPRETATION
Parallel-laminated medium to coarse sand. Some symmetrical ripples.	Outer-fan sheetflood sands with evidence of reworking in a standing-water body.
Upward-fining, erosively based, trough cross-stratified sandstone. Fines up to a current or, in a few cases, a wave rippled top.	Waning flow in a distal fan channel.
Parallel-laminated BIF and quartzitic sandstone. Erosively based with current lineations and scours.	Outer-fan sheetflood (longitudinal bar) deposit.
Multiple intersecting small- to medium-scale troughed sets separated by erosive, undulose laminae. Poorly sorted immature sandstone. Unidirectional palaeocurrents.	Outer-fan braided-channel system.

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Figure 3. Schematic summary of rock types present in the outer-fan facies association.

INTERLOBE FACIES

Description

Upward-fining pebbly sandstones, fine- to coarse-grained sandstones, siltstones and shales occur in 0.3-4m thick units between the conglomerate units. Individual beds are 5-60 cm thick and may be lenticular or continuous over tens of metres. The pebbly and medium-grained sandstones frequently exhibit parallel stratification or trough cross-stratification; the latter occurs as sets 3-10 cm in thickness. Palaeocurrent data from localities 1 and 2 indicate that transport was towards the south and southeast respectively. Fine sandstones and siltstones are parallel-laminated or ripple cross-laminated. No internal structure is visible in the shales.

Interpretation

The sandstones, siltstones and shales are interpreted as streamflow sediments laid down in the intervals between the lobe-forming depositional events. The cross-stratified pebbly sandstones and medium- to coarse-grained sandstones represent minor braided-channel deposits, whereas the parallel- or ripple-laminated sandstones and siltstones are interpreted as sheetflood deposits (Bull, 1972). The structureless shales are considered to represent the suspended load of the channel and sheetflood deposits, though it is recognized that these sediments usually accumulate on the more distal parts of the fan (Wasson, 1977).

OUTER-FAN FACIES ASSOCIATION

This association consists principally of trough cross-stratified braided-stream deposits and parallel-laminated sheetflood sandstone (Fig. 3).

BRAIDED-STREAM FACIES

Description

The braided-stream facies consists of single or stacked sets of trough cross-stratified fine- to very coarse-grained sandstones. The sandstones are composed of BIF fragments, feldspar, and quartz; the latter component being volumetrically more important in the finer grained lithologies. Generally this facies consists of multiple intersecting troughs (5-20 cm in height and up to 1.5 m in width) which together form tabular cosets 0.30-1.5 m thick. Cosets are often separated by a few centimetres of erosively based, parallel-laminated, medium- to coarse-grained sand. Beds of medium- to coarse-grained trough cross-stratified sandstone locally grade up to a fine ripple-laminated sandstone. Symmetrical ripples are preserved on the tops of some beds.

Interpretation

Both single and multiple sets of trough cross-strata are attributed to the migration of dunes in a braided fluvial system (Harms and others, 1983). Trough cross-stratified beds, which fine upwards to a current-rippled top, were probably deposited during relatively short-lived waning-flow events; while the symmetrical ripples suggest local reworking of these deposits by wave processes in standing water. Palaeocurrent data from this facies (Fig. 4) are unidirectional at any stratigraphic level, for a given locality, and indicate that sediment transport took place towards the south and east at localities 1 and 2.

SHEETFLOOD SANDSTONE FACIES

Description

This facies consists of horizontally laminated fine- to coarse-grained sandstones composed of alternating light-coloured and dark ferruginous lithotypes. The light-coloured component consists of angular, very fine- to medium-grained quartz with lesser plagioclase and K-feldspar. Ferruginous lithotypes consist of medium- to coarse-grained BIF fragments and recrystallized hematite with some quartz detritus.

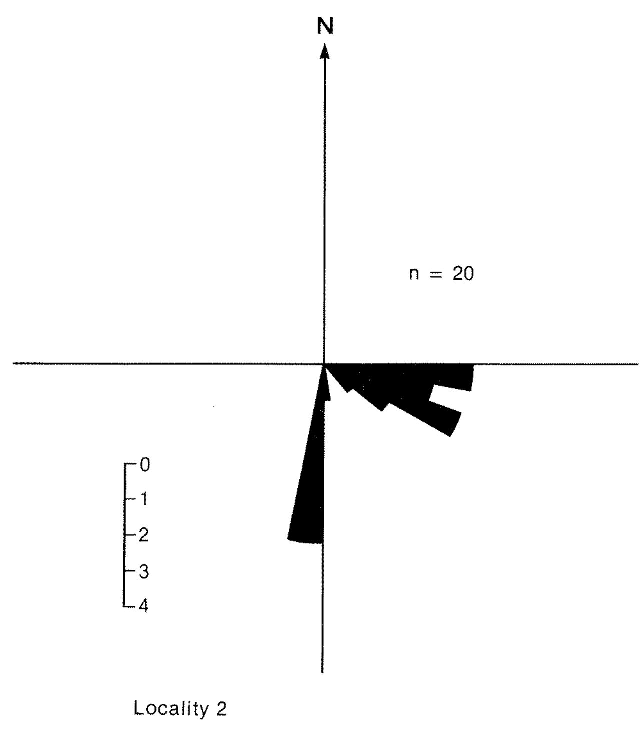


Figure 4. Outer-fan braided-channel palaeocurrents from locality 2.

These sandstones outcrop in tabular units up to 3 m in thickness. Many beds exhibit low-angle scours and occasional current lineations are also observed. In addition, 1-3 cm thick units of current- or wave-rippled, fine-grained quartz sandstone occur interbedded within the facies.

Interpretation

The presence of low-angle scours and current lineations in association with widespread parallel-lamination suggest that this facies was deposited in upper regime flows (Harms and others, 1983). Thin interbeds of current- and wave-rippled sandstone mark periods during which residual reworking by lower flow regime and oscillatory currents took place.

Parallel-laminated sandstones, similar to those described in this work are reported from the Gum Hollow fan delta at Nueces Bay, Texas (McGowen, 1970). Study of this Holocene fan delta shows that while sediment dispersal for much of the year takes place along braided channels, the most active phases of fan growth occur during periods of storm-induced sheetflooding. The sheetflood deposits typically comprise parallel-laminated fine-grained sands, laid down as extensive longitudinal bars over much of the fan surface.

TIDAL CHANNEL AND TIDAL SAND-BAR FACIES ASSOCIATION

TIDAL CHANNEL FACIES

Description

This facies outcrops in erosively based units 1-30 m thick. Individual channels range in width from a few metres to 120 m, with scouring between 0.25 and 8 m into the underlying beds.

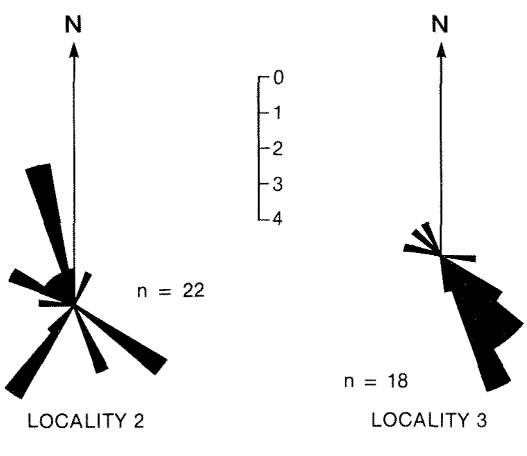


Figure 5. Tidal-channel facies palaeocurrents from localities 2 and 3.

Channels are usually composed of medium- to very coarse-grained quartz sand with local pebble-cobble lags of Woongarra Volcanics detritus in the larger channels. Rarely, the channels may be infilled by silt and shale.

The internal structure of the tidal-channel-facies sandstones is dominated by stacked sets of trough cross-stratification, with individual sets 0.1-1.0 m thick. Palaeocurrent data from this facies at localities 1 and 2 (Fig. 5) show south- and southeast-directed transport, with a minor component towards the northwest. Measurements from locality 3 on the northeastern flank of the Wyloo Dome (Fig. 5) show a predominantly northwesterly directed mode with an additional component indicating a broad southerly flow.

TIDAL SAND-BAR FACIES

Description

Tidal sand-bar deposits, comprising fine- to medium-grained quartz sandstones, occur in sharp-based, tabular units 1-5 m thick (average 2 m). Individual units typically consist of a lower assemblage of trough and tabular cross-stratified sandstones which pass upwards into a ripple-laminated division 0.1-1.5 m thick (Fig. 6).

Much of the lower portion of the sand-bar facies is dominated by 3-15 cm thick sets of gently scoured trough cross-stratification. Troughs exhibit high- or low-angle foresets with a generally low angle of climb, and often contain reactivation surfaces. The tops of individual troughs are commonly planed off and pass upwards into either thin layers of erosively based undulatory lamination, or 5-20 cm thick sets of high-angle planar tabular cross-stratification. Palaeocurrent data from the lower part of the sand-bar assemblage shows a bipolar distribution. At localities 1 and 2 (Figure 1), troughed sets indicate transport towards the northwest and southeast, while tabular sets record a broad northward or southerly directed flow (Figs 7A, B).

The upper ripple-laminated division is typically 5-15 cm thick and often capped by linguoid current-ripple or wave-ripple bedforms. Palaeocurrent directions from the current ripples differ from those in the underlying cross-stratified portion of the bar by 60°-120° (Fig. 7C). Less commonly, the upper division of the bar is up to 1.5 m thick and composed of thin sets of symmetric and asymmetric rippled fine-grained sandstone intercalated with 0.5 to 2 cm thick mudstones and 5 to 10 cm thick horizontal stratified sandstones.

The latter are erosively based and the upper surface may show evidence of wave and current reworking. Wave-ripple crests are oriented east-west (Fig. 7D).

TIDAL CHANNEL—TIDAL SAND-BAR ASSOCIATION: INTERPRETATION

The sedimentary structures and palaeocurrent patterns within this association suggest tidally influenced sedimentation. Johnson (1975), summarizing the work of de Raaf and Boersma (1971), noted that the single most important criterion for recognition of tidal deposits is directional bimodality in both large-scale and small-scale cross-stratified units—a feature which characterizes this part of the Beasley River Quartzite. Other features, singularly not diagnostic, but nevertheless abundant in tidal deposits include:

(a) Coupled arrangement of large-scale and small-scale bedded units.

(b) Presence of fining-upward sequences in which the basal large-scale unit is generally channelled.

(c) Discontinuity (reactivation) surfaces within unidirectional cross-stratified sets.

(d) Various types of sand-mud alterations, *e.g.* lenticular bedding or flaser bedding.

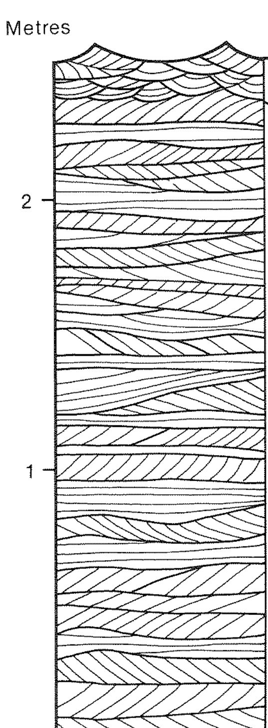
All these above features occur in the tidal channel - tidal sand-bar assocation of the Beasley River Quartzite.

Tidal channels

The following features are consistent with deposition in tidal channels: rounding and compositional maturity of clasts; presence of pebble lags; erosive boundaries; and medium- to large-scale stratification. In this environment, high rates of grain collision, scouring, and deposition under strong currents take place. Palaeocurrent data indicate that the larger channels were principally ebb-dominated (by comparison with the source direction for terrigenous detritus). Smaller channels show a mixture of ebb- and flood-dominated structures and scattered palaeocurrent directions typical of a migrating channel system (Weimer and others, 1982).

Tidal sand-bars

The trough and tabular cross-stratified sets are interpreted as dunes and small sand waves that formed in lower flow-regime conditions on tidal sand-bars. The complex internal organization of this facies is reflective of deposition in response to the time-velocity asymmetry of tidal currents (Klein, 1970a). Rounded upper-set boundaries and



DESCRIPTION	INTERPRETATION
Linguoid current ripples. Quadrimodal palaeocurrent distribution.	Current ripples formed during late-stage run off on bar surface.
Fine quartzitic sandstones, displaying a complex association of trough and tabular cross strata, with washed-out set tops and reactivation surfaces. High - or low-angle forsets. Biplolar-palaeocurrent distribution.	Dunes and small sand waves, formed under lower flow-regime conditions on inter-tidal sand bars. Complex internal structure reflects time-velocity assymmetry of tidal currents.

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Figure 6. Vertical sequence through the sand-bar facies showing variations in sedimentary structures.

reactivation surfaces formed as a result of the destructional phase of tidal-cycle reworking, whereas the bipolar palaeocurrent distribution suggests deposition took place during both ebb and flood stages. The thin cappings of current ripples overlying the complex cross-stratification show a quadrimodal-palaeocurrent distribution, and were formed by late-stage runoff when flow directions were controlled by sand-bar topography (Klein 1970a).

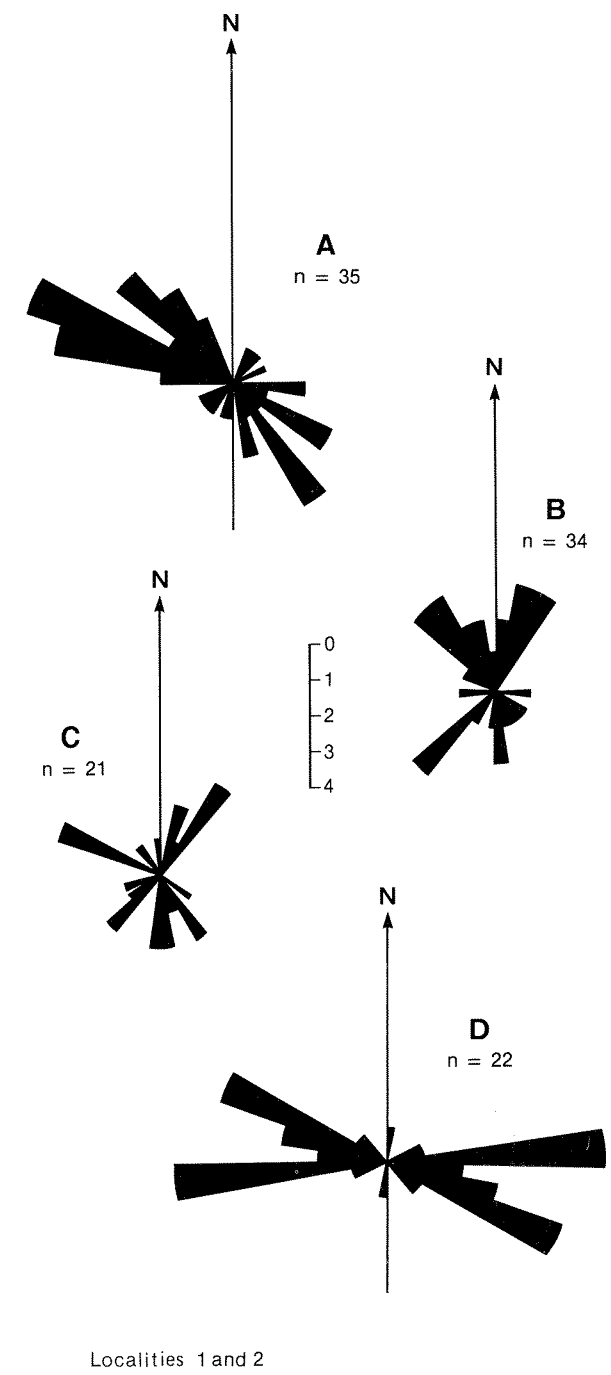

Figure 7. Palaeocurrents from the sand-bar facies at localities 1 and 2.
A—trough cross-strata
B—tabular cross-strata
C—current ripples
D—wave ripple crests

The thicker cappings of wave- and current-rippled sandstones with interbedded shales and parallel-laminated sandstones are interpreted as tidal-flat deposits. The rippled sandstones record deposition during ebb and flood periods, while the shales were deposited from suspension during prolonged periods of slack-water sedimentation. There is no evidence of any fine-grained sandstone-mudstone alteration that may have been formed during a single tidal cycle in a manner envisaged by Klein (1970b). The parallel-laminated erosively based sandstones are interpreted as storm deposits. Similar lithologies were described by Johnson (1975) from the Precambrian of northern Norway.

OFFSHORE-FACIES ASSOCIATION

Description

This association comprises 10-75 m of parallel-laminated shale and siltstone, alternating with minor, 0.5-5 cm thick, parallel- or ripple-laminated fine-grained sandstones.

Interpretation

The shales and siltstones are suspension deposits, laid down in an offshore or prodelta setting (Elliott, 1978). The thin sandstones were probably deposited from turbidity currents during periods of increased terrigenous supply.

DISTRIBUTION OF FACIES ASSOCIATIONS

In the vicinity of the Wyloo Dome, the lower part of the Beasley River Quartzite comprises mid-fan and outer-fan deposits overlain by tidal-channel/sand-bar lithologies, which are in turn succeeded by offshore sediments (Fig. 8). The remainder of the succession consists of a second suite of fan and tidal deposits. At Horseshoe Creek, fan deposits are thin, and overlie shoreline sandstones while offshore deposits are extensive and occur interbedded with the tidal-channel and sand-bar association.

DEPOSITIONAL MODEL FOR THE BEASLEY RIVER QUARTZITE

The interrelation of facies associations can be interpreted by the following depositional model (Fig. 9).

Deposition of the Beasley River Quartzite was initiated following a period of tectonism which affected the southern margin of the Pilbara Craton

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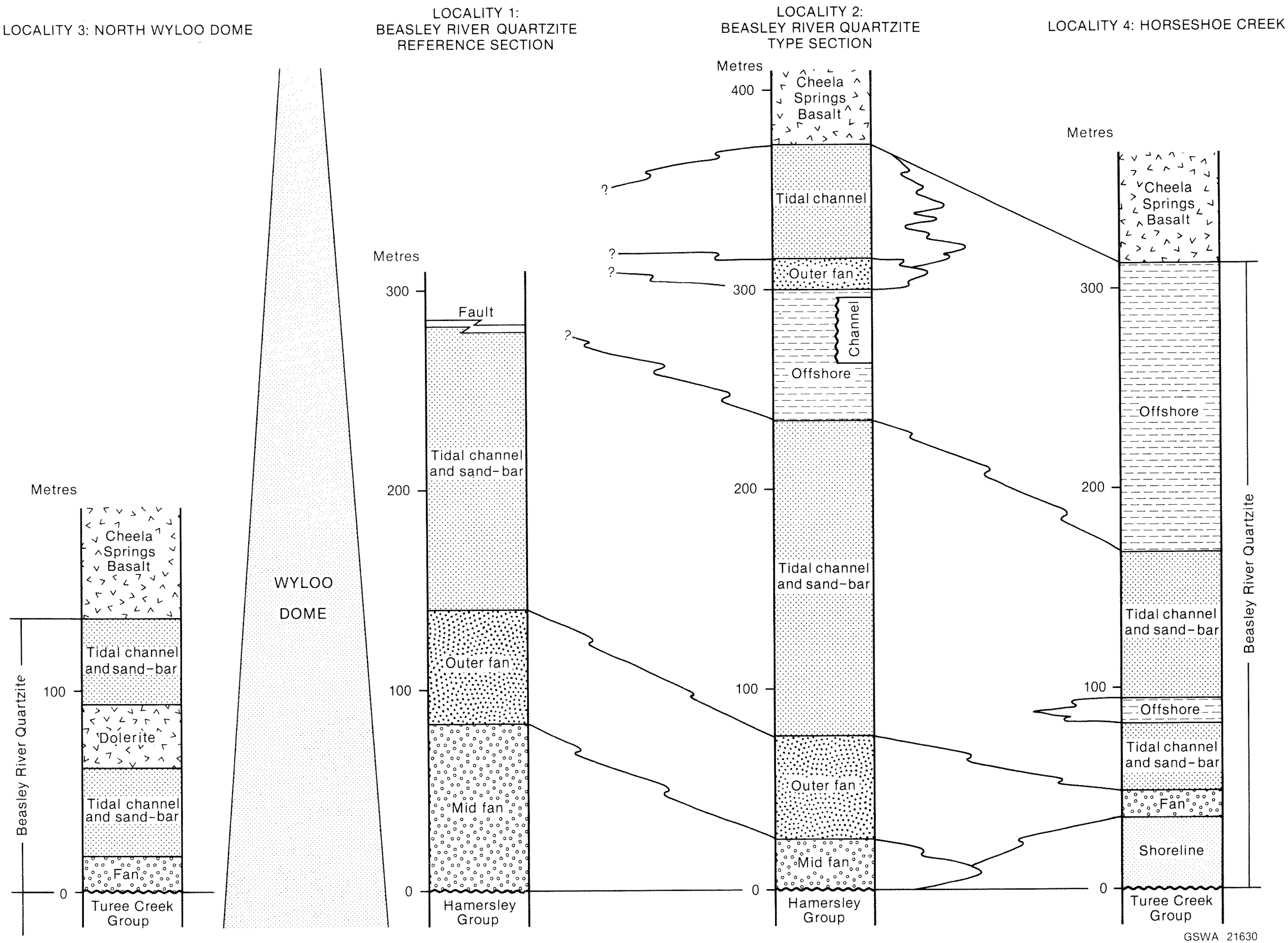

Figure 8. Generalized cross sections through the Beasley River Quartzite showing interrelation of facies associations.

(Seymour and Thorne, in press). Immediately to the north and southeast of the Wyloo Dome, the major controls on sedimentation were:

(a) A tectonically active, upland area of Hamersley and Turee Creek Group rocks adjacent to the eastern closure of the Wyloo Dome. Evidence for this comes from the widespread occurrence of conglomerates in the Beasley River quartzite. In addition, Seymour and Thorne (in press) show that the Wyloo Dome was an active tectonic feature during deposition of most of the Wyloo Group.

(b) A narrow coastal plain to the north and southeast of the Wyloo Dome. The absence of extensive fluvial deposits in the Beasley River Quartzite coupled with a comparison of the dimensions of most modern alluvial fans suggests a 2-5 km wide coastal plain.

(c) A low wave energy (see discussion) tidally influenced shoreline seaward of the coastal plain.

Ephemeral streams emerging from the upland area built a series of alluvial fans on to the coastal plain. Maximum development of fan deposits occur to the southeast of the Wyloo Dome, in the vicinity of locality 1. Here, debris flows were deposited close to the fan apex, whereas braided-streamflow sediments characterize the mid to outer parts of the fan. More distal parts of the subaerial fan experienced longitudinal-bar formation during periods of storm-induced sheetflooding with some reworking of these deposits by coastal processes. Further still from the upland area, the coastline was dominated by tidal channels and sand-bars, with deeper waters offshore.

Following the initial phase of coarse clastic input and fan growth, rates of subsidence slightly exceeded those of deposition resulting in the advance of the shorelines and offshore facies of the subaerial fan. The period of coastal retreat was abruptly terminated by a second phase of fan growth and shoreline migration, which controlled facies distributions until the close of the Beasley River Quartzite deposition.

DISCUSSION

Alluvial fans that prograde into a standing body of water are termed fan deltas (McGowen, 1970). Rust (1979) has criticized the term fan delta on the basis that fans are allegedly dominated by terrestrial processes, and do not show a distinct separation at sea level between subaerial and subaqueous processes. This criticism could be

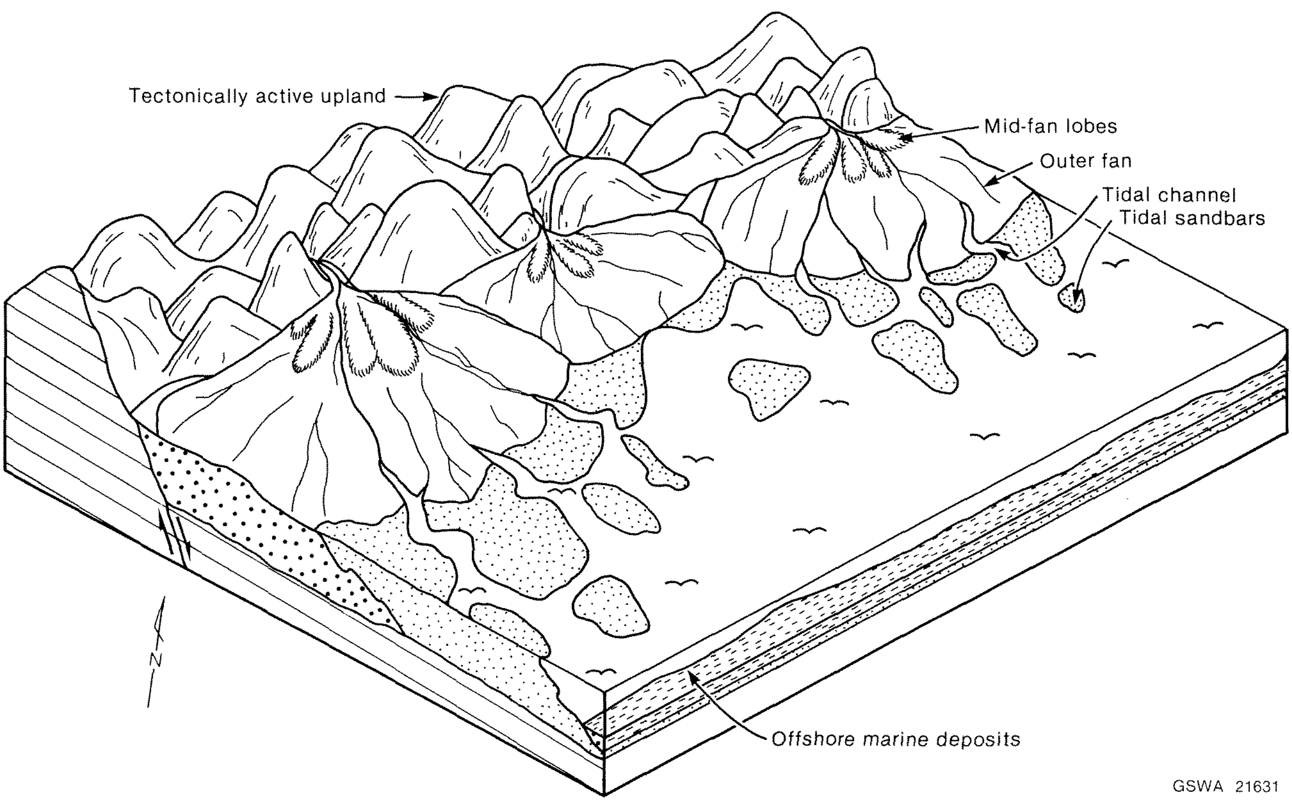

Figure 9. Model showing the principal depositional environments of the Beasley River Quartzite in the vicinity of the Wyloo Dome.

directed at fans that prograde directly on to a steeply sloping shoreline. However, fans that build out on to a gently inclined marine shelf are characterized by distinct zones of alluvial and coastal processes (Wescott and Ethridge, 1980). The Beasley River Quartzite fan-delta system is of this second type. The subaerial portion of the fan was dominated by streamflow, debris-flow and sheetflood processes which resulted in deposition of texturally and mineralogically immature sediments comprising BIF, chert, and quartz detritus. These alluvial lithotypes contrast with the fine quartzitic deposits of the shoreline fan, whose maturity reflects prolonged reworking by tidal and, to a lesser extent, wave processes.

Modern lobate fan deltas, similar to those envisaged for the Beasley River Quartzite, are described from Lower Cook Inlet, Alaska (Hayes and Michel, 1982). These Holocene fan deltas form in protected areas of low-wave energy, and as a result show little beach and shoreface development. The paucity of beach and shoreface deposits in the Beasley River Quartzite suggests that this ancient

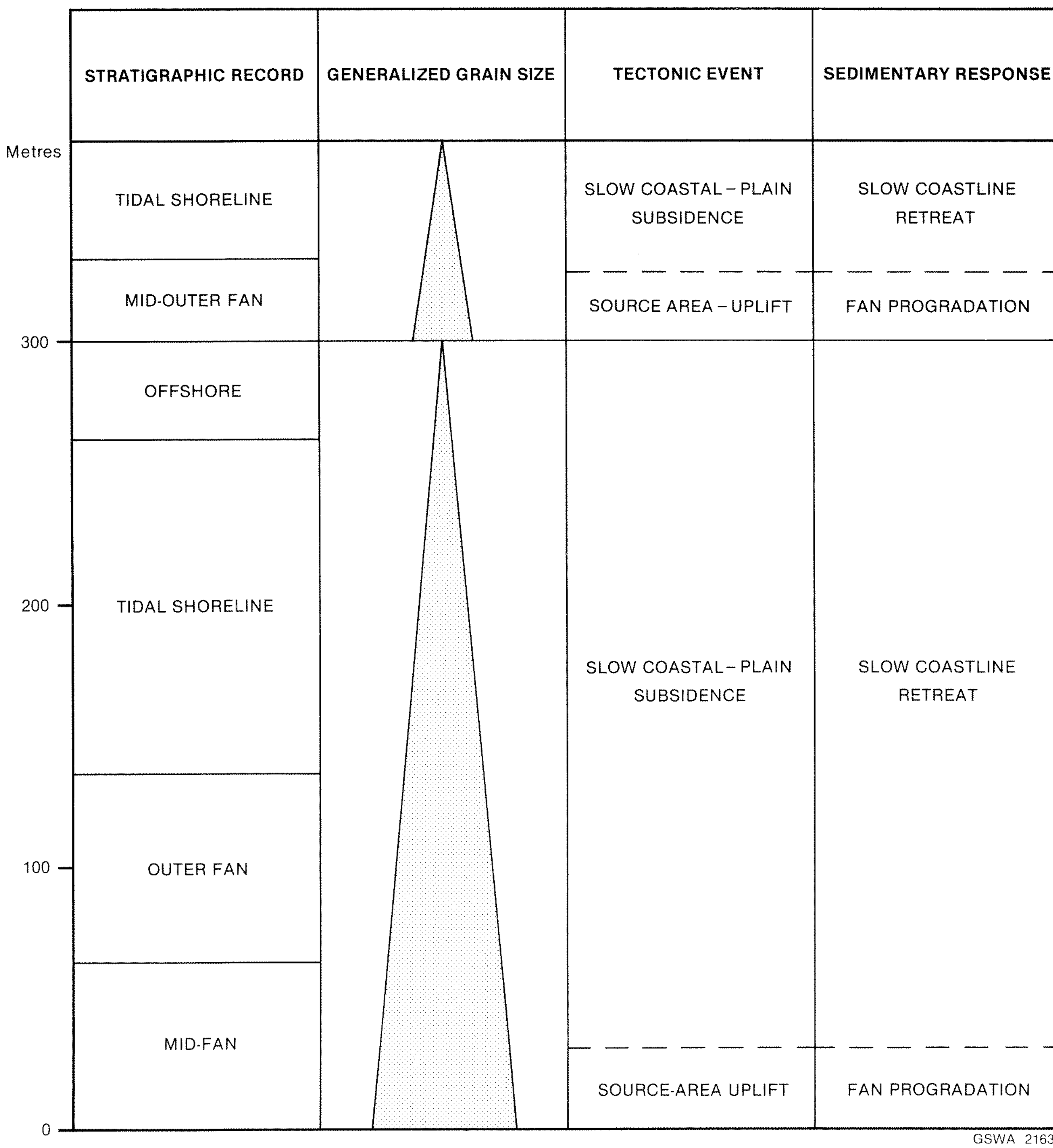

Figure 10. A summary of the depositional history of the Beasley River Quartzite showing the relation between sedimentary facies and tectonism.

fan-delta system also developed in a setting that was at least partially protected from wave action. It is likely that some of this protection could have been provided by an upland area now occupied by the Wyloo Dome.

Klein (1971) has shown how the thickness of a prograding tidal-flat sequence can be used to determine palaeotidal range. The possible discrepancies in this theory were, however, noted by Klein (1971), and these are caused by subsidence, erosion, and compaction. When these sources of error are coupled with the fact that in the Beasley River Quartzite it is difficult to distinguish low tidal-flat sediments from those of migrating tidal channels, it becomes impossible to provide a reliable estimate of palaeotidal range.

The central role played by tectonism in the development of Holocene fan deltas is widely recognized (Wescott and Ethridge, 1980). Synsedimentary tectonism was also a major influence on the depositional history of the Beasley River Quartzite. This succession (summarized in Fig. 10) can be interpreted as the result of two phases of sedimentation. Each phase represents a relatively brief period of tectonically induced fan-delta growth which was followed by a more protracted phase of tectonic quiescence (characterized by tidally influenced shoreline sedimentation). Fan-delta lobe switching, which should have produced a similar sedimentary record, is not considered a likely mechanism to explain the observed facies relationships since the entire fan system was affected by the same pulses in sedimentation. Palaeocurrent data and facies distributions presented in this study suggest that the eastern Wyloo Dome and a northwesterly trending fault system in the adjacent Hamersley Range (Fig. 1) were active elements and the sources of detritus during deposition of the Beasley River Quartzite. This view contrasts with the interpretation of Horwitz (1981, 1982) who invokes a source area occurring to the south of the present-day outcrop.

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AGE AND STRATIGRAPHY OF A SEQUENCE OF METAVOLCANIC AND METASEDIMENTARY ROCKS IN THE PRAIRIE DOWNS-DEADMAN HILL AREA, SOUTHWESTERN MARGIN OF THE SYLVANIA DOME

by I. M. Tyler

ABSTRACT

A sequence of metavolcanic and metasedimentary rocks at the southwestern margin of the Sylvania Dome clearly post-dates the granitoid-greenstone terrain of the dome and can be correlated with the Fortescue and Hamersley Groups of the southwestern Hamersley Basin.

INTRODUCTION

The Sylvania Dome is an inlier of granitoid-greenstone terrain outcropping at the southern margin of the Hamersley Basin (Fig. 1A). The northern margin of the dome is in tectonic contact with the Fortescue and Hamersley Groups, but to the south and east it is unconformably overlain by the Bangemall Group.

At the southwestern margin, in the general vicinity of Deadman Hill and the Prairie Downs Fault, Daniels and MacLeod (1965) identified an area of metabasaltic and metasedimentary rocks which they equated with greenstone sequences of the dome. Recent detailed mapping by the author has confirmed the suggestion made by Horwitz and Smith (1978) that a major part of these rocks belongs to the 2 700 Ma Fortescue Group, and that the metasedimentary rocks at Deadman Hill belong to the overlying Hamersley Group, (Horwitz, 1980, Fig. 1; Morris and Horwitz, 1983, Fig. 2a).

GEOLOGY OF THE PRAIRIE DOWNS-DEADMAN HILL AREA

Basement rocks of the Sylvania Dome

The western part of the Sylvania Dome consists mainly of foliated granitoid rock intruded by undeformed granitoid rock. Both rocks are medium grained and extensively recrystallized. Areas of greenstone belts occur and are intruded and extensively veined by the undeformed granitoid rock. They consist of mafic and ultramafic rocks together with metasedimentary rocks including banded chert and banded iron-formation. They are strongly deformed and show a well-developed metamorphic foliation. The cherts show a phase of isoclinal folding, and this is followed by a phase of tight folding which is widespread in the other rock types. Metamorphic mineral assemblages formed in conditions transitional from greenschist to amphibolite facies.

Both granitoid rocks and greenstones are intruded by numerous north-trending mafic dykes.

The cover succession

A succession of interbedded mafic volcanics, felsic volcanics, and metasedimentary rocks is well exposed along the track from Jillary Well to Prairie Downs Homestead (Fig. 1B). It has a southeasterly strike near Jillary Well, but strikes south to the north and west of Nirran Nirrie Bore. Dips are moderate or steep to the southwest and west. The succession has a minimum thickness of 6.5 km not 2 km as indicated by Horwitz (1980, Fig. 1) and Morris and Horwitz (1983, Fig. 2a). Contacts with adjacent granitoid rocks and greenstones are everywhere faulted.

Three main features distinguish this succession from the greenstones.

(a) The rocks are less deformed, and isoclinal folds are absent; but tight folds with a well-developed cleavage occur. Late, open folds show faults along their axial planes.

(b) The metamorphic grade is lower, and mineral assemblages indicate lower to middle greenschist facies conditions.

(c) The greenstones are extensively veined by granitoid rocks and are cut by north-trending mafic dykes, whereas these intrusions are absent from the cover sequence.

Northwest and west-northwest-trending mafic dykes cut both the granitoid-greenstone terrain and the overlying cover sequence.

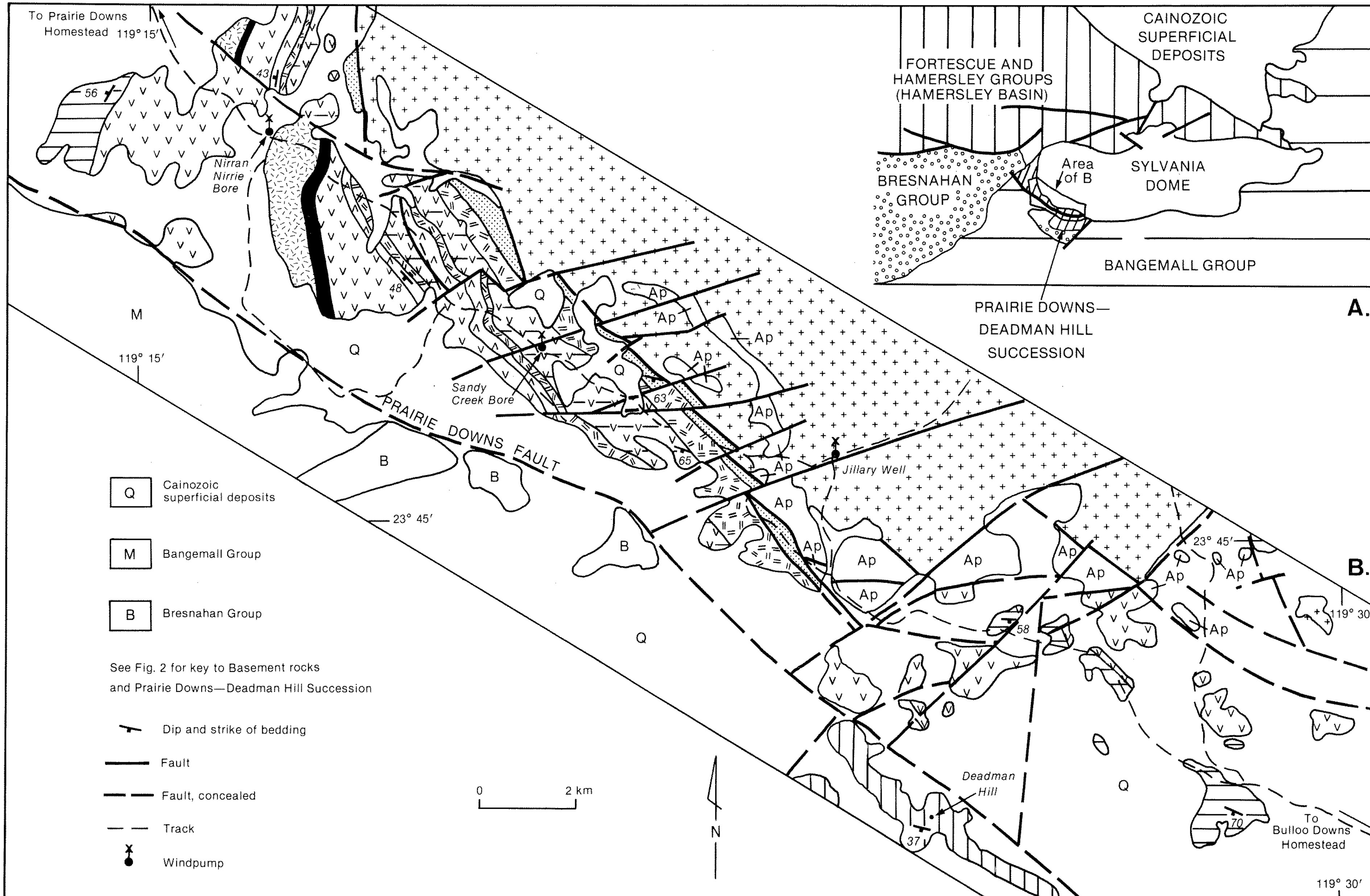

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Figure 1. A—Simplified map of regional geology. Location of Fig. 1B outlined. B—Geological map of the Prairie Downs-Deadman Hill area with mafic dykes omitted.

Six distinct units can be recognized within the succession (Fig. 2). A lower metasedimentary unit consists of interbedded phyllite, quartz-muscovite schist, metasandstone and peloidal carbonate. The metasandstone is generally coarse grained, shows poorly developed cross-bedding, and forms a good marker horizon up to 5 m thick. It is underlain by a purple-brown phyllite. Above the metasandstone, quartz-muscovite schist (possibly derived from felsic tuff) is interlayered with thin 10-15 cm thick, peloidal carbonate units.

This schist is overlain by a coarse metagabbroic sill (with a metapyroxenitic base) and a lower mafic volcanic unit, comprising interlayered mafic lava flows and tuffs which are locally bedded. Individual rock units are in the order of 2-3 m thick. These volcanic rocks are overlain by a second layered metagabbro sill below a felsic pyroclastic unit. This comprises felsic tuff which is locally laminated. A 5-10 m thick banded chert occurs above the tuff.

These rocks are overlain by an upper mafic unit, consisting of a monotonous sequence of metabasalts. However, it includes a komatiite flow with well-developed spinifex textures. An adjacent unit of serpentinite may represent a dunitic base to the flow.

An upper metasedimentary unit, comprising phyllite and silicified mudstone, occurs above the metabasalt.

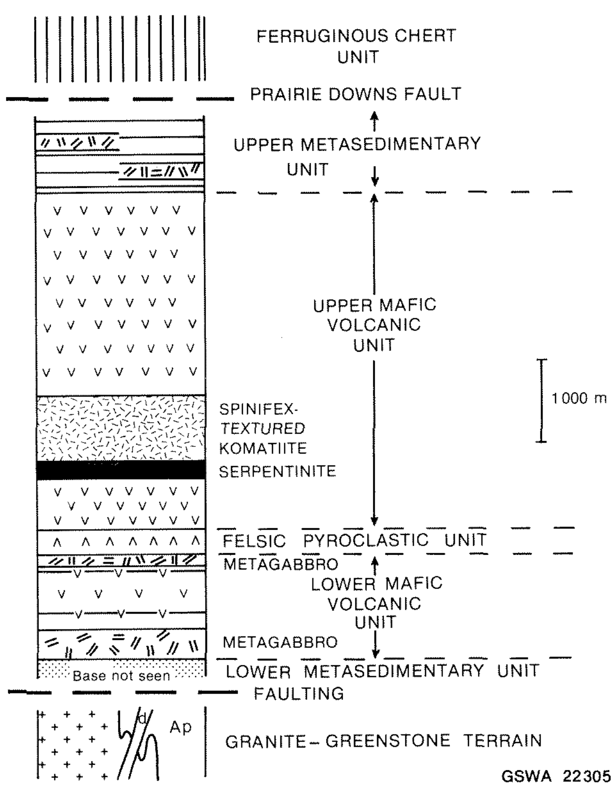

Figure 2. Stratigraphic column of the Prairie Downs-Deadman Hill succession.

To the south, the sequence is truncated by the Prairie Downs Fault, whereas to the west it is covered by Tertiary colluvial deposits.

The area to the south and east of Jillary Well is extensively disrupted by faulting, and a continuous sequence is not preserved. However, individual fault blocks may be correlated with the upper mafic volcanic unit and the overlying upper metasedimentary unit. Several mafic sills are evident within the uppermost unit.

At Deadman Hill, a ferruginous chert unit, characterized by internal podding, is exposed. This is separated from the sequence described above by the southeasterly extension of the Prairie Downs Fault (Fig. 1B).

STRATIGRAPHIC INTERPRETATION

The succession described above clearly post-dates the granitoids and greenstones of the dome. The predominantly mafic character and relatively high metamorphic grade of the succession distinguishes it from the adjacent Bangemall and Bresnahan Groups. As will be discussed below, the succession is similar to that of the Fortescue Group and lower part of the Hamersley Group of the Hamersley Basin, and this is the preferred correlation.

Fortescue Group—Hamersley Group stratigraphy

The stratigaphy of the Fortescue and Hamersley Groups has recently been reviewed and summarized by Hickman (1983) and Trendall (1983). Hickman (1983) produced a standardized nomenclature for rocks on, or adjacent to, the Pilbara Block north of the Fortescue River (Table 1).

In the southwestern part of the Hamersley Basin, the nomenclature of the succession described by de la Hunty (1965) on the Mount Bruce sheet—and extended by Daniels (1968, 1970) to the Turee Creek and Wyloo sheets—has been retained by Seymour and Thorne (in press) for the second edition of the Wyloo sheet. The Mount Roe Basalt, identified by Blight (1985), is included beneath the Hardey Sandstone. The succession is shown in Table 1.

On the Newman sheet, Daniels and MacLeod (1965) did not identify the Mount Jope Volcanics and believed the Jeerinah Formation to lie directly on the Hardey Sandstone. Horwitz (1976), in recording a section through the Fortescue Group adjacent to the northern contact of the Sylvania Dome identified a mafic volcanic unit between the Hardey Sandstone and the Jeerinah Formation which he ascribed to the 'Maddina Volcanics'.

Correlation with the Prairie Downs-Deadman Hill succession

The succession in the Prairie Downs-Deadman Hill area closely resembles that in the southwestern Hamersley Basin and a direct correlation is suggested (Table 1). The Mount Roe Basalt is absent. The lower metasedimentary unit represents the Hardey Sandstone; the lower mafic volcanic unit, the felsic pyroclastic unit, and the upper mafic volcanic unit, represent equivalents to the Boongal Pillow Lava Member, the Pyradie Pyroclastic Member, and the Bunjinah Pillow Lava Member of the Mount Jope Volcanics; and the upper metasedimentary unit represents the Jeerinah Formation. The ferruginous chert recognized at Deadman Hill, although separated from the main succession by the Prairie Downs Fault, is correlated with the Marra Mamba Iron Formation of the Hamersley Group.

The occurrence of a similar succession in the Prairie Downs-Deadman Hill area to that seen in the southwestern Hamersley Basin suggest that it may be possible to extend the stratigraphic nomenclature used by Seymour and Thorne (in press) on the Wyloo Sheet throughout the Fortescue Group of the southern Hamersley Basin. It seems unlikely, in view of the evidence presented in this paper, and of the observations of Horwitz (1976), that the Mount Jope Volcanics are absent from the rest of the Newman sheet.

TABLE 1. FORTESCUE GROUP—LOWER HAMERSLEY GROUP STRATIGRAPHY

Pilbara Block (a)			*Southwestern Hamersley Basin (b)*		*Prairie Downs-Deadman Hill*	
Formation	*Principal lithologies*		*Formation*	*Principal lithologies*	*Unit*	*Lithologies*
			HAMERSLEY GROUP			
Marra Mamba Iron Formation	Shale, chert and BIF		**Marra Mamba Iron Formation**	Chert and BIF	**Ferruginous Chert Unit**	Ferruginous chert
			FORTESCUE GROUP			
Jeerinah Formation	Siltstone, shale and chert, sandstone, local felsic lava, tuff and agglomerate, basalt; mafic sills locally abundant		**Jeerinah Formation**	Shale, chert, BIF, mudstone, quartzite, dolomite and dolerite	**Upper Metasedimentary Unit**	Phyllite, silicified mudstone; mafic sills locally developed
Maddina Basalt	Vesicular and amygdaloidal basalt and andesite	Mount Jope Volcanics	**Bunjinah Pillow Lava Member**	Pillow basalt, vesicular basalt, and volcanic breccia	**Upper Mafic Volcanic Unit**	Metabasalt and komatiite
Kuruna Siltstone	Sandstone and siltstone, local shale, ooidal sediments, pisolitic tuff and banded siliceous limestone					
Nymerin Basalt	Coarse textured basalt		**Pyradie Pyroclastic Member**	Tuff with agglomerate and volcanic bombs, local chert, shale, BIF and pillow lava	**Felsic Pyroclastic Unit**	Felsic tuff and banded chert
Tumbian Formation	Siliceous limestone (stromatolitic) and pisolitic tuff; subordinate tuff, basalt and sediments					
Kylena Basalt	Massive amygdaloidal and vesicular basalt and andesite; subordinate but widespread interbedded sediments		**Boongal Pillow Lava Member**	Pillow basalt, minor carbonate	**Lower Mafic Volcanic Unit**	Metabasalt and mafic tuff
Hardey Sandstone	Sandstone and conglomerate; local tuff, shale, mudstone and basalt		**Hardey Sandstone**	Sandstone, arkose, shale and conglomerate	**Lower Metasedimentary Unit**	Metasandstone, phyllite, quartz muscovite schist, peloidal carbonate
Mount Roe Basalt	Basalt and andesite with thin basal sedimentary rocks		**Mount Roe Basalt**	Basalt, intermediate to mafic tuffs and coarse clastic sediments; local felsic pyroclastics	**Absent**	

(a) is taken from Trendall (1983), Table 3-II

(b) is based on De la Hunty (1965) and Seymour and Thorne (in press)

Recent mapping by the author has confirmed the suggestion by Horwitz and Smith (1978) that volcanic and metsedimentary rocks in the Prairie Downs-Deadman Hill area of the Newman sheet, previously grouped with greenstones of the Sylvania Dome (Daniels and MacLeod, 1965) are part of the Fortescue and Hamersley Groups. A correlation with the succession recorded from the southwestern Hamersley Basin is preferred.

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