ARCHAEAN – EARLY PROTEROZOIC GRANITOIDS, METASEDIMENTS and MYLONITES of SOUTHERN EYRE PENINSULA, South Australia

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PREFACE

Although never the centre of major base metal, gold or copper mining, interest in Eyre Peninsula followed close on the heels of copper mining operations at Wallaroo and Moonta in the late 1850's and early 1860's. From about 1920 until the early 1960's, iron-ore mining in the Middleback Ranges became the mainstay of mineral production in South Australia and, in fact, of iron-ore production in Australia. Today, iron ore is still being mined and Iron Duke, at the southern end of the Middleback Ranges, is about to be commissioned as a new iron-ore mine. Nephrite jade, talc, marble and granite are all being profitably quarried at various locations on the Peninsula, a promising zinc and lead prospect has been discovered at Menninnie Dam north of Kimba, a moderately-sized graphite deposit has been delineated at Mikkira just south of Port Lincoln, and mineral exploration continues throughout.

Not only has economic interest been maintained on Eyre Peninsula for an extended period of time, but academic interest has also continued since Tilley, in some of his many classic papers of the early 1920's, first described the metamorphic and igneous rocks in some considerable detail. The Peninsula has never enjoyed the same intensity of academic interest as Mount Isa and Broken Hill, yet it shares much in common with those regions, and also with the Pine Creek Geosyncline and many, similar, Early Proterozoic provinces throughout the world (e.g. the Lake Superior region of North America).

This excursion guide is a refinement of a 6-day excursion led by Parker, Fanning, and Flint in 1980 at the height of economic interest in the basement geology of the Gawler Craton. The interest in those days was for Alligator River-style uranium mineralization, Broken Hill-style base-metal mineralization and, particularly, for clues to the concealed basement geology of the Olympic Dam region. A much-shortened, 2-day excursion, based on the same guide but led by Dr Robin Oliver of the University of Adelaide, was run in conjunction with the Fourth International Antarctic Symposium in 1982 to compare southern Eyre Peninsula with similar rocks of the Commonwealth Bay region, Antarctica. Then, in 1986 in conjunction with the Eighth Australian Geological Convention and supported by the S.G.T.S.G., the basics of this excursion guide were prepared. The latter was essentially a three-day excursion to study, especially, the superb mylonites and highly-deformed granitic gneisses. These notes and stops have been slightly expanded here to include a few extra stops from the original guidebook. The guidebook is not an absolutely exhaustive treatise on the geology of Eyre Peninsula but it is still the most complete documentation yet compiled. It describes a truly EPIC tour of some of the best coastal outcrops in Australia.

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A.J. PARKER

MAP BASE AND FIELD GUIDES TO ADJACENT AREAS.

The field-guide area is covered by the 1:250 000 map sheet areas of KIMBA, LINCOLN and WHYALLA. Geological maps of LINCOLN and WHYALLA have been published by the S.A. Geological Survey and the Kimba sheet should be published in 1989. The LINCOLN sheet, although previously published (Johns, 1958), is presently under revision and a second edition will be published in the future. Unpublished 1:100 000 geological maps are available for KIMBA and WHYALLA 1:250 000 sheet areas. Three 1:50 000 geological maps Cowell, Mangalo, and Rudall, have been published, and the emphasis of these maps is on the structure of the basement geology. All these maps are available from the South Australian Department of Mines and Energy.

This field guide was initially prepared for an 8th AGC conference excursion organized in conjuction with the SGTSG. The guide has since been extended to include some areas described in a previous excursion guide produced by EPIC Tours for the same region:-

"Archaean to Middle Proterozoic geology of the Southern Gawler Craton South Australia" (Parker, Fanning and Flint, 1981).

There are four field-guides covering the geology of adjacent areas or different aspects of the same area:-

1. "SADME Gawler Ranges Excursion, October 1978" (Blissett and Thomson, 1978) covers the volcanic rocks of the Gawler Range Volcanics and Roopena Volcanics.

2. "Excursion Guide, Tarcoola 1:250 000 Map Sheet Area" (Daly, 1986) covers the Archaean to Middle Proterozoic rocks of the northwest Gawler Craton.

3. "Late Precambrian and Cambrian geology of the Adelaide Geosyncline and Stuart Shelf, South Australia" (Thomson, Daily, Coats, and Forbes, 1976).

4. "Fluvial Sedimentology Field Excursion, Whyalla-Corunna-Depot Creek Area" (Lemon and Gostin, 1983) covers sedimentology of the Corunna Conglomerate and Pandurra Formation.

INTRODUCTION

Eyre Peninsula is the southern part of the Gawler Craton and as such forms part of the Central Australian Orogenic Province (Rutland, 1981) (Fig. 1). The Gawler Craton was cratonised about 1450 Ma and consists of a broad variety of Archaean to Early Proterozoic gneisses, granites, and metasediments, and a range of Middle Proterozoic sediments, volcanics and granites (Parker et al., 1985). To the south it is bound by a series of faults defining the continental margin formerly connected to Antarctica (ca Commonwealth Bay in George V Land); to the east it is bound by the Adelaide Geosyncline now forming the Delamerian Fold Belt with the Torrens Hinge Zone delineating the actual boundary; and to the north and northwest it is bound by the Musgrave Orogenic Domain although the actual boundary is concealed beneath the Officer Basin. Aeromagnetic and gravity data have been variously used to define the latter (Thomson, 1980) and a recent drillhole, into a major linear aeromagnetic anomaly south of Ooldea, (Rankin et al., in press) has delineated a wide mylonite zone.

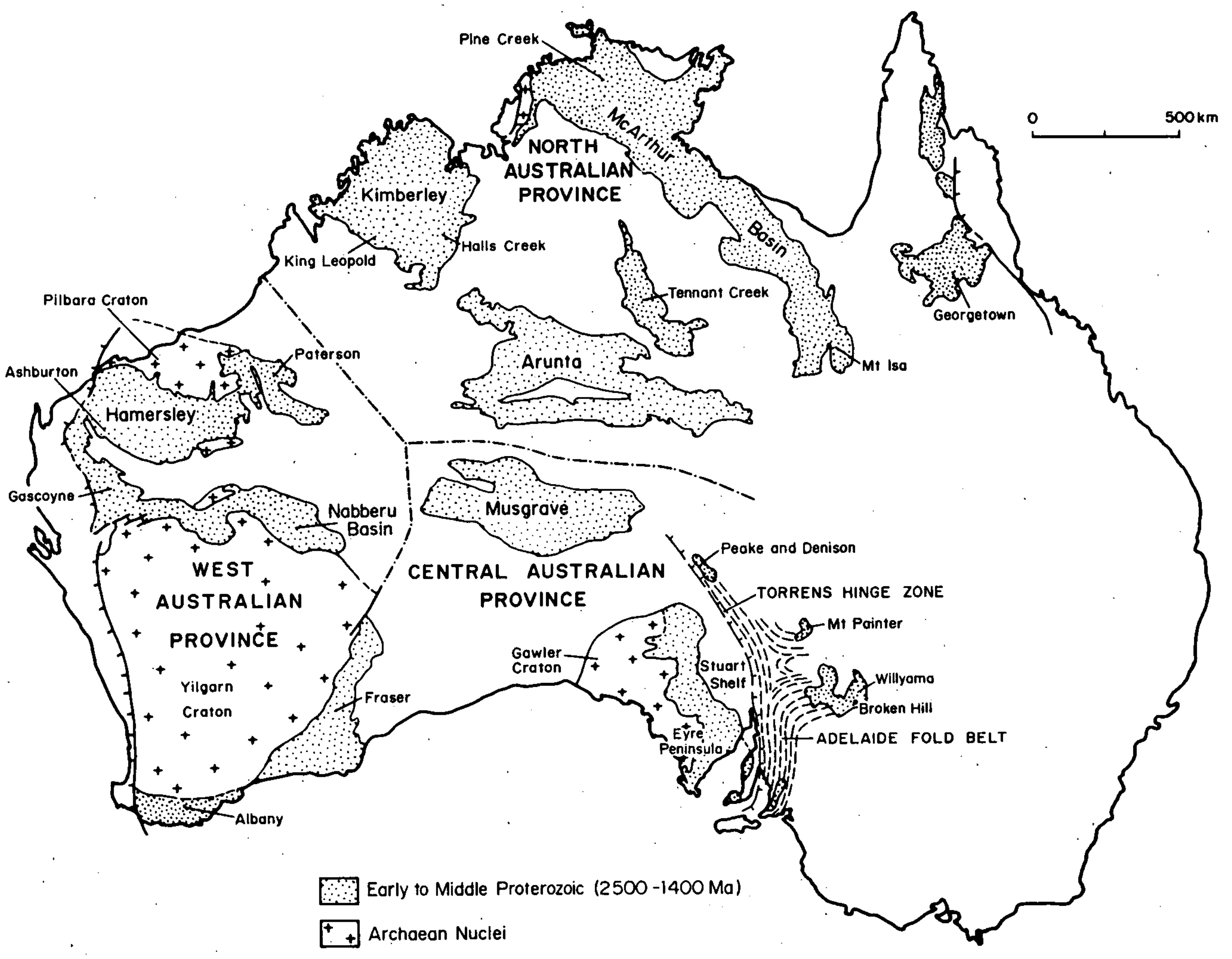


Fig. 1 Setting of the Gawler Craton relative to other Archaean and Early Proterozoic Provinces of Australia, after Webb et al. (1986).

STRATIGRAPHY

Late Archaean/Early Proterozoic Stratigraphy

The Late Archaean to Early Proterozoic basement of southern Eyre Peninsula is known as the Sleaford Complex (Figs. 2 and 3).

It is composed of two distinct elements: a highly metamorphosed supracrustal sequence, the Carnot Gneisses (Fanning et al., 1979 and 1981), and slightly younger, high level granitoids of the Dutton Suite - the Coulta Granodiorite, the Whidbey Granite, and the gneissic Kiana Granite (Webb and Thomson, 1977; Webb, 1978 and 1979; Parker et al., 1981) which are grouped together as the Dutton Suite (Fig. 3).

The Carnot Gneisses form an extensive layered sequence which crops out in the central portion of southernmost Eyre Peninsula. A thinly layered (1-3 cm) garnetiferous quartzofeldspathic gneiss is the dominant lithology. This is often intimately intercalated with layers a few centimetres thick of leucogneiss, biotite-garnet gneiss, hypersthene-bearing felsic gneiss and basic granulite. The hypersthene gneisses (± garnet) are also found as distinct meso-layers within the sequence. Basic granulites typically occur intercalated with, and forming, boudins in the felsic gneisses. Other less abundant but noteworthy felsic lithologies include augen gneiss, plagioclase gneiss, cordierite-garnet gneiss and coarse- to medium-even-grained, garnetiferous granite gneiss. Minor calc-silicate gneiss is also present.

Banded iron formations are notably absent from the Carnot Gneisses. However, there are magnetite-bearing, feldspathic gneisses in drill core on central Eyre Peninsula and there are Archaean banded iron formations in the northern Gawler Craton (Daly et al., 1978).

The high level granitoids of the Sleaford Complex crop out in the southwest Eyre Peninsula and offshore in the Whidbey Islands (see Fig. 2). These granitoids are intrusive into layered quartzofeldspathic gneisses (Wangary Gneisses) on Coffin Bay Peninsula and west of Lake Hamilton, however, their relationship to the Carnot Gneisses has not been observed.

In southwest Eyre Peninsula, Webb (1978) proposed a two-fold subdivision of the granitoids into a northern group which tends to be gneissic with large tabular feldspar phenocrysts (the Kiana Granite), and a southern group which is typically massive and even-grained (Whidbey Granite, see also Webb and Thomson, 1977). This grouping follows closely the distribution of the Rb-Sr analyses of the granitoids on an isochron plot. Those samples assigned to the Kiana Granite define a poorly fitted older isochron, with a calculated age of 2440 ± 58 Ma and initial ratio of 0.7041 ± 0.0043 (13 samples, MSWD = 23.8). The Whidbey Granite samples from the Four Hummocks Islands yield an isochron with little variance outside the expected experimental error (MSWD = 5) and a calculated age of 2337 ± 71 Ma and initial ratio of 0.7019 ± 0.0025 (Webb et al., 1986).

Recent mapping in the Marble Range to Point Drummond region (Fanning, pers. comm.) indicates that there are at least two granitoid types within what was originally assigned to Kiana Granite: an earlier granodiorite phase with abundant mafic xenoliths (the Coulta Granodiorite) and a later, intrusive, tabular feldspar granite phase (the Kiana Granite proper). There are also some granite gneisses of unknown origin. These either represent the basement into which these granitoids intruded, or they could simply be more deformed varieties of the intrusive suite.

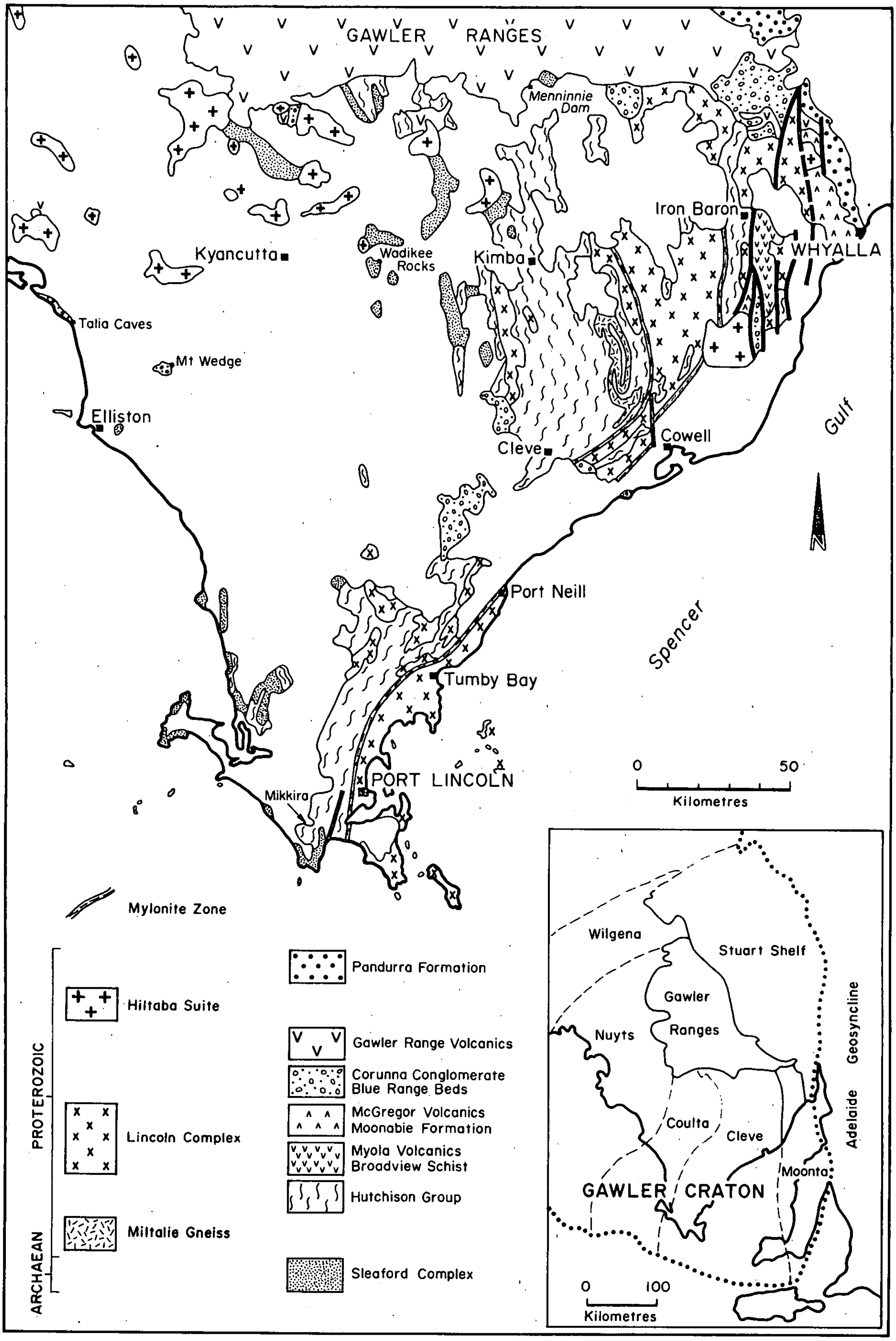


Fig. 2 Regional geological map of Eyre Peninsula with inset showing the tectonic subdivisions of the Gawler Craton.

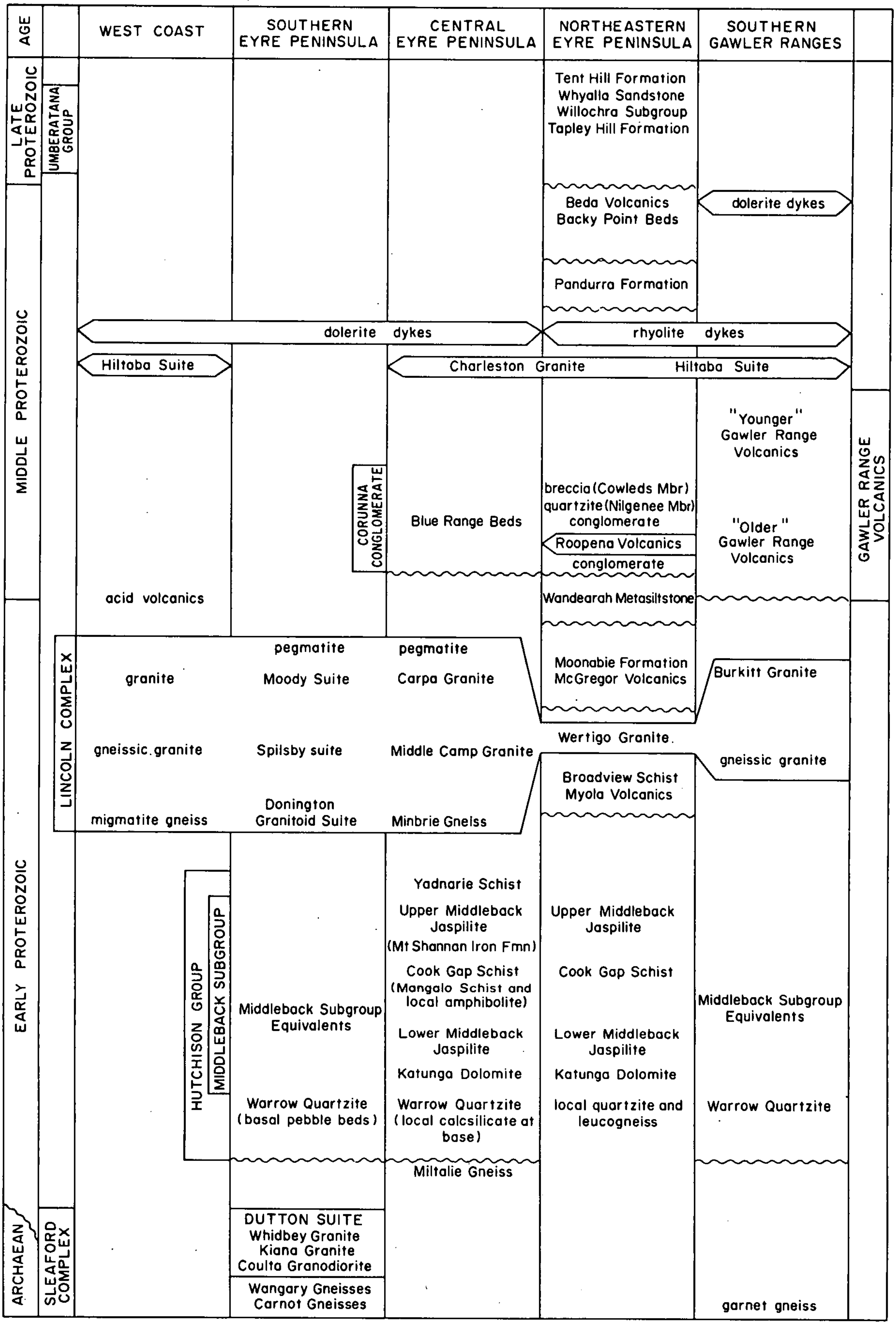


Fig. 3 Precambrian stratigraphy of Eyre Peninsula, modified from Parker *et al.* (1985).

Prominent shear zones are seen at Point Drummond and elsewhere on southern Eyre Peninsula. In these zones massive or foliated granitoids are typically transformed into crenulated schists.

The Sleaford Complex was believed to have been represented in the Cowell - Cleve region by the Miltalie Gneiss, a variably migmatised, grey, medium-grained gneiss of granodioritic composition (Parker et al., 1981). Recent U-Pb zircon dating by Fanning et al. (1986a) has yielded an age of 2014 ± 28 Ma indicating that the Miltalie Gneiss is probably not part of the Sleaford Complex but of some, as yet undefined, younger unit. In the Plug Range area the Miltalie Gneiss is unconformably overlain by Warrow Quartzite hence placing new constraints on the maximum age of deposition of the Hutchison Group in Eyre Peninsula.

Early Proterozoic Stratigraphy

Early Proterozoic metasediments and metavolcanics in eastern Eyre Peninsula are represented by the Hutchison Group, Myola Volcanics, Broadview Schist, McGregor Volcanics, and Moonabie Formation (Fig. 3).

The Hutchison Group is a mixed clastic and chemical sequence consisting of a basal quartzite unit, the Warrow Quartzite, a number of mixed chemical/clastic units dominated by various carbonates and iron formations collectively referred to the Middleback Subgroup, and an upper psammopelitic unit, the Yadnarie Schist (Parker and Lemon, 1982).

At Marble Range and in adjacent areas on southwest Eyre Peninsula, Warrow Quartzite directly and unconformably overlies Sleafordian basement. The quartzites are massive, locally cross-bedded, and near the base contain distinct quartz pebble conglomerate beds. To the east and northeast, sedimentary features have not been observed but north of Cleve, local calc-silicate/dolomite/podded sillimanite gneiss occurs as the base, and in the upper Warrow Quartzite there are a number of pelitic schist interbeds. Parker and Lemon (1982) believe that the Warrow Quartzite represents a fluvial to marginal marine, sandy arkose sequence with fluvial sediments represented in the west and more distal, progradational marine sediments represented in the east.

The pelitic schist interbeds at the top of the Warrow Quartzite represent transgression from east to west across the primeval shelf. They are overlain sequentially by dolomite (Katunga Dolomite), carbonate facies iron formation, silicate facies iron formation and oxide facies iron formation believed to represent progressively deepening water and more distal sedimentary facies. The iron formations are known collectively as the Lower Middleback Jaspilite. Regressive sedimentation following this sequence is represented by the influx of clastic sediments as seen in the Cook Gap Schist. The Upper Middleback Jaspilite represents a second major transgressive cycle and the overlying Yadnarie Schist represents a return to clastic sedimentation. Minor local perturbations superimposed on these macroscopic facies variations are clearly evidenced by the "mesobanding" well developed in the Upper Middleback Jaspilite. In places, alternating dolomite and jaspilitic chert bands several millimetres thick represent minor cyclic pulses superimposed on the regional cycle. The "mesobanding" may be analogous to mesobanding in classical Hammersley Group iron formations (e.g. Trendall, 1976) and "microbanding" within the chert bands would support this analogy.

Acid volcanics are not known on central and southern Eyre Peninsula, but east of the Middleback Ranges there is a prominent sequence of weakly metamorphosed rhyolites and rhyodacites known as the Myola Volcanics. These are associated with conformable gabbroic amphibolites, fine-grained laminated quartzites, and slaty schists of the Broadview Schist. They represent a volcanosedimentary sequence that was erupted/deposited prior to, and deformed by, D_2 of the Kimban Orogeny. A combination of five conventional and one 'single' grain U-Pb analyses of zircons from the Myola Volcanics yields a well defined upper concordia intercept of 1791 ± 4 Ma (MSWD = 1.88) which records the age of crystallisation of the zircons in these volcanics (Fanning et al., 1986a).

A sequence of less deformed volcanics, the McGregor Volcanics, crop out to the south of the Myola Volcanics. These appear to have been folded only by the D_3 Kimban event and have been dated, by U-Pb analyses on zircons, at 1755 ± 12 Ma (Fanning et al., 1986a). They occur as steeply-dipping acidic, welded, ash-flow tuffs derived from melting of a lower crustal source, plus basaltic lava flows derived from a mantle source (Giles et al., 1979).

Overlying the McGregor Volcanics are massive, very poorly-bedded volcanoclastic grits of the Moonabie Formation. These consist of a mixture of acid volcanic and chert clasts in an immature matrix, indicating rapid erosion of the underlying volcanic pile.

Middle Proterozoic Stratigraphy

Middle Proterozoic sediments and volcanics on the Eyre Peninsula are represented by the Corunna Conglomerate, Gawler Range Volcanics, and Pandurra Formation.

The Corunna Conglomerate overlies the Moonabie Formation and was deposited, at least in part, synchronous with eruption of the Gawler Range Volcanics. The conglomerate contains acid volcanic clasts, thin autobrecciated tuffs (locally seen at Tassie Creek) and is also intruded by plugs and dykes of younger acid volcanics. Fanning et al. (1986a) have shown that the Gawler Range Volcanics formed over a relatively short time period at 1592 ± 2 Ma.

In Moonabie Range, the Corunna Conglomerate consists of basal conglomerates overlain by a probable marine, heavy-mineral bearing sandstone which intertongues to the east with a rapidly deposited talus breccia. The breccia coarsens eastwards, with angular clasts up to 0.5 metres in diameter, and was probably deposited from an ancient escarpment perhaps coinciding with the present day Moonabie Scarp (Parker, 1980a).

Probably equivalent to the Corunna Conglomerate are the Blue Range beds. They are unmetamorphosed arenites, outcropping from Gibbon Point on Spencer Gulf, through Blue Range and the "Ningana" area on central Eyre Peninsula to Mount Wedge and Talia Caves on the west coast. This chain of outcrops thus suggests an east-west depositional basin. Various ages have been interpreted, including Middle and Late Proterozoic, Cambrian and Mesozoic. The nomenclature has been equally varied; Blue Range Beds, Blue Range sandstone, Mount Wedge grit, Mount Wedge beds, Talia sandstone and Pandurra Formation all have been used to describe some or all of the outcrops. To overcome this confusion in age and terminology Flint and Parker (1981) have supported Whitten's (1966) interpretation i.e. the arenites should be known as Blue Range Beds and are of Middle Proterozoic age.

The Pandurra Formation is generally regarded as the earliest unit to be deposited on the Stuart Shelf. It unconformably overlies the Roopena Volcanics (basalts) and is, in turn, disconformably overlain by the Beda Volcanics and Backy Point beds, which are regarded as basal Callanna Group in this region (Mason et al., 1978). The Pandurra Formation is a fluvial sequence of arenites, commonly feldspathic, kaolinitic and exhibiting characteristic fossil liesegang weathering bands. It was deposited in a series of northwest-trending grabens, reaches a maximum thickness exceeding 600 metres and was deposited about 1420 Ma (Fanning et al., 1983).

N.B. Contrary to earlier beliefs the Roopena Volcanics are considered to be related to Gawler Range Volcanics (Giles and Teale, 1979) rather than early Adelaidean volcanics. In the Roopena DDH6 drillhole they overlie (though they may interfinger with) conglomerates of Corunna Conglomerate affinity, and these in turn overlie Moonabie Formation grit and "Wandearah-type" metasiltstone.

STRUCTURE

Thomson (1980) and Parker and Lemon (1982) have recognised a number of major tectonic subdivisions of the southern Gawler Craton (Fig. 2). This excursion will traverse two of these subdivisions:

1) the Coulta Subdomain (of western Eyre Peninsula), a cratonic region characterised by scattered, poorly exposed remnants of the older, Archaean to Early Proterozoic basement and probably representing the exposed source region for sediments of the Hutchison Group; and

2) the Cleve Subdomain (of eastern Eyre Peninsula), an Early Proterozoic orogenic belt ("mobile zone") probably representing a shelf or basinal depository for the Hutchison Group prior to its deformation during the Kimban Orogeny.

The Cleve Subdomain may be further subdivided into a western subzone characterised by highly deformed/metamorphosed Hutchison Group sediments, and an eastern subzone characterised by less intensely deformed/metamorphosed acid volcanics and sediments ranging from the Myola Volcanics to the Moonabie Formation. Both subzones are overlain by relatively unmetamorphosed clastics of the Corunna Conglomerate, southern equivalents of the Gawler Range Volcanics, and the Pandurra Formation, perhaps representing the southern extremity of the Stuart Shelf Subdomain.

Two major periods of complex deformation, metamorphism, and plutonism have been recognised in the Cleve Subdomain. They are the Sleafordian Orogeny (SD) which culminated ca 2300 Ma and the Kimban Orogeny which extended from ca 1850 Ma to ca 1700 Ma (Webb et al., 1986; Fanning et al., 1986a; Mortimer et al., 1986). The Sleafordian Orogeny affected basement rocks in both the Coulta and Cleve Subdomains and was a high metamorphic-grade gneiss-forming orogeny accompanied by both mafic (early) and acid (late) plutonism. Because of overprinting by the subsequent Kimban Orogeny, resolution of the exact nature of the Sleafordian Orogeny has not been possible, although some specific plutonic events have been dated.

The Kimban Orogeny locally affected the Coulta Subdomain but was most intense within the Cleve Subdomain. In the western subzone, three main tectonic events can be identified: an early high-grade upper amphibolite to locally granulite facies metamorphic event, M_1 ; a high-grade, isoclinal fold event possibly with associated thrusting, D_2 ; and a lower-grade, open fold event, D_3, with associated development of major mylonite zones, D_m. Principal structural characteristics of each of these events are outlined by Glen et al. (1977), Parker (1978), Parker and Lemon (1982) and Parker et al. (1985) but briefly they are as follows (refer also to Fig. 4):

Event 1 (M_1) - a high-grade fabric-forming event not obviously related to folding. S_1 is always layer parallel and defined by the crystallographic alignment of mica and sillimanite, and by inclusion trails in garnet and andalusite porphyroblasts. S_1 is commonly preserved only in the hinge zones of F_2; elsewhere it is parallel to and indistinguishable from S_2.

Event 2 (D_2) - a high-grade deformational/metamorphic event characterised by very tight to isoclinal folds with pervasive axial-planar fabrics. These fabrics include a strong mica (or amphibole) schistosity, a local segregation schistosity or metamorphically differentiated layering, the platy alignment of elongate or discoid sillimanite aggregates, garnet, feldspar, and quartz, and an often quite strong, quartz-rod lineation. Transposition on a mesoscopic scale is frequent and may also be important on a macroscopic scale.

Event 3 (D_3) - a lower-grade, retrogressive, deformational event characterised by broad, tight to open folds and crenulations, and local, highly-deformed, mylonite zones. By contrast to F_2, D_3 folds generally lack a strong axial-planar schistosity but within the mylonite zones variable fabric development from weak (incipient mica orientation) to strong (slaty ultramylonite) is evident.

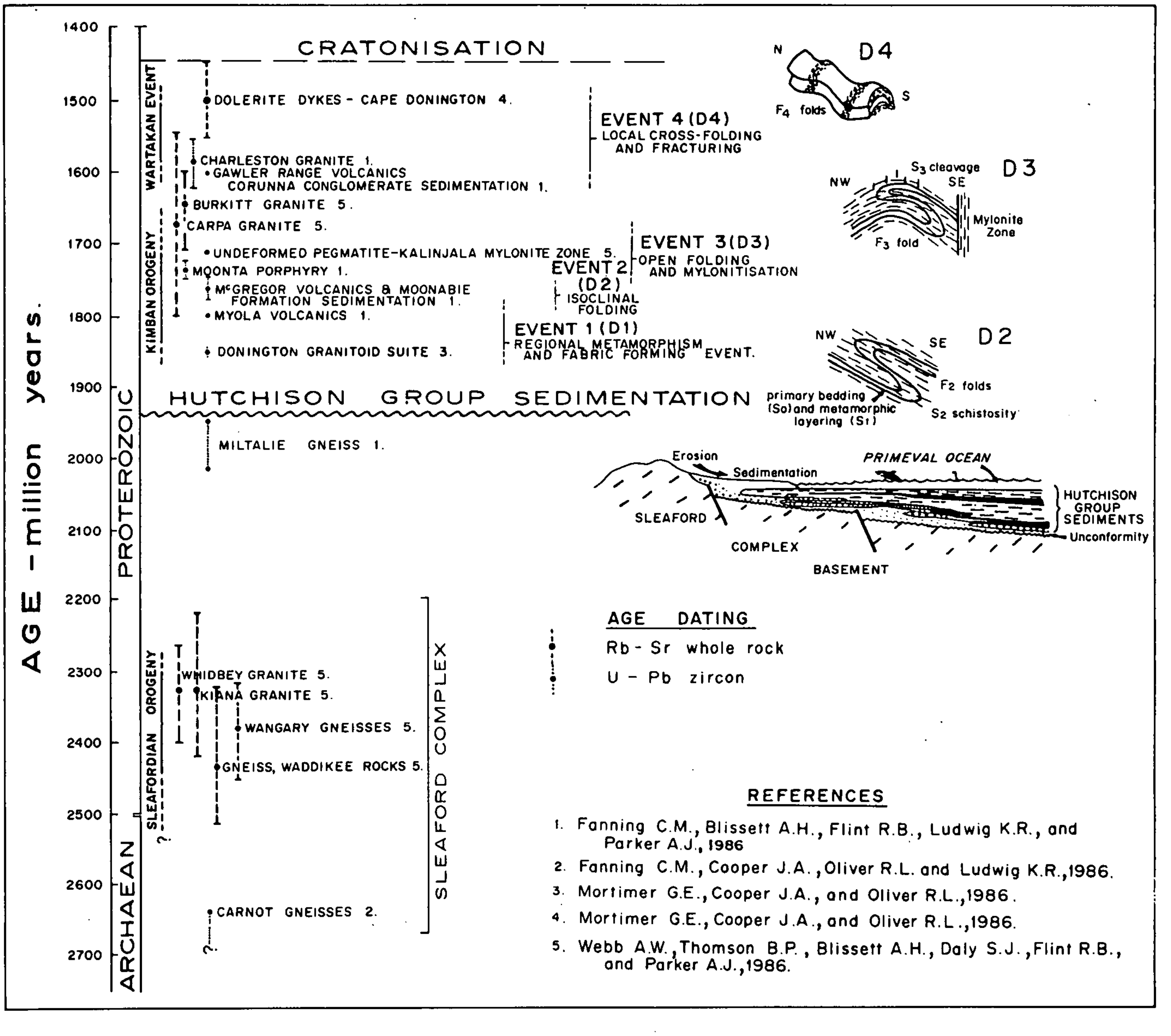


Fig. 4 Geological evolution of Eyre Peninsula.

In the eastern subzone of the Cleve Subdomain the various units record varying degrees of deformation. The Broadview Schist (and Myola Volcanics) is clearly deformed by D_3 folds and crenulations, but while D_3 deformed a strong layer-parallel slaty schistosity, there is no obvious evidence of D_2 or M_1. It is considered, however, that the slaty schistosity is probably related to D_2. McGregor Volcanics and Moonabie Formation grit are locally deformed (open folds) suggesting that they may have been deformed by D_3 prior to Corunna Conglomerate sedimentation. The U-Pb zircon age of the McGregor Volcanics (ca 1755 ±12 Ma) would imply a maximum age for D_3 whereas Rb-Sr isotope measurements, on a little-deformed pegmatite intruding mylonites at Port Neill, give a minimum age of about 1710 Ma for D_3 (Fanning, 1984).

Throughout the Cleve Subdomain there is local evidence for post-D_3 tectonism in the form of cross folding, fracturing and the development of major lineaments and/or shear zones. This event (D_4) is known as the Wartakan Event (Thomson, 1969) and affected not only basement (Sleaford Complex to Moonabie Formation) but also the Corunna Conglomerate. The latter is well displayed at Mt Laura where folding of quartzite, believed to represent Corunna Conglomerate rather than Moonabie Formation as previously mapped (Nixon, 1975; Thomson et al., 1976), occurs about NW-SE trending axes. In the Middleback Ranges and the Cowell-Cleve region, major, often quartz-veined, lineaments are associated with cross-folding about E-W and NW-SE trending axes. Nephrite jade at Cowell was formed during this event.

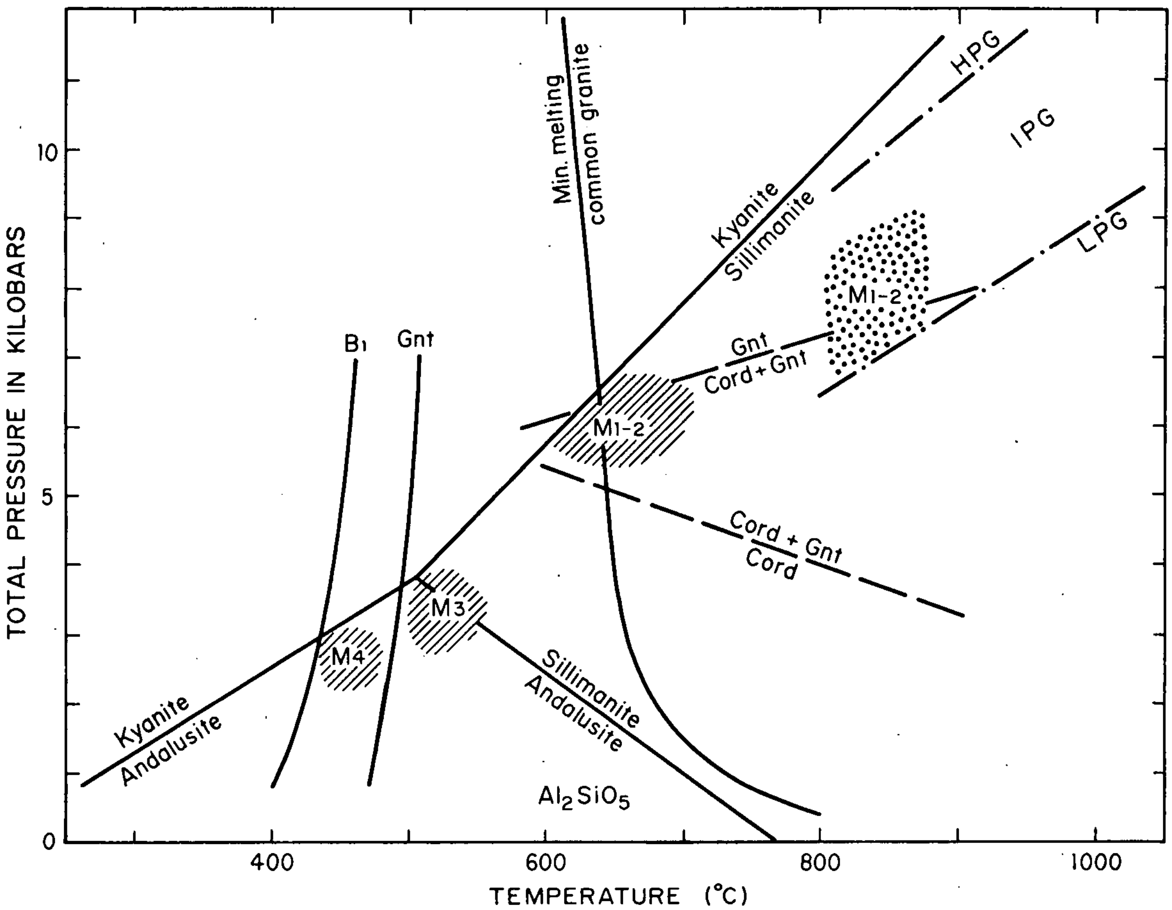
Subsequent tectonism on the southern Gawler Craton is largely restricted to local tensional tectonics such as graben development (e.g. the Pandurra Formation depository trending NW of Whyalla), dolerite dyke intrusion probably during late Pandurra to early Willouran time, and block faulting of graben development during Tertiary to Recent time. Cratonisation of the Gawler Craton is believed to have occurred ca 1500-1450 Ma so these later tensional effects were intracratonic.

METAMORPHISM

In southernmost Eyre Peninsula, the Carnot Gneisses were subjected to a prograde granulite facies event (SM) at ca 2400 Ma (Fanning et al., 1979). Pressure-temperature estimates for this event are 800-900°C at a total pressure of 7-9 Kb (Fig. 5). The muscovite-bearing granitoids (Kiana and Whidbey Granites) intruded grey, layered, quartzofeldspathic gneisses late during the Sleafordian Orogeny. These layered gneisses perhaps reached low to middle amphibolite facies during this time, but this is uncertain due to overprinting, in particular shearing.

The structural and metamorphic development of the Kimban Orogeny is far better preserved due to the stabilisation of the Gawler Craton not long after the orogeny. On central Eyre Peninsula, D_1 and D_2 of the Kimban Orogeny were both high-grade, pervasive metamorphic events of upper amphibolite facies and estimated P-T conditions ca 600-700°C at a total pressure of 5-7 Kb. D_3 and D_4 were retrograde events and lower P-T conditions are evident.

The peak (prograde) metamorphic conditions in southern Eyre Peninsula tend to be higher for the D_1 and D_2 events of the Kimban Orogeny. Mortimer et al. (1979) propose that primary crystallisation of the Donington Granitoid Suite took place at ca 900°C and ca 8 Kb. However, it is likely that only southernmost Eyre Peninsula was subjected to such high-grade conditions early in the Kimban Orogeny.



Al_2SiO_5	Polymorph curves after Holdaway (1971)
Bi	Lowest stability limit of prominent biotite (Winkler, 1976)
Gnt	Lower stability limit for almandine (Winkler, 1976)
Cord + Gnt	Stability field for bulk Fe / (FeO + MgO) ratio of 0·6 (after Currie, 1971)
LPG	Low pressure granulites, olivine + plag. in basaltic compositions (Green and Ringwood, 1967)
IPG	Intermediate pressure granulites, opx + cpx + plag ± qtz (Green and Ringwood, 1967)
HPG	High pressure granulites, presence of garnet in quartz tholeiites (Green and Ringwood, 1967)
M1-2 (hatched)	P-T conditions for Cowell-Cleve area from Parker (1978) M_{1-2}, M_3, M_4 correspond with events D1-4, figure 4
M1-2 (stippled)	P-T conditions for Southern Eyre Peninsula from Bradley (1972) and Fanning (1975)

Fig. 5 Comparison of metamorphic grade in central and southern Eyre Peninsula.

GRANITOIDS

Even during the early days of geological mapping on Eyre Peninsula, a complex plutonic history was recognised. Lockhart Jack (1914) on central Eyre Peninsula and Tilley (1921a) on southern Eyre Peninsula both identified multiple granite intrusions but only in recent times through the application of isotopic dating techniques have temporal relationships been established.

On southern Eyre Peninsula four temporally distinct granitoid suites have been identified: the Late Archaean/Early Proterozoic, Dutton Suite (see Archaean-Early Proterozoic Stratigraphy); the Early Proterozoic, Donington Granitoid Suite; and the Early to Middle Proterozoic, Spilsby and Moody Suites.

The Donington Granitoid Suite comprises a broad spectrum of granitoids which ranges from quartz gabbronorite, through hypersthene granite and late-stage leucogranite (see STOP 3). Mortimer et al. (1979; 1986) consider that these granitoids evolved through a crystal fractionation process and were emplaced ca 1850 Ma during the first tectonic event of the Kimban Orogeny (see Fig. 4). Deformation during later events of the Kimban Orogeny has resulted in the development of a folded gneissic fabric and variable retrogression of the primary pyroxene to hornblende.

The Spilsby Suite crops out in the Sir Joseph Banks Group of islands located in the Spencer Gulf east of Port Lincoln. This suite is composed of hornblende granite and tabular-feldspar granite which are seen to intrude more deformed, megacrystic granite and granite of the Donington Granitoid Suite. The Spilsby Suite is massive to foliated and the presence of shear zones implies emplacement prior to the third tectonic event of the Kimban Orogeny (see Fig. 4).

Granite, porphyritic granite (with tabular feldspar), adamellite, leucogranite and syenite comprise the Moody Suite. These granitoids crop out northwest of Tumby Bay and appear to form a series of related plutons that vary from massive granitoid at Moody Tank to a more foliated porphyritic granite with aligned tabular feldspar phenocrysts at Moreenia (6 km west of Moody Tank, (see Flint and Pain, 1981; and Coin, 1976). The Moody Suite is characterised by xenoliths of gneissic material, and garnet is common in the Moody Tank outcrop. Rb-Sr whole rock measurements do not yield a unique isochron solution but do suggest a possible age of emplacement 1600-1740 Ma (more detailed sampling of individual plutons is proposed). The Moody Suite is considered to have been emplaced late in the Kimban Orogeny.

On central and northeastern Eyre Peninsula there are five main suites of granites and granitic gneisses: an old gneissic suite; an early tectonic suite associated with the first Kimban tectonic event (M_1) and likely equivalent to the Donington Granitoid Suite; two syn-tectonic suites associated with the D_2 and D_3 Kimban events; and a post-tectonic suite correlated with the Hiltaba Suite of the Gawler Ranges.

The old gneissic suite is represented by the medium-grained grey, tonalitic to granodioritic component of the Miltalie Gneiss: As noted earlier, this gneiss was originally considered part of the Dutton Suite but U-Pb zircon dates do not support this correlation. Nevertheless, the Miltalie Gneiss is older than the Hutchison Group.

The Miltalie Gneiss also occurs as rafts or schlieren enclosed within migmatites of the early-tectonic granite suite known in the Cowell region as the Minbrie Gneiss. North of Cowell, but along strike, the Minbrie Gneiss is multiply deformed and contains rafts of both Miltalie Gneiss and Warrow

Quartzite thus dating it as post Hutchison Group but syn- or early Kimban Orogeny. All these granitoids are now gneissic and display a complex deformation history spanning M_1, D_2, and D_3. No dating techniques have yet been successful on this suite, but an age similar to the Donington Granitoid Suite (viz. ca 1850 Ma) is proposed.

Syn-Kimban Orogeny granites are numerous and varied, including the Middle Camp Granite the Carpa Granite and the Wertigo Granite. The Middle Camp and Carpa Granites are associated respectively with the D_2 and D_3 events; the Middle Camp Granite is a foliated biotite granitoid, of granitic to granodioritic composition, deformed by D_3, and intensely veined with pegmatite; the Carpa Granite is a massive, somewhat leucocratic granite and often contains garnet (cf Moody Suite).

The youngest of the granitoids is the post-tectonic suite represented by the circular Charleston Granite pluton. It is a massive, homogeneous, phenocrystic granite and intrudes all lithologies up to and including the Corunna Conglomerate. U-Pb zircon analysis using conventional methods and 'single' - grain analysis yielded an age of 1583 ± 26 Ma (Fanning et al., 1986a) thus representing a minimum age for deposition of Corunna Conglomerate.

NOTES

ROAD LOG AND LOCALITY DESCRIPTIONS

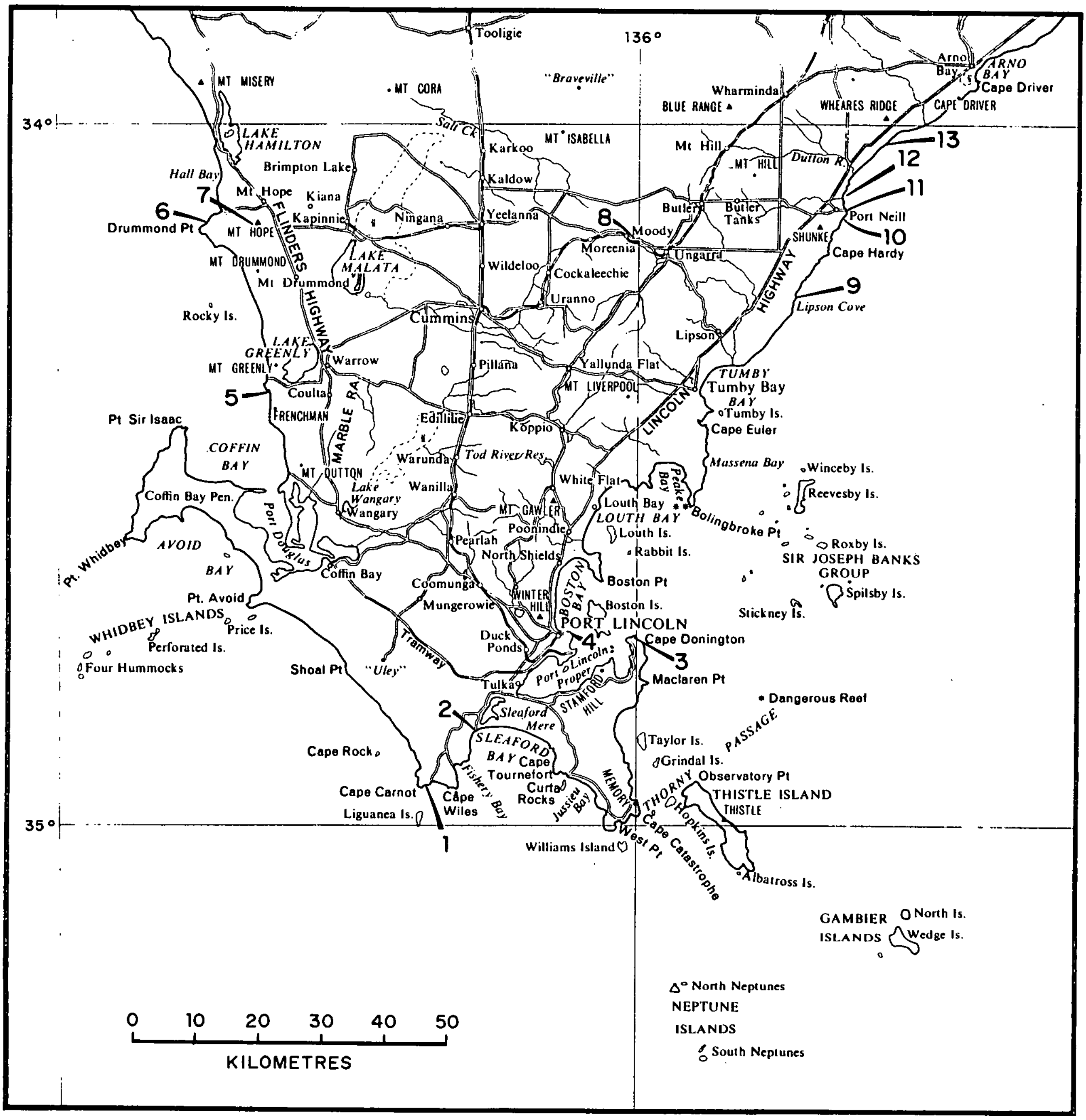


Fig. 6 Locality plan for southern Eyre Peninsula showing excursion stops for days 1-3.

ROAD LOG AND LOCALITY DESCRIPTIONS

DAY ONE

Port Lincoln - Cape Carnot - Cape Donington - Port Lincoln

Distance (kms) - Fig. 6.

0.0 Leave Pt Lincoln and proceed south along Mortlock Tce, from the corner of Tasman Tce, following signs to Whaler's Way (Cape Carnot). N.B. If visiting Cape Carnot (STOP 1) it will be necessary to obtain a key to enter Whaler's Way. Keys may be obtained from Motels, Service Stations and the Travel and Information Center at Port Lincoln.

32.7 At Whaler's Way enter gates and proceed along main dirt road around coastal cliffs to Cape Carnot. The coastal cliffs are composed of a thick sequence of Pleistocene calcreted aeolianite, which directly overlies the Sleaford Complex.

41.5 STOP 1a - CAPE CARNOT (Fig. 7).

The outcrop to be inspected is a small peninsula immediately west of and adjoining Cape Carnot (see Fig. 7).

The condition of the sea at the time will determine whether this particular outcrop can be examined in detail. Four people including a geology student have lost their lives in this area as a result of the frequent large Southern Ocean swells, sometimes referred to as freak waves.

All members of the group are urged to keep close to the tour guides and not wander indiscreetly about the small peninsula.

If time permits or the sea is too rough to cross safely onto the small peninsula, then other aspects of the Carnot Gneisses can be seen at the "Trig Point", Cape Carnot proper, STOP 1b.

Geology

The small peninsula and the adjacent point to the east (see Fig. 7 and Plate 1) form the type area for the Carnot Gneisses. The major lithologies on the small peninsula are a thinly (1-3 cm) layered garnetiferous quartzofeldspathic gneiss and a garnetiferous augen gneiss.

Layered garnet gneiss is the overall dominant lithology of the Carnot Gneisses. In thin section it has a granoblastic texture; the main mineralogy consists of perthitic potash feldspar, plagioclase (oligoclase-andesine) and quartz with lesser amounts of garnet (almandine-pyrope) and biotite. Minerals present in accessory amounts are zircon, monazite, magnetite, ilmenite, rutile, apatite and traces of sillimanite and spinel (hercynite?). Secondary-alteration minerals include sericite and chlorite.

The augen gneiss consists of two types, either with the potash feldspar augen orientated parallel to the regional foliation direction, or in random orientation. Mineralogically it has a similar composition to the layered garnet gneiss, except for the presence of some microcline in the augen gneiss. An inequigranular granoblastic texture is observed in thin section.

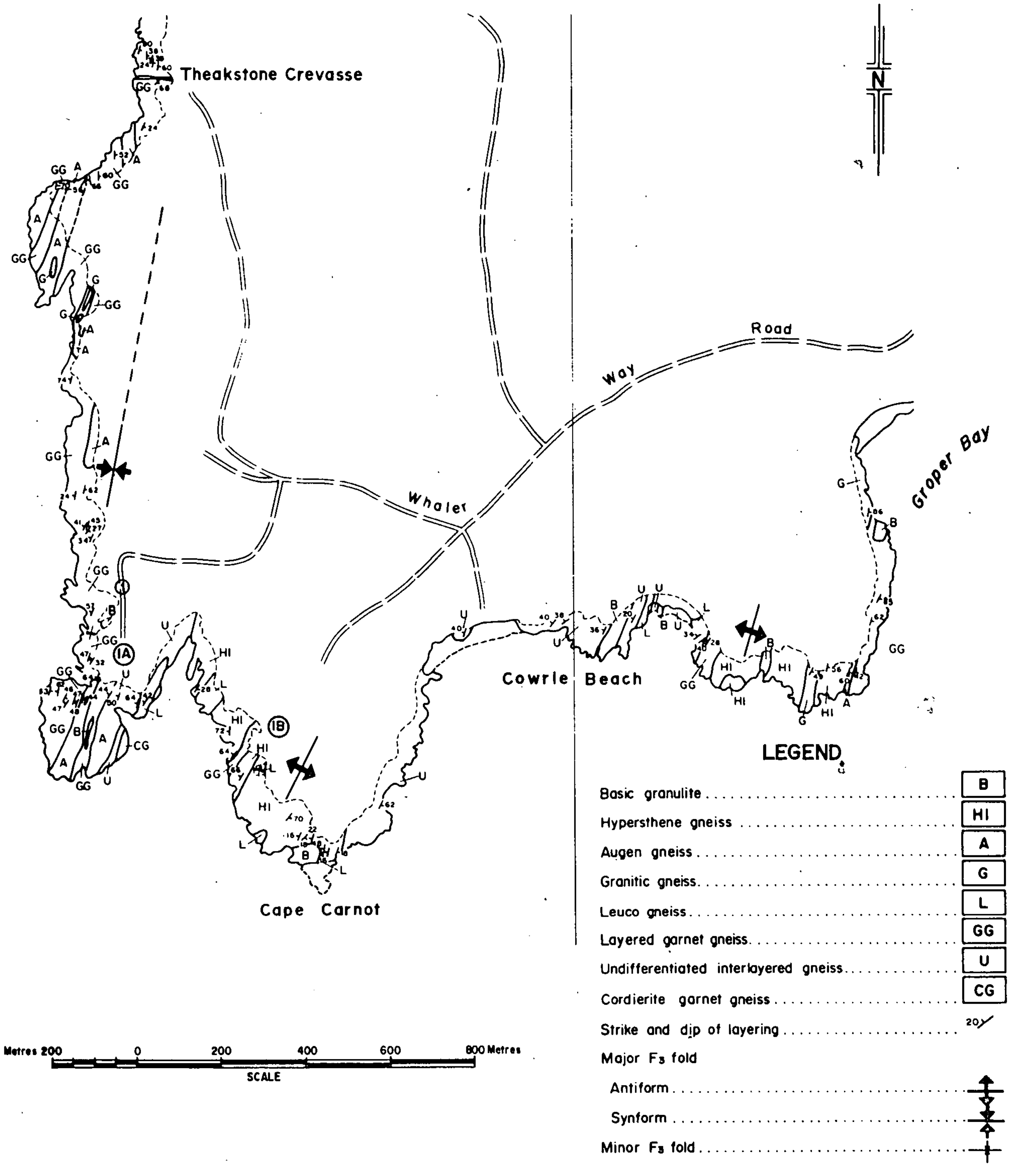


Fig. 7 Geological plan, Cape Carnot area, STOP 1. From Fanning et al. (1981).

Biotite garnet gneiss, characterised by high volume percentages of biotite (10-20%) and garnet (up to 25%), is present as small isolated pods in both the augen and layered garnet gneiss. This lithology may contain cordierite and sillimanite and is interpreted as representing relict pods of pelitic sediment.

The easternmost end of the small peninsula consists of cordierite-garnet gneiss. This is a pale fawn-coloured rock marked by the presence of dark green cordierite porphyroblasts. In thin section the cordierite coexists with garnet, sillimanite, and quartz (± spinel). Potash feldspar and plagioclase are also present, as are the typical accessory minerals mentioned above. Minor pegmatite veins associated with this rock-type contain up to 20% sillimanite.

Between the augen gneiss and cordierite garnet gneiss there is a 15 metre wide zone of intimately interlayered gneisses (shown as undifferentiated interlayered gneisses in Fig. 7). The gneisses that comprise this zone vary in compostion, from leucogneiss (potassium feldspar, plagioclase, quartz with minor biotite and garnet) through layered garnet gneiss (as above) and hypersthene gneiss (potassium feldspar, plagioclase, quartz, hypersthene, garnet ± biotite) to basic granulite (orthopyroxene, clinopyroxene, plagioclase, opaques). They are intimately interlayered to the extent that the lithologies change randomly across strike over a distance of a few centimetres. The compositional layering parallels the local foliation trend.

The basic granulites are thought to be originally intrusive into a principally felsic sequence of supracrustals, although all contacts observed within the Carnot Gneisses are essentially concordant. Deformation has resulted in local boudinaging and wrapping of the layering of the felsic gneisses around the larger (1-2 m thick) basic boudins. Thinner (20-30 cm thick) basic bodies have been involved in F_2 and F_3 folding plus shearing (see Fig. 4 and Plate 2). The basic granulites are equigranular, granoblastic, and foliated as defined by the parallel orientation of elongate ferromagnesian grains. Mineralogically they consist of orthopyroxene (hypersthene), clinopyroxene (diopside), plagioclase (labradorite) and opaques (magnetite and ilmenite). Variations are provided by the presence in some rocks of hornblende or biotite, or both. The hornblende is considered to be a retrograde mineral, replacing pyroxene. Quartz where present is only a minor constituent. Traces of potash feldspar, zircon and spinel have also been observed.

Hypersthene gneisses, similar to those grouped as intimately interlayered gneisses above, also form prominent macro-layers in the sequence. At the "Trig Point", Cape Carnot proper (see Fig. 7), a 200-300 metre wide band of hypersthene gneiss can be seen in the hinge zone of a major F_3 structure. In places, particularly along the western margin of this band, disrupted pods and blocks of basic granulite occur within the hypersthene gneiss suggesting that it is a late-stage development in the sequence, possibly intruded contemporaneously with granulite metamorphism of the Sleafordian Orogeny.

Two suites of pegmatite are present at Cape Carnot; segregations and veins.

Pegmatitic segregations are developed approximately parallel to the layering and consist of extremely coarse potash feldspar augen, up to 20 centimetres in length. The contacts between the segregations and the surrounding gneisses are diffuse, implying that they were formed from the gneisses. It is suggested that shears parallel to the layering during the D_3 deformation (cf Fig. 4) may have acted as loci for the development of coarse feldspar segregation.

PLATE 1. View of the Cape Carnot Peninsula, looking south.
CG - cordierite + garnet gneiss
U - undifferentiated interlayered gneiss
A - augen gneiss

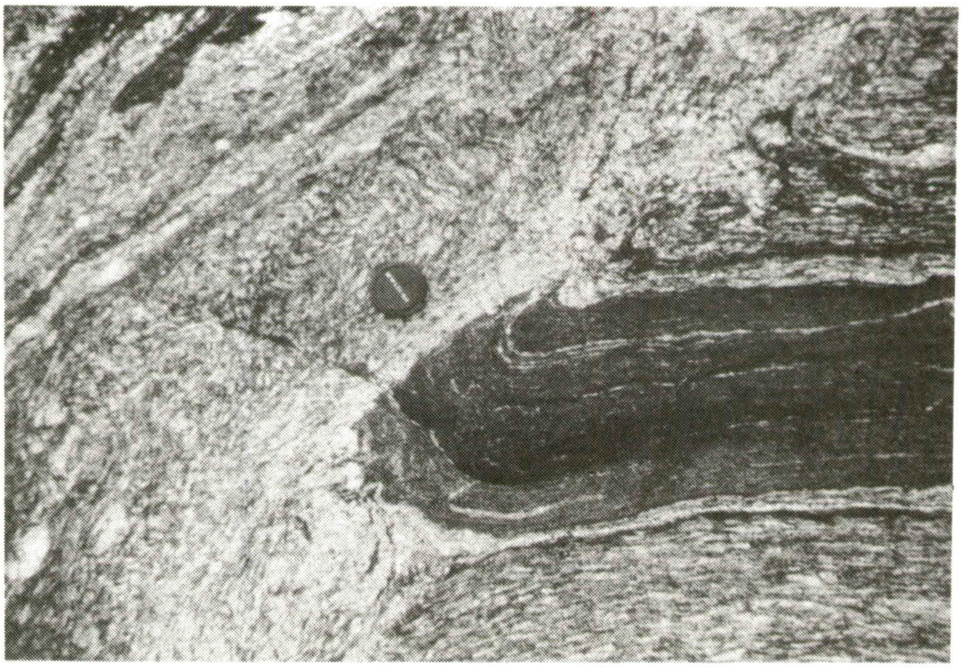


PLATE 2. Narrow mafic layer within deformed augen gneiss truncated by D_3 dextral shears. Note folding within both gneiss and mafic layer. Lens cap is 50 mm wide. Cape Carnot Peninsula.

Pegmatitic veins are ubiquitous and are characteristically developed in axial planes of F_4 crenulations or flexures of the layering. Aplite veins with mafic rims (biotite) are also discordant younger features.

Prograde metamorphic (SM) conditions correspond to those of the intermediate pressure granulite facies. Comparison of the observed mineral assemblages with experimentally determined mineral stability curves indicates that prograde metamorphic conditions probably reached 700-900°C and a pressure of 7-9 kilobars. Coexisting pyroxene geothermometry for the basic granulites indicates temperatures in the range 800 to 860°C (using Wood and Banno, 1973). However, average garnet-biotite geothermometry for the acid gneisses yields lower temperature estimates in the range 600 to 690°C using Perchuk (1977) and 595 to 710°C using Thompson (1976). In places biotite is seen to be oriented parallel to the axial plane of F_3 folds, thus the lower temperature estimates for garnet-biotite pairs are most likely a result of later metamorphic events.

Geochronology

Fanning et al. (1986b) report Rb-Sr total rock measurements for each of nine gneissic compositions, including the basic granulites. The cordierite garnet gneiss yields the only near-perfect fit isochron with a calculated age of 2412 ± 72 Ma and IR of 0.7060 ± 0.0008. An intrusive hypersthene gneiss sampled from near the "Trig Point" records a slightly, but not significantly younger total rock age of 2300 ± 131 Ma with an IR of 0.7092 ± 0.0017. A "grand" regression for 44 of 55 samples analysed at Cape Carnot, excluding the hypersthene gneiss, approximates the well fitted cordierite + garnet gneiss isochron, with an age of 2416 ± 49 Ma and IR of 0.7065 ± 0.0011.

It has previously been proposed by Fanning et al. (1979 and 1981) that this age reflects the time of granulite facies metamorphism during the Sleafordian Orogeny. The other 11 samples plot above this "regional" isochron and it is inferred that this is due to either the presence of older crust not reset at the time of the granulite event, or the presence of heterogeneous IR's or both. In detail, individual isochron plots for a granitic gneiss, intimately interlayered gneisses, and biotite garnet gneisses show sub-parallel younger cross trends, probably a consequence of partial resetting during the Early Proterozoic Kimban Orogeny. This resetting is more apparent on the mineral scale, where total rock-potassium feldspar joins yield a range in ages from 1500-1720 Ma, and biotites record ages of ca 1515 Ma. A high IR of between 0.7060 - 0.7066 for the majority of the Carnot Gneisses is supported by analyses of the basic granulites.

Fanning et al. (1986b) also report U-Pb zircon analyses on four gneissic compositions previously analysed by Rb-Sr geochronology. The augen gneiss and layered garnet gneiss appear to define a crude two-stage chord and it may be interpreted that zircon crystallisation in these gneisses occurred at ca 2300 Ma. Similarly the cordierite + garnet gneiss analyses may indicate crystallisation at ca 2550 Ma. However these chords are characterised by scatter much in excess of experimental error which suggests a more complex lead loss history. The hypersthene gneiss discordia indicates zircon crystallisation at ca 2640 Ma and lead loss at ca 1756 Ma during the Kimban Orogeny. This chord is appropriately aligned above the most concordant points of the other three gneisses and may provide the key to the complex U-Pb systematics. It is argued by Fanning et al. (1986b) that the augen layered garnet, and cordietite garnet gneisses are also about 2640 Ma in age, and that their zircons lost ca 60% and 15% respectively of their original lead during the Kimban Orogeny. These zircons then remained closed until the Late Palaeozoic when episodic lead loss occurred giving rise to their present alignment on the condordia diagram. The hypersthene gneiss zircons do not appear to record this younger event.

Follow road back towards Whaler's Way Entrance.

Turn right into the tourist stop labelled "Trig Point".

42.5 Follow the path towards "Flinders Crevasse and Cave".

STOP 1b - CAPE CARNOT "TRIG POINT" (Fig. 7).

Proceeding east along the coast a 200-300 metres wide band of thinly layered hypersthene gneiss crops out in the hinge of a major F_3 antiform (see Fig. 7). An intrusive origin is inferred from the disrupted blocks (and pods) of basic granulite to be seen within the hypersthene gneiss immediately east of Flinders Crevasse.

Further east, layering in the gneisses is flat-lying in the hinge of the broad F_3 structure. In places small-scale F_2 folds can be seen refolded by F_3.

The hypersthene gneiss from the "Trig Point" yields a U-Pb zircon age of 2637 ± 21 Ma and its U-Pb systematics provide the key to the complex U-Pb and Rb-Sr systematics recorded by the Carnot Gneisses (Fanning et al., 1986b).

Follow road back toward Port Lincoln.

64.9 Turn right at T-junction and proceed to Naval Mine Monument carpark on coast.

68.2 STOP 2 - SLEAFORD BAY (Fig. 8).

In the northwest corner of Sleaford Bay is an 800 metre long coastal platform exposure of Hutchison Group metasediments unconformably overlying Sleaford Complex gneisses (Fig. 8). This section was first described by Tilley (1920, 1921 a and b).

The Hutchison Group can be seen to abut against the Carnot Gneisses at the western end of a sandy beach (Fig. 8). Here the Carnot Gneisses consist of a layered hypersthene gneiss that is characterised by euhedral garnets and garnet-quartz symplectitic intergrowths. The main mineralogy of this gneiss is plagioclase, quartz, hypersthene, garnet, biotite, and variable diopside. Layering of the gneiss strikes approximately 010° and dips steeply to the west.

Immediately east of this gneiss the Warrow Quartzite outcrops, marking the base of the Hutchison Group. Clay-rich bands emphasize the layering of the quartzite which is roughly parallel to the layering in the adjacent Carnot Gneisses. There is a general structural concordance between these two major rock groups in this area.

By analogy with the Hutchison Group sequence of Parker and Lemon (1982), the Sleaford Bay section youngs to the east, away from the quartzite. Near the eastern end of the sandy beach interlayered ferruginous bands, garnet-quartz-feldspar gneiss, forsterite marble and granitic sills crop out in the intertidal zone.

The forsterite marble is a thin, but distinctive lithology and can be equated with the Katunga Dolomite. The adjacent ferruginous bands have been intensely altered and in places silicified, but remnant fine-scale quartzose and iron-rich bands suggest that the primary rock was a banded iron formation.

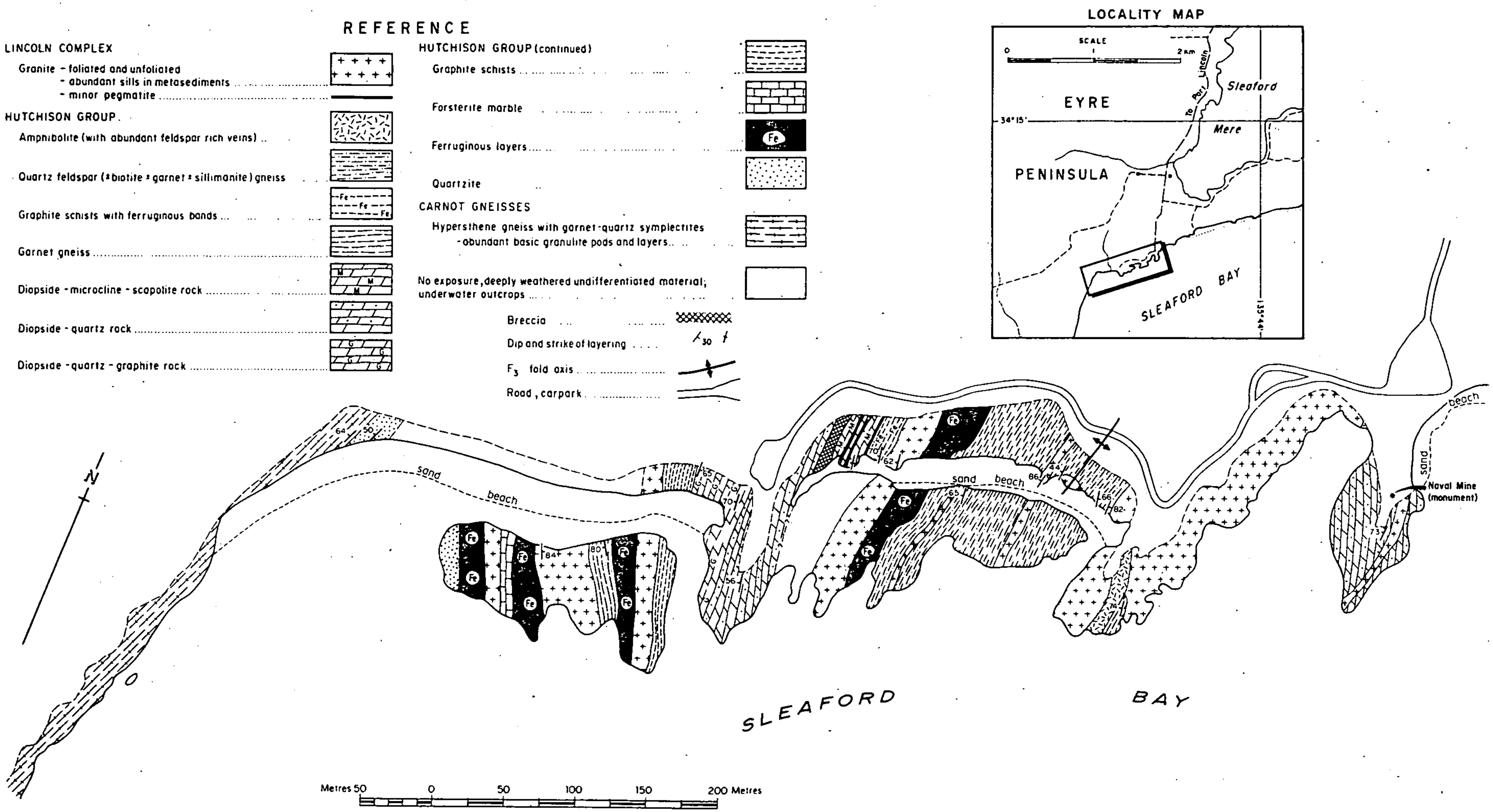


Fig. 8 Geological map of the northwest corner of Sleaford Bay. After Fanning et al (1982).

The prominent headland at the eastern end of this sandy beach consists of quartz-diopside rock. Interlayered bands of pure diopside and quartz, with variable thicknesses, strike roughly north-south and dip steeply to the west. On the western side of this headland, there is a gradation from graphite schist through graphite-quartz-diopside rock to quartz-diopside rock. Hornblende occurs as a minor retrograde metamorphic phase and is also found in tension gashes, perpendicular to the layering, that resulted from the boudinage of the quartz-diopside rock during the D_3 event (see Fig. 4).

Progressing further east, into the next small bay, there is a 10 to 20 metre wide zone of interlayered scapolite-microcline-diopside rock, diopside rock, graphite schist, thin ferruginous bands and garnet quartz feldspar gneiss. The garnets are visibly zoned in hand specimen corresponding to more grossular-rich cores and more almandine-rich rims. A ferruginous layer immediately to the east marks the upper boundary of the Lower Middleback Jaspilite equivalents in this section (see Fig. 3 and Parker and Lemon, 1982).

Layered quartzofeldspathic gneisses comprise the remainder of the Hutchison Group exposed at Sleaford Bay. These gneisses have variable amounts of biotite, garnet and sillimanite, most probably a reflection of variations in pelitic content in the original arkosic sediment.

Foliated and unfoliated granite form the next headland to the east. Minor but distinctive constituents of the granite are hornblende, biotite, sphene, apatite and sillimanite. The intrusive nature of the granite is evidenced by the inclusion of a several metre wide tract of garnet - quartz - feldspar gneiss and a low-angle crosscutting relationship with quartz-diospide rock (repeated due to folding) near the Naval Mine (monument). Granitic sills observed within the Hutchison Group metasediments are mineralogically similar to this larger granite body, and are considered contemporaneous.

Return north along road toward Pt. Lincoln.

78.0 Turn right along dirt road to Cape Donington (sign posted).

106.7 STOP 3 - CAPE DONINGTON (Fig. 9).

Two sections of coastal outcrop of early Lincoln Complex granitoids of the Donington Granitoid Suite are to be examined on the northern tip of the Cape Donington Peninsula (see Fig. 9). The first section covers Cape Donington proper, in the immediate vicinity of the lighthouse. Part of this area was mapped by Flook (1975). The second section is a westward extension of the first, across a small beach towards Cape Colbert (see Fig. 9). Both sections are part of G.E. Mortimer's Ph.D. thesis area and we gratefully acknowledge his generous assistance in the preparation of the locality description.

<u>A. Cape Donington</u>

From the lighthouse walk down a small track to the eastern cove. The outcrop consists of a mafelsic megacrystic granite gneiss,* characterised by the presence of ovoid plagioclase and orthoclase megacrysts, on average 2-3 cm in diameter but ranging up to 4 cm. These feldspars are typically zoned and the cores can be seen in hand specimen to contain mafic minerals. High modal

*(Note: A more leucocratic megacrystic granite gneiss (i.e. lower mafic content than is seen here) is one of the dominant lithologies that outcrops along the eastern coast of southern Eyre Peninsula, extending from Cape Tournefort to Port Neill.)

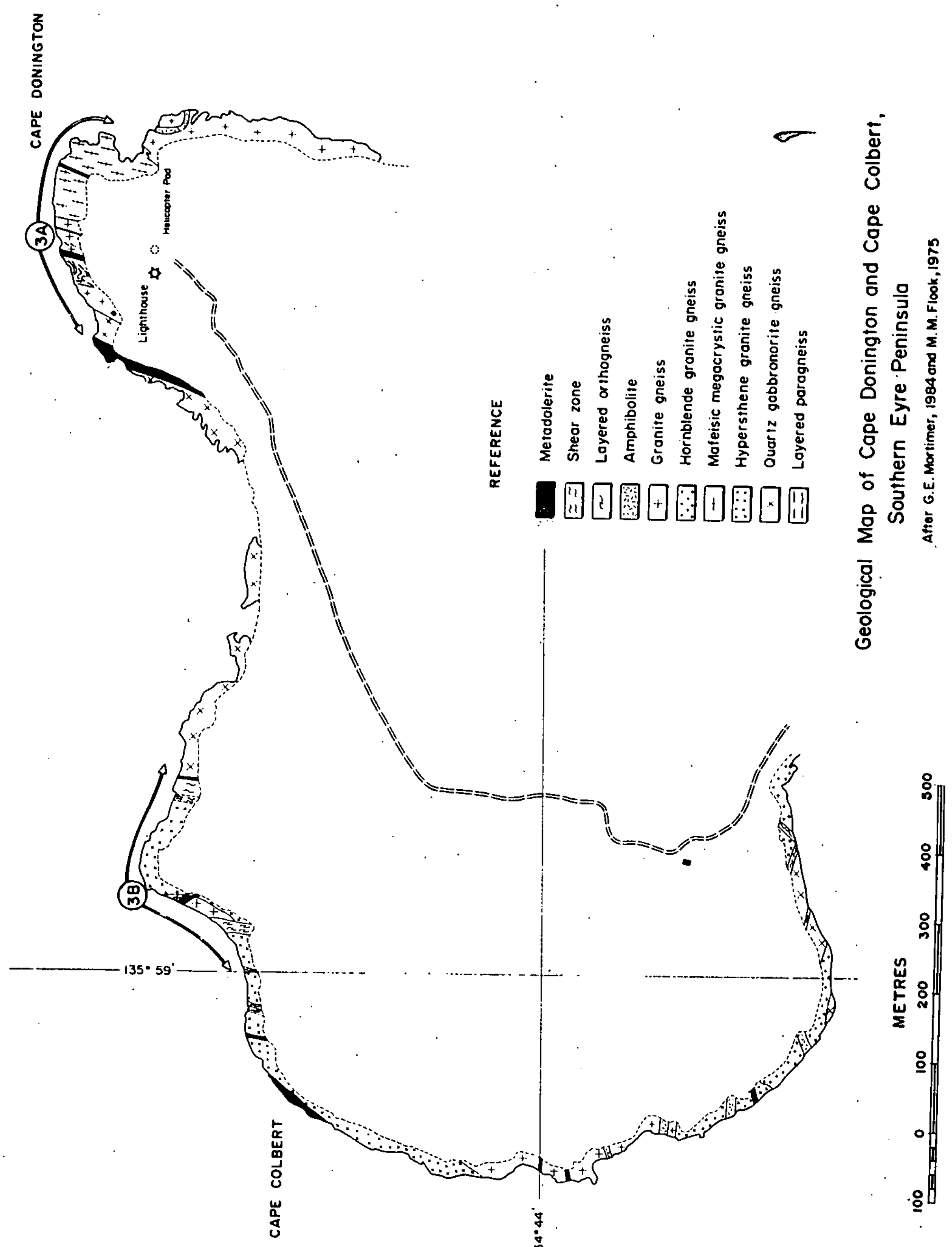


Fig. 9 Geological map of Cape Donington and Cape Colbert. After Mortimer (1984) and Flook (1975).

percentages of hornblende and biotite result in the distinctive dark colouration of this gneiss. Orthopyroxene is generally absent, although there are a few patches which do contain hypersthene (G. Mortimer, pers. comm.). Xenoliths of mafic material, possibly earlier dolerites or amphibolites, are relatively common.

The term gneiss is applied to the megacrystic granite as there is a well developed foliation, striking approximately 010-020°. There are also zones of very intense deformation (shears) which strike at a low angle to the local foliation direction and which are of the order of 10-20 cm in width (Plate 3).

At least two generations of basic rocks crop out along this coastal section; amphibolites and metadolerites (Plate 4). The relative age of the amphibolites is not clear although it may be suggested that the relatively simple planar boundaries of the amphibolites at Cape Donington imply that they were originally intrusive into the granitoids. The main phases of the Kimban Orogeny have affected the amphibolites since they are foliated and in places involved in F_3 folds (see Fig. 4). Mineralogically they consist of plagioclase and hornblende with varying amounts of pyroxene, depending on the extent of retrogression. The metadolerites have ophitic textures and are essentially undeformed.

The mafelsic megacrystic granite gneiss is intruded by an even-grained granite gneiss which in places is very intensely deformed. Principal minerals of the granite are microcline, quartz, plagioclase, and biotite. Accessories include sphene, apatite, zircon, and opaques. In thin section the texture is typically granoblastic seriate, however quartz forms elongate flattened ribbons in areas of intense deformation.

A folded tract of paragneisses consisting of calc-silicate, semi-pelitic and quartzo-feldspathic compositions, occurs as a xenolithic inclusion within the even-grained granite gneiss at this locality. The semi-pelitic and calc-silicate layers have a strong compositional banding and Flook (1975) reports a variety of assemblages:

Quartz - Plag - Diopside - Scapolite (sphene, magnetite, allanite)
Quartz - Plag (sphene)
Quartz - Plag - Hornblende (sphene, allanite, opaques, zircon)
Quartz - Sillimanite - Muscovite - Hornblende - Zoisite
Quartz - Sillimanite - Muscovite
Quartz - Plag - Biotite - Garnet - Sillimanite (fibrolite).

Folded quartzofeldspathic zones containing microcline megacrysts are also present and it is more likely that these have formed through metamorphism and deformation of the granitoids. Nevertheless the paragneisses undoubtedly represent inclusions of pre-existing crust at the time of intrusion of the granitoid suite.

Immediately west of these folded paragneisses, xenoliths of quartz gabbronorite gneiss (Mortimer et al., 1979; cf mafic charnockite of Flook, 1975) occur within the even-grained granite gneiss. The main body of the quartz gabbronorite gneiss crops out immediately to the west and extends around the point and into a bay where we shall examine its western margin. Mortimer et al. (1979) consider this lithology to be the earliest intrusive phase of their Donington Granitoid Suite. Mineralogically it consists of zoned plagioclase phenocrysts (oligoclase), augite, hypersthene, biotite, and quartz. Accessories include zircon, apatite, magnetite, and ilmenite; there is also minor potash feldspar. Blue-green hornblende replacing pyroxene is a result of later retrogression.

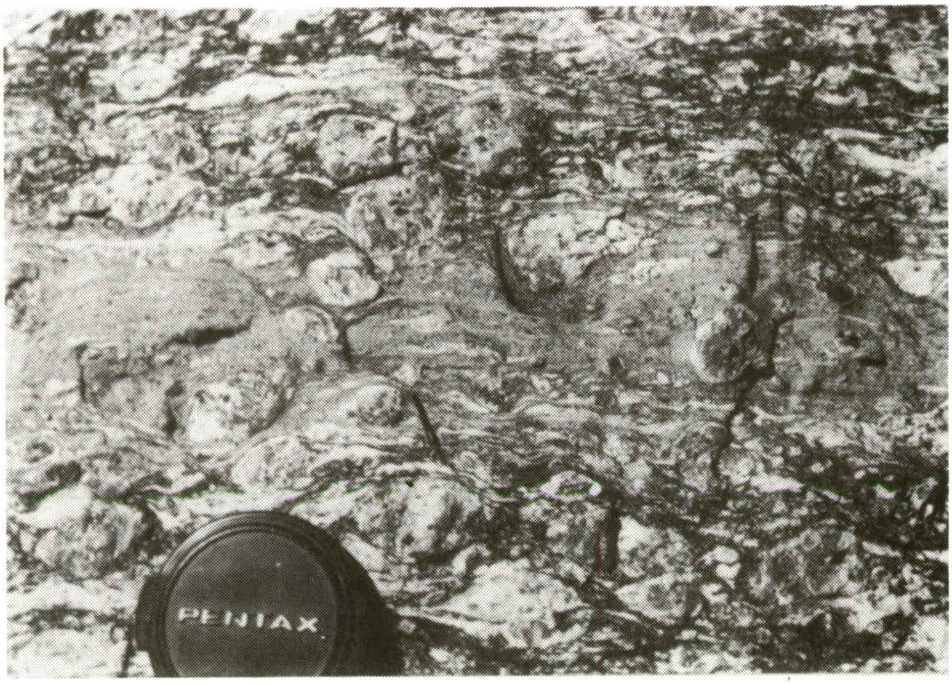


PLATE 3. Narrow band of ultramylonite developed within the megacrystic granite gneiss, Cape Donington. Lens cap is 50 mm wide.

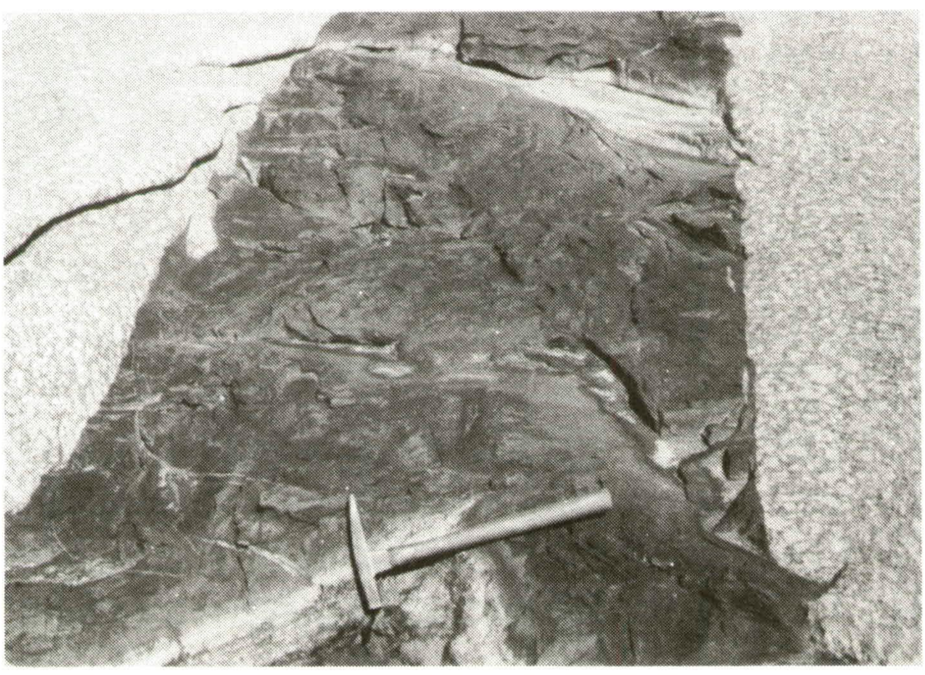


PLATE 4. Metadolerite dyke intruding the megacrystic granite gneiss, Cape Donington.

B. Cape Colbert

This traverse commences at the western boundary of the quartz gabbronorite examined above. The western boundary of this lithology is marked by very intense deformation, grading into a mylonitic zone of approximately five metres in width. This shear zone obscures the contact between the quartz gabbronorite gneiss and a more acidic, hypersthene granite gneiss which is similarly intensely deformed adjacent to the shear. Some granite gneiss may also have been incorporated into the shear.

The hypersthene granite gneiss* consists of primary igneous plagioclase, hypersthene and orthoclase phenocrysts in a partly recrystallised matrix of orthoclase, plagioclase, quartz and biotite. Minor hornblende may also be present. Accessories include apatite, magnetite, ilmenite and zircon. Xenoliths of mafic material can be seen in some outcrops.

The hypersthene granite gneiss is intruded to the west by an even-grained granite gneiss of similar composition to that seen at Cape Donington. Note here the absence of the paragneiss xenoliths. A branching metadolerite dyke can be seen to truncate the granite gneiss - hypersthene granite gneiss contact.

A few metres to the southwest, this same contact is also truncated by a band of well-layered quartzofeldspathic gneiss. The quartzofeldspathic gneiss gradually merges to the west with a more deformed variety of the even-grained granite gneiss and it is suggested that the well layered gneiss was tectonically produced from the surrounding granite gneiss. Thin biotite-rich interlayers occur within the well layered gneiss and these display prominent tight to isoclinal folding. It is possible that the biotite-rich layers were originally amphibolites that have been subjected to intense retrogression during the deformation that formed this well layered zone.

Further west there is another example of the production of a tectonic layering within the granitoids. Through increasing deformation over a distance of two to five metres across strike, quartzofeldspathic veins within the hypersthene granite gneiss are strung out into parallel orientation, to the extent that a new layering is produced.

Hence in the Cape Donington - Cape Colbert area there is evidence for both layered paragneiss and layered orthogneiss. The layered paragneisses formed by deformation of original sedimentary xenolithic tract, whereas the layered orthogneisses are derived from deformation of the intrusive granitoid suite.

C. Timing and conditions of formation of the Donington Granitoid Suite

On the basis of applications of the Wood and Banno (1973) pyroxene geothermometer, and using the composition of orthopyroxene in the granite gneiss, Mortimer et al. (1979) suggest that primary magmatic crystallisation took place at around 8 kb pressure and in excess of 900°C. They also suggest that a subsequent lower grade amphibolite facies retrogression has affected

*(Note: The hypersthene granite gneiss is a prominent lithology within the Donington Granitoid Suite. It occurs as extensive outcrops in the Memory Cove - Cape Catastrophe area, and is also to be seen at Stamford Hill, Boston Island and Cape Euler. The name Memory Cove Charnockite (MCC) may be used for this lithology.)

the suite. The degree of this retrogression is variable, although Mortimer et al. (1979) note a general increase in extent from east to west. This retrogression is typically represented by the replacement of pyroxenes by hornblende.

Seventeen Rb-Sr whole-rock measurements of essentially hypersthene-bearing gneiss varieties indicate a Model 3 isochron age of 1818 ± 13 Ma with an initial ratio of .7055 ± .0005 (MSWD = 2.54) (Mortimer et al., 1979). While more recent U-Pb zircon data yield a precise age of crystallisation and emplacement of the Donington Granitoid Suite of 1843 ± 2 Ma (Mortimer et al., 1986). From calculations using the isochron age and initial ratio, and the mean Rb/Sr ratios, it can be inferred that this suite represents a new addition of material to the continental crust during the Early Proterozoic. It is not derived from the reworking of older continental crust such as the Hutchison Group or Sleaford Complex. However, xenoliths within the suite do indicate the presence of continental crust in the area prior to emplacement.

148.1 Return to Port Lincoln. From the corner of Mortlock Tce head east along Tasman Tce to London St.

150.2 Follow map (Fig. 10) to the boat ramp car park at the northwestern end of STOP 4.

STOP 4 - KIRTON POINT (Fig. 10).

"The most characteristic rock of the Lincoln area is a coarse-grained augen-gneiss, which is developed in the type-section of Kirton Point, near the jetty of Port Lincoln", according to Tilley (1921a) in his description of the gneisses of the Lincoln area (Lincoln Gneisses). These are now included within the Donington Granitoid Suite.

More recently Flook (1975) has mapped the coastal platform adjacent to the Kirton Point Caravan Park. The area also forms part of G.E. Mortimer's Ph.D. thesis area.

A megacrystic granite gneiss crops out at the boat ramp. It consists of large microcline and microcline-perthite megacrysts, on average 2-3 cm in diameter, set in a coarse-grained matrix of quartz, plagioclase, orthoclase and biotite. Hornblende may also be present. Typical accessory minerals include apatite, sphene, zircon, magnetite and ilmenite. The rock has a well developed foliation and appears to have been subjected to at least D_2, D_3 and D_4 of the Kimban Orogeny (cf Fig. 5). The degree of deformation is variable and in places this lithology resembles an augen gneiss (cf Tilley, 1921a).

At least three different types of basic rocks can be recognised at Kirton Point: metadolerites, veined and unveined amphibolites. As at Cape Donington, the metadolerites are essentially undeformed and preserve primary igneous ophitic textures. Clearly these are late-stage basic intrusions.

The relative age of the amphibolites, however, may be the topic of some debate. Adjacent to the boat ramp there is evidence to suggest that a suite of amphibolites existed prior to the emplacement of the megacrystic granite phase. Veins extend from the granite gneiss into the amphibolite (Plate 5) and disrupted blocks of amphibolite can be seen within the granite near the amphibolite margins. Notwithstanding the possibility of back veining of the granite contemporaneous with intrusion of a high-temperature basic magma, it is felt that at least some of the amphibolites predate the granitoids (cf Tilley, 1921a; Flook, 1975). Further support for this concept comes from the presence of mafic xenoliths within the granitoids.

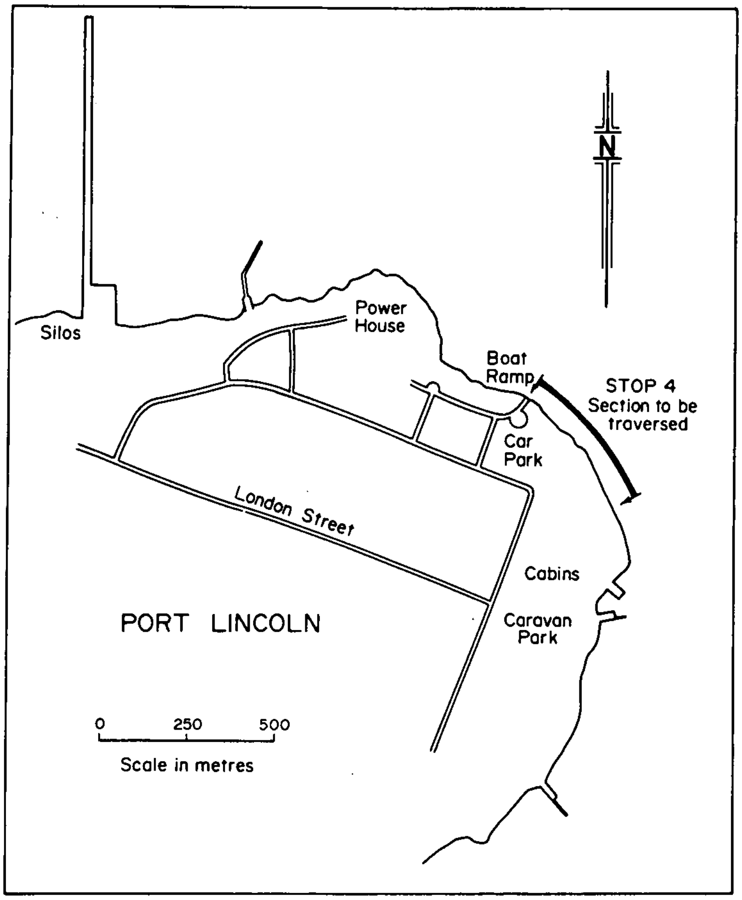


Fig. 10 Locality plan for Kirton Point.

PLATE 5. Megacrystic granite intruding amphibolites at Kirton Point.

Other amphibolites have very sharp contacts with the surrounding granitoids (cf Cape Donington) and the foliation direction in the amphibolites parallels that in the granitoids. It is suggested that these amphibolites are intrusive into the granitoid suite and subsequently metamorphosed and deformed.

Flook (1975) describes two types of amphibolite. One has a hornblende + plagioclase ± quartz ± biotite assemblage and can be seen in outcrop to be intensely veined. The other type is not as intensely veined and contains a hornblende + plagioclase + potash feldspar + quartz ± biotite assemblage. In one instance a boundary between these two amphibolite types is cut by granite gneiss (Flook, 1975). This relationship can be examined.

The other prominent lithology which crops out at Kirton Point is an even-grained granite gneiss. In places this grades into a megacrystic granite gneiss. Mineralogically the even-grained phase consists of microcline, plagioclase (andesine), quartz and biotite, with sphene, magnetite, apatite and zircon present in accessory amounts. Minor green hornblende may also be present. This lithology bears a close resemblance to the granite gneiss seen to intrude the hypersthene granitoids at Cape Donington.

Aplites and pegmatites constitute later intrusive phases, for the most part preceding emplacement of the metadolerites. Pegmatitic veins within the amphibolites outline F_2 folds (see Fig. 4) whereas within the megacrystic granite gneiss some of the pegmatitic segregations seem to have sweated out along the axial planes of F_4 folds.

Metamorphism and Age

For the most part it is envisaged that the primary igneous crystallisation conditions at Kirton Point were essentially similar to those discussed for Cape Donington. The general absence of hypersthene within the gneisses led Mortimer et al. (1979) to suggest that the lower grade amphibolite facies retrogression has had a marked effect at Kirton Point.

Flook (1975) and Cooper et al. (1976) report the results of Rb-Sr measurements on nine samples of the even-grained granite phase. Regression of the nine samples yields a Model 2 isochron age of 1814 ± 22 Ma with an IR of .7045 ± .0017. An MSWD of 7.9 indicates scatter in excess of that attributable to experimental error alone. Most of this scatter appeared to be due to two samples, and regression of the other seven samples results in a perfect fit isochron (MSWD = 0.90). The age calculated from this grouping is 1816 ± 10 Ma with an IR of .7043 ± .0008.

Both of the above age calculations are statistically indistinguishable from those reported by Mortimer et al. (1979) for the hypersthene granitoids at Cape Donington. This supports the interpretation that these ages reflect the time of emplacement and crystallisation of the Donington Granitoid Suite. Note however, there is a slight difference in initial ratios (.7043 ± .0008 compared with .7055 ± .0005).

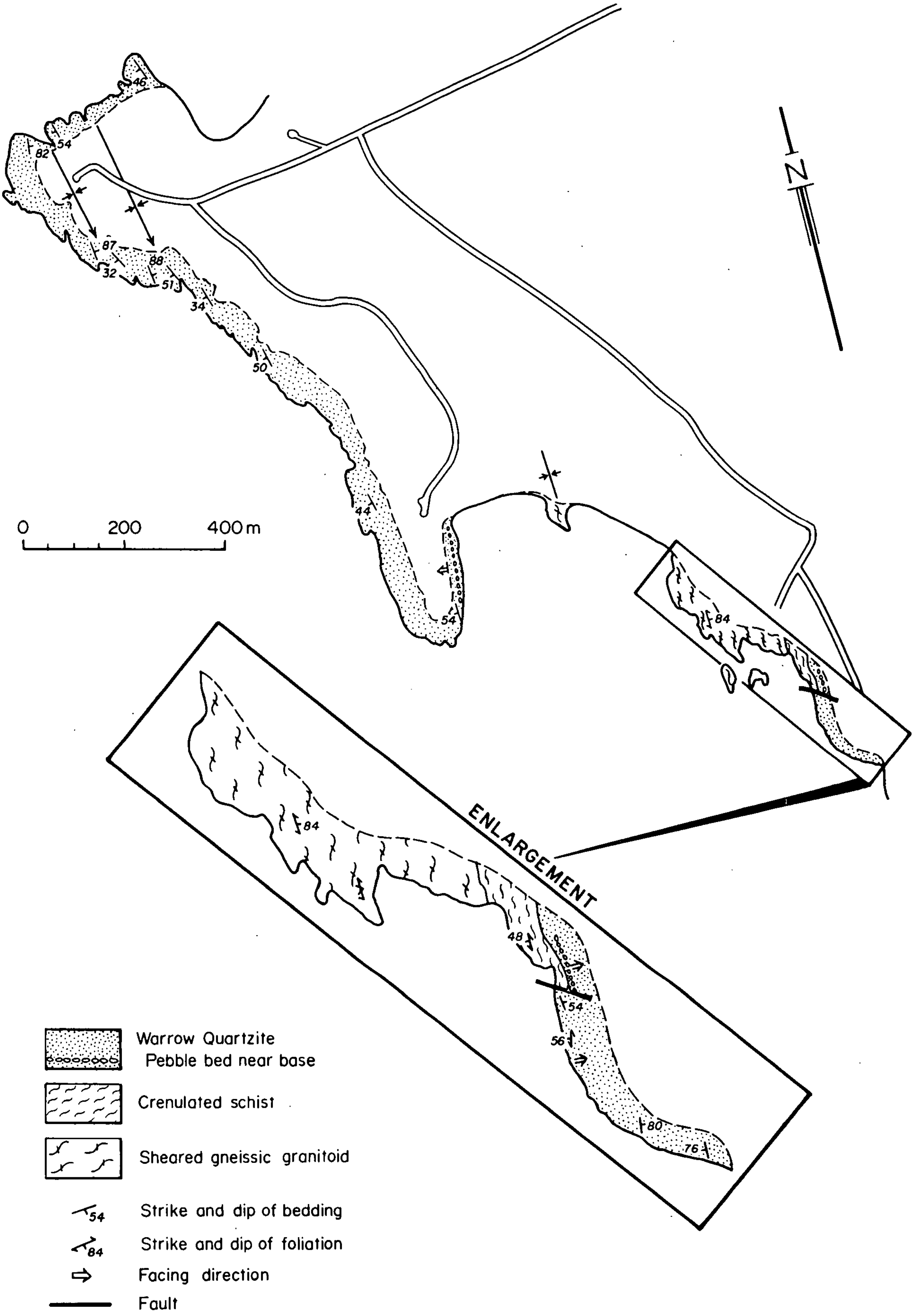


Fig. 11 Geological map of Coles Point.

DAY TWO

Port Lincoln - Coles Point - Drummond Point - Moody Tank - Tumby Bay

Distance (km)

0.0 From Port Lincoln head northwest along the Flinders Highway.

47.6 Mt Dutton on left - composed of Kiana Granite.

50.6 The highway crosses low rises on the southwest edge of Marble Range. These ranges are composed of Kiana Granite overlain but interfolded with Warrow Quartzite.

58.5 Turn left along dirt road.

68.3 Y-junction, turn to left along dirt track through gate. The hill on the right is Mount Greenly which is composed entirely of Warrow Quartzite.

69.2 T-junction, turn left and proceed to southern end of Coles Point (see Fig. 11).

70.6 STOP 5 - COLES POINT (Fig. 11).

A sheared Sleaford Complex granite gneiss crops out along the southern beach at Coles Point (see Fig. 11). The shearing post-dates the aplite veins and some of the pegmatite veins that intrude the granite. These veins can be seen as large augen-like pods within a highly foliated mica-rich quartzofeldspathic gneiss (sheared granite). The foliation strikes approximately north-south and dips fairly steeply to the east. Later pegmatitic veining is discordant to the shearing.

In thin section the granite gneiss consists of recrystallised patches of perthite and microcline-perthite, presumably the relicts of original feldspar phenocrysts as seen at Mt Dutton. Other components of the gneiss are plagioclase, quartz, muscovite and biotite. Tourmaline is also prominent and may be asociated with the ubiquitous veining (cf Mt Dutton). Accessories include apatite, zircon, rutile, and opaques.

East of the sheared granite gneiss, a crenulated andalusite-muscovite-quartz schist crops out and strikes approximately north-south with a steep dip to the west. Attenuated and boudinaged quartzofeldspathic veins occur within the schist. The Warrow Quartzite overlies this schist and the contact can be seen in the cliff exposure. A bed, consisting of quartz pebbles and minor mafic clasts (tourmaline?), occurs at the contact in this outcrop. Note however, at the western end of this beach, the pebble bed occurs as a distinct layer within massive quartzite. In thin section the quartz pebbles are seen to be recrystallised and made up of many strained sub-grains. The interstitial material is also mostly quartz with minor tourmaline, zircon, sphene, and monazite (?). Small muscovite flakes are also present.

Away from the contact the quartzite is massive and the bedding strikes N-S and dips to the east. Cross-bedding indicates that the bedding is the right way up. A prominent cleavage is also observed in the massive quartzite and this has essentially the same orientation as the foliation in the schist. Thus the cleavage in the quartzite and the foliation in the schist were most likely formed during the same deformational event, perhaps D_3 of the Kimban Orogeny (see Fig. 4).

The origin of the schist is likely to cause considerable debate. There are two main hypotheses:

1. It is a lower schist unit within the Warrow Quartzite.
2. It is part of gneissic basement.

In support of the first hypothesis, lithological similarities are observed elsewhere within the Warrow Quartzite. For example, crenulated andalusite muscovite quartz schists occur interbedded with massive quartzite in coastal outcrops south of the Frenchman. These however, are stratigraphically very high or at the top of the Warrow Quartzite.

Crenulated schists do occur below the Warrow Quartzite, in the Marble Ranges within the intervening area between the pebble bed (underlying massive quartzite) and granite gneiss. Here the unconformity appears to lie within a zone of crenulated schistose rock some 10 to 20 metres wide, and it is uncertain where the outcrop ceases to be intensely deformed granite gneiss (evidenced by feldspar content) and passes into mica-rich quartzite. Note here the absence of andalusite porphyroblasts.

Other outcrops of schist are to be seen interleaved with granite gneiss south of the Coles Point beach, immediately west of the Frenchman. These contain biotite, but andalusite is absent. The schists within the basement are most likely a result of local shearing as evidenced by the presence of a crenulated muscovite schist in a shear zone through the tabular feldspar granite cropping out on the coast west of Lake Hamilton. Furthermore, the crenulated schists developed through shearing of the granitoids also commonly contain deformed vein material, a feature which is markedly absent from schists within the Warrow Quartzite in this area.

Therefore the composition of the schists tends to suggest that they form a basal unit to the Warrow Quartzite, whereas the deformation style and the presence of deformed veins tends to suggest that they represent metamorphosed and deformed, weathered granitoid.

82.7 Return to Flinders Highway and proceed north.

118.5 Road junction, turn left toward Drummond Point.

122.6 Road junction, proceed straight ahead.

130.8 STOP 6 - DRUMMOND POINT (Fig. 12).

A 350 m wide shear zone crops out at Drummond Point and separates granodiorite (Coulta Granodiorite) in the east from tabular feldspar granite (Kiana Granite) and even-grained leucogranite in the west.

The tabular feldspar granite is best exposed on the northwest point. Perthite and microcline perthite phenocrysts up to 5 cm in length are a prominent feature of this granite and are typically aligned parallel to the regional foliation. Plagioclase, quartz, muscovite and biotite are also present.

There are several phases of veining; earlier aplite and pegmatite suites which are cut by later pegmatite veins. Both pegmatite suites are characterised by the presence of coarse tourmaline. Some aplites also contain finer-grained tourmaline crystals. The later pegmatite suite contains more coarse muscovite than the earlier pegmatites.

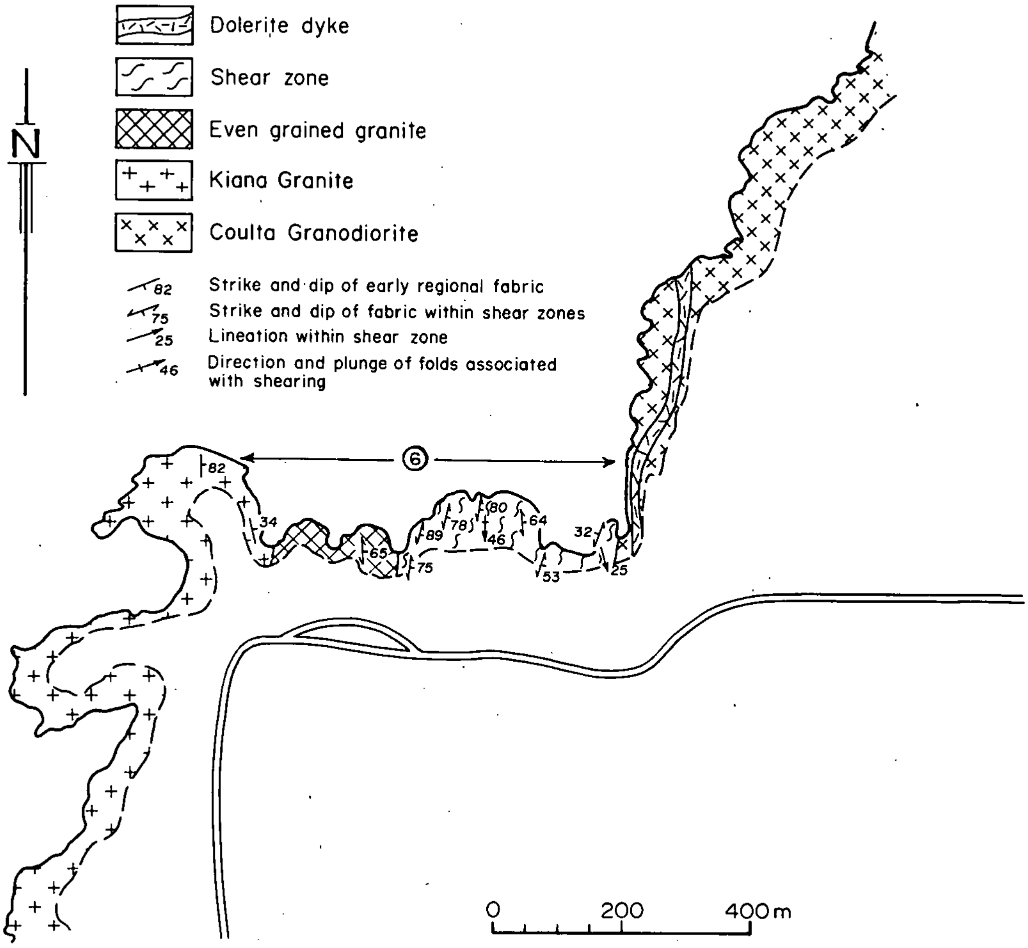


Fig. 12 Geological map of Drummond Point.

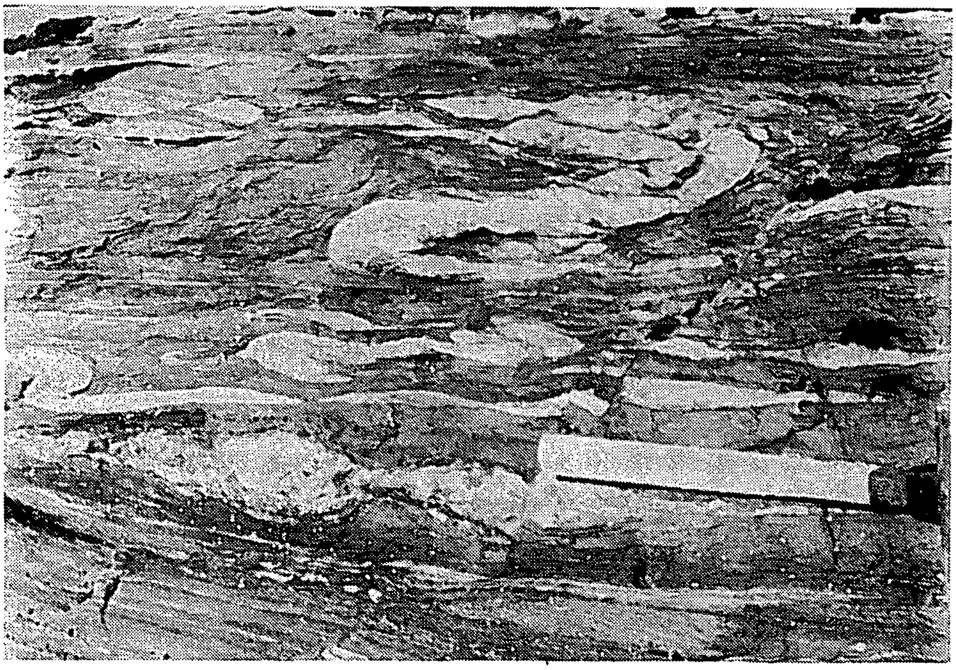


PLATE 6. Isoclinally folded pegmatite veins within micaceous ultramylonite, Drummond Point. Hammer handle is 25 cm long.

The tabular feldspar granite is also intruded by an even-grained leucogranite which in places contains ubiquitous clots of mafic minerals, principally biotite. To the east the even-grained phase dominates and the tabular feldspar granite is preserved as xenolithic rafts or patches. The even-grained leucogranite consists of perthite, microcline perthite, plagioclase, quartz, and muscovite. Biotite is rare except for within the mafic clots.

The north-south striking shear zone is composed for the most part of intensely deformed tabular feldspar granite, even-grained leucoganite and granodiorite (see later). Tight to isoclinally folded pegmatite veins with attenuated fold limbs are a prominent feature (Plate 6), as are a series of conjugate kink folds within the sheared granitoid. Ptygmatically folded veining is also relatively common.

Boudinaged and folded metadolerites can be seen throughout the shear zone. These are relatively late-stage intrusives. In thin section the primary igneous texture appears to be only slightly deformed as shown by the poor parallel alignment of hornblendes. Plagioclase laths have a random orientation.

This shear zone bears a close resemblance to that seen at Coles Point and both of these zones are most likely contemporaneous with the Kalinjala Mylonite Zone in eastern Eyre Peninsula (Parker, 1980b; see STOP 12).

A prominent foliated granodiorite (Coulta Granodiorite) with abundant mafic xenoliths crops out east of the shear zone. Mineralogically it consists of zoned plagioclase phenocrysts, slightly coarser grained than the matrix of quartz, plagioclase, orthoclase and biotite. Apatite, zircon and opaques are the accessories. The xenoliths show variable effects of absorption, but typically consist of hornblende and plagioclase with lesser amounts of quartz and biotite.

An essentially unmetamorphosed dolerite dyke is intrusive into the granodiorite. This dyke is compositionally zoned. Its margins are finer grained and more mafic (hornblende) compared to the core, which is coarser grained and contains greater proportions of plagioclase at the expense of hornblende. Quartz is also more common in the central portion.

138.8 Return to cross roads and turn left.

140.0 Proceed through grid and turn right around southern end of paddock to low rises at southern end of Mt Hope (Good Luck!!).

140.4 STOP 7 - SOUTHERN END OF MT HOPE (Fig. 13).

Mt Hope and Mt Drummond are composed of tabular feldspar granite (Kiana Granite), identical to that seen cropping out on northwestern Drummond Pt. Earlier suites of aplite and pegmatite veins are cut by later pegmatite veins. Mafic xenoliths are also present.

The relationship between the granodiorite and the tabular feldspar granite is obscured at Drummond Pt by the shear zone. However at the southern end of Mt Hope a large xenolith of granodiorite occurs within the tabular feldspar granite (Plate 7). The granodiorite contains mafic xenoliths as at Drummond Pt and has been intruded by an aplite vein prior to incorporation within the granite.

This outcrop provides conclusive evidence that the granodiorite existed prior to the tabular feldspar granite.

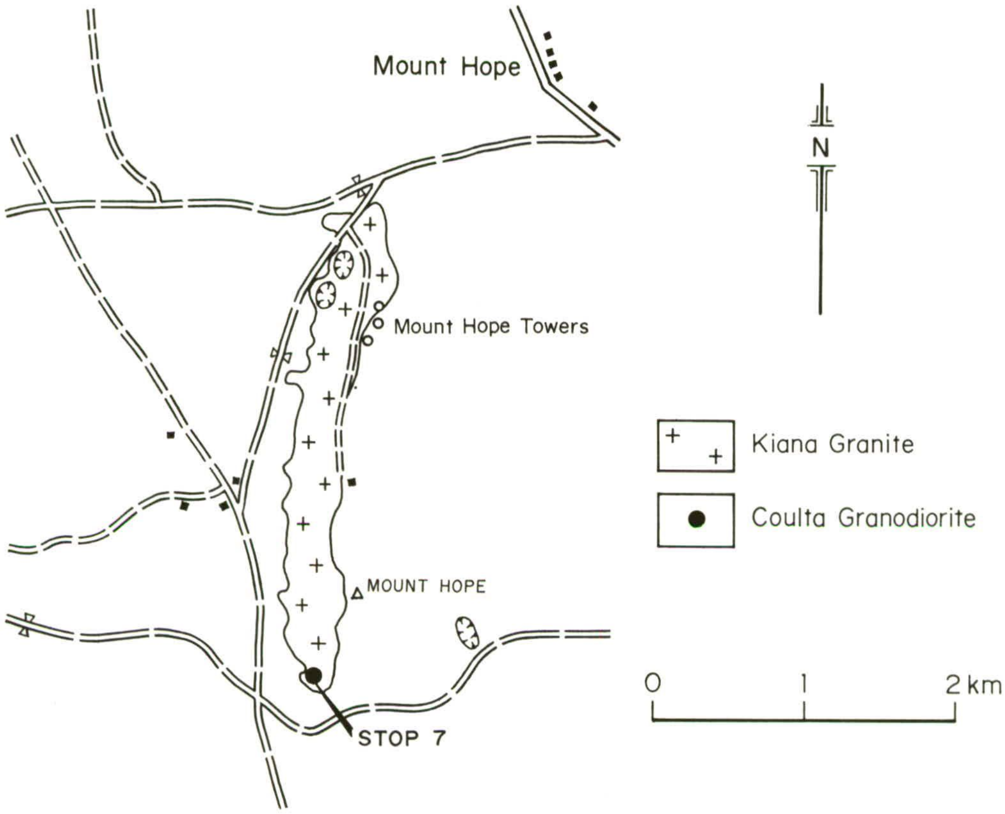


Fig. 13 Locality and geological plan for Mt Hope.

PLATE 7. Contact of Coulta Granodiorite (left) and Kiana Granite (right), Mt Hope. Note mafic xenoliths within Coulta Granodiorite. Sleepy lizard approximately 25 cm long (bite radius 6 cm).

144.9 Return to Flinders Highway, turn right and proceed 100 m then turn left along road to Cummins.

187.3 Township of Cummins, turn left and head north along major road to Yeelana.

200.9 At Yeelana turn right and proceed east along the Ungarra Road.

213.0 Cross road, continue straight ahead.

219.8 Cross road, continue straight ahead.

219.8 Cross road, continue straight ahead.

226.8 T-junction, continue straight ahead.

227.4 Turn left onto dirt track through scrub (50 m) to Moody Tank. The tank was built by the South Australaian Railways c. 1913, utilising the run-off from the granite outcrop.

STOP 8 - MOODY TANK.

Outcropping at Moody Tank is the Moody Tank Adamellite, a light pink to grey, medium-grained, relatively homogeneous adamellite. It is composed of potassium feldspar (40-60%), quartz (25-40%), plagioclase (10-15%), biotite (5-10%) and minor muscovite, with grain size generally less than 5 mm but the feldspars may be locally up to 20 mm resulting in a weakly porphyritic texture. Dispersed garnets are common and some aggregates are up to 10 mm size (Coin, 1976; Flint and Pain, 1981). Within the granite there are xenoliths of grey schistose material which probably represent the Cook Gap Schist. The outcrop is crosscut by a series of randomly oriented diffuse pegmatite veins.

Structural evidence suggests that the granite is a late Kimban Orogeny granite which intruded syn- to post- D_3. The quartz grains show undulose extinction and in some samples the biotite grains are aligned producing a weak foliation which indicates some degree of strain.

The Moody Tank Adamellite is one of several intrusions making up the Moody Suite including several other adamellites, Chinmina Syenite, Uranno Microgranite and the Yunta Well Leucogranite. Mortimer (1984) considered all except the Yunta Well Leucogranite to be I-type granitoids. He believes the Yunta Well Leucogranite was formed by insitu melting of Hutchison Group metasediments just prior to the intrusion of the I-type granitoids.

Rb-Sr analyses of 7 samples from the Moody Tank region do not define a unique isochron, however assuming an initial $^{87}Sr/^{86}Sr$ ratio of 0.703 the probable age is 1600-1700 Ma. K-Ar analysis of biotite from one specimen gives an age of 1562 Ma (Webb _et al._, 1986). Rb-Sr analyses carried out by Mortimer (1986) on all the I-type granitoids of 1709 ± 14 Ma with an initial ratio of 0.7070. This age is a minimum age for D_3 which better fits estimates of the timing of D_3 from U-Pb zircon dating of between 1755 and 1710 Ma.

The granite at Moody Tank is very similar in mineralogy and age to the Carpa Granite west of Cowell (Parker _et al._, 1981).

Continue east along Yeelana - Ungarra road to Ungarra.

233.9 Ungarra township; head south along road to Tumby Bay.

235.0 Y-junction, veer left.

237.4 Y-junction, continue ahead southeasterly.

241.4 Numerous road cuttings of Yunta Well Leucogranite and Cook Gap Schist. The Yunta Well Leucogranite is a light-coloured weakly-moderately foliated quartz, feldspar, muscovite leucogranite.

251.4 Township of Lipson; continue ahead.

252.3 T-junction, turn right and head south along Lincoln Hwy to Tumby Bay.

262.7 Tumby Bay township.

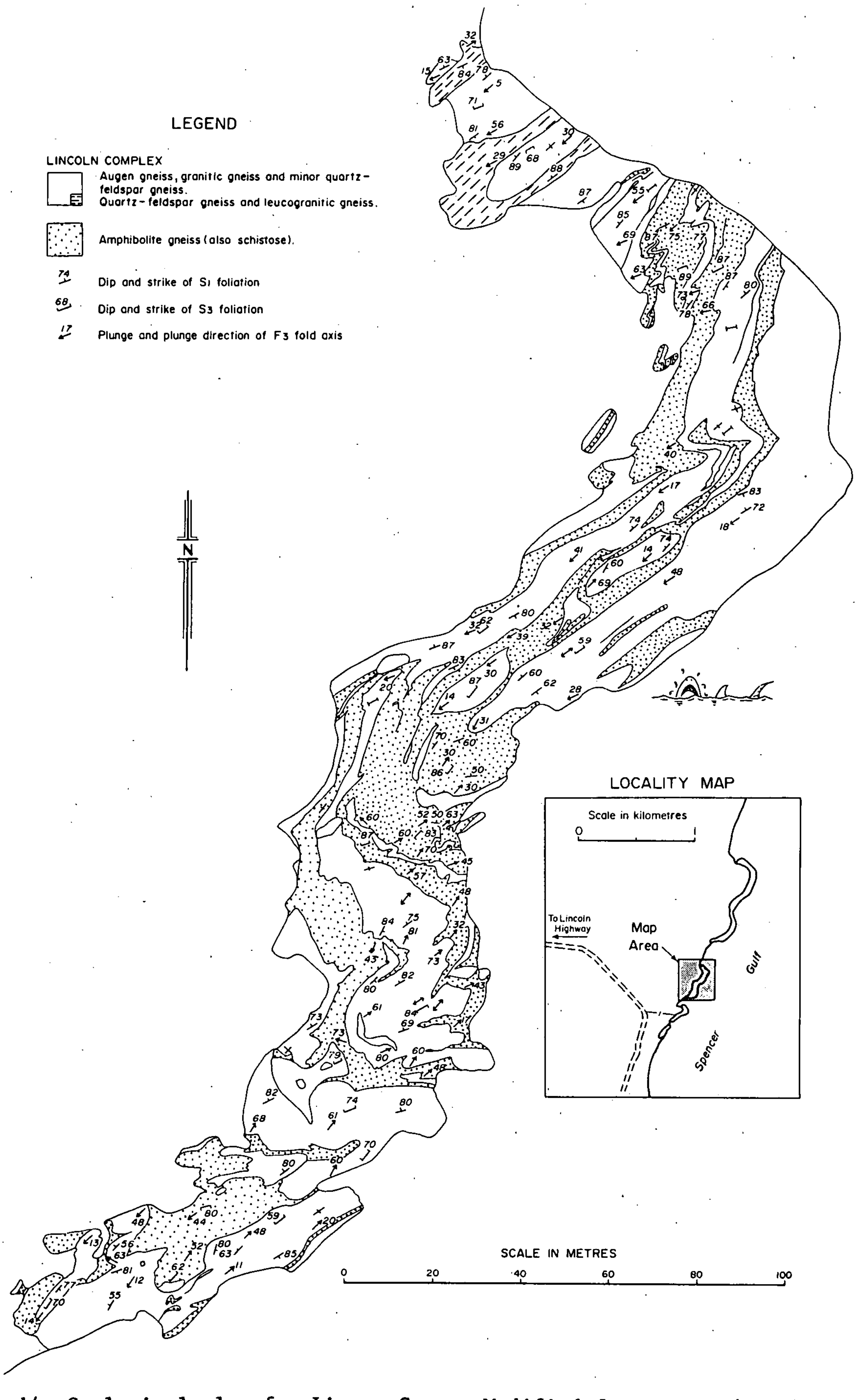


Fig. 14 Geological plan for Lipson Cove. Modified from Cohen (1983).

DAY THREE

Tumby Bay - Lipson Cove - Port Neill - Cleve

Distance (km)

0.0 Leave Tumby Bay and head north along the Lincoln Highway.

16.9 Road junction, turn right along dirt road to Lipson Cove.

23.6 Stop at north end of Lipson Cove near grid and walk down to coastal platform. NB. This coastal traverse should be done at low tide as high tide covers much of the outcrop.

STOP 9 - LIPSON COVE (Fig. 14).

The Lincoln Complex at Lipson Cove consists of intimately interlayered felsic and mafic gneisses exposed continuously for 1.5 km along a narrow (approx. 30 m wide) shore platform. The coastline is at a low angle to the regional structural trend, exposing the geology roughly along strike. This locality is part of P.H. Cohen's Ph.D. thesis area (Cohen, 1983), and his assistance in the preparation of the locality description is gratefully acknowledged.

The felsic gneisses at Lipson Cove can be subdivided into two principal units. The first, and most common, is a coarse-grained quartz + feldspar + biotite (± hornbende) augen gneiss which contains more than 5% mafic minerals. The augen are composed of potassium feldspar and perthite and range from prolate to extreme oblate forms. The second is a medium to coarse-grained leucocratic gneiss with <5% mafic minerals.

Three principal mafic lithologies that have been recognised. The most common a foliated hornblende + feldspar + biotite (±epidote-chlorite) amphibolite which contains large amounts of quartz and migmatite veining. The second is leuco-amphibolite gneiss which has a high feldspar/hornblende ratio, no quartz, and accessory sillimanite. Thirdly, a biotite-rich gneiss which occurs as pods within the amphibolite and as thin (<5 cm) layers within the acid gneisses; this unit contains abundant quartz ribbons.

The high-grade LS tectonic fabric at this locality represents a transition between the low-grade fabric within the augen gneisses at Cape Hardy to the north and the high-grade LS mylonitic fabric developed at Tumby Bay to the south. The four major deformational episodes of the Kimban Orogeny have been recognised in this zone.

D_1 has produced a strong LS fabric parallel to the S_0 gneiss/amphibolite layering. S_1 is the pervasive fabric at this locality. L_1 is a non-penetrative mineral elongation. No F_1 folds have been recognised in this area.

D_2 produced mesoscopic tight to isoclinal folds, in places folding L_1. The plunge of the F_2 fold axes varies considerably due to refolding by F_3 folds. A weak, non-penetrative schistosity (S_2) was produced in some of the fine-grained mafic units.

The third deformation produced folds of variable morphology and intensity, ranging from microscopic crenulations to macroscopic open, asymmetric folds, commonly with doubly-plunging hinges. The F_3 folding is responsible for the regional structural trend at this locality. A weak LS fabric was produced during D_3, with the S_3 schistosity being locally visible

PLATE 8. A tight to isoclinally folded (F_3) amphibolite layer within the megacrystic gneiss at Lipson Cove. The fold occurs within a localised zone of high strain. A protomylonitic foliation has developed axial planar to the fold.

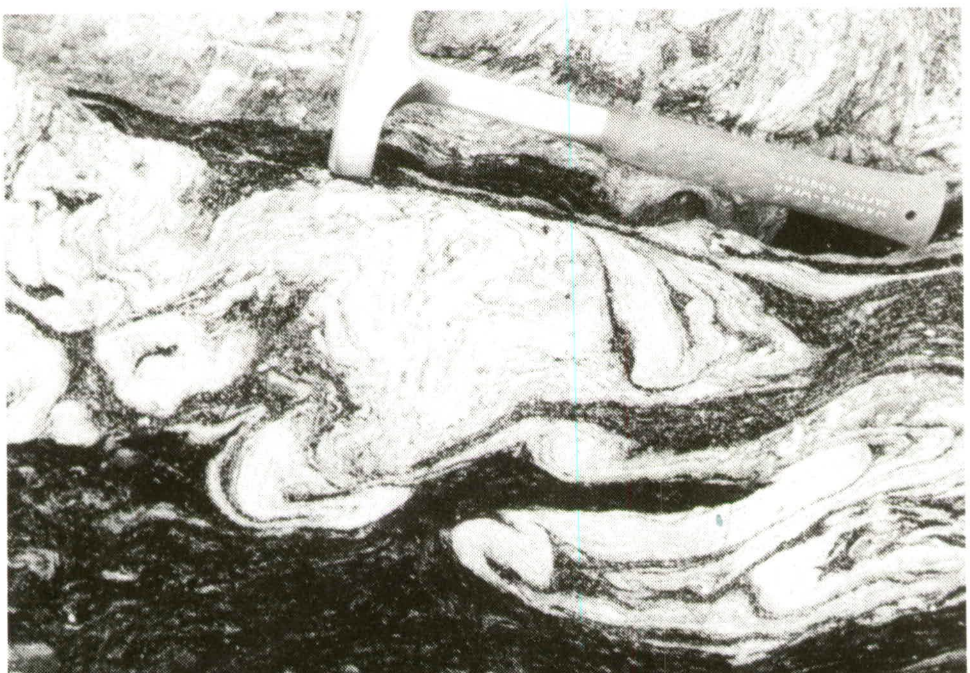


PLATE 9. Intense folding, including possible small-scale sheath folds within highly-stained interlayered augen gneiss and amphibolite at Lipson Cove.

in the hinges of F_3 folds. Several syn-D_3 pegmatites exhibit this schistosity. The L_3 lineation is a non-penetrative elongation lineation, similar to L_1, defined by the elongation of K-feldspar augen within the augen gneiss. Cohen (1983) found no evidence of L_1 being folded about F_3 folds, suggesting that L_1 may have been rotated into parallelism with the F_3 fold axes. Localised crenulation of S_1 by D_3 has also produced a weak crenulation lineation L_3^1. The intense mylonitic fabric produced by D_3 in higher strain areas at Tumby Bay and Port Neill, has not been formed at this locality, although localised zones of intense strain have developed a protomylonitic foliation axial-planar to F_3 folds (Plate 8). Small-scale sheath folds also occur within zones of intense strain (Plate 9).

The final deformational event D_4 produced minor F_4 kink folds and crenulations.

30.3 Return to the Lincoln highway and continue north to Port Neill.

53.0 Pass through Port Neill to southern end of Back Beach (see Fig. 15).

53.8 STOP 10 - CAPE BURR (Fig. 15).

Two coastal traverses at this locality examine deformed Lincoln Complex granitoids and intrusive mafic dykes. The granitoids are dominated by a coarse-grained, grey quartz + feldspar + biotite (± hornblende) megacrystic gneiss, the same unit seen at Lipson Cove, and interlayered even-grained mafic and felsic gneisses, commonly intruded by amphibolite dykes (1-10 m wide).

The granitoids exhibit a steeply dipping, moderately strong LS fabric which results from the modification of an early (?D_1) fabric, by $D_{m/3}$. Similar granites to the south, at Cape Hardy, exhibit a much less intense fabric (S>L), indicating that with increasing distance from the mylonite zone there is less modification of the early fabric by $D_{m/3}$. The amphibolite dykes also exhibit a strong LS fabric but this is probably due entirely to $D_{m/3}$; i.e. the amphibolite probably intruded after D_2. The competence contrast between the amphibolites and granitoids has resulted in heterogeneous concentration of high strain within the amphibolites and the immediately adjacent gneisses.

The orientation of the fabric is roughly parallel to the mylonite zone, which is located 3 km, across strike, to the northwest. The mean strike trend is 030° and mean dip 85°W. The lineation plunges between 0° and 40° towards 210° and parallels both the axes of $F_{m/3}$ folds and the lineation in the mylonite zone.

The $F_{m/3}$ folds are commonly seen as a series of small-scale tight to isoclinal folds, folding and hence modifying an earlier fabric in the granitoids, but do not appear to fold any early fabric within the amphibolites. The fabric in the amphibolites is probably an axial planar fabric resulting from $D_{m/3}$ folding. The vergence of the $D_{m/3}$ folds varies randomly from S, M to Z with no particular vergence dominating.

A later folding event, probably also associated with $D_{m/3}$, folds the early $D_{m/3}$ fabric and lineation, in the amphibolites and granitoids, about steeply plunging (85°-90°) fold axes.

Return to Port Neill and proceed to small headland northeast of the town (see Fig. 15).

54.9 STOP 11 - PORT NEILL (Fig. 15).

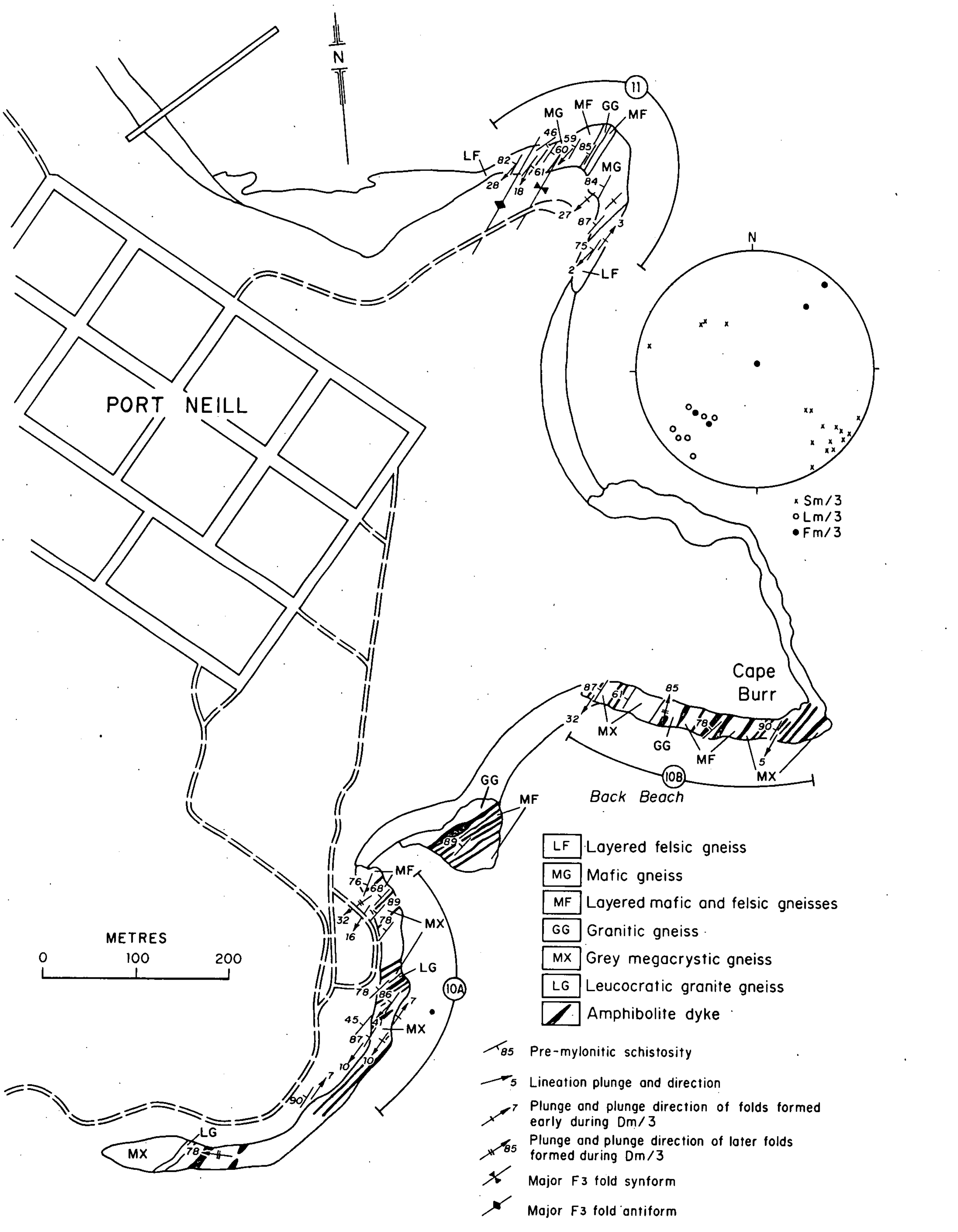


Fig. 15 Structural and lithological plan for the Pt Neill area, STOPS 10 and 11.

The small headland is structurally closer to the mylonite zone than STOP 10. It is composed of a different sequence of Lincoln Complex granitoids with the two dominant lithologies being a medium-grained pink, quartz + feldspar + biotite granite gneiss and a medium-grained mafic, quartz + feldspar + hornblende (± clinopyroxene + garnet) gneiss.

The orientation of the fabric is the same as that seen at STOP 10, but there is a much greater variation in fabric intensities ranging from S>L fabrics to intense L fabrics. Although there is no continuous variation in fabric from SE to NW, the intense L fabrics are seen on the northwest side, i.e. nearer the mylonite zone. The fabric is again a modification of the pre-mylonitic fabric by $D_{m/3}$, with $D_{m/3}$ becoming more intense to the northwest. The lineation on the northwest side of the outcrop is a mineral elongation lineation where the feldspar has preferentially formed along the intersection lineation $L^{I}_{3/m}$. This is most evident in $F_{3/m}$ fold hinges.

The $F_{3/m}$ folds plunge shallowly toward 210° and are evident as small-scale tight folds with varying vergences. A large-scale (50 m) synform-antiform $F_{3/m}$ fold set which plunges 20° toward 210° can be seen on this traverse, the fold has a dextral vergence (viewed from north).

From northern corner of Port Neill proceed along dirt road around NW side of clay pan (see Fig. 15).

56.2 Turn right through fence/grid and proceed around western and northern side of paddock (see Fig. 16).

57.8 STOP 12 - PORT NEILL MYLONITE ZONE (Fig. 16).

The coastal platform to be examined forms the southern portion of the type locality for the Kalinjala Mylonite Zone (Parker, 1980b). The zone separates the upper amphibolite metamorphic facies Lincoln Complex gneisses, seen at STOPS 10 and 11, from a series of felsic and mafic granulites, to the north in the Bratten Cairn region, of as yet undetermined origin.

The Kalinjala Mylonite Zone is a major linear zone of intense ductile deformation extending the length of the Eyre Peninsula, from Sleaford Bay to the western flanks of the Middleback Ranges. At Port Neill the zone is roughly 3 km wide and displays a variety of textures ranging from protomylonite to slaty ultramylonite.

The main lithology present at this locality is quartz+feldspar mylonite, but it contains numerous scattered bands of ultramylonite and also remnants of protomylonite. There is no obvious pattern to the distribution of each of these lithologies and boundaries/textural changes are gradational. Typical mylonite consists of scattered augen-shaped feldspar porphyroblasts (2-50 mm across) with trails of recrystallised feldspar merging into a fine to very fine-grained, strongly foliated matrix with ribbons of finely recrystallised mica, quartz and feldspar.

The orientation of the mylonitic schistosity, S_m, is parallel to the overall orientation of the zone and strikes approximately 030° with mean dip 80-85° west. However this varies considerably due to mesoscopic and macroscopic swirling around large amphibolite pods (or boudins) and widespread pinch and swell structures (Plate 10). The lineation (L_m) developed within the mylonitic schistosity is defined by elongate rods of recrystallised feldspar, quartz and mica, whilst individual feldspar augen are generally equidimensional within the plane of S_m. The orientation of the lineation (10° toward 030° to 40° towards 210°) parallels the axes of F_m folds which are common throughout the mylonite. They fold the pre-S_m gneissosity, with S_m

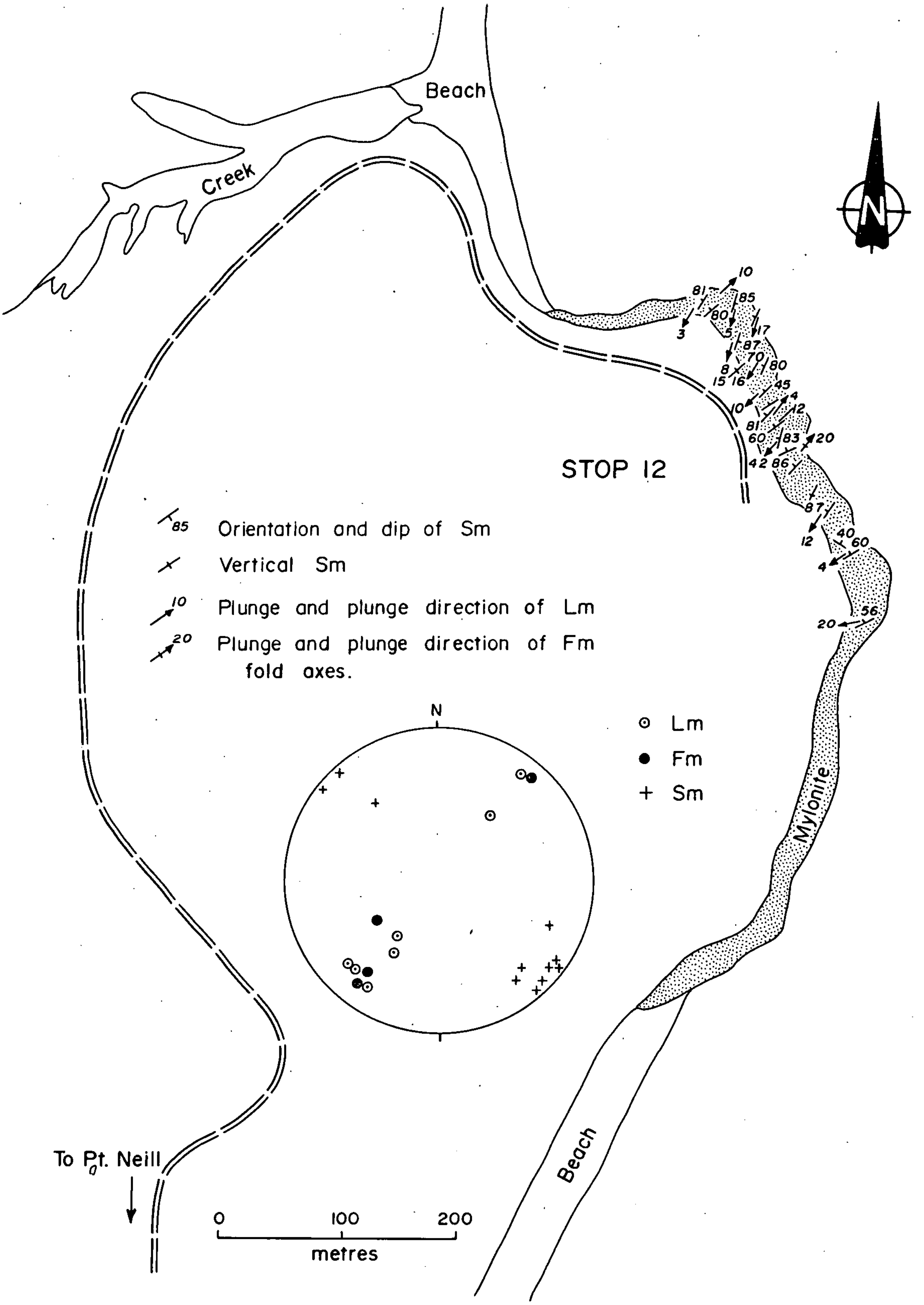


Fig. 16 Structural and locality plan for the northern portion of the Pt Neill mylonite zone.

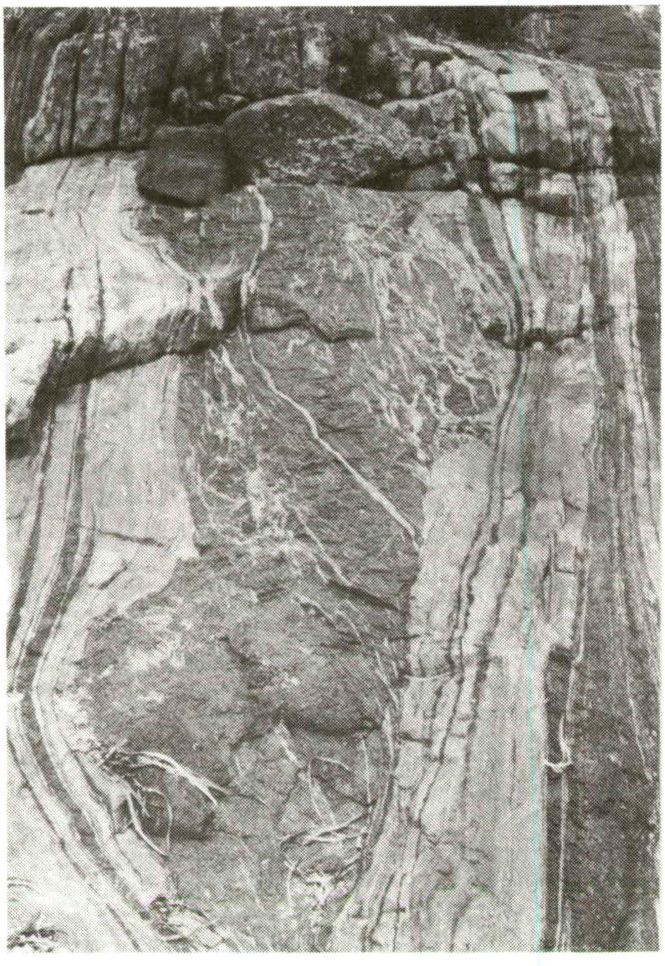


PLATE 10. Mafic and felsic mylonite surrounding amphibolite boudins. Note S-plane foliation within boudin, oblique to the surrounding Sm foliation. Boudin is approximately 0.5 m wide. Port Neill mylonite zone.

PLATE 11. Small-scale folds within late-stage dextral shear zones cross-cutting the principal mylonite zone strike. Port Neill mylonite zone.

defining an axial planar fabric. The axial planar S_m is manifest as spaced cleavage zones in protomylonite and as pervasive mylonitic schistosity in mylonite/ultramylonite. The vergence of the folds varies randomly with no one vergence dominating.

The main mylonitic fabric is crosscut by a series of later thin mylonitic shear zones, which are oriented approximately 065° dipping 85°NW. S_m and L_m are rotated into the shear zones indicating a dextral sense of shear (Plate 11). Although post-dating the main mylonite forming event these shear zones were probably still developed during the $D_{3/m}$ deformation event.

The Kalinjala Mylonite Zone is the most extensive of several parallel mylonite zones along eastern Eyre Peninsula. These have formed in response to intense ductile deformation related to D_3 folding in intervening zones. On all scales from microscopic to macroscopic, homogeneous flattening is at least, if not more, as important as, simple shear. Clark (1976) argues that the apparent rotation of blocks or boudins of amphibolite etc. indicate a major component of simple shear and, in particular, transcurrent movement. However, this evidence is overpowered by the presence of intense folding about shallowly plunging axes in adjacent rocks. The L_m lineation, which has also been interpreted as evidence for transcurrent movement (Clark, 1976), is very closely related to F_m folds, and is at least in part an intersection lineation.

Return to Lincoln Highway and proceed north.

74.5 Turn right at second bend after Bratten Cairn and proceed along dirt track on southern side of fence (see Fig. 17).

76.6 STOP 13 - BRATTEN CAIRN (Fig. 17).

The section to be examined starts at the southern end of the low scrubland. This coastal exposure is probably located 1-2 km, across strike, west of the mylonite zone and consists of a sequence of complexly folded and sheared mafic granulites and felsic gneisses.

The mafic granulites appear as large pods, up to 200 m in length, within the felsic gneisses. A pervasive foliation was developed in the mafic granulites prior to $D_{m/3}$ during a granulite metamorphism. The S_1 fabric remains somewhat constant, striking 120°-130° and steeply dipping, indicating the pods have not undergone significant rotation. The $D_{m/3}$ event produced a pervasive LS tectonite fabric, locally grading to mylonite in the felsic gneisses. $S_{m/3}$, which anastomoses around the mafic granulite pods, completely overprints the early granulite fabric, except in small pockets surrounding these pods, in the felsic gneisses. The $D_{m/3}$ foliation strikes roughly 050°-060° and is steeply dipping. $S_{m/3}$ is commonly folded by small intrafolial folds which change from open to tight as the axes become progressively rotated into parallelism with the $S_{m/3}$ fabric (Clark, 1976).

A conjugate set of narrow shear zones, oriented at 045° and 175°, locally crosscut the dominant $S_{m/3}$ foliation. These late-stage shears probably represent semi-brittle fracture during the waning stages of $D_{m/3}$.

Complex fold patterns and varying orientations of the fabrics seen in the northern part of this traverse probably result from interference patterns produced by the combination of all these deformations.

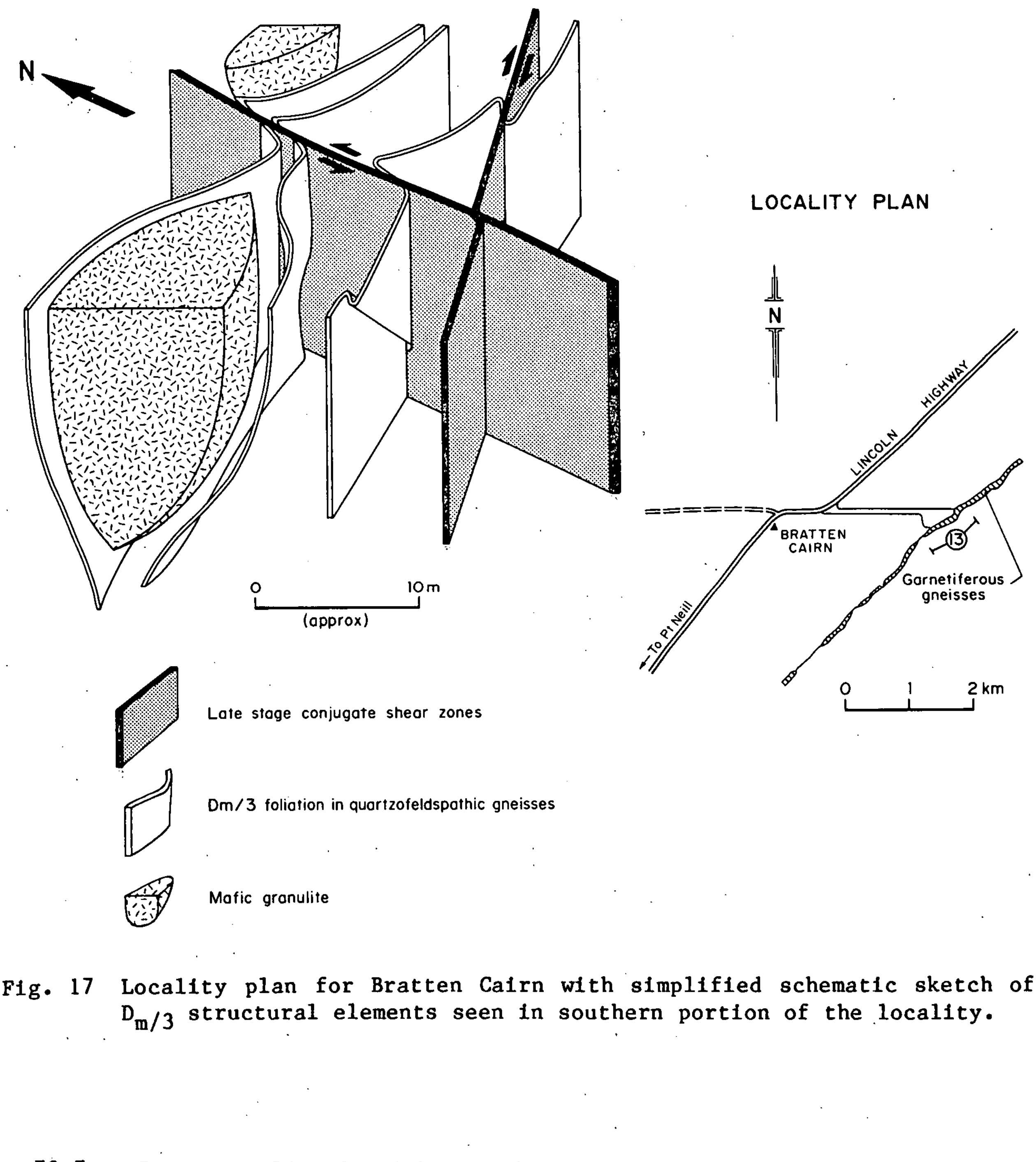


Fig. 17 Locality plan for Bratten Cairn with simplified schematic sketch of $D_{m/3}$ structural elements seen in southern portion of the locality.

78.7 Return to Lincoln Highway and continue north to Arno Bay.

100.2 At Arno Bay turn left along sealed road to Cleve. Although there is minor Hutchison Group outcrop along the road close to Cleve, much of this area was originally covered by Middle Proterozoic post-tectonic sediments of the Blue Range beds (Corunna Conglomerate equivalent). These crop out by the Cleve aerodrome, near Ben Boy, at Point Gibbon, at Port Hughes on Yorke Peninsula, and around the margin of the Polda Basin, suggesting that that basin may have been initiated in the Precambrian.

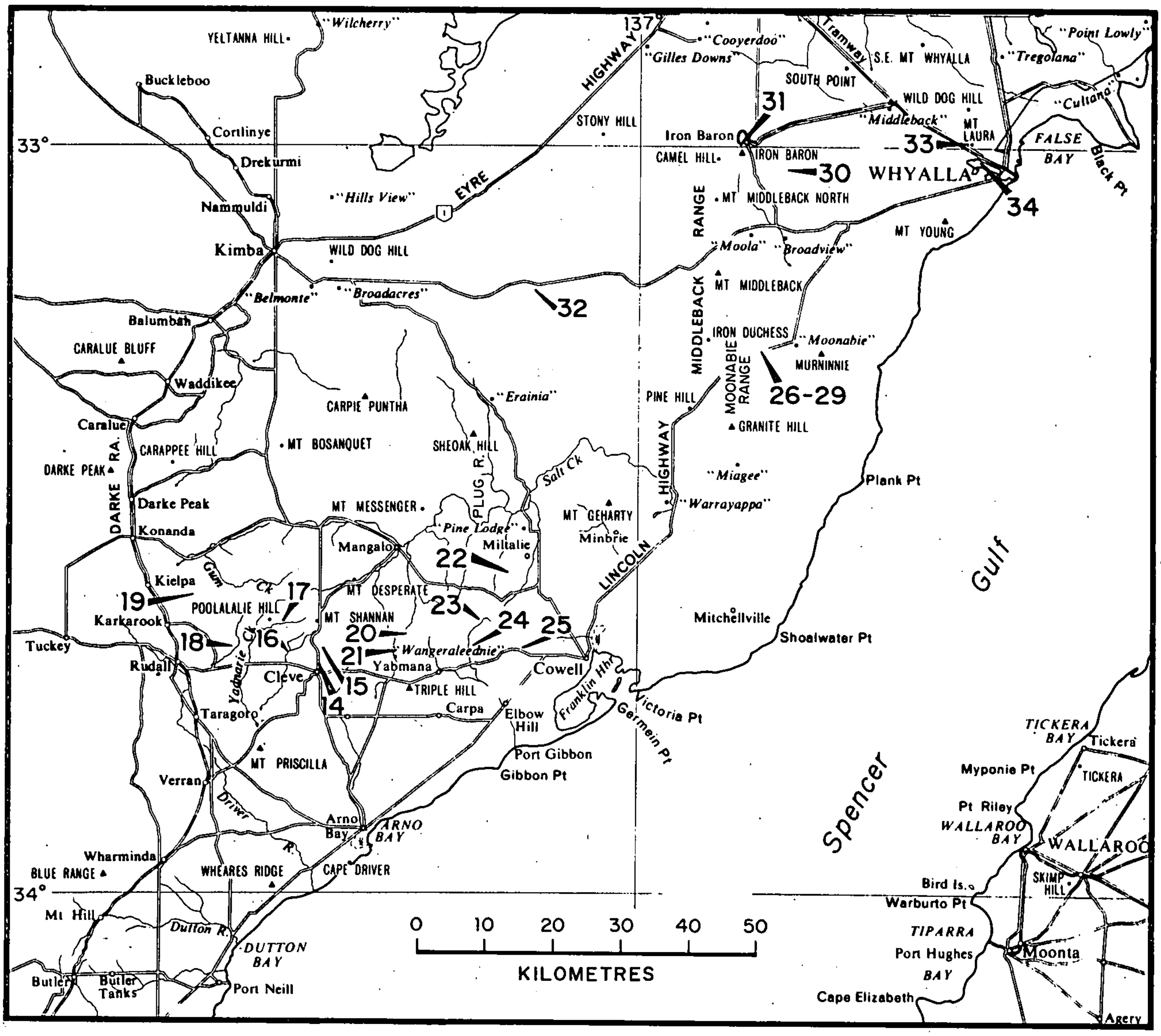


Fig. 18 Locality plan for eastern Eyre Peninsula showing excursion stops for days 4 and 5.

DAY FOUR

Cleve - Mangalo - Miltalie - Cowell

Distance (kms) - Fig. 18.

0.0 Proceed north from Cleve along Cleve-Kimba road parallel to Poornamookinie Creek.

2.4 STOP 14 - POORNAMOOKINIE CREEK (Fig. 19).

On the left hand side of the road the Warrow Quartzite consists of feldspathic and micaceous quartzite with narrow pelitic schist interbands, plus numerous granitic sills. The quartzite-schist interbanding represents bedding (So), with a layer-parallel schistosity in the pelitic layers interpreted as a combined S_{1-2} fabric. This schistosity is axial planar to minor tight F_2 folds within the pelitic layers. A weak L_2 mineral lineation is defined by the alignment of micas within both the pelitic and quartzite layers.

The dominant deformation in this locality is D_3 with the bedding dipping shallowly to the southwest in the core a major D_3 antiform (the "Cleve Antiform"). Toward the northwest end of the outcrop, quartzite bands are locally boudinaged, indicating tensile stress in the antiform core. Partial melting commonly occurred in the pinch zone of the boudins, which are mostly offset in a dextral sense. Local dextral-vergence folding in less competent bands is associated with the boudinaging. Minor mesoscopic D_3 open warps refold the L_2 mineral lineation. A weak crenulation lineation (L_3) has been developed in the more micaceous layers.

Granitic sills (and dykes) of massive, leucocratic, off-white to pink, quartz + feldspar + muscovite + garnet granite equivalent to the Carpa Granite are prolific in this area (Plate 12). The conformable nature of the sills with the S_2 (folded by F_3) layering in the quartzite suggests that the granite was intruded pre-D_3. It is, however, believed that there was a close spatio-temporal relationship between the two. Pegmatite veins are diffuse and often occur at the sill quartzite boundaries. Two generations of pegmatite are recognised; a) a garnetiferous, syn-granite intrusion pegmatite, and b) a later coarse-grained tourmaline-bearing (± andalusite/sillimanite) pegmatite with sharp intrusive contacts.

Continue north along the Cleve-Kimba road.

4.0 T-junction, continue along main road to the northeast, toward Mangalo.

4.4 STOP 15 - CLEVE COUNCIL PIT (Fig. 20).

Overlying the Warrow Quartzite is the Katunga Dolomite, the Lower Middleback Jaspilite and the Cook Gap Schist. Generally in the Cleve area, the Katunga Dolomite is a massive to poorly bedded, off-white dolomitic marble, while the Lower Middleback Jaspilite is represented by silicate-facies "iron formation" and recrystallised graphitic chert both with a low iron oxide content (Parker and Lemon, 1982).

In the Cleve council pit the sequence above the Warrow Quartzite is dolomite and opalite, graphitic and tremolitic quartzite, and semipelitic schist. The semipelitic schist (A) consists of quartz, biotite, muscovite, feldspar and garnet, with scattered psammitic bands and is quartz-veined. The graphitic and tremolitic quartzite unit (B) consists of grey, laminated, very

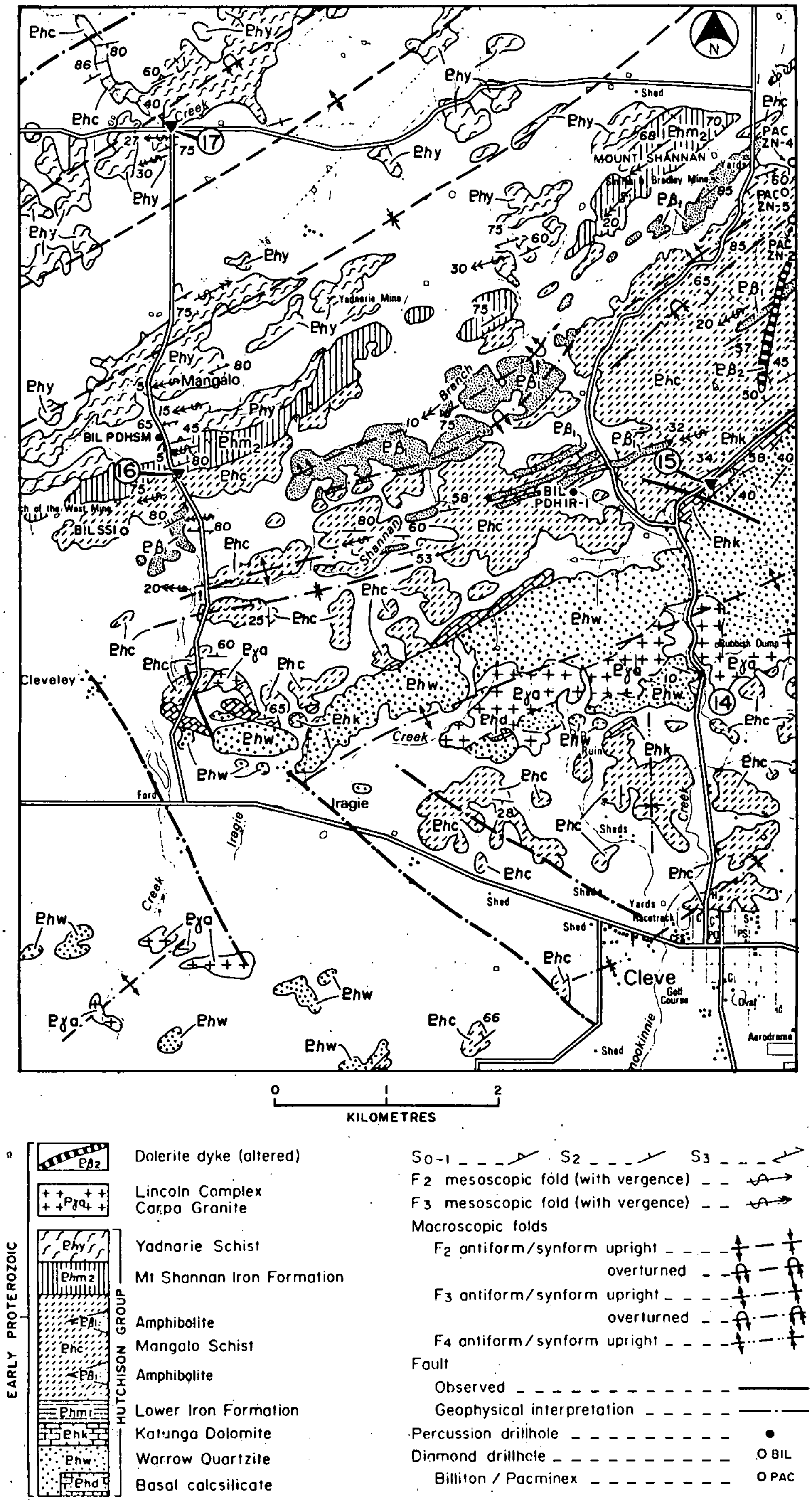


Fig. 19 Geological map of the Cleve-Mangalo Creek area. Modified from Rankin (1987).

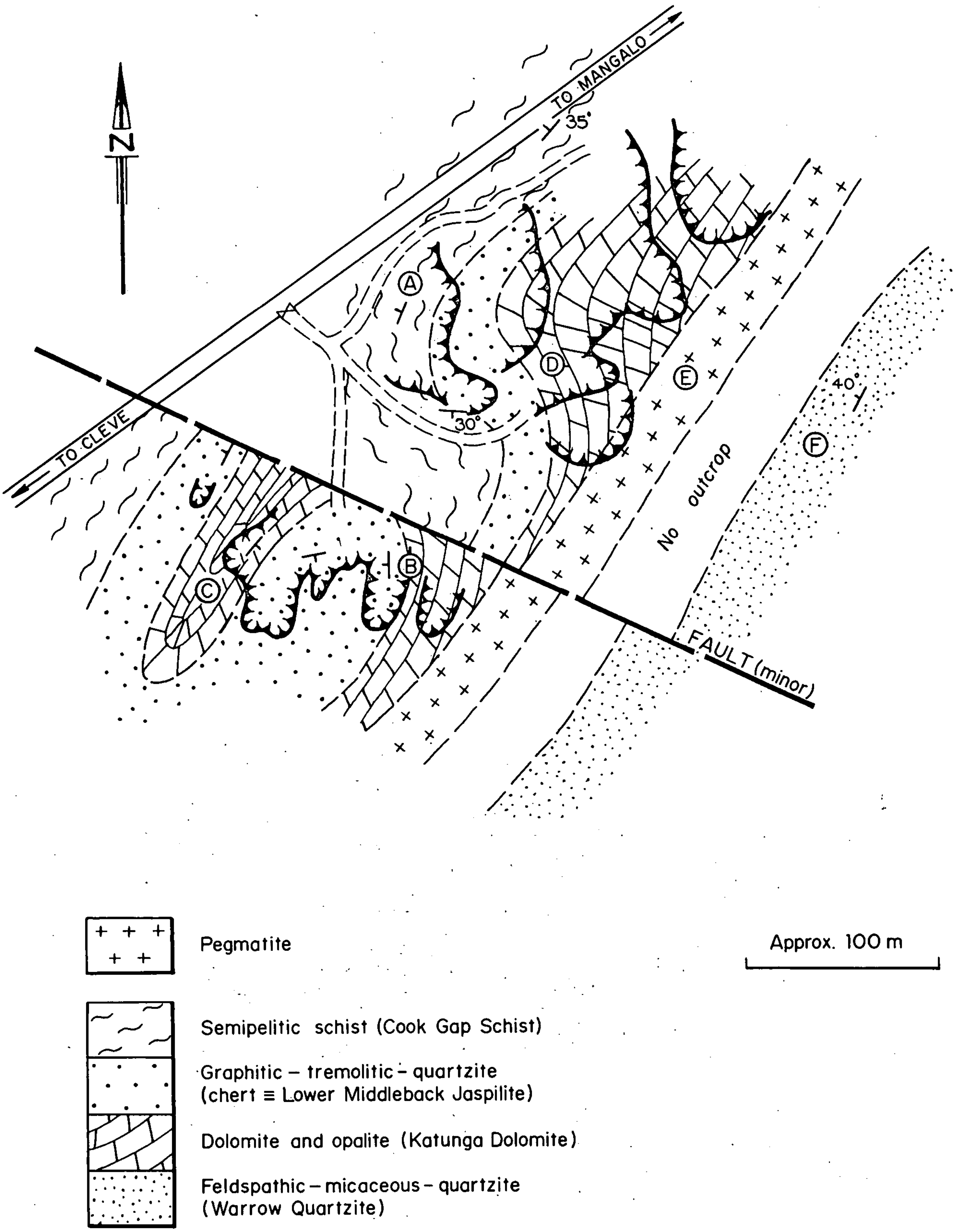


Fig. 20 Geological plan for the Cleve council pit.

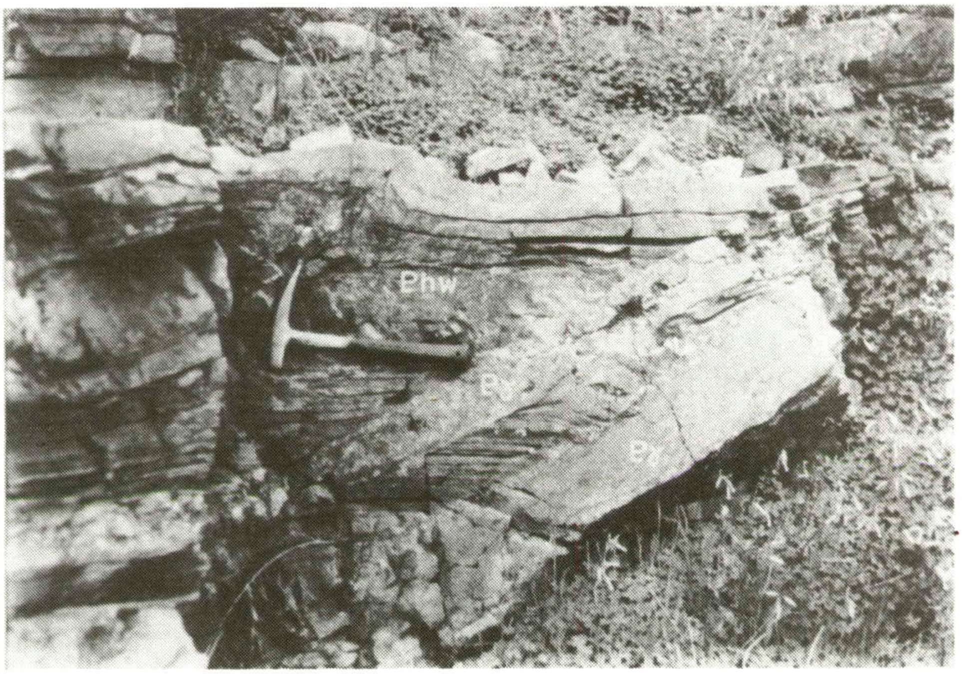


PLATE 12. Gently-folded Warrow Quartzite cross-cut by Carpa Granite-equivalent pegmatite. Poornamookinie Creek.

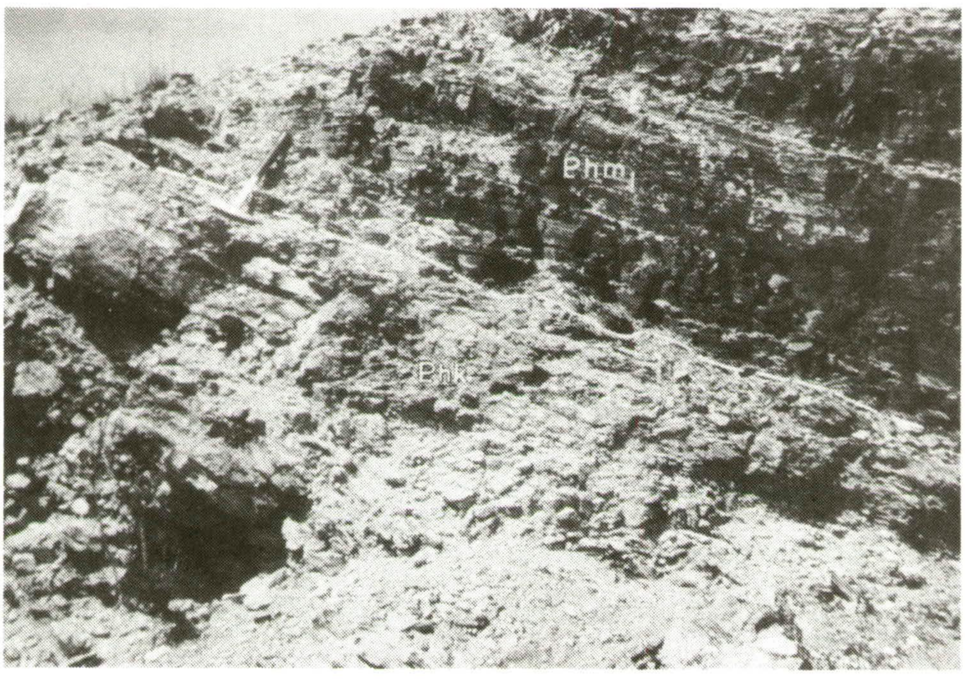


PLATE 13. Contact of Katunga dolomite (Ᵽhk) and Lower Middleback Jaspilite ($Ᵽhm_1$), Cleve Council Pits.

fine-grained quartzite with thin, either graphitic or grunerite-rich bands, and broader interbands (several cm wide) of massive tremolite. The underlying dolomitic marble is best seen at (C) (Plate 13) but elsewhere has been partially replaced by opalite, particularly along bedding planes and joints (D). The boundary of the dolomite and graphitic chert is gradational with interbanding of grey, fine-grained quartzites, tremolite, dolomite and opalite, and occasional pegmatite veins. Broad, open, D_3 buckling of the dolomite and graphite units is well exemplified at (B). East of the pits is a coarse-grained pegmatite (E) separating the dolomite unit from the massive to flaggy outcrops of feldspathic quartzite (Warrow Quartzite) at (F).

Return to Cleve.

8.8 At Cleve Post Office turn right (west) along the main road.

13.7 Turn right at the crossroads and proceed north, parallel to Mangalo Creek. The outcrops immediately to the east in the fields consist of muscovite-feldspar rich quartzites of the Warrow Quartzite. A near complete stratigraphic section of Hutchison Group metasediments is traversed northwards along the road.

14.4 Overlying the Warrow Quartzite, white to pink dolomitic marbles of the Katunga Dolomite are locally exposed in a borrow pit on the eastern side of the road. At this location the unit consists of pink dolomitic marble, with minor fine-grained quartzite bands, capped by Tertiary calcrete. The Lower Middleback Jaspilite, which locally overlies the Katunga Dolomite elsewhere in the Cleve Uplands, does not appear at this locality.

15.1 The Mangalo Schist with F_2 and F_3 folds crops out in road cuttings on the eastern side of the road. Continue to the north.

16.8 STOP 16 - MANGALO CREEK (Fig. 19).

Overlying the Katunga Dolomite, to the north, is the Cook Gap Schist, known locally in the Cleve area as the Mangalo Schist. In Mangalo Creek, it is a sequence of quartz-veined biotite and muscovite schists and psammites consisting of 20-50% quartz, 20-50% plagioclase, 10-45% potassium feldspar, 5-20% biotite, 5-35% muscovite plus minor garnet, tourmaline and opaques. The schist is folded by mesoscopic tight to isoclinal F_2 folds with a strong axial planar S_2 schistosity at a high angle to $S_{o/1}$ in the hinges. S_2 is parallel to and indistinguishable from the $S_{o/1}$ compositional layering in the attenuated limbs of both mesoscopic and macroscopic F_2 folds, the resultant foliation being labelled $S_{1/2}$. The F_2 folds and their accompanying S_2 axial planar fabrics have been refolded by F_3 buckle folds and crenulations (Plate 14). The axial planes for F_2 and F_3 are close to coaxial in strike, though there is a slight angular difference in dip and plunge. Clear overprinting criteria are present.

The Mangalo Schist commonly contains concordant amphibolite bodies (of uncertain origin) which are foliated and multiply deformed. They vary from fine-grained, strongly schistose amphibolites to more massive, medium to coarse-grained varieties. The coarser hornblende crystals are somewhat randomly oriented, however finer hornblende needles locally form a mineral lineation. Amphibolites containing F_3 folds outcrop in Mangalo Creek towards the top of the Mangalo Schist.

PLATE 14. Typical biotite schist with quartz veins of the Cook Gap Schist, Mangalo Creek. Isoclinal F_2 folds defined by the quartz veins are refolded by open to tight F_3 folds. An S_3 crenulation cleavage is visible.

PLATE 15. $S_{0/1}$ compositional layering within interlayered metapelites and metapsammites of the Yadnarie Schist folded by open to tight F_2 folds at Poolalalie Creek. An axial-planar S_2 crenulation cleavage is visible.

To the north of the amphibolite, 8 samples of Mangalo Schist were analysed, with regression of the Rb-Sr analyses producing a Model 2 isochron with an age of 1686±76 Ma, initial $^{87}Sr/^{86}Sr$ ratio of 0.7061±0.0042 and MSWD of 23.1 (Webb et al., 1986). K-Ar dates on 2 samples gave muscovite ages of 1629 and 1600 Ma and biotite ages of 1588 and 1537 Ma (Webb et al., 1986).

The Upper Middleback Jaspilite is known locally in the Cleve area as the Mount Shannan Iron Formation. It includes brecciated cherty and ferruginous quartzites, dolomites with jaspilitic chert bands and graphitic schists. By Mangalo Creek, in a small borrow pit on the east side of the road (Fig. 19), Mangalo Schist at the southern end is overlain to the north by massive to poorly banded, reddish pink to grey dolomite with occasional tremolite-rich layers. The reddish colour is probably due to a moderate Fe content. Further north along the pit, there are highly micaceous crenulated schists, while north of these, adjacent to the creek, are some superb outcrops of interbanded medium to fine-grained dolomites and cherty quartzites. The dolomites are pinkish-red to brown (sideritic), but occasionally grade to white. The cherty quartzites are very fine-grained, sometimes laminated, and range from pinkish-grey to red jaspilitic bands 5-30 mm thick.

The Mount Shannan Iron Formation is overlain to the north by the Yadnarie Schist. It has been recognised only in this area northwest of Cleve, where the unit has a minimum thickness of 1 000 m (the top is not known). The Yadnarie Schist consists of fine-grained muscovite + biotite + quartz + feldspar schists with interlayered psammitic bands up to several centimetres thick. This unit is nearly identical to the Mangalo Schist in petrography and outcrop appearance. The two units are distinguished on tectonostratigraphic criteria.

Continue to the north.

19.8 Cross roads, proceed down to creek.

STOP 17 - POOLALALIE CREEK (Fig. 19).

In Poolalalie Creek, the Yadnarie Schist consists of fine-grained muscovite + biotite + quartz + feldspar schists with interlayered psammitic bands up to several centimetres thick. This locality is in the hinge of a major D_2 syncline and the psammitic bands have been tightly folded with the development, particularly within schist, of a strong axial planar schistosity, S_2 (Plate 15). S_2 varies from a local crenulation cleavage (crenulating S_1) to a strongly developed segregation schistosity. Developed on S_2 are minor fine crenulations (F_{3-4}).

Proceed west.

23.9 T-junction, continue to the west.

25.0 Y-junction, turn right and head northwest.

26.8 Stop at road cutting just south of Sheoak Creek.

STOP 18 - SHEOAK CREEK (Fig. 21).

In the roadcutting 500 m south of Sheoak Creek, well-banded calcsilicates and dolomitic marbles are exposed. This unit locally defines the base of the Warrow Quartzite in the Cleve, Calcookara and Plug Range areas (Rudall and Mangalo 1:50 000 Geological Maps).

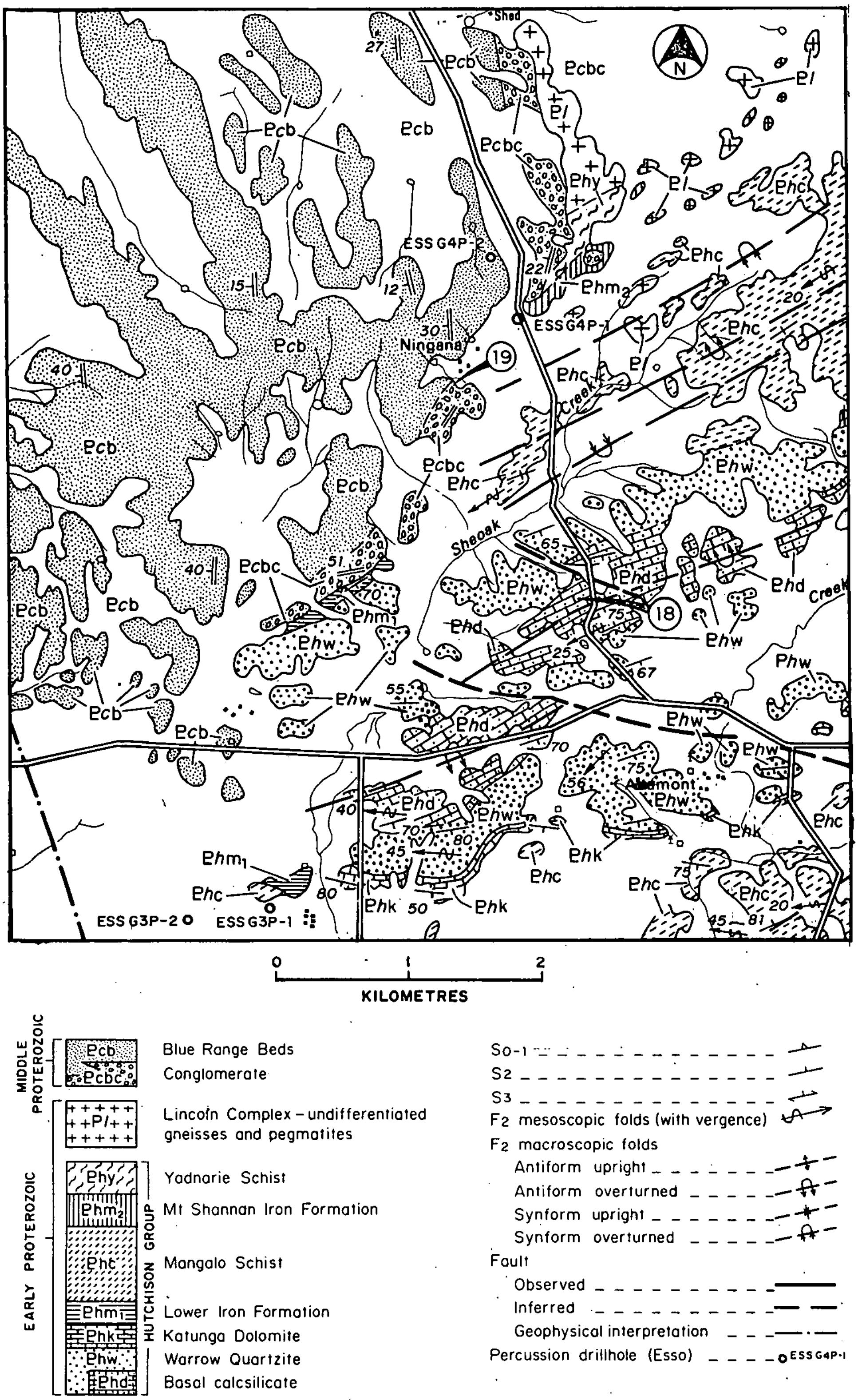


Fig. 21 Geological map of the Ningana-Sheoak Creek area. Modified from Rankin (1987).

At this locality, the calcsilicate unit consists of alternating bands (5-25 cm thick) of dolomite (80-90%) + microcline (10-15%) + sericite/muscovite (5-10%) and dolomite (30-40%) + diopside (20-30%) + microcline (25-30%) + quartz (5-10%) + sericite (2-5%). In thin section, the calcsilicates are medium-grained with granoblastic to granoblastic elongate textures. Diopside is typically intergrown with quartz + dolomite, with optical continuity between grains suggesting that the diopsides were coarse (10-15mm) grained porphyroblasts.

On the northeast side of the road is a spectacular outcrop of intensely folded calcsilicate, with complex tight F_2 folds plunging 30° towards 235° (Plate 16). A weak axial planar fabric is visible in the hinges of the F_2 folds, defined by the alignment of quartz-diopside aggregates and micas. The folds show dominantly Z-vergence, but with some M-vergence folds evident. In the paddock to the west of the road cutting rare S-vergence F_2 folds have been found. The calcsilicate is therefore interpreted to occupy the core of a macroscopic F_2 overturned antiform plunging at this locality to the southwest. Isolated outcrops of this unit can be traced for approximately 10 km to the northeast, where the fold then plunges to the northeast.

A weak F_3 crenulation within mica-rich layers is the only evidence of D_3 in this locality.

Continue north along road.

27.7 Turn left to "Ningana" farm-house. NB Permission must be obtained from the owner before proceeding onto the property.

28.3 From farmhouse walk south across paddock to outcrop on hill.

STOP 19 - "NINGANA" (Fig. 21).

In the Cleve area, unmetamorphosed arenites unconformably overlie Hutchison Group metamorphics. The nomenclature and suggested ages have varied considerably (see section Middle Proterozoic Stratigraphy) but the sediments are known as Blue Range beds and are interpreted as equivalent to Corunna Conglomerate in Moonabie Range.

Near "Ningana" the basal cobbly conglomerate is not exposed though its presence is indicated by well-rounded quartz cobbles up to 20 cm in diameter. The lowermost unit exposed is a pebbly conglomerate with subrounded to rounded quartz and quartzite clasts (Plate 17). This is in turn overlain by interbedded granule conglomerates and sandstones. These conglomerates are poorly sorted with abundant granules in a quartz, feldspar (< 5 mm) and mica matrix, while the sandstones are medium to very coarse-grained with a subangular to subrounded arkosic matrix. The sequence of apparently fluvial sediments is in excess of 2.5 km thick. Foresets indicate a southeasterly source.

A pervasive mauve and off-white colour mottling occurs either patchily or streaked along bedding planes.

The Blue Range beds have been locally folded with dips up to 70°, and intruded by botryoidal hematite veins, and quartz veins (minor). These probably occurred during the Wartakan Event ca 1500-1450 Ma.

Return to Cleve.

39.0 Proceed straight on through Cleve past Post Office towards Cowell (east).

PLATE 16. Tight F_2 folds within the basal calcsilicate member of the Warrow Quartzite at Sheoak Creek.

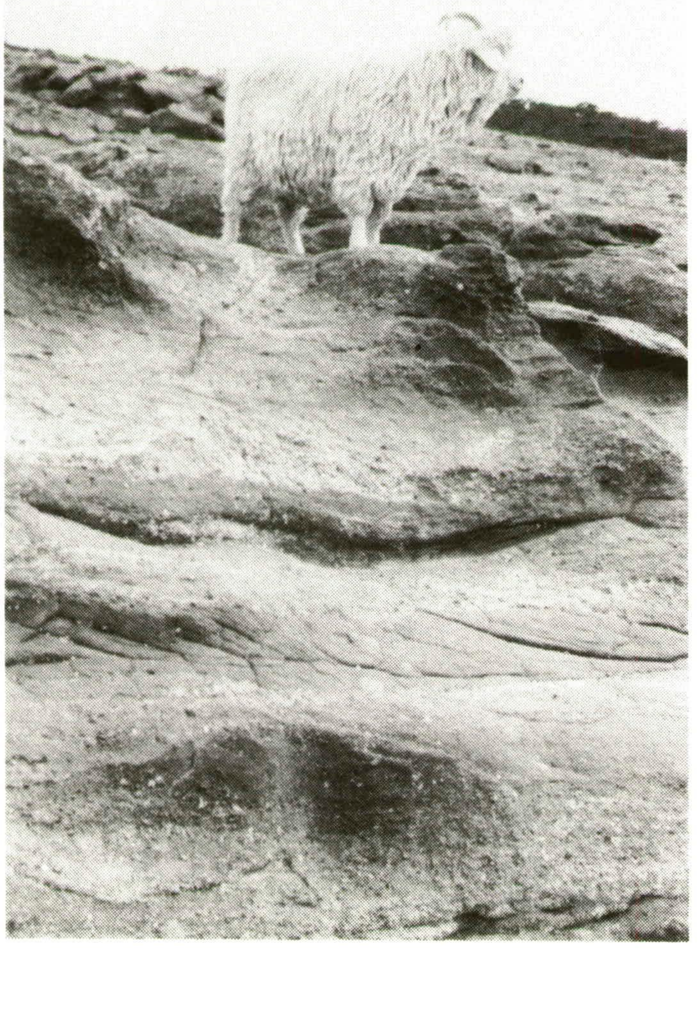


PLATE 17. Quartz pebble beds interbedded with medium-grained, cross-bedded sands within the Blue Range Beds at "Ningana". Scale of goat unknown.

47.2 Y-junction, turn left onto dirt road. The hills to the left comprise Warrow Quartzite forming the core of a major D_3 anticline.

50.3 Straight on.

52.2 Wangaraleednie Homestead, on the left, is on the site of the first white settlement in the region. Dr James McKechnie and his brothers Donald and Peter occupied this site in 1853 and subsequently took up a lease of 43 square miles at a rental of 10 shillings per square mile. Wangaraleednie is believed to be aboriginal for "Hill of the West Wind".

54.0 Turn left onto Range Road. The road crosses a tight D_2 syncline in Cook Gap Schist (Fig. 22).

55.4 Amphibolite, at base of Cook Gap Schist, crops out on left.

55.5 Grey tremolitic marble of the Katunga Dolomite is exposed in road cut.

56.3 STOP 20 - RANGE ROAD LOOKOUT (Fig. 22).

The lookout at the top of Range Road in the Wangaraleednie area is located very nearly on the hinge of a major D_3 antiform known as the Sheoak Hill Antiform. The upper part of the Warrow Quartzite crops out at the lookout and down the road in a number of borrow pits where flaggy to massive, feldspathic quartzite is interbanded with pelitic schist. The quartzite is medium-grained, foliated, strongly rodded, contains scattered coarse-grained quartz lenticles, and consists of quartz (ca 75%), feldspar (ca 15-20%), muscovite, minor biotite and accessories. The rodding is a D_2 feature because it is defined by highly elongate aggregates of quartz or feldspar flattened in S_{1-2} and folded by F_3 (Plate 18). These quartz/feldspar aggregates and lenticules are believed to be of tectonic origin rather than deformed sedimentary clasts as has been suggested.

Silvery, pelitic schist interbeds characterise the upper part of the Warrow Quartzite. They are medium-grained, almost always strongly crenulated and consist of muscovite (60-70%), biotite (15-20%), quartz (10-15%) and minor sillimanite, feldspar, tourmaline, garnet and opaque minerals. Two crenulations are evident - a pervasive, SW-trending D_3 crenulation overprinted by a weaker, more locally developed E-W-trending D_4 crenulation. These deform both the S_{1-2} schistosity and rare, isoclinal, intrafolial F_2 defined by narrow S_1 quartz veins (and also, in thin section, by sillimanite aggregates).

56.5 Turn around at lookout and return down Range Road.

59.0 STOP 21 - WANGARALEEDNIE AREA (Fig. 22).

The creek section represented in Fig. 22 displays an excellent section from the Cook Gap Schist down sequence through amphibolite, the Lower Middleback Jaspilite equivalent, Katunga Dolomite, pelitic schist to typical Warrow Quartzite. It is complicated slightly by a broad, open, shallow southwest plunging F_3 fold, parasitic to the Sheoak Hill Antiform. To the east of the section a major D_2 syncline is defined by a change in F_2 fold vergence from "Z" or dextral vergence at "A" (plunging southwest) through "M" vergence, to "S" or sinistral vergence (plunging southwest) some 600-700 m downstream. F_3 vary in plunge from shallow northeast to shallow southwest.

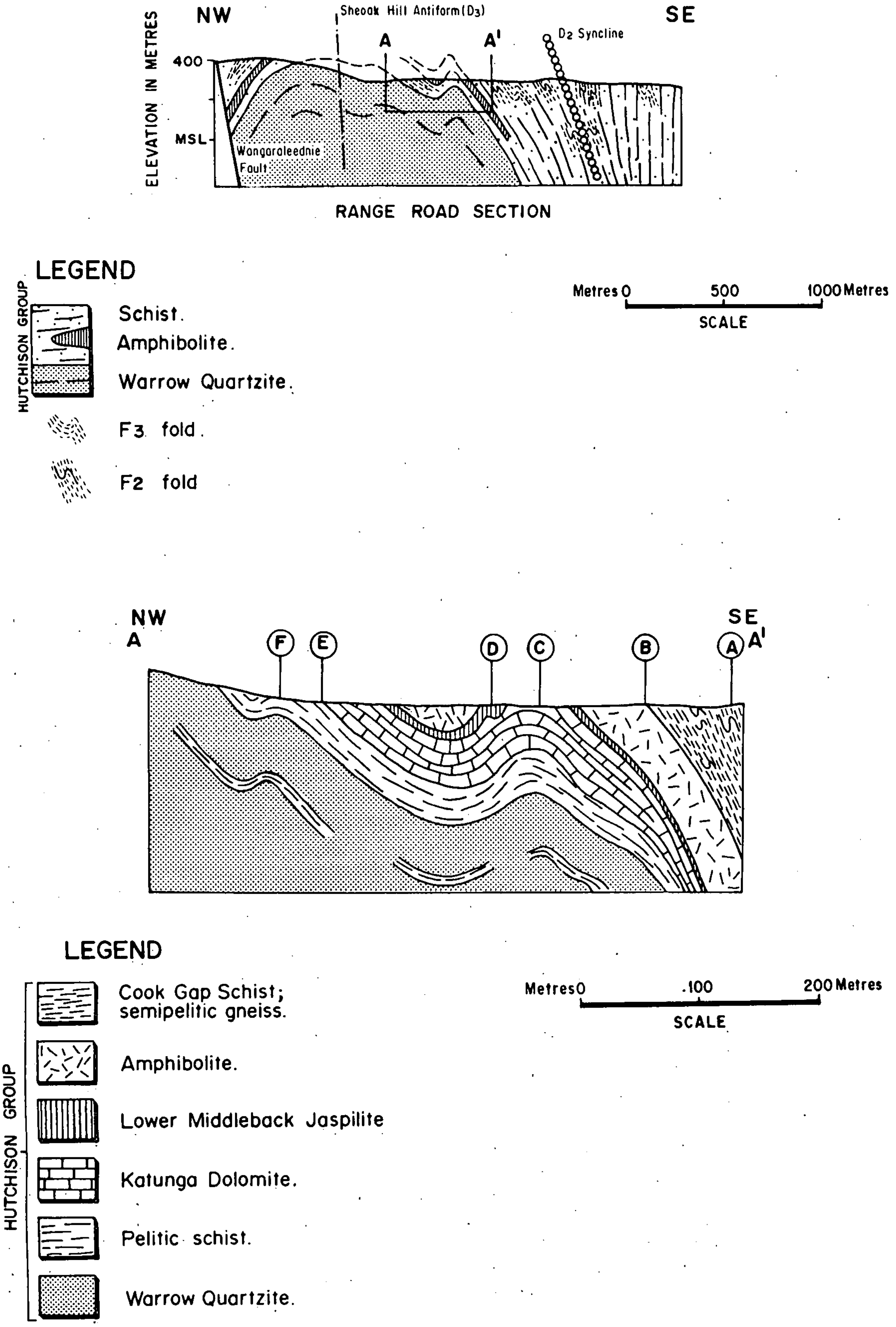


Fig. 22 Geological cross sections for the Wangaraleednie area, STOPS 20 and 21.

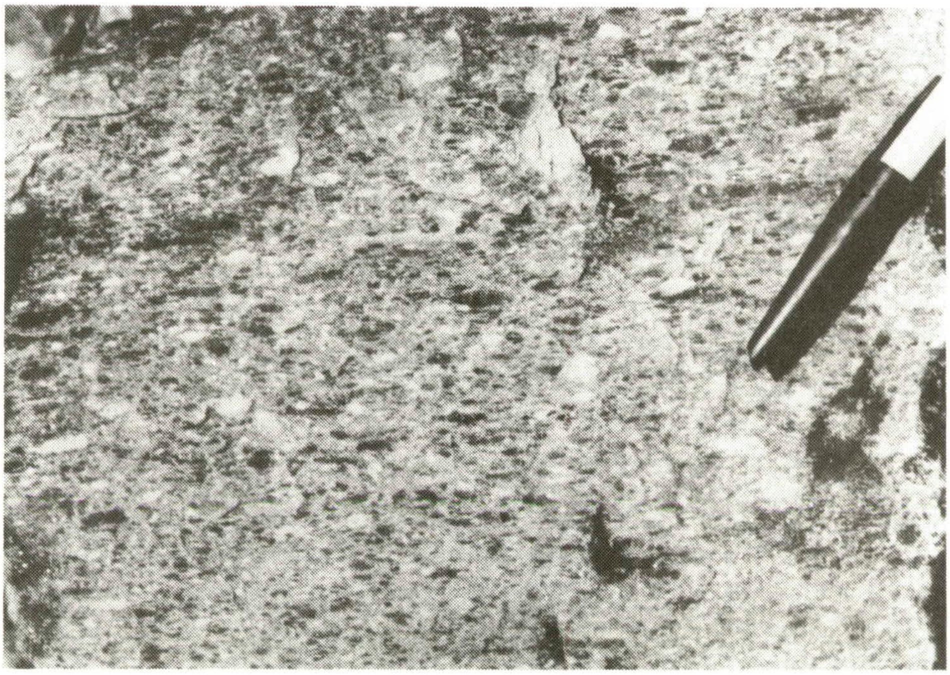


PLATE 18. Warrow Quartzite (near Range Road Lookout) viewed perpendicular to a strong quartz rodding. The quartz rods (dark grey) are flattened in the plane of S_2 (horizontal).

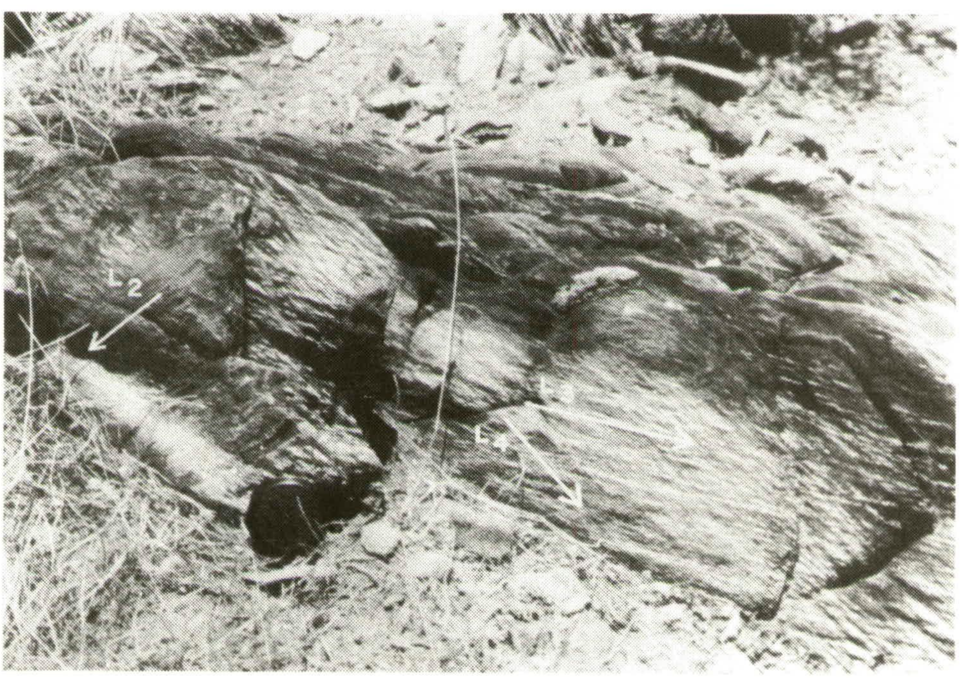


PLATE 19. Multiple lineations in Cook Gap Schist. L_2 (parallel to F_2 fold axes) pitches steeply to the left, F_3 crenulations pitch shallowly to the right, and F_4 crenulation axes pitch steeply to the right. Lens cap is 50 mm wide.

Typical Cook Gap Schist crops out in the creek at "A" (Fig. 22). It is a semipelitic rock of medium grainsize and consists of interbanded quartz-veined schist and psammitic schist with local bands rich in garnet. Quartz (30-60%), plagioclase (10-25%), muscovite (10-30%), biotite (10-20%), garnet (up to 50% in some bands but usually ca 1-2%), microcline, tourmaline, apatite, opaque minerals, sillimanite and zircon are the principal components. Structural features include the strong layer-parallel S_{1-2} schistosity; its crenulation by shallow northeast-plunging F_3 and slightly steeper eastnortheast-plunging F_4 (Plate 19); local, tight to isoclinal F_2 plunging shallowly southeast and characterised by a pervasive axial planar S_2 schistosity; a locally developed S_2 segregation schistosity (metamorphic/tectonic differentiation); and a quartz rodding lineation (L_2) parallel to F_2 axes.

Northwest of "A" and both structurally and stratigraphically below the schist is a massive amphibolite sill "B". This is a concordant body that can be followed along the same horizon for several kilometres. Chemically it corresponds to a quartz tholeiite and its boundary with schist is sharply defined although there is narrow zone a few centimetres wide characterised by mixed hornblende and biotite. The two dominant lithologies are schistose amphibolite and porphyroblastic amphibolite. Both consist of blue-green hornblende (50-60%), plagioclase (35-45%), sphene, opaque mineral, quartz and locally diopside.

Between "B" and "C" part of the sequence is absent due to lack of outcrop (it can be observed further northwest at "D"). Katunga Dolomite at "C" is a massive, white, dolomitic marble with poorly defined layering. It consists of dolomite (85-98%) with minor tremolite, opaque mineral, calcite, phlogopite and quartz.

At the bend in the creek "D" the base of the outcrop shows interbanded dolomite (sideritic) and iron formation here represented by a banded amphibole - magnetite quartzite. This horizon represents the top of the Katunga Dolomite and base of the Lower Middleback Jaspilite. The iron formation (chemical analyses indicate ca 20% total Fe) consists of alternating bands of quartz and amphibole ca 5-15 mm wide, and the common amphibole is grunerite. There is minor magnetite, hematite, garnet and biotite. On the southwest bank of the creek 10 m west, the iron formation is overlain by amphibolite (the same unit as at "B").

Following the left fork of the creek around to the west, one again continues to go down sequence. Below the dolomite at "E" is pelitic schist identical to schist interbeds of the upper Warrow Quartzite (STOP 20). This represents the top of the Warrow Quartzite and is a transitional unit immediately preceding the onset of carbonate deposition. Typical Warrow Quartzite outcrops a few metres further southwest in the creek at "F".

Continue down Range Road to T-junction.

60.0 T-Junction, turn left (northeast).

65.9 Glenville Stud, on the left, is one of the leading merino sheep studs on Eyre Peninsula. It is located at a spring, proclaimed Yabmana (meaning "fresh water"), which was the site of the original police station in 1853.
Turn left (north) at the crossroads.

74.3 T-junction, turn right (east).

76.8 T-junction, continue east.

79.7 T-junction, turn left (north).

81.7 T-junction, continue ahead to the north.

83 T-junction, turn right (east).

83.2 STOP 22 - MILTALIE MINE (Fig. 23).

Mining operations at the Miltalie Silver and Lead Mine ceased in 1910, but restarted several times between 1910 and 1914 when the mine was periodically dewatered. The Government put down three diamond drill holes in 1915-1916 (NB no drillcore has been retained) but results were discouraging and no further work has been done.

The mineralisation consists mainly of galena, malachite, and azurite and is closely associated with a dolomite and calc-silicate horizon separating quartzite to the east from migmatitic granite gneiss and pegmatite to the west. This particular horizon can be traced intermittently to the southwest where it hosts a number of other small silver, lead and copper prospects including Atkinson's Mine from which one parcel of ore assayed at 6,800 oz. fine silver/ton. The horizon is the calc-silicate horizon locally present at the base of the Warrow Quartzite and effectively represents the unconformity between Miltalie Gneiss and Hutchison Group.

Historical records of mining operations are contained in early Mining Reviews (Numbers 12, 15, 17, 18 and 20) and are summarised by Johns (1961). Bore records are contained in Mining Reviews 22-24 and indicate that while no significant mineralisation was intersected neither was the dolomite horizon!

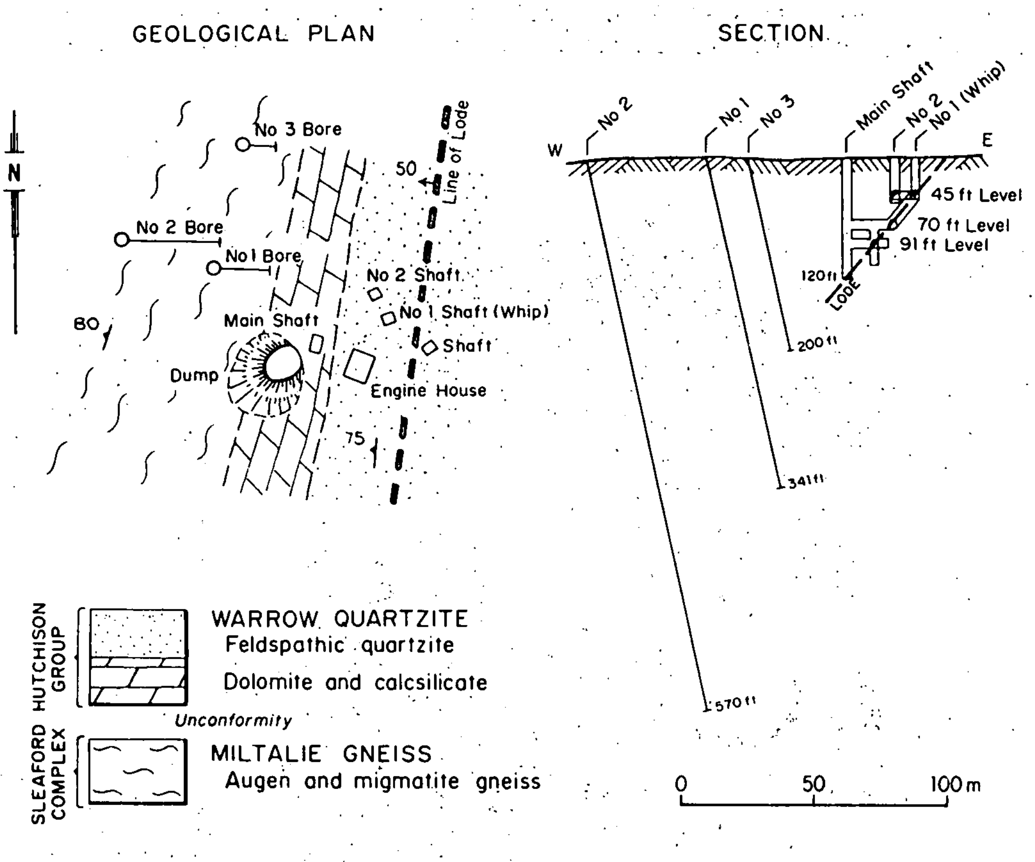


Fig. 23 Miltalie Mine; geological plan and section. Modified from Johns (1961).

83.4 Return to T-junction, turn left heading south.

86.7 T-junction, turn left (southeast).

88.7 T-junction, turn right and head southwest along road.

95.1 STOP 23 - FERNS QUARRY (Fig. 24).

This quarry at Ferns was used by the Highways Department to obtain dolomite suitable for road metal. It is located in a broad 200-400 m wide zone of Katunga dolomite exhibiting an incredible variety of tectonic fold and fold-related structures. The abnormal thickness of dolomite here is tectonic in origin, this region representing the hinge(s) of a major D_2 synclinorium (Fig. 24).

Dolomitic marble is the dominant lithology and ranges from a white, medium-grained, crystalline rock to a fine-grained, grey, recrystallised rock with lenses, bands, and pods of pink microcline and calc-silicate minerals. The microcline and calc-silicate bands behave much more competently than the enclosing dolomite, and have been tightly folded, attenuated and boudinaged during D_2 and D_3 (Plate 20). They exhibit a variety of structures including hinge zones thickened relative to limbs, fold cusps with dismembered limbs, lenticular boudins obviously representing highly attenuated limbs, and many other transposition-related structures. They reflect a marked ductility contrast between calc-silicate and dolomite.

On the southwest wall of the quarry, quartzites and gneisses of the iron formation facies have been folded into the carbonate. They include a banded diopside (±magnetite) quartzite, a massive diopside rock, banded diopside amphibole quartzite and a sideritic calc-silicate. The banded quartzites represent the equivalent lithology to iron formation facies quartzite at

PLATE 20. Folded and attenuated calcsilicate clast within dolomitic marble of the Katunga Dolomite at Ferns Quarry.

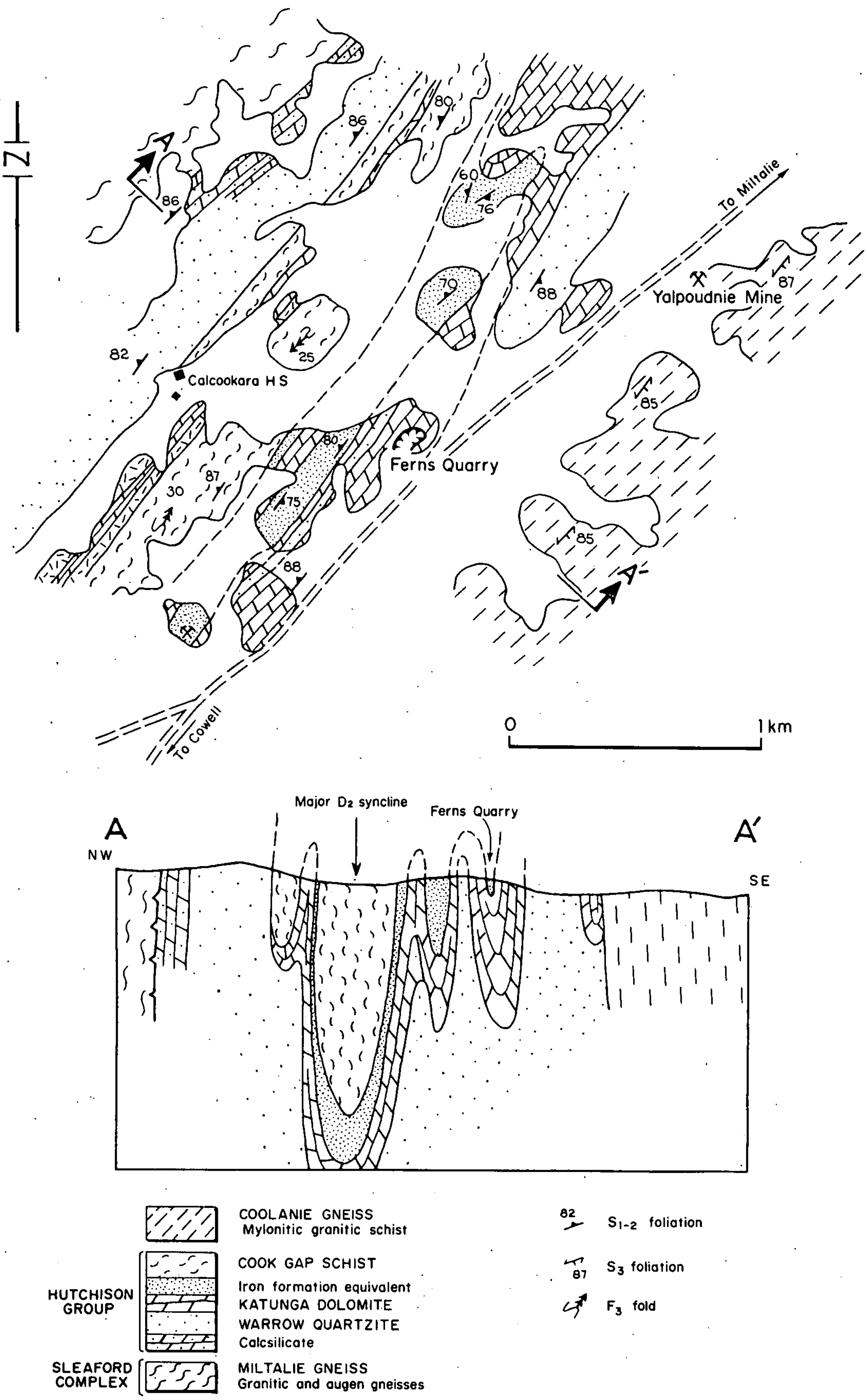


Fig. 24 Geological plan and section of the Ferns Quarry area.

Wangaraleednie, but because of a slightly higher metamorphic grade (the adjacent schists have been incipiently migmatised) diopside has replaced grunerite. Diopside in these rocks is often hedenbergitic indicating absorption of Fe into the pyroxene, and magnetite is generally restricted to the quartz bands. Local F_3 (and probably also F_2) have often localised magnetite and other opaque minerals in hinge zones, an important consideration for exploration in this region. Rb-Sr measurements on dolomites from this quarry are consistent with metamorphic ages reset during the Kimban Orogeny.

95.2 Leave quarry and proceed southwest back along road.

96.7 Y-Junction, turn left.

99.2 Turn left (southeast) and drive across the Calcookara Fault and Coolanie Gneiss zone (NB Calcookara means "brackish water").

100.9 STOP 24a - "PINEROW" - COOLANIE GNEISS (Fig. 25).

Separating high-grade granite/granite gneisses (to the east) from medium-grade metasediments (to the west) is a 3 km wide zone of highly-deformed and recrystallized mylonitic augen gneisses. This zone represents a major crustal shear zone and is known as the Coolanie Gneiss Complex. It was formed during D_3/D_M, at the same time as the Kalinjala Mylonite Zone, but exhibits a more homogeneous fabric deformed only locally by D_4 kink bands and lineaments. The homogeneity of the fabric probably reflects the original lithology which was granitic and possibly equivalent to the Carpa Granite.

While there is no obvious connection between this mylonite zone and the Kalinjala Mylonite Zone at Port Neill, it is possible that the two are, or were, connected. Early thoughts equated another mylonite zone near Cowell with the Kalinjala Mylonite Zone (Parker, 1980b), but structural juxtaposition of high-grade gneisses to the east of Coolanie and metasediments to the west (i.e. similar to relationships at Port Neill), suggest that the Coolanie and Port Neill zones may have been originally connected.

At this locality (beware of electrified fence!!), the Coolanie Gneiss is a coarse-grained augen gneiss with augen ranging up to, and greater than, 10 mm across (Plate 21). The augen are dominantly K-feldspar and have been microscopically deformed, whereas the matrix is composed of recrystallised quartz, feldspar and mica. The augen gneiss is poorly to non-lineated.

Also at this locality, on the northwest side of the outcrop, is some suboutcropping amphibolite. The amphibolite is medium-grained, pre-tectonic (pre-D_M) and is very strongly lineated, a lineation which is subparallel to lineations in augen gneiss at the next stop (Fig. 25).

Continue southeast along road.

101.9 STOP 24b - COOLANIE GNEISS (FIG. 25).

At this locality the Coolanie Gneiss is a fine-grained mylonitic augen gneiss with feldspar augen generally less than 5 mm in size, ovoid to lenticular in shape and enclosed in a ribbon-textured matrix of recrystallised quartz. The very strong fabric varies from an "S" to an "LS", to locally, an "L" fabric and is very constant in orientation, a feature characteristic of this zone. The foliation, $S_{m/3}$, strikes 045° dipping 80°SE and the rodding (defined by elongate quartz and feldspar aggregates) is subhorizontal to plunging 10° to 225° (Fig. 25).

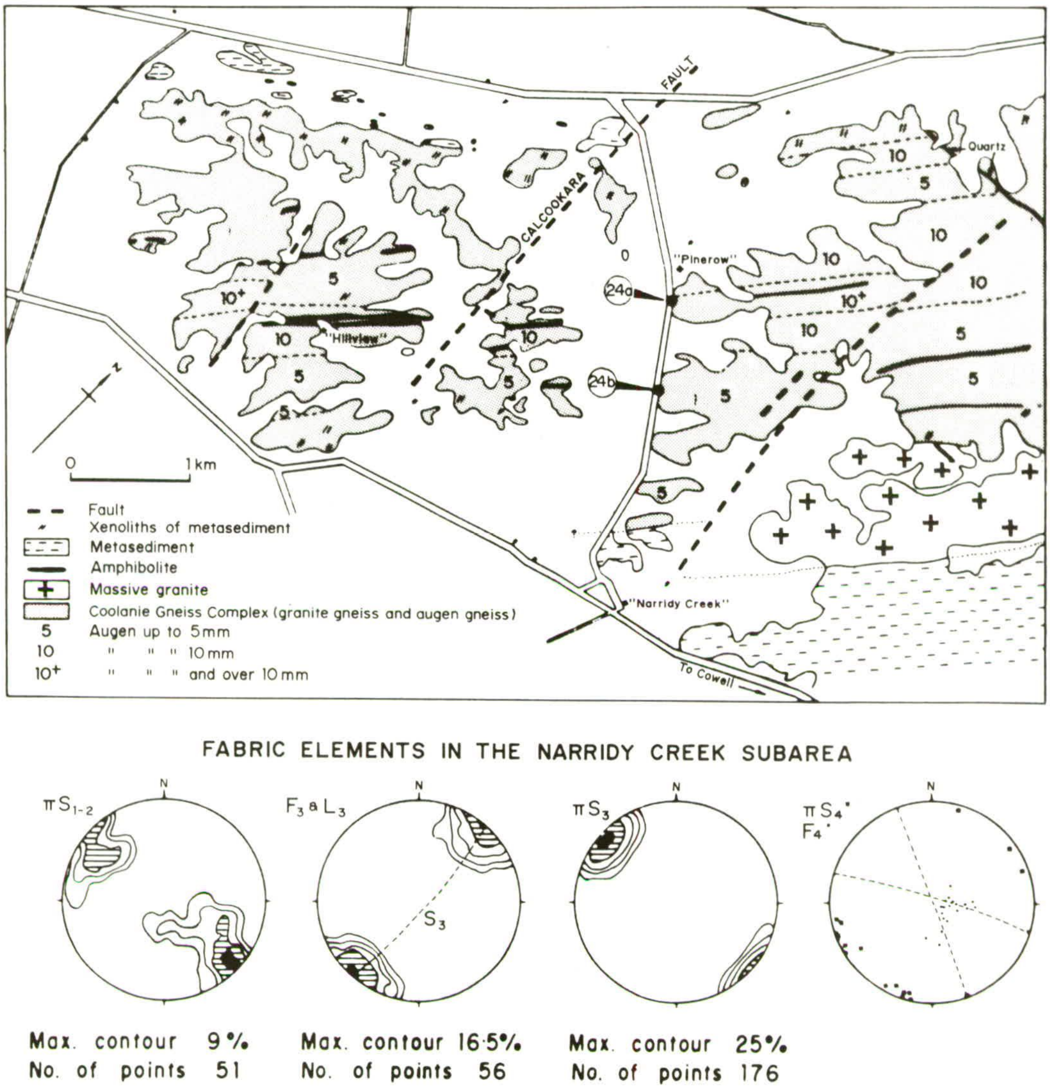


Fig. 25 Geological map of the Coolanie Gneiss in the Pinerow region.

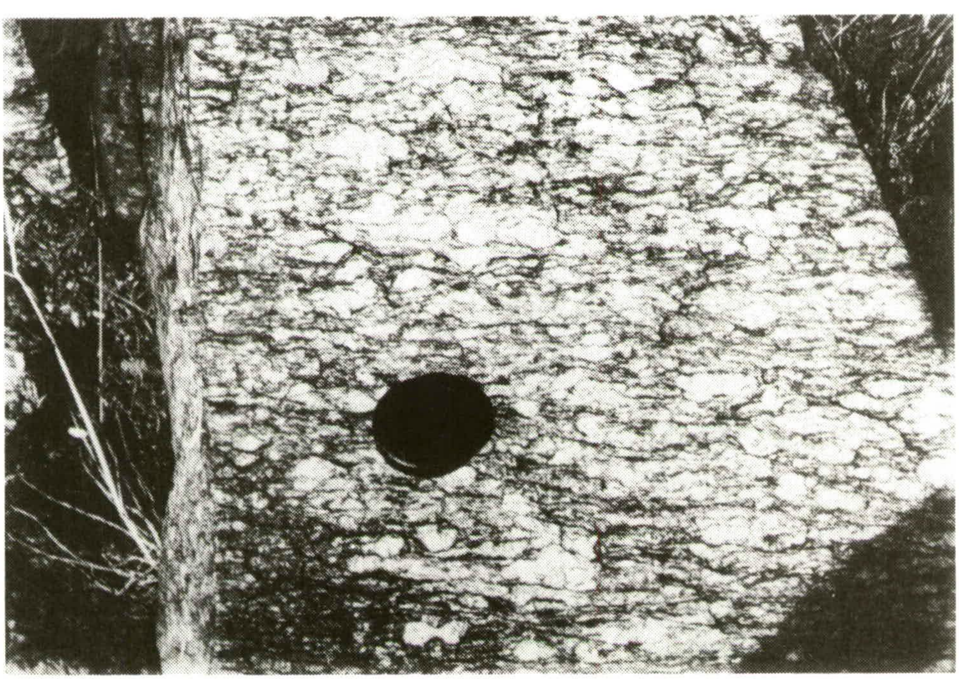


PLATE 21. Coarse-grained augen gneiss (the Coolanie gneiss) at "Pinerow". The lens cap is 50 mm wide.

Interspersed with the mylonitic augen gneiss are one or two bands, a few metres wide, of fine-grained "slaty" granite gneiss (slaty mylonite) in which no feldspar augen remain. These grade into the augen gneiss and represent zones of greater deformation and recrystallisation. By contrast to these bands, there are also bands of weakly gneissic pegmatite which show only minor deformation. There are some scattered, post-D_3 quartz veins which are related to D_4.

Continue southeast along dirt road until intersection with the main Cowell-Cleve highway.

104.3 Turn left (east) towards Cowell. The road winds through the Middle Camp hills (Fig. 26), firstly across Warrow Quartzite, then calc-silicates and dolomitic marbles at the base of Warrow Quartzite, and then across the Middle Camp Granite.

109.1 STOP 25 - MIDDLE CAMP (Fig. 26).

On the south side of the road in a stream bed opposite Middle Camp ruins, typical Middle Camp Granite crops out. It is a grey, medium-grained gneissic granite of adamellite to granodiorite composition and it is characterised by numerous narrow pegmatite veins (Plate 22). Two sets of pegmatites are present: an early veining parallel to the gneissic foliation and folded by F_3; and a later veining closely parallel to F_3 axial planes. Typical composition of the Middle Camp Granite is quartz (20-25%), plagioclase (45-60% and 20-30%), potassium feldspar (10-25% and 30-50%), biotite (5-10%), sphene, apatite, and minor accessories.

100 m downstream there are a few small xenoliths of former dolomite/calc-silicate enclosed within the Middle Camp Granite. The xenoliths are ca 0.5-1 m in size and have been completely altered by local metasomatism to massive tremolite/actinolite rock. They are surrounded by a rim of coarse pegmatite and there are other veins of pegmatite invading the tremolite. From petrographic evidence the alteration of the original dolomite postdates the main M_1 metamorphism. It suggests an early D_2 associated origin for the granite and this is confirmed by the presence of the S_2 gneissic foliation.

Continue east towards Cowell.

112.2 Just as the road winds down the Cowell Fault scarp, it crosses another major mylonite zone (Fig. 26) characterised by mylonitic layered gneisses, quartzite, and marble, and textures ranging from augen gneiss to slaty mylonite. The Cowell Flats are underlain by Cainozoic sediments dating back to Eocene lignites containing trace uranium. They are overlain by Miocene marine limestone and Plio-Pleistocene gravel, sand and clay.

115.0 T-junction, turn left towards Cowell.

221.2 Township of Cowell.

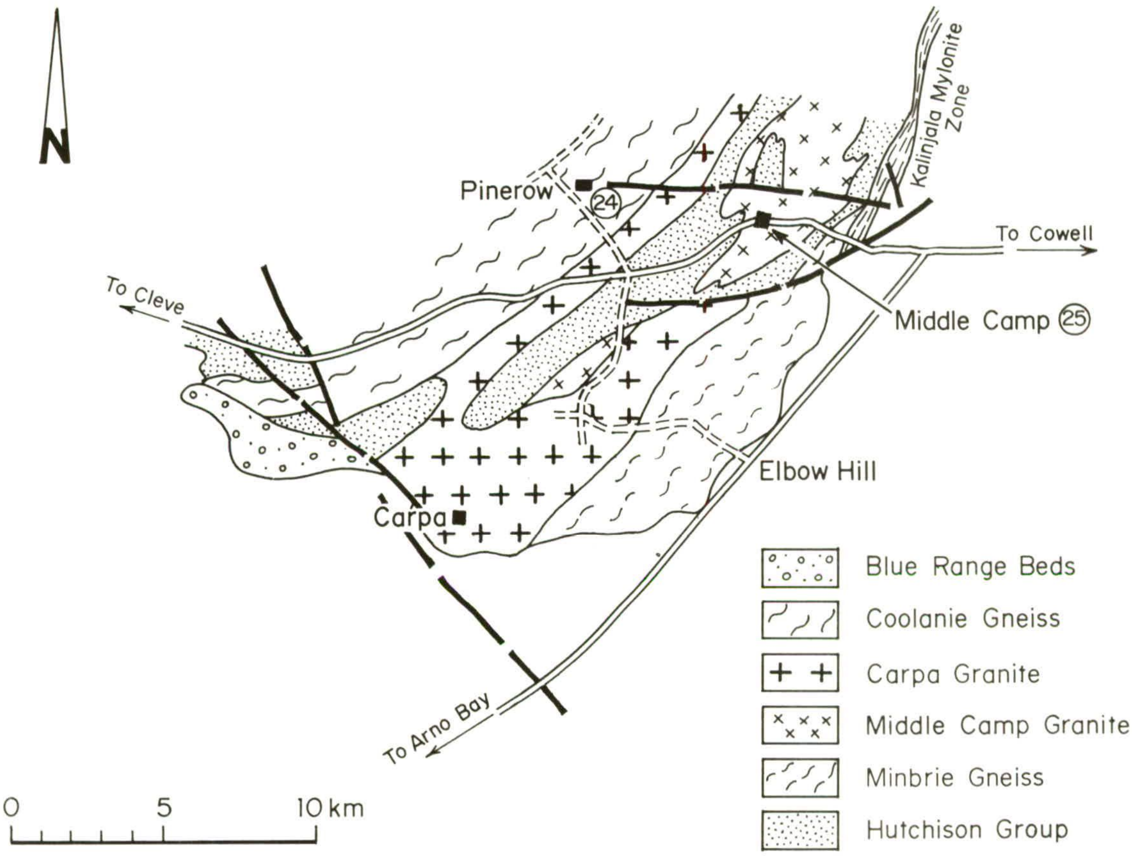


Fig. 26 Geology of the Middle Camp area showing the Middle Camp Granite, Minbrie Gneiss, Carpa Granite and Coolanie Gneiss.

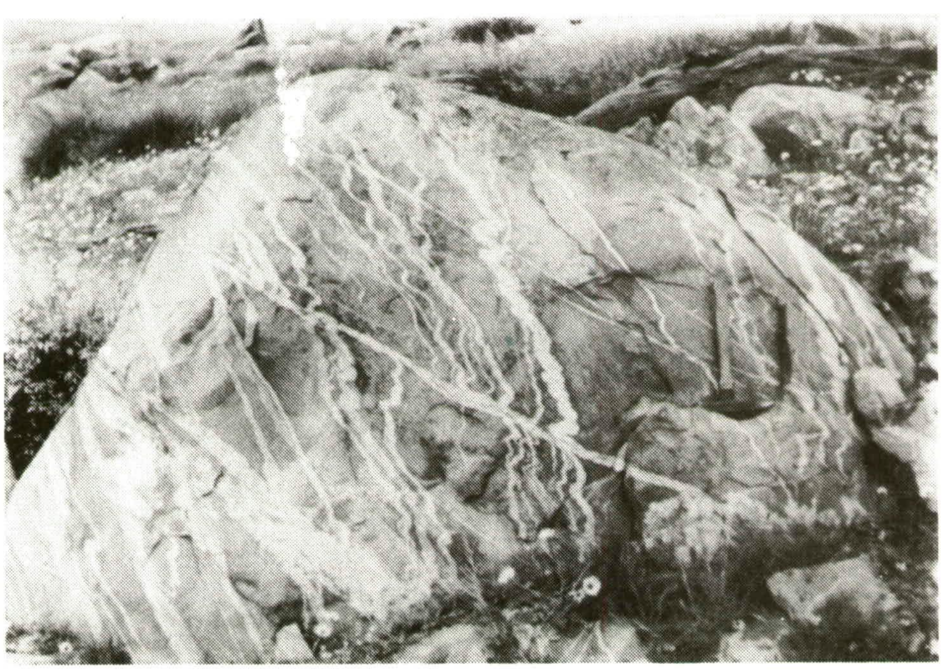


PLATE 22. Pegmatite veining within the Middle Camp Granite, Middle Camp.

DAY FIVE

Cowell - Moonabie - Iron Baron - Whyalla

Distance (km)

0.0 From Cowell head north and follow Alternate Highway 1 towards Whyalla.

27.1 The main road climbs up an E-W fault scarp onto the Charleston Granite (Fig. 2).

38.8 Good outcrop in road cutting of typical Charleston Granite.

44.8 Low hills immediately south and north of the highway (Fig. 27) are comprised of McGregor Volcanics (also formerly known as Moonabie Volcanics and Moonabie Porphyry).
T-junction, turn left (north) along dirt track.

45.2 STOP 26 - McGREGOR VOLCANICS, (Fig. 27).

The McGregor Volcanics are a bimodal suite of acid, ash-flow tuffs and basaltic lava flows. The oldest unit exposed is a rhyolite, exhibiting spherulitic devitrification textures. This is succeeded by a series of both porphyritic and non-porphyritic rhyolitic to dacitic volcanics. Porphyritic dacites contain abundant plagioclase phenocrysts. The groundmass usually consists of a devitrified mosaic after glass showing excellent compaction and flattening of the shards, indicate that the volcanics are welded ash-flow tuffs.

Few of the basalts, which occur as thin interbands, preserve their primary mineralogy. However, they do show a wide textural range including ophitic, intersertal and hyalopilitic textures indicative of their lava-flow origin.

Andesites, which bridge the petrological characteristics of the bimodal suite, are absent. Giles et al. (1979) have suggested that the acid and basic volcanics have different origins. The primary basic magma may have been derived by relatively shallow (<60 km) wet melting of the mantle in which olivine, orthopyroxene and clinopyroxene are residual. The primary dacite magma probably originated by partial melting of a dry, basic granulitic lower crustal source in which plagioclase, clinopyroxene, orthopyroxene and magnetite are residual. (The above descriptions of volcanics and their origins are from Giles et al. (1979) and Giles, pers. comm.).

Intercalated with the volcanics are heavy mineral-laminated quartzites (with dispersed quartz pebbles) and volcaniclastic grits.

The age of the McGregor Volcanics is of considerable interest since they are believed to have formed during the Kimban Orogeny maybe between the D_2 and D_3 deformations.

Rb-Sr whole rock analyses from two separate localities (see Fig. 27) yield the following isochron regressions (Webb et al., 1986).

No. of Samples	MSWD	Age	I.R. $^{87}Sr/^{86}Sr$
4	1.28	1645 ± 15 Ma	0.7159 ± 0.0021
9	1.93	1615 ± 29 Ma	0.7073 ± 0.0010

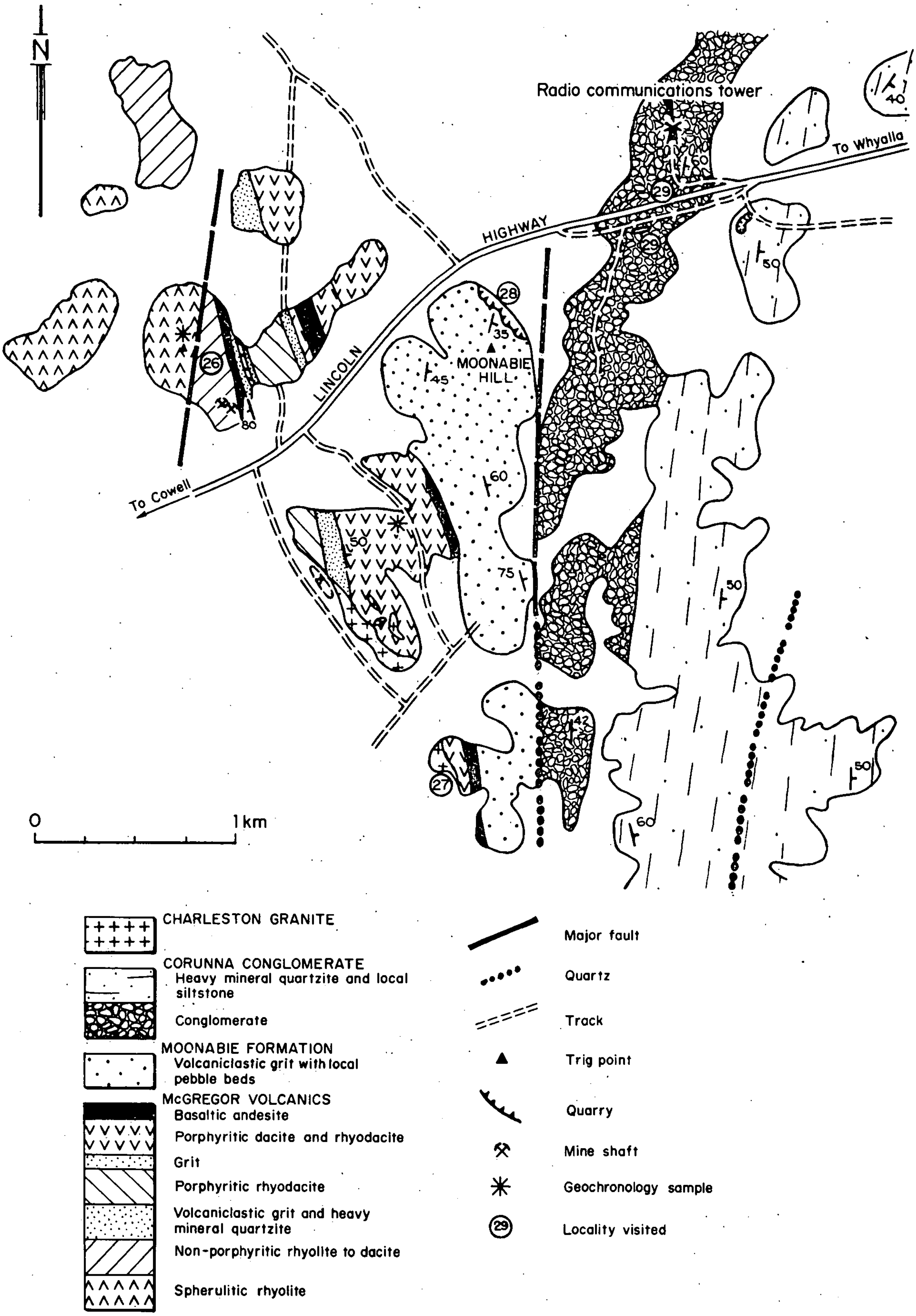


Fig. 27 Geology of the Moonabie Range area, after mapping by N.M. Lemon (pers. comm.), C.W. Giles (pers. comm.) and Parker (1983).

For the first group of 4 samples, the high initial $^{87}Sr/^{86}Sr$ ratio indicates either assimilation of material enriched in ^{87}Sr or updating during a tectonothermal event, maybe related to intrusion of the Charleston Granite. U-Pb zircon analyses for the McGregor Volcanics yield an upper intercept of 1755 ± 12 Ma (MSWD = 23.9) for six conventional and one single grain measurements. This poorly fitted chord is controlled by significantly discordant fractions, thus the nearly concordant analyses would be expected to give a better indication of the true age of the zircon. The three most concordant analyses, including the only slightly discordant "single" grain analysis indicate a younger age of ca 1730 Ma for the McGregor Volcanics.

45.6 Return to the highway, turn right (south-west).

45.7 T-junction, turn left (south) along dirt track. Proceed along dirt track around hillside.

47.6 STOP 27 - CHARLESTON GRANITE (Fig. 27).

The Charleston Granite is a massive, post-orogenic, megacrystic granite forming a large pluton with an exposed surface area covering 300 km^2. Potash feldspar phenocrysts, which are mostly microcline, are up to 3 cm in length and comprise about 30-40% of the rock. Some of the phenocrysts are zoned with plagioclase rims. Other constituents are medium-grained (locally coarse to 1 cm) and consist of white to greenish plagioclase 10-15%, biotite 5% and clear, greyish quartz.

Charleston Granite intrudes the Moonabie Formation and McGregor Volcanics (Miles, 1954; Smale, 1966; Giles et al., 1979). This is evident from aplitic and granitic veining within the volcanics, recrystallisation of the groundmass within the acid volcanics, and hornfelsing of the volcaniclastic grit within the Moonabie Formation. Also, along the contact with the volcanics, there is a zone less than 20 cm wide in which the granite is finer grained and phenocryst-poor. The adjacent volcanics are also veined by quartz and epidote which emanate from the granite.

Hornfelsing, epidotisation and quartz + epidote veining also occur in Corunna Conglomerate at the south end of Moonabie Range indicating that the Charleston Granite postdates sedimentation of the Corunna Conglomerate (Giles et al., 1979).

Geochronology of the Charleston Granite gives conflicting results. Compston et al. (1966), from single total rock, biotite and apatite Rb-Sr analyses, calculated an age of 1556±30 Ma (recalculated value using $\lambda^{87}Rb = 1.42 \times 10^{-11} y^{-1}$) which is in close agreement with three K-Ar biotite dates of 1532, 1567 and 1568 Ma reported by Webb et al. (1986). However, conflicting with these values are some additional Rb-Sr analyses on 8 total rock samples (Webb et al., 1986). These samples define an isochron age of 1445±39 Ma with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7133 ± 0.0058. U-Pb zircon dating, recently completed (Fanning et al., 1986a) supports the older Rb-Sr age.

Conventional U-Pb zircon measurements are significantly discordant, recording a poorly defined upper intercept of 1645 ± 46 Ma (MSWD = 10.5), however carefully selected clear euhedral zircons from the least magnetic -92+72 µm fraction yield a near concordant result and suggest that a more realistic estimate for the Charleston Granite is 1583 ± 26 Ma. This date will be refined by further "single" grain style analyses.

Return to highway.

49.5 Turn right and proceed northeast along highway.

50.6 Turn right (southeast) and proceed along dirt track to quarry.

46.0 STOP 28 - MOONABIE QUARRY (Fig. 27).

Overlying and probably interbedded with extrusives of the McGregor Volcanics are volcaniclastic grits of the Moonabie Formation (Smale, 1966; Giles et al., 1979). The volcaniclastic grits are massive, dark grey and have an immature, medium to coarse-grained matrix of quartz, feldspar, sericite and volcanic rock fragments (Plate 23). Bedding is very poorly defined. Rounded to subrounded pebbles are dispersed in the matrix and rarely form distinct lenses. Volcanic clasts dominate and include porphyritic and non-porphyritic rhyolites to dacites typical of the McGregor Volcanics. Chert clasts are common and vary in grain size from cryptocrystalline to about 0.2 mm.

The volcaniclastic grits are quartz-veined and have been hornfelsed by intrusion of the Charleston Granite.

Return to highway but continue east along dirt track into roadside rest area.

51.7 STOP 29 - MOONABIE RANGE - ROAD CUTTING (FIG. 27).

Warning: please take care and watch for cars.

The Corunna Conglomerate (Miles, 1954; Smale, 1966; Lemon & Gostin, 1975; Parker, 1980a) structurally overlies the Moonabie Formation. The basal conglomerate, well exposed in the road cutting, contains abundant rounded to well-rounded clasts ranging in size up to 40 cm (Plate 24). Clast types include fine- to coarse-grained sandstone and quartzite (not typical of the Warrow Quartzite), banded iron formation (Upper and Lower Middleback Jaspilites), acid volcanic (McGregor Volcanics), volcaniclastic grit (Moonabie Formation), and minor silicified, crenulated schist, mylonite and granite. The matrix is sandy and very poorly sorted. Individual conglomerate beds (up to 20 m thick) are massive to poorly layered and are interlayered with well-bedded, fine-grained sandstones. Developed throughout the unit is a mauve and yellow to off-white mottling.

East and south of here the conglomerates are overlain by a more-mature, cross-bedded heavy-mineral-bearing sandstone with local purple siltstone beds. The sandstone is of likely marine origin but is interbedded to the east with a very rapidly deposited talus breccia suggesting that the basin of deposition may have been fault-bounded along its eastern margin.

Continue northeast along the main highway.

75.4 Turn left (west) towards Kimba and Iron Baron.
Through this region the road traverses a series of gneissic granites (Wertigo Granite) of syn-Kimban origin. Wertigo Rockhole (Miles, 1954) is off to the north of the road intersection and the "Tor" granite (Miles, 1954) outcrops are visible to the right about 2.5 km from intersection.
About 7.6 km from intersection, gneissic equivalents of the Myola Volcanics, crop out on the left side of the road.
Further west, where the power line crosses the road, Broadview Schist crops out in a small quarry just to the north of the gate.

86.0 Follow main road towards Iron Baron.

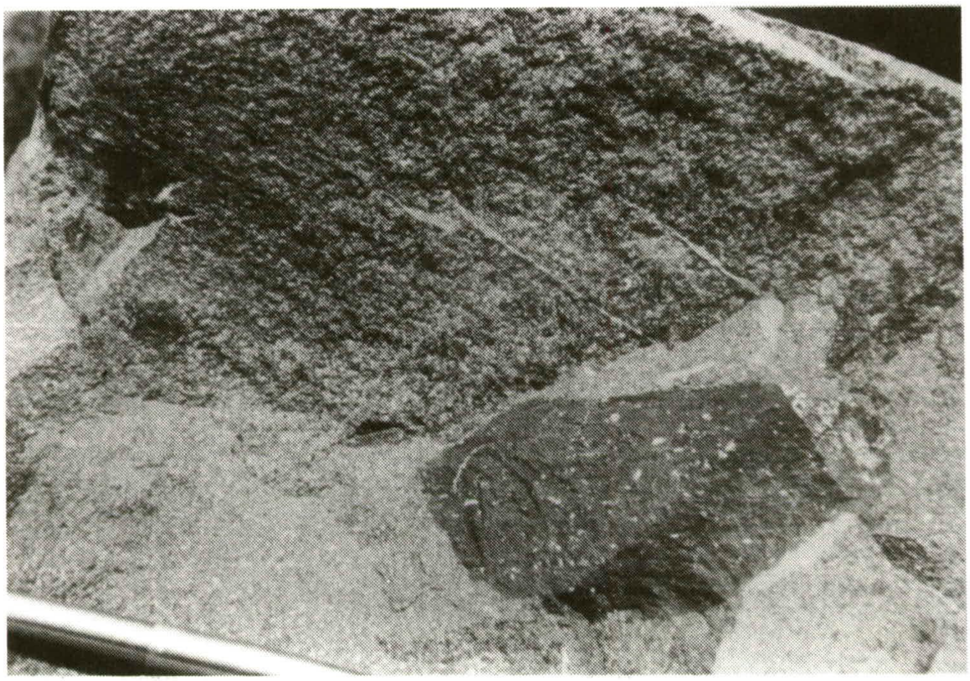


PLATE 23. Angular clast of McGregor Volcanics basalt within volcanoclastic arkose of the Moonabie Formation. Moonabie Quarry.

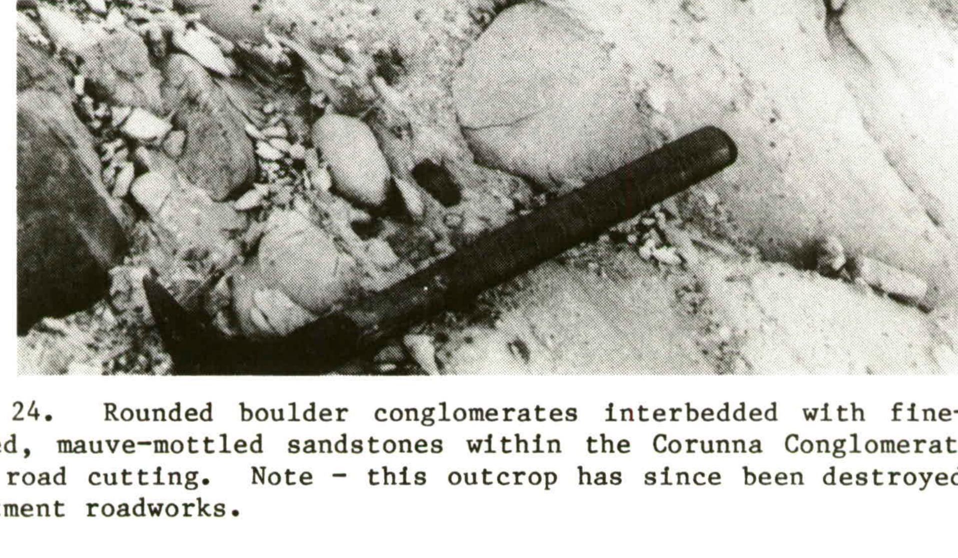


PLATE 24. Rounded boulder conglomerates interbedded with fine- to medium-grained, mauve-mottled sandstones within the Corunna Conglomerate. Moonabie Range road cutting. Note - this outcrop has since been destroyed by Highways Department roadworks.

88.5 Turn right onto track and proceed to Myola H.S. Conformable amphibolites within the Broadview Schist crop out on left of track.

95.1 Myola H.S. - proceed to right and follow map to location A. (Permission to enter should be sought at the homestead before proceeding).

96.8 STOP 30 - MYOLA VOLCANICS (Fig. 28).

This is the type locality, and geochronology sampling site, for the Myola Volcanics (Parker, 1983). At this location, the Myola Volcanics consist of a series of interbanded porphyritic rhyolites (A), fine-grained felsic gneisses and fine-grained amphibolites (B) intruded by a northwest-trending dolerite dyke(C). The rhyolites, gneisses and amphibolites are all foliated, dip steeply to the east and locally are very strongly lineated. Both the foliation and lineation are locally folded by sinistral folds and crenulations plunging shallowly to the north-northwest.

Near the trig point (A) the dominant lithology is a massive, foliated, and often strongly lineated, porphyritic rhyolite. It is grey to greyish-pink in colour, and fine-grained with scattered phenocrysts of potassium feldspar, plagioclase, lesser quartz and rare mafic mineral (now altered to opaques, chlorite and leucoxene). The phenocrysts are at least partly recrystallised and they are contained in a fine-grained groundmass of recrystallised quartz and feldspar. Mineral composition is potassium feldspar (35-40%), quartz (25-30%), plagioclase (25-30%), opaque iron oxide (3-5%) and trace chlorite, leucoxene, carbonate, and zircon (Fig. 28).

To the east, massive rhyolites are interbanded with very strongly foliated and lineated felsic gneisses and amphibolites (B). The felsic gneisses are fine-grained metamorphic rocks, pale pink in colour and consist of potassium feldspar (25-45%), quartz (55-25%), plagioclase (15-25%), and minor iron oxide, muscovite, biotite, zircon, and apatite. They are generally even-grained, but there are scattered aggregates of recrystallised quartz and feldspar which may have been derived from recrystallisation of highly deformed, primary phenocrysts. Petrographically there is little else to confirm a volcanic or sedimentary origin for these gneisses but chemically they are similar to, though slightly more siliceous than, the rhyolites (Fig. 28). The fine compositional layering and rodding obvious in hand specimen is tectonic in origin and defined largely by elongate and flattened aggregates of quartz and feldspar.

Amphibolites interlayered with the gneisses are also fine-grained, foliated and often lineated. They consist of hornblende (35-60%), plagioclase (55-30%), and minor opaque iron oxide, sphene, apatite, sericite, epidote, biotite, quartz and potassium feldspar. Textures suggest that the amphibolites were derived from either fine-grained dolerites or coarse-grained basalts.

On the eastern end of the type outcrop, the Myola Volcanics are intruded by a medium-grained dolerite dyke with well-preserved ophitic textures. The dyke is vertical, trends northwest and cuts across the gneissic layering of the volcanic sequence, clearly post-dating the main tectonic events responsible for the formation of the gneissosity and its subsequent folding. The dyke is probably related to the swarm of northwest-trending dykes evident on the Stuart Shelf and other areas of the Gawler Craton.

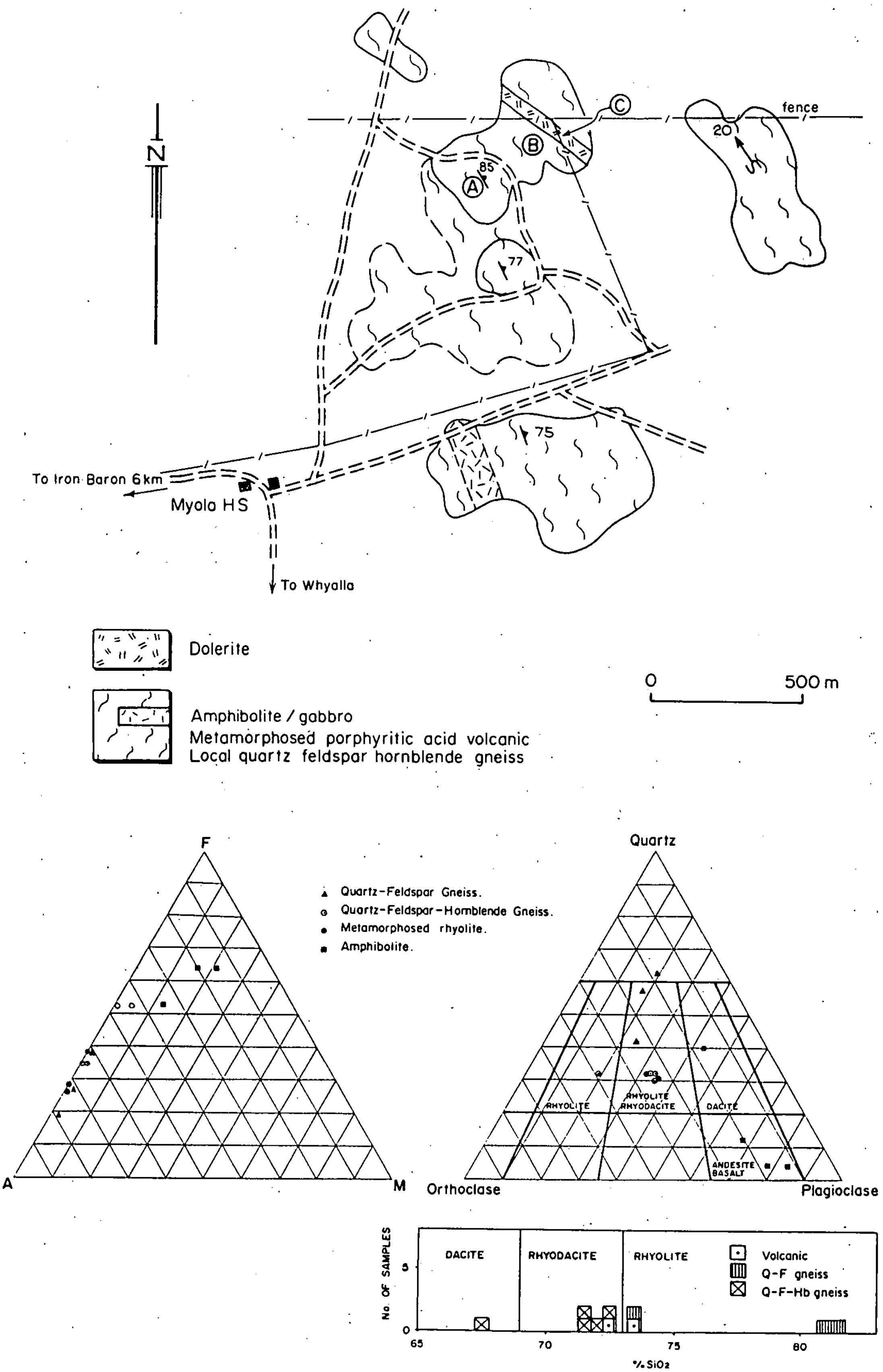


Fig. 28 Type locality plan and, chemical and modal compositional diagrams, for the Myola Volcanics.

Just east of the dyke there is an outcrop of pegmatite. This resembles other pegmatites prominently developed southeast of Myola and represents the development of the gneissic Wertigo Granite. The pegmatites and lenses of gneissic granite form sill-like bodies interspersed with felsic gneisses of the volcanic sequence.

The Myola Volcanics appear to have been deformed by D_2 of the Kimban Orogeny, but not necessarily by D_1. A combination of five conventional and one "single" grain U-Pb analyses of zircons collected from site A yields a well defined upper concordia intercept of 1791 ± 4 Ma (MSWD = 1.88) which records the age of crystallisation of the zircons in these volcanics (Fanning et al., 1986a).

98.5 Return to Myola H.S. and proceed to west towards Iron Baron.

105.3 At Iron Baron continue through township and turn left at Iron Knob turnoff along road into the quarry (Fig. 29).

106.3 STOP 31 - IRON BARON (Fig. 29, Plate 25).

NOTE: to visit Iron Baron, first enquire with BHP.

These notes have been extracted from notes written by John Elestheriou of the BHP Co. Ltd. (in Parker et al., 1981).

The Middleback Ranges Iron Ore Operation

In 1899 the B.H.P. was granted mining leases at Iron Monarch and Iron Knob to mine flux for smelting ore from Broken Hill at Port Pirie. Initially the ore was transported by bullock dray to Port Augusta and then railed to Port Pirie but with the building of a tramway ore was railed to Hummock Hill (later called Whyalla) and barged across the gulf to the smelters at Port Pirie. Leases were extended in 1920 to include the Iron Prince and Iron Baron areas to the south. These deposits became the basis for the establishment of blast furnaces and steelworks at Newcastle (1915) and Whyalla (1941, 1965). Total production of ore from the Middleback Ranges is 170 million tonnes to end of 1980 and ore is currently being mined at approximately 2.7 million tonnes per annum.

Geology

The Early Proterozoic Middleback Subgroup of the Hutchison Group is represented by dolomites, iron formations and schists in the Middleback Ranges. In the Iron Prince area the Middleback Subgroup is underlain by weathered muscovite-sericite-quartz-feldspar granitic gneisses and feldspathic quartzites with, along the western flanks, common cataclastic and microfaulting textures (this represents the northern extension of the Kalinjala Mylonite Zone).

The lowest unit of the Middleback Subgroup is the Katunga Dolomite. It comprises a sequence of ferroan and ankeritic dolomites interbedded with chlorite schists, becoming very iron enriched near the dolomite-iron formation boundary.

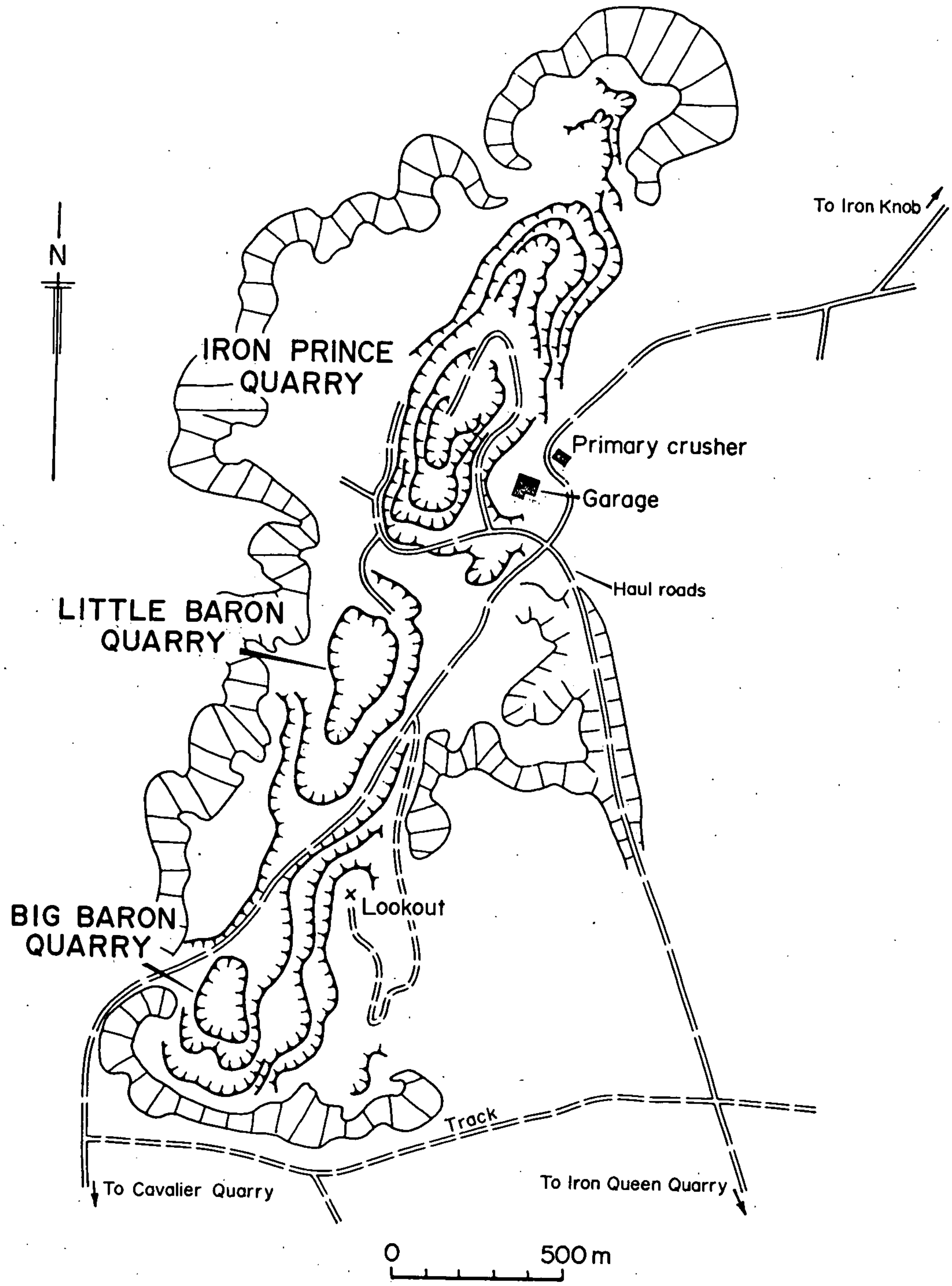


Fig. 29 Mine plan of Iron Baron and Iron Prince, November 1980.

PLATE 25. View of the Iron Baron Mine.

PLATE 26. Well-layered silicate facies iron formation of the Lower Middleback Jaspilite. Iron Baron Mine.

The dolomite is overlain by the Lower Middleback Jaspilite which is a sequence of banded iron formation approximately 350 m thick but exhibiting marked variations along strike. The Lower Middleback Jaspilite comprises pyritic dolomite at the base overlain by carbonate ore (a magnetite-carbonate mixture), magnetite-talc schist, talcose carbonate-magnetite jaspilite, magnetite-amphibole jaspilite, grunerite schist and magnetite-siderite amphibole jaspilite (Plate 26). All these rocks are oxidised and silicified near surface and appear in outcrop as hematite and limonite jaspilites.

High grade iron ore appears to have resulted from supergene enrichment of carbonate ore, magnetite-talc schist and to some extent the dolomite. Typical assays are listed in Figure 30.

South of Iron Baron the Cook Gap Schist and the Upper Middleback Jaspilite overlie the Lower Middleback Jaspilite found in the Iron Prince.

The occurrence of primary ore in the Iron Prince is structurally controlled by a tight north and northeast trending syncline crossfolded on a NW-SE axis resulting in common pitch reversals. Faulting is common along the limbs of the major northerly folds. Sill and dyke-like emplacement of syn- and post-tectonic intermediate and basic rock types (since altered to amphibolites and clays) intrude the mine sequence.

Ore types in the mining areas are blue and red hematite, lesser limonite and minor magnetite with a minimum ore cut-off set at 55% Fe. Potentially beneficiable material below this grade are enriched jaspilites, enriched amphibolites, scree and dumped contact material.

All the iron ore from the Iron Baron area is pelletized before use. Red ore, because of its fine grain size, is restricted to 10% of the pelletizing feed. Impurity levels that require monitoring are alumina, zinc, phosphorous and silica.

Iron Prince and Iron Baron Operations

Production at the Iron Baron and Iron Prince mines is scheduled on a two shift, five day week basis. Blast hole drilling is carried out by a Bucyrus Erie 45R track mounted drill. The blasting operation uses an ANFO mixture to produce approximately 50-100 000 tonnes per blast.

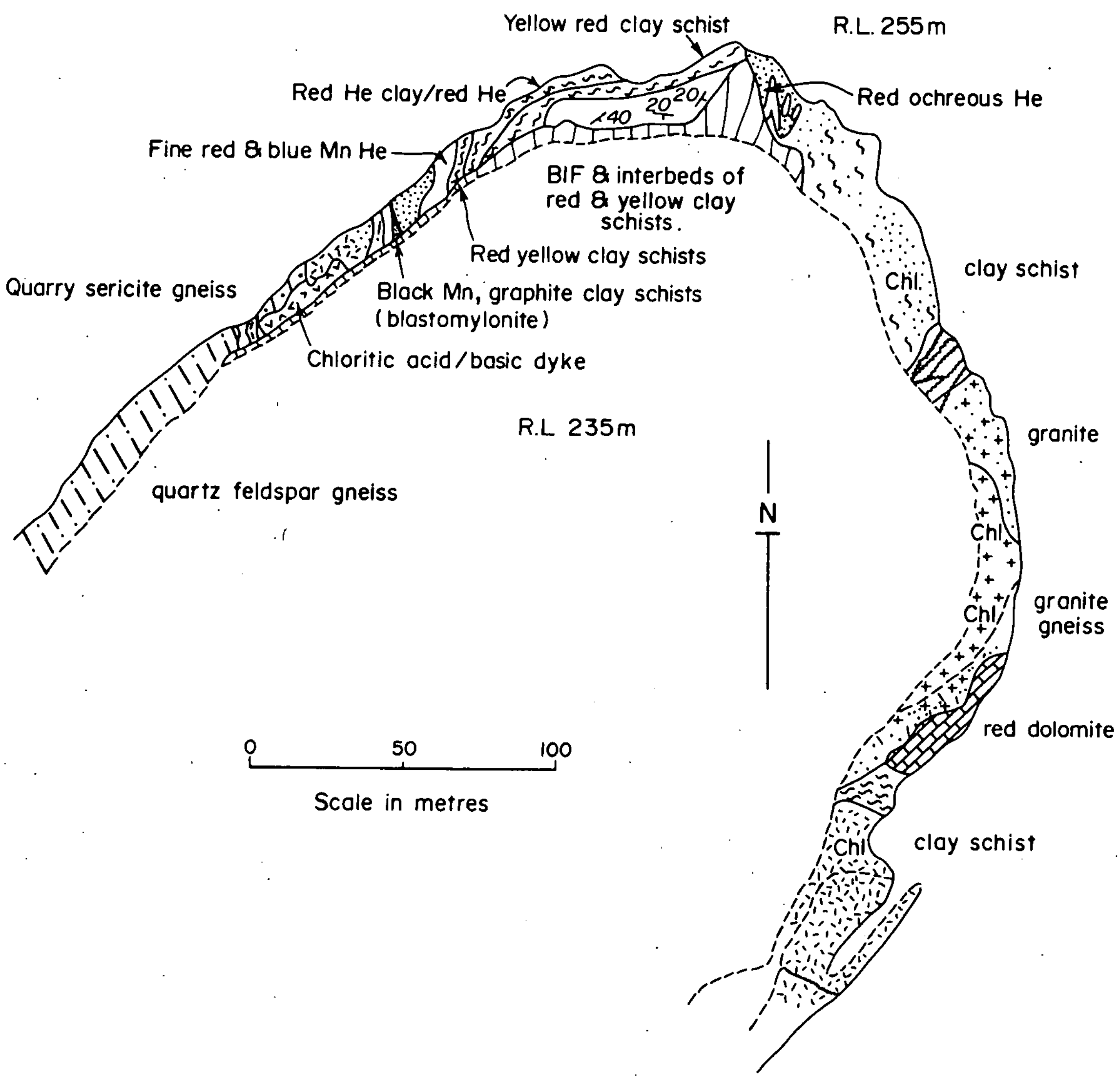
Two 151M Marion 4.5 m^3 electric shovels and two Michigan 475 rubber tyre loaders load broken ore into eleven 65 tonne Haulpack trucks for transport to the crusher. The average haulage distance is 1.6 km.

The ore goes through a primary jaw crusher and two secondary gyratory crushers for reduction to minus 50 mm and is stored in four 1 500 tonne storage bins. This is loaded on 45 wagon ore trains of 2 700 tonnes capacity for the 47 km trip to Whyalla.

The workforce at the Iron Baron/Iron Prince mine is approximately 145. The company township of Iron Baron has a population of 250 people; facilities include company homes, a general store, community hall, swimming pool, primary school, quarters for 100 single men, and a licenced recreation club. Similar facilities exist at the Iron Knob government township which has a population of approximately 450.

Return out of quarry, back through Iron Baron township, and follow main road to Whyalla - Kimba road.

125.5 T-junction, turn right and proceed west toward Kimba.



	Fe	SiO_2	Al_2O_3	LOI	P	CaO	MgO	Mn	S	T_1O_2	Cu	Zn	K_2O
High grade ore	68·6	0·7	0·5	0·7	0·1	0·1	0·01	0·02	0·01	0·05		0·02	0·01
Microgranite	1·2	77	13	3·7	0 01	0·1	2·3	0·1	0·01	0·01	0·015	0·005	
Sericite quartzite	1·8	72	15	2·1	0·02	0·1	1·4	0·1	0·02	0·2	0·005	0·005	
Barren Dolomite	15·2	13·8	4·7	27·4	0 02	16·2	13·7	0·9	0·03	0·32		0 03	
Lean jaspilite	27·4	52·3	2·9	3·3	0·04	0·15	0·42	1·24	0 02	0·19		0·07	
Central Intrusive	22·8	27	19	10·2	0·19	0·1	2·8	1·3	0·01	5·2		0·34	

Fig. 30 North Iron Prince, Middleback Ranges RL235-355 m.

160.0 Turn left of main road into small picnic area.

STOP 32 - REFUGE ROCKS (Fig. 31).

The outcrops at Refuge Rocks consists of Lincoln Complex, Minbrie Gneiss, intruded into Hutchison Group metasediments. The Minbrie Gneiss at this locality is composed of three phases of granitic gneiss, all with a moderately intense gneissic fabric equated with S_{1-2} within the metasediments.

The earliest intrusive phase is a medium, even-grained granodioritic gneiss containing highly deformed mafic xenoliths. This has been intruded by a coarse-grained megacrystic granite, subsequently deformed to produce an augen gneiss. The third phase is a moderately well foliated dioritic gneiss similar in appearence to the Middle Camp Granite. The contacts of this gneiss with the augen gneiss are typically obscured by layer-parallel pegmatites, although at one locality the intrusive contact with the augen gneiss can be observed.

The augen and diorite gneisses exhibit tight, subangular F_{2-3} folds trending towards 015° with wavelengths of 10-30 cm. The augen gneiss exhibits marked strain variation of the megacrysts around the folds, with megacrysts becoming increasingly flattened from the hinge zones to the limbs where zones of banded gneiss can be seen. Near the eastern end of the outcrop, narrow (<1 m) layer-parallel shear zones (D_3?) produce intense zones of banded gneiss within the augen gneiss. The S_{1-2} layering of the gneisses is commonly folded by minor D_4 angular kinks, trending towards 040°.

The gneisses are intruded by at least three generations of pegmatite/aplite veins. The earliest are subcordant to the foliation and are folded by the F_{2-3} folds. The second generation is parallel to the axial planes of the F_{2-3} folds and represents minor localised partial melt. Both of these are intersected by a massive, unfoliated pegmatite. At the western end of the outcrop, Warrow Quartzite has been intruded by both the augen and dioritic gneisses, producing an agmatite of quartzite rafts within pegmatite-veined migmatitic gneisses (Plate 27). The augen gneiss is similar in appearance to the megacrystic granitoids of the Donington Suite (STOP 3), which has been dated at 1850 Ma. By extrapolation, it is suggested that the deposition of the Hutchison Group metasediments occurred prior to 1850 Ma.

Return east along main road to Whyalla.

238 Enter city of Whyalla.

240.6 Turn left on to Norrie Ave.

243.3 Proceed over bridge.

243.6 Turn left at T-junction.

247.7 Turn right into quarry.

STOP 33 - MOUNT LAURA (Fig. 32).

At Mount Laura, Pandurra Formation unconformably overlies folded Middle Proterozoic sediments. The spectacular angular unconformity is best viewed at the northwest end of Mount Laura from the Quarry Industries Ltd. quarry (Section A-B, Fig. 32).

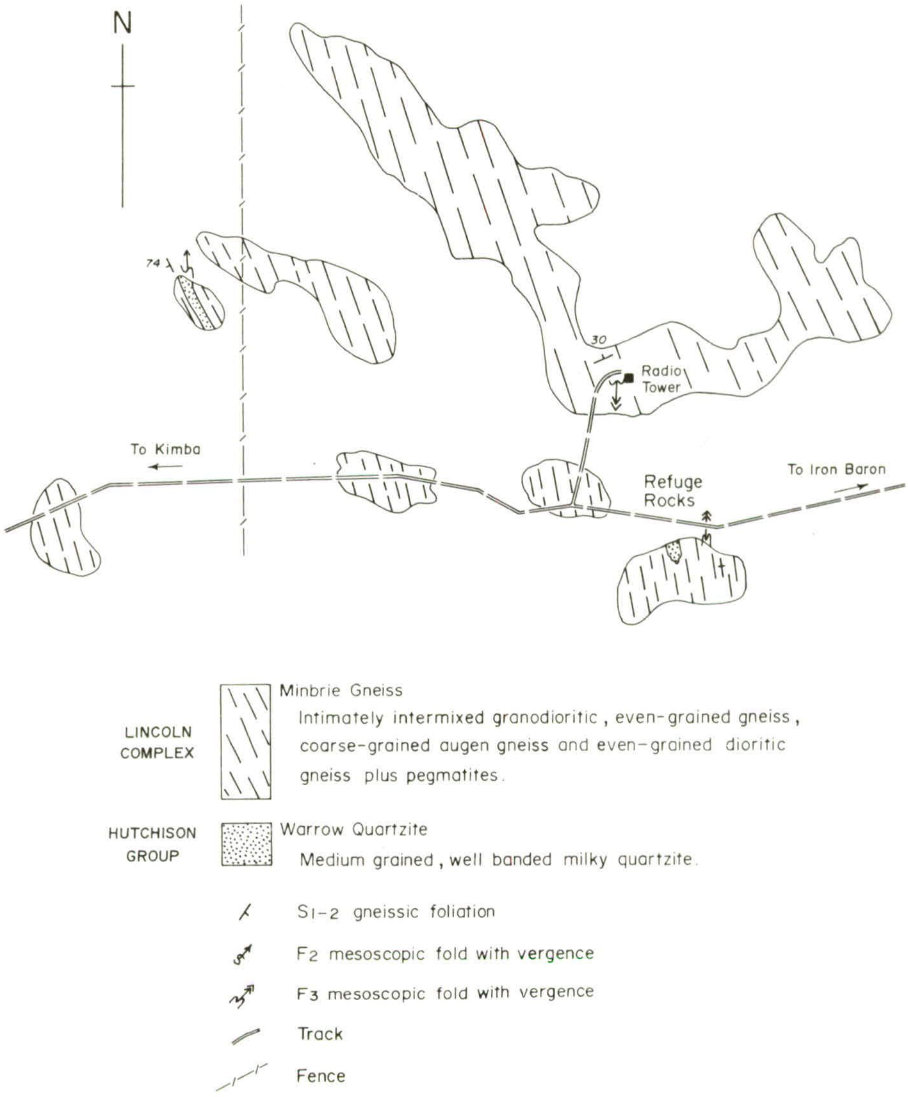


Fig. 31 Geological plan for Refuge Rocks.

PLATE 27. A boudinaged raft of well-layered Warrow Quartzite within foliated granite gneiss of the Minbrie complex. Refuge Rocks.

The Middle Proterozoic sediments have been previously correlated with the Moonabie Formation (Nixon, 1975; Thomson et al., 1976) but although typical Moonabie Formation grits outcrop nearby to the southwest and south (e.g. at Mount Young), only the lowermost unit in the quarry may be equivalent. That unit consists of white to light grey, massive, coarse-grained arkose and feldspathic granule conglomerate with subangular to rounded quartz grains and scattered lithic fragments. Feldspar in the groundmass is commonly kaolinised, as are some clasts which are thought to be of volcanic origin. There are local pebble beds and some cross bedding within the sequence indicating that the bedding is right-way-up.

The majority of sediments in the quarry above the lowermost unit are more likely equivalent to the Corunna Conglomerate sequence. They consist of interbedded massive to flaggy arkose and quartzite. The arkose is less feldspathic than the lower unit, quartz grains are rounded, and lithic fragments are generally absent. Dark-coloured siltstones characterise this sequence and they show fine cross-bedding, heavy mineral lamination, and local ripple marks. Similar lithologies are present in the Corunna Conglomerate at Moonabie Range.

Folding of the Middle Proterozoic sequence(s) is clearly evident in the centre of the quarry's upper bench. It occurs about NW-SE trending axes correlated with the Wartakan tectonic event. It clearly predates deposition of the Pandurra Formation.

The Pandurra Formation capping Mount Laura consists of a series of interbedded conglomerates and sandstones at the base grading upwards and to the northeast into a thick sequence of sandstone with minor red siltstone interbeds. The basal conglomerate is of variable thickness reflecting irregularities in the unconformity surface, is purplish in colour grading up into a pink-brown colour, and contains a variety of generally rounded pebbles and cobbles. Most of the clasts are of quartzite similar to the underlying sequence(s). The upper conglomerates and sandstones are massive, cross-bedded and pink and brown in colour. They contain local grit bands and maroon siltstone interbeds. Cross-bedding and heavy-mineral lamination are present and in a borrow pit 3 km southeast large foresets up to 25-35 m across are present. Local, thin, barytes veinlets are developed in the Pandurra Formation.

Head east along Iron Knob Road back to Norrie Ave.

251.8 T-junction, turn right along Norrie Ave.

252.3 Turn left along Lacey St.

252.7 Turn right into water reserve.

STOP 34 - WATER TANK HILL (Fig. 33).

At this locality dark grey, dense, blocky, feldspathic sandstones of the Moonabie Formation are overlain unconformably by sandstones and conglomerates of the Pandurra Formation. The Moonabie Formation sandstones contain occasional pebbles of quartzite, acid volcanics, gneiss and minor jaspilite, with bedding trending approximately 315/80° SW.

The basal conglomerate of the Pandurra Formation, containing abundant large rounded pebbles of Moonabie Formation, intertongues to the northeast with well bedded red and white quartzite sandstone. The sandstone bedding trends approximately 320/12° NE. The Pandurra Formation was deposited against and onlapping onto an irregular cliff embayment, originally exposed in a

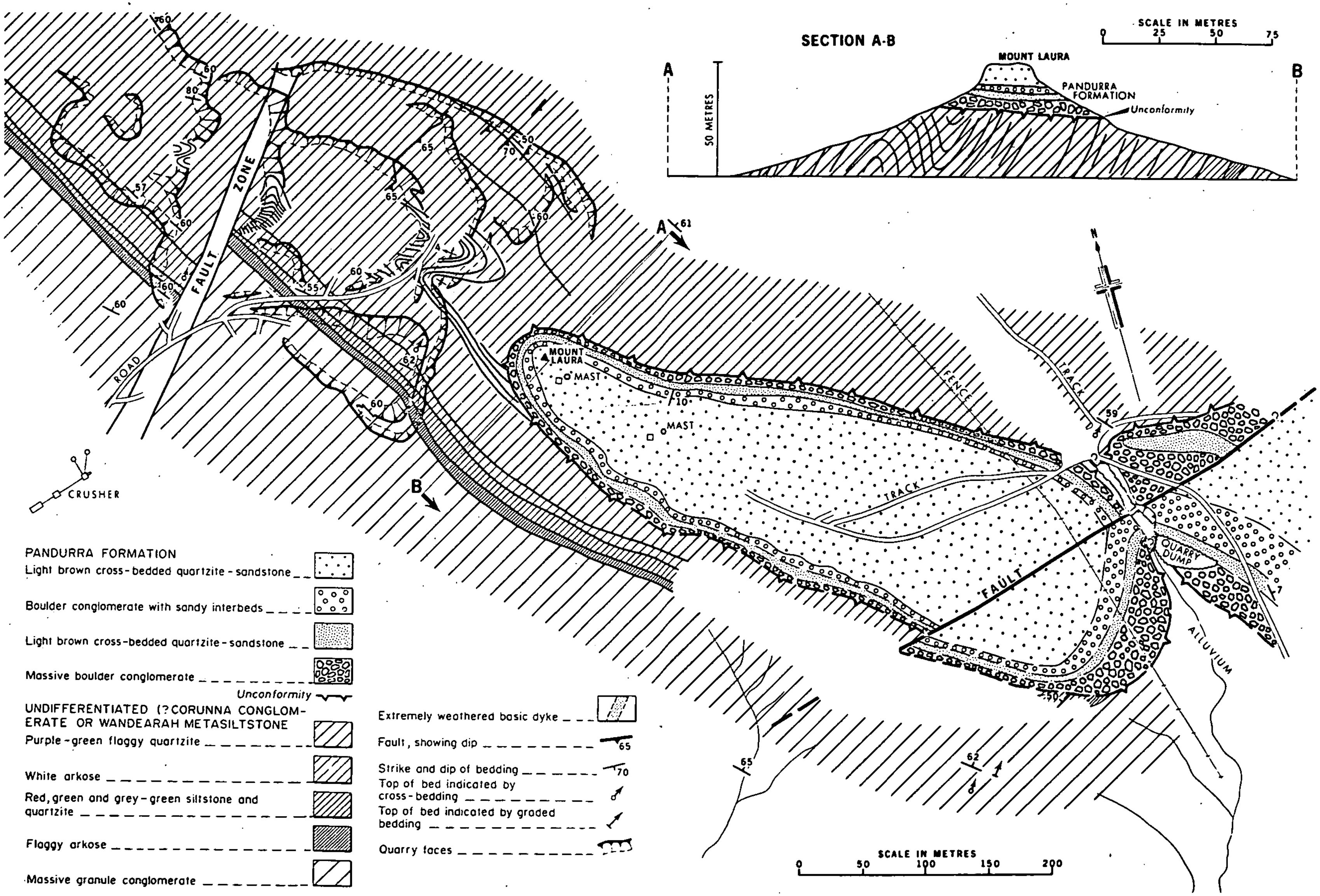


Fig. 32 Geology of the Mount Laura area, after Drexel and Nixon in Thomson *et al.* (1976).

building stone quarry southeast of the watertank. This was originally interpreted as a fossil wave-cut bench and cliff (Miles, 1954). However it is now suggested that the Pandurra Formation was deposited as terrestrial sediments.

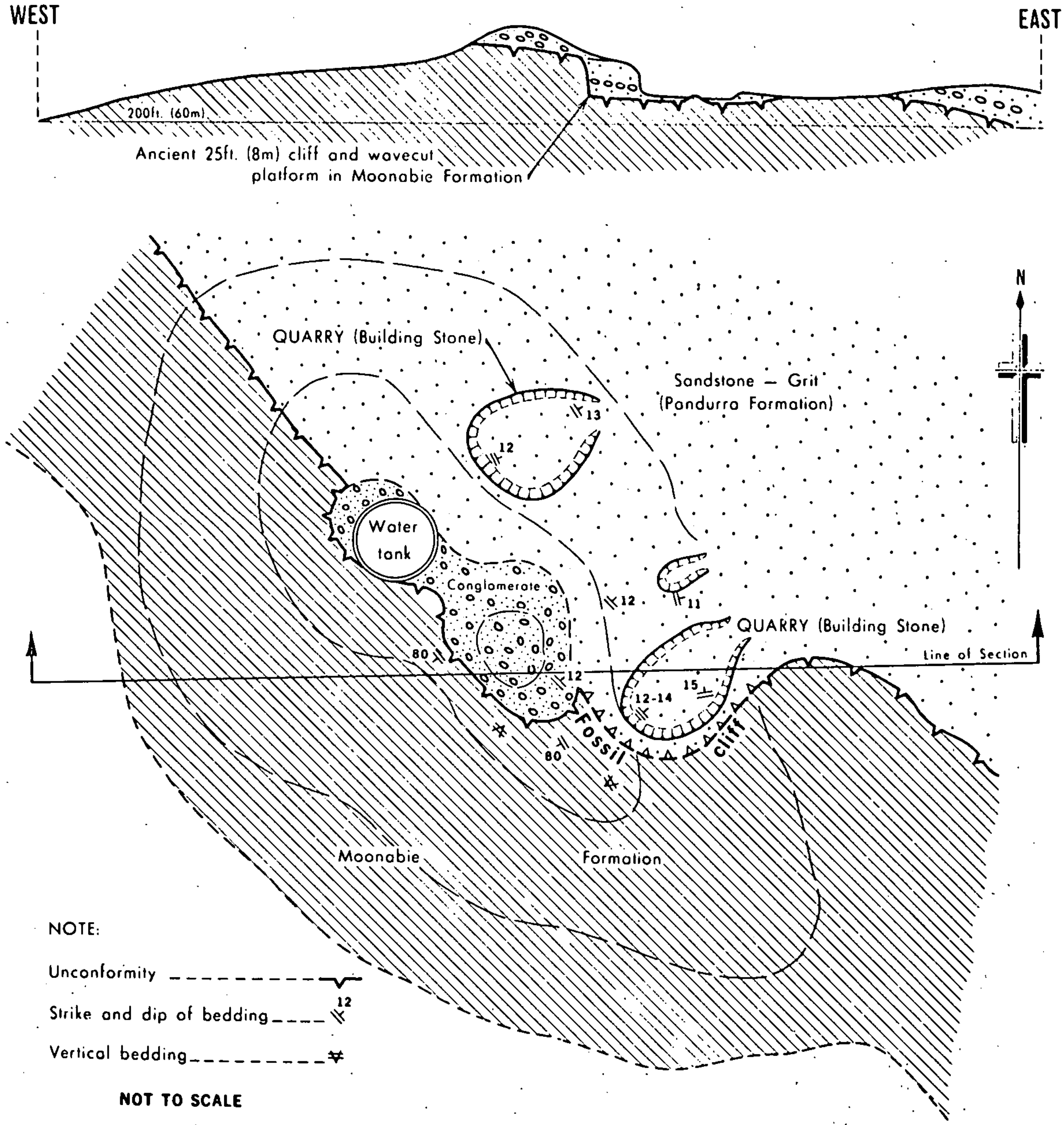


Fig. 33 Plan and section for Water Tank Hill.

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