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FIELD GUIDE TO THE ADELAIDE GEOSYNCLINE AND AMADEUS BASIN, AUSTRALIA

by

Richard J F Jenkins, John F Lindsay and Malcolm R Walter

AGSO
AUSTRALIAN GEOLOGICAL SURVEY ORGANISATION

AGSO RECORD 1993/35

Field Guide to the Adelaide Geosyncline and Amadeus Basin, Australia

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INTRODUCTION

(John F. Lindsay, AGSO, Canberra)

The Australian continent is divided in two by a feature generally referred to as the Tasman Line (Fig. 1). To the east of the Tasman Line rocks are generally Palaeozoic and younger in age and are frequently highly deformed. From the Cambrian through to the Permian the eastern margin of the continent was an active margin which has been referred to as the Tasman Geosyncline. To the west of the Tasman Line the rocks are generally much older. This older craton formed the site for the accumulation of considerable thicknesses of Neoproterozoic and early Palaeozoic sedimentary rocks.

A number of broad, shallow, intracratonic depressions formed on the Australian craton during the late Proterozoic and early Palaeozoic. The development of these basins relates perhaps at first to the assembly and later almost certainly to the breakup of the Proterozoic supercontinent, - basin dynamics thus appears to be largely tied to these global tectonic event (Lindsay and others, 1987). It is likely that an older part of the continent east of the Tasman Line was removed during the breakup leaving the new eastern margin of the continent free to act as an active margin and to accumulate Palaeozoic sediments.

In terms of their general morphology and tectonics the Australian intracratonic basins are all similar. Most basins have one abruptly truncated thrusted margin. Deep sub-basins occur parallel to the thrusted margins of most basins and all shallow to broad gently folded platformal settings along the opposite margin (Lindsay and Korsch, 1989; Lindsay and others, 1987). The Adelaide Geosyncline is somewhat different for, although formed at the same time as the intracratonic basins and probably by the same major tectonic events, it developed along the continental margin.

Some of the basins, such as the Officer Basin, are completely obscured at the surface by a thin veneer of Pleistocene desert dunes. However, the terminal Proterozoic rocks of the Adelaide Geosyncline are well exposed in the northern Flinders Ranges while rocks of similar age are well exposed along the northern margin of the Amadeus Basin in the Macdonnell Ranges, especially in the vicinity of Alice Springs.

Over the next 10 days we will visit sections in both of the these well exposed key areas.

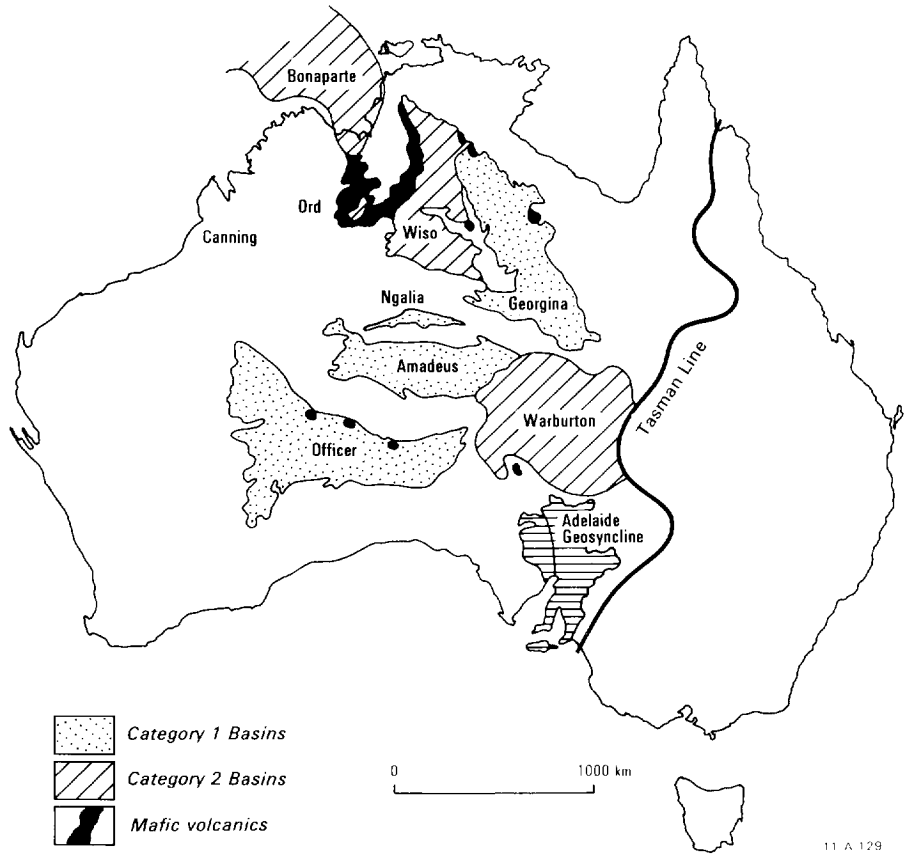


Figure 1. Location of the Adelaide Geosyncline, Amadeus Basin and related intracratonic basins. Stippled basins were initiated about 800 Ma; the cross hatched basins were initiated about 600 Ma (after Lindsay and others, 1987).

ITINERARY

Day 1, May 11: Arrive in Adelaide, South Australia. Afternoon - Introductory remarks and examination of specimens.

Day 2, May 12: Drive to Parachilna in the Flinders Ranges, perhaps via the Clare Valley, the source of some excellent wines.

Day 3, May 13: Overview of the Bunyeroo Gorge section, from the Etina Formation up through the Elatina Formation (Marinoan glacials) up to the Early Cambrian carbonates; succession, ejecta bed, fossils, chemostratigraphy, etc - type section of the Ediacaran and the Ediacarian.

Day 4, May 14: Uratanna Formation (basal Cambrian) at Angepena, putative glacials in the Billy Springs beds (Proterozoic Wonoka Formation) in the Umberatana syncline. Overnight at Arkaroola Tourist Village.

Day 5, May 15: Canyons in the Wonoka Formation, Umberatana syncline. Overnight at Parachilna.

Day 6, May 16: Ediacara assemblage and Parachilna Formation. Marinoan glacial succession near Blinman.

Day 7, May 17: Brachina Gorge to view facies different from those in Bunyeroo Gorge. Contingencies, requests. Lunch at Wilpena. Drive to Port Augusta; board the train for Alice Springs at about 4 pm. Party and discussion, 7 pm +. Overnight on train, sleeper accommodation.

Day 8, May 18: Check in to motel in Alice Springs. Briefing at the Northern Territory Geological Survey. Examination of Wallara 1, well that intersected most of the Proterozoic units in the central part of the Amadeus Basin, including both glacials.

Day 9, May 19: "Battery Flat" - Proterozoic regolith on the Bitter Springs Formation, Pioneer Sandstone (=upper glacials), stromatolite marker at base of Pertatataka Formation. "Hidden Valley"-Pertatataka Formation, ??seismite, unnamed older succession.

Day 10, May 20: Top of Bitter Springs, up through the Pioneer Sandstone, Pertatataka and Julie Formations and Arumbera Sandstone. Olympic Formation (upper glacials or mass flow?) type section and associated succession. Overnight at Ross River Homestead.

Day 11, May 21: Drive to Alice Springs. Ad hoc stops in the Bitter Springs Formation and the Arumbera Sandstone if there is interest.

ICS TERMINAL PROTEROZOIC AND IGCP PROJECT 320 FIELD PROGRAM

11 - 21 MAY, 1993

PART 1

ADELAIDE GEOSYNCLINE

Stratigraphic and tectonic overview of the Adelaide Geosyncline, South Australia

(W.V.Preiss, South Australian Department of Mines and Energy)

The Adelaide Geosyncline contains an extremely thick, folded late Proterozoic to Middle Cambrian sequence deposited initially in rifted troughs and later in broad zones of regional subsidence (Figs. 2 and 3). These are partly analogous to the Mesozoic-Cainozoic system of basins associated with the southern continental margin of Australia, but relationship to an ocean is speculative since no oceanic-crust remnants have been discovered.

Age constraints on deposition are based on dating of rare igneous rocks near the base of the Adelaidean and on a few Rb-Sr whole-rock shale ages. Contradictions between some of the data have yet to be resolved. The Callanna Group commences with platformal deposition of clastics and carbonates followed by regional extensional mafic volcanism (Arkaroola Subgroup); thereafter evaporitic clastics and carbonates of the Curdimurka Subgroup were deposited in rift-valleys with limited marine access. These commonly occur in a disrupted state in the Flinders Ranges diapirs. The Burra Group commences with largely fluvial sandstone and conglomerate, followed by eastward-prograding deltaic cycles alternating with marginal-marine and lagoonal carbonates, including sedimentary magnesite, and with fine-grained basinal sediments.

A regional unconformity separates the older rocks from glaciogenic sediments at the base of the Umberatana Group. Shelf deposits of glaciomarine tillite pass laterally into basinal clastics. Post-glacial marine transgression inundated the geosyncline as well as much of the Stuart Shelf, with uniform deposition of basinal carbonaceous silt. Regression led to marginal carbonate deposition, then renewed clastic influx from the west. During Marinoan-glaciation, the main source of ice was to the north, while conditions on the Stuart Shelf were periglacial.

The Wilpena Group records two major marine transgressive-regressive cycles, during the second of which submarine canyons were cut and, later, the metazoa of the Ediacara assemblage became prolific.

After withdrawal of the sea, renewed marine transgression in the early Cambrian led to wide-spread platform carbonate deposition (Hawker and Normanville Groups). Redbed sedimentation followed in the north, while in the south, the Kanmantoo Trough developed and was rapidly filled with a very thick clastic sequence.

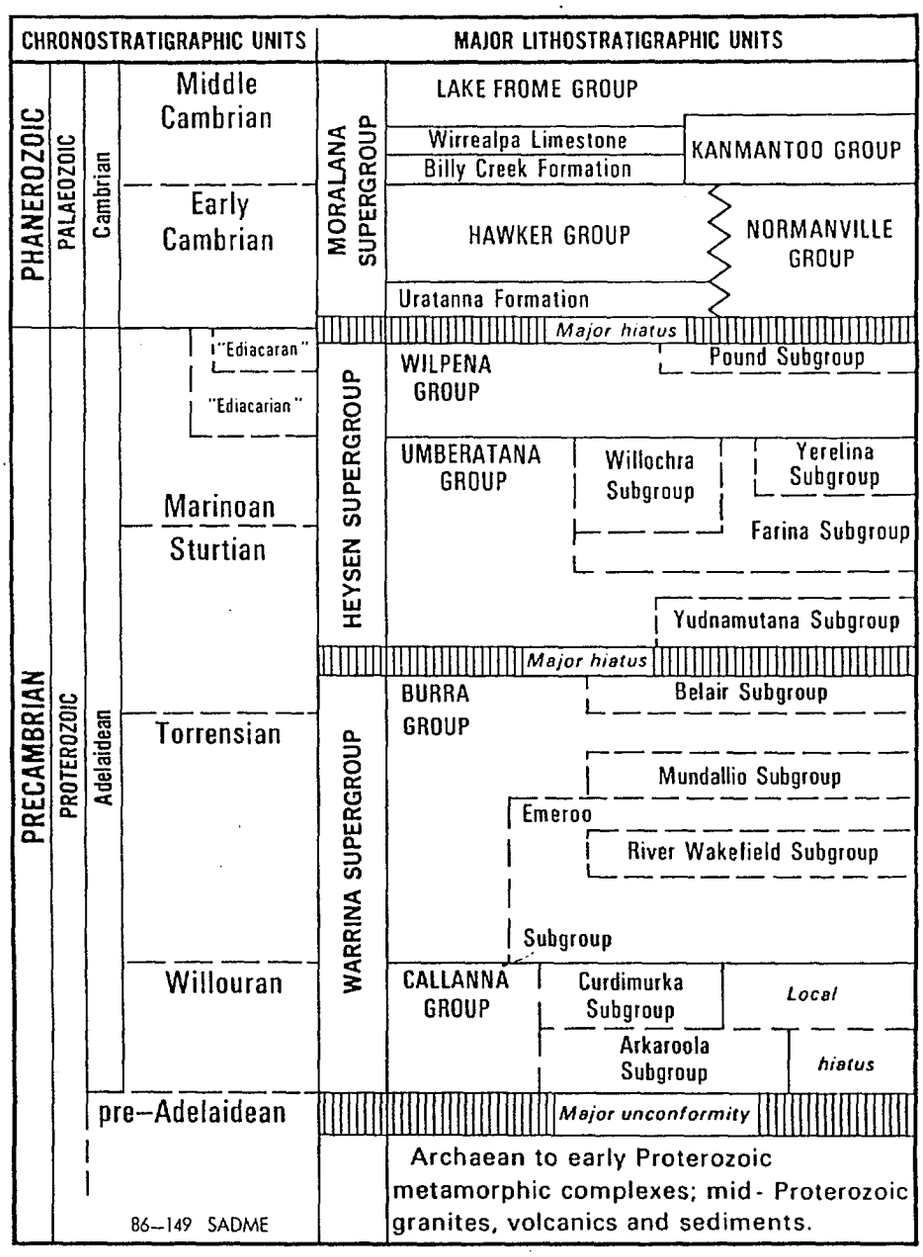


Figure 2. Chronostratigraphic and major lithostratigraphic units in the Adelaide Geosyncline (after Preiss, 1987).

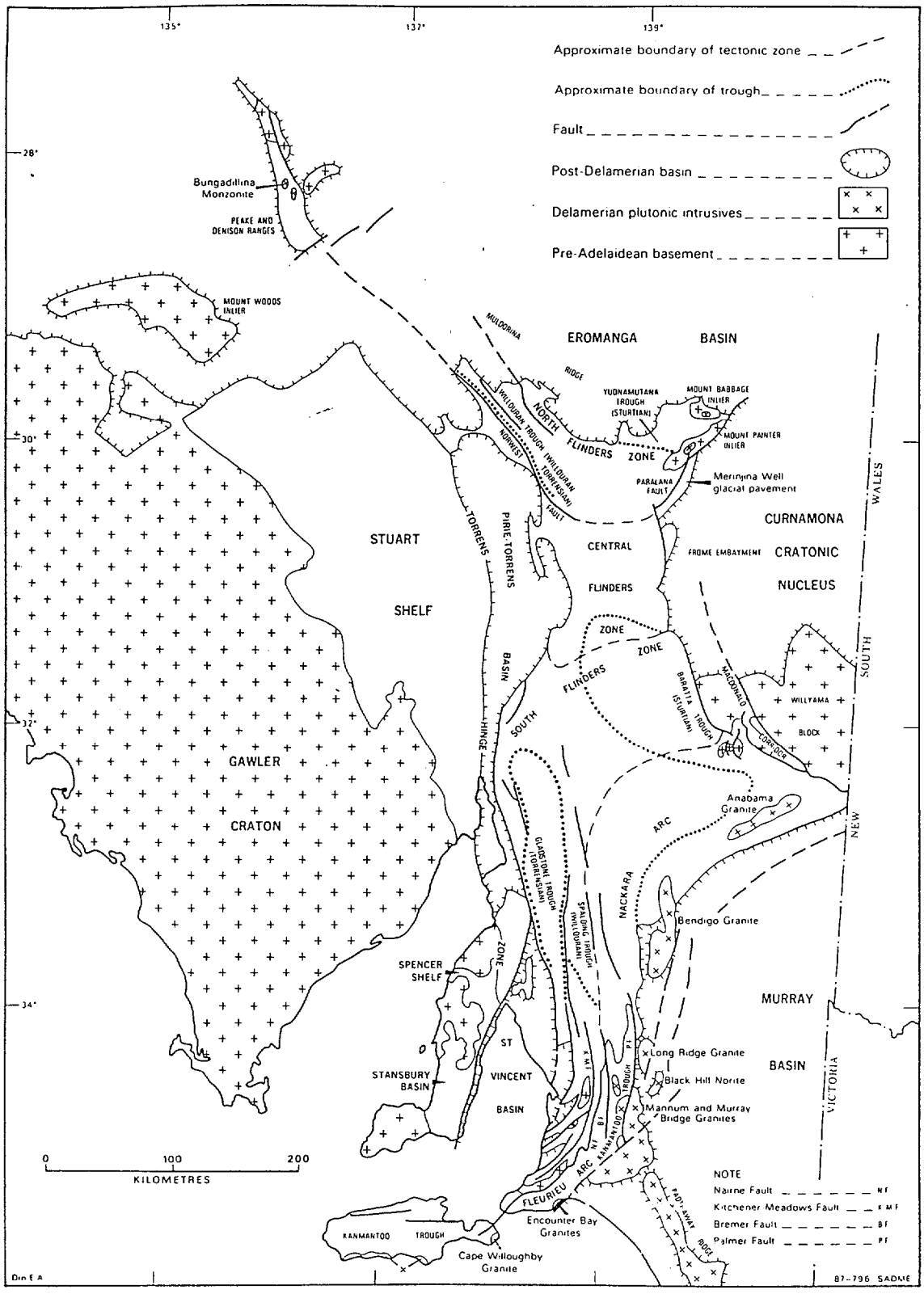
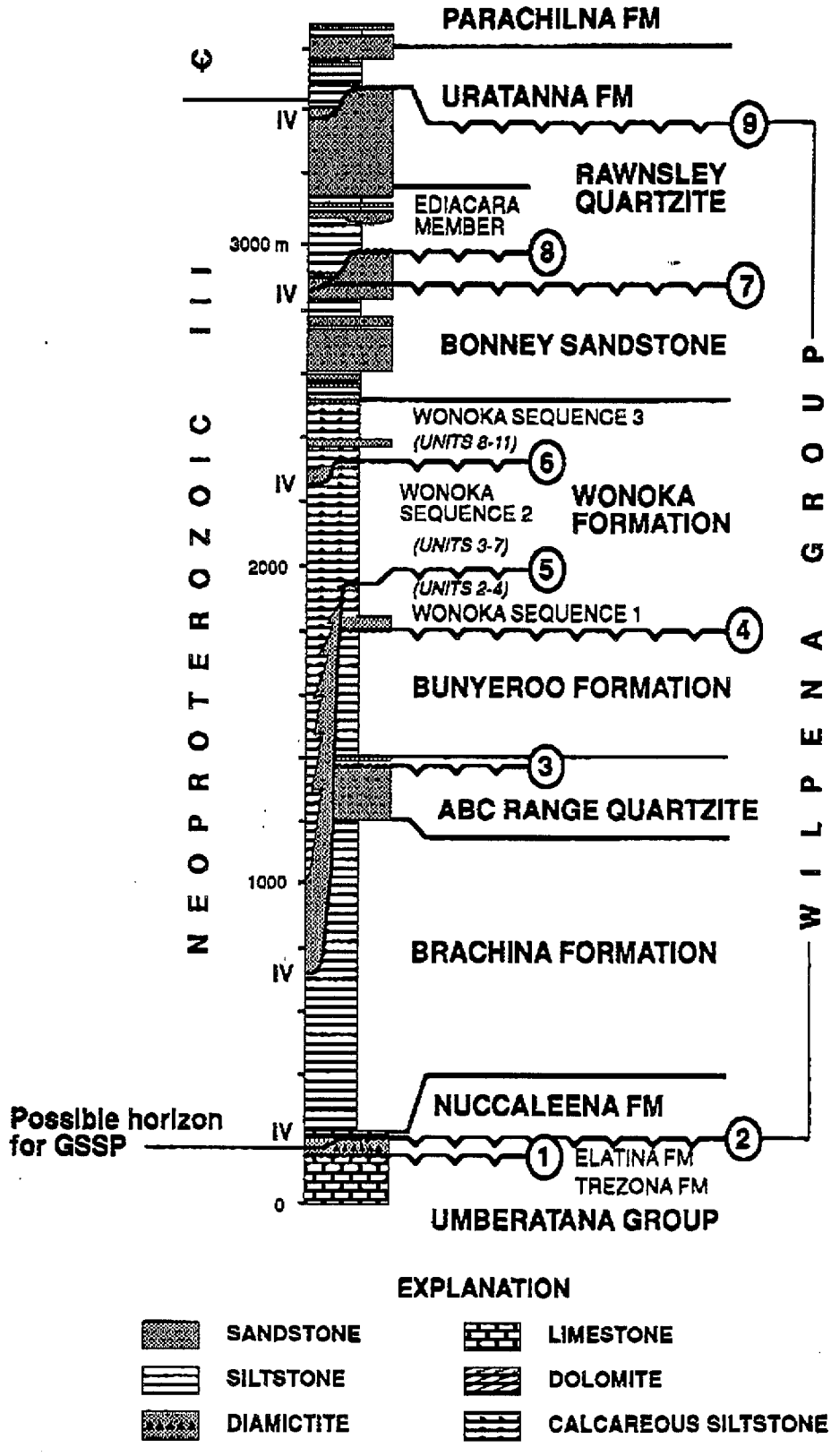


Figure 3. Tectonic elements of the Adelaide Geosyncline and environs.

Figure 4. Generalised stratigraphic section for the Wilpena Group of the central and northern Flinders Ranges, showing the interpreted location of the main sequence boundaries (modified from Christie-Blick and others, 1990). The boundaries are numbered informally, in this figure beginning with an unconformity at the base of the glacial Elatina Formation of the Umberatana Group. Incised valleys are well developed at five horizons (IV). The most prominent, to be examined on day 5 of the excursion, are at sequence boundary 5. This boundary is located within unit 3 (of Haines, 1987) of the Wonoka Formation in the northern Flinders Ranges, but is thought to correspond with a facies discontinuity at the base of unit 5 in the Bunyeroo Gorge area of the central Flinders Ranges (day 1). A candidate horizon for defining a GSSP for the base of a terminal Proterozoic system is sequence boundary 2, the age of the boundary specified as corresponding with the correlative conformity of this surface. In many parts of the Flinders Ranges, including Bunyeroo Gorge, the surface appears to merge with a prominent flooding surface at the base of the Nuccaleena Formation. Locally, incised valleys cut down as much as 150 m+ into the underlying glacial succession, and an intermediate lithic unit (Seacliff Sandstone) is present (Dyson, 1992; I.A. Dyson, pers. comm., 1993).


PARACHILNA FM
Є
URATANNA FM
IV
9
RAWNSLEY QUARTZITE
EDIACARA MEMBER
3000 m
8
IV
7
BONNEY SANDSTONE
WONOKA SEQUENCE 3
(UNITS 8-11)
6
IV
WONOKA SEQUENCE 2
WONOKA FORMATION
(UNITS 3-7)
2000
5
(UNITS 2-4)
WONOKA SEQUENCE 1
4
BUNYEROO FORMATION
3
ABC RANGE QUARTZITE
1000
BRACHINA FORMATION
IV
NEOPROTEROZOIC III
WILPENA GROUP
NUCCALEENA FM
Possible horizon for GSSP
IV
1
ELATINA FM
TREZONA FM
2
0
UMBERATANA GROUP
EXPLANATION
SANDSTONE
SILTSTONE
DIAMICTITE
LIMESTONE
DOLOMITE
CALCAREOUS SILTSTONE

The Delamerian Orogeny commenced towards the end of the Cambrian with folding and the intrusion of granites in the south. Syntectonic and post-tectonic intrusives follow the Nackara and Fleurieu Arcs and a north-northwest lineament that was probably an active basement shear during folding. Deformation was most intense in the south, with high-grade metamorphism also being controlled by this shear.

Overview of Neoproterozoic Sequence Stratigraphy in the Adelaide Geosyncline

(N. Christie-Blick, Lamont-Doherty Geological Observatory, New York)

Beginning in the mid-1980s (Christie-Blick and von der Borch, 1985; von der Borch and others, 1986), attempts have been made to interpret the sedimentary and tectonic evolution of both the Adelaide Geosyncline and Amadeus basin in terms of "sequence stratigraphy" (Lindsay, 1987; Christie-Blick and others, 1988, 1990, and in review; von der Borch and others, 1988; Lindsay and Korsch, 1989, 1991). Sequence stratigraphy is a technique for studying repetitively arranged sedimentary facies with reference to their stratal geometry, particularly the location of regional unconformities termed sequence boundaries (Van Wagoner and others, 1988). Sequence boundaries are present at a great range of scales, and they include surfaces recognised as unconformities in classical stratigraphy, as well as a good many more subtle features not normally so classified but which are nevertheless objectively recognisable. Subtle sequence boundaries are commonly present *within* conventional lithostratigraphic units, which form the basis for virtually all existing interpretations of Neoproterozoic sedimentary rocks but which are fundamentally diachronous entities. The value of sequence stratigraphy in Neoproterozoic geology is 1) to draw attention to the location of stratigraphic discontinuities (and hence hiatuses) in what are commonly regarded as continuous successions; 2) to provide an independent way of establishing relative time-stratigraphy, at least at the scale of a single sedimentary basin; and to the extent that some of the observed discontinuities are due to glacial-eustacy, 3) to provide additional clues about interregional and global correlation of Neoproterozoic "events". Given the focus of this excursion and the fact that sequence stratigraphy is not yet widely applied in Neoproterozoic successions, and because no consensus exists about interpretive details in the basins to be examined, we plan to address both general and specific aspects of sequence stratigraphy in each of the key sections. We emphasise also that there is nothing inherently wrong with the presence of sequence boundaries, even in a section that is a

candidate for a global stratotype. Such surfaces are present in all successions. More relevant is the overall richness and completeness of the preserved stratigraphic record. The large number of discrete sequence boundaries identified in the Neoproterozoic of South Australia is, perhaps ironically, an indication of rich preservation of geological events.

In the terminal Proterozoic succession of the Adelaide Geosyncline, sequence boundaries have been identified at 9 horizons (Fig. 4). All of these are associated at least locally with incised valleys and/or regional facies discontinuities involving abrupt upward *shoaling* of facies, observations that are taken to imply the lowering of depositional base level. That is, these features cannot be due simply to the continuous shifting of the shoreline in response to changes in the local supply of sediment, but require systematic changes in the rate of tectonic subsidence and/or eustatic fluctuation. The valleys range from tens to hundreds of metres deep, and in the case of a horizon at or near the base of the Wonoka Formation, more than 1 km deep (boundary 5 in Fig. 4). Most of the boundaries pass laterally into surfaces lacking erosional relief, and in these places overlying and underlying strata are generally concordant. This inevitably leads to disagreement between those who interpret concordance as an indicator of conformity, and those who infer the presence of cryptic unconformities on the basis of erosional relief and facies discontinuities documented elsewhere. Critical to such arguments is the mechanism and environment of valley incision.

A related difficulty in intracontinental ramp settings, which characterise much of the depositional history of both the Adelaide geosyncline and Amadeus basin, is that in many cases sequence boundaries pass laterally into or amalgamate with marine flooding surfaces associated with abrupt upward *deepening* of facies. On the one hand it is therefore possible not to recognise that a flooding surface is also a sequence boundary, and on the other, to misinterpret a flooding surface as a sequence boundary when no such boundary exists or, if it does, is located at a different, typically lower horizon.

Confident sequence stratigraphic interpretation in ramp settings generally requires the documentation of incised valleys. In this context, the sequence boundary most difficult to locate objectively in the Adelaide geosyncline is the one inferred in the upper part of the ABC Range Quartzite (boundary 3 in Fig. 4). Erosion surfaces described at this level in previous studies have for the most part proven to be related to valleys in the stratigraphically higher Wonoka Formation (boundary 5 in Fig. 4) and/or to be of structural rather than stratigraphic origin.

Concepts of Ediacaran and Ediacarian Systems

(R.J.F. Jenkins, University of Adelaide, South Australia)

Jenkins (1981) formally proposed an Ediacaran System and Period comprising the terminal interval of the Precambrian and based on a stratotype at Bunyeroo Gorge in the Flinders Ranges, South Australia. The Ediacaran system was defined as embracing the Wonoka Formation, Bonney Sandstone and Rawnsley Quartzite. The wider designation of an Ediacarian System made by Cloud and Glaessner (1982) was extended downwards to include the Nuccaleena Formation (Figs. 5 and 6).

These proposals derive from the original idea of an 'Ediacarien' (French spelling) Stage advanced by Termier and Termier (1960). In an extended discussion, Cloud (1968) greatly advanced the concept of an Ediacaran/'(or Ediacarian)' System and Period, particularly stressing the distinctive life forms ascribed to this time interval. Harland (1974) outlined the history of ideas surrounding the Precambrian-Cambrian transition and reviewed characters or events potentially useful for establishing internationally recognised boundaries between division of the late Precambrian and Cambrian; the occurrence of the 'Ediacaran faunas' comprised one event in the sequence discussed. Harland and Herod (1975, p. 205) tabled a suggested 'International Standard Stratigraphic Scale' for the late Precambrian, including an Ediacaran Period. The literature pertinent to the division of the latest Precambrian is extensive (Cloud and Glaessner, 1982; Harland and others 1982) and discussion continues.

Local acceptance of any new time-rock division within the late Precambrian of Australia is likely to be contingent on the priority of the Adelaide System (David, 1922; Mawson and Sprigg, 1950; Thomson and others, 1964), which is based on composite stratotypes in the hills immediately east and south of Adelaide. Recognition of an 'Adelaidean' time interval (Thomson, 1969; Preiss, 1987) is well established within the wider framework of Australian geology.

The formal erection of an Ediacaran System in the Flinders Ranges (Jenkins, 1981) implies that the upper limit of the Adelaidean System and Marinoan (Mawson and Sprigg, 1950) should be restricted to rock intervals represented in their type area; the uppermost Precambrian rocks known in the Adelaide Hills are equivalents of the ABC Range Quartzite and conformably overlying purple siltstone correlative with the Bunyeroo Formation (Jenkins and Gostin, 1983; Dyson, 1992).

During the late 1950s and 1960s the type Marinoan came to be correlated with a thicker section in the Flinders Ranges because of confusion between the significance of the ABC Range Quartzite and

the then designated Pound Quartzite (now Pound Subgroup). This extended division in the Flinders Ranges identified as the Wilpena Group (Dalgarno and Johnson, in Thompson and others, 1964), overlooked a major change in basin geometry, with local erosion occurring to the south, and renewed subsidence in the north accommodating the carbonate-rich Wonoka Formation. The recognition of the Ediacarian (Cloud and Glaessner, 1982) preserves the Wilpena Group as a single entity.

Jenkins (1981) considered the Wonoka Formation as heralding a new major depositional entity because of observing a sharply angular relationship (local) between truncated Bunyeroo Formation and basal Wonoka Formation beds in the Pichi Richi area (southern Flinders Ranges). Moreover, large clasts of the Bunyeroo Formation incorporated in Wonoka Formation olisthostromes showed a spaced cleavage implying not only prior lithification, but some presumably compressive prior orogenic movement, locally expressed by more tightly appressed symmetrical folding of Adelaidean (sensu stricto) rocks.

Prior-lithified Adelaidean clasts are also known to be incorporated in Wonoka Formation 'canyon' fills in the northern Flinders Ranges (Eickhoff and others, 1988) and there have been informal suggestions of such 'canyons' being localised in ?syndepositional synclinal structures (von der Borch, pers. comm.).

There are several reports of supposed indications of metazoans within the Adelaidean *sensu stricto*, but none has so far been sustained by subsequent investigations (Teesdale-Smith, 1956; Branagan, 1979; Jenkins, Plummer and Moriarty, 1981; Jenkins, 1986).

In the central Flinders Ranges, where the tectonic setting was more basinal, deposition between the Bunyeroo and Wonoka Formations seems to have been more or less continuous; a useful field marker for the base of the Wonoka Formation is provided by a thin cupriferous dolomite bed and commonly associated brecciolas (combined thickness 5 cm to 10 m), either conformable or paraconformable above the Bunyeroo Formation, and seemingly continuous over an area in excess of 6000 km^2. This bed identified as the Wearing "Dolomite Member of the Bunyeroo Formation" (Preiss, 1987) apparently signals a ?brief episode of shallowing, and provides a regional 'spike' marking the boundary of the Adelaidean and Ediacaran. Succeeding lower parts of the Wonoka Formation include calcareous and clastic storm-beds and a distal basinal interval, and suggest rapid regional subsidence concomitant with the erosion of deep (1100 m) 'submarine canyons' on the flanks of the basin and basinal highs. Trace fossils, imperfectly preserved 'medusoids' and a possible frond-like fossils, occur in shelf facies

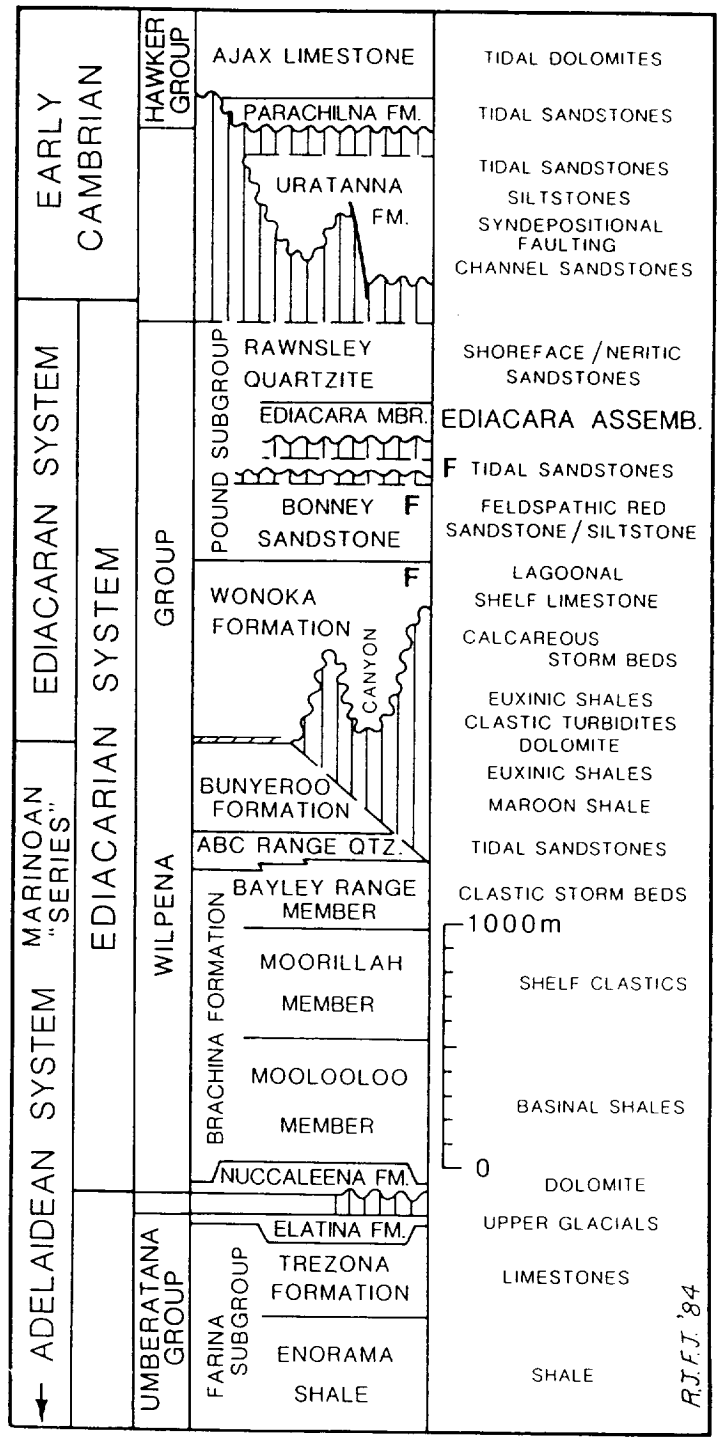


Figure 5. Terminal Precambrian and Early Cambrian stratigraphy of the western Flinders Ranges, South Australia. Old two part division model.

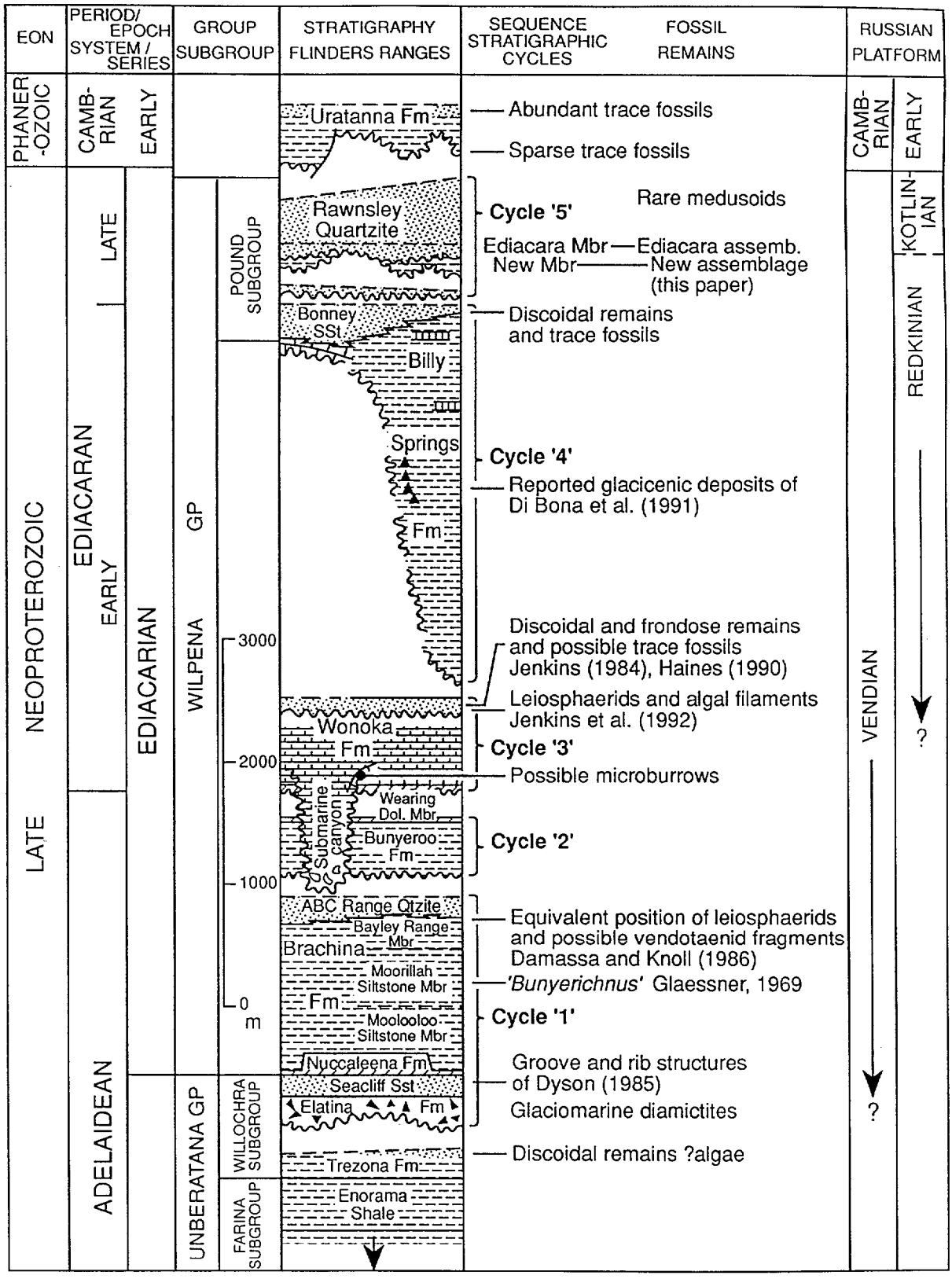


Figure 6. Late Neoproterozoic stratigraphy of the Flinders Ranges, South Australia with a generalized sequence stratigraphy.

near the top of the Wonoka Formation (Jenkins, 1984; Haines, 1987, 1990). The well known Ediacara assemblage occurs within the overlying Pound Subgroup (Jenkins and others, 1983).

The Flinders Ranges Succession

The Neoproterozoic succession of the Flinders Ranges of interest for the present study embraces the upper Umberatana Group and Wilpena Group (Preiss, 1987). In the central Ranges siltstone lithologies predominating in the mid-parts of the Umberatana Group are succeeded by the shallow-water Trezona Formation of carbonates, calcareous shale and arenites. A significant surface of erosion occurs at the base of the overlying glacigenic Elatina Formation (e.g. Lemon and Gostin, 1990). Sequence stratigraphic studies of the later Neoproterozoic succession inclusive of the Elatina Formation and its equivalents have supplanted the earlier notion that the section principally comprises two major depositional cycles (von der Borch and others, 1988; Christie-Blick and others, 1990; Jenkins and others, 1992; Fig. 3). At least five depositional 'sequence' cycles or more now seem evident.

The Elatina Formation, Seacliff Sandstone, Nuccaleena Formation, Brachina Formation and ABC Range Quartzite constitute one cycle, the greater part of which is of highstand aspect (Dyson, 1992). A second highstand cycle includes lithic grits overlying a surface of erosion high in the ABC Range Quartzite, together with the succeeding Bunyeroo Formation. Interbedded thin dolostones and shales of the older part of the Wearing Dolomite 'Member' are a condensed interval within the regressive phase.

Intraclastic dolostones forming upper parts of the Wearing Dolomite 'Member' herald a lowstand systems tract embracing much of the Wonoka Formation. I do not agree with Christie-Blick and others (1990) that a 'cryptic' or hidden systems tract boundary within the Wonoka Formation corresponds to the time of formation of major 'submarine' canyons, as if regional exposure had occurred, carbonate-rich shales deposited on basin slopes should have been susceptible to erosion, or formation of carbonate crusts. Instead, this seems to be the time of maximum flooding in interior parts of the basin. Haines (1987) documents a series of truncating surfaces associated with older ?canyon channels near Beltana and evokes submarine slumping to explain the successive cuts.

An ensuing highstand systems tract includes the c. 2-3 km thick Billy Springs Formation, of the northeastern Flinders Ranges, possibly the topmost beds of the Wonoka Formation further to the south, and the Bonney Sandstone (Fig. 6). The Rawnsley Quartzite

and its fossiliferous Ediacara Member record a terminal Proterozoic cycle.

N.M. Lemon (pers. comm.) of Adelaide University has located discoidal fossil remains showing an 'Ediacaran' style of preservation in arenites forming a member at the top of the Trezona Formation. The structures may be of algal origin but further investigation is clearly required.

The supposed sea-pens which Dyson (1985) described from the Seacliff Sandstone of the southern Flinders Ranges show numerous longitudinal grooves and regular cross-ribbing which varies in its spacing in a cyclical manner unlike biogenic remains. This rhythmical variation in the ribbing is similar on different parts of the structure and the marks may have been made by the bottoming of a bladed agglomeration of ice bobbing along in the waves (Jenkins, 1986). Another rhythmical structure, *Bunyerichnus dalgarnoi* Glaessner, 1969, from the Moorillah Member of the Brachina Formation, may similarly be attributed to a tethered structure skittering in an arcuate manner over the substrate (Jenkins, *in* Jenkins and Gravestock, 1988). Thalli of vendotaenids (c. 10 cm 'tall') are of about the right dimensions to have formed *Bunyerichnus* and fragmentary remains of such megascopic 'algae' are known from a marginally higher stratigraphic level on the Stuart Shelf (Damassa and Knoll, 1986). Rare vertical, cylindrical structures 4 - 15 mm in diameter collected by J.G. Gehling (University of South Australia, Adelaide) from the ABC Range Quartzite at Bunyeroo Gorge show internal disruption of bedding that probably resulted from fluid escape.

Upper parts of the walls of 'submarine' canyons associated with older parts of the Wonoka Formation are commonly 'covered' with a thin veneer of micritic carbonates and a variety of admixed syndepositional brecciolas (von der Borch and others, 1989). Von der Borch (pers. comm.) has found that these micrites include small, sub-cylindrical 'tubules' c. 0.2 - 0.4 mm in diameter filled with a clearer spar than the surrounding matrix. The random orientations of these 'tubules' and their variably indistinct margins are consistent with them being endogenic microburrows. Following ideas of Clark (1964) and Trueman (1975) concerning early metazoans requiring a hydrostatic skeleton in order to displace sediment during burrowing, the 'tubules' were possibly made by small coelomates or pseudocoelomates penetrating settled carbonate muds veneering the walls of the canyons. As the mode of formation of the canyons remains uncertain, the timing of deposition of the carbonate veneer cannot be precisely fixed, but it was certainly early during sedimentation of the Wonoka Formation (with reference to Haines, 1987, possibly corresponding to the time of Unit 3 or early Unit 4), and is thus early 'Ediacaran'.

Within Unit 8 of the Wonoka Formation a widespread narrow interval of carbonate-rich shales located by its proximity to a purple tuffaceous horizon includes curved, dark, banded structures that form widening swaths on bedding planes (Haines, 1990, Fig. 10a, b). Though Haines identified these structures as possibly being of algal origin, they conform closely to published illustrations of *Palaeopascichnus delicatus* Palij, 1976. Recently collected weathered material shows slight relief between the matrix and the 'fill' of the bands, which seems somewhat leached; the bands are sometimes seen to be connected by turn-arounds. These characteristics support my view that the structures are actually thigmotactic trace fossils. 'Branching' forms may reflect places where the organism accidentally encountered a previous burrow and thence formed a new trace away from it. Concomitant with this idea, Haines (1987) describes common dark, micritic, ovoid peloids averaging 0.17 mm in length from Units 9 and 11, and notes that they form the dominant component of some beds. The suggestion that the peloids represent faecal pellets accords with the presence of animals having a gut.

Carbonised microfossils including large leiosphaerids (125 - 175 mm diameter) and filamentous forms (10 - 50 mm wide and up to 1.4 mm long) have been observed in bedding plane thin sections from carbonaceous micrites of upper Unit 8 (e.g. Jenkins and others, 1992).

Unit 10, sandstones or quartzite of shallow-water origin, includes rare discoidal remains of the kind commonly associated with Ediacaran assemblages, and also a possible frondose structure (Jenkins, 1984). However, these structures are so imperfectly preserved as to limit their systematic evaluation. While the notion that they represent metazoan remains is preferred, the frondose form could potentially be an overfolded part of a crinkled algal film and the simple morphology of the discs is similar to that of the circular colonies of living diatomaceous algae illustrated by Glaessner (1969, Fig. 3). At a similar level, small 'dimple' marks resembling those figured by Narbonne and Hofmann (1987, Fig. 10i) closely crowed some bedding planes.

Haines (1990, Fig. 10c) reports a large, foliate fossil form shaped rather like a palm leaf from an equivalent stratigraphic level near Black Range Spring in the central Flinders Ranges; several of these ribbed structures are associated on one large bedding plane and it is intriguing that the material of which they were constructed seems to have been sufficiently brittle for fragments to break away, an indication that desiccated algal fronds may be represented.

Di Bona (1991) reported possible glacigenic rocks in a lower part of the Billy Springs Formation, though the common slumping of the described diamictites has led to questioning as to their mode of formation (M.R. Walter, pers. comm.). My own observations of highly angular to well rounded polymictic clasts including rare basement (granitic) pebbles, and the occurrence of lonestones or dropstones supported in a laminated matrix, accord with possible ice rafting from a local glacial source and support the notion of a climatic cooling. Disjunct exposures of the Billy Springs Formation make it impossible to correlate closely with sections further south (Reid, 1992), but the study of Di Bona and others (1990) suggests it postdates slope deposits within the Wonoka Formation represented by their units E and F, which are apparently equivalents of Unit 10 of Haines (1987, 1990).

Levels in the paralic Bonney Sandstone include moulds of algal films and ring-shaped structures. Short sections of *Palaeohelminthoida* Ruchholz, 1967, sp. occur low in the formation and *Helminthoidichnites tenuis* Fitch, 1850, and possible burrows cf. *Lockeia* U.P. James, 1897, sp. have been located high in the unit. Small discoidal remains were collected by M. Wade (Queensland Museum, Brisbane) from the Bonney Sandstone.

The complex stratigraphy of channel sands, possible tidal deposits and beach or barrier sands constituting older parts of the Rawnsley Quartzite has yet to be adequately elucidated. Several intervals of siltstones and thin to medium bedded, fossiliferous sandstones or quartzite and other, generally more massive, unfossiliferous sandstones were described by Jenkins and others (1983) as the Ediacara Member, a designation extended downwards by Gehling (1987, 1988). On a western pastoral property channel deposits filling erosion valleys cut down into older light-coloured sandstones contain an unusual fauna atypical of the type Ediacara assemblage present in the overlying Ediacara Member. These channel deposits are being described elsewhere as a new member, which will be accorded informal status only here. The upward-shallowing sedimentary cycle of the new member includes dark-red-brown siltstones, storm-deposited sand beds and manganese-stained, thin quartzite beds which locally show abundant mud cracks. An estuarine environment seems to be indicated.

The new fossiliferous member includes rare specimens resembling *Rangea* Gürich, 1929, common discoidal remains, rare larger discs with regularly spaced annulation and referable to *Kullingia* Glaessner, 1979, discoidal forms which bore stems, and abundant tentacular remains of *Hiemalora* Fedonkin, 1982, or closely related taxa. *Dickinsonia costata* Sprigg, 1947, and *Tribrachidium heraldicum* Glaessner, 1959, occur rarely. A rather sparse trace fossil assemblage is present, with *Helminthoidichnites tenuis,*

Planolites montanus Richter, 1937, *Helminthopsis* Heer, 1877, sp., *Palaeopascichnus delicatus*, and epichnial markings comprising radial arrays of branching grooves cf. *Chondrites* von Sternberg, 1883, sp. A form of *Helminthoida* Schafhüntl, 1851, is a rare element, but is better represented by more complete specimens found at about the same stratigraphic level elsewhere in the high western Flinders Ranges (Runnegar, 1992, Fig. 3.9). Specimens of rangeids have also been located at similar levels elsewhere in the Ranges (Gehling, 1991, pl. 3, Fig. 1). At Ediacara Range an unusual massive bed of quartzite a narrow distance stratigraphically below the Ediacara Member (see Wade, 1972b) includes *Pteridinium carolinaense* (St Jean, 1973).

The Ediacara Member, which also constitutes one or several upward shallowing and coarsening sedimentary cycles, includes the well described body fossil and rather less well known trace fossil elements of the type Ediacara assemblage. Approximately 19 distinctive kinds of trace fossils occur in this assemblage, and while these taxa tend to have a scattered distribution on rock surfaces, some lower surfaces of sandstone flags may be closely covered with ichnological remains. In particular, concave moulds of 'Form F' of Glaessner, 1969) (= c.f. *Planolites* Nicholson, 1873) may be especially dense and wetting of the surface may disclose the unweathered fill of cris-crossing specimens. Interestingly, these traces do not penetrate discoidal remains preserved on the same slabs, possibly because the discs were originally buried and the inchnofossils represent lengths of faecal extrusion on the substrate. Very rare *Psammichnites* Torell, 1870, and relatively common 'escape' burrows (resembling *Bergaueria* of Fedonkin, 1985, pl. 22, Fig. 2) indicate animals whose ?contracted bodies were of the order of 1 cm in diameter. Sets of scratching marks made by an animal that had approximately 13 pairs of double-clawed limbs are relatively common (Gehling, 1991, pl. 6, Fig. 3). Subvertical or vertical dwelling burrows are either rare or absent, but a short worm-like animal with toughened elements at its ?head end lived in shallowly oblique burrows in which its body imprint may sometimes be preserved. The backfilled, septate *Taenidium* cf. *serpentium* Heer, 1877, which loops shallowly above and ?below a given bedding plane, almost certainly indicates a coelomate as do the penetrative hypichnial burrows of a large *Chondrites* sp.

Higher in the Rawnsley Quartzite rare discoidal fossil remains are known. They may be loosely assigned to *Ediacaria* or *Cyclomedusa*.

The base of the Uratanna Formation is a widespread erosional surface with channel sands rich in mud clasts commonly developed above (Daily, 1973); other local thick (c. 500 m) expressions of the unit are controlled by syndepositional faults (e.g. Mount, 1989, 1991). Higher parts of the thicker expressions of the Uratanna

Formation show upward shallowing and coarsening with trace fossils of Cambrian aspect such as *Skolithos* Haldemann, 1840, *Rusophycus* Hall, 1852, *Phycodes pedum* Seilacher, 1955, and *Didymaulichnus miettensis* Young, 1972 becoming increasingly abundant (e.g. Jenkins, 1992, Fig. 1). The appearance of these trace fossils evidently shows a relationship to the depth of water in which the sediment accumulated. This poses an important question as to how supposed first appearances of trace fossils due to evolutionary changes may be distinguished from climax colonisation related to eustatic phenomena. Thus the upper part of the Uratanna Formation is validly Cambrian on biostratigraphic grounds, but what age can be assigned to the older part of the unit based on the *non occurrence* of diagnostic traces? Shales of the Uratanna Formation are similar in lithology to those of older parts of the Ediacara Member, but certainly have a quite different (vertical) stratigraphic placement.

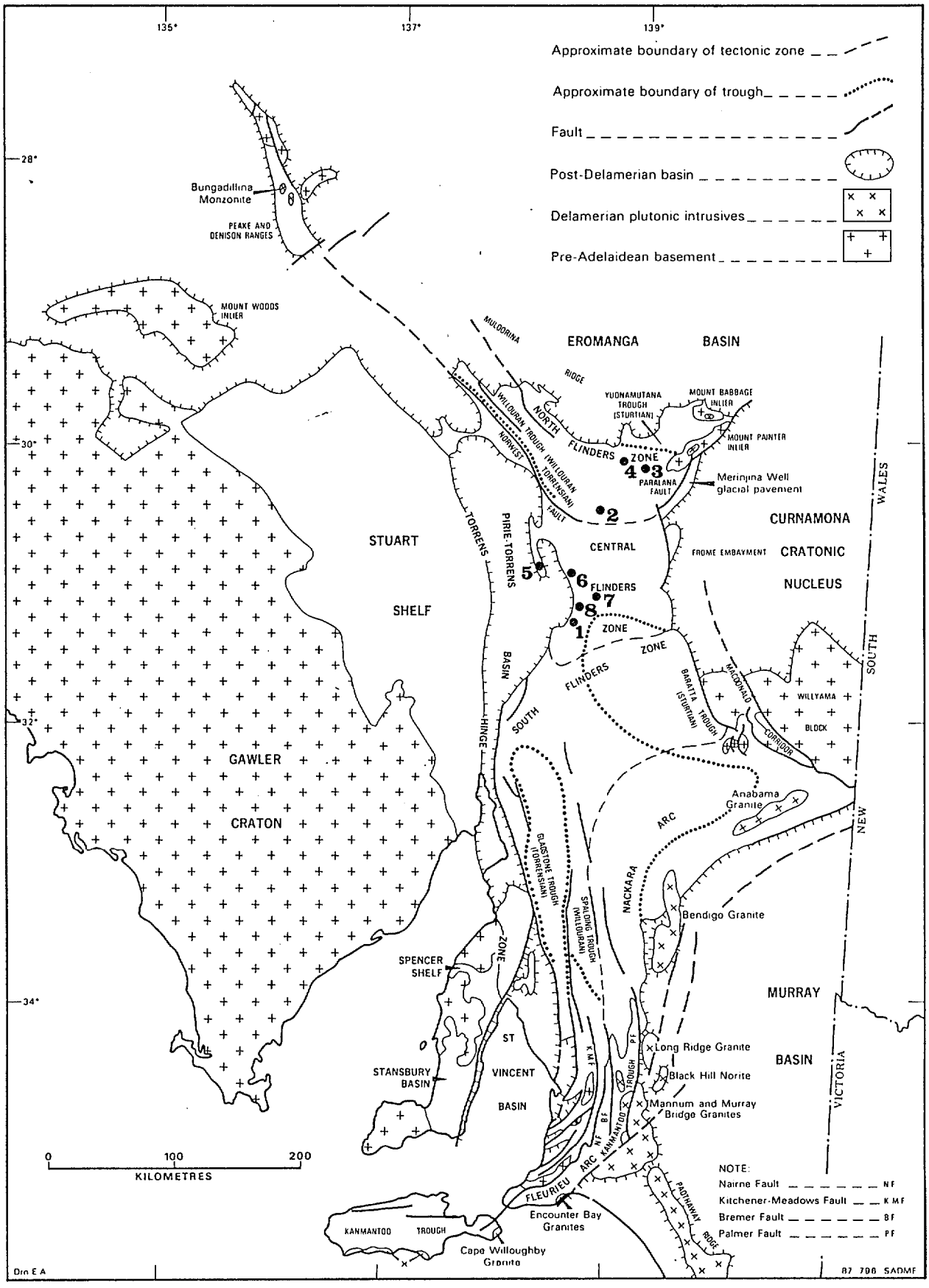


Figure 7. The Adelaide Geosyncline and the location of sites to be visited. 1. Day 3, Bunyeroo Gorge, 2. Day 4, Angepeena, 3. Day 4, Umberatana, 4. Day 5, Umberatana, 5. Day 6, Ediacara fossils, 6. Day 6, Parachilna Gorge, 7. Day 6, Enorma Ck - Bulls Gap, 8. Day 7, Brachina Gorge.

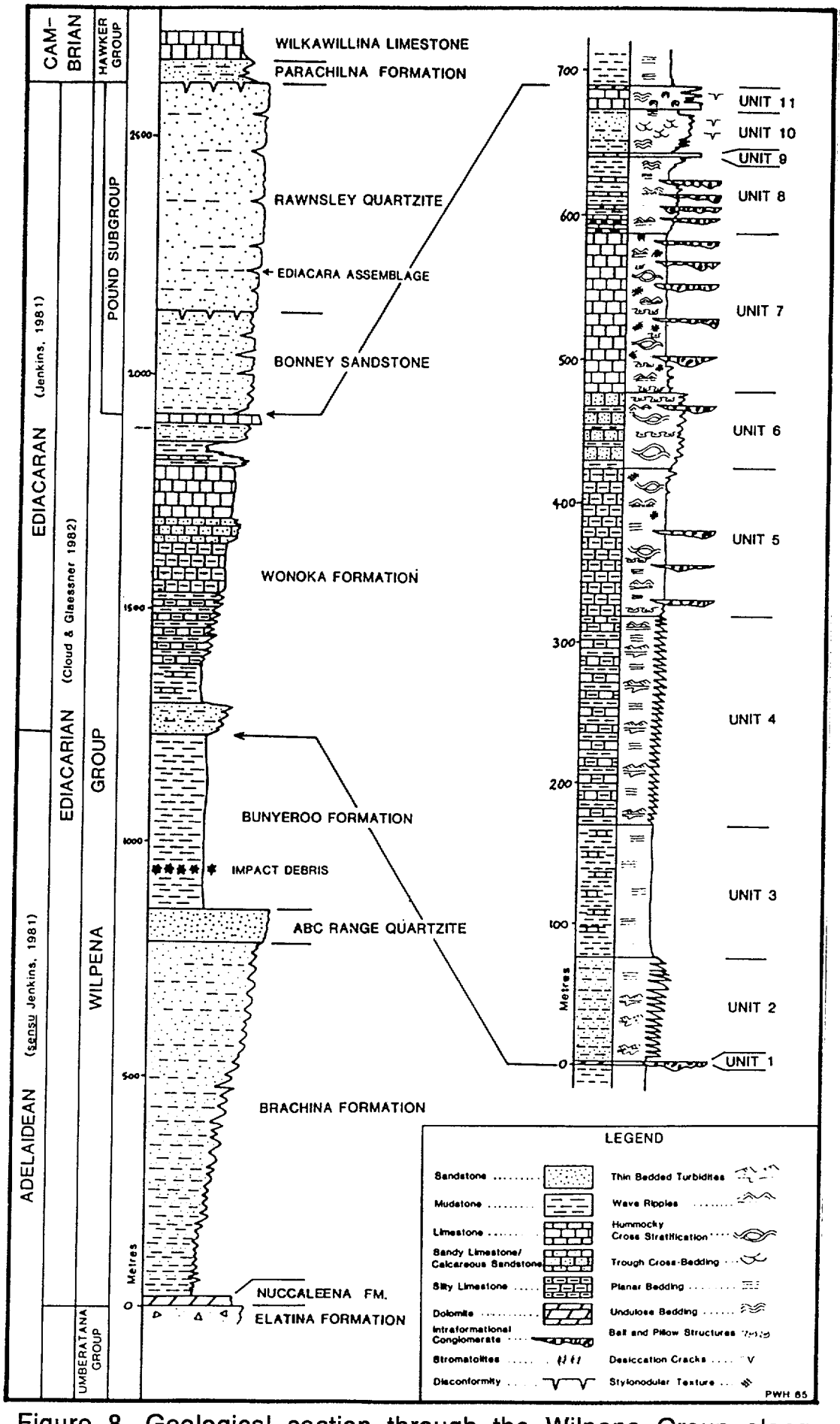


Figure 8. Geological section through the Wilpena Group along Bunyeroo Creek, showing details of the Wonoka Formation (after Haines, 1986).

FIELD GUIDE

Day 1 Tuesday, May 11 -- Arrival

Arrive in Adelaide, South Australia. Afternoon - Introductory remarks and examination of specimens.

Day 2 Wednesday, May 12 -- Travel to Parachilna

Drive to Parachilna in the Flinders Ranges, perhaps via the Clare Valley, the source of some excellent wines.

Day 3 Thursday, May 13th --- Bunyeroo Gorge

This is the formally nominated type section of the Ediacaran (Jenkins, 1981) and Ediacarian (Cloud and Glaessner, 1982) Systems and/or Periods (see Figs. 7 to 11). The stratigraphic interval represented by the Ediacarian is approximately twice that of the Ediacaran. The Ediacarian is inclusive of the Nuccaleena Formation (a thin post-glacial cap dolomite) and embraces the Wilpena Group, which is locally overlain disconformably by the Early Cambrian Parachilna Formation, the latter mostly an impure, bioturbated sandstone. The Ediacaran is inclusive of the Wearing Dolomite 'Member' arbitrarily assigned to the basal part of the Wonoka Formation (Unit 1) of Haines (1987, 1990), and includes the remainder of the Wilpena Group. The thickness of the Ediacarian in the vicinity of Bunyeroo Gorge is c. 3350 m while the Ediacaran is c. 1500 m. In the northeastern Flinders Ranges the Ediacaran attains much greater thickness with the arenaceous rocks of the Pound Subgroup reaching as much as 3000 m in the Gammon Ranges, and equivalents of the uppermost Wonoka Formation and Bonney Sandstone represented by siltstones, sandstones and carbonates of the Billy Springs Formation also recently measured as c. 2.3 km thick.

Stop 3.1 -- Trezona and Elatina Formations

In the vicinity of Brachina and Bunyeroo Gorges (Figs. 7 and 10) the closing phases of deposition of the Umberatana Group are represented by the Trezona and Elatina Formations (e.g. Preiss, 1987). Within this area the typically reddish coloured Trezona Formation comprises fine-grained stromatolitic, oolitic and intraclastic limestones alternating with greenish-grey, laminated calcareous shale and siltstone. The intraclastic facies commonly consist of small, curled carbonate mudflakes set in sparry calcite

and have been described as showing a 'hieroglyphic' texture. A shallow water environment of deposition is represented.

Older parts of the Umberatana Group are commonly dark grey, grey or greenish grey in colour (though pink hues have been noted in an unmetamorphosed occurrence of the Sturtian glacials). The upward change to red coloration can be shown by false colour satellite imagery to be a regional phenomenon with ferric oxides of iron predominating. Later Umberatana Group sediments (predating the 'Elatina glaciation') persistently show abrupt cycles of deepening and shallowing, either linked to local diapirism (Lemon, 1989), or possibly reflecting glacio-eustatic phenomena (comparable cycles are beautifully exposed in the 'Marinoan' type section on the sea-coast immediately south of Adelaide). The change to red coloration may relate to an increasing trend in oxygenation of the contemporary atmosphere (Jenkins, 1991).

The Elatina Formation is recently recognised as marking the start of a new sedimentary cycle (Jenkins, 1990) above a regional surface of disconformity (Lemon and Gostin, 1990). Glacigenic facies are not well developed in the present area, where the Elatina Formation is thin and consists mainly of pink sandstone and feldspathic greywacke. To the north of Brachina Creek glacigenic diamictites include striated clasts and dropstone facies.

Diamictic rocks of the Elatina Formation and more northerly equivalents recognised as the Yerelina Subgroup reflect the last classical Precambrian 'glaciation' represented in the Flinders Ranges and have been correlated with similar facies in southwestern New South Wales, central Australia and north-western Australia (Preiss and others, 1978). However, their relationship with the numerous terminal Precambrian glacigenic 'events' of northern hemisphere continents is unclear, principally because of uncertainties in respect of local geochronological dates and their stratigraphic separation from classical Ediacaran fossil finds. This problem may find resolution with $\partial^{13}C$ carbonate and strontium isotope studies in the Flinders Ranges and central Australia. $^{87}Sr/^{86}Sr$ ratios for underlying rocks are relatively light (c. .707).

The Elatina Formation may be considered as a low-stand depositional systems tract bounded below by a type '1' erosive surface which forms a low-angle unconformity over the Trezona Formation. Thus different parts of the upper stratigraphy of the Trezona Formation are in effect erosional remnants (Lemon and Gostin, 1990). N. Lemon (pers. comm.) of Adelaide University has discovered circular structures showing an 'Ediacaran' style of preservation in an upper arenaceous member of the Trezona Formation southeast of Blinman.

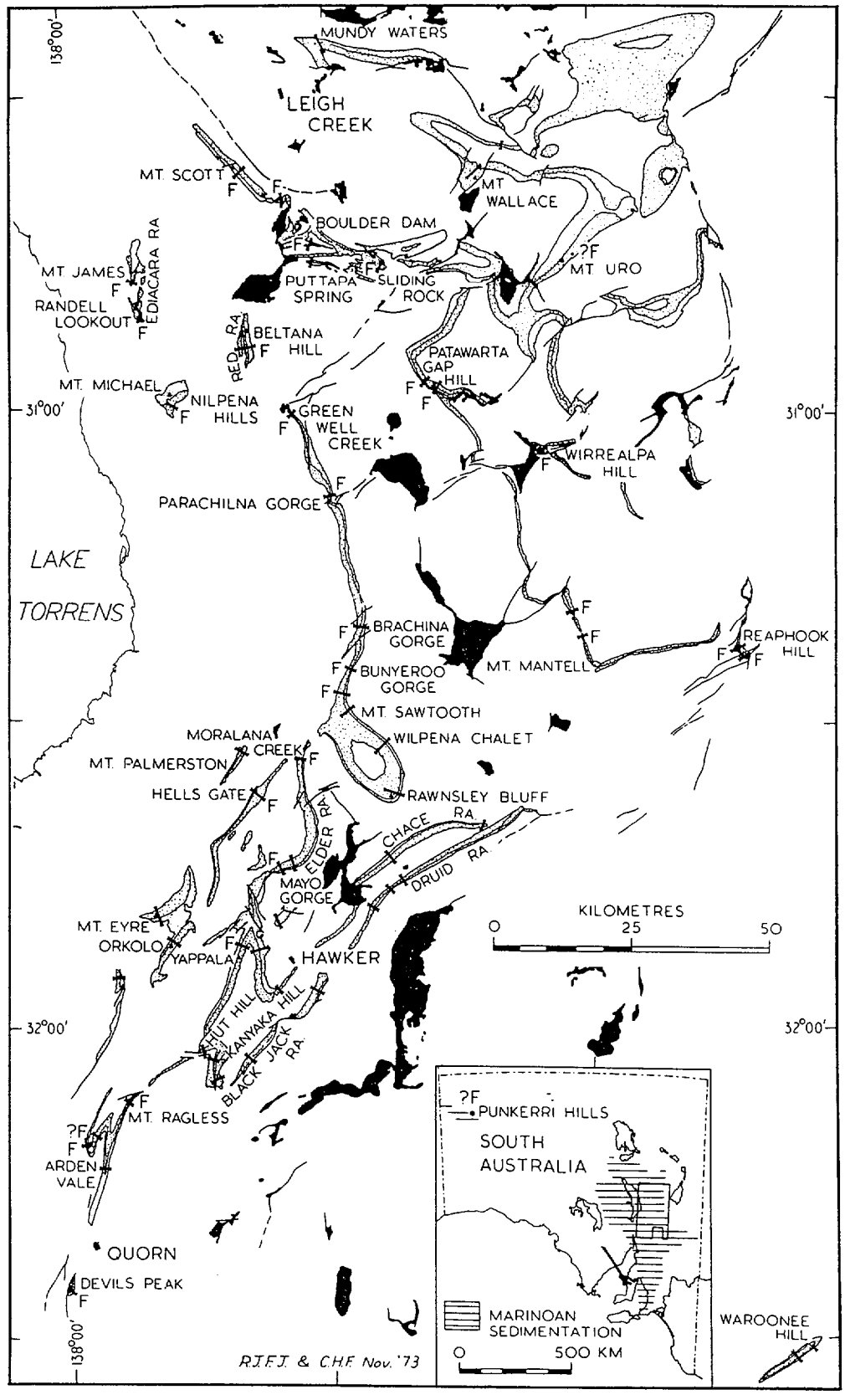


Figure 9. Locality map showing the likely extent of the Marinoan (pre-Ediacaran) and extent of outcrop of the Pound Subgroup in the Flinders Ranges (after Jenkins and others 1983).

Figure 10. Type area of the Wilpena Group, Bunyeroo Gorge, central Flinders Ranges (D. Gravestock, SADME).

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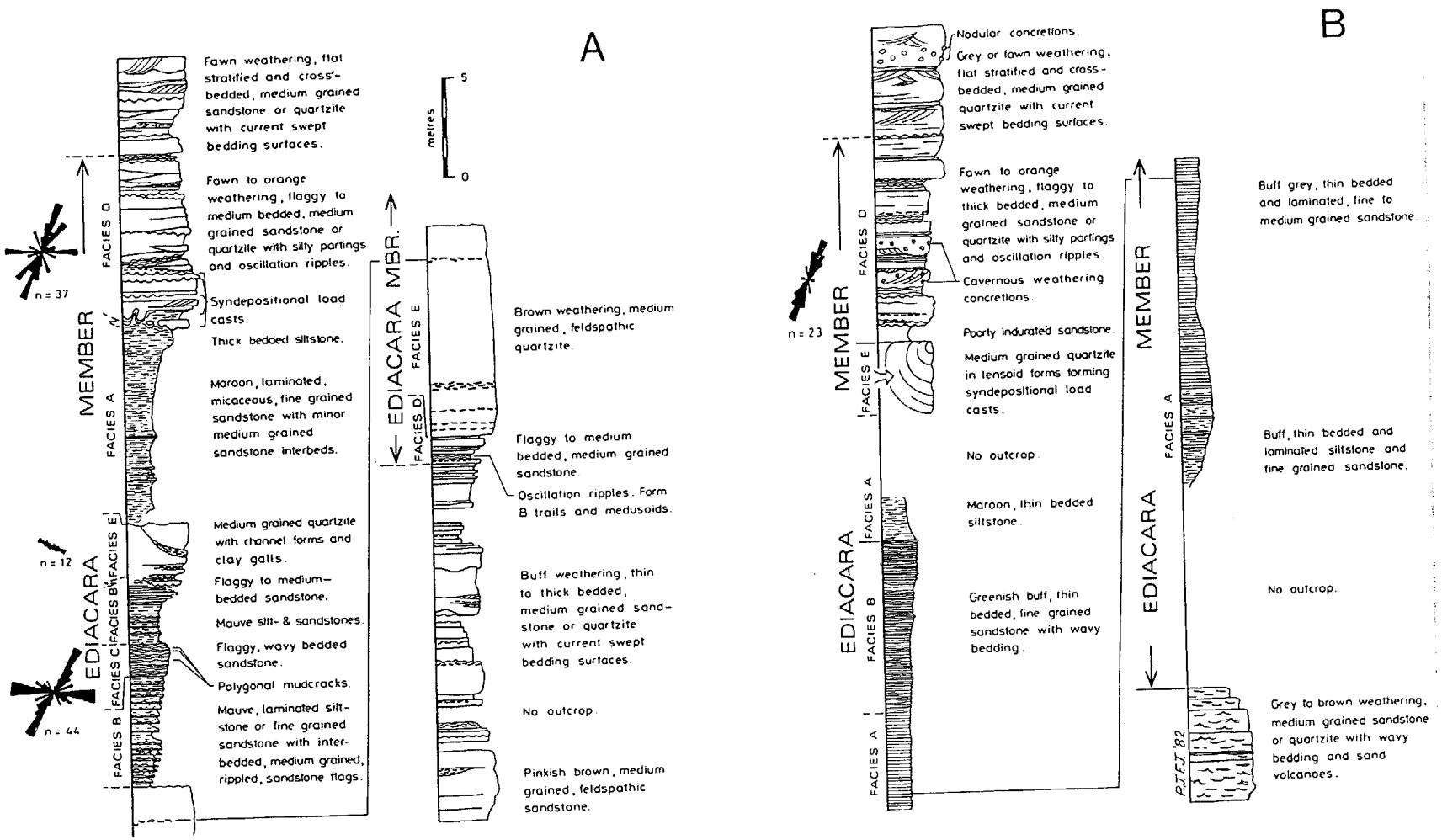


Figure 11. Measured exposures of Ediacara Member at Brachina Gorge (A) and Bunyeroo Gorge (B). Paleocurrents based on ripple-mark crests (after Jenkins and others, 1983).

Probable equivalents of the Elatina Formation in central Australia, the Olympic Formation = Pioneer Sandstone of the Amadeus Basin show similar erosive relationships on older strata. The Olympic Formation overlies offshore maroon siltstones where the section is more complete.

Arenaceous parts of the upper Elatina Formation are sometimes identified as the Seacliff Sandstone Member but there is disagreement as to whether this 'Member' is represented above or below the regional 'cap-dolomite' represented by the Nuccaleena Formation.

Workers at Adelaide University place the Seacliff Sandstone Member below the Nuccaleena Formation, and record dropstone diamictites interleaved between thick sandstone beds. Thus ice rafting persisted during this time.
Dyson (1992) extends an opposing view that in the type area of the Seacliff Sandstone at Hallett Cove, south of Adelaide, several dolomite beds occur within the Seacliff Sandstone which is thus an equivalent of the Nuccaleena Formation. Adelaide University workers would argue that the upper dolomite is the true equivalent of the Nuccaleena Formation.

Cross plot $\partial^{13}C$ and $\partial^{18}O$ carbonate data suggest a direct equivalence between the Elatina Formation and carbonates in a facies resembling the Seacliff Sandstone associated with the ?Olympic Formation at Hidden Valley in the eastern Amadeus Basin.

Stop 3.2 -- Wilpena Group

A classical view has developed that the Wilpena Group of the Flinders Ranges comprises two major sedimentary cycles, completing Precambrian deposition in the region (Fig. 4). Both cycles begin with thin dolomites which are locally of shallow water origin, overlain by basinal shales and turbidites indicating rapid subsidence. Subsequent parts of each cycle show progradation from turbidites to shelf sediments succeeded by sandstones of shallow tidal origin. Transgressions then led to formation of a starved basin (Bunyeroo Fm.), or open marine deposits of tidal and shelf origin (Rawnsley Quartzite).

The older cycle equates with the upper part of the Marinoan 'Series' (Mawson and Sprigg, 1950) in its type area, immediately south of Adelaide, and extends throughout the Adelaide Fold Belt and the adjacent Stuart Shelf. The younger cycle is limited to the Flinders Ranges and has been nominated as the Ediacaran System (Jenkins, 1981).

Modern sequence stratigraphy studies (see Christie Blick and others, 1990) greatly complicate this simplistic picture with some sequence boundaries potentially associated with the Wilpena Group (Fig. 4). While every one of the identified sequence boundaries is observable at places in the field, and can be mapped laterally for some distance or detected by laboratory data (e.g. isotopic changes), the interpretation of the surfaces in terms of classical seismic stratigraphy related to vertical reflector profiling is unclear. This is because on outcrop scale, low angle offlap/onlap relationships are seldom evident or are unknown, and thickness changes in the Flinders Ranges are legion due to growth faulting linked to diapirism.

Nowhere is the controversial genesis of the surfaces better illustrated than by the varied opinions, models and isolated observations linked to the deeply downcutting 'canyons' associated with the older Wonoka Formation. However, opinion related to other surfaces is beginning to become equally divided. Moreover, since some of the surfaces occur within apparent lithostratigraphic entities and may involve either a lengthy diastem or appreciable condensation, their identification with respect to the already complex (and frequently confusing) stratigraphic nomenclature in use poses a continuing dilemma. Perhaps this is the penalty of trying to generate a sequential stratigraphic framework without control of time constrained fossils. This is not the deny that the sequence surfaces may be real and regional, and that the classical lithostratigraphic mapping may also be discriminating broad patterns of basinal geometry. Local erosion is an implicit part of the wider pattern of cumulative sedimentation and one aspect of the question may be a matter of scale; the whole expanse of the Flinders Ranges might constitute but one part of a wide modern continental shelf and slope.

The buff-yellow, thin dolomite of the Nuccaleena Formation is a regional marker showing abrupt lower and upper contacts with the enveloping strata. Local angular relationships occur at its lower boundary in the northeastern Flinders Ranges, where tepee structures and intraclastic facies document shallow water deposition (Preiss, 1987). Fine, millimetric lamination possibly of cyanobacterial origin is common in the formation.
The dolomite has been linked to fairly abrupt warming after the Marinoan glaciation, and may be regressive, resulting from post-glacial isostatic rebound (refs. in Preiss, 1987). The palaeogeography of Nuccaleena time has been illustrated as the initiation of post-glacial transgression (Preiss, 1987, pp. 390 - 391).

Dyson (1992) identifies the base of the Seacliff Sandstone as a sequence boundary and considers this unit and intraclastic dolomites

as a lowstand systems tract. The question is posed here as to how this may be separated from the lowstand implied for the Elatina Formation. Dyson (1992) associates the Nuccaleena Formation (dolomite) with a maximum flooding surface related to the ensuing highstand systems tract embracing the Brachina Formation. This accords well with the fine (parallel) depositional banding sometimes seen in the dolomite and its local upward gradation into siltstones or intercalated siltstones and thin turbidite-like beds of the basal Brachina formation (e.g. at Hallett Cove). However, such a view is not consistent with the commonly observed thickness changes of the Nuccaleena Formation along strike, characteristics that seemingly might be better explained by chemical deposition associated with an irregular surface at the beginning of a transgression (e.g. Preiss, 1987). This would place a sequence tract boundary at the base of the Nuccaleena Formation or within it.

Close spaced samples of little altered parts of the Nuccaleena Formation show remarkably constant $\partial^{13}C$ carbonate isotopic values (c. -2 to -3‰ PDB) which indicate a marginally declining curve Frick, 1991) and are likely to provide a distinctive 'fingerprint' significant for interregional correlation.

Stop 3.3 -- Brachina Formation

The thick interval of shales, common storm-deposited sandstones and less prevalent turbidites constituting the Brachina Formation and its lateral equivalents (Ulupa Siltstone) signals rapid basin deepening or eustatic sea-level rise following the proceeding glaciation. Three members are present in Bunyeroo Gorge, the Moolooloo Siltstone Member, mainly of olive-green, micaceous siltstone, the Moorillah 'Siltstone' Member, chocolate or purple intercalated storm-bed sandstones and shales, and the greenish Bayley Range 'Siltstone' Member, an upward shallowing sequence of shales and thin, interbedded siltstones and sandstones that include mudcracks and common ripple marks.

In the eastern Amadeus Basin the upper glacial to succeeding basinal transition is rather similar. The Elatina Formation has the Olympic Formation (= Pioneer Sandstone) as its equivalent and sandstones also tend to predominate up section. The yellow and white or pinkish, sharp-based cap-dolomite of variable thickness (mainly <6m) is commonly banded and locally intraclastic. Grey limestones above a dissolution surface grade up into green siltstones with thin limestone interbeds. Locally, ferruginous grits or gritty rip-up conglomerates top the cap dolomite or its covering carbonates which record $\partial^{13}C$ carbonate values of -6 to -6.4‰ PDB. Presumably the erosional phenomena are associated with a sequence boundary. The succeeding siltstones with limestone interbeds may be equivalents of the Moolooloo Siltstone Member and the carbonates record a

marked negative isotopic excursion of -6.9 to -7.5‰ PDB. The Moolooloo equivalent is recognised here as a new formation.

Stop 3.4 -- Moorillah Member

The structure *'Bunyerichnus dalgarnoi'* Glaessner, 1969, variously claimed to be the trail of an early coelomate or the imprint of a fossil medusa (e.g. Cloud and Glaessner, 1982) is from the ?lower part of the Moorillah Member in Bunyeroo Gorge. The structure consists of lower and upper counterpart moulds showing a family of nearly concentric curves scribing areas about a third the length of the apparent circumference. Not quite regular, transverse serial impressions extend between several of the outer arcs. These impressions may show adjacent small mounds of displaced sediment and ridges of sediment are displaced from the scribing arcs. The transverse marks of the inner and outer bands do not necessarily correspond and their width varies considerably.

Prising away a thin lamina of sediment (c. 1 mm thick) disclosed a series of transverse marks at a different level in the sediment. The writer considers that the sharp outline of these marks is primary and that they have counter-impressed through the covering laminae to form the more weakly marked overlying band. Thus the two bands of transverse marks seem to have been formed at different times during sedimentation, ruling out the likelihood of the structure being an animal trace. The bed showing the structure formed in a high energy regime shows current lineae extending through the centre of the enclosing arc of circular markings, thus negating the possibility of the structure being the mould of a medusa. The scribed arcs and sub-regular transverse imprints may have been made by the jigging motion of an alga swinging in the current. Fragments suggested to be of vendotaenids occur close to this stratigraphic level in a bore to the west (Damassa and Knoll, 1986).

South of the present region ferruginised bands in the Moorillah Siltstone Member mark weathered tuffs characterised by siliceous laminae composed of devitrified glass shards. This material offers promise for geochronological studies. A study on several close spaced siliceous layers found that any indigenous volcanic zircons are vastly outnumbered by admixed frosted sedimentary ones and hand picking is barely likely to be economical.

South of Wilpena Pound, lower and upper parts of the Moorillah Siltstone Member include numerous pot-casts and gutter-casts with fills of intraclastic material and laminated siltstone and fine sandstone (Jenkins and others 1981). These structures may form deep-conical or even pipe-like shapes, but appear to be of erosional origin. Isolated examples have been located in Bunyeroo Gorge.

Stop 3.5 -- ABC Range Quartzite

The ABC Range Quartzite caps the first major cycle of the Wilpena Group, and represents the regressive phase of a highstand systems tract. The Brachina Formation and ABC Range Quartzite together constitute the major part of the Sandison Subgroup of Dyson (1992).

The ABC Range Quartzite is well exposed where Bunyeroo Creek passes through the ABC Range, although the sharp conformable contact with shales of the overlying Bunyeroo Formation is not exposed here. The ABC Range Quartzite comprises white medium-grained crossbedded quartzite with interbedded fine sandstone and siltstone. Plummer (1978) suggests an intertidal sand-flat depositional environment in this region. Rare vertical tubes up to about 1.5 cm in diameter show disrupted internal remnants of bedding suggestive of water escape.

The top few metres consist of coarse-grained to pebbly ferruginous sandstone sharply overlying a planar surface on finer grained quartzite. This regional surface is locally unconformable on the older sequence (Plummer, 1978). It constitutes a sequence boundary below the highstand cycle represented by the finely laminated, offshore siltstones of the Bunyeroo Formation.

The ABC Range Quartzite is characteristically arkosic, perhaps indicating that its sediments were supplied at a time of cool climate. Isolated pebbles rarely located in older parts of the Bunyeroo Formation may be an indication of (?) seasonal ice rafting. Red stained intervals of the ABC Range Quartzite almost certainly reflect a tuffaceous component.

The labile grits at the top sequence boundary may have been derived from shedding diapirs known to penetrate the formation, and probably include comminuted relicts of the common volcanics encountered in diapiric 'cores'.

Stop 3.6 -- Acraman impact ejecta

A thin (<10 cm) but widespread horizon containing volcanic clasts of largely dacitic composition occurs 80 metres above the base of the Bunyeroo Formation. This horizon has been traced widely (in outcrop and drillcore) along the western margin of the Adelaide Fold Belt. The clasts, which range from sand-size to 30 cm in diameter, show petrographic evidence of intense shattering and shock metamorphism, and display rare micro-shatter cones. Large clasts show evidence of vertical fall (dropstone) emplacement. The dacite petrographically resembles that of the Gawler Range Volcanics, and is of similar age (c. 1600 Ma). It has been suggested that the clasts represent ejecta dispersed from the 30 km diameter Acraman

impact structure in the Gawler Ranges c. 300 km west of Bunyeroo Gorge (Gostin and others 1986). This implies ejection velocities of the order of 2 km per sec. and ablation effects may be expected. Note that the clasts comprise parts of a shattered, chalcopyrite rich ore body. The Bunyeroo Formation (c. 450 m thick at Bunyeroo Gorge) dominantly comprises monotonous maroon shales, although a dark grey pyritic interval is present near the middle of the unit. Concretionary carbonates are present near the top. The environment of deposition is uncertain but a relatively deep setting (below storm wave base), starved of coarse sediment, is indicated.

Stop 3.7 -- Wonoka Formation, Unit 1

The amended (Gostin and Jenkins, 1983) base of the Wonoka Formation, which is also base of the Ediacaran, is marked by a 2 m thick cupriferous dolomitic horizon which forms Unit 1 of Haines (1987, 1990). The thin planar dolomites, interbedded with green dolomitic shales, are locally reworked into lenses of intraformational conglomerate with associated beach imbrication, stromatactic voids, coarse rounded labile grains and rare stromatolitic layering.

The thin planar dolomites may be considered as part of a condensed regressive sequence tract terminating the depositional cycle represented by the Bunyeroo Formation. The intraclastic dolomites are associated with a sequence boundary at the real base of the older Wonoka depositional cycle. The $\partial^{13}C$ carbonate isotope signal for the dolomites is -0.08 to -3‰ but has not been well studied (Jansyn, 1990; Ayliffe, 1992; Urlwin, 1992).

Unit 1 is overlain by 70 m of thin to thick bedded, brown, fine sandy calcareous and dolomitic turbidites, interbedded with maroon shales (Unit 2). Bouma sequences (dominantly BCE in lower beds) and flute moulds are well developed. Palaeocurrents were directed towards the east at this locality. Regional palaeocurrent trends and thickness variations suggest a coalesced fan morphology for this unit, perhaps representing the distal portion of deltas fed from the west (Haines, 1990).

Stop 3.8 -- Wonoka Formation, Units 3 & 4

Unit 2 is overlain by 95 m of reddish calcareous shale containing thin pink limestone interbeds (Unit 3). Palaeocurrent measurements are usually directed at right angles to those obtained from turbidites above and below, indicating the possibility of contouring currents. Randomly orientated tool marks on the top of some beds indicate storm-wave bottoming.

The top of Unit 3 is marked by gradual incoming of thin bedded brown calcareous siltstone and green silty limestone turbidites characteristic of Unit 4. Unit 4 (150 m thick) is marked by a cyclic alternation of green and brown coloration, with green becoming dominant up section. The upper beds of the unit display the first evidence of sediment reworking by storm-waves in the form of oscillation ripples and 'micro-hummocky cross-stratification'.

Deep (c. 1 km) canyon-like structures associated with the older Wonoka Formation commonly show erosive shoulders extending to about the Unit 3 - 4 transition. No obvious sign of a regional sequence boundary has been detected at this level in the central Ranges. In the northern and southern Flinders Ranges the Wonoka canyons erode lithologies lithified at a prior time (Bunyeroo Formation and older). Large rounded eroded blocks of the Bunyeroo Formation in olisthostromes within a southern canyon show a prior cleavage and locally a veneer of carbonates on the canyon wall forms at 60° angular unconformity over the intact Bunyeroo Formation.

In the eastern Ranges an episode of faulting lead to differential erosion of the Bunyeroo Formation prior to the formation of the sequence boundary at the base of the Wonoka Formation. Thus the time span of the condensed stratigraphy at the top of the Bunyeroo Formation and base of the Wonoka Formation may be rather longer than might be suspected from the paraconformable relationship in the central Ranges.

Stop 3.9 -- Wonoka Formation, Units 4 to 7

There is a rapid transition from distal shales, interbedded storm beds and turbidites at the top of Unit 4, into thicker bedded grey-green silty limestones of Unit 5 (105 m thick). This Unit displays abundant evidence of storm reworking, including hummocky cross-stratification (HCS) and associated oscillation ripples. A layer containing large ball and pillow structures is well exposed at the bend in Bunyeroo Creek. Unit 6 (53 m thick) was deposited under similar conditions to Unit 5, but consists dominantly of brown calcareous siltstones and fine sandstones interbedded with minor grey shale. HCS is well developed throughout the unit. Unit 7 (110 m thick) consists mainly of relatively pure grey and greenish limestones displaying planar beds, climbing oscillation ripples and minor HCS. Stylonodular bedding is abundant.

A canyon-like structure located in the vicinity of the Wilpena Chalet reveals considerable syndepositional increase in thickness of Units 3 - 7 within its fill (Jansyn, 1990). The coincidence of timing between this subsidence and the formation of other canyons seems too close to be accidental and there is abundant evidence for the wider collapse of the Wilpena structure being linked to extensional

faulting. Dislocated slide blocks from the shoulders and veneering carbonate on one wall of the structure parallel known characteristics of the other canyons.

Stop 3.10 -- Wonoka Formation, Units 8 & 9

The base of Unit 8 is recognised as a sequence boundary (Christie Blick and others, 1990) and corresponds to a sharp positive trend in $\partial^{13}C$ carbonate isotope values (see Urlwin and others this guide). The extensional movements just described had ceased prior to Unit 8 time.

Unit 8 (54 m thick) represents a rapid return to more distal storm dominated sedimentation. This change was probably related to the formation of partially effective wave barriers of diapiric origin along the platform margin north of Bunyeroo Gorge (Haines, 1987). The Unit consists of thin oscillation rippled grey limestones and thicker HCS limestones, interbedded with an upwardly increasing proportion of green calcareous siltstone and shale. Intraformation conglomerates are abundant and form the base of many beds. The top of Unit 8, composed largely of green siltstone, contains a thin (1 - 2 m) horizon bearing lenticular silty, laminated limestones displaying complex presumed trace fossils of sediment-ingesting organisms on bedding surfaces. These traces are rare at Bunyeroo Gorge but common elsewhere. Unit 8 is overlain by a 1 m thick band of shallow water limestone displaying microbial laminates (Unit 9).

Bedding plane thin sections from a dark carbonate high in Unit 8 in Brachina Gorge include large leiosphaerids and filamentous 'algal' remains (Jenkins and others, 1992).

Stop 3.11 -- Wonoka Formation, units 10 & 11

Unit 10 (31 m) comprises siliciclastic facies identical to parts of the Bonney Sandstone. Greenish shallow marine siltstone and sandstone near the base of the unit grade upward into red, poorly sorted and mineralogically immature sandstones, of marginal marine to possibly terrestrial origin. Desiccation cracks and mud flakes are abundant. Rare medusoids and other fossils of uncertain systematic position have been found in shallow marine sandstones near the top of the unit, which is capped by 4 m of green calcareous siltstone.

Sedimentation of the Wonoka Formation was completed with the deposition of Unit 11, a thin (16 m at Bunyeroo Gorge) but widespread shallow water carbonate. This exposure shows several coarsening upward cycles of thin bedded grey to black micritic and peloidal limestones grading into black calcarenite (dominant particles are often ooids). Coarse particles are often preferentially dolomitized. Several stromatolite horizons are present in the upper

parts of some cycles. The uppermost bed of Unit 11 is composed of yellow dolomite, the top surface of which is a phosphatic hardground. Phosphatic surfaces are also found elsewhere in the unit.

Stop 3.12 -- Bonney Sandstone

The Wonoka Formation is conformably overlain by dominantly red coloured (green at the base) sandstones and siltstones of the Bonney Sandstone. Good exposures of the entire formation (220 m thick) occur along the walls of Bunyeroo Gorge.

The Bonney Sandstone consists dominantly of red, fine- to very fine-grained, haematite-rich sandstones, with subordinate, pink, medium-grained subarkoses. Within the red sandstones, characteristics such as lenticular and wavy bedding, micro-cross-lamination, mud cracks, clay galls, well-rounded granules and possible rain prints indicate an extremely shallow-water environment of deposition with probable subaerial exposure, and led Forbes (1971) to tentatively infer a tidal-flat environment. The subarkoses are well sorted, show persistent low-angle cross-bedding and minor channelling, and are interpreted as representing tidal sand sheets. A trail resembling Form B of Glaessner (1969) and probable small medusoids were collected by Dr Mary Wade in the Bonney Sandstone at Brachina Gorge (Wade, 1970, p. 92).

Stop 3.13 -- Rawnsley Quartzite

Clean quartzites containing minor feldspar comprise the base of the Rawnsley Quartzite, which overlies the Bonney Sandstone at a sharp flat surface with a lag of coarse well rounded granules. A large downcutting channel is associated with the same surface near Mt Michael on the Nilpena pastoral property (Nedin, 1990), indicating the probable presence of a sequence boundary.

The Rawnsley Quartzite is some 470 m thick in Bunyeroo Gorge. The Ediacara fossil horizon occurs about 80 m above the base of the formation within the Ediacara Member. Impressions of fossil medusoids and the annelid worm *Dickinsonia* are present on the base of some sandstone slabs at this locality. Trace fossils are also present at the base of the horizon. The Rawnsley Quartzite is composed largely of clean, current-bedded quartzites and sandstones of active, shoreface environments.

In the southern Flinders Ranges, the cross-bedding of sandstones both above and below the Ediacara Member shows generally low variance and indicates a persistent, eastward trending current. Further north and east, palaeocurrent trends are polymodal, but still with an eastern bias, perhaps reflecting currents of tidal or wind

origin. For the major part of the Rawnsley Quartzite, the likely environment was a transgressive regimen with winnowed upper shoreface sands, laterally migrating tidal channels and inlets, and sandbar facies in a shallow sea, a closely comparable model of deposition to that which Hobday and Tankard (1978) proposed for the lower Palaeozoic Peninsula Formation, South Africa. In Bunyeroo Gorge sandstones with unusual 'dish'-structures occur below the Ediacara Member. In plan view the dish structures may be polygonal, and curling of some thin sandy layers suggests their intermixture with cyanobacterial mats which shrank on desiccation (A. Seilacher, pers. comm.).

In Brachina Gorge and some other areas lower parts of the Rawnsley Quartzite consist dominantly of channel deposits with included clay gall conglomerates and common cross-bedding. A spectacular surface of disconformity below the channel deposits occurs above a leached zone evident in the underlying sandstones. The history of this leached zone is unresearched. Gehling (1987; 1988) recognised this disconformity as occurring at the base of his (downwardly) revised Ediacara Member.

At Parachilna Gorge a channel is erosive on underlying sandstones, developing an observed relief of as much as 10 m at the contact, and clasts identifiable as the underlying sandstone occur in the channels. Thus the contact between the two is a disconformity, apparently with lithification of parts of the older sandstone prior to deposition of the subsequent channel sands.

The whole question of the distinction of the Bonney Sandstone and Rawnsley Quartzite remains problematical because of the presence of pink and while granule-rich sandstones within both.

In Bunyeroo Gorge the Ediacara Member includes a thick expression of facies 'A' (of Jenkins and others 1983), khaki to reddish brown laminated siltstones with scattered coarse granules, facies 'B', intercalated in silt and sandstone beds, and facies 'D', medium to thick bedded clean quartzites with well defined partings and abundant fossil remains. At the base of 'D' is a zone of prominent load casts.

With reference to other sections, Gehling (1982, 1983, 1987) includes channel sandstones and intervening thick siltstones within his wider definition of the Ediacara Member, noting that rare fossil remains occur at varying levels through this facies association.

Stop 3.14 -- Ediacara Member

The distinctive lithologies comprising the Ediacara Member are usually overlain in turn by medium- to coarse-grained, flat-

stratified or trough-cross-bedded sandstones similar to those that may occur below. This upper and major part of the Rawnsley Quartzite is relatively mature and constitutes a blanket sandstone of regional extent. Cross-bedded sets up to 3m thick are common in its upper parts, and thence tend to be succeeded by thin beds with oscillation ripples and occasional pebbles, possibly a regressive facies. Lenticular orthoquartzite bodies up to 20 m thick, usually with clay gall conglomerates at the base of the lenses, characterise much of the Rawnsley Quartzite at Wilpena Pound and the Chace and Druid Ranges, and apparently evidence extensive channelling. The possibility that such facies may be non-marine in origin and fluviatile has yet to be fully explored.

Stop 3.15 -- Parachilna Formation

The Rawnsley Quartzite is overlain disconformably by the Parachilna Formation, which locally marks the base of the Cambrian. If time permits, inspection may be made of beds containing the U-shaped burrow *Diplocraterion* in the Parachilna Formation, and archaeocyath-bearing carbonates of the overlying Wilkawillina Limestone.

Day 4 Friday, May 14th -- Umberatana Syncline

Stop 4.1 -- "Wall plaster" carbonate veneer Locality
(C C von der Borch)

From the Umberatana Station shearers' quarters, proceed through the gate near the northern shed of the homestead complex (Fig. 7). Follow this track towards the airstrip as it winds through low hills of the Ulupa Siltstone. At a distance of about 1.2 km, just before the final ridge of Ulupa Siltstone, take the track to the right (east) for about 0.8 km and stop at the small dam. The section (Fig. 12) is situated 300 m to the east of the dam.

In a regional sense, upper slopes of the lower Wonoka Formation canyons are typically mantled by a thin (0.5-5m) veneer of micritic or locally recrystallised microsparry limestone. This veneer may be brecciated and reworked into cross-bedded calcarenites, while in other localities it comprises in-situ sheets of micritic calcite. In some areas it forms giant (several metres relief) tepee structures. Thin section petrography shows it to be a relatively structureless calcite micrite or calcite microspar. Rare sub-millimetre trace fossils and possible evaporite pseudomorphs occur in some localities. The carbonate veneer contributes some of the many types of clasts that constitute lower canyon wall and canyon floor breccias. The origin of this somewhat diverse "wall plaster" is currently unresolved. It clearly veneers the canyon unconformity

and was most likely deposited during hiatus, when potential terrigenous dilution was minimal. Its depositional setting must, at least in part, have been subaqueous.

Figure 12 shows the section which will be traversed. In this locality the "canyon unconformity" has cut down through bluish grey shales of the Bunyeroo Formation into light grey shales of the upper Ulupa Siltstone. Above the unconformity lies a several metre thick megabreccia of metre size micritic carbonate clasts derived from former "wall plaster", along with clasts of Ulupa Siltstone (unit 1). This is overlain by about 25 m of locally slumped siltstone (unit 2) containing scattered metre-sized clasts of micritic carbonate, again representing resedimented "wall plaster". The overlying 30m (unit 3) mainly comprises a dark grey slumped siltstone with local white microspar breccias and small (several cm) tabular clasts which have undergone dissolution during weathering. Unit 4 is a microspar breccia. Unit 5, approximately 30 cm thick, is a sheet of microsparry limestone, interpreted as the only in-situ fragment of "wall plaster" in this section. Unit 5 caps the canyon wall-mantling complex in this locality, and marks the boundary between this complex and an overlying succession of interbedded cyclic carbonate-shale couplets.

Stop 4.2 -- Billy Springs Formation -- Possible Glacials(?)

(N. Christie-Blick, Lamont-Doherty Geological Observatory, New York)

Stratigraphically more than 2 km above the upper of the two main glacial units in the Adelaide Geosyncline is an enigmatic unit of diamictite in the Billy Springs Formation. The diamictite is as much as 50 m thick and it can be mapped for 10 km along the southern limb of Umberatana syncline (von der Borch and Grady, 1982; DiBona, 1989, 1991). According to DiBona, the diamictite is composed of pebble- to boulder-sized clasts of chert, dolomite, sandstone, rare granite and basalt in a massive silty matrix. It overlies laminated siltstone and sandstone, locally abundantly slumped, and is overlain by similar laminated siltstone that grades upwards into wave-rippled and cross-stratified sandstone and siltstone.

The diamictite has generally been interpreted as a product of down-slope mass flow. The presence of lonestones (dropstones?) in laminites and of "exotic" clasts led DiBona (1989, 1991) to suggest that the rock might be of glacial origin. DiBona's photographs of lonestones are tantalising, but there are nevertheless certain difficulties with the glacial interpretation. 1) Apart from this restricted unit, there is absolutely nothing else in the regional stratigraphy that remotely implies a glacial or glacial-marine

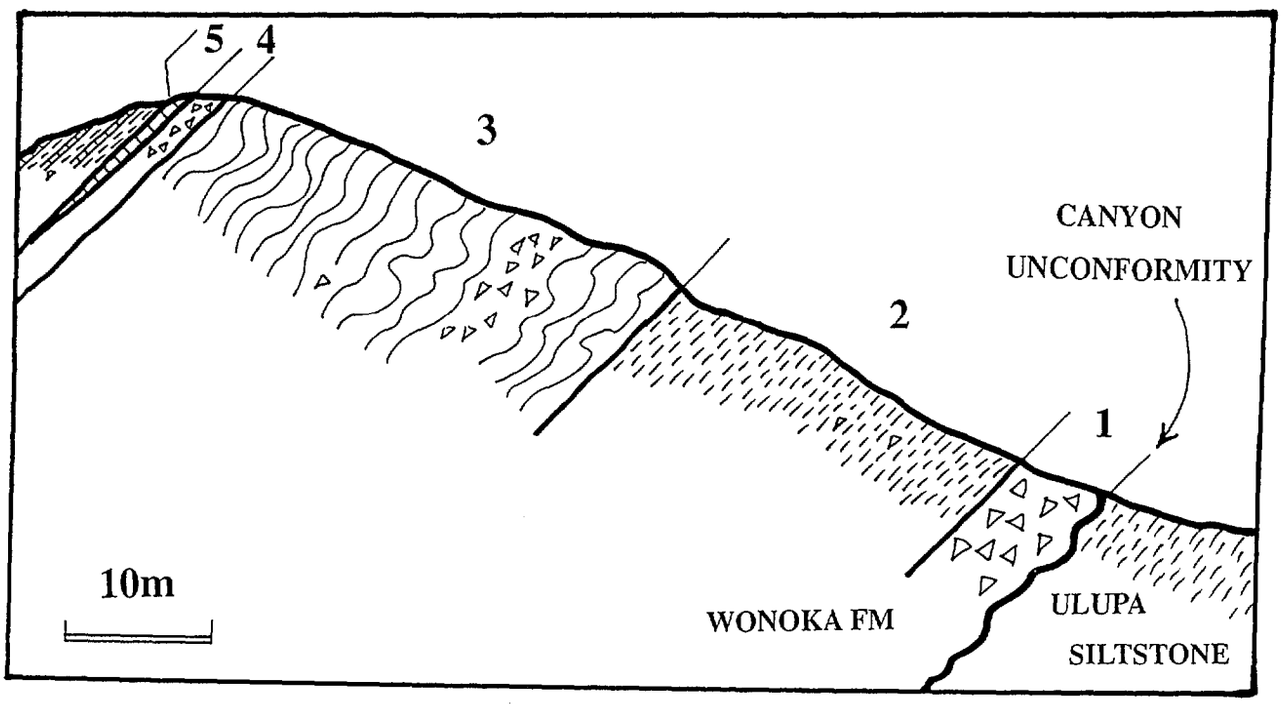


Figure 12. The wall plaster carbonate veneer locality at Umberatana.

environment at this horizon. 2) There is abundant evidence for sediment gravity flow on a paleoslope, which therefore represents a viable alternative for deposition of the diamictite. Nevertheless, in view of the potential importance of this unit if it proved to be glacial, it seemed sensible to include it in the excursion. We should examine the evidence critically and attempt to reach a consensus on the outcrop.

Stop 4.3 -- Uratanna Formation and the base of the Cambrian System, Angepena Syncline.

See Appendix B (J. Mount, Department of Geology, University of California, Davis, California). See Fig. 7.

Day 5 Saturday, May 15th -- Wonoka Canyons, Umberatana Syncline

(N. Christie-Blick, Lamont-Doherty Geological Observatory, New York)

Perhaps the most spectacular surface in the Adelaide Geosyncline is a sequence boundary characterised by a series of kilometre-deep incised valleys at or near the base of the Wonoka Formation (Fig. 4). These valleys were initially interpreted by Thomson (1969), von der Borch and others (1982, 1985) and Haines (1987) as submarine canyons, cut and filled in a relatively deep marine setting, analogous to the Neogene canyons of modern continental margins. Papers by Eickhoff and others (1988), von der Borch and others (1989) and Christie-Blick and others (1990) describe features of the sedimentary fill that cast doubt on this interpretation, specifically the presence of sedimentary structures such as oscillation ripples and hummocky cross-stratification that have been taken to imply sedimentation above storm wave base.

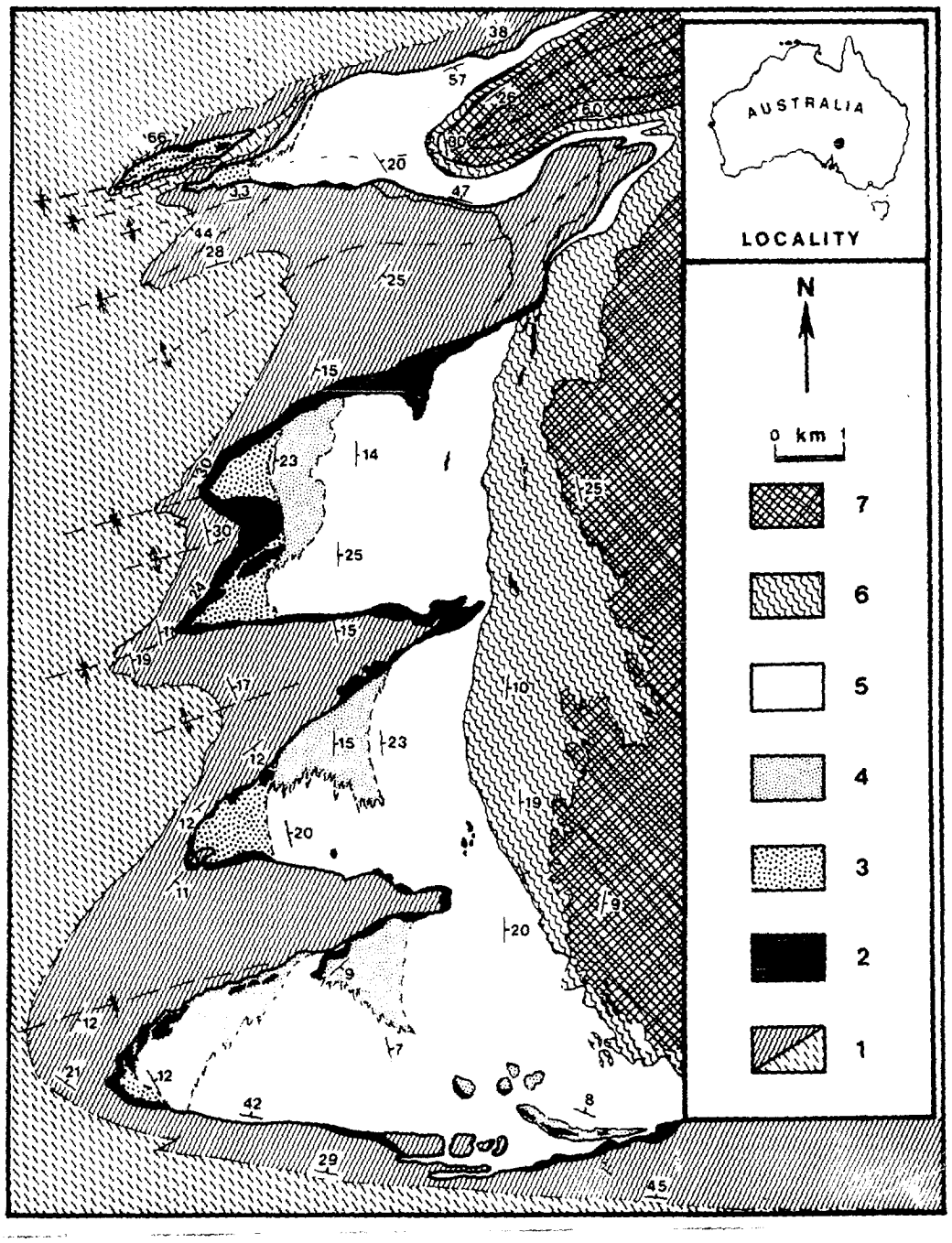
Recent work at Umberatana syncline suggests the presence of higher-order incised valleys within the sedimentary fill of the large-scale valley, and that conglomerates previously interpreted as submarine sediment gravity flow deposits are at least in part fluvial. The reinterpretation of the Wonoka canyons is significant for two reasons. First, if the interpretation is correct, it implies either an unusually large fall of sea level or large-scale uplift of the Adelaide Geosyncline during an interval lacking corroborative evidence for significant crustal deformation. Second, it casts doubt on the interpretation of valleys at other horizons, which until recently were generally ascribed to mass wasting in a marine environment, but which are filled by sedimentary facies comparable to those of the Wonoka Formation. This includes the valleys in the Pound Subgroup in which are preserved the Ediacara fauna (Gehling,

1982), and the valleys at the base of the Uratanna Formation located immediately below the Precambrian-Cambrian boundary (McDonald, 1992). The objective for this part of the excursion is therefore to examine the critical evidence. Advocates for opposing views will be present on the outcrop, in some cases for the first time. In spite of its inaccessibility for a large group, we have elected to visit the locality at Umberatana syncline where the rocks are best preserved and best exposed.

Here, the Wonoka Formation fills a series of three "incisions" into the Brachina Formation (in this area termed Ulupa Siltstone; Fig. 13). Bunyeroo Formation is preserved only locally above the Brachina on the shoulder of the northernmost incision, actually a double incision in the shape of a W. In places, the erosion surface cuts almost to the level of the Nuccaleena Formation, and relief in the preserved Brachina Formation alone is about 750 m. The rocks are broadly folded into an east-plunging syncline, with a number of subsidiary folds indicated in Fig. 13. The structure is generally simple, but complicated by layer-parallel faulting especially on the northern and southern margins of the Wonoka outcrop, areas that we will avoid for the purpose of this excursion. An unusual feature of the Wonoka Formation at this locality is that the incisions appear to correspond to a single sinuous valley, characterised by systematic changes in paleocurrents from one incision to the next (von der Borch and others, 1985; Eickhoff and others, 1988; Fig. 14). This feature, as well as the observed contact relations, make it very hard to argue that the valley is actually not a valley but a result of detached normal faulting (e.g., Jenkins, 1990).

The principal components of the valley fill are calcareous siltstones and sandstones, for the most part fine- to very fine-grained, and ranging from diffusely stratified to pervasively parallel-laminated and cross-laminated. Typical sedimentary structures are flute casts, parting lineation and three-dimensional swaly current ripples and climbing ripples. Less common are large-scale trough cross-stratification, especially in somewhat coarser-grained sandstones; symmetrical oscillation ripples, in some cases flat-topped, double-crested or with cuspate crests, and with tuning-fork bifurcations in plan view; and hummocky cross-stratification. Ripples previously interpreted as due to combined flow (unidirectional flow with a superimposed oscillatory component) are now regarded by Christie-Blick as for the most part current ripples. Subsidiary components of the valley fill are carbonate-clast conglomerate and siltstone-clast diamictite, especially close to the northern margins of each incision, and a thin laminated carbonate veneer, best preserved on the upper valley walls. The diamictite is generally non-stratified and composed of a jumble of blocks as large as boulders derived by mass wasting of the adjacent Brachina Formation, which appears to have been consolidated but not lithified at the time. The

Figure 13. Geology of the Fortress Hill canyon complex in the Wonoka Formation at Umberatana syncline (from von der Borch and others, 1985). The field trip will focus especially on outcrops along the northern margin of the second incision from the south. Explanation of geological units (modified from original): unit 1, undifferentiated Umberatana Group and lower Wilpena Group below the canyon-cutting horizon, here at the base of the Wonoka Formation; unit 2, diamictite derived from the canyon wall and conglomerate that crops out preferentially along the northern side of each incision; units 3 and 4, quartz-rich and calcareous sandstones and siltstones; unit 5, calcareous siltstones and immature sandstones; unit 6, upper part of Wonoka Formation; unit 7, Billy Springs Formation. Recent sequence stratigraphic mapping demonstrates that units 2 to 5 are markedly diachronous facies assemblages. The tendency for the coarsest facies to be biased towards the north is thought to be due to syndepositional northward tilting. Within individual incisions, time-stratigraphic units thicken away from the canyon wall as a result of differential compaction.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conglomerate consists of pebble- to boulder-sized clasts of carbonate, siltstone and sandstone, and a small proportion of exotic clasts such as quartzite, in a poorly sorted sandy matrix. It ranges from disorganised to well stratified and rarely cross-stratified.

In the original interpretation (and in the view of at least some workers today), the clastic facies were regarded as an assemblage of fine-grained turbidites and coarser-grained marine sediment gravity flow deposits. However, there are several difficulties with this view. 1) Although the bulk of the sediment accumulated as event layers, none of the sediment has the overall attributes of classical turbidites. 2) No facies is truly massive. Much of the conglomerate is relatively well stratified, and therefore not consistent with a mass-flow interpretation. 3) It is difficult to reconcile the presence of oscillation ripples and hummocky cross-stratification, even if these are uncommon, with deposition well below storm wave base, or of large-scale cross-stratification with deposition from sediment gravity flows. In the alternative interpretation, the rocks are therefore interpreted as an assemblage of fluvial and shallow estuarine sediments dominated by flood-driven turbulent underflows.

Critical new evidence has been acquired of the vertical and lateral relations between facies in a high-resolution sequence stratigraphic context. Figure 15 is a composite of 24 measured sections along the northern edge of the central incision (Fig. 13). Details of lateral relations in the upper part of this interval about 120 m thick are shown in Figure 16 The conglomerates are not random layers derived from the canyon wall, but instead tend to overlie higher-order erosion surfaces (valleys) aligned parallel to the main valley. In some cases (e.g., Fig. 16), conglomerate *directly* overlies the deepest-water facies (finely laminated carbonate siltstone), an example of a marked facies discontinuity. Conglomerate tends to fine upwards overall, and to become better stratified and/or cross-stratified upwards. It onlaps the margins of the high-order valleys and passes laterally over tens to hundreds of metres into diffusely stratified or massive sandstone. There, the erosion surface loses its identity in a concordant succession of fine-grained predominantly current-rippled sandstone and siltstone. On the basis of these observations, the high-order valleys are interpreted as fluvially incised, and the conglomerate to have accumulated largely from traction in a fluvial environment. The tendency for these prominent but localized erosion surfaces to pass laterally into marine flooding surfaces lacking evidence for subaerial exposure appears to mimic the regional stratigraphic character of the main sequence boundary.

Day 6 Sunday, May 16th -- The Ediacara Fossil Beds and the Elatina Formation.

The Ediacara Fossil Beds

(R.J.F. Jenkins, University of Adelaide, South Australia)

The Ediacara assemblage (Figs. 17 and 18) is associated with sandstones that Mawson (1937) named the Pound Quartzite, a vast lithological body reaching a thickness of 2740 m (Reyner and Pitman, 1955), and which crops out in a broad arc up to 150 km wide and extending some 280 km in a N-S direction. Forbes (1971) formalised a two-member nomenclature for the Pound Quartzite, applying the name Bonney Sandstone Member to the lower red part and terming the upper white sandstones the Rawnsley Quartzite Member.

Ongoing studies of the Pound Quartzite indicate that not only do the Bonney Sandstone and Rawnsley Quartzite Members show characteristic and different lithologies, but suggest that they represent separate cycles of deposition. Jenkins (1975) proposed that they be considered separate formations, the Bonney Sandstone Member becoming the Bonney Sandstone, and the Rawnsley Quartzite Member becoming the Rawnsley Quartzite (e.g., Jenkins and others, 1983). Correspondingly, the Pound Quartzite was elevated in rank to become the Pound Subgroup, which forms the uppermost part of the late Precambrian Wilpena Group.

Wade (1970) showed that in the western flank of the Flinders Ranges the Ediacara assemblage occurs in a distinctive essentially single stratigraphic interval, a few tens to a hundred or so metres above the base of the Rawnsley Quartzite. This relationship is now known to be general for the fossil beds throughout much of the Flinders Ranges. The fossiliferous unit is identified as the Ediacara Member (Jenkins, 1975, 1981) of the Rawnsley Quartzite. Outcrops of the Pound Subgroup in the northeastern Flinders Ranges have yet to be properly explored and to date no fossils have been found in these very thick sections.

Stop 6.1 -- Rawnsley Quartzite

At the site to be studied, the base of the Rawnsley Quartzite is interpreted as the sharp change from purplish red siltstones and sandstones or quartzites of the Bonney Sandstone to overlying white sandstones seen locally to fill a large channel cut into the underlying unit (Figs. 17 and 18). Elsewhere, the contact is marked by a stringer of polished granules at a similar colour change. The older Rawnsley Quartzite contains numerous beds with sub-polygonal, dish-shaped structures is and clean, tabular, sandstone,

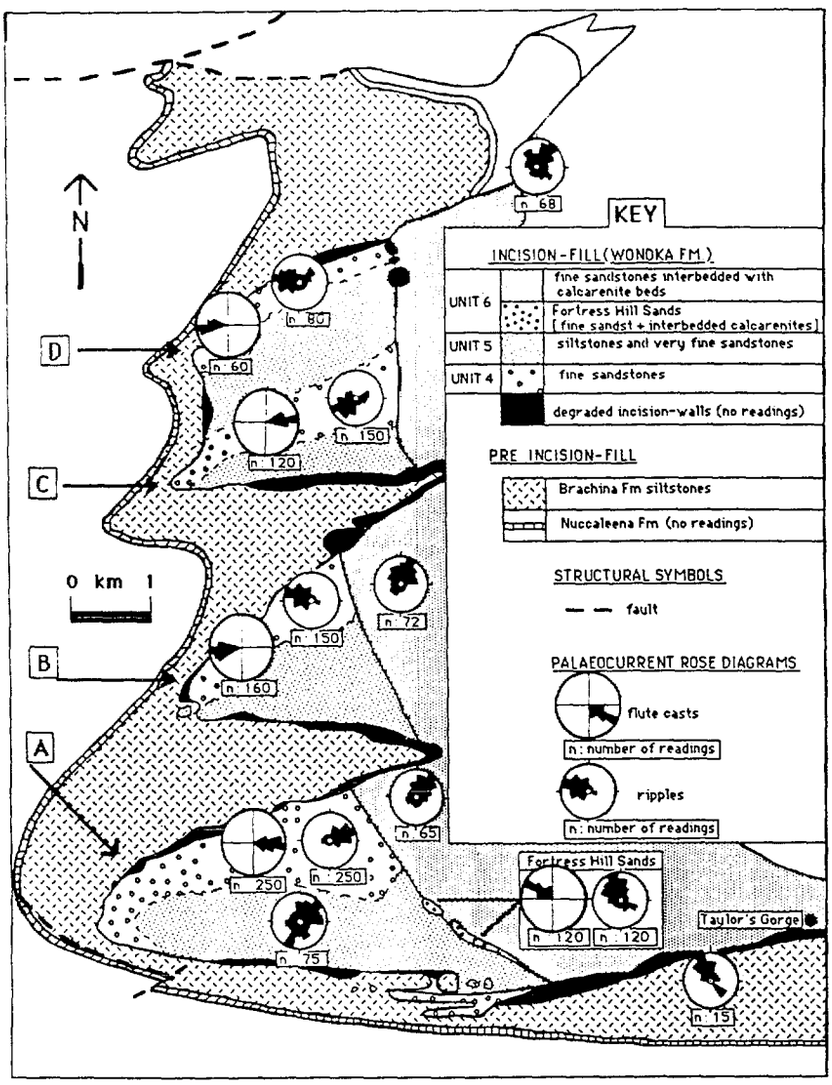


Figure 14. Palaeocurrent data from flute casts and current ripples in the Wonoka Formation (from Eickhoff and others, 1988). The reversals in flow direction from one incision to another indicate the presence of a sinuous valley. Recent work on paleocurrents in a high-resolution sequence stratigraphic context broadly confirms the overall picture. Note that the informal unit nomenclature used in this figure differs from that of Figure 13 and from the nomenclature of Haines (1987) for the Bunyeroo Gorge area.

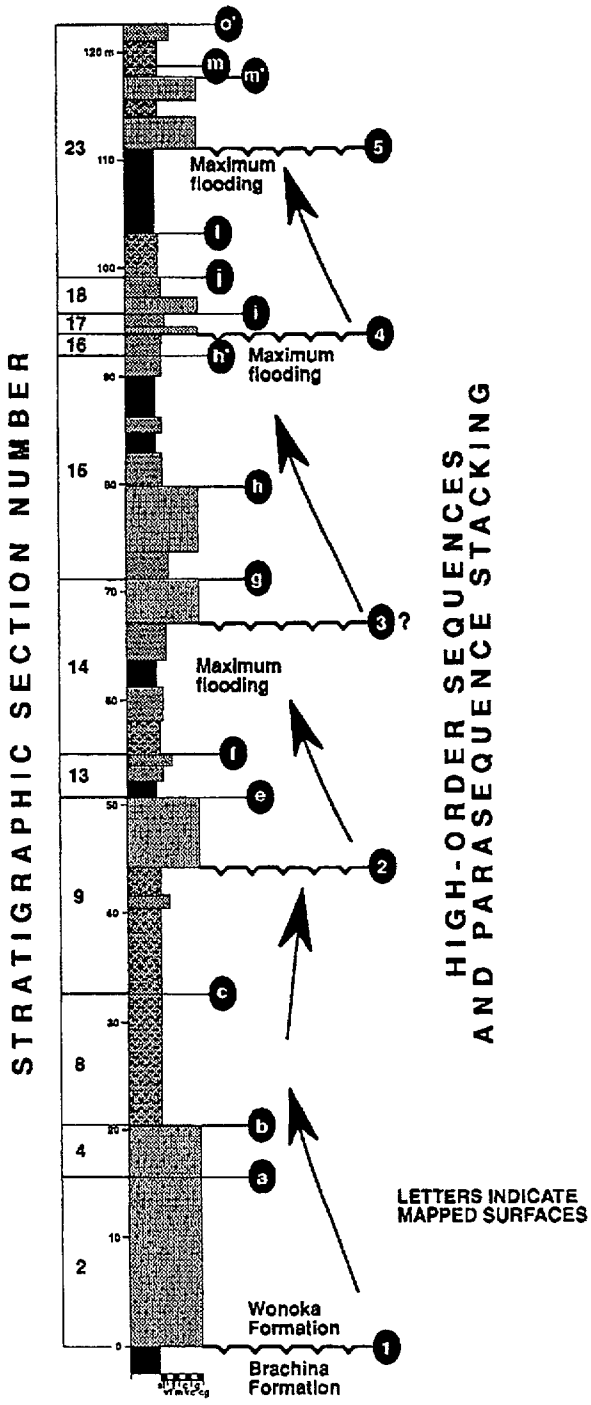


Figure 15. A composite of 24 measured sections along the northern edge of the central incision shown in Figure 13, indicating the presence of five high-order sequences in an interval of about 120 m. Grain size is indicated by column width, from siltstone (sl) to conglomerate (cg). Details in the upper part of this section are shown in Figure 16. Thicknesses are strongly influenced by differential compaction and hence proximity to the canyon wall. Nevertheless, there appears to be a systematic arrangement of facies and parasequence stacking: 1) Individual sequences are strongly asymmetrical with deepest water facies being preferentially located in the upper part. 2) The degree of asymmetry increases upwards from one sequence to another, and the facies discontinuity at corresponding sequence boundaries becomes increasingly pronounced. 3) Overall, the succession fines and deepens upward.

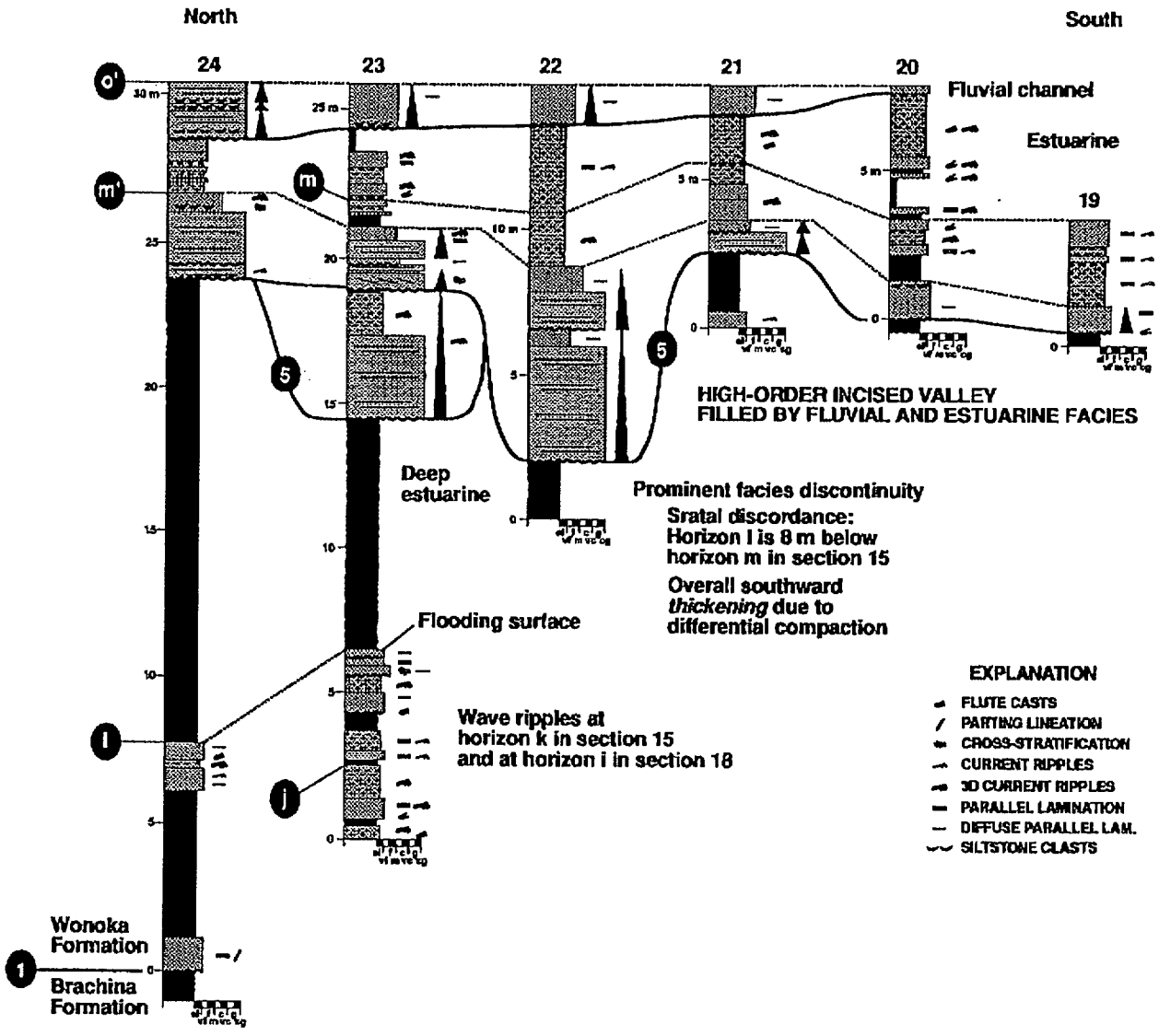


Figure 16. Details of lateral relations in the upper part of the section shown in Figure 15 over a distance of several hundred metres. Note the pronounced facies discontinuity at the erosion surface labelled 5, and the systematic arrangement of facies (both vertically and laterally). Triangles indicate grain-size trends. Letters indicate mapped horizons. Individual time-stratigraphic units tend to thicken southward away from the main canyon wall owing to differential compaction. Thinning in the same direction of the interval between horizons l and m is interpreted as due to overall tilting of the depositional surface towards the north. Sequence boundary 5 passes southward into a cryptic surface that is concordant with bedding in estuarine sandstones and siltstones. Incised valleys at this boundary are oriented in the direction 104° (flow towards 284°), approximately parallel to the main canyon at this locality (084°). Examples of shallow-water indicators are present in the sequence between boundaries 4 and 5.

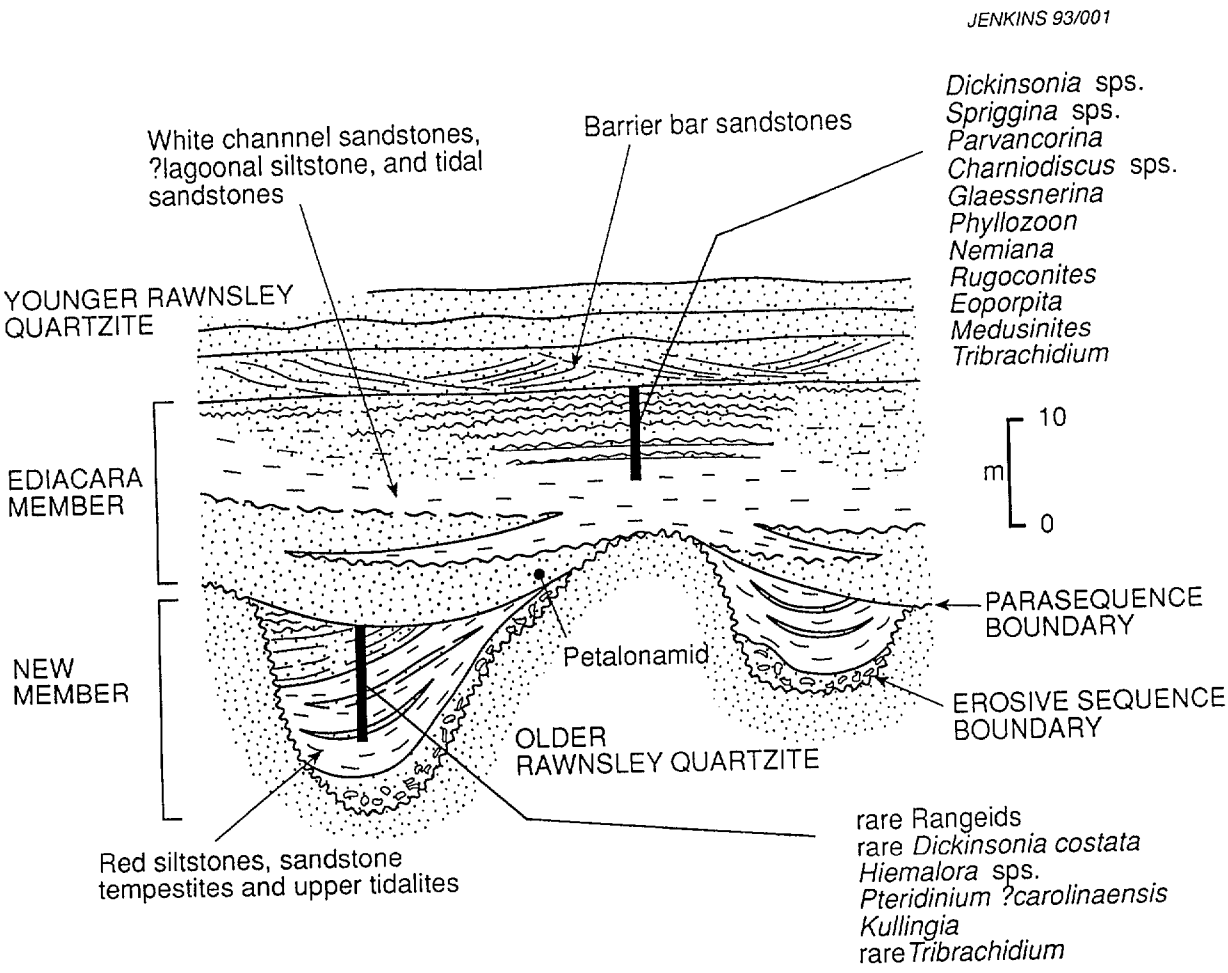


Figure 17. Generalised stratigraphy of the older Rawnsley Quartzite in the western Flinders Ranges, showing the relationship between newly recognised older member and the Ediacara Member, and the placement of biotas.

strata showing spectacular, festoon cross-bedding. It was probably deposited in a shallow shoreface setting. The origin of the dish-shaped structures is unclear, but movement of tidal waters through the sands as a result of alternate flooding and emergence likely played a part.

Stop 6.2 -- Fossiliferous Interval

The fossiliferous intervals comprise two transgressive, upward-shoaling depositional cycles, each overlying channelled downcut surfaces, with the second cycle truncating various parts of the first (Figs. 17 and 19). Locally, these erosive surfaces resemble angular unconformities, but on a wider scale the discordance is only of the order of a few degrees. The two cycles include different lithologies and different fossil assemblages.

The upper cycle mainly comprises lithologies formed by clean, well winnowed, shoreface sands; it contains the characteristic elements of the local Ediacara assemblage. This cycle is readily identified as an expression of the Ediacara Member, which is defined at a type section in Ediacara Range (see Jenkins and others 1983).

The older cycle is associated with channels of the order of 10-20 m deep and which were apparently eroded by fast currents that deposited mud-clast rich sands asymmetrically over the bottom and one wall of several of the larger cuts. Overlying, laminated red-purple siltstones include increasing numbers of thin flaggy sandstone bed upsection. These flags are brownish or greenish coloured internally and relatively impure. Infrequent casts of tool marks at the base of flags are orientated in various directions. Top surfaces show complex 'rib- and furrow' markings indicating interfering oscillatory currents. These characteristics conform closely to those which Grey and Benton (1982) and Seilacher (1982) indicate as diagnostic for storm beds or tempestites. Cleaner sandstones with abundant mudcracks predominate towards the top of the cycle. Impurities in the sandstones cause them to weather to a distinctive dark red-brown colour with abundant black manganiferous stainings.

Notwithstanding that the sandstone flags were deposited at times of energetic disturbance as density currents, they are fossiliferous, including common discoidal remains, tentaculate specimens of the anemone-like *Hiemalora* Fedonkin, 1982, rare indeterminate frondose forms and very rare *Dickinsonia costata* Sprigg 1947, cf. *Rangea* Gürich, 1929 and *Tribrachidium* Glaessner 1958. An unusual trace fossil on the upper surfaces of the intertidal beds forms a radiating branched pattern of burrows resembling *Chondrites* von Sternberg, 1993. It is proposed that this older fossiliferous

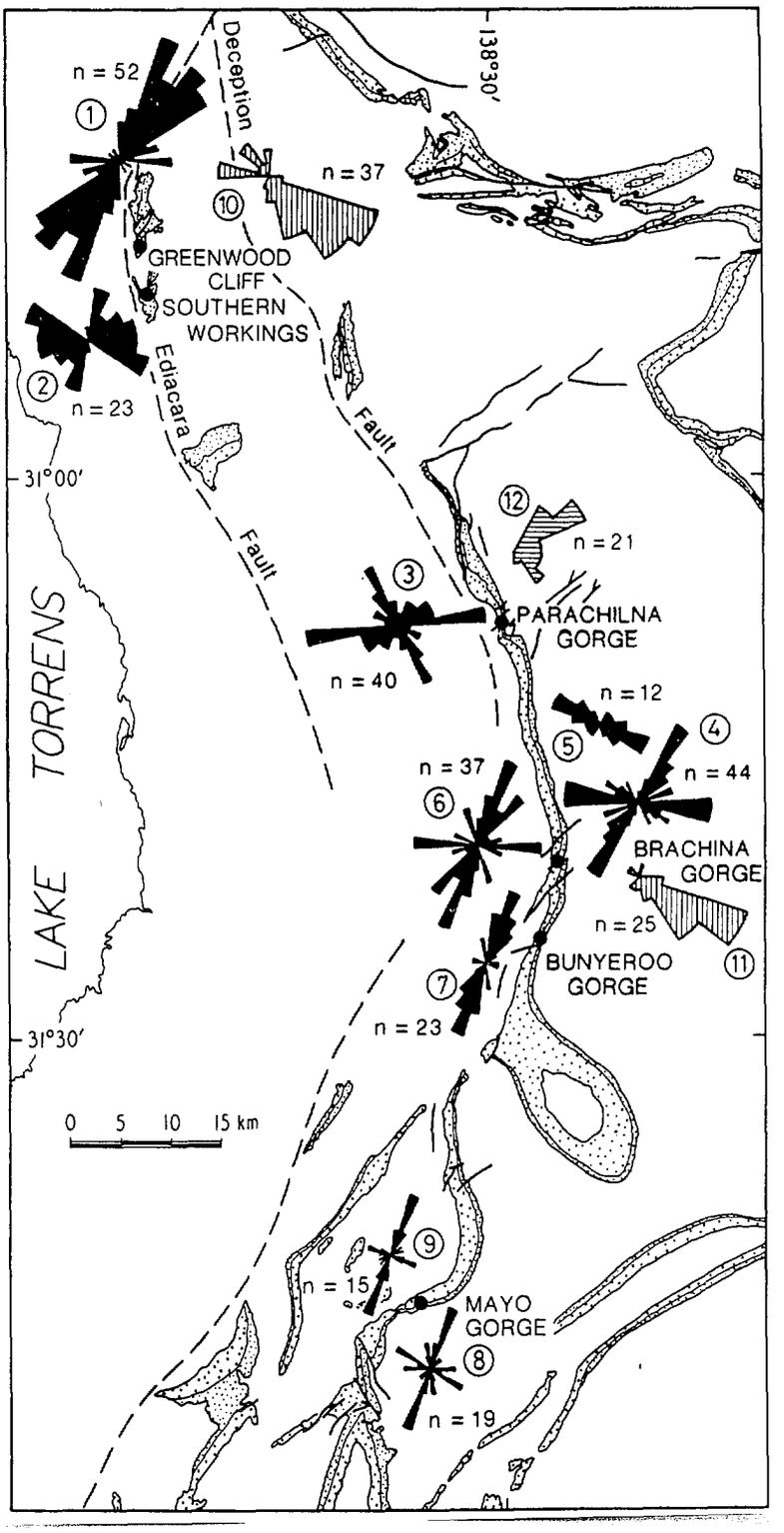


Figure 18. Ripple crest orientations in the Ediacara Member with some papaleocurrent data from the Ediacara Member and the Rawnsley Quartzite (after Jenkins and others, 1983).

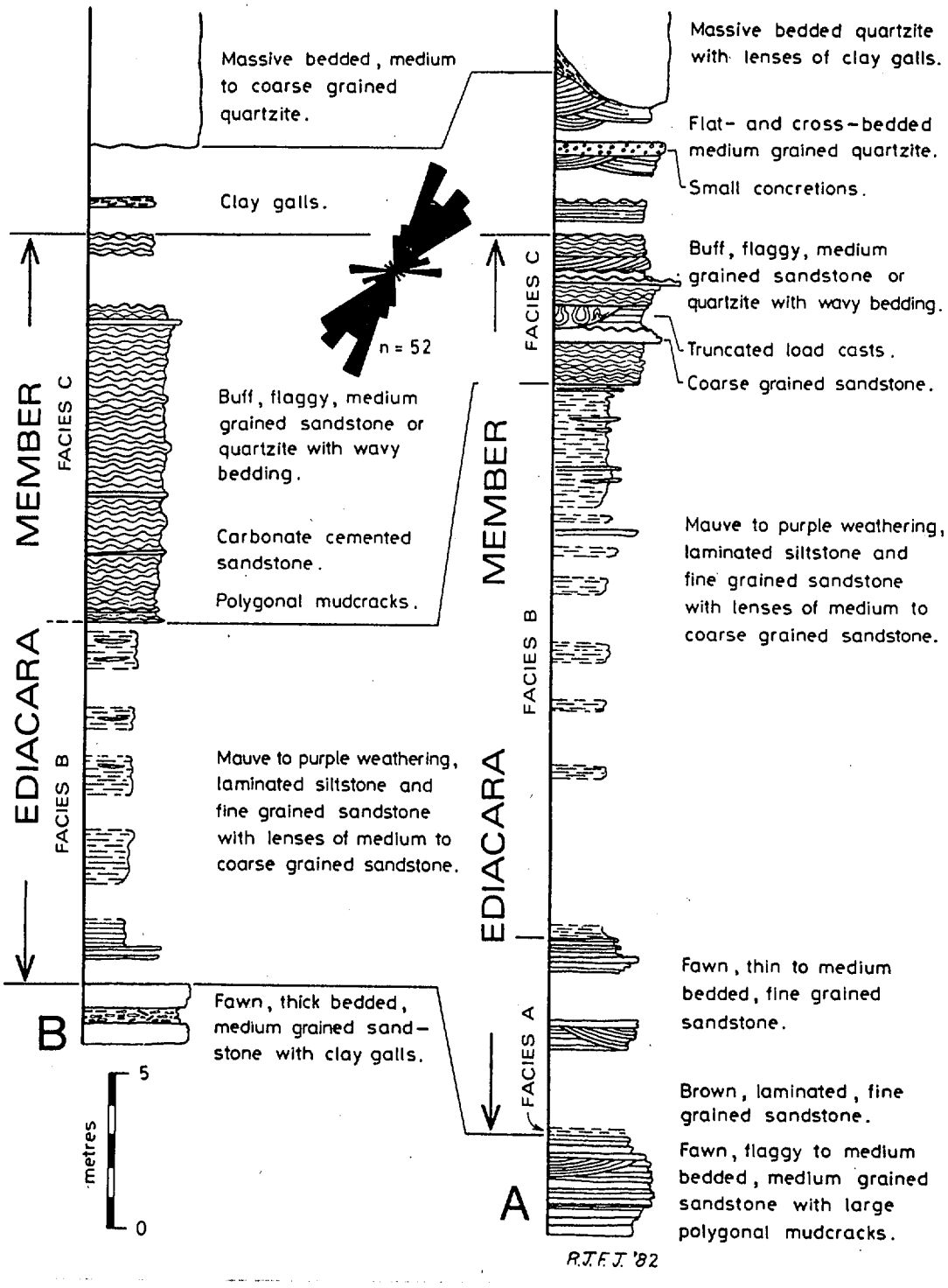


Figure 19. Ediacara Member at Greenwood Cliff, Ediacara Range.

interval is formally recognised as a new member of the Rawnsley Quartzite.

Considerable confusion has come about in the literature concerning the environment of deposition of the Ediacara Member because various workers (Gehling, 1987, 1988; Mount, 1989, 1991) extended this unit well below the mail fossil beds in a way not visualised by Jenkins and others (1983). It seems likely that this downward extension may overlap the newly recognised lower member and J.G. Gehling (pers. comm. Gehling, 1991) has also located *Rangea*-like specimens at lower fossiliferous levels in the Chance Range and in the Heysen Range.

Stop 6.3 -- Ediacara Member

The fossiliferous interval in the Rawnsley Quartzite described and measured by Goldring and Curnow (1967) at Greenwood Cliff, 550m south of Gap Creek, Ediacara Range (Fig. 19) is the designated type exposure of the Ediacara Member (Jenkins, 1981). The rapidly eroding gorges in the western flank of the Flinders Ranges contain the least altered and most complete surface exposures of the beds comprising the Member.

At least five distinctive lithofacies characterise the Ediacara Member at its type exposure, and a similar development of three lithofacies is present at the study site:

1. *Facies A. Siltstone and fine-grained sandstone.*
   Unbedded or thin-bedded and laminated, micaceous siltstone and fine-grained quartz arenites, which in places contain thin, discontinuous lenses or thin beds of medium- to coarse-grained quartz arenite; they commonly show wavy bedding (e.g. Reineck and Wunderlich, 1968) and micro-cross lamination. The rocks are unusually pale green, buff, maroon or red-brown, the coloration evidently reflecting different oxidation states of iron in Fe-rich minerals. Partings produced by silty laminae, or changes in grain size in successive laminae, are rare. The only fossils so far found in this facies are trails referable to Form B (Glaessner, 1969) and rare, indistinct medusoids. The facies may reach thicknesses of 100 m or more in erosion valleys in the central Flinders Ranges.

2. *Facies B. Heterolithic facies of laminated siltstones and interbedded orthoquartzite flags.*

   This facies comprises purple to mauve-weathering laminated or very thin-bedded, micaceous siltstones or fine-grained, quartz arenites and thin, flaggy lenses and interbeds of medium- to coarse-grained quartz arenite. The rocks show fine-grained

partings and are commonly fossiliferous; fossils are particularly numerous on the base of the sandstone flags. The facies may reach 17.5 m in thickness (Ediacara Range).

3. *Facies C. Flaggy, wavy bedded orthoquartzite.*

This comprises light coloured, flaggy, wavy bedded, fine- to medium-grained orthoquartzite with perfect partings or thin, silty interbeds. The top surfaces of the flags typically show wave oscillation and interference-ripple forms. Numerous, flat clay galls occur within the flags, especially near top surfaces. Sinuous mudcracks, irregular or polygonal mudcracks, flat-topped ripples, ladder ripples, and carbonate cemented beds occur within restricted intervals. This facies is the most fossiliferous of the Ediacara Member. The maximum known thickness is 19 m (Ediacara Range).

Mid parts of the single main cycle of the Ediacara Member at the study site include several thin quartzite beds which preserve abundant remains of *Dickinsonia costata* and were probably deposited as a result of local slumping. Upper wave-rippled beds are prolifically fossiliferous. The cycle is topped by a narrow interval (several metres) of unfossiliferous, mudcracked upper-intertidal sands and succeeding low-angle cross-bedded beach or barrier deposits (e.g. Jenkins and others 1983; Gehling, 1987; Mount, 1989).

At other sites vertical disposition of facies in the Ediacara member is generally complex. Coarsening-upward cycles, involving the facies sequence A, C, D or the sequence B, C, D overlain by unfossiliferous (flat-stratified and trough-cross-bedded) sands, are commonly developed. Facies E tends to be interposed at various levels and may be repeated as many as five or six times within the fossil beds (e.g. at Parachilna Gorge).

The fossils of the Ediacara assemblage are chiefly confined to the Ediacara Member. They are found mainly as casts or moulds on the lower surfaces of quartzite flags (Wade, 1968). Very rare remains of *Pteridiunium* Gürich, the only known specimen of *Chondroplon bilobatum* Wade, and a single specimen of *Charniodiscus arboreus* (Glaessner) have been collected from an unusual bed of massive quartzite that crops out below the Member in the Ediacara Range (Goldring and Curnow, 1967, p. 201; Wade, 1968, p. 256 and 1971, pp. 183-4). The preservation of these specimens is atypical, in that they occur as positive and negative counterpart moulds.

Environmental interpretation of Ediacara Member
(R.J.F. Jenkins, University of Adelaide, South Australia)

Facies A: The laminar bedding of the silts or fine sands composing this facies suggests deposition in either quiet water, or suspension sedimentation at times of waning current. Starved ripples attest traction-load transport of a minimal sand supply. Unbedded siltstones constituting parts of this facies contain scattered sand grains up to 1.5 mm in diameter. Modern tidal muds commonly contain similar scattered grains that have been attached to algae or organic matter and held in suspension prior to deposition in association with mud aggregates (Wunderlich, 1969). Facies A shows no evidence of subaerial exposure, and wave influence seems to have been minimal.

Where facies A is overlain by thick sand beds, it is almost invariably distorted by load casts, which commonly reach large dimensions. At Bunyeroo Gorge large pods of sand with vertical dimensions of up to 4 m have subsided into the upper part of a unit of facies A. Kuenen (1958, 1965) showed experimentally that such load casts and pseudonodules form by *in situ* foundering of sand bodies into underlying, water-logged, plastic sediments. The examples in the Ediacara Member are syndepositional, because erosional surfaces truncate the top of the structures. Wade (1970) termed these pseudonodules slump rolls, although she later realised that no lateral motion was involved in their formation (M. Wade, pers. comm.).

In terms of its developed thickness, and its common fine lamination or thin bedding, this facies might be considered to be of neritic, shelf aspect. In the Hawker, Wilpena area, where there is good outcrop control transverse to the main Flinders basin, thick developments of facies A occur within trough-like structures (Gehling, 1982, 1983, 1987).

Facies A tends to be coarser grained with increasing sand content towards the medial part of the Ediacara Member, the environment possibly becoming lagoonal. Silts floor the deeper parts of many modern lagoons behind coastal barrier islands (Elliott, 1978).

Facies B: The laminar siltstones and fine-grained sandstones within this rock-type are not greatly dissimilar to those of facies A. However, the numerous lenses and interbeds of medium- to coarse-grained sandstone document a much larger, current-transported traction load. Different parts of the rock show rhythmically laminated bedding, wavy bedding, or lenticular bedding with connected thick lenses. Wave- and current-ripple forms occur on the top surfaces of the sand beds, as well as interference ripples with strong and weaker ripples superimposed almost at right angles.

Such superimposed or ladder ripples may be due either to the interaction of wave and current action at an angle (Reineck and Singh, 1973) or to changes in wave or current directions during late-stage ebb outflow (Klein, 1977).

Facies C: The sandstones composing this facies show both wavy bedding and well developed silty partings. Some sandstone beds also show internal lenses and films of silt and clay that form flaser bedding. The sand grainsizes in individual beds usually fine upwards, but reverse grading, suggestive of deposition by currents of increasing strength, is also common. Ripple marks occurring on the upper surfaces of beds have been studied previously by Goldring and Curnow (1967). The ripples are wave and current-formed kinds, generally similar to those associated with the sand beds of facies B. Most are symmetrical in form, show inclined cross-laminae, and have a thin layer of sand either fully draped over them or draped over their stoss sides. In a few places, these drapes are multiple or show reversed micro-cross-lamination. The drapes commonly enclose flat clay galls and coarse sand grains, which evidently represent a larger deposit from a winnowing current. The stoss side, sand drapes and reversed ripple foresets indicate reversals of currents, the secondary current having eroded parts of the previously formed ripple. The clay or silt film that commonly blankets the entire ripple form was probably deposited from suspension. A closely comparable set of processes was documented by Reineck and Wunderlich (1969) from studies of modern intertidal deposition. Clay galls are a common characteristic of such environments, especially in tidal channels and on the higher parts of tidal flats.

Cracked mud layers in facies C commonly seem to have been sandwiched between sand beds that have slumped somewhat, causing the enclosed mud film to crinkle, distort and partially overlap its cracked edges; the filling of the cracks is continuous with the sandstone below the cracked layer. Other crack systems apparently were formed in surficial mud films, because crack-separated pieces of the film have been dislocated to form clasts. Some polygonal and approximately polygonal crack systems seem to have been filled from above, and the edges of the cracks are curved slightly upwards. It seems likely that these formed subaerially due to desiccation.

Ripple orientation and current data

Indications of current directions during the time of deposition of the Ediacara Member are limited chiefly to the internal lamination of ripple sets and cross-lamination of megaripples.

Most sets of measurements suggest that the trends of the ripple crests tend to follow two main direction at a high angle to each

other. A major ripple trend is NNE-SSW to NNW-SSE, in most instances broadly parallel to the strike of the outcrop at the sites measured. For facies B and C at Ediacara Range and Brachina Gorge, symmetrical ripples with wavelengths of 10 to 15 cm show cross-lamination, suggesting ESE to SE-trending currents; by analogy with the tidal flats of Sylt, this persistent trend may reflect the latter stages of ebb.

An E-W crest trend (dominant at the Southern Workings, Ediacara Range and Parachilna Gorge) is frequently indicated by undulatory and linguoid current ripples and perhaps parts of ladder sets. Currents trending S to SE are inferred, rarely northward. It is tempting to suppose that these water movements tended to be longitudinal to the coast and associated with the early stages of ebb, or the late phase of flood.

The parallelism between dominant ripple-crest trends and strike generally observed in the Ediacara member suggests a correspondence between the depositional configuration of the basin and tectonic trends, at least for the western Flinders Ranges. The major ripple trends parallel inferred significant faults (Fig. 18) which are probably of ancient origin and comprise part of the major tectonic element bounding the western side of the Flinders trough, the Torrens Hinge Zone (Rutland and others, 1981). The significance of such deep-seated faults in controlling Adelaidean sedimentation has long been recognised (e.g. Coats, 1962, 1965).

Body Fossils

Numerous casts and imprints of soft-bodied animals of the Ediacara assemblage occur on basal surfaces of sandstone beds in facies B, C and D. Their various modes of preservation have been described by Wade (1968) and she considered that the position in which most Ediacara 'medusoids' are preserved indicates that the beds containing them were 'continuously under water when the animals now fossilised were deposited and buried'.

Wade (1972) suggested that some members of the plexus of forms referred to *'Cyclomedusa'* Sprigg may actually have been attached to the substrate at the apex of their conical exumbrella surface as juveniles, and maintained a similar attitude as free-living benthonic adults. New information indicates that *Ediacaria* Sprigg almost certainly lived attached to the substrate by a short stalk and that the various taxa included within *Cyclomedusa* are likely preservational morphs resulting from differing emphasis of particular characteristics due to taphonomic processes at the time of burial. It now seems probable that the majority of specimens of *Ediacaria* and *Cyclomedusa* are preserved in life position and mainly occur as casts of certain parts of the aboral surface. The stalks

may have tended to become filled with sand prior to death, and this fill is probably broken away from the remainder of the casts in most instances.

Wade (1972a) observed that 'medusiform' remains may show evidence of physical battering, with sharp, marginal re-entrants leading into deep creases reminiscent of the deep radial tears into the mesogloea commonly displayed by stranded modern medusoids. This does not in itself document stranding but is consistent with wave damage in an energetic environment subject to period emergence. Jenkins and others (1983) maintained ideas of Sprigg (1947) and Glaessner (1961) that fossilisation of the soft-bodied fauna was principally linked to diurnal movement of sands during tidal deposition. Extending some of the notions of Goldring and Curnow (1967), and Wade's (1968) conclusion that the fossils represent remains interred by sands laid down in some depth of water, Gehling (1982, 1983, 1987, 1988, 1991) considers that deposition occurred in an outer shelf setting below fair-weather wave base, with burial of the organisms by storm-surge sands (cf. Mount, 1989, 1991). A. Seilacher (pers. comm., 1985) also adheres to the idea of storm-related sedimentation, but holds that the presence of ubiquitous wave oscillation and interference ripples in the fossiliferous beds is characteristic of a shoreface. Ideas of storm deposition are opposite to Wade's (1968) conclusion that fossilisation occurred in regions of lessened hydrodynamic energy in the lee of rising diapirs that impeded the predominant wave fetch.

Investigations disclose large polygonal desiccation cracks in highly fossiliferous parts of the section at Ediacara Range and in facies D of Jenkins and others (1983) in part of the Heysen Range. This seems to potentially substantiate the suggestion of deposition in tidal regimes subject to emergence, and the intense concentric crinkling of some of the remains resulted from drying. Probable pelagic forms such as *Rugoconites* Glaessner and Wade, 1966, and *Eoporpita* Wade, 1972, may have been stranded at ebb tide.

Groupings of like organisms in the fossil assemblages may indicate preservation of virtually intact segments of the community, and delicate epichnial trace fossils are common. Hence, fossilisation was initiated by burial of organisms stuck down or clinging to the gluey surface of drying algal films. On the other hand, the more complete benthic communities that are represented must indeed have inhabited subtidal settings (cf. Nedin and Jenkins, 1991).

Stop 6.4 -- Parachilna Formation Stratotype.

Near the western end of Parachilna Gorge a sharp disconformable contact (Dalgarno, 1962, 1964; Dalgarno and Johnson, 1965) separates fine-grained feldspathic Pound Quartzite from the rounded

quartzose basal sands and shales of the stratotype Parachilna Formation. Clasts to 20 cm in length and presumably of Pound Quartzite are visible on the contact. The trace fossil *Phycodes pedum* occurs in the basal few metres (Daily, 1973) and is succeeded by numerous bioturbated sandstones with vertical U-shaped burrows of *Diplocraterion* Torell. The Scandinavian ichnogenus *Plagiogmus* reported by Daily, Twidale and Alley (1969) from higher levels of the formation in nearby Wilpena Pound remains unlocated in the type section. Shale, sandstone, and oolitic and oncolitic limestones occur higher. There is a passage into the shallow-water carbonates of the Woodendinna Dolomite.

The Elatina Formation, Oraparinna area.

(N.M. Lemon, National Centre for Petroleum Geology and Geophysics)

The Elatina Formation is well represented by two sections east of Brachina Gorge (Fig. 20). Accurate locations are given by Australian Map Grid co-ordinates bracketed against locations mentioned below.

Stop 6.5 -- Trezona Camp Site

The section along Etina Creek near the "Trezona Camp Site" turnoff (74860 mE/30400 mN) is that shown as the type section for the Parachilna 1:250,000 map sheet and was measured as Section 19 of the accompanying diagram (Fig. 20). The base of the Elatina Formation at this locality is a pink, "granule train" sandstone in contact with a karst surface developed on the underlying limestones of the Trezona Formation. The irregular nature of the surface, black, botryoidal Mn (Fe) concretions and possible mud-filled caverns identify the karst surface.

One kilometre to the north in Enorama Creek, the section is very similar to Section 19 but here the top contact of the Elatina Formation is well exposed in clean outcrop (74800 mE/31050 mN). Recent creek erosion has scoured a surface on gently dipping, pebbly dropstone diamictite. This facies was subject to current activity at the time of deposition which flushed the finer sediments to leave gravel lags, dominated by basalt, dolerite and dark, cemented sandstone clasts. Rocks equivalent to these can be found outcropping in the nearby Enorama Diapir, which was active at the time of deposition.

The upper contact is gradational over 15 cm between red, ripple cross-laminated sandstones of the Elatina Formation and buff-weathering dolomites of the Nuccaleena Formation. Several deep water tepees, with no associated intraformational conglomerates

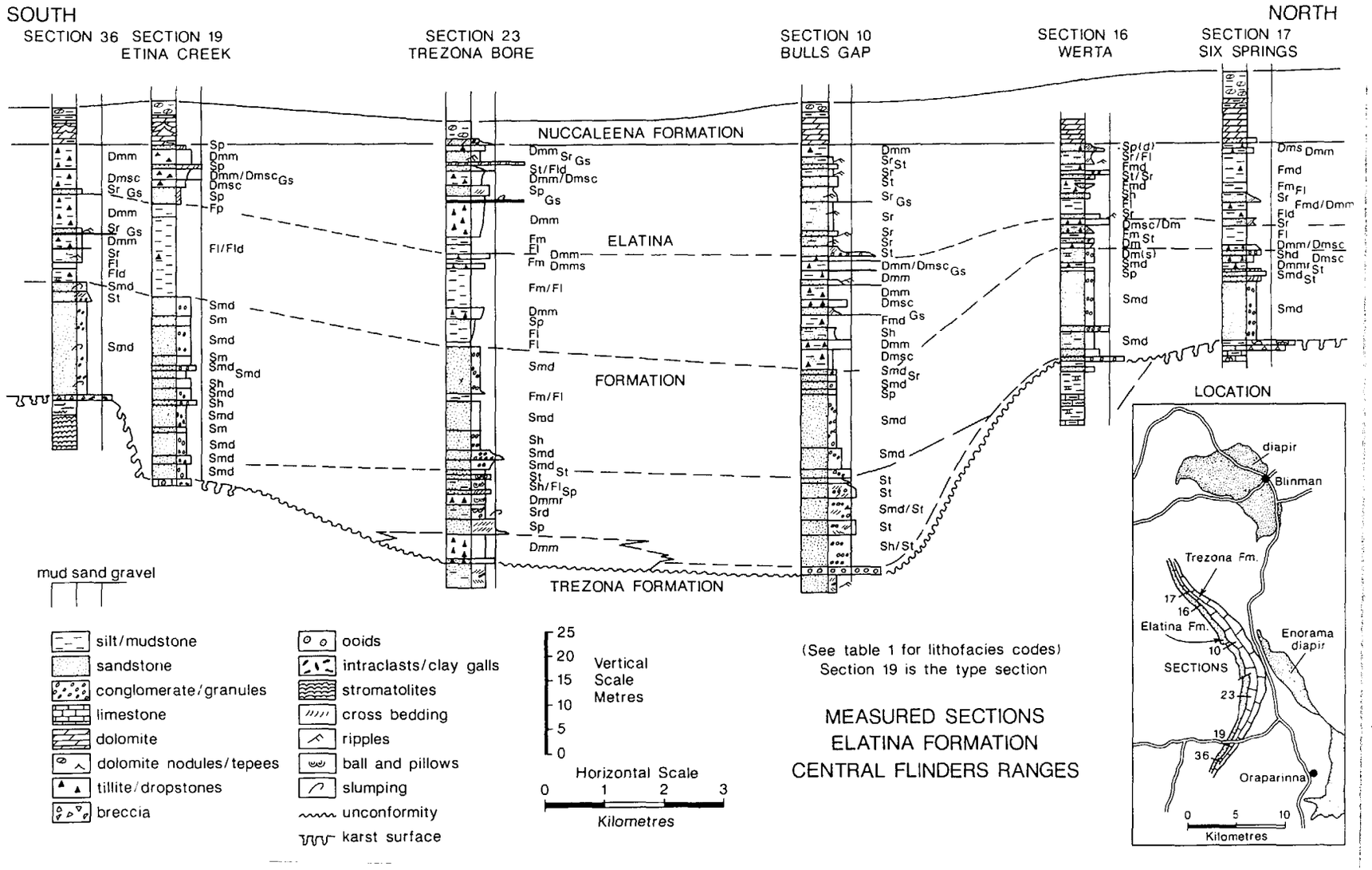


Figure 20. Correlated sections of the Elatina Formation north from the type section.

but evidence of syn-depositional growth, occur in the dolomites. The dolomite becomes siltier up section, eventually grading through a red nodular bed to a red shale. The dolomite appears to have been deposited during rising sea level following de-glaciation.

Stop 6.6 -- Bulls Gap

The section at Bulls Gap (75070 mE/40750 mN) shows the complete Elatina Formation in fair to good exposure and was measured as Section 10 on the accompanying diagram (Fig. 20). Additional units are present in the base of the Elatina Formation and in the top of the Trezona Formation. There is an additional 120 m of section present below the Elatina base when compared to Section 19, 11 km to the south (Fig. 20). Mapping shows progressive truncation of beds south of section 23 along an unconformity between the Elatina and Trezona formations. Much of the additional section is a coarsening-upwards sequence of tidal flat siltstones and channel sandstones.

The contact between the formations, in the area close to the Enorama Diapir where maximum preservation of the sequence occurs, is occasionally marked by large (to 2 m) exotic clasts "bulldozed" into the unlithified sediments of the upper Trezona Formation. Above these is a thin but more extensive conglomerate composed of a 50:50 mixture of reworked Trezona limestone clasts and exotic clasts, probably derived from the Gawler Craton to the west. This is in turn overlain by coarse, yellow sandstone with occasional dropstones. The sequence above this is similar to that near Trezona Camp Site with pink, slumped, "granule train" sandstones overlain by red diamictites, with common dropstones. These are usually of basalt, dolerite or dark, cemented sandstones, some showing clear faceting and striae. Current reworking of the diamictite becomes more obvious up section with gravel lags intercalated with the diamictites and ripple cross-laminated sandstones. The rippled sandstones dominate the upper section of the formation beneath the massive, buff-weathering, pink dolomite of the Nuccaleena Formation. Detail of the upper contact is obscured by scree blocks of the dolomite.

Onset of glaciation forced the regression culminating in the erosion of sequence from the Trezona Formation. Rare large boulders and small pockets of ice-push tillite are the only sediments deposited during peak glaciation. The majority of the Elatina Formation was deposited during waning glaciation.

For background notes see Appendix G.

Day 7 Monday, May 17th -- Brachina Gorge and travel to Alice Springs

Brachina Gorge to view facies different from those in Bunyeroo Gorge. Drive to Port Augusta; board the train for Alice Springs. Overnight on train.

ICS TERMINAL PROTEROZOIC AND IGCP PROJECT 320 FIELD PROGRAM

11 - 21 MAY, 1993

PART 2

THE AMADEUS BASIN

Introduction

(J.F. Lindsay AGSO, Canberra)

A number of broad, shallow, intracratonic depressions formed on the Australian craton during the late Proterozoic and early Palaeozoic (Fig. 1). The Amadeus Basin is typical of these intracratonic basins and contains a varied late Proterozoic to mid-Palaeozoic succession (Fig. 21). The basin's lithostratigraphy is summarised by Wells and others (1970). More recent detailed sequence studies can be found in Lindsay (1987a, 1989), Lindsay and Korsch (1989, 1991), Kennard and Lindsay (1991), Gorter (1991) and Lindsay and others (1993). The current understanding of the sequence stratigraphy of the Neoproterozoic succession is summarised in Figure 22.

Location

The Amadeus basin lies at the centre of the Australian continental block to the south of Alice Springs (Fig. 21). From east to west, along its longest axis, the basin extends for 800 km from the edge of the Simpson Desert in the east to 150 km past the border of Western Australia in the west. From its inception in the Neoproterozoic to its ultimate demise in the mid-Palaeozoic the basin maintained a tenuous contact with the major oceanic water masses.

Morphology

The basin can be subdivided into several distinct morphological features (Fig. 21). Major sub-basins along the northern margin of the basin (Ooraminna, Carmichael and Idirriki Sub-basins) are connected by a shallow trough or saddle (Missionary Plain Trough), and are separated from a much larger platform area to the south and west by a low ridge (the Central Ridge). At times this ridge acted as a barrier to sedimentation. The sub-basins contain up to 14 km of late Proterozoic to mid-Palaeozoic rocks, whereas the platform areas typically have less than 5 km of sedimentary section.

Evolution

The basin evolved in three major stages Lindsay and Korsch, 1989, 1991). Stage 1 began at about 800 Ma with extensional thinning of the crust and possibly with the formation of half-grabens. Thermal subsidence was followed by a long post-thermal phase of minimal subsidence during which sedimentation was probably controlled by small-scale tectonic events (Lindsay and Korsch, 1991; Shaw, 1991). Subsequently, a major compressional event (the latest Neoproterozoic Petermann Ranges Orogeny) affected the southern

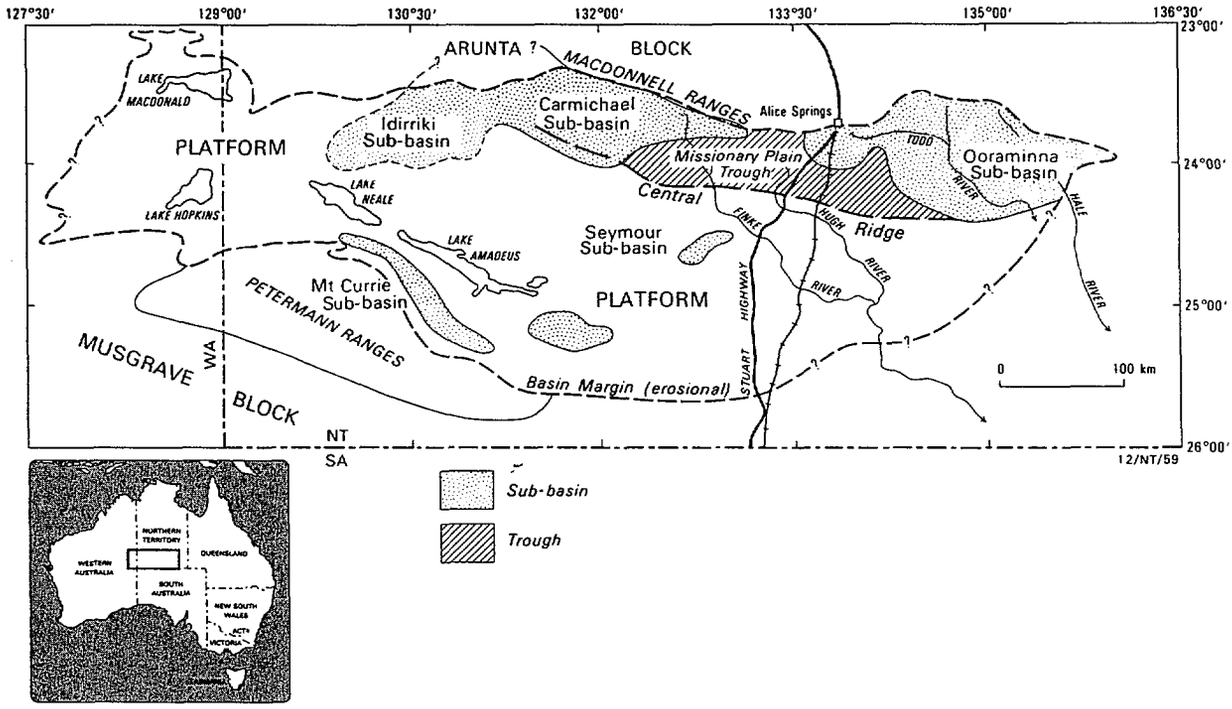


Figure 21. The Amadeus Basin and its main features (sub-basins, trough, platforms, and Lindsay & Korsch, 1989, 1991).

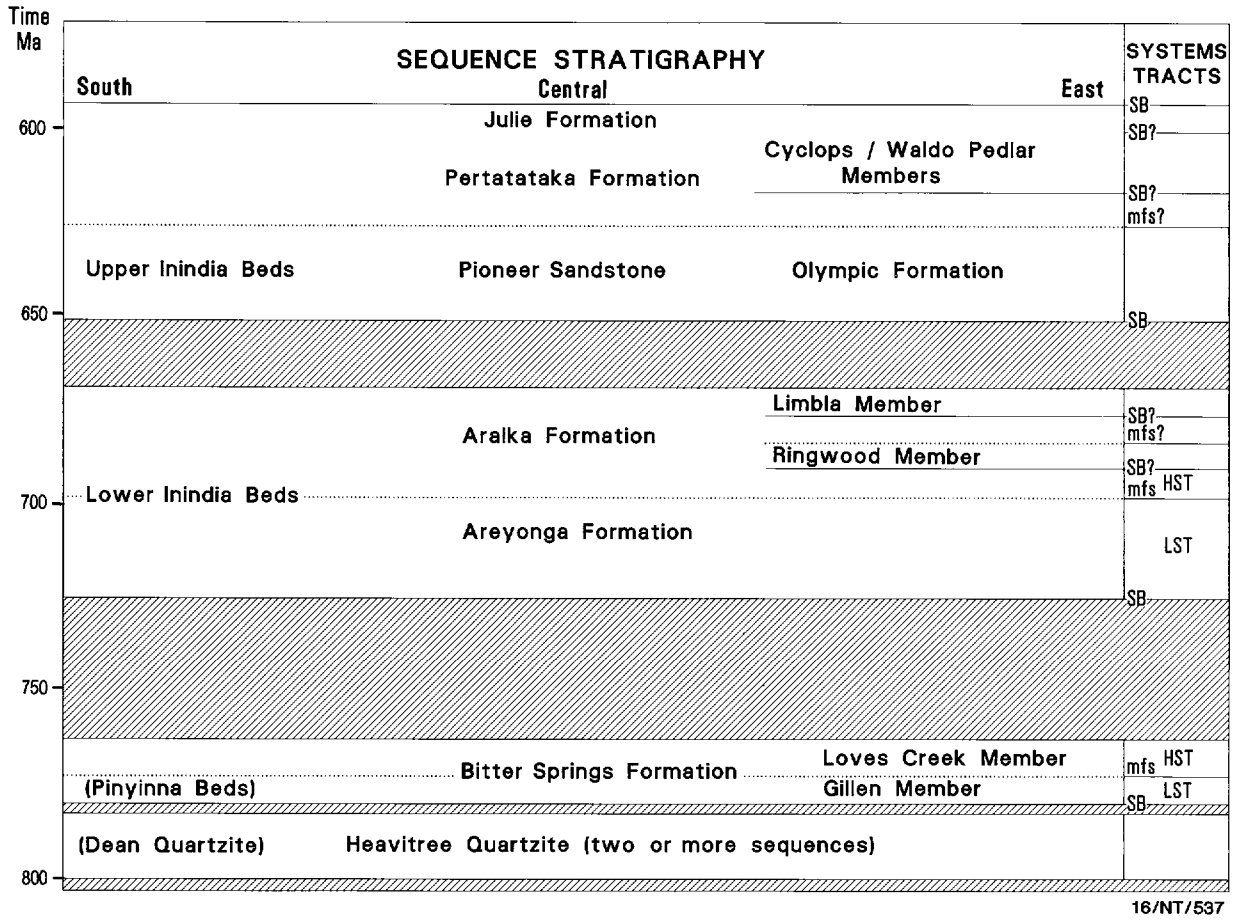


Figure 22. The Neoproterozoic stratigraphy of the Amadeus Basin of central Australia showing major sequence boundaries (after Lindsay, in press). HST = highstand systems tract, LST = lowstand systems tract, SB = sequence boundary, mfs = maximum flooding surface.

margin of the basin, and a less intense phase of crustal extension (Stage 2) occurred in the northern portion of the basin at about 600 Ma. Stage 2 subsidence was followed by a major compressional event beginning at about 450 Ma (Stage 3) in which major southward-directed thrust sheets caused progressive downward flexing of the northern margin of the basin, and sediment was shed from the thrust sheets into a foreland basin (the Late Devonian - Early Carboniferous Alice Springs Orogeny) (Jones, 1972; Korsch and Lindsay, 1989). This event shortened the basin by 50-100 km and effectively terminated sedimentation at approximately 360 Ma.

Sequence Recognition in Intracratonic Settings
(John F. Lindsay, AGSO, Canberra)

The techniques used in sequence interpretation have for the most part been developed on passive margin settings in relatively young sedimentary successions. Here large prograding sequences can be readily discerned so that sequence geometry, the distribution of stratal terminations, sequence boundaries and systems tracts can readily be identified.

However, intracratonic basins present a very different setting. Seismic data gathered in these basins appears almost featureless and suggests layercake stratigraphy such that conventional passive margin seismic models are difficult to apply. However, in spite of these differences the lessons learnt on passive margins are generally applicable to intracratonic settings. The various systems tracts are present within intracratonic sequences, but their geometry (e.g. relative proportions and overall dimensions) are different. The key to using sequence stratigraphy in an intracratonic setting is in defining the geometry of sequences and the factors controlling them (Lindsay, in press; Lindsay and others, 1993).

In an intracratonic setting sequences are mostly thin and poorly differentiated compared to their passive-margin equivalents. Slow subsidence rates, low depositional slopes and shallow water depths collectively result in diminished sediment accommodation within these basins. The sequences are extensive and thin, few have recognisable progradational geometries, and erosional unconformities generally have minimal topographic relief. During relative sealevel lowstands, little or no accommodation space may be available for sediment accumulation, and lowstand deposits may be poorly developed and areally restricted. Thus intracratonic successions commonly comprise stacked transgressive-highstand deposits separated by near planar unconformities or paraconformities; that is, flooding surfaces commonly coincide with sequence boundaries.

Basin dynamics allowed a full range of expression of sequence geometries during Neoproterozoic-Cambrian time in the Amadeus Basin (Lindsay, in press) (Fig. 22). Thus seismic interpretation could be augmented by well-log and outcrop facies analysis.

Rationale and Objectives

(M.R. Walter, Macquarie University, Sydney)

We regard the Officer and Amadeus Basins as sources of very significant data which supplements knowledge gained from the Adelaide Geosyncline. They have the following advantages over the Adelaide Geosyncline:

• Extensive grids of seismic records in both basins permit detailed sequence analysis without reliance only on outcrop characteristics;

• Very low grades of thermal alteration have allowed the exquisite and frequent preservation of rich assemblages of acritarchs;

• The same low grades of thermal alteration have allowed the application of chemostratigraphy with fewer of the complications that arise in the Adelaide Geosyncline.

The Amadeus Basin has good outcrop of the relevant successions, whereas outcrop in the Officer Basin is very poor. For that reason we regard the Amadeus Basin as the main source of supplementary data for characterising Neoproterozoic stratigraphy in Australia.

All three basins are linked by a web of correlations using lithostratigraphy, sequence analysis, chemostratigraphy, and biostratigraphy. In addition, the Acraman ejecta layer occurs in both the Adelaide Geosyncline and the Officer Basin.

Our objectives in including the Amadeus Basin in the excursion are to:

• demonstrate the relevant correlations, including those that are under dispute;

• show you key seismic and well data;

• demonstrate marker beds that have proven to be regionally significant;

• show you the units that have yielded a rich acritarch and isotopic record;

• contrast the two regions.

Like the Adelaide Geosyncline, the Amadeus Basin covers a huge area, so it is impossible on this excursion to visit more than a few critically significant sections.

The excursion will focus on the succession from the base of the upper glacial units (Olympic Formation and Pioneer Sandstone), through the Pertatataka and Julie Formations, to the base of the Cambrian (within the Arumbera Sandstone) (Fig. 23).

Major sequence boundaries are recognised at the base of the glacial units, at or near the base of the "upper marker cap dolomite", and at the base and in the middle of the Arumbera Sandstone (Fig. 22) (Lindsay, 1987a, Lindsay, in press).

FIELD GUIDE

(Malcolm R. Walter, Richard J.F. Jenkins and John F. Lindsay)

Day 8 Tuesday, May 18 -- Alice Springs

Stop 8.1 -- Northern Territory Geological Survey

Briefing at the Northern Territory Geological Survey. Outline of program (Fig. 23), overview of basinal geology, discussion of sequence analysis, biostratigraphy and palaeontology. Examination of seismic sections and well correlation charts.

Stop 8.2 -- Wallara 1

(Clive. R. Calver)

Examination of the core of drill hole Wallara 1 (Fig. 24). This well spudded in Cambro-Ordovician sediments and was fully cored from 350m to its total depth of 2001 m in the Neoproterozoic Bitter Springs Formation. Of particular interest here, it intersected the Arumbera Sandstone, Pertatataka Formation, Pioneer Sandstone, Aralka and Areyonga Formations.

The Aralka Formation abruptly overlies sandy diamictite of the (Sturtian) Areyonga Formation. The Aralka here is relatively thin, and consists of very thinly laminated black shale with abundant early diagenetic nodules and layers of dolomite. The unit is closely comparable to the "lower marker cap dolomite" overlying Sturtian glacials in the Adelaide Geosyncline.

The Pioneer Sandstone is also attenuated relative to elsewhere in the basin. It consists of two units. The lower unit is a medium- to coarse-grained pebbly quartz sandstone, 2m thick, of possibly fluvial or fluvioglacial origin, that is a lithostratigraphic equivalent of the probably Marinoan, glacigene Olympic Formation of the northeastern Amadeus Basin (Preiss and others, 1978). The bottom contact of the sandstone in the drillcore is load-caste, and the underlying Aralka shales were apparently unconsolidated at the time of sandstone deposition. The upper unit of the Pioneer Formation is the "upper marker cap dolomite" (Preiss and others, 1978), here consisting of several metres of pale grey dolomicrite with dark grey shale interbeds. An abrupt contact at the base of the dolomite probably represents a sequence boundary, since regionally there is an abrupt facies change, and locally an angular unconformity, at this level (Freeman and others, 1991). The cap dolomite grades upward into the Pertatataka Formation.

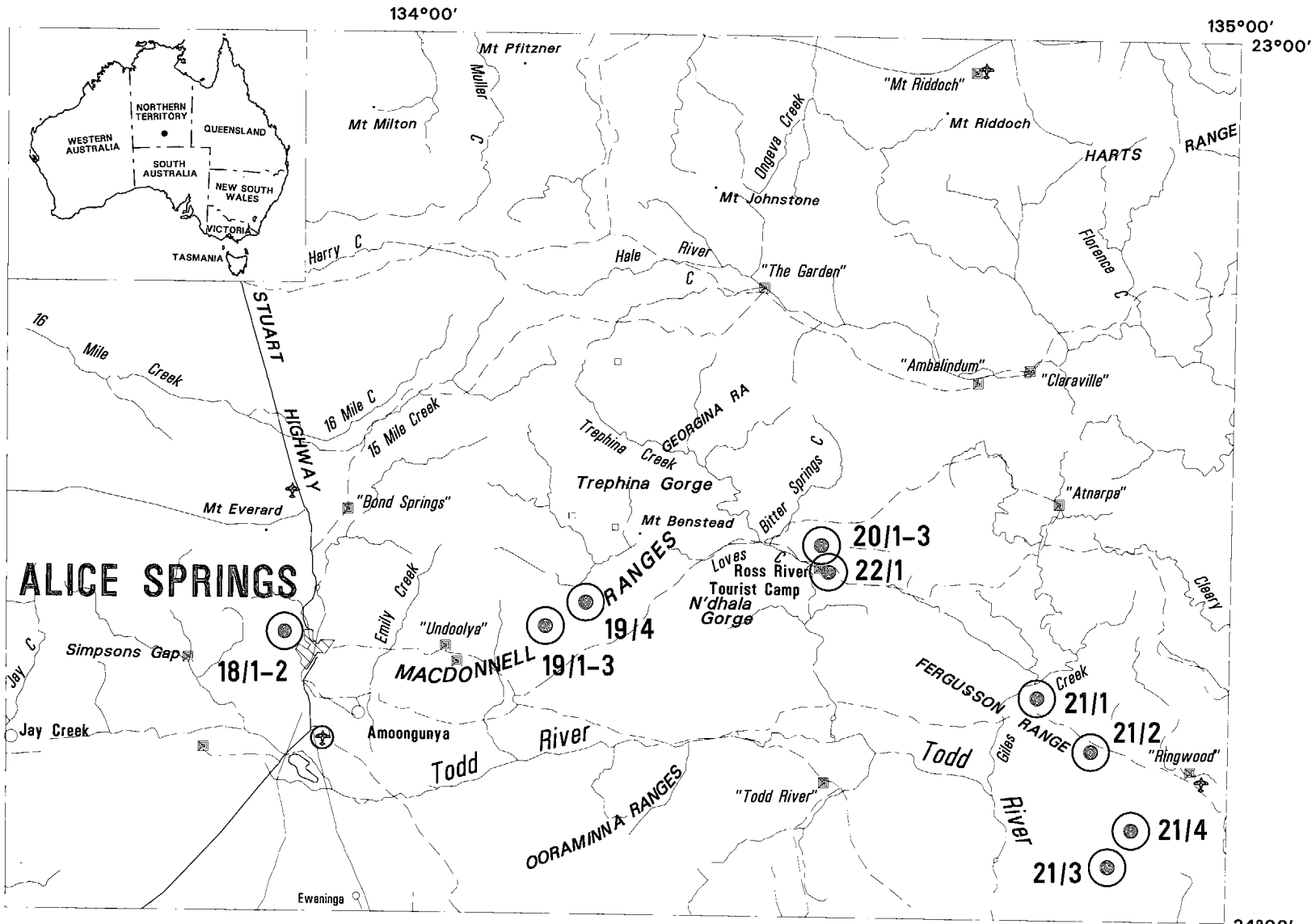
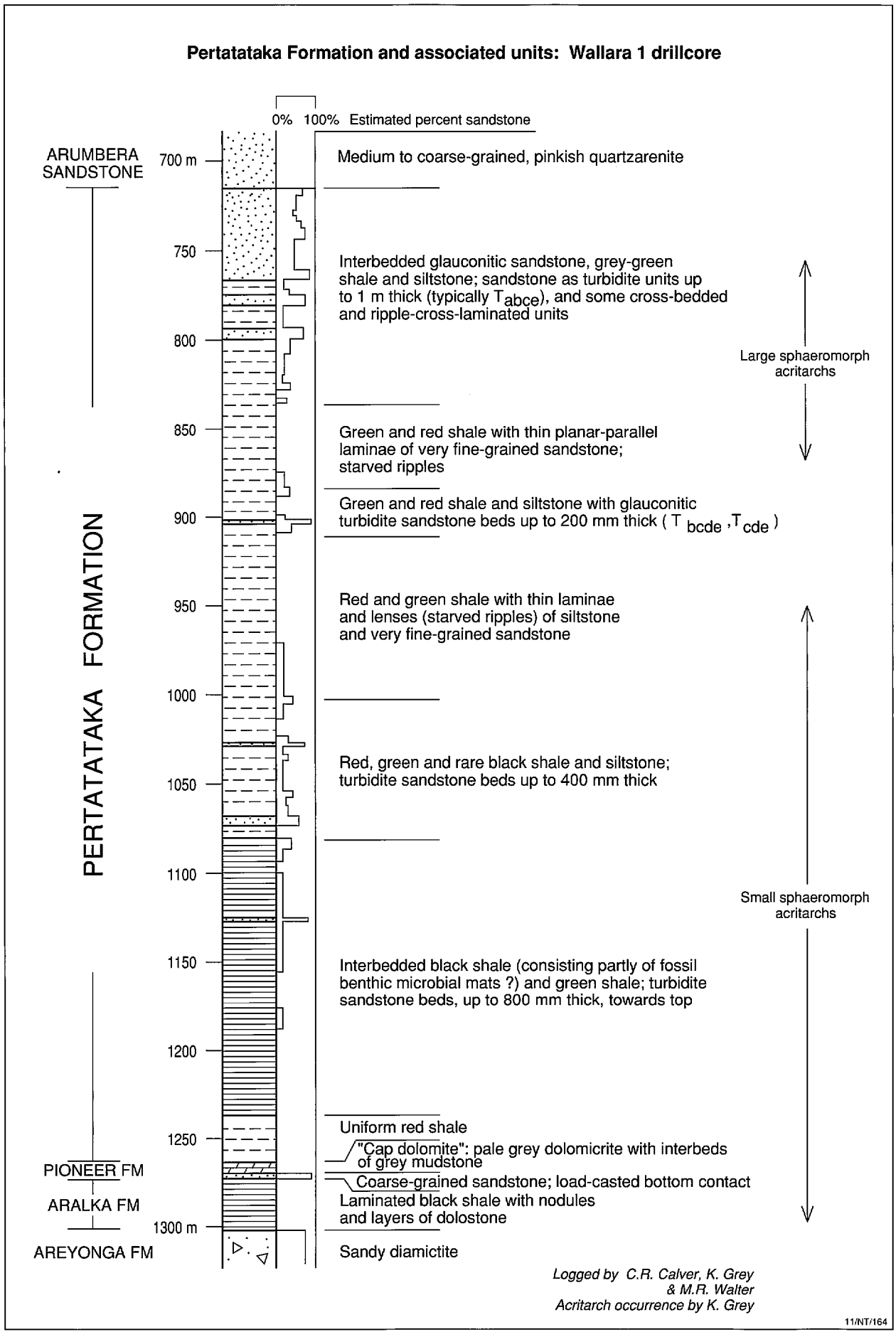


Figure 23. Locality map for the Alice Springs Area. Daily stops are shown a date/number.

Figure 24. Wallera 1 core showing the Pertatataka Formation and associated units.



Chemostratigraphic (Calver, Appendix E) and palynological (Grey, Appendix F) correlations suggest that the upper half of the Pertatataka Formation is missing in Wallara 1, as is the Julie Formation. The Arumbera Sandstone here rests abruptly on Pertatataka Formation. The drillhole is sited close to the lateral facies transition between relatively distal hemipelagic shales and turbidites in the northern part of the basin (the Pertatataka Formation), and more proximal, sandy inner-shelf sedimentary rocks of the southern part of the basin (the Winnall beds). Near the base of the Pertatataka Formation in Wallara 1, uniform red shale passes up into a 150m thick unit of interbedded black and grey-green shale and minor turbiditic sandstone. Abundant, very thin, weakly anastomosing carbonaceous microlaminae in the black shale bands are probably fossil benthic microbial mats (cf. Schieber, 1986). In accord with this interpretation is the distinctly light isotopic composition of organic carbon in the black shales relative to the interbedded green shales. This unit may represent, in sequence-stratigraphic terms, an interval of sediment starvation at the base of a highstand systems tract consisting of the broadly shallowing-up succession of the remainder of the Pertatataka Formation in Wallara 1. Overlying the black shale interval is a thick succession of predominantly grey-green and reddish-grey shales, with siltstone and fine-grained sandstone occurring as thin planar laminae, starved ripples or thin cross-laminated beds; and as thicker beds with partial Bouma sequences. Deposition by turbidity currents (dominantly low-density), and hemipelagic processes is inferred. Depositional environment was probably never very deep (<200 m?) considering the vertical and lateral facies transitions into shallow-water sediments and the intracratonic basin setting. In the uppermost 100m or so of the Pertatataka Formation in Wallara 1, sandstone becomes predominant (see Fig. 24); more complete, coarser-grained Bouma sequences are seen; and high in this interval, cross-bedding and ripple cross-lamination with wavy and flaser bedding suggest deposition within wave base. The contact with the overlying Arumbera Sandstone is abrupt and erosional.

In the Amadeus Basin Wallara 1 and Rodinga 4 drill core and several field sections have been sampled for acritarchs and isotope chemostratigrahy (see appendices E and F). Wallara 1 drillhole contains intervals of well-preserved acritarchs and abundant amorphous tissue. Most acritarchs are simple leiospheres, but some samples contain small, spiny forms. No large spiny forms have been observed, and this supports Calver's interpretation, based on isotope geochemistry, that there is no overlap between Rodinga 4 and Wallara (see Fig. 24). No major breaks in sedimentation are shown by the chemostratigraphic profiles of carbon isotopic composition for the Pertatataka Formation.

Day 9 Wednesday, May 19

Stop 9.1 - Basal Pertatataka Formation

At this locality (Fig. 23) the contact between the stromatolite marker bed ("upper marker cap dolomite", top of the Pioneer Sandstone) and the Bitter Springs Formation is well exposed. In places there is some angular discordance, but the contact is predominantly conformable. There is a weakly developed regolith at the top of the carbonates and calcareous siltstones of the Bitter Springs Formation, expressed as calcrete veins and wisps within the carbonates. There is a patchily developed pebble conglomerate (Pioneer Sandstone) overlain by 2-15 cm of chert with small columnar stromatolites (?*Elleria minuta*).

The cherty ministromatolites of the "upper marker cap dolomite" are interpreted as occurring in a condensed interval marking the maximum flooding surface following the glaciation.

The stromatolites at this locality were collected by Preston Cloud and identified by Cloud and Semikhatov (1969) as *Anabaria juvensis*, which they thought came from the Bitter Springs Formation. The same sample was studied by Vanyo and Awramik (1982), and a sinuous pattern in the orientation of the axes of the columns was interpreted as being heliotropic, i.e, as representing the seasonal variations in the orientation of the sun. From a rhythmicity in the lamination in the stromatolite columns they estimated the number of days in the year (at least 410) about 850 Ma ago (as they accepted that the specimen came from the Bitter Springs Formation). You may wish to assess the regularities or otherwise in the orientations of the columns as seen in outcrop.

Stop 9.2 -- Battery Flat

Here the Pioneer Sandstone and the regolith at the top of the Bitter Springs Formation are well exposed, and there are small outcrops of the "upper marker cap dolomite". This locality has been mapped in detail by Richard Jenkins and Malcolm Walter, and BMR diamond drilling supervised by Peter Southgate (BMR Alice Springs 23) spudded in the Pertatataka Formation and produced continuous core down into the Bitter Springs Formation (Fig. 25).

The contact between the Pioneer Sandstone and the Bitter Springs Formation is an angular unconformity with variable degrees of discordance. Calcrete wisps, veins and pisolites occur in the upper few metres of the Bitter Springs Formation. In what are inferred to be contemporaneous valleys there are breccias with angular clasts of Bitter Springs Formation lithologies, and clasts of what are

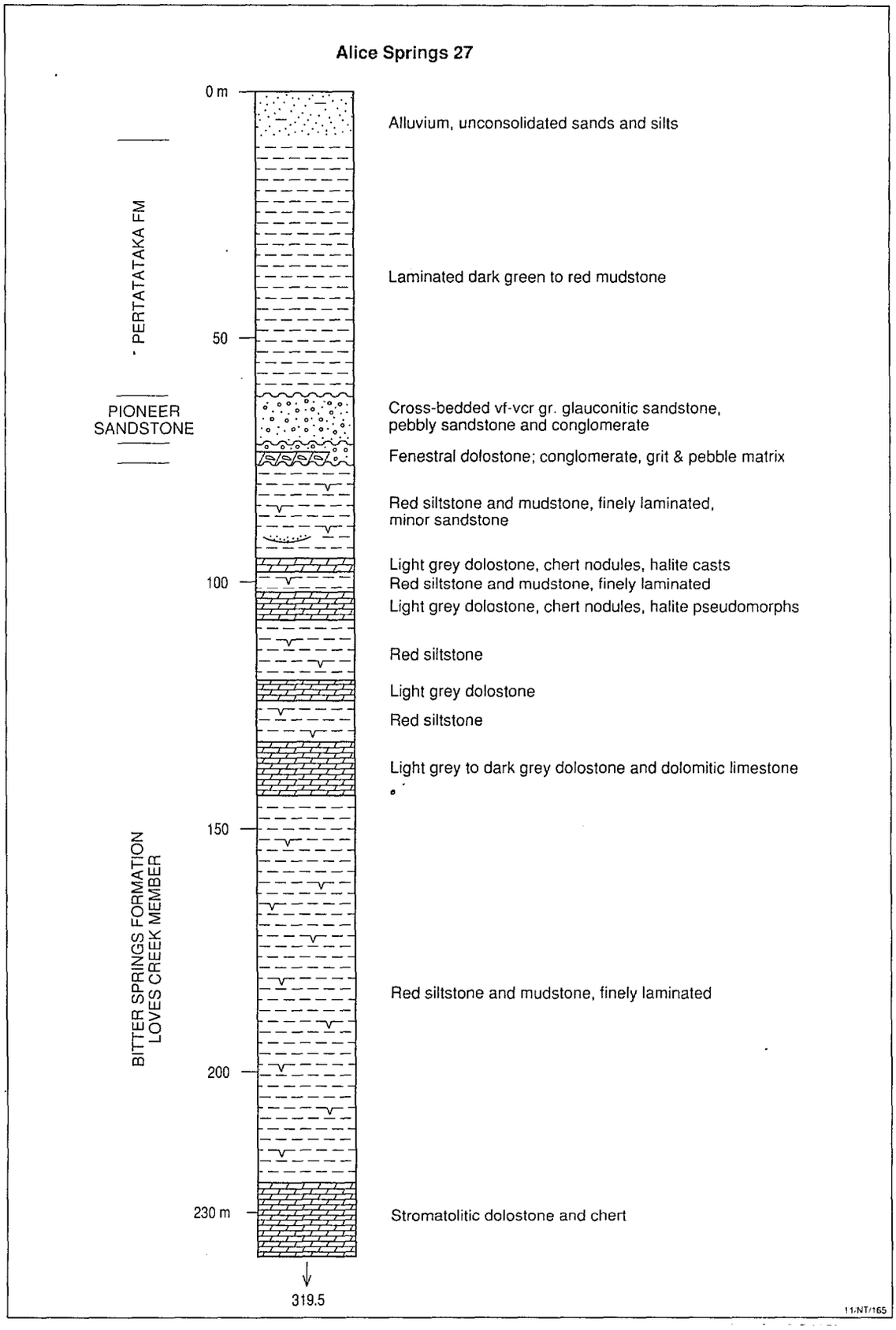
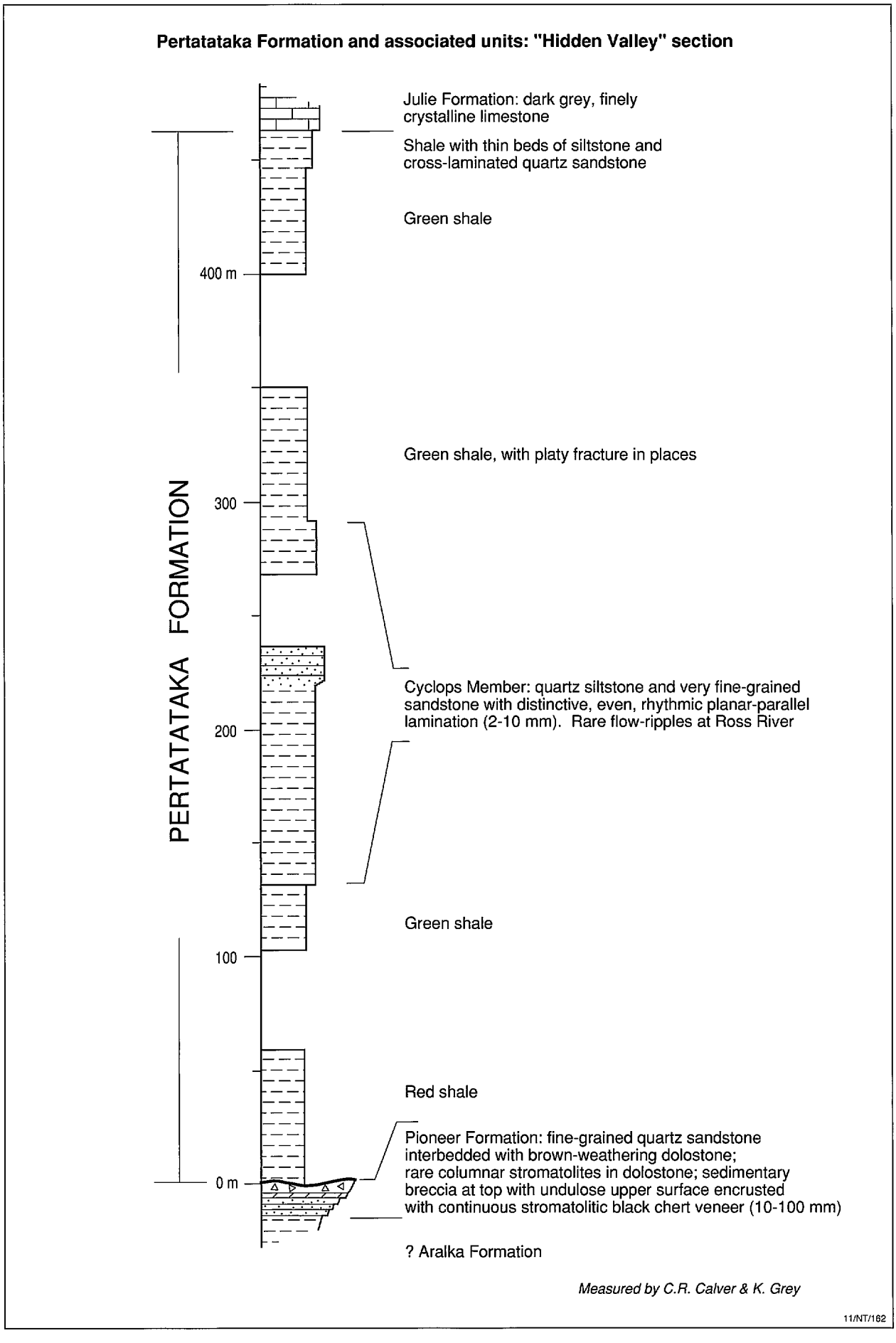


Figure 25. Alice Springs 27 at Battery Flat, showing the Loves Creek Member of the Bitter Springs Formation , the Pioneer Sandstone and the base of the Pertatataka Formation (Southgate, pers. comm.) (Stop 9.2).

Figure 26. Hidden Valley section of the Pertatataka Formation and associated units (Stop 9.3)



considered to be spring travertine. This unit is regarded as a talus breccia. Above the breccia in the valleys, and directly on the Bitter Springs Formation elsewhere, there is 10m of quartz pebble conglomerate and pebbly sandstone of the Pioneer Sandstone. This is overlain abruptly by some 10 cm of red stromatolitic chert like that at Stop 9.1. About a kilometre to the south, brown stromatolitic dolomite occurs at the same stratigraphic level.

Stop 9.3 -- Hidden Valley

At Hidden Valley (Fig. 26), the Pioneer Formation consists of 5-10 m of coarsening-upward, well-bedded quartz sandstone and minor brown-weathering sandy dolostone. Rare, small columnar stromatolites are seen in the dolostone. Capping the sandstone unit is an unusual sedimentary breccia consisting of angular, ragged fragments of white, fine-grained quartz sandstone in a brown-weathering sandy dolostone matrix. The upper contact of the breccia has a complex, mounded topography with up to several metres of relief. The mounds are elongate and in an en echelon pattern. Encrusting the mounds is a continuous, 10-100 mm thick layer of black stromatolitic chert, with closely-spaced small columnar stromatolites (*?Elleria minuta*), like that at stop 9.2. Uniform red shale of the Pertatataka Formation abruptly overlies the chert veneer. Poorly exposed, green and red, uniform silty shale comprises the bulk of the Pertatataka Formation, except for the central 150m or so that comprises the Cyclops Member. This unit forms low strike ridges and consists of quartz siltstone and very fine-grained sandstone with an even, platy fracture that reflects a regular, 2-10 mm planar-parallel lamination. The top and base of the Cyclops Member are gradational over several tens of metres. At Ross River, inferred to have been nearer the northern margin of the basin, the unit is somewhat coarser and ripple-marked bedding planes are seen. A northerly provenance, and deposition in relatively shallow water, are inferred (Korsch, 1986b).

Stop 9.4 -- Cyclops Member

This unit is well exposed at this locality and will be examined briefly to stimulate discussion on its interpretation. Preiss and others (1979) regarded it as a correlative of the ABC Quartzite of the Adelaide Geosyncline and the Grant Bluff Formation of the Georgina Basin. Jenkins (submitted) considers that it is younger than the ABC. The Grant Bluff Formation and the immediately overlying Elkera Formation contain what Walter and others (1986) interpret as the oldest trace fossils known in Australia (see also Jenkins and others, 1992); these are simple sinuous burrows (*Planolites* and *Palaeophycus*). Despite numerous searches, no trace fossils have been found in the Cyclops Member. However, Lindsay

(1991) fround trace sedimentary structures in the much older Heavitree Quartzite which he tentatively suggested to be burrows.

Day 10 Thursday, 20 May

Stop 10.1 -- Bitter Springs Formation

A brief stop will be made to examine the locality from which Elso Barghoorn collected black chert samples that formed a major part of the collection used by Bill Schopf in his study of microfossils of the Bitter Springs Formation. Subsequent research by Peter Southgate has indicated that this part of the Bitter Springs Formation is lacustrine.

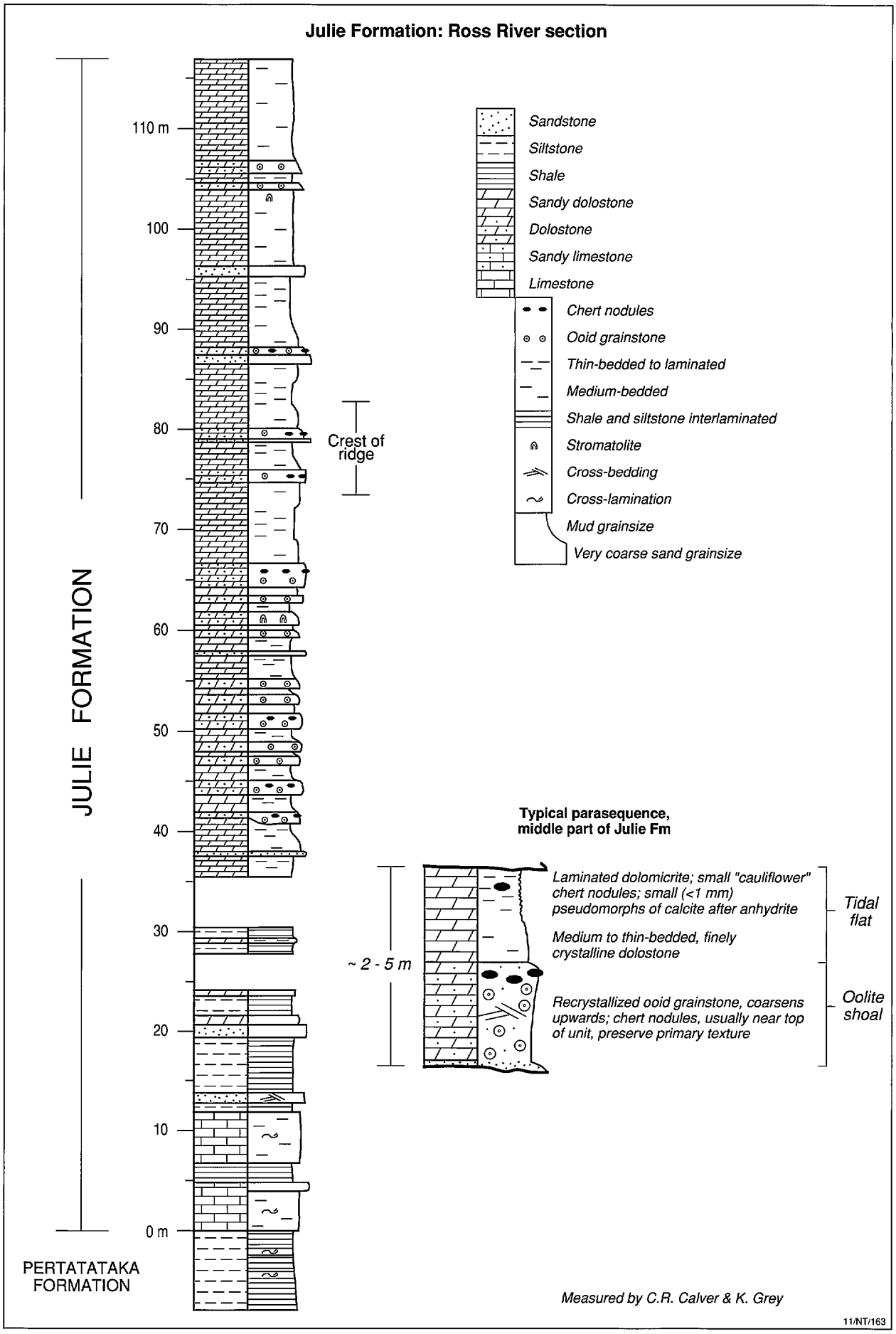
Stop 10.2 -- Pioneer Sandstone

At this locality the unit is a conglomerate. This stop is to demonstrate the eastward coarsening of the formation. The contact with the Bitter Springs Formation is exposed 500 m to the west. The contact with the Pertatataka Formation is not exposed.

Stop 10.3 -- Julie Formation

The Julie Formation (Fig. 27) is well-exposed on the ridge immediately south of the Ross River homestead. There is an apparently abrupt lower contact upon Pertatataka Formation. The lowest 12m of Julie Formation consists of dark grey, sandy and silty, well-bedded limestone, in places with low-angle, polymodal (swaley?) cross-stratification. Then follows a recessive unit (12 - 36 m on illustrated section) of siltstone, cross-bedded coarse-grained quartz sandstone and minor dolostone with small 'cauliflower' chert nodules (probable evaporite replacements). The rest of the formation (36-117 m) consists essentially of alternating sandy, recrystallised ooid dolograinstone and fine-grained dolostone; with minor coarse-grained sandstone. Particularly in the 40-70 m interval, these lithologies are arranged in shallowing-upward cycles or parasequences (see Fig. 27)). The erosional bases of parasequences are typically overlain by a thin (<100 mm) layer of very coarse-grained quartz sandstone. Then follows 1 - 2 m of recrystallised dolograinstone, in places with cross-bedding, that tends to coarsen upward. The original oolite fabric is well-preserved within large (100 - 200 mm) ovoid chert nodules that tend to occur toward the tops of the grainstone units. The upper part of each parasequence consists of medium to thin-bedded, fine-grained pale grey dolostone and dolomicrite, with common small 'cauliflower ' chert nodules and small (<1 mm) calcite pseudomorphs after anhydrite. In places, tepee structures and poorly-preserved

Figure 27. Measured section through the Julie Formation at Ross River Homestead (Stop 10.3).



domical stromatolites are present. Each parasequence probably represents a tidal flat complex prograding over an oolite shoal.

Whole-rock carbon and oxygen stable-isotope compositions of dolostones average +4.5 ‰ and -8 ‰ respectively; of limestones, +4.5 ‰ ($\partial^{13}C$) and -13 ‰ ($\partial^{18}O$). The upper parts of some parasequences display marked departures towards lower $\partial^{13}C$ and higher $\partial^{18}O$, signifying meteoric (exposure-surface) alteration (cf. Allan and Matthews, 1982).

Ringwood Area

Introduction

(M.R. Walter, Macquarie University, Sydney)

Two major questions impede the fuller understanding of the later Neoproterozoic stratigraphy of the northeastern Amadeus Basin,

a) elucidation of the relationships of four recognised arenaceous formations seemingly close in age to or part of the second (upper) glacigenic event: the Olympic Formation (Wells and others, 1967; Preiss and others, 1978), Pioneer Sandstone (Preiss and others, 1978), Gaylad Sandstone (Freeman and others, 1991) and Waldo Pedlar Member of the Pertatataka Formation (Wells and others, 1967); Freeman and others, 1991); and

b) an understanding of the wider correlatives of the Pertatataka Formation outside of the Amadeus Basin.

Recent work of M.R. Walter and R.J.F. Jenkins tracing the lateral extent of the Olympic Formation and Pioneer Sandstone suggests the two are direct equivalents, with the Pioneer Sandstone becoming more pebble rich and thence passing into cobble conglomerates towards the east, and the Olympic Formation including cobble-rich diamictites. Jenkins has observed that the impressive exposure at Mt Capitor (Freeman and others, 1991) greatly resembles a section of the Olympic Formation at Olympic Bore (near Ringwood) on the one hand, and other than for including diamictic horizons, is essentially similar to the type exposure of the Pioneer Sandstone. Unfortunately, Mt. Capitor is too remotely sited for visiting with a large party.

In the central northern Amadeus Basin grey or yellow-white cap-dolomites occur at the top of the Pioneer Sandstone and a widespread bed of small stromatolites forms a regional marker at the base of the grey-green or red shales of the Pertatataka Formation which is paraconformable above.

Stop 10.4 -- Roadside panorama

East of Ross River or Allua Well, gritty feldspathic sands and fluvial sands and conglomerates of the Gaylad Sandstone occur below the sharp base of the Pertatataka Formation. The fluvial conglomerates downcut through the Pioneer Sandstone into the Gillen Member of the Bitter Springs Formation c. 3 km east of Allua Well, suggesting the presence of an important (type '1') sequence boundary.

Stop 10.5 -- Eastern end of Gaylad Range

At the eastern end of Gaylad Range, c. 10 km west of 'Ringwood' homestead, upper parts of the Olympic Formation are poorly exposed in a valley adjacent to a high rampart formed by the Bitter Springs Formation. Here the Olympic Formation includes thin limestone beds in light coloured sandstones and an upper lithic grit which is topped by a thin, intraclastic, distinctive, pink dolomite c. 25 cm thick and characteristic of the regional 'cap-dolomite' (a comparable lithotype is 30 m thick at Mt Capitor). A further 20 cm of haematic grit constitutes a likely surface of reworking below 17 m of grey-green siltstones which include thin beds of dark green-grey carbonate rich sands. The siltstone is succeeded by the coarse grits of the Gaylad Sandstone and while the base of the latter is everywhere obscured by scree, it may be presumed to correspond with the erosive sequence boundary observed farther towards the west along the Gaylad Range.

Stop 10.6 -- Olympic Formation

An exactly analogous set of exposures is present at the section of the Olympic Formation, c. 3 km SE of Olympic Bore (Fig. 28). Here pebbly, diamictic grits constituting the top of the Olympic Formation are succeeded by a c. 3 m thick pink and yellow intraclastic 'cap-dolomite'. Amongst the scree on the steep hillside above small exposures of the grey-green siltstone are present, and traced westward, become substantial outcrops with diagnostic thin olive-green, manganiferous carbonate beds, which give $\partial^{13}C$ carbonate values of -5.13 to -7.43 ‰ PDB. Unconformable above the grey-green silts is a bold outcrop of coarse feldspathic sandstone occupying the same placement as the Gaylad Sandstone west of Ringwood. Here, the sequence boundary at the base of the sandstone is well exposed and is a conglomerate rich in carbonate clasts; it forms an unusual framework rock where the clasts have dissolved out of the sand-rich matrix.

A comparable circumstance occurs at Mt Capitor where a c. 2 m thick grey stromatic limestone occurs at the top of the spectacular 30 m thick 'cap-dolomite' of the Olympic Formation (= Pioneer Sandstone). Large fragments of the limestone are eroded to form a ferruginous conglomerate (cf. 10 km W. Ringwood) at the base of a c. 2 m thick coarse grit which passes into grey-green siltstones. A second erosive boundary with large cobbles of olive-green manganiferous carbonates, occurs at the base of the grey-green platy to thick bedded, storm-bed deposit of the Waldo Pedlar Member of the Pertatataka Formation. Mt Capitor is the defined type section of the Waldo Pedlar Member (Freeman and others, 1991) which itself is considered to be the basal part of the Pertatataka Formation. Thus in this triangle south and west of 'Ringwood' a siltstone

sequence locally with diagnostic olive-green carbonate beds occupies an intervening position between cap-dolomites of the Olympic Formation and several arenites (Gaylad Sandstone ?= Waldo Pedlar Member) disconformable or unconformable above and defined to be basal to the Pertatataka Formation. Where exposed, grits associated with a transgressive surface of reworking occur at the base of the intervening siltstones, which can be considered as a pre-Pertatataka high-stand systems tract.

Stop 10.7 -- Olympic Formation, Ringwood

Some 11 km southeast of 'Ringwood' homestead grey-green siltstones with many thin olive-green manganiferous carbonate interbeds are paraconformable above the cap-dolomite of the Olympic Formation. Locally, red-stained, grey-limestone stromatolite heads are associated with the top of the cap-dolomite. Close examination reveals the stromatolitic carbonates filling dissolution fissures in the cap-dolomite, indicating some ? minor break in deposition. The siltstones with limestone interbeds at this site constitute the type section of the Halfway Dam Formation, and are some 205 m thick where they pass up into red pelites. Several medium to thick bedded dark gritty limestones are present. A $\partial^{13}C$ carbonate value of -6.94‰ was obtained from one of the lower limestone interbeds. Here, the Gaylad Sandstone is not seen, leaving the upper extent of the new Formation an open enigma. Similarly, near the Cleary-Creek Yards a section of perhaps 260 m of the Halfway Dam Formation is exposed. The lower carbonates are relatively thin at the Clearly-Creek Yards and thin storm-bed sandstones occur high in the section.

The Halfway Dam Formation occupies the equivalent position of the Moolooloo Siltstone Member of the Brachina Formation (Sanderson Subgroup) in the Adelaide Fold Belt. Thus, the sequence boundary below the Gaylad Sandstone and Pertatataka Formation is likely to equate with that of one of the sequence tracts postdating the Sanderson Subgroup (presuming that the latter does not itself include cryptic erosional breaks). Lithological similarities permit correlation between the Pertatataka Formation and the Bunyeroo Formation and older Wonoka Formation of the Adelaide Fold Belt, and this is supported by the suggested equivalence of distinctive acritarch microfloras in the late Pertatataka Formation (Zang and Walter, 1989, 1992) and a unit resembling the Wonoka Formation in the Officer Basin (Jenkins and others, 1992). Large leiosphaerids also occur high in the Wonoka Formation in its type section of Brachina Gorge. A difficulty with this idea is that the basal Ediacaran sequence boundary of the Flinders Ranges would seemingly be cryptic in central Australia. Alternatively, the basal Pertatataka boundary could directly equate with the basal Ediacaran, and this would accord well with the storm-deposited arenites of the Waldo

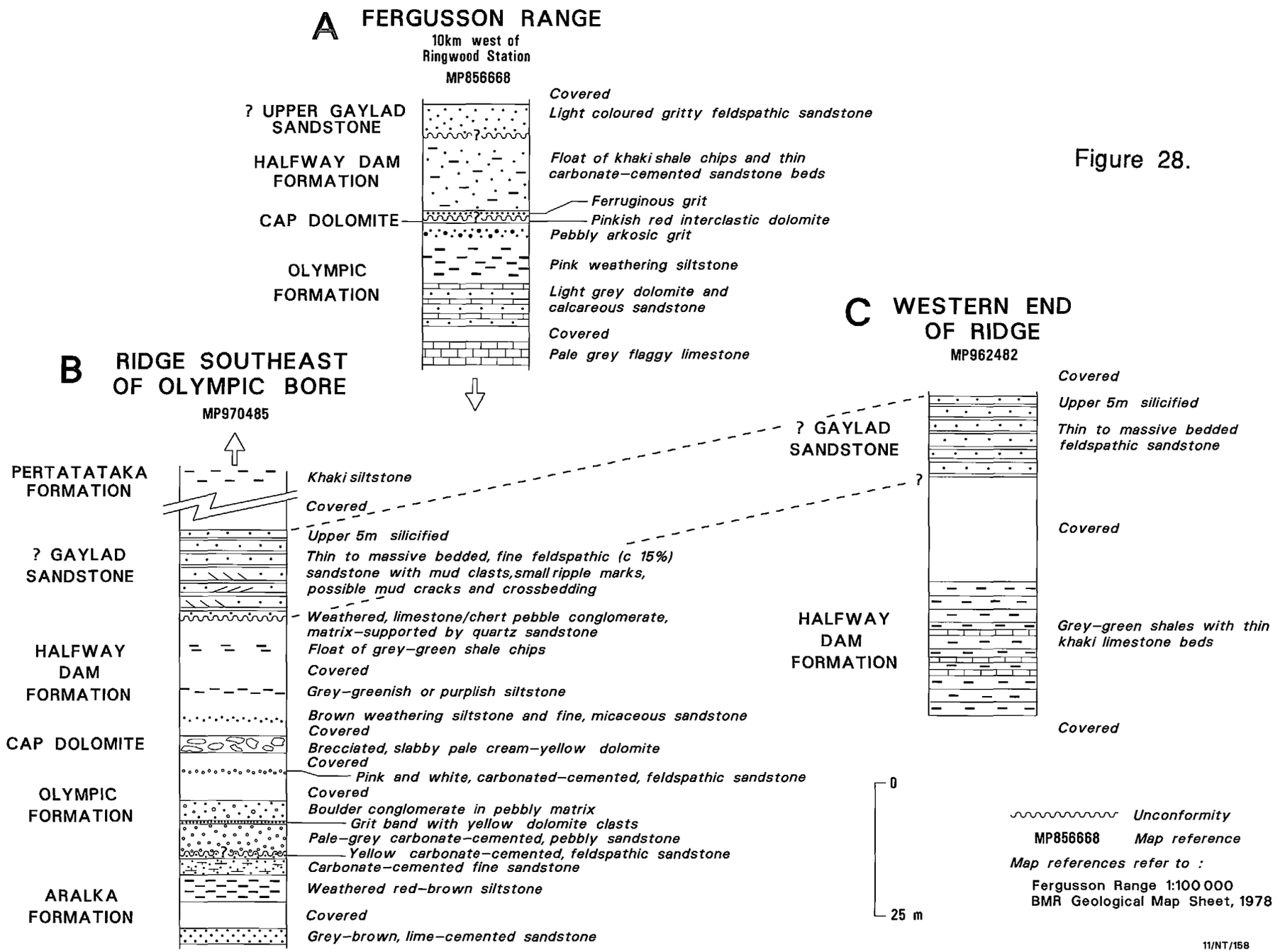

A FERGUSSON RANGE
10km west of Ringwood Station
MP856668
? UPPER GAYLAD SANDSTONE
HALFWAY DAM FORMATION
CAP DOLOMITE
OLYMPIC FORMATION
Covered
Light coloured gritty feldspathic sandstone
Float of khaki shale chips and thin carbonate-cemented sandstone beds
Ferruginous grit
Pinkish red interclastic dolomite
Pebbly arkosic grit
Pink weathering siltstone
Light grey dolomite and calcareous sandstone
Covered
Pale grey flaggy limestone
B RIDGE SOUTHEAST OF OLYMPIC BORE
MP970485
PERTATATAKA FORMATION
? GAYLAD SANDSTONE
HALFWAY DAM FORMATION
CAP DOLOMITE
OLYMPIC FORMATION
ARALKA FORMATION
Khaki siltstone
Covered
Upper 5m silicified
Thin to massive bedded, fine feldspathic (c 15%) sandstone with mud clasts, small ripple marks, possible mud cracks and crossbedding
Weathered, limestone/chert pebble conglomerate, matrix-supported by quartz sandstone
Float of grey-green shale chips
Covered
Grey-greenish or purplish siltstone
Brown weathering siltstone and fine, micaceous sandstone
Covered
Brecciated, slabby pale cream-yellow dolomite
Covered
Pink and white, carbonated-cemented, feldspathic sandstone
Covered
Boulder conglomerate in pebbly matrix
Grit band with yellow dolomite clasts
Pale-grey carbonate-cemented, pebbly sandstone
Yellow carbonate-cemented, feldspathic sandstone
Carbonate-cemented fine sandstone
Weathered red-brown siltstone
Covered
Grey-brown, lime-cemented sandstone
C WESTERN END OF RIDGE
MP962482
? GAYLAD SANDSTONE
HALFWAY DAM FORMATION
Covered
Upper 5m silicified
Thin to massive bedded feldspathic sandstone
Covered
Grey-green shales with thin khaki limestone beds
Covered
0
25 m
Unconformity
MP856668 Map reference
Map references refer to :
Fergusson Range 1:100 000
BMR Geological Map Sheet, 1978
11/NT/158

Figure 28.

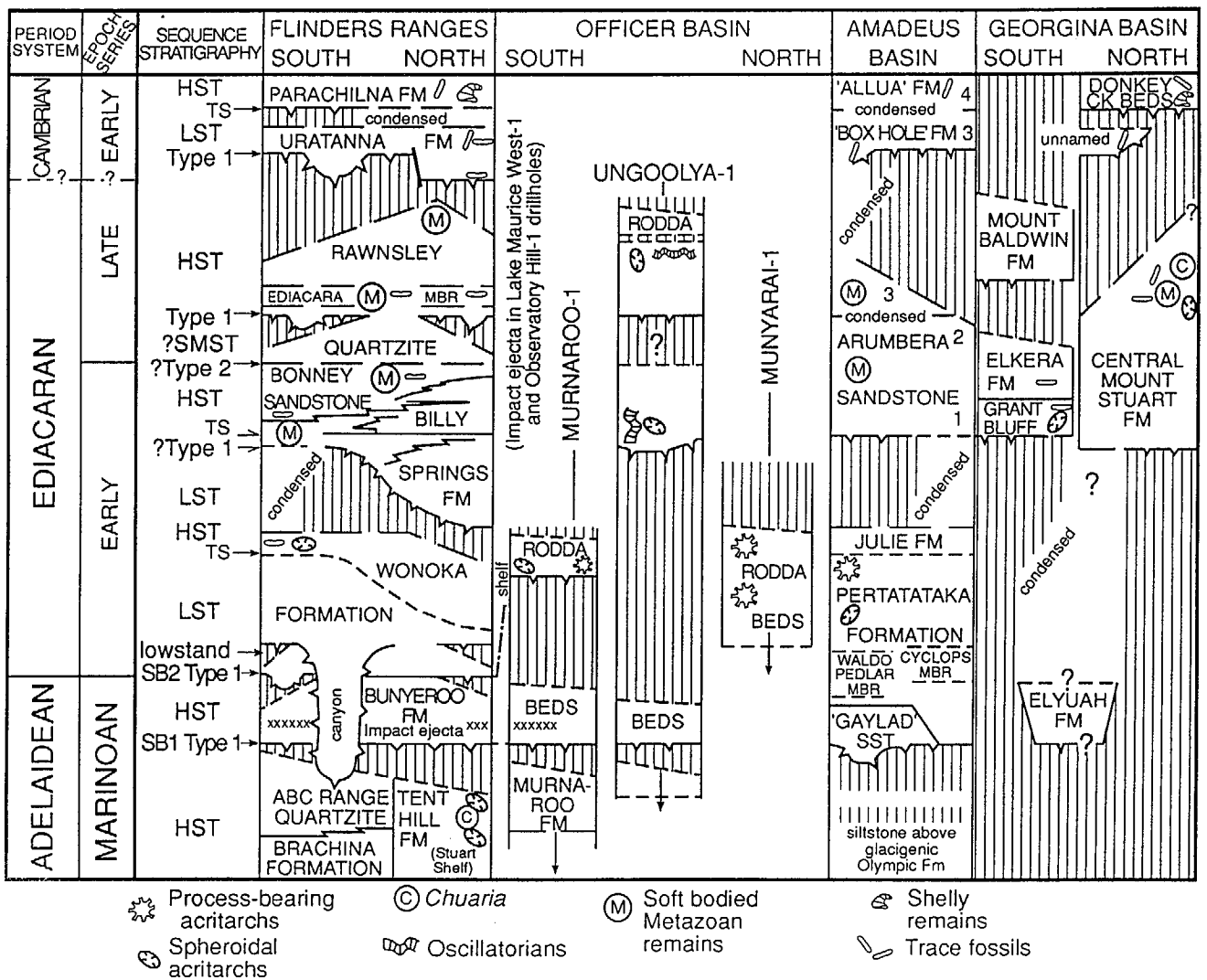


Figure 29. Suggested correlations for the Adelaide Geosyncline and the main intracratonic basins of central and southern Australia. This figure is intended for use in discussions following the last stop (Stop 10.7).

Pedlar Member occupying large erosive channels (Freeman and others, 1991) that mirror the 'canyons' of the older Wonoka Formation, and rather closely resemble the storm-deposited sandstones of Unit 2 of the Wonoka Formation (Haines, 1987, 1990) in the central Flinders Ranges. This possibility is the idea most favoured by Richard Jenkins.

A third possibility is that the Pertatataka Formation and Julie Formation together overlap the Billy Springs Formation. The khaki-green siltstones of the Pertatataka Formation are indistinguishable from the Billy Springs Formation, which includes prevalent oolitic limestones and shallow-water dolomites up-section. Shales and intercalated quartzites in some sections in an upper part of the Billy Springs Formation are red coloured. The strange cyclical layering that is characteristic of the Cyclops Member of the Pertatataka Formation is closely replicated elsewhere in facies in proximity to glacigenic deposits. Thus it may be possible that the Cyclops Member is a lateral equivalent of the reported glaciomarine deposits in the Billy Springs Formation.

A fragmentary $\partial^{13}C$ carbonate isotopic record from the Billy Springs Formation (Pell, 1989; Reid, 1992) is limited by the low-grade regional metamorphism of the unit and an absence of carbonates through much of the sequence. Values vary from c. -2‰ PDB in a lower interval to c. +7‰ in the prominent oolitic carbonate scarp that forms Wild Man Bluff. No acritarch preparations have been made, though local lagoonal siltstones are carbonaceous.

Discussions based on figure 29.

Day 11 Friday, 21 May

Return to Alice Springs.

Most of the group will return at about 7:30 am for the 1 hour drive into Alice Springs. One or more vehicles could remain at Ross River until about 10.30 am so that nearby sights could be briefly examined. These could include the columnar stromatolites of the Bitter Springs Formation or the trace fossils of the Arumbera Sandstone.

Stop 11.1 -- Arumbera Sandstone (Optional Stop)

(J.F. Lindsay, AGSO, Canberra)

The latest Proterozoic and early Paleozoic rocks of the Amadeus Basin are well exposed in the Ross River Gorge. The section, at this locality, occurs in the centre of a major sub-basin, the Ooraminna Sub-basin, and is largely carbonate. In all, 5 sequence have been

identified from the base of the Arumbera Sandstone which is latest Proterozoic in age to the top of the Cambrian (Fig. 30). Those with the time may wish to drive or walk at least part of the section. From the view point of our present objectives the Arumbera Sandstone is most relevant and is well exposed close to the hotel (Fig. 31).

The Arumbera Sandstone and the associated Todd River Dolomite form two depositional sequences that begin abruptly above the platform carbonates of the Julie Formation (Figure 31) (Lindsay, 1987a). In lithologic terms each sequence forms a large-scale coarsening-upward cycles. In general terms, the two sequence are similar lithologically but there are significant differences. The lowermost sequence, sequence 1, begins abruptly above the carbonates of the Julie Formation as fine-grained, red, highly fissile, silty shales. Higher in the section thin irregularly spaced fine-grained sandstone beds begin to appear with increasing frequency. The sandstone beds, which are mostly two to three centimetres thick, all have sharp bases and gradational upper contacts. Internally most are finely laminated and some contain climbing ripples. The sandstone beds all appear to be "distal" turbidites deposited into a somewhat deeper water pelagic environment. The sediments of this part of the sequence are perhaps best described as pelagic or basinal. The thin turbidite units appear to have been generated upslope during extreme storm or flood events.

Higher in the sequence the thin sandstones cease abruptly and are replaced by much thicker sandstone beds which, like their predecessors, have sharp lower contacts and gradational tops. Initially these beds are relatively infrequent and have a maximum thickness of one meter. Higher in the section they become more frequent and have maximum thicknesses of two to three meters. Ultimately the sandstones become the dominant lithology and then begin to oscillate back and forth through more shaly and then more sandy intervals. The lower contacts of these thicker sandstones are not only sharp but bear both flute and load casts. Internally they are poorly laminated and many units have poorly developed hummocky cross stratification. As the sandstones thicken higher in the sequence they tend to become more massive and featureless. Approximately 10% of these sandstone units are highly deformed as a result of soft-sediment failure. Generally the deformed units are less than 50 cm thick. These sandstone units all appear to have been generated by storm or flood events in a slope environment.

The uppermost part of sequence 1 is considerably more variable than the lower part of the sequence. In the Ooraminna sub-basin the sandstones forming the uppermost part of the sequence are thick massive and largely featureless with rare indications that they are in part large scale channels. In the Missionary Plain Trough and the

Carmichael sub-basin the upper unit is not as monolithic and consists of stacked shallowing upward cycles that are predominantly massive sandstone.

The lower or earlier of these small cycles begin in silty shales or poorly sorted silty sandstones that are generally featureless. Better sorted and more regularly bedded sandstones appear above them. Internally the sandstones are well laminated (varvelike) and towards the top they are highly contorted due to both soft-sediment folding and water-escape structures. The intensity of soft-sediment deformation increases upward until more massive cross-bedded sandstones, which are locally intersected by large channels, begin to dominate. Slump folds and mud-chip breccias are common features of the channel sandstones. Soft sediment deformation is more common in the Gardiner Range area on the southern margin of the Carmichael sub-basin there elsewhere perhaps in response to a locally steeper paleoslope. The massive well sorted sandstones near the tops of the cycles form prominent strike ridges. In the west the upper unit is entirely fluvial sandstone. These varvelike and channelled sandstones are interpreted as part of a coastal or delta plain association.

To the south, where the formation onlaps the Central Ridge, major cycles within sequence 1 terminate locally in as much as two meters of poorly sorted conglomerate. The conglomerates are interpreted as braided stream deposits. Locally, at the eastern end of the Gardiner Range, the Arumbera Sandstone appears to interfinger laterally with marine stromatolitic carbonates that were deposited in a shallow interdistributary embayment where terrigenous sedimentation was precluded.

Sequence 2, like sequence 1, begins in fissile featureless siltstones. However, the siltstones are grey green in colour in the lower part of the section only becoming red about half way up the section and do not include the thin turbidite units present in sequence 1. Instead the shales contain thin (one to seven centimetres thick) better sorted sandstone units which are often internally laminated or contain climbing ripples. Their upper and lower contacts are sharp. Some of the thicker sandstone units are much coarser in grainsize and highly glauconitic.

As in sequence 1 thick sandstone units with sharp bases begin to appear about half way through the section. Internally these units contain climbing ripples and well developed hummocky cross stratification. Ultimately thick featureless sandstone units begin to dominate and invertebrate tracks and trails and evidence of bioturbation appear. Within a small stratigraphic interval sandstone predominates and large channels appear. The channels are two to four meters thick and contain relatively well sorted medium-grained

sandstones with some glauconite. The sandstones are often cross bedded and large scale soft-sediment folds appear in the base of the channels apparently as a response to sediment loading. Mudchip breccias likewise are a common feature. Locally traces of copper mineralisation appear near the base of these channelled sandstones in an association similar to that described by Smith and Rose (1985) in the Catskill Delta. As in sequence 1 the lower part of sequence 2 appears to pass upward from a basinal or pelagic facies through the shoreface environment to a coastal plain and deltaic depositional system.

In contrast to sequence 1 the more massive channel sandstones are missing at the top of sequence 2 in the Ooraminna sub-basin such that it passes abruptly upward into laminated shales, silts and carbonates interbedded with 60 to 90 cm thick sandstone units. Festoon cross bedding and mudchip breccias are a common feature of the thicker sandstone units which are laterally quite variable in thickness. Towards the top of the sequence the thicker sandstone units disappear and thinly bedded shales and carbonates dominate. Herringbone cross stratification is present but poorly developed in the thinner units. Occasional halite casts suggest hypersaline conditions in the final phase of deposition. All of these features strongly suggest a tidal flat depositional system. These tidal deposits, while temporally part of the Arumbera sequence, are mapped lithologically as part of the Todd River Dolomite. Intraclastic conglomerates associated with the surface of transgression within the Todd River Dolomite form the upper boundary of sequence 2.

In general the above patterns within the major cycles can be observed throughout much of the basin. However, in detail they are much more complex in that they consist of numerous small scale coarsening upward parasequence similar to coal measure cyclothems but modified in that vascular plants had not evolved at the time the Arumbera Sandstone was being deposited. The parasequences vary in their make up and completeness according to their position in the depositional sequence. Where the dominant environment was in the deeper water basinal facies parasequences are at best difficult to define. However, as the overall depositional environment shallowed parasequences are better developed because small changes in water depth had a significant effect on the sedimentary structures preserved. Thus, as discussed previously, towards the top of the sequence the parasequences are much more prominent as they progress from variably bioturbated, poorly sorted sandstones to more massive well-sorted sandstones of a coastal plain/delatic depositional system.

The trace and body fossils of the Arumbera Sandstone are discussed in Appendix H.

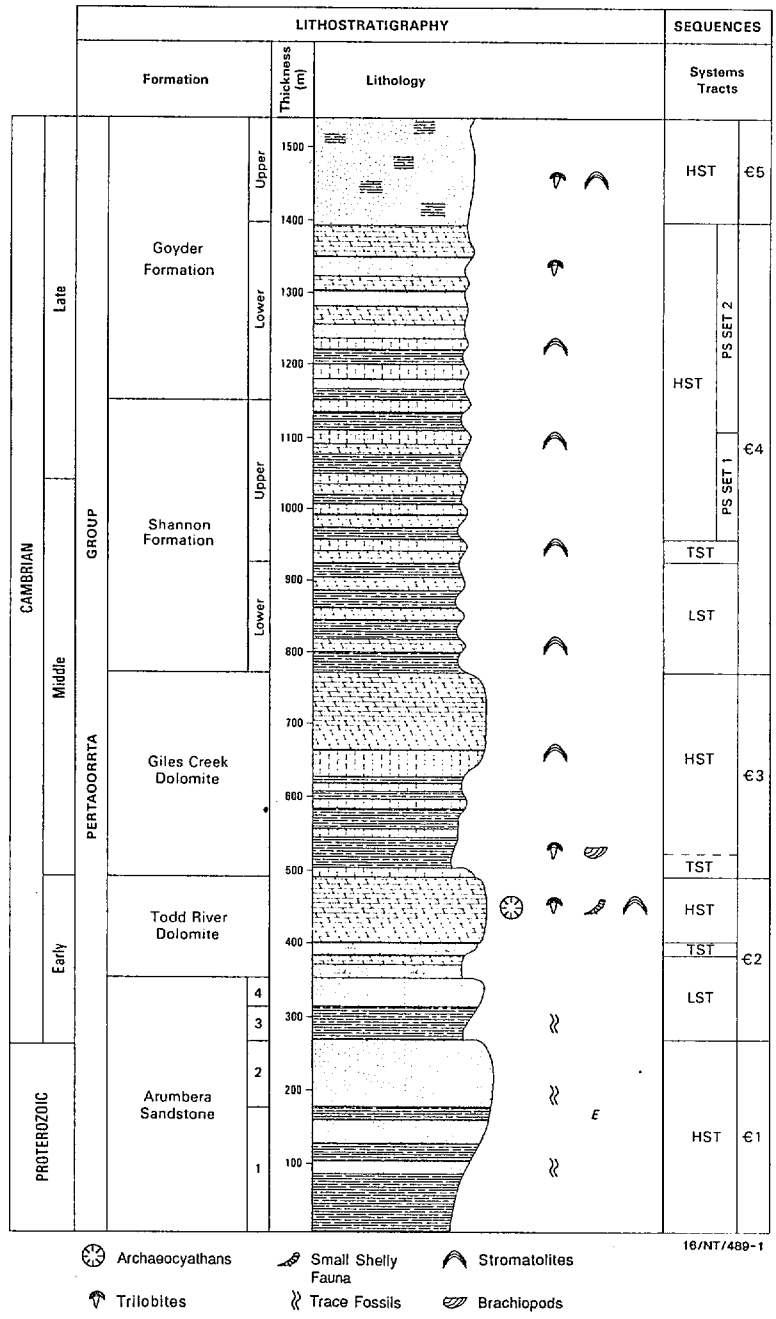


Figure 30. Cambrian sequence stratigraphy of the Ross River section in the Ooraminna Sub-basin of eastern Amadeus Basin, central Australia (after Kennard and Lindsay, 1992). The Cambrian section in this part of the basin is dominated by carbonate sedimentation, further west the section is predominantly clastic in nature. The earliest known phosphatic shelly fauna occurs in the Todd River Dolomite along with the first Archaeocyathans. HST = highstand systems tract, LST = lowstand systems tract, TST = transgressive systems tract, PS = parasequence.

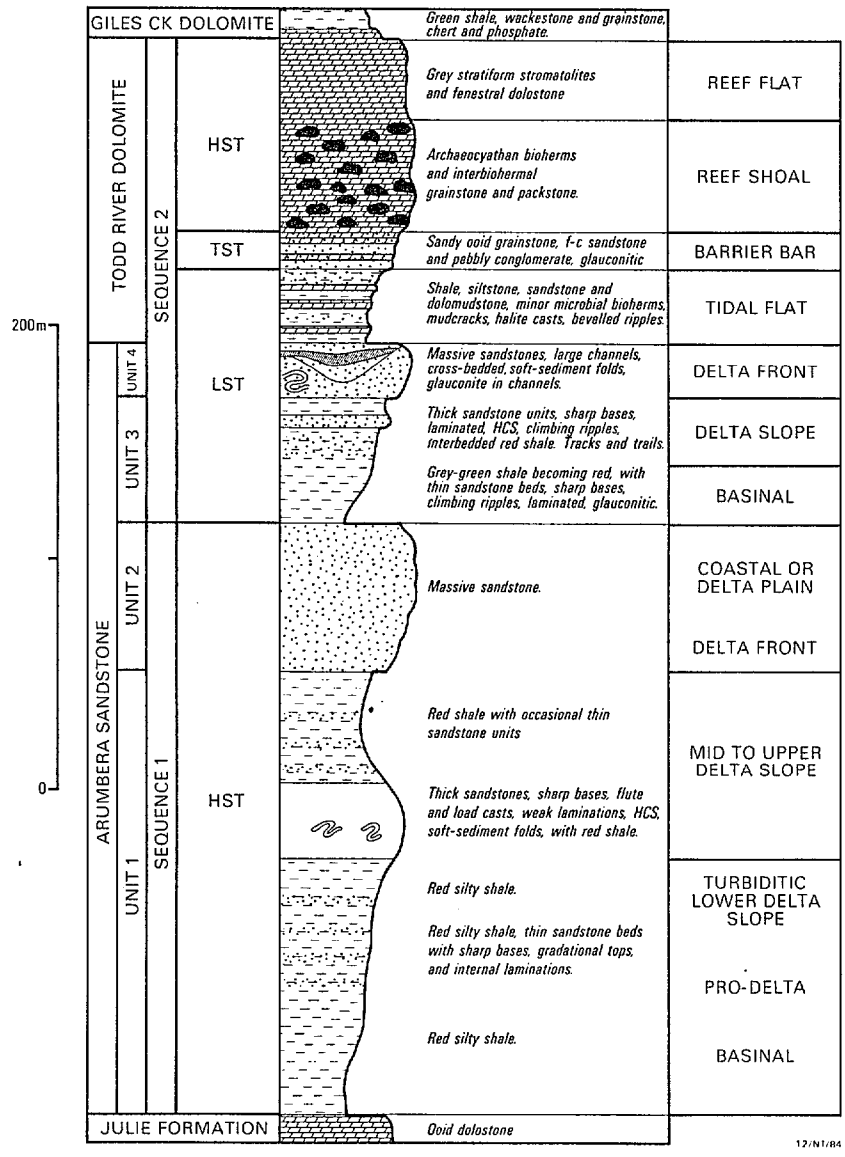


Figure 31. A detailed section through the Arumbera Sandstone at Ross River in the Ooraminna sub-basin (134° 29'E, 23° 36'S) showing the two depositional sequences and the interpreted facies associations. At this locality the clastic sediments of sequence 2 are relatively thin compared with its development closer to the centre of the sub-basin (after Lindsay, 1987a and Kennard and Lindsay, 1992). HST = Highstand systems tract, TST = Transgressive systems tract and LST = Lowstand systems tract.

Depositional Setting

The depositional space into which the Arumbera Sandstone was deposited was created in large part by the initiation of the second extensional episode of basin formation at about 600 Ma which led to the development of the major sub-basins along the basin's northern margin (Lindsay, 1987a,b; Lindsay and others, 1987; Lindsay and Korsch, 1989, 1991). The Petermann Ranges Orogeny modified the southwestern margin of the basin at approximately the same time (Forman, 1966) and further accelerated the development of growth structures and in particular the Central Ridge (Lindsay, 1987b). Sediments derived from upthrust blocks developed during the Petermann Ranges Orogeny to the south and southwest provided the primary clastic source for the formation. Thus, by the time Arumbera sedimentation began at the end of the Proterozoic a shallow, rapidly subsiding, east-west-trending, depositional basin had developed along the basin's northern margin with a broad growth structure, the Central Ridge (McNaughton and Huckaba, 1978; Lindsay 1987a), forming to the south of the main depocenters which were being fed by an active source of sediment to the southwest.

The Arumbera Sandstone was deposited during two major eustatic sealevel cycles one either side of the Proterozoic-Cambrian boundary. Sealevel rose rapidly following the deposition of the platform carbonates of the Julie Formation at the end of the previous sealevel cycle. This rapid rise combine with the rapidly subsiding sub-basins resulted in the development of a single large scale coarsening upward high stand systems tract in all three sub-basins and their connecting troughs. Locally a relatively thin transgressive systems tract developed on the margins of the sub-basins and it is significant that it is this unit that forms the major reservoir interval in the Dingo Field (a point discussed in more detail in the following section). As a consequence of the rapid changes in relative sea level lowstand deposits are of minimal importance in sequence 1. They are not seen either in outcrop or wells logs but are indicated on seismic in the centre of the Carmichael sub-basin as poorly defined mounded units presumably a lowstand fan (Lindsay, 1987a).

Sedimentation occurred in a shallow marine and deltaic or coastal plain environment in a setting very similar to that of the Devonian Catskill Delta of North America. The facies transition from the shallow-water oolitic platform carbonates at the top of the Julie Formation to the deeper-water pelagic shales and turbidites of the basal Arumbera Sandstone is abrupt suggesting that, for a brief period, the combined effects of rising eustatic sealevel and rapid basin subsidence resulted in available depositional space outpacing sediment supply. Sediment was supplied from the southwest by

braided streams. The sediments were probably deposited in a coastal plain setting for the most part with small scale deltas prograding across the carbonate platform sediments of the Julie Formation.

During the accumulation of the second sequence sealevel at first fell rapidly exposing large areas of sediment deposited during the first cycle and restricting sedimentation to the steep-sided, rapidly subsiding sub-basins. Deposition continued in the sub-basins as a series of major deltaic complexes prograded into the sub-basins from their southern and southwestern margins. Deposition was thus in large part progradational. Between the major delta complexes a depositional environment similar to sequence 1 predominated. The two sub-basins remained in communication by way of the deep channel along the northern margin of the Missionary Plain Trough. Thus the lowstand systems tract produced a large scale coarsening upward succession lithologically similar to the highstand systems tract of sequence 1 but with many smaller differences as discussed previously.

As sealevel began to rise again the depositional setting changed abruptly. The supply of clastic sediment was dramatically reduced such that the transgressive and highstand systems tracts are dominated by carbonates deposited in a shallow, often tidal or reefal, setting. The transgressive and highstand systems tracts consist of the Todd River Dolomite and the Namatjira Formation which have been mapped separately from the Arumbera Sandstone as lithologically distinct entities. The controls on sedimentation leading to this abrupt lithologic change are complex and not completely understood. In large part the change results from the growth of the Central Ridge which may have been triggered by flexure in response to sediment loading. Certainly, there is evidence that clastic sediments were being blocked by the Central Ridge and redirected around its western end into the Idirriki sub-basin. The change from clastics to carbonates along the east-west axis of the sub-basins in post Arumbera time is gradational and well documented. The slowing of basin subsidence and the broadening of the regional subsidence is probably also important because an extensive shallow platform began to develop south of the Central Ridge between the sub-basins and their source of clastic sediment thus reducing movement of clastics. The abundance of clastic sediments in the post-Arumbera Cleland Sandstone and other units in the Idirriki sub-basin suggest that the availability of clastics in the source area had not been reduced but that transport of clastics into the two eastern sub-basins was restricted.

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Appendix A

CHEMOSTRATIGRAPHY OF EDIACARIAN SUCCESSIONS IN THE ADELAIDE GEOSYNCLINE, AMADEUS BASIN AND OFFICER BASIN

C.R. Calver, School of Earth Sciences, Macquarie University, NSW

Appendix A: Chemostratigraphy of Ediacarian Successions in the Adelaide Geosyncline, Amadeus Basin and Officer Basin

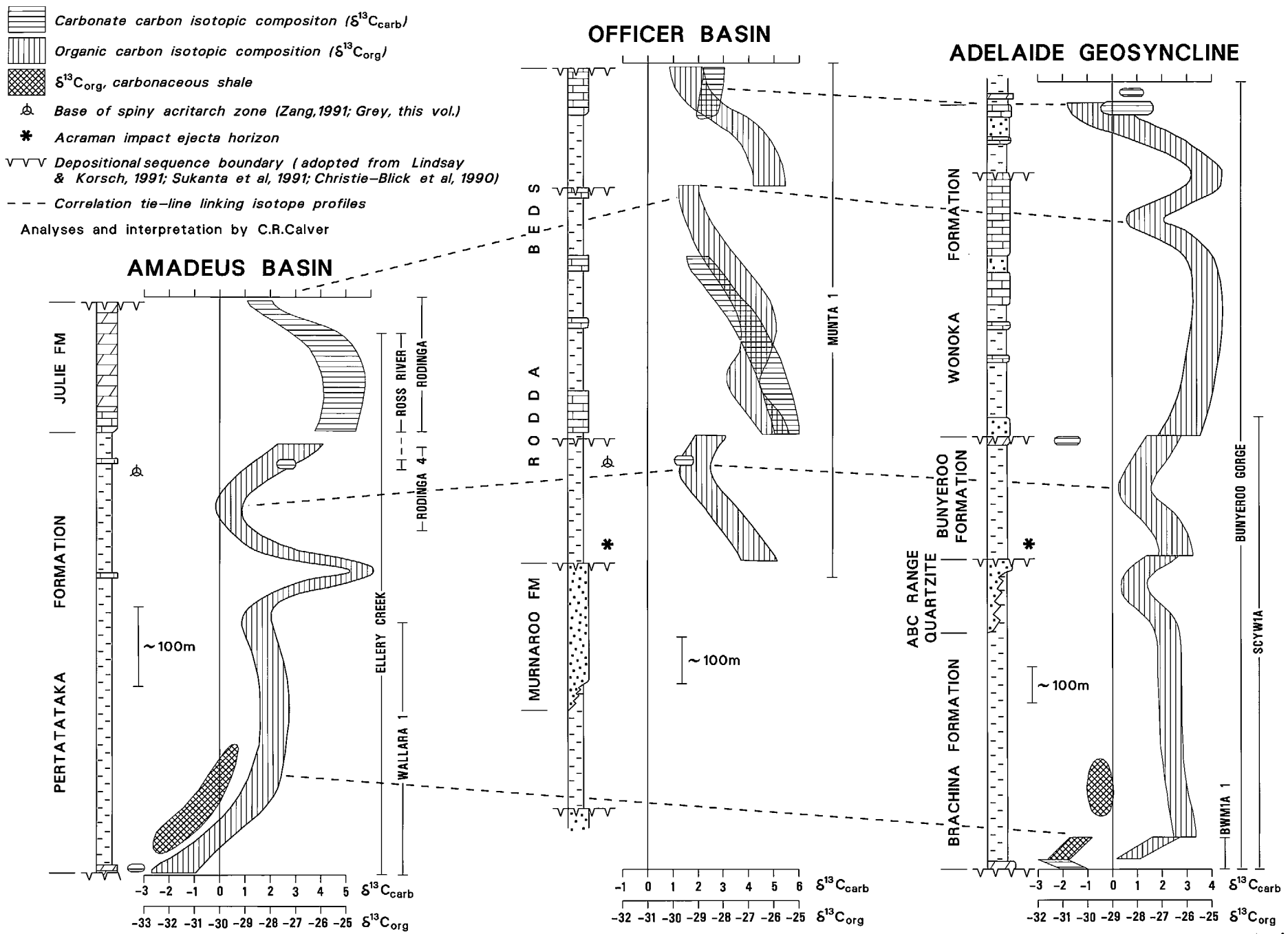
C.R. Calver, School of Earth Sciences, Macquarie University, NSW

Carbon and strontium isotope chemostratigraphy allows time-correlation of Ediacarian rocks of the Amadeus, Officer, Georgina and Savory Basins and the Adelaide Geosyncline. The method gives consistent results within basins and, for the most part, suggests geologically reasonable correlations between basins. Important corroborative information is provided by an associated acritarch biostratigraphic study (Grey, this volume), by the occurrence of the Acraman impact ejecta horizon in the Officer Basin and Adelaide Geosyncline (Wallace and others, 1989), by some sequence-stratigraphic correlations (Christie-Blick and others, 1990; Sukanta and others, 1991) and by broad lithostratigraphic similarities in all basins (e.g., Preiss and others, 1979). Correlations presented here (Fig. 1) are provisional, and contingent on further work.

As primary carbonates are not stratigraphically widespread, much of the chemostratigraphy is necessarily based on organic carbon isotopic composition ($\partial^{13}C_{org}$), and a large mass of $\partial^{13}C_{org}$ data has been accumulated. Stratigraphic profiles of $\partial^{13}C_{org}$ show remarkably little scatter and display consistent and stratigraphically useful patterns of variation. A correction factor is applied to take account of an observed, inherent $\partial^{13}C_{org}$TOC relationship. Variations in thermal maturity within basins are reflected in overall changes in $\partial^{13}C_{org}$, but do not significantly alter patterns of stratigraphic change. Least-altered isotopic compositions of carbonate carbon ($\partial^{13}C_{carb}$) and strontium are selected with reference to trace-element (Sr^{2+}, Mn^{2+}), $\partial^{18}O$ and petrographic data. Sampling intervals are approximately 10 m or less in drillholes and well-exposed outcrop sections in the Amadeus Basin, 20 m in the Officer Basin and 30-50 m in the thicker outcrop sections in the Adelaide Geosyncline. Systematic chemostratigraphy has been restricted to the interval below the Ediacara fauna-bearing Pound Subgroup and its correlates, because of the rarity of suitable rock types in these latter units.

The Nuccaleena Dolomite at the base of the Ediacarian in the Adelaide Geosyncline, and the 'cap dolomite' in the Amadeus Basin, have similarly depleted $\partial^{13}C_{carb}$(-1.5 to -3.5 ‰, Fig. 1). The lower part of the Pertatataka Formation (shale and turbidite, Amadeus Basin) is characterised by a rising trend in $\partial^{13}C_{org}$. The lower part of the Brachina Formation (shale, turbidite and storm-laid beds; Adelaide Geosyncline) has roughly constant, relatively enriched

Fig. 1. Generalised stratigraphic profiles of isotopic compositions of sedimentary carbon in the Amadeus and Officer Basins and the Adelaide Geosyncline. The main outcrop sections and drillholes sampled are indicated to the right of the profiles. Depleted (<2 ‰ $\partial^{13}C_{carb}$) carbonate of Wonoka Formation not included. Compositional fields shown here encompass >95% of data points, and are based on the following approximate total numbers of isotopic analyses: 200 (Amadeus Basin); 70 (Officer Basin); 130 (Adelaide Geosyncline).


Carbonate carbon isotopic compositon ($\delta^{13}C_{carb}$)
Organic carbon isotopic composition ($\delta^{13}C_{org}$)
$\delta^{13}C_{org}$, carbonaceous shale
Base of spiny acritarch zone (Zang,1991; Grey, this vol.)
Acraman impact ejecta horizon
Depositional sequence boundary (adopted from Lindsay & Korsch, 1991; Sukanta et al, 1991; Christie-Blick et al, 1990)
Correlation tie-line linking isotope profiles
Analyses and interpretation by C.R.Calver
AMADEUS BASIN
OFFICER BASIN
ADELAIDE GEOSYNCLINE
JULIE FM
PERTATATAKA FORMATION
~100m
ELLERY CREEK
WALLARA 1
ROSS RIVER
RODINGA
RODINGA 4
RODDA BEDS
MURNAROO FM
MUNTA 1
WONOKA FORMATION
BUNYEROO FORMATION
ABC RANGE QUARTZITE
BRACHINA FORMATION
BUNYEROO GORGE
SCYW1A
BWM1A 1
-3 -2 -1 0 1 2 3 4 5 $\delta^{13}C_{carb}$
-33 -32 -31 -30 -29 -28 -27 -26 -25 $\delta^{13}C_{org}$
-1 0 1 2 3 4 5 6 $\delta^{13}C_{carb}$
-32 -31 -30 -29 -28 -27 -26 -25 $\delta^{13}C_{org}$
-3 -2 -1 0 1 2 3 4 $\delta^{13}C_{carb}$
-32 -31 -30 -29 -28 -27 -26 -25 $\delta^{13}C_{org}$

$\partial^{13}C_{org}$ except in BWM1A-1 drillhole (Curnamona Craton) where a steeply rising $\partial^{13}C_{org}$ trend is seen near the base of the formation. The interval represented by the rising trend in the Pertatataka Formation may therefore be condensed or absent in the Adelaide Geosyncline (a condensed section at the base of the Brachina Formation is inferred on the basis of sequence-stratigraphic interpretations - Dyson, 1992). Both the lower part of the Brachina Formation and the lower part of the Pertatataka Formation have locally-developed carbonaceous shale horizons consisting in part of probable fossil benthic microbial mats. These black shales have anomalously depleted $\partial^{13}C_{org}$ (2 to 7 ‰ lighter than associated grey-green shales presumably containing organic carbon of predominantly pelagic origin). In the Adelaide Geosyncline, a minor negative $\partial^{13}C_{org}$ excursion occurs below thin ABC Range Quartzite at Bunyeroo Gorge, and within thicker (more proximal) ABC Range Quartzite on the Stuart Shelf (drillhole SCYW1A). The succeeding Bunyeroo Formation (shale) is characterised by a falling, then rising $\partial^{13}C_{org}$ trend. A similar pattern is seen in the lower, shale unit of the Rodda beds (Officer Basin) which is chronostratigraphically linked to the Bunyeroo Formation by the occurrence of the Acraman impact ejecta horizon. A distinctive spinose acritarch assemblage first appears within the rising $\partial^{13}C_{org}$ trend in the Rodda beds (Grey, this volume). Similarly, in the Amadeus Basin, this assemblage first appears within a rising $\partial^{13}C_{org}$ trend near the top of the Pertatataka Formation.

Limestone in the Wonoka Formation (Adelaide Geosyncline) is characterised by mostly highly depleted 618O and $\partial^{13}C_{carb}$ (the latter typically -5 to -11 ‰). Organic carbon in the Wonoka Formation is relatively enriched in ^{13}C, with no indication of depletion congruent with $\partial^{13}C_{carb}$. There is no discontinuity at the inferred stratigraphic level of incision of large canyons developed in the Wonoka Formation in the northern Flinders Ranges. These limestones, of subtidal to outer-shelf facies, may have undergone delayed diagenetic stabilisation in the deep-burial environment and consequent secondary alteration of $\partial^{13}C_{carb}$. The uppermost few metres of the Wonoka Formation consists of peritidal limestone with probably near-primary $\partial^{13}C_{carb}$ (-1 to +2 ‰). The Julie Formation (shallow-water carbonate, Amadeus Basin) has enriched primary $\partial^{13}C_{carb}$ (ca. +5 permil), as has the limestone in the middle part of the Rodda beds. Two distinct breaks in the Rodda Beds $\partial^{13}C_{org}$ profile (Fig. 1) correspond to sequence boundaries.

A limestone unit near the top of the Rodda beds in drillhole Munta-1 has a similar $\partial^{13}C_{carb}$ and strontium isotopic composition to the uppermost Wonoka Formation. Correlation is also supported by similar $\partial^{13}C_{org}$ profiles in the upper Wonoka Formation and upper Rodda beds. A hiatus is inferred in the Amadeus Basin.

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Appendix B

URATANNA FORMATION AND THE BASE OF THE CAMBRIAN SYSTEM, ANGEPENA SYNCLINE.

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Appendix B: Uratanna Formation and the base of the Cambrian System, Angepena Syncline.

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Introduction

The base of the Cambrian System in the Flinders Ranges, South Australia has long been considered to lie at the regional disconformity that separates the Ediacaran Pound Subgroup from overlying units that bear Cambrian-aspect trace and body fossils (review in Mount, 1989). However, to date, biostratigraphic criteria necessary for placement of the boundary and its correlation to other boundary sections have been largely based on negative evidence (Daily, 1976a). Recent discovery of the trace fossil *Phycodes coronatum* and unusual, organic-walled body fossils that resemble *Sabellidites cambriensis* within the Uratanna Formation form a key biostratigraphic link between South Australia and the newly-approved global stratotype for the Precambrian-Cambrian boundary in southeastern Newfoundland.

This field trip stop will visit a well-exposed section of the middle and upper portions of the Uratanna Formation near Finke Springs along the northeastern limb of the Agepena Syncline (Fig. 1). The base of this section is covered and appears to be truncated by layer-parallel faults associated with formation of the syncline.

Lithostratigraphy

Throughout the Angepena area the Uratanna Formation is underlain by the Rawnsley Quartzite of the late Proterozoic Pound Subgroup (Jenkins, 1981). Although the Rawnsley Quartzite is exceptionally thick within the syncline, especially near Mudlapena Graben, the Ediacara Member appears to be locally absent. The Parachilna Formation, the basal unit of the Early Cambrian Hawker Group, gradationally overlies the Uratanna Formation.

Three mappable units can be recognised within the Uratanna Formation: a basal channel sandstone unit, a middle siltstone/sandstone unit, and a capping cross-stratified sandstone unit (Fig. 2). Gauld (1976) informally referred to these as the lower, middle and upper members in outcrops near the Angepena Syncline.

The lower member consists of laterally-variable, amalgamated quartz arenite channel-fills that are typically massive or parallel laminated, yield flute casts and tool marks at their base, and contain structures indicative of liquefaction (Mount, in press).

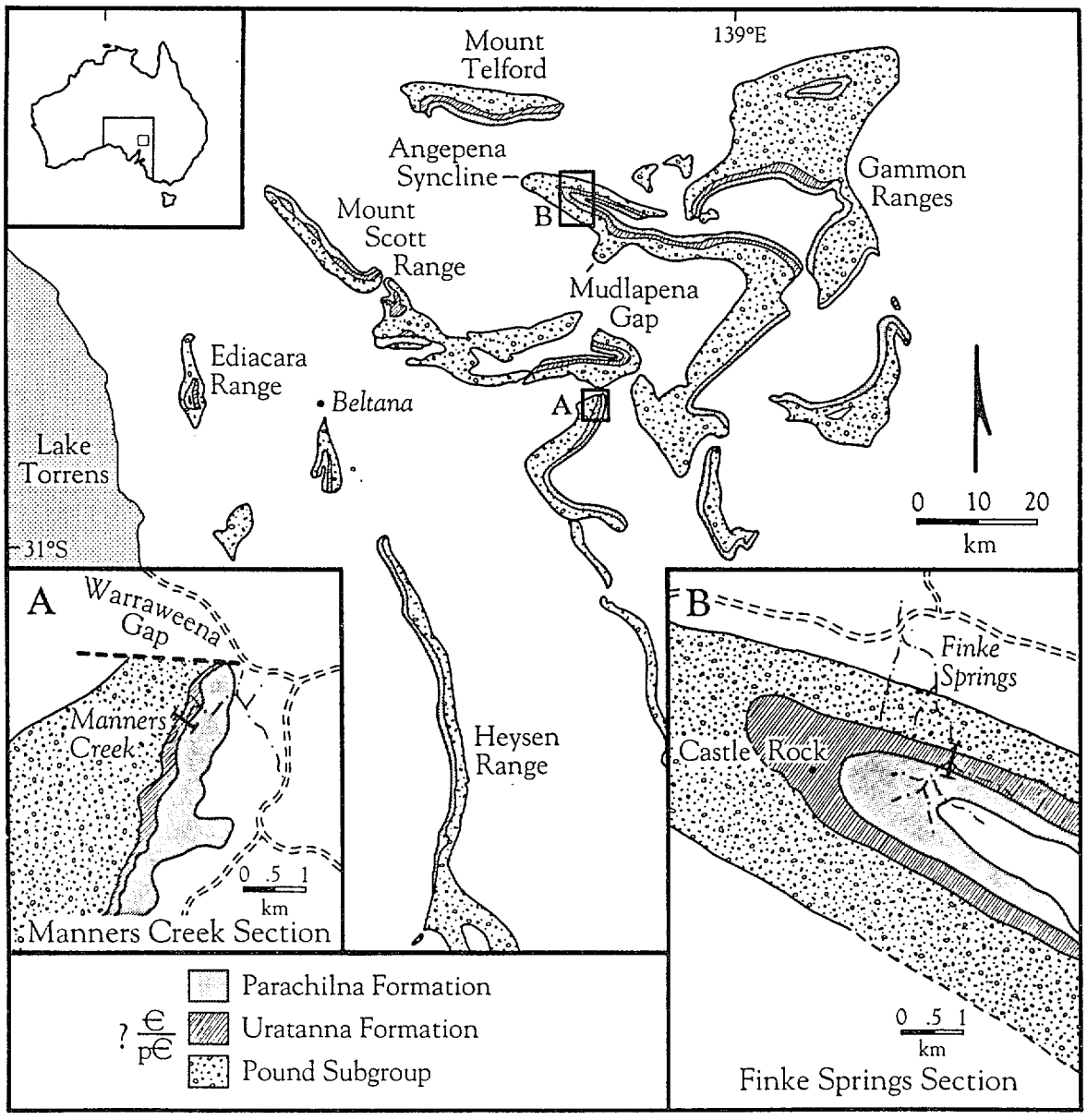

Fig. 1. Geologic and location map for northern Flinders Ranges, South Australia depicting distribution of Pound Subgroup, Uratanna Formation and Parachilna Formation and location of sections noted in text. Geologic map adapted from COPELY 1:250,000 map sheet.

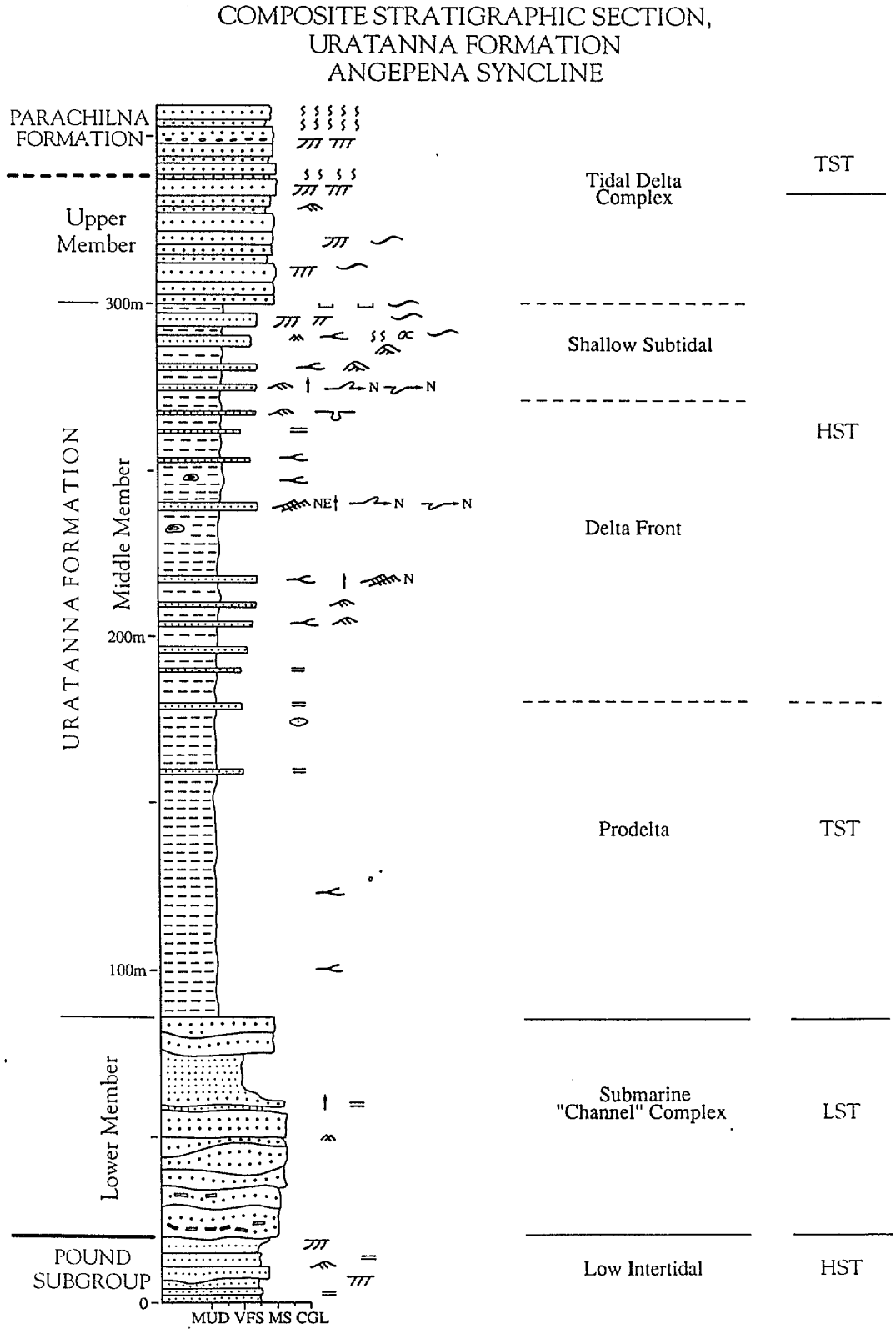

Fig. 2. Generalised stratigraphic column for the Uratanna Formation in the Angepena Syncline. Lower member of the Uratanna is absent at Finke Springs due to layer-parallel faulting associated with syncline formation.

Kaolinitic siltstone intraclasts are locally abundant, forming intraformational conglomerates. Regionally, the thickness of this member is highly variable and appears to have been locally enhanced by syndepositional faulting (Mount, 1991). At this field trip stop the member appears to have been removed by layer-parallel faulting. The sandstones of the lower member are interpreted to record energetic, mass-flow processes on a tectonically-steepened ramp (McDonald and Mount, 1992, McDonald, 1992).

The middle member, which makes up the bulk of the Uratanna Formation in the northern Flinders Ranges, sharply overlies the lower member. The lower and middle portions of this member are dominated by dark-coloured, organic-rich siltstones and mudstones with rare fine-grained sandstones. The siltstones and mudstones are finely laminated and locally contain phosphate nodules; the sandstones contain thin Bouma sequences indicative of turbidite deposition. The upper portions of the member consist of interstratified siltstones and fine- to medium-grained sandstones, with increasing sandstone frequency and thickness near the top of the member. The fine-grained sandstones contain parallel laminae, rare hummocky cross-stratification and small current, combined flow and wave ripples. Medium-grained sands are dominated by parallel laminae, tabular and trough cross-stratification. The lower and middle portions of the middle member are inferred to have accumulated in a prodelta, mud-dominated environment. The upper portions record the overall shoaling of the shelf and the deposition of delta front and delta plain sands.

The middle member grades into the overlying upper member. This member is dominated by medium- to fine-grained quartz arenites with abundant tabular and trough cross-stratification, parallel lamination and local pebble conglomerates. At least three overall thickening-upward, coarsening-upward parasequences can be recognised within the upper member. The parasequences are inferred to record cycles of aggradation/progradation on the Uratanna shelf. Each cycle appears to begin with widespread shallow subtidal deposition which aggrades to energetic low intertidal deposition. Fluvial deposits may cap these cycles locally.

Sequence Stratigraphy

In her regional study of the Uratanna Formation, McDonald (1992) suggested that the Uratanna Formation represents a complete Type 1 depositional sequence. The formation of the basal sequence boundary involved initial widespread subaerial erosion followed by extensive modification in a marine environment during rapid drowning of the shelf. The channel sandstones that make up the lower member of the Uratanna in Angepena are interpreted to represent lowstand incised-channel fill deposits, albeit of marine

origin rather than the more typical fluvial origin. The mudstones and siltstones of the lower middle member appear to form a retrogradational succession and are inferred to reflect a transgressive systems tract. The upper middle member and the entire upper member of the Uratanna Formation contain coarsening- and thickening-upward parasequences that are associated with the development of a highstand systems tract. The transition between the lowstand systems tract and the highstand systems tract in the Angepena region is placed within phosphatic units near the middle of the middle member. The top of the Uratanna depositional sequence is difficult to locate within the Angepena area. McDonald (1992) places the boundary either at the Uratanna/Parachilna contact or within the uppermost Uratanna Formation.

In their analysis of Lower Cambrian sequence stratigraphy of the Arrowie and Officer Basins, Gravestock and Hibburt (1991) included the Uratanna Formation as part of the lowstand deposits of their C1.1 Sequence. However, McDonald (1992) and Mount and McDonald (1992) have proposed that the Uratanna depositional sequence and its associated boundaries can be correlated throughout south and central Australia, including the Arrowie, Stansbury, Officer, Amadeus and Georgina Basins. If confirmed by further study, Gravestock and Hibburt's C1.1 Sequence should be subdivided to include an additional depositional sequence at the base of the Cambrian.

Fossil Occurrences and the Precambrian-Cambrian Boundary

Organic-walled fossils that strongly resemble *Sabellidites cambriensis* were recovered from finely-laminated, organic-rich olive-green siltstones of the middle and lower portions of the middle member (Fig. 2). The best-preserved specimens of *Sabellidites* were collected from outcrops along Manners Creek near Waraweena Gap approximately 10 m above the contact between the middle and lower members (Fig. 1) . In most localities the sabelliditids are difficult to identify due to intense weathering and oxidation of organic material. Poorly-preserved sabelliditids were collected from approximately the same stratigraphic interval in the Finke Springs area. *Phycodes coronatum* was also collected from finely-laminated, weakly-bioturbated olive-green siltstones of the middle portion of the middle member near Finke Springs.

Discussion

A generalised depiction of the stratigraphic range of key trace and body fossils of the Uratanna and Parachilna Formations is shown in Figure 3, which combines the earlier work of Daily (1972, 1973, 1976a,b) and Gauld (1976) with the results of recent collecting. It should be noted that the initial appearance of *Sabellidites* and the

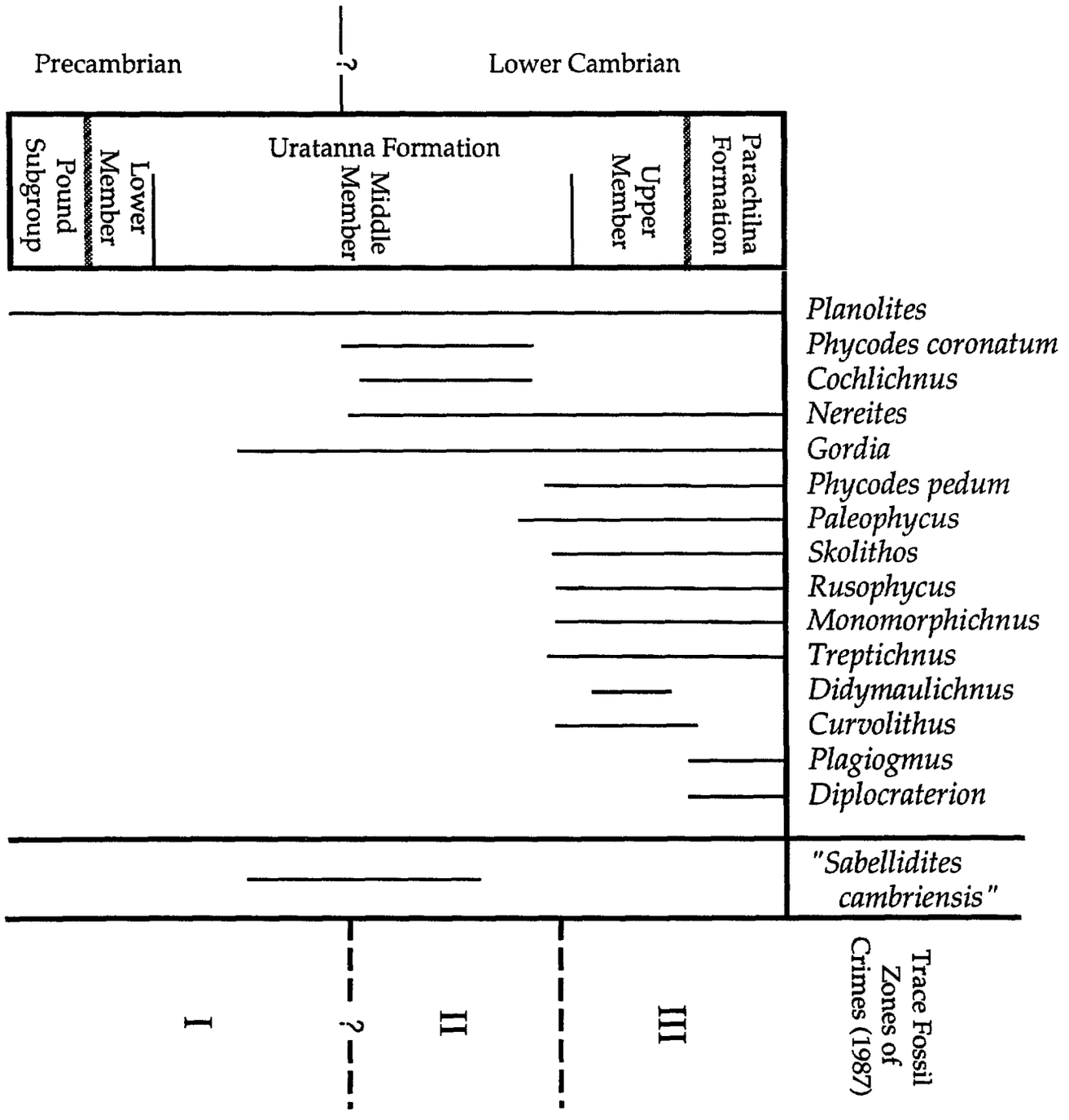

Fig. 3. Generalised stratigraphic column depicting range of important ichnogenera and body fossils. Based on data of this study and Daily (1972, 1973, 1976a,b) and Gauld (1976).

traces *Phycodes coronatum* and *Cochlichnus* in the middle member does not appear to be associated with detectable changes in depositional conditions.

The identification of the co-occurrence of *Sabellidites* and *Phycodes coronatum* may allow the most direct correlation of the Flinders Ranges boundary section with the newly-designated Precambrian-Cambrian boundary stratotype from Burin Peninsula, southeast Newfoundland. The base of the Cambrian System and Placentian Series in the stratotype of southeastern Newfoundland is placed at the base of the *Phycodes pedum* ichnofossil zone, immediately above the base of Member 2A of the Chapel Island Formation (Narbonne and others, 1987; Landing and others, 1989). Although not specifically noted by Narbonne and others (1987), *Phycodes coronatum* was described by Crimes and Anderson (1985) from the *Phycodes pedum* ichnofossil zone.

Body fossils are rare within the Newfoundland stratotype. However, Landing and others (1989) define a *Sabellidites* Zone within the Chapel Island Formation that brackets the Precambrian-Cambrian boundary, spanning an interval from immediately below to immediately above the *Phycodes pedum* ichnofossil zone. Based on the association of *Phycodes coronatum* and *Sabellidites* in the Flinders Ranges, the middle and upper portion of the middle member of the Uratanna Formation is correlated with Member 2A of the Chapel Island Formation at Burin Peninsula in southeastern Newfoundland. The Precambrian-Cambrian boundary in the Flinders Ranges is therefore placed at the first appearance of *Phycodes coronatum* in the middle member. Unfortunately, to date no material has been recovered from the lower middle member or the lower member that would clearly indicate *either* a Cambrian or Precambrian age. Therefore, this boundary placement is tentative.

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Appendix C

A $\partial^{13}C$ Survey of Carbonate in the Early Ediacaran Wonoka Formation

Urlwin, B., Ayliffe, D.J., Jansyn, J., McKirdy, D.M., Jenkins, R.J.F. and Gostin, V.A.

Appendix C: A $\partial^{13}C$ Survey of Carbonate in the Early Ediacaran Wonoka Formation

Urlwin, B., Ayliffe, D.J., Jansyn, J., McKirdy, D.M., Jenkins, R.J.F. and Gostin, V.A.

The Wonoka Formation is a well described example of a late Proterozoic (Early Ediacaran) storm dominated carbonate shelf sequence (Haines, 1988, 1990). The considerable thickness and lateral extent of the formation throughout the northern Adelaide Fold Belt makes it an excellent opportunity for applying the principles of carbon isotope stratigraphy.

Following a study of Eickoff and others (1988) on carbonates veneering the wall of a 'canyon' associated with the lower Wonoka Formation, and several "random" samples analysed by Pell (1989), Jansyn (1990) made a reconnaissance survey of the Wonoka Formation in a deep part of the basin exposed at Bunyeroo Gorge, only 12 km south of the type section at Brachina Gorge.

Here we report ongoing work from stratigraphic sections measured through the Wonoka Formation exposed in the Warraweena area (Urlwin, 1992), located on the boundary of the North Flinders and Central Flinders structural zones (Preiss, 1987), and further south at Pichi Richi Pass (Ayliffe, 1992). Close attention has been paid to possible diagenetic alteration of the carbonates. Samples analysed were selected according to their homogeneity, carbonate content, and visible affects of alteration. Polished thin-sections were examined using transmitted and cathodoluminescent-mode microscopy. Both techniques are essential for determining the carbonate composition of sediments and for assessing the extent of their diagenetic alteration (i.e. recrystallization and cementation).

In addition, the carbonates were geochemically analysed for the trace and major elements Sr, Rb, Mn, Mg, Fe and Ca. Selective dissolution using acetic acid limited extraction of trace elements to the carbonate phase only. Comparisons of the concentrations and ratios of these elements for each sample provides an independent assessment of diagenetic alteration (Brand and Veizer 1980, 1981, Derry and others, 1991) and is therefore another method of assessing the validity of any isotopic results obtained.

Whole-rock isotopic analysis was performed on representative portions of each sample. Several samples were geochemically and/or petrographically identified as having undergone a significant amount of diagenetic alteration. However, these were found to have $\partial^{13}C$ values that did not vary significantly from the isotopic trend

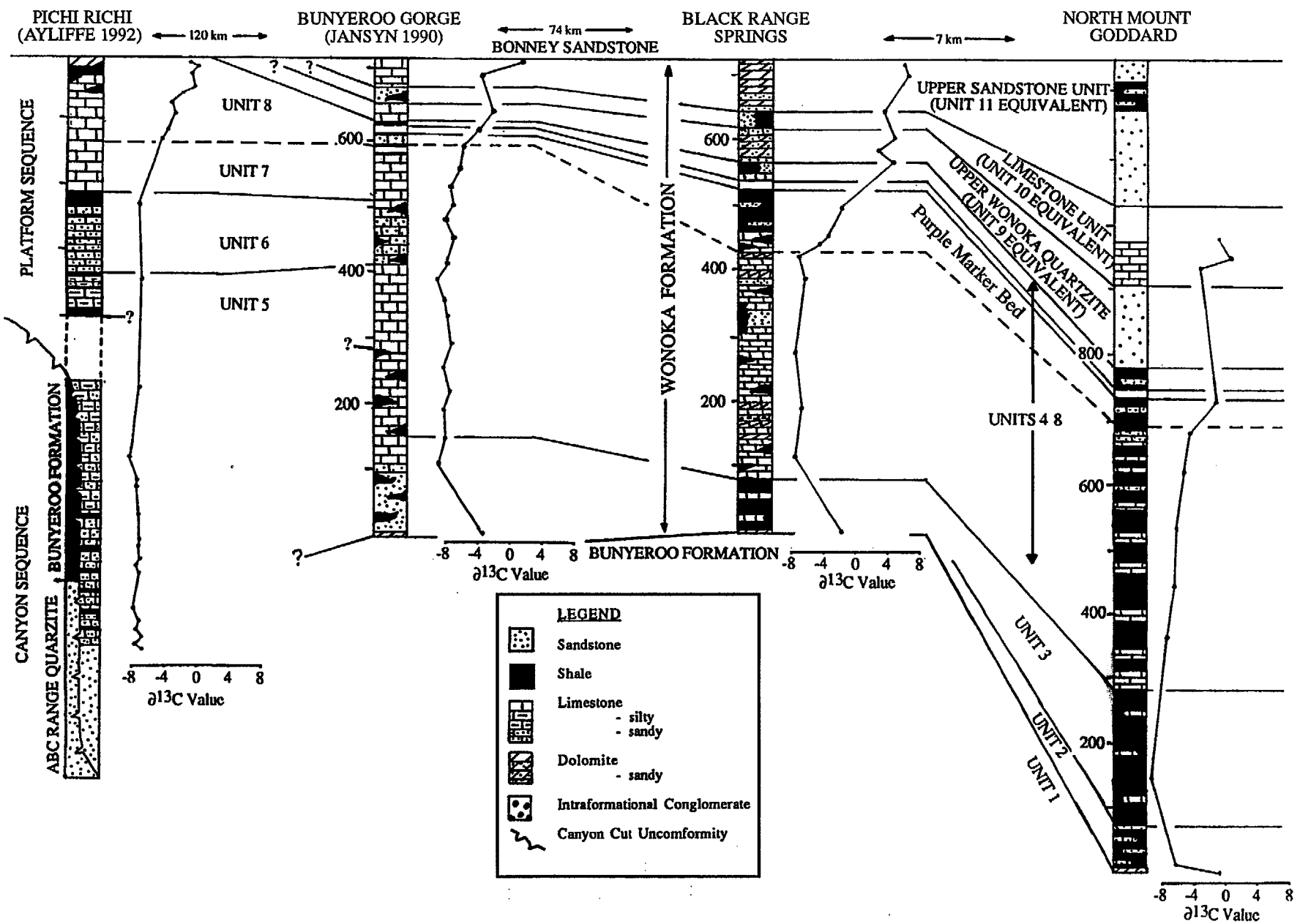

Figure 1: Correlation of carbon isotope stratigraphies of the Wonoka Formation in the northern Adelaide Fold Belt.

defined by the "least altered" samples, and are therefore included in the isotopic curves shown in Figure 1.

XRD analysis revealed that some of our carbonates from the Warraweena area had been partially dolomitized. Analysis of the calcite and dolomite phases determined that dolomitization had not significantly affected the $\partial^{13}C$ of these samples. This is thought to be due to the fact that the bulk of the carbon atoms present in the dolomite were derived from the original carbonate phase. Dolomitization resulted in isotopic shifts ranging from -1.04 to +1.21 ‰ (average @ -0.1 ‰).

Carbonates of the Wonoka Formation were found not to have been strongly affected by metamorphism (lower greenschist facies, chlorite zone: McKirdy and others, 1975) and to contain very little organic matter (TOC values consistently less than 0.05 wt%: Pell, 1989; Jansyn, 1990). Thus thermal maturation of dispersed organic matter is unlikely to have affected the carbon isotopic composition of coexisting carbonates. This view is also supported by analysis of maturation levels of kerogen in the Warraweena area (H/C atomic ratios of 0.46 to 0.56: Lemon and others, 1992). Significant shifts to heavier kerogen $\partial^{13}C$ values occur only at much higher maturation levels (H/C < 0.2: McKirdy and others, 1975). Such minimal TOC values also reduce the possibility of carbonate isotope values having been affected by the incorporation of biogenic CO_2 from the bacterial decay of organic matter during early diagenesis. The fact that most of the carbonates in question are associated with thick sequences of shale and/or are micritic limestones of low initial porosity, minimises their potential for later diagenetic alteration.

Our carbon isotopic data are plotted stratigraphically and combined to produce an intra-basinal correlation diagram (Figure 1). Curves obtained for all sections are readily correlative across the Wonoka basin.

The very negative excursion preserved in the lower two thirds of the Wonoka Formation (Units 2-7: Haines, 1988, 1990) is extremely consistent, with mean $\partial^{13}C$ values ranging from -6.7 ‰ to -7.7 ‰ across the four measured sections. Similar negative values are recorded in the clast and matrix of an intraformational conglomerate ($\partial^{13}C$= -8.3 and -8.9 ‰ respectively: Ayliffe, 1992) and in carbonates veneering canyon walls ($\partial^{13}C$= -8.9 ‰: Eickoff and others, 1988; Ayliffe, 1992). Fracture-filling calcite also gives the same negative signal (Ayliffe, 1992), suggesting that its $\partial^{13}C$ value was inherited from the host carbonate.

The positive excursion apparent in the upper portion of the Wonoka Formation commences just prior to a purple marker bed, near the boundary between Units 7 and 8 (Haines, 1988, 1990), and is

illustrated by the dotted line on Figure 1. The beginning of this excursion correlates with a possible sequence boundary described by von der Borch and others (1988) and Christie-Blick and others (1990). The maximum extent of the excursion is recorded in Unit 11 at Black Range Springs ($\partial^{13}C$ = +6.6 ‰). The more muted expression of this maximum in the same unit at Bunyeroo Gorge ($\partial^{13}C$ = +2.1 ‰) may reflect increased oxidation and recycling of organic matter in restricted, shallow lagoonal environments which contrast with the open marine conditions further north (Haines, 1990). The shift to higher $\partial^{13}C$ coincides with the general shallowing of the sequence and the appearance of thick sandstone, sandy dolomite and sandy limestone units. The onset of the excursion occurs at a consistent stratigraphic level and hence, appears to be a basin-wide phenomenon.

Efforts to interpret the carbon isotopic signatures of late Proterozoic carbonates encounter the problem of overprints associated with secondary, post-depositional alteration. However, the fact that apparently altered samples give $\partial^{13}C$ values consistent with the results obtained from samples most likely to retain primary isotopic values (i.e. minimal trace element contamination and/or recrystallization) suggests that these isotopic signatures are relatively robust and lends credibility to their stratigraphic utility. It is less clear how the values recorded in rocks relate to ambient waters at the time the sediments were deposited or initially altered. There is no model that satisfactorily explains the pronounced negative excursion through the older Wonoka Formation.

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Appendix D

Syn-Sedimentary Influence of Salt Movement in the Type Section of the Neoproterozoic Aralka and Glaciogenic Olympic Formations.

Martin Kennedy, University of Adelaide, South Australia

Appendix D: Syn-Sedimentary Influence of Salt Movement in the Type Section of the Neoproterozoic Aralka and Glaciogenic Olympic Formations.

Martin Kennedy, University of Adelaide, South Australia

Salt movement within the Bitter Springs Formation of the Amadeus Basin had a dramatic syn-depositional effect on the Neoproterozoic sequence (Kennedy, 1993; Oaks and others, 1991, Lindsay 1987). Broad scale folding during the Alice Springs Orogeny and subsequent erosion has exposed a clear cross-sectional view of a "fossil" diapir; the Olympic Structure.

Previous work conducted on the Late Proterozoic stratigraphy in the north-east of the basin has established laterally variable thickness changes within units and regionally paraconformable relationships between units with local angular unconformities. Thickness changes and angular relationships have been used as a basis for dividing the sequence and assigning formation status. They have also been interpreted as evidence for tectonic movements, such as the Olympic and Souths Range Movements and Arltunga Orogeny (Ding and others, 1992, Freeman and others 1991, Field 1991, Shaw and others 1991). Areas with maximum thickness of sections have been used to construct balanced cross sections (Stewart and others 1991) and basin subsidence curves (Korsch and Lindsay, 1989; Shaw et. al 1991) while rapid facies variations or discontinuities are used as evidence for differentiating thrust sheets (Oaks and others, 1991).

The Olympic Structure provides an outcrop example of the localised influence salt movement had on the Neoproterozoic Aralka and Olympic Formations in their type sections. Lithological and structural variation around this structure underscores the need to account for syn-sedimentary salt influence prior to making regional stratigraphic interpretations from single areas.

Background

The outcrop is tilted 50°-90° degrees southward. Only the western flank of the structure is well exposed and discussion will be limited to this area. The local stratigraphy is summarised in figure 1.

The breccia within the region suggested to comprise the salt dome is sourced from the Gillen Member of the Bitter Springs Formation which lies to the north of the structure and penetrates the overlying Neoproterozoic sequence south-eastward (grey coloured area, figure

1). Gillen Member carbonates and sandstones at the base of this region are tightly folded while breccias in the 'shaft of the diapir' are recessive and are only visible in rare stream cuttings.

Previously this zone of brecciated Gillen member and its relationship to the flanking sediments has been interpreted differently in the two structural models proposed for the region (Stewart and others, 1991). In the first model, termed the Mesothrust interpretation, the zone of brecciated Gillen Member resulted from Devono-Carboniferous transport of the Phillipson sheet containing the Aralka, Olympic and Pertatataka Formations over the auchthonous zone of Gillen Member carbonates north of the fault (grey area figure 1). This interpretation requires the Gillen Member and overlying Neoproterozoic sequence to be unrelated during deposition, but structurally superimposed at a much later time. In the second model, the Megathrust interpretation, the flanking sediments and Gillen Member are not separated by a thrust sheet margin, but are part of a larger consistent autochthonous area.

Synsedimentary Evidence presented below supports the Megathrust model by demonstrating the influence salt movement within the Gillen Member had on the overlying Proterozoic rocks. Linkage of these two areas during the Proterozoic implies the flanking sequence could not have been emplaced as part of the separate Phillipson thrust sheet during the Devono-Carboniferous as the Mesothrust model requires.

The generalised affects of salt movement on flanking strata have been summarised in Seni and Jackson (1983) and can be directly compared to features observed in figure 1. Evidence supporting interpretation of an active period of salt dome formation during the Neoproterozoic includes:

1) Dragging of the sequence overlying the salt bearing Gillen Member parallel to the flank of the breccia body. Dragging becomes more exaggerated as it approaches the 'salt dome' with units progressively overturned and pierced by breccia toward the top of the structure (south-east corner of figure 1.).

2) Thickness variations shown in figure 2 indicate periodic subsidence adjacent to the 'salt dome'. The pattern of this subsidence is suggestive of primary and secondary peripheral sink development which formed in response to progressive salt migration from beneath flanking sediments into the diapiric column.

3) Syn-sedimentary normal faults directly adjacent to the 'salt dome' controlled thickness variation within the type section of the Olympic Formation and occur within the proposed secondary

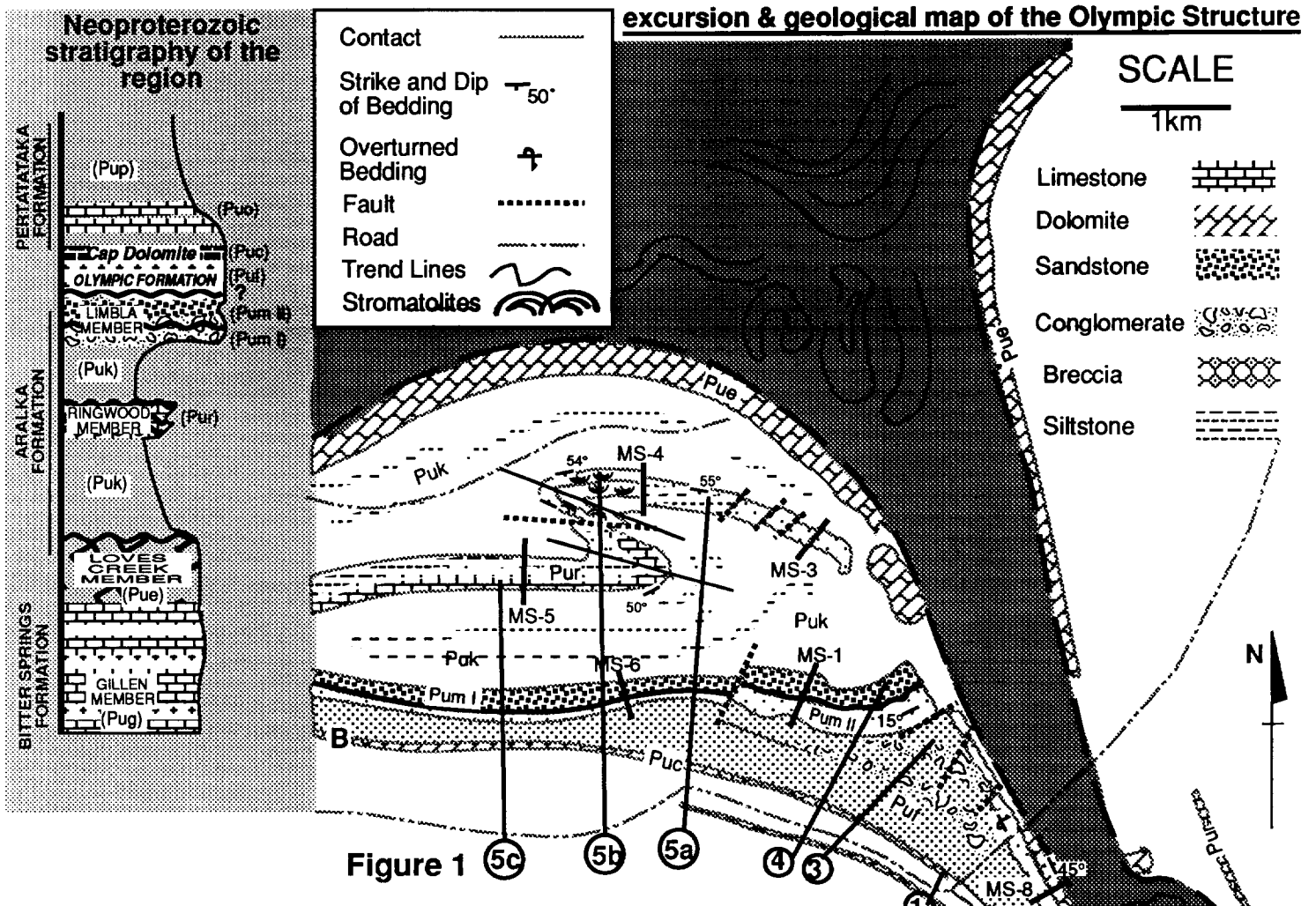
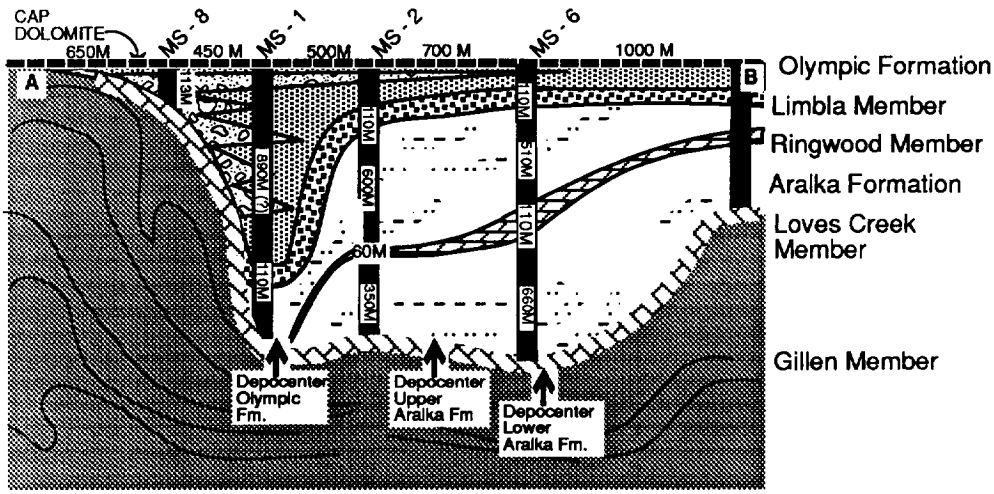
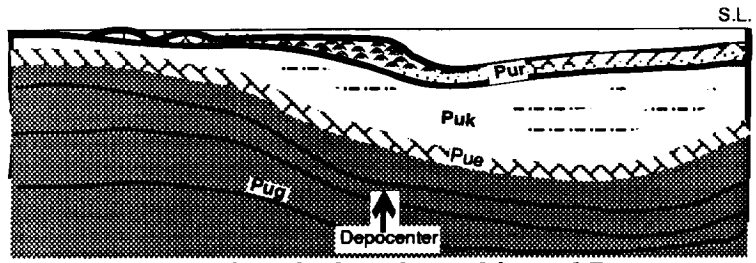


Figure 1



true thickness of flanking sediments from A-B on map

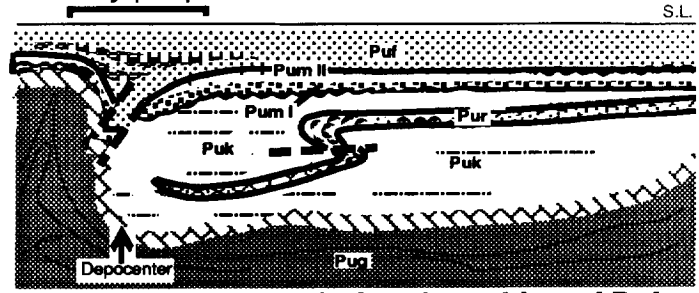
Figure 2



reconstruction during deposition of Pur

Figure 3a

secondary peripheral sink



reconstruction during deposition of Puf

Figure 3b

peripheral sink (figure 3b). These faults sole into the Gillen Member within the 'diapir'.

4) Folding of the Aralka Formation occurs on the outboard flank of the subsiding area interpreted as the peripheral sink. These folds are similar to turtle anticlines described around other salt structures (Seni and Jackson 1983).

5) Erosion of the lower Aralka Formation beneath the Limbla Member occurred in a limited area which corresponded with the crest of the 'salt dome'. Paraconformable relations beneath the Limbla Member exist away from this margin within the 'peripheral sink'.

6) Grain size increases toward the 'salt dome' margin in the Pertatataka Formation, cap dolomite, Olympic Formation and lower Limbla Member indicate the former presence of a topographic high, which may have been the crest of the salt dome.

7) Facies variation within the Ringwood Member changing from higher energy oolitic sands and stromatolitic bioherms to basinal features occurs away from the 'salt dome' toward the 'peripheral sink'.

8) Palaeocurrent was redirected away from the 'salt dome' within the Limbla Member suggesting this area formed a topographic high during deposition.

Locality 1: The Ringwood road crosses a low strike ridge, enabling the study of an excellent section of fine grained dolomicrites composing the post Olympic glacial cap dolomite. These dolomicrites show deeper water features (incomplete Bouma Sequences) while lacking evidence for a shallow water origin such as desiccation features, fenestral fabrics or evidence of storm or wave reworking. They are overlain by 20cm wide iron oxide-rich stromatolites similar to previously described stromatolites of a deeper water origin. These dolomicrites are interpreted as a transgressive sequence tract while the overlying stromatolitic horizon is thought to be a condensed sequence representing a downlap surface. The cap dolomite in this area displays westward directed asymmetric slumping. Grain size increases occur in a southeasterly direction within this unit as it is followed to the crest of the diapir until it becomes conglomeratic near Locality 2.

Locality 2: Follow a small track departing southward from the main road about 100m west of Locality 1. This track parallels the low strike ridge formed by the cap dolomite and continues to a low pass through the grey dolomites of the Bitter Springs Formation. Walk eastward from this pass to Locality 2a.

This area represents the overturned crest of the diapir which may have been extrusive during deposition of the Pertatataka Formation. Examine the grey dolomitic breccias in the low ridge that forms the pass. These may be of a tectonic origin related to later faulting, or they may be extrusive diapiric breccia. Viewing clasts under thin section reveals halite pseudomorphs suggesting a previous halite component.

The cap dolomite in this area (2b) is conglomeratic with rare granitic lithologies. Dark limestone beds showing Bouma A, C, and D units (also with granitic clasts) occur about 150m upsection (2c) within the Pertatataka Formation. Like the cap dolomite, these beds become finer grained NW away from the diapir. Deposition of these limestones may be related to local briney or shallow water conditions around the salt dome, as they appear to wedge out away from the diapiric margin.

Walk north along the margin, until a low ridge is encountered (2d), this is the area where the Loves Creek Member and Limbla Member were pierced when the salt dome made the transition from pillow to diapir stage. The Limbla Member here is very steeply dipping and much thinner than exposures to the north (locality4). Southward along strike, 20m wide channels filled with ripple (herringbone?) cross laminated sandstones (Pum II) scour into the underlying sands. The base of the unit contains numerous beds of cobble conglomerate (PumI) with clasts composed of limestones eroded from the underlying Ringwood Member (locality5). These sandstones and conglomerates were deposited on a slight topographic high corresponding with the crest of the initial salt pillow.

Separating the Loves Creek Member, which maintains a considerably different dip, and the Limbla Member is a 5m thick recessive zone that follows this contact along the flank of the diapir. This zone is composed of a flow-banded breccia which resulted from differential movement between the Loves Creek Member and the overlying sequence, implying the Loves Creek Member acted as a 'sheath' over the mobile Gillen Member.

Locality 3 occurs 1km north of the road where the Limbla Member makes an abrupt change in strike from East to South in what was the secondary peripheral sink. This is a fault controlled basin filled with purple muddy siltstone and numerous conglomeratic mass flow deposits that occur within the Olympic Formation. The mass flows were sourced from the top of the salt dome and wedge out westward away from the margin. Note the clast composition of these conglomerates. In addition to indigenous reincorporated conglomerate clasts suggesting episodic uplift, and extrabasinal

lithologies probably related to glacial transport, there is a large component of limestone clasts derived from the Ringwood Member, suggestive of exposure and erosion of this formation. A conspicuous absence of Limbla Member lithologies (Herringbone Sandstone) can also be noted. Considering there is supposedly a major erosion surface below the Olympic Formation (overlying the Limbla Member) (Freeman and others, 1991, Field and others, 1991), it is noteworthy there are no rip-up clasts of Limbla Member sandstones included in the Olympic Formation conglomerates, whereas clasts of the underlying Ringwood Member are well represented.

It can be shown that the persistence of a distinctive sandy bed conformably overlying the upper Limbla Member is not disrupted by erosion at the base of the Olympic Formation in this region. Erosion of the Lower Aralka Formation and Ringwood Member has clearly occurred prior to, or partially contemporaneous with, the deposition of the lower unit of the Limbla Member (Pum I). This is evidenced by a high percentage of Ringwood lithologies in mass flows within Pum I and 1000m of missing section beneath Pum I on the flank of the diapir. Regressive HCS silt and sandstones underlying Pum I indicate shallowing basinal conditions, while Palaeocurrent data indicate a change in orientation from unidirectional westward movement within Pum I to bi-directional NE-SW during Pum II times. Measured sections in the eastern Amadeus Basin show an angular relationship between Pum I and Pum II and a sharp transition from regressive HCS basinal facies to erosional fluvial facies within Pum I, suggesting stream rejuvenation and down-cutting.

These observations support the hypothesis there was a relative sea level fall during deposition of the Upper Aralka Siltstones, leading to erosion and deposition of fluvial facies in the east and slope deposition in the west as a low stand systems tract within Pum I. Pum I is overlain by an unconformity in the East and a correlative conformity in the west containing a sequence boundary. Tidal sandstones of Pum II were deposited in a transgressive sequence tract, reworking the top sandstones of Pum I.

In the NE corner of the peripheral sink, the purple mudstones of the Olympic Formation reveal a few lonestones of probable dropstone origin. Two Synsedimentary faults, responsible for thickening within this small basin, can also be observed against the margin of the Bitter Springs Formation and so must have been active during Olympic time. They do not offset the overlying cap dolomite.

Locality 4: Locality 4 can be reached by a brief walk over the hill formed by the dip slope of the pink coloured quartz arenites of the Upper Limbla Member. A change in strike around this hill forms a small turtle anticline on the outboard flank of the peripheral sink. On the back side of this slope, conglomerates and gritty calcareous

sandstones of the Lower Limbla Member (Puml) are well exposed. Again, note the clast composition of the conglomerates and their dissimilarity with those of the grey dolomitic Bitter Springs Formation. Very similar lithologies can, however, be observed in the next locality(Locality 5) within the Ringwood Member. The calcareous sandstones of Pum I appear to be transitional, or reworked within Pum II here, and an actual erosion surface defining a sequence boundary is not apparent.

Locality 5: Locality 5 is a traverse along the strike ridge of the Ringwood Member. This area displays the influence the topographic high over the salt dome had on carbonate facies distribution. This area is most easily accessed by walking north-west 2-3km from locality4 or less easily by vehicle. By vehicle, follow a track toward the airstrip which leaves from the south side of the creek at Ringwood Homestead. Follow this track until 1km before Murray Bore and turn eastward on a disused track that eventually runs parallel to the south side of the Ringwood Member strike ridge. Walking westward from the flank of the diapir reveals changes from higher energy oolitic limestones (5a) which once lay over the dome (and are similar to clasts seen in Pum I conglomerates), to stromatolitic build ups that formed as a flanking mound on the side of the dome (5b), to storm dominated shoaling upward cycles (parasequences) deposited in the peripheral sink typical of the Ringwood Member in other parts of the north eastern Amadeus Basin (5c). The large asymmetric anticline which forms the prominent hill midway down the strike ridge may be an exaggerated result of turtle anticline formation (Locality 5b-5c, figure 3a,b) .

The Ringwood Member is included within two thick packages of sub-wave base to storm dominated dolomitic siltstone that form the recessive flats on either side of the strike ridge. Thickness changes within these units indicate salt movement and peripheral sink subsidence during their deposition. The true thickness diagram (adjusted for dip) also indicates the depocentre of the peripheral sink migrated toward the dome as salt was progressively squeezed from the flanks of the salt pillow into the column of the diapir (figure 2).

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Appendix E

COMMENTS ON THE SEDIMENTATION OF THE PERTATATAKA FORMATION

Clive R. Calver, School of Earth Sciences, Macquarie University, NSW

Appendix E: Comments on the Sedimentation of the Pertatataka Formation

Clive R. Calver, School of Earth Sciences, Macquarie University, NSW

The Pertatataka Formation consists of several hundred metres of deep-water shale, siltstone and minor sandstone in the northern and NE Amadeus Basin. In places in the middle of the formation there is a thin, shallow-water sandstone unit, known in the northeast as the Cyclops Member (Wells and others, 1970; Preiss and others, 1978). The Pertatataka Formation undergoes a southward and westward facies transition to cross-bedded sandstone and siltstone of the thicker (2000m), mostly shallow-water Winnall beds of the southern Amadeus Basin (Wells and others, 1970; Freeman and others, 1991).

Turbidite sandstone beds are common in the Pertatataka Formation in the central Amadeus Basin at Areyonga (Korsch, 1986a) and in the drillhole Wallara 1 (see Part 2 of Guide). Distal, prodelta-derived turbidite deposition and N to NE transport directions are inferred at Areyonga (Korsch, 1986a; Korsch, pers. comm., 1991). Sandstone is rare in more northerly sections such as Ellery Creek. North- and east-prograding clinoforms are locally seen in seismic reflection profiles (e.g., Kennard and Lindsay, 1991), but the Pertatataka Formation is generally characterised by weak, parallel seismic reflectors. Depositional environment was probably never very deep (<200m?) considering the lateral and vertical facies transitions into sediments of shallow-water aspect (Winnall beds, Cyclops Member) and the intracratonic setting.

A regional unconformity is present at the base of the Winnall beds and Pertatataka Formation, or beneath thin sandstone (Gaylad Sandstone, Waldo Pedlar Member) or dolomite (upper marker cap dolomite) that locally conformably underlie the Pertatataka Formation. The Gaylad Sandstone comprises incised valley-filling fluvial, and shallow-marine sandstone and conglomerate, of northerly provenance, in the northeastern Amadeus Basin (Freeman and others, 1991). The Cyclops Member consists of shallow-marine, rhythmically-laminated and ripplemarked sandstone, also of northerly derivation (Korsch, 1986b). There is evidence for upward-shallowing at the top of the Pertatataka Formation, but the contact with the overlying shallow-water carbonate of the Julie Formation is usually abrupt.

Breaks in sedimentation or sequence boundaries within the Pertatataka-Julie succession are difficult to detect because of poor exposure, prevailing parallelism of seismic reflectors (Kennard and Lindsay, 1991) and the predominantly deep-water depositional environments of the Pertatataka Formation. Further work is needed to confirm the presence of a sequence boundary at the base of the

Julie Formation. No major breaks in sedimentation are shown by the chemostratigraphic profiles of carbon isotopic composition, which is essentially smoothly varying through most of the succession (see Part 2 of Guide). A sequence boundary may be expected to be associated with the temporary shallowing of depositional environment represented by the Cyclops Member. No chemostratigraphic profiles are yet available through this unit, but the steepest rates of change in $\partial^{13}C$ in the Ellery Creek section coincide with a sandy interval in the middle of the Pertatataka Formation that is a possible correlate of the Cyclops Member.

See also Amadeus Basin Stop 8.2.

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Appendix F

ACRITARCH BIOSTRATIGRAPHY OF THE EDIACARIAN OF THE CENTRALIAN SUPERBASIN

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Appendix F: Acritarch Biostratigraphy of the Ediacarian of the Centralian Superbasin

Kathleen Grey, School of Earth Sciences, Macquarie University, NSW

Introduction

Development of a Proterozoic chronostratigraphy, similar to that of the Phanerozoic, faces many limitations. Numerous fossils - including trace fossils, metazoan body fossils, metaphytes, stromatolites and microfossils (mainly prokaryotes and simple eukaryotes) have been recorded (Schopf and Klein, 1992), but their biostratigraphic usefulness throughout much of the Precambrian is generally restricted by sporadic distribution and low diversity (Knoll and Walter, 1992). Although the fossil record dates back to 3.4 Ga (Awramik and others, 1983; Schopf, 1983), few correlations have been possible. Trace fossils and metazoon body fossils are known only from just below the Precambrian/Cambrian boundary. Metaphytes (mainly consisting of carbonaceous films and up to 2.0 Ga old) are poorly known and still of limited value. Stromatolites (which date back to 3.4 Ga) have provided successful correlations, but are a difficult to work with because of taxonomic problems, their response to environmental influences, and their hybrid nature resulting from sediment accumulation by microbial communities. Most microfossils (also dating back to 3.4 Ga), consisting of prokaryotes and simple eukaryotes, show limited diversity and have conservative morphologies (summarised in Schopf and Klein, 1992).

The potential for biostratigraphic correlation improves considerably in the Neoproterozoic. In particular, the acritarchs, organic-walled microfossils (mostly phytoplankton remains) have considerable promise (Vidal and Knoll, 1983). They have short stratigraphic ranges, are abundant, are morphologically complex, and are widely distributed both geographically and in a variety of lithofacies (Knoll and Walter, 1992). Shale microfossils have been extracted by standard palynological techniques in Eurasia since the early 1950's, whereas researchers elsewhere have tended to concentrate on microfossils in chert (Mendelson and Schopf, 1992). Studies have gained momentum as the result of several recent advances, such as improved preparation techniques, attempts to reconcile the taxonomy of "chert" and "shale" microfossils, and the discovery of large acanthomorphs in the Ediacarian (Zang and Walter, 1989, 1992; Zang, in Shergold and others,1991). The potential of acritarchs to provide good biostratigraphic control is being investigated as part of the study of the Ediacarian of the Centralian Superbasin (Fig. 1).

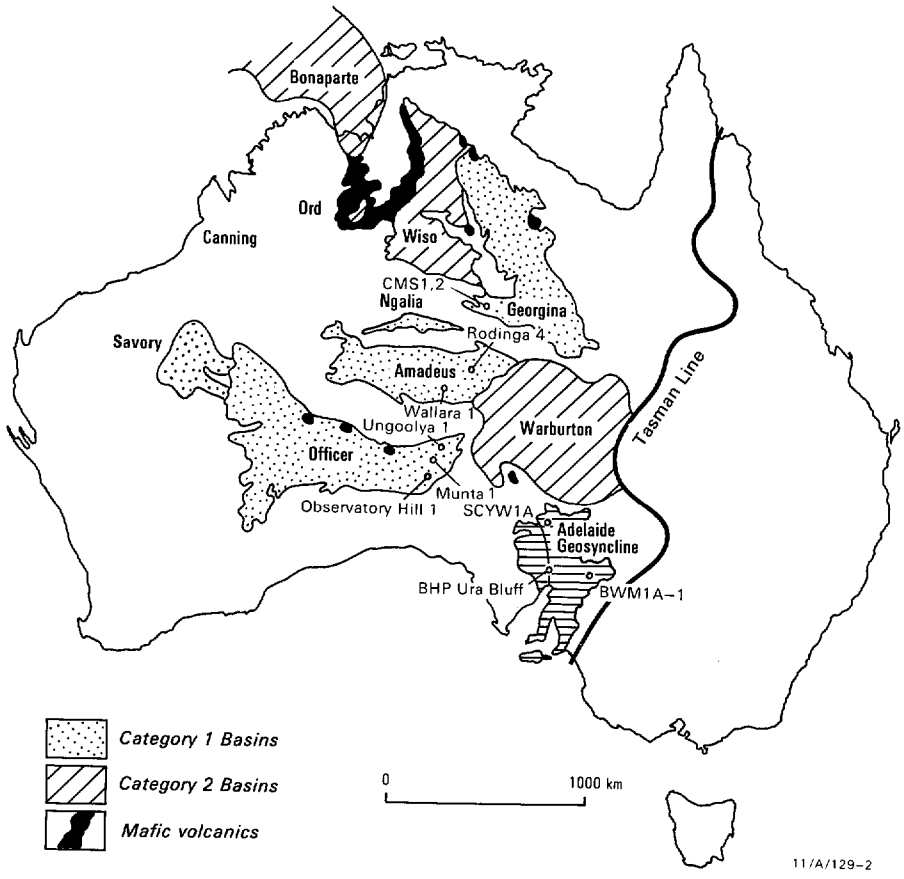

Figure 1. Intracratonic basins and the locations of major drillholes sampled for acritarchs and isotope-chemostratigraphic studies.

PREPARATION TECHNIQUES

Figure 2.

1. MECHANICAL DISAGGREGATION
   - *a. cleaning*
   - *b. bedding - plane splits*
   - *c. gentle crushing to pea size*

2. HCl DIGESTION - carbonate removal
   - *a. wash with distilled water* }
   - *b. gravity settling* } × > 4
   - *c. decant* }

3. HF DIGESTION - silicate removal
   - *a. wash with distilled water* }
   - *b. gravity settling* } × > 6
   - *c. decant* }

4. BOILING HCl TREATMENT - removal of clay sediments
   - *a. wash with distilled water* }
   - *b. gravity settling* } × > 4
   - *c. decant* }

5. KEROGEN STREW MOUNT

6. OXIDATION WITH CONCENTRATED HNO_3
   - *a. wash with distilled water* }
   - *b. gravity settling* } × > 4
   - *c. decant* }

7. FILTRATION - using selected filters
   - *a. 25 µm* } *sizes most commonly*
   - *b. 10 µm* } *employed*

8. SWIRLING TO SEPERATE HEAVY MINERALS

9. MAKE STREW MOUNTS OF VARIOUS RESIDUES

Acritarch palaeobiology

The term acritarch was proposed by Evitt (1963) and is derived from the Greek meaning "of uncertain origin". An acritarch has been defined as a "unicellular, or apparently unicellular, resistant-walled microscopic organic body of unknown or uncertain biological relationship and characterised by varied sculpture, some being spiny and others smooth" (Bates and Jackson, 1980).

Acritarchs are a polyphyletic group, and their classification is an artificial one. Nomenclature follows the International Code of Botanical Nomenclature. In many cases their affinities cannot be determined, although most are probably a resting cyst stage. Some are related to the prasinophyte green algae, while others have morphologies reminiscent of hystrichospheres (dinoflagellate resting cysts), and this is consistent with the identification of dinosterane (Summons and others, 1997).

Several morphological types can be recognised (Mendelson and Schopf, 1992a): acanthomorphs, herkomorphs, netromorphs, oomorphs, polygonomorphs, prismatomorphs, pteromorphs, and a variety of miscellaneous structures such as filaments - which are sometimes classified separately, and sometimes included with the acritarchs.. Some of these groups do not occur until the Phanerozoic, but others have well-established Proterozoic records (Fig. 4). A detailed summary is given by Schopf and Mendelson (1992b).

Phanerozoic rocks may contain thousands of specimens, but Proterozoic material is rarely so abundant. Nevertheless, diverse assemblages have been recorded from about 24 sequences around the world (Knoll and Walter, 1992). Records of Australian Ediacarian assemblages are consistent with the reported distributions elsewhere (Damassa and Knoll, 1986; Zang and Walter, 1989; Jenkins and others, 1992). Many Proterozoic assemblages contain morphologically complex forms, and small spiny forms are common throughout the Neoproterozoic. However, only a few large spiny forms were reported prior to Zang's study of Amadeus Basin assemblages (Zang, 1988; Zang and Walter, 1989, 1992). These include some from the upper Sinian Doushantuo Formation of southern China (Yin and Li, 1978; Yin, 1987; Awramik and others, 1985, Zhang, 1984)), from the Qingbeikouan and Sinian Systems (Xing and Liu, 1982), and a variety of other occurrences (Jankauskas, 1982; Timofeev and others, 1976; Butterfield and Rainbird, 1988; Butterfield and others, 1988; Knoll, 1984; Knoll and Ohta, 1988; Vidal and Ford, 1985; Pyatiletov, 1980; Allison and Awramik, 1989; Jankauskas, 1989; Knoll and others, 1991). Other However, none of these assemblages show the diversity found in the Ediacarian assemblages in the Amadeus Basin, where several new species of extremely large acanthomorph acritarchs occur (Zang and Walter, 1992). Giant acanthomorphs are apparently

restricted to the interval extending from post-Varanginian to the Proterozoic/Cambrian boundary. This restricted distribution can be demonstrated by comparing acritarch diameters in the Pertatataka Formation with those in the older Bitter Springs Formation and the younger Tempe Formation. The Pertatataka Formation shows a continuous spread of diameters ranging up to 425 mm. By contrast both younger and older forms are less than 100 mm in diameter, with the exception of a few large sphaeromorphs ranging in diameter from 300 to 325 mm in diameter.

Preparation techniques

Preparation techniques are critical. Zang (1989) demonstrated that samples needed gentle treatment (e.g. minimal crushing and no centrifuging) to extract large, brittle specimens. By processing between 50 and 150 gm of material per sample he obtained rich assemblages, but this approach has limitations for biostratigraphic and sedimentological studies. Techniques have now been refined to cope with smaller samples, normally about 25 gm, and methods are summarised in Figure 2. Heavy liquid separation techniques were investigated, but were discontinued because pyrite embedded in the tissue prevents flotation. Filtration is also critical, and after considerable experimentation filter cloths of 25 and 10 um were found to produce the best results. In order to monitor possible contaminants a series of "control samples" (usually granitic rocks) were processed with sample batches.

Acritarch biostratigraphy

Zang's preliminary studies suggested that species in the Pertatataka Formation had good biostratigraphic potential. By using Zang's thesis data (Zang, 1988), a preliminary range chart has been constructed (Fig. 3). Although the distributions of some taxa require further confirmation, there does seem to be scope for biostratigraphic subdivision. This is one of the aspects being investigated further in the present study.

Preliminary Results

The current project is designed to investigate the distribution of acritarchs in Centralian Superbasin. Material has been collected from drillholes and some field sections from the Amadeus, Georgina, Officer, Smithton and Savory Basins, and the Adelaide Geosyncline. A list of drillholes and localities sampled is shown in Fig. 1. A total of about 1000 samples have been processed from these areas, and yields have been moderately high. More than 50% of samples contain at least some acritarchs, but not all of these are abundant or well preserved. Some batches of samples have produced extremely well-

RANGE CHART

RODINGA 4

	81 - 83	63 - 65	58.27 - 58.55	55.64 - 55.95	53.86 - 54.14	52.16 - 52.50	44.61 - 44.92	48.38 - 48.67	48 - 49	39.35 - 39.55	37.24 - 37.51	35.00 - 35.30	34.50 - 34.76	34 - 35	32.28 - 32.50	32 - 33	28.00 - 28.25	26.44 - 26.74	23.80 - 25.35	23.02 - 23.20
	007	005	W20	W19	W18	W17	W16	W15	004	W11	W10	W09	W08	003	W07	002	W05	W04	W02	W01
Octoedryxium truncatum	x	-	-	-	-	-	-	-	-	-	-	x								
Comasphaeridium hispidum		x	-	-	-	-	x	x	-	-	-	-	-	x						
Alicesphaeridium medusoidum			x	x	x	x	x	-	-	x										
Baltisphaeridium rarusente				x	-	-	-	x	x	-	-	-	-	-	-	x				
Gorgonisphaeridium curtoblastum					x	-	x	x	-	-	-	x	x	x						
Gorgonisphaeridium pulchrum					x	-	-	x	-	-	-	x								
Trachyhystrichosphaera pedanaspina					x	x	-	-	-	-	-	-	-	-	x	-	-	x		
Trachyhystrichosphaera triangula					x	-	-	x	-	-	-	-	x	-	-	x				
Indigestosphaeridium adiakritum						x	-	-	x	-	-	-	-	x	-	x			x	
Schizofusa rodingia						x	-	-	-	-	-	-	-	-	-	x				x
Trachyhystrichosphaera obtusaspina						x	x	x	-	-	-	-	-	x						
Trachyhystrichosphaera spargaspina						x	-	-	-	x	-	-	-	x	-	-	-	x		x
Trachyhystrichosphaera tribulosa						x	x	x	-	x										
Cymatiosphaeroides? proceropilum							x	x	x											
Cymatiosphaeroides publicopilum							x	x	x											
Schizofusa sp.							x	-	-	x										x
Skiagia prodigiosa							x													x
Skiagia spatiosa							x	x												
Solisphaeridium? araithekum							x													x
Somphosphaeridium amphilaphum							x	-	x	-	-	-	x	x						
Baltisphaeridium plerusente								x	-	x	-	-	x							
Filisphaeridium nassokarum								x	-	-	-	x								
Trachyhystrichosphaera anelpum								x	-	x										x
Variomargosphaeridium litoschum								x											x	x
Variomargosphaeridium multiramum								x	-	-	-	-	-	-	-	-	x			
Cymatiosphaeroides dilutopilum									x	-	-	-	x						x	x
Dictyotidium ambonum									x	-	-	-	-	x					x	x
Goniosphaeridium crebrum									x	-	-	-	x							
Rhopaliophora sp.										x										x
Cymatiosphaeroides pilatopilum											x	-	-	-	-	x			x	
Monosozosphaeridium ekmelum											x	-	x						x	
Polyhedrosphaeridium echinatum												x							x	x
Comasphaeridium crinitum													x							
Gorgonisphaeridium inconstanum													x	-	-	x				
Goniosphaeridium cratum														x						x
Sticcasphaeridium incertum														x						x
Hocosphaeridium scaberfacium																			x	x
Tasmanites? fistulosum																			x	x
Multifronsphaeridium pelorium																				x

Based on data from Zang (1989, unpublished)

11/NT/160

Figure 3. Range chart for achritarchs from the Rodinga 4 core (based on data from Zang, 1989 and unpublished data).

ACRITARCHS

INFORMAL MORPHOLOGICAL GROUPINGS AND NUMBER OF REPORTED TAXA

Age	999	949	899	849	799	749	699	649	599	549
(Ma)	-950	-900	-850	-800	-750	-700	-650	-600	-550	-530
								EDIACHARIAN		
Spaeromorphs	A	A	A	A	A	A	A	A	A	A
Acanthomorphs	6	1	6	-	7	2	4	15	19	321
Herkomorphs	-	-	-	-	1	-	2	-	1	37
Netromorphs	1	-	-	-	-	-	1	1	1	27
Oomorphs	-	-	-	-	-	-	-	-	-	16
Polygonomorphs	-	-	-	-	-	-	-	-	-	19
Prismatomorphs	1	-	-	-	1	3	-	6	-	9
Pteromorphs	-	4	-	-	2	-	-	-	1	9

Figure 4. Informal morphological groupings and number of reported taxa of achritarchs.

preserved, diverse assemblages, including examples of the large acanthomorph microflora. To date efforts have been mainly concentrated on preparation, and detailed systematic logging of samples has only just begun. The results outlined below are based only on quick scans of samples and are of necessity of a very preliminary nature. Nevertheless, the material has demonstrated good potential for future biostratigraphic correlations once taxonomic studies and systematic analysis is completed.

Amadeus Basin

In the Amadeus Basin two drillholes and several field sections were sampled. Rodinga 4 drillhole was resampled to provide comparative material for further studies. Most samples are from the Pertatataka Formation. Zang's distribution patterns can be confirmed from preliminary study of prepared slides. Wallara 1 drillhole contains intervals of well-preserved acritarchs and abundant amorphous tissue. Most acritarchs are simple leiospheres, but some samples contain small, spiny forms. No large spiny forms have been observed, and this supports Calver's interpretation, based on isotope geochemistry, that there is no overlap between Rodinga 4 and Wallara 1 (see Appendix A). Several field sections were sampled, but most have not yet been examined microscopically. Samples from Mount Connor contained a few fragments of amorphous tissue, but were otherwise barren. Samples from the vicinity of Dead Bullock Plain were also barren.

Georgina Basin

Samples were collected from Central Mount Stuart 1 and 2 drillholes (these are labelled Mount Skinner in the core shed), and from Huckitta 3, 4, and 7. Material was collected from the Central Mount Stuart, Elkera, Grant Bluff and Elyuah Formations.

Well-preserved assemblages occur in CMS 1 and 2. No large spiny acritarchs have been observed, but leiospheres - both large and small - are common. Small, spiny acritarchs occur sporadically throughout the drillhole. Filaments are common in many samples, particularly in the Grant Bluff Formation, and many samples contain amorphous tissue fragments. The Huckitta samples have not yet been scanned in any detail, and organic material appears sparse. Acritarchs are present in one or two samples.

Savory Basin

A few field samples were collected from the Savory Basin, but results are not yet available.

Officer Basin (Western Australia)

Material from Yowalga 2, from the Babbagoola Formation, below the Table Hill Volcanics, yielded well-preserved assemblages containing filaments, leiospheres (mainly thin-walled, clusters of spheres, and a distinctive, large, thick-walled multi-folded leiosphere. No spiny acritarchs have been observed so far. Similar specimens were identified in several drillholes in preparations housed with the Geological Survey of WA. The large folded species is present in Hussar 1, Lungkarta 1, and Yowalga 3. No spiny acritarchs have been observed so far. Material collected from Western Mining NJD 1 drillhole is still being processed.

Officer Basin (South Australia)

Sampling concentrated on the Rodda Beds in Munta 1, Observatory Hill 1 and Ungoolya drillholes (Fig. 5).

Ungoolya material is still being processed. Jenkins and others, 1992, recorded large spiny acritarchs in this drillhole, and our more comprehensive sampling is expected to provide a fairly continuous section through the large spiny acritarch assemblage.

There are still some gaps in sample preparation in Munta 1, but from material examined so far filaments and amorphous tissue are common to a depth of 1000 m, but acritarchs are rare or absent. Large spiny acritarchs are present from depths of 1200 to 1810 m. Preservation is variable, but some samples between 1200 and 1300 are extremely well preserved, and appear to have many taxa in common with Rodinga 4 in the Amadeus Basin.

The large spiny acritarch assemblage has also been encountered in the upper part of Observatory Hill 1, down to about 258 m. Below this samples are apparently barren, and the first productive sample occurs about 100 m above the Acraman Impact Ejecta Horizon. Material is especially well preserved near the top of the drillhole. Some taxa appear similar to those in Munta 1 and Rodinga 1, but many others appear new, and have complex morphologies.

Adelaide Geosyncline

Samples were collected from SCY W1A, BHP Uaroo Bluff, and BWM 1. No proper results are available from the last two drillholes, but yields were low - so far only one sample (from BWM) appears to contain leiospheres. Material was also collected from Brachina and Bunyeroo Gorges.

SCY W1A - previously examined by Damassa and Knoll (1986), and sampled at more frequent intervals by us, contains well preserved leiospheres and small spiny acritarchs in some parts of the

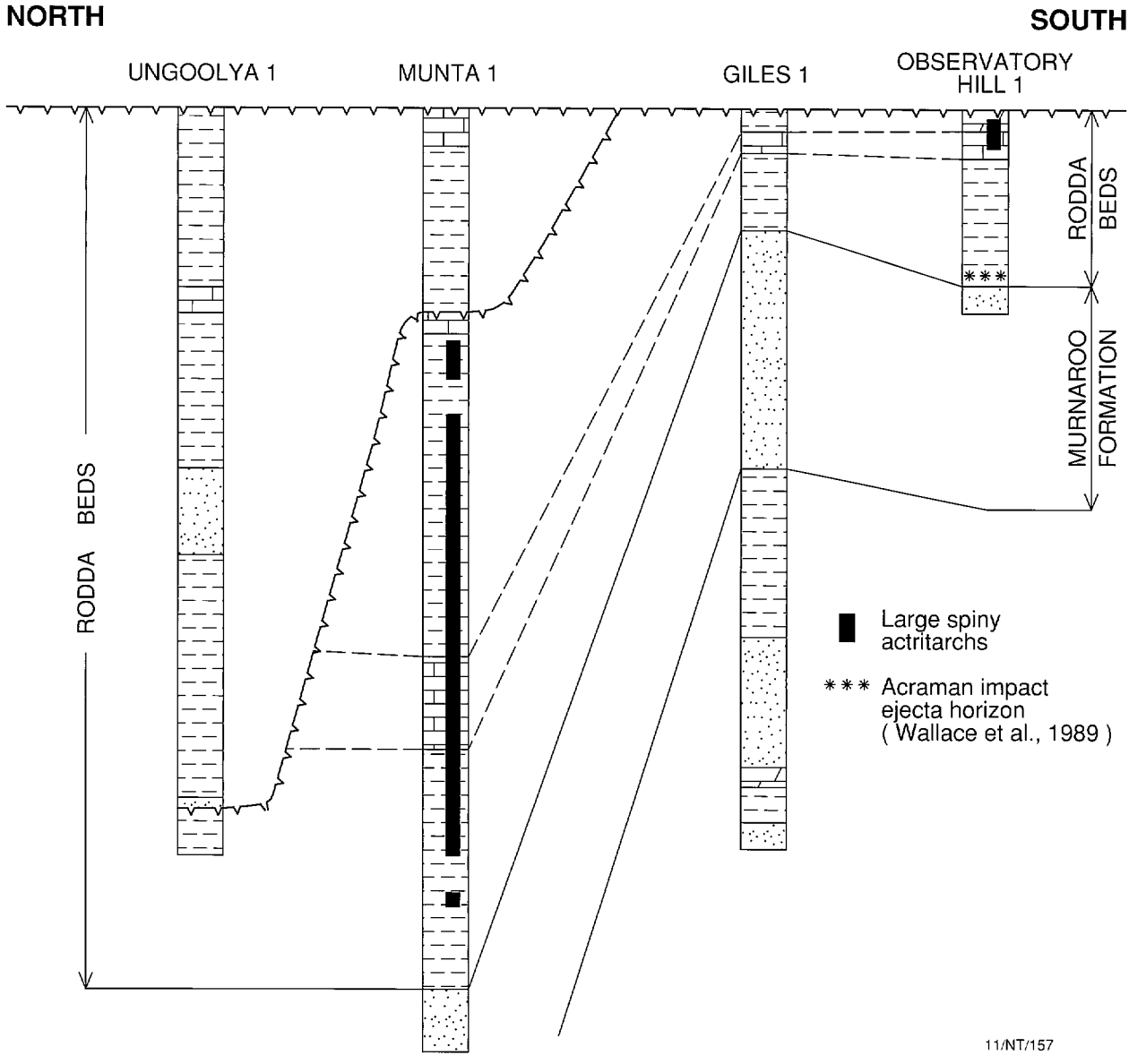

Figure 5. Stratigraphy of Rodda Beds and associated units in the Officer Basin (adapted from Sukanta and others, 1991) and the occurrences of spiny achritarchs. Datum - basal Cambrian unconformity.

drillhole. One or two samples may contain large spiny acritarchs, but these observations are of a very preliminary nature and are yet to be confirmed.

Samples from Brachina and Bunyeroo Gorges have yielded very little organic residue, and what there is is poorly preserved. Poorly preserved acritarchs do occur in the Brachina Formation in Bunyeroo Gorge, but their nature requires further investigation.

Tasmania

Samples from Forest 1 and field localities in the Smithton Basin have yet to be prepared.

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Appendix G

Early Cambrian Carbonates in the Flinders Ranges

David I. Gravestock, Department of Mines and Energy, Adelaide, South Australia

Appendix G: Early Cambrian Carbonates in the Flinders Ranges

D.I. Gravestock, Department of Mines and Energy, Adelaide, South Australia

Palaeogeographic Setting

Ediacaran and Early Cambrian sediments in the Flinders Ranges were deposited in a meridional seaway on the eastern seaboard of the Gawler Craton (Fig. 1). Despite obvious differences in the type of sediment produced due to the sudden appearances of skeletonised metazoans and calcified cyanobacterial/microbial organisms, Ediacaran as well as Early Cambrian rocks share some fundamental attributes. Both display the development of shallow subtidal platform, platform-slope and basinal facies, and both sequences are punctuated by recurrent episodes of submarine erosion. Erosional episodes have resulted in downcutting on a scale ranging from shallow channels 10 m in depth to spectacular canyons 1 km or more deep. Examples from the Ediacaran are Wonoka Formation (Coats, 1965; von der Borch and others, 1982), Rawnsley Quartzite (Jenkins and others, 1983). Cambrian examples are Uratanna Formation (Daily, 1973), Wilkawillina Limestone (Daily, 1976; Roche, 1986) and upper Ajax Limestone (Gravestock, unpublished data).

Von der Borch and Grady (1986) have suggested that major transgressive and regressive sequences from the Marinoan to Early Cambrian may be fundamentally controlled by rates of oceanic crustal accretion and seafloor spreading beyond the Adelaide Geosyncline. However, the numerous disconformities resulting from erosional episodes (one of these being the local Cambrian-Precambrian boundary) are responses on the platform itself, to more localised syndepositional tectonic and diapiric activity. These higher frequency events are superimposed on the broader plate tectonic controls. Not surprisingly, the Ediacaran and Early Cambrian sequences share several morphotectonic features:

1. a broad, shallow marine platform covered by the Flinders Sea;

2. almost coincident, superimposed platform margins in the northern Flinders Ranges, being the sites of inception of submarine canyons (Ediacaran) and slope megabreccias (Early Cambrian). Drillcore data from the Warburton Basin to the northeast confirm the presence of Middle to Late Cambrian carbonate shelf and slope facies northwest of Moomba, and Early Ordovician sediments are also present, escaping the Delamerian Orogeny but being variably affected by the Banambran Orogeny (Gatehouse, 1986; Gatehouse and Cooper, 1986; Cooper, 1986).

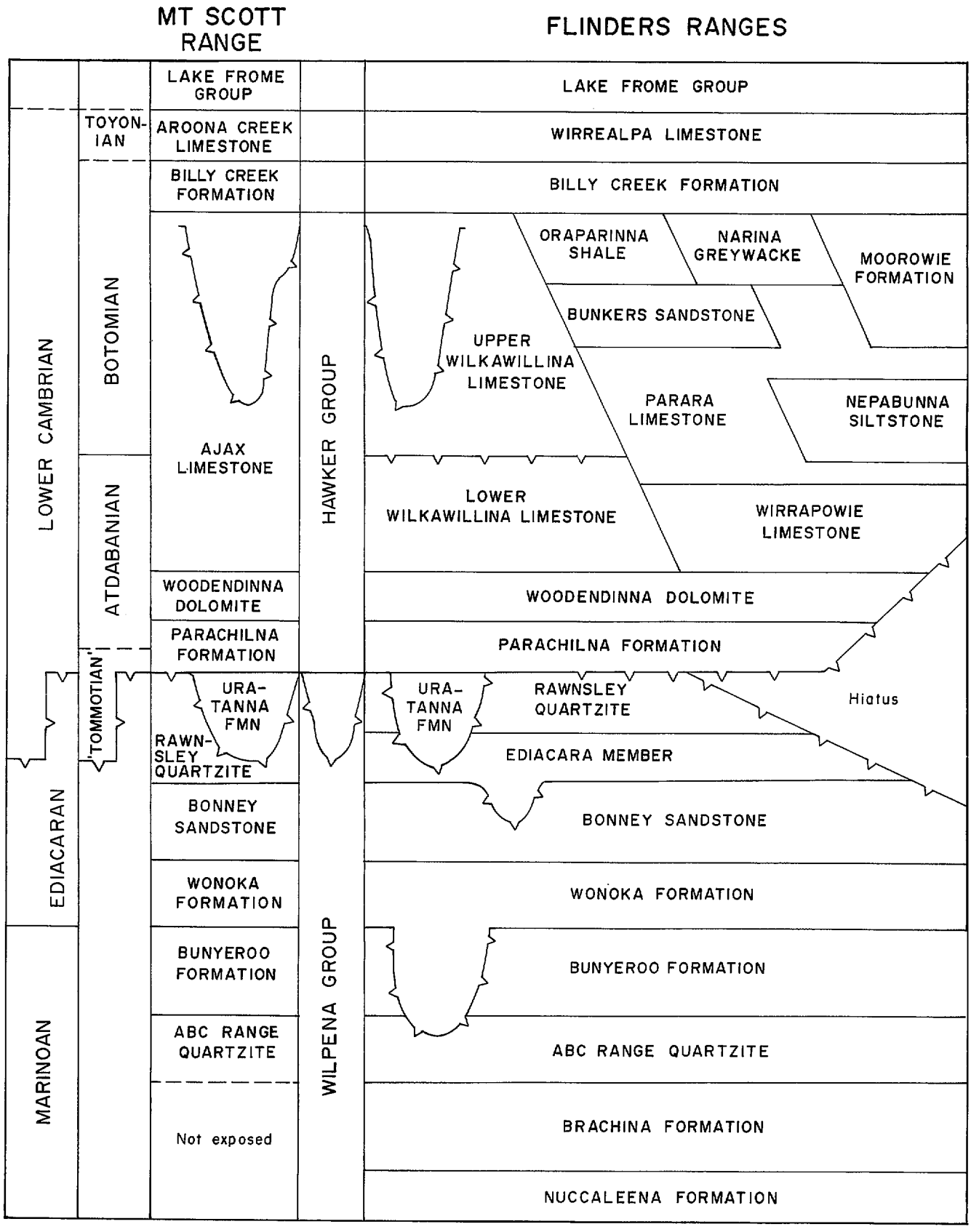

Figure 1 Ediacaran and Early Cambrian succession, Flinders Ranges.

SADME S20145

The Warburton Basin is suggested here to have been a major Early Cambrian (? and Adelaidean) depocentre but deeper drilling is required for proof).

3. During the Early Cambrian a somewhat narrower carbonate shelf developed on the southeastern seabord of the Gawler Craton. Shelf, ramp and basinal facies are proposed for Hawker Group equivalents (Alexander and Gravestock, 1990), succeeded by thick siliciclastics of the Kanmantoo Group which poured from the Gawler Craton into a presumed back-arc basin. Wonoka Formation canyons in the southern Flinders Ranges (Pichi Richi Pass) may have transported detritus to the southeast (Preiss, 1987, 1990) into the same basin.

A consistent palaeogeographic scheme is thus beginning to emerge for the Ediacaran and Early Cambrian platform and periplatform sequences in the Flinders Ranges, not only in their similar overall styles of sedimentation but also in their responses to recurrent tectonic activity, which sets them apart from older Precambrian sediments of the Adelaide Geosyncline.

Hawker Group Carbonates

General aspects of the basal Cambrian Uratanna and Parachilna Formations have been discussed by Daily (1976), and it is proposed here to provide minor background data on the older Cambrian carbonates mapped on the PARACHILNA and COPLEY 1:250,000 sheets as Wilkawillina and Ajax Limestone (Daily, 1956). There are excellent grounds for dividing the Wilkawillina Limestone into two units separated in many areas by a thin but widespread and distinctive exposure surface (Daily, 1976; Clarke, 1986). The lower unit contains shelly fossils assigned by Daily (1956) to his Faunal Assemblages 1 and 2. The assemblages are dominated by Archaeocyathat which have been studied by Gravestock (1984) who correlates them with the upper Bazaikha and Kameshky Horizons of the USSR Altay-Sayan region. However, much of what has been mapped as lower Wilkawillina Limestone is unfossiliferous and can justifiably be separated into Woodendinna Dolomite and Wirrapowie Limestone (Haslett, 1975). These formations range from 25 m to over 500 m in thickness and consist of often dolomitized oolitic and stromatolitic limestone with abundant desiccation cracks and quartz sandstone interbeds (Woodendinna), overlain by fine grained silty limestone which is stromatolitic, less commonly oolitic, lacks sand and shows much less evidence of subaerial exposure (Wirrapowie). In the Mount Scott Range 65 m to 110 m of Woodendinna Dolomite can be separated from the Ajax Limestone but Wirrapowie Limestone is absent. Woodendinna Dolomite conformably overlies the Parachilna Formation and is overlain conformably by Wirrapowie, Wilkawillina or Ajax Limestone depending on locality.

The lower Wilkawillina Limestone is subtidal, massive to nodular but principally well-bedded skeletal packstone conspicuously rich in archaeocyaths, brachiopods and the tannuolinid *Micrina etheridgei* (Tate). The first shelly trilobite remains appear in the upper part of the unit (upper Faunal Assemblage 2). The sediments accumulated in generally wide open platform settings with only local development of basinal carbonates. bioherms, though locally abundant are subordinate to high energy packstone facies and rarely coalesce into massive complexes. Principal biohermal constituents are archaeocyaths *Girvanella, Renalcis* and *Epiphyton* in varying proportions. Hexactinellid spicules are abundant in certain bioherms.

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Appendix H

Appendix H: Trace fossils

M. R. Walter, Macquarie University, Sydney

Appendix H: Trace fossils

M. R. Walter, Macquarie University, Sydney

The following comments have been extracted from Walter and others (1989). Subsequent unpublished work by these and other authors will be discussed during the field trip.

Four successive assemblages of trace fossils are recognisable in the Neoproterozoic to Early Cambrian of Australia. These are informally designated as Assemblages 0-3, of which the older three are Ediacarian, and the fourth is Early Cambrian. The assemblages are defined largely in the Amadeus Basin and the closely adjacent Georgina Basin; between these two basins lithostratigraphic correlations can be established with confidence (Preiss *and others*, 1978; Walter, 1980). One assemblage (2) is defined from the Adelaide Geosyncline, where it occurs with the Ediacara fauna; lithostratigraphic correlations agree with those based on the occurrence of the Ediacara fauna in the Amadeus Basin and allow this assemblage to be placed in context.

"Assemblage" O

The Elyuah Formation of the Georgina Basin, and the lower Pertatataka Formation of the Amadeus Basin, lack definite trace fossils. Furthermore, no trace fossils are known from any correlative rocks or from any pre-Ediacarian rocks in Australia. Potentially fossiliferous rocks are abundant and have been searched, in some examples, repeatedly over several decades (particularly in the Adelaide Geosyncline but also in the central Australian basins and recently in the east Kimberley region of Western Australia). The fossil *Bunyerichnus dalgarnoi* Glaessner occurs at this level in the Adelaide Geosyncline.

Assemblage 1

At present this is represented by only the occurrence of *Planolites ballandus* Webby in the lower Elkera Formation, and *Planolites* and *Palaeophycus* in the Grant Bluff Formation of the Georgina Basin. No other trace fossils are known from correlative units in Australia. These occurrences predate the Ediacara fauna.

Assemblage 2

Glaessner (1969) described six different forms of trace fossils that occur with the Ediacara fauna in the Adelaide Geosyncline, and noted in addition that there are "less common and more obscure kinds of trails".

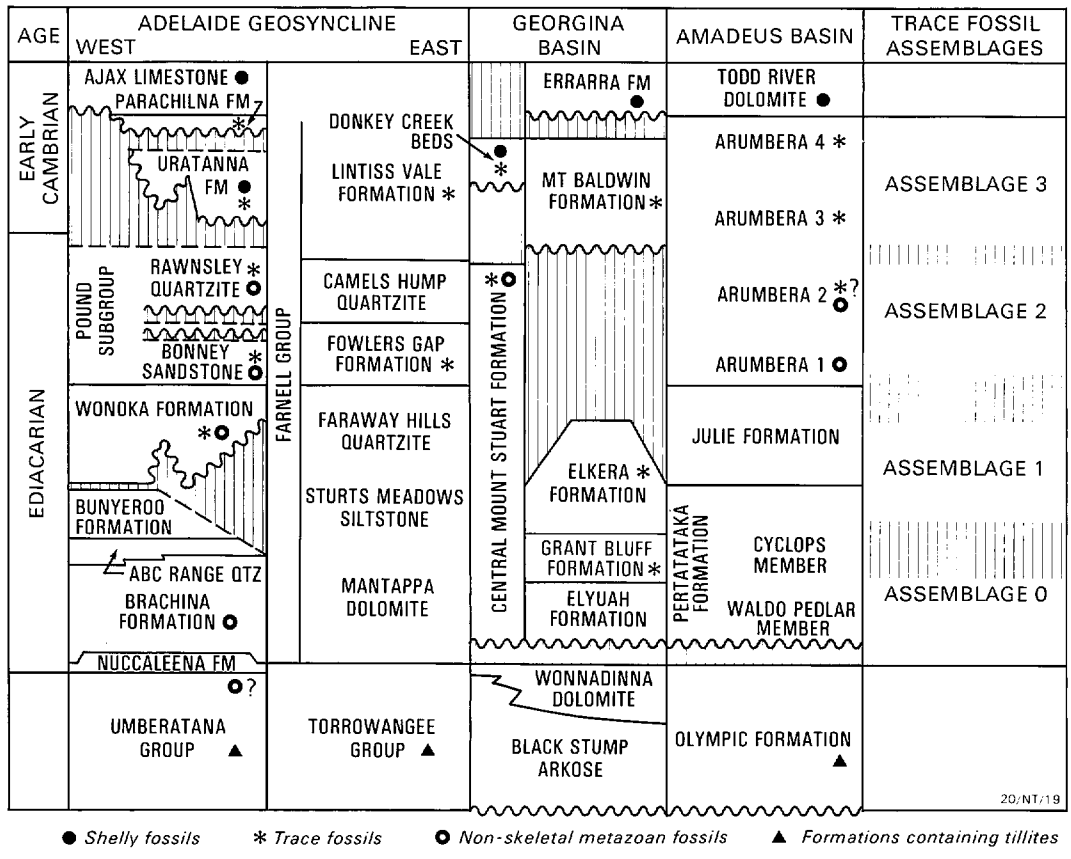

Figure 1. Distribution of trace and body fossils in the Neoproterozoic and Early Cambrian (After Walter and others, 1989).

No trace fossils are known from Arumbera Sandstone 1 or 2, despite the fact these units contain metazoan body fossils. Trace fossils have been reported from time to time but numerous subsequent searches have all failed to confirm the reports. Horizontal trails and steeply inclined burrows have been recorded from the Namatjira Formation, which was suggested to be a correlative of Arumbera 2 (Bradshaw, 1988). However, more recent observations indicate that the Namatjira Formation disconformably overlies the Arumbera Sandstone. Short vertical burrows (*Hormosiroidea*) are described below from the upper Central Mount Stuart Formation, a correlative in the Georgina Basin of the lower Arumbera Sandstone.

The medusae *Hallidaya brueri* Wade and *Skinnera brooksi* Wade occur in the upper Central Mount Stuart Formation of the Georgina Basin (Wade, 1969; Walter, 1980), and *H. brueri* has also been found in the lower Arumbera Sandstone (Wade, 1969; see Glaessner and Walter, 1975 and Kirschvink, 1978a, b, for stratigraphic information on this occurrence). The available data suggest that this occurrence is in Arumbera 2. The frond-like fossil *Charniodiscus* sp. is known from a loose block in the lower Arumbera Formation (Glaessner, 1969, 1984; Jenkins and Gehling, 1978). A medusoid fossil figured and described as *Kullingia* aff. *concentrica* by Føyn and Glaessner (1979) is reported to occur about 250 m above the base of the Arumbera Sandstone and 63 m below a "glauconitic zone" in the Laura Creek area 22 km WSW of Alice Springs. This puts it at very close to the same stratigraphic level as *H. brueri* in the same area. However, it is said by the collector to have occurred with abundant trace fossils, but no trace fossils were found this low in the formation during the study of the dubiofossil *Arumberia banksi* (Glaessner and Walter, 1975) at Laura Creek.

Assemblage 3

Currently only one assemblage is distinguished in Arumbera 3 and 4, despite the systematic, limited stratigraphic distribution of many taxa, because it is at least as likely the distribution is a palaeo-environmental effect as it is the result of evolution or the progressive occupation of new habitats. Occurrences of this assemblage are common in the northeastern Amadeus Basin.

Assemblage 3 is distinguished by its abundance and diversity of taxa - 36 ichnofossil taxa have been recognised. This assemblage is characterised by the presence of deep vertical burrows, and the presence of many different forms of complex subhorizontal burrows, many of which are large (e.g. *Plagiogmus arcuatus, Treptichnus* sp. and *Didymaulichnus miettensis*). *Rusophycus*-like forms are present but extremely rare (in central Australia). This assemblage is con-

spicuous in the field, and readily recognised, in contrast to all older assemblages.

Assemblage 3 also occurs in the Uratanna and Parachilna Formations, and the Lintiss Vale Formation, of the Adelaide Geosyncline (although some of the more characterisitic elements are missing in the latter example). *Rusophycus* may be more abundant in the Uratanna Formation than it is in central Australia.

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Appendix I

Preliminary Results from China

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Appendix I: Preliminary Results from China

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A workshop and field program carried out on the Yangtze Platform of southern China (Fig. 1) as part of IGCP Projects 303 and 320 in September 1993 has allowed a first preliminary attempt at sequence correlation between the Neoproterozoic and Cambrian successions of southern China and central Australia. The following is a brief summary of a report compiled by Lindsay (1993) which attempts to bring together the present state of knowledge.

The Yangtze Platform and associated Sichuan Basin compare well morphologically and structurally with the Amadeus Basin. The northern margins of both basins are truncated abruptly by major overthrusts adjacent to which is a series of deep sub-basins. Southward both basins shallow to a broad platformal setting highlighted by broad later stage anticlinal structures. Thus similar deposition patterns could be expected.

In contrast to the Amadeus Basin, deep weathering profiles and heavy soil and vegetation cover make identification of sequences difficult in the Neoproterozoic and Cambrian sections of the Yangtze Platform. In spite of this some major boundaries can be identified in the Neoproterozoic in clastic and mixed carbonate associations and correlated with similar units within the Amadeus Basin of central Australia. The latest Neoproterozoic is difficult to interpret because of the uniformity of facies in the thick platform carbonate successions (Fig. 1).

The earliest Cambrian rocks are more readily interpreted in terms of their sequence stratigraphy in large part because they are clastic or mixed carbonate-clastic units and can be identified as shoaling upward cycles. The four latest Neoproterozoic and earliest Cambrian sequences of the Yangtze Platform do not correlate simply with the first the Amadeus Basin section (Fig. 2). While there are broad similarities in the two sections a sequence correlation is only partially successful. The results suggest that some deposition sequences reflect events that are of global significance but that other surfaces are of local significance only. However, interpretation of the Yangtze Platform succession can not be carried out completely without resorting to well-logs, cored sections and seismic data, and there is little doubt that some sequences are as yet unrecognised. Similarly, time gaps in the Amadeus Basin succession suggest that other sequences are yet to be identified.

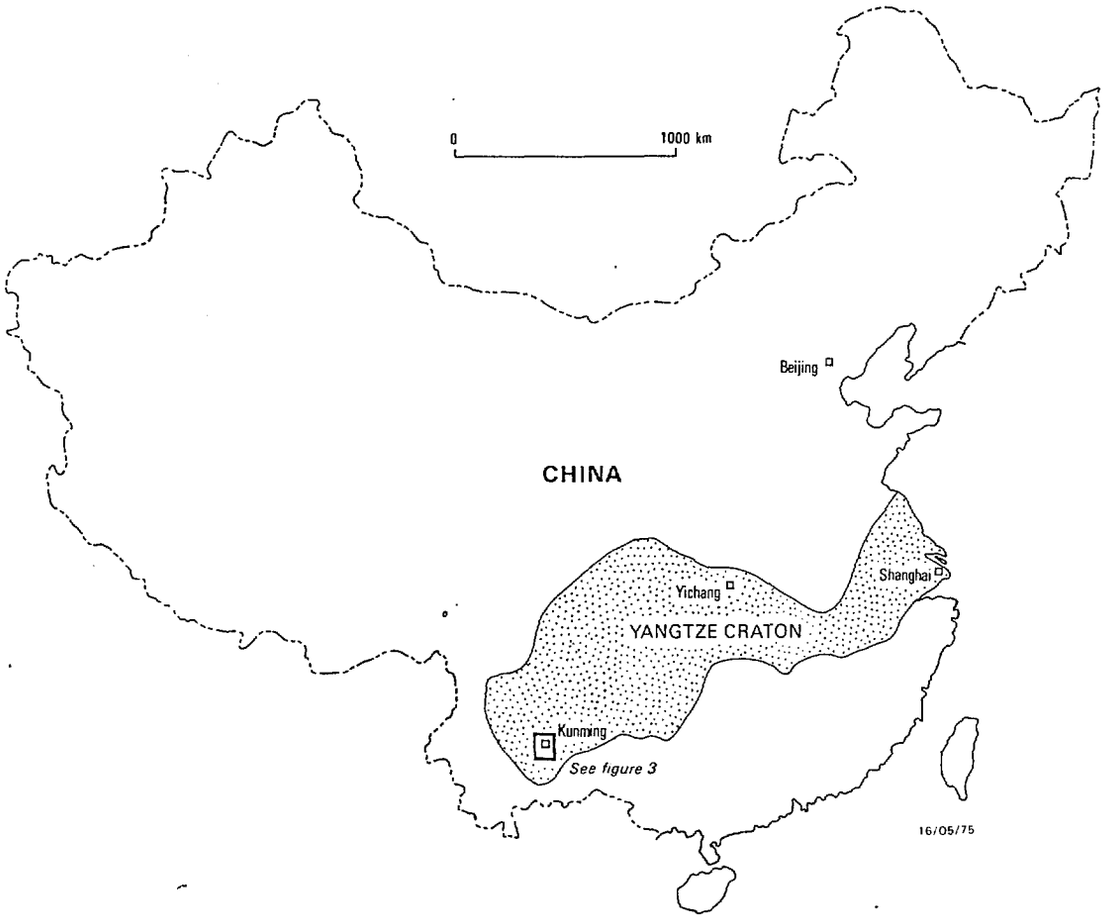

Figure 1. Map of China showing the location of the Yangtze Craton.

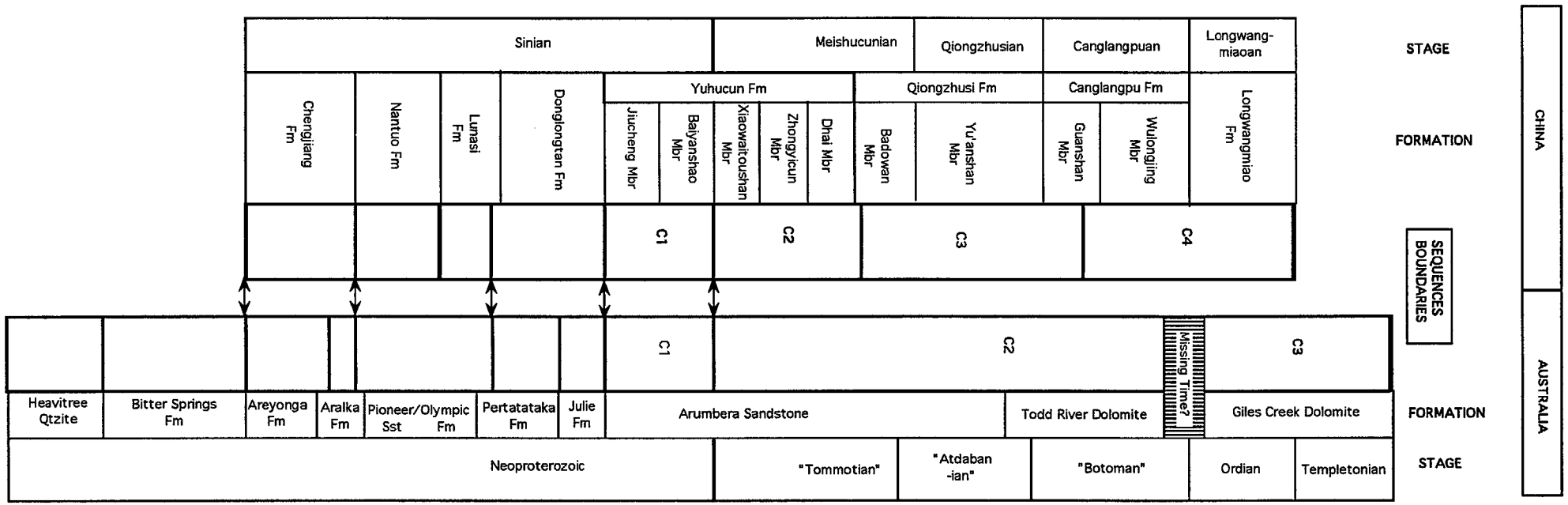

Figure 2. A direct sequence stratigraphic comparison of composite sections for the Yangtze Platform and the Amadeus Basin. Heavy lines down the centre columns show sequence boundaries identified at each section and suggest a possible correlation. The missing time above the Todd River Dolomite may represent a new sequence which includes the Chandler Formation which is present in the section further west.

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