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SEDIMENTOLOGY OF THE LATE PRECAMBRIAN MUNDALLIO SUBGROUP : A CLASTIC-CARBONATE (DOLOMITE, MAGNESITE) SEQUENCE IN THE MT. LOFTY AND FLINDERS RANGES, SOUTH AUSTRALIA.

(VOLUME I)

by

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VOLUME I

SUMMARY

During deposition of the mixed carbonate-clastic sequence of the Mundallio Subgroup, the "Adelaide Geosyncline" was a very shallow, elongate sedimentary basin, flanked to the west and east by older Precambrian basement.

In much of the southern and northern Flinders Ranges, clastic deposition predominated in the lower Mundallio Subgroup. In the north, alternating development of shallow mudflats and sandflats (Nankabunyana Formation) depended on the interplay between the sediment supply and winnowing processes, while dolomite mudstones were locally deposited in the shallowest areas. In the eastern half of the Willouran Ranges, massive shales were deposited as the environment remained persistently below wave base (Camel Flat Shale), but a renewed sand influx led to deposition of the Tilterana Sandstone. In the southern Flinders Ranges, terrigenous clay and silt were deposited on submergent mudflats which shallowed into intermittently exposed dolomite mudflats (Nathaltee Formation). Dolomite mudflats were a more persistent feature in areas more distal from the terrigenous source, and sometimes contained isolated, ephemeral lakes which were sites of magnesite deposition (Yadlamalka Formation). Dolomite and magnesite mudstone deposition of the Yadlamalka Formation became widespread in the northern and southern Flinders Ranges in the upper Mundallio Subgroup, as shallowing and retreat of the basin margin led to the formation of semi-isolated lakes, separated and enclosed by exposed carbonate mudflats. The clastics deposited in association with these carbonate mudstones consisted largely of sand sized detritus, probably derived from the reworking of aeolian deposits. In the eastern Willouran Ranges, the greater influx of sand and the slightly deeper, largely submergent environments, led to the deposition of the sandstones, dolomites and siltstones of the Mirra Formation.

Because of little clastic influx into the northern Mt. Lofty Ranges, shallow to occasionally exposed environments were largely sites of dolomite

deposition (Skillogalee Dolomite). To the south, shales were deposited in slightly deeper environments (Woolshed Flat Shale), although local dolomite deposition occurred in the Adelaide region (Castambul Formation, Montacute Dolomite). In the uppermost part of the subgroup, the area of shale deposition extended northward, encroaching over the dolomite mudflats of the upper Skillogalee Dolomite.

Dolomite, occurring largely as mudstones, is the major carbonate mineral present in the Mundallio Subgroup, but magnesite is also widespread. Limestones are not present. The carbonates experienced minor replacement by early diagenetic chert, initially precipitated as both crystalline and amorphous phases. Within the upper Mundallio Subgroup, the preservation of fine details of the detrital texture of dolomite mudstones and peloidal dolomites, and the high Sr contents of dolomites (largely in the range of 400-650 ppm), suggest that these sediments consisted of Ca-Mg carbonates (protodolomite, Mg-calcite) at the time of deposition. Slightly greater recrystallisation of dolomites in the lower Mundallio Subgroup resulted in their lower Sr and higher Mn and Fe contents.

Magnesite mudstones may have initially precipitated as hydrated Mg-carbonates. Lithification of surface sediments as a result of subaerial exposure, led to the formation of micritic magnesite. Much of this magnesite was subsequently reworked into intraclastic beds.

The carbonate mineralogy of this sequence, and the evidence of only rare sulphates, indicate that the carbonates were precipitated from alkaline, Mg-Ca-CO_3 waters, with a higher carbonate and lower sulphate content than seawater.

STATEMENT OF ORIGINALITY

This thesis contains no material which has been accepted for the award of any other degree or diploma in any University, nor, to the best of my knowledge, does it contain any material previously published or written by any other person, except where due reference is made in the text.

Robin K. Uppill

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PART I

GENERAL INTRODUCTION

CHAPTER 1

AIMS AND SCOPE OF THE STUDY

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An arcuate belt of folded late Precambrian (Adelaidean) and Cambrian sedimentary rocks, outcrops in the Mt. Lofty and Flinders Ranges of South Australia (Fig. 1.1). Isolated outcrops are also present in the Peake and Denison Ranges (Fig. 1.1). The sedimentary sequence, which is largely of shallow water origin, was deposited in an elongate basin, the Adelaide Geosyncline. Many of the sediments deposited in this basin have no modern analogues. For example, thick and extensive horizons of quartz-cemented, mature and supermature sandstones, which are present at several stratigraphic levels within the sequence (e.g. Plummer, 1978b), are not accumulating in modern sedimentary environments. In addition, dolomites are widespread in the Callana and Burra Groups, which comprise the lower part of the sedimentary pile deposited within the Adelaide Geosyncline. However in modern sediments, dolomite forms only small, localized deposits.

This study is aimed at deducing the sedimentary history of the Mundallio Subgroup, one of the dolomite rich intervals within the Burra Group. The Mundallio Subgroup is a mixed carbonate-clastic sequence, in which dolomite is the major carbonate mineral present. However magnesite, an uncommon carbonate mineral in both modern and ancient sediments, is a widespread, although usually minor component. The subgroup has been studied in its two major areas of outcrop, in the northern Flinders Ranges, and in the southern Flinders and Mt. Lofty Ranges. Stratigraphic equivalents present in the Peake and Denison Ranges have not been considered. The major part of this study discusses both the carbonate and clastic facies, and their environmental interpretations. Emphasis is placed on the dolomite-magnesite-sandstone sequence which comprises the upper part of the Mundallio Subgroup. The environmental interpretations are incorporated within a regional palaeo-geographical model and discussion of the nature of the basin of deposition.

The latter part of this study considers particular aspects of the carbonate facies, including their early diagenetic history, geochemistry, and the diagenetic replacement of carbonates by silica. These are combined with the environmental interpretations, in order to determine the origin of dolomite and magnesite within this sequence, and the implications which the presence of this mineral association has for the nature of the basin of deposition. Modern occurrences of dolomite and magnesite, in both marginal marine and lacustrine environments, provide analogues for the chemistry of the basin and methods of carbonate production within it. However because of the major differences in scale, there are no modern analogues for the extensive, shallow basin in which this unusual carbonate sequence was deposited.

CHAPTER 2

GEOLOGICAL SETTING - THE ADELAIDE GEOSYNCLINE

TECTONIC FRAMEWORK

The sediments within the Adelaide Geosyncline were deposited on an older Precambrian cratonic basement, now largely preserved in the Gawler Block to the west, although basement inliers occur as far east as the Broken Hill area in New South Wales (Willyama Inlier, Fig. 2.1). The western boundary of the Adelaide Geosyncline, the Torrens Lineament, separates folded sedimentary rocks from the basement of the Gawler Craton (Daily *et al.*, 1973), on which unfolded to gently folded Adelaidean and Cambrian sedimentary rocks are preserved in the Stuart Shelf area (Fig. 2.1). The eastern margin of the geosyncline is largely covered by Mesozoic and Tertiary sedimentary sequences.

The basement inliers east of the Torrens Lineament represent parts of larger basement blocks which experienced continued instability throughout Adelaidean time (Glenn *et al.*, 1977). Retrograde schist zones which developed in the basement, were related to faults which affected Adelaidean sedimentation, and produced thickness variations (Glenn *et al.*, 1977). Hence stable platforms equivalent to the Gawler Craton may not have existed north and east of the geosyncline (Rutland and Murrell, 1975). However, Thomson (1976), proposes that the inferred Curnamona Cratonic Nucleus (Fig. 2.1) has resisted deformation since the Carpentarian, and controlled fold trends in the adjacent Adelaidean sediments. The Mt. Painter and Willyama Inliers, north and south of the Curnamona Nucleus respectively, were involved in the Delamerian Orogeny which terminated sedimentation in the Adelaide Geosyncline (Thomson, 1969b).

Thick Adelaidean sediments occur in the Peake and Denison Ranges (Ambrose and Flint, 1979), and the geosyncline probably extended into this area. More intermittent connections may have occurred with other basins to the north, such as the Amadeus Basin (Wopfner, 1969; Thomson, *et al.*,1976). The presence of the Muloorina Gravity High has led to the suggestion of a basement ridge (Fig. 2.1) which formed the northeasterly margin of the basin between the Mt. Painter and Denison Inliers (Thomson, 1970), although the gravity high may also represent Delamerian fold trends in Adelaidean sediments (Thomson, 1976). Sedimentological studies in the Willouran Ranges indicate the presence of a source area in the vicinity of the Muloorina Ridge, at least during part of the Adelaidean (Murrell, 1977).

Hence Adelaidean and Cambrian sediments accumulated in a tectonically active zone, with a large stable basement block to the west, and possibly smaller, ephemerally stable blocks to the east and north. Sedimentation was significantly affected by tectonics within the basin of deposition, and as a result there were several cycles of deposition (Table 2.1). Although the basin has been termed the Adelaide Geosyncline, and continued subsidence over a long period of time allowed a thick sequence of sediments to accumulate (average thickness of 10-15,000 m, Rutland, 1976), the depositional environment was in general that of a shallow epeiric basin (Preiss, 1973).

HISTORY OF SEDIMENTATION

Sedimentation within the Adelaide Geosyncline occurred in four major cycles of deposition (Table 2.1), each characterised by a distinctive type of sedimentation and tectonic style, and separated by major disconformities or low angle unconformities. Attempts at dating the commencement of sedimentation have been based largely on radiometric dating of the underlying basement sequences, and basic volcanics within the Callanna Group. Dating of pre-Adelaidean basement metamorphism near Adelaide, gives an age of 849 ± 31 Ma (Cooper and Compston, 1971; Cooper, 1975). Volcanics within the Callana Group have been dated at about 800 Ma (Compston *et al.*, 1966; Cooper, 1975). Revised correlations in the area of the Torrens Lineament, and the Stuart Shelf by Mason *et al.*,(1978), suggests a correlation between the Beda Volcanics, now dated at 1076 ± 34 Ma (Thomson *in* Preiss, 1979a), and volcanics within the Callanna Group. Hence there is some conflict between the available data, but it is likely that Adelaidean sedimentation commenced in the interval between 1100 and 800 Ma ago (Preiss, 1979a). Preiss (1976) has also used stromatolite biostratigraphy to suggest that deposition of the Callanna Group commenced in the Late Riphean (950±50 to 680±20 Ma).

WILLOURAN SEDIMENTATION (CALLANNA GROUP)

The sequence in the type area in the Willouran Ranges has been recently summarised by Murrell (1977), in which he proposed Callanna Group to replace Callanna Beds. Only scattered outcrops of this group occur in the Adelaide Geosyncline, and they frequently exhibit disruption of the stratigraphic sequence. Sedimentation was associated with tensional tectonics and basaltic volcanism (Rutland and Murrell, 1975; Rutland, 1976; Glenn *et al.*, 1977). This may be reflected within the basement by mafic igneous activity (Glenn *et al.*, 1977). Shallow water sedimentation predominated, resulting in the deposition of a clastic-carbonate-evaporite sequence. Although deposition of this group occurred in much of the area characterised by Adelaidean sediments, including the adjacent Gawler Block, Willouran sediments are absent from the Mt. Lofty Ranges and the Olary region, where Torrensian sediments directly overly the basement (Preiss, 1979a).

TORRENSIAN SEDIMENTATION (BURRA GROUP)

Extensive outcrops of this group are present throughout the Adelaide Geosyncline, with the exception of the central Flinders Ranges. The Burra Group is not preserved on the Gawler Craton, although widespread erosion in this area prior to deposition of the overlying Umberatana Group, could have removed any Torrensian sediments deposited on the craton. The Burra Group is also absent from the Broken Hill region (Preiss, 1979a).

The Burra Group consists of a sequence of rather repetitive lithologies of shallow water origin, including sandstones, grey siltstones and shales, and grey dolomites. In contrast with the underlying Callanna Group, evidence of evaporites is rare. The repetitive nature of the sequence may cause correlation problems between isolated areas. However three cycles of deposition may be recognised. The basal cycle is dominated by terrigenous clastics, and shows facies and thickness changes due to proximity to the basin margin and differential subsidence rates. A coarse clastic wedge, probably with fluvial sediments at the base (Preiss, 1979a), was deposited on the western margin of the basin (now preserved between Adelaide and the Emeroo Range), and on both sides of the Willouran Ranges (Murrell, 1977). Away from these regions this sequence becomes finer grained, and has a higher proportion of shales and dolomites, although it usually begins and ends with a sandstone dominated interval.

Within the second cycle of deposition (the Mundallio Subgroup), carbonate deposition, including dolomite and magnesite, was significant. However, interbeds of sandstone and shale are abundant in some areas. The third cycle of deposition is characterized by reduced deposition of dolomite, due to a renewed influx of clastics of sand, silt and clay size. In the southern part of the geosyncline, deltaic sandstones prograded from the west (Preiss, 1979a). Although the sediments tended to be finer grained than those of the first cycle, the coarsest sequences were again deposited on the western margin.

Differential subsidence rates produced substantial lateral thickness variations within the Burra Group, particularly in the Willouran Ranges (Murrell, 1977), where syndepositional tectonism also produced growth folds and faults. Widespread uplift preceding deposition of Sturtian glaciogene sediments, resulted in erosion of the upper parts of the Burra Group. In some areas, including the Mt. Painter area, the Olary area, and east of Hawker, the Burra Group was completely removed, possibly suggesting a considerable time break between deposition of these two sequences.

STURTIAN - MARINOAN SEDIMENTATION (UMBERATANA AND WILPENA GROUPS)

The third major cycle of deposition, which commenced with glacial sediments, was again dominated by shallow water sedimentation. Syndepositional block faulting influenced deposition in some areas (Plummer and Gostin, 1976; Murrell, 1977), while uplift and erosion resulted in the development of local disconformities within the sequence (Plummer, 1978a). Clastic sediments become mineralogically more mature during this cycle. Sandstones within the Umberatana Group are feldspathic (Plummer, 1974; Jablonski, 1975; Miller, 1975), whereas those within the Wilpena Group often have the composition of quartz-arenites (Plummer, 1978b). Red coloured sediments developed throughout this cycle, whereas they were absent from the underlying Torrensian cycle. Carbonates within the Torrensian cycle are predominantly grey dolomite mudstones, while in this cycle both limestones and dolomites are present, and they exhibit a much greater variety of facies, both depositional and diagenetic, on a local and regional scale. Stromatolite morphologies are also more variable (Preiss, 1973).

CAMBRIAN SEDIMENTATION

Mild tectonic activity appears to have been widespread at the end of the Precambrian, with the result that a disconformity or low angle unconformity separates the Precambrian and Cambrian sequences (Daily, 1976; Daily *et al.*, 1976). Initially, shallow water clastics and carbonates were deposited throughout the geosyncline, on the Stuart Shelf, and in the Yorke Peninsula area. Syndepositional tectonism is indicated by disconformities within the sequence (Daily, 1976). In the late Lower Cambrian, increased subsidence relative to the rate of deposition in the southern part of the geosyncline, produced the Kanmantoo Trough (Fig. 2.1). Deeper water shales and greywackes accumulated within this trough, while shallow water sedimentation continued elsewhere.

SOURCE AREAS

Sediments of the Adelaide Geosyncline were predominantly derived from the high grade metamorphic and igneous terrains of the adjacent basement blocks, although the widespread Gawler Range Volcanics and other acid volcanics of the Mt. Painter area and the Curnamona Cratonic Nucleus (Giles and Teale, 1979) were probably also a significant source. The Gawler Block consists of metasediments (quartzites, mica schists), granite gneiss, and intrusive granitic and basic rocks (Thomson, 1969a). Lithologies in the other basement inliers are similar to those of the Gawler Craton (Thomson, 1969a), and hence would have supplied sediment of similar composition. A source of combined igneous plutonic and metamorphic rocks, with variable recycling, is indicated by the arkosic, subarkosic, and quartz-arenitic composition of the sandstones (Folk, 1974a).

The Gawler Craton appears to have been the major source during much of the history of the geosyncline, as indicated by the clastic wedge on the western margin of the geosyncline within the Burra Group, geochemical studies in the lower Umberatana Group (Sumartojo, 1974), and palaeocurrent data from the Brachina Subgroup (Plummer, 1978b). Other source areas were intermittently present, for example, during the Torrensian, sediment was also supplied from northeast of the Willouran Ranges (Murrell, 1977), and the Willyama Block was a minor source area in the lower Sturtian (Sumartojo, 1974). The Curnamana Cratonic Nucleus, and basement exposed within the Mt. Painter area, may also have supplied sediment during the early Sturtian (Ashton, 1973; Link, 1977; Giles and Teale, 1979).

Hence the picture that emerges is one of a stable Gawler Craton to the west, but experiencing adequate uplift to continuously supply sediment to the geosyncline. In other areas to the east and north, tectonics controlled the morphology of the sedimentary basin, and the presence or absence of upfaulted basement blocks, which intermittently supplied sediment to the adjacent areas of the geosyncline (Wopfner, 1969).

DELAMERIAN OROGENY

A major tectonic episode, the Delamerian Orogeny (Thomson, 1969b), terminated the sedimentary history of the Adelaide Geosyncline. The major folding episode occurred about 500-510 Ma ago (Mancktelow, 1979), and produced the major fold structures. Tectonic activity apparently continued for about another 50 Ma (Milnes *et al.*, 1977; Mancktelow, 1979), and included granite emplacement, and the formation of a crenulation cleavage in the southern part of the geosyncline. The underlying basement rocks, now represented as inliers, were affected by the Delamerian Orogeny, and trends within the basement may have controlled the Delamerian fold trends (Glenn *et al.*, 1977).

Metamorphism preceding and associated with the Delamerian Orogeny, was most intense in the Kanmantoo Trough and adjacent regions, and produced a well defined arrangement of metamorphic zones (Offler and Flemming, 1968; Mancktelow, 1979). The remainder of the geosynclinal sequence was metamorphosed to chlorite facies, although biotite and amphibolite facies were produced in the Olary and Mt. Painter regions.

PART II

STRATIGRAPHY, SEDIMENTOLOGY AND PALAEOGEOGRAPHY OF THE MUNDALLIO SUBGROUP

CHAPTER 3

THE MUNDALLIO SUBGROUP : INTRODUCTION AND STRATIGRAPHY

INTRODUCTION AND PREVIOUS INVESTIGATIONS

The Mundallio Subgroup was introduced by Uppill (1979) to refer to the second cycle of deposition within the Burra Group. This cycle which resulted in the deposition of a mixed sedimentary sequence, including dolomites, sandstones, shales and magnesites, had previously been referred to as the Skillogalee Dolomite (Wilson, 1952; Mirams and Forbes, 1964; Forbes, 1971) in all its areas of occurrence with the exception of the Adelaide region, where the Castambul Dolomite and Montacute Dolomite had been used (Mawson and Sprigg, 1950). The revised nomenclature is shown in Figure 3.1. Descriptions of this interval may be found in general references referring to the Burra Group (Mawson, 1941, 1947; Sprigg 1946; Spry, 1952; Preiss and Sweet, 1967; Binks, 1971; Coats and Blissett, 1971; McCarthy, 1974; Fairchild, 1975; Daily *et al.*, 1976, Murrell, 1977; Ambrose and Flint, 1979). These references generally contain useful summaries of the lithologies present in the particular area of study, but environmental interpretations are generally rather limited, with the exception of Spry (1952), Preiss and Sweet (1967), Murrell (1977), and Ambrose and Flint (1979).

Studies specific to the Mundallio Subgroup have been carried out by Forbes (1955, 1960, 1961), and Preiss (1971, 1973). Forbes studied that part of the subgroup characterised by magnesite, and described its distribution and that of the associated facies. A summary palaeogeographical reconstruction was presented (Forbes, 1961, Fig. 5). Preiss (1971, 1973) was concerned mainly with the morphology and occurrence of stromatolites within the Mundallio Subgroup, but also described in greater detail the sequence at Depot Creek (DC). Preiss discussed and enlarged the environmental interpretation of Forbes (1961), but stated that "the origin of dolomite and magnesite remains problematical ..... the association of bedded dolomite and conglomeratic magnesite appears to have no modern analogue" (Preiss, 1973, p. 508). Hence particular attention in this study is given to the dolomite and magnesite facies, their sedimentary characteristics and origin.

Microfossils preserved within diagenetic cherts replacing dolomites, and in particular stromatolitic dolomites, within the Mundallio Subgroup, have been described by Schopf and Barghoorn (1969), Schopf and Fairchild (1973), Fairchild (1975) and Knoll *et al.* (1975). These studies reveal an areally widespread (Peake and Denison Ranges to Mundallio Creek), and diverse microflora. Microfossils present within the Myrtle Springs Formation and the River Wakefield Subgroup have similarities with those of the Mundallio Subgroup. Many Precambrian microfossils appear to be long ranging forms, and this combined with the rarity of well preserved fossils, and the lack of accurate age dates for fossiliferous sequences, means that their biostratigraphic potential is limited (Fairchild, 1975). Hence despite the abundant microflora within the Mundallio Subgroup, it cannot be used to determine the age of the sequence.

DISTRIBUTION OF THE MUNDALLIO SUBGROUP

The distribution of the Mundallio Subgroup and its constituent formations, is shown in Figure 3.2. Extensive outcrops of facies similar to those within the Mundallio Subgroup, are also present in the Peake and Denison Ranges (Ambrose and Flint, 1979). The Burra group is not present within the central Flinders Ranges where basal Umberatana Group sediments abut against disrupted Willouran sediments mapped as diapiric bodies (Dalgarno and Johnson, 1966). Rutland and Murrell (1975) have suggested that the Burra Group was not deposited within the central Flinders Ranges, and this possibility, with respect to the Mundallio Subgroup, will be discussed in Chapter 8. No outcrops of the Mundallio Subgroup are known from the platform sequence which was deposited on the Gawler Craton, although possible subsurface equivalents are preserved within the Torrens Hinge Zone south of Port Pirie (Lynch, 1978).

STRATIGRAPHY OF THE MUNDALLIO SUBGROUP

CASTAMBUL FORMATION

The Castambul Formation refers to the sequence of pale pink to buff coloured massive dolomites, previously referred to as the Castambul Dolomite (Mawson and Sprigg, 1950), and the associated phyllites and minor sandstones, which overlie the Aldgate Sandstone, and is overlain by the Montacute Dolomite (Uppill, 1979). It has a localised occurrence, being confined to the area south and west of the Houghton Inlier, with the type area in Torrens Gorge (TG) near Castambul. To the south, east and north, the Aldgate Sandstone is overlain by the Woolshed Flat Shale (Fig. 3.3).

MONTACUTE DOLOMITE

The Montacute Dolomite, defined by Mawson and Sprigg (1950), consists of a sequence of interbedded grey dolomites, grey to cream magnesites, dolomite-cemented sandstones, and minor phyllites. The formation is largely confined to the same area as the Castambul Formation which it overlies. Scattered outcrops are present north of the Mt. Bold Reservoir (MB), 20 km south of Adelaide. The Montacute Dolomite is overlain by, and replaced laterally by the Woolshed Flat Shale (Fig. 3.3). Hence in the Adelaide area, dolomitic sequences are largely confined to the area west of the Houghton Inlier, being replaced to the east by fine grained clastics (Talbot, 1962; Mancktelow, 1979).

SKILLOGALEE DOLOMITE

The Skillogalee Dolomite, which consists of a lower member of buff recrystallised dolomites, and a thin upper member of dark-grey dolomites, was proposed by Wilson (1952), with the type section in an area of poor outcrop west of Auburn, near Skillogalee Creek (SC). This formation, which is in general very poorly outcropping, extends from its type section 75 km northward, to the northern end of the inlier of River Broughton Beds (Preiss, 1974a), near Spalding (S). Because of poor outcrop the southern extent is not well known, but outcrops occur at least as far south as an area west of Tarlee (T). Possible outcrops are also present on the River Light (RL) south of Stockport, where white, weathered and recrystallised dolomites overlie ferruginised, cross-bedded sandstones[1] which may be Rhynie Sandstone (this sandstone is not shown on the ADELAIDE 1:250,000 geological map sheet, Thomson, 1969c). The Skillogalee Dolomite also outcrops between Scrubby Range (SR) and Burra (B), and the outcrops are more extensive than those shown on the BURRA 1:250,000 geological map sheet (Mirams, 1964).

The Skillogalee Dolomite overlies the Rhynie Sandstone, possibly disconformably in some areas (Preiss, 1974a), and is overlain by the Woolshed Flat Shale. It may also intertongue with the Woolshed Flat Shale, but because of poor outcrop, the precise nature of this relationship cannot be determined.

WOOLSHED FLAT SHALE

This formation, also defined by Wilson (1952), consists of weathered grey shales, with minor sandstones and dolomites in some areas. It overlies dolomites of the Montacute and Skillogalee Dolomites, and is overlain by sandstones of the Stonyfell and Undalya Quartzites. South of Burra, due to the absence of the Undalya Quartzite, the Woolshed Flat Shale cannot be distinguished from the overlying, and lithologically similar Saddleworth Formation.

NATHALTEE FORMATION

This formation, proposed by Uppill (1979), refers to the "silty-quartzitic sequence containing light grey or cream dolomite beds" mapped by Binks (1971, p. 25) in the southern Flinders Ranges, and previously referred

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1. The ferruginization may be a Tertiary weathering effect, however these sandstones are bedded and steeply dipping, and hence unlikely to be an outcrop of the ferruginous, massive sandstones and conglomerates of Tertiary age, which outcrop in this area.

to as the lower part of the Skillogalee Dolomite. In the type section near Depot Creek (DC), the Nathaltee Formation consists of three units:

(1) interbedded grey dolomite mudstones[1] and stromatolitic dolomites, grey siltstones and shales, and sandstones;

(2) interbedded grey-green shales, buff stromatolitic dolomites and dolomite mudstones, and minor sandstones;

(3) a coarsening upwards shale to sandstone sequence.

The Nathaltee Formation outcrops on the western margin of the southern Flinders Ranges and in the Yednalue Anticline. In the Port Pirie area, subsurface occurrences assigned by Lynch (1978) to the Skillogalee Dolomite, may be equivalent to the Nathaltee Formation. A sequence of grey dolomite mudstones, black shales and minor conglomeratic sandstones, similar to Unit 1, overlies conglomerates and sandstones of the Emeroo Quartzite, and is generally overlain unconformably by Sturtian or Tertiary sedimentary rocks. However a sequence of green and grey shales and siltstones, buff and fawn dolomite mudstones, and medium- to coarse-grained sandstones, similar to Unit 2 of the Nathaltee Formation, is sometimes preserved above the lower unit. This sequence, its distribution and relationships, and that of other Adelaidean sediments in this area, are described by Lynch (1980, in prep.).

The stratigraphic position of the Nathaltee Formation, together with the presence of buff coloured dolomites, and grey-green siltstones, suggests a correlation between it, the Castambul Formation, and the lower part of the Skillogalee Dolomite. The Nathaltee Formation is overlain conformably by the Yadlamalka Formation, and field relationships suggest that intertonguing between these two formations also occurs (Fig. 3.4).

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1. The terms dolomite mudstones and magnesite mudstones are used throughout this thesis in the sense of Dunham (1962).

NANKABUNYANA FORMATION

The lower unnamed member of the Skillogalee Dolomite in the northern Flinders Ranges (Coats *et al.*,1969), has been renamed the Nankabunyana Formation by Uppill (1979), and a type section 6 km southwest of Copley (CP) proposed. The formation, which overlies the Copley or Wortupa Quartzites, consists dominantly of terrigenous clastics, including siltstones, shales and fine grained sandstones, with minor dolomite interbeds in the middle part of the formation. The Nankabunyana Formation comprises the lower part of the Mundallio Subgroup in much of the Northern Flinders Ranges, but is replaced laterally by the Camel Flat Shale and the Tilterana Sandstone (Murrell, 1977) northeast of the Norwest Fault in the Willouran Ranges (Figs. 3.2 and 3.5). Because of its stratigraphic position between quartzites of the lower Burra Group and the Yadlamalka Formation (Fig. 3.1), the Nankabunyana Formation appears to be equivalent to the Nathaltee Formation.

CAMEL FLAT SHALE AND TILTERANA SANDSTONE

Northeast of the Norwest Fault in the Willouran Ranges, the lower Mundallio Subgroup consists of a lower unit of dark-grey shale with minor lenticular brown dolomites (Camel Flat Shale), overlain by a sequence of white, fine-grained sandstones, with minor siltstones and dolomites (Tilterana Sandstone, Figs. 3.1 and 3.5). These formations, which were defined by Murrell (1977), are difficult to recognise in the central Willouran Ranges (Mirra Creek, MI, and Rischbieth Hut, R, areas), because of the absence of the Copley Quartzite, and similarities in facies between the lower Burra Group and the Mundallio Subgroup.

YADLAMALKA FORMATION

The Yadlamalka Formation refers to that part of the previously mapped Skillogalee Dolomite in the Flinders Ranges, characterised by interbedded dark-grey dolomites, stromatolitic dolomites, intraclastic magnesite, dolomite-cemented sandstones, and minor shales (Uppill, 1979). This formation, which is the most widespread in the Mundallio Subgroup (Fig. 3.2), has a type section in Depot Creek (DC). Outcrops in the southern Flinders Ranges are overlain by the Undalya Quartzite, the Auburn Dolomite equivalent, or the Saddleworth Formation. In the northern Flinders Ranges, where it is overlain by the Myrtle Springs Formation, the Yadlamalka Formation is present in all areas except the eastern half of the Willouran Ranges, where it is replaced laterally by the Mirra Formation (Figs. 3.1 and 3.5).

MIRRA FORMATION

The Mirra Formation was defined by Murrell (1977), with a type section in Mirra Creek (MI), 30 km southwest of Marree in the central Willouran Ranges. It is distinguished from the laterally equivalent Yadlamalka Formation by a higher clastic content, including dolomitic sandstones, siltstones and shales, and quartz-cemented sandstones, and the almost complete absence of magnesite. Dark-grey dolomites similar to those of the Yadlamalka Formation are present.

CORRELATION AND BOUNDARY PROBLEMS

The Burra Group is a sequence of rather repetitive lithologies, in which statigraphic terminology is based on the recognition of major quartzite horizons, and the presence of carbonate dominated, or clastic dominated sequences. Many of the facies which are present within the Mundallio Subgroup, are not unique to it, and may be found in both the underlying and overlying sequences. As a result, correlation between isolated areas, particularly where outcrop is poor, is sometimes difficult.

For example, the River Wakefield Subgroup contains dolomites similar to those within the Mundallio Subgroup. In particular, the Benbournie Dolomite (Forbes, 1964) contains grey dolomite mudstones with black chert nodules, similar to dolomites within the Yadlamalka Formation. Grey dolomites with black chert nodules also occur in some outcrops of the Auburn Dolomite. In the Carrieton Anticline (C), the sequence mapped as the River Wakefield Subgroup by Binks (1968), contains an interval with facies identical to those of the Yadlamalka Formation. The horizon mapped as the Yadlamalka Formation (Skillogalee Dolomite), is similar to, and enclosed within facies usually included in the Auburn Dolomite equivalent. The Yadlamalka Formation is well developed at Johnburgh (J), 17 km to the east of Carrieton, and similarities between this outcrop, and the horizon included within the River Wakefield Subgroup at Carrieton, suggests that the latter is in fact the Yadlamalka Formation.

The Mundallio Subgroup generally has well defined boundaries, especially on the western margin of the Mt. Lofty and southern Flinders Ranges, where it is enclosed between two distinctive quartzitic units (Aldgate Sandstone, Rhynie Sandstone or Emeroo Quartzite below, and Stonyfell or Undalya Quartzites above, Fig. 3.1). Likewise in the northern Flinders Ranges and most of the Willouran Ranges, the boundary with the underlying Copley or Wortupa Quartzites is distinctive, as is the upper boundary with the sandstone-siltstone sequence of the Myrtle Springs Formation. Although these boundaries may be gradational, they represent distinct lithological changes. There does not appear to be significent intertonguing between the Mundallio Subgroup and the overlying and underlying formations, in the areas mentioned above.

In the central part of the southern Flinders Ranges, the Mundallio Subgroup is represented only by the Yadlamalka Formation. In the underlying Bungaree and Yednalue Quartzites, sandstones form the most conspicuous lithology, but there is a significant component of finer clastics and dolomites. The dolomite interbeds may be lithologically similar to those of the Nathaltee Formation and the Skillogalee Dolomite. Hence intertonguing of the Yednalue and Bungaree Quartzites with the Nathaltee Formation to the west, and the Skillogalee Dolomite to the south, may have occurred. In the Yacka area, the Yadlamalka Formation intertongues to the west with a sequence of quartzites, siltstones and dolomites similar to the Bungaree Quartzite which underlies the Yadlamalka Formation in this area (Fig. 3.6). In the Yacka-Spalding area, both the Mundallio Subgroup and the underlying lower Burra Group, show substantial thickness and facies changes, largely in an east-west direction across the Spalding inlier of River Broughton Beds (Fig. 3.6). This reflects considerable differential subsidence rates during deposition of much of the Burra Group.

South of Adelaide, in the Congeratinga River (CR) area on the southern end of the Myponga-Little Gorge Basement Inlier, the Mundallio Subgroup appears to have been replaced by a sandy sequence. The Burra Group in this area, which has been mapped by Anderson (1975), begins with heavy mineral laminated sandstones, probably equivalent to the Aldgate Sandstone. These are overlain by buff dolomitic sandstones with phyllitic interbeds, with an uppermost unit of siltstones and fine-grained sandstones. This area is well separated from other Burra Group outcrops in the Adelaide region, hence the relationship between the two areas is not known. However the sequence of dolomitic sandstones may represent the same depositional period as the Mundallio Subgroup.

Equivalents of the Mundallio Subgroup in the Peake and Denison Ranges have been recently described by Ambrose and Flint (1979). The sequence contains three major units, and totals 3,600 m in thickness. These units are:

(1) sandstones with interbedded green-grey shales and minor brown dolomites;

(2) sandstones with more interbedded dolomites than (1), minor shales and magnesite. The dolomites are brown, but become dark-grey at the top;

(3) dark-grey dolomites with minor interbedded shales and sandstones. This sequence is comparable with that in the northern Flinders Ranges, in that there is a lower clastic dominated interval, overlain by a dolomite dominated interval, although the boundary between them appears to be gradational.

SUMMARY

The formations within the Mundallio Subgroup may be grouped into three major facies associations, and these form the basis for the subsequent discussion of the sedimentology and depositional history of the Mundallio Subgroup. These associations are:

(1) the generally clastic dominated sequence with some dolomite of the lower Mundallio Subgroup, which is widespread in the northern Flinders Ranges (Nankabunyana Formation, Camel Flat Shale and Tilterana Sandstone), but becomes more lenticular and dolomitic in the southern Flinders Ranges (Nathaltee Formation). This association is characterised by rapid vertical facies changes.

(2) Dolomite-shale association of the Mt. Lofty Ranges, in which few facies are present, and they occur as thick, internally homogeneous units (Castambul Formation, Skillogalee Dolomite and Woolshed Flat Shale);

(3) the widespread association of dark-grey dolomites, intraclastic magnesite and dolomitic sandstone (Yadlamalka Formation and Montacute Dolomite) and the laterally equivalent sequence of dark-grey dolomites, sandstones and siltstones (Mirra Formation). These formations, in particular the Yadlamalka Formation and Montacute Dolomite, are characterised by rapid vertical facies changes.

The environmental interpretations of these facies associations are combined to produce a palaeogeographical history of the Mundallio Subgroup.

CHAPTER 4

CLASTIC DOMINATED SEQUENCES OF THE LOWER MUNDALLIO SUBGROUP IN THE SOUTHERN AND NORTHERN FLINDERS RANGES (NATHALTEE FORMATION, NANKABUNYANA FORMATION, CAMEL FLAT SHALE AND TILTERANA SANDSTONE): FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS.

NATHALTEE FORMATION

INTRODUCTION

Deposition of the Nathaltee Formation was localized in two separated areas of the southern Flinders Ranges (Fig. 3.2), and in both the lower boundary with the underlying quartzite sequences (Fig. 3.1) is gradational. Within the Emeroo Range, this boundary is well exposed in the Depot Creek area (DC), but elsewhere the Nathaltee Formation is often deeply weathered, and the boundary extensively covered by quartzite scree. At Depot Creek, dolomite mudstones, siltstones, and stromatolitic dolomites are interbedded with quartz-cemented sandstones in the uppermost 40 m of the Emeroo Quartzite, and quartz-cemented sandstones persist in the lowermost 10 to 15 m of the Nathaltee Formation (Fig. 4.1). There is no distinctive marker bed on this boundary, and due to variable development of quartzites, siltstones and dolomites along strike, this boundary is best considered as an intertonguing, gradational boundary, positioned at the point above which dolomites and siltstones form more than 50% of the sequence.

In the Port Germein Gorge (PG) and Beetaloo Valley (B) areas, interbedded quartzites with symmetrical ripple marks and desiccation cracks, and deeply weathered siltstones, deposited in a very shallow water environment, form the uppermost unit of the Emeroo Quartzite (McCarthy, 1975; Winsor, 1977). This unit is gradational into a sequence of dolomites and siltstones, with minor sandstones, which forms Unit 2 of the Nathaltee Formation (Fig. 4.1).

In the Yednalue Anticline (YDA) the Nathaltee Formation gradationally overlies the Yednalue Quartzite, which on the west limb of the anticline, consists of an intermittently outcropping sequence of feldspathic, medium-to coarse-grained quartzites, shales, siltstones, and some interbedded silty dolomites (Forbes, 1969). On the east limb, outcrops of quartzite interbeds become less prominent eastward, and the boundaries of the Yednalue Quartzite become difficult to define. As on the western margin of the southern Flinders Ranges, a decrease in sand content, and the predominance of finer grained sediments (siltstones, shales, and dolomite mudstones), mark the lower boundary of the Nathaltee Formation. However the transition is less significant than in other areas, because of the higher siltstone and dolomite content of the Yednalue Quartzite.

The upper boundary of the Nathaltee Formation with the overlying Yadlamalka Formation is generally sharp, occurring at the top of the quartzite horizon forming the upper part of Unit 3 (Fig. 4.1; Uppill, 1979). Where this quartzite is absent, siltstones and shales, with some interbedded dolomites are replaced by the largely dolomitic sequence of the Yadlamalka Formation.

The Nathaltee Formation may be very locally present in the Willow Creek (WI) area, on the west limb of the Worumba Anticline. A sequence of interbedded grey siltstones and grey dolomites which are often stromatolitic, is lithologically similar to the Nathaltee Formation. It is conformably overlain by the Yadlamalka Formation in the central area of the range of hills in which the Mundallio Subgroup outcrops. This sequence has a faulted base. However in the northern part of the range, the Yadlamalka Formation overlies a sequence of siltstones and quartzites, which may be the Yednalue Quartzite (Preiss, 1979b). Preiss has also tentatively correlated the sequence of siltstones and dolomites with the Yednalue Quartzite. However the alternative interpretation of this sequence being the Nathaltee Formation, implies that it intertongues with both the Yednalue Quartzite and Yadlamalka Formation. However, because of the isolated nature of this outcrop, faulting, and the gradational boundary of the Yednalue Quartzite and Nathaltee Formation in other areas, the exact position of this sequence remains open to question.

FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

The Nathaltee Formation contains three units, each with a distinctive association of lithologies. The distribution of these units is apparent from Figure 4.2, and the composition of each in the main areas of study, in Figure 4.1. Within the Depot Creek area, the three units are mappable (Fig. 4.3), although the boundary between Units 1 and 2 is transitional.

Unit 1

The facies association in Unit 1 consists of interbedded siltstones and shales, dolomite mudstones, stromatolitic dolomites and sandstones. All facies are predominantly grey in colour, although at Depot Creek, shales and dolomites are often weathered to reddish colours near the base of the formation.

Siltstones and Shales

Grey to dark-grey laminated siltstones and shales are the most abundant facies in the Yednalue Anticline, being less significent in the Depot Creek area (Fig. 4.1) and south of Port Pirie. Silty laminae are sometimes dolomite cemented, and minor laminae of dolomite mudstone are present. The lamination is flat, to wavy and lenticular, and infrequently cross laminated. Other features include graded laminae, microscopic load structures, small scale slump structures, and small erosional scours attributable to minor traction currents. Desiccation and synaeresis cracks are both of infrequent occurrence.

This facies was deposited in a largely submergent environment, in which sediment was carried in suspension or by weak traction currents. Hence the environment was a low energy, submergent, to infrequently emergent mudflat. Reducing conditions prevailed within the sediment, as indicated by the grey colour, and the presence of minor pyrite.

Sandstones

Fine- to coarse-grained sandstone interbeds (up to 6 m in thickness at Depot Creek, and 2 m in the Yednalue Anticline, although they are largely 10-30 cm in thickness here), generally have sharp boundaries with other facies. Flat, to slightly wavy lamination and thin bedding, are the dominant structures, but may be poorly defined due to little textural contrast between adjacent laminae. Also present are ripple lamination, symmetrical ripple marks (wavelength < 10 cm), minor isolated tabular cross-beds, and rare interference ripple marks. Occasional muddy laminae (dolomite and terrigenous) contain desiccation or synaeresis cracks. The cement within sandstones may be dolomite (predominant in the Yednalue Anticline), or quartz (predominant in the Depot Creek area).

Most significant in determining the depositional environment of this facies, is the predominance of flat to slightly wavy bedding, with associated symmetrical ripple marks. According to Reineck and Singh (1973, p. 105), "evenly laminated sand is generally abundantly distributed on beaches or other sandy areas exposed to wave action". Flat-bedded sands may also form in lower (coarse-grained sand only), and upper flow regimes of unidirectional currents (Allen, 1970, Fig. 2.6), although the latter is generally associated with parting lineation (Reineck and Singh, 1973). Deposition of sand from storm induced sunspension clouds produced by shoaling waves, also results in evenly laminated sand in the submerged parts of sand bars and shoals (Reineck and Singh, 1973).

The presence of flat bedding in fine- to coarse-grained sandstones, in association with symmetrical ripple marks, suggests that this facies was deposited in an environment dominated by wave induced processes, with occasional unidirectional currents producing isolated cross-beds. The sandstones may have been deposited as submergent sand shoals, but some slack water periods were followed by emergence, and the desiccation of thin mud laminae.

Conglomeratic Sandstones

In the Port Pirie area, thin sandstone interbeds (predominantly less than 20 cm in thickness) are concentrated near the base of Unit 1. They contain scattered granules and small pebbles, and may be gradational into clast-supported conglomerates. Their base is often erosional on dolomite mudstones, intraclasts of which may be incorporated within the sandstones. This erosional surface often forms the base of a cycle, 0.1 to 1.5 m in thickness, of conglomeratic sandstone-shale-dolomite mudstone (Fig. 4.4). Unit 1 in this area overlies a very conglomeratic facies of the Emeroo Quartzite, indicating proximity to source.

These thin conglomeratic sandstones may represent the introduction of sediment from a landward source, possibly as the distal margin of alluvial fans, which spread out into a low energy lake or lagoon. Fine sediment was carried further into the lake, and deposited from suspension (Fig. 4.4). As the sediment supply was exhausted, clastic deposition was replaced by dolomite deposition and fine-clastics were trapped in marginal areas. Sandstones and conglomerates decrease in abundance through Unit 1, suggesting that the sediment source, initially close to this area, became more distal.

Dolomite Mudstones

Thin beds to horizons up to 5 m, and rarely 25 m in thickness of light- to dark-grey dolomite mudstones, have sharp and less commonly gradational boundaries with other facies. The dominant sedimentary structure is flat to wavy and lenticular lamination and thin bedding (Plate 4.1a), which is generally defined by the variation in content of terrigenous sediment (sand and silt laminae, or clay partings). Other features include small erosional surfaces (Plate 4.1a), graded lamination, and small soft sediment slumps. Occasional desiccation cracks, and rare small tepees (Depot Creek only), may be associated with lenticular interbeds of intraclastic dolomite. The dolomite, which is often ferroan[1], has the texture of dolomicrosparite (Folk, 1974a). Replacement of dolomite by diagenetic chert nodules (described in more detail in Chapter 11), is common at Depot Creek, but minor elsewhere.

Dolomite mudstones were deposited in a low energy, submergent to intermittently exposed environment, with little influx of terrigenous sediment. More massively outcropping, thinly bedded dolomite mudstones with a low terrigenous content, may have been deposited from suspension following chemical precipitation, and experienced little reworking. However those with more terrigenous laminae experienced greater reworking, as fine-grained sand and silt were introduced by weak currents which sometimes produced small scours in the unlithified dolomite muds. Dolomite mudstones experienced longer periods of exposure than terrigenous mudstones, and the desiccated crusts produced were probably subject to erosion during storms.

Dolomite Packstones and Grainstones

Grain supported dolomitic sediments represent only a small proportion of this unit. They include lenses and thin beds of intraclastic dolomite, rare peloidal dolomites at Depot Creek, and oncolitic dolomites as massive beds up to 1 m in thickness on the west limb of the Yednalue Anticline. The oncoids are small (maximum of 2-3 mm), and their structure has been partly destroyed during recrystallisation.

Stromatolitic Dolomites

Domal stromatolites, either in bioherms or biostromes, or as isolated structures (Plate 4.1b), are most significant in the more dolomitic sequences of the Depot Creek area. They consist of grey dolomite with minor

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1. All thin sections of carbonate facies described in this thesis, have been stained by the method of Dickson (1965), confirming the presence of dolomite and the absence of calcite. X-ray diffraction has been used for those samples suspected of containing magnesite, see Chapter 6.

diagenetic chert nodules, and hence are similar in appearance to the associated dolomite mudstones. Rarely bioherms are separated by intraclastic dolomite. Columnar stromatolites are less abundant at Depot Creek, and have been described as *Tungussia wilkatana* (Preiss 1974b).

In the Yednalue Anticline, columnar stromatolites form biostromes with undulating surfaces, and which are up to 1 m in thickness, or small bioherms enclosed within dolomite mudstones. The stromatolites may begin as domes or flat laminae, which develop upward into columns with vertical to horizontal growth (Plate 4.1c). The interspace sediment is generally mudstone or wackestone.

Stromatolitic dolomites are largely associated with low energy facies (dolomite mudstones, siltstones and shales), and both domal and columnar stromatolites appear to have grown in largely submergent environments, as there is infrequent evidence of desiccation and erosion of algal mats. Aitken (1967) suggests that domal stromatolites reflect a lower degree of turbulence than columnar forms. In a low energy environment, the algal mats may completely colonise the sediment surface (Hardie and Ginsburg, 1977). Deposition of carbonate mud from suspension will then produce a blanketing laminae of carbonate mud on the algal mat, favouring the development of flat laminated or gently domal stromatolites. However if turbulence is sufficient to maintain carbonate mud in suspension, algal mats may trap sediment out of suspension (Hardie and Ginsburg, 1977). This will tend to enhance any irregularities in the mat surface, which may have been initiated by the slightly more energetic environment, and thus leads to columnar growth. This is also suggested by Horodyski (1977), who states that a predominance of biological stabilization of the sediment over mechanical deposition, will favour the growth of columns.

Magnesite

Rare interbeds of intraclastic magnesite are present at both Depot Creek (up to 2 m in thickness) and on the west limb of the Yednalue Anticline (up to 50 cm, Fig. 4.1). Intraclasts, which are well rounded and up to 5 cm in size, form a close packed framework, or are associated with sand and dolomite intraclasts in a more open framework. Rarely is bedded magnesite mudstone from which the intraclasts were derived, associated with the intraclastic beds (Yednalue Anticline).

Unit 1 : Environmental Summary

This unit is dominated by fine-grained lithologies, with sandstones most significant near the boundary with the underlying quartzite sequences. Dolomite mudstones were deposited in lagoons or embayments with fringing exposed carbonate mudflats, and in which stromatolites sometimes grew. Siltstones and shales may represent slightly more offshore environments. Shoaling sequences often developed without a zone of wave agitation, probably due to low gradients causing wave damping, as occurs on the western side of Andros Island during calm periods (Roehl, 1967), and in the Gulf of Pechili in the Yellow Sea (Dunbar and Rodgers, 1957). However sand sized sediment was periodically introduced and deposited as sand shoals. This unit contains evidence of fluctuating water level, but this may be due to sediments building up to water level, and seasonal or longer term climatic factors, rather than tidal action. In general, features indicative of tidal action are lacking, and deposition of coarser sediments may have been dominated by wave processes.

The facies generally have a random vertical arrangement, apart from rare coarsening upward shale to sandstone cycles (0.5 to 7 m in thickness), and the conglomeratic sandstone-shale-dolomite mudstone cycles in the Port Pirie area which have already been discussed. Hence there was not a simple sequence of depositional environments parallel to the shoreline, but rather a complex of environments which varied both along and away from the shoreline, which was unlikely to have been a simple linear feature. Repeated movement of these environments across a low gradient depositional basin, produced vertical sequences with random, but rapid facies variation.

Unit 2

Interbedded greenish-grey shales, buff to pink stromatolitic dolomites and dolomite mudstones form Unit 2. Sandstones are minor except at Port Pirie. This unit is confined to the western margin of the southern Flinders Ranges, where it gradationally overlies Unit 1 in the Emeroo Range and the Port Pirie area, and the Emeroo Quartzite in the Port Germein Gorge-Beetaloo area (Fig. 4.1).

Shales

Horizons of greenish-grey shales which weather to a yellowish-green colour, comprise 50-70% of this unit, and have sharp or gradational boundaries with the interbedded dolomites. Flat to slightly wavy lamination is due to an alternation of clay laminae with lenticular and sometimes rippled laminae of silt and sand (Plate 4.2a). Other features include rare synaeresis and desiccation cracks, and small soft sediment slumps. Possible gypsum pseudomorphs, now filled with quartz containing minor anhydrite inclusions, are preserved in one shale bed in Port Germein Gorge (Plate 4.2b). The sulphates grew within the sediment prior to compaction, as lath shaped crystals and rosettes.

The depositional environment of this facies was essentially the same as that for the siltstones and shales of Unit 1, although the ratio of clay to silt sized material was greater. The green colour is due to chlorite, probably largely of metamorphic origin, but the absence of iron oxides and the presence of minor pyrite, indicates that mildly reducing conditions prevailed within the sediment following deposition. However in contrast to Unit 1, carbonaceous material was completely removed.

Dolomite Mudstones

Dolomite mudstones, or dolomicrosparites in the terminology of Folk (1974a), are present as thin interbeds within shales, to horizons up to 6 m in thickness. Colour varies from light-grey to buff to pink, but becomes more reddish and yellowish on weathering. Flat to wavy lamination and thin bedding is due to variations in the terrigenous content, and in the grain size of the dolomite, which as in Unit 1, is often ferroan. Laminae of dolomite may be internally graded. Terrigenous sediment is up to coarse-grained sand size, and may be present in thin rippled beds (Plate 4.2c). Desiccation cracks and rare tepees (Beetaloo) are evidence of intermittent exposure. Erosion of desiccated beds produced lenses and interbeds of intraclastic dolomite, particularly at Beetaloo.

The environment of deposition was similar to that of dolomite mudstones in Unit 1. Dolomite mud was deposited largely from suspension along with some terrigenous clay, whereas minor sand and silt were introduced by current activity, possibly storm generated. The pale colours reflect the lack of both carbonaceous material, in contrast to Unit 1, and iron oxides, although minor pyrite may be present.

Stromatolitic Dolomites

Massive biostromes of buff to pale-pink dolomite (Plate 4.3a), 0.3 to 2 m in thickness, and frequently with undulating surfaces, contain tuberous columnar stromatolites, with vertical to horizontal growth and divergent branching. The stromatolites have been described in detail by Preiss, 1974b, pp. 197-201). The irregular columns contain smooth, continuous laminae, with a high degree of inheritence, and only slight microuncon-formities. The laminae consist of dolomicrosparite, often with grumous or clotted textures (Plate 4.3b). The origin of this fabric will be discussed in more detail for stromatolites of the Yadlamalka Formation (Chapter 6). Silt and sand laminae are minor. The dolomite, which is often ferroan, has experienced minor replacement by white chert. The biostromes, and rare bioherms, are associated with low energy facies, dolomite mudstones and shales, into which they sometimes pass laterally along strike. The interspace sediment between columns is predominantly mudstone and wackestone (Preiss 1974b), with occasional intraclasts derived from erosion of columns, and minor sand and silt.

The irregular column growth, smooth laminae with a high degree of inheritence, mudstone and wackestone as the interspace sediment, and the absence of desiccation and fenestral features, indicate that stromatolite growth occurred largely undisturbed in a low energy, submergent environment. According to Haslett (1976), columns which grew in sheltered environments during deposition of Cambrian carbonates in the Wirrealpa area, show regular lamination with high inheritence, whereas those of the open tidal flat, have more irregular lamination. However stromatolite growth was occasionally interrupted by periods of desiccation, followed by erosion of the stromatolite columns. The uniformity of stromatolites in this unit, indicatesthat growth occurred under similar conditions in all areas.

Stromatolites are most abundant in the Depot Creek area, where they exceed dolomite mudstones in abundance. The biostromes are generally enclosed within shales. Hence dolomite deposition took place predominantly in association with algal growth, indicating that algal activity, as well as trapping sediment, may have provided a favourable microenvironment for precipitation of dolomite (Schneider, 1977; Zamarreño, 1977).

Grain Supported Sediments : Carbonate and Terrigenous

Sandstones are minor in Unit 2 (Fig. 4.1), except in the Port Pirie area, where they often contain scattered quartz granules. Sandstones are fine- to medium-grained, and occasionally coarse-grained at Port Pirie, and often contain dolomite intraclasts. Intraclastic dolomites, also minor, are present as thin interbeds, but are rarely up to 0.5 m in thickness. They have a sandy matrix. Oncolitic dolomites are rare (Beetaloo, Port Germein Gorge).

Unit 2 : Environmental Summary

Unit 2 has a similar style of sedimentation as Unit 1, with largely submergent terrigenous mud flats passing into slightly shallower dolomite mud flats with fringing exposed areas. However biostromes of columnar stromatolites are more abundant in Unit 2, and are also more abundant in this unit, particularly in the Depot Creek area, than anywhere else in the Mundallio Subgroup. The minor sandstone content reflects the small quantities of sand supplied to the basin, with the exception of the Port Pirie area which may have been closer to source. In addition diagenetic processes within the sediment were sufficient to completely remove organic matter, either by simple oxidation, or by the activity of sulphate reducing bacteria. The presence of minor pyrite, and the lack of iron oxides, suggests that the latter process may have been at least partially involved. The greater removal of organic matter as compared with Unit 1, may reflect a slower rate of deposition of Unit 2, or better circulation so that basin waters were more oxygenated.

A sequence of shoaling upward, shale-dolomite cycles in the Rocknest Formation of Canada has been described by Hoffman (1975). The facies in Unit 2 probably occupy similar relative positions, although grainstones formed in the zone of wave agitation in the Rocknest Formation, are generally absent from Unit 2. Compared with the Rocknest Formation, cyclicity is less well developed in Unit 2, and both shallowing and deepening cycles are present (Fig. 4.5), indicating random variability in subsidence rates, depositional rates, and the influx of fine-grained terrigenous sediment. Shallowing cycles may commence with a thin sandy intraclastic bed which is erosional on the underlying desiccated dolomite mudstones (Depot Creek).

Unit 3

The uppermost unit in the Nathaltee Formation consists of a large (27-103 m in thickness) coarsening upward shale-sandstone cycle. The shale, which generally has a sharp base, is dark-grey, weathering to an olive-green colour. Fine detritus was carried in suspension and deposited in a region of only very weak current activity, resulting in the formation of poorly laminated shales. As deposition continued, the basin shallowed, and increased current activity produced laminated siltstones. This was followed by sand deposition, initially as plane-bounded interbeds within siltstones (Depot Creek) or as scour based channels (Port Germein Gorge, Fig. 4.6). The sandstones at the top of this unit are medium- to very coarse-grained, quartz-cemented, and moderately to well sorted. At Depot Creek, sedimentary structures include flat to low angle bedding, tabular cross-beds, 0.5-2 m in thickness, and symmetrical wave ripple marks with wavelengths up to 5 cm indicating generation by short period waves in a shallow water body of restricted size (Tanner, 1971; Komar, 1974). At Port Germein Gorge and in the Yednalue Anticline, the sandstones contain trough cross-beds (Plate 4.3c) up to 50 cm in thickness, and flat bedding (Figs. 4.1 and 4.6). Flat bedding dominates at Beetaloo. Limited palaeocurrent data from Depot Creek (Fig. 4.7), indicates a northwest-southeast shoreline, with some longshore reworking to the southeast.

On the west limb of the Yednalue Anticline, the poorly laminated shales at the base of Unit 3, and the sandstones at the top, are separated by a sequence of dark-grey to brown, massive or laminated, generally dolomitic shales and siltstones, with some interbeds of dolomite mudstone (Fig. 4.1). This is the only part of Unit 3 represented on the east limb of the Yednalue Anticline (Fig. 4.2). The environment of deposition was similar to that of the underlying massive, non-dolomitic shales, but this facies may reflect the introduction of detrital dolomite mud along with terrigenous.

Coarsening upward cycles similar to those in Unit 2, may be characteristic of prograding linear clastic shorelines (Fig. 6.5 in Selley, 1970), shallow marine sand bodies overlying offshore muds (Harms *et al.*,1975), or small prograding deltas (van Dijk *et al.*, 1978). Unit 3 may have resulted from the progradation of such deltas, and their reworking by waves within the basin. At Port Germein Gorge, trough cross-bedded sandstones were deposited in erosionally based channels, with some subsequent reworking by wave processes to produce flat bedded sandstones as the channels migrated elsewhere. At Beetaloo and Depot Creek, there was greater reworking by wave processes in the shoreline zone, due a position away from the main channels transporting sand into the basin. This produced flat bedded and low angle cross-bedded sandstones, with symmetrical ripple marks. Tabular cross-beds may have originated as offshore or along-shore migrating sandwaves.

NATHALTEE FORMATION : SUMMARY

Most of this formation represents alternating deposition of dolomite mudstones and shales, with minor deposition of sandstones, the abundance of which reflects proximity to source, and the available supply. The low gradients within the depositional environment produced a sequence with rapid vertical facies changes, as sediment buildups, changes in water level and energy level produced environmental, and hence facies changes. Terrigenous mudflats represented on the average, a slightly deeper water environment than carbonate mudflats, which were generally fringed by exposed areas subject to inundation and erosion during storms. The morphology of stromatolites, which grew largely in submerged environments, was influenced by environmental factors, although biologic influences cannot be excluded.

Unit 3 contrasts with the remainder of the Nathaltee Formation, in that it represents a thick coarsening upward sequence, whereas Units 1 and 2 are characterised by more rapid, and often random facies changes. This may reflect tectonic eventswhich produced sediment starved basins in which shales accumulated, but which subsequently produced an increased sand supply, so that small deltaic sand bodies prograded into the basin. A period of rapid subsidence relative to the rate of sediment influx and the rate of deposition, may have initiated these sediment starved, slightly deeper areas, with compensation in the adjacent source area producing uplift and an increased sand supply to the basin.

NANKABUNYANA FORMATION

INTRODUCTION

In the northern Flinders Ranges, the lower part of the Mundallio Subgroup consists of a sequence of interbedded siltstones, shales, sandstones and minor dolomites, the Nankabunyana Formation. This formation extends into the southwest Willouran Ranges, but northeast of the Norwest Fault, it is replaced by the Camel Flat Shale and Tilterana Sandstone (Fig. 3.5). In the Avondale (AV) and Mandarin Hill (MH) areas, the lower part of the Mundallio Subgroup is not exposed (Fig. 3.5).

The Nankabunyana Formation overlies the massive, white, ridge forming quartzites of the Copley and Wortupa Quartzites. The lower-most unit is often deeply weathered, and forms strike valleys (Copley to Myrtle Springs, southwest Willouran Ranges), and as a result the lower boundary is often poorly exposed. This boundary is defined at the top of the last massive quartzite outcrop of the Copley Quartzite. In the Arkaroola area, the boundary occurs at the transition from pale quartzites of the Wortupa Quartzite, to grey siltstones, and more poorly sorted, often muddy grey sandstones of the lower Nankabunyana Formation.

The boundary with the overlying Yadlamalka Formation is usually sharp, occurring at the change from terrigenous clastics at the top of the Nankabunyana Formation, to the more dolomitic sequence of the overlying Yadlamalka Formation. The boundary is marked by the introduction of mallee scrub which often grows on outcrops of the Yadlamalka Formation, in contrast with the relatively treeless areas occupied by the Nankabunyana Formation.

FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

The Nankabunyana Formation contains four discrete units (Uppill 1979), which are generally mappable (Figs. 4.8 and 9). These units are present in all areas, with the exception of the area east of the Paralana Fault at Arkaroola where Unit 2 is absent, and hence Units 1 and 3 have merged (Fig. 4.10).

Unit 1

The lowermost unit is a sequence of shales and sandstones, with rare lenticular conglomerates (Arkaroola only), and rare dolomites. Outcrop is often poor and deeply weathered.

Massive Shales

This facies occurs largely in the southwest Willouran Ranges, where it may form at least 65% of this unit (Fig. 4.11). The shales are weathered, yellowish to greenish in colour, and only indistinctly laminated due to little internal grain size variation. Lenticular laminae and thin interbeds of sand are rare. In the Copley-Myrtle Springs area, a 7-14 m thick horizon of massive grey shale is the only representative of this facies and apparently forms a continuous horizon between Copley and Myrtle Springs (Figs. 4.8 and 11).

Deposition occurred in an environment into which mud was introduced in suspension, and was virtually free from current reworking.

Laminated Shales and Siltstones

In the Copley-Myrtle Springs and Arkaroola areas, the fine-grained component of Unit 1 is generally well laminated. In the former area, this facies is deeply weathered, generally yellowish or occasionally reddish in colour, and may contain limonite pseudomorphs after pyrite. The lamination is generally flat. Lenticular, rippled or plane-bounded laminae and thin interbeds of fine- to medium-grained sand are common (Plate 4.4a).

In the Arkaroola region, siltstones and shales are generally grey due to the formation of metamorphic biotite, although east of the Paralana Fault, outcrops of this facies are very similar to those at Copley. From Munyallina Creek (MY) to Nudlamutana Hut (NH, Fig. 3.2), grey shales and siltstones contain scattered sand up to coarse-grained sand size, and may grade into muddy sandstones. However northward, as the proportion of shales and siltstones increases, this facies becomes lithologically more distinctive from the interbedded muddy sandstones. Sedimentary structures include flat lamination, occasional cross-lamination, and minor desiccation cracks. Metamorphism has resulted in the formation of scapolite in some shales north of Wywyana Creek (WC, Fig. 3.2), while north of the Needles (N) cordierite is present in some horizons, and rarely, others have been metamorphosed to fine mica schists.

Deposition of this facies occurred in a low energy, submergent environment, with intermittent exposure in the Arkaroola area. Weak current activity produced lenticular lamination and cross-lamination. Storm activity may have introduced sand, and this was reworked by wave action to produce isolated ripple lenses and cross-laminated sand (de Raaf *et al.*, 1977).

Medium-Grained Quartz Cemented Sandstones

This facies, which occurs in the Copley-Myrtle Springs area and the southwest Willouran Ranges, forms intervals up to 10 m in thickness (Fig. 4.11). Some sandstone beds have erosional bases on the underlying shales. Sandstones are moderately well, to well sorted, and largely medium-grained, although the grain size varies from fine- to coarse-grained sand. Occasionally mixing of two size fractions in a single bed produced poorer sorting. Grains are rounded, hence these sandstones are mature to supermature. However textural inversions (Folk, 1974a), such as rounded but less well sorted sand, suggests that rounding may not be produced in the final depositional environment.

The dominant sedimentary structure is flat lamination or thin to medium bedding (Plate 4.4b). Cross-beds occur as isolated sets within flat bedded sandstones, and have planar or erosional bases, laminae tangential to the base of the set, and occasional reactivation surfaces (Plate 4.4c). Sets are 10 to 40 cm, and rarely 1 m in thickness. Small scale cross lamination is also preserved. Symmetrical and asymmetrical ripple marks (wavelength of 5 to 15 cm, rarely 40 cm) occur within sandstone beds, or are preserved on their upper surfaces. The rounded profiles of asymmetrical forms, suggest that they are combined wave-current influenced ripple marks (Harms, 1969). Sets of interference ripples are minor.

Sedimentary structures indicate the influence of both oscillatory wave processes and unidirectional currents during the deposition of this facies. Palaeocurrent data is limited, and presented in combination with data for the remainder of the formation (Fig. 4.13). In the Copley-Myrtle Springs area currents were dominantly to the east, with minor northerly and southerly components.

The dominance of flat lamination reflects wave reworking of sand as a major depositional process (Reineck and Singh, 1973). Wave induced suspension clouds (Reineck and Singh, 1973; Button and Vos, 1976) may have been produced by increased turbulence associated with storms (Banks, 1973), which eroded sand and placed it in suspension. As energy levels decreased, the suspension clouds sank, producing flat bedded sandstones (Reineck and Singh, 1972; Goldring and Bridges, 1973; Kumar and Sanders, 1976). As conditions waned further, wave ripples and associated cross lamination may have been produced, although these features may also have resulted from wave action under fair weather conditions. Tangential laminae at the base of cross-beds indicate that they also formed from processes carrying some sand in suspension (Reineck and Singh, 1973). However the presence of cross-beds indicates unidirectional currents. Three main types of currents operate in shallow marine environments, semipermanent currents, tidal currents, and meteorological currents (Banks, 1973; Johnson, 1977). The latter are produced during storms, in which wave generated, unidirectional currents may produce cross-beds (Banks, 1973; Johnson, 1977). One such storm generated process is the offshore storm surge, which is an effective agent of transporting sediment (Freidman and Sanders, 1978), and is capable of transporting sand under dune and sand wave regimes although it may be too short-lived to produce large scale migrating bedforms (McCave, 1973; Johnson, 1977). However dunes and sand waves are produced off the eastern U.S.A. by unidirectional storm-generated currents (Johnson, 1978). Sand bars formed by wave processes may also contain abundant cross-bedding (Greenwood and Davidson-Arnott, 1979, Roep *et al.*, 1979). Hence although dunes and sandwaves, and the cross-beds produced, are more common in tide dominated seas, they may also be produced by wind-driven storm induced, or oceanic currents. The presence of reactivation structures implies bedform migration followed by erosion (Button and Vos 1976), and such structures are apparently common in tidal environments, although not exclusive to them. They may also result from two consecutive high energy events (Johnson, 1978).

Johnson (1978, p. 235) lists the main features indicative of tidal currents in shallow marine sands,

> "(1) bidirectional current-formed structures and bimodal palaeocurrent patterns with the two modes approximately 180° apart, reflecting flow reversals in rectilinear tidal currents, (2) multimodal palaeocurrent patterns reflecting either temporal fluctuations in direction or the rotary nature of tidal currents, possibly with superimposed storm activity, and (3) the abundance of cross-bedding reflecting dunes and sand waves which are today found mainly in tidal seas."

In contrast these sandstones contain dominantly flat bedding, formed by wave and storm processes, and cross-bedding is not abundant. Hence the environment was wave dominated, although it is difficult to determine if tidal processes were completely absent, even though they are not essential to form any of the structures present.

Fine-Grained Muddy or Quartz-Cemented Sandstones

This facies occurs in the Arkaroola area, in the southern part of which sandstones are grey and impure. The matrix of muscovite, chlorite and biotite is of metamorphic origin, but indicates the presence of a muddy matrix in the original sediment. The average grain size is fine-grained sand, but scattered grains up to granule size are often present. Hence sorting is poor to moderate. Northward the average grain size decreases to the coarse-grained silt to fine-grained sand size, with occasional medium-grained interbeds. Sorting also improves. Grey sandstones contain minor alumino-silicates, whereas paler sandstones are quartz-cemented and matrix free. Sedimentary structures include flat to wavy and wispy lamination and thin bedding, with occasional cross-lamination and symmetrical ripple marks (wavelength less than 10 cm).

East of the Paralana Fault, this facies is deeply weathered, but largely quartz-cemented. Sedimentary structures are similar, but occasional cross-beds and desiccation features are present.

The poorly sorted sandstones reflect deposition from intermittent currents capable of transporting sand and granule sized material, but too impersistent to winnow mud. Deposition may have occurred in terrestrial alluvial environments adjacent to the basin, and which were not subject to reworking by wave action. This environment graded northward into a shallow nearshore environment, where moderate turbulence as a result of wave action, produced more winnowed sandstones.

Conglomerates

Terrigenous conglomerates are mainly confined to the area south of Wywyana Creek at Arkaroola, and are well exposed in the Nudlamutana Hut area (Fig. 4.11). Conglomerates are a characteristic feature of the Burra Group in the Arkaroola area, being abundant in the Blue Mine Conglomerate, and significant in the Wortupa Quartzite in localised areas.

Conglomerates occur in isolated channels of rounded granule to pebble sized clasts (less than 3 cm, Plate 4.5a), or as horizons of several superimposed channels up to 2 m thick. The latter are continuous along strike for at least 600 m (outcrop limits). Conglomerates are erosive into the underlying siltstones and sandstones. The fabric is clast-supported, and lacks internal stratification and imbrication, although the latter feature is only common when clast long axes exceed 2 cm (Harms *et al.*, 1975). The occurrence of conglomerate filled channels within poorly sorted siltstones and muddy sandstones, suggests that they did not form in a high energy littoral environment. Rather, they may represent small braided channels which moved across an alluvial plain adjacent to the basin, and transported sediment into it. Here it was winnowed by wave processes, and lenses and thin beds of conglomerate near the top of this unit, may represent lag deposits formed by these processes in a shallow littoral environment (Kumar and Sanders, 1976).

Clasts within the conglomerates include acid volcanics, probably derived from acid volcanics similar to those preserved in the Mt. Painter Inlier and on the Curnamona Cratonic Nucleus (Giles and Teale, 1979), granitic, quartzitic and cherty varieties. The two former types are predominant. The abundance of pebble varieties may vary along strike, indicating localised source areas of exposed pre-Adelaidean rocks.

Dolomites

Dolomites are rare in this unit (Fig. 4.11) and include beds of weathered dolomite mudstone in the Willouran Ranges, and pale recrystallised dolomites at Arkaroola. A 3-4 m thick bed of brownish to reddish, wavy laminated dolomite mudstone, is probably continuous for at least 30 km in the Copley-Myrtle Springs area (Fig. 4.8). At Copley it is overlain by isolated, 2 m thick stromatolite bioherms, but elsewhere by an extensive, massive grey shale, previously described from this area.

Unit 1 : Environmental Summary

The termination of deposition of the Copley and Wortupa Quartzites was associated with a vastly diminished sand influx into the basin. In western areas, deposition of shales was associated with intermittent sand influx, possibly as a result of the reworking of lower Burra Group sands back into the basin. Sandstones and shales in this area do not appear to show any pronounced cyclicity, and both were deposited in submergent environments. Muddy sediments may accumulate in a range of shallow, near shore environments (McCave, 1971), depending on the concentration of mud in suspension, and the degree of turbulence. Mud deposition is however inhibited by persistent agitation (McCave, 1971). Hence the muddy facies accumulated in environments below effective wave base. Such environments could range from protected nearshore, to deeper offshore environments.

Sandstone interbeds probably represent submerged sandbars and sand-sheets, in which deposition under wave agitated, storm conditions was significant, resulting in flat bedded sandstones. The structures present, and the lack of features indicative of nearshore subtidal and intertidal environments, such as herringbone crossbedding, clay drapes, mud intraclasts and desiccation cracks (Klein, 1971; Johnson, 1977), suggests a shallow, nearshore shelf environment of deposition, dominated by wave processes. Unidirectional currents may have been entirely storm induced. However if tidal currents were present, this would imply some connection with the open ocean, as even large enclosed seas have negligible tidal currents (Anderton, 1976).

In the Arkaroola region, a low energy nearshore to terrestrial environment prevailed in the south. Coarse-grained sand, granules and pebbles were introduced from local sources, and deposited along with poorly sorted siltstones and muddy sandstones, on a low energy alluvial plain. This environment graded laterally into shallow nearshore environments. Well sorted fine-grained sandstones were deposited in the shoreline zone during periods of wave agitation. However deposition of finer clastics was also significant, indicating that low gradients may have significantly damped wave action along the shoreline.

Unit 2

This unit is characterised by more abundant dolomite interbeds (Fig. 4.11), and largely dolomite-cemented sandstones. Clastics still predominate however, with the exception of the southern part of the Arkaroola region, where dolomite may form up to 60% of this unit. In the western regions, the boundary between Units 1 and 2 is also defined by the introduction of fresher, greenish-grey shales, and a decrease in the grain size of sandstones. The abundance of dolomite interbeds decreases upwards through Unit 2.

Shales and Siltstones

Shales and siltstones are generally more abundant than other lithologies, and resemble in appearance the shales of Unit 2 of the Nathaltee Formation, largely because of the pale green-grey colour due to the presence of chlorite. At Arkaroola, this facies is grey because the higher metamorphic grade has resulted in the development of biotite, along with muscovite and chlorite. Scapolite, and occasionally cordierite, may also be present.

This facies is well laminated, with the lamination being flat to wavy and more irregular (Plate 4.6a and b). Coarser laminae of silt and sand are often lenticular. Irregular lamination, and small scale slumping and disruption (Plate 4.6a and b), may be the result of several processes including rapid deposition, synaeresis, dewatering, and differential compaction of clay and silt laminae in a saturated sediment. Small erosional scours have a relief of 1-2 cm. Desiccation cracks are fairly common at Arkaroola, and Top Mount Bore (TM) in the Willouran Ranges, but appear to be rare to absent in the Copley-Myrtle Springs area. Synaeresis cracks are present in this latter area (Plate 4.6b).

Flat laminated shales were largely suspension sediments, while weak bottom currents produced more lenticular laminae and small scours. Lenticular sandy beds, some of which are cross-laminated, and starved ripples, were probably deposited during storms (de Raaf *et al.*,1977). The environment was predominantly submergent in the Copley-Myrtle Springs area, but elsewhere, periods of exposure were more frequent.

Sandstones

Sandstone interbeds are generally less then 5 m in thickness, and occasionally up to 12 m, and have sharp, and less commonly gradational boundaries with interbedded shales. Thin interbeds and laminae of shale and dolomite mudstone, sometimes desiccated, are minor. Sandstones, in which quartz cement is dominant, are white in colour, whereas those with dolomite cement are light-grey to grey, and weather to a dark-brown. The average grain size is very fine-grained sand, and sometimes coarse-grained silt, with laminae and interbeds of medium- to coarse-grained sand being most common in the southern part of the Arkaroola region. Coarser beds are often poorly sorted, and contain dolomite intraclasts. However the very fine-grained sandstones are moderately well- to well-sorted. The composition is generally subarkosic, however in the Top Mount Bore area, the finer fraction is more feldspathic than the coarser fractions, possibly suggesting different sources. In the Arkaroola area, increasing metamorphic grade northward, has resulted in the metamorphism of dolomite cemented sandstones to calc-silicates containing scapolite, actinolite, quartz and dolomite.

Sandstones are generally flat or wavy laminated, the latter may occur in association with symmetrical wave ripple marks. Wavy lamination occasionally develops into ripple cross-lamination which may be erosional into flat lamination (Plate 4.5c). The cross-lamination resembles that formed by current ripples in silt sized sediment (Fig. 3.7 in Harms *et al.*, 1975), but there are also similarities to wave influenced ripples (Harms *et al.*, 1975; de Raaf *et al.*, 1977). Ripple forms preserved on bedding planes are generally symmetrical wave ripples, with rounded profiles and straight crests. Slightly asymmetrical ripple marks also have rounded profiles indicating some wave influence (Harms, 1969). Wavelength is small (range of 2-7 cm, but generally less than 5 cm). There are minor sets of interference ripple marks, generally with polygonal patterns. Tabular cross-beds, which occur in coarser grained sandstones, are uncommon. They have amplitudes of less than 30 cm, and planar or slightly erosional bases. Sets are isolated, with only rare cosets (Plate 4.5b). Sand filled desiccation cracks in thin shale or dolomite laminae, are abundant in the Arkaroola region, but less common elsewhere. Synaeresis cracks are present in the western areas.

This facies was deposited in shallow water environments subject to frequent periods of exposure at Arkaroola, but which were more persistently submergent elsewhere. Sandstones represent environments of more persistent turbulence than the interbedded shales, in which mud was winnowed from coarse-grained silt and very fine-grained sand. Slack water conditions preceded periods of exposure, and thin clay laminae were deposited and subsequently desiccated. Both oscillatory wave action, and unidirectional currents producing ripple cross-lamination and rare cross-beds, were active in this environment. Flat lamination may be the result of wave action, possibly storm induced, during which sand was moved in suspension (see discussion for sandstones of Unit 1). The wavelength of ripple marks and set thickness of crossbeds is smaller in Unit 2 than in Unit 1, and may reflect the finer grain size, a shallower environment and smaller wave fetch. The small ripple spacing indicates that short period waves were produced in either a restricted body of water, or a shallow water body with limited fetch (Tanner, 1971; Komar, 1974; Friedman and Sanders, 1978).

Dolomite Mudstones

With the exception of the Arkaroola region, where many dolomite interbeds are significantly recrystallised, mudstones represent 50-75% of the dolomite component of this unit. Colour ranges from light-grey and grey (most common), to buff and orange, generally with a brown outcrop surface. Texturally, the mudstones are dolomicrosparites, but have been recrystallised to finely crystalline dolomite at Arkaroola. The dolomite is often ferroan. Flat to wavy and lenticular lamination to very thin bedding is characteristic, and generally reflects variation in the content of terrigenous impurities (clay to coarse-grained sand). Some laminae are graded, either with a decreasing abundance and grain size of terrigenous material, or with a decreasing grain size of dolomite. Laminae of pelsparite may also be present. Other features include small erosional scours, desiccation cracks (Plate 4.7a), small soft sediment slumps, and rare quartz filled rosettes, pseudomorphing an evaporite mineral, possibly gypsum (Top Mount Bore only, Plate 4.6c).

Dolomite mudstones were deposited on carbonate mudflats subject to frequent periods of exposure. The texture of the mudstones is similar to that of terrigenous shales, indicating deposition from suspension and weak traction currents. Although the dolomite may have originated as a chemical sediment, the similarity in texture to fine grained terrigenous sediments, indicates that much of the dolomite may have been reworked, either by erosion of previously deposited muds, or from the abrasion of intraclasts which are characteristically well rounded. More wavy laminated mudstones may have resulted from the greater influence of algal mats.

Dolomite Intraclastic Grainstones

This facies occurs as thin lenticular interbeds with elongate, angular intraclasts within dolomite mudstones, or as thicker beds (rarely up to 2 m) of rounded granule and small pebble sized intraclasts. The latter are indistinctly bedded, or rarely cross-bedded (Plate 4.7b). They overlie other dolomite facies, dolomite mudstones or stromatolitic dolomites, sometimes erosionally, or terrigenous facies. According to the Folk (1974a) classification, this facies ranges from sandy to extremely sandy intrasparudites, and grades into dolomitic sandstones. The sand is medium- to very coarse-grained and well rounded. Within a single sample, the intraclasts have variable textures, indicating that they were derived from several sources.

The abundance of intraclastic dolomite indicates that dolomite mudflats were subject to more frequent and extended periods of exposure than the environments in which terrigenous sediments were deposited. Beds of angular elongate intraclasts which are commonly associated with desiccated beds, were probably produced when intense storms caused inundation and erosion of the mudflats. However in thicker beds, the rounding and small size of intraclasts, the relatively good sorting, and the association with coarse-grained sand, suggest that they represent turbulent conditions over an extended period, and of higher energy than that producing the interbedded sandstones. Transgressive events, encroaching across the mudflats, may have caused erosion of lithified dolomite mudstones, and the reworking of intraclasts in the surf zone.

Stromatolitic Dolomites

Stromatolites occur in domal or more tabular bioherms. The stromatolites are most frequently tuberous, vertical to inclined columns (Plate 4.7b and c), although wavy laminated and domal forms also developed. Two specimens of columnar stromatolites are illustrated by Preiss (1972, his Figs. 6i and j), and are described as *Baicalia burra*. The columns consist of alternating darker laminae of micrite and microspar, and lighter laminae of finely crystalline dolomite. Terrigenous silt to fine-grained sand is more abundant in the coarser laminae, which may infill desiccation cracks in the micrite laminae. This texture is similar to that described by Horodyski (1976), suggesting that the coarser laminae are detrital, and the finer laminae formed partly by *in situ* precipitation of carbonate within the algal mat. Erosion of the stromatolite columns is indicated by microunconformities.

Interspace sediment between columns includes sandy intrasparudite, dolomitic sandstone, and minor lenses of dolomicrospar. Some of the intraclasts have clearly been derived from the adjacent stromatolite columns (Plate 4.7c). The sand is fine- to coarse-grained, hence it appears that stromatolite growth may have only trapped the finer detritus, or that coarser sand was introduced during high energy events in which erosion of columns rather than growth took place. Stromatolite growth may have occurred under quieter water conditions, during which dolomite mud and minor terrigenous silt and fine-grained sand were available. Desiccation cracks indicate periods of exposure, during which growth was halted, and often the desiccated algal mats were subsequently eroded. The intraclasts deposited between columns are angular, indicating that storm periods were of insufficient duration to round and sort the intraclasts. However stromatolite bioherms sometimes overly high energy intraclastic dolomites with well rounded intraclasts (Plate 4.7b).

Dolomite Marbles

In the Arkaroola area, north of Arkaroola Homestead (AH, Fig. 3.2), all dolomites within this unit have been metamorphosed to white, wavy bedded, dolomite marbles. They consist of coarsely recrystallised dolomite, with minor talc, tremolite, orthopyroxene, phlogopite and sphalerite.

Massive Iron Stained Dolomites

In the southern part of the Arkaroola region, depositional textures in dolomites are generally preserved. However near the base of Unit 2, thick dolomite horizons (25-35 m) consist of massive, recrystallised, brownish dolomites, overlain by flat-bedded, slightly recrystallised dolomite mudstones (Fig. 4.12). The massive dolomites contain an irregular lamination, apparent on weathered surfaces (Plate 4.8a), which is not primary bedding. This lamination is defined by irregular concentrations of aphanitic iron oxides, concentrated on crystal grain boundaries. Depositional textures have been largely destroyed, but when the trace of bedding is preserved, it is cut by the irregular iron rich laminae. Irregular, scattered fenestrae have been filled with iron stained, zoned dolomite crystals. Partial silicification has produced fine-grained quartz, and minor length-slow chalcedony. Within the dolomites, are pods, lenses and irregular areas generally less than 1 m in size, infilled with aphanitic iron oxides containing scattered dolomite crystals, sometimes replaced by quartz, and minor detrital material (Plate 4.8b and c).

The close proximity of recrystallised dolomites to only slightly recrystallised dolomite mudstones, suggests different depositional or diagenetic histories which have enhanced recrystallisation in the massive dolomites, either during diagenesis or metamorphism. Aphanitic iron oxides within the dolomites have been redistributed along grain boundaries during recrystallisation. The iron oxides may have originally been concentrated within the sediment during subaerial exposure and weathering. During exposure, dilute meteoric solutions may have caused recrystallisation, and dissolution and disaggregation of the crystalline dolomite. Irregular solution cavities which formed within the rock were infilled by iron oxides, silt-sized dolomite fragments, and minor terrigenous detritus. Similar sediments have been described by Dzulinski and Kubicz (1971) and Siedlecka (1976). Smaller cavities were infilled with iron stained rhombs, probably precipitated from dilute solutions.

Hence this facies may indicate extended periods of exposure and subaerial weathering of dolomites. Because of metamorphic recrystallisation of dolomites to the north, the extent of the area subject to exposure cannot be determined.

Unit 2 : Environmental Summary

The environments of deposition of sandstones and shales do not appear to have been of significantly different water depth in any given area. This may indicate that the "energy of coastal processes" and sediment supply (Beall, 1968), were important in determining whether finer or coarser sediments predominated in a given area at a given time. However dolomitic mudstones and stromatolitic dolomites appear to have been subject to more extended periods of exposure, and hence occupied more shoreward environments, possibly carbonate lagoons with fringing exposed mud flats (c.f. Nathaltee Formation).

Sand transportation and deposition was effected by a combination of wave-induced processes and minor unidirectional currents, whereas shales and siltstones represent environments in which the energy levels of these processes was much less. Although the environment was shallow subaqueous to emergent, many of the features listed by Klein (1971) as characteristic of subtidal and intertidal environments influenced by tidal transportation processes, are not present. However Klein's model has been proposed largely from studies of modern tidal flats which contain elaborate tidal channels, and have developed behind barrier islands or in protected coastal embayments to which there is a significant sand supply (Thompson, 1975). In contrast, tidal flats associated with the Colorado River Delta lack many of the features typical of tidal environments, despite a large tidal range of 4-5 m (average). Because of the large supply of silt and clay to this area, and the low wave energy, there is insignificant barrier development, and tidal currents move across the flats in a broad uniform flow, rather than forming channels (Thompson, 1968, 1975). As a result, sediments deposited in the subtidal to intertidal zones are fine grained, well laminated suspension deposits, in which features indicative of alternating tidal current bedload and suspension settling during slack water (Klein, 1971) are lacking. Coarser, sandy sediments form near the high water level. During periods of reduced mud supply, erosional transgressions occurred. Wave action had sufficient time to rework muddy sediments, and produced transgressive sand sheets (Thompson, 1968). Hence the alternation of sandy and muddy sediments may largely reflect the rate of supply of muddy detritus, and the ineffectiveness of winnowing processes when the rate of supply is high.

Intertidal and subaqueous mudflats have also developed in areas of high mud supply and low tidal range, and under the influence of moderate wave action, on the Louisiana coastline (Beall, 1968). The environments vary from "tidal mudflat" to "transitional beach" to "normal beach", depending on the interaction between sediment supply and the energy of coastal processes (Beall, 1968, Fig. 9). In both this example, and the Colorado delta tidal flats, progradation of mudflats is occurring. Sandy facies may develop without a change in sea level, but as a result of reduced sediment supply.

Tidal flats may develop under influences other than astronomic tides. For example the Laguna Madre flats are sometimes inundated by tides of 30-50 cm amplitude, which are wind induced (Miller, 1975). A variety of facies are produced in this environment due to slight irregularities in the depositional surface, and variable influx of sand and mud into different areas. Algal mats and carbonate mud interbeds are also present.

Although none of these examples may be directly analogous to the sequence in Unit 2, they illustrate the range of conditions under which very shallow water, mixed muddy-sandy sequences may develop, and the importance of sediment supply and the effectiveness of sedimentary processes. Variation in these factors without changes in water level, may result in facies changes.

The small wavelength of ripple marks in this sequence, suggests either a small body of water or low wave energy, and shallow water. If depositional gradients were very low in parts of the basin, wave energy would have been substantially damped and a surf zone would not have formed. Under such conditions, shale deposition may have occurred in near shore environments. Sediment supplied to the basin was largely clay to very fine-grained sand size. A high rate of influx combined with a low wave energy, may have resulted in deposition of siltstones and shales close to the source. Winnowing processes may have been more effective further from the source, or during periods of reduced sediment influx, thus allowing winnowing of clay from the coarse-grained silt and very fine-grained sand.

As sedimentary structures within the sandstones may be attributed to wave induced processes, and unidirectional currents appear to have been minor agents of sediment transport on the shallow sand and mud flats, tidal currents were probably insignificant. Wind tides may have been effective agents in causing water level fluctuations. Because of low depositional gradients, small changes in water depth would have exposed or inundated large areas.

Unit 3

The abundance of dolomite in Unit 2 decreases upwards (Fig. 4.11), and Unit 2 is replaced by a sequence of clastic sediments with only rare dolomite interbeds (Unit 3). The proportion of sandstone interbeds also increases from Unit 2 to Unit 3.

Shales and Siltstones

This facies is essentially the same as the shales and siltstones of Unit 2, and has a similar suite of sedimentary structures. This includes flat to wavy and irregular lamination (Plate 4.9a), lenticular coarse silt and sandy laminae and thin interbeds which may be cross-laminated, small erosional scours, desiccation cracks (abundant in the lower part of the unit at Arkaroola, but rare elsewhere), synaeresis cracks (Plate 4.9a, Copley to Myrtle Springs), small scale slump folds, and disruption and intermixing of silt and clay laminae.

The environment of deposition of this facies, which occurs as interbeds up to 9 m in thickness, was essentially the same as the equivalent facies in Unit 2.

Sandstones

In the Copley-Myrtle Springs and Top Mount areas, sandstones are generally white, quartz-cemented, and very fine- to fine-grained, with minor laminae and interbeds of medium- to coarse-grained sand. They are slightly coarser grained than the sandstones of Unit 2. Sedimentary structures are essentially the same as in the sandstones of Unit 2, and are dominated by flat to wavy lamination, sometimes developing into ripple cross-lamination (Plate 4.9b). Desiccation cracks are rare to absent.

In the Arkaroola area, both quartz- and dolomite-cemented sandstones are present. The mineralogy of the latter is increasingly altered to calc-silicate minerals, including actinolite and scapolite, north of Wywyana Creek. However lamination and cross-lamination is preserved. Sandstones dominate the outcrop of Unit 3 in the southern part of this area, but shales become more abundant northward. This suggests that the influx of fine clastics was less in the south, and extensive mudflats of low wave energy developed only infrequently.

Sedimentary structures within the sandstones of Unit 3 suggest that the environment of deposition was similar to that of sandstones in Unit 2, although periods of exposure were less frequent, particularly in the western areas.

Unit 3 : Environmental Summary

Unit 3 represents clastic depositional environments similar to those described for Unit 2. The transition from Unit 2 to Unit 3 represents a change from environments fluctuating between shallow water clastic and carbonate deposition, to a situation dominated by clastic deposition. This was associated with less frequent periods of exposure, suggesting that a slight increase in average water depth caused the more restricted environments in which dolomite was deposited, to migrate away from the present areas of outcrop. Wave and current formed structures in this unit are similar in size to those of Unit 2, suggesting that the available wave fetch was not significantly increased.

Unit 4

A poorly laminated shale, which generally has a sharp boundary with Unit 3, forms the lower part of Unit 4. It coarsens upwards into siltstones and fine-grained sandstones with minor dolomite interbeds (Fig. 4.11).

The shale horizon varies in thickness between different areas, but is internally homogeneous and lithologically similar in all areas of outcrop.

It resembles the shale in the lower part of Unit 3 of the Nathaltee Formation. A greenish-grey colour in the western area is due to the presence of muscovite and chlorite, whereas a grey colour at Arkaroola reflects the presence of biotite due to the higher metamorphic grade. Minor quartz silt and authigenic pyrite are present. Rarely interbeds of dolomite mudstone are present, and the shale is slightly dolomitic (Myrtle Springs, Top Mount Bore).

The homogeneity of this shale in all areas, suggests deposition of clay sized sediment from suspension in an area of very weak current activity. Deposition probably occurred below wave base. Hence the transition to slightly deeper water environments from Unit 2 to Unit 3, was continued into the lower part of Unit 4.

At Copley, dolomite-cemented siltstones and very fine-grained sandstones at the top of Unit 4, form the upper part of a simple coarsening upward cycle. They are initially flat laminated, but become largely ripple cross-laminated as the grain size increases. Small cross-beds up to 15 cm in size are present at the top of the unit. The cross-lamination occurs as complexly interwoven, variably directed sets, with scooped shaped erosional bases (Plate 4.9c). Hence ripple profiles are often partly eroded or buried by more sand, so that bedding plane exposures of ripple mark sets are often incomplete. Those preserved have rounded, symmetrical to slightly asymmetrical profiles (Plate 4.10a), although the internal lamination is often unidirectional. Lenses of dolomite mudstone sometimes define a weak flaser lamination.

Ripple cross-lamination originated as combined current and wave influenced structures (Harms, 1969; Harms *et al.*, 1975; de Raaf *et al.*, 1977). Thicker sets of low angle cross strata resemble the hummocky stratification of Harms *et al.* (1975) attributed to storm wave surges. Measurements of ripple mark orientations indicate predominantly easterly flowing currents (Fig. 4.13). These may have been formed by offshore directed storm generated waves, and as energy levels waned, were subsequently covered by dolomite mud, also transported offshore by the storm waves. Hence they had a higher preservation potential.

At Myrtle Springs, medium-grained sandstones with isolated tabular cross-beds are interbedded in finer grained sandstones similar to those described above. In the south-western Willouran Ranges, medium- to coarse-grained sandstones with well rounded dolomite intraclasts, contain tabular cross-beds indicating northeasterly directed currents with minor reversals. The presence of coarser grained sand in these areas may reflect closer proximity to the source.

In the Arkaroola area the upper part of Unit 4 is more lithologically variable, and generally consists of interbedded grey dolomite-cemented sandstones (metamorphosed to calc-silicates in the north), white quartz-cemented sandstones and grey shales, and hence is similar to the underlying Unit 3. However in the Blue Mine Creek area, the massive shale is overlain by white sandstones which coarsen upward from fine- to medium-grained, with ripple cross-lamination at the base, and small scale tabular cross-bedding at the top.

The sandstones forming the upper part of Unit 4 were deposited in submergent environments in response to an influx of sand and as the environment shallowed into one of persistent wave agitation. Transportation was by a combination of uni-directional currents and wave oscillation processes. The former may have been produced by offshore storm surges. Dolomite mud laminae were deposited during slack water periods in more distal environments, and may have been derived from erosion of dolomite mudflats in more landward areas, where deposition of the Yadlamalka Formation had commenced. However at Arkaroola, terrigenous mud was still available, and accumulated in low energy environments.

NANKABUNYANA FORMATION : SUMMARY

Because of the predominance of flat laminated sandstones, and the minor occurrence of cross-beds, palaeocurrent data for the Nankabunyana Formation is limited (Fig. 4.13). However it is generally consistent within a single area. At Copley, unidirectional currents were largely easterly flowing, possibly offshore and associated with ebb-current storm surges (Von Brun and Hobday, 1976; Johnson, 1977), and hence may indicate a north-south shoreline. Minor northerly, along shore currents were also present. Symmetrical wave ripple marks do not provide a unique direction, but indicate east-west wave oscillation. Symmetrical ripple marks at Arkaroola are consistent with those at Copley. Because of the shallow depositional environment in these areas, the orientation of symmetrical ripple marks may have been influenced by the prevailing wind direction. In the southwestern Willouran Ranges, north-northeast to south-southwest wave oscillation suggests that the north-south shoreline in the Copley-Myrtle Springs area had swung around to an east-southeast to west-northwest orientation. The different orientation of symmetrical ripple marks in this area as compared with Copley, suggests that palaeobathymetry and the orientation of the shoreline in the Top Mount Bore area influenced the direction of wave oscillation, and this was not solely controlled by the prevailing wind direction.

Deposition of the Nankabunyana Formation occurred in similar environments in all areas of outcrop, resulting in the development of similar facies, and a similar vertical arrangement of facies. However the environment of deposition was shallowest in the Arkaroola area, as indicated by probable terrestrial facies in Unit 1, extended periods of exposure causing weathering of dolomites in the lower part of Unit 2, and more abundant desiccation cracks in clastic sediments.

The Nankabunyana Formation is predominantly a clastic sequence, with dolomites only abundant in Unit 2. Dolomite facies represent more shoreward environments with little sediment influx, and which were subject to more extended periods of exposure. The desiccated crusts produced were eroded to produce dolomite intraclastic grainstones. In contrast, desiccated terrigenous mudstones were rarely eroded. Clastic sediments consist of an alternation of shales and fine-grained sandstones, the deposition of which was influenced largely by wave-induced processes. Features typical of tidal sequences are absent. Wind tides may have been effective in causing small water level fluctuations. As in the Nathaltee Formation, this sequence is characterised by frequent and often abrupt vertical facies changes, generally with little cyclical arrangement of facies. However shallow water environments of similar water depth were persistently present, reflecting an overall balance between the rates of deposition and subsidence. The alternation between high and low energy facies reflects a complex interplay between sediment supply, and the effectiveness of sedimentary processes (e.g. winnowing by wave agitation), rather than systematic depth changes.

Large scale coarsening upward cycles (Unit 4), are atypical of the remainder of the Nankabunyana Formation. This unit reflects the development of a deeper environment of deposition below effective wave base, where current influence was minimal. Hence the shales deposited are now poorly laminated. This may have been initiated by a period of more rapid subsidence and possibly an expansion in the size of the basin, so that shoreline, wave-agitated environments were not recorded in the present outcrop areas. The relatively rapid transition from an environment below wave base, undisturbed by currents, to the agitated zone above wave base, which occurs in Unit 4 (Fig. 4.11), and in Unit 3 of the Nathaltee Formation, reflects generation of waves in a restricted body of water, analogous to wave generation in lakes. In the marine environment where the available wave fetch is much larger, the sediment bottom can be agitated below wave base (Picard and High, 1972), and the sediments deposited here will contain evidence of weak, intermittent current or wave activity.

CAMEL FLAT SHALE AND TILTERANA SANDSTONE

INTRODUCTION

In the Willouran Ranges, the Nankabunyana Formation is replaced laterally across the Norwest Fault by the Camel Flat Shale and the overlying Tilterana Sandstone (Fig. 3.5). These formations are readily recognizable in those areas where the Camel Flat Shale overlies the Copley Quartzite, that is in the eastern Willouran Ranges from Mt. Norwest H.S. (NW) northward, and in outcrops immediately northeast of the Norwest Fault between Cadnia Hill (CH) and Coronation Bore (CO). In these areas, the Camel Flat Shale forms a strike valley, with an adjacent parallel ridge formed by the Tilterana Sandstone. However in the north-central Willouran Ranges, near Rischbieth Hut (R), Mirra Creek (MI) and Mirra Bore (MR), the Copley Quartzite is absent, and the top of the underlying Witchelina Subgroup consists of a sequence of grey sandstones, siltstones, shales and dolomites, the Willawalpa Formation (Murrell, 1977). These lithologies are similar to those in the overlying Mundallio Subgroup, which is represented largely by the Mirra Formation (Fig. 3.5). This formation contains less sandstone and more dolomite than the Willawalpa Formation. Immediately underlying the Mirra Formation is a ridge of sandstones with minor grey dolomites and siltstones (80-140 m in thickness), which has been mapped by Murrell (1977) as the Tilterana Sandstone. This ridge is underlain by a poorly outcropping interval of shales and siltstones with minor sandstones and dolomites of similar thickness as the sandstone ridge, and which has been mapped as the Camel Flat Shale. However both these horizons contain facies similar to the Willawalpa Formation, which also consists of alternating sandstone ridges and valleys of less massive lithologies. Hence the correlation of these two particular horizons with the Camel Flat Shale and Tilterana Sandstone, is based solely on their stratigraphic position.

In the north-central Willouran Ranges, both the Witchelina Subgroup and the Mundallio Subgroup show marked stratigraphic thinning and facies changes towards areas of outcrop of the Callana Group (Murrell, 1977). Both the Camel Flat Shale and the Tilterana Sandstone are involved in these changes.

CAMEL FLAT SHALE : FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

This formation, which is of variable thickness in the Willouran Ranges (40-240 m, see also Murrell, 1977, Fig. 17), is invariably poorly outcropping due to the presence of shales as the dominant lithology, although creek exposures in the type section near Mt. Norwest H.S. may be relatively continuous. In this area, the Camel Flat Shale transitionally overlies the Copley Quartzite, with the lowermost 20-30 m consisting of grey weathered shales with thin to medium, lenticular and plane bounded beds up to 1 m in thickness, of quartz cemented sandstones (Fig. 4.11). The lenticular beds may have a rippled form, but do not show any internal lamination. Thicker beds contain indistinct flat bedding and cross-bedding. The transition zone is overlain by a homogeneous, dark-grey (although generally weathered greenish-grey), pyritic shale, which is very poorly laminated but is coarsely cleaved. Lenticular beds (30 cm to 1 m thick) of dark-brown to dark-grey dolomite mudstone are also internally homogeneous. In the uppermost 20 m of the Camel Flat Shale, shales coarsen into siltstones which are overlain by the fine-grained sandstones of the Tilterana Sandstone.

Outcrops of the Camel Flat Shale between Cadnia Hill and South Hill are very poor, and consist of weathered shales. However in the Mirra Creek and Rischbieth areas, the formation consists of dark-grey shales and siltstones which are often dolomitic, with minor interbeds of dark-grey dolomite mudstones and stromatolitic dolomites, and grey very fine-grained sandstones with wave ripple marks.

The massive poorly laminated shales are lithologically similar to some shales in the Nankabunyana Formation, including massive shales in Unit 1, and shales forming the lower part of Unit 4. The depositional environment was one starved of silt and sand sized sediment. Clay, winnowed from adjacent shallower more agitated regions, was carried in suspension, and deposited in a reducing environment below wave base where current and wave reworking was negligible. In the north-central Willouran Ranges, the environment was slightly shallower, allowing the deposition of minor wave-rippled sandstones and stromatolitic dolomites.

TILTERANA SANDSTONE : FACIES DESCRIPTION AND DEPOSITIONAL ENVIRONMENTS

The type section of this formation is also located in the Mt. Norwest H.S. area, where the sequence present consists almost exclusively of sandstones (Fig. 4.11). Outcrops in the region immediately northeast of the Norwest Fault contain similar facies, but are more poorly outcropping.

The sandstones are white to light-brown, quartz-cemented, although with minor late stage dolomite cement, very fine- to fine-grained, and moderately well sorted. Medium- and coarse-grained sand is present in thin lenticular and planar interbeds, or as lag deposits above erosional scours (Plate 4.10b). Sedimentary structures include flat lamination, low angle cross-lamination, ripple cross-lamination with complexly interwoven bidirectional sets, and rarely climbing ripples (Plate 4.10b and c). Symmetrical ripple marks and rare sets of interference ripples have small wavelengths (1-5 cm). These sedimentary structures are similar to those in sandstones of similar grain size in the Nankabunyana Formation. Deposition occurred in a shallow, largely submergent environment, as indicated by the rare desiccation cracks, under the influence of wave processes. Oscillatory wave processes may have prevailed under fair weather conditions, while unidirectional currents may have been generated during storm conditions.

Minor shale partings, and thin siltstone beds in sandstones, represent deposition under more slack water conditions. Thicker siltstone interbeds (1-8 m), with thin interbeds of sandstone, may represent more offshore environments, into which sand was occasionally introduced during storms.

In the north-central Willouran Ranges, light-grey quartz-cemented sandstones also dominate the Tilterana Sandstone. However the lamination within them is indistinct, but may be planar to wavy. Occasional symmetrical ripple marks are present. Some sandstone interbeds are dolomite-cemented, and may grade into dolomitic siltstones which are darker grey in colour. Minor interbeds of dark-grey dolomite mudstone are present. Deposition in this area also occurred in a shallow water environment which was persistently agitated, probably by wave processes.

The influx of very fine-grained sand into the north-eastern half of the Willouran Ranges which led to the deposition of the Tilterana Sandstone, may be correlated with the influx of very fine-grained sand into adjacent areas where Unit 2 of the Nankabunyana Formation was deposited. Sandstones in Unit 1 of the Nankabunyana Formation are largely medium-grained in the southwestern Willouran Ranges and the Copley-Myrtle Springs area.

CHAPTER 5

THE MUNDALLIO SUBGROUP IN THE MT. LOFTY RANGES (SKILLOGALEE DOLOMITE, CASTAMBUL FORMATION AND WOOLSHED FLAT SHALE): FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

GENERAL INTRODUCTION

In the Mt. Lofty Ranges, the Mundallio Subgroup consists largely of thick intervals of pale coloured recrystallised dolomites and laminated siltstones and shales, which contain only minor interbeds of other facies. A subdivision of three formations is present, the Skillogalee Dolomite, Castambul Formation, and Woolshed Flat Shale (Figs.3.1, 3.3). In addition, in the Adelaide region, there is a lenticular unit of dark-grey dolomite, magnesite and sandstone, the Montacute Dolomite, which is characterised by more frequent vertical facies changes. This formation will be discussed in conjunction with the Yadlamalka Formation in Chapter 6.

Hence in contrast to both the northern and southern Flinders Ranges where the Mundallio Subgroup is characterised by frequent vertical facies changes, often between clastic and carbonate facies, in the Mt. Lofty Ranges area either carbonate, or fine clastic deposition prevailed over long periods of time.

Apart from occasional road cuttings, quarries, and creek exposures, outcrop of these formations is poor, and sometimes structurally complex, so that the distribution of formations and intertonguing relationships cannot always be precisely defined.

SKILLOGALEE DOLOMITE

INTRODUCTION

In the westerly outcrops, between Spalding (S) and Tarlee (T, Fig. 3.2), the Skillogalee Dolomite overlies the Rhynie Sandstone, which in the Skillogalee Creek (SC)-River Wakefield (RW) area, consists of heavy mineral laminated pink quartzites, with siltstone interbeds and minor dolomites. The contact in this area is generally poorly exposed. Southwest of Spalding, siltstones are more abundant than sandstones in the Rhynie Sandstone, and minor limestone interbeds occur at the top (Preiss, 1974a). The contact with the Skillogalee Dolomite may be a disconformity (Preiss, 1974a). In the area between Burra (B) and Scrubby Range (SR), the base of the Skillogalee Dolomite is not exposed. The formation outcrops in an anticlinal structure centred on a brecciated zone associated with the Kooringa Fault. Blocks of heavy mineral laminated sandstone, lithologically similar to the Rhynie Sandstone, are present in breccias in the Burra Mine area (Wright, 1976), suggesting that the Rhynie Sandstone may have been deposited in this area. The lower boundary of the Skillogalee Dolomite represents an abrupt facies change, from largely clastic deposition to carbonate deposition.

The Skillogalee Dolomite is 200 to 350 m in thickness in the area between Tarlee and Spalding (Fig. 3.3). In the Burra-Scrubby Range area, because the base is not exposed, and due to the presence of complex folding and faulting, accurate thickness estimates are not possible. However the Skillogalee Dolomite is significantly thicker in this area, and may reach 1000 m in thickness.

This formation consists largely of pale recrystallised dolomites, with dark-grey dolomites present as a lenticular unit at the top, or as interbeds within paler coloured dolomites (Scrubby Range, Fig. 5.1). Outcrops occur in areas which have experienced biotite facies metamorphism, and may have been affected by at least two deformations. The first produced the major fold structures, and the second a crenulation cleavage (Mancktelow, 1979) which is present in phyllites of the River Wakefield Subgroup.

FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

Pale Coloured Recrystallised Dolomites

White to buff, and occasionally orange and pinkish recrystallised dolomites, constitute a large part of the Skillogalee Dolomite (Fig. 5.1), and are characterised by the complete obliteration of microscopic sedimentary textures. In the western outcrops, this facies tends to be fairly homogeneous, but in the area south of Burra, the colour and degree of recrystallisation are more variable.

The dolomites are thin to medium bedded (Plate 5.1a), although bedding is irregular as a result of recrystallisation and deformation. Occasional gently domal structures, 20-70 cm wide (Plate 5.1b), which pass laterally into flat bedded dolomites, may represent isolated domal stromatolites. However recrystallisation has destroyed details of microstructure. Other sedimentary structures are minor, and include poorly defined intraclastic textures, and small peaked structures (? tepees). White chert nodules are present, often being concentrated along a particular horizon. They may be isolated nodules up to 25 by 35 cm, or irregular beds up to 30 cm in thickness which continue along strike for several metres. Small centimetre sized nodules concentrated along bedding planes in outcrops near Riverton, may be replaced sulphates. They resemble silicified anhydrite nodules illustrated by Seidlecka (1976, Fig. 3), and Tucker (1976, Fig. 3).

Recrystallisation has produced xenotopic fabrics (Friedman, 1965) of finely to coarsely crystalline dolomite (Folk, 1974a). Chert nodules also have recrystallised fabrics, and contain both equigranular and inequigranular mosaics of polygonal quartz. Isolated euhedral grains, or aggregates of a few grains of quartz, are also common in dolomites. Quartz constitutes the main impurity in these dolomites. Talc and authigenic feldspar are also present. Hence deposition of clay minerals within these dolomites was minor.

Rare outcrops of recrystallised limestone are present at Scrubby Range, but because of poor outcrop the relationships with the associated dolomites could not be determined. Authigenic quartz in one sample contains anhydrite inclusions, again suggesting minor formation of sulphate minerals.

The general lack of sedimentary structures apart from bedding and occasional domal stromatolites, makes detailed interpretation of the nature of the original sediment and its environment of deposition impossible.

However carbonate deposition occurred in an environment protected from the influx of terrigenous clastics. The environment appears to have been oxygenated, as indicated by the pale colour of the sediments as compared with the overlying grey dolomites. It may also have been quite shallow (possible intraclasts, tepees).

Pale Coloured Stromatolitic Dolomites

Columnar stromatolites are rare, and occur in dolomites which are slightly less recrystallised than the facies described above. Occurrences include:

(1) a 1 m thick biostrome within bedded recrystallised dolomites on the River Wakefield south of Undalya (U);

(2) a 3 m horizon of columnar stromatolites interbedded with laminated dolomite at the top of the Skillogalee Dolomite, near the River Wakefield west of Rhynie (RW);

(3) biostromes of columnar stromatolites overlying darker grey dolomites in the Duttons Trough area (DT).

The latter example has been described by Preiss (1972) as *Baicalia burra*. Minor intraclasts (less than 1 cm in size), and rare sand, were deposited in the interspace between the columns. Other occurrences of *Baicalia burra* in the Mundallio Subgroup (see Chapters 4 and 6), and the closely related *Tungussia wilkatana* (Chapter 4), grew in largely submergent environments, but were subject to occasional desiccation, and erosion during storms. These stromatolites may have grown in similar environments.

Grey Dolomite Mudstones

The uppermost part of the Skillogalee Dolomite consists of a lenticular unit of grey dolomite mudstones (dolomicrosparites and finely crystalline dolomites, Fig. 5.1). In the western area, this unit varies from 0-20 m in thickness, but is thicker in the Scrubby Range area (from less than 50 m to 240 m). One major horizon of similar grey dolomite mudstones occurs within pale recrystallised dolomites in Scrubby Range (Fig. 5.1), and may be the southerly extension of the banded dolomite "No. 1 marker bed", also known as the 1C horizon, of the Burra Mine area (Thomson, 1963). Most outcrops are flat laminated and thinly bedded (Plate 5.1c, Fig. 5.2). More domal lamination may represent stromatolites (Plate 5.1c). The lamination is due to an alternation of laminae of dolomicrospar with more impure dolomite laminae containing silt and fine-grained sand. The latter occasionally formed starved ripple lenses indicating reworking by waves or currents. Variation in the grain size of dolomite also defines lamination. Carbonaceous streaks and crinkled carbonaceous laminae may be the relics of algal mats which grew on the sediment surface. Other sedimentary structures include infrequent desiccation cracks and rare small tepees (Plate 5.2a). Within the tepees, laminae of dolomite mudstone have been slightly lithified and disrupted during growth of the tepee. Silty dolomite has infilled the disrupted centre of the tepee, sometimes from below. Hence water escaping through the centre of the tepee may have played a part in their growth (Von der Borch and Lock, 1979). Erosion of desiccated dolomite mudstones and tepees, produced lenses and interbeds of intraclastic dolomite up to 15 cm in thickness.

This facies was deposited on a low energy, subaqueous dolomite mudflat, subject to infrequent exposure. However binding of the surface sediment by algal mats may have inhibited the formation of desiccation cracks, and partly account for their infrequent occurrence. The predominance of flat lamination, with only minor domal stromatolites, and the lack of columnar stromatolites, may reflect a uniform sediment supply into a low energy environment[1]. Thin extensive laminae of dolomite mud were deposited. This allowed complete colonization of the sediment surface by algal mats (Hardie and Ginsburg, 1977),

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1. See discussion concerning domal and columnar stromatolites for Unit 1 of the Nathaltee Formation, Chapter 4.

with isolated areas of preferred growth producing small domes.

Magnesite

Magnesite facies appear to be rare in the Skillogalee Dolomite. Interbeds of intraclastic magnesite were deposited in both the Skillogalee Creek area and Scrubby Range, where they are associated with grey dolomite mudstones. The occurrence of intraclastic magnesite as interbeds within a widespread facies representing a low energy environment, may preclude long distance transport of intraclasts, suggesting a local source. However the only occurrence of magnesite mudstones observed, is a 70 cm bed of inter-laminated magnesite and dolomite (Plate 5.2b), 3 m below the top of the formation on the River Wakefield west of Rhynie.

Terrigenous Clastics

Sandstones and shales are rarely exposed in outcrops of the Skillogalee Dolomite. Rare interbeds of very fine- to fine-grained, dolomite-cemented sandstone, up to 1 m in thickness, are interbedded within grey dolomite mudstones at Scrubby Range. They contain flat to wavy, and ripple lamination (Plate 5.2c). Very rare coarse-grained sandstones are present in the River Wakefield area. Hence the influx of sand into the area of deposition of the Skillogalee Dolomite was minor, and that which is present may have had a local source. The sand was deposited in a submergent environment, with moderately agitated conditions.

Shale interbeds appear to be minor, even where outcrop is relatively continuous. However the shales present are generally weathered. Hence little fine terrigenous detritus reached this area.

ENVIRONMENTAL SUMMARY

The Skillogalee Dolomite represents deposition in an area where the influx of terrigenous detritus was limited. The grey dolomite mudstones were deposited in a low energy environment, and any terrigenous mud entering this area would have been deposited concurrently. The pale recrystallised dolomites which comprise much of the formation, may represent a shallow water environment in which conditions were oxidizing. The more recrystallised nature of the pale dolomites, as compared with the overlying grey dolomite mudstones, must reflect primary differences in the original sediments, as both have experienced the same tectonic history. Quartz nodules within the pale dolomites are also more recrystallised than those within grey dolomite mudstones. Impurities within the latter (terrigenous silt and clay, and carbonaceous material), may have inhibited their recrystallisation. However contrasting diagenetic histories, with greater early diagenetic alteration and recrystallisation in the pale dolomites, may have induced differing behaviour during metamorphism.

Grey dolomite mudstones also represent a shallow water environment, probably as extensive subaqueous mudflats, with intervening exposed, desiccated areas. Hence shallow dolomite mudflats were widespread in the northern Mt. Lofty Ranges. A shoreline position, or a shoreward direction, cannot be determined, although sandstone interbeds in the Scrubby Range area may have been introduced from local sources to the east, as transportation across the mudflat from a westerly source is unlikely.

CASTAMBUL FORMATION

The Castambul Formation comprises a sequence of massive buff to pink dolomites, grey phyllites and siltstones, and minor sandstones (Fig. 5.3). The best exposures are found in the Torrens Gorge area (TG, Fig. 3.2), where Mancktelow (1979) has presented a geological map and cross section. This area is complicated by folding and faulting, hence the thickness shown in Figure 5.3, may not be accurate. The Castambul Formation overlies the Aldgate Sandstone, and was deposited in areas where the Aldgate Sandstone is thinnest, immediately south and west of the Houghton Inlier. To the east and further south, the Aldgate Sandstone is much thicker, and is overlain by the Woolshed Flat Shale (Fig. 3.3).

FACIES DESCRIPTION AND DEPOSITIONAL ENVIRONMENTS

Pale Recrystallized Dolomites

Massively outcropping, buff to pink coloured, fine- to medium-crystalline dolomites (Plate 5.3a), dominate the lower part of this formation (Fig. 5.3). They are thick bedded, and occasionally have a vague horizontal to wavy lamination, possibly of algal origin. The homogeneous appearance, lack of sedimentary structures, and the destruction of microscopic textures due to recrystallisation, makes environmental interpretation of this facies difficult. However the low impurity content indicates an environment free from detrital influx.

Phyllites and Siltstones

Weathered grey, occasionally dolomitic phyllites, siltstones and shales, comprise much of the upper part of the Castambul Formation, and may also occur as interbeds within dolomites (Fig. 5.3). The degree of development of phyllitic textures depends on the silt-clay ratio in the initial sediment. Sedimentary structures include flat lamination, occasional cross lamination and small cross-beds (5-10 cm), and graded lamination. The dominating features are however schistosity and crenulation, which developed during the Delamerian Orogeny.

This facies was deposited in a low energy submergent environment, in which intermittent wave or current activity was sufficient to produce a well laminated sediment with occasional cross laminae, but was insufficient to winnow mud. The thick interval of shales in the upper part of the Castambul Formation, records a period when deposition and subsidence rates were approximately equal, and during which there was a persistent influx of terrigenous detritus.

Sandstones

Two major sandstone units occur within the Castambul Formation (Fig. 5.3). The sandstones are medium- to coarse-grained, quartz-cemented subarkoses. The lower sandstone is generally weathered, and dominantly flat-bedded, and may have been deposited in the wave-agitated zone as a sand sheet prograded across areas in which dolomite had previously been deposited. The upper sandstone, at the top of the Castambul Formation, has a slightly gradational boundary with the underlying siltstones. It contains alternating, 1 m thick cosets of trough cross-bedded and horizontal to low angle bedded sandstones (Plate 5.3b). Trough cross-beds are generally 20-30 cm, and rarely 80 cm in thickness. This unit may represent a shallow coastal sand shoal, formed by migrating dunes (trough-cross beds) on the upper shoreface, while the flat and low angle cross-bedding may represent deposition under a lower flow regime, or deposition on the foreshore and backshore (Harms *et al.*, 1975). Orientation of low angle tabular sets are varied, but south-southeastly to east-southeasterly flowing currents appear to have been dominant.

Hence the sequence at the top of the Castambul Formation is a shallowing upward sequence from shales to sandstones, which is followed by the mixed dolomite-sandstone-magnesite deposition of the Montacute Dolomite. Hence it is analogous to the upper part of the Nathaltee and Nankabunyana Formations (Chapter 4), both of which are overlain by the dolomite-sandstone-magnesite sequence of the Yadlamalka Formation.

ENVIRONMENTAL SUMMARY

Deposition of the lower dolomitic part of this formation may have occurred in an environment similar to the major part of the Skillogalee Dolomite. However the area around Torrens Gorge deepened, either due to transgression or increased rates of subsidence, and offshore terrigenous mud deposition replaced that of dolomite. Introduction of sand and shallowing of the environment, resulted in deposition of a coastal sand shoal at the top of the formation.

WOODSHED FLAT SHALE

The distribution and stratigraphic relationships of the Woolshed Flat Shale have been discussed in Chapter 3 (Figs. 3.2, 3.3). It is an internally homogeneous formation, dominated by fine-grained terrigenous clastics, with minor sandstone and dolomite interbeds. It has a transitional base with the Montacute Dolomite, but may overly the Skillogalee Dolomite with a sharp (River Wakefield, west of Rhynie, RW), or transitional boundary (Skillogalee Creek area, SC).

FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

Siltstones and Shales

This facies consists of alternating laminae of clay and silt to very fine-grained sand (Plate 5.3c). However in the southern areas near Adelaide, alignment of micas to form a schistosity, and the subsequent crenulation, may be the dominant fabrics. Flat, even lamination is dominant, but silt and very fine-grained sand laminae are sometimes lenticular, occur as isolated ripple lenses, or are cross-laminated. Rare low angle erosional truncations are up to 1 cm deep. Boundaries between clay and silt laminae are generally sharp, graded lamination is less common.

This facies represents a similar environment as shales within the Castambul Formation, that is a low energy, submergent environment, with weak current activity. Clay laminae were deposited from suspension, whereas the more lenticular silt laminae indicate weak traction currents or wave action, possibly during storms. The rarity of erosional features suggests that winnowing was virtually inactive.

Grey Dolomite Mudstones

This facies is most abundant where the Woolshed Flat Shale overlies the Montacute Dolomite. Dolomite mudstone interbeds (generally less than 1 m, and rarely 7 m in thickness) are flat laminated with silty lenses, or internally massive. They may be largely detrital in origin, derived from erosion in adjacent shallower areas of dolomite mud production, and were deposited when the influx of terrigenous mud was reduced.

Sandstones

Interbeds of grey, fine-grained, to occasionally coarse-grained sandstone occur in many areas. Wavy, and occasional cross-lamination, is present. They record periods when winnowing was more effective, or transport of sand by storms into more offshore, quiet water areas.

ENVIRONMENTAL SUMMARY

The Woolshed Flat Shale represents an offshore facies of fine-grained detritus, which had bypassed shallower near shore areas. However it was subject to weak, periodic, current or wave activity, and hence probably accumulated in the zone immediately below wave base. The environments represented by this formation, and the other formations in the Mt. Lofty Ranges, are incorporated in a regional synthesis in Chapter 8.

CHAPTER 6

THE DARK-GREY DOLOMITE-MAGNESITE-SANDSTONE (YADLAMALKA FORMATION, MONTACUTE DOLOMITE) AND DARK-GREY DOLOMITE-SANDSTONE (MIRRA FORMATION) FACIES ASSOCIATIONS OF THE UPPER MUNDALLIO SUBGROUP: FACIES DESCRIPTIONS AND DEPOSITIONAL ENVIRONMENTS

INTRODUCTION

The most widespread facies association of the Mundallio Subgroup, is the dark-grey dolomite mudstone, stromatolitic dolomite, intraclastic magnesite and dolomitic sandstone association of the Yadlamalka Formation, which occurs throughout the southern and northern Flinders Ranges (Fig. 3.2). Related to this is the dolomitic sandstone, dolomitic siltstone, quartz-cemented sandstone and dark-grey dolomite mudstone association of the Mirra Formation, which occurs in the central and eastern Willouran Ranges (Fig. 3.2). The Montacute Dolomite represents a local development, within the Adelaide area, of facies similar to the Yadlamalka Formation, from which it is separated by dolomite and shale facies of the Skillogalee Dolomite and Woolshed Flat Shale (Fig. 3.1).

These formations cannot be further subdivided into units. They generally form a monotonous sequence in which all major facies occur throughout, and which are characterized by rapid vertical facies changes. However the proportions of individual facies varies vertically through the sequence in any given area (Figs. 6.1-7). Occasional shale-sandstone couplets (8-14 m in thickness), and shale beds (up to 12 m in thickness) form the only marker horizons (Figs. 6.1 and 2).

BOUNDARIES

The lower boundaries of the Yadlamalka and Mirra Formations are generally sharp where they overly the clastic dominated sequences of the lower Mundallio Subgroup. Sandstones at the top of Unit 3 of the Nathaltee Formation, Unit 4 of the Nankabunyana Formation, and the Tilterana Sandstone, often form ridges overlain by the less boldly outcropping dolomitic sequences. Likewise the Montacute Dolomite has a sharp boundary with sandstones at the top of the Castambul Formation.

In the central part of the southern Flinders Ranges, deposition of the Yadlamalka Formation commenced at the base of the Mundallio Subgroup, and here it overlies the Bungaree and Yednalue Quartzites. These two formations consist of interbedded quartz-cemented sandstones, grey siltstones and poorly laminated shales, and white to grey coloured dolomites, which are generally more recrystallised than those of the Yadlamalka Formation. The boundary with the Yadlamalka Formation is transitional, with dolomites in the lowermost part (up to 30 m) varying in colour from dark-grey to pale-grey to buff. Both dolomite- and quartz-cemented sandstones occur in the transitional zone.

The upper boundaries of the Yadlamalka and Mirra Formations, and the Montacute Dolomite, are transitional, and not characterised by the presence of marker beds. In the southern Flinders Ranges, deposition of the Yadlamalka Formation was halted by the introduction of sand on the western margin of the basin, possibly as a result of renewed tectonic uplift in the source areas. This sandstone unit (the Undalya Quartzite), thins to the east and is replaced laterally by finer grained clastics (Fig. 3.4), indicating that there was probably a deepening of the basin in this area. In the northern Flinders Ranges, the Yadlamalka and Mirra Formations are overlain by the Myrtle Springs Formation, a sequence of sandstones, siltstones and dolomites, which is somewhat similar to the Nankabunyana Formation. This boundary also marks an increase in the influx of terrigenous detritus to the basin.

DOLOMITE FACIES

DOLOMITE MUDSTONES

This facies includes all the mud supported, dolomite sediments in the Yadlamalka and Mirra Formations, and the Montacute Dolomite. Dolomite mudstones are the most abundant facies in the Yadlamalka Formation and Montacute Dolomite, forming 50-80% of the outcrop in all areas, with the exception of the Copley-Myrtle Springs (CP-MS) and Yacka (YE) areas (Figs.

6.1-7). Within the Mirra Formation, this facies is less abundant (13-42%), because of the greater contribution of fine-grained sandstones and siltstones.

Most dolomite mudstones contain less than 25% terrigenous sediment (clay, silt and sand), and some are near pure dolomite. However this facies also includes very impure mudstones, in which clay and silt are as abundant as dolomite. Weathering of impure dolomite mudstones is often greater than those with only minor clay content, and tends to exaggerate their impure nature.

In many areas, dolomite mudstones are exclusively medium- to dark-grey in colour, due to disseminated carbonaceous material, although the carbon content is only a fraction of a percent[1]. Light-grey dolomite mudstones are common at Mundallio Creek (MC), and Top Mount Bore (TM). Yellowish-grey, greyish-orange and light brown dolomite mudstones are interbedded with dark-grey dolomite mudstones at Yednalue (YD), Willow Creek (WI), Johnburg (J), and in the Yednalue Anticline (YDA). The pale coloured dolomites are often very impure.

Dolomite mudstones form horizons up to 25 m in thickness although they are generally much thinner, and are interbedded with all other facies. Boundaries are sharp, and less commonly gradational.

Sedimentary Structures and their Origin

Dolomite mudstones are characteristically laminated and thinly bedded, and outcrops may be flaggy (Plate 6.1a), or more massive. The lamination and thin bedding within the flaggy outcrops is flat, to slightly wavy and lenticular, with associated small erosional scours, and minor cross-laminae. The flaggy nature is probably due to weathering along clay rich seams.

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1. 21 analyses for carbon on insoluble residues, made with a Leco Carbon Determinator, gave an average content of 0.1% (range .4-.008%) in the total rock.

Within massively outcropping dolomite mudstones, the lamination is less conspicuous because of little textural contrast between laminae, and low terrigenous content. The lamination is wavy (Plate 6.1b), and often irregular due to compaction around chert nodules.

The mudstones are largely dolomicrosparites (Folk, 1974a), although coarser textures have developed in areas of higher metamorphic grade (Torrens Gorge, Arkaroola). The microspar may have variable grain size between adjacent sharply bounded laminae, or form graded laminae, probably reflecting primary grain size differences in the carbonate mud. Planar and lenticular laminae of terrigenous detritus are common, and silt or sand may also occur at the base of graded laminae (Plate 6.2a). Some mudstones have a clotted or grumous fabric (Plate 6.2b; Bathurst, 1974). This fabric may grade into more distinctly peloidal fabrics. The latter have not been distinguished in stratigraphic sections from mudstones, but their characteristics are described separately below. Clotted fabrics may have originated from muddy peloidal sediments, in which compaction and diagenesis of largely soft peloids and mud has obscured the fabric. Wispy carbonaceous laminae, which may be the relics of sediment poor algal mats, occur infrequently.

The flat even, to slightly wavy lamination, formed by detrital deposition of dolomite mud and minor terrigeneous detritus. Weak current activity transported silt and sand as bedload, and finer material in suspension. Hence the lamination reflects the periodic influx of sediment. Although algal mats may have colonised the muddy sediment surface during breaks in deposition, their role in trapping sediment was less important than deposition of sediment by physical processes. However in mudstones with a more irregular, wavy lamination, algal trapping may have played a greater role in producing the laminated fabric of the sediment (Zamarreño, 1975).

The lamination within dolomite mudstones was sometimes disrupted by soft sediment deformation structures. Slump folds (Plate 6.1c) affected a 1-20 cm thickness of sediment. Semi-brittle fracturing of the sediment was more common, and resulted in the formation of discrete blocks with their internal lamination in variable orientations (Plate 6.2c). This probably resulted when slump folds fractured along their axes, with the limbs subsequently becoming completely separated. Disrupted lamination was sometimes erosionally overlain by sand or dolomite mud, and hence formed in the uppermost parts of the sediment. Truncated slump folds were also observed by Forbes (1960).

These structures reflect instability in near surface, unlithified, but cohesive muds. Varying degrees of lithification, and differences in water content, may have produced density differences within texturally fairly homogeneous sediment. Small earthquake shocks may have initiated failure of the sediment. Small slump structures, and disrupted lamination, are also characteristic of terrigenous mudstones in the lower Mundallio Subgroup.

Desiccation cracks (Plate 6.3a) are present in dark-grey dolomite mudstones in all areas, but are not abundant as has been previously noted by Forbes (1961). They are rare to absent in the Mirra Formation. However they are abundant in yellowish-grey and light-brown mudstones (Plate 6.3b) in the Yednalue and Willow Creek areas. Desiccation cracks are up to 20 cm in size and may be infilled with sand, silt or dolomite mud. Tepees are uncommon in the Yadlamalka Formation and Montacute Dolomite, and absent from the Mirra Formation. The tepees present are small (Plate 6.3c), with a height of 2-12 cm (rarely larger), and a separation of up to 50 cm. Hence they fall in the embryonic class of Assereto and Kendall (1977). Bedding plane exposures illustrate the characteristic polygonal form of tepees.

Tepees form in marginal carbonate environments, in the supratidal zone, or other areas which experience extended periods of exposure (Davies, 1970; Assereto and Kendall, 1971, 1977; Burri *et al.*, 1973; Evamy, 1973; von der Borch and Lock, 1979). Crystal growth within the sediment leading to expansion, may be the major process in tepee formation (Assereto and Kendall, 1977). Von der Borch and Lock (1979) propose that tepee formation in ephemeral dolomite lakes associated with the Coorong, results from compressional stress produced as upwelling groundwater causes extrusion of mud at polygon boundaries. The disrupted sediment in the centre of tepees, the association with desiccated beds, and the small size (Plate 6.3c), suggests that this process was important in producing tepees in this facies. Hence dolomite mudstones with tepees may have experienced annual, or longer term wetting-drying cycles, in which groundwater recharge was important.

Very small tepees, affecting only one or two thin dolomite mudstone beds, developed when an area was subject to the above conditions for only a short period of time (e.g. several wetting-drying cycles), after which sedimentation of dolomite mudstones was renewed. Larger tepees formed at the top of a dolomite mudstone unit, and were subsequently eroded and overlain by transgressive, higher energy deposits, including sandstones and intraclastic beds.

Diagenetic Quartz and Chert nodules

Chert nodules are ubiquitous in most grey dolomite mudstones, although they represent only a small volume of the total outcrop. The nodules, which vary in colour from white to black depending on the carbonaceous content of the host sediment, are present as thin lenses in the more flaggy outcrops (Plate 6.1a), and as more irregular and often larger nodules in the more massively outcropping dolomite mudstones. The texture and formation of this chert, which preserves the lamination and other structures of the host sediment, is described in Chapter 11.

However rare outcrops of dolomite mudstones at Mundallio Creek (MC), Depot Creek (DC), Yednalue Anticline (YDA), Yatina (YT), Port Germein Gorge (PG), Top Mount Bore (TM), and Warra Warra Mine (WW), contain nodules of white, more coarsely crystalline quartz, the texture and morphology of which is similar to that in nodules described elsewhere as being replacements of evaporites. These nodules are present in a variety of forms, including horizons of centimetre sized nodules arranged in aggregates parallel to bedding (Plate 6.4a), and which may continue along strike for several 100's of metres. Some nodules have a more random distribution. Other forms present include rosettes and lath shaped pseudomorphs (Plate 6.4b and c), which may be gypsum replacements, whereas nodular forms may be replaced anhydrite nodules.

The nodules and pseudomorphs consist of quartz and dolomite spar. The quartz consists of coarse-grained mosaics of often cloudy crystals, with undulose and flamboyant extinction, and highly sutured boundaries. These may grade into aggregates of radiating elongate crystals, which are also undulose. Mosaic chert and length-slow chalcedony are minor components. Small lath shaped pseudomorphs, associated with nodules, are sometimes observable in thin section. Minute rectangular inclusions within some quartz crystals resemble anhydrite, but are too small to identify. Hence direct evidence for the replacement of evaporites is lacking, but nodules containing quartz of similar texture and morphology, and with associated relic sulphates, have been described in the literature as forming by the replacement of sulphates (Siedlecka, 1972, 1976; Chowns and Elkins, 1974; Tucker, 1976; Walker *et al.*, 1977; Milliken, 1979; Young, 1979).

If these nodules are replacements of evaporites which grew within the sediment, they are however rare, with the Mundallio Creek area being the only location where more than two or three occurrences were observed. Sulphate minerals may form interstitially within the sediment in areas where saturated brines are produced, commonly by evaporation in an arid climate.

Precipitation occurs in intertidal to supratidal zones, or in playa lakes (James, 1979; Kendall, 1979). However there must be sufficient influx of brines to maintain the water level near the sediment surface, and a supply of sulphate ions within the brines. The presence of nodular sulphates indicates a very shallow probably exposed environment.

Dolomite Mudstones : Summary

This facies contains evidence of exposure in the form of desiccation cracks and tepees. However these structures are not abundant, and when combined with the lack of fenestral features, may indicate a predominantly submergent environment of deposition (Gill, 1977). However smoothly laminated carbonate sediments which are exposed for more than 80% of the time on the carbonate tidal flats of Andros Island, lack desiccation cracks, apparently due to the pervasive binding of algal mats (Hardie and Ginsburg, 1977). However, horizontal sheet cracks and small fenestral pores provide some evidence of desiccation. Hence the degree of exposure may be difficult to assess when features indicative of exposure are lacking, and it is likely that the sediment surface was colonised by algal mats.

Dolomite mud, along with minor terrigenous silt and sand, were deposited as thin detrital laminae by physical processes, probably including storm settle-out, in largely submergent environments. Algal trapping may have produced more wavy laminated mudstones. Mudstones with abundant desiccation features, consist of alternating detrital laminae (sand, silt, dolomite intraclasts, dolomite mud), and desiccated dolomite mud laminae. These were deposited when exposed mudflats were inundated by broad sheets of water, either storm induced from adjacent submerged areas, or possibly from landward derived sheet floods (Hardie *et al.*, 1978) which supplied the terrigenous detritus. Initial deposits of coarser detritus were followed by settling out of dolomite mud as currents waned, or from ponded water on the mudflats (Hardie *et al.*, 1978).

Dolomite mudstones generally experienced reducing conditions following deposition. In the abundantly desiccated mudstones formed on the exposed mudflats, more oxidising conditions prevailed.

DOLOMITE INTRACLASTIC GRAINSTONES

This is an uncommon facies, either as thin lenses of intraclasts associated with desiccated beds, or as thicker beds (rarely more than 0.5 m in thickness), reflecting the infrequent periods of desiccation of dolomite mudstones. Intraclastic grainstones occur as lenses and beds of rounded sand to granule sized intraclasts, often with a sandy matrix (Plate 6.5a), or as lenses and beds of coarser, elongate and more angular intraclasts, interbedded with dolomite mudstones and sandstones. In the northern Flinders Ranges, the former variety predominates, and is frequently interbedded with very fine-grained sandstones. Beds of coarser intraclasts invariably contain associated sand, and grade into sandstones with scattered intraclasts (Plate 6.5b).

Coarser intraclastic grainstones have experienced little transportation, and formed from the erosion of desiccated dolomite mudstones during high energy events. However beds of rounded, sand to granule sized intraclasts have experienced greater abrasion and reworking, possibly in the zone of wave agitation, where they were subsequently deposited with sand.

PELOIDAL GRAINSTONES AND PACKSTONES

Peloidal dolomites are similar in outcrop appearance to dolomite mudstones, particularly where the peloids are of silt size. However this facies does not appear to be common, except within the central and eastern Willouran Ranges. However many fine-grained sandstones within the northern Flinders Ranges contain dolomite peloids (see descriptions of sandstone facies below). The peloids are elliptical grains of micrite to microspar, less than 0.3 mm in size, grading into more irregular grains up to 1 mm in size, which may be small intraclasts, and they are enclosed in clear microspar and fine spar.

Peloidal carbonate sediments probably have a polygenetic origin (Beales, 1965), although two common origins, faecal pellets and micritization of skeletal fragments by boring algae, cannot be applied to this sequence because of its Precambrian age. Other possibilities include derivation from erosion of pre-existing mudstones (Eugster and Hardie, 1975), inorganic precipitation in an agitated environment (Deelman, 1978a), and organically induced precipitation (Davies, 1970). Peloids produced by these mechanisms cannot be distinguished on textural grounds, although the dark carbonaceous staining in some peloids may suggest some organic influence. However the resultant peloidal sediment was mechanically deposited in a submergent environment, under moderate energy conditions.

OOID GRAINSTONES

Ooid grainstones are rare, and were observed only at Depot Creek (two beds), Top Mount Bore (two beds) and east of the Paralana Fault at Arkaroola. These beds are massive, light-grey oosparites, with ooids 0.5 to 2.0 mm in size. The nucleus may be a dolomite or magnesite intraclast, or a sand grain, but is commonly recrystallised dolomite. The cortex, which may also be partly recrystallised, contains concentric laminae of even thickness. This contrasts with oncoids in which laminae show significant thickness variation and are often discontinuous (see below).

The important conditions in natural ooid forming environments which appear to be important to ooid growth, have been listed by Bathurst (1974, p. 301):

(1) supersaturation of the $CaCO_3$ solution;

(2) available detrital nuclei;

(3) agitation of the grains;

(4) a hydrologic system that retains the grains within the ooid forming environment.

The depositional environment of this sequence appears to have been one in which carbonate formation was favoured, as in most areas carbonate facies predominate. However high energy environments were generally associated with the introduction of terrigenous sediment or the erosion of pre-existing carbonate sediments. Shallow, persistantly agitated environments with minimal introduction of sediment, as in modern ooid shoals, were apparently absent, or very rare.

ONCOID GRAINSTONES

Oncoid grainstones are only a minor component of the Yadlamalka Formation, being most abundant at Depot Creek where this facies forms about 1% of the formation. They were not observed in either the Mirra Formation or Montacute Dolomite (Figs. 6.3-7). Many occurrences of this facies have been replaced by black chert (Plate 6.5c), and the preservation of oncoids is much better in silicified outcrops than dolomitic. Oncoid grainstones form plane bounded thin to thick beds, or rarely laminae, which are internally massive, or rarely cross-bedded. They are interbedded with dolomite mudstones, sometimes with an erosional base, and may also be associated with stromatolitic dolomites and intraclastic grainstones. Some beds may be followed for up to 100 m along strike before outcrop ends, but a 0.5 m thick silicified bed is continuous for at least 12 km at Copley.

The subsequent descriptions are based largely on silicified samples, for within dolomitic samples, compaction and recrystallisation has generally destroyed oncoid form.

Regular Coated Oncoids-I

These oncoids are predominantly elongate grains in cross-section, and hence are probably disc-shaped in 3-dimensions, and less commonly are more circular. They have smooth, lenticular, concentric laminae, which show marked thickening on the ends of grains (Plates 6.6a, 11.5b and c).

The laminae consist of chert or fine-grained quartz, which contain disseminated carbonaceous material. Laminae of clear, fine-grained quartz are less common. In well preserved dolomitic oncoids, the laminae consist of microspar, however most dolomitic oncoids have a distorted shape, and the lamination is indistinct. Dolomite and magnesite intraclasts, and sand grains, form nuclei in oncoids.

Within a given sample these oncoids fall in a small size range, from 0.2 mm to a maximum of 1-2 mm. Compound oncoids form a minor component in some samples, but their maximum size is no larger than the largest single oncoid. In all areas of occurrence, both the appearance and size of these oncoids is remarkably similar.

Regular Coated Oncoids-II

Coated oncoids with a vaguely radial pattern and laminae of more even thickness (Plate 6.6b), are much less common than the coated oncoids described above. Silicified oncoids consist of chert in which the lamination is defined by carbonaceous material (Plate 6.6b), whereas dolomitic oncoids contain alternating laminae of microspar and clear fine spar. The laminae in unsilicified oncoids are much more irregular than the smooth lamination in silicified oncoids, suggesting that prior to diagenesis, dolomitic oncoids now characterised by an irregular lamination, may have been more regular.

Massive Oncoids

Massive oncoids generally are found in association with coated oncoids, and rarely form the major type present. They are approximately equidimensional, and less commonly elongate, irregular grains, often with concave margins (Plates 6.6c, 11.6c). They are larger than coated grains, with a maximum size of 1 cm. Coated oncoids may be incorporated within more massive oncoids. Carbonaceous impurities within massive oncoids have been partly redistributed by the silicification. The shape of massive oncoids has generally been completely destroyed in unsilicified oncoid grainstones within which they are often unrecognizable.

Cements

Silicified samples contain various silica cements, which formed during silicification, and not as a replacement of pre-existing carbonate cements. In partially or unsilicified samples, pore space has generally been reduced by compaction, but that which remains is infilled with clear dolomite spar. Interstitial mud is lacking, hence most oncoid grainstones appear to have been well winnowed during deposition.

Discussion : Origin and Depositional Environments

The use of the term oncoid implies a specific origin, that is, the grains were formed by algal activity. The presence of carbonaceous material, lenticular lamination, and the destruction of oncoid shape during diagenesis unless silicification had occurred, suggest that oncoids were not rigid structures formed by precipitation as in ooids, but consisted of a mesh or aggregate of carbonate mud and algal material. The mud was bound by algae which maintained the shape of the oncoid until the algal material was destroyed during diagenesis. The presence of carbonaceous material in massive oncoids, and their similar destruction during diagenesis, suggest that they may also have contained a significant proportion of algal material.

Oncoids are generally considered to form in the lowest intertidal to subtidal environments (Logan *et al.*, 1964; Aitken, 1967; Peryt, 1977). However the lack of lithification of oncoids to form semi-rigid structures prior to diagenesis, and the lack of carbonate cements suggest that oncoids were not subject to intertidal and supratidal vadose conditions, which would probably have resulted in early lithification and cementation. Hence the oncoids grew and were deposited in largely submergent environments.

Concentrically laminated, slightly elliptical to circular oncoids, may have formed in an agitated environment (Kauffman, 1977), in which the oncoids were placed in suspension and redeposited in various orientations, allowing circular structures to form. Lower energy conditions resulted in more elongate oncoids, as the grains less frequently changed their orientation. Massive oncoids may have grown in environments of only mild agitation, permitting continuous algal growth (Dabanyake, 1977).

Coated oncoids are significantly smaller than most examples discussed in the literature (Kutek and Radwanski, 1965; Aitken, 1967; Dabanyake, 1977; Kauffmann, 1977). This may simply be a reflection of the energy levels of the environment, although the similarity in size and morphology between widely separated areas, may suggest some biological control.

STROMATOLITIC DOLOMITES

Stromatolitic dolomites are present in both the Yadlamalka and Mirra Formations,but generally form less than 1% of the sequence, except at Depot Creek (9%), Yednalue (2%), and in the southwestern half of the Willouran Ranges (1.5%). However stromatolites commonly form extensive, massively outcropping biostromes of dark-grey dolomite with black chert nodules, and hence form conspicuous outcrops.

Columnar Stromatolites

Columnar stromatolites are the most common stromatolite type in all areas, with the exception of the Depot Creek area. The columns formed biostromes (most frequent) and bioherms, and rarely grew as isolated columns, or formed very small groups (Plate 6.7a). The morphology of columnar stromatolites has been described in detail by Preiss (1972), and only a single form, *Baicalia burra*, is present. It contains tuberous and irregular columns, with vertical and inclined growth, and slightly to markedly divergent branching (Plate 6.7b and c).

Biostromes, 0.5-2 m in thickness, are tabular, or have undulating surfaces with relief up to 0.5 m. At Copley, one biostrome is continuous along strike for at least 12 km (Fig. 4.8) but outcrop is generally inadequate to follow single biostromes for anywhere near this distance. Bioherms are more common than indicated by Preiss (1972), and grew in most areas. They are domal to tabular in shape, and generally less than 0.5 m in thickness. Larger structures, up to 5 m by 5 m occur near Rischbieth Hut (R) in the Willouran Ranges. Small bioherms may be enclosed in peloidal, oncoid and intraclastic grainstones, and are less frequently overlain by dolomite mudstones. In contrast biostromes are almost invariably overlain by dolomite mudstones. Bioherms occur in isolation, but often several are present at a single stratigraphic level (Plate 6.8b). An extreme example of this is present at Copley, where an horizon of tabular bioherms, each up to 1.5 by 15 m in size, and separated and overlain by shale, extends from Copley to Myrtle Springs, a distance of 30 km (Fig. 4.8). The stromatolites within biostromes and bioherms often colonised the eroded surface of dolomite mudstones (Plates 6.7a, 6.8a), and erosional surfaces may also occur within biostromes. The stromatolites also grew on intraclastic magnesite and dolomitic sandstone.

Interspace sediment between columnar stromatolites ranges from mudstone (most frequent) to grainstone. Allochems include elongate and irregular intraclasts derived from erosion of stromatolite columns, peloids and oncoids. Terrigenous detritus is rare in those areas where sand forms only a small proportion of the associated sequence, e.g. at Yednalue, but is more common in those areas with a higher sand content, e.g. Copley, Arkaroola and the Willouran Ranges (Preiss, 1973). Where sand is present in the interspace sediment, it is more abundant, and coarser grained than that within the adjacent stromatolite columns. Deposition of interspace sediment often records higher energy events, sometimes introducing sand, than those which normally prevailed during stromatolite growth. Columnar stromatolites were probably capable of trapping sand size sediment, as indicated by peloids and oncoids within columns. The formation of interspace sediment may thus record a destructional event in the history of stromatolite growth, causing their erosion, and the deposition of intraclasts, sand and mud between columns. Such events may have been relatively infrequent, interupting the normal low energy conditions of stromatolite growth, but left a significant record because of the high energy levels. Deposition of mud laminae between columns (Plate 6.8a) records a period in which there was an abundant supply of mud. Each laminae may record a storm, following which mud was deposited from suspension as currents waned.

Small Domal Stromatolites

Small domal stromatolites up to 30 cm in width and 30 cm in height, are most abundant at Depot Creek (Fig. 6.4), and minor elsewhere. Domes may be isolated structures, or form horizons of low relief, regular domes (Plate 6.8c), or more irregular, undulating, linked domes. Gently undulating domes may pass laterally along strike into flat laminated dolomite mudstones. Erosion of linked domes was uncommon, but isolated domes were sometimes eroded, and are associated with intraclastic dolomite. Horizons of domal stromatolites are generally associated with dolomite mudstone, although hollows between domes are sometimes infilled with coarser sediment.

Large Domal Stromatolites

This stromatolite type is rare, and is present as one, or sometimes two horizons near the top of the Yadlamalka Formation in the southwestern Willouran Ranges, and as isolated occurrences at Copley. They are similar to large domal stromatolites described by Fairchild (1975) from the equivalent sequence in the Peake and Denison Ranges. The domes range in size from 0.6-4 m in width, to 0.4-4 m in height. They have grown as unlinked structures on beds of intraclastic magnesite or dolomite mudstone (Plate 6.9a). Laminae are continuous across domes, and may be overturned on the margins, indicating growth by algal trapping, and/or *in situ* precipitation. The lamination has a crinked appearance, and there is no evidence of erosion. Outcrop is generally inadequate to assess the continuity of the horizons of these stromatolites, with the exception of one horizon of 4 m high domes at South Hill (SH), which is continuous along strike for at least 1 km.

Stromatolite Microstructure

Columnar and domal stromatolites are finely laminated, although this if often indistinct on outcrop surfaces (Plate 6.7a and c) because of little textural contrast between laminae, particularly where there has been minor recrystallisation. The lamination results largely from an alternation of darker, finer grained laminae, and lighter, coarser grained laminae (Preiss, 1972). Although some darker laminae have an internally homogeneous texture, other laminae have a clotted or grumous fabric (Plate 6.9b and c). Variable proportions of micrite and spar within adjacent clotted laminae, can also produce an alternation of darker and lighter laminae. However due to recrystallisation, the textures in some stromatolites are now indistinct.

Laminae of homogeneous micrite or microspar are often the predominant laminae type in both columnar and domal stromatolites. These laminae are continuous across columns, except where eroded, although very thin laminae (fraction of a millimeter), may be streaky. Individual laminae are occasionally crinkled, and rarely desiccated. Within a single stromatolite, the grain size of microspar may vary between adjacent laminae, possibly reflecting primary grain size differences in the sediment. The orientation of these laminae in horizontal to steeply inclined positions, indicates that they formed by algal trapping and agglutination, and/or *in situ* precipitation.

Clotted laminae are present in both columnar and domal stromatolites. They consist of irregular, overlapping clots of carbonaceous stained micrite, enclosed in clear microspar and fine spar (Plate 6.9b and c). The latter may have formed in part by recrystallisation. Distinct detrital peloids, and rarely oncoids and terrigenous silt and fine-grained sand, are present in some clotted laminae.

Clotted fabrics are common in modern stromatolites and algal laminated sediments (Monty, 1967, 1976; Davies, 1970, Friedman *et al.*,1973; Monty and Hardie, 1976; Hardie and Ginsburg, 1977). The fabrics result from a variety of processes including:

- deposition of peloids on algal mats, with subsequent dehydration of a mixture of soft and firm peloids, causing them to merge and produce a clotted fabric (Hardie and Ginsburg, 1977);
- trapping of a mixture of peloids and mud by an algal filament complex (Hardie and Ginsburg, 1977);
- precipitation of aragonite peloids within an algal mat due to CO_2 removal (Friedman *et al.*, 1973);
- precipitation of fine-grained carbonate around an algal fabric, and subsequent collapse of the organic framework in deeper parts of the mat (Monty, 1976).

Although some laminae contain detrital peloids, many have a framework of micritic clots, which may have formed by *in situ* precipitation around algal filaments, analagous to that described by Friedman *et al.* (1973). The dark carbonaceous staining of the clots in well preserved samples (Plate 6.9b) indicates that the clots consisted of algal filaments and dolomite micrite. Porosity within the mat was later infilled by inorganic precipitation of clear microspar.

In areas where interbedded sandstones are significant, laminae of silt and very fine-grained sand were deposited on the growing stromatolite columns. Silt and sand grains, oncoids, peloids and small intraclasts may be scattered within laminae of dolomite mud. Fine, crinkled carbonaceous laminae, commonly anastomosing, are rarely preserved in dolomitic stromatolites, but are more common when silicification has occurred. They represent the remains of sediment-poor algal mats, which have collapsed and been compacted.

Microunconformities, as low angle truncations across laminae, formed in columnar stromatolites in all areas, but most commonly in the Willouran Ranges (Preiss, 1972). They probably formed as a result of erosion of poorly indurated sediment. Evidence of desiccation in the form of vertical or horizontal desiccation cracks, is rare to absent in stromatolites. However the presence of elongate intraclasts in the interspace sediment between columnar stromatolites, derived from desiccation and subsequent erosion of algal mats, indicates that they were subject to periods of exposure.

Stromatolitic Dolomites : Discussion

The occurrence of columnar stromatolites in biostromes indicates uniform substrate and depositional environments over extensive areas. Biostromes grew in low energy environments, possibly around the margins of shallow lagoons, where they were occasionally subject to periods of desiccation. Periodic high energy events interrupted stromatolite growth. In environments of slightly more persistent agitation, possibly due to wave action in a more open location, stromatolite growth became restricted to bioherms, the growth of which may have been initiated on surface irregularities. Once stromatolite columns developed, their growth may have been maintained by the ability of algal mats to trap mud out of suspension, while no sediment was deposited on the intervening mat-free areas. Lamination within columns reflects variation in the rate of sediment supply and the nature of this sediment, alternating periods of mechanical deposition and algal trapping, and periodic *in situ* precipitation.

Columnar stromatolites have similar morphologies in all areas (Preiss, 1973) but they are associated with similar facies. Hence similar environments, both in physical and chemical characteristics, were present over wide areas.

Environmental influences may thus account for morphological similarities, although some biological influence cannot be excluded. This contrasts with the Brighton Limestone, which has a much greater local and regional variation in facies, and hence depositional environments, and contains a much wider morphological variety of columnar stromatolites (Preiss, 1973).

Small domal stromatolites represent a lower energy environment than columnar forms, as indicated by the lack of erosion and the absence of coarse sediment deposited contemporeneously with stromatolite growth. In this environment, more continuous colonization of the substrate by algal mats was permitted (Horodyski, 1977). Aitken (1967) and Haslett (1976) also consider that a lower degree of turbulence and a reduced total sediment influx favour the formation of domal stromatolites. However the relative importance of mechanically deposited, as compared with organically stabilised sediment, may also be significant (Horodyski, 1977). The predominance of mechanical deposition of dolomite mud within this sequence, may be one of the factors accounting for the predominance of flat laminated dolomite mudstones, as compared with stromatolitic dolomites.

Both small and large domal stromatolites grew in submergent environments. In the latter form, the regularity and continuity of lamination, and the high degree of inheritence of the lamination, indicates undisturbed growth in a low energy environment.

In the above interpretation, physical factors within the environment of deposition have been considered as influencing stromatolite morphology and abundance. However the abundance of different algal species may also have influenced stromatolite morphology. In addition, due to the predominance of subaqueous deposition of dolomite mudstones, penecontemporeneous cementation of the sediment, which is necessary for structures with relief to develop, may have been inhibited, and partly account for the low abundance of stromatolites.

MASSIVE DOLOMITES WITH DISSOLUTION AND OTHER DIAGENETIC FEATURES

The dolomite facies previously discussed, have experienced a simple diagenetic history, with lithification and compaction occurring without significant destruction of the primary depositional texture. However pervasive recrystallisation may destroy finer detail. In contrast with this, massively outcropping, unbedded to indistinctly bedded grey dolomites, in several areas of the central southern Flinders Ranges, have experienced a more complex early diagenetic history, in which their depositional texture was significantly altered. This facies is most common in the middle part of the Yadlamalka Formation at Yednalue (Fig. 6.4), where it forms horizons 2-27 m in thickness. It is also present in the Yednalue Anticline, particularly the east limb, and at Willow Creek. Isolated horizons are possibly present at Mundallio Creek and Torrens Gorge.

The massive dolomites form horizons continuous for up to 4 km along strike before outcrops end. Boundaries are generally sharp and planar, although some may gradationally overly unaltered dolomite mudstones. One horizon at Yednalue has relief on the upper surface of 4 m, and is overlain by bedded sediments. Rare breccia beds of very irregular, rounded to angular clasts (Plate 6.10a) in a matrix of dolomite silt, overly massive dolomites, from which they were derived by erosion. Some clasts contain wavy lamination resembling that present in calcretes.

Textures and their Origin

Many massive dolomites have the appearance of a breccia, although the fragments are angular and very irregular in shape (Plate 6.10b and c), and are clearly not detrital in origin. The fragments consist of micrite and microspar, and less commonly have a peloidal fabric. They are enclosed in a thin rim cement, and more equant, blocky spar (Plate 6.10b and c). In some samples clear spar has also formed by replacement of the microspar fragments. Lamination is preserved in some samples where brecciation of the original sediment is minor. However in general, the original textures have been destroyed. This fabric appears to have originated by brecciation and dissolution of lithified mudstones, and subsequent cementation of the brecciated sediment.

Other dolomites consist of peloidal grains (less than 1 mm in size) in association with larger, more irregular, but rounded grains (less than 1 cm in size), both of micrite and microspar, with subsiduary coated grains (Plate 6.11a-c). These grains are also enclosed in a rim cement (sometimes absent), and equant blocky spar. Coated grains (less than 1 cm in size), consist of concentric coatings of alternating, slightly wavy laminae of microspar and fine spar, around a nucleus which is often a fragment of homogeneous microspar (Plate 6.11a and b). The pore space between peloids and other grains, is irregular in shape (Plate 6.11a and c), and less commonly laminoid in appearance (Grover and Read, 1978). Rarely, large pores are infilled with homogeneous clear microspar (Plate 6.11c), which may have originated as an internal silt-size dolomite sediment, analogous to the vadose silt of Dunham (1969).

The textures of these peloidal dolomites are also not detrital in origin. They may have formed following dissolution of mudstones, and redeposition of micrite as peloids and coated grains, analogous to the formation of inorganic peloids and coated grains in calcrete profiles (Harrison, 1977; Harrison and Steinen, 1978).

Cements in this facies consist of a thin rim cement (less than 0.1 mm in thickness), followed by blocky, equant spar. The rim cement may have an acicular appearance with radial sperulitic extinction, but is more commonly recrystallised to fine microspar with only a vague acicular texture. Textures indicative of cementation within the vadose zone, such as incomplete rims, meniscus cements, and gravitational cements (Badiozamani *et al.*, 1977), are lacking, and hence cementation probably occurred in the phreatic zone. In modern environments, acicular aragonitic and Mg-calcite cements precipitate from seawater, or other solutions with abundant Mg^{2+} ions (Badiozamani *et al.*, 1977), but these cements are entirely dolomitic. The texture of the blocky spar may have formed during recrystallisation, as abundant veins which cut the fragments of microspar, are recrystallised within the blocky spar. However it was probably initially precipitated from solutions more dilute than those from which the rim cement precipitated.

Discussion

A complex history of solution and precipitation is apparently characteristic of diagenesis within a subaerial environment (Purdy, 1967). Dissolution may occur under subaerial conditions in both the phreatic and vadose zones, while in a marginal marine environment, dissolution results when an influx of fresh waters causes undersaturation (Grover and Read, 1978). Vadose silt is common as an internal sediment in dissolution affected carbonates (Dunham, 1969; Grover and Read, 1978).

Some features of this facies, including brecciated textures, and peloidal and coated grains which have formed *in situ*, are similar to those in modern and ancient calcrete profiles formed during subaerial exposure (James, 1972; Walls *et al.*, 1975; Harrison, 1977; Harrison and Steinen, 1978). However the laminated crusts typical of calcrete horizons are not present except possibly within erosionally derived intraclasts, and these massive dolomites are sometimes much thicker than observed calcrete profiles. However the similarity in texture indicates that similar processes produced these dolomites, and they were subject to diagenesis in the presence of alternately saturated and undersaturated solutions. The formation of coated grains *in situ* is apparently unique to vadose diagenesis under exposed conditions (Harrison, 1977), hence these diagenetic fabrics may have developed during periods of subaerial exposure.

This facies resulted from a complex history of vadose and phreatic diagenesis, associated with periods of exposure, and the introduction of dilute groundwaters which were undersaturated with respect to the carbonate minerals present within the sediment. Dissolution and precipitation of carbonate may occur within a single carbonate profile, due to evaporative concentration in the vadose zone causing saturation and precipitation, while undersaturated conditions are maintained below the water table. Salinity may also vary due to seasonal or longer term climatic changes (Purdy, 1967), producing alternating periods of dissolution and precipitation. Undersaturation also results from mixing of different water types, each of which may be saturated (Runnells, 1969; Wigley and Plummer, 1976).

This facies is in contrast to the majority of carbonate facies within this sequence which were not invaded during early diagenesis by solutions of significantly different chemistry from those in which they originally formed. It is also characterised by a very low terrigenous content, and hence developed during periods of prolonged exposure and non-deposition, following near pure carbonate deposition. Exposed low relief carbonate islands may have developed periodically in the Willow Creek-Yednalue-Yednalue Anticline area, due to minor uplift. A lens of fresher ground water developed below these islands, above more saline groundwaters typical of the normal basin water, resulting in substantial diagenetic changes (Fig. 6.8). The development of a fresh water lens requires a humid climate. In the Willow Creek area, evidence of more significant uplift during deposition of the Yadlamalka Formation, is indicated by the presence of synsedimentary breccias interbedded in facies typical of the Yadlamalka Formation. These breccias, which are described by Preiss (1979b), consist of equant, angular clasts, ranging in size from a few millimetres to 1.5 m and rarely much larger. The clasts include lithologies typical of the Yadlamalka Formation, such as grey dolomite mudstones, black chert, dolomitic sandstones and intraclastic magnesite. They were derived from erosion of uplifted areas of exposed, lithified Yadlamalka Formation, and formed contemporeneously with deposition of this formation. The breccias were deposited in large channels, probably in a subaerial environment (Preiss, 1979b), and five distinct episodes of channel formation occurred, each causing erosion of previously deposited breccias.

MAGNESITE FACIES

Magnesite facies are far exceeded in abundance by dolomite facies in most areas (Figs. 6.1 and 2). They form a significant proportion of the sequence in only a few areas, Depot Creek (11%), Arkaroola (6%), Copley to Myrtle Springs (21-18%), and the southwestern Willouran Ranges (14%, see Table 6.1). In outcrop, magnesite facies are readily distinguished from dolomite facies, by cream to yellow weathering outcrop surfaces, although fresh surfaces may be cream to dark-grey. The presence of magnesite has been confirmed by X-ray analysis, and some samples have been stained using the method of Friedman (1959).

The dominant magnesite facies is intraclastic magnesite (Forbes, 1960, 1961; Figs. 6.3-7), however the facies present will be described in a sedimentary environment, evolutionary series, rather than in order of abundance.

INTERLAMINATED MAGNESITE-DOLOMITE MUDSTONES

Several interbeds of this facies are found in the Montacute Dolomite south of Torrens Gorge, but elsewhere it is very rare (Table 6.1). This facies gradationally overlies dolomite mudstones, or sharply overlies dolomitic sandstones, and is erosionally overlain by intraclastic magnesite or dolomitic sandstones (Plate 6.12a).

Sedimentary Structures and their Origin

This facies is well laminated, and consists of alternating magnesite and dolomite rich laminae, which are 1-10 mm in thickness. The magnesite laminae are light-grey to cream in colour, and consist of micrite or microspar, despite metamorphism to biotite facies in areas of outcrop of the Montacute Dolomite. They may also contain dolomite as a minor component. Dolomite laminae, dark-grey in colour, are recrystallised to fine spar, and may irregularly replace the margins of adjacent magnesite laminae. They also contain quartz silt and sand, magnesite intraclasts, and sometimes magnesite mud. The magnesite intraclasts were derived from the laminae of magnesite mudstone, which may be desiccated. Tepees, 10-40 cm high (Plate 6.12a) are common, but have been attenuated by shortening during the Delamerian Folding, which in this area (Torrens Gorge) produced a slaty cleavage and crenulation in the more pelitic units. Within tepees, magnesite laminae are fractured, disrupted, and occasionally crumpled, whereas dolomitic sediment was apparently unlithified, and infilled the fractures which formed and the centres of tepees. The tops of tepees were truncated by erosion, and intraclasts derived from erosion of magnesite laminae were deposited in hollows.

The magnesite laminae appear to have been micritic crusts formed on top of detrital laminae on an exposed mudflat, where evaporative concentration caused precipitation of micritic magnesite at the sediment surface (c.f. Smoot, 1978). Alternation of micritic crusts with unlithified detrital sediment laminae, indicates deposition of fine grained detrital sediment, alternating with periods of exposure (Hardie *et al.*, 1978). Tepee growth continued over several cycles of deposition and exposure. Precipitation of micritic crusts may have caused expansion of the sediment surface and tepee growth. However the disruption of unlithified dolomitic sediment in the centre of tepees during growth and its incorporation of pieces of micritic crust, and the apparent upward movement of this dolomitic sediment into the centre of some tepees, suggests that upwelling groundwater may have contributed to tepee formation (von der Borch and Lock, 1979).

MAGNESITE MUDSTONES

This facies consists of laminated and thinly bedded, cream (due to weathering) to dark-grey mudstones, which consist largely of magnesite, but contain a few percent dolomite (rarely up to 50%)[1], as an intimate admixture. It represents 1 to 3% of the sequence in the Copley-Myrtle Springs area, but elsewhere is very minor although often present (Table 6.1). Magnesite mudstones generally overlie dolomite mudstones, often with a sharp boundary (Plate 6.13a). They are invariably overlain by intraclastic magnesite, and less commonly are gradational into nodular magnesite. Some interbeds are completely enclosed within intraclastic magnesite. Beds of magnesite mudstone are up to 30 cm, and rarely 1.5 m in thickness.

Sedimentary Structures

The mudstones consist of micrite which appears dark-brown in thin section. Impurity content, which includes detrital quartz silt, authigenic quartz and albite, and talc, is low. Flat to wavy lamination is characteristic (Plates 6.12b, 6.13a), but there is little textural contrast between adjacent laminae. Minor silty laminae are planar or lenticular, and some are erosionally based. Graded laminae of quartz silt to magnesite micrite are rare. Elongate intraclasts form thin lenticular interbeds (Plate 6.12b). Some intraclasts were deposited almost *in situ* with little transportation (Plate 6.12c), but are characterised by a rounded, although elongate shape. Intraclastic lenses resulted from the erosion of desiccated surface crusts and tepees. The latter are small, 4-10 cm and rarely 40 cm in height, and up to 40 cm apart (Plates 6.12b, 6.15a). Some thick mudstone units contain several horizons of tepees (Plate 6.12b). Both bending and fracturing of the laminae occurred during tepee growth. In contrast with tepees in dolomite mudstones, these tepees lack unlithified detrital laminae which infilled the central zone. Buckling and fracturing may have resulted largely from

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1. The percentage dolomite was determined from X-ray diffraction traces, see Appendices 3 and 4, and also roughly approximated from powder films.

expansive crystallisation within the sediment (Assereto and Kendall, 1977) on exposed mudflats, where the water table remained near the surface, thus supplying solutions from which the micrite precipitated within the sediment. Tepee growth was followed by erosion, or further mud deposition and another cycle of tepee growth.

Discussion

Deposition of magnesite mudstones occurred in very low energy environments virtually free from detrital influx. Such an environment may have been shallow peripheral flats shoreward of the zone of wave agitation, or lagoons physically separated from a larger basin. Desiccated mud layers were eroded by small wind-generated waves around the margin of these lagoons. There is little evidence for detrital reworking of magnesite mud as is the case for dolomite, hence the magnesite may have been precipitated from lagoon waters and deposited from suspension. This resulted in homogeneous mud laminae, similar to those of protodolomite forming in ephemeral lakes associated with the Coorong (von der Borch and Lock, 1979). Further interstitial precipitation on exposure produced tepees. The original precipitate was probably a magnesium carbonate, although it may have been a hydrated variety such as nesquehonite or hydromagnesite (Forbes, 1961; see Chapters 9 and 12).

NODULAR MAGNESITE

Laminated grey magnesite mudstones may be partly or completely replaced by a mosic of white nodular, micritic magnesite (Plates 6.12c, 6.13b and c). Rarely nodular magnesite is present in other facies e.g. dolomitic sandstones (Torrens Gorge) or dolomite mudstones (Depot Creek). Completely replaced beds resemble intraclastic beds, as some of these consist of close-packed intraclasts with very little matrix. The presence of regular, rounded intraclasts, and internal bedding, distinguishes the latter from nodular magnesite. However the two facies are intergradational when erosion and transportation of intraclasts derived from nodular magnesite, has been

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minimal. These clasts still have irregular, but rounded shapes. Both magnesite facies may be badly weathered, and this makes distinction between them difficult.

This facies is most abundant in the upper part of the Yadlamalka Formation at Copley (Fig. 6.6). Elsewhere, it is rare, as magnesite is largely represented by intraclastic beds.

Description of Texture

The nodules consist of micritic magnesite, which in thin section has a lustre-mottled appearance, and is much paler than the darker brown of the host micritic magnesite, which also contains minor dolomite (Plate 6.14a and b). Within the nodules, dark stringers of the host micrite are present. The nodules may be isolated botryoidal structures within the host sediment (Plate 6.12c), or sometimes have a "folded" appearance resembling the enterolithic structure of anhydrite nodules (Plate 6.13c). They disrupt and crosscut the lamination within the host sediment, suggesting replacive and displacive growth. Nodules within dolomitic sandstone lack sand grains, indicating displacive growth. With continued growth, the magnesite mudstone may be replaced by a mass of coalescing nodules, separated by stringers of host sediment (Plate 6.13c). This fabric resembles the chicken-wire texture of nodular anhydrite.

In thin section, it appears that nodule growth occurred by the formation of small discrete areas (0.15-0.3 mm) of pale micrite within darker host micrite (Plate 6.14b). With continued growth, these small areas coalesced to form nodules, which internally have a vague granular texture, reflecting the original formation of small, isolated, micritic areas. This granular fabric, and the lustre-mottled appearance (Plate 6.14a and b), is characteristic of all nodular magnesite, and enables easy recognition of intraclasts derived from it (Plate 6.14c). Although this facies is minor, intraclasts derived from it are common, and often the dominant type within intraclastic magnesite, indicating that this facies was much more abundant than its present record indicates. The nodules lack any relic lamination derived from the host sediment. The impurity content is low, and includes talc and authigenic albite.

Discussion : The Origin of Nodular Magnesite

Although the appearance of nodular magnesite is somewhat similar to that of nodular anhydrite (West, 1965; Maiklem *et al.*, 1969; Shearman, 1978), it is likely that these are primary magnesium carbonate nodules. The reasons for this include the similarity of texture in nodules at all stages of growth, the preservation of identical texture in intraclasts reworked from nodular magnesite, the absence of relic sulphates or pseudomorphs after sulphates, and the micritic nature of the magnesite. Nodular anhydrite is more coarsely crystalline, and consists of rectangular cleavage fragments (Shearman, 1978).

The morphology is also similar to nodular magnesite calcretes described by Wells (1977, Figs. 2, 3 and 7) at Gosses Bluff, in the Northern Territory. The magnesite is microcrystalline, and forms irregular nodules. Nodular fabrics are common in calcrete profiles (Reeves, 1976), although they may be associated with pisolites. Nodular fabrics are also preserved within submarine limestones. The nodules consist of carbonate-rich material, enclosed in more impure carbonate with clay and other impurities, and which may form stylolitic zones between nodules (Garrison and Fischer, 1969; Hudson and Jenkyns, 1969, Tucker, 1974). Nodules form during periods of slow deposition, or non-deposition, allowing redistribution of carbonate within the sediment, or preferential cementation in some areas (Jenkyns, 1974; Tucker, 1974). However, because of their environment of formation, these nodules are only rarely eroded to form intraclastic beds (Hudson and Jenkyns, 1969; Tucker, 1974).

The nodular fabrics developed diagenetically within the sediment, probably at and immediately below the sediment surface, as indicated by the frequent erosion of this facies. Growth occurred during periods of non-deposition, following subaerial exposure of magnesite mudstones. Lithification, as a result of nodule growth, protected the sediment surface from aeolian deflation. Evaporative pumping (Hsu and Siegenthaler, 1969) may have provided interstitial solutions in the vadose zone, from which micrite precipitated. Nodular calcrete profiles (1-2 m in thickness) in continental environments where they may be formed largely by percolating rainwater, may take up to 100,000's of years to form (Gardener, 1972; Chapman, 1974). However with continuously available groundwater, and an available supply of ions, growth of nodular fabrics is much more rapid, as is the case for nodular anhydrite. For example, anhydrite nodules have developed in 2-3 ft of supratidal carbonate sediments, in a 3 mile wide zone of the Trucial Coast, in the last 3,000 years (Shearman, 1978).

Hence this facies may record a period of exposure and non-deposition over a period of time of the order of several thousands of years. Nodular magnesium carbonate was precipitated within the sediment by replacement of the laminated mudstones, and by displacive growth within the sediment. Ions were probably provided from both the original sediment and groundwaters. The development of tepees in laminated mudstones, represents shorter periods of exposure, during which the ground water table may have been at and near the sediment surface. In contrast, nodule growth may have occurred largely in the vadose zone. This situation is analogous to a lacustrine sequence described by Tucker (1978), in which tepees formed during extended periods of exposure, while during more prolonged periods of shoreline retreat, calcretes developed.

INTRACLASTIC MAGNESITE

Occurrences of this facies generally far exceed in abundance the other magnesite facies from which they were derived. This facies consists almost entirely of clast-supported beds of rounded magnesite intraclasts. Matrix supported beds are uncommon. The beds are characteristically cream in colour, and are invariably weathered to some degree.

This facies is rare to absent in the Mirra Formation, and forms only a minor component of the Montacute Dolomite, and the Yadlamalka Formation in the southern Flinders Ranges with the exception of the Depot Creek area (Figs. 6.3 and 4). However it is more significant in the northern Flinders Ranges (Fig. 6.5), and is sometimes more abundant than dolomite facies (Fig. 6.6). In areas where this facies is minor, beds average about 0.5 m in thickness, with rare occurrences up to 2-3 m. Where this facies is more abundant, average bed thickness is greater (Table 6.1; Forbes, 1960). Beds thicker than 10 cm are plane bounded, or have an irregular, erosional, lower boundary when they overlie magnesite mudstone (Plate 6.15b). Outcrops are generally continuous for only a few tens or hundreds of metres, but thicker horizons (1-2 m) may be continuous for several kilometres (Copley, Myrtle Springs, Arkaroola). Hence it appears that beds of intraclastic magnesite were extensive sheet deposits (Forbes, 1955, 1960). Lenticular beds, a few metres in extent, are generally less than 10 cm in thickness, and rarely up to 0.5 m (Yacka).

Sedimentary Structures

Intraclastic magnesites are almost invariably clast supported, and texturally are grainstones and packstones. Intraclasts have textures characteristic of nodular magnesite and laminated magnesite mudstones

(Plate 6.14c). Some intraclasts are recrystallised, or may be partly replaced by dolomite on the margins. The maximum intraclast size in individual beds ranges from granule size to 20 cm, although most beds have a maximum of less than 5 cm. The intraclasts are elongate in cross-section (ratios up to 1 : 10) and hence probably discoid, or more equidimensional (Plates 6.15 and 6.16). Elongate intraclasts are commonly curved (Plate 6.15a and b), suggesting derivation from mud-cracked surfaces and tepees. Both elongate and more equidimensional clasts are rounded, although the latter may be irregular (Plate 6.16a and c). However intraclasts produced from desiccated magnesite mudstones are somewhat rounded, even where there has been little transportation (Plate 6.12c). Erosion of nodular magnesite would also produce rounded intraclasts, without significant abrasion.

Sorting is variable. Beds with intraclasts in the granule to small pebble size range, are generally well sorted (Plate 6.16b). Coarser beds are less well sorted, and sometimes bimodal (Plate 6.15a), possibly resulting from mixing of intraclasts from two discrete sources. Elongate intraclasts are largely arranged approximately parallel to bedding (Plate 6.15b), imbrication is less common. In poorly sorted beds, they may have random orientation, indicating rapid deposition and no reworking.

Units of intraclastic magnesite are internally bedded on a medium- to thick-bedded scale (Plates 6.15b and c, 6.16a). The contrast in grain size, sorting and matrix content between adjacent beds, reflects depositional events of contrasting energy levels and persistence. Thin interbeds of grey dolomite mudstone are common in thick horizons of small clast size (Plate 6.15c). Beds of granule to small pebble size, occasionally contain small tabular cross-beds less than 20 cm in thickness, ripple cross-lamination (Plate 6.16b), or symmetrical wave ripple marks. In most areas, there are infrequent inversely graded beds (Plate 6.16c), whereas normally graded beds are rare (Forbes, 1960). Inverse graded beds are characterised by close packed intraclasts at the base, passing upward into coarser, more poorly sorted intraclasts, with a higher matrix content. This feature may be caused by a high concentration of clasts and matrix in the transporting medium, during very high energy events. As a result, larger intraclasts move to regions of least shear away from the sediment surface, while smaller intraclasts remain near the sediment surface (Davies and Walker, 1974). Rapid deposition following a rapid decrease in energy, may preserve this clast size distribution, resulting in inversely graded beds. The deposition of an inversely graded bed (they may be up to 30 cm in thickness) thus records a single depositional event.

Cement or matrix is minor as most beds contain closely packed intraclasts. The cement is finely crystalline dolomite, usually with a recrystallised texture, and it may replace the margins of magnesite intraclasts. Scattered sand and silt are common. A matrix of dolomite and magnesite micrite or microspar is less common than dolomite cement. Talc is present as randomly oriented flakes in both intraclasts and matrix. Although talc is of metamorphic origin, its abundance may vary significantly between adjacent intraclasts (Plate 6.14c), suggesting that its abundance was determined by the composition of the precursor sediment. Talc may have formed from magnesium silicates such as sepiolite or palygorskite (which invert to talc on metamorphism, Velde, 1977) chemically precipitated with magnesite, or even from chemically precipitated talc, which may precipitate from Mg-silica rich solutions at high pH (Velde, 1977).

Replacement chert nodules, authigenic quartz and albite are minor constituents. The texture and origin of chert nodules is discussed in Chapter 11. Occasionally, extensive dolomitization to finely crystalline dolomite may destroy textures.

Discussion : Deposition of Intraclastic Magnesite

Beds of intraclastic magnesite do not in general represent intraclastic pavements formed by desiccation and erosion of surface crusts on exposed mud flats, although lenses and thin beds of intraclasts within laminated mudstone do. Rather they represent erosion of thick intervals (often more than 1 m) of magnesite mudstone or nodular magnesite during extended high energy periods, and the transportation of intraclasts into a range of environments. The occurrence of intraclastic magnesite as sheet deposits, and the lack of channels indicate low depositional gradients, and erosion of magnesite by sheet flow following widespread inundation or transgression of flat, exposed areas.

Deposition of beds that lack a preferred fabric and are poorly sorted, indicates a transportation process in which clasts were not free to move relative to one another (Harms *et al.*, 1975), and hence may represent rapid deposition following a high energy event in which there was a very high concentration of intraclasts in the transporting medium. Inversely graded beds formed when some rearrangement was possible, although the concentration was still high. In both cases intraclasts may have been transported largely in suspension, and hence indicate very high flow strengths, and possibly deposition close to the source. High flow strengths, which sometimes led to the deposition of beds up to 30 or 40 cm in thickness in a single event, were in marked contrast to the low energy conditions which prevailed during deposition of dolomite and magnesite mudstones.

Deposition of finer beds with elongate clasts arranged approximately parallel to bedding, indicates deposition from lower flow strengths, possibly with transportation largely as bedload, and deposition further from the source. Some intraclasts experienced wave and current reworking, which improved sorting, and formed cross-beds and ripples. However the grain size of the intraclasts does not depend entirely on flow strength when they are derived from erosion of nodular magnesite, but is also a function of the size of the nodules within the source bed.

Within an extensive, shallow basin, such as that in which deposition of this facies association occurred, it is difficult to generate high energy processes such as those described above. In areas close to the margin of the basin, land derived, unconfined sheet floods (Hardie *et al.*, 1978) moving across exposed magnesite mudflats, may have caused extensive erosion, and introduced minor sand which was deposited as matrix. Coarse beds were deposited close to source, while finer clasts were transported offshore to topographically lower parts of the basin, where they were deposited as interbeds within dolomite mudstones. However intraclastic magnesite is not solely confined to those areas which form the present outcrop limits of this sequence, and hence were closest to the basin margin, but it occurs throughout the basin. Exposed magnesite mudflats may have been inundated by transgressive events, which increased the size of the basin, and thus wave fetch and wave energy. Wave undercutting during severe storms, analogous to that described by Hardie and Ginsburg (1977), eroded lithified magnesite. Continuation of this process as the transgression continued, may have caused erosion of thick, extensive deposits of lithified magnesite.

TERRIGENOUS FACIES

SANDSTONES

Adelaide Region and Southern Flinders Ranges

Sandstones within the Montacute Dolomite in the Adelaide region, and the Yadlamalka Formation in the southern Flinders Ranges, are largely fine- to medium-grained, dolomite- and rarely quartz-cemented, and light- to medium-grey in colour, although outcrop surfaces weather to a dark brown.

The composition is subarkosic to arkosic. Sandstones are massively outcropping, compared with the interbedded dolomite mudstones. The abundance of sandstones is indicated in Fig. 6.1, and is greatest in the Yacka area where it increases from north to south. This increase is also associated with an increase in the abundance of quartz-cemented sandstones, and it appears that the Yadlamalka Formation is replaced laterally to the south and west, in this area, by the Bungaree Quartzite.

Sandstones form thin interbeds within dolomite mudstones, or thicker horizons up to 3 or 4 m in thickness (rarely 25 m, Yacka). The thicker horizons (1-4 m), may be followed for several kilometres along strike suggesting sheet like bodies, although outcrops in this region are usually very discontinuous.

Texturesand Sedimentary Structures

On the western margin of this area, sandstones are largely medium-grained, but fine- to very coarse-grained interbeds are present. Coarse-grained interbeds are most significant where sandstones are most abundant, probably reflecting greater proximity to source. In the more central areas of the southern Flinders Ranges, sandstones are largely fine-grained, but interbeds of medium to very coarse-grained sandstone are still present.

Sandstones are moderately to well sorted, and occasionally have a bimodal distribution. Grains are sub-rounded to well-rounded, although they are often embayed by dolomite (Plate 6.18a). Hence most sandstones are texturally mature, although less well sorted sandstones may exhibit textural inversions. Dolomite intraclasts, from sand size to 10 cm in size, and less commonly magnesite intraclasts, were deposited in sandstones. Sandstone beds with dolomite intraclasts, are generally coarser grained and less well sorted than adjacent beds which lack intraclasts. This reflects energetic depositional events, which eroded desiccated dolomite mudstones, but were of insufficient duration to eliminate intraclasts by abrasion. More persistent agitation produced better sorted sandstones.

The dominant sedimentary structure is flat to slightly wavy lamination and thin bedding (Plate 6.17a). Frequently there is little textural contrast between adjacent laminae. Cross-beds form isolated sets, up to 30 cm, and very rarely 1.5 to 2 m in thickness. Rare cosets contain only 2 or 3 sets. The cross beds are tabular, or shallow troughs (Plate 6.17b), and the laminae are generally tangential to the lower surface. Hence the sand was transported at least partly in suspension (Reineck and Singh, 1973). Ripple marks exposed on bedding planes are symmetrical, or sometimes slightly asymmetrical, have rounded profiles and straight crests (Plate 6.17a) and are wave formed. The wavelengths are less than 7 cm. Ripples were sometimes buried by dolomite mud to form flaser lamination. Because of internal homogeneity of sand within ripples, they often appear to lack internal lamination. However finer grained sandstones in the central southern Flinders Ranges, contain sets and cosets of ripple cross-lamination (Plate 6.17c). This was formed from wave-modified current ripples, as indicated by rounded profiles and offshoots. Desiccation cracks formed in thin beds of dolomite mudstone within sandstones, and these have commonly been eroded and deposited as intraclasts in the enclosing sandstones.

Discussion : Depositional Environment

Sedimentary structures within these sandstones are similar to those in sandstones of similar grainsize elsewhere in the Mundallio Subgroup. The association of structures present is similar to that present in wave formed sand bars described by de Raaf *et al.* (1977) and Roep *et al.* (1979). Flat lamination was formed by deposition of sand from suspension, possibly during storms. Cross-beds and ripple lamination formed in shallower, more turbulent water. Hence sandstones may have been deposited in submergent, to occasionally emergent wave formed bars or sand shoals. Shoreward of the swash zone, the sand shoals may have passed into very shallow to emergent sand flats. Coarse-grained sand was introduced to these flats during storms. As currents waned the surface of the sand was reworked to form ripple marks. During the subsequent slack water period, thin beds of dolomite mud were deposited, sometimes as drapes across the rippled surface. Exposure resulted in desiccation of the dolomite mud, and this was eroded during the next high energy period with its associated influx of sand. In more persistently agitated parts of the sand shoal, mud deposition was inhibited, and well sorted sandstones were deposited.

Interbedding of sandstone and dolomite mudstone in this facies, occurs on a scale of thick lamination or greater. This is too coarse to attribute to alternating tidal current bed-load transport and slack water deposition. It is more likely to result from severe storms, or longer term variation, causing an alternation of agitated and quiet water conditions. The deposition of dolomite mud rather than terrigenous mud during slack water periods, reflects the scarcity of terrigenous mud as compared with coarser terrigenous detritus.

Northern Flinders and Willouran Ranges

In this region the average grain size is much finer than to the south (Forbes, 1961), falling in the coarse-grained silt to very fine-grained sand size range, and hence is similar to, or slighter finer grained, than sand in the lower Mundallio Subgroup. Sandstones are predominantly dolomite-cemented. Quartz-cemented sandstones are abundant only in the Mirra Formation.

Dolomite-cemented sandstones form interbeds which are generally less than 3 m in thickness, but are rarely up to 10 m. Sand size grains may be exclusively terrigenous, with a subarkosic composition. However many sandstones also contain dolomite peloids (Plate 6.18b). The abundance of dolomite peloids and dolomite cement exceeds that of terrigenous sand in some beds, and hence strictly they are very silty or sandy dolomites, or dolopelsparites. However because there is a continuous gradation from sandstones with only terrigenous grains to very sandy pelsparites, and because they contain the same sedimentary structures, these facies are considered jointly. At Arkaroola, and in the Top Mount Bore area, dolomitic sandstones generally contain dolomite peloids as a significant component. Elsewhere the proportion of sandstone in the sequence is greater, and sandstones with and without dolomite peloids are present. Dolomite is also present as laminae of dolomite mudstone, especially in sandstones with dolomite peloids. On fresh surfaces, the sandstones are light- to medium-grey. They weather to a dark brown on outcrop surfaces but become lighter brown as the proportion of dolomite, as both peloids and cement, increases.

Quartz-cemented sandstones (quartzites) are white to pale-grey on fresh surfaces, weathering to pink or brown. Rarely, interbeds up to medium-grained sand size are present within finer grained sand. In the Arkaroola, Copley-Myrtle Springs, and Top Mount Bore areas, the only major quartzite beds occur in association with shales, forming part of coarsening upward cycles (Fig. 6.2 and 6.5). These are continuous within the existing outcrop of the Yadlamalka Formation at both Arkaroola and Copley-Myrtle Springs (Figs. 4.8, 9). In the West Rischbieth area, there are two major quartzite horizons, each about 20 m in thickness and continuous along strike for at least 5 km. Occasional thinner interbeds are also present. However in the Mirra Formation, quartzites are more abundant, especially in the eastern Willouran Ranges (Fig. 6.7), and they may exceed dolomitic sandstones in abundance.

Textures and Sedimentary Structures

Sandstones are generally well sorted, but grain rounding is frequently obscured by replacement of grain margins by dolomite, or by quartz overgrowths. Where this feature is apparent, grains are sub-rounded to rounded, hence these sandstones are mature to supermature. In dolomitic sandstones, interbeds of less well sorted, fine- to coarse-grained sand, often with rounded sand and granule sized dolomite and magnesite intraclasts, form a minor component.

Dolomitic sandstones contain alternating sets of flat lamination and cross-lamination (Plate 6.18c), with the latter commonly eroding the former. Quartzites are largely flat laminated. Individual sets of cross laminae have amplitudes of 1-2 cm, and rarely up to 10 cm. They occur as complexly interwoven sets (Plate 6.18c), with erosive bases, and tangential laminae. Most cosets indicate variable direction of movement, although one direction may be dominant. Ripple crests are rarely preserved on bedding planes due to the erosive nature of most sets. This ripple cross-lamination is similar to that in sandstones of similar grain size in the Nankabunyana Formation. It resembles cross-lamination formed by wave ripples, and combined wave-current ripples (Harms *et al.*, 1975; de Raaf *et al.*, 1977). Some ripples are gently climbing, usually with only the lee side preserved (Plate 6.19c). This implies low flow strengths and an abundant supply of temporarily suspended sand, due to a fairly rapid change in flow strength (Harms *et al.*, 1975). The absence of larger scale current formed structures is a function of grain size, and the rare tabular cross-beds are found in fine- to medium-grained sandstones.

Ripple marks preserved on bedding planes are largely symmetrical wave ripple marks, with wavelengths of 1.5 to 8 cm. Asymmetrical ripple marks, with wavelength of 2 to 7 cm, have rounded profiles and straight crests, indicating wave modification (Harms, 1969). Ladder ripples indicative of variable wave orientation in shallow water, occur infrequently in dolomitic sandstones, but many ripple marks preserved in quartzites are of the ladder ripple type (Plate 6.19b).

Dolomite mud laminae were sometimes deposited in ripple troughs, and define a flaser lamination. Planar dolomite laminae are also present, and they often contain synaeresis cracks (Plate 6.19c). These cracks have a more irregular margin than desiccation cracks, are sometimes ptygmatic due to subsequent compaction, and may be filled from above or below. Desiccation cracks, forming regular polygonal patterns on bedding planes, are relatively uncommon, being less abundant than synaeresis features. They are rare to absent in the Mirra Formation.

Discussion : Depositional Environment

The grain size of this facies is probably largely a function of the available sediment supply. Even where sandstones are abundant, the grain size is still coarse-grained silt to very fine-grained sand. However the depositional environment was probably similar to the coarser grained counterpart to the south, i.e. submergent to occasionally emergent wave formed bars and shoals, and shallow shoreline sandflats. In general wave agitation was sufficient to winnow dolomite mud from the sand, but muddy laminae were deposited during slack water periods on the shallow sandflats. As in coarser grained sandstones in the south, mud laminae are almost exclusively dolomitic rather than terrigenous.

Although wave and current ripple lamination, and flaser lamination are common in tidal environments, they may also form in wave formed bars in marine environments (de Raaf *et al.*, 1977), and in the shoreline zones of lacustrine environments (Clemmensen, 1978; Link and Osborne, 1978). Hence there is no clear evidence in this facies, or in coarser grained sandstones to the south, for the activity of tidal currents during deposition.

SHALES AND SILTSTONES

Montacute Dolomite and Yadlamalka Formation

The fine grained component of this facies association is dominated by dolomite mudstones. Shales and siltstones are minor (Figs. 6.3-6). The major shale interbeds are 3-15 m thick beds, which are non-dolomitic, grey in colour, with olive green and rusty brown weathering. Lamination is indistinct (shales do not split easily along shale partings), and hence their appearance is similar to shales in Unit 3 of the Nathaltee Formation, Unit 4 of the Nankabunyana Formation, and the Camel Flat Shale. Some of these shale interbeds form the lower part of coarsening upward shale to sandstone cycles (Figs. 6.3-6). These shale beds represent deposition of mud from suspension in a low energy environment free from current reworking. Thinner shale beds, which are laminated and often dolomitic, form a minor component of the Yadlamalka Formation in the southern Flinders Ranges.

However rarely shales may be more significant. At Yatina, the lower part of the Yadlamalka Formation contains about 40 m of grey, dolomitic shales, which are laminated or massive, and contain dolomite interbeds and nodules (Fig. 6.4). These shales may represent a slightly deeper water environment in this area, which was less favourable to dolomite deposition. At Yacka, the uppermost part of the Yadlamalka Formation consists of interbedded grey dolomite mudstones and grey shales (Fig. 6.4), possibly due to intertonguing with the Woolshed Flat Shale.

Mirra Formation

Within the Mirra Creek area, dark grey dolomitic shales and siltstones are conspicuous in the middle part of the Mirra Formation (Fig. 6.7). This facies is probably gradational into very fine-grained dolomitic sandstones. It represents low energy conditions, in which fine terrigenous and carbonate detritus accumulated.

In most areas of outcrop of the Mirra Formation there are large gaps in the sequence which may be occupied by either this facies or more dolomitic mudstones.

CYCLIC SEDIMENTATION

MARKOV CHAIN ANALYSIS

A Markovian process is one ".....in which the probability of the process being in a given state at a given time, may be determined from knowledge of the immediately preceding state" (Harbaugh and Bonham-Carter, 1970, p. 98). Markov Chain Analysis may be used as a means of recognising cyclicity in stratigraphic successions (Miall, 1973; Jones and Dixon, 1976). It may also be used to indicate that a particular facies has a small probability of succeeding another facies, and hence suggesting that it has a random occurrence in the succession (Jones and Dixon, 1976).

Markov Chain Analysis was performed on selected intervals, with few if any gaps in outcrop, of several stratigraphic sections from different areas (Fig. 6.9). The embedded chain method, in which the transitions recorded are between distinct lithological types, was used, rather than recording the lithology at fixed intervals through the succession (Miall, 1973). The matrices produced are listed in Appendix 2, and the results summarised in Figure 6.9.

Example A is the COPLEY MAGNESITE DDH1 section (Fig. 6.6), and the analysis includes transitions on a scale of a few centimetres, not all of which are shown in Figure 6.6 due to the scale of presentation. Markov Chain Analysis reveals a clearly defined cycle, in which all transition probabilities are greater than 0.1 (Fig. 6.9)[1]. Examples of the cycle outlined in Figure 6.9 in the original stratigraphic section, range in

1. A χ^2 test applied to this cycle, allows rejection of the Null Hypothesis, thus indicating that the sequence is not random.

thickness from 1.2-2.4 m, and rarely up to 3.7 m. Imperfect cycles, with one or more facies missing, or with repetitions, have more variable thicknesses. The cycle outlined begins with intraclastic magnesite, and is often erosionally based. It is also a fining upward cycle, and represents a transition from detrital to chemical deposition.

The other examples show less well developed cyclicity, and weaker linkages between facies. In Examples B and C, also from Copley, the number of occurrences of facies 2A, and 4 in particular, is much less than facies 1 and 3. Hence although weak 1 to 3 to 4, and 1 to 2A to 3 to 4 cycles are indicated, the alternation of facies 1 and 3 is equally significant, due partly to the greater frequency of occurrence of these two facies.

Example D, from Yednalue, includes stromatolitic dolomites and non-dolomitic shales, although there are only two occurrences of the latter and hence they cannot be meaningfully considered. Once again there are weak 1 to 2B to 3 to 4, and 1 to 3 to 4 cycles indicated, although the presence of stromatolitic dolomites complicates the cycle.

The matrix derived from the complete section at Depot Creek (Example E), indicates the presence of an absorbing state (Jones and Dixon, 1976), in this case dolomite mudstones, with no transitions between the other states indicating that they may be random. Although the χ^2 test indicates rejection of the null hypothesis that the sequence is random, it is biased by the presence of an absorbing state. When only the lowermost sandy part of the sequence is considered, Example F, dolomite mudstones still form an absorbing state, but weak 1 to 2B to 3 cycles are indicated.

Hence in summary, the only cycle which appears to be present, other than an alternation of two facies, is a 1 to 2A or 2B to 3 to 4 (to 5) cycle, or variations on this.

ORIGIN OF CYCLES AND MORE RANDOM SEQUENCES

Most areas show a more random arrangement of facies, due to the dominance of one facies, usually dolomite mudstone, or an alternation of two facies, such as dolomite mudstone with either dolomitic sandstone or intraclastic magnesite. However where magnesite mudstones are present, they are almost always overlain by intraclastic magnesite, and sandstones and intraclastic magnesite are commonly associated when both are present. The origin of the cycle outlined by Markov Chain Analysis, that is intraclastic magnesite to sandstone to dolomite mudstone to magnesite mudstone, will be discussed in order to indicate the relationships between the different environments in which these facies formed, and the possible causes of facies changes.

This cycle is a mixed carbonate-clastic cycle, and also involves a change in carbonate mineralogy from dolomite to magnesite in the upper part. Carbonate-clastic cycles have been described much less frequently than either pure carbonate, or pure clastic cycles. Hoffman (1975), Beukes (1977), and Siedlecka (1978), describe cycles with an upward increase in carbonate content, and a decrease in clastics, which formed in subtidal to intertidal marine environments. The example described by Hoffman (1975), contains shales as the clastic component, and hence is not applicable here. In the other examples, the cycles resulted from an influx of clastics, followed by progradation. Lacustrine sequences also contain carbonate-clastic cycles (Eugster and Hardie, 1975; Clemmenson, 1978; Link and Osborne, 1978; Surdam and Stanley, 1979). These may result from climatic variations, and periodic uplift in the source area. The development of a more humid climate may increase the amount of detritus supplied to the lake, while a more arid climate may lead to carbonate, and evaporite deposition. Climatic changes will have a greater effect on lacustrine environments, and possibly epeiric seas, than open marine basins.

The ideal cycle is summarised in Figure 6.10. Only the uppermost part of the cycle is shallowing upward. Both sandstones and dolomite mudstones contain occasional desiccation cracks, although rare tepees at the top of dolomite mudstones, record more extended periods of exposure. The middle part of the cycle may record a period when deposition approximately kept pace with subsidence, following the initial rapid deposition of intraclastic magnesite. This cycle also records marked changes in energy levels, from very high energy events capable of transporting magnesite intraclasts in suspension, to low energy environments in which chemical deposition of magnesite predominated. Changes between facies are often abrupt, reflecting the low palaeoslopes within the basin.

Inundation of exposed and lithified magnesite mudflats resulted in their extensive erosion, and the subsequent deposition of intraclastic magnesite. The erosive and depositional processes involved may have included land derived sheet flooding in areas near the basin margin, which eroded magnesite and transported it basinward, while increased wave energy in an expanding basin may have caused erosion of previously exposed areas. The deposition of intraclastic magnesite and the erosive events preceding it, record a major environmental change, and a dramatic increase in energy levels. Because magnesite mudstones were almost invariably subject to erosion, the processes leading to their deposition and subsequent erosion, were probably interdependent. Magnesite mudstones were deposited in the shallowest, most protected areas of the basin. Their deposition led to vertical accretion so that the area became permanently exposed, and a site of non-deposition. In these areas, magnesite mudstones were partly replaced by nodular magnesite. However this was probably associated with a fall in the water table, and hence a lowering of the water level in the basin. Renewed deposition in these areas occurred only when the water level rose. Because of the shallow nature of the basin, it would have been susceptible to climatic changes in the hinterland. A more humid climate may have led to transgressions in the basin and inundation of magnesite mudflats. Increased runoff resulted in basin margin mudflats being subject to sheet flooding as streams adjacent to the basin were reactivated. However hydrodynamic conditions changed with the rise in water level, so that following inundation, these areas were initially subject to erosion rather than deposition. Areas which had previously been low energy mudflats, experienced increased wave energy, possibly due partly to an increase in basin size, and hence wave fetch. Changing orientation of the shoreline with respect to the prevailing wind direction, and changing topography within the basin, may also have led to higher wave energy.

Sand was introduced during and subsequent to deposition of intraclastic magnesite. It forms a matrix and interbeds within intraclastic magnesite, and overlies it. The lack of associated terrigenous mud, and the rounded to well rounded nature of medium- to very coarse-grained sand, suggest that it may have been derived from aeolian deposits adjacent to the basin (Pettijohn *et al.*, 1973, p. 225). An expanding basin may have encroached across dune fields, and the dunes themselves may have prograded into the basin, as occurs on the leeward side of Qatar Peninsula in the Persian Gulf today (Shinn, 1973). Dunes may also have been eroded by streams and transported into the basin. Within the basin sand was reworked largely by waves on shallow sand flats close to source, and into wave formed bars further along shore. Sand probably did not reach more central areas of the basin. In these areas, exposed magnesite mudstones may have been eroded and redeposited, with little associated deposition of terrigenous sediment.

Deposition of sand was replaced by deposition of dolomite mud. The influx of sand may have been reduced due to changes in the source area. Within the basin, the formation of sand shoals would have changed basin topography, and probably led to the formation of semi-isolated lagoons, or very extensive, shallow mudflats, on which wave energy was limited except during severe storms. Stromatolites may have grown in the slightly more agitated environments around the margin of these lagoons. Progradation of dolomite mudflats, or simple vertical accretion, probably led to the formation of isolated ephemeral ponds, physically restricted from larger lagoons. These became sites of magnesite deposition. Increasing aridity may also have led to the formation of isolated ponds, as the water level within the basin fell. These environmental changes are summarised in Figure 6.11.

Areas of the basin underwent alternate periods of expansion and reduction in size with resultant facies changes, as a result of combination of factors. These may have included external factors, such as climatic changes, and tectonic events which changed the nature and position of the source areas. Within the basin, changes in the relative rates of deposition and subsidence, and variation in these factors between different areas, would have changed water depth and basin topography, and induced facies changes.

It is not possible to identify basin wide cycles. Because of the large area of the basin occupied by this facies association, the above factors would have varied across the basin, hence the abundance of each facies, and the nature of the small scale cycles which developed, also varied across the basin. However within the northern Flinders Ranges, different areas may show the same relative changes in the abundance of different facies at the same level. Hence over a period of time in which many small cycles were deposited, different areas were influenced by related events.

As previously indicated, many areas do not show well developed cyclity due to the absence of one or more facies, so that there is a simple alternation of two facies, or dolomite mudstones are dominant. Complete cycles of the type shown in Figure 6.10, are most likely to have developed in flat marginal areas, where there was a periodic sand supply, and magnesite was able to form during periods of shoreline retreat. In other areas adjacent to the basin margin, there was a periodic sand influx, but magnesite only rarely developed. Slightly greater palaeoslopes may have prevented the formation of extensive exposed mudflats, with ephemeral magnesite lagoons. In areas where the influx of sand was reduced, an alternation of dolomite and magnesite facies occurred.

FACIES DISTRIBUTION

The abundance of the major facies in different areas is indicated in Figures 6.1-7, and the abundance of all facies has been indicated in the facies descriptions. The variation in the abundance of different facies depends on the formation of different environments, as a result of changes in the factors discussed above. Because this facies association is characterised by rapid vertical facies changes, and almost all facies are represented within all areas of outcrop, lateral facies changes can only be discussed in terms of variations in the proportion of individual facies, and hence variation in the frequency of development of different environments.

NORTHERN FLINDERS AND WILLOURAN RANGES

In the Arkaroola region, dolomite mudflats predominated during deposition of the lower part of the Yadlamalka Formation. Intraclastic magnesite is largely fine grained (less than 1 cm), and may have been derived largely from outside the present outcrop area. However occasional interbeds of magnesite mudstone record the formation of ephemeral magnesite lagoons, and these are generally associated with coarser intraclasts. The proportion of magnesite varies little through this area, but may decrease near Blue Mine Creek. Coarse-grained silt and very fine-grained sand were periodically introduced, and became more common in the upper part of the formation. This was probably the result of changes in the source area. There is no change in the proportion and nature of magnesite facies in the upper part of the formation, indicating that the average water depth was probably similar throughout the period of deposition of this formation.

In the Copley-Myrtle Springs area, there is an increase in the abundance of dolomitic sandstones from Copley to Myrtle Springs (22-33%), possibly indicating that the Myrtle Springs area was more proximal to the sand source. However there is little change in the abundance of magnesite between these two areas. There is a 50% increase in the thickness of the Yadlamalka Formation from Copley to Myrtle Springs. However the continuity of some thin units in this area, and the lack of facies changes which would indicate significant depth changes in this area, suggest that the rate of deposition in general matched differential rates of subsidence and shallow water environments with low palaeoslopes were maintained. As in the Arkaroola area, dolomite mudflats were the most common environment in the lower part of the formation, and the influx of sand increased through time. Magnesite lagoons and mudflats developed more frequently in the upper part of the formation, and the clast size of intraclastic magnesite also increased. Cycles of intraclastic magnesite to sandstone to dolomite mudstone to magnesite mudstone, were often developed in the upper part of the formation, due to the environmental changes outlined in the previous discussion.

Within the Willouran Ranges, the outcrop occurs in four, distinct, approximately parallel zones (Top Mount Bore-West Mount Hut, West Rischbieth-Coronation Bore, Mirra Creek-Rischbieth Hut-Mirra Bore, and Mount Norwest H.S.-Willouran Hill), within which the proportion of facies remains similar along strike, but varies considerably between adjacent zones (Fig. 6.12). Within the first of these areas, deposition of the Yadlamalka Formation was again dominated by dolomite mudflats in the lower part, with an increasing sand content higher in the sequence. Magnesite is present throughout, except for the uppermost 100 m, which is predominantly dolomite with minor sand. Magnesite mudstones are minor (0.6%), but intraclastic magnesite is significant (13.1%). Hence magnesite mudstones deposited in this area were almost invariably subject to erosion, although some fine intraclastic beds (less than 1 cm), may have been derived from the southwest during periods of expansion of the basin. Northeastward, in the West Rischbieth-Coronation Bore area, the abundance of intraclastic magnesite within the Yadlamalka Formation is less, and it almost invariably contains intraclasts finer than 1 cm. The intraclasts may have been largely derived from the southwest, but rare beds of magnesite mudstone are present, and they are overlain by coarser intraclasts. Sandstones are more abundant in this area, and both this facies and dolomite mudstones appear to have been deposited in largely submergent environments.

The Mirra Formation was deposited northeast of the Bungarider Fault, in areas which were too distal from magnesite mudflats for intraclastic magnesite to be deposited other than very rarely. The Mirra Creek and Rischbieth Hut areas are characterised by rapid thickness changes away from the Bungarider Fault (Figs. 6.12 and 13; Murrell, 1977). Areas now occupied by outcrops of Callana Group were sites of slow deposition, and a thin sequence of submergent to emergent dolomite mudstones, stromatolitic dolomites and sandstones was deposited. In adjacent areas of more rapid subsidence, the environment of deposition was almost exclusively submergent. Dolomite mudstones dominate in the lowermost and uppermost parts of the Mirra Formation in this area, with periodic influxes of sand and silt in the remainder of the sequence. Vertical facies changes are not as rapid as in the Yadlamalka Formation. This may reflect slightly deeper water conditions, so that small changes in water depth did not always produce major facies changes.

In outcrops of the Mirra Formation in the eastern Willouran Ranges, the abundance of sand is greater than elsewhere. These sandstones were deposited as submergent sand sheets and bars. The interbedded dolomite mudstones and silty dolomites, were also deposited in submergent environments.

In the Willouran Ranges, there is a decrease in the abundance of sand from northeast to southwest, suggesting that the major influx of sand was from the northeast. Shallow water environments above wave base were developed throughout this area. Hence much of the basin floor may have been affected by wave action at any one time, allowing sand transport from northeast to southwest. However the two areas of minimal subsidence (Fig. 6.12) may have at times formed shallow barriers to circulation between adjacent parts of the basin. Infrequent influx of sand from the southeast probably occurred also, and this included coarse-grained sand. The sand influx into this area was transported away from the southwesterly margin into the Top Mount Bore-West Mount Hut, and West Rischbieth-Coronation Bore areas, along with granule sized magnesite and occasionally dolomite intraclasts, when the basin in this area was at its maximum extent.

Other outcrops of the Yadlamalka Formation in the northern Flinders Ranges, occur at Mandarin Hill (MH) and Avondale (AV). In the former area, only the uppermost part of the sequence is exposed, and it consists of interbedded sandstones and dolomite mudstones. At Avondale, outcrop is poor, and neither the base nor the top of the formation are exposed. The facies present are dolomite mudstone, dolomitic sandstone, and intra-clastic magnesite. The sequence is similar in appearance to that at Coronation Bore in the Willouran Ranges.

In summary, in the northern Flinders Ranges, a similar spectrum of environments and hence facies, is represented at Arkaroola, Copley-Myrtle Springs, and in the southwestern half of the Willouran Ranges. The only interuption to the deposition of dolomite mudstones, dolomitic sandstones and siltstones, and intraclastic magnesite, was the deposition of widespread shale, or shale-sandstone coarsening upward sequences (Fig. 6.2). These record periods in which the water depth was increased, possibly due to more extensive transgressions than those described for small scale, mixed carbonate-clastic cycles. Basin topography was featureless, so that wide areas were below wave base allowing extensive shale deposition. However gradual infilling of the basin caused a return to shallow water carbonate-clastic deposition.

More significant lateral facies changes occurred in a southwest-northeast direction across the Willouran Ranges. In this area, there were three sub-basins, separated by shallow areas with low subsidence rates.

SOUTHERN FLINDERS RANGES

Within this area, facies changes between areas are also related to the variation in the proportion of major facies, in particular sandstones and intraclastic magnesite (Fig. 6.1). Hence similar environments were developed in all areas, although in different proportions. Major environmental changes resulted in the transition to the laterally equivalent Skillogalee Dolomite and Woolshed Flat Shale to the south.

Within the Yacka area, there was a persistent sand influx throughout deposition of the Yadlamalka Formation. The decreasing abundance of sand from south to north (Fig. 6.14), and the continuation of this trend into the Bundaleer area (Forbes, 1955), suggests a southerly source. Both dolomite mudstones and sandstones contain evidence of exposure, indicating a shallow environment in this area. Magnesite mudstones were infrequently deposited, and were subsequently subjected to extensive erosion. The outcrop at Yatina may be a northerly extension of the Yacka outcrops. Here dolomitic shales are significant in the lower part of the formation, and sandstones and magnesite are minor.

Northwest of Yacka, towards the Crystal Brook-Port Germein Gorge area, the sandstone content also decreases. In this area, sandstone interbeds become thicker and more abundant toward the top of the formation, preceding the transition to the Undalya Quartzite. However dolomite mudstone is the dominant facies in this area, and was deposited on largely submergent mudflats. Restricted ephemeral magnesite ponds surrounded by exposed mudflats developed only infrequently. Hence the rates of deposition and subsidence were probably similar, so that relatively continuous deposition prevailed in largely submergent to occasionally exposed environments.

Further north in the Emeroo Range, both sandstone and magnesite deposition were more abundant at Depot Creek than in the northern and southern ends of the range. At Depot Creek, sandstones are most significant in the lowermost and uppermost parts of the formation, where they are commonly associated with intraclastic magnesite. The latter facies is present throughout the sequence, and often contains intraclasts in the 5-15 cm range, indicating deposition close to source. However magnesite mudstones are very minor, and hence they rarely escaped erosion. In the lower and upper parts of the Yadlamalka Formation at Depot Creek, the inundation and erosion of exposed magnesite mudflats was commonly associated with the introduction of sand. However during deposition of the middle part of the formation, the available sand supply in areas adjacent to the basin was limited. Dolomite mudflats were largely submergent, but were often colonised by domal stromatolites, while oncolites formed in areas of moderately persistent agitation. At Mundallio Creek, intraclastic magnesite is much less abundant, and also finer grained (clasts generally less than 3 cm). Dolomite mudstones predominate, and some contain white quartz nodules which may have formed as a replacement of sulphates, as a minor component. Hence exposed dolomite mudflats contained interstitial pore waters with abundant sulphate ions, a feature generally absent in other areas.

In the central part of the southern Flinders Ranges (Willow Creek, Yednalue, Yednalue Anticline, Johnburg and Carrieton), there are significant variations in thickness of the Yadlamalka Formation between adjacent areas. However similar facies are present in all areas. Exposed dolomite mudflats were a more common feature, especially at Yednalue and Willow Creek, than elsewhere in the Yadlamalka Formation. Sandstone interbeds are less common throughout this area than on the western margin of the southern Flinders Ranges, and form beds generally less than 1 m in thickness. The decreased abundance indicates a more limited sand source supplying these areas. However at Yednalue, the introduction of sand was often associated with the deposition of intraclastic magnesite, possibly following a transgressive event.

A feature of the Yednalue, Willow Creek and Yednalue Anticline (east limb) areas, is the presence of massive dolomites whose texture is largely diagenetic in origin. They formed during exposure, possibly following minor uplift to form low relief carbonate islands surrounded by carbonate mud flats. At Willow Creek, greater uplift produced more elevated areas, possibly with marginal cliffs. Erosion of these islands, contemporeneous with deposition of the Yadlamalka Formation, resulted in the deposition of breccias containing lithoclasts typical of Yadlamalka facies. These massive dolomites and breccias contrast with the remainder of the Yadlamalka Formation in which exposure and non-deposition generally followed the deposition of magnesite mudstones on extensive mudflats. In contrast, this facies has little associated magnesite, although it may occur in other parts of the Yadlamalka Formation in these areas.

Outcrops of the Yadlamalka Formation also occur in the easternmost southern Flinders Ranges, near Weekeroo (W). The sequence here contains little terrigenous detritus, and dolomite mudstone deposition predominated. Massive dolomites may record a similar event as the massive dolomites at

Yednalue, although their texture is recrystallised due to the higher metamorphic grade. However laminated dolomite and magnesite mudstones are still fine grained.

ADELAIDE REGION

Localised dolomite mudflats, which passed laterally into deeper water shales (Woolshed Flat Shale), were the major environment in which the Montacute Dolomite was deposited. These were largely submergent, but storms transported dolomite mud and minor terrigenous sediment onto adjacent exposed mudflats, which were also sites of magnesite deposition. Sandstones, with scattered magnesite intraclasts, often formed the base of cycles which began as transgressive events eroded the lithified magnesite crusts and tepees on these exposed mudflats (Fig. 6.3b).

SUMMARY

Deposition of this facies association occurred in a very shallow basin, in which all areas except perhaps the northeastern half of the Willouran Ranges, were periodically subject to exposure. Facies with features indicative of emergence, such as desiccation cracks, tepees, and intra-clastic interbeds, are not confined to outcrops around the margin of the Adelaide Geosyncline, as defined by present day outcrop limits. However it is not possible to position the margin of the basin in which this facies association was deposited, as nowhere does it thin to zero due to non-deposition, or pass into continental deposits. In addition, clasts typical of this sequence are common in the Sturtian glacial sequence which overlies the Burra Group. They were probably largely derived from erosion in areas outside the present outcrop limits of this facies association. Hence the basin was probably larger than that defined by present outcrops.

It appears that the basin of deposition was a series of shallow perennial, and smaller ephemeral lagoons or lakes, separated by exposed, low relief mudflats. Because of the widespread distribution of very shallow water facies, and the continuity of individual beds, it appears that the basin was characterised by low palaeoslopes. Hence high energy deposits formed extensive sheets rather than channels. Because of the low palaeo-slopes, small changes in water level would have caused flooding or exposure of extensive areas, both around the basin margin, and within the basin. Wind-tides may have been effective in causing small water level variations in local environments. Larger, more widespread water level variations, may have occurred due to changes in the climatic regime, and probably induced facies changes. Other changes, such as sediment buildups, or increased subsidence rates, would also have produced facies changes. In such a shallow basin, shorelines would have been ephemeral, highly variable features (Clemmenson, 1979). Sequences are characterised by frequent and often abrupt vertical facies changes, indicating that the topography of the basin, and the position of shorelines, were in a constant state of flux.

Despite the significant development of low energy environments in widespread areas, terrigenous mudstones are of minor importance. Terrigenous sediment is dominated by coarse-grained silt and sand, clays are minor. It is likely that any fine sediment which entered an extensive very shallow basin, would have been deposited during slack water periods, or in low energy environments. Hence the grain size distribution of the terrigenous sediments probably reflects that of the detritus which entered the basin. The minor contribution of clays may be a function of several factors, including a lack of chemical weathering in the source area, derivation of sediment largely from reworking of lower Burra Group sandstones, and the nature of the sediment transport processes which supplied the basin. Wind transport may have been important in the adjacent continental environments due to the lack of vegetation on the Precambrian land surface. Hence plains adjacent to the basin may have contained extensive dune-fields, from which sediment was subsequently transported into the basin. If streams carrying a wider range of grain sizes from more mountainous areas, did not flow directly into the basin, but fanned out on alluvial plains adjacent to the basin, then fine sediment could have been largely trapped in these areas.

Within the basin, coarse-grained silt and sand were reworked by wave processes. The small wavelength of symmetrical ripple marks in the southern Flinders Ranges (average of 3-4.6 cm in different areas), indicates wave generation in a shallow body of water, and not in the open ocean (Tanner, 1971), and hence is consistent with deposition in shallow lakes or lagoons. The average wavelength of ripple marks in the northern Flinders and Willouran Ranges where grain size is finer, is similar (4-4.3 cm). Studies of ripple marks in silt to very fine-grained sand, are less common than for coarser grain sizes, and hence Tanner's conclusion may not apply. However the same wavelength for finer grained sand should indicate larger waves, deeper water, or greater fetch (Tanner, 1971; Komar, 1974). Sedimentary structures indicate that waves would have been generated in water of similar depth in both areas. Hence topography in the northern area may have been more uniform over larger distances, thus the available wave fetch was greater.

Palaeocurrent data for this facies association is limited, due to the deposition by wave processes resulting in the formation of predominantly flat laminated sandstones with few cross-beds. Ripple marks are dominantly symmetrical, and have crests oriented approximately north-northeast to south-southwest in all areas where data is availabile (Fig. 6.15), except the Willouran Ranges (Fig. 6.16). Although ripple marks in shoaling areas such as shelf or near-shore zones adjacent to the open ocean, may have crestlines oriented approximately parallel to the shoreline (de Raaf *et al.*, 1977; Potter and Pettijohn, 1977), the prevailing wind direction is likely to be the dominant factor in shallow basins with very low palaeoslopes (Stricklen and Amsbury, 1974; Clemmenson, 1978, 1979). In such basins, shorelines would not have been regular, approximately linear features, and would have been of "insignificant geomorphological significance" (Clemmenson, 1979, p. 768). Hence wave ripples were probably oriented perpendicular to the prevailing wind direction, in this case from the east and/or west. However in the Willouran Ranges, the ripple crestlines approximately parallel the facies changes across this area, although some are parallel to the orientations in other areas (Fig. 6.16). The facies changes in this area result from the transition from environments subjected to periodic, and sometimes prolonged exposure in the southwest, to environments which were largely submergent in the northeast. Hence water depth may have at times varied in a systematic, although gradual fashion, so that basin paleobathymetry controlled the orientation of ripple crestlines. However in very shallow areas, with minimal gradients, the prevailing wind direction was the controlling factor.

Palaeocurrent data from cross-beds is limited to Depot Creek and Yacka only (Fig. 6.15). At Yacka, the decrease in sand content in northerly and easterly directions, suggests that the northeasterly and southwesterly modes may have resulted from onshore and offshore migration of megaripples and sandwaves in wave formed bars. The east-southeast mode may reflect longshore movement, related to the prevailing wind direction, as it is consistent with ripple mark orientation. Cross-bed orientations at Depot Creek are also consistent with ripple mark orientations.

CHAPTER 7

TERRIGENOUS CLASTICS : PETROLOGY AND GEOCHEMISTRY

PETROLOGY OF SANDSTONES

The grain size, texture, and major cement types of sandstones within each formation, have been summarised in the previous chapters. The variation in these properties, and sandstone composition, between different formations,and in different outcrops of the same formation, will now be considered.

TEXTURE

The contrast in grain size of sandstones within the Yadlamalka Formation between the northern and southern Flinders Ranges, was noted in Chapter 6. This contrast in grain size is also a feature of the lower Mundallio Subgroup, although sandstone interbeds near the base of the subgroup in the northern Flinders Ranges, are largely fine to coarse-grained. Hence in the northern Flinders Ranges, sandstones within the Mundallio Subgroup, are predominantly very fine-grained. Elsewhere, sandstones are commonly medium-grained, although the grain size is variable.

Grain size depends on the size range of the available material, and the current strength in the local environment (Folk, 1974a). Within the northern Flinders Ranges, currents were at times capable of transporting coarser sediment, as indicated by occasional interbeds of medium to coarse-grained sand within finer sand, and thick beds of intraclastic magnesite with clasts of granule size and coarser. Hence the grain size may largely reflect that of the available sediment supply, and indicate more subdued topography in the source areas which supplied the basin in the area of the northern Flinders Ranges (Forbes, 1961), or more distant source areas.

Very fine-grained sandstones in the northern Flinders Ranges are largely well sorted throughout the subgroup. However mixing of medium- to coarse-grained sand, with very fine-grained sand in a single bed, produces less well sorted sandstones with a coarse tail, or a bimodal distribution.

Sandstones with such grain size distributions, also commonly contain rounded sand and granule size, dolomite and magnesite intraclasts. These deposits resulted from the introduction of coarser sand and minor intraclasts into environments of very fine-grained sand accumulation, during short lived, high energy events.

Fine- to medium-grained sandstones in the southern Flinders and Mt. Lofty Ranges are moderately to well sorted. Those which are moderately sorted contain sand with a wide range of grain sizes, coarse-grained silt to very coarse-grained sand, with associated dolomite intraclasts. The sorting reflects rapid deposition and insignificant reworking, which would have separated the different grain sizes, and eliminated dolomite intraclasts by abrasion. Well sorted sandstones were deposited in wave agitated environments, with greater reworking.

The formation of quartz cement as optically continuous overgrowths and corrosion of grains by dolomite cement, has frequently obscured grain rounding. However where this property is apparent, coarse-grained silt and fine-grained sand grains are sub-angular to rounded. Medium-grained and coarser sand grains are sub-rounded to well-rounded. Rounded to well rounded grains suggest that the sand experienced several depositional cycles before final accumulation. The recycling probably included an aeolian episode (Pettijohn *et al.*, 1973, p. 224; Siedlecka, 1978). Poorly sorted muddy sandstones in Unit 1 of the Nankabunyana Formation at Arkaroola, contain angular and subangular sand, which may have been derived direct from the basement source, without recycling.

COMPOSITION

Sandstones within the Mundallio Subgroup are largely subarkoses. Arkoses and quartz arenites are less common (Fig. 7.2). In the northern Flinders Ranges, finer grained sand is often more feldspathic than coarser when the two components are present in a single sample. This feature has been observed in other feldspathic sandstones (Lobo and Osborne, 1976; Odum *et al.*, 1976), and indicates prolonged abrasion, selective sorting or different sources. Sandstones in the southern Flinders Ranges are more feldspathic than those to the north (Forbes, 1960; Fig. 7.2), and also have variable compositions between different areas. This variability in sandstones deposited in essentially the same environments, may indicate that the more feldspathic sandstones experienced less recycling, or had a contribution from a primary source.

Within all grain sizes, quartz is dominated by monocrystalline grains. Strained and unstrained grains are commonly equally abundant, but the proportion of strained grains increases in areas with greater deformation, suggesting that the undulose extinction of many grains is post-depositional. The dominance of monocrystalline quartz suggests prolonged abrasion. Mineral inclusions within quartz grains are rare, and include micas, tourmaline, zircon and rutile. These could have been derived from a primary source dominated by plutonic, acidic igneous or metamorphic rocks (Folk, 1974a; Lobo and Osborne, 1976).

Feldspar includes orthoclase, microcline and plagioclase (albite). The latter is both twinned and untwinned. The relative abundance of the three types varies between different areas. In the northern Flinders Ranges, orthoclase, and less commonly plagioclase, are the most abundant varieties. Forbes (1960) indicated that plagioclase was most abundant at Copley, Arkaroola and Witchelina (eastern Willouran Ranges). In the southern Flinders and Mt. Lofty Ranges, microcline is invariably the most abundant feldspar, with plagioclase least common. However at Yednalue, plagioclase is often most abundant, suggesting that the sand supplied to this area came from a different source from that supplying the area to the west (Yacka-Port Germein Gorge-Emeroo Range). Arkoses and subarkoses usually have K-feldspar (orthoclase and microcline) as the dominant feldspar, due to derivation from a granitic source, and the greater stability of K-feldspar (Pettijohn *et al.*, 1973; Folk, 1974a). If plagioclase is more abundant, this suggests some contribution from intermediate (e.g. quartz diorite), or even more basic rocks (Pettijohn *et al.*, 1973; Folk, 1974a).

Orthoclase is invariably the most weathered feldspar, with a brownish, turbid appearance. When overgrowths are present, these are often fresh and enclose weathered grains, indicating that weathering was a pre-depositional feature. Some sandstones contain fresh microcline in association with slightly weathered grains, and likewise plagioclase may show variable weathering or serecitization in a single sample. This reflects recycling of sediments, or the presence of both close and distant primary source rocks (Plummer, 1978b).

Lithic fragments are a minor component (Fig. 7.2), and are generally present only in medium-grained and coarser sandstones. They are largely granitic and gneissic fragments, but chert, mica schist, and acid volcanics (common at Arkaroola), are also present. In the southern areas, the very coarse-grained sand to granule size fraction, although minor, consists very largely of monomineralic grains. This indicates a coarse grained source area, or prolonged abrasion eliminating polymineralic fragments.

Muscovite is the only detrital mica present, except for minor biotite and chlorite at Arkaroola, and it is present only as an accessory mineral. In coarse-grained siltstones and very fine-grained sandstones of the Yadlamalka Formation in the northern Flinders Ranges, muscovite is very minor. Micas will be easily lost in aeolian environments (Pettijohn *et al.*, 1973). Thus, the minor quantity of muscovite in these sandstones supports the suggestion that they were derived from aeolian deposits.

Heavy minerals, present in only trace amounts, include tourmaline, zircon, opaques and rutile. Tourmaline, the most abundant, is largely blue and green coloured varieties (schorl). These were probably derived from acid igneous plutonic, pegmatitic, or high grade metamorphic sources (Krynine, 1946; Plummer, 1978b). Yellow and brown tourmalines (dravites), characteristic of low grade metamorphic terrains (Plummer, 1978b), are less common. Colourless, or pale-coloured, authigenic overgrowths are common on tourmaline grains. Very rare grains have rounded overgrowths indicative of recycled sediments. Within some samples, tourmaline grains vary from subangular to rounded, suggesting mixed sources, or variable recycling of the source sediment. Tourmaline is more abundant in some quartz-cemented sandstones than in dolomite-cemented sandstones of the same grain size, indicating more prolonged reworking of the former in the depositional environment, and thus greater concentration of tourmaline.

Zircon grains are also subangular to rounded, and are colourless in thin section. However zircon colour is generally only observable in whole grains (Blatt, *et al.*, 1972). Forbes (1960) states that sandstones in the south contain pink zircons, while those in the north contain colourless zircons. Opaques are largely authigenic in origin, and include pyrite and iron oxides. They have a euhedral form, or occur as aggregates of small grains. Detrital grains are rare.

The composition of sandstones indicates a primary source of acid igneous plutonic, gneissic and other high grade metamorphic rocks. However the primary source which supplied the northern Flinders Ranges and Yednalue, also included more intermediate and basic rocks with abundant plagioclase. At Arkaroola, acid volcanics, probably similar to porphyritic-acid volcanics now preserved in the Mt. Painter Inlier and on the Curnamona Cratonic Nucleus (Giles and Teale, 1979) were also a major source. Compositional and textural evidence supports a multicyclic origin for most sandstones, and this probably included an aeolian episode.

CEMENTATION

Sandstones within the Mundallio Subgroup have both dolomite (most common) and quartz cements (Chapters 4, 5 and 6). Dolomite cement occurs as equigranular mosaics of finely to coarsely crystalline dolomite, with the grain size of the dolomite often similar to that of the detrital sand grains. Within the lower Mundallio Subgroup, the dolomite is often ferroan. Dolomite-cemented sandstones usually contain minor quartz cement, which appears to have preceded the dolomite cement, as indicated by the replacement of quartz by dolomite. Feldspar overgrowths, developed most commonly on orthoclase, also preceded dolomite cement. The feldspar overgrowths generally completely enclose grains, and often have subhedral forms, indicating growth in an uncompacted or only slightly compacted sand, without significant overburden (Stablein and Dapples, 1977). The corrosion and replacement of detrital grains by dolomite cement often partly destroyed the grain supported fabric. Quartz and feldspar overgrowths may have been derived from interstitial solutions, trapped during, or soon after deposition. However with increasing burial, these were replaced by carbonate saturated solutions derived from the interbedded dolomite sediments.

Quartz cements usually formed as optically continuous overgrowths on detrital quartz grains. The shape of the detrital grain is often not apparent, unless defined by a zone of inclusions. Where detrital grains are discernible, pressure solution between detrital grains is uncommon, indicating that quartz cementation was initiated before significant compaction. In areas of higher metamorphic grade and greater deformation, quartz overgrowths may be recrystallise to finer polygonal mosaic quartz, or more sutured boundaries have developed. Overgrowths on feldspar grains most commonly developed on orthoclase grains as in dolomitic sandstones, and often completely enclose grains. Quartz-cemented sandstones often contain minor dolomite cement, and may grade into sandstones in which dolomite and quartz cements occur in approximately equal abundance. Dolomite cement formed after quartz cement, and replaces detrital grains and quartz cement. With increasing burial, and hence increasing temperature, silica becomes more soluble, and carbonate less soluble (Sharma, 1965) and this accounts for the replacement of quartz cement by dolomite. Quartz-cemented sandstones occur largely in carbonate poor sequences, and reflect the lower carbonate content of interstitial solutions.

METAMORPHIC EFFECTS

In all areas, the detrital fabric of sandstones is clearly recognizable, and recrystallisation has been minor, although metamorphic talc flakes are common in dolomitic sandstones. However at Arkaroola, increasing metamorphic grade northward from upper greenschist to amphibolite facies, resulted in the development of new mineral assemblages in dolomitic sandstones, while quartz-cemented sandstones were significantly recrystallised. Dolomitic sandstones within the Nankabunyana Formation north of Wywyana Creek were metamorphosed to assemblages of actinolite (tremolite) - scapolite or clinopyroxene - scapolite, with minor sphene, and some remnant dolomite and quartz. Poikilitic scapolite porphyroblasts developed in dolomitic sandstones of the Yadlamalka Formation, from Wywyana Creek northwards. North of the Needles, they are replaced by calc-silicate assemblages, as in the Nankabunyana Formation. The composition of scapolite changes northwards with increasing metamorphic grade, from varieties with Na much greater than Ca, to varieties with similar Na and Ca contents. This change is a feature of scapolite in regionally metamorphosed terranes (Kwak, 1977). Whole rock analysis of calc-silicates have Na_2O/K_2O ratios greater than one (G. Teale, pers. comm., 1978) indicating that plagioclase was more abundant than K-feldspars.

SHALES AND SILTSTONES

MINERALOGY

The major part of the study area has experienced low grade regional metamorphism to the chlorite zone of the greenschist facies. However the outcrops in the Mt. Lofty Ranges have experienced metamorphism to biotite zone, and at Arkaroola, to biotite zone and lower amphibolite facies. Hence the aluminosilicate mineralogy of shales and siltstones, is a chlorite-muscovite assemblage, which in the absence of other pigments such as iron oxides or carbonaceous matter, results in the characteristic greenish-grey colour of many shales and siltstones. Although these minerals are of metamorphic origin, they occur as small flakes parallel to the lamination. In areas near fold hinges, or with small scale folding, these flakes may be rotated parallel to the axial plane. Shales and siltstones also contain a silt to fine-grained sand fraction of quartz with minor feldspar. Its composition is the same as that of the equivalent grain size in sandstones.

In the Arkaroola area, shales and siltstones contain a biotite-muscovite-chlorite assemblage, although the mineralogy may vary between adjacent laminae and beds, reflecting primary compositional variations. Biotite occurs as irregular grains rather than elongate flakes, and rarely defines a schistose fabric. Hence it may have developed subsequent to the main folding. Muscovite and chlorite form elongate flakes, coarser than those in lower grade areas, but parallel to the lamination. Coarser, randomly oriented flakes of chlorite also occur. Scapolite porphyroblasts (Na rich varieties) developed in siltstones of the Nankabunyana Formation north of Arkaroola H.S. They occur as millimetre sized, poikilitic grains, concentrated in particular laminae, reflecting compositional variation in the primary sediment. The main controlling factor may be the distribution of chlorine (Hietanan, 1967; Ramsay and Davidson, 1970). Cordierite has developed in siltstones and shales of appropriate composition in the Nankabunyana Formation north of the Needles, and marks the transition from greenschist to amphibolite facies (Winkler, 1976). However cordierite is uncommon, as shales are not generally sufficiently aluminous.

GEOCHEMISTRY

As indicated above, shales within the Mundallio Subgroup have a simple mineralogy, and it represents a metamorphic assemblage. Hence major element geochemistry of shales was also studied, in an attempt to provide further evidence of source rocks, the intensity and nature of weathering processes, and the major diagenetic effects. Shales within the Mundallio Subgroup are invariably weathered to some degree, particularly in the Mt. Lofty Ranges, and on the western margin of the Flinders Ranges (Port Germein Gorge to Depot Creek, Copley to the Willouran Ranges). Hence the sample group is small (32 samples). Results are given in Table 7.1, and methods of determining whole rock compositions in Appendix 3.

The factors influencing the composition of fine grained sedimentary rocks have been listed by Spencer *et al.* (1968) as:

(1) composition of source rock;

(2) the environmental conditions at the site of weathering;

(3) nature of the transportation process;

(4) environmental conditions at the site of weathering;

(5) nature and activity of the biomass at the site of deposition;

(6) tectonic and volcanic events during weathering, transportation and the depositional cycle;

and (7) diagenetic factors.

These factors combine to produce the resultant bulk chemistry, and the major problem is distinguishing the effects, if any, of each.

Results and Discussion

The major element composition of a variety of shales is presented in Table 7.2, along with data from this study. When the mean values of major elements of this study are compared with those of the average shale (Wedepohl, 1971) the main differences are:

(1) slightly higher SiO_2 content for the Mundallio Subgroup samples, probably due to more free quartz;

(2) lower Fe_2O_3 and higher MgO content;

and (3) higher K_2O content, and slightly higher Na_2O content, although Na_2O values are very variable (Table 7.1).

The last two differences are also apparent if these samples are compared with other shales from the Adelaide Geosyncline (Table 7.2).

A Fe_2O_3-MgO-TiO_2 plot (Fig. 7.3A), indicates the high MgO/Fe_2O_3 ratio of many of the samples, as compared with other shales and average igneous rocks. Igneous rocks generally have a MgO/Fe_2O_3 ratio of less than one. Analyses of igneous and metamorphic rocks of the Gawler Craton (Bradley, 1972; Flook, 1975; Fanning, 1975; Pedlar, 1976), and the Gawler Range Volcanics (Giles, 1977), indicate a MgO/Fe_2O_3 ratio of less than one. During weathering, iron is enriched with respect to magnesium (Garrels and MacKensie, 1971), and hence detritus derived from a basement of igneous and metamorphic source rocks would be expected to have ratios greater than one.

A CaO-Na_2O-K_2O plot (Fig. 7.3B) indicates the predominance of K_2O over Na_2O, and the non-dolomitic nature of most shales. Hence the high MgO values do not result from the presence of dolomite. The K_2O/Na_2O ratio reflects the dominance of muscovite (probably derived from illite) and orthoclase, over plagioclase.

Correlation coefficients are given in Table 7.3. The marked effect which the presence of dolomite has on the content of other elements, accounts for many of the correlations in Table 7.3A. Hence only the correlations indicated in Table 7.3B, which were calculated excluding three dolomitic shales, will be discussed. SiO_2 is negatively correlated with Al_2O_3, Fe_2O_3, CaO and K_2O, reflecting the inverse relationship between the content of free quartz and clay minerals. The negative correlation of MgO and Na_2O with the detritally influenced Al_2O_3 and K_2O respectively, suggests that MgO and Na_2O contents were not related to the amount of fine grained detritus entering the basin.

The above relationships are also emphasized in a "Principal Component Analysis"[1]. If the complete sample group is considered, the first eigen vector accounts for 85.3% of the total variance (Table 7.4). It represents the ratio of $(SiO_2 + Al_2O_3)/(MgO + CaO + LOI)$, and hence reflects the dolomite content. If the three dolomitic samples are again excluded, a more meaningful result is obtained (Table 7.5). The first eigen vector represents the ratio of free quartz (SiO_2) to finer detritus ($Al_2O_3 + K_2O + Fe_2O_3$). The second represents the ratio of MgO and Na_2O to the oxides SiO_2, Al_2O_3 and K_2O. As the latter three reflect both the clays and silt supplied to the basin, that is the detrital influence on the composition of the sediment, this suggests that MgO and Na_2O were not controlled by the composition of sediment entering the basin, but by other factors. A plot of the first and second transformed variables determined in the "Principal Component Analysis" (Fig. 7.4), partly distinguishes shales from Unit 3 of the Nathaltee Formation and Unit 4 of the Nankabunyana Formation, from the rest of the samples, due to their generally lower MgO contents. This separation is also apparent in Fig. 7.3A.

The composition of shales may be used to measure their maturity, or the amount of weathering in the source area. Ronov and Khlebnikova (1957) have attempted to group shales from three main environments, based on whole rock composition (Table 7.2). These three groups are:

(1) Clays of marine reservoirs and those of salt lakes of arid regions. This group is distinguished from the others by higher MgO, CaO and Fe_2O_3, with K_2O approximately the same as continental clays of the cold belt, and Al_2O_3, TiO_2 and Na_2O intermediate between the other types. This group is close to (2) and also to magmatic rocks (Fig. 7.5).

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1. The "Principal Component Analysis" was performed using the computer programme PRINCA, of Fitzgerald (1975).

(2) Continental clays of the cold and temperate belts. Their average composition most closely approximates that of magmatic rocks. They are distinguished by a higher percentage of SiO_2, Na_2O and K_2O, and lower Al_2O_3, than the other groups.

(3) Continental clays of the hot and humid belts have very high Al_2O_3 and TiO_2, low SiO_2 and very low Fe_2O_3, MgO, CaO, Na_2O and K_2O.

The distribution of these three groups in a triangular plot of $SiO_2 + K_2O$ - $CaO + MgO + Fe_2O_3 + Na_2O + LOI$ - $Al_2O_3 + TiO_2$, is shown in Figure 7.5. Shales from the Mundallio Subgroup plot in the area of overlap of (1) and (2), although almost entirely with group (2). This indicates that the detritus was not subject to extreme weathering conditions prior to deposition.

Another method used to measure the maturity of detrital sediments is the Vogt Index. This index is given by

$$VI = \text{mole \%} \ \frac{(Al_2O_3 + K_2O)}{(MgO+CaO+Na_2O)} ,$$

and measures the residual character, and hence the amount of weathering of the sediments prior to deposition. The concentrations of MgO, CaO and Na_2O decrease regularly with increasing residual character, whereas K_2O and Al_2O_3 increase (Vogt, 1927, in Sumartojo, 1974). The index represents the ratio of illite or muscovite plus orthoclase, to chlorite plus plagioclase (Bjorlykke, 1974b). Chlorite and plagioclase will be eliminated under subtropical and tropical weathering, and/or a low relief terrain (Bjorlykke, 1974a, b). The index will also be influenced by provenance.

The mean Vogt Index for these samples is quite low, although the values for shales in Unit 3 of the Nathaltee Formation, and Unit 4 of the Nankabunyana Formation, are higher than other samples (Table 7.6). The variation in Vogt Indices, and the low values of many samples, are due largely to the MgO values, and to a lesser extent Na_2O (Table 7.6). The low values of Vogt Indices suggest that these shales are very immature, however as previously indicated the MgO and Na_2O content may be influenced by factors other than the composition of the detritus entering the basin.

Chlorite and muscovite (probably formed from illite) are the major aluminosilicates in these samples, and variation in the MgO and K_2O contents reflect variation in the abundance of these two minerals. The moderate Fe_2O_3 contents reflect the absence of iron oxides, and the minor content of sulphides in most samples, so that iron is present largely in aluminosilicates, probably chlorite. Bjorlykke (1974a, b) suggests that the contents of illite (and hence muscovite) and chlorite will be controlled by weathering in the source area and provenance. Chlorite will be produced in significant quantities with either rapid erosion, slow rates of weathering, or a major contribution from basic and ultrabasic rocks. Because most of the source areas were probably dominated by more acidic rocks, the factor most likely to contribute to preferred formation of chlorite, is a slow rate of weathering in the source area. However the enrichment of MgO with respect to Fe_2O_3 over what would be expected from the available source rocks, and most igneous rock types, the relationship of MgO to other elements, and the apparent facies control on MgO, suggests that there was postdepositional enrichment of MgO, and this resulted in an increased chlorite content in the sediments. Hence the Vogt Index for these samples may be a measure of the amount of diagenetic alteration of the composition of shales, as it reverses the trend caused by weathering. The variation in Na_2O, which partially parallels that of MgO, may be due to similar alteration.

Mg^{2+} within the porewaters of fine grained marine sediments, may show depletion compared with seawater (Drever, 1971; Skolkovitz, 1973). Mechanisms involving iron-magnesium exchange within sediments have been invoked to explain this depletion (Drever, 1971; Skolkovitz, 1973; Heller-Kallai and Rozenson, 1978). Iron may be lost from lattice sites in clays (Drever, 1971), or from iron oxides absorbed on clays (Skolkovitz, 1973) under reducing conditions, and be replaced by magnesium. The iron may precipitate as sulphides in the presence of S^{2-}, and thus be retained in the system. Heller-Kallai and Rozenson (1978) propose an alternative mechanism. Removal of iron causes partial breakdown of clays, and magnesium silicates then form by reaction of the degraded clays with interstitial waters. It is possible that the iron lost from clays may migrate out of the system, thus lowering the iron content of the sediments. Braunagel and Stanley (1977), in a study of non-marine variegated redbeds, indicated that green mudrocks of red-green couplets, have a lower iron content (Fe = 1.94%, or 2.77% Fe_2O_3), than red mudrocks (Fe = 4.72%, or 6.75% Fe_2O_3). This was attributed to postdepositional migration of iron from the green layers, and precipitation in an oxidizing environment.

The magnesium enrichment in these shales is attributed to introduction of Mg during diagenesis, coupled with a loss in iron. Samples from Unit 3 of the Nathaltee Formation, and Unit 4 of the Nankabunyana Formation, which have lower MgO contents than other samples, also have slightly higher Fe_2O_3 contents. However the MgO - Fe_2O_3 correlation, although negative, is not significant (Table 7.3B). Hence some iron may have been completely lost from shales, and possibly precipitated within interbedded dolomites. Dolomite interbeds within the Nathaltee and Nankabunyana Formations have an average of about 7000 ppm Fe (1% Fe_2O_3), which precipitated in dolomites largely during diagenesis[1].

The significant enrichment in magnesium during diagenesis may reflect a concentration of Mg^{2+} in interstitial waters greater than that of seawater, as normal marine shales (Table 7.2, Ronov and Khlebnikova, 1957) have lower magnesium contents. The lower values in massive shales at the top of the Nathaltee and Nankabunyana Formations may reflect deposition from waters

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1. This is discussed in Chapter 10.

of lower magnesium concentration, either due to a different composition, or lower salinity.

Hence the Vogt Index is unsuitable as a measure of maturity, but may provide an indication of the amount of diagenetic alteration which the sediments have experienced. However the maximum values of this index, which represent shales with least magnesium enrichment (i.e. Group 4 in Table 7.6), may provide some measure of maturity. The values for this group are higher than those of the Sturt Tillite (Sumartojo, 1974), suggesting that they experienced greater weathering in the source area. This is supported by higher Al_2O_3 and K_2O values (Table 7.2). The low CaO contents of these shales also suggest that weathering was sufficient to significantly lower the content of mobile elements.

Another maturity index has been proposed by Dennen and Moore (1971). Si is plotted against Al/Fe, where Si + Al + Fe = 100%. They suggest that normal chemical weathering rapidly generates a constant Al/Fe ratio of 1.9 ± 0.4. Deviations from this occur in immature, or supermature sediments which have high Si or Al values due to extended chemical weathering (Fig. 7.6). Other shales plotted in Figure 7.6 fall in the mature range, as do the Sturt Tillite and Tindelpina Shale from the Adelaide Geosyncline (Sumartojo, 1974). However Mundallio Subgroup shales exhibit a wide scatter of Al/Fe values, with a mean Al/Fe = 3.4. This variation is due to a low and variable Fe_2O_3 content, and a relatively constant Al_2O_3. This is not due to the shales being supermature, as higher Al_2O_3 (Al) values would be expected, and this would also not be compatible with the subarkosic and arkosic composition of interbedded sandstones. Hence the variability, and high Al/Fe ratios, support the previous suggestion that iron was removed during diagenesis.

Summary

Non-dolomitic shales from the Mundallio Subgroup differ in some respects from other shales described in the literature. The higher MgO and lower Fe_2O_3 values are attributed to substantial magnesium enrichment during diagenesis, associated with the removal of iron, due to the presence of magnesium rich, interstitial solutions. Their compositions indicate deposition in marine or other saline basins, or as continental clays of the cold and temperate belt. However the magnesium enrichment indicates the presence of saline solutions, thus supporting the first alternative. Because of the influence of diagenesis on MgO, Fe_2O_3, and possibly also Na_2O contents, it is difficult to apply maturity indices which will indicate the amount of weathering in the source area. However the maximum values of the Vogt Index, the fact that the Al_2O_3 content is comparable with other shales, the high K_2O contents, and low CaO contents, suggest that the detritus which entered the basin had experienced significant chemical differentiation.

CHAPTER 8

SUMMARY : DEPOSITIONAL HISTORY OF THE MUNDALLIO SUBGROUP

INTRODUCTION

The Mundallio Subgroup outcrops in two distinct areas, (1) the northern Flinders Ranges and Willouran Ranges, and (2) the southern Flinders Ranges and Mt. Lofty Ranges. These two areas are separated by the central Flinders Ranges, in which both the Mundallio Subgroup and the Burra Group are absent. Within the first of these two areas, the stratigraphy of the Mundallio Subgroup is relatively simple, with a lower clastic dominated interval and an upper carbonate dominated interval. However in the second area, the stratigraphy is more complex, with inter-tonguing between carbonate and clastic dominated sequences. The palaeo-geographies of these two regions during deposition of the Mundallio Subgroup are described separately below.

NORTHERN FLINDERS AND WILLOURAN RANGES

Apart from the vertical facies change from clastic-dominated to carbonate-dominated sequences, the only major facies changes present within this area occur from southwest to northeast across the Willouran Ranges. Parallel facies changes are also present in the underlying Witchelina Subgroup in this area (Murrell, 1977). There are no time markers within this sequence, hence palaeogeographic maps representing a single time plane cannot be produced. Instead generalised distributions of environments for different parts of the sequence are presented.

PRE-MUNDALLIO SUBGROUP

Deposition of sandstones with some shale intervals, generally preceded the deposition of the Mundallio Subgroup. Within the Willouran Ranges, the underlying Witchelina Subgroup consists of an initially transgressive, fining upward and basinward sequence of sandstones and shales, which is overlain by a regressive sandstone facies, the Copley Quartzite (Murrell, 1977). The Copley Quartzite appears to have accumulated as thick, deltaic sandstone wedges, which prograded into the basin from the southwest and

and northeast margins. Finer sandstones and shales were deposited in the centre of the basin in this area (the Willawalpa Formation, Murrell, 1977). In the Copley-Myrtle Springs area, the Copley Quartzite, the base of which is not exposed, also underlies the Mundallio Subgroup. In the Arkaroola region, the Wortupa Quartzite which underlies the Mundallio Subgroup, is also a sandstone dominated sequence, with minor shales and conglomerates. Detailed sedimentological and palaeoenvironmental studies of these sandstone sequences have not been made. Desiccation features and halite casts have been recorded (Coats, 1973), suggesting a shallow water environment, however the transportational and depositional processes, and the nature of the depositional basin, are not known. These sequences record a widespread, and relatively continuous influx of sand-size detritus into the basin.

LOWER MUNDALLIO SUBGROUP : CLASTIC DEPOSITION

Deposition of the Mundallio Subgroup commenced as the influx of sand to the basin of deposition was diminished. Lowering of gradients in the source area, and thus a reduced stream carrying capacity, or a more arid climate with associated reduced denudation rates and sediment supply, may have caused the reduction in sand supply. In the western areas (Willouran Ranges, Copley-Myrtle Springs), the base of the subgroup may also reflect a transgressive event. This was most marked in the eastern Willouran Ranges, where poorly laminated shales, devoid of coarser sediment (Camel Flat Shale), were deposited below wave base in an environment of negligible current reworking. However in the central Willouran Ranges, the environment periodically shallowed above wave base, allowing rare deposition of sandstones in the Camel Flat Shale. In the area from the southwest Willouran Ranges to Copley, sandstone interbeds deposited under wave-agitated conditions as sand bars and sheets, are interbedded with shales deposited below wave base (Unit 1 of the Nankabunyana Formation). These shales are indistinctly laminated and massive in the Willouran Ranges, whereas at Copley and Myrtle Springs they are generally well laminated. The presence of interbedded sandstones and massive shales representing more offshore environments below wave base, indicates that in some areas there was a very abrupt transition from environments of wave agitation, to those of very limited wave and current reworking. In marine environments the zone below wave base is subject to periodic agitation, and hence laminated silts and muds are deposited. However in lakes, or water bodies of limited size, the zone below wave base will be quiet and undisturbed (Picard and High, 1972). The area of the transition zone from agitated to quiet, undisturbed environments will be small, although this will also depend on the depositional slope. Hence the presence of interbedded massive shales and sandstones, indicates wave generation in a basin of limited size.

In the Arkaroola region, sand deposition was also replaced by deposition of mud and silt with lesser sand. However this was not associated with a transgression, possibly due to a reduced rate of subsidence in this area. An alluvial plain in the south, on which poorly sorted siltstones and sandstones, and minor channel conglomerates were deposited, passed northward into shallow submergent mudflats. The lack of a permanent wave-agitated shoreline or beach on the margin of the basin, indicates low palaeoslopes, probably lower than in western areas. The environments in the lowermost Mundallio Subgroup, are summarised in Figure 8.1.

The transition to Unit 2 of the Nankabunyana Formation and the Tilterana Sandstone, represents a shallowing of the basin in the western areas, and a change in the nature of sand-size detritus supplied to the basin. Medium-grained sand, previously predominant, was subordinate, and sandstones were now largely very fine-grained. In much of the Willouran Ranges, the basin was infilled above wave base, and very fine-grained sandstones were deposited under the influence of wave processes (Tilterana Sandstone).

This continued for the remainder of the lower Mundallio Subgroup (Figs. 8.2-4). Elsewhere interbedded sandstones and shales of Unit 2 of the Nankabunyana Formation, were deposited in very shallow environments subject to periodic exposure. Mudflats formed in areas of limited winnowing and high sediment influx. Sandflats formed in areas in which winnowing was more effective, either further from the source, or during periods of reduced sediment influx. The mud and sand were introduced by rivers flowing directly into the basin, but this built out as mudflats rather than deltas, because of the very shallow nature of the basin, low depositional slopes, and low wave energy. Lagoons developed on the landward side of extensive mudflats, sandflats, or sand bars, and were sites of deposition of dolomite mud. These lagoons were fringed by exposed dolomite mudflats, and stromatolites grew around the margin. Well rounded coarse-grained sand deposited between stromatolite columns, or with dolomite intraclasts eroded from the exposed dolomite mudflats, may have been derived from local sources, such as aeolian deposits reworked from older Burra Group sandstones landward of the lagoons. Deposition of dolomite mudstones was most significant in the southern part of the Arkaroola region. Dolomite deposition in this area was followed by prolonged exposure, non-deposition and subaerial weathering on at least two occasions, before dolomite deposition was renewed.

Dolomite facies become less common through Unit 2, and are minor in Unit 3 of the Nankabunyana Formation. Likewise features indicative of exposure in the clastic facies also become more infrequent, indicating a slightly deeper environment in the areas of outcrop, possibly due to an expansion of the basin (Fig. 8.3).

Continued deepening in most areas, except the central and eastern Willouran Ranges, resulted in the widespread deposition of poorly laminated shales (Unit 4, Nankabunyana Formation) below wave base in more offshore environments (Fig. 8.4). However gradual infilling of the basin above wave base, resulted in the deposition of sandstones in most areas. At Arkaroola, shales and sandstones similar to Unit 3 were deposited in the upper part of Unit 4, except in the north (Blue Mine Creek) and east of the Paralana Fault, where only sandstones were deposited. Minor dolomite mud and intraclasts deposited in sandstones at the top of Unit 4 (Copley, Myrtle Springs, Top Mount Bore), may have been derived from erosion of dolomite mudflats in more landward areas where deposition of the Yadlamalka Formation had commenced. The presence of shales deposited in quiet, undisturbed environments below wave base, and their gradation upward into sandstones deposited by wave processes, again suggest wave generation in a water body of limited size, so that there was a fairly abrupt transition from undisturbed to wave agitated environments.

Because of limited palaeocurrent data, and the presence of environments of similar water depth in different areas over the same period of time, determination of the precise orientation of the shoreline, and its approximate position during deposition of the lower part of the Mundallio Subgroup, is difficult. At Arkaroola, the presence of continental facies in the south, where there is also evidence for prolonged periods of exposure, suggest a southerly or southeasterly margin in this area. However the northern area (Blue Mine Creek) received a localised sand influx in the uppermost part of the Nankabunyana Formation, suggesting source areas to the north or east. A north-south shoreline may have existed west of the Copley-Myrtle Springs area, as indicated by an increase in the abundance of dolomitic facies in Unit 2 towards Myrtle Springs, and palaeocurrent data. In the Willouran Ranges, sediment was supplied from both the southwest and northeast. The latter source area may have been in the vicinity of the "Muloorina Ridge", and continued eastward to the Arkaroola area. Hence the basin in this area probably had an arcuate shape, although the shorelines were unlikely to have been simple linear features. The northwesterly limb of this basin may have continued into the area of the Peake and Denison Ranges, where similar, although sandier facies, were deposited in the lower part of Mundallio Subgroup (Ambrose and Flint, 1979).

UPPER MUNDALLIO SUBGROUP

Deposition of the upper Mundallio Subgroup was associated with significant deposition of dolomite, and in many areas, magnesite. This occurred as the basin became shallower, and probably reduced in size. The single, persistent water body, with some marginal dolomite lagoons, was probably replaced by several shallow and extensive lakes, separated and enclosed by exposed mudflats in which there were smaller ephemeral lakes. The size and position of these lakes varied significantly through time, as did the energy levels of the depositional processes within them.

The Yadlamalka Formation was deposited over much of this area (Fig. 8.5). The major facies present is dolomite mudstone, deposited on shallow, but largely submergent mudflats in areas of low wave energy. However as water level fell, or shorelines retreated, mudflats were extensively exposed, and smaller ephemeral lakes formed in which deposition of magnesite occurred. The area of outcrop from the southwestern Willouran Ranges to Copley, and the area to the southwest and west of these outcrops, often became a series of isolated, ephemeral lakes in which magnesite was deposited. At Arkaroola, ephemeral magnesite lakes were a less common feature, but occurred more frequently to the east of the present outcrops. With a continued fall in water level, areas of magnesite deposition were subjected to prolonged periods of exposure, during which magnesite was lithified, and partly replaced by nodular magnesite. The lithified magnesite was extensively eroded during high energy events associated with rising water level and expansion of lake margins. In the southwestern Willouran Ranges to Copley area, much of the magnesite was eroded and reworked as intraclastic beds. In the Arkaroola area, centimetre sized intraclasts were transported westward into the present outcrop area. Erosion of intraclastic magnesite was commonly associated with the introduction of sand into the basin, especially in the Copley and Myrtle Springs area. The sand was reworked by waves on shallow sandflats, or in wave formed bars.

In the Willouran Ranges, the environment of deposition became more persistently submergent to the northeast (West Rischbieth-Coronation Bore area), and exposed magnesite mudflats were infrequently developed. Intraclastic magnesite in this area was derived largely from shallower areas to the southwest. In the central and eastern Willouran Ranges, in areas of deposition of the Mirra Formation, magnesite was rarely deposited. Deposition in the Willouran Ranges occurred in three major sub-basins, separated by areas of minimal subsidence on which thin sequences were deposited (Fig. 8.6). These areas, now sites of outcrop of Callana Group, were at times exposed, so that the basin in this area consisted of two or three separated water bodies. However periods of more complete circulation also occurred. In the southwesterly sub-basin, deposition of the Yadlamalka Formation occurred as described above. In the other sub-basins, sandstones and dolomite mudstones of the Mirra Formation were deposited in largely submergent environments. In the central sub-basin, siltstones and shales were also deposited in the slightly deeper, central areas.

In the lower Mundallio Subgroup, low energy environments were sites of terrigenous mud deposition, whereas carbonate mudstones predominated in the upper part of the subgroup. However coarse-grained silt and very fine-grained sand were still supplied to most areas of the basin. Although the grain size of sand is slightly finer grained than in the lower part of the subgroup, the mineralogy is similar, suggesting a similar source. Hence the boundary between the clastic and carbonate dominated sequences also represents a marked change in the ratio of sand plus coarse-grained silt, to clay, in the sediment supplied to the basin. Streams and rivers carrying mud, silt and sand, probably flowed directly into the basin during deposition of the lower subgroup. However as the basin decreased in size, and shallow ephemeral lakes and exposed mudflats developed around the margin, permanent rivers probably no longer flowed into the basin. Stream capacity in the adjacent hinterland may also have been reduced. The coarse-grained silt and sand deposited in the upper part of the subgroup, were derived largely from aeolian deposits adjacent to the basin. These were apparently widespread as indicated by the presence of sand in all areas. However the most persistent sand influx was from northeast of the Willouran Ranges.

Within the basin, sand was redistributed by waves, the orientation of which was largely determined by the prevailing wind direction, although in the area of the Willouran Ranges, palaeobathymetry may at times have been the controlling factor.

POST-MUNDALLIO SUBGROUP

The Yadlamalka and Mirra Formations are overlain by the Myrtle Springs Formation, a sequence of sandstones, siltstones and dolomites, somewhat similar to the Nankabunyana Formation. This sequence was deposited due to an expansion of the basin, and a slight increase in water depth, with the subsequent elimination of extensive exposed mudflats and ephemeral lakes. A renewed supply of terrigenous mud to the basin, in addition to sand, resulted in low energy environments again becoming sites of deposition of terrigenous muds.

SOUTHERN FLINDERS AND MT. LOFTY RANGES

The stratigraphy of the Mundallio Subgroup is more complex in this area, and contains more significant lateral facies changes. The correlations within this area, as summarised in Chapter 3 (Figs. 3.1, 2 and 3), are based on the lithologies present within both the Mundallio Subgroup and the overlying and underlying formations. Outcrop in many of these areas is poor, and there are no time markers within the sequence. Hence the correlations made, and on which the palaeogeographic reconstructions are based, may not be the only ones possible.

PRE-MUNDALLIO SUBGROUP

On the western margin of this area, from Adelaide to the Emeroo Range, deposition of sandstones, sometimes with conglomerates and shales, preceded the Mundallio Subgroup. In more central areas deposition of sandstones was accompanied by significant deposition of finer clastics, and some dolomites. Studies of these sequences in isolated areas (Port Germein Gorge, McCarthy, 1974; Depot Creek, Preiss and Sweet, 1967; and the Yednalue Anticline, New, 1978), indicates deposition as submergent to occasionally emergent sand bodies. The major sand source occurred west of the basin, although local sources were probably present to the east. For example, minor gravels, sand and finer clastics were introduced into the Olary and Weekeroo regions.

LOWER MUNDALLIO SUBGROUP

A reduced sand influx, as in the northern area, marks the beginning of deposition of the Mundallio Subgroup. In most areas this does not appear to be associated with significant changes in water depth, however the average energy levels in the depositional environment were reduced. Within the southern Flinders Ranges, in areas in which the terrigenous detritus supplied was predominantly of clay and silt size, deposition of the Nathaltee Formation occurred. In other areas, the terrigenous influx was minor, except at Yacka, and was dominantly sand sized. This sand was deposited along with dolomite and minor magnesite, as the Yadlamalka Formation.

Deposition of the Nathaltee Formation occurred in an approximately linear zone along the western margin of the southern Flinders Ranges, with the exception of the Mundallio Creek area (Fig. 8.6). Initially grey dolomites and shales, with minor sandstones, were deposited in the Emeroo Range and south of Port Pirie (Unit 1). In the Port Germein Gorge and Beetaloo Valley area, better circulation and hence more oxidising conditions, resulted in deposition of pale coloured dolomites and shales. Deposition of these more oxygenated facies subsequently extended northward into the Emeroo Range, and westward into the Port Pirie area (Fig. 8.7). At Mundallio Creek there was little influx of fine terrigenous detritus, and deposition of the Yadlamalka Formation followed that of the Emeroo Quartzite.

Deposition of the Nathaltee Formation also occurred in the Yednalue Anticline, and possibly locally at Willow Creek. The area of deposition of grey shales with lesser dolomites and sandstones, of the Nathaltee Formation in the Yednalue Anticline, was separated from the area of deposition of the Nathaltee Formation on the western margin of the Flinders Ranges, by a shallow area consisting of emergent to submergent dolomite mudflats, with occasional isolated lakes in which magnesite deposition occurred (Yadlamalka Formation, Figs. 8.6 and 7). Circulation and transportation of sediment across this area would have been limited, suggesting that the areas of deposition of the Nathaltee Formation had separate source areas. Hence the terrigenous sediment deposited in the area of the Yednalue Anticline may have been derived from the north or northeast. Sandstones within the Mundallio Subgroup at Yednalue and in the Yednalue Anticline, have different feldspar compositions from sandstones on the western margin of the southern Flinders Ranges, also indicating a different source.

The Nathaltee Formation may have been deposited adjacent to areas into which permanent, low energy rivers transported clay and silt.

However because of the low depositional slopes and very shallow nature of much of the basin, this sediment was deposited close to the site at which it was introduced. Dolomite mudflats were developed in areas of deposition of the Nathaltee Formation along shore from river outlets, and in protected lagoons and embayments. However they became a persistent feature in areas more distal from the source, resulting in deposition of the Yadlamalka Formation. Sandstones are a minor, and rarely a significant component in that part of the Yadlamalka Formation laterally equivalent to the Nathaltee Formation. Some of this sand may have been derived from aeolian deposits adjacent to the basin, as suggested for the Yadlamalka Formation in the northern Flinders Ranges.

The uppermost part of the Nathaltee Formation contains a coarsening and shallowing upward shale-sandstone sequence, analogous to that at the top of the Nankabunyana Formation. A reduced sediment influx may have caused deepening of the basin, and deposition of poorly laminated shales below wave base. Small deltaic sand bodies subsequently prograded into the basin, where they were partially reworked by wave processes (Fig. 8.8).

In the Mount Lofty Ranges, the influx of sand was minimal. Widespread deposition of dolomite occurred in shallow areas with little influx of clay or sand (Skillogalee Dolomite, lower part of Castambul Formation, Fig. 8.6). However deposition of shales was also widespread (Woolshed Flat Shale), and subsequently replaced dolomite deposition in the Adelaide region (upper part of Castambul Formation, Figs. 8.7 and 8).

Hence the lower Mundallio Subgroup is characterised by variable facies within this region. This was due to variable sediment influx, both in abundance and grain size, variable basin topography and water depth, and variable energy levels. Palaeoslopes within the basin did not have a regular gradient in any particular direction, but were more irregular, particularly in the southern Flinders Ranges, where deposition of shallow water facies was widespread.

UPPER MUNDALLIO SUBGROUP

As in the northern Flinders Ranges, low energy environments now became almost exclusively sites of carbonate mud accumulation, with the exception of parts of the Mt. Lofty Ranges where deposition of shales continued, probably in slightly deeper environments than those in which dolomite deposition was predominant (Fig. 8.9). However the shales and siltstones within the Woolshed Flat Shale are well laminated, indicating that the areas in which it accumulated were subject to weak current or wave activity. In the southern Flinders Ranges deposition of dolomite mudstones alternated with deposition of sandstones and magnesite facies (Yadlamalka Formation), while in the northern Mt. Lofty Ranges, only dolomites were deposited (Skillogalee Dolomite, Fig. 8.9). The Yadlamalka Formation accumulated in similar environments as those described for the northern Flinders Ranges. However, although deposition of magnesite was less frequent with the exception of the Depot Creek area, it occurred in all areas of outcrop of the Yadlamalka Formation, and also rarely within the Skillogalee Dolomite. Facies similar to the Yadlamalka Formation were also locally deposited in the Adelaide region (Montacute Dolomite).

Sand sized detritus was supplied in greatest abundance on the western margin of the basin. Within the basin the sand was reworked by wave processes, however transportation of sand into the basin away from source areas may have been limited. For example the very sandy sequences of the Yadlamalka Formation in the Yacka area, are now only about 4 km distant from outcrops of the Skillogalee Dolomite near Spalding, in which sandstones are not present (Fig. 3.6). These two sequences may however have been slightly further apart prior to folding and faulting. Hence those areas in which sandstones are abundant, were probably close to source. Transport of sand over longer distances may have occurred when aeolian dunes migrated over exposed mudflats, and were subsequently reworked when the area experienced a transgression.

The area of shale deposition in the south expanded in the uppermost part of the subgroup, and replaced areas of dolomite deposition in the Mt. Lofty Ranges (Fig. 8.10). Subsequently, there was renewed progradation of sand from the west, which halted carbonate deposition in the southern Flinders Ranges, and shale deposition in the Mt. Lofty Ranges (Fig. 8.11). However in the eastern half of the southern Flinders Ranges, these sandstones (Undalya Quartzite), are very thin or absent. Carbonate deposition in this area was replaced by deposition of shales, probably winnowed from areas of sandstone deposition. This change was associated with a deepening of the basin, which had now developed a regular east-west asymmetry of facies and probably water depth, a feature which was not present during deposition of the Mundallio Subgroup.

CENTRAL FLINDERS RANGES

As the Burra Group does not outcrop in the central Flinders Ranges, Rutland and Murrell (1975) proposed that it was not deposited in this area. The Umberatana Group and younger sediments overlie and are cut by zones of disrupted Willouran sediments. During the Sturtian and Marinoan these disrupted sediments were periodically exposed, and shed detritus into the adjacent regions (Dalgarno and Johnson, 1966; Plummer, 1978b). In the northernmost part of the southern Flinders Ranges, and the Arkaroola area, there was extensive, and sometimes complete erosion of the Burra Group prior to Sturtian deposition. Hence the lack of occurrence of the Burra Group in the central Flinders Ranges may be due to pre-Sturtian erosion in this area.

The alternative, that of non-deposition, is difficult to prove from the facies present within the Mundallio Subgroup. Throughout the Mundallio Subgroup, and in particular in the upper part, the lack of a regular arrangement of facies in a lateral sense, with the shallowest facies confined to the basin margin and passing offshore into deeper water facies in the central areas of the basin, makes it difficult to precisely define the basin margin, or determine areas which are approaching it. In addition, due to the very low depositional slopes within the basin, the extent of submergent and exposed areas varied greatly during deposition of the subgroup. At Copley, the abundance of magnesite and sandstone in the upper part of the sequence suggests that the basin margin may have been close to this area at that time. However in the lower part of the subgroup, it was probably further to the west, and apparently had a north-south orientation. In some of the northernmost outcrops of the southern Flinders Ranges, there is evidence of uplift and exposure, and in some cases extensive penecontemporeneous erosion, during deposition of the Mundallio Subgroup (Yednalue, Yednalue Anticline and Willow Creek, Fig. 8.9), which may support the existence of an exposed area in the vicinity of the central Flinders Ranges. However the Mundallio Subgroup in these three areas is thick (700-1500 m), and the uplift and exposure was not associated with the influx of sediment, other than that derived from the erosion of previously deposited Mundallio Subgroup sediments. Hence this uplift may have been a localised phenomenom. However sediment was supplied to the Yednalue Anticline and Yednalue areas, from a landmass to the northeast during deposition of parts of the subgroup. This may have been part of a continental area extending from east of Arkaroola to the Olary region (Curnamona Cratonic Nucleus).

The similarity in carbonate mineralogy and facies in outcrops of the Mundallio Subgroup in the northern and southern Flinders Ranges, suggests that these areas were part of a single depositional system, with similar chemical and hydrological regimes. Hence subsidence and deposition probably occurred in the area of the central Flinders Ranges, as it is unlikely to have been an exposed, positive topographic feature which permanently separated two basins. Uplift and erosion subsequent to Burra Group sedimentation removed the record of deposition in this area.

THE BASIN OF DEPOSITION

The Mundallio Subgroup was deposited in a linear basin with a dominantly north-south orientation. In the vicinity of the Willouran Ranges, this changed to a southeast-northwest trend, which may have continued into the Peake and Denison Ranges (Preiss, 1973). A continental land mass existed on the western margin of the basin. There is also evidence for a continental land mass to the east and northeast of the basin, extending from the Olary-Broken Hill region, to east of the Arkaroola area, to the area northeast of the Willouran Ranges. Similarities in the stratigraphy and deformational history of basement of the Gawler Block and that in the Olary-Broken Hill region (Glenn *et al.*, 1977), suggest that cratonic basement was continuous beneath the basin. The nature and extent of the basin, and the existence of continental areas east and south of the Mt. Lofty Ranges, are not known. However a positive gravity anomaly on the western margin of the Murray Basin, extending from near Broken Hill to east of Murray Bridge, has been attributed by McInerney (1974) to a basement ridge. This may have been an extension of the continental areas in the Olary-Broken Hill region, and formed a margin to the basin in this area, locally supplying sand to the basin (e.g. Scrubby Range).

Hence the basin of deposition was a shallow intracratonic depression. It may have been the shallow arm of a larger marine basin, or a completely continental feature such as an inland sea or large lake. The main depocentre was in the northern Flinders and Willouran Ranges, where in excess of 1000 m of sediment accumulated, although this area was probably also the shallowest part of the basin. Because of the shallow nature of of the basin, the extent of water filled areas varied considerably, especially during deposition of the Yadlamalka Formation. Exposed carbonate mudflats were a common feature in much of the basin. Clastic and carbonate sediments were deposited in varying proportions in different areas and through time. Carbonate sediments were predominantly dolomitic, however magnesite deposition was also widespread, as has been previously indicated by Forbes (1961).

The persistently shallow character of this basin, and its elongate shape, with continental landmasses to the west and east, would have caused effective damping of astronomical tidal currents, even if there was a connection with a larger marine basin in the south. Although Klein (1977) and Klein and Ryer (1978) suggest that ancient epeiric seas had significant tidal effects, this requires that they had an extensive and unrestricted connection to the open ocean. Sedimentary structures within sandstones include ripple marks, ladder ripple marks, desiccation cracks, ripple cross-lamination, flaser lamination, cross-bedding, and intraclastic lenses, features which are common in modern tidal environments, but not exclusive to them. All these features have also been observed in lacustrine environments (Eugster and Hardie, 1975; Clemmenson, 1978; Link and Osborne, 1978). They are rather an indication of an environment of fluctuating water levels, in which the fluctuations may be caused by wind tides, storms, or seasonal climatic variations, as well as tidal currents (Laporte, 1975). Sandstones of similar grain size within the Mundallio Subgroup contain similar sedimentary structures, all of which may be attributed to meteorologically induced wave processes.

PART III

CARBONATE MINERALS IN THE MUNDALLIO SUBGROUP : ORIGIN AND DIAGENESIS

CHAPTER 9

DOLOMITE AND MAGNESITE MUDSTONES : TEXTURES AND EARLY DIAGENETIC HISTORY

INTRODUCTION

The textures of carbonate facies within the Mundallio Subgroup have been described in Chapters 4, 5 and 6. Carbonate deposition occurred largely as mudstones, although these were often subject to extensive erosion and reworking (e.g. intraclastic magnesite). The textures of fine grained carbonate facies (dolomite mudstones, peloidal dolomites, magnesite mudstones and nodular magnesite), as observable with both the petrographic microscope and the scanning electron microscope (SEM), are considered in more detail below. Determination of depositional and early diagenetic processes from observations of textures, may provide clues to the timing of formation of the existing carbonate minerals, dolomite and magnesite, and the nature of any precursor minerals. Within the Mundallio Subgroup, magnesite and dolomite were deposited in close association. Examples of this include mudstones with alternating dolomite and magnesite laminae, laminae and thin beds of dolomite mudstone within intraclastic magnesite, beds containing both dolomite and magnesite intraclasts, magnesite intraclasts within a dolomite matrix, and magnesite intraclasts as nuclei within dolomitic oncoids and ooids. This suggests that dolomite and magnesite, or minerals with very similar composition, existed within surface sediments.

DOLOMITE MUDSTONES

INTRODUCTION

Dolomite mudstones within the Nathaltee, Nankabunyana, Yadlamalka and Mirra Formations, are largely dolomicrosparites (microspar is 5-15μ, Folk, 1959), with the exception of those in the Arkaroola region where regional metamorphism has resulted in recrystallisation to coarser grain sizes. However mudstones within the lower Mundallio Subgroup (Nathaltee and Nankabunyana Formations) are generally slightly coarser grained (often coarse microspar), than those of the upper Mundallio Subgroup (Yadlamalka and Mirra Formations), which are micrites and fine microspars. Within the Castambul Formation, and the Skillogalee and Montacute Dolomites, dolomite mudstones have often been recrystallised to grain sizes coarser than microspar.

The origin of microspar in limestones has been extensively discussed by Folk (1965) and Bathurst (1974), with some specific examples by Longman (1977), Longman and Mench (1978) and Steinen (1978). Folk (1965) indicates that microspar is a neomorphic product of micrite, and forms by replacement in the solid state, but with the aid of interstitial solutions. Although this was supported by Bathurst (1974), he cautioned that it would be premature to regard all microspars as being of neomorphic origin. Folk's evidence for neomorphic origin is as follows (as summarised in Bathurst, 1974, pp. 513-514).

"(1) Allochems float in three dimensions in the microspar so that it cannot be a cement.
(2) Though having a uniform crystal size in any small area, microspar commonly passes by gradual reduction of crystal size, into micrite.
(3) Microspar is commonly concentrated around allochems in an otherwise micritic matrix.
(4) Some microspar adjacent to allochems has a radial fibrous fabric.
(5) Some faecal pellets embedded in microspar, have been replaced by identical microspar so that the only remaining evidence of their existence, is an elliptical, brown organic stain."

In addition, Folk (1965, p. 40), states

"if microspar were a carbonate silt, it should follow bedding planes, occur as lenses, cross beds, etc. Different grain sizes of microspar should be sorted into layers, not grade centifugally in grain size from the centre of a patch outward."

Hence Folk (1965) maintains that microspar forms by coalescive neomorphism of micrite. However if microspar also forms by lithification of a carbonate silt, then other processes may be involved, similar to those operative in the lithification of micrite. This includes solution of small grains, the formation of overgrowths, as well as the replacement of smaller grains by larger (Folk, 1965). Precipitation into open spaces, created or enlarged by dissolution, may be important in some cases (Steinen, 1978).

In limestones, the micrite to microspar transition is enhanced by the removal of Mg ions, trapped interstitially within the micrite during early diagenesis (Folk, 1974b; Longman, 1977). The Mg ions may be removed by flushing the system with fresher water, or by their adsorption onto clay minerals. In contrast, saline interstitial solutions probably inhibit the formation of microspar (Longman and Mench, 1978). Porosity may also assist the formation of microspar which continues in a subsurface environment long after deposition (Longman, 1977). Removal of saline waters may enhance the conversion of dolomicrite to dolomicrospar, but it is less likely that the removal of Mg ions would be significant as is the case for calcite.

The similarities in textures of calcareous microspar and dolomicrospar, suggest that similar processes are operative in their formation. Longman and Mench (1978) describe dolomicrospar produced by the development of overgrowths on micritic dolomite rhombs, which formed by the replacement of aragonite in the supratidal zone (Longman and Mench, 1978). In the supratidal zones of the Persian Gulf and Andros Island, micritic dolomite rhombs have incompletely replaced pelloidal aragonitic muds, but the sediment is still porous (Illing *et al.*, 1965, Shinn *et al.*, 1965). Dolomitization tends to destroy the pelloidal texture of these muds, producing a more homogeneous micrite or microspar. With complete lithification and dolomitization, a mosaic of micrite or microspar with anhedral grains would form, and contain little evidence of the early formed dolomite rhombs.

The texture of a compact, non-porous micrite or microspar, may have little resemblance to the primary carbonate mud. Euhedral grains will be eliminated as porosity is reduced, and dissolution will remove the smallest grains. However the texture may record the diagenetic history, with variations in texture recording different diagenetic histories (Longman and Mench, 1978).

DESCRIPTION OF TEXTURES AND THEIR SIGNIFICANCE

Mineralogy of dolomite mudstones has been confirmed by staining of thin sections using the method of Dickson (1965). XRD analysis has been used to distinguish dolomite and magnesite. These methods confirm that dolomite is the only carbonate mineral within dolomite mudstones, apart from extremely rare samples in which calcite has partly replaced dolomite. X-ray diffraction patterns show that the dolomite is well ordered, and its composition close to being stiochiometric.

Dolomite mudstones are well laminated, and the grainsize of the microspar often varies between adjacent laminae, reflecting grain size differences in the original mud (Plate 9.1a). Graded laminae contain coarse microspar or finely crystalline dolomite at the base, sometimes with scattered terrigenous silt-sized grains, passing upwards into finer microspar (Plate 6.2a). The variation in grain size between adjacent laminae may be preserved when the dolomite has been recrystallised to material coarser than microspar.

Most dolomite mudstones contain a minor component of terrigenous silt and sand, as scattered grains in homogeneous microspar or concentrated in laminae. In the latter case, the enclosing dolomite exhibits a wider range of grain sizes. In laminae which are not grain supported, this dolomite includes microspar and finely crystalline material which often partly replaces quartz grains. In grain supported laminae, the dolomite is largely finely crystalline, and partly replacing quartz grains. This finely crystalline dolomite probably formed at the same time as dolomite cement in interbedded sandstones, and may have originated partly as a cement, and partly as a neomorphic replacement of a pre-existing dolomite matrix. However microspar within silty laminae may have been originally deposited as silt sized grains.

Clay mineral flakes are often too fine grained to be visible with the petrographic microscope, but may be clearly observed with the SEM. Irregular single flakes, 0.5 to 5μ (Plate 9.2a), or irregular flaky aggregates (Plate 9.2b), are located along dolomite grain boundaries. The grain size of dolomite within a single sample, is similar in clay rich and clay poor areas.

Grey and dark-grey dolomite mudstones contain carbonaceous material, largely concentrated along grain boundaries, and sometimes obscuring them. The carbonaceous content may vary between adjacent laminae, although the grain size in these laminae is often similar. In mudstones with clotted or grumous textures, the carbonaceous material is concentrated in the clots (Plate 6.2b). These textures are only distinctive in very dark grey, carbonaceous dolomite mudstones, and are obscured where weathering has caused oxidation of the carbonaceous material. Hence clotted textures may have been more abundant than they now appear. In mudstones with clotted textures, the carbonaceous stained clots are finer grained (micrite to finer microspar) than the intervening clear areas (finely crystalline spar). This may be due to carbonaceous material inhibiting recrystallisation, although this was not the case in mudstones with laminae of more homogeneous microspar. The present grain size distribution probably reflects the primary grain size distribution, in which the clear, coarser dolomite originated at least in part, as a cement. Some areas of clear dolomite show an increase in grain size toward the centre, or rarely rhombs growing towards the centre which has been infilled with chert. This supports the origin of the clear dolomite as a cement. Clotted areas probably originated as a mixture of mud and peloids, the individuality of which was destroyed during lithification and compaction (Bathurst, 1974). In dolomites with a clearer peloidal texture, the grain size distribution is similar to that in mudstones with clotted textures (Plate 9.1b), and the boundaries of peloids may be quite sharp. Hence the depositional texture, and primary grain size distribution in these sediments have been well preserved.

Carbonate diagenesis involving mineral inversion of aragonite to calcite, does not preserve grain for grain detail, even when this occurs via thin solution films, although general features of the depositional texture may be preserved (Sandberg, 1975; Pingitore, 1976). Hence it is unlikely that the inversion of aragonite or low Mg-calcite to dolomite, would preserve the primary grain size distribution, as has occurred in laminated dolomite mudstones, and those with clotted and peloidal textures. In present day environments in which dolomite is forming from aragonite after deposition, fine details of texture are destroyed (Illing *et al.*, 1965; Shinn *et al.*, 1965).

Because of the fine grain size of dolomite mudstones, their textures have also been examined with the SEM. Grain size has been estimated from SEM photographs, with magnifications of 1000 to 3000 times. Average grain size and sorting, using the sorting parameter of Folk and Ward (1957), were determined on small areas of some samples. These values are given in Table 9.1, and some representative grain size distributions in Figure 9.1. As most samples are inhomogeneous due to variation in grain size between laminae or clotted fabrics, determination of these properties in different areas of the sample would give slightly different results. It is apparent from Table 9.1 that most samples are well sorted. This is a characteristic feature of micrite and microspar (Folk, 1965), and is the result of competition between space and growth (Flugel *et al.*, 1968). The sorting is not a reflection of hydrodynamic conditions as in clastic sediments.

Examination of fracture surfaces of microspar with the SEM, indicates that it forms compact mosaics. In those samples with a grain size of micrite to fine microspar, the individual grains are equant to subequant, regular polyhedral blocks. They are anhedral, and only rarely have rhombohedral boundaries (Plate 9.2c and d), or fracture parallel to cleavage planes. Grain surfaces are plane to slightly curved, hence grain boundaries are simple, either straight or gently curvilinear. Some samples contain largely straight boundaries (Plate 9.2c), but most samples contain a mixture of both.

With increasing grain size to coarse microspar, the grains become more irregular, although they are still approximately equant to subequant (Plate 9.2e and f). They also exhibit a greater tendency to fracture across grains, or parallel to cleavage, rather than along grain boundaries. The tendency for coarser carbonates to fracture along cleavage steps has been previously illustrated by Harvey (1966).

The compact, non-porous fabric of the microspar, indicates that it is a secondary fabric. The main mechanisms involved in the formation of this fabric may have included aggrading neomorphism, in which finer grains are replaced by coarser (Folk, 1965; Bathurst, 1974), and syntaxial rim-cementation producing an interlocking grain mosaic. The variation in grain size between adjacent laminae and within graded laminae, and the presence of coarser lenticular laminae, indicate that the laminae within the primary mudstones consisted of sediment of different grain sizes. Preservation of this grain size distribution during lithification, suggests that rim cementation, similar to that described by Longman and Mench (1978), has been significant.

Angular grains indicate the importance during diagenesis, of overgrowth mechanisms (rim cementation), which resulted in crystal interlocking (Flugel *et al.*, 1968; Mimran, 1977). Rim cementation should also produce plane intercrystalline boundaries (Bathurst, 1974). Neomorphic fabrics may be characterised by curved or wavy boundaries, although *in situ* neomorphic growth at crystal-solution film interfaces, may produce plane boundaries (Bathurst, 1974). Sparry dolomites formed during metamorphism will tend to have planar boundaries (Bathurst, 1974). Most microspars contain a significant proportion of straight boundaries, supporting rim cementation as a major process of lithification. The absence of sutured or amoeboid boundaries within microspars, indicates that pressure solution has not been important. This, and the rarity of even fine stylolites, are due to significant lithification of dolomite mudstones at shallow depths (Scholle, 1977).

Hence the existing fabrics in dolomicrites and fine dolomicrospars, may be attributed largely to early lithification, which involved rim cementation, dissolution of small grains, and probably to a lesser extent, neomorphism. This preserved the primary grain size distribution of the sediment, and probably substantially reduced its porosity. Hence, although these processes probably continued during late diagenesis and low grade regional metamorphism, their subsequent effect on substantially lithified sediments was minor. However in coarser microspars, the increase in grain size and alteration of texture, producing more irregular grains, has been greater during late diagenesis and metamorphism.

The degree of early lithification may have influenced the amount of recrystallisation during metamorphism in higher grade areas. For example, in the Skillogalee Dolomite, grey dolomite mudstones are still largely dolomicrosparites and finely recrystallised dolomites. These may have been more significantly lithified during early diagenesis, than the closely associated pale-coloured dolomites, which have been more coarsely recrystallised during metamorphism.

MAGNESITE MUDSTONES AND NODULAR MAGNESITE

DESCRIPTION OF TEXTURES AND THEIR SIGNIFICANCE

Laminated magnesite mudstones and nodular magnesite are both largely micritic, as are intraclasts derived from these facies. Hence magnesite facies are finer grained than the interbedded dolomite mudstones. Average grain sizes determined from SEM photos are micritic, and may vary within a single sample. Hence in petrographic thin sections, the magnesite appears aphanitic with individual grains barely visible. Coarser magnesite is found only in the Arkaroola and Torrens Gorge areas, where it has been recrystallised to microspar and finely crystalline magnesite due to higher the metamorphic grade.

Grain size variation between laminae in magnesite mudstones is minor, but some have a streaky or patchy texture (Plate 9.1c), in which darker areas are slightly finer grained than lighter. The coarser fraction, which has a grain size of coarse micrite to fine microspar, appears to have formed by replacement of the finer micrite. Sorting values given in Table 9.2, indicate that magnesite micrites are less well sorted than dolomite microspars. This is due to the magnesite consisting of a mixture of grain sizes from fine micrite to fine microspar, as in Plate 9.3a, in which the distribution appears bimodal. Some samples contain scattered micrite grains, less than 1μ in size, or aggregates of grains of this size, enclosed within coarser micrite and fine microspar. Less commonly the micrite is well sorted, and texturally resembles the fine dolomicrospars (Plate 9.3b).

The fine micrite occurs as rhombs, platy rhombs, or more rounded grains (Plates 9.3a, 9.3c). Areas with grains largely of this size, tend to have a slightly porous texture (Plate 9.3c). Replacement of the fine micrite may have produced the coarse micrite, which occurs as polyhedral grains, or less commonly as rhombs, or irregular grain which have fractured parallel to the cleavage. Because of the range of grain sizes present, the texture is more irregular than that of fine dolomicrospars (Plate 9.3a), and often appears slightly porous. However samples lacking grains of fine micrite, form more compact mosaics of polyhedral grains (Plate 9.3b), similar to dolomicrospars. Grains have largely plane faces, others have gently curved faces.

Nodular magnesites are also micritic. Fracture surfaces examined with the SEM have an apparently slightly porous mosaic of anhedral and minor rhombic grains (Plate 9.3d). Examination of polished and etched surfaces, indicates that these micritic mosaics contain areas of aligned, elongate grains (Plate 9.3e and f). The direction of elongation varies within a single sample (Plate 9.3e), indicating that this is not a deformational feature. The grains are largely anhedral, but elongate step like rhombs are also present (Plate 9.3e and f). The elongation direction parallels the long axis of the rhombs. Boundaries between the adjacent grains are largely planar. It is possible that adjacent, parallel grains have similar crystallographic orientations, resulting in the areas observable in thin section in which aggregates of micritic grains have similar birefringence. This is most readily apparent when the grains have their C-axis approximately normal to the thin section, resulting in small areas with low birefringence. Areas in which micritic grains have similar orientation, may correspond to the small primary nodules, less than a millimetre in size, which coalesce to form the larger nodules replacing magnesite mudstones (Chapter 6).

Micron sized grains of magnesite, some of which were rhombohedral, may have formed by replacement of a primary magnesium carbonate mud (with minor dolomite or protodolomite), resulting in a semi-lithified sediment.

Plate 9.3c may be a relic of this very early diagenetic sediment. This fine micritic mud experienced further diagenesis to form coarser micrite, but this appears to have been incomplete in some cases, resulting in mudstones which are a mixture of fine and coarse micrite (Plate 9.3a). This process may have involved dissolution of the finest grains, which if incomplete, may leave rounded micron and minimicron (Folk, 1974b) sized grains (Plate 9.3c). Rim cementation, and neomorphism in which aggregates of micron sized grains are replaced by single coarser grains, were probably also involved. When the formation of coarse micrite was complete, this produced well lithified compact mosaics (Plate 9.3b). The more irregular texture of some magnesite mudstones, their wider range of grain sizes than dolomicrospars, and relics of very early diagenetic sediment, indicates that they are texturally more immature than dolomicrospars. Hence they have probably experienced an even more limited diagenetic history, due to substantial lithification in near surface environments.

Nodular magnesite may also have formed a well lithified micritic fabric during early diagenesis. In fact this probably developed prior to erosion, as fabrics in eroded and uneroded nodular magnesite are identical. It is unlikely that the mosaics of elongate, aligned, both anhedral and rhombic grains, would have formed by coalescive neomorphism of a finer grained fabric, in which coarser grains develop from nuclei scattered throughout the primary sediment as described by Folk (1965). This would produce a mosaic of more equant, largely anhedral grains (Bathurst, 1974). Outward growth from an initial point, or set of points, along the direction of elongation is more likely. Growth was porphyritic, but individual porphyroblasts developing from scattered points, were not single grains, but aggregates of micritic grains, which grew by the addition of further similar grains. probably in similar orientations. The grain size of nodular magnesite is largely coarse micrite, hence it may have largely formed after the development of micron sized rhombs in magnesite mudstones during very early diagenesis (Fig. 9.3).

The texture preserved within nodular magnesite appears to be primary, suggesting that the nodules originally formed as magnesite, rather than as other metastable magnesium carbonate minerals. The fine grain size of some magnesite mudstones, and the presence of micron sized rhombs, indicates that magnesite was also present in this facies during very early diagenesis.

CHAPTER 10

GEOCHEMISTRY OF CARBONATE FACIES

INTRODUCTION

The abundance of trace elements within the lattice of carbonate minerals appears to be both primary and diagenetic facies controlled, as indicated by numerous geochemical studies (Veevers, 1969; Veizer and Demovic, 1974; Fruth and Scherreiks, 1975; Rao and Naqvi, 1977; Veizer, 1977a and b; Veizer *et al.*, 1977, 1978). Hence geochemical studies may provide further means of delineating facies variations. This is particularly important in unfossiliferous Precambrian sequences, in which fossil evidence distinguishing non-marine, normal marine, and hypersaline environments, is lacking.

Chemical analyses are most appropriate for samples containing one carbonate component with a single origin. In carbonate grainstones, allochems and cements have different origins, and hence different chemical compositions. Thus bulk rock analyses of multi-component systems will produce an average of the chemical composition of the different components (Veizer *et al.*, 1978). However the majority of the samples analysed in this study are dolomicrosparites, which contain only one carbonate mineral, and one textural component. Only the carbonate component of the sample, and not the total sample including the insoluble residue, has been analysed. The analytical procedures and the complete set of analytical results are given in Appendices 3 and 4 respectively. All elemental analyses are calculated on total carbonate (insoluble residue free) basis. In the subsequent discussion, only the means of sample groups, based on facies and location (Tables 10.1 and 2), are considered, rather than individual samples.

IRON AND MANGANESE

INTRODUCTION

The chemistry of iron and manganese are closely analogous (Krauskopf, 1967; Hem, 1972), and hence these two elements are considered jointly. The main controls on the Fe and Mn phases present in solution, or forming as precipitates, are the Eh and pH of the solutions, and the concentration of the other ions present (Krauskopf, 1967). From his Figure 9.2, it is apparent that Fe^{2+} ions may be transported in solution for long distances, if the solution remains slightly acidic and reducing. However in more alkaline conditions, and in the presence of carbonate, silicate or sulphide ions, precipitation of iron minerals occurs. In seawater, iron precipitates as oxides if the water is aerated, as silicates or carbonates if it is mildly reducing, or as sulphides if the redox potential is low and sulphur abundant.

Although the solution chemistries of iron and manganese are similar in that both elements experience oxidation/reduction under the present environmental conditions at the earth's surface, and form insoluble oxy-hydroxides, sulphides, and carbonates, the conditions under which these occur are different (Callender and Bowser, 1976). Manganese is more soluble under many conditions than iron (Hem, 1972, Fig. 8). Ferrous ions are more readily oxidised than manganous ions, and ferrous sulphides are more insoluble than manganous sulphides (Hem, 1972). Hence the concentration of manganese may be increased relative to iron in reducing environments with moderate to high sulphate concentrations. The stability field of $MnCO_3$ is also larger than that for $FeCO_3$. However iron and manganese in solution form organic complexes and bicarbonate-sulphate-organic complexes respectively, and this complicates the precipitation of iron and manganese compounds.

PRECIPITATION OF IRON AND MANGANESE WITHIN CARBONATES

The coprecipitation of these elements in carbonate minerals will depend on the following factors (Bencini and Turi, 1974):

(1) supply of these elements to the basin of deposition;

(2) physicochemical conditions of the basin waters in the depositional environment;

(3) mineralogy;

(4) diagenetic processes.

In metamorphosed sediments, the metamorphic grade is also significant. Above lower greenschist facies the iron and manganese contents of carbonate minerals may be increased, as increasing reduction of Fe^{3+} to Fe^{2+} occurs (Rosenberg, 1967, 1968).

Iron and manganese are supplied in solution, and in the form of oxides, hydroxides, and clay minerals. Because of adsorption effects, there is often a correlation between clay minerals and the iron and manganese content of carbonate rocks (Fruth and Scherreiks, 1975). During humid climatic periods, the amount of iron and manganese supplied to sedimentary basins will increase, due to greater continental discharge, and a shift in factors such as pH, Eh, the content of organic matter, and hydration, in the direction of higher solubility (Veizer, 1978). The Mn/Fe ratio will also increase due the greater solubility of manganese compounds.

The amount of iron and manganese incorporated within carbonate minerals also depends on the appropriate distribution coefficients. The only coefficient that is well established is $k_{Mn}{}^{C}$, with a value of approximately 16 (Pingitore, 1978). Hence calcite in equilibrium with seawater should contain about 30 ppm Mn (Veizer, 1977a). Aragonite may have a slightly lower distribution coefficient (Pingitore, 1978; Ragland *et al.*, 1979). Quoted values for $k_{Fe}{}^{C}$ range from 1 (Richter and Fuchtbauer, 1978) to 21 (Veizer, 1974). Fe and Mn may substitute for Mg to a slightly greater extent than for Ca, hence the distribution coefficients of dolomite may be slightly greater than those of calcite.

Iron and manganese contents of carbonate sediments are commonly increased during diagenesis, and the amount of increase will depend on the Eh and pH, the composition of the interstitial pore waters, and whether the system is open or closed (Pingitore, 1978). In particular low Eh environments are favourable for the formation of Fe^{2+} ions, and their incorporation within carbonates. However the low Eh is often produced by bacterial decay and reduction of SO_4^{2-} to S^{2-} or H_2S (Richter and Fuchtbauer, 1978). Hence in marine pore fluids, which have an initially low iron content, abundant S^{2-} in reducing environments inhibits high ferrous ion concentration, and iron will not be incorporated within carbonates during early diagenesis (Richter and Fuchtbauer, 1978). However the presence of continental pore waters with a lower SO_4^{2-} concentration during diagenesis, may favour the formation of ferroan calcite. Hence the conditions favourable to the formation of ferroan calcite are a reducing environment, and a lack of organic matter combined with anaerobic sulphate reducing bacteria. The ferrous ions required could be provided from iron oxides or silicates, which are partially dissolved during diagenesis in a reducing environment (Richter and Fuchtbauer, 1978).

RESULTS

Means and standard deviations of Fe, Mn and Mn/Fe are given in Table 10.1 for each formation in each of the major outcrop areas. The data is plotted in histograms in Figures 10.1 and 2. Correlation coefficients are given in Table 10.3. Fe and Mn are strongly correlated for all formations (Table 10.3, Fig. 10.3). Although the Fe and Mn values are determined on the acid soluble carbonate component, during the acid leaching of sample preparation, some Fe and Mn could have been leached from clay minerals. In the Nathaltee and Nankabunyana Formations, there is no correlation of either Fe or Mn with IR, suggesting that any Fe and Mn derived from acid leaching of clays is a minor component as compared with that in the dolomite lattice. However for samples from the Yadlamalka and Mirra Formations, which contain significantly less Fe and Mn, there is a correlation with IR (Tables 10.1 and 3, Figs. 10.4 and 5). Hence the Fe and Mn leached from clays could form a significant component as compared with that in the dolomite lattice, although derivation of the Fe and Mn within carbonates from exchange with clays during diagenesis, could also produce this correlation. The intercepts of regression lines of plots of Fe and Mn versus IR for samples from the Yadlamalka and Mirra Formations, are 774 ppm Fe and 120 ppm Mn respectively. This gives the average minimum amount of Fe and Mn within the dolomite lattice, for the case in which the IR is dominated by clays from which the remainder of the Fe and Mn are derived. However because the IR usually contains significant quartz and feldspar, the amount within the dolomite lattice may be greater.

Fe and Mn values for the Nathaltee and Nankabunyana Formations are greater than those of the other formations. The Fe and Mn mean values of the Yadlamalka and Mirra Formations sample group are significantly different from those of both the Nathaltee and Nankabunyana Formations at greater than a 99.9% confidence level, as indicated by t-tests. The Fe mean value for the Montacute Dolomite differs from those of the Nathaltee and Nankabunyana Formations at the 98% confidence level, while Mn values differ at the 99.9% level.

The higher values for the Nathaltee and Nankabunyana Formations are probably related to the abundance of interbedded shales and siltstones (Button *in* Trusswell and Eriksson, 1975; Land *et al.*, 1975). The other formations studied contain only minor fine grained terrigenous interbeds. This relationship is supported by vertical variation in the Nathaltee Formation in the Depot Creek area. Dolomites from Unit 1 have lower Fe (4849 ppm) and Mn (817 ppm) values than Unit 2 (Fe = 7365 ppm, Mn = 1363 ppm), which is characterised by a higher proportion of interbedded shales. Fe and Mn values in the Yednalue Anticline, where shales and siltstones are more abundant than elsewhere, are higher than the mean values for the Nathaltee Formation from all areas (Table 10.1). However within the Nathaltee Formation, the iron and manganese contents do not have a systematic relationship with the reducing conditions which prevailed during early diagenesis. Darker grey dolomites have both higher (YDA) and lower values

(Unit 1, DC) than pale dolomites (Unit 2), suggesting that the availability of iron and manganese from interbedded shales is a more important factor.

Within the Yadlamalka and Mirra Formations sample group, most areas have similar mean Fe (in range 2000-3000 ppm) and Mn (in range 150-3000 ppm) values (Table 10.1), with the exception of the Arkaroola and Yacka areas. In the Arkaroola area, the higher metamorphic grades west of the Paralana Fault probably resulted in the higher values here (Rosenberg, 1967, 1968), although the IR content is also above average. East of the Paralana Fault, the extreme values (Fe = 11097, Mn = 1367) may reflect the close proximity to the pre-Sturtian unconformity, which cuts the Yadlamalka Formation in this area. Weathering prior to deposition of the Sturtian sequence, may have enriched the Yadlamalka Formation in Fe and Mn, and produced the extensively ferruginised horizons which are present immediately below the unconformity. In the Yacka area, the higher means are due to three anomalous samples (YE029, YE070, YE072, Appendix 4), which are associated with shale interbeds at the top of the formation. Within any given area, the sampling interval is generally too large to assess any significant vertical variation of Fe and Mn, with the exception of the Yednalue section (Fig. 10.6). Here low IR, Fe and Mn values in the middle of the formation, correspond to massive dolomites, deposited in an environment with little terrigenous influx, and which have experienced complex diagenetic histories (Chapter 6). At Depot Creek, Copley and Arkaroola, the highest Fe and Mn values are at the base of the Yadlamalka Formation, immediately above the shaly sequences of the lower Mundallio Subgroup. Dolomite mudstones, and intraclastic dolomites derived from them, have higher Fe and Mn values than other dolomite facies which have a lower content of terrigenous material, especially clays (Table 10.2).

Mean Fe and Mn values for magnesite samples, and the correlation of these elements with other variables, are strongly influenced by samples from Arkaroola (Tables 10.1 and 3). The Arkaroola samples have experienced a higher grade of metamorphism and have higher IR and higher dolomite contents. The first factor in particular will increase the Fe and Mn contents of the carbonate minerals (Rosenberg, 1967, 1968). Excluding the Arkaroola samples, the Fe content is similar, and the Mn content lower than dolomite samples (Table 10.1). This suggests that the relative substitution of the Mn^{2+} ion (r = 0.8Å) with respect to the Fe^{2+} ion (r = 0.74Å) for Mg^{2+} (r = 0.66Å), is less than for Ca^{2+} (r = 0.99Å), producing lower Mn/Fe ratios in magnesite than dolomite.

DISCUSSION

The iron and manganese contents of carbonate sediments may show considerable variation. The contents of the dolomite and magnesite samples in this study are comparable with those determined in several studies of Phanerozoic carbonates (Bencini and Turi, 1974; Fruth Scherreiks, 1975; Rao and Naqvi, 1977; Supko, 1977; Veizer *et al.*, 1978). However early Proterozoic carbonates commonly contain a few percent Fe and Mn (Truswell and Eriksson, 1975; McLennan *et al.*, 1979), and this has been attributed to lower atmospheric free oxygen (Veizer, 1978).

Seawater and any other alkaline aerated water would precipitate only minor amounts of Fe and Mn in carbonates, and this is shown by the low values in Recent and sub-Recent carbonates (Supko, 1977; Pingitore, 1978; Ragland *et al.*, 1979). Hence carbonates formed in marine environments or saline lakes will contain little Fe and Mn, unless conditions are reducing and there is little S^{2-} present. However the wide variation in Fe and Mn values for grey and dark grey dolomites within different formations of the Mundallio Subgroup, indicates that the availability of Fe and Mn during deposition and from interstitial and interbedded clays during diagenesis, was the major factor rather than Eh. Most Fe and Mn was probably introduced post-depositionally, and the higher values may also correspond with the more altered facies in which Fe and Mn were introduced over a longer period of time, in some cases continuing into the metamorphic realm. For example, dolomite mudstones of the Nathaltee and Nankabunyana Formations which have higher Fe and Mn contents, are slightly more recrystallised than those of the Yadlamalka and Mirra Formations. The tendency for Fe and Mn to increase during diagenesis, is due to the distribution coefficients being greater than one and possibly the higher Fe and Mn concentrations in diagenetic pore solutions than in the basin waters from which the carbonates were originally precipitated. Areas of deposition of the Nathaltee and Nankabunyana Formations were supplied by runoff waters transporting clay and silt. These runoff waters may have had higher Fe and Mn contents in solution, as fine precipitates or adsorbed on clays, than runoff and groundwaters reaching areas of largely dolomite deposition. This may also partly account for the higher Mn and Fe contents of dolomites within these two formations.

The lowest Fe and Mn values may approach those within the initial precipitates (Veizer, 1974). The lowest Mn values are in the range 30-70 ppm, which is not much different from that expected for precipitation from seawater. The lowest Fe values measured in these samples fall in the range 290-380 ppm, and are somewhat higher than that expected for seawater precipitation (approximately 70 ppm for calcite and slightly higher for dolomite). Hence the Fe content may have been higher, and the Mn/Fe ratio less in the basin waters than in present day seawater. This may reflect the influence in this basin of continental waters, which have higher Fe concentrations and lower Mn/Fe ratios (Turekian, 1972), rather than precipitation of dolomite occurring from waters of normal marine chemistry.

SODIUM

PRECIPITATION WITH CARBONATES

Analysis of the sodium content of carbonate rocks is of interest because it may provide additional information on the palaeosalinity of the depositional and diagenetic environment (Fritz and Katz, 1972; Land and Hoops, 1973; Rao and Naqvi, 1977; Veizer *et al.*, 1977, 1978). Sodium within carbonate rocks may be attributed to three main sources (Fritz and Katz, 1972):

(1) sodium introduced during crystallisation as dry or liquid inclusions;

(2) sodium exchanged during crystal growth for Ca and Mg sites within the lattice;

(3) sodium associated with mineral impurities such as clays.

Determination of Na content within the acid soluble carbonate phase will determine the first two components, and possibly part of the third due to acid leaching of clay minerals.

Precipitation of Na within the carbonate minerals aragonite, calcite and Mg-calcite, has been studied by White (1977, 1978). Incorporation of Na within the carbonate lattice depends on the aqueous activity ratio of Na^+/Ca^{2+}, and hence on salinity, as this ratio generally increases with salinity. With increasing Na^+/Ca^{2+}, precipitation of sodium bicarbonate may occur. Incorporation of Na also increases with pH. White (1978) also suggests that the presence of organic substances increases the Na concentration in calcite.

There does not appear to be significant variation in the Na content of carbonate sediments due to mineralogical variation. Land and Hoops (1973), Rao and Naqvi (1977) and Veizer *et al.* (1977, 1978) indicate that ancient limestones and dolomites have similar Na contents, or that dolomites have slightly less Na than associated limestones. Aragonite may contain slightly more Na than calcite (Ragland *et al.*, 1979).

Modern carbonate sediments contain several thousand ppm Na (Land and Hoops, 1974; White, 1978; Ragland *et al.*, 1979). However this is decreased markedly during diagenesis due to recrystallisation in the presence of pore waters that are less saline and/or less alkaline than those from which the sediment initially formed. White (1978) suggests that this decrease may also be related to the replacement of organically produced marine calcite or aragonite, by inorganically formed calcite during diagenesis.

Despite the significant reduction in Na, it appears that variations in Na content imposed by the original salinity or very early diagenetic conditions, are preserved (Fritz and Katz, 1972; Veizer *et al.*,1977, 1978). Fritz and Katz indicate that dolomites of supratidal origin (determined from petrographic evidence and from their association with evaporites), contain more Na than diagenetic dolomites, and quote values of

200-900 ppm	supratidal dolomites
70-200 ppm	"early" diagenetic dolomites
< 150 ppm	"late" diagenetic dolomites
< 100 ppm	hydrothermal dolomites.

Veizer *et al.*(1977, 1978) distinguished normal and hypersaline carbonate facies (both limestones and dolomites) in a Canadian Palaeozoic sequence, on the basis of faunal, petrographic and evaporite evidence. These facies were also distinguished on the basis of Na values, with normal marine carbonates containing less than 230 ppm average Na, and hypersaline carbonates containing greater than this value. However original variability in facies, and differences in post-depositional history causes this boundary to be diffuse, with an intermediate zone between the two facies at 150-300 ppm (Veizer *et al.*, 1978).

RESULTS AND DISCUSSION

The major feature of the sodium content of the dolomite sediments within the Mundallio Subgroup, is the general similarity of mean values both between formations, and in different areas of outcrop of the same formation. In addition the standard deviation is commonly small (Table 10.1, Fig. 10.7). Na does not show a significant correlation with other elements (Table 10.3). The lack of correlation with IR indicates that any Na leached from clay minerals during acid dissolution, is minor as compared with that leached from carbonates.

Within the Yadlamalka Formation, Na values are similar in all areas (121-186 ppm), with the exception of the Arkaroola area. Here the lower mean value west of the Paralana Fault, may be related to the higher metamorphic grade. East of the Paralana Fault, the lower metamorphic grade (lower greenschist facies) accounts for a Na value (135 ppm) comparable with other areas of outcrop of the Yadlamalka Formation. Likewise, the lower Na values in the Montacute Dolomite (66 ppm) may be due to the higher metamorphic grade which it has experienced (upper greenschist facies). However the pale recrystallised dolomites of the Skillogalee Dolomite do not have a particularly low Na content (146 ppm), although it is lower than that of the overlying grey dolomite mudstones which are less altered (242 ppm). The sodium contents of dolomites within the lower shaly sequences of the Mundallio Subgroup are greater than those of overlying Yadlamalka Formation in some areas, although the reverse situation also occurs (Table 10.1). However the differences are small.

If this data is compared with that of Veizer *et al.* (1977, 1978), it falls in the group of normal marine salinity, or intermediate between this group and that with hypersaline tendencies. The rare indications of evaporite minerals within the Mundallio Subgroup support the interpretation that these rocks did not form in environments of extreme salinity. However comparison of data from sequences from different areas, with different ages and diagenetic, metamorphic and tectonic histories, may not be valid. A salinity scale for each sequence may need to be established, using samples with associated evaporites, and others known to have formed in environments of normal salinity. For example if carbonate samples from entire Adelaidean and overlying Cambrian sequence were considered, they could be used to determine a salinity scale for normal, intermediate and hypersaline environments, and unknowns compared with it.

In summary, dolomite sediments within the different formations of the Mundallio Subgroup formed in environments of somewhat similar salinity. This may have been close to normal marine (35‰), or slightly greater. However Na measurements of other Adelaidean carbonate rocks with associated evaporites are needed to confirm this.

The sodium contents of magnesite facies within the Mundallio Subgroup are lower than those of dolomite facies (Table 10.1). However the mean values are not significantly different (t-test). From the discussion in Chapter 6, it was indicated that magnesite formed in shallow frequently desiccated lakes, in which saline pore waters probably formed during exposure. In addition magnesite is less recrystallised than many dolomites, and may have been expected to have retained a greater proportion of its coprecipitated sodium during diagenesis. These results suggest either that the average salinity in magnesite lakes was not significantly different from that in environments of dolomite formation, or that less sodium is incorporated in the magnesite lattice than in that of dolomite.

POTASSIUM

PRECIPITATION WITHIN CARBONATES

The potassium content of carbonate sediments may also indicate salinity variations during deposition and very early diagenesis (Fritz and Katz, 1972; Veizer *et al.*, 1978). White (1977), in an experimental study of the precipitation of sodium and potassium in aragonite, indicated that potassium coprecipitation is related to the aqueous activity ratio of potassium to calcium, analogous to the situation for sodium. However the amount of potassium incorporated is much less than sodium (Ragland *et al.*, 1979). Increasing aqueous activity of sodium may inhibit the coprecipitation of potassium in aragonite (White, 1977). Hence potassium may to some extent reflect salinity variations, but to a lesser extent than sodium.

Within carbonate sediments, particularly mudstones, associated aluminosilicates, illite and muscovite, contain substantial potassium. Measurements of potassium within carbonate rocks often show correlation of K with both aluminium and IR (Fritz and Katz, 1972; Veizer *et al.*, 1978). Some of the potassium measured may be derived from acid leaching of micas (Veizer *et al.*, 1978).

RESULTS AND DISCUSSION

From Table 10.1 and Figure 10.8, it is apparent that potassium values are much more variable than sodium values. K is almost invariably correlated with IR, often at better than a 99.9% confidence level (Table 10.3). The only exception is for the pale-coloured dolomites of the Skillogalee Dolomite in which the IR frequently consists largely of quartz. Elsewhere the IR contains quartz, feldspar, and aluminosilicates (muscovite, chlorite and talc). It is likely therefore that some potassium has been derived from leaching of clays, but this component cannot be precisely differentiated from that leached from carbonates. A plot of K versus IR (Fig. 10.9) has a cluster of points with K less than 300 ppm, and IR less than 20%. In these samples the K measured may largely represent that derived from dolomite.

The correlation of Fe with K in most sample groups (Table 10.3) suggests that as is the case for K, some Fe is derived from leaching of clays. However a plot of Fe versus K for all dolomite samples (Fig. 10.10), indicates that the correlation is not particularly good. Mn and K are not correlated, suggesting that little Mn has been derived from acid leaching of clays.

The potassium content is thus mainly an indication of the amount of fine grained terrigenous detritus within dolomite sediments. Dolomites within the Yadlamalka Formation generally contain less K than those of the Nathaltee and Nankabunyana Formations, reflecting their lower alumino-silicate content. Within the Yadlamalka Formation, dolomite mudstones and intraclastic dolomites derived from them, have the highest K contents (Table 10.2). The K content is similar in most areas of outcrop of the Yadlamalka Formation, with the exception of the Yacka, Yatina and Arkaroola areas. In the first two areas, this is due to a greater amount of deposition of clays in association with dolomite mud. At Arkaroola, the dolomites are often silty, and contain abundant muscovite and quartz. Variation in the K content across the Willouran Ranges (Table 10.2), reflects the greater deposition of fine clastics in the Mirra Creek-Rischbieth area than in adjacent areas. Abundant muscovite in some dolomite samples of the Montacute Dolomite accounts for the high K values in this formation. The abundance of fine clastics in these dolomites reflects the close proximity to areas of concurrent deposition of the Woolshed Flat Shale.

The potassium content of magnesite samples is similar to that of dolomites, and is also correlated with IR.

STRONTIUM

PRECIPITATION WITHIN CARBONATES

The strontium content of carbonate rocks is of particular interest because of the significant mineralogical control of Sr contents in modern carbonate sediments, and the markedly reduced Sr content of ancient carbonate sediments as compared with modern (Kinsman, 1969a; Veizer and Demovic, 1974; Morrow and Mayers, 1978). In addition, Sr is a major trace element in carbonate minerals, and the contribution from IR will be low (as compared with Fe, Mn and K).

The strontium content of carbonate sediments reflects properties of the sediment, including the mineralogy, texture, and diagenetic history. Veizer and Demovic (1974), Veizer (1977b) and Veizer *et al.* (1978) argue that primary mineralogy is the major factor controlling the Sr content of carbonate rocks, whereas Morrow and Meyers (1978) claim that groundwater chemistry during diagenesis, and porosity are most important.

Mineralogy

Different carbonate minerals have significantly different distribution coefficients (k_{Sr}), where

$$k_{Sr} = \frac{mSr^{2+}}{mCa^{2+}}\text{(solid)} \Big/ \frac{mSr^{2+}}{mCa^{2+}}\text{(liquid)}.$$

Aragonite has $k_{Sr}^{A} = 1.17$ at $16^{o}C$ (Kinsman, 1969a) and it increases with temperature, but appears to be fairly insensitive to rates of precipitation and solution composition. Aragonite precipitated in equilibrium with seawater contains approximately 9000 ppm Sr. In contrast the distribution coefficient of calcite is significantly less than 1, $k_{Sr}^{C} = 0.14$ at $25^{o}C$ (Kinsman, 1969a). This value applies to direct precipitation of calcite from solution, and for the case of seawater, the resultant calcite should contain about 1100 ppm Sr. However Katz *et al.* (1972) determined a value of 0.055 at $40^{o}C$, and 0.058 at $98^{o}C$, for the wet transformation of aragonite to calcite. Morrow and Meyers (1978) support the latter values for the transformation of aragonite to calcite on Barbados and Barbuda.

Dolomite has not been experimentally precipitated at low temperatures characteristic of sedimentary environments, hence its distribution coefficient is not known. However Sr (ionic radius = 1.16Å) may substitute largely for Ca in dolomite, and to only a minor extent for the smaller Mg ion (Behrens and Land, 1972; Jacobson and Usdowski, 1977), suggesting that dolomite would contain about half the amount of Sr as calcite precipitated under the same conditions. Hence dolomite precipitated from seawater should contain about 550 ppm Sr. Katz and Matthews (1977) in an experimental study of the dolomitization of aragonite at 252-295^{o}C, determined a distribution coefficient for dolomite which was about half that of magnesian calcite (35.6 mol% Mg) formed under the same conditions. However proto-dolomite in which the crystal structure is only partially ordered, may contain Sr values higher than those expected for well ordered dolomite (Jacobson and Usdowski, 1977).

Nature of Solutions

As is apparent from the distribution coefficient, the Sr content of carbonate minerals also depends on the mSr^{2+}/mCa^{2+} of the solutions from which they were precipitated. The ratio in seawater is 0.86×10^{-2} (Kinsman, 1969a). Variation in this ratio on evaporative concentration depends on the minerals which precipitate.

It may vary in a complex manner, especially where sulphate minerals precipitate (Butler, 1973). However if aragonite is the major mineral to form, the ratio will not be significantly affected because k_{Sr}^{A} is close to one. However when minerals with a distribution coefficient signficantly less than one precipitate in a closed basin or a closed diagenetic system, the ratio will rise.

Continental waters generally have lower mSr^{2+}/mCa^{2+} ratios than sea-water (Kinsman, 1969a; Turekian, 1972; Muller and Wagner, 1978). However groundwaters in carbonate terrains, especially where aragonite is present, may have higher ratios (Plummer *et al.*, 1976; Morrow and Mayers, 1978).

Diagenesis

Diagenesis within carbonate sediments results in mineral inversion from metastable phases, such as aragonite, Mg-calcite, and protodolomite, to the stable phases calcite and dolomite, and often in further recrystallisation of these stable phases. The change in strontium content of the sediment depends on the mSr^{2+}/mCa^{2+} ratio in the interstitial pore waters, and the nature of the diagenetic system, in particular how open or closed it is (Kinsman, 1969a; Pingitore, 1978). In a closed system, for example during the aragonite to calcite transition, the Sr content of the interstitial pore waters will rise as the mineral conversion proceeds. The bulk Sr content of calcite following a complete transition will be the same as that of the precursor aragonite (Pingitore, 1978). In an open system, the resultant Sr content will depend on the distribution coefficient, and mSr^{2+}/mCa^{2+} of the pore waters, as the excess Sr is removed from the system.

How open or closed the diagenetic system is depends on a number of factors, especially the sediment porosity and permeability, timing of diagenesis, and the nature of the mineral inversion/recrystallisation process. Morrow and Mayers (1978) anticipate that grainstones will experience greater Sr loss than mudstones, which have lower porosity and permeability. Early diagenesis occurs in both the vadose and phreatic zones. In the vadose zone, mineral inversion by thin film processes may act as a temporarily closed system until it is influenced by percolating solutions. Hence Sr contents are variable and relatively high (Pingitore, 1976). In the phreatic zone, ion exchange will be better and the system more open, especially if there is a development of secondary porosity. However mineral inversion may still occur by thin film processes, with the result that the precipitating minerals are not in equilibrium with the bulk of the aquifer water (Plummer *et al.*, 1976; Pingitore, 1976), and high Sr values may result (Veizer *et al.*, 1978).

The extent of diagenesis will also influence the Sr values. Carbonate sediments with a limited diagenetic history and one which occurred under the influence of solutions similar to those in which the original sediment formed, will not experience large reductions in Sr content. However when diagenesis occurs under the influence of dilute meteoric waters over a long period of time, the Sr content of the original sediment will be substantially reduced (Morrow and Mayers, 1978).

Discussion

The relative importance of the above variables needs to be determined in order to assess the Sr values of ancient carbonates, in particular those of dolomites for which Sr data may be useful in determining the dolomitization history. The Sr contents of some modern and ancient dolomites are given in Table 10.4, and it is apparent that most ancient dolomites have low Sr contents. Veizer and Demovic (1974) and Veizer *et al.* (1978) suggest that dolomites have a bimodal Sr distribution, and that this reflects their original mineralogy. The dolomite group with low Sr values (20-110 ppm) could not be derived from $CaCO_3$ precursors with greater than 1500 ppm, and hence were originally calcite. The high Sr group (80-700 ppm) formed from precursors with higher Sr contents, such as aragonite (Fig. 10 in Veizer *et al.*, 1978). The Sr content of the dolomite depends on that of its precursor because the dissolution-reprecipitation occurs by thin film processes. The chemistry of the dissolving phase influences the chemistry of the thin film which may not be in equilibrium with the bulk of the aquifer water (Veizer *et al.*, 1978). Direct precipitation of protodolomite could also produce high Sr values. If progressive ordering of the dolomite occurs in a near surface environment, and hence under the influence of the same waters from which the protodolomite was precipitated, similar or slightly lower Sr values may be retained.

Morrow and Mayers (1978) in a computer simulation of limestone diagenesis under the influence of meteoric groundwater, based on data from Barbados and Barbuda, suggest that water chemistry including acidity, porosity and permeability, were the main factors determining the amount of Sr reduction. However their method involved the inversion of all metastable phases (aragonite, Mg-calcite) to calcite, at which time early diagenesis was considered to have ceased. The number of pore volumes required to achieve this varied considerably depending on the original mineralogy, as well as the porosity. According to Pingitore (1978) similar results are produced for the aragonite to calcite transition by various different combinations of the above factors.

RESULTS

Dolomite Facies

The mean Sr contents for each formation in each area are given in Table 10.1 and frequency histograms in Figure 10.11. It is apparent that the dark-grey dolomites of the Yadlamalka and Mirra Formations, the Montacute Dolomite, and the upper part of the Skillogalee Dolomite (subsequently referred to as the high-Sr group), have higher Sr contents than dolomites of the Nathaltee and Nankabunyana Formations, and the lower Skillogalee Dolomite (subsequently referred to as the low-Sr group). A t-test shows that the mean Sr value of samples from the Yadlamalka and Mirra Formations, is significantly different from means of both the Nathaltee and Nankabunyana Formations, at greater than the 99.9% confidence level. Likewise mean values of buff and grey dolomites within the Skillogalee Dolomite are signficantly different. However all samples fit into the early diagenetic or penecontemporenous dolomite group of Veizer and Demovic (1974) and Veizer *et al.* (1978).

Low-Sr Group

Samples within this group are largely dolomite mudstones (dolomicrosparites) and stromatolitic dolomites. The only facies controlled Sr variation appears to be higher Sr values for grey dolomites as compared with pale coloured dolomites. The more recrystallised dolomites of the Skillogalee Dolomite have similar Sr values to the dolomicrosparites of the other formations. As the main impurity in these dolomites is quartz, there may have been little exchange of Sr between the dolomite and other mineral phases present during recrystallisation. Similarly, the more recrystallised dolomites of the Nankabunyana Formation at Arkaroola have only slightly lower values than elsewhere. Within this sample group, Sr is not generally significantly correlated with any other element (Table 10.3).

High-Sr Group

This sample group consists largely of grey and dark-grey dolomite mudstones, but grainstones and stromatolitic dolomites are also present (Table 10.2, Appendix 4). Mean values for different areas range from 316-852 ppm, clustering in the 400-650 ppm range. These values are substantially higher than many other dolomites (Table 10.4). However they are only slightly less than the Sr contents of some modern dolomites (Table 10.4).

Dolomite and magnesite within the mudstones of the formations within this sample group, formed as either primary or penecontemporeneous minerals at or near the sediment interface (Chapters 6 and 9). Many dolomite mudstones have a detrital texture, and the detrital mud and peloids may have been dolomite, or a closely related mineral, at the time of deposition. Textural evidence also indicates that dolomite mudstones had a limited early diagenetic history, and this may partly account for their high Sr values. However oncoid and ooid grainstones, and to a lesser extent stromatolitic dolomites and intraclastic grainstones, are more recrystallised than dolomite mudstones, but they have similar Sr values (Table 10.2). This suggests that the recrystallisation occurred as a result of late diagenesis and metamorphism, during which the system remained closed with respect to Sr. However massive dolomites, with a more complex diagenetic history involving dissolution under the influence of dilute pore waters and the precipitation of cements, have lower Sr values (Table 10.2).

There is little systematic variation between areas. However the Sr values increase across the Willouran Ranges from southwest to northeast (Table 10.1) towards areas in which dolomite deposition was exclusively submergent. There is also a tendency for those areas with the highest proportion of magnesite (Depot Creek, Copley, Arkaroola, Top Mount Bore), and hence probably the shallowest and most frequently exposed areas, to have the lowest Sr mean values.

The only significant correlations are negative with both Fe and Mn. This may support the suggestion of Veizer (1977b), that decreasing Sr content is related to increasing substitution by Fe.

Discussion

The two groups of samples have distinct Sr values. Factors which may have caused the variation between the two groups include differences in primary mineralogy, differences in the mSr^{2+}/mCa^{2+} of the precipitating solutions, and nature and duration of diagenesis. Both sample groups fall in the high Sr group of Veizer *et al.* (1978) which would suggest a penecontemporeneous or early diagenetic origin from an aragonitic precursor. Alternatively the sediments may have initially precipitated as protodolomite, and experienced stabilization to more ordered dolomite in a diagenetic environment at and near the sediment interface, and hence under the influence of pore solutions similar to those of the overlying basin waters.

Textural evidence indicates that dolomite mudstones were significantly lithified prior to compaction, and the very fine grain size indicates that diagenesis occurred under the influence of solutions of similar composition to those from which the primary sediment was precipitated. More dilute meteoric solutions were not, in general, present during diagenesis. However dolomite mudstones in the Nathaltee and Nankabunyana Formations, may be slightly more recrystallised (coarse microspar) than those in the Yadlamalka and Mirra Formations (fine microspar) and this may reflect a slightly longer diagenetic history, during which Sr contents experienced a greater reduction. This is reflected in the higher Mn and Fe values, which increased progressively during diagenesis as Sr was reduced. However the Sr values determined during diagenesis were not significantly altered during metamorphism, whereas Fe and Mn were further increased. This is shown in plots of Sr vs Mn (Fig. 10.12), and to a lesser extent for Sr vs Fe (Fig. 10.13). The hyperbolic curve indicates early Sr decrease, and slight Mn increase, with further Mn increase associated with a lesser change in Sr.

In a closed basin, increased concentration of waters on evaporation associated with precipitation of Ca-Mg carbonates, will cause an increase in the mSr^{2+}/mCa^{2+} ratio of the water (Rothe *et al.*, 1974; Muller and Wagner, 1978), and hence a small increase in the Sr content of the carbonates may result. However this increase is associated with a larger Na increase (Muller and Wagner, 1978, Table 3). From discussions in Chapters 4, 5, 6 and 8, it appears that dolomites of the high Sr group (Yadlamalka Formation, Mirra Formation, Montacute Dolomite and upper part of Skillogalee Dolomite), formed in a basin in which circulation was more restricted than that in which dolomites of the low Sr group formed. However there is no Na variation between these two groups, and Na does not correlate with Sr. This may suggest that variation in mSr^{2+}/mCa^{2+} of the basin waters was not a major factor in causing the variation between the two groups, or that Na variations as a result of any salinity variations were not large enough to be retained during diagenesis.

Hence the major reason for the Sr variation between the two groups, and probably within the two groups, may be the degree of early diagenetic alteration of different facies, with a greater degree of alteration producing a greater Sr reduction. Within the high Sr group, dolomite mudstones deposited in areas of predominantly submergent environments may have experienced the least amount of alteration (e.g. eastern half of the Willouran Ranges-MI, R, NW; Yatina, YT; and areas in which dark grey dolomites of the Skillogalee Dolomite were deposited, see Table 10.1), thus accounting for the particularly high values in these areas.

Magnesite Facies

The strontium content of magnesite facies is much less than that of the associated dolomites. This is due to the tendency for Sr to substitute to only a very minor extent for Mg. The negative correlation between Sr and the percentage magnesite of the total carbonate within the sample (Fig. 10.14) supports this, as does the extremely low content of Sr in samples which have close to 100% magnesite. The samples of 100%, or close to 100% magnesite are nodular magnesite, or intraclastic magnesite derived from this facies. The nodular magnesite formed by replacement of magnesite mudstones which usually contained some dolomite. Hence it appears that this occurred in an open diagenetic system, allowing removal of both Ca and Sr.

Sr ISOTOPES

Sr^{87}/Sr^{86} ratios for several samples from the Mundallio Subgroup along with other Adelaidean samples are published in Veizer and Compston (1976). These ratios, and those of the River Wakefield Subgroup, are higher than samples from younger Adelaidean carbonates, and from the Bitter Springs Formation (Table 10.5), and are higher than the apparent ratio for seawater at that time. The authors indicated that these values may be too high. They discounted selective leaching of radiogenic Sr from the silicate fraction, but suggested that they were dealing with a mixed $^{87}Sr/^{86}Sr$ from two end members, the carbonate and silicate phases, due to some degree of equilibration during deep burial, as indicated by signs of recrystallization and strong diagenesis, accompanied by Sr loss. However dolomites from the Mundallio and the River Wakefield Subgroups, have not experienced strong diagenesis as indicated by their high Sr values (Table 10.4) and fine grain size. In addition because of the high Sr values in the carbonate phase, the contribution of Sr from the silicate fraction will form only a small component of total Sr. An alternative explanation is that these dolomites did not form from seawater, but precipitated from waters with a greater continental influence, i.e. non-marine, which would produce the higher initial ratios.

SUMMARY

The geochemical studies have enabled different chemical groups to be distinguished (largely using Sr, Mn, Fe), and these groups correspond to formations previously defined on lithological grounds. Similar facies, including dolomite mudstones and stromatolitic dolomites, occur in different groups, and hence have different chemistries. The controlling factor does not appear to be depositional facies, but is rather the extent of alteration during diagenesis, and to a lesser extent metamorphism, which tend to increase Mn and Fe, and reduce Sr. However the amount of Mn and Fe incorporated within the carbonate lattice, also depends on their availability within interstitial solutions, as shown by higher values in dolomite mudstones of the Yadlamalka and Mirra Formations than in the associated stromatolitic dolomites and dolomite grainstones within these formations. However different facies in these two formations have similar Sr contents, reflecting similar primary mineralogy, and similar early diagenetic histories. Following early diagenesis, the system remained closed to Sr, although further recrystallisation during metamorphism occurred in some facies, and in some areas.

The Sr contents, in particular for the Yadlamalka and Mirra Formations, are much higher than those of other dolomites (Table 10.4). This reflects the very early and possibly primary nature of the dolomite, and significant mineralogical stabilization to well ordered dolomite during very early diagenesis under the influence of interstitial pore waters similar in composition to the overlying basin waters. The diagenetic model used by

Morrow and Mayers (1978), involving the inversion of metastable to stable phases during the passage of many pore volumes of dilute meteoric waters, with further recrystallisation of stable phases under similar conditions, is clearly not applicable. The diagenetic history of these dolomites may be even more simple than that of the penecontemporeneous and early diagenetic dolomites formed from aragonite in hypersaline environments, as described by Veizer *et al.* (1978). These early diagenetic dolomites contain some associated calcite.

Similar high values were measured by Veizer and Compston (1976) for the dolomites of the River Wakefield Subgroup, whereas early diagenetic dolomites of the Bitter Springs Formation have much lower Sr contents (Table 10.4). Sr isotopic ratios of dolomites within the Mundallio and the River Wakefield Subgroups (Veizer and Compston, 1976), may indicate that these dolomites did not form in a marine environment, although the amount of data is small.

The sodium contents of dolomites from all formations is similar, indicating that average salinity did not vary markedly during deposition of the Mundallio Subgroup.

CHAPTER 11

OCCURRENCE AND ORIGIN OF DIAGENETIC CHERT NODULES WITHIN CARBONATE FACIES

INTRODUCTION

Chert nodules, consisting predominantly of microcyrstalline quartz, are a minor component of carbonate sediments within the Mundallio Subgroup, in particular within the Yadlamalka and Mirra Formations, and the Skillogalee and Montacute Dolomites. They are clearly of secondary origin, as they cross-cut bedding, lamination and other structures within the host carbonate sediment. Moreover the morphology and petrology of chert nodules are related to the nature of the host sediment, and these properties will be described in detail below. However a discussion of the forms of non-detrital silica which are found in low temperature, sedimentary environments, and the aqueous geochemistry of silica, is presented first.

FORMS OF SILICA

QUARTZ

Quartz has a variety of microscopic forms distinguished by different optical properties, grain size, and morphology. They all have the diffraction pattern of quartz, and hence its crystal structure.

Chalcedony : This is the term generally applied to the microcrystalline, "fibrous" form of quartz. It may have slightly lower and more variable refractive indices than normal quartz. It is characterised by fibrous extinction, but may also be more undulose and irregular (Plate 11.3a; Smith, 1960). The common direction of fibre elongation is along one of the polar $[11\bar{2}0]$ directions of quartz, a direction which is a two fold symmetry axis (Frondel, 1978). This produces the common length fast variety of chalcedony. Quartz fibres may less commonly be elongated along the quartz C-axis, that is the $[0001]$ direction. This produces length slow chalcedony, also known as quartzine. Its occurrence has been considered diagnostic of former evaporitic or hypersaline environments (Folk and Pittman, 1971). More recent work indicates that although this is a common association, it is not unique (Oehler, 1976). The fibre direction may also be oriented at approximately 30° to the C-axis, this is the variety lutecite (Folk and Pittman, 1971).

The elongation of quartz fibres along one of the polar [1120] directions, is believed to be controlled by dislocations implanted during nucleation and growth, causing preferred growth along one of three otherwise equivalent directions (Frondel, 1978). The C-axis lies perpendicular to the fibre axis, and it is often rotated around it, causing the development of banded chalcedony (Plate 11.3a).

Fibrous quartz has two main fabrics, either random or spherulitic (Frondel, 1978). In the latter case, crystallisation generally occurred at an interface, resulting in divergently fibrous, or parallel fibrous crusts (Plate 11.3b). In nodular chert, more random orientation of isolated fibres occurs, although there may be some degree of organisation around the sites of nucleation, producing partly spherulitic forms (Plate 11.3a; Frondel, 1978).

<u>Microcrystalline Granular Quartz</u> : Microcrystalline quartz also occurs in a granular form, characterised by small (less than 10μ), polygonal equant grains, which show pinpoint extinction (Plate 11.3b). However because of the small grain size, and hence overlapping polarisation effects, grain size is difficult to determine optically. This form of quartz is characterised by the adjacent grains having significantly different crystallographic orientations, whereas in the more fibrous varieties, misorientations between the adjacent grains are much smaller, resulting in their undulose extinction. However gradations between the two forms occur.

<u>Megaquartz</u> : Coarser varieties of quartz range in grain size up to 1 mm or more. Megaquartz contains polygonal grains, with planar and more irregular boundaries (Plate 11.3b and c). Megaquartz mosaics may be equigranular,

have more randomly variable grain size, or a drusy fabric. Other quartz varieties microcrystalline granular quartz and chalcedony, are often gradational into megaquartz.

OPAL

The various forms of opal (opal-A, opal-CT and opal-C) which have been described by Jones and Segnit (1971), are other crystallographic forms of silica which form in low temperature sedimentary environments. However they are commonly replaced during diagenesis by one of the microcrystalline forms of quartz, and are preserved only in relatively young sediments (Mesozoic to Tertiary). Opal-A is an amorphous form of silica, its major occurrence being in siliceous skeletons. Opal-CT forms spherical bodies called lepispheres, and relic spherical structures may be preserved in microcrystalline quartz formed by replacement of opal-CT.

LOW TEMPERATURE GEOCHEMISTRY AND THE PRECIPITATION OF SILICA

SOLUBILITY

Different forms of silica have different solubilities in aqueous solution, with amorphous silica being the most soluble. The solubility of amorphous silica is pH independent for pH's below 9, with a value of 100-140 ppm at 25°C (Krauskopf, 1967; Millot, 1970; Berner, 1971). When the pH rises above 9, the solubility of silica increases sharply. The solubility of amorphous silica in seawater is very similar to that in fresh water, as the solubility is largely unaffected by the presence of the other ions in seawater. Although Al^{3+} ions may decrease the solubility of silica, their concentration in seawater is too low to be significant (Millot, 1970). Mg^{2+} ions may also decrease the solubility, but only at high pH, or high Mg concentrations, at which Mg-silicates, such as sepiolite and palygorskite precipitate.

The solubility of quartz is much less than that of amorphous silica. Quoted values fall in the range of 4-10 ppm, with the higher values determined from extrapolation of higher temperature studies, and the lower values determined from experimental solubility measurements (Siever, 1972). Other crystalline phases of silica have solubilities significantly less than that of amorphous silica, but higher than quartz. For example, α-cristobalite has a solubility of 25-30 ppm (Meyers, 1977). The solubility of all forms of silica increases with temperature.

SiO_2 IN NATURAL WATERS

Seawater has variable silica concentrations depending on water depth, with up to 12 ppm in deeper oceanic waters. However seawater is generally undersaturated with respect to quartz (Berner, 1971). River waters entering the ocean often have higher silica concentrations, but the silica may be rapidly removed by organisms, or used in silicate equilibria (MacKenzie and Garrels, 1966; Lancelot, 1973). Silica concentrations of river waters and non-alkaline lakes are often greater than that of quartz saturation, but less than amorphous silica saturation. However alkaline lakes, in which Na^+, HCO_3^-, and CO_3^{2-} are the major ions present, become saturated and supersaturated with respect to amorphous silica as evaporative concentration causes the silica concentration and pH to increase (Garrels and Mackenzie, 1967; Jones *et al.*, 1967). The wide variation in silica concentration in natural waters is due to the variation in the amount received from weathering of silicate rocks and from hot springs and other groundwaters, dilution by rainwater, and removal biologically or by chemical reactions (Siever, 1972).

PRECIPITATION OF SILICA

The factors determining which form of silica will precipitate include the nature of the precipitating solutions (concentration, pH, and the other ions present), and the properties of the site of precipitation, including, in the case of silica forming by replacement, the nature of the material being replaced (Millot, 1970). Variation in these factors on a microscopic scale and with time, results in the close association of different forms of silica. For example, crystallisation within a localised gel or portion of the interstitial pore solution, could cause changes in the composition of the remaining gel or solution, and hence systematic changes in the nature of the silica phase being precipitated (Jeans, 1978).

Quartz growth is favoured by dilute solutions and slow growth rates, which allow the more internally ordered quartz to form (Jones and Segnit, 1972). The presence of suitable nucleating sites probably also assists quartz precipitation, and this property has been used by Mackensie and Gees (1971) to precipitate quartz experimentally at conditions similar to those at the earth's surface. More concentrated solutions and more rapid growth rates will favour the growth of the more disordered forms of silica, chalcedony, opal CT, and amorphous silica (Oldershaw, 1968; Meyers, 1977; Robertson, 1977). The presence of impurities may also be significant in producing the more disordered forms of silica (Millot, 1970; Lancelot, 1973; Meyers, 1977).

Amorphous silica inverts to other forms of silica, probably by different paths depending on the presence of other minerals, and the conditions during diagenesis and burial. For deep sea cherts, the following maturation sequence has been proposed,

opal-A → opal-CT → chalcedony or cryptocrystalline quartz

(Kastner *et al.*, 1977). However Oehler (1976), following low temperature hydrothermal experiments involving crystallisation of a silica gel, proposed that amorphous silica could invert to chalcedony, and thence to polygonal quartz, without an intervening opal-CT phase. Remnant microspherular structures preserved within the polygonal quartz are due to the former presence of spherulitic chalcedony.

Precipitation of silica within carbonate sediments is favoured by the variations in pH which occur in some carbonate environments (Peterson and von der Borch, 1965). In the Coorong ephemeral lakes, a high pH is produced by organic activity in winter and spring, and silica concentration increases. Subsequent decay of organic matter within the sediment, causes pH to fall, and the precipitation of silica within the carbonate mud. A decrease in the pH of interstitial solutions within carbonate sediments, will also favour their dissolution, and hence replacement by silica, if the interstitial solutions are silica saturated. The presence of CO_2 may also enhance silica precipitation (Lovering and Patten, 1962). Organic matter within the sediment also provides favourable nucleating sites for silica precipitation (Chanda *et al.*, 1976; Oehler, 1976).

However according to Knauth (1979), pH values above 9 which cause a marked increase in silica solubility, are not common in sedimentary environments. Hence he proposes a mixing zone model to explain the features of chert nodules present in typical Phanerozoic carbonates formed in shallow water environments. Mixing of seawater and groundwater may cause undersaturation with respect to calcite and supersaturation with respect to opal-CT and quartz, and hence the replacement of calcite by silica in the mixing zone (Knauth, 1979).

In summary, all forms of silica will precipitate from aqueous solution in low temperature sedimentary and diagenetic environments. The more disordered phases will probably precipitate more easily than quartz. The amorphous and opaline phases subsequently invert to microcrystalline quartz, especially with increased temperature. This inversion is also favoured in carbonate sediments, particularly where they are free of other impurities (Lancelot, 1973; Kastner *et al.*, 1977; Jeans, 1978). This is apparently due to the presence of Mg-hydroxide nuclei, for which the Mg has been supplied from seawater, and the alkalinity from dissolution of carbonate (Kastner, *et al.*, 1977).

OCCURRENCE OF CHERT NODULES WITHIN CARBONATE FACIES OF THE MUNDALLIO SUBGROUP

DOLOMITE MUDSTONES

The macroscopic form of the chert nodules reflects the nature of the host sediment. In well laminated, fissile dolomite mudstones, chert forms elongate lenses, parallel to, although cross-cutting the lamination (Plate 6.1a). In the more massively outcropping dolomite mudstones, chert nodules are more irregular (Plate 11.1a). This may reflect greater homogeneity due to little textural variation between laminae, within the more massive dolomite mudstones. Some horizons are preferentially silicified, with chert nodules present for several tens of metres along strike, while the enclosing beds are unaffected. Lamination within dolomite mudstones may be disrupted and compacted around chert nodules (Plate 11.1b), indicating the formation and lithification of silica, before the enclosing dolomite mudstone was completely compacted. However in some occurrences, there are no compaction features within the enclosing sediment. As well as being controlled by the lamination, the distribution of chert is sometimes controlled by other structures. For example in Plate 11.1c, the polygonal chert pattern may reflect the presence of desiccation cracks.

Chert nodules within dark grey dolomite mudstones are black due to the presence of carbonaceous matter. As the colour of the host sediment becomes paler, the chert also changes to light grey and white, indicating formation after substantial oxidation of organic matter.

The microscopic texture of the chert nodules may be simple and homogeneous, or more heterogeneous. The simple chert nodules consist of microcrystalline granular quartz (Plate 11.4b), which preserves the lamination of the host sediment. The degree of silicification is variable, from near complete with only minor dolomite inclusions, to much less complete, resulting in an intimate mixture of microscrystalline quartz and dolomicrospar. The amount of remnant dolomite often varies within a single nodule. Boundaries between chert nodules and the host dolomite sediment may be sharp, although slightly irregular in detail, or more gradational (Plate 11.4b).

The microcrystalline quartz has a grain size of 1-2µ in parts of some nodules, ranging up to 4-6µ. Coarser grain sizes, greater than 10µ, have developed in areas where metamorphic grade is higher than lower greenschist facies. Fracture surfaces of microcrystalline granular quartz, studied with the Scanning Electron Microscope (SEM), exhibit compact mosaics of regular polyhedral grains, with triple point junctions (Plate 11.3a). This texture is similar to those illustrated by Keller *et al.* (1977, Fig. 1), and le Roux and Jackson (1978), which have formed due to recrystallization with increasing temperature (up to 140^{o}C, Keller *et al.*, 1977). Thus the quartz has a recrystallisation texture, probably formed during late diagenesis and metamorphism. However the lack of distortion between these nodules and the enclosing dolomite, and the preservation of undistorted textures within the nodules, suggests that the quartz initially precipitated in a crystalline form.

Other cherts have a more irregular fabric, with variable silicification on a microscopic scale. Often the more silicified areas are rich in carbonaceous material derived from degraded organic matter, while the associated more dolomitic areas contain less carbonaceous material, and the dolomite within them is recrystallised. This may suggest that the presence of organic matter favoured silicification on a microscopic scale (Chanda *et al.*, 1976), although some carbonaceous material may have been removed during recrystallisation of the dolomite. This chert also consists largely of microcrystalline granular quartz, but rarely chalcedony spherulites are present.

Some irregular nodules also contain evidence of brecciation and shrinkage following silicification. For example in Plate 11.3c, a silicified dolomite mudstone has been disrupted as a result of brittle fracture, to form angular fragments. The clasts were then cemented by chalcedony and megaquartz. Irregular sheet cracks parallel to lamination, have developed in some cherts due to shrinkage. Within more massive cherts, approximately polygonal shrinkage cracks formed (Plate 11.4a). The areas of replaced dolomite consist of brown, carbonaceous stained, microcrystalline granular quartz and chalcedony, with some remnant dolomite and quartz silt grains, while the shrinkage cracks are infilled with clear, fine-grained, polygonal quartz, which sometimes increases in grain size towards the centre of the cracks. Hence although the silica formed as a replacement of dolomite, it was initially partially hydrated and amorphous (Gross, 1972), and shrinkage cracks developed during the inversion to a less hydrated, and probably more crystalline form. Cementation of the cracks occurred as quartz was precipitated directly from solutions, probably more dilute than those causing the initial silicification. Replacement of dolomite, shrinkage, and cementation, apparently occurred prior to compaction of the sediment, in an unlithified or only partly lithified dolomite mud. Shrinkage of silica, and subsequent compaction of dolomite mud around a rigid silica nodule, resulted in distortion of the lamination in the enclosing dolomite.

The sequence of silicification within dolomite mudstones is summarised in Figure 11.1. It appears that the distribution of chert nodules has been influenced by properties of the original sediment, including the distribution of organic matter, and possibly the homogeneity of the sediment and its porosity and permeability.

STROMATOLITIC DOLOMITES

Both columnar stromatolites and small domal stromatolites have experienced some replacement by irregular chert nodules (Plates 6.7a, 11.2a). As in dolomite mudstones, the colour of the chert nodules depends on that of the host sediment. The shape of the nodules is often controlled by lamination within the stromatolites, but all nodules are cross-cutting to some extent.

Within some columnar stromatolites, both the columns and the adjacent interspace sediment are silicified, while in others, parts of columns have been preferentially silicified (Fig. 12e, in Preiss, 1974b). However the chert nodules appear to be randomly distributed within the bioherms or biostromes, of which only a volumetrically small amount has been replaced. Distortion around chert nodules is insignificant, hence silification post-dated early cementation or lithification of stromatolites.

Within nodules, silicification is generally incomplete. Nearly completely silicified laminae or sets of laminae, alternate with dolomite rich laminae or sets of laminae in which silicification is less complete. The relic dolomite within nodules has often been finely recrystallised subsequent to the silica replacement. Wispy carbonaceous laminae are a common feature of the nodules (Plate 11.4c).

Microcrystalline granular quartz is the most common form of silica present. Chalcedony, largely of replacement origin, and less frequently as a pore filling, is also present. Microcrystalline granular quartz and chalcedony are sometimes present in alternating irregular laminae. In black cherts, the microcrystalline granular quartz has a carbonaceous staining, while the chalcedony is often, but not always, free of carbonaceous material. The chalcedony has a somewhat granular appearance, with some areas of more well defined radial extinction (Plate 11.3a). The latter is commonly banded. The chalcedony appears to consist of microcrystalline quartz, in which adjacent grains have very similar crystallographic orientation, hence grain boundaries are indistinct. Their orientation is such that grain aggregates have fibrous, spherulitic and undulose extinction.

Examination of fracture surfaces of chert with the SEM, reveals that microcrystalline quartz forms a compact, polyhedral, mosaic, with plane grain faces, and triple point junctions (Plate 11.8b), similar to that in quartz replacing dolomite mudstones. Grain size is largely in the range of 2-6μ, but may be finer (1-2μ), or coarser due to recrystallisation to fine megaquartz. Some samples fracture across grains rather than along grain boundaries (Plate 11.8c) because of the strongly interlocking fabric produced during recrystallisation. Chalcedony cannot be distinguished with the SEM from microcrystalline granular quartz, because the preferred crystallographic orientation cannot be observed. This is only revealed with the petrographic microscope. Chalcedony appears to have a granular fabric, morphologically similar to microcrystalline granular quartz. This probably also formed during recrystallisation from a finer grained precursor, which had better defined fibrous extinction. This extinction pattern has been partly preserved during recrystallisation to a more granular fabric by adjacent grains having similar crystallographic orientations.

Outlines of microspherular structures, 10-30μ in size, are rarely preserved within the chert nodules (Plate 11.4c). They have displaced carbonaceous material with growth, and have subsequently recrystallised to microcrystalline granular quartz with poor radial extinction. The structures may have initially precipitated as chalcedony spherulites, which would indicate the presence of a viscous, colloidal solution, saturated with respect to silica (Oehler, 1976).

Although the present morphologies of microcrystalline granular quartz and chalcedony are the result of recrystallisation, they also reflect primary differences in the morphology of the silica which replaced the dolomite. Oldershaw (1968) lists several factors important in determining which of these two forms will precipitate. These factors include the presence of impurities and the concentration of silica in solution, higher values of which will favour the precipitation of chalcedony. The structure of chalcedony reflects dislocations incorporated during nucleation and growth (Frondel, 1978), and hence these factors must have some influence on the manner of nucleation of silica. Variations in properties of the carbonate sediment being replaced, could also influence nucleation. These include the porosity and permeability, the availability of suitable nuclei, and the rate of dissolution of carbonate. Microcrystalline granular quartz commonly occurs in the more carbonaceous zones, possibly indicating that the carbonaceous material provided suitable nucleating sites. Hence this form of quartz may tend to form in a more homogeneous, permeable sediment with an abundance of nucleating sites, whereas the formation of chalcedony indicates more difficult nucleation (Spry, 1969; Oehler, 1976). However alternating zones of microcrystalline granular quartz are also present within pale cherts, now lacking carbonaceous material. Lower pH also favours precipitation of microcrystalline granular quartz (White and Corwin, 1961). Hence initial variations within the host sediment, and local variations in concentration of interstitial solutions, determined the form of silica to precipitate.

This does not exclude the possibility that other forms of silica, such as opal CT, precipitated in some laminae, and subsequently inverted to chalcedony (Meyers, 1977). The precipitation of associated microcrystalline granular quartz and opal CT, will also depend on the concentration and nature of the silica-rich solution (Jeans, 1978).

The sequence of silification in stromatolitic dolomites is summarised in Figure 11.2.

ONCOID GRAINSTONES

This facies forms a minor component of the Yadlamalka Formation, and a very rare component of the Nathaltee and Nankabunyana Formations. Within the Yadlamalka Formation, this facies is notable for its extensive and often near complete replacement by silica (Plate 6.5c).

Regular Coated Oncoids

These oncoids have been replaced by microcrystalline granular quartz which contains few dolomite inclusions (Plates 6.6a and b, 11.5). Some have an almost isotropic appearance due to the high content of carbonaceous material (Plate 11.5c). Clear laminae of fine megaquartz are present in some oncoids. In others the microcrystalline quartz has recrystallised to megaquartz, but the carbonaceous laminae are still present, and their distribution is unrelated to the present quartz grain boundaries. Hence it appears that carbonaceous material is relatively immobile during recrystallisation. Rarely radial or more irregular shrinkage cracks are present within the oncoid cortex (Plate 11.6a). The cracks are infilled with silica cement, similar to that forming the interstitial cement between oncoids.

Carbonate intraclasts present as nuclei within oncoids are less completely silicified than the oncoid cortex (Plates 6.6a, 11.5b and c). They contain less carbonaceous material, and the quartz is slightly coarser grained. Preferential silicification of the carbonaceous oncoid cortex has also occurred when oncoids are enclosed in dolomite mudstone, which has been only slightly silicified (Plate 11.5a).

Fracture surfaces studied with the SEM have a compact mosaic of polyhedral grains, 2-6μ in size, similar to that in dolomite mudstones and stromatolitic dolomites. Strongly interlocking fabrics which fracture across grains rather than along grain boundaries, are also present. Rarely less recrystallised quartz is present, with poorly defined grains about 1μ in size (Plate 11.8d). Areas of cryptocrystalline silica enclosed within aggregates of polyhedral quartz grains are also rarely present (Plate 11.8e).

Most silicified oncolites contain a silica cement, but when this is absent, the oncoids have been significantly distorted by compaction

(Plate 11.5b and c). Cemented and uncemented zones exist in close proximity (Plate 11.5b and c). The distortion of the silicified oncoid cortex, indicates that the silica initially formed in a semi-plastic state.

Oncoids, prior to silicification, consisted of a mixture of dolomite and organic matter derived from degraded algal material. Replacement of dolomite by silica may have been enhanced by the presence of the organic matter. Decomposition of the algal material would have produced CO_2, thus decreasing the pH of interstitial solutions, and favouring the dissolution of dolomite (Siever, 1962). Subsequent precipitation of silica would have been favoured by the presence of CO_2 (Lovering and Patten, 1962), and the presence of organic matter to act as nuclei (Oehler, 1976). The influence of organic matter is indicated by the less complete silicification of carbonaceous poor areas, such as dolomite intraclasts and dolomite mud.

The occasional presence of shrinkage cracks and distortion of silicified oncoids, indicates that at least some of the primary silica was a partly hydrated and probably amorphous phase, formed due to the presence of supersaturated solutions. Subsequent recrystallisation may have initially formed a more rigid structure of cryptocrystalline quartz, with further recrystallisation to the existing microcrystalline granular quartz, as the temperature increased during late diagenesis and metamorphism (Fig. 11.3).

Massive Oncoids

These oncoids also show near complete silicification with only minor remnant dolomite (Plates 6.6c, 11.6b and c), but the grain size of the quartz is more variable with microcrystalline quartz and megaquartz commonly present within individual oncoids. Carbonaceous material has been removed during recrystallisation of some oncoids, while in others it has remained immobile, and been retained despite recrystallisation to megaquartz.

Some massive oncoids contain aggregates of microspherular structures, 20-50μ in diameter, outlined by carbonaceous rims (Plate 11.6b). They consist of microcrystalline quartz, sometimes with a poorly defined spherulitic extinction and may have originated as chalcedony spherulites. Other oncoids with a more diffuse carbonaceous staining, commonly contain shrinkage cracks (Plates 6.6c and 11.6c). These are similar to those described by Gross (1972) and Dimroth and Chauvel (1973) within silicified intraclasts of iron formations. Carbonaceous material is concentrated along the margins of the cracks, analagous to the concentrations of haematite dust in the silicified clasts of iron formations (Dimroth and Chauvel, 1973). The shrinkage cracks are infilled with silica cement. Hence a large proportion of the massive oncoids appear to have been replaced by hydrated, amorphous silica, which was dehydrated and recrystallised prior to compaction and cementation.

Cementation

A variety of cements are present, including length fast chalcedony, microcrystalline granular quartz, and megaquartz with a drusy or equigranular fabric (Plates 11.3b, 11.5c). The boundary between silicified oncoids and cement is sharp, and occasionally a rim of euhedral dolomite crystals is present around the margin of the oncoid, partially replacing it (Plate 11.6b). The chalcedony has a somewhat granular appearance, as in chert nodules present in other facies, but has fibrous extinction oriented perpendicular to oncoid boundaries, a common feature of chalcedony precipitated as open space filling. Transitions from chalcedony to megaquartz are common, probably indicating a decrease in the silica concentrations of the interstitial solutions during cementation.

The development of shrinkage features within oncoids as the original silica precipitate experienced dehydration, and their subsequent infilling with a silica cement, indicates a break between silicification of oncoids, and cementation.

Summary

The texture of silicified oncolites indicates that silicification and subsequent cementation occurred prior to compaction, and hence in the uppermost part of the sediment pile. However silicified oncolites did not experience reworking after silicification. Initial silicification of oncoids may have been favoured by the high organic content, and the greater permeability, as compared with the interbedded dolomite mudstones.

OTHER DOLOMITE FACIES

Intraclastic dolomites have sometimes been partially silicified. This may largely effect the intraclasts (Plate 11.2b), with minor silicification of the matrix or precipitation of quartz cement. The latter sometimes postdates a dolomite rim cement. Silicification of peloidal grainstones sometimes preserves the peloidal texture more clearly than in the host unsilicified sediment. Peloids are often replaced by microcrystalline granular quartz, and enclosed in an unsilicified matrix. However the latter has sometimes experienced replacement by granular chalcedony. Continuing silicification through diagenesis is sometimes indicated by zoned nodules, in which the peloidal texture is well preserved in the centre of the nodules, but is less distinct on the margins.

INTRACLASTIC MAGNESITE

Silicification is an uncommon feature within this facies, and magnesite mudstones appear to have been rarely, if ever, silicified. Silicification of intraclastic magnesite has occurred in two main forms, and both clearly occurred subsequent to deposition of the intraclasts. Silica occurs as small nodules and lenses, replacing only a minor part of the intraclastic bed (Plate 6.15c), while other beds have been extensively replaced (Plate 11.2c). At Copley, two silicified beds of intraclastic magnesite, separated by 2 m of unaltered dolomite, continue for at least 12 km along strike. However most extensively replaced beds cannot be followed for this distance.

The chert contains a variety of silica forms, and preservation of the original intraclastic texture is variable. Poor preservation is generally more characteristic of the extensively replaced beds. In chert nodules in which the texture is well preserved, the intraclasts are commonly replaced by chalcedony spherulites, 0.05 to 0.2 mm in diameter. They begin as isolated spherulites, which coalesce to completely replace the intraclast (Plate 11.7a and b). The resultant chalcedony consists of incomplete, interfering zones with spherulitic extinction, which may grade into areas with a more granular fabric (Plate 11.7b). Some intraclasts contain abundant, micron sized inclusions of magnesite, scattered throughout the intraclast, and these obscure the texture of the silica. Less commonly intraclasts have been replaced by microcrystalline granular quartz. The texture of the chalcedony and microcrystalline quartz as observed with the SEM, is similar to that replacing dolomite sediments. Minor inclusion filled megaquartz is present, and the cements include length-fast chalcedony and megaquartz.

There is no evidence for hydrated or amorphous silica as a primary precipitate within these nodules, but the abundance of chalcedony spherulites suggests the presence of concentrated solutions during silicification. The development of chalcedony spherulites from scattered nucleating sites indicates poor nucleation. This may have been due to the well lithified nature, and hence low permeability of the intraclasts, inhibiting migration of silica rich solutions, and the removal of magnesium in solution following magnesite dissolution. Hence the interstitial solutions may have remained impure, also favouring spherulitic growth (Oehler, 1976).

Magnesite intraclasts adjacent to chert nodules have sometimes been compacted and distorted around the nodules, indicating the formation of rigid chert nodules in which silicified intraclasts had been cemented by quartz, prior to complete compaction.

The more extensively silicified beds generally have poorly preserved textures with indistinct intraclasts (Plate 11.7c), and in some cases also resemble nodular magnesite. The unsilicified zones within these beds are commonly extensively replaced by finely crystalline dolomite, with minor remnant centimetre sized patches of cream-coloured, micritic, magnesite. The silica consists of fine polygonal megaquartz, coarser than 10μ, associated with medium crystalline polygonal megaquartz often with abundant magnesite inclusions (Plate 11.7c). These are micron sized, and sometimes have a zoned distribution within individual quartz grains. The abundance of magnesite inclusions within this polygonal quartz, indicates that the quartz is a primary precipitate, and not a replacement of finer quartz. Precipitation occurred from solutions saturated with respect to quartz, and with variable saturation with respect to magnesite. There is little evidence of the timing of silicification of this facies, but it may post date the formation of smaller nodules in which chalcedony predominated, but preceded the dolomitization of the magnesite.

DISCUSSION : THE DIAGENETIC FORMATION OF SILICA

Distribution

The formation of diagenetic silica is confined to carbonate sediments of the Mundallio Subgroup, apart from the formation of quartz cement in sandstones. Dolomite facies which are grey to dark-grey in colour have experienced more frequent silicification than paler coloured dolomites. Pale coloured dolomites within the Nathaltee Formation contain minor chert at Depot Creek, but in other outcrops of this formation, they are rarely silicified. Dolomites within the Nankabunyana Formation are only very rarely silicified. However some beds of pale recrystallised dolomite within the Skillogalee Dolomite are extensively replaced by quartz.

Within the Yadlamalka Formation, the amount of silicification varies widely between areas. Many outcrops of dolomite facies of this formation in the Emeroo Range, particularly at Depot Creek, contain chert, although usually as only a minor component. In other areas of outcrop in the southern Flinders Ranges, chert nodules are more scattered, with oncoid grainstones, stromatolitic dolomites, and massively outcropping dolomite mudstones being preferentially silicified, as compared with the more fissile dolomite mudstones, particularly those which are impure. In the northern Flinders Ranges, chert is most abundant within the Yadlamalka Formation in the Copley-Myrtle Springs area, but it is much less common here than at Depot Creek. Chert is very minor in the Arkaroola region. The abundance of chert nodules is related to the facies which are present. Areas with little chert have minor amounts of stromatolitic dolomites, oncoid grainstones, and massively outcropping dolomite mudstones.

Source of Silica

Within Mesozoic and Tertiary sediments, the source of silica in diagenetic cherts is generally attributed to the dissolution of the amorphous silica provided by siliceous micro-organisms within the sediment pile (Wilson, 1966; Middlemiss, 1978). However siliceous micro-organisms have not been observed within Precambrian sediments, and in particular, not in this sequence. Dissolution of quartz and other silicate minerals is an alternative source. Detrital quartz is present as a minor component in most carbonate sediments, and within interbedded sandstones. Most quartz grains have been etched and replaced by dolomite on their margins, but this appears to be a late diagenetic feature, postdating the precipitation of quartz cement within sandstones (Chapter 7). Hence evidence of any very early diagenetic dissolution is obscured. Quartz grains preserved within cherts have irregular margins due to the nucleation of microcrystalline quartz on them, also obscuring any irregularities which may have been present due to dissolution.

Silica may also be introduced in solution, either within surface waters or within groundwaters entering the basin. Within Precambrian basins, more of the silica may have been retained in solution due to the absence of siliceous micro-organisms, than in modern marine basins. Hence waters within Precambrian basins may have been saturated with respect to silica (Oehler and Logan, 1977). In closed basins with alkaline waters, high silica concentrations are produced on evaporation (Garrels and Mackensie, 1967; Jones *et al.*, 1967). The development of shallow, and sometimes isolated lagoons or lakes with limited circulation, and in which abundant carbonate formation indicates the presence of alkaline waters, was a feature of deposition of parts of the Mundallio Subgroup, in particular the Yadlamalka Formation and Montacute Dolomite. Hence the waters within these lakes may at times have been saturated with respect to silica.

However volumetrically, the amount of silica replacement of dolomite or magnesite is small, and a massive silica source is not required. However the presence of diagenetic chert nodules indicates processes which redistributed silica within the sediment, causing its concentration and precipitation in localized areas.

Timing

Silica is present either as a replacement of a pre-existing carbonate sediment, or as cement. There is no evidence of silica precipitation other than in a pre-existing sediment. Silicification commenced prior to lithification and compaction in some sediments, and in others, continued during compaction. Hence silicification commenced in the uppermost parts of the sediment column, but no evidence for the erosional reworking of silicified sediments has been observed.

Micro-organisms are preserved within some cherts (Schopf and Barghorn, 1969; Fairchild, 1975). Within outcrops of stratigraphic equivalents of the Mundallio Subgroup in the Peake and Denison Ranges, some microfossils are preserved oriented or aggregated in apparent living position (Fairchild, 1975), indicating that silicification occurred prior to bacterial degradation and oxidation of organic matter. However in most cherts, the morphology of micro-organisms was destroyed prior to silicification, leaving wispy laminae, or zones of structureless, disseminated carbonaceous material.

<u>Precipitation of Diagenetic Silica</u>

The nature of silica replacement suggests that provided there was some silica within pore waters, the factors leading to the concentration of silica and its precipitation with associated carbonate dissolution, depended on local variations within the sediment. Replacement may have been localised where chemical conditions were suitable, and within the more permeable horizons. The concentration of silica in solution at the site of precipitation varied, resulting in the precipitation of different forms of silica, although the morphology was also controlled by the nature of the sediment being replaced. Solutions were commonly saturated with respect to amorphous silica, as this was often the primary silica precipitate.

Carbonate facies, particularly those within the Yadlamalka Formation, were deposited in shallow lakes or lagoons, in which algal mats frequently colonised the sediment surface. Prolific photosynthetic activity may have increased the pH of surface waters above 9, as occurs in ephemeral lakes associated with the Coorong (von der Borch, 1965). The associated increase in silica solubility may have caused quartz dissolution, and a rise in silica concentration (Peterson and von der Borch, 1965). In more isolated lakes in which evaporation played a greater role, silica concentration and pH may have also risen due to evaporative concentration (Jones *et al.*, 1967). If these silica rich solutions were trapped interstitially within the sediment where pH subsequently fell due to the presence of decaying organic material, supersaturation may have resulted in silica precipitation.

Some amorphous silica may have initially formed as a finely disseminated precipitate in dolomite mudstones. However during diagenesis, the amorphous silica may have been dissolved, and precipitated as more crystalline forms in localized areas, due to the establishment of concentration gradients and variations in sediment properties. However in some sediments, for example oncoid grainstones, amorphous silica was initially precipitated in isolated areas, due to chemical conditions being concurrently suitable for dolomite dissolution and silica precipitation. Such conditions may have developed in sediments rich in organic matter, in which the pH of interstitial solutions was low, and nucleating sites for the silica were readily available. Incomplete mixing and equilibration of pore waters due to controls on their composition by the local environment, probably favoured silica replacement occurring as nodules. Following initial silica precipitation, continued replacement within the same localized area was favoured as the silica already present provided nuclei for further precipitation.

Dolomite facies containing chert nodules generally experienced few other early diagenetic changes, as indicated by their textures and geochemistry (Chapters 4-6, 9 and 10). Hence the model for the origin of chert nodules in limestones proposed by Knauth (1979), which involves the mixing of meteoric water and seawater, is unlikely to apply to this sequence, as the dolomite sediments containing chert would contain evidence of more early diagenetic changes. Massive dolomites in the Yadlamalka Formation whose textures are the result of diagenetic change, with little relic depositional textures (Chapter 6), are not silicified.

Magnesite mudstones are generally not silicified. However they are likely to have formed in the most alkaline environments, in which silica concentrations were higher than in dolomite environments. The absence of silicification may result from several factors, including near surface lithification of this facies in exposed environments, and hence a very early reduction in porosity and permeability; little pH variation within interstitial solutions; or the precipitation of silica with magnesium as Mg-silicates, now preserved as talc, because of the high concentration of magnesium in solution.

CHAPTER 12

ORIGIN OF DOLOMITE AND MAGNESITE IN THE MUNDALLIO SUBGROUP

SEDIMENTARY DOLOMITE AND MAGNESITE : PROBLEMS OF FORMATION

DOLOMITE

Dolomite is a significant component in ancient carbonate rocks, and its abundance apparently increases with age (Engelhardt, 1977, Fig. 5.75). Some dolomite is found in association with $CaCO_3$ minerals, with clear evidence of the formation of dolomite as a replacement of $CaCO_3$ precursors. In dolomites in which depositional textures are well preserved and where the distribution of dolomite is facies controlled, or dolomite is present without associated $CaCO_3$ minerals and there is no equivocal evidence for their former presence, the origin of dolomite is more problematical. Such dolomites appear to be primary, or very early diagenetic in origin.

Despite the abundance of ancient dolomite occurrences, the lack of an adequate mechanism to explain dolomite formation, particularly where it appears to be a primary sediment, arose for the following reasons as summarised by Lippman (1973, p. 148):

> "The problem of the origin of dolomite arose and persists today for two main reasons: (1) In spite of the simple composition, Ca Mg $(CO_3)_2$, it has not been possible to synthesize the mineral from appropriate solutions in the range of ordinary temperatures to over 100°C, i.e. at temperatures characteristic of the sedimentary cycle. (2) Dolomite is rare in Recent and Pleistocene sediments in proportion to the total amount of carbonate sediments of that age, and also in comparison with the abundance of dolomite in ancient sedimentary rocks."

Dolomite is found in small amounts in a number of modern environments, including marginal marine locations in Florida and the Bahamas (Shinn *et al.*, 1965), Bonaire (Delfeyes *et al.*, 1965), the Persian Gulf (Illing *et al.*, 1965), Broad Sound in Queensland (Cook, 1973), and in a subtidal environment in Baffin Bay, Texas (Behrens and Land, 1972). Lacustrine environments in which dolomite is present in Recent and sub-Recent sediments, include shallow lakes associated with the Coorong lagoon in South Australia (Alderman and Skinner, 1957; Skinner, 1963; von der Borch, 1962, 1965, 1976) several lakes in Texas (Reeves and Parry, 1965) including Salt Flat (Friedman, 1965), Deep Springs Lake, California (Peterson *et al.*, 1966; Clayton *et al.*, 1968), Lake Balkhash in the U.S.S.R. (Strakhov, 1970; Torovskii and Sheko, 1974), and Lake Balaton, Hungary (Muller and Wagner, 1978).

Recent dolomite occurrences provide models for water chemistry and mechanisms of carbonate production, but because such occurrences are small, the depositional environments may not be comparable to those of ancient extensive dolomite deposits with respect to sedimentation processes. The lack of extensive, modern dolomite occurrences arises partly because modern environments are not completely representative of environments which were present in the past. For example shallow epicontinental seas, probably a major locus of dolomite deposition, appear to have covered much of the continental areas in the past (Heckel, 1972). In addition there has been a progressive shifting of the dominant mode of carbonate deposition from shallow marine, often restricted environments, to the deep sea, from the Mesozoic onwards (Veizer, 1978).

Formation of Dolomite : Chemical Factors

Evidence on the solubility of dolomite is contradictory, largely because of the difficulties in attaining equilibrium conditions. A range of values for the solubility product of dolomite have been published (Hsu, 1967), and it is generally accepted that dolomite is less soluble than both magnesite and calcite (Lippman, 1973). The solubility products indicate that seawater is saturated with respect to dolomite, calcite and magnesite, and that the dolomite is the stable phase in the system $CaCO_3$-$MgCO_3$-CO_2-H_2O at 25^{o}C and atmospheric pressure, for the mMg^{2+}/mCa^{2+} conditions of seawater (Lippman, 1973; Bathurst, 1974). However under natural conditions, the metastable phases aragonite and Mg-calcite are the major carbonate precipitates from seawater, although low Mg-calcite is significant in non-marine environments.

Because of the problems of synthetic precipitation of dolomite, and from studies of modern environments, it appears that dolomite forms by the replacement of pre-existing carbonate minerals, and not as a primary precipitate (Usdowski, 1968; Muller *et al.*, 1972; Lippman, 1973; Moller and Kubanek, 1976). However Deelman (1975a) suggests that dolomite may precipitate directly from solution. Although dolomite clearly replaces both aragonite and calcite in early and late diagenetic environments, the penecontemporeneous or very early diagenetic formation of dolomite in surface and near surface sediments, may be more likely if Mg-calcite is available as a precursor (Muller *et al.*, 1972; Moller and Kubanek, 1976; Kelts and Hsu, 1978). Ca-Mg carbonates with a composition approaching that of dolomite, but which lack or have only low intensity dolomite ordering reflections, may also be precursors to dolomite, to which they invert by solid state diffusion, or by a recrystallisation process. Those which lack ordering reflections are strictly Mg-calcites, but minerals with partial ordering are referred to as protodolomites (Gaines, 1977). The classification of minerals with the composition of dolomite, but which do not have its crystal structure of alternating sheets of $CaCO_3$ and $MgCO_3$, has recently been discussed by Gaines (1977, 1978), Deelman (1978b, 1979), Gidman (1978) and Muller and Wagner (1978). High Mg-calcites with a composition approaching that of dolomite, and protodolomites, appear to be forming by direct precipitation from solution in some modern environments (Skinner, 1963; von der Borch, 1965, 1976; Clayton *et al.*, 1968; von der Borch and Jones, 1976; Eugster and Hardie, 1978; von der Borch and Lock, 1979). Protodolomite also forms diagenetically from Mg-calcite. Experimental precipitation of protodolomite has been achieved, but with temperatures slightly higher than those of sedimentary environments, and in some cases may have occurred via a Mg-calcite phase (Liebermann, 1967; Gaines, 1974; Ohde and Kitano, 1978; Fischbeck, 1979).

Studies of modern occurrences of dolomite and its relationship to co-existing carbonate minerals, and experimental and theoretical studies, have determined the significant factors in the formation of dolomite. The major factor inhibiting dolomite growth is the difficulty of dehydrating the $Mg^{2+}(H_2O)_6$ ion (Lippman, 1973; de Boer, 1977). Precipitation of dolomite from supersaturated solutions is hampered by a decrease in crystal ordering during growth, and hence in supersaturated solutions, Mg-calcite or protodolomite precipitate (de Boer, 1977).

Elevated Mg/Ca ratios, greater than in normal seawater, appear to favour dolomite formation (von der Borch, 1965; Muller *et al.*, 1972; Butler, 1973; Moller and Kubanek, 1976; Eugster and Hardie, 1978). In marine environments, increased salinity leading to precipitation of aragonite and gypsum, increases the Mg/Ca ratio and may lead to dolomite formation. In alkaline lakes, precipitation of calcite on evaporative concentration, increases the Mg/Ca ratio, and in turn the Mg concent of the calcite, and may eventually lead to protodolomite precipitation (Eugster and Hardie, 1978), either as a direct precipitate or by replacement of Mg-calcite (Muller *et al.*, 1972; Muller and Wagner, 1978). Mg/Ca ratios greater than 6-12 favour dolomite formation in modern environments (von der Borch, 1965; Muller *et al.*, 1972; Eugster and Hardie, 1975). However elevated ratios may only be necessary in highly saline, supersaturated solutions. At lower saturations, and lower salinity, dolomite may form at Mg/Ca ratios less than seawater (Hanshaw *et al.*, 1971; Folk and Land, 1975; de Boer, 1977).

An adequate supply of CO_3^{2-} ions is also important in dolomite formation (Lippman, 1973; Davies *et al.*, 1977; Kelts and Hsu, 1978). Lippman (1973) suggests that the disproportionately low concentrations of CO_3^{2-} in seawater as compared with the cations Ca^{2+} and Mg^{2+}, inhibits dolomite formation, even though its solubility product is exceeded. This is supported by Lahann (1978), who suggests that in ionic-salt precipitation from solution, crystal growth is controlled only by the concentration of the deficient ion, in this case

CO_3^{2-}. A high concentration of carbonate or bicarbonate ions may assist in dehydration of the Mg^{2+} ion (Lippman, 1973; Lahann, 1978). Russian workers, in studies of lacustrine carbonates, also emphasize the importance of the availability of carbonate (Teodorovich, 1961; Strakhov, 1970; Turovskii and Sheko, 1974).

The formation of dolomite is also influenced by pCO_2, because of the increased solubility of carbonate minerals with increasing pCO_2. High pH may favour dolomite formation (von der Borch, 1965; Liebermann, 1967; Kelts and Hsu, 1978), because of the relationship between CO_3^{2-}, HCO_3^-, pCO_2 and pH. The pH may be increased due to the removal of CO_2, either by photosynthetic activity (von der Borch, 1965), or by increasing salinity (Eugster and Jones, 1979), and this increases the CO_3^{2-} concentration.

The presence of Na^+ and Li^+ ions may assist in the dehydration of the Mg^{2+} ion, and thus promote dolomite formation (Gaines, 1974; Deelman, 1978b; Ohde and Kitano, 1978). This may account for the promotion of dolomite formation at high salinities in marine environments.

Different combinations of the above factors may be favourable to dolomite formation either as a primary (protodolomite), or secondary mineral. Hence dolomite forms by several different mechanisms, in different sedimentary environments, as summarised in Figure 12.1.

MAGNESITE

Magnesite is an uncommon carbonate mineral in both modern and ancient carbonate sediments. Recent occurrences are found largely in lacustrine environments, in which magnesite is generally present in association with dolomite and other carbonate minerals. Examples in Recent and sub-Recent sediments include Glacial Lake Bonneville in the U.S.A.. (Graf *et al.*, 1961), ephemeral lakes associated with the Coorong lagoon in South Australia (von der Borch, 1965, 1976), several lakes in Turkey (Muller *et al.*, 1972), Lake Balkhash in the U.S.S.R. (Turovskii and Sheko, 1974), and lakes of the Basque Basin in British Columbia (Nesbitt, 1974, *in* Eugster and Hardie, 1978). Precipitation of magnesite from sea water is rare, but it is present in sabkhas of the Trucial Coast in association with anhydrite and has formed diagenetically within the sediment (Kinsman, 1969b; Bush, 1973).

Magnesite occurrences in sedimentary rocks are minor, and have commonly experienced alteration and destruction of primary textures, so that few environmental interpretations have been made. Association with dolomite is common. Occurrences have been described from Manchuria (Nishihara, 1956), Washington State, U.S.A. (Fox and Rinehart, 1968), Rhodesia (MacGregor and Bliss, 1968), India (Valdiya, 1968), eastern Europe (Lesko, 1972) and the west Pyrenees of Spain (Petrascheck *et al.*, 1977; Kralik and Hoeffs, 1978). None of these occurrences is as extensive, or contains as well preserved sedimentary textures, as the magnesite of the Mundallio Subgroup. Magnesite is also found in association with evaporite minerals, for example anhydrite (Strakhov, 1962), or even more saline minerals within the Zechstein Formation (Braitsch, 1971; Dean, 1978) and Green River Formation (Eugster and Hardie, 1978). Magnesite is generally only the dominant carbonate in such sequence when it is present in association with evaporite minerals.

Formation of Magnesite

Seawater is apparently saturated with respect to magnesite because its solubility product is exceeded (Sayles and Fife, 1973). However as is the case for dolomite, magnesite does not precipitate from normal seawater, and may be synthesized experimentally only at temperatures significantly above those characteristic of sedimentary environments (Lippman, 1973; Sayles and Fife, 1973). Magnesite is the stable magnesium carbonate mineral under earth surface conditions. Hydrated magnesium carbonates, such as nesquehonite ($MgCO_3 \cdot 3H_2O$) and hydromagnesite ($Mg_4(CO_3)_3(OH)_2 \cdot 3H_2O$)

are metastable (Christ and Hostetler, 1970), but may be synthesized at low temperatures by mixing the appropriate concentrations of a magnesium salt and alkali carbonate (Lippman, 1973). Hydration of the magnesium ion inhibits magnesite precipitation at low temperatures (Christ and Hostetler, 1970; Lippman, 1973), but this will be less significant in the formation of metastable hydrated phases.

The two important factors in dolomite formation, the supply of Mg^{2+}, usually expressed as the Mg/Ca ratio, and the supply of CO_3^{2-}, will be significant in magnesite precipitation. Magnesite forms in lacustrine sediments at higher Mg/Ca ratios than dolomite (von der Borch, 1965; Muller *et al.*, 1972; Nesbitt, 1974 *in* Eugster and Hardie, 1978). Muller *et al.* (1972) suggest that with increasing Mg/Ca ratios, the diagenetic sequence high Mg-calcite → dolomite → huntite (Ca Mg_3) $(CO_3)_4$ → magnesite develops with magnesite forming at Mg/Ca ratios greater than 40. They anticipate that hydrous magnesium carbonates may form as primary minerals at extremely high ratios, but this was not observed within the lakes in their study. A similar sequence was observed by Nesbitt (1974, *in* Eugster and Hardie, 1978) in the Basque Lake No. 2, but here sulphates are precipitated in association with protodolomite and magnesite because of the higher SO_4^{2-} content in the inflow waters. An increase in Mg/Ca in interstitial waters of sabkhas of the Persian Gulf due to gypsum and anhydrite precipitation, produces a similar diagenetic sequence with magnesite precipitating at the highest Mg/Ca ratios (Bush, 1973).

A high concentration of CO_3^{2-} may also enhance magnesite precipitation, either as a direct precipitate, or as a diagenetic product of the dehydration of hydrated Mg-carbonates, because of its effect in breaking the hydration envelope of Mg^{2+} (Lippman, 1973). Christ and Hostetler (1970) suggest that increasing salinity will favour magnesite precipitation. The associated decrease in water activity results in a higher proportion of less strongly hydrated Mg^{2+} ions. However Sayles and Fife (1973) suggested that positive catalysis of reaction rates in solutions of high ionic strength, i.e. high salinity, which was observed in the synthesis of magnesite at 126°C, may be equally as important as diminished hydration effects in promoting the formation of magnesite at high salinity. They also found that increased pCO_2 enhanced reaction rates at this temperature.

FORMATION OF DOLOMITE AND MAGNESITE IN THE MUNDALLIO SUBGROUP

SEDIMENTARY PROCESSES

Textural evidence and the association of the two minerals, indicates that both dolomite and magnesite existed as discrete minerals at the sediment surface. Within dolomite mudstones and peloidal dolomites, the preservation of fine details of the detrital texture, particularly in samples from the upper Mundallio Subgroup, suggests that these sediments were dolomite, or a Ca-Mg carbonate with a chemistry similar to that of dolomite, rather than calcite or aragonite, at the time of deposition. Textural evidence indicates that, during early diagenesis, major mineral inversions such as aragonite → calcite, and aragonite or calcite → dolomite probably did not occur. Early diagenesis involved the formation of a dolomite microspar, the grain size of which was coarser in the lower Mundallio Subgroup, reflecting a more prolonged period of diagenetic recrystallisation and alteration. This is supported by lower Sr values, and higher Mn and Fe values in dolomite mudstones of the lower Mundallio Subgroup, than in those of the upper subgroup. Some dolomitic sediments including stromatolitic dolomites, oncoid and ooid grainstones, do not show such clear textural preservation. However the Sr values are similar to those in dolomite mudstones, suggesting that the primary mineralogy was similar, and that the more extensive recrystallisation may have occurred largely during metamorphism.

Within carbonate sequences, the carbonate sediments are derived from three mechanisms:

(a) detrital input;

(b) inorganic precipitation;

(c) organic production.

The first source is generally minor. In modern shallow marine environments, the organic component is generally dominant, and carbonate muds may be derived largely from calcareous algae (Bathurst, 1974). However within some modern lacustrine environments, inorganic precipitation, which may however be biologically induced, plays a significant role (Kelts and Hsu, 1978). Likewise in the Precambrian, inorganic precipitation must have been the major process of carbonate production.

The dolomite mudstones within the Mundallio Subgroup often have a detrital fabric, however the dolomite is unlikely to have had an extrabasinal source. Detrital fabrics within carbonate mudstones have been described by other workers (Eugster and Hardie, 1975; Wanless, 1975; Zamarreño, 1975; Hardie and Ginsburg, 1977; Smoot, 1978). In the example described by Hardie and Ginsburg on the Three Creeks Tidal Flats of Andros Island, onshore storms and hurricanes place carbonate mud and peloids from the Great Bahama Bank in suspension, and transport them onto the tidal flats, forming laminae and thin beds of carbonate mud. This is the major process of sedimentation on the tidal flats. There is minor direct precipitation of carbonate as algal tufa domes and paper thin crusts in the freshwater algal marsh of the tidal flats. However most of the sediment has an offshore derivation. By contrast in the Wilkins Peak Member of the Green River Formation (Eugster and Hardie, 1975; Smoot, 1978), dolomite derived from erosion of crusts, caliches and tufas around the lake margin, was transported basinward during infrequent storm induced flooding, and deposited as laminated and thinly bedded peloidal mudstones.

The laminated, detrital fabric in the above examples results from periodic events which transfer sediment from the site of production to adjacent areas. However in small, physically separated ephemeral lakes associated with the Coorong, the precipitated carbonate mud is stirred by wind driven circulation. This produces a water saturated slurry which forms layers of homogeneous, structureless micrite or "yoghurt" muds, up to several centimetres in thickness, as the water level in the lakes falls (von der Borch and Lock, 1979).

Within the Mundallio Subgroup, it is difficult to ascertain whether mud and peloids were derived from erosion of pre-existing sediments, or formed by direct precipitation at the site of deposition where they experienced some reworking. A range of processes is likely to have been operative. Thin beds of homogeneous microspar with little detrital component, may have originated by the latter process, analagous to the "yoghurt" muds of the Coorong lakes. The association of terrigenous silt and fine-grained sand with dolomite peloids, and the presence of silt and sand at the base of graded dolomite mud laminae, suggest that dolomite mud and peloids, and terrigenous detritus, were introduced jointly from areas where both were available. Basinward transport of sand across dolomite mudflats by sheet flooding, may have caused their erosion to produce dolomite peloids. Small scours are common within dolomite mudstones, indicating erosion and reworking of unlithified muds.

Stromatolitic dolomites, although having formed at least in part by trapping of carbonate mud and minor terrigenous detritus, also formed in part by *in situ* precipitation. Erosion of stromatolites, as indicated by the presence of microunconformities, probably contributed some mud and peloids. Erosion of lithified dolomite crusts and tepees on exposed mudflats may also have contributed sediment to submerged mudflats. However these facies are minor, and hence would not have contributed a large amount of sediment.

Inorganic precipitation of carbonate muds from a standing body of water, occurs in several modern lacustrine environments (Kelts and Hsu, 1978; Muller and Wagner, 1978; von der Borch and Lock, 1979). The precipitation may be induced by biogenic factors such as photosynthetic activity removing CO_2, or by changes in a number of physico-chemical factors causing supersaturation and precipitation. These include evaporative concentration which also causes CO_2 degassing and a pH increase in alkaline waters (Eugster and Jones, 1979), temperature increase, and mechanical release of CO_2. Supersaturation may also be produced by the mixing of alkaline brines with dilute inflow waters (Eugster and Maglione, 1979). Many of these factors vary seasonally due to climatic changes, and cause annual precipitation.

Much of the dolomite within this sequence must have formed by inorganic precipitation, due to variation in one or more of the above factors on a seasonal or longer term basis. Precipitation may have been greatest in the shallowest environments, subjected to the highest, but probably variable salinity. The sediment was then subjected to several periods of transportation, deposition and re-erosion, as energy levels and water levels varied. In shallow slightly agitated environments, peloids rather than mud may have formed.

X-ray diffraction measurements indicate that the dolomite now has a uniform composition i.e. it is well ordered, and there is little variation in the Mg/Ca ratio. The primary precipitates are likely to have been more variable. They probably consisted of protodolomite and high Mg-calcite. Variation in the Mg/Ca ratio within the basin waters probably caused variations of this ratio in the precipitates, but complete inversion to stoichiometric, ordered dolomite subsequently occurred. The absence of calcite within the sequence other than as a very rare, late stage replacement of dolomite, suggests that primary low Mg-calcite and aragonite must have been of only minor occurrence within the area of study.

In contrast to dolomite mudstones, magnesite mudstones experienced little current reworking prior to lithification, although the subsequent erosion of lithified magnesite mudstones following exposure was substantial. Some magnesite formed as thin micritic crusts overlying detrital laminae of silty dolomite mud. During exposure, evaporative concentration may have increased pH and CO_3^{2-} concentration leading to dolomite and then magnesite precipitation within the crusts. However most magnesite formed within shallow isolated ephemeral ponds and lakes with little detrital influx. These lakes may have been partly fed by groundwater seepage from adjacent larger water bodies (Fig. 12.2). The water level rose and fell due to seasonal changes in evaporation rates, and in the shallowest and smallest lakes, the water level probably fell below the lake sediment surface during periods of increased evaporation. Storm flooding across the mudflats may have also provided inflow, and in very shallow depressions within the mudflats, may have been the only water inflow. From the discussion of the chemical considerations of dolomite formation, it is likely that the protodolomite and high Mg-calcite were precipitated from waters with a Mg/Ca ratio greater than one. With continued precipitation and evaporative concentration, the Mg/Ca ratio would have increased, because the ratio in solution was greater than in the precipitating minerals (approximately one), and may have become high enough to lead to magnesite precipitation. Although magnesite mudstones generally contain minor dolomite, boundaries between magnesite mudstones and underlying dolomite mudstones are sharp. Hence it is likely that much of the magnesium carbonate was precipitated as a primary phase from the lake waters, rather than as a diagenetic replacement of dolomite within the sediment. The initial precipitate from the Mg-CO_3 rich waters, may have been a hydrated Mg-carbonate. During periods when the ponds and lakes were dry, and thin intraclastic lenses and tepees formed, further evaporative concentration, and thus reduced water activity, may have resulted in the inversion of the hydrated Mg-carbonates to magnesite within the sediment. Efflorescent crusts of saline minerals may have formed during exposure, but were removed on flooding of the lake.

The transition from dolomite to magnesite deposition was initiated by the formation of isolated ponds within mudflats, to which there was a continued influx of Mg-CO_3 rich waters. Repeated flooding and drying led to the precipitation of beds up to 1.5 m (maximum preserved thickness) or more in thickness. Because some dolomite was precipitated with the magnesite, the Mg/Ca ratios within the waters of these lakes may have been retained at high levels, allowing the continued precipitation of magnesite. For example with the precipitation of 90 wt% magnesite and 10 wt% dolomite, as long as the Mg/Ca mole ratio of the waters was on average greater than 27/2, continued precipitation would not have decreased the ratio within the precipitating solutions. Variation in the amount of co-precipitated dolomite was probably due to variation of the Mg/Ca ratio of the water. However dolomite and magnesite may not have co-precipitated, but subsequent stirring of surface muds by wind induced wave action would have mixed the two minerals, as has been proposed for the formation of a homogeneous aragonite-hydromagnesite assemblage in one of the ephemeral Coorong lakes (von der Borch, 1962).

Many areas in which magnesite mudstones accumulated were subject to prolonged periods of exposure, during which the water table remained below the sediment surface. This inhibited the formation of surface evaporites. Magnesite mudstones were partly replaced by nodular magnesite, possibly by dissolution and reprecipitation of magnesite within the sediment.

IMPLICATIONS FOR WATER CHEMISTRY

A sequence of widespread and extensive dolomite of "primary" origin with associated magnesite (up to half as abundant as dolomite, although usually forming a smaller proportion), but in which sulphates appear to have been a minor constituent, and in which calcite is absent, is unlikely to form from waters of marine chemistry. Limited Sr-isotope data also supports a non-marine chemistry (Chapter 10). Theoretical considerations indicate that progressive evaporation of seawater should result in precipitation of carbonates (calcite and/or dolomite), calcium sulphate, halite, K- and Mg-bearing sulphates and chlorides, and Mg-salts (Dean, 1978). Hence although widespread dolomite deposits have formed in shallow marine environments, because of the excess of Ca^{2+} and Mg^{2+} over CO_3^{2-} plus HCO_3^-, and the abundance of SO_4^{2-} in seawater, magnesite is only a minor component, and its precipitation is preceded by or occurs in association with Ca-sulphates, or even more soluble salts. Extensive sequences of dolomite and Ca-sulphates have formed in marine basins (Hite, 1970; Gill, 1977), but magnesite, if present, is minor. The mineral sequence forming on evaporation in the system $Ca\text{-}Mg\text{-}CO_3\text{-}SO_4\text{-}H_2O$ at 25°C, is also shown in Figure 32 in Eugster and Hardie (1978). From this it is apparent that extensive deposits of dolomite and magnesite will form without coprecipitation of sulphate minerals only if CO_3 is present in solution in significant amounts as compared with SO_4. However magnesite within the Mundallio Subgroup does not contain associated sulphate minerals, and there is little evidence that they were ever present.

The SO_4^{2-} content of interstitial pore waters and anaerobic bottom waters within closed basins, is lowered by sulphate reducing bacteria, and H_2S and S^{2-}, which is often precipitated as sulphides, are produced (Jones, 1966; Jones and van Denburgh, 1966; Neev and Emery, 1967; Deelman, 1975b). CO_3^{2-} is also produced by this process and may be precipitated as carbonates. The H_2S may be recycled as sulphate after it diffuses into the oxidizing zone. In deep evaporitic basins (e.g. the Dead Sea), this process may result in some removal of gypsum and precipitation of aragonite (Neev and Emery, 1967). However in shallow lakes and exposed sabkhas with hypersaline waters from which magnesite and gypsum would precipitate, this process is unlikely to completely suppress gypsum precipitation if the waters present have a marine chemistry. In addition, dolomite mudstones in this sequence contain only minor pyrite, and it is rare in magnesite mudstones.

Dilute continental waters, and the saline brines produced from them on evaporation in lakes and closed basins, exhibit a wider range of composition than seawater, and hence produce a variety of mineral suites as a result of evaporative concentration and precipitation. Most dilute continental waters contain Ca^{2+} and HCO_3^- as the major ions (Garrels and Mackenzie, 1971). However variation in the proportion of these and the other ions present, leads to vastly different sequences of precipitated minerals, and different compositions of the resultant brines (Hardie and Eugster, 1970; Eugster and Hardie, 1978; Eugster and Jones, 1979). These authors illustrate the main evolutionary paths on evaporation, during which water composition varies as a result of mineral precipitation, selective dissolution of efflorescent crusts, exchange and sorption reactions on active surfaces, degassing and redox reactions (Eugster and Jones, 1979). Because alkaline earth carbonates are the first minerals to precipitate, the ratio of $Ca^{2+} + Mg^{2+}$ to HCO_3^- in the dilute inflow waters, significantly influences the evolutionary path of the resultant brine on evaporation (Eugster and Hardie, 1978). They consider three types of dilute inflow waters:

1. $HCO_3^- \gg Ca^{2+} + Mg^{2+}$;
2. $HCO_3^- \ll Ca^{2+} + Mg^{2+}$;
3. $HCO_3^- \geqslant Ca^{2+} + Mg^{2+}$.

These lead to different mineral assemblages and brine types on evaporation. Although alkaline-earth carbonates are the initial precipitate in all cases, only for the third type, will the amount of carbonate to precipitate be extensive. Often Ca^{2+} is greater than Mg^{2+} and hence calcite is the first mineral to precipitate, but as the Mg/Ca ratio increases, Mg-calcite, protodolomite and eventually magnesite will be precipitated. Because of the significant carbonate alkalinity of this water type, protodolomite may form at fairly low salinities and low Mg/Ca ratios.

Modern examples fed by waters of type 3 include Deep Springs Lake, Lake Chad, Lake Magadi and Lake Balkhash. Following alkaline-earth carbonate precipitation, Na-CO_3 (-SO_4-Cl) rich solutions develop, and Na-carbonates often precipitate. The Green River Formation was deposited in an extensive alkaline lake, in which calcite and dolomite precipitation was followed by trona precipitation, as Na-CO_3 brines developed. Ephemeral lakes associated with the Coorong are one of the few examples in which dolomite and magnesite are present without more saline minerals. During summer evaporation, concentrated brines are produced, but retreat of the groundwater level below the lake surface inhibits the surface formation of evaporites (Lock, 1976). Ephemeral lakes which contain dolomite, magnesite and hydromagnesite, are fed by bicarbonate rich groundwaters (von der Borch, 1962). The lakes contain significantly higher "total carbonate" (bicarbonate plus carbonate) than sea water, as well as having higher Mg/Ca ratios. The total carbonate content is generally much greater than the Ca content, although it is usually less than the total of Mg + Ca (von der Borch, 1962). This is also apparent for analyses of Kingston Lake which contains dolomite and calcite in surface sediments (Skinner, 1963) and in which the Mg^{2+} + Ca^{2+}/CO_3^{2-} ratio of the lake waters is less than in seawater. In particular the Ca^{2+} content is sometimes less than CO_3^{2-}.

Hence the dilute waters entering the basin in which the Mundallio Subgroup was deposited contained Mg^{2+}, Ca^{2+} and HCO_3^{2-} as major ions, with the Mg^{2+} plus Ca^{2+} content approximately balanced by HCO_3^{2-}. Because of the absence of calcite, the Mg/Ca ratio must have exceeded one. The high Mg content is also supported by the apparent uptake of Mg by clay minerals within shales during early diagenesis (Chapter 7). The sulphate content of the waters relative to carbonate, was probably lower than that of seawater, and some sulphate removal by bacterial reduction within the basin may have maintained the sulphate concentration below gypsum saturation. However the inflow water composition probably varied slightly between areas. For example, in the Mundallio Creek area a higher sulphate concentration than elsewhere is indicated by some evidence of the precipitation of Ca-sulphates in this area.

If the waters entering the "Mundallio basin" were derived solely from the surrounding land areas by surface runoff and groundwater in flow, their composition would have been determined by the gross lithology of the drainage areas, and the time of contact with these rocks (Jones and van Denburgh, 1966). Weathering of silicate minerals in igneous and metamorphic rocks produces carbonate rich waters, with the cation composition depending on which silicates are involved. Ca, Na and Mg are usually the major cations. Weathering of sediments with evaporites introduces SO_4^{2-} and Cl^-, whereas solutions derived from pyritic shales and mineralized terrains will also introduce SO_4^{2-}. Weathering of silicate minerals involves the attack by CO_2 dissolved in water:

primary silicate + H_2O + CO_2 = clay + cations + HCO_3^- + SiO_2

(Jones, 1966; Garrels and Mackenzie, 1971). Mafic mineral phases with high Si/Al are preferentially attacked and their weathering introduces silica in solution, whereas weathering of feldspars will contribute little, if any silica (Jones, 1966). The diagenetic silica within carbonates was probably introduced to the basin in solution following weathering of mafic silicates. Sandstones with the Mundallio Subhroup are sub-arkosic and arkosic, and are probably largely recycled. Hence chemical weathering of non-mafic silicates may have been limited. In contrast mafic minerals such as biotite are rare in this sequence.

The source areas for this basin included high grade metamorphic and igneous terrains with both acidic (predominant) and basic components, although clastic sediments, dolomites and basic volcanics of the underlying Callana Group may also have been exposed in some areas adjacent to the basin. Hence the waters generated would have contained CO_3^{2-} as the dominant cation. However waters with Mg more abundant than Ca are produced only if there is a significant contribution by ultramafic rocks. In modern lakes, generation of waters with Mg > Ca is generally achieved by calcite precipitation. This process has also been applied to the Wilkins Peak

Member of the Green River Formation (Eugster and Hardie, 1975). Calcite was rapidly precipitated as a result of the capillary evaporation of groundwaters in alluvial fans adjacent to the Green River basin, while dolomite was precipitated on mudflats and in the playa lake, where waters were more saline and Mg-rich. During deposition of the lower Mundallio Subgroup, permanent rivers from the basement hinterland may have flowed directly into different areas of the basin. However in the upper part of the subgroup, in which magnesite largely occurs, inflow into much of the basin may have been by local surface runoff and groundwaters. Hence precipitation of calcite in areas adjacent to the basin, analagous to that which occurred within the Green River Basin, may have resulted in inflow waters being more Mg-rich when they reached the basin. However such a hypothesis cannot be proved, because there is no preserved record of the Mundallio Subgroup, other Burra Group sediments or equivalent continental deposits outside the present outcrops of folded Adelaidean sedimentary rocks. Hence, although it is clear that the basin waters were enriched in Mg relative to Ca, the source of the Mg is not certain. A similar problem exists for the groundwaters entering the ephemeral Coorong lakes.

DISCUSSION

The uniform carbonate mineralogy of dolomite and magnesite, and the similarity in the carbonate facies over wide areas of the basin, indicates uniformity of sedimentary processes and water chemistry throughout a very shallow basin, at least 600 km in length (and probably extending another 150 km to the Peake and Denison Ranges) and 100-200 km in width. Magnesite is present, although often in minor amounts, in all areas of the basin, with the exception of the southeastern most part, and the area west of Yacka, where terrigenous facies dominate. Hence most inflow waters in all areas must have had similar compositions, although the Mg/Ca ratio may have varied, producing different amounts of magnesite in different areas. This requires similar rock types in source areas, which appears to have been the case, and similar climatic conditions and weathering processes. Hence this sequence was deposited in an extensive, shallow lake or inland sea fed by Mg^{2+}, Ca^{2+}, CO_3^{2-} bearing waters, and probably had limited or no connection with the open ocean, a possibility which has been previously suggested by Spry (1952). Forbes (1961) proposed that deposition occurred in marine and paralic environments, with magnesite forming due to the influx of alkaline continental waters which reacted with seawater in marginal areas. However magnesite was deposited as mudstones in most areas of the basin (Table 6.1).

A more complex model has been proposed by Murrell (1977) who studied the Burra Group in the Willouran Ranges. This model involves mixing of three water types:

(a) saline basin water related to oceanic water but enriched in Mg by mixing of water with type (b);

(b) hypersaline water enriched in Mg, Na, SO_4, Cl following gypsum precipitation on a shallow shelf isolated from a continuous deeper basin, and the reflux of these brines into the deeper basin;

(c) river water containing Ca, HCO_3.

This model is based on the discussion of carbonate shelves with adjacent refluxing evaporite basins of Hite (1970). However Hite's model is not compatible with the interpretation of the Mundallio Basin as a very shallow basin, consisting at times of a series of shallow lakes, and in which most areas of the basin were subjected to periods of exposure. Because of the very shallow widespread nature of carbonate mudflats, it is unlikely that two-way reflux of waters between marginal evaporite basins and the basin proper, would have operated. In addition, the contribution of sulphate from (a) and (b) would have been expected to produce greater evidence of sulphate deposition. The carbonate and evaporite mineralogy produced would have been more variable than that recorded within the Mundallio Subgroup, which is notable for its simple, although rather unusual nature, and similarity over wide areas, and for the similarity of depositional processes which produced it.

Within the remainder of the Burra Group, dolomite is generally the only carbonate mineral present, although rare magnesite is present in the River Wakefield Subgroup in the Yednalue Anticline. The dolomites are largely mudstones which often contain diagenetic chert nodules. Hence the chemistry of the basin waters may have been similar throughout deposition of the Burra Group, but the basin was shallowest during deposition of the Mundallio Subgroup resulting in the widespread deposition of magnesite. In contrast with dolomites of the Mundallio Subgroup which appear to be "primary" precipitates, formed in a lacustrine environment (Coorong-type dolomite, von der Borch, 1976), dolomites in the Callana Group contain abundant evidence of former evaporites in the form of gypsum pseudomorphs, and chert nodules after anhydrite nodules, and show indications of being early diagenetic dolomites analogous to those forming in the Persian Gulf (von der Borch and Lock, 1979).

These differences reflect climatic variations as well as changes in water chemistry. During deposition of the Mundallio Subgroup, a humid climate must have prevailed in the source areas, which may have had an elevated topography, so that the inflow into the basin was sufficient to maintain a permanent water body. However on the basin floor and in the immediately surrounding areas, conditions were more arid. This led to evaporative concentration in the shallowest areas, and magnesite precipitation, while in the immediate hinterland, permanent rivers were not always present and dune fields developed. However evaporation was not intense enough for the widespread formation of saline minerals, although surface crusts which were subsequently redissolved, may have formed in exposed areas. Hence it is likely that the basin was located outside tropical areas. Conditions may have been somewhat similar to those described for the Green River Basin during deposition of the least saline parts of the Wilkins Peak Member, and the Laney Member, although lake waters were more magnesium rich than in the Green River Basin (Eugster and Hardie, 1975, 1978; Surdam and Stanley, 1979).

SUMMARY

During deposition of the Mundallio Subgroup, a shallow, elongate, intracratonic basin existed in the area of the Mt. Lofty and Flinders Ranges, and was the site of both clastic and carbonate deposition. The palaeogeography of this basin has been summarised in Chapter 8. This basin had a non-marine chemistry, and limited or no connection with the open ocean, particularly during deposition of the upper part of the Mundallio Subgroup. This resulted in the deposition of dolomite and magnesite carbonate sediments, without associated deposition of limestones. Within the basin, there was little variation in water depth over wide areas (see also Preiss, 1973), and environments which experienced subaerial exposure, developed at times in most areas.

Dolomite was derived largely from the inorganic precipitation of mud and peloids within the basin, with the initial precipitates probably consisting of protodolomite and Mg-calcite which subsequently inverted to dolomite during very early diagenesis. The dolomite mud frequently experienced reworking and transportation before final deposition. Erosion of stromatolitic dolomites, and lithified dolomite crusts in exposed areas, also supplied dolomite mud and peloids for deposition at other locations. Magnesite mud, which may have initially precipitated as hydrated Mg-carbonates, experienced only minor stirring by wave action, and was deposited in essentially the same site as that of the initial precipitation. Prolonged periods of exposure led to the lithification of the magnesite muds, following which they were extensively eroded into intraclastic beds.

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