SEDIMENTOLOGY AND MINERALOGY OF SOME JURASSIC-CRETACEOUS SEDIMENTS OF THE SOUTHERN EROMANGA BASIN

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awarded 11.4.78

This thesis is the result of research undertaken submitted by the author in part fulfillment of the requirement for the Master of Science Degree in the University of Adelaide 1988.
STATEMENT

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TANH VAN ĐOAN
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CHAPTER 1. INTRODUCTION

1.1 LOCATION OF THE AREA

The Eromanga Basin is one of the most extensive sedimentary basins in Australia. It is located in the centre of the Australian continent and stretches from about 20° to 32° south and from 132° to 147° east and has an area of approximately 1.5 million km². The northern part of the Eromanga Basin is separated from sequences of similar age in the Carpenteria Basin by the Euroka Arch and the eastern part is separated from the Surat Basin by the Nebine Ridge (Fig.1). The Jurassic-Cretaceous sediments of interest in the southern Eromanga Basin include the Bulldog Shale, Cadna-owie Formation, and Algebuckina Sandstone. The Bulldog Shale and facies equivalents which contain both lonestones (i.e. erratic boulders) and boulder conglomerate, are the subject of this study. These are regionally distributed throughout the centre-north of South Australia and in western New South Wales (Flint et al., 1980).

1.2. AIMS OF STUDY

The aims of this study are as follow:

1. To determine the geological history of the Bulldog Shale and related units in the Eromanga basin through field observation.

2. To determine whether the environment of deposition of the Cretaceous clast-bearing shales included glaciation, through the use of scanning electron microscopy on quartz-sand surface textures and clay minerals.

3. To determine the mineralogical composition, in particular the clay mineralogy, and locate the source region for the sediments.

4. To elucidate the diagenesis and burial history and the formation of authigenic minerals in these sediments.

The first part of the study involved section measurement and outcrop examination in parts of the central north of South Australia. In this study, stratigraphy, sedimentary features, fossils and lithologies were examined by field study methods. Samples were collected for examination of mineralogy and quartz sand surface features. Mineralogical studies were carried out by X-ray diffraction, petrography, scanning electron microscopy and electron microprobe analysis. Quartz sand grains were studied and particular attention was paid to their surface textures. The results of samples studied are presented statistically in histograms and tables. A comparison was made between the area studied and various world-wide localities described in the literature.
Pre-Mesozoic basement.

Jurassic-Cretaceous basin margin

Basement ridges defining

Jurassic-Cretaceous basins.

0 100 200km.

SCALE

LEGEND:

Figure 1: Regional setting of the Eromanga Basin
(after Exon and Senior, 1976)
1.3. PREVIOUS WORK

The search for groundwater led to early geological studies in the Eromanga Basin (e.g., Rawlinson, 1878). Many of these studies were carried out for the Geological Survey of South Australia during the 1890's by H.Y.L. Brown. Water-worn boulders in the Bulldog Shale were first reported by Brown (1894) and his work from 1899 to 1908 sparked curiosity about the origin of these boulders (David, 1906). Woolnough and David (1926) noted the presence of apparent glacial boulders in the Upper Cretaceous Winton Beds. These boulders showed faint marks which resembled glacial striae and they suggested these had been largely erased by further aqueous action. They also noted fragments of quartz porphyry that have similarities with Proterozoic volcanic rocks found in the Gawler Ranges, South Australia.

Jack (1931) described the stratigraphy and the structure of the Great Artesian Basin. This work provided a basis for regional mapping and led to an approximation to the current lithostratigraphic division. The first palaeontological work was by Glaessner and Rao (1955), who dated the plant-bearing strata at Mt. Babbage in the Flinders Ranges as early Cretaceous in age. Publication of the Gairdner and Moolawatana 1:250,000 geological maps in 1961 added significantly to understanding of the local stratigraphy. However, the present stratigraphic nomenclature was established by Forbes in 1965 when he published the 1:250,000 Marree geological map. From this work, Forbes offered a comprehensive account of Precambrian, Mesozoic and Tertiary stratigraphy. Ludbrook (1966) published a major paper on the biostratigraphy of the Mesozoic sequence in the Southern Eromanga Basin. Freytag et al. (1967) produced a major assessment of the Jurassic-Cretaceous successions in the Oodnadatta 1: 250,000 geological map. Correlations were made between type sections in the Oodnadatta and the Marree area which were described by Forbes (1966) and Ludbrook (1966). Wopfner et al. (1970) made a major regional assessment of the Jurassic-Cretaceous sequence on the southern margin of the Eromanga Basin.

Exon and Senior (1976) correlated the Mesozoic sequences of the Eromanga and Surat Basins. This work detailed the sedimentary environments of the regional Jurassic-Cretaceous succession in sedimentary basins of central Australia.

Flint et al. (1980) reported on fossiliferous Devonian boulders in the Bulldog Shale of central-north South Australia. In a discussion of the Jurassic-Cretaceous boulder shales in the southern margin of the Eromanga Basin, Frakes and Francis (1988) proposed a model for seasonal climates in the Cretaceous. This work gave evidence for high latitude cold regions at sea level during the Cretaceous.
CHAPTER 2. REGIONAL GEOLOGY

2.1. STRATIGRAPHY

Whitehouse (1954) suggested the name "Great Artesian Basin", which is applied both in the sense of a geological structure and as a hydrological basin, to include what is now known as the Eromanga sedimentary basin, plus the Surat Basin and the Carpenteria Basin. The stratigraphic nomenclature of the Eromanga Basin is summarized in Figure 2. The lower part of the Cretaceous Eromanga sequence has been described by Forbes (1966) as the Marree Formation, including the Algebuckina Sandstone, Cadna-owie Formation and the Bulldog Shale. The Bulldog Shale is overlain by the Coorikiana Sandstone in the southwest Eromanga Basin. Ludbrook (1966), Freytag et al. (1967) and Wopfner et al. (1970) described the sedimentology and stratigraphy and confirmed that the Jurassic-Cretaceous sequence unconformably overlies the Palaeozoic-Precambrian basement in northern South Australia. Forbes (1966) referred to rocks equivalent to the Cadna-owie and the Bulldog Shale in the northern Flinders Ranges as the Pelican Well and Village Well Formation, respectively. The most widely-exposed rocks in the Eromanga Basin are those of the weathered and silicified Tertiary cover.

Ambrose et al. (1979) described the type section of the Algebuckina Sandstone (north of the Oodnadatta Track). The section consists of a basal sequence of kaolinitic cross-bedded sandstones, including highly weathered porphyry, quartz and agate clasts. The upper part of the section consists of well-sorted, gritty sandstone and siltstone. Ambrose et al. also used a silcrete which occurs at the top of the Algebuckina Sandstone to distinguish it from the overlying Cadna-owie Formation. Ambrose et al. (1984) used this silcrete as a criterion to distinguish the two parts of the Mesozoic sequence in the Preliminary Curdimurka 1:250,000 Sheet.

Barker (1983) used the term Marree Formation to describe a transitional sequence in the Coober Pedy area. The sequence contains the agglutinated foraminifera, Texturalia acoonaensis, diagnostic of lowermost Aptian age (Ludbrook, 1966; Lindsay, 1975). The overlying sequence (Marree Formation) was considered by Barker (1983) as equivalent to the Bulldog Shale of a similar sequence described by Pitt (1976).

Benbow (1982), in the Explanatory Notes for the Coober Pedy 1: 250.000 geological map summarized detailed-stratigraphic nomenclature for the Mesozoic lithofacies of the Great Artesian Basin. This is shown in table 1.
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<tr>
<th>CENO.</th>
<th>SOUTH AUSTRALIA</th>
<th>MARREE-COOLEY GENERAL (Forbes, 1982)</th>
<th>WAVERLY MARGIN (Freitag et al., 1966; Wayte et al., 1970)</th>
<th>SOUTH AUSTRALIA</th>
<th>STUART CREEK OPAL FIELD (Vuk, 1978; Niall, 1979)</th>
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<td>ALBIAN</td>
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<td>OODHADATTA FORMATION</td>
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<td>MARREE</td>
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<td>CADNA-Owie FORMATION</td>
<td>MARREE</td>
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**Figure 2.** Stratigraphic nomenclature of the southern Eromanga Basin
<table>
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<th>LITHOLOGY</th>
<th>REMARK</th>
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<tbody>
<tr>
<td>Tertiary</td>
<td>Unnamed (Arkaringa Paleosol, Benbow, 1982)</td>
<td>Kaolinite and bleached profile. Possibly represents multiphase events. Widespread.</td>
<td>less than 60 metres in thickness</td>
</tr>
<tr>
<td>Cretaceous</td>
<td>Bulldog Shale (Freytag et al. 1966)</td>
<td>Grey to dark grey shale and silty shales, minor sandstone lenses, strongly bioturbated minor concretionary limestone with mollusc casts. Basal part of grey shale with thin lenses and silty shales, cone in cone and sandy limestones. Horizons of well rounded cobbles and boulder of predominantly quartzite, acid porphyry and banded chalcedony. Rare quartzite clasts with Devonian bivalves and brachiopods.</td>
<td>Marine, deeply bleached and weathered by multiphase events. Less than 110 metres in thickness</td>
</tr>
<tr>
<td>Cretaceous</td>
<td>Cadnaowie Formation - Mount Anna Sandstone Member (Wopfner, et al. 1970)</td>
<td>Conglomeratic and yellow-brown feldspathic sandstone, containing abundant clasts of quartz and quartzite, porphyritic rhyolites and dacites. Calcareous in part.</td>
<td>Fluvial commonly ferruginised black. Outcrops around the southern margin of the Bulldog Shale. Contact with the Bulldog Shale is sharp. Thickness less than 50 metres</td>
</tr>
<tr>
<td>Jurassic</td>
<td>Algebuckina Sandstone (Wopfner, et al., 1970)</td>
<td>Crossbedded white kaolinitic sandstone, gritty clay and clays. Conglomeratic in part with pebbles and cobbles of rounded to angular, white to blue quartz and various basement lithologies. Dark grey to grey carbonaceous and massive clays, sandy in part, with rare fossil plant stem fragments.</td>
<td>Fluvial Outcrops on the Gawler Block. Thickness less than 60 metres</td>
</tr>
</tbody>
</table>

**TABLE 1** Summary of Mesozoic lithostratigraphy of the Great Artesian Basin (after Benbow, 1982)
2.1.1. Pre-Mesozoic basement.

The pre-Mesozoic basement is found only on the southern margins of the Eromanga Basin, in the Peake and Denison Ranges and in deep bore holes (Fig. 1). The basement consists of Precambrian-Cambrian Adelaidean rocks and late Palaeozoic strata of the Arkaringa and Cooper Basins. Forbes (1966) and Wopfner et al. (1970) described the Burra Group above the Callana Beds and the Wakefield Group near Davenport Spring, south of Lake Eyre. Nicol (1979) described basement rocks of Cambrian age in the area of the Stuart Creek Opal Field. The Andamooka Limestone and Yarra Wurta Shale overlie the Wilpena Group, which is in turn underlain by the Umberatana Group (Aberoona Formation and Elatina Formation). At Trinity Well the basement consists of the Umberatana Group (Tapley Hill Formation and Sturt Tillite). Near Pelican Well the basement consists of the Proterozoic Terrapinna Granite in contact with volcanics (Phillips, 1983). The Umberatana Group unconformably overlies the igneous sequence.

Pre-Mesozoic basement also includes the Permian to Triassic fluvial, lacustrine and possibly restricted marine sediments which form part of the Arkaringa and Cooper Basin. These sequences are unconformably overlain by the Jurassic-Cretaceous sediments of the Eromanga Basin.

2.1.2. Mesozoic sediments

Mesozoic sediments were deposited on the surface depression of the pre-Mesozoic basement. The stratigraphic nomenclature of the southern Eromanga basin in South Australia was summarized in Figure 2.

2.1.2.1 ALGEBUCKINA SANDSTONE

The name Algebuckina Sandstone was first used by Freytag et al. (1967) on the 1:250,000 Oodnadatta geological map and formally defined by Wopfner et al. (1970) when describing the section near the abandoned Algebuckina railway siding. Harris (1970) suggested the age of this sequence was late Jurassic to early Cretaceous (Neocomian) from studies of the microflora.

The Algebuckina Sandstone, including the Mt. Anna Sandstone member, unconformably overlies the pre-Mesozoic basement and represents the basal unit of the southern Eromanga Basin. The rock consists mainly of white kaolinitic medium-grained sandstone and contains conglomerate with well-rounded quartz pebbles at the base. Laterally discontinuous shale and siltstone units commonly display climbing
ripple laminations, and contain plant fossils. There is an absence of marine fauna, suggesting a fluvial environment (Wopfner et al., 1970), Ambrose (1980) suggests in particular a braided stream environment.

The Mount Anna Sandstone Member is part of the Algebuckina Sandstone and reflects a trough of fluvial deposition north of Lake Torrens (Wopfner et al., 1970). The sequence consists of coarse-grained feldspathic sandstone and contains interbedded conglomerate indicating a fluvial-deltaic environment of the latest Neocomian regression (Flint et al., 1980). Forbes (1982) noted the derivation of porphyry boulders from the Proterozoic Gawler Range Volcanics.

2.1.2.2. CADNA-OWIE FORMATION

The name Cadna-owie was also first used by Freytag et al. (1967) and the formation was defined by Wopfner et al. (1970). The type section is located 4 km southwest of the abandoned Algebuckina railway siding. The unit generally disconformably overlies the Algebuckina Sandstone (Wopfner et al., 1970). The contact may be transitional above the Algebuckina Sandstone, or there may be a low angle unconformity (Forbes, 1982). The Cadna-owie ranges in age from Neocomian to Aptian (Early Cretaceous) and represents brackish to marine environments of the first marine transgression in the Eromanga Basin.

The upper part of the Cadna-owie Formation consists of laminated, silty to medium-grained brownish sandstone with intercalations of shale and siltstone. The sandstone is locally calcitic, feldspathic, pyritic and micaceous. Quartzite and siltstone clasts of pebble to boulder size are also distributed throughout the sequence. Carbonaceous shale is locally developed in the sequence.

The common presence of pyrite and iron sulphide facies indicates that much of the sediment was deposited under reducing conditions (Wopfner et al., 1970). The Cadna-owie Formation appears to overlap the Algebuckina Sandstone, and anoxia may have originated during transgression. The Pelican Well Formation, described by Forbes (1966), is the equivalent of the Cadna-Owie Formation in the northern Flinders Ranges.

2.1.2.3. BULLDOG SHALE

The marine Bulldog Shale conformably overlies the Cadna-Owie Formation. It was first described by Freytag (1966) from a section 8 km south of Bulldog Creek, east of the Peake and Denison Ranges. It is considered to be of Aptian to about middle Albian Age (Ludbrook, 1966; Morgan, 1980). However, Alley (1987) in a study of
the palynology of the base of Trinity Well section suggested a marine transgression in that area during the early Neocomian.

Lithologically the Bulldog Shale is a dark grey to black shale and siltstone with thin interbeds of sandstone. Krieg (1982) noted that the base of the Bulldog Shale is a dark, coffee-brown shale containing small lenses of fine to coarse sand. The sediments are locally carbonaceous, silty, glauconitic and pyritic in a shaley-mudstone. The shaley-mudstone contains some concretionary limestone and cone-in-cone structures. The glauco-pyritic shale is indicative of alternating oxic-anoxic conditions (McKirdy et al., 1986). The Bulldog Shale was deposited in marginal marine to shallow marine environments, probably less than 200 metres deep, as indicated by foraminifera and pelecypods.

Boulders of mainly quartzite and siltstone are distributed throughout the Bulldog Shale but concentrated near the base (Forbes, 1982; Flint et al., 1980). Fossilised wood occurs sporadically and large rounded boulders are distributed throughout the sequence.

2.1.2.4. OODNADATTA FORMATION

Conformably overlying the Bulldog Shale is the Oodnadatta Formation (Freytag et al., 1967). The basal Coorikiana Member of the Oodnadatta Formation was described in the area of Coorikiana Creek 40 km southwest of Oodnadatta. Lithologically it is typically a fine-grained quartz, feldspathic and glauconitic sandstone with some coarser-grained pebbly lenses. Thomson (1980) and Forbes (1982) used the name Coorikiana Sandstone for this unit. The overlying Wooldridge Limestone Member of the Oodnadatta Formation consists of calcareous and sandy siltstone with limestone concretions and contains a marine micro- and macro-fauna of late Albian age (Ludbrook, 1966). The age of the Oodnadatta Formation thus is Albian. The environment of the Oodnadatta Formation is shallow marine with a low rate of sedimentation and hardly affected by tectonic instability (Krieg, 1982). The Mount Alexander Sandstone Member is the uppermost part of the Oodnadatta Formation, consisting of dark-grey silty shale and claystone with intercalations of ferruginous carbonaceous and micaceous sandstone. This unit is thought to be equivalent to the basal part of the Blanchewater Formation in the Marree area (Forbes, 1982).

The uppermost Mesozoic unit is the non-marine Winton Formation, which conformably overlies the Oodnadatta Formation and is in turn overlain by Early Tertiary deposits above an erosional surface. The latter are referred to as the Eyre Formation (Wopfner et al., 1974).
2.2. REGIONAL STRUCTURAL GEOLOGY AND TECTONIC SETTING OF THE BASIN

Several subsurface basement rises separate the Jurassic-Cretaceous basins of South Australia, New South Wales and Queensland. The Nebine Ridge separates the Eromanga Basin and the Surat Basin. The Euroka Arch separates the Eromanga basin and Carpenteria Basin. The southern Eromanga margin is limited by the Adelaidean rocks of the northern Flinders Ranges. To the southeast the Eromanga Basin overlies and abuts against the Broken Hill Block; to the northwest it is limited by the Arunta Block and to the northeast by the Mount Isa Block (Fig. 1).

The Eromanga Basin strata generally thicken and deepen from the southern margin in a northerly direction. The structure of the Eromanga Basin is characterised by broad, generally low-amplitude folds and northeast trending faults. A number of breached regional anticlines and troughs can be recognized in the southwest margin and central area of the Eromanga Basin. Many of the structures of the basin are defined by seismic, gravity and core data. Kreig (1982) described the Dahousie Anticline, a broad structure of the southern Eromanga Basin, as a growth structure of Late Cretaceous-Cainozoic age. Seismic profiles show its origin dates from the base of the Bulldog Shale, in the form of a simple, steady, curving surface with several folds and faults superimposed on the anticline form (Kreig, 1982).

In the northeast, there are a number of anticlines, domes, synclines and faults overlying the Cooper Basin. These structures are formed by compaction over basement ridges and troughs which were affected strongly by the Late Palaeozoic orogeny. The trend is generally parallel to the north-northeast direction developed in the Cooper Basin and the basement (Exon and Senior, 1976). A major downwarp movement occurred during the Cretaceous, therefore the area over the Cooper Basin was a depocentre, and the Jurassic-Cretaceous units are generally thin elsewhere.

Deformation of the Mesozoic and Early Tertiary sequence occurs within Cretaceous formation (Exon and Senior, 1976). Most folds have a decreasing amplitude in the younger Cretaceous sequence, and the movements on faults become less. The Tertiary sequences are also folded and displaced in the Dalhousie Anticline region. There is evidence of movement both during and after Cretaceous time (Exon and Senior, 1976; Kreig, 1982).

In general terms, the structural pattern is dominated by normal faults and related anticlines. These anticlines are commonly drape folds over deeply buried basement faulted blocks. There was intermittent movement during sedimentation, which resulted in thinning over these blocks (Exon and Senior, 1976).
Figure 3.1: Sample locations of the area studied in the Eromanga Basin. (Map after Forbes, 1982)

LEGEND:
- Undifferentiated Tertiary and Quaternary sediments
- Jurassic-Cretaceous outcrops
- Palaeozoic
- Late Precambrian-Cambrian
- Older Precambrian and volcanics
- Limits of the Eromanga Basin
- Sample location associated with clast bearing shale
CHAPTER 3. LITHOFACIES DESCRIPTION

3.1. INTRODUCTION

Several outcrops were examined and sections measured from the southern margin of the Eromanga Basin during joint field excursions with fellow researchers. The sections are described in this chapter. The effects of gypsification, silicification and calcification made sections difficult to describe. Moreover, some of the original sedimentary structures are poorly preserved due to bioturbation and compaction. The sections presented are mainly from exposures at Stuart Creek Opal Field, Trinity Well and Pelican Well from the southern Eromanga Basin (Fig. 3.1). In addition, Mesozoic sediments in one bore hole (Toodla #1) were sampled and examined. (chapter 5)

3.2. STUART CREEK OPAL FIELD SECTIONS

Basal sandstone and boulder conglomerates of the Mesozoic sequence crop out at Yarra Wurta Cliff and Stuart Creek Opal Field to the north of Lake Torrens (Fig. 3.2). These are the most southerly known outcrops and mark the boundary of the southern Eromanga Basin with the Willouran Ranges.

In research on the precious opal deposits in the area, Vnuk (1978), Barnes and Scott (1979) and Nicol (1979) made detailed studies of the geological history of the area. The Mesozoic sediments are deeply weathered and have been kaolinised. For this study, three sections were measured in the area. To the NNE of Lake Torrens, Mesozoic outcrops occur as "mesa" and "pimple" land forms which are the result of intense weathering processes. Measured sections from Yarra Wurta Cliff and Stuart Creek Opal Diggings are shown on Figure 3.3.

3.2.1. Stuart Creek Beds

The Stuart Creek Beds (Vnuk, 1978) overlie the Cambrian basement at Yarra Wurta Cliff. The base of the beds is marked by an unconformity and a conglomerate layer on the undulating surface of the Cambrian shale (Plate 1a). At the base is a conglomerate which contains green silt and quartzite clasts of pebble and small cobble sizes in a silty-sandy matrix. The conglomerate beds range from 0.3 to 1 metre in thickness. Above the conglomerate bed is a sequence of dark green shale grading up into a sequence of white or yellow to pale brown, micaceous, glauconite, silty to very fine sand. The sand is generally friable and shows patchy stains of iron oxide. In places, the Stuart Creek Beds contain thin lenses of coarse sand to pebble
Figure 3.2: Location of Measured sections, Stuart Creek Opal Field area (after Johns et al., 1966) Geological Survey of South Australia.

LEGEND:
- Tertiary and Quaternary sediments
- Marree Formation
- Cambrian rocks
- Precambrian rocks
- Measured section
- Creek
Figure 3.3: Correlated sections and sedimentary facies between Yarra Wurta Cliff, Stuart Creek Opal Diggings and Charlie Swamp.
Plate 1a.
Stuart Creek Beds overlie the undulating surface of the Precambrian basement (dash line). The contact is marked by a lag gravel deposit and a thin ferruginous sandy layer. Yarra Wurta Cliff, Stuart Creek Opal Field. Scale bar is 1m.

Plate 1b.
Rounded cobbles distributed throughout the Marree Formation, Stuart Creek Opal Digging.

Plate 1c.
Agglutinated form of foraminifera *Textularia anacooraensis*, Lowermost Aptian Marree Formation, Yarra-Wurta Cliff, Stuart Creek Opal Fields (SEM secondary electron image).

Plate 1d.
Marree Formation at Stuart Creek Opal Diggings. Note heavy gypsification in fractures and joints and a red (dark) zone of biomottling.

Plate 1e.
Scour and fill channel of the upper part of the Pelican Well Formation, Pelican Creek, Pelican Well.

Plate 1f.
Intermittent conglomerate layers from the upper part of the Pelican Well Formation, Pelican Creek, Pelican Well.

Plate 1g.
Tidal beds from the Trinity Well Sandstone. Pelican Well. Note: The herring-bone cross stratifications.

Plate 1h.
Crossbedding in the Trinity Well Sandstone, Pelican Well.
conglomerate or patches of dark brown silty sand. The sequence is overlain by the Bulldog Shale which contains foraminifera of lowermost Aptian age (Ludbrook, 1966; Lindsay, 1975). Thus the age of the Stuart Creek Beds is Aptian or older.

3.2.2. Bulldog Shale (Marree Formation equivalent)

The Bulldog Shale overlies the Stuart Creek Beds in the Yarra Wurta Cliff and Stuart Creek Opal Digging sections or laps directly onto the undulating surfaces of the Precambrian basement. The base of the Bulldog Shale is marked by a conglomerate layer. The conglomerate layer consists of quartzite, siltstone, and shale pebbles with closely-packed coarse sand and a silty matrix.

Above the conglomerate layer is a sequence of green-grey to pink-white poorly-bedded kaolinite-and alunite-rich silty shale with erratic boulders or lonestones, up to 1 metre in diameter. Most of the boulders have rounded and polished surfaces and are randomly oriented. They are distributed throughout the formation (Plate 1b). A rounded quartzite boulder was measured at 2.65x1.00x0.80 metres by Nicol (1979).

The silty shale facies is bioturbated and mottled. These structures were described and discussed by Barnes and Scott (1978, 1979) and Nicol (1979). Nicol (1979) used the term biomottling to describe this texture. The present author also uses the term "biomottling" sensu Nicol.

Dark green-grey layers contain agglutinated brackish-water foraminifera (Plate 1c). The foraminifera were identified as *Textularia anacoooraensis* (Dr. B. Mc.Gowran. pers. coms.). According to Ludbrook (1966) and Lindsay (1975) this fossil is diagnostic of the lowermost Aptian. At Stuart Creek Opal Diggings, the Bulldog Shale is deeply weathered and contains secondary gypsum and halite. These minerals infill joints and fractures (Plate 1d) or have penetrated the unconformity surface between the basement and Mesozoic strata.

The Bulldog Shale is overlain by crossbedded, (fluvial) sandstone of Tertiary age. This starts at the base with a medium to coarse kaolinitic sandstone layer. The sediment source was possibly from the Marree Formation.

3.3 PELICAN WELL SECTIONS

The Mesozoic sequence in this area was described by Forbes (1966) when he published the Marree 1:250,000 geological map and in a review paper in 1982. Sequences in the area (between Pelican Well and Village Well, around Western Spur, west of Mt Freeling; Fig. 3.4) were also correlated. In the present study several sections were measured along Pelican Creek. These sections were PWN₁, PWN₂,
PWN₃ and VW. The sections were also correlated and are shown in figure 3.5. Exposures of these sediments are characteristically infrequent and poor. The major Mesozoic sequence occurs in segments and beneath flat plains. Since no fossils were found the sections were correlated on a lithological basis.

3.3.2. Pelican Well Formation

A section (approximately 25 metres) of the Mesozoic sediment was measured in the Pelican Well area, along Pelican Creek. This sequence lies unconformably on Precambrian basement at PWN₁ and PWN₂. The basal part consists of a conglomerate layer representing a lag gravel deposit, and is laterally continuous throughout the section. It contains rounded clasts such as quartzite, siltstone and shale derived from the Precambrian basement (Adelaidean), within a grey, yellow silty sandy matrix. The conglomerate layer ranges in thickness from 0.8 to 1.3 metres. At the contact plane with the Adelaidean there is a thin (few centimetres), ferruginised sandy layer. This layer is a distinct band which can be easily recognized at the unconformity surface.

Above the basal conglomerate layer is a sequence of green, grey, yellow, silty sandstones containing pebbles, cobbles and boulders. The boulders are concentrated near the base of the sequence. However, there are large numbers of clasts randomly distributed throughout the sequence. The largest measured boulder in the Pelican Creek section is about 1.5 metres in diameter.

Most of the silty layers have small bioturbated traces or burrows, which are infilled with a darker silty-very fine sandstone. Organic matter is abundant throughout the section. However, no fossil plants were found. There are several conglomerate layers and channel fills, and fining-upward sequences were also recorded near the upper part of the sequence (Plate 1e, 1f).

3.3.3. Trinity Well Sandstone

Above the dark green-grey silty bioturbated sequence is a unit of medium to coarse feldspathic-kaolinitic sandstone. Its colour varies from white to grey and it is stained with iron oxides. Good outcrops of the Trinity Well Sandstone are seen at PWN₃. At Pelican Creek, the type section, the unit consists of large cross bedding sets and tidal units (Plate 1g) (Lemon, 1988) and can be interpreted as indicating an estuarine environment (N. Lemon, pers. coms.). Cross-bedding is the most prevalent structure in the Trinity Well Sandstone (Plate 1h). The thicknesses of cross beds range from 50 cm to several metres. The cross beds are predominantly planar but some trough cross-bedding was seen at the Pelican Creek Section. Sedimentary structures in this unit
Figure 3.4: LOCATIONS OF MEASURED SECTIONS, TRINITY WELL, PELICAN WELL AND VILLAGE WELL, EAST OF MT.FREELING. (after Forbes, 1965)

LEGEND:
- Tertiary and Quaternary sediments
- Marree Formation
- Pelican Well Formation
- Village Well Formation
- Precambrian rock
- Well
- Creek
- Measured section
Figure 3.5: Correlated sections of Pelican Well, Trinity Well Village Well and Western Spur
include scour-based channels and cross-bedded sandstone yielding paleocurrent directions ranging from 240° to 270°.

The basal part of this unit is marked by an intermittent boulder-conglomerate layer. Boulders and pebbles characteristically are well-rounded with occasional very angular clasts. The clasts have a similar composition to the lag gravel deposit above the Adelaidean. At the Village Well and Western Spur sections, there is an equivalent body of sandstone overlying the upper part of the Pelican Well Formation. The basal part is also marked by a conglomerate layer resting on an unconformity surface. In thin section the Trinity Well Sandstone consists of quartz, feldspar, muscovite and kaolinite and other minor components including biotite and heavy minerals (Plate 2a, 2b).

3.3.4. Marree Formation (Bulldog Shale equivalent)

Above the Trinity Well Sandstone member is the Marree Formation, which is a weathered white to grey silty-shale which is overlain by Tertiary silcrete. The Marree Formation is an onlap unit of the last marine transgression in the area. There are well correlated sections of the Mesozoic sediments in this area (Pelican Well, Village Well and Western Spur area; Forbes, 1982). Correlated sections are shown on figure 3.5. Therefore mineralogical studies and environmental interpretation of the studied area will be discussed together with the Trinity Well sections.

3.4. TRINITY WELL SECTIONS

Trinity Well is located about 20 km NE of Pelican Well (Fig. 3.4) Most of the geological studies in the area have been undertaken by Forbes of the (SADME) Geological Survey of South Australia (1966; 1982). On the MARREE sheet the name Trinity Well Formation was used for sandstone in the upper Marree Formation. Sections were measured in an area of 25km².

3.4.1. Pelican Well Formation

The Pelican Well Formation overlies Precambrian basement above an unconformity surface.(Plate 2c, 2d ) The base of the type section is marked by a conglomerate layer (Plate 2e) but elsewhere the basal part of the formation consists of a dark brown feldsparitic coarse sandstone. Lag gravel deposits above the unconformity surface are one to two metres in thickness. At Recorder Hill, the Pelican Well Formation consists of a yellow to grey pebbly mudstone, which in part is silicified, calcified or gypsified. The clasts vary in size from pebbles to boulders and the biggest
are up to 2 metres in diameter. In thin section, the calcareous conglomerate samples consist of rounded-to-subrounded pebbles, cobbles and coarse sands in a cemented calcareous silt matrix. The clasts consist of siltstone, quartzite and shale, possibly from the Precambrian basement. Correlated sections are shown in figure 3.5.

The bioturbated mudstone contains small mixtures of glauconite grains, and darker mud layers or sparse pyrite peloids. At the bottom, the sequence is marked by a calcareous pebbly sandstone with worm burrows (Thallasinoides type). In thin section, the calcareous pebbly sandstone consists of rounded-to-subrounded pebbles, cobbles and coarse sands in a cemented calcareous silt matrix. The clasts consist of siltstone, quartzite and shale, possibly from the Precambrian basement. Bioturbation structures are patchily distributed and the burrows vary from 2 mm to a few millimetres in diameter. Thus, there is evidence of faunal activity in the sediments (fecal pellets and bioturbation). However, no fossil fauna was found in the Pelican Formation. Fossil leaves and pollen were found in the middle part of the Recorder Hill section. Alley (1987) studied the palynology of the Pelican Well Formation and concluded that the sequence contains pollen diagnostic of early Neocomian age. In the uppermost parts of the Pelican Well Formation carbonaceous material appears in calcareous, hummocky cross-stratified beds (Plate 2f) and cone-in-cone structure is seen in the TR10, TR6, CMS1 and CMS2 sections.

3.4.2. Trinity Well Sandstone

Conformably overlying the Pelican Well Formation is the Trinity Well Sandstone Member. The basal part of this unit is also marked by a conglomerate layer. The sandstone is fine-to-medium grained, poorly sorted and contains abundant sub-angular to angular quartz grains. At the base, there are several noticeably coarser, occasionally pebbly and bouldery layers. Matrix material is usually silty but less abundant and imparts a porous nature to the sand body. Minor components include coarse mica flakes and weathered feldspar grains. Most of the sandstone is cross-bedded with current directions ranging from 145° to 295°, but in general, most of the paleocurrent directions trend toward the north west. No fossils were recorded in the Trinity Well Sandstone Member in this area.
Plate 2a.
General view of the Trinity Well Sandstone in thin section. Quartz (Q) and feldspar (F) are the predominant features in the sample, Trinity Well Sandstone, Pelican Well.

Plate 2b.
Closer look at a detrital grain of microcline (M); the grain is overgrown by authigenic microcline, Pelican Well.

Plate 2c.
Mesozoic sequence unconformity overlies a dolomitic sandstone unit of the Precambrian basement at Trinity Well.

Plate 2d.
Angular unconformity between the Precambrian basement and the overlying Mesozoic sediment, Trinity Well

Plate 2e.
Basal part of the Pelican Well Formation is marked by an unconformity surface and a conglomerate layer at Trinity Well.

Plate 2f.
Fine-grained sandstone forms hummocky cross-stratified beds. The rock contains calcareous-cement. Trinity Well Sandstone Member, Trinity Well Section.

Plate 2g.
Kaolinite (K) infills pore space in the Trinity Well Sandstone, Pelican Well Formation.

Plate 2h.
Rounded glauconite (G) grain in the Bulldog Shale, Toodla #1, Oodnadatta
3.4.3. Marree Formation (Bulldog Shale equivalent in age)

The Marree Formation unconformably overlies the Trinity Well Sandstone. The unconformity surface is found at the base of a 1 or 2 metres layer of red iron oxide-cemented sandstone. These indicate an old weathering surface of possible early Aptian age (N. Lemon pers. coms). However, at TR₁ the red sandstone layer was eroded and Bulldog Shale lies on the unconformity surface, suggesting the red sandstone. The Marree Formation onlaps the Trinity Well Sandstone in the area. The Marree Formation consists of dark grey to pale grey glauconitic shale with some pyritic components. Overlying the Marree Formation is the Tertiary Telford Gravel. The thickness of the gravel ranges from one to several metres.

Lithologies of the Pelican Well and Trinity Well Sections are similar. The dominant minerals in the silty shale facies of both the Bulldog Shale and the Pelican Well Formation in the Trinity Well area include quartz, glauconite, authigenic potassium feldspar, illite/smectite and pyrite. In addition, there are secondary minerals such as gypsum and halite. Detailed sample descriptions are given in appendix 1.
CHAPTER 4. SCANNING ELECTRON MICROSCOPY OF QUARTZ-SAND SURFACE TEXTURES

4.1. INTRODUCTION

The study of sand-grain surface textures has become an established technique in the study of clastic sediments. Through this method lithological characteristics which are extremely important in environmental reconstruction can be recognized, leading to improved interpretation of ancient environments.

Superior depth of focus and wide ranges of magnification are the great advantages of the scanning electron microscope (SEM). Indeed, the use of SEM to study sand-grain surface textures is a new sedimentological method providing valuable information for environmental reconstruction (Krinsley and Marshall, 1987). The instrument is proving more useful over time. With scanning electron microscopy of sand-grain surface textures, as with many other forms of geological work, sampling and laboratory work must be done carefully. The study of sand-grain surface textures must be used in conjunction with statistical methods in recognizing depositional environments.

4.2. PREVIOUS WORK

Several early papers were published on the electron microscopic examination of sand grain-surface textures (e.g., Beiderman, 1962; Krinsley and Takahashi, 1962a, 1962b, Krinsley et al., 1964). These workers were concerned with finding surface textures that were indicative and diagnostic of particular environments.

Research increased when the scanning electron microscope (SEM) became popular in the late sixties-early seventies. Krinsley and co-workers systematically studied quartz sand. Quartz sand grains from various environments were collected and experimentally examined in the laboratory. The results gave correlations of textures with environments, and diagenetic, mechanical and chemical transportation factors were recognized. The method was used widely in the late seventies and early eighties (Higg, 1978; Whalley, 1978; Cheng, 1978; Middleton and David, 1979; Whalley and Langway, 1980; Mazzullo et al., 1982). These works have contributed significantly to environmental reconstructions. There has been a recent revival of research on sand-grain surface textures (Sharp and Gomez, 1986; Rai, 1987) and the most important was the publishing of "Clastic Particles" by Mazzullo and co-workers (1987). This volume included experimental work and examined clastic particles in various environments. It was also confirmed that the examination of sand surface texture is
one of the modern methods which shows remarkable results compared with traditional sedimentological techniques.

The quartz-sand surface texture techniques are most effective when combined with statistical analysis (Margolits and Kennett, 1971; Baker, 1976; Strass, 1977; Bull, 1978; Higg, 1978; Hill and Nadeau 1984; Rai, 1987). The statistical method gives more reliable results and minimises erroneous conclusions about characteristics inherited from earlier depositional events. But statistical analysis cannot produce new insights unless the origin and development of sand surface textural data are well understood.

4.3. SCANNING ELECTRON MICROSCOPE METHODS

4.3.1. Method and sample preparation

Samples were collected from both outcrops and cores from various locations (Fig. 3.1). These were then dispersed and washed clean of clays and cements. Sand grains from the 65-200 µm fraction were oven dried at 60°C until entirely dry. Their general features were first studied with binocular microscope in order to estimate percentages of polycrystalline grains and the dominant mineralogy in the size fraction. About 5 to 15 gr of the sand fraction (65-250 µm) were boiled for ten to fifteen minutes in concentrated hydrochloric acid to remove carbonate, then rinsed thoroughly with distilled water. Iron oxide was removed when necessary by boiling in 20% stannous chloride solution for 20 minutes. Organic materials were removed by a strong oxidizing solution of 1.5 gr of potassium dichromate and potassium permanganate dissolved in 15 ml of concentrated sulfuric acid. In using this technique to remove organic debris, samples required further boiling with 20% hydrochloric acid to remove stains caused by the potassium dichromate and potassium permanganate. The samples were then washed in distilled water several times and oven-dried until entirely dry.

Quartz grains from this material were randomly selected and mounted on aluminium specimen stubs, which were coated with double-sided adhesive tape. The stubs were then coated under a standard vacuum evaporator. The use of gold-palladium for coating tended to give the best results.

4.3.2. Interpretation of quartz sand grains

Quartz-sand surface textures are observed on the premise that grain morphology is diagnostic of the environmental deposition history from which the grain has come (Krinsley and Doornkamp, 1973). The environment types that can be differentiated on
this basis are shown in figure 4.1. This conceptual model shows the interrelation among surface features, source materials, and four major environments: glacial, aeolian, littoral and diagenetic environments. Krinsley and Doornkamp (1973), and Krinsley and Marshall, (1987) indicated that interpretation of environment must follow certain rules:

1. No single feature seen on the surface of one or more sand grains is sufficient to make an interpretation of the environmental conditions.
2. Any diagnostic characteristic must be observed on several grains.
3. Interpretations must be based on statistical methods to ensure the characteristic observed represents a statistically significant sample.

In addition, overprinted features must be accounted for (to be discussed in 4.5.6).

Many workers confirm that common features have been found among grains from many environmental conditions (Margolis and Kennett, 1971; Krinsley and Doornkamp, 1973; Margolis and Krinsley, 1974). These include conchoidal fracture patterns, upturned cleavage plates and V-shaped patterns. These features are produced by chemical and mechanical action. Therefore the interpretation of quartz-sand surface textures becomes an assessment of the presence or absence of those features and the degree of natural alteration of these features.

4.4. SUMMARv OF QUARTZ-SAND SURFACE TEXTURES

4.4.1. Conchoidal fracture patterns

Conchoidal fracture patterns consist of a series of irregular broken surfaces, commonly stepped in a series of shell-like ridges or arc-step markings. Krinsley et al. (1964) showed that these are probably caused by acoustical wave phenomena. Margolis and Krinsley (1974), confirmed by Sharp and Gomez (1986) showed that as quartz is compressed in static fatigue tests, cracks may propagate in irregular motion, each producing a microfracturing event associated with ultrasonic radiation. The formation of conchoidal fracture patterns is strongly controlled by crystallographic orientation. The size of the area affected can vary from 1 μm to as much as one third of the grain surface. Conchoidal fractures of varying dimensions and abundance have commonly been observed on sand grains from many environments: fresh and weathered granite or granite gneiss, glacial environments, aeolian environments, colluvium and also by mechanically fracturing of quartz in the laboratory.
Fig. 4.1 Conceptual model showing interrelation among surface features and environmental contents

**SOURCE MATERIAL**
- Angular and poor sorting
- Conchoidal fractures
- Diagenetic features inherited from earlier sediments
- Crystal moulds
- Solution and precipitation of silica

**GLACIAL**
- Extremely angular and poorly sorted grains
- Large conchoidal fractures greater than 1 μm
- High relief
- Irregular outline
- Flat cleavage plates
- Step like fractures
- Arc steps

**AEOLIAN**
- Rounded and well sorted
- Mechanical upturned plates
- Low relief features
- Cracks
- Frosting

**SUBAQUEOUS**
- Rounded grains and well sorted
- May show polished surfaces
- Variable rounding of corners
- Small conchoidal fractures <1μm
- Mechanical V-shaped patterns
- Oriented etch pits
  (low energy beach)

**DIAGENESIS**
- Irregular pitted surface
- Oriented etch pits
- Frosting (aeolian burial)
- Precipitation of silica:
  - Irregular
  - Regular (crystal faces)
- Other diagenetic forms
  (e.g., crystal moulds)
4.4.2. Upturned cleavage plates

Some mechanical features are ubiquitous on quartz sand grains found in natural environments: upturned cleavage plates consist of a series of thin, parallel plates, often continuous and observed to be parallel to the cleavage direction. (Krinsley and Doornkamp, 1973). They are commonly found in small grains (200 μm in diameter or less). The plates, if unaffected by solution and precipitation, are jagged, irregular in height and generally broken on top. Upturned cleavage plates appear to be oriented along traces of cleavage planes (generally r 1011 or z 0111; Krinsley and Doornkamp, 1973) and on crystallographic faces. However, diagenetic or other chemical action usually affect the plate form, either by solution or precipitation. In acid conditions, silica precipitates on quartz to form a series of disconnected small crystal prism faces. They are often overprinted on the earlier-formed mechanical surfaces (Baker, 1976; Rai, 1987). Thus the height and thickness of cleavage faces are closely related to precipitation and solution. The size of the grain is thereby changed along with the appearance of a new grain image where there was previously only a flat surface.

Flat freshly-upturned cleavage plates are features commonly found in glacial environments and are possibly caused by glacial grinding and crushing actions. Brittle fracture of quartz can give rise to them. Chemically formed upturned-cleavage plates are common features of active diagenetic environments within which the solution and precipitation rate of silica is at least moderate.

4.4.3. V-shaped patterns

V-shaped patterns were first described by Krinsley and Takahashi (1962). They distinguished between features formed by chemical and mechanical action, and described the mechanical V-shaped patterns as the result of subaqueous impacts between grains.

Mechanical V-shaped patterns are characteristic of beach action (Krinsley et al., 1964; Krinsley and Donahue, 1968). Margolis and Kennett (1971) reported that the density of V-shaped patterns increases with distance of transport from the source. However, mechanically formed V-shaped patterns are also found on quartz sand grains from continental shelves, intertidal environments and turbidities in deep sea sediments (Baker, 1976; Hampton et al., 1978; Middleton and David, 1979; Mazzullo et al., 1982; Middleton and Anderson, 1987).

Mechanical V-shaped patterns are essentially notches cut in the tops of cleavage plates and may be produced by removal of small chips as the result of collision with other grains. If these mechanical V-shaped patterns are aligned they may form curved
grooves. Abrasion causes shearing of fragments from the grain during tangential movement, when static and dynamic forces acting on the grains in contact exceed grain strength. Early workers suggested that dense mechanical V-shaped patterns are a function of turbulent currents, this is shown in figure 4.2 (Margolis and Kennett, 1971; Krinsley and Doornkamp, 1973). However, Middleton and Anderson (1987) explained that the mechanical V-shaped patterns are related to grain size parameters. Their statistical study shows that the mechanical V-shaped pattern density is related to the degree of exposure of grains to impacting due to fluctuations of hydraulic conditions in the suspension zone. Thus, small grains which spend a long time in the suspension zone have higher mechanical V-shaped pattern density than larger grains. A high density of V-shaped patterns has been recognised (Heezen, 1963; Mazzullo et al., 1982) on sand grains from turbidites from abyssal depths. Mazzullo et al (1982) reported that sand grains from shallow water shelf turbidites also have high densities of mechanical V-shaped patterns.

Chemical V-shaped patterns are the result of silica dissolution or etching. In comparison with mechanical V-shaped patterns, most chemical V-shaped patterns have sharp corners, deep grooves, and show oriented etch pits. Furthermore, the chemical components show silica precipitation and dissolution, which are formed on upturned cleavage plates and chemical flat cleavage plates in a complex diagenetic environment.

4.5. CRITERIA FOR VARIOUS ENVIRONMENTS

Krisnley and Margolis (1968) described criteria for various environments and these were confirmed by Krinsley and Doornkamp (1973). Recognition of diagenesis and four sedimentary environments is basic to the application of the statistical method to the study of sand grain surface textures. These are source material; glacial; littoral and aeolian environments.

4.5.1. Source material

The original source of most quartz grains is granite and granite gneiss. Weathered bedrock also yields crystalline quartz. If quartz grains delivered from these rocks are unweathered after separation. The grains may have characteristics such as conchoidal breakage patterns on large grains and flat cleavage plates where the grains are small. Since all surfaces of a grain are not exposed to weathering simultaneously, in some part, a grain may have extremely flat faces unaffected by chemical action, while other parts may contain features resulting from chemical solution and precipitation (Krisnley and Doornkamp, 1973). The results, upturned cleavage plates, chemical V-shaped
Figure 4.2 Evolution of subaqueous impact features on quartz sand grain surfaces. The production of mechanical V-shaped patterns increases from river to high energy beach and turbidite environments (Margolis and Kennett, 1971; Krinsley and Doornkamp, 1973).
etching and smoothing of some surfaces, are common features occurring during weathering of the source rocks.

The environmental interpretation of quartz-sand surface textures may be complicated by factors controlling weathering; temperature, humidity, pH, Eh and mineralogical composition of the source rocks. Also, quartz sand surfaces are often interpenetrated by other minerals, leaving crystal moulds. Chemical features may appear in up to 95% of the sample with an extremely complex mixture of solution and precipitation phenomena occurring in the source material.

4.5.2. Diagenetic Features

Diagenesis results from the processes of physical and chemical change in a sediment during lithification. Morphology of grains can vary as a function of the relationship between textures created by the previous chemical or mechanical effects by the diagenetic environment in which the grains have been deposited (Krinsley and Doornkamp, 1973).

Chemical action is manifested on quartz-sand-surface textures by solution and precipitation; the results of chemical action are regularly found as triangular etching pits, irregular solution pits, deep grooves, precipitated plates, crystal overgrowths and other features of silica precipitation (Krinsley and Doornkamp, 1973).

Detrital quartz sand grains are strongly affected by chemical action, and pressure solution is most important in the movement of silica, resulting in the formation of smooth surfaces or upturned plates (Fournier, 1960; Pittman, 1972). Ancient glacial grains differ slightly from modern glacial grains by showing the effects of diagenesis: etch pits, smooth surfaces and featureless fracture planes. Chemical activity related to the presence of interstitial water reduces glacial features through time (Margolis, 1968). Pittman (1972) observed a phenomenon of abrasion solution of quartz in sandstone and suggested etching and cavities may result from pressure solution. As explained by Krinsley and Doornkamp (1973), the solution of silica is caused by insufficient silica concentration in the environment; the enrichment of silica in the environment may cause local supersaturation leading to the precipitation of silica.

Precipitation was found to be affected by the particle size and porosity of the rock. On the other hand, the silica precipitation rate controls the quartz-sand surface texture. At high rates, as in aeolian conditions, silica forms an undulating surface without any sharp projections. At moderate rates silica will form on upturned plates or may create a new set of upturned plates. At low rates, silica precipitates on the projections on the grain surfaces or forms quartz growth crystal faces where there is sufficient pore space between grains (Pittman, 1972; Krinsley and Doornkamp, 1973).
4.5.3. Glacial environment

A variety of glacial sand grain-surface textures have been studied in the literature since the work of Krinsley and Takahashi (1962). These include Krinsley and Margolis (1964); Margolis and Kennett (1971); Blackwelder and Pilkey (1972); Krinsley and Doornkamp (1973); Krinsley and Margolis (1969); Whalley and Krinsley (1974); Whalley (1978); Middleton and David (1979); Mazzullo et al. (1982, 1984); Hill and Nadeau (1984); Gomez and Small (1983), Rai (1987) and Hodel et al. (1988). The principal surface textures observed on quartz sand grains from contemporary glacial environments from part of these publications are summarized in table 3. Margolis and Kennett (1971) explained that these features are the result of glacial grinding and crushing and a combination of fatigue fractures. This was shown to result from "surface characteristic brittle fractures" by Sharp and Gomez (1986). The quartz-sand grain surface features found on recent ice-rafted grains are also consistent with features on grains of known ancient glacial environments.

Whalley and Krinsley (1974) observed quartz sand grains from subareas of the glacial environment, and demonstrated that no surface textures were observed which could characterize any particular glacial subenvironment. However the formation of general glacial features were recognized in their study.

Recent statistical studies on sand-grain surface features from glacial environments include Strass (1978), Hill and Nadeau (1984) and Mazzullo et al. (1984). The statistical studies characterize glacial sand grains by the presence of unique features and help to distinguish imprinting features from various environments. Mazzullo and Anderson (1987) showed that surface features from tillites and glacial-marine sand grains in Antarctica were different. Previous investigations give a firm foundation for further research on ancient and modern glacial features.

4.5.4. Littoral environment

Characteristic features of the littoral environment include the formation of mechanical V-shaped patterns, polished surfaces and small blocky conchoidal breakage patterns. Mechanical V-shaped patterns and polished surfaces are important criteria for littoral environments. The effect of abrasion is cumulative and the source sands often have former impact pits from older transportation cycles. The result of overprinting cannot be avoided (this will be discussed in the section on combined environments).

Grains from high-energy beach environments contain mechanical V-shaped patterns (see V-shaped patterns) and conchoidal breakage patterns. However, these
<table>
<thead>
<tr>
<th>Reference</th>
<th>Environment Sampled</th>
<th>Angular Outline</th>
<th>High Relief</th>
<th>Concoidal Breakage</th>
<th>Stepped Surfaces</th>
<th>Breakage Blocks</th>
<th>Grooves, Striae Indentation</th>
<th>Bounding + &quot;Edge Abraison&quot;</th>
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<tr>
<td>KRINSLEY &amp; SMALLEY (1972)</td>
<td>Glacial</td>
<td>X</td>
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<td>KRINSLEY &amp; DOORHKAMP (1973)</td>
<td>Glacial</td>
<td>X</td>
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<td>KRINSLEY &amp; MARGOLIS (1969)</td>
<td>Glacial</td>
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<td>WHALLEY &amp; KRINSLEY (1974)</td>
<td>Supraglacial</td>
<td>X</td>
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<td>WHALLEY &amp; KRINSLEY (1974)</td>
<td>Eglacial</td>
<td>X</td>
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<td>X</td>
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<tr>
<td>WHALLEY (1978)</td>
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<td>X</td>
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<td>X</td>
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<td>WHALLEY &amp; LANGWAY (1980)</td>
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<td>GOMEZ AND SMALL (1983)</td>
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<td>Low Level</td>
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<tr>
<td>RAI (1987)</td>
<td>Glacial</td>
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<td>X</td>
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<tr>
<td>KODEL et al. (1988)</td>
<td>Glacial</td>
<td>X</td>
<td>X</td>
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Table 3. Summary of principal surface textures observed on quartz sand grains from contemporary glacial environments and experimental glacial studies.
conchoidal breakage patterns differ from glacial counterparts, by being smaller (1 to 2 μm) and less abundant. Grains from low energy beaches contain chemical V-shaped patterns or V-etch pits which result from chemical action.

Grain size is a critical parameter which affects the formation of littoral features (Middleton and Anderson, 1987) due to the fact that small grains spend more time in suspension and hence are more pitted by abrasion than larger grains. Low relief and the degree of rounding are also considered to be common features of littoral environments.

4.5.5. Aeolian environment

Grain size parameters are also important in the formation of aeolian features. Large aeolian sand grains are strongly affected by wind abrasion. Abrasion causes impact between grains, leading to the production of mechanical upturned plates with rounded corners, edges and low relief. Conchoidal breakage patterns also appear in the aeolian features. However, Krinsley and Doornkamp (1973) explained conchoidal breakage patterns in aeolian environments as due to a single event during violent wind abrasion. These tend to be smaller in size compared with the conchoidal breakage patterns produced by glacial environments.

The porosity of aeolian deposits is larger than in any other environment, hence aeolian features are often affected by diagenetic features. Frosting is one of the most common features. Frosting can be used as diagnostic features of aeolian sand grains. It consists of raised platforms of roughly polygonal outline (Plate.3a) created chemically.

The artificial abrasion of large grains of quartz in wind tunnels produces upturned plates very similar to those formed in natural environments (Beideman, 1960). The height and spacing of the plates are related to energy input and velocity of the environment.

4.5.6. Combined environments

4.5.6.1. GLACIAL AND SUBAQUEOUS ENVIRONMENTS COMBINED

Sand often is transported into the marine environment by subaqueous turbulent flows or by ice-rafting. Sand grains from glacial environments are reworked by aqueous mechanisms, leading to the superimposition of aqueous abrasion on features from the former glacial episode. If grains have been transported for a long time under the high flow regime, it may not be possible to recognize a former glacial episode in surface textures. However, in short term transport by water currents, some grains in the sample will show both episodes in the complex history.
Blackwelder and Pilkey in 1972 showed that grains with glacial features are present in many samples from the continental shelf north of Delaware, eastern USA. The shelf shows a combination of two material sources. The northern source is characterized by glacial features and the southern one by non-glacial features (beaches, river and some aeolian). The combined features are interpreted by Blackwelder and Pilkey as due to multiple cycles of abrasion. Recent research shows that quartz sand grains often display a combination of several sources (Strass, 1978; Mazzullo et al., 1982 & 1987; Hill and Nadeau, 1984; Rai, 1987).

4.5.6.2. SUBAQUEOUS AND AEOLIAN COMBINED

Coastal dune sand is the most typical environment for the study of sand-grain surface textures of combined subaqueous and aeolian environments. Most grains have not travelled far from the beach, hence the abundance of subaqueous features, such as mechanical V-shaped patterns. Baker (1976) studied coastal dunes and confirmed this result. Baker also showed a decrease of subaqueous environmental features inland, while those thought to be characteristic of the aeolian processes increased. However, simple statistics prove that it is not possible to distinguish beach from coastal dune sand by the use of diagnostic features of quartz sand grain-surface textures alone (Baker, 1976).

River and aeolian environments commonly occur together in seasonal climates where wet and dry seasons are significantly different during the year. Sand grains from the Colorado River contain some rounded grains with frosted surface textures, indicating that these grains were of aeolian origin. Diagenetic features are predominant in this environment, resulting from a high silica concentration in the water that contacts the quartz. However, reworking of sand grains from river beds in the wet season causes subaqueous mechanical features. If quartz grains have been moved by water far from the source rocks, mechanical V-shaped patterns of subaqueous features will be found overprinting aeolian features.

4.6. RESULTS OF QUARTZ SAND SURFACE TEXTURES IN THIS STUDY

SEM examination of sand-grain surface textures of Cretaceous sediments from the southern Eromanga Basin, central-north of South Australia, led the author to recognize five sediment types. The environmental interpretation of these sediment types is briefly discussed below.
Type I. The surface marking on quartz sand grains of type one represents an aeolian environment. Aeolian features are predominant in this grain type and include rounded grains, particle flatness, large surface depressions and mechanical upturned cleavage plates. These features are the result of wind abrasion. Chemical features include frosting and adhering-clay particles on grain surfaces which are also diagnostic features indicating aeolian environment (Plate 3b, 3c).

Type II. Sand grains from this environment have a distinct distribution of surface features that are caused by the mechanical abrasion on the beach and surfzone. These sand grains contain well-rounded, V-shaped patterns, curved grooves and polished surfaces typical of littoral environments. Chemical action has slightly affected the precipitation of silica. But the mechanical features of subaqueous environments can be recognized on the grain surface (Krinsley and Doornkamp, 1973). Type II represents littoral environment (Plate 3d, 3e, 3f).

Type III. Surface textures of type III consist of glacial grains which have features such as highly angular outline, conchoidal fractures, parallel stepped surfaces, high relief, well developed elongate prismatic grains, etc. due to glacial grinding and crushing. The glacial features are often partly affected by diagenetic processes, but overall the grains still exhibit the major features of a former glacial episode (Plate 3g, 3h, 4a, 4b, 4c, 4d, 4e, 4f).

Type IV. This type represents diagenetic features, which include grains which have been completely overgrown by silica or other authigenic minerals, and show strong dissolution of original grain surface features as grains have been subjected to repeated, prolonged periods of chemical weathering. Type IV also represents some problematical grains (Plate 4g, 4h, 5a, 5b, 5c, 5d, 5f).

Type V. Type V has the combined features of glacial and subaqueous grains. Glacial features are overprinted by latter subaqueous features (Plate 5e, 5f, 5g, 5h)

4.7. STATISTICAL ANALYSIS

Semiquantitative analysis of the quartz-sand surface textures was described by Mazzullo et al. (1984) and Hodel et al. (1988). The results of studies of quartz-sand surface textures of the Jurassic-Cretaceous sediments are presented in table 4 and the representative samples are shown in histograms of frequency versus grain types (Fig. 4.3). These histograms show for each sample, the percentage of grains exhibiting the characteristic features of various environments. Certain differences from environment columns are readily apparent. The abundance of grain types is identified directly on the frequency axis on the left hand side and the proportion of the grain types can be read on the relative frequency on the right hand side of the histogram. The
Plate 3a.
Frosting features of an aeolian sand grain (sample 889.85.08).

Plate 3b.
Extremely rounded grain shows aeolian features of a typical type I grain. (sample 889.86.5).

Plate 3c.
Closer look at the surface of an aeolian grain. Note adhering-clay particles and mechanical upturned cleavage plates, on the surface of the grain indicating a typical type I grain. (sample 889.86.5).

Plate 3d.
Rounded grain and polished surface with a number of mechanical V-shaped patterns. Indicates subaqueous action, and is typical of type II grain (sample 889.86.14).

Plate 3e.
Rounded, polished edges of grain have been subjected to both mechanical and chemical action, probably the result of low energy beach action (sample 889.86.10).

Plate 3f.
Closer look at rounded surface of a type II grain. There are numbers of mechanical V-shaped patterns, which are arranged in irregular patterns (sample 889.86.8).

Plate 3g.
Extremely angular grain with conchoidal breakage patterns. The grain shows a series of parallel striaete lines indicating a glacial source, thus typical of type III grain (sample 889.86.25).

Plate 3h.
General view of a glacial grain shows extremely angular, irregular surface, conchoidal breakage patterns, high relief, indicating typical type III grain (sample 889.86.25).
Plate 4a.
Extremely angular small chip of quartz shows flat cleavage plate, conchoidal breakage patterns are steps and irregular outline. These characteristics indicate a glacial origin, type III (sample 889.195.8).

Plate 4b.
Large grain of quartz shows extremely angular outline, conchoidal breakage patterns and parallel arc steps, possibly the result of mechanical grinding and crushing, type III grain (sample 889.86.30).

Plate 4c.
Irregular outline and conchoidal breakage patterns on large grain. The top surface of the grain shows a cleavage plane. The front surface displays a series of irregular conchoidal breakage patterns. Suggest glacial origin, type III grain (sample 889.86.30).

Plate 4d.
Broken cleavage surfaces and conchoidal fractures patterns, type III grain (sample 889.86.25).

Plate 4e.
Flat cleavage plate with irregular outline and conchoidal breakage patterns. The top surface shows glacial features with slight silica overgrowth, type III grain (sample 889.86.25).

Plate 4f.
Extremely angular outline, conchoidal breakage pattern, adhering particles and flat cleavage plate. The grain shows a fresh surface, probably caused by glacial grinding or crushing. These are good criteria for glacial determination (sample 889.86.30).

Plate 4g.
Precipitation of silica, flat cleavage plates and conchoidal breakage patterns. The upper surface shows flat cleavage plate. Below, on the left and right surfaces appear precipitates of silica, but with the extremely angular, high, flat cleavage plates which suggest glacial origin (sample 889.86.27).

Plate 4h.
Etching of subaqueous grain. The grain shows complex features such as irregularity in one size and semi-circularity in the other size. This suggests, the grain originally came from a beach environment, and was broken mechanically, possibly by glacial grinding and crushing. In the latter phase, the grain was extremely etched, possibly by calcite or authigenic feldspar, type IV (sample 889.85.04)
Plate 5a.
Split image of an early-authigenic feldspar (F). The authigenic component is overgrown on the surface of an aeolian grain. In addition, precipitation of silica also can be seen on the enlarged photo, type IV grain (sample 889.86.29)

Plate 5b.
An early stage of silica diagenesis. The grain shows silica overgrowth on the surface. But the angular features and conchoidal breakage patterns can be recognized on the left of the photo, type IV grain (sample 889.86.25)

Plate 5c.
Precipitation on quartz-crystal prism faces, individual crystal of quartz formed crystal prism face on a flat surface, type IV grain, Trinity Well sandstone, Pelican Well.

Plate 5d.
Quartz growth crystal prism faces. The overgrowth has fully covered the detrital component, type IV grain (889 85.08).

Plate 5e.
A general view of the Trinity Well Sandstone; Pelican Well (sand fractions). Several grain types are recognized in the sample: angular grain with conchoidal breakage patterns; rounded grains showing glacial and subaqueous combined; diagenetic grains (sample 889.86.10).

Plate 5f.
An elongated-grain shows that conchoidal breakage patterns have been rounded by subaqueous action. Possibly from shore line environment, type IV grain (sample 889.86.25).

Plate 5g.
A rounded grain of subaqueous environment, on the higher portions showing mechanical V-shaped patterns and polished edges and corners. On the lower portions, dissolution and conchoidal breakage patterns occur. Whereas, the lower portions have been protected by higher portions, type IV grain (889.86.27)

Plate 5h.
Split image of a rounded grain appears with a polished surface and latter precipitation of silica. This grain has undergone a long period of subaqueous action and precipitation of silica, type IV grain (889.86.11).
Figure 4.3. Histograms of frequency versus grain types of the clasts-bearing sediments of the Cretaceous Southern Eromanga Basin

Type 1 = aeolian  Type 2 = subaqueous  Type 3 = glacial
Type 4 = diagenesis  type 5 = glacial & subaqueous combined
Table 4. Comparison of probability of glacial population of the Cretaceous sediment to glacial occurrence world wide

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<tr>
<td>(P3 ranges from) 0.29 to 0.60</td>
<td>0.30 to 0.61</td>
<td>0.39 to 0.63</td>
<td>0.10 to 0.40</td>
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* P3 = Probability of glacial grains in the sample
relative frequency axis also shows the distribution of the environmental population. For example, sample 889.86.14 shows that the frequency of glacial grains is 14, and the relative frequency (R(f)) of glacial grains is equal to 0.40. In another words, the probability is 0.40 that a quartz sand grain choosen randomly will give glacial features. But, if the probability of glacial and combined glacial and subaqueous are added the total probability of glacial grain will be 0.6 to 0.8 (1.0 is certain).

Glacial, subaqueous and diagenetic features were the predominant features observed in these samples. These were mainly affected by dissolution and precipitation of silica on the primary surface textures. Hence, the mobilization of silica has reduced the number of glacial and subaqueous features in the sediments.

Most of the samples showed diagenetic features but some representative samples (with less diagenesis) taken from beneath the lonestones and boulder beds showed the probability of glacial features up to 0.40.

A comparison with various histograms from the literature reveals similarities to glacial environments elsewhere, as shown in Table 4. From this table we can see that the probability of glacial environments for the Cadna-owie Formation and the Bulldog Shale (0.10-0.40) is similar to those for modern sediments of the Canadian Beaufort Shelf (Hill and Nadeau, 1984), where R(f) ranges from 0.30 to 0.61, and for modern coastal glacials, which ranges from 0.39 to 0.63 (Strass, 1978).

Furthermore, quartz-sand surface textures of the Bulldog Shale, Davenport Spring area studied by Fennessy (1984) provided a table of grain morphologies. From this table the probability of glacial environment ranges from 0.28 to 0.60.

In natural environments, most samples contain two or more grain characteristics (Blackwelder and Pilkey, 1972; Krinsley and Doornkamp, 1973; Hill and Nadeau; 1984; Mazzullo and Anderson, 1987). The presence of aeolian grains in a number of samples indicates subsequent erosion and deposition in a marine environment of grains subjected to wind abrasion processes and reworking in dune beach environment (Mazzullo and Anderson, 1987). A dry climate may have allowed the development of local dune fields and generation of the aeolian features. These aeolian grains thus are consistent with paleoclimatic evidence which suggest a cold and dry environment like the north west of Canada and Northern slope of Alaska.

Sample 889.85.8 contains subrounded grains showing subaqueous and glacial features (type V = 28%, II = 33% and III =22% and type IV=16%). This sample has undergone a substantial amount of high energy subaqueous transport in the littoral environment, which has overprinted abundant glacial features. However, sample 889.86.8 (type V = 27%, II =20% and III =20% and type IV=20%) contains a number of fresh glacial grains and also subaqueous grains and imprinting features; this sample has undergone less reworking and may have been deposited within the fluvially
influenced part of the Trinity Sandstone. The relative proportions of glacial and reworked grains indicate the relative influence of littoral environment (Hill and Nadeau, 1984). Thus a greater influence of reworking near shore must be, as expected, in the form of polished grains mixed with angular grains.

4.8. SUMMARY AND DISCUSSION.

Scanning electron microscopy was effectively used to study quartz-sand surface textures of the Jurassic-Cretaceous sediments in the southern Eromanga Basin. The method provided valuable information concerning the provenance history of clastic sediments. Semiquantitative methods can be successfully used in the study of quartz-sand surface textures. Grain type classification is an important tool and leads to the determination of regional sources of the clastic sediments and environmental interpretation.

About 93% of quartz grains exhibit composite features, as they contain surfaces attributed to glacial, subaqueous, aeolian and diagenetic activities. Among primary glacial features as mentioned earlier, conchoidal fracture patterns, angular outlines, high relief and arc steps are the prominent features in the sediment. Besides these features, it is interesting to note that many fractured and unfractured grains show small adhering particles which appear similar to precipitation platelets characteristic of glacial environments. A considerable number of samples exhibit both diagenetic and glacial features. In addition, some samples show that glacial features tend to have angular shape (elongate, triangular, quadrahedral) and fractures indicating that the grains were deposited while being broken mechanically and had undergone little reworking.

Quartz grains from the Cadna-owie Formation and the Bulldog Shale illustrated both glacial and subaqueous features (Pelican Well and Trinity Well Sections). The glacial features were attributed either to crushing and grinding of the basement rocks or reworking of nearby ancient glacials (Precambrian and Permian tillites). However, it is difficult to see how some grains with glacial features could have escaped the continuous buffeting which occurs in nearshore marine environments. Although the sediments from group V shows subaqueous features associated with glacial features (the suites of textures superimpose of II and III), therefore the environments of modification is probably the combination of the two. It is unlikely that group III represents reworked or slumped slope deposits of Group II sediments.
The slight rounding of many quartz grains, the degree of imprinting of aqueous impact pits and diagenetic features on glacial grains, suggest very little reworking following their formation during a glacial episode. The presence of several grain types in most of the samples, including glacial, subaqueous and aeolian grains, is evidence that sediments have been subjected to varied environments in the southern Eromanga Basin during the Cretaceous.

In most of the samples studied, the effects of silica solution increase with decreasing grain size, and are apparently most prevalent in the Bulldog Shale samples. Stuart Creek Opal Field quartz shows extreme dissolution. The study of quartz-sand surface textures in carbonate-cemented sandstones, led to the conclusion that secondary quartz and authigenic K-feldspar are the major factors which destroy quartz-sand surface textures.

The sediments were brought into the area by subaqueous transport mechanisms and by rafting. The superimposition of subaqueous actions and chemical features on glacial grains is the result of subsequent transport mechanisms and diagenetic processes in the study area.

If some of the sand from the Bulldog Shale was transported into this part of the basin by ice rafting it seems likely that glaciation occurred regionally in the Cretaceous. Some early Neocomian to Aptian sand grains were eroded by subaqueous or glacial erosion, underwent fluvial and nearshore marine cycles of aggradation and eventually were deposited in the area studied.
CHAPTER 5 DIAGENESIS

5.1. INTRODUCTION

Diagenesis has been defined by many workers in different fields of study, such as Burley et al. (1987), who defined diagenesis for clastic diagenetic environments as "the sum of those processes by which sedimentary clastic assemblages attempt to reach equilibrium with their environment" (pp.189). They also distinguished three types of diagenetic environment: depositional environment or surface chemistry or eogenesis; burial or subsurface conditions or mesogenesis; and weathering of surface conditions or telogenesis. The diagenetic study and environmental reconstruction must therefore be based on the mineralogical composition of clastic sediments. In the present work, authigenic minerals were studied to provide important information on palaeo-weathering or diagenetic changes within the sediments. Diagenesis revealed in most sediments results from the dissolution of unstable minerals, and controls the formation of authigenic minerals.

Depositional environmental reconstruction is related to the eogenetic mineral assemblages, these minerals equilibrating between sedimentary geochemical conditions and interstitial pore fluids. For example, in the fluvial lacustrine environments of the Late Jurassic (Algebuckina Sandstone), the mineral assemblage reflects a low pH and oxidizing conditions. However, in the transgressive marine environments of the Cretaceous (Cadna-owie Formation and Bulldog Shale), the authigenic mineral assemblage reflects a high pH and reducing conditions.

Mesogenesis is a more complex process involving pressure and temperature, leading to an increase of reaction rates. The processes cause losses of interstitial pore fluid and large amounts of fluid are released into the environment. The released fluid may serve to transport a large amount of material (i.e., numbers of ions, gas, oil, etc.). The degree of mesogenesis can be predicted as a function of depth. In addition there are other factors involved in the formation of mesogenetic mineral assemblages.

Sediment exposed under weathering conditions by uplift or desiccation leads to alteration in the telogenetic realm. These processes are influenced by meteoric water and climatic conditions.

Scanning electron microscopy is very important in paleoenvironmental analysis. The value of textural properties and mineralogical composition of the rock as indicators of palaeoenvironment is well known. Hence, the effects of diagenesis on textural properties of rocks must be well understood.
5.2. AIMS OF DIAGENETIC STUDY

The aims of this part of the study are
-to determine the mineralogical composition of the Mesozoic sediments of outcrop and Toodla #1 core, and possibly to determine the source region.
-to determine differences between the diagenetic regimes of burial and surface conditions.

The study involved logging of the Toodla #1 core, study of hand specimens and SEM descriptions of both outcrop and core samples. The samples were collected for quartz-sand surface texture examination and mineralogical studies, and taken at random depths (Toodla #1) throughout the Mesozoic sequence. Additional mineralogical and quartz sand surface texture studies were made on various samples from sites throughout the southern margin of the Eromanga Basin. These include Davenport Springs, the Peake and Denision Ranges and Moolawatana (see sample location map figure 3.1). The core log is shown in figure 5.1. The studies also included detailed sample descriptions under the SEM. The results of the described representative samples are shown in appendix 1.

The mineralogical study of core samples places particular emphasis on clay mineralogy. The silty shale facies sediment is a very friable material and tends to disintegrate in water during laboratory processes. Hence, thin sections could not be made. SEM and X-ray diffraction were the only useful methods to examine the mineralogical composition. Residue sand fractions were prepared by polished block methods for chemical analysis under the electron microprobe.

5.3. MINERALOGICAL STUDIES

5.3.1. Quartz

Quartz was identified by X-ray diffraction of the bulk rock. Authigenic quartz overgrowths are the major siliceous diagenetic feature, which is widespread throughout the sequence. It occurs mainly as partial or complete overgrowths of detrital grains. The authigenic quartz has euhedral crystal prism faces and aggregates with authigenic kaolinite. Details of silica overgrowth are discussed in chapter 4.

Quartz overgrowths are a major porosity-reducer in all samples from both outcrops and core samples, and unlike the authigenic feldspar, siderite and dolomite, which may often be dissolved away to form secondary porosity, silica overgrowths reduce porosity permanently.
Figure 5.1. Core log of Toodla #1 Oodnadatta region

LEGEND:
- g: Glauconite
- Bivalve shells
- B: Bivalve shells
- P: Pyrite
- L: Laminated
- C: Crossbedded
- A: Ripple Cross lamination
- Z: Zoophycus
- C: Carbonate concentration
- S: Sand filled horizontal burrows
- O: Core loss
- L: Slump
- L: Lignite
- f: Pebble
- L: Intraclast
- Ca: Ca CO₃
- B: Bioturbation
- R: Reddening surface
- C: Channel
- X: Cone in cone structure
- D: Scour base
- *: Sampled
Quartz is the most common mineral in all samples from various outcrops and different stratigraphic levels. Quartz was identified from X-ray diffractograms showing sharp symmetrical peaks at about 4.27Å, 3.34Å, 2.46Å, 2.28Å, 2.22Å, 2.24Å, 2.13Å, and 1.81Å (according to JCPDS). Under the SEM, quartz grains show a strong degree of diagenetic alteration. (see chapter 4)

5.3.2. Authigenic feldspar

5.3.2.1. INTRODUCTION

Authigenic feldspar is a ubiquitous feature of the Algebuckina Sandstone and Trinity Well Sandstone. It overgrows detrital feldspar and primary quartz. The authigenic components have the euhedral crystal faces of monoclinic potassium feldspar (andularia habit) (Waugh, 1978), and are pure end members of potassium feldspar.

Precipitation of the authigenic potassium feldspar takes place together with formation of other authigenic minerals (illite/smectite, quartz, zeolite and kaolinite). The optical discontinuity of overgrowth and host rock is significant in distinguishing between detrital and authigenic components.

The presence of authigenic feldspar indicates intrastratal dissolution of detrital grains from which a supply of potassium, aluminium and silica is derived. Pore-water chemistry, hydrolysis, local weathering and groundwater conditions are the dominant factors involved in the formation of authigenic feldspars.

5.3.2.2. OCCURRENCE OF AUTHIGENIC FELDSPAR.

Authigenic potassium feldspar (microcline) is a predominant feldspar of the Jurassic-Cretaceous sediments in the southern Eromanga Basin. It occurs mainly as partial overgrowths covering detrital grains, or is developed among cemented clays, in infilled pore spaces and as completely developed euhedral grains. The most common cases of overgrowth are those on detrital grain surfaces.

The large authigenic microcline in the Algebuckina Sandstone and the lower part of the Cadna-owie Formation is associated with kaolinite, while the authigenic microcline in the upper part of the Cadna-owie Formation is overgrown with illite/smectite or associated with zeolite. Outcrop samples from the Cadna-owie Formation in the Peake and Denison Ranges contain authigenic microclines associated with zeolitic-glaucnitic-quartzose silty sandstone. Authigenic microclines in the Bulldog Shale in
Toodla #1 are small and associated with zeolitic-pyritic-glauconic-quartzose silty shale.

5.3.2.3. DEVELOPMENT OF AUTHIGENIC FELDSPAR.

The typical morphology of authigenic K-feldspar is seen as euhedral crystal faces usually developed along the axes of grains and aggregated as small oriented crystals on detrital grain surfaces. If the overgrowths on a grain surface are advanced, this will be indicated by a high degree of crystallization and the overall shape will be that of a well-formed authigenic microcline with euhedral crystal faces (Plate 6a).

Early stages of authigenic feldspars are recognizable as small-flattened rhombohedral crystals. These are combinations of the basal pinacoid and crystal prism faces, which form individual crystals with hairing as the preferred orientation (Plate 6b). This early overgrowth was regarded by Baskin (1956) and Deer et al. (1966) as an andularia habit in morphological terms (Waugh, 1978; Ali and Turner, 1982).

The next stage in the development of authigenic feldspar is an increase in numerical density of the individual crystals, leading to the combination of groups of small flattened rhombohedral crystals which form larger crystal faces (Plate 6c). At a further stage of development, growth occurs over large areas until the entire grain surface is covered by well-crystallised faces of authigenic microcline (Plate 6d). The overgrowth crystals have 2V angles ranging from 0 to 68°. This is indicative of sanidine, microcline and orthoclase habits (Kastner, 1971).

5.3.2.4. CHEMISTRY

Kastner (1971) determined the chemistry of authigenic feldspars by using the electron microprobe. This method was confirmed by Waugh (1978), Ali and Turner (1982) and Burley et al. (1987). Electron microprobe analyses provide the only way of characterizing many phases of the individual grains. In addition, precise identification is supported by X-ray diffraction and the EDS semiquantitative analysis.

Electron microprobe results of both detrital and authigenic feldspars are shown in appendix 2 and table 5. From the results, we can see that authigenic overgrowths occur under similar conditions to authigenic K-feldspar from different ages and places around the world (e.g., The Bromsgrove Sandstone, Triassic, Central England; British Permo-Triassic Sandstones; Waterton formation, Late Precambrian, Alberta, Canada; Trenton Formation, Ordovician, New York, U.S.A.). The authigenic microcline is virtually an end member of potassium feldspar. The authigenic microcline has less than 1 mole percent of NaAlSi$_3$O$_8$ ranging from 0 to 0.6 %, and contains from 100% to
Plate 6a.
Authigenic K-feldspar (F) overgrowths (microcline) on a detrital grain which is surrounded by silica and kaolinite (K). Identification is based on the EDS analysis giving the major elements Si, Al and K corresponding to the K-feldspar chemical composition. The overgrowth almost completely covers the detrital grain surfaces (889.86.05).

Plate 6b.
Split image of early stage of authigenic K-feldspar overgrowing detrital feldspar. The individual overgrowths have euhedral crystal faces and prefered orientations (889.86.12).

Plate 6c.
Authigenic K-feldspar (F) developed among detrital quartz (889.86.05).

Plate 6d.
A fully developed authigenic K-feldspar (889.86.05).

Plate 6e.
Detrital grain of kaolinite. The grain shows face to face book stacks, rounded edges and corners. Identification is based on the EDS analysis yielding the major elements Si and Al at nearly equal peak heights (889.85.07).

Plate 6f.
Authigenic kaolinite partly filling pores. Kaolinite developed among illite. Identification is based on EDS analysis and a series of face to face stacks of pseudohexagonal plates (889, 258.5).
99.4% mole of KAlSi$_3$O$_8$. This clearly shows a similarity to the theoretical composition of the end member of potassium feldspar (Kastner, 1971; Waugh, 1978).

5.3.2.5. FORMATION OF AUTHIGENIC FELDSPAR

Authigenic feldspar is related to the hydrolysis of unstable detrital grains within the stability series of Goldich (1938), in which the solution of ferromagnesian minerals, plagioclase and volcanic rock fragments resulting from non-equilibrium with interstitial water, provides sufficient silica, aluminium and potassium, and can lead to the formation of authigenic feldspar. This type of occurrence was attributed by Kastner and Siever (1979) to the influence of volcaniclastic materials in the sediment. Scanning electron microscopy shows that hydrolysis has also affected minerals higher in the stability series. Dissolution of K-feldspar and the precipitation of pore-filling minerals (kaolinite, illite/smectite, zeolite) indicate that interstitial pore water has played an important role in diagenesis.

Authigenic potassium feldspar from this sedimentary sequence of the basin has not previously been described in the area studied. The formation of authigenic feldspar in the Jurassic-Cretaceous sediment conforms to the kinetic model of chemically interrelated minerals in sedimentary environments proposed by Helgeson et al. (1969) (Fig. 5.2).

**Fig. 5.2** Activity diagram for the system K$_2$O–Al$_2$O$_3$–SiO$_2$–H$_2$O at 0 °C, unit activity of water, and one atmosphere. The shaded areas labeled M and A represent the compositional range of surface and deep ocean waters, respectively, and those annotated N and B designate, respectively, the average composition of world streams and the compositional range of interstitial waters in deep sea and shelf sediments—see text. In addition to the data cited in the caption of Fig. 2, heat capacities taken from Chass and Cobble (1964), Kelley (1960, 1962), and Pankratz (1961) were used to calculate the positions of the stability field boundaries. Estimated heat capacities of the ions and H$_2$SiO$_4$ at 0°C were used in the calculations. The methods and uncertainties involved in calculating equilibrium constants for this system (and those represented in Figs. 10 and 11) at temperatures other than 25°C have been discussed elsewhere (Garrels and Christ, 1965; Helgeson, 1967b, 1969).
### TABLE 5 Comparision authigenic K-feldspar from the Eromanga Basin with World wide authigenic K-feldspar

<table>
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<tr>
<td>SiO2</td>
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<td>64.41</td>
<td>64.90</td>
<td>66.14</td>
<td>67.57</td>
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<td>Al2O3</td>
<td>18.32</td>
<td>18.54</td>
<td>18.45</td>
<td>18.46</td>
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<tr>
<td>Na2O</td>
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<tr>
<td>K2O</td>
<td>16.93</td>
<td>16.94</td>
<td>16.90</td>
<td>14.95</td>
<td>14.29</td>
</tr>
<tr>
<td>Total:</td>
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<td>99.89</td>
<td>100.35</td>
<td>99.55</td>
<td>100.14</td>
</tr>
</tbody>
</table>

2. Waugh (1978)
3. The Trenton Formation, Ordovician, Glenfall, NewYork, U.S.A. (Kastner, 1971)
4. Cadna-owie Formation, Toodla #1, Oodnadatta Formation
5. Trinity Well Sandstone, Trinity Well, central north of South Australia
This model is confirmed in the diagenetic literature and particularly as regards the formation of authigenic feldspar. From this model we can see that formation of authigenic feldspar results from the relation between the \( \text{K}^+ \) and \( \text{H}_2\text{SiO}_4 \) ratio. The higher the \( \text{H}_2\text{SiO}_4 \) concentration, the greater the rate of precipitation of K-feldspar.

The purity and well-developed crystal faces of authigenic feldspar lead to the suggestion that precipitation rates were slow, but with a consistent supply of element sources.

Kastner and Siever (1979) proposed that authigenic feldspar forms in the early stage of diagenesis. However, the occurrence of authigenic feldspar in the Bulldog Shale clearly indicates that the minerals were precipitated during and following the cementation of the rock (authigenic feldspar contains calcite inclusions and calcite cement has also been overlapped by authigenic microcline) and at the same time or after the authigenic clays had developed in clastic sediments (authigenic feldspar displays overgrowths on illite/smectite). Authigenic feldspar filling secondary pore spaces indicates formation during a late phase of diagenesis.

### 5.3.3. Kaolinite

#### 5.3.3.1. INTRODUCTION

Kaolinite in the Jurassic-Cretaceous sediments is a characteristic product of diagenetic processes. It is an ubiquitous feature in the Algebuckina Sandstone and at the base of the Cadna-owie Formation (Wopfner et al., 1970, Exon and Senior, 1976). Most of authigenic kaolinite in the Algebuckina Sandstone sequences shows a high degree of crystallization. Kaolinite often fills primary pore spaces, or secondary pore spaces created as a result of hydrolysis and dissolution of the unstable minerals. The presence of authigenic kaolinite indicates acidic and oxidizing conditions in the late Jurassic sequence and the early Neocomian surface weathering of telogenetic regime.

#### 5.3.3.2. OCCURRENCE OF KAOLINITE

Kaolinite consists of detrital and authigenic components. At low magnification detrital kaolinite often has rounded edges and corners. However, at high magnification detrital kaolinite shows the face-to-face book stack of pseudohexagonal crystals (Plate 6e). Detrital components are rarely found and occur only in residual non-carbonate outcrop samples of the Cadna-owie Formation. The detrital kaolinite grains are silt-sized and display rounded edges and corners.
The authigenic component shows infilled pore spaces and secondary pore spaces and commonly replaces other minerals, such as illite/smectite and feldspar. The authigenic component consists of a series of face-to-face stacks of pseudohexagonal plates (Plate 6f). This also yields characteristics of the kaolinite spectrum under the energy dispersive system (EDS) which have equal peak heights of silica and aluminium corresponding to the chemical composition of kaolinite \(\text{Al}_4[\text{Si}_4\text{O}_{10}](\text{OH})_2\). They also have similar chemical compositions to Stuart Creek Opal Field kaolinites.

However, kaolinite from the Algebuckina Sandstone Toola #1 and Trinity Well Sandstone reflects a higher degree of crystallization than weathered kaolinite of the Bulldog Shale in the Stuart Creek Opal Field. The morphology and EDS spectrum of authigenic components are also characteristic of a typical kaolinite mineral (Plate 7a, 7b) (Wilson and Pittman, 1977).

5.3.3.3. CHEMISTRY

Chemical analyses of kaolinite were performed on the electron microprobe. The electron microprobe analyses give detailed indications of kaolinite chemical composition, and are shown in Appendix 2. From these results (table 6.) we can see that the chemistry of kaolinite from the Algebuckina Sandstone in Toola #1 and the Trinity Well Sandstone are similar to kaolinite from worldwide occurrences described in the literature. The electron microprobe analyses also yield a low elemental total (about 81%). This can be explained by the presence of \(2\text{H}_2\text{O} ,(\text{OH})\) ions in the structural formula of kaolinite or perhaps by an interlayer of water.

The chemical composition of kaolinite in the Algebuckina Sandstone (appendix 2) shows very little variation. Analyses show very few substitutions in the chemical composition. Deer et al., (1966) suggested that the formula of kaolinite can be written in terms of oxides such as \(\text{Al}_2\text{O}_3.2\text{SiO}_2.2\text{H}_2\text{O}\). This composition is also similar to halloysite (\(\text{Al}_2\text{O}_3.2\text{SiO}_2.4\text{H}_2\text{O}\)) but, X-ray diffraction studies confirmed the crystal structure of kaolinite. On the X-ray diffractogram of the bulk samples, kaolinite (3T) was identified by two major peaks at 7.19Å and 3.58Å (according to JCPDS). These peaks collapse and become non-crystalline when heated to 550-600°C for about one hour. However, X-ray diffraction of the 2 μm fraction shows a stronger intensity of kaolinite peaks (a higher degree of crystallization). Thus, there is higher proportion of kaolinite (60%) in the clay fractions (Wilson and Pittman, 1977). Kaolinite in the Algebuckina Sandstone and the lower part of the Cadna-owie Formation from Toola #1 is dominated by kaolinite type 3T.
Table 6. Chemical Comparison of Kaolinite from the Eromanga Basin with occurrence of kaolinite world wide.

<table>
<thead>
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<tbody>
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<td>SiO2</td>
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<td>49.01</td>
<td>46.5</td>
</tr>
<tr>
<td>Al2O3</td>
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<td>23.50</td>
<td>30.91</td>
<td>39.5</td>
</tr>
<tr>
<td>TiO2</td>
<td>-</td>
<td>0.63</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>FeO</td>
<td>0.18</td>
<td>1.71</td>
<td>0.50</td>
<td>-</td>
</tr>
<tr>
<td>Fe2O3</td>
<td>0.57</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>MgO</td>
<td>0.14</td>
<td>0.86</td>
<td>0.29</td>
<td>-</td>
</tr>
<tr>
<td>MnO</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>CaO</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Na2O</td>
<td>-</td>
<td>0.15</td>
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<td>-</td>
</tr>
<tr>
<td>K2O</td>
<td>0.03</td>
<td>1.05</td>
<td>0.12</td>
<td>-</td>
</tr>
<tr>
<td>H2O</td>
<td>14.09</td>
<td>13.98</td>
<td>81.11</td>
<td>14.0</td>
</tr>
<tr>
<td>Total</td>
<td>100.77</td>
<td>101.43</td>
<td>81.11</td>
<td>100.00</td>
</tr>
</tbody>
</table>

1. Hydrothermal kaolinite from Niigata, Japan (Nagasawa, 1953. Quoted from Deer et al. (1966)).
2. Kaolinite from early stage of hydrothermally-altered granite (Mewnier and Velde, 1982).
3. 889.86.12, Trinity Well Sandstone, Trinity Well, centre north of South Australia.
Plate 7a.
Close look at the morphology of authigenic kaolinite. Individual kaolinite crystals form a series of pseudohexagonal crystal faces and book stacks. EDS analysis shows the major elements of Si and Al at nearly equal peak heights (sample 889.258.5).

Plate 7b.
Authigenic kaolinite developed among illitic layers. Morphology and EDS spectrum are diagnostic of kaolinite (sample 889.258.5).

Plate 7c.
Split image of authigenic kaolinite infilling secondary porespaces. Dissolution of detrital grains yields secondary porosity (dissolution porosity), infilled by authigenic kaolinite (sample 889.258.5).

Plate 7d.
Part dissolution of detrital feldspar (F) and development of authigenic kaolinite (K) (sample 889.258.5)

Plate 7e.
Split image of a detrital grain of illite (I), EDS analysis shows major elements Si, Al, K, Ti and Fe similar to composition of glauconite (G) (sample 889.195.8)

Plate 7f.
Closer look at the morphology of detrital illite (I). Individual crystals appear as flakes, and aggregate among pyrite crystals. This morphology and EDS analysis are the common characteristics of detrital components (sample 889.195.8).
Authigenic kaolinite is the principal clay in the Trinity Well Sandstone and was possibly formed during diagenesis. The kaolinite was identified by powder X-ray diffraction and has similar characteristics to the Stuart Creek kaolinite. However, under the SEM, kaolinite appeared as euhedral crystals of an authigenic origin often filling secondary pore spaces (dissolution porosity), as the result of hydrolysis and dissolution of feldspar. (Plate.7c, 7d).

5.3.3.4. FORMATION OF KAOLINITE

Kaolinite forms under low temperature; low pH; high Eh and low alkali conditions (Marshall, 1977). The degree of crystallization of kaolinite is increased by the presence of iron in the system, hence the crystallization is related to the ferric minerals (Hurt and Kunkle, 1985). The formation of kaolinite in the rock leads to the suggestion of a fluvial type of environment within an oxygenated environment (according to Hurt and Kunkle, 1985).

Formation of authigenic kaolinite follows the kinetic model of authigenic minerals in sedimentary environments (Helgeson et al., 1969). In burial conditions, the water pressure and concentration of $H_4SiO_4$ requisite for kaolinite are maintained, leading to the precipitation of kaolinite. On the other hand, kaolinite tends to be transformed into illite/smectite and other minerals under high pH and alkaline conditions.

The presence of kaolinite reflects the early Neocomian fluvial environment (Exon and Senior, 1976; Griffith, 1979) and the weathering conditions of the Tertiary arid-semi-arid climate of the central north of South Australia (Wopfner, 1974). The kaolinite consists of both detrital and authigenic components.

In surface weathering, dissolution of unstable minerals also contributes to the formation of kaolinite. Barker (1983) showed that the formation of kaolinite in the Andamooka and Coober Pedy Opal Field results from leaching of alkalis and alkali earths under surface conditions and also leads to the transformation of smectite/illite and other phyllosilicates to kaolinite.

5.3.4. SMECTITE/ILLITE

5.3.4.1. INTRODUCTION

The most abundant clays in the upper part of the Cadna-owie Formation and the Bulldog Shale are authigenic smectite and illite and allogenic mixed-layer illite/smectite (I/S) series. The clays become rare in the bottom part of the sequence (Fig.5.1)
Crystal morphologies and chemical composition of the clays are similar to those from clays described in several places in the world. Authigenic illite and smectite are the predominant clays in the silty-sandy layers and in bioturbated zones of the Cadnawie Formation and the Bulldog Shale.

5.3.4.2. OCCURRENCE OF SMECTITE/ILLITE

Authigenic smectite and illite (allogenic mixed layer I/S) are mainly recognized in the marine transgressive sequences of the southern Eromanga Basin. Scanning electron microscopy of the lower unit of the Bulldog Shale in Toodla #1 at 234.75 m depth (I/S morphology in interbedded silts and shale) showed that allogenic I/S is predominant in the shale layers while authigenic smectite and illite dominate in the silty layers and are strongly developed as pore linings and pore fillings. These probably are a result of variable compaction between shale and silt-sandstone.

Moreover, in some bioturbated layers there are other forms of detrital illite or biogenic pellets (fecal pellets). In hand specimens, they have a green color indicative of glauconite. They also have the spectrum and show the illite morphology under the SEM (Plate 7e, 7f). This evidence indicates a transitional phase from illite to glauconite. The burrows are also filled with pyrite frambooids.

5.3.4.3. CHEMISTRY AND MORPHOLOGY OF SMECTITE/ILLITE

The smectite/illite mixed-layer clay mineral is represented in the powder X-ray diffraction of clay fractions. The illite is identified by a small sharp peak at 10.05 Å, and is more abundant in core samples in both the 2 μm fraction and whole rock analyses than in outcrop samples. This possibly results from burial conditions. Smectite is identified by 15.5 Å-17.1 Å peaks and referred to as 17 Å.

Allogenic smectite/illite occurs with a flaky morphology. It often has a preferred orientation and lies parallel with the silty layers. However, bioturbation has changed the fabric of some shales (Plate 8a). The study of Bulldog Shale samples shows that increasing potassium content and flaky I/S lead to decreasing proportions of expandable layers. The allogenic clays in the sediment of the marine transgression have mixed layers ranging from 65-45% expandable layers.

The authigenic smectite in the Bulldog Shale has cellular, oakleaf and cornflake morphologies. Individual crystals of smectite cannot be resolved under the SEM. They appear as a thin honeycomb crust on detrital sand grains or growths outward from the grains in pore linings (Plate 8b). Authigenic illite occurs mainly as pore linings within silty layers. It has scallop morphology and curled edges. Its size ranges from 2-8 μm,
and it often aggregates with pyrite frambooids (Plate 8c). Authigenic smectite and illite have a high degree of purity (i.e. 80-90% of smectite, 80-90% of illite) (Plate 8d). Semiquantitative analysis of the authigenic smectite yielded major elements corresponding to the composition of Na-Ca smectite \((1/2\text{Ca,Na})_0.7 (\text{Al,Mg,Fe}) (\text{Si,Al})_8\text{O}_{20} (\text{OH})_4 n\text{H}_2\text{O}\), while illite shows a true potassium-illite member (Plate 8e) similar to that described by Welton (1984).

### 5.3.4.4. FORMATION OF SMECTITE/ILLITE

The allogenic I/S in the Early Cretaceous sequence is clearly from the pre-Mesozoic Basement and also indicates volcanic sources. Study of I/S has shown the low to moderate iron content in smectite possibly reflects an andesitic or felsic volcanic clastic. In addition, I/S is associated with chemically immature sediments of the Aptian and Albian. This suggests the sequences were either rapidly buried in marine conditions or accumulated under extremely unfavourable climates for chemical weathering, such as cold or arid conditions.

Conversion of smectite to illite is well known as illitization (Perry and Hower, 1970; Keller et al., 1987). The process of illitization in the Jurassic-Cretaceous sediments in the southern Eromanga Basin possibly relates to the dissolution, K-fixation and morphological conversion of smectite. Other factors are illitization, pressure, temperature and burial history. The illitization from smectite releases a large amount of intergrain water by the dissolution of smectite. The water may carry a large quantity of material leading to the formation of oil, gas and authigenic minerals (Burley et al., 1987).

### 5.3.5. Zeolite

#### 5.3.5.1. INTRODUCTION

Clinoptilolite is the dominant mineral of the zeolite facies in the Cretaceous sediments. It commonly replaces detrital plagioclase or infills pore space. Zeolite in several places throughout the world is associated with volcanic activity, and it is also considered as an authigenic mineral in sandstone and shale (Kastner and Siever, 1979). Clinoptilolite and several other zeolite minerals in the Cadna-owie Formation and Bulldog Shale are possibly representative of the alteration of volcanic clastics in the Cretaceous sediments. Also, Ca-clinoptilolite can be related to the dissolution of plagioclase in the rocks. The mineralogy of zeolite in the Cretaceous sediments in the southern Eromanga Basin is similar to several other zeolite occurrences in silicic
Plate 8a.
A worm burrow has changed the fabric of illitic shale (I) in the bottom left corner of the photo; this was later infilled by massive pyrite (P) framboids.

Plate 8b.
Split image of authigenic smectite (S). The smectite infills pore space as pore lining. EDS analysis yields Na, Ca, Si, Al, Fe and a little Mg. This corresponds with the composition of the Na-Ca smectite series.

Plate 8c.
Authigenic illite (I) and pyrite framboids (P). Illite shows cornflake and hair-like morphologies, the crystals grow as linings on the surface of detrital grains.
EDS shows the major elements of a true illite Si, Al, K, Mg and Fe.

Plate 8d.
Split image of authigenic smectite (S). The smectite shows honeycomb features and individual smectite crystal cannot be resolved. EDS analysis shows the spectrum of a Na-Ca smectite series.

Plate 8e.
Illite (I) shows cornflake morphologies and has a composition of potassium end members of true illite. Estimated 80% of illitic layers.

Plate 8f.
Split image of authigenic clinoptilolite(C); the authigenic infills primary pores and dissolution porosity. Identification based on the EDS analysis yields the major elements Si, Al, Na, Ca and Fe corresponding to the chemical composition of clinoptilolite (see text).

5.3.5.2. OCCURRENCE OF ZEOLITE

Zeolite is a ubiquitous feature in the shaley layers of the upper part of the Cadnaowie Formation and the Bulldog Shale in the southern Eromanga Basin (Norton, 1983). The most common occurrence of authigenic clinoptilolite was as the euhedral crystal shapes, and lining of pore spaces or pore filling in silty layers of the Bulldog Shale (Plate 8f, 9a). At high magnification clinoptilolite crystals are tabular or coffin-shaped. Some individual crystals have dimensions of about 1x5x20 μm (Plate 9c). They are aggregated among authigenic K-feldspar crystals and are usually coated with a thin layer of authigenic smectite/illite. With this combination, identification of the morphology of the two minerals is often difficult. However, EDS analyses show that authigenic K-feldspar is a pure end member of K-feldspar and yields the major elements potassium, silica and aluminum. On the other hand, clinoptilolite contains all the major elements, such as sodium, silica, aluminium, calcium and potassium, with the addition of iron and magnesium coming from the clay coatings (Plate 9d).

5.3.5.3. CHEMISTRY

The identification of clinoptilolite is based on the morphology and semiquantitative analysis by EDS and X-ray diffraction. However, application of X-ray diffraction techniques was difficult due to the small proportion of clinoptilolite and the overlapping of peaks from other minerals. Hence, SEM and electron microprobe were the major techniques used in the study. Chemical composition of various zeolites are shown in appendix 2. From this result we can see that the dominant component of zeolite is clinoptilolite.

5.3.5.4. FORMATION OF ZEOLITE

The presence of clinoptilolite in the Jurassic-Cretaceous sediments of the southern Eromanga Basin possibly indicates a volcaniclastic origin for the source rock, even though the genesis of zeolite is the result of Gawler Ranges volcanic alteration. It is not apparently associated geographically with the regional sources. Norton (1983) reported an abundance of volcanic fragments in the Cadna-owie Formation and the Bulldog Shale (Toodla #1). Volcaniclastic material is relatively unstable at the earth's surface.
Consequently its constituents react readily with their pore waters to form a variety of authigenic minerals. Alteration of volcaniclastic andesite yields smectite from augite and plagioclase (Glasmann, 1982). It is well known that the alteration of basalt produces smectite, clinoptilolite and other zeolite facies (Benson and Teague, 1982). These researchers also reported small amounts of mordenite replacing the clinoptilolite. The alteration of basalts to these various authigenic minerals can be related to low-temperature ground-water conditions, and the formation of zeolite therefore is possible within 1000 m of the earth's surface and without involving high temperatures (Benson and Teague, 1982; Burley et al., 1987).

Kastner and Siever (1979) proposed a model for volcaniclastic components in marine conditions (Fig. 5.3). In natural sea water, the trapped sea water in volcaniclastic components reacts with a mixture of volcaniclastic material and pelagic clay, leading to the release of silica, aluminium, potassium and sodium. In other words, dissolution of feldspar silica, aluminium, potassium and sodium. In other words, dissolution of feldspar and other unstable minerals releases large mobile ions from the host rocks. The release of K⁺ may increase the K⁺/H⁺ ratio into the K-feldspar stable field of Na⁺ and K⁺, or it may be increased by the hydrolyzing glass as it converts to zeolite and smectite. However, at low temperature the effect of releasing K⁺ is small and would not alter the Na⁺/K⁺ ratio which could lead to the formation of zeolite in marine sediments.

![Figure 5.3 Diagenetic model for marine-volcanic clastic formation of authigenic mineral (Kastner and Siever, 1979).](image-url)
Thus, the presence of zeolite in the Cadna-owie Formation and the Bulldog Shale means that there is proportionally less authigenic K-feldspar in marine volcaniclastic conditions. The critical factors of conversion of these minerals are not known. However, in sediments older than about 200 my., zeolites and smectites are rarely found as the result of further alteration to an illite-chlorite-feldspar assemblage.

5.3.6. Glendonite

5.3.6.1. INTRODUCTION

Glendonite belongs to a group of calcite pseudomorphs of ikaite and is regarded as a syngenetically-formed mineral from organic rich muds at sub-zero temperature within high latitudes or polar environments, such as Greenland and Antarctica (Danish, 1963; Kemper and Schmitz, 1975, 1981; Suess et al., 1982; Kemper, 1987).

The occurrence of glendonite within the Cadna-owie Formation and the Bulldog Shale led to a detailed study being carried out as part of this project; the results support the palaeoenvironmental reconstruction of Fennessy (1984). The occurrence of glendonite indicates a cold climate similar to the polar arctic region (Canada, Siberia). It is necessary to describe the glendonite mineral, which is suitable for evidence as a thermometer in geological history.

5.3.6.2. PREVIOUS STUDIES

David (1905) described glendonite as a pseudomorph mineral occurring in Permian glacial sediments and possibly the Cretaceous ice rafted unit in New South Wales. However the original mineral was known as glauberite.

In 1963, Pauly discovered that ikaite was the original mineral of glendonite, when he found ikaite at the surface at low tide in the waters of Ika Fjord, Greenland. The samples rapidly disintegrated under warmer conditions. Pauly believed ikaite was formed by bicarbonate springs at the bottom of Ika Fjord.

Brook et al., (1980) described ikaite as a low temperature calcium carbonate mineral, but it is unstable and at normal temperature becomes calcite and water. Brook et al also suggested that ikaite might have widespread occurrence in polar marine and high latitude cold-water regions.

Suess et al., (1982) found ikaite on the sea bottom at sub-zero temperatures in Antarctica. They described crystallization of the mineral as monoclinic-symmetrical crystals which disintegrate rapidly under surface conditions. Stein and Smith (quoted by Shearman and Smith, 1985) found ikaite in the sediment of the Nankai Trough in
Plate 9a.
Aggregates of clinoptilolite (C) among pore space and dissolution porosity (secondary porospace). Identification based on the EDS analysis yields the major elements Si, Al and Na corresponding to the chemical composition of clinoptilolite (sample 889.234.75).

Plate 9b.
Possible phillipsite (PL) crystals; note that the euhedral crystals infill secondary pore spaces. Identification based on the EDS analysis shows spectrum of Na, Ca, K, Si, Al and little Fe corresponding to the chemical composition of phillipsite (sample 889.234.75).

Plate 9c.
Close look at authigenic clinoptilolite crystals (C). Individual clinoptilolite crystals are small (5-15 μm), tabular shaped (sample 889.195.9).

Plate 9d.
A contrast of authigenic clinoptilolite (C) and an authigenic K-feldspar (F) EDS analysis on K-feldspar yield the major elements Si, Al and K, whereas clinoptilolite yields Si, Al, Na. In addition Fe and Ca are yielded by clay coating (sample 889.218.40).

Plate 9e.
Microglendonite (GL) coats the surface of detrital quartz grains; note the polar-euhedra crystals. Identification is based on the EDS and the XRD (sample 889.266.5) (see text).

Plate 9f.
Split image of a hedgehog microglendonite (GL) (sample 889.266.5).
Japan at a great depth under cold conditions, and also proposed a stable field of ikaite under natural conditions. Kemper and Schmitz (1975, 1981) and Kemper (1987) compared the results of palaeontological studies on glendonites of Cretaceous age in Siberia and the Canadian Arctic Region. From an account of all known glendonites, Kemper (1987) concluded that glendonite is the pseudomorph of calcite, and that glendonite can be used as evidence of a polar marine depositional environment. These workers thus showed that glendonite is a geological thermometer.

5.3.6.3. OCCURRENCE OF GLENDONITE IN THE EROMANGA BASIN

Micro-siderite concretions occur in the Cadna-owie Formation (Toodla #1). These have a similar morphology to the macroscopical-glendonite which appears in outcrop samples near Coober Pedy. Powder X-ray diffraction studies of these glendonites show that siderite and bassanite are the major pseudomorph minerals. Macroglendonite shows very high crystallization in the X-ray diffractogram. Major peaks appear at 6.02 Å; 3.47Å; 3.00Å; 2.72Å; 1.84Å. According to JCPDS these peaks are identified as bassanite. However, this could be a result of the dehydration of gypsum under arid conditions in the area. Thus, the original pseudomorph mineral is more likely to have been gypsum than bassanite. In addition, dehydrating gypsum (CaSO$_4$) nH$_2$O at temperatures above 45°C would give bassanite Ca$_2$(SO$_4$).nH$_2$O (Rowe and Bigelow, 1972. Quote by JCPDS).

The most common appearance of microglendonite is as the polar euhedra pyramidal and hedgehog crystals of siderite pseudomorph. Scanning electron microscopy of samples from the Cadna-owie Formation (889.273.5) shows that most of the detrital grains are coated by microglendonite (Plate.9e, 9f), ranging in size from a few μm up to 30 μm. They occur as embedded crystals in fine-grained clay sediments and parent crystals grew displacively. Hence the original mineral is considered of authigenic origin and was possibly generated by reactions with interstitial waters.

5.3.6.4. MORPHOLOGICAL STUDIES

Macroglendonite appears as "hedgehog" concretions on macroscopical studies. They range in colour from white-brown to yellowish-grey and their sizes vary from 0.5 to 5 cm. They occur as spherical clusters. The terms pineapple" (David et al., 1905) "pola euhedra" and "hedgehog" (Kemper and Schmitz, 1975) were used as genetic terms. The morphology of glendonite in the Cretaceous sediments is also similar to those described by Shearman and Smith (1985) as summarized in the
previous studies. Overall, individual crystals have a pyramidal-appearance which is seen as asymmetry in the crystal shapes, and incline slightly in opposite directions. The pyramidal crystals are clearly recognizable in all the samples, and are roughened by series of gypsum crystals. This is evidence of material in the pseudomorphous crystalline discontinuity (David et al., 1905). Measurement of the angles between pyramidal crystals give 95°, 60°, and 125° respectively, suggesting a monoclinic system for the macroglendonite.

Scanning electron microscopy gives the best microphotographic reproduction of microglendonite. At high magnification, individual pyramidal shapes are recognizable as elongated crystals, which are similar to those crystals described on the macro scale by Shearman and Smith (1985) and Kemper (1987). The hedgehog form of many of either pinacoids or prismatic faces (Plate 10a) occurs as in the macro scale. The crystallised system does not correspond to that of the trigonal class of symmetry, but either the tetragonal or the orthorhombic system. Some of the individual pinacoidal crystals have crossing angles which might indicate the presence of either twinning or parallel intergrowths (Plate 10b).

5.3.6.5. FORMATION OF GLENDONITE

Glendonite (pseudomorph of ikaite) has been described as a syngenetic shallow epicontinental sea mineral, and the presence of glendonite appears to be a reliable indicator of low temperature (0°C or less).

Glendonite is often accompanied by glaciomarine sediments, which were widespread during the Cretaceous in the Arctic regions (Kemper, 1987). However, Shearman and Smith (1985) believed that ikaite was formed under cold conditions at great depth.

Marland (1975) experimentally studied the stability of ikaite facies and proposed the stability field relations of ikaite with respect to temperature and pressure (Fig.5.4). During diagenesis, a change from the original environment would cause rapid dehydration and conversion of monoclinic ikaite to calcite. This was considered as the process of formation of glendonite pseudomorphs (Suess et al., 1982).

![Figure 5.4 Stability relations of ikaite with respect to pressure and temperature (After Marland, 1975).]

\[ \text{IKAITE} \quad (\text{CaCO}_3.6\text{H}_2\text{O}) \quad (\text{Monoclinic}) \]
\[ \text{ARAGONITE} \quad (\text{CaCO}_3) \quad (\text{Or thorohombic}) \]
\[ \text{CALCITE} \quad (\text{CaCO}_3) \quad (\text{Hexagonal/Trigonal}) \]
Plate 10a.
A hedgehog form of microglendonite (GL); note the hedgehog formed by small individual prismatic crystals (sample 889.266.5)

Plate 10b.
Polar euhedra crystals show twin crystals of microglendonite (GL). (sample 889.266.5)

Plate 10c.
Split image of a worm burrow; note that it is infilled by massive pyrite frambooids (P) (sample 889.195.8).

Plate 10d.
Split image of a worm burrow, infilled by small octahedral pyrite crystals (P) (sample 889.189.1).

Plate 10e.
Closer look at surface features of pyrite frambooids (P). Note pyrite frambooid formed by small individual euhedral-octahedral crystals (sample 889.195.8).

Plate 10f.
Weathered pyrite framboid (Fe). The framboid has been corroded by weathering processes and yields only Fe under the EDS analysis. It appears among smectite. (sample 889.85.08).
Furthermore glendonite occurs in Cretaceous sediments of the southern Eromanga Basin within a probable depth range of 50 to 200 metres (chapter 2). The presence of glendonite in the Cretaceous sediments is good evidence of a polar cold climate during at least some parts of the Cretaceous.

5.3.7. Pyrite

5.3.7.1. INTRODUCTION

Sedimentary pyrite is characteristic of authigenic minerals in anoxic muds and sulphate-reducing conditions. It is widespread in the Cretaceous marine-transgressive sequences of the Eromanga Basin. Pyrite normally forms during early diagenesis and it is one of the best known authigenic minerals (Berner, 1981).

5.3.7.2. OCCURRENCE OF PYRITE

In hand specimens and thin sections, pyrite occurs as a thin, black mass adjacent to detrital quartz or clays. Pyrite-infilled worm burrows from the Bulldog Shale and the upper part of the Cadna-owie Formation sediments were examined by X-ray diffraction. On the X-ray diffractogram, the d-spacings of 3.13Å, 2.70Å, 2.21Å and 1.91Å were identified as pyrite (JCPDS). However, scanning electron microscopy shows that smectite/illite occurs together with pyrite crystals. The pyrite occurs either as framboids or single euhedral crystals. Pyrite framboids are round, occasionally oval-shaped and often infill worm burrows. Pyrite framboid aggregates are composed of small, closely-packed octahedral crystals. Small euhedral forms of pseudooctahedral pyrite are also randomly distributed through the rock. In addition, large pseudooctahedral pyrite crystals are sometime aligned, possibly replacing worm tube burrows. Their size ranges from a few μm to 15 μm (Plate 10c, 10d). At high magnification, pyrite framboids consist of a cluster of many small individual pseudooctahedral pyrite crystals (Plate 10e). The euhedral crystal faces of octahedral pyrite show its authigenic origin. In most outcrop samples, the pyrite shows highly weathered surfaces, corroded into crystal shapes (surface oxidation of telogenesis) and yields iron oxide traces (possibly goethite) under the EDS (Plate 10f)
5.3.7.3. FORMATION OF PYRITE

Formation of pyrite is well known as originating in the reaction between $\text{H}_2\text{S}$ from bacterial sulphate reduction and reactive detrital iron minerals (Berner, 1964, 1970, 1972; Goldhaber and Kaplan, 1974).

The amount of pyrite formed in a sediment depends on three factors (Fig. 5.5). These are, the availability of dissolved sulfate, the concentration of organic matter which can be metabolized by sulfate reducing bacteria, and the concentration of iron minerals in the environment. These three factors vary from fresh water to marine and brackish water environments. In fresh water conditions, formation of pyrite is limited by low concentration of dissolved sulphate.

In addition, oxidation of organic matter in the fresh water environments tends to inhibit the precipitation of pyrite. However, in normal marine conditions, formation of pyrite is also limited by the amount and reactivity of organic debris buried in sediment.

The reaction between pyrite and organic debris is considerable. This reaction may partly dissolve pyrite, leading to the formation of minor pyrite framboids in the laminated black shale facies. But, in anoxic marine conditions (brackish lagoonal facies, etc.) the precipitation of pyrite is greatly enhanced by a plentiful supply of organic debris and $\text{H}_2\text{S}$ plus the abundance of iron minerals (smectite, glauconite,) brought to the site of pyrite deposition.

Nevertheless, brackish water sediment represents an intermediate situation between fresh water and marine environments. If the water is insufficiently saline, sulfate will be dissolved. The result is that siderite has a high probability of forming (Berner, 1981). Berner (1974) reported formation of siderite in the Baltic Sea similar to the described environment during the last glaciation. Thus, the occurrence of pyrite is
evidence of sea level change during the Cretaceous in the southern Eromanga Basin (from fresh water to marine and brackish environment).

5.3.8.  Glauconite

5.3.8.1. INTRODUCTION

Glauconite is common throughout the upper part of the Cadna-owie Formation and in the Bulldog Shale. The term glauconite as used here, was defined by Millot (1970) as "a variable mixture of minerals in which illite, montmorillonite, chlorite or various mixed layer clays can be defined. The total aspect remains that of green clay minerals, but it is not known whether one part or another yet merits the name glauconite". pp 205. The Occurrence of glauconite has been used both as a stratigraphic tool and as an indicator of depositional environment.

5.3.8.2. OCCURRENCE AND CHEMISTRY OF GLAUCONITE

Glauconite is abundant in the marine sequence of the Cadna-owie Formation and Bulldog Shale (sample 389.189.5 contains up to 30% glauconite). The first glauconite was recorded at a depth of 258.5m in the core and becomes more abundant in the Bulldog Shale and Oodnadatta Formation.

Glauconite in the Cretaceous sediments can be grouped into two components (a) those which are biogenic or fecal pellet and (b) those which have the morphology of detrital illite/smectite (glauconitic illite). Fecal pellet is the common type found in outcrops samples and in some areas can make up to 20% of the sediment. They are generally of medium sand size, and are considered as palaeontological evidence of marine or paralic environments (Exon and Senior, 1976).

Powder X-ray diffraction results (chapter 3) show peaks at 10Å and 16Å after glycolation, indicating a clay with highly variable hydration properties. Under the SEM, glauconite shows two types of morphologies: fecal pellets and detrital illite. At high magnification, the fecal pellets show very well rounded, polished surfaces and yield a glauconite spectrum under EDS. Detrital illite also seems to occur as well rounded particles. But at high magnification it appears flaky, or cornflake-like, which is typical of the detrital illite morphologies series, and yields a glauconite spectrum under the EDS (Plate 11a, 11b). However, X-ray diffraction of both types does not show the peaks for glauconite (only smectite/illite/). Thus, glauconite in these Cretaceous sediments is crystallographically poorly defined. Chemical analyses of glauconite (table 7 and appendix 2) shows that glauconite from the transgressive marine
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<td>100.03</td>
<td>94.33</td>
<td>91.69</td>
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1. Deer et al. (1966)
4. Bulldog Shale, Toodla #1, Oodnadatta, central north of South Australia
sequence of the Mesozoic sediments lies in-between a true glauconite and illite (Deer et al., 1966).

5.3.8.3. FORMATION OF GLAUCONITE

These results suggest that the glauconite from the Bulldog Shale of the Eromanga Basin consists of detrital mixed-layer illite/smectite, which possibly came from the basement rocks. The occurrence of similar types of glauconite is also reported by Hower (1961) and Bornhold and Giresse (1985).

5.3.9. Gypsum

5.3.9.1. INTRODUCTION

Secondary gypsum is a dominant feature in the studied area. It only appears in surface and outcrop samples, and has been well described by many workers such as Jones and Segnit (1966); Nicol (1979) and Barker (1983).

5.3.9.2. OCCURRENCE OF GYPSUM

There is widespread occurrence of gypsum in a large area of central-north South Australia. Scanning electron microscopy has shown that gypsum crystals infill pore spaces, replace clay minerals and have euhehedral shapes.

Powder X-ray diffraction of gypsified silty-shale samples (Marree Formation) yields major peaks at about 7.58Å, 4.25Å. These peaks indicate the presence of well crystallised gypsum. Under the SEM, gypsum shows a fibrous appearance in the form of patchy rosettes or fan shapes. These are formed by long prismatic lath-shaped crystals ranging from 5 to 20 µm (Plate 11c, 11d). Moreover, supplementary identification by EDS yields calcium and sulfur of nearly equal peak heights. These peaks correspond to the chemical composition of gypsum (CaSO₄).

5.3.9.4. FORMATION OF GYPSUM

Secondary gypsum of the Mesozoic sediments in the southern Eromanga Basin relates to the weathering of pyrite-rich mudstone and limestone (Dr. J.B. Jones pers. coms.). Ground water chemistry of the Coober Pedy region features very high concentrations of calcium and sulphate (Barker, 1983). Gypsum infills pore spaces and secondary pore spaces. It replaces wood fragments and overgrows most minerals in the samples.
Plate 11a.
Glauconite (G) forming a fecal pellet. Note the rounded and polished surfaces, possibly created by bioturbation (sample 889.85.7).

Plate 11b.
Split image of glauconite (G) The detrital grain is aggregated among smectite/illite. The morphology of illite is identified by flaky crystals, but EDS analysis yields a small fraction of Ti and Fe peaks higher than true illite (sample 889.224.45).

Plate 11c.
Split image of fibrous gypsum (Gy) forming fans of individual lath-shaped gypsum crystals. EDS analysis yields major elements of Ca and S corresponding to the composition of gypsum (sample 889.DBC).

Plate 11d.
Secondary gypsum (Gy) infills pore spaces. The gypsum is formed by individual fibrous gypsum crystals. Close look at gypsum fibres shows that individual gypsum crystals have rod shapes which are parallel to each other.

Plate 11e.
A complex composition of the Bulldog Shale: a twin authigenic K-feldspar (F) on top of the photo; an erionite (E) on the left; quartz (Q) growth-crystal prism faces on the bottom left (large and small crystals); detrital ilmenite (IL); dissolution of K-feldspar and authigenic smectite (S) cover on top.(sample 889.224.45).

Plate 11f.
A detrital grain of muscovite (M) shows angular features and good cleavage (sample 889.CPB 1/2).
Thus, the occurrence of gypsum is as a secondary mineral rather than a syngenetic form, and clearly represents a product of the telogenetic regime.

Garrels and Lerman (1981) proposed a genesis model for sulfide species and considered that weathering of sedimentary pyrite and CaSO₄ contributed to the ocean reservoir. The ocean forms the same minerals. The redox imbalance resulting from a net shift of sulphur between the CaSO₄ and pyrite reservoirs is taken care of by concurrent shifts between organic carbon and calcium carbonate reservoirs. This can be chemically expressed by the reaction:

\[15H₂O + 8CaSO₄ + 2Fe₂O₃ + 7MgSiO₃ \rightarrow 4FeS₂ + 8CaCO₃ + 7MgCO₃ + 7SiO₂ + 15H₂O.\]

From the reverse equation, we can see that the oxidation of pyrite and calcium carbonate leads to the formation of gypsum in the studied area. The reaction is known as the processes of sulphur/sulphate reaction (Berner, 1984).

5.3.10. Minor components

The minor components consist of detrital apatite, detrital ilmenite and micas. The minor components were mainly identified by EDS under the SEM (Plate 11e, 11f, 12a, 12b, 12c, 12d). Both ilmenite and apatite show variable sizes, shape and degree of rounding. Mica is dominated by muscovite, but biotite was also recorded in shale facies at a depth of 258m. Although these minerals occur in small numbers and have been little affected in the sedimentation, their occurrence provides more evidence about the depositional environment.

Powder X-ray diffraction analyses of bulk samples from the Stuart Creek Opal Diggings also indicate that muscovite is abundant in the Marree Formation. It is identified by strong peaks at 9.97Å, 4.94Å, 4.45Å, 3.34Å, 2.55Å, 2.29Å (according to JCPDS). Under the SEM muscovite has mainly angular shapes. Individual cleavage plates of mica appear as flakes and the EDS spectrum contains major elements of silica, aluminium, potassium and iron which correspond to the chemical composition of muscovite \(K₂Al₄Si₆Al₂O₂₀(OH,F)₄\) (Plate 12e).

Muscovite is the predominant mica in all samples from the Trinity Well Sandstone, the Pelican Well Formation and the Marree Formation. Powder X-ray diffraction analysis of the sand fractions from PWN₃ yielded well-defined muscovite peaks of 9.98Å (according to JCPDS). Scanning electron microscopy demonstrated that the crystal morphology of biotite and muscovite are similar. Individual cleavage layers of mica appear as flakes. It was possible to differentiate between biotite and muscovite with the EDS. Biotite consists of silica, aluminium, potassium, magnesium, iron, and titanium, whereas muscovite contains only a minor amount of iron and does not show
Plate 12a.
Detrital ilmenite (IL). Identification is based on the EDS analysis which yields major elements Ti and Fe corresponding to the chemical composition of ilmenite (sample 889.218.5).

Plate 12b.
Calcium phosphate (CP) developed among smectite (S) (sample 889.CPB. 1/2)

Plate 12c.
Detrital apatite (A) with illite (I), note the elongated crystal shaped and rounded edges, possibly created by volcanic activity (sample 889.224.45).

Plate 12d.
A volcanic fragment (V) with detrital quartz and dissolution feldspar (F). The volcanic is identified by the needle shapes of small TiO₂. Electron microprobe, backscattered electron image (sample 889.195.8).

Plate 12e.
Intergrowths of muscovite (M) and chlorite (Chl); note that muscovite appears brighter under backscattered electron image corresponding to the composition of muscovite. Pyrite frambooids (P) appear rounded in shape and extremely bright; dissolution of feldspar (F) is infilled by clays (sample 889.224.45).

Plate 12f.
Detrital muscovite (M) is infilled by authigenic feldspar (F) and silica (deformation of detrital shape). EDS analysis yields major elements Si, Al, K and Fe corresponding to the composition of muscovite (sample 889.86.5).
magnesium in the spectrum (Welton, 1984). The muscovite is often overgrown by other authigenic minerals (Plate 12f) such as potassium feldspar, illite and silica or quartz. Calcareous-cemented-fine-grained sandstone (889.12) from the hummocky cross stratified unit at Trinity Well shows detrital grains overgrown by authigenic feldspar.

Siderite is identified by X-ray diffraction with peaks about 3.59Å, 2.79Å, 2.34Å, 2.13Å, 1.96Å and 1.79Å. Under the SEM the siderite is represented as euhedral crystals forming glendonite nodules (as mention earlier). Identification of these minerals was made by the methods mentioned above.

5.4. DISCUSSION AND CONCLUSION

The Jurassic-Cretaceous sediments consist of two mineralogic assemblages and cover a wide range of depositional facies. Furthermore, diagenetic features can be distinguished from primary ones (Burley et al., 1987).

The classical interpretation of the change from kaolinite to montmorillonite is related to the change from fluvial to marine environments (Wopfner et al., 1970; Exon and Senior, 1976; Griffith, 1979). Weaver (1967) notes that a change from low to high kaolinite content represents a change from a relatively more open sea to a near-shore marine environment. However, Segall et al. (1988) detailed the study of clay mineral indicators of geological subaerial modification near surface Tertiary sediments on the northeastern Grand Banks of Newfoundland. This work has demonstrated that clay mineralogy responds to sea level changes during geological time. Segall et al. noted that fall in sea level led to the alteration of marine sediments (smectite) by acid leaching to form a kaolinite enriched weathered zone; a transgressive stage led to the reworking and redeposition of early weathered zones. The clay mineral assemblage input by glaciers from northern latitude provenances includes illite, chlorite, smectite and minor components of kaolinite. These sediments are also associated with sands and gravels from ice rafting. A model for the formation of the Jurassic Cretaceous sediments on the southern margin of Eromanga Basin is suggested and summarized in figure 5.6.

From this model, two chemical environments have been identified on the basis of clay mineralogical analyses: a fluvial zone exemplified by an acidic environment for the Late Jurassic-Neocomian; and a marine shelf deposition zone indicated by an alkaline environment during the deposition of the thick sequence of the upper part of the Cadna-owie Formation and the Bulldog Shale.

Clay minerals from the Algebuckina have a kaolinite component, which is absent in the overlying late Aptian-Albian. The high kaolinite concentrations are suggested to
Figure 5.6: Schematic section of basal facies relationships from the late Jurassic to the Tertiary period of southern margin of the Eromanga Basin
have been formed from the alteration of Precambrian sediments. The Algebuckina Sandstone sediments are clay-poor sands and gravels that were deposited during the regressive event. Local differences in stratigraphy and sedimentary textures compared with later marine strata from the area are also confirmed to be of fluvial type, subaerially exposed during the Late Jurassic.

In marine environments (Bulldog Shale), alkaline sea water gave little potential for chemical reaction between the detrital mineral phases and normal marine pore water. Hence minerals in both fine-and-coarse-grained sediments were affected only by changes in pore water chemistry during initial burial or by reaction of the less stable detritus, including ferromagnesian minerals and feldspars (Kastner and Siever, 1979; Ali and Turner, 1982).

Diagenesis in the marine Bulldog Shale probably commenced immediately following deposition (Burley et al., 1987), and probably at the sediment water interface (Berner, 1981). At first, bacterial oxidation of organic matter released bicarbonate ions into solution. Further, sulphate-reducing bacteria also liberated bicarbonate ions into pore waters and continued to reduce pH by production of H₂S. Hydrolysis and dissolution of unstable minerals also occurred at this stage. As a result, authigenic minerals were rapidly formed near the sediment-water interface. Pyrite may form after these reactions (Berner, 1974) together with authigenic-mineral suites including illite/smectite, chlorite, K-feldspar, quartz and zeolite (Kastner, 1971; Segall et al., 1988). Thus, surface chemistry (eogenetic) of the Bulldog Shale diagenesis conforms to a predictable pattern which is summarized in figure 5.7.

The feldspathic greywackes contain authigenic smectite, and illite formed either by grain dissolution or directly from sea water (Elverhoi and Bjorlykke, 1978). The precipitation is probably the result of an increase in the ionic concentration of the sea water or the pore water as silica- and alumina-bearing solutions migrated from adjacent (or underlying sequence) shales to combine with potassium from interstitial sea water (Kastner and Siever, 1979). On the other hand, the breakdown of unstable detrital mineral grains overrode the subtler effects outlined above (Jones and Fitzgerald, 1987). Silica, aluminium, sodium, calcium potassium and magnesium ions may be released into solution during grain dissolution, leading to complex diagenesis in the marine Bulldog Shale. Authigenic chlorite may be formed as a result of the above released ions. But overall the I/S tends to be precipitated in preference to more illite. Glaucocrite (detrital I/S) may become more potassium-rich and evolve and mature until reaching an end member which is chemically and mineralogically a potassium mica.

Following the initial stage of alteration in the depositional environment (eogenesis), the rocks underwent further changes during diagenesis. This requires a change in physicochemical conditions, including temperature, pressure and pore fluid
FIGURE 5.7. SCHEMATIC OF DIAGENETIC PATHWAYS OF THE MARINE BULLDOG SHALE

- **Anoxic Ground Water**
  - Chlorite
  - Quartz overgrowths
  - Siderite

- **Oxygenated Ground Water**
  - Kaolinite
  - Quartz & feldspar overgrowths
  - Non-ferroan calcite

- **Brackish Marine Pore Water**
  - Smectite
  - (Smectite/illite conversion series)
  - Illite
  - Zeolite, feldspar & quartz
  - Siderite or non-ferroan calcite

**Introduction of Formation Water**

- Ferroan calcite and clays
chemistry under burial conditions (mesogenesis). Petrological study of the Jurassic-Cretaceous sediments indicate that the principle change from the eogenetic to the mesogenetic regime is a change from friable to slightly cemented and compacted sediment.

Illite/smectite and probably glauconite follow the path shown in figure 5.8. Illitization releases silica, calcium, sodium, iron and magnesium into the pore waters which may be capable of transporting dissolved ionic species from mud rock into sandstone. Silica and aluminium may be precipitated as quartz overgrowths and authigenic kaolinite, whilst iron and magnesium may contribute to the formation of chlorite and siderite.

Shale and silt, when initially deposited, have porosities in excess of 80% (Burst, 1969). Much of this porosity is lost during early burial and water is expelled by mechanical compaction. Thus illitization and compaction led to water (carrying ionic species) migrating upward under the pressure gradient and precipitating minerals at higher levels in the section (Burst, 1969; Parson et al., 1982). Interbedded shale and silt layers react in different ways. During compaction the orientation of clay particles in shale changes from a random to an oriented direction which is parallel to basal spacings. The spaces appear to have undergone intense compaction.

Zeolite and I/S cements are the most common features in the shaley layers of the upper part of the Cadna-owie Formation and the Bulldog Shale. It is inferred that the shaley facies have less permeability due to large percentage of fine matrix which clogs pores. The prevention of water movement or basinal compaction (buried condition of Toodla #1) are important factors in the reduction of porosity of the Bulldog Shale. However, in the sand rich facies, pore fluids which moved through the sandstone pore spaces precipitated calcite and siderite in acid to moderate pH condition. In alkaline conditions (marine), smectite and illite freely infill pore spaces and secondary pore spaces. Thus, smectite and illite are the major porosity destroyers in these silty-sandy
marine sediments. The formation of matrix and compaction precluded the development of cements that are common in the Mesozoic sediments of the southern Eromanga Basin.

Rocks exposed at the surface suffer from weathering processes. Most minerals formed under burial conditions (elevated temperature and pressure) become unstable in oxidic and acidic conditions. Alteration first affects organic matter and organic matter deposited under euxinic conditions during burial becomes oxidized and carbon is removed from the rocks (Curtis, 1976). The Trinity Well Sandstone is a highly permeable sandstone itself and meteoric water would have easily penetrated into the friable sediments. As a result, most of the organic matter was dissolved in water solutions. Siderite and pyrite oxidized to iron oxides, eventually producing limonite and goethite, whilst the oxidation of sulfides resulted in the formation of sulfate as gypsum during sub-aerial exposure beneath the Tertiary unconformity (Jones and Segnit, 1966; Baker, 1983). In part, the rock was exposed under oxygenated water conditions. This could explain the presence of kaolinite and alunite occurring as secondary-authigenic minerals in the weathered zone of the Bulldog Shale.

Another form of kaolinite in the oxidation zone is authigenic kaolinite of the Trinity Well Sandstone. Oxidization of detrital feldspars in a fresh-water environment leads to the formation of authigenic kaolinite (called kaolinitic sandstone), and it is the only potential source of acidity.

The alteration of the Jurassic-Cretaceous sediment at the surface represents not only the end of one cycle, but also the beginning under the present clastic diagenesis of arid and semi-arid conditions.
CHAPTER 6. ENVIRONMENTAL INTERPRETATION.

The overall environmental interpretation of the Jurassic-Cretaceous sediments is based on field studies, mineralogical examination and the result of the quartz sand surface texture studies of both outcrop and core samples. Although field work and the basic studies give an outline of environmental deposition, laboratory techniques allow refinement of the results from field work. In this interpretation, scanning electron microscopy, electron microprobe analysis, X-ray diffraction and petrography play a major role in palaeoenvironmental reconstruction.

6.1. ALGEBUCKINA SANDSTONE (VILLAGE WELL FORMATION EQUIVALENT)

The environmental interpretation of the Algebuckina Sandstone has been well described (Forbes, 1966, 1982; Wopfner et al., 1970; Exon and Senior, 1976; Griffiths, 1979; Norton, 1983). Wopfner et al. (1970) considered that the Algebuckina Sandstone was formed by a terrestrial-fluvial regime in the late Jurassic. The medium to coarse sandstone is shown here to contain kaolinite and other features characteristic of terrestrial-fluvial environments (see chapter 3). In addition, the appearance of quartz overgrowths, pore-filling kaolinite, altered muscovite, abundant precipitation of authigenic K-feldspar and dissolution of unstable grains such as detrital feldspars are typical features of the Algebuckina Sandstone.

The study of clay mineralogy of the Algebuckina Sandstone showed that the most common clay mineral is authigenic kaolinite, since matrix content is one of the most potent petrographic criteria in discrimination of various fluviodeltaic and marine environments (Davies and Ethridge, 1975; Wilson and Pittman, 1977). It can be concluded that the matrix of the Algebuckina Sandstone formed diagenetically by extensive dissolution of unstable minerals and precipitation of new stable minerals in a new geochemical condition. Detrital quartz and feldspars have to reach stable equilibrium with pore solution as the result of fluctuation of silica concentration in the pore water. The results are that both quartz and feldspar may have precipitated directly from pore water. These diagenetic processes have affected the mineralogy and chemistry of the sandstone (see chapter 5). The fluvi-deltaic sands partly change both macroporosity and grain size parameters.
6.2. CADNA-OWIE FORMATION (PELICAN WELL FORMATION EQUIVALENT)

Environmental interpretation of this formation is rather difficult due to the large variation in rock types. Wopfner et al. (1970), Griffiths (1979) and Kreig (1982) regard the Cadna-owie Formation as a transitional sequence of the Early Cretaceous. The lower boundary of the typical section of the Cadna-owie Formation is commonly marked by a disconformity representing the onset of the Lower Cretaceous marine transgression.

The first effect is that of widespread marine transgression. The marine environment gradually predominates together with a decrease of authigenic kaolinite in the Cadna-owie Formation as described in the Davenport Spring and Pelican Well sections and in Toodla #1 (see chapter 3). Exon and Senior (1976) stressed that the marine transgression also extended to the east, forming the Bungil Formation of the Surat Basin.

Microgldononite later pseudomorphed by siderite was formed in the early stage of marine transgression. This represents one of the major factors confirming climates in cold high latitude regions in the Early Cretaceous (Kemper and Schmitz, 1975, 1981; Kaplan, 1979; Kemper, 1983, 1987). Calcite is completely absent from the surrounding rock. The solubility of calcium at low temperatures is well known and ikaite is well known as a subzero-temperature mineral.

The studied hummocky cross stratification (HCS) sequence of the Cadna-owie Formation (Davenport Spring) shows a recurrent prograding of the shoreline and, as with HCS described by Bose et al. (1988), the Cadna-owie Formation HCS indicates the depositional environment was affected by diurnal tides (flood and ebb tide).

The fine-grained sandstone with shaley lenses occasionally contains scoured structures, but is more commonly structureless due to bioturbation (Pelican Well section), possibly indicating a lagoonal environment. Shallow water found behind barriers is ideal for the accumulation of organic matter, and favours worm burrows in brackish conditions. Flood-tidal deltas are clearly recognized as delta foreset stratification of the inlet mouth from the tidal beds. This could possibly indicate diurnal tides as today. Overall diverse facies ranging from boulder siltstone to biotubated organic rich silty-shale are suggested as

The presence of siderite and the lack of pyrite indicate an anoxic-nonsulphidic environment in the early marine transgression stage with fresh water influence and low salinity. Sulfate is dissolved close to the sediment water interface. There is a high concentration of iron in the environment leading to the formation of siderite (Berner,
Thus siderite is considered as an early diagenetic mineral. The marine environment fully controlled the later phases and is marked by the predominance of pyrite, because under marine conditions sulphate can only dissolve at greater depths and the concentration of sulphate is very high. As a result, pyrite formed in anoxic-sulphidic sediments instead of siderite.

Berner (1974) suggested that siderite may also have formed in Black Sea sediments during the last glacialiation when salinities were much lower than at present. Holdren (1977) (quoted by Berner, 1981) reported the same mineral assemblage from the brackish sediments of Chesapeake Bay, USA. Suess (1979) also reported a similar-mineral assemblage in the Baltic Sea.

The brackish lagoonal water of the Cadna-owie Formation represent an intermediate condition between fresh water and a marine environment. In the Neocomian fresh water, or perhaps low salinity lagoonal environment, the sedimentary facies were controlled. However, in the early Aptian (possibly Neocomian) the rise in sea level led to the formation of marine facies.

The classical transitional sequence of Wopfner et al. (1970) is the most suitable for the environmental interpretation of the lower part of the Cadna-owie Formation. However, at the top of the sequence its lithology is almost indistinguishable from the lower part of the Bulldog Shale. For this reason, the marine trangression stage will be discussed together with the formation of the Bulldog Shale.

The Stuart Creek Beds have been interpreted as a trangressional unit of early Aptian age. Brackish-water agglutinated foraminifera and marine glauconitic sandstones are indications of the early trangression in the southern Eromanga Basin. Moreover, confinement of the unit to the undulating surface of the Precambrian-Cambrian basement and the conformable relationship beneath the Marree Formation, suggest the Stuart Creek Beds are equivalent to the Cadna-owie Formation and/or the Pelican Well Formation.

6.3. BULLDOG SHALE (MARREE FORMATION EQUIVALENT)

The Bulldog shale is a marginal marine sediment (Wopfner et al., 1970). Sediments were deposited in shallow marine and lagoonal environments (along the shoreline) (Fig. 6). Both fine-and coarse-grained sediments were transported into the basin by stream and rafting by river-and-shore winter ice (to be discussed later). The sequence apparently became finer and there were fewer drop stones to the north. As true glacial deposits (tillites) have not been recorded. The glaciation was not intense. The epicontinental sea possibly communicated with the open ocean through a northern sea way in the Carpenteria Basin (Exon and Senior, 1976), and possibly to the
Southern Ocean through the Lake Eyre region to the Great Australian Bight (David, 1932). Thus, the Aptian-Albian sea of the Eromanga basin covered a large area in which fine-grained sediments were the predominant feature. This epeiric basin provides striking examples of various lithologic facies which may be models for other basins of the same age. Benthic fauna are found in the Bulldog Shale, but they probably reflect cold water conditions (Kemper and Schmitz, 1975) or an adaption which avoided strong currents (Reading, 1982).

Trace fossils are abundant. Locally high concentrations of organic matter in the Bulldog Shale indicate low oxygen levels and poor substrate conditions on the sea bed. They provide ideal hydrocarbon source rocks under burial conditions.

There possibly was volcanic activity in the Cretaceous. This is evidenced by the deposition of sand, silt and shale with volcanic fragments. Other evidence includes a considerable proportion of the illite/smectite-zeolite clay assemblages in the sand grade glauconite which are characteristic of the volcaniclastic Bulldog Shale and the upper part of the Cadna-owie Formation. These probably originated from the alteration of the pre-Mesozoic basement and much of the acid rhyolite and porphyry from the Gawler Range Volcanics, or probably from contemporaneous volcanic debris.

6.4. ENVIRONMENTAL INTERPRETATION OF CLAST BEARING SHALE (A DISCUSSION AND CONCLUSION)

6.4.1. Possible transport mechanism of the sediments

The origin (transportation and deposition) of the clast bearing shale (boulder shale) has been discussed by many workers in the literature. These include glacial origin (H.Y.L. Brown, 1894; Jack, 1915; Woolnough and David, 1926; Frakes and Francis, 1988). Erratic boulders could have been derived from reworking of older glacial deposits, such as the Sturtian Tillite which crops out near Marree (David et al., 1923; Howchin, 1928; Parkin, 1956). Emery (1955) showed that erratic boulders can be rafted by drift wood (trees, kelp, etc..) Wopfner et al. (1970) suggested the boulders originated on the shorelines, were reworked and transported downslope by slow sediment creep. Barnes and Scott (1979) interpreted the sandy conglomeratic beds as fluvial and suggested that erratic boulders were moved by wind gliding across frictionless mud surface. Flint et al. (1980) suggested that coarse sediment reworked along the shoreline was transported by low angle debris flows.

Organic rafting may have transported a large number of erratic clasts during the Late Phanerozoic (Emery, 1955, 1963). Emery proposed that clasts up to 3 to 4 metres in diameter have been found in the roots of trees floating in ocean currents. Since
fossilised wood is often found in the boulder shale, this type of raft could have operated. Kelp has also long been understood to carry erratic clasts. A large quantity of kelp can carry a large number of clasts for long distances. In cold water conditions, kelp can survive a longer time than in warm water. (Emery, 1955).

The organic rafting theory is unsatisfactory as the coarse sediment is far too abundant to have been transported solely by trees and kelp. Frakes and Francis (1988) show that the weight of many erratic boulders greatly exceeds that of the fossil wood. For example, to float the largest boulder (3 metres in long dimension) we need 9-10 m³ of driftwood.

The Cretaceous sediment contains a large amount of silt and sand. It is hard to explain how boulders could be carried in tree roots for several kilometres, without the silty-sandy soils being washed out by currents and wave action. Moreover, the study of quartz-sand surface textures indicates that the sediments were transported by ice-rafting and subaqueous action. It is unlikely that the abundance of conchoidal fractures and other mechanical forms of quartz sand surface features (chapter 4) in the sediments were created by driftwood or kelp.

Wind gliding across frictionless mud surfaces (Barnes and Scott, 1979) implies subaerially exposed facies, but there is a lack of evidence for this (mud cracks, ripple marks and oxidized deposits). The nature clast-bearing shale is a bioturbated marine sediment. Furthermore, it is unlikely that large boulders could be moved by wind over long distances.

The shallow marine sediment and the spread of erratic boulders over large areas tend to discredit the theory that boulders were transported downslope by slow sediment creep. It is unlikely that the Cretaceous sediment was transported by debris flow of reworked material from the shoreline. Neither debris flow nor downslope sediment creep can explain the boulder distribution on a regional scale (over hundreds of kilometres). The lack of sorting and the low slope angle of the shoreline in the southern Eromanga Basin do not support the theory of either debris flow or downslope movement. In addition, the widespread distribution of these boulders does not relate to the proximity of basement highs. There are no outcrops of fossiliferous Devonian quartzites in South Australia and furthermore, lithologies prominent in the Permian are lacking among clasts (Benbow, 1982).

6.4.2. Evidence of cold climate and ice rafted sediments in the Cretaceous

Study of fossil plants in the early Cretaceous indicate that there was global warmth during the Cretaceous (Fischer, 1982). However, Barron and Washington (1982) used
Cretaceous geography in order to ascertain the climate in the Cretaceous. They show that increased polar ocean temperature did not cause warming of the continental interiors at high latitudes. The greatest variability of temperature gradients would be concentrated in the area of continental margins (ocean water and continental interface). Barron and Washington also predicted winter temperatures of about -3 °C for central Australia. The Arctic ocean winter temperature ranges from -13 to 7°C and Central Siberia can have temperatures as low as -23 °C. This was also confirmed by Schneider et al. (1985). Frakes and Francis (1988) proposed a model to support the possibility of early Cretaceous glaciers both at high latitudes and higher elevations. Such hypothetical late Mesozoic glaciers would lie beyond about 78° paleolatitude. Deposits from ice-rafting occur between 65 and 78° degree palaeolatitude.

Winter ice has been recognised in the high latitudes of the northern hemisphere by many workers. Dalland (1977) reported transportation of winter ice which formed erratic clasts in the lower Tertiary deposits of Svalbard. Pickton (1981) confirmed Dalland in the study of clast-bearing-shale deposits in Spitsbergen during the Cretaceous-Paleogene period. According to Pickton (1981), winter ice is non-selective of grain sizes and shapes, and capable of transporting large volumes of sediment. Based on high latitudes, Pickton also believed there was a greater seasonal temperature variation during the Cretaceous. Moreover, winter ice today occurs at lower latitudes and iceberg form in the Norwegian fiords capable of rafting large amounts of sediments.

The most obvious high latitude early Cretaceous winter-ice deposits in central Siberia have been described by many Soviet researchers (Epshteyn, 1978; Chumakov, 1981). Chumakov noted that Cretaceous sediments in northern Asia consist of silty sandy sediments with scattered pebbles and cobbles. The clasts are usually rounded but may be angular (non-selective characteristic of winter ice deposits). The deposits, often nonbedded, were interpreted as seasonal-marine ice deposits from mountains and rivers. Ice rafting events during the Cretaceous in the southern Hemisphere are not well documented. A number of diamictons and clast bearing shales have been described in New Zealand and Chile, but these may have been the result of debris flow and other density currents.

6.4.3. Ice-rafting in the southern Eromanga Basin

Ice rafted boulders in the Eromanga Basin were first reported by H.Y.L. Brown (1894) when he described water-worn boulders scattered on surface in the central-north of South Australia. Brown proposed that the erratic boulders had been
transported by drift-ice at the time the basin was accumulating sediment. Further work was carried out in 1902 and 1905. Jack (1915) proposed the presence of drift-ice in the Cretaceous when he studied erratic boulders in the area west of the Peake and Denison Ranges. He also noted that the direction of movement of drift-ice was from south to north, as indicated by the decrease from south to north in abundance of erratic boulders. Woolnough and David (1926) argued that erratic boulders caused by drift-ice may have occurred in the Cretaceous.

Woolnough and David also made a detailed study of the problem of Cretaceous glaciation in Central Australia (Moolawatana and Dalhousie areas). They proposed that the Gawler Range highlands probably had been an ice-source area during the middle Cretaceous. The work shows that there was an abundance of water-worn erratic boulders these probably derived from a distance source rocks area. According to Woolnough and David, clasts consist of quartzite, porphyry, siltstone and shale common to the Proterozoic and Cambrian. The clasts are also found in the northern Flinders Ranges area and range in size from 2 feet (0.6 m) to 5 feet (1.5 m). Some boulders bear faint striae and these were interpreted as having been partially erased by aqueous action.

Whitehouse (1928) correlated the Cretaceous marine deposits of Australia and stressed the lines of evidence for the cold climate of the epicontinental sea. These include, a) lack of carbonate fossils of warm climate types: reef coral, Rudistid lamellibranchs, large types of foraminifera and equatorial types of ammonite. The Cretaceous (Roma series) fauna is most like that of Patagonia and North Russia. b) Fresh, angular feldspars occur in the sequence (Aptian). c) Erratic boulders, some of which show faint striae and are up to 5 feet (1.5m) in diameter, penetrating the underlying shales at White Cliffs, New South Wales and the opal-fields of the Stuart Range, Central-North South Australia. d) Glendonites occurring as pseudomorphs after glauberites (chapter 5) in the marine Cretaceous are similar to known glendonites occurring in the Permian glacial rocks of Australia.

David (1932) agreed with Whitehouse about the significance of the fauna correlation, and suggested that glaciation was not intense (based on faint striae of erratic boulders) but possibly the work of shore ice and river ice, like the seasonal ice in northern Canada today.

Supporting the glacial hypothesis, Vnuk (1978), Nicol (1979) and Fennessy (1984) interpreted the erratic boulders in the Stuart Creek Opal Field and Davenport Spring respectively as having been transported by shore ice and under stormy conditions. The reworking of rounded boulders in part included Cretaceous ice rafted materials (Benbow, 1982). Clast measurement in the Stuart Creek Opal Field led to the
conclusion that the Bulldog Shale sediments were transported by an ice rafting mechanism (Nicol, 1979). He also noted that the random orientation of erratic boulders indicates the absence of dominant-current direction or strong unidirectional marine currents.

Frakes and Francis (1988) have made the latest comment about the clast-bearing facies of the Cretaceous, southern Eromanga Basin. They noted that glacial features rarely occur on clasts and that rare clasts truncate underlying mud layers. The clasts consist of rounded quartzite and siltstone appearing in both lestone and boulder beds. The lonestones were interpreted as dropstones transported by shore ice and river ice. The boulder beds were formed by winnowing by storm waves and currents.

In addition, mineralogical and quartz-sand surface textures studies support the hypothesis of ice-rafted sediments in the Cretaceous, Eromanga Basin. The clay mineral assemblage found in the upper part of the Cadna-owie Formation and the Bulldog Shale sediments is dominated by allogenic clays, illite/smectite, chlorite and minor kaolinite. This mineral assemblage can be taken as an indication of ice transport from high latitudes (Bornhold and Giresse, 1985). The occurrence of glendonite in the Cretaceous sediments suggests that they were deposited under polar and arctic condition (chapter 5). Evidence of some mixing of minerals between the Bulldog Shale and the transitional unit of the Cadna-owie Formation suggests reworking possibly by ice rafting or storm-driven currents.

Studies of quartz-sand surface textures indicate that the sediments were transported by ice-rafts (chapter 4). This agrees with the hypothesis that a cold climate exist, at least in the high latitudes or polar regions (H.Y.L. Brown, 1894; Jack, 1915; Woolnough and David, 1926; David, 1932; Vnuk, 1978; Nicol, 1979; Pickton, 1981; Benbow, 1982; Fennessy, 1984; Frakes and Francis, 1988).
CHAPTER 7 CONCLUSIONS

With the widespread occurrence of clast-bearing shales in the southern Eromanga Basin, there is obvious evidence of a regional depositional environment which controls the distribution of the erratic boulders in the area. The works of H.Y.L. Brown (1894); Jack (1915); Woolnough and David, (1926); David (1932); Vnuk (1978); Nicol (1979); Benbow (1982); Fenessy (1984); Frakes and Francis (1988) and this study lead to the conclusion that the Bulldog Shale and the equivalent facies which contain lonestones were deposited by shore ice and river ice, and the conglomerate beds were formed by reworking of the shale containing lonestones and producing lag gravel deposits. To understand the seasonal winter-ice rafting, we must modify the classical glacial model. Included in the characteristics of boulders are:

* Faint striae, uncharacteristic of glacial ice (Woolnough and David, 1926; David, 1932);
* Truncated, indented under layers were considered to be caused by dropstones (Frakes and Francis, 1988);
* Non-sorted grain sizes and shapes characteristic of winter ice (Pickton, 1980);
* Random orientations, typical of dropstones (Nicol, 1979);
* The boulders occur within a quiet environment evidenced by closed shell bivalves, silt and shale sediments, glacial features of quartz sand grains and a high latitude mineral assemblage. These features provide support for an ice transport mechanism.

Based on field observations, mineralogical composition and the results of quartz-sand surface textures, occurrence of the Cretaceous sediments in the southern Eromanga Basin can be summarized as follows:

1) A thin, patchy fluvial-lacustrine deposit of the initial stage of the Late Jurassic-early Cretaceous Algebuckina Sandstone -equivalent to the Village Well Formation which deposited kaolinitic sandstone.

2) Moderate thickness of a transitional sequence (Neocomian-Aptian), the Cadna-owie Formation - Pelican Well Formation which deposited kaolinitic-sideritic fine-grained sandstone at first, followed by marine glauconitic sand and silty shale.

3) A thick sequence of Aptian-Albian marine Bulldog Shale -equivalent to the Marree Formation consisting of normal and bituminous shale.

The Algebuckina Sandstone (clay-poor sands) was deposited by fluvial processes in the late Jurassic, and reworked by the transgressive event of the Neocomian-Albian.(Cadna-owie Formation and the Bulldog Shale).
Figure 6. Schematic diagram of the southern Eromanga Basin margin during deposition of the Cadna-owie Formation and the Bulldog Shale.
Buried sediments underwent mesogenesis, while exposed Cretaceous sediments were under conditions of the telogenetic regime, leading to the formation of a complex and unique mineral assemblage.

In genetic terms, there are two mineral types were clearly identified - detrital and authigenic. Detrital sand grade components are characterized by angular to subrounded grains as a result of transport. These unstable sediments were heavily affected by diagenesis. The most common diagenetic effects completely or partially dissolved grains, some of which were replaced by later authigenic minerals. Identification of the two components (detrital and authigenic) is very importance to the environmental interpretation.

Although smectite-zeolite and illite derived from smectite (chapter 5), and were likely derived from the Proterozoic Gawler Ranges Volcanics, not all of these are geographically with the source region. However, Gottardi and Galli (1985) stressed that at larger distances from the source, only smectite remains in the sediment. The Bulldog Shale and the lower part of the Cadna-owie Formation were deposited during a transgression and as a consequence, nearshore sands fringing the margin became the zone of concentrated coastal erosion of the Aptian-Albian seasonal deposits.

The presence of glauconite-rich deposits may be the result of feldsic volcanic debris, and alteration of smectite/illites from the Precambrian basement which were easily glauconised close to the water/sediment interface of the bioturbated sediment.

Although scanning electron microscopy of quartz-sand surface textures and applied statistical methods provide certain clues to the understanding of Cretaceous sediments, the entire methodology is yet to be perfected. However, on the basis of previous studies and this study of quartz-sand surface textures, it is possible to conclude that glacial features predominated in the sand fractions from the Cadna-owie Formation and the Bulldog Shale. These features are best developed in tillites and their occurrence in shale suggests reworking from tillite, possibly including some of Cretaceous age. The statistical results on glacial features of the sand fraction from the Cretaceous sediments are similar to Quaternary and modern glacial sediments (worldwide). It is known that the process of glacial grinding and crushing produces mechanically broken surfaces on grains. The reworked features can be seen in several samples and the proportion increases in the near shore zone. The presence of aeolian together with glacial grains indicates cold and possibly dry climatic conditions similar to the Northern Slope of Alaska today.

The model proposed in figure 6 is based on field observations, the study of mineralogy and quartz sand surface textures of patchy outcrops and core. Further research is needed, including detailed geological mapping, micro-paleontological study and on the genesis of the sediments to reduce the uncertainty of the stratigraphy in the
area. The comparison of the depositional environment of the Cretaceous Southern Eromanga Basin to high northern and southern latitudes (South America, New Zealand and Antarctica) in terms of fossils, mineralogical studies and quartz-sand surface textures may provide more evidence on the palaeoclimate of the southern hemisphere during the Cretaceous.
Acknowledgements

I am indebted to my supervisors, Professor L.A. Frakes and Dr J.B. Jones of the Department of Geology and Geophysics for their helpful advice and discussion during the course of my research, and for their constructive criticisms, encouragement and interest.

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Other staff member Drs R.A. Both, V.A. Gostin, B. McGowran and R. Oliver also gave advice and encouragement to me in many aspects of my studies and thesis preparation.

I extend my appreciation and thanks to all other technical staff and fellow research students of the Department of Geology and Geophysics; Wayne, Geof, Rick, Evert, John, Christine and others.

I am grateful to Australia which gave me the opportunity to live and study in this country, my brothers and sisters for their support, and finally my wife Phung and daughter Thuan for their helpful companionship, endurance and patience during the preparation of this thesis.
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ADDITIONAL REFERENCE


TABLE OF SAMPLE LOCATIONS AND THEIR STRATIGRAPHIC
POSITION IN THIS STUDY

889.85.04 Sandy-mudstone; 4m above the basal conglomerate layer, Pelican Well section, Pelican Creek, Pelican Well Formation.

889.86.05 Kaolinitic coarse grained sandstone; 1m from the top part of the clast bearing mudstone, Pelican Well section, Trinity Well Sandstone (Plate 1h).

889.85.07 Mudstone; 2m above the base of the Pelican Well section, Pelican Creek, Pelican Well Formation.

889.85.08 Sandy-mudstone; 6m above the basal conglomerate layer, Pelican Well section, Pelican Creek, Pelican Well Formation.

889.86.10 Coarse grained sandstone; at the base of the Trinity Well Sandstone, Pelican Creek.

889.86.11 Kaolinitic mudstone, 2m above the Cambrian basement, Stuart Creek Bed, Stuart Creek Opal Digging, Stuart Creek Opal Field.

889.86.12 Fine grained sandstone; 30m above the Pre-cambrian basement, Trinity Well section, Trinity Well Sandstone, Trinity Well.

889.86.14 Mudstone; 10m above the basal section, Trinity Well section, Trinity Well, Pelican Well Formation.

889.86.22 Sandy-mudstone; 13m above the basal conglomerate layer, Trinity Well Section, Trinity Well, Pelican well Formation.

889.86.25 Pebby-mudstone; 15m above the basal conglomerate layer, Trinity Well Section, Trinity Well, Pelican well Formation.

889.86.29 Mudstone; Matrix of the clast bearing shale; 17m above the basal conglomerate layer, Trinity Well Section, Trinity Well, Pelican well Formation.

889.86.30 Mudstone; Matrix of the clast bearing shale; 16m above the basal conglomerate layer, Trinity Well Section, Trinity Well, Pelican well Formation.

889.DBC.1 Pebby Mudstone; Cadna-owie Formation?, Coober Pedy.

889.CP1/2.LSIS. Sandstone; 0.5m below a limestone, Cadna-owie Formation, Coober Pedy.

889.CP3/1.LIB. Blackshale; 2m above the sandstone bed (Cadna-owie Formation?), Bulldog Shale, Coober Pedy.

889.BLD.B1 Glaucnitic silty shale; Bulldog Shale Coober Pedy.

889.153.65 Mudstone; 153.65m depth, Upper part of the Bulldog Shale, Toodla #1, Oodnadatta.

889.189.1 Bioturbated-mudstone; 189.1m depth, Bulldog Shale, Toodla #1, Oodnadatta.

889.195.85 Mudstone; 195.85m depth, Bulldog Shale, Toodla #1, Oodnadatta.

889.218.4 Shaley-mudstone; 218.4m depth, Bulldog Shale, Toodla #1, Oodnadatta.

889.218.5 Shaley-mudstone; 218.5m depth, Bulldog Shale, Toodla #1, Oodnadatta.

889.224.45 Mudstone; 224.45m depth, Bulldog Shale, Toodla #1, Oodnadatta.

889.234.75 Bioturbated-mudstone; 234.75m depth, lower part of the Bulldog Shale, Toodla #1, Oodnadatta.
889.258.25 Glauconitic-sandstone; 258.25m depth, lower part of the Bulldog Shale, Toodla #1, Oodnadatta.

889.266.5 Bioturbated-sandstone; 266.5m depth, upper part of the Cadna-owie Formation, Toodla #1, Oodnadatta.

889.266.6 Sideritic-sandstone; 266.6m depth, upper part of the Cadna-owie Formation, Toodla #1, Oodnadatta.
APPENDIX 1
MINERALOGICAL DESCRIPTION
Sample. 889.DTW1. At the base of the Pelican Well Formation, Trinity Well.

Hand specimen.

Dark grey calcitic pebbly mudstone, consisting of rough bands of very coarse-angular pebbles in a fine-silty matrix, Detrital grains are poorly sorted and from poorly defined sources.

SEM observation.

Angular-quartz sand grains are the predominant features in this sample. many of the quartz grains are overgrown by silica, including crystal prism faces and silica spheres. On analysis, detrital muscovite shows a variety of compositions and has good cleavage under the secondary electron image. Detrital plagioclase grains are affected by dissolution and hydrolysis, and often overgrown by authigenic K-feldspar. The authigenic K-feldspar is widely developed within calcitic cement and displays euhedral crystals. These feldspar grains contain small calcite inclusions and overlap the calcitic cement. Much of the clay has been replaced by calcite crystals. In part, the cement remains illitic in its morphology. The pebble consist of 60% of siltstone and shale; 30% of quartzite and 10% of porphyry and quartz.

Grain size 20-0.01mm
Composition.
Quartz 30-40
Feldspar: 10-15
Muscovite: 5-10
Calcite: 10-20
Illite: 13-10
Accessories: 2-5
(Iron oxide, illmenite...)

Diagnosis. Cobble-pebbly mudstone with calcitic-silty cement (diamicton).
Sample 889.DBC.1.Cadna-owie Formation?, Coober Pedy
Hand specimen.

Grey-green pebbly mudstone. Pebbles range from rounded to angular, and sizes range from a few mm to 3-4 cm. Diagenesis the predominant feature in the sample which includes silicified and gypsified haloes and fractures. Bioturbation is infilled by glauconitic silty-shale.

SEM observations
Smectite is the major clay mineral in this sample. It consists of Na-Ca members. Detrital illite/smectite shows rounded corners and edges. The clay has chemical composition similar to glauconite grains and possibly originated from the Precambrian Basement. Both cemented clays and calcite crystals are overgrown by authigenic K-feldspar. Pyrite is deeply weathered to iron oxide (possible goethite). Authigenic kaolinite is aggregated among silica overgrowths, with poorly shown image and low degree of crystallization (possible weathering effects). Zeolite is predominated by clinoptilolite and analcite, and shows a deeply weathered surface. Gypsum fibres overgrow detrital and authigenic grain surfaces.

Grain size: 20-0.01mm
Composition.
Siltstone pebbles: 25-30
Quartzite pebbles: 10-15
Quartz: 10-15
Feldspar: 10-12
Allogenic S/I: 10-15
Zeolite: 5-10
Gypsum: 4-5
Accessories: (Kaolinite, muscovite etc.) 2-3

Diagnosis. Pebbly mudstone with a smectitic-zeolitic-glauconitic silicified silty matrix.
Sample 889. BDL.B1 Bulldog Shale Coober Pedy Region
Hand specimen.
Greyish-yellow glauconitic-micaeous siltstone containing vertical and horizontal bioturbated burrows. These are infilled by pyritic shale. Gypsum and halite infill other pore spaces.

SEM observation.
Detrital glauconite grains in an illite/smectite matrix. Grains show rounded edges and corners. Authigenic K-feldspar infills pore spaces. Detrital muscovite appears both parallel and perpendicular to the bedding. Detrital apatite shows rounded edges and corners. Weathered pyrite replaced by iron oxide (goethite). Gypsum fibres infill primary and secondary pore spaces, and are often associated with halite. Silica sphere overgrows the surface of detrital sand grains and aggregates among illitic/smectitic clay.

Detrital grain size 0.5-0.05 mm

Composition.
Quartz: 25-35
Glauconite: 10-15
Feldspar: 15-20
Muscovite: 3-5
Allogenic I/S: 10-15
Gypsum: 5-7
Accessories: 2-5
(apatite, jarosite etc.)

Diagnosis. Marine-glauconitic silty shale, possibly deposited in shallow marine environments, beach zone, shoreline environments.
Hand specimen.
Dark grey-black shale, very fine laminae of clay layers.
SEM observation.

The rock contains massive shale layers, and consists of illite/smectite. Illite shows scallop morphologies and high potassium components (estimated 65% of illitic layer). Detrital muscovite, chlorite, biotite, quartz and organic fragments appear throughout the sample. The euhedral pyrite framboids are well preserved and randomly distributed among illitic/smectitic layer.

Grain size: 0.1mm-μm
Composition:
Allogenic I/S:  75-80
Detrital grains:  15-17
Muscovite, biotite, quartz, organic fragments and pyrite:  2-5.

Diagnosis: Pyritic-illitic/smectitic organic-rich shale. The rock was deposited in an anoxicic environment.
Hand specimen:

Yellowish-green feldsparthic medium sandstone. The fabric is characterized by laminated dark-green layers of glauconite and white layers of quartz and feldspar. These layers are loosely cemented by clays.

SEM observation.

Smectite infills spaces between detrital feldspar, quartz and glauconite, whose size ranges from a few mm to 1cm. Smectite showed cornflake morphology and draping around detrital feldspar and clinoptilolite crystals. Fine-grained clinoptilolite infills secondary pore space of detrital plagioclase. Analcite appears with well developed crystal faces, and have cubic shapes. Analcite often having twin crystals. EDS shows major elements such as Si, Al, Na, K and minor of Mg (which is possibly due to contamination of other clay), with sizes ranging from 10-25 μm. Glauconite grains are rounded, some are enwrapped by gypsum fibres. Dissolution of detrital grains has led to replacement authigenic clays.

Grain size: 1.2-0.01mm

Composition:

Quartz: 25-30
Glauconite: 15-20
Feldspar: 10-15
Smectite: 10-15
Clinoptilolite: 5-10
Gypsum: 2-7
Accessories: 3-5
(Muscovite, halite, apatite etc.).

Diagnosis. Glauconitic-quartzo-feldsparthic medium sandstone. possibly deposited in shallow marine, shoal or beach environments.
889. 86.18 Pelican Well Formation, Pelican Well section.

Hand specimen.

Greyish-green feldspathic fine-grained bioturbated sandstone with white spots of muscovite

SEM observation.

Authigenic S/I is the predominant clay in the sample. It is euhedral in shape and shows a high degree of crystallization. It occurs as infilling in pore space or draped around detrital feldspar grains. Detrital sand grains are often overgrown by authigenic K-feldspar. Most of the detrital illite grains are replaced by silica. Dissolution of plagioclase forms resorbed feldspar. Pyrite frambooids are formed by clusters of small individual octahedral pyrite crystals and often infilled worm tube burrows.

Grain size. 0.8-0.06 mm

Composition:

Feldspar: 20-25
Quartz: 15-20
Glauconite: 5-10
Kaolinite: 10-15
Smectite/illite: 10-15
Pyrite: 2-5
Accessories: 5
(micas, apatite, ilmenite etc.)

Diagnosis. Quartzo-glaucotypic-feldspathic bioturbated greywacke with authigenic kaolinite infilling secondary pore spaces.
889.153.65, Toodla #1, Upperpart of the Bulldog Shale, Oodnadatta.
Hand specimen.
Greyish-green bioturbated glauconitic micaceous sandstone with abundance of organic fragments.
SEM observation.
The rock is dominated by detrital glauconite and quartz grains (size ranging from 0.1-0.6 mm) with minor feldspar, clinoptilolite and illite/smectite components. Glauconite is pitted by the diagenetic process and has illite-like morphology. Detrital quartz is overgrown by silica and authigenic feldspar. Authigenic illite/smectite occurs as pore linings or infills pore space, and has pure end members of the illite/smectite conversion series. Dissolution of detrital plagioclase grains is often infilled by authigenic K-feldspar and zeolite. Pyrite frambooids occur throughout the sample ranging in size from 8-15 μm. Large octahedral pyrite crystals infill worm tubes burrows. The octahedral crystals have sizes up to 100 μm.

Grain size: 0.6-0.05 mm
Composition.
Glauconite: 20-25
Quartz: 15-20
Feldspar: 10-15
Muscovite: 5-10
Smectite/illite: 20-25
Pyrite: 5-10
Accessories: 3-5 (apatite, ilmenite etc.)
Diagnosis. Glauconitic-quartzo-feldsparitic-pyritic siltstone with smectite/illite in the matrix.
889.195.85 Tocdla #1, Bulldog Shale, Oodnadatta.

Hand specimen.

Slump structure in green mudstone. There are cut and fill channels of fine-grained glauconite with good sorting and low degrees of rounding.

SEM observation.

The matrix consists predominantly of authigenic smectite. The smectite consists of both Ca-Na members, and shows honeycomb textures, estimated at about 90% of the expandable layer. Pyrite is aggregated among mixed layer allogenic S/I. The pyrite framboïds are formed by clusters of small individual octahedral crystals of a microsphere (10-25 μm). The small octahedral crystals range in size from 1 to 1.5 μm. Dissolution of plagioclase is often infilled by zeolite (possible clinoptilolite) and K-feldspar. Clinoptilolite and K-feldspar show euhedral crystals which reflect an authigenic origin. They also infill pore spaces or drape around detrital sand grains. Glauconite has composition and morphology similar to detrital illite/smectite grains.

Grain size. 1.0-0.1 mm

Composition.

<table>
<thead>
<tr>
<th>Component</th>
<th>%</th>
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</thead>
<tbody>
<tr>
<td>Glauconite</td>
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</tr>
<tr>
<td>Quartz</td>
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<tr>
<td>Feldspar</td>
<td>10-15</td>
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<td>Smectite/illite</td>
<td>20-25</td>
</tr>
<tr>
<td>Clinoptilolite</td>
<td>2-5</td>
</tr>
<tr>
<td>Pyrite</td>
<td>5-7</td>
</tr>
<tr>
<td>Accessories</td>
<td>3-5</td>
</tr>
</tbody>
</table>

(muscovite, apatite etc.)

Diagnosis. Glauconitic-quartzo-feldspathic medium sandstone with smectite dominant in the matrix.
889.234.75. Toodla #1, lower part of the Bulldog Shale, Oodnadatta.

Hand specimen.

Medium to coarse-grained glauconitic sandstone. The glauconite formed laminated bands with heavy bioturbation which was infilled by white grey feldspathic sand grains.

SEM observation.

The rock consists of well-sorted glauconitic-feldspathic medium sandstone, and is cemented by authigenic smectite. Morphology of the smectite was estimated at about 80% of the expandable layer. The smectite consists of Ca-Na end members. Pyrite appears as euhedral framboids or octahedral crystals and infills burrows. Dissolution of plagioclase is infilled by authigenic K-feldspars. Mordenite appears in pore space or aggregates among smectitic/illitic layers. Glauconite shows surface textures of detrital illite and is pitted by diagenetic processes, in sizes ranging from 50 μm to 0.8mm. Detrital apatite and muscovite show very little rounding.

Grain size. 1.0-0.05 mm

Composition:
- Glauconite: 25-30
- Quartz: 10-15
- Smectite/illite: 25-30
- Zeolite: 5-10
- Accessories: 2-5
  (muscovite, apatite etc.).

889.258.25. Toodla #1, lower part of the Bulldog Shale, Oodnadatta.

Hand specimen.

Grey-green medium sandstone in bioturbation with micaceous and white feldspathic layers.

SEM observation.

Glaucobic-feldspathic grains are predominant in the detrital components. Dissolution of detrital plagioclase is infilled by authigenic potassium feldspar and clinoptilolite. Most of the secondary quartz develops crystal prism faces. Smectite with cornflake morphology infills pore space of the rock. Pyrite framboid infills burrows, sizes range from a few μm to 15 μm. The pyrite consists of a large number of small octahedral crystals.

Grain size: 1.5-0.2 mm

Composition:
Quartz: 15-20
Glaucinite: 15-20
Smectite/illite: 35-45
Feldspar: 10-15
Accessories: 4-7
(Muscovite, clinoptilolite etc.)

889.266.5. Toodla #1, upper part of the Cadna-owie Formation, Oodnadatta.

Hand specimen.

Light to green grey intense bioturbated medium sandstone. Patches of altered-feldspar grains with micro-micaceous silty matrix.

SEM observation.

The rock consists of rounded and subrounded quartz and detrital feldspar. Many of the detrital grains are partly overgrown by authigenic feldspar. Partial dissolution of detrital plagioclase grains, infilled by well crystallised authigenic feldspar and clinoptilolite. Euhedral mordenite aggregates among authigenic kaolinite. Pyrite is less common than in sample 889.258. Authigenic quartz overgrows on detrital grain surfaces. Na-Ca-smectite infills pore spaces, and are estimated at 90% of smectite layer. Detrital muscovite shows random orientation and slight abrasion occurs at corners and edges. Rounded glauconite has the composition of illite, and appears to be unoriented and isolated throughout the sample.

Grain size: 1.2 - 0.04 mm
Composition:
- Glauconite: 7-10
- Feldspar: 10-15
- Quartz: 20-30
- Smectite: 20-25
- Kaolinite: 5-10
- Muscovite: 2-5
- Accessories: 2-5
  (apatite, organic fragments etc.)

Diagnosis: Bioturbated feldspathic-glauconitic medium-grained sandstone. Authigenic smectite, zeolite and authigenic K-feldspar infilled pore space and matrix. Possibly deposited in shallow marine or particularly lagoonal environment.
889.266.6. Toodla #1, upper part of the Cadna-owie Formation, Oodnadatta.

Hand specimen.

Brownish-yellow, medium-grained sandstone with a calcareous cement. Laminated dark mud layers contain feldspathic and white kaolinite dots which appear in patches throughout the sample.

SEM observation.

Subrounded, medium-grained detrital feldspar and quartz grains are covered by thin authigenic siderite crystals. Some siderites form rose shapes and euhedral aggregated crystals which are similar to microglendonite. Detrital muscovite shows angular edges and corners. Authigenic feldspar overgrows surface of the detrital sand grains and infills pore space. Authigenic quartz has well developed crystal prism faces, and appears among authigenic kaolinite. The kaolinite shows a very high degree of crystallization. Dissolution of plagioclase and ferromagnesian minerals form a significant secondary pore space. Small numbers of pyrite frambooids aggregate among smectite/illite layers (dark mud layers) and are often associated with clinoptilolite.

Grain size: 1.2-0.3 mm

Composition:

- Quartz: 30-40
- Feldspar: 10-15
- Siderite: 10-15
- Kaolinite: 15-20
- Smectite: 10-15
- Accessories: 5
  (Muscovite, pyrite, clinoptilolite etc.).

Diagnosis. Sideritic-kaolinitic medium grained sandstone with kaolinite and smectitic/illitic clay. Possibly deposited in marginal marine environment (estuary).
APPENDIX 2
CHEMICAL ANALYSES
QUANTITATIVE ELECTRON MICROANALYSIS:

Polished thin sections and polished blocks were coated with carbon in a standard vacuum evaporator. The composition of selected grain minerals was determined by using the JEOL 733 Superprobe model electron microprobe analyser, a KEVEX 7000 series energy dispersive system attached to a JEOL 733 analyser. All determinations use the standard oxide and silicate analysis (Griffin, 1979) based on 15 elements as follow: Na, Mg, Al, Si, P, S, Cl, Fe, K, Ca, Ti, V, Cr, Mn and Ni. The analytical conditions were: voltage was 15 kV; the beam current was 5 nA, beam size 3-5 μm in diameter, counting was fixed at 60 seconds. The drift factor was very small, and for in each grain several analyses were taken.
### Authigenic feldspar from Trinity Well Sandstone PW3

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**Structural formulae (based on 8 oxygen per formula unit)**

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### Authigenic feldspar from Trinity Well Sandstone Pelican Well section

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**Structural formulae (based on 8 oxygen per formula unit)**

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### Authigenic feldspar from Trinity Well Sandstone,

#### Trinity Well section

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Structural formulae (based on 8 oxygen per formula unit)

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Detrital feldspar from Cadna-Owie Formation, Pelican Well section

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Structural formulae (based on 8 oxygen per formula unit)

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Ca: Na: K = 1.7: 98.3: 0

Structural formulae (based on 8 oxygen per formula unit)

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**Structural formulae (based on 24 oxygen per formula unit)**

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**Structural formulae (based on 24 oxygen per formula unit)**

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Structural formulae (based on 18 oxygen per formula unit)

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Structural formulae (based on 36 oxygen fer formula unit)

- P: 0.03
- Si: 6.93
- Al: 7.12
- Fe: 5.06
- Mn: 0.07
- Mg: 5.16
- Ca: 0.03
- K: 0.07
- Na: 0.23
### Glauconite at depth 258.5 m, Bulldog Shale Toodla #1, Oodnadatta

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**Structural formulae (based on 18 oxygen per formula unit)**

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Illite analyses from the Bulldog Shale and the Cadna-owie Formation,
Southern Eromanga Basin, South Australia.

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Structural formulae (based on 24 oxygen)

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1: 889.232.2
2: 889.86.12
3: 889.199.5
4: 889.205.4
5: Deer et al. (1966)
Biotite from Bulldog Shale, Toodla #1, Oodnadatta.

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Structural formulae (based on 24 oxygen)

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