PETROLOGY AND PROVENANCE OF PERMIAN GLACIOGENIC SEDIMENTS OF SOUTHERN AUSTRALIA

By

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To my dearest parent and my wife Parvin

STATEMENT OF ORIGINALITY

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

A. Granmayeh

ABSTRACT

The Permian glaciogenic sediments of southern Australia well exposed at Hallett Cove, Cape Jervis, Kings Point and Kangaroo Island, consist of a variety of sands, clays, till and many erratics. Samples from the Permian sands, erratics, and probable source rocks from East Antarctica, Encounter Bay and Kanmantoo Group were studied by petrographic microscope, and were also analysed by XRF and microprobe so as to relate the Permian sands and erratics to their source rocks and thus interpret their provenance.

Glaciated bedrock indicates that the direction of ice movement was from the S-S E. Surface texture of sand size quartz and garnet were studied by S.E.M. to interpret the depositional history of the sediments. This showed that transportation of the Permian glaciogenic sediments was due to three phases of solid (ice), liquid (water) and gaseous (wind). The garnet grains in the Permian sands have chattermark trails on their surfaces, indicating that they have experienced a very long distance of transport by ice. The surface of quartz grains contain parallel, subparallel steplike, arc-shaped step and conchoidal breakage faces that are caused by glacial grinding or mechanical action. The grain morphology of the Permian sands indicate that well rounded aeolian grains are present together with angular and subangular glacially transported grains.

The rock samples of East Antarctica, mainly from Commonwealth Bay and Windmill Island, consists of mostly metamorphic rocks ranging from medium to high grade. The rock samples from Kanmantoo Group are mainly metasediments, and rock samples from Encounter Bay are mostly composed of granitic rocks

The provenance of the Permian glaciogenic sediments of southern Australia have been interpreted from both the petrology of sediments and the petrology/mineralogy of the probable source rocks. Detailed physical and chemical analyses of heavy and light minerals were undertaken to ascertain their individual character for the source of heavy and light minerals in the Permian glaciogenic sediments. At the same time the rock

samples from East Antarctica, Encounter Bay and Kanmantoo Group have been studied in the same manner to compare with the Permian glaciogenic sediments.

The characteristics of quartz grains, such as elongation, number of crystal units in polycrystalline quartz grains, trace elements and CL colours in the quartz grains of Permian sands, and from rock samples in different localities such as East Antarctica, Kanmantoo Group and Encounter Bay Granite, have also been studied. This comparison showed that the quartz grains in metamorphic rocks have very different characteristics from the quartz grains of plutonic rocks, and importantly they indicate that about 60% of the detrital quartz grains of Permian sands have been derived from plutonic rock samples.

Composition of detrital feldspar grains in the Permian sands also have been studied by microprobe analysis and were compared with the feldspar grains of different rock samples. This indicates that most of the calcic feldspars were derived from volcanic rocks, and the more common potassic feldspars came from metamorphic and plutonic rocks. Therefore, it suggests that more than 97% of the detrital feldspar grains in the Permian sands were derived from gneisses or granitic rocks.

Heavy minerals are amongst the most sensitive indicators of the nature of source areas, they have frequently been used in studies of provenance of the sediments. Compositional analysis of minerals or mineral groups through the use of the electron microprobe is presently receiving much attention. Studies of the heavy minerals in the Permian sands using a microprobe showed that they are dominated by garnets, which constitute 73% of the total assemblage, and the rest of the minerals are rutile, sillimanite, zircon, staurolite, ilmenite, tourmaline, epidote and trace of sphene. Garnets that are rich in Fe and Mg were recorded from gneisses, and those rich in Mn and Fe were recorded from phyllite rock samples of East Antarctica. However, the detrital garnet grains in the Permian sands shows heterogeneity in composition and suggest different source rocks but, mostly from high grade metamorphic rocks such as gneisses.

The Hf contents of detrital zircon grains in the Permian sands were compared with different rock samples, this shows a polymodal distribution in the Permian sands indicating multiple sources. The optical properties such as colour of detrital tourmaline

grains of Permian sands suggests mostly granitic source rocks. Chemical composition of detrital epidote grains in the Permian sands shows that they have been mostly derived from metamorphic sources. Composition of detrital staurolite in these sands also suggests local source, such as the Kanmantoo Group metasediments of southern Australia.

The chemical composition of ilmenites shows that the ilmenite grains in the metamorphic rocks are richer in TiO2 than those from igneous rocks. MgO less than 0.4wt% and MnO less than 5wt% have been recorded from metamorphic rocks, those with more have been found in plutonic rocks. The chemical composition of detrital ilmenite grains of the Permian sands suggests that more than 95% have been derived from metamorphic source rocks. Rutile grains have not been recorded from granite, volcanic and low grade metamorphic rocks, but are common in high grade metamorphic rocks. The presence of detrital rutile in the Permian sands suggests high grade metamorphic source rocks. The chemical composition of detrital sillimanite grains in the Permian sands shows mostly a viridian composition, that is very similar to the sillimanite grains from gneissic rock samples of East Antarctica.

Trace amounts of sphene and pyroxene have been found in the Permian sands. This suggests that small amounts of volcanic rocks have contributed to the Permian sands. Trace elements in the porphyritic rock samples of Mt Monster in southeast South Australia were compared with the Permian porphyritic erratics, but big differences between the trace elements of these two rocks indicate that the Permian porphyritic erratics were not derived from this area.

This study concludes that the Permian glaciogenic sediments of southern Australia have been derived mostly from high grade metamorphic rocks that are very similar to the rock types of East Antarctica. This implies that part of East Antarctica was uplifted as an ice dome, and debris transported towards southern Australia has supplied most of the minerals to the Permian glaciosediments of southern Australia.

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CHAPTER ONE

INTRODUCTION

The term provenance is defined as the place of origin. The specific problem in question here, is the nature of source rocks that could supply sufficient quantities of detritus representing Permian glaciogenic sediments. In sedimentary petrology, provenance refers to the nature of the source rocks in a particular area and the identification of these source rocks through the study of sedimentary detritus derived from them. Tectonic events produce highlands and provide source rocks that shed sediments into the surrounding basins. Although these highlands serve as the primary control on the eventual sandstone composition, the relief, climate, transport mechanisms, depositional environment and diagenesis can all be important secondary factors (Dickinson et al. 1983).

Because the goal of all provenance studies is to determine the source of the clastic sediments, and to decipher earth history through palaeogeographic reconstructions of the earth crust, it is important not only to understand the above processes and their effects, but also to understand the influence of source rocks on the detrital grains themselves.

Most recent work concerning tectonics and sandstone mineralogy concentrates on thin section petrology of the light minerals. Ternary diagrams are produced using quartz, felspar and rock fragments. Detrital percentages calculated from point counts are plotted on these diagrams to indicate specific tectonic settings (Dickinson and Suczek 1979; Dickinson 1985; Zuffa 1985). Although these studies have led to a greater understanding of the relationship between provenance and sandstone composition, they have failed to consider the importance of heavy minerals.

Heavy minerals, the minor, high density accessory minerals of sands, have traditionally been used for provenance determination (Boswell 1933; Krumbein and Pettijohn 1938; Stow 1939; Milner 1962), and as stated by Stategger (1987) the study of these minerals allows for an adequate interpretation of plate tectonic setting where analysis of light mineral based petrology has failed (Mack 1984).

However, because heavy minerals are sensitive to the processes of weathering, transportation, deposition. and intrastratal solution, the heavy mineral suite found in a sandstone does not necessarily reflect the source area mineralogy. Although the effects of weathering in the source area and during transport may be minimal (Morton 1985a), 34hydraulic sorting and intrastratal solution can severely restrict their usefulness (Pettijohn 1941; Rittenhouse 1943; Morton 1985a).

The most effective method to counteract the problems of intrastratal solution and hydraulic sorting is to examine the relative abundance of the physical and chemical varieties of one mineral.

Classical methods of varietal studies have involved optical differentiation based on colour, shape, or other physical properties of grains. Minerals especially scrutinised in this manner have included tourmaline (Krynine 1946) and zircon (Poldervaart 1955; Vitanage 1957). With the advent of the electron microprobe it is now possible to make further advances in these studies by determining quickly and accurately the compositional variation of a particular mineral species. This method has been particularly successful in the study of garnets (Morton 1985b), although this approach may also be applied to a variety of other detrital minerals including pyroxenes, amphiboles, epidotes, tourmalines and detrital opaque minerals such as ilmenite and magnetite.

In order to use the composition of a mineral as a provenance indicator the mineral should satisfy the following requirements:

1- be common in a variety of parent rocks

2- be variable in composition

3- be relatively stable

Tourmalines, because of their ultrastability should offer the best opportunity in provenance determination, but because of the presence of boron in the structure, which is not detectable with an energy dispersive system, accurate compositional analysis is difficult to achieve (Morton 1985a).

Ilmenite and magnetite appear to be ideal minerals for the determination of provenance. This is because both occur in a wide variety of igneous rocks, both extrusive and intrusive, as well as in pegmatites, other vein rocks and metamorphic rocks

(Ramdohr 1980). Both show compositional variations based on source rock paragenesis (Hutton 1950; Buddington and Lindsley 1964), both are relatively abundant in sands and sandstones and are relatively stable (Pettijohn 1941), both are easily separated using magnetic techniques (Rosenblum 1958; Force 1976; Lumpkin and Zaikowski 1980). Previous studies of detrital ilmenite (Darby 1984; Darby and Tsang 1987) and magnetite (Luepke 1980) have suggested that variation in their composition is sufficient to provide an unmistakable signature to determine sediment source.

Thus one of the aims of this thesis is to test the utility of detrital heavy and light minerals as provenance indicators in glaciogenic sediments.

Location of the study areas

The principal area of study for this thesis is in South Australia, specifically in the Hallett Cove area, Fleurieu Peninsula (Cape Jervis, Kings Point, Inman Valley), and on Kangaroo Island. These areas form part of widespread Permian sands in southern Australia and are now deeply dissected by rivers and streams.

In order to compare the chemical composition of some heavy and light minerals in the Permian sands with possible source rocks some samples were selected from eastern Antarctica from Windmill Island to Commonwealth Bay, and from South Australian areas: the local Kanmantoo Group of the Fleurieu Peninsula and Encounter Bay.

Previous investigations

One of the earliest references to the geology of the Permian sands in South Australia was made by Selwyn (1859), who described glacial sediments, and mentioned the occurrence of striated pavements near Adelaide. Late Palaeozoic sediments have long been known from the Adelaide region, at Hallett Cove, Cape Jervis and elsewhere on Fleurieu and Yorke peninsulas as a result of great interest in the associated glaciation (Howchin 1895, 1926). In Australia this glaciation is now known to have covered a large part of the continent (Crowell and Frakes 1971; BMR Paleogeographic Atlas; Veevers 1990).

The only detailed investigation of the area was made recently by Bourman and Alley (1990). They found the direction of ice movement of the Late Palaeozoic ice was towards the north and northwest, thus indicating a southern and southeastern provenance.

Geographic distribution

The basal Permian sections of many areas in all states of Australia contain the diagnostic glacial assemblage of pavement, tillite with striated stones, and dropstone-bearing shale. David (1950) briefly described many of these materials and referred to publications then available. Subsequent reports include Clarke and others (1951), Kenley (1952), Banks and others (1955), Glaessner and Parkin (1958), McWhae and others (1958), Playford (1959), Veevers and Wells (1961), Banks (1962), Banks and Ahmad (1962). Wopfner (1970) suggests that in South Australia Permian glaciers developed on uplifted highlands as ice cap glaciers.

The tillites and other subordinate glacial deposits are thin and discontinuous across the low-relief stable platform of southern and central Australia, but thicken and interfinger with, and are overlain by marine sediments around the western, northwestern, and eastern margin of the continent. The general direction of ice flow was northwestward in southern Australia.

The environments of deposition of Permian sands vary between solid and wet environments. The erratics of different sizes in these sediments were deposited under a rigorous and over changings glacial regime. Sediments are characteristically unsorted and unbedded, whereas some contain dropstones that indicate the passage of abundant icebergs. The sand units in these sediments have been transported in fluvioglacial and fluvial environments of high energy. This has also affected the heavy mineral assemblages, as will be shown later. Polycrystalline quartz is relatively unstable and as such would not survive long in such a sedimentary mill. Hence the decrease in this proportion supports the idea of a great distance of transport and a vigorous environment of deposition. In contrast, the clay size sediments have been deposited in lakes with very low energy and in paralic water environments.

In summary, the Permian sands in southern Australia are interpreted as a sequence of glaciogenic deposits on the evidence of the abundant variety, and size of erratics, and on the surface texture of quartz and garnet surfaces.

Setting of southern Australia within Gondwanaland

Du Toite was probably the first to publish a map of the "Great ice-cap" overlying Gondwana during the Late Carboniferous. Figure.1.0 is a reconstruction of Gondwanaland modified from du Toite (1937) to account for new data from Antarctica and other regions, and provide a high minimum latitude for glacial deposit.

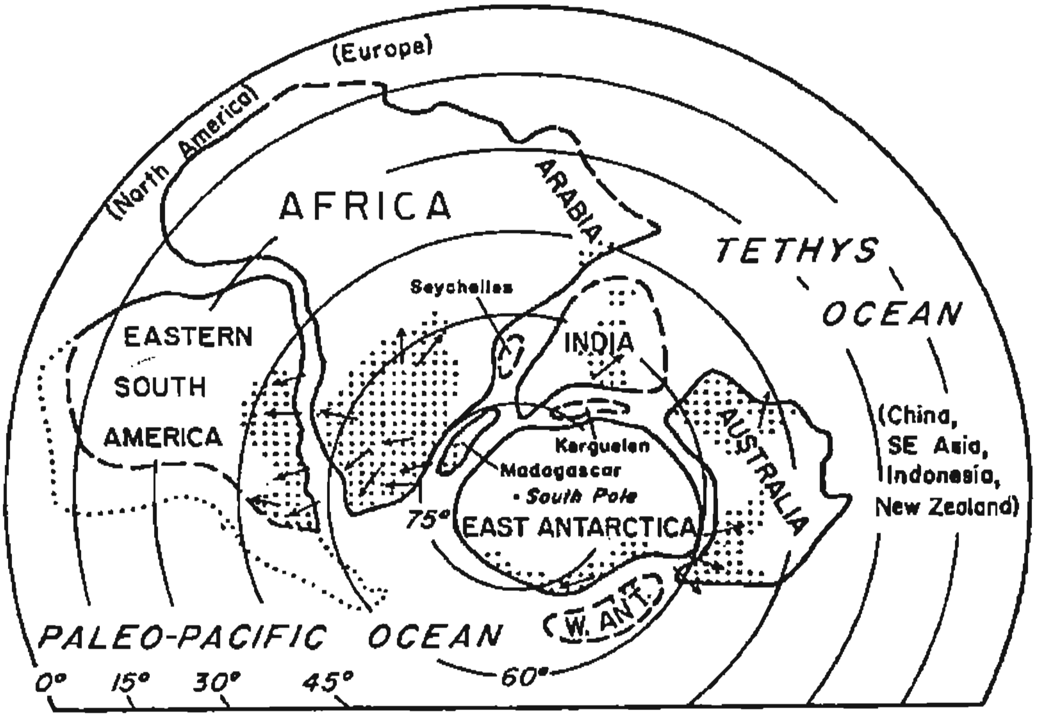


Fig 1.0. Reconstruction of Gondwanaland at the beginning of Permian time, showing paleolatitude~ stipple marks the regions of known tillite, and arrows show directions of ice flow determined from glacial pavements. Solid lines are margins of present continental masses changed little in internal shape since Permian time. Dashed lines show margins of continental masses much deformed since the Permian (South America, Africa, India) and possible Permian positions of continental blocks now very different in shape (Seychelles and Kerguelen ridges, West Antarctica). Outer margin of present South America is shown for reference by dots, but entire Andean belt may have been at low latitude in Early Permian (after du Toite 1937).

Regions glaciated during the Late Carboniferous were then parts of the large continent of Gondwanaland and lay at middle and high southern latitudes. Various positions for Australia have been proposed by Lawver and Scotese (1987), Dewitt et al.(1988), Veevers (1990). One of the latest developments is that by Tewari and Veevers (1993), (Fig. 1.2).

A linkage between Antarctica and Australia in the former Gondwana supercontinent has been established in general beyond any doubt, but there are considerable differences of opinion as to matching details. The greater part of the segment between longitudes 145 E° and 153 E° comprises the Wilkes Basin, identified from geophysics, interpreted as having developed due to east west tension between cratonic rocks to the west and the younger metasedimentary groups now seen in the Transantarctic Mountains (Steed 1983). The onset of this extensional regime was suggested to have commenced during the Palaeozoic (Steed and Drewry 1982) but when exactly rifting took place is not apparent.

The broad scale structure of northern Victoria Land is strikingly similar to that of southeastern Australia. The conformity in postulated crystallisation age of the megacrystic granitoids in southern Eyre Peninsula at Commonwealth Bay and Adelie Land in Antarctica (Fig. 1.3) is considered to be evidence in support the former juxtaposition of these terrains.

The late Proterozoic geologic record along the palaeo-Pacific margin of Australia and Antarctica further indicates more or less uniform geologic evolution, manifested by sedimentation of the Adelaidean, the Berry Group, and Priestley Group. The enigmatic fault-bounded Cape Hunter phyllites at Commonwealth Bay were discussed by Grew (1982) as possible Adelaidean equivalents (see also, Stüwe and Oliver 1989), though this is, by no means certain.

Aims of this study

The basic aim of this study is to determine the provenance of the Permian sands of southern Australia and thereby to better reconstruct the palaeogeography of southern Australia and Antarctica. More specifically:

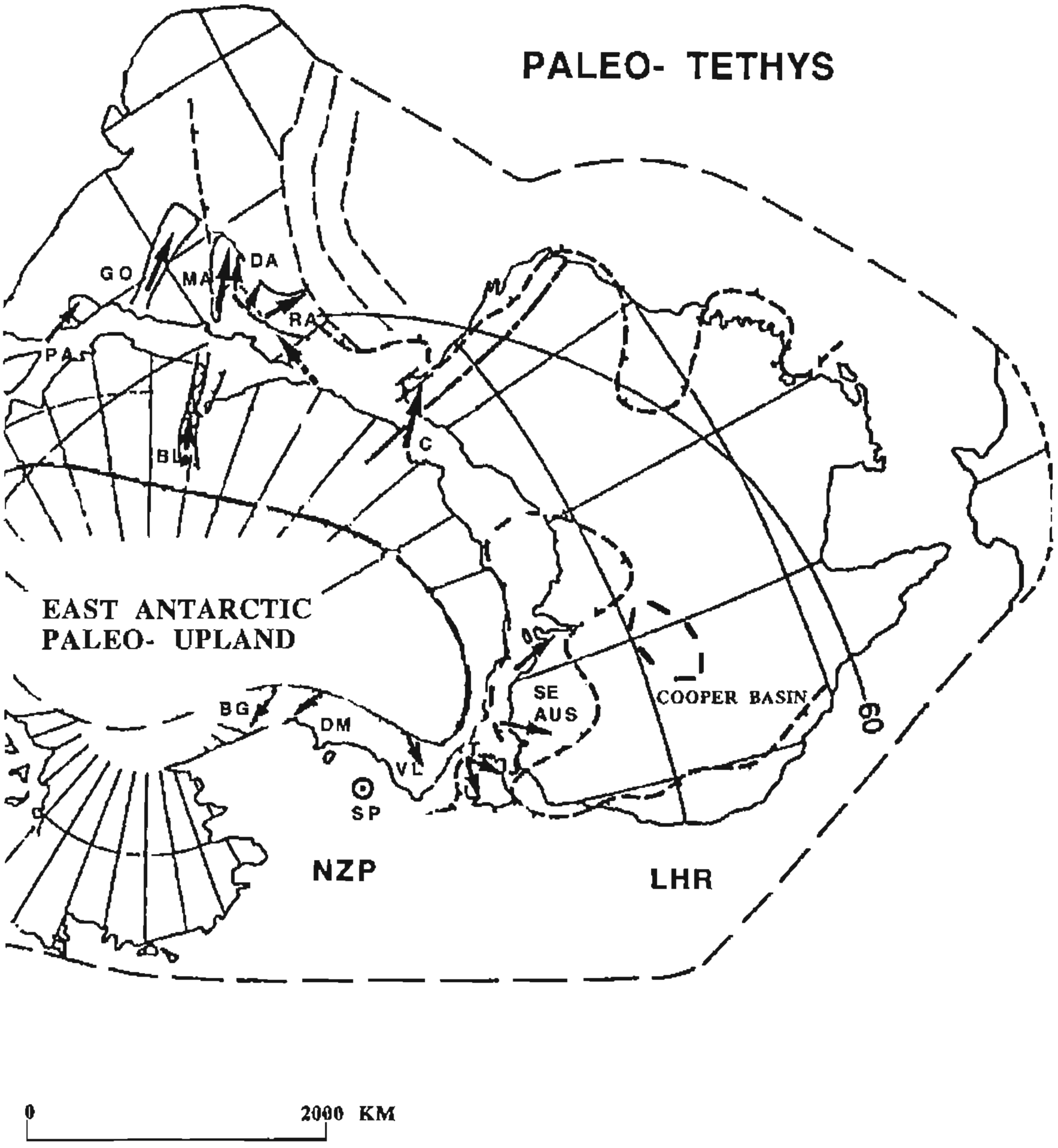


Fig. 1.2. Reconstruction and Paleo-latitude of Early Permian eastern Gondwanaland between Paleo-Tethys and Panthalassa (after Te wari and Veevers 1993). Permian and Triassic fluvial azimuth (full arrow). NZP: New Zealand Plateau, LHR: Lord Howe Rise, SP: South Paleo-Pole, BG: Beardmore Glacier, DM: Darwin Mountains, VL: Victoria Land, BL: Beaver Lake, C: Collie, PA: Palar southern India, GO: Godavari, MA: Mahanadi, DA: Damodar, RA: Rajmahal. Dashed line marks the Permo-Carboniferous basins: See details on Fig. 2.

1- To determine the provenance of the Permian glaciogenic sediments of southern Australia, in as great detail as possible using both heavy and light minerals.

2- To correlate the provenance of the Permian sands with the known geology and tectonics of East Antarctica.

3- To determine those features of light and heavy minerals that are most useful as provenance indicators.

The chemical composition of some heavy minerals of the eastern part of the Antarctic craton can be compared with the same heavy minerals of southern Australia to verify the reconstruction of Antarctica and Australia (Chapter 7). Ideas concerning the tectonic development of East Antarctica (Flöttman and Oliver in press) will be tested. The idea of a "Wilkes" domal uplift and the subsequent rift basin separating the Early Proterozoic crystalline bedrock on the northwestern side (opposite South Australia) from the mixed Late Proterozoic (Adelaidean) metasediments on the southeastern side can be tested by means of the light and heavy mineral suites that derived from either the Early Proterozoic crystalline basement, or from the Late Proterozoic (Adelaidean) metasediments (see Chapter 4). Bimodal acid and basic volcanics may have erupted during the rifting of the Wilkes basin. However, no basalts have yet been found in the Permian erratics. The volcanics found include quartz-feldspar porphyries, rhyolites, andesites, and welded tuffs, but these volcanics could also be of older Devonian to Silurian age. The glacial transport directions and the orientation of deep glacial valleys (eg. Inman Valley) indicate a radial transport outwards from the "Wilkes" domal uplift and rift basin edges (ie. northwestward in South Australia, eastward, north eastward, and northward in Victoria, and eastward in Tasmania (see Chapter 2). Current literature presents the Wilkes Basin as essentially Tertiary in age. However most Tertiary basins in southern Australia have underlying Permian infrabasins, and the Wilkes Basin need be no exception.

The concept of a small ice cap centred to the west of modern Tasmania, southwest of Victoria, and southeast of Fleurieu Peninsula was expressed many years ago (Wopfner 1974). In simple terms this would most likely result in common assemblage of erratics and common types of light and heavy minerals. As the assemblages are however

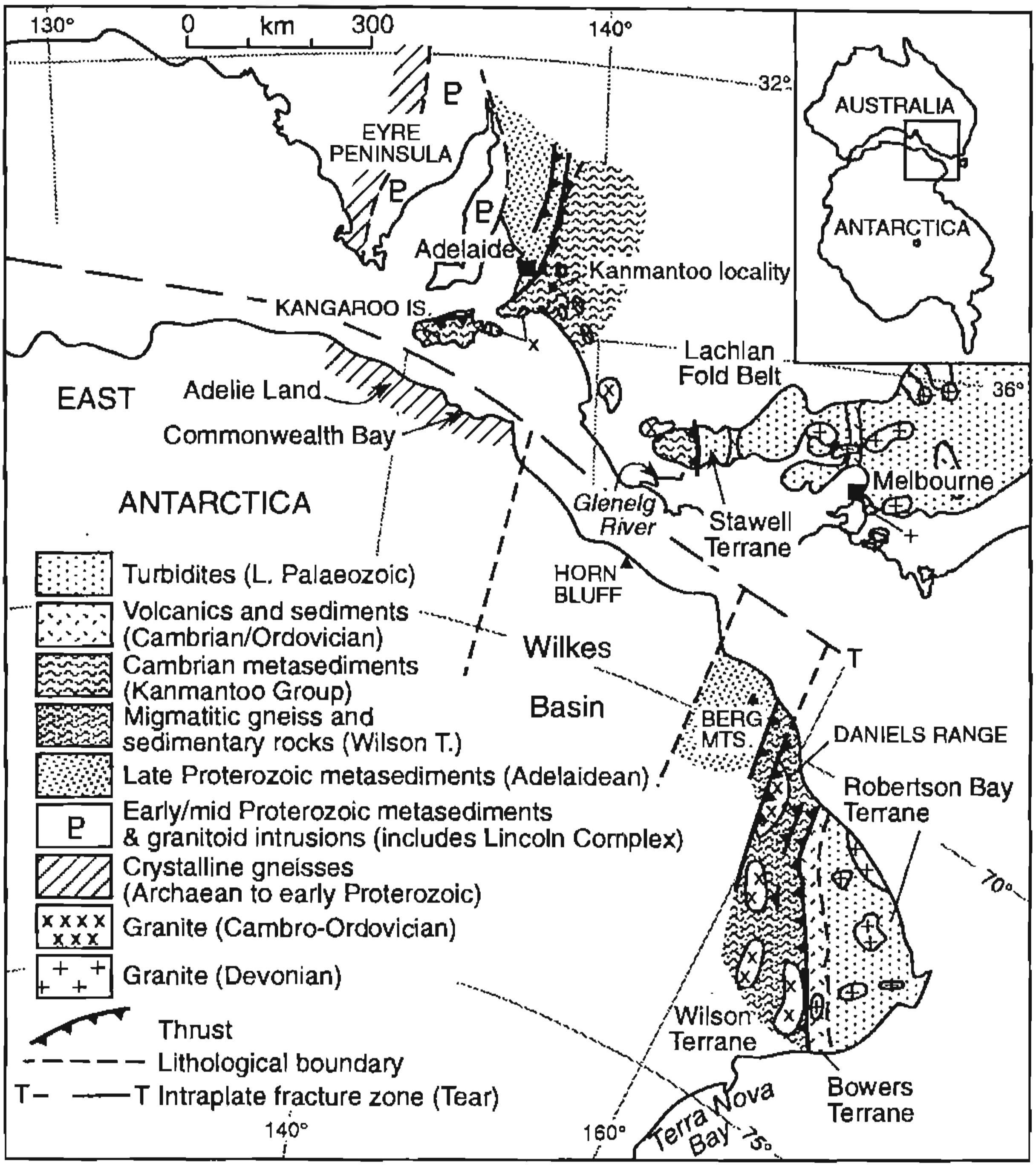


Fig. 1.3. Tectonic Map of Southern Australia and Eastern Antarctica (after Flöttmann & Oliver in press)

distinct, a new model of glacial provenance will be presented; one based on the contrasting geology of the rifted uplifted shoulders of Wilkes Basin in East Antarctica.

Method of study

The Permian sediments in the Hallett Cove area are divided into three units and at Cape Jervis into five members (see Chapter 2). These sediments rest on the Precambrian sediments at Hallett Cove, and on Cambrian Kanmantoo Group metasediments at Cape Jervis, Inman Valley, Kings Point and Encounter Bay. Different lithologies have been assigned to a variety of environments ranging from glacial to fluvioglacial in the Early Permian with rare paralic marine sediments. Therefore it was planned that comparison of heavy and light minerals between the Permian sands and those samples from East Antarctica, Kanmantoo Group and Encounter Bay would be made.

Samples of Permian sands have been collected from several localities from widely spaced outcrops (Fig. 1.4). The Antarctic, Encounter Bay and Kanmantoo Group samples came from the collections of R. L. Oliver, A. R. Milnes, and from the Tate Museum of the Department of Geology and Geophysics University of Adelaide. The samples from Antarctica range from metamorphic acid and basic gneiss, charnockite, porphyritic granite, pegmatite and migmatite and the localities of samples is shown in (Fig. 1.5). These rocks were collected widely over the eastern Antarctica including the Windmill Island, and their ages range from 1477±73 to 1465±34 m.a (Williams et al. 1983) and Commonwealth Bay, whose ages range from 1740 to 2300 m.y (Oliver and Fanning in preparation). The samples from Kanmantoo Group are mostly metasediment in composition and their age is between 523 to 531 m.y (Milnes 1973). The samples from Encounter Bay are mostly granitic in composition and an Early Paleozoic age for many of these granites has been established in recent years as the result of radiometric age dating techniques (Foden et al. 1990b).

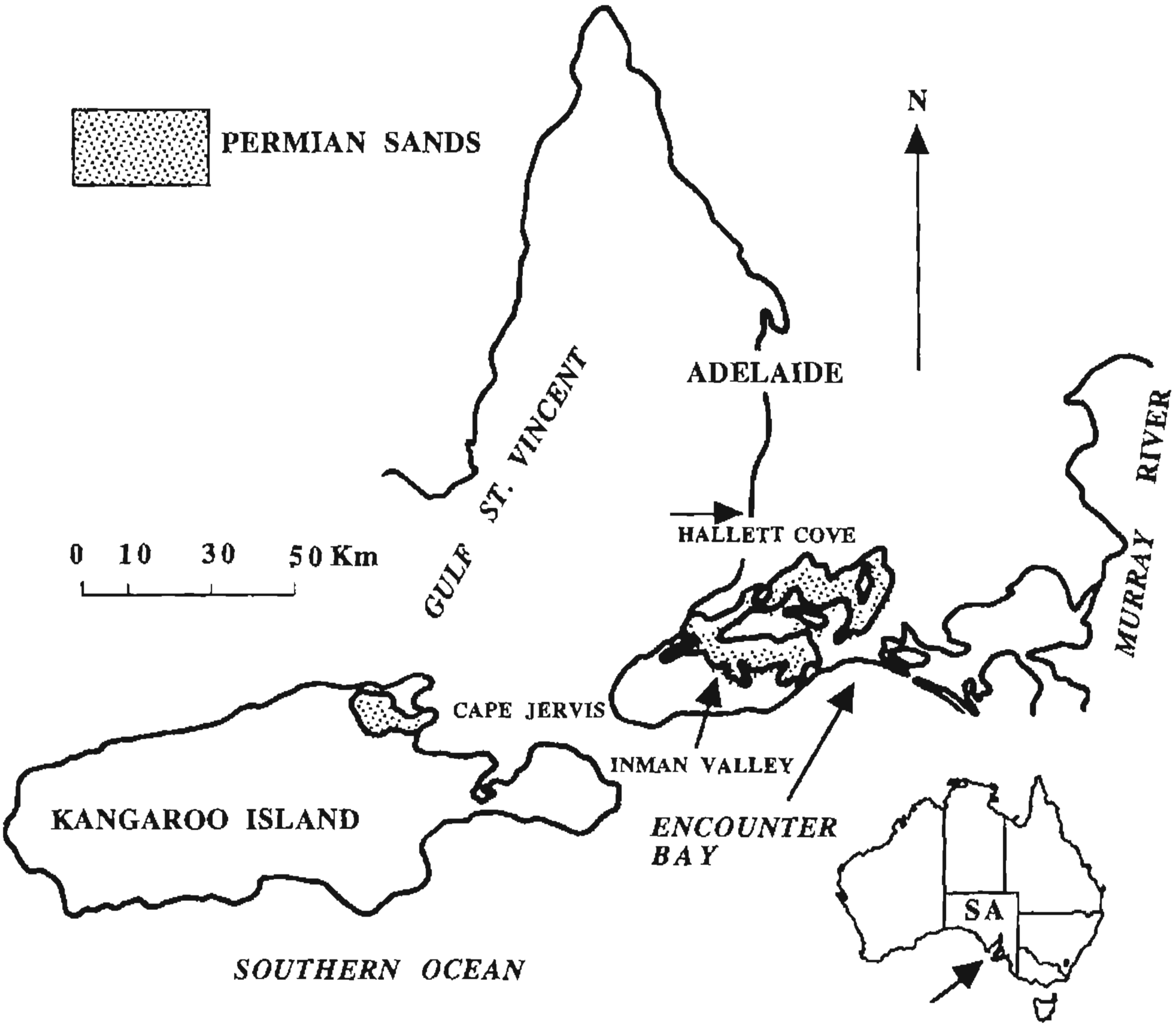


Fig. 1.4 Localities of sample collection from Permian sands of southern Australia.

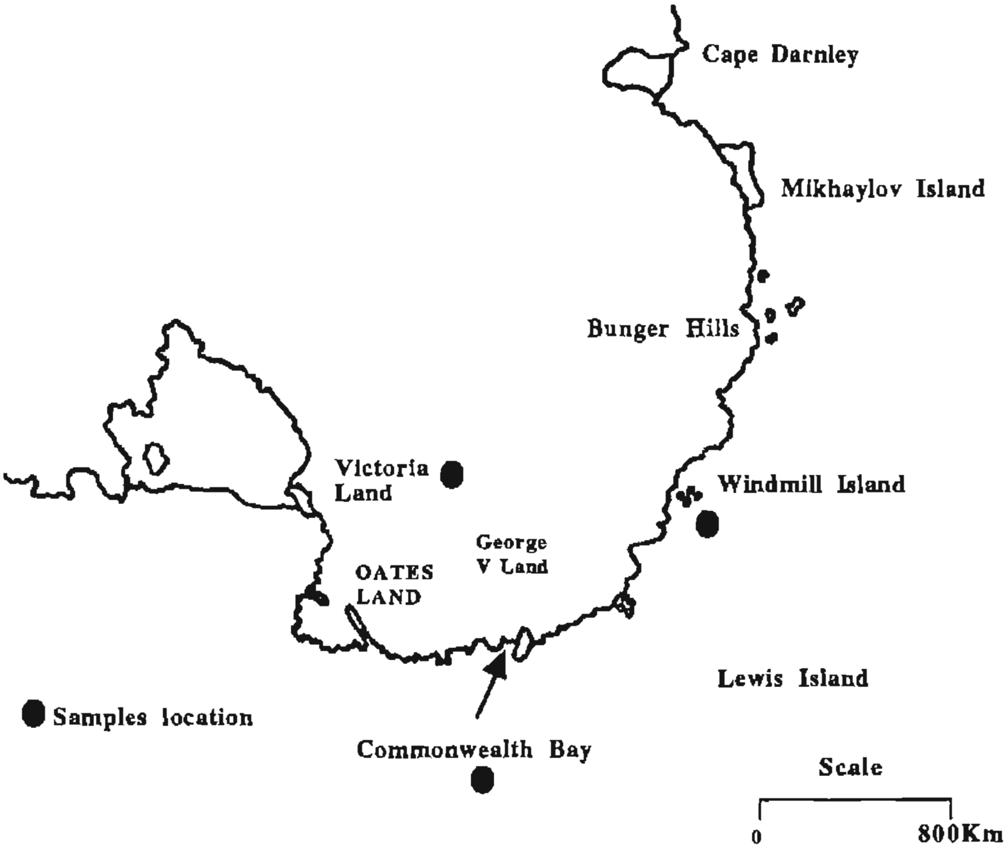


Fig. 1.5. The location of rock samples from East Antarctica.

Heavy Minerals

With regards to provenance of the sediments, different minerals have been studied by Boswell (1933), Milner (1962), and Morton (1985). Epidote, zircon and tourmaline were studied following the studies of Krumbein and Pettijohn (1938), Krynine (1946). The characteristics of rutile were studied following the studies of Krumbein and Pettijohn (1938), Milner (1962), Hubert (1962), and Force (1980). Garnet is the main heavy mineral studied, but ilmenite, zircon, tourmaline, epidote, sillimanite, apatite and staurolite have also been used. The most abundant heavy mineral that we found was garnet. Thus the chemical composition of garnets, following the technique of Morton (1985b), were compared with garnets from various source rocks. The Hafnium contents of zircons, following the techniques of Owen (1987) and Morton (1991) were compared with zircons from various sources. The chemical composition of staurolites were compared, following the methods of Kepezhinskas and Koryluk (1973) and Morton (1984a,b). The chemical composition of epidotes as used by Morton (1984a) and Yokoyama et al. (1990) were also compared with epidotes from various samples.

Trace elements of ilmenite and magnetite grains have previously been studied by Hutton (1950), Blatt (1967b), Luepke (1980), Darby (1984), Darby et al. (1985) and Darby and Tsang (1987). In this study, the trace elements of ilmenite grains has also been used to interpret the source of the Permian sands, and were compared with the trace elements of ilmenites from various sources.

Light Minerals

Quartz grains were initially used by Mackie (1896) and more recently by many sedimentary petrologists such as Keller and Littlefield (1950), Bokman (1952), Doty and Hubert (1962), Blatt and Christie (1963), Basu et al. (1975), Young (1976), Blatt, Middleton and Murray (1980). Thus the number of crystal units in polycrystalline quartz were used for the interpretation of the provenance following the method of Blatt et al. (1980) and Basu et al. (1975). Bokman (1952) used the elongation of quartz grains to interpret the source of sediments, and the same method has been applied in this thesis to interpret the source of the Permian sands.

Trace elements of quartz grains for provenance purposes have been used first by Dennen (1964,1967) and by Suttner and Leininger (1972), so the same procedure was used for quartz from Permian sands and erratics, from East Antarctica and from Encounter Bay samples to compare with each other.

Cathodoluminescence is a useful tool in studying provenance (Sprunt et al. 1978; Owen 1984, 1991; Matter and Ramseyer 1985; and Ruppert 1987). The CL colour of quartz grains have been used to interpret the provenance of the Jackfork sandstone (Arkansas) by evaluating the relative proportions of brown and blue CL quartz in that formation, and in several coeval sandstones for which provenance information was already available (Owen, 1984; Owen and Carozzi 1986; Miller 1988). The CL colours of quartz grains has also been used in this study and compared with the CL colours of quartz grains from Antarctica, erratics and from Encounter Bay samples to interpret the provenance of the Permian sands of southern Australia.

Feldspars were used by many workers including Boles and Coombs (1975), Odom (1975), Stablein and Dapples (1977), Hubert and Reed (1978), Sibley and Pentony (1978) and Waugh (1978) who gave average of microprobe analyses of detrital feldspar. Recently the composition of feldspar grains has been used to interpret the provenance by Trevena and Nash (1979, 1981). In this study the composition of detrital feldspar grains has also been used in the Permian sands of southern Australia and these were compared with the composition of feldspar grains from various sources.

Surface Texture

S.E.M. work on the surface of garnet for the chattermark trails from the glacial sediments has been used by Folk (1975); Bull (1977, 1980) and Gravenor et al. (1978) and quartz grain surface texture by Mahaney (1990) and by Williams and Morgan (1993) for evidence of glacial sediments. Similar S.E.M. analysis was conducted for garnets and quartz grains of the Permian sands in order to confirm their glacial origins and interpret their specific transport histories.

CHAPTER TWO
PERMIAN SEDIMENTS OF SOUTHERN AUSTRALIA

INTRODUCTION

Late Palaeozoic glaciation affected Gondwanaland for over 100 m.y. and glaciogenic deposits can be found on all of the Gondwana continents (Crowell 1983). The thickest accumulations of Late Palaeozoic glacially influenced strata (Lynes Group, 2500m), occur within the Carnarvon Basin of western Australia Veevers and Powell (1987). The Late Palaeozoic glaciation has left its mark on many parts of Australia from the Bonaparte Gulf Basin to Tasmania and from central Queensland to the Perth Basin. The Early Permian in southern Australia was continental in character, the sediments of this age being exclusively of terrestrial origin, having been laid down by land ice and melt water following very severe glacial conditions. Since then southern Australia has undergone important erosion and disturbance by block-faulting, which resulted in the removal of great parts of these glacial deposits. In protected situations, however, remnants have been preserved and still cover extensive areas. Important exposures of these include, many on Fleurieu Peninsula, on the eastern side of Kangaroo Island and at Hallett Cove.

The grooved striated and polished pavements and Permian glaciogenic sediments, recorded by Selwyn (1859) at Inman valley, and by Tate (1877) at Hallett Cove in the Troubridge Basin, were the earliest observations of glaciation in Australia. The South Australian Permian has been studied from scattered exposures of glaciogenic sediments and from land forms, many of which were recorded as being of glacial origin (eg, Tate 1887; Howchin 1898,1903, 1910a,b, 1926,1929a).

Boulder pavements present are the result of direct ice contact, as the ice contact of a basal tillite is subjected to glacial scouring, and to the concentration of clasts from washing by basal melt water (Muller 1983). The surfaces of boulder pavements are commonly striated (Plate 2.1A).

Age of glaciation and glaciogenic sedimentation

All Permian sediments within South Australia studied by Ludbrook (1969) and Wopfner (1970), show that the glaciation was well developed during the Sakmarian. Most of the well dated Gondwana glacial deposits are of very Early Permian age. At least local glaciation occurred during the Late Carboniferous and also during the later Early Permian. The age of the glaciogenic sediments has long been established as Early Permian (Ludbrook 1956, 1967; Harris and McGowran 1971). A Permian or Permo-Carboniferous age had been favoured by earlier workers (Howchin 1898, 1899, 1900, 1910, 1924; Campana and Wilson 1955). Ludbrook (1956, 1967) described Early Permian marine foraminifera in glaciogene sediments from Cape Jervis, localities on Yorke Peninsula, and in cores from the Murray Basin, and suggested that the sediments were at least in part deposited in marine environment. According to recent work (Veevers 1989) the continental glaciation of Australia was most likely Late Carboniferous in age. Thus continental glaciation in the south was partly contemporaneous with alpine glaciation in the northeast (see BMR 1990).

In South Australia, foraminifera (Ludbrook 1956, 1967, 1969), spores (Evans 1964; Martin 1967; Paten 1969), and fragmented leaf impressions (W.K.Harris 1961) indicate a Sakmarian age for most of the glaciogenic and related sediments, but with the base of the section locally as old as Stephanian. Harris and McGowran (1971) described palynomorph assemblages of Permian age in glaciogenic sediments from several localities adjacent to Fleurieu Peninsula. Present evidence suggests that, within the region of the Arckaringa Basin, the Permian glaciation came to an end by about Mid Sakmarian time. If this is correct, the duration of the glaciation period would have been about 7m.y.

Glacial pavements

The evidence for Late Palaeozoic glaciation most impressive to the majority of geologists is probably the widespread occurrence of polished and striated pavements below the sediments. The best of the Australian glacial pavements is that polished brightly upon hard quartzite and exposed discontinuously for a quarter of mile along the top of the sea cliff extending north from Hallett Cove, south of Adelaide in South

Australia. Many of the other exposures of glacial pavements in South Australia are in Inman River Valley, also south of Adelaide, where topography with a relief of (540m) was overridden by ice moving northwestward (David 1950).

Striae and ice movements

Transport directional indicators are most important for reconstruction of the palaeogeography (Fig. 2.0). It is therefore essential to measure as many directions as possible if a reliable regional picture with prevailing directions is to be obtained. The direction of ice movement at Hallett Cove measured by Howchin (1926), Crowell and Frakes (1971) and Bourman and Alley (1990), on one of the two glaciated pavements in this area indicated the direction of ice movement was towards the north and northwest. In South Australia, tillite and outwash debris lie upon clearly striated pavements within glacial valleys, and show that ice sheets with valley tongues moved northward from sources now in a position occupied by deep ocean south of Australia. The underlying Precambrian and Cambrian rocks, eroded by glaciers, show striae, grooves, *roches moutonnées* , modified pre-glacial valleys, and moulded rock-cored streamline forms, all of which provide evidence of ice-movement towards the north and northwest (Frakes and Crowell 1969).

According to Wopfner (1970) ice issued in the Permian from conjugate Antarctica towards South Australia where an elevated block formed a median ridge that divided it and provided foci for local ice accumulation. Evidence presented from areas adjacent to Cape Jervis indicates that movement of Late Paleozoic ice was generally westward over Fleurieu Peninsula. Till fabric data at Cape Jervis, however, show that local ice movement was N.N.W, and it is likely that this was related to the orientation of the bedrock controlled palaeoslope (Alley and Bourman 1984). Similar controls over local ice movement have also been suggested at Hallett Cove to the north of the Fleurieu Peninsula (Sprigg 1942; Milnes and Bourman 1972).

Striae on the western side of Backstairs Passage at Smith Bay, Kangaroo Island are used to infer a westward and northward direction of ice movement (Bourman and Alley 1990). At some localities elongate clasts have a preferred

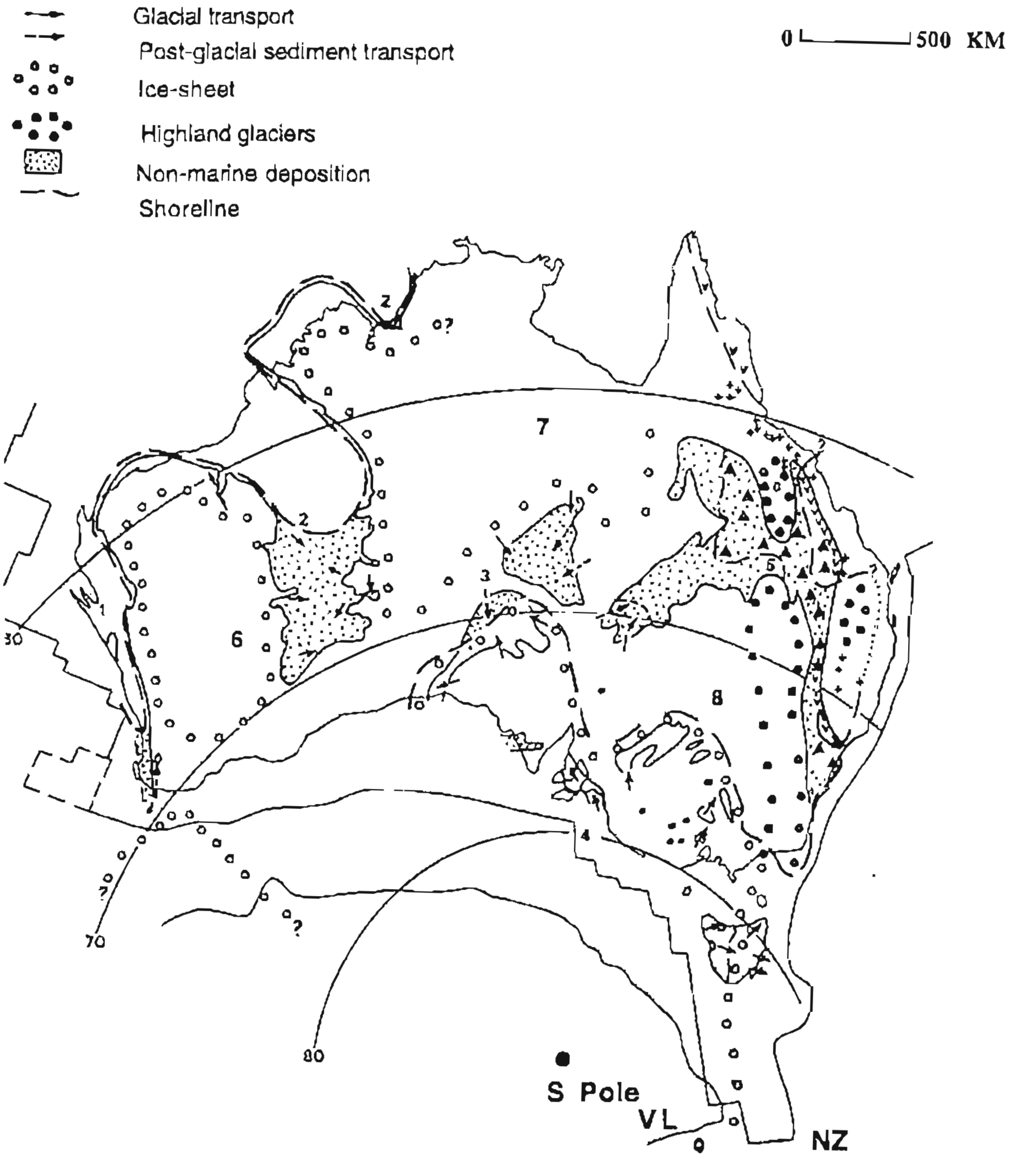


Fig. 2.0. Palaeogeography about the Late Carboniferous/Permian boundary (300 to 280 Ma ago. Depositional provinces (1-5) and platform upland (6-7-8). Also shown are sediment transport directions and distribution of ice (after Veevers 1984).

orientation, their long axes tending to lie subparallel to the pavement striae. Such an orientation is probably parallel to the direction of ice flow (Bourman and Alley 1990). Striae direction of the Hallett Cove pavements are anomalous when compared with those measured from the numerous pavements to the south, moreover Hallett Cove is separated from the pavement localities of Fleurieu Peninsula by at least two major faults along which there has been marked vertical displacement during the Tertiary (Thomson and Horwitz 1962). Evidence for direct glacial deposition is seen in striated boulder pavements overlain by tillite that contains large Victor Harbor granitic boulders, and Kanmantoo Group metasediments that have been derived from the local and underlying basement. The first possibility requires that north trending striae initially developed on bedrock surfaces on Fleurieu Peninsula were obliterated in all localities except Hallett Cove by younger westerly moving ice. Alternatively, the ice movement directions may have been strongly affected by local topography, at least during the retreating and thinning phases. This may imply the concept of multiple glaciations, but there is as yet no unequivocal evidence for this in South Australia.

Over-deepened glacial valleys

A glaciated valley can be fashioned either by a glacier or by an ice sheet. A valley glacier flows in a rock valley. It may originate in an ice field or a cirque. Valleys bearing the stamp of glacial modification occur in many parts of the world, and in some regions exist by the hundreds. In general, a glaciated valley has a steplike long profile, steeper in the headward part than that of a stream valley.

Glacial alteration of a stream valley involves both deepening and widening. In some valleys the volume of rock excavated by deepening exceeds that excavated by widening, but in many valleys widening has been the more important process (Matthes 1930; Crosby 1928; Cotton 1942). In most valleys these changes result in a cross profile of pronounced U-shape, approaching that of a parabola. In the glaciated U valley the side slopes expose much bare bedrock.

Campana and Wilson (1955) showed that the over-deepened floors of several Permian basins in South Australia are comparable in their dimensions to Lake Geneva in

Switzerland. Many basins however, are structurally controlled and their high relief floors are also likely to be the result of differential subsidence during faulting (e.g. Youngs 1975; Wopfner 1972, 1981; Veevers 1984; see also Storey 1991). Veevers and Powell (1987) suggested that ice began to spread westwards and invade these basins during the Late Carboniferous (mid-Namurian), though the onset of widespread glacial deposition only commenced in the Stephanian.

The Permian glaciogenic sediments in southern Australia had different paleo-flow directions: northwestward in South Australia, northeastward in Victoria and eastward in Tasmania along various paleo-valleys. It Inman Valley the glacial valleys have topographic relief of 300m or more, and friction cracks and crescentic gouges document ice flow from east to west. At Hallett Cove the margin or wall of a small valley is clearly exposed and at several places, especially in the vicinity of Glacier Rock, basement knobs are up to 25m in height and are probably modified *roches moutonnées* .

On Fleurieu Peninsula glaciers carved and deepened an irregular system of valleys originally delineated by streams. In view of the vast spread of ice, deep erosion in the southern coast region would be expected when the ice reached its northernmost extent; but instead, details of glacial topography are still preserved in the south, and glacial valleys are infilled with till, outwash, and lake and rare marine sediments.

In Tasmania during the Early Permian times, highlands on the west and northeast occurred with lowland near sea level facing southeastward to the sea in between. A buried U-shaped valley has been recognized near Cradle Mountain and near Karoola (Banks and Clarke 1987).

Lithology and Depositional Environment

The lithology of sediment units is important as it facilitates the correlation between various outcrops. Grain size, mineralogical composition, sedimentary structures, deformations and palaeocurrent indicators are also helpful tools if the depositional history is to be reconstructed. The mixing of detritus during transport by ice results in poor sorting. The Permian glaciogenic sediments comprise three distinct lithologies: diamictite, massive and well laminated sand, laminated clays and silts. The surface exposure of glacial sediments vary in compactness and colour, from unconsolidated buff or light grey deposits easily dug with a shovel, to solidly lithified bluish or greenish grey tillites. These sediments are mostly massive unsorted deposits consisting of pebbles and boulders in a muddy to sandy matrix. The clasts consist of varied granites, gneisses, volcanic rocks, and local geologists have recognized many rock types of known source area, but some clasts have no known source region within Australia (Kenley 1952). The size and proportion of pebbles and boulders vary widely, also their rounding is variable, and large clasts tend to be more rounded than small ones (Plate 2.1F). Such differences may occur due to variations in length of transport, the prevailing transport mechanism and the position of clasts within the ice. The largest erratics in these sediments are mostly granite and metasediments, while the intermediate size erratics sediments are mainly rhyolite, rhyodacite, various porphyritic and metamorphic rocks. The erratics of sedimentary rocks such as sandstones are generally smaller in sizes. In addition, the sediments in South Australia contain numerous dropstones, and these are often visible in stratified sediments typically depressing the underlying laminae (Plate 2.1C). The dropstone diamictites are present in the majority of deposits and require a relatively large open body of water, either lacustrine or marine, before deposition can take place. Permian deposits in southern Australia are generally poorly sorted, immature sediments. Some are quartz rich, produced by reworking by streams during deglaciation.

Permian glaciogenic sediments include a laminated to indistinctly layered facies of poorly consolidated silt and fine sand, in places interbedded with layered to cross bedded sand. Even the most finely laminated sediments may contain numerous dropstones. The concentration of dropstones varies from scattered to abundant.

Structures on tops of some sand beds are ripple marked with some cross bedding. Some of the sand beds are rhythmites and may be turbidites formed by diurnal and annual meltwater turbidity flows along the bottom of lakes (Plate 2.1C). In some areas slump structures are common. Laminated shale in places contain thin sandstone layers with graded bedding and current bedding. Crowell and Frakes (1971a) suggested that sedimentation occurred in a discontinuous marine and lagoonal epeiric sea, but much could be lacustrine.

At Hallett Cove, lacustrine sediments lap onto a Precambrian basement ridge, on which isolated patches of till occur. The basin fill appears to be a mixture of fluviatile and lacustrine deposits with some aeolian sands.

DESCRIPTION OF SAMPLED AND OTHER LOCALITIES

HALLETT COVE

The glacial sediments at Hallett Cove form part of the Late Palaeozoic Troubridge Basin. Hallett Cove is an outstanding locality for Permian glacial pavements and sediments in South Australia. At Hallett Cove, glacial grooves, striae, friction cracks, and percussion marks are cut into an undulating polished surface of Precambrian quartzites and slates which exhibit highly polished and striated pavements (Howchin 1918; Segnit 1940; Sprigg 1942). Boulders of quartzite, granite, gneiss, porphyry, volcanic rocks, and local Precambrian diamictite are present. Some of these boulders are foreign to this region and apparently have been carried more than 80 km from regions to the south. Near the mouth of the Field River which enters the sea at the southern side of the Cove, the beach is littered with erratics that have been washed out of the glacial clay, some of which are of great size (Plate 2.1F)). Among the boulders can be recognized that of Victor Harbor granite, some are very well rounded, soft fine grained micaceous rocks of the Kanmantoo Series, and Precambrian tillite from the Onkaparinga river, as well as other stones characteristic of the country in the south-southeastern direction. The glaciogenic sediments comprise three main facies: lodgement till, subaquatic flow till complex and glaciolacustrine sediments (Bourman and Alley 1990).

Lodgement till facies: The lowest unit in the stratigraphic sequence is a very compact diamicton with a blue-grey, sandy clay matrix exposed on the shore platform at low tide and extending seawards at least 40m.

Subaquatic flow till complex: Overlying the till on the shore platform in the Cove proper, and overlying bed rock south of the Cove, is a unit referred to as subaquatic flow till complex. These diamictons are much stonier than the lodgement till and contain larger clasts of igneous and metamorphic rocks.

Glaciolacustrine facies: These sediments comprise relatively flat lying, fine to medium sand, silt and clay containing occasional thin beds of fine gravel, dropstones and angular clasts of bedded clay. Graded bedding is common in the sandy beds, particularly in the lower part of the sequence, and undoubtedly the beds represent turbidites.

In a review of the Late Palaeozoic glaciation of Australia, Crowell and Frakes (1971 a,b) described glacial features at Hallett Cove and concluded that they were formed in two episodes, during which ice flow was generally to the North. Harris (1971, unpublished Honours thesis) examined the petrology of erratics in order to determine their provenance, and described the erosional effects of glacier ice. He disagreed with the view of Crowell and Frakes that crossing striae necessarily indicated multiple glaciations at Hallett Cove.

Various summaries of, and references to, the geology of the Hallett Cove area have been made, but these do not add to knowledge of the evolution of the glaciogenic sediments. These references include (Glaessner and Parkin 1958; Nesbitt 1969; Oliver and Daily 1969; Talbot and Nesbitt 1968; Cooper et al. 1970., Daily et al. 1979 and Toteff 1983).

Three informal stratigraphic names have been used for the Cape Jervis Formation at Hallett Cove (Harris 1971):

1 **Waterfall Creek beds,**

2 **Amphitheatre beds and**

3 **Sugarloaf sand.**

The Waterfall Creek beds consist of alternating grey green shale beds, thick and thin beds of fontainbleau sandstone (sandstone cemented by calcite crystals). The Waterfall Creek beds rest directly on bed rock. The tops of the sandstone beds are often ripple marked, showing a north-easterly current direction. Erratics are common towards the base of this unit. This is overlain by 12m of Amphitheatre beds. The main lithologies of these beds are purple mudstone and yellow sand. This unit shows turbidite features. The Sugarloaf sand is a massive, coarse yellow sand. It is unconformably overlain by Pliocene limestone. The Sugarloaf sand rests with abrupt irregular, possibly erosional contact on the Amphitheatre beds at the Sugarloaf. The tops of sand beds are ripple marked by westerly currents. Slump folds of isolated shale and angular shale clasts occur. Sand becomes more abundant towards the top, forming beds up to 6m thick. Although the rhythmically bedded clay and sand at many localities here have been regarded as varves by Mawson (1926), the beds are not laterally persistent and do not exhibit the characteristic graded couplets of true varves. Other workers interpreted the clay containing dropstones, and thin sandy diamicton lenses directly overlying the bed rock, as till (Bowen 1958; Toteff 1983) and this may imply multiple glaciation for the Hallett Cove area. The highest outcrop of sediments in the Hallett Cove area, previously unreported, consists of 4m of horizontally laminated lacustrine silt with minor coarse sand lenses, granules and pebbles exposed in the Lonsdale railway cutting at 80m ASL (Bourman and Alley 1990).

FLEURIEU PENINSULA

CAPE JERVIS

Exposure of the Late Palaeozoic glaciogenic sediments at Cape Jervis has long been known; the significance of these beds was first established late last century (David and Howchin 1897). A report on the Cape Jervis area (Glacial Research Committee 1898) considered the glaciogenic sediments there to represent the most extensive occurrence of a genuine till yet discovered in South Australia. Over large areas of Fleurieu Peninsula, and on Kangaroo Island compelling evidence for Late Palaeozoic glaciation is provided by exposures of striated bedrock surfaces exhibiting friction cracks, chattermarks, grooving, polish, gouging and stoss-and lee features, together with extensive boulder clay deposits containing numerous exotic and local erratics, the largest of which are invariably granitic. The till was described as an unstratified deposit, varying in colour from grey to almost black and full of boulders of all sizes. The glaciogenic sediments rest unconformably on smoothed and striated surfaces of basement, Adelaide Supergroup or of Kanmantoo Group metasedimentary rocks. These sediments consist of boulder sands and silts, associated with rhythmically bedded thin sands and clays containing dropstones. Erratics, occur within the sediments and are randomly strewn over the present land surface. Many of them have been locally derived from outcrops of Encounter Bay Granites, although several exotic types can be recognized. During the course of regional geological investigations on Fleurieu Peninsula, Campana et al. (1955) noted the exposure at Cape Jervis and described the beds as moraines consisting mainly of unsorted, unstratified boulder clay and sandy clay containing erratics of all sizes and degree of angularity. A detailed description of this locality is given in Ludbrook (1967) and more recently by Alley and Bourman (1984). Exotic erratics include granites, gneisses, rhyolites, dacites, tuff, a large Victor Harbor granite boulder, and arkose to subarkoses. Here tills of units 1 and 2 (see below) are slumped, they show sharp contacts at the base of boulder bearing beds with similar non-boulder bearing claystones. The slumped beds contain contorted lenses of sandstone and exotic erratics. Bourman and Alley (1990) suggest that these sediments are related to only one glacial advance. The Cape Jervis Formation lies unconformably upon easterly dipping greywackes of the Cambrian Kanmantoo Group. These beds are extensively developed over southern Fleurieu Peninsula and are subject to gully erosion.

The Cape Jervis Formation has been divided into five units on the basis of sedimentological and genetic differences:

1- **Fluvioglacial and glaciolacustrine.**
2- **Lodgement till.**
3- **Fluviolacustrine beds.**
4- **Flow till complex.**
5- **Glaciomarine sediments.**

INMAN VALLEY

The passage of glacier ice over South Australia has long been known with the first discovery of glacial action in Australia being made in the Inman Valley South of Adelaide on the Australian continent by Selwyn (1859a). In Inman Valley the striated pavements show that the ice travelled in a north-westerly direction. Erratics up to 6m in length, in their longer axes, occur in countless numbers in the bed of the River Inman and along the sides of the valley. Most of the erratics in these localities are granite.

In Inman Valley, eight exposures of striated basement floors occur. The deposits are more or less irregularly bedded and include several meters of sand and silt containing boulders of porphyritic granite and thin bedded and laminated sequences of sandstone and mudstone. The beds carry abundant pebbles, mostly water worn, and some are ice-marked, faceted and polished or striated. Above the Inman Bridge erratics are particularly plentiful, both in the bed of the river and on either side of the valley. Large granites occur within a short distance of each other. The quartz of the granites show the bluish opalescent colouring common to the Victor Harbor granite.

KINGS POINT

This locality was first described by Howchin (1910a) and has been commented on by Ludbrook (1967) and Crowell and Frakes (1971a). The Permian sediments occur in a strike valley 30-40m deep, oriented at NE-SW in Kanmantoo Group metasediments. Exotic boulders of erratics are abundant and are mainly gneisses, granites and sandstones.

KANGAROO ISLAND

Kangaroo Island comprises a diverse association of both rocks and land forms that reflect a long and varied geological history. Daily et al. (1979) compiled the first comprehensive account of the Island's geology, whilst numerous other investigators have concentrated on particular aspects of the diverse geology. Some of the more important contributions have been those of Daily and Milnes (1971a,b); Daily et al. (1979); Flint (1978); Flint and Grady (1979) and Milnes et al. (1983).

The generalized tectonic and geological maps indicate that the framework of the Island is based on Cambrian rocks now deformed into an accuate belt that is an extension of Fleurieu Peninsula.

The scarcity of till in the Permian is surprising and the majority of sediments are predominantly shales or sands. This suggests that most till deposited by the ice was reworked by meltwater streams during glacial retreat and redeposited as fluvioglacial sediments.

YORKE PENINSULA

This locality was first recorded by Howchin (1900). A few small and poor outcrops, scattered erratics, and two stratigraphic bores document the occurrence of Permian beds on Yorke Peninsula (Crawford 1965). The Permian sands are generally covered by cliff falls of Quaternary and Tertiary sediments. The best exposures occur at Edithburgh and Waterloo Bay, where 5-7m of brown, massive claystone, with rare erratics may be seen. Where it is fresh, the claystone is black and carbonaceous but on weathering turns brown and blue. This sediment yielded a Lower Permian microflora which has been assigned to Evan's Stage 2, acritarchs, the arenaceous foraminifera Hemidiscus balmei and reworked Upper to Middle Devonian spores (Harris and McGowran 1967). The preservation of the palynomorphs is suggestive of non marine conditions, but the presence of arenaceous foraminifera probably indicates marine influence. The claystone is overlain by a lens of yellow silty very fine sand. Normally the contact between the two is abrupt, but at one place the upper part of the claystone contains a boulder bearing sand 15cm thick. On Yorke Peninsula, bores and poor outcrops record Permian beds which locally contain granite boulders that have a foreign source and resemble granites exposed along the south coast of Fleurieu Peninsula (Crawford 1965) . However several small exposures of blue-grey diamictite without bedding have also been seen in this region. (Crowell and Frakes 1971a).

SOURCE OF ERRATICS

There are three main sources for the debris transported by glaciers and ice caps, i.e. material eroded from the substratum (and if present, valley walls: Larsen and Mangerud 1981; Rastas and Seppaelae 1981), detritus falling from nunataks (Reheis 1975; Lautridou and Ozouf 1982) on the ice surface, and particles that were supplied by the wind. Erratics that are found in the basal glaciogenic sediments of Hallett Cove, Cape Jervis, Kings Point and other localities are mostly similar in composition. The ice movement data for the Permian suggest the existence of an ice sheet centred South of

Portland, Victoria, and west of Tasmania, off the present Australian continental shelf edge. This has suggested the existence of a former land mass in this area (Officer, Balfour and Hogg 1896; Howchin 1926; David 1950). Earlier workers have either postulated mountains raised from the sea floor (Howchin 1926) or the presence of an extension of Antarctica (Officer et al. 1896). The theory of plate tectonics would suggest that Antarctica and Australia were joined in Palaeozoic times.

In South Australia, the data would suggest that the ice flowed over the area now overlain by the sediments of Murray and Otway basins (see Map inside front flap). Therefore it would seem likely that exotic erratics found on Fleurieu Peninsula had been derived in part from the basement underlying the Murray and Otway basins, areas now lying on continental shelves and beyond.

Granitic clasts are most abundant in the lowermost of Permian sediments are probably derived locally from Victor Harbor rocks. Cambrian metasediment clasts have also been found as erratics, probably derived from local basement rocks. Megaclasts (dropstones-lonestones) which occur throughout the Upper Carboniferous and Permian, and massive boulder-bearing sediments have since 1836 been regarded as having been dropped from icebergs that were derived from land based glaciers. The source of dropstones need not be close to the parent rocks, as ice and icebergs can transport over long distances before melting. Howchin (1924) recorded the presence of Victor Harbor granite, feldspar porphyry, Aldgate Sandstone, Sturt Tillite, Tapley Hill Shale, Mitcham Quartzite and Marinoan shales. Large erratics of Howchin's "earthy and siliceous limestones" occur near the mouth of Hallett Creek. They were derived from the upper calcareous part of the Tapley Hill Formation. Boulders and pebbles of igneous and metamorphic rocks in the Permian sediments in southern Australia are similar in character to the Precambrian igneous and metamorphic rocks in the eastern Antarctica and Encounter Bay Granite (see Ch. 4).

Interpretation

The general evolution of Permian depositional environments in South Australia may be interpreted as follows:

occurred, because the sediments are mostly diamictite containing clasts up to 50cm in diameter. Numerous exotic rock types including granites, sedimentary rocks, and various metamorphic varieties are present.

Periglacial or progacial phase: Well laminated, shale, and silty sandstone represent deposition associated with glaciation but not deposited directly by ice. These units are interpreted to be deposits which accumulated in proglacial lakes. It is concluded that in the beginning of the Permian the climate was very cold. Following the glaciation, the climate became cool and humid.

PERMIAN SEDIMENTS IN VICTORIA

Late Palaeozoic sediments of glaciogenic origin were first recorded in Victoria by Selwyn (1859) and McCoy (1861). Since their discovery, a steady stream of geologists has studied various aspects of these sediments. Most workers studied individual outcrop areas; for example Robbins (1973 a, b) examined the Derrinal area outcrops whereas Officer and others (1896), Jacobson and Scott (1937) and Davis and Mallet (1981) examined outcrops in the Bacchus Marsh area. Reviews of this work that included some new information have appeared from time to time (Summers 1923; Kenley 1952; Singleton 1967; Spencer-Jones 1969; Bowen and Thomas 1976, 1988; O' Brien 1981b). The Late Palaeozoic glacial sediments beneath the Murray Basin have never been named though O' Brien (1981b) equated them with the Cape Jervis Formation of the nearby Troubridge Basin. This usage has proved inadequate so he proposed the name Urana Formation (O' Brien 1985).

The most extensive study of the Permian glacial sediments of Victoria is included in Bowen (1959), an unpublished Ph.D thesis. The origin of the Permian glaciogenic sediments has been discussed by Bowen (1959), erratics are abundant, many are faceted. Jacobson and Scott (1937) reported that most were derived from local Ordovician sandstone and quartzite. Less common are igneous and metamorphic rocks, granite pegmatite, greisen, quartz and feldspar porphyry, rhyolite, reef quartz, gneiss, schist, cordierite hornfels, slate and phyllite. Heavy minerals are largely of local origin. Some erratics are exotic and not matched in Victoria, thought they may of course have come from pre-Permian terrain in possible source areas to the southwest, now mantled by younger rocks. Scott (in Jacobson and Scott 1937) listed a variety of erratics of rocks not recorded from the Ballan Graben of Victoria, apparently derived from the southwest and west. In Victoria, Scott (in Jacobson and Scott 1937) has recorded the presence of garnet granites, tourmaline and topaz bearing gneisses, quartz and feldspar porphyries, gneiss, tourmaline rock, schist and hornfels. He was unable to assign sources to many of these, although some were quite distinctive.

In Victoria, striae on bed rock pavements and within the glacial sequences dominantly trend between 010 and 090. Various indications of transport of sediments: imbrication of boulder pavements, cross bedding and imbrication in conglomerate, show transport directions from southwest to northeast (Bowen et al. 1976).

According to Crowell and Frakes (1971a) during the Permian time, continental glaciers occupied Victoria, and moved from the south and southwest toward the north and northeast across undulating terrain that locally had relief of over 100m, but was very near sea level.

A Sakmarian age for the Permian beds in Victoria is indicated collectively by *Gangamopteris leaves* (Mc Moy 1861; Chapman 1927; Kenley 1952 p.59; Singleton 1967 p.190), and spores (Douglas 1969).

PERMIAN SEDIMENTS IN TASMANIA

The earliest published record of the effects of ice transport in Tasmania, and probably in continental Australia is that of Joseph Milligan (1849 p.18), who explained the presence of a granite boulder in clay formed by *in situ* weathering of Permian sedimentary rocks near Southport as due to glaciers, or more probably icebergs. Such boulders, as noted by Banks 1981 are common in marine Permian rocks in Tasmania. More than 600m of marine and nonmarine Permian strata, only gently deformed, occur in Tasmania.

All clasts are markedly larger in grain size than the enclosing marine sediment, some are demonstrably dropstones, and a very few are faceted and striated. All observers have recognized them as dropstones, most regarded them as having been transported by and dropped from icebergs.

Age and palaeogeography

The Permo-Carboniferous glaciogenic deposits of the Tasmania Basin were deposited entirely under glaciomarine conditions. Tasmania glaciogenic sediments crop out over the western two-thirds of the state (Fig. 2). The age of the Wynyard Tillite ranges from Stephanian, as shown by *Rhacopteris* and other plant fragments from low in the section (Gulline 1967) to within the Sakmarian, in as much as late Sakmarian fossils occur in the overlying Quanby and Golden Valley Groups. Dropstones of probable glacial origin continue on up into the Kazanian (Banks 1962). During the Late Carboniferous (Stephanian) much of the present island was covered by an ice sheet flowing into Tasmania from the present west. The ice flowed over an irregular topography, and may have flowed into fjords in places and left nunataks in others. For much of this time it was covered by shallow cold shelf areas in which marine sediments with dropstones were deposited, and an abundant, but low diversity benthic foram and molluscan fauna thrived (Banks & Clarke1987). Available evidence strongly suggests that the Late Carboniferous glaciation of Tasmania was predominantly sheet glaciation (Dickins 1985). During the Late Carboniferous and throughout the Permian, Tasmania lay in high southern latitudes (Smith et al. 1981).

Plate 2.1

A- Crescentic gouges and glaciated pavement formed from the movement of Permian glacier over the basement rocks (Precambrian) at Hallett Cove of South Australia.
(Photo V. Gostin)

B-Massive tillite in the lower part of Permian glacial sediments at Hallett Cove that is covered by stratified Permian sands. (Photo V. Gostin)

C-Dropstone in the interbedded of sands and muds (varves) in the glaciolacustrine deposits of Permian glacial sediments at Hallett Cove. (Photo V. Gostin)

D- Massive tillite with lenses of coarse grained sands within the Permian glacial deposits at Cape Jervis.

E- Angular unconformity between Precambrian rocks and overlying Permian glacial sediments at Hallett Cove.

F- Boulder size of well rounded Victor Harbor granites as a erratics in the Permian glaciogenic sediments at Hallet Cove of South Australia.

A

B

C

D

E

F

CHAPTER THREE

TEXTURE OF THE PERMIAN SANDS AND THEIR BEARING ON THE NATURE OF GLACIATION

INTRODUCTION

Textural characteristics of the sediments refer to the grain size, grain morphology (sphericity, roundness and surface texture) and fabric. The use of grain shape to identify sedimentary environments assumes that grain morphology reflects environmental history (Krinsley and Doornkamp 1973; Bull 1981). The scanning electron microscope (SEM) can be used to distinguish surface texture of grains from different sediments. It can also be used to recognize certain geological deposits and transport mechanisms (e.g. Krinsley and Doornkamp 1973). Detrital quartz and garnet are common in the Permian sands of southern Australia and contain characteristic surface textures that have been produced during erosion and transportation. Identification of a suite of surface textures on quartz grains gives some information about their depositional environments (Krinsley and Doornkamp 1973; Margolis and Krinsley 1974; Eyles 1978, 1988; Williams and Morgan 1993). Among the depositional environments, glacial deposits are included (Krinsley and Takashashi 1962; Coch and Krinsley 1971; Eyles et al. 1983, 1993). Whereas the mineralogy of the Permian sands and erratics, indicates the provenance of the Permian sediments, the stratigraphic changes in mineralogy, and the surface texture of light and heavy minerals give useful information concerning history of transportation, the persistence of glaciation and the specific environments of deposition.

Textural maturity and surface texture of the Permian sand grains indicate the presence of well rounded aeolian grains together with angular and subangular glacially transported grains with glacially formed chattermark trails, and the more common fluvial features. In addition, detailed study of the Permian sands has revealed a prolonged history of grain transport suggesting multiple glaciations. This chapter explores this theme. The main objective of this study however, is to clarify and refine depositional models for sand accumulation in different environments by determining the source, and transport history of quartz and garnet grains in samples that have been collected at different localities.

CHATTERMARK TRAILS ON GARNET GRAINS

The only natural features resembling these tracks are glacial chattermarks (Chamberlain 1888; Gilbert 1905) which also form trails of parallel concave marks. These are formed by glacial scour over fixed bedrock that results in rhythmic release of strain with regular fracturing as the glacier passes over the surface. Folk (1975) confirmed that the chattermark trails on garnets were due to glacial transport. He also mentioned that the garnets from nonglaciated areas contain no chattermark trails.

Gravenor et al. (1978) searched for similar chattermark trails on quartz, zircon, tourmaline, and epidote minerals, but did not find any such trails. Perhaps the chattermarks occur on garnet because the garnet, at 7.5 on Mho's scale, is one of the hardest minerals available and is the only common isotropic heavy mineral that has no selective breakage directions (Folk 1975). In a more recent study of heavy mineral suites from the Late Paleozoic glacial sediments of Africa, Australia and Antarctica, Gravenor (1979) found that garnet is the dominant heavy mineral in these sediments and about 33% of the garnets are unetched and have chattermark trails on their surfaces. Folk (1975) and Gravenor et al. (1978) suggested that the percentage of garnets with chattermark trails is related to the distance of glacial transport and that these were probably transported over long distances by glaciers of continental dimensions. Garnets that are fully etched, such as those found in the Late Paleozoic of Antarctica and in the Ordovician glacial deposits of Sierra Leone and South Africa (Gravenor 1980), do not have any chattermarks on their surfaces so, they may have been chemically etched following burial.

The chattermarks on the grain surfacs

The percentage of chattermarked garnets increases with the length of time that the grain is held in englacial transport, however, breakage during transport will remove some of the pre-existing trails. As grains are about to be deposited, the amount of breakage will decrease and the number of chattermark trails on the surface increase (Dreimanis and

Vagners 1971).

In this study nearly 100 grains of garnets in the Permian sands were examined by S.E.M and most of the grains showed chattermark trails on their surfaces. The chattermark trails on the surface of garnet grains in these sediments owe their origin to stick slip motion of particles rubbing against one another within the ice or particles held fast in the ice and being ground against the underlying bedrock. Some of these grains had two to several chattermark trails (Plate 3). The percentage of garnets with chattermark trails in the Permian sands is high, and based on these tracks, it is concluded that most of detrital garnets in the Permian glaciogenic sediments have been transported for a very long distance. This evidence also indicates that the grains underwent intensive glaciation over a prolonged period. Thus, the source of most garnet grains in the Permian sands should not be close to the site of deposition. Some of the angular garnets grains in the Permian sands did not showed any chattermark trails on their surface. Therefore, they were probably transported for a short distance from the source area prior to deposition, or may have been derived from local sources.

Rounding and chattermark trails on the garnet grains indicate that the garnets of the Permian glaciogenic sediments of southern Australia were not picked up directly from the shield rocks of Gondwanaland, and were not deposited immediately by the ice. These features of garnet grains in the Permian sands suggests that the rounded grains were either inherited or first eroded from the shield rocks, transported and deposited by ice, and subsequently reworked by a readvance of the ice. At this stage, the chattermark trails developed on the rounded grain surfaces. This sequence of events was undoubtedly repeated many times during the Upper Palaeozoic glaciation which may have lasted as long as 75 my. (Crowell and Frakes 1975). The etched garnets of the Permian sands are angular with sharp corners indicating post-sedimentary dissolution. As such they are not discussed further in this thesis.

SURFACE MICROTEXTURE OF QUARTZ GRAIN

As mentioned earlier, the scanning electron microscope has frequently been used to study the microtexture of quartz sand grains and this has also proved to be a valuable tool in analysis of sedimentary environments. The quartz was chosen because of its resistance to mechanical and chemical breakdown and apparent lack of cleavage.

Study of quartz sand grains in the Permian sands by S.E.M. exhibits a mixture of at least two quartz sand types, each characterized by distinctive morphologies. The quartz grains of the first type are highly angular to subangular and contain arc-shaped, steplike and conchoidal fractures on their surface. These indicate that they were transported into the study area by glaciers and subsequently reworked in subaqueous environments (Plate 3.1A).

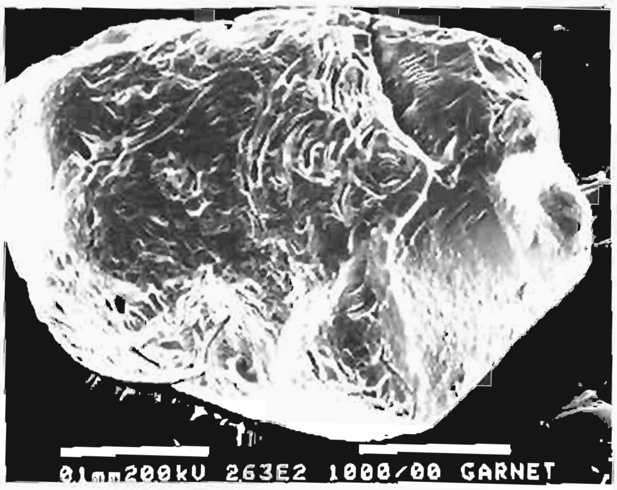
The second type of quartz grains are rounded to subrounded. This type of morphology is formed by mechanical processes that are operating within fluvial and especially beach and aeolian environments (Plate 3.1A). The texture of Permian sands can support the hypothesis of local redistribution of sand grains as a result of coastal or aeolian activity during deposition of the Permian glaciogenic sediments. As seen in Plates 3.1A,E and 3.1F, the pitted microtexture suggests an aeolian mode of transport. This is very similar to Mahaney et al. (1988a) statement that a large amount of glacially crushed quartz grains, primarily from unweathered glacial deposits of Wisconsin (Ontario Basin) appears to have been locally redistributed by aeolian transport.

A number of different fractures such as parallel and subparallel steplike, arc-shaped step, and conchoidal breakage face were observed on the quartz grains of the Permian sands. These grain surface textures usually show evidence of intense wear and abrasion which is caused by glacial grinding or mechanical action. This clearly indicates that most of the Permian sands were derived from glacial ice. Parallel striations of varying length are more common and is one of the best criteria for glaciation (Plate 3.1B) (Krinsley and Doornkamp 1973). Long parallel ridges are one of the most important glacial characteristics, and can have resulted from either fracture or gouging. Subparallel steps on quartz grains of the Permian sands which are shown in (Plate 3.1F) are probably inherited characteristics which are thought to be related to a fracture direction; probably representing one of the main quartz fractures (Frondel 1962). Some of the quartz grains show silica precipitation on grain surfaces (Plate 3.1E). Above a pH value of 9, 'amorphous' silica readily goes into solution (Ller 1955). Small amounts of capillary water in compacting sediments might well give suitable conditions for the dissolution of silica from quartz grains which have a disrupted lattice. These condition are especially likely to occur where a feldspar grain is in high stress contact with quartz grain (Whalley and Krinsley 1974). Evidence seen in Plate 3.1E suggests that some overgrowths were formed before glacial transport.

DISCUSSION

More than 60% of the quartz grains in the South Australian Permian sands are angular to subangular in shape. They show crushing and abrasion on grain surfaces resulting from ice activity. The rest of the quartz grains are rounded to subrounded and have no sign of any crushing, which may indicate fluvial to aeolian activities. The study of surface texture of quartz and garnet grains in the Permian sands has shown that a large part of the sediments have been transported by ice movement. The edge sharpness with arch-shaped steps on quartz grains in the Permian sands were produced by continental glaciation (Plate 3.1C). Overall, most of the glacially crushed quartz grains have the edge sharpness when compared with quartz grains from coastal and aeolian environments (Krinsley and Doornkamp 1973). The quartz grains studied here also have surface features very similar to quartz grains from glacial sediments of Greenland (Whalley and Krinsley 1974), and North America and East Africa (Mahaney and others 1988a). Mazzullo and Anderson (1987) interpreted the quartz shape variation from angular to subrounded as due to derivation from either till or glacio-marine sediments. They also noted that most of the glacio-marine grains were angular and irregular in appearance, with only a small proportion subrounded in shape. Arc-shaped steps and crescentic gouges in the Permian sands (Plate 3.1B) are thought to be abrasion features by ice. These types of abrasion features were observed on many quartz grains of the Sunnybrook diamicton in America (Krinsley and Takahashi 1962), and they may reflect the pressure within the ice during transport, or at the time of subglacial emplacement (Mahancy et al. 1988a). The shape analyses of quartz grains (Plate 3.1A) in the Permian glaciogenic sediments indicate three general shapes ranging from rounded, subangular and angular grains. The types of surface texture (arc-shaped steps, conchoidal breakage face and parallel to sub-parallel steplike fractures), and the degree of crushing and abrasion, in quartz grains of the Permian sands, indicate different transport histories. These grains exhibit compound glacial and fluvial to aeolian surface textures. Glacial textures include parallel, subparallel and conchoidal breaks on grain surfaces and nonglacial texture shows generally minor smoothing that can result by stream activity and the well rounded grains by aeolian erosion. On the bases of surface textures and the shapes of quartz grains, it is suggested that the Permian sands have been transported by at least three stages. The first stage was in the solid phase that produced the angular grains with different kind of crushing on the surface, the secondary stage was aqueous when the grain shapes changed from angular to subrounded and rounded, and the third stage was the time of wind action that made the grains well rounded. Some rounded grains, however, were probably derived from older deposits with rounded grains.

Plate 3. SEM photomicrographs of chattermark trails on the surface of sand size garnet grains in the Permian sands of southern Australia. This indicates that the garnets have passed through a glacial environmental systems.


0.1mm200kV 263E2 1000/00 GARNET

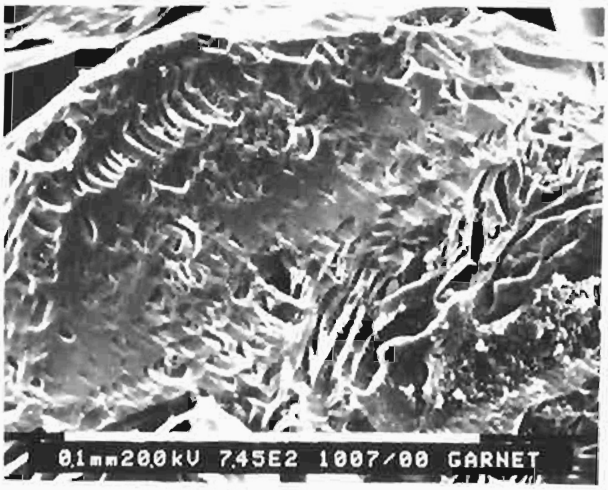

0.1mm20.0kV 7.45E2 1007/00 GARNET

Plate 3.1. Representative SEM photomicrographs of angular to rounded quartz grains showing glacial crushing and non-glacial features in the Permian sands of southern Australia.

A-Very angular to rounded quartz grains (q) and subangular feldspar grains (f). Differences in quartz shape suggests different depositional environments, ranging from glacier (very angular) to subangular (glaciolacustrine) to rounded (aeolian).

B- Subangular quartz grain with parallel step-like (p), arc-shape step-like (a) and gouges (g) fractures.

C-Angular quartz grain with conchoidal fracture (c) that contains arc-shape fractures (a). Parallel step-like (p) and gouges (g) fractures are also present.

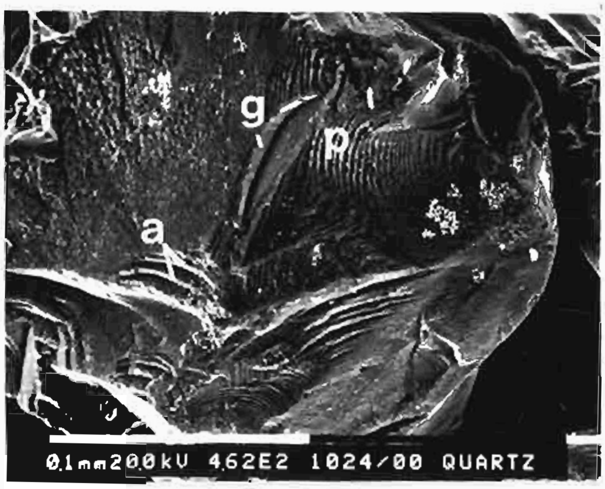
D- V-shaped fracture (v) on the surface of a very angular quartz grain.

E- Fractures on the surface of detrital quartz with silica overgrowth (o) suggesting that fracture has take place after precipitation of silica cement, because part of the cement has also broken during transportation in glacial environment.

F-Step-like fractures (s) on the surface of central quartz grains. Half of the quartz grain on the left contains pits that have probably formed during transportation in aeolian environment.


f
q
1mm200kV 750E1 1017/00 QUARTZ

A


g
p
a
0,1mm200kV 462E2 1024/00 QUARTZ

B


g
c
a
0,1mm200kV 406E2 1021/00 QUARTZ

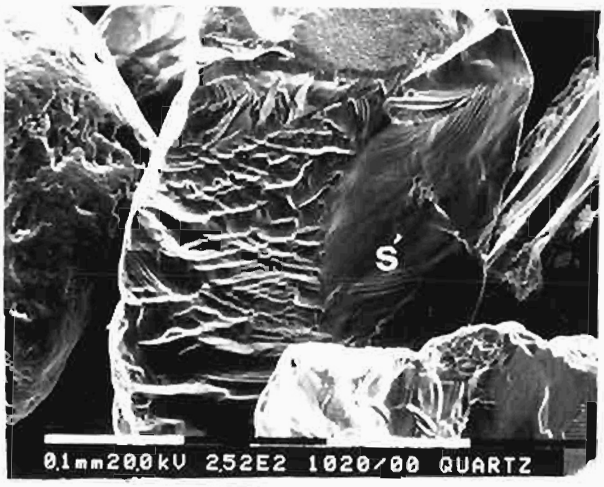
C


v
0,1mm201kV 388E2 1010/00 QUARTZ

D


o
0,1mm200kV 388E2 1023/00 QUARTZ

E


s
0,1mm200kV 252E2 1020/00 QUARTZ

F

CHAPTER FOUR

GEOLOGY OF SOURCE AREAS FOR THE PERMIAN SANDS

INTRODUCTION

The potential source areas for the Permian sands of southern Australia are composed of a variety of igneous, sedimentary, and metamorphic rocks from different localities that have been studied in this thesis. The geology of each locality will be described in the following pages. The palaeocurrent data indicate that the source direction of the sediments was from the South and southeast. Therefore, the latter is likely to provide part of the source for the Permian sands of southern Australia. Different rock samples from eastern parts of Antarctica, Encounter Bay and Kanmantoo Group on Fleurieu Peninsula have been selected for this study as the source areas for the Permian sands. The samples from East Antarctica (long; 110°E, 142°E lat), in the localities of Windmill Island and Commonwealth Bay are composed of medium to high grade metamorphic rocks with some granite and volcanic intrusions. The samples from Encounter Bay are mostly granitic in composition and the Kanmantoo Group are metasediments. Investigations of light and heavy minerals characteristic of various source rock types at different localities were undertaken to interpret possible sources for the Permian sands. These samples were analysed by the same methods that was used for the Permian sands, counting the same number of grains and documenting the same characteristics. The results are illustrated as the Tables and Figures in this chapter.

Method

Hand specimens and thin sections for study were obtained from the collections of R. L. Oliver, Milnes and Fleming held at the Department of Geology and Geophysics and Tate Museum at the University of Adelaide. These samples represent a wide variety of rock types from different localities. More than 30 samples from each rock type have been used. The samples were broken down by two stages: Jaw crushing that was followed by a final grinding in a rotary plate crusher. Then the samples were sieved by using 3 Ø to

2.5 0 size sieves. Finally, the heavy minerals have been separated from the light minerals and then mounted on glass for petrographic and microprobe analyses (see Appendix 5).

ENCOUNTER BAY

General geology

The earliest geological investigation of the Encounter Bay Granites is a brief petrological description of a "biotite granite" from Granite Island, in Encounter Bay by Moulden (1895). Subsequently Gartrell (1903) published a paper about mineralogy and chemistry of the potash feldspar megacrysts from Granite Island. He recorded an anomalously high Ca content in a megacrystic granite of Granite Island.

The first detailed descriptions of the Encounter Bay Granites and their field relationships were published by Tilley (1919a, 1919b), and concern the outcrops at Cape Willoughby in Kangaroo Island. Tilley recognized various types of granite, including a major megacrystic granite and subordinate aplites and vein-like albitites. Browne (1920) described the field relationships and petrology of the Encounter Bay Granites in the Encounter Bay area. He also found that a megacrystic granite was the major granite type in this area. Both Browne and Tilley recognized the relationship between the granites in the Encounter Bay area and at Cape Willoughby. Based on the opinion of Howchin (1906, 1910) that the country rock metasediments in the Encounter Bay area were of Cambrian age, and also on the fact that the granites are overlain by glaciogenic sediments of presumed Permo-Carboniferous age, Browne supported Tilley's suggestion of a Palaeozoic age for the granites.

A lower Palaeozoic age for many of these granites has been established in recent years as the result of radiometric age dating techniques (Richards 1962; Compston, Crawford and Bofinger 1966; White, Compston and Kleeman 1967; Thomson 1970; Dasch, Milnes and Nesbitt 1970; Foden et al. 1990b).

The Encounter Bay Granite outcrops is one of two intersecting zones of foliated and non-foliated granitic rocks in southeastern- South Australia and southwestern Victoria. Granitic outcrops along the coast at Encounter Bay, in particular at Rosetta Head and Granite Island, have been suggested to form the northwestern wall of an extensive granitic mass (Milnes 1990).

General description of rocks

The major granite type in the Encounter Bay area is coarse grained granite (more than 2mm). The coarse grained opalescent blue quartz is a distinctive characteristic of these granites. Rock types from different localities, such as Port Elliott, Granite Island, Wright Island, Rosetta Head and West Island, have been studied. For detailed petrographic study, at least 300 grains were identified and point counted per thin-section. The mineralogy of these rocks is simple and composed of quartz, feldspars and micas. The major minerals in this assemblage are quartz (33%), K.feldspar (49%), plagioclase (13%) and micas (2.4%). Accessories form about 2% of total grains and include zircon, tourmaline, garnet and opaques.

LIGHT MINERALS IN ENCOUNTER BAY GRANITES

Quartz

The mean percentage of the total quartz in these rocks is about 33%. The proportion of monocrystalline quartz is 6.4% and polycrystalline quartz is 93.6%. Most of the polycrystalline grains have straight crystal boundaries and a uni-modal crystal size distribution. The number of components within polycrystalline grains varies from 61% with less than five crystals to 39% with more than five crystal units. Distribution for the number of components is shown in (Fig. 4.0). Distribution of elongation coefficients in these samples has shown that about 40% of grains have a mean less than 1.5 (Table 6.2). Almost all quartz grains show a blue colour under CL (Plate. 6.1C).

Feldspars

The mean percentage of total feldspar grains in the rock samples from this locality is 51%. Feldspars have been divided into two groups: plagioclase and K feldspar. The potassium feldspars are the dominant group in these rocks . The composition of feldspar is shown in Table. 4.0 and ranges from $An_2Ab_1Or_{97}$ to $An_{37}Ab_{62}Or_1$ (Fig. 4.2). The mean ratios expressed as percentages of K, Na and Ca in these rocks are 52%, 38% and 10%.

HEAVY MINERALS IN ENCOUNTER BAY GRANITES

Heavy mineral assemblages in Encounter Bay granites include mostly tourmaline, zircon, ilmenite and garnet. Garnet is mostly deep red in colour and the chemical composition of garnet indicates that they are rich in Mn and Fe, and mostly have almandine -spessartine composition with relatively low contents of pyrope component. The mean percentage of almandine is 47%, spessartine 23%, pyrope 5% and grossular 22% (Fig. 4.4). Composition of garnet grains is shown in Table. 4.1 and Fig. 4.5, and the concentration of different elements and oxides in the garnet grains is shown in Fig. 4.6. Zircon grains in these rocks are mostly colourless and euhedral in shape. Chemical composition of zircon indicates that the mean percentage of hafnium content in zircon grains is 1.57 (Table. 4.2 and Fig. 4.8). The ilmenite grains show a wide range of TiO2 content between 47 and 54% (Table. 4.3). The concentration of MnO and MgO in the ilmenite grains show that about 75% of grains have more than 5wt% MnO and about 64% have more than 0.4wt% MgO (Fig. 4.9 to 4.11). The chemical composition of heavy minerals is shown in Tables. 4.1 to 4.3 as a representative of typical heavy mineral assemblage.

MOUNT MONSTER

In the southeast of South Australia there are several basement inliers. Among them is Mount Monster, nine km south of Keith, where an exposure of a red brown, quartz-feldspar, porphyry forms a low ridge. Mawson and Dallwitz (1944) interpreted this porphyry to be a shallow sub-surface intrusive, while others (eg Rochow 1971) have postulated it to be extrusive. The rock samples in this locality are mostly volcanic and range from rhyolite to rhyodacite. The samples consist of quartz, feldspar and groundmass with accessories of sphene, opaques and zircon. The composition of feldspar ranges from An5Ab95Or0 to An2Ab61Or37 (Table. 4.4 and Fig. 4.3). The chemical composition and trace elements of the rock samples from Mt Monster is shown in Table. 4.5. The trace elements of these samples have been compared with the trace elements of Permian volcanic erratics at different localities and suggest that the Mount

Monster volcanic rocks are quite distinct from the volcanic erratics in Permian sediments (Fig. 4.7). Therefore, the derivation of volcanic erratics in the Permian glaciogenic sediments remains in question.

KANMANTOO GROUP

General geology

The Mt. Lofty Ranges and Kangaroo Island consist of an arcuate belt of folded and metamorphosed Precambrian to Lower Palaeozoic sediments that are present in a broad structural high between Cainozoic rocks to the east and west. The Kanmantoo Group is a thick repetitious sequence of metasedimentary rocks (Sprigg 1942; Thomson and Horwitz 1962) forming the major part of the Mount Lofty Ranges metamorphic belt to the east of the Older Proterozoic basement inliers and on the southern parts of Fleurieu Peninsula and on Kangaroo Island.

The age of the Kanmantoo group and its relationships with rocks of known Lower Cambrian and Proterozoic age (Howchin 1907, 1910; Tille 1919b) indicate that it is probably Cambrian. Sprigg and Campana (1953) also supported the age of Kanmantoo Group as Cambrian.

General description of rocks

The Kanmantoo Group is a sequence of impure arkoses, greywackes and pelitic rocks, with minor calcareous rocks and sulphide horizons (Sprigg and Campana 1953). The Kanmantoo Group rocks contain both andalusite and sillimanite (Offler 1966). Majority of samples from the Fleming collection are metasedimentary rocks such as schists and quartz feldspathic schists. These rocks are common in Mt Lofty Ranges (e.g. Mills 1963; Offler 1966; White 1966). The major mineral assemblages observed in these rocks are mostly quartz, plagioclase and mica. The accessory minerals consist of garnet, staurolite, andalusite, apatite, zircon, sphene, rutile, opaques, sillimanite and trace of tourmaline.

Semi-pelitic schists

These rocks are a very common rock type in this area and contain andalusite or staurolite. The major minerals constituent in semi-pelitic schists are quartz with an average of 48%, plagioclase 19%, K feldspar 5%, mica 21%, staurolite 4%. The common accessory minerals are garnet, apatite, zircon and tourmaline.

Pelitic schists

The pelitic schists are very similar to the semi-pelitic schists, except that the potassium feldspars are rare. Semi-pelitic and pelitic schists contain andalusite and staurolite, and they are very common in most of the rock samples. The mean percentage of quartz is 43%, plagioclase 27% and micas 30%.

Andalusite schists

The major minerals in these rocks are quartz with an average of 29%, plagioclase 19% , micas 41% and andalusite 11%. K feldspar has been recorded in a few samples. Pale pink garnet is present but generally in low amounts. Accessory minerals include apatite, tourmaline, opaques and zircon.

Staurolite schists

Quartz with an average of 48%, plagioclase 2.5%, micas 35%, staurolite 13% and garnet 2% are the major minerals in these rocks. Quartz is generally coarser in these rocks as compared with the andalusite schists. The accessory minerals include zircon, tourmaline and opaques.

Andalusite-staurolite schists

The major minerals in these rocks are quartz 37%, plagioclase 16%, micas 34%, andalusite 11% and staurolite 2%. According to Fleming (1971) who studied these metamorphic rocks of the Kanmantoo Group, the composition of plagioclase in the quartz feldspathic schists ranges from An5 to An15 and in semi-pelitic schists range from An9 to An25. The accessory minerals are garnet, zircon, ilmenite and tourmaline, Garnet is rare and appears as pink in colour. Tourmaline is rare and is commonly zoned with blue-green to pale greenish pleochroic pleochroism.

LIGHT MINERALS IN KANMANTOO GROUP

Quartz

The mean percentage of quartz grains in these samples varies from 29% to 48% with an average of 41%. Most of the quartz grains are polycrystalline with sutured crystal boundaries and a bi-modal size distribution.

Feldspars

The feldspar content ranges from 3% to 27% and can be divided into two groups: plagioclase (17%) and K feldspar (3.2%). The composition of feldspar ranges from $An_{33}Ab_{64}Or_3$ to $An_{35}Ab_{65}Or_0$ as shown in Table 4.6. and Fig. 4.2, that indicates generally an andesine antiperthite.

HEAVY MINERALS IN KANMANTOO GROUP

Heavy mineral assemblages in the Kanmantoo Group metasediments consist mostly of staurolite, sillimanite, garnets, zircon, tourmaline and ilmenite. The composition of each heavy mineral is illustrated in several tables and figures placed at the ends of chapters 4 & 7. The composition of garnet grains in these rocks demonstrate that the almandine content constitutes 77%, spessartine 9%, pyrope 11% and grossular 3% (Fig. 4.4) and (Table 4.7). Comparison of the composition of garnet in the Kanmantoo Group with the Permian sands indicates that the garnets of the Kanmantoo Group are different with respect to the proportions of FeO, MgO, MnO and CaO (Fig. 4.6). This suggests that the garnets in the Permian sands should be derived from sources other than the Kanmantoo Group. The chemical composition of rutile is shown in Table. 4.8. In addition, the composition of staurolite in Kanmantoo Group metasediments shows similarity with the staurolite grains in the Permian sands and indicates that the staurolite in the Permian sands may have been derived from the Kanmantoo Group (Table. 4.10, Fig. 7.7). Hafnium content in the zircon grains ranges from 0.9% to 1.9% with a mean percentage of 1.4% (Fig. 4.8).

The TiO_2 in ilmenite grains shows a narrow range content between 49% to 52% (Fig. 4.9) and (Table. 4.9).

The concentration of MnO and MgO in ilmenite grains from these rock samples shows that the concentration of MnO with less than 5wt% is 99% and MgO with less than 0.4wt% is 96% (Fig. 4.10 and 4.11). The chemical composition of apatite and sillimanite are shown in Tables. 4.11 and 4.12.

EAST ANTARCTICA

General geology

Along the coastline of East Antarctica, from the Amery Ice Shelf at 70°E to Cape Adare at 170°E, geophysical evidence has to date shown that three Phanerozoic sedimentary basins are present on the continental shelf. The ages of these are unknown, but Cretaceous and Early Tertiary sequences are suspected (Truswell 1982). Further east, radio echo-sounding delineation of subglacial topography has suggested that the extensive Wilkes Basin of the continental interior reaches the East Antarctic coast in the vicinity of the Ninnis Glacier and Cook Ice Shelf. The area that is drained by the Mertz Glacier may represent a separate sedimentary basin. The recently reported discovery of in-situ continental siltstones of Cretaceous age in the offshore from the Ninnis Glacier supports the concept of sedimentary basins in the area (Truswell 1982). The Wilkes Basin extends about 1400 km south from the George V Coast, where it is about 600 km wide, to lat 83° S, where is only 100km wide.

COMMONWEALTH BAY

Commonwealth Bay is situated on the coast of East Antarctica (at about 67° 24′ S, 142° 40′ E) and occurs approximately 1500km to the East of the Windmill Islands, on the westernmost edge of George V Land. The East Antarctic Shield is one of the largest continuous granulite facies of the metamorphic terranes in the world. Although different parts of the shield experienced peak metamorphism at different times between the Middle Archaean and the Late Proterozoic. It is interesting to note that different parts of the terrain followed a variety of types of retrograde pressure-temperature paths from a surprisingly similar metamorphic peak geotherm (Stüwe and Powell 1989).

Regional geology

The geology of Adelie Land shows that this part of East Antarctica has undergone multiple high grade metamorphism with different grades. These metamorphic events occurred at different times in different areas around Commonwealth Bay. From Commonwealth Bay to Pointe G'eologie, four units are recognized and all of them show high grade metamorphic paragenesis (Monnier et al. 1993). "Granitic gneisses", outcropping from Port Martin to Cape Denison, are embedded in garnet bearing migmatitic paragneisses that feature in the Pointe G'eologie and Cape Gray areas. The predominant rock type at the sampling locality at Cape Dension is a granitic to granodioritic orthogneiss of amphibolite facies which has been intruded by aplitic and amphibolitic dykes (Stillwell 1918; Stüwe and Oliver 1989).

Charnockitic gneisses are found in the easternmost part of Commonwealth Bay. At Cape Hunter, east of Port Martin, most of the rocks are phyllites. Cape Hunter consists of phyllites that have been metamorphosed in the mid greenschist facies, and a specimen was recently collected from this area which forms a prominent coastal outcrop on the west side of Commonwealth Bay, George V Land. In thin section this phyllite consists mainly of quartz, micas, chlorite and feldspars with accessories of high Mn garnet, sphene and opaques mostly ilmenite. West of Port Martin, garnet-bearing granitic and dioritic gneisses are the most common rock types. Around Dumont d'Urville and Pt. G´eologue, the continuation of these gneisses are interrupted by minor layers and lenses of pelitic cordierite-rich gneisses.

Outcrops of Cape Denison rocks are mainly composed of charnockitic, orthopyroxene-quartz-feldspar gneisses and garnet bearing granite gneisses. Based on U-Pb, Oliver et al. (1983) obtained a age of 2600-2400 Ma for the orthogneiss at Cape Denison. Menot et al. (1993) petrological data from North to South in this area indicates that a prominent granodioritic massif (I. des Empereurs) is progressively invaded by coarse grained pink granitoid dykes (I. des Manchots). The diversity of the rock-types in ascending order are as follows:

1- Folded migmatitic gneisses.

2- Granodioritic massif.

3- Pink granitoids.

4- Granitic dykes.

General description of rock samples from Commonwealth Bay

For the purpose of interpreting the source area for the Permian glaciogenic sediments, rock samples from Commonwealth Bay have been studied by petrographic microscope. The major rock types in Commonwealth Bay are granite to leuco gneisses and granodiorite gneisses with some phyllite at Cape Hunter. These rocks have been intruded by aplite and amphibolite dykes. Majority of minerals in Commonwealth Bay rock samples are quartz, feldspars and micas. The dominant accessory minerals includs zircon, epidote, opaques and garnet.

Granites

These rocks consist essentially of coarse grained quartz and alkali feldspars. Quartz is mostly polycrystalline and unimodal with an average of 39.8%. K feldspar is also dominant with an average of 54%. Plagioclase composes of about 3% of these rocks and mica is present with an average of 2.5%. Accessories are mainly opaques and form less than 1% of the rocks.

Leuco gneisses

These rocks are mostly composed of quartz, feldspars, micas and chlorite. Quartz grains are mainly polycrystalline with an average of about 30% and always show undulose extinction. Potassium feldspars are dominant with 33%, and plagioclase is about 22%. Micas and chlorite are present and form about 12% of these rocks. Accessories are mostly opaques and zircon, forming about 2%.

Granodiorite gneisses

Quartz in these rocks is polycrystalline grains and forms about 30% of the rocks. Quartz grains show bimodality in size. Feldspar is mostly plagioclase with an average of about 29% and K feldspars about 22%. Micas also form about 8% of these rocks. Accessories include opaques with an average of about 2%, and traces of zircon and hornblende.

WINDMILL ISLAND

The first published document about the geology of the Windmill Island was by the Russian workers (Ravich and Voronov 1958; Ravich 1960; Voronov and Krasik 1963). Rocks at the Windmill Island Antarctica (long. 110E, lat. 66S) consist of a layered sequence of schists, gneisses, granitic gneisses and migmatites that are intruded by a charnockite and a porphyritic granite and are also cut by two swarms of east-west trending dolerite dykes (Blight 1975).

The heavy minerals in these rocks between longitude of 140-150°E are mostly garnet, amphibole and epidote, and at longitude 110°E, heavy minerals are also garnet, amphibole and epidote (Von der Borch and Oliver 1968).

Regional geology

Rocks in the Windmill Islands make up part of the Antarctic shield, a poly metamorphic terrain which is called by Harrington (1955) the "basement complex". This terrane consists mostly of high grade metamorphic rocks with some lower grade green schists. These metamorphic rocks occur as gneisses, migmatites and schists that are mostly sedimentary in origin and are intruded by acid and basic rocks. Large intrusive charnockitic bodies are common.

Many age determinations have been carried out on the metamorphics of this basement complex. More recently, Arriens (1973), has reported the following Rb/Sr whole rock ages.

George V Land	1500-1700 m.y
Vestfold Hills	2500 m.y. and 500 m.y
Prince Charles Mountains	2000-2800m.y, 1200-1000m.y. and 700-500 m.y.

General description of rock samples from Windmill Island

The rock types in Windmill Island are composed mostly of metamorphosed acid gneiss or granite gneiss, migmatite gneiss, and some pegmatite and charnockite. Acid gneiss makes up approximately 50 percent of the total outcrop in Windmill Island (Blight

1975). The major minerals in these group of rocks are quartz, feldspars and micas. The dominant accessory minerals include garnet, sillimanite, rutile, zircon, epidote and opaques (essentially ilmenite), with lesser amounts of pyroxene and staurolite.

Acid gneiss

Quartz is mostly polycrystalline, with an average of 50%, and bimodal with sutured grain boundaries. Feldspars are mostly plagioclase, with the mean percentage of 30% and microcline 8%. Microclines mainly show simple twinning. Micas are mostly biotite 5% that is partly altered to chlorite and muscovite 1%.. Accessory minerals are mostly garnet and zircon that form about 1%, and apatite and epidote that form less than 1% of the rock.

Mafic gneiss

Quartz grains in these rocks are mostly bimodal polycrystalline with sutured boundaries and form about 6% of the rocks. Feldspars are mostly plagioclase, with a mean percentage of 50%. Potassium feldspar is rare. Micas are mostly biotite with the mean percentage of 20%. Pyroxene forms about 15% of the rocks. Accessories includes ilmenite and garnet that form about 1% of the rocks.

Ribbon gneiss

Here, quartz is mostly polycrystalline with an average of 45%. Feldspars are mostly K feldspar and form about 45%, and plagioclase about 8%. Mica is mostly biotite with a mean percentage of 4%. Accessories include opaques 2%, garnet and zircon about 1% with a trace of epidote.

Granite gneiss

Quartz grains are mostly polycrystalline with an average of 40%. Feldspars are mostly potassium feldspar 43%, and plagioclase 15%. Accessory minerals are opaques and zircon, forming less than 1% of the total rock.

Aplite

Quartz is mostly polycrystalline and unimodal grain size with an average of 40%. Feldspars are mostly K feldspar, with a mean percentage of 45%, and plagioclase about 10%. Accessories are opaques (essentially ilmenite) form about 2%, garnet and zircon with less than 1%.

Basic schist

Quartz is almost all polycrystalline and bimodal in grain size with a mean percentage of 10%. Feldspars are mostly plagioclase with averages of 50%. Mica is mostly biotite with a mean percentage of 24%. Hornblende forms about 15%. Accessories are mostly ilmenite with a mean percentage of 2%.

Charnockite

Charnockite is mostly composed of polycrystalline quartz 24%, plagioclase 40%, microcline 25%, hornblende 10%, pyroxene 2%. Accessories include opaques 4% and zircon 1%.

Porphyritic granite

Feldspars are the main constituent of these rocks and consist of microcline 60% and plagioclase 10%. Quartz forms about 20% of the rocks. Micas are biotite 2% and muscovite 1%. Accessories are composed of opaques 1%, zircon and apatite less than 1%.

Basic gneiss

Quartz is rare in these group of rocks and the major minerals are feldspars, mostly plagioclase with a mean percentage of 63%, and hornblende about 35%. Accessories are composed of opaques 1% and zircon less than 1%.

LIGHT MINERALS IN EAST ANTARCTICA SAMPLES

Quartz

The mean percentage of quartz grains in the rock samples from Windmill Island of East antarctica is shown in (Table. 4.13). Number refer to the number of rock samples that has been examined.

Quartz	Polycrystalline	Monocrystalline	Samples
50.6	97.7	2.3	acidic gneisses 33
1.5	98.6	1.4	basic gneisses 27
33.4	91.8	9.2	granites 25

Table. 4.13. Mean percentage of quartz grains in the rock samples from Windmill Island of East Antarctica.

Most of the polycrystalline quartz grains in the metamorphic rock samples show suture boundaries, and bi-modal type crystal size distribution for components within polycrystalline quartz grains. The number of components within polycrystalline grains vary from 34.4% with less than five crystals, to 65.6% with more than five crystal units in the metamorphic rock samples. Distribution of the number of components in polycrystalline quartz grains is shown in Fig. 4.0 where we can see there are considerable differences between the metamorphic and plutonic rock samples. Distribution of elongation coefficients of monocrystalline quartz grains in the metamorphic rock samples shows that about 5% of the grains have less than 1.5 mean, and 95% have value of greater than 1.5 mean. In the plutonic rock samples nearly 69% of the quartz grains have an elongation coefficients more than 1.5 mean, and 31% have value of less than 1.5 mean (Fig. 4.1). The CL colour of quartz grains in the metamorphic rock samples is brown and in the plutonic rock samples it is blue (plate. 6.1B).

Feldspars

The mean percentage of feldspar grains in metamorphic and plutonic rock samples from Windmill Island of East Antarctica is shown in Table. 4.14.

Feldspars	Plagioclase	K feldspar	Samples
56	30.7	70.3	acidic gneisses
63	100	0	basic gneisses
60.5	13	87	granites

Table. 4.14. Mean percentage of feldspar grains in the rock samples from Windmill Island of East Antarctica

As the Table shows, the potassium feldspars are the dominant feldspar group in the Antarctic rock samples (Table. 4.15). The composition of feldspars in the gneisses samples ranges from An2Ab0Or98 to An2Ab98Or0, in the granite gneisses it ranges from An7Ab93Or0 to An1Ab0Or99, and in the granites it ranges from An4Ab0Or96 to An5Ab95Or0 see (Fig. 4.12).

HEAVY MINERALS IN ANTARCTIC ROCKS

The chemical composition of heavy mineral assemblages in the Antarctic rock samples have been studied and the results are illustrated in Tables and Figures. The percentage of each heavy mineral, in these samples, are different and is related to the rock types.

Metamorphic rock samples are composed mainly of garnet, rutile, sillimanite, ilmenite, zircon, tourmaline, epidote and trace of chromite and titanium rich SiO_2.

Garnet is the dominant accessory mineral in the metamorphic rock samples essentially in gneisses and shows different colour ranges from pink to colourless. The chemical composition of garnet is related to the rock types and is shown in Fig. 4.13 and Tables. 4.16. Comparison of garnet in the gneisses and phyllite rock samples indicates considerable differences in composition. The mean percentage of different garnet groups in the gneisses and granitic gneisses rock samples of Windmill Island demonstrates that the almandine component constitutes 67%, spessartine 10%, pyrope 22% and grossular

3% (Fig. 4.14). The mean percentage of garnet groups in gneissic rock samples of Commonwealth Bay shows that almandine constitutes 60.4%, spessartine 5%, pyrope 31.2% and grossular 3.4%. The phyllite rock samples from Commonwealth Bay show dissimilar composition of garnet grains when compared with gneiss samples and demonstrates that almandine constitutes 32%, spessartine 58%, pyrope 4% and grossular 6% (Fig. 4.14). The concentration of different elements and oxides in the garnet grains from each rock types is shown in Fig. 4.15. Therefore the chemical composition of garnet in the gneiss rock samples is rich in Fe and Mg and thus is almost almandine-pyrope in composition with relatively low contents of spessartine and grossular components but, in phyllite rock samples it is rich in Fe and Mn and almost almandine-spessartine in composition with relatively low contents of pyrope and grossular components (Fig. 4.14). The concentration of MgO and MnO in ilmenite grains from metamorphic samples shows that the concentration of MnO with less than 5wt% is about 88% and MgO with less than 0.4wt% is about 99% (Figs. 4.10 and 4.11).

In the granitic rock samples the colour of garnets ranges from deep red to red. The chemical composition of garnets in the plutonic rock samples demonstrates that the almandine constitutes 47%, spessartine 33%, pyrope 16% and grossular 3% (Fig. 4.14). Therefore the chemical composition of garnet in the granite rock samples is rich in Fe and Mn, and thus is almost almandine-spessartine in composition with relatively low contents of pyrope and grossular components. The concentration of MnO, TiO2 and MgO in the ilmenite grains in the granite rock samples are much different from metamorphic rock samples (Tables 4.17 and Fig. 4.16). Concentration of MnO with less than 5wt% is 34% and MgO with less than 0.4wt% is 85% (Fig. 4.10 and 4.11).

The mean percentage of Hf content in zircon grains from metamorphic rock samples is 1.65 and in granite rock samples is 1.37 (Fig. 4.17 and Table. 4.19). The chemical composition of tourmaline, staurolite, epidote, rutile and sphene in the rock samples of East Antarctica are shown in Tables. 4.18, 4.20, 4.21, 4.22 and 4.23. These Tables are representative of a typical heavy mineral assemblage.

DISCUSSION

Comparisons of heavy and light minerals in the Permian sands in Chapters 6 and 7 with the source rocks can be summarised in the following paragraphs:

Detrital quartz grains in the Permian sands can be mostly matched with quartz grains in the plutonic and metamorphic rock samples. This can be supported by the elongation, number of crystal units and the CL colours. The chemical composition of feldspar grains in granite rock samples of Encounter Bay and East Antarctica and gneissic samples of Antarctica are composed mostly of potassium feldspar and can be compared with the feldspar grains in Permian sands.

Heavy mineral suites, such as garnets, staurolite, sillimanite, rutile, epidote, ilmenite rich in TiO2, and purple zircon in the metamorphic samples may have provided the heavy mineral suites for the Permian sands. The chemical composition of garnet from Kanmantoo Group metasediments has a different composition to the garnet of Permian sands and Antarctic rock samples. The composition of garnet in Encounter Bay rock samples shows dissimilarity with the garnet grains of Antarctica and Kanmantoo Group rock samples. The garnet composition in phyllite rock samples from Commonwealth Bay in East Antarctica has a different composition with the garnet grains in the gneissic rock samples of Antarctica. Though volcanics are present as erratics, the contribution of volcanic sources to Permian sands is probably very small. It is concluded that garnet, rutile, sillimanite, epidote, ilmenite and zircon from high to medium grade metamorphic rock samples in East Antarctica, and staurolite and zircon grains from Kanmantoo Group metasediments, with quartz and feldspar grains from granite and gneisses rock samples can be the sources for the Permian sands of southern Australia.

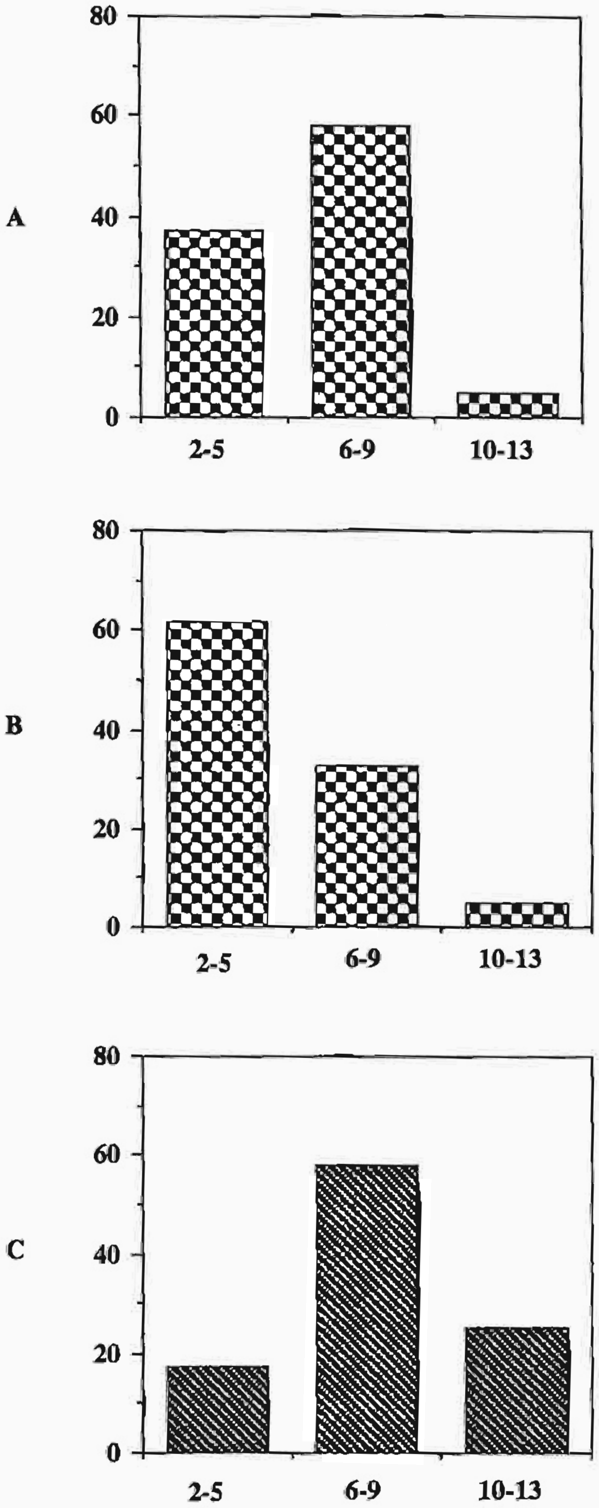


Fig. 4.0. Number of crystal units in polycrystalline quartz grains of different rock samples.
A: granite from Permian erratics
B: granite from Encounter Bay
C: gneisses from Antarctica

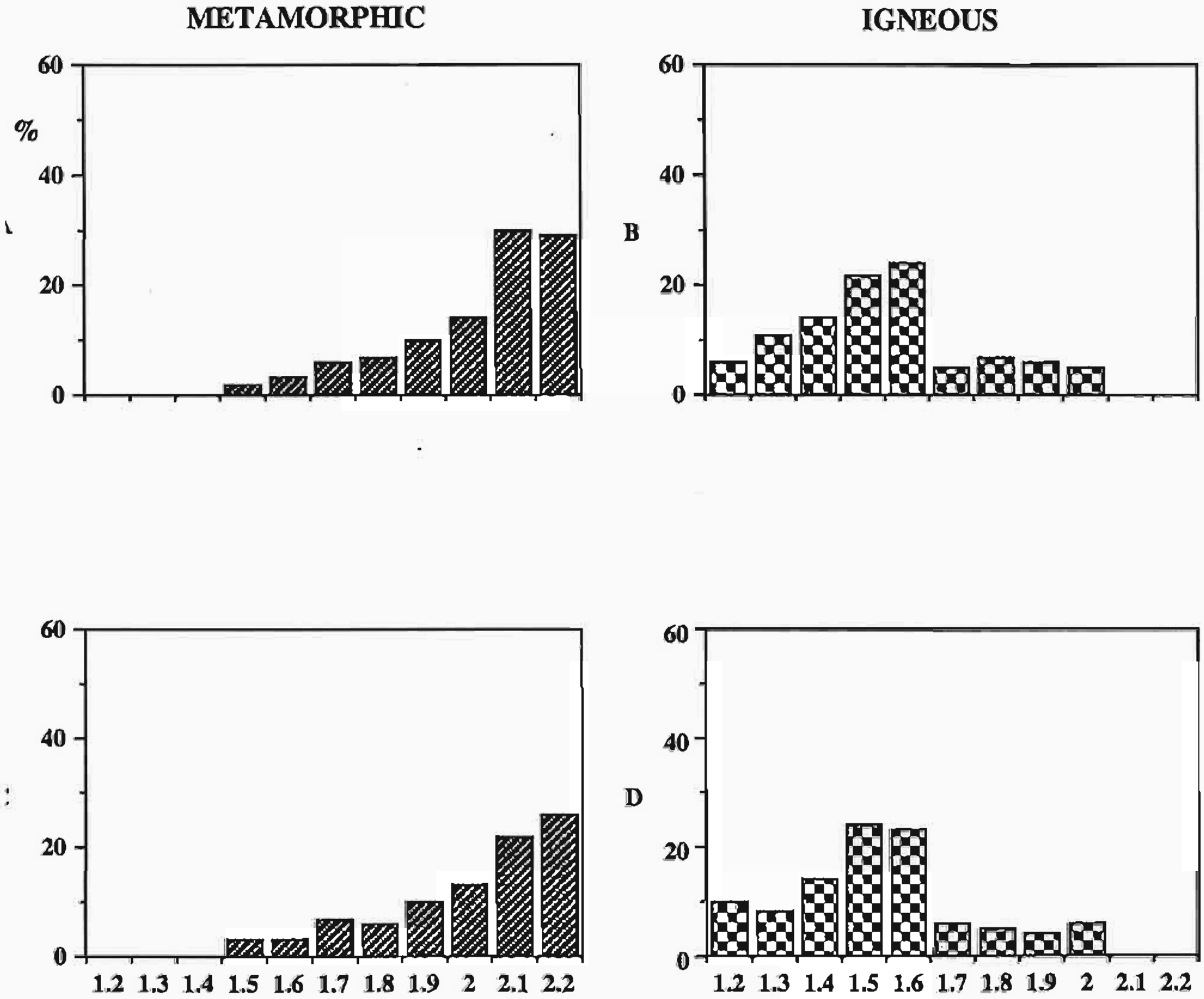


Fig. 4.1. Elongation of quartz grains in the metamorphic and igneous rock samples.
A: gneissic of Permian Erratics
B: granitic of Permian Erratics
C: gneissic of East Antarctica
D: granitic of East Antarctica

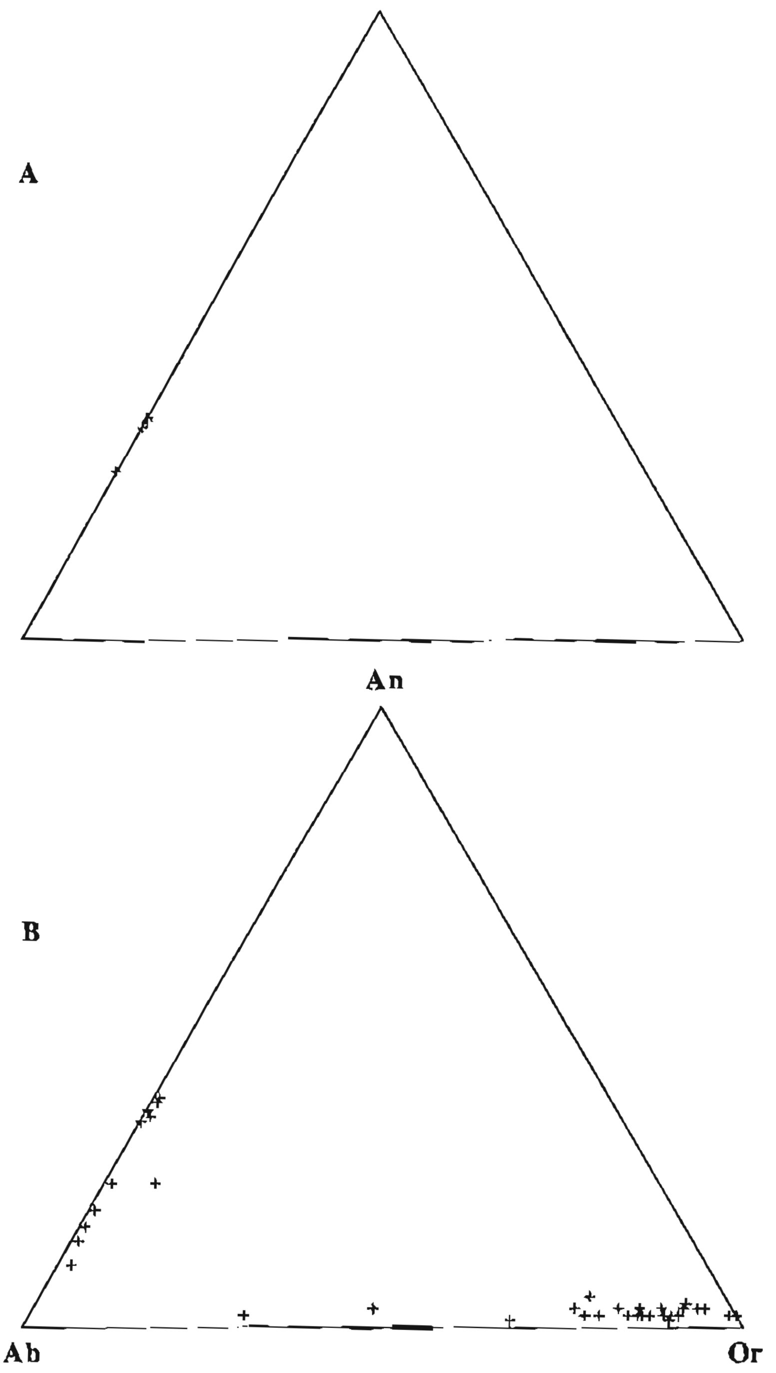


Fig. 4.2. Composition of feldspar grains (weight %) in Kanmantoo Group metsadiments (A) and Encounter Bay Granite (B) as determined by electron microprobe analyses displayed on a triangular diagram: An (anorthite), Ab (albite) and Or (orthoclase) on the base of 32 (O).

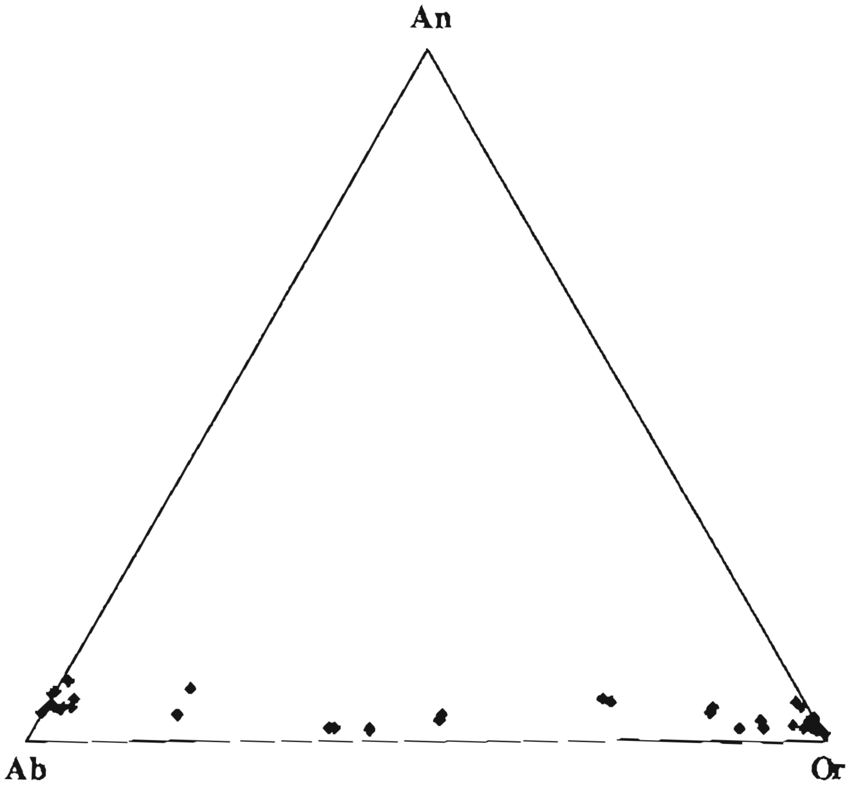


Fig. 4.3. Composition of feldspar grains (weight %) in the porphyritc volcanic rock samples from Mt Monster of southern Australia as determined by microprobe analyses desplayed on a triangular diagram: An (anorthite), Ab (albite) and Or (orthoclase) on the base of 32 (O).

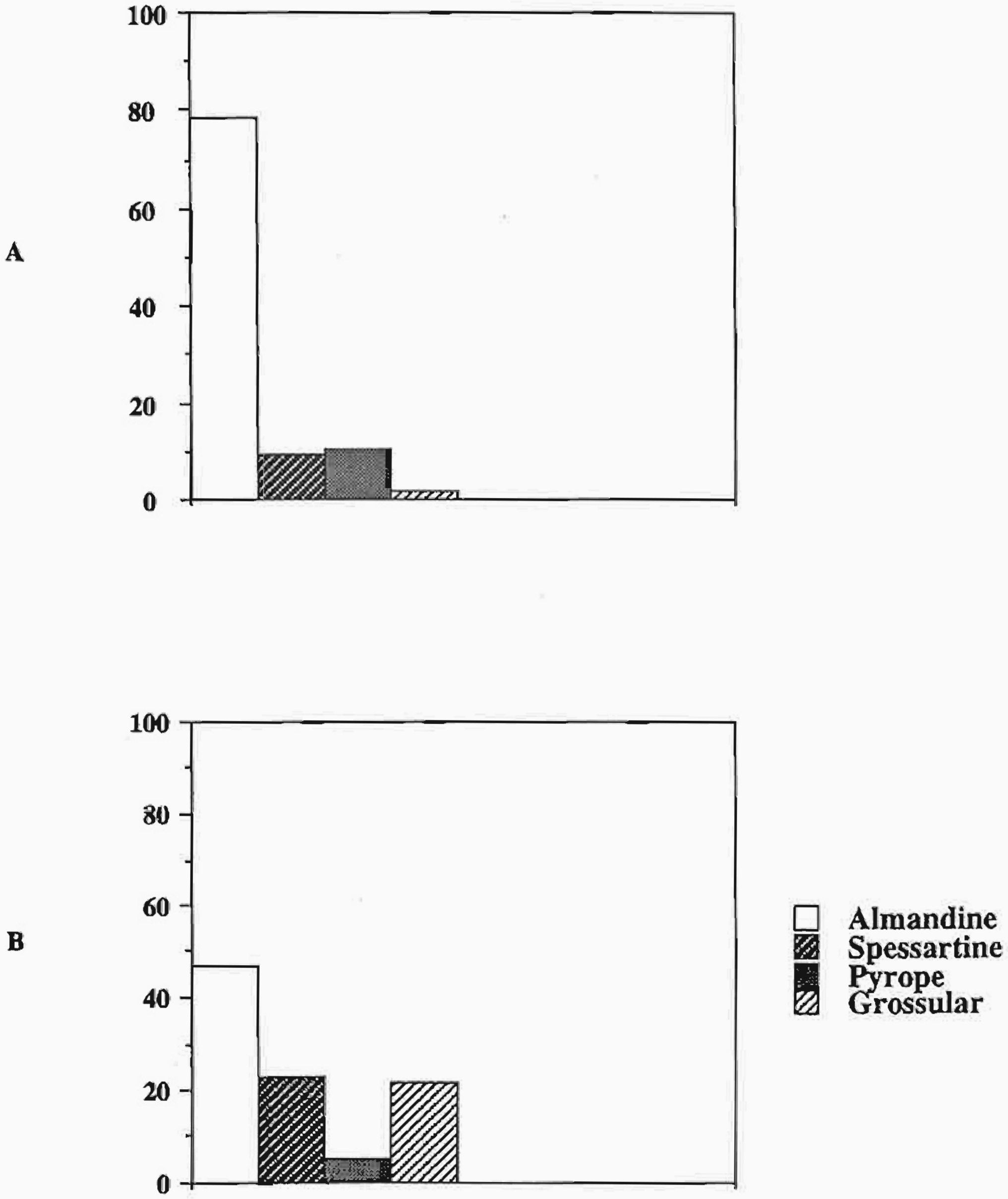


Fig. 4.4. Frequency distribution of garnet groups in the Kanmantoo Group metasediments (A) and Encounter Bay Granites (B) rock samples of southern Australia.

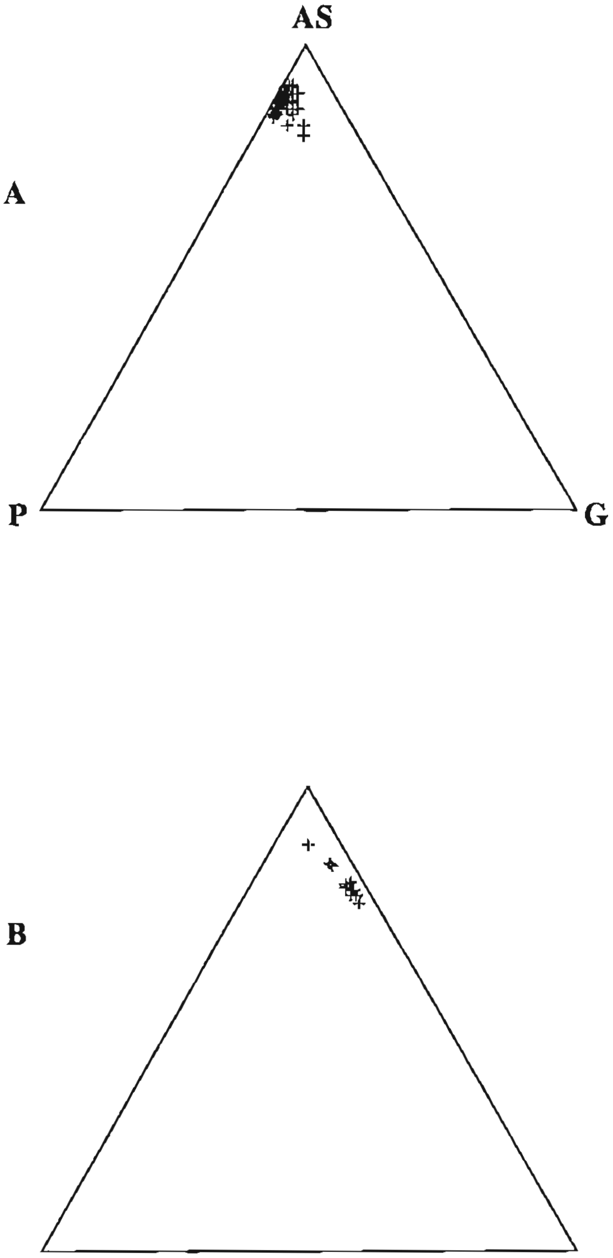


Fig. 4.5. Garnet geochemistry in the rock samples of Kanmantoo Group (A) and Encounter Bay (B), displayed on a triangular diagram: AS=Almandine+spessartine; P= Pyrope and G= Grossular.

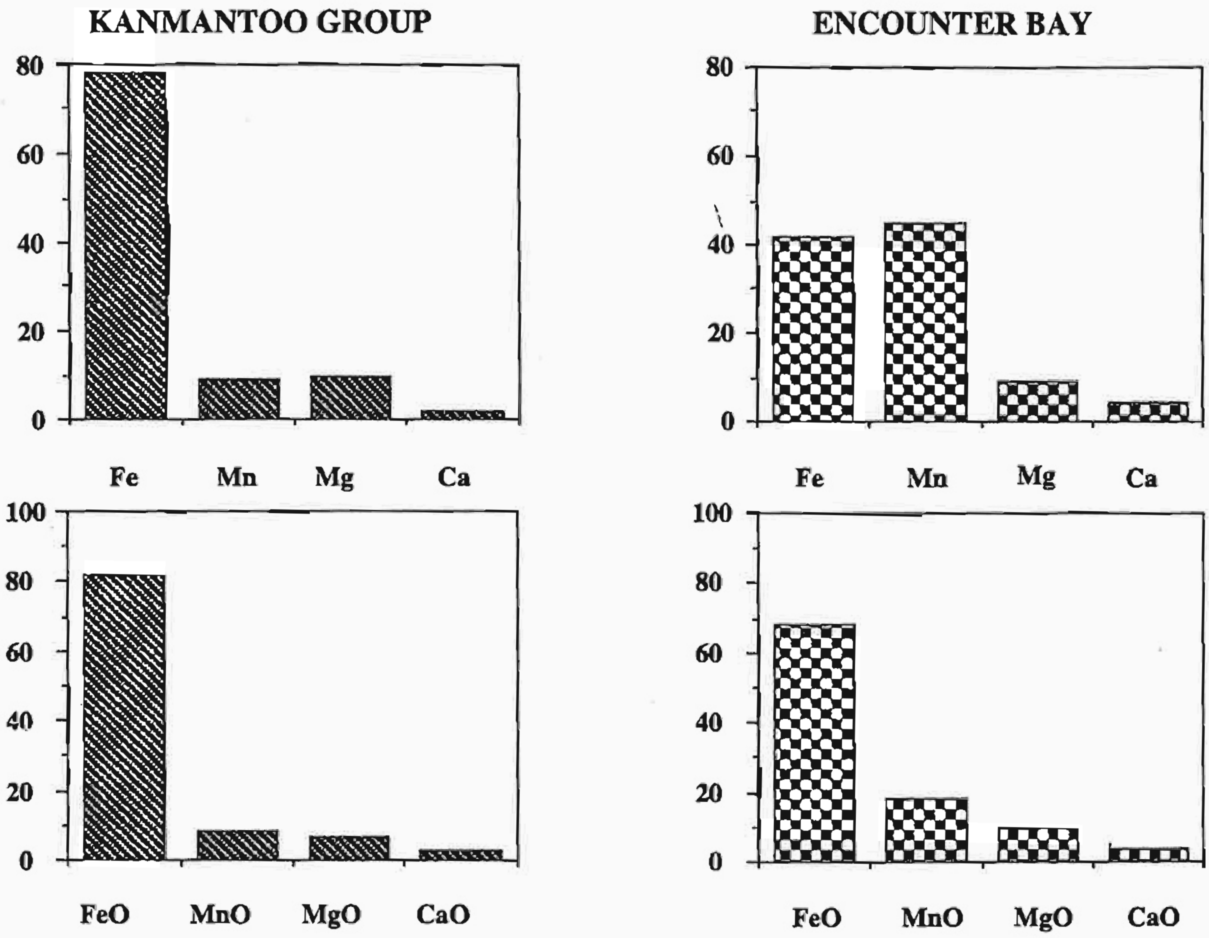


Fig. 4.6. Weight percentage on the base of 24 (O) of different elements and oxides of Fe, Mn, Mg and Ca in the garnet grains in metamorphic and igneous rock samples from Kanmantoo Group metasediments and Encounter Bay Granitic rock samples.

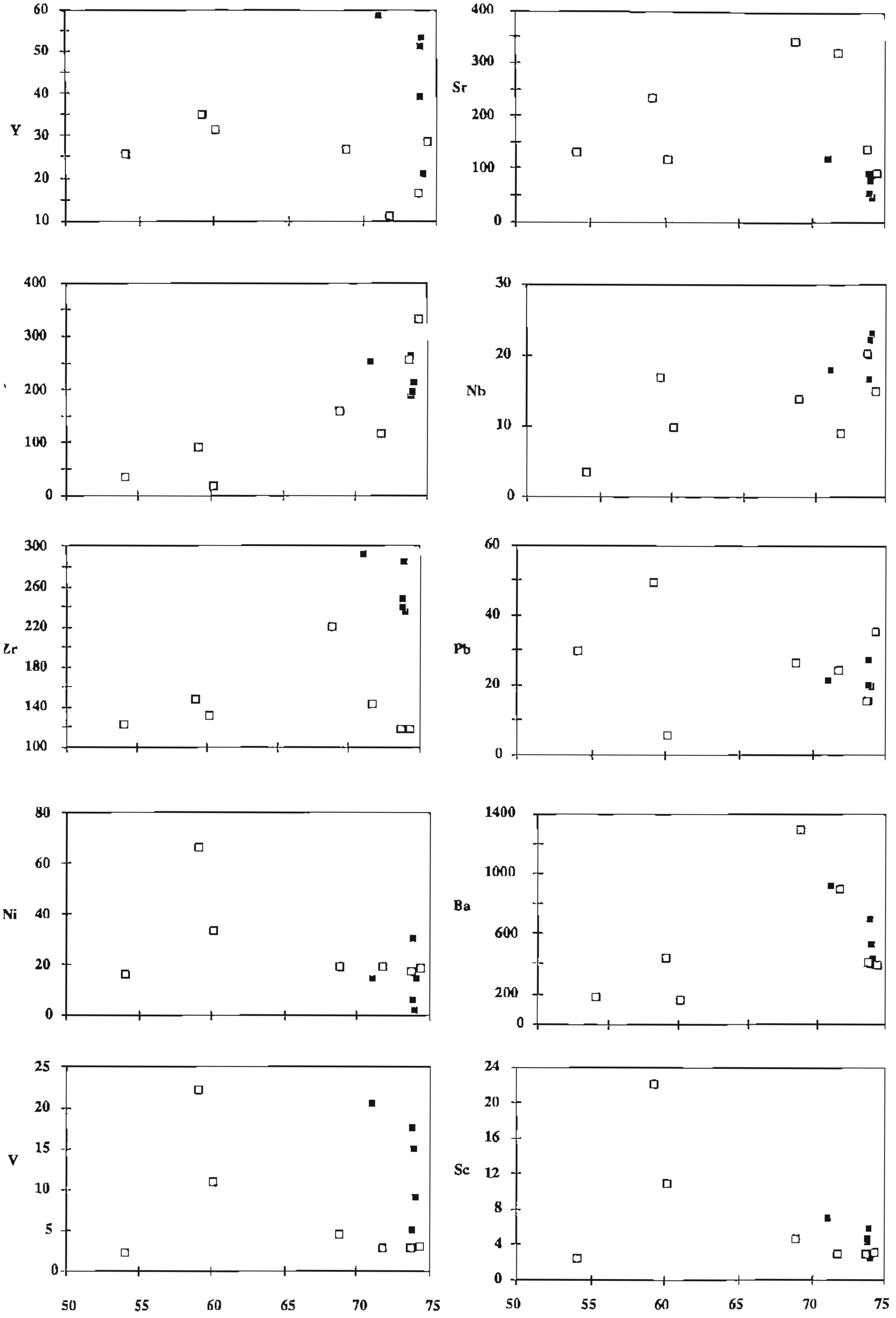


Fig. 4.7. Comparison of trace elements of volcanic erratics in the Permian sediments and volcanic rock samples of Mt Monster areas.

- □ Mt Monstervolcanic
- ■ volcanic of Permian erratics

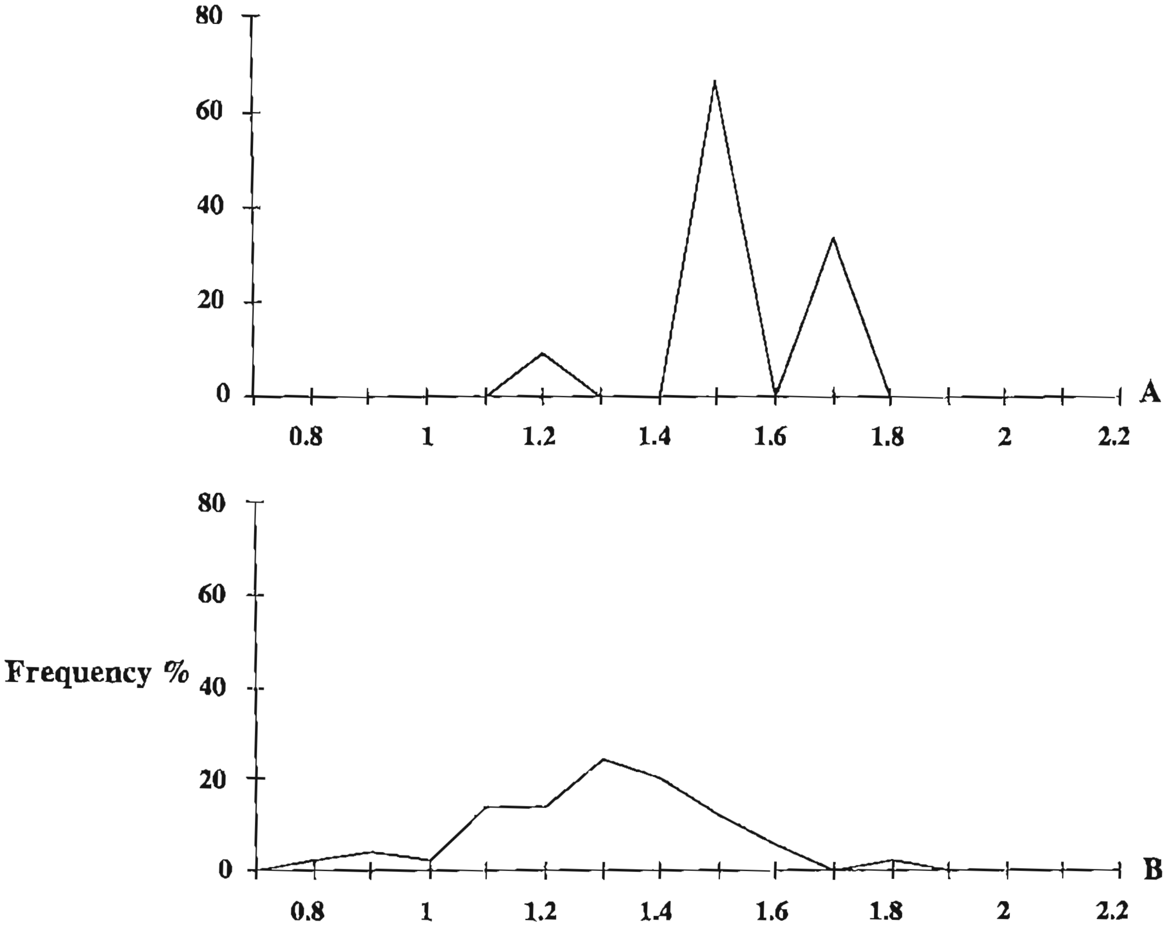


Fig. 4.8. Percentage of hafnium content in zircon grains of Encounter Bay Granite and Kanmantoo Group metasediments rock samples:

A: Encounter Bay

B: Kanmantoo Group

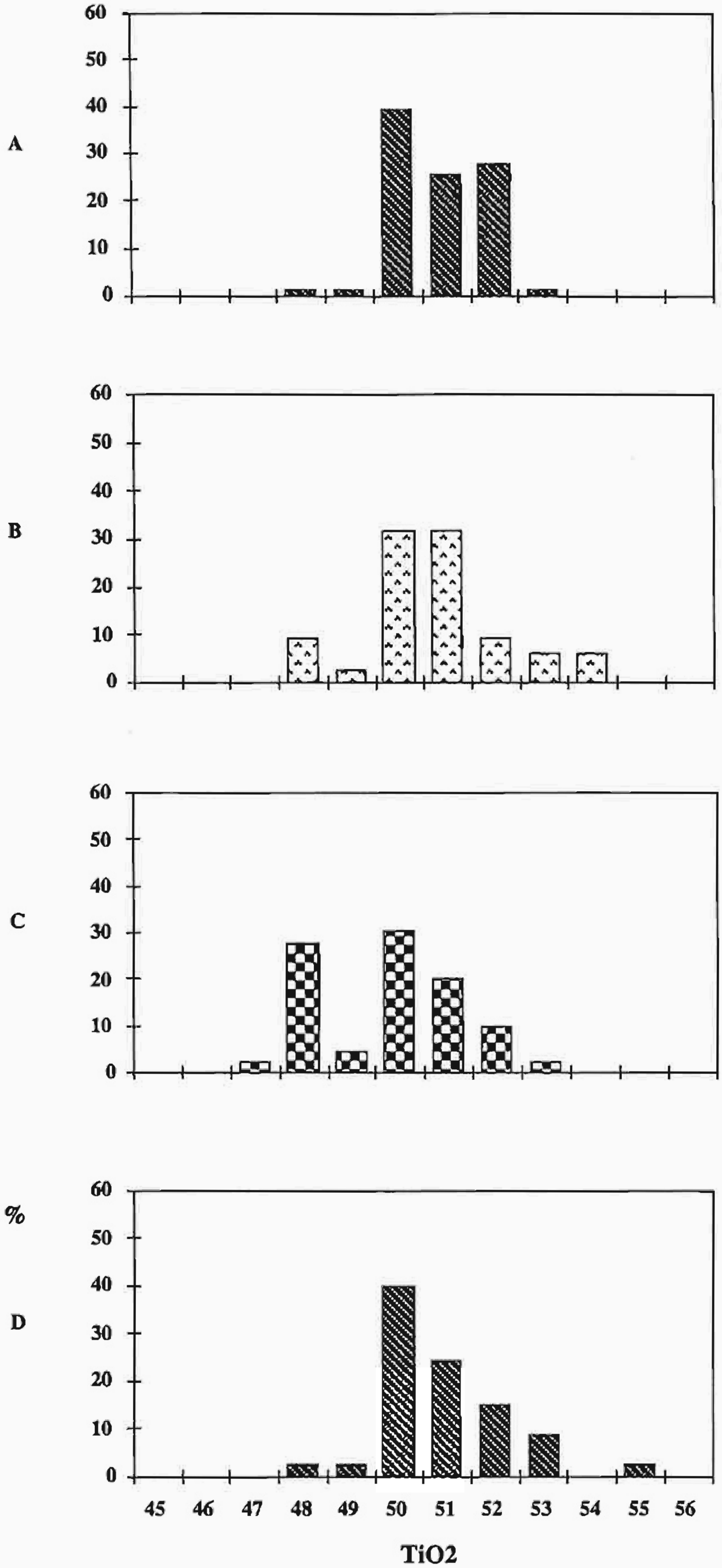


Fig. 4.9. Concentration of TiO2 in the ilmenite grains from Kanmantoo Group and Permian erratics.
A: Kanmantoo Group metasediments
B: Permian volcanic erratics
C: Permian granitic erratics
D: Permian gneisseic erratics

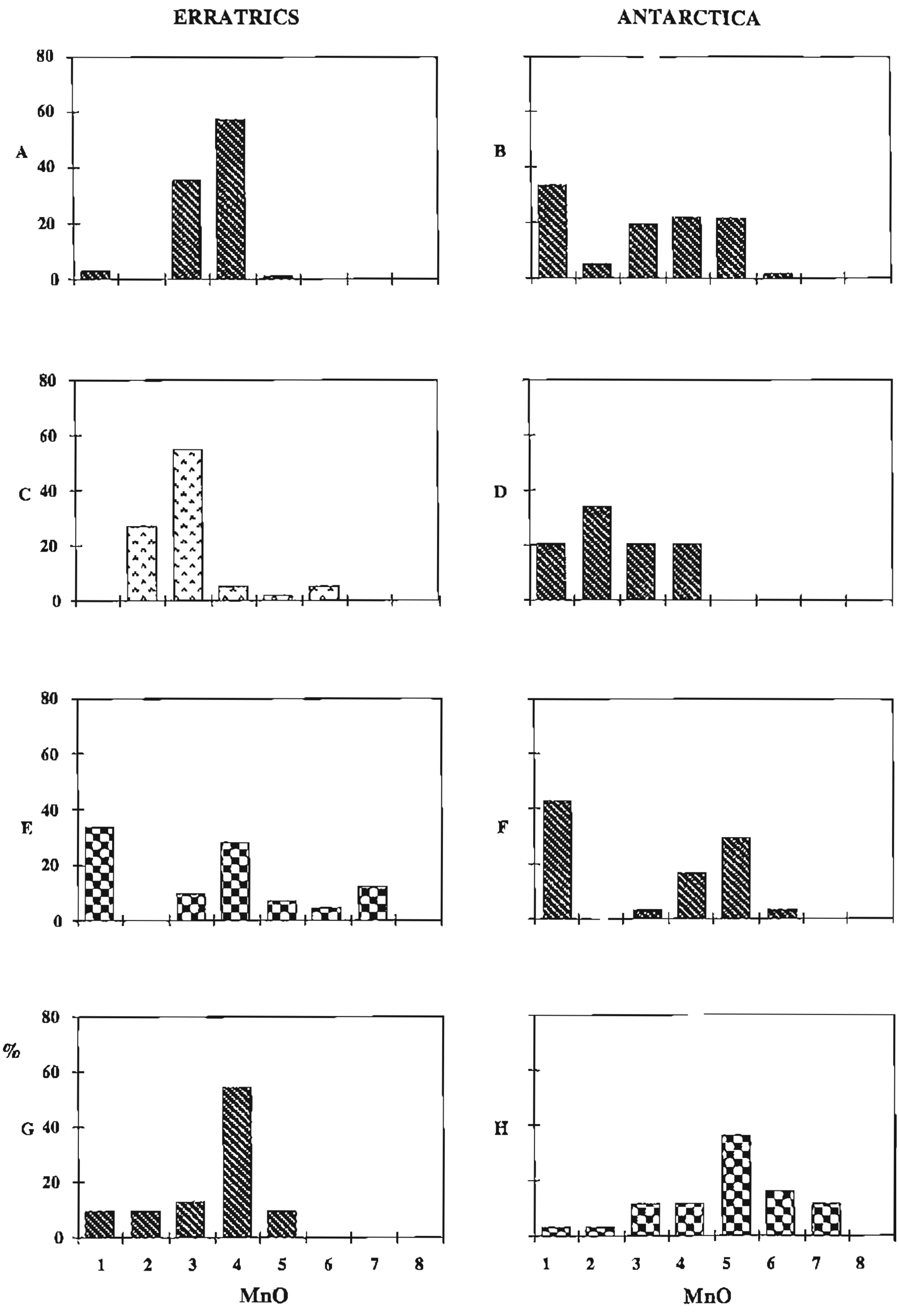


Fig. 4.10. Concentration of MnO in the ilmenite grains from Kanmantoo Group, Permian erratics and the rock samples of East Antarctica.
A: Kanmantoo Group
B: granitic gneisses of Antarctica
C: Permian volcanic erratics
D: basic gneisses of Antarctica
E: Permian granitic erratics
F: gneisses of Antarctica
G: Permian gneissic erratics
H: granites of Antarctica

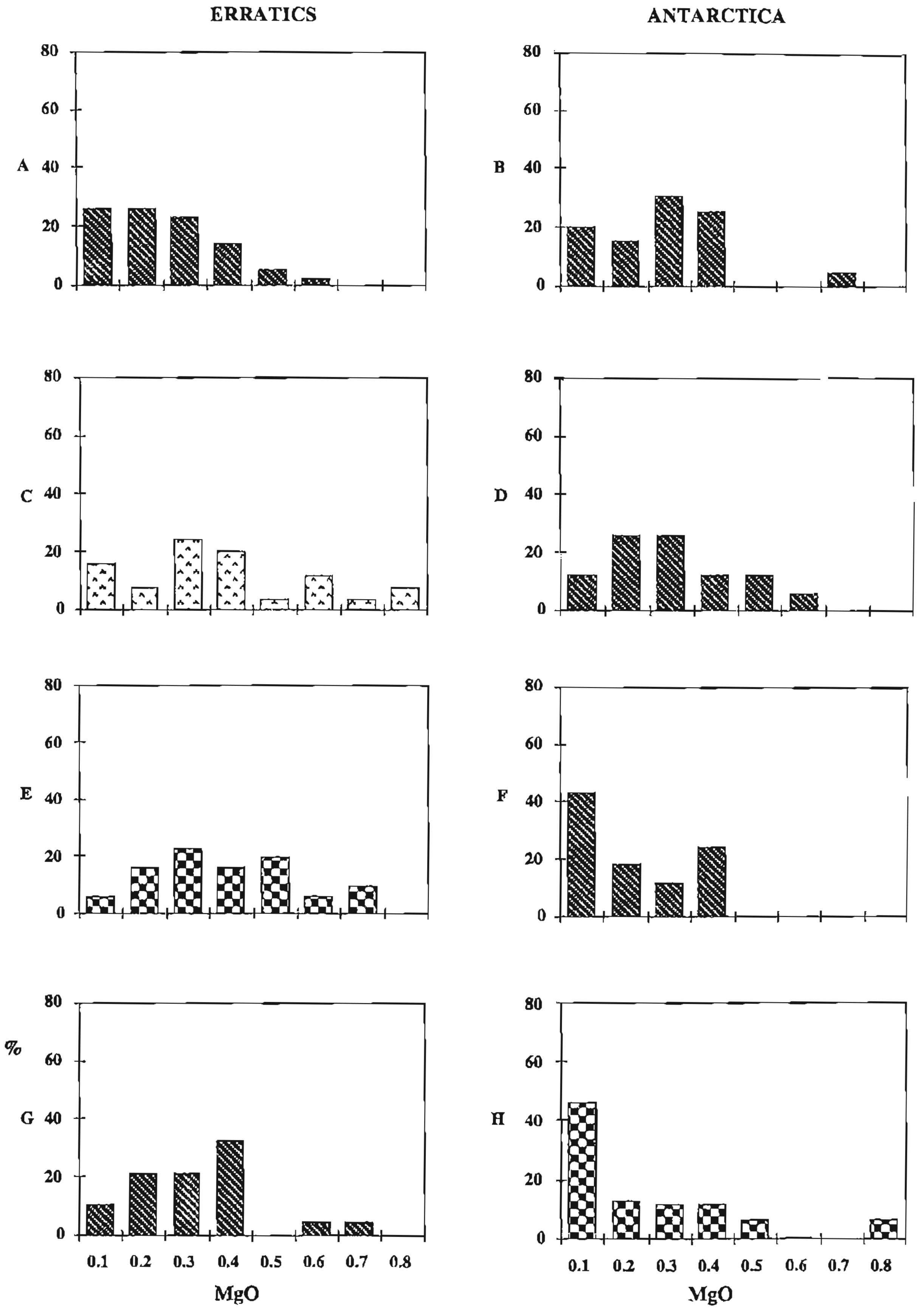


Fig. 4.11. Concentration of MgO in the ilmenite grains from Kanmantoo Group, Permian erratics and the rock samples of East Antarctica.
A: Kanmantoo Group metasediments
B: granitic gneisses of Antarctica
C: Permian volcanic erratics
D: basic gneisses of Antarctica
E: Permian granitic erratics
F: gneisses of Antarctica
G: Permian gneisses erratics
H: granites of Antarctica

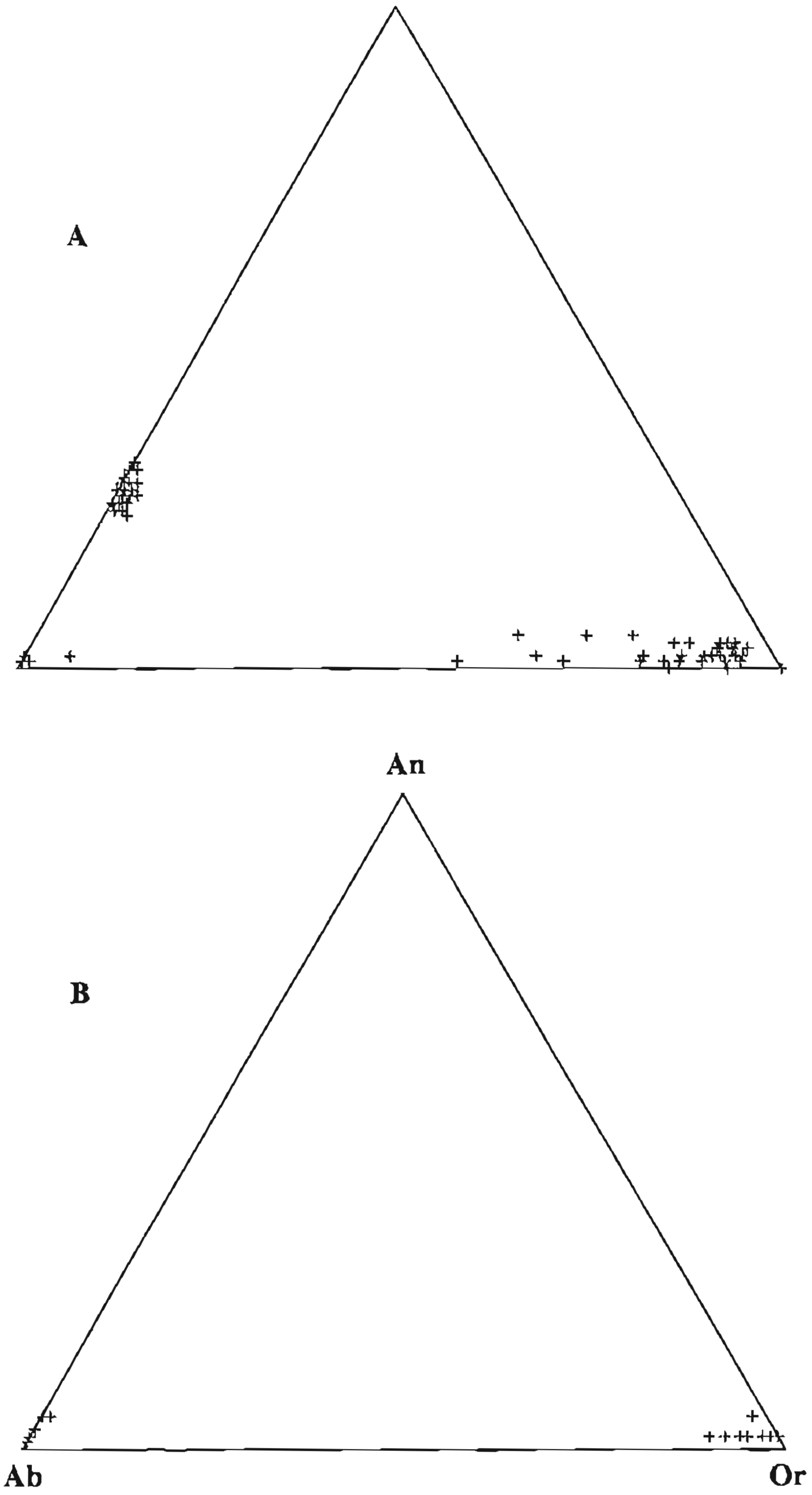


Fig. 4.12. Composition of feldspar grains (weight %) in the gneisses (A) and granites (B) rock samples of East Antarctica as determine by electron microprobe analyses displayed on a triangular diagram: An (anorthite), Ab (albite) and Or (orthoclase) on the base of 32 (O).

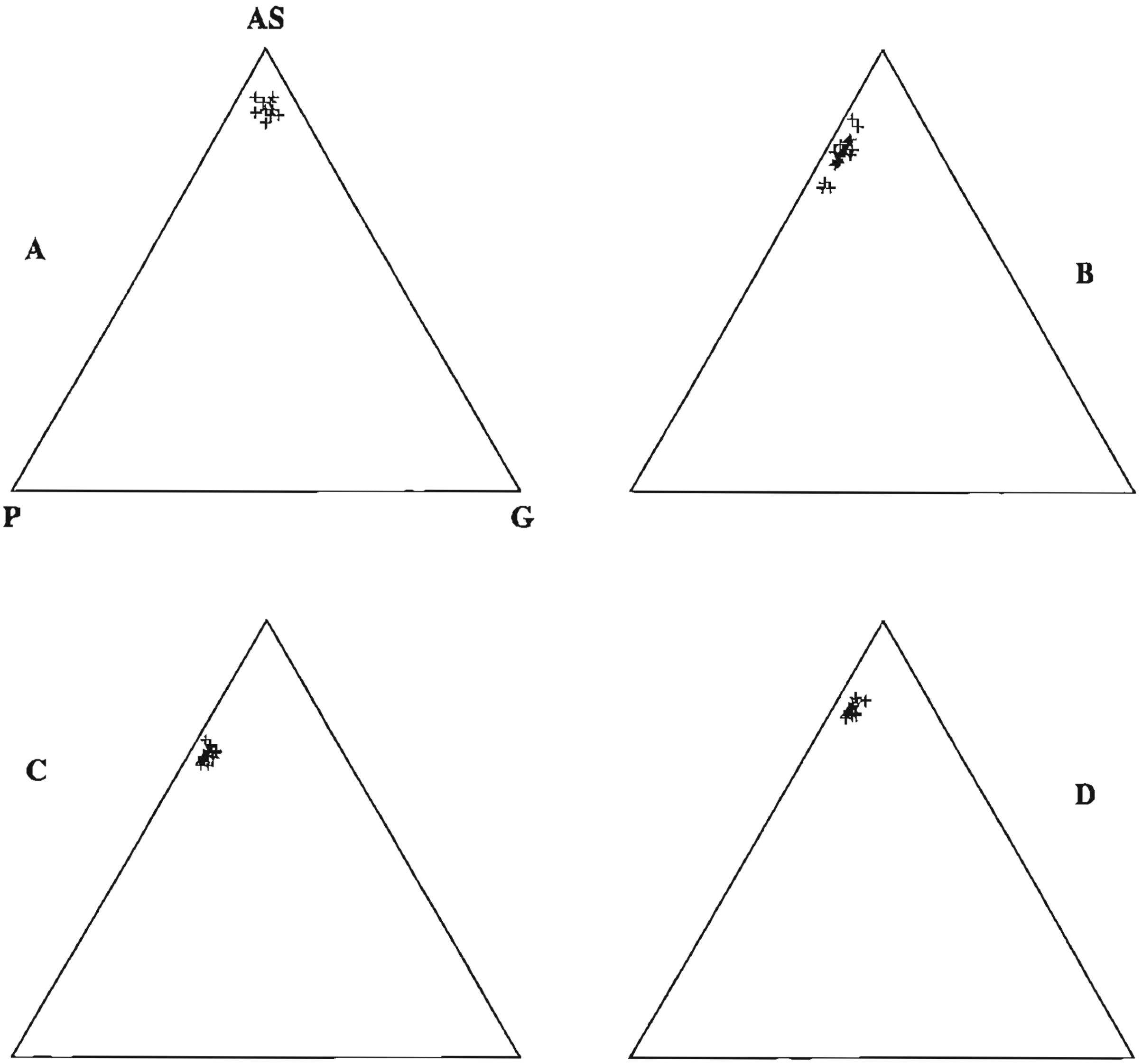


Fig. 4.13. Garnet geochemistry in the rock samples of East Antarctica displayed on a triangular diagram: AS= Almandine+ spessartine; P= Pyrope; G= Grossular.
A: phyllites of Commonwealth Bay; B: granitic gneisses of Windmill Island; C: gneisses of Windmill Island; D: granites of Windmill Island.

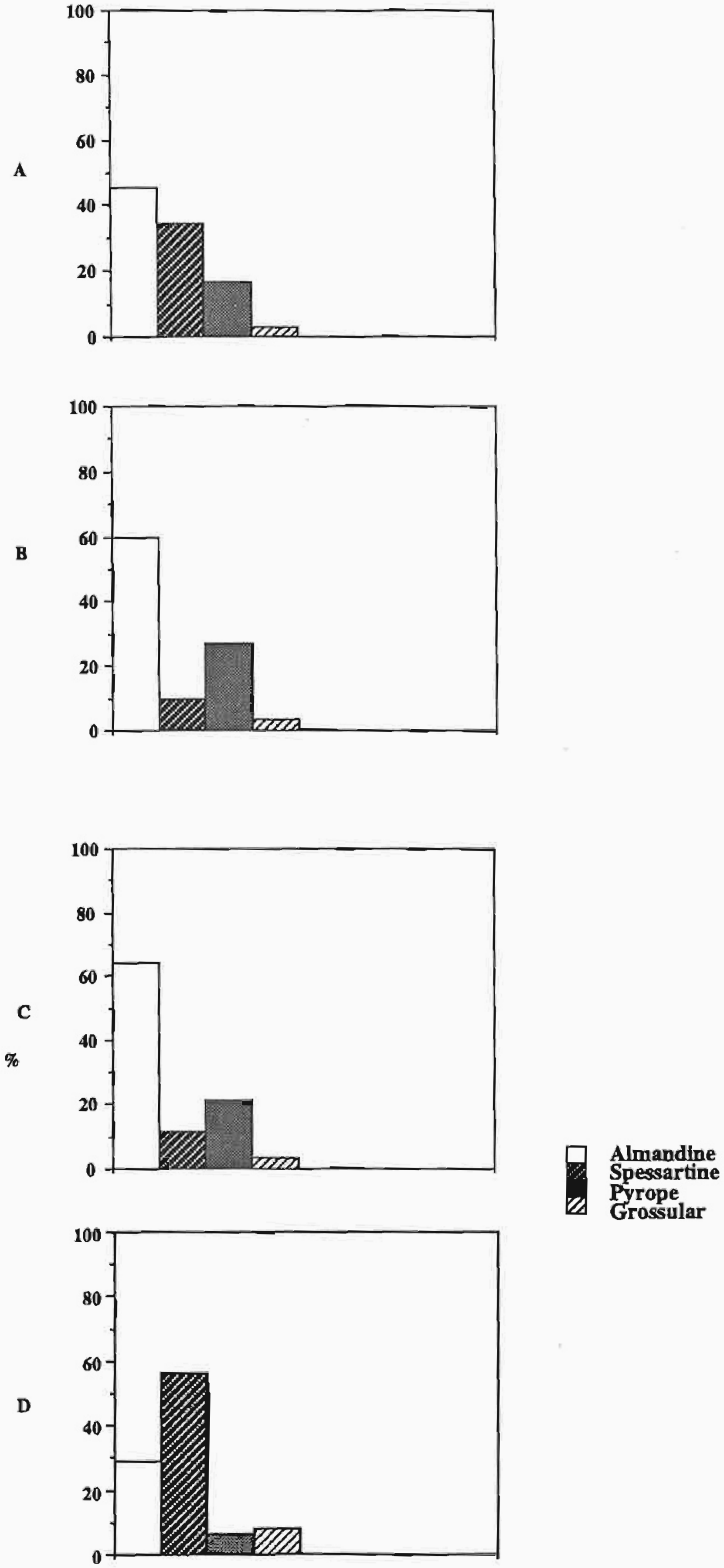


Fig. 4.14. Frequency distribution of garnet group in the rock samples of East Antarctica.
A: granites
B: gneisses
C: granitic gneisses
D: phyllites

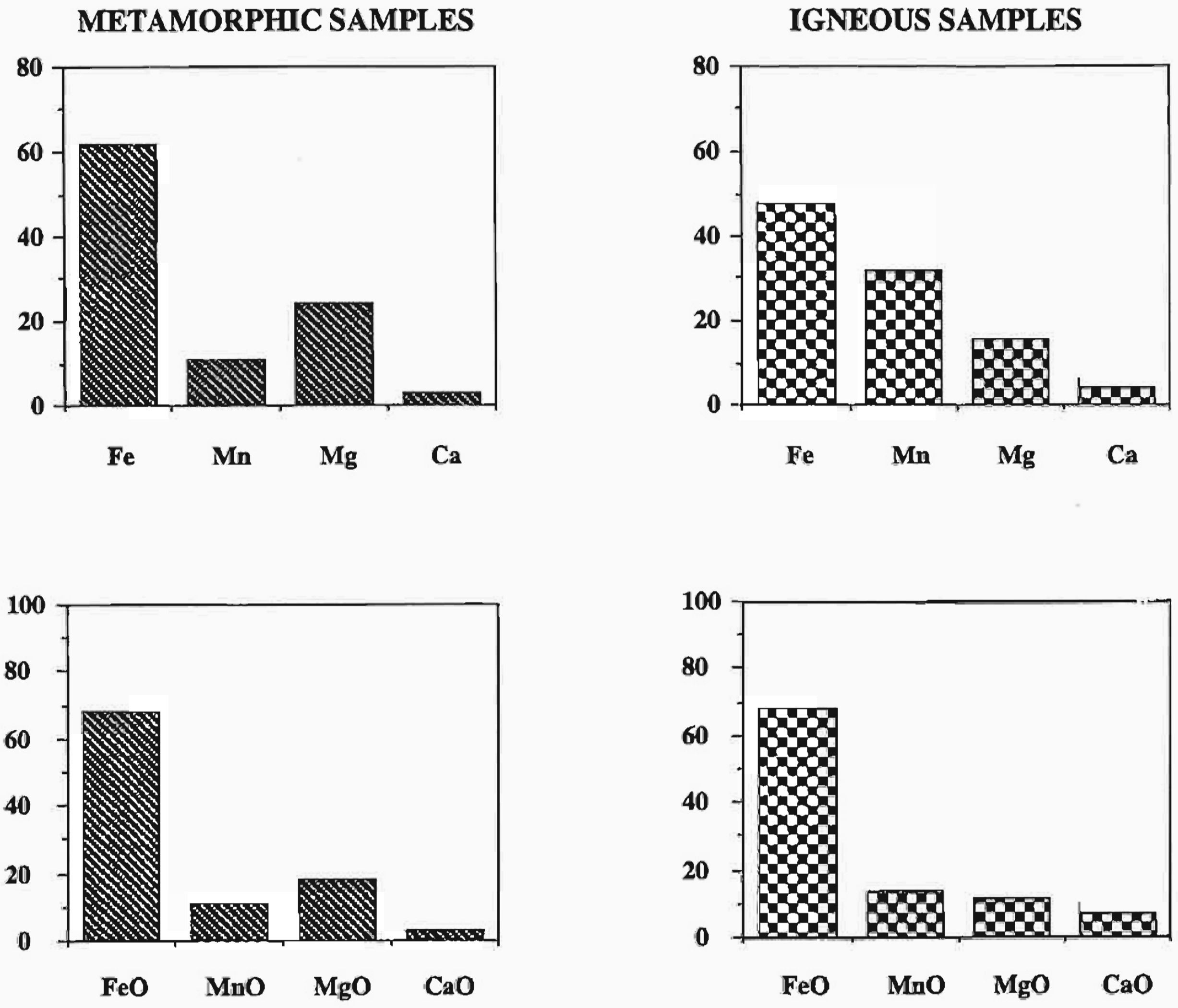


Fig. 4.15. Weight percentage of Fe, Mn, Mg and Ca elements and oxides on the base of 24 (O) in the garnet grains of East Antarctica rocks.

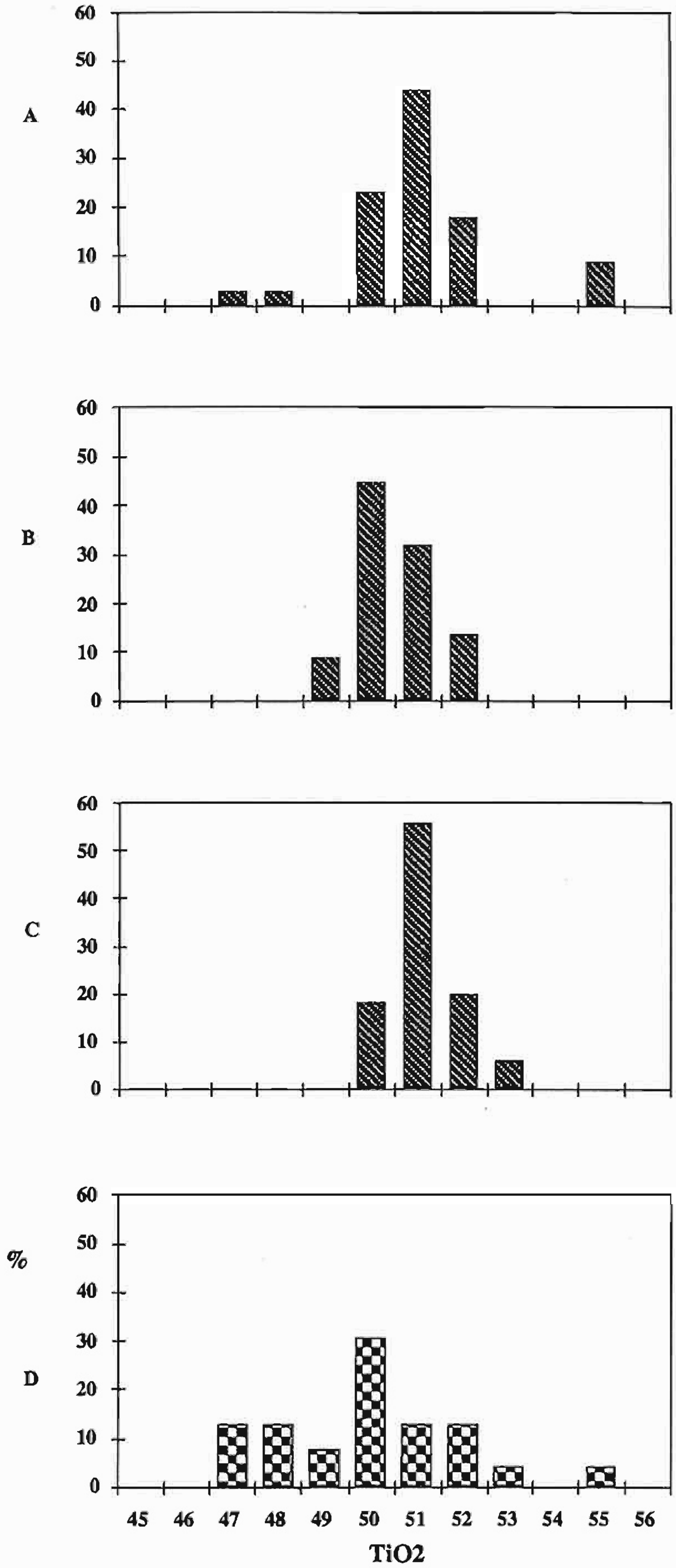


Fig. 4.16. Concentration of TiO2 in the ilmenite grains from the rock samples of East Antarctica.
A: granitic gneisses
B: Basic gneisses
C: gneisses
D: granites

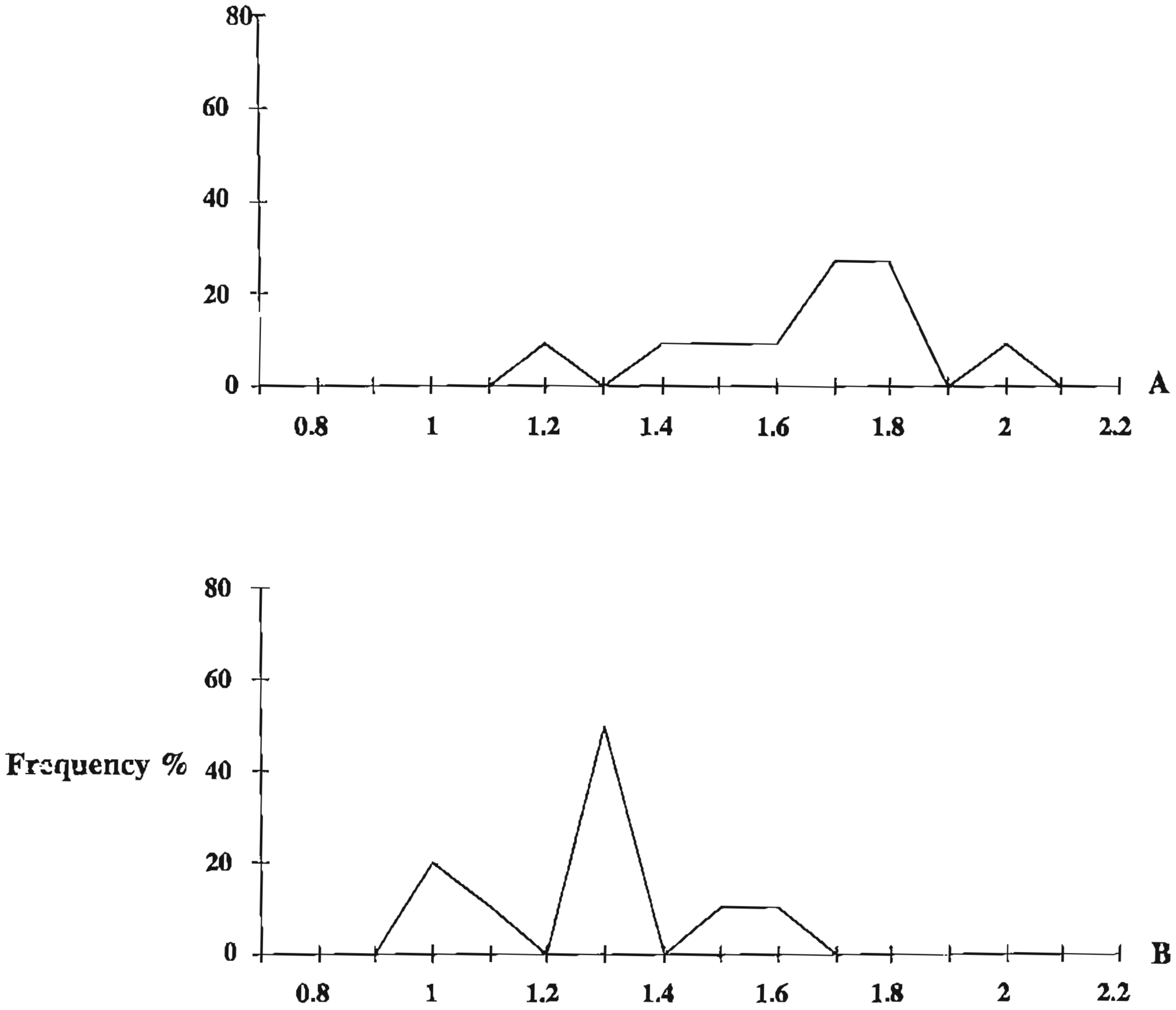


Fig. 4.17. Percentage of hafnium content in zircon grains from East Antarct rock samples.
A: gneisses
B: granites

Table. 4.0. Composition of feldspar grains in Encounter Bay Granite of southern Australia.

SiO2	Al2O3	K2O	Na2O	CaO	MgO	MnO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
65.6	19.4	8.1	5.6	0.7	0	0	0	0	0	99.2	3	50	47
65.5	19.2	11.6	3.6	0.4	0	0	0	0	0	100.3	1	32	67
65.4	19.3	5.1	7.9	0.5	0	0	0	0	0	98.3	2	68	30
65.4	22.3	0.3	10.2	2.1	0	0.3	0.2	0.2	0.1	100.7	10	88	2
65.3	19.1	13.4	2.1	0.4	0	0	0	0.1	0.1	100.5	2	19	79
64.7	19.2	15	0.8	0.6	0	0	0	0	0.1	99.1	3	7	90
64.5	18.6	15.8	0.6	0.7	0	0	0	0.2	0	100.3	3	5	92
64.1	18.9	13.5	2.6	0.6	0	0.1	0	0	0.3	100.1	3	22	75
64.1	18.1	14.8	1.1	0.6	0	0	0	0.1	0.1	98.89	2	10	88
64.1	23.1	0.1	9.1	3.8	0	0	0	0	0.2	100.2	19	80	1
63.9	18.1	14.1	1.4	0.7	0	0.2	0	0.3	0	98.8	2	13	85
63.9	18.8	13.1	2.5	0.7	0.2	0.1	0	0.3	0	99.6	2	21	77
63.9	18.3	14.3	1.3	0.6	0	0.1	0	0.1	0	98.1	2	12	86
63.8	18.1	14.9	0.7	0.9	0	0	0	0.4	0.3	99.2	4	6	90
63.8	18.7	15.5	1.3	0.8	0	0	0	0.5	0.4	100.9	2	10	88
63.8	23.2	0.1	9.7	3.5	0	0.1	0	0	0	100.2	16	83	1
63.7	18.4	13.8	1.7	0.4	0	0	0	0	0.2	98.5	2	15	83
63.7	18.8	14.6	0.7	0.6	0.2	0	0	0.3	0	99.1	3	5	92
63.7	23.1	0.7	8.7	2.7	0	0	0.1	0	0	99.9	14	85	1
63.6	18.4	13.1	1.5	0.7	0	0.2	0	0.4	0.1	98.2	3	16	81
63.5	19.1	14.1	1.9	0.6	0.2	0.1	0	0.4	0	99.9	3	16	81
63.5	18.9	14.6	1.7	0.5	0.1	0	0	0	0	99.5	2	15	83
63.5	18.7	15.1	1.2	0.4	0	0	0	0.1	0	99.7	2	10	88
63.2	18.4	14.2	1.4	0.6	0.3	0.2	0	0.4	0.2	98.9	2	12	86
63.2	18.3	13.9	1.7	0.8	0	0	0	0.7	0	98.6	2	15	83
63.2	18.2	16 .5	0	0.4	0	0	0	0	0	98.4	2	0	98
63.1	18.4	13.5	2.1	0.8	0.1	0	0	0.2	0.2	98.4	5	19	76
63.1	18.7	14.4	1.1	0.3	0.2	0.1	0	0.5	0	98.3	1	10	89
63.1	18.5	14.2	1.5	0.6	0	0	0	0	0.2	98.2	3	13	84
63.1	18.7	14.2	1.3	0.4	0.3	0	0	0	0.1	98.2	2	12	86
63.1	18.1	15.4	0.9	0.7	0	0	0	0.4	0	98.8	2	8	90
63.1	24.1	0	8.4	5	0	0.2	0.3	0.3	0	101.2	23	70	7
62.7	18.7	15.8	0.4	0.7	0.3	0.1	0	0.2	0	98.8	3	4	93
62.7	18.6	14.3	1.1	0.6	0.2	0	0	0.6	0	98.1	3	10	87
62.6	18.4	14.5	1.1	0.6	0	0	0	0.6	0.3	98.3	2	10	88
62.6	18.7	15.7	0.1	0.7	0	0	0	0.3	0.2	98.4	2	1	97
62.6	24.3	0.2	8.7	4.9	0	0	0.2	0	0.1	100.7	23	76	1
62.5	18.3	15.5	1.2	0.7	0	0	0	0.2	0	99.1	2	9	89
62.3	18.7	14.2	1.7	0.5	0.5	0	0	0.6	0.1	98.7	2	14	86
62.1	18.4	15.9	0.7	0.6	0	0	0	0.4	0	98.2	3	7	90
59.6	25.1	0	7.4	6.9	0	0.1	0.2	0.2	0	99.6	33	67	0
59.1	25.3	0.2	7.2	7.2	0.3	0.2	0	0	0.2	99.5	37	62	1
59.1	26.1	0.2	7.5	7	0	0.1	0.1	0.1	0	99.8	34	65	1
58.9	25.5	0	7.6	7.1	0	0	0	0.1	0.1	99.4	35	65	0
58.4	25.9	0	7.3	7.3	0	0.2	0.1	0.3	0.2	99.9	35	65	0
58.2	25.9	0.2	7.5	7.7	0	0	0.2	0	0.4	100.1	36	63	1
58.2	25.7	0	7.4	7.3	0.3	0.2	0	0	0	99.1	35	65	0
65.6	19.4	8.1	5.6	0.7	0	0	0	0	0	99.2	3	50	47
65.5	19.2	11.6	3.6	0.4	0	0	0	0	0	100.3	1	32	67
65.4	22.3	0.3	10.2	2.1	0	0.3	0.2	0.2	0.1	100.7	10	88	2
65.3	19.2	13.4	2.1	0.4	0	0	0	0.1	0.1	100.5	2	19	79
64.7	19.1	15.1	0.8	0.6	0	0	0	0	0.1	100	3	7	90
64.5	18.6	15.8	0.6	0.7	0	0	0	0.2	0	100.3	3	5	92
64.1	23.1	0.1	9.1	3.8	0	0	0	0	0.2	100.2	19	80	1
63.9	18.3	14.3	1.3	0.6	0	0.1	0	0.1	0	98.1	2	12	86
63.8	23.2	0.1	9.7	3.5	0	0.1	0	0	0	100.2	16	83	1
63.7	23.1	0.7	8.7	2.7	0	0	0.1	0	0	99.9	14	85	1
63.2	18.2	16 .5	0	0.4	0	0	0	0	0	98.4	2	0	98
63.1	24.2	0	8.4	5	0	0.2	0.3	0.3	0	101.2	23	70	7
62.6	24.1	0.2	8.7	4.9	0	0	0.2	0	0.1	100.7	23	76	1
59.1	26.1	0.2	7.5	7	0	0.1	0.1	0.1	0	99.8	34	65	1

Table. 4.1. Composition of garnet grains in the rock samples in the Encounter Bay of southern Australia.

SiO2	36.4	36.8	36.4	36.7	36.6	36.4	36.7	36.4	36.5	38.9	37.8	37.8	38.1
TiO2	0.1	0.1	0.3	0.3	0.1	0.2	0.1	0.1	0	0.4	0.3	0	0
Al2O3	21.2	21.2	21.2	22.1	21.6	21.6	21.5	21.6	21.4	21.4	22.4	22.8	22.8
FeO	20.1	23.3	19.1	18.5	23.5	22.2	23.1	27.5	27.4	25.3	27.3	27.2	26.5
MnO	12.5	5.1	14.2	16.3	10.4	10.6	10.2	7.4	7.8	3.7	5.1	3.1	4.4
MgO	1.2	1.6	1.1	1.1	1.7	1.5	1.6	2.2	2.1	7.2	6.7	7.7	7.1
CaO	8.1	5.1	9.1	7.9	6.6	7.2	6.5	4.1	4.5	1.6	1.5	1.5	1.5
Na2O	0	0.1	0	0.1	0	0.1	0.1	0.1	0.1	0	0	0	0
K2O	0	0	0	0.1	0	0	0	0.1	0	0	0	0	0
Cr2O3	0	0.1	0.1	0	0.1	0	0.1	0	0	0.6	0	0	0.4
Total	99.8	100.4	100.7	101.7	100.6	99.8	99.6	99.7	99.4	99.1	101.1	100.1	100.9

No. of ions on the basis of 24 (O)

Si	5.9	5.9	5.8	5.8	5.9	5.8	5.8	5.8	5.8	5.9	5.9	5.9	5.9
Ti	0.01	0.01	0.03	0.03	0.01	0.02	0.01	0.01	0	0.05	0.04	0	0
Al	4	4	3.8	4	4	4	4	4	4	4	4.1	4.1	4.1
Fe	2.7	3.1	2.2	2.1	3.1	3.1	3.1	3.6	3.6	3.3	3.5	3.6	3.4
Mn	1.7	1.6	2.4	2.4	0.9	0.9	0.9	0.8	0.8	0.5	0.7	0.5	0.6
Mg	0.3	0.4	0.4	0.3	0.5	0.4	0.4	0.9	0.8	1.6	1.6	1.7	1.6
Ca	1.4	1.4	1.6	1.4	1.4	1.4	1.4	0.7	0.6	0.2	0.2	0.2	0.2
Cr	0	0.01	0.01	0	0.01	0	0.01	0	0	0.05	0	0	0

Mol.% end-members

Al	44	48	33	34	52	53	53	60	62	59	58	59	59
S	28	25	36	39	15	16	16	13	14	9	12	8	10
P	5	6	6	5	9	7	7	15	14	28	27	29	28
G	23	21	25	22	24	24	24	12	10	3	3	3	3

Table. 4.2. Composition of zircon grains from rock samples of Encounter Bay and Permian erratics of southern Australia.

	*	*	**	**	***	***
SiO2	29.2	28.1	29.5	29.6	28.9	28.2
ZrO2	67.2	67.3	69.5	70.5	67.5	67.4
Al2O3	0	0.1	0	0	0	0.1
FeO	0	0.3	0.6	0.5	0.1	0.2
MnO	0	0	0	0	0	0
MgO	0	0	0	0	0	0
Hf2O3	1.5	1.7	1.4	1.3	1.3	1.5
Th2O3	0.1	0.1	0.1	0.1	0.1	0.2
Total	97.9	97.8	101.1	101.9	97.9	97.4

Structure formula on base of 16 (O)

Si	3.7	3.7	3.6	3.7	3.6	3.5
Zr	4.2	4.2	4.2	4.3	4.3	4.3
Al	0	0.01	0	0	0	0.01
Mg	0	0	0	0	0	0
Hf	0.05	0.06	0.05	0.05	0.05	0.06
Fe	0	0.04	0.07	0.05	0.02	
Mn	0	0	0	0	0	0

*	Encounter Bay
**	Granitic gneisses of Permian erratics
***	Volcanic of Permian erratics

Table. 4.3. Composition of ilmenite grains in the rock samples of Encounter Bay.

TiO2	FeO	MnO	MgO	Al2O3	SiO2	CaO	Na2O	K2O	Cr2O3	Total
52.6	42.1	4.9	0.6	0.4	0	0.1	0	0	0	100.8
51.6	40.3	6.2	0.7	0	0.2	0	0	0	0.2	99.9
51.4	40.1	6.4	0.2	0.2	0	0	0	0	0	99.3
50.9	41.7	7.2	0.5	0.2	0.1	0	0.4	0	0	101.3
50.9	41.7	7.2	0.5	0.2	0.1	0	0.4	0	0	101.3
50.2	43.1	5.5	0.4	0.2	0	0.1	0.2	0.1	0.3	100.2
48.8	48.4	0.7	0	0	0	0	0	0	0.1	98.2
48.8	48.4	0.7	0	0	0	0	0	0	0.1	98.2
48.7	45.1	1.5	0.3	0	0.2	0	0.4	0	0	98.2
48.2	46.9	1.9	0.7	0.3	0.3	0	0.4	0	0	99.6
47.8	48.9	1.2	0	0	0.1	0	0	0	0	98.2
47.8	48.9	1.2	0	0	0.1	0	0	0	0	98.2

Table. 4.4. Composition of feldspar grains in the Mt Monster porphyry rock samples of southern Australia.

SiO2	Al2O3	K2O	Na2O	CaO	FeO	MgO	MnO	TiO2	Cr2O3	Total	An	Ab	Or
70.3	17.4	6.1	3.8	0.5	0.8	0	0.2	0	0	99.3	4	46	50
69.9	14.7	9.7	2.2	0.9	0	0	0	0.4	0	99.1	6	25	69
68.9	18.7	0	10.3	0.9	0	0	0	0.3	0	99.2	5	95	0
68.4	17.6	6.6	5.7	0.3	0.1	0.5	0	0.3	0	99.4	2	56	42
67.7	21.1	0.1	10.3	1.1	0	0	0	0.3	0.2	100.9	5	94	1
66.4	19.9	0.5	10.5	1.2	0	0	0	0	0.3	98.9	5	92	3
66.4	20.9	0	10.8	1.6	0	0.1	0.1	0	0	99.9	7	93	0
65.2	20.2	6.7	7.1	0.4	0	0	0.1	0.2	0	99.9	2	61	37
65.2	19.7	0.4	11.4	1.1	0.3	0	0.1	0	0.2	98.5	5	93	2
65.2	20.7	0	10.6	1.1	0.3	0.1	0	0	0.2	98.3	5	95	0
64.4	20.9	0.2	11.1	1.9	0.1	0	0	0	0	98.6	9	90	1
63.2	18.2	16.6	0.1	0.5	0	0.3	0.3	0.4	0	99.7	3	1	96
63.1	18.5	16.7	0	0.7	0	0.5	0	0.1	0.1	99.7	1	0	99
63.1	18.2	16.9	0.1	0.3	0	0.1	0.2	0	0.1	98.9	2	1	97
62.9	17.9	15.2	1.5	0.9	0	0	0.2	0.1	0	98.8	5	12	83
62.9	18.2	16.2	0.3	0.5	0	0.1	0	0.3	0	98.6	2	2	96
62.5	17.5	16.8	0	0.7	0	0.3	0	0.2	0	98.1	2	0	98
62.5	18.5	16.4	0.1	1.2	0	0.1	0	0.2	0	99.1	5	1	94
62.4	18.1	16.8	0	0.2	0	0.1	0.1	0.2	0.2	98.2	1	0	99
62.1	18.1	16.9	0	0.6	0	0	0	0.6	0.5	98.8	2	0	98
61.9	18.2	15.9	0.7	0.6	0	0.1	0.3	0.8	0	98.7	2	7	91
61.2	20.1	15.1	1.1	0.4	0	0	0.3	0.1	0	98.3	2	10	88

Table. 4.5. Chemical composition and trace elements in the Mt Monster volcanic rocks.

SiO2	74.1	71.1	73.9	68.9
Al2O3	13.2	14.5	12.9	15.7
Fe2O3	2	2.3	1.9	3.4
MnO	0	0	0	0.1
MgO	0.1	0.2	0.3	0.4
CaO	0.2	0.3	1	1.9
Na2O	3.2	3.5	2.9	3.8
K2O	5.6	6	5.3	5.1
TiO2	0.2	0.4	0.3	0.3
P2O5	0	0.1	0.1	0.1
SO3	0	0	0	0
LOI	0.8	0.7	0.3	0.4
Total	99.4	99.2	99.1	99.9
Y	20.8	58.5	39	26.4
Sr	43.8	117.2	89.5	341.9
Rb	213.1	251.2	264	158.4
Nb	23	17.8	16.6	13.7
Zr	233.6	291.9	238.6	220.2
Pb	19.4	21.3	26.9	26.3
Th	21.3	26.9	26	17
U	6	6.2	5	6.7
Ga	19.4	16.5	14.5	20.6
Cu	6	12	6	14
Zn	43	37	37	68
Ni	14	14	30	19
Ba	425	914	696	1284
Sc	2.5	7	4.2	4.5
V	9.1	20.4	17.6	11.4
Co	0.8	2.7	1.6	2.7
Ce	28	173	109	103
Nd	15	102	49	45
La	13	84	55	56

Table. 4.6. Composition of feldspar grains from Kanmantoo Group metasediments rock samples of southern Australia.

SiO2	57.8	60.2	59.9	58.3
TiO2	0	0	0	0
Al2O3	25.6	25.2	25.1	24.9
FeO	0.3	0	0.3	0.1
MnO	0	0	0	0.4
MgO	0.4	0	0	0
CaO	6.8	6.7	5.3	7.1
Na2O	6.9	7.1	8.1	7.4
K2O	0.1	0.3	0.1	0.1
Cr2O3	0.4	0.4	0.2	0
Total	98.3	99.8	99.1	98.3
		Structure formula on the base of 32 (O)		
Si	10.5	107	10.8	10.6
Al	5.5	5.4	5.4	5.4
Ca	1.3	1.3	1.1	1.4
Na	2.4	2.4	2.9	2.7
K	0	0	0.01	0.01
An	35	35	27	34
Ab	64	65	73	66
Or	0	0	0	0

Table. 4.7. Composition of garnet grains in Kanmantoo Group metasediments of southern Australia.

SiO2	36.2	36.4	36	36.4	36.5	35.2	35.6	36	37.9	36.4	37.5	36.6
TiO2	0	0	0	0.2	0	0	0.2	0	0	0.2	0	0
Al2O3	21.4	21.8	21.7	20.8	21.6	20.7	21.4	21.5	22.2	21.1	22.1	21.7
FeO	34.9	34.9	36	33.9	34.2	34.2	35.6	34.8	27.9	25.9	23.4	28.8
MnO	4.3	4.4	3	5.2	4	4.3	3.1	4.6	2.3	10.1	0.6	8.9
MgO	2.4	2.7	2.7	3	2.4	2.3	2.9	2.8	7.8	2.5	6.4	3.1
CaO	0.5	0.3	0.5	0.5	0.6	0.9	0.8	0.7	1.1	2.4	10.9	1.7
Na2O	0.4	0.1	0	0.5	0	0	0.3	0.1	0	0.4	0	0.4
K2O	0	0	0	0.1	0.2	0	0	0	0	0	0	0
Cr2O3	0.3	0	0.1	0.1	0.6	0	0.1	0.1	0.4	0.4	0.4	0
Total	100.6	101.1	100.6	100.8	100.3	98.6	100.1	100.8	99.4	99.5	100.8	100.8

No.of ions on the basis of 24 (O)

Si	5.9	5.9	5.8	5.9	5.9	5.8	5.8	5.9	5.9	5.9	5.8	5.9
Ti	0	0	0	0.02	0	0	0.02	0	0	0.03	0	0
Al	4	4.1	4	3.9	4	4	4.1	4	4.1	4	4	4
Fe	4.8	4.7	4.7	4.6	4.8	4.7	4.8	4.7	3.6	3.5	3	3.8
Mn	0.6	0.6	0.5	0.7	0.5	0.6	0.5	0.6	0.3	1.4	0.08	1.2
Mg	0.6	0.6	0.6	0.7	0.6	0.6	0.7	0.7	1.7	0.6	1.5	0.7
Ca	0.09	0.04	0.09	0.08	0.1	0.1	0.1	0.1	0.2	0.4	1.8	0.3
Cr	0.04	0	0.01	0.02	0.07	0	0.02	0.01	0.04	0.05	0.04	0

Mol.% end-members

Al	78.8	79.1	79.7	75.6	80	78.3	78.7	77
S	9.8	10.1	8.5	11.5	8.3	10	8.2	9.8
P	9.8	10.1	10.2	11.5	10	10	11.5	11.5
G	1.5	0.7	1.7	1.3	1.7	1.7	1.6	1.6

Table. 4.8. Composition of rutile grains in the Kanmantoo Group metasediments of southern Australia.

TiO2	FeO	MnO	MgO	Cr2O3	Al2O3	CaO	Na2O	K2O	SiO2	Total
97.5	0.6	0.4	0	0.2	0.1	0	0	0	0	98.8
98.5	0.7	0.4	0.4	0.8	0	0	0	0	0	100.8
97.8	0.7	0	0	0.3	0	0.1	0	0	0	98.9
97.5	0	0.3	0.1	0.2	0.2	0.3	0.6	0	0	99.2
96.1	0.6	0	0	0.3	0.1	0	0.5	0	0	97.6
97.8	0.7	0	0	0.9	0.1	0	0	0	0	99.5
97.6	0	0.6	0	0.7	0	0	0	0	0	98.9
96.2	0.9	0.3	0.1	0.1	0.4	0.2	0	0.1	0.4	98.7

Table. 4.9. Composition of ilmenite grains in Kanmantoo Group metasediments of southern Australia.

TiO2	FeO	MnO	MgO	Al2O3	Cr2O3	V2O3	SiO2	Total
53.2	40.3	4.1	0.5	0.1	0	0.1	0	98.3
52.7	41.5	4.1	0	0.1	0	0.3	0	99.1
52.5	40.3	4.7	0	0.1	0.1	0.2	0	98.2
52.4	41.3	4.7	0.3	0.1	0	0.1	0	100.1
52.3	43.1	4.6	0	0	0	0.3	0.3	100.6
52.3	40.3	4.7	0.3	0.1	0	0.2	0.2	98.4
52.2	44.1	3.4	0.3	0.2	0	0.2	0	100.2
52.2	40.8	4.6	0.3	0.1	0.2	0.1	0	98.7
52.2	40.8	4.6	0.3	0.1	0.2	0.2	0	98.7
52.1	42.1	4.2	0.4	0.2	0	0.2	0.3	100.3
52.1	41.3	4.2	0.4	0.1	0.2	0.2	0	98.8
52.1	41.1	4.6	0.2	0	0.1	0.2	0	98.1
52.1	41.1	4.1	0	0	0.1	0.2	0.3	98.1
51.8	44.1	4.1	0	0	0.1	0.2	0	100.4
51.7	43.8	3.8	0.6	0	0	0.3	0.1	100.3
51.7	43.8	3.6	0.1	0	0.1	0.2	0	99.7
51.7	39.8	5.6	1.1	0.2	0.4	0.3	0	100.4
51.5	40.1	3.8	0.5	0	0	0.2	0.8	98.2
51.4	43.3	3.7	0.1	0	0	0.3	0	98.6
51.3	43.1	4.1	0	0.3	0.2	0.2	0	99.1
51.2	44.1	3.7	0	0	0.1	0.2	0.1	99.4
51.2	42.4	3.9	0.2	0.2	0.1	0.3	0	98.9
51.2	45.1	3.9	0	0	0.2	0.3	0	100.1
51.1	44.2	4.1	0	0.1	0	0.2	0	99.7
51.1	45.1	3.7	0.1	0	0	0.2	0	100.2
51.1	41.8	4.8	0	0.4	0	0.3	0	98.7
50.9	43.6	4.4	0.1	0	0.2	0.3	0	100.1
50.9	43.4	3.4	0	0	0.2	0.3	0	98.5
50.8	42.8	4.5	0	0.2	0.2	0.3	0	99.1
50.8	42.7	3.5	0.2	0	0	0.2	0	98.5
50.8	43.7	4.1	0	0	0	0.2	0	99.3
50.6	43.8	3.8	0.3	0.2	0.2	0.2	0.4	99.4
50.6	43.8	4.1	0.2	0.1	0.1	0.3	0	98.8
50.5	45.1	1.5	0.1	0.1	0.1	0.2	0	98.4
50.4	44.1	4.1	0	0.2	0.3	0.3	0.1	99.5
50.4	47.6	0.5	0.4	0.4	0.3	0.3	0.2	100.2
50.3	44.8	4.1	0.2	0	0.4	0.3	0	99.8
50.2	44.4	4.2	0.2	0.1	0.4	0.2	0.1	100.1
50.2	43.8	3.6	0.2	0.2	0.1	0.3	0	98.1
50.2	44.9	4.1	0.3	0.1	0.1	0.3	0	99.9
50.1	43.1	3.4	0	0	0.1	0.3	0	97.8
50.1	45.1	3.5	0.1	0.1	0.1	0.3	0	99.3
50.1	44.8	3.6	0.1	0.2	0	0.3	0	99.4
50.1	43.8	3.8	0	0.6	0.3	0.2	0.2	99.5
50.1	44.2	3.8	0.1	0	0.1	0.3	0.2	98.1
50.1	41.1	4.6	0.2	0.1	0.4	0.2	0	97.8
49.8	42.3	4.4	0.5	0.3	0.6	0.2	0	98.7
49.8	43.8	4.1	0.2	0.2	0.2	0.3	0	99.7
48.7	43.6	4.5	0.4	0	0.5	0.2	0	98.4

Table. 4.10. Composition of staurolite in Kanmantoo Group metasediments of southern Australia.

Al_2O_3	54.3	54.1	54.1	54.1	53.3	53.2	53.1	52.9	53.1	54.3	54.3	53.5
SiO_2	28.5	28.8	27.1	26.9	28.7	27.1	27.6	27.5	27.6	27.5	27.9	26.7
FeO	13.5	14.6	14.2	15.9	14.9	14.8	15.4	14.1	14.2	15.1	14.5	15.1
MgO	1.8	1.4	1.4	1.1	1.3	1.3	1.5	1.3	1.3	0.8	1.4	1.7
MnO	0.2	0	0.4	0.1	0	0.4	0.5	0.6	0.3	0	0.1	0.2
TiO_2	1.1	0.5	0.3	0.4	0.4	0.4	0.5	0.6	0.4	0.3	0.6	0.4
CaO	0.2	0	0.2	0	0.2	0	0.1	0.1	0.3	0.2	0.1	0
Na_2O	0.5	0	0	0	0.7	0	0.1	0	0.2	0	0.2	0.1
K_2O	0	0	0	0	0	0	0	0	0	0	0	0
Cr_2O_3	0.3	0.5	0.1	0.4	0.4	0.3	0.5	0.1	0	0.2	0.1	0.4
Total	99.4	99.8	98.4	99.2	99.6	98.2	99.7	98.1	97.2	98.6	99.4	98.2

Structrural formula on base of 48(O)

Al	18.6	18.5	18.5	18.5	18.3	18.3	18.3	18.2	18.4	18.6	18.5	18.5
Si	8.1	8.2	7.9	7.8	8.1	7.9	7.9	7.9	7.8	7.8	7.8	7.6
Fe	3.4	3.6	3.5	3.7	3.7	3.7	3.6	3.5	3.5	3.7	3.5	3.7
Mg	0.9	0.7	0.7	3.3	0.4	0.4	0.7	0.4	0.5	0.4	0.6	0.7
Mn	0.02	0	0.05	0.03	0	0.05	0.05	0.05	0.08	0	0.03	0.04
Ti	0.2	0.1	0.04	0.08	0.08	0.08	0.1	0.1	0.09	0.08	0.1	0.09

Table. 4.11. Composition of apatite grains in Kanmantoo Group metasediments of southern Australia.

CaO	54.7	54.8	54.2	54.8	54.7	54.9	54.7	54.6	54.7	54.9	54.4	54.4
P2O5	41.3	41.9	42.3	42.7	42.1	41.4	41.4	42.6	42.4	41.9	42.5	40.5
TiO2	0.3	0.2	0.2	0.3	0.6	0.5	0.5	0.2	0.4	0.1	0	0
FeO	0.1	0	0.2	0.3	0.4	0.2	0.6	0.3	0.3	0.4	0.3	0.3
MnO	0	0	0.4	0	0	0.3	0.2	0.4	0.3	0.3	0.3	0.4
MgO	0	0	0	0	0	0.4	0.2	0	0	0.3	0.3	0.3
Na2O	0.2	0	0	0	0	0.4	0.3	0.3	0	0.2	0	0
K2O	0	0.1	0	0.1	0	0	0.2	0.1	0.2	0	0	0
SiO2	0	0	0	0	0	0	0	0	0	0	0	0.1
Cr2O3	0	0.1	0	0	0	0	0	0	0	0	0	0
Al2O3	0.1	0.2	0.1	0.2	0.2	0.3	0.2	0.1	0	0	0.1	0.1
Total	97.3	97.1	97.4	99.2	98.6	98.4	98.4	98.6	98.2	98.2	97.9	96.2

Structure formula on the base of 26 (O)

Ca	10.4	10.5	10.3	10.5	10.4	10.3	10.4	10.3	10.5	10.3	10.4	10.4
P	6.2	6.2	6.1	6.1	6.1	6.2	6.2	6.1	6.1	6.2	6.1	6.1
Fe	0.01	0	0.02	0.04	0.04	0.02	0.05	0.03	0.04	0.06	0.05	0.05
Mn	0	0	0.03	0	0	0.03	0.05	0.04	0.04	0.05	0.05	0.05
Mg	0	0	0	0	0	0.03	0.04	0	0	0.08	0.08	
Na	0.02	0	0	0	0	0.03	0.04	0.04	0	0.06	0	0

Table. 4.12. composition of sillimanite grains in the Kanmantoo Group metasediments of southern Australia.

SiO_2	37.6	36.6	36.2
TiO_2	0	0	0
Al_2O_3	60.9	61.9	61.9
FeO	0.8	0.8	0.1
MnO	0	0	0.1
MgO	0	0	0.5
CaO	0.2	0	0
Na_2O	0	0	0
K_2O	0	0	0
Cr_2O_3	0	0	0.3
Total	99.5	99.3	99.1

Table. 4.15A. Composition of feldspar grains in the granitic rock samples of East Antarctica.

SiO2	Al2O3	Na2O	K2O	CaO	MgO	FeO	MnO	TiO2	Cr2O3	Total	An	Ab	Or
67.9	19.7	10.8	0.1	0.5	0	0	0	0	0.1	99.2	2	98	0
67.7	19.8	11.2	0.2	0.7	0.4	0	0	0	0	100.1	3	97	0
67.7	19.3	11.5	0.2	0.4	0	0	0	0	0.3	99.6	2	98	0
67.5	19.9	12.1	0.1	0.2	0.1	0.2	0.1	0.2	0	100.5	1	99	0
67.4	19.9	11.1	0.2	0.8	0	0	0.1	0	0	99.5	5	94	1
67.2	20.2	11.1	0.1	0.9	0	0	0	0	0	99.4	5	95	0
67.2	20.1	11.4	0.1	0.5	0	0.3	0	0	0	99.6	2	98	0
67.2	20.1	11.4	0.1	0.5	0	0.3	0	0	0	99.6	2	98	0
66.7	20.7	10.2	0.1	1.1	0	0	0	0	0	98.9	5	94	1
66.6	20.7	10.2	0.1	1.1	0	0	0	0	0	98.9	5	94	1
66.4	20.1	11.7	0.1	0.5	0.1	0	0.2	0	0.1	99.7	2	98	0
66.4	20.1	11.7	0.1	0.5	0.1	0	0.1	0	0.1	99.7	2	98	0
63.7	18.5	0.1	15.9	0.7	0	0	0.4	0.7	0	100.1	2	0	98
63.5	18.8	1.1	15.8	0.8	0.3	0	0.1	0.2	0	100.6	2	9	89
63.5	18.2	0.2	16.3	0.5	0	0	0	0.2	0.2	99.3	2	2	96
63.5	18.8	1.1	15.8	0.8	0.3	0	0.1	0.2	0	100.5	2	9	89
63.4	18.1	0.2	16.4	0.8	0	0	0	0.3	0.3	99.5	2	2	96
63.4	18.8	0.4	15.8	0.5	0	0	0	0.5	0	99.9	2	2	96
63.2	18.5	0.1	16.3	0.4	0	0	0	0.2	0.2	99.2	2	1	97
63.2	18.4	0.5	16.1	0.6	0.3	0	0	0.3	0.1	99.5	2	5	93
63.1	18.7	0.1	15.9	0.7	0.1	0	0.2	0.2	0	99.1	2	0	98
63.1	18.2	0.4	16.1	0.5	0.1	0	0.1	0.2	0	98.7	2	2	96
63.1	18.6	0.7	15.9	0.5	0	0	0.3	0.6	0.1	99.9	2	7	91
63.1	18.6	0.3	16.2	0.5	0	0	0.2	0.5	0	99.4	2	2	96
63.1	18.8	0.1	15.9	0.7	0.1	0	0.3	0.9	0	99.9	2	0	98
62.9	18.5	0.7	15.9	0.4	0.1	0	0.3	0.3	0	99.1	2	7	91
62.9	18.2	0.5	16.2	0.5	0	0	0	0.4	0.4	99.4	2	4	94
62.9	18.5	0.2	16.2	0.5	0.1	0	0	0	0.2	98.6	2	1	97
62.7	18.5	0.1	15.9	0.7	0	0	0.4	0.7	0	99.1	2	0	98
62.6	19.7	0.5	15.5	0.5	0	0	0	0.5	0.1	99.5	2	5	93
62.6	18.2	0.3	15.5	0.8	0.2	0	0	0.4	0	98.2	5	2	93
62.5	19.1	0.3	16.5	0.4	0.1	0	0	0.7	0	99.8	2	2	96
62.5	18.1	0.2	16.3	0.5	0.2	0	0	0.2	0.2	98.3	2	2	96
62.4	19.4	0	16.3	0.4	0	0	0	0.9	0.2	99.1	2	0	98
62.4	18.3	0.2	16.4	0.8	0	0	0	0.4	0.3	98.6	2	2	96
62.1	18.3	0.07	16.1	0.6	0.5	0	0	0.5	0.4	98.6	2	1	97

Table. 4.15B. Composition of feldspar grains from Victoria Land of East Antarctica.

SiO2	Al2O3	K2O	Na2O	CaO	FeO	MgO	MnO	TiO2	Cr2O3	Total	An	Ab	Or
67.5	19.8	0.1	11.4	0.6	0	0	0.1	0.4	0.3	100.3	2	98	0
64.8	20.7	0.1	8.4	3.5	0.4	0	0.2	0	0.1	98.3	19	80	1
63.6	21.3	0.1	9.6	20.3	0.4	0.1	0	0	0.4	98.2	12	88	0
63.2	21.4	3.7	7.5	2.6	0.1	0	0	0.5	0.4	99.5	13	67	20
62.8	18.1	16.8	0.1	0.5	0	0.1	0	0.3	0.2	98.9	2	0	98
61.9	22.7	3.1	8.3	1.4	0.8	0.5	0.3	0.1	0.1	99.8	6	76	18
61.9	22.7	0.1	7.1	5.5	0.5	0	0	0	0.3	98.3	28	72	0
60.5	22.4	0.1	9.1	5.7	0.2	0.1	0	0	0.1	98.4	26	74	0

Table. 4.15C. Composition of feldspar grains in phyllite rock samples from Comonwealth Bay

SiO2	Al2O3	K2O	Na2O	CaO	FeO	MgO	MnO	TiO2	Cr2O3	Total	An	Ab	Or
62.7	0.1	17.2	0	0	0	0.2	0	16.4	0.2	97.5	1	0	99
62.2	0.2	17.1	0	0.1	0	0.3	0	16.3	0.1	67.1	1.5	0	98.5
62.8	0.2	17.3	0	0.2	0	0.4	0.1	16.1	0	97.7	1.5	0.5	98
63.3	0.2	17.3	0	0	0	0.4	0	16.2	0.1	98.1	2	0	98
67.3	0	19.2	0	0.1	0	0.1	9.9	0.1	0	97.1	0.5	99	0.5
66.5	0	20.8	0.2	0.2	0	0.7	9.9	0	0	98.4	3	97	0

Table. 4.15D. Composition of feldspar grains in the gneissic rock samples of East Antarctica.

SiO2	Al2O3	K2O	Na2O	CaO	MgO	MnO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
68.1	20.4	0.1	10.5	0.2	0.1	0.3	0	0.1	0.4	100.2	1	98	1
68.1	20.1	0.1	11.7	0.4	0.1	0.3	0	0	0.2	101.8	2	98	0
67.5	19.3	0.2	11.4	0.4	0.1	0	0.2	0	0.1	99.6	2	98	0
67.5	20.1	0.1	10.7	0.4	0	0	0	0	0.1	99.4	2	98	0
67.2	19.7	0.1	11.1	0.3	0	0	0.1	0.2	0	98.7	1	99	0
67.2	19.8	0.1	10.8	0.2	0	0	0.2	0.3	0.2	99.7	1	99	0
65.3	20.6	0.9	10.3	0.5	0	0.3	0.4	0.1	0.1	98.5	2	92	6
64.1	18.3	12.2	2.5	0.9	0.2	0	0	0.4	0	99.4	5	23	72
63.8	19.1	14.5	1.3	0.3	0	0	0	0.7	0	99.8	1	13	86
63.7	19.1	14.5	1.3	0.3	0	0	0	0.7	0	99.8	4	12	84
63.6	18.2	11.5	3.2	0.3	0	0	0	0.2	0.3	97.3	1	28	71
63.6	18.8	12.1	3.6	0.7	0.3	0	0	0.3	0.2	99.4	2	31	67
63.5	19.1	15.2	0.3	0.5	0.1	0.4	0	0.4	0	99.9	3	3	94
63.4	18.7	15.4	0.6	0.5	0.1	0	0	0.4	0.1	99.2	2	5	93
63.4	18.7	15.4	0.6	0.5	0.1	0	0	0.4	0.1	99.2	4	6	90
63.3	18.3	15.3	0.7	0.2	0	0.2	0	0.1	0	98.4	1	5	94
63.3	18.3	9.7	4.7	0.1	0	0.4	0	0	0.2	97.1	1	42	57
63.2	19.1	15.1	0.9	0.4	0.2	0.1	0.1	0.3	0	100.3	2	8	90
63.2	18.2	15.3	0.6	0.6	0.1	0.2	0	0.7	0	99.1	3	7	90
63.2	18.2	13.8	1.9	0.3	0	0	0	0	0.3	98.1	2	17	81
63.2	19.1	10.5	3.5	1.1	0.1	0.2	0	0	0	98.6	5	32	63
63.1	18.3	14.1	1.7	0	0	0	0	0	0.3	97.6	0	16	94
63.1	18.7	15.9	0.5	0.7	0.2	0.1	0	0.5	0.4	100.4	2	4	94
63.1	18.3	15.3	0.6	0.6	0	0.1	0	0.4	0	98.5	2	5	93
63.1	18.8	16.1	0.5	0.7	0.3	0.1	0	0.5	0.4	100.4	4	5	91
63.1	19.1	11.6	3.2	0.1	0	0.2	0.1	0	0	97.5	1	28	71
62.8	19.1	15.7	0.5	0.5	0.3	0	0	0.4	0	99.2	2	5	93
62.8	19	15.8	0.5	0.5	0.3	0	0	0.4	0	99.2	4	4	92
62.7	18.4	15.3	0.7	0.5	0	0	0	0.7	0	98.2	2	7	91
62.7	18.4	13.4	1.7	0.2	0	0.1	0.1	0.2	0.3	97.1	1	15	84
62.7	18.1	13.9	1.8	0.5	0.2	0	0	0.2	0.1	97.3	2	17	81
62.4	17.9	15.4	0.7	0	0.2	0.1	0	0.1	0	97.1	0	7	93
62.4	18.5	14.7	1.3	0.3	0	0	0	0.2	0.3	97.6	1	10	89
62.4	18.6	14.8	1.1	0.8	0.4	0.1	0	0.5	0	98.8	4	10	86
62.3	18.4	16.4	0.5	0.7	0.1	0.3	0	0.3	0	99.2	2	5	93
62.3	18.4	16.4	0.5	0.7	0.1	0.3	0	0.3	0	99.2	3	5	92
62.3	18.7	13.5	1.9	0.2	0.2	0.1	0	0.3	0	97.4	1	18	81
62.1	18.1	15.6	0.1	0.1	0	0.2	0	0	0	97.1	0	0	100
62.1	17.9	15.6	1.1	0.6	0.1	0.3	0	0.9	0	99.4	2	9	89
62.1	18.8	15.3	1.3	0.5	0.3	0.4	0	0	0	98.6	2	12	86
62.1	18.8	14.6	1.8	0.7	0	0.2	0	0.4	0	99.8	2	17	81
62.1	18.6	14.8	1.9	0.9	0	0	0	0.3	0.2	97.7	5	17	78
61.8	18.7	15.9	0.9	0.2	0.2	0	0	0	0	98.3	1	7	92
61.8	18.7	16.1	0.9	0.2	0.2	0	0	0	0	98.3	1	7	92
61.8	23.6	0.3	8.1	5.4	0	0	0	0	0.1	99.3	26	72	2
61.6	23.8	0.2	8.1	5.6	0	0.2	0	0	0.1	99.6	28	71	1
61.6	23.7	0.4	8.1	5.3	0	0.1	0	0	0.2	99.5	26	72	2
61.5	23.3	0.3	8.1	5.6	0	0.4	0.2	0	0.3	99.9	27	71	2
67.3	20.4	0.2	10.7	1.1	0	0	0.2	0	0.2	100.1	5	94	1
67.1	20.2	0.1	11.1	0.6	0	0	0.4	0	0.2	99.7	3	96	1
66.9	19.5	0.1	11.2	0.7	0.1	0.6	0	0.1	0.2	99.4	2	97	1
66.9	20.2	0	10.6	0.6	0	0	0	0	0	98.3	3	97	0
66.8	20.2	0.1	11.3	0.7	0.3	0.2	0	0.3	0	99.9	2	97	1
66.6	20.3	0.2	11.1	0.5	0	0	0	0	0.5	99.2	2	97	1
66.3	19.7	0.2	10.5	0.5	0.2	0.6	0	0	0.1	98.1	2	97	1
66.3	20.1	0	11.4	0.7	0.1	0	0	0	0.1	98.7	3	97	0
66.1	19.1	1.5	10.8	0.5	0	0	0.3	0.1	0	98.4	2	91	7
65.9	20.9	0	11.1	1.1	0.1	0.2	0	0	0.2	99.5	4	96	0
65.7	20.4	0.2	11.2	0.9	0	0	0	0	0	98.4	5	94	1
65.2	19.2	9.1	4.3	0.6	0	0.2	0	0.3	0	98.9	3	40	57
63.7	18.3	15.1	0.4	0.5	0.4	0.5	0	0.7	0.1	99.7	3	6	91
63.7	18.4	16.1	0.5	0.7	0.2	0.6	0	0.7	0.3	101.2	2	4	94
63.6	18.4	15.3	0.3	0.7	0.4	0.1	0	0.6	0.1	99.5	3	3	94
63.2	18.1	16.3	0.3	0.7	0	0.1	0	0.2	0	98.9	2	2	96
63.2	17.9	16.7	0	0.5	0	0	0	0.6	0	98.9	2	0	98
63.2	18.8	16.4	0	0.7	0.1	0.2	0	0.4	0.3	100.1	3	0	97

Table. 4.16A. Composition of garnet grains in the granitic rock samples of East Antarctica.

SiO2	37.3	37.5	37.1	37.4	36.9	37.8	37.3	37.6	37.1	37.6	37.5	37.1
TiO2	0	0	0	0	0.1	0	0.1	0	0	0	0	0
Al2O3	22.6	21.5	21.3	22.1	21.3	21.9	21.4	21.9	21.3	21.9	22.5	22.1
FeO	20.9	20.9	21.9	21.9	21.8	21.8	21.8	20.1	21.8	21.8	20.9	20.6
MnO	14.4	14.6	14.8	14.4	15.2	14.8	14.2	15.6	14.5	14.4	14.5	14.4
MgO	4.4	3.9	3.6	3.9	4.1	4.2	4.5	3.7	4.1	4.5	3.9	4.5
CaO	1.2	1.2	1.2	1.2	1.3	1.2	1.2	1.2	1.2	1.2	1.3	1.3
Na2O	0	0	0	0	0	0	0	0	0	0	0	0
K2O	0	0	0	0	0	0	0	0	0	0	0	0
Cr2O3	0.1	0	0.1	0	0	0.1	0	0	0.3	0	0	0
Total	100.8	100.7	100.4	100.6	100.5	101.6	100.8	100.4	100.8	101.2	100.7	100.6

No. of ions on the basis of 24 (O)

Si	5.9	6	5.9	6	5.9	5.9	5.9	5.9	5.9	5.9	6	5.9
Ti	0	0	0	0	0.01	0	0.01	0	0	0	0	0
Al	4.1	4.1	4.1	4	4.1	4	4.1	4.1	4.1	4.1	4.1	4.1
Fe	2.7	2.8	2.9	2.9	2.8	2.8	2.8	2.8	2.9	2.9	2.7	2.8
Mn	1.9	1.9	2	1.9	2	2	1.9	2.1	1.9	1.9	1.9	1.9
Mg	1.1	0.9	0.9	0.9	0.9	1	1	0.9	0.9	1.1	0.9	1
Ca	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Cr	0.01	0	0.01	0	0	0.01	0	0	0.04	0	0	0

Mol.% end-members

Al	47	48	48	49	48	47	47	47	49	48	47	47
Sp	33	34	34	33	34	33	33	35	33	32	33	33
P	17	15	15	15	15	17	17	15	15	17	17	17
G	3	3	3	3	3	3	3	3	3	3	3	3

Table. 4.16B. Cmoposition of garnet grains in the aplite rock samples of East Antarctica.

SiO2	37.3	37.5	37.5	37.1	37.5	37.1	37.8	38.5	37.2	37.8	37.4	37.7	37.1	37.1	37.5	37.1	37.5	37.5	37.3
TiO2	0.2	0	0	0.1	0	0	0	0	0	0	0	0	0	0	0	0.1	0	0	0
Al2O3	21.1	22.1	21.9	21.8	22.2	21.1	21.3	22.2	21.6	21.1	21.3	21.2	22.2	22.7	21.9	21.8	21.8	21.1	21.1
FeO	21.7	22.1	21.1	21.8	20.4	21.8	22.9	21.1	21.8	22.3	21.5	21.2	21.8	21.7	21.7	21.8	21.1	22.1	21.7
MnO	15.1	13.8	14.6	14.5	14.9	15.5	13.8	14.3	15.3	14.9	14.4	14.9	14.9	15.4	15.1	14.5	14.6	13.8	14.9
MgO	3.7	4.3	4.8	3.7	4.1	3.5	3.9	4.3	4.2	4.1	4.3	3.6	4.4	3.7	4.1	3.7	4.8	4.3	3.7
CaO	1.1	1.3	1.3	1.1	1.2	1.1	1.1	1.1	1.4	1	1.2	1.4	1.1	1.1	1.1	1.1	1.3	1.3	1.1
Na2O	0	0	0	0.2	0	0.1	0	0	0	0	0	0	0	0	0	0.2	0	0	0
K2O	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Cr2O3	0	0.1	0.1	0	0	0.1	0	0	0	0	0	0	0	0	0	0	0.1	0	0
Total	100.1	100.2	101.4	100.3	100.2	100.4	100.5	101.7	101.6	101.6	100.5	100.3	101.4	101.7	101.2	100.3	101.2	100.1	99.8

No. of ions on the basis of 24 (O)

Si	6	6	5.9	5.9	5.9	5.9	6	6	5.9	5.9	5.9	6	6	5.9	6	5.9	5.9	6	5.9
Ti	0.03	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.01	0	0	0
Al	4.1	4	4	4.1	4.1	4.1	4.1	4.1	4.1	4.1	4.1	4.1	4.1	4.2	4	4.1	4	4.1	4.1
Fe	2.8	2.9	2.8	2.9	2.8	2.9	2.9	2.7	2.8	2.9	2.8	2.8	2.8	2.8	2.8	2.9	2.9	2.9	2.8
Mn	2	1.8	1.9	1.9	2	2.1	1.9	1.9	2	2	1.9	1.9	2	2	2	1.9	1.9	1.8	1.9
Mg	0.9	1	1.1	0.9	0.9	0.8	0.9	1	1	1	1	0.8	1	0.8	0.9	0.9	1.1	1	0.9
Ca	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Cr	0	0	0.01	0	0	0.01	0	0	0	0	0	0	0	0	0	0	0.01	0	0

Mol.% end members

Al	47	49	47	49	48	48	49	46	47	47	47	48	47	48	47	48	47	48	48
Sp	34	31	32	32	33	35	32	34	33	33	32	34	33	34	34	35	35	34	34
P	15	17	18	15	15	13	15	17	17	17	18	14	17	15	16	14	15	15	16
G	4	4	3	3	3	3	4	3	3	3	3	3	3	3	3	3	3	3	2

Table. 4.16C. Comoposition of garnet grains in the gniessic rock samples of East Antarctica.

SiO2	37.5	37.8	38.1	37.8	37.9	37.9	36.9	36.7	38	37.5	36.7	37.6	37.9	37.3	38.7	37.5	37.9	38.7	37.8
TiO2	0.3	0	0.1	0	0	0	0	0.1	0	0.2	0.2	0.1	0.1	0.2	0	0	0	0.1	0
Al2O3	22.2	22.6	21.9	22.5	21.9	22.2	22.1	22	21.9	22.6	21.7	22.3	22.1	22.3	22.4	22.8	22.1	21.3	22.5
FeO	27.5	27.2	27.5	26.3	28.2	28.2	27.3	27.2	27.4	27	31.2	29.5	29.5	27.9	27.9	27.1	27.8	27.8	28.4
MnO	4.5	4.5	3.8	4.2	4.8	4.5	4.5	4.1	4.3	4.7	5.8	5.6	5.9	6.4	4.1	4.6	4.3	4.8	4.3
MgO	7.3	7.6	6.9	7.3	6.9	7.2	6.8	7.5	7.2	7.1	3.9	4.3	4.5	4.1	7.1	6.8	7.4	7.2	6.6
CaO	1.1	1.3	1.4	1.3	1.2	1.4	1.4	1.6	1.3	1.1	1.3	1.6	1.3	1.1	1.4	1.3	1.4	1.7	1.5
Na2O	0	0	0	0	0	0	0	0	0.5	0	0	0.4	0	0.4	0	0	0	0	0
K2O	0	0	0	0	0	0	0	0	0	0	0.1	0.2	0	0	0	0	0	0	0
Cr2O3	0	0.5	0	0	0	0	0	0.3	0.1	0.1	0	0	0	0.4	0	0.1	0	0.1	0
Total	100.7	101.7	100.3	99.4	101.1	101.6	99.3	99.9	100.8	100.3	100.9	101.6	101.3	100.1	101.8	100.2	100.9	101.7	101.1

No of ions on the basis of 24 (O)

Si	5.8	5.8	5.9	5.9	5.9	5.9	5.8	5.8	5.9	5.9	5.8	5.9	5.9	5.9	5.9	5.9	5.9	5.9	5.9
Ti	0.04	0	0.01	0	0	0	0	0.01	0	0.03	0.02	0.01	0.01	0.02	0	0	0	0.01	0
Al	4.1	4.1	4	4.1	4	4.1	4.1	4.1	4	4.2	4	4.1	4.1	4.1	4.1	4.1	4.1	4	4.1
Fe	3.6	3.5	3.6	3.4	3.7	3.6	3.6	3.6	3.6	3.5	4.1	3.9	4	3.7	3.6	3.6	3.5	3.6	3.6
Mn	0.6	0.6	0.5	0.5	0.6	0.6	0.6	0.5	0.6	0.6	0.8	0.7	0.8	0.8	0.5	0.6	0.5	0.6	0.6
Mg	1.7	1.7	1.6	1.7	1.6	1.6	1.6	1.7	1.7	1.6	1.7	1	1	1	1.6	1.6	1.7	1.6	1.5
Ca	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.3	0.2	0.2	0.2	0.2	0.2	0.3	0.2
Cr	0	0.06	0	0	0	0	0	0.04	0.01	0.01	0	0	0	0.05	0	0.01	0	0.01	0

Mol.% end-members

Al	59	58.3	61	58.6	60.6	60	60	60	59	59.3	63.1	66.1	66.7	70	61	60	59	58	62
Sp	9.8	10	8.5	8.6	9.8	10	10	8.3	9.8	10.2	12.7	11.9	13.3	16	8	10	8	10	9
P	27.9	28.3	27.1	29.3	26.2	25.7	26.7	28.3	27.9	27.1	26.9	16.9	16.7	11	27	27	29	27	26
G	3.3	3.3	3.4	3.4	3.3	3.3	3.3	3.3	3.3	3.4	3.2	5.1	3.3	3	3	3	3	5	3

Table. 4.16E. Composition of garnet grains in the gniesses sample of East Antarctica.

			*								*	*		
SiO2	38.4	37.2	37.5	37.9	36.7	38.2	37.6	37.9	37.9	38.2	38.7	38.7	37.4	37.6
TiO2	0.1	0	0.4	0	0.3	0	0	0.3	0	0	0	0.2	0.3	0
Al2O3	21.6	21.8	21.7	21.8	21.5	22.2	22.3	22.2	22.2	21.8	22.1	22.1	21.8	22.1
FeO	27.8	26.8	27.9	28.2	27.9	27.7	27.2	27.2	27.4	27.3	22.6	26.5	29.5	28.3
MnO	4.4	4.5	4.6	4.2	4.8	3.9	4.2	4.1	4.3	4.2	0.7	0.6	5.6	5.8
MgO	7.5	6.5	7.5	6.5	6.5	7.3	6.9	7.3	6.6	6.7	11.5	10.5	4.8	4.7
CaO	1.3	1.2	1.1	1.6	1.7	1.6	1.5	1.5	1.3	1.2	1.3	1.9	1.4	1.4
Na2O	0	0	0	0	0	0.2	0	0	0	0	0.9	0.1	0	0
K2O	0.1	0	0	0	0	0	0	0	0	0	0.1	0.2	0	0
Cr2O3	0	0.3	0	0.3	0.4	0	0	0.3	0.3	0	0.3	0	0	0
Total	101.5	98.7	101.3	100.6	99.8	101.6	100.2	101.2	100.3	99.4	98.2	100.8	100.8	99.9

No of ions on the basis of 24 (O)

Si	5.9	5.9	5.8	5.9	5.8	5.9	5.9	5.8	5.9	6	6	5.9	5.9	5.8
Ti	0.02	0	0.04	0	0.04	0	0	0.03	0	0	0	0.02	0.03	0
Al	3.9	4.1	4	4	4.01	4.1	4.1	4	4.1	4	3.9	4	4.1	4.1
Fe	3.6	3.6	3.6	3.7	3.7	3.6	3.6	3.5	3.6	3.6	3.1	3.3	3.8	3.8
Mn	0.6	0.6	0.6	0.5	0.6	0.5	0.6	0.5	0.6	0.6	0.07	0.07	0.7	0.8
Mg	1.7	1.5	1.7	1.5	1.5	1.7	1.6	1.7	1.5	1.6	2.4	2.3	1.1	1.1
Ca	0.2	0.2	0.2	0.3	0.3	0.3	0.3	0.2	0.2	0.2	0.2	0.3	0.2	0.2
Cr	0	0.03	0	0.03	0.05	0	0	0.03	0.04	0	0.03	0	0	0

Mol.% end-members

Al	59	61	59	62	61	59	60	59	61	60	54	55	65	64
Sp	10	10	10	8	10	8	10	8	10	10	1	2	12	14
P	28	26	28	25	24	28	27	29	25	27	42	38	19	19
G	3	3	3	5	5	5	3	3	4	3	3	5	3	3

Table.

Table. 4.16F. Composition opf garnet grains in phyllite rock samples in Comonwealth Bay of East Antarctica.

SiO2	36.5	36.3	35.8	35.9	36.8	35.8	35.7
TiO2	0	0.1	0	0.1	0.2	0	0.1
Al2O3	20.9	20.9	21.1	20.3	21.2	20.6	20.5
FeO	15.2	14.8	15.1	12.9	15.3	15.1	12.9
MnO	22.8	23.9	23.7	23.7	23.3	23.8	24.8
MgO	1.9	1.5	1.6	1.4	2.2	1.8	1.3
CaO	2	2.5	1.2	3.7	2.2	2.1	2.9
Na2O	0	0	0	0	0	0.2	0
K2O	0	0	0	0	0	0	0
Cr2O3	0	0.1	0	0.2	0.1	0.1	0.2
Total	99.5	100.3	98.5	98.3	101.4	99.4	98.3

No. of ions on the basis of 24 (O)

Si	5.9	5.8	5.8	5.9	5.8	5.8	5.8
Ti	0	0.008	0	0.008	0.02	0	0.02
Al	4	4	4	3.9	3.9	3.9	4
Fe	2	2	2	1.8	2	2	1.8
Mn	3.2	3.2	3.2	3.4	3	3.2	3.4
Mg	0.4	0.3	0.4	0.3	0.5	0.4	0.3
Ca	0.3	0.4	0.2	0.6	0.3	0.3	0.5
Cr	0	0.02	0	0.02	0.02	0.02	0.02

Mol.% end-members

Al	35.7	33.9	34.5	29.5	34.4	33.9	30
Sp	57	54	55	55.7	51.7	54.2	56.7
P	7	5	6.9	4.9	8.6	6.8	5
G	5	6.8	3.4	9.8	5	5	8.3

Table. 4.17A. Composition of ilmenite grains in the granitic rock samples of East Antarctica.

TiO2	FeO	MnO	MgO	Al2O3	V2O3	Cr2O3	SiO2	Total
50.9	41.7	7.2	0.5	0.2	0.4	0	0.1	101.3
56.7	37.9	3.8	0.2	0.3	0.2	0	0.3	100.7
51.7	44.2	3.2	0	0.3	0.2	0.1	0	100.7
48.1	42.8	6.3	0.3	0.3	0.4	0	0	100.5
50.9	43.4	5.7	0.1	0	0.3	0	0	100.4
50.2	43.1	5.5	0.4	0.2	0.3	0.3	0	100.2
53.1	141.3	4.9	0	0	0.3	0	0	100.1
51.1	44.2	3.5	0	0	0.3	0.1	0	99.9
52.3	40.7	5.9	0.1	0	0.2	0	0	99.8
49.6	43.9	5.5	0	0.3	0.2	0	0	99.7
52.2	44.2	2.9	0	0	0.3	0	0	99.5
52.8	41.1	4.8	0.3	0	0.2	0	0	99.1
50.2	42.1	5.5	0.2	0	0.3	0	0	99.1
47.4	41.5	7.5	0	0.3	0.3	0.2	0	98.7
50.8	41.6	5.7	0.1	0.1	0.3	0	0	98.5
50.3	42.9	5.1	0	0	0.3	0	0	98.4
51.2	40.6	4.3	0.1	3	0.2	0	0	98.3
48.1	43.7	5.1	0.1	0.1	0.3	0.1	0.1	98.2
49.9	40.9	6.6	0	0.2	0.3	0.2	0	97.7
47.3	42.2	7.7	0.1	0.2	0.4	0.3	0	97.7
50.4	39.9	6.2	0	0.2	0.3	0.1	0.5	97.3
48.1	40.9	6.5	0.4	0.1	0.2	0.1	0	97.2
47.6	43.1	5.9	0.1	0.2	0.2	0.1	0.1	97.2
55.2	37.3	1.5	0.8	0.7	0.3	0.6	0.2	96.6

Table.4.17B. Composition of ilmenite grains in the charnokite rock samples of East Antarctica.

TiO2	FeO	MnO	MgO	Al2O3	V2O3	Cr2O3	SiO2	Total
52.3	43.1	1.2	0.3	0	0.2	0	0	97.1
52.2	44.4	1.2	0.2	0.1	0.3	0.1	0	98.4
52.1	45.1	0.8	0.4	0	0.3	0	0	98.6
51.9	44.5	1.1	0.4	0	0.4	0	0	98.4
51.9	44.1	1.7	0.1	0	0.3	0.1	0	98.1
51.8	44.2	1.3	0.2	0	0.2	0	0	97.7
51.7	44.1	1.2	0.3	0.1	0.2	0	0	96.7
51.4	44.7	1.1	0.2	0	0.2	0.1	0	97.7
51.3	44.5	1.1	0.1	0	0.2	0.1	0	97.5
50.7	45.8	1.1	0.1	0	0.3	0	0	98.1

Table. 4.17C. Composition of ilmenite basic gneisses rock samples of East Antarctica.

TiO2	FeO	MnO	MgO	Al2O3	V2O3	Cr2O3	SiO2	Total
49.6	44.1	4.5	0.6	0.1	0.3	0.2	0	99.4
51.5	44.9	2.8	0	0.1	0.2	0	0.3	99.8
52.6	43.6	1.8	0	0.3	0.2	0	0	98.5
50.8	44.7	3.7	0.5	0	0.2	0	0	99.9
51.2	45.2	2.9	0.3	0.3	0.2	0.2	0	100.3
50.8	44.1	1.8	0.1	0	0.3	0.1	0.1	97.3
52.6	43.1	1.9	0	0.2	0.2	0	0.1	98.1
49.2	44.4	1.9	0.3	0.6	0.2	0	0	96.6
50.3	45.1	3.1	0.5	0.5	0.3	0	0	99.8
50.6	42.7	3.9	0	0.1	0.2	0.1	0.1	97.7
50.1	43.1	4.3	0	0	0.2	0	0	97.7
51.7	42.2	4.3	0.3	0.1	0.2	0	0.3	99.1
50.7	43.5	4.1	0	0	0.2	0.3	0	98.8
50.7	43.4	3.7	0	0	0.3	0	0	98.1
51.4	45.7	2.7	0.4	0.1	0.2	0.2	0	100.7
50.8	45.8	2.2	0.2	0	0.2	0	0	99.2
50.7	45.3	2.9	0	0	0.2	0.3	0	99.4
50.1	43.8	4.1	0	0	0.3	0	0	98.3
52.9	46.1	2.3	0	0	0.2	0	0.1	101.6
51.9	44.9	2.5	0.1	0.2	0.3	0	0.2	100.1
51.4	42.8	3.8	0.3	0.3	0.1	0.1	0	98.8
54.1	41.6	1.9	0.2	0.2	0	0.1	0	98.1
51.2	44.1	2.1	0.4	0.3	0.5	0.2	0	98.8

Table. 4.18. Composition of tourmaline grains in the granitic gneisses of East Antarctica.

SiO2	35.5	34.4	36.1
TiO2	1.2	0.9	2.1
Al2O3	35.2	28.1	27.2
FeO	4.4	9.6	9.3
MnO	0.1	0.2	0
MgO	6.9	7.5	8.4
CaO	1.3	2.7	1.6
Na2O	1.5	1.1	2.1
K2O	0	0.1	0
Cr2O3	0.8	0.1	0.3
Total	86.9	84.7	87.1

Table. 4.19. Composition of zircon grains in the gneisses rock samples of East Antarctica.

SiO2	29.1	29.1	29.1	29.9	28.9	29.3	29.7	29.9	29.4	29.4
ZrO2	66.4	67.8	66.7	65.5	67.1	66.3	66.5	66.3	65.8	67.5
Al2O3	0	0	0	0	0	0.1	0	0	0	0
FeO	0	0.1	0.1	0.1	0.1	0.2	0.1	0	0.1	0
MnO	0	0	0	0	0	0	0	0	0	0
MgO	0	0	0	0	0	0	0.1	0	0	0
Hf2O3	1.7	1.4	1.7	2	1.8	1.5	1.8	1.6	1.8	1.2
Th2O3	0.1	0.1	0.1	0	0.1	0.1	0.1	0.1	0.1	0.2
Total	97.4	98.5	97.8	97.5	98.1	97.5	98.2	97.9	97.2	98.2

Structure formula on base of 16 (O)

Si	3.7	3.7	3.7	3.8	3.7	3.8	3.8	3.7	3.8	3.7
Zr	4.2	4.2	4.2	4.1	4.2	4.1	4.1	4.2	4.1	4.2
Al	0	0	0	0	0	0.009	0	0	0	0
Mg	0	0	0	0	0	0	0	0	0	0
Hf	0.06	0.05	0.06	0.07	0.06	0.05	0.06	0.06	0.07	0.04
Fe	0	0.01	0.01	0.01	0.01	0.02	0.01	0	0.01	0
Mn	0	0	0	0	0	0	0	0	0	0

Table. 4.20. Staurolite composition in the metamorphic samples of East Antarctica.

Al2O3	SiO2	FeO	MgO	MnO	CaO	Na2O	K2O	TiO2	Cr2O3	Total
55.5	29.5	14.4	1.8	0.2	0	0.4	0.1	0.7	0	101.6
55.2	27.7	15.4	2.1	0.2	0.1	0	0	0.4	0.2	101.2
55.1	28.3	13.5	2.5	0.1	0.3	0	0	0.4	0	100.9
54.9	27.7	14.6	1.1	0	0.3	0	0	0.4	0.3	99.3
54.6	28.7	14.2	1.8	0.1	0.5	0.6	0	0.7	0.3	101.3
54.3	27.5	14.1	0.6	0.1	0.1	0.1	0	0.4	0.5	98.3
54.2	27.8	14.1	1.9	0.1	0.1	0.4	0	0.6	0	99.8
54.1	28.3	13.6	1.2	0	0.2	0.2	0	0.7	0.2	98.2

Table. 4.21. composition of sillimanite grains in the gneisses of East Antarctica.

	*	*	*	*	
SiO2	38.4	37.9	37.1	37.1	43.2
TiO2	0.3	0.1	0	0.7	0
Al2O3	57.9	59.8	60.8	60.8	56.3
FeO	1.6	1.3	1	0.9	0
MnO	0	0	0.7	0	0
MgO	0.7	0.8	0	0.5	0.2
CaO	0.3	0.3	0.5	0.1	0.2
Na2O	0.3	0	0.1	0.4	0.2
K2O	0.7	0.5	0.5	0	0
Cr2O3	0	0.4	0	0.1	0
Total	100.2	101.1	100.7	100.6	100.1

* Viridine in composition.

Table. 4.22. Composition of epidote grains in granitic gneisses of East Antarctica.

SiO2	37.1	36.9	40.5	37.2	37.9	43.1	39.6	39.9	50.3	36.7	36.7	38.1
TiO2	0.4	0	0	0	0.3	0.2	0	0	0.1	0.2	0	0.2
Al2O3	23.1	24.1	23.7	20.6	23.4	20.1	23.7	22.6	17.8	24.1	24.7	22.7
FeO	12.7	12.6	10.2	16.1	11.8	12.8	11.8	12.8	10.8	13.3	12.3	12.2
MnO	0.4	0	0	0	0.7	0	0	0	0.5	0.5	0.5	0.2
MgO	0	0	0	0	0.6	0	0	0	0	0	0	0
CaO	22.7	22.7	22.5	22.6	23.1	21.1	23.5	22.6	17.9	23.8	22.1	23.1
Na2O	0.3	0	0	0.3	0	0	0.3	0.2	0.3	0.1	0	0.4
K2O	0	0	0.1	0.1	0	0.2	0	0	0	0	0.1	0
Cr2O3	0.1	0.5	0.3	0.2	0	0.2	0.2	0	0.3	0.1	0.2	0
Total	96.8	96.8	97.3	97.1	97.8	97.7	99.1	98.1	98	98.8	96.6	96.9

Strcture formula on base of 13 (O)

Si	3.2	3.1	3.3	3.2	3.2	3.6	3.2	3.3	4	3.1	3.2	3.2
Al	2.3	2.4	2.4	2.1	2.4	2	2.4	2.3	1.7	2.4	2.4	2.3
Fe	0.9	0.9	0.7	1.2	0.8	0.9	0.8	0.9	0.7	0.9	0.9	0.9
Mn	0.03	0	0	0	0.05	0	0	0	0.03	0.03	0.04	0.01
Mg	0	0	0	0	0.07	0	0	0	0	0	0	0
Ca	2.1	2.1	2	2.1	2.1	1.9	2.1	2.1	1.5	2.2	2	2.1

* Victoria Land

Table. 4.23. Composition of rutile grains in the gneisses of East Antarctica.

TiO_2	FeO	MnO	MgO	Cr_2O_3	Al_2O_3	CaO	Na_2O	K_2O	SiO_2	Total
99.3	0.9	0	0	0.3	0.1	0	0	0	0	100.6
99.1	0.4	0	0.3	0	0.3	0	0	0	0.1	100.2
98.6	0.4	0	0	0	0	0	0.3	0	0	99.3
98.6	0.2	0	0	0	0.1	0	0.5	0.1	0	99.6
98.6	0.7	0.6	0.4	0	0.1	0	0.7	0.2	0	101.4
98.6	0	0	0	0	0.2	0	0	0	0	98.8
98.6	0.4	0	0	0	0	0	0.3	0	0	99.5
98.6	0.2	0	0	0	0.1	0	0.5	0.1	0	99.6
98.6	0.7	0.6	0.4	0	0.1	0	0.7	0.2	0	101.4
98.4	0.9	0	0	0	0.1	0	0.1	0	0	99.5
98.1	0	0	0	0.5	0.3	0	0.3	0	0	99.2
98.1	0.3	0	0.3	0	0.2	0.1	0.1	0	0	99.1
98.1	0	0	0	0.5	0.3	0	0.3	0	0	99.3
98.1	0.3	0	0.3	0	0.2	0.1	0.1	0	0	99.1
97.9	0.5	0	0	0.3	0	0.1	0.3	0	0	99.1
97.9	0.1	0	0.2	0	0.5	0	0.5	0.2	0	99.4
97.9	0.5	0	0	0.3	0	0.1	0.3	0	0	99.1
97.8	0	0	0	0.1	0	0	0.2	0	0	98.2
97.8	0.3	0	0	0	0.3	0	0.2	0	0	98.6
97.8	0.1	0	0.2	0.3	0.4	0	0.2	0	0.2	99.2
97.8	0	0	0	0.1	0	0	0.2	0	0	98.2
97.8	0.3	0	0	0	0.3	0	0.2	0	0	98.6
97.8	0.1	0	0.2	0.3	0.4	0	0.2	0	0.2	99.2
97.7	0	0	0	0.7	0	0	0.7	0	0	99.1
97.7	0.1	0.3	0	0.3	0	0	0.5	0	0	98.9
97.7	0	0	0	0.7	0	0	0.7	0	0	99.1
97.2	0.8	0.1	0.2	0.3	0.1	0	0.4	0.1	0	99.2
97.4	0.3	0.9	0.4	0	0.7	0.1	0.5	0	0.1	100.4

Table. 4.24. Composition of sphene grains in the rock samples of East Antarctica.

SiO2	TiO2	CaO	Al2O3	FeO	MgO	MnO	Na2O	K2O	Cr2O3	Total
*										
30.1	27.9	24.4	6.8	3.9	4.9	0.2	0	0	0.6	99.9
28.1	37.7	27.2	1.1	1.1	0	0.3	0	0.1	0.1	98.1
31.1	31.5	28.8	4.4	2.1	0.6	0	0.5	0	0.1	99.5
**										
24.1	40.1	23.1	1.6	7.6	0.1	2.4	0.8	0	0.3	101.1
28.7	39.4	28.1	0.9	0.2	0.5	0	0.2	0	0.3	98.8
28.8	35.3	28.8	3.5	0.9	0	0	0	0	0.3	98.2
27.9	41.5	27.1	0.7	0.7	0	0	0	0	0	98.5
28.5	36.1	26.3	3.9	2.3	0	0	0.6	0	0	98.2
30.8	31.4	27.8	5.9	1.9	0.3	0	0	0	0	98.4
***										
21.2	42.9	19.2	1.6	7.8	0	5.6	0	0	0	98.8
16.8	62.5	16.1	0.2	1.6	0.1	0.1	0.5	0	0	95.4
23.4	39.4	22.5	0.9	6.8	0	4.1	0	0	0.4	98.1
29.1	32.5	28.2	5.2	1.2	0	0	0.1	0.2	0.6	98.1
28.9	36.3	27.6	2.3	2.4	0	0	0	0	0.2	97.1
21.2	42.9	19.2	1.6	7.8	0	5.6	0	0	0	98.5
16.8	62.5	16.1	0.2	1.6	0.1	0.1	0	0	0	98.1
11.9	46.9	13.1	0.3	15.6	0	10.3	0	0	0	98.2
29.1	32.5	28.2	5.2	1.2	0	0	0	0	0	98.4

**	**Windmill Island granitic samples**
*	**Victoria Land**
***	**Windmill Island gneisses samples**

CHAPTER FIVE

PETROLOGY OF ERRATICS IN THE PERMIAN SEDIMENTS OF SOUTHERN AUSTRALIA

INTRODUCTION

Erratics have long been known in the Permian glaciogenic sediments of southern Australia and presumably were derived from all areas across which the ice sheet travelled. The wide distribution of glacial erratics in the Permian sediments suggests that during this time, South Australia was possibly covered by a very thick ice sheet. The shape of erratics ranges from angular to well rounded. Angular erratics indicate that for them, only minor reworking has taken place in glaciofluvial environment. The majority of erratics however, in the Permian sands, are well rounded. Within a tillite, well rounded clasts suggest that these erratics have probably been deposited and reworked by successive glacial advances. Petrological studies of erratics indicate that they are composed of granites, gneisses, rhyolites, dacites, tuffs, arkoses and subarkoses. The erratic samples, that have been used in this study were collected from different localities in southern Australia by R. Harris (Honours thesis 1971), and myself. Location of samples are shown on (Fig. 1.4). Statistical counts of rock type proportions were not made since bias during collection may not have been excluded.

Methods of study

More than 100 thin sections of the Permian erratics were petrographically studied from different localities. Modal analyses was carried out by point counting, and about 300 points were counted in each thin section .

Although some relationship between size and composition of the erratics was observed, no systematic study of this feature was carried out. The largest clasts are mostly of boulder sizes and consist of granite and quartzite. Volcanic and sedimentary clasts are generally pebble to cobble size. Erratics are dominantly of this size and volcanics appear to form the major rock type (Fig. 5.0). Some of the erratics in the Permian glaciogenic sediments such, as Encounter Bay Granites and Kanmantoo Group metasediments, have local sources, therefore they have only been carried for short distances. In contrast, the porphyritic, volcanic and gneissic erratics have been carried for long distances perhaps from East Antarctica.

IGNEOUS SUITE

The existence of exotic igneous erratics in the Permian sands was first noted by Howchin (1910a), who recorded exotic granites and porphyries at Kings Point near Victor Harbor. Based on this study, this suite makes up about 74% of the total Permian erratics. This assemblage of rock types is reminiscent of the igneous association found in the Victorian Upper Devonian ring complexes. The only locally derived igneous erratics are of Victor Harbor Granite, which was transported northward and northwestward from outcrops at Victor Harbor, Cape Willoughby and Kangaroo Island. Such erratics occur in Yorke Peninsula, Waterloo Bay, Lake Fowler, and Port Vincent. They also occur at Hallett Cove, Cape Jervis and Inman Valley.

Mineral assemblages in plutonic erratics

Petrological studies of plutonic erratics showed that they do not contain any rutile or epidote, so the mineralogy of the plutonic erratic suite is relatively simple and nearly uniform through the Permian sediments The major minerals in these rocks consist of quartz, plagioclase, potassium feldspars and micas (biotite and muscovite). The common accessory minerals are zircon, apatite, sphene, tourmaline and opaques (essentially ilmenite). The following mineral assemblages have been recorded in each group of rocks.

Granite

Granite is mainly composed of quartz (mostly unimodal in size) with average of 36.8%, feldspars mainly microcline and orthoclase with 45.2%, and plagioclase 11.3%. Micas in these rocks are mostly biotite, with average of 3.2% and chlorite less than 1%. The accessory minerals include zircon, apatite, sphene, tourmaline, garnet and opaques.

Volcanic erratics

The volcanic erratic suites in the Permian glaciogenic sediments of southern Australia are comprised principally of rhyolite, rhyodacite, tuff and dacite porphyry that make up an average of 67% of the erratic types in the sediments (Fig.5.0).

Elements	Permian volcanic erratics	Mt Monster volcanic
Y	24.7 (8.3)	44.5 (15)
Sr	194.6 (102.7)	75.7 (29.5)
Rb	142.3 (115.3)	222.1 (34.1)
Nb	12.5 (5.6)	19.9 (2.7)
Zr	142.8 (36)	259.2 (27.2)
Pb	26.4 (13.9)	20.5 (4.3)
Ni	26.8 (18.2)	13.2 (10.7)
Ba	533 (410)	588.2 (217.9)
Sc	6.9 (7.3)	4.8 (1.7)
V	40.4 (5.6)	13.4 (6.3)
Nd	28.3 (13.9)	62.8 (32.7)

Table.5.0. Means and standard deviation of trace elements in the Permian volcanic erratics and Mt Monster volcanic rock samples.

The Permian volcanic erratic samples from different localities have been analysed by XRF and compared with the volcanic rock samples of Mt Monster areas in southeastern South Australia (see Chapter. 4 and Fig. 4.7). As the figure shows there are a big differences between the Permian volcanic erratics and Mt Monster volcanic rock samples. The means and standard deviations of the values of each element were calculated for the Permian volcanic erratics and volcanic rock samples of Mt Monster rock areas (Table. 5.0).

The provenance of the volcanic erratics have been done by a comparison of trace element compositions (Zr, Y, Sr, Rb, Zn, Nb, Pb, Ni, Ba, V, Sc and Nd) of volcanic erratics and volcanic rock samples of Mt Monster area. The volcanic erratics contain highly significant difference in Sr, Zr, V, Nd and Rb when compared to the Mt Monster areas. This indicates that the Permian volcanic erratics were not derived from the volcanic rocks of Mt Monster areas. The location of the source terrane is unknown, but palaeocurrent features indicates these erratics have been carried totally from south and southeast areas. Therefore, the volcanic erratics in these sediments may have been derived from some pre rift-related volcanic activity , rocks now buried under the present continental margins.

Welded tuff

The mean percentage of quartz is 16.5%, Kfeldspar 17.2%, plagioclase 6.6%, mica 3.2%, groundmass 48.5% and pumice 5%. The accessories in these rocks are mostly opaques with a trace amount of sphene and zircon. The minor amount of biotite is usually present in these rocks, but it is mostly altered to chlorite.

Rhyolite

Rhyolite consists of quartz with average of 18.8%, Kfeldspar 19.4%, plagioclase 3.6%, mica 5.1% and groundmass 50%. Accessory minerals are opaques, zircon and sphene.

Rhyodacite

These rocks contain 12.1% quartz, 17.3% Kfeldspar, 16.6% plagioclase, 6% mica and 43.7% groundmass. The accessories in rhyodacite are opaques, zircon, apatite and sphene.

Dacite

These rocks contain 12.5% quartz, 3% Kfeldspar, 26% plagioclase, (mostly zoned crystals), 8.8% mica and 46.2% groundmass. The accessories are mostly opaques, and rarely apatite and zircon are present. Some of the plagioclase in the volcanic rocks show alteration to sericite.

METAMORPHIC SUITE

Different metamorphic rock types can be seen as erratics in the Permian glaciogenic sediments. Gneisses and granitic gneisses rocks are the most common exotic erratics in these sediments. This suite makes up about 22% of total erratics. Most of these suites have a simple mineralogical composition and they are usually composed of quartz, feldspar and micas. Some of the metamorphic suites are very distinctive and have bright red potassium feldspar porphyroblasts. Such rocks can not be matched with any known area along the outcrop belt. It is unlikely that many of these suites were derived from the Lower Proterozoic metamorphic inliers in the Mt. Lofty Ranges (R. Harris 1971). Metasediments of Kanmantoo Group also occur as erratics in the Permian glaciogenic sediments. This indicates a local source for such metasediment erratics.

Mineral assemblages in the metamorphic erratics

Petrographic studies have shown that the metamorphic erratics consist mainly of quartz, K feldspar, plagioclase and micas. The accessories include garnet, rutile, sillimanite, epidote, zircon and opaques. The following mineral assemblages have been recorded in each group of metamorphic rocks.

Gneiss

This rock consists of quartz with an average of 52%. This is mostly polycrystalline and shows bimodality in shape. Feldspars are mostly potassium feldspar with the mean percentage of 43%. Micas are mostly muscovite with the mean percentage of 4%. The accessories include ilmenite, garnet, rutile, sillimanite, epidote and zircon make up less than 1% of the total rock.

Granite gneiss

These rocks are composed of quartz, feldspars and micas. Quartz, with an average of 37.4% is mostly polycrystalline and shows bimodality in shape. Feldspars consist mostly of potassium feldspar, with an average of 44%, and plagioclase 12.8%. Mica is mostly biotite with an average of 2.5%. The accessories include garnet, zircon, epidote, sphene, tourmaline and opaques (mostly ilmenite).

Augen gneiss

Quartz in these rocks is mostly bimodal polycrystalline with an average of 36.7%. Feldspars are mostly potassium feldspar with an average of 42.8% and plagioclase 12.7%. Mica is mostly biotite with an average of 4.5%, and chlorite with less than 1%. Accessories include zircon, sphene, apatite, garnet, epidote and opaques.

SEDIMENTARY SUITE

Clastic sedimentary rocks form an important part of the Permian exotic erratic assemblages. They show a wide range of compositional and textural variation, ranging from quartz arenite to subarkose, arkose, as well as finer grained sediments. They are scattered throughout the Permian sediments, and mechanically resistant types are predominant. Sedimentary erratics form about 4% of total erratics. They are characterized by reddish colours, presence of blue quartz grains with volcanic and metamorphic rock fragments. Blue quartz is very common in the Permian porphyritic erratics. The heavy mineral suites in these erratics consist mostly of garnet, staurolite, rutile, tourmaline, sphene and ilmenite. This suggests that the area from which the metamorphic and igneous erratics were derived possibly formed the source area for these sedimentary erratics. The age of sedimentary suites is unknown, but is possibly Lower or Middle Palaeozoic (Harris 1971).

Mineral assemblages in sedimentary erratics

The major minerals in these suites are quartz and potassium feldspar. Plagioclase does not occur in many samples and probably has been altered during diageneis. Accessory minerals are rare and include garnet, staurolite, biotite, zircon, ilmenite and traces of rutile, sphene and apatite. Study of thin sections from these suites have shown that the surface of most quartz grains have been coated with iron oxides.

Quartz arenites

The quartz arenites consists essentially of quartz grains with an average of 97.7%, so mineralogically it is supermature. The mean grain size is 0.35mm. The shape of quartz grains ranges from subrounded to very well rounded.

Sub-arkoses

These rocks are mineralogically mature and well sorted with a mean grain size of 0.65mm. The mineral composition of these rocks is relatively simple and include mostly quartz and potassium feldspars with averages of 94.5% and 3.5% respectively. Accessories include opaques, zircon, staurolite and tourmaline with less than 1%, and some rock fragments.

Arkoses

These rocks are much coarser grain than the previous arenites and the mean grain size is about 0.8mm. They are mostly composed of quartz 48.3%, potassium feldspar 40%, micas 1.7% and rock fragments fragments 8%. Accessories are zircon, sphene, apatite and opaques.

Cement

Sedimentary erratics are generally well lithified, and the cement consists of silica, hematite and carbonate in various proportions.

LIGHT MINERALS IN THE PERMIAN ERRATICS

Quartz

The proportion of quartz grains in the Permian erratic samples varies and depends on the rock types. The mean percentage of quartz grains in each group of erratics is shown in Table. 5.1.

Samples	Quartz	K feldspar	Plagioclase
Plutonic	34.6	45.2	11.3
Volcanic	13.8	11.5	14.3
Metamorphic	42.3	42	13
Sedimentary	71.4	21	8.5

Table. 5.1. Mean percentage of quartz and feldspars in the Permian erratic samples.

The proportion of monocrystalline to polycrystalline quartz also varies and is related to the rock types. Metamorphic erratic samples have high percentages of polycrystalline quartz grains when compared with igneous erratic samples. The boundary between crystals in polycrystalline quartz grains in metamorphic samples is mostly sutured and crystal size has a bimodal distribution. The boundary between polycrystalline quartz grains in plutonic erratics is straight to sutured, and most have a uni-modal type grain size distribution for components within polycrystalline grains. The number of components within polycrystalline quartz grains varies from 62% with less than five crystals, to 38% with more than five crystal units in plutonic rock samples. The metamorphic rock samples show about 18% with less than five crystals to 82% with more than five crystal units within polycrystalline quartz grains. Distributions for the number of components is shown in Fig. 4.0. Distribution of elongation coefficients in metamorphic samples show that about 94% have more than 1.5 mean, whereas in the igneous samples about 68% have more than 1.5 mean. The results are illustrated in Fig. 4.1. Quartz crystals from plutonic and metamorphic rock samples have shown a variation on CL colours. Quartz crystals in the metamorphic samples have brown CL colours and in the igneous samples have blue CL colours.

Feldspars

The mean percentage and chemical composition of feldspar grains in the erratic samples is related to the rock types. The mean percentage of each feldspar types from different erratics is shown in Table. 5.1. The chemical composition of feldspar grains have been identified by microprobe in different erratic samples and show some variation within rock samples. The chemical composition in the sedimentary samples ranges from An1Ab66Or33 to An2Ab98Or0, in the plutonic samples ranges from An2Ab0Or98 to An1Ab99Or0, in porphyritic samples ranges from An38Ab62Or0 to An3Ab93Or4 and in metamorphic samples ranges from An1Ab0Or99 to An3Ab97Or0 (Fig. 5.1) and in Tables5.2.

HEAVY MINERALS IN THE PERMIAN ERRATICS

Heavy mineral assemblages in erratic samples are mostly related to the rock types. However, the heavy mineral suites of erratics have been identified both physically and chemically. The composition of heavy mineral suites has been studied by micro-probe analyses.

Heavy mineral suites in the metamorphic erratics include garnet, ilmenite, rutile, purple zircon, sillimanite, epidote and staurolite. Garnet is the dominant heavy mineral and the colour ranges from pink to colourless. The composition of garnets in the metamorphic erratics show that most garnets are rich in Fe and Mg and have almost almandine-pyrope composition with relatively low contents of the spessartine and grossular components (Fig. 5.2). The chemical composition of garnets in these samples demonstrates that the almandine constitute 65%, spessartine 10%, pyrope 22% and grossular 3% (Fig. 5.3).

Hf content in zircon of the metamorphic erratics ranges from 1.32% to 1.4%, in plutonic samples it is about 1.5% and in the volcanic samples about 1.33%.

Ilmenite grains in the metamorphic samples are rich in TiO2 and show a narrow range from 49% to 54% (Fig. 4.9). The concentration of MgO and MnO in the ilmenite of metamorphic erratics show that about 94% has less than 0.4wt% MgO and 86% has less than 5wt% MnO in composition (Figs. 4.10 and 4.11).

Heavy mineral suites in the plutonic erratics include tourmaline, zircon, apatite, ilmenite and garnet. Garnets in the plutonic erratics range in colour from deep red to pink. The composition of garnets in the granitic rocks are almost almandin-spessartine and grossular with a relatively low content of pyrope component. However garnets from granitic rocks are compositionally distinct from those in the metamorphic rock samples. The chemical composition of garnet in these samples demonstrates that the almandine constitute16.9%, spessartine 65.1%, pyrope 4.8% and grossular 13.1% (Fig. 5.2). The TiO2 content in the ilmenite grains from plutonic samples show wide ranges from 34% to 48% (Fig. 4.9). The concentration of MnO in ilmenite of plutonic erratics show that about 34% have less than 5wt% (Fig. 4.10). The concentration of MgO in ilmenite from these erratics show that about 85% have less than 0.4wt% (Fig. 4.11).

Heavy mineral suites in the sedimentary samples, mostly sandstones, include staurolite, garnet, ilmenite, zircon, tourmaline, and traces of rutile. The chemical composition of garnets shows that almandine constitute 86.8%, spessartine 4%, pyrope 9% and grossular 2.3% (Fig. 5.2). Volcanic erratics contain large proportions of ilmenite, pyroxene with traces of sphene and zircon. Tourmaline and rutile have not been found in these erratics. Therefore, this suggests that most volcanic samples did not contribute any heavy mineral grains to the Permian sands probably because these rocks are generally too fine grained to supply any sand size grains. The chemical composition of heavy minerals is shown in Tables. 5.3 to 5.7.

DISCUSSION

The Permian glaciogenic sediments of southern Australia contain different erratic rock types, they range from metamorphic, plutonic, volcanic and sedimentary. Among the erratics, volcanic rock samples are the dominant rock types (Fig. 5.0), although N.Alley (pers.comm.) claims a predominance of sedimentary types. As the Permian glaciogenic sediments have been carried mostly by ice, the erratics in these sediments should be derived from all the areas over which the ice sheet has passed.

The composition of garnet in the metamorphic erratics when compared with metamorphic samples from different localities indicates that the metamorphic rock samples from East Antarctica are very similar in composition. This suggests that the gneisses and granitic gneisses of Permian erratics may have been derived from East Antarctica. The characteristics of quartz and feldspars grains in the granitic erratics are very similar to the quartz and feldspars grains of Encounter Bay Granites. Therefore, the source of these granitic erratics would be from Encounter Bay areas. The volcanic erratics of Permian glaciogenic sediments are not however similar to the volcanic samples of Mt Monster areas, and it is therefore unlikely that southewestern Victoria was a source of glacial erratics or sands.

It is concluded that some of the erratics in the Permian glaciogenic sediments have a local source such as Encounter Bay Granite, Kanmantoo Group metasediments but some of them have been carried for long distances, and may have been derived from East Antarctica.

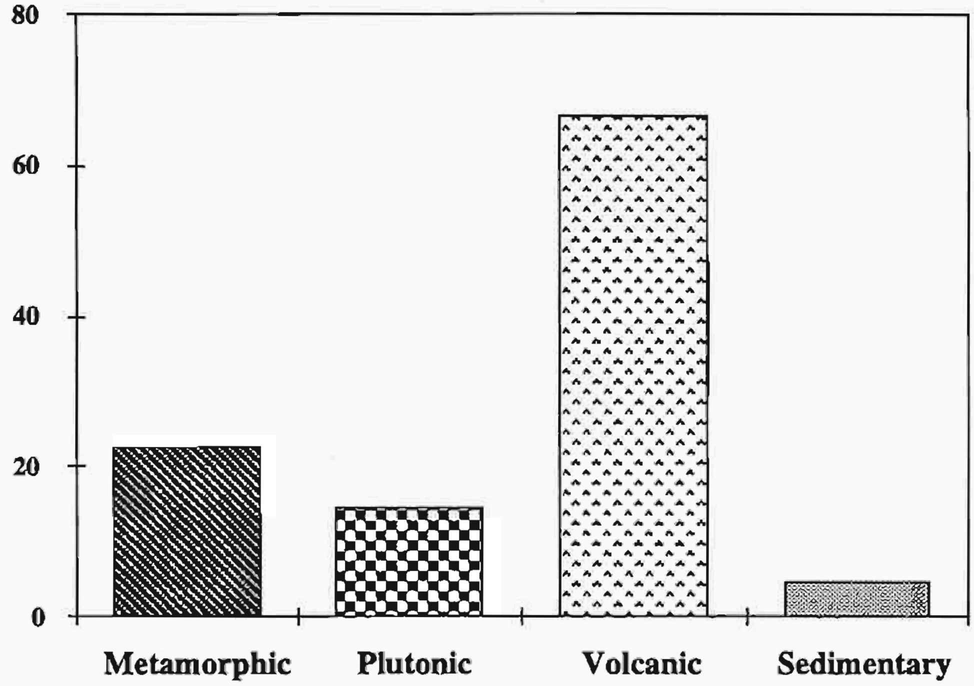


Fig.5.0. Frequency distribution of erratics in the Permian glaciogenic sediments of southern Australia.

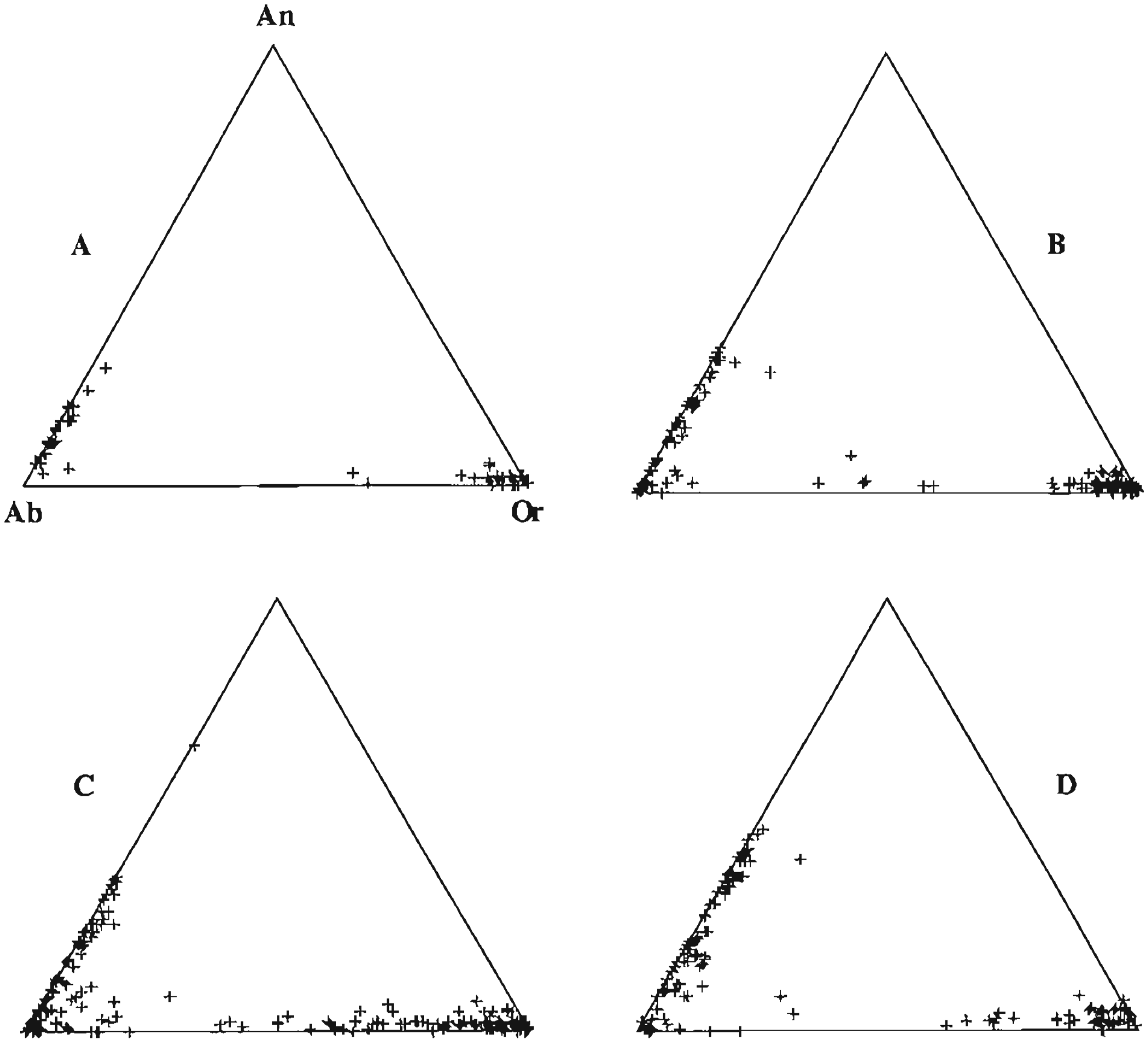


Fig. 5.1. Composition of feldspar grains (weight %) in the Permian erratics of southern Australia as determined by electron microprobe displayed on a triangular diagram: An (anorthite), Ab (albite) and Or (orthoclase) on the base of 32 (O).

A: gneisses

B: granites

C: andesite porphyry

D: dacite

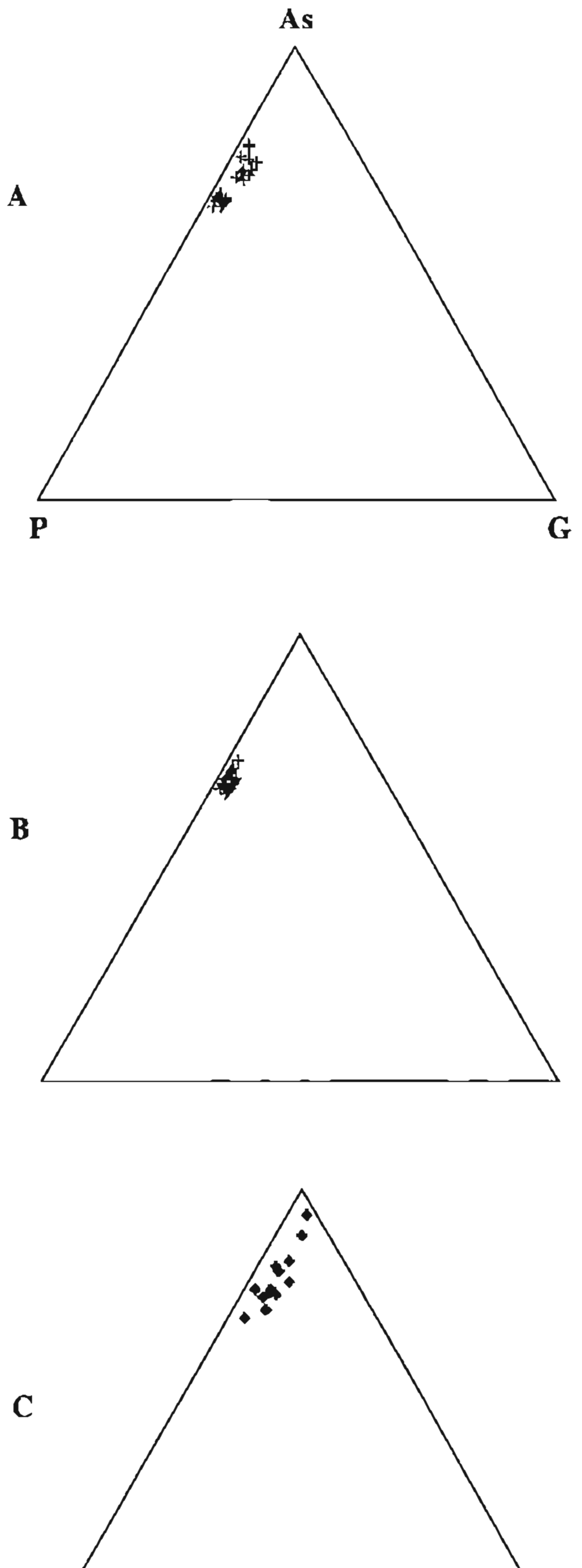


Fig. 5.2 . Garnet geochemistry in the Permian erratic samples of southern Australia displayed on a triangular diagram: AS= Almandine + spessartine; P= Pyrope; G= Grossular.
A: graintic gneisses; B: gneisses; C: granites

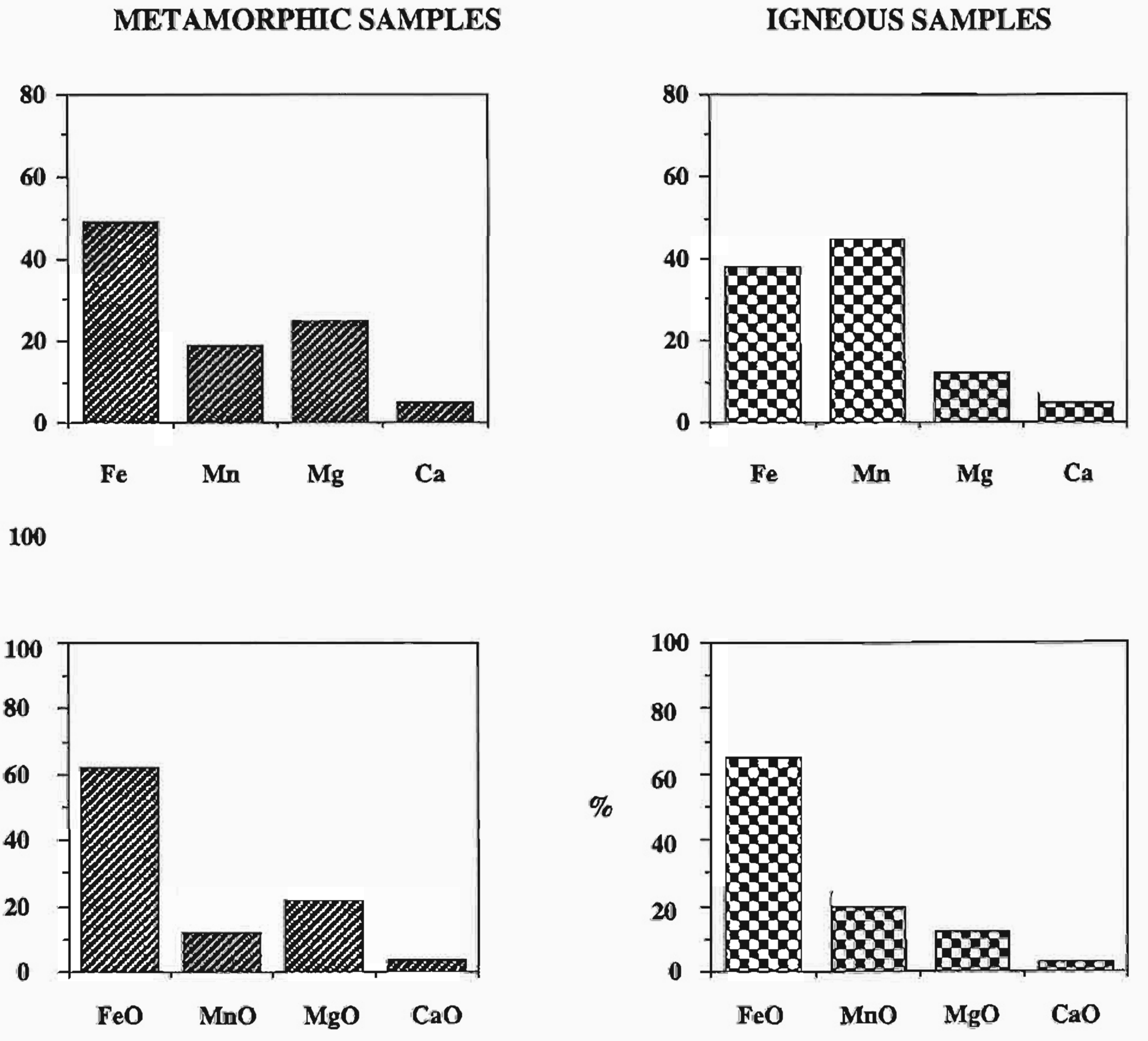


Fig. 5.3. Weight percentage of Fe, Mn, Mg and Ca elements and oxides in the garnet graines on the base of 24 (O) in the Permian erratic samples of southern Australia.

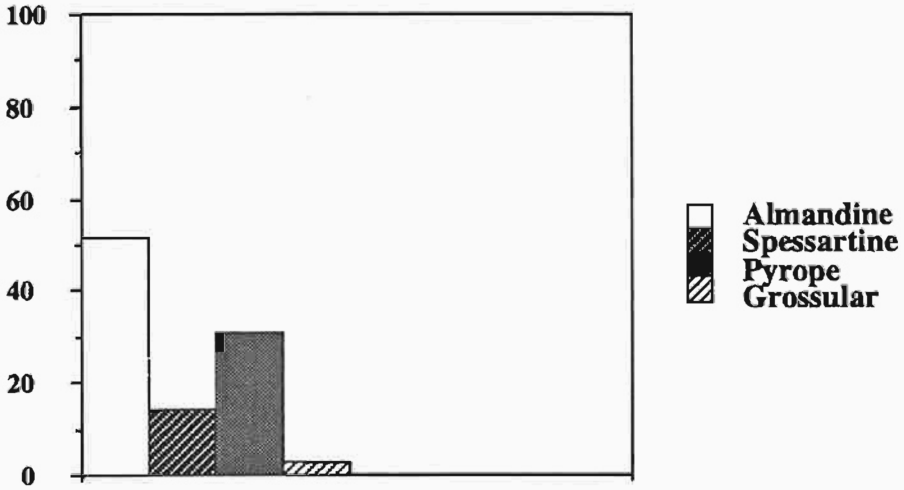


Fig. 5.3a. Garnet group from gneisses in the Permian erratics of southern Australia.

Table. 5.2A.Feldspar composition in plutonic samples from Permian erratics of southern Australia.

SiO_2	Al_2O_3	K_2O	Na_2O	CaO	MgO	MnO	FeO	TiO_2	Cr_2O_3	Total	An	Ab	Or
66.9	20.5	0	10.3	1.4	0	0	0	0	0	99.6	7	93	0
65.4	20.5	0.1	10.5	0.9	0.4	0	0	0.3	0.2	99.5	4	90	6
64.7	18.9	8.1	6.3	0.6	0	0.3	0	0.3	0.3	100.1	3	53	44
64.4	18.4	13.9	1.4	0.7	0	0	0	0	0.4	99.4	2	13	85
64.2	20.9	0	10.6	1.7	0	0	0	0	0.2	98.01	8	92	0
63.8	18.3	16.2	0.4	0.6	0.1	0.1	0	0.2	0.3	100.2	2	2	96
63.6	17.7	15.6	1.2	0.2	0	0.3	0	0.3	0	100.7	1	11	88
63.5	17.5	16.5	0	0.2	0	0.2	0	0.2	0.1	100.1	1	0	99
63.5	18.1	16.5	0.5	0.8	0	0	0	0.4	0.2	100.2	5	5	90
63.4	18.3	16.6	0	0.2	0	0.3	0	0	0	100.1	1	0	99
63.4	18.2	15.6	1.3	0.6	0	0	0	0	0	99.2	2	12	8.6
63.2	18.2	16.8	0	0.2	0.1	0	0	0.5	0	100.4	1	0	99
63.1	22.4	0	9.8	3.6	0.1	0	0.3	0	0.1	99.7	17	83	0
62.9	17.8	16.7	0.5	0.3	0	0.1	0	0	0.5	100.3	1	4	95
62.7	18.2	16.2	0.5	0	0.1	0.5	0	0.2	0	100.5	0	5	95
62.7	17.9	15.3	0.7	0.7	0	0	0	0.1	0.3	99.1	3	7	90
62.6	16.9	16.1	0.6	0.4	0	0.1	0	0.6	0.3	100.4	2	5	93
62.6	17.6	16.9	0.2	0	0	0	0	0.2	0.3	99.9	0	1	99
62.6	18.3	16.2	0.5	0.3	0	0.1	0	0.5	0	99.8	2	4	96
62.6	23.1	0	9.2	4.4	0.2	0	0	0.2	0	99.9	20	80	0
62.3	17.5	16.9	0.1	0.3	0.2	0	0	0	0.1	98.9	2	1	97
62.2	18.3	16.6	0.3	0.2	0	0.2	0	0.1	0.2	99.7	1	2	97
62.1	17.8	16.1	0.2	0.3	0	0	0.1	0.1	0.3	98.3	2	2	96
62.1	18.1	16.4	0.3	0	0	0	0	0	0.4	98.8	0	3	97
61.7	17.4	16.5	0.8	0.5	0	0	0	0.5	0	100	2	7	91
61.6	18.2	16.1	0	0.5	0	0.1	0	0	0.1	98.5	2	0	98
61.3	22.8	0	8.5	4.4	0	0	0.3	0.1	0.3	98.1	22	78	0
59.8	24.7	0.2	7.8	6.5	0.6	0	0	0	0.5	99.1	31	68	1
59.2	24.7	0	7.7	6.2	0	0	0.4	0.3	0.4	99.4	31	69	0
59.2	25.3	0	7.8	6.4	0.1	0	0.3	0	0.3	99.6	31	69	0
58.8	24.7	0.2	7.8	6.5	0.5	0	0	0	0.5	99.1	31	68	1
58.7	23.9	0.3	8.4	5.8	0.2	0.1	0	0.3	0	99.1	27	72	1
67.3	19.8	0	11.3	0.1	0	0.1	0	0	0	98.8	1	99	0
66.1	20.6	0	11.5	0	0.1	0.1	0	0.1	0.1	98.9	0	100	0
63.5	18.3	15.9	0.4	0	0	0.2	0	0.2	0.1	98.9	0	5	95
62.6	18.3	16.2	0	0.5	0	0	0	0.3	0.1	99.2	2	0	98
62.4	17.7	16.4	0	0.2	0	0	0	0.2	0	98.9	1	0	99
62.3	17.3	15.5	0.7	0.5	0	0.6	0	0.8	0.1	98.4	2	6	92
62.3	17.4	15.5	0.7	0.5	0	0.6	0	0.8	0.1	99.4	2	4	94
62.3	18.5	15.6	0.7	0.5	0	0.1	0	0.2	0	99.4	2	4	94
61.8	18.1	16.5	0.4	0.6	0	0	0	0.2	0	99.7	3	3	94
61.2	18.8	15.5	0.2	0.9	0.5	0.1	0	0.6	0	99.1	4	2	94
61.2	18.8	15.5	0.2	0.9	0.5	0.2	0	0.6	0.2	98.1	5	2	93
61.1	18.1	16.3	0.3	0.8	0	0.1	0.4	0.3	0	98.8	4	3	93

Table. 5.2B. Composition of feldspar grains in the Permian gneisses erratics of southern Australia.

SiO2	Al2O3	K2O	Na2O	CaO	MgO	MnO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
62.6	18.7	16.2	0	0.1	0.1	0.4	0	0.02	0	98.1	0.5	0	99.5
61.6	18.1	16.8	0.4	0.1	0.3	0	0	0.2	0.7	98.7	0	2	98
62.1	17.9	16.5	0	0.5	0	0	0	0.2	0	97.9	2	0	98
61.8	17.6	16.5	0	0.3	0	0	0	0	0.2	97.5	2	0	98
61.6	17.8	15.9	0	0.4	0	0	0	0.2	0.2	97.4	2	0	98
61.2	17.8	16.1	0	0.5	0	0	0	0.2	0.2	97.2	2	0	98
62.4	18.1	15.9	0.4	0.2	0	0	0	0	0.4	97.7	1	2	97
62.2	17.4	15.6	0.3	0.1	0	0	0	0	0.2	96.9	0	3	97
61.1	17.9	16.1	0.4	0	0.1	0	0	0.2	0.4	96.3	0	3	97
61.8	17.3	17.2	0.4	0.5	0	0	0	0	0.4	97.8	2	2	96
61.6	18.3	15.8	0.2	0.4	0	0.2	0	0.2	0	97.8	2	2	96
61.5	17.4	15.7	0.4	0.1	0	0	0	0	0.2	96.8	0	5	95
62.1	17.8	15.4	0.5	0.3	0	0.1	0	0	0.3	98.2	1	5	94
61.2	17.6	15.2	0.6	0.6	0	0	0	0.3	0	96.6	2	4	94
61.4	17.5	16.3	0.7	0.3	0	0	0	0.4	0.3	98.1	1	5	94
61.7	17.9	16.1	0.8	0.1	0.4	0	0	0	0.1	97.5	0	7	93
62.1	17.6	15.4	0.9	0.4	0.1	0	0	0.4	0.1	97.2	2	7	91
61.2	18.2	15.5	0.5	0.9	0.5	0.1	0	0.5	0.2	98.4	5	5	90
61.4	17.6	15.9	0.5	0.5	0	0	0	0	0.2	97.1	2	5	93
61.4	17.7	15.4	0.9	0.2	0.1	0.1	0	0	0	97.4	1	7	92
62.1	17.9	15.6	0.5	0.8	0	0	0	0	0	97.6	5	5	90
63.1	17.8	1.3	11.7	3.6	0	0	0	0.4	0.3	98.5	1	32	68
61.9	18.2	16.4	0.5	0.1	0	0	0	0.5	0.2	98.7	4	89	7
60.2	23.8	0.4	8.2	6.1	0	0	0.1	0	0.5	99.3	27	70	3
61.2	23.3	0.4	9.1	4.5	0	0	0.6	0	0.4	99.9	22	76	2
64.1	21.1	0.2	10.5	1.3	0.5	0.2	0.2	0	0.03	98.1	5	94	1
64.1	19.9	0.2	10.6	1.2	0	0.4	0.5	0	0.1	97.1	5	94	1
63.1	21.1	0.1	10.3	2.2	0	0.3	0	0	0.3	97.6	10	89	1
64.2	19.7	0.1	11.2	1.2	0.3	0	0.1	0	0.2	96.9	5	95	0
63.8	21.1	0	9.8	1.9	0	0.3	0.3	0.03	0.07	97.9	10	90	0
63.5	21.1	0	9.8	2.1	0	0.4	0	0	0	97.1	10	90	0
63.4	20.1	0	11.1	1.4	0.2	0.3	0	0	0.3	97.3	7	93	0
63.3	21.6	0	9.6	2.4	0	0.5	0	0	0	97.8	13	87	0
63.1	17.7	15.9	0.3	0.5	0	0	0	0.3	0.3	98.1	2	2	96
63.4	19.1	16.5	0.7	0.6	0	0.1	0	0	0	100.4	2	5	93
62.7	22.5	0	8.9	3.7	0	0.2	0	0	0.1	98.1	15	85	0
62.4	22.6	0.3	8.9	3.4	0	0.1	0	0	0	97.7	16	82	2
62.8	19.1	15.1	1.1	0.3	0	0	0	0.3	0	98.7	1	10	89
63.3	23.2	0.1	9.1	3.7	0	0.2	0	0	0.2	99.8	18	81	1
62.4	22.3	0	9.9	3.2	0	0.2	0.2	0.1	0.2	98.5	15	85	0
61.8	18.3	15.1	1.4	0.6	0	0.4	0	0	0	97.6	2	12	86
62.5	23.1	0.1	9.4	3.7	0.2	0.3	0.2	0.2	0.4	100.1	18	81	1

Table. 5.2C. Composition of feldspar grains in the Permian volcanic erratics of southern Australia.

SiO2	Al2O3	K2O	Na2O	CaO	MnO	MgO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
74.1	12.5	10.9	0	0.5	0.1	0.2	0.1	0.9	0	99.3	3	0	97
65.9	17.8	11	2.8	0.8	0.1	0.2	0	0.7	0	99.4	4	26	70
65.1	21.8	0.5	10.5	2.1	0.1	0.5	0.1	0.1	0	100.8	9	88	3
64.9	17.9	14.1	1.7	0.7	0	0	0	0	0	99.5	3	15	82
64.4	18.1	15.5	0	0	0	0	0	0.2	0.2	98.6	0	0	100
64.4	21.8	0.7	8.9	3.4	0	0	0.3	0	0	99.7	15	79	5
64.1	21.1	0.5	9.3	2.3	0.1	0	0.9	0.5	0	98.8	11	86	3
64.1	18.1	16.7	0.4	0.9	0	0	0	0.4	0.1	100.6	5	2	93
64.1	18.1	16.7	0.4	0.9	0	0	0	0.4	0.1	100.6	5	2	93
64.1	18.1	14.9	0.8	0.3	0	0	0	0	0	98.3	1	7	92
63.8	18.8	15.9	0	1	0	0.1	0	0.3	0	99.9	5	1	94
63.8	18.1	16.6	0	0.9	0	0	0	0	0.2	99.5	5	0	95
63.8	22.5	0.3	8.9	3.1	0	0	0.4	0	0.2	99.2	16	82	2
63.7	17.5	16.6	0.5	0.4	0.1	0	0	0.4	0.5	99.7	2	5	93
63.7	23.1	0	9.2	3.4	0	0	0.2	0.1	0	99.7	16	84	0
63.7	18.1	16.5	0.7	0.6	0	0	0	0.1	0.5	100.4	2	4	94
63.7	21.4	0.9	9.1	3.8	0	0	0.1	0	0.2	99.3	17	78	5
63.7	18.1	16.5	0.7	0.6	0	0	0	0.1	0.5	100.4	2	4	94
63.6	18.4	15.4	0.5	0.5	0	0	0	0	0.3	99.2	2	8	90
63.6	18.1	16.5	0.7	0.7	0	0	0	0	0.2	99.6	2	4	94
63.4	22.9	0	9.2	4.1	0.4	0	0.8	0	0	100.8	20	80	0
63.4	22.7	0.1	8.5	4.1	0.1	0.2	0.5	0.1	0.2	99.9	19	80	1
63.3	18.1	15.6	1.4	0.8	0	0.3	0.1	0	0.5	99.8	3	11	76
63.3	22.3	0.09	8.6	4.3	0.1	0	0	0	0	98.8	21	78	1
63.3	22.9	0.2	9.5	3.8	0	0	0.3	0	0.3	100.2	18	81	1
63.3	18.1	15.6	1.4	0.8	0	0.3	0.1	0	0.5	99.8	3	11	76
63.2	18.8	11.8	3.1	0.4	0.1	0	0.1	0	0.5	98.1	2	28	70
63.1	18.2	15.8	0.7	0.9	0	0	0	0.6	0	98.9	4	6	90
63.1	18.2	15.9	0.7	0.5	0	0	0	0.2	0	98.9	2	5	93
63.1	18.7	15.7	0.5	0.2	0	0.4	0	0	0.2	99.2	1	5	94
63.1	18.2	15.8	0.7	0.9	0	0	0.1	0.6	0	98.9	4	6	90
63.1	18.7	15.7	0.5	0.2	0	0.4	0	0	0.2	99.2	1	5	94
62.9	17.8	16.4	0.5	0.3	0	0	0	0	0.2	98.2	1	5	94
62.8	18.3	16.4	0.7	0.7	0.3	0.1	0	0	0.1	99.4	3	6	91
62.8	18.2	16.6	0	0.8	0	0	0	0	0	98.4	5	0	95
62.6	18.3	15.9	0.9	0	0	0.2	0	0.2	0.1	98.4	0	7	93
62.6	18.7	16.6	0.1	0.6	0.3	0	0.1	0.3	0.2	99.5	2	1	97
62.3	18.2	16.2	0	0.9	0.3	0	0	0	0.2	98.1	5	0	95
62.3	19.1	15.2	0.8	1	0	0.5	0	0.5	0.4	99.8	5	7	88
61.7	23.2	0.3	9	5.3	0	0	0.1	0.5	0.3	100.1	23	74	2
60.7	23.4	0	8.7	5.9	0	0	0.7	0.1	0	99.8	27	73	0
60.1	25.1	0.2	6.9	7.6	0.2	0	0.1	0.2	0	100.5	36	62	2
60.1	25.7	0.3	6.2	7.5	0.2	0.2	0	0	0.3	100.4	39	59	1
58.9	24.9	0.4	7.6	7.6	0	0	0.2	0	0.1	99.9	36	62	2
58.7	25.3	0.3	6.5	7	0.1	0	0.1	0	0.3	98.3	36	61	3
58.4	26.1	0.1	6.8	7.7	0	0.2	0.3	0	0	99.7	39	60	1
58.4	26.4	0.4	6.4	7.9	0.3	0	0	0.1	0	99.9	39	58	3
58.1	25.9	0.3	7.4	7.7	0	0	0	0	0.4	99.7	36	62	2
57.4	26.2	0.1	6.3	8.5	0	0	0.5	0.1	0	99.1	42	57	1
74.1	12.5	10.9	0	0.5	0.1	0.2	0.1	0.9	0	99.3	3	0	97
65.9	17.8	11	2.8	0.8	0.1	0.2	0	0.7	0	99.4	4	26	70
65.1	21.8	0.5	10.5	2.1	0.1	0.5	0.1	0.1	0	100.8	9	88	3
64.9	17.9	14.1	1.7	0.7	0	0	0	0	0	99.5	3	15	82
64.4	21.8	0.7	8.9	3.4	0	0	0.3	0	0	99.7	15	79	5
64.1	21.1	0.5	9.3	2.3	0.1	0	0.9	0.5	0	98.8	11	86	3
64.1	18.1	16.7	0.4	0.9	0	0	0	0.4	0.1	100.6	5	2	93
64.1	18.1	14.9	0.8	0.3	0	0	0	0	0	98.3	1	7	92
63.8	18.8	15.9	0	1	0	0.1	0	0.3	0	99.9	5	1	94
63.8	18.1	16.6	0	0.9	0	0	0	0	0.2	99.5	5	0	95
63.8	22.5	0.3	8.9	3.1	0	0	0.4	0	0.2	99.2	16	82	2

Table. 5.2D. Composition of feldspar grains in the Permian arkosic erratics of southern Australia.

SiO2	Al2O3	Na2O	CaO	K2O	MgO	MnO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
63.1	17.5	0.4	0.7	16.6	0	0	0	0	0.1	99.9	2	2	96
61.7	17.9	0.7	0.9	15.6	0.1	0	0	0.3	0.4	99.7	5	5	90
63.2	17.9	0	0.5	17.2	0.4	0	0	0.1	0	100.1	2	0	98
64.2	21.5	9.9	2.7	0.1	0.1	0.1	0.4	0.4	0	99.2	13	86	1
65.7	20.7	10.1	1.3	0.2	0	0.1	0	0	0.3	99.6	5	94	1
68.2	19.1	10.8	0.1	0	0	0.1	0	0	0.5	99.1	1	99	0
60.7	18.2	0.8	0.2	14.7	0.1	0.2	0.7	0	0.3	99.3	1	8	91
67.1	19.7	11.1	0.1	0.1	0.4	0	0.4	0.1	0.1	99.6	0	99	1
63.1	17.5	0.4	0.7	16.6	0	0	0	0	0.1	99.9	2	2	96
61.7	17.9	0.7	0.9	15.6	0.1	0	0	0.3	0.4	99.7	5	5	90
63.2	17.9	0	0.5	17.2	0.4	0	0	0.1	0	100.1	2	0	98
64.2	21.5	9.9	2.7	0.1	0.1	0.1	0.4	0.4	0	99.2	13	86	1
65.7	20.7	10.1	1.3	0.2	0	0.1	0	0	0.3	99.6	5	94	1
68.2	19.1	10.8	0.1	0	0	0.1	0	0	0.5	99.1	1	99	0
60.7	18.2	0.8	0.2	14.7	0.1	0.2	0.7	0	0.3	99.3	1	8	91
67.1	19.7	11.1	0.1	0.1	0.4	0	0.4	0.1	0.1	99.6	0	99	1
63.1	17.5	0.4	0.7	16.6	0	0	0	0	0.1	99.9	2	2	96
61.7	17.9	0.7	0.9	15.6	0.1	0	0	0.3	0.4	99.7	5	5	90
63.2	17.9	0	0.5	17.2	0.4	0	0	0.1	0	100.1	2	0	98
64.2	21.5	9.9	2.7	0.1	0.1	0.1	0.4	0.4	0	99.2	13	86	1
65.7	20.7	10.1	1.3	0.2	0	0.1	0	0	0.3	99.6	5	94	1
68.2	19.1	10.8	0.1	0	0	0.1	0	0	0.5	99.1	1	99	0
60.7	18.2	0.8	0.2	14.7	0.1	0.2	0.7	0	0.3	99.3	1	8	91
67.1	19.7	11.1	0.1	0.1	0.4	0	0.4	0.1	0.1	99.6	0	99	1
68.2	19.1	10.8	0.1	0	0	0.1	0	0	0.5	99.1	1	99	0
67.1	19.7	11.1	0.1	0.1	0.4	0	0.4	0.1	0.1	99.6	0	99	1
65.7	20.7	10.1	1.3	0.2	0	0.1	0	0	0.3	99.6	5	94	1
64.2	21.5	9.9	2.7	0.1	0.1	0.1	0.4	0.4	0	99.2	13	86	1
63.2	17.9	0	0.5	17.2	0.4	0	0	0.1	0	100.1	2	0	98
63.1	17.5	0.4	0.7	16.6	0	0	0	0	0.1	99.9	2	2	96
61.7	17.9	0.7	0.9	15.6	0.1	0	0	0.3	0.4	99.7	5	5	90
60.7	18.2	0.8	0.2	14.7	0.1	0.2	0.7	0	0.3	99.3	1	8	91

Table. 5.3A. Composition of garnet grains in granitic gneiss samples in the Permian erratics of southern Australia.

SiO2	37.1	37.3	38.8	37.5	37.1	38.7	37.4	38.5	38.5	38.3	38.1	37.2	37.6	38.5	36.9	38.3	37.6	38.3	37.7
TiO2	0	0	0	0	0.1	0	0	0	0.1	0	0	0	0.3	0.3	0	0	0	0.1	0
Al2O3	21.8	22.2	21.7	22.1	22.3	21.9	22.1	21.5	21.5	22.1	22.9	22.1	21.1	22.7	21.4	22.3	22.7	22.7	21.9
FeO	24.3	25.9	24.9	24.3	25.3	24.3	24.1	25.1	24.5	24.1	24.1	24.6	24.7	24.1	25.4	25.1	24.1	25.4	24
MnO	7.5	6.2	6.8	6.1	5.9	6.3	6.2	5.8	5.1	5.1	5.4	6.4	7	6.1	6.3	6.8	6.4	5.8	7.6
MgO	8.1	6.6	7.2	8.2	7.2	7.9	7.3	8.7	8.8	8.4	8.8	8.5	8.1	7.7	7.6	8.1	7.5	7.3	7.7
CaO	0.9	0.9	1.1	0.9	1.1	0.9	1.1	0.8	0.9	1.1	0.9	0.9	1.2	1.3	0.9	0.9	0.8	0.9	1.1
Na2O	0	0.1	0	0	0	0.4	0	0	0	0	0	0	0	0.7	0	0	0.2	0	0.1
K2O	0	0	0	0	0	0	0	0	0	0	0	0	0.08	0.03	0	0.08	0.1	0	0
Cr2O3	0	0.3	0.1	0	0	0	0	0	0.1	0	0	0	0.04	0	0	0	0.1	0.1	0
Total	100.1	101.5	100.8	100.2	100.3	100.1	99.4	100.6	100.7	100.2	100.5	100.9	100.1	101.4	98.6	101.7	99.6	101.8	100.6

No. of ions on the basis of 24 (O)

Si	5.8	5.8	6	5.8	5.8	6	5.8	6	6	5.9	5.9	5.8	5.9	5.9	5.9	5.9	5.9	5.9	5.9
Ti	0	0	0	0	0	0	0	0	0.01	0	0	0	0.03	0.03	0	0	0	0.01	0
Al	4	4.1	4	4	4	4	4.1	3.9	4	4	4	4.1	3.9	4.1	4.02	4	4.2	4.1	4
Fe	3.2	3.4	3.2	3.1	3.2	3.1	3.2	3.2	3.1	3.2	3.1	3.2	3.2	3.1	3.4	3.2	3.1	3.2	3.1
Mn	1	1	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.8	0.7	0.8	0.9	0.8	0.8	0.9	0.8	0.7	1
Mg	2	1.5	1.8	1.8	1.8	1.8	1.8	2	1	2	2	1	1.9	1.7	1.8	1.8	1.7	1.7	1.8
Ca	0.1	0.2	0.2	0.2	0.1	0.1	0.2	0.1	0.1	0.2	0.1	0.1	0.2	0.2	0.1	0.1	0.1	0.1	0.2
Cr	0	0.02	0.01	0	0	0	0	0	0.01	0	0	0	0.004	0	0	0	0	0.02	0

Mol% end-members

AL%	51	57	54	54	54	55	54	52	52	52	54	54	51.6	53.4	55.7	53.3	54.4	56.1	50.8
SP%	16	17	13	13	13	13	13	13	13	13	10	14	14.5	13.8	13.1	15	14	12.2	16.4
PY%	32	32	30	30	30	30	30	32	32	32	33	30	30.6	29.3	29.5	30	29.8	29.8	29.5
GR%	1	3	3	3	3	2	3	3	3	3	3	2	3.2	3.4	1.6	1.7	1.7	1.7	3.3

Table. 5.3B. Composition of garnet grains in the granitic samples from Permian erratics of southern Australia.

SiO2	37.6	35.7	36.1	37.7	40.7	39.6	35.8	36.4	37.3	36.1	36.1	36.1	36.4	36.6
TiO2	0	0.1	0.1	0	0	0.1	0.8	0.2	0	0	0.3	0.3	0.1	0
Al2O3	21.9	18.1	20.2	21.8	23.5	23.1	20.9	21.4	22.1	18.1	21.5	20.7	21.1	21.5
FeO	18.2	8.1	7.9	30.2	20.9	20.9	14.3	14.4	26.3	8.2	15.3	7.9	5.1	6.9
MnO	18.7	30.8	26.3	2.1	0.8	0.8	17.8	18.6	8.5	31.1	18.1	26.6	30.4	25.9
MgO	2.8	1.3	0	6.2	12.9	12.4	0.6	0.4	4.1	1.1	0.5	0.1	0.8	4.4
CaO	2.1	2.7	5.8	1.4	3.3	3.4	6.6	7.1	2.5	2.8	6.6	5.6	4.4	2.1
Na2O	0.1	0.8	0	0.3	0.1	0	0.5	0.4	0	0.4	0.2	0.5	0.8	0.8
K2O	0	0.2	0	0	0	0	0	0.1	0	0.2	0	0	0.1	0.2
Cr2O3	0	0.1	0	0.2	0.1	0.1	0	0	0	0	0	0	0	0.1
Total	101.5	99.1	97.8	100.1	101.3	100.6	98.2	99.1	100.6	98.7	98.5	98.1	99.3	98.7

No. of ions on the basis of 24 (O)

Si	5.9	6	6	5.9	5.9	5.9	5.9	5.9	5.9	6.1	5.9	5.9	5.9	5.9
Ti	0	0.01	0.01	0	0	0.01	0.1	0.03	0	0	0.04	0.03	0.01	0
Al	4.1	3.6	4	4.1	4.1	4.1	4	4.1	4.1	3.6	4.1	4	4	4.1
Fe	2.4	1.1	1.1	3.9	2.6	2.6	2	2	3.5	1.2	2.1	1.1	0.7	0.8
Mn	2.5	4.4	3.7	0.3	0.1	0.1	2.5	2.6	1.1	4.4	2.5	3.7	4.2	3.6
Mg	0.6	0.3	0	1.5	2.8	2.8	0.1	0.1	1	0.3	0.1	0.03	0.2	1.1
Ca	0.3	0.5	1	0.2	0.5	0.5	1.2	1.2	0.4	0.5	1.1	1	0.8	0.4
Cr	0	0.02	0	0.01	0.01	0.01	0	0	0	0	0	0	0	0.01

Mol% end-members

Al	41.3	17.5	19	66.1	43.3	43.3	34.4	33.9	58.3	18.7	36.2	18.8	11.9	13.6
As	43.1	69.8	63.8	5.1	1.7	1.7	43.1	44.1	18.3	67.5	43.1	63.5	71.2	61
P	9.7	4.8	0	25.4	46.7	46.7	1.7	1.7	16.7	4.7	1.7	0.5	3.4	18.6
G	5.2	7.9	17.2	3.4	8.3	8.3	20.7	20.3	6.7	7.8	18.9	17.2	13.6	6.8

Tabl. 5.4A. Composition of ilmenite grains in the Permian gneissic erratics of southern Australia.

TiO2	FeO	MnO	MgO	V2O3	Al2O3	Cr2O3	SiO2	Total
53.7	42.3	3.4	0.4	0.2	0.4	0	0	100.2
53.5	40.9	4.6	0	0.1	0	0	0	99.3
53.2	43.4	2.7	0.2	0.2	0	0.3	0	101
53.2	43.4	2.7	0.2	0.2	0	0.3	0	100.4
53.1	41.8	4.5	0.1	0.4	0.2	0	0	100.3
53.1	40.1	4.7	0	0.2	0	0	0	98.06
52.7	42.3	3.4	0.4	0.3	0.4	0	0	99.4
52.6	42.8	1.5	0.2	0.3	0	0	0	97.4
52.6	43.7	2.9	0	0.2	0	0.1	0	99.8
52.6	43.7	2.9	0	0.2	0	0.1	0	99.8
52.4	44.9	2.4	0.3	0.2	0.2	0	0	100.6
52.3	44.5	1.1	0.2	0.3	0	0	0	98.4
52.1	40.1	4.9	0.3	0.2	0	0.3	0	99.7
52.1	44.9	2.7	0	0.2	0.2	0	0	100.4
52.1	44.9	2.7	0	0.2	0.2	0	0.1	100.4
51.9	43.3	2.5	0.3	0.3	0.4	0.1	0.2	99.1
51.9	43.3	2.5	0.3	0.2	0.4	0.1	0.2	98.8
51.6	43.2	4.4	0	0.2	0	0	0	99.3
51.5	42.6	4.9	0	0.3	0.1	0	0	99.9
51.5	42.3	4.6	0.4	0.3	0	0	0	98.9
51.3	41.9	3.1	0.6	0.2	0	0	0	99.5
51.2	44.3	2.7	0.1	0.2	0.2	0	0	98.6
51.1	42.6	4.9	0	0.3	0.1	0	0	98.9
51.1	41.6	4.5	0.4	0.3	0	0.5	0	99.9
50.9	43.8	4.6	0.2	0.2	0.1	0	0	101.1
50.9	43.7	4.9	0.1	0.3	0	0	0.1	99.8
50.9	40.9	4.9	0.2	0.2	0	0	0	98.3
50.9	44.7	3.2	0.3	0.3	0	0.1	0	99.6
50.9	44.7	3.2	0.3	0.3	0	0.2	0	99.6
50.8	41.9	4.5	0.4	0.3	0.6	0	0.5	99.7
50.7	42.5	5.2	0	0.2	0.1	0	0	99.4
50.7	42.5	5.3	0	0.2	0.1	0	0	99.4
50.4	44.1	4.1	0.5	0.2	0.1	0	0	99.8
50.4	44.1	4.1	0.5	0.3	0.1	0	0	99.9
50.3	43.8	4.4	0.3	0.2	0	0	0	99.2
50.2	44.2	3.5	0	0.2	0	0	0	98.4
50.2	44.2	3.5	0	0.2	0	0	0	98.4
50.2	45.7	2.3	0	0.3	0	0	0	99.5
50.2	43.1	5.5	0.4	0.2	0.2	0.3	0	100.2
50.1	44.3	4.1	0.8	0.2	0.2	0	0	100.2
50.1	43.3	5.5	0	0.3	0	0	0	99.9
49.8	46.8	1.1	0.3	0.3	0	0	0	98.3
48.7	44.6	4.1	0.4	0.3	0.1	0	0.1	98.7

Table. 5.4B. Composition of ilmenite grains in the Permian granitic erratics of southern Australia.

TiO2	FeO	MnO	MgO	V2O3	Al2O3	Cr2O3	SiO2	Total
53.1	40.1	4.7	0	0.2	0	0	0	98.1
52.6	42.1	4.9	0.6	0.3	0.4	0	0	100.8
52.6	42.8	1.5	0.2	0.3	0	0	0	97.4
52.6	48.9	1.2	0	0	0	0	0	98.2
52.3	44.5	1.1	0.2	0.3	0	0	0	98.4
51.8	42.7	3.8	0.3	0.2	0.3	0	0	100.3
51.6	40.3	6.2	0.7	0.2	0	0.2	0.2	99.9
51.6	43.2	4.4	0	0	0	0.2	0	99.3
51.6	48.9	1.2	0	0	0	0	0	98.2
51.4	40.1	6.4	0.2	0.2	0.2	0	0	99.3
51.4	38.5	7.5	0.1	0.3	0.1	0	0.4	98.3
51.4	48.4	0.7	0	0	0	0	0.1	98.2
51.1	41.6	4.5	0.5	0	0.4	0.1	0	99.9
50.9	41.7	7.2	0.5	0.2	0.2	0	0.1	101.3
50.9	41.7	7.2	0.5	0.3	0.2	0	0.1	101.3
50.9	48.4	0.7	0	0	0	0	0.1	98.2
50.9	46.9	1.9	0.7	0	0.4	0	0	99.6
50.8	41.9	4.5	0.3	0.6	0.4	0.3	0.5	99.7
50.5	43.5	4.8	0.2	0.3	0.3	0.2	0	99.7
50.5	42.6	4.5	0.4	0.3	0.5	0	0	99.3
50.3	43.8	4.4	0.1	0	0.3	0	0	99.2
50.2	43.1	5.5	0.4	0.2	0.2	0.3	0	100.2
50.2	45.1	1.5	0.3	0	0.4	0	0	98.2
50.1	44.1	3.7	0.2	0.2	0	0	0.1	98.1
50.1	43.3	5.5	0.3	0	0	0.3	0	99.9
49.8	46.8	1.1	0.3	0.3	0	0	0	98.3
49.8	43.3	3.3	0	0.2	0.3	0.3	0.2	97.5
48.8	48.4	0.7	0	0.2	0	0.1	0	98.2
48.8	48.4	0.7	0	0.3	0	0.1	0	98.2
48.8	43.1	5.5	0.4	0.1	0.2	0.1	0.3	100.2
48.8	42.1	4.9	0.6	0.1	0	0	0	100.8
48.7	45.1	1.5	0.3	0.2	0	0	0.2	98.2
48.7	44.6	4.1	0.3	0.1	0.4	0	0.1	98.7
48.7	45.1	1.5	0.4	0	0.3	0	0.2	98.2
48.7	41.7	7.2	0.5	0	0.4	0	0	101.3
48.5	46.4	3.7	0.4	0.3	0	0.3	0	99.5
48.2	46.9	1.9	0.7	0.2	0.3	0	0.3	99.6
48.2	41.7	7.2	0.5	0	0.4	0	0	101.3
47.9	45.9	4.1	0.5	0.2	0	0.2	0	98.9

Table. 5.4C. Composition of ilmenite grains in the Permian volcanic erratics of southern Australia.

TiO2	FeO	MnO	MgO	V2O3	Al2O3	Cr2O3	SiO2	Total
54.7	39.7	3.4	0	0.2	0	0	0.1	99.6
54.1	39.1	4.5	0.5	0.3	0.2	0.3	0.2	99.7
53.5	40.9	3.2	0.6	0.3	0.2	0.1	0.2	99.7
53.3	43.3	3.1	0.6	0.3	0	0	0	100.9
52.5	42.9	2.2	0.4	0.3	0.2	0.1	0.3	100.4
52.3	43.7	3.1	0.4	0.3	0	0	0	100.2
51.7	43.1	3.4	0.3	0.3	0.1	0	0	98.9
51.6	45.4	3.3	0.3	0.2	0	0	0	100.8
51.5	42.3	2.9	0	0.2	0.7	0.6	0.4	99.6
51.5	41.5	5.2	0.1	0.2	0.2	0	0	99.1
51.4	45.5	3.2	0	0.2	0	0.6	0	100.6
51.4	43.7	3.8	0	0.2	0.4	0.2	0	100.3
51.2	43.6	3.6	0.7	0.3	0.3	0	0	99.6
51.2	43.6	3.1	0.4	0.3	0.1	0	0.1	98.6
51.1	46.8	2.5	0.2	0.3	0.4	0	0.2	100.9
50.8	44.9	3.4	0	0.2	0	0	0	100.4
50.5	44.9	2.8	0.1	0.2	0	0	0	99.4
50.5	44.4	4.1	0.5	0.2	0.7	0	0.2	100.8
50.4	43.9	2.8	0.9	0.3	0	0.2	0	98.6
50.3	45.6	3.9	0.4	0.3	0.1	0	0	100.9
50.2	45.2	2.6	0.8	0.3	0	0	0.8	100.2
50.1	46.3	3.1	0	0.2	0.2	0	0.4	100.7
50.1	45.2	2.6	0.4	0.3	0	0.3	0.3	99.3
49.8	45.8	3.7	0.1	0.2	0	0.1	0.1	100.1
48.4	47.2	3.5	0	0.2	0.6	0.3	0.1	100.4
48.2	47.3	2.9	0.3	0.2	0.4	0	0	100.3

Table. 5.5. Composition of rutile grains in the Permian gneisses erratics of southern Australia.

TiO2	FeO	MnO	MgO	Cr2O3	Al2O3	CaO	Na2O	K2O	SiO2	Total
99.9	0.2	0	0.1	0	0.1	0.1	0	0	0.1	100.5
99.7	0.4	0.2	0.1	0.2	0.1	0.3	0	0	0.1	101.1
99.3	0.6	0	0.2	0	0.3	0.1	0.1	0	0.2	100.9
99.3	0	0.4	0.2	0	0	0.1	0	0	0.1	100.1
99.3	0	0.5	0.2	0	0.1	0	0	0	0.1	100.2
99.3	0	0.5	0.1	0	0	0.1	0	0	0.1	100.1
99.3	0.9	0.1	0	0	0	0	0	0.7	0.1	101.2
99.2	0.3	0.3	0.1	0	0.1	0	0	0	0.1	100.1
99.2	0.1	0.1	0	0.2	0	0	0.1	0	0	99.7
99.1	0.8	0	0.5	0.3	0	0.5	0	0	0.2	101.4
99.1	0.8	0	0.5	0.3	0.5	0	0	0	0.2	101.4
99.1	0.3	0.3	0.1	0	0	0.1	0	0	0.1	100.1
99.1	0.2	0.3	0.2	0	0.1	0	0	0	0.1	100.1
99.1	0.2	0.1	0	0.2	0	0	0.5	0	0	100.1
99.1	0.4	0.3	0	0.3	0	0.7	0	0	0.1	99.9
98.6	0.1	0	0	0	0	0.4	0	0.2	0	99.4
98.6	0	0.2	0	0	0	0	0	0	0	98.8
98.6	0.4	0	0	0	0	0.3	0.1	0	0	99.5
98.5	0.9	0	0	0	0	0.1	0	0.1	0	99.6
98.4	0.9	0	0.4	0	0	0.2	0.1	0.1	0	100.1
98.4	0	0	0	0	0	0	0	0.2	0	98.6
98.3	0.5	0	0.3	0.1	0	0.2	0	0.3	0	99.8
98.2	0.1	0.4	0	0.2	0	0.2	0	0.3	0.2	99.3
98.1	0	0	0	0	0	0.4	0.1	0	0	98.2
98.1	0	0.3	0	0	0	0.3	0	0.4	0	99.3
98.1	0.4	0	0	0	0.1	0.4	0	0.3	0	99.2
98.1	0.3	0.2	0	0.4	0.1	0.1	0	0	0	99.1
97.9	0.1	0.2	0	0	0	0.3	0	0	0	98.7
97.9	0.1	0.5	0	0.2	0	0.5	0.2	0	0	99.4
97.9	0.1	0.2	0	0	0	0.3	0	0	0	98.5
97.8	0.1	0	0.1	0.6	0.2	0.1	0	0.1	0	99.6
97.8	0.4	0	0.2	0.5	0	0	0.3	0	0.1	99.4
97.8	0.4	0	0.2	0.5	0	0	0.3	0	0.1	99.3
97.7	0.1	0	0.3	0	0	0.5	0	0.3	0	99.1
97.5	0.3	0	0.3	0.1	0	0.1	0.3	0.1	0	99.1
97.5	0.3	0	0.3	0.1	0	0.1	0.3	0	0	98.6
97.4	1.2	0.4	0	0.2	0.3	0.4	0.1	0.1	0	100.1
97.3	1.2	0.4	0	0.2	0.3	0.4	0.1	0.1	0.1	100.1
97.3	0.3	0	0.2	0.2	0.1	0.1	0	0	0	98.5
97.2	0.1	0	0	0.2	0.1	0	0.2	0	0	98.9
97.2	0.8	0.1	0.1	0.2	0	0.4	0.1	0.3	0.1	99.3
97.2	0.9	0	0.4	0.2	0.1	0.2	0.1	0	0	99.1
97.1	0.9	0	0.4	0.2	0.1	0.2	0.1	0	0	100.1
97.1	0.8	0.5	0.3	0.2	0	0	0.3	0.1	0.5	99.8
97.1	0.3	0.1	0	0	0.1	0	0.1	0	0	98.1
96.7	1.5	0.4	0.1	0.1	0	0	0	0	0	98.8
96.5	1.6	0	0.4	0	0.1	0.2	0.7	0	0	99.5
97.5	0	0	0	0	0	0	0.4	0	0	97.9
98.7	0.2	0	0	0	0	0	0.2	0	0	99.1
98.3	0.1	0	0	0	0	0	0	0	0	98.4

Table. 5.6. Composition of magnetite grains in the Permian erratics of southern Australia.

Fe2O3	FeO	Al2O3	V2O3	Cr2O3	MnO	MgO	TiO2	SiO2	Total	
86.1	9.6	0	0	0.1	0.3	0	2.8	0	98.6	*
83.3	11.5	0.3	0.1	0	0.1	0	2	0.1	97.4	
91.8	9.1	0.1	0.3	0.1	0.1	0	0	0	100.9	
91.8	7.2	0.1	0.2	0.1	0.2	0	0	0	100.4	
91.9	8.2	0.1	0.2	0	0	0	0	0	101.4	
91.5	8.1	0.2	0.1	0.1	0.1	0	0	0	99.9	**
92.2	8.3	0.1	0.3	0.2	0.1	0	0	0	101.5	
91.5	8.6	0	0.3	0.1	0.1	0	0	0	101.6	
80.4	9.1	1.2	0.2	0	3.2	0	9.3	0	99.7	
83.1	9.2	1	0.2	0	0.2	0.1	9.2	0	98.3	
81.9	8.2	1.7	0.1	0	2.7	0	11.1	0	98.7	
81.1	9.1	0.8	0.1	0	1.1	0	8.6	0	99.3	
72.9	8.1	1.1	0.2	0	3.7	0	9.9	0	99.4	***
76.2	8.5	1.9	0.2	0	4.1	0	11.1	0	99.1	
81.6	9.1	1	0.1	0	3.1	0	8.4	0	99.1	
72.5	12.7	1.7	0.2	0	4.1	0	10.9	0	99.3	
72.4	9.1	0.9	0.2	0	3.1	0	9.8	0	99.7	

*	**Plutonics**
**	**Volcanics**
***	**Metamorphics**

Table. 5.7. Composition of sphene grains in the Permian volcanic erratics of southern Australia.

SiO2	TiO2	CaO	Al2O3	FeO	MgO	MnO	Na2O	K2O	Cr2O3	Total
29.2	32.8	27.2	4.8	1.5	0	0.7	0	0	0	96.8
29.1	31.9	29.1	5.1	1.4	0.2	0	0	0	0	97.1
29.4	35.5	28.3	2.5	1.9	0	0.3	0.1	0	0	98.1
29.8	33.7	28.4	3.1	1.9	0	0.6	0.1	0.1	0	97.6
29.8	31.9	28.2	5.1	1.8	0.2	0.5	0.1	0.1	0	97.7
30.1	33.1	27.7	4.1	1.3	0	0.4	0.5	0	0	97.3
29.5	35.8	27.6	2.4	1.6	0.1	0.1	0.2	0	0	97.5
29.5	34.6	28.8	3.9	0.7	0.6	0.5	0.1	0	0	98.7
29.9	35.8	28.1	3.9	1.2	0	0	0.3	0	0.6	99.8
31.1	32.7	28.1	4.9	0.2	0.8	0.1	0.5	0.5	0.4	99.3

CHAPTER SIX
LIGHT MINERAL ANALYSIS OF THE PERMIAN SANDS

INTRODUCTION

The provenance of sediments may be determined in many ways. A mineralogical analysis of the sediments is useful for the determination of the source area (Zandstra 1983). This is important for palaeogeographical reconstructions because it allows transport routes to be found (eg. Di Labio and Shilts 1979). Unfortunately, minerals most likely to defy destruction by weathering, transport, and diagenesis tend to be rare in primary igneous or metamorphic rocks that are often the sources of sediment material. The only common resistant light mineral is quartz. Feldspar comes a poor second.

The overall mineral content of sand provides a direct estimate of the nature of the source rocks. Sands of fluvioglacial environments associated with continental glaciations provide a direct insight into the geology of the glaciated area, for areas larger than that of a single drainage basin. This chapter deals with the general mineral content of the Permian sands of Fleurieu Peninsula. It then focuses on details of the quartz types including their optical characteristics, CL and trace elements signatures. Finally feldspar types are analysed and discussed.

Sample preparation_

At least one sample was collected from each sand bed unit at different localities of Permian outcrop. All the specimens cited in this thesis are held within the Geology and Geophysic Department of Adelaide University under accession number 759. The sand size fraction (2 mm to 0.063 mm) was removed by wet sieving into three size fractions, and light minerals were separated from the heavy. The prepared samples were mixed with Epoxy and allowed to harden. Thin sections were cut from these artificially cemented sands and stained for potassium feldspar. [see Appendix. 2 for details]

Method of study

Approximately 60 representative samples were chosen from Permian sands for detailed petrologic study.

Detailed microscopic analysis of specimens were carried by point counter. Permian sand point counts were performed using the procedures outlined by Dickinson (1970), and discussed by Gazzi et al. (1973) and Ingersoll et al. (1984). At least 300 framework grains were identified and counted per-sample at a spacing of 0.5mm. Probable errors in counting and curves for their estimation are presented by Krumbein and Pettijohn (1938). They showed that counting of 300 to 400 grains gives a satisfactory degree of accuracy. Each Permian sand of southern Australia was counted for Q (quartz), F (K feldspar and plagioclase) and L (lithic fragments). As quartz was the most abundant light mineral in the Permian sands, 300 quartz grains per-thin section were identified and counted as either non-undulose monocrystalline quartz, undulose monocrystalline quartz, polycrystalline quartz with two to five crystals per-grain, or polycrystalline quartz with greater than five crystals per-grain. Polycrystalline quartz fragments were noted as either bimodal or unimodal in character.

COMPOSITION OF THE PERMIAN SANDS

There are various ways of evaluating the composition of the sediments. The composition of the Permian sands is shown in a triangular diagram similar to the one that was established by Dickinson (1970) which has three end-members of quartz, feldspars and rock fragments. In the past this type of triangular diagram has been used for interpretation of sand composition by many workers such as McBride (1962), Folk (1980), Casshyap (1969), Okada (1971) and Pettijohn (1972).

Thin sections of the light minerals have been point counted and the following components have been identified: quartz, feldspars and rock fragments. In all Permian sands quartz is the dominant grain (average 91%) and the remaining constituents are feldspars 8.3% and rock fragments and accessories 1% (Q91F8L1). Composition of the

Permian sands are presented in (Fig. 6.0) and sumarised in (Table. 6.0). As the division between arkosic and subarkosic sand is drawn at the standard value of 25% feldspars, where a Permian sand has 8.3% feldspar the term sub-arkosic sand is appropriate.

Quartz

Details of quartz grain types are discussed later in this chapter. However, in all sandy lithotypes, both monocrystalline and polycrystalline quartz grains are important constituents. The mean percentage of quartz grains in the Permian sands of southern Australia is 90.4% which includes 61.7% monocrystalline and 38.3% polycrystalline quartz (Table. 6.0). The ratio of monocrystalline to polycrystalline quartz is 1.6. The dominance of monocrystalline quartz grains within the Permian sand suggests that prolonged abrasion has taken place and destroyed the polycrystalline quartz grains (see chapter 3 on surface texture). The shape of monocrystalline quartz grains vary widely, and range from rounded to subangular . The former variety is mostly undulatory, and nearly 75% show undulose extinction. Inclusions are occasionally seen within the quartz grains, but where inclusions are present, they consist of zircon, tourmaline, rutile, garnet and opaques. Intracrystalline boundaries in polycrystalline quartz grains size distribution in these sands vary from sutured to straight. The size distribution of the polycrystalline quartz grains ranges from unimodal to bimodal; about 87% of the polycrystalline quartz grains in the Permian sands have a unimodal size distribution.

Feldspars

Feldspar is less abundant and makes up only 8.3% of the Permian sands (Table. 6.0). It consists mostly of K feldspar with smaller amounts of plagioclase. K feldspar includes microcline and orthoclase. The plagioclases commonly have an albitic composition. Variation of feldspar content in the Permian sands at different localities is very low and ranges from 7.2% at Cape Jervis, 8.7% at Hallett Cove and 9.1% at Kings Point. The dominant potassium feldspars are almost fresh and show cross hatched twinning (Plate. 7E and 7F). This freshness indicates either a cold or dry climate and may suggest proximity to source rocks. Microcline is in fact, one of the low temperature feldspars more resistant to weathering and alteration (Pettijohn 1975), and this feature could account for its abundance in the Permian sands. The ratio of K feldspar to Na feldspar ranges from 1.8 to 2.6 and indicates the abundance of K feldspar in the source rocks, or it may be more resistant than Na feldspar. Potassium feldspar is more resistant to glacial comminution (Krauskopf 1979 p.86). Haldorsen (1977) reported that Na feldspar in till from southeastern Norway is very susceptible to glacial comminution because of the presence of fractures.

The increasing abundance of K feldspar in the Permian sands seems to be the result of the following factors: (i) glacial transport where K feldspars have greater resistance to glacial comminution than Na feldspar, (ii) source rocks that have more K feldspar than Na-Ca feldspars, and (iii) the fact that K feldspar is more resistant to weathering and alteration than Na feldspar. Calculations of plagioclase to total feldspar P/F ratios from data for sand and sandstone given in Dickinson and Suczek (1979) give mean P/F values of 0.39 for sand of continental block provenance, 0.47 for sand of foreland uplift provenance, and 0.66 for sand of collision orogen provenance, and 0.81 for sand of magmatic arc provenance. The plagioclase to total feldspar ratio of the Permian samples was studied and has a mean of 0.7. This value is lower than that for most sand derived exclusively from volcanic terranes and is fairly typical of sand derived from mixed plutonic and metamorphic terranes of granitic composition (Dickinson 1970; Dickinson and Rich 1972; Ingersoll 1978).

ACCESSORIES

Micas, pyroxene and amphibole are present within all the Permian sands. They make up usually less than 0.1% of total minerals. Volumetrically they are unimportant because these minerals are not stable and they will alter during weathering and transportation of the sediments. Heavy minerals are also present in the Permian sands (see the following chapter).

ROCK FRAGMENTS

Lithic rock fragments are only present in trace amounts and include chert, granite, gneisses and porphyritic fragments; however the mean percentage of rock fragments is

1%. Volcanic lithic rocks occur in low percentage but granitic and metamorphic lithic rock fragments, such as phyllites and gneisses, occur in highly variable amounts.

Q monocrystalline 61.7%	Q polycrystalline 38.3%	Total quartz 90.4
Na feldspar 2.2%	K feldspar 6.1%	Total feldspars 8.3
Rock fragments 1%	Accessories 0.1%	Total 1.1

Table.6.0. Mean percentage of detrital quartz, feldspars and rock fragments in the Permian sands of southern Australia.

QUARTZ TYPE ANALYSES

Quartz is the most abundant and durable mineral in all sediments representing almost all common parent rocks. Durability and abundance of quartz also assure that nearly all parent rocks containing quartz are represented by detrital quartz in their daughter sediments. According to Basu (1985), quartz has the greatest potential of all detrital minerals for reading the provenance of sediments. Therefore, study of detrital quartz, at least theoretically, can provide an insight to the ultimate sources of arenites.

Quartz occurs in sandstones as monocrystalline and polycrystalline or composite grains that may be derived from plutonic, volcanic and metamorphic rocks or recycled from sediments. Since Sorby (1880), first attempted to identify in thin sections features of quartz from different source rocks, numerous attempts were made to use quartz as a guide to provenance such as Mackie (1896), Gilligan (1920), and Dake (1921). Krynine (1940, 1946) did a comprehensive study based on grain shape, inclusions, and extinction. Other studies in grain shape include Ehrlich and Weinberg (1970), Brown et al. (1980) and Hudson and Ehrlich (1980). Pettijohn (1949), Keller and Littlefield (1950), Bokman (1952), Blatt and Christie (1963) Conolly (1965), Folk (1980) Blatt (1967a,b) studied undulatory extinction and the shape of polycrystalline quartz from known source rocks. Basu et al. (1975) first established a new method of distinguishing source areas by the analysis of quartz grain types. They studied quartz from known source rocks and measured the undulosity of monocrystalline quartz, and coupled that with the quantity of polycrystalline quartz and number of crystal units per-grain. This led to the distinction between plutonic, low rank and high rank metamorphic rocks. This method of Basu et al. (1975) has been applied in this thesis. With regard to provenance, early studies by Mackie (1896) divided the quartz grains of the sands and sandstones of eastern Moray Scotland into four groups on the basis of their inclusions:

1- Regular inclusioned

2- Circular inclusioned

3- Irregular inclusioned

4- Inclusionless

Keller and Littlefield (1950) examined inclusions in the quartz of 35 igneous and 16 metamorphic rocks and (with slight modifications) agreed with Mackie's findings. Keller and Littlefield (1950), Bokman (1952), Doty and Hubert (1962) worked on the grain shape and inclusions of quartz grains, and have deduced significant information about the nature of their source rocks. Dake (1921) and Feniak (1944) have published measurements of the size of quartz grains in igneous and metamorphic rocks. Blatt and Christie (1963) studied thin sections of 119 igneous and metamorphic rocks, and 44 clastic rocks, in an attempt to evaluate the use of undulatory extinctions and polycrystallinity in detrital quartz grains for the determination of provenance of sediments. Bokman (1952) used the elongation of quartz grains for the source of this mineral in the clastic rocks.

Blatt, Middleton and Murray (1980) indicate that the average number of crystal units in sand size grains of polycrystalline quartz varies depending on the source rocks of the quartz. In general, they note that polycrystalline quartz in plutonic rocks has a more coarser crystal size units (2-5 crystal units per grain) than polycrystalline quartz from gneissic rocks (five crystal units per-grains). According to Tortosa et al. (1991) and Young (1976) who studied thin sections of 122 metamorphic rocks and medium sand size Holocene fluvial sand, and suggested that detrital polycrystalline quartz derived from low grade metamorphic rocks are characterised by high polycrystallinity indices, and those from medium to high grade metamorphic rocks have moderate to low polycrystallinity indices. Polycrystalline quartz grains are also directly useful in the interpretation of source rocks. Folk (1980) has identified a continuum of undulosity and polycrystallinity in detrital quartz grains from different source rocks. Folk continued to reiterate Krynine's faith in the shape of quartz grains, especially that of the subgrains in polycrystalline quartz as an indicator of source rock. Currently however, the cathodoluminescence of quartz is receiving much attention and used for characterisation of source area lithology (Matter & Ramseyer 1985; Owen and Carozzi 1986; Baker 1988).

CLASSIFICATION OF QUARTZ GRAINS

Study of thin sections of the Permian sands, has shown that all three distinctive types of quartz grain are present. These are very similar to what has been reported by other workers in other areas, see (Krynine 1940, 1946; Hubert 1960; Blatt and Christie 1963; Folk 1980). The three distinctive quartz types are:

1- Nonundulatory monocrystalline quartz

2- Undulatory monocrystalline quartz

3- Polycrystalline quartz

MONOCRYSTALLINE QUARTZ GRAINS

Monocrystalline quartz is the most abundant detrital mineral, averaging 61.7% of the detrital grains in the Permian sands. According to Blatt and Christie (1963) monocrystalline quartz grains occur in both igneous and metamorphic rocks. The monocrystalline quartz grains were separated into non-undulose and undulose varieties using the method of Basu et al. (1975).

Non-undulatory quartz grains: Blatt and Christie (1963) define non-undulatory quartz grains as quartz crystal units showing no visible evidence of strain. Non-undulatory monocrystalline quartz grains are more common in the Permian sands and their abundance is about 52% of total quartz grains.

Undulatory monocrystalline quartz: This refers to monocrystalline quartz grains showing visible evidence of being strained. Undulatory extinction was first used by Mackie (1896), Basu et al. (1975) to distinguish the source rock type.

The presence of undulatory monocrystalline quartz in the Permian sands could have been formed by the disaggregation of polycrystalline quartz during transportation of the sediments from a long distance. This hypothesis is also supported by the fact that polycrystalline undulose quartz is not stable when it is transported and soon breaks down to monocrystalline grains with undulatory extinction (Blatt and Christie 1963). The abundance of undulatory quartz is about 48% in the Permian sands.

POLYCRYSTALLINE QUARTZ GRAINS

Polycrystalline quartz grains are those containing two or more crystal units with different optical orientations.

The most common type of polycrystalline quartz appears to contain less than five sub-grains. The boundaries between adjacent crystal units vary from sutured to straight, and most polycrystalline quartz grains in the Permian sands are undulatory. However, various kinds of polycrystalline quartz grains are present in Permian sands, including those with sutured subgrain boundaries and strong undulatory extinction, and those with straight boundaries and less undulatory extinction. Such variations in quartz types are not uncommon in granitic rocks (Basu et al. 1975). The boundaries between subgrains of polycrystalline quartz may be used as indicator for the source of sediments. Sutured boundaries generally indicate metamorphic sources (Wilson 1973; Folk 1980). The abundance of polycrystalline quartz grains in sedimentary rocks has been generally accepted as indicating a derivation from gneissic rocks (Blatt 1963). Polycrystalline quartz averages 38.3% of the detrital grains in the Permian sands.

NUMBER OF CRYSTAL UNITS IN POLYCRYSTALLINE QUARTZ GRAINS

Blatt, Middleton and Murray (1980) and Basu et al. (1975) showed that the average number of crystal units in a single polycrystalline quartz grain varies depending upon the source rock of the quartz and the grain size of the sediment (Blatt 1967a, Fig. 23).

Polycrystalline quartz is categorised as consisting of grains that contain two to five crystals per-unit, and grains that are composed of more than five crystals per-unit. The ratio of monocrystalline to polycrystalline quartz grains in the Permian sands is 1.6, so that the amount of polycrystalline quartz in these sands in comparison with the amount of polycrystalline quartz from different kinds of rocks suggests probably a predominantly granitic source.

In this study the number of crystal units in polycrystalline quartz grains of the Permian sands have been compared with those from Permian erratics, Antarctic gneiss and Encounter Bay Granite. The results have been shown in Fig. 6.1 and Table 6.1. As shown in these figures the greatest abundance of polycrystalline quartz grains with more than five components occur in the metamorphic samples from erratics and from East Antarctica. This suggests that most of the polycrystalline quartz grains in the Permian sands have a plutonic source, rather than metamorphic source. The size distribution of polycrystalline quartz in these sand grains is mostly unimodal. The percentage of different size distribution of polycrystalline quartz was determined for the Permian sands. It has been shown that about 76.8% of the polycrystalline quartz grains are unimodal and 23.2% are bimodal. This indicates that the source of polycrystalline quartz was plutonic. This is very similar to what has been described by Blatt (1967a); viz, that the distinct bimodality of crystal size in polycrystalline quartz suggests a metamorphic and unimodality suggests a plutonic source.

ELONGATION OF QUARTZ GRAINS

The form of quartz grains may give a clue to their derivation. Quartz elongation principally has been used by Bokman (1952) to interpret the provenance of sediments and he considered that metamorphic quartz was more elongated than igneous quartz. This concept, however, was not supported by the work of Blatt and Christie (1963).

Quartz elongation has been studied in the Permian sands and has been compared with samples from the possible source areas. The ratio of elongation has been determined by the formula, $E = A/B$ where the A is the long and B the intermediate axes. These ratios are given in (Table. 6.2). The relationship between quartz elongation and quartz type is given in Fig. 6.1 and shows that the elongation coefficient for metamorphic samples is greater than 1.5, and for plutonic samples is smaller than 1.5. However, the mean elongation coefficient for Permian sands is less than 1.5 and this indicates that plutonic rocks were probably the source for quartz grains.

Localities	%< 5 C. U.	%> 5 C. U.
Permian sands	71	29
Encounter Bay Granite	61	39
Plutonic samples of erratics	48	52
Plutonic samples of Antarctica	57	43
Metamorphic samples of erratics	82	18
Metamorphic samples of Antarctica	76	24
Kanmantoo Group samples	73	27

Table.6.1. Number of crystal units (CU) in polycrystalline quartz grains in the Permian sands of South Australia and in source rock samples.

Localities	%>1.5 E	%<1.5 E
Permian sands	68	32
Encounter Bay Granite	72	28
Plutonic samples of erratics	48	52
Plutonic samples of Antarctica	69	31
Metamorphic samples of erratics	6	94
Metamorphic samples of Antarctica	5	95
Kanmantoo Group samples	8	92

Table.6.2. Elongation coefficient (E) of monocrystalline quartz grains in the Permian sands of South Australia and in source rock samples.

CATHODOLUMINESCENCE MICROSCOPY OF QUARTZ GRAINS

The CL colour of quartz in sand and sandstone may provide a signature of the source rock from which individual grains, or populations of grains, were derived. Cathodoluminescence colours of quartz were first detected in vein quartz and smoky quartz by Goldstein (1907) during experiments on the nature of cathode rays in evacuated discharge tubes.

Early applications of cathodoluminescence (CL) to geology (Smith and Stenstrom 1965; Long and Agrell 1965; Sippel 1965) demonstrated that quartz and feldspar luminesce when exposed to an accelerated electron beam, and that different grains of the same minerals species may exhibit different colours of emission. Smith and Stenstrom (1965) and Sippel (1965) used this property as a petrological tool for the distinction of detrital quartz grains of igneous or metamorphic origin from their overgrowth cement. Quartz from most igneous rocks has a blue luminescence (Smith and Stenstrom 1965; Zinkernagel 1978; Sprunt and Nur 1979). Quartz from extrusive igneous rocks generally luminesces lighter blue than quartz from intrusive rocks although this relationship is not consistent and depends on the later thermal and deformational history of the rock (Zinkernagel 1978; Matter and Ramseyer 1985; Owen and Garson 1990).

Quartz from low grade metamorphic rocks generally luminesces reddish-brown (Zinkernagel 1978; Sprunt et al. 1978; Sprunt 1981). Quartz from high grade metamorphic rocks luminesces dull blue or indigo (Zinkernagel 1978; Sprunt et al. 1978; Owen 1984; Matter and Ramseyer 1985). Owen (1984) found that only above garnet grade does quartz CL become homogenised to a uniform reddish-brown colour. Sippel (1968) used CL colour contrast between detrital quartz grains and their authigenic overgrowths to assess the origin of quartz cement in sandstones. This technique was utilised more fully by Sibley and Blatt (1976) in their study of the quartz cement budget in the Tuscarora quartz arenite (Pennsylvania). Sedimentary petrologists have long sought petrographic criteria capable of reliably discriminating between quartz grains from igneous and low or high grade metamorphic sources but such efforts have led to more controversy than success (Pettijohn et al. 1987, ch.7). To date, most applications of CL to varietal provenance studies of sandstone have employed quartz. This is probably a response to the difficulty in assessing diversity in monocrystalline quartz using other techniques. Optically not only is one quartz grain nearly indistinguishable from the next, but quartz exhibits only modest chemical diversity as well (Dennen 1967). However, CL offers the potential of distinguishing quartz from different source rock types.

Several studies have been done on the diversity of CL colours of quartz in an effort to relate CL colour to source rock type (Zinkernagel 1978; Füchtbauer et al. 1982; Owen 1984,1991; Matter and Ranseyer 1985). Füchtbauer et al. (1982) based most of their study on diversity in quartz CL, relating different grain types to inferred source rock types. Owen (1984), Miller (1988), Owen and Carozzi (1986) addressed provenance of the Jackfork Sandstones (Arkansas) by evaluating the relative proportions of "brown" and "blue" CL quartz in that formation and several coeval sandstones for which provenance information was already available in more esoteric study. Owen and Anders (1988) used CL to discount a predominantly volcanic source for shock-metamorphosed quartz grains at the Cretaceous Tertiary boundary in Colorado and New Mexico.

CL is a useful tool in studying provenance of sand and sandstone because CL colour differences permit discrimination between quartz grains that were originally crystallised in igneous environments from those recrystallised under metamorphic conditions (Zinkernagel. 1978). Quartz grains in sandstone display a variety of CL colours that are generally representative of the source rocks from which the grains were derived (Matter and Ramseyer 1985; Owen and Carozzi 1986; Ruppert 1987; Owen 1991). Bluish-black luminescence has been observed in plutonic quartz (Matter and Ramseyer 1985). The luminescence intensity of blue, violet and brown CL colours displayed by igneous and metamorphic quartz remains stable or changes only slightly during electron bombardment. These may be termed long lived CL colours (Ramseyer et al. 1988).

The causes of CL in quartz are not yet fully determined (Zinkernagel 1978; Tovey and Krinsley 1980; Sprunt 1981; Waychunas 1988., Walker 1990).

Several factors have been suggested for different CL colour in quartz grains, including trace elements (Grant and White 1978; Sprunt 1981), crystal structural deformation (Zinkernagel 1978; Tovey and Krinsley 1980), and intrinsic defects (Walker

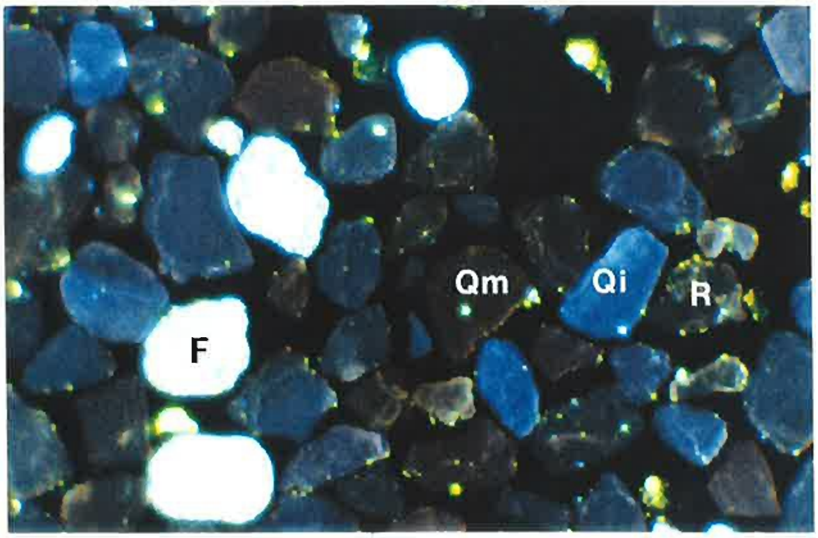
Plate 6.1

A: Cathodoluminescene photomicrographs on quartz grains of Permian sands shows the different colour in quartz grains. Qi (igneous quartz), Qm (metamorphic quartz), F (feldspars), R (rock fragments).

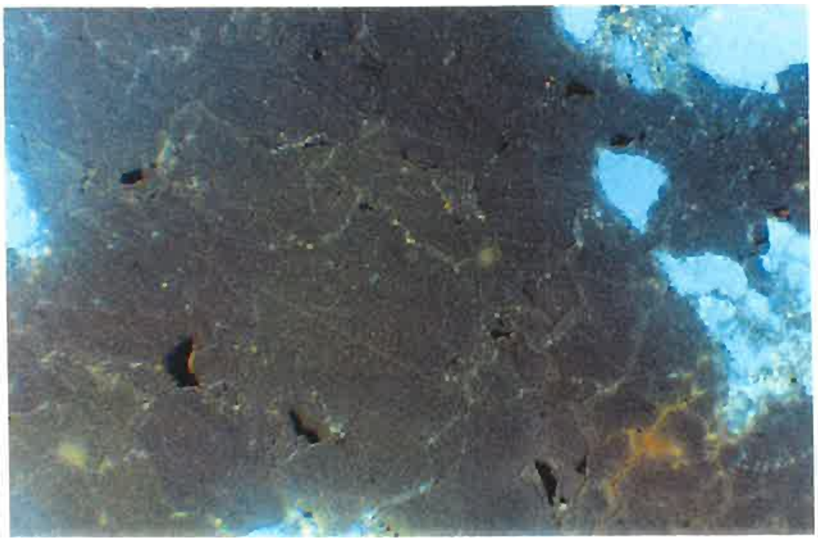
B: CL colour of quartz grains within the gneissic rock samples of East Antarctica.

C: CL colour of quartz grains in the granitic rock samples of Encounter Bay.

Scale. 80 x.



A



B

C

1990). No reliable correlation has yet been established for the abundance of any specific trace element and quartz CL.

INTERPRETATION OF PROVENANCE

CL studies of quartz grains can be used in studies of the provenance of the Permian sands from which the grains have been derived. Thus, the relative abundance of brown and blue CL quartz may provide a fingerprint of the ultimate source area from which a sand was derived.

Interpretation of luminescence has been done on a colour photo and point count analysis can be easily undertaken using the printed colour to find the ratio of different CL colours on quartz grains from different samples. Quartz occurs in the Permian sands as monocrystalline and polycrystalline or composite grains. They are ultimately derived from a huge mix of plutonic, volcanic, metamorphic and sedimentary rocks. Quartz grains in the Permian sands have been studied by CL and showed different luminescent colours ranging from blue to brown luminescence because individual grains were derived from different sources (Plate. 6.1). The ratio of blue and brown CL quartz may provide a fingerprint of the ultimate source area from which the Permian sands were derived. Provenance study using CL can provide guidance to proximate sources, and the ratio of brown to blue acts as a sort of fingerprint that can identify most of the source rocks that have produced these sands. The distributions of the CL colours of quartz in the Permian sands were used by point counting approximately 200 grains on the CL photos. The results indicate that the percentage of blue CL quartz is 52.9% and brown CL quartz is 47.1% and the ratio of brown to blue is 1.12. Comparison of these data with the samples from Antarctica, erratics and Encounter Bay, indicate that the quartz grains from plutonic samples have a bluish luminescence (Plate 6.1C) and that quartz phenocrysts in the volcanic erratics have blue through violet luminescence colours. The similarity between plutonic and volcanic quartz is not suprising because phenocrysts in volcanic rocks crystallised early under hypabyssal or plutonic conditions. Quartz grains from metamorphic samples show reddish brown luminescence colours (Plate 6.1B). Since

Permian sands have a heterogeneous mixture of these two CL colours and the ratio of blue to brown colours is very close to unity, this therefore indicates approximately equal derivation from both plutonic and high grade metamorphic sources.

TRACE ELEMENTS IN QUARTZ GRAINS

Physical properties of quartz, either optical or morphometric, are limited in their use as provenance indicators. The possible utilisation of the trace element content of quartz in provenance interpretation was first noted by Dennen (1964, 1967). However Stavrov (1961) and Dennen (1964, 1967) showed that the quartz grains in different arenites vary considerably in the kinds and proportions of their contained trace elements. Trace element content of quartz is determined principally by the physico-chemical environment during crystallization of a quartz grain (Dennen 1967). Suttner and Leininger (1972) have shown that the trace element distribution in detrital quartz derived exclusively from a shallow level granodioritic batholith is different from that found in detrital quartz derived from the volcanic cover of the same batholith. Consequently, the trace element content of quartz can be different for different granite masses (Stavrov 1961; Dennen 1964), for granite and associated pegmatite (Tatekawa 1954; Walenezak 1966), and for various contact metamorphic facies associated with the same intrusive body (Perry 1971). Although the trace element contents of a population of detrital quartz grains in the sand may be difficult to interpret, it may not be as difficult to interpret the trace element content of individual quartz grains. Standard deviations for plutonic quartz are 14ppm for Fe, 3 ppm for Mg, 8 ppm for Ti and for metamorphic and volcanic quartz 73 ppm for Fe, 18 ppm for Mg and 24 ppm for Ti (Dennen 1964). Moreover less amounts of Fe and Mg are found in plutonic quartz than in the volcanic and metamorphic quartz (Dennen 1964). This study will provide some clues to differences in trace element content of quartz from Permian sands and possible source rocks. Quartz trace elements from Permian sands, erratics, Antarctica and Encounter Bay samples could be clearly identified using the content of aluminium, calcium, iron, lithium, magnesium and manganese. The mean and standard deviation of trace elements in the quartz grains of Permian sand, erratics, and Antarctica are shown in Table. 6.3.

Mn ppm	Fe ppm	Li ppm	Al ppm	Ca ppm	Mg ppm	Localities	samples
32 (11)	470 (333)	3.8 (4)	4132 (227)	3775 (236)	1050 (623)	Antarctica	gneisses
25 (14)	598 (379)	3 (4)	25576 (43)	5090 (405)	497 (108)	Antarctica	granites
28 (42)	411 (299)	3 (4)	13722 (14777)	6156 (104)	2148 (568)	Erratics	volcanics
8 (5)	130 (61)	1.2 (1.2)	7697 (342)	250 (28)	7755 (1048)	Erratics	granites
2 (1)	188 (66)	0 (0)	330 (40)	823 (108)	753 (950)	Erratics	gneisses
7 (4)	123 (62)	1 (1)	7468 (325)	286 (31)	6784 (113)	Encounter.B	granites
5 (4)	113 (34)	1 (0.3)	488 (352)	43 (33)	731 (460	Permian	sands

Table.6.3. Mean and standard deviation (S.D.) of trace elements in the quartz grains from Permian sands of southern Australia and from the source rock samples.

The Table shows that lower amounts of Mn are found in plutonic samples, and higher amounts of Mn are found in volcanic and metamorphic samples. The amounts of Fe in volcanic, plutonic and metamorphic rocks are lower in granitic erratics and higher in gneisses. Lower amounts of Mg are found in metamorphic rocks compared with plutonic and volcanic samples. High amounts of Ca and Al may result from the incomplete separation of quartz from feldspar as this was the most difficult step in the analytical procedure. The result from trace-element chemistries of quartz showed that quartz from different plutonic and metamorphic rocks have different proportions of trace elements. Assuming that plutonic and high grade metamorphic rocks are the dominant source rocks for the Permian sands we expect that the standard deviations in the Permian sands result from the mixture of two sources. The mean percentage and standard deviation of trace elements in quartz grains from the mixture of plutonic and metamorphic rock samples were measured as 14.8 ppm for Mn [standard deviation of 1 ppm], 301.8 ppm for Fe [S.D. 193 ppm], 1.8 ppm for Li [S.D. 1 ppm] and 2369 ppm for Mg [S.D. 151 ppm]. Therefore, comparing the trace elements of quartz grains in the Permian sands with the mixture of both rock samples suggest that plutonic and high grade metamorphic rocks were the chief sources of detrital quartz grains in these sands.

DISCUSSION

Quartz grains comprise 91% of the Permian sands and provide clear evidence with regard to provenance. Distinct unimodality of crystal size in polycrystalline quartz of the Permian sands suggests a plutonic source (Plate. 7.1E). The number of grains with less than five components in the polycrystalline quartz grains of Permian sands when compared with the various rock samples suggests mostly plutonic source rocks. The form of quartz grains may give a clue to their derivation. Quartz from metamorphic rocks has an elongation coefficient greater than 1.5 (Bokman 1952). The elongation coefficient of the monocrystalline quartz grains in the Permian sands have shown about 68% of the monocrystalline quartz grains have less than 1.5, and about 32% have greater than 1.5, suggesting both plutonic and high grade metamorphic sources (Table. 6.2 and Fig. 6.1). The trace element chemistries of quartz from the Permian sands are in such good agreement as to suggest that a mixture of plutonic and high grade metamorphic rocks was the chief source of detrital quartz grains. Quartz grains in Permian sands showed both blue and brown CL colours because individual grains were derived from different sources. The almost equal ratio of blue to brown CL also suggests mixed plutonic and high grade metamorphic sources.

FELDSPAR FOR PROVENANCE

Feldspar is a useful provenance indicator because it is abundant in many sedimentary basins and because the variation in its physical and chemical properties has genetic implications. Feldspar which formed in the plutonic regime, for example, has inherent physical and chemical properties which are different from those of feldspar which formed in volcanic and metamorphic environments. Feldspar has also been commonly used in chemical and physical studies. Gorai (1951) made careful studies of twinning in plagioclase and its abundance in igneous and metamorphic rocks. Oscillatory zoning, progressive zoning and lack of zoning have also been shown to be important indicators of provenance (Pittman 1963). Chemical zoning is well documented in both plagioclase and alkali feldspar (Smith 1974); however, only zoning in plagioclase has received much attention as a possible provenance indicator. Pittman (1963) examined 90 plutonic, volcanic, hypabyssal and metamorphic rocks in order to determine the relative abundance of progressively-zoned, oscillatory-zoned, and unzoned plagioclase in each. The results of this study showed that the presence of any type of zoning is a strong indicator of an igneous origin for plagioclase. In particular, oscillatory zoning suggests either a volcanic or hypabyssal source. Metamorphic rocks are characterised by the predominance of untwined-unzoned plagioclase (Pittman 1970). Rimsaite (1967) studied the varietal properties of feldspar in potential source rocks and defined nine classes based on zoning, intergrowth habits, twinning and fracturing. Suttner and Basu (1977), Plymate and Suttner (1979) determined the structural studies of alkali feldspar by single grain X-ray diffraction and showed how this information can be used in provenance interpretation. They found sanidine to be highly disordered indicating rapid cooling (volcanic origin) whereas disorder in microcline and orthoclase was less, which they correlated with the slower cooling history of a batholith or metamorphic belt. In the past, feldspar compositions were determined by tedious wet chemical methods that were too time consuming to be of practical use to the sedimentary petrologist. The application of electron microprobe in the 1960s made possible the rapid and precise chemical analysis of individual feldspar grains. Trevena and Nash (1981) plotted the microprobe composition of over 2000 feldspars from igneous and metamorphic rocks and compared them with sandstone of known volcanic provenance. Plots on a triangular diagram of the system Ca-Na-K indicate that the potassium content is a useful parameter for determining the origin of both plagioclase and alkali feldspars. There is a gradual decrease in the maximum potassium content of plagioclase from volcanic to plutonic and metamorphic environments. Emphasis in this chapter is given to those properties which are most useful to the sedimentary petrologist interested in deciphering the provenance of Permian sands. Because the detrital feldspar grains in Permian sands may have been derived from a number of different sources, the properties of an individual feldspar grain are useful only in determining the provenance of that particular grain. Therefore the determination of feldspar provenance in the Permian sands was made on a grain by grain basis.

CHEMICAL COMPOSITION OF FELDSPARS

Feldspar is a useful mineral for provenance studies because it is the second most abundant mineral in terrigenous detrital rocks. Thus determining the chemical composition of detrital feldspar is important in studies of sandstone provenance and diagenesis. Chemical study of detrital feldspar can be useful in provenance analyses to determine palaeogeographies and to test palaeotectonic models. The chemical composition of individual detrital feldspar grains can be determined in 30 seconds with an electron microprobe.

The electron microprobe has become a standard tool in provenance studies. Chemical composition of feldspars is also given particular emphasis in the light of recent electron microprobe studies (Trevena and Nash 1981; Putnam and Pedskalny 1983) that have proven the utility of microanalyses in the delineation of possible source rocks. Despite the microprobe capability for rapidly analysing individual grains, there appears to be relatively few chemical analyses of detrital feldspar that have been published to date. These are by Boles and Coombs (1975), Odom (1975), Stablein and Dapples (1977), Trevena and Nash (1979, 1981), Hubert and Reed (1978), Sibley and Pentony (1978) and Waugh (1978). Minor elements analysis of feldspar can also be useful as a provenance tool, especially where a potential source rock might contain unusually high levels of a minor element (Trevena and Nash 1981). This has not been used in this thesis.

Therefore, in this thesis feldspars from Permian sands were analysed and compared with feldspar from different possible source rocks.

ANALYSIS OF FELDSPAR

MAJOR ELEMENTS

The composition of detrital feldspar grains from Permian sands is illustrated in Fig.6.0 and can be compared with Figs. 4.2, 4.12 and 5.1.

The weight percentage of anorthite (An), orthoclase (Or), and albite (Ab) was determined by electron microprobe. Volcanic alkali feldspar varies widely in potassium content with its composition ranging from An1Ab12Or87 to An13Ab74Or13, while alkali feldspar from both plutonic and metamorphic rocks are potassium rich. Alkali feldspar from plutonic rocks is rather potassic ranging from An0Ab2Or98 to An1Ab40Or59 and from metamorphic rocks ranges from An0Ab2Or98 to An2Ab98Or0 (Evans 1978; Trevena and Nash 1981). Detrital grains of alkali feldspar that are more potassic than Or88 orthoclase molecule are derived from metamorphic and plutonic sources, and alkali feldspar more sodic than Or51 weight percent albite molecule are almost certainly derived from volcanic source rocks (Trevena and Nash 1981). Calcic plagioclase with composition up to An54 (Labradorite) with alkali feldspars and only moderately high potassium contents, fall within the known compositional range of volcanic feldspar (Trevena and Nash 1981; Helmold 1985). Trevena and Nash (1981) interpret the high K plagioclase to be mostly of volcanic origin, and the low K plagioclase to be derived from plutonic or metamorphic rocks. The potassium rich alkali feldspar is most likely derived from plutonic or metamorphic rocks, while the intermediate alkali feldspar is probably volcanic or plutonic in origin (Trevena and Nash 1981). Plymate and Suttner (1979) used the frequency of cross-hatch twinning in K feldspar to determine the provenance of Holocene stream alluvium from southwest Montana. The distribution of cross-hatch twinned K-feldspar in the alluvium very clearly reflected the distribution of microcline -rich rocks in the source terrane. Alkali feldspar with the composition of Ab16 are mostly from the granulite facies (Bohlen and Essene 1977; Stormer and Whitney 1977).

DISCUSSION

Feldspar grains comprise 8.3% of the Permian sands and can be used as a provenance indicator using microprobe analysis. This allows discrimination between volcanic and metamorphic or plutonic sources, especially when the composition was determined grain by grain. Alkali feldspar grains have provided some evidence about the nature of the source. Calcic feldspars have supported a volcanic source.

Detrital feldspars of the Permian sands contain both K feldspar and lesser amounts of plagioclase. The more calcic feldspars were derived from volcanic rocks, and the potassic feldspars came from metamorphic or plutonic samples, when compared with other rock samples in the source areas. Detrital feldspar grains from 60 Permian sands were analysed and have been compared with the source rocks; the results have been illustrated in Figs. 4.12, 5.1 and 6.0A. The plagioclase grains from these samples follows a low K trend, An8Ab92Or0 to An6Ab94Or0 characteristic of plagioclase derived from non-volcanic sources. Plagioclase with almost non K composition trend probably were derived from plutonic and metamorphic sources. Rarely is albite present in Permian sands. Composition of the alkali feldspar within the Permian sands ranges from An0Ab8Or92 to An3Ab0Or97, and a few feldspar grains range from An2Ab23Or75 to An12Ab0Or88. The alkali feldspar grains of Permian sands showed mostly Or>90% molecule. In this study there are no alkali feldspars in the Permian sands that have a composition between 17% to 55% Or. The percentage of K, Na and Ca from different localities have been measured in the Permian sands and the average is 80.2%K, 17%Na and 2.7%Ca. The composition of detrital feldspar in the Permian sands indicates a mixed provenance for the feldspar grains. Compositional analysis of feldspar in the Permian sands (Table. 6.4) showed that the most common feldspars are potassium rich alkali feldspar, principally orthoclase-potassium rich sanidine to microcline (An0Ab6Or94). Comparison of the composition of feldspars in the Permian sands with other sources indicates that most of the feldspars have been derived from gneisses or granitic rocks rather than volcanic sources.

However the presence of some sanidine and calcic plagioclase in the Permian sands imply a minor volcanic source.

CONCLUSION

The general composition of Permian sands indicates a derivation from igneous and high grade metamorphic cratonic sources. Detailed analyses of quartz grains showed that the dominant nature of the source areas for the Permian sands are plutonic and metamorphic rocks. This can be supported by the form of quartz grains, number of crystal units in polycrystalline quartz grains, trace elements in quartz grains and their CL colour.

Analyses of feldspar grains indicate these to be mostly potassic in composition and suggest metamorphic and plutonic sources. The microcline with cross hatched twinning and almost no alteration seems to be a result of granite and gneiss source rocks and glacial transport. This is enhanced by its greater resistance to glacial comminution than plagioclase.

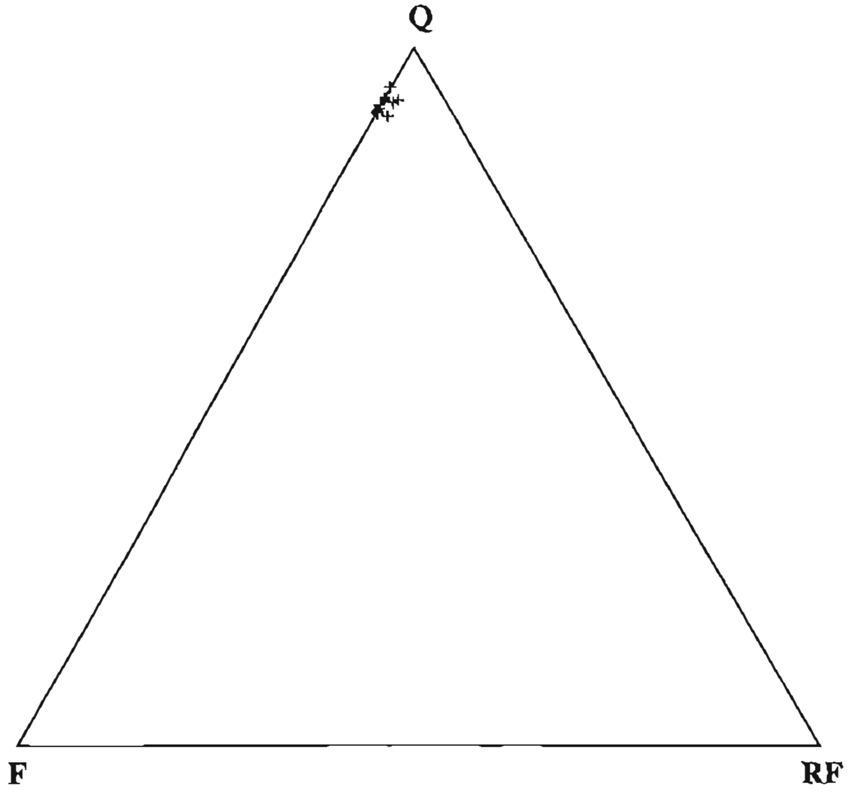


Fig. 6.0. Composition of Permian sands from southern Australia displayed on a triangular diagram: Q= Quartz; F= feldpars and RF= rock fragments.

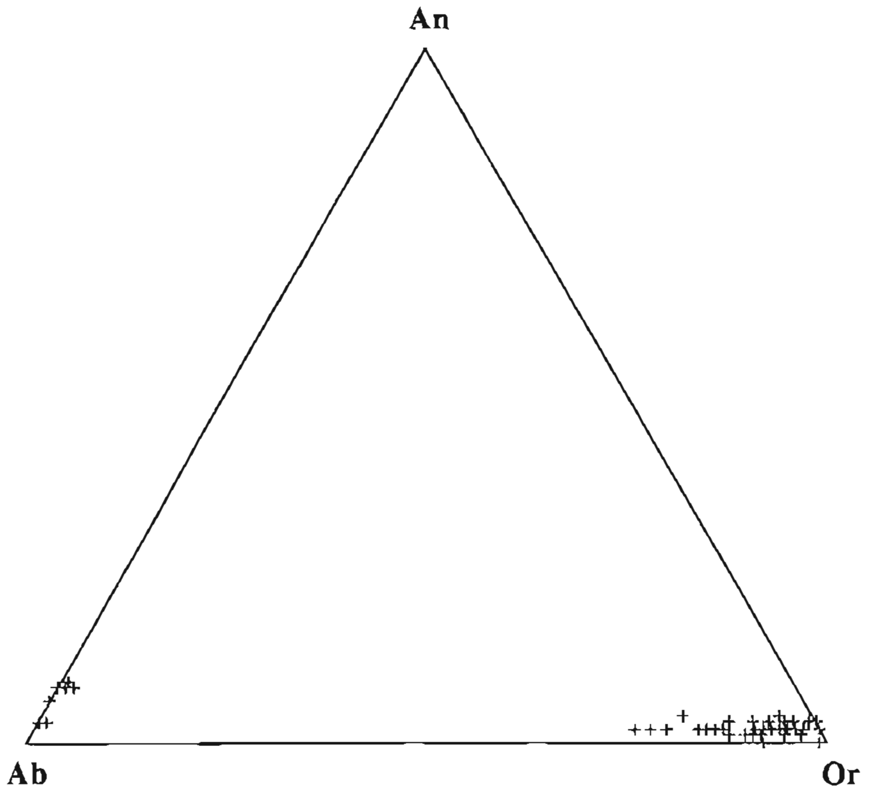
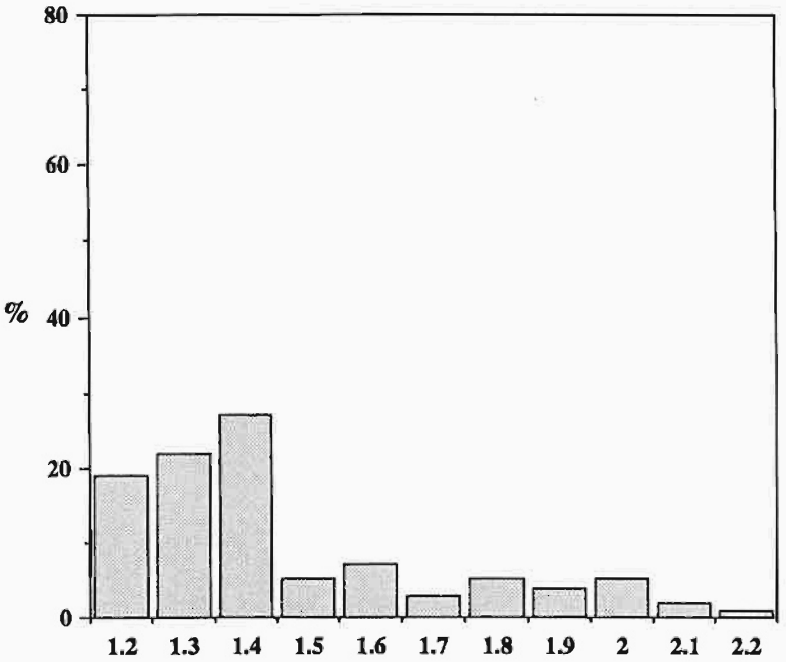


Fig.6.0A. Composition of detrital fedspar (weight %) grains in the Permian sands of southern Australia as determined by microprobe analyses displayed on a triangular diagram: An (anorthite), Ab (albite) and Or (orthoclase) on the base of 32 (O).



Elongation of detrital quartz grains in the Permian sands of southern Australia.

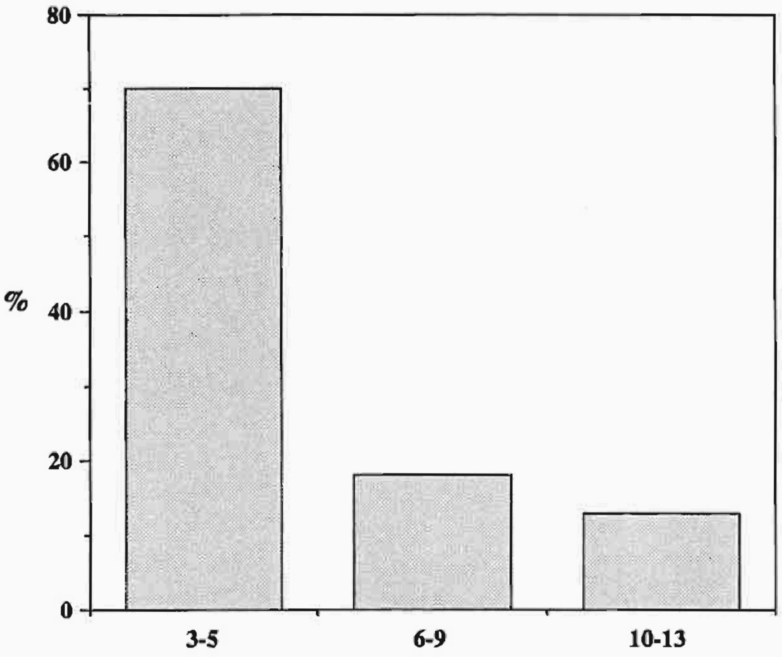


Fig. 6.1 Number of crystal units in polycrystalline quartz grains from Permian sands of southern Australia.

Table. 6.4. Composition of detrital feldspar grains in the Permian sands of southern Australia.

SiO2	Al2O3	K2O	Na2O	CaO	MgO	MnO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
65.5	18.9	16.9	0.2	0.1	0	0	0	0.2	0	101.8	0	1	98.5
64.1	18.3	16.1	0.1	0.5	0	0	0	0.2	0	99.1	2	0	98
63.5	18.8	16.5	0.1	0.4	0	0	0	0.3	0.1	99.6	2	0	98
64.1	18.4	16.6	0.1	0.6	0	0.1	0	0.1	0	99.9	3	0	97
64.7	19.1	16.9	0.1	0.5	0	0	0	0	0.1	101.2	3	0	97
64.5	18.7	16.4	0.1	0.6	0	0.3	0	0.2	0.1	100.7	3	0	97
63.6	18.6	16.8	0.1	0.6	0	0	0	0	0.2	99.9	3	1	96
63.8	18.3	15.3	1.2	0.3	0	0.2	0	0	0.1	99.1	1	3	96
65.6	18.5	16.1	0.3	0.4	0	0.1	0	0.1	0.3	101.3	2	2	96
64.7	18.6	15.7	0.2	0.4	0	0	0	0	0.1	99.8	2	2	96
64.8	19.2	16.1	0.5	0.5	0.1	0	0	0.2	0	101.5	2	4	94
63.5	18.8	16.2	0.4	0.4	0	0	0	0.9	0	100.3	2	4	94
64.4	18.8	15.9	0.5	0.5	0	0.2	0	0	0.1	100.5	2	4	94
62.6	18.7	16.1	0.5	0.7	0.1	0.3	0	0.3	0.4	99.8	3	4	93
65.5	19.2	15.7	0.6	0.5	0	0.2	0	0	0.1	101.9	2	5	93
63.8	18.5	15.8	0.6	0.5	0	0	0	0	0	99.2	2	5	93
64.2	18.8	15.3	0.6	0.4	0	0.2	0	0.2	0	99.8	2	6	92
65.1	19.2	15.8	0.6	0.6	0	0.2	0	0	0	101.4	2	6	92
64.2	18.5	15.4	0.5	0.8	0	0.3	0	0.1	0	99.9	4	4	92
65.2	18.1	15.8	0.7	0.1	0	0	0	0.6	0	100.4	0	8	92
64.9	19.1	15.6	0.7	0.5	0	0	0	0.2	0	100.9	2	6	92
63.9	18.8	15.4	0.6	0.5	0	0.2	0	0.3	0.1	99.8	2	6	92
64.7	18.4	15.5	0.7	0.5	0	0.2	0	0.2	0.2	100.6	3	6	91
64.2	18.7	15.2	0.9	0.2	0	0	0	0	0	99.1	1	8	91
64.4	18.8	15.3	1.1	0.2	0	0.1	0	0.2	0.2	100.3	1	9	90
65.3	18.8	15.7	1.1	0.5	0	0	0	0.1	0	101.4	2	8	90
63.9	18.6	15.2	1.1	0.2	0	0	0	0	0	99.1	1	10	89
63.8	19.6	14.2	1.4	0.2	0	0	0	1	0.2	100.1	1	12	87
64.1	19.1	14.4	1.3	0.7	0	0	0	0.3	0	99.8	3	11	86
64.1	18.8	14.8	1.4	0.4	0	0	0	0.2	0	99.6	2	12	86
65.7	19.2	14.8	1.5	0.4	0	0.1	0	0.1	0	101.8	2	13	85
64.3	19.1	14.3	1.6	0.4	0	0	0	0.1	0.2	100.2	2	14	84
63.4	18.8	14.8	2.1	0.8	0	0	0	0	0.1	99.9	4	16	80
61.4	18.5	15.7	2.5	0.6	0	0	0	0.3	0	99.2	2	19	79
64.6	19.1	13.7	2.4	0.4	0	0	0	0.4	0.1	100.7	2	21	77
64.6	19.1	13.1	3.1	0.5	0	0	0	0.3	0	100.7	2	23	75
66.7	21.4	0.1	10.4	1.8	0	0	0.1	0	0.1	100.7	9	90	1
65.6	20.1	0.1	12.1	0.7	0	0.1	0.3	0	0	98.9	3	96	1
66.1	21.6	0.1	10.6	1.9	0	0.1	0.1	0.1	0.1	100.9	9	90	1
66.3	18.9	0.1	12.5	0.8	0.1	0.1	0.1	0.1	0.1	99.2	3	97	0
67.6	20.9	0.1	11.1	1.3	0	0.1	0	0	0.1	101.1	6	94	0
67.9	20.1	0.1	11.3	0.7	0	0	0.2	0.1	0.1	100.2	3	97	0
66.7	21.4	0.1	10.4	1.8	0	0	0.1	0	0.1	100.7	9	90	1
66.4	21.4	0.2	11.1	1.9	0	0.1	0.2	0.2	0.1	101.6	8	90	2
66.1	21.6	0.1	10.6	1.9	0	0.1	0.1	0.1	0.1	100.9	9	90	1
65.7	19.2	14.8	1.4	0.4	0	0.1	0	0.1	0	101.8	2	13	85
65.7	19.1	14.2	1.7	0.4	0.2	0	0	0.3	0.1	101.9	2	15	83
65.7	21.3	0.1	11.3	1.7	0	0.1	0.4	0	0	100.5	8	92	0
65.6	18.5	16.1	0.3	0.4	0	0.1	0	0.1	0.3	101.3	2	2	96

Table. 6.4. Continue...

SiO2	Al2O3	K2O	Na2O	CaO	MgO	MnO	FeO	TiO2	Cr2O3	Total	An	Ab	Or
65.6	20.1	0.1	12.1	0.7	0	0.1	0.3	0	0	99.1	3	97	0
65.3	18.8	15.7	1.1	0.5	0	0	0	0.1	0	101.4	2	8	90
65.2	18.1	15.8	0.7	0.1	0	0	0	0.6	0	100.4	0	8	92
65.1	18.7	15.9	0.5	0.2	0	0	0	0	0	100.4	1	5	94
65.1	19.1	15.8	0.7	0.5	0	0.2	0	0	0	101.4	2	6	92
65.1	19.1	15.8	0.7	0.6	0	0.1	0	0	0	101.4	2	6	92
64.9	19.1	15.9	0.5	0.8	0	0.1	0	0	0.1	101.1	4	4	92
64.9	19.1	15.6	0.7	0.5	0	0	0	0.2	0	100.9	2	6	92
64.9	19.4	15.5	0.6	0.6	0	0	0	0.2	0	101.2	3	8	89
64.8	19.2	16.1	0.5	0.5	0.2	0	0	0.2	0	101.5	2	4	94
64.8	18.5	15.2	1.7	0.5	0	0	0	0.4	0.1	101.2	2	14	84
64.7	18.4	15.4	0.7	0.5	0	0.2	0	0.2	0.2	100.6	3	6	91
64.4	18.6	16.4	0.1	0.3	0	0	0	0.6	0.1	100.5	2	0	98
64.4	18.8	15.3	1.1	0.2	0	0.1	0	0.2	0.2	100.3	1	9	90
64.3	19.1	14.3	1.6	0.5	0	0	0	0.1	0.2	100.2	2	14	84
64.2	18.8	16.8	0.3	0.5	0	0	0	0.2	0	100.6	2	3	95
64.2	18.8	15.3	0.6	0.4	0	0.2	0	0.2	0	99.8	2	6	92
64.2	18.5	15.4	0.5	0.8	0	0.3	0	0.1	0	99.9	4	4	92
64.1	18.7	16.1	0.9	0.4	0	0.2	0	0.1	0	100.5	2	7	91
63.9	18.5	16.2	0.4	0.5	0	0	0	0.1	0.1	99.8	3	3	94
63.8	18.3	15.3	1.2	0.3	0	0.2	0	0	0.1	99.1	1	3	96
63.7	18.3	16.1	0.1	0.5	0	0	0	0.2	0	99.1	2	0	98
63.7	18.8	15.7	0.2	0.3	0.1	0	0	0.3	0.1	99.2	2	4	94
63.7	19.6	14.2	1.4	0.2	0	0	0	0.1	0.2	100.3	1	12	87
63.6	18.6	16.7	0.1	0.6	0	0	0	0	0.2	99.9	3	1	96
62.6	18.7	16.1	0.5	0.7	0.1	0.3	0	0.3	0.4	99.9	3	4	93
61.4	18.4	15.7	2.5	0.6	0	0	0	0.3	0	99.1	2	19	79

CHAPTER SEVEN

HEAVY MINERAL ANALYSIS OF THE PERMIAN SANDS

INTRODUCTION

The heavy mineral assemblage of the Permian glacial sediments at different localities (Hallett Cove, Cape Jervis, Kings Point and Kangaroo Island) consists of garnet, rutile, sillimanite, staurolite, zircon, tourmaline, epidote, with a trace of sphene and spinel. The opaque fraction consists of ilmenite and magnetite. Problems of interpreting provenance from heavy mineral distribution may be overcome to a large extent by characterizing the variations of morphology, internal texture, other crystallographic properties, and chemical compositions of a single mineral species or a single mineral group. Heavy mineral analysis is a sensitive and well proven technique for determining the provenance of clastic sediments. The interpretation of the data is considerably enhanced by determining the composition of individual detrital grains, and comparison with possible source rocks. Many heavy mineral species, including pyroxene, amphibole, epidote, staurolite, garnet, zircon, tourmaline, monazite and spinel, show significant variations in composition that are related to the conditions under which their parent rocks were formed (Morton 1988). It should be kept in mind, however, that the mineralogical composition of a glaciogenic sediment is almost always the resultant of erosion in the source area, erosion during ice movement, and even some erosion in the neighbourhood of the final depositional site, thus giving a mixture of assemblages, each of which must be recognized as such. Mineralogical analyses are commonly restricted to heavy minerals which give rather reliable and easily obtainable results (Brodzikowski et al. 1991).

Heavy minerals: the minor, high density accessory detrital minerals of sand have been used as classical indices of provenance (Boswell 1933; Milner 1962). Because heavy minerals are amongst the most sensitive indicators of the nature of source areas, they have frequently been used in studies of provenance of their sediments.

Prior to 1949, many studies on heavy minerals in sediments were carried out in an attempt to define source areas of sediments, and for correlation purposes. Unique features such as shape, colour, composition have been used to identify the grains that were derived from a particular source rock or area. Several recent studies have focused upon the variability of a single heavy mineral species (Byerly at al. 1975; Morton 1985a,b). Heavy minerals are usually emphasised over light minerals because their greater mineralogical variety supplies more petrological information, so heavy mineral analysis was at its most popular in the early part of this century.

Minerals especially scrutinised, where variation in colour or grain shape with source rock have been emphasised, include tourmaline and zircon (Mackie 1927; Krynine 1946; Poldervaat 1955; Vitanage 1957; Callender and Folk 1958; Owen 1987), garnet (Dreimanis et al. 1957; Connally 1965) and hornblende (Van Andel and Poole 1960).

Compositional analysis of a mineral or mineral group through the use of the electron microprobe is presently receiving much attention. Studies to date have concentrated on amphiboles (Mange-Rajetsky and Oberhansli 1982), pyroxenes (Nisbet and Pearce 1977; Cawood 1983) and garnets (Morton 1984, 1985a,b, 1987a,b, 1988). Current interest in the opaque Fe-Ti oxides and chromite is also significant (Luepke 1980; Darby 1984; Basu and Hood 1985; Darby and Tsang 1987; ~~Shukla 1988;~~ Grigsby 1988, 1989, 1990, 1992; Basu and Molinaroli 1989, 1991).

Compositional variations of detrital garnet in the North Sea sandstones have proved to be a useful tool for provenance interpretation. These sandstones were derived essentially from metamorphic basement sources, either directly or through successive recycling episodes (Morton 1985). Garnet geochemistry has been used to evaluate provenance elsewhere, notably of the Palaeogene McKee Formation, which occurs on and offshore of the North Island, New Zealand (Smale and Morton 1988), Jurassic sandstones offshore mid-Norway (Gjelberg et al. 1987), Triassic sandstones of southern Germany (Borg 1986) and Devonian sandstones of the Midland Valley of Scotland (Haughton and Farrow 1989).

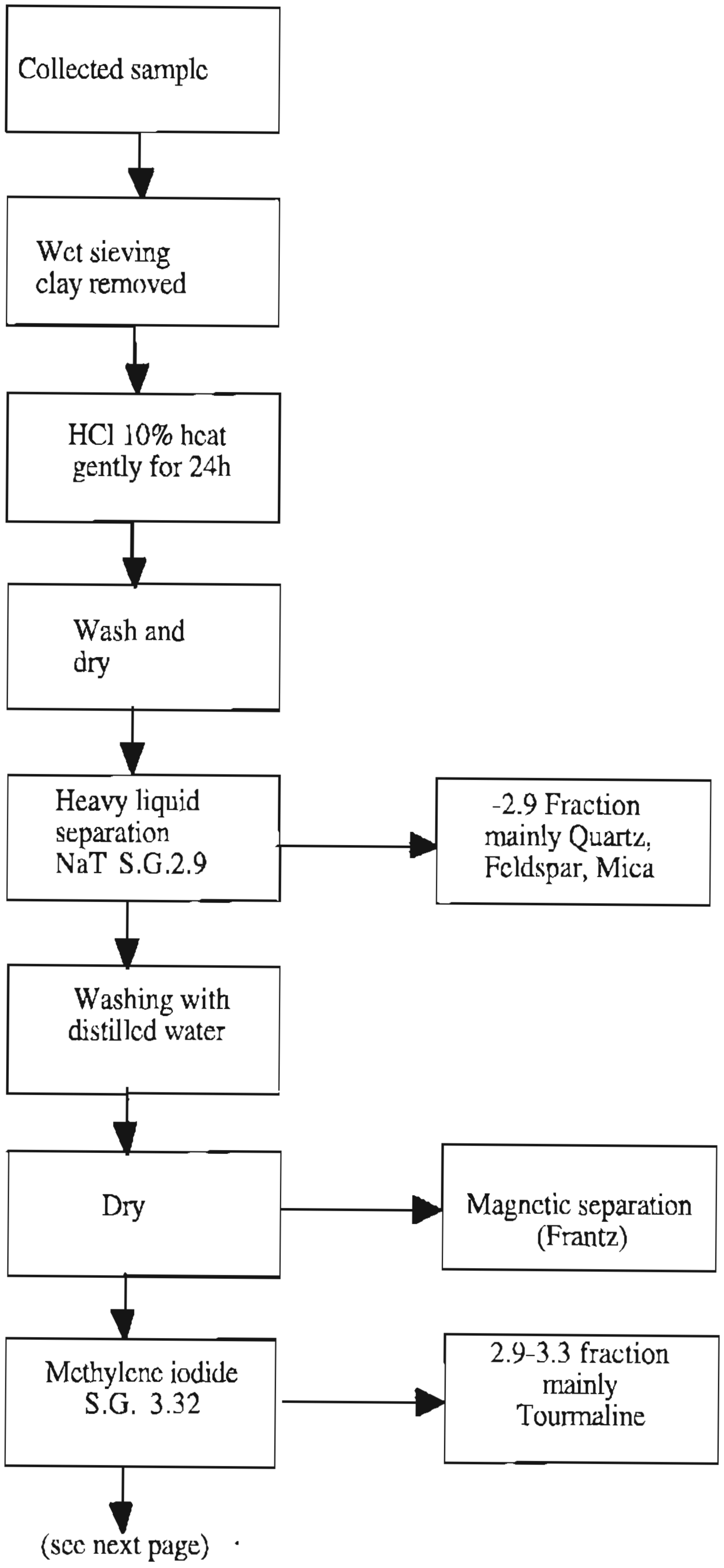
The possibility of dating individual detrital zircon grains to decipher their provenance obviously has special utility (Compston and Pidgeon 1986; Kober 1987; Cliff et al. 1990).

As mentioned in the previous chapter, during Early Permian time, the direction of ice flow was from the south and southeast, therefore Permian sands should contain heavy minerals from the Antarctic, Kanmantoo Group, Encounter Bay rocks, and adjacent sources. Some of the heavy minerals, in the Permian sands of southern Australia have been transported for several hundred kilometres from their source areas. This transport by glacial ice involved crushing and abrasion. Crushing may have increased the number of fragments of brittle and hard minerals (e.g. garnet), but abrasion decreased the size and the number of mineral grains, particularly the softer ones, such as mica, amphibole and pyroxene.

The mineralogy of the heavy minerals of the Permian sands in southern Australia has been investigated for this thesis by optical and electron microprobe analysis. The advent of rapid micro analysis equipment such as electron microprobe has added a new dimension to provenance studies, because it is now possible to determine, accurately and quickly, the compositional variation of a particular species in a heavy mineral assemblage. A considerable number of heavy mineral species that have been identified in the Permian sands are potentially useful in provenance studies. The data obtained are illustrated in the following pages. As results of this work broadly agree with the derivation of heavy minerals from the Antarctic, Encounter Bay and Kanmantoo Group areas, so the mineralogy of these potential source rocks has been examined in the same procedure.

Methods

Details of sample preparation and separations are presented in Fig. 7.0 and Appendix. 5.


Collected sample
Wet sieving
clay removed
HCl 10% heat
gently for 24h
Wash and
dry
Heavy liquid
separation
NaT S.G.2.9
-2.9 Fraction
mainly Quartz,
Feldspar, Mica
Washing with
distilled water
Dry
Magnetic separation
(Frantz)
Methylene iodide
S.G. 3.32
2.9-3.3 fraction
mainly
Tourmaline
(see next page)

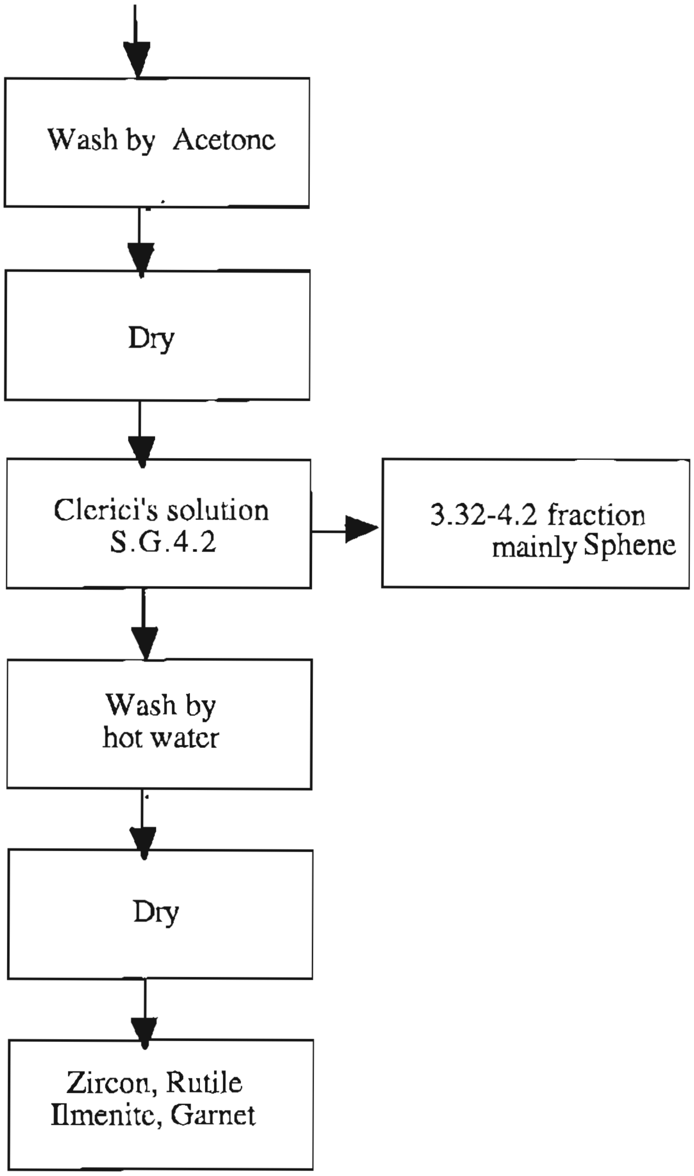


Fig. 7.0. Flow chart of sample treatment procedures involving heavy liquids and magnetic separations

Results

Investigation of heavy mineral assemblages in the Permian sands show complex variations. In the present study, Permian sands contain mainly garnet, rutile, sillimanite, staurolite, zircon, tourmaline and epidote. These seven most abundant non-opaque heavy minerals were studied and the mean percentage of each mineral in the Permian sands is shown in Table.7.0. These are in order of importance but vary significantly in abundance between samples. Minor accessory minerals such as sillimanite, staurolite and rutile make up less than 2% of the suite in the Permian sands. Ilmenite appears to dominate the opaque minerals, and in some samples comprises nearly 1.5wt% of the total concentrate. The heavy mineral data outlined in Table. 7.0 have been plotted graphically on a triangular diagram (Fig. 7.1). The three end members selected are: zircon, tourmaline and rutile (representing the ultra stable group), garnet (representing an intermediate stability), and others (representing lesser stability). In this triangle the heavy minerals occupy a position closer to the garnet end point illustrating the abundance of garnet in the source rocks.

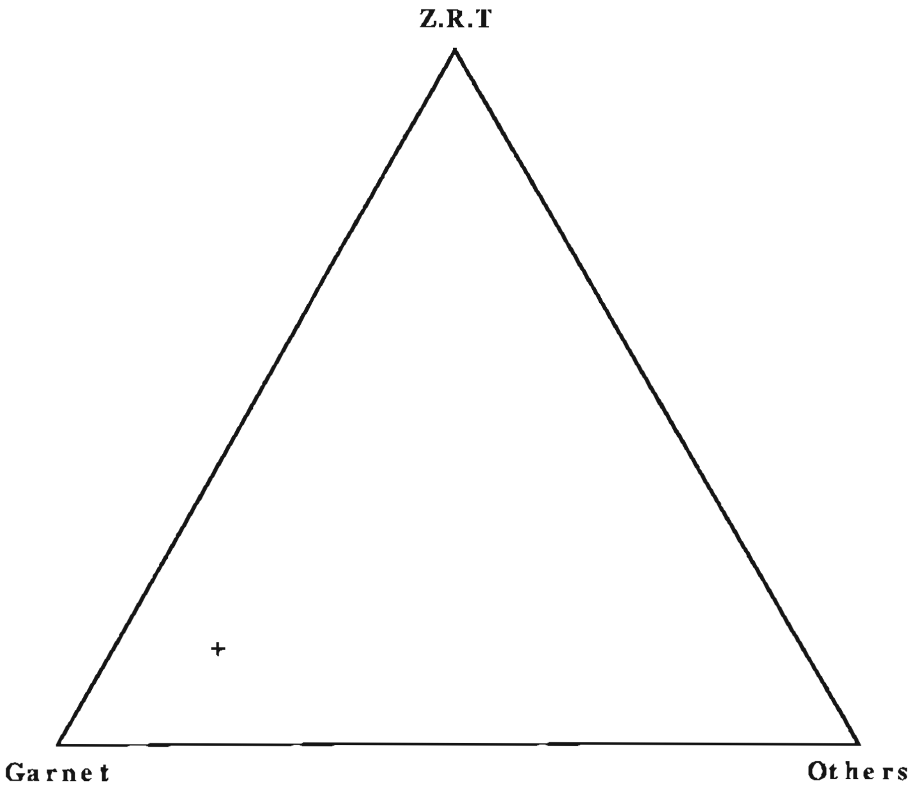


Fig. 7.1. Average composition of heavy minerals in Permian sands of southern Australia (Z.R.T: zircon, rutile and tourmaline).

Minerals	Percentage %
Garnet	73
Ilmenite	8
Tourmaline	7
Zircon	4
Rutile	3
Staurolite	2
Sillimanite	0.6
Epidote	0.5
Apatite	0.1
Unknown	1.2
Pyroxene+Amphibole	T
Spinel	T
Micas	T

Table. 7.0. Percentage of heavy minerals in the Permian sands of southern Australia.

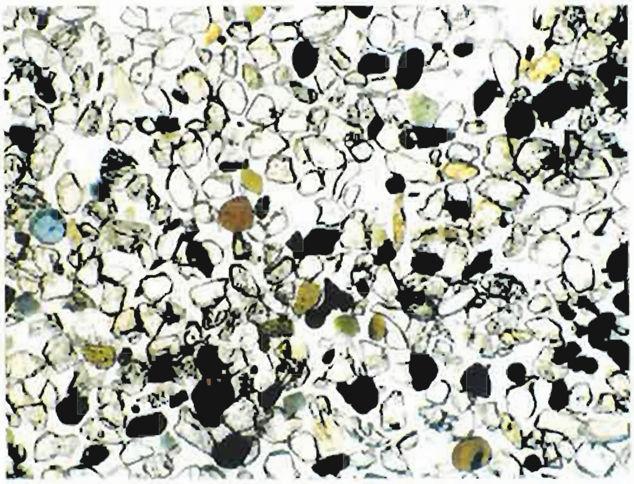
Usually weathering alters the mineral composition of sediments in source areas, particularly in broad regions of low relief (Stanley 1965). Investigation of heavy minerals in the Permian sands however revealed that more than 14 heavy mineral species are present. They are not altered, probably due to their preservation under cold and glacial conditions during and after transportation. Kyanite and sphene are absent in the Permian sands, but sphene has been found in the Permian erratics, so that the absence of these two minerals could well be the result of intrastratal solution because of their chemical instability.

GARNET

Garnet grains are subangular to rounded and many display a mammillary or knobbly surface texture (Plate 7.1A to D). Four different species of garnet were observed in the Permian sands, with colour ranges from colourless through pink to deep red. In the granitic rock samples, garnets have clear and pink to deep red colour, whereas in the metamorphic rock samples garnets are colourless to pink in colour. Detailed scanning electron microscopy of garnet grains have shown that the chattermark trails were caused by ice movements, and the other kinds of surface texture were caused by mechanical abrasion and chemical etching (see chapter 3). The heavy minerals of the Upper Palaeozoic tillites of Gondwanaland are dominated by garnets, a fact that has been noted by a number of geologists (e.g de Villiers and Wardaugh 1962). Many of the garnets in these tillites are well rounded and support very delicate surface chattermark trails (Folk 1975; Gravenor et al. 1978). The heavy minerals of the Permian tillites of South Australia were previously studied by Gravenor and Gostin (1979) who concluded that there has been an actual enrichment of garnet in the tillites of this age. The percentage of garnets in the heavy mineral suites from the intensively glaciated regions in Australia, South Africa and Antarctica ranges from 21-99% with a mean of 79% (Gravenor 1979). Abundant garnet in the heavy mineral suites of Permian fluvio-glacial sediments has been reported by number of geologists (e.g. Ludbrook 1967; de Villiers and Wardaugh 1962; Jacobson and Scott 1938). The Permian sands of southern Australia are rich in garnet with an average of 73%. At Hallett Cove location, garnets constitute 80% or more of the total assemblage (Gravenor 1979). This enrichment in garnets is due to mechanical stability of this mineral during periods of transport and deposition, compared to the abrasion of unstable minerals such as pyroxene and amphibole during transportation. Studies made on the chemical stability of garnet have provided a variety of results. The general consensus, however, is that garnet is highly resistant to mechanical abrasion, and only moderately stable under soil forming processes; but once buried, garnet shows high resistance to further weathering by intrastratal solutions, (Dryden 1946; Goldich 1938; Gravenor 1954; Pettijohn 1941; Sindowsky 1949). Less mechanically stable minerals in the glacial sediments during transportation were broken down leaving concentrates of garnets. As the ice readvanced, the garnet concentrates were picked up by the ice, reworked and transported. After many cycles then, the heavy minerals in the glacial deposits became enriched in garnets with a high percentage showing chattermark trails (Gravenor 1979). In this thesis it will be shown that the chemical composition of garnets in the Permian sands indicate that they were derived mostly from various metamorphic

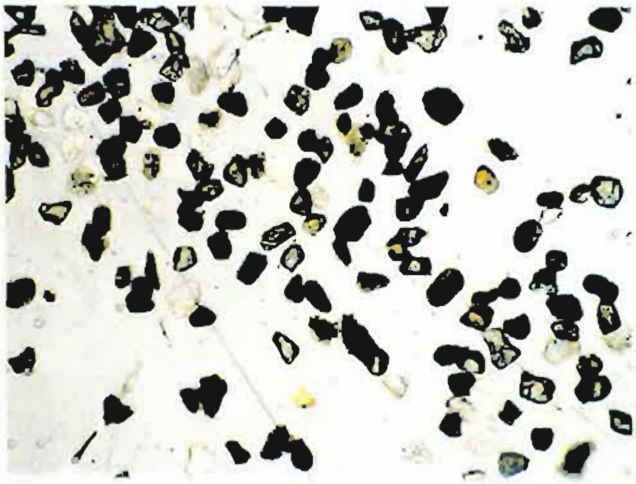
Plate. 7.1:

A, B, C and D plates examples of heavy minerals from Permian sands.
Clear: garnet; Black: ilmenite; Blue and brown green: tourmaline; Purple: zircon and Foxy red: rutile. Scale. 75 x.
E and F plates showing unimodal polycrystaline quartz grains (pq) and microcline with cross hatched twinning (f) in the Permian sands. Scale. 100 x.

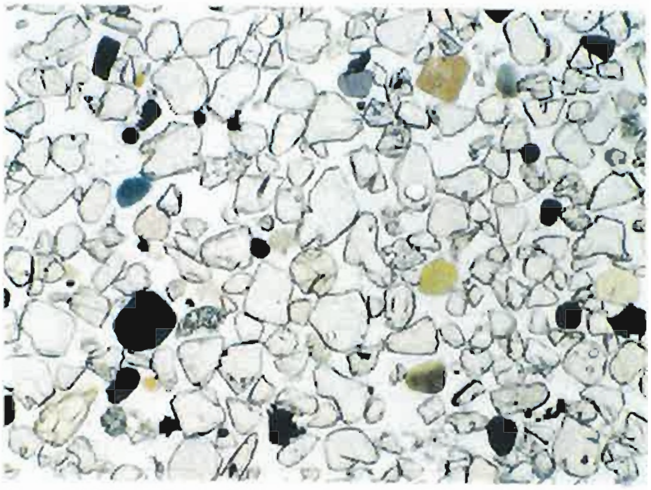


A

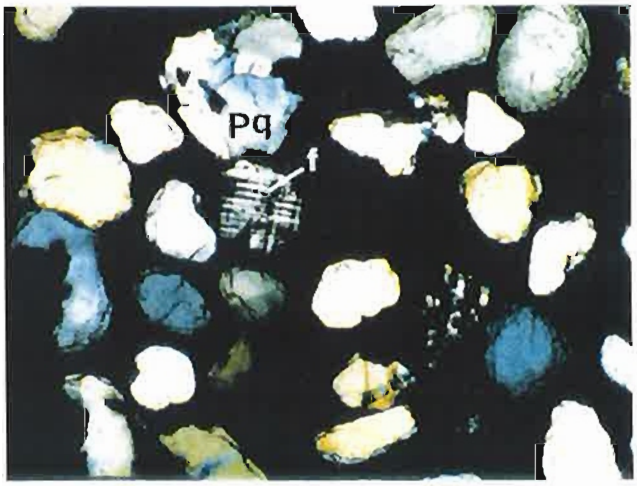
B



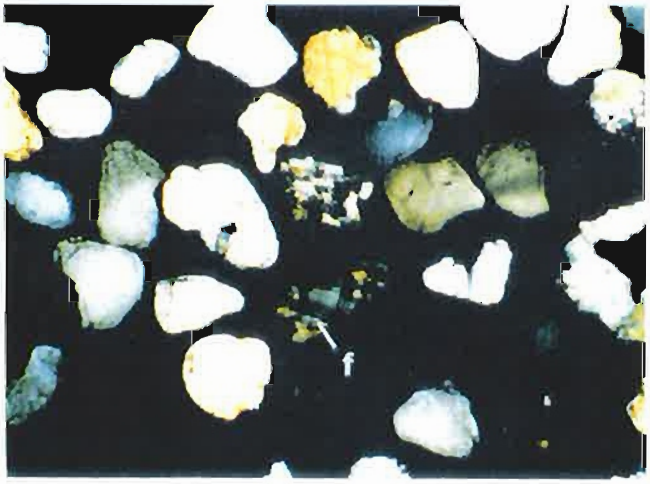
C



D



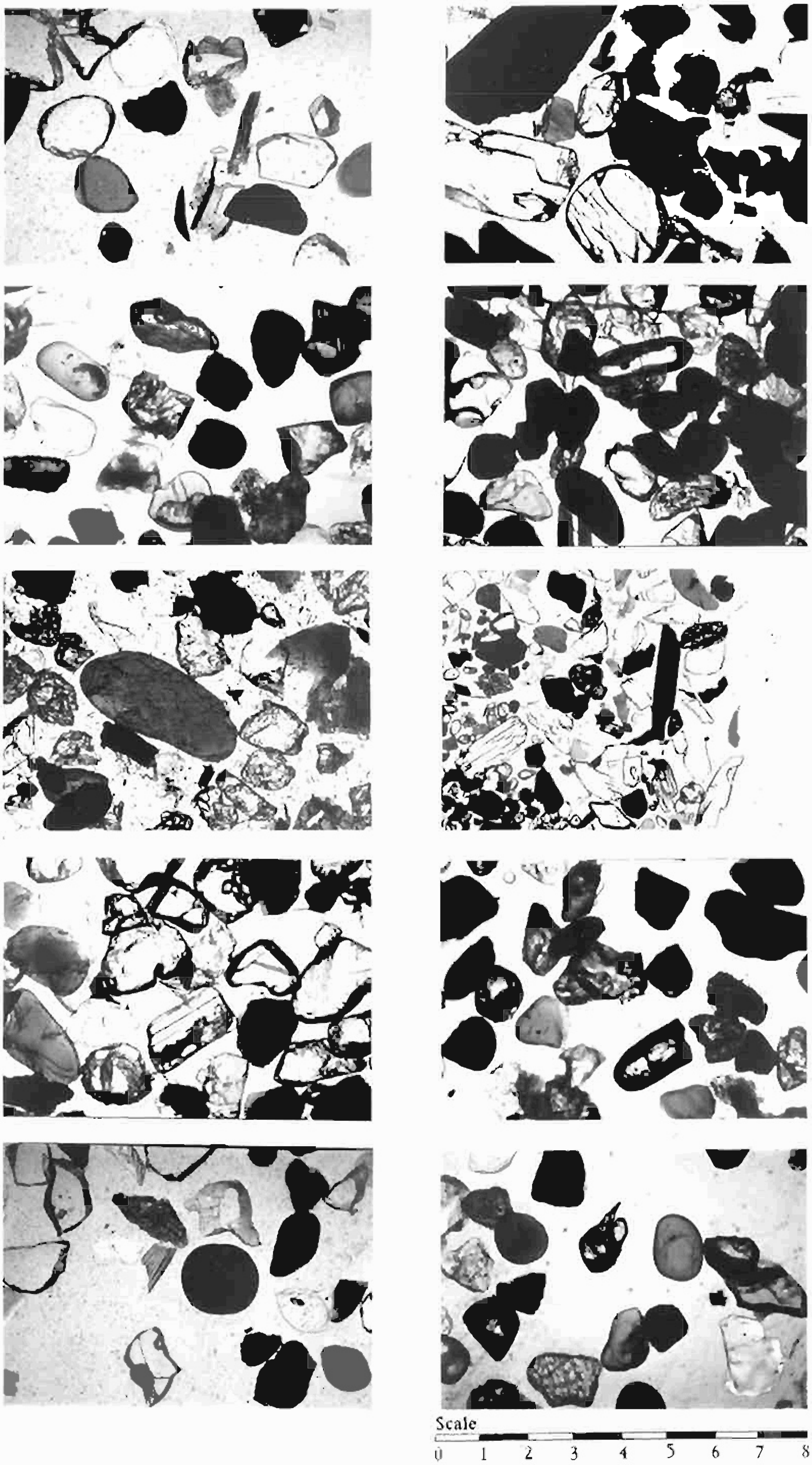
E



F

Plate. 7.1A

Examples of heavy minerals in the Permian sands showing different shapes ranging from rounded to angular and euhedral shapes. Scale. 100 x.


Scale
0 1 2 3 4 5 6 7 8

rocks. Analyses on many garnet grains from the Antarctica samples, mostly from gneisses, indicate similar compositions to some of the garnets in the Permian sands.

The ratio of pink colour to colourless garnet from Permian sands is very similar to the samples from erratics and Antarctica. The ratio of pink to colourless garnets from Permian sands is 4.8, metamorphic samples in Antarctica is 4.6 and erratics is 5.6. The percentage of colourless garnet varies from 84.2% in the metamorphic rocks from Antarctica to 85% in the Permian sands, and to 86.4% in the metamorphic erratics. It is rare to find garnet in Permian volcanic erratics.

Geochemistry of garnets

Electron microprobe analysis of detrital garnet grains is, potentially one of the most powerful techniques presently applicable to provenance studies because, detrital garnets are relatively stable and have a fairly limited density range. Therefore, variation in garnet composition can be ascribed to changes in provenance with more confidence than can variations in the total heavy mineral suite (Morton 1987). After this mineral was identified using a petrographic microscope, its chemical properties was studied by microprobe analysis. The following components were determined: SiO2, Al2O3, FeO (all Fe reported as FeO), MgO, MnO, CaO, TiO2, Cr2O3, K2O and Na2O. Of the minor elements, K2O and Na2O were less than 0.3% by weight. No garnets analysed in the Permian sands and parent rock samples contained appreciable amounts of Cr and Ti, because their Cr2O3 and TiO2 contents are invariably below one percent and generally below 0.5%. Variations in uvarovite and andradite are considered to be insignificant. Therefore the garnets can essentially be characterised in terms of only four end-members, almandine, spessartine, grossular and pyrope.

For comparisons, the compositions of garnets in the Permian sands and samples from East Antarctica, Encounter Bay, Permian erratics and Kanmantoo Group were analysed and plotted on triangular diagrams and histogram (Fig. 7.2 and 7.3). Representative analyses are given in Tables. 7.1.

The heterogeneity of the Permian garnet association is an indication of the great variety of metamorphic rocks available in the eastern part of Antarctica, Kanmantoo Group and

Permian erratics. The major elements of garnets in the Permian sands are Fe and Mg, and minor elements are Mn and Ca, thus, the garnet compositions are expressed in terms of four end members with all Fe calculated as almandine (Fe3Al2Si3O12), all Mg as pyrope (Mg3Al2Si3O12), all Ca as grossular (Ca3Al2Si3O12), and all Mn as spessartine (Mn3Al2Si3O12). The mean percentage of each major element on the basis of 24 (O) is shown in Fig. 7.3. The mean weight percentage of different oxides are shown in Fig. 7.4.

As these Fig. 7.3 and Table. 7.1 show, almandine is the dominant component in the Permian sands as indicated by the Fe, with the mean percentage of 58.2% , the pyrope component as indicated by the Mg has a mean percentage of 25.8%, the spessartine component indicated by Mn seldom exceeds 10%, has a mean percentage 6%, and the grossular component indicated by the Ca has a mean percentage 9.9%. Most of the garnets in the Permian sands contain major Fe and Mg and minor Mn and Ca, thus are in the almandine-pyrope ([Fe, Mg]3 Al2Si3O12) series. Increasing grade of metamorphism is usually accompanied by an increase in the (Fe+Mg) content of the garnet as noted by other investigators (e.g. Brown 1967). The mean and standard deviation of each group of garnets from different localities of Permian sands and parent source rock samples shown in Table. 7.2, confirms that there is little difference between each localitiy of Permian sands, and suggests similar sources and good mixing.

Almandine+spessartine	Pyrope	Grossular	Samples
3.5 (0.8)	1.6 (0.8)	0.4 (0.4)	Permian sands at Hallett Cove
4.0 (0.7)	1.4 (0.7)	0.6 (0.4)	Permian sands at Kings Point
3.8 (0.7)	1.5 (0.7)	0.5 (0.4)	Permian sands at Cape Jervis
3.7 (0.65)	1.6 (0.7)	0.5 (O.5)	mixing average of Permian sands
4.2 (0.2)	1.7 (0.3)	0.2 (0.1)	granitic gneisses of erratics
4.1 (0.1)	1.8 (0.2)	0.2 (0.05)	gneisses of Permian erratics
4.7 (0.1)	1.1 (0.1)	0.2 (0.1)	gneisses in Antarctica
4.6 (0.2)	1.1 (0.2)	0.2 (0.04)	granitic gneisses of Antarctica
4.8 (0.09)	0.9 (0.1)	0.2 (0.0)	granite in Antarctica
5.2 (0.05)	0.4 (0.09)	0.4 (0.1)	phyllites of Commonwealth Bay
4.2 (0.1)	1.5 (0.2)	0.2 (0.04)	gneisses at Victoria Land
5.2 (0.1)	0.6 (0.1)	0.1 (0.1)	Kanmantoo Group
4.6 (0.2)	0.6 (0.5)	2.3 (0.6)	Encounter Bay granites
4.7 (0.2)	0.3 (0.4)	0.8 (0.2)	granite of Permian erratics

Table. 7.2. Mean and standard deviation (S. D.) in garnet groups on the base of 24 (O) in the Permian sands from different localities of southern Australia, and in parent source rock samples.

Almandine+spessartine	Pyrope	Grossular	Samples
64.1	27	0.9	Permian sands
78.3	18.3	3.3	gneisses in Antarctica
68.9	27.9	3.2	gneisses of Permian erratics
71.2	25.4	3.4	gneisses of Victoria Land
86.7	6.7	6.6	phyllites of Commonwealth Bay
88.1	10.2	1.7	Kanmantoo Group
81.3	15.2	3.4	granite in Antarctica
70.8	9.2	20	Encounter Bay granites
81	5.2	13.8	granite of Permian erratics

Table. 7.3. Mean percentage of garnet group in the Permian sands of southern Australia, and in parent source rock samples.

The varieties of garnet groups is shown in Table 7.3 and indicate that the Kanmantoo Group samples are rich in almandine. The metamorphic samples from erratics and Antarctica are rich in almandine with significant pyrope, and the plutonic samples from Antarctica, erratics and Encounter Bay are rich in almandine with significant spessartine and the phyllites of Commonwealth Bay are rich in almandine with significant amounts of spessartine. Comparison between possible sources in Table 7.3 and 7.4 then confirm that the gneissic rocks are most likely to be the main source rocks for the Permian sands.

The mean and standard deviation of FeO, MnO, MgO and CaO of Permian sands and parent rock samples is shown in Table 7.4.

FeO	MnO	MgO	CaO	Samples
26.7 (0.7)	4.4 (0.3)	7 (0.3)	1.4 (0.2)	East Antarctica (gneisses)
29.3 (0.7)	4.8 (0.8)	5.2 (0.8)	1.4 (0.2)	East Antarctica (granitic gneisses)
21.4 (0.5)	14.6 (0.5)	4.2 (0.3)	1.2 (0.1)	East Antarctica (granites)
21.7 (0.6)	14.8 (0.5)	4.1 (0.3)	1.2 (0.1)	East Antarctica (aplites)
13.7 (1.5)	24.5 (1.5)	1.6 (0.3)	2.4 (0.6)	Commonwealth Bay (phyllites)
22.7 (3.2)	10.5 (3.5)	1.6 (0.4)	6.6 (1.7)	Encounter Bay (granites)
34.7 (2)	4.3 (2.4)	1.6 (0.7)	0.7 (0.5)	Kanmantoo Group (metasediments
11.4 (6.9)	22.6 (9.7)	2.4 (3.5)	4.3 (1.4)	Permian erratics (granites)
24.7 (0.6)	6.4 (0.6)	7.9 (0.6)	0.9 (0.1)	Permian erratics (granitic gneisses)
26.6 (4.2)	2.4 (4.4)	7.2 (3.3)	3.4 (2.4)	Permian sands

Table. 7.4. Mean and standard deviation (S.D) of different oxides in garnet grains in the Permian sands of southern Australia and source rock samples.

As can be seen from this table, the chemical composition of garnets in the Permian sands, when compared with different source rocks, indicates mostly metamorphic source rocks. The composition of garnets from Permian sands may be matched quite closely with garnet compositions from gneissic samples from Permian erratics and from Antarctica. Similar garnets have also previously been recorded from gneisses and migmatites of Commonwealth Bay (Stillwell 1918). Based on the Fe, Mn, Ca and Mg contents, (Fig.7.4) indicates that the high Fe and Mg has been recorded from metamorphic rocks rather than plutonic rocks from Antarctica, erratics, Kanmantoo Group and Encounter Bay samples. Comparison of the chemical analyses of garnet concentrated from the Permian sands with the oxides of MgO and MnO, from different samples shows a similar composition pattern to that from metamorphic sources. The concentrate of low manganese content less than 5% by weight and high magnesium contents greater than 15% by weight have been recorded from medium to high grade metamorphic rocks rather than plutonic rocks (Tables 4.1,4.7,4.16,5.3, Fig 4.6,4.13,5.3). Therefore, medium to high grade metamorphic rocks seem to be the most likely source of garnets in the Permian sands. The pink to colourless garnets have an Mg-rich composition similar to the composition of the majority of the garnets within the gneissic samples, and these garnets are not common in the granitic samples. Garnet with almandine and pyrope compositions with very little Ca and Mn substitution are also common in the Permian sands and from the gneissic samples.

The compositions of the detrital garnet suites from Permian sands and in the Antarctica, Encounter Bay, erratics and Kanmantoo Group samples are plotted on triangular diagrams using pyrope (P), grossular (G), and almandine+spessartine (AS). This association contains several distinct groupings: the main grouping is almandine-pyrope with low Ca and Mn contents. Subsidiary groups include almandine-spessartine garnets, low in Mg and Ca, and grossular garnet with variable Mg, Fe and Mn. (Fig. 7.5) shows that the garnet at different localities in the Permian sands has a similar composition. This suggests a similar provenance for the garnet suites at different localities, but this provenance consisted of various rock types.

Garnet with high Ca or Mn contents are typical of granitic pegmatites and skarns, and grossular is also typical of metamorphosed limestones (Deer et al. 1982). The pyrope indicates mostly high grade metamorphic source, whereas the spessartine suggests low grade metamorphic source. Thus the most likely source for the detrital garnets in the Permian sands seems to be high to medium grade metamorphic rocks such as gneisses. The garnet in the granites and pegmatites samples from Antarctica are always almandine in composition. Almandine has been found in plutonic erratics, Encounter Bay, and Antarctica samples. This study shows the remarkable consistancy of one type of garnet in each parent rock samples when compared with the Permian sands. Comparison the chemical composition of garnets in the parent rock samples with Permian sands suggesting that the garnets in the Permian sands were derived mostly from metamorphic rocks in Antarctica, similar to the metamorphic Permian erratics, rather than from Encounter Bay granites and the Kanmantoo Group. Consequently, the study of garnet geochemistry suggests that the Permian sands was not derived from a single source. The heterogeneity of the Permian garnet association is an indication of a great variety of metamorphic and igneous rocks available as seen in the erratics, Kanmantoo Group and in the Antarctica samples.

GARNET COMPOSITION AS RELATED TO REGIONAL GEOLOGY

The garnet composition in Permian sands can be used for understanding the regional geology of source area. The Encounter Bay igneous complex of granitic rocks is characterized by the presence of almandine and spessartine with small amounts of MgO . The Kanmantoo Group is a complex of generally low grade metamorphic rocks characterized by the presence of garnet with a high content of FeO. The East Antarctica complex association of granites, granitic gneisses , gneisses and phyllites is associated with amphibolite dykes. Garnet in phyllites and granitic samples of East Antarctica are characterized by high MnO and in gneisses and granitic gneisses samples specially with silimanite grains represent by high MgO content (Table 7.4.). Therefore garnets are compositionally distinct from each other. Garnet in Permian metamorphic erratic samples has shown similar composition with metamorphic samples of Antarctica and dissimilarity with Kanmantoo Group samples. Table. 7.5 shows the maximum and minimum of different oxides in the garnet grains of Permian sands and parent rock samples and confirms that the geology of the source area is mostly medium to high grade rocks rather than low grade metamorphic rocks with less amount of granite rocks.

FeO	MnO	MgO	CaO		
Max- Min	Max- Min	Max- Min	Max- Min	Number of grains	Samples
35.8-10.9	29.4-0	14.1-0	14.3-0.4	500	Permian sands
31.1-24.1	7.5-3.7	8.8-4.6	1.8-0.7	160	Permian metamorphiic erratics
29.1-25.8	5.3-3.9	7.5-6.1	1.7-0.9	110	Antarctica gneisses
31.1-28.8	6.4-5.2	5.5-3.3	1.5-1.2	150	Antarctica granitic gneisses
22.7-20.6	16.1-14.4	4.8-3.7	1.6-1.1	100	Antarctica granite
38.4-23.2	10.1-1.2	7.8-1.7	10.9-0.2	200	Kanmantoo Group samples
11.7-5.3	31.5-24.1	4.5-0.1	6.1-2.1	25	Permian plutonic erratics
27.4-18.5	16.3-5.1	2.2-1.1	9.1-4.1	20	Encounter Bay samples
15.4-10.8	27.8-22.7	2.2-1.1	3.7-1.2	17	Commonwealth Bay phyllite Antarctica
22.9-20.4	15.5-13.8	4.8-3.5	1.4-1	25	Aplite in Antarctica

Table. 7.5. Maximum and minimum percentage of different oxides in the garnet grains from the Permian sands of southern Australia and parent rock samples.

TOURMALINE

Tourmaline is the second most abundant heavy mineral among non-opaque heavies in the Permian sands. It is a stable mineral and the shape ranges from well rounded to euhedral, with most being euhedral. Tourmaline constitutes 7% of the heavy mineral assemblage in these sediments. The chemical composition of tourmaline in the Permian sands and source rocks is shown in Table 7.6. However, accurate tourmaline analyses are less easy to achieve than those of garnet, because of the presence of boron in the structure. Boron is not detectable with an energy-dispersive system, and although it can be accuratley determined with a wavelength-dispersive analyser the geometry of the target area is far more critical.

Nearly all colour varieties of the main types of tourmaline were found. The colour varies from brown to yellow to olive green, and shows a pleochroic range from green to brown, and yellow to brown. Many names are used for tourmaline varieties, depending on the chemical composition and colour. The only correlations between colour and source of sediments (on which all writers on the subject agree) are yellow and brown colours that occur almost exclusively in metamorphic rocks (Owen 1984). According to Krynine (1946) the colour of detrital tourmaline can be used as an indicator of its mode of origin. Tourmaline with blue paleochroic has a pegmatite origin. Granitic terrains are indicated by green, brown and pink to black colour tourmaline. Whilst pleochroic colourless to yellow and light brown tourmaline is typical of metamorphic terrains. In the high grade metamorphic rocks tourmaline is typically dark brown and green in colour. Comparing the tourmaline grains on the basis of colour with different rock samples indicate a mostly granitic source.

ZIRCON_

Zircon is an ultrastable mineral commonly found in sediments (Marshall 1967; Morton 1985). During the last 70 years several papers have appeared dealing with accessory minerals and using many varietal features of zircon for basing some of the more important conclusions regarding the origin of the sediments. Brammal (1928) described zircon and other accessory minerals from Dartmoor granites. They were followed in such practice by many like Mackie (1927) and Groves (1931). Bruce and Jewitt (1936) used zircon in studing the Precambrian intrusives in the Canadian Shield. Reliance continues to be placed on the morphological features of zircon, such as colour, size, shape, zoning, growth and elongation. Roundness has been most used in establishing the detrital nature of zircons. Armstrong (1922) used roundness in zircons as a criterion of igneous or sedimentary origin. Groves (1931), Poldervaart (1955) have all written about the difference in zircons in sediments and in igneous rocks. Some of the conflicting opinions about the roundness of zircons in granites and granitic gneisses have been listed by Poldervaart (1955) where he quotes Armstrong (1922), Mackie (1927) and others. Poldervaart's conclusion on this aspect is that the intrusive granites contain a majority of euhedral zircons. Eckelmann and Kulp (1956) have used the size and shape of zircons in certain granitic rocks to account for their genesis. They used the following assumptions, which truly reflect the accepted ideas of most workers today.

During transportation, zircon grains will be abraded and gradually become more and more rounded. In igneous rocks, the majority of zircons will be euhedral and in many sediments and metamorphic rocks zircons may remain unchanged. Zircon occurs in much variety in the Permian sands and constitutes 4% of the heavy mineral suite. Although many zircons with different morphologies are found, they can be divided into four distinct groups of angular to rounded, subhedral and euhedral grains. Zircon occurs in a variety of colours (pink, purple, pale yellow and brown) and colourless. There are also a variety of inclusions in zircon grains. Zircon survives erosion, transportation, deposition, with much less alteration than most other accessory minerals. Consequently, detrital zircons may be recycled many times, so the Permian sands may contain zircons from a variety of ultimate source rocks. The use of zircon morphology as a provenance indicator is difficult; however following the reasoning of Poldervaart (1955), the morphologies of most euhedral, colourless zircons from the West Water Canyon Member (U.S.A) indicate that these zircons were derived from acid igneous rocks. According to the classification of Pupin (1976, 1980), zircon grains in the plutonic samples from erratics, Encounter Bay and Antarctica are mostly euhedral and colourless, and are similar to zircons in the Permian sands thereby suggesting a plutonic source. On the other hand, most well rounded zircons have been recycled from sedimentary rocks. Therefore the rounded zircon grains in the Permian sands suggests a sedimentary source. The ubiquitous presence of inclusions in euhedral zircons within the Permian sands indicates that they came from a rhyolitic source as inclusions are not common in granitic zircon (Poldervaart 1955). Purple grains of zircon have been recorded in the Permian sands and metamorphic samples. As first noted by Mackie (1927), the zircons of very old rocks, especially from the Precambrian, tend to be purple coloured. Such coloured, subhedral zircons were probably derived from older, probably Precambrian gneissic rocks in the source area (Tomita 1954), because much of the colour is the result of prolonged exposure to radioactive elements in the zircon lattice.

CHEMICAL COMPOSITION OF ZIRCON

Zircon is a mineral whose optical characteristics were used to identify provenance. However, the chemical composition of zircons has also shown variety. Accurate determination of only trace amounts of Hf can not be achieved using energy dispersive analytical systems, and thus wave length dispersive techniques must be employed. Zircon analyses were performed using an electron microprobe W.D.S. Analytical precision was estimated by maintaining the electron beam on a single spot.

Zircon is one of the end members of the solid solution of the series zircon-hafnon (Correia Neves et al. 1974). High-hafnium zircon, zirconian hafnon, and true hafnon have been identified, but they are rare (Levison and Borup 1960; Correia Neves et al. 1974). The mean Hf content of 463 zircons tabulated by Ahrens and Erlank (1969) is 1.7 weight percent, corresponding to a Zr/Hf ratio of about 40/1. This is approximately the value of their relative abundance in the crust of the Earth, the Moon, and chondritic meteorites (Ehmann and Chyi 1974; Erlank et al. 1978; Speer 1980). Hf/Zr ratio is not a simple function of Hf content however, because a wide variety of elements besides Hf substitute for Zr (Erlank et al. 1978; Speer 1980). The normal range of Hf weight percent for almost all naturally occurring zircons is 0.6-3% (Ahrens and Erlank 1969). The Zr and Hf elements behave nearly identically and always occur together; all Zr minerals contain some Hf (Elinson and Petrov 1969). Despite the similarities between Zr and Hf, slight segregation may occur during crystallization in igneous rocks. Numerous studies have reported an increase in Hf content from basic to acidic igneous rocks (Kosterin et al. 1958, 1960; Gottfried and Waring 1964; Erlank et al. 1978).

The Hf composition of detrital zircons may thus be used in provenance study in a manner similar to varietal studies of other accessory minerals (see Morton 1985).

RESULTS

A large number of zircon grains was analyzed in the Permian sands and parent source rocks to show the diversity in composition of the zircon population. There is a little significant difference in Hf content between samples from the granitic gneiss of Antarctica (1.35%), Permian erratics (1.36%) and this amount is comparable with the Permian sands (1.38%). The mean and standard deviation of Hf_2O_3 and ZrO_2 in Permian sands and each group of samples has been shown in Table. 7.7. As the table shows, the mean percentage and standard deviation of Hf2O3 in Permian sands is very close to the Kanmantoo Group metasediments and granitic gneiss samples. The chemical composition of zircon in Permian sands and the source rocks has been presented in Tables 4.2,4.19 and 7.8. Mean ZrO2/Hf2O3 in zircon from Kanmantoo Group samples is 47.4, Encounter Bay samples is 45.9, Antarctica granitic gneiss samples is 49.7, metamorphic erratics is 49.4, volcanic erratics is 50.7 and in the Permian sands is 48.9. This indicates that the ratio between Zr and Hf content in the Permian sands also can support Kanmantoo Group and granitic gneisses.

As shown in the Fig 7.6 the frequency distribution of hafnium (weight percent) in the Permian sands zircon populations appears to have a Hf distribution that is very similar to the Kanmantoo Group metasediments and granitic gneiss of East Antarctica, but it is unlike the Encounter Bay granites. This distribution suggests that the Kanmantoo Group metasediments and granitic gneisses were the source rocks for most of the zircon populations in the Permian sands of southern Australia.

ZrO2	Hf2O3	Samples
67.7 (0.9)	1.4 (0.2)	Mean and (S.D.) in Permian sands
67.2 (0.8)	1.6 (0.2)	East Antarctic (granitic gneisses)
67.5 (0.8)	1.5 (0.2)	Permian erratic (granitic gneisses)
67.8 (0.6)	1.5 (0.2)	Kanmantoo Group (metasediments)
66.7 (0.5)	1.7 (0.1)	Encounter Bay (granites)
66.9 (0.8)	1.4 (0.1)	Permian erratic (volcanics)
66.5 (0.7)	1.6 (0.2)	East Antarctica (gneisses)

Table. 7.7. Mean and standard deviation of Hf2O3 and ZrO2 in zircon grains from Permian sands of southern Australia and parent rock samples.

RUTILE

High to medium grade metamorphic rocks are the common sources for rutile Milner (1962), Force (1980). Granitic igneous rock: Krumbein and Pettijohn (1938), Milner (1962), Hubert (1960), Pettijohn, Potter and Siever (1987). Mafic igneous rocks:

Krumbein and Pettijohn (1938), Pettijohn (1987). Volcanic rocks in general contain no rutile (Force 1980). Also rutile is rare in granitic pegmatites (Pettijohn 1972). Rutile is unstable under low grade metamorphism but stable at higher grades (Force 1976). Dark coloured to black rutile grains are sometimes opaque and often regarded as being more highly ferriferous varieties. Force (1980) has shown that rutile is supplied to most sediments by the erosion of high rank metamorphic terranes, with only very minor amounts supplied by granitic and low rank metamorphic. In granitic rocks, rutile is extremely sparse, most igneous rocks lack rutile except where they are hydrothermally altered (see Herz 1976 a, b; Herz et al. 1970). Rutile in Permian sands constitutes about 3% of the heavy mineral suite and occurs with a deep red-foxy colour to deep brown. The chemical composition of rutile is shown in Table. 7.9. Rutile has been recorded from mostly high to medium grade metamorphic erratics, samples of Kanmantoo Group and from Antarctica, with the mean percentage of 3.1%. In this study rutile has not been found in any volcanic and plutonic rock samples from different localities. The presence of rutile in the Permian sands and comparison with different rock types strongly suggest that the high to medium grade metamorphic rocks can be the source for rutile and this can be support by the presence of sillimanite, staurolite and garnet in the same rocks.

SILLIMANITE

Aluminous silicates Al2SIO5 are present throughout the Permian sands and the grains are sillimanite and staurolite.

Sillimanite in the Permian sands and source rocks has been studied by the petrographic microscope and microprobe micro-analyses. Sillimanite in the Permian sands and parent rock samples often occurs in long slender crystals or in fine needles. Sillimanite constitutes 2% of the heavy mineral suite in the sands and their colour ranges from brown to grayish, white, greyish-green and pale olive green. The composition of detrital sillimanite grains in the Permian sands and parent source rocks is shown in Table. 7.10.

Sillimanite in the Permian sands has a mostly viridine composition and is very similar to the sillimanite from high grade metamorphic rocks of Antarctica and Permian erratics. Thus the source of the sillimanite in Permian sands was probably from high grade

Thus the source of the sillimanite in Permian sands was probably from high grade metamorphic rocks.

STAUROLITE

Low grade metamorphic rocks are characterized by minerals such sphene, pyroxene and staurolite. These minerals have been described from many parts of the world (Irwin and Coleman 1972).

The percentage of staurolite in the Permian heavy minerals is about 2%. The staurolite in the sands shows pleochroic colours ranging from pale reddish brown to dark brown. In general, the staurolite grains are subrounded. Study of the Kanmantoo Group rocks has shown that the staurolite is the most common heavy mineral in these rocks, and may have been the source of staurolite for the Permian sands. Comparison of the analysis of detrital staurolite in the Permian sands with the Kanmantoo Group samples demonstrates that there are not any variations in composition related to provenance. The staurolite in the Permian sands has usually Mg/Mg+Fe ratio lower than 0.2 which points to formation at relatively low pressure. Fe rich staurolite may occur with quartz at pressures as low as 1 Kb. However Mg-staurolite is not stable at 11Kb and 800 C (Schreyer & Seifert 1969), implying that Mg-content of staurolite is pressure dependent. The chemical compositions of staurolite were indentified by microprobe analyses and are shown in Fig. 7.7 and Table. 7.11. Chemical composition of staurolite grains from Permian sands and Kanmantoo Group have been plotted on triangular diagram on the base of 48 (O) for the elements of Al, Mg and Fe and indicates similar distributions. Therefore the presence of detrital staurolite grains in the Permian sands suggests that they have been derived from local Kanmantoo Group rock samples.

EPIDOTE

Grains of epidote are moderately well rounded to anhedral, and their colour ranges from almost colourless through pale green to strong green or brownish green. Their index of refraction indicates that they are high in ferric iron. The percentage of epidote is very low (0.5%) in the heavy mineral assemblages of the Permian sands. The chemical composition of epidote groups is shown in Table. 7.12. This group comprises the minerals clinozoisite, piemontite and epidote a notable feature of the epidote group. The mean percentage and standard deviation of Al2O3, FeO and MnO in the epidote group in Permian sands and parent rock samples is shown in Table. 7.13. As the table shows, Permian sands and metamorphic rock samples are rich in Al2O3 and indicate mostly clinozoisite in composition. Comparison of the (Fig. 7.8) indicate that the epidote groups in the Antarctica metamorphic samples and Permian erratic samples are nearly similar in composition but they show dissimilarity with the Permian sands, this suggests that some of the epidote grains in Permian sands should have come from other sources. According to Pettijohn (1941), epidote groups are quite unstable minerals during transportation and may have been removed.

Al2O3	FeO	MnO	Samples
23.7 (4.2)	5.1 (2)	2.4 (3)	Permian sands
23.2 (2.3)	11.4 (2)	0.2 (0.2)	Antarctic granitic gneisses
21.6 (1)	14 (1.1)	0.4 (0.3)	Permian erratic (granitic gneisses)
23.9 (0.3)	11.6 (0.8)	0.4 (0.3)	Kanmantoo Group metasediments
20.6 (3.6)	14.3 (1.4)	0.4 (0.3)	Victoria Land (gneisses)

Table. 7.13. Mean and standard deviation of epidote group in the Permian sands of southern Australia and parent rock samples

APATITE

Apatite constitutes 0.6 % of the heavy mineral population in the Permian sands. The lower percentage of apatite is not due to provenance changes, but rather to the instability of apatite under acidic conditions (Nickel 1973). Apatite appears as colourless to slightly pleochroic pink or brown under the petrographic microscope. Significantly, more apatite has been found in some source areas rather than in Permian sands. The chemical composition of apatite in Permian sand and parent rock samples is shown in Table. 7.14. The source of apatite in the Permian sands is most likely to be granitic rocks.

AMPHIBOLE AND PYROXENE

Amphibole and pyroxene are present in the Permian erratics and from Antarctic samples, but only trace amounts of them have been identified in the heavy mineral assemblages of the Permian sands. Probably because they are unstable and were removed by action of postdepositional intrastratal solutions, or were lost prior to deposition by mechanical abrasion, or a combination of both (Gravenor and Gostin 1979). Unstable minerals, such as amphibole and pyroxene, are sometimes removed by weathering and thus bias the determination of source rock lithology (Pettijhon et al. 1987). However the only trace of unweathered pyroxene (sodium rich hedenbergite and especially actinolite in the Permian sands see Table. 7.15 suggests that the Permian sands also had a local origin such as the basic igneous erratics, and this has not been affected by the action of intrastratal solutions.

MICAS

The only trace of biotite and muscovite in Permian sands (these are easily destroyed during long distance transport) confirms that they were only carried a short distance and should have local sources. The chemical composition of micas is shown in Table. 7.16.

The chemical composition of spinel and sphene are shown in Tables7.17 and 7.18.

DISCUSSION

Heavy mineral analysis is a sensitive and well proven technique for determining the provenance of clastic sediments. The interpretation of the data is considerably enhanced by determining the composition of individual detrital grains. This study have shown that microprobe analysis of detrital heavy minerals, especially garnet grains, adds considerable sophistication to provenance determinations.

Based on this study, the dominant heavy minerals in the Permian sands suggests that they were derived mainly from medium to high grade metamorphic and granitic rocks. The sources of the heavy minerals in the Permian sands were most probably the rocks of eastern Antarctica, together with local South Australian Kanmantoo Group and Encounter Bay Granite, and recycled grains.

The data obtained in this investigation indicate that the source of garnets in the Permian sands were mostly from medium to high grade metamorphic rocks. The presence of garnet, sillimanite and rutile in the Permian sands, and comparison of the chemical composition of these minerals with the Antarctica metamorphic rocks supports the theory that part of the Permian sands should be derived from these rocks. The garnet concentrate in the Permian sands has shown a similar composition pattern to that from gneissic rock samples in Antarctica and in the Permian erratics. In this concentrate many garnets grains have (Fe+Mg) greater than 5.4 % and the maximum MgO content is 14.1 % which also supports a gneissic source when compared with igneous and phyllite source rocks. The petrography of the East Antarctica rocks has shown that these rocks contain abundant garnet, zircon, tourmaline, sphene, rutile and sillimanite, whereas Kanmantoo Group samples contain garnet, staurolite, rutile, epidote, opaques zircon and tourmaline. However, the variation in the percentage of heavy mineral proportions in the Permian sands is largely independent of weathering and is therefore, the result of derivation from different source areas. So the primary conclusion of the chemical and microscopic characteristics to the proposed source areas for the Permian sands in the southern Australia reveals that the source rocks for these sediments indicates derivation mostly from metamorphic and plutonic rocks rather than from volcanic rocks.

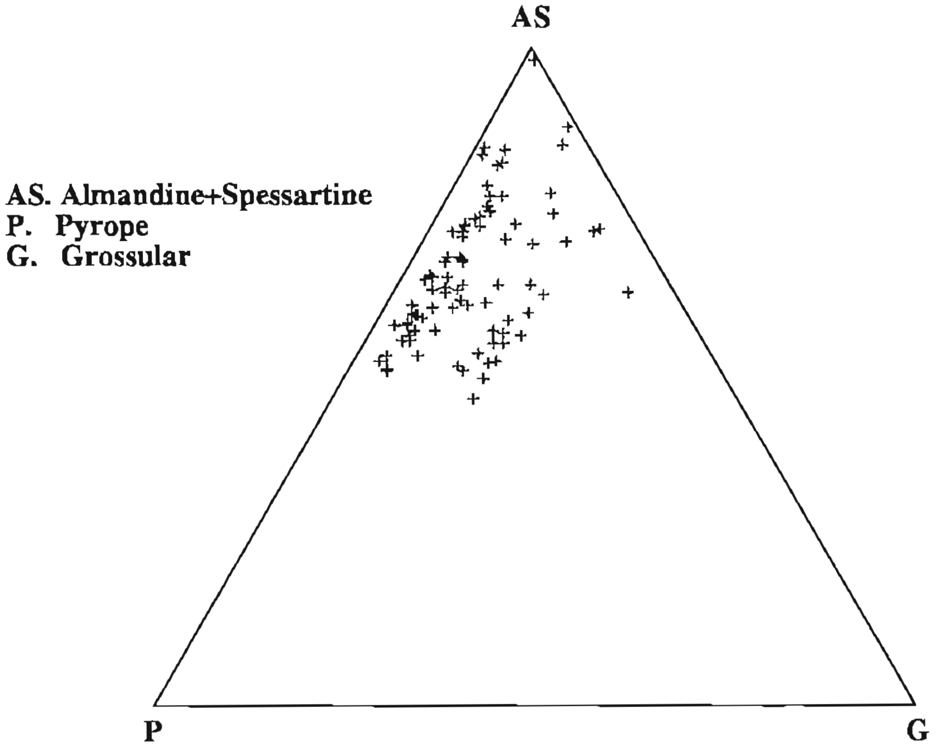


Fig. 7.2. Garnet geochemistry in the Permian sands of southern Australia displayed on a triangular diagram: AS= Almandine+Spessartine; P= Pyrope and G= Grosslar.

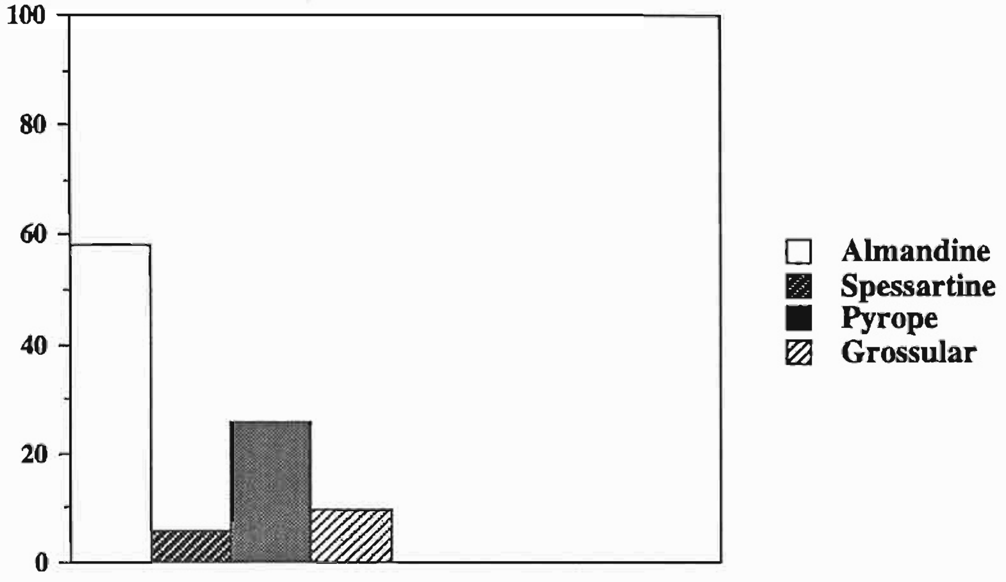


Fig. 7.3. Distribution of garnet groups in the Permian sands of southern Australia.

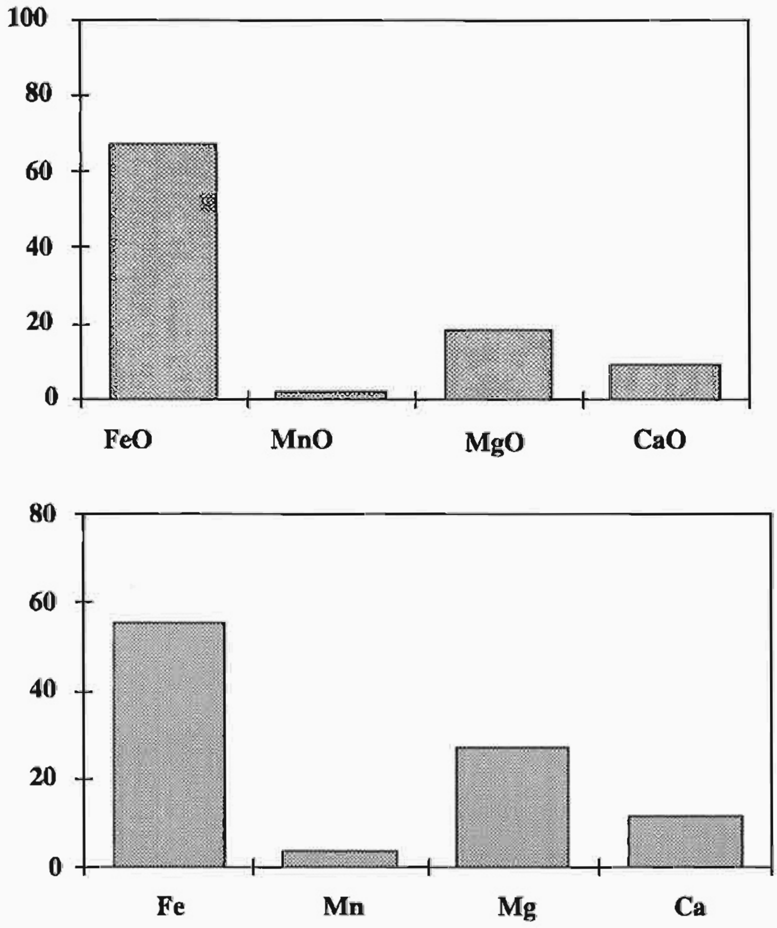


Fig. 7.4. Weight percentage of different elements and oxides on the base of 24 (O) of Fe, Mn, Mg and Ca from detrital garnet grains in the Permian sands of southern Australia.

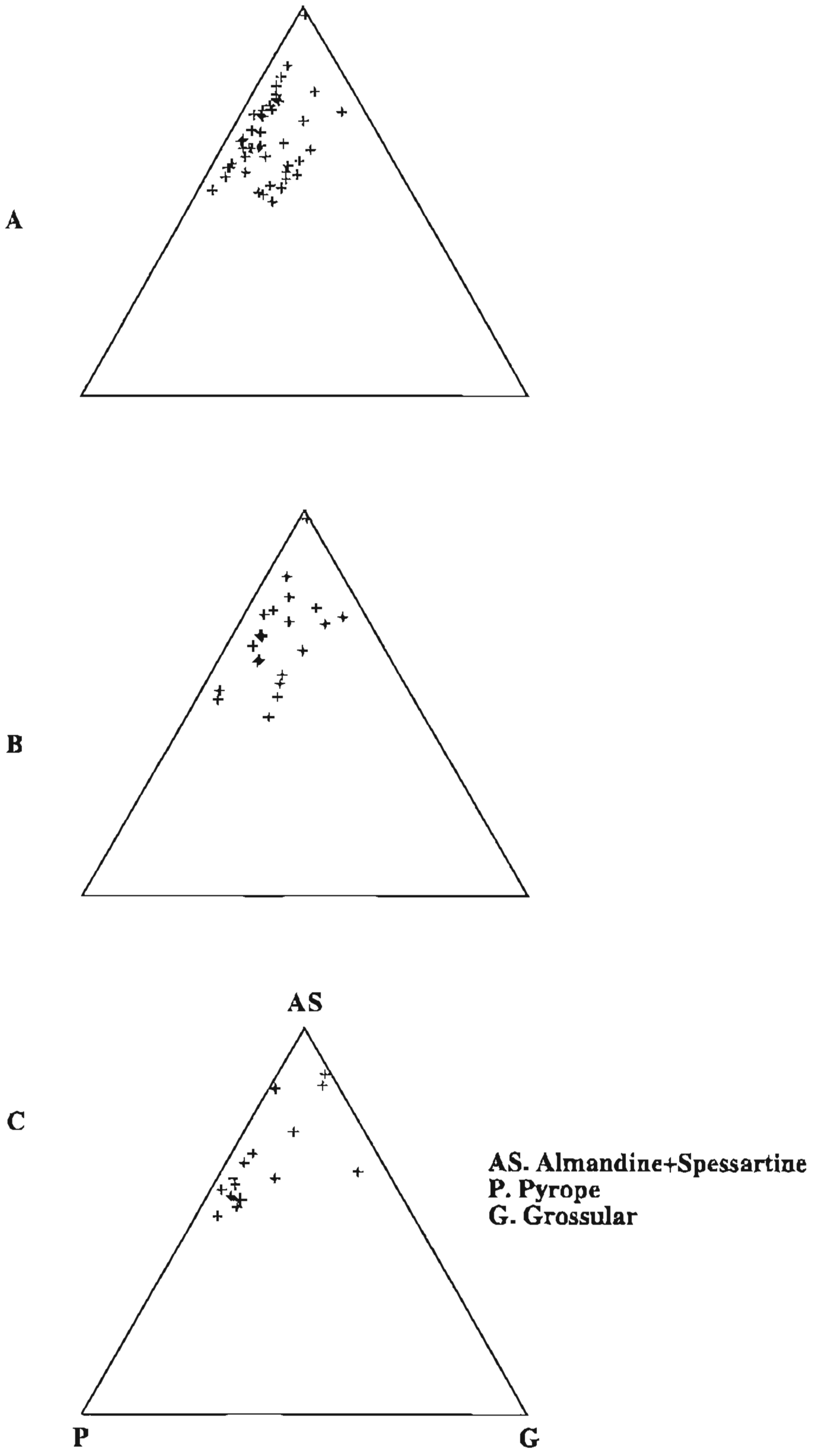


Fig. 7.5. Garnet geochemistry from different localities in the Permian sands of southern Australia displayed on a triangular diagram: AS= Almandine+Spessartine; P= Pyrope and G=Grossular.
A: Cape Jervis, B: Kings Point, C: Hallett Cove.

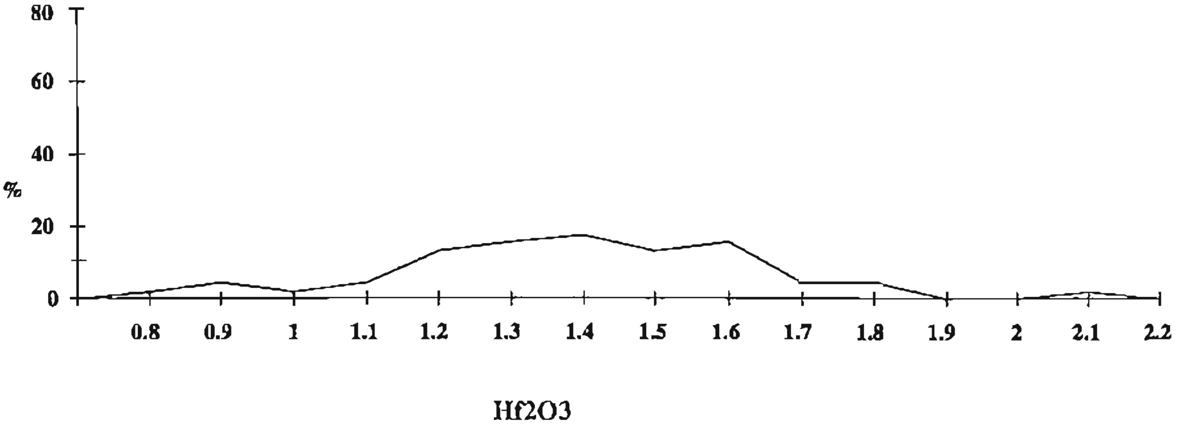


Fig. 7.6. Percentage of hafnium content in the detrital zircon grains from Permian sands of southern Australia.

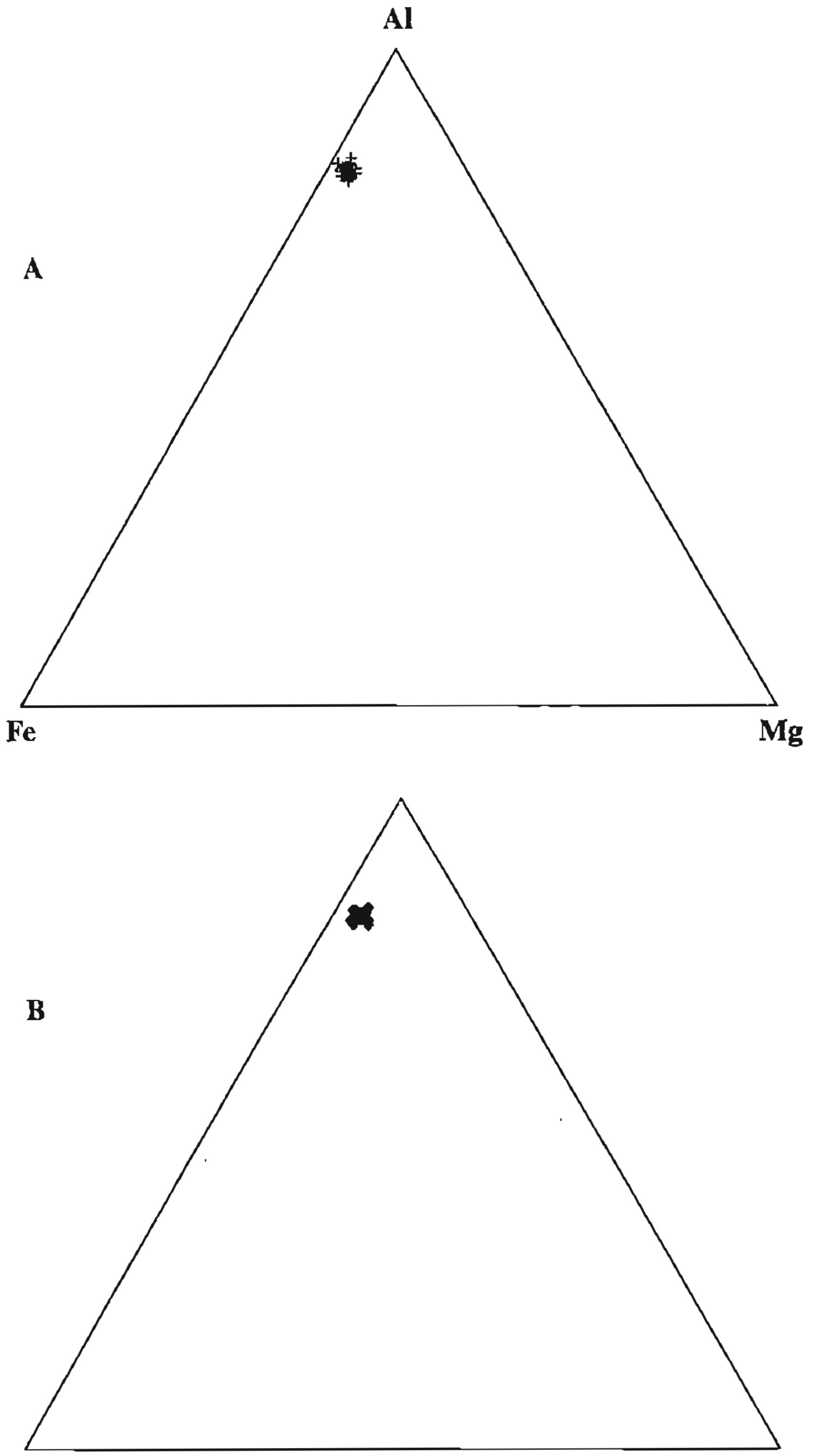


Fig. 7.7. Staurolite geochemistry displayed on Al, Fe and Mg elements on the base of 48 (O) in Kanmantoo Group metasediments and Permian sands of southern Australia.
A: Kanmantoo Group
B: Permian sands

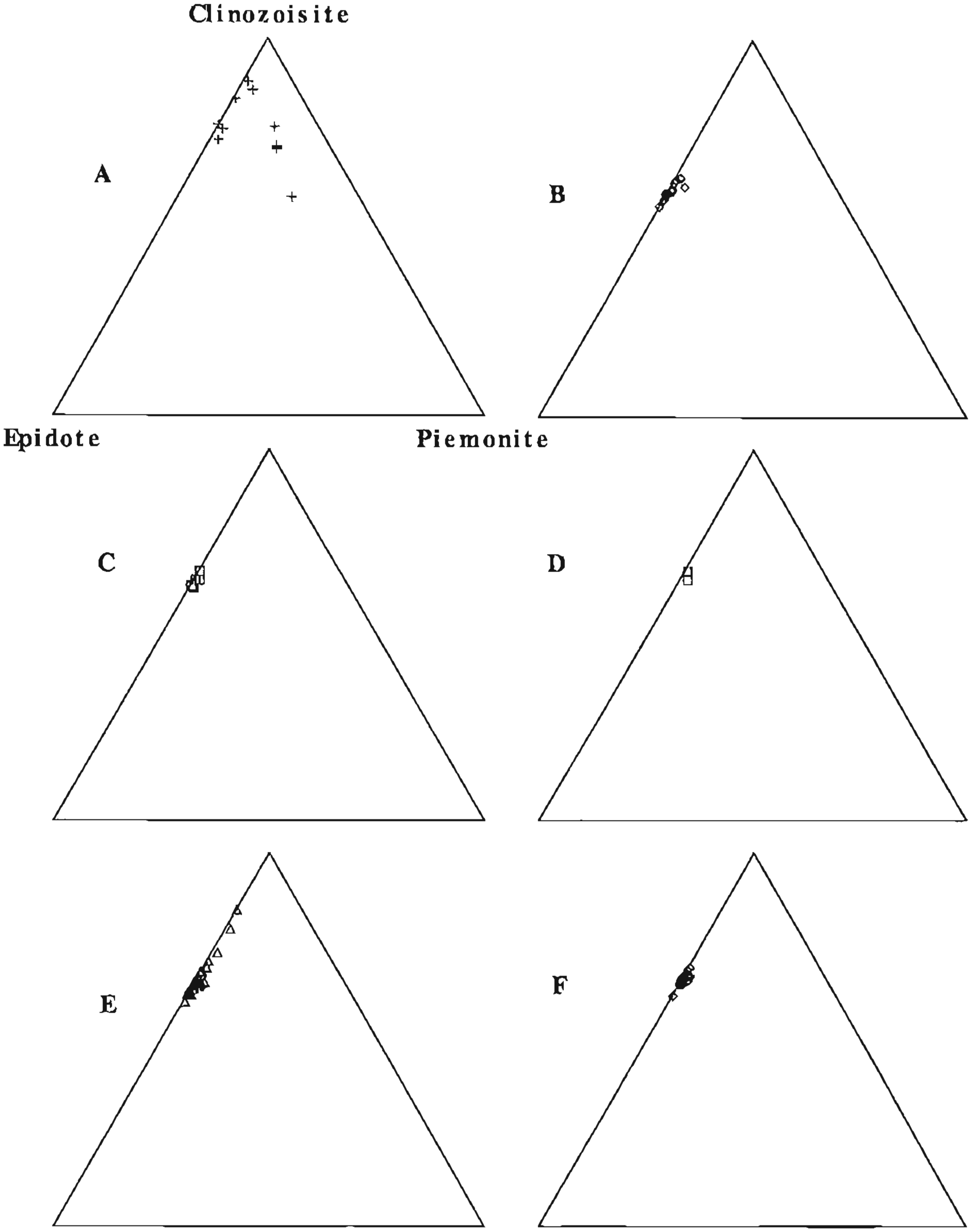


Fig. 7.8. Epidote geochemistry in the Permian sands of southern Australia and source rock samples.
A: Permian sands; B: granitic gneisses of Permian erratic; C: granitic of East Antarctica; D: granitic from Encounter Bay; E: granitic gneisses of East Antarctica; F: basic gneisses of East Antarctica.

Table.7.1. Composition of detrital garnet grains in the Permian sands of southern Australia.

SiO2	38.3	38.3	41.3	39.6	37.8	39.7	37.2	38.3	38.9	39.4	40.1	38.9	37.6	38.8	38.6	38.3	38.1	37.3
TiO2	0.5	0.2	0.5	0	0	0.2	0	0	0	0	0	0	0	0	0	0	0	0.4
Al2O3	22.5	22.2	23.6	22.9	21.6	23.3	21.8	22.7	23.2	22.5	21.8	22.1	21.1	22.1	23.1	22.9	21.3	22.1
FeO	26.5	24.2	14.3	22.9	28.4	25.3	32.9	25.1	27.4	27.1	26.8	24.7	32.1	27.5	25.9	24.6	29.1	25.1
MnO	0	0.6	0.2	0.1	4.7	0.7	1.8	2.9	0.4	0.1	0.1	0.7	0.3	1.1	1.4	1.1	1.4	1.1
MgO	11.2	10.7	15.5	14.2	1	9.5	5.2	6.8	9.5	10.3	2.5	7.8	5.2	5.3	8.6	8.2	2.7	6.3
CaO	1.3	3.7	5.1	0.9	6.9	1.8	1.2	4.5	2.2	0.9	10.5	6.9	2.5	6.3	3.6	5.2	7.5	7.5
Na2O	0	0	0	0.1	0	0	0.9	0	0	0.2	0	0	0.3	0	0	0.3	0	0
K2O	0.2	0	0	0	0.3	0	0.1	0	0	0	0.1	0	0	0	0	0	0	0
Cr2O3	0.2	0.2	0.3	0.4	0.2	0	0	0	0	0	0	0	0.4	0	0	0	0	0.1
Total	100.7	100.1	100.8	101.1	100.9	100.5	101.1	100.3	101.6	100.5	101.9	101.1	99.5	101.1	101.2	100.6	100.1	99.9

No. of ions on the basis of 24 (O)

Si	5.8	5.8	6	5.9	6	6	5.8	5.9	5.9	6	6	6	5.9	6	6	6	6	5.9
Ti	0.06	0.02	0.06	0	0	0.02	0	0	0	0	0	0	0	0	0	0.03	0	0.04
Al	4.01	4	4.01	4.01	4.9	4.2	3.9	4.1	4.1	4.04	4	4	4.1	4.1	4.2	4	3.9	4.1
Fe	3.3	3.1	1.7	2.8	3.9	3.3	4.2	3.2	3.5	3.4	3.5	3.1	4.2	3.6	3.2	3.2	3.8	3.2
Mn	0	0.08	0.02	0.01	0.6	0.09	0.2	0.4	0.05	0.08	0.08	0.08	0.04	0.1	0.2	0.1	0.2	0.1
Mg	2.5	2.4	3.3	3.1	0.3	2.1	1.2	1.5	2.1	2.3	0.6	1.7	1.2	1.2	1.9	1.8	0.6	1.4
Ca	0.2	0.6	0.8	0.1	1.2	0.3	0.2	0.7	0.3	0.1	1.7	1.1	0.4	1.1	0.6	0.8	1.3	1.2
Cr	0	0.03	0.03	0	0.02	0	0	0	0	0	0	0	0	0	0	0	0	0.01

Mol.% end-members

Al	55	50	29.2	46.6	65	57	72.4	55.2	58.8	58.5	59.3	56.6	72	60	54.2	54.2	66	54
As	0	1.3	0.3	0.2	10	1.5	3.4	6.9	0.8	0.2	1.3	1.4	0.7	16.7	3.4	1.7	3.4	1.7
P	42	38.8	56.7	51.6	5	36.3	20.7	25.9	35.3	39.6	10.2	10.9	20.5	20	32.2	30.5	10.7	23.7
G	3	9.7	13.7	1.7	20	5.2	3.4	12.1	5	1.7	28.8	31	6.8	18.3	10.2	13.6	22	20.3

Table.7.1. Continue...

SiO2	47.8	38.9	38.1	36.1	38.5	39.4	39.6	38.2	37.4	37.3	37.7	38.5	37.1	38.5	36.9	37.8	37.8	37.2
TiO2	0.9	0.3	0.1	0	0	0.1	0.1	0	0.1	0	0	0.1	0.3	0	0.2	0	0	0
Al2O3	9.2	23.2	21.8	22.8	21.5	23.9	22.2	21.9	21.4	21.6	23.6	22.6	22.6	22.6	21.2	21.4	21.3	21.5
FeO	11.6	21.6	32.9	28.2	28.5	23.6	24.6	25.7	13.5	31.9	29.7	28.8	25.6	29.8	26.4	26.1	26.1	32.7
MnO	0.4	0.8	1.8	2.4	0.9	0.6	0.3	0.3	22.8	0.7	0.3	1.4	3.3	1.1	8.3	10.2	4.6	0.5
MgO	15.2	5.5	3.7	7.1	7.1	11.1	11.6	7.2	0.6	1.6	6.1	7.1	7.7	7.2	0.3	3.2	5.1	2.7
CaO	12.3	10.1	2.8	3.1	2.4	2.2	1.5	5.4	4.4	7.2	3.3	3.1	4.1	1.6	6.3	1.6	6.5	3.9
Na2O	0.4	0.5	0	0	0.2	0	0	0.1	0	0	0	0	0	0	0	0	0	0
K2O	0.4	0	0	0	0.2	0	0	0	0	0	0	0	0	0	0	0	0	0
Cr2O3	0	0	0	0.2	0.5	0	0.2	0.3	0.4	0.3	0.6	0	0.1	0	0.1	0	0.1	0
Total	98.5	100.9	101.2	99.9	99.8	100.9	100.1	99.1	100.6	100.6	101.3	101.6	100.8	100.8	99.7	100.3	101.5	98.5

No. of ions on the basis of 24 (O)

Si	7.2	6	6	5.9	6	6.1	6	6	5.9	5.9	5.9	6	5.9	6	5.8	5.9	5.9	6
Ti	0.1	0.01	0.01	0	0	0.01	0.01	0	0.01	0	0	0.01	0.03	0	0.02	0	0	0
Al	1.7	4	4	4	4	4.2	4	4	4	4.1	4.2	4	4	4	4	4	4	4
Fe	1.6	2.8	4.3	3.7	3.7	3.1	3.1	3.2	3.3	4.2	3.8	3.6	3.3	3.8	3.4	3.4	3.4	4.2
Mn	0.04	0.08	0.1	0.2	0.08	0.07	0.04	0.04	0.03	0.1	0.03	0.2	0.4	0.1	1.1	1.3	0.6	0.06
Mg	3.4	1.2	0.8	1.6	1.6	2.4	2.5	1.6	1.7	0.4	1.4	1.6	1.7	1.6	0.1	0.7	1.1	1.1
Ca	2	1.6	0.5	0.6	0.4	0.4	0.2	0.8	0.6	1.2	0.5	0.5	0.7	0.3	1.1	0.3	1.1	0.6
Cr	0	0	0	0.02	0.04	0	0.02	0.02	0.02	0.04	0.07	0	0.01	0	0.01	0	0.01	0

Mol. % end-members

Al	22.7	49.3	75.4	60.6	62.7	52.5	53.4	57.1	58.9	71	66	61	54	66	60	60	54	70.5
As	0.5	1.4	1.7	3.3	0.3	0.1	0.1	0.1	0.1	2	1	3	7	2	19	23	10	1
P	48.3	21.1	14	26.2	27.1	40.6	43.1	28.6	30.3	7	24	27	28	28	2	12	18	18.4
G	28.4	28.2	8.8	9.8	6.8	6.8	3.4	14.3	10.7	20	9	9	11	4	19	5	18	10

Table. 7.6. Composition of detrital tourmaline grains in the Permian sands of southern Australia.

SiO2	34.4	35.1	36.5	34.1	36.1	34.2	34.6	34.8	37.8	37.6	40.3	34.9
TiO2	0.9	0.8	0.5	0.3	0.6	1.1	0.8	1.1	0.5	1.3	0	1.1
Al2O3	28.1	29.2	34.2	35.1	29.6	31.3	34.6	34.9	36.1	36.4	39.2	30.3
FeO	9.6	11.3	8.8	10.2	9.3	13.1	10.6	10.5	10.1	12.7	6.3	8.5
MnO	0.2	0.7	0	0.1	0	0.1	0.2	0	0.3	0.2	0	0
MgO	7.5	7.2	5.2	2.7	8.4	3.4	3.5	3.9	3.8	4.1	8.1	6.9
CaO	2.7	2.1	1.5	1.4	3.2	0.5	0.4	0.9	0	0.1	1.2	1.6
Na2O	0.9	0.8	0.5	0.1	0.9	2.9	1.8	1.6	3.2	2.1	2.2	2.2
K2O	0.1	0	0	0	0	0	0	0	0	0	0.2	0
Cr2O3	0.1	0	0.3	0.2	0.3	0.1	0.2	0	0.5	0	0	0
Total	84.7	88.1	86.3	84.3	88.1	86.7	87.1	87.8	92.3	94.7	97.5	85.6

Table. 7.6. (Cont.) Composition of detrital tourmaline grains in the Permian sands of S.A.

SiO2	34.1	33.5	35.1	34.9	35.8	35.4	34.8	34.6	34.2	35.4
TiO2	0.3	1.3	1.7	1.1	1.2	1.3	0.5	0.8	1.1	1.3
Al2O3	35.1	34.2	28.2	30.3	35.3	34.2	33.3	34.6	31.3	34.2
FeO	10.2	8.8	11.3	8.5	5.2	7.2	6.2	10.6	13.1	7.2
MnO	0.2	0	0.7	0	0	0.1	0.2	0.2	0.1	0.1
MgO	2.7	5.2	7.2	6.9	5.7	5.9	6.7	3.5	3.4	5.9
CaO	0.4	0.5	2.1	1.6	1.1	1.1	0.1	0.5	0.5	1.1
Na2O	1.1	1.8	1.9	2.2	2.6	1.6	2.4	1.8	2.9	1.6
K2O	0	0	0	0	0	0	0	0	0	0
Cr2O3	0.2	0.4	0	0	0	0	0	0	0	0.1
Total	84.3	86.1	88.2	85.5	86.9	86.8	84.2	86.5	86.6	88.1

Table. 7.8. Composition of detrital zircon grains in the Permian sands of southern Australia.

SiO2	29.1	29.8	29.8	29.7	29.1	29.3	29.3	29.3	29.1	28.9	29.1	29.9	28.9	29.7	29.2	28.9
ZrO2	67.6	66.7	66.6	66.7	68.1	69.2	69.1	68.1	69.4	67.9	69.1	70.5	71.2	66.8	66.8	66.8
Al2O3	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
FeO	0.4	0.3	0.1	0.2	0.1	0.2	0.2	0.1	0.1	0.1	0	0.1	0.1	0.1	0.1	0.1
MnO	0	0	0	0	0	0	0	0	0	0	0	0	0	0.1	0	0
MgO	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Hf2O3	1.3	1.2	1.5	1.5	1.4	1.1	1.5	1.6	1.6	1.7	1.6	1.3	1.2	1.4	1.5	1.7
Th2O3	0	0.1	0	0.1	0.1	0	0	0	0.1	0	0.1	0	0.1	0	0.1	0
Total	98.5	98.3	98.1	98.2	98.7	99.8	100.1	99.1	100.2	98.6	99.8	101.9	101.5	98.1	97.8	97.6

Structure furmula on the basis of 16 (O)

Si	3.7	3.7	3.7	3.7	3.7	3.7	3.7	3.7	3.7	3.7	3.7	3.7	3.6	3.7	3.7	3.7
Zr	4.2	4.2	4.2	4.2	4.2	4.2	4.2	4.2	4.3	4.2	4.3	4.3	4.3	4.2	4.2	4.1
Al	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Mg	0	0	0.001	0	0	0	0	0	0	0	0	0	0	0	0	0
Hf	0.05	0.04	0.05	0.05	0.05	0.04	0.05	0.06	0.06	0.06	0.06	0.05	0.04	0.05	0.06	0.06
Fe	0.05	0.03	0.01	0.02	0.007	0.02	0.02	0.008	0.01	0.007	0	0.01	0.01	0.01	0.01	0.01
Mn	0	0.001	0.002	0.003	0.002	0	0.005	0.004	0	0.004	0	0	0	0.01	0	0

Table. 7.8. continue...

SiO2	28.1	29.1	28.4	29.6	28.9	28.7	28.9	29.9	29.4	29.4	29.4
ZrO2	66.5	67.9	68.1	68.1	68.6	67.3	66.1	67.1	67.1	67.9	68.1
Al2O3	0.1	0	0	0	0	0	0	0	0	0	0
FeO	0.4	0.4	0.2	0.1	0.3	0.1	0.4	0.2	0.2	0.2	0.1
MnO	0.1	0	0	0	0	0	0	0	0	0	0
MgO	0	0	0	0	0	0.01	0	0	0	0	0
Hf2O3	1.8	1.2	1.4	1.4	1.5	1.3	1.6	1.4	1.3	1.2	0.9
Th2O3	0.1	0	0	0.1	0.1	0	0	0	0.1	0	0.1
P2O5	0	0	0	0	0	0	0	0	0	0	0
Total	97.1	98.5	98.1	99.4	99.4	97.5	97.2	98.7	98.1	98.8	98.7

Structure formula on the basis of 16 (O)

Si	3.6	3.7	3.7	3.7	3.7	3.7	3.7	3.8	3.7	3.7	3.7
Zr	4.2	4.2	4.3	4.2	4.2	4.2	4.2	4.2	4.2	4.3	4.3
Al	0.01	0	0	0	0	0	0	0	0	0	0
Mg	0	0	0	0	0	0.002	0	0	0	0	0
Hf	0.07	0.04	0.05	0.05	0.05	0.05	0.06	0.05	0.05	0.04	0.03
Fe	0.04	0.04	0.02	0.02	0.03	0.005	0.04	0.02	0.02	0.02	0.005
Mn	0.01	0	0	0	0	0	0	0	0	0	0.002

Table 7.9 Composition of detrital rutile grains in the Permian sands of southern Australia.

TiO2	FeO	MnO	MgO	Al2O3	SiO2	CaO	Na2O	K2O	Cr2O3	Total
99.8	0	0	0	0	0	0	0	0.1	0	99.9
99.7	0	0	0	0.2	0	0	0.2	0	0	100.1
99.9	0.3	0.1	0.2	0	0	0	0	0	0.2	100.7
99.8	0	0	0	0	0	0.3	0	0	0	100.1
99.7	0.1	0.6	0	0.2	0	0	0	0	0	100.4
99.3	0.4	0.1	0	0.2	0	0	0	0	0	100.1
99.3	0.5	0.1	0	0.2	0	0	0	0	0	100.1
99.3	0.3	0.1	0	0.2	0	0	0	0	0	99.9
99.3	0	0	0	0	0.1	0.2	0	0.1	0.1	99.8
99.2	0.3	0	0	0.2	0.2	0	0	0	0	99.9
98.9	0	0.4	0	0	0	0	0	0	0.1	99.6
98.8	0	0	0	0.2	0	0	0	0	0	99.2
98.8	0.2	0.3	0	0.2	0	0	0	0	0	99.7
98.8	0	0.2	0.2	0.2	0.2	0.2	0.6	0	0.4	101.5
98.7	0	0.6	0	0.2	0	0	0	0	0	99.7
98.7	0.7	0.2	0.2	0.2	0	0	0.7	0	0.2	100.9
98.7	0	0.2	0	0	0	0.2	0	0.2	0	99.3
98.7	0.6	0	0	0	0	0	0	0	0.7	100.1
98.7	0	0.2	0	0	0	0.2	0	0.1	0	99.3
98.6	0.8	0.2	0	0.1	0	0	0	0	0.5	100.3
98.6	0.9	0	0	0	0	0.3	0	0.1	0.1	101.1
98.5	0	0.2	0	0.3	0.2	0	0	0	0.5	99.9
98.5	0.3	0.3	0	0	0	0	0	0	0.4	99.7
98.5	0	0	0.3	0.2	0.1	0	0	0	0	100.1
98.4	0	0.2	0	0.3	0.2	0	0	0	0.5	99.9
98.3	0	0	0.2	0.2	0	0	0	0	0	99.4
98.3	0	0	0.2	0.2	0	0	0.3	0	0	99.4
98.3	0.1	0.1	0.2	0.2	0	0	0.3	0	0	99.1
98.2	0.2	0.2	0	0	0.2	0	0.4	0	0	99.5
98.2	0	0.4	0	0	0	0	0	0	0.2	98.8
98.1	0.6	0	0	0	0	0	0.4	0	0	99.2
98.1	0	0.1	0	0	0	0	0	0	0.2	99.1
98.1	0.4	0.1	0	0	0	0.1	0	0	1	99.2
97.8	0.6	0.3	0	0.2	0.2	0.1	0	0	0.3	99.7
97.7	0.6	0	0	0.3	0.3	0	0.2	0	0	99.2
97.7	0.3	0	0.3	0.4	0.8	0	0.4	0	0.5	100.4
97.6	0.1	0.4	0.3	0.2	0.1	0.2	0	0	0.7	99.4
97.6	0.6	0	0.3	0.3	0.1	0	0	0	0	98.9
97.4	0	0	0.3	0.6	0.6	0	0.7	0	0	99.8
97.2	0.9	0.6	0	0	0.5	0	0	0	0	99.3
97.2	0.1	0.9	0.2	0.2	0	0.2	0.1	0	0.3	98.8
97.1	0	0	0	0.2	0	0	0	0	0	98.1
97.1	0.9	0.2	0.2	0.2	0	0	0	0	0.3	98.7
98.1	0.4	0	0.1	0.3	0	0	0	0	0.2	99.1
96.7	0.7	0	0.1	0.2	0.1	0	0	0	0.4	98.2
99.5	0.5	0.1	0	0	0	0	0	0	0	100.1
97.4	0.6	0	0.2	0.5	0	0.1	0	0	0	99.2
98.1	0	0	0.4	0.1	0	0.1	0.3	0	0.1	99.1
97.1	0	0.5	0	0.5	0.3	0	0.1	0	0.4	98.9
98.7	0.9	0	0	0	0	0.3	0	0.1	0.1	100.1
**										
98.1	0.9	0	0.5	0.3	0.3	0	0	0	0	100.1

** **Kangaroo Island**

Table. 7.10A. Composition of detrital sillimanite (viridian) grains in the Permian sands of southern Australia.

Al2O3	SiO2	FeO	MnO	MgO	TiO2	CaO	Na2O	K2O	Cr2O3	Total
61.9	37.8	1.3	0	0	0.4	0.3	0	0	0.2	101.8
61.9	37.7	1.3	0	0	0.4	0.3	0	0	0.2	100.8
61.9	37.7	1.3	0	0	0.4	0.3	0	0	0.1	100.8
61.7	37.2	1.4	0	0	0	0.3	0.2	0	0	100.8
61.7	37.2	1.4	0	0	0	0.3	0.2	0	0	100.8
61.7	37.2	1.4	0	0	0	0.3	0.2	0	0.1	99.8
61.7	36.5	1.3	0	0	0.3	0.2	0	0	0	100.1
61.6	36.7	1.4	0.6	0	0.3	0	0	0	0.1	100.6
61.6	35.9	1.3	0.1	0	0	0.1	0.2	0	0.3	99.7
61.5	36.7	1.3	0	0	0.2	0	0	0	0	99.7
61.4	37.1	1.3	0.1	0	0.1	0	0	0	0	99.9
61.4	36.4	1.1	0.2	0	0.3	0.1	0.1	0	0.2	100.1
61.3	38.2	1.4	0	0	0.1	0.1	0.2	0	0.2	101.4
61.3	38.1	1.4	0	0	0.2	0.1	0.2	0	0.2	101.4
61.3	37.5	1.5	0.1	0	0	0	0	0	0	100.7
61.3	36.4	1.2	0	0.1	0	0.2	0	0	0	99.1
61.2	37.9	1.1	0	0	0.3	0.4	0	0	0	100.9
61.2	37.4	1.2	0.3	0	0	0.3	0.1	0	0	100.6
61.2	37.4	1.2	0.3	0	0	0.2	0.1	0	0	100.5
61.2	37.4	1.2	0.3	0	0	0.3	0.1	0	0	100.6
61.2	37.4	1.2	0.3	0	0	0.3	0.1	0	0.2	101.4
61.2	37.1	1.5	0	0	0	0.1	0	0	0.1	99.9
61.1	37.6	1.2	0.2	0	0	0.1	0	0	0.4	100.6
61.1	37.1	1.4	0.3	0	0.3	0.1	0.1	0	0.1	100.3
61.1	36.9	1.2	0.2	0	0.1	0.2	0	0	0.5	100.3
61.1	36.9	1.2	0	0	0	0.1	0.2	0	0.4	99.9
61.1	36.9	1.2	0	0	0	0.1	0.2	0	0.4	99.9
61.1	36.9	1.2	0	0	0	0.1	0.2	0	0	99.3
61.1	36.9	1.1	0	0.3	0	0.2	0.1	0	0	100.1
61.1	36.8	1.6	0.3	0	0	0.1	0.2	0	0.2	100.6
61.1	36.7	0.1	0.2	0	0	0.1	0.1	0.1	0.7	99.4
61.1	36.1	1.2	0.2	0	0	0.1	0.3	0	0	99.3
60.9	37.4	1.7	0	0	0	0.1	0	0	0.5	100.7
60.7	37.2	1.4	0.1	0	0.3	0.1	0	0.1	0	99.9
60.7	36.5	1.2	0	0	0	0	0	0	0.5	99.1
60.5	36.6	1.1	0.1	0	0.1	0.1	0	0	0	98.5
60.5	36.1	1.2	0	0	0.2	0.2	0.1	0	0	98.8
60.4	37.1	1.4	0.1	0	0.1	0.1	0	0	0.1	99.3
60.4	37.1	1.4	0.1	0	0.1	0.1	0	0	0.2	99.4
60.4	37.1	1.4	0.1	0	0.1	0.1	0	0.1	0	99.9
60.2	36.7	1.1	0.4	0	0.3	0	0	0	0.6	100.3
60.2	36.3	1.1	0.3	0.1	0	0.3	0	0	0.1	98.6
60.1	36.5	1.5	0	0.1	0.3	0	0	0	0	98.6
59.9	37.5	1.5	0.1	0	0.3	0.2	0.2	0	0	99.8

Table. 7.10B. Composition of detrital sillimanite grains in the Permian sands of southern Australia.

SiO2	Al2O3	FeO	MnO	MgO	CaO	Na2O	K2O	TiO2	Cr2O3	Total
38.2	59.9	0.9	0.3	0.5	0.2	0.1	0	0.2	0.2	100.7
38.1	60.6	0.7	0.2	0.1	0.3	0	0	0	0.3	100.4
37.8	60.5	0.6	0.1	0	0.1	0	0	0.3	0	99.6
37.4	61.3	1.8	0	0	0.1	0	0	0.2	0.1	100.8
37.1	60.6	0.8	0.1	0	0.1	0	0	0.2	0.1	99.1
37.1	61.4	0.6	0.1	0.1	0.1	0.2	0	0.1	0.1	99.8
36.5	60.8	0.7	0.1	0	0.1	0.1	0	0	0.3	98.6
36.4	62.1	0.5	0	0.2	0.3	0.2	0	0	0.2	99.8
36.4	61.9	0.6	0	0.2	0.3	0.2	0	0	0.1	99.1
35.9	59.9	0.4	0	0	0.2	0	0	0.5	0.6	97.5
36.3	61.4	0	0.1	0	0.2	0	0	0	0.3	98.3 *
36.1	61.3	0.4	0	0	0.1	0	0	0.1	0.3	98.3
36.9	60.7	0.1	0	0	0.3	0	0	0.5	0.3	97.8
36.9	61.1	0	0.1	0	0	0	0	0.1	0.2	97.9
36.1	61.2	0.4	0.1	0	0	0	0	0.1	0.5	98.4
36.9	60.4	1.8	0	0	0.1	0	0	0.1	0	99.2
36.9	60.9	0.4	0	0	0.2	0	0	0.5	0.6	99.5
35.1	61.3	0.4	0	0	0.1	0	0	0.1	0.4	97.4
35.9	60.7	0.1	0	0	0.3	0	0	0.5	0.4	97.9
36.9	60.1	0	0.2	0	0	0	0	0.1	0.3	97.6
35.1	61.2	0.3	0.1	0	0	0	0	0.1	0.5	97.3
37.5	60.2	0.3	0	0	0	0.1	0	0.2	0.4	98.7 **

* Kangaroo Island

** Erratics gneisses

Table. 7.10C. composition of detrital sillimanite grains in the Permian sands of Kangaroo Island.

SiO2	36.3	36.1	36.9	36.9	36.1	36.9	36.9
TiO2	0	0.1	0.5	0.1	0.1	0.1	0.5
Al2O3	61.4	61.3	60.7	61.1	61.2	60.4	60.9
FeO	0	0.4	0.1	0	0.4	1.8	0.4
MnO	0.1	0	0	0.1	0.1	0	0
MgO	0	0	0	0	0	0	0
CaO	0.2	0.1	0.3	0	0	0.1	0.2
Na2O	0	0	0	0	0	0	0
K2O	0	0	0	0	0	0	0
Cr2O3	0.3	0.3	0.3	0.2	0.5	0	0.6
Total	98.3	98.3	97.8	97.9	98.4	99.2	99.5

Table. 7.11. Composition of detrital staurolite grains in the Permian sands of southern Australia.

Al2O3	SiO2	FeO	MgO	MnO	CaO	Na2O	K2O	TiO2	Cr2O3	Total
55.7	28.7	12.8	2.3	0.3	0	0.2	0	0.7	0.3	101
55.2	28.7	11.5	1.4	0.7	0	0.1	0	0.6	0.1	98.7
54.9	28.7	13.2	1.6	0	0.2	0	0	0.5	0.3	99.4
54.9	28.7	13.2	1.6	0	0.2	0	0	0.5	0.3	99.4
54.8	29.3	13.1	3.2	0.2	0.1	0.2	0	0.9	0.4	101.7
54.8	28.6	14.3	1.8	0.1	0	0	0	0.5	0.1	100.3
54.8	28.5	15.1	0.9	0	0.1	0.1	0	0.5	0.1	100.2
54.8	28.4	13.4	1.7	0.2	0.1	0.3	0	0.6	0.2	99.7
54.8	28.4	13.4	1.7	0.2	0.1	0.3	0	0.6	0	99.7
54.7	28.9	12.9	2.5	0.5	0.2	0.1	0	0.7	0.1	100.6
54.7	28.7	12.8	2.3	0.3	0	0.1	0	0.7	0.3	100.7
54.7	28.5	14.7	1.3	0.2	0.1	0	0	0.3	0.1	99.9
54.6	29.6	14.5	1.4	0.9	0.2	0	0	0.1	0.1	101.4
54.6	28.6	14.5	1.4	0.9	0.2	0	0	0.1	0.1	100.4
54.4	28.8	13.9	1.1	0.2	0	0	0	0.5	0.2	99.1
54.4	28.8	13.9	1.1	0.2	0	0	0	0.5	0.2	99.1
54.4	28.4	13.1	1.9	0.4	0	0	0	0.6	0.1	99.2
54.4	28.4	13.1	1.9	0.4	0	0	0	0.6	0.1	99.2
54.3	28.9	12.4	1.3	0.4	0	0	0	0.9	0.2	99.1
54.3	28.3	13.7	1.1	0.2	0	0	0	0.7	0	98.3
54.2	28.6	12.5	1.9	0.4	0.3	0	0	0.5	0	98.4
54.2	28.4	14.1	1.3	0.1	0.1	0	0	0.6	0.2	99.2
54.1	28.5	13.6	1.1	0	0.1	0.1	0	0.6	0.2	98.9
53.9	28.9	13.4	3.1	0	0.3	0.7	0	0.7	0.1	101.1
53.9	28.9	13.4	3.1	0	0.3	0.7	0	0.7	0.1	101.1
53.8	28.3	12.9	3.2	0.2	0.1	0.2	0	0.9	0.4	100.4
53.7	30.4	13.6	0.9	0.1	0.1	0	0	0.5	0.1	99.4
53.7	28.4	13.7	1.5	0.1	0	0.1	0	0.7	0.2	98.6
53.6	28.9	14.1	1.3	0.8	0	0.2	0	0.6	0.1	99.7
53.4	29.5	12.7	2.3	0	0	0.2	0	0.9	0.6	100.8
53.4	28.5	13.1	2.1	0.6	0.2	0.5	0	0.5	0	98.8
53.1	28.7	12.9	2.4	0.2	0.2	0	0	0.5	0.2	99.2
56.8	27.3	15.3	2.1	0.1	0.3	0	0	0.7	0.1	101.5
56.6	27.3	14.1	1.3	0	0	0	0	0.8	0.4	100.6
56.6	27.3	14.1	1.3	0	0	0	0	0.8	0.4	100.6
55.8	27.3	14.1	2.1	0.1	0.3	0	0	0.7	0.1	100.5
55.6	27.8	13.1	2.5	0.6	0	0.3	0	0.5	0.1	100.6
55.3	26.2	14.5	1.1	0.3	0.1	0	0	0.6	0.3	98.5
55.2	27.6	15.6	1.7	0.2	0.2	0.1	0.1	0.8	0	101.8
54.7	28.1	13.9	1.4	0.1	0.1	0.2	0	0.9	0.1	98.5
54.7	27.8	13.8	1.9	0.1	0.1	0	0	1.1	0.2	99.8
54.7	28.1	11.6	2.1	0.5	0.3	0.4	0	0.4	0.3	98.5
54.5	27.5	13.2	1.5	0.4	0	0.6	0	0.6	0.1	98.4
54.5	27.4	13.5	2.1	0.5	0.2	0	0	0.8	0	99.3
54.4	27.4	13.4	2.6	0.3	0.2	0.1	0	0.3	0.3	99.1
54.4	27.2	15.4	1.1	0.4	0.2	0	0	0.2	0	98.7
54.4	27.1	13.8	1.3	0.4	0.3	0.1	0	0.1	0.1	98.9
54.2	27.6	12.7	2.4	0.6	0	0	0	0.7	0.3	98.7
54.2	27.6	15.6	1.7	0.1	0.2	0.1	0.1	0.8	0	101.1
54.2	26.9	15.1	1.4	0.3	0	0.1	0	0.8	0.4	99.1

Table. 7.11. Continue...

Al2O3	SiO2	FeO	MgO	MnO	CaO	Na2O	K2O	TiO2	Cr2O3	Total
53.9	27.7	12.5	2.4	0.7	0	0	0	0.9	0	98.5
53.8	27.8	14.6	2.2	0	0.1	0	0	0.8	0	99.4
53.8	27.8	14.5	1.6	0.1	0	0	0.1	0.7	0.3	99.5
53.8	28.2	13.8	1.7	0.5	0	0	0	0.4	0.1	99.1
53.7	27.1	15.1	1.7	0.5	0.1	0	0	0.7	0	99.3
53.7	27.1	15.1	1.7	0.5	0.1	0	0	0.7	0	99.3
53.5	28.1	13.8	1.7	0.3	0.1	0	0	0.8	0.3	98.6
53.5	27.8	13.8	1.7	0.2	0.1	0.4	0	0.7	0.2	98.2
53.4	28.2	13.8	1.9	0.1	0.2	0.4	0	0.4	0.5	99.4
53.4	28.1	13.8	1.9	0.1	0.2	0.4	0	0.4	0.5	100.4
53.3	27.7	14.1	2.1	0.4	0.2	0	0.1	1.3	0.1	100.1
53.3	27.7	14.4	1.7	0.4	0.4	0.4	0	0.8	0.3	99.7
53.3	27.6	14.4	1.7	0.4	0.4	0.4	0	0.7	0.3	99.7
53.3	27.6	14.1	0.2	2.1	0	0	0	1.3	0.2	100.1
53.1	27.6	12.9	2.9	0	0	0.1	0.1	0.6	0.1	98.2

Table. 7.12. Composition of detrital epidote grains in the Permian sands of southern Australia.

SiO2	37.7	38.7	39.3	39.6	38.6	38.3	38.3	38.2	37.7	37.7	38.3	38.3	37.7	38.3	37.7
TiO2	0.6	0.4	0.4	0.4	0.3	0	0.4	0	0.8	0.4	0.4	0.4	0.4	0	0.6
Al2O3	20.9	19.9	21.7	31.3	27.5	27.3	28.1	29.2	19.7	19.9	21.7	21.1	19.8	29.3	20.9
FeO	5.5	5.8	3.7	3.6	9.5	8.1	3.7	5.3	2.3	5.8	3.7	3.6	5.7	5.4	5.5
MnO	9.5	0	4.9	0.4	0.6	0.5	4.9	0.1	0.7	0	4.9	4.9	0	0	9.6
MgO	0.1	0.7	0.1	0.2	0.5	0	0.1	0.2	0.4	0.8	0.1	0.1	0.7	0.1	0.1
CaO	24.7	35.8	31.7	23.5	32.2	23.1	31.7	24.8	34.8	35.8	30.7	30.7	34.8	24.8	24.7
Na2O	0	0.3	0.3	0.2	0.5	0	0.3	0.2	0	0.3	0.3	0.3	0.3	0	0
K2O	0	0.1	0	0	0	0	0	0	0	0	0	0	0	0	0
Cr2O3	0.4	0.4	0.5	0	0.1	0	0.5	0.2	0.2	0.4	0.4	0.5	0.4	0.2	0.4
Total	99.4	100.3	100.7	99.4	100.1	97.3	100.7	99.1	98.2	100.6	100.7	100.7	100.6	99.1	99.5

No of ions on the basis of 13 (O)

Si	3.2	3.1	3.2	3.1	3.1	3.1	3.1	3.1	3.2	3.2	3.1	3.2	3.1	3.2	3.2
Ti	0.03	0.02	0.02	0.02	0.02	0	0.02	0	0.04	0.02	0.02	0.02	0.02	0	0.03
Al	2.1	1.9	2.1	2.8	2.3	2.2	2.4	2.3	2.1	2.1	2.1	2.1	1.9	2.2	2.1
Fe	0.4	0.4	0.2	0.2	0.5	0.4	0.3	0.4	0.2	0.4	0.3	0.2	0.4	0.4	0.4
Mn	0.7	0	0.3	0.1	0.2	0.1	0.3	0.01	0.1	0	0.3	0.3	0	0	0.7
Mg	0.01	0.1	0.01	0.01	0.1	0	0.01	0.01	0.1	0.1	0.01	0.01	0.01	0.01	0.01
Ca	2.2	3.1	2.8	2.1	2.9	2.1	2.8	2.2	3.1	3.1	2.8	2.8	3.1	2.2	2.2

Table.7.14. Composition of apatite grains of Kanmantoo Group metasediments rock samples.

													*	**	**	**	***
CaO	54.7	54.8	54.2	54.8	54.7	54.9	54.7	54.6	54.7	54.9	54.4	54.4	54.9	54.2	54.6	54.9	53.2
P2O5	41.3	41.9	42.3	42.7	42.1	41.4	41.4	42.6	42.4	41.9	42.5	40.5	40.1	43.1	40.4	40.6	40.7
TiO2	0.3	0.2	0.2	0.3	0.6	0.5	0.5	0.2	0.4	0.1	0	0	0.2	0	0	0	0.7
FeO	0.1	0	0.2	0.3	0.4	0.2	0.6	0.3	0.3	0.4	0.3	0.3	0.8	0.4	0.4	0.3	0.6
MnO	0	0	0.4	0	0	0.3	0.2	0.4	0.3	0.3	0.3	0.4	0.4	0.9	1.1	1.5	0.9
MgO	0	0	0	0	0	0.4	0.2	0	0	0.3	0.3	0.3	0	0.1	0.2	0.1	0.6
Na2O	0.2	0	0	0	0	0.4	0.3	0.3	0	0.2	0	0	0.1	0.1	0	0	0
K2O	0	0.1	0	0.1	0	0	0.2	0.1	0.2	0	0	0	0.1	0.1	0.1	0	0.1
SiO2	0	0	0	0	0	0	0	0	0	0	0	0.1	0.6	0	0	0.1	0
Cr2O3	0	0.1	0	0	0	0	0	0	0	0	0	0	0.3	0	0	0	0.1
Al2O3	0.1	0.2	0.1	0.2	0.2	0.3	0.2	0.1	0	0	0.1	0.1	0	0.2	0	0	0
Total	97.3	97.1	97.4	99.2	98.6	98.4	98.4	98.6	98.2	98.2	97.9	96.2	97.5	99.1	96.8	97.5	96.9

Structure formula on the base of 26 (O)

Ca	10.4	10.5	10.3	10.5	10.4	10.3	10.4	10.3	10.5	10.3	10.4	10.4	10.5	10.2	10.3	10.5	10.1
P	6.2	6.2	6.1	6.1	6.1	6.2	6.2	6.1	6.1	6.2	6.1	6.1	6.1	6.2	6.1	6.1	6.2
Fe	0.01	0	0.02	0.04	0.04	0.02	0.05	0.03	0.04	0.06	0.05	0.05	0.03	0.05	0.05	0.05	0.05

*	**Permian sands.**
**	**Gneisses of East Antarctica.**
***	**Granitic gneisses of Permian erratics**

Table 7.15A. Composition of detrital amphibole and pyroxene grains in the Permian sands of southern Australia.

SiO_2	FeO	Al_2O_3	MgO	CaO	Na_2O	MnO	Cr_2O_3	K_2O	TiO_2	Total
55.6	12.7	13.5	0.5	16.4	0	0.4	0.3	0	0.5	100.1
53.6	2.1	8.7	11.1	17.3	4.5	0	0.3	0	0.2	98.2
53.5	0.4	28.2	0.2	10.8	5.2	0	0.3	0	0	98.7
53.4	2.1	8.6	11.6	18.1	4.1	0	0.7	0	0	99.4
51.8	26.6	16.7	1.9	2.4	0.4	0.5	0	0.2	0	100.1
51.1	5.6	22.9	0.2	12.5	5.7	0	0.4	0	0	98.3
49.4	34.1	0.8	13.5	0.7	0.6	1.7	0	0	0	100.2
48.9	33.5	0.6	13.4	0.9	0	1.6	0.1	0	0	99.1
46.9	14.7	11.9	12.3	11.1	0.9	0.8	0	0.4	0.7	99.9
45.6	14.1	11.2	12.5	11.9	1.5	0.2	0.2	0.5	0.6	98.3
43.1	19.7	11.1	10.4	11.2	1.7	0.4	0	1.8	0.2	99.6
41.3	6.3	39.1	8.1	1.3	2.2	0	0	0.3	0	100.4
40.1	20.1	22.1	9.3	6.7	0	0.9	0	0	0.2	99.4
40.1	26.2	22.2	9.3	1.6	0	0.9	0	0	0.2	100.4
38.1	30.8	22.8	6.1	1.8	0	1.1	0.2	0.1	0.1	100.3
37.6	27.4	22.4	9.1	1.6	0	1.3	0.3	0.4	0	100.1
37.5	33.7	21.1	1.9	3.9	0	1.2	0	0	0.3	100.2
37.4	24.9	23.6	8.9	2.3	0	0.3	0.3	0	0	100.4
37.3	33.9	21.9	5.1	0.8	0	1.1	0	0	0	100.2
36.9	32.7	21.7	3.4	2.2	0	2.5	0.3	0	0.4	100.7
36.6	8.2	21.9	2.6	7.4	0.2	24.1	0.4	0	0.2	100.2
36.6	25.8	21.2	1.2	13.4	0	0.5	0	0	0	98.3
36.5	33.8	21.8	4.4	2.4	0.3	1.1	0.4	0	0	100.5
35.9	23.9	22.3	4.9	10.5	0	1.1	0.4	0	0	99.3
35.1	35.6	20.8	4.3	1.1	0	2.1	0.1	0	0.4	100.1

Table. 7.16. Composition of detrital micas with unknown minerals in the Permian sands of southern Australia.

	*	***	***	***	***	**	**	**	**	**	**	**	**	**	**	**	**	**
SiO2	52.6	41.1	41.8	37.1	36.3	47.5	29.4	17.3	51.8	47.5	60.3	44.6	26.7	19.6	50.1	38.4	49.3	0
TiO2	0.2	1.4	2.1	1.8	2.9	49.6	64.8	78.5	42.4	49.6	38.3	51.1	73.4	75.2	38.2	57.1	0.4	34.6
Al2O3	8.3	39.1	10.8	14.4	14.2	0.4	0.9	0.1	1.6	0.4	0.5	1.78	0.2	0.2	6.9	1.1	39.7	16.2
FeO	2.8	4.9	19.3	27.8	27.4	0	2.6	0.3	0	0	0.1	0.3	0	1.2	0.4	0.1	10.4	48.1
MnO	0.2	0.7	0.2	0.3	0.5	0	0	0.3	0	0	0	0	0	0	0	0	0.1	2.8
MgO	11.3	9.5	9.2	6.6	5.8	0.2	0.6	0.2	0.3	0.2	0.1	0.3	0.1	0.4	0.3	0.2	0.9	0
CaO	16.9	1.2	11.4	0.3	0.2	0.5	0.2	0.3	0.6	0.5	0.5	0.6	0	0.2	0.4	0.4	0.2	0
Na2O	4.4	2.2	1.6	0	0	0.2	0.2	0.5	0.1	0.2	0.3	0	0	0	0.3	0.9	0.1	0.3
K2O	0	0	1.9	9.3	9.4	0.1	0	0	0.3	0.1	0	0.4	0	0.3	2.1	0.4	0	0
Cr2O3	0.4	0	0.1	0.1	0.4	0.1	0	0	0	0.1	0	0.2	0.1	0	0	0.2	0.2	0
Total	97.1	100.1	98.4	97.7	97.1	98.6	98.7	97.5	97.1	98.6	100.1	99.3	100.5	97.1	98.7	98.8	101.3	101.8

*	Composition of detrtial hedenbergite grains in Permian sands of southern Australia.
**	unknown mineral in the Permian sands of southern Australia.
***	Composition of detrital mica grains in the Permian sands of southern Australia.

Table. 7.1[illegible]A Composition of sphene grains in the rock samples of East Antarctica.

SiO2	TiO2	CaO	Al2O3	FeO	MgO	MnO	Na2O	K2O	Cr2O3	Total
*										
30.1	27.9	24.4	6.8	3.9	4.9	0.2	0	0	0.6	99.9
28.1	37.7	27.2	1.1	1.1	0	0.3	0	0.1	0.1	98.1
31.1	31.5	28.8	4.4	2.1	0.6	0	0.5	0	0.1	99.5
29.9	36.6	28.5	2.8	0.2	0	0.3	0.5	0	0	98.9
30.8	37.1	28.7	2.8	0	0.4	0.6	0.1	0	0	100.6
29.9	38.3	27.7	2.3	0.3	0.1	0.6	0.1	0		99.3
**										
28.5	35.2	26.3	3.9	2.3	0	0	0.6	0	0	96.8
27.9	40.5	27.1	0.7	0.7	0	0	0.7	0	0	97.6
29.9	35.3	27.8	3.5	0.9	0	0	0	0	0.3	97.7
24.1	40.1	23.1	1.6	7.6	0.1	2.4	0.8	0	0.3	101.1
28.7	39.4	28.1	0.9	0.2	0.5	0	0.2	0	0.3	98.8
28.8	35.3	28.8	3.5	0.9	0	0	0	0	0.3	98.2
27.9	41.5	27.1	0.7	0.7	0	0	0	0	0	98.5
28.5	36.1	26.3	3.9	2.3	0	0	0.6	0	0	98.2
30.8	31.4	27.8	5.9	1.9	0.3	0	0	0	0	98.4
***										
21.2	42.9	19.2	1.6	7.8	0	5.6	0	0	0	98.8
16.8	62.5	16.1	0.2	1.6	0.1	0.1	0.5	0	0	95.4
23.4	39.4	22.5	0.9	6.8	0	4.1	0	0	0.4	98.1
29.1	32.5	28.2	5.2	1.2	0	0	0.1	0.2	0.6	98.1
28.9	36.3	27.6	2.3	2.4	0	0	0	0	0.2	97.1
21.2	42.9	19.2	1.6	7.8	0	5.6	0	0	0	98.5
16.8	62.5	16.1	0.2	1.6	0.1	0.1	0	0	0	98.1
11.9	46.9	13.1	0.3	15.6	0	10.3	0	0	0	98.2
29.1	32.5	28.2	5.2	1.2	0	0	0	0	0	98.4
****										
29.4	39.9	27.7	0.2	2.4	0	0	0	0	0.2	99.8

*	Victoria Land
**	Windmill Island granitic gneis samples
***	Windmill Island gneisses samples
****	Windmill Island basic gneisses samples

Table. 7.16. Composition of sphene grains in the Permian volcanic erratics of southern Australia.

SiO2	TiO2	CaO	Al2O3	FeO	MgO	MnO	Na2O	K2O	Cr2O3	Total	
29.2	32.8	27.2	4.8	1.5	0	0.7	0	0	0	96.8	
29.1	31.9	29.1	5.1	1.4	0.2	0	0	0	0	97.1	
29.4	35.5	28.3	2.5	1.9	0	0.3	0.1	0	0	98.1	
29.8	33.7	28.4	3.1	1.9	0	0.6	0.1	0.1	0	97.6	
29.8	31.9	28.2	5.1	1.8	0.2	0.5	0.1	0.1	0	97.7	
30.1	33.1	27.7	4.1	1.3	0	0.4	0.5	0	0	97.3	
29.5	35.8	27.6	2.4	1.6	0.1	0.1	0.2	0	0	97.5	
29.5	34.6	28.8	3.9	0.7	0.6	0.5	0.1	0	0	98.7	
29.9	35.8	28.1	3.9	1.2	0	0	0.3	0	0.6	99.8	
31.1	32.7	28.1	4.9	0.2	0.8	0.1	0.5	0.5	0.4	99.3	
29.4	37.2	27.1	1.8	1.9	0	0.6	0.6	0	0	98.6	
29.7	37.7	27.1	2.1	1.5	0.1	0	0	0	0	98.2	
29.3	35.4	28.8	3.6	0.8	0	0	0.2	0	0	98.1	
28.6	37.7	28.6	2.5	1.1	0	0	0.1	0	0	98.6	
29.5	35.9	27.6	2.4	1.6	0.1	0.1	0	0	0	97.2	
29.5	34.6	28.8	3.9	0.7	0.6	0.5	0.1	0	0	98.7	
29.9	35.8	28.1	3.9	1.2	0	0	0.3	0	0.6	99.9	
30.4	35.4	27.7	2.1	0.8	0.2	0.3	0.2	0	0.3	97.5	*
29.8	31.7	28.1	5.4	0.6	0	0	0	0	0.4	96.1	

* Permian sands.

Table.7.1[illegible]Composition of detrital sphene grains in the Permian sands of southern Australia.

SiO2	TiO2	CaO	Al2O3	FeO	MgO	MnO	Na2O	K2O	Cr2O3	Total
30.4	35.4	27.7	2.1	0.8	0.2	0.3	0.2	0	0.3	97.5
29.8	31.7	28.1	5.4	0.6	0	0	0	0	0.4	96.1

Table. 7.18. Composition of detrital spinel grain in the Permian sands of southern Australia.

Al2O3	MgO	FeO	MnO	Na2O	Cr2O3	CaO	SiO2	K2O	TiO2	Total	
71.7	26.7	1.1	0.3	0.1	0.1	0	0	0	0	100.1	Gahnite

CHAPTER EIGHT

OPAQUE MINERALS AS PROVENANCE INDICATORS

INTRODUCTION

Detrital opaque heavy minerals, occurring as accessory minerals in most sands (Pettijohn et al. 1987), are commonly multiphase grains and may be thought of as rock fragments. Detrital ilmenite grains carry unique chemical fingerprints that can be used in provenance research. The purpose of this investigation is to test the characteristics of detrital ilmenite so as to identify those with the greatest potential for provenance interpretation. It has been shown that detrital opaque Fe-Ti oxide minerals (DOPQ) principally consist of magnetite, ilmenite and hematite, either as monomineralic grains or as polymineralic grains with exsolution and intergrowth textures. The varieties of composition and texture of DOPQ are useful as provenance indicators (Riezebos 1979; Basu and Hood 1985). Ilmenite does not appear in metabasites until mid greenschist or lower amphibolite facies (Peacock and Norris 1989). Ilmenite and magnetite are the major Fe-Ti oxides in metamorphic rocks, with rutile and hematite being somewhat less common (Frost and Lindsley 1991). Ilmenite is generally absent from high-Al basalts and basaltic andesites and is rare in andesitic lavas (Brophy 1984). In all rock types ilmenite tends to be absent in the lowest grade metamorphic rocks (Wise 1959; Trommsdorf and Evans 1980).

Ilmenite is relatively abundant in sands and sandstones and is relatively stable (Pettijohn 1941). Ilmenite appears to be the ideal mineral for the interpretation of provenance because ilmenite can occur in a wide variety of igneous rocks, both intrusive and extrusive, as well as pegmatite, and other vein rocks, and even some metamorphic rocks, especially gneisses (Ramdohr 1980). Ilmenite show compositional variations depending on source rock paragenesis (Hutton 1950; Buddington and Lindsley 1964). This mineral can be easily separated using magnetic techniques (Rosenblum 1958; Force 1976; Lumpkin and Zaikowski 1980; Darby 1984). Previous studies of detrital ilmenite (Darby 1984, Darby and Tsang 1987; Basu and Molinaroli 1989) and magnetite (Luepke 1980; Grigsby 1988, 1990, 1992) have suggested that variation in their composition is sufficient to provide an unmistakable signature to determine sediment source.

Opaque minerals have not been extensively used in heavy mineral studies for two reasons:

1-They are difficult to identify by mineral species in an ordinary grain- mount.

2-They are so ubiquitous that they are commonly believed to reveal little about source. But as Blatt (1967b) indicated, for provenance studies, opaque minerals have not received the attention they deserve.

Method (see in Appendix. 7)

ILMENITE

In this investigation the characteristics and chemical composition of detrital ilmenite grains in the Permian sands were compared with samples of erratics, Antarctica, Kanmantoo Group and Encounter Bay. This chapter summarises the comparative study of detrital opaque minerals, especially ilmenite, in igneous and metamorphic rocks of different localities. The average percentage of opaque minerals within the samples is about 8%, and as ilmenite is the dominant opaque, the high percentage of this mineral can be readily used for provenance interpretation. Ilmenite compositions are not necessarily simple and therefore, the chemical composition of detrital ilmenite were used to estimate quantitatively the provenance of Permian sands. Note however, that pure ilmenite is rare in nature; it commonly occurs in solid solution with hematite, and replacement of major cations (Fe, Ti) by Mg, Mn, Al and Cr is common. Furthermore pedogenic and diagenetic alteration may affect the original chemical composition of detrital grains (Basu and Molinaroli 1991).

Major elements

Bivariate plots of the abundance of Al2O3, Cr2O3, MgO and MnO versus TiO2 in the Permian sands and parent rock samples are shown in Figs. 8.0, 8.1, 8.2 and 8.3. The distribution of Cr2O3 show no distinction between source rocks; however, MnO, MgO and Al2O3 show some significant variations.

Plate. 8.1 Detrital opaque Fe-Ti oxide grains in the Permian sands of Cape Jervis.

A-Photomicrographs taken through an ore microscope showing heterogenous detrital ilmenite, light colour is hematite (hm) and dark colour is ilmenite (il).

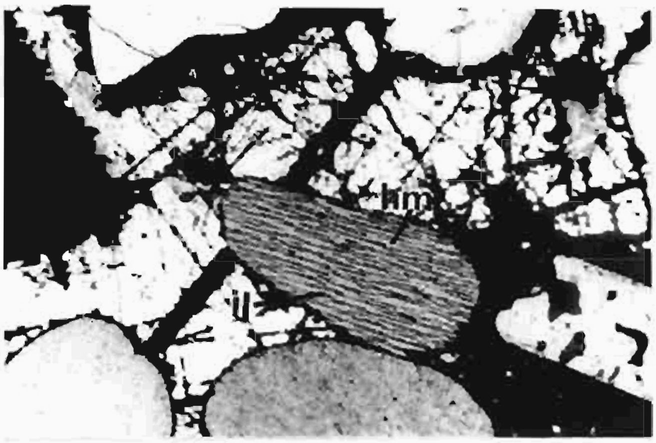
B- Backscattered electron micrograph of heterogenous detrital ilmenite.

C- Photomicrograph taken through an ore microscope of heterogenous ilmenite indicating ilmenite (il) and hematite (hm) have formed intergrowth.

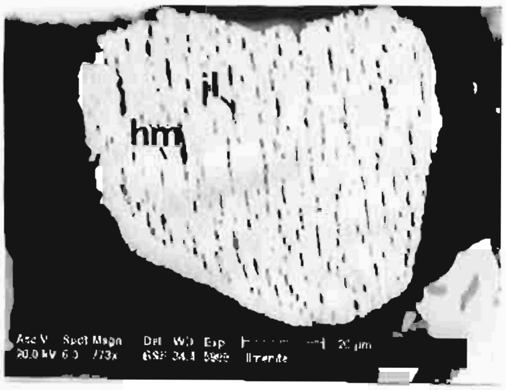
D- Scattered electron micrograph of heterogenous ilmenite (il) and hematite (hm).

E- Photomicrograph taken through an ore microscope of heterogenous (h) and homogenous (ho) of detrital ilmenite.

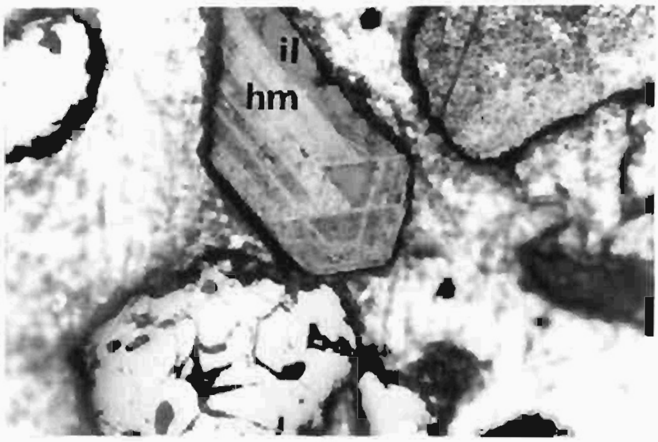
F- Backscattered electron micrograph of heterogenous (h) and homogenous (ho) detrital ilmenite. Scale. 100 x.


hm
il

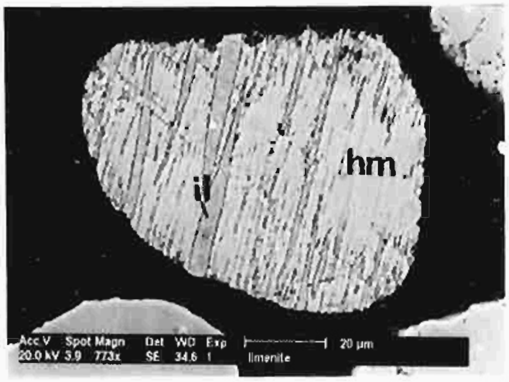
A


il
hm
Acc.V Spot Magn Det WD Exp 20 µm
20.0 kV 6.0 773x BSE 34.1
Ilmenite

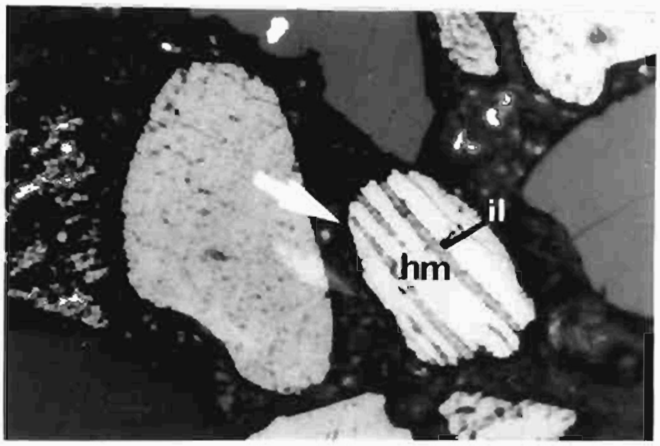
B


il
hm

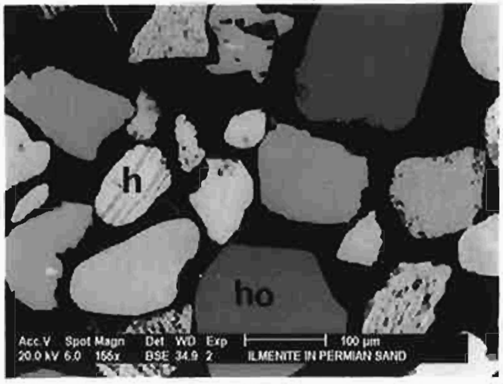
C


hm
il
Acc.V Spot Magn Det WD Exp 20 µm
20.0 kV 3.9 773x SE 34.6 1
Ilmenite

D


il
hm

E


h
ho
Acc.V Spot Magn Det WD Exp 100 µm
20.0 kV 6.0 155x BSE 34.9 2
ILMENITE IN PERMIAN SAND

F

Ilmenite from metamorphic source rocks are richer in TiO2 than those from igneous rocks (Basu and Molinaroli 1989), TiO2 content from metamorphic sources show a narrow range of content, and from igneous sources show a wider range (Basu and Molinaroli 1989; Grigsby 1992). The ilmenite composition in the Permian sands indicate a high Fe content may be due to inclusion of hematite, and low Cr values in ilmenite indicate non mafic rock sources (Pearre and Heyl 1960; Darby 1984). The mean and standard deviation of TiO2, FeO, MnO and MgO in Permian sands and parent rock samples is shown in Table.8.0. As the table shows, the mean percentage of TiO2 and MnO in Permian sands can be compared with the metamorphic rock samples. Detrital ilmenite grains from Permian sands show a narrow range of TiO2 content between 49 to 54%. Ilmenites of greater purity occur in metamorphic rather than in igneous source rocks, the latter showing a wider range of TiO2 content (30 to 48%) (Figs. 8.1 to 8.3). The chemical composition of detrital ilmenite grains in the Permian sands is shown in Table. 8.1. Chemical composition of detrital ilmenite grains from metamorphic parent rocks indicate that grains with TiO2 content greater than 50% are most common, occurring in 87% of all cases. For comparisons the detrital ilmenite grains from Permian sands show grains with TiO2 content greater than 50% occurring in 91% of all cases, therefore, TiO2 content of detrital ilmenite in the Permian sands indicate mostly metamorphic sources rather than igneous sources. [Adelaidean sandstones do carry some ilmenite probably derived from older granite gneisses, but sandstones rich in ilmenites form only a tiny percentage of outcrops available to Permian glaciers.]

TiO_2	FeO	MnO	MgO	Samples
53.1 (2.4)	43.5 (3.6)	2.1 (1.9)	0.3 (0.3)	**Permian sands**
51.1 (0.4)	42.3 (2.2)	3.7 (2.3)	0.1 (0.1)	**East Antarctic (granitic gneisses)**
49.1 (1.4)	42 (1.1)	6.8 (0.8)	0.2 (0.1)	**East Antarctic (granite)**
51.1 (0.8)	42.3(1.8)	0.9 (0.3)	0.2 (0.2)	**East Antarctic (basic gneisses)**
51.6 (1.4)	43. (2.1)	3.1 (1.8)	0.2 (0.1)	**Permian erratics (granitic gneisses)**
49.8 (1.6)	44.6 (3.5)	3.7 (2.7)	0.3 (0.3)	**Permian erratics (granite)**
52.1 (1.5)	44.1 (2.1)	3.3 (0.6)	0.3 (0.3)	**Permian erratics (volcanic)**
51.1 (1.)	43. (1.7)	4 (0.8)	0.2 (0.2)	**Kanmantoo Group (metasediments)**
50.7 (2.3)	42.3 (2.9)	3.7 (1.3)	0.2 (0.2)	**Encounter Bay (granites)**

Table. 8.0. Mean and standard deviation of different oxides from detrital ilmenite grains in the Permian sands of southern Australia and parent rock samples

Minor elements

Common minor elements in detrital ilmenite are Mn, Mg, Cr, Ti, Si, V and Al. Bivariate plots of the abundance of MgO, MnO, Al_2O_3 and Cr_2O_3 versus TiO_2 in ilmenites from Permian sands of southern Australia and source rock samples are shown in Figs. 8.0, to 8.3. Basu et al. (1989) showed that the TiO_2 content of detrital ilmenite in Holocene sands from igneous sources 48.3% is slightly less than those from metamorphic sources 50.8%. MgO concentrations above 0.5wt%, Al_2O_3 concentrations above 0.4wt% and MnO concentrations above 6.5wt% are found only in ilmenite derived from igneous source rocks. Cr_2O_3 concentrations show no distinction between these source rocks. Minor element contents of ilmenite from andesites with MgO between 1 and 4wt% and MnO from 0.5 to 1wt% (Kyle 1981).

The most magnesian ilmenite was found in metamorphic samples of erratics, Kanmantoo Group and Antarctica. Minor element concentrations of MgO above 0.4wt%, Al_2O_3 concentration above 0.4wt% and MnO concentration above 4.5wt% are found only in the ilmenites from igneous rocks in different localities as shown in Tables in

Chapter. 4 and 5 . However ilmenites from granite samples from East Antarctica generally contain less than 0.4wt% MgO (Table 4.17A). MgO concentrations of less than 0.4wt% with MnO less than 4.5wt% are found in metamorphic rocks.

In the Permian sands, concentrations of MnO less than 4.5wt% is 92%, and MgO less than 0.4wt% are 73.5%, and concentrations of Al2O3 less than 0.4wt% is 83.7% (see Tables. 8.2, 8.3, 8.4 and 8.5). The abundance of V2O3 and Cr2O3 do not show any distinction between metamorphic and igneous provenance.

1 -4%	4 -8%	8 -12%	Samples
81.8	18.2	0	Metamorphic samples of East Antarctica
64.8	35.2	0	Kanmantoo Group metasediments
73.3	26.7	0	Metamorphic samples of Permian erratics
14.3	85.7	0	Igneous samples of Permian erratics
40.9	54.5	4.5	Igneous samples of East Antarctica
79.6	18.4	2.1	Permian sands of southern Australia
96.1	3.9	0	Volcanic samples of Permian erratics

Table. 8.2. Weight percentage of MnO in the ilmenite grains from Permian sands of southern Australia and parent source rocks.

0.1-0.4%	0.4-0.7%	Samples
88.3	11.7	Metamorphic samples of East Antarctica
87	13	Kanmantoo Group metasediments
86.6	13.4	Metamorphic samples of Permian erratics
66.1	35.9	Igneous samples of Permian erratics
68.2	31.8	Igneous samples of East Antarctica
73.5	26.5	Permian sands of southern Australia
73.1	26.9	Volcanic samples of Permian erratics

Table. 8.3. Weight percentage of MgO in the ilmenite grains from Permian sands of southern Australia and source rock samples.

0.1-0.4%	0.4-0.7%	Samples
92.2	7.8	Metamorphic samples of East Antarctica
98.1	1.9	Kanmantoo Group metasediments
94.6	5.4	Metamorphic samples of Permian erratics
92.8	7.2	Igneous samples of Permian erratics
91	9	Igneous samples of East Antarctica
83.7	16.3	Permian sands of southern Australia
88.6	11.4	Volcanic samples of Permian erratics

Table. 8.4. Weight percentage of Al_2O_3 in the ilmenite grains from Permian sands of southern Australia and source rock samples

44 -48%	48 - 52%	52 -56%	56 - 60%	Samples
0	94.4	5.6	0	Kanmantoo Group
0	96.2	3.9	0	Metamorphic rock samples of East Antarctica
0	94.8	5.1	0	Permian metamorphic erratics
4.5	63.6	32	0	Igneous samples of East Antarctica
8.9	80.3	10.7	0	Permian igneous erratics
7.3	45.6	47	0	Permian sands
0	76.9	23.1	0	Permian volcanic erratics

Table. 8.5. Weight percentage of TiO_2 in the ilmenite grains from Permian sands of southern Australia and source rock samples.

The abundance of MnO in ilmenite tends to decrease markedly with increasing metamorphic grade in both metabasites and metapelites (Cassidy and Groves 1988).

From the limited published analyses of ilmenite composition from known rock types, significant differences exit in at least Mn and Mg values, and possibly in Ti and Fe. For example Best (1982) found Mn/Mg ratio of 0.23 in ilmenite from a gabbro, and Herz et al. (1970) found ratio 5-20 in heavy mineral assemblages dominated (75%) by ilmenite from charnockites and granulites bordering the anorthosite massif of Roseland District in Virginia. The detrital ilmenite grains of Permian sands and ilmenite grains from East Antarctica, Kanmantoo Group and Permian erratic rock samples have shown different ratio and indicate metamorphic and igneous rocks have different ilmenite composition. The ratio of Mn/Mg in the low grade metamorphic samples of Kanmantoo Group is 26, and in the medium to high grade metamorphic samples from Antarctica and Permian erratic ranges from 24 to 12, and in the igneous samples in Antarctica and Permian erratic ranges from 34 to 59. The ratio of Mn/Mg in the Permian sands is 14. Comparing the ratio of Mn/Mg in ilmenite grains of Permian sands and different rock types suggests that the metamorphic rock are most similar in ilmenite composition rather than igneous rocks.

In the Permian sands MnO content of ilmenite ranges from less than 0.2 wt% to more than 10.5wt% and MgO content ranges from 0 to 2.2wt%. Variation of each minor elements content from different samples is shown in Fig.8.4. Chemical composition of detrital magnetite in the Permian sands is shown in Table. 8.6.

CONCLUSIONS

The results of these investigations have confirmed that the ilmenite grains from igneous and metamorphic parent rocks are chemically distinct.

The relevance of an eight element chemical analysis in detrital ilmenite to provenance interpretations has shown that TiO2 content of detrital ilmenite in the Permian sands indicates that 88% of the grains have more than 50% TiO2 and have a narrow range. MgO concentrations show that about 73.5% have less than 0.4wt%. MnO concentrations show about 92% have less than 5wt%, and Al2O3 concentrations show that about 83.7% have less than 0.4wt%. The results of such determinations have shown that using the composition of detrital ilmenite in the Permian sands is useful tool for the source of the sands. The concentration of ilmenite with similar composition to the metamorphic samples in the Permian sands is 76.7%. Therefore the results are in agreement that more than 76% of the detrital ilmenite in the Permian sands have been derived from metamorphic rocks. This feature is also independently indicated by the presence of rutile, sillimanite, staurolite, garnet and epidote in the sands.

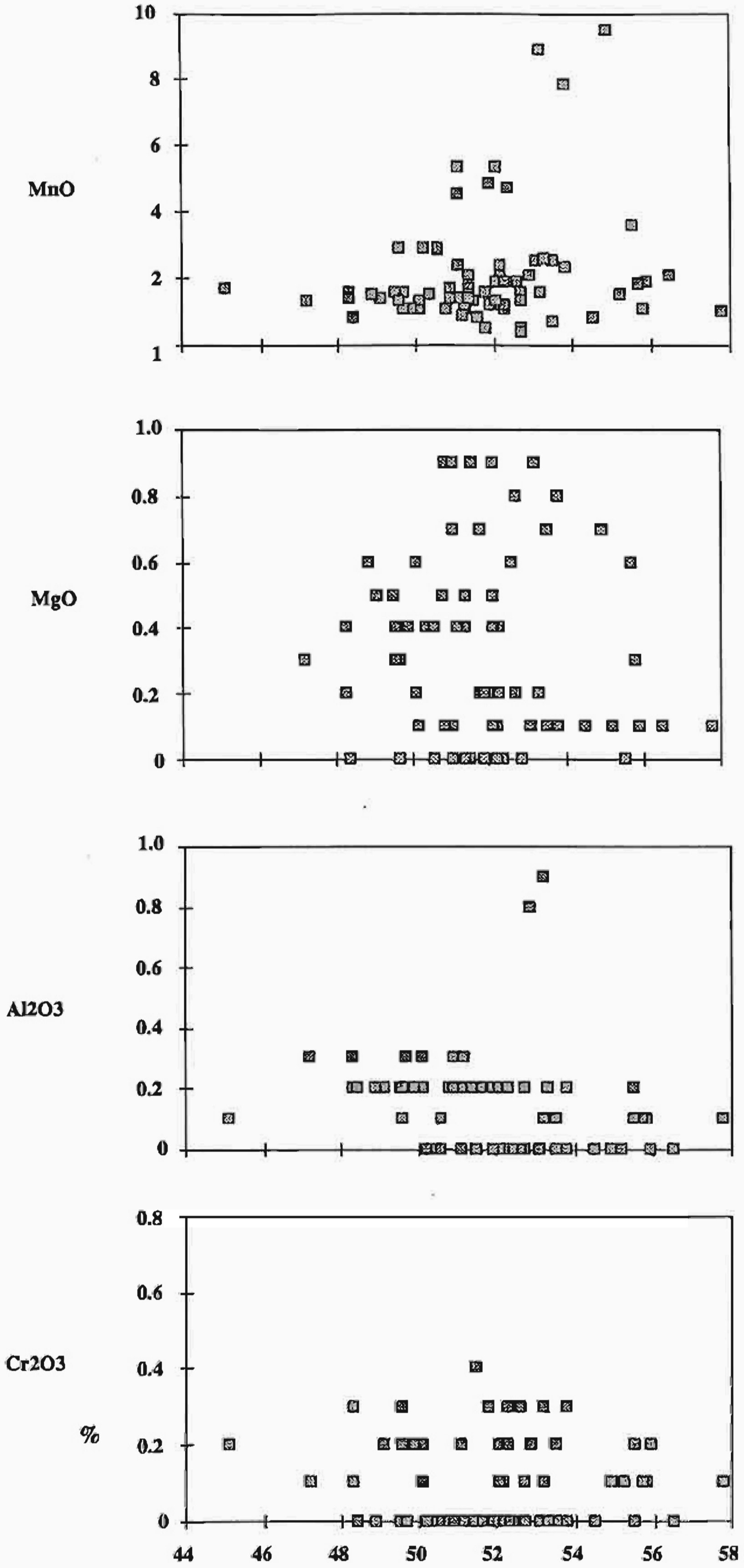


Fig. 8.0. Scatter plots of abundance of MnO. MgO, Al2O3 and Cr2O3 versus TiO2 in the detrital ilmenite grains from Permian sands of southern Australia.

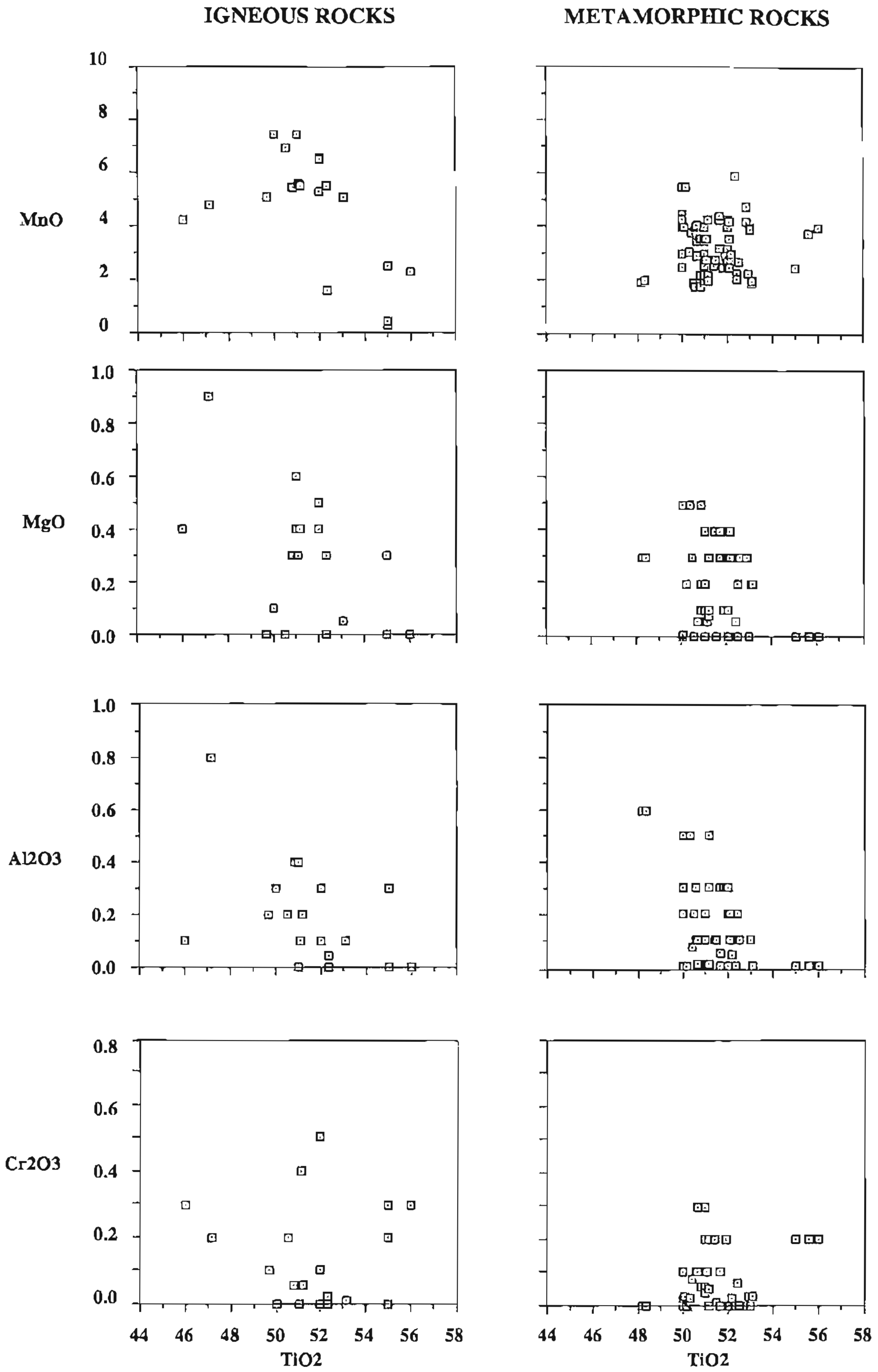


Fig. 8.1. Scatter plots of abundance of MnO, MgO, Al2O3, Cr2O3 versus TiO2 in the ilmenite grains of East Antarctica rock samples.

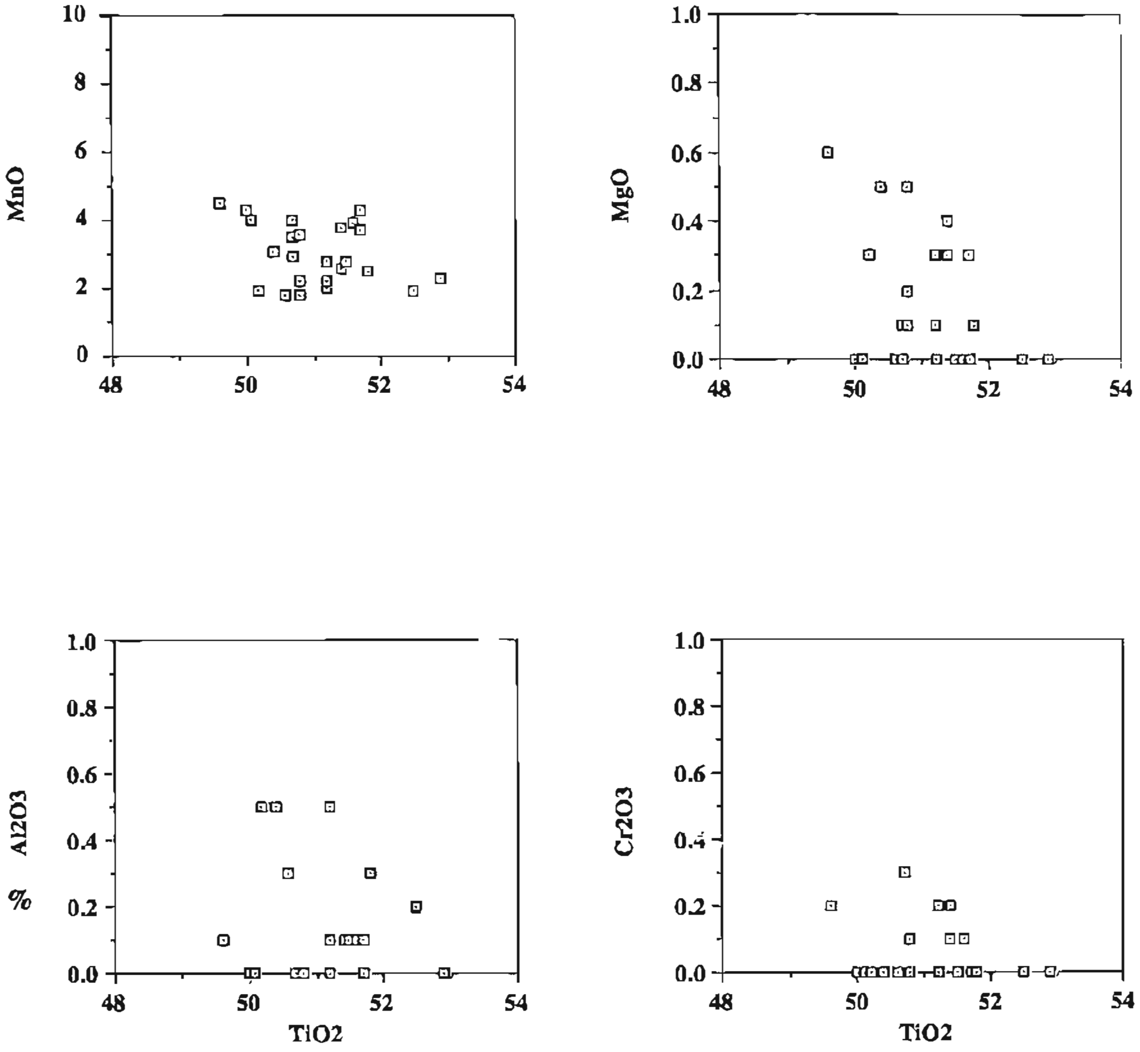


Fig. 8.1a. Scotter plots of an abundance of MnO, MgO, Al2O3 and Cr2O3 versus TiO2 in the ilmenite grains from basic gneisse rock samples of East Antarctica.

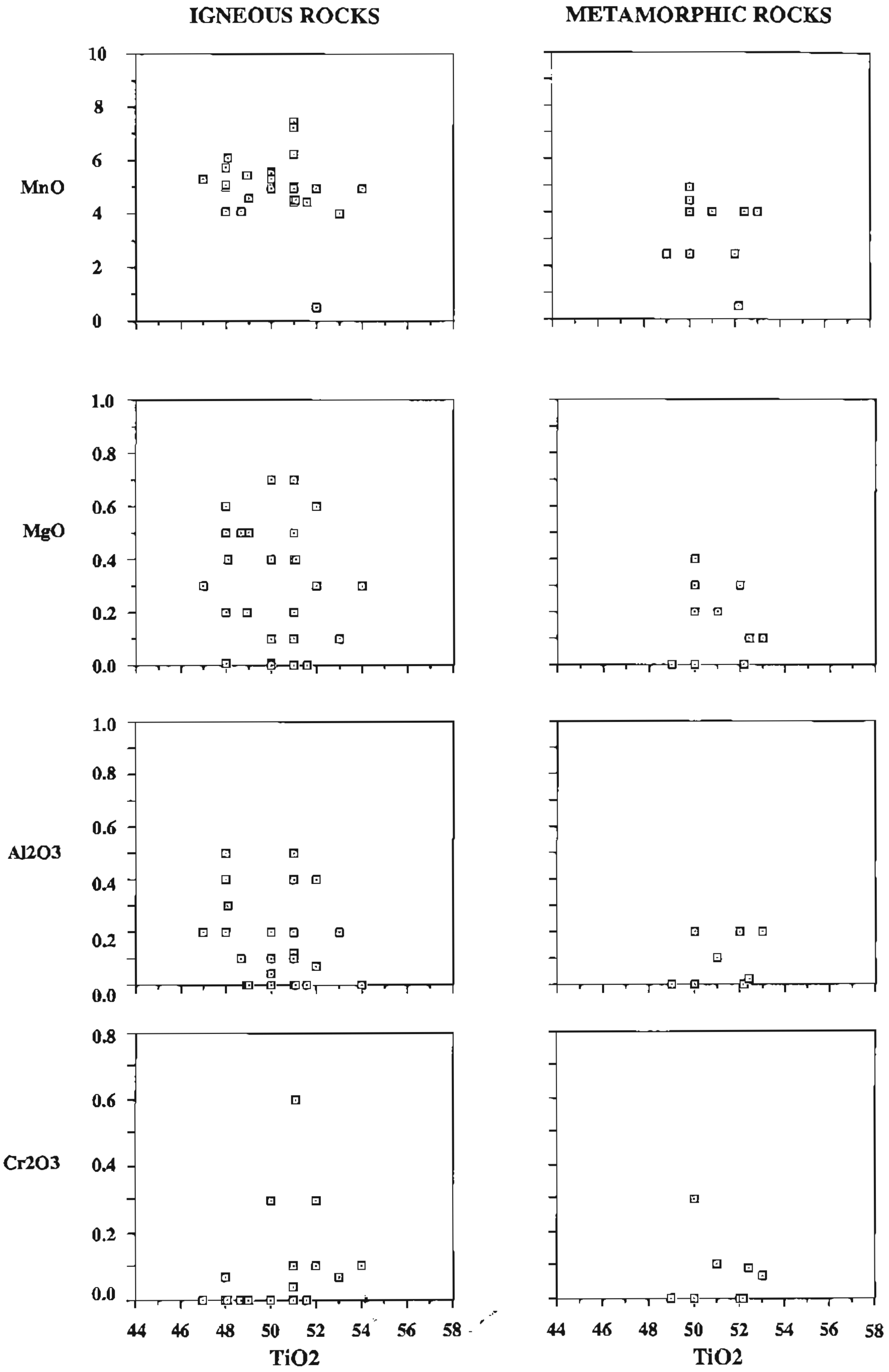


Fig. 8.2. Scatter plots of abundance of MnO, MgO, Al2O3, Cr2O3 versus TiO2 in the ilmenite grains from Permian erratics of southern Australia.

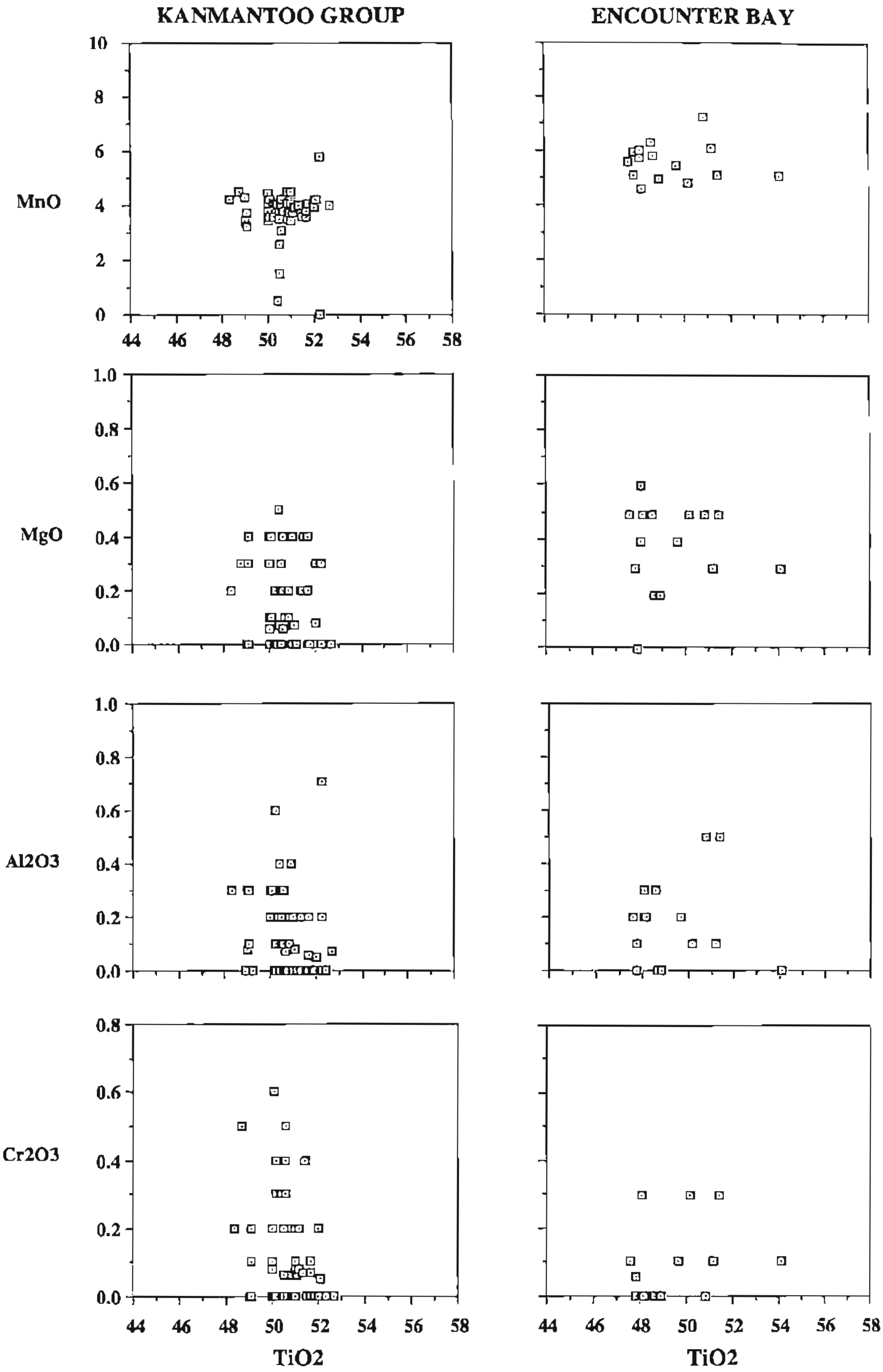


Fig. 8.3. Scatter plots of abundance of MnO, MgO, Al2O3, Cr2O3 versus TiO2 in ilmenite grains from Kanmantoo Group metasediment and Encounter Bay Granite of southern Australia.

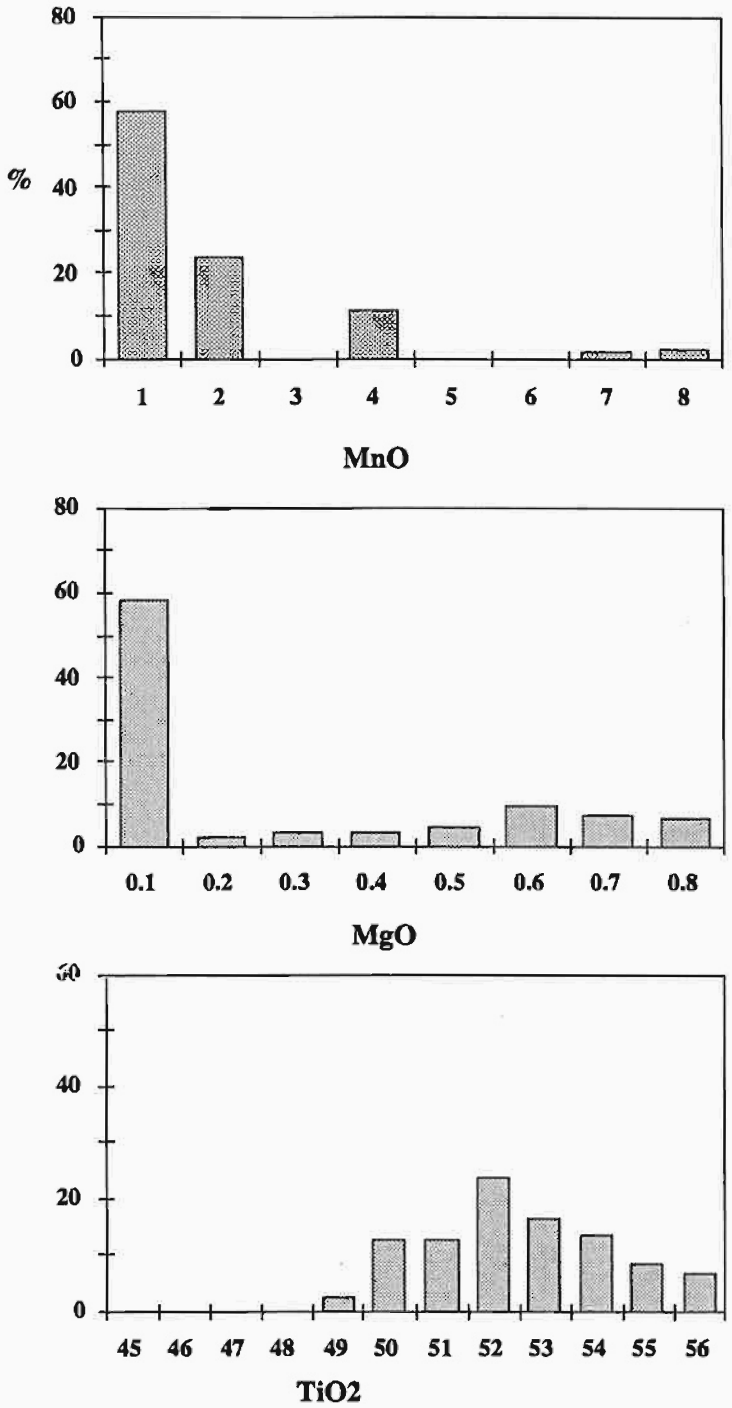


Fig. 8.4. Concentration of MnO, MgO and TiO2 in the detrital ilmenite grains in the Permian sands of soutrhern Australia.

Table. 8.1. composition of detrital ilmenite grains in the Permian sands of southern Australia.

TiO2	FeO	MnO	MgO	Al2O3	V2O3	Cr2O3	SiO2	Total	
56.5	41.7	2.1	0.1	0	0.3	0	0	100.7	
55.9	39.9	1.9	0.1	0	0.3	0.2	0	98.5	
55.8	41.9	1.1	0.3	0.1	0.3	0.1	0	99.7	
55.7	40.3	1.8	0.6	0.1	0.3	0.1	0.2	99.1	
55.2	42.6	1.5	0.1	0	0.2	0.1	0.1	100.5	
54.9	40.3	3.5	0.7	0	0.4	0.1	0	98.9	
54.5	44.1	0.8	0.1	0	0.2	0	0	99.8	
53.8	43.2	2.3	0.1	0	0.2	0	0	99.4	
53.5	44.8	0.7	0.1	0	0.2	0	0	99.5	
53.2	45.8	1.6	0.9	0.1	0.1	0.1	0	101.4	
53.1	44.1	2.5	0.1	0	0.2	0	0	100.1	
52.9	43.9	2.1	0	0.8	0.2	0.2	0.2	100.1	
52.7	44.9	0.5	0.8	0	0.5	0.1	0	99.6	
52.7	44.4	0.4	0.8	0	0.5	0.1	0	99.1	
52.7	44.9	1.6	0.2	0	0.4	0	0	99.9	
52.6	44.9	1.9	0.6	0	0.1	0.3	0	100.7	
52.4	43.1	4.7	0	0	0.2	0	0	100.4	
52.2	44.6	2.4	0	0	0.3	0	0	99.6	
52.2	43.4	2.1	0.1	0	0.2	0.1	0	98.2	
52.1	45.8	1.9	0.1	0	0.2	0	0	100.2	
51.9	42.6	4.8	0	0	0.2	0	0	99.6	
51.8	46.6	0.5	0.7	0.2	0.2	0.3	0	101.2	
51.5	44.1	1.3	0	0	0.3	0.4	0	98.3	
51.1	43.1	4.5	0	0	0.4	0	0	99.1	
51.1	44.3	1.4	0.7	0.3	0.2	0	0.3	100.8	
50.9	45.8	1.7	0.7	0.3	0.2	0	0	100.4	
50.6	48.1	2.8	0	0.1	0.3	0	0	101.9	
50.4	48.5	1.5	0.4	0	0.1	0	0	99.9	
50.2	45.5	2.9	0.1	0	0.3	0	0	99.1	
50.1	47.1	1.1	0.6	0.3	0.2	0.2	0	100.1	
49.7	47.4	1.6	0	0.3	0.2	0	0	99.4	
41.4	55.1	0.4	0.6	0.5	0.4	0.1	0	98.1	
45.1	47.6	1.7	3.6	0.1	0.2	0.2	0.1	99.8	*
53.2	31.6	0.7	0.7	0.6	0.4	0.3	0.5	98.4	
55.5	38.9	3.6	0	0.2	0	0.2	0	98.6	
53.5	39.2	2.5	1.4	0.1	0.4	0.2	0	98.2	
52.1	40.3	5.3	0.9	0.2	0.3	0.2	0.1	99.3	
49.6	44.1	2.9	0.4	0.1	0.3	0.3	0	97.8	

* Kangaroo Island

Table. 8.6. Composition of detrital magnetite grains in the Permian sands of southern Australia.

Fe2O3	FeO	Al2O3	V2O3	Cr2O3	MnO	MgO	TiO2	SiO2	Total
92.7	8	0.1	0.2	0	0.2	0	0.1	0	101.9
93.2	6.4	0.2	0.2	0.4	0.1	0	0.1	0	101.3
91.1	9.1	0	0.3	0.1	0.1	0	0	0	100.2
96.4	1.7	0.1	0.2	0.1	0	0	0.1	0	101.8
83.1	12.6	0.9	0.1	0.1	0.1	0.2	0.1	0	97.8
91.8	7.3	0.1	0.2	0.1	0.1	0	0.1	0	101.2
83.4	12.8	0	0.1	0.1	0.1	0.6	0.2	0	97.3
86.1	10.7	0.8	0.1	0.1	0.1	0.3	0.1	0	97.7
81.8	9.1	0.2	0.2	0.1	0.1	0	6.7	0	96.7
85.5	9.5	0	0.1	0.1	0.1	0	3.6	0	98.8
81.3	9.1	0.1	0.3	0.2	0.1	0	7.9	0	98.9
83.8	9.8	0.2	0.1	0.1	0	0.1	4.7	0	98.7

CHAPTER NINE

CONCLUSIONS

The Permian sediments in southern Australia were deposited in basins of glacial and fluvioglacial to shallow water sediment accumulation. Evidence from the striae indicate that the major sources of the detrital grains supplied to the Permian glaciogenic sediments lay to the south to southeast of the basin. Sediments deposited over southern Australia during the Permian were probably derived from the present continental shelf of Australia and formerly adjacent East Antarctica. An uplifted ice dome in eastern Antarctica at the present site of the Wilkes Basin may have provided a source for glacial sediments of southern Australia. The predominant sediments are tillite and fluvioglacial sands, therefore they are mineralogically and texturally immature. All contain far-travelled megaclasts that have been dropped from the continental ice.

S.E.M study of sand-size quartz and garnets revealed that the quartz grains are round to subangular, and the garnets are mostly rounded. The angular to subangular quartz grains show crushing and abrasion on grain surfaces that resulted from ice movements. The rounded to subrounded quartz grains do not have any crushing and indicate that they were deposited in glaciofluvial and aeolian environments. Chattermark trails on the surface of garnet grains and rounding of some of the garnet grains indicate that the Permian sands of southern Australia have been transported for a long distance from the source areas by the ice.

Some of the erratics, to date found in the Permian glaciogenic sediments are mostly of rock types known in Encounter Bay and Kanmantoo Group areas, but some of them have not been derived from these localities. Therefore they have been studied and compared with the rock types in eastern Antarctica and the results suggested possible Antarctica sources (see Chapter 4).

Bourman and Alley (1991) presented the paleoflow data that indicate the Permian glaciogenic sediments of southern Australia were derived from south and southeast source areas. In that direction the Encounter Bay Granite, Kanmantoo Group metasediments occupy the area. As the general ice flow was from this direction so, all the

Permian glaciogenic sediments should contain heavy minerals from those localities and beyond. The results of this investigation for the provenance of the Permian glaciogenic sediments, with particular emphasis on both light and heavy minerals as a provenance indicator, lead to the following conclusions:

Quartz is the most abundant light mineral in the Permian sands and in almost all common parent rocks. Therefore, quartz has the greatest potential for reading history of provenance for Permian sands.

Polycrystallinity of quartz grains has been used for interpretation of the source area. Polycrystalline quartz grains in the Permian sands have shown about 71% of them having less than five crystal units and about 29% have more than five crystal units. Compared with the plutonic and metamorphic rock samples, a mostly igneous source for quartz grains is confirmed. The size distribution within the polycrystalline quartz grains in the Permian sands ranges from unimodal to bimodal and it was shown that about 87% of the polycrystalline quartz grains in the Permian sands have unimodal size distribution. The elongation coefficient and the size distribution of polycrystalline quartz grains in the Permian sands when compared with the plutonic and metamorphic rock samples, confirmed that plutonic rocks were the major source for quartz grains.

Trace elements, such as Fe, Mn, Mg, Ca, Li and Al, in quartz grains from the Permian sands (Chapter 6) show such good agreement as to suggest that a mixture of plutonic and high grade metamorphic rocks was the chief source of detrital quartz grains.

Catholuminescence studies of sand size quartz were also used for the Permian sands and showed that they are blue and brown in colour. Since the blue CL quartz grains usually derive from plutonic and the brown CL quartz derive from high grade metamorphic rocks. The ratio of blue to brown CL of quartz grains of Permian sands is about 1.1, which means the quartz grains were probably derived from both these sources. Determining the chemical composition of detrital feldspar using microprobe analysis of the Permian sands has also been used in the study of provenance. Feldspar from Encounter Bay rock sample is rather potassic, ranging in composition from An2Ab1Or97 to An37Ab62Or1. Feldspar in granite and gneisses of Permian erratics are potassic but range in composition from An2Ab0Or98 to An1Ab99Or0 in granite and from

An1Ab0Or99 to An3Ab97Or0 in gneisses rock samples. Feldspar in Permian volcanic erratics is mostly calcic in composition ranging from An38Ab62Or0 to An3Ab93Or4. Feldspar in granite rock samples of East Antarctica is mostly potassic, ranging in composition from An4Ab0Or96 to An5Ab95Or0 and in the gneisses rock samples is also potassic, ranging in composition from An2Ab0Or98 to An2Ab98Or0 . Detrital feldspar grains of Permian sands are potassic in composition ranging from An0Ab8Or92 to An3Ab0Or97. Therefore, this suggests that the detrital feldspar grains in the Permian sands may have been derived from the same source as the granitic or gneissic rock samples (Chapter 6).

Heavy minerals are the most sensitive indicators of the nature of sediments source areas, and have frequently been used in studies of provenance. Amongst the heavy minerals, Permian sands contain garnet, tourmaline, zircon, rutile, staurolite, sillimanite, ilmenite and epidotes, with a trace of apatite, sphene, amphibole and biotite.

Permian sands of southern Australia are rich in garnets. This enrichment in garnets is due to mechanical stability of this mineral during periods of transport and deposition, compared to the abrasion of unstable minerals such as mica, pyroxene and amphibole during transportation. Garnets in the Permian sands show different colours ranges, from colourless to pink and deep red colours and suggests different source rocks.

The microprobe analysis of detrital heavy mineral suites within the Permian sands were also used for the provenance studies, because it is possible to determine accurately and quickly the compositional variation of each minerals, particularly of opaque minerals.

Variations in garnet geochemistry were related to the source rocks and revealed these points: Garnets that are rich in Fe,Mg were recorded from gneisses rock samples of East Antarctica and granitic gneisses of Permian erratics. Garnets that are rich in Fe and Mn have been recorded from phyllites and granitic rock samples of East Antarctica. Based on these analyses, it is concluded that most of the garnets in these sands were derived from medium to high grade metamorphic rocks that have shown similar compositions to gneisses and granitic gneisses rock samples of East Antarctica and to the Permian erratics. They also have distinct composition from the garnet grains of Kanmantoo Group metasediments and phyllite rock samples of East Antarctica. Most of the detrital garnet grains collected from different localities in the Permian sands were derived from similar source rocks.

Zircon is among the most stable minerals commonly found in the clastic rocks. The Hf contents of detrital zircon grains in the Permian sands have been used for provenance study. Mean percentage of Hf is 1.4 in Permian sands and 1.5-1.6 in granitic gneisses and 1.5 in Kanmantoo Group rocks. The percentage of Hf in the Permian sands and Kanmantoo Group has shown a bimodal distribution and therefore the detrital zircon grains of Permian sands may have been derived from both Kanmantoo Group metasediments and granitic gneisses (Chapter 7).

The optical properties such as colour of detrital tourmaline grains of Permian sands suggests mostly granitic source rocks. Chemical composition of detrital epidote grains in the Permian sands have shown a wide compositional range. The chemical composition of epidote grains (Chapter 7) indicate that the epidote grains in these sands were largely derived from metamorphic rock sources, such as Kanmantoo Group and granitic gneisses of East Antarctica.

Investigations of chemical composition of detrital staurolite grains in the Permian sands have shown a distribution of Fe, Al and Mg elements similar to the staurolite grains of Kanmantoo Group. This suggests that the source of detrital staurolite in these sands should be from Kanmantoo Group rocks.

Detrital ilmenite grains carry unique chemical fingerprints that have been used in provenance research. Ilmenite appears to be the ideal mineral for the interpretation of provenance because ilmenite can occur in a wide variety of igneous rocks, both intrusive and extrusive, as well as pegmatite, and other vein rocks, and even some metamorphic rocks, especially gneisses (Chapter 8). Ilmenite show compositional variations depending on source rock paragenesis. Ilmenite in metamorphic source rocks are richer in TiO2 than those from igneous rocks. MgO less than 0.4wt% and MnO less than 5wt% have found from metamorphic rocks of East Antarctica, Kanmantoo Group and Permian erratics. MgO more than 0.4wt% and MnO more than 5wt% are found from igneous rocks such as Encounter Bay Granites, granites of East Antarctica and Permian erratics. Ilmenite grains from metamorphic and igneous source rocks are chemically distinct and can be used for provenance indicators This thesis suggests that more than 95% of all detrital ilmenites within the Permian sands have been derived from metamorphic source rocks (Chapter 8).

The chemistry of detrital sillimanite grains in the Permian sands are mostly viridian in composition and are similar to the composition of sillimanite grains from gneisses rock samples of East Antarctica. Therefore the source rocks for the detrital sillimanite grains of Permian sands should be from similar high grade metamorphic rocks.

Rutile in the Permian sands appears mostly in foxy red colours. Rutile has not been recorded from granitic, volcanic rock samples of East Antarctica and Permian erratics, but is common in gneissic rock samples of East Antarctica and from similar erratics. The presence of detrital rutile in the Permian sands suggests that they may have been derived from high grade metamorphic rock such as gneisses. The Permian sands contain very small amount of pyroxene and sphene. This indicates that other sources, such as volcanic rocks may have contributed to the Permian sands.

The source of Permian sands of southern Australia was thus largely high to medium grade metamorphic and plutonic rocks that made up the exposed rocks at Kanmantoo Group, Encounter Bay and East Antarctica. This conclusion is based on the study of quartz, feldspars and heavy mineral suites in the Permian sands and different rock samples. These results confirmed that the Permian sands have been derived mostly from high to medium grade metamorphic rocks such as gneisses and granitic gneisses which were similar in composition to those of eastern Antarctica.

This conclusion is very significant, because it implies that parts of East Antarctica were uplifted and formed an ice dome and sediments were transported toward southern Australia, forming the heavy and light mineral assemblages of the Permian sands. Divergent directions of ice flow probably resulted from rifted areas where the rift shoulders became the ice divides. Therefore, the source areas for the Permian glacial sediments of South Australian would have been different to the source areas for the Victoria and Tasmania sediments. Further research on this proposal is obviously necessary.

Paleo-ice flow directions, and provenance data for the Permian sands of southern Australia indicate that deposition has taken place in a stable cratonic region, that was very far from the paleo-Pacific margin of East Antarctica. These data also support the reconstructions of the Pacific margin of Gondwana which place together southern Australia and the Commonwealth Bay area in East Antarctica.

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APPENDIX 1

Surface texture analysis

Sample preparation: Garnet and quartz grains in the samples of the Permian sands and parent rocks were separated from other minerals. For the purity of garnet and quartz, a needle with glue was used to picked them up under the binocular microscope, then dissolved the glue in acetone. Finally the grains were washed and dried. The next step is that the grains were boiled with 5N HCl for ten minutes to remove any iron stain on the grain surfaces. The grains were washed many times and then put in an ultrasonic cleaner to clean the grain surfaces and then dried. The grains have been mounted and coated with gold for use in the S.E.M. An energy dispersive X-ray analysis (EDX) was used to check that representative grains were indeed quartz or garnet. The scanning electron microscope allows examination of the surface morphology of the samples with excellent resolution (down to 40 angstrom units) and a great depth of field (300 times greater than of a light microscope). Techniques such as back-scattered electron imaging and X-ray analysis allow specimen composition to be determined, and diffraction patterns obtained by electron channeling can provide information about crystal structure and orientation. The electron beam is generated using either a hot Tungsten filament or a thermally heated Lanthanum Hexaboride crystal.

APPENDIX 2

Petrography

Technique of modal analyses: Approximately 60 thin sections from Permian sands of southern Australia have been studied by petrographic microscope. At the same time for comparison of the Permian sands with parent source rocks, about 300 thin sections of rock samples from Permian erratics, East Antarctica , Encounter Bay and Kanmantoo Group were used in the same manner. Modal analyses for each thin section were carried out by point counting. In order to increase the accuracy and reproducibility of this study, I used a dual operator method of point counting (as utilized by Graham et al. 1976). I counted one half of each thin section, and the results were compared whit re-counts of the other half of the thin section. There was not any significant differences between two parts. For analysis of the polycrystalline detrital quartz grains I have used The methodology recommended by Basu et al. (1975).

Method of staining: Each thin section has been treated for ten second with hydrofluoric acid (HF) fumes and then stained with Na cobaltinitrite solution for one minute. After treatment the K feldspar grains are intensely yellow and plagioclase are white.

APPENDIX 3

Cathodoluminescence (CL)

In the vacuum chamber of the CL stage, the electron beam is adsorbed by the surface it hits, much of energy of the incident electron beam is adsorbed by the specimen molecules causing an increase in the energy levels of adsorbing electrons. Normally the excited atoms (also termed Cathodoluminescence centers) return to the ground state by transfer of the excess energy to adjacent atoms by inelastic collisions. Under certain circumstances, the adsorbed energy is re-emitted as light energy in the visible range before these collisions can take place. The conditions for luminescence often occur in impure crystalline substances where the impurities act as the luminescent centers. The intensity of light emitting from any particular point will be proportional primarily to the surface density of luminescent centers. The electron energy is readily adsorbed near the surface, and little luminescence is emitted from below the surface.

CL stage uses an electron beam to produce Cathodoluminescence in susceptible materials. The CL 8200 Mk2 stage is used to generate an electron beam rated at 20kV and when applied to the electron gun, this generated a current of 0.5 mA. This electron beam is capable of producing X-rays from any material it strikes. The stage is designed to adsorb this radiation, but regular periodic inspections are important to ensure that leaks have not developed. Mini Instruments Ltd makes a compact radiation detector suitable for testing the CL stage. This equipment design has been checked for X-ray safety. X-rays production was measured using a Philips monitor type PW 4517 placed directly above the top window. Polished sections, about 100 micron thick, were made from each sample. Samples were not coated with conducting material. The extra thickness is needed for heat dissipation since the standard thin sections have a tendency to crack and burn. Abnormal heating of thinner sections can alter the colour of cathodoluminescence. CL photomicrographs were taken with Kodak high speed Ekta-Chrome at an effective ASA of 1600 reversal colour film using exposure time for quartz that varies from 10.45 to 38.34 minutes depending up on the magnification of the objective used.

Interpretation of luminescence have been done on the colour print and point counting analysis can also be easily undertaken using the colour print to find the ratio of different CL colour on quartz grains from different samples.

APPENDIX 4

Atomic Absorption

Sampling and analysis: Quartz grains from forty five samples of Permian erratics, Antarctica, Encounter Bay and Permian sands have been separated from feldspar grains using a sodium polytungstate liquid, made into a solution of density to the level that the feldspar was floated in the liquid. For purity of quartz grains they were placed under a 10X binocular microscope and about 1mg clean quartz has been picked out by hand and treated with cold concentrated hydrochloric acid for 18h to remove any iron oxides on the grain surface.

Details of the digestion procedure

Following procedures for wet chemical analysis have been done for each samples.

1- Weighting the samples: The weight ranges from 0.974 to 2.4914 gram in teflon beakers.

2- Add the acids required for digestion: 1ml of concentrate HCl (70%) and add 10ml of concentrate HF (50%).

3- Heating the solutions for 24 hours to ensure that the solution has completely evaporated.

4- Remove beakers from hot plate and allow to slightly, cool then add 10ml HCl and add 10ml La/K solution.

5- Make up to final volume(100ml).

Finally, the solutions were analysed by Atomic Absorption to find the trace elements in each sample and then compared with each others.

APPENDIX 5

Heavy minerals analysis

Analysis: Analysis took place in four stages: sampling, preparation, separation and data treatment.

Sample collection

Samples were collected from different localities in the Permian glaciogenic sediments. About 500grm of sand samples from each bed unit were taken and at the same time samples of Permian erratics were collected. Rock samples from eastern Antarctica, Encounter Bay and Kanmantoo Group have also been chosen for analysis. These samples were crushed to sand size to compare the physical and chemical characteristics of the heavy and light minerals with the Permian sands.

Preparation: At this stage, the sample was washed by wet sieve through 65 micron to remove all of the remaining clay and silt, then dried at low temperature. The size fractions of each sample chosen for further examinations were 2.5Ø, 3Ø and 3.5Ø. The surface of sand fractions were cleaned with 10% HCl to dissolve out the iron oxides on the grain, and then washed and dried. Now the samples was ready for separation into light and heavy fractions.

Separation: In the laboratory, the heavy were separated from light minerals by using different heavy liquids. The scheme adopted is based on heavy liquid and magnetic separations and is summarised in the main text (Fig. 7.0). Magnetic separations were carried out to facilitate the identification of opaque heavy minerals. Although the Frantz electromagnetic separator can be used for separating heavy minerals, it is slow and necessitates preliminary sizing to produce good results. In the present study, to find the weight percentage of non-opaque and opaque minerals the Frantz electromagnetic separator was used. First, each sample was magnetically separated by using a bar magnet, then an electromagnetic separator was used. All samples from Permian sands and parent rocks were mounted with resin together prior to grinding and polishing the sample in preparation for microscopic and electron microprobe analyses.

The light minerals were separated from heavy by using a sodium polytungstate (3Na2WO4 9(WO3)H2O, an inorganic, non-toxic solution with density of 2.9. For further studies the methylene iodide solution with a density of 3.32 and Clerici's solution with a density of 4.2 were also used. As a precaution against toxic fumes, during the procedure of separation, the apparatus was placed inside a fume cupboard. Gloves were worn when handling Clerici's solution and extra care was taken to avoid spillage. Further information on the handling of the heavy liquids can be found in Hauff & Airey (1980). Once separation has been achieved, the heavy minerals were mounted on glass slides and examined under a petrological microscope. Counting a total of 150 detrital non-opaque grains is usually considered to give a reasonable estimations of the mineral proportions. Based on this assumption, 150 detrital non-opaque grains were counted.

Data treatment:

Electron microprobe analyses (EMPA): The chemical composition of heavy minerals were identified by electron microprobe analyses. The electron microprobe operates on the same principles as the scanning electron microscope, but has been optimized for use as an analytical instrument. Spectrometers operating on either the wavelength or energy dispersive principle are used to measure the characteristic X-rays emitted from a sample under electron irradiation, and a computer provides on line data reduction. Analyses from micron-sized areas of a sample can be obtained for elements ranging from Boron to Uranium, at levels down to 100 ppm or better.

The JEOL-733 Superprobe installed in the CEMMSA, Adelaide University, features three dual crystal wavelength dispersive spectrometers, as well as a kevex energy dispersive X-ray analyser, and may be operated in a variety of modes, from fully computer controlled (unattended) to manual operation. Special thin sections or polished sections were made for Electron Microprobe Analyses. Carbon-palladium was used as a standard preparation for a conducting layer.

The latter is computerized to carry out structural formula determination using oxygen number input (Griffin 1979).

Heavy minerals composition

Following separation of heavy minerals by gravity settling, they were mounted and coated with carbon and analysed with electron microprobe (E.D.S). The chemical composition of selected heavy minerals and their various types were determined by electron microprobe to assist the interpretation of provenance, because this is a rapid and effective method of confirming the optical identification of minerals, particularly of rare or problematic grains. Each grain was analysed for a period of 70sec´, and total of 50 grains garnets, sillimanite, staurolite, rutile, tourmaline, epidote, sphene and apatite grains were analysed per-sample. This was sufficient to delineate the full compositional ranges in each case. Most studies of this kind are made on polished thin sections whereas this study is based on analysis of grain surface. Analysis of grain surfaces allows a relatively rapid determination of the compositional ranges of the minerals. The center of each mineral were also analysed.

APPENDIX 6

Feldspar composition

Analytical procedures: More than one hundred grains of detrital feldspars were analysed for Ca, K and Na with the electron microprobe analyser. The microprobe was operated at 30-nA sample 200uA emission current and 15-Kv excitation voltage. Counting time was approximately 20 sec at constant integrated beam current.

Two analyses were made for each grain: one near the grain center and the other at the edge of the grain. The chemical composition of these two points are nearly similar. Analyses that totalled less than 97% or greater than 102% feldspar molecule (weight percent) were eliminated. The remaining analyses of feldspar were recalculated for the percentage of An, Or and Ab in each grain on the base of 32 (O).

APPENDIX 7

Analytical techniques for opaque minerals

Electron microprobe analysis is particularly valuable as a confirmatory test for opaque minerals. Without such chemical analysis for example it would be extremely difficult to distinguish ilmenite from magnetite.

Separation:

Ilmenites were magnetically separated without using any heavy liquid. Minor amounts of magnetite were removed by a hand magnet and the 0.0 amp settling on the Frantz separator, using a forward and side slope of 20°. The 0.1amp to 0.3amp of Franz separation was used to separate ilmenite grains in all the samples. In addition, because nearly all separations contained hematite, as shown by X.R.F, and because pure ilmenite is rare, all samples that were separated by the Frantz have been treated for 18 hours with cold, concentrated HCl to remove hematite.

Electron microprobe analysis

Electron probe microanalysis was carried out for eight elements (Ti, Al, Mg, Mn, , Cr, V, Fe, Si) with an electron microprobe analyses at the University of Adelaide. All analyses were performed with a Jeol model 50A electron microprobe with a 15Kv electron beam regulated at beam current of 0.12 microamps. All Fe was calculated as FeO for analysis. Approximately 60 grains were chosen from each slide for analysis. Two points per grain were analysed to reduce the effects of possible variability of the chemical compositions of individual grain. These points were averaged, giving the results reported in chapter seven. Counting time was approximately 6 minutes for each grain at constant integrated beam current.

Chemical composition

In this study, detrital ilmenite grains were analysed by microprobe (W.D.S).

Electron probe microanalysis (E. P. M) was conducted on 60 grains of detrital ilmenite for each sample and each grain took about six minutes to be analysed.

APPENDIX 8

Geochemistry

Geochemistry of Permian sands

Chemical analyses were made for Permian sands from different localities of southern Australia using x-ray fluorescence except for carbonate. Result from the chemical composition of the Permian sands which are shown in table. S have more Al2O3 and more K2O than CaO and MnO. the chemical composition of these sands also provides an opportunity to see how sand composition relates to plate tectonics as well as how depends on climate, and also on maturity.

	*	**	***
Sio2	96.5	96.5	95.5
Al2o3	1.4	1.3	1.6
Fe2o3	0.5	0.5	0.9
Mno	0.0	0.0	0.0
Mgo	0.0	0	0.1
Cao	0.1	0.0	0.0
Na2o	0	0	0
K2o	0.6	0.6	0.7
Tio2	0.1	0.0	0.1
P2o5	0	0	0.0
So3	0	0	0
Loi	0.1	0.6	0.2
Total	99.2	99.1	99.1

Table. S. Chemical composition in the Permian sands at different localities of southern Australia.

***: Kings Point**
****: Cape Jervis**
*****: Hallett Cove**

Geochemistry of erratics

The chemical composition of these suites have been studied and the average of these suites are presented in Table. E. Specimens of Permian erratics selected for analyses were first cleaned of weathering products with a cut saw. These samples were then crushed in a jaw crusher to small pea size fragments and then were crushed to a fine powder using a chrome steel grinding vessel. Chemical analyses were determined on this powder by X-Ray fluorescence using a modified Philips X-Ray.

Table. E. Chemical composition and trace elements in the Permian erratics of southern Australia.

SiO2	98.2	97.8	96.1	96.4	74.1	71.1	73.9	68.9	74.4	71.8	73.8	59.2	60.2	54.1
Al2O3	0.8	0.7	1.6	1.7	13.2	14.5	12.9	15.7	12.8	15.5	13.6	19.6	20.8	2.7
Fe2O3	0.4	0.6	0.6	0.1	2	2.3	1.9	3.4	1.8	1.5	1.4	6.8	1.5	0.7
MnO	0	0	0	0	0	0	0	0.1	0	0	0	0.1	0	0.4
MgO	0	0	0	0.1	0.1	0.2	0.3	0.4	0.2	0.3	0.2	3.7	4.5	0.4
CaO	0	0	0	0	0.2	0.3	1	1.9	0.5	2.3	0.6	2.3	0.5	21.8
Na2O	0.2	0.2	0.2	0.2	3.2	3.5	2.9	3.8	2.7	3.7	3.3	2.6	8.7	0.3
K2O	0.5	0.4	0.9	0.8	5.6	5	5.3	5.1	5.6	4.1	5.6	1	0.6	0.9
TiO2	0.1	0	0.1	0.1	0.2	0.4	0.3	0.3	0.2	0.2	0.3	0.8	0.4	0.2
P2O5	0	0	0	0	0	0.1	0.1	0.1	0.1	0	0	0.1	0.1	0
SO3	0	0	0	0	0	0	0	0	0	0	0	0	0	0
LOI	0.2	0.1	0.2	0.4	0.8	0.7	0.3	0.4	0.6	0.5	0.5	2.1	2.3	17.6
Total	100.4	99.8	99.6	99.7	99.4	99.2	99.1	99.9	98.9	99.8	99.3	99.3	99.7	99.1
Y	1.6	2.8	2.9	3.4	20.8	58.5	39	26.4	28.4	11.1	16.2	34.7	31.2	25.5
Sr	9.3	10.8	20.1	17.4	43.8	117.2	89.5	341.9	90.2	318.7	134.2	230.7	115.3	131
Rb	11	10.4	21.7	21.3	213.1	251.2	264	158.4	330.1	116	254.7	89	15.9	32.2
Nb	2.5	2.8	4.3	2.4	23	17.8	16.6	13.7	14.8	8.8	20.1	16.7	9.8	3.4
Zr	78.3	99.9	163	145.1	233.6	291.9	238.6	220.2	118.3	143.3	118.1	146.7	131.2	122
Pb	2.1	2.7	4	5.1	19.4	21.3	26.9	26.3	35.1	24	15.4	49.1	5.5	29.7
Th	0.6	0.1	0.8	0.2	21.3	26.9	26	17	28.5	7.2	18.2	17.2	14.3	4.2
U	1.5	1.2	1.8	1.8	6	6.2	5	6.7	5.9	3.1	3.2	5.3	3.2	2.7
Ga	2.4	2.3	3	2.7	19.4	18.5	14.5	20.6	17.5	17.1	19.2	27.8	18.9	3.7
Cu	8	15	24	14	6	12	6	14	17	12	11	18	9	2
Zn	9	10	13	10	43	37	37	68	35	28	23	83	17	18
Ni	12	15	19	0	14	14	30	19	18	19	17	56	33	16
Ba	95	77	168	164	415	914	696	1284	383	890	407	433	152	180
Sc	0.6	0.5	0.1	0.5	2.5	7	4.2	4.5	3	2.8	2.9	22.1	10.9	2.3
V	6.4	7.7	9.3	6.7	9.1	20.4	17.6	11.4	10.7	12.3	24.5	152.6	53	18.3
Co	17.5	5.2	0.6	87.6	0.8	2.7	1.6	2.7	1.1	1.7	0.7	21.5	3.6	5
Ce	0	0	0	0	28	173	109	103	95	18	50	78	30	27
Nd	4	1	2	2	15	102	49	45	45	8	23	34	17	26
La	0	0	0	0	13	84	55	56	47	11	29	33	15	13
	granite	**gneisses**	**rhyolite**	**porphyry**	**rhyolite**	**andesite**	**volcanic**	**volcanic**	**rhyolite**	**volcanic**	**andesite**	**porphyry**	**porphyry**	**dacite**

Table. E. Chemical composition of unknown minerals in the Permian sands, erratics and East Antarctica.

	*	*	*	*	**	***
SiO_2	60.3	29.4	17.3	52.7	75.1	42,3
TiO_2	38.3	64.8	78.5	44.9	22,9	55.1
Al_2O_3	0.5	0,9	0.1	0	0,8	0.2
FeO	0.1	2.6	0.2	0	0	0.3
MnO	0	0	0.3	0.8	0	0
MgO	0.1	0.2	0.2	0.3	0.1	0
CaO	0.5	0.2	0.3	0.2	0.5	0.3
Na_2O	0.3	0.2	0.5	0.8	0.3	0
K_2O	0	0	0	0.2	0.4	0
Cr_2O_3	0	0	0	0	0.3	0
Total	100.1	98,1	98.1	100.2	99.8	98,4

*	**Permian sands**
**	**gneisses of Permian erratics**
***	**gneisses of East Antarctica**

	Kings Point	Cape Jervic	Hallett Cove
Sio2	96.5	96.5	95.5
Al2o3	1.4	1.3	1.6
Fe2o3	0.5	0.5	0.9
Mno	0	0	0
Mgo	0	0	0.1
Cao	0.1	0	0
Na2o	0	0	0
K2o	0.6	0.6	0.7
Tio2	0.1	0	0.1
P2o5	0	0	0
So3	0	0	0
Loi	0.1	0.6	0.2
Total	99.2	99.1	99.1

Table. S. Chemical composition of the Permian sands in different localites of southern Australia.

List of representative rock samples.

Locality	Number	Name	Locality	Number	Name	Locality	Number	Name
East Antarctica	335/400	Aplite	**Encounter Bay**	283/319	Granite	**Victoria Land**	957/7227	Gneisses
	335/72	Gneisses		283/425	Granite		957/7229	Gneisses
	335/18	Granite		283/336	Granite		957/7230	Amphibole
	335/13	Gneisses		283/341	Granite		957/7228	Gneisses
	335/219	Granite		283/342	Granite		957/7225	Greywacke
	335/629	Charnockite		283/445	Granite		957/7224	Gneisses
	335/640	Granite gneiss		283/444	Granite		975/7226	volcanic
	3235/219	Granite		283/450	Granite		957/7231	Granite
	335/117	Granite		283/913	Granite		957/7223	Volcanic
	335/15	Aplite						
	335/32	Gneisses						
	335/1	Gneisses						
	335/35	Gneisses						
	335/100	Gneisses						
	335/642	Gneisses						
	335/144	Granite gneiss						
	335/642	Basic gneiss						
Kanmantoo Group	281/226	Pelitic schists	**Permian erratics**	345/18	Phorphyrye	**Permian erratics**	345/21	Granite
	282/220	Staurolite schists		345/468	Granite		345/260	Granite
	282/221	pelitic schists		345/2	Arkosic		957/1	Siltstone
	282/20	Pelitic schists		345/3	Basic volcanic		957/12	Granite
	281/232	Staurolite schists		345/4	Granite gneiss		957/11	Basic volcanic
	281/219	pelitic schists		345/252	Porphyrye		957/19	Gneiss
	281/225	Pelitic schists		345/160	Porphyrye		957/4	Granite
	F.54a	Staurolite schists		345/300	Porphyrye		957/3	Arkose
	F.12	pelitic schists		345/169	Porphyrye		957/7	Siltstone
				345/200	Granite		957/14	Basic volcanic
				345/5	Granite		957/10	Granite
Commonwealth Bay	T.51	Granite		345/266	Gneiss		957/5	Gneiss
	T.54	Granite		345/476	Geniss		957/15	Granite
	T.62	Gneiss		345/500	Gneiss		957/17	Granite
	T.64	Granite		345/100	Porphyrye		957/30	Porphyrye
	T. 63	Gneiss		345/6	Porphyrye		957/31	Granite
	T.44	Phyllite		345/429	Porphyrye		957/32	Gneiss
	T.47	Phyllite		345/208	Porphyrye		957/33	Aplitic
	T.59	Gneiss		345/276	Granite		957/34	Porphyre
	T.60	Granite		345/462	Granite		957/35	Granite
				345/440	Granite		957/36	Porphyrye
				345/1	Porphyrye		957/37	Gneiss
				345/177	Porphyrye		957/38	Granite
Mt Monster	957/104	Volcanic		345/176	Porphyrye		957/39	Arkosic
	957/103	Andesite		345/162	Porphyrye		957/40	volcanic
	957/211	Porphyrye		345/7	Porphyrye		957/41	Andesite
	957/101	Porphyrye		345/283	Volcanic		957/42	Rhyolite
	957/102	Rhyolite		345/207	Gneiss		957/43	Porphyrye
				345/162	Andesite		957/44	Porphyrye
				345/22	Basic volcanic		957/45	Alkali granite
				345/208	Acidic volcanic		957/46	Porphyrye
				345/8	Acidic volcanic		957/47	Andesite
				345/9	Porphyrye		957/48	Gneiss
				345/10	Basic volcanic		957/49	Porphyrye
				345/11	Basic volcanic		957/50	volcanic
				345/190	Granite gneiss		957/51	Andesite
				345/12	Porphyrye		957/52	Volcanic
				345/266	Porphyrye		957/53	Gneiss
				345/23	Alkali granite		957/212	Rholite
				345/60	Basic volcanic		957/213	Dacite
				345/26	Porphyrye		957/214	Gneiss
				345/13	Siltstone			
				345/14	Granite			
				345/15	Tuff			
				345/16	Andesite			
				345/17	Volcanic			
				345/18	Granite porphyrye			
				345/19	Basic volcanic			
				345/20	Granite			
				345/21	Granite			

Table. 4.15A . Composition of clinopyroxene in the rock samples from Victoria Land of East Antarctica.

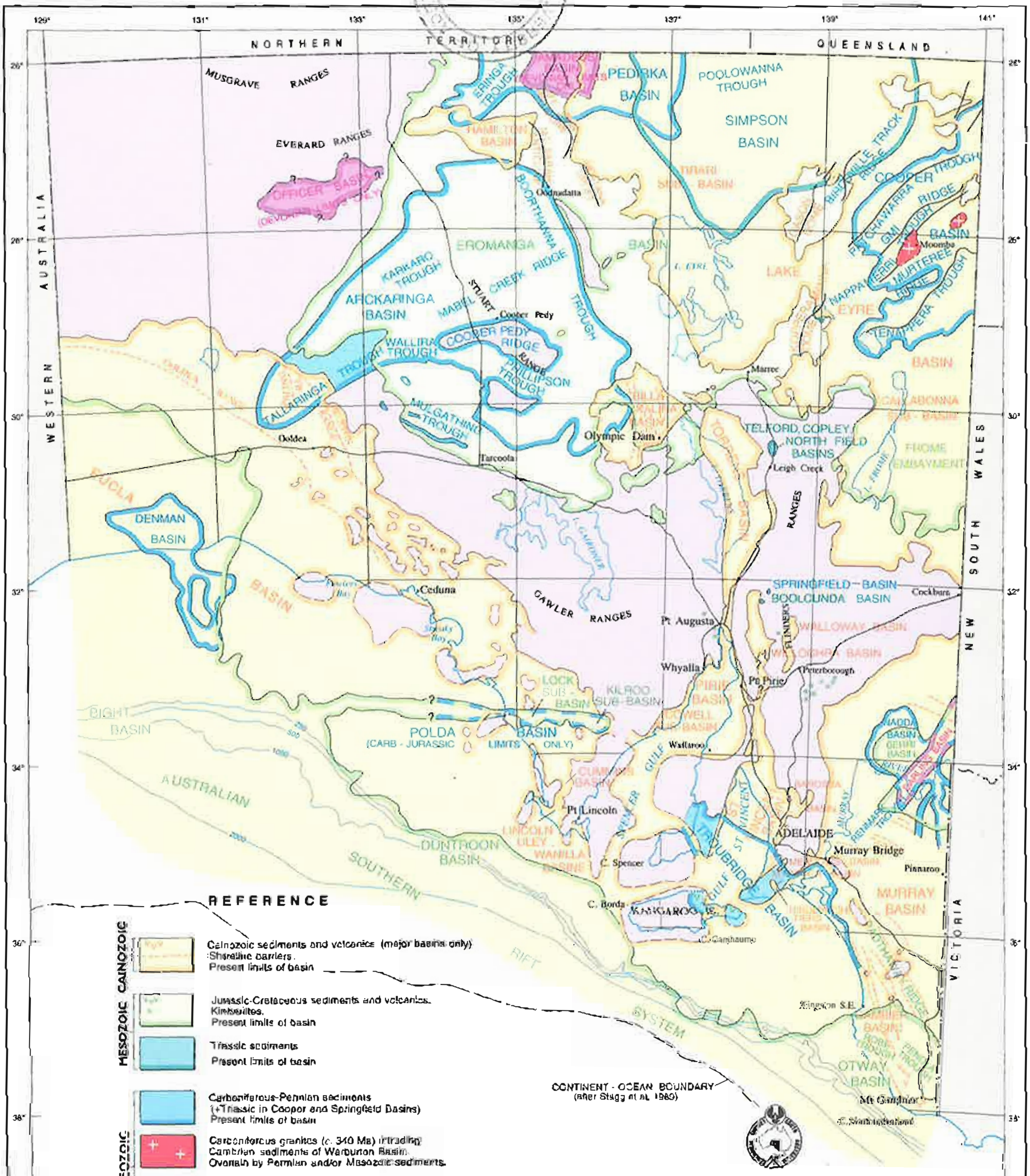
SiO2	51.4	51.2	51.1	50.3	51.6	48.3	52.1	51.5	51.9	51.9	49.1	51.4
TiO2	0	0.3	0.3	0.1	0	0.5	0	0	0.2	0	0.1	0.2
Al2O3	0.8	1.7	1.6	1.6	1.3	7.1	0.6	1.1	0.6	0.8	6.9	1.3
FeO	12.7	13.6	13.9	17.9	13.6	18.7	13.4	14.4	12.5	13.5	20.4	16.9
MnO	0.3	0.4	0.5	0.6	0.3	0.3	0.4	0.6	0.3	0.4	0.6	0.6
MgO	10.1	11.4	10.9	10.2	11.1	11.2	10.9	10.9	11.1	10.8	10.1	9.9
CaO	24.5	21.8	21.5	20.7	23.1	11.6	23.2	22.2	24.4	23.3	12.2	19.7
Na2O	0.1	0.2	0	0.2	0	0.7	0.1	0.5	0	0	0.4	0.5
K2O	0	0	0	0	0	0.5	0.1	0	0	0.2	0.4	0
Cr2O3	0.1	0	0	0.3	0	0.2	0.1	0.1	0.2	0	0	0
Total	99.9	100.6	99.8	101.8	101.1	99.1	101.1	101.3	101.3	100.8	100.2	100.6

Table. 4.15A. continue

SiO2	51.9	51.8	50.7	51.5	50.8	51.1	48.4	50.5	52.3	51.5	51.1	47.2
TiO2	0.1	0	0.5	0	0	0.1	0.1	0.4	0	0.1	0	0.9
Al2O3	0.5	1.1	0.9	0.6	0.9	1.8	7.8	1.4	0.6	0.5	5.2	7.9
FeO	13.2	14.4	13.8	13.1	14.2	14.3	20.3	13.9	13.1	12.7	18.1	19.6
MnO	0.8	0.7	0.6	0.7	0.8	0.7	0	0.8	0.5	0.6	0.4	0.4
MgO	10.7	9.9	11.1	11.3	11.1	10.5	9.7	10.6	10.9	11.1	11.9	9.9
CaO	24.2	23.2	21.7	23.6	22.4	21.8	12.3	22.2	24.2	24.1	12.9	12.4
Na2O	0.1	0	0.6	0.1	0.2	0.2	0.2	0.1	0	0	0	0.7
K2O	0.2	0	0.1	0	0	0	0.8	0	0	0	0.1	0.5
Cr2O3	0	0	0	0.1	0	0.2	0	0.3	0	0	0	0
Total	101.8	101.1	99.9	100.9	100.4	100.6	99.4	100.1	101.7	100.7	99.7	99.5

Table. 4.15A. continue

SiO2	51.8	51.4	49.8	50.4	51.5	51.7	51.8	51.1	50.9	51.4	50.4	51.9
TiO2	0	0	0.8	0.1	0.2	0	0.6	0.2	0.5	0.3	0.2	0.2
Al2O3	0.9	0.7	5.5	1.7	1.8	1.1	2.5	1.7	1.6	0.7	0.6	0.9
FeO	12.6	13.4	18.9	15.1	14.2	13.8	15.9	16.3	13.9	12.9	13.3	13.6
MnO	0.5	1.1	0.3	0.6	0.5	0.5	0	0.4	0.4	0.5	0.7	0.5
MgO	11.4	11.6	11.8	11.3	11.1	11.4	11.7	9.3	11.1	11.1	11.1	11.1
CaO	23.7	23.1	11.6	20.1	22.1	23.1	18.1	21.2	21.8	23.6	22.8	23.3
Na2O	0.1	0	0.6	0	0.3	0	0.1	0	0.2	0	0.4	0.1
K2O	0	0	0.1	0	0	0	0	0	0	0.1	0	0
Cr2O3	0	0	0	0	0.2	0.1	0	0	0	0	0	0
Total	100.1	101.3	99.3	99.3	101.8	101.7	100.7	100.2	100.5	100.6	99.3	101.8



REFERENCE

CAINOZOIC	🟨	Cainozoic sediments and volcanics (major basins only) Shoreline barriers. Present limits of basin
MESOZOIC	☐	Jurassic-Cretaceous sediments and volcanics. Kimberlites. Present limits of basin
	🟦	Triassic sediments Present limits of basin
PALAEOZOIC	🟦	Carboniferous-Permian sediments (+Triassic in Cooper and Springfield Basins) Present limits of basin
	🟥	Carboniferous granites (c. 340 Ma) intruding Cambrian sediments of Warburton Basin. Overlain by Permian and/or Mesozoic sediments.
	🟪	Mid-Palaeozoic sediments Present limits of basin
		Delamerian Orogeny (c. 500 Ma).
PRECAMBRIAN	☐	Precambrian to Cambro-Ordovician basement, commonly with thin Cainozoic cover (including palaeochannel sediments).

Geological boundary.................
Fault.................
Major fold axial trace.................
Isobath (metres).................

SCALE

0 50 100 150 200

KILOMETRES

MIDDLE PALAEOZOIC TO CAINOZOIC BASINS OF SOUTH AUSTRALIA

GEOLOGICAL SURVEY OF SOUTH AUSTRALIA
DEPARTMENT OF MINES AND ENERGY

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