Antarctic Science 3 (2): 205–216 (1991)

U-Pb zircon dating of Proterozoic igneous charnockites from the Mawson Coast, East Antarctica

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Abstract: We report ion-microprobe U-Pb zircon ages from charnockites of a large Proterozoic composite batholith, Mawson Coast, Australian Antarctic Territory. The charnockites crystallized from orogenic magmas of intermediate composition (mainly 54–68% SiO_2) intruded into a granulite-facies metasedimentary gneiss sequence between the second and third recognized deformations. A sample of low-Ti charnockite provides an age of 954 ± 12 Ma and a high-Ti charnockite is dated at 985 ± 29 Ma (all ages quoted at 95% confidence). The age difference is not significant at the 95% confidence level. Both these ages were obtained from zircons with igneous zoning and/or morphology and thus are thought to date igneous crystallization. Zircons from a felsic gneiss xenolith within the charnockite have cores of various ages, many from 1.7 to 2.0 Ga, but with other grains between 1.0 and 1.5 Ga and a single 2.5 Ga zircon. These zircon cores are direct evidence for an early to middle Proterozoic age for the supracrustal basement sequence in this mobile belt. Many of these zircon cores are concordant but abundant discordant grains suggest a complex history of multiple Pb-loss events. Zircon rims grew at 921 ± 19 Ma, probably during the post-charnockite deformation (D_3). Previously obtained Rb-Sr dates for charnockite of 886 ± 48 Ma and 910 ± 18 Ma were probably also reset during D_3. A Rb-Sr isochron date of 1061 ± 36 Ma previously reported for high-Ti charnockite from Mawson Rock is thought to be erroneous, and a new date of 959 ± 58 Ma (consistent with both the igneous and reset dates above) is interpreted from those data.

Received 12 March 1990, accepted 10 December 1990

Key words: batholith, deformation, ion-microprobe, metamorphism, U-Pb geochronology.

Introduction

True orogenic magmas of Grenville age (*ca.* 1000 Ma) are considered rare throughout Proterozoic mobile belts of the world: anorogenic models are favoured for many rapakivi granites and charnockites found in association with massif anorthosites (Emslie 1978, Morse 1982, McLelland 1986). An important exception is the very large intrusive complex of charnockitic (orthopyroxene-bearing) granitoids at Mawson Coast, Antarctica (Crohn 1959, Trail 1970, Sheraton 1982, Young & Ellis 1990). Although geochemically similar in some trace element abundances to anorogenic or A-type granites (Sheraton & Black 1988), the Mawson charnockites are distinctly more mafic and indeed have unusual compositions compared to most crustal granites (Young & Ellis 1990). These granitoids were intruded into basement gneiss between the second and third deformations recognized in the Mawson Coast area (Clarke 1988) and are thus orogenic in nature. Several suites of granitoid are recognized from petrography and geochemistry and are grouped into low-Ti and high-Ti series (Young & Ellis 1990, 1991).

We have undertaken a U-Pb zircon study by ion-microprobe to determine more precisely the emplacement age of selected charnockites and constrain the ages of deformational events and the basement gneiss. Precise dating of the charnockites also is needed to provide the igneous reference point for a Nd and Sr isotopic study (D.N. Young, M.T. McCulloch & D.J. Ellis, unpublished data) which shows many charnockites to have very evolved isotopic compositions. The Sensitive High-Resolution Ion Microprobe (SHRIMP) of the Australian National University is ideal for dating in a complex terrain because it can analyse 30 μm spots on zircon cores and later overgrowths. This has allowed many recent studies to resolve distinct Precambrian igneous and metamorphic events in which zircon grew (Black *et al.* 1986, Kinny 1986, Kinny *et al.* 1988, Compston *et al.* 1986, Compston & Kröner 1988), and to place constraints on sediment deposition ages (Froude *et al.* 1983, Schiotte *et al.* 1988, Kröner & Compston 1988, Compston *et al.* 1987, Kröner *et al.* 1988)

Geological background

The Mawson Charnockite is a composite batholith made up of many plutons of foliated, orthopyroxene-dominated granitoids cropping out over at least 3130 km^2 and perhaps more than 5000 km^2 (Fig.1). The plutons range in composition from mafic to felsic but are dominantly intermediate, with SiO_2 from 54 to 68% (Young & Ellis 1991, Sheraton 1982, Sheraton & Black 1983, 1988) in marked contrast to the

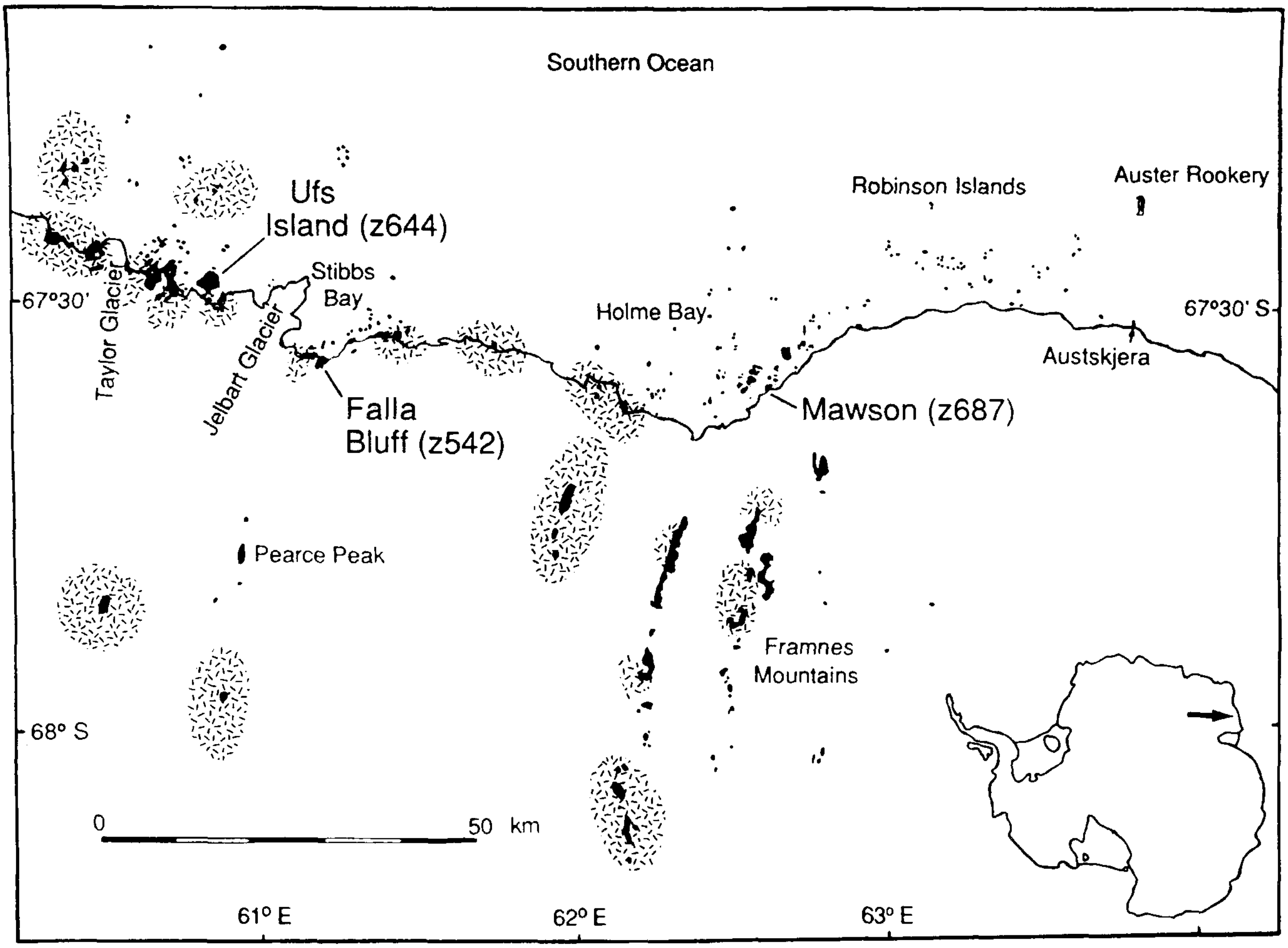

Fig. 1. Map of Mawson Coast, East Antarctica, showing areas of rock outcrop (black) and sample locations at Falla Bluff, Ufs Island and Mawson Station. Ornamented areas are basement gneiss, other outcrops are charnockite

bimodal felsic and mafic magmatism which characterizes many other Proterozoic terrains (Etheridge *et al.* 1987). Mineralogy and texture of the charnockites vary widely. Clinopyroxene occurs only in metaluminous mafic samples (peraluminosity increases with SiO_2) but orthopyroxene occurs in all compositions. Clinopyroxene, orthopyroxene and biotite are igneous, as seen from euhedral mineral inclusions in feldspars (Young & Ellis 1990). Some metamorphic biotite also formed, along with quartz, as a hydration product of orthopyroxene and feldspars. Hornblende was developed as a retrograde metamorphic product after pyroxenes in some metaluminous mafic samples. Garnet developed in felsic, peraluminous rocks as metamorphic coronas around orthopyroxene and ilmenite where they were originally in contact with plagioclase. Igneous garnet, indicated by euhedral morphology and rare earth element evidence, is found only in one suite of rocks (Young & Ellis 1990). Clinopyroxene and garnet were not observed to coexist in the same rock, except for one unusually Fe-rich and Mg-poor enclave.

The charnockites intruded isoclinally-folded granulite-facies gneiss after D_2 but before D_3 (Clarke 1988). They are foliated and contain xenoliths and enclaves which were flattened by D_3 but show neither the effects of D_2 nor the intense flattening which characterizes the D_1 structures dominant in the basement gneiss. The second and third deformations, and possibly also the first, may be ascribed to the Rayner event of earlier authors (Sheraton *et al.* 1980, Ellis 1983).

Previous dating studies

The charnockite at Mawson Station has been dated by Rb-Sr isochron and conventional U-Pb zircon methods (Black *et al.* 1987). A Rb-Sr isochron date of 1084 ± 37 Ma (P.A. Arriens unpublished data, quoted by Sheraton 1982 and Tingey 1982) has been recalculated to 1061 ± 36 Ma (16 data points used: group A in Fig. 2) using the decay constant of Steiger & Jäger (1977). Those samples assigned to group A by Arriens are mainly charnockites whereas those assigned to group B are xenoliths and enclaves and were not used in calculating the 1061 Ma isochron by Arriens. However, a younger and probably more meaningful age of 959 ± 58 Ma is obtained from Arriens' group A data by discarding one high and one low Rb/Sr sample which are unlikely to be cogenetic with the other samples. This leaves 14 samples of charnockite which have a restricted range of Rb/Sr but are clearly all of the one rock type.

A good isochron which fits the data within error (Model 1 of McIntyre *et al.* 1966) is obtained by leaving out one of the 14 cogenetic data points from the regression, giving the 959 ± 58 Ma age. Xenolithic charnockitic rocks analysed by Arriens (group B on Fig. 2, plus the low Rb/Sr sample from group A) display a range of initial $^{87}Sr/^{86}Sr$, assuming that they are of the same age, and might represent various rock types not directly related to the host charnockites.

Black *et al.* (1987) obtained an 886 ± 48 Ma Rb-Sr whole rock isochron from charnockites at Mawson Station which have lower Rb/Sr than those of Arriens' group A and which are the medium-grained granodiorite xenoliths described by Sheraton (1982). The samples were collected from a small volume (1 m^3) and are almost certainly cogenetic and of the same age; the scale of sampling by Arriens is not known exactly but is likely to have been significantly greater because of the wide variety of rock types (mainly xenoliths) collected.

The 886 ± 48 Ma date probably represents resetting of the Rb-Sr system over the scale of at least a metre or two during post-igneous deformation and/or metamorphism. The revised 959 ± 58 Ma date is more likely to represent the age of igneous crystallisation, because the samples were apparently more widely spaced apart and the re-equilibration of Sr isotopes during metamorphism is known to be limited in areal extent (e.g. Black *et al.* 1987), but it is still within error of the 886 Ma date.

An isochron date of 910 ± 18 Ma, recalculated using the decay constant of Steiger & Jäger (1977), was obtained for charnockite at Mount Burnett, 25 km south of Mawson (P.A. Arriens, unpublished). Considering the small scale of sampling (from <1 m^3, R.J. Tingey, personal communication 1986) this isochron is likely to have been reset by D_3 deformation and metamorphism.

A U-Pb zircon age of 935 Ma determined by Black *et al.* (1987) for the Mawson Station charnockite is only a minimum emplacement age because the analyses are discordant. Thus, the original crystallization age of the charnockitic intrusives was not known before this ion-probe U-Pb zircon study, although Black *et al.* (1987) had inferred it to be older than 962 Ma.

The ages of the first two deformations and the basement gneiss are not well constrained. Tingey (1982) reported Rb-Sr isochron dates, obtained by P.A. Arriens, of 1254 ± 31 Ma and 1153 ± 47 Ma (here recalculated using the decay constant of Steiger & Jäger 1977), for banded gneiss from two outcrops in the Framnes Mountains, south of Mawson. It is uncertain whether these represent partial or total metamorphic overprints of the protolith age.

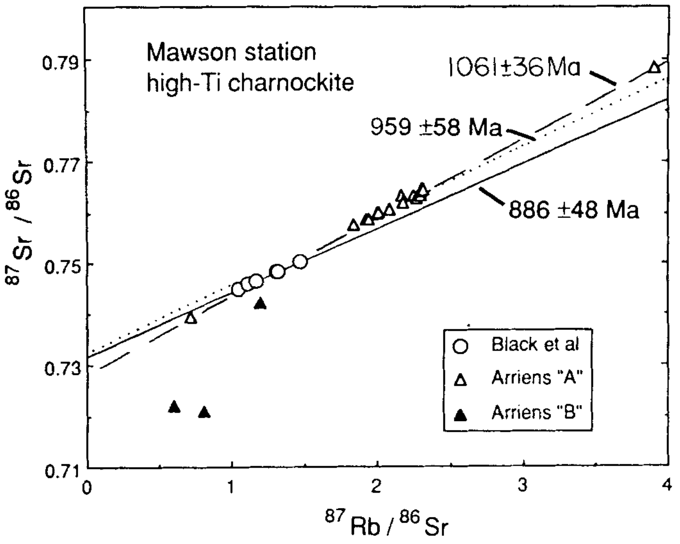

Fig. 2. Rb-Sr isochron diagram for charnockite samples from Mawson Station (Black *et al.* 1987, P.A. Arriens unpublished).

Sample description

To determine the igneous age of the charnockite batholith, zircons were separated from two charnockites representing both the low-Ti and high-Ti series identified by Young & Ellis (1991). Both charnockites are intermediate in composition and have orthopyroxene as the main mafic mineral, but sample z644 has markedly higher contents of Y, Ti, Nb and P, and lower Ni, Cr and Mg/Mg+Fe than z542 (Table I). These compositional differences at a given silica content distinguish the low-Ti from the high-Ti charnockite series, although in detail there are distinct sub-groups or suites recognized within the two main groups (Young & Ellis 1990). Some suites of low-Ti charnockite are depleted in heavy rare earth elements (HREE) which is thought to indicate high pressure crustal melting with garnet in the residue.

Sample z542 is a mafic example of a HREE-depleted low-Ti charnockite. In contrast, no high-Ti rock (z644 included) has any depletion in HREE.

Sample z644 is fine-grained (0.1–1 mm, occasional feldspars up to 5 mm) whereas z542 is medium-grained (0.2–2 mm, feldspars often up to 8 mm). No hornblende is present in either. Minor K-feldspar is present in both rocks, mainly as equant intergrowths within large antiperthite crystals which probably represent exsolutions. Plagioclase also occurs as smaller grains with no exsolved K-feldspar. Biotite, quartz, apatite, zircon and ilmenite are found in both rocks, whereas z542 also has magnetite.

Both charnockite samples are well foliated and have even-grained texture. Both the low-Ti z542 and the high-Ti z644 are tonalite according to the normative scheme of Barker (1979). They have small granular crystals of orthopyroxene and clinopyroxene that are free from any coarse exsolutions. Only a few of the least-deformed two-pyroxene rocks contain coarsely exsolved relict igneous pyroxenes which would indicate high igneous temperatures (cf. Bohlen & Essene 1978, Ranson 1986, Ollila *et al.* 1988). Most pyroxenes, however, totally recrystallized during D_3, including those in z542 and z644. The only direct evidence of high, igneous temperatures in z542 and z644 are relict phenocrysts of antiperthite which have ternary bulk compositions of 8–15 mol.% Or (Young & Ellis, 1991). These indicate temperatures of at least 950–1000°C using the ternary feldspar geothermometer of Furhman & Lindsley (1988). High temperatures are required if orthopyroxene is to be an igneous phase in charnockitic granites (Naney 1983).

The mineral chemistry of both rocks, typical of the foliated Mawson Coast charnockites, was metamorphically re-equilibrated during the D_3 overprint. Garnet-orthopyroxene-plagioclase-quartz geobarometry (Perkins & Chipera 1985) of more felsic charnockite samples gives pressure estimates of 5.5–6.5 kbar, which records D_3 metamorphic conditions and is a minimum estimate of igneous emplacement pressure (Young & Ellis 1991). Two-pyroxene temperatures (Lindsley 1983) of mafic charnockites are usually 650–750°C, indicating metamorphic re-setting.

In addition, a sample of basement gneiss was chosen to investigate the age of deformation and metamorphism. Orthogneiss rather than paragneiss was thought to be more appropriate on the basis of ease of interpretation. A paragneiss is likely to have a wide range of ages and morphologies of detrital zircon (eg. Kröner *et al.* 1988). In contrast, an orthogneiss is more likely to have restricted zircon populations, such as zoned igneous zircons which might contrast with later, unzoned metamorphic overgrowths (e.g. Kinny *et al.* 1988). The sample chosen for study (z687) was a large (10 x 4 m) felsic granitic xenolith from Mawson Station which has a strong S1 foliation and is interbanded with mafic gneiss in the xenolith. It was chosen because of well-developed core/rim relationships in its zircon population. A chemical analysis is shown in Table I. The high CaO at 70% SiO_2, the coarse granitic texture and the massive, homogeneous nature of the felsic gneiss bands are evidence for an igneous origin of the protolith. Interestingly, the high Ca and high K/Na of the pre-D_1 gneiss are similar to those of the distinctive post-D_2 charnockites of Mawson, which might suggest that similar charnockitic magmatism occurred before D_1. However, in most respects the gneiss, with its higher SiO_2, lower TiO_2 and P_2O_5 (Table I), is quite different from the high-Ti charnockites of the Mawson area.

Table I. X-ray fluorescence whole-rock geochemical analyses of samples in this study.

	z644 Maw64	z542 Maw16	z687 Maw121
Major Elements (weight %)			
SiO_2	54.63	57.40	71.44
TiO_2	2.54	1.07	0.69
Al_2O_3	14.71	17.09	13.08
Fe_2O_3	1.73	1.92	0.24
FeO	7.90	5.60	3.52
MnO	0.15	0.12	0.08
MgO	3.58	4.79	1.35
CaO	6.75	6.47	3.22
Na_2O	2.29	2.47	1.89
K_2O	2.80	1.77	3.98
P_2O_5	1.03	0.25	0.02
S	0.08	0.08	0.02
H_2O-	0.24	0.19	0.14
H_2O+	0.57	0.49	0.27
CO_2	0.49	0.52	0.38
Total	99.49	100.23	100.32
less O=S	0.04	0.04	0.01
Total	99.45	100.19	100.31
mg	44.7	60.4	40.6
MG	40.3	53.8	39.2
A/CNK	0.87	1.00	0.99
Trace elements (parts per million by weight)			
Ba	1120	655	1170
Rb	108	98	128
Sr	398	329	157
Pb	22	9	21
Th	10	<1	0.8
U	2	<1	0.8
Zr	505	139	374
Nb	31	8.5	11.0
Y	54	10	7
La	83	26	-
Ce	197	55	38
Sc	22	22	-
V	136	150	39
Cr	97	162	17
Mn	1150	965	-
Co	30	31	17
Ni	37	68	12
Cu	32	32	7
Zn	155	95	49
Ga	23.5	23	14.2

"Maw" prefixes indicate sample numbers referred to by Young & Ellis (in press). "z" numbers refer to the ANU SHRIMP catalogue. mg = 100.Mg/(Mg+Fe^{2+}), MG = 100.Mg/(Mg+FeTotal), and A/CNK is the peraluminosity index, corrected for Ca in apatite and calculated as Al_2O_3/(CaO - 10/3.P_2O_5 + Na_2O + K_2O) using molecular proportions.

Analytical Methods

Samples weighing at least 2 kg were split, crushed for 20 seconds in a tungsten carbide swing mill, screened to <200 µm and then washed in water to remove fines. Zircon was separated from other minerals by standard heavy liquid and magnetic techniques, then hand-picked under a binocular microscope, mounted in epoxy and sectioned in half by polishing. Only elongated, near-euhedral zircons from the two charnockite samples were selected for mounting in epoxy because the age of igneous crystallization was the primary concern, rather than that of older protolith, assimilated material or younger metamorphic zircons. For the gneiss

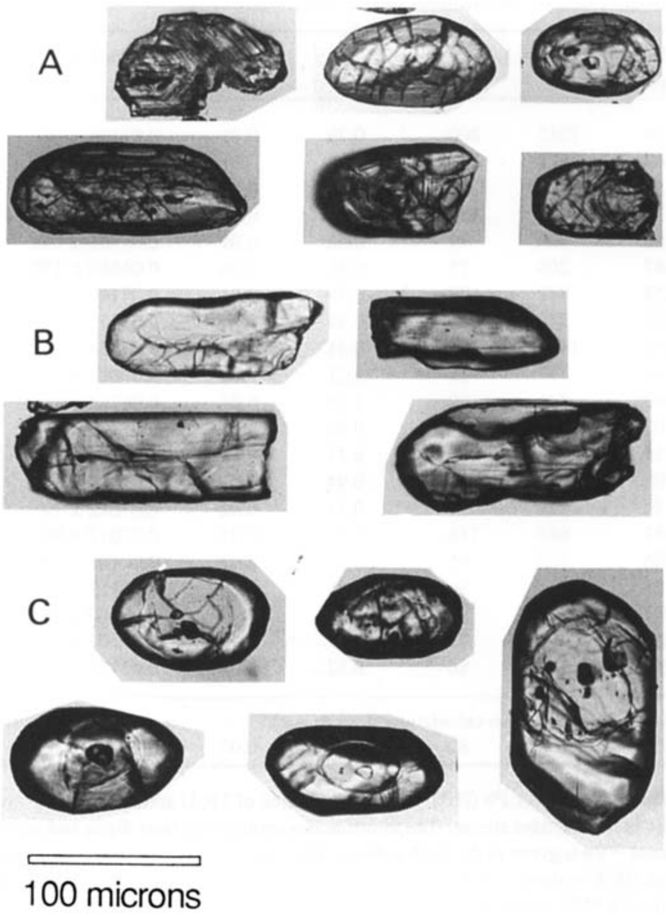

Fig. 3. Photomicrographs of typical zircons from the three samples analysed: (**a**) 6 zircons from sample z542, showing euhedral grains and zoning; (**b**) 4 zircons from sample z644, showing near-euhedral elongated grains, some with mineral inclusions; and (**c**) 5 zircons from sample z687, showing rounded zircons with cores of older zircon.

sample a selection of different zircon sizes and shapes was chosen. After analysis, the zircon mounts from all samples were etched in HF fumes for 15–30 seconds to accentuate zoning, thought to be typical of igneous zircon. Photographs of typical zircons are presented in Fig. 3.

Using the ANU ion-microprobe (SHRIMP), U, Pb and Th isotopic compositions were measured from positive secondary ions sputtered from a 30 micron spot by negative oxygen ions in a 10 kV primary beam. The ion-microprobe results are less precise than conventional U-Pb analyses but the statistical pooling of several analyses can give weighted mean ages with errors as low as 10 Ma (95% confidence) or less. Further details of the methods are available elsewhere (Compston *et al.* 1984, Kinny *et al.* 1988). A standard zircon was measured after every four analyses and used to calibrate the unknowns. No significant mass fractionation of secondary Pb+ ions from the standard was recognized, hence $^{207}Pb/^{206}Pb$ ages were determined directly from the observed results with a small correction for common Pb.

Common Pb is only a very small percentage of total Pb in the zircons of this study, but it was given careful consideration. The amount of common Pb was monitored using either the measured ^{204}Pb or ^{208}Pb and appeared to be contained within the zircons rather than being a surface contaminant. Plots of $^{204}Pb/^{206}Pb$ vs uncorrected $^{207}Pb/^{206}Pb$ were used to estimate common Pb composition by projecting a regression line from the data to the Model 111 growth curve of Cumming & Richards (1975) and using the common Pb ratios at the intersection. However, for all zircons in this study, a restricted range of $^{207}Pb/^{206}Pb$ made it difficult to estimate reliably the common Pb composition. For the zircon rims of z687, the line of best fit projected to the Cumming & Richards (1975) growth curve at the inferred age of the zircons (approx. 0.9 Ga), whereas for z542 it indicated old common Pb (3.0 Ga). Pb from sample z644 appeared to project to radiogenic initial Pb, but this was loosely constrained because of a lack of spread in the data, and common Pb of 1.0 Ga age was used. The cores of z687 were corrected assuming that common Pb was average crustal Pb (Cumming & Richards 1975) for the approximate original age of each zircon.

Analyses were corrected using the unradiogenic ^{204}Pb isotope where there appeared to be disturbance of Th, U and daughter Pb isotopes. Analyses were corrected for common Pb using ^{208}Pb if they plotted within error of the isochron of appropriate age on a $^{208}Pb/^{206}Pb$ vs. Th/U plot, if Th/U was less than 1, and if there were no other indications of a disturbed Th-U-Pb system. In calculating mean $^{207}Pb/^{206}Pb$

Table II. Isotopic data for zircons from sample z542.

Grain	Th ppm	U ppm	Pb* ppm	Th/U	Comm. Pb%	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$ age (Ma)	Conc. %
2.1	356	1243	200	0.29	0.01	0.07108 ± 54	0.1613 ± 18	1.581 ± 23	960 ± 16	100
3.1	260	678	108	0.38	0.01	0.07073 ± 80	0.1566 ± 18	1.527 ± 26	950 ± 23	99
4.1	422	1838	274	0.23	0.01	0.07158 ± 43	0.1519 ± 17	1.499 ± 20	974 ± 12	94
5.1	776	911	153	1.17	0.01	0.07166 ± 75	0.1604 ± 18	1.585 ± 29	976 ± 26	98
6.1	150	300	49	0.50	0.04	0.06961 ± 138	0.1540 ± 18	1.478 ± 36	917 ± 41	101
7.1	447	206	75	0.46	0.06	0.06960 ± 170	0.1628 ± 19	1.563 ± 44	917 ± 51	106
8.1	258	747	122	0.35	0.01	0.07142 ± 77	0.1615 ± 19	1.590 ± 27	969 ± 22	100
8.2	550	242	91	0.44	0.03	0.06875 ± 134	0.1614 ± 19	1.529 ± 36	891 ± 41	108
9.2	616	1498	245	0.44	0.01	0.07134 ± 55	0.1589 ± 18	1.563 ± 23	967 ± 16	98
11.1	90	177	31	0.51	0.05	0.07019 ± 193	0.1641 ± 20	1.588 ± 50	934 ± 57	105
12.1	493	541	90	1.10	0.04	0.06735 ± 168	0.1534 ± 18	1.424 ± 41	849 ± 53	108
13.1	402	488	91	0.82	0.01	0.07171 ± 130	0.1637 ± 19	1.619 ± 37	978 ± 38	100
14.1	514	209	77	0.41	0.01	0.07125 ± 111	0.1444 ± 17	1.418 ± 29	965 ± 32	90
15.1	908	964	186	0.94	0.01	0.07089 ± 100	0.1648 ± 19	1.610 ± 31	954 ± 29	103
16.1	54	514	79	0.11	0.02	0.07072 ± 79	0.1617 ± 19	1.577 ± 27	949 ± 23	102
17.1	361	680	118	0.53	0.03	0.07017 ± 89	0.16341± 19	1.581 ± 29	933 ± 26	105
18.1	346	522	95	0.66	0.02	0.07105 ± 106	0.1666 ± 19	1.632 ± 33	959 ± 31	104
19.1	123	344	56	0.36	0.03	0.07113 ± 118	0.1593 ± 19	1.562 ± 34	961 ± 34	99
20.1	1217	187	79	0.15	0.02	0.07100 ± 86	0.0669 ± 08	0.654 ± 12	957 ± 25	44
20.2	250	436	73	0.57	0.01	0.07101 ± 118	0.1568 ± 18	1.535 ± 33	958 ± 34	98
21.1	244	468	80	0.52	0.02	0.06771 ± 127	0.1622 ± 19	1.514 ± 35	860 ± 40	113
Additional data not included in calculation of mean age:										
10.1	456	103	80	0.23	0.03	0.08235 ± 138	0.1789 ± 21	2.031 ± 44	1254 ± 33	85

All ratios refer to radiogenic Pb (Pb*). The concentration of Th, U and radiogenic Pb is given in parts per million by weight. U is accurate to ca. 20% owing to variability in the standard zircon. The uncertainties apply to the later digits and are the standard error of the mean. The percentage of common Pb in the total measured ^{206}Pb is given in the sixth column. The final column gives the concordance of analyses on a standard $^{206}Pb/^{238}U$ - $^{207}Pb/^{235}U$ concordia plot (greater than 100% is above concordia). For z542 a ^{208}Pb correction is used for common Pb, except for eight samples (5.1, 7.1, 8.2, 10.1, 12.1, 14.1, 20.1, 21.1) for which a ^{204}Pb correction is used. For z644 and the zircon cores of z687 a ^{204}Pb correction is used for common Pb. For the rims and younger grains of z687 a ^{208}Pb correction is used for common Pb, except for five grains (1.1, 5.1, 12.2, 13.2, 40.1) which are ^{204}Pb corrected.

ages individual analyses were weighted as to the inverse of their variance, emphasising the most precise data. The observed standard error of the population will be no greater than the expected standard error multiplied by $\sqrt{2}$ if the zircons represent a single population

Results

Low-Ti charnockite (z542)

A ^{208}Pb correction was used for common Pb except for analyses which had either high Th/U or some disturbance of Th-U-Pb (see Appendix). Most zircon analyses plot on or just below concordia (Fig. 4) and 21 of the 22 analyses define a mean $^{207}Pb/^{206}Pb$ age of 954 ± 12 Ma (95% confidence level). Excluded is one older zircon which is 85% concordant with a minimum age of 1254 ± 33 Ma (Table II). It may be a xenocryst or restite from the source rock which melted to form the charnockite magma.

Two grains (12.1, 21.1; Fig. 4) plot above concordia when plotted using one standard error (Table II). However, their $^{207}Pb/^{206}Pb$ age is within error of the other zircons and the analyses are within two standard errors of concordia. The age of 954 ± 12 Ma is likely to be that of igneous crystallization because the grains are elongated and near-euhedral and commonly show euhedral zoning (Fig. 3).

High-Ti charnockite (z644)

A ^{204}Pb correction was used for sample z644 because Th/U is high and the U-Th-Pb system appears to have been disturbed slightly. If one discordant analysis and one reverse discordant analysis are excluded, the remaining 19 zircons define a population with a mean $^{207}Pb/^{206}Pb$ age (95% confidence) of 985 ± 29 Ma (Table III). This age cannot be considered as reliable as that of z542 because, on a concordia plot, there is more scatter about the mean age (Fig. 5). Most analyses plot near concordia, but there is a general correlation between concordance and apparent $^{207}Pb/^{206}Pb$ age, with reversely discordant zircons giving young apparent ages and those below concordia giving old apparent ages (Fig. 6). It is likely that either some random error is involved or that the system has been disturbed.

The result of 985 ± 29 Ma is interpreted as the igneous age because the zircons are elongated and nearly euhedral. There is only minor rounding, presumably caused by

Table III. Isotopic data for zircons from sample z644. (explanation as in Table II)

Grain	Th ppm	U ppm	Pb* ppm	Th/U	Comm. Pb%	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{208}Pb/^{206}Pb$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{206}Pb$ age (Ma)	conc. %
1.1	297	185	39	1.60	0.05	0.0726 ± 25	0.1561 ± 38	1.562 ± 69	0.5080 ± 70	0.0495 ± 14	1004 ± 70	93
3.1	268	206	44	1.30	0.02	0.0707 ± 20	0.1703 ± 41	1.659 ± 65	0.3961 ± 55	0.0518 ± 15	948 ± 57	107
3.2	263	196	39	1.34	0.05	0.0676 ± 22	0.1547 ± 38	1.443 ± 62	0.4152 ± 62	0.0478 ± 14	858 ± 70	108
4.1	175	160	33	1.09	0.04	0.0677 ± 27	0.1708 ± 42	1.595 ± 78	0.3206 ± 70	0.0502 ± 17	861 ± 85	118
5.1	254	247	48	1.03	0.02	0.0729 ± 21	0.1632 ± 40	1.640 ± 65	0.3109 ± 55	0.0494 ± 15	1011 ± 59	96
6.1	191	229	41	0.84	0.02	0.0757 ± 31	0.1570 ± 38	1.639 ± 82	0.2556 ± 78	0.0480 ± 19	1088 ± 83	86
7.1	1029	435	108	2.37	0.02	0.0705 ± 11	0.1599 ± 38	1.554 ± 46	0.7342 ± 40	0.0496 ± 12	943 ± 31	101
8.1	408	217	50	1.88	0.03	0.0732 ± 19	0.1605 ± 39	1.621 ± 61	0.5863 ± 59	0.0500 ± 13	1021 ± 54	94
9.1	188	214	38	0.88	0.07	0.0701 ± 21	0.1546 ± 38	1.494 ± 61	0.2725 ± 54	0.0480 ± 15	930 ± 63	100
10.1	623	277	67	2.25	0.04	0.0728 ± 19	0.1555 ± 38	1.562 ± 58	0.7218 ± 62	0.0500 ± 13	1009 ± 53	92
11.1	296	181	39	1.63	0.03	0.0772 ± 25	0.1582 ± 39	1.683 ± 72	0.5194 ± 72	0.0503 ± 14	1125 ± 65	84
12.1	161	237	37	0.68	0.04	0.0709 ± 23	0.1409 ± 34	1.378 ± 58	0.2170 ± 56	0.0451 ± 16	955 ± 67	89
13.1	143	112	22	1.28	0.08	0.0724 ± 46	0.1544 ± 39	1.542 ± 109	0.3990 ± 118	0.0482 ± 19	997 ± 134	93
14.1	351	194	42	1.81	0.04	0.0748 ± 27	0.1513 ± 37	1.560 ± 72	0.5634 ± 79	0.0472 ± 14	1063 ± 75	85
14.2	306	191	39	1.60	0.05	0.0736 ± 31	0.1508 ± 37	1.530 ± 78	0.4866 ± 84	0.0458 ± 14	1031 ± 86	88
15.1	219	144	31	1.52	0.04	0.0755 ± 29	0.1588 ± 39	1.653 ± 80	0.4832 ± 82	0.0504 ± 15	1082 ± 80	88
16.2	471	255	57	1.85	0.02	0.0730 ± 18	0.1579 ± 38	1.590 ± 58	0.5809 ± 57	0.0497 ± 13	1014 ± 50	93
17.1	165	199	35	0.83	0.03	0.0712 ± 20	0.1544 ± 38	1.515 ± 59	0.2564 ± 51	0.0497 ± 15	963 ± 57	96
18.1	572	267	62	2.14	0.04	0.0744 ± 21	0.1532 ± 37	1.571 ± 62	0.6781 ± 68	0.0485 ± 13	1052 ± 58	87
Additional data omitted from calculation of mean age:												
2.1	512	227	56	2.26	0.01	0.0777 ± 13	0.1599 ± 39	1.712 ± 53	0.7090 ± 51	0.0502 ± 13	1139 ± 33	84
16.1	187	190	35	0.98	0.05	0.0664 ± 23	0.1545 ± 38	1.414 ± 63	0.2982 ± 59	0.0469 ± 15	817 ± 73	113

deformation and/or metamorphism. Another igneous feature is that of many mineral inclusions (mainly biotite, feldspar and quartz) in these zircons. Igneous zoning is absent even after etching in HF, but this may be due to the low U contents. Euhedral zoning patterns are typical of high-U igneous zircons where there is an alternation of high-U and low-U zones (e.g. Kinny 1988).

Orthogneiss xenolith (z687)

The gneissic xenolith (z687) has a great diversity of zircons. Overgrowths of zircon on massive cores are present in many grains. Both the cores and rims are rounded, which probably indicates that they grew and/or partially dissolved during a metamorphic or deformational event (Fig. 3). The rounding of the zircon cores might be due to either detrital or metamorphic processes, although no pitting typical of sedimentary transport can be seen. The lack of igneous zoning is not strong evidence against an igneous origin because U is low and any zoning present might not be visible. On the basis of their rounded shape the zircon rims are thought to be of metamorphic origin.

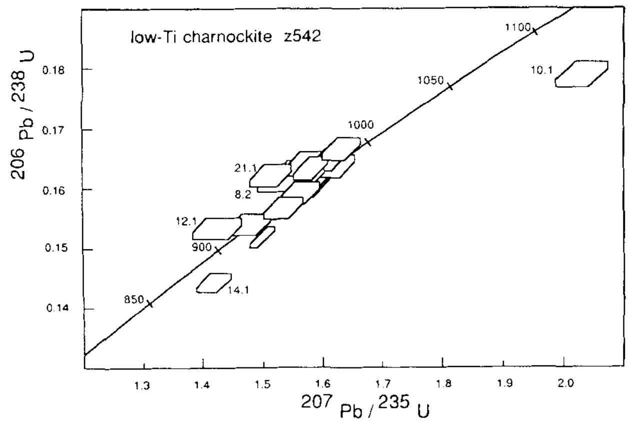

Fig. 4. Concordia plot of zircon isotopic results for low-Ti charnockite z542. Error boxes for this and other concordia plots are one standard error.

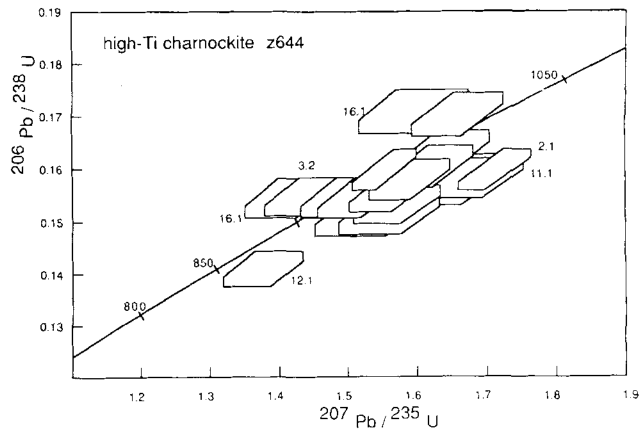

Fig. 5. Concordia plot of zircon isotopic results for high-Ti charnockite z644.

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Table IV Isotopic data for zircon cores from sample z687 (explanation as in Table II).

Grain	Th ppm	U ppm	Pb* ppm	Th/U	Comm. Pb%	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{208}Pb/^{206}Pb$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{206}Pb$. age (Ma)	Conc. %
1.2	43	87	29	0.50	0.06	0.1184 ± 19	0.3109 ± 62	5.08 ± 0.14	0.1477 ± 42	0.0916 ± 33	1932 ± 29	90
2.1	76	195	74	0.39	0.01	0.1231 ± 08	0.3533 ± 69	6.00 ± 0.13	0.1213 ± 16	0.1096 ± 27	2002 ± 12	97
4.1	164	192	67	0.85	0.01	0.1090 ± 09	0.3018 ± 59	4.54 ± 0.10	0.2428 ± 22	0.0860 ± 19	1783 ± 15	95
4.3	351	358	119	0.98	0.01	0.1038 ± 07	0.2752 ± 55	3.94 ± 0.09	0.2948 ± 19	0.0827 ± 18	1693 ± 13	93
5.2	64	260	101	0.25	0.01	0.1195 ± 06	0.3789 ± 73	6.25 ± 0.13	0.0700 ± 12	0.1077 ± 29	1949 ± 10	106
6.2	81	183	34	0.45	0.04	0.0811 ± 17	0.1817 ± 35	2.03 ± 0.06	0.1262 ± 39	0.0516 ± 19	1224 ± 42	88
7.1	108	195	36	0.55	0.04	0.0833 ± 15	0.1728 ± 33	1.99 ± 0.06	0.1545 ± 35	0.0483 ± 15	1277 ± 36	80
8.2	114	322	46	0.35	0.02	0.0751 ± 11	0.1413 ± 27	1.46 ± 0.04	0.1057 ± 25	0.0424 ± 13	1072 ± 29	80
9.2	62	142	43	0.44	0.04	0.0957 ± 14	0.2906 ± 57	3.83 ± 0.10	0.1181 ± 30	0.0780 ± 26	1542 ± 27	107
10.1	58	111	37	0.53	0.03	0.1140 ± 15	0.3084 ± 61	4.85 ± 0.12	0.1460 ± 33	0.0857 ± 27	1864 ± 24	93
11.1	64	217	87	0.30	0.01	0.1246 ± 06	0.3823 ± 74	6.57 ± 0.14	0.0902 ± 11	0.1162 ± 28	2023 ± 09	103
11.3	76	222	89	0.34	0.01	0.1242 ± 07	0.3797 ± 76	6.50 ± 0.14	0.0996 ± 13	0.1099 ± 28	2018 ± 10	103
12.1	92	166	50	0.55	0.01	0.1012 ± 09	0.2760 ± 54	3.85 ± 0.09	0.1582 ± 21	0.0790 ± 19	1646 ± 17	95
13.1	43	87	33	0.49	0.03	0.1203 ± 15	0.3498 ± 70	5.80 ± 0.14	0.1434 ± 33	0.1015 ± 32	1961 ± 23	99
14.1	58	150	26	0.39	0.04	0.0781 ± 17	0.1698 ± 33	1.83 ± 0.06	0.1087 ± 39	0.0478 ± 20	1150 ± 44	88
15.1	67	161	50	0.42	0.02	0.1045 ± 10	0.2932 ± 57	4.23 ± 0.10	0.1209 ± 21	0.0848 ± 23	1706 ± 18	97
16.1	90	337	81	0.27	0.01	0.0952 ± 07	0.2382 ± 46	3.13 ± 0.07	0.0751 ± 14	0.0668 ± 19	1532 ± 14	90
16.2	115	351	68	0.33	0.01	0.0860 ± 07	0.1882 ± 36	2.23 ± 0.05	0.1011 ± 16	0.0583 ± 15	1339 ± 17	83
17.1	52	122	19	0.43	0.03	0.0778 ± 20	0.1535 ± 30	1.65 ± 0.06	0.1198 ± 46	0.0429 ± 19	1141 ± 52	81
18.1	41	86	22	0.48	0.06	0.0874 ± 21	0.2496 ± 50	3.01 ± 0.10	0.1144 ± 47	0.0599 ± 28	1369 ± 46	105
19.1	48	92	39	0.52	0.02	0.1153 ± 13	0.3990 ± 80	6.34 ± 0.16	0.1323 ± 28	0.1021 ± 32	1885 ± 21	115
19.2	33	221	52	0.15	0.02	0.1003 ± 11	0.2378 ± 48	3.29 ± 0.08	0.0500 ± 22	0.0804 ± 40	1630 ± 21	84
20.1	56	143	30	0.39	0.03	0.0871 ± 15	0.2031 ± 40	2.44 ± 0.07	0.1108 ± 33	0.0572 ± 21	1362 ± 33	88
21.1	61	112	52	0.54	0.02	0.1213 ± 11	0.4222 ± 84	7.06 ± 0.16	0.1573 ± 24	0.1222 ± 32	1975 ± 16	115
22.1	77	164	33	0.47	0.06	0.0941 ± 21	0.1897 ± 38	2.46 ± 0.08	0.1449 ± 47	0.0583 ± 23	1510 ± 42	74
23.1	92	197	34	0.47	0.11	0.0762 ± 28	0.1694 ± 34	1.78 ± 0.08	0.1055 ± 66	0.0381 ± 25	1100 ± 75	92
24.1	54	140	51	0.38	0.03	0.1075 ± 15	0.3511 ± 72	5.20 ± 0.14	0.1010 ± 33	0.0924 ± 37	1757 ± 27	110
25.1	43	91	33	0.47	0.03	0.1156 ± 16	0.3323 ± 69	5.30 ± 0.14	0.1455 ± 34	0.1025 ± 35	1889 ± 25	98
26.1	51	342	65	0.15	0.02	0.0880 ± 10	0.1925 ± 38	2.34 ± 0.06	0.0562 ± 20	0.0719 ± 30	1382 ± 22	82
27.1	206	317	117	0.65	0.01	0.1104 ± 06	0.3313 ± 66	5.04 ± 0.11	0.1885 ± 14	0.0962 ± 21	1806 ± 10	102
28.1	57	127	55	0.45	0.02	0.1203 ± 13	0.4026 ± 83	6.68 ± 0.16	0.1282 ± 27	0.1154 ± 37	1961 ± 20	111
29.1	85	411	60	0.21	0.02	0.0742 ± 11	0.1493 ± 30	1.53 ± 0.04	0.0547 ± 23	0.0396 ± 19	1047 ± 30	86
30.1	63	137	23	0.46	0.03	0.0845 ± 18	0.1608 ± 33	1.87 ± 0.06	0.1515 ± 42	0.0529 ± 19	1304 ± 42	74
31.1	95	180	67	0.53	0.02	0.1181 ± 11	0.3434 ± 69	5.59 ± 0.13	0.1523 ± 24	0.0993 ± 27	1927 ± 17	99
32.1	73	200	44	0.36	0.01	0.1002 ± 10	0.2117 ± 42	2.92 ± 0.07	0.1056 ± 21	0.0613 ± 18	1628 ± 19	76
33.1	57	140	55	0.40	0.02	0.1214 ± 13	0.3657 ± 75	6.12 ± 0.15	0.1173 ± 26	0.1062 ± 34	1977 ± 19	102
34.1	55	126	38	0.44	0.03	0.1119 ± 17	0.2790 ± 57	4.31 ± 0.11	0.1415 ± 36	0.0905 ± 31	1831 ± 27	87
35.1	44	90	30	0.49	0.06	0.1098 ± 22	0.3178 ± 66	4.81 ± 0.15	0.1269 ± 50	0.0821 ± 38	1796 ± 37	99
36.1	51	185	54	0.28	0.03	0.1124 ± 13	0.2858 ± 58	4.43 ± 0.11	0.0807 ± 26	0.0833 ± 33	1838 ± 21	88
37.1	88	264	54	0.33	0.01	0.0957 ± 09	0.1977 ± 39	2.61 ± 0.06	0.1052 ± 19	0.0625 ± 18	1542 ± 19	75
38.1	36	474	101	0.08	0.02	0.0842 ± 08	0.2247 ± 44	2.61 ± 0.06	0.0178 ± 15	0.0528 ± 46	1298 ± 18	101
41.1	51	99	33	0.52	0.03	0.1072 ± 18	0.3080 ± 64	4.55 ± 0.13	0.1409 ± 40	0.0836 ± 31	1752 ± 32	99
42.2	82	288	53	0.29	0.03	0.0803 ± 12	0.1844 ± 37	2.04 ± 0.05	0.0723 ± 25	0.0466 ± 19	1205 ± 29	91
43.1	62	202	45	0.31	0.03	0.0936 ± 13	0.2167 ± 43	2.79 ± 0.07	0.0930 ± 28	0.0660 ± 25	1499 ± 27	84
44.1	68	178	96	0.38	0.01	0.1604 ± 07	0.4910 ± 99	10.86 ± 0.23	0.1172 ± 12	0.1515 ± 37	2460 ± 08	105
45.1	74	399	65	0.18	0.01	0.0738 ± 08	0.1677 ± 33	1.71 ± 0.04	0.0557 ± 18	0.0506 ± 19	1036 ± 23	96
46.1	66	112	37	0.59	0.04	0.1043 ± 18	0.3004 ± 62	4.32 ± 0.12	0.1624 ± 40	0.0831 ± 28	1701 ± 32	100
47.1	98	468	81	0.21	0.02	0.0808 ± 09	0.1759 ± 35	1.96 ± 0.05	0.0583 ± 18	0.0491 ± 19	1217 ± 21	86

For the cores a ^{204}Pb correction for common Pb was used because there was a wide range of grain ages and evidence for a complex Pb loss history. The young zircons (rims on older cores, and young grains without older cores) were corrected for common Pb mainly using ^{208}Pb because of the agreement within error of the expected isochron on a plot of $^{208}Pb/^{206}Pb$ against Th/U. However, five zircons had slightly disturbed Th-U-Pb systems, indicated by this plot, and were corrected using ^{204}Pb.

The zircon cores have a large spread of $^{207}Pb/^{206}Pb$ ages, from 1036 to 2460 Ma (Table IV). Most analyses plot on or slightly below concordia, but a few plot above concordia, perhaps due to having gained radiogenic Pb during metamorphic disturbance (e.g. Williams *et al.* 1984). The data cluster mainly around 1700–2000 Ma, with a clear trend of Pb loss towards concordia at approximately 900–1000 Ma (Fig. 7a). Interpretation must be cautious, however, because other concordant zircons plot at 1300–1400 Ma and there

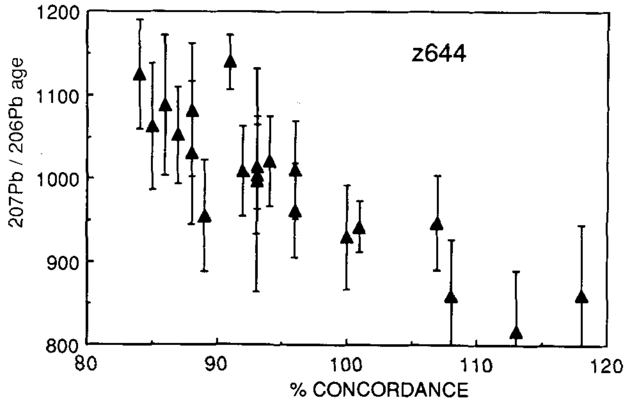

Fig. 6. Plot for high-Ti charnockite z644 showing percentage concordance of zircon analyses versus $^{207}Pb/^{206}Pb$ age (bars represent only one standard error). Note that analyses with greater than 100% concordance plot above concordia and those with less than 100% plot below concordia.

may have been additional Pb loss in recent times. Three of the old zircon cores (4.1, 4.3, 27.1; see Table IV) have relatively high Th/U, but most are lower in Th/U than the zircon rims.

There is no obvious cluster of zircon ages to indicate precisely the igneous age of the granite or volcanic rock which was later deformed and metamorphosed into orthogneiss. The spread of apparent ages could be explained by some zircons coming from older rocks, which melted to form the granite, and some forming during later metamorphism. It is possible that metamorphic disturbance of the cores took place at around 1000 Ma because there is a trend of Pb-loss discordia towards this age, and some zircon cores are near-concordant at approximately 1000 Ma. However, the data are spread sufficiently that it is not possible to resolve between this possibility and that of disturbance at the younger age of 921 ± 19 Ma, obtained for D_3 (see below).

The zircon rims and grains without older cores define a single population with a mean $^{207}Pb/^{206}Pb$ age of 921 ± 19 Ma (95% confidence) if one grain is excluded (Fig. 7b, Table V). The old grain (5.1) formed before the bulk of the population, perhaps during D_1. The age given by the rims is too young to be that of D_1 or of original igneous crystallization of the orthogneiss, because it is younger than that of the post-D_2 charnockites. It is likely that the zircons grew during D_3 deformation and accompanying metamorphism, because of their rounded shape and lack of euhedral zoning. The 921 ± 19 Ma date from zircon rims agrees within 95% confidence with the 959 ± 59 Ma Rb-Sr isochron result of P.A. Arriens (unpublished data) and also with the 886 ± 48 Ma Rb-Sr isochron age of Black *et al.* (1987).

Discussion

This study provides ages for charnockitic plutonism in the Mawson Coast area of 985 ± 29 and 954 ± 12 Ma. The

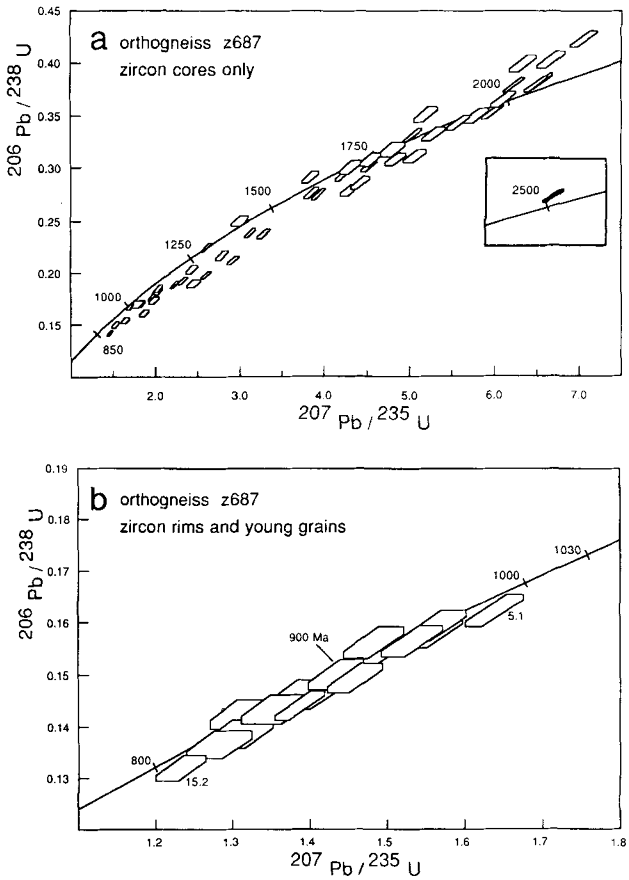

Fig. 7. Concordia plot of zircon isotopic results for xenolith of basement gneiss, z687, showing (**a**) cores, and (**b**) rims (overgrowths) on older cores and young grains without cores.

difference between the two ages is not significant at the 95% confidence level. These igneous ages contrast with younger Rb-Sr isochron ages from previous studies, implying that the younger rages represent resetting during metamorphism. The third deformation (D_3) produced a penetrative foliation in the charnockites and was probably responsible for this resetting of Rb-Sr systems over a scale of at least one to two metres. Our best estimate of the age of D_3, obtained from zircon overgrowths in a gneissic xenolith, is 921 ± 19 Ma, compared to an estimate of 963 Ma (Black *et al.* 1987) which combined zircon U-Pb ages from an area of some hundreds of square kilometres. It is possible that some of the individual results of Black *et al.* (1987) could be equivalent to D_2 at Mawson.

A previously reported age of 1061 ± 36 Ma for the Mawson Station charnockite is believed to be erroneous because non-cogenetic samples were grouped together to give effectively a 3-point isochron. The cogenetic samples are reinterpreted here to give an age of 959 ± 58 Ma which is within error of a metamorphically reset age of 886 ± 48 Ma (Black *et al.* 1987). A conventional U-Pb zircon age of 935 Ma was obtained by Black *et al.* (1987) but their analyses were closely grouped and discordant and provide a minimum

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Table V. Isotopic data for zircon rims and young grains from z687 (explanation as in Table II).

Grain	Th ppm	U ppm	Pb* ppm	Th/U	Comm. Pb%	^{207}Pb.^{206}Pb	^{206}Pb/^{238}U	^{207}Pb/^{235}U	^{207}Pb/^{206}Pb age (Ma)	Conc. %
1.1	181	342	58	0.53	0.02	0.07107 ± 101	0.1595 ± 31	1.563 ± 39	959 ± 29	99
2.2	185	250	44	0.74	0.02	0.07131 ± 116	0.1583 ± 31	1.556 ± 42	966 ± 34	98
3.1	102	106	18	0.96	0.03	0.06748 ± 221	0.1423 ± 28	1.324 ± 53	853 ± 70	101
4.2	235	312	52	0.75	0.01	0.06934 ± 101	0.1500 ± 29	1.434 ± 37	909 ± 30	99
6.1	168	177	30	0.95	0.02	0.06862 ± 165	0.1433 ± 28	1.356 ± 44	887 ± 51	97
8.1	123	144	23	0.86	0.03	0.06845 ± 175	0.1386 ± 27	1.308 ± 44	882 ± 54	95
9.1	352	355	65	0.99	0.01	0.06986 ± 115	0.1547 ± 30	1.491 ± 40	924 ± 34	100
11.2	141	153	26	0.93	0.02	0.06914 ± 177	0.1461 ± 28	1.393 ± 47	903 ± 54	97
12.2	227	340	58	0.67	0.03	0.07099 ± 98	0.1565 ± 31	1.532 ± 39	957 ± 29	98
13.2	277	312	54	0.89	0.01	0.07084 ± 97	0.1494 ± 29	1.459 ± 36	953 ± 28	94
15.2	297	360	54	0.82	0.01	0.06771 ± 109	0.1320 ± 25	1.232 ± 33	860 ± 34	93
22.2	358	423	75	0.85	0.02	0.06886 ± 104	0.1562 ± 31	1.483 ± 39	894 ± 31	105
39.1	63	414	57	0.15	0.01	0.06980 ± 66	0.1441 ± 28	1.387 ± 32	922 ± 20	94
40.1	145	215	32	0.67	0.03	0.06808 ± 165	0.1366 ± 27	1.282 ± 43	871 ± 51	95
Additional data not used in calculation of mean age:										
5.1	255	326	60	0.78	0.02	0.07317 ± 71	0.1624 ± 31	1.639 ± 37	1019 ± 20	95

age only.

The age of the basement gneiss, as represented by zircons from a single sample of xenolithic orthogneiss, ranges from at least 2.0 Ga to 1.3 Ga with only one zircon of late Archaean age. This is consistent with the model of Black *et al.* (1987) that this mobile belt, known further west from here as the Rayner Complex, is derived mainly from Proterozoic material. Reworking of Archaean gneisses is recognized elsewhere in the mobile belt (Sheraton *et al.* 1980, 1987, Clarke 1988) but this is probably not the case at the Mawson Coast.

The orthogneiss sample developed extensive zircon overgrowth at 921 ± 19 Ma, but there is no well-defined zircon population to indicate the primary age of the rock. Because of the spread in ages it is possible that some or all of the zircon cores are sedimentary, but an igneous origin, followed by multiple disturbance and Pb loss, cannot be discounted. There appears to have been loss of Pb from zircon cores in this orthogneiss at approximately 900–1000 Ma, and possibly also during recent times. There are many discordant analyses of zircon cores, mainly between 1.0 Ga and 2.0 Ga. Some grains appear to have gained radiogenic Pb during disturbance and plot above concordia.

The age of D_1, the major isoclinal folding event in the region, is not apparent from the study of this gneiss despite its strong S_1 fabric. It is likely that zircon grew during the lower-grade and less intense D_3 event rather than during D_1. D_1 and D_2 must both be older than the charnockites dated at 985 ± 29 and 954 ± 12 Ma.

Thick crust at the time of magma generation has been invoked (Ellis 1987, Young & Ellis 1990, 1991) to explain the REE geochemistry of some charnockites. No high-Ti rocks in this study are HREE-depleted, whereas some of the low-Ti rocks (including z542, the sample dated) are depleted in HREE and require a higher pressure of generation from within thickened crust (Young & Ellis 1991). HREE-depleted I-type granitoids require garnet to be stable in the residue of partial melting in order to explain the REE pattern, and this is only possible at high pressures (>15 kb). This suggests that crust of normal thickness could not have melted to produce magmas of this distinctive composition. If the inferred thick crust was produced by compressional forces during D_1 isoclinal folding, then the crust may have remained thick until the post-D_2 generation of Mawson charnockites by partial melting of crustal material, the genetic model favoured by Young & Ellis (1991).

However, an alternative model (Sandiford 1989, Harley 1989) is that recumbent structures in granulite terranes can be produced by extensional thinning of already overthickened crust. If the recumbent D_1 structures at Mawson (Clarke 1988) were produced by extensional collapse, then the crust would presumably have been of normal thickness after D_1. If this were the case, the REE geochemistry of some charnockites would require that crustal melting took place during or before D_1. We might then expect that charnockite intrusion (as opposed to initial magma generation) took place reasonably soon after D_1 and D_2. In such a case, D_1 may have taken place shortly before 985 ± 29 Ma, the age of a HREE-depleted low-Ti charnockite. Further geochronological study at Mawson should again focus on the ages of D_1 and D_2 and whether the two types of charnockite have resolvably different ages.

Conclusions

The ages of charnockites determined in this study, and interpreted as igneous emplacement ages, are 985 ± 29 Ma for a low-Ti charnockite and 954 ± 12 Ma for a high-Ti charnockite. A xenolithic sample of gneiss provides evidence for a predominantly Proterozoic rather than Archean basement age. The gneiss has zircon core ages from 1.0–2.5 Ga, but most are clustered around 1.7–2.0 Ga. Pb-loss from the zircons appears from loosely-constrained discordia to have taken place at around 1.0 Ga and in recent times, but may well be more complex than this. The first two deformations were not periods of zircon growth in this particular sample and hence have not been dated. D_3, the deformation that produced a gneissic foliation in all the charnockites, is believed to have taken place at 921 ± 19 Ma on the basis of U-Pb analysis of zircon overgrowths in the gneissic sample. This age is consistent with previously obtained Rb-Sr isochron ages (910 ± 18 Ma, 959 ± 58 Ma and 886 ± 48 Ma) from charnockites that probably had their Rb-Sr systems reset during D_3.

Acknowledgements

The Australian Antarctic Division and the expeditioners at Mawson are thanked for their invaluable support. Prof. W. Compston kindly made available analytical facilities for this project. The ion-probe team at Canberra, including Ian Williams, John Foster, Peter Kinny, Liz Jagodzinski, Mike Wiedenbeck and Rod Page provided helpful assistance and discussion. David Ellis gave constructive criticism of the manuscript and, along with Prof. D.H. Green from the University of Tasmania, helped plan and support fieldwork for D.Y. Bob Tingey made available unpublished Rb-Sr data, originally obtained in the 1970's by P.A. Arriens for the Australian BMR. Chris Gray assisted with comments and regression programs. We thank Simon Harley, Bob Morrison, Alfredo Camacho and in particular John Sheraton and Bob Pankhurst for their comments and/or reviews. This paper is published with the permission of the Director of the Bureau of Mineral Resources.

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