

Nature and Origin of A-Type Granites with Particular Reference to Southeastern Australia

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Abstract. In the Lachlan Fold Belt of southeastern Australia, Upper Devonian A-type granite suites were emplaced after the Lower Devonian I-type granites of the Bega Batholith. Individual plutons of two A-type suites are homogeneous and the granites are characterized by late interstitial annite. Chemically they are distinguished from I-type granites with similar SiO₂ contents of the Bega Batholith, by higher abundances of large highly charged cations such as Nb, Ga, Y, and the REE and lower Al, Mg and Ca: high Ga/Al is diagnostic. These A-type suites are metaluminous, but peralkaline and peraluminous A-type granites also occur in Australia and elsewhere.

Partial melting of felsic granulite is the preferred genetic model. This source rock is the residue remaining in the lower crust after production of a previous granite. High temperature, vapour-absent melting of the granulitic source generates a low viscosity, relatively anhydrous melt containing F and possibly Cl. The framework structure of this melt is considerably distorted by the presence of these dissolved halides allowing the large highly charged cations to form stable high co-ordination structures. The high concentration of Zr and probably other elements such as the REE in peralkaline or near peralkaline A-type melts is a result of the counter ion effect where excess alkali cations stabilize structures in the melt such as alkali-zircono-silicates. The melt structure determines the trace element composition of the granite.

Separation of a fluid phase from an A-type magma results in destabilization of co-ordination complexes and in the formation of rare-metal deposits commonly associated with fluorite. At this stage the role of Cl in metal transport is considered more important than F.

Introduction

Granites of the southeastern part of the Lachlan Fold Belt of eastern Australia (Fig. 1) may be grouped into suites with characteristic chemical, petrographic and field characteristics (Hine et al. 1978; Griffin et al. 1978; White and Chappell 1982). Suites may consist of one pluton as does the Wullwey Granodiorite of the Berridale Batholith, or as many as fourteen separate mappable units as in the case of the Glenbog Suite of the Bega Batholith (Beams 1980).

If enough elements are precisely determined, it can be shown that the set of rocks belonging to any one suite cannot be derived from those of another suite by processes such as fractional crystallization or contamination and hence it has been postulated that different source rocks are the cause of variation from one suite to another.

Source compositions can be modelled for those suites that show a large linear dispersion in composition when rock compositions are plotted onto Harker variation diagrams (White and Chappell 1977). For suites from the Berridale, Kosciusko and Moruya Batholiths, Chappell and White (1974), White and Chappell (1982) and elsewhere, showed that the first-order subdivision of granitic rocks according to source composition was into those derived from sedimentary sources and those from igneous sources. Chappell and White (1974) called these S-types and I-types.

Study of a wider area (Fig. 1) has led to the recognition of four suites of another distinct granite type in the southeastern Australian province. These are the Gabo, Mumbulla, Monga and Wangrah Suites. The first two of these are here briefly described. Granites with similar characteristics have been described from elsewhere in the world where, like the southeastern Australian examples, they are last in the sequence of granitic intrusions. They have been called A-types by Loiselle and Wones (1979) because they are somewhat alkaline, anorogenic and anhydrous. A genetic model based on our new data as well as data from elsewhere is here proposed for the A-type granite suites.

Gabo Suite

Six homogeneous granite plutons (Watergums (28 km²), Naghi (3.5 km²), Nagha (5.3 km²), Howe Range (14 km²), Gabo Island (1.6 km²) and Carlyle (4.6 km²) with a total areal extent of 57 km² (Fig. 1) constitute the Gabo Suite. These granites intrude Upper Devonian rhyolites of the Boyd Volcanic Complex (Collins 1977; Fergusson et al. 1979), as well as folded Ordovician metasediments. This complex is bimodal in that it also contains transitional type basalts; it is thought to occupy a graben (McIlveen 1975). The rhyolites are considered to be comagmatic with the granites of the Gabo Suite and both are overlain unconformably by Upper Devonian sediments, mostly terrestrial in nature. The Gabo Suite granites are therefore Upper Devonian and distinctly later than the Lower Devonian granites of the adjacent Bega Batholith (Beams et al. 1982). Petrographic features indicative of near surface intrusion, and

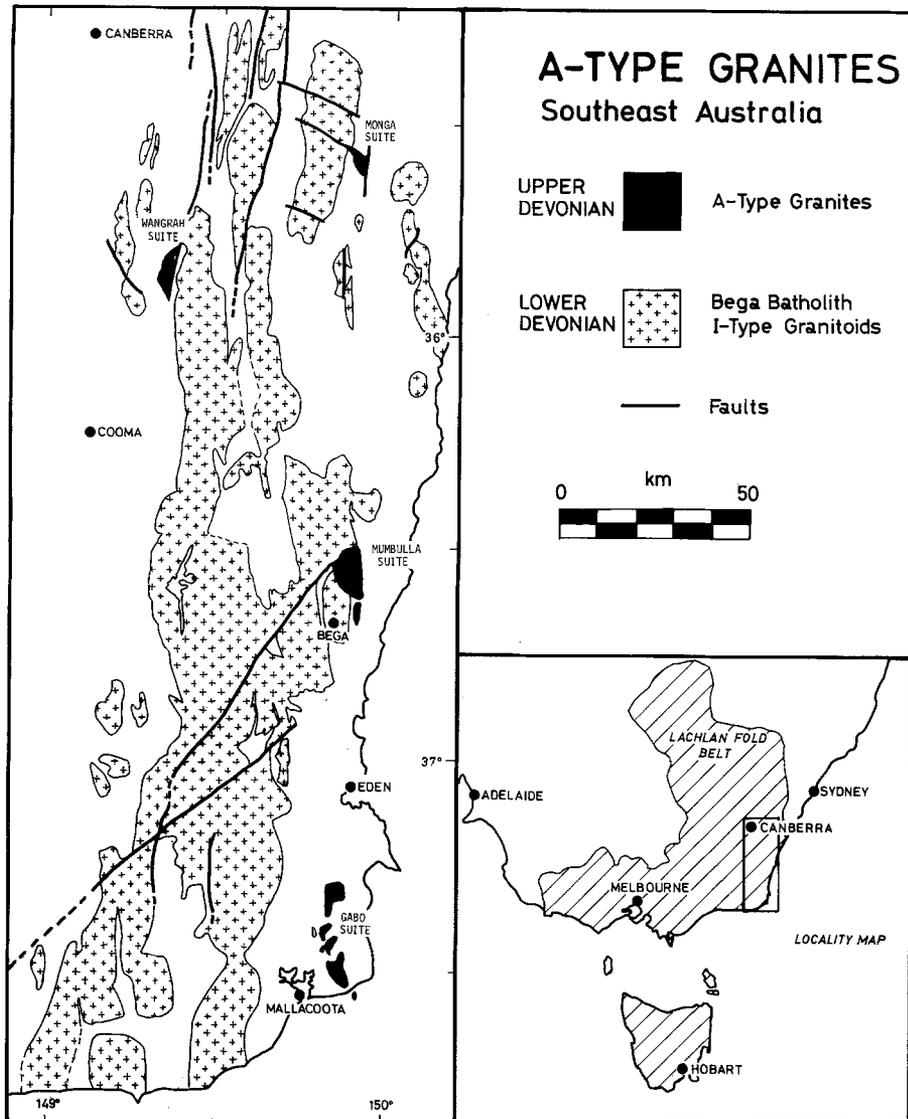


Fig. 1. Locality map showing the occurrence of A-type granites in the eastern part of the Lachlan Fold Belt, southeastern Australia

hence supporting the temporal relationship with the rhyolites, are the presence of miarolitic cavities, granophyric intergrowths and embayed quartz phenocrysts. The narrow contact aureoles (10–20 m wide) within the Ordovician country rocks are also characteristic of subvolcanic granites.

All granites of the suite are even, medium to fine grained, felsic rocks ranging from granite (*sensu stricto*) to adamellite using Streckeisen's classification (Table 1 and Fig. 2). Specimens range in colour from grey to brick red depending upon the degree of alteration. K-feldspar ($Or_{60}Ab_{40}$) is the most abundant mineral, forming tabular microperthitic crystals or granophyric intergrowths with quartz. It is accompanied by normally zoned plagioclase ($An_{30}-An_5$) with a composition mostly in the oligoclase range. Quartz also appears as clusters of small grains between K-feldspar or as large embayed crystals having the β -quartz form. Biotite (α = pale yellow, $\beta = \gamma$ = dark brown) is mostly interstitial to K-feldspar but may form subhedral grains within the granophyric parts of the rock. Microprobe analysis shows that it is annite. Fluorite sometimes occurs as small lenses interleaved in biotite. Hastingsite (α = colour-

less, β = light brown, γ = khaki) is subordinate to biotite but crystallized before it and forms long prisms. In places it is interstitial to K-feldspar. It typically contains inclusions of zircon, apatite and opaque minerals. A distinctly blue amphibole (α = aqua-blue, γ = pale to inky-blue) occurs as small irregular grains around the hastingsites in granites of the Watergums intrusion; microprobe analysis indicates that these are riebeckite-arfvedsonite solid solutions. Zircon, fluorite, and magnetite (with exsolved ilmenite) are constant accessory minerals whereas apatite and allanite are less common. Relicts of fayalite are rare.

Mumbulla Suite

This suite comprises two plutons, the Mumbulla (outcrop area 56 km²) and Dr George (13 km²) both of which intrude granites of the Bega Batholith as well as Ordovician metasediments in the region northeast of Bega (Fig. 1).

Rocks of the Mumbulla pluton are medium to coarse, even-grained granites (*sensu stricto*) dominated by pink K-feldspar and quartz and with lesser amounts of plagioclase that is greenish because of alteration. Biotite (α = pale yel-

Table 1. Modes and locality data. Grid references are for the Australian Metric Grid, 1:100,000 sheets: Bega 8824, Eden 8823, Mallacoota 8822. Sample numbers refer to the La Trobe University catalogue

Mumbulla suite									
Pluton	Mumbulla						Dr. George		
Sample	4473	4474	4475	4476	4477	4478	4479	4480	
Map sheet	8824	8824	8824	8824	8824	8824	8824	8824	8824
Grid references	571500	568497	566486	588468	564517	612464	597382	610431	
Quartz	34.7	32.2	34.8	35.6	34.7	34.2	39.4	31.0	
K-feldspar	46.4	47.7	45.0	43.2	46.2	47.3	46.2	51.3	
Plagioclase	15.4	15.9	16.6	16.9	14.1	15.5	12.8	14.3	
Biotite	3.0	4.0	3.3	4.2	4.7	2.7	1.6	3.1	
Opaques	0.3	0.2	0.2	0.1	0.2	0.2	<0.1	0.2	
Apatite	<0.1	<0.1	0.1	<0.1	0.1	0.1	<0.1	0.1	
Allanite	0.2	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	<0.1	
Gabo suite									
Pluton	Watergums			Naghi	Nagha	Howe Range			Gabo Island
Sample	4464	4465	4466	4467	4468	4469	4470	4471	4472
Map sheet	8823	8823	8823	8823	8823	8823	8823	8823	8822
Grid reference	554666	532648	513684	496564	534532	532520	530476	533460	567396
Quartz	27.6	31.4	31.3	31.6	27.0	30.3	25.9	28.3	36.0
K-feldspar	43.0	38.8	44.7	37.8	46.7	41.1	36.7	43.6	39.0
Plagioclase	24.4	24.4	21.3	25.3	20.0	24.0	31.2	24.8	20.0
Biotite	3.8	2.7	1.5	4.1	3.6	2.7	2.7	2.6	4.0
Amphibole		1.3	0.7	<0.1	0.9	<0.1	2.2	0.2	0.2
Opaques	0.8	0.7	0.2	1.1	1.6	1.7	0.4	0.5	0.5
Sphene	0.1								
Apatite	<0.1	<0.1	<0.1	<0.1	0.1	0.1	<0.1	<0.1	<0.1
Fayalite	0.1	0.1	0.1		0.1	<0.1	0.2		0.1
Fluorite	0.2	<0.1	<0.1	0.1	<0.1	0.1	<0.1	<0.1	0.1
Stilpnomelane	<0.1	0.6	0.2	<0.1	<0.1	<0.1	0.7	<0.1	0.1

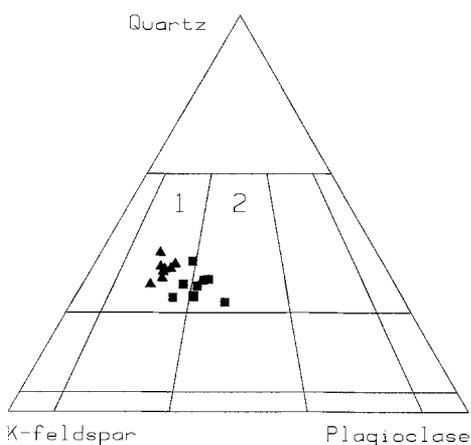


Fig. 2. Plot of modal compositions of quartz, K-feldspar and plagioclase from the Gabo Suite (squares) and the Mumbulla Suite (triangles). The classification used is modified from that proposed by the IUGS Subcommittee on the systematics of Igneous Rocks (Streckeisen 1973). Field 1 is granite, 2 is adamellite: the lower limit for the quartz content of adamellite and granite is taken as 25% of the total quartz + feldspar

low, $\beta = \gamma$ = dark brown) is the sole primary mafic mineral and is always less than 5% of the mode; it forms irregular-shaped crystals which are commonly interstitial to quartz and K-feldspar. Chlorite is a common secondary alteration

product of biotite whereas red-brown alteration products presumably after fayalite are rare. Accessory minerals include magnetite, fluorite, zircon, apatite and allanite.

The Dr. George Granite is finer grained and has abundant miarolitic cavities, but is otherwise texturally and mineralogically similar to the Mumbulla Granite. Both have granophyric intergrowths of K-feldspar and quartz, like the Gabo Suite rocks. Dr. George Granite shows much more deuteric alteration than the Mumbulla rocks.

Chemistry of the Gabo and Mumbulla Suites

Chemical analyses (Tables 2 and 3) show that all of the A-type granites in these two suites have high $(\text{Na}_2\text{O} + \text{K}_2\text{O})/\text{Al}_2\text{O}_3$; however none are peralkaline. The Gabo Suite is distinctly more mafic than the very felsic Mumbulla Suite. Both are fairly homogeneous in composition. For the Gabo Suite this is important in terms of its occurrence as several discrete plutons spread laterally over a distance of some 30 km. In the case of the Mumbulla Suite, it is of interest because of the significant area of that pluton (56 km^2) with the analyzed samples coming from a vertical interval of 450 m, implying a large and uniform volume of rock.

It is instructive to compare the two A-type suites with I-type rocks of similar SiO_2 -content from the Bega Batholith: there are no S-types in this region and hence these are not considered. In Table 4 we list average compositions

Table 2. Analyses of Mumbulla suite granites

Sample	4473	4474	4475	4476	4477	4478	4479	4480
SiO ₂	77.00	76.84	77.51	77.77	77.14	77.12	77.00	77.33
TiO ₂	0.15	0.14	0.14	0.12	0.13	0.12	0.12	0.12
Al ₂ O ₃	11.83	11.76	11.76	11.49	11.91	11.79	11.78	11.99
Fe ₂ O ₃	0.40	0.38	0.27	0.34	0.32	0.40	0.14	0.65
FeO	1.05	0.95	1.04	0.89	0.77	0.64	1.14	0.34
MnO	0.04	0.03	0.03	0.03	0.03	0.02	0.05	0.02
MgO	0.04	0.03	0.05	0.06	0.06	0.04	<0.03	0.04
CaO	0.61	0.56	0.34	0.38	0.50	0.34	0.11	0.24
Na ₂ O	3.06	3.06	3.04	3.01	3.14	3.07	3.10	3.15
K ₂ O	4.98	4.86	4.94	4.87	4.91	5.12	5.16	5.19
P ₂ O ₅	0.02	0.02	0.02	0.02	0.01	0.02	0.01	0.02
S	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02
H ₂ O ⁺	0.47	0.58	0.59	0.42	0.70	0.85	0.38	0.50
H ₂ O ⁻	0.18	0.18	0.17	0.18	0.21	0.26	0.13	0.20
CO ₂	0.07	0.20	0.18	0.12	0.12	0.14	0.69	0.10
rest	0.20	0.21	0.20	0.19	0.19	0.19	0.18	0.20
	100.10	99.80	100.28	99.89	100.14	100.12	99.99	100.09
Trace elements (ppm)								
Ba	655	645	640	555	575	520	510	500
Cs	10.7	9.0	10.3	10.6	11.8	7.7	3.7	6.1
Rb	230	229	232	242	255	248	240	261
Sr	50	53	43.5	43.0	45.0	37.5	31.0	37.5
Pb	33	36	37	35	36	48	32	38
Th	21.8	25.9	26.8	24.0	25.2	25.7	25.1	26.9
U	5.1	5.5	5.6	5.4	5.8	5.6	4.7	5.3
Zr	187	176	173	159	169	171	165	163
Hf	6.3	6.4	6.8	5.8	6.6	6.1	6.2	6.4
Nb	18.0	18.5	20.0	17.0	18.5	19.0	19.0	21.0
Ta	3.1	2.8	2.9	2.9	3.2	3.3	3.2	3.2
Y	100	87	84	79	77	94	100	99
La	59	70	59	55	51	61	54	95
Ce	132	150	134	124	120	130	119	196
Nd	66	71	65	61	58	63	58	93
Sm	16.4	16.7	15.2	14.9	14.1	15.6	14.4	21.1
Eu	1.61	1.68	1.57	1.40	1.43	1.29	1.21	1.59
Gd	15.3	14.0	14.6	13.8	11.9	13.4	14.5	18.7
Ho	4.2	3.4	3.3	3.2	3.0	3.6	3.7	3.8
Tm	1.4	1.1	1.2	1.2	1.1	1.2	1.3	1.3
Yb	10.6	9.1	8.8	8.7	8.6	9.1	9.7	9.3
Lu	1.57	1.39	1.35	1.33	1.31	1.38	1.49	1.41
Sc	12.2	13.1	12.8	13.5	12.1	12.5	12.5	13.1
V	2	3	2	2	2	2	2	2
Cr	<1	<1	<1	<1	<1	<1	<1	<1
Co	1	1	1	1	1	2	2	1
Ni	<1	<1	<1	1	<1	1	1	1
Cu	8	9	5	9	12	17	7	8
Zn	106	112	116	135	101	151	138	118
Ga	20.0	20.0	20.0	19.8	20.4	20.2	20.6	20.6

of the Gabo and Mumbulla Suites, along with averages of I-type compositions from the Bega Batholith. For the latter, we present an average of 48 analyses encompassing the same SiO₂ range as the Gabo Suite analyses (71.46–73.71%). There is only one I-type analysis which has SiO₂ as high as Mumbulla so we have averaged the 20 analyses of Bega Batholith rocks containing more than 75% SiO₂. Compared with the Bega Batholith, the Gabo Suite granites have lower Al₂O₃, MgO and CaO and higher total FeO. Na₂O + K₂O is slightly higher. It is these differences in chemical composition that are reflected by the presence of annite, hastingsite and the lower An content of the plagioclase in the Gabo Suite rocks.

Some trace elements¹, e.g. Rb, Sr, Th and U have comparable abundances in the Gabo Suite and the I-type granites. V, Cr, Co and Ni are less abundant whereas Sc and Zn (Fig. 3) are much more abundant than in I-type granites of comparable SiO₂ content. Ba and Pb are more abundant but this is less marked. However, higher contents of a chemically coherent group of highly-charged cations in the Gabo Suite is most distinctive; these include Zr, Nb, Y, the rare-earth elements (REE) and Ga (Fig. 3). The combination

¹ Ba, Rb, Sr, Pb, Zr, Nb, Y, V, Cr, Co, Ni, Cu, Zn and Ga measured by x-ray spectrometry, and Cs, Th, U, Hf, Ta, REE and Sc by instrumental neutron activation analysis

Table 3. Analyses of Gabo Suite granites

Sample	4464	4465	4466	4467	4468	4469	4470	4471	4472
SiO ₂	72.06	73.34	73.60	73.62	73.13	73.71	71.46	73.34	73.06
TiO ₂	0.38	0.36	0.34	0.38	0.39	0.35	0.42	0.37	0.37
Al ₂ O ₃	12.43	12.58	12.44	12.72	12.66	12.59	12.87	12.77	12.54
Fe ₂ O ₃	1.61	1.38	1.42	1.99	1.96	1.67	1.40	1.74	1.47
FeO	1.55	1.66	1.49	0.82	1.33	1.20	2.67	1.26	1.60
MnO	0.08	0.10	0.08	0.04	0.07	0.08	0.14	0.07	0.08
MgO	0.40	0.42	0.27	0.20	0.32	0.26	0.35	0.43	0.36
CaO	0.93	0.99	1.24	0.83	0.47	0.59	1.61	0.65	1.29
Na ₂ O	3.94	3.58	3.53	3.62	3.87	3.78	3.60	3.77	3.59
K ₂ O	4.13	4.21	4.23	4.11	4.21	4.16	3.73	4.04	4.15
P ₂ O ₅	0.09	0.09	0.07	0.08	0.08	0.07	0.10	0.08	0.08
S	<0.02	<0.02	<0.02	0.07	<0.02	<0.02	0.02	<0.02	<0.02
H ₂ O ⁺	0.89	0.93	0.56	0.73	0.73	0.87	0.84	0.53	0.71
H ₂ O ⁻	0.33	0.35	0.34	0.29	0.36	0.38	0.29	0.37	0.22
CO ₂	0.83	0.11	0.14	0.29	0.17	0.16	0.28	0.56	0.15
rest	0.24	0.28	0.26	0.27	0.27	0.26	0.30	0.26	0.25
				100.06			100.08		
0 = S				0.03			0.01		
	99.89	100.38	100.01	100.03	100.02	100.13	100.07	100.24	99.92
Trace elements (ppm)									
Ba	725	875	710	745	740	735	920	720	730
Cs	2.8	3.7	11.7	4.1	4.4	3.2	6.2	4.1	10.0
Rb	159	168	201	166	165	172	136	165	169
Sr	95	187	142	161	123	127	205	146	150
Pb	22	29	31	30	27	30	27	29	31
Th	19.7	19.8	22.3	20.1	20.1	19.8	19.6	20.2	20.0
U	5.0	5.2	6.0	5.6	5.5	5.3	4.7	5.3	5.3
Zr	460	465	472	496	520	484	589	457	464
Hf	11.1	10.9	11.7	11.6	11.7	11.1	13.7	11.6	11.3
Nb	24.5	24.5	28.0	26.5	27.0	25.0	23.5	24.5	24.5
Ta	2.9	3.0	3.4	3.2	3.0	2.7	2.5	2.9	3.1
Y	79	84	83	81	91	87	73	89	78
La	64	68	54	69	64	69	76	67	72
Ce	132	136	136	135	141	154	156	140	137
Nd	64	67	62	65	69	71	75	67	66
Sm	14.9	15.6	15.0	15.7	15.5	16.5	16.2	15.8	14.8
Eu	2.42	2.49	2.41	2.68	2.69	2.73	3.48	2.59	2.46
Gd	12.4	13.4	12.9	12.8	12.8	14.1	13.2	13.7	12.7
Ho	3.2	3.2	3.3	2.9	3.2	3.1	2.8	3.4	3.0
Tm	1.1	1.1	1.2	1.0	1.2	1.1	1.1	1.2	1.2
Yb	8.2	8.2	8.7	8.0	8.8	8.5	7.8	8.8	8.2
Lu	1.27	1.26	1.36	1.22	1.37	1.30	1.20	1.35	1.25
Sc	15.2	15.1	14.8	14.7	15.5	15.0	16.8	14.8	14.8
V	8	6	5	7	7	6	6	7	6
Cr	2	2	2	2	2	2	2	2	2
Co	4	2	3	2	2	2	3	4	3
Ni	1	<1	2	<1	<1	1	2	2	<1
Cu	4	4	5	5	5	4	4	4	4
Zn	125	123	121	139	142	149	157	125	112
Ga	20.2	20.8	21.2	21.6	21.6	20.4	20.8	20.4	20.4

of high Ga and low Al relative to I-type granites is diagnostic (Fig. 4).

The very felsic Mumbulla Suite shares most of these very distinctive trace element features. Again, most of the highly-charged cations are more abundant than in the felsic I-types. Exceptions are Zr and Nb which are only very slightly more abundant (Fig. 3). The different Zr content is the most marked chemical difference between the two A-type suites. The Hf difference is less pronounced, with

a marked difference in Zr/Hf, averaging 42 for Gabo and 27 in Mumbulla.

REE are high in both A-type suites with abundances dropping from some 200 times chondrites for La to 40 times for Lu (Figs. 5 and 6). These abundances are some 2 times and 3 times higher, respectively, than average values of 105 and 13 times chondrites for 37 I-type granites of the Bega Batholith (B.W. Chappell, unpublished data). The negative Eu anomaly for the Gabo Suite is modest with

Table 4. Average compositions of Gabo and Mumbulla Suite granites and of felsic I-type granites from the Bega Batholith

	1 Mumbulla	2 I-type	3 Gabo	4 I-type
SiO ₂	77.21	76.03	73.04	72.50
TiO ₂	0.13	0.11	0.37	0.31
Al ₂ O ₃	11.79	12.64	12.62	13.63
Fe ₂ O ₃	0.36	0.46	1.63	0.70
FeO	0.85	0.70	1.51	1.65
MnO	0.03	0.03	0.08	0.06
MgO	0.04	0.24	0.33	0.72
CaO	0.39	0.80	0.96	2.32
Na ₂ O	3.08	3.43	3.70	3.09
K ₂ O	5.00	4.46	4.11	3.67
P ₂ O ₅	0.02	0.02	0.08	0.08
Trace elements				
Ba	575	331	767	577
Rb	242	212	167	160
Sr	43	67	148	165
Pb	37	24	28	19
Th	26	25	21	20
U	6	6	5	4
Zr	170	95	490	147
Nb	19	11	25	9
Y	90	46	83	29
La	64	30	62	33
Ce	150	68	153	71
Sc	16	6	17	9
V	2	6	6	30
Cr	<1	1	2	5
Co	1	2	3	6
Ni	<1	<1	1	2
Cu	9	2	4	3
Zn	122	20	133	37
Ga	20	14	21	15

1. Average of 8 analyses of Mumbulla Suite granites (see Table 2)
2. Average of analyses of the 20 most felsic I-type granites from the Bega Batholith
3. Average of 9 analyses of Gabo Suite granites (see Table 3)
4. Average of analyses of the 48 analyzed samples of I-type granite from the Bega Batholith with SiO₂ in the range 71.46–73.71% SiO₂, the SiO₂ range of the Gabo Suite

All trace element analyses averaged in this table were made by X-ray spectrometry, whereas, some analyses in Tables 2 and 3 were made by activation analyses (see footnote in text)

an average Eu/Eu* of 0.56. For the Mumbulla Suite the anomaly is more pronounced with an average Eu/Eu* of 0.30.

A-Type Granites Elsewhere

Granites with characteristics similar to those recorded from southeastern Australia occur throughout the world. Some are associated in space and time with alkaline complexes which include biotite granites, peralkaline granites, quartz syenites and nepheline syenites; examples being Kola Peninsula, USSR (Gerasimovsky et al. 1974) and Gardar, South Greenland (Upton 1974). The ekerites of the Oslo province have the geochemical characteristics of A-types including high Ga/Al and high Nb: they are most comparable with the Gabo Suite in having exceptionally high Zr, REE and Y (Dietrich et al. 1965). In addition they have large Eu depletions (Neumann et al. 1977). The "alkaline group" granites of Massachusetts, USA are also associated

with syenites but these are all near-peralkaline (Buma et al. 1971). Relative to chondrites, these granites have a high, flat REE pattern but with a strong Eu depletion. The younger (Jurassic) granites of Nigeria also have similar REE patterns (Bowden and Whitley 1974).

In summary, the granites such as those discussed above, having A-type chemical characteristics also have characteristic mineralogy. The biotites are annite-rich and the feldspar is mostly alkali feldspar, often albite-orthoclase solid solutions or intergrowths. We suggest that most, if not all, of the hypersolvus granites of Tuttle and Bowen (1958) are A-types. Micrographic intergrowths of quartz and alkali feldspars are very common. Annite is late in the crystallization sequence occurring as interstitial grains between alkali feldspars and quartz: it commonly contains strips of fluorite or is intimately associated with that mineral.

All granites with A-type affinities have intruded late in the magmatic cycle, or have intruded crystalline basement that had undergone previous partial melting. These granites are commonly associated with tensional regimes in continental blocks (e.g. Western U.S.A., Lipman et al. 1972), but may occur in areas that do not appear to be related to any tectonic setting (e.g. Pikes Peak Batholith, Barker et al. 1975).

Summary of Chemical Characteristics of A-Type Granites.

Compared with typical I-type granites of similar silica contents, A-types in general contain lower abundances of MgO and CaO, but higher Na₂O + K₂O, although I-types with very high SiO₂ approach the values found for these elements in A-types. Over the complete SiO₂ range, some large, highly charged elements, particularly Nb and Ga and other elements not measured on southeastern Australian rocks such as W and Mo are significantly higher than for other granites. Preliminary data (Chappell, unpub.), indicate that Sn contents of the A-type suites in southeastern Australia are twice as high as the associated I-types of the Bega Batholith. In metaluminous and peralkaline A-types, Zr, Y, REE, Pb and Zn, are also relatively abundant. There is a considerable overlap of Sr and Rb with those of siliceous I-types, and the transition elements, V, Ni, Co and Cr have much lower abundances.

Limited chemical and mineralogical data on A-type granites indicate that F and Cl could have been exceedingly high in the magmas with Cl possibly being more abundant in peralkaline types and fluorine in metaluminous and peraluminous types. The alkaline Younger Granites of Nigeria contain F contents in excess of 1,000 ppm and some values exceed 4 wt.% (cf. Imeokparia 1982). These granites contain inclusions of fluorite in biotite, a feature of the biotites from the Watergums pluton in the Gabo Suite. Most F analyses on A-type granites indicate contents in excess of 1,000 ppm. However, the variation in the amount of fluorine actually measured in A-type granites is considered to result from anion exchange with later circulating waters, and may not be indicative of original magmatic contents.

Petrogenesis of A-Type Granitoids

Derivation by Fractional Crystallization from a Syenitic Magma

Rocks such as the ekerites (A-type granites) of the Oslo province have traditionally been regarded as products of

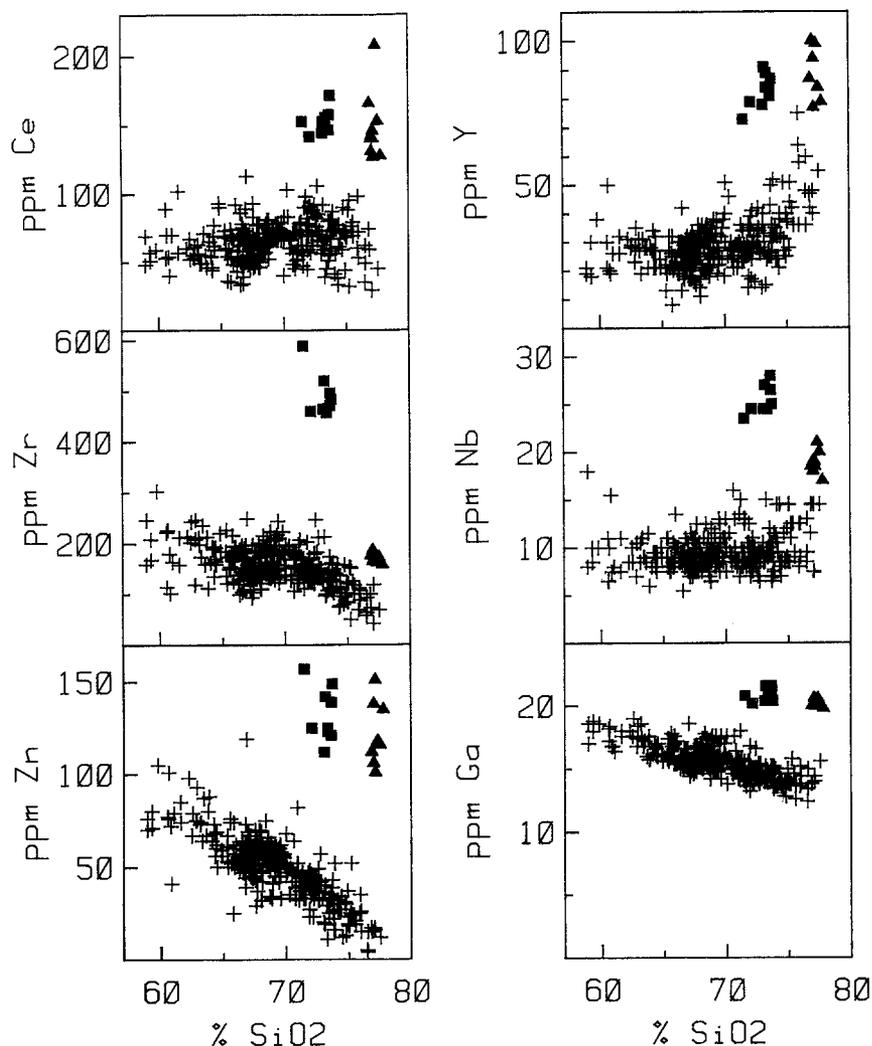


Fig. 3. Harker variation diagrams showing the abundance of some highly-charged cations, and Zn for the Gabo Suite (*squares*) and the Mumbulla Suite (*triangles*), compared with abundances in I-type granites of the Bega Batholith

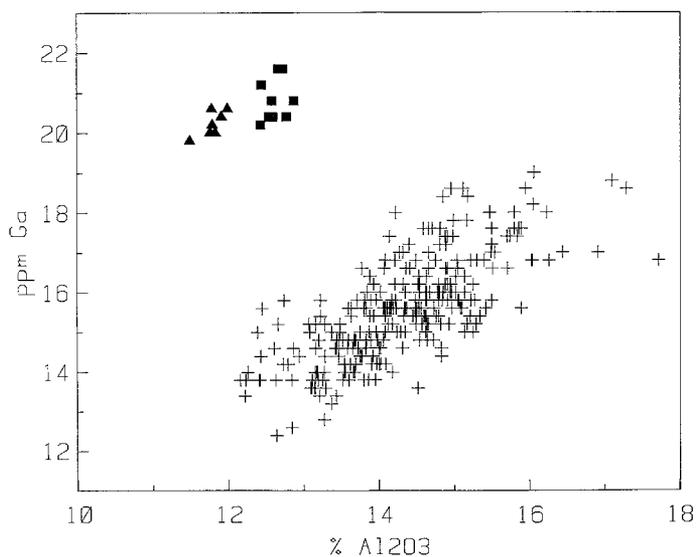


Fig. 4. Plot showing high Ga relative to Al_2O_3 for A-type granites in contrast with normal I-type values of the Bega Batholith. Symbols as for Fig. 3

fractionation from a fairly mafic syenitic magma with a SiO_2 content near 57% (Barth 1945); the sequence being from kjasite \rightarrow larvikite \rightarrow nordmarkite \rightarrow ekerite. The biotite granite of the Oslo region is not considered to be part of the sequence (Neuman et al. 1977).

Petrographic studies indicate that the magmas that crystallize to give A-type granites were completely molten. Most other granites of southeastern Australia, including those of the Bega Batholith, contain some, and commonly large amounts of "restite" derived from the source of partial melting (e.g. White and Chappell 1977; Beams 1980). Restite is absent in the A-type granites, which means that there is no petrographic evidence against fractional crystallization.

The high concentration of elements such as Nb and the low abundance of others such as Ca is consistent with a fractionation model. However, Dietrich et al. (1965) in their study of the Oslo ekerites point out that K/Rb, an "indicator of differentiation", is not as low, and Cs is not as high as expected; they explained the relatively high K/Rb by suggesting that the Oslo province has a higher "K/Rb ratio than the crustal average". But the K/Rb in many A-types from various parts of the world including those

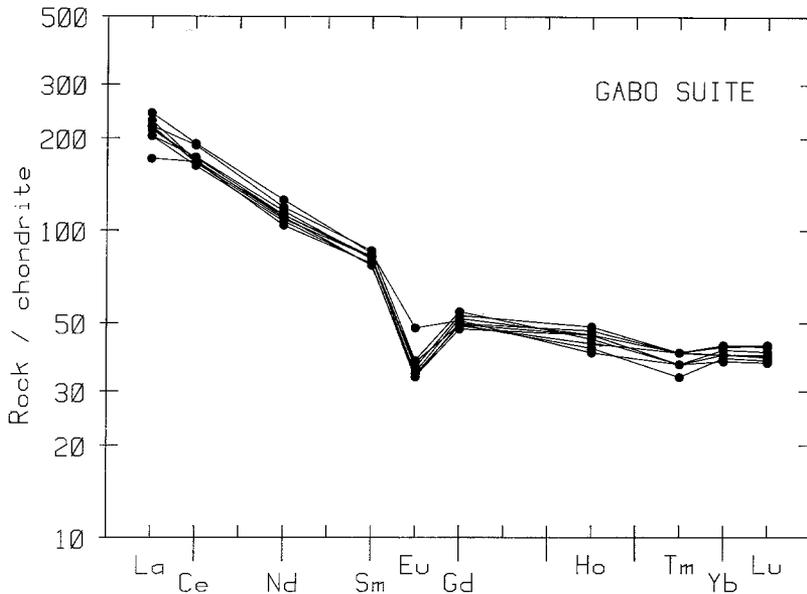


Fig. 5. Chondrite-normalized plot of rare-earth element abundances in 9 granites of the Gabo Suite

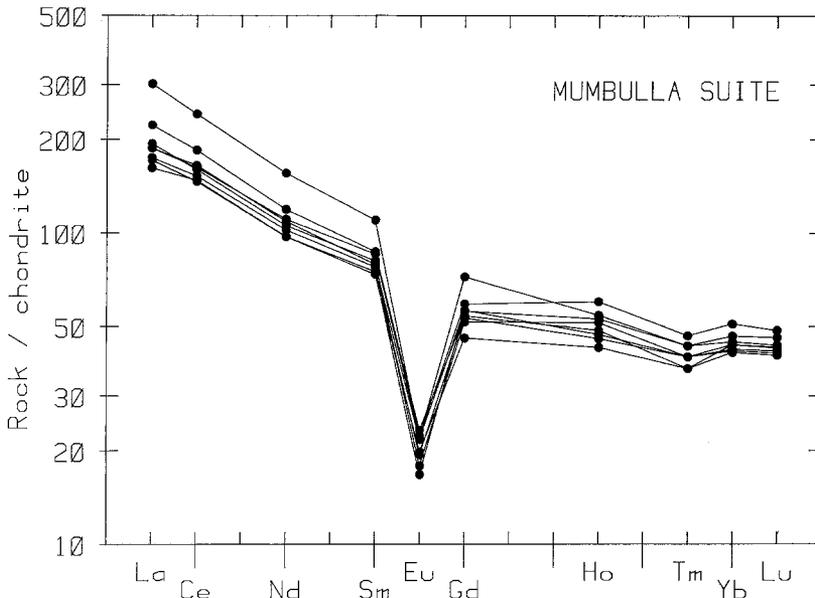


Fig. 6. Chondrite-normalized plot of rare-earth element abundances in 8 granites of the Mumbulla Suite

from southeastern Australia do not have the low K/Rb necessary for their production from syenites by feldspar fractionation. However, crystallization of an A-type magma involving both alkali and plagioclase feldspar fractionation may occur. In southeastern Australia, the subtle, regular variations over a limited SiO_2 interval of certain elements (see Fig. 3) and elemental ratios (e.g. K/Rb) within A-type suites supports fractional crystallization, but only on a minor scale. In contrast, these ratios remain remarkably constant in Bega Batholith I-type granites which result from restite separation.

A-type granites may occur as large bodies (plutons) comparable in size to other granite plutons. The majority are not associated in space and time with intermediate rocks; most of those A-types occurring in normal granitic terrains have probably been overlooked and as more are recognised, the number not associated with syenites should increase. Some A-type granites are associated with gabbros

or basalts (e.g. Fergusson et al. 1979). However, there is a Chayes (1963) gap in the sequence; intermediate rocks are rare.

A-type granites have extremely variable initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratios, ranging from 0.703 (e.g. Evisa Complex, Corsica, Bonin et al. 1978) to over 0.720 (Younger Granites of Nigeria, van Breemen et al. 1975). Low initial ratios support a model of fractional crystallization from mafic to intermediate sources, and the general observation that low ratios characterize A-type granites that are associated with syenites and more mafic rocks, provides additional evidence for this model. However, using a fractionation model, high initial ratios cannot be explained without appealing to mixing or contamination of mantle and crustal material. If a lower crustal source is invoked, then isotopic variation may result from different source rocks and from the time lapse between production of the source and generation of A-type granites. Furthermore, it is possible that metaso-

matic processes have a greater effect on changing initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratios in A-type granites, relative to other granites (van Breemen et al. 1975). Isotopic evidence relating to the origin of the Gabo and Mumbulla Suites will be presented at a later date.

These data suggest that although some A-type granites may be produced by crystal fractionation from a syenitic or even lower SiO_2 source magma, such a mechanism is not a general explanation for their origin.

Differentiation of an I-Type Magma

We argue against production of the A-type melts by fractional crystallization of I-type melts for two reasons. First, the A-type melts were dry or almost anhydrous, as evidenced by the late precipitation of biotite and amphibole as interstitial crystals. Any fractionation from a felsic I-type melt would lead to a hydrous melt because of early, anhydrous phase crystallization. Second, the low Rb content and fairly high Sr content of both southeastern Australian A-type suites is not consistent with their production by extensive fractionation involving feldspars. It has been shown quite conclusively, e.g. McCarthy and Hasty (1976) that fractionated granite melts have Rb and Sr contents respectively very much higher (~ 700 ppm) and very much lower (< 10 ppm), than the values we report here.

Presumably because of the absence of evidence for feldspar fractionation, certain rhyolitic volcanic rocks of the western USA having similar compositions to A-type granites, have been explained by a process of "thermogravitational diffusion" (Shaw et al. 1976). This model requires that the large highly charged cations such as Nb concentrate at the top of an I-type magma chamber by a process involving diffusion. However, A-type plutons have similar size and shape to those of adjacent I-types and they are not always associated with volcanic rocks. Also, the metamorphic aureoles around A-type plutons are identical to those of any other granite emplaced at similar structural levels. This, together with the observation that the rocks of the Mumbulla intrusion show no variation in composition over a vertical interval of at least 450 m, leads to the conclusion that A-types are not differentiation products of I-type magmas at the top of a magma chamber.

Partial Melting

If direct partial melting is a viable process for the formation of A-type melts, the only constraints that can be placed on the nature of the source are that (a) it must contain quartz + K-feldspar + plagioclase either as separate phases or as normative components necessary to form any granite and (b) it must be fluorine and/or chlorine-rich but poor in water. This source composition may be satisfied if the source rock has already had a granitic magma extracted from it. A residual source was suggested by Barker et al. (1975) for the Pikes Peak Batholith of Colorado. A residual source also explains the occurrence of A-type magmatism later than the main S- or I-type granite magmatism.

It has been shown that the degree of partial melting necessary to obtain an I-type granite magma from a dioritic lower crust, is about 25% (Compston and Chappell 1979). Melting is initiated by vapour absent breakdown of biotite and amphibole which provides the water necessary for melt formation (Burnham 1979). According to the restite model

of White and Chappell (1977), these I-type granite magmas consist of unmelted residual material (restite) plus a felsic melt; the magmas may progressively separate restite upon ascent and emplacement into the upper crust. An important implication of this model is that magmas ranging from tonalites to granites may all be formed at relatively low temperatures ($< 850^\circ\text{C}$). These can be referred to as "minimum" and "near-minimum" melts.

The amount of "minimum" or "near-minimum" melt produced is controlled by the release of volatiles and the melting of plagioclase, K-feldspar and quartz. If one of these components is exhausted, the production of enough melt necessary for magma movement is dependent upon dissolution of one or more of the remaining phases and this can only occur at higher temperatures and a "non-minimum" melt is produced. In the case of minimum melt granitoids there is always enough of each of the melt components in the source region to permit production and movement of granite magma at minimum temperatures. The residue remaining after production of a "minimum" or "near-minimum" I-type is normally felsic and contains the mineral assemblage, quartz + plagioclase + K-feldspar + orthopyroxene \pm clinopyroxene. If amphibole remains as a stable residual phase, it is F-rich (Holloway and Ford 1975). Likewise any residual biotite is most probably F-rich (Munoz and Ludington 1974). Lesser amounts of Cl would be retained in biotite and amphibole. Other residual minerals include apatite, zircon \pm sphene \pm magnetite. This residual rock is a granulite (Brown and Fyfe 1970; Nesbitt 1980).

In southeastern Australia, most of the early Devonian I-type granites that formed in the same region as late Devonian A-type granites are of "minimum-melt" type. This means that the early melting episode left a residue in the deep crust containing quartz + feldspar + mafic minerals. Formation of another melt from such a residue must be at higher temperatures than for first melt production. For this to occur at a reasonable temperature it is necessary for the melt to have at least a small amount of dissolved volatile component. This could be achieved by the high temperature, vapour absent breakdown of the F- and Cl-rich amphiboles and biotites residual from the previous melting event. Apart from quartz and feldspars other components that may also melt include accessory phases. A high proportion of the large highly charged cations may be contributed by melting of these phases with minor contributions from the 6-fold (and higher) coordination sites within mafic silicates.

Nature of A-Type Melts

Based on the above considerations, A-type magmas should contain larger amounts of fluorine, and chlorine, than I-type granite magmas. HF and HCl behave similarly to H_2O in a granite melt by reacting with the bridging oxygens of the aluminosilicate tetrahedra. F^- , Cl^- and OH^- break the Si—O bridges and depolymerise the melt resulting in a decrease in viscosity (Burnham 1979). Whereas HF is more soluble than H_2O in the melt, HCl is less soluble (Wyllie and Tuttle 1964). F forms stable Si—F bonds relative to Si—OH bonds, but Si—Cl bonds are much weaker and longer. Therefore, both F and Cl result in considerable distortion of the aluminosilicate structure within the melt. The increase in disorder and distortion of the melt frame-

work should produce a large range of sites of various size and coordination allowing the large highly charged cations to form higher-order coordination structures than in normal granitic melts. Bonding between these large highly charged cations, adjacent halogens, OH and alkalis in the melt produces structures somewhat analogous to those of lower temperature fluid phases. These structures result in the relative retention and ultimate enrichment of the large highly charged cations during progressive crystallization.

Although the distortion of the Si–O framework by F and Cl could produce octahedral and higher coordination sites capable of accommodating Ni, Co, V and Cr, these elements are most readily accommodated in residual pyroxenes because they have high crystal field stabilization energies, and like Mg have a compatible ionic radius for the M_1 octahedral site of pyroxene (Burns 1970). They are therefore likely to be of low abundance in the melt. Any subsequent fractionation involving pyroxene or amphibole will further deplete the magma in these elements. Like the large highly charged cations, Zn, Pb and possibly Cu are not readily accommodated in pyroxene structures, resulting in the retention of these elements in the melt during crystallization.

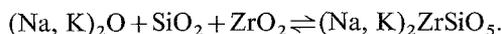
Perhaps the most diagnostic feature of A-type granites is their high Ga/Al ratios. Preferential retention of An-rich plagioclase in the source-region during partial melting could lead to melts with high Ga/Al ratios because Ga is preferentially excluded from the anorthite structure, relative to Al (Goodman 1972). However, of more fundamental significance is the effect of fluorine upon Ga retention in aluminosilicate melts. The stability of the octahedral GaF_6^{3-} complex at high temperatures is well established (e.g. Cotton and Wilkinson 1980), but AlF_6^{3-} structures only appear to be stable in water-saturated or near-saturated aluminosilicate melts at magmatic temperatures (Manning et al. 1980). Thus, during production of an A-type magma, Ga would partition into the melt more readily than Al, and would also be preferentially retained in that melt during crystallization because the melt is relatively dry. In contrast, chloro-complexes of Al and Ga are not stable at magmatic temperatures (Cotton and Wilkinson 1980) and accordingly do not affect Ga/Al ratios in A-type melts.

Dissolved volatile components exert a strong influence on the crystallization of amphibole in granitic magmas (e.g. Burnham 1979). Petrographic evidence indicates amphibole may crystallize early in A-type magmas even through these magmas are relatively anhydrous. Crystallization of amphibole results from stabilization of the amphibole structure by F^- , which enters the 'OH sites'. In fluorine-rich magmas therefore, Ca-amphiboles (e.g. hastingsite) are early crystalline phases, resulting in the magma becoming not only enriched in the alkali feldspar component, but ultimately peraluminous with progressive crystallization. This trend could be enhanced by the increasing stability of the AlF_6^{3-} structure in the melt if the magma becomes more hydrous during crystallization (Velde and Kushiro 1978; Manning et al. 1980) which is consistent with the late micas found in pegmatites. If fluorine is relatively low in the melt, early amphibole crystallization may not occur and anorthite-rich plagioclase could be the dominant crystallizing phase, resulting in the "plagioclase effect" of Bowen (1945) and a peralkaline trend. Fluorine concentration in the melt is therefore considered to be critical in determining fractionation trends in A-type magmas.

Fluorine in an A-type melt forms stable ionic, high coordination complexes with the large, highly charged (d^0-d^2) metal ions in preference to Ca^{2+} . Low CaO contents in A-type magmas and the high solubility of fluorite in high temperature alkali chloride solutions (Richardson and Holland 1979) indicate that CaF_2 is not stable under these magmatic conditions. The small size and high degree of electronegativity of F^- enhances the stabilization of octahedral and higher order coordination compounds with the large highly charged cations (e.g. MoF_6 , WF_6). Halides of the heavy transition metals may also have metal metal interactions resulting in higher coordination numbers, as is often found in the 'subhalides' of Nb and Ta (e.g. $(TaF_5)_2$) and the trihalides of Mo and W (e.g. MoF_3 , WCl_3). The lanthanide and actinide trifluorides (e.g. CeF_3 , UF_3) form stable clusters of 5- and 6-fold trigonal structures, also by metal-metal interactions (Fergusson 1974). The REE may also form stable CaF_2 type structures such as LaOF in 8-fold coordination. The fact that magmatic and pegmatitic fluorite may be enriched in the REE and Y, which substitute for Ca, indicates these complexes are stable in the melt.

The heavy transition metals, as well as the lanthanides and actinides, may form high-coordination complexes by the addition of alkali metal counter ions. The coordination number and the stability of these structures is dependent upon the size of the counter ion and complexes such as Na_3TaF_8 , Na_2NbF_7 , Na_2UF_6 and $KCeF_4$ would be produced.

The high concentration of Zr in per alkaline melts (Bowden 1966) relative to peraluminous melts is also considered to result from counter ions (e.g. Na, K) associated with high coordination structures within an A-type aluminosilicate melt. The excess of alkali metals over aluminium in peralkaline melts increases the solubility and therefore the relative mobility of Zr resulting in the formation of polymeric dodecahedral alkali-zircono-silicate complexes according to a reaction of the form:



Therefore, an excess of alkalis favours development of the complex, which increases the concentration of Zr in the melt. The presence of alkali cations within the Zr polymer prevents development of the zircon structure thereby inhibiting zircon crystallization. The constant 2:1 molar ratio of excess alkalis to ZrO_2 in alkaline melts (Watson 1979) is consistent with this model. In addition, alkali zircono-fluoride complexes are known to be stable at high temperatures and exhibit high coordination numbers as well as a great variety of coordination polyhedra (Fergusson 1974). The presence of excess alkalis in A-type magmas should promote the formation of melt complexes such as Na_2ZrF_6 and Na_3ZrF_7 , particularly if the concentration of fluorine is high. In fact, just as for Ga, much of the Zr may exist as fluoride complexes. Crystallization of amphibole and biotite will remove F^- from the melt, destabilizing the zircono-fluoride complexes, resulting eventually in the precipitation of zircon.

Zinc forms complexes with fluorine at magmatic temperatures, e.g. ZnF_2 (Cotton and Wilkinson 1980). High Zn contents in some A-type granites can be explained by the presence of excess alkalis in the melt. Fluorozincate complexes may be stabilised by the counter-ion effect; the distorted octahedral ZnF_2 structure changes to a perovskite

structure (e.g. NaZnF_3 , KZnF_3 , Aylett 1973) with no distorted octahedra.

In peraluminous melts, alkali cations are strongly bonded in the feldspar structure and are not available as counter ions. Alkali-zircono-fluoride or silicate complexes will therefore not form and zircon crystallization will commence early, relative to peralkaline melts. This process explains the low Zr content of high- SiO_2 , peraluminous A-type granites (Collins, unpublished data) as well as the abundance of refractory zircon in the restite of other granite types. Neither S- nor I-type granites are strongly alkaline and therefore zircon will not readily 'dissolve' into the melt and will therefore remain as a residual phase during partial melting (see also, Watson 1979).

The counter-ion effect may also be a crucial factor in retaining other highly charged cations in alkaline melts. Indeed, the higher proportion of these cations in alkaline and peralkaline melts (e.g. peralkaline nepheline syenites and A-type granites) relative to peraluminous melts supports this contention.

Separation of a Fluid Phase

The coordination sites of the large highly charged cations in the melt form as a result of distortion of the aluminosilicate tetrahedra by complexing with ligands other than OH. The cations are initially coordinated by the oxygen and halide anions of the aluminosilicate complex, forming hybrid metal-oxygen bonds and more ionic metal-halide bonds. Various complexes are formed depending upon the coordination number of the cation and the number of different ligands in the complex. At higher halide concentrations, either as a result of source rock characteristics or progressive crystallization, discrete fluoro-chlorocomplexes would begin to form as separate structures within the melt.

The large size of the Cl^- ion precludes the development of strong Si-Cl bonds in an aluminosilicate melt. Accordingly, any chloride complex in the melt will be readily hydrolysed if an aqueous phase develops and Cl^- will partition strongly into that phase. The metal transporting capacity of chlorides in magmatically derived hydrothermal solutions has been well established experimentally (e.g. Kilinc and Burnham 1972; Holland 1972) and is well supported by fluid inclusion studies (e.g. Sawkins and Scherckenbach 1981). Alkali chlorides are very stable in aqueous solutions, but the high stability of the FeCl_2 complex (Chou and Eugster 1977) will also result in high iron chloride concentrations during vapour phase evolution (Burnham 1979). Cl^- would also be a particularly effective scavenger of the metal ions with weak crystal field stabilization energies. It forms stable tetrahedral complexes of the M(II)X_4^{2-} type with Mn, Pb, and Zn (Holland 1972) and also forms distorted tetrahedral and octahedral complexes with Cu^{2+} ; this is a result of the Jahn-Teller effect (e.g. Cotton and Wilkinson 1980). Both Cl^- and F^- form stable trigonal pyramidal complexes with Sn^{2+} in a fluid phase, particularly under conditions of low pH and low f_{O_2} (Eadington and Giblin 1979). At high halide concentrations, Cl^- may also form octahedral SnCl_6^{2-} anions, which are more stable than the distorted SnF_6^{2-} structure in acid solutions (Fergusson 1974).

The genetic link of ore deposition with F-bearing fluids has been established by several workers, e.g. Gunow et al.

(1980), Kwak and Askins (1981) who also identified a magmatic source for the fluorine. However, the fluid evolved from an A-type magma will probably be dominated by H_2O because F is incorporated in early amphiboles as well as interstitial biotite. This means that the role of F^- as a complexing agent in the fluid phase may be significantly less than the role of F^- in the melt. If fluoro-complexes of the highly charged cations exist in the melt at this late magmatic stage they will be destabilized by hydrolysis because the heavy transition metals such as Nb, Ta, W and Mo generally form polymeric oxo-anions in aqueous solutions particularly under acid conditions (Cotton and Wilkinson 1980). The lanthanides and actinides are also hydrolysed, generally forming high coordination structures involving H_2O molecules as well as halides (Moeller 1972). Increase in pH associated with dilution and decrease in temperature of the hydrothermal system as it moves away from the host A-type magma destabilizes any remaining rare-metal fluoro-complexes, resulting in the precipitation of F-rich minerals such as micas, topaz and fluorite. These minerals occur in sheet-like greisens at the top of the granite pluton, with disseminated ore in the granite, or within mineralized veinlets above the granite. Gunow et al. (1980) described the association of various F-bearing minerals with different alteration zones and mineralization at the Henderson molybdenite deposit, Colorado. The general association of F-bearing ore deposits with "alkalic igneous rocks" in the western U.S.A. is compatible with this model (cf. Lamarre and Hodder 1978).

Acknowledgements. Financial support was received from the Australian Institute of Nuclear Science and Engineering to cover neutron irradiation costs. The ready cooperation of staff of the Australian Atomic Energy Commission in carrying out the irradiation is acknowledged. Valuable technical support was provided by Mr. R. Freeman, Mr. Z. Wasik, and Mrs. E. Webber at the ANU.

V.J. Wall constructively criticized an early draft of this paper.

We are grateful to H.W. Nesbitt for stimulating discussions on the concepts presented and to the teachings of C.W. Burnham.

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