

CHAPTER 3

HILLGROVE PLUTONIC SUITE

3.1 INTRODUCTION

The granitoids of the New England Batholith have recently been grouped into several suites according to the nature of the inferred source rocks (Flood and Shaw, 1975; Chappell, 1978). These authors recognized two major granitoid types, 'I- and S-type', depending on the igneous or sedimentary nature of the respective source rocks. The Hillgrove Suite is one of two suites of S-type granitoids which have been distinguished in the New England Batholith. The other is the Bundarra Suite, an extensive group of coarsely porphyritic cordierite-bearing adamellites which flank the western margin of the New England Batholith from the Queensland border to just north of Bendemeer (Fig. 3.1).

The Hillgrove Suite comprises 21 intrusions ranging in composition from adamellite to granodiorite. The majority of these intrusions occur within a small, northeast-trending arcuate belt which constitutes the eastern and southeastern margin of the New England Batholith. A secondary 'belt' of intrusions trending in a southeasterly direction occurs ~ 50 km east of the main belt and comprises at least three plutons. Members of both belts occur as relatively small, individual, often elongate bodies displaying a well-developed tectonite fabric in addition to an earlier foliation. Many plutons are truncated or transected by prominent N-S and NE-SW trending faults.

Characteristic mineralogical features of the Hillgrove Suite granitoids include ubiquitous graphite, coexisting reddish-brown and green biotite two generations of plagioclase feldspar, alkali feldspar which is highly ordered (i.e. microcline), ilmenite as the sole iron-titanium oxide, coexisting colourless and blue varieties of quartz, Fe-Mg and calciferous amphiboles and the not uncommon occurrence of modally minor almandine-rich garnet. The mineralogy and the characteristic deformation microstructures of these granitoids distinguish them from other granitoids of the New England Batholith, particularly from the Bundarra Suite which generally lacks amphibole, has only one obvious generation of plagioclase, a colourless quartz, and typically contains either fresh or extensively pinitized cordierite.

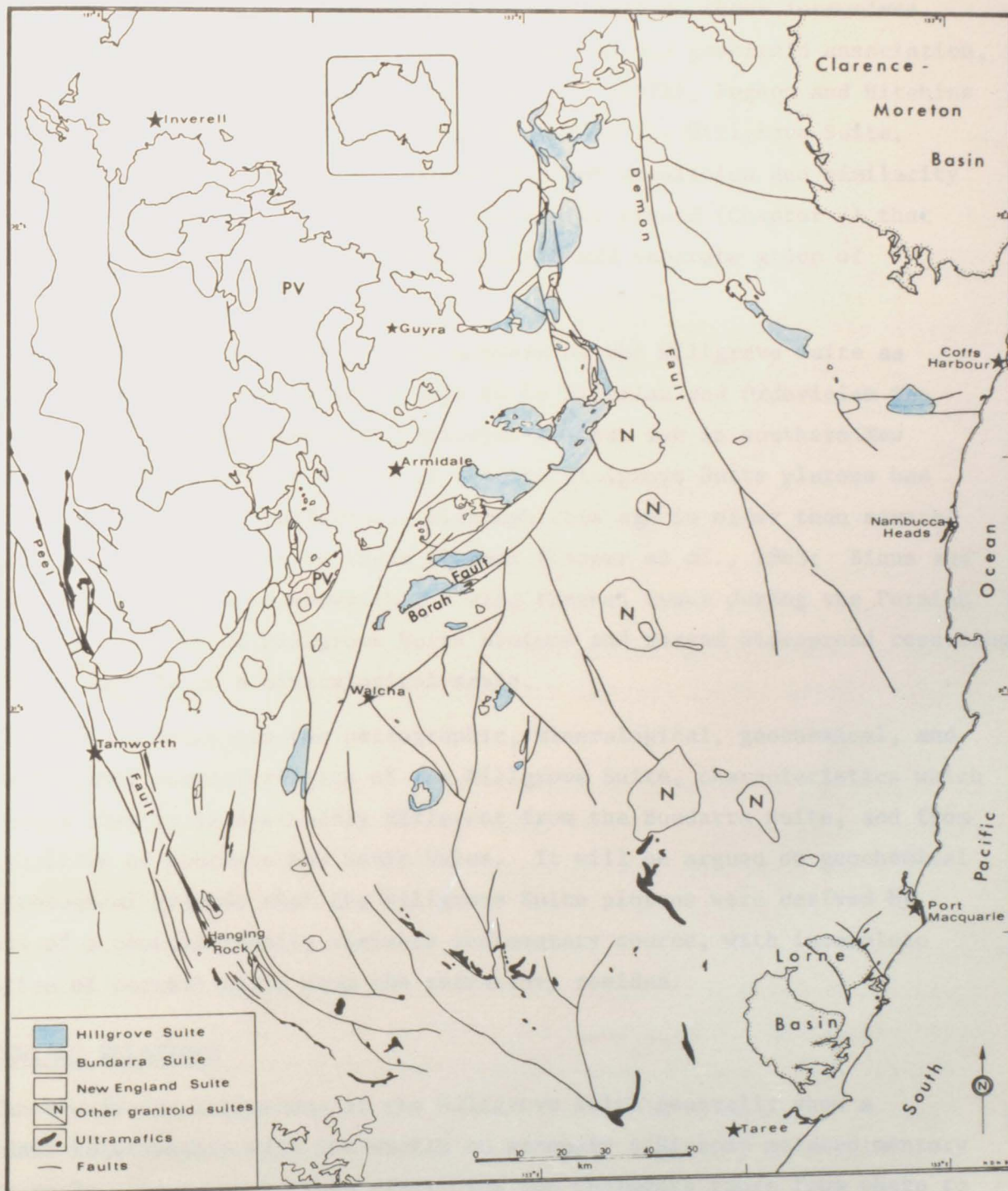


Figure 3.1 Distribution of plutons belonging to the Hillgrove Suite.

Closely associated with the foliated and deformed granitoids of the Hillgrove Suite are numerous, small, heterogenous, mafic intrusions of tholeiitic and calc-alkaline magmatic affinity. Because these intrusives have been difficult to assign to a specific New England granitoid association, Binns (1966), Binns *et al.* (1967), Leitch *et al.* (1971), Pogson and Hitchins (1973) and Scheibner (1973, 1976) grouped them with the Hillgrove Suite, largely on the basis of field association with these granitoids and similarity in degree of deformation. It will be subsequently argued (Chapter 7) that these mafic rocks in fact constitute a specific and separate group of intrusions.

Browne (1929) classified certain members of the Hillgrove Suite as "synchronous granites" and believed them to be Silurian and Ordovician in age, by analogy with the stressed granitoids of that age in southern New South Wales. However, Rb-Sr dating of all the Hillgrove Suite plutons has shown them to be Mid-Carboniferous. Although this age is older than several earlier K/Ar determinations on these plutons (Cooper *et al.*, 1963; Binns and Richards, 1965) it will be shown that a mild thermal event during the Permian affected the majority of Hillgrove Suite plutons and caused widespread resetting of isotopic systems on a mineralogical scale.

This chapter documents the petrographic, mineralogical, geochemical, and geochronological characteristics of the Hillgrove Suite, characteristics which demonstrate that it is distinctly different from the Bundarra Suite, and from the batholiths of southern New South Wales. It will be argued on geochemical and petrological grounds that the Hillgrove Suite plutons were derived by anatexis of a compositionally variable sedimentary source, with incomplete separation of partial melts from the refractory residua.

3.2 CONTACT RELATIONS

The twenty-one intrusions of the Hillgrove Suite generally show a concordant relationship with the weakly to strongly schistose metasedimentary country rocks. Contacts between granitoids and sediments range from sharp to gradational. Where contacts are sharp, the sedimentary rocks show minimal metamorphic effects. Where contacts are gradational, the relationship to the metasediments may be complex. For example, at the south-western margin of the Argyll Granodiorite, the western margin of the Abroi Granodiorite and the

northern margin of the Tia Granodiorite, the contacts are gradational over at least 20 m. Country rocks coarsen and partially melt close to the plutons. Along the southern margin of the Rockvale Granodiorite the partial melts were concentrated within compositionally favourable beds, resulting in arterites. At the southern margins of the Hillgrove and Wollomombi Adamellites the contacts with the metasediments show relatively extensive lit-par-lit interleaving of granitic and aplitic rocks with sediment. Similar lit-par-lit injection involving microgranodiorite and numerous aplite veinlets and quartzo-feldspathic schists, is developed around parts of the Tia Granodiorite.

Mild metamorphic effects at the sharply defined contacts and the absence of chilled margins suggest that emplacement of the granitoids occurred at comparatively low temperatures. Some authors (e.g. Ramberg, 1952; Stephansson, 1975) would argue that it could also reflect emplacement of the granitoids by diapiric uprise. Flood (1971) suggested that the Kilburnie Adamellite continued to rise into the crust after crystallization. He cited textural features, e.g. cracked and bent feldspars, as evidence for this intrusive mechanism. By analogy with the Kilburnie Adamellite, the juxtaposition of the Hillgrove and Enmore Adamellites against weakly contact-metamorphosed country rocks might be similarly explained by continued diapiric uprise after crystallization.

The dips of contacts of the Hillgrove, Kilburnie, Rockvale and Enmore plutons are generally steep to vertical; other plutons such as the Kookabookra Adamellite and the Abroi, Tia and Argyll Granodiorites, apparently possess flatly dipping and irregular contacts. The dips of many contacts are difficult to ascertain. For example at the northwestern margin of the Kookabookra Adamellite the granitoid demonstrably overlies schistose metapelites. Although the overlying sheet is now only 10 m thick, and restricted in extent, scattered outcrops suggest that it may have been originally much thicker. An irregular, but essentially flat-lying contact, is also indicated at the contact between the Abroi Granodiorite and high-grade metasediments at its northern and northwestern margins. Numerous "windows" of the Abroi Granodiorite emerge through the high-grade and migmatitic country rocks.

Apart from two possible exceptions, members of the Hillgrove Suite are rarely intrusive into each other; hence cross-cutting field relationships cannot be applied to ascertain the sequence of intrusion. The two exceptions are between the Abroi Granodiorite and the Tobermory Adamellite and between

the Abroi and Rockvale Granodiorites (see Fig. 3.1). In both cases, contacts are marked by contrasting outcrop patterns, i.e. from whalebacks (Abroi Granodiorite) to smaller, rubbly corestones (Rockvale Granodiorite and Tobermory Adamellite), as well as by slight differences in the colour indices of these granitoids. Finer-grained variants occurring within each of these plutons, and indeed within many of the Hillgrove Suite plutons, are interpreted as fragments of already crystallized rock which have been incorporated into later pulses of the same or similar magma, rather than forceful intrusion into each other.

3.3 STRUCTURAL RELATIONS

The outcrop patterns of the Hillgrove Suite plutons generally parallel the direction of major regional faults; for example, the Abroi Granodiorite and Enmore Adamellite are elongated parallel to the N-S trending Wongwibinda and NE-SW trending Borah Faults respectively (Fig. 3.1). This elongate character of S-type granitoids is particularly well displayed throughout the southern part of the Tasman Orogenic Province, e.g. the Bega, Kosciusko and Murrumbidgee Batholiths.

Many faults are intimately associated with the Hillgrove Suite. However, despite the fact that some exceed 75 km in length, e.g. Rowleys Fault, their precise attitudes and their direction of movement are difficult to determine. Some appear to be at least partly strike-slip as evidenced by the horizontal dislocation of granitoids. Severe granulation and mylonitization is manifest in both the granitoids and the adjacent country rocks. For example, cataclasis increases progressively in all rock units with decreasing distance to the Wongwibinda Fault, and earlier fabrics, especially those in the Abroi Granodiorite, are partly or completely obliterated. Within tens of meters from this fault the primary foliation, which elsewhere throughout the elongated portion of the pluton subtends an angle of 30-40° to the secondary cataclastic foliation, is totally obliterated. At a distance of several meters from the fault all rocks appear blastomylonitic and the distinction between the juxtaposed, low-grade Lower Permian greywackes of the Dyamberin Beds on the eastern side and the granitoids on the western side of the fault is very difficult. At its northernmost point the Wongwibinda Fault is marked by intensely brecciated Henry River Adamellite "healed" by anastomosing veinlets of hyalomylonite. Deformation has been similarly intense along the NE-SW trending Borah Fault, where a zone over 200 m wide contains rocks which are mostly ultramylonites of unrecognizable parentage.

The development of internal fabric (foliation and lineation) in the Hillgrove Suite plutons varies considerably, from almost isotropic to mylonitic. The foliation and lineation directions also vary considerably, with changes in direction being more pronounced *between* plutons than within plutons. Small changes in these directions within plutons (domains) can usually be related to changes in the attitude and direction of nearby major faults.

The variation in foliation and lineation directions observed in the Hillgrove Suite granitoids would probably be understood better if detailed structural data on the country rocks were known more precisely. For example, structures in the country rocks generally reflect at least two deformations and in the Wongwibinda and Moona Plains areas the rocks may show up to three deformations (Stephenson and Korsch, 1977; Hensel, 1972). In both these areas the deformation intensity increases towards a major structural dislocation and in both areas successive deformations become progressively more dominant.

Foliation in the Hillgrove Suite granitoids is ubiquitous, generally defined best by the alignment of predominantly medium-grained biotite flakes and more-or-less elliptical granules of quartz. Rare streaks, schlieren and aggregates of biotite, schistose metasedimentary xenoliths and accessory muscovite enhance this foliation but alkali feldspar phenocrysts and alternating leucocratic and melanocratic bands do not contribute to the development of foliation in these plutons. The intensity of foliation in the Hillgrove Suite granitoids is related to three main factors: (1) grain size, (2) biotite content and (3) proximity to a major fault. Fine-grained granitoids display at best only a mild foliation suggesting that they were apparently less susceptible to deformation than their coarse-grained counterparts. Similarly, very felsic granitoids appear to show only a vague foliation. However, it is clearly evident in thin-section that many of these felsic variants have been intensely deformed; indeed more so than most of the adamellites and granodiorites.

3.3.1 Deformation Textures

It is not certain at what stage in their history the Hillgrove Suite plutons developed their deformation textures. The textures (and therefore foliation) in many of the plutons have previously been interpreted by Binns *et al.* (1967) as predominantly protoclastic, i.e. resulting from deformation of minerals in an incompletely crystallized magma. Whether or not these textures are primary, developed during emplacement of granitoids into

geosynclinal or orogenic environments, has been a matter for some debate. Thus, Waters and Krauskopf (1941) interpreted the textures of the Colville Batholith, Washington, to be predominantly protoclastic, formed when the more mobile and partly liquid central mass of the batholith intruded its own, nearly solid, periphery. Such peripheral effects clearly do not occur in the stressed Hillgrove Suite granitoids.

Sound evidence in support of primary protoclastic texture, such as fractures in phenocrysts filled by unstrained quartz, and bent but unknicked biotite, is lacking in the Hillgrove Suite granitoids. Instead, all petrographic features point to the conclusion that the textures are essentially cataclastic (i.e. formed by deformation of solid rock). These features include subgrain development, rodding and streaking out of quartz, fracturing and dislocation of twin planes in feldspars and severe kinking and fraying of biotite.

Penetrative *s*-surfaces, defining a second (cataclastic) foliation, are clearly exhibited by some plutons, e.g. Abroi Granodiorite. This secondary foliation can also be found superimposed on the original foliation or schistosity of xenoliths occurring in the granitoids. In the case of large xenoliths the intersection angle between the original and the secondary foliation in the xenoliths can vary widely, principally because large xenoliths do not commonly parallel the early foliation direction. In fact the early foliation may be often either truncated at the xenolith margin or alternatively it may wrap around the xenolith.

The development of a tectonite fabric is particularly evident in the Kilburnie, Hillgrove and Enmore Adamellites. All these plutons show considerable strain, yet the degree of foliation is poor. Foliation is defined here, not by the alignment and parallelism of biotite flakes, but by stringers of biotite moulded around the felsic minerals. The direction of this foliation and resultant protomylonitic fabric always parallels the direction of the major faults.

Flood (1971) demonstrated a progressive increase in cataclasis in the Kilburnie Adamellite with increasing proximity to the Kilburnie Fault by studying changes in the structural state of alkali feldspars. Along the western margin of this pluton where deformation was very mild, alkali feldspar obliquities are low; however, close to the eastern margin, where the body is truncated by the N-S trending Kilburnie Fault, cataclasis was intense and feldspar obliquities approach maximum values ($\Delta \sim 0.90 - 0.95$).

It is highly probable that the Hillgrove Suite granitoids were deformed while they were warm. This is supported by (a) the recovery and subsolidus recrystallization of quartz grains in the majority of plutons and (b) more importantly, the minimal degree to which other minerals have recrystallized, biotite in particular.

In many of these granitoids it is common to find polygonal quartz aggregates where individual grains meet at $\sim 120^\circ$ and show no evidence of strain, yet associated biotite shows severe distortion and kinking. As will be shown in section 3.7, isotopic studies indicate that interpretations made on the basis of textural data as above may be substantially incorrect and that in all probability these granitoids were heated to at least 300°C after consolidation and emplacement.

3.4 PETROGRAPHY OF THE HILLGROVE SUITE GRANITOIDS

3.4.1 Classification

The classification adopted in this thesis for all intrusive rocks is basically that recommended by the IUGS Subcommittee on the Systematics of Igneous Rocks (Streckeisen *et al.*, 1973). However, there is one notable exception, the retention of the adamellite field. The adamellite field is particularly relevant in the case of the Hillgrove Suite granitoids since the major mineralogical variation in this suite is in the relative proportions of the feldspars. The term trondhjemite has been retained for low-K granitoids which are associated in the field with mafic rocks, are essentially devoid of alkali feldspar, have colour indices between 0 and 10 and may have andesine as the plagioclase feldspar.

3.4.2 Textures and Descriptive Petrography

The adamellites and granodiorites of the Hillgrove Suite are typically medium-grained (1-3 mm), have a hypidiomorphic inequigranular texture and consist essentially of quartz, alkali feldspar, plagioclase, biotite and ilmenite. Amphibole, garnet and muscovite are significant but volumetrically minor additional phases. Apatite, zircon, monazite and more rarely sphene and allanite are the principal accessories. In the adamellites the felsic minerals are in approximately equal proportions, while in the granodiorites the feldspar content is expectedly weighted towards plagioclase.

The felsic variants of the granitoids have allotriomorphic equigranular to interlobate textures, average 0.5 - 2.0 mm in grain size, and consist essentially of quartz (35-40%) and perthitic alkali feldspar (45-55%). Minor sodic plagioclase and iron-rich biotite are the remaining phases.

The plagioclase of the Hillgrove Suite granitoids is generally subhedral, more-or-less equidimensional and shows a wide range in grain size (0.2 - 5.0 mm). Two generations can usually be recognized in the majority of samples. The first shows incipient submicroscopic clouding, has a calcic composition (An_{35-47}), and occurs either as broken (or fragmented) crystals or as cores to the second generation of plagioclase. The latter is usually clear, anhedral and relatively sodic (An_{20-35}). Albite twinning predominates in both generations. However, combined albite-carlsbad twinning with blurred, bent and displaced twin planes are not uncommon in the early generation. Patchy and oscillatory zoning and albite overgrowths are very rare.

Unlike the alkali feldspar of the massive New England Suite granitoids, which is invariably orthoclase, the alkali feldspar of the Hillgrove Suite is always microcline. It occurs both as small "interstitial" crystals between larger grains of plagioclase and quartz and as conspicuous, generally anhedral crystals which tend to enclose other minerals. It is characterized by ubiquitous cross-hatched twinning, perthitic exsolution and frequently, considerable strain. The exsolution of albite in these microclines appears to be more prevalent in rocks which are notably deformed, as for example in granitoids truncated by, or in close proximity to, faults. Where developed, alkali feldspar phenocrysts are whitish, rarely larger than 2 cm in length and sparsely distributed throughout the rock.

Biotite (*mg* 35-40) is the dominant and sometimes sole ferromagnesian phase in the Hillgrove Suite plutons. It occurs most frequently as ragged, bent and kinked grains (0.5 - 3.0 mm in diameter), enclosing stumpy inclusions of apatite, zircon and monazite and sometimes displaying abundant acicular, presumably exsolved ?rutile needles (Plate 3.1). Deeply-coloured pleochroic haloes surround the inclusions. More rarely, sphene occurs as thin rods along the biotite cleavage. The biotite is usually fresh, despite its high degree of deformation. Minor incipient chloritization is, in fact, found more readily along some of the cleavage traces in *less* deformed samples. The colour of the biotite is highly distinctive and ranges from medium-brown with a reddish tinge to deep reddish-brown. In this respect alone these biotites differ grossly from biotites of massive granitoid suites from the Tasman Orogenic Province (e.g. Chappell, 1978; Hine *et al.*, 1978; Griffin *et al.*, 1978).

In addition to the reddish-brown variety of biotite, granitoids belonging to the Hillgrove Suite characteristically contain a green biotite. This green



Plate 3.1 Deep reddish-brown biotite, characteristic of Hillgrove Suite granitoids, showing exsolution of abundant ? rutile needles. 40 x.

variety, which is always more magnesian (*mg* 38-45) than the coexisting reddish-brown biotite, occurs throughout the granitoids as small flakes and stringers in apparent continuity with the wispy and frayed extremities of strongly deformed reddish-brown biotite. Less commonly it is located at the grain boundaries between felsic minerals. Although the green biotite is most probably secondary (i.e. post-crystallization) there appears to be no obvious genetic relationship between it and the small amount of muscovite present in most Hillgrove Suite granitoids. Indeed, there is a suggestion of an inverse correlation between these minerals in most samples.

Amphibole occurs in most of the Hillgrove Suite plutons. Its presence here is unusual because amphibole is characteristically absent from S-type granitoids (Chappell and White, 1974; Chappell, 1978). Fe-Mg and calciferous amphiboles both occur in these granitoids, though they rarely coexist. The pale greenish to pale greenish-brown calciferous amphibole is always actinolitic and more magnesian (*mg* 37-58) than the coexisting turbid, brownish cummingtonite (*mg* 38-44). Both amphibole varieties are scattered sporadically throughout the granitoids as either discrete well-formed grains or in mafic clots where their habit is typically ragged and fibrous.

Garnet is a minor constituent in nearly all Hillgrove Suite plutons. Crystals are usually pinkish and xenoblastic although clusters of euhedral garnets are not uncommon (e.g. Blue Knobby Adamellite). Garnet compositions are variable and apparently largely independent of bulk rock compositions. Most are predominantly almandine (> 80 mol.%); however, outer rims may be strongly enriched in the spessartine end-member (up to 40 mol.%).

Manganian ilmenite, containing 3-11% MnO, is the only iron-titanium oxide present in these granitoids. It is usually stumpy rod-like in shape with prominent embayments, and is most frequently enclosed by reddish-brown biotite.

A final and very significant petrographic feature of the Hillgrove Suite is the ubiquitous occurrence of two varieties of quartz, namely blue and colourless. Although the overall bluish-grey colour of the granitoids is derived directly from the blue quartz, the clear colourless variety predominates and occurs as small anhedral grains "interstitial" to the relatively coarse plagioclase and blue quartz. Earlier explanations of the coloured quartz were based largely on the nature of inclusions, usually thought to be rutile (e.g. Binns, 1966). However, Nassau and Prescott (1977) have shown that the bluish colouration of quartz is probably a colour centre which is due to either, or both, radiochemical damage or crystal lattice defects. Since quartz was a

likely residual phase during granitoid formation it is suggested that the blue quartz may be a restite phase from the ultrametamorphic episode which culminated in the generation of the Hillgrove Suite granitoids whereas the colourless variety is interpreted as having crystallized directly from the melt.

Unlike many other S-type granitoids from the Tasman Orogenic Province [e.g. Bundarra Suite (Flood and Shaw, 1975); Kosciusko Batholith (Griffin *et al.*, 1978)], the Hillgrove Suite plutons do not carry cordierite, primary muscovite nor Al_2SiO_5 polymorphs. The absence of these phases generally indicates relatively low-Al sedimentary parent rock. However, considering the moderately aluminous biotite and the not uncommon presence of garnet it is concluded that a delicate balance with respect to Al prevailed during ultrametamorphism in both the source rocks and the resultant granitoids.

3.4.3 Xenoliths

The majority of xenoliths occurring in the Hillgrove Suite granitoids are essentially extensions of their hosts in terms of mineralogy and petrography. Most are psammo-pelitic (greywacke) and although finer-grained than their hosts they have identical textures, i.e. hypidiomorphic inequigranular. The main difference is that the xenoliths usually have higher colour indices than the respective hosts, reflecting the higher modal abundance of biotite. Relict sedimentary features (e.g. bedding) are commonly preserved in the xenoliths despite extensive recrystallization and development of a prominent metamorphic fabric. Porphyroblast development is extremely rare and there has been no obvious reaction between these xenoliths and their granitoid hosts.

Less common xenoliths include impure cherts and metabasalts. Their textures indicate that both types of xenoliths have been thoroughly recrystallized. In addition, there is strong evidence that metasomatism, replacement and retrograde alteration accompanied recrystallization in the mafic xenoliths.

The distribution of xenoliths in the Hillgrove Suite plutons is not systematic. This fact, combined with the apparent absence of mass reaction between xenoliths and hosts and their commonly nebulous outline, suggest that these xenoliths* are remnants of country rock which survived ultrametamorphism and partial melting during the formation of the Hillgrove Suite.

* Some authors (e.g. Stephenson, 1973, 1974) prefer to use the term skialith for this type of intrusion.

3.5 MINERALOGY

3.5.1 Biotite

Biotite is a ubiquitous and sometimes sole ferromagnesian phase in the Hillgrove Suite granitoids. Of the two distinct but often coexisting varieties, namely reddish-brown and green, the former is clearly dominant.

3.5.1.1 Reddish-brown Biotite

(a) Major Elements. Representative analyses of reddish-brown biotites from all Hillgrove Suite plutons are listed in Table 3.1 together with structural formulae calculated on the basis of 22 oxygens (anhydrous). Also listed are the results of ferrous iron and trace element measurements carried out on biotite separates used for Sr isotopic determinations.

Although all these biotites are broadly similar in composition there are significant variations among specific chemical parameters (e.g. Al^{vi} ranges from 0.18 to 0.85 atoms per formula unit). The principal chemical variations of these biotites are tabulated below and compared with biotites from other S-type granitoids of the New England and Murrumbidgee Batholiths. Also included in Table 3.2 are selected compositional ranges for biotites from the New England Suite plutons and from high-grade psammo-pelitic metamorphic rocks from the Wongwibinda Complex.

Table 3.2 Ranges of selected chemical parameters for reddish-brown biotites from the Hillgrove Suite granitoids compared with biotites from other granitoids and related metamorphic rocks.

	Hillgrove Suite	Bundarra Suite	Murrumbidgee Batholith (Joyce, 1973)	Wongwibinda high-grade metamorphics	New England Suite
Al^{iv} *	2.24 - 2.69	2.32 - 2.62	2.30 - 2.68	2.44 - 2.81	2.10 - 2.51
Al^{vi} *	0.18 - 0.85	0.13 - 0.34	0.26 - 0.75	0.44 - 1.05	0 - 0.44
mg	29.5 - 42.7	34.2 - 38.1	38.9 - 52.6	27.1 - 57.4	28.4 - 71.4
Ti(max)*	0.56	0.51	0.46	0.53	0.68
Mn(max)*	0.07	0.05	0.06	0.07	0.09
Fe^{3+}/Fe^{2+}	0.02 - 0.13	-	0.02 - 0.17	0.03 - 0.12	0.05 - 0.20

* cations per formula unit

Table 3.1 Electron microprobe and trace element analyses and structural formulae of representative reddish-brown biotites from the Hillgrove Suite. Two analyses of biotites from the Bundarra Suite are also included for comparison.

Sample No. †	(1) HR	(2) K14	(3) TOB	(4) AB	(5) RVGD core	(6) RVGD mid	(7) RVGD rim	(8) RVT7	(9) RVT7 xen	(10) RVT10	(11) RVT10	(12) HGA core	(13) HGA rim	(14) BK	(15) ENA
Catalogue No.	46026	46027	46030	46032	46036	46036	46036	46039	46039	46038	46038	46040	46040	46053	46054
SiO ₂	36.2	35.4	34.5	35.3	36.4	35.7	35.8	36.3	34.7	35.0	36.4	34.3	35.1	35.8	35.7
TiO ₂	3.12	3.28	3.24	2.96	3.73	3.78	.51	3.74	3.16	4.16	4.13	4.59	1.50	3.62	3.5
Al ₂ O ₃	15.7	14.7	16.8	16.6	16.8	17.0	19.0	16.4	18.3	16.1	13.8	15.4	17.7	16.6	16.0
FeO*	23.2	24.2	22.2	22.3	21.5	21.0	20.2	20.3	21.0	22.3	24.2	23.0	22.4	21.6	22.5
MnO	.30	.28	.31	.40	.21	.19	.21	.22	.26	.15	.16	.55	.60	.36	.23
MgO	7.33	7.65	7.90	8.13	8.36	8.77	9.24	8.27	7.79	7.30	7.63	7.70	8.60	7.60	7.76
CaO	.11	.20	.18	.19	-	-	-	-	-	-	-	.11	.16	-	-
Na ₂ O	-	.20	-	-	-	-	-	.27	.25	.30	-	-	.19	.40	-
K ₂ O	8.98	8.96	9.29	9.23	9.89	9.60	9.24	9.29	9.54	8.93	9.59	9.31	8.73	9.26	9.23
Total	94.90	94.83	94.39	94.08	96.93	94.98	94.56	94.83	93.72	94.21	95.83	92.44	91.73	95.45	94.86
mg	36.1	36.1	38.9	39.4	40.9	42.7	54.8	42.1	39.8	36.9	36.0	37.4	40.6	39.7	38.1
<i>Structural formulae (based on 22 oxygens per formula unit)</i>															
Si	5.638	5.580	5.413	5.485	5.517	5.505	5.492	5.591	5.367	5.483	5.667	5.383	5.442	5.516	5.554
Al ^{IV}	2.362	2.420	2.587	2.515	2.483	2.495	2.508	2.409	2.633	2.517	2.333	2.617	2.558	2.484	2.446
Ti	.362	.389	.382	.347	.426	.323	.059	.433	.368	.490	.484	.542	.175	.419	.408
Al ^{VI}	.524	.303	.520	.521	.523	.605	.953	.566	.704	.464	.203	.226	.673	.527	.491
Fe	3.016	3.181	2.905	2.898	2.731	2.705	2.588	2.614	2.716	2.919	3.148	3.018	2.902	2.780	2.924
Mn	.039	.037	.041	.052	.028	.026	.028	.028	.035	.020	.021	.074	.078	.035	.030
Mg	1.700	1.795	1.846	1.885	1.890	2.018	2.184	1.897	1.795	1.705	1.773	1.799	1.983	1.829	1.801
Ca	.019	.034	.030	.031	-	-	-	-	-	-	-	.019	.026	-	-
Na	-	.062	-	-	-	-	-	.081	.076	.090	-	-	.055	.119	-
K	1.782	1.799	1.857	1.832	1.913	1.891	1.810	1.823	1.882	1.786	1.906	1.863	1.726	1.821	1.833
<i>Σ (cations)</i>	15.445	15.559	15.580	15.566	15.510	15.567	15.623	15.441	15.575	15.475	15.534	15.570	15.644	15.529	15.486
<i>F</i>	23.00	23.95	22.05	22.11		21.40				23.50		23.00		22.05	23.05
FeO	21.15	21.75	20.99	21.25		20.70				23.05		21.55		21.35	20.35
Fe ₂ O ₃	2.06	2.45	1.16	0.96		0.78				0.50		1.61		0.78	3.00
<i>Trace Elements (ppm)</i>															
Zn	385	415	350	350		315				300		345		375	195
Cr	190	200	210	210		220				170		290		220	90
Ni	70	70	140	70		100				80		75		120	70
Cu	60	54	22	18		24				26		22		36	50
Co	65	65	40	45		45				55		50		40	10
Li	204	156	196	148		172				200		180		160	148

† See Appendix I

* Total Fe as FeO

F estimated average total Fe as FeO from all microprobe analyses

Table 3.1 cont. Electron microprobe and trace element analyses and structural formulae of representative reddish-brown biotites from the Hillgrove Suite. Two analyses of biotites from the Bundarra Suite are also included for comparison.

Sample No.†	(16) WOLL core	(17) WOLL rim	(18) WINT	(19) KA	(20) KILB	(21) INGA	(22) ARG	(23) ARG xen	(24) GK6	(25) TM4	(26) TM2	(27) DUND	(28) DUND xen	(29) BS	(30) BS
Catalogue No.	46035	46035	46060	46061	46062	46063	46064	46064	46071	46069	46068	46049	46049	46074	46074
SiO ₂	35.6	36.2	34.6	35.1	37.0	36.1	36.1	36.9	35.5	34.3	35.7	34.7	35.4	33.8	36.1
TiO ₂	3.45	1.99	3.88	4.24	4.79	3.54	3.46	1.97	3.05	2.82	3.18	4.43	2.31	3.69	4.26
Al ₂ O ₃	17.0	18.3	15.2	15.0	13.6	16.3	15.4	16.9	15.9	17.6	16.4	14.1	14.9	15.8	13.1
FeO*	22.4	21.4	24.1	23.2	24.2	24.3	21.7	20.4	23.5	21.2	20.9	23.4	23.8	24.2	23.5
MnO	.17	.30	.25	.22	.21	.13	.73	.96	.34	.35	.35	.53	.75	.38	.40
MgO	7.17	7.46	7.52	7.61	6.50	5.68	9.08	9.44	7.29	8.11	8.71	7.18	8.19	7.06	8.09
CaO	.16	.19	.29	-	.16	.24	.35	.47	.17	.16	-	-	-	-	-
Na ₂ O	-	-	-	.25	-	-	-	-	.52	-	.32	-	-	-	-
K ₂ O	9.65	10.00	8.52	9.31	9.08	8.75	9.28	9.24	9.60	9.68	9.72	9.38	9.70	9.52	9.01
Total	95.59	95.84	94.37	97.22	95.55	92.11	96.11	96.27	95.87	94.23	95.28	94.73	95.05	94.45	94.46
mg	36.3	38.3	35.8	36.9	32.4	29.5	42.7	45.2	35.6	40.5	42.6	34.4	38.0	34.2	38.1
<i>Structural formulae (based on 22 oxygens per formula unit)</i>															
Si	5.502	5.544	5.466	5.506	5.761	5.637	5.535	5.590	5.526	5.369	5.515	5.507	5.576	5.381	5.688
Al ^{iv}	2.498	2.456	2.534	2.494	2.239	2.363	2.465	2.410	2.474	2.631	2.485	2.493	2.424	2.619	2.312
Ti	.401	.229	.461	.500	.561	.415	.399	.225	.357	.333	.370	.529	.274	.442	.505
Al ^{vi}	.603	.845	.300	.285	.257	.624	.324	.612	.446	.622	.499	.153	.350	.338	.124
Fe	2.892	2.747	3.183	3.043	3.146	3.165	2.781	2.578	3.057	2.780	2.701	3.238	3.138	3.224	3.097
Mn	.032	.039	.033	.029	.027	.018	.095	.123	.045	.047	.046	.075	.101	.052	.054
Mg	1.651	1.706	1.772	1.779	1.508	1.322	2.073	2.131	1.691	1.895	2.004	1.700	1.925	1.676	1.903
Ca	.026	.032	.048	-	.027	.039	.058	.076	.028	-	-	-	-	-	-
Na	-	-	-	.078	-	-	-	-	.157	-	.096	-	-	-	-
K	1.902	1.956	1.718	1.861	1.804	1.742	1.813	1.785	1.906	1.935	1.913	1.901	1.951	1.934	1.814
Σ (cations)	15.498	14.554	15.515	15.574	15.331	15.325	15.543	15.530	15.688	15.639	15.628	15.592	15.738	15.665	15.436
F	22.15			22.60		24.45	20.95				20.75	24.05			
FeO	21.70			21.45		22.45	20.55				20.15	23.70			
Fe ₂ O ₃	0.50			1.28		2.22	0.44				0.68	0.39			
<i>Trace Elements (ppm)</i>															
Zn	385		185	280		465	380				375	360			
Cr	180		190	250		90	220				220	130			
Ni	70		150	65		30	120				70	40			
Cu	34		22	20		22	40				32	20			
Co	65		65	65		30	45				60	45			
Li	188		96	174		896	178				232	180			

† See Appendix I

* Total Fe as FeO

P estimated average total Fe as FeO from all microprobe analyses

Significant chemical variations within individual biotite grains are few; however, some biotites [e.g. Rockvale Granodiorite (anal. 5-7) and Hillgrove Adamellite (anal. 12-13)] show pronounced compositional zoning, especially Ti. This depletion of Ti from core to rim probably reflects the non-availability of Ti at a stage when most of the Ti presumably had been partitioned into ilmenite, whose crystallization range was in all likelihood less than that of the coexisting biotite.

Perhaps the best illustration of compositional variation in biotite, in terms of the principal components, is by a Mg-Al^{vi}-(Fe + Mn) plot. Figure 3.2 shows the reddish-brown Hillgrove Suite biotites occupying a large elliptical field, separate from the coexisting green biotites, the Wongwibinda metamorphic biotites and the biotites from the New England Suite. The relationship between Mg, Fe and Al^{vi} in biotite can be best understood from ionic considerations and charge balance requirements (Foster, 1960). Foster showed that where trivalent (Fe³⁺, Al^{vi}) cations and Ti⁴⁺ substitute for divalent (Fe²⁺, Mg²⁺) cations in the octahedral layer, a charge compensation is required. This takes two forms, both operating simultaneously; (a) Al substituting for Si in the tetrahedral layer and (b) the development of vacancies in the octahedral layers.

The distribution of Al^{iv} and Al^{vi} in biotite is generally attributed to the nature of the coexisting aluminous phases (Nockolds, 1947; Harry, 1950; Foster, 1960; Haslam, 1968; Joyce, 1973; Albuquerque, 1973). The variation of Al^{iv} and Al^{vi} in the Hillgrove Suite biotites is basically consistent with this generalization, i.e. Al^{iv} and Al^{vi} are low when there is coexisting amphibole and moderately high when there is coexisting muscovite. However, instead of displaying high to very high Al^{iv} and Al^{vi} when there is coexisting garnet, as suggested by the above authors, Hillgrove Suite biotites show only moderate Al^{iv} and Al^{vi}.

On an Al^{iv} v. Al^{vi} plot (Fig. 3.3) the reddish-brown biotites from the Hillgrove Suite occupy an extensive field which shows only a limited overlap with biotites from the New England Suite. The difference between these two fields may be attributed primarily to the preferential partitioning of Al^{vi} into coexisting amphibole during crystallization of the New England Suite granitoids. Also evident from Figure 3.3 is the large overlap between the reddish-brown and the coexisting green biotites from the Hillgrove Suite.

Biotite from a seemingly typical sample of the S-type Bundarra Suite was also analyzed and compared to the reddish-brown Hillgrove Suite biotites - it was low in Al^{vi}, ranging from 0.13 to 0.34 atoms per formula unit.

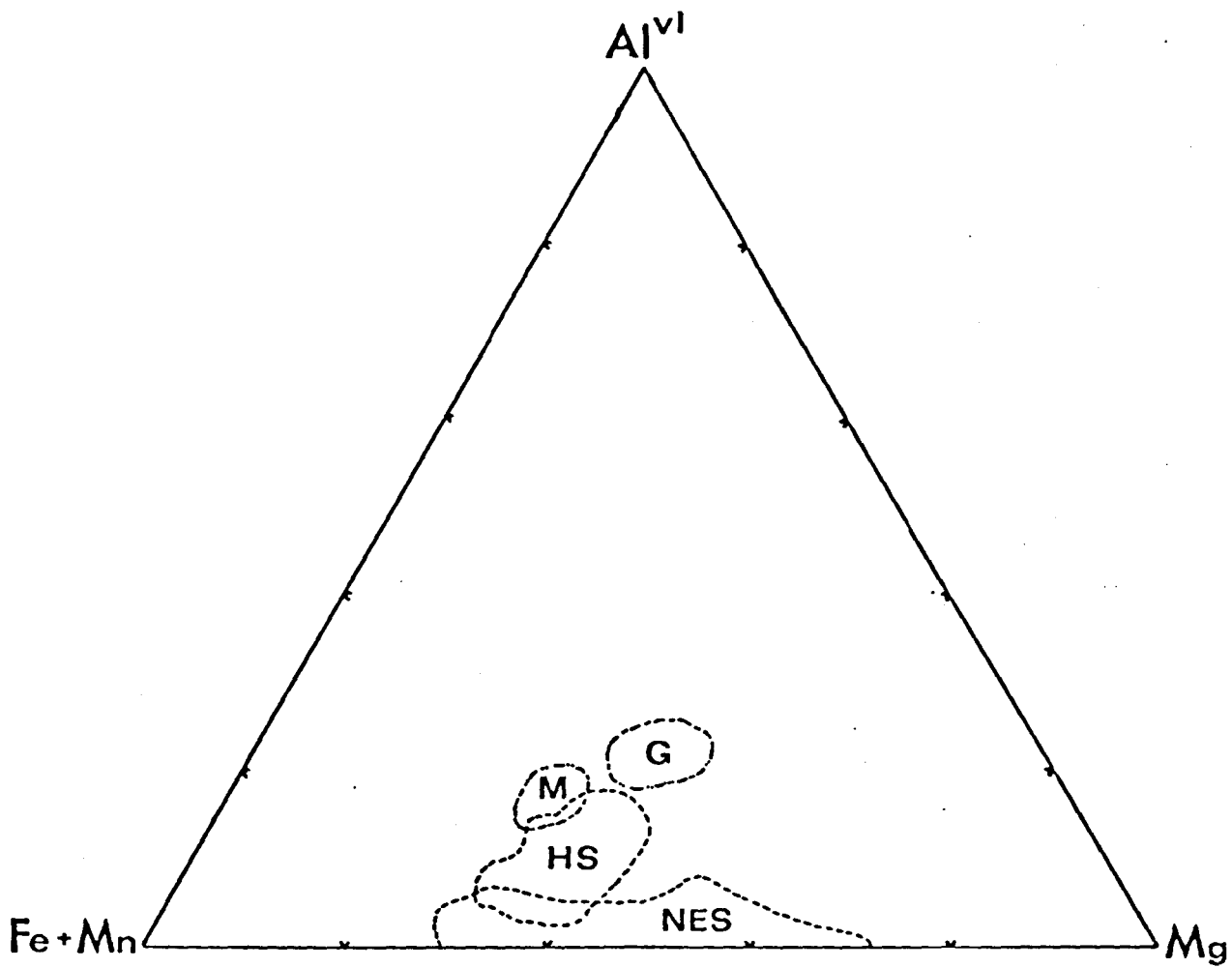


Figure 3.2 Plot of $Mg-Al^{VI}-(Fe+Mn)$ for reddish-brown biotites from the Hillgrove Suite granitoids. Also included are fields occupied by coexisting green biotites (G), biotites from high-grade metamorphic rocks from the Wongwibinda Complex (M) and biotites from plutons of the New England Suite (NES).

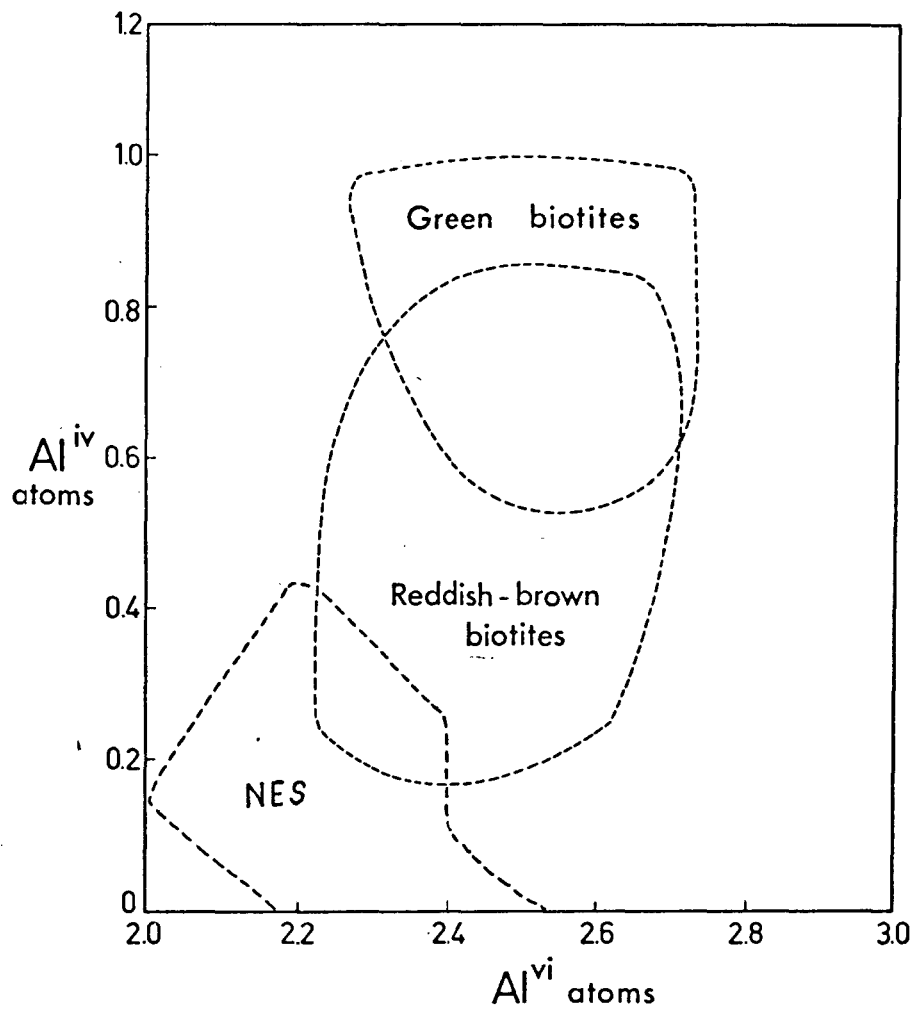


Figure 3.3 Plot of Al^{iv} against Al^{vi} for Hillgrove Suite reddish-brown biotites, showing for comparison the fields occupied by co-existing green biotites and biotites from the New England Suite (NES).

These low values are surprising in view of the coexistence of large crystals of cordierite (now pinitized), minor garnet and muscovite, and contrast with values of Al^{vi} reported by O'Neil *et al.* (1977) for Bundarra Suite biotites. A source for this discrepancy may have been alluded to by Chappell (1978) who suggested that some of the biotite and much of the coexisting muscovite were probably due to the alteration of cordierite. Consequently, mineral separates, such as those of O'Neil *et al.* may show excess Al^{vi} .

(b) Trace Elements. Table 3.3 compares the ranges of Zn, Cr, Ni, Cu, Co, Li, Rb and Sr for the reddish brown biotites from the Hillgrove Suite with biotites from other S-type granitoids and from high-grade metasediments from the Wongwibinda Complex.

Table 3.3 Trace element abundances for biotites.

Element (ppm)	Hillgrove Suite	Bundarra Suite	Murrumbidgee Batholith (Joyce, 1973)	Wongwibinda Metamorphics
Zn	185 - 465	250	215 - 444 (855)*	190 - 580
Cr	90 - 290	60	73 - 476	20 - 330
Ni	30 - 150	40	24 - 119	45 - 200
Cu	18 - 60	34	1 - 121	20 - 32
Co	10 - 65	40	39 - 59	30 - 90
Li	96 - 896	688	70 - 350	40 - 672
Rb	516 - 1054	1013	443 - 901	409 - 990
Sr	5 - 43	5	4 - 23	7 - 47

Trace element abundances for these biotites show a surprisingly large variation, considering the limited variation in the major element and host rock compositions. However much of this large variation is due to the felsic Ingleba Adamellite, being enriched in Zn, Li and Rb. Indeed the chemical similarity between the biotites from the Ingleba Adamellite and plutons belonging to the S-type Bundarra Suite is striking. The similarity between the remaining Hillgrove Suite biotites and the 'average' biotite from the Murrumbidgee Batholith (Joyce, 1973) is also very close.

* One anomalously high value.

Somewhat surprising are the high Cr and Ni values for the Hillgrove Suite biotites. This may reflect the relatively immature nature of the volcanogenic detritus from which the Hillgrove Suite granitoids are thought to have been derived. As a comparison, biotites from the high-grade psammpelitic metamorphic rocks at Wongwibinda were also analyzed, since these metasediments were clearly immature greywackes before reconstitution. Table 3.3 shows that there is nearly a complete overlap between the biotites from the Hillgrove Suite and those from the metamorphic rocks. Thus, on biotite trace element abundances alone any hypothesis relating to the generation of the Hillgrove Suite from their surrounding high-grade metamorphic rocks or less-metamorphic equivalents is acceptable.

3.5.1.2 Green Biotite

From the analyses presented in Table 3.4 it is clear that the green biotites from the Hillgrove Suite plutons are chemically distinct from the dominant reddish-brown biotite. Compared to the latter the green biotites contain significantly lower Ti, notably higher Al^{vi} and *mg* values, and slightly but consistently higher Mn, Ca and Na.

It was not possible to determine the Fe^{3+}/Fe^{2+} ratios of the green biotites because of difficulties in mineral separation. However, the colour suggests that these biotites probably have high Fe^{3+}/Fe^{2+} ratios (Deer *et al.*, 1962). This conclusion is based partly on theoretical and quantitative experimental data (Osborn, 1962; Wones and Eugster, 1965; Best and Mercy, 1967) which indicate that if crystallization is not buffered increasing P_{H_2O} in the dwindling melt may yield increased fO_2 . The likelihood that the crystallization of green biotite may have been unbuffered is suggested by its occurrence at the deformed extremities of the reddish-brown biotite and along quartz-alkali feldspar grain boundaries.

The low Ti and the relatively high Al^{vi} , Mn, Na and Ca contents of the green biotite also suggest late, or even subsolidus, crystallization. Late in the magmatic history of a granitoid Al activity may often be enhanced relative to Si (Albuquerque, 1973). This generally results in an increase in Al^{vi} , particularly since there is no competition from Ti. Similarly, the activity of Mn also usually increases slightly under late-stage magmatic conditions, as shown by the zoning in garnet (see 3.5.5) and the rim compositions of the large reddish-brown biotite. Hence, Mn contents of the green biotites are usually higher than for the reddish-brown variety.

3.5.1.3 Biotite from Felsic Hillgrove Suite Variants

Analytical data for two typical samples are presented in Table 3.5. Both samples come from highly felsic rocks which occur as segregations in two large granitoids, namely the Tia Granodiorite and Hillgrove Adamellite. Neither sample contains muscovite and only the Tia Granodiorite contains garnet. The samples have been included in this study to illustrate the effect of fractionation on biotites precipitated from granitic melts.

Compared to the reddish-brown biotite from the main plutons, the biotites from the felsic samples are characterized by lower Ti and *mg* values and higher Al^{vi}, Mn, Fe and Na. The relatively lower Ti contents and *mg* values are consistent with biotites crystallized late in differentiated rock series, e.g. the Guadelupe Complex (Best and Mercy, 1967) and Murrumbidgee Batholith (Joyce, 1973), and simply reflect changes in the compositions of successive liquid fractions.

Table 3.5 Electron microprobe analyses and structural formulae of biotites from felsic Hillgrove Suite variants.

Sample number Catalogue number	(1) LTM2 46070	(2) LTM2 46070	(3) LHGA 46046	(4) LHGA 46046
SiO ₂	34.5	35.6	37.2	35.1
TiO ₂	2.61	1.59	1.77	.74
Al ₂ O ₃	17.3	17.4	20.9	18.3
FeO*	23.0	23.3	23.2	28.1
MnO	.59	.63	.20	.39
MgO	6.82	7.95	4.51	4.04
CaO	-	-	.11	-
Na ₂ O	.23	.32	.59	.41
K ₂ O	9.50	9.65	7.94	9.23
Total	94.55	96.49	96.42	96.21
<i>mg</i>	34.6	37.9	25.7	20.4
<i>Structural formulae (based on 22 oxygens per formula unit)</i>				
Si	5.435	5.492	5.597	5.529
Al ^{iv}	2.565	2.508	2.403	2.471
Ti	.309	.185	.201	.088
Al ^{vi}	.647	.655	1.314	.919
Fe	3.027	2.997	2.923	3.693
Mn	.079	.083	.025	.051
Mg	1.599	1.827	1.013	.947
Ca	-	-	.018	-
Na	.070	.095	.172	.126
K	1.908	1.897	1.526	1.853
Σ (cations)	15.639	15.738	15.192	15.677

* Total Fe as FeO.

1,2 Tia Granodiorite;

3.5.2 Amphibole

According to Chappell and White (1974) calciferous amphiboles are usually absent from, or only a very minor phase, in S-type granitoids. The occurrence of Fe-Mg and calciferous amphiboles in the Hillgrove Suite is therefore significant, suggesting that either the Hillgrove Suite plutons crystallized under P/T conditions different from other S-type granitoids, i.e. within the stability field of amphibole, or that the bulk composition of the source rocks differed from that of the sedimentary source rocks in the Lachlan Fold Belt, hence producing chemically different partial melts.

Structural formulae were calculated on an anhydrous basis of 23 oxygens. This method was preferred to recalculation on the basis of either 13 or 15 cations (e.g. Stout, 1972) because of the inherent difficulty of assigning the cations to their correct structural sites.

3.5.2.1 Cummingtonite

(a) Occurrence and Chemistry. Cummingtonite occurs in only four of the Hillgrove Suite plutons. Its distribution and habit vary from pluton to pluton; it may occur as discrete crystals, as fine lamellar intergrowths with actinolite (e.g. Kilburnie Adamellite), or as a felted mass replacing ferrohypersthene (e.g. Enmore Adamellite). Its modal abundance never exceeds 1% and its colour ranges from pale greenish-brown to very pale green.

Chemical analyses of three Hillgrove Suite cummingtonites are presented in Table 3.6. Except for the cummingtonite from the Argyll Granodiorite, compositions are typical of cummingtonites from igneous rocks (Deer *et al.*, 1962). Also typical are their *mg* values which are always slightly lower than the coexisting calciferous amphibole. The amphibole from the Argyll Granodiorite is usually enriched in MnO (up to 8.1%) and compares favourably with analyses of dannemorites (manganoan cummingtonite; Deer *et al.*, 1962). Cummingtonite also occurs in the Kilburnie Adamellite; however, because it is finely intergrown with actinolite, an acceptable analysis could not be obtained.

3.5.2.2 Calciferous Amphibole

(a) Occurrence and Chemistry. Although calciferous amphiboles are more widely distributed among the Hillgrove Suite plutons than cummingtonite, their abundance in specific rocks never exceeds 2%. They are usually pale greenish, only weakly pleochroic, and generally occur as small ragged crystals.

Table 3.6 Electron microprobe analyses and structural formulae of cummingtonites from the Hillgrove Suite.

	(1)	(2)	(3)	(4)
<i>Sample number</i>	RVT10	ARG	ARG	DHZ-15
<i>Catalogue number</i>	46038	46064	46064	
SiO ₂	51.0	53.0	51.4	50.7
TiO ₂	.19	.16	.27	.06
Al ₂ O ₃	1.6	1.2	1.0	.88
Cr ₂ O ₃	-	.31	.31	-
FeO*	30.2	27.3	25.0	25.75
MnO	1.51	3.44	8.09	7.38
MgO	10.3	11.8	9.74	10.57
CaO	2.21	.79	1.07	2.00
Na ₂ O	.48	-	-	.22
K ₂ O	.11	-	.11	.08
Total	97.58	98.03	97.43	97.93
<i>mg</i>	37.8	43.6	40.9	36.1
<i>Structural formulae</i> (based on 23 oxygens per formula unit)				
Si	7.886	8.027	7.991	7.829
Al ^{iv}	.134	-	.009	.160
Ti	.022	.018	.032	.007
Al ^{vi}	.157	.214	.174	-
Cr	-	.037	.038	-
Fe	3.900	3.458	3.256	3.320
Mn	.197	.441	1.065	.964
Mg	2.363	2.675	2.257	2.430
Ca	.365	.128	.178	.331
Na	.144	-	-	.066
K	.022	-	.022	.016
Σ (cations)	15.170	14.998	15.022	15.123

* Total Fe as FeO

1 Rockvale Granodiorite (northern margin)

2,3 Manganoan cummingtonite from Argyll Granodiorite

4 Dannemorite (Deer *et al.*, 1962; anal. 15, Table 36)

Table 3.7 shows that with only one exception the calciferous amphiboles of the Hillgrove Suite are actinolite or actinolitic hornblende (after Leake, 1968). The exception is euhedral, olive-green to brownish-green ferro-hornblende which coexists with pale-green actinolite in the Kilburnie Adamellite. This ferrohornblende is notably lower in Si and Ca, but higher in Ti, Al, Na and K than the coexisting actinolite.

If it is accepted that amphibole compositions closely reflect bulk rock compositions as suggested by Nockolds and Mitchell (1948), Chappell (1966), Dodge *et al.* (1968), Haslam (1968), Rhodes (1969) and Albuquerque (1974), it is surprising that the calciferous amphiboles of the Hillgrove Suite show such a large variation in *mg* numbers (39.2 - 61.6) compared with the limited compositional variation of their host rocks (28.6 - 45.2). The large difference in the *mg* numbers of the coexisting hornblende (39.2) and actinolite (53.6) from the Kilburnie Adamellite is a clear example of disequilibrium. However, whether this is due to disequilibrium crystallization or to a metastable preservation of relict crystals cannot be resolved immediately.

To investigate the possibility that the Hillgrove Suite amphiboles may be "relict" from the ultrametamorphic period of melt generation, the high-grade metamorphic rocks surrounding the granitoids were examined. Surprisingly, none of the metagreywackes contained amphibole, despite the fact that P/T conditions attained during metamorphism were appropriate to upper amphibolite facies. However, small calcareous lenses and pods in the metamorphosed psammo-pelitic beds contain abundant amphibole. The sporadic distribution of these lenses could possibly be correlated with the sporadic distribution of amphibole in the granitoids. Thus, variations in precursor metamorphic bulk rock compositions could satisfactorily account for the wide compositional range of the igneous actinolites.

The fibrous and ragged crystal habit of the igneous actinolites suggests that they may be disequilibrium phases. An approach towards equilibrium between coexisting mineral phases, can often be verified by the distribution pattern of major elements, e.g. Fe/Mg. The range of K_D values for the Hillgrove Suite actinolite/biotite pairs (see Table 3.8) far exceeds the range of K_D values for most other rock suites. For example, Aregos pairs (Albuquerque, 1974) and Scottish Caledonian pairs (Nockolds and Mitchell, 1948) vary from 0.56 - 0.64 and 0.75 - 0.87 respectively. Apart from the Dundurrabin mineral pair, which comes from a microxenolith, the Hillgrove Suite mineral pairs appear to be

Table 3.7 Electron microprobe analyses and structural formulae of actinolites and hornblendes from the Hillgrove Suite.

Sample No. [†]	(1) HGA	(2) KA	(3) KA	(4) KILB	(5) KILB	(6) KILB	(7) DUNDX	(8) DUNDX	(9) ARG	(10) ARG	(11) RVT10
Catal.No.	46040	46061	46061	46062	46062	46062	46051	46051	46064	46064	46038
SiO ₂	54.1	52.8	54.5	53.0	47.1	52.7	51.2	51.4	54.4	52.3	48.4
TiO ₂	.15	-	-	.17	1.35	-	-	-	.15	.20	.86
Al ₂ O ₃	.8	.3	1.0	1.3	6.1	3.2	1.1	2.4	1.0	1.1	5.7
Cr ₂ O ₃	.41	-	-	-	.20	.24	-	-	.35	.45	.12
FeO [*]	16.2	19.0	15.4	23.1	22.1	17.2	21.8	19.4	16.6	20.2	21.0
MnO	.74	.50	.45	1.63	.69	.44	2.20	1.28	1.82	3.20	.47
MgO	13.3	12.3	13.8	10.7	8.0	11.1	10.6	10.0	12.1	11.8	9.3
CaO	11.8	11.3	11.8	7.7	10.5	11.6	9.2	11.2	10.9	7.8	10.5
Na ₂ O	-	.18	-	.16	1.33	.18	-	.33	-	-	1.28
K ₂ O	.10	-	.09	.15	.62	.81	-	.20	.08	.11	.60
Total	97.60	96.38	97.04	97.81	97.99	97.47	96.10	96.21	97.40	97.16	98.23
mg	59.4	53.6	61.6	45.2	39.2	53.6	46.5	47.9	56.5	50.9	44.2
<i>Structural formulae (based on 23 oxygens per formula unit)</i>											
Si	7.918	7.934	7.974	7.938	7.169	7.781	7.840	7.789	7.998	7.865	7.272
Al ^{iv}	.082	.052	.026	.062	.831	.219	.160	.211	.002	.135	.728
Ti	.016	-	-	.019	.155	-	-	-	.016	.022	.097
Al ^{vi}	.058	-	.142	.160	.263	.330	.045	.211	.172	.057	.281
Cr	.048	-	-	-	.024	.029	-	-	.041	.053	.014
Fe	1.986	2.384	1.877	2.889	2.808	2.124	2.789	2.463	2.041	2.543	2.639
Mn	.092	.064	.055	.207	.089	.054	.285	.165	.227	.407	.060
Mg	2.909	2.754	3.009	2.381	1.808	2.450	2.425	2.268	2.653	2.636	2.089
Ca	1.855	1.826	1.850	1.240	1.714	1.841	1.514	1.826	1.720	1.263	1.697
Na	-	.052	-	.046	.392	.052	-	.096	-	-	.372
K	.018	-	.017	.028	.121	.152	-	.038	.015	.021	.116
Σ	14.981	15.006	14.951	14.969	15.374	15.032	15.058	15.068	14.885	15.001	15.364
<i>(cations)</i>											

† See Appendix I.

* Total Fe as FeO.

bimodal, with a narrow range of K_D values for the adamellites and a similarly small, but different, range for the granodiorites.

Low K_D values have been regarded as indicating low-pressure crystallization at relatively low temperatures (e.g. Albuquerque, 1974). The low K_D 's of the adamellite mineral pairs (0.42 ± 0.01) could therefore be interpreted as due to late-stage crystallization at shallow crustal levels, whereas the granodiorite pairs ($K_D = 0.73 \pm 0.01$) could represent early attainment of equilibrium at the site of melting.

Table 3.8 Distribution coefficients ($K_D^{Fe/Mg}$) for actinolite/biotite mineral pairs from the Hillgrove Suite granitoids.

	Fe/Mg_{Bt}	Fe/Mg_{Act}	$K_D^{Fe/Mg}_{Act/Bt}$
Rockvale Granodiorite	1.71	1.26	0.74
Kilburnie Adamellite (fine-grained)	2.09	0.87	0.42
Kilburnie Adamellite (coarse-grained)	1.71	0.87	0.43 (av.)
Hillgrove Adamellite	1.68	0.68	0.41
Argyll Granodiorite	1.34	0.97	0.72
Dundurrabin Granodiorite	1.90	1.09	0.57

3.5.3 Orthopyroxene

(a) Occurrence and Chemistry. The only occurrence of pyroxene in the Hillgrove Suite granitoids is in an extremely fresh, fine-grained variant of the Enmore Adamellite. The surface of this rock is pitted, due to the preferential weathering of mineral aggregates containing large orthopyroxenes surrounded by a felted mass of cummingtonite. It is not certain from textures whether the cummingtonite has partially replaced the orthopyroxene or if the orthopyroxene has crystallized from the breakdown of cummingtonite.

Analyses of orthopyroxenes are listed in Table 3.9. Total cations in the structural formulae exceed the ideal 4.00 which suggests a significant Fe^{3+} content. Despite an extensive search in the literature few orthopyroxenes have been recorded in felsic orogenic granitoids. However, the Hillgrove Suite orthopyroxenes are not dissimilar from orthopyroxenes occurring in charnockites from Antarctica (Blight, 1975), U.S.A. (Bohlen and Essene, 1978) and South Australia (G. Mortimer, *pers. comm.*) and in central Australian granulites (Woodford and Wilson, 1976). Compared to these, the Hillgrove Suite orthopyroxenes are low in Al_2O_3 (1.2 - 2.1%) and very low in CaO (< 0.21%) suggesting relatively low temperatures and pressures during their crystallization. On the other hand, TiO_2 is surprisingly high (0.24 - 0.39%) compared to average values of $\sim 0.1\%$ TiO_2 in orthopyroxenes from granulites and charnockites (Mortimer, *pers. comm.*).

Table 3.9 Electron microprobe analyses and structural formulae of orthopyroxenes from the Enmore Adamellite (Hillgrove Suite).

Sample number Catalogue number	(1) ENA 46054	(2) ENA 46054
SiO_2	48.9	48.9
TiO_2	.39	.24
Al_2O_3	2.1	1.2
FeO^*	31.5	33.6
MnO	.82	.96
MgO	16.1	14.6
CaO	.20	.21
Na_2O	-	.23
mg	47.7	43.7
<i>Structural formulae (based on 6 oxygens per formula unit)</i>		
Si	1.913	1.938
Al^{iv}	.087	.062
Ti	.011	.007
Al^{vi}	.012	-
Fe	1.029	1.114
Mn	.027	.032
Mg	.939	.863
Ca	.008	.009
Na	-	.018
Σ (cations)	4.026	4.036

* Total Fe as FeO.
Totals normalized to 100%.

3.5.4 Ilmenite

3.5.4.1 Occurrence and Chemistry

Ilmenite is the sole opaque oxide phase in all Hillgrove Suite granitoids and never exceeds 0.4% of the mode. It occurs most commonly as small, subhedral, elongate to lozenge-shaped crystals within reddish-brown biotite. Less commonly, it forms inclusions in plagioclase. Textures suggest that ilmenite was probably the first mineral to crystallize from the melt. However, it may also represent a relict phase which survived and which reacted with the melt to form the surrounding biotite.

The ilmenite contains no visible inclusions or exsolution lamellae, shows no alteration and is homogeneous.

Analyses and structural formulae of Hillgrove Suite ilmenites are presented in Table 3.10. Apart from MnO:FeO variation there is a remarkable chemical similarity between all ilmenites. The presence of minor SiO₂ and Al₂O₃ could be due to either submicroscopic silicate inclusions or to incorrect microprobe calibration. Carmichael (1967) and others have suggested that small amounts of SiO₂ and Al₂O₃ may substitute into the ilmenite structure. Some of the MgO analyses are probably real (as geikelite component), although the 'correct' MgO values for the Hillgrove Suite ilmenites are likely to be between the probe detection limit and the indicated values.

The MnO content of Hillgrove Suite ilmenites ranges from 3.1% to 10.9%. The average MnO content (~ 6%) is higher than the upper limit of 5% MnO, reported by Buddington and Lindsley (1964), but this is low compared to 12-16% MnO in ilmenites from other adamellites and granites, e.g. Snetsinger (1969) and Tsusue (1973). The amount of MnO present in Hillgrove Suite ilmenites can be directly and consistently related to the nature of the enclosing mineral. For example, ilmenite enclosed by plagioclase has low MnO compared to ilmenite enclosed by biotite or to ilmenite occurring as discrete grains between quartz and feldspar.

Table 3.10 Electron microprobe analyses and structural formulae of ilmenites from the Hillgrove Suite.

Sample No. †	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
Catalogue No.	HRA	K14	ABGD	TOBA	RVGD	RVT10	WOLL	HGA	BKA	ENA	ENA
SiO ₂	(.8)	(.7)	-	-	(.7)	(.7)	(.9)	-	(.6)	-	-
TiO ₂	52.7	52.1	52.3	52.5	52.2	52.1	52.4	52.6	51.9	52.0	52.1
Al ₂ O ₃	-	-	-	-	.2	.3	.2	-	.3	.1	-
FeO*	40.3	41.8	41.0	39.5	40.6	43.1	41.7	42.4	40.1	42.0	43.5
MnO	6.26	5.28	6.64	8.04	6.10	3.41	4.39	5.00	6.67	5.75	4.21
MgO	-	.12	-	-	.16	.21	.22	-	.19	.15	.16
Structural formulae (based on 6 oxygens per formula unit)											
Si	(.037)	(.034)	-	-	(.034)	(.036)	(.043)	-	(.030)	-	-
Ti	2.012	1.994	1.990	1.993	1.999	1.989	2.000	1.998	1.983	1.978	1.982
Al	-	-	-	-	.011	.017	.014	-	.018	.006	-
Fe	1.708	1.776	1.736	1.669	1.711	1.832	1.773	1.791	1.705	1.777	1.840
Mn	.269	.228	.285	.344	.263	.147	.189	.214	.287	.247	.181
Mg	-	-	-	-	.012	.016	.017	-	.014	.011	.012
Σ(cations)	3.989	3.998	4.010	4.007	3.996	4.001	3.993	4.002	4.007	4.019	4.015
X _{Ilm} ^{Mn/Fe}	.16	.12	.16	.21	.15	.08	.11	.12	.17	.14	.10
X _{Bt} ^{Mn/Fe}	.013	.012	.018	.014	.013	.007	.008	.025	.013	.009	.010
K _D ^{Mn/Fe} _{Ilm/Bt}	.082	.091	.109	.068	.086	.086	.075	.206	.075	.066	.105
Sample No. †	(12)	(13)	(14)	(15)	(16)	(17)	(18)	(19)	(20)	(21)	
Catalogue No.	WINT	ARG	ARGX	KA	KILB	INGA	TM4	TM2	DUND	GK6	
SiO ₂	-	-	-	-	-	-	-	-	-	-	
TiO ₂	52.4	52.6	52.2	52.8	52.0	53.1	52.3	52.1	52.6	52.3	
Al ₂ O ₃	-	-	0.1	-	0.3	-	-	0.3	-	.17	
FeO*	43.9	40.5	36.1	41.6	41.2	41.4	39.1	37.1**	37.1	40.5	
MnO	3.46	6.60	10.6	5.60	4.17	4.49	8.43	9.42	10.0	6.13	
MgO	.12	.13	-	-	-	-	-	.35	-	.37	
Structural formulae (based on 6 oxygens per formula unit)											
Si	-	-	-	-	-	-	-	-	-	-	
Ti	1.991	1.997	1.992	2.002	2.011	2.025	1.988	1.974	1.994	1.982	
Al	-	-	.007	-	.015	-	-	.016	-	.010	
Fe	1.857	1.711	1.529	1.757	1.774	1.757	1.655	1.579	1.567	1.709	
Mn	.148	.282	.456	.239	.182	.193	.361	.402	.428	.262	
Mg	.009	.010	-	-	-	-	-	.036	-	-	
Σ(cations)	4.006	4.000	3.977***	3.998	3.982	3.975	4.005	3.995	3.990	3.990	
X _{Ilm} ^{Mn/Fe}	.08	.17	.30	.14	.10	.11	.22	.25	.27	.15	
X _{Bt} ^{Mn/Fe}	.010	.034	.048	.010	.009	.006	.019	.017	.022	.015	
K _D ^{Mn/Fe} _{Ilm}	.130	.206	.160	.070	.085	.053	.086	.056	.081	.096	

Totals normalized to 100%

† See Appendix I

* Total Fe as FeO

** 0.3% Fe₂O₃

*** Does not include 0.18 atoms Cr (0.4% Cr₂O₃)

The excellent 1:1 relationship between MnO and FeO shown by the Hillgrove Suite manganoan ilmenites (Fig. 3.4) is not evident from the MnO and FeO contents of the host rocks. Indeed, the MnO content of the ilmenite is quite independent of host rock MnO content. A similarly independent MnO_{ilm}/MnO_{rock} relationship is shown by the high-grade metamorphic ilmenites of Broken Hill (cf. Plimer, 1978) indicating that the MnO content of Hillgrove Suite ilmenites is a function of *both* modal proportion and bulk rock composition. However, because of the sporadic distribution of small modal proportions of ilmenite in the rocks, this relationship can be only rarely demonstrated.

Unlike granitic suites in Japan (e.g. Tsusue, 1973), the MnO content of Hillgrove Suite ilmenites shows no correlation with increasing D.I.* For example, ilmenite from the Ingleba Adamellite (D.I. = 88) contains only 4.5% MnO, whereas ilmenite from the Tia Granodiorite (D.I. = 79) contains up to 10.9% MnO. This lack of correlation may be primarily due to the likelihood that variation of D.I. in the Hillgrove Suite granitoids is *not* a function of differentiation.

Cationic Ti is always in excess of cationic Fe; hence the R_2O_3 component in the Hillgrove Suite ilmenites is very small. Highest R_2O_3 values occur in samples of the Enmore Adamellite (1.3% - 1.6%) compared with < 1% R_2O_3 in the remaining ilmenites. To confirm this, Fe_2O_3 has been measured on ilmenite separated from the Tia Granodiorite (TM2, anal. 19; Table 3.10) and found to be less than 0.3 wt.%. While this value is considerably less than usually recorded in igneous ilmenites (e.g. Carmichael, 1967; Duchesne, 1972; Mathison, 1975) it is consistent with low Fe_2O_3 contents in other Hillgrove Suite minerals and exemplifies the highly reduced nature of the host granitoids.

3.5.4.2 Distribution of Mn-Fe Between Ilmenite and Biotite

Since only some of the total Mn, Ti and Fe in the rock is contained in ilmenite, the remaining Mn, Ti and Fe must be distributed between coexisting biotite and, where present, garnet or amphibole. Figure 3.5 illustrates the

* Differentiation Index (Thornton and Tuttle, 1960), i.e. the sum of normative *qz*, *or*, *ab*, *lc*, *ne* and *ks*.

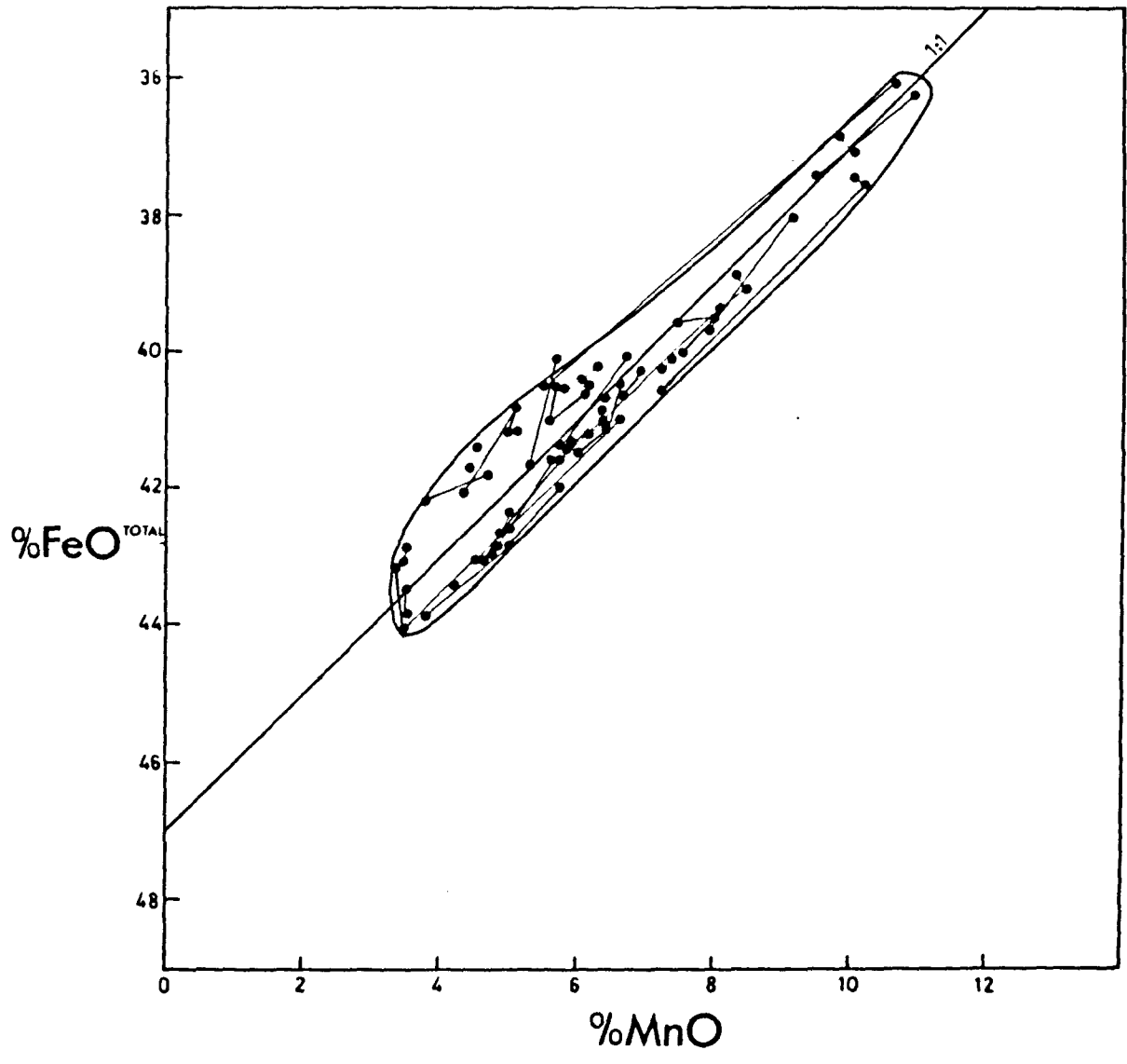


Figure 3.4 Relationship between MnO and FeO in manganian ilmenites from the Hillgrove Suite. Tie lines connect analyses of ilmenites from the same slide.

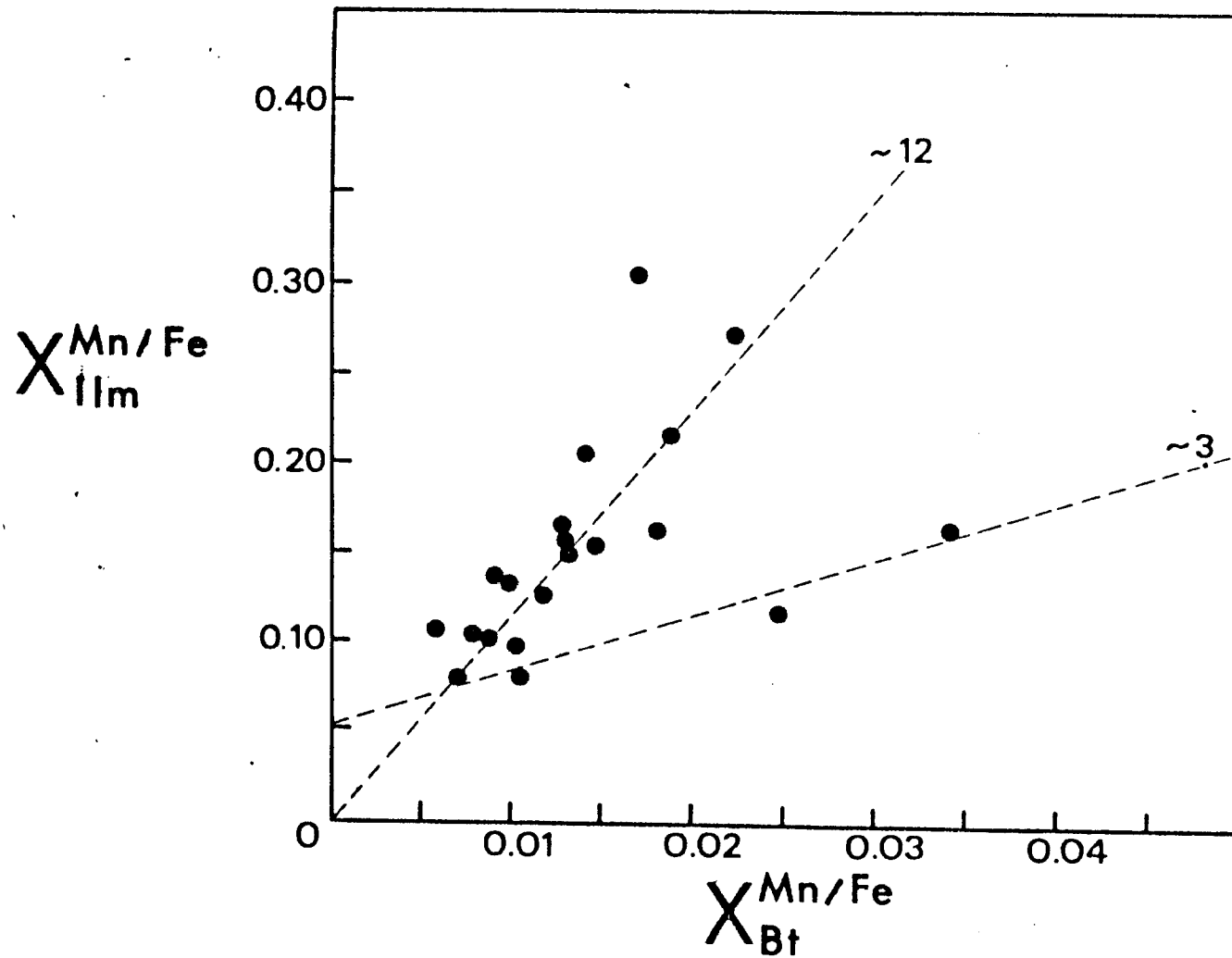


Figure 3.5 Distribution of Mn-Fe between ilmenite and co-existing biotite from Hillgrove Suite plutons.

distribution of Mn-Fe between ilmenite and biotite and shows two trends; a major trend with $K_D \sim 12$, and one with $K_D \sim 3$. The precise reason for this difference is not immediately known but appears to be related to a preferential enrichment of Mn (over Fe) in biotites from the Hillgrove Adamellite and Argyll Granodiorite biotites.

3.5.5 Garnet

3.5.5.1 Occurrence and Chemistry

Garnet is a minor constituent in many of the Hillgrove Suite plutons. It rarely exceeds 1% of the mode; however, some phases of the Abroi Granodiorite contain up to 5% (Binns, 1966). Three types of garnet may be recognized, namely types A, B and C. Table 3.11A summarizes the salient petrographic features of these garnets.

Table 3.11A Summary of petrographic features of Hillgrove Suite garnets. For comparison, Bundarra Suite garnets have also been included.

Petrographic feature	Type A	Type B	Type C	Bundarra Suite
colour	pale pink	pale pink	v.pale brown	pale pink
maximum size	6 mm	3 mm	1 mm	5 mm
shape	anhedral	euhedral	anhedral	anhedral to subhedral
distinctive textural feature	poikiloblastic	occurs in clots	derived from xenoliths	ranges from single poikiloblasts to clots
associated minerals	green biotite	relict cordierite	rimmed by pale yellowish-brown biotite	cordierite and muscovite
relation to foliation	cataclastic foliae wrap around garnet	?	?	—

Analyses of seven garnets from the Hillgrove Suite are listed in Table 3.11B. The most obvious aspect of these garnets is their considerable compositional range; for example, CaO ranges from 8.69% to 0.88%; MnO from 18.2% to 1.09%; MgO from 6.59% to 1.41% and FeO from 32.5% to 22.1%. The Bundarra Suite garnet which has been included for comparison contains more iron (34.6% FeO), is low in CaO (~ 1.0%) and shows only minor variation in MgO and MnO.

Table 3.11B Electron microprobe analyses and structural formulae of representative garnets from the Hillgrove Suite. Included for comparison is a typical garnet from the Bundarra Suite.

Sample No.†	Type A							Type B					Type C			Bundarra Suite		
	(1) RVGD core	(2) RVGD rim	(3) ENA core	(4) ENA rim	(5) FENA	(6) TM4 core	(7) TM4 rim	(8) BKA core	(9) BKA rim	(10) LTM2 core	(11) LTM2 ½out	(12) LTM2 ¼out	(13) LTM2 rim	(14) RTV7 core	(15) RTV7 midway	(16) RTV7 rim	(17) BS core	(18) BS rim
Catalog.No.	46036	46036	46054	46054	46055	46069	46069	46053	46053	46070	46070	46070	46070	46039	46039	46039	46074	46074
SiO ₂	36.9	36.8	37.2	37.0	36.8	36.9	36.5	37.8	37.0	36.8	36.7	36.5	35.9	37.4	37.5	37.3	35.8	36.2
TiO ₂	-	-	.21	.11	.17	-	-	-	-	-	-	-	-	-	-	-	.11	-
Al ₂ O ₃	22.1	21.8	20.9	21.1	21.6	21.7	21.5	21.8	21.5	21.8	21.7	21.4	21.2	21.7	21.9	22.1	21.2	21.3
FeO*	32.5	31.3	31.5	31.4	32.5	29.4	26.5	30.5	30.0	31.8	30.3	26.5	21.6	22.1	27.9	31.6	34.6	34.3
MnO	1.77	4.88	3.15	4.70	1.09	5.92	11.3	2.20	7.32	3.33	5.70	11.7	18.2	7.34	6.59	2.54	3.03	4.15
MgO	5.63	4.25	5.42	3.88	6.61	4.50	2.59	6.20	-2.95	4.84	4.22	2.62	1.41	2.16	3.37	4.88	3.70	3.18
CaO	1.12	0.98	1.25	1.28	1.15	1.28	1.12	1.00	0.89	0.88	.089	0.89	1.05	8.69	2.45	1.01	0.94	0.96
mg	23.6	19.5	23.5	18.0	26.6	21.4	14.8	26.6	14.9	21.3	19.9	15.0	10.4	14.9	17.7	21.6	16.0	14.2
<i>Structural formulae (based on 8 oxygens per formula unit)</i>																		
Si	2.930	2.948	2.974	2.982	2.920	2.956	2.963	2.988	2.987	2.949	2.953	2.963	2.946	2.984	2.995	2.971	2.916	2.935
Al ^{iv}	.070	.052	.026	.018	.080	.044	.037	.012	.013	.051	.047	.037	.054	.016	.005	.029	.084	.065
Ti	-	-	.013	.077	.010	-	-	-	-	-	-	-	.010	-	-	-	.007	-
Al ^{vi}	1.997	2.006	1.944	1.987	1.940	2.005	2.020	2.019	2.033	2.008	2.011	2.011	1.996	2.025	2.057	2.046	1.952	1.971
Fe	2.159	2.094	2.106	2.117	2.157	1.967	1.798	2.016	2.024	2.134	2.037	1.801	1.481	1.475	1.861	2.107	2.360	2.324
Mn	.119	.331	.213	.321	.073	.402	.779	.147	.501	.226	.388	.806	1.267	.496	.446	.171	.209	.285
Mg	.666	.507	.646	.466	.782	.537	.313	.730	.355	.578	.506	.317	.172	.257	.401	.579	.449	.384
Ca	.095	.084	.107	.111	.098	.110	.097	.084	.077	.076	.077	.077	.092	.743	.210	.086	.082	.083
Σ(cations)	8.036	8.022	8.029	8.009	8.060	8.021	8.007	7.996	7.990	8.022	8.019	8.012	8.018	7.996	7.975	7.989	8.059	8.047

† See Appendix I.

* Total Fe as FeO.

Analyses recalculated to 100%.

3.5.5.2 Compositional Zoning

Hillgrove Suite garnets all show at least moderate core-rim variation as illustrated on a triangular plot of Ca - (Fe + Mg) - Mn (Fig. 3.6). Type A and Type B garnets have essentially the same element variation, i.e. as MnO increases toward the margin there is a concomitant decrease in FeO, CaO, and most significantly, MgO. This is the reverse of that described for many metamorphic garnets, e.g. Sturt (1962), Atherton (1968), but is similar to that generally recorded in felsic igneous rocks (Type 2, Birch and Gleadow, 1974; Green, 1977) and from the high-grade metamorphic rocks associated with the Hillgrove Suite plutons. One conclusion that might be reached from the elemental abundances and the zoning pattern of the high-grade metamorphic garnets (in particular those from the Wongwibinda Complex), is that the Type A and Type B garnets in the Hillgrove Suite, conceivably *could* be refractory relict phases unaffected by any later reaction with melt.

In contrast, Type C garnets zone strongly from very grossular-rich (25 mol.%) and spessartine-rich (17 mol.%) cores, to grossular- and spessartine-poor rims (3 and 6 mol.% respectively). The complementary variation in these Type C garnets is mainly in FeO, although MgO also shows a significant, though relatively small, increase.

Zoning patterns similar to Type C garnets have not been documented for other naturally occurring garnets nor grossular-rich experimental garnets. In the latter, Mn usually varies antipathetically with Ca, i.e. as Ca decreases outwards, Mn increases, the opposite to that observed in Type C garnets.

The extent of spessartine zoning in Type A and B garnets is the same in most cases (~ 12-13 mol.%), regardless of composition, e.g. garnets from the Blue Knobby Adamellite zone outward from 5 mol.% spessartine in the core to 17 mol.% at the rim while garnets from the Tia Granodiorite zone from 13 mol.% to 26 mol.% spessartine. However, in the felsic veins from the Tia Granodiorite, where host rock compositions are relatively Mn-rich, maximum zoning in the garnets ranges from 7.5 mol.% spessartine in the cores, to 42 mol.% at the rims.

3.5.5.3 Garnet Origin and Stability in Granitoids

The garnets in the Hillgrove Suite granitoids, in particular Types A and B, are remarkably similar in their chemistries and in their compositional zoning patterns, to other garnets occurring in felsic igneous rocks throughout the Tasman Orogenic Province (B. Chappell *pers.comm.*; T. Green *pers.comm.*). This close compositional similarity would appear to preclude an accidental origin (Birch and Gleadow, 1974; Green, 1977) and hence implies a genetic relationship between the garnets and their hosts.

Although the experimental work by Green and Ringwood (1968, 1972) and Green (1976, 1977), on garnet-bearing felsic volcanic rocks and on rocks with pelitic compositions, has been particularly useful in delineating garnet compositions at certain P/T conditions, the data cannot be used to establish conclusively whether almandine-rich garnets in granitoids are refractory crystals remaining after ultrametamorphism and melt generation or whether they have crystallized directly from the melt.

Based on experimental data, Green (1976, 1977) suggested that garnets with compositions similar to Types A and B indicate crystallization pressures ranging from ~ 5 to ~ 7 kb (18-25 km), but stressed that these pressure estimates are directly dependent on the proportion of spessartine end-member in these garnets. According to Green, the addition of significant Mn to the experimental system drastically expands the stability field of almandine-rich garnet (cf. Weisbrod, 1973), so that garnets with 20-25 mol.% spessartine may be stable in felsic melts at pressures as low as 3 kb. This is consistent with the interpretation that garnet-bearing felsic veins from the Tia Granodiorite represent low-pressure (2-3 kb) fractionation products of the granodiorite. In fact, the last garnet to crystallize in these veins contains 42 mol.% spessartine (see Fig. 3.6).

By comparison, the high grossular content of Type C garnet (25 mol.%) suggests crystallization pressures of ~ 20 kb according to Green and Ringwood (1968) and Green (1977). This pressure is clearly inconsistent with the mineralogy, chemistry, and field association of the host adamellite; hence the garnet is suspected to be a relict phase from a calcareous xenolith. Chemical evidence to support the possibility that Type C garnets are relicts is provided by the garnets from high-grade calc-silicate rocks adjacent to the Abroi Granodiorite. These metamorphic garnets have the same zoning patterns as the Type C garnets, except for Mg which shows no obvious core-rim variation. Even during ultrametamorphism, garnets within calcareous metasediments presumably change compositions only slightly; however, any garnets coming into contact with granitic melt, as for example in the Rockvale sample, would probably react and become rimmed by almandine-pyrope (anal. 14-16); Table 3.11. This conclusion is consistent with the observed physical break-up of xenoliths in this adamellite.

In contrast to Type C garnets, there is no convincing textural evidence to provide some insight into the precise origin for Type A and Type B garnets. Both types show zoning patterns which are consistent with crystallization from a cooling melt and both are irregularly distributed throughout the granitoids. Thus the most valid method of discriminating between Type A and Type B garnets with respect to their origin appears to be grain shape. Accordingly, it is concluded that Type B garnets are likely to have crystallized directly from a melt whereas Type A garnets are likely to be refractory.

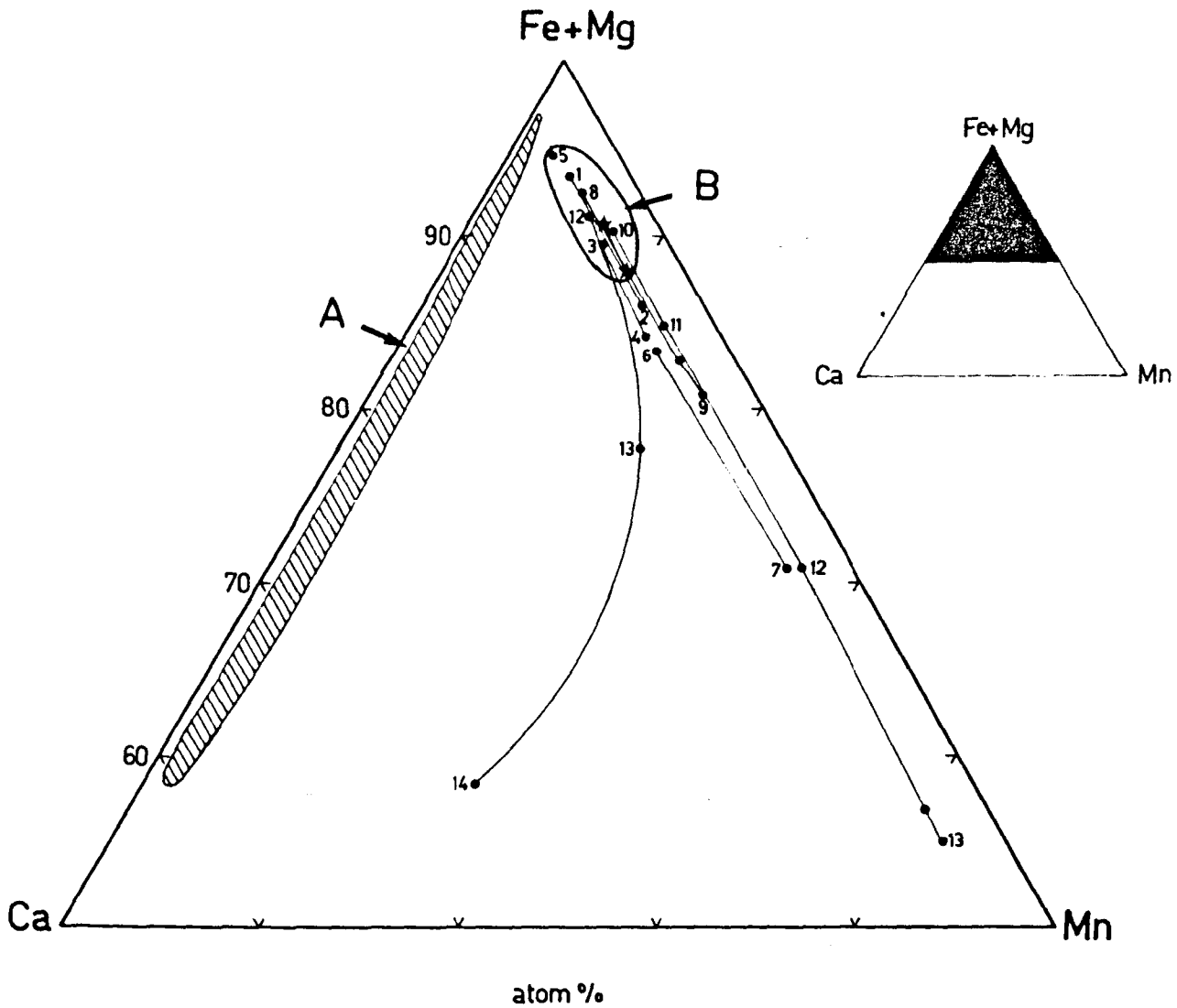


Figure 3.6 Plot of Ca-(Fe+Mg) - Mn for garnets from Hillgrove Suite plutons. Tie lines connect different garnet zones. Numbers refer to analyses in Table 3.11B. Bundarra Suite garnets (stars) fall within the field of analyzed garnets from this Suite (Green, 1977). Experimental garnets from rocks of pelitic composition (A; Green, 1977) are included for comparison.