

## CHAPTER 4

### NEW ENGLAND PLUTONIC SUITE

#### 4.1 INTRODUCTION AND AIMS

The New England (Plutonic) Suite is defined here as those plutons which constitute the bulk of the outcrop area of the New England Batholith (Fig. 4.1), excluding the Hillgrove and Bundarra Suites. It is characterized by numerous, areally extensive, Mid to Late Permian plutons which display textures, mineralogies and rock compositions that generally differ from those of other suites. In addition to their very large compositional range, members of the New England Suite are characterized by strong enrichment of  $K_2O$  in the mafic and intermediate members, variable  $Fe^{3+}/Fe^{3+} + Fe^{2+}$  ratios, pronounced REE enrichment in the felsic granitoids and exceptionally high 100 Mg/Mg + Fe ratios for both host rocks and minerals.

Plutons of this suite have been designated as I-type on the basis of their chemistry, mineralogy, Sr isotopes, textures and field relationships (Flood and Shaw, 1975, 1977; Chappell, 1978). However, in a number of aspects many members of this suite are indistinguishable from members of the Hillgrove Suite.

The original purpose for studying these rocks was to examine whether a sharp contrast existed between the New England Suite and the granitoids of the Hillgrove Suite. During the course of this investigation it became apparent that the rocks of the New England Suite may in fact represent one end of a spectrum of compositions that is continuous with rocks of the Hillgrove Suite. Indeed, O'Neil *et al.* (1977) suggested that there may perhaps be a transition between these suites.

On the basis of chemical and isotopic data the New England Suite has been subdivided recently into a number of smaller but nevertheless distinct suites, e.g. the Uralla and Moonbi Suites (O'Neil *et al.*, 1977), and the Inlet, Bendemeer, Attunga Creek and Looanga Suites (Chappell, 1978). However, for comparative purposes it was considered more convenient not to subdivide the New England Suite into these lesser suites.

The major objectives of this chapter are to document the textural, mineralogical, chemical and isotopic characteristics of the New England Suite and use these data to establish precisely any relationships between this Suite and the Hillgrove Suite. In addition this chapter examines the applicability of the 'S- and I-type' granitoid classification to the New England Suite.

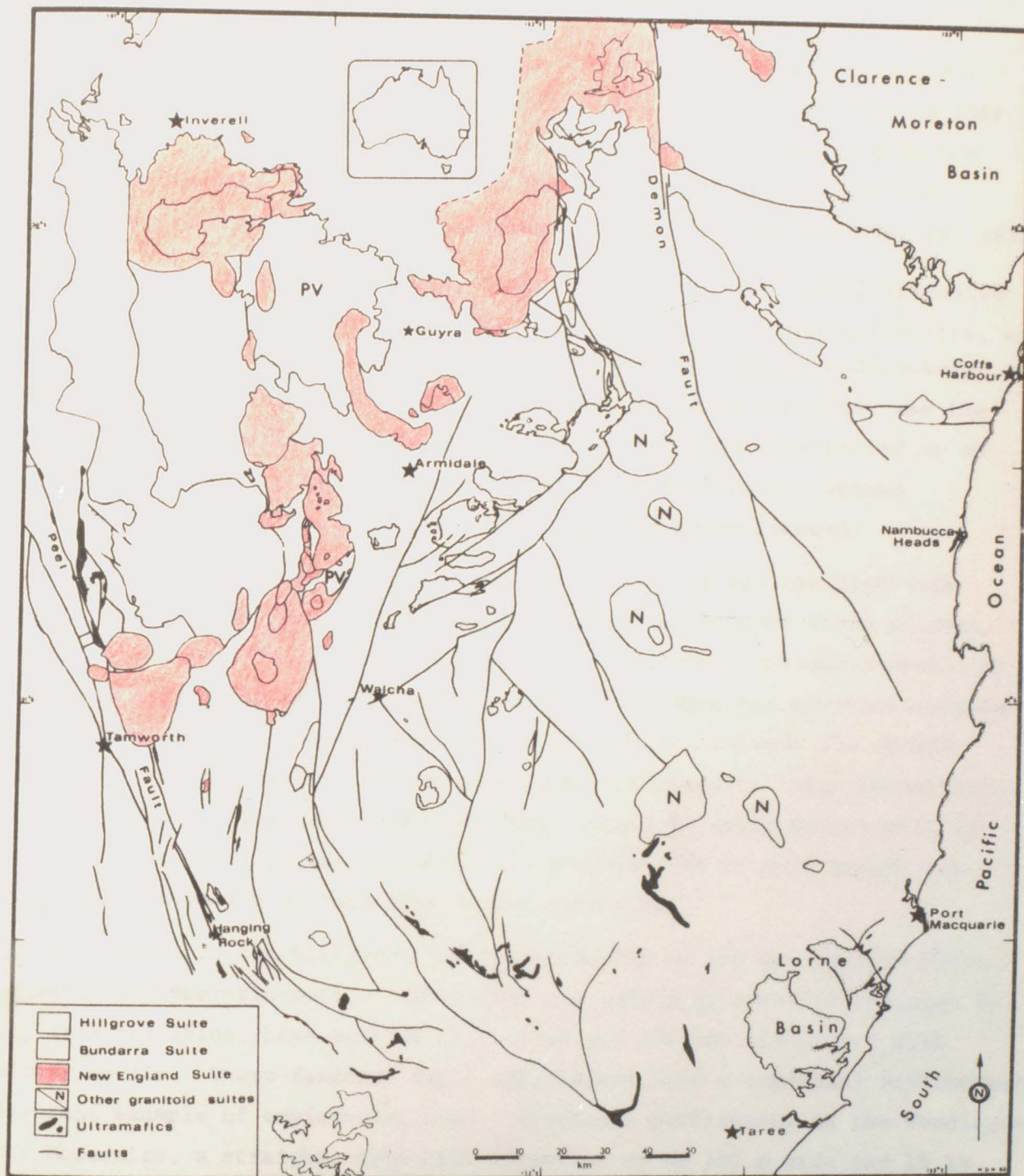


Figure 4.1 Distribution of plutons belonging to the New England Suite.

#### 4.2 CONTACT RELATIONS AND STRUCTURAL ASPECTS

Most plutons of the New England Suite are structurally discordant and cut across the general strike of the country rock. The one exception to this is the Shalimar Tonalite which displays a conformable relationship with the surrounding sediments. Contacts between the majority of plutons and the country rocks are sharp and steeply dipping. Nevertheless determining the relative ages of adjacent plutons is generally restrained by outcrop. Local intrusive complexity is evident in some areas; for example, the Glenburnie Adamellite, with its rudimentary foliation, mild deformation and abundant, wispy sedimentary inclusions, has been intruded by several plutons, of which the first was the Terrible Vale Microtonalite. This microtonalite, recently interpreted as an extrusive (Flood *et al.*, 1977), occurs as three distinct masses around Kentucky and Uralla and has itself been intruded and metamorphosed.

With the exception of the Wards Mistake Adamellite and the Highlands Monzonite, chilled margins are not observed in the New England Suite plutons. Two distinct marginal facies are associated with the Wards Mistake Adamellite: the first is relatively mafic in composition and has abundant microphenocrysts of plagioclase set in a quartzo-feldspathic groundmass, whereas the second resembles the typical texture of the Adamellite but contains large phenocrysts of alkali feldspar. The development of large alkali feldspar phenocrysts in association with small pegmatitic veins and segregations at this margin may suggest a concentration of volatiles during intrusion.

In contrast to the Hillgrove Suite the plutons of the New England Suite are essentially post-kinematic. Intrusions are generally ovoid or circular in shape, lack foliation, have massive structures and are not associated with prominent faults. These features imply emplacement into a tensional environment. A different example of emplacement into a tensional environment is the Woodlands Quartz Monzonite, a straight, dyke-like intrusion up to 200 m wide and 15 km long transecting the Kookabookra Adamellite.

Xenoliths, inclusions and clots of ferromagnesian minerals are common in New England Suite plutons. Most xenoliths are mafic (diorites and quartz diorites), and although they frequently show evidence for disaggregation there never appears to be any reaction with the host rock. Felsic xenoliths (leucotonalite to leucoadamellite) are comparatively rare. Xenoliths of pelitic and psammitic country rock are also rare but where present they are confined to the pluton margins.

#### 4.3 ASSOCIATED METAMORPHIC ROCKS

Narrow 200-600 m wide thermal aureoles are developed around several of the New England Suite plutons and may be recognized in the field as prominent hornfels ridges, e.g. Shalimar Tonalite. Migmatites and regional metamorphic rocks are never associated with this Suite.

Mineral assemblages in the country rocks surrounding these plutons, especially in the interbedded metabasalts, transform abruptly from those appropriate to lower greenschist facies to those of the hornblende hornfels facies. The occurrence of a two-pyroxene hornfels between the Walcha Road Adamellite and the Banalasta Adamellite probably represents the highest grade of contact metamorphism attained in New England. Other occurrences of pyroxene hornfels are restricted to basic xenoliths within the mafic Kentucky Ponds and Kentucky Diorites. Andalusite-almandine and andalusite-cordierite assemblages are not uncommonly developed in pelitic interbeds adjacent to the New England Suite plutons. Sillimanite is uncommon in these aureoles but is developed intensely in some pre-Permian volcanics overlying the Kentucky Ponds Diorite.

#### 4.4 PETROGRAPHY OF THE NEW ENGLAND SUITE

##### 4.4.1 Introduction

This suite of plutons is characterized by much greater textural and mineralogical variation than the Hillgrove Suite. Plutons range from melanocratic gabbroic diorite (e.g. Kentucky Diorite) to leucoadamellite (e.g. Oban River Leucoadamellite). Average grain sizes range from 0.5 mm in chilled samples (usually mafic and often microporphyritic) to 6 mm in felsic granitoids. However, the majority of plutons are medium-grained and sparsely porphyritic with plagioclase phenocrysts up to 12 mm. Some adamellites are coarsely porphyritic and contain large crystals (megacrysts) of alkali feldspar up to 5 cm in length (e.g. Walcha Road Adamellite).

##### 4.4.2 Textures and Descriptive Petrography

###### (a) Leucoadamellite (Oban River Leucoadamellite)

This rock type is coarse-grained (4-6 mm), hypidiomorphic-granular, mildly porphyritic and consists predominantly of large, simply-twinned, sub-hedral crystals of orthoclase microperthite, anhedral quartz, small to large,

stumpy, tabular crystals of polysynthetically-twinned, subhedral to euhedral, sodic plagioclase ( $An_{28-12}$ ). Scattered, straw-yellow (X) to dark brown (Y=Z) biotite ( $mg \sim 30$ ) is the most common ferromagnesian mineral. Small, subhedral, strongly pleochroic, dark-green prisms of ferrohornblende ( $mg \sim 34$ ) are sporadically disseminated throughout, although there is also a tendency for the amphibole to occur in patches. Manganiferous ilmenite with up to 10.4% MnO is the dominant iron titanium oxide even though small octahedra of magnetite are not uncommon. The ilmenite occurs mainly in three forms:

- (a) discrete crystals, commonly at plagioclase-amphibole grain boundaries;
- (b) ragged, cigar-shaped grains enclosed by biotite; and
- (c) as lamellar intergrowths with magnetite.

Large, superbly zoned, euhedral crystals of variably metamict allanite and euhedral to subhedral zircon are common accessories.

- (b) Adamellites, Granodiorites and Tonalites (Walcha Road, Glenburnie and Wards Mistake Adamellites; Balala and Uralla Granodiorites; Khartoun, Wilhelmshohe and Shalimar Tonalites)

These rocks are typically medium-grained (1-4 mm), with a hypidiomorphic inequigranular to seriate texture. In addition to the major variation in modal plagioclase and alkali feldspar, there is a marked increase in ferromagnesian phases but only a small decrease in quartz from adamellite to tonalite.

Plagioclase, ranging from  $An_{19}$  to  $An_{69}$ , occurs as fine to coarse, complexly zoned, polysynthetically and Carlsbad-twinned laths. In fresh hand specimens the broad laths are dark grey, but with incipient alteration, a greenish colouration may appear. Clear untwinned albite rims ( $An_3$ ) are not uncommon in some of the adamellites. Zoning is normal, uneven and oscillatory. One very significant feature of plagioclase in the granodiorites, tonalites and some of the adamellites, e.g. Wards Mistake Adamellite, is the presence of an evenly-spaced, usually narrow zone one-third to one-half the distance from core to rim. This zone is notably more calcic than the bulk crystal composition, and strongly zoned from very calcic on the inner side to relatively sodic on the outer side. Very rarely, two such zones occur within individual crystals. By comparison, plagioclase cores are usually less calcic than these zones and in fact may even be more sodic than subsequent oscillations.

The alkali feldspar, which may occur as large phenocrysts in some adamellites, e.g. Walcha Road Adamellite, is predominantly orthoclase although margins and small interstitial crystals may show some cross-hatched twinning.



The large phenocrysts are usually strongly perthitic, but smaller grains generally lack these exsolution features.

Reddish-brown to brown (Y=Z) and straw-yellow to tan (X) biotite (*mg* 38-57) is the dominant ferromagnesian mineral in these rocks. It occurs as either small, ragged crystals or as large, euhedral plates enclosing numerous, acicular needles of apatite and subrounded to euhedral crystals of zircon. Ilmenite commonly forms cores to the reddish-brown biotite of the tonalites.

The most major common amphibole in these rocks is greenish, generally subhedral and only moderately aluminous (4-8%  $Al_2O_3$ ). It frequently displays cores which are ragged, and consist of either a felted mass of pale yellow-brown or pale yellow-green cummingtonite (*mg* 42-61), or of fibrous pale-green actinolite (*mg* 55-78). The fibrous amphibole varieties are both presumably secondary after orthopyroxene and clinopyroxene, respectively. Like the biotite, the dominant amphibole is often riddled with acicular apatite, and less commonly, zircon. Large, strongly zoned ferroan pargasites, zoned firstly to edenitic magnesiohornblende then to low-Al magnesiohornblende, are not uncommon in these rocks, e.g. Wards Mistake Adamellite.

Both clinopyroxene and orthopyroxene are present in the tonalites, the former as rare cores to amphibole and as inclusions in calcic plagioclase, and the latter as discrete, ovoid, pale-coloured grains.

Ilmenite is the sole iron-titanium oxide in all these rock types except for the Walcha Road Adamellite, which contains in addition, small equant granules of magnetite.

(c) Microtonalites (Microgranodiorites) (Terrible Vale and Back Creek Microtonalites)

These rocks are characterized by a microporphyritic texture in which microphenocrysts of orthopyroxene, clinopyroxene and plagioclase are set in a very fine-grained equigranular quartzo-feldspathic groundmass. Their textures resemble those of some felsic extrusives, e.g. rhyodacites from the Marysville Igneous Complex, Victoria (McLaughlin, 1976) but not those of the type ignimbrite from Owharoa, New Zealand (Marshall, 1932).

Plagioclase, the largest and most abundant phenocryst, is usually polysynthetically twinned, up to 3 mm in length, and zoned from clear unaltered cores of  $An_{62}$  to rims of  $An_{35}$ . Although zoning in the plagioclase is generally normal with only minor oscillations, a poorly developed, inner calcic zone may

also be present, particularly in varieties lacking clinopyroxene. The plagioclase phenocrysts frequently appear to be broken or fractured which, according to Flood (1971) suggests a xenocrystal origin. The composition of the groundmass plagioclase is particularly interesting, being similar to the rims of the phenocrysts.

Alkali feldspar (Or<sub>77-86</sub>) is usually restricted to the groundmass, although some anhedral microphenocrysts occur in the Terrible Vale samples. In some facies of the Back Creek Microtonalite the alkali feldspar is sufficiently abundant and coarse-grained to develop a monzonitic relationship towards the plagioclase.

Orthopyroxene is the most abundant ferromagnesian phase, forming phenocrysts up to 3 mm in length. Core compositions range from En<sub>64</sub> to En<sub>40</sub> depending on the host rock chemistry. Iron-enrichment at the rims is only slight. Clinopyroxene is usually more magnesian than the coexisting orthopyroxene and occurs as smaller, subhedral to euhedral phenocrysts frequently mantled by brownish biotite. Rare cores of low-Ca, moderately aluminous (5.1% Al<sub>2</sub>O<sub>3</sub>), and relatively magnesian pyroxene (*mg* 74) occur in specific samples of the Back Creek Microtonalite. Compared to biotite, which is ubiquitous but ranges in colour, habit and composition from chocolate-brown poikilitic plates (*mg* 45) (Back Creek samples) to reddish-brown flakes in the groundmass (*mg* 71) (Terrible Vale samples), amphibole is restricted to only a few samples. Cummingtonite forms clear, fibrous cores (*mg* 57-59) to subhedral, pale to moderately green, zoned calciferous amphibole (*mg* 64-68) in rocks free of orthopyroxene. Vanadiferous magnetite (~ 0.5% V<sub>2</sub>O<sub>3</sub>) and ilmenite coexist in the Back Creek Microtonalite, but magnetite is absent from the Terrible Vale samples.

(d) Quartz Monzonites and Monzonites (Woodlands and Clovernook Quartz Monzonites; Highlands Monzonite)

Textures of these rock types vary from monzonitic (Highlands Monzonite) to hypidiomorphic inequigranular. Average grain size is fine to medium (0.5-3 mm). The important characteristic of this group is the abundance of alkali feldspar, as tabular to equant, simply-twinned crystals, or as large prisms heavily occluded with plagioclase, pyroxene, biotite, and to a lesser extent, amphibole. The alkali feldspar is usually strongly perthitic, with a bulk composition between Or<sub>75</sub> and Or<sub>80</sub>. Plagioclase generally occurs as subequant, polysynthetically-twinned, clear, pale-grey crystals and varies from An<sub>48</sub> to An<sub>10</sub>. Pale-brownish quartz is interstitial to the feldspars.

Amphibole and biotite are the major ferromagnesian minerals; however, large subhedral crystals of orthopyroxene ( $En_{70-61}$ ) exsolving clinopyroxene along (100) are prominent in the monzonite. The biotite in the monzonite (*mg* 54-65) varies from olive-green to dark reddish-brown and contains numerous inclusions of euhedral zircons, monazite, stumpy apatite, and occasionally, also ferromagnesian and iron-titanium minerals. Both the moderately green subaluminous hornblende (*mg* 59-70) and the biotite form rims around pyroxenes, particularly the clinopyroxene, which compared to the orthopyroxene, forms smaller, subhedral to euhedral prisms (*mg* 68-77) and shows one well developed set of thin exsolution lamellae of Ca-poor pyroxene parallel to (100). In the quartz monzonites, pyroxenes are restricted to cores within amphibole. Many of these cores are partly hydrated and replaced by numerous fine blades of actinolite. However, large euhedral crystals of actinolite, in apparent equilibrium with more aluminous hornblende, are also sporadically distributed throughout these rocks.

A second characteristic feature of these rocks is the ubiquitous presence of coexisting iron-titanium oxides. Vanadiferous magnetite frequently occurs as solitary octahedra whereas ilmenite forms ragged crystals of variable composition, probably due to oxidation. Sphene is a major accessory mineral and occurs as rims to ilmenite and magnetite, or as large, euhedral to small anhedral crystals. Stocky apatite and zircon are additional accessories.

(e) Diorites and Gabbroic Diorites (Kentucky and Kentucky Ponds Diorites)

These very mafic melanocratic plutons are characterized by a ferromagnesian mineral assemblage which consists of amphibole, biotite, clinopyroxene and orthopyroxene in decreasing order of abundance. Average grain size ranges from 1 mm to 4 mm; however, the Kentucky Diorite has several, very coarse (up to 12 mm), felsic pegmatites. Textures range from typically dioritic, i.e. albite- and Carlsbad-twinned plagioclase interlocking with large euhedral to subhedral amphibole and dark reddish-brown biotite, to a texture which may be best described as an accumulation of decussate sub-idiomorphic mafic minerals and randomly oriented, thick laths of well-twinned, clear, poorly zoned and sometimes porphyritic plagioclase. Zoning in the plagioclase is usually negligible, and occasionally reverse, with cores of  $An_{53}$  and rims of  $An_{57}$ . Alkali feldspar is rare, but sometimes present as anhedral, simply-twinned grains up to 3 mm, but more commonly as interstitial grains in close association with accessory quartz.



Amphiboles, comprising medium-green to pale-brown, twinned magnesiohornblende (*mg* 63-71), brown ferroan pargasite (*mg* 62-71), and pale-green cummingtonite (*mg* 59-61), are the most abundant mafic minerals (up to 40%). Their abundance and habit appears to be inversely proportional to the amount of clinopyroxene in the sample, i.e., where it is abundant amphiboles are restricted to small anhedral grains, or to overgrowths on pyroxene, but where it is scarce, amphiboles are large and euhedral. Orthopyroxene (*En* 55-65) is colourless and frequently altered along transverse cracks. Clinopyroxene occurs in two forms: (a) smaller, usually clear, unzoned crystals (*mg* 70-73) and (b) as large, slightly turbid, low-Ca, Cr-rich, zoned diopsides (*mg* 77-82).

Straw-yellow (X) to deep reddish-brown (Y=Z) biotite (*mg* 54-59) occurs as subhedral to ragged plates, associated or intimately intergrown with amphibole, clinopyroxene, orthopyroxene and ilmenite. Ilmenite and acicular apatite are frequent inclusions in the more discrete biotite. Additional accessories in these mafic rocks include sphene, rutile, pyrrhotite, penlandite and chalcopyrite.

#### 4.5 MINERALOGY

##### 4.5.1 Biotite

###### 4.5.1.1 Occurrence and Chemistry

Biotite occurs in all plutons and is commonly the most abundant ferromagnesian mineral. It occurs as:

- (a) decussate aggregates, in which both pyroxenes, amphibole and ilmenite are closely associated;
- (b) discrete crystals, containing abundant inclusions of apatite, zircon, allanite and ilmenite; and
- (c) overgrowths on pyroxene, sometimes directly, but usually in intimate association with, amphibole.

In contrast to the Hillgrove Suite biotites, the biotites from the New England Suite plutons never contain rutile lamellae, nor are they ever associated with either pale-green Ti-poor biotite or muscovite.

The biotite analyses listed in Table 4.1 are arranged in order of approximately increasing bulk rock SiO<sub>2</sub>. Specific chemical features which readily distinguish the New England Suite biotites from those of the Hillgrove Suite include:

Table 4.1 Representative electron microprobe and trace element analyses of biotites from New England Suite plutons.

Sample No.†	(1) KD	(2) KDX	(3) KPD	(4) HM208	(5) HM207	(6) HM207X	(7) CQM	(8) WQM	(9) WQM	(10) TVPT	(11) HGPT	(12) BCKT3	(13) BCKT2	(14) BCKT4	
Catalogue No.	46100	46100	46101	46123	46124	46124	46113	46126	46126	46115	46114	46104	46103	46105	
SiO <sub>2</sub>	36.9	37.1	37.4	38.6	39.0	38.9	38.7	37.7	37.6	39.0	37.9	37.4	40.2	36.2	
TiO <sub>2</sub>	6.02	5.65	4.59	4.28	4.45	2.51	2.20	3.37	4.76	4.50	4.18	4.31	3.24	4.53	
Al <sub>2</sub> O <sub>3</sub>	13.7	13.5	14.2	13.3	12.4	12.2	14.9	12.7	14.4	12.6	13.5	13.5	12.2	13.4	
FeO*	17.01	15.82	16.54	14.70	17.24	15.36	17.98	27.21	23.36	11.93	16.69	18.37	13.73	21.72	
MnO	-	-	-	-	-	-	-	.32	.35	-	-	-	-	-	
MgO	12.70	13.46	13.07	15.61	13.59	15.77	12.52	6.05	6.44	16.72	13.06	11.48	15.67	10.01	
CaO	-	-	-	-	-	.09	.10	-	.24	.12	.11	-	.30	-	
Na <sub>2</sub> O	-	-	-	-	-	-	.19	.43	.40	-	.25	-	-	-	
K <sub>2</sub> O	9.00	8.64	9.19	9.97	8.95	8.52	9.63	9.01	8.51	9.23	9.18	9.04	8.84	9.15	
Total	95.33	94.56	95.00	96.46	95.63	93.34	96.22	96.79	94.07	94.10	94.87	94.10	94.19	94.94	
mg	57.1	60.3	58.5	65.4	58.4	64.7	55.4	28.4	33.0	71.4	58.3	52.7	67.0	45.1	
<i>Structural formulae (based on 22 oxygens per formula unit)</i>															
Si	5.557	5.596	5.636	5.692	5.840	5.901	5.781	5.883	5.622	5.788	5.726	5.741	5.991	5.617	
Al <sup>4</sup>	2.437	2.401	2.364	2.308	2.160	2.099	2.219	2.117	2.378	2.205	2.274	2.259	2.009	2.383	
Ti	.682	.641	.521	.475	.501	.287	.247	.395	.565	.503	.476	.498	.361	.529	
Al <sup>6</sup>	-	-	.154	.008	.022	.080	.412	.207	.303	-	.126	.177	.133	.061	
Fe	2.145	1.996	2.087	1.816	2.158	1.950	2.247	3.549	3.083	1.483	2.110	2.358	1.711	2.820	
Mn	-	-	-	-	-	-	-	.042	.047	-	-	-	-	-	
Mg	2.855	3.026	2.939	3.436	3.032	3.567	2.789	1.405	1.516	3.704	2.945	2.626	3.460	2.316	
Ca	-	-	-	-	-	.014	.016	-	.040	.020	.018	-	.049	-	
Na	-	-	-	-	-	-	.055	.131	.123	-	.073	-	-	-	
K	1.731	1.663	1.768	1.878	1.710	1.650	1.835	1.792	1.714	1.750	1.770	1.770	1.679	1.813	
Σ (cations)	15.417	15.324	15.468	15.613	15.423	15.548	15.601	15.521	15.391	15.471	15.519	15.428	15.415	15.538	
est. av.															
FeO <sub>t</sub>	16.85	17.30													
FeO	15.65	14.45													
Fe <sub>2</sub> O <sub>3</sub>	1.33	3.17													
<i>Trace elements (ppm)</i>															
Zn	190	190													325
Cr	760	870													140
Ni	180	160													150
Cu	24	18													58
Co	70	65													65
Li	22	58													124

† See Appendix I

\* Total Fe as FeO.

Table 4.1 continued. Representative electron microprobe and trace element analyses of biotites from New England Suite plutons.

	(15)	(16)	(17)	(18)	(19)	(20)	(21)	(22)	(23)	(24)	(25)	(26)
Sample No. †	WT	KT	SHT1	SHT2	UGD	EGD	WMA	WMA	WRA	WRAX	GG1	ORLA
Catalog. No.	46116	46112	46110	46108	46117	46119	46125	46125	46122	46122	46120	46129
SiO <sub>2</sub>	36.8	38.0	36.9	38.8	36.7	37.0	34.8	35.9	37.9	38.2	36.6	34.7
TiO <sub>2</sub>	4.21	4.99	3.23	4.23	3.56	3.77	5.04	4.72	4.00	2.76	3.45	3.84
Al <sub>2</sub> O <sub>3</sub>	13.5	13.1	14.2	13.7	14.2	14.1	13.0	13.3	13.8	13.3	12.9	13.3
FeO *	17.57	17.35	19.51	18.28	18.44	22.40	23.74	21.55	17.88	16.13	23.29	27.91
MnO	-	-	.17	-	.16	.20	.27	.21	-	.15	.29	.66
MgO	12.89	12.44	10.85	12.05	12.04	9.45	8.29	10.68	12.94	14.69	8.55	6.85
CaO	.11	-	.14	-	.15	.16	-	-	.11	.21	-	-
Na <sub>2</sub> O	-	-	-	-	-	-	-	-	-	-	.19	.21
K <sub>2</sub> O	9.41	9.08	9.11	9.42	9.37	9.09	9.56	9.78	9.48	9.32	9.26	9.16
Total	94.50	94.97	94.72	96.49	94.62	96.18	94.70	96.14	96.12	94.77	94.52	96.64
mg	56.7	56.1	49.8	54.0	53.8	42.9	38.4	46.9	56.3	61.9	39.6	30.4
Structural formulae (based on 22 oxygens per formula unit)												
Si	5.633	5.740	5.663	5.789	5.628	5.671	5.519	5.523	5.688	5.765	5.769	5.495
Al <sup>4</sup>	2.367	2.260	2.337	2.211	2.372	2.329	2.432	2.422	2.312	2.235	2.231	2.492
Al <sup>6</sup>	.061	.082	.238	.194	.192	.221	-	-	.130	.132	.153	-
Fe	2.248	2.194	2.505	2.284	2.364	2.869	3.148	2.776	2.242	2.037	3.066	3.696
Mn	-	-	.022	-	.021	.026	.037	.028	-	.020	.039	.088
Hg	2.939	2.803	2.483	2.683	2.752	2.156	1.961	2.451	2.892	3.305	2.007	1.617
Ca	.019	-	.024	-	.024	.026	-	-	.017	.034	-	-
Na	-	-	-	-	-	-	-	-	-	-	.057	.063
K	1.836	1.752	1.786	1.794	1.832	1.776	1.934	1.922	1.813	1.795	1.859	1.849
Σ(cations)	15.587	15.398	15.500	15.430	15.595	15.508	15.631	15.674	15.566	15.636	15.588	15.758
est. av.												
Fe <sub>T</sub>					18.65		22.35		18.45			27.05
FeO					17.80		20.60		16.30			23.70
Fe <sub>2</sub> O <sub>3</sub>					0.96		1.94		2.39			3.72
Trace elements (ppm)												
Zn					255		275		305			475
Cr					680		140		380			100
Ni					170		20		130			40
Cu					18		36		52			52
Co					60		65		70			50
Li					118		280		300			544

† See Appendix I

\* Total Fe as FeO.

- (a) higher  $\text{TiO}_2$ ,  $\text{SiO}_2$ , Cr, Ni, *mg* and  $\text{Fe}^{3+}/\text{Fe}^{3+} + \text{Fe}^{2+}$ ; and
- (b) lower  $\text{Al}^{\text{vi}}$  and Li.

CaO, Na<sub>2</sub>O and MnO are usually near the microprobe detection limit for these biotites. Zoning is rare, but where it occurs FeO, TiO<sub>2</sub> and MnO increase in the rims and SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> decrease.

In addition to the comparisons made with biotites from S-type granitoid suites (Table 3.2), the New England Suite biotites are less oxidized, generally contain less Al<sub>2</sub>O<sub>3</sub>, and have more TiO<sub>2</sub>, SiO<sub>2</sub>, Cr, Ni and higher *mg* numbers than biotites from Lachlan Fold Belt I-type granitoids (Rhodes, 1969).

The structural formulae for the New England Suite biotites occasionally have a deficiency in the Z group. Although some workers, e.g. Flood (1971), have suggested that this deficiency may be satisfied with Ti<sup>4+</sup>, it is considered to be sufficiently small to infer slightly erroneous SiO<sub>2</sub> and Al<sub>2</sub>O<sub>3</sub> contents.

Chappell (1966) and Rhodes (1969) suggested that Ti and Fe<sup>3+</sup> occupancy in granitoid biotites is limited to about 0.46 and 0.50 atoms per formula unit, respectively, if there is coexisting magnetite and ilmenite (or sphene). However, even though Fe<sup>3+</sup> falls well within their limit, Ti occupancy in the biotite from the Woodlands Quartz Monzonite reaches a maximum of 0.57 atoms.

The compositions of the biotites from the two suites reflect the host rock chemistry and the coexisting mineralogy. For example, where amphibole is a significant phase in the mode, as in the relatively more calcic plutons of the New England Suite, Al<sup>vi</sup> appears to be preferentially partitioned into the amphibole, hence the deficiency of Al<sup>vi</sup> in the New England Suite biotites. Figure 4.2 illustrates the major chemical variations within the New England Suite biotites, the relationship between bulk rock composition and biotite composition (in terms of Mg/Mg + Fe) and the relationship between these biotites and those of the Hillgrove Suite in terms of Al<sup>vi</sup>, Al<sup>iv</sup> and Ti. In all cases the New England Suite biotites occupy a field distinct from the Hillgrove Suite biotites.

The trace element contents (Table 4.1) also reflect the host rock compositions, i.e. as the host rock becomes more felsic, Cr and Ni decrease and Li, Rb and Zn increase.

#### 4.5.1.2 Crystallization Conditions of New England Suite Biotites

If *mg* values can be taken as a rough guide to crystallization temperatures, New England Suite biotites crystallized at significantly higher temperatures

than biotites of the Hillgrove Suite. The highest biotite  $mg$  values are recorded in the microtonalites, rather than in the pyroxene-rich mineral assemblage of the mafic diorites. This may reflect different  $fO_2$  conditions during crystallization of the respective rock types. When plotted on a  $Fe^{3+}$ - $Fe^{2+}$ -Mg diagram (Fig. 4.3) the biotite from the Kentucky Diorite may be shown, using the experimental data of Wones and Eugster (1965), to have crystallized under  $fO_2$  conditions similar to those for some Hillgrove Suite biotites, i.e. slightly above the trend for the QFM buffer but less than that of the Ni-NiO buffer.

Plotted according to experimentally determined biotite stability relations at  $P_{H_2O} = 2.07$  kb (Wones and Eugster, 1965), New England Suite biotite compositions display a trend which suggests that  $fO_2$  conditions were oxidizing at both the initial and concluding stages of crystallization but reducing for the major part (Fig. 4.3). Extrapolation of the biotite data also indicate that biotite crystallization temperatures are highest in the mafic rocks ( $880^\circ C$ ) and lowest in the leucoadamellite ( $715^\circ C$ ) at  $fO_2$  of  $10^{-13}$  and  $10^{-15}$  bars respectively (Table 4.2). These values are consistent with values obtained by other authors for similar rock types, i.e. Albuquerque (1973). However, the inferred trend of Figure 4.4 assumes that plutons of the New England Suite are related by fractional crystallization, an assumption which is not totally consistent with other mineralogical, chemical and isotopic data.

Table 4.2 Estimates of crystallization temperature and  $fO_2$  for New England Suite biotites (after Wones and Eugster, 1965).

Rock Type	Sample number*	$mg_{bt}$	T° C	$fO_2$
Diorite	(KD)	57.1	880	$10^{-13}$
Gabbroic diorite	(KPD)	58.5	860	$10^{-12}$
Granodiorite	(UGD)	53.8	860	$10^{-13.5}$
Adamellite	(WRA)	56.3	850	$10^{-13.5}$
Adamellite	(WMA)	38.4	775	$10^{-14.5}$
Leucoadamellite	(ORLA)	30.4	715	$10^{-15}$

\* See Appendix I

Piwinskii (1968) also experimentally determined P/T relationships for Sierra Nevada biotites at conditions above the Ni-NiO buffer and found that the

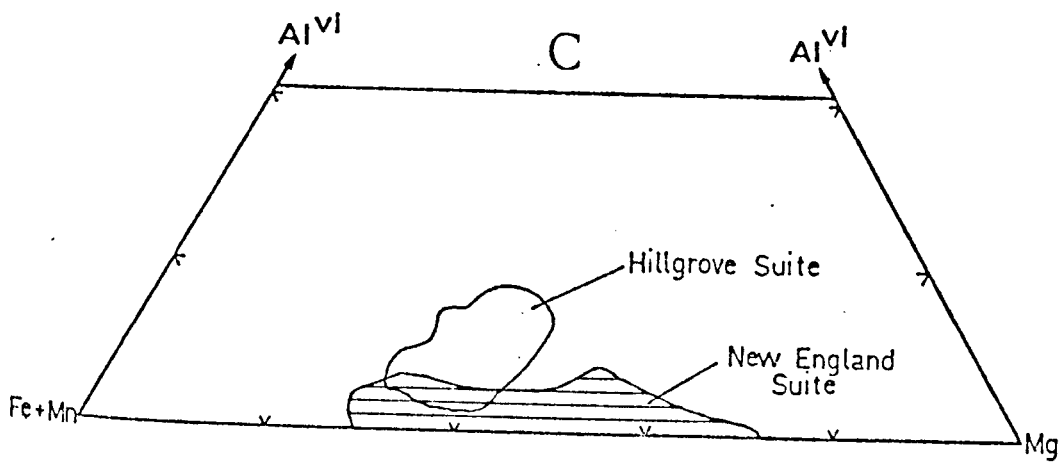
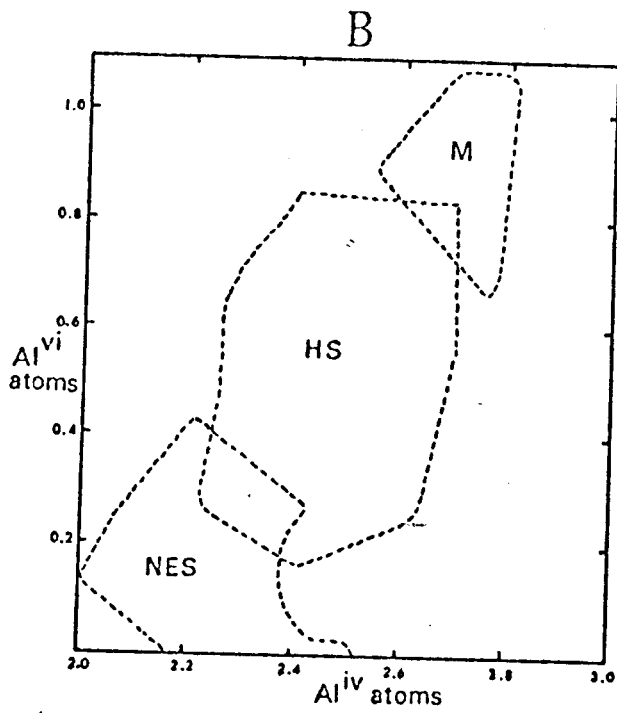
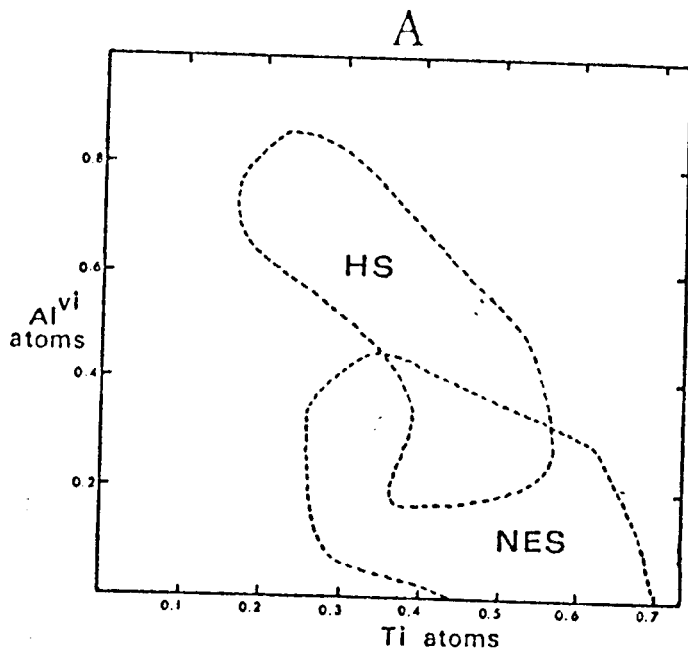


Figure 4.2 Major chemical variations in New England Suite biotites. A =  $Al^{VI}$  v.  $Ti$ ; B =  $Al^{VI}$  v.  $Al^{IV}$ ; C =  $Mg - Al^{VI} - (Fe + Mn)$ . For comparison biotites from Hillgrove Suite granitoids and high-grade New England metamorphics are also included.



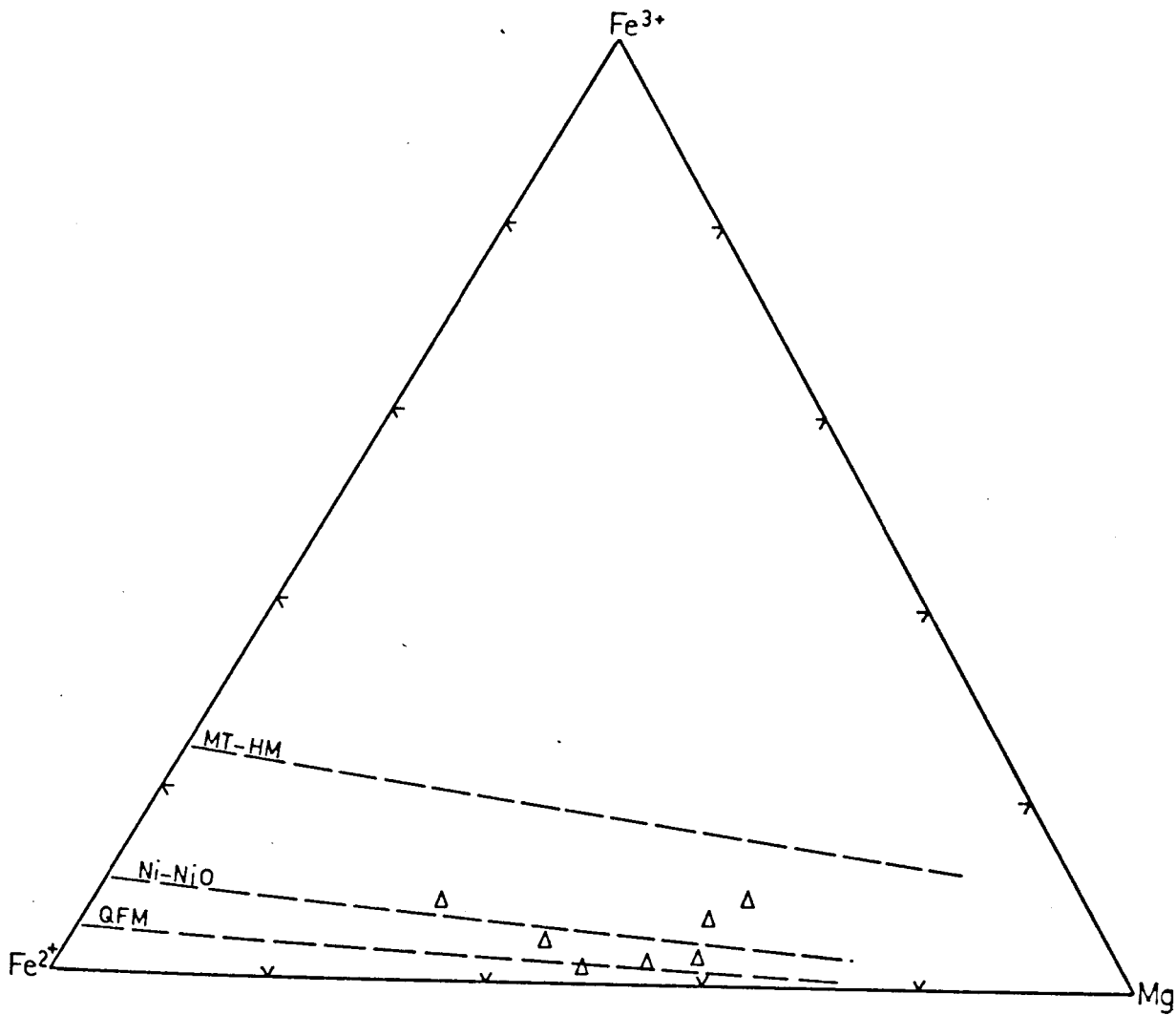


Figure 4.3 Compositions of New England Suite biotites (triangles) plotted in terms of  $Fe^{3+} - Fe^{2+} - Mg$  and compared to trends approximating the buffers magnetite-hematite (MT-HM), nickel-nickel oxide (Ni-NiO) and quartz-fayalite-magnetite (QFM).

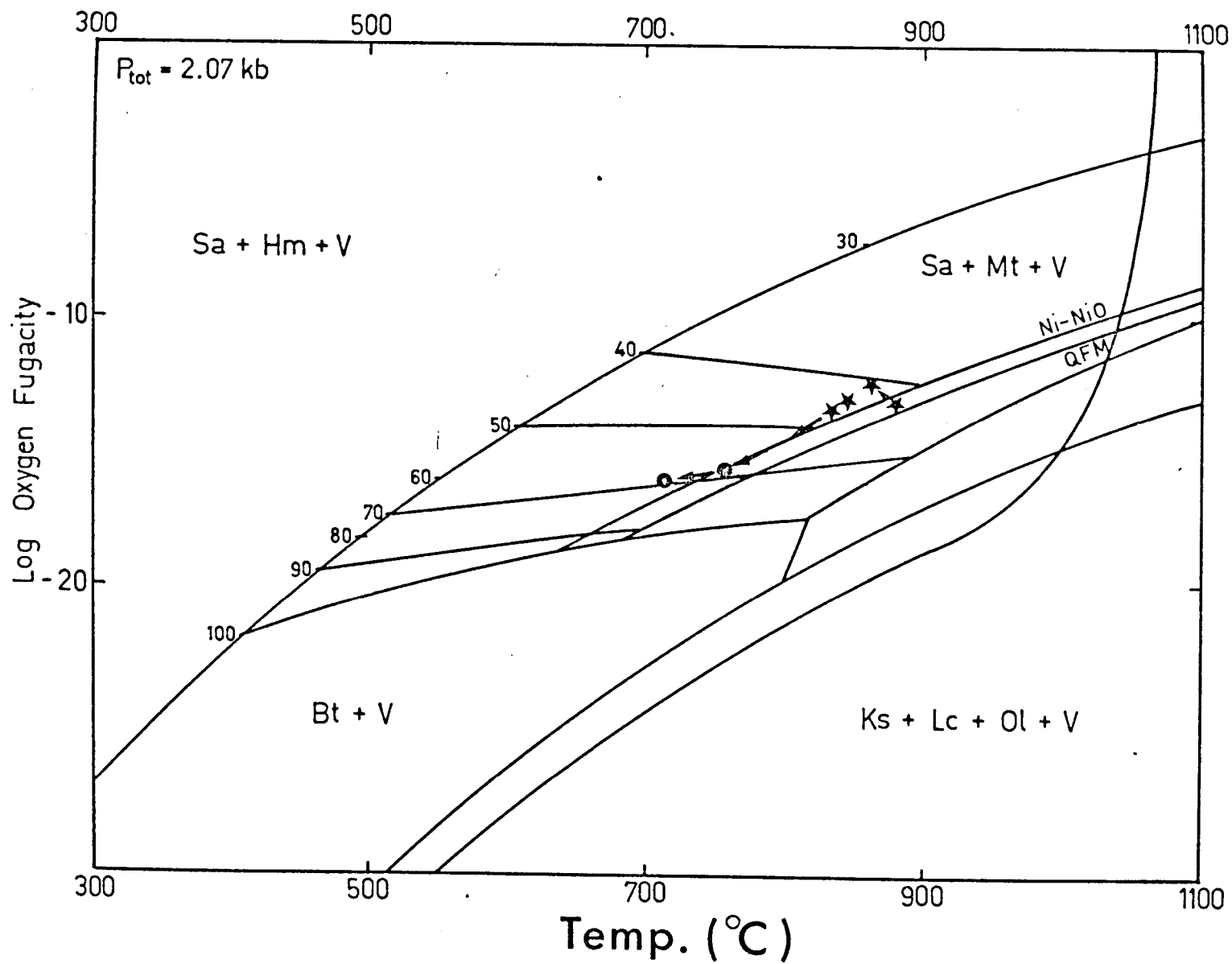


Figure 4.4 Compositions of New England Suite biotites compared to experimental data on biotite stability relations (after Wones and Eugster, 1965; Fig. 4). Arrows represent trend of increasing  $\text{SiO}_2$  in host rocks.

biotite-disappearance curves closely matched calculated biotite stability curves (Dodge *et al.*, 1969). The conclusion by Piwinski that the Sierra Nevada biotites with  $mg \leq 50$  became unstable at  $\sim 850^\circ\text{C}$  at  $P_{\text{H}_2\text{O}} \sim 2 \text{ kb}$  is totally consistent with the temperature estimates for the New England Suite, biotites especially since the experimental data further show that the biotite stability field is displaced to higher temperatures with higher  $mg$  and/or lower  $f_{\text{O}_2}$ .

#### 4.5.2 Amphibole

##### 4.5.2.1 Cummingtonite

(a) Occurrence and Chemistry. Cummingtonite occurs as six plutons, namely the Kentucky Ponds Diorite, Back Creek Microtonalite, Khartoun Tonalite, Uralla and Balala Granodiorites and Glenburnie Adamellite. In most cases it forms ragged, sometimes bladed cores within larger hornblende crystals. It is always pale in colour and often shows incipient(?) chloritic alteration or iron staining.

The cummingtonites of the New England Suite (Table 4.3) have a compositional range which exceeds that for the coexisting calciferous varieties ( $mg$  42-85). This is in sharp contrast to the restricted compositional range exhibited by the Hillgrove Suite cummingtonites. The cummingtonites from both suites are similar however, in that they are typically less magnesian than the coexisting calciferous amphiboles. One of the cummingtonites from the Uralla Granodiorite (anal. 5; Table 4.3) is a notable exception. Its highly magnesian composition, combined with its occurrence within a mafic clot suggests that it may have been derived from a highly magnesian orthopyroxene (see 4.5.3). The existence of highly magnesian cummingtonites in the Uralla Granodiorite was suggested originally by Vernon (1961).

The New England Suite cummingtonites differ from those of the Hillgrove Suite by containing no  $\text{TiO}_2$ , substantially less  $\text{Al}_2\text{O}_3$ , and more  $\text{CaO}$  than their Hillgrove Suite counterparts. However, with increasing  $\text{SiO}_2$  in the host rock (roughly analogous to decreasing  $MG$ ) there is a marked convergence in composition between the cummingtonites from both suites.

##### 4.5.2.2 Calciferous Amphiboles

(a) Occurrence and Chemistry. The presence (and occasional abundance) of calciferous amphiboles in most plutons is a characteristic feature of the New England Suite. The only exceptions are some of the microtonalites, e.g.

Table 4.3 Electron microprobe analyses and structural formulae of cummingtonites from the New England Suite.

Sample number <sup>†</sup> Catalogue number	(1) KPD 46101	(2) BCKT3 46104	(3) KT 46112	(4) UGD 46117	(5) UGD 46117	(6) BGD 46119	(7) GG2 46120
SiO <sub>2</sub>	54.8	54.6	54.3	53.9	56.3	52.8	52.6
Al <sub>2</sub> O <sub>3</sub>	.9	.4	.6	.6	3.6	.5	.5
Cr <sub>2</sub> O <sub>3</sub>	-	-	.13	-	.22	-	-
FeO*	21.0	21.7	22.6	23.3	7.61	26.8	28.3
MnO	1.06	1.10	1.35	1.50	-	2.59	2.20
MgO	18.1	17.6	16.8	16.7	23.7	12.6	11.6
CaO	2.22	1.24	1.27	1.27	.52	2.20	1.32
Na <sub>2</sub> O	-	-	-	-	-	-	.40
K <sub>2</sub> O	-	-	-	-	.07	-	.10
Total	98.08	96.54	97.05	97.27	92.02	97.49	97.02
mg	60.5	59.1	57.0	56.1	84.7	45.6	42.1
<i>Structural formulae (based on 23 oxygens per formula unit)</i>							
Si	7.925	8.026	7.996	7.959	8.020	7.978	8.031
Al <sup>iv</sup>	.075	-	.004	.041	-	.022	-
Al <sup>vi</sup>	.083	.072	.093	.064	.597	.061	.088
Cr	-	-	.015	-	.025	-	-
Fe	2.542	2.662	2.784	2.871	.907	3.393	3.610
Mn	.130	.138	.168	.188	-	.331	.284
Mg	3.899	3.846	3.688	3.666	5.034	2.839	2.629
Ca	.344	.196	.200	.201	.079	.359	.215
Na	-	-	-	-	-	-	.117
K	-	-	-	-	.012	-	.020
Σ (cations)	14.996	14.938	14.948	14.989	14.675	14.980	14.994

<sup>†</sup> see Appendix I.

\* Total Fe as FeO.

Terrible Vale Microtonalite, where it is totally absent.

At least two varieties of amphibole may be distinguished on the basis of colour - pale-green actinolite and brownish-green hornblende. Both varieties may show continuous or sharply defined colour zoning from core to rim. Texturally, the amphiboles of this Suite display the same features as outlined for the biotites, i.e. they occur in decussate aggregates, as discrete and often well-formed crystals, and as overgrowths on pyroxene. Another important feature of these amphiboles is the high proportion of accessory minerals that they enclose. Thin, prismatic apatite inclusions are most abundant but zircon, monazite and allanite are also present.

The majority of the amphiboles belong to the actinolite-hornblende series (Table 4.4). Si cations range from 6.76 to 7.90 per formula unit, where  $\Sigma \text{Ca} + \text{Na} + \text{K}$  is less than 2.5 cations. However, several analyses have  $\Sigma \text{Ca} + \text{Na} + \text{K} > 2.5$  cations; this indicates that two additional, coupled substitutions may have influenced amphibole compositions. Although the tschermakitic substitution ( $\text{R}^{3+}\text{Al}^{\text{IV}} \Leftrightarrow \text{Mg Si}$ ) usually predominate over the edenitic substitution ( $\text{Na}^{\text{A}}\text{Al}^{\text{IV}} \Leftrightarrow \square^{\text{A}}\text{Si}$ ) in amphiboles from calc-alkaline rocks (Leake, 1968). This relationship is not obvious in the New England Suite amphiboles, mainly because edenitic hornblendes occur as intermediate zones within individual amphiboles, e.g. in the Wards Mistake Adamellite (anal. 32; Table 4.4).

$\text{Al}_2\text{O}_3$ ,  $\text{TiO}_2$  and  $\text{Na}_2\text{O}$  show large continuous variations, 0.7%-10.1%, < 0.1%-2.78% and < 0.1%-2.27% respectively. In addition, MnO in these amphiboles increases sharply from mafic to felsic host rock compositions. These variations are more extensive than usually observed in calc-alkaline plutonic sequences, e.g. Chappell (1966), Haslam (1968) and Rhodes (1969), and exceed earlier estimates for these rocks, e.g. Flood (1971). However, it does confirm the observation of Vernon (1961) that some New England Suite plutons, e.g. the Uralla Granodiorite, contain amphiboles which range widely in their *mg* numbers, i.e. from *mg* 85 to *mg* 56. This wide variation in amphibole compositions within individual samples results in considerable scatter on a *mg*<sub>amph/rock</sub> correlation diagram (Fig. 4.5). Nevertheless, two features are clearly illustrated in Figure 4.5; firstly, amphibole *mg* numbers are usually higher than the corresponding *MG* number of the host rock, and secondly, actinolites are generally more magnesian than the coexisting aluminous amphibole, particularly where they occur as pale cores within hornblendes. The textural relationship displayed by actinolite and hornblende is analogous to the occurrence of remnant clinopyroxene

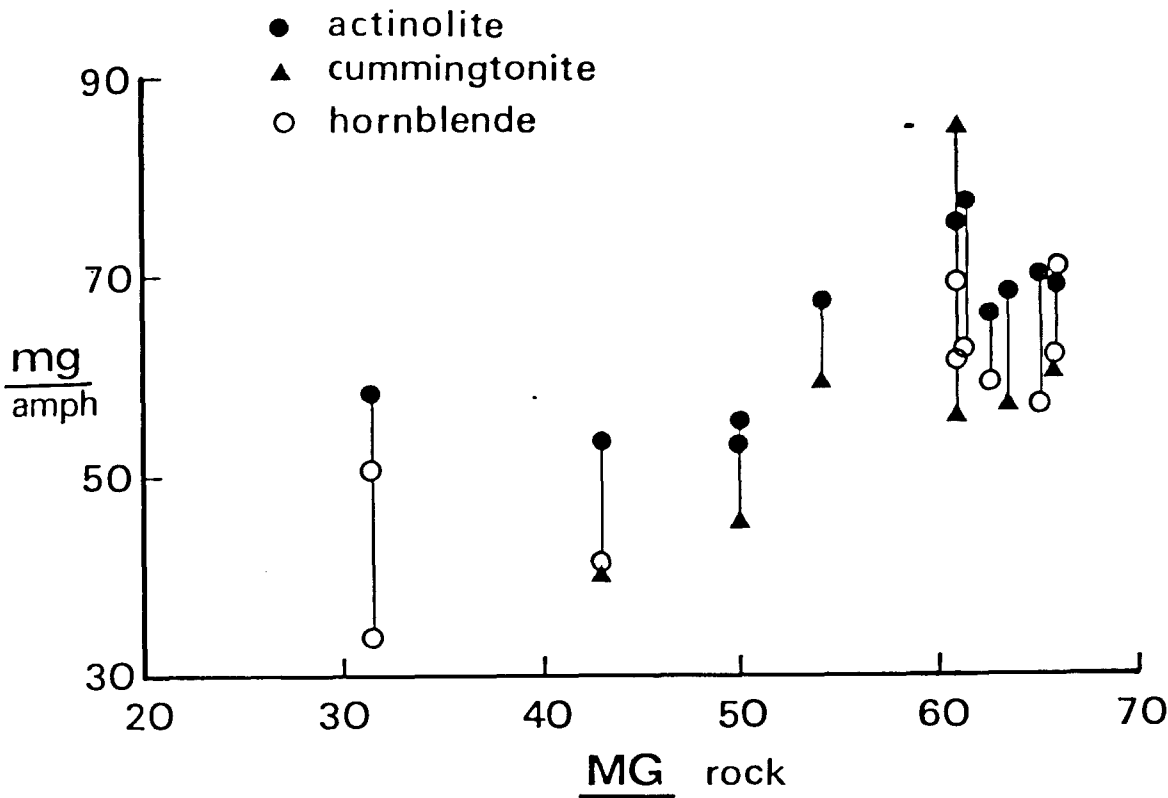


Figure 4.5 Plot of  $mg_{\text{amph}}$  v.  $MG_{\text{rock}}$  for New England Suite amphiboles.

within hornblende in many New England Suite plutons, e.g. the Shalimar and Wilhemshohe Tonalites. This textural evidence, combined with the fact, as shown in Figure 4.5, that  $mg$  numbers for actinolite and clinopyroxene are essentially the same, strongly infers that the actinolite has replaced the original clinopyroxene. A similar conclusion was reached by Albuquerque (1974) for actinolites from Aregos granitoids, Portugal.

(b) Distribution of Mg/Fe Between Amphibole and Biotite in the New England Suite Plutons. Using only the most common amphibole in apparent textural equilibrium with coexisting biotite,  $K_D^{\text{Fe-Mg}}$  values were calculated for amphibole-biotite pairs (Table 4.5). The data show a more-or-less continuous increase from the mafic rocks, which are reasonably uniform (0.60 - 0.63), to the felsic rocks (0.88). This is similar to the variation calculated from the data of Nockolds and Mitchell (1948) for the Scottish Caledonian granitoids and also for the granitoids from the Feather River area of the Sierra Nevada Batholith (Dodge *et al.*, 1968, 1969).



Table 4.4 Representative electron microprobe analyses of calciferous amphiboles from New England Suite plutons.

Sample No. <sup>†</sup>	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)	(16)	(17)	
	KD	KPD	KPD core	KPD rim	KPD	HM208	HM207	BCKT2	BCKT3	BCKT4	CQM	CQM	HGPT	WQM core	WQM mid	WQM rim	WQMX	
Catalogue No.	46100	46101	46101	46101	46101	46123	46124	46103	46104	46105	46113	46113	46114	46125	46125	46125	46125	
SiO <sub>2</sub>	49.0	45.0	55.1	47.8	46.0	50.1	48.5	46.7	53.6	50.8	51.9	47.9	51.1	50.9	44.3	46.0	54.3	
TiO <sub>2</sub>	1.19	2.40	-	1.27	1.83	.70	1.09	1.44	.12	.70	.51	.04	.71	.45	2.30	1.32	-	
Al <sub>2</sub> O <sub>3</sub>	5.7	9.7	1.5	6.7	9.9	4.1	5.7	8.2	1.3	4.1	4.4	6.7	4.5	3.5	10.0	7.0	1.7	
Cr <sub>2</sub> O <sub>3</sub>	.18	.17	-	.12	.61	-	-	-	-	.11	-	-	.19	.27	.10	-	-	
FeO*	13.14	13.99	13.50	14.06	10.36	12.37	15.15	14.43	13.01	14.45	12.88	14.98	13.49	15.80	17.64	22.65	15.26	
MnO	-	.15	.37	.24	-	.14	.15	.17	.21	.36	.33	.46	.26	.48	.27	.69	.76	
MgO	14.45	12.41	16.77	13.21	14.35	16.02	13.59	12.92	15.06	13.39	14.10	12.29	14.71	12.36	9.78	6.65	14.00	
CaO	11.57	12.17	10.48	11.61	12.65	11.09	11.31	11.30	12.17	11.63	12.11	11.80	11.27	11.60	11.16	10.29	11.08	
Na <sub>2</sub> O	.87	1.34	-	.73	1.41	1.10	1.22	1.45	-	.63	.86	1.43	.87	1.00	2.27	1.90	.59	
K <sub>2</sub> O	.50	.95	-	.64	1.06	.44	.54	1.31	.16	.40	.38	.69	.44	.31	.87	.75	.12	
Total	96.61	98.24	97.67	96.35	98.14	96.10	97.16	97.89	95.59	96.51	97.50	97.24	97.50	96.72	98.79	97.21	97.85	
mg	66.2	61.2	68.9	62.6	71.2	69.8	61.5	61.5	67.4	62.3	66.1	59.4	66.0	58.2	49.7	34.4	62.0	
<i>Structural formulae (based on 23 oxygens per formula unit)</i>																		
Si	7.217	6.628	7.888	7.105	6.671	7.382	7.180	6.888	7.883	7.502	7.531	7.116	7.436	7.572	6.610	7.093	7.883	
Al <sup>4</sup>	.783	1.372	.112	.895	1.329	.618	.820	1.112	.117	.498	4.69	.884	.564	.428	1.390	.907	.117	
Ti	.132	.266	-	.142	.199	.078	.122	.160	.013	.077	.056	.105	.078	.050	.258	.153	-	
Al <sup>6</sup>	.204	.304	.136	.290	.360	.097	.166	.317	.112	.215	.288	.296	.203	.188	.373	.363	.179	
Cr	.021	.020	-	.014	.070	-	-	-	-	.013	-	-	.022	.032	.012	-	-	
Fe	1.619	1.724	1.617	1.749	1.256	1.524	1.877	1.781	1.602	1.787	1.563	1.861	1.642	1.965	2.200	2.923	1.852	
Mn	-	.018	.046	.030	-	.018	.019	.021	.026	.046	.041	.058	.032	.061	.034	.090	.094	
Mg	3.172	2.724	3.581	2.927	3.103	3.517	3.000	2.843	3.305	2.951	3.049	2.720	3.191	2.738	2.174	1.531	3.028	
Ca	1.826	1.921	1.608	1.850	1.966	1.750	1.795	1.789	1.919	1.842	1.884	1.878	1.759	1.847	1.783	1.702	1.722	
Na	.249	.383	-	.211	.397	.313	.352	.414	-	.180	.241	.413	.245	.287	.657	.569	.166	
K	.094	.178	-	.122	.196	.083	.102	.246	.030	.075	.070	.130	.083	.058	.165	.147	.022	
Σ (cations)	15.318	15.538	14.988	15.325	15.547	15.381	15.432	15.567	15.005	15.186	15.190	15.461	15.255	15.227	15.656	15.477	15.063	

<sup>†</sup> See Appendix I.

\* Total Fe as FeO.

Table 4.4 continued. Representative electron microprobe analyses of calciferous amphiboles from New England Suite plutons.

Sample No. †	(18)	(19)	(20)	(21)	(22)	(23)	(24)	(25)	(26)	(27)	(28)	(29)	(30)	(31)	(32)	(33)	(34)	(35)	(36)	(37)	(38)
	WT	KT rim	SHT1 core	SHT1 mid	SHT2 core	SHT2 rim	UGD	UGD core	UGD rim	EGD core	GG1 rim	GG1 rim	GG1 core	WRA core	WRA mid	WMA rim	WMA	WMA core	WMA mid	WMA rim	ORLA
Catalogue No.	46116	46112	46110	46110	46108	46108	46117	46117	46117	46119	46120	46120	46120	46122	46122	46125	46125	46125	46125	46125	46129
SiO <sub>2</sub>	51.2	52.7	52.8	49.0	55.5	51.0	45.5	55.0	50.1	53.0	50.5	52.1	45.1	53.2	45.8	49.5	50.0	41.7	43.1	45.5	43.2
TiO <sub>2</sub>	.57	.53	.22	.70	-	.49	1.64	-	.45	-	.49	-	1.59	-	1.20	.69	-	2.78	1.76	1.17	1.27
Al <sub>2</sub> O <sub>3</sub>	4.1	3.3	2.2	4.9	1.5	4.3	9.4	.7	4.1	1.3	5.6	1.9	7.5	2.0	7.8	5.0	2.8	10.1	8.9	6.5	7.6
Cr <sub>2</sub> O <sub>3</sub>	.19	-	-	-	.50	.17	.57	-	-	-	.13	-	-	.12	.13	-	-	-	.14	.14	-
FeO*	14.81	13.17	14.76	16.55	9.19	14.99	11.17	10.54	14.80	17.29	16.92	18.61	20.64	12.31	16.22	15.17	18.63	16.45	16.18	17.15	24.31
MnO	.32	.27	.53	.61	.28	.28	.22	.33	.61	.54	.76	.80	.65	.38	.44	.47	.89	.32	.34	.50	1.37
MgO	14.63	15.41	13.65	12.05	17.72	13.60	13.96	17.40	13.91	11.91	10.67	11.40	8.01	15.90	12.13	13.55	11.14	10.77	11.66	11.59	6.75
CaO	10.87	11.05	11.75	10.71	12.23	10.59	12.73	13.45	11.79	12.45	11.16	11.36	10.87	12.23	11.72	11.51	11.65	10.93	10.88	11.20	10.54
Na <sub>2</sub> O	.92	.43	.33	.98	.41	.56	1.33	-	.56	-	1.10	.60	1.72	.33	1.51	1.06	.44	2.16	1.97	1.46	1.86
K <sub>2</sub> O	.32	.31	.18	.43	.15	.48	.95	.06	.38	.13	.41	.16	.94	.08	.90	.45	.14	.97	.86	.70	.90
Total	97.91	97.13	96.41	95.99	97.41	96.45	97.48	97.50	96.75	96.58	97.77	96.87	96.92	96.51	97.92	97.42	95.66	96.17	95.73	93.89	97.79
mg	63.8	67.6	62.2	56.5	77.5	61.8	69.0	74.6	62.6	55.1	52.9	52.2	40.9	69.7	57.1	61.4	51.6	53.9	56.2	54.6	33.1

Structural formulae (based on 23 oxygens per formula unit)

Si	7.451	7.643	7.783	7.378	7.871	7.535	6.681	7.869	7.426	7.901	7.453	7.803	6.938	7.751	6.840	7.308	7.621	6.396	6.610	6.960	6.756
Al <sup>4</sup>	.549	.357	.117	.622	.129	.465	1.319	.121	.574	.099	.547	.197	1.062	.249	1.160	.692	.379	1.604	1.390	1.040	1.244
Ti	.062	.058	.024	.079	-	.055	.182	-	.051	-	.054	-	.185	-	.135	.076	-	.321	.203	.135	.149
Al <sup>6</sup>	.162	.198	.270	.252	.118	.278	.311	-	.145	.120	.184	.130	.292	.091	.220	.179	.223	.220	.209	.134	.164
Cr	.022	-	-	-	.056	.019	.066	-	-	-	.015	-	-	.014	.015	-	-	-	.016	.017	-
Fe	1.804	1.597	1.820	2.083	1.091	1.850	1.371	1.261	1.834	2.155	2.088	2.332	2.658	1.501	2.025	1.873	2.376	2.110	2.076	2.195	3.182
Mn	.040	.034	.066	.078	.034	.035	.027	.039	.076	.068	.095	.101	.085	.047	.056	.058	.115	.042	.044	.065	.181
Mg	3.175	3.331	2.999	2.702	3.750	2.992	3.054	3.711	3.072	2.646	2.347	2.546	1.839	3.454	2.699	2.980	2.533	2.463	2.666	2.643	1.575
Ca	1.697	1.716	1.857	1.726	1.859	1.675	2.001	2.063	1.872	1.989	1.765	1.823	1.794	1.911	1.874	1.820	1.903	1.797	1.787	1.837	1.767
Na	.259	.120	.094	.286	.112	.160	.377	-	.160	-	.315	.174	.513	.094	.438	.304	.132	.644	.586	.432	.565
K	.059	.057	.034	.083	.028	.091	.179	.012	.071	.024	.078	.031	.184	.015	.172	.084	.026	.190	.169	.137	.180
Σ (cations)	15.279	15.110	15.063	15.290	15.047	15.155	15.567	15.076	15.280	15.002	15.191	15.136	15.548	15.127	15.633	15.375	15.207	15.788	15.757	15.594	15.763

† See Appendix I.

\* Total Fe as FeO.

Table 4.5 Fe/Mg distribution coefficients for New England Suite amphibole/ biotite pairs. Numbers in brackets denote values obtained for less commonly occurring amphiboles.

Sample number	Catalogue number	Fe/Mg <sub>amph</sub>	Fe/Mg <sub>bt</sub>	K <sub>D</sub> Fe/Mg <sub>amph/bt</sub>
KD	46100	.51	.85	.60
KPD	46101	.60	.82	.60 (.55-.77)
HM208	46123	.43	.69	.61
HM207	46124	.63	.71	.89
CQM	46113	.51	.81	.63
BCKT2	46103	.63	.49	1.29
BCKT3	46104	.61	.90	.68 (.53)
KPD7	46139	.51	.71	.72
WQM	46126	1.91	2.03	.94 (.35-.49)
WQMX*	-	.61	2.53	.24
WT	46116	.57	.81	.70
KT	46112	.48	.76	.63
SHT1	46110	.77	1.00	.77 (.61)
SHT2	46108	.62	.89	.69 (.32)
UGD	46117	.60	.86	.69 (.39-.52)
BGD	46119	.89	1.39	.64 (.58)
GG1	46120	1.45	1.53	.95 (.60)
WRA	46122	.62	.78	.79 (.55-.97)
WMA	46125	.86	1.13	.76 (.69)
ORLA	46129	2.02	2.28	.88

$$K_D = \frac{[\text{Fe/Mg}]^{\text{amph}}}{[\text{Fe/Mg}]^{\text{bt}}}$$

where Fe = total iron as Fe<sup>2+</sup>

† see Appendix I.

\* mafic clots within adamellite.

The often large differences in the K<sub>D</sub> values between the dominant amphibole (usually actinolitic hornblende) and the more aluminous amphiboles (e.g. pargasite) suggest that the ferromagnesian mineral assemblages of most New England Suite plutons are probably not in equilibrium. This conclusion is supported by the abrupt chemical variations sometimes observed in individual amphiboles.

Based on the mineral assemblages from Aregos granitoids, Albuquerque (1974) suggested that low amphibole/biotite  $K_D$ 's imply crystallization at relatively low temperatures and pressures. If this is so it infers that the mafic diorites and monzonites of the New England Suite crystallized at lower pressures and temperatures than the felsic granitoids. However, a number of mineralogical and petrological constraints, including (a) biotite and pyroxene equilibration temperatures, and (b) experimental phase equilibria in the "granite" tetrahedron (4.8), are not consistent with this view.

#### 4.5.3 Pyroxenes

##### 4.5.3.1 Orthopyroxene

###### (a) Occurrence and Chemistry

Orthopyroxene is a widespread, on occasions prominent, phenocryst in members of the New England Suite, often accompanied by clinopyroxene. The orthopyroxene occurs principally in two textural forms: (a) as discrete, euhedral to subhedral, prominent, and commonly pleochroic crystals up to 3 mm in length, and (b) as clear, untwinned, anhedral crystals surrounded by a reaction rim of pale, calciferous amphibole. In samples containing the second textural form, the rock surface is often pitted, where these mafic clots have preferentially weathered. The mantling of the orthopyroxene by amphibole strongly suggests that the magma had become sufficiently hydrous for amphibole to crystallize as a stable phase. Other examples of textural relationships which possibly indicate disequilibria include the mantling of orthopyroxene by Ca-rich pyroxene or intergrowths involving orthopyroxene with amphibole and/or biotite.

The orthopyroxene analyses (Table 4.6) are typical of orthopyroxenes from calc-alkaline hosts (e.g. Ewart, 1966; Best and Mercy, 1967; Haslam, 1968; Bryan *et al.*, 1972). The majority of compositions are hypersthene ( $En_{51-69}$ ; Fig. 4.6) but bronzites with compositions as magnesian as  $En_{87}$  are found in the Wilhemshohe and Khartoun Tonalites, near Uralla. Such magnesian compositions have not been documented in any other granitoids on a world-wide basis. Even in calc-alkaline extrusives bronzites approaching this composition have only been recorded from Japanese bronzite andesites and from a Hawaiian rhyodacite (Bauer *et al.*, 1973). Iron-rich orthopyroxenes are rare in the New England Suite plutons. Their only occurrence is in the iron-enriched variant of the Back Creek Microtonalite where ferrohypersthene ( $mg$  40) is recorded.

Table 4.6 Representative electron microprobe analyses, structural formulae and atomic proportions of orthopyroxenes from New England Suite plutons.

Sample No. †	(1) KD core	(2) KD rim	(3) KPD core	(4) KPD rim	(5) KPD	(6) HM208 core	(7) HM208 rim	(8) HM208	(9) HM207 core	(10) HM207 rim	(11) BCKT2 core	(12) BCKT2 rim	(13) BCKT2	(14) BCKT4	(15) HGPT rim	(16) TVPT core	(17) TVPT rim	(18) KT core	(19) KT	(20) KT rim	(21) WT core	(22) WT rim	(23) WT core	(24) WT rim
Catalogue No.	46100	46100	46101	46101	46101	46123	46123	46123	46124	46124	46103	46103	46103	46105	46114	46115	46115	46112	46112	46112	46116	46116	46116	46116
SiO <sub>2</sub>	51.9	51.9	52.7	52.3	53.5	53.0	52.6	53.9	53.1	52.1	53.0	51.9	54.0	50.5	52.3	52.1	52.6	57.5	57.5	55.6	55.6	55.2	53.7	52.6
TiO <sub>2</sub>	-	.17	.19	-	.11	.36	.30	.27	.35	.38	.32	.31	.29	.13	-	.32	-	-	-	-	.11	-	.12	.13
Al <sub>2</sub> O <sub>3</sub>	.8	.7	.9	.9	.4	.8	.7	1.0	.9	.7	1.1	1.3	1.5	.6	-	.4	.8	1.3	1.2	1.2	1.0	.9	1.4	.6
Cr <sub>2</sub> O <sub>3</sub>	.45	.11	-	-	.13	-	-	-	-	-	-	-	-	-	-	.14	.39	.40	.50	.45	.45	.30	.35	.19
FeO*	25.07	26.42	24.51	26.58	21.50	20.36	22.85	18.51	22.00	26.14	24.11	26.78	19.65	32.25	28.22	27.22	24.46	8.08	7.76	14.46	13.71	18.32	20.99	22.43
MnO	.60	.85	.66	1.02	.29	.45	.44	.42	.46	.88	.50	.74	.36	1.48	1.41	.81	.58	-	-	-	.16	.33	.41	1.20
MgO	20.19	18.98	20.08	18.20	22.53	23.81	21.85	24.71	21.83	18.31	19.62	17.33	22.28	13.76	17.23	18.16	20.72	32.08	32.37	27.56	27.22	23.62	22.07	18.59
CaO	.93	.87	1.04	1.02	.54	1.29	1.32	1.24	1.43	1.45	1.27	1.27	1.89	1.04	.88	.85	.47	.71	.73	.70	1.41	1.17	1.02	.96
Na <sub>2</sub> O	-	-	-	-	-	-	-	-	-	-	-	.23	-	.25	-	-	-	-	-	-	.30	.19	-	.25
K <sub>2</sub> O	-	-	-	-	-	-	-	-	-	.06	.10	.12	.07	-	-	.06	-	-	-	-	-	-	-	.08
mg	58.9	56.1	59.4	55.0	65.1	67.6	63.0	70.4	63.9	55.5	59.2	53.6	66.9	43.2	52.1	54.3	60.2	87.6	88.1	77.3	78.0	69.7	65.2	56.5
Structural formulae (based on 6 oxygens per formula unit)																								
Si	1.967	1.979	1.985	1.994	1.980	1.960	1.968	1.972	1.978	1.984	1.991	1.983	1.986	1.990	2.001	1.990	1.980	1.993	1.992	1.989	1.990	2.012	1.985	1.998
Al <sup>4</sup>	.033	.021	.015	.006	.020	.040	.032	.028	.022	.016	.009	.017	.014	.010	-	.010	.020	.007	.008	.011	.010	-	.015	.002
Ti	-	.005	.005	-	.003	.010	.008	.008	.010	.011	.009	.009	.008	.004	-	.009	-	-	-	-	.003	-	.003	.004
Al <sup>6</sup>	.003	.009	.023	.032	.040	-	-	.015	.016	.017	.040	.041	.051	.016	-	.007	.015	.045	.040	.039	.032	.039	.044	.024
Cr	.014	.003	-	-	.004	-	-	-	-	-	-	-	-	-	-	.004	.012	.011	.014	.013	.013	.009	.010	.006
Fe	.794	.842	.772	.847	.665	.630	.716	.567	.685	.833	.758	.855	.605	1.063	.908	.871	.770	.234	.225	.432	.410	.559	.649	.808
Mn	.019	.027	.021	.033	.009	.014	.014	.013	.014	.028	.016	.024	.001	.049	.046	.026	.019	-	-	-	.005	.010	.013	.039
Mg	1.140	1.078	1.128	1.034	1.242	1.314	1.220	1.348	1.212	1.040	1.099	.986	1.222	.808	.988	1.035	1.163	1.658	1.672	1.468	1.451	1.284	1.216	1.053
Ca	.038	.036	.042	.042	.021	.051	.053	.049	.057	.059	.051	.052	.075	.044	.036	.035	.019	.027	.027	.027	.054	.046	.041	.039
Na	-	-	-	-	-	-	-	-	-	-	-	.017	-	.019	-	-	-	-	-	-	-	-	-	-
K	-	-	-	-	-	-	-	-	-	.003	.005	.006	.003	-	-	.003	-	-	-	-	.021	.013	-	.019
Σ (cations)	4.008	4.000	3.991	3.987	3.985	4.013	4.009	3.999	3.994	3.991	3.978	3.991	3.975	4.003	3.989	3.991	3.997	3.975	3.978	3.980	3.990	3.971	3.977	3.994
Mg	57.8	55.1	58.1	53.8	64.4	65.9	61.4	68.7	62.0	53.8	57.6	52.1	64.3	42.2	51.1	53.3	59.6	86.4	86.9	76.2	75.8	68.0	63.8	55.4
Ca	1.9	1.8	2.2	2.2	1.1	2.6	2.7	2.5	2.9	3.1	2.7	2.8	3.9	2.3	1.9	1.8	1.0	1.4	1.4	1.4	2.8	2.4	2.1	2.1
Fe	40.3	43.1	39.7	44.0	34.5	31.5	35.9	28.8	35.1	43.1	39.7	45.1	31.8	55.5	47.0	44.9	39.4	12.2	11.7	22.4	21.4	29.6	34.1	42.5

All analyses normalized to 100%

† See Appendix I

\* Total Fe as FeO

- not detected

All orthopyroxene analyses are low in  $\text{Al}_2\text{O}_3$ ,  $\text{TiO}_2$  and  $\text{Na}_2\text{O}$ . Ti:Al ratios are also low (usually  $< 1:6$ ) and decrease even further with increasing  $\text{Cr}_2\text{O}_3$ . The low Ti:Al ratios in these New England Suite orthopyroxenes suggest that some  $\text{R}^{2+}\text{Cr}^{3+}\text{SiAlO}_6$  component (Boyd, 1971) may be present in these pyroxenes.  $\text{Cr}_2\text{O}_3$  indeed shows a mild correlation with  $\text{Al}^{\text{vi}}$  and ranges from below detection limit in many samples to 0.50%. In contrast, MnO increases with increasing *mg* number from  $< 0.1\%$  MnO in the magnesian bronzites, to 1.48% in the ferro-hypersthene. CaO is also low through the compositional range and defines a singular trend that is close to and essentially parallel with the Mg-Fe join. Because the New England Suite orthopyroxenes are chemically homogeneous and often lack exsolution of a Ca-rich phase, this trend is consistent with crystallization at low temperatures. However, Nwe (1976), Coleman (1978) and von Gruenewaldt and Weber-Diefenbach (1977) demonstrated that a "typically calc-alkaline" trend may also be obtained for Ca-poor pyroxenes from tholeiitic layered intrusions if the pyroxenes have re-equilibrated at subsolidus conditions.

(b) Zoning

Most orthopyroxenes are moderately zoned and show a progressive decrease in *mg* from core to margin. The only exception is orthopyroxene from the Terrible Vale Microtonalite which shows an increase of 6 atom % *mg* towards the margins. Although the average core-rim variation is about 6-8 atom % *mg*, slightly higher variations (up to 11%) are recorded in the more magnesian pyroxenes of this Suite.

4.5.3.2 Clinopyroxene

(a) Occurrence and Chemistry

Clinopyroxene also occurs in many New England Suite plutons. It is generally subhedral and ranges from colourless to pleochroic jade-green. Four textural varieties of clinopyroxene are readily recognized:

- (a) discrete microphenocrysts up to 3 mm in length;
- (b) very small, anhedral "globules" enclosed by plagioclase;
- (c) intergrowths with, or overgrowths on, orthopyroxene; and
- (d) ragged crystals forming cores to many large, well-developed amphiboles.

As in the case of the orthopyroxenes, it is not possible to consistently relate these textural varieties to specific rock types. However, the "globules"



occur only in the siliceous granitoids, e.g. Wards Mistake Adamellite, whereas the microphenocrysts are best developed in the fine-grained monzonites and mafic tonalites.

Clinopyroxene is present in all rocks containing orthopyroxene. Its apparent absence in the Khartoun Tonalite (Table 4.7) reflects its patchy distribution throughout the pluton (*see* Vernon, 1961).

Representative clinopyroxene analyses and structural formulae are listed in Table 4.7 and displayed in terms of Ca-Mg-Fe in Figure 4.6. Most analyses cluster around where the diopside-salite-augite fields meet. This is typical for calc-alkaline clinopyroxenes, although the New England Suite pyroxenes are perhaps slightly more magnesian than average. The Kentucky Ponds Diorite pyroxene which is distinctly diopsidic (*mg* 82), the iron-enriched pyroxene (*mg* 58) of the Back Creek Microtonalite (sample BCKT4), and the augites from the Highlands Monzonite and Back Creek Microtonalite (sample BCKT3) are notable exceptions to this rather restricted field.

Other typical features of the New England Suite clinopyroxenes are high CaO (usually > 21 wt. %) and variable but usually low TiO<sub>2</sub>, alkalis and Al<sub>2</sub>O<sub>3</sub>. These low minor element contents parallel those of the coexisting orthopyroxene. However, Cr<sub>2</sub>O<sub>3</sub> concentrations in the clinopyroxenes are somewhat anomalous because they are frequently *lower* than abundances in the coexisting orthopyroxenes and thus contrary to the known partitioning behaviour for Cr<sub>2</sub>O<sub>3</sub> between coexisting pyroxenes. An additional features of the New England Suite clinopyroxenes is their low cation totals in the structural formulae (rarely > 4.00); implying small or negligible Fe<sup>3+</sup>.

(b) Zoning

The maximum compositional zoning of clinopyroxenes from this Suite is ~ 10 atom % *mg* from core to rim. Zoning is usually normal, with respect to Mg and Fe, although small compositional fluctuations within individual crystals are not uncommon. Carmichael (1964) suggested that crystallization of iron-titanium oxides may temporarily reverse Fe/Mg relations in the melt and result in intra-crystal fluctuations of *mg*. However, the petrographic evidence to support the mechanism proposed by Carmichael is lacking, firstly because magnetite is absent from many of the samples, and secondly, because it cannot be confidently demonstrated that the modally insignificant ilmenite actually crystallized from the melt. In the absence of additional data it is suggested that the small intra-crystal composition fluctuations in some of the pyroxenes may reflect their possibly relict nature.

Table 4.7 Representative electron microprobe analyses, structural formulae and atomic proportions of clinopyroxenes from New England Suite plutons.

Sample No. †	(1) KD	(2) KPD core	(3) KPD mid	(4) KPD core	(5) HMS core	(6) HMS rim	(7) 207	(8) 207 core	(9) BCKT2 core	(10) BCKT2 rim	(11) BCKT3 core	(12) BCKT3 rim	(13) BCKT4	(14) SHI2 mid	(15) WT	(16) HGPT rim	(17) TVPT core	(18) TVPT rim	(19) WRA	(20) WMA	(21) WMA
Catalogue No.	46100	46101	46101	46101	46123	46123	46124	46124	46103	46103	46104	46104	46105	46108	46116	46114	46115	46115	46122	46125	46125
SiO <sub>2</sub>	52.8	52.8	53.6	53.5	51.9	51.6	52.8	52.1	51.0	53.2	51.5	52.9	51.9	53.4	53.4	53.4	53.1	53.1	53.0	51.9	52.5
TiO <sub>2</sub>	.11	.47	.12	-	.84	.81	-	.49	.62	-	.91	-	.12	-	-	-	.11	.13	-	-	-
Al <sub>2</sub> O <sub>3</sub>	.9	4.2	1.6	.3	2.1	1.8	.5	1.8	5.1	1.4	5.3	.7	1.2	.9	.3	.4	.7	.7	-	.9	1.1
Cr <sub>2</sub> O <sub>3</sub>	.19	.73	.39	.29	.39	.50	.29	.27	.30	.17	.32	.26	-	.36	.23	.28	.32	.31	.30	.15	.12
FeO*	10.19	7.18	6.30	9.17	8.74	9.28	10.88	11.43	7.74	11.48	9.36	10.46	13.99	8.72	10.11	9.77	9.08	11.56	9.27	11.85	10.29
MnO	.25	.14	-	-	-	.13	.24	.16	-	.22	.12	.51	.67	.28	.30	.21	.16	.24	.61	.81	.67
MgO	13.18	17.05	16.25	13.52	16.61	15.19	13.02	14.15	14.01	12.38	14.58	12.01	10.71	13.06	13.90	13.67	14.08	13.66	12.63	12.29	13.32
CaO	22.14	16.75	21.73	23.24	19.25	20.44	21.86	19.25	20.69	20.86	16.66	22.60	21.25	23.20	21.59	22.27	22.45	20.38	23.81	22.07	22.09
Na <sub>2</sub> O	.22	.56	-	-	.16	.20	.32	.33	.33	.34	.98	.42	.19	-	.21	-	-	-	.30	-	-
K <sub>2</sub> O	-	.16	-	-	-	-	.09	.09	.06	-	.27	.12	-	.12	-	-	-	-	.08	-	-
mg	69.7	81.0	82.2	72.4	77.2	74.4	68.1	68.8	76.4	65.8	73.5	67.2	57.7	72.8	71.1	71.3	73.4	67.8	70.8	64.9	69.8
<i>Structural formulae (based on 6 oxygens per formula unit)</i>																					
Si	1.981	1.923	1.967	1.998	1.920	1.924	1.989	1.951	1.883	1.994	1.990	1.993	1.980	1.994	1.998	1.996	1.983	1.991	1.996	1.970	1.972
Al <sup>4</sup>	.019	.077	.033	.002	.080	.076	.011	.049	.117	.006	.010	.007	.020	.006	.002	.004	.017	.009	-	.030	.028
Ti	.003	.013	.003	-	.023	.023	-	.014	.023	-	.025	-	.004	-	-	-	.003	.004	-	-	-
Al <sup>6</sup>	.021	.103	.038	.012	.014	.005	.011	.029	.105	.056	.219	.026	.033	.032	.011	.013	.012	.020	-	.010	.019
Cr	.006	.021	.011	.009	.001	.015	.009	.008	.009	.005	.009	.008	-	.011	.007	.008	.009	.009	.009	.004	.004
Fe	.320	.219	.193	.287	.271	.290	.343	.358	.239	.360	.289	.330	.446	.272	.316	.306	.284	.363	.292	.376	.323
Mn	.008	.004	-	-	-	.004	.008	.005	-	.007	.004	.016	.022	.009	.010	.007	.005	.008	.019	.026	.021
Mg	.737	.926	.889	.753	.916	.845	.731	.790	.772	.692	.801	.675	.609	.727	.776	.762	.783	.764	.709	.695	.746
Ca	.890	.654	.855	.931	.763	.817	.882	.773	.819	.839	.658	.913	.869	.928	.866	.893	.898	.819	.961	.897	.890
Na	.016	.040	-	-	.012	.015	.023	.024	.024	.025	.070	.031	.014	-	.015	-	-	-	.022	-	-
K	-	.008	-	-	-	-	.004	.004	.003	-	.013	.006	-	.006	-	-	-	-	.004	-	-
Σ (cations)	4.001	3.988	3.989	3.992	4.010	4.012	4.010	4.006	3.992	3.985	3.997	4.005	3.997	3.984	4.000	3.990	3.994	3.986	4.012	4.008	4.003
Mg	37.9	51.5	45.9	38.2	47.0	43.3	37.4	41.1	42.2	36.6	45.8	35.2	31.7	37.7	39.6	38.9	39.9	39.3	36.1	37.7	38.1
Ca	45.7	36.4	44.1	47.2	39.1	41.9	45.1	40.2	44.8	44.3	37.6	47.6	45.1	48.2	44.2	45.5	45.7	42.1	49.0	45.8	45.4
Fe	16.4	12.2	10.0	14.5	13.9	14.8	17.5	18.6	13.1	19.0	16.5	17.2	23.2	14.1	16.2	15.6	14.4	18.6	14.9	16.6	16.5

All analyses normalized to 100%.

† See Appendix I

\* Total Fe as FeO

- not detected

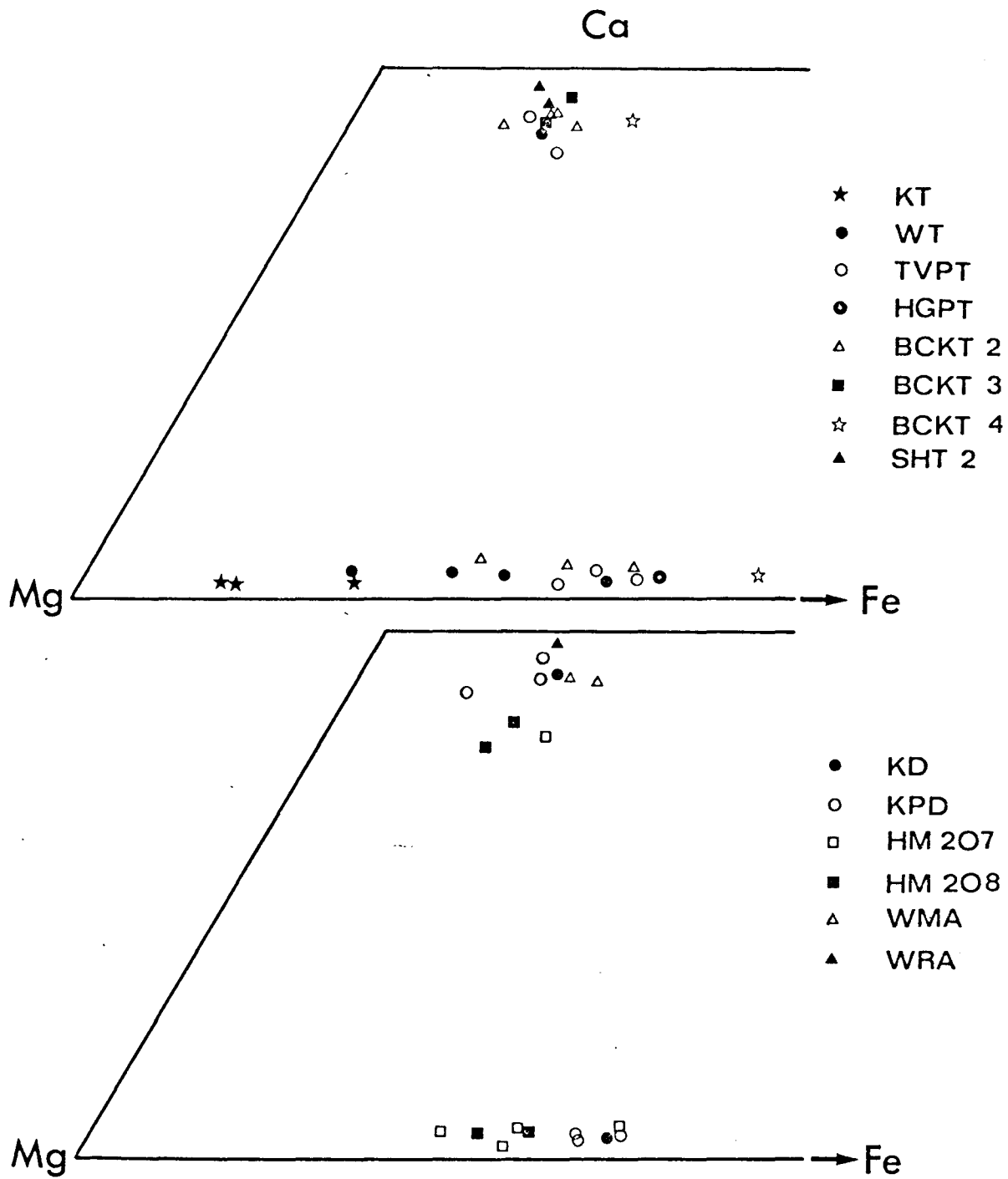


Figure 4.6 Compositions of coexisting clinopyroxenes and orthopyroxenes from New England Suite plutons plotted on pyroxene quadrilateral. Tie lines joining coexisting pairs have been omitted for clarity.

#### 4.5.3.3 Exsolution in New England Suite Pyroxenes

Two sets of very fine ( $\sim 1\mu$ ), exsolution lamellae are commonly present in the New England Suite clinopyroxenes. The exsolution closely resembles that of granulitic pyroxenes with the exception that, in the latter, the additional development of a second set of pigeonite lamellae not quite parallel to (100) is possible (Jaffe *et al.*, 1975).

It is generally accepted that lamellae exsolved from clinopyroxene parallel to (100) are orthopyroxene and any less prominent set parallel to (001) are pigeonite (Poldervaart and Hess, 1951). That pigeonite should be an exsolved phase at conditions well below that of the orthopyroxene-pigeonite inversion curve seems anomalous. However, Jaffe *et al.* (*loc.cit.*) proposed a mechanism for the metastable nucleation of pigeonite in calc-alkaline, plutonic pyroxenes which is supported by optimal phase boundary constraints and atomic structure considerations. Nevertheless, the possibility that pigeonite may have a lower stability field under certain conditions should also not be discounted (Boyd and England, 1965; Kuno, 1966; Smith, 1969; Grover, 1972).

By comparison, the orthopyroxenes display only one set of exsolution lamellae, parallel to (100). These lamellae are usually very fine ( $\sim 1\mu$ ) but vary in density and distribution from pluton to pluton and even from grain to grain. The well-developed, single set of (100) lamellae in the Highlands Monzonite orthopyroxene was interpreted by Neilson (1970) as indicating that the orthopyroxenes were probably primary and not inverted pigeonites.

A second and minor form of exsolution in clinopyroxenes can also be recognized in some mildly metamorphosed intermediate to acid members of the New England Suite, e.g. the marginal facies of the Terrible Vale Microtonalite. The exsolution resembles a schiller texture but it may also be interpreted as the "tweed structure" recently recognized in lunar pyroxenes (Nord *et al.*, 1973, 1976; Champness and Lorimer, 1976) and in a dolerite from the Whin Sill (Copley, 1973). This particular texture has two modulations parallel to (100) and (001) and is thought to result from the process of spinoidal decomposition (Cahn, 1968; Hilliard, 1970). The decomposition mechanism in atomic structures which are similar (such as pigeonite and augite) results in the precipitation of the exsolved phase in the peaks and troughs of sinusoidal composition waves. Whereas 'normal' lamellae were interpreted in classical terms as products of nucleation and growth under conditions of slow cooling, "tweed structure" is probably related to rapid cooling. However, the discovery of this structure in the Whin Sill and its assumed presence in metamorphosed samples of the

Terrible Vale Microtonalite implies that it may also develop under conditions of less rapid cooling, or possibly by rapid metamorphism of a previously homogeneous clinopyroxene.

#### 4.5.3.4 Distribution of Mg and Fe in New England Suite Pyroxenes Pairs

##### (a) Introduction

Although considerable attention has been directed to elemental distribution between coexisting pyroxenes, most studies to date have been confined to either high-grade metamorphic environments or to layered tholeiitic intrusions and very little has been published on intermediate and felsic calc-alkaline rocks.

##### (b) Mg-Fe Distribution

Table 4.8 lists the  $K_D$  values for New England pyroxene pairs. The most important feature of this tabulation is that the presentation of the data according to specific mineral domains, i.e. cores and rims. As shown in Table 4.8 the differences in  $K_D$ 's between these domains are often very large and with only one exception (Terrible Vale Microtonalite) core  $K_D$ 's are markedly higher

Table 4.8 Mg-Fe distribution between New England Suite pyroxene pairs

Rock Unit	Sample number	core $K_D^{\text{Mg-Fe}}$ opx-cpx	rim $K_D^{\text{Mg-Fe}}$ opx-cpx	$M_G$ rock
Kentucky Diorite	(KD)	.62	.56	64.1
Kentucky Ponds Diorite	(KPD)	.56 - .71	.47	65.7
Highlands Monzonite	(HM208)	.61	.50	66.9
Highlands Monzonite (chilled margin)	(HM207)	.80	.57	60.1
Back Creek Microtonalite	(BCKT2)	.75	.60	61.2
Back Creek Microtonalite	(BCKT4)	.56	-	47.3
Terrible Vale Microtonalite	(TVPT)	.56	.72	65.1
Harnham Grove Microtonalite	(HGPT)	.44	-	60.8
Wilhemshohe Tonalite	(WT)	.76	.53	60.0

than rim  $K_D$ 's. Considering the gross similarities and close association of all the host rocks, the combined range of  $K_D$  values for both domains is surprisingly wide (0.44 - 0.80) and virtually embraces all published  $K_D$  data, including metamorphic pyroxenes. The average rim  $K_D$ 's of the New England Suite pyroxene pairs compare most closely to granulites and charnockites, particularly the charnockitic series of Madras (0.51 - 0.56; Howie, 1955), granulite facies banded gneisses of the Pennsylvania and Delaware piedmont province (Clavan *et al.*, 1954), two-pyroxene plagioclase gneisses from Scotland (Muir and Tilley, 1958), as well as the granulitic rocks described by Subramaniam (1962), Leelanandam (1967), Himmelberg and Phinney (1967), Ward (1969), and the well-documented Australian granulite terranes of Broken Hill (Binns, 1962, 1967) and Quairading, Western Australia (Davidson, 1968). By comparison,  $K_D$  calculations of pyroxene pairs from diorites, monzonites, monzodiorites, tonalites and more felsic intrusives, comparable to the New England Suite samples, are exceedingly rare. Best and Mercy (1967) reported  $K_D$ 's as low as 0.57 for pyroxene pairs from calc-alkaline gabbros from the Guadelupe Complex but argued on the basis of textural data that these "low" values resulted from subsolidus re-equilibration.

(c) Temperature and Pressure Dependence on Mg-Fe Partitioning

Although stressing the importance of assuming equilibrium crystallization (Kretz, 1963) attributed the difference in  $K_D$ 's between metamorphic and igneous pyroxenes mainly to temperature, citing as evidence the difference between metamorphic and igneous parageneses. Many early workers favoured this primary dependence on temperature; however crystal chemistry studies (Burns, 1968) have shown that the dependence may not be as strong as originally assumed. A comparison of Tables 4.8 and 4.9 shows that the correlation between  $K_D$  and respective crystallization temperatures is only moderately good.

Atkins (1967) suggested that pressure differences (supported by field evidence) may have been largely responsible for variations in  $K_D$  between the Bushveld and Skaergaard pyroxenes. Kretz (1963) had also suggested that  $K_D$  values would be depressed by high pressure but agreed with Ramberg and De Vore (1951) that although pressure dependence was potentially significant, the range in crustal pressures was too small to have a dramatic effect on Mg-Fe distribution in pyroxenes such as those of the New England Suite plutons.

(d) Effect of Site Occupancy on  $K_D$

In order to relate  $K_D$  to pressure and temperature it is assumed that pyroxenes display ideal solid solution behaviour (Sahama and Torgeson, 1949;



Hess, 1952; Kuno, 1954; Brown, 1960; Mueller, 1960, 1961; Kretz, 1961, 1963; and others). With advances in pyroxene crystal chemistry, this assumption has received much criticism. For example, Binns (1962) pointed to the non-ideality of Broken Hill basic granulite pyroxenes, and similarly Matsui and Banno (1965) and Banno and Matsui (1966) described a strong compositional dependence for their  $K_D$ 's. In a study of pyroxene distribution coefficients from Quairading basic granulites (Western Australia), Davidson (1968) showed that  $K_D$  depended on the composition of the orthopyroxene, particularly in the case of orthopyroxenes more iron-rich than  $En_{60}$ . This implied that the distribution of Mg and Fe became concentration dependent at a certain stage, departed from ideality, and resulted in a sigmoidal distribution curve.

Crystal field studies of pyroxenes (Ghose, 1965; Burns, 1968; Virgo and Hafner, 1970; Saxena and Ghose, 1971) have also confirmed the importance of site ordering to the distribution of Mg and Fe. From these studies it has been shown that the non-ideal behaviour in the distribution may be attributed primarily to the fact that both orthopyroxene and clinopyroxene have two non-equivalent sites ( $M_1$  and  $M_2$ ). The  $M_1$  site in orthopyroxene is small, and therefore occupied mainly by  $Mg^{2+}$ . This means that essentially all the  $Fe^{2+}$  is in the relatively large and distorted octahedral  $M_2$  site. By comparison, the  $M_2$  site in clinopyroxene contains all the  $Ca^{2+}$ . Because the competitive nature of  $Ca^{2+}$  for this site restricts the entry of  $Fe^{2+}$  into  $M_2$  it directly influences the distribution of Mg and Fe between the  $M_1$  and  $M_2$  sites, which in terms of intracrystalline ion exchange may be expressed as  $Fe(M_2) + Mg(M_1) \leftrightarrow Fe(M_1) + Mg(M_2)$  (Saxena, 1973).

(e) Influence of Bulk Rock Composition on  $K_D$

Maxey and Vogel (1974) proposed that, because pyroxene compositions usually reflect host rock compositions, the latter must be the determining factor in  $K_D$  variation, and not the temperature of crystallization. They furthermore suggested that  $K_D$ 's may be sensitive to variations in the oxidation state of the rock. For the New England Suite pyroxene pairs neither alternative appears to influence the Mg-Fe distribution, as emphasized by the fact that there are significant variations in the Mg-Fe distribution between different crystal domains.

(f) Anomalies in Mg-Fe Distribution

Because the apparently anomalous  $K_D$ 's of some pyroxene pairs (Table 4.8)

cannot be readily attributed to the above influencing factors, it is suggested that the particular pyroxene pairs simply represent disequilibrium assemblages. Textural and chemical evidence supporting this suggestion include:

- (a) distinctive pyroxene 'generations' within one rock type;
- (b) reaction relationship between pyroxene(s) and hydrous melt; and
- (c) overgrowths of clinopyroxene on orthopyroxene.

Consequently, it is critical that a pyroxene assemblage, which most closely approaches equilibrium, be selected when calculating the Mg-Fe distribution. For example an extreme range of  $K_D$  values (0.53 - 1.41) may be obtained by combining disequilibrium pyroxenes from the Wilhelmshohe Tonalite. This is illustrated by a plot of  $K_D$  versus  $mg_{\text{Opx}}$  (Fig. 4.7) which stresses the importance of using low- $mg$  orthopyroxenes in  $K_D$  calculations.

The reversed trend of  $K_D$ 's exhibited by the core and rim compositions of the Terrible Vale Microtonalite pyroxenes is also anomalous. If temperature is an influencing factor on  $K_D$ 's (assuming an equilibrium pyroxene assemblage), a reversal could imply that temperature increased during intrusion (see below).

The unusually low  $K_D$  for Harnham Grove Microtonalite pyroxenes (0.44) may be attributed to subsolidus equilibration resulting from mild metamorphism by the adjacent Kentucky Ponds Diorite. However, Mori (1978) cautioned against this approach, suggesting that present knowledge of activity composition relations at low temperatures is too limited to make this assumption.

#### 4.5.3.5 Pyroxene Thermometry

Temperatures based on the distribution of elements between coexisting clinopyroxenes and orthopyroxenes are presented in Table 4.9. The Wood and Banno (1973) thermometer is favoured over that of Wells (1977) despite the fact that recent experimental data on site occupancies, and on the diopside-enstatite miscibility gap (Nehru and Wyllie, 1974; Mori and Green, 1975, 1976; Lindsley and Dixon, 1976), suggest that some of the empirical corrections used by Wood and Banno may be slightly incorrect. However, as Stormer and Whitney (1977) pointed out, the more sophisticated thermometry models do not give better results below 900°C, and indeed, may give geologically unreasonable temperatures at times.

Table 4.9 lists temperatures for various combinations of core-rim pyroxene compositions. Pyroxenes were selected for their representivity and apparent

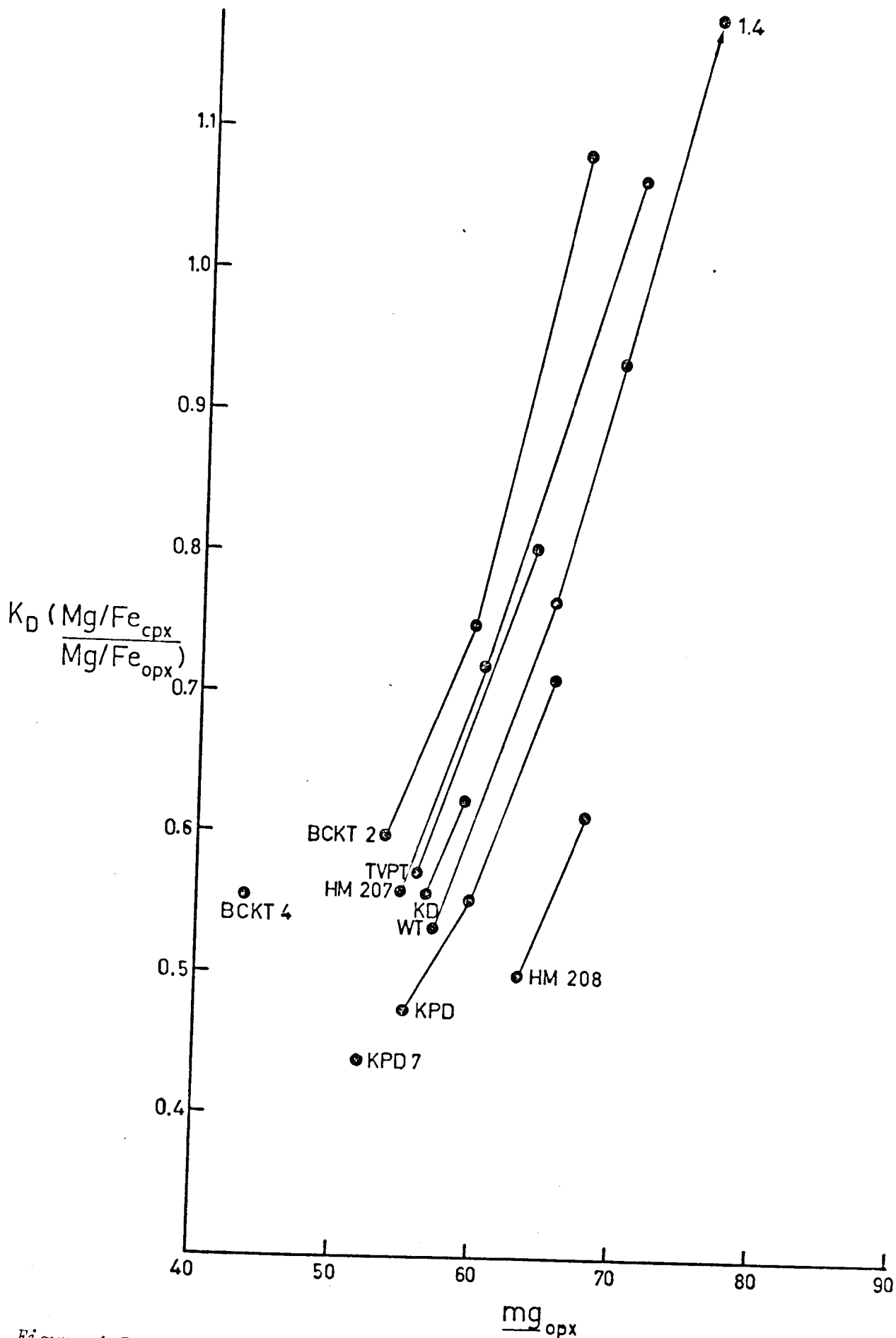


Figure 4.7 Plot of Mg-Fe distribution coefficient between coexisting pyroxene pairs in New England Suite plutons against  $mg_{opx}$  to illustrate the effect of compositional zoning within orthopyroxene on  $K_D$ .