Comparative amphibole chemistry of the Monteregian and White Mountain alkaline suites, and the origin of amphibole megacrysts in alkali basalts and lamprophyres*

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Abstract

Amphiboles in dykes from the dominantly silica-undersaturated Monteregian series range from pargasitic megacrysts and xenocrysts to kaersutitic, pargasitic and hastingsitic phenocrysts, groundmass prisms and reaction rims. Amphiboles in dykes and plutons from the silica-oversaturated White Mountain Magma Series range from kaersutite, through hornblende, hastingsite and edenite, to sodic–calcic and sodic varieties. This contrast between the amphiboles from the two series is probably a reflection of differing melt silica activity and is a useful petrologic discriminant. In most cases, pargasitic amphibole megacrysts from Monteregian monchiquites reflect the Mg numbers of their host rocks and are considered cognate. The megacrysts are lower in Ti and higher in Mg and Al than their phenocryst mantles and rims. This is probably a result of higher pressures of formation.

KEYWORDS: amphiboles, basalts, lamprophyres, Monteregian Province, White Mountain, Quebec, New England.

Introduction

THE mafic magmas of the Mesozoic Monteregian and White Mountain alkaline plutonic suites of southern Quebec and New England have compositions that straddle the plagioclase–olivine–clinopyroxene thermal divide (Yoder and Tilley, 1962). This divide separates magmas that fractionate to silica oversaturated residua (Coombs Trend) from those that evolve towards silica undersaturated differentiates (Kennedy Trend; Miyashiro, 1978). Fractionation of amphibole may allow derivative magmas to cross this thermal divide. Amphibole megacrysts and phenocrysts occur in Monteregian mafic dykes, and amphibole is a common phase in both the Monteregian and White Mountain suites. Consequently, knowledge of the compositions and occurrences of amphibole are essential adjuncts to the erection of petrogenetic models.

In this report I present textural and compositional data for amphiboles from Monteregian and White Mountains dykes. Published analyses from the associated plutonic rocks will also be considered. Over 125 dykes were examined; of these, 22 Monteregian dykes (1 alnöite, 1 mica peridotite, 1 aillikite from Ile Bizard, 4 camptonites, 5 monchiquites, 10 intermediate and leucocratic dykes) and 9 dykes from the White Mountain Magma Series were analysed for their constituent minerals. The petrography and pyroxene chemistry of the dykes is described in more detail elsewhere (Bédard, 1985*a*; Bédard *et al.*, 1987 and 1988). Petrographic descriptions and the full data set will be supplied upon receipt of a $5\frac{3}{4}''$ IBM-formatted diskette.

The Monteregian Petrographic Province and the White Mountain Magma Series

Triassic rifting of the central North Atlantic is associated with continental tholeiites (e.g. Weigand and Ragland, 1970) and localized alkaline magmatism (McHone, 1978; McHone and Butler, 1984; Bédard, 1985*b*). The most abundant Triassic and Jurassic igneous rocks in New Hampshire are the granitic and syenitic plutons and ring-complexes of the White Mountain Magma Series (Fig. 1; Chapman and Williams, 1935; Chapman, 1976), while mafic alkali dolerite dykes predominate in coastal Maine, New Hampshire and Massachusetts.

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Alkaline magmatism recurred in the Cretaceous and led to the emplacement of the Monteregian Petrographic Province in southern Quebec (Adams, 1903; Philpotts, 1974), the younger White Mountains complexes, and the Kelvin Seamounts. In this paper, the term 'White Mountain Magma Series' (WMMS) is taken to include both Triassic and Jurassic members.

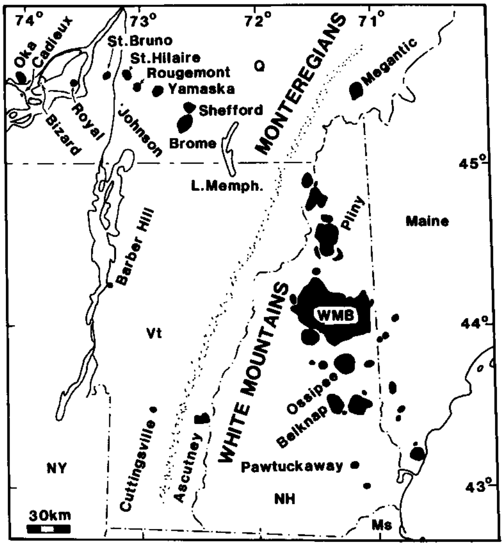


FIG. 1. Map of Monteregian-White Mountains area adapted from McHone (1978) and Bédard (1985*b*). Mesozoic plutonic rocks are in black. WMB = White Mountain Batholith, Q = Quebec, NH = New Hampshire, NY = New York, Vt = Vermont. The stippled band marks the proposed boundary between the Monteregians and WMMS.

The Monteregian-WMMS alkaline provinces exhibit a marked geographic compositional polarity (see Hodgson, 1969; Philpotts, 1974; and Bédard *et al.*, 1987 and 1988 for references and more details). In the west (Fig. 1), around the Oka complex and towards the south and southwest in Pennsylvania and New York State, there are strongly silica-undersaturated carbonatites, kimberlites, ijolites, jacupirangites and mica peridotites, with dykes of alnöite, monchiquite, camptonite and aillikite (carbonate-rich ultramafic lamprophyre; Rock, 1986). The Ile Bizard intrusion has been referred to in the literature as a kimberlite, a central complex kimberlite, an alnöite and an aillikite. Pending resolution of the nomenclatural problems, I will refer to it as an aillikite, following Eby (1985). Towards the east the most common rock types are alkali pyroxenite, yamaskite (kaersutitite or hornblendite), alkali gabbro to essexite, feldspathoidal diorite, monzonite and syenite, and foyaite. Orthopyroxene and quartz-bearing gabbros and syenites are subordinate and are linked to crustal contamination. Their abundance increases towards the east. Basanite-monchiquite (BM) and camptonite dykes occur throughout the Monteregian Province and its southward extensions. In contrast, intermediate and leucocratic dykes are concentrated around intrusive complexes and mirror the differentiation trend of the plutonic rocks (Hodgson, 1969).

The WMMS is dominated by sub-alkaline, metaluminous quartz syenite and granite (A-type), with subordinate nepheline syenite, riebeckite granite, granodiorite, quartz monzonite, diorite, gabbro and norite (see Chapman and Williams, 1935; McHone and Butler, 1984; and Bédard *et al.*, 1987 for details and additional references). Tholeiitic to alkali basalt lavas and dykes are potential parental magmas (McHone, 1978; McHone and Trygstad, 1982; Hermes *et al.*, 1984; Bédard *et al.*, 1987). Here too the intermediate and leucocratic dykes and lavas mirror the differentiation trend of the plutonic rocks. Calc-alkaline lamprophyres are concentrated around the intrusive complexes.

Monteregian dykes. Most Monteregian mafic dykes can be divided into two groups: Basanite-Monchiquites (BM), and Alkali Basalt-Camptonites (ABC). The division is set at a ratio of groundmass feldspar/(feldspar + feldspathoids + carbonate + zeolites + glass) of $\frac{1}{2}$, following the precedent set by Hodgson (1969). This definition does not correspond to that proposed by Rock (1977), where monchiquites entirely lack feldspar. By Rock's definition, many of the more feldspar-rich 'monchiquites' of this study would be analcime camptonites. Some of the BM dykes lack amphibole phenocrysts and are basanites, as was pointed out by Eby (1985). Since no attempt was made to determine the groundmass mode, it is not possible to determine how many are sannaites (alkali feldspar-rich lamprophyres). For simplicity and continuity's sake, Hodgson's (1969) original division is retained and the feldspar-poor mafic dykes are referred to as BM-dykes or simply monchiquites.

Basanite-monchiquite dykes contain euhedral phenocrysts of clinopyroxene (titan-augite or salite) that may enclose green, Na-Fe-rich and Mg-Cr-poor clinopyroxene cores (Bédard *et al.*, 1988). Many dykes carry megacrysts of hastingsitic to pargasitic amphibole that are enveloped by kaersutitic rims (Figs. 2*a*, *b*). Hastingsitic to pargasitic amphiboles may also be enveloped by salitic phenocrysts, and some are partially to completely decomposed (Fig. 2*c*). The more magnesian dykes contain olivine phenocrysts ($Fo_{82\text{-}90.5}$), and rare

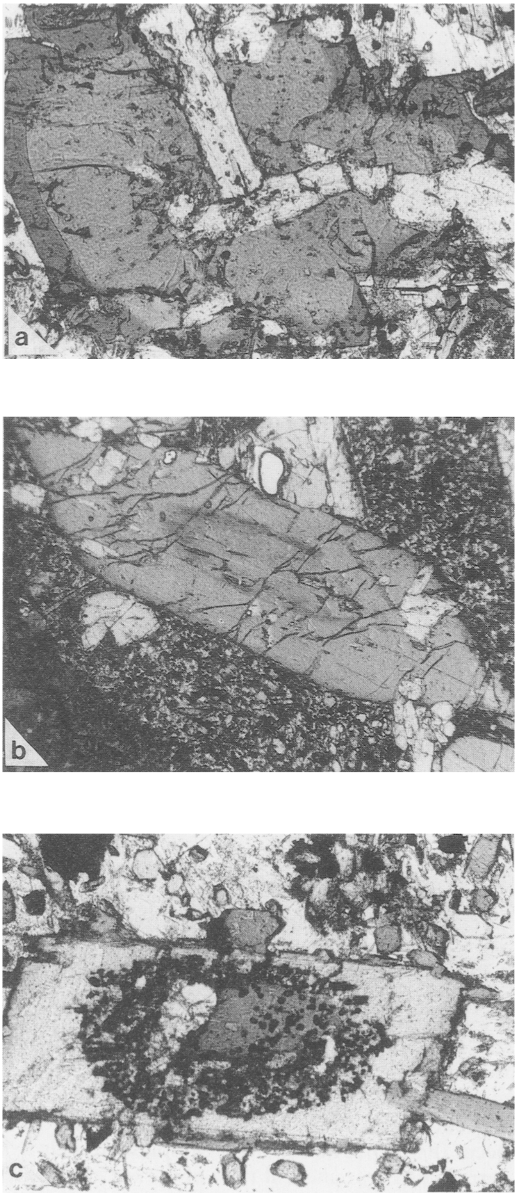


FIG. 2. (*a*) Amphibole megacryst in Monteregian monchiquite JP8786. Note the sharp interface between the megacryst core and the phenocryst rim, and the euhedral clinopyroxene inclusions. Grain is 0.5 mm long. Plane light. (*b*) Continuously-zoned amphibole xenocryst-megacryst-phenocryst from monchiquite JP8797. The turbid rim of the xenocryst is truncated by the enveloping megacryst-phenocryst. Grain is 1.2 mm long. Plane light. The microprobe traverse of Fig. 3 was done on this grain. (*c*) Partly decomposed amphibole xenocryst(?) enveloped by a salitic pyroxene phenocryst in monchiquite JP8792. Grain is 0.5 mm long; plane light.

examples contain anorthoclase megacrysts. The panidiomorphic groundmass is composed of euhedral clinopyroxene, kaersutite, Mg-Al-rich titano-magnetite and apatite, with interstitial zeolites and carbonate, and minor plagioclase, alkali feldspar, mica and glass. Carbonate-rich amygdales are common and grade into feldspar-rich ocelli (Philpotts and Hodgson, 1968). Deuteric alteration products include chlorite, calcite, sphene and actinolite.

As the proportion of feldspar/(feldspar + glass + zeolites + carbonate + feldspathoid) exceeds $\frac{1}{2}$, the BM dykes grade into camptonites. Camptonites typically have higher proportions of amphibole to pyroxene, but amphibole megacrysts have not been observed. In general, camptonites are less porphyritic than the BM dykes. They may contain Al-Mg-rich titano-magnetite phenocrysts. A few camptonites contain plagioclase phenocrysts, marking a gradation into alkali basalts.

The ultramafic lamprophyres generally do not carry amphibole and so will not be described. The Ile Bizard intrusion does, however, contain mantle-derived amphibole-bearing peridotite xenoliths (Marchand, 1970).

Plagioclase is the dominant phenocryst species in Monteregian intermediate dykes. It commonly zones out to perthitic rims. Zoned phenocrysts of Ti-augite to aegirine augite may be surrounded and partly replaced by amphibole coronas. Kaersutitic, to pargasitic, to hastingsitic amphibole also occurs as discrete, zoned phenocrysts. Ilmenite and Ti-magnetite phenocrysts commonly have sphene or biotite reaction rims. Apatite, sphene and biotite phenocrysts also occur. The groundmass is rich in feldspar and commonly contains feldspathoids.

Leucocratic dykes are variable. Strongly silica undersaturated tinguaites may contain phenocrysts of sanidine, sodalite, nepheline, aegirine, kaersutite to hastingsite and Ca-Ti-eudyalite. Trachytes and nepheline microsyenites carry rare phenocrysts of plagioclase, alkali feldspar, nepheline, sodalite, aegirine-augite, hastingsitic amphibole and oxides. Bostonites have rare alkali feldspar (+/− plagioclase, aegirine-augite, hastingsite or biotite) phenocrysts in a trachytic alkali feldspar to anorthoclase +/− quartz groundmass. Sölvsbergites (alkali granites) are principally composed of granular anorthoclase, acmite and riebeckite.

WMMS dykes. WMMS mafic dykes and lavas are dominantly Coombs-trend alkali basalts that carry phenocrysts of plagioclase and Ti-Al-salite or augite (+/− olivine and spinel). They grade into a doleritic, sometimes orthopyroxene-bearing facies. Camptonites and monchiquites are rare (McHone, 1978; McHone and Trygstad, 1982; Hermes *et al.*, 1984). Some carry mantle-derived, sodian hastingsite-bearing xenoliths (Leavy and Hermes, 1979).

In the WMMS, calc-alkaline lamprophyres (spessartites and kersantites) are more common than monchiquites and camptonites. In the more magnesian examples, green amphibole and mica occur as reaction rims on olivine and pyroxene(?) phenocrysts (Bédard *et al.*, 1987). The more evolved examples typically contain phenocrysts of green hornblende and plagioclase (+/− olivine, oxides and apatite). Quartz is a common groundmass phase.

The intermediate dykes and lavas include basaltic andesite, hawaiite and mugearite. Plagioclase and augitic pyroxene (+/− olivine, apatite and oxides) are the dominant phenocryst types. Kaersutite occurs in association with Mg-Al-rich titanomagnetite and carbonate in some hawaiites from Mégantic (Bédard *et al.*, 1987). Other hawaiites from Mégantic are essentially carbonate-free and contain abundant green edenite to hastingsite and biotite as coronas upon olivine, pyroxene and oxides, and in the groundmass with quartz and magnetite. The carbonate-free hawaiites are texturally transitional into calc-alkaline lamprophyres.

Leucocratic dykes include hypersolvus syenites and quartz porphyries, aplite, biotite granite, bostonite, sölvsbergite and comendite. Alkali feldspar, sodic plagioclase, quartz, hedenbergitic pyroxene, calcic or sodic-calcic amphiboles, apatite, magnetite and ilmenite are common phenocryst species. Cogenetic plutonic rocks may also carry cumulus fayalite, low-Ca pyroxene, zircon, allanite and monazite.

Textural characteristics of amphiboles in dykes

Amphibole phenocrysts and groundmass prisms from Monteregian and WMMS mafic dykes are typically composed of euhedral, dark brown (in thin section), pleochroic kaersutite. In intermediate and silica undersaturated leucocratic dykes the brown cores commonly have a cloudy appearance due to the presence of innumerable fine-grained inclusions of Ti-oxide (schiller or sagenetic texture), and many have greenish rims. Amphibole also occurs as a patchy replacement of pyroxene, as coronas on pyroxene or as anhedral to euhedral groundmass grains. In silica-oversaturated intermediate and felsic rocks, yellow-green, to grey-green to blue-green calcic and sodic-calcic amphiboles occur as euhedral to anhedral plates and/or as reaction rims around anhydrous silicates. Grunerite reaction rims around fayalite occur in some WMMS syenites (Creasy, 1974). Blue sodic amphiboles occur as veinlets or prismatic overgrowths on earlier amphiboles in peralkaline granitic rocks of the WMMS (Billings, 1928; Chapman, 1942; Creasy, 1974; Bédard *et al.*, 1987) and in the Shefford Nordmarkites (Frisch, 1970).

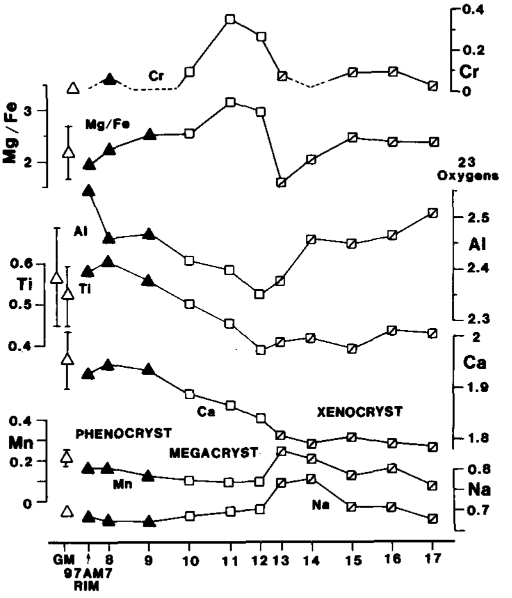


FIG. 3. Microprobe traverse through the amphibole xenocryst–megacryst–phenocryst of Fig. 2*b*, from monchiquite JP8797. The horizontal scale is arbitrary. The numbers represent analyses from 97AM17 (core, see Table 1) to 97AM7 (rim). GM = range of groundmass amphiboles in this sample. The microprobe traverse is represented on Figs. 4 and 5 by an arrowed line.

Amphibole megacrysts are composed of pale to medium brown (in thin section), pleochroic, titanian pargasite or magnesio hastingsite. Most megacrysts are ovoid, with euhedral rims of darker kaersutite (Fig. 2*a*); but euhedral examples also occur (Fig. 2*b*). The megacryst amphibole is always surrounded by euhedral rims of darker kaersutite. Megacryst and rim amphibole are generally separated by a sharp interface (Fig. 2*a*), but gradational zonations also occur (e.g. in JP8797, Figs. 2*b* and 3). The megacrysts may enclose carbonate, zeolite, sulfide or oxide inclusions or infiltrations(?). Euhedral clinopyroxene (Fig. 2*a*), apatite(?) and olivine inclusions also occur. In JP8797 the amphibole megacrysts envelop subeuhedral, zoned, amphibole 'xenocrysts' and truncate the zoning patterns of these xenocrysts (Fig. 2*b*).

Amphibole xenocrysts occur either as anhedral intergrowths with pyroxene that are surrounded by euhedral kaersutite rims (e.g. in dyke JP8782), as inclusions in amphibole megacrysts (Fig. 2*b*), or as partially to completely decomposed grains of amphibole enclosed within salite phenocrysts (Fig. 2*c*).

Amphibole chemistry

Amphibole phenocrysts from Monteregian dykes are mostly kaersutites (commonly sub-silicic and potassian) that show normal zoning towards Fe-Mn-rich and Cr-Mg-poor rims (Fig. 3; Table 1). Amphibole phenocrysts from camptonites tend to be poorer in Al than are those from BM dykes (Fig. 4) and the rim and groundmass amphiboles in the camptonites generally show Ti-enrichment. Conversely, in the BM dykes, the outermost rims of phenocrysts and groundmass amphiboles are commonly enriched in Al and depleted in Ti (Figs. 3 and 4; Table 1). Kaersutites from the Maine (McHone and Trygstad, 1982) and Vermont (McHone and

TABLE 1 REPRESENTATIVE AMPHIBOLES

	Monchiquite Dyke					Camptonite Dykes			
	Xenocryst-Megacryst-Phenocryst					Phenocryst			
	Core.....Rim		Core....Rim		Gmass	Core.....Rim		Gmass	Deuter
	797AM17	797AM13	797AM12	797AM7	797AM3	771AM7	771AM4	771AM1	778AM7
SiO_2	40.32	40.37	41.36	39.10	39.61	39.60	38.23	39.51	54.54
TiO_2	3.92	3.72	3.56	5.16	4.51	4.11	5.38	5.41	.24
Al_2O_3	14.68	13.77	13.66	14.50	14.24	13.76	13.23	12.65	2.45
FeO	9.74	12.60	8.36	10.26	10.52	10.78	11.03	9.67	9.12
MnO	.06	.19	.07	.13	.14	.15	.24	.15	.25
MgO	13.96	12.46	14.65	12.32	12.97	13.35	12.33	14.27	17.73
CaO	11.46	11.53	11.76	12.05	11.99	11.53	12.06	12.12	12.46
Na_2O	2.42	2.71	2.50	2.37	2.50	2.55	2.51	2.52	.34
K_2O	1.29	1.14	1.40	1.26	1.24	1.32	1.01	1.01	.06
Cr_2O_3	.03	.06	.22	<d.l.	.05	<d.l.	<d.l.	<d.l.	.06
NiO	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.
V_2O_3	.20	.23	.20	.26	.25	.22	.24	.27	<d.l.
Total	98.08	98.78	97.74	97.44	98.02	97.36	96.26	97.57	97.25
Si	5.8392	5.9009	6.0165	5.8200	5.8332	5.8410	5.7882	5.8228	7.7171
Al^{iv}	2.1618	2.0991	1.9835	2.1800	2.1668	2.1590	2.2118	2.1772	0.2829
Al^{vi}	0.3444	0.2736	0.3580	0.3639	0.3040	0.2337	0.1492	0.0206	0.1258
Ti	0.4270	0.4089	0.3887	0.5769	0.4989	0.4551	0.6117	0.5997	0.0255
Cr	0.0031	0.0079	0.0269	--	0.0063	--	--	--	0.0001
V	0.0251	0.0300	0.0237	0.0317	0.0301	0.0269	0.0302	0.0348	--
Fe^{3+}	0.4591	0.3760	0.1681	--	0.0993	0.3672	--	0.1864	0.2356
Fe^{2+}	0.7197	1.1647	0.8494	1.2775	1.1963	0.9619	1.3966	1.0051	0.8436
Mn	0.0078	0.0237	0.0079	0.0158	0.0174	0.0190	0.0302	0.0190	0.0300
Mg	3.0138	2.7152	3.1773	2.7341	2.8477	2.9363	2.7821	3.1345	3.7387
Ca	1.7784	1.8060	1.8328	1.9210	1.8911	1.8221	1.9559	1.9130	1.8892
Na^{M4}	0.2216	0.1940	0.1672	0.0791	0.1089	0.1778	0.0441	0.0869	0.0933
Na^{A}	0.4581	0.5748	0.5375	0.7622	0.6054	0.5507	0.6931	0.6330	0.0000
K	0.2386	0.2131	0.2591	0.2393	0.2328	0.2481	0.1954	0.1899	0.0001
Tot	15.6904	15.7879	15.7966	15.8433	15.8382	15.7988	15.8885	15.8229	14.9564
Mgn	0.7188	0.6380	0.7574	0.6815	0.6873	0.6884	0.6658	0.7246	0.7760
Ca+ Al^{iv}	3.9402	3.9051	3.8163	4.1010	4.0579	3.9811	4.1677	4.0902	2.1721
Si+ Na+K	6.7575	6.8828	6.9803	6.7424	6.7803	6.8176	6.7208	6.7326	7.8105

	Monteregians				WMMS (Megantic)				
	Interm. Dyke	Leuc. Dyke	Tinguaite Core.....Rim		Alkali Basalt Dyke	Hawai-ite Dyke	Quartz Syenite Pluton	Alkali-Granite Dyke	
	824AM1	840AM4	863AM1	863AM3	200AM1	65AM1	35AM4	207AM9	207AM12
SiO_2	39.92	38.08	38.37	40.70	40.13	43.45	42.59	48.82	51.81
TiO_2	3.85	3.15	5.53	2.30	4.61	1.46	.81	.67	.14
Al_2O_3	11.75	12.46	13.84	9.63	12.30	8.60	5.77	2.06	.57
FeO	14.73	18.33	13.61	21.22	13.91	29.33	33.94	32.10	35.55
MnO	.47	.54	.37	1.79	.20	.27	1.06	1.88	.07
MgO	11.65	8.21	11.17	7.21	12.38	9.99	.72	2.32	1.41
CaO	11.48	10.94	11.42	9.27	11.14	11.56	9.68	6.17	.11
Na_2O	2.81	2.79	3.41	3.96	2.84	2.43	1.49	3.96	7.67
K_2O	1.02	1.88	1.03	1.76	.74	.64	.93	.95	.21
Cr_2O_3	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	<d.l.	n.d.	n.d.
NiO	<d.l.	<d.l.	n.d.	n.d.	<d.l.	<d.l.	<d.l.	n.d.	n.d.
V_2O_3	.18	.16	n.d.	n.d.	.26	.20	.05	n.d.	n.d.
ZrO_2	n.d.	n.d.	.09	.48	n.d.	n.d.	n.d.	n.d.	n.d.
F	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	.94	.87
Cl	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	<d.l.	<d.l.
Total	97.86	96.55	98.61	98.29	98.51	97.93	98.04	99.46	97.44
Si	5.9722	5.9429	5.7182	6.2983	5.8908	6.5580	6.8912	7.8130	7.9776
Zr	--	--	0.0064	0.0337	--	--	--	--	--
Ti^{iv}	--	--	--	--	--	--	0.0089	--	--
Al^{iv}	2.0278	2.0571	2.2754	1.6680	2.1092	1.4419	1.0999	0.1870	0.0224
Al^{vi}	0.0438	0.2353	0.1556	0.0886	0.1980	0.0877	--	0.2017	0.0808
Ti^{vi}	0.4327	0.3692	0.6194	0.2684	0.5087	0.1665	0.0893	0.0809	0.0167
Cr	--	--	--	--	--	--	--	--	--
V	0.0222	0.0225	--	--	0.0298	0.0236	--	--	--
Fe^{3+}	0.4063	--	0.0543	0.4350	0.5912	0.4233	0.9051	--	1.6816
Fe^{2+}	1.4370	2.3919	1.6421	2.3100	1.1170	2.0171	3.6865	4.2968	2.8961
Mn	0.0602	0.0706	0.0464	0.2346	0.0251	0.0345	0.1457	0.2555	0.0015
Mg	2.5978	1.9103	2.4822	1.6634	2.7083	2.2473	0.1734	0.5538	0.3233
Ca	1.8402	1.8301	1.8228	1.5364	1.7522	1.8704	1.6783	1.0583	0.0182
Na^{M4}	0.1598	0.1701	0.1772	0.4636	0.2479	0.1296	0.3217	0.5530	1.9818
Na^{A}	0.6549	0.6743	0.8087	0.7241	0.5607	0.5818	0.1462	0.6751	0.1295
K	0.1950	0.3740	0.1952	0.3471	0.1382	0.1225	0.1918	0.1936	0.0789
Tot	15.8499	15.0483	16.0039	16.0712	15.6989	15.7043	15.3380	15.8687	15.2084
Mgn	0.5849	0.4440	0.5940	0.3773	0.6132	0.4794	0.0364	0.1142	0.0650
Ca+ Al^{iv}	3.8680	3.8872	4.0982	3.2044	3.8614	3.3123	2.7782	1.2453	0.0406
Si+Na +K	6.9819	7.1613	6.8993	7.8331	6.8376	7.3919	7.5509	9.2347	10.1678

Table 1 Representative amphibole analyses. Analyses were done at McGill University on a wavelength-dispersive Cambax/Micro electron microprobe using Cameca's MBXCOR/Micro software package. Counting times were 20 or 25 seconds for most elements except F and Cl which were counted for 100 seconds. An accelerating voltage of 15Kv and a beam intensity of 7nA was employed. n.d. = not determined, <d.l.- below detection limits, -- = not applicable. Mgn- Mg/(Mg+Fe). The structures were calculated after the method of Laird and Albee (1981). Total cations less (Ca+Na+K) were normalized to 13, Al^{iv} = 8-Si, Al^{vi}=Al_{total}-Al^{iv}, Fe^{3+}=46-total cation charge using all ferrous iron, Fe^{2+}=Fe_{total}-Fe^{3+}, Na^{M4}=7-(M1+M2+M3 cations), Na^{A}=Na-Na^{M4}.

797AM17 = titanian hastingsite xenocryst core and 797AM13= titanian hastingsite xenocryst rim, 797AM12 = potassian titanian hastingsite megacryst core and 797AM7 - kaersutite phenocryst rim, 797AM3 = kaersutite groundmass prism (monchiquite JP8797). 771AM7 - titanian hastingsite core and 770AM4 = kaersutite phenocryst rim, 770AM1 = groundmass kaersutite (camptonite JP8770). 778AM7 = deuteric actinolite (camptonite JP8778). 824AM1 = titanian hastingsite phenocryst rim (JP8824), 840AM4 = potassian titanian ferro-pargasite phenocryst core (JP8840). 63AM1 = subsilicic kaersutite core and 63AM3 = sodian potassian titanian ferro-hastingsite rim of phenocryst in tinguaite (JP8863). Megantic: 200AM1 = brown euhedral groundmass kaersutite in alkali basalt (JB83-200), 65AM1 - ferro-edenite core of anhedral green plate in hawaiite (DD6065B), 35AM4 - green corona of ferrian ferro-hornblende on pyroxene in syenite (DD5035), 207AM9 - green grain of ferro-winchite with 207AM12 - blue riebeckite overgrowth in alkali granite dyke (JB83-207).

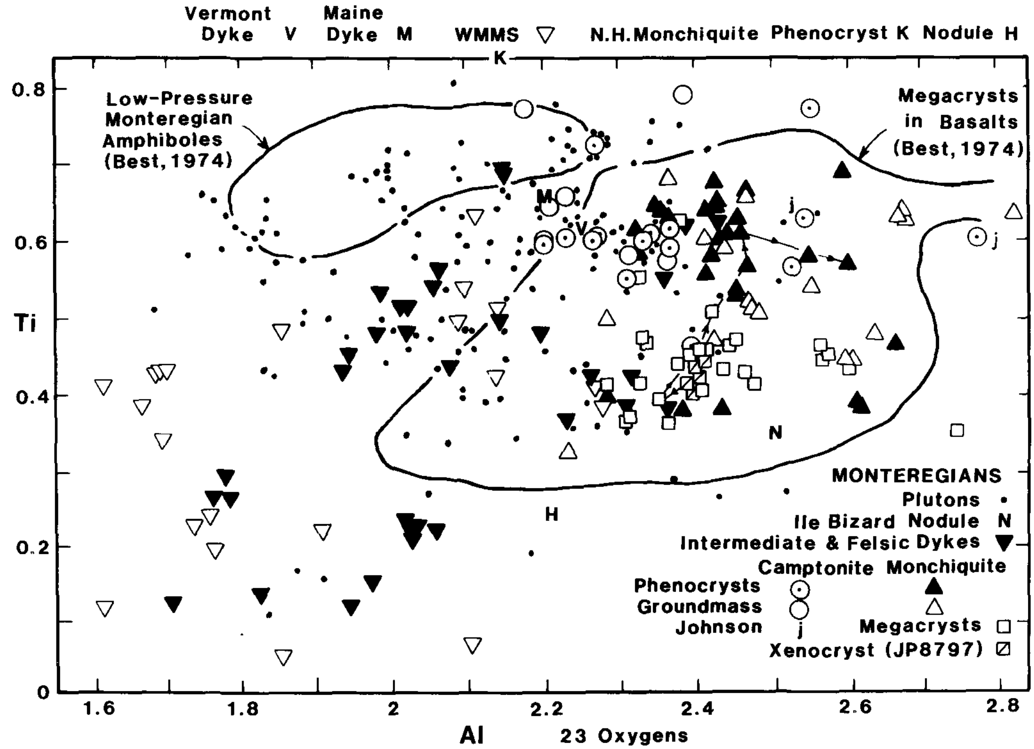


FIG. 4. Al *vs.* Ti diagram for Monteregian and WMMS amphiboles. Amphibole structures calculated by the method of Laird and Albee (1981) or Rock and Leake (1984) based on 23 oxygens. Data from this study; Pajari (1967), Marchand (1970), Frisch (1970), Gunn (1972), Gélinas (1972), Creasy (1974), Woussen (1974), Valiquette and Pouliot (1977), Czamanske *et al.* (1977), Loiselle (1978), Leavy and Hermes (1979), McHone and Corneille (1980), McHone and Trygstad (1982), Greenwood (1983), Currie *et al.* (1986), K. Currie (1986, pers. comm.), and Gold *et al.* (1986). The arrowed line is the core to rim traverse of the composite amphibole xenocryst-megacryst-phenocryst of Figs. 2B and 3. Only the xenocryst from JP8797 is marked differently from the megacrysts.

Corneille, 1980) dykes are compositionally similar to Monteregian kaersutites (M and V on Figs. 4 and 5). An amphibole phenocryst from the monchiquite studied by Leavy and Hermes (1979) is richer in Ti and has a lower Si + Na + K content (K on Figs. 4 and 5); and appears to belong to the WMMS amphibole suite.

The rims of phenocrysts from the magnesian dykes, and the phenocrysts and groundmass amphiboles of the intermediate Monteregian dykes are composed of kaersutite (commonly sub-silicic and ferroan), pargasite (commonly ferroan and titanian) or magnesian ferro-hastingsite (commonly titanian). They are less aluminous and richer in Si + Na + K (Fig. 5) than amphiboles from the BM dykes and camptonites. Pale, green, deuteric amphiboles are silica-rich actinolites (Table 1). The peralkaline tinguaite dykes have phenocrysts of kaersutite that zone out to titanian sodian potassian manganoan ferro-hastingsite rims.

There is a fairly good bulk compositional control on amphibole Mgn [cationic Mgn = Mg/(Mg + Fe)] (Fig. 6). The relationship can be described as:

$$\text{Mgn amphibole} = 0.253125 \times \text{Mgn whole rock} + 0.5425$$

for rocks with Mgn > 0.4 and as:

$$\text{Mgn amphibole} = 1.6 \times \text{Mgn whole rock}$$

for amphiboles coexisting with rocks of Mgn < 0.4.

In the WMMS (and rare silica-rich Monteregian rocks), compositional evolution is from kaersutite in the basaltic dykes, to kaersutite, edenite, hastingsite, and to edenitic and hastingsitic hornblendes in the hawaiite dykes and gabbros. These variations are characterized by decreasing Mg, Ti, Na and Al, increasing Si and Fe, and almost constant Ca (Table 1). Variations in amphibole composition from the intermediate dykes and gabbros to the ferro-edenites and hastingsites of the subalkaline syenites and granites, to the ferrowinchite, ferro-barroisite, ferro-richterite, arfvedsonite and riebeckite of the peralkaline syenites and peralkaline granites (comendites) are also characterized by decreasing Mg, Ti and Al, and by increasing Si and Fe. However, the abundance of Ca decreases, while that of Na increases (Table 1). Amphiboles from the WMMS also exhibit a strong bulk compositional control on their Mgn (Fig. 6), but there is little or no overlap with Monteregian amphiboles on the $Ca+Al^{iv}$ *vs.* $Si+Na+K$ diagram (Fig. 5).

Amphibole megacrysts from Monteregian BM dykes are composed of magnesio hastingsite, pargasite (typically potassian and titanian), or rarely kaersutite. In one case (JP8797) they zone continuously out to a kaersutitic rim identical to phenocryst and groundmass amphibole (Figs. 2*b*, 4 and 5). The Mgns of the megacrysts also reflect a bulk compositional control (Fig. 6). The amphibole megacrysts generally have higher contents of Mg and Al, but lower contents of Ti and $Ca+Al^{iv}$ than the phenocryst and groundmass kaersutites, and plot in a compositionally distinct field (Figs. 4 and 5). Amphiboles from the Brossard intrusion (Gunn, 1972) have compositions typical of megacryst amphiboles (B on Fig. 5C).

The amphibole xenocryst from monchiquite JP8797 (Fig. 2*a* and 3) is normally zoned from a titanian magnesio hastingsite or pargasite core to a turbid, Fe-rich, titanian magnesian hastingsite rim. This xenocryst has Ti, Al, $Ca+Al^{iv}$ and $Si+Na+K$ contents typical of megacryst amphiboles, but is richer in Fe and Mn (Fig. 3; Table 1). Conversely, a xenocryst from dyke JP8782 is more magnesian than coexisting phenocrysts, but still plots with the megacrysts. Amphibole from a mantle-derived peridotite xenolith in the Ile Bizard aillikite (Marchand, 1970) is still poorer in Ti and richer in Al than are the megacryst amphiboles from the BM dykes (N on Figs. 4 and 5). An amphibole from a mantle xenolith in the New England monchiquites studied by Leavy and Hermes, 1979 (H on Figs. 4 and 5), is also depleted in Ti, but is less aluminous than the mantle amphibole from Ile Bizard.

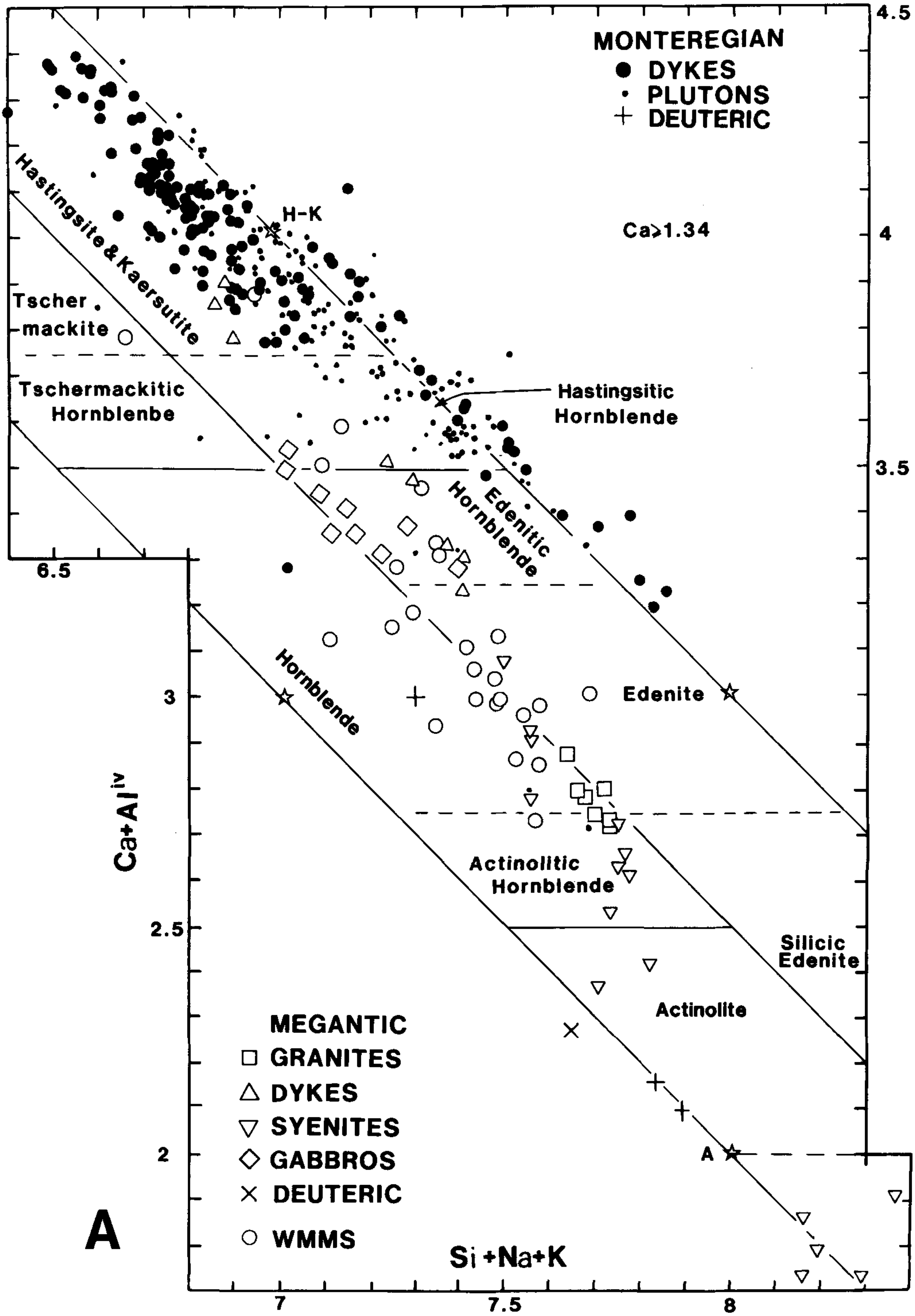
Discussion

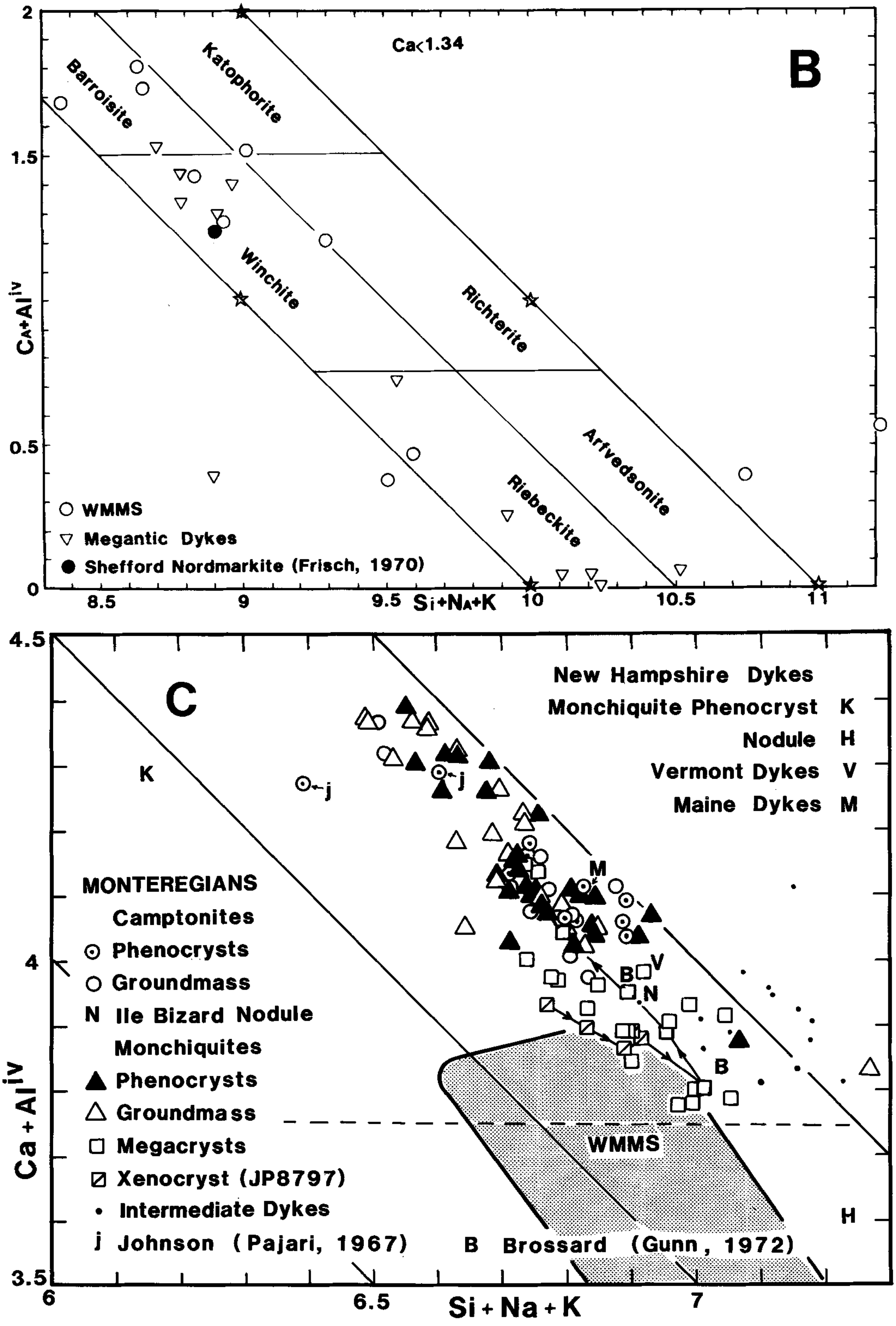
Phenocryst and groundmass amphiboles. Amphibole phenocrysts from the Monteregian series range from kaersutites to hastingsites and display a limited range of Ti, Al, $Ca+Al^{iv}$ and $Si+Na+K$ (Figs. 4 and 5). The dominant substitution mechanisms in amphiboles from silica-undersaturated alkaline rocks are thought to be $Ti+O \rightleftharpoons Fe^{3+}+OH^{-}$, $\square Ti \rightleftharpoons Na^{A}Al^{iv}$, $Mg \rightleftharpoons Fe^{2+}$, $Ca(Mg,Fe^{2+}) \rightleftharpoons Na^{M}Fe^{3+vi}$ and $CaAl^{iv} \rightleftharpoons Na^{M}Si$ (Giret *et al.*, 1980; Stephenson and Upton, 1982). The lack of significant Ti *vs.* Al covariation in Monteregian amphibole phenocrysts from mafic dykes (Fig. 4) suggest that the temperature-dependent Ti-substitution mechanism is $Ti+O \rightleftharpoons Fe^{3+}+OH^{-}$. Monteregian magmas probably had low silica activities. This would inhibit substitutions such as $CaAl^{iv} \rightleftharpoons Na^{M}Si$ and should result in a limited range of $CaAl^{iv}$ *vs.* $Si+Na+K$ variation, as is observed. Monteregian amphiboles show limited Na-enrichment and little Ca-depletion. This implies that substitutions like $Ca(Mg,Fe^{2+}) \rightleftharpoons Na^{M}Fe^{3+vi}$ did not occur to any great extent. Groundmass and phenocryst rim amphiboles from BM dykes are generally enriched in Al and depleted in Ti, whilst the opposite pattern prevails in the camptonites (Fig. 4). This may reflect early saturation and fractionation of plagioclase from the camptonite magmas, leading to Al-depletion and Ti-enrichment in the co-precipitating amphibole.

Amphiboles from the WMMS, and from the Si-rich Shefford Nordmarkite, exhibit very different trends from those of the silica-undersaturated rocks of the Monteregians (Figs. 4 and 5). The dominant substitution mechanisms in amphiboles from mafic to intermediate silica-saturated and oversaturated alkaline rocks are thought to be $(Na,K)^{A}Al^{iv} \rightleftharpoons \square^{A}Si$, $Ca(Mg,Fe^{2+}) \rightleftharpoons Na^{M}Fe^{3+vi}$ and $Mg = Fe^{2+}$ (Neumann, 1976; Giret *et al.*, 1980; Stephenson and Upton, 1982). The reduced nature of WMMS melts (Bédard *et al.*, 1988) would probably inhibit substitutions such as $Ca(Mg,Fe^{2+}) \rightleftharpoons (Na,K)^{M}Fe^{3+vi}$. Conversely, increases of silica activity in the melt should favour the $Na^{A}Al^{iv} \rightleftharpoons \square Si$ reaction and lead to reductions in Al^{iv} with little change in Ca or $Si+Na$. This is the pattern of variation observed in WMMS amphiboles and is consistent with the observations of Bédard *et al.* (1987) that WMMS mafic and intermediate melts show progressive Si-enrichment.

The dominant substitution mechanisms in felsic, silica-saturated to oversaturated alkaline rocks are thought to be $CaAl^{iv} \rightleftharpoons Na^{M}Si$ or $(Na,K)^{A}Al^{iv} \rightleftharpoons$

FIG. 5. A, $Ca+Al^{iv}$ *vs.* $Si+Na+K$ (cationic) diagram of Giret *et al.* (1980) for calcic amphiboles from the Monteregians and WMMS. Data from this study and *as per* caption of Fig. 4. Note the almost perfect discrimination of WMMS and Monteregian amphiboles. The stars mark end-member compositions; A = Actinolite, H = Hastingsite, K = Kaersutite. B, $Ca+Al^{iv}$ *vs.* $Si+Na+K$ (cationic) diagram for sodic-calcic and sodic amphiboles. C, $Ca+Al^{iv}$ *vs.* $Si+Na+K$ (cationic) diagram for calcic amphiboles from mafic Monteregian dykes. The arrowed line is the microprobe traverse of Fig. 3.


MONTEREGIAN
● DYKES
• PLUTONS
+ DEUTERIC
Ca>1.34
H-K
Hastingsite & Kaersutite
Tscher . -mackite
Tschermackitic Hornblenbe
Hastingsitic Hornblende
Edenitic Hornblende
Edenite
Hornblende
Actinolitic Hornblende
Silicic Edenite
Actinolite
A
MEGANTIC
□ GRANITES
△ DYKES
▽ SYENITES
◇ GABBROS
× DEUTERIC
○ WMMS
A
Ca+Al^{iv}
Si+Na+K
6.5
7
7.5
8
2
2.5
3
3.5
4
4.5


B
Ca<1.34
Barroisite
Katophorite
Winchite
Richterite
Arfvedsonite
Riebeckite
C_A+Al^{iv}
Si+N_A+K
2
1.5
0.5
0
8.5
9
9.5
10
10.5
11
○ WMMS
▽ Megantic Dykes
● Shefford Nordmarkite (Frisch, 1970)
C
K
j
M
B
N
V
H
WMMS
New Hampshire Dykes
Monchiquite Phenocryst K
Nodule H
Vermont Dykes V
Maine Dykes M
MONTEREGIANS
Camptonites
⊙ Phenocrysts
○ Groundmass
N Ile Bizard Nodule
Monchiquites
▲ Phenocrysts
△ Groundmass
□ Megacrysts
▨ Xenocryst (JP8797)
• Intermediate Dykes
j Johnson (Pajari, 1967)
B Brossard (Gunn, 1972)
Ca + Al^{iv}
Si + Na + K
4.5
4
3.5
6
6.5
7

$\square^4$Si under reducing conditions, and either $Fe^{2+}Si \rightleftharpoons Fe^{3+vi}Al^{vi}$, $CaFe^{2+} \rightleftharpoons Na^{M}(Fe^{3+},Al)^{iv}$, or $(Na,K)^4Fe^{2+} \rightleftharpoons \square^4Fe^{3+vi}$ if conditions are oxidizing (Giret *et al.*, 1980; Stephenson and Upton, 1982; Strong and Taylor, 1984). Sodic-calcic and sodic amphiboles from the WMMS show Ca, Ti and Al-depletion, and Si, Fe and Na-enrichment. These compositional variations are consistent with progressive increases in the melt's silica activity, sodium activity, and f_{O_2}. Formation of riebeckite and arfvedsonite are considered to represent late-stage precipitation and/or replacement due to the action of a low-temperature (500–600 °C) fluid or vapour phase (Billings, 1928; Ernst, 1962; Ferguson, 1978; Strong and Taylor, 1984).

There is little or no overlap between WMMS and Monteregian amphiboles (Figs. 4 and 5). The distribution of amphiboles from Mégantic is identical to that of the WMMS and clearly differs from that of the Monteregians, supporting the proposal of Bédard *et al.* (1983 and 1987) that Mégantic should be considered a member of the WMMS. Very few Monteregian amphiboles plot with the WMMS, and most of those are from the more quartz-rich intrusions, or are late-stage deuteric actinolites. The amphibole variations in the Monteregian–WMMS provinces supports the concept advanced by Giret *et al.* (1980), who proposed that the silica activity of the melt controls the amphibole evolution trends. However, many feldspathoidal syenites contain sodic-calcic and sodic amphiboles (Nash and Wilkinson, 1970; Larsen, 1976; Mitchell and Platt, 1978; Grapes *et al.*, 1979; Nielsen, 1979; Jones, 1984), a paragenesis considered typical of silica-oversaturated rocks by Giret *et al.* (1980). Mitchell and Platt (1978) proposed that the

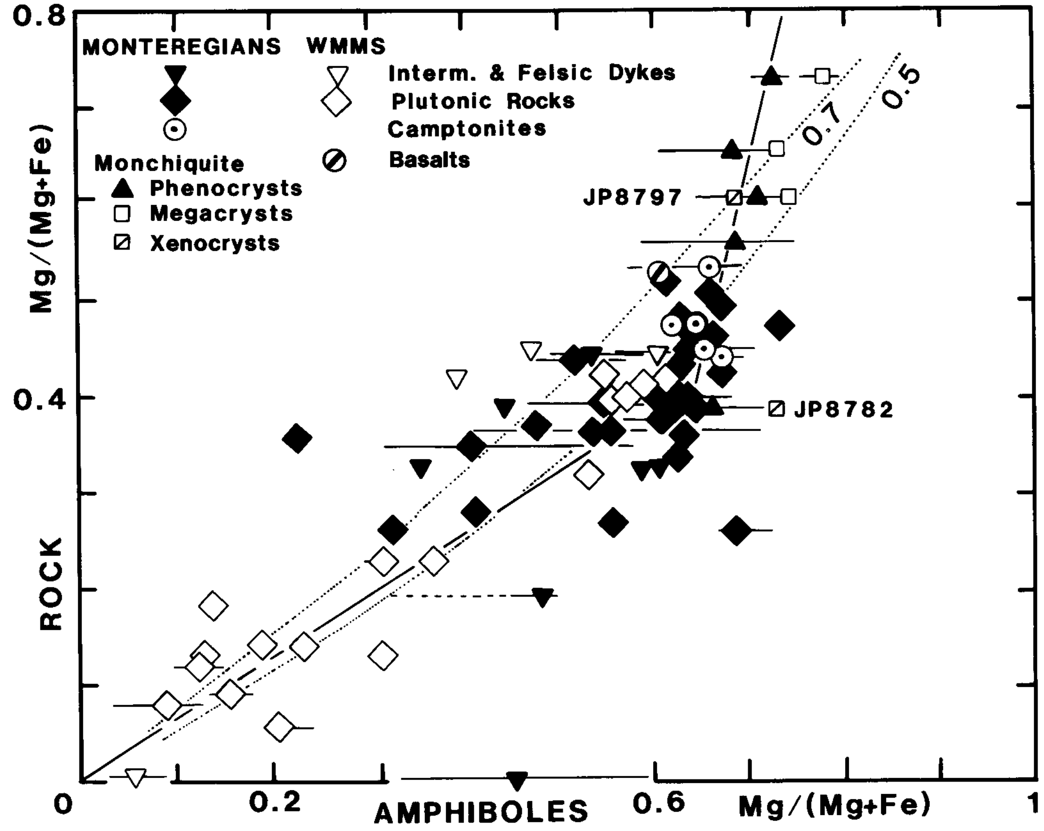


FIG. 6. Amphibole *vs.* whole rock Mg/(Mg + Fe) (cationic) diagram for Monteregian and WMMS amphiboles. Data from this study, Pajari (1967), Czamanske *et al.* (1977), Loiselle (1978), McHone and Trygstad (1982), Greenwood (1983) and K. Currie (1986, pers. comm.). Symbols represent average analyses, horizontal bar marks the observed range. The solid lines are defined as Mgn amphibole = 0.253125 × Mgn whole rock + 0.5425 for rocks with Mgn (Mg/(Mg + Fe)) > 0.4, and Mgn amphibole = 1.6 × Mgn whole rock for rocks with Mgn < 0.4. The dotted lines are for Kds of 0.5 and 0.7. $Kd = [FeO^t_{amph.}/MgO_{amph.}] \times [MgO_{host\ rock}/FeO^t_{host\ rock}]$, FeO^t and MgO are in wt. %, FeO^t = all iron as FeO.

development of sodic-calcic and sodic amphiboles is principally the result of the development of peralkalinity in the magma. Yet peralkaline dykes from the Monteregians do not contain such amphiboles. This problem has yet to be resolved.

The amphiboles reflect the Mgn of the melt (Fig. 6) to a much greater degree than do pyroxene phenocrysts (cf. Bédard *et al.*, 1988). Consequently, amphibole fractionation should be less 'efficient' than pyroxene or olivine fractionation and is more likely to produce a continuous compositional spectrum of residual magma types. The empirical correlation of Fig. 6 corresponds fairly well to Kd values of 0.5 to 0.7, except for the more mafic dykes which have Kds around 1 ($\mathrm{Kd} = [\mathrm{FeO}^{t}_{\mathrm{amph.}}/\mathrm{MgO}_{\mathrm{amph.}}] \times [\mathrm{MgO}_{\mathrm{host}}/\mathrm{FeO}^{t}_{\mathrm{host}}]$, FeO^{t} and MgO in wt. %, FeO^{t} = all iron as FeO). Fractionation of relatively Si-poor phases like amphiboles would tend to increase the melt's Si-content and provides a way of generating silica-oversaturated felsic residua from critically undersaturated basaltic parents (Gunn, 1972; Brooks and Platt, 1975; Cawthorn, 1976; Bédard *et al.*, 1987).

Amphibole megacrysts in Monteregian dykes are generally composed of titanian potassian pargasite or magnesio hastingsite. As in other megacryst suites (Best, 1974; Francis, 1976*a*, *b*; Becker, 1977; Vinx and Jung, 1977), they tend to have higher Mg and Al, and lower Ti than the phenocryst and groundmass amphiboles (Figs. 4 and 5). Amphibole in mantle-derived peridotite xenoliths from the Ile Bizard aillikite has still lower Ti and higher Al-contents. This suggests that there is a pressure control on the $\square \mathrm{Ti} \rightleftharpoons \mathrm{Na}^{4}\mathrm{Al}^{\mathrm{vi}}$ substitution mechanism. Ti-solubility in amphiboles is known to be pressure-sensitive (Cawthorn, 1976), and so the compositional variations among Monteregian amphiboles support a high-pressure, lower crustal or uppermost mantle environment for megacryst formation. The Mgns of amphibole *megacrysts* also reflects that of the host magma (Fig. 6) and is evidence for a cognate relationship. They are, however, more magnesian than the phenocryst rims. In contrast to the megacrysts, amphibole xenocrysts are either too magnesian (JP8782) or marginally too Fe-rich (JP8797) to have formed in equilibrium with their host magmas. Their presence implies contamination or hybridization. Some amphiboles are partly decomposed (Fig. 2*c*), suggesting incorporation into a melt hotter than the one with which they had previously equilibrated.

The pressure at which these megacrysts and xenocrysts crystallized is only partly constrained by the parageneses and available experimental data. Numerous experiments (e.g. Yoder and Tilley, 1962; Holloway and Burnham, 1972; Knutson and Green, 1975; Presnall *et al.*, 1978) have shown that the stability field of plagioclase is reduced by increasing the pressure or water-content. The extreme rarity of plagioclase megacrysts and xenocrysts, together with the abundance of cognate amphibole megacrysts, imply that the magmas were water-rich and that crystallization took place at a relatively high pressure ($P \geqslant$ 5–12 kbar). The upper limit for amphibole stability is thought to be around 25 kbar for temperatures of about 1000 °C (Yoder, 1976), although Holloway and Ford (1975) report stable fluor-hydroxy pargasite at 35 kbar and 1285 °C. The absence of orthopyroxene megacrysts implies that pressures were less than 25 kbar (Fig. 12 in Ellis, 1976). The rarity of mantle-derived xenoliths in the Monteregian suite also suggests a crustal environment for formation of the amphibole megacrysts.

It is common to see a sharp optical and compositional boundary (Fig. 2*b*) between the megacryst core and the phenocryst rim. Whether the rounded outlines reflect partial resorption, partial re-equilibration, or simply non-faceted growth is uncertain. In any case, the sharp discontinuity suggests a sudden transition from one growth environment to another and may reflect rapid ascent or a sudden change in equilibrium caused by volatile exsolution. The occurrence of carbonate and zeolite inclusions in the amphibole megacrysts would then record decompressive volatile exsolution during ascent.

Conclusions

Most Monteregian amphiboles follow the silica-undersaturated evolution trend on the $\mathrm{Ca} + \mathrm{Al}^{\mathrm{iv}}$ *vs.* Si + Na + K diagram, whereas WMMS amphiboles follow the silica-oversaturated trend. The distribution of amphiboles from Mégantic and the WMMS are indistinguishable on these diagrams (Fig. 5), supporting the idea that Mégantic is part of the WMMS. Pargasitic megacrysts from Monteregian basanite–monchiquite dykes have higher Mg and Al and lower Ti-contents than enveloping kaersutitic phenocrysts and rims. This is interpreted to be a pressure effect. Both phenocrysts and megacrysts show a good bulk compositional control on Mgn; this implies that the megacrysts are cognate. The similarity in Mgn between amphibole and melt (whole rock) implies that amphibole fractionation would be less 'efficient' than pyroxene fractionation at decreasing the Mg-content of the melt. The silica-undersaturated nature of amphiboles in the mafic dykes of both suites implies that amphibole fractionation could lead to the development of silica-oversaturated residua from critically silica-undersaturated parental magmas. The occurrence of euhedral amphibole xenocrysts suggests that the host magmas have been contaminated through mixing with a more evolved melt. A similar origin was attributed to the Fe-rich green pyroxene cores to pyroxene phenocrysts (Bédard *et al.*, 1988). The presence of these mineralogical relics implies that the dykes are mixtures or hybrids. Consequently, dyke chemistry cannot simply be inverted to determine the composition of the mantle source.

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