

THE PETROGENESIS AND EMPLACEMENT OF THE NEW HAMPSHIRE PLUTONIC SUITE

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ABSTRACT. A longstanding problem associated with the Acadian orogeny in New England is whether or not the orogeny involved a redistribution of mass and energy within the midcrust without the addition of magmas and/or anomalous heat from the mantle and/or lower crust. Several closed system models suggest the New Hampshire Plutonic Suite (NHPS) resulted from essentially *in situ* partial melting of the metasedimentary rocks of the Central Maine terrane (CMT) at midcrustal levels. Crustal thickening models, shear heating along the decollement, and the presence of U-enriched metasedimentary rocks have been suggested as heat sources for anatexis. While applicable to numerous NHPS plutons, there are several difficulties with these models as explanations for the NHPS as a whole: 1) the syntectonic Bethlehem, Kinsman, and Spaulding members are too CaO- and Na₂O-rich to have been derived from the exposed metapelites; deeper seated metagraywackes may account for some of the NHPS source rocks but not for the metaluminous Spaulding Tonalite and other more mafic NHPS rocks; 2) some NHPS $\delta^{18}\text{O}$, Nd, and Sr isotopic values are outside the range of CMT and Bronson Hill terrane metasedimentary rocks, requiring a more primitive mantle-like component; 3) the temperatures of the Kinsman and the Spaulding magmas were higher than those attained during regional metamorphism associated with crustal thickening, greater than 850°C; 4) magmatic epidote and hornblende geobarometry for the Winnepesaukee pluton of the Spaulding Tonalite suggest intratelluric crystallization, initially at pressures ≥ 6 Kbars, deeper than the inferred decollement between the allochthonous CMT terrane metasedimentary rocks and underlying basement. These data suggest that while the NHPS contains a significant component from the CMT metasedimentary rocks, the metaluminous Spaulding rocks, mafic enclaves in the Kinsman, and mafic rocks in northeastern Vermont, western Maine and southeastern New Hampshire provide evidence of heat and mass beyond that associated with closed, midcrustal system processes.

INTRODUCTION

Numerous models have been proposed to explain the origin of granitic magmatism in orogenic belts and the New Hampshire Plutonic Suite (NHPS) in particular. Several models suggest that lower crustal materials partially melt 30 to 50 Ma after crustal thickening because of the delay of high temperature isotherms to sweep upward through the colder, depressed crust (England and Thompson and others, 1984; Chamberlain and England, 1985; Zen, 1988). Bradley and others (2000) show a progression in age of Acadian deformation across the state of Maine from 423 Ma in the south to 387.5 Ma in the north. Projecting the times of deformation along strike to the locations of the NHPS plutons of interest suggests the deformation front progressed through central New Hampshire between 418 to 409 Ma. The ages of the syntectonic NHPS range from 413 ± 5 Ma to 398 Ma (Lyons and Livingston, 1977; Barreiro and Aleinikoff, 1985; Eusden and Barreiro, 1988; Lyons and others, 1997), overlapping the range of Acadian deformation and crustal thickening, not 30 to 50 Ma younger as predicted by crustal thickening models. Thus these models explain well post-tectonic plutons, but do not address the association of syntectonic magmatism that is contemporaneous with crustal thickening.

Those who interpret the syntectonic plutons as crustal melts call upon additional heat sources beyond that of simple crustal thickening. One suggestion for this heat source is that some of the sediments of the CMT of Northern New England were

deposited in an anoxic basin where U was reduced and precipitated to enrich the sediments in radioactive elements (Chamberlain and Sonder, 1990). During orogenesis, this extra heat source is proposed to have contributed to high grade metamorphism and partial melting in the middle crust, thereby explaining the spatial relation between the NHPS plutons and high grade metamorphic rocks in New England.

A modified crustal thickening model has been proposed by Brown and coworkers (Brown and Pressley, 1999; Brown and Solar, 1999 and references therein). They suggest that in transpressional orogenic belts such as the New England Appalachians, melting occurred along the Central Maine belt shear zone with pulsed magma flow taking place through structurally controlled channels. Melting occurred farther to the east where crustal thickening had already taken place (Bradley and others, 2000) at deeper levels along the eastward dipping shear zone.

This model explains well some of the peraluminous NHPS such as the Phillips pluton, but cannot account for the entire suite. For example, the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of several anomalous Bethlehem and Kinsman samples of Lyons and Livingston (1977) range from 0.7036 to 0.704. These values are lower than other samples from these plutons and are considerably more primitive than the CMT metasediments whose time equivalent initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios average 0.7198 (Lathrop and others, 1996; Cullers and others, 1997). As a result of this primitive component found in the NHPS and other orogens, many researchers take an open system view for the origin of orogenic magmatism, suggesting that even in compressional tectonic settings, mantle input exists, either in the form of above normal heat flux or by the input of basaltic magmas, contributing to crustal anatexis and granitic magmatism (Nielson and others, 1976; Lyons and Livingston, 1977; Wickham and Oxburgh, 1985; Duke and Rumble, 1986; Lux and others, 1986; Lyons, 1988; Dorais and Paige, 2000).

This paper addresses the origin of the synplutonic magmatism in New Hampshire during the Acadian orogeny, the tectonic implications for the NHPS spatial and compositional relations, and the methods of emplacement of the three syntectonic NHPS members and the post-tectonic granites. We test the models of Chamberlain and Sonder (1990) and Brown and Solar (1999) that the Acadian orogeny involved redistribution of mass and energy within the middle crust without the influence of anomalous heat and/or magmas from the mantle/lower crust.

GEOLOGIC SETTING

The New Hampshire Plutonic Suite (NHPS) of the Acadian orogeny consists of four members, the Bethlehem Granodiorite (410 ± 5 Ma, Lyons and others, 1997), Kinsman Granodiorite (411 ± 19 Ma, Lyons and Livingston, 1977; 413 ± 5 Ma, Barreiro and Aleinikoff, 1985), Spaulding Tonalite (393 ± 5 Ma, Lyons and Livingston, 1977; 398 Ma, Eusden and Barreiro, 1988), and ~ 365 Ma Concord Granite (fig. 1; Billings, 1956; Lyons and others, 1997). The first three members are syntectonic; the Concord Granite is post-tectonic, having been emplaced up to 50 Ma after tectonism (Lyons and Livingston, 1977; Harrison and others, 1987).

In New Hampshire, most of the NHPS plutons intruded metasedimentary rocks of the CMT. The majority of the sediments were shed from western sources into a deepwater depositional basin that closed in the Middle Devonian by convergence of North America and the Avalonian plates during the Acadian orogeny (Boone and others, 1970; Hall and others, 1976; Hatch and others, 1983; Moench and Pankiwskyj, 1988). The New Hampshire portion of the CMT consists of rocks that are dominantly metapelitic, forming the Silurian Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations. Overlying these formations is the Lower Devonian Littleton Formation, which consists of turbidites and volumetrically minor amounts of volcanic rocks derived from an eastern source (Roy and Mencher, 1976). Southeastward in Maine,

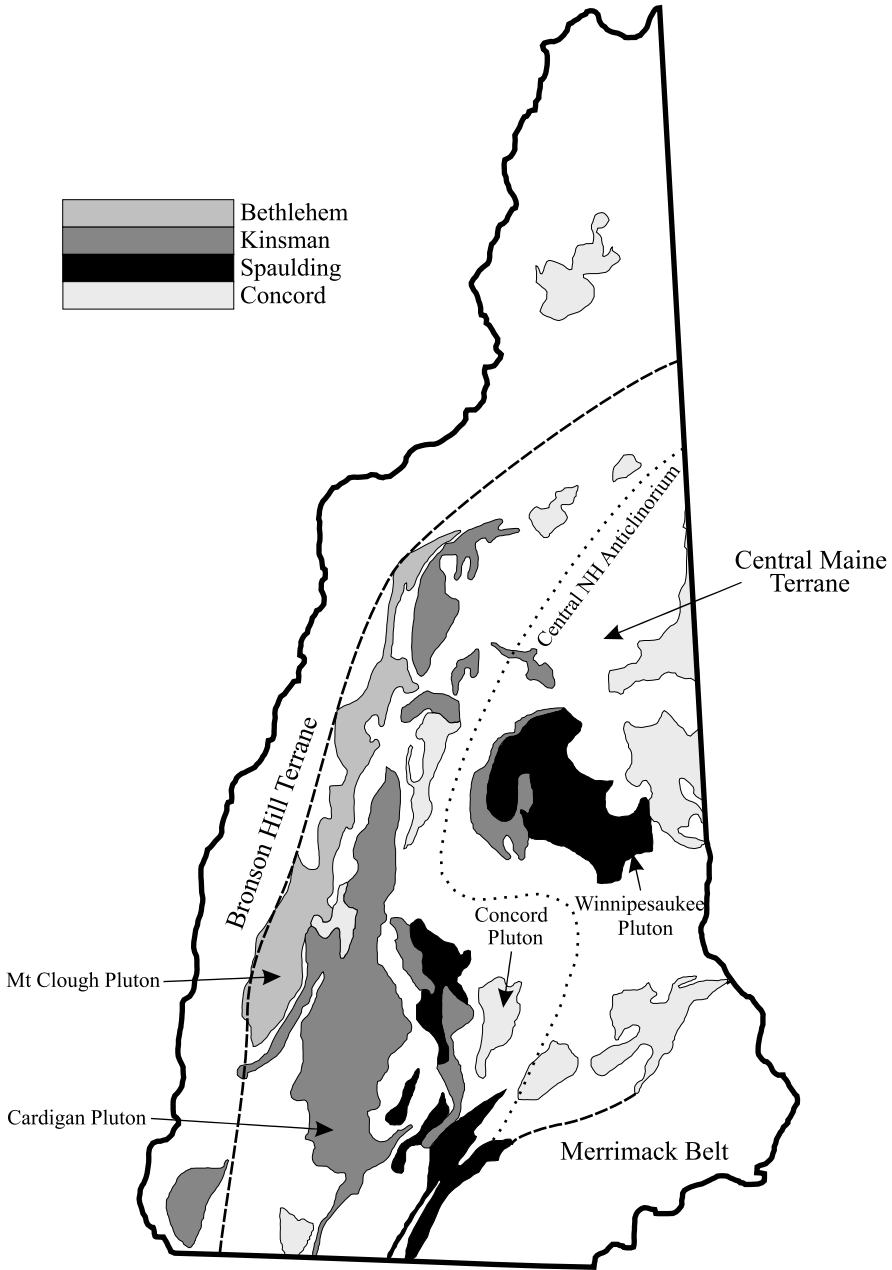


Fig. 1. Map of New Hampshire showing the location of the four members of the New Hampshire Plutonic Suite with respect to the Central New Hampshire Anticlinorium or dorsal zone (after Lyons and others, 1997).

metagraywackes are more abundant, particularly in the Hutchins Corner Formation (formally known as the Vassalboro Formation, Osberg, 1988).

The New Hampshire CMT metasedimentary rocks experienced multiple episodes of folding during the Acadian orogeny (Englund, 1976; Nielson, 1981; Lyons, 1988). F1 consists of east and west verging nappes that are separated by a pop up or dorsal

zone (Eusden and others, 1987; Eusden and Lyons, 1993) called the Central New Hampshire anticlinorium with an axial trace extending north-northeast across central New Hampshire (fig. 1). The westward verging nappes subsequently experienced isoclinal to recumbent folding with northeast trending axes (F2). Broad open folds with northwest to westerly trending axes define F3. F4 is defined by open to isoclinal folds with northeast trending axes.

The CMT metasedimentary rocks are allochthonous. The terrane was transported to the west during the Acadian orogeny, followed by underthrusting of the basement during the Alleghanian orogeny (Zartman, 1988; Stewart, 1989; Wintsch and others, 1992). A decollement is inferred to lie between the CMT rocks and the underlying basement (Brown and Solar, 1999). The thickness of the CMT metasedimentary rocks above the decollement has been estimated to be as thin as 3 kilometers by Eusden and Lyons (1993), other estimates range up to 10 kilometers (Thompson and others, 1993; Stewart and others, 1993).

While field relations and isotopic data clearly indicate that the Bethlehem Granodiorite and the Kinsman Granodiorite are older than the Spaulding Tonalite, interpretations of the age relations of the Bethlehem and Kinsman in the literature are contradictory. Much of this confusion resulted from interpretations of the observation of Quinn (1944) who reported that xenoliths of the Meredith Porphyritic Granite, a Kinsman Granodiorite equivalent, are present in the Winnepesaukee Tonalite. The Winnepesaukee Tonalite has been considered by some to be Bethlehem equivalent (Nielsen and others, 1976; Lathrop and others, 1996), hence the Kinsman would be the older of the two units. We present data below to show that the Winnepesaukee Tonalite is actually a Spaulding Tonalite equivalent, thereby removing the field evidence that the Kinsman is older than the Bethlehem. The geologic map of New Hampshire (Lyons and others, 1997) lists the age of the Winnepesaukee Tonalite as 392 ± 3 Ma which is also compatible with the Spaulding, not the Bethlehem member.

The Bethlehem Granodiorite is located in the western regions of the state and farthest from the dorsal zone. This member of the NHPS is the most tectonized of the three syntectonic members, traditionally having been called the Bethlehem Gneiss (Billings, 1956). The rock is a medium-grained biotite gneiss that is predominantly strongly to weakly foliated but in some locations, is unfoliated and massive. The largest Bethlehem intrusion is the Mt. Clough pluton (fig. 1).

The dominant rock type in the Bethlehem is granodiorite, but granite is also abundant. Minor amounts of rocks as mafic as quartz diorite are also present. The Bethlehem consists of quartz, oligoclase-andesine, biotite, K-feldspar, and minor amounts of muscovite. Many grains are up to 5 millimeters in diameter, but these generally are aggregates of smaller grains from post-solidification deformation.

The Kinsman Granodiorite is located east of the Bethlehem plutons, closer to the dorsal zone. The largest Kinsman intrusion is the Cardigan pluton (fig. 1; Clark and Lyons, 1986; Plank, 1987). There are several other large Kinsman-type plutons and in this paper, we present new data on the Meredith Porphyritic Granite of the Winnepesaukee pluton (Quinn, 1944) to compliment the Cardigan pluton analyses from Clark and Lyons (1986).

The Kinsman Granodiorite exhibits variable degrees of deformation but in general, not as much as the Bethlehem Granodiorite. It is a medium to coarse grained, massive to strongly foliated rock consisting of oligoclase-andesine, quartz, Kspar, biotite, muscovite and variable amounts of garnet. The rock is very distinctive with Kspar crystals up to 15 centimeters in length. The Kinsman is predominantly granodiorite, but a continuum exists between minor amounts of quartz diorite and relatively abundant granite. Some outcrops have garnetite pods consisting of up to 60 percent garnet (Clark and Lyons, 1986; Plank, 1987). Clark and Lyons (1986) present evidence

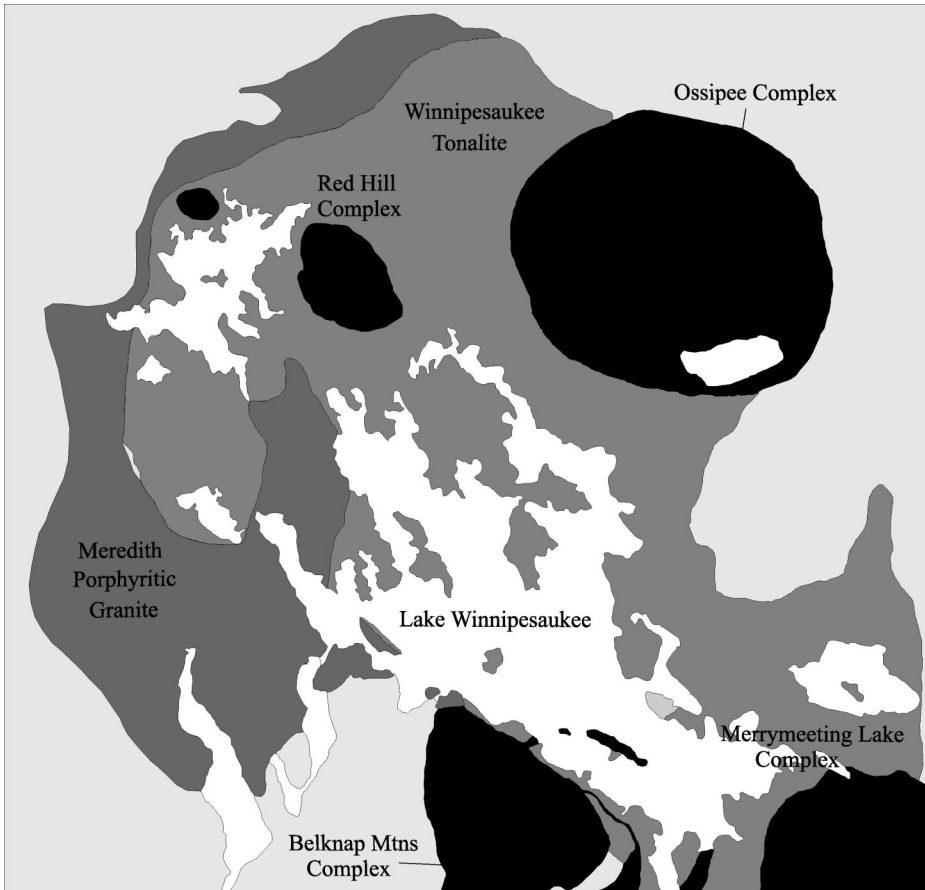


Fig. 2. Geologic map of the Winnepesaukee Pluton showing the locations of the Meredith Porphyritic Granite (dark gray), the Winnepesaukee Tonalite (intermediate gray), Mesozoic White Mountain Magma Series intrusions (black), country rocks (light gray), and Lake Winnepesaukee and other lakes (white).

that the garnetite pods are residua of partial melting and that the compositional variation of the pluton is controlled by restite unmixing.

The Meredith Porphyritic Granite phase of the Kinsman Granodiorite is granodiorite to granite in composition, but minor amounts of more mafic rocks in the form of mafic enclaves and mingled zones are present, consisting primarily of tonalitic rocks. These mafic rocks display typical quench and mingling features (Reid and others, 1983; Dorais and others, 1990).

The Spaulding comprises several moderately sized plutons located along a north-south axis in the center of the state, primarily adjacent to the dorsal zone (fig. 1; Duke, 1978). The largest of these is the 393 ± 2 Ma Winnepesaukee Tonalite (fig. 2) which, on the Geologic Map of New Hampshire, is listed as its own distinct plutonic member that is probably coeval with the Spaulding Tonalite (Lyons and others, 1997). Based on similar field relations, and mineralogic and geochemical characteristics that are discussed below, we group the Winnepesaukee Tonalite as part of the Spaulding Tonalite.

The Spaulding Tonalite is the most mafic of the NHPS. The Spaulding Tonalite ranges from quartz diorite to granite in composition (Quinn, 1944; Fowler-Billings,

1949; Duke, ms, 1978). The Winnepesaukee Tonalite covers this same compositional range and displays a distinct geographic variation in composition. Quartz diorite is located along the western margin, tonalite in the central regions, and granodiorite to granite in the eastern portions of the pluton.

In common with the other syntectonic members of the NHPS, the Winnepesaukee Tonalite exhibits variable degrees of deformation, but is generally less deformed than the Bethlehem and Kinsman, ranging from localized, moderately sheared rocks to undeformed rocks at many localities where deformation is not noted in outcrop or thin section. The nonfoliated, granitic portions of the pluton are similar to the post-tectonic Concord Granite.

The mineralogy of the Winnepesaukee Tonalite varies because the rocks span the range from metaluminous to peraluminous compositions. The more mafic portions of the Winnepesaukee Tonalite consist of minor amounts of quartz diorite with hornblende, plagioclase, and biotite as dominant minerals. Magmatic epidote is present in the hornblende-bearing rocks. With increasing silica content, hornblende proportions diminish, as biotite becomes the dominant mafic silicate. Muscovite is present in the most felsic rocks.

The Concord Granite (Lyons and others, 1997) is a post-tectonic, two mica granite named after the type pluton at Concord, New Hampshire (Hitchcock, 1877). The Concord Granite ranges from 30 to 50 Ma younger than the syntectonic NHPS plutons (Lyons and Livingston, 1977; Harrison and others, 1987; Lyons and others, 1997). Unlike the three syntectonic members of the NHPS, the Concord Granite shows no spatial pattern in emplacement, having intruded the CMT metasedimentary rocks across New Hampshire (fig. 1), in the Northeast Kingdom batholith of Vermont, and across western Maine. The Concord two-mica granites have been treated in greater detail in another publication (Dorais and Paige, 2000) and are only briefly discussed in this paper.

ANALYTICAL METHODS

Mineral compositions were determined at Indiana University using a Cameca SX50 electron microprobe with 15 KV accelerating voltage and 20 uA beam current for garnet and amphibole and 10 uA for micas and feldspar. Data were reduced using the PAP correction scheme of Pouchou and Pichoir (1985). Bulk-rock major and selected trace element analyses were conducted by XRF techniques at Michigan State University. Analyses of additional trace elements were obtained by INAA at the Phoenix Memorial Laboratory at the University of Michigan.

Sm and Nd concentrations and the $^{147}\text{Sm}/^{144}\text{Nd}$ ratio were measured at the University of California Santa Cruz utilizing a mixed ^{149}Sm - ^{150}Nd tracer. Digestions were carried out in Teflon bombs at 180°C for 72 hours to assure trace mineral dissolution. Chemical separation of Sm and Nd were carried out in a class 1000 clean laboratory and follow procedures modified after Richard and others (1976). Nd isotope ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. All data were collected on a 9-collector VG54-WARP in static mode. La Jolla Nd isotopic standard yielded $^{143}\text{Nd}/^{144}\text{Nd} = 0.511858 \pm 10$ ($n = 5$) over the period of analyses. Sr isotopic ratios were measured in peak-jumping mode. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. Sr standard reference material 987 yielded $^{87}\text{Sr}/^{86}\text{Sr} = 0.710235 \pm 15$ ($n = 5$).

MINERALOGY

Amphibole.—Of the NHPS rocks in the CMT in New Hampshire, amphibole is only present in the Spaulding Tonalite (Billings, 1956). The Winnepesaukee Tonalite contains quartz diorite with approximately 40 percent hornblende. The modal proportion of hornblende decreases in rocks to the east where it is replaced by biotite as the

TABLE 1
Representative amphibole analyses, Winnepesaukee Pluton

	WQD-1 Amph-1a	WQD-1 Amph-2a	WQD-1 Amph-5a	WQD-1 Amph-1b	WQD-1 Amph-1c	WQD-1 Amph-2	WQD-1 Amph-3	WQD-2 Amph-1a
SiO ₂	42.05	42.81	42.08	41.72	43.05	42.16	42.71	43.10
TiO ₂	1.32	1.44	1.10	0.89	0.94	0.83	1.01	0.74
Al ₂ O ₃	10.68	10.16	10.87	12.34	11.70	11.66	11.48	11.09
MgO	10.02	10.36	9.71	8.77	9.66	8.99	9.40	9.80
CaO	11.81	11.62	11.89	11.43	11.42	11.48	11.29	11.82
MnO	0.23	0.27	0.36	0.28	0.22	0.34	0.30	0.52
FeO	18.83	18.06	18.92	19.65	19.17	20.00	18.97	18.56
Na ₂ O	1.20	1.41	1.26	1.40	1.34	1.32	1.39	1.22
K ₂ O	0.94	0.94	1.06	0.99	0.87	0.93	1.03	1.03
F	0.14	0.15	0.21	na	na	na	na	na
Cl	0.28	0.23	0.29	0.42	0.36	0.37	0.40	0.13
total	97.50	97.45	97.75	97.89	98.73	98.08	97.98	98.01
O=F	-0.06	-0.06	-0.09	-	-	-	-	-
O=Cl	-0.06	-0.05	-0.06	-0.10	-0.08	-0.08	-0.09	-0.03
TOTAL	97.38	97.34	97.60	97.79	98.65	98.00	97.89	97.98
Liq Mg#	32.90	33.10	30.60	28.60	32.70	29.60	30.80	31.40

	WQD-2 Amph-1b	WQD-2 Amph-2a	WQD-2 Amph-3a	WQD-13 Amph-2a	WQD-13 Amph-3a	WQD-13 Amph-3b	WQD-13 Amph-4a	WQD-13 Amph-4b
SiO ₂	43.22	43.99	43.29	44.26	44.15	44.11	45.29	44.75
TiO ₂	0.82	1.01	0.77	1.34	1.19	0.98	1.25	1.18
Al ₂ O ₃	11.14	10.90	11.81	9.88	10.28	10.27	9.14	9.64
MgO	9.57	9.68	9.37	12.40	12.35	12.39	13.10	12.75
CaO	11.65	11.61	12.04	11.90	12.20	11.90	12.20	12.07
MnO	0.43	0.51	0.44	0.49	0.43	0.37	0.47	0.43
FeO	18.56	18.40	18.78	14.33	14.46	14.51	13.80	14.15
Na ₂ O	1.16	1.24	1.19	1.20	1.18	1.18	1.17	1.15
K ₂ O	1.09	1.07	1.09	0.86	0.78	0.88	0.72	0.69
F	na	na	na	0.24	0.22	0.28	0.15	0.20
Cl	0.12	0.08	0.10	0.27	0.20	0.24	0.20	0.21
total	97.76	98.49	98.88	97.17	97.44	97.11	97.49	97.22
O=F	-	-	-	-0.10	-0.09	-0.12	-0.06	-0.08
O=Cl	-0.03	-0.02	-0.02	-0.06	-0.04	-0.06	-0.04	-0.05
TOTAL	97.73	98.47	98.86	97.01	97.31	96.93	97.39	97.09
Liq Mg#	30.4	30	28.4	43.40	42.50	44.50	44.80	45.00

rocks become more felsic. The amphibole is tschermakite in both the quartz diorite and the tonalite (table 1; Leake and others, 1997). Amphibole Mg/(Mg+Fe) values are the same in quartz diorite sample WQD-1 and tonalite sample WQD-2, regardless of the different bulk-rock SiO₂ contents, as is the calculated Mg# of the liquids from which they crystallized (table 1).

Mica.—Biotite is the most abundant mafic silicate in the Winnepesaukee pluton. Compositions are plotted in Siderophyllite-Eastonite-Phlogopite-Annite space in figure 3 and listed in table 2. The composition of biotite in the Winnepesaukee Tonalite varies widely as rock compositions range from metaluminous to peraluminous. Biotite coexisting with hornblende is relatively Mg-rich and Al-poor and plots in the biotite

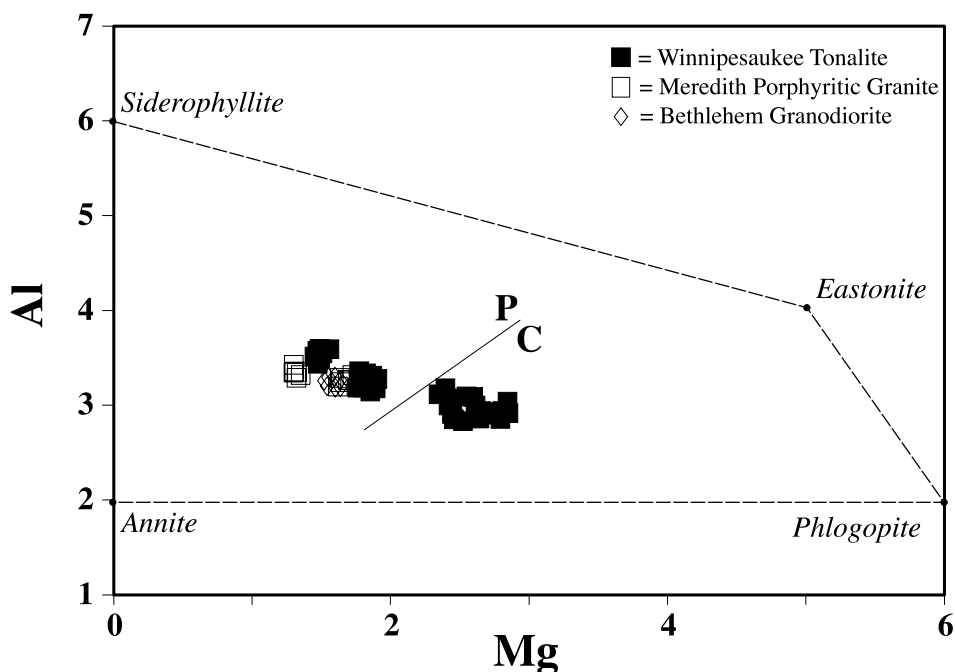


Fig. 3. Biotite Al versus Mg contents (atoms per formula unit calculated to 22 oxygens). The Winnepesaukee Tonalite displays a range of biotite compositions (filled squares), extending from the field of biotite in calc-alkaline rocks (labeled C) to the peraluminous field (labeled P). Biotite in the Meredith Porphyritic Granite (open squares) and the Bethlehem Granodiorite (open triangles) also plot in the peraluminous field.

from calc-alkaline rocks field (Abdel-Rahman, 1994; Stussi and Cuney, 1996). The two-mica granite phase of the Winnepesaukee Tonalite has biotite compositions reflective of peraluminous magmas. Biotite in the Meredith Porphyritic Granite is also indicative of peraluminous compositions and is identical in composition to biotite in the Bethlehem Granodiorite.

Muscovite is present in the most felsic portions of the Winnepesaukee Tonalite and in all Meredith Porphyritic Granite rocks except the tonalitic enclaves. Compositions are rich in Al (total) and poor in (Fe + Mg), characteristic of muscovite in strongly peraluminous granites (fig. 4A) as is muscovite in the Bethlehem Granodiorite. Paragonite contents range from ~ 5 percent in the two-mica rocks of the Winnepesaukee Tonalite to about 10 percent in the Meredith Porphyritic Granite (fig. 4B). Muscovite in the Bethlehem Granodiorite is paragonite-poor compared to muscovite in the Winnepesaukee pluton, containing <5 percent.

Feldspar.—Feldspar compositions are given in table 3 and illustrated in figure 5. Plagioclase ranges from An₄₀ to An₀ in the Winnepesaukee Tonalite, An₃₅ to An₅ in the Meredith Porphyritic Granite, and spans only a narrow range of An₃₆ to An₃₀ in our limited number of Bethlehem Granodiorite samples. Potassium feldspar is present in all three syntectonic members of the NHPS but is absent in the most mafic portions of the Winnepesaukee Tonalite. Kspar compositions range from Or₈₅ to Or₉₈. The large phenocrysts of microcline in the Meredith Porphyritic Granite range from Or₉₀ to Or₉₂.

The quartz dioritic phase of the Winnepesaukee Tonalite contains relatively sodic plagioclase (An₃₅) for a rock with as little as 47 percent SiO₂. The plagioclase is no

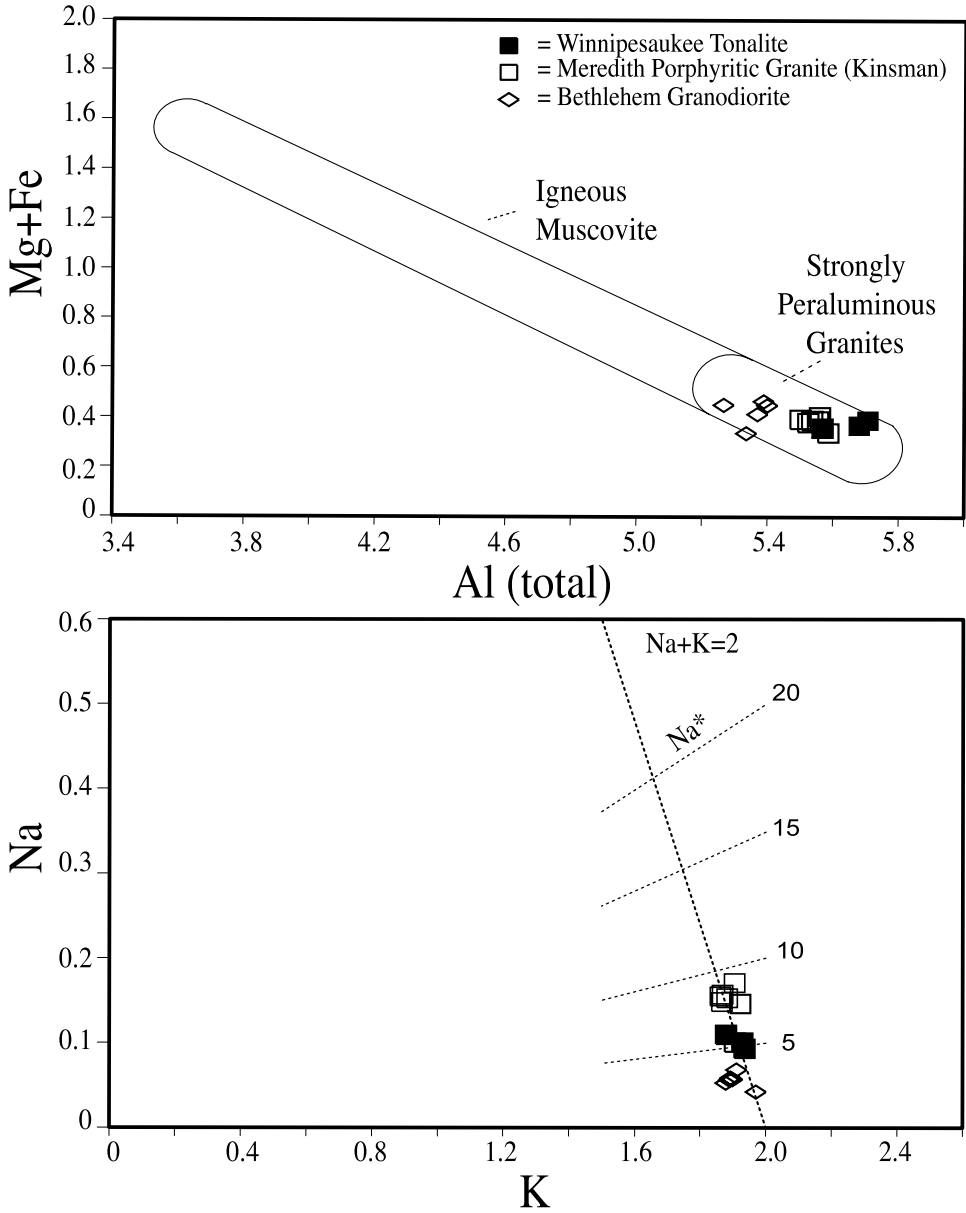


Fig. 4. A) Fm (total Fe and Mg) versus total Al (atoms per formula unit calculated to 22 oxygens) for muscovite (after Zane and Rizzo, 1999) of the Winnepesaukee Tonalite, Meredith Porphyritic Granite, and Bethlehem Granodiorite. The figure shows the fields of igneous muscovite with muscovite from strongly peraluminous granites plotting at low values of Fm and high Al contents. All the muscovite of this study plot in the strongly peraluminous field. B) Muscovite Na versus K contents (after Zane and Rizzo, 1999). The negatively sloped Na + K line represents full occupancy; $Na^* = 100 Na / (Na + K)$. Muscovite in the Meredith Porphyritic Granite is richest in Na^* at values between 7-9. Muscovite in the Winnepesaukee Tonalite has values ~5 whereas muscovite in the Bethlehem Granodiorite is the poorest in Na^* .

TABLE 2
Representative mica analyses, Winnepesaukee Pluton, Mt. Clough pluton

	WQD-1 Bio-1a	WQD-1 Bio-1b	Wqd-1 Bio-2a	WQD-2 Bio-1a	WQD-2 Bio-1b	WQD-2 Bio-2a	WQD-3 Bio-1a	WQD-3 Bio-1
SiO ₂	36.02	36.75	37.11	36.14	35.73	35.98	34.96	33.63
TiO ₂	1.94	2.02	2.02	2.31	2.26	2.11	3.34	3.40
Al ₂ O ₃	16.43	16.55	16.60	16.21	15.95	16.22	17.75	17.40
MgO	12.39	12.77	12.20	11.33	11.09	11.33	8.03	8.17
CaO	0.02	0.02	0.01	0.02	0.00	0.02	0.08	0.00
MnO	0.09	0.06	0.11	0.30	0.33	0.33	0.48	0.46
FeO	18.47	18.78	18.82	20.37	20.17	19.57	23.10	23.61
Na ₂ O	0.20	0.22	0.19	0.19	0.19	0.16	0.21	0.30
K ₂ O	9.70	9.73	9.74	9.67	9.76	9.46	9.48	9.50
F	0.31	0.27	0.17	0.28	0.38	0.25	0.69	0.75
Cl	0.17	0.19	0.29	0.11	0.12	0.10	0.05	0.06
total	95.74	97.36	97.26	96.93	95.98	95.53	98.17	97.28
O=F	-0.13	-0.11	-0.07	-0.12	-0.16	-0.10	-0.29	-0.32
O=Cl	-0.04	-0.04	-0.06	-0.03	-0.03	-0.02	-0.01	-0.01
TOTAL	95.57	97.21	97.13	96.78	95.79	95.41	97.87	96.95

	WQD-3 Bio-1b	WQD-7 Bio-1a	WQD-7 Bio-1b	WQD-7 Bio-2a	WQD-8 Bio-1b	WQD-8 Bio-2a	WQD-8 Bio-3a	WQD-10 Bio-1a
SiO ₂	34.67	34.78	34.93	35.10	34.21	34.01	33.87	32.94
TiO ₂	3.15	2.29	2.13	2.82	2.98	3.33	3.20	1.92
Al ₂ O ₃	17.75	17.60	16.82	16.66	17.93	18.29	18.13	19.17
MgO	7.86	10.48	11.23	10.62	8.24	7.87	8.07	6.43
CaO	0.01	0.05	0.01	0.04	0.00	0.01	0.01	0.00
MnO	0.32	0.31	0.33	0.26	0.36	0.28	0.23	0.57
FeO	23.49	20.55	20.97	21.43	22.02	22.02	22.20	25.14
Na ₂ O	0.31	0.24	0.24	0.24	0.17	0.20	0.17	0.29
K ₂ O	9.44	9.37	9.53	9.50	9.70	9.62	10.00	9.45
F	0.68	0.68	0.67	0.61	0.55	0.37	0.57	0.67
Cl	0.00	0.04	0.16	0.05	0.02	0.04	0.01	0.03
total	97.68	96.39	97.02	97.33	96.18	96.04	96.46	96.61
O=F	-0.29	-0.29	-0.28	-0.26	-0.55	-0.15	-0.24	-0.28
O=Cl	0.00	0.01	-0.04	-0.01	-0.02	-0.01	-	-0.01
TOTAL	97.39	96.11	96.70	97.06	95.61	95.88	96.22	96.32

more calcic than that present in the tonalitic to granodioritic phase (up to An₄₅) of the Winnepesaukee Tonalite.

Fe-Ti Oxides.—Many of the more felsic rocks of the Winnepesaukee Tonalite only contain minor amounts of Fe-Ti oxides. Oxides are relatively abundant in the quartz diorite, which contains both ilmenite and magnetite, most commonly in the form of 0.5 millimeter ilmenite grains with magnetite exsolution lamellae. The ilmenite has compositions of approximately 0.9 X_{ilm} with the lamellae containing approximately 0.35 X_{usp} (table 4).

Epidote.—Epidote occurs as a magmatic phase in the hornblende-bearing quartz diorite and tonalite of the Winnepesaukee Tonalite. Where rimmed by hornblende or biotite, epidote has euhedral morphology with idiomorphic zoning (fig. 6). Adjacent

TABLE 2
(continued)

	WQD-10 Bio-1b	WQD-10 Musc-1a	WQD-10 Musc-1b	WQD-12 Bio-1a	WQD-12 Bio-1r	WQD-12 Bio-2b	MPG-2 Bio-1a	MPG-2 Bio-1b
SiO ₂	34.24	46.07	45.65	36.54	36.75	35.79	33.51	33.39
TiO ₂	1.86	0.34	0.32	2.26	1.69	1.64	3.28	3.29
Al ₂ O ₃	19.00	35.58	35.43	16.61	16.88	17.21	18.06	18.12
MgO	6.40	0.75	0.71	11.02	11.62	11.44	5.80	5.58
CaO	0.00	0.00	0.03	0.00	0.01	0.00	0.01	0.00
MnO	0.59	0.00	0.03	0.30	0.23	0.27	0.23	0.41
FeO	23.78	1.90	1.76	20.32	19.58	19.70	26.32	26.18
Na ₂ O	0.21	0.42	0.39	0.19	0.19	0.15	0.19	0.13
K ₂ O	9.55	11.10	11.17	9.51	9.27	9.77	9.67	9.37
F	0.68	0.21	0.29	0.39	0.42	0.39	0.79	0.69
Cl	0.04	0.01	-	0.13	0.04	0.04	0.13	0.06
total	96.35	96.38	95.78	97.27	96.68	96.40	97.99	97.22
O=F	-0.29	-0.09	-0.01	-0.16	-0.18	-0.16	-0.33	-0.29
O=Cl	-0.01	-	-	-0.03	-0.01	-0.01	-0.03	-0.01
TOTAL	96.05	96.29	95.77	97.08	96.49	96.23	97.63	96.92

	MPG-2 Bio-2a	MPG-2 Musc-1a	MPG-2 Musc-1b	MPG-2 Musc-1c	MPG-3 Bio-1a	MPG-3 Bio-1b	MPG-3 Musc-1a	MPG-3 Musc-1b
SiO ₂	33.52	44.35	44.35	44.66	33.97	34.24	44.42	44.10
TiO ₂	3.28	1.03	1.05	1.14	3.02	3.28	1.35	1.39
Al ₂ O ₃	18.14	34.79	34.53	35.00	17.59	17.46	34.41	34.48
MgO	5.68	0.81	0.73	0.76	7.00	7.01	0.83	0.87
CaO	0.01	0.01	0.00	0.00	0.00	0.00	0.06	0.12
MnO	0.29	0.05	0.00	0.11	0.25	0.14	0.02	0.08
FeO	25.71	2.15	2.12	2.12	24.20	24.46	2.01	1.78
Na ₂ O	0.18	0.57	0.65	0.59	0.17	0.14	0.58	0.56
K ₂ O	9.49	10.78	11.00	10.87	9.25	9.69	10.85	11.09
F	0.64	0.22	0.28	0.24	0.84	0.86	0.30	0.25
Cl	0.11	0.00	0.00	0.00	0.09	0.16	0.01	0.00
total	97.05	94.76	94.71	95.49	96.38	97.44	94.84	94.72
O=F	-0.27	-0.09	-0.12	-0.10	-0.35	-0.36	-0.12	-0.11
O=Cl	-0.02	-	-	-	-0.02	-0.04	-	-
TOTAL	96.76	94.67	94.59	95.39	96.01	97.04	94.72	94.61

	B-3 Bio-1a	B-3 Bio-2a	B-8 Bio-1a	B-8 Bio-1c	B-9 Bio-1a	B-9 Bio-2a	B-9 Musc-1a	B-9 Musc-1b
SiO ₂	35.14	34.80	34.56	34.45	35.14	35.14	45.30	44.67
TiO ₂	3.35	3.55	3.46	2.82	3.50	3.50	1.36	1.15
Al ₂ O ₃	17.80	17.73	17.59	17.91	18.17	17.55	33.80	33.62
MgO	6.88	6.98	6.90	6.95	6.68	6.96	0.94	1.10
CaO	0.01	0.00	0.01	0.03	0.00	0.00	0.00	0.00
MnO	0.32	0.43	0.31	0.29	0.22	0.27	0.03	0.02
FeO	23.09	22.56	23.77	23.79	23.50	23.55	1.98	2.11
Na ₂ O	0.10	0.08	0.05	0.07	0.05	0.10	0.27	0.19
K ₂ O	9.54	9.57	9.49	9.26	9.64	9.72	10.98	11.26
F	0.57	0.48	0.52	0.59	0.47	0.58	0.24	0.23
Cl	0.00	0.02	0.02	0.03	0.03	0.05	0.00	0.00
total	96.80	96.20	96.68	96.19	97.40	97.42	94.90	94.35
O=F	-0.09	-0.24	-0.22	-0.25	-0.20	-0.21	-0.10	-0.10
O=Cl	0.00	-0.01	-0.01	-0.01	0.00	-0.01	0.00	0.00
TOTAL	96.71	95.95	96.45	95.93	97.20	97.20	94.80	94.25

TABLE 3

Representative feldspar analyses, Winnepesaukee pluton, Mt. Clough pluton

	WQD-1 Plag-1a	WQD-1 Plag-1b	WQD-1 Plag-2a	WQD-1 Plag-2b	WQD-2 Plag-1a	WQD-2 Plag-2a	WQD-2 Plag-1	WQD-2 Plag-2
SiO ₂	59.38	59.22	59.19	58.32	59.21	59.22	59.8	59.01
Al ₂ O ₃	26.22	26.35	26.35	26.54	26.7	25.79	25.34	25.41
CaO	7.43	7.29	7.68	7.84	7.67	6.98	7.02	7.41
FeO	0.06	0.09	0.2	0	0.16	0.07	0.05	0
Na ₂ O	7.13	7.18	7.12	6.86	7.07	7.2	7.38	7.31
K ₂ O	0.06	0.06	0.06	0.05	0.16	0.21	0.2	0.22
Total	100.28	100.19	100.6	99.61	100.97	99.47	99.79	99.36
Ab	63.22	63.33	62.42	61.1	61.95	64.32	64.77	63.28
Or	0.36	0.36	0.36	0.29	0.92	1.26	1.18	1.24
An	36.42	35.81	37.21	38.61	37.13	34.42	34.05	35.48

	WQD-3 Plag-1a	WQD-3 Plag-1b	WQD-3 Plag-1c	WQD-3 Plag-2b	WQD-4 Plag-1b	WQD-4 Plag-2a	WQD-4 Plag-2b	WQD-7 Plag-1a
SiO ₂	57.39	57.79	62.45	62.64	62.90	62.71	63.21	61.56
Al ₂ O ₃	26.68	26.06	23.71	23.34	24.27	24.16	23.86	24.73
CaO	8.30	7.43	4.75	4.55	4.57	5.06	4.59	5.93
FeO	0.00	0.03	0.00	0.07	0.06	0.00	0.00	0.00
Na ₂ O	6.69	7.06	8.56	8.85	8.66	8.39	8.58	7.96
K ₂ O	0.20	0.23	0.35	0.26	0.40	0.35	0.39	0.36
Total	99.26	98.60	99.82	99.71	100.86	100.67	100.63	100.54
Ab	58.66	62.37	74.87	76.68	75.62	73.50	75.33	69.30
Or	1.15	1.35	1.99	1.50	2.30	2.01	2.26	2.09
An	40.19	36.28	22.95	21.78	22.04	24.49	22.26	28.57

	WQD-4 Kspar-1b	WQD-7 Kspar-1a	WQD-7 Kspar-1b	WQD-7 Kspar-4	WQD-7 Kspar-5	WQD-7 Kspar-6	WQD-7b Kspar-1a	WQD-7b Kspar-1b
SiO ₂	65.34	64.14	64.13	64.91	64.60	64.25	64.78	64.56
Al ₂ O ₃	18.55	18.64	18.88	18.80	18.74	18.58	18.52	18.75
CaO	0.00	0.03	0.01	0.00	0.03	0.00	0.02	0.00
FeO	0.17	0.04	0.00	0.05	0.03	0.06	0.06	0.02
BaO	0.35	1.40	1.20	0.96	1.58	1.00	0.38	0.44
Na ₂ O	0.56	1.67	1.42	1.24	1.36	1.18	1.01	1.17
K ₂ O	16.26	14.33	14.79	14.93	14.60	15.11	15.32	15.11
Total	101.23	100.25	100.43	100.89	100.94	100.18	100.09	100.05
Ab	4.97	14.67	12.47	10.99	12.04	10.43	9.03	10.48
Or	94.40	82.72	85.34	87.29	84.99	87.78	90.18	88.73
An	0.00	0.14	0.05	0.00	0.15	0.00	0.09	0.00
Cs	0.63	2.47	2.13	1.72	2.83	1.79	0.69	0.79

to tectosilicates, epidote is anhedral and highly resorbed. Complexly zoned allanite cores are common. Epidote compositions commonly contain approximately 25 percent of the pistacite molecule $\{Fe^{3+}/(Fe^{3+} + Al^{VI})\}$, table 5).

Garnet.—Garnet in the Meredith Porphyritic Granite is generally less than 1 modal percent; no garnetite pods as found in the Cardigan pluton (Clark and Lyons, 1986;

TABLE 3
(continued)

	WQD-7 Plag-1b	WQD-7 Plag-4a	WQD-7 Plag-4b	WQD-7b Plag-1a	WQD-7b Plag-1b	WQD-7b Plag-2a	WQD-8 Plag-1a	WQD-8 Plag-4b
SiO ₂	61.46	57.28	58.28	61.19	61.33	61.77	60.99	57.28
Al ₂ O ₃	25.03	27.45	26.68	24.65	24.49	25.01	25.05	26.95
CaO	5.84	8.97	8.58	5.79	5.63	5.85	6.25	9.00
FeO	0.04	0.00	0.00	0.00	0.00	0.00	0.00	0.01
Na ₂ O	7.89	6.30	0.13	8.09	8.08	8.01	7.84	6.25
K ₂ O	0.27	0.16	6.66	0.29	0.29	0.25	0.18	0.16
Total	100.53	100.16	100.33	100.01	99.82	100.89	100.31	99.65
Ab	69.85	55.42	57.51	70.34	70.98	70.21	68.64	55.16
Or	1.59	0.90	1.35	1.65	1.70	1.46	1.02	0.95
An	28.56	43.60	40.93	27.80	27.32	28.33	30.23	43.89

	WQD-8 Kspar-1a	WQD-8 Kspar-1b	WQD-9 Kspar-1a	WQD-9 Kspar-1b	WQD-10 Kspar-1a	WQD-10 Kspar-1	MPG-2 Kspar-1a	MPG-2 Kspar-1b
SiO ₂	64.45	64.49	64.72	64.48	64.67	65.07	65.02	64.58
Al ₂ O ₃	18.99	18.71	18.60	18.63	18.59	19.03	18.89	18.94
CaO	0.00	0.01	0.00	0.00	0.00	0.03	0.02	0.05
FeO	0.08	0.09	0.01	0.01	0.00	0.00	0.00	0.00
BaO	1.14	1.06	0.30	0.30	0.17	0.29	0.17	0.02
Na ₂ O	0.84	0.56	0.90	0.68	0.95	1.03	0.95	0.92
K ₂ O	15.53	16.03	15.82	16.12	15.70	15.62	15.83	15.98
Total	101.03	100.95	100.35	100.22	100.08	101.07	100.88	100.49
Ab	7.47	4.98	7.92	5.99	8.42	9.03	8.30	8.01
Or	90.49	93.07	91.54	93.47	91.27	90.29	91.29	91.73
An	0.00	0.06	0.00	0.00	0.00	0.17	0.10	0.22
Cs	2.04	1.88	0.54	0.54	0.31	0.52	0.31	0.04

	WQD-8 Plag-3a	WQD-9 Plag-1a	WQD-9 Plag-1b	WQD-9 Plag-3a	WQD-10 Plag-1a	WQD-10 Plag-1b	WQD-10 Plag-2a	WQD-10 Plag-3a
SiO ₂	58.76	60.63	59.68	60.15	68.18	64.81	65.59	65.32
Al ₂ O ₃	26.38	25.02	25.16	25.03	19.72	22.07	22.43	22.31
CaO	7.83	6.46	6.87	6.72	0.25	2.95	3.14	3.11
FeO	0.01	0.04	0.00	0.00	0.00	0.00	0.00	0.04
Na ₂ O	6.89	7.65	7.46	7.66	11.43	9.68	9.59	9.59
K ₂ O	0.17	0.27	0.16	0.17	0.09	0.21	0.12	0.13
Total	100.04	100.07	99.33	99.73	99.67	99.72	100.87	100.50
Ab	60.78	67.11	65.60	66.70	98.18	84.51	83.97	84.16
Or	0.97	1.55	0.94	0.97	0.54	1.22	0.68	0.73
An	38.17	31.33	33.38	32.32	1.20	14.25	15.20	15.10

Plank, 1987) were observed in the Meredith Porphyritic Granite. Garnet occurs as anhedral grains, averaging 0.5 millimeter in diameter. Garnet is predominantly almandine in composition (~70 – 80%) with ~5 to 20 percent spessartine, ~5 to 10 percent pyrope and ~1 to 2 percent grossular (table 6) but is zoned from ~5 percent spessartine in the cores to ~20 percent at the rims.

TABLE 3
(continued)

	MPG-4 Kspar-1a	MPG-4 Kspar-1b	MPG-4 Kspar-1d	MPG-4 Plag-2b	MPG-4 Plag-2d	MPG-4 Plag-5a	B-3 Plag-3a	B-3 Plag-3b
SiO ₂	64.54	63.58	64.68	60.59	67.56	59.49	60.49	60.30
Al ₂ O ₃	18.35	18.70	18.61	25.16	20.34	25.30	25.33	25.19
CaO	0.00	0.00	0.01	6.07	1.04	6.27	6.66	6.78
FeO	0.10	0.09	0.01	0.02	0.00	0.00	0.08	0.04
Na ₂ O	0.69	0.85	0.78	7.78	10.79	7.77	7.59	7.39
K ₂ O	16.07	15.50	15.55	0.13	0.07	0.15	0.05	0.17
Total	99.75	98.72	99.64	99.75	99.80	98.98	100.20	99.87
Ab	6.12	7.70	7.07	69.34	94.61	68.57	67.13	65.71
Or	93.88	92.30	92.86	0.76	0.38	0.86	0.31	1.00
An	0.00	0.00	0.07	29.91	5.02	30.50	32.56	33.29

	B-8 Kspar-1a	B-8 Plag-1a	B-8 Plag-2a	B-9 Plag-2a	B-9 Plag-2b	B-9 Kspar-4b	B-9 Kspar-4c	B-9 Kspar-5a
SiO ₂	63.39	59.97	59.72	59.15	58.77	64.58	64.58	63.98
Al ₂ O ₃	18.74	25.21	25.27	25.86	25.85	18.63	18.43	18.54
CaO	0.03	6.72	6.42	7.48	7.30	0.00	0.00	0.00
FeO	0.01	0.11	0.00	0.04	0.00	0.05	0.00	0.00
Na ₂ O	1.03	7.51	7.38	7.10	7.13	0.86	0.89	0.84
K ₂ O	15.13	0.12	0.14	0.23	0.26	15.33	15.53	15.58
Total	98.33	99.64	98.93	99.86	99.31	99.45	99.43	98.94
Ab	9.35	66.45	66.97	62.39	62.87	7.89	7.98	7.53
Or	90.50	0.70	0.87	1.30	1.53	92.11	92.03	92.47
An	0.15	32.85	32.20	36.31	35.60	0.00	0.00	0.00

GEOCHEMISTRY

Major and Selected Trace Elements

Figure 7 illustrates plots of selected major element compositions for the Winnepesaukee Tonalite, the Meredith Porphyritic Granite (table 7), the Spaulding Tonalite (analyses from Duke, ms, 1978), and the Kinsman Granodiorite (analyses of the Cardigan pluton, Clark and Lyons, 1986). Because granitic compositions tend to converge at high SiO₂ contents, only Spaulding and Kinsman samples from the literature with less than 70 weight percent SiO₂ are plotted in order to clearly define the separate trends.

Two main trends are shown in figure 7. With only a few samples as exceptions, the Meredith Porphyritic Granite forms distinct trends from those of the Winnepesaukee Tonalite, being richer in Fe₂O₃, TiO₂ and K₂O and poorer in CaO. The Kinsman Granodiorite of the Cardigan pluton is identical to the Meredith Porphyritic Granite. Likewise, the Bethlehem Granodiorite samples from the Mt. Clough pluton are identical to those of the Meredith Porphyritic Granite and the Kinsman Granodiorite. The Spaulding Tonalite trend encompasses the Winnepesaukee Tonalite trend.

Selected trace element abundances are plotted versus SiO₂ in figure 8. The Winnepesaukee Tonalite samples again are distinct from the Meredith Porphyritic Granite, being richer in Sr and poorer in Rb, Zr, and Y. The Bethlehem Granodiorite is identical in these trace element abundances to the Meredith Porphyritic Granite.

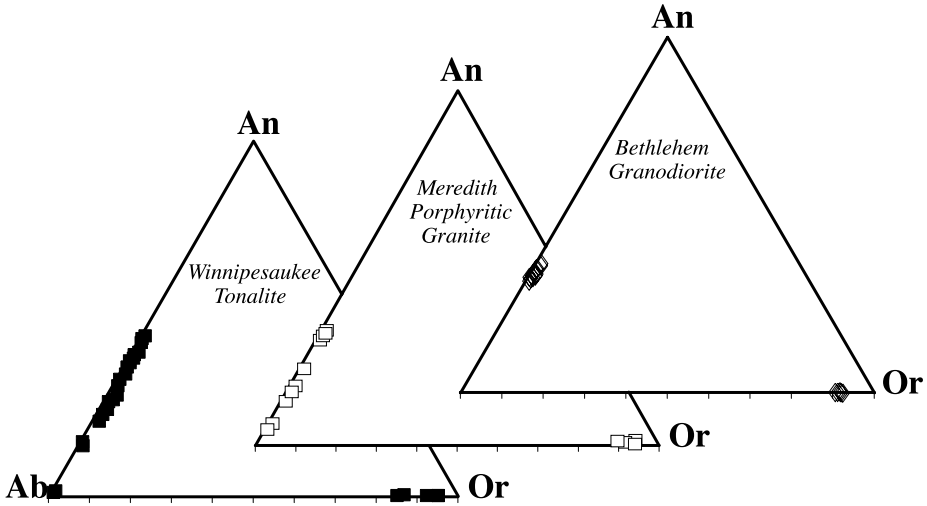


Fig. 5. An-Ab-Or diagrams for the syntectonic members of the New Hampshire Plutonic Suite. Plagioclase in the Winnepesaukee Tonalite and the Meredith Porphyritic Granite display a range of compositions between An_{50} and nearly pure albite. In contrast, the Bethlehem Granodiorite displays a limited range, clustered at An_{35} .

Rare Earth Elements

Chondrite-normalized REE patterns for selected Winnepesaukee Tonalite samples have positive Eu anomalies indicative of plagioclase enrichment (fig. 9A). Chondrite-normalized REE patterns of other Winnepesaukee pluton samples lack or have very small Eu anomalies (fig. 9B). Three tonalitic samples with SiO_2 contents of less than 60 percent are LREE poor and HREE rich compared to three granodioritic to granitic samples with SiO_2 contents between 68 to 71 percent. The tonalites have $(La/Yb)_N$ values of ~ 7 compared to the granodiorites and granitic samples with $(La/Yb)_N$ values of ~ 50 .

Chondrite-normalized REE patterns for Meredith Porphyritic Granite samples are plotted in figure 9C, as are samples of the Kinsman Granodiorite from the Cardigan pluton (Clark and Lyons, 1986) and the Bethlehem Granodiorite from the Mt. Clough pluton (table 7). In general, REE abundances decrease with increasing SiO_2 contents. Two leucogranite samples from the Kinsman and a high silica granite from the Meredith Porphyritic Granite are REE-poor compared to all other samples. With the exception of these samples and two Meredith Porphyritic Granite samples that have flatter LREE patterns, the Meredith Porphyritic Granite, Kinsman Granodiorite, and Bethlehem Granodiorite patterns are similar in shape and slope.

Normalized multi-element diagrams for the five Winnepesaukee Tonalite samples that lack positive Eu anomalies are shown in figure 10A. As expected, abundances of incompatible elements, particularly Rb, Th, Ta, La, and LREE, increase in rocks of higher SiO_2 contents. Troughs of P and Ti are evident for the granodioritic to granitic samples. Figure 10B shows elemental compositions for the Meredith Porphyritic Granite and Bethlehem Granodiorite. The Meredith Porphyritic Granite shows a greater range in values than the limited number of Bethlehem samples, but most of the Meredith samples and the Bethlehem samples are similar.

Oxygen, Nd, and Sr Isotopes

Bulk-rock $\delta^{18}O$ versus SiO_2 values are plotted in figure 11. The Winnepesaukee Tonalite samples display a wide range in $\delta^{18}O$, from 6 to 7 for quartz dioritic to

TABLE 4
Representative Fe-Ti oxide analyses, Winnepesaukee tonalite

	WQD-1 Mag-1a	WQD-1 Mag-2a	WQD-1 Mag-2b	WQD-1 Mag-1a1	WQD-1 Mag-1b1	WQD-1 Ilm-1a	WQD-1 Mag-2a
SiO ₂	0.02	0.04	1.60	0.02	0.00	0.00	0.03
TiO ₂	0.05	0.03	0.16	12.03	11.97	47.58	10.94
Al ₂ O ₃	0.09	0.10	0.38	0.03	0.05	0.00	0.02
Cr ₂ O ₃	0.06	0.00	0.00	0.10	0.15	0.00	0.13
Fe ₂ O ₃	68.67	68.81	67.45	43.12	43.19	10.07	79.03
MgO	0.00	0.02	0.04	0.00	0.00	0.02	0.00
CaO	0.02	0.02	0.09	0.00	0.01	0.00	0.00
MnO	0.00	0.06	0.03	0.16	0.21	1.80	0.12
FeO	31.08	30.98	30.80	40.95	40.84	40.91	9.79
Total	99.99	100.06	100.55	96.41	96.42	100.38	100.06

	WQD-1 Mag-2a	WQD-1 Ilm-2b	WQD-1 Mag-1a	WQD-1 Mag-1b	WQD-1 Ilm-2	WQD-1 Ilm-3a	WQD-1 mag-3
SiO ₂	0.00	0.02	0.00	0.00	0.00	0.01	0.00
TiO ₂	9.79	43.39	11.02	10.74	45.28	49.92	19.48
Al ₂ O ₃	0.01	0.01	0.05	0.02	0.00	0.00	0.02
Cr ₂ O ₃	0.07	0.06	0.10	0.12	0.06	0.01	0.01
Fe ₂ O ₃	46.73	18.32	43.33	44.74	15.20	6.63	28.69
MgO	0.00	0.00	0.04	0.00	0.03	0.00	0.02
CaO	0.00	0.01	0.01	0.02	0.00	0.00	0.02
MnO	0.10	1.72	0.16	0.16	1.64	1.94	0.57
FeO	38.56	37.30	39.15	39.35	39.03	42.93	47.33
Total	95.26	100.83	93.86	95.15	101.24	101.44	96.14

tonalitic samples to 10 for the two-mica granites. The Meredith Porphyritic Granite has higher $\delta^{118}\text{O}$ values, ranging up to 11.5 per mil. Bethlehem Granodiorite samples average 10 per mil.

ϵ_{Nd} (at 400 Ma) values for six Winnepesaukee Tonalite and two Meredith Porphyritic Granite enclaves are plotted against initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in figure 12. Additional plotted samples are Kinsman and Bethlehem rocks from Lathrop and others (1996). The Winnepesaukee samples of this study have ϵ_{Nd} between 0.1 and -5.4 (table 8). The Meredith Porphyritic Granite enclaves have even more primitive values of 4.7 and 5.2. Most samples have primitive initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of ~ 0.704 . Thus, these samples have a strong mantle/lower crustal signature, considerably more primitive than the CMT metasedimentary rocks. One Meredith Porphyritic Granite enclave (MPG-10) has a relatively elevated initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.706. In contrast to the presence of a crustal component shown by its Sr isotopic value, the enclave has a primitive ϵ_{Nd} (at 400 Ma) of 4.7. Holden and others (1987) demonstrated that Rb metasomatism preferentially equilibrates the enclave Rb-Sr isotopic system to host magma compositions whereas the Sm-Nd system is relatively immobile, preserving initial enclave magma Nd isotopic compositions. Sample WQD-10 is a two-mica granite with a relatively high initial $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.7066 and ϵ_{Nd} (at 400 Ma) of -5.4 , plotting adjacent to the CMT field. One additional Meredith Porphyritic Granite sample from Foland and Allen (1991) plots in the CMT field. Several Kinsman and Bethlehem samples (from Lathrop and others, 1996) have relatively primitive values, but most plot in the CMT field.

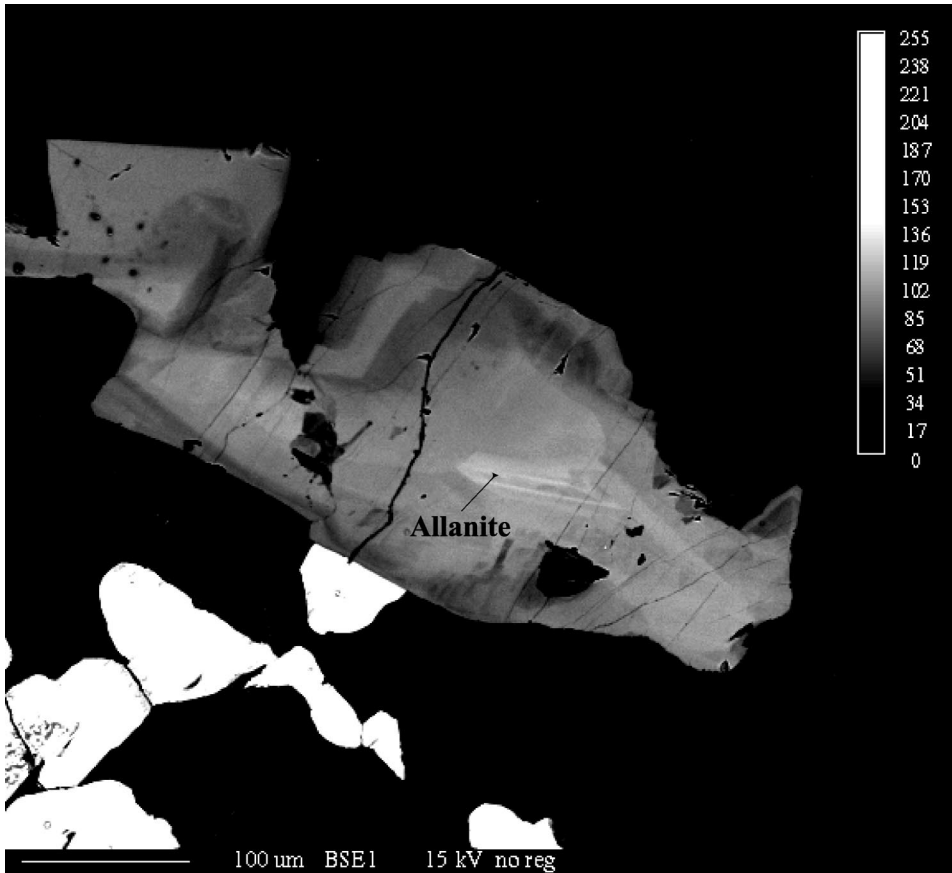


Fig. 6. Backscatter electron image of armored epidote with allanite core in the Winnepesaukee Tonalite showing idomorphic zoning characteristic of igneous growth.

ϵ_{Sr} (at 400 Ma) versus $\delta^{18}\text{O}$ (fig. 13) show that while some Winnepesaukee samples plot within or near the fields defined by CMT and Bronson Hill metasedimentary rocks (Lathrop and others, 1994, 1996), several more mafic samples plot outside the metasedimentary range nearer mantle-derived magmas.

DISCUSSION

Bulk-Rock Compositions, Relationships Among NHPS Members

The similar ages and major and trace element compositions of the Winnepesaukee Tonalite and the Spaulding Tonalite and the compositional distinctions between the Winnepesaukee Tonalite and the Bethlehem Granodiorite indicate that the Winnepesaukee Tonalite is Spaulding and not Bethlehem equivalent. Thus the Meredith Porphyritic Granite xenoliths in the Winnepesaukee Tonalite do not constrain the Kinsman to be older than the Bethlehem.

Several textural, mineralogic, and bulk-rock geochemical features of the Winnepesaukee Tonalite are indicative of cumulate compositions. Given the syntectonic emplacement of the Winnepesaukee Tonalite, we suspect that the Eu enrichment and similar compositional range of plagioclase, amphibole, and biotite in rocks ranging

TABLE 5
Representative epidote and allanite analyses, Winnepesaukee tonalite

	WQD-1 Epi-2c	WQD-1 Epi-2b	WQD-1 Epi-2a	WQD-1 Epi-3b	WQD-1 Epi-3c	WQD-1 Epi-3d	WQD-1 Epi-3e	WQD-1 Epi-1a
SiO ₂	37.26	37.89	37.66	37.89	37.26	37.27	37.41	37.60
TiO ₂	0.02	0.05	0.09	0.05	0.02	0.08	0.12	0.02
ThO ₂	0.00	0.10	0.10	0.10	0.00	0.04	0.00	0.00
Al ₂ O ₃	24.20	24.07	23.81	24.07	24.20	24.31	24.00	24.42
Y ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.03	0.00	0.00
La ₂ O ₃	0.00	0.02	0.01	0.02	0.00	0.00	0.01	0.04
Ce ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	23.77	23.61	23.74	23.61	23.77	23.21	23.27	23.15
MnO	0.18	0.11	0.15	0.11	0.18	0.28	0.11	0.07
Fe ₂ O ₃	12.69	13.32	12.66	14.81	12.69	12.74	12.96	13.16
Total	98.12	99.17	98.22	100.66	98.12	97.96	97.88	98.46
(Fe/Fe+Al)	0.26	0.26	0.25	0.29	0.25	0.25	0.26	0.26

	WQD-1 Epi-1c	WQD-13 Epi-1b	WQD-13 Epi-1c	WQD-1 Allanite- 3a	WQD-1 Allanite- 1a	WQD-1 Allanite- 1b	WQD-13 Allanite- 1a	WQD-13 Allanite- 2a
SiO ₂	37.75	37.92	37.78	34.15	33.59	34.03	33.08	32.34
TiO ₂	0.11	0.13	0.10	0.19	0.14	0.20	0.40	0.25
ThO ₂	0.00	0.09	0.08	0.00	0.00	0.01	0.30	0.00
Al ₂ O ₃	24.45	24.03	24.28	20.69	19.46	19.83	18.56	18.41
Y ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00
La ₂ O ₃	0.01	0.00	0.01	2.96	4.13	4.63	5.19	5.46
Ce ₂ O ₃	0.00	0.00	0.00	6.85	8.21	9.05	10.05	8.69
CaO	23.09	22.93	23.27	16.89	15.69	13.18	14.25	14.95
MnO	0.18	0.12	0.20	0.17	0.16	0.20	0.37	0.27
Fe ₂ O ₃	12.85	13.06	12.65	13.20	12.70	13.01	13.62	13.29
Total	98.44	98.28	98.37	95.10	94.08	94.14	95.83	93.66
(Fe/Fe+Al)	0.25	0.26	0.25					

from quartz diorite to tonalite are the result of squeezing residual liquids from early formed crystals of plagioclase, hornblende, and biotite, leaving quartz diorites with compositions dominated by cumulus phases.

The Winnepesaukee Tonalite samples that lack positive Eu anomalies are more indicative of liquid compositions and show an increase in LREE and depletion in HREE with increasing SiO₂ contents. The depletion in HREE abundance is likely related to hornblende fractionation from parental, tonalitic magmas. Samples ranging from quartz diorite to granodiorite have relatively primitive isotopic compositions with initial ⁸⁷Sr/⁸⁶Sr ratios between 0.70371 to 0.70396, δ¹⁸O values ~7, and ε_{Nd} (at 400 Ma) values of ~ -1 (figs. 12 and 13). These values suggest the lack of, or at most, a minor amount of CMT metasedimentary component in the quartz diorite - tonalite -

TABLE 6
 Representative garnet analyses, Meredith porphyritic granite

	MPG-2 Gar-1a Core	MPG-2 Gar-1b Rim	MPG-2 Gar-1c Core	MPG-2 Gar-1d Int	MPG-2 Gar-1e Int	MPG-2 Gar-1f Int
SiO ₂	36.90	36.62	36.93	36.38	36.55	36.23
TiO ₂	0.07	0.00	0.00	0.06	0.02	0.06
Al ₂ O ₃	21.09	21.01	21.17	21.14	21.07	21.24
Cr ₂ O ₃	0.02	0.00	0.02	0.05	0.02	0.00
MgO	2.64	1.25	2.55	2.40	2.11	1.69
CaO	0.74	0.68	0.73	0.64	0.62	0.65
MnO	2.43	7.37	2.78	2.93	3.63	4.83
FeO	36.08	33.27	36.26	36.30	35.72	34.80
Total	99.97	100.20	100.44	99.90	99.74	99.50

	MPG-2 Gar-1h Rim	MPG-2 Gar-1i Core	MPG-2 Gar-2a Core	MPG-2 Gar-2b Int	MPG-2 Gar-2c Rim	MPG-2 Gar-2d Rim
SiO ₂	36.28	36.31	36.93	36.97	37.08	36.92
TiO ₂	0.03	0.07	0.03	0.01	0.03	0.02
Al ₂ O ₃	21.27	21.59	21.01	20.88	20.93	20.95
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.04	0.00
MgO	1.16	2.87	1.66	1.50	1.09	1.12
CaO	0.67	0.60	0.74	0.73	0.72	0.76
MnO	6.39	2.05	5.02	6.06	8.53	8.68
FeO	33.95	36.34	34.92	33.89	31.56	31.59
Total	99.75	99.83	100.31	100.04	99.98	100.04

granodiorite portion of the Winnepesaukee pluton. However, the two mica granite (WQD-10) has a $\delta^{18}\text{O}$ value of 10.5, an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70663 and ϵ_{Nd} value (at 400 Ma) of -5.4 and plots at the primitive end of the CMT metasedimentary field (fig. 12), indicating the presence of a CMT component in the felsic portions of pluton.

The Meredith Porphyritic Granite is texturally, mineralogically, and compositionally equivalent to the Cardigan pluton of the Kinsman Granodiorite. It contains the same Kspar megacrysts, and is garnet-bearing as is the Kinsman Granodiorite. Likewise, major, minor and trace element abundances (figs. 7, 8, 9C, and 10B) are very similar to those of the Cardigan pluton, supporting earlier conclusions that the Meredith Porphyritic Granite is Kinsman equivalent (Quinn, 1944).

Thompson and others (1968) proposed that the Bethlehem Granodiorite is compositionally equivalent to the Kinsman Granodiorite with the difference in mineralogy simply reflecting a difference in temperature and $f_{\text{H}_2\text{O}}$. The Meredith Porphyritic Granite and the Bethlehem Granodiorite samples are similar in all measured elements (figs. 7, 8, 9C, and 10B), supporting the assertion of Thompson and coworkers that the two rock types crystallized from the same or similar magmas.

With few exceptions, mineral compositions in the Bethlehem Granodiorite and the Meredith Porphyritic Granite are also similar. Biotite compositions in the two NHPS members are similar (fig. 3), however, the Bethlehem Granodiorite contains muscovite that is paragonite-poor compared to muscovite in the Meredith Porphyritic Granite (fig. 4B) and the plagioclase range in the Bethlehem is quite limited (fig. 5).

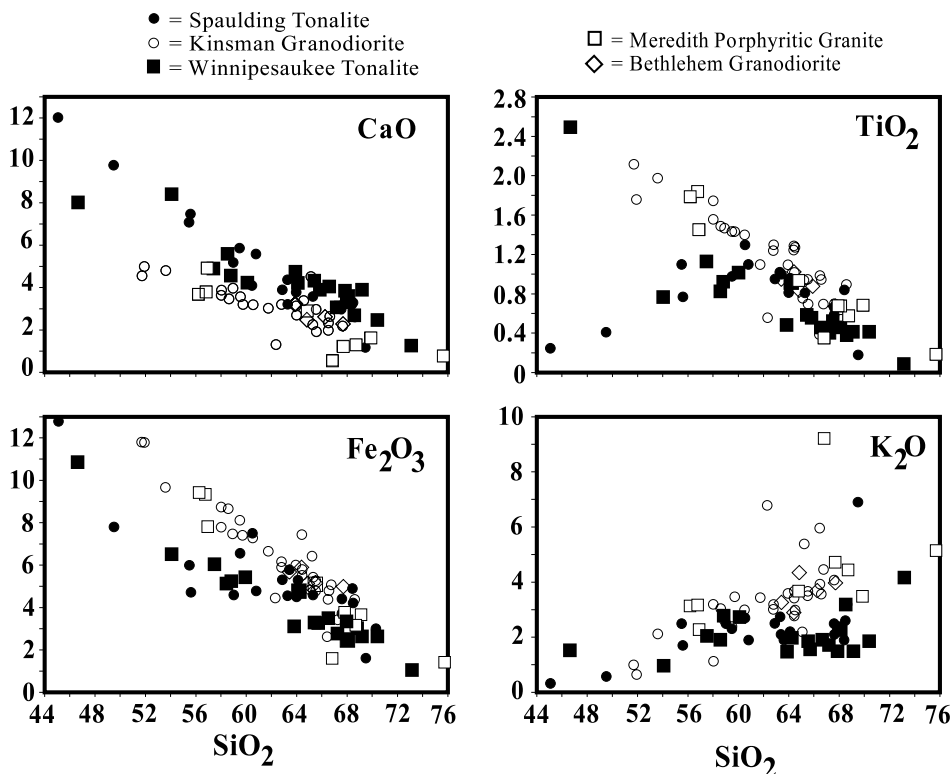


Fig. 7. Harker diagrams (wt. %) showing the compositions of the Winnepesaukee Tonalite, Meredith Porphyritic Granite, Kinsman Granodiorite of the Cardigan pluton (samples of Clark and Lyons, 1986), and Bethlehem Granodiorite. The Winnepesaukee Tonalite is richer in CaO and poorer in TiO₂, Fe₂O₃, and K₂O compared to the Meredith Porphyritic Granite, the Kinsman Granodiorite, and the Bethlehem Granodiorite, which have the same major element compositions.

This lack of compositional diversity may be a reflection of a paucity of analyses for the Bethlehem Granodiorite. Alternatively, plagioclase almost always shows zoning indicative of Rayleigh fractionation, hence the limited anorthite range may be a product of metamorphism following emplacement, which may have homogenized initial igneous zoning. Equilibration of muscovite with plagioclase of An₃₅ in the Bethlehem Granodiorite may have led to relatively paragonite-poor compositions compared to the more Na-rich muscovite in the Kinsman Granodiorite and Winnepesaukee Tonalite which may have equilibrated with more albitic plagioclase that crystallized from late-stage, residual liquids. The large span of plagioclase compositions in individual samples of both the Meredith Porphyritic Granite and Winnepesaukee Tonalite suggests typical igneous zoning, and that these rocks were not subjected to the same duration of metamorphism as the Bethlehem Granodiorite. We view this feature as supportive evidence that the Bethlehem Granodiorite is older than the Meredith Porphyritic Granite and other Kinsman equivalents, an interpretation that is compatible with P-T results and structural data of Spear (1992).

Source Rocks for the NHPS Magmas

Several researchers have considered specific NHPS plutons to represent essentially *in situ* mid-crustal melts of the CMT metasedimentary rocks (Harrison and others,

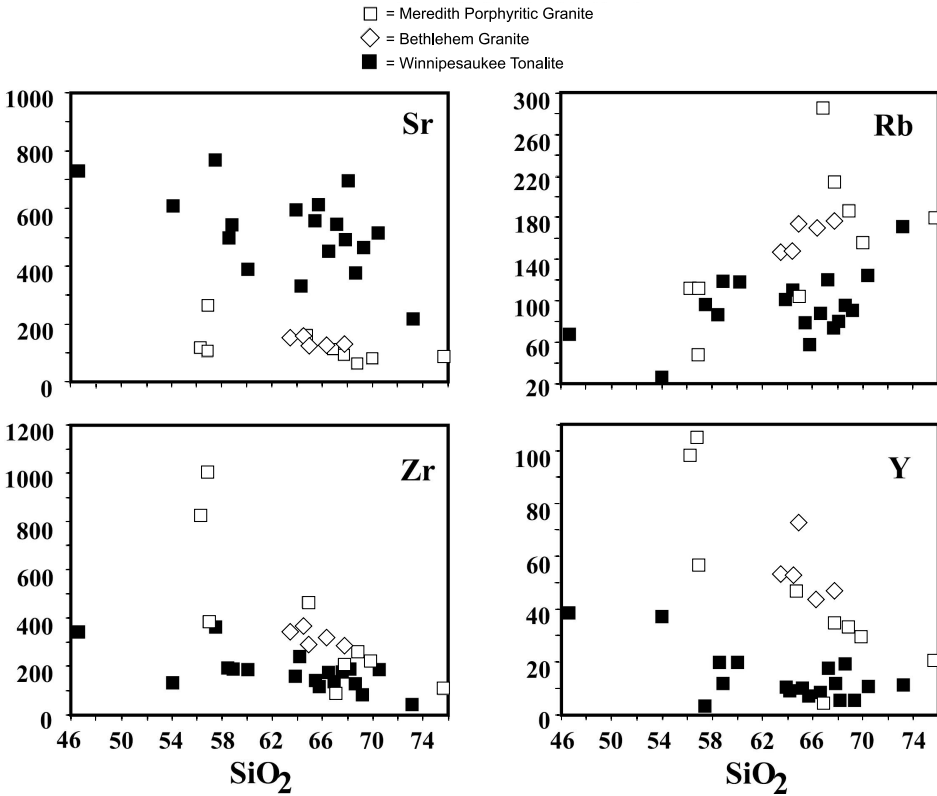


Fig. 8. Sr, Rb, Zr, and Y (ppm) versus SiO₂ (wt. %) showing the compositions of the Winnepesaukee Tonalite, Meredith Porphyritic Granite and Bethlehem Granodiorite. The Winnepesaukee Tonalite is richer in Sr and poorer in Rb, Zr, and Y than Meredith Porphyritic Granite and Bethlehem Granodiorite rocks of equivalent silica contents. The Meredith Porphyritic Granite and Bethlehem Granodiorite compositions overlap.

1987; Allen, 1992; Lathrop and others, 1994, 1996; Brown and Pressley, 1999; Brown and Solar, 1999). While these models are applicable to some NHPS plutons, they cannot account for the compositional diversity of the entire suite. For example, analyses of CMT metasedimentary rocks (Twyman, ms, 1983; Lathrop and others, 1994; and Dorais and Wintsch, unpublished data) are plotted in figure 14 along with Kinsman samples from Clark and Lyons (1986). All the metapelites are too CaO-poor to be Kinsman source rocks, Kinsman compositions require a metagraywacke/metapelite component such as the Hutchins Corner Formation (fig. 14). The metaluminous, hornblende-bearing Spaulding Tonalite is even richer in CaO than the Kinsman (fig. 7) and none of the CMT metasedimentary rocks can be the source of these Spaulding rocks as such rocks cannot produce metaluminous melts (for example, Patino Douce, 1999).

Inferences of the nature of the source rocks can be deduced from figure 15, which plots Winnepesaukee, Meredith, and Bethlehem samples in Al₂O₃ + FeO + MgO + TiO₂ versus Al₂O₃ / (FeO + MgO + TiO₂) (after Patino Douce, 1999). Also shown are fields for magmas derived from felsic pelites, mafic pelites and graywackes. Many of the Meredith and Bethlehem samples plot in the graywacke source field, but several

TABLE 7

Major and trace element and stable isotopic analyses, Winnepesaukee pluton, NH

	WQD-1	WQD-2	WQD-3	WQD-4	WQD-7	WQD-7B	WQD-8	WQD-9
SiO ₂	46.60	58.52	67.16	70.39	68.06	68.54	63.85	69.17
TiO ₂	2.50	0.83	0.41	0.42	0.48	0.39	0.49	0.42
Al ₂ O ₃	18.50	19.36	17.63	15.19	16.37	15.53	19.65	16.15
Fe ₂ O ₃	10.84	5.13	2.80	2.69	2.47	2.92	3.14	2.65
MnO	0.14	0.09	0.06	0.07	0.04	0.05	0.05	0.06
MgO	3.95	2.89	1.01	1.07	1.22	1.01	1.11	1.40
CaO	8.10	5.64	3.06	2.51	3.56	2.70	4.76	3.89
Na ₂ O	3.12	3.84	4.85	4.27	3.90	3.75	4.30	3.72
K ₂ O	1.53	1.94	1.70	1.87	2.29	3.20	1.49	1.50
P ₂ O ₅	1.10	0.24	0.07	0.10	0.15	0.14	0.15	0.18
Total	96.38	98.48	98.75	98.58	98.54	98.23	98.99	99.14
Large Cations								
Rb	69	87	121	125	80	94	101	92
Cs	2.6	2	6.8	7.1	5.1	4.7	5.6	6.1
Sr	734	498	545	519	700	381	600	469
Ba	308	194	193	76	1063	453	284	171
Rare Earth Elements								
La	48.4	6.5	43.3	62.3	50.2	20.5	21	6
Ce	104.1	18.4	85.5	113.4	86.1	46.6	41.7	11.1
Sm	7.83	3.9	5.07	6.22	3.07	3.83	3.06	1.41
Eu	2.08	1.17	1.28	1.38	1.02	1.01	1.44	0.88
Tb	0.95	0.53	0.51	0.51	0.2	0.57	0.31	0.16
Dy	6.47	2.69	1.4	1.59	2.06	3.14	1.54	0.97
Yb	2.8	1.76	1.33	0.74	0.5	0.9	0.59	0.47
Lu	0.42	0.24	0.25	0.13	0.05	0.15	0.07	0.07
(La/Lu) _N	11.6	2.5	21.7	56.7	67.6	15.3	24	8.6
High Valence Cations								
Zr	351	192	178	198	189	130	169	87
Y	39	20	18	11	6	19	11	6
Hf	9.1	5.6	4.6	5.4	12.7	12.5	4.8	4
Nb	8.4	2.4	6.6	0	0	0	5.5	2.3
Ta	1.9	0.59	0.94	0.81	5.4	4.1	0.22	0.53
Th	1.6	0.2	15.9	20.1	15.6	14.4	12.7	6.2
U	1.77	0.79	5.85	2.69	3.35	2.99	0.64	1.39
Ferromagnesian Elements								
Sc	19	20	6	5	5	6	5	5
Cr	4	16	5	5	17	7	4	9
Co	29	27	3	4	6	5	5	7
Ni	0.7	84	0	0	0	0	0	0
Zn	84	51	74	62	55	59	58	47
¹⁸ O	7.1	6.7	10.2				9.7	9.6
T _{zir}	770	776	796	809	796	766	785	735

TABLE 7
(continued)

	WQD-10	WQD-11	WQD-12	WQD-13	WQD-14	WQD-15	WQD-16	WQD-17
SiO ₂	73.15	67.79	65.40	57.43	54.04	64.21	59.97	58.81
TiO ₂	0.09	0.52	0.58	1.13	0.77	0.92	1.01	0.92
Al ₂ O ₃	14.68	15.79	16.64	18.37	18.05	17.12	19.23	19.76
Fe ₂ O ₃	1.10	3.25	3.29	6.04	6.49	4.77	5.41	5.23
MnO	0.03	0.04	0.04	0.06	0.18	0.07	0.08	0.08
MgO	0.22	1.62	1.83	2.44	5.61	1.81	2.05	2.25
CaO	1.29	3.82	4.27	4.94	8.41	4.08	4.23	4.57
Na ₂ O	4.05	3.63	3.60	3.79	3.07	3.78	4.51	4.46
K ₂ O	4.18	1.55	1.86	2.07	0.97	2.00	2.71	2.80
P ₂ O ₅	0.16	0.17	0.16	0.47	0.27	0.38	0.38	0.43
Total	98.79	98.18	97.67	96.74	97.86	99.14	99.58	99.31
Large Cations								
Rb	172	77	78	96	26	112	119	120
Cs	2	4.3	7	3.4	1.5	3.5	2.6	3.1
Sr	213	496	558	771	610	334	392	546
Ba	224	296	256	725	41	361	462	422
Rare Earth Elements								
La	5.6	18.8	11.4	18.1	24.6	27.5	16	15.5
Ce	11.7	38.3	23	35.7	65.9	57.1	34.6	34.4
Sm	1.33	1.98	1.45	2.02	6.91	4.57	3.81	4.75
Eu	0.34	0.98	1.1	1.24	2.07	1.61	1.55	1.2
Tb	0.23	0.15	0.14	0.08	0.87	0.56	0.34	0.33
Dy	2.32	2.15	2.02	2.18	4.78	2.03	2.09	1.25
Yb	0.7	0.51	0.3	0.95	2.53	1.05	0.99	0.97
Lu	0.12	0.08	0.05	0.14	0.43	0.17	0.15	0.16
(La/Lu) _N	5.4	24.8	25.6	12.8	6.6	17.6	10.9	10.8
High Valence Cations								
Zr	50	173	147	363	134	248	187	190
Y	11	13	9	3	38	10	20	12
Hf	1.7	4.8	3.1	8.6	4.1	6.7	5.4	5.1
Nb	0	3.9	0	0	0	9.2	12.5	11.4
Ta	0.5	0.55	0.47	0.36	0.66	0.88	1.11	0.85
Th	1.8	2.7	0.4	2.3	3.3	1.5	0.4	0.4
U	1.61	2.15	0.84	1.52	1.91	1.41	0.9	1.3
Ferromagnesian Elements								
Sc	3	8	6	4	27	11	13	11
Cr		11	12	17	144	138	15	17
Co	1	8	8	11	7	10	11	12
Ni	0	0	0	82	59	3	0	2
Zn	58	37	37	73	76	71	59	68
¹⁸ O	10.5		8.1	7.8	7.4			
T _{zir}	697	790	771	837	709	816	779	777

TABLE 7
(continued)

	WQD-18	WQD-19	MPG-2	MPG-3	MPG-4	MPG-6	MPG-7	MPG-10
SiO ₂	66.56	65.73	75.63	66.78	64.73	67.61	69.83	56.86
TiO ₂	0.46	0.56	0.19	0.36	0.90	0.57	0.69	1.46
Al ₂ O ₃	17.40	16.78	12.89	16.58	16.40	15.74	14.05	18.42
Fe ₂ O ₃	3.49	3.32	1.44	1.63	5.10	3.40	3.70	7.83
MnO	0.04	0.03	0.04	0.02	0.10	0.04	0.06	0.14
MgO	1.49	1.48	0.25	0.46	1.35	0.77	0.96	2.55
CaO	4.05	3.92	0.83	0.58	3.06	1.26	1.64	4.98
Na ₂ O	4.10	4.29	2.61	2.06	3.40	3.08	2.84	4.34
K ₂ O	1.91	1.61	5.18	9.24	3.69	4.75	3.48	2.25
P ₂ O ₅	0.19	0.13	0.16	0.21	0.14	0.26	0.23	0.60
Total	99.69	97.85	99.22	97.92	98.87	97.22	97.48	99.43
Large Cations								
Rb	87	59	182	287	106	215	157	50
Cs		1.5	2		1.4	3.1	2.5	
Sr	458	619	89	113	160	96	89	269
Ba	123	256	264	501	395	373	220	370
Rare Earth Elements								
La		7.7	17.6		93	42.6	52.1	
Ce		14.6	40.9		211.3	100.4	112.5	
Sm		0.97	3.59		18.22	8.56	10.6	
Eu		0.98	0.97		2.2	1.38	1.23	
Tb		0.16	0.61		1.86	1.01	1.04	
Dy		1.17	3.09		8.45	6.04	5.08	
Yb		0.29	1.75		5.12	2.35	1.49	
Lu		0.06	0.24		0.73	0.36	0.21	
(La/Lu) _N		17.9	6.8		12.2	12.2	23.6	
High Valence Cations								
Zr	146	125	112	104	468	206	228	383
Y	8	8	21	5	47	35	30	57
Hf		0.91	2.62		2.22	2.9	3	
Nb	5.1	2.8	5.9	11.9	20.4	18	17.6	23.6
Ta		0.31	0.66		1.6	1.5	1.2	
Th		0.51	9		43.8	17.3	20	
U		0.91	2.62		2.22	2.9	3	
Ferromagnesian Elements								
Sc		10	5		17	9	7	
Cr		8	5	1	16	16	19	14
Co		7	2		8	6	6	
Ni	0	0	73	2	0	5	5	11
Zn	35	35	34	47	76	71	76	105
¹⁸ O			11.7				11.8	
T _{zir}	771	756	768	752	879	822	834	827

TABLE 7
(continued)

	MPG-13	MPG-14	MPG-17	B-2	B-3	B-7	B-8	B-9
SiO ₂	56.75	56.16	68.73	67.70	66.28	63.42	64.84	64.39
TiO ₂	1.84	1.79	0.58	0.76	0.88	0.95	0.84	1.02
Al ₂ O ₃	17.44	17.26	14.99	15.54	15.68	16.18	15.43	16.23
Fe ₂ O ₃	9.38	9.46	3.20	4.53	5.15	5.65	5.12	5.89
MnO	0.16	0.16	0.05	0.08	0.08	0.08	0.07	0.08
MgO	2.06	2.03	0.71	1.31	1.54	1.60	1.38	1.72
CaO	3.86	3.70	1.31	2.28	2.62	3.19	2.47	3.21
Na ₂ O	3.96	3.96	3.77	2.87	2.87	3.18	2.77	3.17
K ₂ O	3.18	3.14	4.48	3.98	3.70	3.26	4.35	2.90
P ₂ O ₅	0.73	0.68	0.11	0.19	0.22	0.22	0.22	0.21
Total	99.36	98.34	97.82	99.05	99.02	97.73	97.49	98.82
Large Cations								
Rb	113	113	188	177	171	147	174	148
Cs	3.5	3.7	4.9	9.6	7.7	6.4	8.4	6.8
Sr	113	123	60	131	131	154	129	161
Ba	361	391	232	667	606	510	669	576
Rare Earth Elements								
La	46.5	53.2	42.7	47	41.6	62.9	50.3	47.1
Ce	106.9	122.6	95.3	105.2	96.1	132.2	101.6	103.5
Sm	16.36	16.96	9.72	9.02	8.68	12.01	10.04	9.35
Eu	2.27	2.5	1.04	1.52	1.52	2.03	1.74	1.83
Tb	3.08	3.1	1.05	1.08	1.26	1.58	1.49	1.46
Dy	20.3	5.36	17.54	6.54	6.18	8.73	9.36	8.58
Yb	9.43	9.24	3.3	4.25	3.99	4.12	4.62	4.88
Lu	1.29	1.23	0.44	0.62	0.58	0.58	0.65	0.67
(La/Lu) _N	3.3	3.9	8.7	7.5	7	10.3	7.3	6.5
High Valence Cations								
Zr	1010	833	264	289	323	346	292	373
Y	106	99	33	47	44	53	73	53
Hf	27.1	25.1	6.9	8	9.1	10.8	9.3	11
Nb	31	34	19	15	16	17	16	19
Ta	2.8	3.2	1.4	1.1	1	1	1	1.4
Th	6.4	9.3	21.1	15.8	17.4	19.5	16.2	19.8
U	1.5	2	4.6	3.2	3.3	2.5	3.8	2.9
Ferromagnesian Elements								
Sc	24	24	7	13	15	16	14	18
Cr	6	7	9	34	35	40	32	39
Co	15	18	6	9	10	12	10	13
Ni	18	15	4	16	15	20	27	23
Zn	150	145	79	73	78	84	77	83
¹⁸ O		9.5	11	10.2			9.9	9.8
T _{zir}	934	914	834	847	854	851	839	863

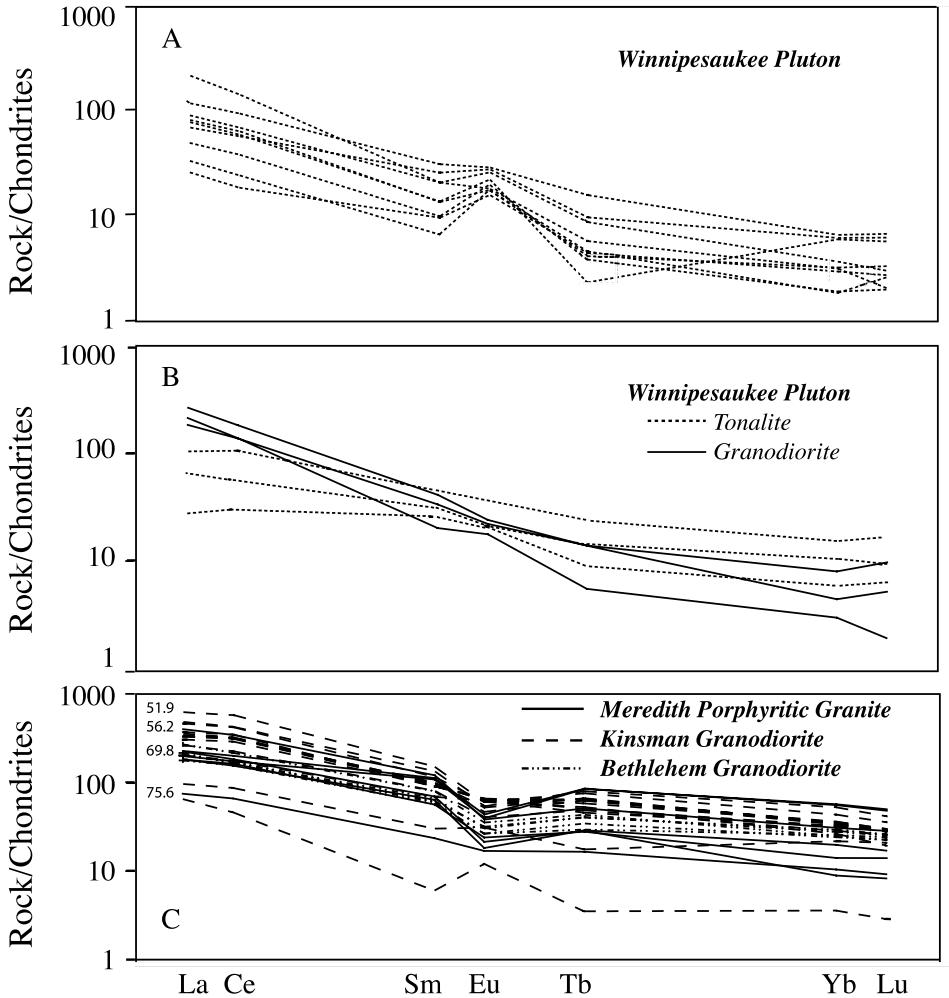


Fig. 9. A) Chondrite-normalized REE patterns (after Anders and Ebihara, 1982) for selected Winnepesaukee Tonalite samples. These samples have positive Eu anomalies indicating plagioclase accumulation. B) Chondrite-normalized REE patterns for selected Winnepesaukee Tonalite samples that lack positive Eu anomalies. The tonalites are LREE poor and HREE rich compared to the granodiorites. C) Chondrite-normalized REE patterns for Meredith Porphyritic Granite, Kinsman Granodiorite (Cardigan pluton samples from Clark and Lyons, 1986), and Bethlehem Granodiorite samples. Total REE abundances decrease with increasing silica contents (selected silica contents listed to the left of the patterns). With the exception of two silica-rich Kinsman samples and a Meredith Porphyritic Granites sample, the patterns are similar for all three rock units.

Winnepesaukee and Meredith samples plot at $\text{Al}_2\text{O}_3 + \text{FeO} + \text{MgO} + \text{TiO}_2$ values beyond those of graywacke partial melts requiring more primitive source rocks.

Many SiO_2 -rich Winnepesaukee Tonalite, Meredith Porphyritic Granite, and Bethlehem Granodiorite samples have elevated $\delta^{18}\text{O}$ values (fig. 11) indicating a dominant sedimentary component to the melts. Kinsman and Bethlehem Granodiorite samples of Lathrop and others (1996) plot within the CMT metasedimentary field (fig. 12); thus there is no question that there is a major contribution from the CMT metasedimentary rocks as proposed by many researchers (Lathrop and others, 1994, 1996; Brown and Solar, 1999; Brown and Pressley, 1999). However, the majority of

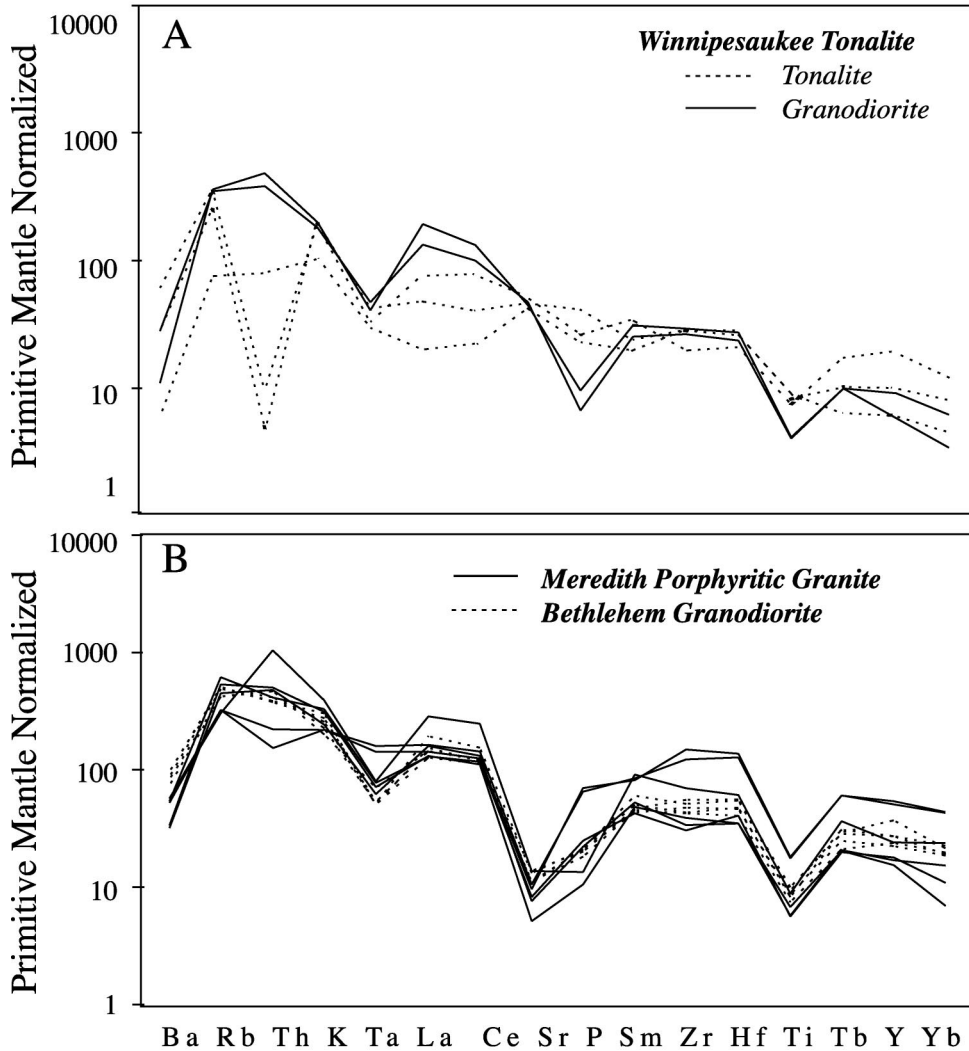


Fig. 10. A) Normalized multi-element diagram (after Thompson and others, 1984) for the samples that lack positive Eu anomalies plotted in figure 9B. The tonalites are poorer in Rb, Th, and LREE but richer in P, and HREE compared to the granodiorites. B) Normalized multi-element diagram of Meredith Porphyritic Granite and Bethlehem Granodiorite samples. The Meredith Porphyritic Granite shows a larger range in trace element compositions compared to our limited number of Bethlehem Granodiorite samples, but several samples of both rock units are similar in composition.

CMT metapelites have values between 10 to 19 (Chamberlain and Rumble, 1988; Lathrop and others, 1994, 1996), which along with their CaO and Al₂O₃ + FeO + MgO + TiO₂ compositions discussed above, indicate that metapelites cannot be the exclusive source of the NHPS magmas. A few metapelite samples have δ¹⁸O values as low as 7 to 8 (Lathrop and others, 1994, 1996), but are also too CaO-poor to serve as direct sources of the magmas. No oxygen isotopic data are available for the Hutchins Corner Formation, but metagraywackes such as these are a more plausible source rock for the NHPS crustal component.

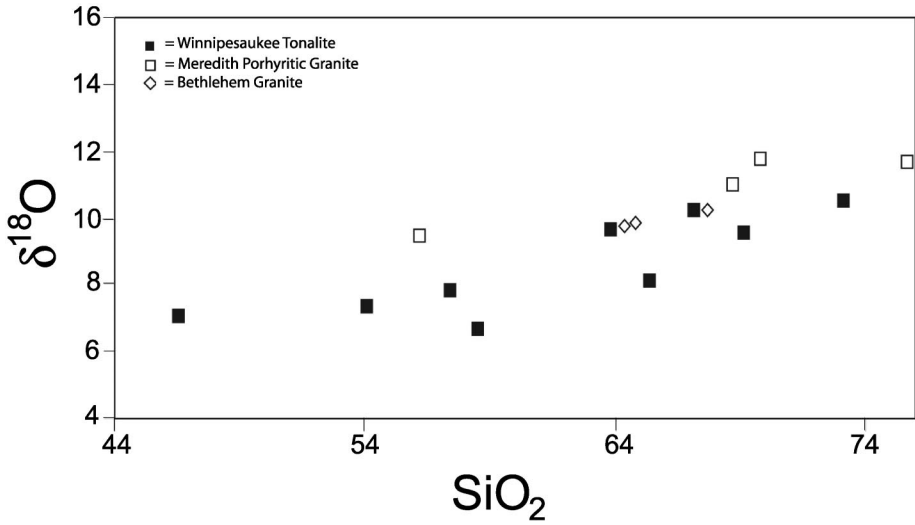


Fig. 11. $\delta^{18}\text{O}$ versus SiO_2 contents for Winnepesaukee Tonalite, Meredith Porphyritic Granite and Bethlehem Granodiorite samples.

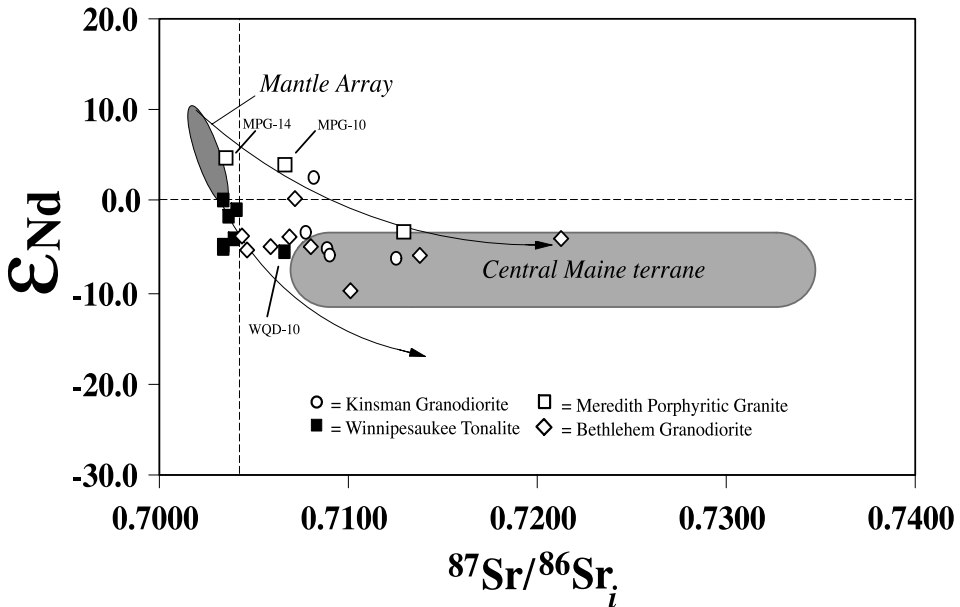


Fig. 12. ϵ_{Nd} (@ 400 Ma) versus $^{87}\text{Sr}/^{86}\text{Sr}_i$ for Winnepesaukee Tonalite samples, Meredith Porphyritic Granite enclave samples, Kinsman Granodiorite and Bethlehem Granodiorite (Kinsman and Bethlehem samples from Lathrop and others, 1994, 1996). Several Winnepesaukee and Meredith samples plot on or near the mantle array, indicating a large mantle component to the Winnepesaukee pluton. Other Winnepesaukee samples and those of the Kinsman and Bethlehem contain a significant CMT component (CMT field from Lathrop and others, 1996).

TABLE 8
Nd and Sr isotopic compositions, WQD and MPG mafic enclaves

Sample	Sm(ppm)	Nd(ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_{\text{cor}}$	error	E_{Nd_0}	E_{Nd} (400 Ma)	T_{DM}
WQD-1	6.75	37.28	0.1095	0.512414	10	-4.4	0.1	927
WQD-2	5.01	17.33	0.1747	0.512340	19	-5.8	-4.7	2809
WQD-9	2.41	10.56	0.1381	0.512211	19	-8.3	-5.3	1674
WQD-10	1.76	7.08	0.1503	0.512241	15	-7.7	-5.4	1924
WQD-13	4.51	28.62	0.0953	0.512292	11	-6.7	-1.6	972
WQD-14	7.33	34.35	0.1289	0.512248	15	-7.6	-4.1	1426
MPG-10	7.84	33.38	0.1420	0.512363	10	-5.4	4.7	1441
MPG-14	23.94	99.09	0.1461	0.512390	10	-4.8	5.2	1469

	Rb ppm	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	error	$^{87}\text{Sr}/^{86}\text{Sr}_i$
WQD-1	69	734	0.272	0.704968	16	0.703419
WQD-2	87	498	0.505	0.706251	17	0.703374
WQD-9	92	469	0.568	0.706557	20	0.703322
WQD-10	172	213	2.336	0.719934	13	0.706628
WQD-13	96	771	0.360	0.705764	14	0.703713
WQD-14	26	610	0.123	0.704656	16	0.703955
MPG-10	50	269	0.538	0.709898	15	0.706833
MPG-14	113	123	2.474	0.717577	17	0.703485

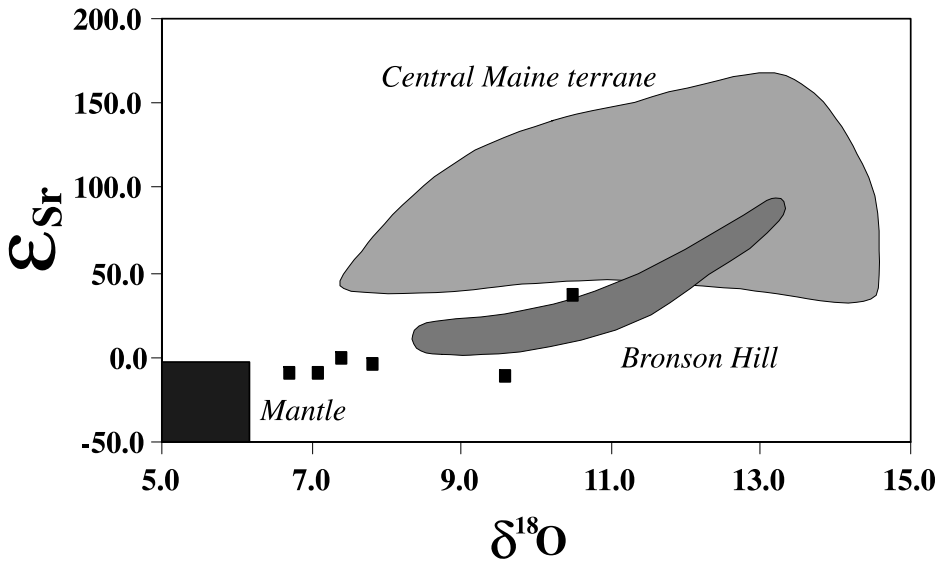


Fig. 13. ϵ_{Sr} (@ 400 Ma) versus $\delta^{18}\text{O}$ for Winnepesaukee Tonalite samples. The more mafic Winnepesaukee Tonalite samples plot outside the fields defined for the CMT and Bronson Hill terranes (after Lathrop and others, 1996) and show a dominant mantle/lower crustal component.

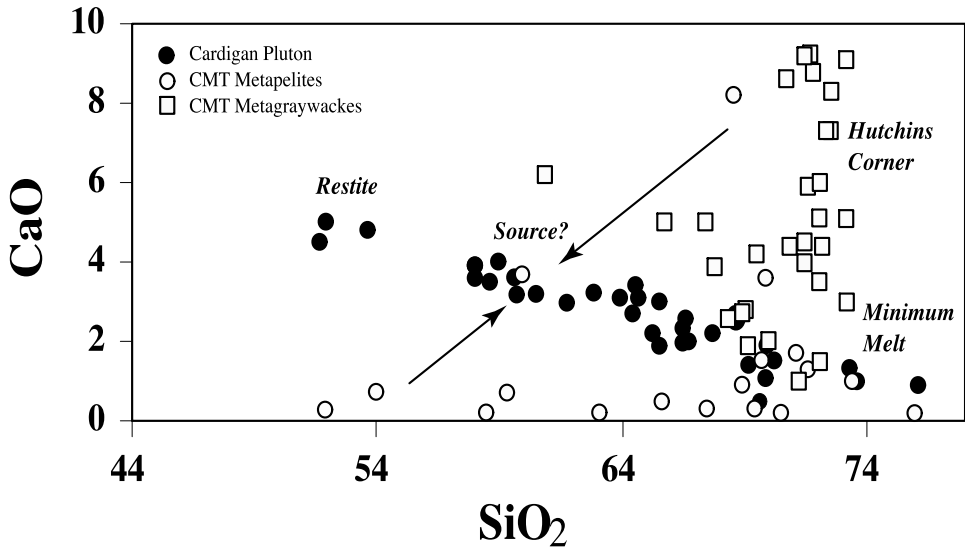


Fig. 14. CaO versus SiO₂ (wt. %) diagram for Kinsman Granodiorite (Cardigan pluton samples from Clark and Lyons, 1986) and metasedimentary rocks of the Central Maine terrane. (Lathrop and others, 1994; Ferry, 1982, 1983, 1988; Twyman, 1983; and Dorais and Wintsch, unpublished data). The garnet-rich rocks of the Kinsman Granodiorite plot along the SiO₂-poor extension of the Kinsman rocks, the trend of which was interpreted by Clark and Lyons (1986) to represent restite unmixing. Potential source rocks should plot along the trend (White and Chappell, 1977). No Central Maine terrane metapelites are rich enough in CaO to have served as source rocks for the Kinsman Granodiorite. The CaO-rich Hutchins Corner Formation mixed with metapelites could be source rocks.

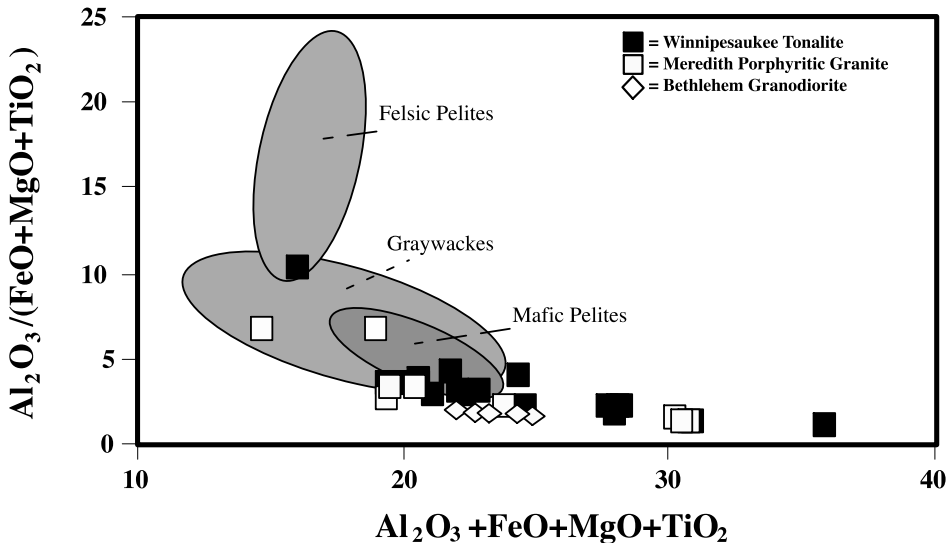


Fig. 15. Al₂O₃+FeO+MgO+TiO₂ versus Al₂O₃/(FeO+MgO+TiO₂) (wt. %) showing the fields of granitic melts derived from felsic pelites, mafic pelites, and graywackes (after Patino Douce). Some of the NHPS rocks are consistent with partial melts of graywackes or mafic pelites, but many are richer in Al₂O₃+FeO+MgO+TiO₂ than melts from these metasedimentary rocks.

Other more mafic Winnepesaukee samples and Meredith enclaves (MPG-10, MPG-14) have $\delta^{18}\text{O}$ between 6 and 8, ϵ_{Nd} (at 400 Ma) values from 0.1 to 5.2, and initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7035 (table 8), indicating a mantle or lower crustal origin (fig. 13). Precambrian T_{DM} ages (table 8) indicate either contamination of mantle-derived magmas or, given the metaluminous compositions, that the source rocks were lower crustal amphibolites. If the latter is the origin of the mafic NHPS, amphibolite dehydration melting requires temperatures of greater than 1000°C (Beard and Lofgren, 1991; Wolf and Wyllie, 1991), temperatures that exceed all the closed system models discussed above.

Geothermometry and Geobarometry

The Spaulding Tonalite is the youngest of the NHPS syntectonic members and as such, the degree of deformation is considerably less than the Bethlehem Granodiorite. Samples chosen for thermobarometric study are from portions of the Winnepesaukee pluton that preserve igneous textures and have no evidence of shearing or metamorphism.

Several geothermometers were used to estimate the temperatures of the NHPS magmas. The Winnepesaukee Tonalite contains hornblende, plagioclase and Fe-Ti oxides. The amphibole - plagioclase geothermometer of Holland and Blundy (1994) gives 770°C for the quartz dioritic sample WQD-1 for adjacent plagioclase - hornblende rims. Fe-Ti oxide pairs in the form of magnetite exsolution lamellae in ilmenite give temperatures as high as 800°C at the NNO buffer for this sample (Anderson and others, 1993). Hornblende - plagioclase pairs in tonalitic sample WQD-2 give temperatures of 715°C.

Assuming that garnet cores and isolated matrix biotite grains preserved magmatic compositions, Plank (1987) used the biotite - garnet geothermometer (Ferry and Spear, 1978) to determine the temperatures of the Cardigan pluton of the Kinsman Granodiorite. The margins of the pluton cooled the quickest, preserving a record of temperatures as high as 850° to 900°C. The interior of the pluton cooled slower, giving lower temperatures of 750° to 800°C. Matrix biotite and garnet cores of interior samples of the Meredith Porphyritic Granite give temperatures of 720° to 730°C (Holdaway and others, 1997).

No mineral geothermometers are available that preserve magmatic temperatures for the Bethlehem Granodiorite. However, Thompson and others (1968) suggest that the Bethlehem magma is a lower temperature, more hydrous, Kinsman equivalent. The presence of muscovite in the Bethlehem versus garnet in the Kinsman can be explained as a function of temperature and $f_{\text{H}_2\text{O}}$ according to the reaction: $\text{musc} + \text{bio} + 3\text{qtz} = 2\text{Kspar} + \text{gar} + 2\text{H}_2\text{O}$ where the low temperature assemblage is characteristic of the Bethlehem Granodiorite and the high temperature, garnet-bearing assemblage represents the Kinsman magmas.

Metamorphic pressures of 3 to 4 Kbars are recorded in the metasedimentary rocks of the CMT of central New Hampshire (Chamberlain and Lyons, 1983; Day and Chamberlain, 1989). An aluminosilicate triple point isobar has been mapped through New Hampshire (Thompson and Norton, 1968; Hodges and Spear, 1982), passing through the Meredith Porphyritic Granite along the western margin of the Winnepesaukee pluton (Lyons and others, 1997). These P-T conditions were reached during the Acadian orogeny concomitantly with the intrusion of the syntectonic members of the NHPS and are considered to be the emplacement pressures of the plutons.

Amphibole thermobarometry gives a wide range of pressures for the Winnepesaukee Tonalite. For rocks containing the appropriate mineral assemblage, the Al-in-hornblende geobarometer of Johnson and Rutherford (1989) gives near emplacement pressures of 4 Kbars for many grains, others yield pressures exceeding 5 Kbars and a few greater than 6 Kbars. The Schmidt geobarometer (1992) gives pressure of 6.5

Kbars for many grains with a few as high as 7.8 Kbars. Using temperatures obtained from adjacent hornblende - plagioclase rims (Holland and Blundy, 1994), the Anderson and Smith (1995) thermobarometer gives pressures as high as 6.5 Kbars. The empirical thermobarometer of Ernst and Liu (1998) give pressures as low as 4 Kbars for some grains with many others giving pressures greater than 6 Kbars. These higher pressures obtained by all the amphibole geobarometers are greater than the pluton's emplacement pressure, suggesting polybaric amphibole crystallization.

The presence of magmatic epidote in the Winnepesaukee Tonalite places some qualitative constraints on the higher pressure, pre-emplacement history of the pluton. In general, magmatic epidote is thought to require relatively high pressures for crystallization (Zen and Hammarstrom, 1984). In the presence of quartz, the epidote-out reaction curve has a positive slope (Liou, 1973) that intersects the solidus to give an estimation of minimum pressure of crystallization. Schmidt and Thompson (1996) demonstrated that high f_{O_2} increases the pistacite molecule $\{\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al}^{\text{VI}})\}$ in epidote, thereby increasing its thermal stability such that the epidote-out reaction intersects the solidus at pressures as low as 3 Kbars. In assemblages buffered at NNO, epidote contains ~25 percent pistacite and the intersection of the tonalite solidus and epidote stability field occurs at 5 to 6 Kbars.

If the amphibole - plagioclase temperatures of 770°C and the Fe-Ti oxide temperatures of 800°C at the NNO buffer reflect solidus temperatures at the emplacement pressure of 3.8 Kbars for sample WQD-1, the intersection of the epidote stability curve for epidote with 25 percent pistacite molecule (table 5) and the solidus occurs at approximately 7 Kbars (fig. 16). A large error is associated with the intersection of the two curves because the solidus temperature is strongly dependent on the H₂O contents of the magma. A tonalite solidus at ~800° and 3.8 Kbars clearly would have to have been H₂O undersaturated (Johannes and Holtz, 1996), a condition that is difficult to verify. This inferred intersection pressure is considerably higher than the emplacement pressure but matches the 6 to 7 Kbars pressure derived from some hornblende grains. Sample WQD-2, with an amphibole - plagioclase temperature of 715°C, gives an epidote-out curve and solidus intersection at ~6 Kbars, matching the epidote stability experimental results of Schmidt and Thompson (1996) and the higher pressure range of amphibole thermobarometry for this sample. These pressure estimates exceed that of emplacement and suggest that epidote is intratelluric, owing its ability to survive resorption at lower pressures because it was armored by hornblende and/or biotite (Brandon and others, 1996). Portions of epidote grains adjacent to tectosilicates were not armored and display resorption textures. The armored epidote displays internal idiomorphic zoning, which is truncated at the resorbed edges of grains. The truncation of internal zoning indicates an average of 200 um of epidote dissolution occurred in the unarmored epidote.

An additional constraint on the pressure of differentiation of the Winnepesaukee pluton is displayed in figure 17, which plots CaO+Al₂O₃ versus CaO/Al₂O₃ contents. Patino Douce (1999) determined melt compositions produced by experimental hybridization of a high-Al tholeiite with graywackes, the dominant sedimentary component of the Winnepesaukee pluton, as a function of pressure. The curves represent high pressure (~12-15 Kbars) and low pressure (<5 Kbars) differentiation with Winnepesaukee samples plotting along the high pressure trend.

The combination of amphibole geobarometry, magmatic epidote, and congruity between bulk-rock compositions and experimentally determined granitic magma compositions at moderate pressures (Patino Douce, 1999) suggests that the Winnepesaukee Tonalite preserves a record of pressures \geq 6 Kbars. However, irrespective of geobarometric constraints, the metaluminous compositions of the Winnepesaukee Tonalite require a source below the metapelitic to metagraywacke CMT.

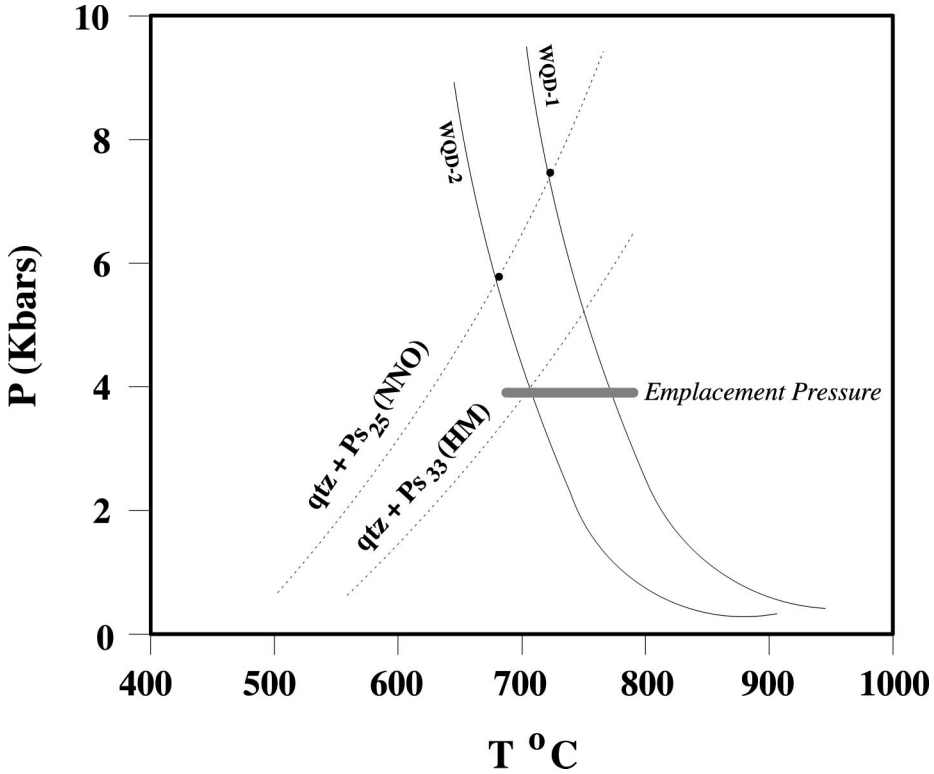


Fig. 16. P-T diagram showing the epidote-out curves at pistacite contents of 25 and 33 percent and f_{O_2} at the NNO and HM buffers. These two curves intersect the estimates solidi of Winnepesaukee Tonalite samples WQD-1 and WQD-2 at ~ 7 and ~ 6 Kbars respectively. See text for discussion.

Review of Thermal Modeling

Jameson and others (1998), Engi and others (2001) and Roselle and others (2002) modeled the effects of tectonically accreted radioactive material (TARM) on subduction zone metamorphism and anatexis. These studies indicate that accreted sediments enriched in heat-producing elements and located in an accretionary wedge or a tectonically thickened continental margin sequence lack the necessary heat to produce significant anatexis and granitic plutonism in orogens. In contrast, partial subduction of upper crustal materials to the uppermost mantle produces lower crustal temperatures as high as 800° to 900°C , within the range of temperatures required to melt the inferred source rocks of this study. These temperatures are reached only in the lower crust, far deeper than the depths to which the metasedimentary rocks of the Central Maine terrane were buried.

If the metaluminous Spaulding magmas were differentiates from mantle derived melts rather than partial melts of lower crustal amphibolites, then the redistribution of mass and energy would have exceeded that proposed by the subducted TARM models. Others have presented thermal balance calculations to determine the limits of crustal contamination of mantle-derived melts and anatexis (Reiners and others, 1995; Knesel and Davidson, 1998). Such calculations for the NHPS would be entirely speculative as there are no constraints on the masses of crustal rock, crustal melt, crystals or basalt required to produce the lower crustal component of the NHPS. However, any input of mantle-derived magmas would have caused even larger amounts of lower crustal

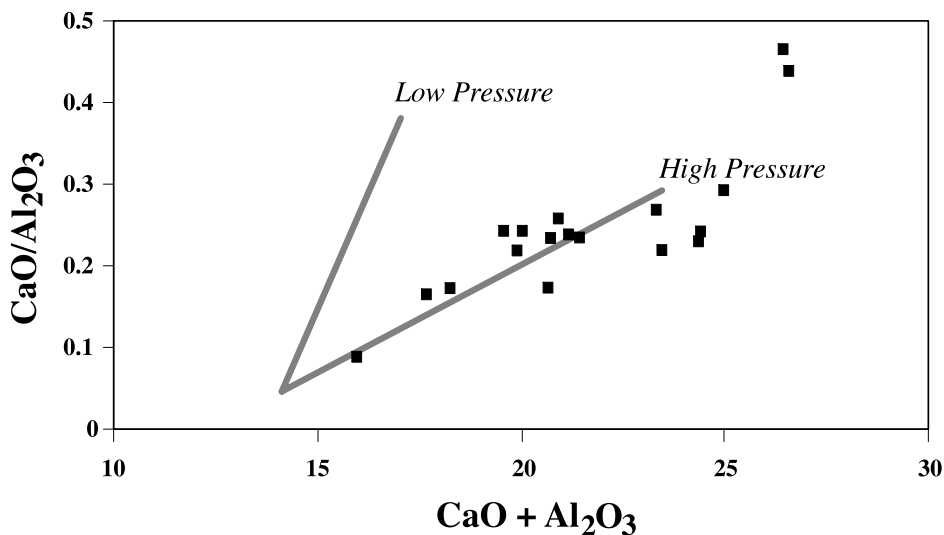


Fig. 17. $\text{CaO} + \text{Al}_2\text{O}_3$ versus $\text{CaO}/\text{Al}_2\text{O}_3$ (wt. %) showing differentiation trends of tholeiitic basalts mixed with graywackes at high ($\sim 12\text{--}15$ Kbars) and low (< 5 Kbars) pressures (after Patino Douce, 1999). The Winnepesaukee samples plot along the high pressure differentiation trend.

anatexis than those predicted by the subducted TARM models. Thus thermal models support the geochemical and isotopic data presented above that the Acadian orogeny involved redistribution of mass and energy transfer from the lower to midcrust.

Structure and Emplacement of the NHPS Magmas

One plausible origin of Acadian magmatism is that some NHPS magmas were generated by shear heating along the thrust decollement as proposed for the Philips pluton (Brown and Solar, 1999). This model of shear heating and partial melting has also been proposed by Deniel and others (1987) and by France-Lanord and Le Fort (1988) for the origin of the Manaslu Leucogranite of the High Himalayas, and has been adapted by Barbarin (1996) as the characteristic melting mechanism for two-mica granites. Following the model proposed by Brown and Pressley (1999), we infer that similar, shear heating and partial melting of metagraywackes may have produced the earliest Bethlehem magmas of the Acadian orogeny in New Hampshire. These low temperature Bethlehem magmas may have migrated along the decollement and up the dorsal zone and were carried by thrust sheets to the far western portions of the state as proposed by Spear (1992; fig. 18A).

The Kinsman rocks preserve high temperatures for felsic plutonic rocks, up to $\sim 850^\circ\text{C}$ (Plank, 1987). Mafic enclaves with mantle and/or primitive lower crustal isotopic signatures (figs. 12 and 13) in the Meredith Porphyritic Granite phase of the Kinsman Granodiorite demonstrate a heat source external to the CMT, one that contributed to the high magmatic temperatures of the Kinsman magmas. Ponding of mantle-derived, mafic magmas at depth could have caused partial melting of deeper seated metagraywackes, thereby generating magmas that are too CaO-rich and isotopically too light to be exclusively melts of the CMT metapelitic rocks. These magmas followed the Bethlehem up the dorsal zone and were carried with thrust nappes to the west (Spear, 1992; fig. 18B). Density filtering prevented the mafic magmas from ascending up the dorsal zone; only minor amounts of mafic enclaves are present in the Meredith Porphyritic Granite.

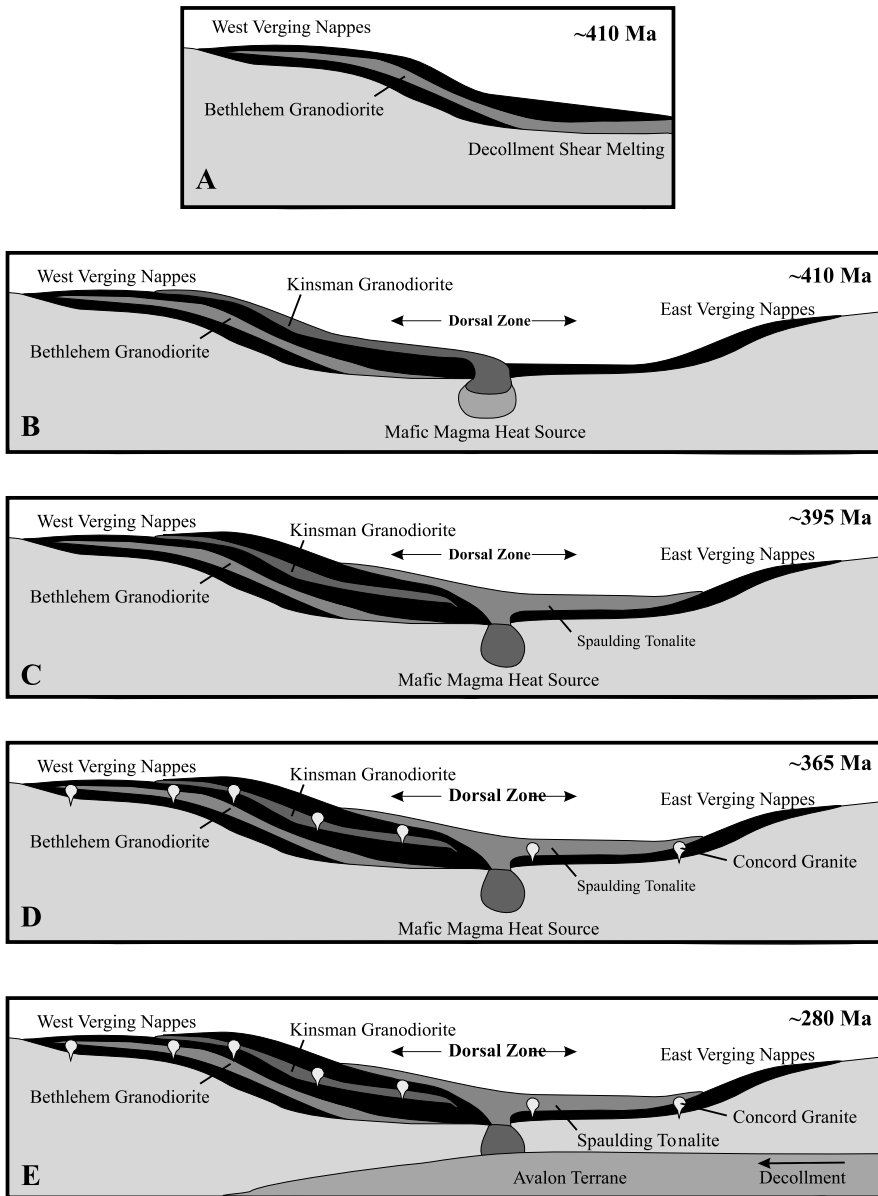


Fig. 18. Schematic cross sections depicting the emplacement of the NHPS (after Spear, 1992; Eusden and Lyons, 1993). A) The Bethlehem Granodiorite magmas, perhaps generated by shear heating along the decollement (Brown and Solar, 1999), ascended up the dorsal zone and were carried along thrust sheets (shown in black) to the western portions of the state. B) The Kinsman Granodiorite contains mafic enclaves with primitive isotopic signatures, indicative of a mantle/lower crustal heat source for the Kinsman. These magmas followed the Bethlehem up the dorsal zone and were carried to the west, now occupying intermediate locations between the Bethlehem and the dorsal zone. C) The last of the syntectonic units, the Spaulding Tonalite, followed its NHPS precursors, and was carried away from the dorsal zone by both east and west verging nappes. The last magmas up the dorsal zone are the metaluminous tonalites and quartz diorites with primitive isotopic characteristics, suggesting that density filtering inhibited emplacement of the mafic heat source for NHPS magmatism. D) Approximately 50 My after thrusting, high temperature isotherms swept upwardly through the depressed crustal rocks causing anatexis. These Concord Granite plutons were emplaced across the state without the structural controls of the syntectonic NHPS members. E) At approximately 280 Ma, the basement was underthrust during the Alleghanian orogeny, thereby detaching the roots of the NHPS plutons (Wintsch and others, 1992; Dorais and others, 2001).

Amphibole geobarometry and the presence of magmatic epidote in the Winnepesaukee Tonalite yield information about the pre-emplacment history and transportation mechanisms of the Spaulding magmas. Brandon and others (1996) experimentally determined the dissolution rate of epidote in granodioritic and tonalitic melts at pressures below the stability field of epidote. Given that epidote is stable only at high pressures but is found as armored grains in shallow level plutons, Brandon and others suggested that the presence of magmatic epidote in shallow level plutons requires rapid ascent of magma to prevent epidote dissolution, thereby preserving evidence of an earlier, higher pressure crystallization history.

These experimental results can be used to estimate the ascent rate of the Winnepesaukee Tonalite and implications of emplacement mechanisms. Brandon and others (1996) determined the rate of epidote dissolution and showed that in tonalitic magmas at low pressures and 800°C, 200 μm of epidote will dissolve in 2 years. Our estimates of the pressures of epidote crystallization in the Winnepesaukee Tonalite range from 6 to 7 Kbars, pressures, which are supported by amphibole geobarometry. The maximum pressure of 7 Kbars gives a difference between initial crystallization and emplacement pressures as 3 Kbars, which equates to a transportation distance of approximately 10 kilometers. If the epidote rims required approximately 2 years to undergo the observed degree of resorption as the magma ascended from 7 to 4 Kbars, then the ascent rate was 5000 m/yr. A minimum pressure estimate of 6 Kbars or 2 Kbars deeper than emplacement pressures gives an ascent rate of 3300 m/yr. These calculations do not consider any epidote resorption that may have occurred after emplacement of the pluton prior to solidification and should be viewed as maximum ascent rates, but sample WQD-1 was taken near the pluton's margin, which should have minimized the time for post-emplacment solidification.

These ascent rates are well within the range of models of magma transport by fracture propagation and dike networking (Clemens and Mawer, 1992; Petford and others, 1993; Petford, 1995). Regardless of the imprecision of the calculations, the range of ascent rates is very fast compared to estimated ascent rates of ~ 0.3 m/yr for diapiric emplacement. At first approach, such a model may seem implausible because dike transport *sensu stricto* may be occluded in contracting orogenies, but diking *sensu lato* is present in transpressive orogens such as the northern Appalachians during the Acadian (van der Pluijm and van Staal, 1988; Swanson, 1992; de Roo and van Staal, 1994; van Staal and de Roo, 1995; van Staal and others, 1998; Brown and Solar, 1999).

According to Eusden and Lyons (1993), the allochthonous CMT sequence may be as thin as 3 kilometers. As the current erosional level of the CMT in central New Hampshire records pressures of ~ 4 Kbars which is also the emplacement pressure of the Winnepesaukee Tonalite, a 3 kilometer thick CMT sequence would equate to only ~ 5 Kbar pressure at the base of the terrane. Pressures of 6 to 7 Kbars for the Winnepesaukee Tonalite indicate that the magma had an extensive magmatic history below the allochthonous CMT cover rocks. This pressure constraint, combined with the metaluminous, $\delta^{18}\text{O}$, Nd and Sr isotopic compositions of the Winnepesaukee Tonalite and Meredith Porphyritic Granite enclaves, indicates that the more mafic magmas cannot have been partial melts of the CMT metasedimentary rocks. A more conservative 10 kilometers thick CMT (Thompson and others, 1993; Stewart and others, 1993) would equate to ~ 7 Kbars pressure at the decollement, which would put the level of epidote and early hornblende crystallization in the Winnepesaukee Tonalite at the interface between the CMT metasedimentary rocks and basement.

The Winnepesaukee Tonalite and other Spaulding Tonalite plutons are located closer to the dorsal zone than the older, syntectonic members. The lowest density peraluminous magmas were the first of the Winnepesaukee Tonalite magmas up the

dorsal zone and were carried eastwardly in east verging thrust sheets to form the easternmost two-mica granites in the pluton. Progressively more mafic magmas migrated up the dorsal zone in the form of dikes to pressures of 3.8 Kbars where they accumulated with the earlier emplaced, more felsic peraluminous magmas to form the zoned Winnepesaukee Tonalite (fig. 18C). Additional Spaulding magmas were carried to the west (fig. 1).

There may be a relative paucity of mafic rocks in central New Hampshire leading others to propose that the Acadian orogeny involved redistribution of mass and energy within a closed midcrustal system, but mafic magmas of this same age are more abundant in Maine and Vermont (Doll and others, 1961; Osberg and others, 1985; Dorais and Paige, 2000) and in the Merrimack belt of southeastern New Hampshire (Watts and others, 2000). Density filtering may have hindered the ascent of these magmas in the thicker portions of the orogen, thereby explaining their minor abundance in central New Hampshire.

Concord Granite plutons were emplaced some 30 to 50 Ma later than the Spaulding Tonalite and are randomly distributed across the CMT in New Hampshire (fig. 18D). The Concord magmas may have originated by a delayed melting of thickened continental crust resulting from thrusting during the Acadian orogeny (England and Thompson, 1984; Chamberlain and England, 1985; Zen, 1988). The random distribution indicates that by the time of emplacement, the tectonic forces had ceased, thereby having no control on the geographic distribution of Concord Granite emplacement as was the case with the syntectonic members of the NHPS.

Finally, the CMT cover rocks have been decoupled from the basement since the Acadian orogeny (Stewart, 1989; Wintch and others, 1992; Kohn and Spear, 1999) as Avalon underplated the margin of Laurentia. The inferred mafic roots of these syntectonic NHPS plutons may no longer be under the dorsal zone, having been transported westwardly during the Alleghanian orogeny (fig. 18E).

CONCLUSIONS

The Bethlehem Granodiorite was the earliest emplaced New Hampshire Plutonic Suite (NHPS) member, perhaps having formed from partial melting along the active decollement during thrusting, and migrated up the dorsal zone to the far western portions of New Hampshire. This member is the most heavily deformed of the NHPS, having been subjected to Acadian metamorphism during emplacement and crystallization. The Kinsman Granodiorite appears to be compositionally equivalent to the Bethlehem, but was considerably dryer and hotter. The presence of mafic enclaves with primitive isotopic compositions in the Meredith Porphyritic Granite provides evidence of heat augmentation and the presence of mantle/lower crustal magmas in the NHPS. The Kinsman magmas also migrated up the dorsal zone to a location closer to the dorsal zone than the Bethlehem Granodiorite. The Spaulding Tonalite, which includes the Winnepesaukee Tonalite, is the youngest syntectonic member of the NHPS and is located almost exclusively adjacent to the dorsal zone. Hornblende geobarometry, the presence of magmatic epidote suggesting intratelluric crystallization, and bulk-rock compositions congruent with high pressure differentiation, all suggest a magmatic history at pressures as great as 7 Kbars that was below the CMT decollement. Nd, Sr, and O isotopic compositions for a range of Winnepesaukee rocks indicate a mantle or a lower crustal component to the pluton. These metaluminous magmas were the last up the dorsal zone, having been previously arrested in their ascent by density filtering. The hornblende-bearing metaluminous rocks provide evidence of higher temperature magmas, indicating heat augmentation from lower crustal/mantle sources. While the geochemical characteristics of the majority of the NHPS plutons in central New Hampshire provide evidence of major CMT input to the magmas, there is no compelling reason to suggest that either the Spaulding or the Kinsman magmas were

the products of an essentially *in situ* heat source and that the Acadian orogeny was closed to mantle/lower crustal heat and/or magmatism.

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