Supra–subduction zone extensional magmatism in Vermont and adjacent Quebec: Implications for early Paleozoic Appalachian tectonics

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ABSTRACT

Metadiabasic intrusions of the Mount Norris Intrusive Suite occur in fault-bounded lithotectonic packages containing Stowe, Moretown, and Cram Hill Formation lithologies in the northern Vermont Rowe-Hawley belt, a proposed Ordovician arc-trench gap above an east-dipping subduction zone. Rocks of the Mount Norris Intrusive Suite are characteristically massive and weakly foliated, have chilled margins, contain xenoliths, and have sharp contacts that both crosscut and are parallel to early structural fabrics in the host metasedimentary rocks. Although the mineral assemblage of the Mount Norris Intrusive Suite is albite + actinolite + epidote + chlorite + calcite + quartz, intergrowths of albite + actinolite are probably pseudomorphs after plagioclase + clinopyroxene. The metadiabases are subalkaline, tholeiitic, hypabyssal basalts with preserved ophitic texture. A backarc-basin tectonic setting for the intrusive suite is suggested by its LREE (light rare earth element) enrichment, negative Nb-Ta anomalies, and Ta/Yb vs. Th/Yb trends. Although no direct isotopic age data are available, the intrusions are broadly Ordovician because their contacts are clearly folded by the earliest Acadian (Silurian–Devonian) folds. Field evidence and geochemical data suggest compelling along-strike correlations with the Coburn Hill Volcanics of northern Vermont and the Bolton Igneous Group of southern Quebec. Isotopic and stratigraphic age constraints for the Bolton Igneous Group bracket these backarc magmas to the 477–458 Ma interval. A tectonic model that begins with east-dipping subduction and progresses to outboard west-dipping subduction after a syncollisional polarity reversal best explains the intrusion of deformed metamorphosed metasedimentary rocks by backarc magmas.

Keywords: metadiabases, supra–subduction zone, Rowe-Hawley belt, Dunnage zone, Bolton Igneous Group, Coburn Hill Volcanics, Mount Norris Intrusive Suite.

INTRODUCTION

The pre-Silurian tectonic belts of the northeastern Appalachians are composed of rocks that were deposited in an ancient ocean basin (Iapetus) that developed east of ancestral North America (Laurentia) during Late Proterozoic to Ordovician time; these rocks originated within the continental margin, main ocean basin, or supra–subduction zone (including forearc, arc, and backarc). In southern Quebec, these rocks are assigned to the Humber and Dunnage zones (Fig. 1) depending on whether they have continental-margin or oceanic (including supra–subduction zone) affinity, respectively (e.g., Williams, 1978). Rocks of the Dunnage zone were juxtaposed against those of the Humber zone during the Early to Middle Ordovician closure of Iapetus. The Dunnage zone and part of the easternmost Humber zone of southern Quebec connect along strike to the south with the northern Vermont Rowe-Hawley belt of New England. That belt has not been formally subdivided into continental and oceanic (including supra–subduction zone) domains as in southern Quebec because the belt, as a terrane, has been defined to include lithologies that were assembled in an arc-trench gap tectonic setting (e.g., Stanley and Ratcliffe, 1985) during the closure of Iapetus. Thus, the belt contains elements of both southern Quebec zones.

The bedrock geology of the northern Vermont Rowe-Hawley belt occupies a critical position in the northern Appalachian orogen because it links well-preserved early Paleozoic ophiolite and accretionary-wedge sequences in southern Quebec with correlative rocks in New England. Pre-Silurian metadiabasic intrusions of the Mount Norris Intrusive Suite, found within specific metasedimentary thrust slices in the Rowe-Hawley belt, are particularly important links. These intrusions have a supra–subduction zone geochemical signature and, thus, also provide crucial information about the tectonic setting of the New England and Quebec Appalachians during Ordovician (Taconic) orogenesis. The purpose of this paper is to describe the geologic setting and geochemistry of the Mount Norris Intrusive Suite metadiabases in northern Vermont, make correlations with analogous mafic rocks in Vermont and southern Quebec, and formulate tectonic models for the region on the basis of these data and other geologic constraints.

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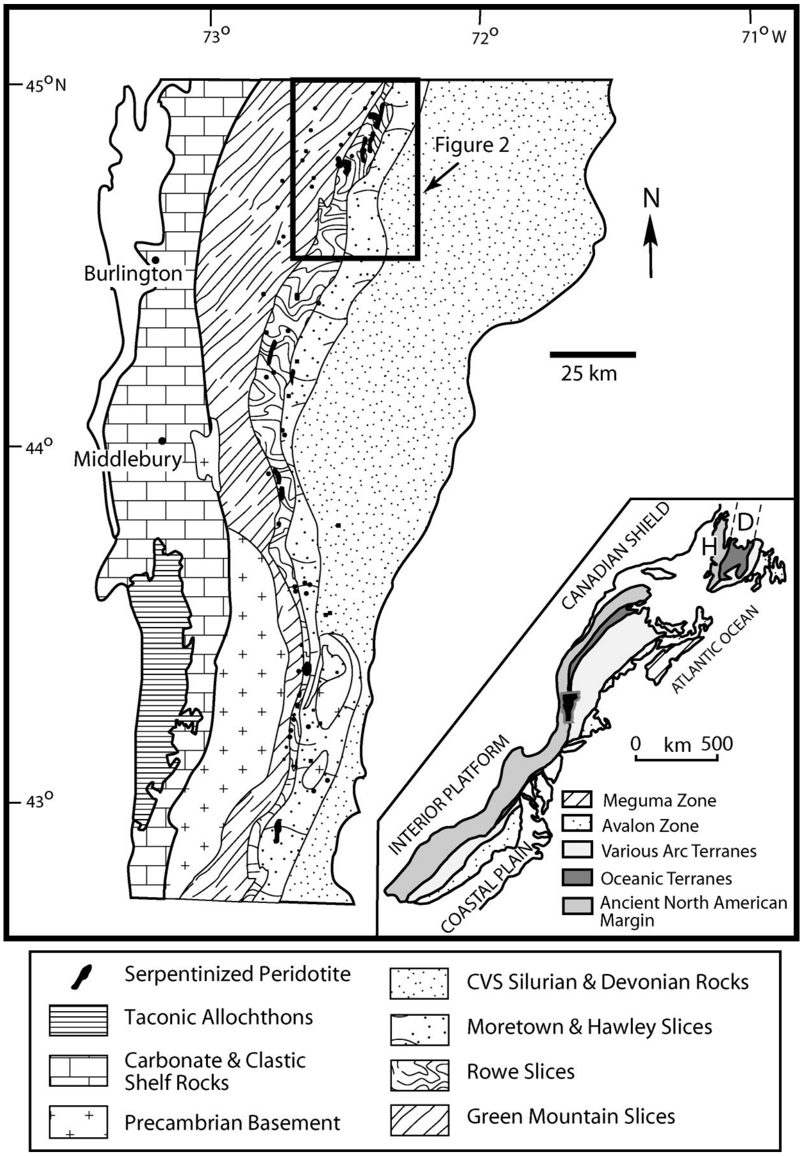


Figure 1. Generalized geologic map of Vermont (after Doll et al., 1961; Stanley and Ratcliffe, 1985) and generalized tectonostratigraphic map (inset) of the Appalachians (Williams, 1978). CVS—Connecticut Valley Sequence. Note that in the text, the Rowe-Hawley belt includes both the Rowe slices and Moretown and Hawley slices on this map. In inset map, H and D refer to Humber and Dunnage zones, respectively, as defined by Williams and modified by later workers (e.g., Waldron and van Staal, 2001).

REGIONAL GEOLOGIC SETTING

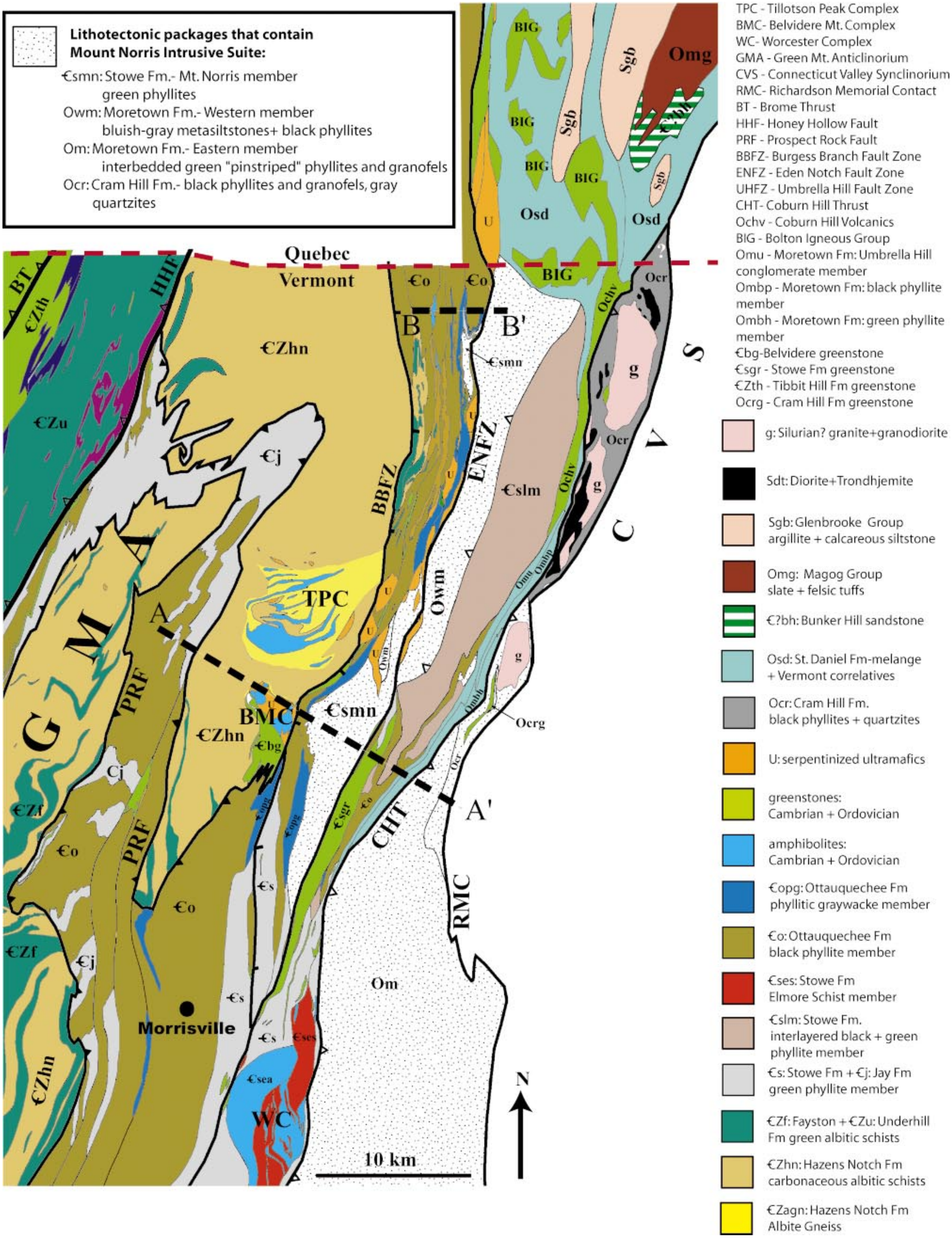
Western Vermont exposes continental-margin rocks deformed by collision of a volcanic arc with the ancient Laurentian continent during closure of segments of the Early Paleozoic Iapetus Ocean (Stanley and Ratcliffe, 1985; Rankin, 1994). Precambrian (Laurentian) basement makes up the core of the Green Mountains. Carbonate and clastic rocks in western Vermont represent an early Paleozoic continental-shelf sequence (Fig. 1). A sequence of rift-related clastic rocks of Late Proterozoic to early Paleozoic age is preserved in a group of thrust sheets, called the Green Mountain slices (Fig. 1). Oceanic and supra–subduction zone rocks occur in the Rowe and the Moretown and Hawley slices, known collectively as the Rowe-Hawley belt. The belt in Vermont includes the Ottauquechee, Stowe, Moretown, and Cram Hill Formations in the northern half of the state; intrusive and extrusive rocks of the Barnard Gneiss and North River Igneous Suite (Armstrong, 1995; Ratcliffe et al., 1998) are part of the Rowe-Hawley belt in the southern half of Vermont. The Connecticut Valley Sequence lies to the east of the belt and consists mainly of Silurian and Devonian rocks. Although the Rowe-Hawley belt and Green Mountain slices were first deformed and metamorphosed during the Ordovician Taconic orogeny, the overprint from the Silurian–Devonian Acadian orogeny varies in intensity from modest to severe and generally increases in intensity to the east and in the vicinity of Acadian faults.

The Mount Norris Intrusive Suite crops out in part of the Rowe-Hawley belt; in particular, representatives of the suite appear most frequently in the Stowe and Moretown Formations (Fig. 2). Tectonic models for the Rowe-Hawley belt in New England (e.g., Stanley and Ratcliffe, 1985; Stanley and Hatch, 1988; Kim and Jacobi, 1996) generally portray the Stowe and Ottauquechee Formations as distal continental-shelf or rise sedimentary deposits that were incorporated into an accretionary prism during eastward-dipping subduction in the Ordovician, whereas the Moretown and Cram Hill Formations were situated to the east of the accretionary prism and were intruded by arc-source magmas in a supra–subduction zone setting.

LOCAL GEOLOGIC SETTING

The field area covering the northern Vermont Rowe-Hawley belt extends from the Vermont-Quebec border southward to the latitude of the Town of Morrisville and is bounded to the west by the Burgess Branch fault zone and to the east by the Richardson Memorial Contact, a major unconformity separating pre-Silurian from Silurian–Devonian rocks (Fig. 2). The northern Vermont Rowe-Hawley belt consists primarily of metasedimentary and metaigneous rocks from the Ottauquechee, Stowe, Moretown (western and eastern members), and Cram Hill Formations (lithologic descriptions are given in Fig. 2). The Belvidere Mountain, Tillotson Peak, and Worcester structural complexes are also considered to be part of the northern Vermont Rowe-Hawley belt.

Ottauquechee (Ꞓo) and Stowe Formation (Ꞓs) rocks are commonly juxtaposed with one another in fault-bounded lithotectonic packages. Likewise, Moretown (Owm, Om) and Cram Hill Formation (Ocr) lithologies form lithotectonic packages. Ultramafic rocks in the northern Vermont Rowe-Hawley belt occur only in lithotectonic packages containing Ottauquechee, Stowe, and Moretown Formation rocks. Rowe-Hawley belt rocks sit structurally


Lithotectonic packages that contain Mount Norris Intrusive Suite:
€smn: Stowe Fm.- Mt. Norris member green phyllites
Owm: Moretown Fm.- Western member bluish-gray metasiltstones+ black phyllites
Om: Moretown Fm.- Eastern member interbedded green "pinstriped" phyllites and granofels
Ocr: Cram Hill Fm.- black phyllites and granofels, gray quartzites
TPC - Tillotson Peak Complex
BMC- Belvidere Mt. Complex
WC- Worcester Complex
GMA - Green Mt. Anticlinorium
CVS - Connecticut Valley Synclinorium
RMC- Richardson Memorial Contact
BT - Brome Thrust
HHF- Honey Hollow Fault
PRF - Prospect Rock Fault
BBFZ- Burgess Branch Fault Zone
ENFZ - Eden Notch Fault Zone
UHFZ - Umbrella Hill Fault Zone
CHT- Coburn Hill Thrust
Ochv - Coburn Hill Volcanics
BIG - Bolton Igneous Group
Omu - Moretown Fm: Umbrella Hill conglomerate member
Ombp - Moretown Fm: black phyllite member
Ombh - Moretown Fm: green phyllite member
€bg-Belvidere greenstone
€sgr - Stowe Fm greenstone
€Zth - Tibbit Hill Fm greenstone
Ocrg - Cram Hill Fm greenstone
g: Silurian? granite+granodiorite
Sdt: Diorite+Trondhjemite
Sgb: Glenbrooke Group argillite + calcareous siltstone
Omg: Magog Group slate + felsic tuffs
€?bh: Bunker Hill sandstone
Osd: St. Daniel Fm-melange + Vermont correlatives
Ocr: Cram Hill Fm. black phyllites + quartzites
U: serpentinized ultramafics
greenstones: Cambrian + Ordovician
amphibolites: Cambrian + Ordovician
€opg: Ottauquechee Fm phyllitic graywacke member
€o: Ottauquechee Fm black phyllite member
€ses: Stowe Fm Elmore Schist member
€slm: Stowe Fm. interlayered black + green phyllite member
€s: Stowe Fm + €j: Jay Fm green phyllite member
€Zf: Fayston + €Zu: Underhill Fm green albitic schists
€Zhn: Hazens Notch Fm carbonaceous albitic schists
€Zagn: Hazens Notch Fm Albite Gneiss
Quebec
Vermont
Morrisville
10 km
N

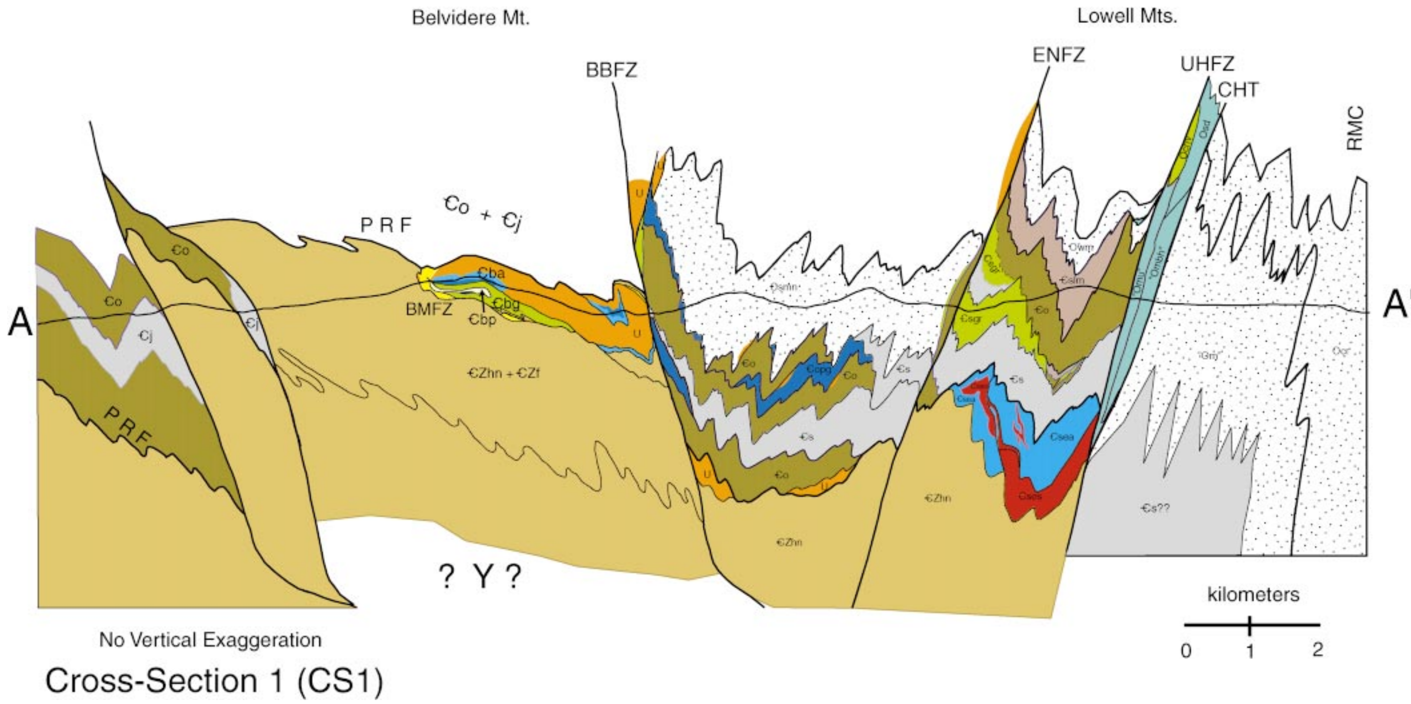


Figure 3. Cross section A–A′ modified from Kim et al. (1999). See Figure 2 for line of section. Dot pattern identifies lithotectonic packages that have metadiabases (Csmn, Owm, Om, and Ocr). BMFZ—Belvidere Mt. Fault Zone, Cba—Belvidere amphibolite, Cbp—Belvidere politic schist, Csea—Elmore amphibolite member of Stowe Formation.

above albite porphyroblast–bearing rocks of Hazens Notch (ЄZhn) and Fayston (ЄZf) Formations (Thompson and Thompson, 2003) as well as the intervening Belvidere Mountain (ophiolitic) and Tillotson Peak (blueschist-bearing) mafic complexes (Laird et al., 1984; Gale, 1986; Kim et al., 2001; Laird et al., 2001). The base of the Rowe-Hawley belt has been interpreted to be the Taconic-age Prospect Rock Fault (Fig. 2)—a west-directed thrust (Thompson et al., 1999; Thompson and Thompson, 2003). The Burgess Branch fault zone (Fig. 2) is a steeply east-dipping down-to-the-east normal fault of presumed Silurian age that reactivated an earlier thrust surface (Kim et al., 1999); detailed microstructural analysis has verified the normal-fault interpretation (Lamon and Doolan, 2001). The name ''Burgess Branch fault zone'' has supplanted the name ''Belvidere Mountain Fault zone'' of Stanley et al. (1984) (Kim et al., 1999). The Belvidere Mountain Fault zone was specifically redefined to represent only the sole thrust at the base of the Belvidere Mountain Complex (Kim et al., 1999). The Eden Notch Fault zone, of uncertain age, is a steeply west-dipping fault surface with equivocal slip direction (Fig. 2).

Detailed mapping has delineated fault-bounded lithotectonic packages that contain Mount Norris Intrusive Suite metadiabases in the study area (Stanley et al., 1984; Kim, 1997; Kim et al., 1998, 1999) (Fig. 2). None of the metadiabase-bearing slices is found east of the unconformity between pre-Silurian and Silurian–Devonian rocks (Richardson Memorial contact). Ultramafic bodies are found in fault contact with metadiabase-bearing lithologies in some locations.

A geologic cross section modified from Kim et al. (1999) (Fig. 3) shows that the metadiabases are found only in fault-bounded Stowe (Єsmn), Moretown (Om, Owm), and Cram Hill (Ocr) lithologies that lie at the highest structural levels. Similarly, a cross section modified from R.S. Stanley (1990, personal commun.) across the northern part of the study area shows that the metadiabases occur in the same Stowe and Moretown (western member) lithologies (Fig. 4A); however, instead of lying at the highest structural level, these lithotectonic packages were emplaced along early faults and subsequently were complexly infolded with Ottauquechee Formation lithologies (Єo). The metadiabases do not have a uniform distribution throughout any individual lithotectonic package. For example, Figure 4B shows the distribution of individual metadiabase bodies in the Big Falls synform represented in the cross section of Figure 4A (Stanley et al., 1984; Evans, 1994).

STRUCTURAL GEOLOGY AND METAMORPHISM

The field area (Fig. 2) lies in the overlap zone between Taconic and Acadian structural fabrics, and the rocks have generally been subjected to biotite-grade metamorphism, although the Belvidere Mountain, Tillotson Peak, and Worcester Complexes have undergone garnet and higher grades of metamorphism (Laird et al., 1984, 2001). The dominant foliation west of the Burgess Branch fault zone is a generally steeply east-dipping Ta-

Figure 2. Geologic map of the northern Vermont study area (modified from Kim et al., 1999) and southern Quebec (modified from Slivitsky and St-Julien, 1987). Primarily white areas with light-colored dots are lithotectonic packages in the northern Vermont Rowe-Hawley belt that have Mount Norris Intrusive Suite metadiabasic intrusions as detailed in upper-left key.

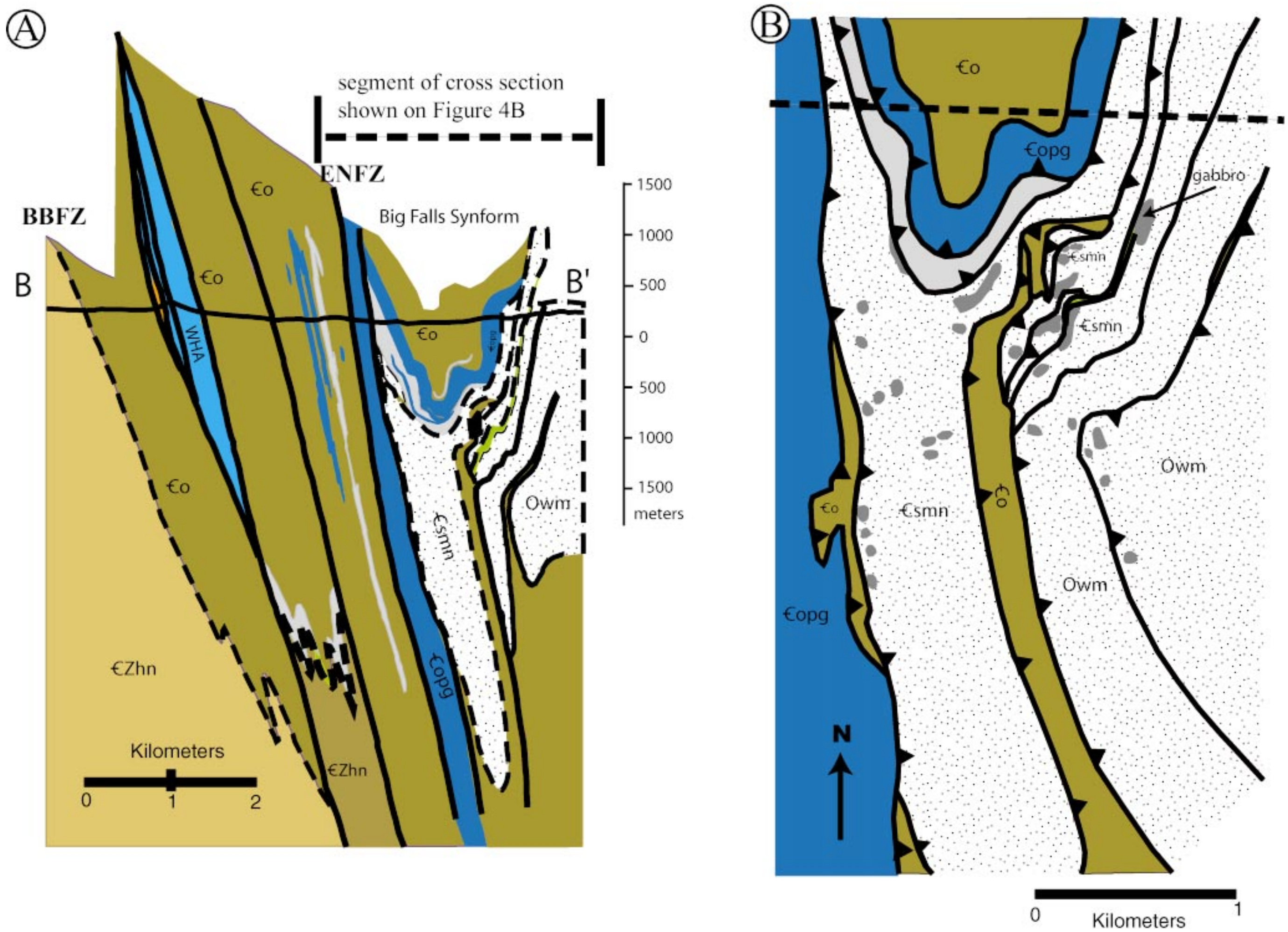


Figure 4. (A) Cross section B–B′ modified from R.S. Stanley (1990, personal commun.). Dot pattern identifies lithotectonic packages containing Mount Norris Intrusive Suite metadiabases. WHA—Warner Hill Amphibolite. (B) Map of Big Falls synform area modified from Dick (1989), showing distribution of Mount Norris Intrusive Suite metadiabase bodies. Csmn and Owm are the only units that contain metadiabase (dot pattern).

conic S1-S2 composite foliation that has been folded by gently plunging asymmetric F3 folds (Green Mountain folds) with a locally developed steeply west-dipping crenulation cleavage (S3). East of the Burgess Branch fault zone, the dominant foliation is a composite S2/S3 fabric in which S2 has been reoriented into the attitude of S3 and has been overprinted by a strongly developed S3 spaced cleavage (Kim et al., 1999). The S3 cleavage is known to be Silurian–Devonian because this cleavage can be traced across strike into Silurian–Devonian rocks where it is the earliest cleavage.

In the northern Vermont Rowe-Hawley belt, $^{40}Ar/^{39}Ar$ total-fusion ages on amphiboles indicate that the age of the Taconic metamorphism generally ranges from 471 to 460 Ma, whereas muscovite and biotite total-fusion ages indicate that the Acadian metamorphism ranges in age from 386 to 355 Ma (Laird et al., 1984, 1993). These data are consistent with the suggestion that the S1/S2 composite fabric is probably 471–460 Ma in age and the S3 fabric is probably 386–355 Ma in age.

Laird et al. (1993) obtained an $^{40}Ar/^{39}Ar$ plateau age of 505 $\pm$ 2 Ma on barroisitic amphibole from the Belvidere Mountain Complex amphibolite; this age implies that this amphibolite was deformed and metamorphosed prior to juxtaposition with the surrounding metasedimentary rocks at ca. 470–460 Ma. A total-fusion age of 468 $\pm$ 6 Ma on glaucophane from the adjacent blueschist-bearing Tillotson Peak Complex was reported by Laird et al. (1984).

Recent comprehensive $^{40}Ar/^{39}Ar$ work on more than 100 samples in southern Quebec indicates that metamorphic ages on muscovite and amphiboles range from 469 to 460 Ma in the Dunnage zone and from 430 to 411 Ma in the internal part of the Humber zone (Castonguay et al., 2001). The separation between the two metamorphic-age domains is the Silurian, down-to-the-east, St. Joseph normal fault, which is locally coincident with the Baie-Verte–Brompton Line that separates Dunnage zone from Humber zone rocks (Williams and St-Julien, 1982; Castonguay et al., 2001). The oldest metamorphic ages in the Dunnage zone are associated with the S1-S2 composite foliation generated by east-over-west thrusting, whereas the Silurian ages are associated with S3, which is either a west-over-east backthrust-related fabric or down-to-the-east normal-fault fabric. Both the Dunnage and easternmost part of the Humber zones are also affected by

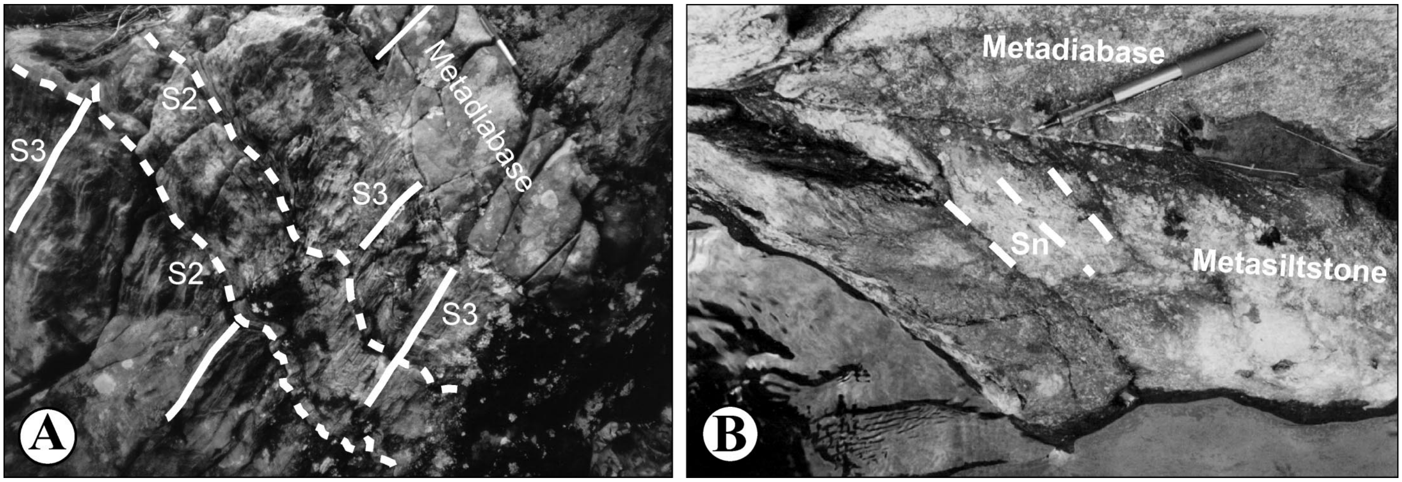


Figure 5. (A) Metadiabasic sill on Hadley Mountain that intruded parallel to the dominant foliation (S2) in grayish-green phyllites of the Stowe Formation (Csmn). Note pen for scale. Open asymmetric F3 folds (Acadian) clearly fold the metadiabase; note that the accompanying S3 cleavage is emphasized. (B) Close-up of metadiabase contact with bluish-gray metasiltstones of the western member of the Moretown Formation (Owm) in Lowell, Vermont. This metadiabase cuts the earliest foliation preserved in the metasiltstones foliation (Sn).

Silurian–Devonian (Acadian) deformation and metamorphism.

DESCRIPTION OF MOUNT NORRIS INTRUSIVE SUITE

Field Relationships

The Mount Norris Intrusive Suite metadiabases are characteristically gray, massive, rounded, granular, weakly foliated rocks with distinct buff-colored weathering rinds. Some, but not all, metadiabases have plagioclase phenocrysts (Fig. 5). Fresh metadiabase surfaces commonly have secondary calcite. An intrusive origin is based on the presence of chilled margins, rare xenoliths of metasedimentary lithologies, and sharp contacts with surrounding metasedimentary units. The metadiabase contacts may be either coplanar with the dominant foliation (Fig. 5A) or cut it at some angle (Fig. 5B). At one location in the western member of the Moretown Formation metasedimentary rocks (Owm), a large (5 m wide) metadiabase dike in the metasiltstones cuts the earliest recognizable composite foliation (probably Taconic) at a high angle (Fig. 5B); this early foliation is folded by the Acadian F3 upright asymmetric folds. Intrusion of the dikes/sills is always clearly pre-S3 (Acadian) and therefore is pre- or syn-S2 (Fig. 5A). The metadiabases are frequently boudinaged within the dominant foliation (Taconic S2) such that individual dikes/sills can usually not be traced on the ground over large distances. The boudinage indicates that the original contact relationship between the metadiabases and the host metasedimentary rocks can be obscured because the contact may have been rotated into parallelism with the dominant foliation. The metadiabases do not cross any known fault.

Mineral Assemblage

Although highly altered, the metadiabases exhibit an ophitic texture that is observable both in outcrop and hand sample. In thin section, the ophitic texture manifests itself as intergrowths of albite + actinolite that may be pseudomorphs after plagioclase + clinopyroxene. In many instances, the plagioclase has been replaced by epidote. The overall mineral assemblage of the metadiabases is albite + actinolite + epidote + chlorite + calcite + quartz (Dick, 1989; Evans, 1994; Kim and Coish, 2001).

AGE

Regional Age Information

There are no igneous crystallization ages for any igneous units in the study area, so along-strike extrapolation is necessary to establish age control. The minimum age for the Moretown Formation in southern Vermont has been established by 496 ± 8 Ma and 486 ±3 Ma U-Pb zircon ages on a trondhjemitic and a tonalitic intrusion, respectively (Ratcliffe et al., 1997). A metafelsite within the Cram Hill Formation in southern Vermont has been dated at 484 ± 4 Ma by U-Pb in zircon (Ratcliffe et al., 1997).

In the Dunnage zone of Quebec, the along-strike counterpart to part of the northern Vermont Rowe-Hawley belt, the U-Pb zircon age of plagiogranites in the Thetford Mines and Mount Orford ophiolites is 479 +3/–2 Ma (Dunning and Pedersen, 1988) and 504 ± 3 Ma (David and Marquis, 1994), respectively. More recently acquired U-Pb zircon ages of granitoids in the Thetford Mines ophiolite are 469 ± 4 Ma and 470 +5/–3 Ma (Whitehead et al., 2000). Metarhyolites from the Ascot Complex yielded U-Pb zircon ages of 441 +7/–12 and 460 ± 3 Ma (David and Marquis, 1994).

Doolan et al. (1982) traced black and gray slates interbedded with pillowed and massive greenstones that are part of the St. Daniel Formation mélange in southern Quebec across the international border into Vermont where they are mapped as part of the Cram Hill Formation and Coburn Hill Volcanics, respectively. Although there is no fossil age control in the northern Vermont Rowe-Hawley belt, there are zone 12 graptolites (ca. 458 Ma on the 1999 Geological Society of America Time Scale) near the base of the Magog Group in southern Quebec (Berry, 1962; Harwood and Berry, 1967). Because the Magog Group unconformably overlies the St. Daniel Formation mélange in southern Quebec (Cousineau, 1990), the mélange is ostensibly pre–Late Ordovician in age. In addition, the St. Daniel Formation unconformably overlies the ca. 477 Ma (obduction age) Thetford Mines ophiolite

TABLE 1. MAJOR AND TRACE ELEMENT DATA FOR METADIABASE DIKES OF THE MOUNT NORRIS INTRUSIVE SUITE, NORTHERN VERMONT

Field no.	6-13-96-1	6-26-96-1	6-28-96-2-2	7-8-96-8	6-23-96-1	7-8-96-2	7-10-96-1	7-11-96-1	6-17-96-4	8-10-96-2	8-15-96-1	8-14-96-1-2	7-25-96-1
SiO_2	51.87	54.47	54.08	51.75		54.33	52.98	52.81	53.53	48.71	49.40	53.31	54.66
TiO_2	1.54	1.64	1.45	1.51		1.35	1.48	1.13	1.11	1.05	0.95	1.29	1.30
Al_2O_3	15.52	15.73	15.09	14.81	14.89	15.01	15.97	15.57	15.41	16.13	16.74	14.65	15.42
Fe_2O_3(t)	11.90	13.30	11.93	11.86	10.91	12.06	11.55	10.27	11.11	10.51	9.23	10.58	12.31
MnO	0.17	0.35	0.16	0.19	0.27	0.18	0.20	0.18	0.18	0.16	0.16	0.17	0.21
MgO	5.84	5.13	5.93	6.73		6.37	6.79	8.34	6.86	6.79	7.64	5.41	5.85
CaO	9.63	7.14	7.80	10.36		10.41	9.48	9.36	9.94	11.57	12.16	9.12	8.71
Na_2O	3.89	1.55	0.91	2.79	1.13	1.30	1.39	2.75	0.72	2.23	2.14	2.66	3.25
K_2O	0.31	0.01	0.03	0.25	2.13	0.03	0.02	0.38	0.03	0.17	0.42	0.76	0.08
P_2O_5	0.17	0.18	0.10	0.10	0.12	0.13	0.17	0.14	0.12	0.12	0.12	0.15	0.15
Total	100.83	99.52	97.48	100.35		101.17	100.02	100.92	99.00	97.44	98.96	98.10	101.94
LOI	13.79	4.10	5.36	2.65	14.15	7.08	7.06	10.69	3.71	2.41	2.74	1.95	2.69
Sc	42	46	46	33	47	40	44	50	38	37	44	35	48
V	326	372	328	256	296	278	300	296	287	264	248	282	345
Cr	116	58	137	147	124	128	183	407	251	211	316	139	150
Ni	67	45	66	218	65	52	87	109	56	102	91	58	52
Cu	42	25	80	48	54	67	38	88	57	67	110	49	40
Sr	234	284	214	229	210	319	241	171	236	187	178	193	165
Y	35	36	35	31	24	30	31	23	28	24	20	32	29
Zr	104	106	101	92	61	88	98	73	93	65	49	103	70
Ba	26	245	16	10	159	7	7	29	18	29	30	92	10

Field no.	6-12-96-1	6-25-96-1	187	ME-94-11	ME-94-12	ME-94-13	ME-94-14	ME-94-15	ME-94-16	ME-94-17	ME-94-18	ME-94-19	ME-94-20
SiO_2	52.23	53.81	48.27	51.48	53.10	50.07	54.56	55.89	53.98	51.72	50.71	49.26	49.72
TiO_2	1.17	1.34	0.84	1.45	1.44	1.48	1.54	1.67	1.26	1.09	1.20	1.56	1.04
Al_2O_3	15.32	15.69	19.34	15.21	15.09	14.78	14.30	14.70	14.91	16.55	14.94	15.90	15.01
Fe_2O_3(t)	11.18	11.68	8.61	10.76	10.49	11.45	12.11	12.60	10.40	10.41	10.46	12.62	10.06
MnO	0.18	0.25	0.14	0.19	0.17	0.28	0.22	0.30	0.19	0.22	0.27	0.35	0.19
MgO	6.45	7.58	6.52	6.47	6.04	6.36	4.64	4.30	5.64	6.12	7.10	6.41	6.63
CaO	8.45	6.67	11.88	10.15	9.16	9.70	8.04	6.48	8.42	9.38	10.89	9.80	10.22
Na_2O	3.65	3.96	2.07	3.00	2.72	2.06	2.07	2.29	3.51	3.57	2.09	1.29	3.19
K_2O	0.33	0.24	0.77	0.03	0.01	0.00	0.04	0.10	0.06	0.12	0.06	0.00	0.28
P_2O_5	0.13	0.14	0.11	0.16	0.17	0.16	0.20	0.24	0.17	0.16	0.12	0.17	0.12
Total	99.09	101.36	98.55	98.90	98.40	96.33	97.72	98.57	98.55	99.34	97.84	97.36	96.47
LOI	3.35	8.26		2.59	2.70	8.30	2.93	4.23	2.24	2.71	3.81	7.23	2.42
Sc	40	42	34	41	40	42	41	42	38	38	45	43	43
V	288	307	220	298	303	298	303	319	258	227	289	317	274
Cr	110	222	192	199	149	192	86	115	172	197	201	194	149
Ni	52	74	78	64	55	77	39	32	61	88	66	70	59
Cu	92	79	47	46	27	25	47	94	34	16	96	18	16
Sr	292	110	184	289	326	276	291	316	237	331	273	341	205
Y	25	30	19	36	36	36	36	32	30	31	24	33	23
Zr	66	80	52	107	113	101	118	136	130	130	85	123	81
Ba	35	17	167	6	6	4	9	14	126	39	9	6	47

Field no.	ME-94-21	ME-94-22	ME-94-23	ME-94-24	GD8813	GD8814	GD8815	GD8816	GD8818	GD8840	GD8841	GD8847	GD8850
SiO_2	51.83	49.75	50.31	48.99	51.99	52.23	56.16	52.7	50.82	50.51	53.47	54.7	50.78
TiO_2	1.33	1.32	1.09	1.27	1.69	1.69	0.54	1.67	0.75	1.92	0.41	1.26	1.27
Al_2O_3	15.02	14.87	15.06	15.57	14.03	13.72	18.96	14.42	17.66	14.16	20.54	14.98	15.08
Fe_2O_3(t)	11.06	11.37	10.85	10.75	12.95	12.66	9.35	13.08	7.89	13.06	5.84	10.51	11.62
MnO	0.21	0.21	0.21	0.21	0.21	0.21	0.08	0.2	0.15	0.21	0.1	0.18	0.19
MgO	6.62	7.71	7.43	7.60	6.15	6.02	5.41	6.51	8.84	6.32	5.52	5.21	7.67
CaO	9.29	11.00	10.75	11.25	10.04	9.5	1.84	8.4	9.95	12.61	7.1	9.18	10.55
Na_2O	2.82	2.68	2.83	2.71	2.98	3.11	7.28	3.35	3.36	1.38	5.35	2.59	3.21
K_2O	0.26	0.24	0.29	0.19	0.05	0.08	0.05	0.08	0.01	0.09	2.04	1.11	0.11
P_2O_5	0.11	0.13	0.11	0.12	0.19		0.11						
Total	98.55	99.28	98.93	98.66	100.28	99.22	99.78	100.41	99.43	100.26	100.37	99.72	100.48
LOI	2.76	2.61	2.42	2.80									
Sc	43	46	45	48	42	44	31	45	34	43	27	39	46
V	338	308	273	295	360	369	189	384	193	371	153	334	332
Cr	149	129	144	253	95	88	114	109	360	69	73	104	187
Ni	66	63	70	79	53	60	43	61	162	58	67	58	76
Cu	53	26	21	40	43	25	7	63	98	149	75	21	72
Sr	202	206	217	242	315	299	132	267	248	211	149	430	227
Y	24	24	20	24	34	30	20	37	11	36	13	34	31
Zr	91	83	71	83	114	127	89	127	70	91	14	129	86
Ba	32	19	46	17	6	10	6	9	8	22	172	59	10

TABLE 1. *(Continued)*

Field no.	6-13-96-1	6-26-96-1	6-28-96-2-2	7-8-96-8	6-23-96-1	7-8-96-2	7-10-96-1	7-11-96-1	6-17-96-4	8-10-96-2	8-15-96-1	8-14-96-1-2	7-25-96-1
Field no.	GD8853	GD8854	GD8855	528	170	176	ME-94-25	ME-94-37	ME-94-38	ME-94-39	ME-94-40	ME-94-41	ME-94-42
SiO_2	50.55	51.95	53.35	51.86	52.14	50.25	50.64	51.38	52.05	46.79	50.69	49.89	52.41
TiO_2	1.97	1.47	1.64	1.6	0.6	1.18	1.05	1.67	1.15	1.32	1.43	1.41	1.17
Al_2O_3	15.1	15.98	14.26	14.17	15.19	16.18	15.40	14.07	14.44	16.32	15.05	14.79	14.79
Fe_2O_3(t)	13.42	11.41	11.87	12.17	12.37	11.04	9.88	12.84	10.82	11.77	11.92	11.74	11.21
MnO	0.22	0.19	0.2	0.21	0.21	0.2	0.20	0.22	0.22	0.20	0.26	0.21	0.22
MgO	6.98	6.44	6	6.64	5.98	8.23	7.37	6.50	6.51	7.63	7.01	6.97	6.68
CaO	8.92	9.14	9.44	10.81	6.91	9.45	10.30	9.23	10.11	11.85	9.90	10.57	10.36
Na_2O	1.81	2.89	3.07	2.49	5.05	2.43	3.22	2.64	3.38	2.16	2.50	2.97	3.38
K_2O	0.79	0.89	0.25	0.22	0.17	0.06	0.33	0.09	0.10	0.15	0.09	0.07	0.09
P_2O_5					0.11	0.11	0.11	0.17	0.15	0.14	0.15	0.15	0.14
Total	99.76	100.36	100.08	100.17	98.73	99.13	98.50	98.81	98.91	98.33	99.01	98.77	100.45
LOI							2.73	2.69	2.19	3.00	2.85	2.38	2.03
Sc	48	42	43	43	48	45	43	44	46	48	45	44	45
V	422	339	373	329	195	304	253	318	277	314	319	307	267
Cr	94	198	115	107	460	264	193	134	88	204	166	157	121
Ni	57	74	59	50	103	110	60	55	50	58	75	56	51
Cu	49	75	52	109	41	93	62	52	16	44	46	26	18
Sr	228	264	300	179	120	280	232	207	233	212	265	234	238
Y	41	34	34	32	15	26	20	34	28	31	31	30	27
Zr	143	110	109	82	30	75	74	122	102	103	108	103	100
Ba	77	60	27	16	16	21	48	20	7	30	7	7	7

(e.g., Schroetter et al., 2001, 2002). Therefore, the St. Daniel Formation has an age range of 477–458 Ma. The fact that the Cram Hill Formation is correlative with parts of the St. Daniel indicates that the Cram Hill is also pre–Late Ordovician.

The Umbrella Hill conglomerate member (Omu—Fig. 2) of the Moretown Formation is a deformed and metamorphosed polymict quartz-cobble and metamorphic-rock-fragment conglomerate with a phyllitic matrix (e.g. Badger, 1979) that unconformably overlies lithotectonic packages containing Stowe and Ottauquechee Formation rocks (Fig. 2). Kim et al. (1999) interpreted this unconformable contact to be faulted (and called it the Umbrella Hill Fault zone). Under the assumption of east-facing stratigraphic continuity, early workers interpreted the Umbrella Hill conglomerate member to be the base of the Moretown Formation (Konig and Dennis, 1964). Recent mapping, however, shows that it is interlayered to the east with green phyllites (Ombh) that are truncated by the Coburn Hill Thrust (Kim et al., 1999); this faulting has disrupted the stratigraphic continuity within the Moretown Formation such that the original relationship between the Umbrella Hill conglomerate member and the main body of the Moretown (Om) is uncertain. The Umbrella Hill Conglomerate can be traced continuously northward until it is interlayered with greenstones of the Coburn Hill Volcanics; these relationships indicate a general equivalence in stratigraphic position with both the Coburn Hill Volcanics and St. Daniel and Cram Hill Formations.

Age of the Mount Norris Intrusive Suite

Although the metadiabases cannot be dated directly, the following criteria strongly support a pre-Silurian age: (1) The metadiabases either cut or are parallel to the earliest foliations (Ordovician) in the Stowe and Moretown Formation rocks. (2) The metadiabases are confined to distinct fault-bounded thrust slices in the northern Vermont Rowe-Hawley belt. Where exposed, the faults that border the thrust slices are folded by the earliest Silurian–Devonian (F3) folds. (3) The earliest Acadian-age (F3) folds clearly fold the metadiabases. (4) The metadiabases do not cut faults of any age. (5) The metadiabases are folded by pre-F3 folds in the Big Falls synform area of northern Vermont (Stanley et al., 1984). (6) The metadiabases are not found in any Silurian–Devonian metasedimentary rocks to the east.

GEOCHEMISTRY

Analytical Methods

Samples selected for analysis were as fresh and unweathered, homogeneous, and free of significant veining as possible. Samples were cut into ~2 cm cubes with a water-cooled saw, air dried, passed through a ceramic jaw crusher, and powdered in a shatterbox. The powders were ignited at 1000 °C in graphite crucibles and were fused and dissolved via lithium metaborate methods (Coish and Sinton, 1992). Concentrations of the 10 major elements (SiO_2, Al_2O_3, TiO_2, MgO, Fe_2O_3, CaO, Na_2O, K_2O, MnO, and P_2O_5) and concentrations of Cr, Ni, V, and Sc were determined on a Jarrell Ash ICP-OES (inductively coupled plasma–optical emission spectrometer) unit at Middlebury College, Middlebury, Vermont. Concentrations of the 14 rare earth elements (REEs) and of Rb, Sr, Ba, Th, Nb, Ta, Zr, Hf, and Y were determined on a VG Instruments Plasmasquad ICP-MS (inductively coupled plasma–mass spectrometry) at Union College, Schenectady, New York. All geochemical data are shown in Tables 1 and 2. Analytical accuracy was evaluated by running U.S. Geological Survey standards as unknowns before and after each 10-sample run (Table 3).

Chemical Mobility

Metamorphic mineral assemblages in the metadiabases and surrounding metasedimentary rocks indicate that, in the study area, the Rowe-Hawley belt underwent lower greenschist–facies metamorphism that reached a maximum of biotite grade (Laird et al., 1984). Because the primary mineral assemblage of the metadiabases has been severely altered and loss-on-ignition (LOI) values are relatively high, chemical effects of alteration should be evaluated prior to presenting petrogenetic and tectonic interpretations.

TABLE 2. RARE EARTH ELEMENT, Hf, Ta, Th AND Nb VALUES FOR SELECTED METADIABASE DIKES OF THE MOUNT NORRIS INTRUSIVE SUITE

Field no.	6-13-96-1	6-26-96-1	6-28-96-2-2	7-8-96-8	6-23-96-1	7-8-96-2	7-10-96-1	7-11-96-1	6-17-96-4
La	10.13	11.67	11.30	8.72	5.14	7.88	9.24	6.74	8.72
Ce	23.37	25.25	23.11	20.03	12.02	18.13	20.89	15.22	19.43
Pr	3.12	3.34	3.06	2.77	1.65	2.43	2.81	2.05	2.67
Nd	14.73	15.47	14.62	13.08	8.34	11.77	13.11	9.46	12.06
Sm	4.11	4.18	3.95	3.60	2.51	3.24	3.69	2.66	3.34
Eu	1.21	1.31	1.22	1.21	0.91	1.08	1.17	0.96	1.11
Gd	4.93	5.11	4.84	4.65	3.29	4.10	4.53	3.20	4.08
Tb	0.84	0.86	0.85	0.79	0.61	0.69	0.78	0.55	0.73
Dy	5.40	5.49	5.38	5.05	3.76	4.61	5.04	3.69	4.51
Ho	1.15	1.17	1.08	1.05	0.81	0.98	1.02	0.85	1.01
Er	3.48	3.52	3.40	3.18	2.38	2.87	3.07	2.35	2.74
Tm	0.51	0.52	0.50	0.47	0.36	0.43	0.47	0.36	0.42
Yb	3.43	3.42	3.23	2.92	2.42	2.90	3.18	2.29	2.73
Lu	0.52	0.54	0.51	0.48	0.37	0.45	0.48	0.35	0.43
Hf	2.84	2.78	2.55	2.31	1.65	2.22	2.65	1.93	2.37
Ta	0.50	0.46	0.43	0.39	0.19	0.30	0.54	0.51	0.84
Th	1.84	2.04	1.94	1.35	0.90	1.50	1.67	1.29	1.59
Nb	4.5	5.3	5.2	3.9	2.8	3.6	5.0	4.5	5.3

Field no.	8-10-96-2	8-15-96-1	8-14-96-1-2	7-25-96-1	6-12-96-1	6-25-96-1	GD8816	GD8847	ME-94-22
La	5.82	4.56	11.71	4.14	8.75	6.59	9.1	10.3	6.20
Ce	12.96	11.06	26.41	10.40	19.12	15.70	21	17	15.00
Pr	1.92	1.56	3.47	1.54	2.54	2.16			
Nd	9.42	7.88	15.50	8.38	11.36	10.90	10	13	10.00
Sm	2.76	2.22	4.10	2.74	3.06	3.31	3.01	2.66	2.73
Eu	1.00	0.83	1.20	0.96	1.02	1.12	0.95	0.87	1.04
Gd	3.54	2.77	4.68	3.72	3.72	4.20			
Tb	0.62	0.48	0.78	0.65	0.63	0.75	0.8	0.6	0.60
Dy	3.98	3.13	4.89	4.22	3.95	4.77			
Ho	0.82	0.69	1.01	0.90	0.82	1.03			
Er	2.47	2.02	3.07	2.77	2.51	3.05			
Tm	0.37	0.30	0.44	0.41	0.39	0.47			
Yb	2.34	1.95	3.04	2.72	2.51	2.99	2.8	2.33	2.63
Lu	0.39	0.29	0.46	0.40	0.41	0.48	0.43	0.35	0.41
Hf	1.78	1.27	2.54	1.74	1.57	2.17			1.80
Ta	0.83	0.30	0.95	0.43	0.24	0.36			0.50
Th	1.00	0.82	2.14	0.54	1.57	1.16			1.10
Nb	3.0	2.1	5.2	4.7	2.4	3.9			

Several researchers have shown that major elements such as K_2O, Na_2O, MgO, CaO, and SiO_2 and trace elements such as Rb, Sr, and Ba are mobile during seawater-influenced metamorphism (e.g., Humphris and Thompson, 1978; Mottl, 1983; Wilson, 1989). In an effort to evaluate major element mobility, we utilized a method that calculates a numerical index of alteration $= 100[(MgO + K_2O)/(MgO + K_2O + CaO + Na_2O)]$ where indices of 36 ± 8 represent relatively unaltered rocks (Hashiguchi et al., 1983). The index of alteration for the 53 Mount Norris Intrusive Suite metadiabase samples ranges from 31.3 to 42 (average = 35.6; median = 35.5; standard deviation = 2.8). Thus, none of the metadiabases falls outside the unaltered envelope in this method.

An alternative method to evaluate mobility is to plot elements against a high field strength element (HFSE), such as Zr, not thought to be mobile during metamorphism (e.g., Floyd and Winchester, 1978; Pearce and Norry, 1979; Rollinson, 1993). Strong linear trends on such diagrams indicate relative immobility of the element. Key elements—Th, Nb, Ce, and Y—used in petrogenetic interpretations were plotted against the immobile element Zr in order to evaluate their mobility (Fig. 6). Although there is some scatter in the plots, a strong positive linear trend for the majority of the samples probably directly reflects the original igneous fractionation processes. If metasomatic alteration significantly affected these HFSEs, one would not expect such a distinct linear pattern for these incompatible elements. Furthermore, as seen in the following sections, the geochemical groups maintain their integrity throughout the various discrimination methods.

Th is thought to be mobile under some circumstances (Mélançon et al., 1997) but is an important chemical discriminant. In the Mount Norris Intrusive Suite samples, it forms linear trends with Zr (Fig. 6) and also with Ce and Nb (not shown); furthermore, there is no correlation between LOI and Th, indicating no systematic variation with alteration. Finally, in discrimination diagrams used in later sections, Th plots in tight groupings typical of petrogenetically related igneous rocks. Thus, we consider Th to be have been relatively immobile during metamorphism of the Mount Norris Intrusive Suite rocks.

Classification

SiO_2 concentrations in the metadiabases range from 48% to 56%, indicating that the rocks may be basaltic to basaltic andesite in composition (Table 1). Because SiO_2 is mobile during metamorphism, other classification schemes using immobile elements are more reliable for metavolcanic rocks. Accordingly, we use the Zr/TiO_2 vs. Nb/Y diagram (Floyd and Winchester, 1978) and the alkali index vs. Al_2O_3 plot (Middlemost, 1975) to classify the Mount Norris Intrusive Suite samples (Figs. 7A, 7B). The general coherence of the Mount Norris samples in these classification diagrams suggests that these elements have not been greatly affected by metamorphic alteration; hence the diagrams may be used to determine igneous origins. From the Zr/TiO_2 vs. Nb/Y diagram, we conclude that the Mount Norris samples are subalkaline basalts (Fig. 7A). Subalkaline basalts can be further classified as tholeiitic or calc-alkaline basalts. Clearly, the Mount Norris Intrusive Suite samples are tholeiitic basalts (Fig. 7B).

Geochemical Characteristics of Mount Norris Intrusive Suite Rocks

Although all samples from the suite are basalts or basaltic andesites, there is some variation in major and trace element chemistry. Samples range from fairly primitive basalts (MgO = 8.5%, Ni = 160 ppm, Cr = 400 ppm) to fractionated basalts (MgO = 4.3%, Ni = 35 ppm, Cr = 50 ppm). Furthermore, there are systematic trends from the most primitive to the most fractionated samples; specifically, Ni, Cr, and Ca decrease with decreasing MgO, whereas Ti, P, Fe, Zr, and Y increase. These trends are typical of fractionation of early-formed minerals, such as olivine, pyroxene, and/or plagioclase, from basaltic magmas.

The rare earth element abundances are also typical of basalts; light rare earth element (LREE) abundances are between 20 and 40 times chondritic abundances, whereas heavy rare earth elements (HREEs) range from 10 to 20 times chondritic abundances (Fig. 8A). The REE patterns show enrichment in LREEs, display flat HREE abundances, and have slight negative Eu anomalies (Fig. 8A). The negative Eu anomalies likely reflect the fractionation of plagioclase. Relative to mid-ocean ridge basalts (MORBs), the Mount Norris Intrusive Suite samples have highly irregular LILE

TABLE 3. ANALYTICAL UNCERTAINTY IN CHEMICAL ANALYSES

	MRG-1‡	*MRG-1*§	*ΔMRG-1*# (%)	*BCR-2*‡	*BCR-2*§	*ΔBCR-2*# (%)
ICP-OES†—Middlebury College						
SiO_2	40.07	39.86	*2*	54.72	54.1	*0.3*
TiO_2	3.71	3.74	*2*	2.26	2.26	*1.0*
Al_2O_3	8.36	8.62	*3*	13.33	13.5	*0.4*
FeO	18.01	17.99	*3*	14.05	13.8	*0.5*
MnO	0.16	0.17	*4*	0.20	0.2	*0.7*
MgO	13.34	13.68	*1*	3.53	3.59	*0.3*
CaO	14.57	14.98	*2*	6.98	7.12	*0.4*
Na_2O	0.68	0.72	*3*	3.13	3.16	*2.0*
K_2O	0.16	0.17	*6*	1.74	1.79	*3.0*
P_2O_5	0.06	0.06	*5*	0.35	0.35	*0.5*
Sc	53.3	54	*5*	32	33	*3*
V	549	530	*2*	413	416	*3*
Co	90.2	90	*6*	39	37	*3*
Cr	472	475	*4*	17	18	*4*
Cu	140	135	*4*	17	19	*9*
Ni	202	195	*2*			
Sr	280	265	*4*	342	346	*1*
Y	15.6	16	*8*	34	37	*4*
Zr	95.5	105	*10*	165	188	*11*
Ba	47.9	50	*5*	678	683	*2*
	BHVO-1‡	*BHVO-1*§	*BHVO-1*# (%)	*MRG-1*‡	*MRG-1*§	*MRG-1*# (%)
ICP-MS—Union College						
Rb	10.84	11	*4*	8.66	8.5	*4*
Sr	391.09	403	*2*	282.28	266	*3*
Y	26.99	27.6	*4*	13.84	14	*5*
Zr	178.19	179	*2*	109.67	108	*4*
Nb	18.96	19	*2*	20.02	20	*4*
Ba	141.26	139	*2*	52.92	61	*9*
La	16.04	15.8	*4*	9.28	9.8	*8*
Ce	38.91	39	*3*	25.98	26	*6*
Pr	5.40	5.7	*5*	3.80	3.4	*7*
Nd	25.52	25.2	*4*	18.66	19.2	*5*
Sm	6.27	6.2	*4*	4.24	4.5	*6*
Eu	2.01	2.06	*5*	1.49	1.39	*7*
Gd	6.32	6.4	*5*	4.11	4	*5*
Tb	0.95	0.96	*3*	0.57	0.51	*5*
Dy	5.13	5.2	*3*	2.97	2.9	*6*
Ho	0.96	0.99	*5*	0.53	0.49	*7*
Er	2.47	2.4	*4*	1.19	1.12	*5*
Tm	0.33	0.33	*4*	0.14	0.11	*6*
Yb	2.01	2.02	*4*	0.82	0.6	*8*
Lu	0.30	0.291	*4*	0.12	0.12	*8*
Hf	4.46	4.38	*3*	3.66	3.76	*3*
Ta	1.21	1.23	*2*	0.82	0.8	*3*
Th	1.15	1.08	*2*	0.83	0.93	*4*

†Analyses recalculated on a volatile-free basis.
‡Average of eight analyses on rock standards.
§Recommended values (Canmet Report 79-35, and U.S. Geological Survey reference sheets).
#Precision (% relative standard deviation of replicate analyses).

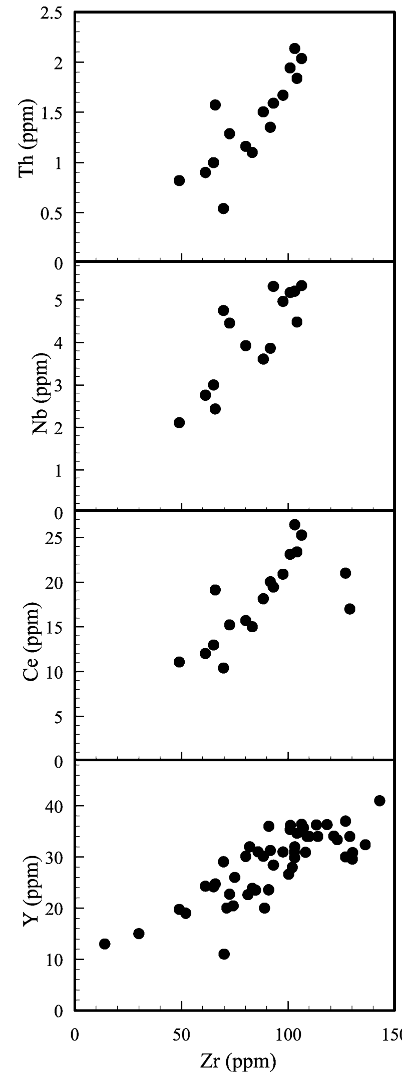


Figure 6. Covariation diagrams of relatively immobile elements for the Mount Norris Intrusive Suite metadiabases: Th, Nb, Ce, and Y vs. Zr.

(large ion lithophile element) abundances, which are indicative of significant metasomatic alteration; also, the samples have distinctive negative Ta and Nb anomalies and generally flat patterns, near or above unity, for elements P to Yb in Figure 8B.

Tectonic Environment Inferred from Geochemistry

Geochemical fingerprints in ancient mafic rocks are not always diagnostic, but can be suggestive, of tectonic environment. In this section, normalized-element diagrams and tectonic-discrimination diagrams are used to suggest a tectonic environment that is shown herein to be consistent with the geology of the region.

Mafic rocks with slightly LREE-enriched patterns, as shown by the Mount Norris Intrusive Suite rocks, can be found in various tectonic settings such as backarc basins (BAB; Fig. 8A) or, more broadly, supra–subduction zone (SSZ) basins, continental rifts, and certain enriched mid-ocean ridge (E-MORB) environments. On MORB-normalized extended-element diagrams (e.g., Fig. 8B), patterns with negative Ta-Nb anomalies relative to adjacent Th and Ce, however, suggest involvement of a mantle source that has been affected by subduction. Marginal basins (interarc, backarc, forearc) are perhaps the most common environment for eruption of volcanic rocks with such characteristics. Specifically, basalts from marginal basins are chemically similar to mid-ocean ridge basalts, except many marginal-basin basalts show enrichment in LILEs, depletion in Al and Ti, negative Nb-Ta anomalies, and higher $^{87}Sr/^{86}Sr$ ratios for a given $^{143}Nd/^{144}Nd$ ratio relative to MORB (e.g., Keller et al., 1992; Hawkins, 1995; Leat et al., 2000). In detail, marginal-basin basalts are not all alike but rather exhibit a range of compositions. For example, in the northern section of the Mariana Trough, basalts vary from MORB-like in mature spreading areas in southern regions to those indistinguishable

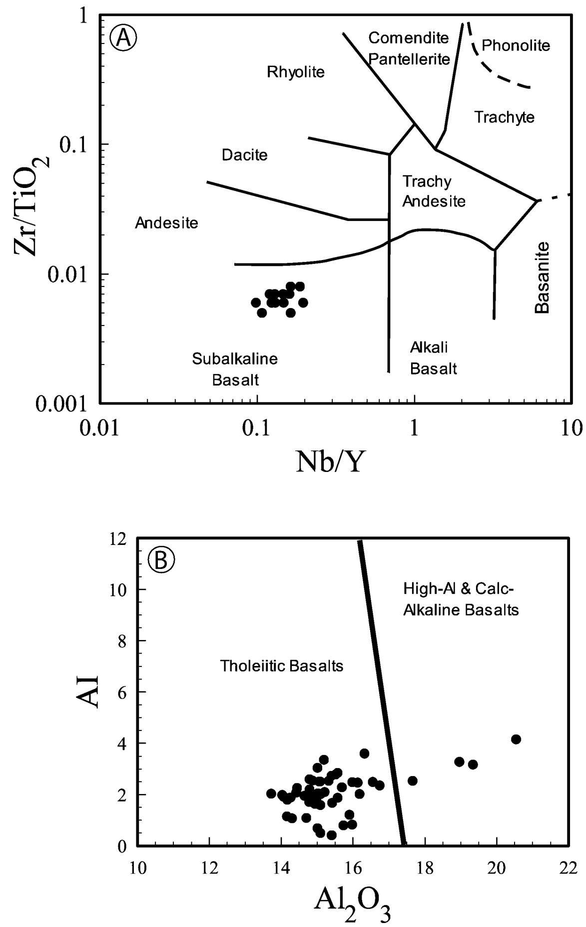


Figure 7. (A) Mount Norris Intrusive Suite data plotted on Zr/TiO_2 vs. Nb/Y discrimination diagram (Floyd and Winchester, 1978). Note that all samples are subalkaline basalts. (B) In an alkali index [AI = $(Na_2O + K_2O)/((SiO_2–43) \times 0.17)$] vs. Al_2O_3 discrimination diagram (Middlemost, 1975), the Mount Norris Intrusive Suite rocks mostly plot as tholeiitic basalts.

from island-arc basalts in the north where spreading is incipient (Gribble et al., 1996, 1998). Likewise, in the Scotia Sea, backarc volcanic compositions vary in detail with location along the spreading-center segments, but in general, samples have slightly enriched LREE abundances (relative to their HREE abundances) with small or no Ta-Nb anomalies (Leat et al., 2000). One factor called on to explain compositional variation of backarc basalts is the amount of subduction component added to depleted subarc mantle; the subduction component can be delivered in fluids or as partial melts from the subducting plate. The amount of the subduction component added may be controlled by the proximity of the subducting plate to the site of marginal-basin spreading, which in turn may be related to the maturity of the basin; early stages of spreading may produce magmas indistinguishable from island-arc volcanic rocks, whereas once spreading becomes established, subduction-related components become less abundant.

Following from the foregoing discussion, the negative Nb-Ta anomalies coupled with the flat MORB-normalized abundances of the elements from P to Yb (in Fig. 8B) in Mount Norris Intrusive Suite metadiabases suggest the tectonic environment of a SSZ basin. In fact, their overall concentrations of HFSEs and negative Ta-Nb anomalies are similar to those characteristics in many basaltic rocks from western Pacific marginal basins (Figs. 8A and 8B). Neither the Nb-Ta anomalies nor the LREE enrichment is extreme. This fact may indicate that the metadiabases formed in a mature rather than an incipient SSZ basin (Allan and Gorton, 1992) but not so mature as to produce MORBs without any subduction-zone signature.

Tectonic-discrimination plots of Mount Norris Intrusive Suite samples are also consistent with their origin in a marginal basin. In most tectonic-discrimination diagrams, modern marginal-basin basalts plot in the same field as mid-ocean ridge basalts or in the overlap fields between island-arc tholeiites and marginal-basin basalts. Rocks from the Mount Norris Intrusive Suite also consistently plot in MORB or BABB (backarc-basin basalt) fields in many discrimination diagrams (e.g., Fig. 9A). A small number of samples fall in the arc field, further suggesting a connection with a subduction environment.

The Ta/Yb vs. Th/Yb diagram (Pearce, 1983) can be used to more convincingly show the presence of a subduction component in volcanic rocks. Mafic samples derived from the mantle and unaffected by later processes fall within a mixing zone (array) between a depleted-mantle source (DMS = MORB) and an enriched-mantle source (EMS = oceanic-island basalt [OIB] source) (two parallel lines in Fig. 9B). Subsequent igneous processes and/or the introduction of contaminants will drive a sample away from the mixing array (Fig. 9B): S represents the direction a magma composition moves by addition of a subduction component, C represents the addition of a continental-crust component, f represents fractional crystallization, and W represents within-plate source variations. Although two of the Mount Norris Intrusive Suite metadiabase samples plot within the mixing array, the majority of samples plot above the mixing array in a region that may indicate the addition of a subduction component in either an oceanic island-arc or active continental-margin tectonic setting. The fact that some samples plot in the oceanic island-arc (OIA) field or in the overlap zone between the oceanic island-arc field and active continental-margin (ACM) field makes a subduction component more likely and thus consistent with a backarc-basin origin for the metadiabases.

Summary of Geochemistry

The Mount Norris Intrusive Suite metadiabases in the Rowe-Hawley belt of northern Vermont are tholeiitic, basaltic, hypabyssal intrusions. Because it is difficult to distinguish chemically among mid-ocean-ridge, backarc-basin, and island-arc basalts, more than one interpretation of the tectonic origin of the Mount Norris Intrusive Suite may be possible. However, the following geochemical criteria strongly indicate an origin in a supra–subduction zone: (1) negative Ta-Nb anomalies on MORB-normalized spider diagrams, (2) displacement along a subduction trajectory away from a normal-mantle mixing trend in a Ta/

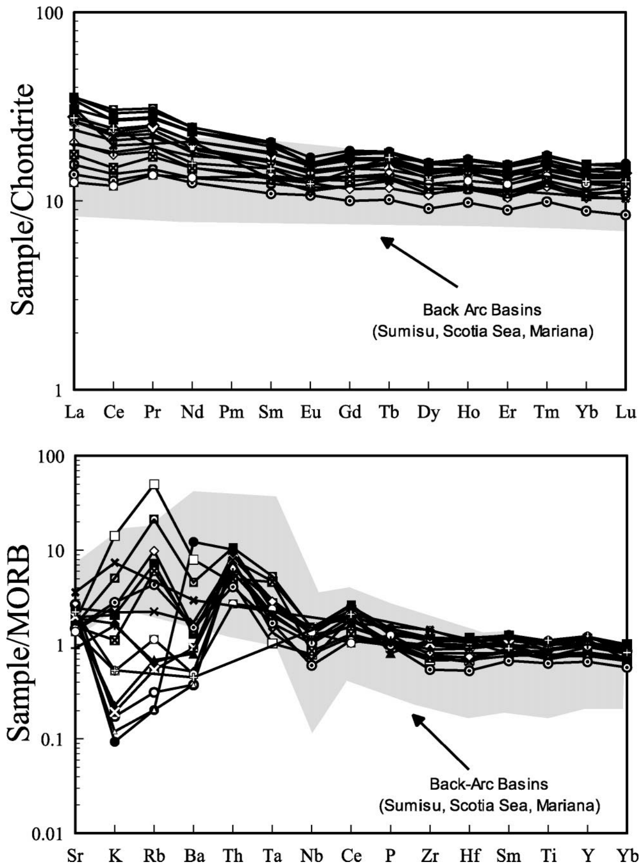


Figure 8. Chondrite-normalized (Sun et al., 1980) REE diagram of Mount Norris Intrusive Suite metadiabases. MORB-normalized (Pearce, 1982) spider diagram of Mount Norris Intrusive Suite metadiabases. Shaded areas show range of values for selected samples from several marginal basins: Scotia Sea (Leat et al., 2000), Mariana backarc-basin region (Hawkins et al., 1990; Gribble et al., 1996, 1998), and Sumisu basin (Fryer et al., 1990; Hochstaedter et al., 1990).

Yb vs. Th/Yb discrimination diagram, and (3) slight LREE enrichment as is common in many backarc-basin environments.

REGIONAL GEOCHEMICAL AND TECTONIC CONTEXT

The Mount Norris Intrusive Suite metadiabasic intrusions can be correlated to rocks north and south along strike. To the south in central Vermont, metadiabasic intrusions occur in the Rowe-Hawley belt (Cua, 1989; Martin, 1994); however, limited geochemical work precludes detailed comparison with the Mount Norris Intrusive Suite. To the north, the Coburn Hill Volcanics in northern Vermont (Gale, 1980; Evans, 1994) and the Bolton Igneous Group in southern Quebec (Mélançon et al., 1997) may be equivalents, albeit extrusive rather than intrusive. If the Mount Norris Intrusive Suite metadiabases are considered collectively with these along-strike correlatives, then the magmatic episode was regionally extensive, and not merely an isolated igneous event, in the northeastern Appalachians. It is thus important to examine these correlative rocks and make the case on field, geochemical, and geochronologic grounds that the Mount Norris Intrusive Suite should be considered as equivalent.

Coburn Hill Volcanics

The Coburn Hill Volcanics, situated on the eastern side of the Rowe-Hawley belt in northern Vermont (Fig. 2), are a thick sequence of pillowed greenstones of presumed Ordovician age in stratigraphic contact with black phyllites that locally have breccia horizons. These mafic rocks were originally called the Bolton Igneous Group (Doll, 1951), but later were mapped as the Coburn Hill Member of the Moretown Formation and were stratigraphically correlated with the Bolton Igneous Group of southern Quebec (Cady et al., 1963). Doolan et al. (1982) demonstrated that both the black phyllites and the associated mafic rocks of the Coburn Hill Volcanics in northern Vermont can be mapped continuously northward into the respective black ‘‘slates’’ of the St. Daniel Formation and Bolton Igneous Group mafic metavolcanic rocks of southern Quebec (Fig. 2). Although the Mount Norris Intrusive Suite metadiabases are found within lithotectonic packages that lie to the south and west of the Coburn Hill Volcanics and associated black phyllites, all these rocks are thought to be correlative (Kim et al., 1999; Kim and Coish, 2001) because (1) Coburn Hill Volcanics and interlayered Cram Hill Formation black phyllites have been mapped continuously southward into black phyllites (also classified as Cram Hill) that contain metadiabases (Fig. 2) and (2) bluish-gray metasiltstones found in slices of the western member of the Moretown Formation to the west are also intimately associated with Cram Hill black phyllites to the east (Gale, 1980); the slices of the western member of the Moretown Formation probably root to the east in the Cram Hill Formation.

On the basis of major and trace element geochemistry, Gale (1980) and Evans (1994) determined that the Coburn Hill Volcanics (called Bolton volcanics by Evans) were tholeiitic to calc-alkaline basalts. A representative tectonic-discrimination diagram using their data shows that the mafic rocks of the Coburn Hill Volcanics have an ocean-floor (MORB or BABB) affinity (Fig. 10A). Evans (1994) showed that these metabasalts have LREE-enriched patterns and spider-diagram patterns with significantly elevated Th (7–8 times MORB) abundances, elevated Ce (~2 times MORB), and flat patterns from P to Yb (in Fig. 10B). Their trace element patterns (Evans, 1994) are nearly identical to those of the Mount Norris Intrusive Suite metadiabases (Fig. 10B); however, Nb was not analyzed by Evans (1994) and thus a critical component cannot be compared. Nevertheless, on the Ta/Yb vs. Th/Yb diagram (Fig. 9B), the two samples of Evans (1994) plot within the grouping of Mount Norris Intrusive Suite samples, near the overlap zone between oceanic island arcs and active continental margins.

Bolton Igneous Group

As previously discussed, the mafic rocks of the Bolton Igneous Group of southern Quebec are a northern continuation of the Coburn Hill Volcanics of northern Vermont (Gale, 1980; Doolan et al., 1982) (Fig. 2). The contact re-

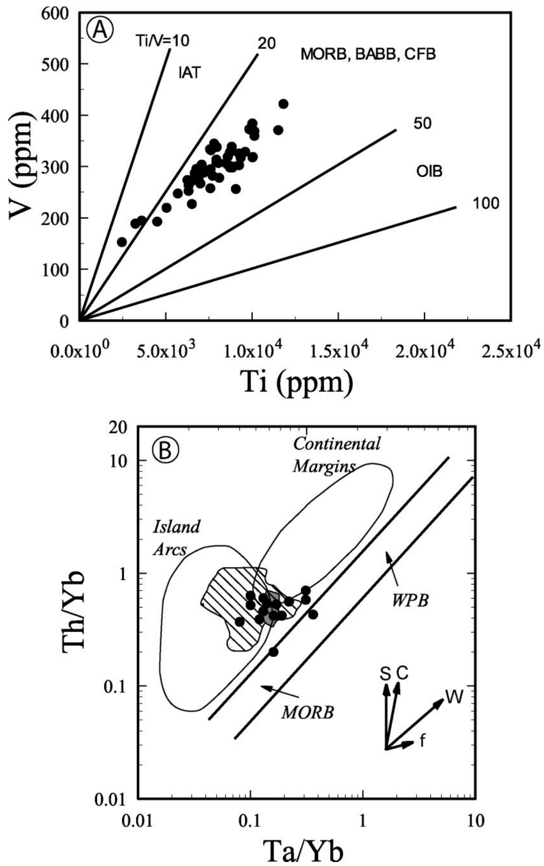


Figure 9. Tectonic-discrimination diagrams for Mount Norris Intrusive Suite rocks. (A) Ti vs. V diagram (Shervais, 1982). (B) Ta/Yb vs. Th/Yb diagram (Pearce, 1983) plotting study-area rocks and showing the field of selected samples from the Bolton Igneous Group of southern Quebec (Mélançon et al., 1997) (ruled area) and the field of Coburn Hill Volcanics (Evans, 1994) (darker area) for comparison. WPB—within-plate basalts, IAT—island-arc tholeiite, MORB—mid-ocean ridge basalt, BABB—backarc-basin basalt, CFB—continental flood basalt. Arrows show direction of change that results from contamination by the following components: S—subduction, C—continental crust, W—within-plate source, f—fractional crystallization.

lationship between the Bolton mafic rocks and the surrounding St. Daniel Formation mélange can be stratigraphic or tectonic. Doolan et al. (1982) observed conformable contacts with interbedding of mafic metavolcanic rocks and black phyllites over short distances. Because the Bolton mafic bodies have highly sheared contacts with the surrounding St. Daniel Formation mélange and because they lack feeder dikes, Mélançon et al. (1997) concluded that the Bolton mafic rocks were fault-bounded tectonic blocks. Furthermore, ultramafic slivers juxtaposed with Bolton mafic bodies also suggest tectonic contacts (Stanley et al., 1984).

Although there are no direct age constraints on the mafic rocks of the Bolton Igneous Group, there are isotopic and stratigraphic age constraints on units adjacent to the St. Daniel Formation mélange. The mélange unconformably overlies the Thetford Mines ophiolite (e.g., Schroetter et al., 2001, 2002) and stratigraphically (unconformably) underlies the Late Ordovician Magog Formation (e.g., Doolan et al., 1982). Because a 479 Ma U-Pb zircon age on a plagiogranite constrains the age of crystallization of the Thetford Mines ophiolite (Dunning and Pedersen, 1988) and a 477 $\pm$ 5 Ma Ar/Ar age on amphibole in the metamorphic sole of the ophiolite constrains the date of obduction (Whitehead et al., 1996), the base of the unconformably overlying the St. Daniel Formation mélange is ostensibly younger than 477 Ma. Biostratigraphic control (graptolites) suggests that the base of the Magog is Caradocian or approximately pre–Late Ordovician (e.g., Berry, 1962). Thus, the Bolton mafic rocks are essentially bracketed in the age interval from 479 to 458 Ma. If the correlation between the Mount Norris Intrusive Suite metadiabases and the Bolton Igneous Group and Coburn Hill Volcanics is correct, then the Mount Norris is also pre–Late Ordovician. The case for correlation is strengthened by comparing the geochemistry of the Bolton mafic rocks with the Mount Norris diabases.

Mafic rocks of the Bolton Igneous Group of Quebec (Mélançon et al., 1997) exhibit trace element geochemical signatures nearly identical to those of the northern Vermont Mount Norris Intrusive Suite metadiabases (Fig. 10B). The Bolton mafic samples are LREE enriched, and their MORB-normalized spider diagrams show a negative Ta-Nb anomaly relative to Th and Ce. Ti vs. Zr (not shown) and Zr vs. Zr/Y tectonic-discrimination diagrams also show that the Bolton Igneous Group rocks plot in MORB or BABB fields. Figure 9B—a Ta/Yb vs. Th/Yb diagram—shows that Bolton Igneous Group mafic rocks also plot above the mantle mixing array along a subduction-signature trend suggesting a SSZ basin origin for the Bolton Igneous Group. Also on the Ta/Yb vs. Th/Yb diagram, the field defined by the Bolton samples overlaps with Mount Norris Intrusive Suite and Coburn Hill Volcanics samples.

Mélançon et al. (1997) interpreted the mafic rocks of the Bolton Igneous Group to have formed as transitional MORB in the Iapetus Ocean, but acknowledged that they could also be backarc related. The St. Daniel Formation mélange is currently interpreted to have been deposited in a ‘‘piggyback’’ basin unconformably overlying the Thetford Mines ophiolite (e.g., Schroetter et al., 2001, 2002; Tremblay et al., 2001) rather than in an accretionary-prism tectonic setting. This interpretation favors a backarc or, more generally, an SSZ basin origin for the Bolton Igneous Group.

Summary of Correlations

The Mount Norris Intrusive Suite intrusions can be correlated with the Coburn Hill Volcanics in northern Vermont and the Bolton Igneous Group in southern Quebec on the basis of similar stratigraphy and geochemistry. First, the Mount Norris is correlated with the Coburn Hill mainly because of their identical geochemistry but also because they both lie in the highest structural slices in the region. Second, the Coburn Hill can be directly linked to the Bolton rocks because outcrops can be traced between the two. Third, black phyllite units interbedded with the Coburn Hill Volcanics in the Cram Hill Formation of Vermont

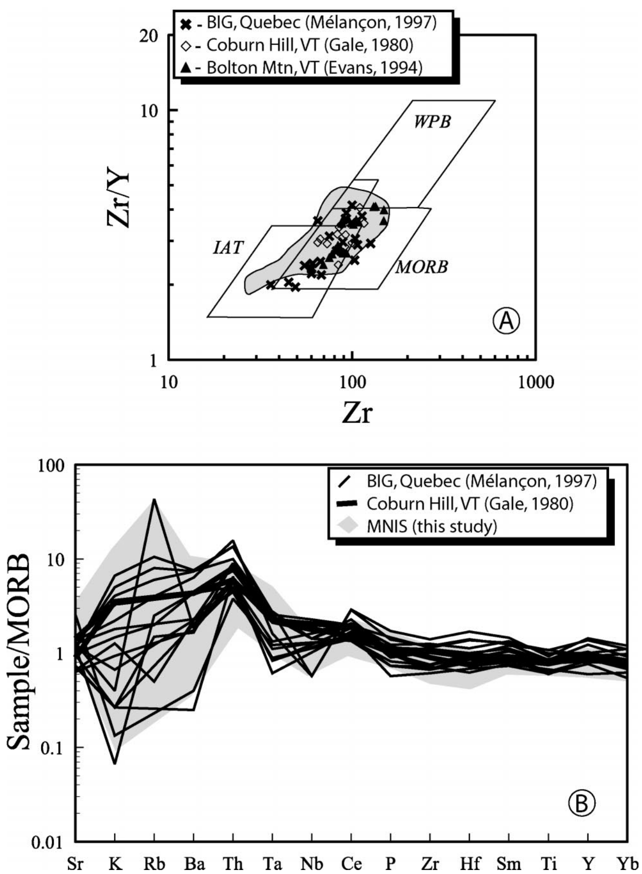


Figure 10. (A) Zr vs. Zr/Y diagram used to compare Mount Norris Intrusive Suite data (gray field) with Coburn Hill Volcanics data (Gale, 1980; Evans, 1994) and Bolton Igneous Group (BIG and Bolton Mountain) (Mélançon et al., 1997). (B) MORB-normalized (Pearce, 1982) spider diagram plotting selected Bolton Igneous Group (BIG) mafic rocks (Mélançon et al., 1997) compared to Bolton Mountain, Vermont, data (Evans, 1994) and the Mount Norris Intrusive Suite (MNIS) rocks.

can be traced into phyllites interbedded with the Bolton in the St. Daniel Formation mélange of southern Quebec. Finally, by using age information from southern Quebec, we infer that the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics were formed in the Middle Ordovician, between the obduction of the Thetford Mines ophiolite at 477 Ma (Whitehead et al., 1996) and the deposition of the base of the Magog Formation at ca. 458 Ma (Berry, 1962).

TECTONIC MODELS

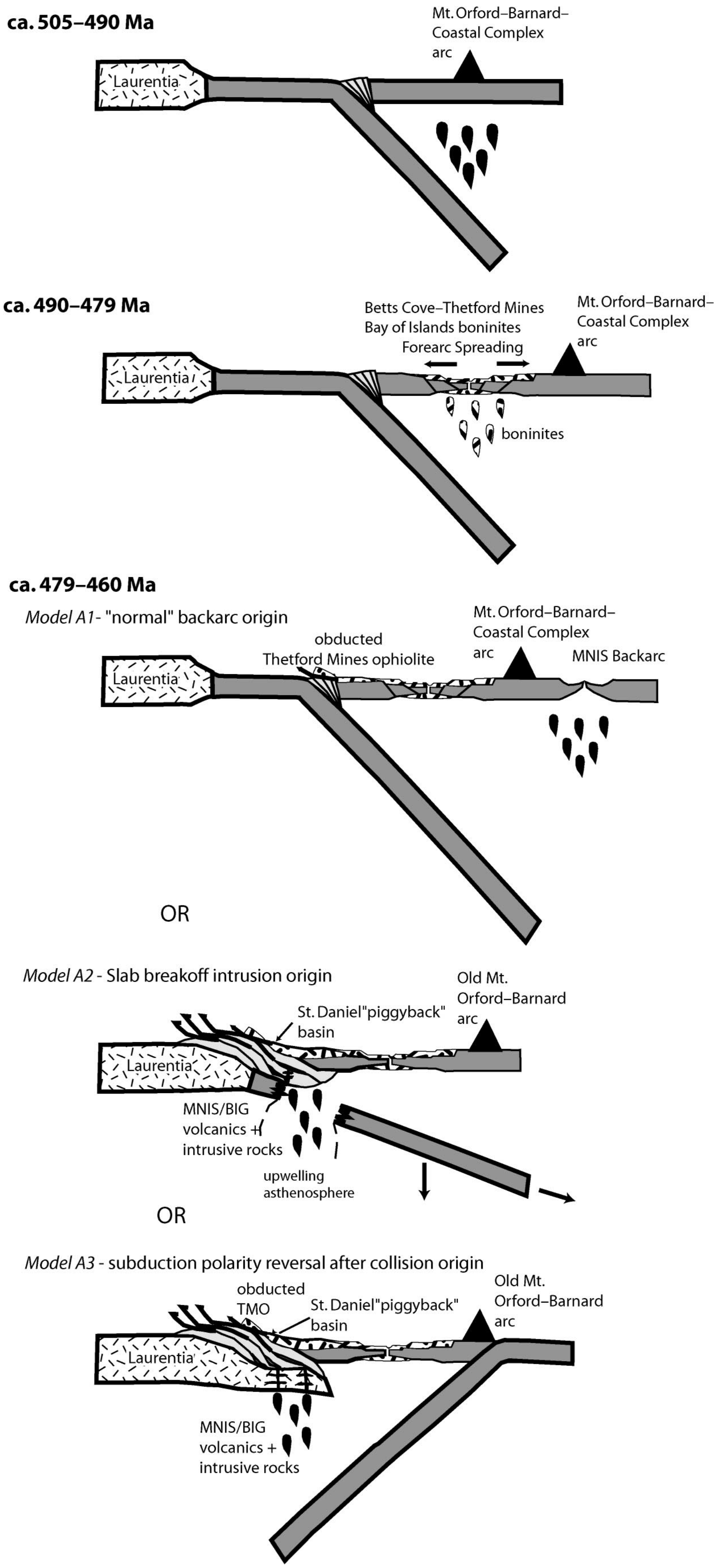
Regional Considerations

It is generally accepted that the Late Proterozoic–early Paleozoic Iapetus Ocean began to close through subduction in the Late Cambrian. This closure culminated in an arc-continent collision by which parts of the passive-margin sequence, Laurentian basement, early rift-facies rocks, and oceanic and supra–subduction zone terranes were thrust westward over autochthonous passive-margin sedimentary rocks—i.e., what is called the Taconic orogeny. Most existing tectonic models (e.g., Stanley and Ratcliffe, 1985; Pinet and Tremblay, 1995; Karabinos et al., 1998) begin with eastward-directed subduction; however, there is controversy over the duration and polarity of the subduction zone(s) that led to the Taconic orogeny. The controversy focuses on whether there was protracted evolution of a single arc system (Bronson Hill arc) above an east-dipping subduction zone (Stanley and Ratcliffe, 1985) or two independent arcs wherein the first arc (Early Ordovician Shelburne Falls arc) developed above an east-dipping subduction zone and the second arc (Late Ordovician Bronson Hill arc) developed above a west-dipping subduction zone (Karabinos et al., 1998). In the single-arc model, arc volcanism lasted from ca. 500 Ma until 440 Ma and continued past the time of arc-continent collision (ca. 450 Ma) (Stanley and Ratcliffe, 1985; Ratcliffe et al., 1998). In the two-arc model, eastward-directed subduction resulted in arc-continent collision at ca. 470–460 Ma and was followed by a reversal in subduction polarity and construction of the Bronson Hill arc above a west-dipping subduction zone (Karabinos et al., 1998). Alternatively, other workers have proposed that the Bronson Hill arc was built off the coast of Gondwana on the east side of Iapetus and accreted to Laurentia in the Late Ordovician (van Staal et al., 1998). Following accretion of the arc(s) to Laurentia in the Late Ordovician, the Iapetus Ocean continued to close and eventually Laurentia collided with the Avalon microcontinent in the Devonian (Acadian orogeny).

Although ''Taconic'' arc magmatism in the northeastern Appalachians spanned from ca. 505 to ca. 440 Ma, there appear to have been different pulses of arc magmatism that correspond to different tectonic events. The oldest arc magmas, generated from the initiation of eastward-directed subduction outboard of Laurentia (Baie-Verte oceanic tract of van Staal et al. [1998]), ranged from ca. 505 Ma to ca. 490 Ma and include the Mount Orford ophiolite at 504 Ma (David and Marquis, 1994), the 496 Ma Barnard Gneiss of the Rowe-Hawley belt in southern Vermont (Ratcliffe et al., 1998), and the Coastal Complex of Newfoundland at 505 Ma (Jenner et al., 1991) (Fig. 11). Several workers have proposed that after the arc system became established, a major episode of forearc extension occurred to generate boninitic magmas (Hibbard, 1983; Bédard and Kim, 2002; Kim and Jacobi, 2002). Examples are found in the Betts Cove ophiolite (488 Ma—Dunning and Krogh, 1985; Coish, 1989; Bédard et al., 1998), Bay of Islands Complex (484 Ma—Jenner et al., 1991), and Thetford Mines ophiolite (479 Ma—Dunning and Pedersen, 1988) (Fig. 11). After obduction of the Thetford Mines ophiolite at ca. 477 Ma (Whitehead et al., 1996) and during deposition of the overlying St. Daniel Formation mélange in a proposed ''piggyback'' basin environment (Schroetter et al., 2001, 2002), the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics story begins.

Tectonic Models for Origin of Mount Norris Intrusive Suite

Any tectonic model to explain the origin of the Mount Norris Intrusive Suite should account for (1) production of magma with a



supra–subduction zone (backarc) geochemical signature, (2) intrusion of the magma into thrust slices containing already-deformed lithologies, and (3) transportation of the Mount Norris Intrusive Suite westward along thrust sheets. Given these restrictions and the foregoing regional magmatic and tectonic context, we present three possible models to explain the origin of the Mount Norris Intrusive Suite: (1) intrusion in a backarc associated with east-dipping subduction (Fig. 11, model A1), (2) intrusion following slab breakoff (collisional delamination) (Fig. 11, model A2), and (3) intrusion in a backarc associated with west-dipping subduction (Fig. 11, model A3). We argue the merits of each model and provide reasons why we prefer model A3.

In model A1, the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics magmas were generated at a backarc spreading center and intruded into Moretown, Stowe, and Cram Hill Formation sedimentary rocks behind an early island arc developed above an eastward-dipping subduction zone (Fig. 11, model A1). Thrust sheets from the backarc would have to be deformed, metamorphosed, and transported westward during arc-continent collision. A shortcoming of model A1 is that field evidence indicates that the Mount Norris Intrusive Suite cuts Taconic foliations internal to the thrust slices that contain them or that have xenoliths of foliated metasedimentary rock indicating that the host rocks were deformed and metamorphosed prior to intrusion. With the exception of the accretionary prism, supra–subduction zone (forearc, arc, and backarc) regions generally behave quite rigidly and are not significantly ductilely deformed prior to orogenesis (Hamilton, 1988). The accretionary prism can be intruded by arc magmas only if significant trench rollback has occurred (C.R. van Staal, 2003, personal commun.) or if the angle of subduction is nearly vertical. Furthermore, prior to arc-continent collision, the arc region is extensional, not compressive (e.g., Hamilton, 1988). Once arc-continent collision begins, however, convergence may be translated from the forearc and trench region to the backarc and result in the formation of backarc/retro-arc thrusts (Silver et al., 1983; Rangin et al., 1995). Only if the Mount Norris Intrusive

Figure 11. Possible models for the tectonic setting of the Mount Norris Intrusive Suite (MNIS) metadiabases (and Bolton Igneous Group [BIG]). See text for discussion.

Suite intruded during backarc thrusting would model A1 be considered viable.

In model A2 of Figure 11, incipient arc-continent collision is followed by subduction of the leading edge of the continent, and ensuing collisional delamination (slab breakoff) as the continental lithosphere is separated from oceanic lithosphere by strong buoyancy differences (von Blanckenburg and Davies, 1995). Delamination of the lithosphere creates space into which asthenospheric mantle flows, melts lithospheric mantle, and generates magmas that intrude the overlying stack of deformed and metamorphosed thrusts. The major strength of the model is that it provides a mechanism for intrusion of magmas into deformed rocks formed by a collisional event. The major flaw is that the magmas produced, at least in the European example of von Blanckenburg and Davies (1995), are highly potassic tholeiitic, calc-alkaline, and alkaline lavas, whereas the Mount Norris Intrusive Suite are unimodal, tholeiitic basalts with a mild subduction-related geochemical signature.

In a third model, the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics represent supra–subduction zone magmas generated above a westward-dipping subduction zone that formed by subduction-polarity reversal following ophiolite obduction at ca. 477 Ma (Fig. 11, model A3). A plausible tectonic sequence would be (1) ophiolite obduction at a Laurentian continental promontory, (2) ''jamming'' of the east-dipping subduction zone by the continent, (3) thrusting and collisional delamination, and (4) breakthrough of a new, outboard, west-dipping subduction zone. The flip in subduction-zone polarity thus transforms the older forearc into a backarc (Fig. 11, model A3) and allows a series of deformed and metamorphosed thrust slices from the earlier arc-continent collision to be intruded by supra–subduction zone magmas. Van Staal et al. (1998) proposed a model like this to explain the Notre Dame arc of west-central Newfoundland whereby an earlier oceanic arc (Baie-Verte oceanic tract; 490–480 Ma) that had originally developed above an east-dipping subduction zone was intruded by a temporally (480–465 Ma) and geochemically distinct generation of arc plutons derived from a second westward-dipping subduction zone. Likewise, Karabinos et al. (1998) proposed a similar model for the evolution of the Shelburne Falls arc in New England. Although there is little published evidence for the suture of a westward-dipping subduction zone in New England or southern Quebec, van Staal et al. (1998) proposed that the Boil Mountain ophiolite of Maine marks this suture along the east side of the Connecticut Valley synclinorium. Furthermore, van Staal et al. (1998) interpreted the Ascot Complex (460–441 Ma) of southern Quebec to represent the arc above this westward-dipping subduction zone. In this scenario, the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics would have formed in a backarc tectonic environment behind the Ascot arc.

Recent geochronologic data from southern Quebec support model A3. Isolated 470–469 Ma granitoids that intruded the peridotite of the Thetford Mines ophiolite (Whitehead et al., 2000) indicate that some felsic magmatism postdated the obduction of the ophiolite at 477 Ma. Whitehead et al. (2000) suggested that these granitoid magmas could have been ''derived by melting of the Laurentian margin over a west-dipping subduction zone or by shear heating at the base of an obducting ophiolite nappe'' (p. 926).

Although we think that each of the tectonic models proposed herein for the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics have merits and pitfalls, we think that the subduction-polarity-reversal model (Fig. 11, model A3) best accommodates the field and geochemical constraints. Variations of this model have been proposed both by van Staal et al. (1998) and Karabinos et al. (1998). Hamilton (1988) elaborated about his work in the Pacific that ''long-continuing, steady-state subduction systems are atypical; that complex sequences of collision, aggregation, reversal, and internal deformation are the rule; and that aggregates of collided bits can be assembled far from their final resting places'' (p. 1518). The subduction-polarity-reversal model allows intrusion of supra–subduction zone magmas into deformed and metamorphosed thrusts slices by turning the former outer forearc region and collision zone into a backarc. In this model, the Mount Norris Intrusive Suite–Bolton Igneous Group–Coburn Hill Volcanics are backarc intrusions and volcanic rocks that formed roughly coevally with the Ascot Complex arc and forearc igneous rocks.

CONCLUSIONS

Five general conclusions result from a detailed field and geochemical study of the Mount Norris Intrusive Suite: (1) Mafic rocks of the Mount Norris Intrusive Suite intruded metamorphosed sedimentary rocks during the Early–Middle Ordovician, probably sometime between 480 and 460 Ma. (2) The intrusions were tholeiitic basalts that were metamorphosed to greenschist facies and deformed to varying degrees; remnant igneous textures are preserved in many samples. (3) Geochemistry indicates that the basalts may have formed in an extensional region (marginal basin) of a supra–subduction zone environment; Nb-Ta anomalies, Th/Yb relationships, and the abundance patterns of REEs indicate that their source was probably depleted mantle modified by a subduction component. (4) Rocks of the Mount Norris Intrusive Suite are correlated with the Coburn Hill Volcanics in Vermont and the Bolton Igneous Group in Quebec on the basis of stratigraphy and geochemistry. (5) Although several tectonic models are plausible, the conclusion that the intrusions are subduction-related tholeiitic basalts that cut metamorphosed and deformed sedimentary rocks leads to a preferred model in which westward-directed subduction resulted in the development of extensional basins in terranes accreted to Laurentia by earlier eastward-directed subduction.

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