3-2000

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## Recommended Citation

J. Dykstra Eusden, Jr., Chris A. Guzofski, Alexander C. Robinson, and Robert D. Tucker, "Timing of the Acadian Orogeny in Northern New Hampshire," The Journal of Geology 108, no. 2 (March 2000): 219-232.

# Timing of the Acadian Orogeny in Northern New Hampshire

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**ABSTRACT**

New U-Pb geochronology constrains the timing of the Acadian orogeny in the Central Maine Terrane of northern New Hampshire. Sixteen fractions of one to six grains each of zircon or monazite have been analyzed from six samples: (1) an early syntectonic diorite that records the onset of the Acadian; (2) a schist, a migmatite, and two granites that together record the peak of the Acadian; and (3) a postkinematic pluton that records the end of the Acadian. Zircon from the syntectonic Wamsutta Diorite gives a $^{207}Pb/^{206}Pb$ age of circa 408 Ma, the time at which the boundary between the deforming orogenic wedge and the foreland basin was in the vicinity of the Presidential Range. This age agrees well with the Emsian position of the northwest migrating Acadian orogenic front and records the beginning of the Acadian in this part of the Central Maine Terrane. We propose a possible Acadian tectonic model that incorporates the geochronologic, structural, and stratigraphic data. Monazite from the schist, migmatite, Bigelow Lawn Granite, and Slide Peak Granite gives $^{207}Pb/^{206}U$ ages, suggesting the peak of Acadian metamorphism and intrusion of two-mica granites occurred at circa 402–405 Ma, the main pulse of Acadian orogenesis. Previously reported monazite ages from schists that likely record the peak metamorphism in the Central Maine Terrane of New Hampshire and western Maine range from circa 406–384 Ma, with younger ages in southeastern New Hampshire and progressively older ages to the west, north, and northeast. Acadian orogenesis in the Presidential Range had ended by circa 355 Ma, the $^{207}Pb/^{235}U$ age of monazite from the Peabody River Granite. From 408 to perhaps at least 394 Ma, Acadian orogenesis in the Presidential Range was typical of the tectonic style, dominated by synkinematic metamorphism, seen in central and southern New Hampshire, Massachusetts, and Connecticut. From no earlier than 394 Ma to as late as 355 Ma, the orogenesis was typical of the style in parts of Maine dominated by postkinematic metamorphism.

## Introduction

The Acadian orogeny is one of the best known tectonic events in the Northern Appalachians. It primarily affected the rocks of central and eastern New England and portions of the Canadian Maritime provinces. The details of the stratigraphy, structure, metamorphism, and plutonism have been studied by legions of researchers. However, the tectonics of the Acadian have always been controversial due to a lack of key lithotectonic assemblies such as ophiolites and blueschists.

In the past 10–15 yr, high-precision geochronology has allowed researchers of the Acadian to more precisely pinpoint the timing of orogenesis and move toward better, but necessarily more complex, tectonic reconstructions. In addition, detailed structural analyses of the complexly deformed rocks, at scales of 1 : 5000 or larger, have revealed complex variations in the sequences and transitions of Acadian deformation throughout the orogen. Though our understanding of Acadian tectonism in portions of the orogen is better, much still remains to be discovered.

In this article we present new U-Pb geochronology that constrains the timing of the Acadian Orogeny in the Presidential Range of northern New Hampshire. The Presidential Range, which includes Mount Washington (the highest peak in the northeastern United States), lies in the heart of the Acadian orogen and within a significant transition zone where the style of Acadian orogenesis changes

Manuscript received May 12, 1999; accepted November 2, 1999.

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[The Journal of Geology, 2000, volume 108, p. 219–232] © 2000 by The University of Chicago. All rights reserved. 0022-1376/2000/10802-0006$01.00

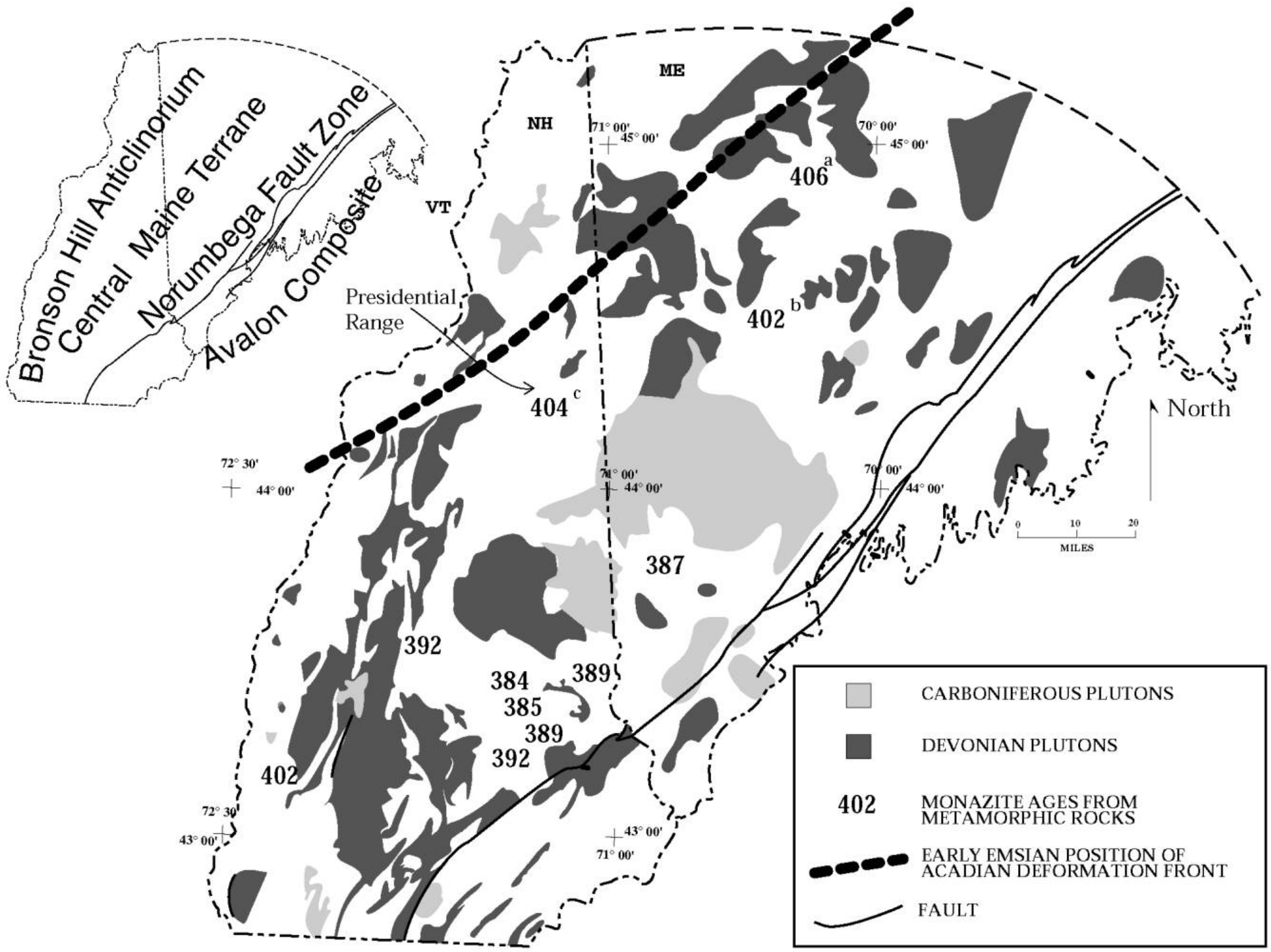


**Figure 1.** Location map showing the study area, Central Maine Terrane, Bronson Hill Anticlinorium, Norumbega Fault Zone, and the Emsian position of the Acadian deformation front from Bradley et al. (1998). Numbers with alpha superscripts (e.g., *404*[c]) are the oldest U-Pb monazite ages in millions of years from Silurian and Devonian schists and migmatites. Key for sources of ages shown by superscript: *a*, Solar et al. (1998); *b*, Smith and Barreiro (1990); *c*, this article; the rest, without superscripts are from Eusden and Barreiro (1988). Pluton ages taken from Lyons et al. (1997) and Bradley et al. (1998).

dramatically. From the Presidential Range south through the rest of New Hampshire, Massachusetts, and Connecticut, the Acadian style is dominated by syntectonic metamorphism and plutonism, whereas to the north in Maine, postkinematic metamorphism and plutonism dominate. These variations are fortunately now exposed because the Acadian orogen in New England is tilted with shallower, upper-crustal rocks exposed to the northeast and deeper, middle-crustal rocks exposed to the southwest.

Most important, the new U-Pb data presented here allow us to assign an absolute age to several phases of deformation and metamorphism previously only relatively known in the Presidential Range. This geochronology is related to our long-term mapping project to redefine the stratigraphy, structure, and metamorphism in the Acadian transition zone. Furthermore, these ages relate well to several recent geochronologic studies from New Hampshire and Maine, enabling us to evaluate (1) the onset of orogenesis as the Acadian deformation front moves through; (2) the diachroneity of timing of peak metamorphism; and (3) the duration of the entire Acadian.

## Geological Setting

The Presidential Range is located on the western flank of the Central Maine Terrane (CMT; fig. 1).

The Silurian and Devonian cover rocks of the CMT correlate with both the Central Maine Basin of Bradley et al. (1998) and the Merrimack Belt of Robinson et al. (1998). The Silurian and Devonian cover rocks of the CMT stretch from Connecticut to New Brunswick and are bounded to the southeast by composite Avalonian rocks along the Maine, New Hampshire, and Massachusetts coasts and to the northwest by the Bronson Hill, Boundary Mountains, and Lobster Mountain anticlinoria (Lyons et al. 1997; Bradley et al. 1998; Robinson et al. 1998). Within the CMT of New Hampshire, the major structural features, from west to east, are the Bronson Hill anticlinorium, the Kearsarge–Central Maine synclinorium, the Central New Hampshire anticlinorium, and the Lebanon antiformal synclinorium (Eusden and Lyons 1993; Lyons et al. 1997).

The CMT contains Silurian metasedimentary cover rocks that are interpreted as an eastward thickening sequence of deepwater turbidites deposited in either a passive margin basin (Moench and Pankiwskyj 1988; Robinson et al. 1998) or a forearc basin associated with a northwest dipping subduction complex (Hanson and Bradley 1989; Eusden et al. 1996*a*; Bradley et al. 1998). The Silurian rocks in contact with and adjacent to the Bronson Hill and Boundary Mountains anticlinoria are thin, nearshore conglomerates and calcareous turbidites of the Clough, Fitch, and portions of the Rangeley Formations. These thicken to the southeast into deeper-water turbidites of the Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations (Hatch et al. 1983; Moench and Pankiwskyj 1988; Hanson and Bradley 1989, 1993). Conformably overlying these rocks are Devonian deepwater turbidites of the Littleton, Carrabassett, and Seboomook Formations. Paleocurrent directions in the Carrabassett show overall northerly flow (Hanson and Bradley 1993). These formations were deposited in a foreland basin setting associated with either a southeast dipping subduction system that overrode the Silurian northwest dipping subduction system (Bradley et al. 1998) or simply the same northwest-dipping Silurian subduction system that persisted into the Devonian (Eusden et al. 1996*a*).

The Central Maine Terrane has experienced intense ductile deformation, high-grade metamorphism, and a protracted period of pre-, syn-, and postkinematic granitic plutonism. In general, the deformation in the northeast part of the CMT (Maine and New Brunswick) is dominated by upright structures and lower-grade postkinematic contact metamorphisms associated with synkinematic, but largely postkinematic plutons (Moench and Pankiwskyj 1988; Guidotti 1989; Osberg et al. 1989). In the southwest (portions of western Maine, New Hampshire, Massachusetts, and Connecticut) structures are generally recumbent, multiply deformed, and accompanied by synkinematic higher-grade metamorphisms and associated intrusion of predominately synkinematic granitic plutons (Eusden and Lyons 1993; Lyons et al. 1997; Robinson et al. 1998). The transition between these different styles of the Acadian tectonism occurs in a zone only 75–100 km long, as measured along the strike of the CMT, and represents a transition from shallower crustal levels to deeper crustal levels through the now exposed orogen (Carmichael 1978; Osberg et al. 1989). Recent modeling of granite ascent in convergent orogenic belts by Solar et al. (1998) and Solar and Brown (1999) suggests that the synchronous nature of deformation, metamorphism, and plutonism may also be present in the high-*T*, low-*P* metamorphism of western Maine. Bradley et al. (1998) have also shown that many 408–404-Ma plutons in Maine are syntectonic. However, many workers have demonstrated the static nature of the bulk of metamorphism in this same region (e.g., Holdaway et al. 1982; DeYoreo et al. 1989; Guidotti 1989). The fact that there is controversy over the nature of orogenesis in western Maine may reflect the complex effects one would expect to find within and adjacent to the Acadian transition zone and may also be related to the scale of observation and methods of research employed by these researchers.

## Previous Geochronology in the CMT

Using U-Pb ages of plutons, conodont and palynomorph ages of deformed strata, and tectonic analysis of depositional basins, Bradley et al. (1998) have documented a migration of the Acadian deformation front and the adjacent foreland basin to its west that began near the Maine coast during the Late Silurian and swept northwestward across central and western Maine and New Hampshire during the Early Devonian. Bradley et al. (1998) placed the Acadian deformation front in the vicinity of the Presidential Range during the Early Emsian (406–407 Ma). Solar et al. (1998) used U-Pb zircon and monazite ages from plutons and schlieric granite within migmatites in western Maine, approximately 100 km northeast of the Presidential Range, to constrain the granite crystallization to 408–404 Ma. A study of the metamorphism in the Rumford 7.5′ quadrangle of western Maine, about 50 km northeast of the Presidential Range, using U-Pb monazite ages from schists, revealed two groups of ages (Smith and Barreiro 1990), most likely related to the M2 and M3 pulses of high-grade metamorphism recognized by Guidotti (1989). The initial pulse lasted from ~405 to 398 Ma, and the second pulse lasted from ~370 to 365 Ma. A study of peak metamorphism in a broad region of central New Hampshire, about 100 km southwest of the Presidential Range, using monazite ages from schists, migmatites, and two-mica granite sheets, revealed a period of protracted high-grade metamorphism(s) lasting from ~384 to 402 Ma (Eusden and Barreiro 1988).

## Geology of the Presidential Range

Using as a foundation the excellent work of previous geologists who have mapped the rocks in the Presidential Range (Billings 1941; Billings et al. 1946, 1979; Hatch and Moench 1984; Hatch and Wall 1986) and collaborating with those who are actively working there (Allen 1992, 1996; Wing 1996), we are in the midst of a mapping project to redefine the geology in the range. Figure 2 shows the geologic map we have compiled to date, as well as the sample locations for this article.

We have subdivided the Devonian Littleton Formation into 16 members, recognized both the Silurian Madrid and Smalls Falls Formations, subdivided the Rangeley Formation into nine members, and recognized an episode of sedimentary disruption in the Rangeley, which we interpret as a migmatized olistostromal mélange (Eusden et al. 1996*a*). The rationale for treating the Rangeley migmatite as a stratigraphic unit is described in Eusden et al. (1996*a*).

The sequence of deformation is interpreted to comprise five events, D1–D5. The D1 event is characterized by east-verging isoclinal nappes, which are macroscopic at high elevations and mesoscopic at low elevations, while D2 is characterized by the Clay klippe and Greenough Spring thrust (see Eusden et al. 1996*a*). The D3 folds result in anomalous easterly dips of bedding (S0) and D1 foliation (S1), macroscopic refolding of the D2 thrusts, and definition of the great Chandler Ridge Dome. Folds of D4 are the most common structural features in the Presidential Range and vary in scale from mesoscopic at high elevations to mesoscopic and microscopic at lower elevations. Event D5 is principally a crenulation restricted to the Pinkham Notch region. The many vertical and lateral variations in structural style that the phases of deformation exhibit further reflect the complex nature of the deformation that occurred in the Acadian crustal transition (Eusden et al. 1996*a*).

Many pulses of metamorphism have been recognized in the Presidential Range. The following summary is based on Eusden et al. (1996*b*), Wing (1996), Wall (1988), and Allen (1996). The first, M1, is characterized by aligned andalusite, much of which is now preserved as pseudomorphs and occurred during D1 nappe-stage folding, and M2 is characterized by sillimanite zone metamorphism in the Littleton schists and migmatization in the Rangeley gneisses and occurred during the later part of D1 nappe-stage folding. The migmatites, which are all characterized by sillimanite zone metamorphism with potassium feldspar-absent melting, occur in two types of sharp contacts with the surrounding schists. The first type is represented by the Greenough Spring thrust fault, a discontinuity that offsets the stratigraphy and early folds and also coincides exactly with the sharp metamorphic transition between schists outside of the klippe and migmatites within. The second type of migmatite contact, observed sporadically throughout the Presidential Range, is one in which the pelitic schists and quartzites, calc-silicate granofels, and rusty schists of the Littleton, Madrid, and Smalls Falls Formations, respectively, are in conformable stratigraphic contact with the migmatites of the Rangeley Formation. There is no stratigraphic or structural discontinuity recognized at this metamorphic transition, which is again sharp, occurring directly at the contact with the Rangeley.

The later events, M3 and M4, are contact-metamorphic events, both reaching staurolite grade. Event M3 occurred prior to D4 folding, based on the observation that granites related to this phase of metamorphism are folded by F4 folds. The M4 metamorphism occurred after D4 deformation and is related to the latest stage of posttectonic-granite intrusion and D5 crenulation, while M5 is a retrograde metamorphism producing scattered occurrences of chlorite and/or sericite alteration in the schists and gneisses.

The earliest plutonism in the Presidential Range is characterized by one rare dioritic stock, the Wamsutta Diorite (Guzofski 1997). This pluton has a weak S1 foliation but also cuts across the S1 fabric in the metasedimentary rocks. We interpret these observations to mean that the Wamsutta Diorite intruded during the waning stages of F1 nappe-stage folding. The diorite thus represents the earliest synkinematic intrusion in the range and is probably in part synchronous with M1. Two other small diorite intrusions were mapped on the Nineteenmile Brook in the Carter Dome 7.5′ quadrangle, 0.5 km east of Route 16 (fig. 2; Billings and Fowler-Billings 1975). A widespread scattering of sills, veins, and small plutons of two-mica granite intruded sub-

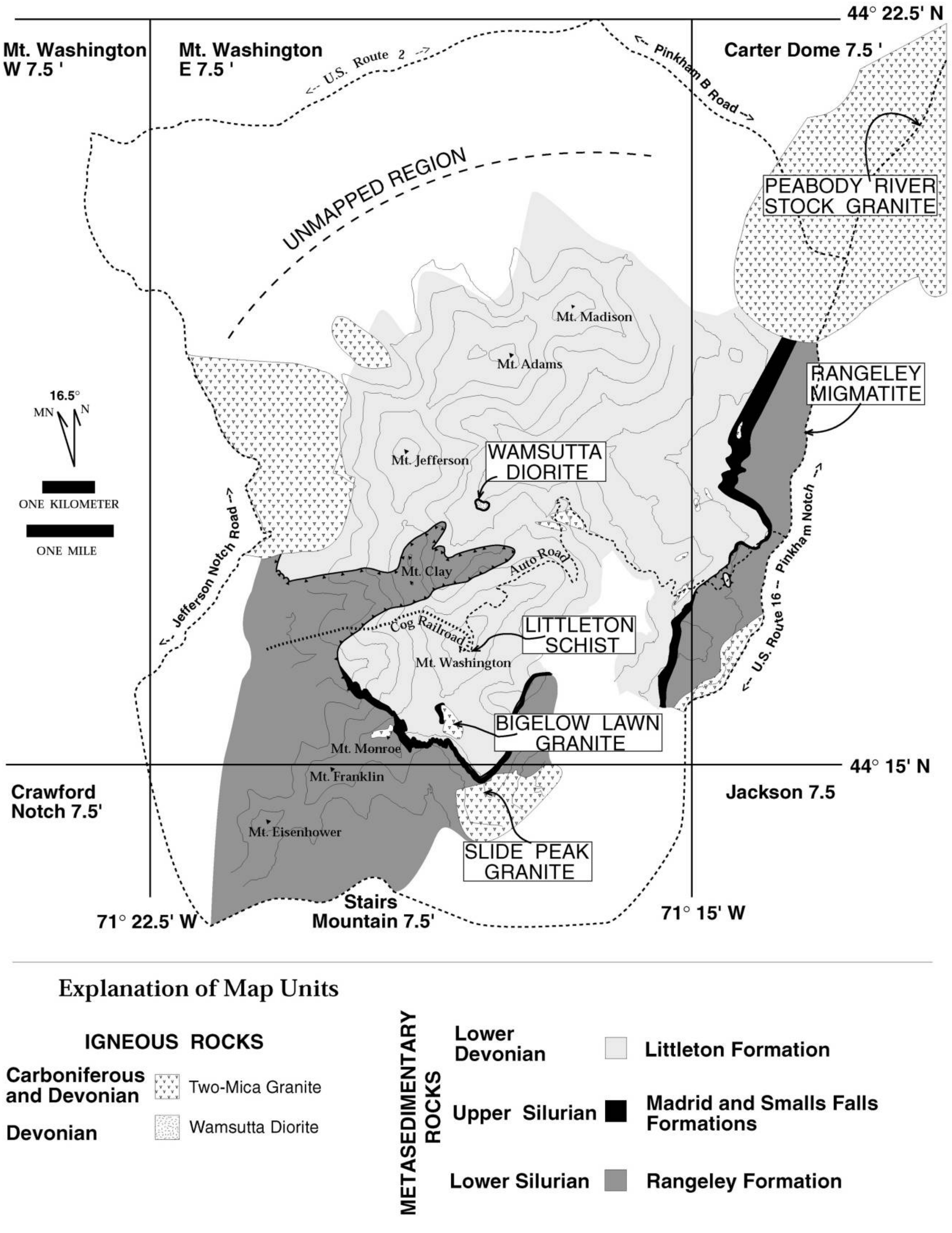


**Figure 2.** Simplified geologic map of the Presidential Range, New Hampshire, showing sample locations (at tip of arrows) and 7.5′ quadrangles.

sequently. The M2 and M3 metamorphisms were associated with the intrusion of these granites. These granites are deformed by F4 folds. These small plutons are normally not foliated but may have a weak foliation near the contact with the metasedimentary rocks and/or migmatites. They cut across F1 folds, thereby postdating D1. Allen (1992, 1996) mapped the largest of these granites near the Wildcat Ski Area immediately adjacent to the Presidential Range. The Wildcat granite occurs in two phases: a medium-grained, two-mica granite and a coarser-grained, biotite-rich granitoid with calc-silicate pods. The second phase of granite is interpreted as having formed from partial melting of the Rangeley Formation (Allen 1992, 1996). Late-stage posttectonic granites are restricted to the Peabody River Stock previously identified by Allen (1992, 1996), Wall (1988), and Billings and Fowler-Billings (1975). This granite probably caused M4 contact metamorphism and possibly D5 crenulations, which are restricted to the areas surrounding its contact. The youngest intrusions are volcanic vent agglomerates and diabase dikes, part of the Jurassic-Cretaceous White Mountain Magma Series, found in small numbers throughout the study area (Allen 1992, 1996; Billings and Fowler-Billings 1975).

## U-Pb Analytical Methods

U-Pb analyses were performed in the geochronology lab at Washington University, St. Louis. Six samples of 25–50 kg were taken from outcrops of two different metamorphic and four igneous rocks in the Presidential Range. Monazite and zircon were extracted using standard crushing, Rogers table, sieving, heavy liquids, and magnetic separation techniques. Monazites and zircons were then handpicked under a binocular microscope based on grain size, shape, clarity, and color. Analyses were performed on two to six fractions of one to six crystals each. All analyses were air abraded (Krogh 1982) and then cleaned in warm 4N $HNO_3$, water, and distilled acetone to remove common Pb components. Samples were then dissolved in TFE Teflon bombs, spiked with $^{205}Pb/^{235}U$ tracer, and digested in 48% HF and 7N $HNO_3$. Following conversion to chloride form, U and Pb were extracted using ion exchange techniques (Krogh 1973). Isotope ratio measurements of Pb and U were made using a VG Sector-54 automated thermal ionization mass spectrometer with seven collectors and a Daly-type detector.

Errors for the $^{238}U/^{206}Pb$, $^{235}U/^{207}Pb$, and $^{207}Pb/^{206}Pb$ ages were estimated using the method of Ludwig (1980) and all age uncertainties are quoted at the 95% confidence level. Cited ages for zircon and monazite are the mean $^{207}Pb/^{206}Pb$ and $^{207}Pb/^{235}U$ ages, respectively, of concordant or slightly discordant analyses weighted according to the inverse variance of each analysis (Ludwig 1992). The quoted age error is the standard error of the average value calculated using the assigned error for each analysis. The reliability of the cited ages may be evaluated by the mean square of the weighted deviates (MSWD), which in all cases is <1.5, indicating that the assigned errors may be somewhat overestimated. Concordia diagrams are shown in figure 3 and U-Pb isotope-dilution analyses are reported in table 1. The time scale of Tucker et al. (1998) was used in the discussion.

## Results

U-Pb ages for each sample are given below. For detailed sample descriptions and locations, the reader is referred to Robinson (1997).

***Wamsutta Diorite.*** The Wamsutta Diorite is a weakly foliated, medium-grained, quartz diorite sampled from an outcrop along the Peabody River at Wamsutta Falls, 2.8 km N 08° E of Mount Washington (fig. 2). Six analyses of different zircon fractions were performed (table 1). Analyses 1, 3, and 4 are clear, colorless zircon needles; analysis 2 is a clear, colorless zircon tip; and analyses 5 and 6 are clear, colorless, rounded zircon prisms. Analyses 1–3 plot on concordia, and analysis 4 plots slightly below concordia (fig. 3*A*). These four analyses yield an upper intercept age (using the $^{206}Pb/^{238}U : ^{207}Pb/^{235}U$ ratios and the concordia plot) of 408.4 ± 1.9 Ma, with the lower intercept forced through the present. The weighted mean of the $^{207}Pb/^{206}Pb$ ages of these four analyses gives an age of 408.2 ± 2.0 Ma. Both of these ages agree within error. Analyses 5 and 6 are normally discordant, plotting well below concordia, with upper discordia intercepts of 1207 ± 6 Ma and 1487 ± 6 Ma (fig. 3*A*).

***Bigelow Lawn Granite.*** The Bigelow Lawn Granite (Peters 1992) is a well-foliated, fine-grained, light gray to white, two-mica granite and was sampled from Bigelow Lawn 1.2 km S 80° W of the summit of Mount Washington (fig. 2). Two monazite fractions of two grains each were analyzed (table 1). Both monazite fractions consisted of clear, yellow, rounded grains. Analysis 1 yielded a $^{207}Pb/^{235}U$ age of 402.5 ± 0.5 Ma and plots on concordia (fig. 3*B*). Analysis 2 yielded a $^{207}Pb/^{235}U$ age of 424.9 ± 0.8 Ma and also plots on concordia.

***Slide Peak Granite.*** The Slide Peak Granite (Johnson 1993) is a well-foliated to nonfoliated, medium-

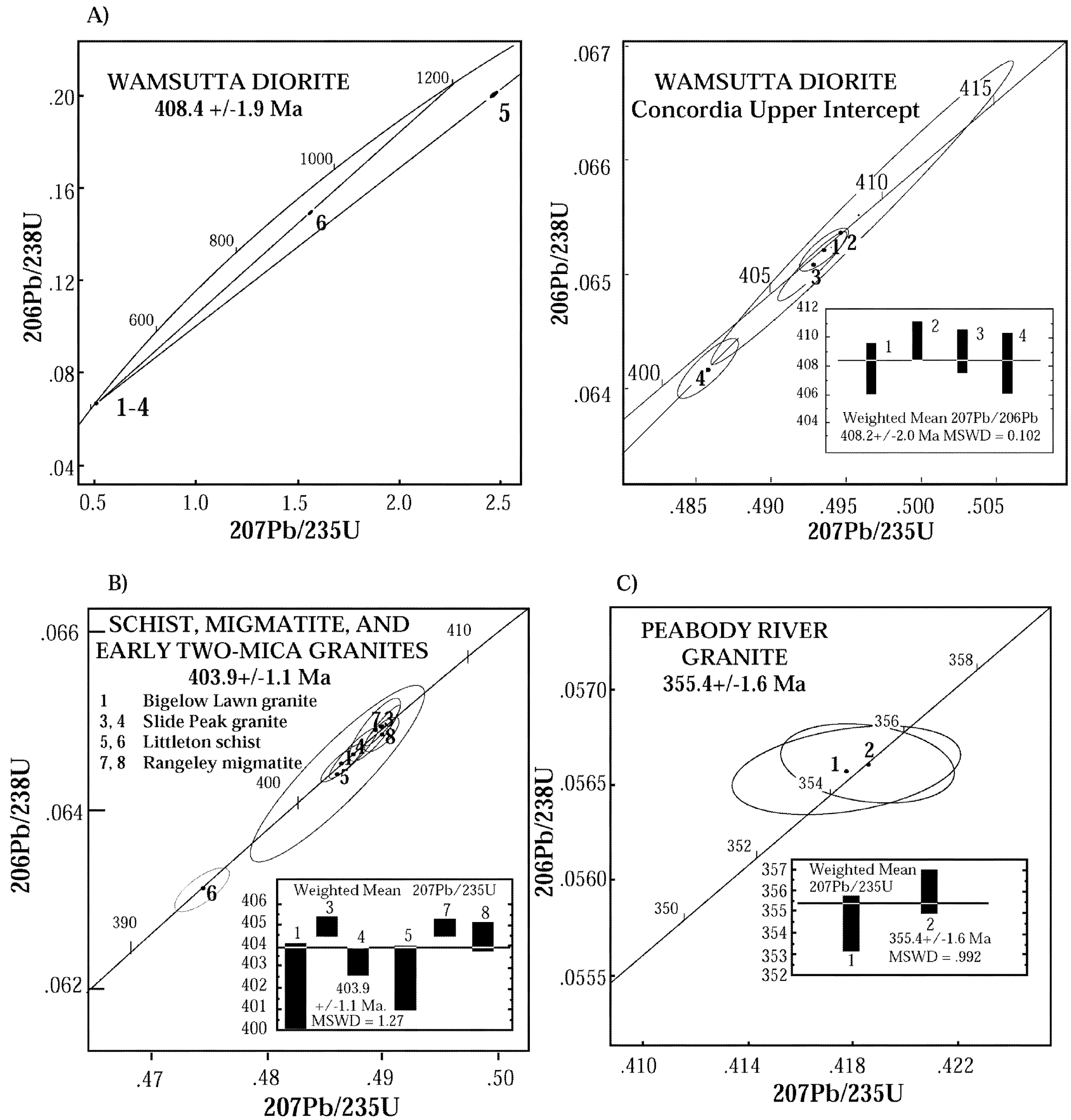


**Figure 3.** U-Pb concordia diagrams. See table 1 for data. *A*, Zircon analyses from the Wamsutta Diorite. *B*, Monazite analyses from metamorphic rocks and early two-mica granite plutons. *C*, Monazite analyses from the Peabody River Granite.

to fine-grained, light gray to white, two-mica granite with lenses of well-foliated schist or gneiss (Johnson 1993) and was sampled from outcrop along the Glen Boulder Trail above the Gulf of Slides, 0.3 km N 60° W of Slide Peak (fig. 2). Two monazite fractions of one grain each were analyzed. Both monazite fractions consisted of a single clear, yellow, rounded grain. Analysis 1 yielded a $^{207}Pb/^{235}U$ age of 404.9 ± 0.5 Ma, and analysis 2 yielded a $^{207}Pb/^{235}U$ age of 403.3 ± 0.6 Ma (table 1). Analysis 1 is reversely discordant, and analysis 2 plots on concordia. These two ages do not agree within error from the $^{207}Pb/^{235}U$ ratios. However, the two analyses do overlap on the concordia plot, which is derived from both the $^{207}Pb/^{235}U$ ratio and the $^{206}Pb/^{238}U$ ratio (fig. 3*B*).

***Littleton Schist.*** The sample from the Littleton Formation is a unit of well-bedded schist and

**Table 1.** U-Pb Isotope Dilution Analyses

| Sample analyses | Concentrations | | | | Atomic ratios | | | | | Ages (Ma) | | | Discordance (%) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | # min[a] | U, ppm | Pb (rad), ppm | Th, ppm | $^{206}Pb^{b}/^{238}U$ | $^{207}Pb^{b}/^{235}U$ | $^{207}Pb^{b}/^{206}Pb^{b}$ | $^{208}Pb^{b}/^{206}Pb^{b}$ | Th/U | $206^{b}/238$ | $207^{b}/235^{b}$ | $207^{b}/206^{b}$ | |
| Wamsutta Diorite: | | | | | | | | | | | | | |
| 1 | 1 z | 593.07 | 37.57 | 137.16 | .06522 | .49361 | .05489 | .0747 | .231 | 407.28 | 407.36 | 407.82 | .14 |
| 2 | 1 z | 355.21 | 22.67 | 83.94 | .06554 | .49607 | .05490 | .0760 | .236 | 409.20 | 409.04 | 408.09 | −.28 |
| 3 | 1 z | 495.98 | 30.82 | 85.87 | .06508 | .49280 | .05492 | .0558 | .173 | 406.42 | 406.81 | 409.05 | .66 |
| 4 | 1 z | 764.89 | 46.17 | 92.76 | .06418 | .48588 | .05490 | .0391 | .121 | 401.03 | 402.10 | 408.25 | 1.82 |
| 5 | 1 z | 285.93 | 59.81 | 113.94 | .20013 | 2.46064 | .08918 | .1258 | .398 | 1176.00 | 1260.60 | 1407.90 | 18.01 |
| 6 | 1 z | 225.56 | 34.67 | 83.31 | .14954 | 1.56580 | .07594 | .1191 | .369 | 898.36 | 956.77 | 1093.58 | 19.12 |
| Bigelow Lawn Granite: | | | | | | | | | | | | | |
| 1 | 2 m | 4119.54 | 849.60 | 34,055.64 | .06453 | .48652 | .05468 | 2.6393 | 8.267 | 403.14 | 402.54 | 399.07 | −1.05 |
| 2 | 2 m | 1101.98 | 237.12 | 8861.85 | .0681 | .51957 | .05533 | 2.597 | 8.042 | 424.71 | 424.86 | 425.70 | .24 |
| Slide Peak Granite: | | | | | | | | | | | | | |
| 3 | 1 m | 5524.34 | 878.92 | 30,854.19 | .06494 | .48995 | .05472 | 1.7795 | 5.585 | 405.60 | 404.87 | 400.74 | −1.25 |
| 4 | 1 m | 5022.66 | 888.51 | 33,125.87 | .06463 | .48757 | .05471 | 2.1106 | 6.595 | 403.72 | 403.25 | 400.58 | −.81 |
| Littleton Schist: | | | | | | | | | | | | | |
| 5 | 1 m | 5096.43 | 788.94 | 27,321.06 | .06440 | .48603 | .05474 | 1.7264 | 5.361 | 402.30 | 402.2 | 401.65 | −.17 |
| 6 | 1 m | 4343.93 | 897.07 | 36,895.70 | .06311 | .47442 | .05452 | 2.7276 | 8.494 | 394.51 | 394.24 | 392.61 | −.50 |
| Rangeley Migmatite: | | | | | | | | | | | | | |
| 7 | 1 m | 5527.05 | 753.91 | 23,745.70 | .06486 | .48994 | .05478 | 1.6793 | 4.296 | 405.31 | 404.87 | 403.32 | −.47 |
| 8 | 2 m | 7533.06 | 1023.05 | 32,398.85 | .06489 | .48933 | .05469 | 1.3680 | 4.301 | 405.28 | 404.45 | 399.75 | −1.43 |
| Peabody River Stock Granite: | | | | | | | | | | | | | |
| 1 | 1 m | 751.98 | 335.78 | 18,859.87 | .05658 | .41778 | .05355 | 8.0542 | 25.080 | 354.79 | 354.46 | 352.34 | −.71 |
| 2 | 1 m | 804.00 | 393.97 | 22,155.86 | .05674 | .41960 | .05364 | 8.9121 | 27.557 | 355.76 | 355.77 | 355.80 | −.01 |

[a] # = number of grains in analyzed fraction; min = mineral analyzed; z = zircon; m = monazite.
[b] Radiogenic lead corrected from common lead using the isotopic ratios 204 : 206 : 207 : 208 = 1 : 17.753 : 15.572 : 37.580 (Stacey and Kramers 1975).

quartzite with 1–1.5 m couplets containing about 30% quartzite. The schist sampled comes from a roadcut on the Mount Washington Auto Road near the summit of Mount Washington (fig. 2; same locality as Stop 1 in Eusden et al. 1996*b*). Two monazite fractions of one grain each were analyzed. Each fraction consisted of one yellow, rounded monazite. Analysis 1 yielded a $^{207}Pb/^{235}U$ age of 402.2 ± 2.1 Ma, and analysis 2 yielded a $^{207}Pb/^{235}U$ age of 394.2 ± 0.6 (table 1). Both analyses plot on concordia (fig. 3*B*).

***Rangeley Migmatite.*** The Rangeley migmatite was sampled from a roadcut along the east side of U.S. Route 16, 2.2 km north of the base of the Mount Washington Auto Road (fig. 2). The rock consists of well-foliated quartz-muscovite-biotite-sillimanite-garnet migmatitic gneiss (Wall 1988). Two monazite fractions were analyzed. Fraction 1 consisted of one yellow, rounded grain. Fraction 2 consisted of two yellow, rounded grains. Analysis 1 yielded a $^{207}Pb/^{235}U$ age of 404.9 ± 0.4 Ma, and analysis 2 yielded a $^{207}Pb/^{235}U$ age of 404.5 ± 0.7 Ma (table 1). Both analyses plot on concordia with overlap (fig. 3*B*).

***Peabody River Stock.*** The Peabody River Stock (Billings and Fowler-Billings 1975) is a nonfoliated, light to medium gray, fine-grained, two-mica granite and was sampled from a roadcut on the east side of U.S. Route 16, 7.4 km north of the base of the Mount Washington Auto Road (fig. 2). Two monazite fractions consisting of one grain each were analyzed (table 1). Both fractions were yellow, rounded grains. Analysis 1 yielded a $^{207}Pb/^{235}U$ age of 354.5 ± 1.3 Ma. Analysis 2 yielded a $^{207}Pb/^{235}U$ age of 355.0 ± 1.0 Ma. Both analyses plot on concordia and overlap each other (fig. 3*C*).

## Discussion

***Early Emsian Plutonism: Implications for Sedimentation, Deformation, and Tectonism.*** The Wamsutta Diorite is likely a Spaulding-type diorite for which Lyons and Livingston (1977) have reported Rb-Sr ages of 402 ± 5 Ma. This agrees within error with the weighted mean $^{207}Pb/^{206}Pb$ age of 408.2 ± 2.0 Ma (fig. 3*A*). The Wamsutta is also correlative to and lies within a belt of Emsian plutons that are syntectonic with respect to the first phase of local Acadian deformation (Bradley et al. 1998). This belt extends northeast-southwest through Maine and into New Hampshire (see fig.1). Amazingly, the Littleton Formation, into which the Wamsutta intrudes, has essentially the same age of deposition based on an Emsian U-Pb zircon age of 407 ± 2 Ma (R. Tucker and D. Rankin, unpub. data cited in Bradley et al. 1998), from an interstratified tuff collected only 40 km west of the diorite! Therefore, the boundary between the deforming orogenic wedge and the depositional site in the foreland basin must have been between these two sites at 407–408 Ma. These two sites are, of course, much closer now than then, due to Acadian shortening. This supports the model of Bradley et al. (1998), who stated that Littleton deposition is diachronous across strike in this area, similar to their findings in Maine. Littleton deposition occurred at approximately 410–409 Ma in the Presidential Range, 408–407 Ma in the Beaver Brook–Gale River area, and slightly younger still in the type section of Littleton, New Hampshire. The first wave of Acadian deformation chased this deposition, also in a westerly direction, lagging behind by 1–2 m.yr. or so.

Bradley et al. (1998) suggested that the migrating deformation front is linked with an A-type subduction zone (one in which continental crust is being subducted). This evolved from a B-type system (one in which oceanic crust is being subducted) with a southeast-dipping subduction zone. The earliest structures are shown as northwest vergent in their model. However, in the Presidential Range the earliest structures, F1 nappes, were clearly easterly verging. The F4 folds were also east-vergent, and the regional dip is westerly, implying an overall top-to-the-east shear for this part of the orogen (Eusden et al. 1996*a*). Furthermore, and more fundamentally for tectonic reconstructions of the Acadian, this same structural geometry is also found throughout the CMT in central and eastern New Hampshire (Eusden and Lyons 1993; Lyons et al. 1997) and southwestern Maine (Eusden et al. 1987). The distribution of east- and west-verging nappes in the CMT has been summarized by Eusden et al. (1996*a*). Any Acadian tectonic model that only accounts for the west-verging nappes (Bradley et al. 1998) observed in the Connecticut Valley region (Robinson et al. 1991) and portions of Maine (Osberg et al. 1989) is an incomplete model for the CMT. Thus, the details of the structural geology in the Presidential Range and throughout much of the CMT of New Hampshire and Maine do not fit the Bradley et al. (1998) tectonic model. Yet, the new geochronology presented here fits beautifully with the timing of Acadian migration proposed by Bradley et al. (1998).

In keeping with the requirement that an area can only have one geological history where all evidence must fit together into a credible model, we suggest that it has to be possible to have both a northwest-migrating deformation front and a thrust-nappe belt that is southeasterly verging at the same time during the Acadian. Figure 4 is an attempt to schematically show such a tectonic model along a cross section through the Presidential Range. No attempt has been made to model another line of section, approximately 100 km along strike, that would address the west-vergent nappes recognized in southwest New Hampshire.

In the model, the precollisional geometry involved two subduction zones as originally suggested by Bradley (1983) and also discussed, but not endorsed, by Robinson et al. (1998). This represents a compromise of sorts between the model of Eusden et al. (1996*a*) and the favored model of Robinson et al. (1998) and Bradley et al. (1998). The Pre-Ludlovian geometry is characterized by B-type subduction along the Laurentian margin (Bradley et al. 1998) and Ordovician-to-Silurian transpressional orogenesis within Composite Avalon now exposed along the Maine Coast (Rankin 1994; Stewart et al. 1995). To account for the lack of Acadian-aged arc volcanics in the Bronson Hill Anticlinorum—or at least the lack of agreement over whether the scattered volcanics are arc (Bradley 1983) or rift related (Hon et al. 1992)—the slab dip angle was possibly too shallow to trigger partial melting and arc volcanism (so-called buoyant subduction of Cross and Pilger 1982). Sometime in the Ludlovian (423–419 Ma), southeast-dipping subduction developed along the leading edge of Composite Avalonian; shortly after, this continental crust made contact with the trench and A-type subduction commenced (Bradley et al. 1998). This subduction system would account for the Coastal Volcanics that erupted on Composite Avalon from the late Llandovery to early Lochkovian. The Acadian deformation front advances in the direction of the dip (northwesterly), chasing the Devonian sedimentation of the Littleton Formation. The structures in the orogenic wedge would verge southeasterly with younger thrust nappes overlying older ones propagating toward Bronson Hill.

A present-day tectonic analog to this two-subduction-zone model is the Molucca Sea in the Indonesian region, as Bradley (1983) has previously discussed. A modern analog for just the southeast subduction along Composite Avalon and the out-of-sequence thrust-nappe structures is the Australian-Pacific plate boundary in the South Island of New Zealand. In North Canterbury, New Zealand, Barnes (1996) reports a west-verging, but eastward-younging (i.e., out of sequence), fold-thrust belt emerging from the sea. The subduction polarity is known and dips west as the Pacific plate subducts beneath the Australian plate along the Hikurangi trench. This geometry is essentially a mirror image

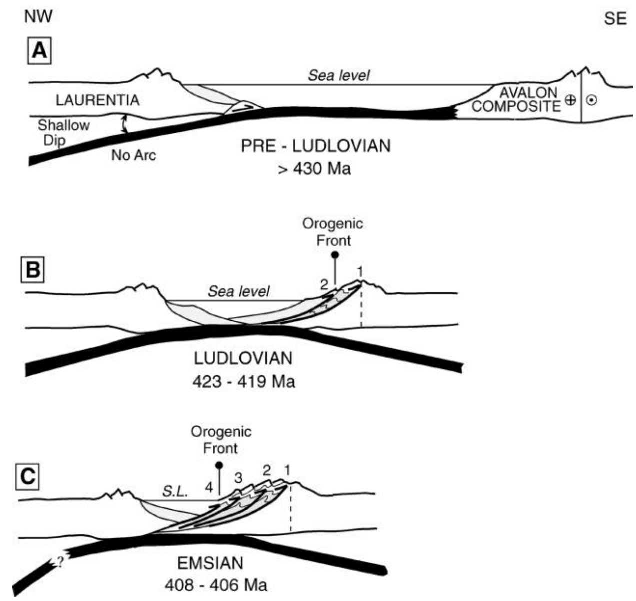


**Figure 4.** Schematic plate tectonic model oriented approximately northwest-southeast through the Presidential Range. *A*, Pre-Ludlovian B-type subduction with an older buttress developed in Composite Avalon. *B*, Ludlovian initiation of A-type subduction and Acadian deformation, earliest thrust nappes (1) developed along the present Maine Coast, stacked out of sequence (2) to the northwest. *C*, Emsian position of the deformation front as it advances to the northwest while the thrust nappes continue to verge to the southeast.

of that proposed in figure 4*B* for Avalon Composite. It is also interesting to note that in the New Zealand example, while tectonic uplift accompanies the active fold-thrust belt on land and immediately offshore, subsidence and deposition occurs just a bit farther offshore; thus, this analog could also provide a mechanism for basin subsidence and Littleton-Seboomook deposition in the Devonian.

The out-of-sequence thrust-nappe propagation proposed in figure 4 could have been created by a tectonic buttress (dashed vertical line within Avalon composite in fig. 4*B*, 4*C*) that formed in Pre-Ludlovian times along the Maine coast as the St. Croix and Ellsworth terranes accreted (Rankin 1994; Stewart et al. 1995). This orogenic belt, formed just prior to and flanking the southeast margin of the Acadian Orogeny, may have served as a barrier precluding southeast migration of the Acadian orogenic front.

This northwest subduction along the Laurentian margin in figure 4 is consistent with the recent tectonic interpretations of Karabinos et al. (1998) for the Bronson Hill Arc in Vermont and New

Hampshire and van Staal (1994) for the Brunswick subduction complex in New Brunswick. Developed principally to illustrate the complexities of Taconian orogenic effects in the Ordovician, both models show a west-dipping slab geometry and B-type subduction during the latest Ordovician and earliest Silurian along the Bronson Hill. If this slab geometry persisted throughout at least some of the Acadian, one would expect east-vergent, syncollisional, thrust-nappe structures and olistostromal facies along the Bronson Hill, as Eusden et al. (1996*a*) have previously reported for the Presidential Range. Though it is not within the scope of this article to explain the west-vergent structures seen 100 km along strike in the Connecticut Valley region (Robinson et al. 1991), we suggest that the New Zealand plate boundary can again be used as an analog. Pettinga and Wise (1994) suggest a flower structure exists due to transfer along the Alpine Fault from the Puysegur trench with east-dipping subduction to the Hikurangi trench with west-dipping subduction. In this model, both east- and west-verging structures exist, separated by approximately only 100 km along strike, close to the geometry of thrust nappes observed in the CMT of New Hampshire.

***Middle Emsian Peak of Metamorphism and Migmatization.*** Figure 3*B* shows a single concordia plot with the Rangeley migmatite, Littleton schist, Slide Peak Granite, and Bigelow Lawn Granite samples plotted. All the analyses (with the exception of analyses 2 and 6) fall within a very short period of time (397–407 Ma, with most of the ages between 402 and 406 Ma) with substantial amounts of overlap. This figure graphically depicts the synchronous nature of the metamorphism and migmatization recorded in the Rangeley and Littleton Formations and the igneous events represented by the Slide Peak and Bigelow Lawn granites. The weighted average $^{207}Pb/^{235}U$ age for the samples with ages between 402 and 405 Ma is 403.9 ± 1.1 Ma. This we take to be the best estimate of the timing of peak metamorphism, which was synchronous with the earliest two-mica granite plutonism.

One of the Bigelow Lawn Granite fractions gave an age of 424.9 ± 0.8 Ma, which cannot date the crystallization of granite, as this age is older than the metasedimentary rock it intrudes. We offer two possible explanations for this age: (1) it could be contamination or (2) a detrital monazite (from the Bronson Hill Anticlinorium?) that somehow survived the peak of metamorphism without being entirely reset (unlikely, given the arguments of Smith and Barreiro [1990] on monazite systematics). One single-grain monazite fraction from the Littleton schist gave a concordant age of 394.2 ± 0.6 Ma (fig. 2), a full 10 m.yr. after the peak of metamorphism. This could be related to one of the other subsequent metamorphisms (e.g., M3) but that is speculative based on a single analysis. It does at least suggest that some high-grade metamorphic monazites were either being formed or reset 10 m.yr. after the presumed peak of metamorphism at circa 404 Ma.

The synchroneity at 404 Ma of peak metamorphism in schists and migmatization linked with crystallization of granites in the Presidential Range strongly supports the migmatite evolution model of Allen (1992, 1996) developed for exposures at the base of the range in Pinkham Notch. Allen's model calls for localized migmatization to be produced by an anatectic partial melting process driven by infiltration of magmatic fluids. Large localized migmatite zones and metamorphic "hot spots" are interpreted as the conduits through which magma passed as it migrated from sources deeper in the crust to be emplaced as plutons at higher crustal levels (Allen 1992, 1996). Allen suggests that these conduits are also structurally controlled; thus, the metamorphism, migmatization, and associated plutonism were synkinematic. The model of granite ascent and synkinematic regional metamorphism in west-central Maine by Solar et al. (1998) and Solar and Brown (1999) is also likely applicable to the Presidential Range.

A compilation of the oldest monazite ages from throughout Maine and New Hampshire is shown in figure 1. Most of these ages fall between circa 384–408 Ma (Eusden and Barreiro 1988; Smith and Barreiro 1990; Solar et al. 1998) with a later pulse at circa 370–363 Ma in the Rumford, Maine, area (Smith and Barreiro 1990). The data set, though meager and representing a mixture of different techniques (e.g., single-grain analysis, bulk analysis), does show a general pattern of younger ages (<385 Ma) in southeastern New Hampshire that get progressively older (400–405 Ma) to the west, north, and northeast (fig. 1). The pattern is likely the result of the complex interactions of a polyphase Acadian orogeny and the post-Acadian history of cooling, tilting, and exhumation.

***Early Carboniferous Postkinematic Plutonism and Contact Metamorphism.*** The Peabody River Stock granite yields a weighted mean $^{207}Pb/^{235}U$ age of 355.4 ± 1.6 Ma (fig. 3*C*). This age is interpreted to be not only the crystallization age of the pluton but also the age of M4 contact metamorphism (Wall 1988; Eusden et al. 1996*a*) and D5 crenulation (Guzofski 1997) around the pluton. Pending future work, this age may also effectively be the younger limit of Paleozoic orogenesis in the Presidential

Range. This is the only age that might be related to the Neo-Acadian (circa 360 Ma) metamorphic overprint (Robinson et al. 1998) seen in southern New Hampshire and central Massachusetts. This zone is sandwiched between the Massabesic Gneiss Complex and the Pelham Dome in Massachusetts and dies out to the north somewhere in southern New Hampshire.

Middle Devonian to Early Carboniferous granites of this age are common throughout the CMT of New Hampshire and Maine. Many two-mica and biotite granitoids are concentrated in the migmatite belt of central New Hampshire (Lyons et al. 1997), although others are widely scattered throughout New Hampshire and Maine (Osberg et al. 1985). They are posttectonic plutons that we consider post-Acadian. Thermal modeling of Chamberlain and England (1985) suggests that magmatism of this age could be the result of thermal relaxation following the earlier peak metamorphic event.

## Summary

The new U-Pb ages presented here constrain the time of several key events in the history of the Acadian orogeny of northern New Hampshire: (1) the boundary between the deforming orogenic wedge and the depositional site in the foreland basin was in the vicinity of the Presidential Range at 408 Ma; (2) southeast-verging thrust nappes formed at circa 408 Ma. Plate tectonic models for the Acadian must be modified to account for both a northwest-migrating deformation front and a southeasterly verging thrust-nappe belt; (3) peak metamorphism in schists, migmatization, and intrusion of early two-mica granites occurred at circa 404 Ma; (4) metamorphic monazite ages from northern New Hampshire are similar in age to those of southwestern New Hampshire and western Maine (circa 400–405 Ma) and are distinctly older than metamorphic monazites from southeastern New Hampshire (circa 385–390 Ma); and (5) the younger limit of Paleozoic orogenesis in the Presidential Range is at 355 Ma, as recorded by the postkinematic Peabody River Granite. From circa 408 to 404 Ma, and perhaps as late as circa 394 Ma, the Acadian orogeny in the Presidential Range was typical of the synkinematic style seen in central and southern New Hampshire, Massachusetts, and Connecticut. From no earlier than circa 394 Ma to as late as circa 355 Ma, the tectonism was typical of the postkinematic style in parts of Maine. All of these constraints confirm the presence of a complex system of temporal and spatial variations in the styles of tectonometamorphism that occurred within the Acadian transition zone of the Northern Appalachians. Future work integrating detailed structural mapping at large scales, high-precision geochronology to constrain the timing of the orogenesis, and studies of the complex metamorphisms and intrusions will greatly help in developing better tectonic models for the Acadian transition zone exposed in New Hampshire and Maine.

## ACKNOWLEDGMENTS

We thank the Howard Hughes Medical Institute and Bates College for providing financial support for the research. We also thank the Appalachian Mountain Club and Mount Washington Auto Road for logistical support during fieldwork. D. Bradley and T. Allen are thanked for helpful conversations about this research. Thanks to three anonymous reviewers for providing very helpful comments.

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