GEOLOGICAL NOTES

Emsian Synorogenic Paleogeography of the Maine Appalachians

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ABSTRACT

The Acadian deformation front in the northern Appalachians of Maine and New Hampshire can now be closely located during the early Emsian (Early Devonian; 408–406 Ma). Tight correlations between paleontologically and isotopically dated rocks are possible only because of a new 408-Ma time scale tie point for the early Emsian. The deformation front lay between a belt of Lower Devonian flysch and molasse that were deposited in an Acadian foreland basin and had not yet been folded and a belt of early Emsian plutons that intruded folded Lower Devonian rocks. This plutonic belt includes the newly dated Ore Mountain gabbro (U/Pb; 406 Ma), which hosts magmatic-sulfide mineralization. Along the deformation front, a 407-Ma pluton that locally truncates Acadian folds (Katahdin) was the feeder to volcanic rocks (Traveler Rhyolite; 406–407 Ma) that are part of the foreland-basin succession involved in these same folds. The Emsian igneous rocks thus define a syncollisional magmatic province that straddled the deformation front. These findings bear on three alternative subduction geometries for the Acadian collision.

Introduction

The Acadian orogeny in Maine was diachronous, younger toward the craton (Donohoe and Pajari 1973). Recent work on the Devonian time scale, and more precise geochronology and paleontology in northern New England and adjacent Canada, have greatly clarified the chain of Acadian events—"Acadian" being used here in a broad sense (e.g., Bradley et al. 2000, p. 42). Near the Maine coast (figs. 1, 2), deformation was already underway by Ludlow time (Late Silurian) (West et al. 1992; Tucker et al. 2001). In the Eastern Townships of Quebec, 240 km to the northwest, Acadian deformation did not occur until the Givetian or Frasnian (Late Devonian) (Bradley et al. 2000).

This article focuses on a single paleogeographic snapshot of the Acadian orogeny—the early Emsian (fig. 3)—when the migrating orogenic front was midway across Maine, about halfway between its initial and final positions. The early Emsian is worthy of special attention because the paleogeography of northern New England is now particularly clear (Bradley et al. 1998, 2000). The early Emsian picture turns out to have been complicated by a magmatic province that straddled the proximal part of the foreland basin, the deformation front, and the actively deforming orogenic wedge. This is rare but not unprecedented in the modern world, and it reveals a previously unappreciated facet of Acadian orogenesis that bears on alternative plate models.

The Devonian time scale of Tucker et al. (1998) includes a new Emsian tie point that allows use of both paleontological and geochronological data to reconstruct the paleogeography. The Sprout Brook ash, which was discovered by Chuck ver Straeten near the base of the Esopus Shale in upstate New York, yielded a concordant U/Pb zircon age of 408.3 ± 1.9 Ma (five concordant fractions; thermal ionization mass spectrometry age). The Esopus Shale is placed in the early Emsian based on its brachiopod fauna and its stratigraphic position between fossiliferous Pragian and Eifelian rocks. Using this and other tie points, Tucker et al. (1998) interpolated the boundaries of the Emsian at around 409 and 394 Ma.

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Emsian Foreland Basin

Thick accumulations of Lower Devonian flysch cover large parts of northwestern New Brunswick, northern Maine, southeastern Quebec, western New Hampshire, and eastern Vermont (Boucot 1970) (fig. 3). These strata range from Lochkovian to early Eifelian age and include the widespread Seboomook Group and Littleton and Temiscouata Formations (Bradley et al. 2000). Rocks yielding Emsian fossils or Emsian isotopic ages are known from six general areas. In northwestern New Brunswick, the Temiscouata Formation has yielded Emsian brachiopods (St. Peter and Boucot 1981). In northern Maine, two samples from the Fish River Lake Formation have yielded spores of late Pragian–early Emsian age (see Bradley et al. 2000). In the Eastern Townships of Quebec, the Compton Formation has yielded Emsian plant fossils (Hueber et al. 1990). Near its type area, the Littleton Formation has yielded brachiopods of Emsian age from four collections, and interstratified rhyolite at Gale River has a U/Pb zircon age of 407 Ma (Rankin and Tucker 2000). All of the Emsian fossils just listed are from turbiditic sandstones and siltstones. By interpolation, some of the unfossiliferous flysch in this belt is undoubtedly also Emsian.

Lower Devonian molasse—some of it Emsian in age—crops out in a relatively narrow belt southeast of the Emsian flysch. Near Moosehead Lake (figs. 3, 4), the Tomhegan Formation consists of deltaic sandstones containing brachiopods of late Emsian age in its upper part (Boucot and Heath 1969). The Tomhegan both overlies and grades laterally into the Kineo Rhyolite, which is equivalent to the Traveler Rhyolite (407 Ma; as in "Geologic Relations Near the Deformation Front"); the lower part of the Tomhegan Formation is therefore early Emsian. Near Mount Katahdin (figs. 3, 4), the Matagamon Sandstone, a deltaic sequence (Pollock et al. 1988), is mostly Pragian based on its brachiopod fauna but is probably earliest Emsian in its uppermost part because of its relations with the Traveler Rhyolite. The Matagamon conformably underlies the Traveler; its uppermost few meters contain rhyolite pebbles identified as Traveler equivalent, and it was injected upward as clastic dikes into the Traveler (Rankin 1968, 1994).

Sedimentary facies, thickness, subsidence rates, and paleocurrent patterns all suggest that these Lower Devonian strata were deposited in a foreland basin (Bradley 1983; Bradley and Hanson 2002). This depositional setting is also supported by evidence, discussed next, for a belt of Emsian deformation just southeast of Emsian molasse facies.

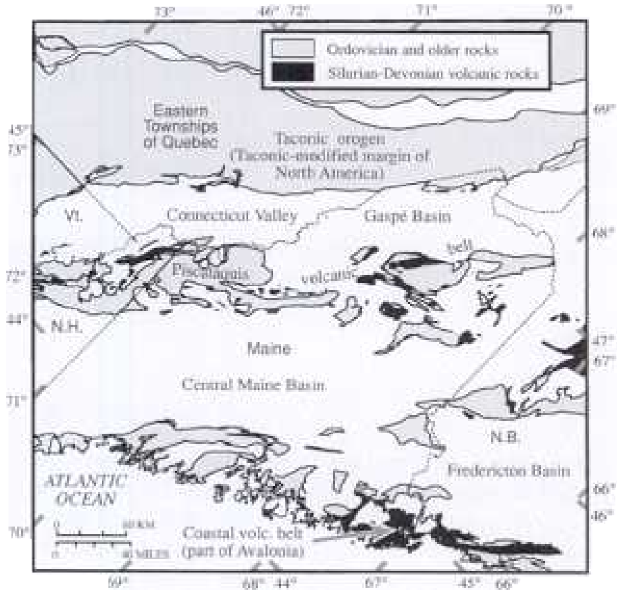

Figure 1. Map of Maine, adjacent parts of northeastern United States, and eastern Canada, identifying the principal tectonic features.

Acadian Orogenic Wedge Defined by Syntectonic and Posttectonic Plutons

Nearly two dozen syntectonic and posttectonic Emsian plutons define a northeast-southwest belt that can be traced across the area of figure 1 and beyond. Until a few years ago, there had been some indication of plutonic activity around 400–410 Ma along this trend but little to suggest that almost all of the plutons along the belt would fall in an extremely narrow age range (404–408 Ma). In central Maine, three of the Emsian plutons quite clearly truncate Acadian folds and associated cleavage, as shown, for example, by the map pattern of the 406-Ma Russell Mountain pluton (Ludman 1978) (fig. 5). Contact-metamorphic rocks on the pluton's northwestern margin have a schistose foliation and aligned chiastolite with biotite strain shadows (fig. 9*a* in Bradley et al. 2000). Because metamorphic biotite is incipient or absent outside the aureole (Ludman 1978), some deformation must have taken place while the pluton was being emplaced, around 406 Ma. There can be little doubt that Acadian deformation was underway in this part of the orogen by the time plutons such as the Russell Mountain were emplaced (Ludman 1978); the new U/Pb geo-

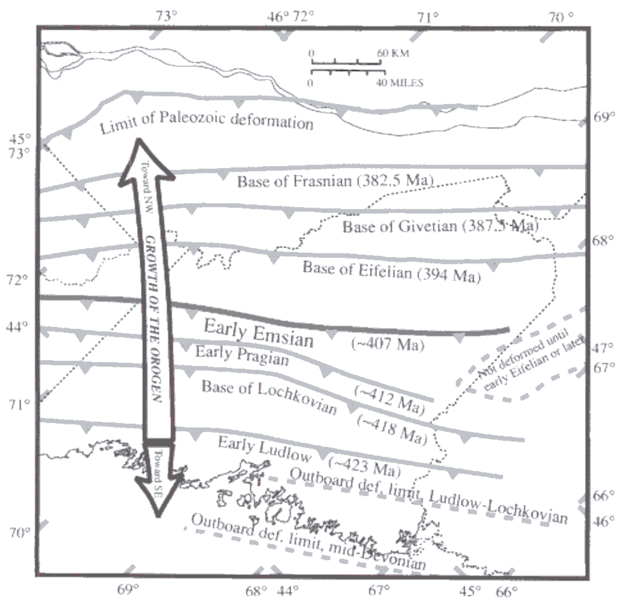

Figure 2. Nonpalinspastic map of Maine and adjacent areas showing sequential positions of the Acadian deformation front. Adapted from Bradley et al. (2000).

chronology securely dates some of the deformation as early Emsian. In western Maine and northern New Hampshire, Solar et al. (1998) and Eusden et al. (2000) obtained early Emsian ages on several plutons that they similarly interpreted as syntectonic (e.g., Phillips and Wamsutta plutons in fig. 3).

Geologic Relations Near the Deformation Front

A boundary between deformed and undeformed rocks must have lain somewhere between the Emsian foreland basin and the syntectonic to posttectonic Emsian plutons. The deformation front was south of the 406–407-Ma Traveler Rhyolite (concordant U/Pb zircon ages; Rankin and Tucker 1995), which conformably overlies and was folded along with the foreland-basin siliciclastic sequence. The Traveler consists of about 3200 m of rhyolitic, subaerial ash-flow tuffs that occupy the remnants of a caldera. The Traveler is intruded along its southern contact by the Katahdin pluton, which has been dated at 406.9 ± 0.4 Ma (Rankin and Tucker 1995). The Katahdin and Traveler have long been regarded as the volcanic and plutonic parts of the same igneous complex (e.g., Hon 1980; Rankin 1994), and this view has been confirmed by the concordant zircon dates. Rankin and Hon (1987) have estimated that the Traveler-Katahdin magmatic center was 65 km long and 15–25 km wide.

Contact relations of the Katahdin pluton are inconsistent. Neuman and Rankin (1994) interpreted the Katahdin pluton as posttectonic because the granitic rocks lack a tectonic fabric and the pluton truncates regional-scale folds that deform Lower Devonian rocks (fig. 6). Along the pluton's northeastern contact at Kelloch Mountain (fig. 6), however, Neuman (1967) mapped a zone of plastically deformed metasedimentary rock fragments, set in a gneissic granitic matrix. We observed C-S fabrics here that suggest syntectonic intrusion: granitic melt defines the "C" shear surfaces, and a biotite foliation defines the oblique "S" surfaces. South of Harrington Lake (fig. 6), field relations are best interpreted in terms of pretectonic emplacement. Here, the pluton intrudes nearly flat-lying, contact-metamorphosed Lower Devonian siltstones and sandstones. Acadian cleavage is pervasive in the

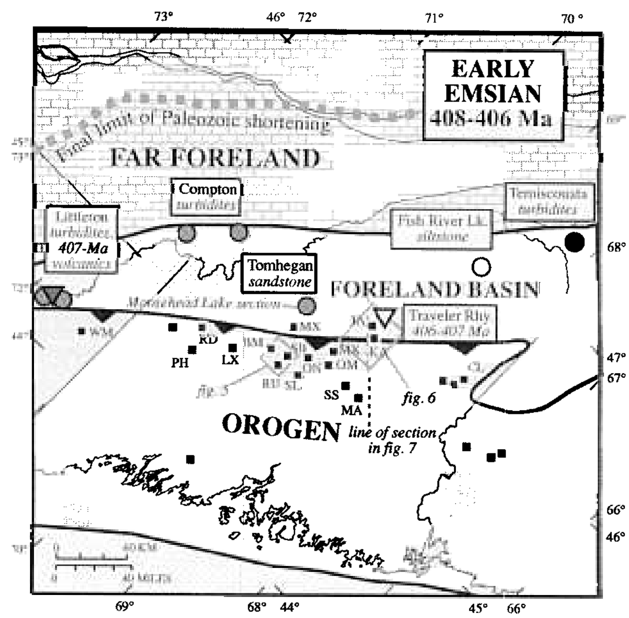

Figure 3. Nonpalinspastic paleogeographic map of Maine and adjacent areas showing the positions of the Acadian orogen and foreland basin during early Emsian time. Toothed line shows inferred position of deformation front, which was probably a diffuse zone of blind thrusts rather than a single emergent thrust. Circles show Emsian fossil locations, and triangles show dated volcanic rocks. Black squares identify selected Emsian plutons with the following abbreviations: *WM*, Wamsutta; *PH*, Phillips; *RD*, Redington; *BM*, Bald Mountain; *RU*, Russell Mountain; *SB*, Shirley-Blanchard; *SL*, Sebec Lake; *ON*, Onawa; *MX*, Moxie; *OM*, Ore Mountain; *HA*, Harrington; *KA*, Katahdin; *SS*, Seboeis; *MA*, Mattamiscontis; *CL*, Cochrane Lake. Adapted from Bradley et al. (2000).

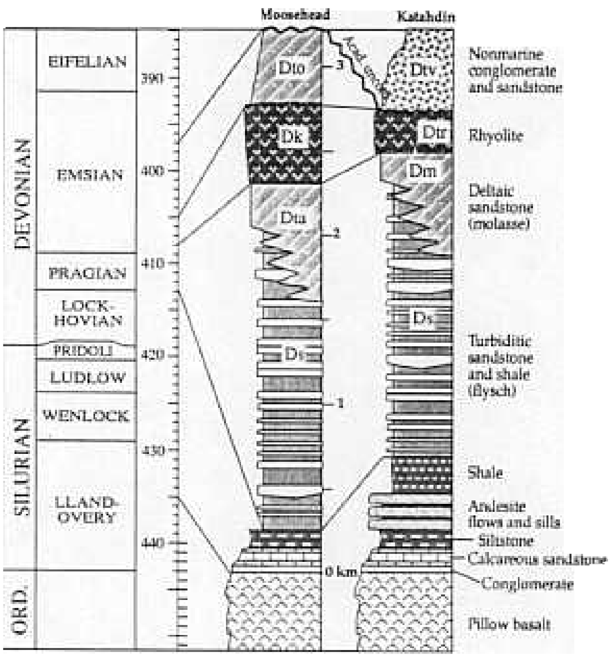

Figure 4. Stratigraphic columns for the Katahdin and northern Moosehead Lake areas. Figure 3 shows locations of Moosehead column; Katahdin column is a composite from northwest of the Katahdin pluton in the area of figure 6. Abbreviations for rock units: *Dto*, Tomhegan Formation; *Dk*, Kineo Rhyolite; *Dta*, Tarratine Formation; *Ds*, Seboomook Group; *Dtv*, Trout Valley Formation; *Dtr*, Traveler Rhyolite; *Dm*, Matagamon Sandstone. Time scale is from Tucker et al. (1998) and Tucker and McKerrow (1995).

region, but within the Katahdin's contact aureole, delicate sedimentary structures are preserved that show no signs of any cleavage, pre-, syn-, or post-hornfels. We thus suggest that these rocks were baked before ever having acquired the regional cleavage. Taken together, conflicting field relations around the Katahdin pluton are best interpreted in terms of emplacement within a broad deformation front.

The Emsian deformation front also can be closely located in two other areas. Near Littleton, New Hampshire (fig. 3), it must have lain between the volcanic member of the Littleton Formation (part of the predeformation foreland succession, dated at 407 Ma by Rankin and Tucker 2000) and the 408-Ma, syntectonic Wamsutta diorite some 40 km to the east (Eusden et al. 2000). At Moosehead Lake (fig. 3), the deformation front must been between the Kineo Rhyolite, which is part of the predeformation foreland stratigraphy, and the Moxie pluton 20 km to the south, which yielded a $^{40}Ar/^{39}Ar$ biotite plateau age of 406.3 ± 3.8 Ma (Bradley et al. 2000). Contact relations of the Moxie are quite variable along its ~90-km strike length (fig. 6). The pluton appears to be pretectonic at Jo Mary Mountain, where nearly flat-lying, contact-metamorphosed Devonian flysch preserves delicate sedimentary structures and not a hint of cleavage (Hanson and Bradley 1989). The pluton appears to be posttectonic at Moosehead Lake (fig. 6), where contact-metamorphic andalusite overgrew an axial-planar cleavage associated with nearly recumbent mesoscale folds. Finally, at the pluton's western end, it appears to be locally syntectonic because contact-metamorphic rocks are strongly foliated and contain aligned andalusite porphyroblasts with biotite strain shadows (Bradley et al. 2000, p. 29). As for the Katahdin, the variable contact relations of the Moxie pluton suggest emplacement within a broad deformation front.

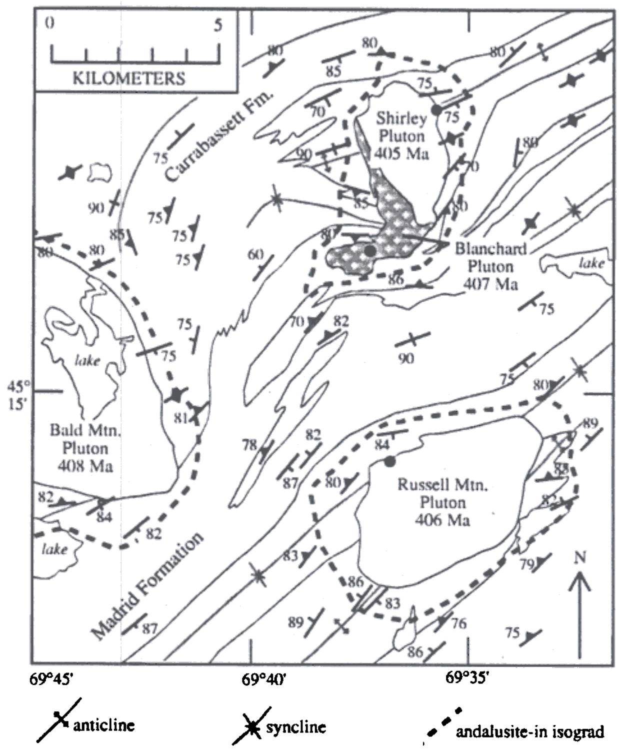

Figure 5. Simplified geologic map of the Shirley-Blanchard and Russell Mountain plutons showing that the Emsian plutons truncate tight Acadian folds and axial-planar cleavage in the Lower Devonian Madrid and Carrabassett Formations. Adapted from Espenshade and Boudette (1967) and Ludman (1978).

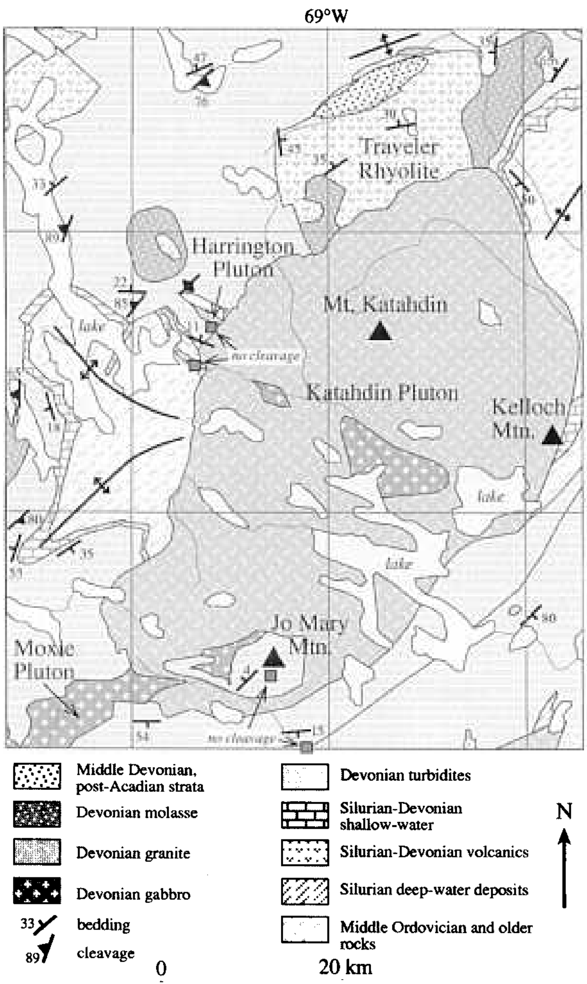

Figure 6. Geologic map of Katahdin region showing locations mentioned in text. "No cleavage" localities near the Katahdin pluton are contact-metamorphosed strata rocks in which delicate sedimentary structures are preserved. Adapted from Osberg et al. (1985), Griscom (1976), and Rankin (1961).

Early Emsian Reconstruction of the Orogenic Wedge

The cross section in figure 7 shows an early Emsian transect through the Katahdin area. Within the orogen, plutons were emplaced into already-deformed Devonian rocks. The pluton depicted is shown at a depth of about 6.5 km based on *PT* calculations from the contact aureole of the Onawa pluton (Symmes and Ferry 1995). The deformation front, separating orogen from foreland basin, was probably not a sharp boundary at an emergent frontal thrust. Rather, we envision a diffuse zone of blind thrusts beneath growth folds, further complicated by volcanism at the surface and plutonism at depth. Meanwhile, flysch and molasse were deposited in a foreland basin that lay to the northwest. Bradley et al. (2000) estimated this (telescoped) width of early Emsian foreland basin at about 90 km; subject to restoration of post-Emsian shortening, the original width was perhaps 180 km. The best-fit distribution of Lower Devonian climatically sensitive sediments from hundreds of localities throughout Euramerica suggests that all this took place at about lat 40°S (Witzke and Heckel 1988). The height of the Acadian mountains, the taper of the orogenic wedge, and the synorogenic climatic regime have not been investigated.

Without an improved Devonian time scale (Tucker et al. 1998), it would have been impossible to construct the early Emsian paleogeographic map and cross section (figs. 3, 7). According to time scales that were in general use until a few years ago (e.g., Palmer 1983; Harland et al. 1990), the 404–408-Ma igneous rocks would be Lochkovian in age and thus would be coeval with an episode of shallow-marine carbonate sedimentation in northern Maine. Instead, we are able to correlate the plutons with flysch and molasse derived from the Acadian orogen. Conversely, using the older time scales, Emsian foreland-basin strata dated by fossils would be mistakenly correlated with a ~390-Ma magmatic lull in the orogen. Any time-scale calibration problem will likewise have repercussions on tectonic interpretations of two separate time intervals. This is not meant to cast aspersions on the older time scales but to point out that the time scale is always a critical

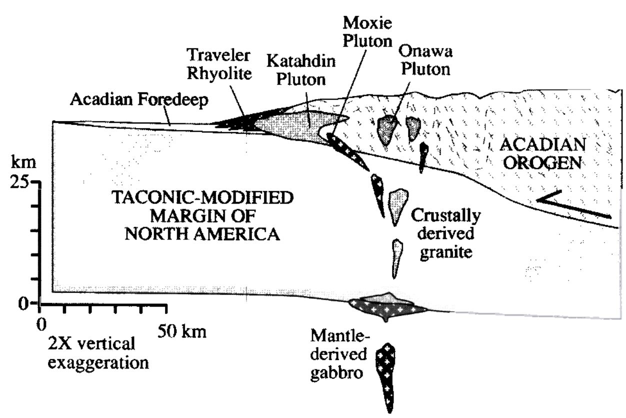

Figure 7. Schematic cross section through the Acadian deformation front during early Emsian time, in the area of Mount Katahdin. Key rock units are exaggerated in extent.

link—and sometimes a very weak one—in tectonic analysis that involves both isotopic and fossil ages. The revised time scale also satisfies problems with Acadian chronology in New Hampshire that were first perceived by Armstrong et al. (1992).

Paleogeographic Setting of Magmatic-Sulfide Mineralization

Two gabbroic plutons of north-central Maine—the Moxie and Ore Mountain—are known to host magmatic sulfide accumulations (Thompson 1984, p. 477–478). The Ore Mountain pluton contains an estimated 180 million tonnes of pyrrhotite (Thompson 1980). A gossan developed by weathering of this ore body was mined at Katahdin Iron Works between the 1830s and 1920s (Sawtelle 1988). New U/Pb geochronology demonstrates that magmatic-sulfide mineralization took place in the early Emsian.

The Ore Mountain pluton is a small stock located a few kilometers northeast of the Onawa gabbro-granodiorite pluton. It also has been referred to as the Katahdin gabbro (Thompson 1984) and as the Silver Lake pluton. Surface exposures are limited, but Mike Foose of the U.S. Geological Survey was able to provide us with drill core samples of gabbro for dating. The emplacement age was determined by the U-Pb zircon method using refinements of the procedures developed by Krogh (1973, 1982) and described previously by Tucker et al. (1998). Interested readers are referred to these articles, as well as footnotes a–f in table 1, for a more complete description of our analytical techniques. We report two results. One analysis, consisting of four grains of faceted zircon, is concordant with a $^{207}Pb/^{206}Pb$ age of 406.1 ± 1.9 Ma (1σ). A second analysis, consisting of five grains, yielded a less precise but similar $^{207}Pb/^{206}Pb$ age of 403.9 ± 3.7 Ma (1σ). As shown in figure 8, both analyses overlap concordia and define a weighted mean $^{207}Pb/^{206}Pb$ date of 405.6 ± 3.3 Ma (95% confidence). This is our estimate of the age of gabbro emplacement and magmatic-sulfide mineralization at Ore Mountain.

The high-precision U/Pb date allows us to correlate the Ore Mountain gabbro with the other early Emsian rocks described above. From the regional relations, we infer that pluton emplacement took place in the physical settings shown in figures 3 and 7; the plutons were emplaced into the orogenic wedge about 30 km behind the contemporaneous deformation front. The Ore Mountain pluton, being of the same age and occupying a similar across-strike position as the Onawa pluton, was probably emplaced at about the same depth, about 6.5 km.

Cause of Emsian Magmatism

Several key points follow from the Emsian paleogeography highlighted in this article and other aspects of the Late Silurian to Late Devonian record as discussed by Bradley et al. (2000): (1) Whatever their local contact relations, the Emsian plutons and volcanics are broadly synorogenic. Acadian orogenesis lasted from at least Ludlow (Late Silurian) to Givetian (Late Devonian) in northern New England and adjacent Canada. During this ~40-m.yr. interval, the deformation front advanced some 240 km toward the craton (present distance; at least twice as far on a palinspastic base) (Bradley et al. 2000). The early Emsian magmatic pulse took place about midway through this lengthy process. (2) The 408–404-Ma igneous province includes both

Table 1. U-Pb Isotope Dilution Analyses for Ore Mountain Pluton

	Fractions		Concentrations			Atomic ratios					
No.	Properties[a]	Weight (μg)[b]	Pb rad (ppm)[b]	U (ppm)[b]	Pb com (pg)[c]	Th/U[d]	$^{206}Pb/^{204}Pb$[e]	$^{207}Pb/^{206}Pb$[f]	$^{207}Pb/^{235}U$[f]	$^{206}Pb/^{238}U$[f]	Age (Ma): $^{207}Pb/^{206}Pb$[f]
	4; cl, c, pr, f	7	49.3	707	1.6	.593	12,649	.05485 ± 5	.4915 ± 7	.06499 ± 9	406.1 ± 1.9
2	5; cl, c, pr, f	5	17.8	261	3.2	.506	1668	.05480 ± 9	.4910 ± 10	.06499 ± 10	403.9 ± 3.7

[a] All analyses are of zircon. Cardinal number indicates the number of zircon grains analyzed (e.g., 10 = 10 grains). All zircon grains were selected from nonparamagnetic separates at 0° tilt at full magnetic field in a Frantz magnetic separator; +200 = size in mesh (>75 μm); c = colorless; cl = clear; f = fragment; pr = prismatic. All grains were air abraded following Krogh (1982).
[b] Concentrations are known to ±30% for sample weights of about 30 μg and to ±50% for samples <3 μg.
[c] Corrected for 0.0125 mol fraction common-Pb in the $^{205}Pb/^{235}U$ spike.
[d] Calculated Th/U ratio assuming that all ^{208}Pb in excess of blank, common-Pb, and spike is radiogenic ($\lambda^{232}Th = 4.9475 \times 10^{-11}$ yr^{-1}).
[e] Measured, uncorrected ratio.
[f] Ratio corrected for fractionation, spike, blank, and initial common-Pb (at the determined age from Stacey and Kramers 1975). Pb fractionation correction = 0.094%/amu (±0.025%, 1σ); U fractionation correction = 0.111%/amu (±0.02%, 1σ). U blank = 0.2 pg; Pb blank ≤10 pg. Absolute uncertainties (1σ) in the Pb/U and $^{207}Pb/^{206}Pb$ ratios calculated following Ludwig (1980). U and Pb half-lives and isotopic abundance ratios from Jaffey et al. (1971).

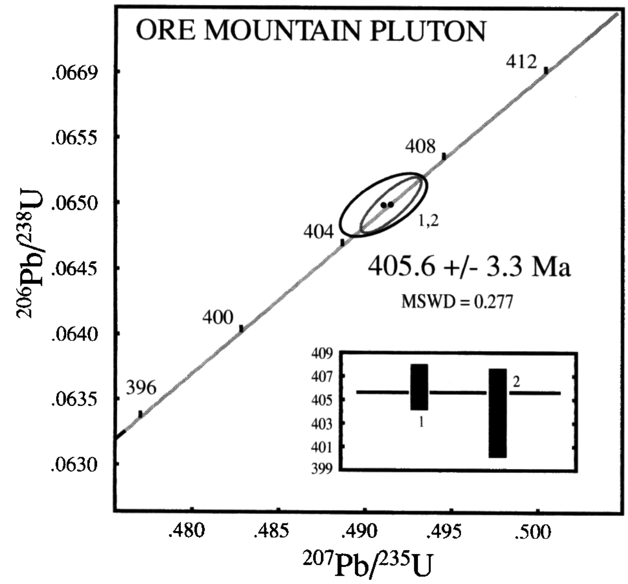

Figure 8. U/Pb concordia diagram of zircon from gabbro, Ore Mountain pluton, Maine. See table 1 for analytical data.

gabbros (e.g., Moxie) and granites (e.g., Katahdin). The granites, which are calc-alkaline, have been interpreted in terms of partial melting of moderately radiogenic continental crust (Ayuso 1986), but the gabbros, which may have provided some of the heat to form the granite magmas, clearly must have a mantle component. Emsian magmatism therefore cannot be solely a consequence of collision-induced thickening of continental crust. (3) Emsian magmas originated within and beneath the orogen but were extruded across the deformation front into the foreland basin. The Quaternary volcanic fields in the collisional forelands of Papua New Guinea (Hamilton 1979) and southern Turkey (Pearce et al. 1990) are thus comparable, at least in their physical setting. (4) The 4-million-year Emsian magmatic pulse was one of three relatively short-lived igneous episodes during the Acadian; intervening times were magmatically quiet. Synorogenic magmatism also took place near the Maine coast at around 423–417 Ma (Ludlow to Lochkovian) and in the Eastern Townships of Quebec around 384–375 Ma (Givetian to Frasnian). In each case, plutons intruded into the frontal part of the orogenic wedge, where deformation was either ongoing or had taken place very recently (Bradley et al. 2000). Plutonism continued throughout New England from about 375 to 360 Ma. Although these younger plutons are commonly referred to as "Acadian," they postdated the documented cratonward advance of the deformation front and so cannot be readily linked to Acadian plate convergence. (5) The Emsian volcanic centers are part of a Silurian-Devonian volcanic trend along the northwestern margin of the central Maine deep-water basin (Piscataquis volcanic belt of Rankin 1968 but used here in the broader sense of Bradley 1983). Ludlow, Pridoli, Lochkovian, and Emsian volcanics in this belt are all synorogenic, erupted in both distal and proximal parts of the foreland (Bradley et al. 2000).

Figure 7 is fairly specific about the physical setting of early Emsian magmatism but deliberately vague about its plate-tectonic context. Three alternative tectonic scenarios (fig. 9) seem to provide a plausible rationale for Emsian magmatism.

One model (fig. *9A*, *9B*) is that Emsian magmatism was related to a subduction zone that dipped to the southeast beneath Avalonia. Presumably, this originated as a B-type subduction zone (one in which oceanic lithosphere is subducted; Bally and Snelson 1980), but by Emsian time, it had evolved into an A-type subduction zone (one in which continental lithosphere is subducted). Because it straddles the deformation front, the Emsian magmatic belt could not conceivably represent the magmatic arc situated 100 km above a southeast-dipping slab; this would be akin to arc volcanics somehow erupting on a subducting oceanic plate. Melting thus would have required some other cause, such as slab break-off or, as shown in figure *9B*, lithospheric delamination of the lower plate during collision (Robinson et al. 1998; Tucker et al. 2001). The older (Ludlow through Lochkovian) foreland volcanism has no single compelling explanation, but there are several possibilities: (1) Geochemical discriminant plots, taken at face value, have been interpreted to show that these volcanic rocks were "primarily erupted in a within-plate extensional environment that terminated with limited subduction" (Hon et al. 1992, p. 163). (2) Keppie and Dostal (1994) attributed the older volcanics to so-called transpressive rifting. The transform margin shown in figure *9A* is a variant on Keppie and Dostal's (1994) model. (3) The volcanics are analogous to collision-related volcanics in the orogenic foreland of southern Turkey (Pearce et al. 1990), thought to have formed as a result of extension related to collisional impact (Sengor 1976). (4) Magmatism might conceivably have resulted from flexure of the foreland plate (e.g., Bradley and Kidd 1991). In the latter two scenarios, it is unclear why volcanism due to collision would have been widespread in the distal foreland yet absent from what was then the proximal foreland (Central Maine basin) (fig. *9A*).

A second model (fig. *9C*, *9D*) would place the

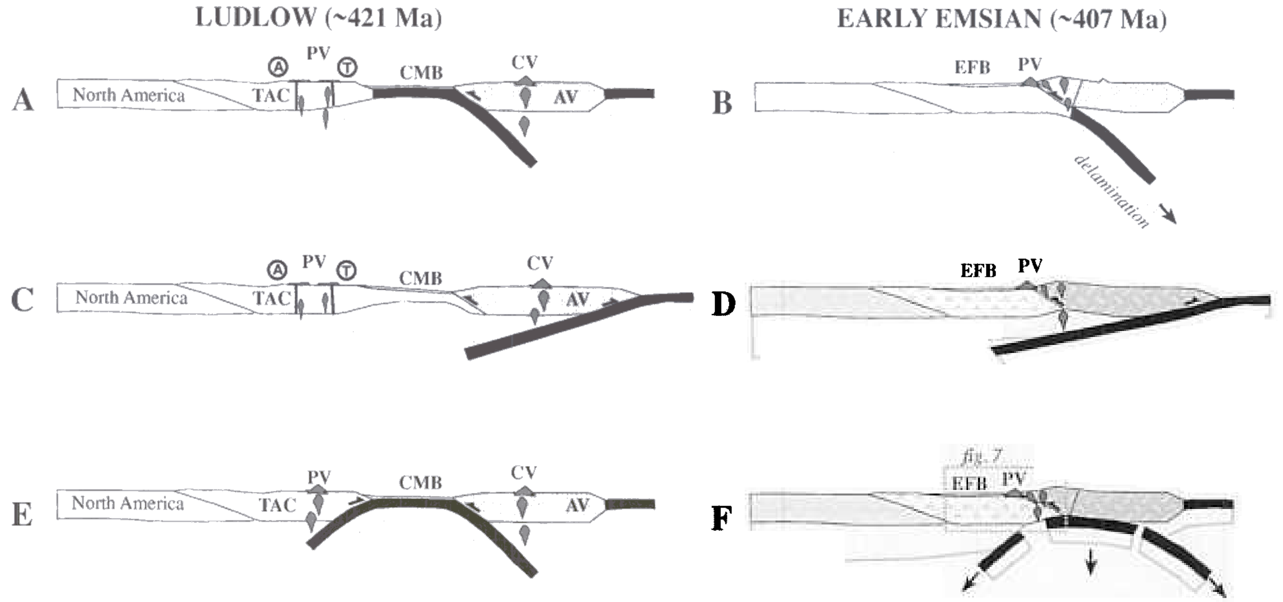

Figure 9. Alternative plate geometries for the Acadian orogeny discussed in the text. In each case, the Ludlow configuration on the left leads to the Emsian configuration on the right. *TAC*, composite arc terranes that collided with North America during the Ordovician Taconic orogeny. *AV*, Avalon composite terrane. *CMB*, Central Maine basin, depicted in (*A*) as a region of thinned continental or arc crust and in (*C*) and (*E*) as oceanic crust. *EFB*, Emsian foreland basin. *PV*, Piscataquis volcanic belt in the sense of Bradley (1983), thus including Ludlow through Emsian rocks. *CV*, Llandovery through Lochkovian Coastal Volcanic belt. *T*, *A*, Strike-slip motion toward and away from viewer, respectively.

Emsian igneous rocks and orogenic belt in a magmatic arc setting—specifically, in the compressional back-arc of a convergent margin. The corresponding "Laramide-type" subduction zone would have dipped gently beneath both Avalonia and North America from a surface trace somewhere in the present Atlantic Ocean (Murphy et al. 1999). This model would require one of the above explanations for the Ludlow through Lochkovian foreland volcanism.

A third possibility (fig. 9*E*, 9*F*) is that the Acadian collision involved two opposed subduction zones (McKerrow and Ziegler 1971; Bradley 1983; Hanson and Bradley 1989; Ludman et al. 1993; Eusden et al. 2000). This scenario thus includes an oceanic plate that has now completely vanished beneath the isoclinally folded flysch of the Central Maine basin. The Emsian magmatic belt would represent the dying phase of an arc above a northwest-dipping subduction zone, which, by the time in question, was being overridden by the accretionary wedge of the other, southeast-dipping subduction zone. The short duration of the Emsian magmatic event might reflect a one-time event such as the break-off (fig. 9*F*) or delamination of the northwest-dipping slab. Ludlow through Lochkovian foreland volcanism along the Piscataquis belt would be related to the soon-to-be-extinct northwest-dipping subduction zone. Intra-arc extension might explain the supposedly extensional geochemical signatures. Such extension could have been related to releasing bends in an arc-parallel strike-slip system (Bradley 1983) or to subduction rollback.

As should be apparent, none of these models are compelling, and the Acadian tectonic framework is no less controversial than it was in the early 1980s. The problem is to find definitive tests. The two-subduction-zone model has been difficult to test because of a dearth of modern examples with this plate geometry—the only two are the Molucca Sea and eastern Papua New Guinea. Moreover, we lack a clear picture of what an ancient system like this ought to look like, especially an exhumed, sediment-choked variant that would be appropriate for the Acadian. Because interpretations of Silurian-Devonian igneous rocks are critical to each of the alternative scenarios, a petrogenetic approach may eventually prove to be the key but not until geochemical discriminant analysis can become sophisticated enough to embrace the various unusual tectonic settings shown or implied in figure 9.

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