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Migration of the Acadian Orogen and Foreland Basin across the Northern Appalachians[1]

By

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ABSTRACT

We reconstruct seven sequential positions of the Acadian deformation front and foreland basin as these features migrated northwestward across Maine from Late Silurian to Middle Devonian time. The reconstructions are based on (1) U/Pb and $^{40}Ar/^{39}Ar$ ages of pre-, syn-, and post-tectonic plutons; (2) conodont and palynomorph ages of key strata that predate or postdate the local age of deformation; and (3) interpretations of the depositional settings of these strata: far-foreland, foreland-basin, inner trench slope, and orogenic hinterland. Tight correlations between isotopically and paleontologically dated events are made possible by recent improvements in the Silurian-Devonian time scale.

In early Ludlow time (~423 Ma), the deformation front must have lain near the present midline of the Fredericton basin, between the post-tectonic Pocomoonshine pluton (423 Ma) to the southeast, and the graptolitic Smalls Falls Formation of the same age, 70 km across strike to the northwest. During early Lochkovian time (~417 Ma), the deformation front was near the present midline of the Central Maine basin, as recorded by the northwestward advance of slope deposits of the Carrabassett Formation across axially-transported turbidites of the Madrid Formation. The early Emsian (406-407 Ma) deformation front lay along or near the Lobster anticlinorium; its position is bracketed between the post-tectonic Russell Mountain pluton (406 Ma) to the southeast and the brachiopod-bearing Tomhegan molasse of the same age, 50 km across strike to the northwest. At the Emsian-Eifelian boundary (~394 Ma), the deformation front was located along the Pennington-Munsungun anticlinorium, northwest of the post-tectonic Mapleton and Trout Valley Formations, but southeast of the youngest nonmarine clastics in the pretectonic Fish River Lake Formation. At the Eifelian-Givetian boundary (~387.5 Ma), the deformation front was probably somewhere near the midline of the Connecticut Valley-Gaspé basin, northwest of a belt of post-tectonic plutons in Québec, including the 384-Ma Scotstown pluton, but southeast of three occurrences of deformed Eifelian carbonates along the basin's northwestern margin. At the Givetian-Frasnian boundary (~382.5 Ma), the deformation front was somewhere to the northwest of these Eifelian carbonate outcrops. The orogen thus migrated toward the northwest about 240 km across strike (present distance) in about 40.5 million years. Meanwhile, on the outboard (southeasterly) side of the orogen, a boundary between deformed and undeformed rocks moved toward the southeast at least 50 km, probably during the Early and (or) Middle Devonian.

The migration pattern of the orogen and foreland basin suggests that during collision, a southeasterly plate that included the Acadian orogenic wedge and its Avalonian backstop overrode the Taconic-modified margin of North America. The implied minimum plate convergence rate is about 6 mm/yr. The actual rate must have been considerably faster because the base map is nonpalinspastic; a more accurate estimate will have to await a careful assessment of Acadian shortening. If shortening reduced Maine to half its pre-Acadian width (a conservative estimate in light of the regional-scale tight to isoclinal folding), this would imply a plate convergence rate of 12 mm/yr, at the slow end of the normal range of modern plate motions. The reconstructed positions of the orogen and foreland basin also constrain the setting of several Silurian and Devonian episodes of volcanism and plutonism, and certain post-D1 Acadian deformations in the Acadian orogenic hinterland.

INTRODUCTION

Three decades after Wilson's (1966) first plate-tectonic interpretation of the Appalachians, a consensus on the plate geometry that led to the Acadian orogeny still has not emerged — though not for lack of trying. In this paper, we present new findings and review old evidence bearing on orogenic timing on a regional scale. By focusing on the syncollisional history, we can make some new headway on the Acadian problem without getting bogged down in controversies regarding the precollisional plate geometry.

Donahoe and Pajari (1973) presented evidence that the Acadian orogeny was diachronous across strike, beginning in the Early Devonian along the Maine-New Brunswick coast, and ending in the Middle Devonian in Québec, near the Gulf of the St. Lawrence. This finding was based on the age and distribution of fossiliferous strata known to predate and postdate deformation, and on isotopic ages of post-tectonic plutons. Five developments make it worthwhile to revisit this topic. First, the Silurian-Devonian time scale has undergone major revisions (Tucker and McKerrow, 1995; Tucker, and others, 1998). The new time scale is calibrated by U-Pb zircon ages of ash beds with good biostratigraphic control; most of the series and stage boundaries are 10-20 million years older than on the time scale available to Donahoe and Pajari (1973), and the durations of individual series and stages are very different as well. Second, many Acadian plutons have now been reliably dated by U/Pb and $^{40}Ar/^{39}Ar$ methods. The new dates, with errors of 1-3 million years, reveal that most of the Rb/Sr and conventional K/Ar ages used in Donohoe and Pajari's (1973) analysis were about as far off as the time scale. Third, certain key stratigraphic units that bear on the position of the deformation front through time have now been dated by conodonts or spores with much greater precision and accuracy than was ever possible using brachiopods, corals, and plants. Fourth, paleocurrent directions and depositional environments are now known for several key stratigraphic units, such as the Madrid and Carrabassett Formations (Hanson and Bradley, 1989; 1993); this has helped us to locate the foreland basin during two time intervals — and by inference, to locate the deformation front at the same times. Finally, as will be shown, current understanding of orogenic wedges, flexural foreland basins, and orogenic backstops makes it possible to glean more information from conventional stratigraphic evidence about orogenic timing than was possible two decades ago.

New U/Pb, $^{40}Ar/^{39}Ar$, conodont, and palynological data, which were presented by Bradley and others (1998), are documented here. These new results, and previously published data, enable us to track six sequential positions of the Acadian deformation front as a wave of D1 deformation migrated cratonward across the northern Appalachians during Late Silurian to Middle Devonian time. This information has implications for (1) the rates, trajectories, and amounts of plate convergence, (2) the syncollisional plate geometry, (3) strike-slip and thrust partitioning during collision, and (4) the relationship between various suites of igneous rocks and the orogenic belt as it existed at the time of magmatism.

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REGIONAL GEOLOGY

The study area is the Acadian-deformed belt of Maine, plus adjacent parts of Vermont, New Hampshire, the Eastern Townships of Québec, and New Brunswick. A good description of the regional geology was given by Osberg and others (1989). The area of figure 1 is subdivided into ten map units, seven of which are relevant to Acadian timing. Middle Ordovician and older rocks are grouped together in figure 1, because they provide no useful constraints on the age of Acadian tectonism.

Deep-water Silurian strata define four basins: the Connecticut Valley-Gaspé basin, Aroostook-Matapedia basin, Fredericton basin, and Central Maine basin (fig. 2). The latter two have been identified by some workers (Bradley, 1983; Ludman and others, 1993) as the former site of an ocean whose closure resulted in the Acadian orogeny; this interpretation remains controversial.

Flanking the deep-water basins are four tracts where pre-Silurian rocks are now exposed. These belts — Avalonia, the Bronson Hill-Pennington anticlinorial belt, Miramichi, and the Taconic orogen — were either undergoing active erosion during at least part of the Silurian, or were the sites of shallow-water deposition. The shallow-water deposits include carbonate and locally-derived siliciclastic rocks that range in age from Llandovery to Eifelian. One focus of the present study was to obtain precise conodont ages from the youngest carbonate rocks in various key stratigraphic sections.

From Late Silurian to Middle Devonian time, much of the area of figure 1 was inundated by a thick succession of clastic rocks, which we regard as the fill of a migrating Acadian foreland basin (Bradley, 1983, 1987, 1997; Hanson and Bradley, 1989). The foreland-basin succession includes all of the Devonian flysch and Devonian molasse, and part of the deep-water Silurian sequence. For want of sedimentological studies, the Silurian deep-water deposits cannot yet be confidently subdivided into foreland-basin and pre-foreland-basin successions. A second focus of the present study was to obtain palynological ages from key foreland-basin deposits.

Silurian and Devonian volcanic rocks occur in two broad belts (figs. 1 and 2). The Coastal volcanic belt erupted into basement of the Avalonian terrane along the coast of Maine and New Brunswick. Farther north lies a second belt of Silurian-Devonian volcanics, which crop out along parts of the Bronson Hill-Pennington anticlinorial belt, Aroostook-Matapedia basin, Miramichi anticlinorium, and Connecticut Valley-Gaspé basin. The northerly belt has been called the Piscataquis volcanic belt (or magmatic belt) in New England (Bradley, 1983) and the Tobique volcanic belt in Canada (Keppie and Dostal, 1994). The Piscataquis volcanics were erupted in a belt that had been accreted to North America during the Ordovician.

In all but the northernmost part of figure 1, the rocks were deformed during a Silurian to Devonian orogeny that can be termed "Acadian" in the loose sense favored here. Where deformation was polyphase, the age of the first deformation (D1) is of greatest interest for present purposes. The style and intensity of Acadian deformation is quite variable across the vast area of figure 1. The common thread, however, is that it was dominantly contractional. In the low-grade rocks of northern and central Maine, Acadian structures are mostly upright, tight to isoclinal folds. At higher metamorphic grades in central New Hampshire, east-directed fold nappes predominate (Eusden and others, 1996), but in southern New Hampshire, thrust- and-fold nappes have westerly vergence (Thompson and Robinson, 1968). Some strike-slip faulting also occurred during Devonian times, as will be discussed below.

Silurian and Devonian (Acadian) plutons, ranging in composition from gabbro to granite, are widespread in the area of figure 1; many are demonstrably syn- or post-tectonic. A major focus of the present study was to date key plutons across a broad swath from the Maine coast to the

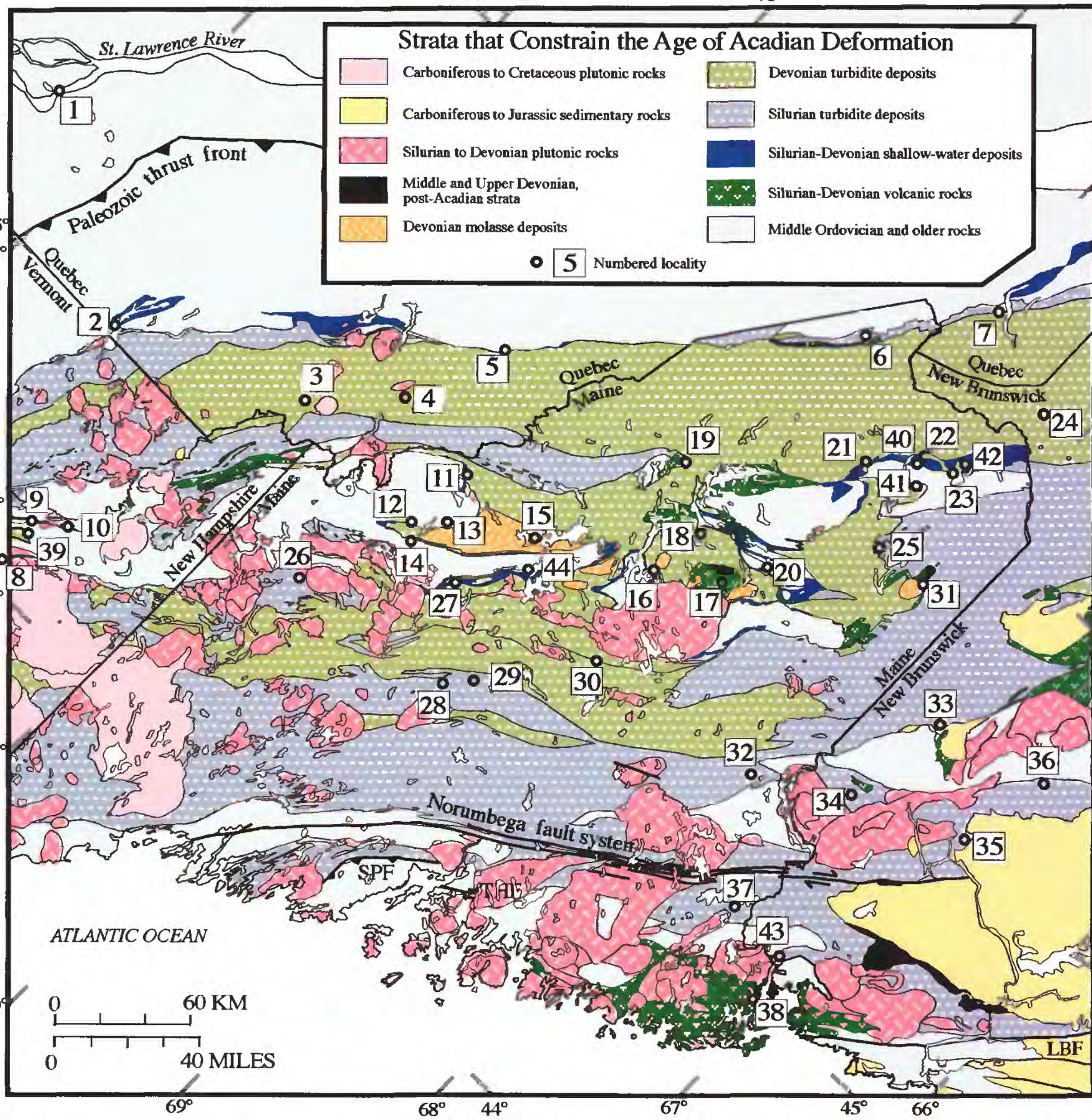


Figure 1. Geologic map of Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, emphasizing the distribution of strata that help constrain the age of Acadian foreland-basin sedimentation and (or) Acadian deformation. Numbered localities are keyed to figure 4. Selected faults are abbreviated as follows: LBF - Lubec-Belleisle fault; SPF - Sennebec Pond fault; THF - Turtle Head fault.

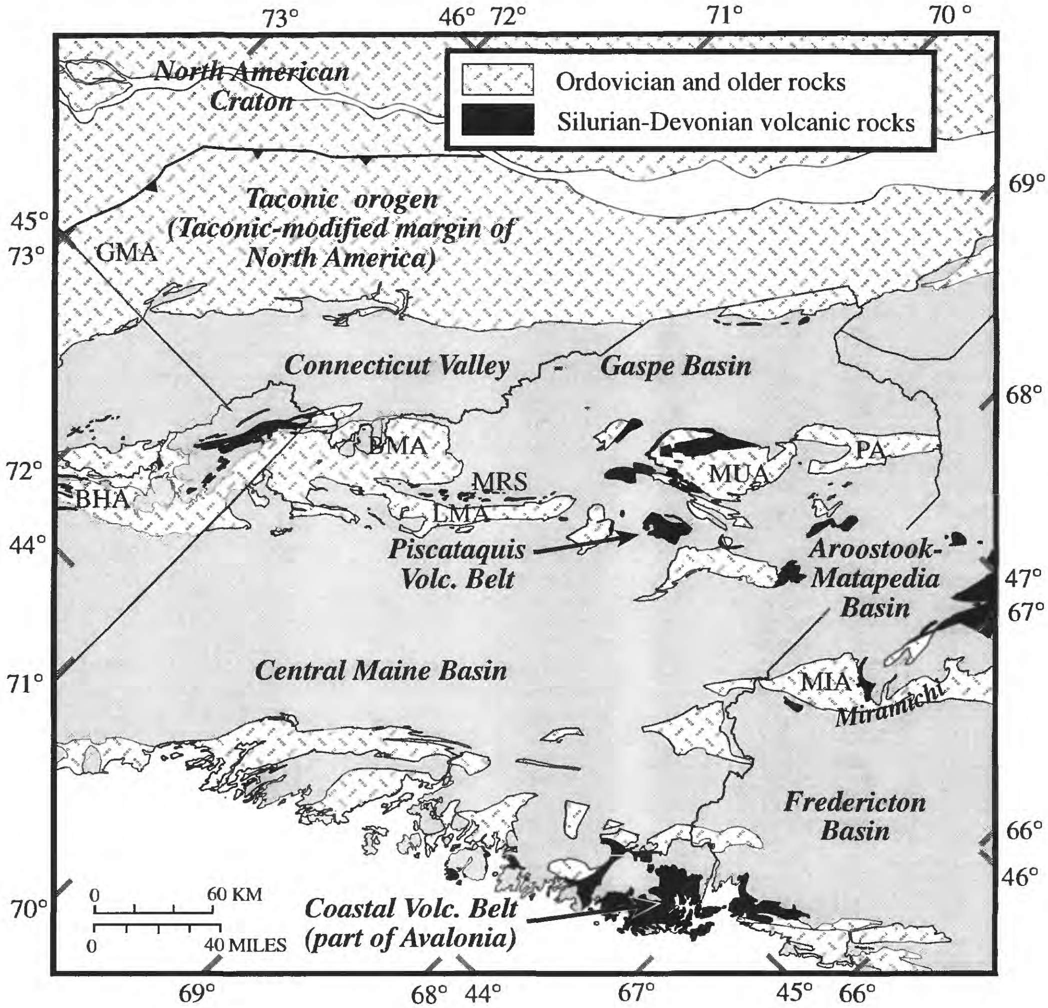


Figure 2. Map of the study area showing the distribution of major Silurian-Devonian paleogeographic elements (labeled in italics). Selected structural features are abbreviated as follows: BHA - Bronson Hill anticlinorium; BMA - Boundary Mountains anticlinorium; GMA - Green Mountain-Sutton Mountain anticlinorium; MRS - Moose River synclinorium; LMA - Lobster Mountain anticlinorium; MIA - Miramichi anticlinorium; MUA - Munsungun anticlinorium; PA - Pennington anticlinorium.

Maine-Quebec border, using the U/Pb and $^{40}Ar/^{39}Ar$ methods. As will be shown, the locus of plutonism shifted through time.

Of the various post-Acadian rocks in the area of figure 1, only Middle and Upper Devonian sedimentary and volcanic rocks that overlie Acadian-deformed rocks are relevant to Acadian orogenic timing. In this paper, we present new palynological ages for two key post-Aacadian units, ages that previously had been documented in unpublished reports of the Geological Survey of Canada. Other post-Acadian rocks in the area of figure 1 include Carboniferous to Cretaceous plutons, and Carboniferous to Jurassic sedimentary rocks.

DETERMINING THE TIMING OF COLLISIONAL OROGENY FROM THE ROCK RECORD

Traditionally, orogenesis in a given place is judged to be younger than the youngest deformed strata, but older than the oldest unit that unconformably overlies the deformed rocks. For most formations in the present study area, we know considerably more than whether or not they participated in Acadian deformation: interpretations of the tectonic setting of the dated rocks can supplement the more traditional evidence for timing of orogeny.

A contractional orogenic system has three components: (1) an *underriding plate*, (2) an *overriding plate*, and between them, (3) an actively deforming *orogenic wedge* (fig. 3). Two broad depositional regimes exist side-by-side on the underriding plate. The *foreland basin* is a marine or nonmarine sedimentary basin that flanks the orogen and is typically filled with orogenically-derived sediment (the terms *foreland basin* and *foredeep* are sometimes used interchangeably, but a *foredeep,* more properly, is an underfilled foreland basin characterized by deep-marine flysch sedimentation). The term *far-foreland* is coined here for the region beyond the limit of orogen-derived sedimentation. A *forebulge* — abroad, gentle uplift formed by lithospheric flexure — may or may not be present in the distal part of the foreland basin, or in the far-foreland.

The *deformation front* is the surface expression of the boundary between the underriding plate and orogenic wedge; it may be a sharp boundary at an emergent thrust-fault, or a broad zone of folding. Much, but not all, of the convergence between the underriding and overriding plates typically occurs at and just to the rear of the deformation front, and is manifested as the D1 deformation event that is of greatest interest here. Some additional convergence may be taken up further to the rear, within the orogenic wedge; this can account for some D2 and younger deformations. The boundary between the actively deforming orogenic wedge and the relatively rigid — or even extensional —*hinterland* on the overriding plate is called the *backstop.*

The model just outlined is sufficiently generalized that it applies to several plate-tectonic regimes, such as arc-passive margin collision, Andean-style foreland contraction, and arc-arc collision. Figure 3 might be thought of as a single frame of a moving picture. Most of the relevant Silurian-Devonian strata in the Acadian orogen can be placed with at least fair confidence in figure 3. In the descriptions of stratigraphic sections that follow, we focus not only on the conventional stratigraphic evidence (pre-orogenic versus post-orogenic), but also on datable events that enable us to place each section in the far-foreland, foreland-basin, orogenic wedge, or hinterland at a particular time. We use only those age constraints provided by fossils *where they occur*; the New England tradition of extrapolating fossil age determinations across strike prevents the recognition of diachronous facies.

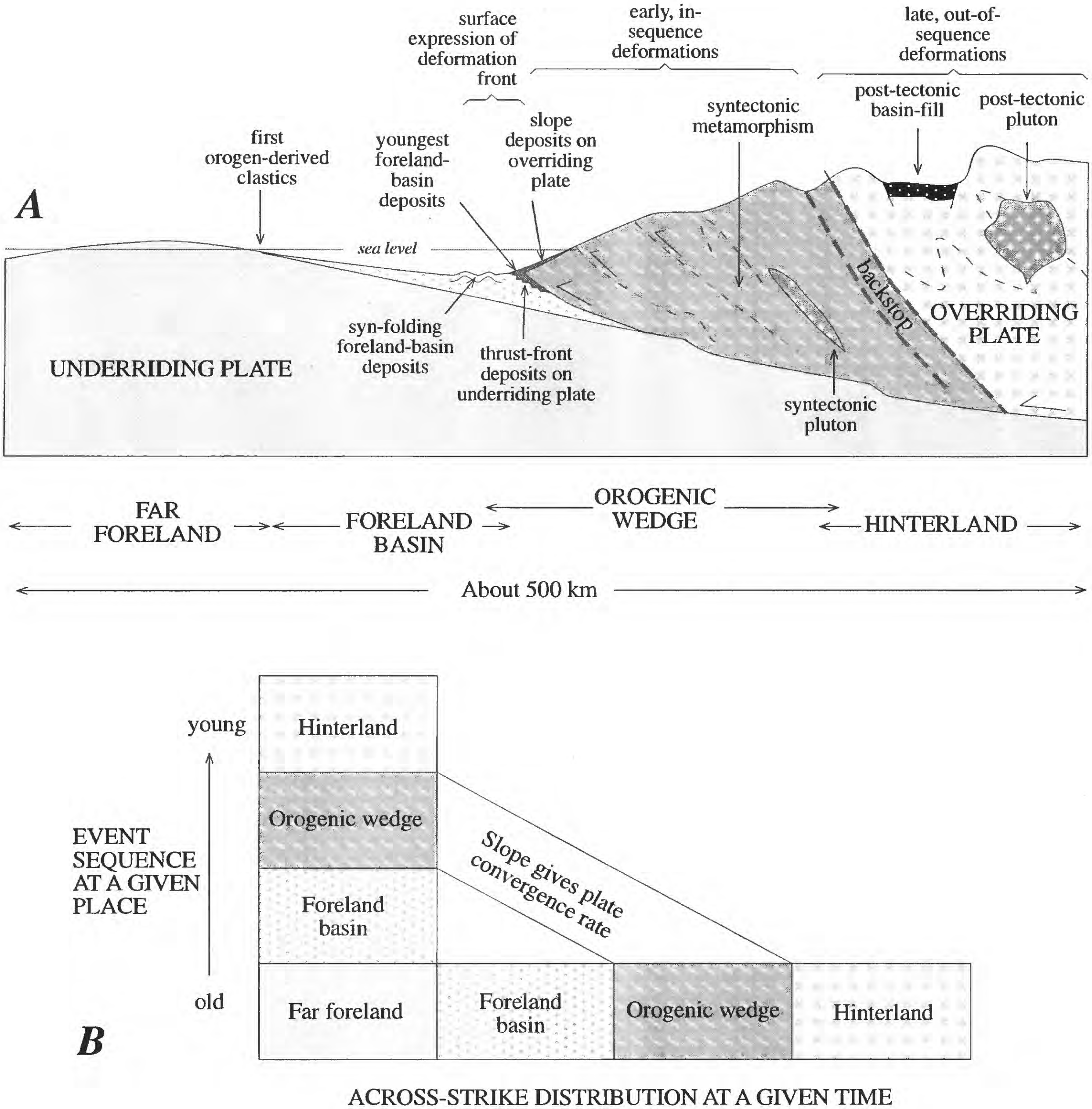


Figure 3. Conceptual model of a two-plate collision showing the positions of the far foreland, foreland basin, orogenic wedge, and hinterland, as discussed in the text. Plate convergence causes each of these paleogeographic elements to migrate to the left through time, at a rate similar to the plate convergence rate. This gives rise to the succession of depositional and deformational events in a given rock sequence.

PALEONTOLOGICAL AND STRATIGRAPHIC EVIDENCE FOR TIMING OF ACADIAN OROGENESIS

PASSAMAQUODDY BAY, MAINE AND NEW BRUNSWICK

The Coastal volcanic belt section of southeastern Maine (figs. 1 and 4, loc. 38) and southwestern New Brunswick consists of 8 km of interbedded bimodal volcanic and fossiliferous sedimentary rocks, of late Llandovery to Lochkovian age. The following highlights (from Gates and Moench, 1981) are key to Acadian tectonic problems. The late Llandovery Quoddy Formation, mostly graptolitic shale, contains a few tuffaceous horizons that record the first known pulse of Silurian volcanism in the belt (fig. 4, col. 38). Mixed volcanism and ever-shallowing marine sedimentation continued through the Wenlock, Ludlow, Pridoli, and the first part of the Lochkovian (Dennys, Edmunds, Leighton, Hershey, and lower Eastport Formations). In New Brunswick, the upper part of the Eastport Formation consists of fluvial sandstone, siltstone, and conglomerate (Pickerill and others, 1978). The Eastport Formation is the youngest pre-Acadian stratified unit; it has been assigned a Lochkovian age, based on a restricted fauna of Lingulas, gastropods, pelecypods, and ostracodes (Berdan, 1971; Pickerill and Pajari, 1976). It should be noted in passing that the Castine Volcanics in the Penobscot Bay region, which were once regarded as Silurian and thus part of the Coastal Volcanic Belt, have recently yielded a U-Pb zircon age of 505 Ma (R.D.Tucker unpublished data) and are thus not very relevant to the Acadian orogeny. All of the Coastal volcanic units were involved in Acadian folding and development of a regional ENE-striking cleavage (Pickerill and others, 1978, p. 42). The Red Beach pluton has yielded two conflicting isotopic ages, both of them suspect: a Givetian Rb/Sr isochron age of 385±6 Ma (Spooner and Fairbairn, 1970; age recalculated herein), and a Lochkovian U/Pb age of 415±6 Ma (Jurinski, 1990). The nominal U/Pb age is likely too old, because the pluton truncates regional-scale Acadian folds that deform the Lochkovian Eastport Formation (Abbott, 1991); on the other hand, Rb/Sr ages from coastal Maine plutons have consistently proven to be too young. Accordingly, the age of folding cannot be determined with confidence. The Perry Formation unconformably overlies both the Eastport Formation and the Red Beach pluton. The basal conglomerate of the Perry Formation contains boulders of the unmistakable Red Beach granite. Plant fossils suggest a general Late Devonian age for the Perry Formation (Kasper and others, 1988, p. 127); an ash bed in the Perry Formation was processed for datable zircons but was barren.

FREDERICTON TROUGH, MAINE AND NEW BRUNSWICK

The thick, isoclinally folded turbidites of Fredericton Basin are devoid of fossils in easternmost Maine (figs. 1 and 4, loc. 37) (Ludman and others, 1993). Nonetheless, some useful information regarding the timing of Acadian deformation can be gleaned. The youngest turbidite unit, the Flume Ridge Formation of presumed Silurian age (Ludman and others, 1993), was penetratively deformed prior to being intruded by the 423±2 Ma Pocomoonshine pluton (see below). Graptolites from turbidites at loc. 35 (figs. 1 and 4) reveal, however, that sedimentation was still taking place in the northwestern part of the basin at this time. Specifically, the Burtts Corners Beds have yielded graptolites ranging in age from the zones of *C. linnarssoni* (Wenlock) to *M. nilsonni* (early Ludlow) (Fyffe, 1995, p. 352-353).

CANTERBURY INLIER, NEW BRUNSWICK

The Canterbury inlier (figs. 1 and 4, loc. 34) lies along the southeastern flank of the Miramichi anticlinorium. It is separated by the Pokiok batholith from the main part of the Fredericton basin, and the stratigraphy is quite different. The Silurian-Devonian section is poorly fossiliferous, but the youngest stratified unit (Hartin Formation), which is of greatest interest here, is dated reasonably well. The Hartin Formation is composed of sandstone and slate, plus minor limestone, conglomerate, and felsic volcanic rocks. A brachiopod fauna indicates a Helderbergian age (that is, Lochkovian)(Boucot cited in Venugopal, 1979, p. 17). The Hartin Formation has been contact

Figure 4 (next two pages). Stratigraphic sections bearing on the age of Acadian deformation and (or) foreland-basin sedimentation in the area of figure 1. Numerical ages of the various stage boundaries are from the time scale of Tucker and others (1998). Spore zones are from Richardson and McGregor (1986). Abbreviations for rock units are as follows. Dbk—Beck Pond Ls.; Dbr—Bear Pond Ls. Mbr.; Dc—Carrabassett Fm.; Dch—Chapman Ss.; Dco—Compton Fm.; De—Eastport Fm.; Deh—Edmunds Hill Andesite; Df—Famine Lmst.; Dfp—Frost Pond Fm.; Dh—Hildreths Fm.; Dha—Hartin Fm.; Dhr—Hersey Fm.; Dhv—Hedgehog Volcanics; Dk—Kineo Rhyolite; Dl—Littleton Fm.; Dm—Matagamon Ss.; Dmc—McKenney Ponds Ls. Mbr.; Dmi—Millimagassett Lake Fm.; Dmp—Mapleton Fm.; Dnd—Nadeau Thoroughfare beds (informal); Dpe—Perry Fm.; Ds—Seboomook Gr. (formerly Fm.); Dsl—Square Lake Ls.; Dsw—Swanback Fm.; Dta—Tarratine Fm.; Dte—Temiscouata Fm.; Dtf—The Forks Fm.; Dto—Tomhegan Fm.; Dtr—Traveler Rhyolite; Dtu—Touladi Ls.; Dtv—Trout Valley Fm.; Du—Unnamed Devonian strata; Dw—Wapske Fm.; DSac—Ayres Cliff Fm.; DScm—Costigan Mtn. Fm.; DSfl—Fish River Lake Fm. lower part; DShl—Hersey and Leighton Fms.; DSm—Madrid Fm.; DSpb—Parker Bog Fm.; DSsp—Spider Lake Fm.; DSwb—West Branch Volcanics; Sb—Burtts Corners beds; Scc—Cross Creek beds; Scl—Canterbury Limestone; Scq—Clough Quartzite; Sd—Dennys Fm.; Se—Edmunds Fm.; Sf—Fitch Fm.; Sfr—Flume Ridge Fm.; Sgls—Grand Lake Seboeis Fm.; Sh—Hardwood Mountain Fm.; Shb—Hayes Brook beds; Sl—Leighton Fm.; SOc—Cabano Fm.; SOm—Matapedia Gr.; Spe—Perham Fm.; Spm—Perry Mtn. Fm.; Spw—Pocowogamis Conglomerate; Sq—Quoddy Fm.; Sr—Rangeley Fm.; Srp—Ripogenus Fm.; Ssa—Sangerville Fm.; Ssf—Smalls Falls Fm.; Ssp—Spragueville Fm.; Sss—Scott Siding Slate; Stx—Taxis River Beds; Su—Unnamed Silurian strata. References for columns: 1—Boucot and others (1986); 3, 4—Hueber and others (1990); 5, 7—Uyeno and Lesperance (1997); 9, 10—Boucot and Arndt (1960); 11—Albee and Boudette (1972) and this study; 12—Boucot and others (1959) and this study; 13, 14—Boucot and Heath (1969) and this study; 15—Boucot and Heath, 1969; x16, 17—Griscom (1976), Rankin (1968), and this study; 18—Hall (1970); 20—Hibbard (1993); 21, 22—Boone (1970) and this study; 23—this study; 24—St. Peter and Boucot (1981); 26—Moench and Pankiwksyj (1988); 27—Marvinney (1984) and this study; 28, 29—Pankiwskyj and others (1976), Ludman 1978 and this study; 31—Boucot and others (1964) and this study; 33—St. Peter (1982); 34—Venugopal (1979); 35—Fyffe (1995); 37—West and others (1992); 38—Gates and Moench (1981).

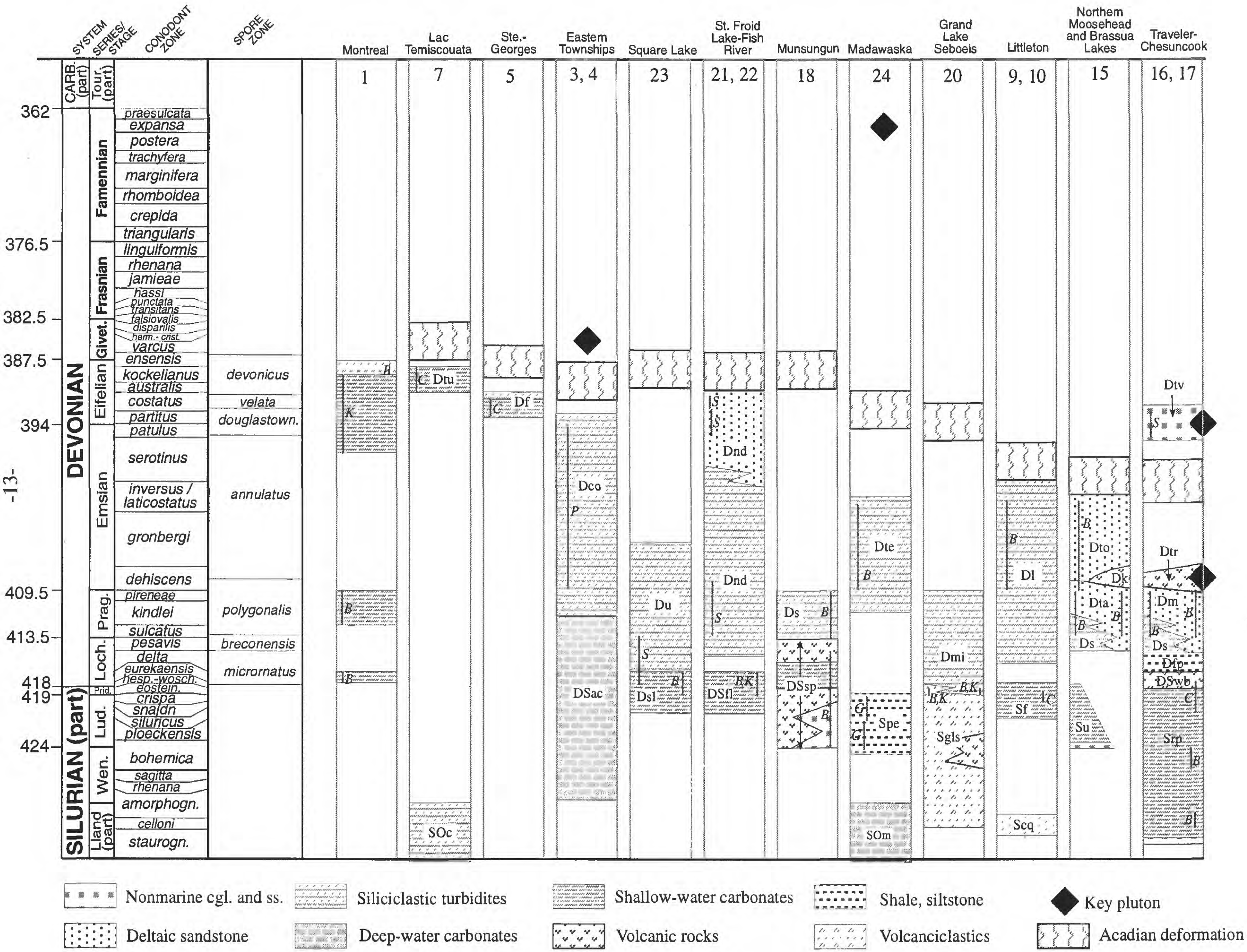
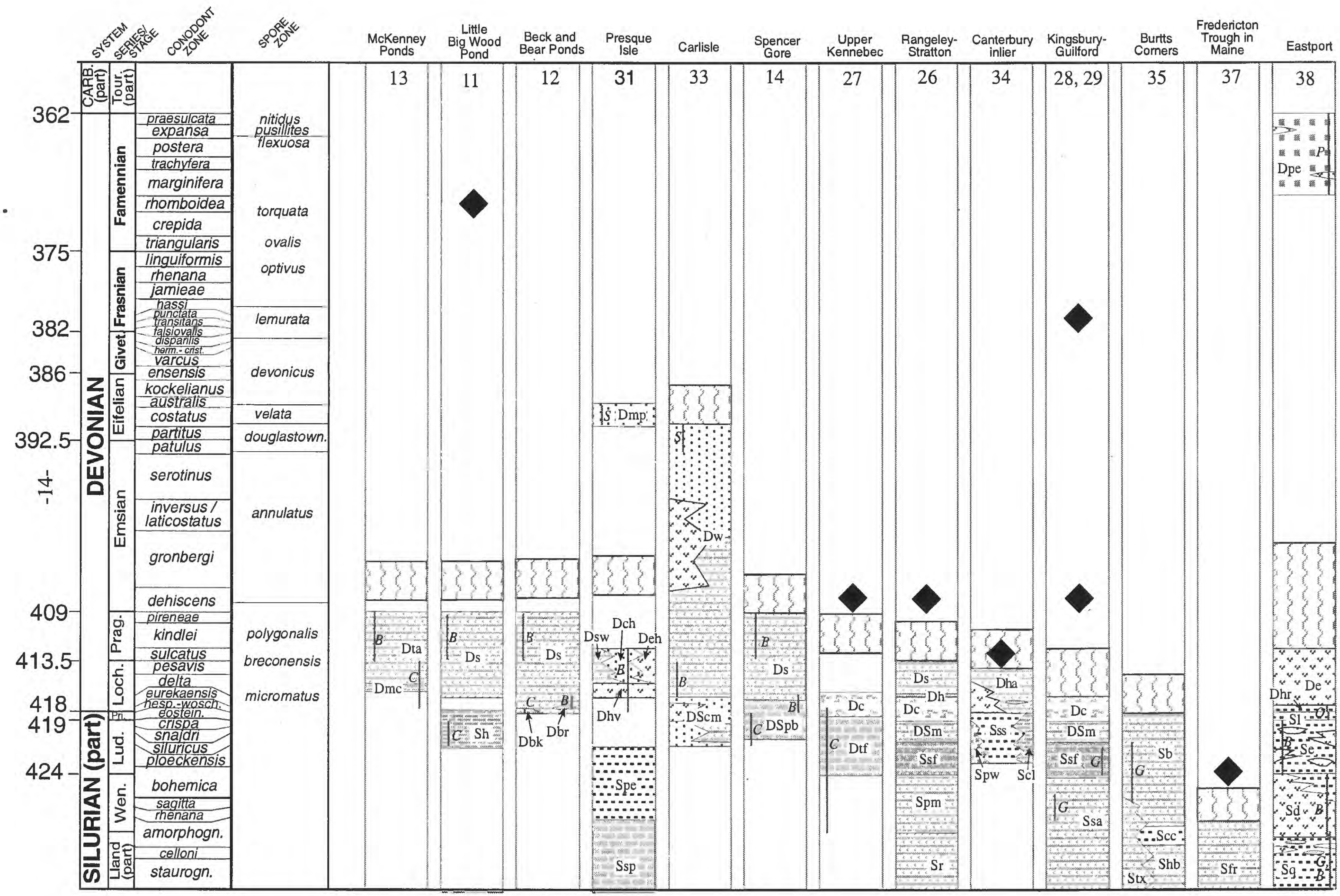


Figure 4, sheet 1 of 2



Bars in columns show age constraints by fossil groups:

B, brachiopods	C, conodonts
G, graptolites	K, corals
S, palynomorphs	P, pelecypods

Figure 4, sheet 2 of 2

metamorphosed by the Harkshaw and Skiff Lake phases of the Pokiok batholith (Venugopal, 1979, p. 28), which have yielded U/Pb ages of 411±2 (sphene) and 409±2 Ma (zircon), respectively (Bevier and Whalen, 1990a). Along strike in easternmost Maine, the Skiff Lake phase truncates Acadian structures that deform Silurian-Devonian strata (Ludman, 1990), and hence appears to be broadly post-tectonic. Acadian deformation in this area must therefore have occurred in the latter part of the Lochkovian or in the Pragian. Given that Acadian deformation must have followed closely after deposition of the Hartin Formation, a foreland-basin depositional setting seems likely.

CARLISLE AREA, NEW BRUNSWICK

Acadian deformation is anomalously young in the Aroostook-Matapedia basin in western New Brunswick (figs. 1 and 4, loc. 33). The Late Ordovician and Silurian were times of deep-water sedimentation (Carys Mills and Smyrna Mills Formations; St. Peter, 1982). The Smyrna Mills Formation is overlain by the Costigan Mountain Formation of predominantly felsic volcanic rocks; it is undated but regarded as probably Lower Devonian based on correlations with fossiliferous rocks to the northeast (St. Peter, 1982, p. 35). The youngest pre-Acadian unit is the Wapske Formation, estimated to be 4 km thick. It consists of slate, siltstone, sandstone, and conglomerate, plus isolated mafic volcanic lenses (St. Peter, 1982). Clasts in Wapske conglomerates can be traced to the Smyrna Mills, Costigan Mountain, and other older units (St. Peter, 1982, p. 40). The lower part of the Wapske Formation has yielded brachiopods from five localities, which according to Boucot (in St. Peter, 1982, p. 45) indicate a late Helderbergian, Becraft-Port Ewen age. This interval corresponds approximately to that of the *delta* and *pesavis* conodont zones of the Lochkovian. Spores of late Emsian or early Eifelian age have been recovered from the upper, nonmarine part of the formation (McGregor, in St. Peter, 1982, p. 478). We suggest that the Wapske Formation was deposited in a foreland-basin setting, and that it was derived, at least in part, from an Acadian orogenic source to the south, which must have existed by about 411 Ma.

PRESQUE ISLE QUADRANGLE, MAINE

The Presque Isle-Mapleton section section (figs. 1 and 4, loc. 31) provides age constraints on the timing of Acadian foreland-basin sedimentation, the main Acadian folding, and post-orogenic hinterland sedimentation. Deep-water sedimentation (slate, deep-water limestone, siliciclastic turbidites) was already underway by the Caradoc, and continued into the Ludlow (Carys Mills and Spragueville Formations and Perham Group; Roy and Mencher, 1976). A gap in the megafossil record, spanning the late Ludlow, Pridoli, and early Lochkovian, corresponds to the "Salinic Disturbance" of Boucot and others (1964). Whatever its cause, this hiatus spanned no more than six million years, judging from the new time scale (Tucker and others, 1998). The next rocks to be deposited were volcanics (Dockendorff Group), deltaic sandstones (Chapman Sandstone), and prodeltaic turbidites (Swanback Formation). Boucot and others (1964) showed these three units to be lateral equivalents of Lochkovian ("New Scotland") age. Roy (1980) interpreted the Swanback and Chapman Formations as easterly-derived flysch and molasse, respectively, which we assign to the Acadian foreland basin. Regional Acadian folding took place some time after deposition of the Lochkovian strata. The folded rocks are unconformably overlain by the Mapleton Formation, a local accumulation of nonmarine conglomerate and sandstone (Boucot and others, 1964). Conglomerate clasts in the Mapleton include fossiliferous clasts from the Chapman Sandstone and various older units (Boucot and others, 1964, p. 61-62; Roy and White, 1987). On the basis of plant fossils, the Mapleton was previously assigned a Middle Devonian, perhaps Givetian age (Kasper and others, 1988). We now assign the Mapleton a slightly older age, early to middle Eifelian (*Grandispora velatus-Rhabdosporites langii* Zone), on the basis of spores from material submitted by Eli Mencher to McGregor in the early 1960's (McGregor, 1962; new slides examined 1989; table 1). The following forms were identified: *Apiculiretusispora gaspiensis*, *Ancyrospora* sp., *Calamospora* cf. *C. atava* (Naumova) McGregor, *Calyptosporites* sp., *Emphanisporites annulatus*, *E. rotatus*, *Grandispora velatus*, *Stenozonotriletes* sp., *Dibolisporites echinaceus*, and *Acinosporites macrospinosus*?. The Mapleton is folded into a gentle syncline of middle Eifelian or younger age, which is nearly coaxial with an older, tighter syncline of the main phase of Acadian deformation (Boucot and others, 1964, p. 73).

CENTRAL MAINE BASIN

The stratigraphy of the Central Maine basin is best known from its northwestern flank in the Rangeley and Phillips quadrangles (figs. 1 and 4, loc. 26), where the Silurian-Devonian deep-water section is about 8 km thick, and most of the formations have their type localities (Moench and Pankiwskyj, 1988). Key sedimentary facies interpretations and paleocurrent data for these units, however, are from 150 km to the east in the Jo Mary Mountain area (Hanson and Bradley, 1989) (fig. 1, loc. 30). Complicating matters still further, the fossil control that is critical for present purposes is from yet a third area, Kingsbury-Guilford (figs. 1 and 4, locs. 28 and 29), which prior to Acadian shortening lay many tens of kilometers from the basin's northwest margin. The following discussion draws on information from all of these places.

Along the northwestern flank of the basin, the Silurian-Devonian section is divisible into lower and upper parts based on provenance and paleocurrents (Hanson and Bradley, 1993). Rocks assigned to the lower sequence (Greenvale Cove, Rangeley, Perry Mountain, and Smalls Falls Formations; fig. 4), are known or inferred to have been derived from the northwest. A northwesterly provenance is well established for the Llandovery-age Rangeley Formation at loc. 26 (figs. 1 and 4), which contains conglomerate clasts from the Attean pluton some 50 km to the northwest, near loc. 12 (fig. 1) (Moench and Pankiwskyj, 1988). Meager paleocurrent data suggest that the Smalls Falls is the youngest unit of northwest derivation (Hanson and Bradley, 1993). It consists of rusty-weathering turbiditic sandstones; the type Smalls Falls Formation is assigned a Ludlow age on the basis of fossils in broadly similar strata from a more southeasterly strike belt, as discussed below. The upper sequence was derived from outboard sources. It consists of two widespread units — the Madrid and Carrabassett Formations — plus poorly understood younger strata of local distribution. The Madrid Formation, in central Maine, is a thick succession of sandstone-dominated, siliciclastic turbidites; it becomes increasingly calcareous to the southwest in western Maine and New Hampshire. Paleocurrents show very clear southwesterly direction of flow, along the basin axis (Bradley and Hanson, 1989). The gradationally overlying Carrabassett Formation consists of another kilometer or so of chaotic, pelitic olistostrome plus subordinate, coherent sand- and silt turbidites. Paleocurrents show overall northerly flow (Hanson and Bradley, 1993). Hanson and Bradley (1989) interpreted the Madrid and Carrabassett Formations as having been deposited along a foreland-basin axis and on a north-facing slope, respectively. Unfortunately, neither the Madrid nor the Carrabassett have yielded useful fossils. The Madrid Formation is early Ludlow or younger. The Carrabassett Formation has yielded non-diagnostic brachiopods and plant fragments consistent with an Early Devonian age (Espenshade and Boudette, 1967, p. F12-F13), and it is cut by the earliest Emsian Sebec Lake pluton.

The depositional history of the central part of the Central Maine basin is not as well understood as the northwestern flank, because detailed sedimentological studies are lacking, and because deformation and metamorphism are more intense. A few observations bear on the timing of Acadian deformation. Most of the area is underlain by the Sangerville Formation, an isoclinally folded succession of siliciclastic and calcareous turbidites many kilometers thick. The Sangerville has yielded graptolites from 13 localities (Pankiwskyj and others, 1976); the most diagnostic collection (figs. 1 and 4, loc. 28), is middle Wenlock. The overlying Smalls Falls Formation has also yielded graptolites from a number of localities; the most diagnostic collection (figs. 1 and 4, loc. 29) is early Ludlow (Pankiwskyj and others, 1976). The overlying Madrid and Carrabassett Formations are the youngest units along the central part of the basin. The Carrabassett Formation is absent in the most southerly strike-belt of Madrid Formation, either because it has since been eroded away (Moench and Pankiwskyj, 1988), or was never deposited (Hanson and Bradley, 1989).

LITTLETON AND VICINITY, NEW HAMPSHIRE

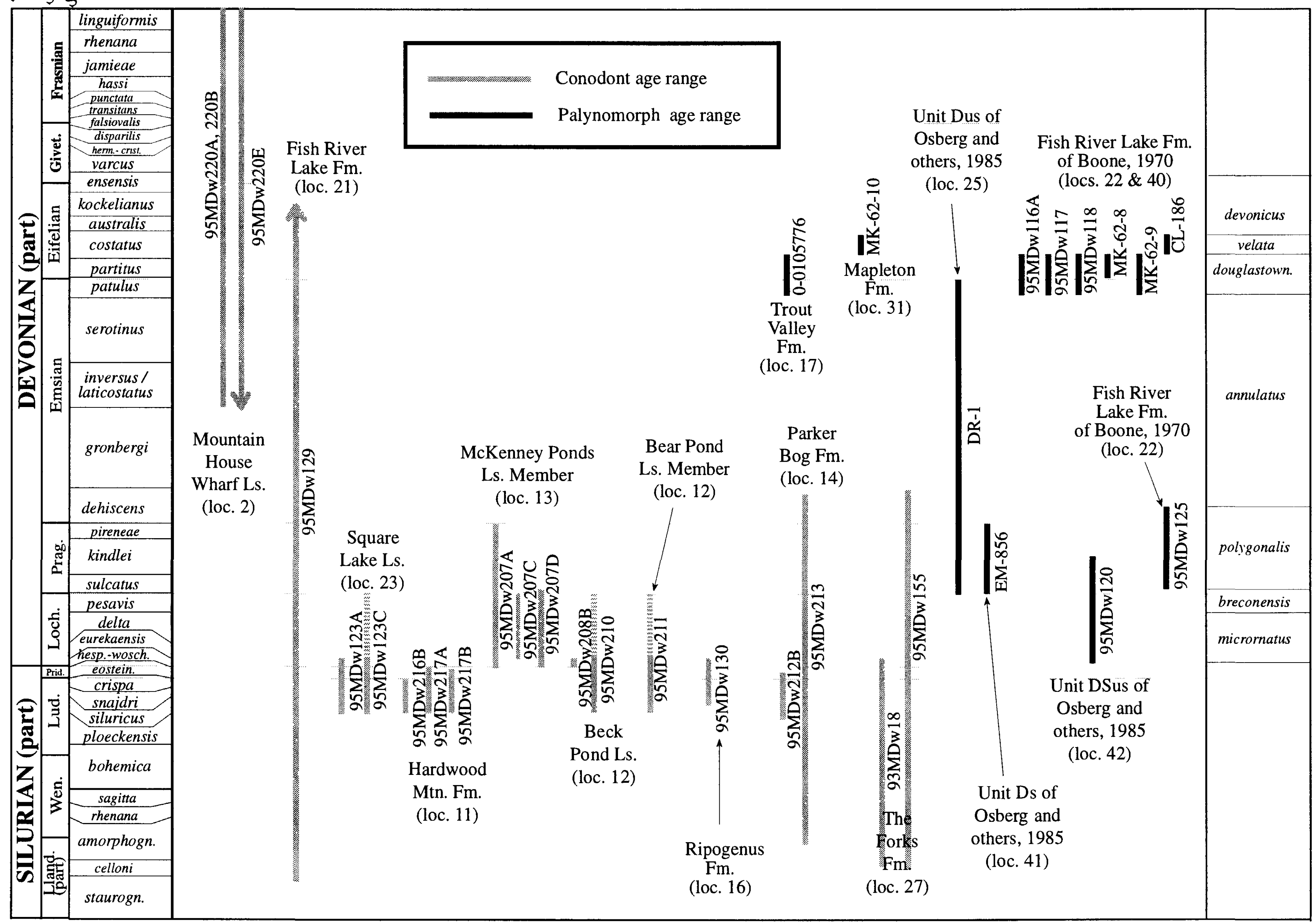
The classic Silurian-Devonian section near Littleton, New Hampshire (figs. 1 and 4, loc.9) (Billings, 1937) constrains the timing of both Acadian foreland-basin sedimentation and deformation. The Lower Silurian Clough Quartzite unconformably overlies Ordovician rocks and is followed by the Fitch Formation, which consists of calcareous metasiltstone, quartzite, granofels, and some limestone. Harris and others (1983, p. 731) reported an age of late Ludlow to mid-Pridoli for the Fitch Formation in the Littelton area, based on a conodont fauna representing the *Ozarkodina remscheidensis eosteinhornensis* Zone; along strike in Massachusetts, strata assigned to the Fitch Formation are as young as the *woschmidti* to *eurekaensis* Zones of the Lochkovian (Elbert and others, 1988). Both the Clough and Fitch are shallow-marine deposits; the Fitch has been interpreted as a far-foreland deposit (Bradley, 1983). The overlying Littleton Formation is a ~1.6-km-thick flysch succession (minimum thickness, as the top is not seen) deposited in a foreland-basin setting (Bradley, 1983). It has yielded Emsian brachiopods at two localities in the Littleton area (figs. 1 and 4, loc. 9), and at a third locality in Whitefield (figs. 1 and 4, loc. 10), where the fossils occur only a few tens of meters above the base (Boucot & Arndt, 1960, p. 41-43). At Gale River (figs. 1, loc. 39) in the next strike belt to the east, a tuff from low in the Littleton Formation has yielded an Emsian U-Pb zircon age of 407±2 Ma (R.D. Tucker and D. Rankin, unpublished data). Still further east at Beaver Brook (fig. 1, loc. 8), a brachiopod fauna suggests a slightly older age ("Oriskany, possibly Esopus"; Boucot and Rumble, 1980, p. 192-194) than at locs. 9 and 10. The Littleton thus appears to be diachronous across strike at the latitude of its type area.

SOUTHWESTERN MOOSE RIVER SYNCLINORIUM, MAINE

In the southwestern part of the Moose River synclinorium, western interior Maine, local carbonate-bearing units overlie Ordovician and older rocks, providing good age constraints on the onset of Acadian foreland-basin sedimentation. At Beck Pond (figs. 1, 4, and 5, loc. 12), the post-Ordovician section begins with the Beck Pond Limestone (Boucot and others, 1959). It includes reef limestone intercalated with boulder conglomerate derived from the nearby Attean pluton, of Ordovician age. A rich brachiopod fauna initially suggested an "upper Helderberg, New Scotland" age (Boucot and Heath, 1969). However, conodonts from the upper part of the Beck Pond (member 5 of Boucot and others, 1959) include *Belodella devonica*, *Decoriconus* sp., *Dvorakia* sp., coniform elements of *Icriodus* sp. indet., and *Ozarkodina remscheidensis remscheidensis*, indicating, instead, an early Lochkovian age (table 2). The Beck Pond Limestone is overlain by a considerable thickness, probably 2-3 km, of Lower Devonian flysch assigned to the Seboomook Group (formerly Formation; Pollock, 1987). Boucot and others (1959) interpreted the contact between Member 5 of the the Beck Pond and overlying slate of the Seboomook as an angular unconformity. Upon digging out this contact (at loc. 10 of Boucot and others, 1959), we observed that reefal limestone debris interfingers with black silty shale. The conodonts thus date the top of the Beck Pond Limestone, the base of the Seboomook flysch sequence, and the onset of Acadian foreland-basin sedimentation.

About 1 km to the north (figs. 1, 4, and 5, loc. 12), the Bear Pond Limestone (Boucot and others, 1959) is a carbonate lens or buildup not far above the base of the Seboomook Group. It is exposed at but a single outcrop; the known stratigraphic thickness is only about 10 meters. A rich brachiopod fauna again indicates an "upper Helderberg, New Scotland" age (Boucot and Heath, 1969) for the buildup and, by implication, for the immediately overlying and underlying Seboomook siltstones and sandstones. A large conodont sample yielded a meager conodont fauna of *Belodella devonica*, *Ozarkodina remscheidensis eosteinhornensis*, and fragments of *Icriodus* sp. Indet. or *Pedavis* sp. indet., allowing a possible age range from late Ludlow to early Pragian (table 2). Nearby, Boucot and Heath (1969, p. 36) reported several brachiopod occurrences that suggest a Pragian age for all but the lowest beds of the Seboomook Group in the Moose River synclinorium.

Figure 5. Conodont and palynological age ranges of samples dated in this study, and highlights from reports by McGregor (1962, 1963, 1968, 1992).



The Seboomook Group is the youngest pre-Acadian stratified unit in the area; it not not known to extend upward beyond the Pragian.

At loc. 13 (figs. 1, 4, and 5), the post-Ordovician section begins with the McKenney Ponds Limestone Member of the Tarratine Formation. Boucot and Heath (1969) reported a "Becraft-Oriskany" (late Lochkovian to Pragian) age based on brachiopods. This assignment can be somewhat refined based on our conodont collections (table 2). The most diagnostic of three samples, located 2 m stratigraphically below the lowest siliciclastic beds in the Tarratine Formation, yielded *Belodella devonica*, *Icriodus* sp. indet., *Ozarkodina remscheidensis remscheidensis*, and *Pseudooneotodus beckmanni*. A Lochkovian, but not earliest Lochkovian age is indicated by the icriodids, which are of a post-*I. woschmidtii* morphotype. The limestone member is gradationally overlain by sandstone of the main body of the Tarratine Formation, which both interfingers with and overlies the Seboomook Group, and likewise is interpreted as part of the foreland-basin fill. The Tarratine Formation has yielded Oriskany-age (Pragian) brachiopods from several nearby locations (Boucot and Heath, 1969).

NORTHEASTERN MOOSE RIVER SYNCLINORIUM, MAINE

This area provides crucial stratigraphic constraints on the age of the youngest foreland-basin deposits. In the area of northern Moosehead and Brassua Lakes (figs. 1 and 4, loc. 15), Ordovician and older basement rocks are overlain by poorly dated, Silurian calcareous rocks and Devonian(?) red shale (Boucot and Heath, 1969). The Acadian foreland-basin succession begins with the Tarratine Formation, which is about 2 km thick in the area (Boucot and Heath, 1969, plate 19). The Tarratine Formation has yielded abundant brachiopods of Oriskany age (Boucot and Heath, 1969, p. 27), that is Pragian. The Kineo Rhyolite, sandwiched between the Tarratine Formation, below, and the Tomhegan Formation, above, represents a pulse of silicic magmatism in the foreland basin. It has yielded only discordant zircon ages, marked by inheritance. It is comparable in stratigraphic position, composition, and eruptive environment to the 406-407 Ma Traveler Rhyolite (Rankin and Tucker, 1995), 65 km to the northeast (figs. 1 and 4, loc. 17); the Kineo and Traveler are presumably coeval. The Kineo varies in thickness from a feather edge at its northern limit to about 1200 m just a few kilometers to the south. The Kineo is overlain by the main body of the Tomhegan Formation, consisting of cross-bedded, probably deltaic sandstones (L. Hanson and D. Bradley, unpublished field observations). Boucot and Heath (1969, p. 17) estimated the thickness at about 1800 m — a minimum, as the top is not preserved. Brachiopods in the Tomhegan indicate a "Schoharie" age (late Emsian) (Boucot and Heath, 1969, p. 20). The Tomhegan is the youngest pre-Acadian unit in this part of Maine, and it shows that the Acadian deformation here was no earlier than the late Emsian.

TRAVELER-CHESUNCOOK AREA, MAINE

Along the west side of the Katahdin batholith (figs. 1, 4, and 5, loc. 16), Silurian sedimentation began with deposition of shallow marine conglomerate, limestone, and siltstone assigned by Griscom (1976) to the Ripogenus Formation. Various localities within the Ripogenus Formation have yielded brachiopods of Llandovery C3-C5, late Wenlock, and possible Ludlow ages (Boucot and Heath, 1969, p. 53). A conodont sample (95MDW130) from Ripogenus Dam (table 2; figs. 1 and 4, loc. 16) yielded *Decoriconus* sp. indet., *Dvorakia* sp. indet., *Oulodus* sp. indet., *Ozarkodina excavata*, and *Panderotus* sp. indet., permitting an age range of late Ludlow to early Lochkovian. The Ripogenus is gradationally overlain by siltstone and andesite (West Branch Formation), followed by red shale (Frost Pond Shale); these units are barren of fossils but are probably Lochkovian. The overlying Seboomook Group begins the Acadian-derived clastic succession. It consists of thin-bedded turbidites that have yielded Oriskany-age (Pragian) brachiopods from a locality at Chesuncook Lake (R.B. Neumann, in Griscom, 1976, p. 147-148). The Seboomook is overlain progradationally by the Matamagon Sandstone, which has also yielded Pragian brachiopods (see, for example, R.B. Neumann, in Griscom, 1976, p. 151-152). North of the Katahdin pluton (figs. 1 and 4, loc. 17), the Seboomook and Matagamon together comprise a prodelta and delta deposited by west-directed paleocurrents (Hall and others, 1976; Pollock and others, 1988). The Matagamon Sandstone is overlain conformably by the Traveler Rhyolite. There appears to have been no hiatus between the two because clastic dikes of Matagamon Sandstone intrude upward into the basal rhyolite (Rankin, 1968). Concordant U/Pb zircon ages of 406 Ma from the Black Cat Member and 407 MA from the Pogy Member (Rankin and Tucker, 1995) indicate an early Emsian age for the Traveler Rhyolite and most likely, as well, for the uppermost beds of the Matagamon Sandstone. The Silurian through upper Emsian strata were involved in the regional Acadian folding, which produced a regional northeast-striking cleavage.

The Traveler Rhyolite is overlain by essentially undeformed nonmarine sandstone and conglomerate of the Trout Valley Formation. Although an angular unconformity cannot be demonstrated at outcrop, 1:62,500-scale mapping shows that the Trout Valley Formation truncates major structures in the Traveler (Rankin, 1968). Kasper and others (1988) reviewed the history of age assignments for the Trout Valley based on plant fossils. Palynomorphs from the Trout Valley (McGregor, 1992) indicate a latest Emsian to earliest Eifelian age, corresponding to the *douglastownense-eurypterota* spore assemblage zone, or the *patulus* conodont zone (table 1; fig. 5). The Trout Valley is therefore significantly younger than the Traveler Rhyolite.

MUNSUNGUN ANTICLINORIUM, MAINE

The Munsungun anticlinorium is cored by Ordovician and older rocks, which are unconformably overlain by a complex assortment of locally-derived clastic, volcanic, and carbonate rocks (figs. 1 and 4, loc. 18). Hall (1970) assigned these to the East Branch Group and to the laterally equivalent Spider Lake Formation. Corals and brachiopods from a number of fossil occurrences give ages of Ludlow, Pridoli, and "middle or upper Helderberg" (Hall, 1970, p. 35). Next follows the Seboomook Group, which is the youngest rock unit in the area, having yielded brachiopods of Oriskany (Pragian) age from three places (figs. 1 and 4, loc. 18) (Hall, 1970, p. 41). Acadian deformation must be Emsian or younger along the Munsungun anticlinorium.

A similar succession has been reported from the Millimagassett Lake area, at the southern end of the Munsungun anticlinorium (figs. 1 and 4, loc. 20). The Grand Lake Seboeis Group consists of slate, sandstone, conglomerate, calcareous slate, and volcanic rocks. Hibbard (1994, p. 216) cited nine fossil collections that give a range of "Late Silurian to earliest Devonian" (no details were published). The Grand Lake Seboeis Group grades up into, but may interfinger with, the Millimagassett Lake Formation, which has yielded Late Silurian, most likely Pridoli brachiopods and corals near its base (Hibbard, 1993); presumably its upper part extends at least into the Lochkovian, if not the Pragian. According to Hibbard (1994), the Millimagassett Lake Formation is a flysch succession resembling the Seboomook Group. It appears to be the oldest such flysch yet identified along the anticlinorial belt, possibly 2-3 million years older than the Seboomook Group at Beck Pond (figs. 1 and 4, , loc. 12), and 5-10 million years older than along the Munsungun anticlinorium (figs. 1 and 4, loc. 18). In the Millimagassett Lake area, Hibbard (1994) reported evidence for an early Acadian sinistral shear deformation that was followed by the main, regional Acadian folding on northeast-strike axes. Both phases of deformation are Lochkovian or younger.

PENNINGTON ANTICLINORIUM, MAINE

To determine the timing of Acadian foreland-basin sedimentation and deformation along the northwestern limb of the Pennington anticlinorium (figs. 1, 4, and 5, locs. 21, 22, and 23), it is first necessary to unravel some stratigraphic problems. The Silurian to Devonian Fish River Lake Formation of Boone (1970) consists of sandstone, siltstone, conglomerate, mafic and silicic volcanic rocks, and impure limestones. Unnamed unit DSus, as shown on the bedrock geologic map of Maine (Osberg and others, 1985), occurs along strike from the Fish River Lake Formation, at the same broad stratigraphic level. New information reported here suggests that both the Fish River

Lake and unit DSus are made up of two unrelated successions: an older one that includes volcanic and calcareous rocks, and a younger one of plant-bearing siltstone, sandstone, and conglomerate.

Carbonate rocks from the Fish River Lake Formation have yielded brachiopods of Late Silurian ("Cobleskill", Pridoli) and Early Devonian ("New Scotland", Lochkovian) age (A.J. Boucot, in Boone, 1970, p. 33-34). We processed a shaly limestone from one of Boucot's Cobleskill-age locations (figs. 1 and 4, loc. 21) for conodonts, but the sample yielded only nondiagnostic Silurian-Devonian forms (table 2). Within unit DSus, the informally named Square Lake Limestone occurs in a single lakeshore exposure (figs. 1, 4, and 5, loc. 23). Boucot (in Boone, 1970) placed the brachiopod and coral fauna "definitely within the lower Helderberg series" (that is, *woschmidti* to *eurekaensis* Zones of the Lochkovian). Conodonts from the most diagnostic of two samples of reef limestone give an age range of late Ludlow to early Lochkovian (table 2, figs. 1 and 4, loc. 23).

Plant-bearing sandstones and siltstones from Boone's Fish River Lake Formation and unit DSus have yielded spores, some of which are considerably younger than the brachiopods (McGregor, 1962, 1963, 1968, and 1996) (table 1, fig. 5). This conclusion is based on material submitted without lithologic descriptions to McGregor by Eli Mencher in the 1960s, and by Bradley in 1995. In unit DSus, the most closely dated of two siltstone samples is undoubtedly Early Devonian, probably Lochkovian or early Pragian in age (table 1). Bedding and a strong cleavage (presumably Acadian) in these rocks both are subvertical. In the Fish River Lake Formation, two samples of dark, thin-bedded siltstone are of late Pragian-early Emsian age (*polygonalis-emsiensis* Zone). These rocks, which resemble the Seboomook Group, are also folded and cleaved. Three sandstone samples from 1995 and two samples of unknown lithology from the 1960's are of late Emsian to early Eifelian age (*douglastownense-eurypterota* Zone). The 1995 samples of this age are from a succession of fluvial or marginal-marine plant-bearing sandstone and conglomerate at Nadeau Thoroughfare (figs. 1, 4, and 5, loc. 22), which dip moderately (50°) but lack cleavage. One of the 1995 samples yielded acritarchs, hence is of marine origin. Finally, one sample of unknown lithology submitted to McGregor by Mencher in the 1960s is early to middle Eifelian (*velatus-langii* Zone).

Piecing together these various facts, we suggest that far-foreland carbonate sedimentation was underway during the late Silurian and at least part of the Lochkovian. An influx of orogenically-derived siliciclastics began in late Lochkovian or early Pragian time. Deposition of the plant-bearing sandstones probably lasted from about late Emsian into the early or middle Eifelian. Taken together, the palyniferous strata can be interpreted as components of a prograding foreland-basin sequence, akin to the somewhat older Seboomook-Matagamon succession farther south.

MADAWASKA REGION, NEW BRUNSWICK

Two useful constraints on Acadian timing are available from northwesternmost New Brunswick, where the Aroostook-Matapedia and Connecticut Valley-Gaspé basins come together (figs. 1 and 4, loc. 24). The Upper Ordovician and Silurian are represented by various deep-water facies (St. Peter, 1977). The youngest Silurian rocks are mostly green slate and calcareous siltstone of the Perham Formation, which has yielded graptolites as young as middle to late Ludlow (St. Peter, 1977, p. 30). There follows a gap in the fossil record spanning the Pridoli, Lochkovian, and Pragian. The overlying Temiscouata Formation consists of several kilometers of siliciclastic turbidites characterized by abundant slump folds; Hesse and Dalton (1995) interpreted the Temiscouata as a submarine foreland-basin succession. It has yielded brachiopods now assigned an Emsian age (St. Peter and Boucot, 1981, p. 91). The Temiscouata Formation has been penetratively deformed; Acadian deformation must be Emsian or younger in age.

LAPATRIE AND STE. CECILE DE WHITTON, EASTERN TOWNSHIPS, QUÉBEC

In the Eastern Townships of Québec, the axial region of the Connecticut Valley-Gaspé basin is occupied by a thick succession of Acadian-deformed sandstone and slate assigned to the Compton Formation (Slivitski and St.-Julien, 1987). Sandstones of the Compton Formation have yielded Emsian plant fossils from locs. 3 and 4 (figs. 1 and 4) (Hueber and others, 1990). Although the age of onset of foreland-basin sedimentation in this area is unknown, the presence of Emsian foreland-basin deposits suggest that the orogen was not far off at the time. Acadian deformation in the Eastern Townships must have taken place between 409-394 (possible age range of the Compton Formation) and the 384±2 Ma Scotstown pluton, which is the oldest post-folding pluton that intrudes the Compton Formation (Simonetti and Doig, 1990).

NORTHWEST MARGIN OF THE CONNECTICUT VALLEY-GASPÉ BASIN, QUÉBEC

Three localities along the northwest margin of the Connecticut Valley-Gaspé basin in Québec provide important contstraints on the age of Acadian foreland-basin sedimentation and deformation. The most informative strata are at Saint-Georges (figs. 1 and 4, loc. 5), where the Eifelian Famine Formation unconformably overlies the Ordovician Magog Group (Uyeno and Lesperance, 1997). The Famine begins with a basal conglomerate, overlain by limestone, which is coral-bearing at the bottom but becomes progressively finer-grained and shalier upsection. The maximum preserved thickness is 215 meters. Uyeno and Lesperance (1997) have documented a conodont fauna of *costatus* age, a refinement on the previously quoted Eifelian age (Oliver, 1971), which was based on corals and brachiopods. The Famine Formation is the youngest unit at Saint-Georges; it was involved in regional Acadian deformation which thus must postdate the *costatus* Zone.

In the Lac Temiscouata area (figs. 1 and 4, loc. 7), the situation is similar to that at Saint-Georges. The key unit, comparable to the Famine, is the Touladi Limestone. It rests with slight angular unconformity on various older units ranging in age from possibly Ordovician to Late Silurian (Uyeno and Lesperance, 1997). The Touladi consists of mostly bioclastic limestone with a maximum preserved thickness of 105 meters. Recent conodont studies by Uyeno and Lesperance (1997) indicate a position in either the *australis* Zone or the *kockelianus* Zone of the Eifelian. This refines but does not significantly alter the Eifelian age quoted by Boucot and Drapeau (1968).

Devonian limestones at Lac Memphremagog (fig. 1, loc. 2) suggest a similar interpretation, although biostratigraphic control is not as tight. There, the Mountain House Wharf Limestone is a dark, slaty limestone surrounded by, and probably faulted against, Ordovician rocks (Boucot and Drapeau, 1968). It has yielded a brachiopod and coral fauna of Eifelian age (Boucot and Drapeau, 1968, p. 9). Three of four large conodont samples yielded only long-ranging forms of middle Emsian to Frasnian age (table 2, fig. 5).

The coral-bearing shallow-water limestones at Saint-Georges, Lac Temiscouata, and Lac Memphremagog were probably deposited in a far-foreland setting. The upward-deepening trend at Saint-Georges most likely records the beginnings of foreland-basin subsidence. Reckoning that an interval of foreland-basin sedimentation would have preceded deformation, the Acadian orogeny most likely took place at Saint-Georges during the Givetian, or conceivably during the latter part of the Eifelian. Defomation at *Lac* at Temiscouata probably took place a few million years later than at Saint-Georges.

MONTREAL, QUÉBEC

Devonian strata do not crop out anywhere north of the Mohawk River in central New York, but were once present in Montreal, 330 km to the north (figs. 1 and 4, loc. 1). The St. Helenes Breccia (Boucot and others, 1986) is a Cretaceous diatreme that was emplaced into a region of gentle foreland deformation about 20 km cratonward of the Paleozoic thrust front. The diatreme contains an assortment of xenoliths that reveal the former existence of a Devonian succession similar to that in eastern and central New York State. The diatreme, in the words of J.B. Thompson, is a "drill hole in the sky". The inferred sequence is Lochkovian (Helderbergian) limestone, Pragian (Oriskany-age) sandstone and limestone, Emsian (Bois Blanc- or Schoharie-age) limestone, Eifelian (early and late Onondagan) limestones, and Givetian (Hamilton-age) micritic limestones, siltstones, and sandstones. The latter siltstones and sandstones resemble the Hamilton, which in the Catskills of eastern New York State comprises the base of the main foreland-basin sequence. By implication, the leading edge of the foreland basin was near Montreal during Givetian time.

ISOTOPIC AGES OF ACADIAN PLUTONS

When the present study was first conceived, only a handful of high-precision U/Pb and $^{40}Ar/^{39}Ar$ ages were available for Silurian and Devonian plutons in the area of figure 1 — not enough to reveal a clear regional age pattern to the plutons. Accordingly, we set out to date all the significant Acadian plutons in a swath from the Maine coast to the Maine-Québec border. New U/Pb and (or) $^{40}Ar/^{39}Ar$ ages are reported here for 17 plutons. Other new ages from the U/Pb lab at Washington University were reported by Tucker and others (in press) for plutons near and on the Maine coast, by Rankin and Tucker (1995) for the Katahdin area, and by Solar and others (1998) for western interior Maine. Figure 6 and table 3 summarize the best available emplacement ages and clearly reveal plutonic belts around 420 Ma near the coast and 406-407 Ma further inland; these are essential in positioning the deformation front. Other plutons are significantly younger than ~406 Ma and some of these also occur in discrete belts, but they are not as useful in tracking the Acadian front.

Acadian plutons fall into three broad and somewhat overlapping categories: pre-, syn- and post-tectonic with respect to the *local* first phase (D1) of Acadian deformation. A pluton might be classified as post-tectonic, even though the Acadian orogeny was still underway elsewhere during the pluton's emplacement. We considered three types of anecdotal evidence in assessing the age of plutonism relative to Acadian deformation: (1) map-scale relations to Acadian folds and faults; (2) plutonic rock fabrics; and (3) metamorphic textures in the contact aureole.

U/Pb ANALYTICAL METHODS

U/Pb analyses (fig. 7 and table 4) were performed at the geochronology laboratory at Washington University, St. Louis. Zircon was extracted from 5-10 kg samples using standard techniques of density and magnetic separation. Zircon grains were selected for analysis on the basis of size, color, clarity, and morphology (table 4). Sample sizes varied, depending on U content, grain size, and age, but most analyses were performed on fractions of 5 to 20 crystals, generally with less than 200 ng of radiogenic Pb. All analyses were air-abraded (Krogh, 1982) and cleaned sequentially in warm 4N HNO_3, water, and distilled acetone to remove components carrying common Pb. Washed and weighed zircon fractions were then loaded in TFE Teflon bombs, spiked with a mixed ^{205}Pb-^{235}U tracer solution, and digested in 48% HF and 7N HNO_3 for 72 hours at 210°C. Following digestion and conversion to chloride form, Pb and U were purified using ion-exchange techniques described in Krogh (1973). Total-procedure blanks for Pb and U, measured during the period of analysis, ranged between 1-8 pg and <1 pg, respectively; total common-Pb abundances were reported for each analysis in table 4. Initial-Pb compositions result in insignificant changes to the calculated ages.

Isotope ratio measurements of Pb and U were made in a VG Sector-54 thermal ionization mass spectrometer with enhanced pumping capacity, seven moveable collectors, and a Daly-type detector with ion-counting capability. Lead and uranium were loaded together on outgassed single-Re filaments with silica gel and phosphoric acid, and all measurements were made by the method of peak-hopping in ion-counting mode. Ion-beam intensities ranged between 0.5 x 10-13A and 1.5

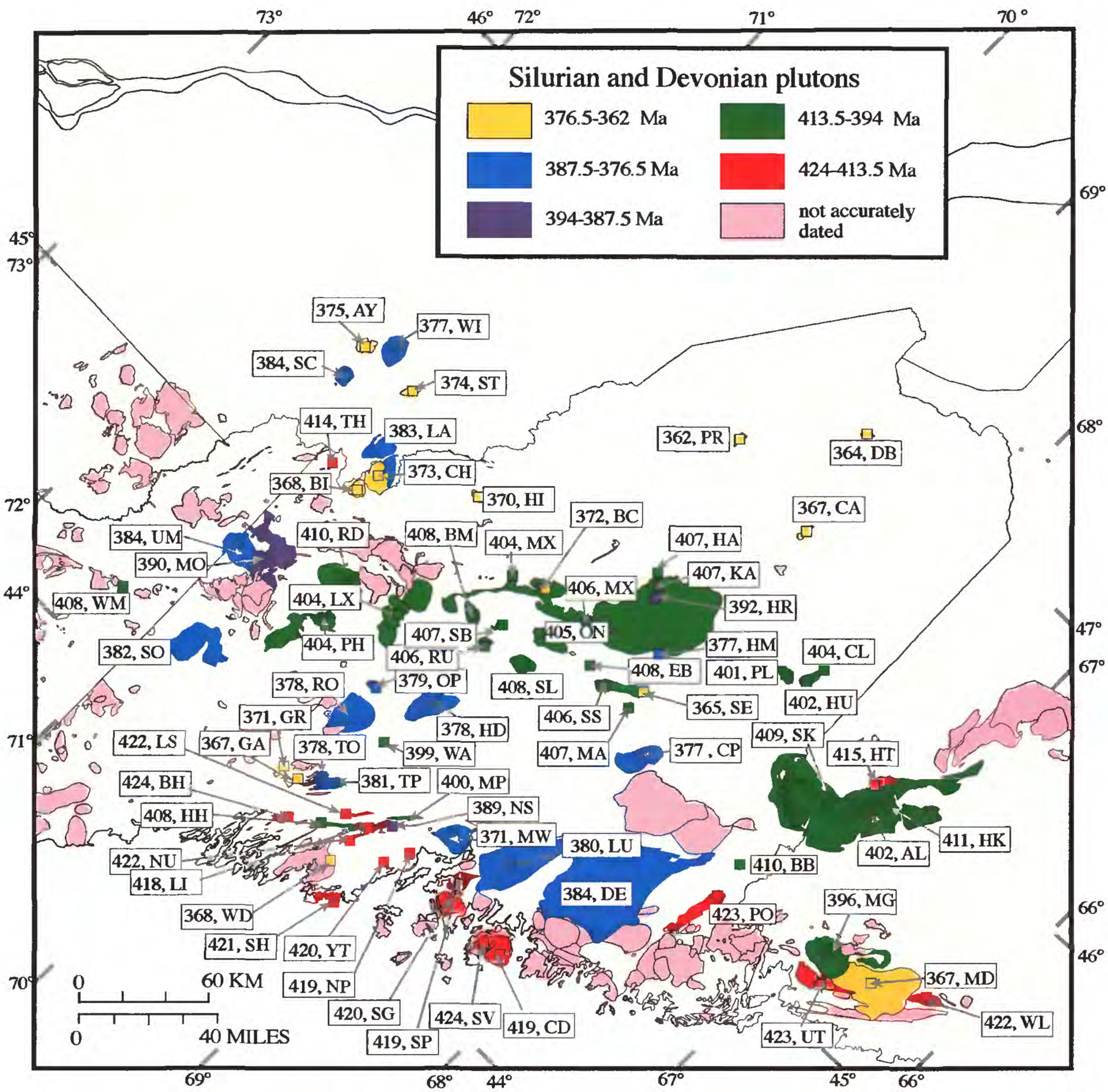
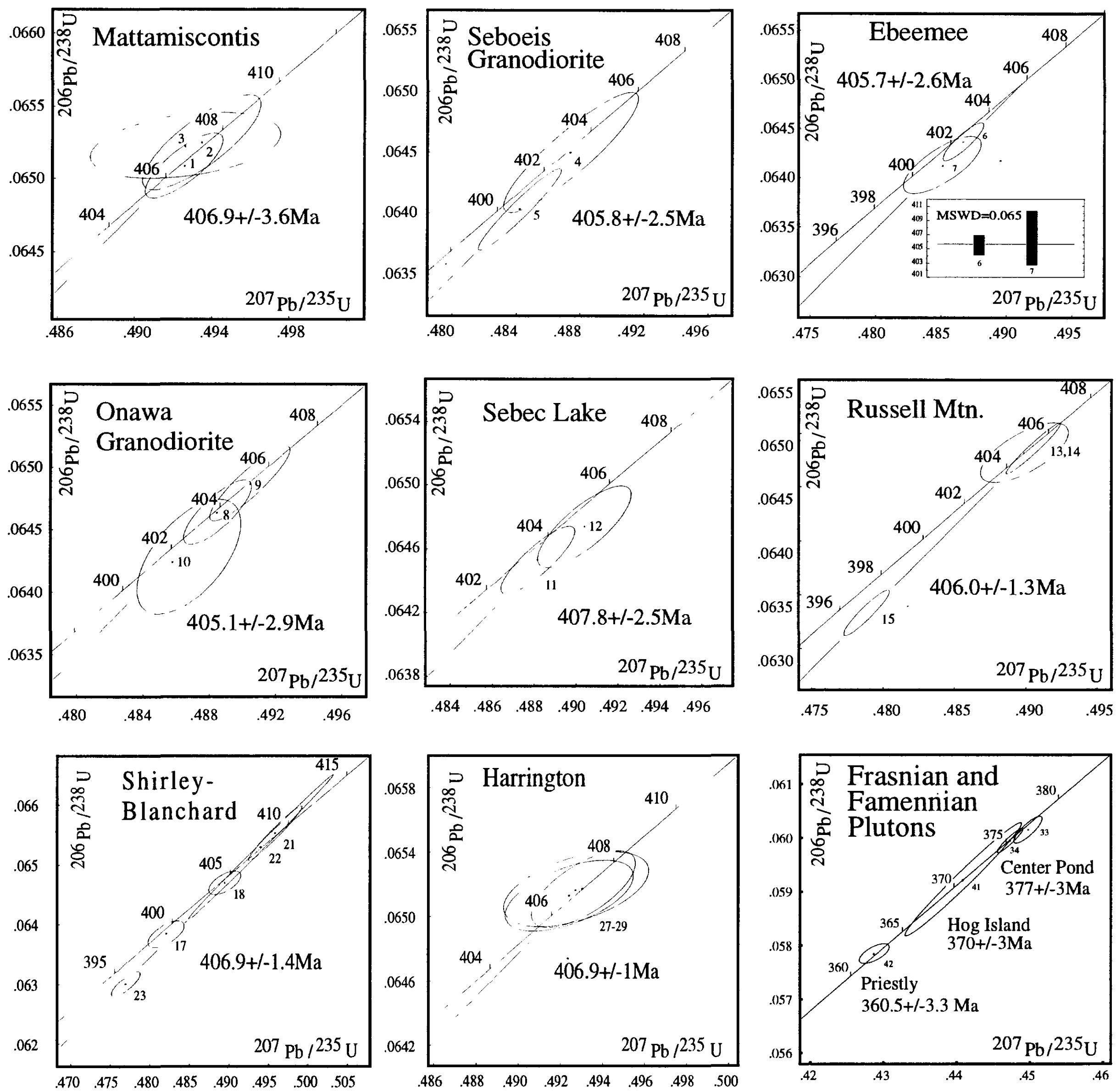


Figure 6. Map of Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing distribution of Late Silurian, Devonian, and earliest Carboniferous plutons. Boxes show best estimate of age of intrusion (concordant U/Pb or 40Ar/39Ar plateau) and two-letter abbreviation, keyed to table 3. Dates are not shown for plutons that have been dated only by less reliable techniques: Rb/Sr, conventional K/Ar, and older U/Pb.



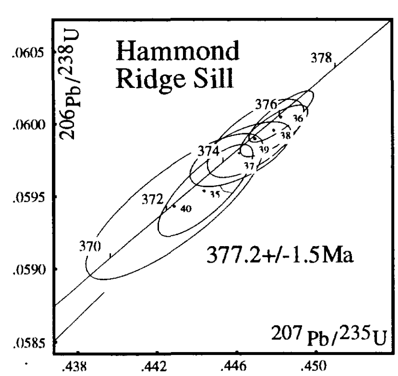


Figure 7. U/Pb concordia plots of igneous rocks dated in this study.

x 10-13A for $^{206}Pb^+$, and between 0.5 x 10-13A and 2.5 x 10-13A for ^{235}U (measured as UO_2^+). Daly bias and nonlinearity were periodically monitored with NIST and CBNM isotopic reference materials, and correction factors and errors for Daly gain were used in data reduction.

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); all age uncertainties are quoted at the 95% confidence level. Cited ages are the mean $^{207}Pb/^{206}Pb$ age 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 reliablity of all cited ages may be evaluated by the MSWD (mean square of the weighted deviates) which is a measure of the observed scatter to that predicted by the assigned errors to each analysis. In all cases, the MSWD is much less than one, indicating that assigned errors may be somewhat overestimated.

$^{40}Ar/^{39}Ar$ ANALYTICAL METHODS

$^{40}Ar/^{39}Ar$ analyses (fig. 8 and table 5) were performed at the geochronology laboratory at University of Maine, Orono; table 5 lists analytical data. Samples were separated using standard magnetic and density separation techniques. The purity of the samples was estimated to be greater than 99.5%. Samples, flux monitors (SBG-7 inter-lab standard), and K and Ca salts were encapsulated in Al foil and sealed in silica glass vials. These were irradiated in the L67 postion of the Ford Nuclear Reactor at the University of Michigan. Micas and flux monitors weighed approximately 35 mg. Samples were heated in a molybdenum crucible within the ultra-high vacuum system on line to the mass spectrometer using radio frequency induction. Temperature estimates have an estimated uncertainty of ± 50°C. Inert gases were purified using standard gettering techniques. The isotopic composition of Ar was measured digitally using a Nuclide 6-60-SGA 1.25 mass spectrometer. All data were corrected for mass discrimination and interfering argon isotopes produced during irradiation (Dalrymple and others, 1981). The decay constants recommended by Steiger and Jager (1977) were used to calculate the ages. Error calculations included both the uncertainty in the analytical measurement and the uncertainty in the J-value and are reported at the 2σ level. Each sample was analyzed in a minimum of eight increments and no more than 15 increments. A plateau age represents the mean of ages in consecutive increments that are not different based on 2σ analytical uncertainties.

LUDLOW, PRIDOLI, AND LOCHKOVIAN PLUTONS

Previous findings. Until a few years ago, the numerous gabbroic to granitic plutons along the Maine coast were all assigned to the Devonian, but in light of modern $^{40}Ar/^{39}Ar$ and U/Pb dating and the new time scale, the oldest group of plutons now is known to range from Ludlow to Lochkovian. The first pluton in Maine to yield an undoubted Silurian age was the Pocomoonshine gabbro-diorite (PO in fig. 6). It yielded a $^{40}Ar/^{39}Ar$ amphibole plateau age of 422.7±3 Ma (West and others, 1992). The pluton is post-tectonic: it intruded into already deformed turbidites of the presumably Lower Silurian Flume Ridge Formation, and "stitches" the fault contact between the Fredericton basin and the St. Croix belt to the south (West and others, 1992). New zircon ages from the Utopia (423.0±1.0 Ma; UT in fig. 6) and Welsford (422.0±1.0 Ma; WL in fig. 6) plutons (M.L. Bevier, written communication, 1996) extend the belt of known Ludlow-age plutons into southern New Brunswick. A number of plutons in the Penobscot Bay region have also yielded concordant zircon ages of 424-417 Ma, that is, Late Silurian to earliest Devonian. These include the Somesville (424.0±2.0 Ma; SV in fig. 6), Blinn Hill (424.0±2.0 Ma; BH in fig. 6), Lake St. George (422.0±2.0 Ma; LS in fig. 6), North Union (422.0±2.0 Ma; NU in fig. 6), Spruce Head (421.0±1.0 Ma; SH in fig. 6), Youngtown (420.0±2.0 Ma; YT in fig. 6), Sedgwick (419.5±1.0 Ma; SE in fig. 6), South Penobscot (419.2±2.2 Ma; SP in fig. 6), Cadillac (419.0±2.0 Ma; CD in fig. 6), Northport (419.0±2.0 Ma; NP in fig. 6), and Lincoln plutons (417.7±1.0 Ma; LI in fig. 6) (table 3;

Figure 8. 40Ar/39Ar age spectra of igneous rocks dated in this study.

Tucker and others, in press; Seaman and others 1995; Stewart and others, 1995). The North Union and Lake St. George bodies are granitic gneisses inferred to have been emplaced syntectonically (Tucker and others, in press). West and others (1995) reported widespread Late Silurian to earliest Devonian $^{40}Ar/^{39}Ar$ ages from metamorphic hornblendes outboard of the Sennebec Pond fault, essentially coeval with the plutons listed above.

EARLY EMSIAN PLUTONS

Previous findings. Nearly two-dozen closely dated of Emsian age define a northeast-southwest belt that can be traced across the entire strike length of figure 6, and beyond, into Massachusetts. Prior to the present study, there was some indication of plutonic activity around 400-410 Ma along this trend, but little to suggest that most plutons along the belt would turn out to be in the same narrow age range. In western New Brunswick, Bevier and Whalen (1990a) reported U/Pb ages 409±2 Ma for the Skiff Lake phase of the Pokiok batholith, and 402±1 for the Allandale phase (SK and AL, respectively, in fig. 6). In eastern Maine, Hubacher and Lux (1987) obtained $^{40}Ar/^{39}Ar$ ages, which are now known to be Emsian, from three plutons that truncate regional Acadian structures: the Pleasant Lake (400.5±4.5 Ma; PL in fig. 6), Hunt Ridge (401.5±4 Ma; HU in fig. 6), and Cochrane Lake plutons 404.3±3.7 Ma; CL in fig. 6). In western Maine, Solar and others (1998) recently reported Emsian U/Pb ages from the S. Roxbury granite (408.2±2.5 Ma; SR in fig. 6), Swift River granite (407.9±1.9 Ma; SW in fig. 6), Redington pluton (407.6±4.6 Ma; RD in fig. 6), N. Roxbury granite (404.3±1.9 Ma; NR in fig. 6), Lexington pluton (404.3±1.8 Ma; LX in fig. 6), and various phases of the Phillips pluton (405.3±1.8, 403.6±2.2, and 403.5±1.6 Ma; PH in fig. 6). In the Presidential Range of northern New Hampshire, the Wamsutta pluton (WM in figure 6), which was emplaced during F1 Acadian folding, has yielded a U/Pb zircon age of 408±1.9 Ma (J.D. Eusden, written communication, 1998). In New Hampshire, outside the area of figure 6, Emsian U/Pb ages have been obtained for the Spaulding pluton (408.4±1.9 Ma; zircon; Robinson and Tucker, 1996), the Ashuelot pluton (403±3; monazite; Robinson and Tucker, 1996), and the Cardigan pluton (404±2; zircon; R.D. Tucker, unpublished data). Finally, the Prescott gabbro, which post-dates the nappe phase of regional Acadian deformation in Massachusetts, has a U/Pb age of 407±3/2 Ma (Tucker and Robinson, 1990). Far outboard of the main belt of Emsian magmatism are two outliers: the Haskell Hill pluton (408±5 Ma; U/Pb; HH in fig. 6; Tucker and others, in press), and the Berry Brook gabbro-diorite (410 Ma; $^{40}Ar/^{39}Ar$; BB in fig. 6; Ludman and Idleman, 1998).

Russell Mountain pluton. The Russell Mountain pluton (RU in fig. 6) is a nonfoliated, two-mica granodiorite. Mapping by Ludman (1978) showed that the pluton truncates regional folds in the Madrid and Carrabassett Formations, and on this basis, it must post-date at least some Acadian deformation. Contact-metamorphic rocks on the northwestern margin, however, have a schistose foliation and aligned chiastolite with biotite strain shadows (fig. 9*A*). Because metamorphic biotite is incipient or absent outside the aureole (Ludman, 1978), some deformation must have taken place while the pluton was being emplaced. It is not known whether deformation in the contact aureole was caused by pluton emplacement, or was a regional event. Two concordant zircon fractions yield a mean $^{207}Pb/^{206}Pb$ age of 406.0±1.3 Ma (fig. 7). Biotite yielded a $^{40}Ar/^{39}Ar$ plateau age of 393.4±3.0 Ma (fig. 8).

Sebec Lake pluton. The Sebec Lake pluton (SL in fig. 6) is a nonfoliated biotite grandiorite that , like the Russell Mountain pluton, truncates regional folds involving the Madrid and Carrabassett Formations (Griffin, 1971). On the basis of its map-scale relations, it must post-date at least some Acadian deformation. Contact-metamorphic textures, however, are synkinematic. Metamorphosed Carrabassett Formation in the aureole contains large andalusite porphyroblasts with asymmetric strain shadows made up of biotite and chlorite aligned in the foliation (fig. 9*B*). Outside the contact aureole, the rocks are at sub-biotite grade; hence the biotite and andalusite are both related to plutonism. It is not known whether deformation in the contact aureole was caused by pluton emplacement, or was a regional event. We obtained two concordant zircon fractions that yielded a

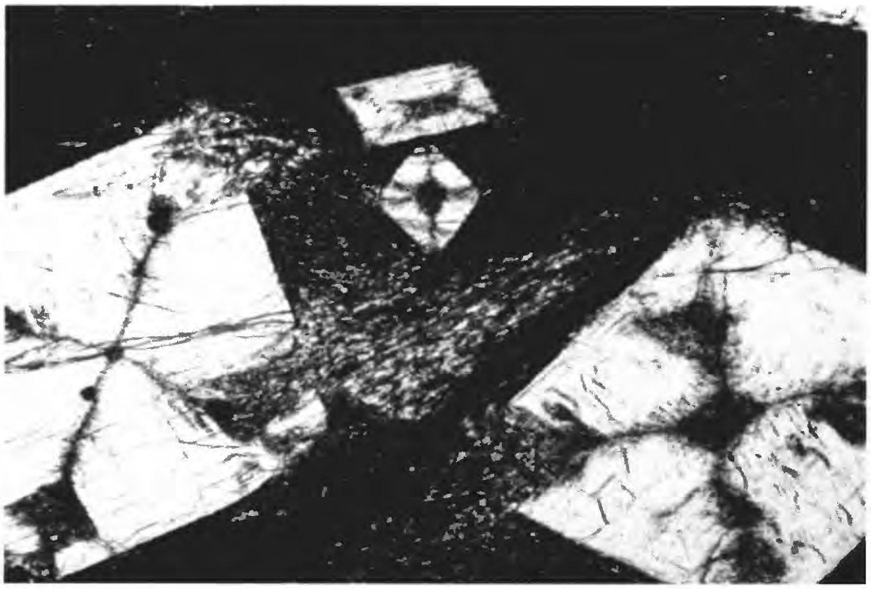


Figure 9*A*

Figure 9 (this and next two pages). Thin section photomicrographs of contact-metamorphosed rocks adjacent to Acadian plutons. All sections are cut normal to foliation and are viewed in plane-polarized light. Field of view in all cases is about 4 mm across. ***A***, Syntectonic metamorphic textures in the aureole of the Russell Mountain pluton. Large, randomly oriented chiastolite phenocrysts are flanked by strain shadows containing aligned flakes of biotite, implying some deformation during contact metamorphism. ***B***, Syntectonic metamorphic textures in the aureole of the Sebec Lake pluton. Chiastolite phenocrysts are flanked by strain shadows containing aligned chlorite and biotite, implying some deformation during contact metamorphism. ***C***, Static metamorphic textures in the aureole of the Bald Mountain pluton, showing randomly oriented, retrograded chiastolite phenocrysts and randomly oriented biotite. This rock shows no vestige of the pervasive regional foliation found outside the contact aureole, and therefore may have been metamorphosed prior to the regional deformation. Alternatively, contact metamorphism might have been intense enough to have obliterated all preexisting fabrics. ***D***, Syntectonic metamorphic textures in the aureole of the Hog Island pluton. Large cordierite phenocrysts overgrow a preexisting foliation defined by aligned biotite and white mica; this first foliation now survives only within the cordierite grains. A second, post-cordierite foliation is defined by biotite in pressure shadows. The metamorphic textures in this rock are broadly syntectonic but it is likely, based on regional relations, that the first Acadian deformation preceded emplacement of the Hog Island pluton by 15-20 million years. ***E***, Post-tectonic metamorphic textures in the aureole of the Priestly pluton. Cordierite phenocrysts overgrow a weak foliation that lies at a high angle to bedding; biotite in the groundmass is randomly oriented.

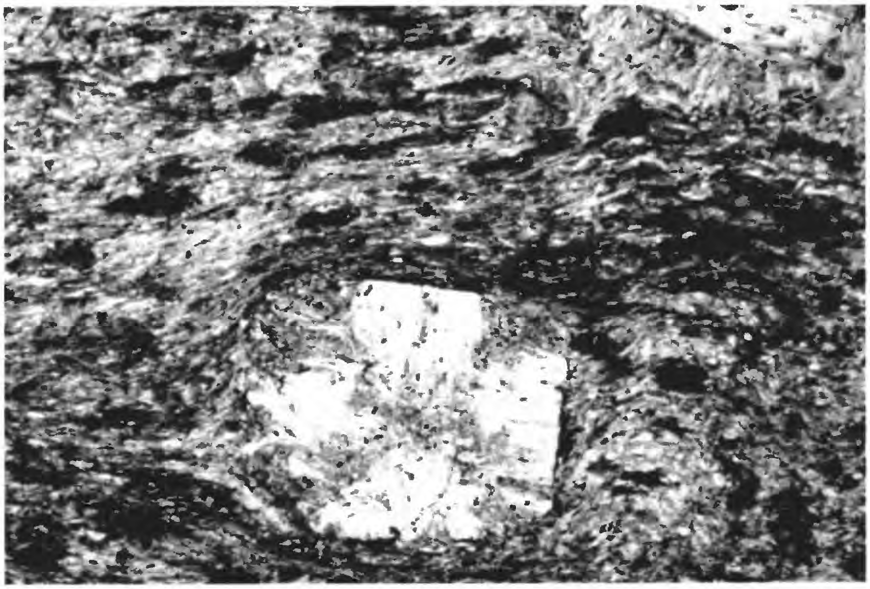


Figure 9*B*

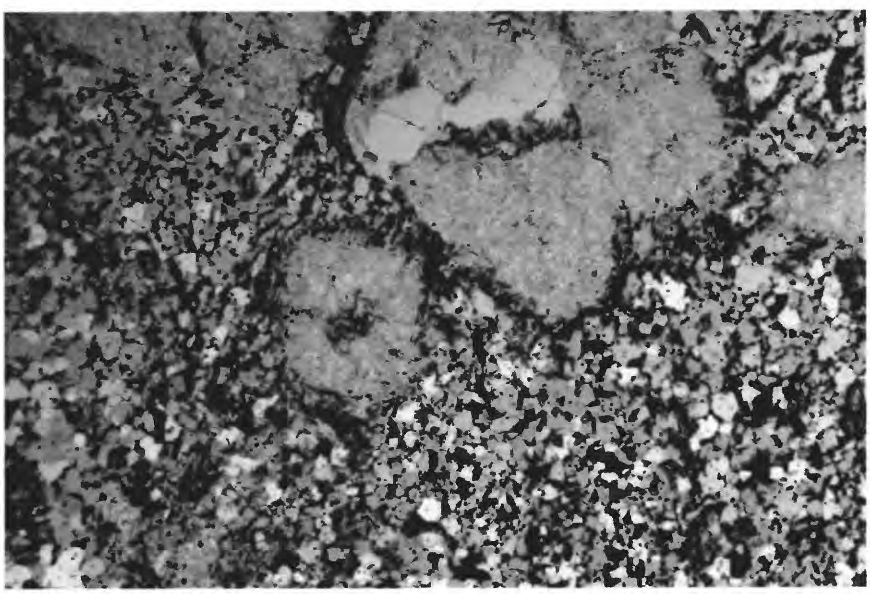


Figure 9*C*

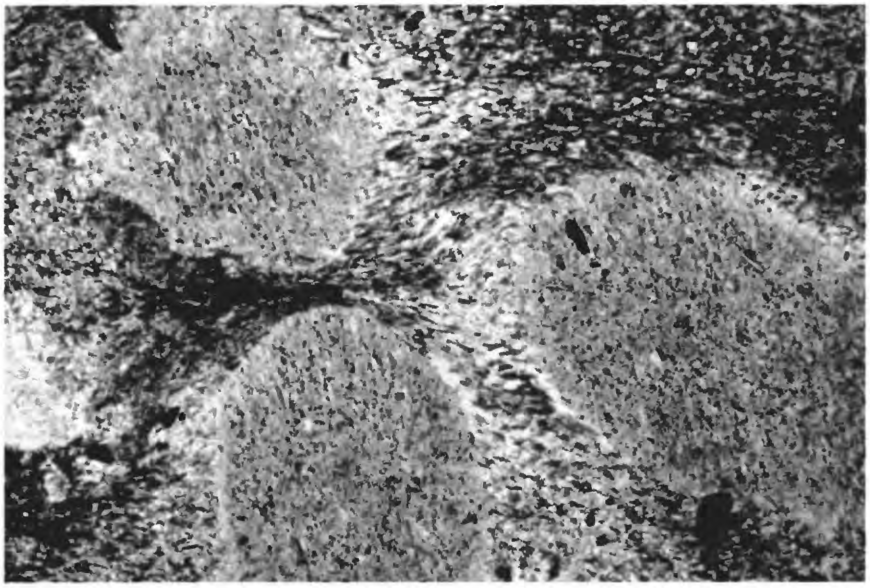


Figure 9*D*

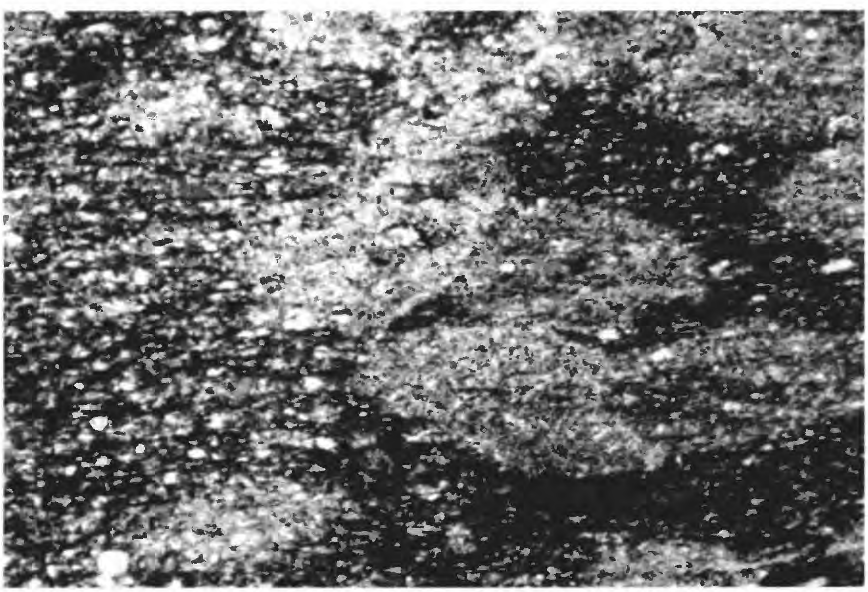


Figure 9*E*

mean $^{207}Pb/^{206}Pb$ age of 407.8±2.5 Ma (fig. 7), and $^{40}Ar/^{39}Ar$ plateau ages of 392.5±2.5 Ma from biotite and 377.2±4.4 Ma from muscovite (fig. 8).

Bald Mountain pluton. The Bald Mountain pluton (BM in fig. 6) is a nonfoliated, two-mica granodiorite; it is in the shape of an ellipse with its long axis normal to regional strike. Mapping by Espenshade and Boudette (1967) and Ludman (1978) showed that the pluton truncates regional folds in the Madrid and Carrabassett Formations and on this basis would appear to post-date at least some Acadian deformation. Unoriented chiastolite and biotite porphyroblasts in steeply dipping contact-metamorphosed strata also suggest post-tectonic emplacement (fig. 9*C*). A single concordant zircon fraction indicates an emplacement age near 408 Ma (table 4). $^{40}Ar/^{39}Ar$ ages were obtained from three biotite separates, two of which yielded plateaus: 390.9±3.0, and 394.3±3.7 Ma, giving a mean age of 392.6±2 Ma for cooling through the closure temperature of argon in biotite (fig. 8).

Onawa pluton. The Onawa pluton (ON in fig. 6) is a zoned body with a granodiorite core and a gabbroic rim. It truncates and deflects the regional structural grain defined by steeply dipping bedding in the Carrabassett Formation. Thin-section observations reported by Van Heteren and Kusky (1994) suggest that andalusite in the contact aureole overgrew a preexisting cleavage; hence the pluton is post-tectonic, at least in part. We obtained three concordant zircon fractions that yielded a mean $^{207}Pb/^{206}Pb$ age of 405.1±2.9 Ma from the granodiorite (fig. 7), and a $^{40}Ar/^{39}Ar$ plateau age of 400.1±3.7 Ma on biotite from the gabbro (fig. 8).

Shirley-Blanchard pluton. The Shirley-Blanchard pluton (SB in fig. 6) is a composite pluton that includes a more mafic phase of pyroxene-hornblende granodiorite (Blanchard body), and a more felsic phase of granodiorite (Shirley body)(Espenshade and Boudette, 1967). The pluton truncates regional-scale tight- to isoclinal folds involving the Madrid and Carrabassett Formations. From the Shirley body, we obtained two concordant zircon analyses with a mean $^{207}Pb/^{206}Pb$ age of 404.9±4.4 Ma (fig. 7), and a $^{40}Ar/^{39}Ar$ plateau age of 390.3±3.0 Ma on biotite (fig. 8). From the Blanchard body, we obtained two concordant zircon fractions that yielded a mean $^{207}Pb/^{206}Pb$ age of 407.2±1.5 Ma (fig. 7). The weighted mean 207/206 age of all zircon analyses from the Shirley-Blanchard body is 406.9±1.4 Ma, our preferred age for this composite pluton.

Mattamiscontis pluton. The Mattamiscontis pluton (MA in fig. 6) is an unfoliated biotite granite; field work for this study was limited to geochronological sampling in the pluton and brief examination of outcrops in the contact aureole. Reconnaissance mapping compiled by Osberg and others (1985) shows the pluton cutting a syncline cored by Carrabassett Formation; similarly, at one outcrop in the contact aureole, tightly folded metapelite displays a weak axial-planar cleavage that is overgrown by cordierite. A thin section from another outcrop, however, shows oriented cordierite surrounded by biotite strain shadows (the regional metamorphic grade is chlorite). These observations together suggest that the pluton postdates Acadian folding, but that some deformation — either regional or pluton-related — accompanied pluton emplacement. A granite sample yielded three concordant zircon analyses with a mean $^{207}Pb/^{206}Pb$ age of 406.9±3.6 Ma (fig. 7), and a $^{40}Ar/^{39}Ar$ biotite plateau age of 402.1±2.8 Ma (fig. 8).

Ebeemee pluton. The Ebeemee pluton (EB in fig. 6) is a small, unfoliated stock of biotite granite; field work for this study was limited to sampling in the pluton and fruitless search for informative outcrops in the contact aureole. Reconnaissance mapping compiled by Osberg and others (1985) shows the pluton cutting a doubly plunging anticline cored by Madrid Formation and flanked by Carrabassett Formation; on this basis, it would appear to post-date Acadian folding. Two concordant zircon analyses yielded a 207/206 age of 405.7±2.6 Ma, and a biotite separate yielded a $^{40}Ar/^{39}Ar$ plateau age of 407.8±2.4 Ma (fig. 8).

Seboeis pluton at Schoodic and Seboeis Lakes. The Seboeis pluton (SS in fig. 6) is shown on the bedrock geologic map of Maine (Osberg and others, 1985) as a large, 5 by 30 km body that cuts the Madrid and Smalls Falls Formations, and faults between them. Because it has not been mapped in detail and was sampled only in reconnaissance fashion, the following observations are necessarily preliminary. We dated four different phases. Weakly foliated granodiorite at Schoodic Lake yielded two concordant zircon fractions that give a mean $^{207}Pb/^{206}Pb$ age of 405.8±2.5 Ma (fig. 7). Near Dudley Rips, an unfoliated granite yielded a $^{40}Ar/^{39}Ar$ muscovite plateau age of 401.3±2.8 Ma (fig. 8). This granite intrudes a tectonically foliated diorite that yielded a slightly hump-shaped spectrum with a total-gas age of 396.3±3.8 Ma (fig. 8). Evidently, the diorite was intruded prior to 401 Ma and prior to at least some of the regional deformation. A much younger biotite age from rocks that had been previously mapped as the Seboeis pluton at East Branch Lake is discussed below in the section on Middle to Late Devonian plutons.

Moxie pluton. The northwestern (inboard) part of the Emsian plutonic belt is dominated by two very large intrusions: the Moxie gabbro and the Katahdin quartz monzonite. The intrusive relations of the Moxie (MX in fig. 6) are quite different from one end to the other. In the Jo Mary Mountain area at the pluton's eastern end (fig. 1, loc. 30), the contact aureole contains remarkably pristine, gently dipping strata that display perfectly preserved cross laminae (Hanson and Bradley, 1989). There is no evidence that contact-metamorphic minerals overgrew a regional slaty cleavage, which is seen everywhere outside the aureole. Mid-length along the pluton at loc. 44 (fig. 1), however, contact-metamorphic andalusite overgrows an axial-planar cleavage associated with nearly recumbent mesoscale folds. At the pluton's western end, rocks of the contact aureole show a strong foliation defined biotite and white mica, and by the preferred orientation of andalusite with asymmetric strain shadows of biotite. Thus, at three places where we have anecdotal observations, the pluton shows local evidence of pre-dating regional Acadian folding, post-dating Acadian folding, and of synkinematic emplacement. We dated the Moxie pluton at its eastern and central sections, and obtained $^{40}Ar/^{39}Ar$ biotite plateau ages of 404.4±3.4 and 406.3±3.8 Ma, respectively (fig. 8).

Katahdin pluton. The Katahdin pluton (KA in fig. 6) is a large (60 by 35 km) body of unfoliated quartz monzonite. It has long had an enigmatic age relationship to the Acadian deformation, but this problem can now be resolved — or at least narrowed — with a better time scale and precise U/Pb ages (Bradley and others, 1996; Rankin and Tucker, 1995). The Katahdin pluton intrudes rocks as young as the Seboomook Group (Pragian), Matagamon Sandstone (Pragian and earliest Emsian), and its own volcanic carapace, the Traveler Rhyolite (406 Ma; Rankin and Tucker, 1995). Near Harrington Lake (HA in fig. 6), the Seboomook Group near the pluton consists of gently dipping hornfelsed siltstones and sandstones that preserve pristine cross bedding. These rocks show no sign of ever having acquired the regional, northeast-striking, axial-planar cleavage, which is pervasive just outside the contact aureole, and which typically obscures or obliterates delicate sedimentary structures. In this area, the rocks in the contact aureole were baked and never cleaved, suggesting that the pluton is locally pretectonic. Along the pluton's northeastern border, in contrast, Neumann (1967) mapped a breccia zone consisting of plastically deformed metasedimentary rock fragments, set in a gneissic granitic matrix. These rocks may have formed along a syntectonic intrusive contact. The Katahdin pluton resisted many early attempts to date it accurately, but Denning and Lux (1989) reported a $^{40}Ar/^{39}Ar$ biotite plateau age of 400.1±1 Ma. More recently, Rankin and Tucker (1995) reported a U/Pb age of 406.9±0.4 Ma that is the weighted average of 5 concordant zircon fractions.

Harrington pluton. The Harrington pluton (HA in fig. 6) is a laccolith-like satellite to the Katahdin pluton (Griscom, 1976). Its map pattern (Griscom, 1976) implies that it may have been intruded into a folding sedimentary section, because its basal contact is planar and concordant with homoclinally dipping beds in the subjacent Seboomook Group, whereas its upper contact is concordant with regionally-folded bedding in the overlying part of the Seboomook Group. The

Harrington pluton yielded three concordant zircon analyses with a mean $^{207}Pb/^{206}Pb$ age of 406.9±1 Ma (fig. 7), which suggests an Emsian age for the folding.

LATE EMSIAN TO EIFELIAN PLUTONS

Previous findings. A few plutons in the age range 400-387.5 Ma are scattered across the area of figure 6, but form no obvious pattern. The Magagauadavic pluton in southern New Brunswick has a U/Pb age of 396±1 Ma (MG in fig. 6; M.L. Bevier, written communication, 1996). A biotite granite phase of the Mooselookmeguntic pluton (MO in fig. 6) yielded a U/Pb age of 388.9±1.6 Ma (Solar and others, 1998). The North Searsmont granitic gneiss (NS in fig. 6) yielded a U/Pb age of 389.0±2.0 Ma (Tucker and others, in press). The Horserace quartz-diorite (HR in fig. 6), a small body within the Katahdin batholith, yielded a concordant U/Pb zircon age of 392±3.1 Ma (R. D. Tucker, unpublished data); a $^{40}Ar/^{39}Ar$ hornblende age of 374.9±3.2 Ma on this pluton (Denning and Lux, 1989)

GIVETIAN TO EARLY TOURNAISIAN PLUTONS

Plutons in this age range are widespread in the area of figure 6. In hindsight, they are not particularly relevant to the main point of this paper, but they *are* important to late-Acadian events, and their place in the scheme of things could only have been learned by obtaining reliable ages.

Previous findings. The best-defined group consists of Frasnian-age plutons in central Maine. Tucker and others (in press) recently reported U/Pb ages for four of these: the Hartland (379±1 Ma; HD in fig. 6), Old Point (379±3 Ma; OP in fig. 6), Rome (378±1 Ma; RO in fig. 6), and Togus (378±1 Ma; TO in fig. 6) plutons. New ages are reported below for the Center Pond pluton and a sill at Hammond Ridge, which clearly belong in this group. Two other plutons that may belong in the same group crop out near the Maine coast: the Lucerne (380±4 Ma; LU in fig. 6; Zartman and Gallego, 1979) and the Deblois (384±5; DE in fig. 6; A. Ludman, written communication, 1998). Both have been dated by U/Pb zircon, but both ages have fairly large errors.

A cluster of seven previously dated plutons in the Eastern Townships of Québec and adjacent Maine fall in the Givetian to Famennian age range. Those dated by the U/Pb method include the Scotstown (384±2 Ma; SC in fig. 6), Lac aux Araignees (in Québec, or Spider Lake, in Maine; 383±3 Ma; LA in fig. 6); Winslow (377±6 Ma; WI in fig. 6); Aylmer (375±3 Ma; AY in fig. 6); and Ste-Cecile (374±1 Ma; SC in fig 6) (Simonetti and Doig, 1990). Plutons in this group dated by the $^{40}Ar/^{39}Ar$ method include the Chain of Ponds (373.3±2 Ma; CH in fig. 6) and Big Island Pond plutons (367.7±1.3 Ma; BI in fig. 6) (Heitzler and others, 1988). The newly dated Hog Island and Beaver Cove, discussed below, appear to belong in this group. Three other new ages from the Priestly, Deboullie, and Chandler plutons, also discussed below, are slightly younger but occur in the same general strike belt

Five previously dated plutons of Famennian age form a belt near the coast. In Maine, these include the Waldoboro pluton (368±2 Ma, WD in fig. 6), the Mt. Waldo pluton (371±2 Ma, MW in fig. 6), and two small pegmatite bodies dated at 367±1 and 371±1 Ma (GA and GR in fig. 6; Tucker and others, in press; all ages are U/Pb zircon ages). Similarly, M.L. Bevier (written communication, 1996) reported a U/Pb age of 366.5±1 Ma from the Mt. Douglas pluton in southern New Brunswick (MD in fig. 6).

Center Pond pluton. The Center Pond pluton (CP in fig. 6) is a hornblende-bearing granodiorite that intrudes steeply dipping, calcareous siltstones of Silurian age (Scambos, 1980). Two concordant zircon analyses yielded a mean $^{207}Pb/^{206}Pb$ age of 377±3 Ma (fig. 7). The pluton is cut in half and offset dextrally about 2 km by the Center Pond fault (Scambos, 1980). A biotite separate from a penetratively dextrally sheared granitoid near the fault yielded a $^{40}Ar/^{39}Ar$ age of 368.7±2.9 Ma (fig. 8), which may date the strike-slip event.

Hammond Ridge sill. At Hammond Ridge in the aureole of the Katahdin pluton (HM in fig. 6), a 50-cm-thick, tabular, concordant body of felsite interlayered with hornfels of the Carrabassett Formation yielded five concordant zircon fractions with a mean $^{207}Pb/^{206}Pb$ age of 377.2±1.5 Ma (fig. 7). This sill was at first mistaken for an ash bed that would have dated the Carrabassett Formation (the reason we analyzed so many fractions), but its young age requires that it be a sill.

Hog Island pluton. The Hog Island pluton (HI in fig. 6) intrudes Ordovician granodiorite, Silurian calcareous rocks, and Devonian flysch of the Seboomook Group at the northern end of the Boundary Mountains anticlinorium. The pluton was mapped by Albee and Boudette (1972) who interpreted the contact metamorphism as either syn- or post-regional metamorphism; we suggest the latter based on thin-section observations. Figure 9*D* shows a metapelite that contains biotite and cordierite, both formed during the contact-metamorphic event (the regional metamorphic grade is chlorite). Cordierite porphyroblasts overgrew a weak foliation, which we suggest formed during an earlier Acadian deformation; a second foliation is defined by aligned biotite, some of which occurs in strain shadows around cordierite. Although probably much younger than the first Acadian deformation, pluton emplacement was nonetheless synkinematic. A single concordant zircon fraction yielded a $^{207}Pb/^{206}Pb$ age of 371±3 (fig. 7); biotite yielded a $^{40}Ar/^{39}Ar$ plateau age of 368.6±3.1 Ma (fig. 8).

Beaver Cove pluton. The Beaver Cove pluton (BC in fig. 6) is a small, unfoliated stock of leucocratic biotite granodiorite; it is surrounded by the Moxie pluton. We obtained a $^{40}Ar/^{39}Ar$ plateau age of 372.4±2.7 Ma from a biotite separate (fig. 8), suggesting that the Beaver Cove is as much as 34 million years younger than the Moxie, and unrelated to it (compare with Gabis and others, 1994).

Deboullie pluton. The Deboullie pluton (DB in fig. 6) is a composite body that includes syenite, granodiorite, and monzonite phases (Boone, 1958). Boone's (1958) mapping showed that the pluton truncates steeply dipping bedding and cleavage in the Seboomook Group, and hence is younger than the main regional cleavage. Textures from the contact aureole, however, indicate that plutonism was followed by some additional shortening: contact-metamorphic biotite is cut by a strong pressure solution cleavage that parallels the regional fabric. Two samples of pink monzonite yielded $^{40}Ar/^{39}Ar$ plateau ages of 365.4±3.2 and 363.2±3.0 Ma on biotite. Similarly, biotite from granodiorite yielded a $^{40}Ar/^{39}Ar$ age of 364.2±3.2 (fig. 8).

Priestly pluton. The Priestly pluton (PR in fig. 6), a stock of unfoliated biotite-hornblende granodiorite, intrudes flysch of the Seboomook Group (Boudette and others, 1976). Textures in the contact aureole suggest that the pluton was emplaced into already foliated slates. At the northern end of the pluton, cordierite porphyroblasts contain inclusion trails of an older slaty cleavage (fig. 9*E*). Biotite is randomly oriented and indicates static conditions of contact metamorphism; cordierite porphyroblasts grew preferentially in the direction of the older foliation, but apparently also under static metamorphic conditions. A single concordant zircon fraction yielded a $^{207}Pb/^{206}Pb$ age of 362±3 (fig. 7). $^{40}Ar/^{39}Ar$ plateau ages of 363.2±2.3 and 360.3±2.7 Ma were obtained from biotite (fig. 8).

Chandler pluton. The Chandler pluton (CH in fig. 6) intrudes flysch of the Seboomook Group and Ordovician mafic igneous rocks along the eastern limb of the Munsungun anticlinorium. As far as we are aware, the pluton and surrounding rocks have never been mapped in detail, and outcrop is not plentiful. A total-gas age of $^{40}Ar/^{39}Ar$ 366.5±3.7 Ma was obtained from biotite (fig. 8); the age spectrum is slightly hump-shaped.

Seboeis pluton at East Branch Lake. As mentioned above, the Seboeis pluton is shown on the Geologic Map of Maine (Osberg and others, 1985) as a single body, but appears to include granodiorite, granite, and diorite of Emsian age, as well as a much younger phase. At East Branch Lake (SE in fig. 6), an unfoliated biotite granodiorite yielded a $^{40}Ar/^{39}Ar$ plateau age of 364.7±4.3Ma (fig. 8).

THE SILURIAN-DEVONIAN TIME SCALE: THE BASIS FOR CORRELATIONS BETWEEN PALEONTOLOGICALLY AND ISOTOPICALLY DATED EVENTS

We are able to make precise — and in some cases unexpected — correlations between isotopically and paleontologically dated events because of a parallel effort to improve the Silurian-Devonian time scale (Tucker, Bradley, and others, 1998). Four new tie-points require significant changes to previous time scales. Each is based on replicate, concordant, isotope-dilution U/Pb zircon analyses of ashes or volcanic strata, dated to within one or two conodont zones. The key dates are as follows: (1) The Bald Hill ash from the early Lochkovian Kalkberg Formation of eastern New York yielded a concordant U/Pb zircon age of 417.6±1.0 Ma. (2) The Sprout Brook ash from the early Emsian Esopus Formation of eastern York yielded a concordant U/Pb zircon age of 408.3±1.9 Ma. (3) The Eifelian Tioga ash from Wytheville, Virginia, yelded a concordant U/Pb zircon age of 391.4±1.8 Ma, in agreement with the concordant U/Pb monazite age of 390.0±0.5 Ma reported for the Tioga ash in Pennsylvania by Roden and others (1990). (4) The Little War Gap ash from the Frasnian part of the Chattanooga Shale of eastern Tennessee yielded a concordant U/Pb zircon age of 381.1±1.3 Ma. (5 and 6) A closely dated palyniferous horizon in the Famennian Piskahegan Group of New Brunswick is bracketed below and above, respectively, by rhyolites that yielded concordant U/Pb zircon ages of 363.8±2.2 and 363.4±1.8 Ma. The vertical axis in figure 4 is numerically calibrated according to the new time scale of Tucker and others (1998), which is contrasted with that of Harland and others (1990) in table 6.

These revisions are critical to the present study. Using the new time scale, most of the closely dated plutons listed in table 3 are now correlated with a younger stage than on the Harland and others (1990) time scale. Among the plutons that are critical to positioning the deformation front, the new time scale shows that the 424-417 Ma plutons in southern Maine are Ludlow to Lochkovian rather than Ludlow only; the 404-408 plutons of interior Maine are Emsian rather than Lochkovian; and the Scotstown pluton of the Eastern Townships is Givetian rather than Eifelian. Looking at it another way, the new time scale implies that the Sebec Lake, Onawa, Russell Mountain, and Blanchard plutons are the same age as the Tomhegan Formation, rather than being coeval with the Beck Pond Limestone, as implied by the Harland and others (1990) time scale.

LOCATION THROUGH TIME OF THE ACADIAN DEFORMATION FRONT AND FORELAND BASIN

In this section, we synthesize the foregoing results with the aid of two figures (10 and 11) which track the migration of the leading edge of the foreland basin, and of the deformation front. Six time slices are depicted, each spanning a few million years. The gross paleogeography and key stratigraphic localities are shown in figure 12 for the four best-constrained time slices.

In constructing these figures, the leading edge of the foreland basin at a given time was interpolated between the locations of foreland-basin and far-foreland deposits of that age. Likewise, the deformation front was interpolated between fossiliferous foreland-basin deposits laid down prior to regional deformation, and dated plutons of the same age which evidently cut Acadian structures. In some cases, as will be discussed, figures 10 and 11 were used interdependently to help fill in gaps in knowledge — but for the most part, the two figures are based on independent data, which nontheless lead to the same general conclusions. For the purposes of figure 10, the details of locally complex Acadian deformation histories are not that crucial, because the point is merely to locate a boundary between rocks that had already

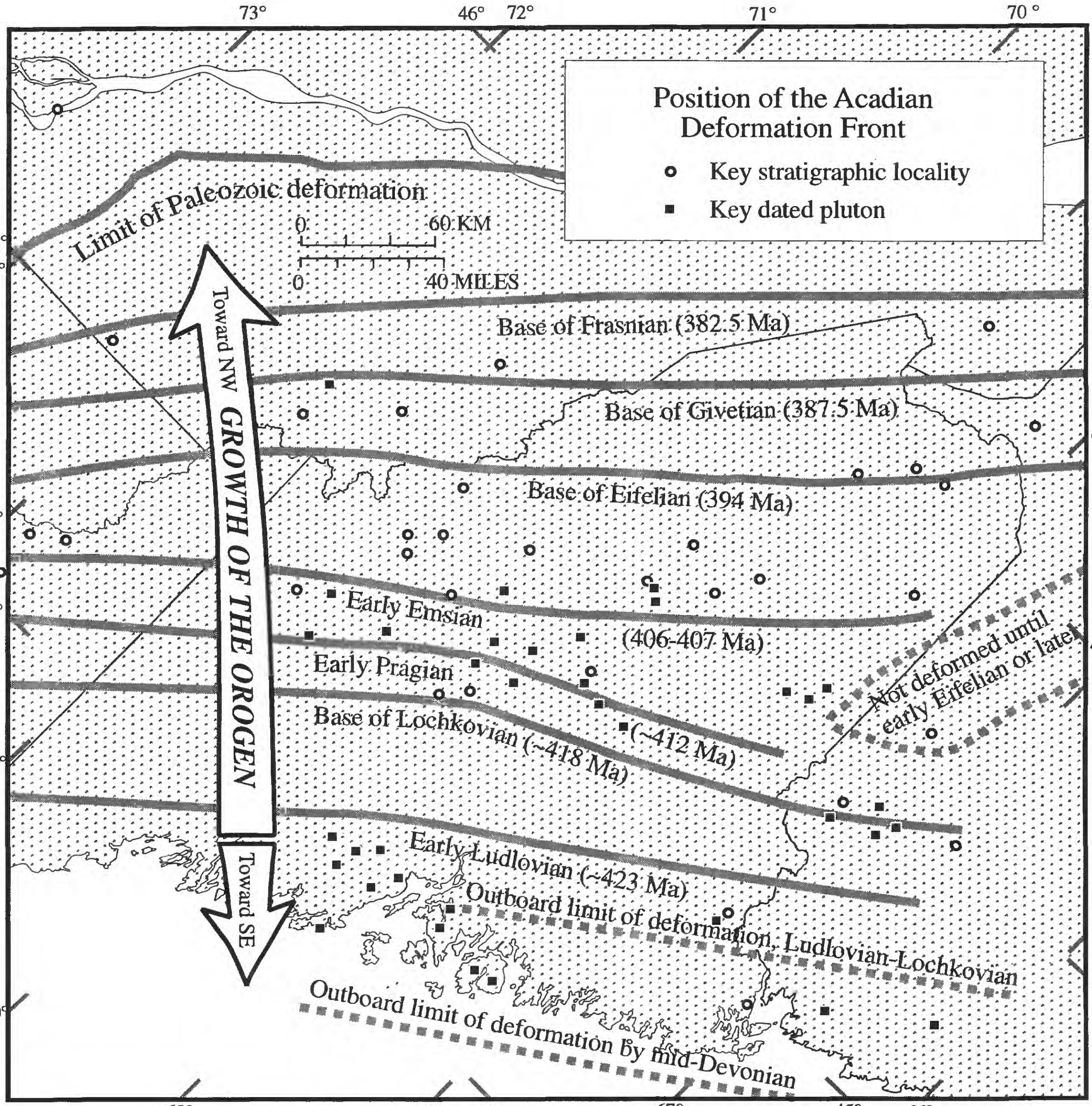


Figure 10. Map showing successive positions of the Acadian deformation front as it migrated across the northern Appalachians.

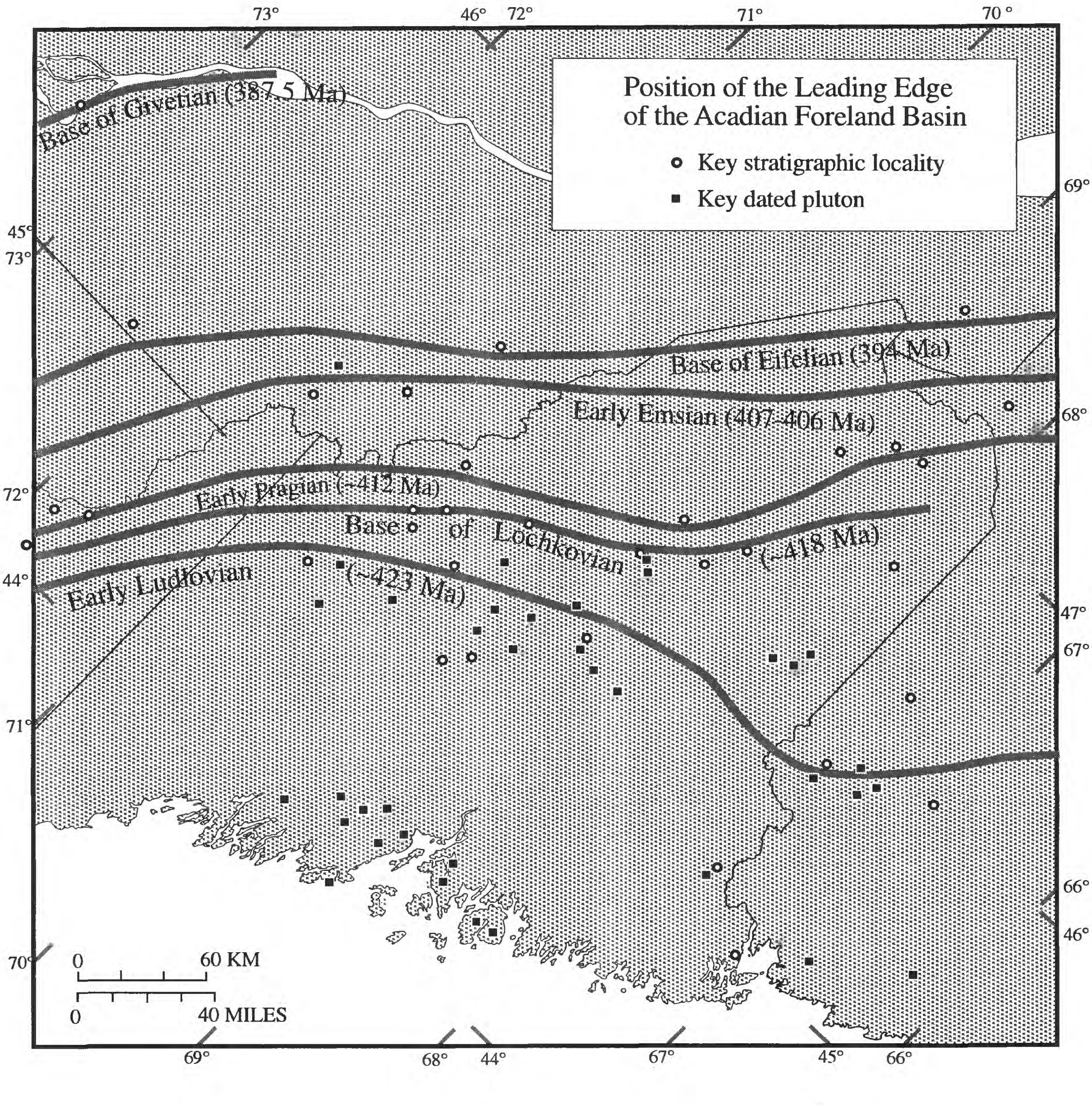


Figure 11. Map showing successive positions of the leading edge of the Acadian foreland basin, constructed following the approach used by Bradley (1989) for the Taconic foreland basin.

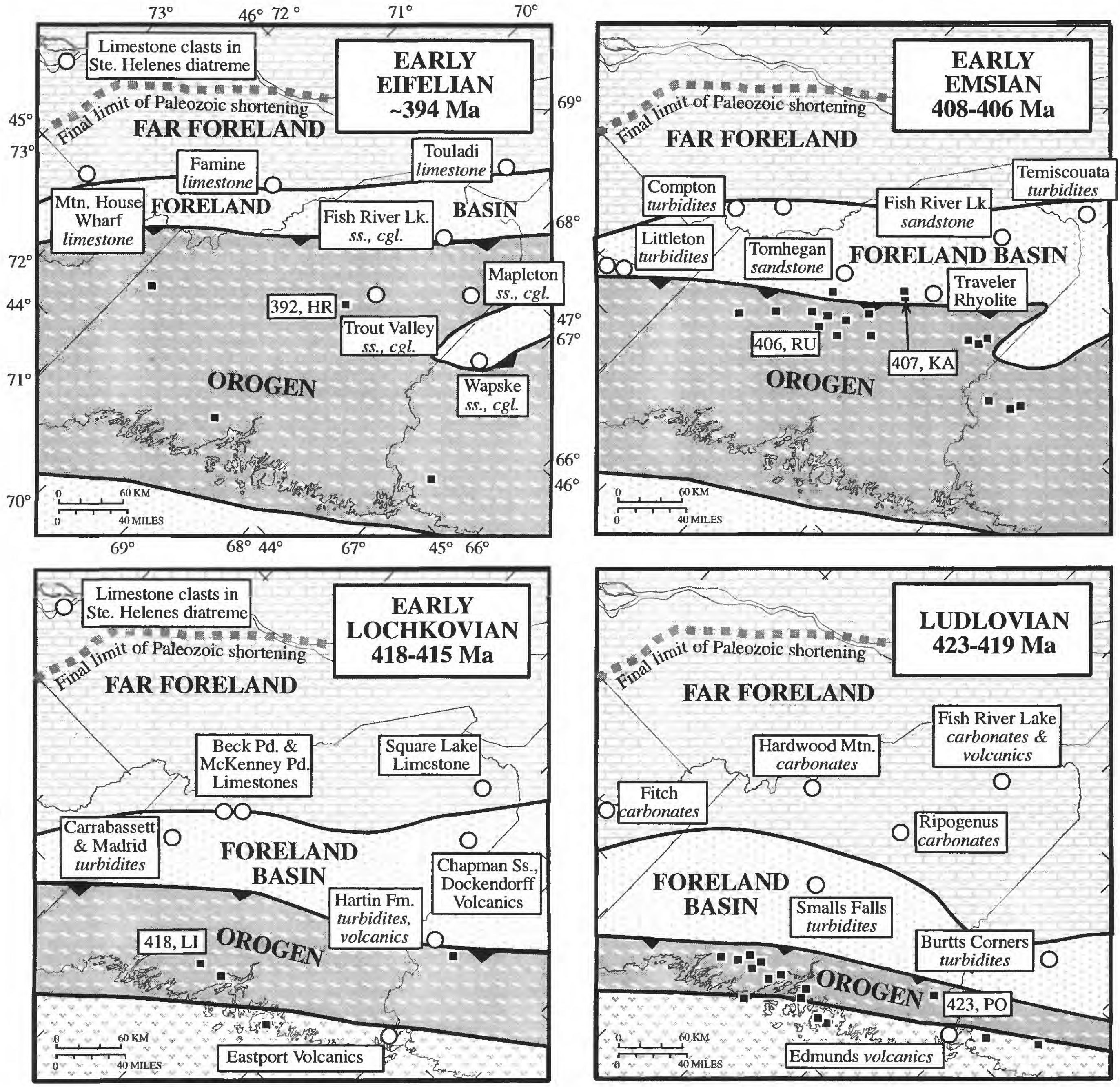


Figure 12. Schematic, non-palinspastic paleogeographic maps of the study area showing migration of the orogen and foreland basin during four time slices. Open circles represent fossil locations in stratified rocks of the appropriate age range, except the Traveler Rhyolite, which is dated isotopically. Black squares represent plutons in the appropriate age range. The age and two-letter abbreviation is shown for a few selected plutons, keyed to table 3 and figure 6. The deformation front is represented by a toothed line that probably was a blind thrust rather than an emergent thrust during much of the Late Silurian to Middle Devonian interval. The far foreland was an area of complex paleogeography, and for present purposes is meant to merely show the area beyond the limits of the Acadian foreland basin. Paleogeography is unknown for the area south of the orogen (present coordinates) during the Eifelian and Emsian.

experienced at least some deformation, from those that still had not. Likewise, unresolved aspects of the pre-, syn-, or post-tectonic nature of various plutons are not crucial for present purposes, so long as evidence exists that some regional deformation occurred in a particular plutonic belt prior to pluton emplacement.

PALINSPASTIC CONSIDERATIONS

The key localities are plotted on a present-day base map; no attempt has been made to remove the effects of Acadian folding and thrust-faulting, post-Acadian shortening, or dextral displacement across the Norumbega or other strike-slip faults. We originally had intended to use a palinspastic base, but found it impossible to construct such a map on the basis of published maps and cross sections. There are simply too few good marker units to unravel the structure, and the exposure in Maine is too poor.

We can, however, make three comments in support of our use of the present day base. First, none of the rocks in the area of figure 1 have been displaced enormous, immeasurable distances across terrane-bounding faults since the beginning of the Ludlow. By that time, the Pocomoonshine pluton had linked the Fredericton basin with the St. Croix belt (West and others, 1992), which in turn was linked by clasts in the Silurian Oak Bay conglomerate to the Coastal Volcanic belt (Ludman, 1981)(fig. 1, loc. 43). The Fredericton basin and Miramichi anticlinorium were linked by conglomerate clasts in the Taxis River beds by the Early Silurian (fig. 1, loc. 36). The Aroostook-Matapedia basin was two-sided as it has provenance linkages to anticlinoria both west and east (Berry and Osberg, 1989; Hopeck, 1991). Specifically, the Llandovery Frenchville Formation near the basin's western margin was derived from the Pennington anticlinorium just to the west (Roy and Mencher, 1976)(fig. 1, loc. 25), whereas the conglomeratic Daggett Ridge Formation (probably Late Ordovician to middle Silurian; Ludman, 1990) on the eastern margin was derived from the Miramichi (fig. 1, loc. 36). The Central Maine basin is linked stratigraphically to older rocks on the northwest: conglomerate in the Llandovery Rangeley Formation (fig. 1, loc. 26) includes clasts traced to the Attean pluton along the Boundary Mountains anticlinorium (Moench and Pankiwskyj, 1988). Finally, the Connecticut Valley-Gaspé basin can be seen as a foreshortened, two-sided basin, because Late Ordovician to Early Silurian conglomerates of the Cabano and Depot Mountain Formations along its northwestern margin in Québec and northernmost Maine were derived from Taconic highlands that lay to the northwest (Roy, 1989)(fig. 1, loc. 6), and along the opposite basin margin in northern Maine, the Fish River Lake Formation includes sandstone and conglomerate that can be traced to sources in the Pennington anticlinorium, just to the east (Boone, 1970)(fig. 1, loc. 21). The contact between the Central Maine and Aroostook-Matapedia basins is probably a major thrust fault (A. Ludman, written communication, 1998), but the two shared a common northwesterly source region in Silurian and hence are not exotic relative to one another. Similarly, the Central Maine and Fredericton basins are in fault contact and may not have been continuous with one another as some have suggested (McKerrow and Ziegler, 1971; Bradley, 1983). Nonetheless, they *are* linked by the chain of provenance relationships discussed above.

Second, evidence for the amount of fold-related Acadian shortening can be gleaned from some 1:62,500-scale quadrangle maps. We estimated shortening on fifteen cross sections, which were selected because they show at least one readily mapped contact, such as a carbonate-flysch contact. Using the sinuous-bed-length method, the ratio of the palinspastic cross-sectional width to width in the deformed state was found to range from 1.30 to 2.78 (fig. 13), with a mean value of 1.83. These results, if taken as being representative of the entire orogen, would suggest that the pre-Acadian across-strike width of figure 1 north of the Sennebec Pond fault was about twice the present width. The actual initial width was most likely even greater, because we only were able to estimate shortening for those exceptional places with simple, tractable structure. In the southeasterly part of the Central Maine basin in Maine (Osberg, 1988), and in the northwestern part of the same basin in New Hampshire (Eusden and others, 1996), early recumbent folds imply much greater amounts of shortening. Our estimate of shortening also does not account for the effects of penetrative strain.

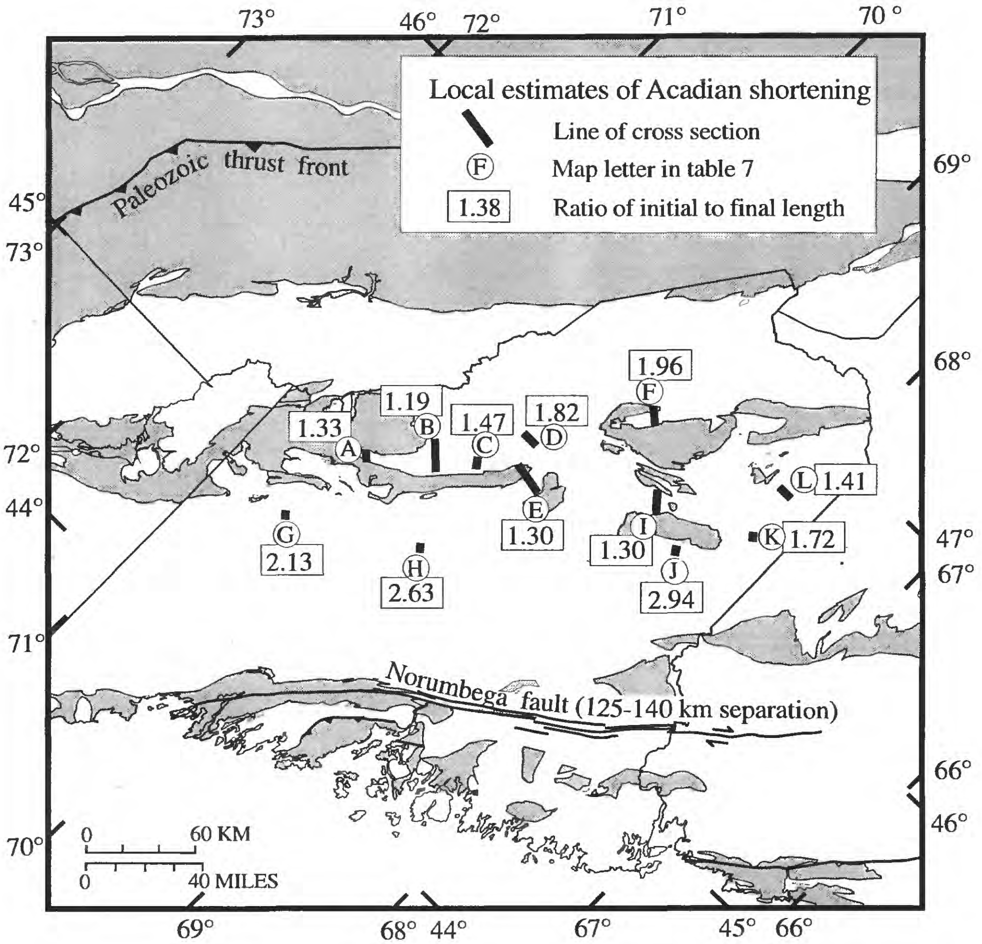


Figure 13. Map showing local estimates of Acadian and younger shortening, determined by the sinuous-bed-length method, across twelve published cross sections. Numbers in boxes are the ratios of initial to final cross-sectional lengths; a value of 2.63, for example, corresponds to a cross section that is now 5.4 km long but is inferred to have been 14.1 km long originally, by unfolding one or more marker horizons. Sources of data are listed in table 7.

Third, the only potentially problematic post-Acadian strike-slip fault in the area of figure 1 is the Devonian to Carboniferous-age Norumbega fault system. The displacement on the Norumbega may be large, but not so large that the rocks on either side are unrelated. Near the Maine-New Brunswick border, turbidites assigned to the Lower to middle Silurian Flume Ridge Formation occur at the same chlorite grade both northwest and southeast of the fault (Ludman, 1981). Map relations reviewed by Ludman and West (in press) suggests 125-140 km of cumulative dextral displacement across the Norumbega fault system.

In summary, the successive deformation fronts in figure 10 are probably more than twice as close together as they would be on an accurate palinspastic base. Also, localities outboard of the Norumbega fault were likely about 125-140 km farther northeast during the Acadian orogeny. Because the Norumbega fault is roughly parallel to Late Silurian paleogeographic belts, however, it does not have a major impact on any conclusions presented below.

EARLY LUDLOW

The Acadian orogeny was underway by the beginning of the Ludlow. The deformation front in figure 10 was placed between the post-tectonic Pocomoonshine pluton at loc. 37 and the graptolite-bearing Smalls Falls Formation and Burtts Corners beds at locs. 29 and 35. (All numbered localities in the ensuing discussion are keyed to figure 1.) The boundary between the foreland basin and far foreland was placed northwest of locs. 29 and 35 but southeast of shallow marine carbonate units such as the Hardwood Mountain, Spider Lake, Ripogenus, Grand Lake Seboeis, and Fish River Lake Formations (locs. 11, 18, 16, 20, and 21).

Although the main emphasis in this paper is on tracking the northwestward migration of the Acadian deformation front, there is also evidence that a second, outboard boundary between deformed and undeformed rocks migrated a much shorter distance to the southeast. In the Coastal Volcanic Belt, volcanism and sedimentation continued without interruption through most of the Silurian and into the Lochkovian. The southeastern boundary between deformed and undeformed rocks must have lain somewhere between the Pocomoonshine pluton and loc. 38 by early Ludlow time. It does not appear to have moved southeast until after Lochkovian time, that is, the age of the youngest Acadian-deformed strata at Eastport (loc. 38).

EARLY LOCHKOVIAN

During early Lochkovian, the Acadian deformation front was located along the northwestern flank of the Central Maine basin, and near the eastern margin of the Aroostook-Matapedia basin. Its position is bracketed by foreland-basin and far-foreland strata to the northwest, and syn- and post-tectonic plutons to the southeast. Plutonic rocks within the range 418.5 to 415 Ma delimit the zone of active deformation. These include the syntectonic Lincoln sill (417 Ma), and the Hartfield pluton (415 Ma).

A Lochkovian foreland basin flanked the plutonic belt on the northwest. Strata assigned to this stage of the migrating foreland basin include the Hartin Formation at loc. 34, the Chapman Sandstone and Swanback Formation at loc. 31, a part of the Millimigassett Lake Formation at loc. 20, and the Madrid and Carrabassett Formations in the Central Maine basin (for example, locs. 26 and 30). The early Lochkovian position of the leading edge of the foreland basin (fig. 11) is accurately placed at locs. 12 and 13, where the Beck Pond and McKenney Ponds limestones grade into or interfinger with the Seboomook Group and Tarratine Formation, respectively. Carbonate rocks of Lochkovian age include parts of the lower Fish River Lake Formation (loc. 21), Spider Lake Formation (loc. 18), and Square Lake Limestone (loc. 23). These presumably were deposited beyond the limits of the Lochkovian foreland basin.

PRAGIAN

Two lines of evidence show that the Pragian foreland basin migrated perhaps 15-20 km (present distance) cratonward from its early Lochkovian position (fig. 11). Along the Munsungun anticlinorium (loc. 16), the Pragian-age Seboomook flysch encroached over Lochkovian far-foreland carbonates. Similarly, in the Pennington anticlinorium (loc. 23), Pragian and younger flysch-like siltstones of unit Dus overlie the Square Lake Limestone. Within the foreland basin itself, the deltaic Matagamon Sandstone prograded over prodelta turbidites of the Seboomook Group. Although there is no direct evidence for migration of the deformation front from its Lochkovian position, in figure 10, we have assumed that it migrated the same distance as the foreland-basin margin did from its previous position (fig. 11). This assumption seems reasonable because the deformation front and foreland-basin front *did* migrate in tandem during those intervals, such as Ludlow to Lochkovian, for which have adequate data.

EARLY EMSIAN

The clearest picture of Acadian paleogeography in Maine is in the early Emsian, at 407-406 Ma (Bradley and others, 1996). We can identify a foreland basin and orogenic wedge. Superimposed on both, and straddling the deformation front, was a volcanic-plutonic belt. Many Emsian plutons truncate Acadian map-scale structures, including the newly dated Sebec Lake, Russell Mountain, and Shirley-Blanchard (SL, RU, and SB in fig. 6). The deformation front was plotted to the northwest of such plutons. The early Emsian foreland basin includes the younger strata of the vast "Devonian slate belt" of Boucot (1970). Foreland-basin flysch units that have yielded Emsian fossils include the Temiscouata Formation at loc. 24, the Compton Formation at locs. 3 and 4, and the Littleton Formation at locs. 9, and 10. It seems likely, though unproven, that parts of the Seboomook Group in the Connecticut Valley-Gaspé basin in northwestern Maine, and the Gile Mountain Formation of Vermont, are also Emsian. Early Emsian molasse deposits in the area of figure 1 include the Tomhegan Formation at (loc. 15), and, as suggested above, the uppermost part of the Matagamon Sandstone at loc. 17. These molasse facies must have been laid down near the proximal, southeastern margin of the foreland basin. The cratonward boundary of the early Emsian foreland basin can be positioned using negative evidence. As discussed below, the three occurrences of Eifelian-age limestones along the northwestern margin of the Connecticut Valley-Gaspé basin rest everywhere on pre-Emsian rocks, so it seems highly unlikely that Emsian foreland-basin deposits — which are kilometers thick, where present — were ever deposited that far northwest.

The deformation front must have corresponded to the Lobster Mountains anticlinorium (LMA in fig. 2), between the site of Tomhegan molasse deposition and the site of the plutons that were emplaced into a deforming or recently deformed orogenic wedge. The Moxie and Katahdin plutons (KA and MX in fig. 6) were emplaced in the general vicinity of the deformation front (Bradley and others, 1996).

LATE EMSIAN TO EARLY EIFELIAN

By the time of the Emsian-Eifelian boundary (~394 Ma), the Acadian deformation front had migrated some distance cratonward, to the approximate position of the Pennington anticlinorium (PA in fig. 2). Southeast of the deformation front in figure 10, late Emsian and younger strata post-date Acadian deformation. To the northwest, foreland-basin and far-foreland deposits of this age range predate Acadian deformation. At loc. 22, the upper siliciciclastic part of Boone's (1970) Fish River Lake Formation ("Nadeau Thoroughfare beds") were deposited in a foreland-basin setting. At locs. 2, 5, and 7, Eifelian-age limestones were deposited in a far-foreland setting. The Mapleton Formation and Trout Valley Formation are similar in age to the upper part of the Fish River Lake

Formation. Both, however, postdate the main phase of Acadian deformation and hence we place them southeast of the deformation front.

The stratigraphy of the Carlisle area (loc. 33) complicates an otherwise straightforward paleogeographic picture. Although it is the same age as the Mapleton Formation, the Wapske Formation *was* involved in the same regional "Acadian" folding as the rest of the Silurian to Devonian section. One possible explanation is that the Carlisle area escaped Acadian deformation until the late Eifelian or later. A second possibility is that the first phase of Acadian deformation, which in the Presque Isle area produced the unconformity below the Mapleton Formation, bypassed the Carlisle area entirely. In this scenario, the folding at Carlisle would be related to the same, younger event that caused gentle, late- or post-Acadian folding of the Mapleton Formation at Presque Isle. A third possibility is that the Wapske Formation was deposited farther to the northwest and was thrust into its present position during a late-stage Acadian event. The first option is illustrated in figures 10 and 12.

LATE EIFELIAN TO EARLY GIVETIAN

By the time of the Eifelian-Givetian boundary (~387.5 Ma), the Acadian deformation front had again migrated cratonward from its previous position, based especially on a belt of plutons in the Eastern Townships of Québec. These intrude already deformed Silurian and Devonian strata of the Connecticut Valley-Gaspé basin; the oldest and hence most significant is the Scotstown pluton (384±2 Ma; SC in fig. 6) (Simonetti and Doig, 1990). The deformation front must have been located somewhere to the northwest by about 384 Ma.

Just a few million years before emplacement of the Scotstown pluton, Eifelian carbonates were still being deposited at loc. 5. These strata are penetratively deformed, which shows that the final cratonward limit of the *Acadian* deformation front was at least that far northwest (the *Paleozoic* deformation front shown on figure 1 could be as old as Ordovician). Exactly when the deformation took place is unknown, but it was likely at least a few million years after deposition of the last carbonates (the youngest being the Famine Limestone at loc. 5), long enough to allow for an interval of foreland-basin sedimentation before deformation.

The late Eifelian to early Givetian stratigraphic record in the area of figure 1 is confined to a few clasts in the St. Helenes diatreme, near Montreal (loc. 1). The clasts include calcareous siltstone, sandstone, and some limestone of late Eifelian to early Givetian age (Boucot and others, 1986). By comparison of these rock types with the Catskill clastic wedge in New York, Montreal was probably near the cratonward edge of the Acadian foreland basin around the Eifelian-Givetian boundary (fig. 11).

BASE OF FRASNIAN

The position of the deformation front at the beginning of Frasnian time (fig. 10) is based on slender evidence. Deformed Eifelian carbonates at locs. 2, 5, and 7 show that the Acadian orogeny was felt at least that far to the northwest. The Frasnian deformation front in figure 10 was extrapolated from its position at the beginning of the Givetian, assuming a roughly constant convergence rate. The final northwestward limit of Acadian shortening is unknown; it might correspond to the Paleozoic thrust front (fig. 10), or alternatively, it might lie somewhere between the Frasian deformation front and the Paleozoic thrust front. Along the Paleozoic deformation front near Albany, New York, strata as young as Eifelian *were* involved in thrust faulting, probably Acadian but conceivably Alleghanian.

IMPLICATIONS

AGE OF THE ACADIAN OROGENY

A zone of active deformation migrated across Maine and adjacent parts of New England and Canada from early Ludlow through Givetian time. Initially, the orogen was a narrow belt, but it widened through time, eventually encompassing the entire State of Maine (fig. 10). The overall pattern of cratonward migration of the orogen is similar to that suggested by Donahoe and Pajari (1973), but is much changed in light of our new data from Maine and our improved time scale.

The Acadian orogeny has traditionally been regarded as a Devonian event. In the past decade, however, new pluton ages have been interpreted to show that the supposedly Devonian orogenesis in parts of the Northern Appalachians actually occurred in the Silurian (Bevier and Whalen, 1990b; West and others, 1992; West and others, 1995; Stewart and others, 1995a). Now, in light of revisions in the time scale that fix the Silurian-Devonian boundary at around 418 Ma, some of the purportedly Silurian plutons are back once again in the Early Devonian (for example, North Pole Stream and Mt. Elizabeth plutons of Bevier and Whalen, 1990b). The rest fall within the last 5 million years of the Silurian.

The northwestward migration of the deformation front in figure 10 might be interpreted in two ways. One possibility is that the orogeny occurred in pulses, each pulse separated by tectonic stability and each with some different cause. Some workers (for example, Boucot and others, 1964; Stewart and others, 1995a) have applied the term "Salinic" to late Silurian orogenic events in Maine — an orogeny that, in this view, is distinct from the Devonian Acadian orogeny. In like manner, other names for orogenies might be coined for other positions of the deformation front in figure 10. The alternative that we prefer is that the pattern in figure 10 resulted from a single plate-convergent event that swept the region over a long period of time. Existing data cannot resolve the two possibilities.

RATE AND TRAJECTORY OF DEFORMATION-FRONT MIGRATION

In figure 10, the deformation front is shown to have migrated about 240 km across strike in about 40.5 million years. Assuming continuous rather than episodic convergence, this implies a rate of about 6 km/m.y. (0.6 cm/yr). The actual rate must have been considerably faster because, as noted above, the base map is nonpalinspastic. Assuming, for example, that Acadian shortening reduced Maine to half its original across-strike width, the rate of thrust-front migration would be about 12 km/m.y. (1.2 cm/yr).

It is likely, but not assured, that the rate of thrust-front migration roughly equaled the rate of plate convergence (Bradley and Kusky, 1986; Bradley, 1989). This is the general case at oceanic subduction zones and in the snowplow analogy: a snowplow pushes a wedge of snow ahead of the plow-blade, which correspond to the orogenic wedge and backstop, respectively. Once a steady-state wedge shape is achieved, the deformation front in the snow migrates at about the same rate as the plow advances. An alternative possibility, however, is that the advance of Acadian thrusts was directly linked not to plate convergence, but rather to extensional orogenic collapse, as in the Carpathians (Royden and Burchfield, 1989). Although little is yet known about Acadian hinterland tectonics between about 420 and 380 Ma (see below on hinterland deformation regimes), there is no evidence for wholesale regional extension of the magnitude required by this mechanism.

Based on the *along-strike* diachronism of foreland-basin deposits in the northern Appalachians (younger toward the southwest), Bradley (1987, 1997) suggested that Acadian plate convergence had a dextral component. The migration of black-shale depocenters across the Acadian foreland-basin in the central Appalachians led Ettensohn (1987) to the same conclusion. The present study area, however, does not extend far enough along strike to provide clear evidence either for or against along-strike diachronism.

PHYSICAL SETTING OF MAGMATISM

One by-product of this study has been to show that many of the Acadian plutons of Maine occur in distinct age belts (fig. 6). Moreover, figure 12 shows how the various plutons and volcanic rocks were spatially related to the orogen as it existed at the time of magmatism.

Coastal volcanic belt, and Ludlow to Lochkovian plutonic belt. On geochemical grounds, Gates and Moench (1981), and Hogan and Sinha (1989) suggested that the Ludlow to Lochkovian volcanic rocks and plutons of coastal Maine formed in an extensional setting. The migration pattern of the deformation front (figs. 10 and 12), however, implies that magmatism took place in an overall setting of plate convergence. The same conclusion is suggested on a more local scale by the fact that the Lochkovian-age Lincoln sill was emplaced synkinematically (Tucker and others, in press). Judging from figure 12, magmatism took place in the upper plate of a two plate system, presumably above an A-type subduction zone.

Piscataquis volcanic belt and Emsian plutonic belt. The cause or causes of Ludlow to Emsian volcanism along the Piscataquis belt has long been debated, and although a resolution is beyond the present scope, our findings do provide an important new constraint: *all of the volcanic rocks are broadly syncollisional and were erupted in a foreland setting.* Most of the Ludlow to Lochkovian volcanics (East Branch Group, West Branch Volcanics, and Fish River Lake, Spider Lake, Grand Lake Seboeis, and Allagash Lake Formations) were erupted in a *far*-foreland position. In addition, several volcanic units of Lochkovian and Emsian age are interbedded with or grade laterally into foreland-basin deposits, and hence formed in a *proximal* foreland position: the Hedgehog Volcanics, Edmunds Hill Andesite, Hartin Formation, Traveler Rhyolite, and Kineo Rhyolite. Three alternative interpretations have been advanced for part or all of the Piscataquis magmatic belt. (1) On the basis of basalt discriminant plots, Hon and others (1992) and Keppie and Dostal (1994) interpreted the Spider Lake and West Branch volcanics as having formed in an intracontinental rift setting, related to so-called "transpressive rifting". In light of our findings, any purported extension would necessarily have taken place in a collisional foreland setting; extension in forelands is generally either caused by lithospheric flexure (Bradley and Kidd, 1991), or by Baikal-style cracking of the lower plate (Sengör and others, 1978). (2) Tucker and others (in press) have suggested that the Emsian- foreland-basin rhyolites and the Emsian plutons just to the southeast were a product of lower-lithospheric delamination of the downgoing (North American) plate, which was being subducted toward the southeast. (3) Bradley (1983) interpreted these units as arc volcanics formed over a northwest-dipping subduction zone. This model would explain magmatism in all three settings — far foreland, proximal foreland, and orogenic wedge — as the consequence of an orogenic wedge above a southeast-dipping subduction zone colliding with (and overriding) a magmatic arc formed over a northwest-dipping subduction zone. (An updated version of this tectonic model was illustrated by Hanson and Bradley, 1989.)

SYN-COLLISIONAL VERSUS PRE-COLLISIONAL PLATE GEOMETRY

Tectonic interpretations of the Acadian orogeny have long been debated. Part of the controversy regarding Acadian plate geometry may be sidestepped for present purposes by separately considering the plate geometry during two intervals: prior to initial impact of the converging plates, and during the actual collision. The paired migration of the orogen and foreland basin shown in figures 10 through 12 suggests that, *during* collision, an upper plate consisting of the growing orogenic wedge and its Avalonian backstop overrode the Taconic-modified margin of North America. The implied A-type subduction zone that operated during collision presumably evolved naturally out of a B-type subduction zone of the same southeast-dipping polarity, where ocean floor that once existed between Avalonia and North America was consumed during late Llandovery and Wenlock.

What remains to be settled is the nature of the northwestern margin of the Central Maine basin *prior to* collision. This is a much-debated topic and the coauthors of the present paper are themselves divided on the subject. In the two main tectonic models, both of which have variants, it was either (1) a "passive" or rifted margin on the backside of an arc that had collided with North America during Ordovician time (for example, Tucker and others, in press; Robinson and others, 1998), or (2) a convergent margin that took up some of the motion between Avalonia and North America during Silurian time (Bradley, 1983). In the second model, the northwestern subduction complex and arc was overridden by, and downflexed beneath, the southeastern one described in the preceding paragraph (Bradley, 1983). In the one-subduction-zone model (Robinson and others, 1998), the northwestern passive margin was overridden by, and downflexed beneath, a southeastern subduction-collision complex. Although the two models differ significantly up until the time of collision, they are not very different after that.

HINTERLAND DEFORMATION REGIMES AND REGIONAL PARTITIONING OF ACADIAN DEFORMATION

The present study has focused on tracking the *first* Acadian deformation across strike. In most places, however, Acadian deformation was polyphase. Figure 10 provides a framework for understanding those post-D1 deformations that occurred between about 423 and 382.5 Ma, that is, the ages of the oldest and youngest deformation fronts. Three examples will be discussed here.

Stewart and others (1995a) described a complex history of motion on a network of northeast-striking dextral faults in Penobscot Bay. The oldest of these, the Penobscot Bay-Smith Cove-North Blue Hill fault, displaces isograds around the Segwick pluton (419.5±1 Ma), but is cut by the South Penobscot Intrusive Suite (419.2±2.2 Ma). Later, the South Penobscot Intrusive Suite was cut by the dextral Turtle Head fault, which was intruded, in turn, by the Lucerne Pluton (380±4 Ma, Wones 1991). Where plate convergence has an oblique component, deformation can be partitioned into thrusting in the orogenic wedge and strike-slip faulting further to the rear (Dewey, 1980). Applying this concept to the latest Silurian episode of dextral strike-slip faulting on the Maine coast, subduction of Central Maine basin crust beneath the Coastal Volcanic Belt evidently had a dextral-oblique component.

Near Waterville (WA in fig. 6), a 399-Ma intermediate dike cuts the first phase Acadian folds but is deformed by second generation of folds that must be older than a static metamorphic event dated at 380 Ma (Osberg, 1988; Tucker and others, in press). The second folding must therefore have taken place when the Acadian deformation front was somewhere cratonward of its early Emsian position (fig. 10), perhaps 100 km to the northwest of Waterville. Accordingly, this contractional deformation was an out-of-sequence, hinterland event.

The Horserace pluton is a small body with a U/Pb age of 392 Ma (R.D. Tucker, unpublished data), within the 407-Ma Katahdin batholith. It was emplaced, probably synkinematically (Hon, 1980, p. 73), along a northwest-striking high-angle fault (the West Branch fault) that sinistrally offsets the western margin of the Katahdin pluton. This is one of many cross faults that disrupt the bedrock map pattern in northern and central Maine (Osberg and others, 1985). Referring to figures 10-12, at around 392 Ma, the deformation front lay perhaps 70 km to the northwest when the West Branch fault was active. Like the second folding at Waterville, sinistral motion on the West Branch fault took place in the Acadian hinterland, while the orogen was still advancing toward North America.

FORELAND DEFORMATION REGIMES

Figures 10-12 also provide insights into pre-Acadian events that took place in what was then the orogenic foreland. Two northeast-striking high-angle faults at Spencer Mountain (fig. 1, loc. 12) cut the Hardwood Mountain Formation (Burroughs, 1979), to which we assign a late Ludlow to Pridoli age (table 2). The map pattern indicates that one of these faults is overlain by the Hobbstown Conglomerate, which ranges from possibly as old as Pridoli to definitely as young as Pragian (Boucot and Heath, 1969). The sense of motion is not well established. Based solely on the relative age of faulted strata, Burroughs (1979) showed both faults as having down-to-northwest motion; evidence is lacking that would bear on a strike-slip component. Whatever the exact age of faulting and the sense of motion, it *is* clear that faulting took place in a foreland setting, and that the high-angle faults paralleled the approaching orogenic front. On this basis, they would appear to resemble normal faults of the Taconic foreland of New York, which formed in response to lithospheric flexure (Bradley and Kidd, 1991).

SUGGESTIONS FOR FURTHER WORK

Since an influential paper by Naylor (1970), the Acadian has often been described as "an abrupt and brief event". This appears to be half true. Although the main, first phase of deformation was indeed abrupt, lasting but a few million years *at a given place*, it took some 40 m.y. for the deformation front to make its way across Maine. Our study thus confirms the findings of Donahoe and Pajari (1973), in general if not in particular. With the benefit of a new time scale, we can see in hindsight that their correlations between specific poorly dated plutons and specific intervals of the Devonian were flawed.

Figures 10 through 12 could be improved with better age control for a number of key units. The Flume Ridge, Madrid, and Carrabassett Formations are particularly significant, but are so loosely dated by fossils as to be almost useless for present purposes; a dedicated search for fossils, especially palynomorphs, and (or) datable pyroclastic horizons in these units would be worth the effort. Tighter paleontological control is needed, but less urgently, for the Hartin, Perry, and Swanback Formations, the Chapman Sandstone, and the Seboomook Group in northernmost Maine. Likewise, modern geochronological studies are needed for the West Branch, Spider Lake, and Hedgehog Volcanics, the Red Beach and Lucerne plutons in Maine, and the many Devonian plutons in the Northeast Kingdom of Vermont. The Devonian time-scale still has holes to be filled in the Pragian, late Emsian, and Givetian; until these intervals are better calibrated, correlations between isotopically and paleontologically dated rocks in these age ranges will remain equivocal.

The depositional environments, paleocurrents, and provenance of unnamed unit Dsus and the Sangerville, Smalls Falls, Hartin, Wapske, Fish River Lake, and Swanback Formations warrant detailed study; we have assigned these units to a foreland-basin setting on the basis of lithology and regional relations alone. The Wapske is of special interest because it was deformed about 15 million years later than rocks along strike in Maine (fig. 10). Accordingly, the Wapske Formation may have been deposited in a "piggy-back" basin, a possibility that might be evaluated using detailed sedimentological studies.

Our assessments of the pre-, syn-, or post-tectonic age of plutons are admittedly cursory, based on anecdotal observations of a few thin sections and one or two traverses from country rock into pluton. Detailed investigations would be especially useful for those plutons, such as the Moxie and Katahdin, that appear to have been emplaced very close to the deformation front.

The present study has emphasized evidence bearing on the northwestward migration of the Acadian orogen, but we have barely touched on its migration to the southeast. Did it migrate in a step-wise fashion, or gradually? Does the Eastport Formation of coastal Maine and New Brunswick preserve a record of synorogenic sedimentation off the back side of the growing Acadian mountains?

A palinspastic base will be essential for discerning geographic relationships between correlative rocks at the time they were formed, and for quantifying the rate at which the orogen and its foreland advanced during collision.

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Table 1, Palynomorph data from Devonian rocks in Maine.

Rock Unit	GSC #	Station #	Loc. # in Figure 1	Lat	Long	Quad	Location	Lithology and Structure	Age	Reference	Collector
Fish River Lake Formation	0-106531	95MDw116A	22	47°00'52"	68°35'18"	Eagle Lake 7.5'	Nadeau Thoroughfare railroad cuts, 90-120 meters northeast of bridge.	Medium sandstone, friable, plant bearing, 65° bedding dip, no cleavage.	Most likely, and no older than, mid-late Emsian to very early Eifelian (*douglastownense-eurypterota* Zone). Spores from succeeding *velata- langii* Zone, early Eifelian, are absent.	McGregor (1996)	Bradley
Fish River Lake Formation	0-106532	95MDw117	22	47°01'04"	68°35'05"	Eagle Lake 7.5'	Nadeau Thoroughfare railroad cuts, 490-520 meters northeast of bridge.	Medium sandstone, micaceous, friable, plant bearing, 51° bedding dip, no cleavage.	Most likely, and no older than, mid late Emsian to very early Eifelian,*douglastownense-eurypterota* Zone. Acritarchs indicate marine waters.	McGregor (1996)	Bradley
Fish River Lake Formation	0-106533	95MDw118	22	47°01'14"	68°34'53"	Eagle Lake 7.5'	Nadeau Thoroughfare railroad cuts, 910-950 meters northeast of bridge.	Medium sandstone, micaceous, friable, plant bearing, 50° bedding dip, no cleavage.	Probably assignable to, and no older than, mid late Emsian to very early Eifelian, *douglastownense-eurypterota* Zone. Acritarchs indicate marine waters.	McGregor (1996)	Bradley
Fish River Lake Formation	0-106536	95MDw124	22	47°02'30"	68°33'48"	Eagle Lake 7.5'	Eagle Lake, south shore, near Camps of Acadia.	Gray micaceous siltstone, 65° bedding dip, strong cleavage.	Spores present, Late Silurian or younger.	McGregor (1996)	Bradley
Fish River Lake Formation	0-106537	95MDw125	22	47°00'52"	68°35'44"	Eagle Lake 7.5'	Nadeau Thoroughfare, under Route 11 bridge.	Gray siltstone, outcrop-scale folds with axial-planar cleavage.	Probably middle Pragian to earliest Emsian, *polygonalis-emsiensis* Zone. Could be next younger zone, but less likely.	McGregor (1996)	Bradley
Fish River Lake Formation	6274	MK-62-8	22	47°58'37"	68°38'09"	Island Pond 7.5'	St. Froid Lake, west shore, exact location uncertain.	Lithology of analyzed sample unknown, but strongly cleaved, gray, micaceous siltstone observed near here in 1995.	Early Eifelian, *douglastownense-eurypterota* Zone.	McGregor (1968)	Mencher
Fish River Lake Formation	6309	MK-62-9	22	47°58'35"	68°38'09"	Island Pond 7.5'	St. Froid Lake, west shore, 30 meters S of MK-62-8; exact location uncertain.	Lithology of analyzed sample unknown, but strongly cleaved, gray, micaceous siltstone observed near here in 1995.	Late Emsian to Early Eifelian, *douglastownense-eurypterota* Zone.	McGregor (1962)	Mencher
Fish River Lake Formation	6612	CL-186	40	46°58'26"	68°34'32"	Winterville 7.5'	Rte 11, 480 meters north of Winterville; float sample.	Unknown	Early to middle Eifelian, *velata-langii* Zone.	McGregor (1963)	Mencher
Fish River Lake Formation	7754	EM-82	22	47°02'58"	68°30'02"	Eagle Lake 7.5'	Eagle Lake, north shore, 80 meters from eastern boundary of quadrangle.	Unknown	Middle Pragian or early Emsian.	McGregor (1968)	Mencher
Fish River Lake Formation	7760	EM-980	22	47°57'24"	68°39'38"	Island Pond 7.5'	Red River, 640 meters south of mouth of Labbe Brook.	Unknown	Emsian.	McGregor (1968)	Mencher
Mapleton Formation	6275	MK-62-10	31	46°46'51"	68°04'57"	Presque Isle 15'	Same locality as Stop 2 of Roy and White (1987).	Unknown	Early to middle Eifelian, *velata-langii* Zone.	McGregor (1992)	Mencher

Trout Valley Formation	0-105776	Unknown	17	Unknown	Unknown	Unknown	Unknown	Unknown	Late Emsian to early Eifelian, *douglastownense-eurypterota* Zone.	McGregor (1992)	Mencher
Unit DSus of Osberg and others (1985)	0-106534	95MDw119	23	47°06'36"	68°22'57"	Square Lake W 7.5'	2.2 km north of Burnt Landing on Square Lake access road.	Tan laminated siltstone, 90° bedding dip, weak cleavage.	Small spores suggest Late Silurian to Lochkovian.	McGregor (1996)	Bradley
Unit DSus of Osberg and others (1985)	0-106535	95MDw120	23	47°07'26"	68°22'52"	Square Lake W 7.5'	4.7 km north of Burnt Landing on Square Lake access road.	Dark siltstone, 80° bedding dip, weak cleavage.	Undoubtedly Early Devonian, probably Lochkovian or early Pragian.	McGregor (1996)	Bradley
Unit Ds of Osberg and others (1985)	7758	EM-856	41	**46°57'**	**68°27'**	**Portage 15'**	Lumber road near W. branch Beaver Brook.	Unknown	Probably Pragian.	McGregor (1968)	Mencher
Uni t Dus of Osberg and others (1985)	7766	DR-1	42	**Pending**	**Pending**	Ashland 15'	Ashland Cemetery. **Wait for info from Dave Roy**	Unknown	Probably Pragian or Emsian.	McGregor (1968)	Mencher

Table 2. Conodont collections from Silurian and Devonian rocks, Maine and Quebec

[See figure 5 for the Silurian and Devonian conodont zonation and graphic display of the age range of each sample.]

LOC. NO. IN FIG. 1	FIELD NO. (USGS collection number)	COUNTY QUADRANGLE LATITUDE/ LONGITUDE	FORMATION, LITHOLOGY, AND GEOGRAPHIC AND GEOLOGIC SETTING	CONODONTS	AGE	BIOFACIES	CAI	REMARKS
23	95MDw123A (12511-SD)	Aroostook Square Lake West 47°03'52"/ 68°22'57"	Square Lake Limestone — prominent lake shore outcrop of biostromal limestone containing abundant brachiopods, corals, crinoids, and bryozoans.	145 *Belodella devonica* (Stauffer) 21 *Decoriconus* sp. 4 *Dvorakia* sp. *Ozarkodina excavata* (Branson and Mehl)? 1 Sa and 1 Sb element fragments *Ozarkodina remscheidensis remscheidensis* (Ziegler) 10 Pa, 5 Pb, 2 M, and 2 Sc elements (chiefly fragments) 9 *Pseudooneotodus beckmanni* (Bischoff and Sannemann) 42 indet. bar, blade, platform and coniform fragments	late Ludlovian-early Lochkovian (*siluricus* Zone to at least *woschmidti* Zone).	Belodellid biofacies: this collection, with an abundance of coniform elements, is typical of reefoid deposits in the Silurian and earliest Devonian.	1.5	8.7 kg processed; 4.02 kg +20 mesh and 220 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly pyrite euhedra and euhedral clusters, ferruginous quartz grains and ferruginous quartzose siltstone grains.
	95MDw123C (12512-SD)		Square Lake Limestone — grainstone, from low outcrops at water line about 50 m west and stratigraphically above 95MDw123A.	1 Pb *Oulodus* sp. indet. *Ozarkodina remscheidensis* (Ziegler) 8 Pa (*O. r. remscheidensis*), 1 Pa (*O. r. eosteinhornensis*), 1 Pb, 1 Sa, and 1 Sc elements (chiefly fragments) 15 indet. bar, blade, and platform fragments	late Ludlovian through earliest Pragian (*siluricus* Zone into lowermost *sulcatus* Zone	Indeterminate— too few generically determinate conodonts. Conodonts indicate high-energy, normal-marine depositional setting.	1.5	3.8 kg processed; 400 g +20 mesh and 62 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered pyrite euhedra and euhedral clusters and ferruginous quartz grains.
21	95MDw129	Aroostook Fish River Lake 46°50'52"/ 68°46'02"	Fish River Lake Formation — silty limestone containing crinoid and stromatoporoid debris.	2 Sc element fragments of post-Ordovician Paleozoic morphotype several phosphatic bryozoan pearls	Middle Ordovician-Mississippian on the basis of the bryozoan pearls.	Indeterminate— too few conodonts.	3	5.2 kg processed; 3.3 kg +20 mesh and 466 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered pyrite euhedra and euhedral clusters and ferruginous quartz grains.
16	95MDw130 (12513-SD)	Piscataquis Harrington Lake 45°52'55"/ 69°10'38"	Ripogenus Formation — limestone. Sample is from 2-m-thick limestone with lenticular shaly partings separated from the big conglomerate by about 1 m of calcareous sandstone.. Cliffs in woods on inside of 90° bend in Frost Pond Road. Highest bed in cliff is a 5-m-thick limestone conglomerate.	2 *Decoriconus* sp. indet. 6 *Dvorakia* sp. indet. 1 Pb element *Oulodus* sp. indet. *Ozarkodina excavata* (Branson and Mehl) 1 Pa, 1 M, 1 Sa, 6 Sb, and 4 Sc elements 59 *Panderodus* sp. indet. 34 indet. bar, blade, and platform fragments	Wenlockian-early Lochkovian.	Panderodid biofacies: normal-marine, probably relatively shallow-water depositional environment.	chiefly 5.5; minor 5 and 6	11.5 kg processed; 4.7 kg +20 mesh and 195 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: composition not noted.
27	95MDw155 (12514-SD)	Somerset The Forks 45°19'58"/ 69°58'05"	The Forks Formation — calcareous sandstone to sandy limestone. Large roadcut on east side of US 201, 0.3 mile south of Kennebec River bridge; sample from prominent 20-cm-thick, white calcareous sandstone. Sample stratigraphically higher than 95MDw180.	*Ozarkodina excavata* (Branson and Mehl) 5 Pa, 1 Pb, 1 M, and 1 Sc elements 46 indet. bar, blade, and platform fragments	late Llandoverian-early Emsian.	Indeterminate— too few conodonts. Normal-marine, relatively high-energy depositional setting.	6.5	12.1 kg processed; 4.9 kg +20 mesh and 1.2 kg 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered pyrite-bearing quartzose sandstone and quartzose siltstone and minor talc.

27	93MDw18 (12334-SD)	Somerset The Forks 45°18.3'/ 69°59.2'	The Forks Formation from roadcut on east side of US 201, 4.9 miles north of Caratunk. Recessive, conspicuously calcareous horizon 75 ft. north of telephone pole no. 234,445,131 and 10 ft. down and to left of spray-painted graffiti labeled "aout 80".	3 *Decoriconus* sp. indet. *Ozarkodina excavata* (Branson and Mehl) 5 Pa, 2 M, and 1 Sb elements 2 *Panderodus* sp. indet. 33 indet. bar, blade, and platform fragments Phosphatized steinkerns: 1 gastropod and 1 pelmatozoan ossicle	late Llandoverian-early Lochkovian.	Indeterminate—too few conodonts. Preservation and composition of fauna suggest a high-energy, normal-marine depositional setting.	5.5-6.5	9.9 kg processed; 4.02 kg +20 mesh and 737 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly pyrite euhedra and euhedral clusters and composite ferruginous and pyrite-bearing quartz grains.
13	95MDw207A (12515-SD)	Somerset Enchanted Pond 45°27'15"/ 70°12'23"	McKenney Ponds Limestone Member of Tarratine Formation. Outlet of McKenney Ponds dies out into a karst depression with three cave entrances oriented NE.-SW. 95MDw207A from middle cave and from ~5 m below lowest exposure of Tarratine Formation siltstone.	14 *Belodella devonica* (Stauffer) 8 *Dvorakia* sp. indet. 34 P element fragments of *Icriodus* sp. indet. of Lochkovian-Pragian morphotype 3 *Pseudooneotodus beckmanni* (Bischoff and Sannemann)	Lochkovian-Pragian.	Post-mortem transport within or from the icriodid biofacies; relatively high-energy normal-marine depositional setting.	5	11.4 kg processed; 200 g +20 mesh and 96 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly phosphatic brachiopod fragments and euhedral pink zircons.
	95MDw207C (12516-SD)		McKenney Ponds Limestone Member of Tarratine Formation. Sample is from the southwesterly of the three caves and from an irregularly bedded bioclastic limestone, 3.5 m stratigraphically above 95MDw207D.	37 *Belodella devonica* (Stauffer) 67 P (all fragments) and 5 coniform elements of *Icriodus* sp. indet. of Lochkovian, but not early Lochkovian morphotype (not *I. woschmidti*) 2 Pa elements *Ozarkodina remscheidensis remscheidensis* (Ziegler) 2 *Pseudooneotodus beckmanni* (Bischoff and Sannemann)	Lochkovian, but not early Lochkovian.		5	19.5 kg processed; 20 g +20 mesh and 77 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly weathered and fresh pyrite, composite ferruginous grains and phosphatic brachiopod fragments.
	95MDw207D (12517-SD)		McKenney Ponds Limestone Member of Tarratine Formation. Sample is from the southwesterly of the three caves and from an irregularly bedded bioclastic limestone, about 3.5 m below the stratigraphically highest limestone (95MDw207C), and from approximately the same stratigraphic level as sample 95MDw207A.	14 *Belodella devonica* (Stauffer) coniform elements 1 small P element fragment *Icriodus* sp. indet. *Ozarkodina remscheidensis eosteinhornensis* (Walliser) 23 Pa and 1 Pb elements 1 *Pseudooneotodus beckmanni* (Bischoff and Sannemann) 32 indet. bar, blade, and platform fragments	Lochkovian.	Ozarkodinid biofacies: normal-marine, shelf depositional environment	5	10.7 kg processed; 180 g +20 mesh and 29 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly composite ferruginous grains, dolomite(?), and lesser phosphatic brachiopod fragments.
12	95MDw208B (12518-SD)	Somerset King and Bartlett Lake 45°22'05"/ 70°20'18"	Beck Pond Limestone. Carbonate matrix conglomerate with clasts of granite and reef debris. This is outcrop 10 of Boucot and others (1959). This sample is from the youngest member (Boucot and others' Member 5) of the Beck Pond.	110 *Belodella devonica* (Stauffer) 14 *Decoriconus* sp. 2 *Dvorakia* sp. 6 coniform elements *Icriodus* sp. indet. *Ozarkodina remscheidensis remscheidensis* (Ziegler) 75 Pa, 17 Pb, 16 M, 1 Sa, 3 Sb, and 6 Sc elements 75 indet. bar, blade, and platform fragments	early Lochkovian (the overlap of icriodids and *Decoriconus* indicates the early Lochkovian).	Ozarkodinid biofacies: normal-marine, shallow-water shelf depositional environment.	4.5-5	41.2 kg processed; 7.6 kg +20 mesh and 1.7 kg 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly phosphatic zooecial linings and pearls of bryozoans, conodonts, detrital muscovite, biotite, chlorite, weathered pyrite, and euhedral zircon.
	95MDw210 (12519-SD)	Somerset King and Bartlett Lake 45°22'11"/ 70°20'41"	Beck Pond Limestone. Sandy limestone containing abundant bryozoans and corals. This is the same locale as outcrop 32 of Boucot and others (1959) and is from the oldest member (Boucot and others' Member 1) of the Beck Pond.	5 Pa elements *Ozarkodina remscheidensis eosteinhornensis* (Walliser)	late Ludlovian-early Lochkovian (*siluricus* Zone to at least *woschmidti* Zone).	Indeterminate—too few conodonts.	4.5-5	8.3 kg processed; 340 g +20 mesh and 208 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly composite ferruginous grains and flakes.

LOC. NO. IN FIG. 1	FIELD NO. (USGS collection number)	COUNTY QUADRANGLE LATITUDE/ LONGITUDE	FORMATION, LITHOLOGY, AND GEOGRAPHIC AND GEOLOGIC SETTING	CONODONTS	AGE	BIOFACIES	CAI	REMARKS
12	95MDw211 (12520-SD)	Somerset Spencer Lake 45°22'40" 70°20'34"	Bear Pond Limestone Member of Seboomook Group (formerly Seboomook Formation). This is outcrop 42 of Boucot and others (1959).	1 *Belodella devonica* (Stauffer) 1 Pa element *Ozarkodina remscheidensis eosteinhornensis* (Walliser) 3 P element fragments *Icriodus* sp. indet. or *Pedavis* sp. indet.	late Ludlovian-earliest Pragian (no older than *siluricus* Zone)	Indeterminate—too few conodonts.	5	11.1 kg processed; 1.43 kg +20 mesh and 2.2 kg 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly phosphatic zooecial linings of bryozoans and phosphatic tubes, ferruginous grains, and rare euhedral zircon.
14	95MDw212B (12521-SD)	Somerset King and Bartlett Lake ~45°18.7'/ ~70°19.6'	Parker Bog Formation — limestone interbedded with siliceous siltstone or tuff. From south bank of Spencer Stream; detailed outcrop map in Boucot and others (1958). From the thickest carbonate bed ~7m downstream from a prominent point.	*Oulodus elegans* (Walliser) 1 Pa, 1 Pb, 1 M, 1 Sb, and 1 Sc *Ozarkodina remscheidensis* (Ziegler) subsp. indet. 29 Pa, 5 Pb, and 2 M element fragments 2 *Panderodus* sp. indet. 4 *Pedavis* sp. indet. coniform elements 85 indet. bar, blade, and platform fragments	late Ludlovian into late Pridolian (from at least *snajdri* Zone into late Pridolian).	Ozarkodinid biofacies; normal-marine, shelf depositional environment. Condition of conodonts indicates relatively high-energy depositional regime.	chiefly 5.5 and minor 5 and 6	8.5 kg processed; 1.56 kg +20 mesh and 973 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly ferruginous siltstone and detrital mica.
11	95MDw216B (12523-SD)	Somerset Jackman 45°38'32"/ 70°20'50"	Hardwood Mountain Formation from outlet stream of Sugar Bertg Pond, near Little Big Wood Pond. Rubble crop of coral-bearing limestone.	*Corryssognathus dubius* (Rhodes) 1 Sb and 3 coniform elements *Oulodus* sp. indet. 1 Pa, 4 Pb, 3 Sb, and 1 Sc elements *Ozarkodina excavata excavata* (Branson and Mehl) 5 Pa and 1 M elements *Ozarkodina remscheidensis remscheidensis* (Ziegler) 8 Pa, 2 M, 1 Sb, and 3 Sc elements 1 *Panderodus* sp. indet. UNASSIGNED ELEMENTS: 1 Pb, 1 M, and 1 Sa 87 indet. bar, blade, and platform fragments	late Ludlovian (late Gorstian through Ludfordian in terms of the British stages). *Corryssognathus dubius* is the most biostratigraphically diagnostic species and is restricted to the late Ludlovian.	Ozarkodinid-oulodid; relatively shallow-water, moderate- to high-energy shelf depositional setting.	5	16.6 kg processed; 100 g +20 mesh and 126 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered pyrite euhedra and ferruginous argillaceous grains and rare subrounded to euhedral zircon.
11	95MDw217A (12524-SD)	Somerset Jackman 45°38'35"/ 70°20'58"	Hardwood Mountain Formation from Sugar Berth Pond. Shaly limestone, 5 m above 95MDw216B and youngest sample collected at 95MDw216 or 95MDw217.	2 Sb elements *Oulodus* sp. indet. 2 Pa elements *Ozarkodina confluens* Branson and Mehl *Ozarkodina remscheidensis remscheidensis* (Ziegler) 8 Pa, 1 Pb, and 1 Sb robust elements 2 *Panderodus* sp. indet. 27 indet. bar, blade, and platform fragments	late Ludlovian-Pridolian.	Indeterminate—too few conodonts. Normal-marine, relatively high-energy depositional environment.	5	24.0 kg processed; 3.97 kg +20 mesh and 220 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly ferruginous micaceous and argillaceous flakes and common phosphatic bryozoan pearls.
11	95MDw217B (12525-SD)	Somerset Jackman 45°38'35"/ 70°20'58"	Hardwood Mountain Formation near Little Big Wood Pond. Shaly limestone 10 m below 95MDw216B.	*Oulodus* sp. indet. 1 Pa, 1 Pb, 1 M, 1 Sb, and 1 Sc elements *Ozarkodina remscheidensis remscheidensis* (Ziegler) 12 Pa, 5 Pb, and 1 M elements (mostly fragments) UNASSIGNED ELEMENTS: 1 Pb, 1 Sb, and 1 Sc 21 indet. bar, blade, and platform fragments	late Ludlovian.	Ozarkodinid; normal-marine shelf depositional setting.	5	16.9 kg processed; 2.36 kg +20 mesh and 202 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly ferruginous argillaceous flakes and fresh and weathered pyrite.

2	95MDw220A (12526-SD)	Quebec, Canada Lake Memphre-magog 45°03'40"/ 72°16'50"	Mountain House Wharf Limestone. Calcareous slate and limestone. At north end of Chemin Mountain. 95MDw220A is from lake shore cliffs, lowest horizon at extreme left end of outcrop.	20 deformed P elements *Icriodus* sp. indet. of late Emsian-Frasnian morphotype (short-platform icriodids) Icriodids of this type are speciated on the basis of platform margin outline. Because these specimens are moderately to strongly deformed, they could not be determined to species.	late Emsian-Frasnian.	Post-mortem transport within or from the icriodid biofacies. These specimens originated in a relatively high-energy reefal or biohermal environment.	upper 5	11.1 kg processed; 8.72 kg +20 mesh and 570 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered euhedral pyrite and pyrite-bearing, carbonaceous semischist fragments.
	95MDw220B (12527-SD)		Mountain House Wharf Limestone. From 30 m below old dam and about 120 m up from lake along small stream that flows past tennis courts—from farthest downstream horizon exposed in outcrop.	4 deformed P elements *Icriodus* sp. indet. of late Emsian-Frasnian morphotype (short-platform icriodids)	late Emsian-Frasnian (latest Early-early Late Devonian)	Indeterminate—too few conodonts. Icriodids indicate derivation from relatively high-energy reefal or biohermal environment.		10.2 kg processed; 2.12 kg +20 mesh and 320 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered euhedral pyrite and pyrite-bearing, carbonaceous semischist fragments.
	95MDw220E (12528-SD)		Mountain House Wharf Limestone. About 8 m structurally above and 30 m right of 95MDw220A.	2 P element fragments *Icriodus* sp. indet.	Devonian	Indeterminate-too few conodonts.		7.1 kg processed; 4.58 kg +20 mesh and 124 g 20-200 mesh insoluble residue. Heavy-mineral concentrate: chiefly fresh and weathered euhedral pyrite and pyrite-bearing, carbonaceous argillite fragments.

Table 3. Emplacement ages of plutons shown in figure 6.

Pluton	State /Prov	Abbr. in Fig. 6	Age	Error	Method	Mineral	Comments	Reference
Allandale pluton, Pokiok batholith	NB	AL	402.0	1	U/Pb	monazite	Mean 207/235 age; 2 concordant fractions	Bevier and Whalen 1990a
Aylmer pluton	QU	AY	375.0	3	U/Pb	monazite	206/238 age of 1 nearly concordant fraction	Simonetti and Doig 1990
Bald Mtn pluton	ME	BM	408.0		U/Pb	zircon	207/206 age; 1 concordant fraction	This paper
Beaver Cove pluton	ME	BC	372.4	2.7	40/39	biotite	Plateau age	This paper
Berry Brook gabbro-diorite	ME	BB	410.0		40/39	hornblende		Ludman and Idleman, 1998
Big Island Pond	ME	BI	367.7	1.3	40/39	hornblende	Average of 2 plateau ages: 367.7±2.1 and 367.6±2.1	Heizler and others, 1988
Blinn Hill granite gneiss	ME	BH	424.0	2	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Tucker and others, in press
Cadillac pluton	ME	CD	419.0	2	U/Pb	zircon	No details given	Seaman and others, 1995
Center Pond pluton	ME	CP	377.0	3	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions	This paper
Chain of Ponds pluton	ME, QUE	CH	373.3	2	40/39	hornblende	Average of 6 plateau ages: 371.6, 371.1, 375.4, 373.2, 372.5, 375.9, all ±2.1	Heitzler and others, 1988
Chandler	ME	CA	366.5	3.7	40/39	biotite	Total gas age; almost a plateau	This study
Cochrane Lake	ME	CL	404.3	3.7	40/39	hornblende	Average of 3 plateaus: 401±4, 405±5, and 407±7	Hubacher and Lux, 1987
Deblois, northeastern arm	ME	DE	384.0	5	U/Pb	zircon	No details given	A. Ludman, written communication, 1998; date by T. Lanzirotti
Deboullie pluton, granodiorite	ME	DB	364.2	3.2	40/39	biotite	Plateau	This paper
Deboullie pluton, monzonite	ME	DB	363.9		40/39	biotite	Avg of two plateaus: 362.3±3 and 365.4±3.2	This paper
Ebeemee pluton	ME	EB	407.8	2.4	40/39	biotite	Plateau age	This paper
Ebeemee pluton	ME	EB	405.7	2.6	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions	This paper
Hammond Ridge sill	ME	HM	377.2	2.5	U/Pb	zircon	Mean 207/206 age; 6 concordant fractions	This paper
Harkshaw pluton, Pokiok batholith	NB	HK	411.0	1	U/Pb	sphene	207/206 age; 1 concordant fraction	Bevier and Whalen 1990a
Harrington pluton	ME	HA	406.9	1	U/Pb	zircon	Mean 207/206; 3 concordant zircon fractions	This paper
Hartfield pluton, Pokiok batholith	NB	HT	415.0	1	U/Pb	sphene	207/206 age; 1 concordant fraction	Bevier and Whalen 1990a
Hartland pluton	ME	HD	378.9	1.3	U/Pb	zircon and spene	Mean 207/206 age; 4 concordant fractions	This paper

Haskell Hill pluton	ME	HH	408.0	+5/-4	U/Pb	zircon	Mean 207/206 age; 1 concordant fraction	Tucker and others, in press
Hog Island	ME	HI	370.0		U/Pb	zircon	207/206 age; 1 concordant fraction	This paper
Horserace quartz diorite	ME	HR	392.0		U/Pb	zircon	207/206 age; 1 concordant fraction	Tucker, unpublished data
Hunt Ridge pluton	ME	HU	401.5	4	40/39	clinopyroxene and hornblende	Average of 2 plateau ages: 401±4 and 402±5.	Hubacher and Lux 1987
Katahdin pluton	ME	KA	407.0	0.4	U/Pb	zircon	Mean 207/206; 5 concordant zircon fractions	Rankin and Tucker, 1995
Lac aux Araignees (Spider Lake) pluton	QU	LA	383.0	3	U/Pb	zircon	207/206 age; 1 concordant fraction.	Simonetti and Doig 1990
Lake Saint George granite gneiss	ME	LS .	422.0	2	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Tucker and others, in press
Lexington pluton	ME	LX	404.2	1.8	U/Pb	zircon	Mean 207/206 age; 1 concordant fraction	Solar and others, 1998
Lincoln shonkinite	ME	LI	417.7	1	U/Pb	zircon	Mean 207/206 age; 4 concordant fractions	Tucker and others, in press
Lucerne pluton	ME	LU	380.0	4	U/Pb	zircon	No details given	Zartman and Gallego, 1979
Magaguadavic pluton	NB	MG	396.0	1	U/Pb	zircon	No details given	M.L. Bevier, written commun., 1996
Mattamiscontis pluton	ME	MA	406.9	3.6	U/Pb	zircon	Mean 207/206; 2 concordant zircon fractions	This paper
Mixer Pond granite gneiss	ME	MP	400.0	3	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions	Tucker and others, in press
Mooselook-meguntic two-mica leucogranite	ME	MO	370.3	1.1	U/Pb	monazite	Mean 207/206 age; 2 concordant fractions	Solar and others, 1998
Mooselookmegun-tic pluton, biotite granite	ME	MO	388.9	1.6	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Solar and others, 1998
Mount Waldo	ME	MW	371.0	1.9	U/Pb	zircon	Mean 207/206 age; 4 concordant fractions	Stewart and others, 1995b
Moxie pluton, central part	ME	MX	404.4	3.4	40/39	biotite	Plateau age	This paper
Moxie pluton, eastern part	ME	MX	406.3	3.8	40/39	biotite	Plateau age	This paper
Mt. Douglas	NB	MD	366.5	1	U/Pb	monazite	Mean of two U/Pb ages	M.L. Bevier, written commun., 1996
North Roxbury two-mica granite	ME	NR	404.3	1.9	U/Pb	zircon and monazite	Mean 207/206 age; 3 concordant fractions	Solar and others, 1998

North Searsmont granite gneiss	ME	NS	389.0	2	U/Pb	zircon	Mean 207/206 age; 4 concordant fractions	Tucker and others, in press
North Union granitic gneiss	ME	NU	422.0	2	U/Pb	zircon	Mean 207/206; 2 concordant zircon fractions	Tucker and others, in press
Northport pluton	ME	NP	419.0	2	U/Pb	zircon	No details given	D.B. Stewart, written communication, 1996. Date by Tucker.
Old Point pluton	ME	OP	379.0	3	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions	Tucker and others, in press
Onawa pluton, gabbro	ME	ON	401.1	3.7	40/39	biotite	Plateau age	This paper
Onawa pluton, granodiorite	ME	ON	405.1	2.9	U/Pb	zircon	Mean 207/206; 3 concordant zircon fractions	This paper
Pegmatite, Gardiner	ME	GA	367.0	1	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Tucker and others, in press
Pegmatite, Greeley Corner	ME	GR	371.0	1	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions	Tucker and others, in press
Phillips pluton	ME	PH	403.6	2.2	U/Pb	zircon and monazite	Mean 207/206 age; 3 concordant fractions	Solar and others, 1998
Phillips pluton schlieritic granite	ME	PH	405.3	1.8	U/Pb	zircon	Mean 207/206 age; 1 concordant fraction	Solar and others, 1998
Phillips pluton, leucogranite	ME	PH	403.5	1.6	U/Pb	zircon and monazite	Mean 207/206 age; 3 concordant fractions	Solar and others, 1998
Pleasant Lake pluton	ME	PL	401.3	4.7	40/39	biotite	Average of 3 plateau ages: 404±5, 397±9, 403±7	Hubacher and Lux, 1987
Pleasant Lake pluton	ME	PL	400.5	4.5	40/39	hornblende	Average of 2 plateau ages: 400±5 and 401±4.	Hubacher and Lux, 1987
Pocomoonshine gabbro-diorite	ME	PO	422.7	3.0	40/39	hornblende	Average of two plateau ages, 423.5±2.7 and 421.5±3.0. Dates intrusion of more northerly mafic phase.	West and others, 1992
Priestly pluton	ME	PR	361.7		40/39	biotite	Avg of two plateaus, 360.3±2.78 and 363.2±2.3	This paper
Priestly pluton	ME	PR	361.0	3	U/Pb	zircon	207/206 age; 1 concordant fraction	This paper
Redington biotite granite	ME	RD	407.6	4.7	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Solar and others, 1998
Rome pluton	ME	RO	378.0	1	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Tucker and others, in press
Russell Mtn. Pluton	ME	RU	406.0	1.3	U/Pb	zircon	No details given	This paper
S. Roxbury two-mica granite	ME	SR	408.2	2.5	U/Pb	zircon and monazite	Mean 207/206 age; 4 concordant fractions	Solar and others, 1998

Scotstown pluton	QU	SC	384.0	2	U/Pb	zircon	Mean 207/206 age of 2 slightly discordant fractions	Simonetti and Doig 1990
Sebec Lake pluton	ME	SL	407.8	2.5	U/Pb	zircon	Mean 207/206; 2 concordant zircon fractions	This paper
Seboeis granodiorite at East Branch Lake	ME	SE	364.7	4.3	40/39	biotite	Plateau	This paper
Seboeis pluton, granodiorite	ME	SS	405.8	2.5	U/Pb	zircon	Mean 207/206; 1 concordant zircon fraction	This paper
Sedgwick pluton	ME	SG	419.5	1.4	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Stewart and others, 1995b
Shirley-Blanchard composite pluton	ME	SB	406.9	1.4	U/Pb	zircon	Mean 207/206; 4 concordant zircon fractions	This paper
Skiff Lake pluton, Pokiok batholith	NB	SK	409.0	2	U/Pb	zircon	207/206 age of most concordant fraction.	Bevier and Whalen, 1990a
Somesville pluton	ME	SV	424.0	2	U/Pb	zircon	No details given	Seaman and others, 1995
Songo pluton	ME	SO	382.0	3	U/Pb	zircon and sphene	Mean 207/206 age; 2 slightly discordant fractions	Lux and Aleinikoff, 1985; Lux and others, 1989
South Penobscot pluton	ME	SP	419.2	2.2	U/Pb	zircon	Supercedes date in Stewart and others, 1995.	D.B. Stewart, written communication, 1996. Date by Tucker.
Spruce Head granite	ME	SH	421.0	1	U/Pb	zircon and sphene	Mean 207/206 age; 4 concordant fractions	Tucker and others, in press
Ste-Cecile pluton	QU	ST	374.0	1	U/Pb	zircon	Mean 207/206 age; 3 concordant or nearly concordant fractions	Simonetti and Doig 1990
Swift River two-mica leucogranite	ME	SW	407.9	1.9	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Solar and others, 1998
Thrasher Peaks intrusions	ME	TH	414.0		U/Pb	zircon	207/206; 1 concordant fraction	Moench and others, 1995
Threemile Pond pluton	ME	TP	381.0	1	U/Pb	zircon and sphene	Mean 207/206 age; 4 concordant fractions	Tucker and others, in press
Togus pluton	ME	TO	378.0	1	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions	Tucker and others, in press
Umbagog pluton	ME	UM	384.0	6	U/Pb	zircon	No details given	Aleinikoff and Moench, 1987
Utopia pluton	NB	UT	422.7	1	U/Pb	zircon	No details given	M.L. Bevier, written commun., 1996
Waldoboro binary granite	ME	WD	367.8	1.6	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Tucker and others, in press
Wamsutta pluton	NH	WM	408.0	1.9	U/Pb	zircon	No details given	J.D. Eusden, Jr., written communication, 1998.
Waterville granitic dike	ME	WA	399.4	0.7	U/Pb	zircon	Mean 207/206 age; 2 concordant fractions. Intrudes first-generation Acadian folds but is involved in second-generation folds.	Tucker and others, in press
Welsford pluton	NB	WL	422.0	1	U/Pb	zircon	No details given	M.L. Bevier, written commun., 1996
Winslow pluton	QU	WI	377.0	6	U/Pb	sphene	Mean 207/206 age; 2 nearly concordant fractions	Simonetti and Doig 1990
Youngtown pluton	ME	YT	420.0	2	U/Pb	zircon	Mean 207/206 age; 3 concordant fractions	Tucker and others, in press

Table 4. U-Pb isotope dilution analyses of igneous rocks in Maine.

FRACTIONS		CONCENTRATIONS					ATOMIC RATIOS				AGE [Ma]
No. in fig. 7	Properties (1)	Wt. [_g] (2)	Pb rad [ppm] (2)	U [ppm] (2)	Pb com [pg] (3)	Th/U (4)	$^{206}Pb/^{204}Pb$ (5)	$^{207}Pb/^{206}Pb$ (6)	$^{207}Pb/^{235}U$ (6)	$^{206}Pb/^{238}U$ (6)	$^{207}Pb/^{206}Pb$ (6)
Bald Mountain [95MDw140] *Complete analytical data not available as of Dec. 7, 1998. Contact dbradly@usgs.gov for more information.*											
†19	20 gr,-200,cl,c,p	32	24.9	401	2.9	0.249	18,028	0.05555± 3	0.4878 ±6	0.06369 ±7	434.3 ± 1.1
†20	6 gr,-200,cl,c,t-p	18	37.9	385	3.5	0.223	11,979	0.07231 ± 8	0.9540 ±22	0.09570 ±24	994.6 ± 2.3
21											
22											
Blanchard [94MDw98]											
21	4 gr,-200,cl,pb,s-p	14	431	6630	7.12	0.306	57,749	0.05488 ±4	0.4958 ±14	0.06553 ±19	407.2 ± 1.4
22	10 gr,-200,cl,cr,sk	29	62.3	879	2.15	0.641	48,679	0.05487 ±3	0.4940 ±37	0.06530 ±49	406.9 ± 1.0
23	7 gr,-200,cl,c,n	11	32.1	509	2.81	0.331	7,653	0.05490 ±5	0.4769 ±7	0.06301 ±9	408.3 ± 2.0
Center Pond [95MDw157]											
33	2 gr,-200,cl,c,t-p	5	31.6	511	1.4	0.431	7,219	0.05423 ±5	0.4497 ±8	0.06015 ±10	380.7 ± 2.2
34	21 gr,-200,cl,c,t-p	40	27.1	434	5.7	0.485	11,804	0.05411 ±3	0.4473 ±7	0.05995± 9	375.8 ± 1.2
Debouille [95MDw115]											
†36	10 gr,-200,cl,c,s-p	28	33.2	451	3.7	1.146	13,092	0.05463 ±4	0.4515 ±8	0.05993 ±11	397.3 ± 1.5
Ebeemee [95MDw167]. *Analytical data not available as of Dec. 7, 1998. Contact dbradly@usgs.gov for more information.*											
6											
7											
Hammond Sill [95MDw184]											
35	2 gr,-200,cl,pb,p	3	157	2710	10.8	0.243	2,918	0.05414 ±8	0.4444 ±9	0.05954 ±13	376.8 ± 3.2
36	2 gr,-200,cl,pb,p	3	238	4065	21.7	0.257	2,200	0.05414 ±4	0.4483 ±6	0.06005 ±7	377.0 ± 1.8
37	4 gr,-200,cl,pb,fp	6	139	2406	57.1	0.236	913.3	0.05413 ±10	0.4462 ±10	0.05979 ±9	376.4 ± 4.0
38	2 gr,-200,cl,pb,p	3	138	2340	12.2	0.277	2,239	0.05419 ± 5	0.4479 ±7	0.05995 ±7	378.9 ± 2.3
39	2 gr,-200,cl,pb,p	2	224	3800	17.9	0.284	1,657	0.05412 ±9	0.4469 ±10	0.05990 ±11	376.1 ± 3.7
40	2 gr,-200,cl,pb,p	3	141	2375	19.6	0.333	1,416	0.05405 ±13	0.4429 ±18	0.05943 ±20	373.0 ± 5.5
Harrington [94MDw48] *Analytical data not available as of Dec. 7, 1998. Contact dbradly@usgs.gov for more information.*											
27											
28											
29											
Hog Island [95MDw205A]											
41	5 gr,-200,cl,c,lp	14	36.2	567	13.7	0.619	2,142	0.05399± 7	0.4409 ±32	0.5923 ±43	370.6 ± 3.0
Mattamiscontis [95MDw164A]											
1	10 gr,-200,cl,c,t-p	17	15.4	236	2.2	0.355	7,530	0.05488 ±6	0.4925 ±8	0.06509 ±9	407.4 ± 2.3
2	8 gr,-200,cl,c,t-p	10	18.8	285	3.4	0.366	3,677	0.05485 ±8	0.4935 ±13	0.06525 ±13	406.3 ± 3.3
3	2 gr,-200,cl,c,t-p	6	19.2	292	13.1	0.367	557.6	0.05478 ±21	0.4926 ±20	0.06523 ±9	403.1 ± 8.6

Onawa [95MDw68]										
8	4 gr,-200,cl,c,cr,fr	13	28.7	400	2.2	0.723	9,468	0.05481 ±6	0.4885 ±8	0.06464±10	404.4 ± 2.4
9	10 gr,-200,cl,cr,c,fr	21	28.7	396	5.3	0.708	6,433	0.05484 ±5	0.4905 ±9	0.06487 ±12	405.6 ± 2.0
10	1 gr,-200,cl,cr,fr	3	26.2	364	2.3	0.771	2,180	0.05485 ±14	0.4858 ±13	0.06424 ±19	405.9 ± 5.7
Priestly Lake [95MDw108]										
42	6 gr,-200,cl,c,p	14	18.2	289	5.6	0.646	2,716	0.05375 ±8	0.4287 ±8	0.05785 ±7	360.5 ± 3.3
Russell Mountain [95MDw98]										
13	20 gr,-200,cl,c,cr,n	43	36.4	578	6.9	0.228	15,074	0.05485 ±2	0.4905 ±7	0.06486 ±10	406.2 ± 0.9
14	3 gr,-200,cl,c,cr,n	5	30.3	469	3.0	0.319	3,013	0.05481 ±10	0.4899 ±13	0.06482 ±11	404.6 ± 4.2
15	15 gr,-200,cl,cr,p	30	54.1	770	4.9	0.723	19,181	0.05484 ±3	0.4788 ±6	0.06333 ±9	405.7 ± 1.1
†16	22 gr,-200,cl,cr,c,n	34	35.7	572	10.7	0.242	7,417	0.05510 ±9	0.4873±18	0.06414 ±25	416.2 ± 3.5
Sebec Lake [95MDw71]										
11	20 gr,-200,cl,c,n	26	29.5	473	12.2	0.219	4,143	0.05487 ±4	0.4882 ±7	0.06453 ±9	406.9 ± 1.7
12	8 gr,-200,cl,c,n	12	23.0	367	6.5	0.220	2,854	0.05494 ±6	0.4904 ±9	0.06474 ±10	409.7 ± 2.6
Seboeis [95MDw132]										
4	8 gr,-200,cl,cr,pb,n	13	246	3800	71.1	0.340	2,802	0.05481 ±9	0.4874 ±17	0.06450 ±20	404.3 ± 3.5
5	2 gr,-200,cl,cr,pb,n	5	177	2800	6.8	0.289	8,458	0.05485 ±3	0.4842 ±11	0.06403 ±14	406.0 ± 1.4
Shirley [95MDw90B]										
†17	6 gr,-200 cl,c,n	11	36.4	562	4.0	0.326	6,511	0.05513 ±4	0.4941 ±7	0.06499 ±8	417.6 ± 1.7
†18	4 gr,+200,cl,c,a,fr	42	32.6	473	4.7	0.258	18,777	0.05776 ±4	0.5581 ±9	0.07008 ±12	520.6 ± 1.4

Notes:

(1) Cardinal number indicates the number of zircon grains analyzed (*e.g.* 35 grains); all grains were selected from non-paramagnetic separates at 0^{o} tilt at full magnetic field in Frantz Magnetic Separator; +200 = size in mesh (> 75µm); c = colorless; cl= clear; el = elongate; e = equant; eu = euhedral; f = faceted; lp = long prismatic; n = 5:1 prismatic needles; p = prismatic; pb = pale brown; s = stubby; s-p = short-prismatic; t = tips from prisms. NA = non-abraded fraction; all other grains were air-abraded following Krogh (1982). † = analyses rejected from regression analysis because of inheritance.

(2) Concentrations are known to ± 30% for sample weights of about 20 µg and ± 50% for samples 5 µg.

(3) Corrected for 0.0215 mole fraction common-Pb in the ^{205}Pb-^{235}U spike.

(4) Calculated Th/U ratio assuming that all ^{208}Pb in excess of blank, common-Pb, and spike is radiogenic (λ $^{232}Th = 4.9475 \times 10^{-11} y^{-1}$).

(5) Measured, uncorrected ratio.

(6) 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). For sample D-01 (†) this column lists the $^{206}Pb/^{238}U$ age (± 1σ absolute) which is considered more accurate. U and Pb half-lives and isotopic abundance ratios from Jaffey and others (1971).

Table 5. $^{40}Ar/^{39}Ar$ data for igneous rocks from Maine.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	moles ^{39}Ar*	% total ^{39}Ar	% 40Ar rad	K/Ca	apparent age Ma	Error
Bald Mtn. pluton, granodiorite		95MDw143			biotite			J=.007489	
760	32.92	0.017	0.0068	745.4	7.2	93.8	28.3	375.4	3.7
900	32.78	0.009	0.0019	1746.9	16.8	98.3	54.0	389.8	3.6
1010	32.98	0.013	0.0013	1303.1	12.6	98.8	38.7	393.8	3.9
1090	32.99	0.018	0.0014	1899.3	18.3	98.7	28.0	393.6	3.6
1130	32.94	0.024	0.0006	1967.7	19.0	99.4	20.0	395.5	4.6
1230	32.98	0.026	0.0012	1338.0	12.9	98.9	18.9	394.3	3.6
1310	33.09	0.045	0.0008	803.1	7.7	99.2	10.9	396.6	3.9
FUSE	33.26	0.129	0.0021	564.9	5.4	98.1	3.8	394.5	4.1
TOTAL			10368.3	100.0				392.4	3.9
PLATEAU AGE								394.3	3.7
Bald Mtn. pluton, granodiorite		95MDw146			biotite			J=.007697	
760	33.32	0.028	0.0295	220.8	1.5	73.8	17.6	312.8	9.7
900	31.72	0.006	0.0025	3063.0	21.4	97.6	86.4	385.7	3.5
1010	31.91	0.008	0.0022	2536.4	17.7	97.9	58.1	388.8	3.7
1090	31.90	0.009	0.0015	1948.3	13.6	98.5	55.8	390.9	3.6
1130	31.72	0.009	0.0009	2259.6	15.8	99.1	56.9	390.8	3.5
1230	31.61	0.013	0.0007	2345.5	16.4	99.3	38.5	390.3	3.5
1310	31.70	0.019	0.0006	1299.2	9.1	99.3	25.4	391.6	3.6
FUSE	31.85	2.337	0.0012	668.7	4.7	99.4	0.2	394.0	3.7
TOTAL			14341.5	100.0				388.3	3.7
PLATEAU AGE								390.9	3.0
Beaver Cove pluton, granodiorite		95MDw195			biotite			J=.007539	
760	59.90	0.122	0.1505	67.5	0.7	25.8	4.0	198.6	24.8
900	31.77	0.018	0.0086	907.6	9.2	91.9	27.0	359.0	3.6
1010	31.21	0.015	0.0043	1486.4	15.0	95.9	33.3	367.1	3.4
1090	31.26	0.018	0.0036	1894.4	19.1	96.5	27.2	369.7	3.4
1130	31.22	0.035	0.0027	1502.4	15.2	97.4	14.2	372.5	3.5
1230	30.90	0.074	0.0016	1793.3	18.1	98.4	6.7	372.5	3.4
1210	30.82	0.080	0.0012	1599.4	16.1	98.8	6.2	372.8	3.5
FUSE	30.89	0.137	0.0017	664.3	6.7	98.3	3.6	371.9	3.5
TOTAL			9915.1	100.0				368.8	3.6
PLATEAU AGE								372.4	2.7
Center Pond pluton, granodiorite		95MDw158A			biotite			J=.00787	
770	34.084	0.01791	0.04073	10.0	2.9	64.6	27.4	288.5	13.1
870	30.933	0.00298	0.01092	26.2	7.6	89.5	164.3	355.7	3.5
950	29.676	0.00541	0.00304	67.3	19.5	96.9	90.6	368.1	3.3
1010	29.466	0.00603	0.00209	70.0	20.3	97.9	81.3	368.9	3.3
1085	29.484	0.01491	0.00208	46.4	13.4	97.9	32.9	369.2	3.4
1145	29.548	0.02216	0.00209	37.0	10.7	97.9	22.1	369.9	3.4
1230	29.429	0.04037	0.00115	61.3	17.7	98.8	12.1	371.8	3.4
FUSE	29.385	0.03921	0.00139	27.4	7.9	98.6	12.5	370.4	3.3
TOTAL			345.6	100.0				366.2	3.6
PLATEAU AGE								368.7	2.9

Chandler pluton, granodiorite		95MDw106			biotite			J=.007527	
760	29.22	0.106	0.0524	67.0	0.9	47.0	4.6	177.3	20.8
900	31.35	0.019	0.0085	404.1	5.3	91.9	25.2	354.2	3.9
1010	31.24	0.019	0.0052	547.3	7.2	95.0	25.4	363.7	3.8
1090	31.33	0.023	0.0030	677.5	8.9	97.1	21.0	371.9	3.6
1130	30.94	0.024	0.0017	1034.6	13.7	98.4	20.2	372.1	3.5
1230	30.50	0.018	0.0009	1659.7	21.9	99.0	26.9	369.6	3.5
1310	30.26	0.013	0.0006	2066.3	27.3	99.3	37.8	367.9	3.4
FUSE	30.45	0.019	0.0013	1117.5	14.8	98.7	25.2	368.0	3.8
TOTAL			7573.9	100.0		366.5		3.7	
Deboullie pluton, monzonite		95MDw112			biotite			J=.007521	
760	34.57	0.021	0.0201	216.1	2.2	82.7	22.9	351.5	6.3
900	30.90	0.006	0.0041	677.8	6.9	96.0	77.5	363.3	3.5
1010	30.50	0.004	0.0018	677.5	6.9	98.2	132.6	366.5	3.7
1090	30.45	0.006	0.0018	784.5	8.0	98.2	82.5	366.0	3.5
1130	30.51	0.005	0.0012	686.7	7.0	98.8	106.7	368.6	3.7
1230	30.18	0.004	0.0008	1650.2	16.8	99.1	127.8	366.2	3.4
1310	30.05	0.005	0.0007	2874.0	29.3	99.3	100.7	365.1	3.4
FUSE	30.16	0.008	0.0011	2235.7	22.8	98.9	62.9	365.0	3.4
TOTAL			9802.3	100.0				365.3	3.5
PLATEAU AGE								365.4	3.2
Deboullie pluton, monzonite		95MDw114			biotite			J=.007917	
760	94.97	0.015	0.2690	33.3	0.3	16.3	32.7	208.5	28.3
900	32.73	0.004	0.0174	564.7	4.9	84.2	133.8	356.1	3.7
1010	29.57	0.001	0.0045	1027.9	8.9	95.4	692.6	363.9	3.4
1090	29.06	0.004	0.0019	915.2	8.0	98.0	129.2	367.0	3.6
1130	28.74	0.004	0.0010	1170.6	10.2	98.9	112.5	366.3	3.4
1230	28.38	0.003	0.0005	2513.7	21.9	99.4	145.9	363.6	3.4
1310	28.25	0.004	0.0004	3387.2	29.5	99.5	126.6	362.5	3.3
FUSE	28.37	0.008	0.0005	1883.9	16.4	99.4	61.6	363.5	3.3
TOTAL			11496.4	100.0				363.0	3.5
PLATEAU AGE								363.2	3.0
Deboullie pluton, granodiorite		95MDw115			biotite			J=.007839	
760	35.11	0.086	0.0312	131.0	3.2	73.7	5.7	333.2	12.9
900	30.86	0.045	0.0094	368.3	9.1	91.0	11.0	359.0	5.6
1010	30.18	0.052	0.0050	437.5	10.8	95.1	9.4	366.1	5.7
1090	29.93	0.061	0.0032	352.8	8.7	96.8	8.0	369.1	6.4
1130	29.47	0.068	0.0019	454.2	11.2	98.1	7.2	368.4	3.8
1230	28.80	0.052	0.0007	1076.7	26.6	99.3	9.5	364.8	3.4
1310	28.75	0.047	0.0007	927.4	22.9	99.3	10.5	364.2	3.4
FUSE	29.02	0.071	0.0018	301.5	7.4	98.1	6.9	363.5	5.0
TOTAL			4049.4	100.0				363.9	4.6
PLATEAU AGE								364.2	3.2

Ebeemee pluton, granodiorite		95MDw169			biotite			J=.007458	
760	31.52	0.055	0.0131	35.7	6.7	87.7	9.0	338.0	6.3
900	33.59	0.020	0.0046	74.7	14.1	95.9	24.4	388.4	4.0
1010	34.40	0.016	0.0033	82.5	15.5	97.1	30.7	401.4	4.1
1090	34.65	0.015	0.0022	89.4	16.8	98.1	32.7	407.6	3.7
1130	34.49	0.011	0.0015	93.3	17.6	98.6	43.7	408.0	3.7
1230	34.47	0.031	0.0015	96.3	18.1	98.7	15.9	407.8	4.1
1310	35.10	0.069	0.0049	31.2	5.9	95.9	7.1	404.1	4.1
FUSE	35.98	0.053	0.0074	27.9	5.2	93.9	9.3	405.3	4.2
TOTAL			531.0	100.0				399.0	4.1
PLATEAU AGE								407.8	2.4
Hog Island pluton, granodiorite		95MDw205			biotite			J=.00792	
760	30.04	0.027	0.00947	27.6	4.2	90.6	17.9	352.2	3.2
900	28.98	0.018	0.00292	75.9	11.6	97.0	28.0	362.6	3.3
1010	29.13	0.014	0.00251	70.2	10.7	97.4	34.5	365.7	3.3
1090	29.19	0.015	0.00186	76.3	11.6	98.1	31.7	368.7	3.3
1130	29.06	0.021	0.00114	96.6	14.7	98.8	23.2	369.6	3.3
1230	28.83	0.027	0.00068	144.7	22.1	99.3	18.3	368.5	3.3
1310	28.86	0.046	0.00101	105.6	16.1	98.9	10.7	367.8	3.4
FUSE	28.94	0.059	0.00103	58.8	9.0	98.9	8.3	368.6	3.3
TOTAL			655.7	100.0				366.9	6.7
PLATEAU AGE								368.6	3.1
Mattamiscontis pluton, granodiorite		95MDw164B			biotite			J=.007498	
880	34.89	0.029	0.0172	374.6	8.5	85.4	16.6	363.6	3.9
1020	34.12	0.028	0.0046	793.9	18.0	96.0	17.5	396.2	3.6
1110	34.05	0.038	0.0023	1187.4	27.0	97.9	13.1	402.5	3.9
1180	33.96	0.043	0.0023	915.5	20.8	98.0	11.5	401.8	3.7
1310	33.85	0.098	0.0019	741.9	16.9	98.3	5.0	402.0	3.8
FUSE	33.85	0.124	0.0035	388.4	8.8	96.9	3.9	396.7	4.4
TOTAL			4401.8	100.0				397.3	3.8
PLATEAU AGE								402.1	2.8
Moxie pluton, gabbro		95MDw187			biotite			J=.007449	
760	40.39	0.033	0.0269	168.3	1.8	80.3	15.0	390.3	10.9
900	35.91	0.023	0.0076	533.9	5.6	93.7	21.4	403.4	4.2
1010	35.14	0.024	0.0038	2118.6	22.3	96.7	20.7	407.2	3.8
1090	34.82	0.023	0.0018	960.6	10.1	98.4	21.7	410.2	4.0
1130	34.43	0.018	0.0013	1410.1	14.9	98.8	26.6	407.6	3.7
1230	34.08	0.011	0.0007	1942.7	20.5	99.3	43.6	405.8	3.7
1310	34.10	0.019	0.0007	1658.5	17.5	99.3	26.1	405.8	3.7
FUSE	34.30	0.027	0.0014	694.2	7.3	98.8	17.9	405.9	4.0
TOTAL			9487.0	100.0				406.4	3.9
PLATEAU AGE								406.3	3.8

Moxie pluton, gabbro		95MDw198			biotite			J=.007962	
760	88.47	0.229	0.23360	2.4	0.6	22.0	2.1	259.7	41.4
900	37.51	0.085	0.02270	20.4	5.3	82.1	5.8	395.6	3.9
1010	34.38	0.086	0.00941	34.6	9.0	91.9	5.7	404.8	3.6
1090	33.25	0.083	0.00474	47.2	12.3	95.8	5.9	407.7	3.7
1130	32.47	0.059	0.00284	67.1	17.4	97.4	8.4	405.2	3.6
1230	32.00	0.038	0.00140	104.1	27.1	98.7	13.0	404.5	3.6
1310	31.88	0.038	0.00103	76.4	19.8	99.0	12.9	404.5	3.6
FUSE	32.07	0.048	0.00194	32.6	8.5	98.2	10.2	403.5	3.7
TOTAL			384.7	100.0				403.6	3.9
PLATEAU AGE								404.4	3.4
Onawa pluton, gabbro phase		95MDw68B			biotite			J=.007633	
760	40.59	0.048	0.0586	19.9	2.3	57.3	10.1	294.7	6.4
900	33.78	0.020	0.0073	126.0	14.5	93.5	24.6	389.9	4.0
1010	33.43	0.016	0.0041	176.3	20.2	96.4	31.3	396.6	3.6
1090	33.57	0.020	0.0026	104.0	11.9	97.7	24.9	402.9	3.7
1130	33.15	0.026	0.0014	144.1	16.5	98.7	19.0	402.3	3.7
1230	32.69	0.015	0.0005	97.2	11.1	99.5	33.6	400.2	3.6
1310	32.80	0.013	0.0006	115.2	13.2	99.4	37.7	401.0	3.9
FUSE	32.87	0.038	0.0014	89.4	10.2	98.7	12.8	399.2	3.9
TOTAL			872.1	100.0				396.2	3.8
PLATEAU AGE								400.1	3.7
Priestly pluton, granodiorite		95MDw107			biotite			J=.00786	
760	31.01	0.010	0.0159	356.9	3.4	84.8	49.2	338.9	4.5
900	29.47	0.013	0.0061	756.1	7.2	93.8	37.7	354.7	3.5
1010	29.33	0.022	0.0043	938.9	9.0	95.6	22.7	359.3	3.5
1090	29.18	0.051	0.0033	1102.0	10.5	96.6	9.6	361.2	3.6
1130	28.88	0.060	0.0020	1497.7	14.3	97.9	8.1	362.0	3.4
1230	28.47	0.038	0.0012	2528.4	24.2	98.7	12.9	360.1	3.3
1310	28.42	0.029	0.0010	2166.4	20.7	98.9	16.9	360.0	3.3
FUSE	28.59	0.032	0.0014	1118.9	10.7	98.5	15.4	360.8	3.4
TOTAL			10465.4	100.0				359.3	3.4
PLATEAU AGE								360.3	2.7
Priestly pluton, granodiorite		95MDw108			biotite			J=.007509	
760	50.26	0.111	0.0911	4.2	1.1	46.4	4.4	291.2	26.3
900	31.69	0.017	0.0087	26.7	7.0	91.8	28.9	356.4	4.1
1010	30.84	0.018	0.0044	38.0	9.9	95.8	28.0	361.3	3.8
1090	30.75	0.029	0.0034	38.6	10.0	96.7	17.1	363.5	3.9
1130	30.70	0.083	0.0024	43.4	11.3	97.7	5.9	366.4	3.6
1230	30.18	0.073	0.0013	74.1	19.3	98.7	6.7	364.2	3.4
1310	29.92	0.052	0.0008	124.6	32.5	99.2	9.3	363.0	3.3
FUSE	30.45	0.059	0.0025	34.4	9.0	97.6	8.3	363.3	5.8
TOTAL			384.0	100.0				362.3	4.0
PLATEAU AGE								363.2	2.3

Russell Mtn. pluton		95MDw80			biotite			J=.00755	
760	31.44	0.014	0.0079	639.2	5.0	92.6	34.7	358.3	3.6
900	32.11	0.005	0.0021	1722.9	13.5	98.0	98.7	384.4	3.5
1010	32.40	0.006	0.0019	1649.9	13.0	98.2	85.6	388.3	3.5
1090	32.71	0.004	0.0016	1721.9	13.5	98.5	116.0	392.8	3.6
1130	32.62	0.005	0.0010	2327.8	18.3	99.1	97.3	393.8	3.5
1230	32.44	0.006	0.0006	2508.1	19.7	99.3	81.1	392.9	3.6
1310	32.53	0.008	0.0007	1530.2	12.0	99.3	65.1	393.9	3.6
FUSE	32.66	0.011	0.0012	619.5	4.9	98.8	44.0	393.4	3.8
TOTAL			12719.5	100.0				389.7	3.6
PLATEAU AGE								393.4	3.0
Russell Mtn. pluton		95MDw98			muscovite			J=.007668	
770	65.509	0.04333	0.12239	3.7	1.1	44.8	11.3	366.0	22.8
870	38.853	0.01887	0.02433	9.3	2.7	81.5	26.0	392.0	6.3
950	33.449	0.00837	0.00506	19.2	5.5	95.5	58.6	395.2	3.8
1010	32.466	0.01020	0.00212	50.8	14.5	98.0	48.0	393.9	3.6
1085	32.277	0.00489	0.00162	70.3	20.1	98.5	100.2	393.5	3.6
1150	32.107	0.00234	0.00096	53.5	15.3	99.1	209.2	393.7	3.5
1230	32.111	0.00589	0.00096	83.8	24.0	99.1	83.2	393.8	3.5
FUSE	32.472	0.03211	0.00241	58.6	16.8	97.8	15.3	393.1	3.8
TOTAL			349.3	100.0				393.4	3.9
PLATEAU AGE								393.6	2.7
Sebec Lake pluton, granodiorite		95MDw71			muscovite			J=.007477	
850	35.78	0.020	0.02633	14.4	5.1	78.2	24.1	342.8	4.5
950	32.19	0.031	0.00872	20.6	7.2	92.0	16.0	360.7	11.5
1010	32.41	0.017	0.00436	34.4	12.0	96.0	28.3	377.2	3.6
1090	32.31	0.021	0.00356	42.6	14.9	96.7	24.4	378.8	3.4
1130	32.04	0.016	0.00317	39.4	13.8	97.0	30.9	377.1	3.4
1210	31.77	0.017	0.00266	43.6	15.3	97.5	29.5	375.8	3.4
1270	31.89	0.013	0.00167	45.2	15.9	98.4	38.9	380.3	3.5
FUSE	31.42	0.019	0.00181	45.1	15.8	98.2	25.7	374.7	3.4
TOTAL			285.4	100.0				374.4	4.1
PLATEAU AGE								377.2	4.4
Sebec Lake pluton, granodiorite		95MDw72			biotite			J=.007885	
750	15.616	0.00261	0.00658	88.8	14.4	87.4	187.4	184.5	1.9
900	31.175	0.00810	0.00360	77.1	12.5	96.5	60.5	384.2	17.5
1090	31.219	0.00277	0.00122	147.4	23.8	98.8	176.6	392.7	3.5
1130	31.093	0.00313	0.00090	154.2	24.9	99.1	156.3	392.4	3.5
1230	31.269	0.00634	0.00046	79.5	12.9	99.5	77.3	395.9	3.6
1300	31.153	0.01663	0.00077	69.3	11.2	99.2	29.5	393.5	3.5
FUSE	31.031	0.08028	0.01556	2.1	0.3	85.1	6.1	341.4	17.5
TOTAL			618.4	100.0				362.0	5.1
PLATEAU AGE								392.5	2.5

Seboeis pluton, diorite		95MDw135A			biotite			J=.007937	
770	39.935	0.07719	0.07671	6.4	1.4	43.2	6.3	231.6	11.5
870	33.444	0.02203	0.01627	26.3	5.8	85.6	22.2	369.3	4.2
950	32.429	0.01091	0.00642	47.2	10.5	94.1	44.9	391.3	3.5
1010	32.378	0.01232	0.00461	56.9	12.7	95.7	39.8	396.9	3.6
1085	32.314	0.01106	0.00221	73.2	16.3	97.9	44.3	404.3	3.6
1145	32.100	0.00865	0.00150	79.1	17.6	98.6	56.6	404.2	3.6
1230	31.710	0.00868	0.00105	101.8	22.6	99.0	56.5	401.3	3.6
FUSE	31.956	0.00836	0.00204	58.5	13.0	98.1	58.6	400.7	4.3
TOTAL			449.5	100.0				396.3	3.8
Seboeis pluton, granite		95MDw135B			muscovite			J=.007465	
760	46.10	0.030	0.04967	7.7	1.5	68.1	16.4	380.0	9.3
900	35.11	0.008	0.00680	39.4	7.7	94.2	61.0	398.2	4.2
1010	34.19	0.003	0.00263	95.5	18.8	97.7	183.5	401.6	3.6
1090	34.02	0.007	0.00220	88.3	17.3	98.0	69.4	401.1	3.6
1130	33.87	0.003	0.00143	95.0	18.7	98.7	162.4	401.9	3.6
1230	33.80	0.004	0.00156	94.8	18.6	98.6	135.0	400.8	3.6
1310	33.84	0.006	0.00153	57.8	11.3	98.6	137.9	401.3	3.6
FUSE	33.44	0.013	0.00315	31.0	6.1	97.2	36.9	391.8	4.5
TOTAL			509.4	100.0				400.2	6.7
PLATEAU AGE								401.3	2.8
Seboeis pluton, granodiorite		95MDw139B			biotite			J=.007437	
760	47.15	0.068	0.0896	3.5	0.6	43.8	7.2	258.0	26.8
900	32.15	0.015	0.0086	45.0	8.0	92.0	33.6	358.8	4.5
1010	31.29	0.015	0.0044	73.8	13.2	95.8	33.2	363.0	3.7
1090	31.03	0.012	0.0026	76.5	13.7	97.5	40.7	366.2	3.4
1130	30.77	0.010	0.0019	94.4	16.9	98.2	50.0	365.6	3.7
1230	30.49	0.008	0.0013	115.4	20.6	98.7	58.4	364.2	3.4
1310	30.40	0.009	0.0013	106.3	19.0	98.7	56.2	363.2	4.0
FUSE	30.86	0.015	0.0021	44.8	8.0	98.0	33.4	365.9	3.7
TOTAL			559.8	100.0				363.4	3.9
PLATEAU AGE								364.7	4.3
Shirley pluton		95MDw90B			biotite			J=7.888001E-03	
720	40.59	0.020	0.0529	7.5	0.9	61.4	24.8	323.9	15.9
900	31.36	0.005	0.0038	107.4	12.5	96.4	103.0	385.7	3.6
1010	31.15	0.007	0.0019	135.5	15.8	98.2	74.9	389.9	3.6
1090	31.13	0.008	0.0015	129.7	15.1	98.5	63.0	391.0	3.5
1130	30.86	0.007	0.0009	178.2	20.8	99.1	75.2	389.8	3.5
1230	30.91	0.011	0.0008	167.2	19.5	99.2	45.3	390.6	3.6
1310	31.11	0.028	0.0009	83.1	9.7	99.1	17.3	392.8	3.7
FUSE	31.38	0.036	0.0018	49.0	5.7	98.3	13.8	392.8	4.1
TOTAL			857.6	100.0				389.5	3.7
PLATEAU AGE								390.3	3.0
moles $^{39}Ar^{*}=4.5x10^{-14}$									

Table 6. Comparison between time scales

Stage or series boundary	Harland and others, 1990	Tucker and others, 1998
Base of Tournaisian (= base of Carboniferous)	362.5	362
Base of Famennian	367.0	376.5
Base of Frasnian	377.4	382.5
Base of Givetian	380.8	387.5
Base of Eifelian	386.0	394
Base of Emsian	390.4	409.5
Base of Pragian	396.3	413.5
Base of Lochkovian (= base of Devonian)	408.5	418
Base of Pridolian	410.7	419
Base of Ludlovian	424.0	424

Table 7, Data used to estimate Acadian and younger shortening in figure 13.

Map letter in fig. 13.	Azimuth of cross section	Location	Stretch	1/stretch (plotted in fig. 13)	Final width (km)	Initial width (km)	Notes	Reference
A	130	Spencer Stream 15' quadrangle, Maine	0.75	1.33	6.5	8.6	Folded contact at base of Seboomook Group.	Boucot and Heath, 1969, Plate 14, part of section U-U'.
B	133	Long Pond and Brassua Lake 15' quadrangles, Maine	0.84	1.19	18.4	21.9	Folded contact between Seboomook Group and overlying Tarratine Formation.	Boucot and Heath, 1969, Plate 14, part of section L-L'.
C	143	Brassua Lake 15' quadrangle, Maine	0.68	1.47	6.8	10	Folded contact between Tarratine Formation and overlying Tomhegan Formation (including Kineo Rhyolite).	Boucot and Heath, 1969, Plate 14, part of section F-F'.
D	90	Caucomgomoc Lake 15' quadrangle, Maine	0.55	1.82	9.6	17.5	Folded contact between Northeast Carry Formation and overlying strata of the undivided part of the Seboomook Group.	Pollock, 1985, part of section D-D'.
E	100	Lobster anticline and Roach River Syncline, Maine	0.77	1.30	18.9	24.6	Folded contact at base of Silurian rocks.	Boucot and Heath, 1969, Plate 14, part of section D1-D1'.
F	129	Munsungun Anticlinorium	0.51	1.96	11.2	21.8	Folded base of Spider Lake Volcanics.	Hall, 1970, part of section A-A'.
G	140	Phillips 15' quadrangle, Maine	0.47	2.13	5.3	11.3	Folded contact between Smalls Falls and Madrid Carrabassett Formations.	Moench, 1971, section B-B'.
H	140	Kingsbury 15' quadrangle, Maine	0.38	2.63	5.4	14.1	Folded contact between Madrid and Carrabassett Formations. Ignoring parasitic folds, stretch is 0.49 and initial width is 10.9 km.	Ludman, 1978, part of section A-A'.
I	138	Shin Pond 15' quadrangle, Maine	0.77	1.30	13.8	17.8	Folded contact between unnamed Silurian strata and Seboomook Group.	Neumann, 1967, part of section A-A'.
J	145	Island Falls 15' quadrangle, Maine	0.34	2.94	5.6	16.5	Folded contact between Mattawamkeag Formation and overlying Allsbury (to the west) and "Rocks of Island Falls" (to the east).	Eckren and Frischknecht, 1967, part of section A-A'
K	150	Maple & Hovey Mtn. area, Maine	0.58	1.72	0.9	1.6	Folded contact of base of manganiferous beds. Constrained by exploratory drilling.	Pavlides, 1962, section A-A'.
L	90	Presque Isle 15' quadrangle, Maine	0.71	1.41	8.7	12.25	Weighted average of measurements across Chapman Syncline (base of unnamed Silurian limestone) and Dudley Syncline (base of Perham Fm.). Post-Mapleton stretch is 0.86.	Boucot and others, 1964, parts of section A-B.