**Dating slate belts using $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and zircon ages from
crosscutting plutons: A case study from east-central Maine, USA**

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Abstract

We report the ages of cleavage development in a normally intractable lower greenschist facies slate belt, the Central Maine-Aroostook-Matapedia belt in east-central Maine. We have attacked this problem by identifying the minimum ages of muscovite in a regional Acadian cleavage (S_1) and in a local ductile fault zone cleavage (S_2) using $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and the ages of crosscutting plutons. Our success stems from the regional low-grade metamorphism of the rocks in which each crystallization event preserves a $^{40}\text{Ar}/^{39}\text{Ar}$ crystallization age and not a cooling age. Evidence for recrystallization via a pressure solution mechanism comes from truncations of detrital, authigenic, and in some rocks S_1 muscovite and chlorite grains by new cleavage-forming muscovite and chlorite grains. Low-blank furnace age spectra from meta-arkosic and slaty rocks climb from moderate temperature Devonian age-steps dominated by cleavage-forming muscovite to Ordovician age-steps dominated by a detrital muscovite component. S_1 - and S_2 -cleaved rocks were hornfelsed by granitoids of ~ 407 and 377 Ma, respectively. The combination of these minimum ages with the maximum metamorphic crystallization ages establishes narrow constraints on the timing of these two cleavage-forming events, ~ 410 Ma (S_1) and ~ 380 Ma (S_2). These two events coincide in time with a change in the plate convergence kinematics from the arrival of the Avalon terrane (Acadian orogeny), to a right-lateral transpression arrival of the Meguma terrane in the Neoacadian orogeny.

1. Introduction

Low-grade metasedimentary slate belts form important parts of many orogenic belts (Robinson and Merriman, 1998; Poblet and Lisle, 2011), but their lack of fossils and geochronometers stymies determination of the time of their deposition and

deformation events. Thus, tectonic interpretations must rely on often-uncertain regional correlations and relative timing constraints (e.g. Freeman et al. 1997). Low-grade slate belts also lack reliable geothermometers and geobarometers (Robinson and Merriman, 1998), and their dating is still subject to many problems (e.g., Cliff, 1985), which obstructs assessing their thicknesses and maximum depths of burial. As a result, questions of the age of deposition, the ages of cleavages, and times and rates of exhumation are commonly left unanswered. However, $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology has successfully constrained ages of deformation and fabric development in lower greenschist facies rocks when combined with detailed textural and structural analyses (e.g., Kunk, et al., 2005; Castonguay et al., 2007; Beltrando et al., 2009; Wintsch et al., 2010; Growdon et al., 2013, McWilliams et al., 2013).

The low-grade metasedimentary rocks that underlie most of northeastern Maine are a classic example of this problem. The early Silurian to Early or Middle Devonian metasedimentary rocks of the Central Maine-Aroostook-Matapedia slate belt (Fig. 1) were folded and cleaved prior to the intrusion of middle Paleozoic plutons (Bradley et al. 2000; Tucker et al., 2001). However, the ages of deposition are not firmly established, and the ages of cleavages were unknown. This study solves this problem by dating the muscovites that define those cleavages using the $^{40}\text{Ar}/^{39}\text{Ar}$ method with step heating in a low-blank furnace. The ages of a detrital component and of two cleavages were constrained, and the results compared with the predictions of Bradley et al. (2000) and placed in a regional context.

[Insert Figure 1 here]

2. Geologic setting

The geology of east-central Maine (Fig. 1) is dominated by metasedimentary rocks that were deposited in what were previously described as the Central Maine sequence and the southeastern lobe of the Aroostook-Matapedia sequence (Fig. 1a). The two sequences are now known to have been deposited in a single large basin (Ludman et al., *in review*) thought to be one of several extensional basins that formed following the accretion of Ganderia to composite Laurentia and before the arrival of Avalonia (Fig. 2 and Table 1) in a period of crustal extension (Osberg et al., 1989; Tremblay and Pinet, 2005; Rankin et al., 2007). Bradley et al. (2000) proposed a model for late Silurian to Middle Devonian sedimentation during which deposition of deep-water sediments in the sedimentary basins from local sources was followed by deposition of a thick succession of clastic rocks in a migrating Acadian foreland basin sourced from the east.

[Insert Figure 2 here]

[Insert Table 1 here]

The units of interest to this study are the Madrid and Carrabassett Formations from the Central Maine sequence and the Smyrna Mills Formation from the Aroostook-Matapedia sequence (Fig. 1b). The early Silurian Madrid Formation comprises thickly bedded (up to 2 meters) typically well sorted, coarse-to fine grained turbiditic quartzofeldspathic sandstones with subordinate siltstones and shales (Bradley et al., 2000). The basal part of the Madrid Formation is thinly bedded (2-15 cm) and more calcareous than the overlying parts. In contrast, the Early Devonian Carrabassett Formation is composed of alternating beds of massive pelite, greywacke, and thin-bedded pelite/sandstone sequences in varying proportions (Hanson, 2013). The Silurian Smyrna Mills Formation is heterogeneous and composed mainly of weakly

metamorphosed mudstone interlayered with less abundant calcareous and non-calcareous siltstone and fine sandstone in variable proportions (Ludman 2010; 2013).

These strata were multiply deformed, regionally metamorphosed to greenschist facies, and intruded by plutons during the Acadian orogeny (Holdaway et al, 1982; Osberg et al. 1985). The classic Acadian deformation in east-central Maine is characterized by two phases of folding; the first phase produced recumbent folding associated with thrust faulting, but no cleavage, and the second phase produced map-scale upright folds with characteristic axial planar cleavage described as S_1 (Tucker et al., 2001).

Two plutons in the Lincoln area (Fig. 1b) are of importance in this study; namely, the early Devonian Mattamiscotis Mountain pluton (# 1 in Fig. 1b) and the Middle-Late Devonian Center Pond pluton (# 2 in Fig. 1b). Three concordant zircon analyses establish a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 407 ± 4 Ma for Mattamiscotis Mountain granite, and a mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 377 ± 3 Ma from two concordant zircon analyses establishes the age of Center Pond granodiorite (Bradley et al., 2000).

The Central Maine Boundary fault (CMBF, Fig. 1b) separates the Central Maine and the Aroostook-Matapedia sequences in east-central Maine. It is interpreted as an east-over-west thrust fault that was active during the early stages of the Acadian orogeny (Ludman et al., 1983). A five km-wide zone of intense dextral shearing east of the CMBF was first described by Ludman (2010) and named the Chester shear zone (CSZ, Fig. 1b). Deformation associated with the CSZ barely straddles the CMBF (Fig. 1b) and is mainly observed in the intensely disrupted beds of the Smyrna Mills Formation to the east, with limited deformation of the Madrid Formation to the west. It is characterized by asymmetric dextral and less common sinistral minor folds, and a

highly penetrative, closely spaced, near-vertical local cleavage Ludman, 2010) that will be described in this study as S_2 . The shear zone is proposed to be a local effect of the CMBF, which suggests that the fault was reactivated as a strike-slip fault after the earlier thrust motion (Ludman, 2010; Ghanem et al., 2013).

3. Previous attempts to date the cleavages

A scenario for the evolution ages of the Acadian orogeny in Maine and adjacent areas was proposed by Bradley et al. (2000). They reconstructed seven sequential positions of the foreland basin and Acadian deformation front (Fig. 3) and illustrated the migration of the orogenic belt to the northwest across Maine and adjacent areas from late Silurian to Middle Devonian time. The interpretation was based on (1) U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of pre-tectonic, syn-tectonic, and post-tectonic plutons; (2) ages, determined using conodonts and palynomorphs, of key strata that predate or postdate the deformation; and (3) interpretations of the depositional settings of these strata (e.g., far foreland, foreland basin, etc.). Their results suggest that regional Acadian deformation took place in the study area in central Maine between ~418 to 412 Ma (Fig. 3).

[Insert Figure 3 here]

Bradley et al. (2000) inferred the time of the regional Acadian deformation and the S_1 cleavage associated with its upright folding, but did not date the cleavage directly, nor did they address the younger deformation associated with the CSZ and the associated S_2 cleavage. Ludman (2010) proposed that if the CSZ was the result of a late-stage reactivation of the CMBF, the shearing must be post-Early Devonian, but there have been no previous attempts to date the S_2 cleavage prior to this study. As no plutons

truncate the CSZ structures observed in the field, the age of S₂ cleavage has therefore remained uncertain.

4. Strategy and methods

In order to constrain the ages of the metamorphic fabrics by directly dating the muscovite grains that define them using ⁴⁰Ar/³⁹Ar dating methods, samples for detailed petrography and ⁴⁰Ar/³⁹Ar geochronology were collected across the CMBF (Fig. 4a). Mineral content of the fine to very fine-grained rocks was determined using X-ray powder diffraction (XRD), and their microstructures were examined using optical and electron petrography.

[Insert Figure 4 here]

Fieldwork, including sampling, structural observations, and optical petrography were followed by analytical work. Mineralogical compositions of whole-rock samples and separates prepared for ⁴⁰Ar/³⁹Ar dating were analyzed by X-ray powder diffraction (XRD) using a Bruker D8 Advance X-ray diffractometer equipped with a Sol-X point detector at Indiana University. The powders were front mounted in cavities in Ti mounts. Analyses were run using Cu K α radiation from 2-70° with a 0.02° 2 θ step, counting for 2 sec/step. Identification of major phases was carried out using EVA software (Bruker AXS). Rietveld quantitative analyses (Bish and Howard, 1988) were performed using Topas 4.2 (Bruker AXS) to estimate mineral modes. During the refinement process, unit-cell parameters, and specimen-related parameters (e.g., lattice strain, crystallite size, preferred orientation of grains) were varied along with other factors in a least-squares procedure until the calculated powder pattern best matched the observed pattern (Post and Bish, 1989; Bish and Post, 1993). Preferred orientation of phyllosilicates on the X-ray mounts was modeled using the March function preferred

orientation correction, assuming preferred orientation on the 001 plane, consistent with the primary phyllosilicate cleavage direction. Electron petrography was done using an FEI Quanta FEG 400 scanning electron microscope at Indiana University, operating at 15 kV accelerating voltage under high-vacuum settings. The composition of some minerals in the samples studied was determined semi-quantitatively by energy-dispersive spectroscopic (EDS) spot analysis. $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology was conducted in Argon Thermochronology Lab at the U.S Geological Survey in Reston, VA. The preparation process began with crushing, grinding, and sieving, followed by standard magnetic and heavy-liquid separation, hand picking and paper shaking if needed, and then ultrasonic cleaning. Carbonates, when present, were removed with cold 10% HCl. The muscovite laths in the phyllitic samples proved to be too fine to prepare pure mineral separates; instead, muscovite-rich “matrix” of each sample was produced from the 60-80 mesh size fraction. Samples were irradiated at the USGS TRIGA reactor in Denver, Colorado (Dalrymple et al., 1981). Subsequent step-heating experiments were carried out by incrementally heating the sample in a low-blank furnace for ten minutes per step, followed by removing the reactive gases from each heating step and analyzing them in a VG MM 1200B or MAP 216 mass spectrometer. Data reduction was done with an updated version of computer programs ArAr* (Haugerud and Kunk, 1988) and Mass Spec (Deino, 2001), with decay constants recommended by Steiger & Jäger (1977).

5. Results

5.1. *Field and Petrographic Analyses*

Several outcrops across the CMBF (Fig. 4a) were visited to explore the spatial effect of the CSZ, describe its structural features, and collect samples for petrography

and $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology. Unfortunately, the sampling process was biased by poor exposure (less than 1%), limited access, and difficulty in sampling mostly pavement exposures. Results presented here come from two groups of outcrops and samples. One group records only the earlier regional deformation and associated foliation (S_1), whereas the second group displays structures characteristic of the CSZ transposing earlier regional fabrics. In some of these outcrops and samples, two foliations (S_1 and S_2) were recognized.

5.1.1. Macroscopic and mesoscopic structures related to the regional deformation and associated S_1 foliation

A single variably developed foliation (S_1) dominates most of the outcrops west of the CMBF (Fig. 4). The roadcuts along I-95 (Fig. 4a) display well-developed spaced cleavage in the metasediments and slates from the upper Madrid Formation (e.g., outcrops 19I and 19G) and Carrabassett Formation (e.g., outcrop 19H). The more pelitic basal Madrid Formation (e.g., outcrop K-C) displays closely spaced cleavage commonly associated with strongly transposed bedding (Fig. 5a). The orientations of these fabrics are summarized on Fig. 4b.

[Insert Figure 5 here]

The bedding-cleavage relationship changes from thin ~ 1 cm transposed layering parallel to cleavage (Fig. 5a) to ~ 15 cm thick graded beds perpendicular to cleavage best observed on cleavage-parallel surfaces. The cleavage-bedding relationship changes due to the different positions of the outcrops in the regional scale fold nose shown in Figure 4a (Osberg et al., 1985). The thick layers in fold noses are perpendicular to the cleavage, whereas the thinned beds in the fold limbs are transposed to an orientation

parallel or sub-parallel to the axial planar cleavage. Beds oblique to cleavage represent the part of the limb close to the fold nose. The intersection of bedding, striking 030° and dipping 64° SE, with a foliation plane striking 220° and dipping 87° NW, produces an intersection lineation that represents the hinge lines of a regional-scale F_1 folds. This hinge line trends 39° and plunges 18° NE and this folding is consistent with the deformation attributed to the NE Acadian deformation (Ludman, 2010). Small-scale upright folds are also present (Fig. 5b). The dominant S_1 cleavage is axial planar, strikes NE-SW, and dips at variable angles (65°-90°) to the NW or SE (Fig. 4b). S-folds that locally affect S_1 cleavage are present in the outcrops along I-95 and in outcrop K-C (Fig. 5c).

5.1.2. Macroscopic and mesoscopic structures related to the CSZ and associated S_2 foliation

The CSZ is exposed along a NE-striking belt >15 km long and about five km wide (Fig. 4a). Most of the CSZ deformation is localized in the thinly bedded, dominantly pelitic Smyrna Mills Formation east of the CMBF but adjacent parts of the Madrid formation are also affected. Characteristic features of the CSZ are (1) the dominant foliation S_2 (Fig. 5d) and parallel transposed bedding (Fig. 5e), (2) dominantly dextral asymmetric folds (Fig. 5e), although a few sinistral folds are also present, and (3) moderately to severely disrupted bedding (Fig. 5e). Although no systematic across-strike strain gradient was identified across the CSZ, S_2 foliation and asymmetric dextral folds are more strongly developed closer to the CMBF in the NW, and they are absent near the SE edge of the zone. Phyllosilicate-rich layers commonly display a well-developed close-spaced penetrative S_2 cleavage, whereas the more competent fine sandstone beds are sheared (Fig. 5e), fractured, or boudinaged (Fig. 5f). Kinematic

indicators such as S-C fabrics (Fig. 5f) and asymmetric folds (Fig. 5e) demonstrate the dextral sense of most shearing in the CSZ.

Earlier S_1 foliation is best preserved in microlithons in F_2 fold noses (Fig. 5d) where it is parallel to bedding and perpendicular to S_2 . S_2 cleavage ranges from scaly, closely spaced, and undulating appearance in pelitic rocks to a disjunctive cleavage in silty-sandy beds. The general trend of the foliation and shear bands is NE-SW with near-vertical dip mostly to the NW (Fig. 4b). S_2 and transposed bedding are weakly folded which makes them appear wavy. However, they may be locally folded into tight dextral asymmetric folds (F_3), with near-vertical axial planes striking N-S. Some folded sandy beds are boudinaged (black arrow, Fig. 5e), showing that extension (probably during D_1) preceded F_2 folding. Quartz and calcite veins are present in some outcrops. They are thickest where they cut the calcareous sandstone at a high angle, but as they pass into phyllite, they become thin and nearly parallel to the foliation. These relationships demonstrate that the veins must have formed before or early in the formation of S_2 .

5.2. Mineralogy, petrography and microstructures

5.2.1. S_1 microstructures and relationships to mineralogy

S_1 regional foliation is the only cleavage observed in thin sections of all samples collected outside the CSZ (Fig. 4a) from the Carrabassett Formation (19H) and the Madrid Formation (19F, 19G, 19I, and K-C). Samples 19F, 19G, and 19I from the upper parts of the Madrid Formation display weakly developed disjunctive cleavage. Their thin sections show the volumetric dominance of poorly sorted and angular detrital grains of quartz, with generally less plagioclase and muscovite (Fig. 6a). This was supported by the XRD modal analyses, which indicate <30% phyllosilicate content (Table 2) and the dominance of quartz. The high concentrations of plagioclase shows

most of these samples to be arkoses. Clast size varies but is generally $<250\text{ }\mu\text{m}$ in diameter ranging from silt size to fine sand size with a fine calcareous matrix. Some samples contain $\sim 1\text{-}2\text{ mm}$ thick, phyllosilicate-rich layers alternating with phyllosilicate-poor layers. Detrital muscovite flakes are commonly bent (Fig. 6b) and exhibit undulose extinction. The elongate detrital muscovite flakes define a strong preferred orientation (Fig. 6b) defining bedding planes (S_0).

[Insert Table 2 here]

[Insert Figure 6 here]

Samples K-C (from the basal part of the Madrid Formation) and 19H (from the Carrabassett Formation) contain $>50\%$ phyllosilicate minerals. Sample 19H contains thinly interbedded muddy siltstone and slightly calcareous fine-grained arkosic sandstone. The muddy siltstone contains detrital muscovite and muscovite-chlorite stacks $<100\text{ }\mu\text{m}$ thick, which are truncated by fine chlorite and muscovite ($<10\text{ }\mu\text{m}$) that define S_1 folia. In the fine sandstone beds, S_1 is defined by opaque seams of fine phyllosilicate (Fig. 6c).

S_1 fabric becomes increasingly well developed in the basal part of the Madrid Formation. In sample K-C, thin silty and muddy phyllosilicate-rich beds are folded with a well developed axial-planar and muscovite-rich cleavage (S_1 ; Figs. 6d and 6e) that truncates older detrital and authigenic mica (Fig. 6f). In general, the (00l) planes of interlayered chlorite-muscovite stacks are oblique or perpendicular to the orientation of the S_1 folia. Some stacks are folded (Fig. 6f), and crenulation cleavage is locally developed in muscovite- and chlorite- rich layers.

5.2.2. S_2 microstructures and relationships to mineralogy

Samples collected from the CSZ (Fig. 4a) display a variety of textures and mineralogy. Phyllosilicate-rich samples display two foliations (S_1 and S_2), whereas samples collected from transposed phyllosilicate-poor sandy beds display only S_1 . Samples from the east side of the CSZ differ by having biotite (samples 30-6A and 19C) or biotite + garnet (sample 19D). The mineralogy, petrography, and microstructures are presented for two broad groups in the CSZ: the chlorite grade rocks and the biotite-garnet grade rocks.

5.2.2.1. *Chlorite Grade*

Samples K-A, K-B, K-D, K-E, K-F, 19B and 19E are chlorite-grade metasandstones and slates that were collected from outcrops that display the characteristic CSZ macrostructures described previously. The dominantly pelitic nature of samples K-D, K-E, K-F, 19B and 19E from the Smyrna Mills Formation was confirmed by both petrographic and XRD analyses (Table 2). The chlorite + muscovite contents were 35 to >75 wt.%. All samples contain well-developed S_1 cleavage parallel to transposed bedding, and both are in turn crenulated and truncated by a well-developed S_2 foliation (e.g., Figs. 7a-d). Some samples display transposed bedding parallel to S_2 foliation and sheared lens-shaped microlithons (Fig. 7e) of slate that preserve the earlier S_1 fabric.

[Insert Figure 7 here]

The S_2 fabric-forming chlorite and muscovite are characterized by strongly preferred shape orientation, very fine grain size, and high aspect ratio (Figs. 7c and e). S_2 folia form domains rich in phyllosilicate (P-domains, Figs. 7b and c) that alternate with domains rich in detrital quartz and albite and/or carbonate minerals (QF-domains or microlithons; Figs. 7b and c). S_2 is locally crenulated in some samples, but no younger foliation is observed in which muscovite might have recrystallized. Rocks from the

eastern border of the CSZ and east of the biotite zone (Fig. 4a), are relatively rich in muscovite (e.g., sample 19E; Fig. 4a and Table 2). Petrographic analysis shows that S_2 folia are thick and almost entirely defined by muscovite and bright needles of rutile.

Phyllosilicate-poor siltstones and arkoses from within the CSZ are different from the phyllosilicate-rich rocks described above; e.g., samples K-A and K-B are both from transposed, sheared, folded, and boudinaged beds, but with textures and structures that still preserve S_1 foliation and parallel transposed sedimentary layering. The mineralogy of sample K-A is dominated by silt-sized detrital grains of quartz and albite with much less muscovite and chlorite (<30% Table 2; Fig. 7f). Fine laths (<10 μm) of diagenetic muscovite and chlorite were found randomly oriented between larger grains. S_1 foliation is defined by muscovite and chlorite, weakly developed, and truncates older detrital and authigenic populations of muscovite and chlorite. BSE petrography of sample K-B (Fig. 7g) reveals many textures including small random laths of chlorite, sutured albite boundaries, and fine folia of muscovite and chlorite that truncate the older diagenetic chlorite-muscovite stacks. The recrystallized muscovite and chlorite define a weak foliation sub-parallel to the bedding (S_0).

5.2.2.2. *Biotite-Garnet Grade*

Non-pelitic rocks in the biotite $\pm$ garnet zone are arkosic (samples 19C, 19D, and 30-6A, Table 2) and the role of biotite in the fabric is variable. Biotite occurs with oligoclase plagioclase in all samples, but some contain phyllosilicate-rich silty beds that locally preserve the earlier muscovite-chlorite fabric with no biotite. The textures vary depending on grain size and layer composition. For example, biotite is concentrated in some samples in alternating domains that may represent earlier P domains separated by QF domains (Fig. 8a). Sandy layers (QF-domains) display randomly oriented biotite

(Fig. 8a), whereas in phyllosilicate-rich layers (P-domains) biotite may define a moderate preferred orientation sub-parallel to the chlorite and muscovite that define S_2 foliation (Fig. 8b).

[Insert Figure 8 here]

Biotite zone sample (30-6A) is phyllosilicate-rich, contains no carbonates, and contains much less biotite (8 wt.%) than muscovite (40 wt.%, Table 2). The earlier S_1 fabric is crenulated and truncated by the later S_2 fabric (Fig. 8c). S_1 foliation is defined by muscovite and minor chlorite whereas S_2 foliation is defined almost entirely by muscovite. Biotite flakes may be randomly oriented or parallel to S_2 foliation (Fig. 8c). Biotite laths may also be concentrated around stretched and folded quartzo-feldspathic layers.

One sample (19D) from the garnet zone (Fig. 4a) was selected for detailed analysis. It contains 11% biotite, 30% muscovite, and minor garnet (Table 2). Small garnet porphyroblasts (typically with a diameter of $<75\ \mu\text{m}$) and small biotite porphyroblasts ($<50\ \mu\text{m}$) both truncate crenulated S_1 and S_2 fabrics (Fig. 8d). The fabrics are defined by chlorite and muscovite (Fig. 8d). Garnet does not occur in the arkosic beds but is restricted to phyllosilicate-rich folia.

5.3. $^{40}\text{Ar}/^{39}\text{Ar}$ Results

5.3.1. $^{40}\text{Ar}/^{39}\text{Ar}$ results from samples with S_1 foliation

Five samples that only display S_1 fabric were selected for $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology to constrain the ages of muscovite defining the regional S_1 foliation. These samples are 19H from the Carrabassett Formation, and 19F, 19G, 19I and K-C

from the Madrid Formation (Fig. 9a to e). Isotopic data are presented in Table 1 in the supplementary material.

Step-heating experiments on samples 19G, 19H, 19I, and K-C yielded sigmoidal age spectra (Fig. 9b, c, d, and e). These age spectra are characterized by a hump shape in low-temperature steps followed by steadily climbing apparent ages from ~430 Ma to ~440 Ma in sample 19G, from ~425 Ma to ~455 Ma in sample 19H, from ~428 Ma to ~458 Ma in sample 19I, and from ~411 Ma to ~450 Ma in sample K-C. Steps with * are not considered to be geologically meaningful (see discussion below). The age spectrum of sample 19F behaves differently and climbs steadily from a minimum apparent age of ~412 Ma to ~433 Ma. No hump shape is observed in the low-temperature age steps.

[Insert Figure 9 here]

5.3.2. $^{40}\text{Ar}/^{39}\text{Ar}$ results samples with S_1 and S_2 foliations

Seven samples from the CSZ were selected for $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology. These samples are K-A, K-D, K-E, K-F, 19B, and 19E from the chlorite-grade rocks of the Smyrna Mills Formation, and sample K-B from the Madrid Formation (Fig. 9f to l). Step-heating experiments conducted on the chlorite-grade samples yielded sigmoidal age spectra (Fig. 9f to l). These age spectra begin with hump shapes in low-temperature steps, followed by steadily climbing apparent ages as low as 384 Ma in sample K-F, to as high as 466 Ma in sample K-B. The age spectrum of sample 19E behaves differently and displays a hump in the low-temperature steps followed by climbing steps up to ~394 Ma with about 40% of the gas on three steps, two of them are ~394 Ma and one of them is ~391 Ma. The ages then drop in two steps from 382 Ma and 381 Ma with the release of another 40% of the gas. The last small steps climb abruptly to 396 Ma then drop again to ~379 Ma.

6. Discussion

6.1. Summary and discussion of the petrography

6.1.1. Chlorite-grade fabrics

The samples can be classified into two broad textural groups: the first is phyllosilicate-poor meta-arkosic with a single weakly developed cleavage (S_1) in which phyllosilicates are only locally concentrated into folia; the second group of samples displays two foliations (S_1 and S_2).

Muscovite belongs to three textural populations as identified based on the orientation, shape, size, and truncation relationships. Some studies have successfully distinguished between different populations of muscovite based on their chemistry using electron microprobe (EMP) analyses (e.g. McAleer et al., 2016). However, the very fine grain size of metamorphic muscovite and its intergrowth with chlorite at a very fine scale (Figs. 7c and d) made it inappropriate to be analyzed using the EMP. Thus, based on textural relationships, the detrital muscovite population was identified to include grains that tend to be large (up to 150 μm long), sub-parallel to bedding and to one another, have frayed ends, and are typically bent or kinked (e.g., Fig. 6b). Diagenetic muscovite (and chlorite; Fig. 7g) occurs in two textural forms: the first is small ($\leq 10 \mu\text{m}$) interlocking muscovite and chlorite flakes filling spaces between detrital feldspar and quartz grains, which suggests a diagenetic or low-grade metamorphic occurrence; the second is as stacks of intergrown muscovite and chlorite typically with a diameter of $>50 \mu\text{m}$ (e.g., Figs. 6c and 6f). Some stacks have their ends commonly truncated by the muscovite and chlorite defining S_1 and S_2 fabrics, and have internal layering that is oblique to the fabric orientation. This type of chlorite-muscovite intergrowth has been interpreted as a growth of secondary minerals that replace detrital (biotite?) grains

during diagenetic-metamorphic processes (Clark and Fisher, 1995). Truncation of these stacks by S_1 fabric-forming muscovite and chlorite suggests that they are pre-tectonic and that the deformation mechanism responsible for S_1 fabric was dissolution of detrital/diagenetic grains and precipitation of new fabric-forming grains.

The third textural variety of muscovite and chlorite is clearly metamorphic and defines the one or two generations of foliation in the rocks (e.g., Figs. 7b-e). This population of muscovite and chlorite is characterized by strongly preferred shape orientation, very fine grain size ($\sim 25\ \mu\text{m}$ long), high aspect ratio, and by the fact that they truncate both muscovite- chlorite stacks and detrital muscovite grains. The truncations represent evidence that both S_1 and S_2 fabrics were developed by a dissolution-precipitation deformation mechanism rather than by mechanical rotation. In some samples, the S_2 fabric is also locally crenulated but no evidence was observed for recrystallization accompanying this event.

6.1.2. Biotite- and garnet- grade fabrics

Samples from the biotite zone display S_1 and S_2 fabrics (Fig. 8c), and both samples contain biotite that either parallels S_2 (Fig. 8c) or occurs as randomly oriented flakes that truncate earlier foliations. Sample 19D displays two fabrics and contains biotite and garnet that truncate both earlier foliations (Fig. 8d).

Garnet crystals locally crosscut S_1 and S_2 foliations and some preserve those fabrics in inclusion trails. These are clearly younger than both foliations. However, in some samples biotite flakes are parallel to S_2 foliation, whereas others truncate all foliations. The simplest interpretation of these relationships is that the biotite forming reactions began during the development of S_2 foliation but that the heat outlasted the deformation so that biotite and garnet growth continued after deformation stopped.

The occurrence of oligoclase plagioclase associated with biotite suggests that a Ca-bearing phase was involved in the reaction. Although this reaction may not be the only biotite forming reaction, the abundance of calcite veins in the outcrop supports the occurrence of calcite as a secondary phase as well as a primary one in calcareous beds, this suggests the following reaction:

This reaction differs from the biotite-forming reaction proposed by Ferry (1976, reaction 1) in that it does not involve chlorite as a reaction product consuming ambient H₂O. Indeed, chlorite persists in rocks of the biotite zone (Table 2), and in the absence of obvious textures suggesting a reaction relationship, chlorite is not considered a reactant or product in this reaction. In contrast, oligoclase rather than albite was best suited for Rietveld refinement analyses for biotite bearing samples (Table 2), and thus is identified as a reaction product consuming excess Al made available by the breakdown of muscovite.

Biotite defining S₁ foliation is absent in the rocks of Central Maine (Fig. 4a) indicating the lower greenschist grade of metamorphism in central Maine. This is supported by the local presence of the classic chlorite zone assemblage of K-feldspar + chlorite (e.g. Spear, 1993) in some arkosic sediments (Table 2). Thus, the occurrence of biotite and garnet suggests a thermal metamorphism event associated with the emplacement of igneous intrusions, most likely, associated with the Center Pond pluton in this locale. The trend of the isograds at a high angle to the contact of the Center Pond pluton suggests that the presence of a stock of magma slightly below the surface of the biotite zone rocks.

6.2. Discussion and interpretation of the $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology

The interpretation of $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra obtained from muscovite concentrates from phyllites and slates requires an understanding of the variables affecting the apparent age of an age step, and the shape of the overall age spectrum. These variables include: 1) the presence of impurities; 2) ^{39}Ar recoil associated with finely intergrown chlorite and muscovite; 3) multiple populations of muscovite; and 4) the presence of excess ^{40}Ar . A discussion of these variables and their effect on the age spectra of our samples is summarized below.

6.2.1. Impurities

The presence of impurities is a common problem when studying very fine-grained samples because the mechanical extraction of a pure separate of muscovite is usually impossible. Thus, if the matrix of the rock is used for dating, large, unwanted grains are removed, but small grains of impurities may remain. Low radiogenic yield of age steps, and anomalously low K/Cl and K/Ca ratios associated with them, are inferred to indicate high concentrations of gas released from impurities. Age steps so affected typically occur at low and high temperatures. Results from powder XRD analysis of the matrix separates used for dating indicate that impurities including albite, quartz, chlorite, and rutile are present in the samples at >1% level, (Table 2) and others may be present in smaller amounts. Steps interpreted to have been affected by impurities are indicated by * on the age spectra (Fig. 9), and their apparent ages are not given any geological significance.

6.2.2. ^{39}Ar Recoil

During the irradiation process, ^{39}Ar may recoil from its host K-rich mineral into an adjacent K-poor mineral (McDougall and Harrison, 1999), which would not necessarily affect the total gas ages, but rather the shape of the age spectra (Hess and Lippolt, 1986). The ^{39}Ar recoil effect is commonly seen in age spectra of slaty samples in which muscovite is finely intergrown with chlorite (e.g. Kunk et al. 2005). The presence of non-separable muscovite-chlorite laths allows the ^{39}Ar produced from ^{39}K to recoil from muscovite into adjacent chlorite, which concentrates ^{39}Ar in chlorite and depletes it from muscovite. In step-heating experiments, ^{39}Ar is released from chlorite at lower temperatures in two major pulses (550 °C and 700-900 °C; Lo and Onstott, 1989), whereas muscovite degases at higher temperature (900-1050 °C; Sletten and Onstott, 1998). This difference in temperatures of degassing allows the interpretation that the gas released from heating steps at low temperature is enriched in recoiled ^{39}Ar from chlorite and is thus artificially too young. The subsequent rapid climb of apparent ages at moderate temperatures can be attributed to gas released from fine grains of muscovite affected by the depletion in ^{39}Ar , and providing ages that are consequently too old. The effect of recoil fades with increasing temperatures of heating steps, and the series of steadily climbing steps from a minimum step age can be attributed to gas released from recoil-free muscovite.

Dong et al. (1995) showed that ^{39}Ar recoil effect in fine minerals like clays, was strongly controlled by their degree of crystallization. The degree of crystallization of fine muscovite laths in this study is eliminated as a factor affecting the ^{39}Ar recoil as evidenced from the XRD patterns (Fig. 1 in the supplementary material), which show sharp 00l peaks so metamorphic 2M1 muscovite is present rather than illite or I/S.

The only sample that produced an age spectrum with no hump shape at low-temperature steps is sample 19F. XRD results (Table 2) indicate that this sample has no detectable chlorite, which supports the association of hump-shaped low-temperature steps in the age spectra with the presence of chlorite. BSE images show that chlorite and muscovite are finely intergrown down to a scale of one μm in all other samples. Therefore, the hump-shaped part of the age spectra at low temperatures seen for other samples, is interpreted to be the result of ^{39}Ar recoiled from muscovite into adjacent chlorite, and the ages of these steps are discounted.

6.2.3. *Multiple Populations of Muscovites*

The third variable that affects the shape and interpretation of age spectra is the presence of multiple age populations of muscovite, e.g., detrital, authigenic, and multiple generations of cleavage-forming muscovite. Detrital muscovites are found to retain their apparent ages throughout the metamorphism in Central Maine. That is, they were not reset by heating to temperatures above the closure temperature of Ar in muscovite ($\sim 350 \pm 50$ °C, McDougall and Harrison, 1999). This implies that the diagenetic and low-grade metamorphic populations retain their crystallization ages.

Several factors control the rates of Ar release during *in vacuo* heating experiments including chemical composition, structural integrity, and grain size (Sletten and Onstott, 1998). Therefore, the sigmoidal-shaped age spectra are attributed to variations correlated with the presence of different populations defined by different grain sizes. This implies that fine grains of muscovite preferentially degas at lower temperatures compared with coarse muscovite that dominates most of the gas at higher temperatures. BSE images indicate that the detrital muscovite is relatively coarse, up to 120 μm , and that the fabric-forming muscovite is finer grained (typically ~ 25 μm). Thus,

the recrystallized finer muscovite is expected to degas in lower-temperature steps than the detrital population.

6.2.4. *Excess Ar*

The concentration of ^{40}Ar in minerals may exceed that generated by *in situ* decay of K. Excess Ar could be inherited by contamination with pre-existing Ar, or it could be introduced to the system by fluids or melt at the grain boundary (Kelley, 2002). This problem can be detected from inverse isotope correlation diagrams by determining the initial $^{40}\text{Ar}/^{36}\text{Ar}$ ratio of the sample (McDougall and Harrison, 1999). However, this method is applicable only to samples in which the isotope ratio within the samples is homogeneous, and it therefore cannot be used for samples with multiple generations of muscovite (Kunk et al., 2005). Saddle-shaped age spectra or those with a climb in age in their high-temperature steps are interpreted in some cases as indicators of excess Ar (McDougall and Harrison, 1999; Forster and Lister, 2004). Although the presence of excess Ar cannot be ruled out in any individual sample, it does not appear to have a significant effect in the interpretation of any of the muscovite age spectra in this study (Fig. 9).

6.3. *Discussion of the $^{40}\text{Ar}/^{39}\text{Ar}$ ages*

Petrographic results discussed earlier reveal the presence of multiple populations of muscovite: detrital muscovite, two generations of metamorphic muscovite defining S_1 and S_2 foliations, and a minor diagenetic population of muscovite. The latter is too small to affect the results of $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology, thus making it possible to constrain, if not precisely pinpoint, the ages of the other three significant populations of muscovite.

6.3.1. *Constraining the age of detrital muscovite*

Petrographic observations indicate that samples 19G, 19H, 19I, K-A, and K-B are fine (albite-rich) arkosic sandstone beds dominated by quartz, albite, and detrital muscovite with weakly developed S_1 fabric. The truncations of detrital muscovite by S_1 are taken as evidence for destruction of those grains by dissolution whereas the truncating muscovite in S_1 is taken as evidence for new crystallization of muscovite. The climbing pattern of steps in the age spectra of these samples is interpreted to result from dating a mixture of old detrital muscovite and young metamorphic muscovite (S_1). As the proportions of each population that was dated are impossible to quantify, the maximum age step in each age spectrum is interpreted to be the minimum cooling age of the larger size fraction of muscovite, namely the detrital muscovite. Comparing the maximum age step of all the samples (Fig. 9), the oldest maximum apparent age is taken as the minimum cooling age of the detrital muscovite; in this suite of samples, it is ~466 Ma from sample K-B (Fig. 9f).

6.3.2. *Constraining the age of S_1 muscovite*

Samples 19F, 19G, 19H, 19I, K-B, and K-C all contain detrital muscovite and variably developed S_1 folia. Sample K-C displays the best example of S_1 muscovite truncating the earlier detrital muscovite grains (Fig. 6f). As discussed in the previous section, the truncations of detrital muscovite are taken as evidence for the crystallization of new muscovite defining S_1 , and the matrix used in the step-heating experiments is a mixture of both the detrital muscovite and the younger S_1 metamorphic muscovite. The minimum age step in each age spectrum is interpreted to be the maximum crystallization age of S_1 muscovite. Sample K-C produced the youngest

geologically meaningful age step (~ 411 Ma) and contains the highest proportion of S_1 cleavage-forming muscovite. Thus, the age of S_1 is Early Devonian, ~ 411 Ma or younger.

6.3.3. *Constraining the age of S_2 muscovite*

Almost all samples in the CSZ display well-developed S_2 foliation. Samples 19B, 19E, K-D, K-E and K-F are phyllosilicate-rich in which S_2 fabric truncates and crenulates S_1 fabric (e.g., Figs. 7b to e). Remnants of detrital muscovite are not common but may be preserved in the microlithons. The age spectra of these samples yield younger ages than samples containing only S_1 foliation. This is interpreted to reflect the low concentration of detrital grains in these samples and the addition of a younger S_2 population to the now ternary mixture of muscovite populations. Because petrography documents that, the S_2 population of muscovite in these samples is the youngest of the three, the minimum apparent ages in the climbing age spectra are interpreted as the maximum crystallization age of S_2 muscovite. However, the age steps with the maximum apparent age may represent a mixture of detrital and S_1 muscovite, and they probably further underestimate the age of the detrital component. Therefore, among the age spectra of samples K-D, K-E, K-F, 19B, and 19E (Fig. 9h-l), the minimum age step of ~ 381 Ma from sample K-F is interpreted as the maximum crystallization age of the youngest muscovite defining S_2 fabric.

Arkosic beds tend to be dominated by detrital muscovite. For example, sample K-A from a silty arkose bed in the Smyrna Mills Formation contains detrital muscovite and displays a moderately developed S_1 foliation and weaker S_2 foliation (Fig. 7f). The age spectrum of the sample climbs from ~ 403 Ma to ~ 461 Ma. The maximum apparent age indicates the dominance of detrital muscovite with a minimum age of Late Ordovician, whereas the minimum apparent age is a mixture of detrital, S_1 muscovite, and S_2

muscovite populations. Therefore, the youngest age step is interpreted to represent degassing from a ternary mixture of these three populations of muscovite, and the age of this step is interpreted as a maximum crystallization age of the muscovite defining S₂ fabric.

6.3.4. *Thermal and Tectonic Evolution of east-central Maine*

⁴⁰Ar/³⁹Ar thermochronologic results combined with petrographic observations were not only successful in constraining the ages of S₁ and S₂ cleavages, but they also provide constraints on the provenance and metamorphic grade of rocks in east-central Maine. These data, together with the cooling ages of amphibole and biotite from the Center Pond pluton (Ghanem et al., 2013 and Ghanem, 2015), place constraints on the thermal and deformational history of east-central Maine.

The ambient grade of metamorphism at the time of regional deformation and younger local deformation in the CSZ was lower to middle greenschist facies. The lack of regional biotite suggests that temperatures never exceeded those necessary of the biotite isograd (300 to 400°C; Spear, 1993). The ⁴⁰Ar/³⁹Ar data support this interpretation because the sigmoidal age spectra indicate that detrital muscovite was not reset by thermal metamorphism. This places a thermal maximum of ~350°C for the Devonian regional metamorphism and deformation.

The oldest maximum apparent age of detrital muscovite is Middle Ordovician (in this suite of samples, it is ~466 Ma from sample K-B). This age is interpreted as the minimum cooling age of detrital muscovite). This indicates a Middle Ordovician or older cooling through the closure temperature of Ar diffusion in muscovite (350°C ± 50, McDougall and Harrison, 1999). This datum provides a starting point on the T-t diagram of Figure 10.

The rates of Ordovician exhumation in the provenance area can be estimated from previous studies of the depositional ages of the Smyrna Mills, Carrabassett, and Madrid Formations. The Smyrna Mills is dated by fossils as middle Silurian (Pavlides and Berry, 1966). Detrital zircons in the Madrid Formation are as young as $\sim 433 \pm 2$ Ma (Ludman et al., 2014), which limits the maximum age of deposition of the Madrid Formation to also be middle Silurian, and the overlying Carrabassett Formation is mapped as Early Devonian (Dorais et al., 2009). Therefore, the age of deposition of the sedimentary sequences ranges between uppermost Silurian (Pridolian) and Early Devonian. A deposition age of ~ 425 Ma indicates that the host rock of the detrital muscovite was cooled, exhumed, eroded, and then deposited within ~ 40 Ma. This suggests a cooling rate from $\sim 350^\circ\text{C}$ to the surface temperature (near zero) of almost $10^\circ\text{C}/\text{m.y.}$ This rate is consistent with tectonic exhumation and the timing suggests exhumation following the Taconic orogeny.

The protoliths of the Carrabassett and Madrid Formations were buried, folded, and metamorphosed to lower greenschist facies conditions before the intrusion of the Mattamiscotis Mountain pluton at 407 ± 4 Ma (Bradley et al. 2000). The depth of the intrusions is ~ 9 km as suggested by Ghanem (2015) based on estimated pressure from the contact aureole of the Center Pond pluton of ~ 0.25 GPa (2.5 kbar). If the rocks were buried to a depth of 9 km (roughly between 433 Ma and 407 Ma), then a normal rate of deposition of ~ 0.5 mm/year is implied. These relationships are shown on Figure 10.

The age of the muscovite that defines S_1 was constrained earlier to be ≤ 411 Ma, but $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology cannot constrain the time at which crystallization of S_1 muscovite ended. However, the combination of the $^{40}\text{Ar}/^{39}\text{Ar}$ crystallization age of S_1 and the emplacement ages of the Mattamiscotis Mountain and the Seboeis Lake

plutons that intrude rocks of this study (Fig. 1b), help to constrain the time of S_1 foliation to a narrow span. Bradley et al. (2000) reported crystallization ages of 406 ± 3 Ma for a foliated granodiorite unit in the Seboeis Lake pluton and 407 ± 4 Ma for unfoliated granite of the Mattamiscotis Mountain pluton. Hornfels collected from the contact aureole of the Mattamiscotis Mountain pluton contains sillimanite and andalusite and lacks any foliation or evidence of deformation (Ghanem, 2015). Therefore, development of S_1 foliation is older than the intrusion age of 407 ± 4 Ma. Together, these data constrain the time of S_1 crystallization between ~ 411 and ~ 407 , or Early Devonian (Fig. 10).

The development of the S_2 fabric in the CSZ rocks east of the CMBF is constrained to be younger than ~ 381 Ma, the inferred maximum age of S_2 muscovite. As with the S_1 fabric, the $^{40}\text{Ar}/^{39}\text{Ar}$ apparent ages do not provide a minimum age of the S_2 fabric. However, the S_2 fabric must be older than the contact metamorphism around the Center Pond pluton because the metamorphic assemblages overprint the S_2 fabric, as shown above. Thus, the age of the Center Pond pluton (377 ± 3 Ma; Bradley et al., 2000) establishes a minimum age for S_2 . These ages constrain the time of muscovite crystallization in S_2 in the CSZ to be between ~ 381 and ~ 377 Ma, or Late Devonian (Fig. 10).

[Insert Figure 10 here]

6.3.5. *Implications of the results on the regional geology of Maine*

The axial planar S_1 cleavage (Fig. 6d, e, and f) represents deformation in the early contractional phase of Acadian deformation. Our study presents direct isotopic evidence for the age of S_1 muscovite in addition to previous constraints established from isotopic ages of plutons that intrude the regionally folded rocks. The estimated time of the

development of S_1 fabric (411 - 407 Ma) is very close to the arrival time of the Acadian deformational front estimated by Bradley et al. (2000) as 418-412 Ma (Fig. 3). This is also in agreement with the age range 419-404 Ma of the early Acadian contractional deformation of Tucker et al. (2001). Their suggested minimum age limit is based on the age of the Lexington pluton of 404 ± 2 Ma that intruded the regionally deformed Carrabassett Formation in south-central Maine.

In contrast, the Late Devonian dextral deformation in the CSZ coincides in time with dextral transpressive shortening associated with the Norumbega fault system (Fig. 1; Ludman et al., 1999; West, 1999). Indeed, the times of the ~ 30 Ma age gap between the peaks of the regional contractional S_1 and younger dextral transpressive S_2 deformation found in the Lincoln area agree with the times of the two kinematically distinct tectonic episodes of the Acadian orogeny in SW Maine suggested by Gerbi and West (2007). Such regional correlations are to be expected if Late Devonian deformation can be attributed to the regional accretion of Meguma (Fig. 2) to composite Laurentia in the Neocadian orogeny (van Staal and Barr, 2012). Thus, it is likely that the CMBF was reactivated as a dextral fault in the Neocadian orogenic episode during a regional transpressive event in the northern Appalachians.

7. Conclusions

The timing of two mid-Paleozoic deformation events in an east-central Maine slate belt has been determined through a combination of $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and careful characterization of textures and structures in fine-grained lower greenschist facies metamorphic rocks. The time spans during which foliation-forming muscovite grains crystallized were estimated by combining $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology with

previously reported U/Pb crystallization ages of syn-tectonic and post-tectonic plutons.

Applying these results to east-central Maine leads to the following conclusions:

- The ambient metamorphic temperature in the Lincoln area did not exceed the closure temperature of Ar retention in muscovite ($\sim 350^{\circ}\text{C}$; McDougall and Harrison, 1999).
- Axial planar cleavage (S_1) defined by newly crystallized muscovite in the Lincoln area is constrained to be to Early Devonian (411-407 Ma). This age provides a good minimum estimate for the timing of regional Acadian folding.
- Muscovite in the foliation of the CSZ (S_2) crystallized during the Late Devonian (381-377 Ma).
- Fabric-forming deformation in the CSZ ended by 377 ± 3 Ma, the age of intrusion of the nearby Center Pond pluton.
- Strain-induced dissolution-precipitation that crystallized S_2 muscovite was triggered by strain localization that probably reactivated the CMBF as a dextral strike slip fault. This strain may be associated with early deformation on the Norumbega fault system that is well developed to the southeast along coastal Maine.
- The lower greenschist facies rocks of east-central Maine provide evidence for two deformation events. These coincide in time with a change in the plate convergence kinematics in the Middle-Late Devonian from shortening, reflecting the arrival of the Avalon terrane, to right-lateral transpression coincident with the arrival of the Meguma terrane (e.g., van Staal and Barr, 2012).

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[Figure 1]. Geologic map of east central Maine, edited after Ludman and Hopeck (2012). Dashed box is the location of Figure 4. Numbered plutons are (1) the Mattamiscotis Mountain pluton; (2) the Center Pond pluton.

[Figure 2]. Map of a part of the northern Appalachians displaying the remnants of different peri-Gondwanan terranes (Ganderia, Avalonia and Meguma) that were accreted to both Laurentia and associated peri-Laurentian terranes. Modified from Hibbard et al. (2006). Outline of Maine is shown on the map as a reference.

[Figure 3]. Sequential positions of the foreland basin (top) and Acadian deformation front (bottom). Edited after Bradley et al. (2000). Three plutons, the town of Lincoln, and the area of the map in Figure 1b are labeled. Dsl: the Devonian Sebobeis Lake pluton (406 ± 3 Ma); Dcp: the Devonian Center Pond pluton (377 ± 4 Ma). Both ages are from Bradley et al. (2000).

[Figure 4]. Geologic maps of the Lincoln area. (a) A geologic map of the Lincoln area with the metamorphic isograds identified. Locations of outcrops are given by 'x', and samples with labels are those discussed in the text. (b) The same geologic map of the Lincoln area with structural data measurements (using the right-hand rule). Equal-area and lower-hemisphere projections summarize poles to planes of measured foliations (orange dots) from outcrops NW and SE of the CMBF (dominated by S_1 and S_2 , respectively). Calculated conical best-fit values (black dots) show that S_1 and S_2 are parallel. The projection was prepared using Stereonet software (Allmendinger et al. 2012; Cardozo and Allmendinger, 2013). CMBF: Central Maine Boundary fault; CSZ: Chester shear zone. Legend for stratigraphy is the same as in Figure 1.

[Figure 5]. Field observations and structures. (a) Transposed thinned bedding parallel to the regional cleavage S_1 ; pen is 12 cm long. (b) Small upright symmetrical fold (F_1) in the regionally deformed rocks; the tip of the pen is 2 mm. (c) Asymmetric S-fold crenulating S_1 cleavage in the regionally deformed rocks; the tip of the pen is 2 mm. (d) A bed of fine sandstone partially transposed by S_2 with a low-strain F_2 fold nose preserving S_0 and S_1 fabrics; from outcrop K-F; the tip of the pen is 2 mm. (e) Sandy beds (s.s) display dextral folds (F_2). Long limbs form boudins (black arrow); outcrop K-F. (f) Domino-type boudins in sandstone with S-C fabric indicating dextral shear sense; from outcrop K-D.

[Figure 6]. Textures from regionally deformed rocks that display S_1 foliation. (a) A crossed polarized light photomicrograph from sample 19G displaying poorly sorted and rounded sandstone from Madrid Formation, the photomicrograph shows angular detrital quartz, plagioclase, and muscovite in a calcareous matrix. (b) A crossed polarized light photomicrograph of strongly oriented elongate detrital muscovite grains parallel to bedding planes (S_0); some grains are bent (from sample 19I). (c) A plane-polarized light photomicrograph of dark seams of fine phyllosilicate truncating muscovite-chlorite stacks, from sample 19H from the Carrabassett Formation. (d) A plane-polarized light photomicrograph of an inverted F_1 -folded graded calcareous arkose bed (yellow dashed line) within muddy beds (brown layers). Axial planar cleavage (S_1) is well developed in micaceous beds (from sample K-C). (e) A crossed-polarized photomicrograph displaying the sharp basal contact between the calcareous arkose bed (right) and phyllosilicate-rich bed that displays S_1 (yellow dashed line). (f) A back-scattered electron image of folded and bent muscovite (dark lamellae)-chlorite (light lamellae) stacks truncated by S_1 folia. Bedding (S_0) is sub-horizontal.

[Figure 7]. Textures from the CSZ in which S_1 cleavage is well preserved. (a-d are from sample K-F). (a) A plane-polarized light photomicrograph showing S_2 cleavage (sub-vertical) crenulating S_1 cleavage (dipping to the lower right). (b) A back-scattered electron image of a microlithon (ML) of a broken calcareous layer (BC) surrounded by P-domains defined by S_2 muscovite. Within the microlithon and the low strain areas, crenulated S_1 cleavage is dominant. Notice that the pyrite (Py) shape is elongated in S_2 folia and is euhedral in the low-strain microlithon. (c) A back-scattered electron image of another microlithon bracketed by strongly developed P-domains defining S_2 . (d) Enlarged view of the area in (b), showing strongly crenulated S_1 muscovite truncated by S_2 muscovite. (e) A back-scattered electron image of a well-developed S_2 fabric with few sigmoidal shaped microlithons displaying S_1 fabric in sample 19B. Chlorite is concentrated in the pressure shadow. (f) A crossed-polarized light photomicrograph displays detrital muscovite (DMs) oriented parallel to bedding in sample K-A. (g) A back-scattered electron image of a silty layer in sample K-B. The image shows many textures including small random laths of chlorite (circled), sutured albite boundaries (Ab), and S_1 folia of recrystallized muscovite (RMs) and chlorite defining a weak foliation sub-parallel to the bedding (S_0).

[Figure 8]. Back-scattered electron images of textures in the biotite zone (sample 19C (a and b), sample 30-6A (c) and the garnet zone (d from sample 19D). (a) A randomly oriented biotite replacing phyllosilicates in the P-domains separated by QF-domains (sample 19C). (b) Muscovite-rich layer with biotite flakes that are both random and sub-parallel to S_2 . Two composite grains of intergrown quartz and calcite are surrounded by biotite. (c) S_1 foliation is crenulated and truncated by S_2 foliation. Biotite flakes occur

parallel to S_2 . (d) Garnet porphyroblasts with inclusion trails (indicated by the pink line) consistent with S_2 and crenulated S_1 foliations observed in the matrix.

[Figure 9]. Age spectra for matrices of five samples from the regionally deformed rocks with S_1 cleavage only (a to e) and seven samples from the Chester shear zone (f to l). Minimum and maximum age steps are labeled. Asterisk (*) indicates steps with low K/Cl ratios. K/Ca and K/Cl scales are logarithmic (0.1-100).

[Figure 10]. T-t diagram summarizing the cooling and tectonic history of the Chester shear zone in the Lincoln area, as established by geochronologic data from this study. Vertical bars reflect the age and closure temperatures of amphibole, muscovite, biotite, and K-feldspar with errors. Time intervals of orogenies are from van Staal and Barr (2012).

[Table 1]. Summary of Paleozoic orogenies in the northern Appalachians (from van Staal and Barr, 2012).

[Table 2]. Modal composition (in wt.%) measured by XRD analysis and Rietveld refinement. The weight percentages are rounded to nearest integer.

Orogeny	Time	Cause	Notes
Taconic	Late Cambrian-Late Ordovician (500-450 Ma)	van Staal et al. (2007) redefined the Taconic orogenic cycle to comprise subduction and deformation related to accretion of oceanic and continental arcs in the peri-Laurentian realm.	-Includes three events: Taconic 1, Taconic 2, and Taconic 3.
Penobscot	Cambrian-Ordovician (486-479 Ma)	Closure of the Penobscot back-arc basin and obduction of back-arc ophiolites onto the Gander margin.	-Overlaps with early phases of Taconic 2, but restricted to Ganderian rocks.
Salinic	Silurian (445-422 Ma)	Collision between Gander margin and composite Laurentia following the terminal closure of an intervening, wide backarc basin.	- Pre-collisional deformation (445-430 Ma). -Collision-related orogenesis (430-422 Ma). -Sinistral oblique convergence.
Acadian	Late Silurian-Early Devonian (421-400 Ma)	Closure of the Acadian seaway by collision of Avalonia with composite Laurentia.	-Early stages overlap in time, although not in space, with the waning stages of Salinic orogenesis in parts of Ganderia. - Bradley et al. (2000) proposed that the migrating retro-arc deformation front and associated foreland basin continued to ~380 Ma and advanced into the Laurentian hinterland at a time when Meguma had already begun to collide with Laurentia. - Involved a combination of orogen-normal shortening and dextral movements along orogen-parallel structures.
Neoacadian	Late Early Devonian – Early Carboniferous (400-350 Ma).	Docking and collision of the Meguma terrane to composite Laurentia.	-The dominant structural style is characterized by a regional set of NW-trending shallowly plunging upright folds and associated shear zones.
Alleghanian	Mississippian to Permian (340-260 Ma)	Closure of Rheic ocean and amalgamation/collisions of composite Laurentia with the main mass of Gondwana, which led to the formation of Pangea.	-The Rheic ocean was closed in a zipper-like fashion with southwest-directed motion and clockwise rotation of Gondwana.

<i>Modal mineralogy of S₁ samples</i>											
<i>Sample</i>	<i>Formation</i>	<i>Ab</i> ¹	<i>Qtz</i>	<i>Chl</i>	<i>Ms</i>	<i>Carbonates</i>	<i>Sulfides</i>	<i>Bt</i>	<i>Oli</i>	<i>Others</i> [†]	<i>Total</i>
19F (WR²)	Madrid	22	46		20	10 (Ank)	2				100
19G (WR)	Madrid	16	54	8	12	9 (Cal+Dol)				1	100
19H (WR)	Carrabassett	17	30	21	30	1 (Cal)				1	100
19I (WR)	Madrid	22	46	9	19	2 (Cal)				1	100
19I (Sep³)	Madrid	21	45	11	22					1	100
K-C (WR)	Madrid	15	32	20	32	1 (Cal)					100
K-C* (Sep)	Madrid	15	17	20	46					1	100
<i>Modal mineralogy of the CSZ samples</i>											
19B (WR)	Smyrna Mills	10	5	9	74					2	100
19B (Sep)	Smyrna Mills	10	5	9	74					2	100
19C (WR)	Smyrna Mills	6	33	8	12	3 (Cal)		21	16		100
19D (WR)	Smyrna Mills	14	27	16	30			11		2 (Grt)	100
19E (WR)	Smyrna Mills	17	26	8	44					3	99
K-A* (WR)	Smyrna Mills	25	47	10	17					1	100
K-A (Sep)	Smyrna Mills	13	6	4	78					0	100
K-B (WR)	Madrid	17	33	11	17	20 (Cal)				1	100
K-B (Sep)	Madrid	12	14	19	54					1	100
K-D* (WR)	Smyrna Mills	29	25	22	23					1	100
K-D (Sep)	Smyrna Mills	16	2	24	58					0	100
K-E (WR)	Smyrna Mills	21	38	9	26	5 (Cal+Dol)					99
K-E (Sep)	Smyrna Mills	11	14	12	60		1			2	100
K-F* (WR)	Smyrna Mills	10	24	12	35	17 (Cal)	1			1	100
K-F* (Sep)	Smyrna Mills	9	23	28	37					3	99
30-6A (WR)	Smyrna Mills	5	32	10	40			8	3	2	100

¹ Abbreviations: Ab: albite; Ank: ankerite; Bt: biotite; Cal: calcite; Chl: chlorite; Dol: dolomite; Grt: garnet; Ms: muscovite; Oli: oligoclase plagioclase; Qtz: quartz.

² WR: Whole-rock samples.

Sep: Separates prepared for ⁴⁰Ar/³⁹Ar.

[†] Includes accessory minerals of low percentage ≤ 2% (e.g., rutile, apatite, ilmenite, or K-feldspar).

* Samples with trace amounts of K-feldspar coexisting with chlorite.