Late Ottawan orogenic collapse of the Adirondacks in the Grenville province of New York State (USA): Integrated petrologic, geochronologic, and structural analysis of the Diana Complex in the southern Carthage-Colton mylonite zone

Graham B. Baird
Department of Earth and Atmospheric Sciences, University of Northern Colorado, Campus Box 100, Greeley, Colorado 80639, USA

ABSTRACT

Crustal-scale shear zones can be highly important but complicated orogenic structures, therefore they must be studied in detail along their entire length. The Carthage-Colton mylonite zone (CCMZ) is one such shear zone in the northwestern Adirondacks of northern New York State (USA), part of the Mesoproterozoic Grenville province. The southern CCMZ is contained within the Diana Complex, and geochemistry and U-Pb zircon geochronology demonstrate that the Diana Complex is expansive and collectively crystallized at 1164.3 ± 6.2 Ma.

Major ductile structures within the CCMZ and Diana Complex include a northwest-dipping penetrative regional mylonitic foliation with north-trending lineation that bisects a conjugate set of mesoscale ductile shear zones. These ductile structures formed from the same 1060–1050 Ma pure shear transition-to-a top-to-the-SSE shearing event at ~700 °C. Other important structures include a ductile fault and breccia zones. The ductile fault formed immediately following the major ductile structures, while the breccia zones may have formed at ca. 945 Ma in greenschist facies conditions.

Two models can explain the studied structures and other regional observations. Model 1 postulates that the CCMZ is an Ottawan orogeny (1030–1035 Ma) thrust, which was later reactivated as a tectonic collapse structure. Model 2, the preferred model, postulates that the CCMZ initially formed as a subhorizontal mid-crustal mylonite zone during collapse of the Ottawan orogen. With continued collapse, a metamorphic core complex formed and the CCMZ was rotated into its current orientation and overprinted with other structures.

INTRODUCTION

Crustal-scale shear zones are fundamental features of orogens. They allow juxtaposition of different terranes or crustal blocks, can be important fluid conduits, and accommodate minor to significant crustal deformation of any or/multiple kinematics (e.g., Hanmer and McEachern, 1992; van der Pluijm et al., 1994; Mahan et al., 2006). Deciphering their tectonic history can be exceedingly difficult because varying and commonly repeating tectonic processes localized along shear zones can eradicate the evidence needed to interpret their history. The Carthage-Colton mylonite zone (CCMZ; Geraghty et al., 1981; Fig. 1) is one such fundamental crustal-scale shear zone. Located in the northwestern Adirondacks (northern New York State, USA), a southern exposure of the Mesoproterozoic Grenville province (e.g., Rivers, 2008; McLelland et al., 2010), the complete history and significance of the CCMZ remains unclear. The structure's later history does include tectonic collapse at the end of the Ottawan orogeny (ca. 1060–1030 Ma; Selleck et al., 2005) and exhumation of the Adirondack Highlands, the terrane to the southeast of the zone (Fig. 1). However, a number of key zone characteristics are not yet understood in this extensional model for the CCMZ. These include: (1) the lack of an obvious lithologic discontinuity in sections of the zone; (2) the identification of sinistral oblique reverse–sensed kinematic indicators associated with mylonitic fabrics in the south-central part of the zone (Baird, 2006); (3) transcurrent Ottawan-aged ductile fabrics in the central part of the zone (Johnson et al., 2004); (4) the apparently folded and steep character of the zone in its northern half (Fig. 1B; Johnson et al., 2004); and (5) the zone's relationship to the region's cooling pattern and other structures inside and outside of the zone (Fig. 1C).

This contribution extends the work presented in Baird (2006) and tests whether a lithological discontinuity exists across a portion of the zone. Further, this work fleshes out aspects of a rotated detachment model for the CCMZ presented in Baird (2008) and incorporates the recent high-quality microstructural and isotopic results of Bonamici et al. (2011, 2014, 2015). The tectonic history of the adjacent terranes is also considered, resulting in the construction of tectonic models that propose solutions to any apparent discrepancies with variable Ottawan-aged kinematic indicators identified along the structure. This work further emphasizes the complexity of crustal-scale shear zones and the need to study such zones by integration of structural, petrologic, and geochronologic data sets along their entire length.

GEOLOGIC BACKGROUND

The Grenville province, broadly described as a series of terranes separated by complex crustal-scale shear zones (e.g., Rivers et al., 1989), occurs in northern New York State as the Adirondacks
In general, the Adirondack segment of the Grenville province formed by protracted accretionary, collisional, and magmatic events ranging in age from ca. 1.35 Ga to ca. 0.98 Ga (Rivers, 2008; McLelland et al., 2010). However, recent work within the Grenville province also emphasizes the importance of tectonic collapse late in its history, as exemplified by work in the Adirondacks, the Morin terrane, the east side of the Mékinac-Taureau domain, and the Ottawa River Gneiss Complex (Selleck et al., 2005; Wong et al., 2011; Rivers and Schwerdtner, 2015; Soucy La Roche et al., 2015; Schwerdtner et al., 2016; Dufréchou, 2017; Regan et al., 2019).

Two terranes are exposed in the Adirondacks, the Adirondack Lowlands (Lowlands) and the Adirondack Highlands (Highlands), which are separated by the CCMZ (Fig. 1B). The geology of the Lowlands includes upper amphibolite grade (~640–680 °C, –6.0–7.8 kbar; Bohlen et al., 1985; Kitchen and Valley, 1995), complexly folded metasedimentary rocks and some metaigneous and igneous rocks (e.g., McLelland and Isachsen, 1986). Many ductile shear zones exist throughout the Lowlands, but little detailed work has focused on these structures, so nearly all have unknown deformation histories (Carl and deLorraine, 1997; Baird and Shrady, 2011). Highlands geology is generally similar to that of the Lowlands but includes a higher, granulite facies metamorphic grade (~675–850 °C, ~6.5–9.0 kbar; Bohlen et al., 1985; Kitchen and Valley, 1995; Storm and Spear, 2005) and a higher proportion of metaigneous and igneous rocks relative to metasedimentary rocks (e.g., McLelland and Isachsen, 1986). Multiple generations of folding are also evident in the Highlands and produce the terrane’s dominant structural grain that follows major fold axial traces (McLelland and Isachsen, 1980; Fig. 1B).

A tectonic model outlining Adirondack geology is provided by McLelland et al. (2010; 2013, and references therein). Major events include magmatism, metamorphism, and deformation associated with the Shawinigan orogeny, which was driven by the closure of a complex backarc basin now found as forming the Lowlands and perhaps portions of the Highlands (ca. 1200–1150 Ma; Chiarenzelli et al., 2010; Baird and Shrady, 2011). Immediately following and partially overlapping the Shawinigan events was widespread intrusion of the amphiromite-mangerite-charnockite-granite magmas (AMCG suite) thought to have been created by mantle and crustal partial melting initiated by delamination or slab breakoff (ca. 1160–1145 Ma; McLelland et al., 1996, 2004; Regan et al., 2011). Based on similarity of deformation style between portions of the CCMZ and the Lowlands, some argue that Shawinigan-aged thrusting may have occurred along the CCMZ (Baird and Shrady, 2011; McLelland et al., 2013). Subsequently, the collision of Amazonia with Laurentia produced the Ottawan orogeny (ca. 1090–1035 Ma; McLelland et al., 2001). This event only affected the Adirondack Highlands, because the Adirondack Lowlands were part of an orogenic
lid, or a region shallower in the crust that escapes later high-grade metamorphism, magmatism, and ductile deformation experienced by lower-crustal rocks such as in the Highlands (Fig. 1; Rivers, 2008). Recognition of the Lowlands as part of an orogenic lid emphasizes that despite the apparent similarity of geology found in the Lowlands and Highlands, the difference in tectonic history suggests that appropriate correlation of pressure-temperature conditions and deformation features between the two terranes always requires temporal constraints. Orogenic collapse of the Ottawa orogen occurred synchronously with intrusion of the Lyon Mountain Granite Gneiss into the Adirondack Highlands and extension along both the CCMZ and East Adirondack shear zone at ca. 1060–1030 Ma (Selleck et al., 2005; Bickford et al., 2008; Wong et al., 2011; Chirenczelli et al., 2017; Fig. 1B).

Metamorphic Core Complexes: Geological Overview

The Adirondacks may be part of a deeply eroded symmetric metamorphic core complex, as the Highlands are bound by the generally northwest-dipping normal-sensed CCMZ and the °20° southeast-dipping normal-sensed East Adirondack shear zone (Isachsen and Geraghty, 1986; Bickford et al., 2008; Wong et al., 2011). In order to fully appreciate major features, characteristics, and processes associated with metamorphic core complexes and how they might fit with the deformational and metamorphic history of the CCMZ presented here, it is worthwhile to provide an outline of the relevant features associated with crustal extension and metamorphic core complex generation. Many sources provide excellent reviews and detailed studies of metamorphic core complexes (e.g., Lister and Davis, 1989; Whitney et al., 2013; Platt et al., 2014), which are a widespread crustal structure and have formed throughout much of Earth history (e.g., Whitney et al., 2013).

A general model for metamorphic core complex formation includes an initial period of crustal thickening. This thickened crust becomes gravitationally unstable as the lower crust becomes weak in a few tens of millions of years due to heating and partial melting (e.g., Vanderhaeghe and Teyssier, 1997). The gravitational instability is alleviated by brittle extension in the upper crust and exhumation of ductile lower crust, which in turn may generate (more) melt through decompression (Teyssier and Whitney, 2002).

The characteristic feature of metamorphic core complexes is generally dome-shaped ductile lower-crustal rocks overlain by an extension-dominated brittle to ductile shear zone, commonly called the detachment. Above the detachment is normal-faulted upper crust (e.g., Whitney et al., 2013). Though metamorphic core complexes can be symmetric, as proposed for the Adirondacks (Isachsen and Geraghty, 1986; Bickford et al., 2008; Wong et al., 2011), most are asymmetric with one dominant detachment. If the lower crust was migmatitic at the time of exhumation, the lower crust would commonly have granitic plutons and complicated foliation patterns defining multiple sub-domes (e.g., Whitney et al., 2004, 2013).

The lower crust may also include a mid-crustal mylonite zone that is not the detachment; however, this mylonite zone is warped over and through the exhumed dome, merging with and overprinted by the detachment on one side of the core complex (e.g., Lister and Davis, 1988; Cooper et al., 2010; Platt et al., 2014). This mid-crustal mylonite zone is the ductile sole of upper-crustal normal faults produced by extension preceding doming. Lister and Davis (1989) termed this zone the “mylonite front,” while Cooper et al. (2010) called this zone the “localized-distributed transition.” Depending on the geometry and history of the normal faults that sole into this zone, both kinematics and total displacement can be quite variable along the length of the mid-crustal mylonite zone (Cooper et al., 2010).

The Adirondack Highlands and other metamorphic core complex examples proposed for the Grenville province do differ from the classic Mesozoic examples of the western United States. Most obvious is the greater size of the Highlands in comparison. This difference could be simply due to deeper exposure of the Proterozoic complexes producing a greater aerial size. Rivers (2011) also argues that Proterozoic orogens, the Grenville province in particular, were much hotter compared to more recent examples and resulted in a more ductile lower crust. This would have led to greater vertical thinning and extension and more complicated strain patterns upon collapse.

The Carthage-Colton Mylonite Zone

Geraghty et al. (1981) defined the CCMZ as a zone of high strain and mylonitic textures in the northwestern Adirondacks. Mezger et al. (1991, 1992) first demonstrated the approximate 100 m.y. difference in U-Pb sphene cooling ages on either side of the CCMZ in the Highlands (ca. 1050–990 Ma) and Lowlands (ca. 1150–1090 Ma). This, combined with the impossibly high thermal gradients found across the zone, indicates that the CCMZ is a terrane boundary and experienced one or more kilometers to tens of kilometers of shearing events across it (Mezger et al., 1992). Other workers, using different isotopic and minerals systems (Streepey et al., 2000, 2004; Dahl et al., 2004), defined separate cooling paths for the Highlands and Lowlands, which helped shape tectonic models explaining that through CCMZ shearing, both terranes experienced Shawinigan tectionism, while only the Highlands experienced Ottawa tectonism.

Figure 1C summarizes how the 40Ar/39Ar hornblende age young to the southeast across the Lowlands from ca. 1100 Ma, to a range of ages (ca. 1025–925 Ma) within the CCMZ, and reaching an approximately consistent young age of ca. 950–900 Ma in the Highlands. These data suggest that the Lowlands were either tilted to the northwest after cooling, or cooled toward the southeast via inclined isotherms. Further, the Highlands were exhumed by one or more episodes of normal-sensed motion along the CCMZ in the ca. 1100–940 Ma interval (Dahl et al., 2004).

Streepey et al. (2001) and Johnson et al. (2004, 2005) were the first to provide complete structural, petrologic, and geochronologic analysis of the Dana Hill metagabbro segment of the CCMZ (Fig. 1B). Results indicate that early ductile structures resulted from dextral strike-slip motion under –740 ± 30 °C conditions at ca. 1090–1050 Ma. This was followed by normal-sensed dip-slip deformation at 1050–1020 Ma during fluid infiltration.
at ~700 °C. Deformation ended in the Dana Hill metagabbro with brittle deformation at greenschist facies conditions sometime during 1000–930 Ma.

The Diana Complex occurs along the southwestern CCMZ and, as mapped by Hargraves (1969), extends across the zone and into both the Lowlands and the Highlands. Therefore, it is unclear whether the complex is a Lowlands or a Highlands unit, and no recent work has investigated whether the currently identified Diana Complex is indeed a single magmatic complex. If large displacements occurred across the CCMZ, a lithological discontinuity should exist along all portions of the CCMZ, but this has not been tested in the Diana Complex.

The complexity of deformation and resulting implications within part of the Diana Complex segment of the CCMZ were considered by Baird and MacDonald (2004), while Bonamici et al. (2011, 2014, 2015) provided a robust microstructural and isotopic analysis of spinel from two expansive outcrops in this area (“east,” here called H1; and “west,” Fig. 2). Results included identification of two shear fabrics that crosscut a penetrative deformation fabric. These shear fabrics formed under rapidly cooling conditions from ~700 to 500 °C in a few million years at ca. 1050 Ma. All fabrics were presumed to have formed during peak Otawan tectonism and subsequent collapse and extension along the CCMZ (Bonamici et al., 2015). The following work focuses on a portion of the Diana Complex, including those outcrops studied by Bonamici et al. (2011, 2014, 2015). Because this research area fully contains the CCMZ (Fig. 1), it allows integration of this work’s results with 40Ar/39Ar thermochronology, U-Pb titanite geochronology, metamorphic petrologic investigations, and detailed structural analysis data sets published for portions of the CCMZ and across the Adirondacks.

PETROLOGY AND U-Pb ZIRCON GEOCHRONOLOGY

The Diana Complex has long been identified as a deformed and metamorphosed granitoid member of the AMCG suite (McLelland et al., 1996). The northwestern contact between the Diana Complex and typical marbles, quartzites, and related rocks of the Lowlands is a multiply folded intrusive contact (Wiener, 1983; Fig. 2). Wiener (1983) suggested that the Diana Complex intruded synchronously with the second folding event identified in the Lowlands adjacent to the complex. Hargraves (1969), building on the work of Buddington (1939), identified two main lithologies for the Diana Complex based on the dominant accessory mineralogy: either predominately pyroxene-bearing or predominately hornblende-bearing. Pyroxene-bearing units tend to be green in outcrop while hornblende-bearing rocks tend to be pink. Cartwright et al. (1993) and Lamb (1993) collectively reported that clinopyroxene, orthopyroxene, and hornblende are in variable proportions at most locations within the Diana Complex.

Smyth and Buddington (1926) described the Diana Complex rocks as syenitic because other workers of the time provided such terminology when describing similar rocks throughout the Adirondacks. This classification is nearly universally used today but is not without issues given the reintegrated ternary composition (Fuhrman et al., 1988; Lamb, 1993) of the mesoperthite that currently dominates the rock. Other similar pyroxene-bearing rocks are classified, depending on technique, as mangerite-charnockite, monzonite, or monzosyenite (Fuhrman et al., 1988).

To the southeast of the complex is a coarse hornblende granite gneiss, termed the Lowville granite by Buddington (1939), which has been identified as a member of the ca. 1100 Ma Hawkeye Granite Gneiss suite (e.g., Chiairenzelli and McLelland, 1991; McLelland et al., 1996). However, the age of this suite has been questioned (Walsh et al., 2016). Also on the southeastern border of the complex is a body of anorthositic gabbro, plus a hornblende-and biotite-bearing aplitic granite gneiss that locally appears to grade into the Diana Complex (Buddington, 1939; Fig. 2).

Geochemistry

To help evaluate the relationships between the granitoids in the study area, and investigate the possibility for a lithological discontinuity related to offsets across the CCMZ after AMCG suite intrusion, bulk-rock geochemistry of one representative sample each of pyroxene-bearing and hornblende-bearing Diana Complex units and of the apatitic granite gneiss was studied. Rock analysis was performed by the mineral laboratories division of Bureau Veritas Mineral Laboratories (Acme Labs; Vancouver, British Columbia, Canada) (Table 1). Analysis procedure included pulverization of ~650 g of rock, followed by lithium borate fusion, acid digestion, and analysis by inductively coupled plasma–mass spectrometry (trace elements) and inductively coupled plasma–optical emission spectrometry (major elements).

Normative analysis of these analyses and of additional Diana Complex bulk chemical analyses presented by Smyth and Buddington (1926), Buddington (1939), and Hamilton et al. (2004) were alternatively calculated by a Microsoft Excel spreadsheet constructed by K. Hollacher (http://minerva.union.edu/hollochk/c_petrology/other_files/norm4.xlsx) following the routine of Johannsen (1931). Results are presented in Figure 3A. For consistency, all analyses shown in Figure 3A assumed only Fe2+ was present. Some analyses reported Fe3+ content; inclusion of Fe3+ in the normative analysis produced results essentially identical to those presented here.

Using mineral norms with International Union of Geological Sciences (IUGS) classification provides a means by which to compare bulk-rock chemistry to the X-ray diffraction–determined modes presented by Hargraves (1969). Buddington (1939) compared modal and normative analyses for these rocks and found them to be very comparable. Revealed by the normative analysis presented in Figure 3A is that the most mafic, pyroxene-bearing members of Diana Complex can be classified generally as monzodiorite to monzonite. The most felsic, hornblende-bearing rocks are generally quartz monzonite to granite. The apatitic granite gneiss normative analysis plots with the felsic members of the Diana Complex. Lithological variation between these rocks is mostly caused by the indirect relationship between plagioclase and quartz at a generally consistent alkali feldspar content (Hargraves, 1969). The studied rocks all plot with other AMCG crustal melt rocks (MCG—mangerite-charnockite-granite).
Figure 2. Geologic map of the study area, Adirondacks of northern New York (USA). "West" and "east" refer to outcrops focused on by Bonamici et al. (2011, 2014, 2015). Map is based on Hargraves (1969), Isachsen and Fisher (1970), Geraghty et al. (1981), Wiener (1983), and this research. Site locations referred to in the text and figures are underlined. Geographic reference frame is UTM zone 18T, NAD27 datum, in thousands of meters. CCMZ—Carthage-Colton mylonite zone.
### U-Pb Zircon Geochronology

The Diana Complex was previously dated via U-Pb zircon geochronology, and results provide magmatic crystalization ages of 1155 ± 4 Ma (thermal ionization mass spectrometry [TIMS]; Greely et al., 1986), 1118 ± 3 Ma (single- and multi-grain TIMS; Basu and Premo, 2001), and 1164 ± 11 Ma (sensitive high-resolution ion microprobe [SHRIMP]; Hamilton et al., 2004). The two results from TIMS work are questionable due to reverse discordance, excessive data scatter and Pb loss, and the identification of multiple zircon growth domains. Therefore, the SHRIMP-obtained age of Hamilton et al. (2004) from the H1 outcrop is likely the most robust, but given the geographic extent and lithological diversity within the Diana Complex and neighboring rocks of igneous origin, additional magmatic ages across the complex will help confirm the extent of the complex.

Zircon was separated from a sample each of the hornblende-bearing and the aplitic granite gneiss rocks through standard rock crushing and magnetic and density separation procedures done at the University of Northern Colorado (Greeley, Colorado, USA) (see Baird et al., 2014). Zircons isolated from both samples were typically transparent, light yellow to brown, euhedral, stubby to elongate grains, ~200–400 microns in length, with the aplite sample having somewhat smaller grains on average. Picked grains were mounted in epoxy, polished to expose cross-sections, and imaged with cathodoluminescence (CL) to assess internal zoning patterns. For both samples, nearly all grains contain oscillatory zoning consistent with magmatic crystallization. A few grains possess dark featureless rims thick enough for analysis.

Uranium-lead (U-Pb) analysis of homogenous zones free of inclusions and cracks based on CL pattern was completed with the SHRIMP−reverse geometry (SHRIMP-RG) instrument at Stanford University (Stanford, California, USA) using standard laboratory procedures (see Baird and Shrady, 2011, and references therein). The spot size was ~21 μm, and data reduction, analysis, and presentation utilized programs based on Ludwig (2009, 2012). Table 2 contains the results of the SHRIMP-RG analysis. All presented concordia diagrams display 1σ error ellipses, while 2σ Pb/Pb age plots display 2σ errors. Regression lines were calculated by York’s algorithm (default method) and includes decay-constant errors. Reported aggregated ages are reported at 95% confidence weighted by assigned errors; also reported is the mean square of weighted deviates (MSWD).

#### Sample CCMZ-J1

Sample CCMZ-J1 is a typical sample of a strongly foliated and lined hornblende-bearing Diana Complex rock (Table 1). U-Pb zircon data have a modest spread, with most analyses ranging

| Sample | CCMZ-J1 | CCMZ-C6 | CCMZ-812 |
|--------|---------|---------|---------|
| Lithology* | HGSG | PQSG | AGG |
| SiO₂ | 65.18 | 57.44 | 69.03 |
| TiO₂ | 1.15 | 1.37 | 0.46 |
| Al₂O₃ | 13.82 | 16.84 | 13.97 |
| Fe₂O₃ | 6.76 | 7.86 | 4.67 |
| MnO | 0.12 | 0.15 | 0.05 |
| MgO | 0.83 | 1.2 | 0.57 |
| CaO | 2.25 | 4.47 | 1.03 |
| Na₂O | 3.79 | 4.58 | 3.56 |
| K₂O | 4.91 | 4.43 | 5.1 |
| P₂O₅ | 0.36 | 0.55 | 0.08 |
| LOI | 0.50 | 0.7 | 1.2 |
| Total | 99.67 | 99.59 | 99.72 |
| Rb | 48.5 | 131.9 | 192.5 |
| Sr | 633.7 | 278 | 105.3 |
| Ba | 1534 | 795 | 440.3 |
| Ta | 0.4 | 1.1 | 1.4 |
| Nb | 13.5 | 22.1 | 28.7 |
| Pb | 1.2 | 6.1 | 3.4 |
| Hf | 19.2 | 21.3 | 18.8 |
| Zr | 887.8 | 899.6 | 741.2 |
| Y | 54.6 | 75.6 | 64.3 |
| Sc | 9 | 7 | 6 |
| Ni | 0.7 | 1 | 0.5 |
| Zn | 117 | 73 | 110 |
| Mo | 1.5 | 2.1 | 5.1 |
| Co | 7 | 7.7 | 2.6 |
| Sn | <1 | 3 | 2.0 |
| Sb | 0.4 | 0.3 | 0.2 |
| W | <0.5 | 1.4 | 0.6 |
| V | 19 | 49 | 9.0 |
| Cu | 3.3 | 7.8 | 2.9 |
| Th | 1 | 9 | 19.4 |
| U | 0.2 | 3 | 6.1 |
| Ga | 25.2 | 24.5 | 26.7 |
| La | 45.2 | 78.2 | 95.0 |
| Ce | 106 | 165.7 | 193.4 |
| Pr | 14.57 | 21.12 | 22.44 |
| Nd | 63.5 | 85.9 | 84.20 |
| Sm | 13.44 | 16.35 | 15.28 |
| Eu | 4.4 | 2.97 | 1.77 |
| Gd | 12.28 | 14.24 | 13.49 |
| Tb | 1.97 | 2.31 | 2.13 |
| Dy | 10.49 | 13.53 | 11.68 |
| Ho | 2.1 | 2.75 | 2.33 |
| Er | 6.05 | 8.24 | 6.68 |
| Tm | 0.8 | 1.23 | 1.00 |
| Yb | 5.43 | 7.82 | 6.53 |
| Lu | 0.82 | 1.24 | 0.99 |

*Note: See Figure 2 for sample locations. Oxides are in weight %; elements are in ppm. UTM zone 18T NAD 27 easting and northing in meters: CCMZ-J1 471073, 4869121; CCMZ-C6 47104, 4884166; CCMZ-812 471400, 4876202. 
†As mapped by Hargraves (1969). HGSG—hornblende quartz syenite gneiss, PQSG—pyroxene quartz syenite gneiss, AGG—aplite granite gneiss. 
‡All Fe reported as Fe₂O₃. 
§LOI—loss on ignition.
Figure 3. Diana Complex geochemical data; symbols for all figure parts are found in A. (A) Compilation of Diana Complex powdered X-ray mode data and normative analysis of multiple bulk chemistry data sets shown on the International Union of Geological Sciences phaneritic rock classification diagram (after Le Maitre, 2002). “MCGs” include the mangerite-charnockite-granite-type rocks near the Marcy massif anorthosite complex in the Adirondack Highlands (Fig. 1). Q—normalized quartz mode; A—normalized alkali feldspar mode; P—normalized plagioclase feldspar mode; q—quartzolite; qg—quartz-rich granitoid; ag—alkali feldspar granite; g—granite; gd—granodiorite; t—tonalite; qas—quartz-alkali feldspar syenite; qs—quartz syenite; qmd/qmg—quartz monzodiorite and quartz monzogabbro; qd/qg/qa—quartz diorite, quartz gabbro, and quartz anorthosite; as—alkali feldspar syenite; s—syenite; m—monzonite; md/mg—monzodiorite and monzogabbro; d/g/a—diorite, gabbro, and anorthosite. (B) Alkalis versus silica diagram of Le Maitre (2002). (C) Ferroan-magnesian diagram of Frost et al. (2001). FeO* is all iron as FeO. (D) Harker-type diagrams. (E) Alkalis-FeO*-MgO (AFM) diagram (Irvine and Baragar, 1971). (F) Rare earth element plot normalized to chondrites (Sun and McDonough, 1989).
from reversely discordant with 206Pb/207Pb ages in the 900 Ma range, through to concordant or near-concordant 206Pb/238U ages of ca. 1150 Ma (Fig. 4). A few analyses were rejected due to reverse discordance, large errors, or the analysis being an outlier. The resulting nine, tightly clustered 206Pb/207Pb age analyses are concordant to near concordant and produce a chord with lower and upper intercepts of 118 ± 620 Ma and 1161 ± 23 Ma, respectively. Because the lower intercept, within error, includes the origin, the 206Pb/238U weighted mean age of 1158 ± 6 Ma (MSWD = 0.28; Fig. 4A inset) from these nine analyses is interpreted to be the best estimate of the igneous crystallization age of this rock. Dark CL rims observed on some zircons do not appear to possess an age different from that of brighter oscillatory-zoned zircon (Fig. 4). Rare inherited cores were not large enough to allow analysis.

Sample CCMZ-812

Sample CCMZ-812 is a typical sample of a moderately foliated aplitic granite gneiss found south of...
the rocks classically considered part of the Diana Complex (Hargraves, 1969; Table 1). Most U-Pb data cluster near concordia at ca. 1150 Ma, but a handful of analyses are quite discordant or provide a much younger 207Pb/206Pb age (Fig. 4B). Fitting a regression line through the seven analyses with the most overlap in 207Pb/206Pb age produces a chord with lower and upper intercepts of 20 ± 550 Ma and 1182 ± 36 Ma. With the lower intercept of the chord essentially zero, the 207Pb/206Pb weighted mean of 1181 ± 25 Ma (MSWD = 0.30; Fig. 4B inset) from these seven analyses is considered to be the best estimate of the igneous crystallization age for this rock. No conclusive inherited cores were analyzed, and dark CL rims are commonly highly discordant and do not appear to provide a coherent age, likely due to radiation damage and significant Pb loss.

**Aggregate Age**

Results for both rocks dated here and that of Hamilton et al. (2004) are in good agreement collectively on a concordia diagram (Fig. 5), so all analyses from the three rocks were used to calculate an aggregate age for the Diana Complex. The combined 27 analyses confirm 1164.3 ± 6.2 Ma as the age of the Diana Complex as a whole, including the aplitic granite gneiss.

**Petrogenesis**

The bulk chemistry and U-Pb zircon geochronology work presented here emphasizes the petrologic relationship of the aplitic rock and the rocks classically identified as belonging to the Diana Complex. All of these rocks are chemically similar to other granitoid members of the AMCG suite in the Highlands (Fig. 3).

Hargraves (1969) and Buddington (1939) both envisioned that the Diana Complex formed from a large, at least 6 km thick, stratified sheet-like intrusion whereby the pyroxene-bearing and hornblende-bearing units are related by fractional crystallization. With this model, Buddington (1939) believed the average Diana Complex composition

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**Figure 4.** (A) Concordia diagram for sample CCMZ-J1; inset is the 207Pb/206Pb ages for each analysis with the best estimate of the magmatic crystallization age (see text for details). Analysis symbols in gray were not used in any age calculation. (B) Concordia diagram for sample CCMZ-812; inset is the 207Pb/206Pb ages for each analysis with the best estimate of the magmatic crystallization age (see text for details). Analysis symbols in gray were not used in any age calculation. On the right are representative cathodoluminescence images with 207Pb/206Pb spot ages (in Ma); blue analysis locations are CCMZ-J1 zircons, red analysis locations are CCMZ-812 zircons. MSWD—mean square of weighted deviates.
characterized the transition between the coarse hornblende granite and hornblende-bearing rocks that are part of the complex as “imperceptible.” This together with the field observations of this work suggest that the coarse hornblende granite gneiss (Lowville granite) too is part of the Diana Complex (Fig. 2; cf. Smyth and Buddington, 1926) and suggests that the body is even larger than originally estimated.

Regardless of some of the finer details regarding the evolution of parental magma(s), this work demonstrates that the Diana Complex crystallized at ca. 1164 Ma and includes the aplitic granitic gneiss and probably the coarse hornblende granite. This result confirms that no major lithological discontinuity exists across the CCMZ within the study area and that the Diana Complex indeed shares affinities with both the Lowlands and the Highlands through intrusive, geochronologic, and petrological relationships. The Diana Complex is a spatially extensive magmatic suite, and this likely allowed the postulated kilometer- to tens of kilometers-scale displacements to occur across the CCMZ without producing a lithological discontinuity.

## STRUCTURES

### Magmatic Fabric

Original magmatic structure has largely been reoriented by subsequent deformation (e.g., Hargraves, 1969), however in a few locations the aplitic granite gneiss does show layering that lacks any associated grain-shape or mineral-aggregate preferred alignment (Fig. 6). This indicates that this fabric is preserved magmatic layering. Orientation of such structures is essentially north-south, an orientation inconsistent with trends of deformational structures in the area.

### Regional Mylonitic Fabric

The regional mylonitic fabric (Hargraves, 1969; Geraghty et al., 1981; Baird and MacDonald, 2004; Fig. 2) in the Diana Complex is nearly ubiquitous within the CCMZ and is penetrative in outcrops throughout the field area. Outside the study area, this fabric appears to exist as an axial planar foliation to some Adirondack Lowlands folding (Wiener, 1983), but ductile deformation that postdates AMCG suite emplacement in most of the Lowlands has yet
to be conclusively demonstrated (Baird and Shrady, 2011). If it does exist, it may be restricted to areas bordering the CCMZ. To the southeast of the study area in this paper, preliminary work suggests that this fabric also at least weakly affects the rocks of the Highlands for some distance to the southeast.

Within the CCMZ study area, the regional mylonitic fabric does vary in intensity. At the CCMZ’s margins, this fabric is weak, whereas in the central CCMZ, this fabric is quite strong (Fig. 7). Though unrecrystallized feldspar porphyroclasts can be identified, all of the regional mylonitic fabric displays some amount of material that is finer-grained than the protolith owing to dynamic recrystallization and other processes (discussed below). Even in nearly fully recrystallized rocks, the boundaries of once-coherent feldspar grains are inferred from the distribution of mafic minerals and quartz that mantled the feldspar in the protolith.

Study area foliation typically dips ~30°–60° to the northwest, except immediately along the contact of the Diana Complex with the Lowlands metasedimentary rocks and in the extreme southeastern part of the field area (Figs. 2 and 8). Mineral and stretching lineations are typically oriented ~40°–350° and are best developed in areas with strong foliation (Fig. 8). Locations without a strong lineation commonly possess a well-developed S-C fabric.

The highest-strain regional mylonitic fabrics and associated microstructures can be mapped out as a band ~1–3 km wide within the central portion of the CCMZ (Fig. 2). This band was identified by A.F. Buddington (Smyth and Buddington, 1926, their map plate), and the northwestern boundary mapped here is nearly identical to that of Buddington, whereas the southwestern boundary is somewhat different owing to the more gradual transition moving out of the zone to the southwest. Thin sections taken from a traverse across the CCMZ show a progressive increase and then decrease in strain intensity as indicated by degree of recrystallization and strength of fabric (Fig. 9). Thin sections of the regional mylonitic foliation reveal two distinctly different sets of microstructures. The spatial transition between the sets is in some places abrupt and is generally in close proximity to the transition from pyroxene- and hornblende-bearing units and delineates the northwestern edge of the high-strain zone (Figs. 2 and 9). This transition separates rocks with characteristics typical of Diana Complex rocks adjacent to Lowlands paragneisses from those typical of Highlands granitoids to the southeast. For the purposes of this work, this transition is a convenient and reasonable feature taken as the Lowlands-Highlands boundary.

Figure 6. Magmatic folding of schlieren in the aplitic granite gneiss. The strike-dip symbol marked on a limb is nearly north-south, with a vertical dip, which is the approximate orientation of this fabric type.

Figure 7. Examples of Diana Complex regional mylonitic fabrics within the Carthage-Colton mylonite zone. All are subhorizontal surfaces with north approximately toward the upper right. (A) Moderately developed protomylonitic fabric in the southeastern part of the study area. (B) High-strain statically recrystallized mylonitic fabric in the central Carthage-Colton mylonite zone. (C) Weakly developed statically recrystallized mylonitic fabric in southeastern part of the study area.

Figure 10 summarizes the variation in percent feldspar and quartz neoblasts, average quartz grain diameter, and average feldspar grain diameter from >30 thin sections from across the CCMZ (Table 3). Plot distance was determined by the perpendicular distance from the Lowlands-Highlands boundary and the line that delineates the abrupt change in microstructural type (Fig. 2), with negative distances measured approximately to the northwest and positive distances being measured approximately to the southeast. Percent neoblasts correlates well with the observed fabric strength (Figs. 2 and 9; Table 3).

The northwestern microstructures, termed Lowlands microstructures, from northwest to southeast, show a profound increase in percent neoblasts such that the rocks range from protomylonite to ultramylonite (Passchier and Trouw, 2005; Figs. 9A, 9B, 9C, and 10). This change is accompanied by a modest increase in both quartz and feldspar neoblast...
Figure 8. Equal-area projection stereonets. (A) Structural pattern for the regional mylonitic fabric (after Baird, 2006). (B) Geometry of mesoscale ductile shear zones of the study area, not including the H1 outcrop; pole of best-fit plane is 45°±5° (after Baird, 2006). (C) Geometry of mesoscale ductile shear zones in the H1 outcrop; pole of best-fit plane is 64°±39°. Undeformed clinopyroxene-bearing hydrothermal vein orientations are also shown. (D) Ductile shear zone kinematics and lineation orientation (after Baird, 2006). For each lineation, the plane of the shear zone is traced by the line through the lineation symbol. (E) Ductile shear zone poles (n = 35) from locations that have definable regional lineation and foliation. All data have been rotated such that regional lineation (L) is horizontal east-west and regional foliation (S) is vertical east-west. In this view, regional shear is sinistral. Shear zone pole density is contoured with a 4% contour interval. Note the prominent conjugate set of shear zones, both oriented ~30° from the regional foliation, which bisects the acute angle of the set.
Figure 9. Full thin-section scans of the regional mylonitic fabric across the Carthage-Colton mylonite zone. A–E are ordered in an approximately north-to-south traverse across the Diana Complex (Fig. 2). All are cross-polarized light; each thin-section orientation is parallel to lineation and perpendicular to foliation, with foliation approximately horizontal in the thin section with the view to the northeast or east depending on local fabric. Location references are shown in Figure 2. A–C are examples of rocks with Lowlands microstructure, while D–E are examples of Highlands microstructure; see text for detail. (A) H1 location; note the prominent undulose extinction (examples marked by U) and the Fe-Ti oxide–filled tension gashes oriented upper right to lower left (examples marked by TG). (B) K3 location; protomylonitic S-C fabric with S surfaces oriented upper right to lower left and C surfaces oriented subhorizontally, indicating top-to-the-southeast transport. (C) WP7 location; well-developed mylonitic fabric with extensive recrystallization. (D) WP6 location; note the high strain and near-complete recrystallization with quartz forming prominent ribbons constructed of adjacent elongate grains (example marked by QR). Upper-central portion of the thin section displays a recrystallized S-type porphyroclast with asymmetry indicating top-to-the-SSE transport. (E) WP18 location; note the similar microstructures compared to D, but of lower strain overall. S-C fabric is present with S surfaces oriented upper right to lower left and C surfaces oriented subhorizontally, indicating top-to-the-SSE transport.
grain size. Additional characteristics of Lowlands microstructure include feldspar porphyroclasts with strong undulose extinction that commonly contain micro–tension gashes filled with Fe-Ti oxides (Baird, 2001; Figs. 9A, 9B, 9C, and 11). Feldspar porphyroclasts also commonly have narrow bands of neoblasts traversing them. Porphyroclast boundaries are always jagged and are surrounded by dynamically recrystallized quartz and feldspar neoblasts in a polygonal mosaic, some of which displays foam texture (Fig. 11). Individual neoblasts have uniform extinction, and quartz and feldspar neoblasts can be distinguished from each other because quartz grains are distinctly larger in size (Fig. 10) and are commonly strung out into ribbons. Quartz, except in rare low-strain rocks (e.g., the H1 outcrop), has completely recrystallized into equant polygonal grains (foam texture). In the low-strain rocks, relatively large relict quartz grains display undulose extinction and subgrains of similar size to the surrounding quartz neoblasts, indicating that subgrain rotation is the dominant recrystallization mechanism (Fig. 11; Passchier and Trouw, 2005; Trouw et al., 2010). Rarely, and typically near the high-strain zone, quartz will show signs of static recrystallization by high-temperature grain-boundary migration, as indicated by quartz ribbons being partially constructed of wide single grains (Fig. 11D; Passchier and Trouw, 2005; Trouw et al., 2010). Many feldspar porphyroclasts, particularly those in the furthest northwest, lack subgrains with the suturing of the grain boundaries equal in size to the surrounding feldspar neoblasts, indicating recrystallization by low-temperature grain-boundary migration, also called bulging (Fig. 11C; Passchier and Trouw, 2005; Trouw et al., 2010). In areas closer to the margin of the high-strain zone in the core of the CCMZ, feldspars do show some subgrains of equal size to the surrounding feldspar neoblasts, indicating recrystallization by subgrain rotation (Fig. 11E).

The microstructures of the highest-strain zone, extending to the southeastern edge of the Diana Complex, are distinctive and are termed Highlands microstructures (Figs. 9D, 9E, and 12). Except on the southeasternmost edge of the Diana Complex, these rocks reflect more complete recrystallization and are classified as ultramylonite with rare relict...
porphyroclasts (Fig. 10). The porphyroclasts that do exist typically have straight extinction. The surrounding feldspar neoblasts commonly display a relatively coarse grain size with equant polyg- nal grains (foam texture) and show signs of static recrystallization by recovery, high-temperature grain-boundary migration leading to grain-bound- ary area reduction, and likely grain-size coarsening. This is indicated by grain-size variability (Fig. 10), straight grain boundaries forming near-120° triple junctions, and larger grains displaying concave grain boundaries adjacent to smaller grains with convex boundaries (Fig. 12; Passchier and Trouw, 2005). Individual quartz grains have uniform extinction and always form ribbons constructed by a series of elongate grains, such that the ribbons commonly are only one grain thick wide (Fig. 12). This indicates that extensive high-temperature grain boundary migration occurred during deformation and/or sub- sequent to deformation (Trouw et al., 2010).

The characteristics that most clearly distinguish the two different microstructure types are that in the Lowlands type, the relict feldspar porphyro- clasts are highly strained as indicated by: strong undulose extinction, narrow bands of neoblasts traversing them, Fe-Ti oxide-filled micro–tension gashes, and highly sutured grain boundaries. This is in contrast to the feldspar porphyroclasts in the Highlands microstructure, which are more recrystal- lized, with the neoblasts showing evidence for grain-boundary area reduction. Other differences, though not as universal, are the much larger grain size for both feldspar and quartz in the Highlands microstructure (Fig. 10) and obvious evidence for high-temperature grain-boundary migration dis- played by quartz in the ribbons.

**Kinematic Indicators**

Kinematic indicators associated with the regional mylonitic fabric are found in multiple locations. The Lowlands microstructures display S-C fabrics indicating top-to-the-SSE kinematics in a number of locations (Figs. 9B and 13A). In other localities, well-defined ς-type or more rarely δ-type kinematic indicators are observed. However, common are weaker fabrics with poorly developed or no kinematics indicators. S-C fabrics, C-C′-type or more rarely δ-type kinematic indicators are common in Highlands microstructure rocks, and these show the same top-to-the-SSE transport determined from the S-C fabrics in the northwestern part of the study area (Figs. 9D, 9E, 13B, and 13C).

**Timing and Temperature Conditions**

Hornblende-plagioclase thermometry (Holland and Blundy, 1994) was used to assess conditions of deformation. Analyses of recrystallized horn- blende and plagioclase were conducted with a JEOL 8900 electron microprobe at the University...
of Minnesota–Twin Cities (Minneapolis, Minnesota, USA) using wavelength-dispersive spectrometry. The analysis procedure used appropriate natural mineral standards, 15 keV accelerating voltage, 20 nA beam current, and 2 µm beam diameter. Counting times on element X-ray peaks were 10–20 s, with 5–10 s counting time on upper and lower backgrounds. The volatile elements of Na and Cl were analyzed first and monitored to ensure no appreciable loss during analysis. Analyses used to calculate temperature were from neighboring or closely located hornblende and plagioclase grains (Fig. 14A). Chosen grains possessed a recrystallized polygonal texture, consistent with these grains reaching equilibrium during deformation and metamorphism. The hornblende and plagioclase compositions were similar for each phase for all analyses and were within the composition guidelines prescribed for the thermometer (Holland and Blundy, 1994). Typical analyses from the regional mylonitic fabric are in Table 4 and summarized in Figure 14B. Results produce consistent deformation temperatures of 700–800 °C across the CCMZ at 7–8 kbar (Bohlen et al., 1985).

Bonamici et al. (2014) identified the regional mylonitic fabric as S1. No direct age constraint on this fabric exists other than that it must be younger than ca. 1164 Ma. If this fabric is correlative with the fabric near Indian River Road in the southern CCMZ (Fig. 1B), it is older than an undeformed ca. 1040 Ma Lyon Mountain Granite Gneiss dike that crosscuts it there (Selleck et al., 2005; Johnson et al., 2005). Though its age cannot be tightly constrained, this fabric likely developed during Ottawan prograde metamorphism (Bonamici et al., 2014, 2015) at ~700 °C.

**Mesoscale Ductile Shear Zones**

Throughout the study area, though more common in the pyroxene-bearing rocks of the Diana Complex, ~1–2-cm-wide (mesoscale) ultramyxylonite-cored ductile shear zones are found as individuals or in networks (Figs. 15A, 15B), such as in the H1 outcrop (Wiener, 1983; Lumino, 1986; Heyn, 1990; Cartwright et al., 1993; MacDonald, 1994; 1995; 1996). This study focuses on the Carthage-Colton mylonite zone.
In many of these shear zones, the ultramylonite core is flanked by a transitional zone where the strain gradient results in a rotation of the surrounding fabric toward parallelism with the shear zone; this rotation can be used to indicate relative displacement. For some shear zones, the transitional zone is discolored compared to lower-strain rock, suggesting that fluids altered this rock (Cartwright et al., 1993). In particular, at the H1 outcrop, high-temperature fluids deposited clinopyroxene-rich veins that both are sheared by and crosscut the shear zones (Fig. 16; Bonamici et al., 2014). Ultramylonite cores of shear zones have a grain size of ~10 μm in the northwestern part of the study area to ~100 μm in the southeast of the field area. Due to limited exposures, terminations of mesoscale ductile shear zones are rarely observed, but those observed widen and merge with the regional mylonitic fabric, suggesting that the regional mylonitic foliation and ductile shear zones formed at the same time (Fig. 15C).

The orientations of a combined 145 mesoscale ductile shear zones in the H1 outcrop and in other locations in the study area are shown in...
With the wall-rock fabric curving into the ultramylonite core of the shear zone, it is reasonable to conclude that the shear zones postdate the regional mylonitic fabric. Though straight-walled shear zones are common (Fig. 15A) and clearly crosscut the regional mylonitic fabric of the surrounding rock, some shear zone walls have undulations that require shear zone wall deformation by shear associated with the regional mylonitic foliation. The undulation pattern is not found in the core of the shear zone nor on the opposite wall of the shear zone, and is best explained by synchronous dynamic shear forming both the regional mylonitic fabric and the mesoscale ductile shear zone (Fig. 15D). This is in agreement with the observations regarding shear zone terminations.

**Kinematic Indicators**

Due to near-complete recrystallization of shear zone ultramylonite cores, lineation in the core could only be identified in thin sections cut parallel to the fabric in the core. Lineation was identified by alignment of hornblende porphyroclast tails, opaque- or mafic-mineral train alignment, and zircon alignment (Fig. 17A). It is assumed that the lineation approximates the slip vector within the ductile shear zone. Shear zone kinematic interpretation was based on thin sections cut parallel to the lineation in the zone core and perpendicular to the plane of the shear zone. Criteria used to determine kinematics included \( \sigma \)- and \( \beta \)-type asymmetric porphyroclasts and the rotation of wall-rock fabric through the shear zone transition zone (Figs. 17B, 17C). Figure 8D shows the mesoscale ductile shear zones as S\(_1\). Detailed chemical, microstructural, and O, U, and Pb isotopic analysis of sphenes from within the shear zones indicate that these formed at \(-700 ^\circ C\) and rapidly cooled to \(-500 ^\circ C\) ca. 1054–1047 Ma (Bonamici et al., 2011, 2014, 2015).

**Kinematic Model**

Bonamici et al. (2014) described a conjugate set of shear zones within the H1 outcrop, which is also displayed in the data of Figure 8C, where the regional mylonitic foliation bisects the two orientations. This pattern is also observable in the rest of the study area but only when the variation in the regional mylonitic fabric orientation is accounted for. Figure 8E shows the mesoscale ductile shear poles from locations that have a well-defined regional mylonitic foliation and lineation (seven different locations). The shear zones have been rotated on the stereonet such that the regional lineation for that location is displayed as horizontal east-west and the foliation is vertical east-west. The pattern that results demonstrates that the...
mesoscale ductile shear zones in fact largely form a conjugate set, with the regional mylonitic foliation ~30° from each shear zone set. Therefore, the girdle pattern observed in Figure 8B is misleading regarding the true nature of the shear zone geometry. The conjugate nature of the shear zones is clear at the H1 outcrop because the regional mylonitic fabric has a consistent orientation across the outcrop. However, the geometry between these structures is slightly different at the H1 outcrop than in the rest of the study area because the regional foliation bisects the obtuse angle of the conjugate set at an angle of ~55°–60°. Minor differences between the orientations of the shear zones at the H1 outcrop and across the field area are likely in part controlled by the shear zones developing preferentially along the generally steeply NW-SE–oriented clinopyroxene-bearing veins common here (Fig. 8C).

Conjugate shear zone networks are common and can be formed by either pure- or simple-shear-dominated bulk deformation (e.g., Choukroune and Gapais, 1983; Castro, 1986; Ring, 1999; Fossen and Cavalcante, 2017). Figure 18B summarizes the observed geometric and kinematic relationships between the coeval regional mylonitic fabric and mesoscale ductile shear zones. The observed pattern is very similar to a synthetic one used by Hudleston (1999) to model the strain across shear zone networks. However, natural examples of this pattern have not been described, but a similar

### TABLE 4. REPRESENTATIVE MINERAL ANALYSES AND TEMPERATURE CALCULATIONS FOR THE REGIONAL MYLONITIC FABRIC

| Lithology†| Pyroxene syenite gneiss | Pyroxene syenite gneiss | Hornblende quartz syenite gneiss | Hornblende quartz syenite gneiss |
|-----------|--------------------------|--------------------------|-------------------------------|-------------------------------|
| Site      | CCMZ-C6                  | CCMZ-C6                  | CCMZ-C6                      | CCMZ-C6                      |
|          | 03-13                    | 03-15                    | 03-35                        | 03-15                        |
| Phase     | Amphibole                | Plagioclase              | Amphibole                    | Plagioclase                  |
|          | N/A                      | N/A                      | N/A                          | N/A                          |
| SiO₂      | 40.01                    | 62.96                    | 42.11                        | 64.75                        |
| TiO₂      | 1.32                     | N/A                      | 0.85                         | N/A                          |
| Al₂O₃     | 11.76                    | 23.14                    | 9.68                         | 23.79                        |
| FeO       | 22.37                    | 0.43                     | 18.40                        | 0.20                         |
| MnO       | 0.30                     | N/A                      | 0.41                         | N/A                          |
| MgO       | 7.25                     | <0.00                    | 10.58                        | <0.00                        |
| CaO       | 11.22                    | 4.43                     | 11.05                        | 3.33                         |
| Na₂O      | 1.48                     | 9.21                     | 1.53                         | 7.86                         |
| K₂O       | 1.98                     | 0.08                     | 1.58                         | 0.09                         |
| Cl        | 0.83                     | N/A                      | 0.45                         | N/A                          |
| Total     | 98.52                    | 100.24                   | 96.64                        | 100.02                       |
| Cation**|                           |                           |                              |                              |
| Si        | 6.154                    | 2.784                    | 6.409                        | 2.831                        |
| Al        | 2.131                    | 1.206                    | 1.737                        | 1.226                        |
| Ti        | 0.153                    | N/A                      | 0.098                        | N/A                          |
| Fe²⁺      | 2.000                    | 0.016                    | 1.187                        | 0.007                        |
| Fe³⁺      | 0.877                    | N/A                      | 1.155                        | N/A                          |
| Mn        | 0.039                    | N/A                      | 0.053                        | N/A                          |
| Mg        | 1.663                    | 0.000                    | 2.401                        | 0.000                        |
| Ca        | 1.848                    | 0.210                    | 1.801                        | 0.195                        |
| Na        | 0.440                    | 0.790                    | 0.451                        | 0.666                        |
| K         | 0.389                    | 0.004                    | 0.306                        | 0.005                        |
| Sum       | 15.695                   | 5.010                    | 15.598                       | 4.891                        |
| Pressure  | Rxn A (°C)               | Rxn B (°C)               | Rxn A (°C)                   | Rxn B (°C)                   |
| 0         | 829                      | 709                      | 824                          | 725                          |
| 5         | 788                      | 727                      | 766                          | 735                          |
| 10        | 746                      | 746                      | 709                          | 745                          |
| 15        | 705                      | 765                      | 652                          | 754                          |
| Notes: See Figure 2 for sample locations. N/A—not analyzed; <0.00—below detection limit. Reactions for temperature calculations: Rxn A—edenite + 4 quartz = tremolite + albite; Rxn B—edenite + albite = richterite + anorthite.  
†As mapped by Hargraves (1969).  
*Amphibole recalculations are done by the technique outlined in Holland and Blundy (1994).  
†Plagioclase is recalculated based on eight oxygens per formula unit.
Figure 15. Images of mesoscale ductile shear zones. Location references are shown in Figure 2. (A) Slabbed rock containing a planar mesoscale ductile shear zone (after Baird and Hudleston, 2007). The transition zone of this shear zone is slightly discolored compared to less-deformed rock. Scale bar is 1 cm (site 03-14). (B) Network of intersecting mesoscale ductile shear zones, outlined in blue for clarity (after Baird, 2006). Strike-dip-lineation symbol provides regional mylonitic fabric attitude in this view; north is toward upper right. Shear zone kinematics shown are based on the rotation of fabric in the transition zone and are only relative. A core sample of a shear zone from this outcrop confirms that shear zone transport is oblique to outcrop surface; see text for details. Lens cap for scale is 52 mm diameter (site 03-18). (C) Mesoscale ductile shear zone merging with regional mylonitic fabric; boundaries of the shear zone are outlined (after Baird, 2006). Undulation in the outcrop surface causes some of the apparent variation in feature orientation. Pen is 14 cm in length (site 03-36). (D) Dotted line highlights the undulating margin of ultramylonite found within a mesoscale ductile shear zone (above line; after Baird, 2006). Outcrop surface is subhorizontal; weathering produces the mottled appearance in the upper right of the photo. Shear zone ultramylonite fabric orientation is indicated by the strike-dip-lineation symbol. Mechanical pencil width is 8 mm (site 03-11).

Figure 16. Features of the H1 outcrop (see Fig. 2 for location). (A) One of many hydrothermal veins comprising quartz, feldspar, and clinopyroxene with a myriad of accessory minerals, most notably sphene (see e.g., Bonamici et al., 2014). Field book for scale is 19 cm long. (B) Mesoscale ductile shear zone (MSDSZ); such features commonly localize along hydrothermal veins, but here it is shown cross-cutting a hydrothermal vein (HTV). Note the shoe at the bottom for scale. (C) Ultramylonite mesoscale ductile shear zone cut by unsheared hydrothermal clinopyroxene (CPX).
pattern is found in shear zone systems formed by pure shear (Choukroune and Gapais, 1983; Castro, 1986). Common to the studied and described systems is a symmetrical conjugate set of shear zones that intersect parallel to the regional lineation. The system studied here differs from the natural pure shear–produced systems because the pure shear systems have opposing shear zone kinematics and lineations perpendicular to the intersection of the shear zones. To explain this discrepancy, a two-step deformation model is used to explain the studied system.

The first step in the model is the initial formation of the regional mylonitic fabric and conjugate shear zones from bulk pure shear (Fig. 18A). This initial step established the geometric relationship observed between the regional mylonitic fabric and conjugate shear zones. During this initial stage, strain accumulation must not have been great because beyond establishing the geometric relationships, no other evidence of this deformation can be confirmed.

Then, in the second step, the deformation transitions to bulk simple shear (Fig. 18B), which more strongly develops the regional mylonitic fabric and kinematic indicators and overprints the kinematics of the mesoscale ductile shear zones to what is observed. The shear zones are likely to remain active after the kinematic transition because they are parallel to the bulk slip direction—a favorable orientation for continued shear. With continued simple shear, more shear zones likely develop with locally complex strain patterns where they interlink (Hudleston, 1999; Bhattacharyya and Hudleston, 2001; Baird and Hudleston, 2007). The complex nature of solitary and interlinking shear zones in bulk shear can account for the variation in shear zone orientation, lineation orientation, and kinematics compared to the ideal geometric model presented in Figure 18B.

Possible Later Ductile Fabrics

Bonamici et al. (2014) identified an S₃ fabric that folds the mesoscale ductile shear zones (S₂) at the outcrop scale. The one example found in this work is shown in Figure 19. Stereographic analysis of this later fabric (Fig. 19C) shows that its orientation is consistent with mesoscale ductile shear zone orientations found throughout the area.
Kinematic analysis of the zone based on the sense of deflection of the mesoscale ductile shear zone into the strong fabric suggests that shear sense for the strong fabric is dextral-oblique normal with top to the north. However, a poorly developed $\sigma$-type kinematic indicator from the strong fabric suggests sinistral-oblique reverse kinematics with top to the south (Fig. 19D). Given the near parallelism of structures, a small error in an orientation measurement would provide erroneous kinematic determinations stereographically. Dating of sphene connected to such fabrics (Bonamici et al., 2015) provides an age of ca. 1047 Ma, essentially identical to that of the mesoscale ductile shear zones, and further supports the likelihood that a dynamic pure shear transition to continuous top-to-the-SSE simple shear event can explain all deformational structures discussed thus far. Alternatively, if the normal-sensed kinematics on the late ductile fabrics holds true, matching up with the high-strain zones investigated by Lumino (1986), then this potentially suggests a rapid kinematic switch resulting in these scattered late ductile fabrics.

Ductile Fault

Hargraves (1969) identified a “ductile fault,” or more appropriately, a ductile shear zone (Fig. 2), in one outcrop and traced out the extent of this zone.

### TABLE 5. REPRESENTATIVE MINERAL ANALYSES AND TEMPERATURE CALCULATIONS FOR THE MESOSCALE DUCTILE SHEAR ZONES

| Lithology/Phase | Pyroxene syenite gneiss 03-14 | Pyroxene syenite gneiss 03-21 | Hornblende quartz syenite gneiss 03-15 |
|----------------|-------------------------------|-------------------------------|--------------------------------------|
| SiO$_2$       | 35.88                         | 64.63                         | 40.55                                |
| TiO$_2$       | 1.36                          | 1.36                          | 1.20                                 |
| Al$_2$O$_3$   | 11.67                         | 10.44                         | 11.81                                |
| FeO           | 23.65                         | 20.42                         | 26.16                                |
| MnO           | 0.54                          | 0.54                          | 0.63                                 |
| MgO           | <0.00                         | <0.00                         | 4.59                                 |
| CaO           | 11.10                         | 11.39                         | 10.35                                |
| Na$_2$O       | 1.72                          | 1.62                          | 1.54                                 |
| K$_2$O        | 1.85                          | 1.20                          | 1.29                                 |
| Cl             | 0.70                          | 1.41                          | 1.30                                 |
| Total          | 98.31                         | 100.13                        | 98.61                                |
| Cation$^+$    |                               |                               |                                      |
| Si             | 6.138                         | 6.270                         | 6.116                                |
| Al             | 2.133                         | 1.856                         | 2.189                                |
| Ti             | 0.159                         | 0.147                         | 0.161                                |
| Fe$^{2+}$     | 2.338                         | 1.821                         | 2.548                                |
| Fe$^{3+}$     | 0.729                         | 0.755                         | 0.893                                |
| Mn             | 0.033                         | 0.069                         | 0.084                                |
| Mg             | 1.489                         | 2.103                         | 1.077                                |
| Ca             | 1.844                         | 1.840                         | 1.744                                |
| Na             | 0.517                         | 0.474                         | 0.499                                |
| K              | 0.366                         | 0.402                         | 0.379                                |
| Sum            | 15.746                        | 15.737                        | 15.689                               |
| Pressure (kbar)| Rxn A (°C)                   | Rxn B (°C)                   | Rxn A (°C)                           |
| 0              | 864                           | 712                           | 884                                  |
| 5              | 817                           | 730                           | 817                                  |
| 10             | 770                           | 747                           | 749                                  |
| 15             | 723                           | 765                           | 682                                  |

Notes: See Figure 2 for sample locations. N/A—not analyzed; <0.00—below detection limit. Reactions for temperature calculations: Rxn A—edenite + 4 quartz = tremolite + albite; Rxn B—edenite + albite = richterite + anorthite.

$^*$As mapped by Hargraves (1969).

$^\dagger$Amphibole recalculations are done by the technique outlined in Holland and Blundy (1994).

$^\ddagger$Plagioclase is recalculated based on eight oxygens per formula unit.
to the southwest (see also Geraghty et al., 1981; Baird and MacDonald, 2004). The shear plane of the ductile fault is oriented ~070, 69 S, and based on the wrapping of the regional mylonitic fabric, has top-to-the-northwest kinematics (Figs. 20A and 20B). Timing of the ductile fault is unclear, but the structure does deform the regional mylonitic foliation and likely formed prior to deformation transitioning to a brittle style (see below).

Breccia Zones

Hargraves (1969) also identified a number of ~20–30-m-wide or wider breccia zones that strike generally NE-SW and commonly mark topographic lineaments, the most prominent of which is named the Diana lineament (Fig. 2). Rocks along such zones are cataclasites and show evidence for veining which has also been brecciated (Figs. 20C, 20D). In locations where brittle deformation is not penetrative, the earlier ductile fabrics can be observed. These fabrics are finely laminated (cf. Hargraves, 1969) and have geometry and kinematics consistent with those of the regional mylonitic fabric and are thus considered the same fabric (Fig. 20E). The breccia zones are the latest structures recognized, but exact timing of their formation, or whether non-Grenvillian reactivation has occurred, is unclear. The orientation of such zones is also unclear, but given that the Diana lineament is closely related to the highest-strain fabric in the core of the CCMZ and near the transition between the pyroxene- to hornblende-bearing lithologies, at least this breccia zone is likely subparallel to the regional mylonitic foliation. Localized brittle shearing can be found throughout the length of the CCMZ, and in the Dana Hill metagabbro (Fig. 2), brecciation developed under greenschist facies conditions at ca. 1000–930 Ma (Johnson et al., 2004).

Chronology of Deformation and Related Events

Figure 21 shows a schematic vertical cross-section through the CCMZ with the major features...
depicted. In the text below, numbers in parentheses are keyed to features in the figure. Earliest deformational structures include the penetrative regional mylonitic foliation, which typically manifests as an S-C fabric within the more mafic pyroxene-bearing rocks along the northwest of the zone, to higher-strain ribbon gneisses with various kinematic indicators in hornblende granitic rocks on the southwestern side of the zone (0 and 1). Generally cross-cutting the regional fabric, with local evidence for coeval shearing, are the mesoscale ductile shear zones (2), which are best developed in the pyroxene-bearing rocks. Fluid infiltration, which precipitated localized pyroxene-bearing hydrothermal veins (2), was synchronous with mesoscale ductile shear zone development. Localized later shear fabrics (3) may in some places deform the mesoscale ductile shear zones, so they may be part of this structural suite. The work presented here, augmented and supported by geochronology, demonstrates that the regional mylonitic fabric, mesoscale ductile shear zones, pyroxene-bearing hydrothermal veins, and possibly later shearing of the mesoscale ductile shear zones, were all driven by top-to-the-SSE oblique-sinistral reverse shearing at 1054–1047 Ma (Bonamici et al., 2015) during rapid cooling from ~700 °C to ~500 °C, likely over just a few million years (cf. Figs. 14B and 18). This was followed by development of the ductile fault (4) under a different deformation paradigm from above, because this structure could not have been formed by the same shear kinematics as for the regional mylonitic foliation and mesoscale ductile shear zones. The latest structures are the breccia zones (5), with presumably all examples formed at the same time. Timing and kinematics on the breccia zones are unknown here, hence the query symbol used in Figure 21.

**FIGURE 20.** Photos and analysis of other structures in the study area. Location references are shown in Figure 2. (A) Looking approximately east (mirror image from reality to make the relative view and kinematics the same as all other figures shown in this report) at one of the shear zones that compose the ductile fault in a road outcrop; solid curved red line traces regional mylonitic foliation that can be seen as alkali feldspar–rich layers above the red line; dashed red line traces ductile fault shear plane. Brunton compass for scale (arrow) (site 03-34). (B) Stereonet construction of slip direction of the ductile fault based on shear plane and folded lineation orientation (Ramsay, 1967). Slip direction is calculated to be 56°–101°. Arrows indicate rotation direction from typical regional mylonitic fabric orientations. Data are from this work and Hargraves (1969). (C) Breccia along the Diana lineament (between sites WP6 and WP18). (D) Breciated quartz vein along the Diana lineament (520 m north of site 03-34). (E) Billet of high-ductile-strain fabrics found along the Diana lineament; view to the northeast; one top-to-the-SSE δ-type kinematic indicator is highlighted (site WP4).

**DISCUSSION**

Bulk-rock chemistry, U-Pb zircon geochronology, and field relationships demonstrate that the Diana Complex traverses beyond both margins of
the CCMZ, is expansive, and has affinities to both the Adirondack Lowlands and Highlands. Therefore, the inferred kilometer to tens of kilometers of CCMZ shearing after ca. 1164 Ma complex intrusion did not result in a lithological discontinuity. The complex records a penetrative top-to-the-SSE, rapidly cooling, high-temperature shearing event at ca. 1054–1047 Ma (Bonamici et al., 2011, 2014, 2015). No evidence exists for an earlier, late-Shawinigan-aged deformation of the Diana Complex as previously postulated.

Though the microstructures studied here may be interpreted to have formed under different temperature conditions, with higher temperatures to the southeast, the temperature estimate for deformation across the zone presented here is consistent with peak temperature estimates of ~700 °C by Bonamici et al. (2011, 2014, 2015) along the northwestern edge of the zone. Uniform temperature estimates of the deformation all come from recrystallized grains within the regional mylonitic foliation and mesoscale ductile shear zones, suggesting that deformation temperature is at least broadly similar across the CCMZ and cannot explain the differing microstructures of the Lowlands and Highlands microstructure types.

Many laboratory experiments have investigated deformation mechanisms and the resulting microstructures of deformed pure quartz rocks (e.g., Stipp et al., 2006, and references therein). In general, mylonitic rocks with fine-grained neoblasts, such as those found in the Lowlands microstructures, could be produced not only by lower-temperature ductile deformation, but also by some combination of fast strain rates, high differential stress, and dry deformation conditions. Alternatively, the Highlands microstructures could have been produced by some combination of higher temperatures, metamorphism outlasting deformation, low strain rates, low differential stress, and wet deformation conditions. Interpretation of microstructures is further complicated when considering polymetamorphic rocks or varying mineral modes, including the potential for varying mineral compositions (e.g., Herwegh et al., 2011). However, the close connection between lithology, here summarized as either pyroxene-bearing or hornblende-bearing (Fig. 2), and microstructures is clear and suggests a fundamental difference between these rocks.

Clearly the modal proportions of pyroxene and hornblende are not high enough to have much of an influence on the whole-rock rheology, so the contrasting microstructure is likely controlled by rock differences other than accessory mineralogy. Charnockitic and related crustal-derived rocks form from dry melts (Emslie et al., 1994), thus leading to accessory magmatic pyroxene (Hamilton et al., 2004) and predominately a single (now exsolved) feldspar. However, the hornblende-bearing rocks require additional explanation. One explanation is that the melt that produced the hornblende-bearing rocks was hydrous, resulting in primary hornblende (Buddington, 1939). Alternatively, the hornblende could have been a result of selective hydration of a pyroxene-bearing rock prior to or during deformation, such that during deformation all pyroxene in the hydrated rocks was lost in favor of hornblende. In either case, the hornblende-bearing rocks would have been initially more hydrous, or became so, and the water content could explain the microstructural difference. However, the hornblende-bearing rocks are higher in quartz and poorer in plagioclase compared to the pyroxene-bearing rocks (Hargraves, 1969; Fig. 3A). This mineralogical difference, either in tandem with or independently of the water content, could have influenced rheology, microstructures, and structures formed in each rock type as well.

The tectonic significance of the late ductile fault and breccia zones clearly indicates significant changes in the deformation style following the penetrative ductile event and must be accounted for in any model. Similarly, any model for the CCMZ must successfully incorporate differing kinematics along the CCMZ from those presented here (e.g., Johnson et al., 2004), the cooling histories of the Highlands and Lowlands (Streepey et al., 2001; Dahl et al., 2004), and the steepness and complex sinusoidal (folded?) map pattern of the northern half of the zone (Fig. 1B). Below, two tectonic models are considered to best explain the collective knowledge regarding the Diana Complex and its deformation history associated with the CCMZ and surrounding terranes.
Model 1

Model 1 considers the CCMZ as an Ottawan-aged top-to-the-SSE oblique thrust that was later reactivated with one or more normal-sensed shearing events that exhumed the Adirondack Highlands (Fig. 22). The initial event of this model is the ca. 1050 Ma creation of the CCMZ by pure shear transition to oblique thrusting (Fig. 18) at peak metamorphic conditions (~700 °C; Bonamici et al., 2011; this work) during collision of Amazonia into Laurentia (Fig. 22A). This thrusting included kilometers or more of displacement, possibly along a listric CCMZ that tilted the Lowlands down to the northwest, leading to inclined isotherms (Dahl et al., 2004).

In this model, the Lowlands largely escaped Ottawan tectonism because it was at a shallower crustal level where thermal, magmatic, and penetrative ductile deformation associated with this orogeny were limited. If this were the case, then later normal-sensed motion would have been very significant, suggesting that structures producing this motion would be more prominent, which is not observed.

Alternatively, isotherms may have been inclined across the Adirondacks prior to oblique thrusting along the CCMZ (Dahl et al., 2004). This thermal structure may have been produced by a growing Ottawan orogen toward the northwest from the locus of Amazonia collision to the southeast of the study area. Oblique CCMZ thrusting occurred when the Diana Complex rocks reached ~700 °C temperature, with higher temperatures to the southeast in the Highlands and lower temperatures to the northwest in the Lowlands. As the Lowlands moved to higher structural levels, this may have largely protected it from the thermal, penetrative ductile deformation and magmatic effects of this orogeny.

Regardless of the isotherm details, continued thrusting led to localized back-thrusting as recorded by the ductile fault in the study area and steepening and/or folding of the northern half of the CCMZ (Fig. 22B). The Lowlands then began to cool as isotherms relaxed. Thrusting of the Adirondack Highlands over the Adirondack Highlands is consistent with rapid cooling of the Lowlands side of the zone (Bonamici et al., 2011; 2014; 2015) against the cooler, shallower crustal portions of the Highlands. This thrusting was closely followed by Lyon Mountain Granite Gneiss intrusion and exhumation of the Adirondack Highlands along the CCMZ at ca. 1045–1037 Ma (Selleck et al., 2005; Fig. 22C) and the East Adirondack shear zone at ca. 1050–1025 Ma (Wong et al., 2011), thus forming a symmetrical metamorphic core complex. Within the study area, the normal-sensed late shear fabrics and/or the breccia zones may record this event. Such zones may correlate with the greenschist facies brecciation found in the Dana Hill metagabbro that may have formed ca. 1000–930 Ma during cooling through ~300 °C and final juxtaposition of the two terranes (Streepey et al., 2001; Dahl et al., 2004).

This model is largely consistent with all data and tectonic models that require crustal thickening preceding tectonic collapse (see summary in Rivers [2011]), though it does not easily explain the ductile dextral strike-slip motion recorded in the Dana Hill metagabbro at similar peak conditions (710–770 °C) and timing (1050–1030 Ma; Johnson et al., 2004, 2005) to the ductile deformation studied here. A possible explanation is that this dextral motion may have been of only local importance for the Dana Hill metagabbro deformation, but not to the CCMZ as a whole. Further, normal-sensed ductile deformation during this interval is evident in the Dana Hill body, but may be completely lacking in the Diana Complex, or at minimum, is very poorly developed. This is problematic if the CCMZ facilitated Adirondack Highlands exhumation, as this should be a penetrative deformation event. Possible solutions to this conundrum include that such extensional structures (1) were recorded outside of the study area to the northwest or southeast, (2) are poorly exposed within the study area, or (3) have been eradicated by brecciation.

Model 2

Model 2 considers the CCMZ as originating as a subhorizontal mid-crustal mylonite zone associated with the ca. 1060–1050 Ma tectonic collapse of the now-built Ottawan orogen formed during Amazonia-Laurentia collision (Fig. 23A). The Lowlands would be structurally higher than the Highlands in a region that received little thermal, magmatic, and penetrative ductile effect due to the Ottawan orogeny, while the Highlands received extensive Ottawan
tectonism. The CCMZ would develop between the Lowlands and the Highlands as the subhorizontal sole of a complex series of extensional shear zones of various geometries at the initial stages of tectonic collapse. Such a subhorizontal mid-crustal mylonite zone has been described by, e.g., Cooper et al. (2010) and Platt et al. (2015) and, depending on the geometry of the extensional shear zones, would have variable kinematics and displacement along the structure. The discrepancy between the essentially normal-sensed overprinting, Lyon Mountain Granite intrusion.

ca. 1050-1040 Ma - CCMZ domed over Highlands as core complex develops, formation of ductile fault, perhaps folding of CCMZ, and local normal-sensed overprinting, Lyon Mountain Granite intrusion.

ca. 945 Ma - Low-grade brecciation along the CCMZ and final juxtaposition of the Lowlands and Highlands.

Figure 23. Tectonic model 2 explaining the history of the Carthage-Colton mylonite zone (CCMZ), based on figures in Cooper et al. (2010) and Platt et al. (2014). Figure is highly schematic and is of an approximately northwest-southeast cross-section through the Adirondack Lowlands and Highlands. See text for added detail. LL—Adirondack Lowlands; HL—Adirondack Highlands; EASZ—East Adirondack shear zone.

This model is largely consistent with all data sets, as is model 1. However, model 2 additionally can
account for the differing kinematics found between the Diana Complex structures studied here and the dextral transcurrent structures of the Diana Hill metagabbro due to local variation in extensional structures before doming. Further, it suggests that the shear zones found throughout the Adirondack Lowlands (Fig. 1B) may have had an extensional or transcurrent history during Ottawaan collapse and may sole into the CCMZ at depth below the Lowlands. This model also suggests that a lack of normal-sensed structures found in the study area may be expected because such structures are not required to form during Highlands exhumation. The overall top-to-the-SSE kinematics is consistent with the warped ca. 1070–1060 Ma Marcy Massif detachment zone described by Regan et al. (2019) and suggests that this deformation style may be pervasive throughout the Highlands.

However, a potential drawback of model 2 is that if the Highlands were at least partially migmatitic at the time of exhumation (Bickford et al., 2008), it is expected that a dome structure, potentially with sub-domes (e.g., Whitney et al., 2013), should be evident in the Highlands, as is seen in the Morin and Mekinac-Taureau domes to the north (Dufrêchéou, 2017). To date, no domes or sub-domes that could be connected to exhumation have been described in the Highlands. Current mapping depicts the Highlands possessing an arcing northeast-southwest to east-west to northwest-southeast structural grain from west to east (Fig 1B; McLelland and Isachsen, 1980). With more detailed mapping in the Highlands, structures consistent with the asymmetric metamorphic core complex model (model 2) may be identified (see Regan et al., 2019). However, perhaps the extent of partial melting was not significant enough in the Highlands to allow such dome or sub-dome structures to develop.

### CONCLUSIONS

There are a variety of conclusions regarding the origin of the Diana Complex and its deformation history:

1. The deformaional history found in the southern CCMZ is significantly different than for other locations studied in detail along the CCMZ; this emphasizes the need to study multiple segments of crustal-scale shear zones in order to fully understand their tectonic history and importance.

2. The Diana Complex is a member of the AMCG suite, spans the CCMZ, and crystalized at 1164.3 ± 6.2 Ma. Original estimates of complex minimum thickness of 6 km likely underestimate its extent, because this work demonstrates the inclusion of, at minimum, the aplitic granite gneiss to the complex. Therefore, the inferred kilometer to tens of kilometers of offsets that occurred across the CCMZ did not result in a lithological discontinuity across the CCMZ.

3. The Diana Complex within the CCMZ contains a northwest-dipping penetrative regional mylonitic foliation that bisects a conjugate set of mesoscale ductile shear zones. These structures initiated in a pure-shear event that transitioned to oblique-sinistral, top-to-the-SSE shearing that mostly developed the structures.

4. Strain associated with the regional mylonitic foliation is greatest in the central CCMZ; this zone of high strain is associated with a change in microstructure. This change is a marker that can be used as the boundary between the Adirondack Lowlands and Highlands.

5. Ductile deformation of the Diana Complex occurred at ca. 1054–1047 Ma (Bonamici et al., 2015) at ~700 °C. Later structures include a ductile fault and breccia zones. Conditions and kinematics of brecciation are less clear, but they may have formed during greenschist facies extension at ca. 945 Ma (Streepey et al., 2000).

6. Two models can explain the development of the studied structures and other data sets from the Adirondacks:

   a. Model 1 depicts the CCMZ as forming from top-to-the-SSE thrusting during Amazonia collision. The CCMZ was reactivated during minimum ductonic collapse as the Highlands exhumed, resulting in a symmetrical metamorphic core complex.

   b. Model 2 depicts the CCMZ as initially forming as a subhorizontal mid-crustal mylonite zone during tectonic collapse. The collapse led to exhumation of the Highlands and doming of the mid-crustal mylonite zone over the Highlands to form the CCMZ (in the northwest) and the East Adirondack shear zone (in the southeast), resulting in an asymmetrical metamorphic core complex. During doming, the CCMZ was overprinted with additional structures. The top-to-the-SSE preserved kinematic indicators in the southern CCMZ originated during the subhorizontal shearing before doming. Model 2 is the preferred model because it best integrates all data.

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